Coupled landscape–lake evolution in High Arctic Canada

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Abstract: We profiled five ice-covered lakes and two ice-covered fiords of Ellesmere Island at the northern limit of High Arctic Canada to examine their environmental characteristics, and to evaluate the long-term limnological consequences of changes in their surrounding landscape through time (landscape evolution). All of the ecosystems showed strong patterns of thermal, chemical, and biological stratification with subsurface temperature maxima from 0.75 to 12.15 °C; conductivities up to 98.1 mS cm^{-1} (twice that of seawater) in some bottom waters; pronounced gradients in nitrogen, phosphorus, pH, dissolved inorganic and organic carbon, manganese, iron, and oxygen; and stratified photo-synthetic communities. These ecosystems form an inferred chronosequence that reflects different steps of landscape evolution including marine embayments open to the sea, inlets blocked by thick sea ice (Disraeli Fiord, Taconite Inlet), perennially ice-capped, saline lakes isolated from the sea by isostatic uplift (Lakes A, C1, C2), and isolated lakes that lose their ice cover in summer. The latter are subject to entrainment of saline water into their upper water column by wind-induced mixing (Lake Romulus; Lake A in 2000), or complete flushing of their basins by dilute snowmelt (Lake C3 and Char Lake, which lies 650 km to the south of the Ellesmere lakes region). This chronosequence illustrates how changes in geomorphology and other landscape properties may influence the limnology of coastal, high-latitude lakes, and it provides a framework to explore the potential impacts of climate change.

Résumé : Cinq lacs et deux fjords recouverts de glace ont été profilés sur l'Île Ellesmere à la limite nord de l'Extrême-Arctique canadien afin d'examiner leurs caractéristiques environnementales et d'évaluer les conséquences limnologiques à long terme des changements, dans le temps, des paysages voisins (évolution du paysage). Tous les écosystèmes montraient de forts patrons de stratification thermique, chimique et biologique avec des températures maximales sous la surface de 0,75 à 12,15 ºC, des conductivités atteignant 98,1 mS cm^{-1} (deux fois celle de l'eau de mer) dans certaines eaux de fond, des gradients prononcés d'azote, de phosphore, de pH, de carbone dissous organique et inorganique, de manganèse, de fer et d'oxygène ainsi que des communautés photosynthétiques stratifiées. Ces écosystèmes forment une chronoséquence inférée qui reflète les différentes étapes de l'évolution du paysage, incluant les baies marines ouvertes à la mer, des bras bloqués par d'épaisses glaces marines (fjord Disraeli, bras Taconite), des lacs salins couverts de glace de façon permanente, isolés de la mer par le relèvement isostatique (lacs A, C1 et C2), et des lacs isolés qui perdent leur couvert de glace durant l'été. Ces derniers sont soumis à l'entraînement d'eau saline dans leur colonne d'eau supérieure par le brassage induit par le vent (lac Romulus; lac A en 2000) ou au lessivage complet de leur bassin par de l'eau de fonte diluée (lac C3 et le lac Char qui se trouve à 650 km au sud de la région des lacs d'Ellesmere). Cette chronoséquence montre comment les changements de la géomorphologie et des autres propriétés du paysage peuvent influencer la limnologie des lacs côtiers de haute latitude et elle fournit un cadre pour explorer les impacts potentiels du changement climatique.

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Introduction

Lake ecosystems are strongly coupled to features of their surrounding landscapes such as geomorphology, lithology, vegetation and hydrological characteristics (Wetzel 2001). Many temperate lakes owe their origin to glacial action in the past that shaped their lake basin and surrounding land-scape, and that distributed glaciogenic deposits of different sources and mineral composition across their catchments. In the polar regions, glacial and periglacial processes continue today. High-latitude lakes, wetlands and rivers are likely to show ongoing responses to recent deglaciation, variations in ice-cover, and, for coastal waters, recent or current changes in their linkages with the sea.

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Fig. 1. Map of the sampling sites. (A) Canadian Arctic Archipelago (from http://maps.grida.no/arctic/); (B) study area on northern Ellesmere Island; (C) the Taconite Inlet area (shaded area in panel B), with catchment areas marked for Lakes C1, C2, and C3 (see Table 1 for note on Lake C3).
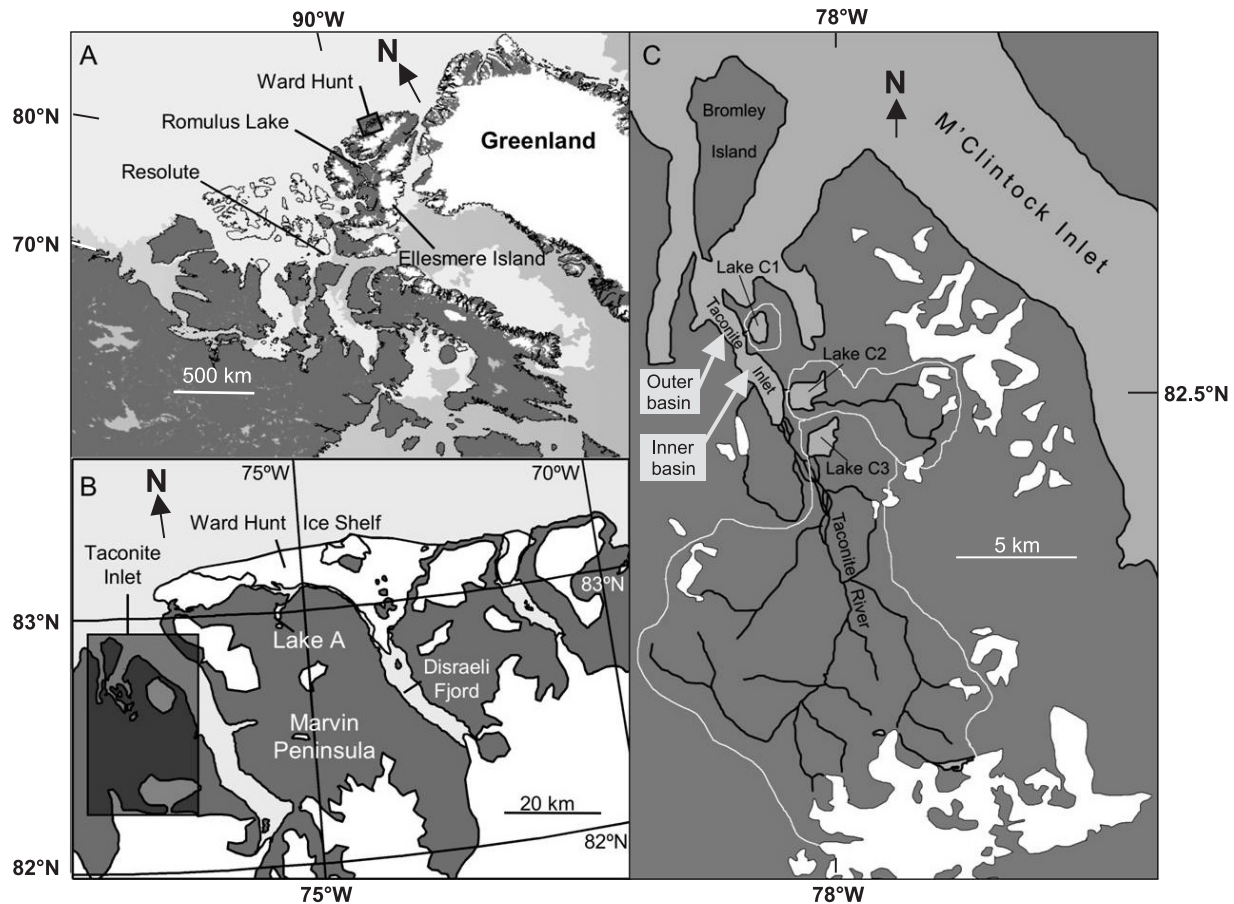
The high arctic landscape has been greatly influenced by the last glaciation. In these northernmost parts of the North American continent, the landscape is still evolving today in response to those glacial influences, mainly by isostatic adjustment of landmasses after glacial retreat. At the last glacial maximum (18 ka BP), the thick Innuitian Ice Sheet covered the Canadian High Arctic and extended to the Greenland Ice Sheet, closing Nares Strait and the Kane Basin (Blake 1970; England 1999; England et al. 2004; Tushingham 1991). Paleo-studies have shown that, after the glacial retreat and the opening of Nares Strait ca. 9.5 ka BP, there was regional warming of northern Ellesmere Island from around 8 ka BP to around 3–4 ka BP. This was followed by a period of cooling (Smol 1983; Bradley 1990; Smith 2002) that brought about the formation of an extensive ice shelf (Ellesmere Ice Shelf) along the northern coast of Ellesmere Island (see Vincent et al. 2001). The most recent period of warming appears to have begun in the mid 19th century (Douglas et al. 1994; Smol et al. 2005) and resulted in major loss of the Ellesmere ice shelves over the course of the 20th century (Vincent et al. 2001). An accelerated warming trend has been observed in this region from the 1980s to the present (Mueller et al. 2003; Antoniades et al. 2005).

