Limnol. Oceanogr., 54(6, part 2), 2009, 2371–2385

High Arctic lakes as sentinel ecosystems: Cascading regime shifts in climate, ice cover, and mixing

Derek R. Mueller,[a,1] Patrick Van Hove,[b,2] Dermot Antoniades,[b] Martin O. Jeffries,[c] and Warwick F. Vincent[b,*]

[a] Geography Department, Trent University, Peterborough, Ontario, Canada
[b] Centre for Northern Studies (CEN), Laval University, Quebec City, Quebec, Canada
[c] National Science Foundation, Arlington, Virginia

Abstract

Climate and cryospheric observations have shown that the high Arctic has experienced several decades of rapid environmental change, with warming rates well above the global average. In this study, we address the hypothesis that this climatic warming affects deep, ice-covered lakes in the region by causing abrupt, threshold-dependent shifts rather than slow, continuous responses. Synthetic aperture radar (SAR) data show that lakes (one freshwater and four permanently stratified) on Ellesmere Island at the far northern coastline of Canada have experienced significant reductions in summer ice cover over the last decade. The stratified lakes were characterized by strong biogeochemical gradients, yet temperature and salinity profiles of their upper water columns (5–20 m) indicated recent mixing, consistent with loss of their perennial ice and exposure to wind. Although subject to six decades of warming at a rate of 0.5°C decade^{-1}, these lakes were largely unaffected until a regime shift in air temperature in the 1980s and 1990s, when warming crossed a critical threshold forcing the loss of ice cover. This transition from perennial to annual ice cover caused another regime shift whereby previously stable upper water columns were subjected to mixing. Far northern lakes are responding discontinuously to climate-driven change via a cascade of regime shifts and have an indicator value beyond the regional scale.

There is a broad consensus that the world is entering a period of accelerated climate change (Solomon et al. 2007). Anthropogenic emissions of greenhouse gases and the resultant increase in net radiative forcing are widely accepted as the cause of this rapidly changing climate (Hansen et al. 2006). We are now faced with the challenge of evaluating the extent and pace of this change and its effect on the biosphere.

Computer models and direct observations (Serreze and Francis 2006) indicate that the highest amplitude and most rapid climate changes are occurring in the Arctic. Although there is considerable variability among regions, the annual average air temperature in the North American continental Arctic increased by 1.06°C decade^{-1} during the last two decades (Comiso 2003), well above the global average of 0.2°C decade^{-1} during the same period (Hansen et al. 2006). The thinning and decrease in extent of Arctic sea ice cover (Maslanik et al. 2007) will lead to even more pronounced changes through an albedo-driven positive feedback that would cause Arctic Ocean warming, a further decrease in sea ice extent, and more amplification of high latitude warming.

Many reports have shown significant climate change and associated ecosystem effects in the Arctic since the beginning of the short observational period in this region (Serreze et al. 2000). Long-term trends, however, have been difficult to resolve in the Arctic, in part because of the lack of site-specific data and substantial interannual variability in climate, which is exacerbated by the Northern Annular Mode. These factors act to obscure underlying trends, particularly those that are important over short timescales. Furthermore, calculating trends in climate variables may not adequately capture temporal change when data do not respond in a linear fashion. Reporting trends for select subperiods where a linear response is noted is an arbitrary (and potentially biased) solution to this problem. Another method for describing change in climate over time is to evaluate where a significant shift in the mean of a given variable occurs over a complete data series. These shifts indicate the timing and severity of a rapid switch between stable states or regimes (Scheffer et al. 2001; Rodionov 2004).

An analogous approach toward evaluating the effects of climate change is to identify critical thresholds in ecosystems where further change results in abrupt, discontinuous shifts in ecosystem properties. The phase change of ice to water, for example, has a single temperature threshold, and crossing that threshold can alter the structure of ecosystems (e.g., excess meltwater runoff, loss of lake ice cover), thereby amplifying the effects of small temperature changes. As thresholds in the annual energy budget are crossed, either through decreased ice formation during the winter or via increased melt over the summer period, discontinuous ecosystem change will become apparent. Some of these thresholds for Arctic aquatic ecosystems have been identified (ACIA 2005), including changes in snow cover, ice cover thickness and duration, and the frequency (or occurrence) of mixing, as well as dissolved organic carbon loading associated with tree line migration. If these abrupt changes are from one stable state to

* Corresponding author: warwick.vincent@fsg.ulaval.ca

Present addresses:
[1] Canadian Ice Service, Environment Canada, Ottawa, Ontario, Canada
[2] GreenFacts, Brussels, Belgium

another, they can be considered environmental or ecological regime shifts (Smol et al. 2005; Grebmeier et al. 2006) and may be more compelling climate sentinels than long-term records of continual change that are obscured by natural scales of variability.

The northern coast of Ellesmere Island lies at the northern limit of the Canadian high Arctic within a region projected to experience the greatest annual warming in Arctic North America over the next eight decades (ACIA 2005). This region contains many types of rare aquatic ecosystems, including meltwater lakes on the surface of ice shelves (Mueller et al. 2006), ice-dammed fiords (Veillette et al. 2008), and perennially ice-covered meromictic (permanently stratified) lakes (Van Hove et al. 2006). These ecosystems are attracting increasing attention as climate change indicators because of their strong dependence on the presence of ice.

Abrupt shifts due to climate change have been discerned throughout the Arctic region, including effects on lakes and ponds from the tree line (Rühland et al. 2003) to the polar desert (Smol et al. 2005). Sedimentation rates in the Arctic are low, and published high-resolution paleolimnological records are rare. Owing to the small size and relative sensitivity to change of typical freshwater Arctic sites, many expressed their most significant biological shifts more than a century ago (Smol et al. 2005). In addition, long-term (>60 yr) climate data for some regions are nonexistent, making it difficult to assess these effects in the context of global change.

In this paper, we present a synthesis of our recent observations on the lakes and climate of northern Ellesmere Island. We first present in situ water-column profiling data spanning four decades, which characterize the long-term stratification regime and record the extent of recent change in these environments. We then use a time series of synthetic aperture radar (SAR) images to examine summer lake ice loss over the last two decades. Finally, we evaluate the hypothesis that significant climate warming trends and regime shifts have occurred in this region, inducing cryospheric (perennial ice) and limnological (water-column) regime shifts in response to this warming. We discuss the nature of the cascading regime shifts within these systems and assess the value of lake ice cover and water-column structure as global sentinels of climate warming.

Materials

Description of study sites—Lakes A (83°00′N, 75°05′W), B (82°58′N, 75°26′W), and C1 (82°51′N, 78°08′W) are meromictic lakes that were first described by Hattersley-Smith et al. (1970; Fig. 1). Two nearby lakes (Lake C2 and C3, located south of Lake C1) were first profiled in 1985 (M. O. Jeffries unpubl. data). The C lakes form a continuum from highly stratified Lake C1, with a small (3.3 km^2) unglacierized catchment; to Lake C2, with a 23.5 km^2 glacierized catchment area; to freshwater Lake C3, which can receive variable amount of flow from the Taconite River (glacierized catchment, 11 to 260 km^2; Bradley et al. 1996; Van Hove et al. 2006). Lakes A and B have unglacierized catchment areas of 37 km^2 (Van Hove et al. 2006) and approximately 5 km^2, respectively. A perennial ice cover prevents wind-induced mixing in all these lakes, while steep salinity gradients (with the exception of Lake C3) prevent convection in their water columns (Ludlam 1996; Vincent et al. 2008). Consequently, the lakes have developed deep thermal maxima over many decades of solar heating (Vincent et al. 2008). In addition to evidence of long-term lake ice covers from the existence of thermal maxima, these lakes were observed to have a refrozen candled ice cover prior to the onset of melt, which suggests that the ice cover was perennial (Hattersley-Smith et al. 1970; Belzile et al. 2001). Between 1969 and 1998, these lakes were never reported to have lost their ice cover during the summer beyond melting around the lake shore that created a "moat" of variable but limited extent (Bradley et al. 1996; Ludlam 1996).

From early-season profiling, the maximum ice thickness was approximately 2 m for Lake A (Hattersley-Smith et al. 1970; Jeffries et al. 1984; Belzile et al. 2001), 2 m for Lake B (Hattersley-Smith et al. 1970; Jeffries and Krouse 1985), and from 1.1 to 2 m for the C lakes (M. O. Jeffries unpubl. data and this study). A 1999 survey on Lake A found an average snow depth of 52 cm with a depth-integrated density of 0.21 g cm^{-3} (Belzile et al. 2001). Snow depths recorded elsewhere on these lakes during early-season profiling were between 42 and 50 cm (M. O. Jeffries unpubl. data and this study). Lake B is the most sheltered of the lakes in this study and is flanked on three sides by mountains. Lake A is the largest and deepest of these lakes and occupies a broader valley with several inflowing streams, whereas lakes C1, C2, and C3 are more exposed to wind and sun in general but are surrounded by steep shores.

Owing to the relative water-column stability of these lakes, they have been the focus of ongoing geochemical and ecological study. Steep biogeochemical gradients, primarily associated with the halocline, have created stratified microbial habitats for a variety of taxa (Van Hove et al. 2008; Pouliot et al. 2009; Antoniades et al. in press). The ecology of these lakes may be altered abruptly if their biogeochemical gradients are disrupted.

Water-column profiling—The water-column temperature and conductivity profiles prior to 2003 were obtained from previous studies (Table 1), and complete methods are described in these papers. Briefly, Hattersley-Smith et al. (1970) used Knudsen bottles, reversing thermometers, and an automatic temperature–depth–salinity recorder; Jeffries et al. (1984), Jeffries and Krouse (1985), and M. O. Jeffries unpubl. data used reversing thermometers and measured aliquots from Knudsen bottles with an Endeco refracting salinometer; Ludlam (1996) used a 2-Hz Seacat SBE 19-03 profiler; and Belzile et al. (2001) and Van Hove et al. (2006) both used a Hydrolab Surveyor 3 profiler. Unpublished profiles provided by M. Retelle were obtained using a temperature and salinity profiler (YSI) down to a depth of 60 m and from a Kemmerer bottle water for salinity at greater depths. Profiles obtained after 2001 were measured at 1 Hz by lowering an XR-420 conductivity–temperature–

Table 1. Temperature and depth of thermal maxima in the lakes of northern Ellesmere Island at each profile date.

