

OBSERVATIONS ON THE ROLES OF SNOW AND ICE ON THE CHEMISTRY OF AN ARCTIC LAKE

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ABSTRACT

The spring snowmelt period (May-June), 1983-85, was studied for a naturally acidified (pH 3.8), high arctic lake, Colour Lake, Axel Heiberg Island (79°N). Snow meltwater was found to be significantly more dilute and alkaline than late winter Colour Lake water. The chemical differences between the two water masses were used to trace meltwater pathways through the Colour Lake system and determine the influence of snow and ice meltwater runoff on the lake each spring. It was found that a substantial amount of the meltwater each spring did not influence the existing lake water at all and was discharged directly to the lake outflows. Meltwater either flowed over the ice surface through a littoral moat or chemical and thermal density differences allowed the incoming meltwater to sheet over the resident lake water and be discharged via the outflow. Colour Lake, below the ice, was not diluted from the addition of snowmelt water but did become distinctly less acidic (32%), from convective circulation of meltwater, drainage of meltwater from the ice surface and possibly from particulate buffering during sedimentation.

Introduction

The annual cycles of concentration of all elements in arctic lakes are largely controlled by freeze-thaw cycles. The electrolyte concentrations of high arctic lakes are generally higher than their inflows, and while ice-covered, than their outflows. Schindler et al (1974), postulated that the preferred loss of dilute, low density snow and ice meltwater during the spring runoff, followed by inputs of more concentrated runoff water and evaporation during the open water season, would lead to a concentrating of arctic lakes over their inputs until some equilibrium level was reached. Bergmann and Welch (1985) found that the Theoretical Flushing Time of a lake:

$$FT(\text{yrs}) = \text{lake volume } m^3 / \text{lake outflow } m^3$$

(Hutchinson 1975), which assumes complete mixing of all inputs, underestimated the real flushing time of a tritiated lake (P&N Lake, Saqvaquac, N.W.T., 60°39'N) by as much as 2X depending upon the timing and magnitude of the yearly runoff occurring during the ice covered period. The segregation of meltwater from lakewater by density differences has important implications for aquatic geochemical and nutrient cycling processes, particularly in lakes with thick, persistent ice covers and in areas where snowmelt is the single dominant hydrologic event. Colour Lake, Axel Heiberg Island (78°45'N 90°35'W, 176m a.s.l.), is highly acidic (pH 3.8), and of high ionic strength (specific conductance 550 uS @ 25°C), while its major snowmelt inflows are relatively dilute (specific conductance 65-150 uS @ 25°C) and moderately alkaline (pH 6.5, TIP alkalinity up to 286 ueq/l), (Fig. 1). Colour Lake was studied intensively during the spring melt

periods (May-June) of 1983-85 to determine how it could maintain its ionic state in relation to its snowmelt inputs. The extreme chemical differences between the two water masses allowed us to trace the meltwater pathways through the Colour Lake system and measure the change in ionic state of the lake as a result of the spring snowmelt.

The most distinctive feature of Colour Lake is that it is a naturally acidified lake occupying a sedimentary basin whose geology indicates a substantial buffering capacity. The value of studying such a system comes from the marked chemical differences between runoff and lake water which allow detailed examination of mixing processes which are typical of headwater arctic lake systems.

Study Area

The following is largely taken from Caflish (1970 and 1979). Colour Lake is situated in the west-central region of Axel Heiberg Island (Fig. 2). The 80 ha. catchment (Fig. 3) is not glacierized today but was probably occupied by a side arm of the White Glacier during the last major glacial advance in the area (ca. 10,000 B.P.). The relatively deep basin ($Z_{max} = 24m$ and 20m below the final outlet of the catchment) suggests that the basin was formed by a combination of glacial scour and dead ice melt (Fig. 4).