Isostatic rebound and sea-level adjustments over the last few thousand years have caused an uplift of land relative to sea level by 80 to 150 m, resulting in gradual isolation of seawater-containing basins from the coastal marine environment (England 1999; Lemmen 1989). This has produced a series of saline lakes and fiords with freshwater overlying seawater. Although comparatively well described and common in coastal antarctic regions (Gibson 1999), saline lakes (i.e., with salinity >3 at least at some depth in the water column) have received little attention in arctic environments. In the Canadian Arctic, 12 saline lakes have been reported in the literature to date: Lakes A, B, C1, C2, C3, Tuborg, and Romulus on Ellesmere Island, Lake Sophia (Cornwallis Island), Garrow Lake (Little Cornwallis Island), and Ogac, Qasigialiminiq, and Tariujarusiq lakes (Baffin Island). These saline waters are covered by a thick layer of ice through most of the year, and on some lakes and years, the ice persists throughout all seasons.

The objectives of the present study were to characterize the limnological properties of marine-derived ecosystems on Ellesmere Island at the northern limit of Canada, and to evaluate their water column characteristics in relation to known changes in landscapes over the course of the Holocene. We profiled five lakes and two fiords, with further measurements at Char Lake, a well-known high arctic lake further to the south in the Canadian Arctic Archipelago. From this ensemble of measurements, we aimed to define the consequences of long-term climate change and landscape evolution on the limnology of this remote, high arctic lake district.

Study sites

Two stratified fiords (Taconite Inlet and Disraeli Fiord) and four meromictic lakes (Lakes A, C1, C2, C3) were sampled on the northern coast of Ellesmere Island. Additional measurements were obtained at meromictic Romulus Lake further to the south on Ellesmere Island near Eureka, and at Char Lake, a freshwater lake in Resolute Bay on Cornwallis Island. The mean annual air temperatures at these sites are –19.4 °C at Alert on the northern coast of Ellesmere, –19.7 °C at Eureka station and –16.4 °C at Resolute Bay (Meteorological Service of Canada, www.ec.gc.ca). Annual precipitation is much lower at Eureka (64 mm) than at Alert (154 mm) or Resolute (135 mm).

Lake A is a deep meromictic lake within a rocky, unglacierized catchment that rises to 600 m (Fig. 1). The lake is thought to have been cut off from the sea 2500–4000 years BP by the isostatic uplift of the northern Ellesmere coast (see Jeffries and Krouse 1985). It has been visited several times over the last forty years (Hattersley-Smith et al. 1970; Jeffries et al. 1984; Retelle 1986; Ludlam 1996a; Van Hove et al. 2001) and has shown a remarkable stability in its stratification throughout this period. It was visited twice over the course of the present study, on 3–8 June 1999 and on 1–4 August 2001.

Lakes C1, C2, and C3 are situated only a few kilometres apart on the edge of Taconite Inlet (Fig. 1), on rocky, mountainous terrain that rises to 1200 m and contains glaciers. Radiocarbon dating of the bottom waters of Lake C1 (Ludlam 1996a) and Lake A (Lyons and Mielke 1973) indicate a similar time of isolation from the sea, within the period 2500–4000 years BP. All three C-series lakes are situated near sea level (4, 1.5, and 10 m above sea level (asl), respectively), also suggesting a common isolation period. They have similar areas and basin depths, but differ substantially in the size of their catchments (Table 1). These lakes were extensively studied in a 1990–1992 paleolimnological research program centered on understanding sedimentary processes in Lake C2 (Bradley et al. 1996). Within the present study, they were visited on 20–28 July 2001.

Romulus Lake is a hypersaline meromictic lake similar in shape, area and basin size to Lake A, but in rolling lowland terrain on the Fosheim Peninusula of Ellesmere Island, 300 km to the south of the northern Ellesmere Island sites. This lake has the lowest water inputs per unit area of all the systems given its small catchment-to-lake area ratio plus its low precipitation regime (Table 1). It lies 8 m above sea level, and ^{14}C-dating of marine shells retrieved from 70 cm in sediment cores indicate that prior to 3600 BP Romulus Lake was connected to the sea via an extension of nearby Slidre Fiord (Davidge 1994). It was first surveyed in 1989 (Eggington and Hodgson 1990) and was described in more detail by Davidge (1994) and Jackson et al. (2004). During the course of this study, Romulus Lake was visited before the melt season on 20–25 May 2000.

Disraeli Fiord (Fig. 1B) is a deep, narrow, marine embayment that at the time of sampling on 9 June 1999 had a surface layer of freshwater dammed by the Ward Hunt Ice Shelf. At that time, this "epishelf" lake was one of the few known examples of such systems in the northern hemisphere. The freshwater surface layer subsequently drained following a

Table 1. Location, date of visit, average ice thickness and snow depth (±1 SD for n sites), maximum depth (Z_{max}), and lake area of the ice-covered lakes and fiords.

Site	Latitude	Longitude	Date of visit	Ice thickness (m)	Snow depth (m)	Z_{max} (m)	Lake area (km²)	Catchment area (km²)	Ratio Basin: Lake	Water load (m)	Chl a (µg L⁻¹)
Lake A	83°00′N	75°30′W	Jun 1999	1.97±0.01 (n = 5)	0.52±0.05 (n = 20)	128	5	37	7.4	1.3	0.53
			Aug. 2001	1.00±0.09 (n = 12)	0						0.32
Lake C1	82°51′N	78°12′W	July 2001	1.05±0.07 (n = 23)	0 (note 1)	65	1.1	3.3	3	0.6	0.79
Lake C2	82°50′N	78°05′W	July 2001	1.27±0.02 (n = 2)	0	84	1.8	23.5	13	2.2	0.67
Lake C3	82°48′N	78°05′W	July 2001	1.05 (n = 1)	0	51	1.7	180[a]	106	16.5	1.02
Romulus Lake	79°50′N	85°00′W	May 2000	2.51±0.01 (n = 3)	0.09±0.08 (n = 125)	60	4.4	28	6.4	0.5	0.47
Char Lake	74°42′N	94°50′W	June 1999	2.27 (n = 1)	0–0.20 (note 2)	28	0.53	4.4	8.3	1.3	0.22
Disraeli Fiord	82°50′N	73°40′W	June 1999	2.60 (n = 1)	0.55±0.11 (n = 6)	>400	143	2100	14.7	2.4	0.27
Taconite Inlet	82°50′N	78°15′W	July 2001	1.5 (n = 1)	0	>80	10	307	30.7	4.9	1.32

Note: This new snow disappeared gradually and was gone by 31 July. Lake A (1999), Disraeli Fiord and Taconite Inlet had multi-year ice cover while annual ice cover prevailed at the other sites. Ice measurements were made at mid-lake sites in ice auger or ice corer holes. Snow depth was measured along >10 m transects near the lake center. Water load was estimated as the product of the total catchment area (catchment plus lake or inlet area) and annual precipitation divided by the lake area. Chlorophyll a (Chl a) values are given as an index of trophic status, and are the maximum concentrations observed in the water column. No snow was present on 20–26 July, then a snowfall on July 27 left 2–4 cm of snow. Approximately 20% of the lake surface was free of snow.

[a]This assumes some input from the Taconite River; Bradley et al. (1996) note that very little of this river currently enters the lake and their estimates of the drainage basin area are 10.8 km² (Taconite River excluded,) and 260 km² (including the full drainage basin of the Taconite River).

partial breakup event on the Ward Hunt Ice Shelf (Mueller et al. 2003) and our measurements, therefore, provide a record of this system prior to major change.

Taconite Inlet (Fig. 1C) is an arm of M'Clintock Inlet in which freshwater has built up behind the multi-year sea ice at its seaward end, giving rise to stratified conditions. The Taconite River and all three C lakes drain into this fiord. The inlet is separated in two basins by a shallow sill and the inner basin is known to have a stratified water column with an anoxic bottom layer (Ludlam 1996*b*). Sampling in the present study was in the outer basin near Lake C1.