Lake	Date	Temperature (°C)	Depth (m)	Reference
Lake A	13 Jul 2007	8.7	18.4	present study
	30 May 2006	8.6	17.5	present study
	26 May 2005	8.4	18.1	present study
	04 Aug 2004	8.6	17.6	present study
	01 Aug 2003	8.7	18.4	present study
	01 Aug 2001	8.8	17.0	Van Hove et al. 2006
	05 Jun 1999	8.5	17.3	Belzile et al. 2001
	26 May 1993	8.9	16.3	Ludlam 1996
	05 Jun 1985	8.6	15.0	M. Retelle unpubl. data
	14 May 1983	7.9	20.0	Jeffries and Krouse 1985
	10 May 1982	8.3	15.0	Jeffries et al. 1984
	01 May 1969	7.6	16.8	Hattersley-Smith et al. 1970
Lake B	03 Aug 2005	10.6	16.5	present study
	04 Aug 2004	10.3	16.8	present study
	26 May 1993	10.2	16.5	Ludlam 1996
	06 Jun 1985	9.3	15.0	M. Retelle unpubl. data
	15 May 1983	9.8	15.0	Jeffries and Krouse 1985
	01 May 1969	8.7	17.3	Hattersley-Smith et al. 1970
Lake C1	04 Apr 2008	13.1	17.4	present study
	03 Aug 2005	12.6	16.3	present study
	01 Jul 2001	12.2	15.6	Van Hove et al. 2006
	29 May 1992	11.4	16.2	Ludlam 1996
	21 May 1985	10.7	15.0	M. O. Jeffries unpubl. data
	01 May 1969	10.1	15.8	Hattersley-Smith et al. 1970
Lake C2	04 Apr 2008	3.4	23.1	present study
	03 Aug 2005	5.0	1.6	present study
	01 Jul 2001	3.4	24.0	Van Hove et al. 2006
	26 May 1992	3.8	19.2	Ludlam 1996
	21 May 1985	3.7	10.0	M. O. Jeffries unpubl. data
Lake C3	04 Apr 2008	3.3	48.2	present study
	03 Aug 2005	4.9	3.2	present study
	01 Jul 2001	4.7	1.5	Van Hove et al. 2006
	03 Jun 1992	4.1	32.9	Ludlam 1996
	22 May 1985	4.0	10.0	M. O. Jeffries unpubl. data

depth sensor (RBR) through natural or drilled holes in the lake ice. This instrument had an accuracy of ±0.002°C, ±0.003 mS cm^{-1}, and ±17 mBars. To correct for the lag in response of the thermistor (time constant of 3 s), a time derivative was estimated by a local least squares slope (Crease et al. 1988). Conductivity measurements (e.g., profiles from Ludlam 1996; Van Hove et al. 2006; and the present study) were converted into salinity (g L^{-1} derived from practical salinity units) using the equations in Fofonoff and Millard (1983) since ionic composition in most of these lakes is similar to standard sea water (Van Hove et al. 2006). Although this assumption is not strictly true for Lake C3 (Van Hove et al. 2006), the conversion allowed for a multiyear relative comparison in the same units as earlier studies. If more than one profile was available in a given year, we considered the earliest (if profiles were taken in different months) or longest (if profiles occurred at approximately the same time) profile in our analysis. Clearly erroneous data, caused by calibration error (low salinity values below the halocline in the 1985 Lake A and B profile) or sampling bottle malfunction (anomalously low salinity at 15 m in 1983 Lake A profile), were excluded in subsequent analyses. Lake stability was calculated from some Lake A profiles using the method proposed by Idso (1973) for the water column above the observed mixing depth in the lake, ignoring bathymetry.

Remote sensing—SAR satellite images obtained between 1992 and 2007 were used to determine the amount of open water present at the end of summer. The images were acquired in standard beam (resolutions: *Radarsat-1* = 32 m, *ERS-1* = 26 m, and *JERS-1* = 28 m) and in *Radarsat-1* fine beam (8-m resolution) and ScanSAR wide B (75-m resolution) beam mode and subsequently georeferenced in an Albers equal-area projection.

The backscatter difference between new and perennial ice was used to determine the proportions of ice and open water at the end of summer. Figure 2 illustrates the principle behind the use of SAR for this purpose. When acquired after freeze-up, the images show that new ice, which formed on open water at the end of the melt season, has a dark and textureless tone. Backscatter from this new ice is very low due to specular reflection off the smooth upper and lower surfaces, and there is negligible volume scattering due to the absence of internal reflectors such as bubbles (Jeffries et al. 2005). In contrast, residual ice has a more gray and textured tone because backscatter from the ice is higher due to scattering from a rough candled surface and volume scattering from internal melt features (Jeffries et al. 2005).

A total of 43 SAR images of lakes A and B and 47 images of lakes C1, C2, and C3 were examined to evaluate the state of the ice cover. Those images that provided the clearest distinction between new and old ice were used to determine the end-of-summer ice and open water areas by digitizing polygons in ArcInfo (ESRI).

Climate analysis—Daily climate data from Alert and Eureka, Nunavut, were obtained from Environment Canada from 01 July 1950 to 02 October 2006 and 01 May 1947 to the end of 2007 (www.climate.weatheroffice.ec.gc.ca; last accessed 14 July 2008). Missing daily values were linearly interpolated if they were preceded and followed by valid data, which amounted to ≤0.21% of observations (depending on the variable). Individual years and seasons were rejected from further analysis if they contained missing data after the interpolation procedure.

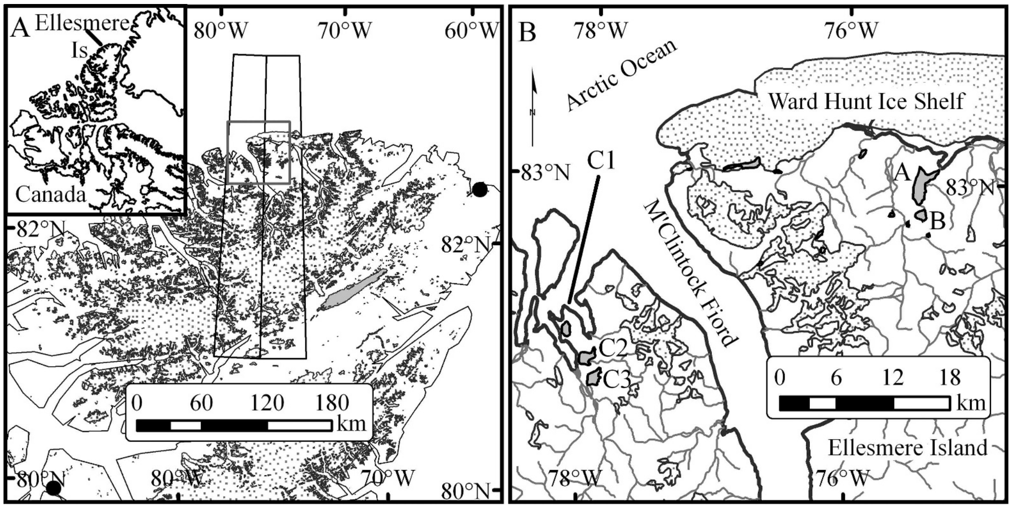

Fig. 1. Maps of the study area. (A) Location (Inset) and the northern coast of Ellesmere Island. Climate reanalysis grid cells used in this study are indicated with vertical trapezoids. Eureka and Alert are indicated by black circles in the lower left and upper right, respectively. The square indicates the region covered by Fig. 1B. (B) Location of the ice-covered lakes on the northern coast of Ellesmere Island.

Since Alert and Eureka are distant (230 km and 380 km) from our study area on the northern coast of Ellesmere Island, we also employed National Centers for Environmental Prediction and National Center for Atmospheric Research daily surface air temperature reanalysis data. This product is made by assimilating climate data from a variety of sources and interpolating to a $2.5^\circ \times 2.5^\circ$ grid (Kalnay et al. 1996). Reanalysis data from 01 January 1948 to 31 December 2007 were obtained for two adjacent grid cells containing lakes A and B, and lakes C1, C2, and C3 (Fig. 1A). Reanalysis data from appropriate grid cells were compared with Alert and Eureka daily air temperatures from Environment Canada and were found to adequately reflect air temperatures at these two stations. The reanalysis data gave slightly lower surface temperatures than observed at the climate stations (a paired t-test indicated significant differences in mean air temperatures of 0.14°C for Alert and 1.6°C for Eureka), perhaps because of the difference between the mean elevation of the grid cell and the stations at sea level. It should be noted that the overall warming trends for reanalysis data were between two and three times higher than the observed trend. This may be due to the coastal location of the stations (the cold ocean–warm land pattern of warming; Wallace et al. 1996), although the exaggeration of warming trends at high latitudes has been noted for several reanalysis products, particularly in the upper atmosphere. Reanalysis data are optimized for synoptic-scale accuracy, and this does little to remove small biases that may interfere with long-term trends (Christy et al. 2006). Given these shortcomings, it may be more appropriate to characterize reanalysis surface air temperatures with regime-shift methods, rather than trend analyses, particularly where these temperatures are poorly constrained by observational data.

Total melting degree days (MDD) and freezing degree days (FDD) were calculated by summing the daily average air temperature for all days of the year where the average air temperature was above (or below) the freezing point. The length of the melt season was calculated by median filtering air temperature data with a 13-d window and counting the number of days where the filtered air temperature remained above 0°C. The annual variance of air temperature data was calculated, and yearly total rainfall, snowfall, and precipitation were determined for climate station data. Trends and their significance in all variables were calculated by regression using the entire available range of annual or seasonal data.