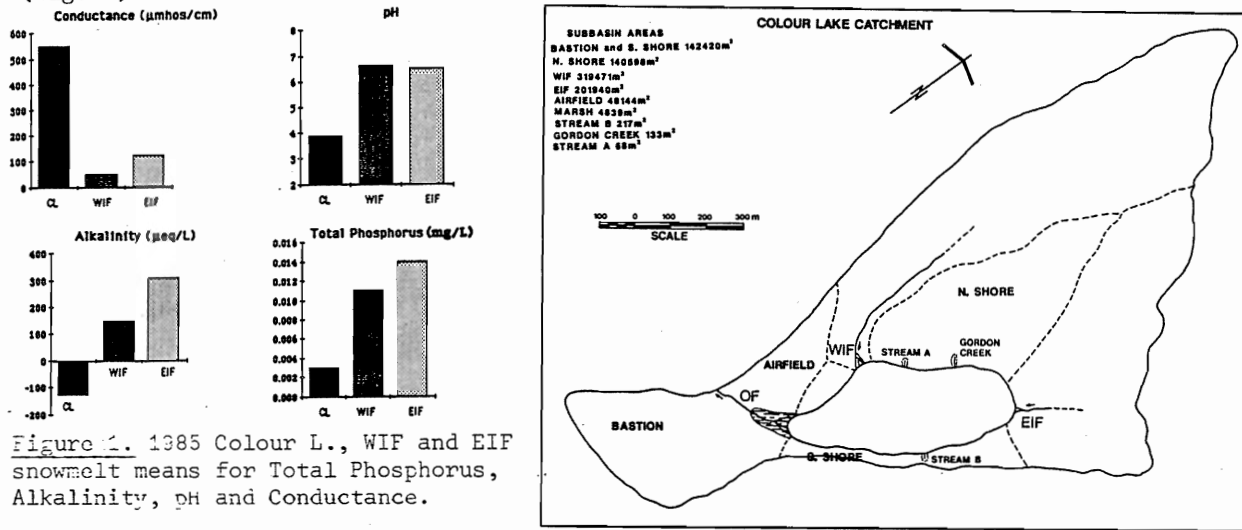


Figure 1. 1985 Colour L., WIF and EIF snowmelt means for Total Phosphorus, Alkalinity, pH and Conductance.

Figure 3. (above) Colour Lake catchment and subbasin areas.

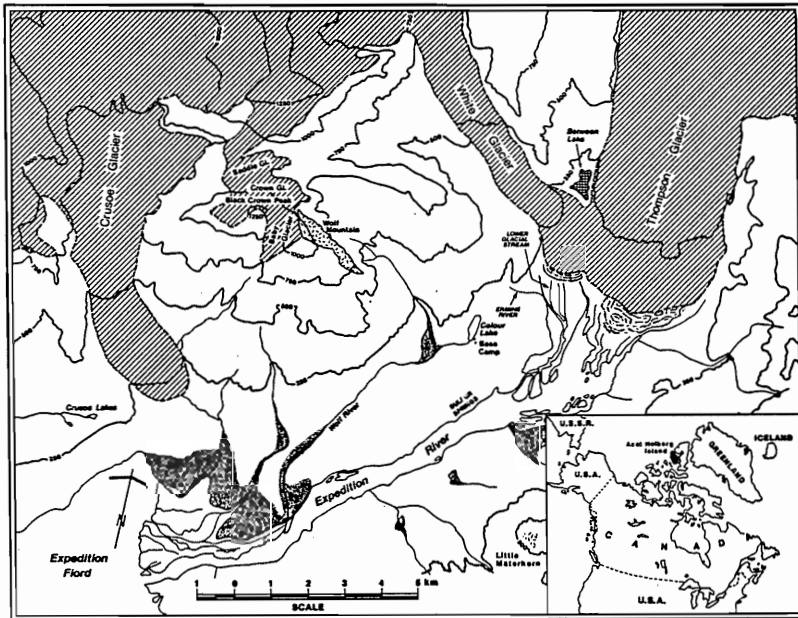


Figure 2. (left) Expedition Fiord area west-central Axel Heiberg Island, N.W.T., Canada.