Char Lake is well known as a result of the International Biological Program study at this site in the late 1960s and 1970s (Schindler et al. 1974). It is a 28 m deep, freshwater, high arctic lake that is used as a drinking water supply for the hamlet of Resolute Bay. It is smaller than the C-series lakes and has an intermediate catchment-to-lake ratio (Table 1). It currently lies 34 m asl and is thought to have emerged from the sea about 6000 years ago (Schindler et al. 1974, and references therein; Michelutti et al. 2003). Sampling was in June before a moat formed around the ice.

Materials and methods

All the aquatic environments were covered by a thick layer of ice (1.0–2.6 m) at the time of our visits. Sampling was conducted from a mid-lake site through a 22 cm diameter hole bored through the ice with a battery-powered drill (Strikemaster). Ice thickness data for the lakes sampled in June 1999 and May 2000 are representative of the annual maximum. The thinner ice measured in late July and early August 2001 was representative of the ice remnant at the end of the melt season of that year.

Profiles for temperature, specific conductivity, pH, and dissolved oxygen were measured with a Surveyor 3 profiler (Hydrolab Corporation) that was lowered through the water column. Profiles of the beam attenuation coefficient at 660 nm (beam c(660)) were measured using a Sea Tech transmissometer (WETLabs, Inc., Philomath, Oregon, USA) measuring transmission of a collimated light beam over a 10 cm path length. Sampling depths were selected from the profile data and water samples for the different analyses were obtained from these depths with an opaque 2 L Kemmerer sampling bottle.

Water for chemical analysis was filtered through 0.45 μm Millipore membranes and subsequently analyzed for dissolved organic carbon (DOC), dissolved inorganic carbon (DIC), nitrate and nitrite (NO_3^- + NO_2^-), reactive Si (SiO_2), soluble reactive phosphorus (SRP), and major ions (Na^+, K^+, Mg^{2+}, Ca^{2+}, Ba^{2+}, Cl^- and SO_4^{2-}). Additional samples were stored cold and unfiltered, and subsequently analyzed for total nitrogen (TN), phosphorus (TP), iron (Fe) and manganese (Mn). Chemical analysis of these samples was conducted by the National Laboratory for Environmental Testing (Burlington, Ontario, Canada) according to their standard protocols, as described in Gibson et al. (2002).

To determine chlorophyll *a* concentrations (Chl *a*), samples from the oxic waters were filtered onto 25 mm diameter Amersham GF/F equivalent filters and stored frozen prior to analysis. Pigments were extracted using 8 mL of boiling ethanol (Nusch 1980). Fluorescence was measured in a Sequoia-Turner model 450 fluorometer before and after acidification with HCl, and concentrations of Chl *a* were calculated using the equations of Jeffrey and Welschmeyer (1997). Ethanol extracts of pigments from the anoxic waters of Lake A (June 1999) were analyzed by spectrophotometry and the absorption spectra were converted to bacteriochlorophyll *e* concentrations using the molar absorption of Borrego et al. (1999).

Water samples from selected depths in Lake C1 and Lake A (2001) were vacuum-filtered in duplicate through 25 mm GF/F filters and stored at –20 °C until measurement (three weeks after collection) of spectral absorption by particles (a_p, m^{-1}) The absorbance of particles concentrated onto the filters was measured over the spectral range 390–800 nm according to Roesler (1998) using a Hewlett-Packard 8452A diode array spectrophotometer equipped with an integrating sphere (Labsphere RSA-HP-84; Hewlett-Packard Corp.). Absorbance values were then converted to a_p using the algorithm of Roesler (1998).

Results

Physical properties

Temperature and conductivity profiles underscored the remarkable diversity of stratification regimes in the Canadian High Arctic (Fig. 2). A mid-water-column temperature maximum can be seen in all lakes. Maximum temperatures ranged from 0.38 °C in hypersaline Romulus Lake to 12.15 °C in Lake C1. Conductivities ranged from 0.18 mS cm^{-1} in the freshwater Lake C3 to 98 mS cm^{-1} at the bottom of hypersaline Romulus Lake. The steepness of the halocline varied between lakes, from gradual in Lakes A and C1, which had shallow mixed layers, to the steep gradient found in Romulus Lake. This likely reflects wind-induced mixing during the longer period of ice-free conditions that characterizes this latter lake. In a 1992 study of Romulus Lake, Davidge (1994) observed complete ice-out by July 28, and open water likely persisted for at least 4 weeks. The 1999 profile in Lake A closely resembled those obtained by earlier investigations (Ludlam 1996*a*). However, in the subsequent year, a change in the surface profile was observed, indicating a mixing event where freshwater mixed into the top of the salinity gradient, evening out the salt concentration in a layer from 3.5 to 10 m. In 1999 conductivity in this zone ranged from 0.27 mS cm^{-1} to 2.08 mS m^{-1}, while in 2001, this zone was isohaline with a conductivity of 0.96 mS cm^{-1}.

Differences in the stratification patterns in Lakes C1, C2, and C3 were observed, with evidence of mixing down to 7 m in Lake C1, 20 m in Lake C2, and to 35 m in Lake C3 (Fig. 2). Lake C3 is essentially a freshwater lake, with a dilute solute content and less than twofold difference in conductivity between the surface and bottom waters compared with the many orders-of-magnitude difference between the surface and deep waters of other meromictic lakes. This marked difference relative to adjacent Lakes C1 and C2 is most likely because of the much greater catchment-to-lake ratio, and hence water loading (Table 1), and the impact of the Taconite River that partially runs through the lake and contributes to water column mixing. It should be noted that the water load is also influenced by evapotranspiration, surface

Fig. 2. Conductivity–temperature profiles for the study sites. Conductivity is indicated by open triangles (note the different scales for Lake C3 and Romulus Lake). Temperature profiles are indicated by black circles (note the different scales for all of the environments).
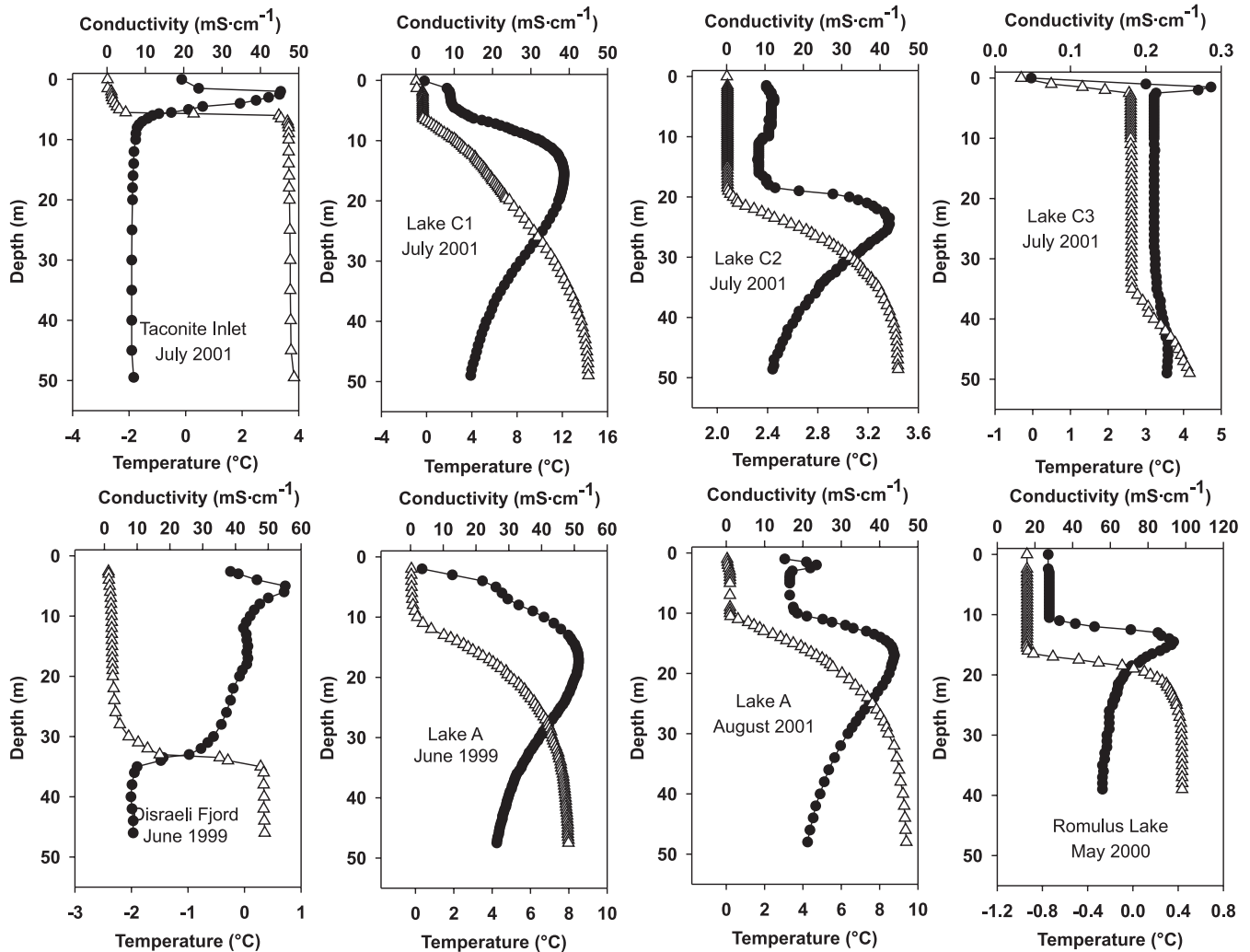
storage, and groundwater storage, and these factors are also likely to show variations among lakes.