Regime shifts were detected in annual surface air temperature, MDD, FDD, and melt season duration from reanalysis data of the study sites using the program provided at www.beringclimate.noaa.gov/regimes/ (last accessed 25 July 2008). The degree of serial autocorrelation in each variable was estimated using the inverse proportionality with four corrections (IP4) method with a subsample size of 10 (Rodionov 2006). This estimate of the autoregressive parameter was used to filter out red noise from the data, if it was detected (known as prewhitening; Rodionov 2006). The filtered data were then examined by a sequential algorithm that flags departures from the mean that exceed a critical value based on the t-distribution. A decision is made at each time step (year) to accept or reject H_0 (no regime change occurred), or to keep testing. The strength of the detected regime shifts are denoted by the regime-shift index (RSI), and a probability of obtaining this value using random data can be used to determine the significance of a regime shift (Rodionov 2004). The algorithm requires two parameters: the target significance level required to detect a regime shift (set at 0.1) and the cutoff length, which determines the minimum length of regimes to include (set at 10 yr). All regimes that are more significant than the target level are selected, and regimes shorter than the cutoff length may be detected if they are highly significant. A down-weighting of outliers based on their distance from the mean of the current regime

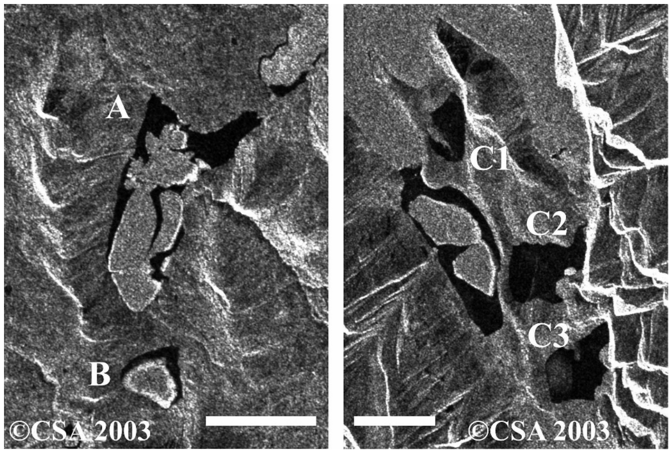

Fig. 2. *Radarsat-1* SAR image subscenes of lakes A and B (left), and lakes C1, C2, and C3 (right) on 11 September 2003. Perennial ice appears textured and gray in contrast to areas of new ice, which appear featureless and dark. The white scale bar represents 2 km.

was considered for data exceeding two standard deviations (Huber parameter set at 2; Rodionov 2006).

Results

Water-column structure—All of the meromictic lakes in this study (lakes A, B, C1, and C2) had monimolimnia that were close to the salinity of seawater (25 to 33 g L^{-1}) overlain by a 7–20-m-thick layer of fresher water (Figs. 3 and 4). Lake C3 had relatively weak vertical structure, with a maximum salinity of 0.3 g L^{-1} at its bottom (Fig. 4D), making it unsuitable for recording mixing events. Above the stable main halocline in the other four lakes, salinity profiles varied interannually from a smooth monotonic increase with depth (indicating long-term stability in the water column) to a stepped profile indicating the presence of a mixed layer(s) within the upper water column (Fig. 5). Slight increases in salinity at the very top of the water column along with the freshening of water just above the halocline provided further evidence of mixing (e.g., 2001 profile, arrow 1, Fig. 5C).

The extreme upper sections of the water columns of all lakes had variable temperature profiles that were characterized by the presence or absence of a temperature peak directly under the ice (Figs. 3 and 4). This peak was likely related to solar heating or the inflow of warm water from the catchment and typically became more prominent in the late summer. We did not consider the upper water column (top ~2 to 5 m) temperature when comparing interannual changes to the lake, as a result of this seasonal artifact. As observed with the salinity data, the shape of the temperature profile in the upper water column indicated whether the water was stable for many years (smooth profile) or recently mixed (stepped profile).

It is possible to determine the timing and maximum depth of mixing events in the water column using salinity

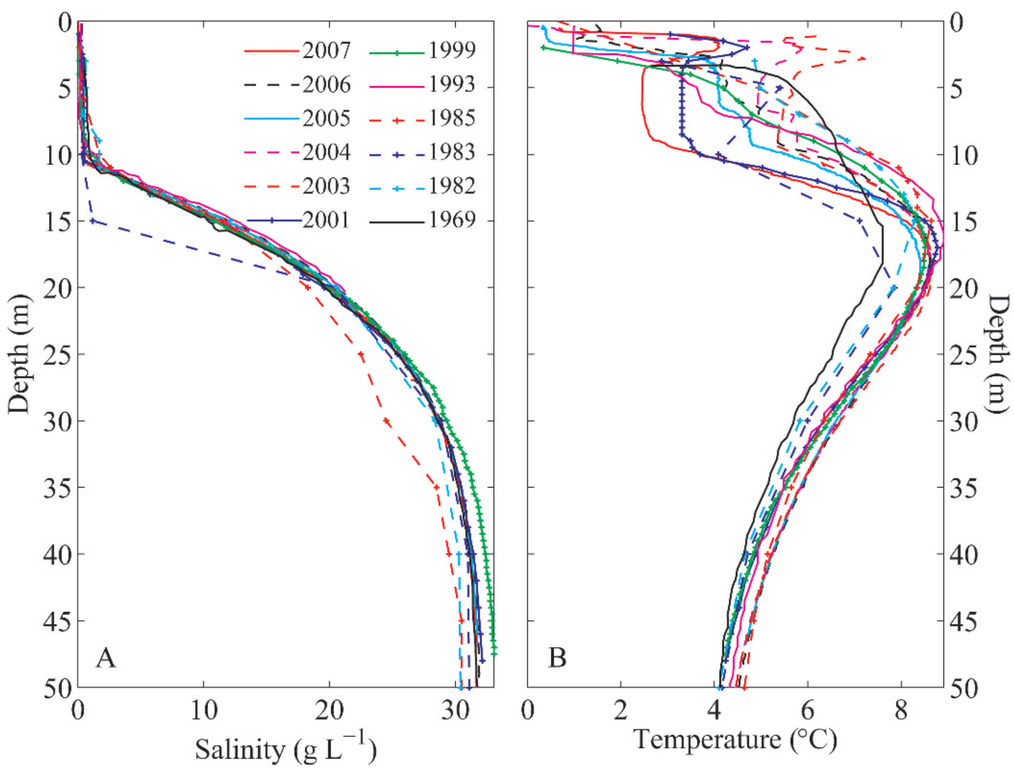

Fig. 3. (A) Water-column salinity and (B) temperature profiles for the upper 50 m of Lake A. The salinity profile is very stable at depth over all the years on record. Departures in the 1983 and 1985 profiles are likely due to instrument calibration error. Temperature profiles are more variable but record changes in the heat storage in the lake (i.e., the depth and temperature at the mid–water-column thermal maximum) and mixing events in the upper water column (e.g., 2001 and 2007).

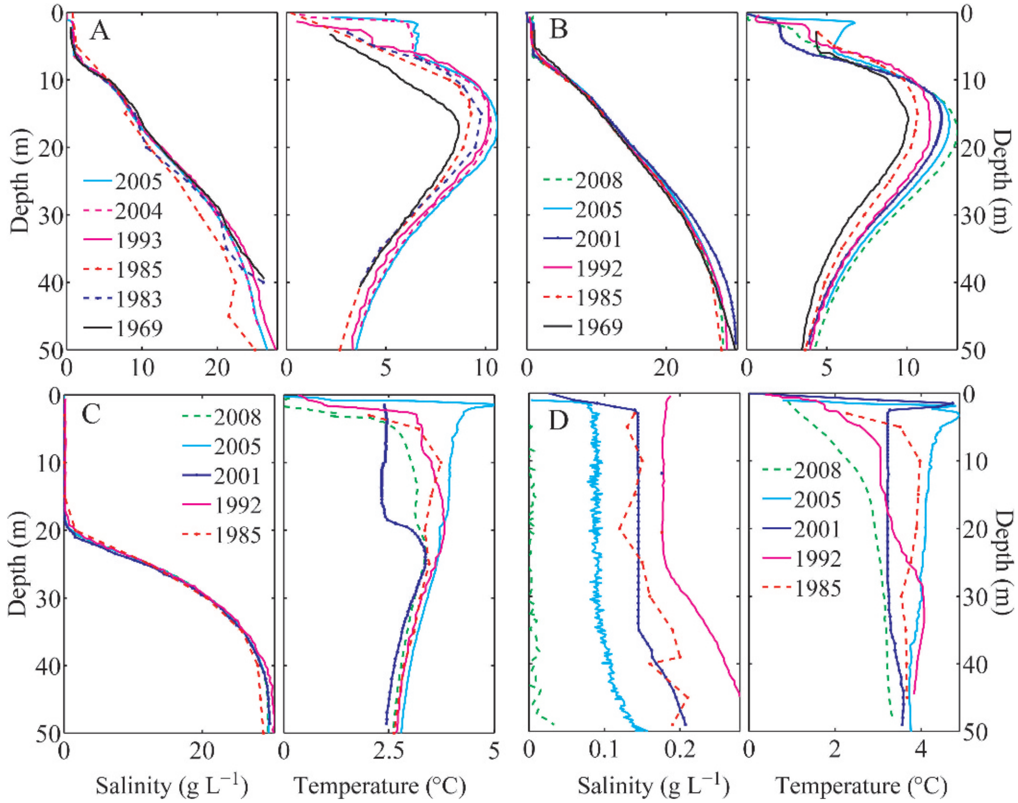

Fig. 4. Water-column salinity (left) and temperature (right) profiles for the upper 50 m of (A) Lake B, (B) Lake C1, (C) Lake C2, and (D) Lake C3.

and temperature profiles. Evidence of mixing remains until obliterated by a deeper mixing event or the profile eventually smoothes out over many years due to diffusion. Therefore, it can be assumed that observed changes occurred between the profile recording mixing and the previous profile. Given that lakes were always profiled prior to ice out, any mixing would have occurred, at the latest, by September of the previous year.

In the 1969 salinity and temperature profiles from Lake A and B, smooth increases from under the ice to the thermal maxima suggest a long period of stability (Fig. 5A,B). The profile of Lake C1, however, indicates that a mixing event may have occurred in the 1960s (arrow 1, Fig. 5F). The 1993 profile in Lake B indicates that a mixing event occurred between 1985 and 1992 (arrow 1, Fig. 5B). Profiles from 2001 show lakes A, C1, and C2 had mixed since previous water-column observations (arrow 1, Fig. 5C; arrow 2, Fig. 5F and Fig. 4C). Lake A's profile in 2004 indicates that mixing occurred in 2003 (arrow 2, Fig. 5C), while more recent mixing is indicated by the 2007 profile in Lake A and the 2008 profile in Lake C1 (arrow 3, Fig. 5E,F).