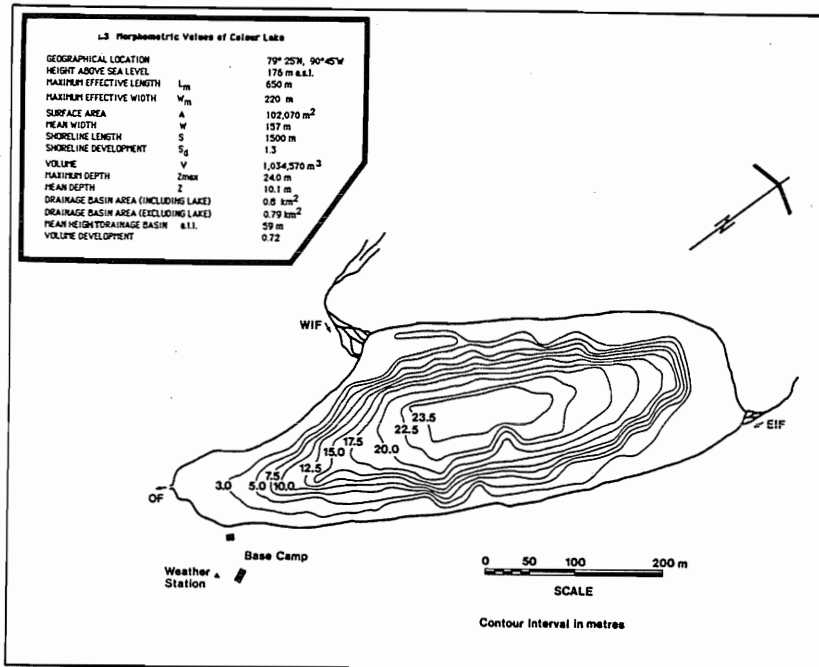


Figure 4. Bathymetric map and morphometric values of Colour Lake.

The Colour Lake catchment is located in the axial portion of the Sverdrup Basin, where the total accumulated thickness of strata exceeds 8,000m. With the exception of Upper Paleozoic anhydrite evaporites, the strata consist mainly of clastic sediments (sandstones, siltstones and shales with minor intercalations of carbonitic rocks, conglomerates and coal), which range in age from Triassic to Early Tertiary. Tectonic movements in the Early Tertiary resulted in strong folding and faulting of the area, and were accompanied by intrusions of Upper Paleozoic evaporites (Fricker 1963a and 1963b and Kranck 1963). Sulphur-salt springs are found to the south of the catchment on the shores of the Expedition River (Fig. 2), but are believed not to influence Colour Lake itself (Beschel 1963, C. Allan and M. English unpublished data).

The climatology of the Expedition Fiord area has been reported in Müller (1963), Müller and Roskin-Sharling (1967) and Ohmura (1981). The mean annual temperature for nearby Eureka (80°N 85°W) is -19.4°C with a mean wind speed of 11.6 kph and a mean total precipitation of 58.4mm of which 22.4mm falls as rain. Eureka's weather has been found to give a reasonably close approximation of that in Expedition Fiord (Blatter 1985).

Permafrost underlies the entire basin to a depth of 600m⁺ with an active layer of approximately 60cm. Plant cover in the catchment ranges from <1% on the steep talus slopes of the north and south sides of the basin to 90% in the flat eastern and western portions of the catchment. Dominant plants include Kobresia myosuriodes, Dryas integrifolia, Saxex arctica and Carex nardina.

Caflish (1970) describes the thermal regime of Colour Lake as having a permanent inverse stratification, or isothermy with one period of circulation due to the short ice-free period, (cold monomictic, Hutchinson 1975). Mean ice thicknesses range from 1.5 to 2.1m and ice is present for 9 to 12 months of the year.

Inflows are generally small and diffuse with numerous small rivulets carrying water over frozen or partially thawed ground during spring runoff. As the active layer deepens, substantial subsurface runoff occurs, particularly on the steep north and south shores of the lake. Occasional July or August rainstorms can lead to intense fluvial erosion as a result of the steepness, high frost table, scanty vegetation and large shale areas with low permeability (Rudberg 1963). The lake outflows are broad and shallow, passing through a marshy area before exiting the catchment through a gap in a basaltic dyke and flowing westward to Expedition Fiord (Figs. 2 and 3).

Methods

Hydrologic

In order to define the duration and timing of the spring runoff and the response of the lake to these inputs several of the inflows and the two outflows of Colour Lake were gauged. The shallow diffuse nature of the inflow and outflow streams and the absence of gauging structures made instantaneous discharge measurements difficult and continuous flow measurements impossible. The larger streams (EIF, WIF and the two outflow channels) were sectioned and the cross sectional area of flow, determined. Velocities were determined by timing dye or clay particle spikes over a measured distance. The area and velocity of each section were multiplied and summed and an instantaneous discharge calculated. In the smaller stream channels (GC and StA), volumetric discharges were obtained by pressing a plastic bag into the stream channel for a measured time period and then determining the volume of water with a graduated cylinder. During the snowmelt period, three or more discharge measurements were made daily. Linear interpolation was used to fill in periods between measurements.