Disraeli Fiord and Taconite Inlet showed stratification patterns caused by the thick ice dams near their seaward ends. Disraeli Fiord had a steep halocline at 33 m, the depth where seawater came in from under the thick ice shelf that blocked the outflow of the surface water (Vincent et al. 2001; Mueller et al. 2003). Taconite Inlet stratification was similar to that of Disraeli Fiord but with a shallower halocline (5.5 m) reflecting the thinner multi-year sea ice at its mouth.

Our measurements at Char Lake illustrate the strong limnological contrast between northern Ellesmere aquatic ecosystems and this classic high arctic lake. Char Lake (Fig. 3) had a well-mixed 28 m deep water column, with low conductivity, cold temperatures, high oxygen concentrations, and ultra-oligotrophic nutrient and chl *a* concentrations. Temperature, conductivity, pH and dissolved oxygen varied little through most of the water column, implying regular mixing during open-water conditions each year. Some variation was recorded immediately under the ice, with lower conductivity and higher DOC and DIC than the rest of the

water column, probably as a result of the early season inflow of meltwaters from the surrounding catchment.

Chemical properties

All of the Ellesmere Island lakes showed the typical meromictic profile with anoxic bottom waters. The oxycline was located at depths ranging from 13 m in Lake A to 47 m in Lake C3 (Fig. 4). The surface waters were well oxygenated and, in some cases, supersaturated. In Lake C1, >200% oxygen saturation over the depth interval 7–14 m persisted for two weeks in mid-July, most likely as the combined result of intense photosynthesis and stable stratification. The Mn-rich layer in Lake A (up to 9.7 and 8.2 mg L^{-1} in 1999 and 2001 respectively) extending from 10 to 29 m has been previously described (Gibson et al. 2002), and occurred in a region that was devoid of both O_2 and H_2S. Our profiles show that similar Mn-rich layers occurred in Lake C1 (3.4 mg L^{-1}), Lake C2 (11.4 mg L^{-1}), and Romulus Lake (6.7 mg L^{-1}). Extremely high concentrations of Fe occurred in anoxic water below the Mn maximum, at 50 m depth in Lake C2 (12.1 mg L^{-1}) and as a sharp peak at 30 m depth in both years of sampling at Lake A (up to 2.0 mg L^{-1}). In

Fig. 3. Water column profiles at Char Lake, June 1999. **(left)** conductivity (open triangles) and temperature (black circles); **(centre)** dissolved oxygen (black diamonds) and pH (black circles); **(right)** dissolved organic carbon (DOC, open diamonds) and dissolved inorganic carbon (DIC, open triangles).
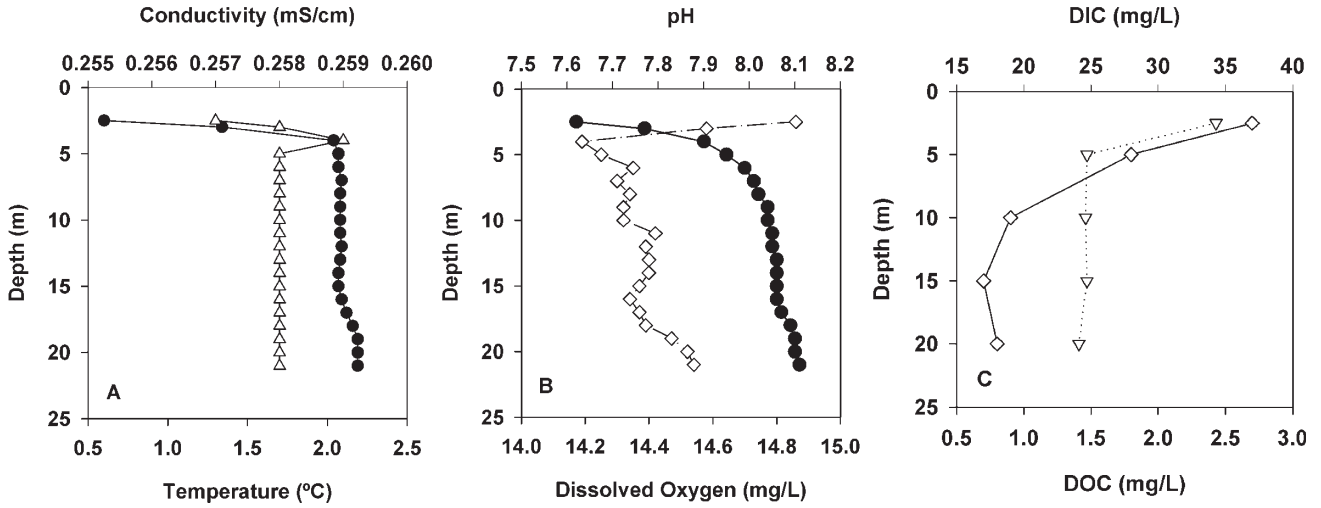
Fig. 4. Profiles of redox indicators. Dissolved oxygen profiles are shown by black circles; the profiles extend down to anoxic layer or end of cable. In Lake C1, the dissolved oxygen concentration exceeded the sensitivity of our sensor (>200% of saturation). Concentrations of total manganese (open diamonds) and total iron (open triangles) are also plotted. Note different scales for each site.
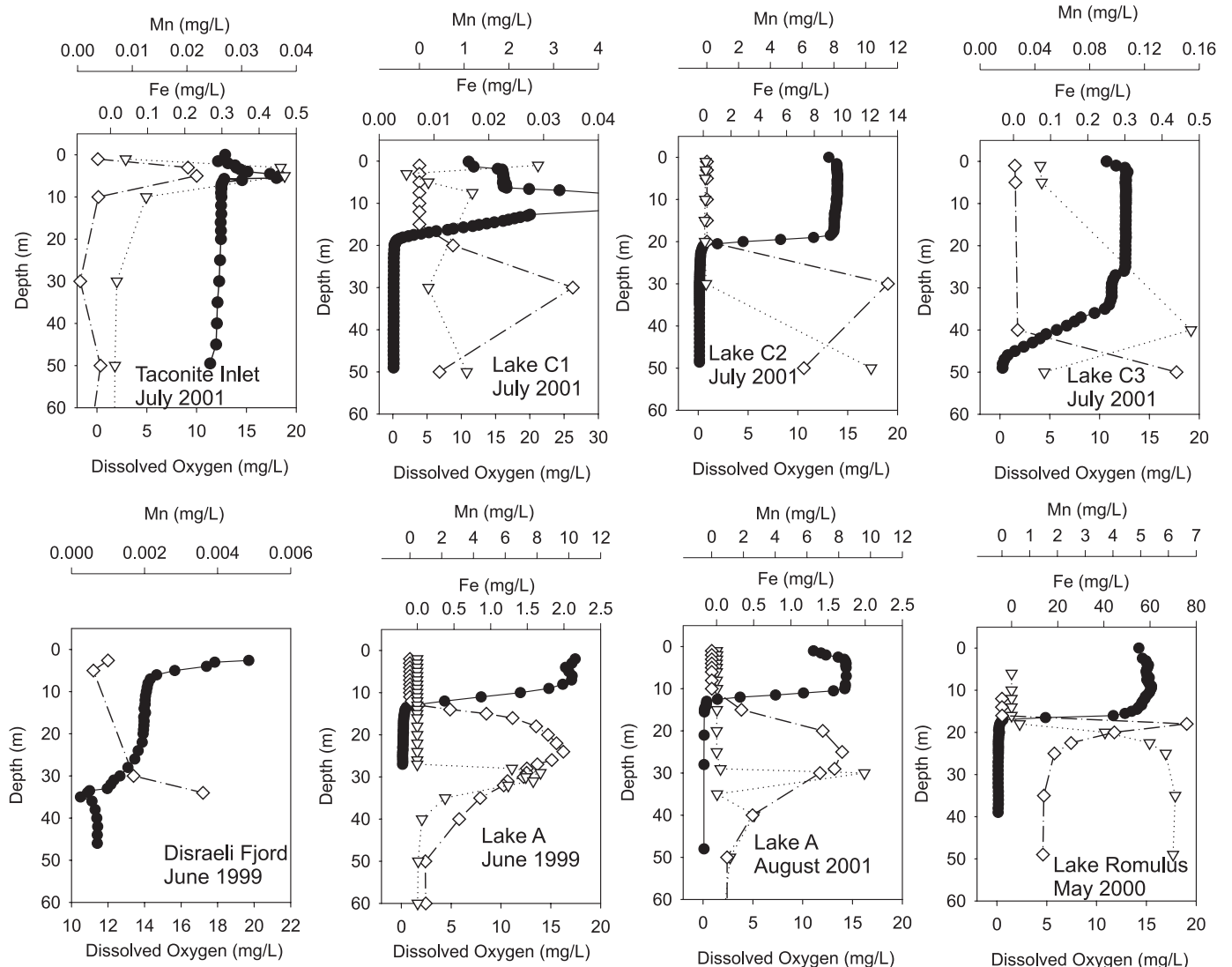
Fig. 5. Profiles for pH (black circles), dissolved organic carbon (DOC, open diamonds), and dissolved inorganic carbon (DIC, open triangles). pH values are given on the same scale for all environments, except Disraeli Fiord; DIC and DOC scales differ among sites.
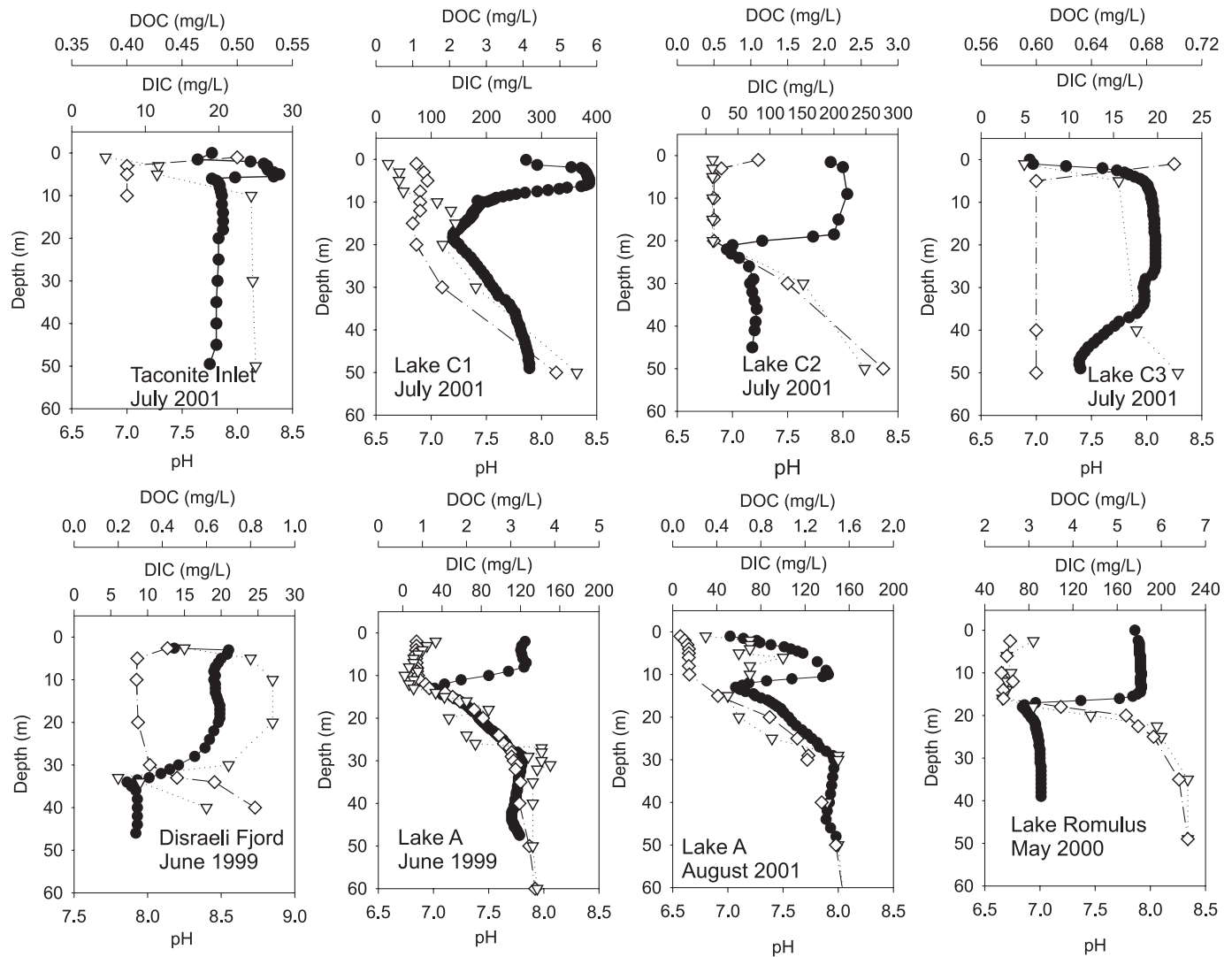
Lake Romulus, Mn peaked at the top of the anoxic monimolimnion, while Fe continued to rise to a plateau with maximum concentrations of 71 mg L^{-1} towards the bottom of the water column. The two fiord sites had oxygenated deep waters reflecting tidal exchange with the offshore marine environment, and low values of Mn (<0.005 mg L^{-1}) and Fe (<0.001 mg L^{-1}) typical of oxic waters.