Lakes A, B, and C1 had deep-water thermal maxima between 7.6°C and 13.1°C, located about 10 m below their haloclines (Table 1; Figs. 3, 4). The maximum water temperature in these lakes increased since 1969, which is consistent with the long-term storage of solar energy (Vincent et al. 2008). Lake C2 also exhibited a deep-water thermal maximum in 2001 and relatively small maxima above the halocline in other years (Fig. 4C; Table 1). In comparison, Lake C3, which is freshwater throughout, had a weak but noticeable pycnocline at 25 m in 1992, which was accompanied by a small rise in water temperature just below this depth (Fig. 4D). Although profiles from Lake C3 in other years showed slight salinity increases, no deep-water thermal maximum was observed.

In summary, the monimolimnia of these lakes were very stable, with fairly constant thermal maxima. One upper water-column mixing event occurred prior to 1969, another occurred between 1985 and 1992, and at least seven mixing events took place after 1992. The data indicate that mixing events are now relatively common, rather than isolated occurrences as in the past, suggesting that a new regime of frequent water-column mixing began in the 1990s.

Ice cover—SAR data indicate that lakes A, B, and C1 retained their ice covers in the majority of years, whereas lakes C2 and C3 lost their ice covers in over half the years on record (Table 2). SAR observations were less frequent during the first 10 yr of the record, with missing data from 1992 to 1994 (depending on the lake; Table 2). It is possible that ice losses occurred during these years; however, a lack of ice loss in 1992 and 1994 from the lake most prone to ice breakup (Lake C3) suggests that other lake ice covers remained intact during those years (Table 2). Apart from the first year of our record (1991), there was no evidence to suggest that the lakes had any ice loss before 1998 (an exceptionally warm year, both in the Arctic and globally).

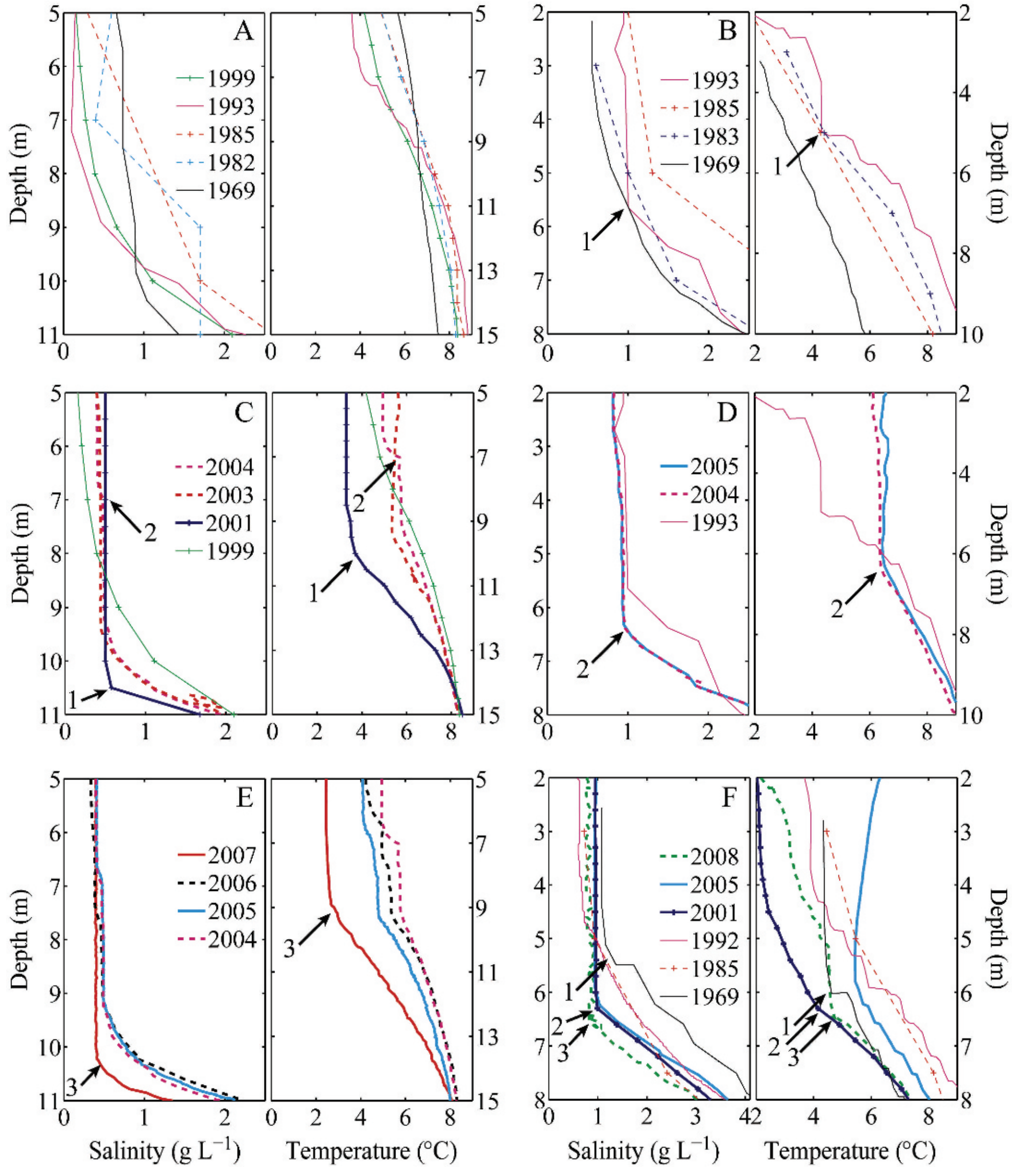

Fig. 5. Evidence of upper water-column mixing from salinity and temperature profile changes over time. (A), (C), and (E) Lake A salinity (left) and temperature (right) changes. (B) and (D) Lake B salinity (left) and temperature (right) changes. (F) Lake C1 salinity (left) and temperature (right) changes. Profiles were divided into three subplots for Lake A and two subplots for Lake B to improve legibility. Note that the depth scales for salinity and temperature are not necessarily the same between left and right panels. Evidence of mixing events is indicated with sequentially numbered arrows. Profiles taken during the regime of frequent water-column mixing (from ca. 2000 onward) are indicated by heavier lines.

Before 1998, three losses of lake ice were observed on all five lakes combined, in contrast to the period 1998 to 2007, when lake ice loss occurred a total of 30 times. This suggests that a new regime of frequent summer ice loss began in 1998 that supplanted a regime of rare ice loss on these lakes.

Climate—Our climate datasets indicate that the northern Ellesmere Island region has warmed over the last six decades at a rate between 0.14°C ($p = 0.530$) and 0.48°C ($p < 0.002$) per decade at Alert and on the northernmost coast, respectively (Table 3). Seasonally, the autumn (September, October, November [SON]) air temperature trend was the highest, followed by winter (December, January, February [DJF]), and then spring (March, April, May [MAM]). The summer (June, July, August [JJA]) air temperature warming rate, although positive, was the least rapid and was found to be nonsignificant in all cases except for Eureka reanalysis data. The variance in daily air temperatures fell for all datasets over the period of observation, while the total precipitation rose at both Alert (not significant) and Eureka ($p = 0.012$). The MDD increased for all datasets except Alert,

Table 2. Proportion of ice cover lost (proportion of open water present) by the end of summer determined from post–freeze-up SAR imagery. Dash indicates no detectable ice loss, and n.a. indicates that no data are available. The SAR image ID indicates the satellite platform (e, *ERS*; j, *JERS*; r, *Radarsat*) orbit number, beam mode (std, standard; swb, ScanSAR), and scene number.

	Lake ice loss (%)						
Summer	A	B	C1	C2	C3	SAR image ID	Acquisition date
1991	13	—	—	76	100	e1_03582_std_214	23 Mar 1992
1992	n.a.	n.a.	n.a.	n.a.	—	e1_09623_std_213	19 May 1993
1993	n.a.	n.a.	n.a.	n.a.	n.a.	no imagery	
1994	n.a.	n.a.	—	—	—	e1_17133_std_213	25 Oct 1994
1995	—	—	—	—	—	j1_19166_std_240	13 Aug 1995*
1996	—	—	—	—	—	r1_06088_swb_241	03 Jan 1997
1997	—	—	—	—	—	r1_11452_swb_211	13 Jan 1998
1998	—	—	100	89	100	r1_14863_swb_241	09 Sep 1998
1999	—	—	—	5	18	r1_20756_st5_210	26 Oct 1999
2000	100	100	100	100	100	r1_25172_swb_211	30 Aug 2000
2001	—	—	—	—	—	r1_30775_st7_208	26 Sep 2001
2002	15	—	7	18	100	r1_35820_st7_208	14 Sep 2002
2003	39	38	86	98	69	r1_40992_st1_215	11 Sep 2003
2004	—	—	—	—	—	r1_47467_swb_211	07 Dec 2004
2005	—	—	—	17	28	r1_53240_st1_214†	15 Jan 2006
2006	52	—	100	57	100	r1_56870_st1_215‡	26 Sep 2006
2007	51	55	18	34	100	r1_61816_swb_209	07 Sep 2007

* NB, mid-August observation.
† Lakes C1, C2, and C3 imaged on 20 April 2006 (r1_54598_st1_214).
‡ Lakes C1, C2, and C3 imaged on 20 July 2007 (r1_58529_fn1_209).

where it decreased (but not significantly; Table 3). The FDD decreased significantly, which is consistent with higher autumn and winter air temperatures. The melt season length increased near the lakes, but not significantly, and the trend in this variable was positive at Alert and negative in Eureka (the reverse was true for reanalysis data; Table 3).

The trend analysis gave an overview of the changes in the climate data over the entire observational record, but long-term linear trend analysis cannot resolve short-term or abrupt warming (e.g., recent summer air temperature increases) that was resolved by the regime-shift methodology. In the Lake A and B region, annual mean air

Table 3. Trends per decade in climate variables for all years on record (reanalysis data, 1948–2007; Environment Canada, Alert, 1950–2005, and Eureka, 1948–2007). Bold type indicates a significant trend ($p < 0.05$).