Lake levels for 1984 were determined with a level and stadia rod placed at the lake water surface through a hole drilled through the ice cover and related to a bench mark with an assumed elevation of 100m. 1985 lake levels were recorded continuously with a Stevens F recorder and related to the same bench mark.

Ice thicknesses were determined by drilling a hole through the ice cover and hooking a graduated pole to the underside of the ice and reading its thickness (Adams and Prowse 1978). A 50m grid was established on the ice surface and ice thicknesses were measured at each point to obtain a mean thickness for the lake. Ice densities of .855 for white ice and .917 for black ice were used in the mass calculations (Adams 1981).

Water Sampling and Analyses

Temperature and conductivity profiles were determined in situ at one metre intervals with a YSI Conductivity/Temperature meter. Temperatures were checked against a mercury thermometer and measured conductivity was adjusted to 25°C with the following equation:

$$\text{measured conductance}/1-(.022(25-\text{measured } T^{\circ}\text{C}))$$

(Edwards et al 1975).

Lake water chemistry and oxygen samples were collected in acid-washed bottles with a peristaltic pump. Oxygen concentrations were determined on site by a modified Winkler method (APHA 1965). In 1983 all chemical parameters were measured at the Ontario Ministry of Environment (O.M.E.) laboratory, Rexdale, Ontario, following Outlines of Analytical Methods (O.M.E. 1981). In 1984 all chemical parameters were again measured at the O.M.E. laboratory except pH which was measured on site with an Orion Mk III digital pH meter standardized to buffers of pH 4 and pH 7. In 1985 Total Unfiltered and Filtered Phosphorus and Total Kjeldahl Nitrogen were again measured at the O.M.E. laboratory. In 1985 major cations (Ca, Mg, Na and K) were determined by atomic absorption at Trent University (Stainton et al 1977). In 1985 anion analyses (SO₄, Cl, F and NO₃) were performed at the Canada Centre for Inland Waters, Burlington, Ontario by ion chromatography (Small et al 1975). 1985 pH measurements were performed on site with a Fisher Acumet Model 602 digital pH meter standardized to buffers of pH 3 and pH 7. Particulate and dissolved phosphorus fractions were determined by subtracting total phosphorus values of filtered subsamples from total phosphorus values of unfiltered subsamples. Filtered subsamples were passed through a Schlecher and Schnell Antlia filter system utilizing a prerinsed 1.2 um Whatman GFC glass fibre filter. Inflow and outflow samples were collected in acid-washed bottles and treated in the same manner as the lake water samples except for conductivity which was measured on site with a Barnstead Hind conductivity bridge corrected to 25°C.

Ice samples were collected with a SIPRE corer in 1983 and 1984 and with an ice auger in 1985. Samples were melted and treated in the same manner as the other water samples.

Chemical mass balances in the lake were calculated as follows. For the unfrozen water column, the concentration at the midpoint of each metre depth was computed by interpolation or extrapolation from the two known concentrations, multiplied by the lake volume at each metre interval, summed and divided by the total volume to give a mean concentration per litre. The chemical content of the lake ice and its change in volume along with lake level changes were taken into account when calculating the whole lake mass changes. Because of the variety of analytical facilities used over the three field seasons, between year differences should be viewed with caution, but within year results should be valid.

Results

Physical

Mean ice thickness ranged from 1.5m to 2.0m during the three years of measurement. Ice occupied $133,673\text{m}^3$ to $210,461\text{m}^3$ or 15% to 20% of the lake volume. With a latent heat of fusion 79.7 cal/g this represents $9,600\text{--}15,066\text{ cal/cm}^2$. As shown by Schindler et al (1974), the latent heat of melting rather than heating is the most important term in the arctic lake heat budget.

The temperature profiles in both 1984 and 1985 (Figs. 5a and 5b) indicate late winter bottom temperatures in excess of 4°C , probably as a result of residual summer heat flux from the sediments resulting in the downslope movement of near-sediment water (Mortimer and Mackereth, 1958 and Welch, 1985). It is evident that during the spring runoff there is a general heating throughout the water column indicating a convective circulation during the spring melt, as molecular diffusion alone cannot account for the heat gain in the lower water strata. In 1984 the bottom strata actually lost heat from the late winter condition (Fig. 5a) as cooler surface water circulated downwards. This feature was not noted in 1985 (Fig. 5b), perhaps because of the later sampling date, circulation of this layer may have already occurred.