Dissolved organic and inorganic carbon concentrations, as well as pH, were similarly characterized by strong vertical gradients (Fig. 5). In general, both DOC and DIC concentrations increased with depth. Unusually high DIC concentrations occurred in the hypolimnion of the saline lakes and ranged from 160 mg L^{-1} in Lake A to 364 mg L^{-1} in Lake C1. Lower DIC occurred in the freshwaters (22 mg L^{-1} in Lake C3) and in oxic fiords (typical seawater values of 25 mg L^{-1} in Taconite Inlet and Disraeli Fiord). DOC concentrations in the surface waters of the meromictic lakes varied over an order of magnitude, from 0.2 mg L^{-1} in Lake A in 2001 to 2.5 mg L^{-1} in Romulus Lake. DOC concentrations increased with depth in all lakes (except the weakly stratified Lake C3), and reached concentrations up to 6.5 mg L^{-1} in the

hypersaline bottom waters of Romulus Lake. DOC concentrations in Taconite Inlet and Disraeli Fiord showed less vertical variation, with values in the range 0.3–0.8 mg L^{-1}. There was a decrease in the DOC concentrations in the anoxic zone measured in Lake A in 1999 relative to 2001. This may be due to seasonal or inter-annual variations, but could also result from contamination or filtration artifacts.

Nutrients also varied markedly among environments and with depth (Table 2). Nutrient concentrations were low in the surface freshwater of the meromictic lakes, although higher than in ultra-oligotrophic Char Lake. Total nitrogen ranged from 0.029 mg L^{-1} at the bottom of Lake C3 to 25.4 mg L^{-1} at the bottom of Lake C1, with generally higher values in anoxic water likely reflecting high concentrations of ammonium. Total phosphorus ranged from the ultra-oligotrophic level of 0.004 mg L^{-1} in the surface of Disraeli Fiord up to a maximum of 4.02 mg L^{-1} in the bottom of Lake C1, again with markedly higher concentrations in the anoxic bottom waters.

The TN:TP ratio varied greatly from site to site, with extreme values ranging from 2.21 (by weight) in the bottom

Table 2. Concentrations of reactive nutrients in the high arctic aquatic environments.