	Reanalysis data				Environment Canada	
Variable	Lakes A and B	Lakes C1, C2, and C3	Alert	Eureka	Alert	Eureka
Annual mean surface air temperature (°C)	**+0.48***	**+0.48***	**+0.43***	**+0.45**	+0.14	**+0.22**
DJF mean surface air temperature (°C)	**+0.68**	**+0.68**	**+0.60***	**+0.42**	+0.14	**+0.30**
MAM mean surface air temperature (°C)	**+0.42**	**+0.41**	**+0.40**	**+0.43***	+0.21	+0.12
JJA mean surface air temperature (°C)	+0.08	+0.08	+0.03	**+0.21**	+0.02	+0.06
SON mean surface air temperature (°C)	**+0.70***	**+0.70***	**+0.66***	**+0.68***	**+0.29**	**+0.36**
Annual variance of daily surface air temperatures	**−6.90***	**−6.86**	**−6.44***	**−3.59***	−1.85	−3.16
Annual total melting degree days (°C d)	+5.16*	+5.22*	+2.42	**+9.29***	−1.21	+5.22
Annual total freezing degree days (°C d)	**−170***	**−170***	**−154***	**−154**	**−53.6**	**−74.0**
Melt season length (d)	+0.78	+1.00*	−0.58	**+3.82***	+0.41	−0.27
Annual total rainfall (mm)	—	—	—	—	−0.54*	+2.33*
Annual total snowfall (cm)	—	—	—	—	**+13.97**	**+4.97**
Annual total precipitation (mm)	—	—	—	—	+0.96*	**+4.45***

* A nonnormally distributed variable.

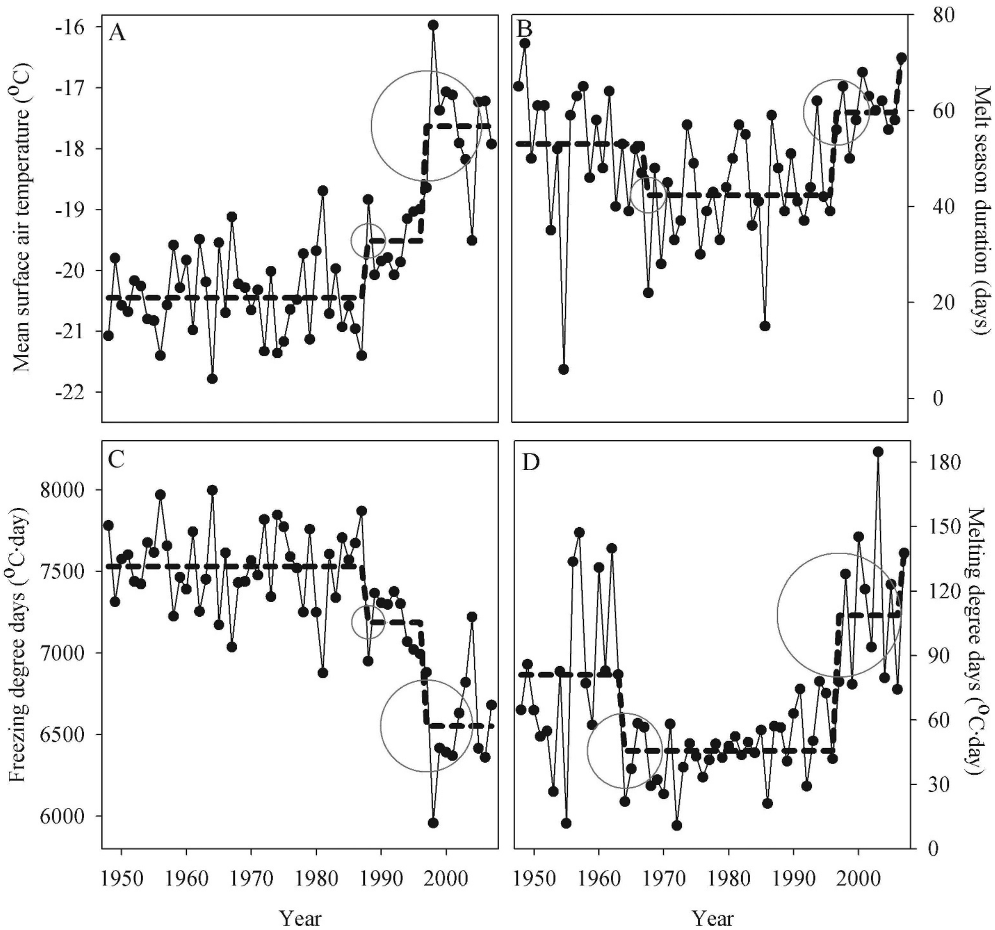

Fig. 6. Climate reanalysis data and regime for the lakes A and B region. (A) Mean surface air temperature. (B) Melt season duration. (C) Total freezing degree days. (D) Total melting degree days. Each variable is shown with a solid circle and line, the regime mean and shift timings are shown with a heavy dashed line and the regime-shift index (RSI) is proportional to the diameter of the circle plotted at the start of new regimes. Only significant ($p \leq 0.05$) regime-shift indices are shown.

temperature remained in the same regime (at an average air temperature of −20.4°C) for 40 yr until a significant shift (RSI = 0.35, $p < 0.0001$) to a new mean of −19.5°C in 1988 (Fig. 6A). This regime lasted for 9 yr, when in 1997 the mean air temperature jumped to −17.6°C (RSI = 1.1, $p < 0.0001$). The same overall pattern was observed for the C lakes region, but with the timing of the second regime shift delayed until 1998 (Fig. 7A). Melt season duration was elevated for the first 20 yr on record (regime mean = 53 d) until a shift in 1968 (RSI = −0.36, $p = 0.016$) to a 29-yr period with less prolonged melt. In 1997, a new, even longer (60 d) melt season regime began (RSI = 0.67, $p < 0.0001$) followed by an insignificant shift to a 71-d-long melt season in the last year on record (Fig. 6B). The C lakes data followed the same pattern, but the first regime shift to shorter melt seasons occurred 5 yr earlier, in 1963 (Fig. 7B). The first portion of the climate record had an average MDD of 81°C d, which dropped by almost half in 1964 (RSI = −0.76, $p = 0.001$) only to rise to 108.7°C d in 1997 (RSI = 1.25, $p < 0.0001$; Fig. 6D). The timings of MDD regimes in the C lakes data did not differ from those of lakes A and B (Fig. 7D). The regime pattern for total freezing degree days mirrors that of the air temperature (including the timing of shifts), as expected, due to the contribution of autumn, winter, and spring warming to the annual average (Figs. 6C, 7C).

Discussion

The timing of climate, ice cover, and water-column regime shifts is summarized in Fig. 8. Prior to 1988, the climate was colder, in situ observations indicate a perennial ice cover for the five lakes (Hattersley-Smith et al. 1970; Jeffries and Krouse 1985; Jeffries et al. 1984), and upper water-column mixing was observed only once (in the late 1960s following a period of warm summers). In 1988, a regime shift in mean air temperature and FDD was associated with the onset of infrequent lake ice loss (recorded first in 1991 and immediately followed by several years without SAR observation; Table 2). While there is no remote sensing or other direct evidence of ice loss in Lake B during this period, water-column mixing suggests that the

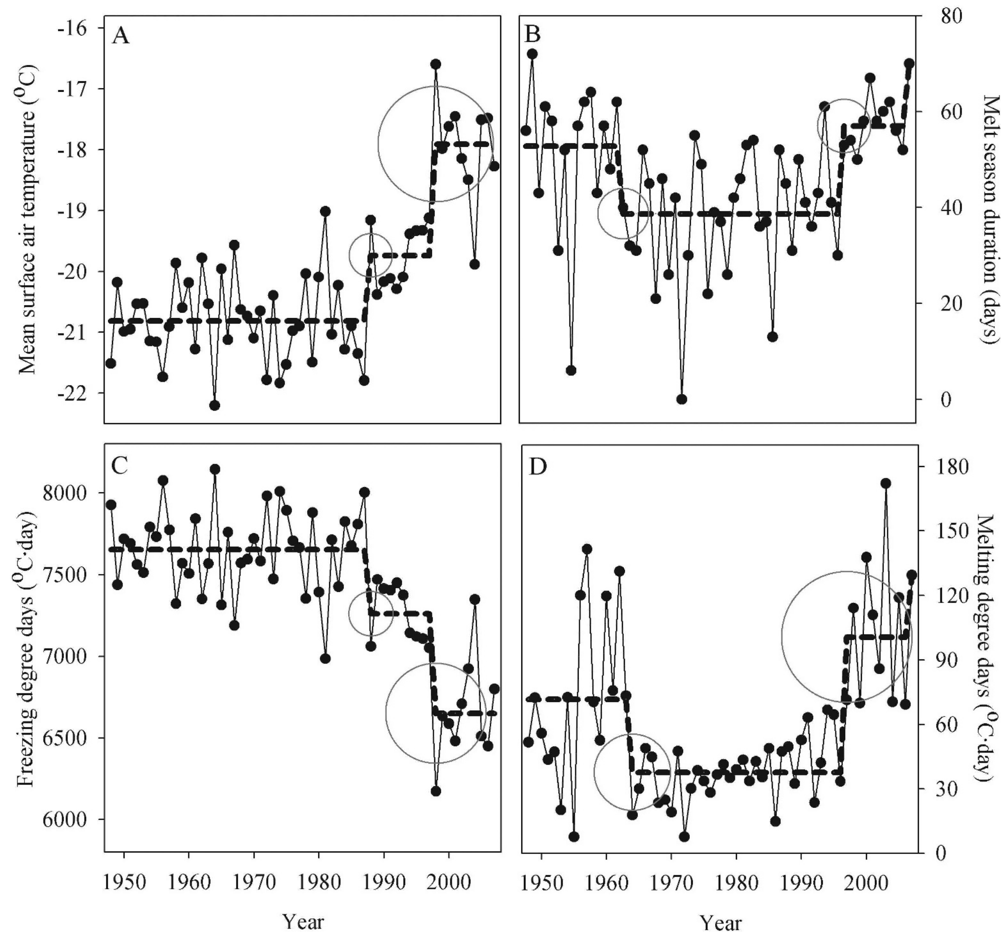

Fig. 7. Climate reanalysis data and regime for the lakes C1, C2, and C3 region. (A) Mean surface air temperature. (B) Melt season duration. (C) Total freezing degree days. (D) Total melting degree days. Each variable is shown with a solid circle and line, the regime mean and shift timings are shown with a heavy dashed line and the regime-shift index (RSI) is proportional to the diameter of the circle plotted at the start of new regimes. Only significant ($p \leq 0.05$) regime-shift indices are shown.

ice may have melted in 1992 (hence the question mark in Fig. 8). Between 1997 and 1998, a second climate regime shift instigated a change in lake ice phenology from infrequent to frequent summer loss in all lakes. This, in turn, initiated a regime where wind-driven water-column mixing took place on a regular basis. For lakes C1 and C2, the onset of this newest mixing regime may have occurred following ice losses in 1998, but there was no direct evidence of mixing until 2001 (Fig. 8). These data provide a clear example of cascading regime shifts in which one regime shift leads to a series of others (Kinzig et al. 2006). In this system, regime shifts in air temperature induced shifts in lake ice phenology, which in turn brought about new regimes in water-column mixing that supplanted previous long-term stability. Evidence of regime shifts in the forcing (i.e., air temperature) variables (Figs. 6, 7) are supported not only by statistical analyses, but also by resultant changes further down the regime-shift cascade (Fig. 8).