Figure 5a. 1984 Colour Lake Temperature Profiles

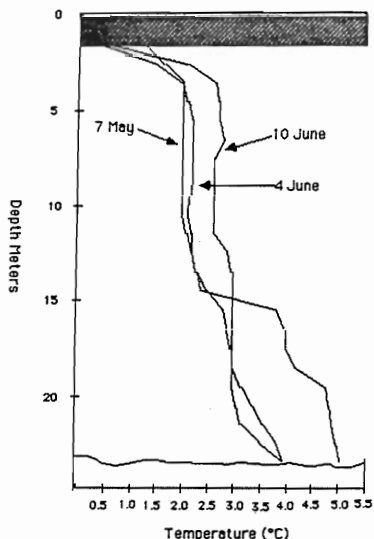
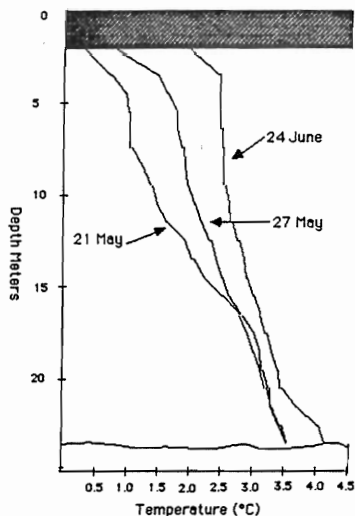


Figure 5b. 1985 Colour Lake Temperature Profile



The 24 hour continuous input of solar radiation in late May and early June leads to a quick snowmelt. Streams show a marked daily cycle as a result of the high relief in the basin shading different portions of it throughout the day. The highest gauged inflow discharge during the 1985 spring melt was only 30 l/sec , in the West Inflow. The low potential energy of the inflow streams precludes any large scale turbulent mixing by streamwater in this system.

In early June, runoff floods onto the ice surface, covering it to a depth of several centimetres. During this time, if the ice surface is breached, either naturally

by wind-deposited sediments melting through the ice cover, or by auger holes, a large portion of this above-ice water can enter the lake in a vertical, turbulent vortex (Fig. 6). The vortex shown transferred cold, dilute, above-ice water to a depth of 5.5m. Clearly, this phenomenon can cause significant perturbations in the local water column. It is likely however that this water rises back to the under-ice surface because of its lower density in relation to the resident lake water. Two days after the pictured vortex ceased, no perturbations of the water column (using conductance data), could be detected except for a thin (70cm) layer adjacent to the underside of the ice (Fig. 7a and 7b). The mechanisms which drive these vortices are a combination of an initial negative hydrostatic water level and the buoyant return of the ice sheet which is depressed by terrestrial runoff water.

Figure 6. WATERHOLE VORTEX CONDUCTANCE AND TEMPERATURE PROFILE MAY 27, 1985

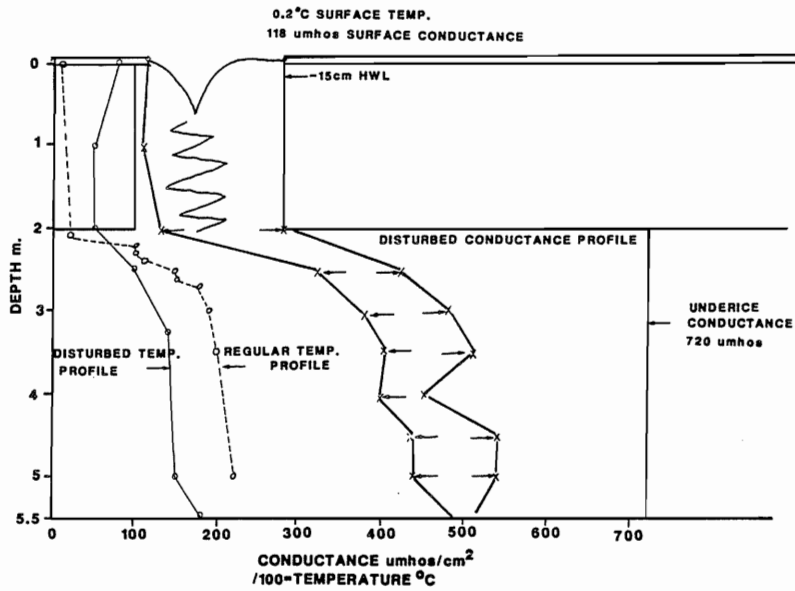


Figure 7a.