Lake		Depth (m)	TN (mg L^{-1})	TP (mg L^{-1})	TN:TP ratio	SiO$_2$ (mg L^{-1})	NO$_3$-N (mg L^{-1})	SRP (mg L^{-1})
Lake A (1999)	Surface	2	0.118	0.005	23.6	0.64	0.022	0.009
	Deep	100	10.136	1.32	7.7	12.6	0.036	1.44
Disraeli Fiord	Surface	2.5	0.120	0.004	30.0	0.88	0.023	0.008
	Deep	40	0.240	0.041	5.9	0.51	0.014	0.036
Lake C1	Surface	3	0.191	0.0056	34.1	1.46	0.056	0.0025
	Deep	50	25.821	4.02	6.4	8.34	0.021	n.a.
Lake C2	Surface	3	0.087	0.01	7.70	0.86	<0.010	0.0056
	Deep	50	24.2	0.78	31.0	11.8	0.02	n.a.
Lake C3	Surface	5	0.054	0.0054	10.0	1.57	0.025	n.a.
	Deep	50	0.262	0.0102	26.7	4.08	0.214	0.0032
Taconite Inlet	Surface	3	0.086	0.0176	4.9	0.91	0.032	0.0027
	Deep	75	2.73	<0.1	27.3	2.54	0.19	n.a.
Romulus Lake	Surface	2.5	0.289	0.0108	26.8	0.45	0.04	0.0109
	Deep	49	17.21	0.0449	383.3	3.29	<0.010	0.066
Char Lake	Surface	2.5	0.295	0.008	n.a.	n.a.	<0.010	<0.002
	Deep	20	0.070	0.008	n.a.	n.a.	<0.010	<0.002

Note: TN, total nitrogen; TP, total phosphorus; SiO$_2$, silicate; NO$_3$-N, nitrate; SRP, soluble reactive phosphorus; n.a., not available.

waters of the outer basin of Taconite Inlet, indicating marine depletion of N relative to P, and up to 383 in the deep waters of Romulus Lake, likely influenced by the high ammonium concentrations at depth. Silica ranged from 0.45 mg L^{-1} in the surface waters of Romulus Lake to 12.6 mg L^{-1} in the bottom waters of Lake A. Nitrate-N concentrations ranged from 0.014 mg L^{-1} in the marine waters of Disraeli Fiord to 0.214 mg L^{-1} in the bottom waters of Lake C3, suggestive of strong or prolonged nitrification in the latter. Soluble reactive phosphorus concentrations ranged from 0.001 mg L^{-1} in Taconite Inlet to 1.44 mg L^{-1} in the anoxic bottom waters of Lake A. The latter was slightly above the TP value, likely reflecting analytical error in the two separate analyses but indicating that the phosphorus pool in the bottom waters of Lake A (and in anoxic waters of the other meromictic lakes) was dominated by dissolved reactive phosphorus.

The ionic composition of water samples from the different environments was compared with standard seawater to provide an index of the relative contribution of marine- and land-influenced processes (Table 3). The different enrichment ratios were calculated using the following equation:

$$[1] \qquad E_x = ([x]_s/[\text{Cl}^-]_s)/([x]_{sw}/[\text{Cl}^-]_{sw})$$

where $[x]$ is the concentration of the ion of interest in the sample (s) and standard seawater (sw) and $[\text{Cl}^-]$ is the concentration of chloride ions in the sample and in standard seawater, where it is the dominant ion. There were large differences in the enrichment ratio of the different ions, but many of the values for the deep waters were near 1.0 for the fiords, as expected given their connections with the sea, and also for Lakes A, C1, C2, and Romulus Lake, indicating their marine origins. Lake C3 was a notable exception, with enrichment of all ions relative to chloride, reflecting the strong terrestrial rather than marine influence on this lake and its present-day flushed, freshwater conditions. The surface

waters at most sites were enriched in calcium and barium indicating the terrigenous influence (Gibson et al. 2002), and there was a major decrease (>70%) in these enrichment ratios in Lake A between 1999 and 2003, consistent with an entrainment of deeper, marine-derived water into the upper water column. Sulfate enrichment ratios were below 1.0 in all the deep samples, with evidence of substantial depletion, probably by sulfate reducing bacteria, in Lake C1. The Fe and Mn enrichments were generally high at all depths, likely influenced by the inclusion of particulate forms in these analyses and indicative of a long-term terrigenous influence superimposed on the marine properties of the waters.

Biological properties

Maximum chlorophyll *a* concentrations ranged from 0.27 to 1.3 µg L^{-1} and were generally observed in the upper part of the oxygenated water column, close to the ice, where light availability for photosynthesis was maximal. These values are typical of ultra-oligotrophic to oligotrophic systems. In the anoxic waters of meromictic Lake A, Lake C1, and Romulus Lake, a community of H$_2$S-dependent photosynthetic bacteria was observed. The sulfur particles associated with these bacteria caused a turbid yellow colouration of the water and likely contributed to the marked peaks in beam attenuation (Figs. 6A, 6B; see also Belzile et al. 2001). The turbidity peak in Lake A also corresponded to the maximum in total Fe, suggesting that microbial production of precipitated iron compounds might additionally contribute to particulates at this depth. Particle absorption spectra of samples from the upper anoxic zones of Lakes A and C1 showed a peak at 715 nm indicative of bacteriochlorophyll *e*, a photosynthetic pigment that is characteristic of green sulfur bacteria (Figs. 6C, 6D). The presence of these green sulfur bacteria represents a striking difference with the autotrophic community of Char Lake and other Arctic freshwater lakes. Analysis of samples from Lake A (June 1999) gave bacterio-

Table 3. Enrichment ratios of major ions and indicator constituents (Ba, total Mn, and total Fe) compared with seawater in surface and sub-halocline waters.

Environment	Na$^+$	K$^+$	Mg^{2+}	Ca^{2+}	SO$_4^{2-}$	Ba^{2+}	Total Mn	Total Fe
Surface ratios								
Taconite Inlet	1.06	1.28	1.21	5.8	1.5	8.77	1974	2787
Lake C1	0.98	1.09	1.41	7.3	1.2	21.64	71.3	125.9
Lake C2	1.01	1.08	2.12	23.5	2.2	66.52	6100	25658
Lake C3	2.71	19.73	61.41	644.9	156.6	1910	209618	217597
Disraeli Fiord	0.98	1.15	1.06	2.3	1.0	n.a.	n.a.	n.a.
Lake A (1999)	1.07	1.45	2.48	24.6	1.9	112.9	837.8	n.a.
Lake A (2001)	0.98	1.25	1.39	5.0	1.1	42.8	217.5	235
Romulus Lake	1.02	0.96	0.97	1.6	0.9	1.36	5.8	5.6
Sub-halocline ratios								
Taconite Inlet	0.94	0.95	0.92	0.93	0.77	0.06	3.92	3.3
Lake C1	1.04	0.93	1.05	0.59	0.25	1.12	1396	5.8
Lake C2	0.99	0.89	1.02	1.03	0.87	1.66	18755	4369
Lake C3	1.43	4.92	14.89	240.41	28.75	695.57	3131102	39437
Disraeli Fiord	0.97	1.08	0.98	0.98	0.92	n.a.	n.a.	n.a.
Lake A (1999)	0.98	1	0.97	0.90	0.7	0.62	2677	2.3
Lake A (2001)	0.99	0.95	0.98	0.93	0.8	0.68	2787	12.8
Romulus Lake	0.9	0.79	0.95	0.92	0.85	0.33	1570	8739

Note: n.a., not available.

chlorophyll *e* concentrations up to 4.7 µg L^{-1} (at 30 m), nearly one order of magnitude higher than the Chl *a* concentrations in the oxic surface waters. A similar magnitude for bacteriochlorophyll *e* concentration in Lake C1 and Lake A (August 2001) relative to Chl *a* concentration in the oxic waters is suggested by the particle absorption curves (Figs. 6C, 6D).

Discussion

The limnological observations reported here underscore the diversity of physico-chemical conditions to be found in high arctic lakes and fiords. Conditions ranged from freshwater to hypersaline, and from continuously warm to near-freezing temperatures that persist even in summer. The variety of water column structures among these ecosystems (Fig. 2) is particularly striking. Despite their fundamental differences, however, the environments sampled in the present study share a number of common features. They all contain low Chl *a* concentrations, placing them in the oligotrophic to ultra-oligotrophic range. Their high-latitude position means that they are all covered by a thick layer of ice for most of the year and experience the strong seasonal light–dark contrast of polar environments. Many of the waters are highly stratified because of isolation from the wind by ice cover, and as a result of strong salinity gradients that resist mixing. The four meromictic systems are characterized by anoxic and sulphidic bottom waters that have major ion ratios indicative of a seawater origin, and that contain high concentrations of phosphorus (mostly SRP), nitrogen (likely to be mostly ammonium), Fe, and Mn. The meromictic lakes also all contain photosynthetic green sulfur bacteria in their upper anoxic zones.

The mid-depth thermal maxima in the meromictic lakes show similarities in position and magnitude to those in Ant-arctic meromictic lakes (Spigel and Priscu 1998; Gibson 1999). The diversity of thermal regimes of the high arctic lakes is similar to the high variability in Antarctica, for example, in the McMurdo Dry Valleys where water column maxima range from 2 to 25 °C. Energy balance calculations show that these unusual temperatures result from solar energy that penetrates the ice and water column. The heat is gradually accumulated over hundreds of years to millennia in the lower water columns of the lakes that are stabilized by salt gradients (Spigel and Priscu 1998). The observed divergence in thermal characteristics between individual lakes is likely to arise from differences in snow and ice cover (which affect the transmission of solar radiation), attenuation properties of the water column (which affect depths of maximum absorption of solar radiation; Spigel and Priscu 1998), the surrounding topography (which may reduce solar input through shading), and water column history (Gibson 1999).