These regime shifts are compelling sentinels of climate warming that are easy to detect and causally unambiguous. They also underscore the overarching importance of climate to numerous processes in lake ecosystems. The resultant cascade of abrupt and discontinuous change between stable states has wide-ranging limnological implications, including the potential disruption of biogeochemical stratification and microbial habitats in these lakes. Lake A contains the most northerly population of Arctic char in North America (J. Babaluk pers. comm.) and these fish likely feed on zooplankton, such as the copepods *Limnocalanus macrurus* and *Drepanopus bungei* (Van Hove et al. 2001). These meromictic lakes also contain a variety of habitats from freshwater to saline and from oxic to anoxic with steep ionic gradients in the water column (Van Hove et al. 2006). A diverse assemblage of protists such as chlorophytes, cryptophytes, dinoflagellates, and ciliates was found in these lakes in addition to cyanobacteria, photosynthetic sulfur bacteria (Belzile et al. 2001; Antoniades et al. in press), and Archaea (Pouliot et al. 2009). Since the ecology of these lakes is dependent on the water-column regime, these complex habitats and communities are vulnerable to climate forcing and could themselves be

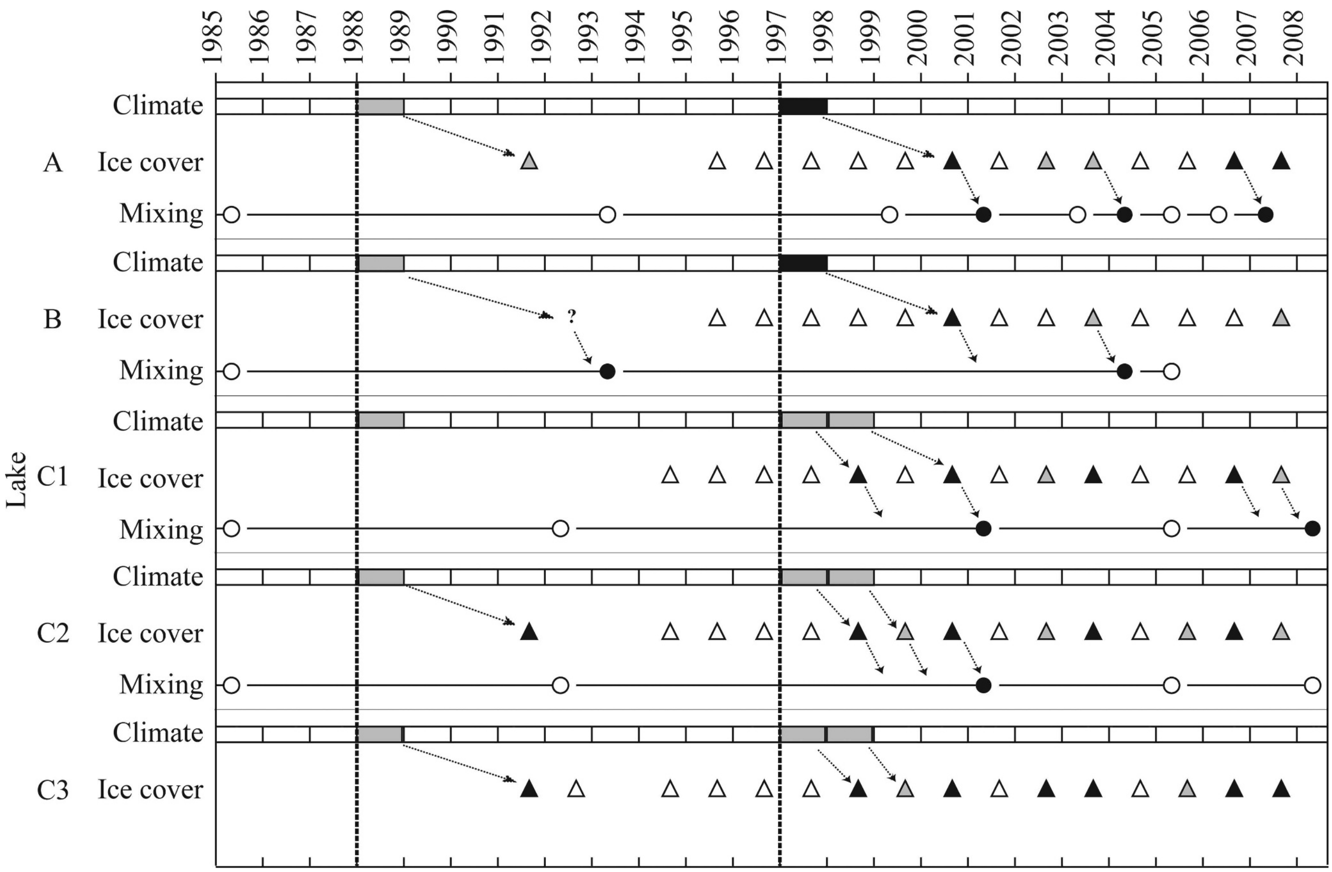

Fig. 8. Regime shifts in climate, ice cover loss, and mixing events since 1985. The first year of a climate regime shift is indicated by rectangles (solid, shift in mean air temperature, MDD, FDD, and melt season; gray, shift in only two of these variables; open, no shift). Ice loss is shown by triangles (solid, >50% of lake area; gray, <50% of lake area; open, no loss). Solid circles represent profiles that recorded evidence of water-column mixing since the previous profile (i.e., a mixing event(s) occurred at some time along the horizontal line to the left of each of these circles). Hollow circles represent profiles with no evidence of past mixing (i.e., horizontal lines to the left of these circles represents periods of water-column stability). Arrows indicate causal links from air temperature increases, to ice loss and mixing. Vertical dashed lines show the onset of regime-shift cascades through the ecosystem starting with a new climate regime that forces a regime change in lake ice cover and in turn initiates new regimes in water-column mixing. Prior to 1988 climate was colder, lake ice was perennial, and mixing rare. From 1988 to 1997, climate was warmer, lake ice loss was infrequent, and mixing rare. From 1997 onward, air temperature increased, lake ice loss became frequent, and lake water columns mixed on a regular basis.

considered a level in the regime-shift cascade. Ice-covered meromictic lakes are also found in several parts of Antarctica (Vincent and Laybourn-Parry 2008), and the cascading effects of climate change on Ellesmere Island lakes may provide insights into eventual climate-driven effects on these south polar ecosystems.

Examining regime changes in combination with climate trends gave a more complete understanding of how the high Arctic climate is changing and confirmed the timing of major environmental shifts in limnetic ecosystems. Simple linear trends calculated from complete climate datasets are misleading because the 1970s and 1980s were cooler relative to early and late periods on record. The regime analysis and climate plots show that annual and winter warming has occurred in two stages, while summer warmth increased abruptly during the last decade.

Reanalysis data helped overcome the lack of long-term climate records for the northern Ellesmere Island region. Using data from the closest Environment Canada climate stations, Alert and Eureka, may not be appropriate since the Alert climate is likely influenced by the Lincoln Sea and the Greenland Ice Cap, whereas Eureka is surrounded by mountains and has a relatively continental climate. Reanalysis data are based on interpolations of sparse data, and each grid cell represents the climate over a large area (Kalnay et al. 1996; Christy et al. 2006). Two-thirds of the grid cells that we used to represent lake climate cover ice caps and mountains, with the Arctic Ocean occupying the northern portion (Fig. 1A). Despite this, there was general agreement between the trends and regimes of the lake grid cells and the nearest climate station data. Therefore, the actual climate at the lakes is likely bounded between the reanalysis estimates and the observations at Alert and Eureka.

The compilation of SAR images presented here shows that the lakes of northern Ellesmere Island have undergone substantial environmental change over the last two decades. The SAR methodology for detecting ice loss worked well in all years, with the possible exception of summer 1991 where one image (acquired in March 1992) suggested a complete loss of ice on Lake C3 and a partial loss on lakes A and C2. Losses from recent years were less ambiguous since they were confirmed using several early winter or autumn SAR scenes. However, field observations from June 1992 indicated that at least partial ice cover loss on Lake C2 occurred in 1991, which contrasted with no evidence of loss the previous year (M. Retelle pers. comm.).

Freshwater ice phenology is a complex function of interactions and feedbacks among meteorological factors, physical characteristics of lakes and the ice itself (Jeffries et al. in press). However, over the long term, freshwater ice phenology is primarily a function of air temperature and a strong indicator of climate variation and change (Jeffries et al. in press). Ice phenology also provides insight into how climate change affects aquatic environments since it has major implications for physical, biogeochemical, and food web processes. The synchronous shifts in reanalysis air temperature and lake ice phenology in these lakes further demonstrate the suitability of lake ice cover as a sentinel of climate change, particularly in regions where measured climate data are lacking.

A significant proportion of warming occurred in the fall, winter, and spring, resulting in the reduction of FDD over time. One consequence of lower FDD is reduced ice growth (and higher temperatures within the ice), which results in a weaker ice cover. For instance, using a modified Stefan's formula (U.S. Army Corps of Engineers 2006) and assuming no change in other parameters of the energy balance of the lake ice, we calculated a 7% reduction in ice thickness between the first and last FDD regime for these lakes. Even small shifts in summer air temperature and melt season duration may now be enough to melt the lake ice cover or increase meltwater production beyond a critical threshold, leading to ice breakup.