1984 COLOUR LAKE CONDUCTANCE PROFILES

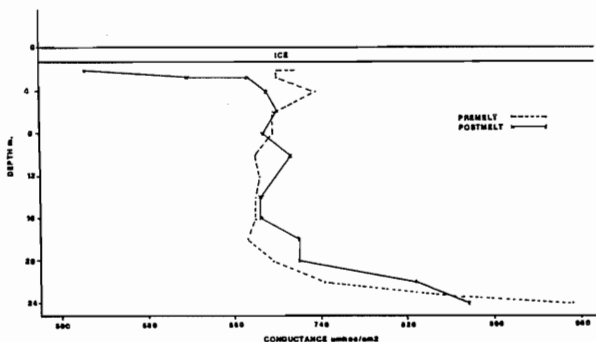
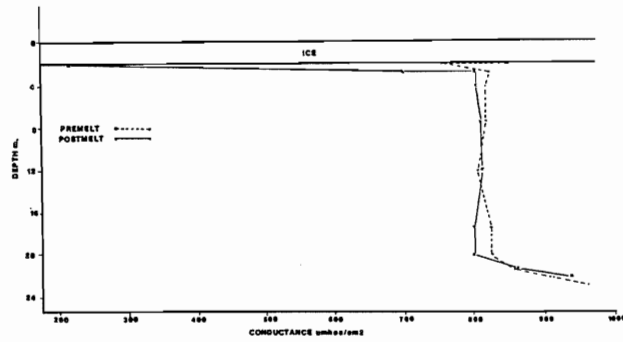


Figure 7b.

1988 COLOUR LAKE CONDUCTANCE PROFILES

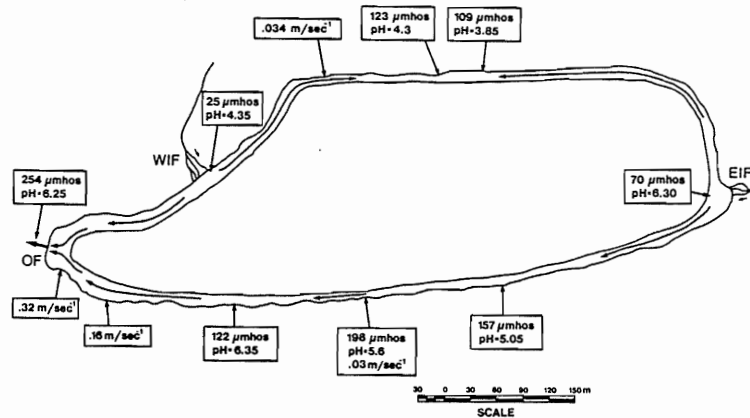


As the ice cover warms to ca. -1°C , microfissures will evolve at the individual ice crystal boundaries which will allow any remaining surface water to drain through the ice cover (Brownman 1974). As the lake level rises, the ice sheet is pushed upwards causing terrestrial meltwater to pool on top of the ice sheet along the shore-fast margins forming a moat (Heron 1985). This moat gradually deepens and widens until the ice sheet is floating free. Terrestrial meltwater entering the lake basin before the ice sheet is

floating free can be diverted around the lake margins on top of the ice surface and out the outflow without coming into contact with the underlying lake water (Fig. 8). Colour Lake moatwater moved in a somewhat complex circulation pattern and was not homogeneous, but rather reflected the chemistry of the various inputs entering it along its course. Timed Rhodamine dye spikes indicated moat velocities from .03m/sec near the outflow. A travel time of five hours was determined for water entering the moat from the East Inflow and exiting the outflow during the spring of 1984. The importance of the moat in transporting meltwater from the lake basin varies from year to year. During the spring of 1985 the Colour Lake moat was incompletely formed and generally stagnant during snowmelt, possibly as a result of the low snow accumulations from that winter. No water elevation differences could be detected at any time between the lake proper and the moat indicating that the two water bodies were linked.

Once the snowdrifts which occur each year in the outlet regions of Colour Lake are breached, water from the ice surface begins to leave the basin through Outflow 1. As the lake level rises above the outlet point of the basin, Outflow 2 begins to flow, fed by sub-ice lake and meltwater.