The chemical characteristics of the Ellesmere Island lakes also find parallel in Antarctica. Hypersaline bottom waters are found in both polar regions, for example up to three times that of seawater in Lake Vanda in the McMurdo Dry Valleys. However, the Dry Valley lakes have a more complex biogeochemical history (Lyons et al. 1997), and ionic ratios deviate substantially from seawater. Total Fe and Mn concentrations are also high in Dry Valley lakes (up to 3.4 mg Mn L^{-1} and to 1.1 mg Fe L^{-1} in Lake Vanda; Green et al. 1989) although below the extreme values recorded here in Ellesmere lakes, likely reflecting differences in catchment geochemistry. Sediment laden underflows are known for Lake C2 (Retelle and Child 1996), and may contribute to the deep delivery of Mn- and Fe-rich particles.

The ice-dependent ecosystems of northern Ellesmere Island are likely to be highly responsive to small shifts in climate once their ice-covers warm to near 0 °C. Some evidence of this sensitivity can be seen by comparing Lake A, which has

Fig. 6. Beam attenuation ($c660$), temperature profiles and particle absorption (a_p) spectra at selected depths in Lake C1 on 19 June 2001 (panels A and C), and Lake A on 3 August 2001 (panels B and D). The numbers in panels C and D refer to the sample depths in metres. The dashed line marks the depth limit of the oxic zone.
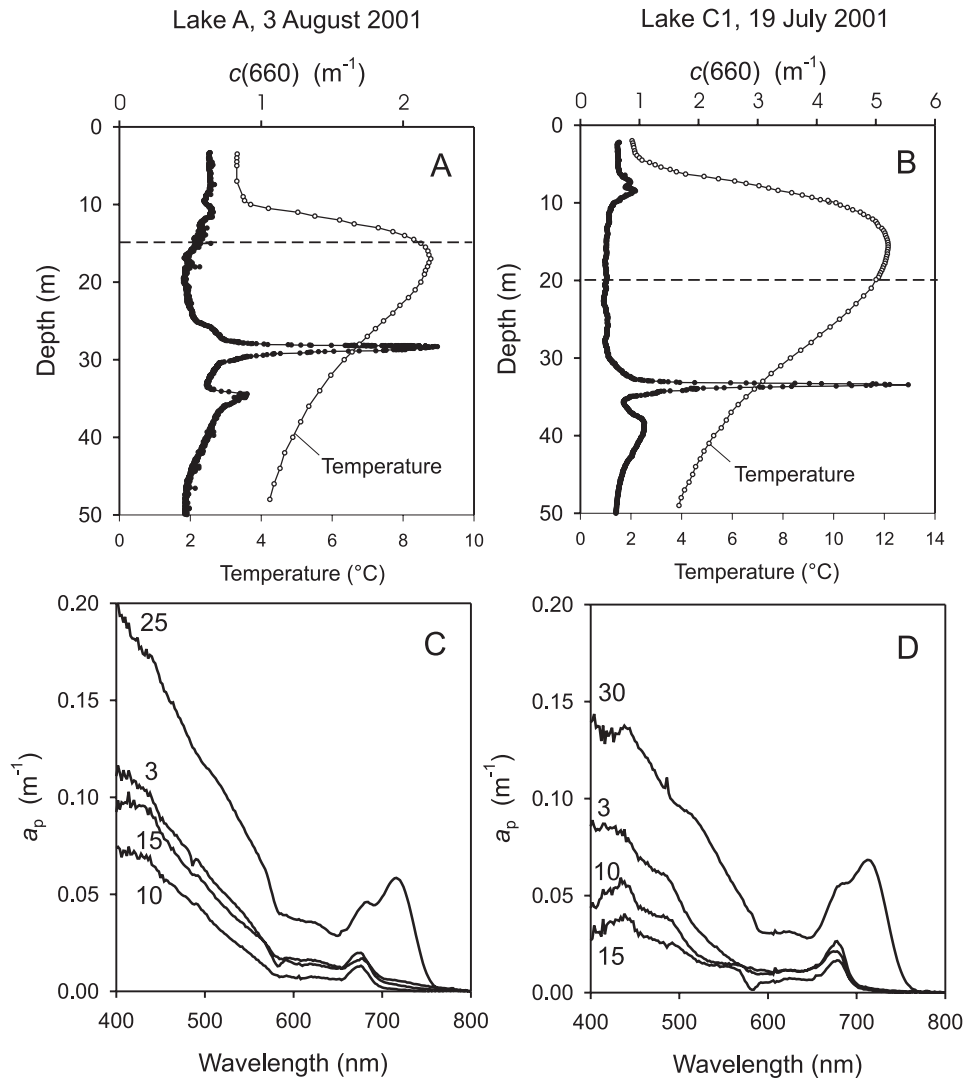

a more persistent ice cover, and Romulus Lake, situated three degrees of latitude to its south, where the ice cover is lost for several weeks each year. Despite its more northerly location, Lake A exhibits a far higher maximum water column temperature (8.78 °C) than Lake Romulus (0.36 °C). Lakes C1, C2 and C3 are adjacent lakes at the same latitude, yet have strikingly different thermal maxima (12.15, 2.45, 3.59 °C), reflecting the importance of not only persistent ice-cover, but also the strength and position down the water column of salinity gradients that promote stable stratification and that allow the long-term trapping and storage of heat.

The marine-derived lakes of Arctic Canada can be placed within an evolutionary sequence of landscape change, with individual lakes representing different steps in the transition from marine to freshwater or hypersaline ecosystems (Fig. 7). All lakes were originally marine basins with complete tidal connection to the open ocean. As sea level fell at the end of the last ice age because of reduced ice loading on the land (isostatic rebound), pockets of seawater would have been progressively cut off, eventually resulting in seasonal

and then complete isolation of the lakes. Lakes which still receive occasional marine input in Arctic Canada include lakes Ogac (fed by spring tides; McLaren 1967), Qasigialiminiq, and Tariujarusiq, all on Baffin Island (Hardie 2004). There are many examples elsewhere including in Scandinavia (Ström and Klaveness 2003) and Antarctica (Gibson 1999). These lakes (lagoons) are typically stratified, with sea water at the bottom of the water column and fresh water at the surface. Eventually, such lakes become completely isolated from the ocean but may retain some seawater in their basin; e.g., Lake Romulus (present study and Davidge 1994), Sophia Lake (Ouellet et al. 1987), and Garrow Lake (Ouellet et al. 1989; Markager et al. 1999).

A specialized route for lake formation is illustrated by Disraeli Fiord and Taconite Inlet, at the highest latitudes of the Canadian Arctic. Build-up of thick, landfast ice at the mouths of the long fiords of northern Ellesmere Island results in dams that trap the inflowing meltwaters behind them, with the depth of the freshwater layer constrained by the thickness of the ice shelf (Vincent et al. 2001). The less