In the summer, the heat received by a given ice cover is a function of the energy balance, as well as the inflow of warm water from the catchment. Changes in air temperature (and related climate variables) were the focus of the present study, but other climate variables could also drive changes at the watershed scale. Increases in snowfall, as observed at both Alert and Eureka, can act to insulate and reflect solar energy away from the ice cover. Modeling results indicate that, as on all frozen lakes, Lake A winter conductive heat flux is very sensitive to snow depth (Vincent et al. 2008), which suggests ice thickness would likely decrease under higher snowfall regimes. Changes in cloud cover can affect both solar and long-wave fluxes to the catchment, while wind speed and direction, along with air temperature and humidity, influence sensible and latent heat fluxes. Once ice covers are weakened through melting, other factors such as wind direction, duration, and intensity are likely to play major roles in subsequent physical breakup of the ice cover.

Changes in climate variables must be viewed with respect to the physiography of each watershed to understand why certain lakes lose their ice cover more readily than others. Lakes with glacierized catchments (lakes C2 and C3) may be easily influenced by abundant meltwater in warm summers. In contrast, lakes with small watersheds without glaciers or inflowing streams receive far lower meltwater inputs (e.g., lakes B and C1). Lake B is sheltered from both wind and sun by topography, which acts to minimize ice loss, whereas Lake A is the largest and deepest of the five lakes, which favors its retention of ice. Finally, our climate analysis indicates that the C lake region was, and still is, slightly ($<1^{\circ}$C) cooler than Lake A and B, which may help to overcome landscape differences that enhance ice loss. Some of the factors that augment ice loss are also implicated in water-column mixing, such as lack of topography (i.e., wind fetch), greater catchment area, and enhanced hydrologic input. Lakes that are less protected are likely to undergo wind-induced mixing to greater depths (e.g., compare mixing depths in Lake A vs. Lake B; Fig. 5E,D).

Most perennially ice-covered lakes are in contact with glaciers, which act to stabilize their ice covers (Doran et al. 1996*b*). Perennially ice-covered lakes without glacial contact must therefore be relatively sensitive to climate forcing and are found only in the coldest climates (at our northern Ellesmere Island study site and Antarctica). Lakes with occasional residual ice covers are more common, such as Lake Hazen (538 km^2 at 81.7°N), which had some residual ice covers in the 1950s (Hattersley-Smith 1974), and Stanwell-Fletcher Lake, which was known to have residual ice covers in the early 1960s (292 km^2 at 72.7°N; Coakley and Rust 1968). Colour Lake on Axel Heiberg Island (1993 mean annual air temperature: -15.2°C) had five documented residual ice covers over the period 1959–1995 (Doran et al. 1996*a*). The analysis of this record revealed that two consecutive residual ice covers are not likely to occur because the ice cover traps heat in the water-column, which reduces ice formation in the subsequent winter (Doran et al. 1996*a*). The colder climate at the north coast of Ellesmere Island likely overcame this negative feedback for the lakes in our study; however, these lakes have now switched to a new regime where annual ice loss is common. In this new regime, accompanied by a mixed upper water column with a lower heat content, seasonal ice may even attain greater thicknesses due to this feedback mechanism.

Climate-driven freshwater ice phenology change has occurred widely in the Northern Hemisphere since the mid-19th century (Magnuson et al. 2000), often with a nonlinear relationship to air temperature change (Weyhenmeyer et al. 2004). Lake ice changes have induced major, rapid shifts in lake ecosystem properties in remote regions in the Arctic (Smol et al. 2005), in the Antarctic (Quayle et al. 2002), and in the Canadian boreal forest (Schindler et al. 1990).

Recent climate warming in the Arctic has caused perennially ice-covered lakes to break up, which exposes their water columns to wind-driven mixing. Water-column profiles therefore provide evidence of major discontinuous shifts within these lakes due to climate forcing. Salinity and water temperature structure are valuable sentinels because they show changes within a lake that cannot be detected by remote sensing and they represent the integration of changes that occurred between successive profiles. Given the severity of the mixing events that are recorded here and the coincidence with ice loss on the lakes, other mechanisms, such as overflow of hydrologic inputs from the catchment or saltwater exclusion from ice cover (which cannot be reconciled with the magnitude of increase in surface salinity or the decrease in salinity close to the halocline from one profile to the next), cannot adequately explain the observed changes.

An indication of the relative magnitude of the wind energy required during the brief ice-free period to mix the upper water column of Lake A can be obtained from the stability, or minimum work required for complete mixing. To mix the lake to a depth of 40 m (well below the halocline) it would take 43,000 J m^{-2}. This value is approximately equivalent to the stability of 27-m-deep Soap Lake in Washington (Walker 1974). Using the 1999 profile, it took 58.5 J m^{-2} to mix the water column to a depth of 10.5 m (the bottom of the mixed layer observed in 2001). A recently mixed salinity profile will recover very slowly with the upward diffusion of salt and the influx of fresh water at the surface. Temperature profiles in the upper water column recover faster but still take several years to regain the same shape as before (Vincent et al. 2008). Once a mixed layer is formed, it takes less energy to remix this weakly stratified water in subsequent years. With no ice cover to prevent mixing, it would have taken 15.6 J m^{-2} to mix the 2003 profile to 7 m and 18.3 J m^{-2} to mix the 2006 profile to 10 m. However, 61% of the lake was ice covered in 2003 (48% in 2006). Therefore, based on the portion of open water available to receive wind energy, it took 40.0 J m^{-2} and 35.2 J m^{-2} (i.e., 156% and 92% more energy than an ice-free lake) to accomplish the observed mixing during those years, respectively. This underscores the importance of residual ice pans (or covers), which act as a barrier and keep surface water at 0°C, no matter how thin or weakened they become. To mix the water column, the combination of wind and extent of ice loss must exceed the stability threshold, which explains why in some years there was ice loss but no mixing (Fig. 8).

As these lakes become further influenced by a warming climate and ice-free conditions prevail, the upper portions of the water column will become regularly mixed. At this point, it may be necessary to examine the deeper portions of water temperature profiles to evaluate the changes in lake heat content. Vincent et al. (2008) showed how long-term seasonal ice loss will modify the Lake A water-column temperature profile. The resulting heat loss in this scenario was projected to reduce the temperature of the deep-water thermal maximum, while increasing its depth. Our results do not show any systematic change in the thermal maximum over the last decade (Table 1), but it is conceivable that these changes will occur with seasonal ice loss under future warming scenarios. Long-term change may also lead to further loss of water-column stratification, which is unprecedented for these 2500–4000-yr-old lakes (Jeffries and Krouse 1985).

The major limnological changes documented here are consistent with other climate-related events along this northern coastline of Ellesmere Island. The Ellesmere ice shelves, which are the largest of their kind in the Arctic, owe their origin to climate cooling that occurred 3000–4500 yr ago. The collapse of the ice shelves has accelerated in recent years (Mueller et al. 2003, 2008) and has raised concern about the loss of unique, ice-associated microbial ecosystems (Mueller et al. 2006). The collapse of the ice shelves is also leading to the drainage of rare epishelf lakes (freshwater overlying tidal seawater) and the extinction of these ice-dependant ecosystems (Mueller et al. 2003; Veillette et al. 2008).

The lacustrine environments of northern Ellesmere Island are currently undergoing an abrupt transition from permanently ice-covered to seasonally melting ice cover. These ecosystems appear to be highly sensitive to the threshold effects of small step changes in the air temperature regime, and are therefore valuable long-term monitoring sites for global change. In this study, we have documented cascading regime shifts that started with shifts in air temperature, leading to transformations in lake ice cover. Once the lake ice was removed, the water column was exposed to wind-driven mixing, resulting in a new regime of water-column structure and stability. Our four decade-long record suggests that lake salinity and temperature structure are valuable sentinels of climate change. The timing of profile shifts matched periods of documented ice loss, which increased in the years following regime shifts detected in climate variables. Lake changes are occurring in concert with the latest episode (c. 2002 to 2008) of ice shelf breakup and loss along the coast of northern Ellesmere Island (Mueller et al. 2006, 2008). These transformations are consistent with major climate-related changes elsewhere in the Arctic cryosphere, such as on permafrost, sea ice, and glaciers (Serreze et al. 2000), and they are indications of change beyond the regional scale.

Acknowledgments

We thank Parks Canada for use of facilities and Polar Shelf, Natural Resources Canada (this is Polar Continental Shelf Project publication number 02309), the Canadian Rangers, and the Northern Scientific Training Program for logistical support. National Centers for Environmental Prediction and National Center for Atmospheric Research reanalysis data were provided by the National Oceanic and Atmospheric Administration, Earth System Research Laboratory, Physical Sciences Division, Boulder, Colorado (www.cdc.noaa.gov). Water-column profile data from 1985 were provided by Mike Retelle (Bates College), and satellite imagery was provided through a National Aeronautics and Space Administration data grant, processed by the Alaska Satellite Facility, and analyzed using resources at the Arctic Region Supercomputing Center. Field assistance was provided by Julie Veillette, Sébastien Roy, Denis Sarrazin, Jeffrey Kheraj, Donald Burke, Joeli Qaunaq, and Pitisulaq Ukuqtunnuaq. We thank Peter Adams and two anonymous reviewers for their insightful comments on the manuscript. Funding assistance was provided by ArcticNet, the Canada Research Chair program, Natural Sciences and Engineering Research Council of Canada, Fonds québécois de la recherche sur la nature et les technologies, and Trent University. The contribution of M.O.J. is based upon work supported by Individual Research and Development funds while serving at the National Science Foundation (NSF); any opinion, findings, and conclusions or recommendations expressed in this material are those of the authors and do not necessarily reflect the views of NSF.

References

[ACIA] ARCTIC CLIMATE IMPACT ASSESSMENT. 2005. Arctic climate impact assessment. Cambridge Univ. Press.

Antoniades, D., J. Veillette, M.-J. Martineau, C. Belzile, J. D. Tomkins, R. Pienitz, S. Lamoureux, and W. F. Vincent. In press. Bacterial dominance of phototrophic communities in a High Arctic lake and its implications for paleoclimate analysis. Polar Sci.

Belzile, C., W. F. Vincent, J. A. E. Gibson, and P. Van Hove. 2001. Bio-optical characteristics of the snow, ice, and water column of a perennially ice-covered lake in the High Arctic. Can. J. Fish. Aquat. Sci. **58:** 2405–2418, doi: 10.1139/cjfas-58-12-2405.