Figure 8. 1984 Colour Lake Moat
June 7-8.



Chemical

Colour Lake's late winter mean chemical parameters are presented in Tables 4 and 5. Charge balance for all Colour Lake samples are within 10% except those from the

Table 4. 1984 Colour Lake Mass Changes

	May 7	May 21	June 11	% Change to end of Terrestrial Snowmelt June 12
H ⁺ (moles)	142810.41	127279.92	-	-
Ca (kg/moles)	1551.08	-	1,454.93	- 6.8
Mg (kg/moles)	661.77	-	603.45	- 8.8
Na (kg/moles)	218.25	-	166.22	-23.8
K (kg/moles)	98.96	-	98.27	- 0.7
Cl (kg/moles)	18.68	-	14.43	-22.7
SO ₄ (kg/moles)	2123.85	-	2,032.68	- 4.3
Si (kg/moles)	83.61	-	78.06	- 6.6
Fe (kg/moles)	6.15	-	9.55	+55.3
Conductance uS/cm ²	497	509	509	+ 2.4
Specific heat cal/cm ³	- 9.98	- 10.0	- 9.38	+ 6.0

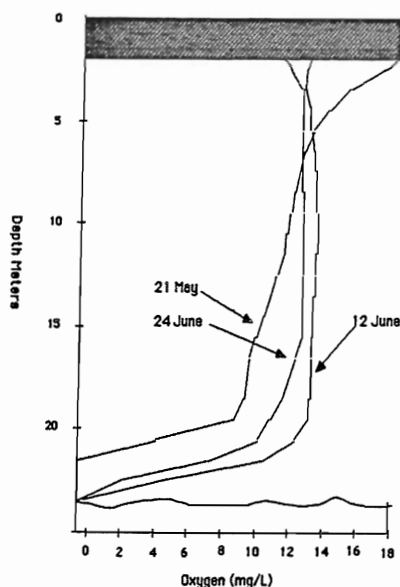
Table 5 1985 Colour Lake Mass Changes

	May 21	27	30	June 3	June 11	% Change to End of Terrestrial Snowmelt (June 11)
H ⁺ (moles)	163,968.30	-	-	141,099.47	111,523.40	-32.0
O ₂ (kg/moles)	349.17	-	-	-	363.95	+ 4.2
Ca (kg/moles)	1,445.77	-	-	1,425.70	1,452.65	+ 0.5
Mg (kg/moles)	557.50	-	-	570.54	577.55	+ 3.6
Na (kg/moles)	180.90	-	-	181.08	180.18	- 0.4
K (kg/moles)	98.70	-	-	106.48	105.98	+ 7.4
Cl (kg/moles)	44.65	-	-	46.32	46.39	+ 3.8
SO ₄ (kg/moles)	2,160.36	-	-	2,181.69	2,231.55	+ 3.3
F (kg/moles)	10.02	-	-	8.14	13.44	- 6.5
PPUT (kg/moles)	0.10	-	-	0.06	0.13	+34.1
TKN (kg/moles)	3.91	-	-	3.29	3.49	-10.8
Conductance uS/cm ²	552.6	560.6	544.7	542.8	554.3	- 0.3
Specific Heat cal/cm ³	- 14.15	-13.17	-	- 8.14	- 7.33	+48.0

anoxic bottom layer, which contained high iron concentrations which were not included in the charge balance equation. The bottom samples had hydroxide alkalinities up to 319 ueq/l and left a red precipitate in the sample bottles (likely $\text{Fe}(\text{OH})_x$).

Oxygen concentrations were greatest just under the ice cover when it was thickest in late May (Fig. 9), at which time a 0.5m to 3m anaerobic stratum was present above the lake bottom. There is evidence of an internal circulation of oxygen with the near-ice concentrations declining and the deep anaerobic zone becoming more aerated (Fig. 9). The mechanism of this deep oxygen circulation is likely the same as for the deep heat circulation mentioned earlier.

Figure 9. 1985 Colour Lake Oxygen Profiles



Conductivity profiles (Figs. 7a and 7b) show a maximum in the anaerobic hypolimnion. A slight conductivity maximum adjacent to the under-ice surface was measured in 1984 but not in 1985. The absence of a substantial near-ice conductivity peak may be a result of our late winter sampling date. Ice growth had all but ceased at the start of each field season thus ions excluded by freezeout during ice growth could have been mixed downwards into the underlying water column by convective circulation (Welch 1985), or by density currents (Schindler et al 1974). The bottom conductivity maxima can only account for a small percentage (< 6%) of the ions released during ice growth, indicating an overwinter whole water column circulation, rather than the presence of a stagnant hypolimnetic sink.