Fig. 7. Postulated evolutionary sequence for coastal, high-latitude landscapes, embayments and lakes.
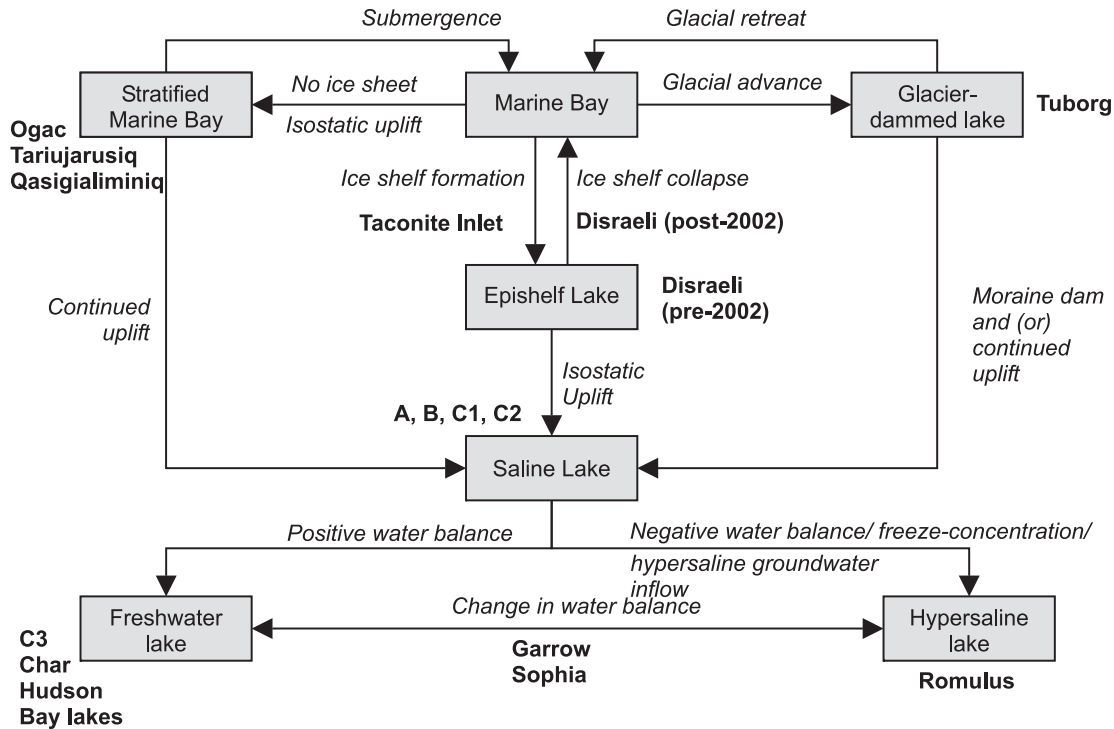
dense freshwater floats on the denser seawater, and these epishelf lakes are tidal. The salinity profile of Taconite Inlet, with its 6 m thick freshwater layer, indicates that the minimum thickness of the ice sheet damming the lake is only ca. 6 m. In 1999 Disraeli Fiord presented a more mature system, with depth of the freshwater layer over 30 m. Epishelf lakes also occur in Antarctica (Gibson and Andersen 2002 and references therein), where the freshwater can be up to 100 m thick. This ecosystem type is entirely dependent on the integrity of the ice dam, as illustrated by Disraeli Fiord in 2002, when a major crack formed across the Ward Hunt Ice Shelf and resulted in almost complete drainage of the surface freshwaters (Mueller et al. 2003). The epishelf lake structure was completely lost at this time and the basin reverted to a marine inlet that is at present fully connected to the Arctic Ocean. The epishelf systems contrast with Lake Tuborg, also an ice-dammed saline lake in the Canadian High Arctic, but formed by a glacial surge that trapped seawater at the landward end of a marine embayment. The glacier holding back Lake Tuborg is grounded throughout, and therefore the lake is not tidal. It is meromictic, with a layer of freshwater overlying marine-derived saltwater (Smith et al. 2004).

With ongoing uplift and decreasing sea level, the tidal waters of epishelf lakes can be cut off to form meromictic lakes. Lake A is believed to have been isolated from a coastal arm of Disraeli Fiord epishelf lake 2500–4000 years ago (see Jeffries and Krouse 1985). Gibson et al. (2002) provided geochemical evidence that significant dilution of seawater in the lake basin occurred when the lake was still tidal prior to final separation. Lakes C1, C2, and C3 may have had a similar origin, being isolated from an earlier epishelf lake in Taconite Inlet.

The ecosystems sampled in this study illustrate that once

isolated from the ocean, a saline lake can follow divergent pathways (Fig. 7). If significant freshwater flow occurs through the lake, all salt can be flushed out resulting in a freshwater lake. Examples of such lakes in the Canadian High Arctic are Char Lake, Lake C3, and Stanwell-Fletcher Lake on Somerset Island (Rust and Coakley 1970). In these cases, it would be expected that the lakes would have a large drainage basin to lake area ratio and water loading. Table 2 shows this to be the case, particularly for Lake C3, which has a potential annual water loading that amounts to 25% of the maximum depth of the lake. Flushing of salt from wind-mixed water columns will be more rapid in areas of higher precipitation or if there is extensive ice or snowfields in the drainage basin that provide a regular source of water. The transition from saline to freshwater conditions is also well known from coastal areas elsewhere and has been the object of detailed paleolimnological studies at many locations, including subarctic Canada (Saulnier-Talbot et al. 2003).

An alternative development route occurs in basins with negative water balance; i.e., in which the input through precipitation and inflow is less than the loss due to evaporation and outflow. In these cases, the initially marine lakes will become hypersaline (Fig. 7). Lake Romulus in our data set provides a compelling example. If the process continues long enough a balance may be reached between evaporation (which decreases with increasing salinity) and input, such as that found for Deep Lake in Antarctica (Ferris and Burton 1988). Freeze-concentration processes both in the catchment (Ouellet et al. 1987) and within the lake (Davidge 1994) are also likely to play a role in the development of hypersaline conditions. Under changing hydrologic conditions, a hypersaline lake could lose its salt though flushing and evolve into a freshwater lake, although no such examples are known from the Arctic.

Lakes that are neither fresh nor hypersaline can exist in a complex field in which changes in climate can have significant affects on the lake. For example, Lake A retains near-seawater salinity in its monimolimnion underneath a ca. 10 m layer of fresher water. The ice on Lake A appears to be going through a climate-induced transition from perennial to seasonal ice cover. The permanent ice cover precluded wind mixing in the lake and lessened the intensity of ice formation. We predict that the occurrence of a period of open water every year will lead to an increase in salinity in the surface water due to entrainment of deeper, more saline water through wind mixing, and that the annual ice formation will result in the production of more saline bottom water through the formation of brines at the lake margins (Ferris et al. 1991; Davidge 1994). Therefore, the bottom waters of the lake may become hypersaline, as in Romulus Lake (Davidge 1994). A similar fate may await Lakes C1 and C2, and it possible that Lakes Sophia and Garrow may have already moved some way down this path. These latter lakes lose their ice cover for several weeks each summer but contain hypersaline water at depth that may be due in part to surface ice formation, but also to the freeze-out of salt from a saline aquifer (talik) under the lake (Ouellet et al. 1987, 1989).

The Canadian High Arctic contains a set of present-day aquatic ecosystems that represent each step of the inferred landscape–lake chronosequence. An analogous set of landscape evolutionary processes leading to different lake types has been described to some extent in Antarctica, specifically for an Antarctic fiord situated in the Vestfold Hills, an unglacierized coastal area near the Australian Davis Station (Gallagher et al. 1989). Ellis Fiord contains a stratified basin, with a hypersaline layer at its bottom. Gallagher et al. (1989) examined the physico-chemical processes and concluded that the formation of the hypersaline layer depended on salt freeze-out during the formation of ice cover. Paleolimnological studies in the Vestfold Hills area have provided a detailed record of environmental change in Antarctic meromictic lakes, and similar studies are required for the Ellesmere Island lakes and fiords, building on the studies at Taconite Inlet (Bradley et al. 1996) and Tuborg Lake (Smith et al. 2004).

The Canadian High Arctic is currently experiencing changes in climate that seem to be unprecedented relative to the last few millennia (Douglas et al. 1994; Smol et al. 2005), and these have begun to have a major impact on the aquatic ecosystems of Ellesmere Island (Mueller et al. 2003; Antoniades et al. 2005). The predicted continuation and amplification of these effects (ACIA 2004) are likely to cause major limnological shifts in the high arctic systems described here, and to accelerate their evolution towards the end points as shown in Fig. 7. Lake C3 is likely to continue to freshen given its large water load and minimal stratification. Lakes A, C1, and C2 have relatively small catchments and deep water columns and are more likely to move towards conditions seen in Romulus Lake, with increased open-water conditions favoring entrainment of saline waters into the surface layer and the production of hypersaline brines by freeze-up. The evolution of all these lakes will also depend on the effect of climate change on precipitation. Global circulation model predictions for this region of Arctic Canada out to 2050 indicate that warming will be accompanied by an increase in summer precipitation in the form of rain and a shift in the

hydrologic regime from nival (discharge controlled largely by snowmelt) to pluvial (discharge controlled by rainfall) (ACIA 2004). The net effect on lake evolution, beside a general increase of flushing rate by increased precipitation, will depend on the timing of snow melt and rainfall. Other climate-related changes forecasted in the Canadian High Arctic include the thermal degradation of permafrost and the increase in the active layer depth, which will affect water storage in soils and have direct impacts on landscape stability and runoff, in turn influencing lake evolution (Quesada et al. 2006). Break-up of the Ward Hunt Ice Shelf has already led to the draining of the Disraeli Fiord epishelf lake, and it has reverted to a marine embayment. A similar fate probably awaits the incipient epishelf lake in Taconite Inlet. Finally, isostatic rebound is likely to continue independent of modern climate change, leading to the eventual isolation of the Baffin Island lagoons that are tenuously connected to the ocean, and perhaps initiating the coupled landscape–lake evolutionary process in other marine bays that become increasingly isolated from the sea.

Acknowledgments

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