Bradley, R. S., M. J. Retelle, S. D. Ludlam, D. R. Hardy, B. Zolitschka, S. F. Lamoureux, and M. S. V. Douglas. 1996. The Taconite Inlet lakes project: A systems approach to paleoclimatic reconstruction. J. Paleolimnol. **16:** 97–110.

Christy, J. R., D. J. Seidel, and S. C. Sherwood. 2006. What kinds of atmospheric temperature variations can the current observing systems detect and what are their strengths and limitations, both spatially and temporally? p. 29–46. *In* T. R. Karl, S. J. Hassol, C. D. Miller, and W. L. Murray [eds.], Temperature trends in the lower atmosphere: Steps for understanding and reconciling differences. Climate Change Science Program and the Subcommittee on Global Change Research.

Coakley, J. P., and B. R. Rust. 1968. Sedimentation in an Arctic lake. J. Sediment. Petrol. **38:** 1290–1300.

Comiso, J. C. 2003. Warming trends in the Arctic from clear sky satellite observations. J. Clim. **16:** 3498–3510.

Crease, J., T. M. Dauphinee, P. L. Grose, E. L. Lewis, N. P. Fofonoff, E. A. Plakhin, K. Striggow, and W. Zenk. 1988. The acquisition, calibration and analysis of CTD data. UNESCO.

Doran, P. T., C. P. McKay, W. P. Adams, M. C. English, R. A. Wharton, and M. A. Meyer. 1996*a*. Climate forcing and thermal feedback of residual lake-ice covers in the high Arctic. Limnol. Oceanogr. **41:** 839–848.

———, ———, M. A. Meyer, D. T. Andersen, R. A. Wharton, Jr., and J. T. Hastings. 1996*b*. Climatology and implications for perennial lake ice occurrence at Bunger Hills oasis, East Antarctica. Ant. Sci. **8:** 289–296.

Fofonoff, N. P., and R. C. Millard. 1983. Algorithms for computation of fundamental properties of seawater. UNESCO Division of Marine Sciences.

Grebmeier, J. M., and others. 2006. A major ecosystem shift in the northern Bering Sea. Science **311:** 1461–1464, doi: 10.1126/science.1121365.

Hansen, J., M. Sato, R. Ruedy, K. Lo, D. W. Lea, and M. Medina-Elizade. 2006. Global temperature change. Proc. Natl. Acad. Sci. USA **103:** 14288–14293, doi: 10.1073/pnas.0606291103.

Hattersley-Smith, G. 1974. North of latitude eighty. Defence Research Board.

———, J. E. Keys, H. Serson, and J. E. Mielke. 1970. Density stratified lakes in Northern Ellesmere Island. Nature **225:** 55–56.

Idso, S. 1973. On the concept of lake stability. Limnol. Oceanogr. **18:** 681–683.

Jeffries, M. O., and H. R. Krouse. 1985. Isotopic and chemical investigations of two stratified lakes in the Canadian Arctic. Z. Gletsch.kd. Glazialgeol. **21:** 71–78.

———, ———, M. A. Shakur, and S. A. Harris. 1984. Isotope geochemistry of stratified Lake "A," Ellesmere Island, N.W.T., Canada. Can. J. Earth Sci. **21:** 1008–1017.

———, K. Morris, and C. R. Duguay. In press. Lake and river ice. *In* R. S. Williams, Jr., and J. G. Ferrigno [eds.], State of the Earth's cryosphere at the beginning of the 21st century: Glaciers, snow cover, floating ice, and permafrost and periglacial environments. Satellite Image Atlas of Glaciers of the World. U.S. Geological Survey Professional Paper 1386-A.

———, ———, and N. Kozlenko. 2005. Ice characteristics and processes, and remote sensing of frozen rivers and lakes, p. 63–90. *In* C. R. Duguay and A. Pietroniero [eds.], Remote sensing in northern hydrology. American Geophysical Union.

Kalnay, E., and others. 1996. The NCEP/NCAR 40-year reanalysis project. Bull. Am. Meteorol. Soc. **77:** 437–470.

Kinzig, A. P., P. Ryan, M. Etienne, H. Allison, T. Elmqvist, and B. H. Walker. 2006. Resilience and regime shifts: Assessing cascading effects. Ecol. Soc. **11:** 20, http://www.ecologyandsociety.org/vol11/iss1/art20/.

Ludlam, S. D. 1996. The comparative limnology of high arctic, coastal, meromictic lakes. J. Paleolimnol. **16:** 111–131.

Magnuson, J. J., and others. 2000. Historical trends in lake and river ice cover in the Northern Hemisphere. Science **289:** 1743–1746.

Maslanik, J. A., C. Fowler, J. Stroeve, S. Drobot, J. Zwally, D. Yi, and W. Emery. 2007. A younger, thinner Arctic ice cover: Increased potential for rapid, extensive sea-ice loss. Geophys. Res. Lett. **34:** L24501, doi: 10.1029/2007GL032043.

Mueller, D. R., L. Copland, A. Hamilton, and D. R. Stern. 2008. Examining Arctic ice shelves prior to 2008 breakup. EOS Trans. Am. Geophys. Union **89:** 502–503.

———, W. F. Vincent, and M. O. Jeffries. 2003. Break-up of the largest Arctic ice shelf and associated loss of an epishelf lake. Geophys. Res. Lett. **30:** 2031, doi: 10.1029/2003GL017931.

———, ———, and ———. 2006. Environmental gradients, fragmented habitats and microbiota of a northern ice shelf cryoecosystem, Ellesmere Island, Canada. Arct. Ant. Alp. Res. **38:** 593–607.

Pouliot, J., P. E. Galand, C. Lovejoy, and W. F. Vincent. 2009. Vertical structure of archaeal communities and the distribution of ammonia monooxygenase A gene variants in two meromictic High Arctic lakes. Environ. Microbiol. **11:** 687–699, doi: 10.1111/j.1462-2920.2008.01846.x.

Quayle, W. C., L. S. Peck, H. Peat, J. C. Ellis-Evans, and P. R. Harrigan. 2002. Extreme responses to climate change in Antarctic lakes. Science **295:** 645.

Rodionov, S. N. 2004. A sequential algorithm for testing climate regime shifts. Geophys. Res. Lett. **31:** L09204, doi: 10.1029/2004GL019448.

———. 2006. Use of prewhitening in climate regime shift detection. Geophys. Res. Lett. **33:** L12707, doi: 10.1029/2006GL025904.

Rühland, K., A. Priesnitz, and J. P. Smol. 2003. Paleolimnological evidence from diatoms for recent environmental changes in 50 lakes across the Canadian Arctic treeline. Arct. Ant. Alp. Res. **35:** 110–123.

Scheffer, M., S. Carpenter, J. A. Foley, C. Folke, and B. Walker. 2001. Catastrophic shifts in ecosystems. Nature **413:** 591–596.

Schindler, D. W., K. G. Beaty, E. J. Fee, D. R. Cruikshank, E. R. Debruyn, D. L. Findlay, G. A. Linsey, J. A. Shearer, M. P. Stainton, and M. A. Turner. 1990. Effects of climatic warming on lakes of the central boreal forest. Science **250:** 967–970.

Serreze, M., and J. Francis. 2006. The arctic amplification debate. Clim. Change **76:** 241–264, doi: 10.1007/s10584-005-9017-y.

———, and others. 2000. Observational evidence of recent change in the northern high-latitude environment. Clim. Change **46:** 159–207.

Smol, J. P., and others. 2005. Climate-driven regime shifts in the biological communities of arctic lakes. Proc. Natl. Acad. Sci. USA. doi: 10.1073/pnas.0500245102.

SOLOMON, S., D. QIN, M. MANNING, Z. CHEN, M. MARQUIS, K. B. AVERYT, M. TIGNOR, AND H. L. MILLER [eds.]. 2007. Climate change 2007: The physical science basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge Univ. Press.

U.S. ARMY CORPS OF ENGINEERS. 2006. Engineering and design: Ice engineering. http://140.194.76.129/publications/eng-manuals/em1110-2-1612/change2.pdf.

VAN HOVE, P., C. BELZILE, J. A. E. GIBSON, AND W. F. VINCENT. 2006. Coupled landscape-lake evolution in the coastal high Arctic. Can. J. Earth Sci. **43:** 533–546, doi: 10.1139/E06-003.

———, K. SWADLING, J. A. E. GIBSON, C. BELZILE, AND W. F. VINCENT. 2001. Farthest north lake and fjord populations of calanoid copepods *Limnocalanus macrurus* and *Drepanopus bungei* in the Canadian high Arctic. Polar Biol. **24:** 303–307, doi: 10.1007/s003000000207.

———, W. F. VINCENT, P. E. GALAND, AND A. WILMOTTE. 2008. Abundance and diversity of picocyanobacteria in High Arctic lakes and fjords. Algol. Stud. **126:** 209–227.

VEILLETTE, J., D. R. MUELLER, D. ANTONIADES, AND W. F. VINCENT. 2008. Arctic epishelf lakes as sentinel ecosystems: Past, present and future. J. Geophys. Res. **113:** G04014, doi: 10.1029/2008JG000730.

VINCENT, A. C., D. R. MUELLER, AND W. F. VINCENT. 2008. Simulated heat storage in a perennially ice-covered high Arctic lake: Sensitivity to climate change. J. Geophys. Res. **113:** C04036, doi: 10.1029/2007JC004360.

VINCENT, W. F., AND J. LAYBOURN-PARRY [eds.]. 2008. Polar lakes and rivers—limnology of Arctic and Antarctic aquatic ecosystems. Oxford Univ. Press.

WALKER, K. F., 1974. The stability of meromictic lakes in central Washington. Limnol. Oceanogr. **19:** 209–222.

WALLACE, J. M., Y. ZHANG, AND L. BAJUK. 1996. Interpretation of interdecadal trends in Northern Hemisphere surface air temperature. J. Clim. **9:** 249–259.

WEYHENMEYER, G. A., M. MEILI, AND D. M. LIVINGSTONE. 2004. Nonlinear temperature response of lake ice breakup. Geophys. Res. Lett. **31:** L07203, doi: 10.1029/2004GL019530.

Editor: Everett Fee

Received: 15 September 2008
Accepted: 05 May 2009
Amended: 18 May 2009