Conductivity profiles obtained from the snowmelt period indicate a thin (ca. 70cm) meltwater layer adjacent to the under-ice surface (Figs. 7a and 7b). This meltwater is a combination of surface ice and snow and terrestrial runoff water. This dilute layer was found to be consistent lakewide in 1984 and 1985. For this meltwater layer to exist some density difference, either salinity or thermally based, must exist between the lake and runoff water. Using an empirically derived equation:

$$\text{TDS mg/l} = 66.071 + 0.411 (\text{cond } \mu\text{S @ } 25^\circ\text{C}), (r^2 = .94),$$

for Colour Lake water, it appears that ionic density gradients are approximately the same order of magnitude as thermal density differences ($1 \times 10^{-4} \text{g/cm}^3$). Both types of density gradient appear to support the maintenance of this layer.

A sampling transect running outwards from a small streamlet towards the lake centre was sampled on May 31 and June 4, 1985 (Figs. 10a,b and 11a,b). Fig. 10a shows a thin dilute meltwater layer moving under the ice for approximately seven metres before layering out over the underlying lake water. The June 4 sampling date (Fig. 10b) shows much the same pattern, but with a much more dilute nearshore zone as a result of surface

ice melt in the moat area. Temperature isotherms from the transect indicate a substantial nearshore warming as a result of the reduced albedo of the moat in relation to the lake centre on the second sampling date. Temperatures have increased and conductance has decreased over time in the deeper zones of the transect indicating a downslope movement of water below the meltwater layer.

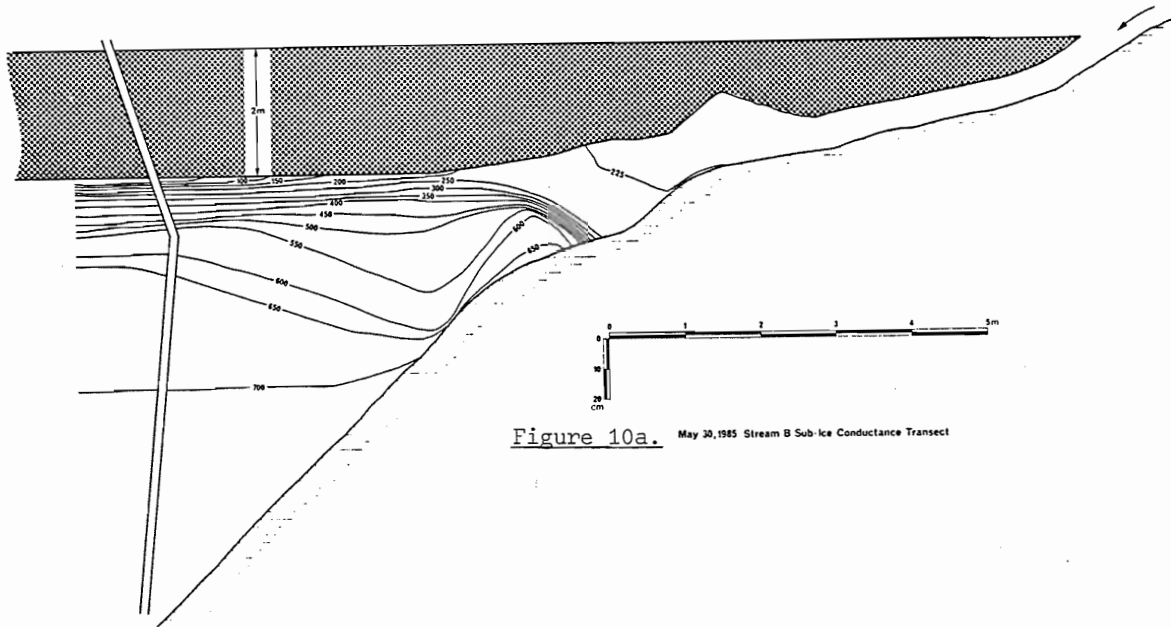


Figure 10a. May 30, 1985 Stream B Sub-Ice Conductance Transect

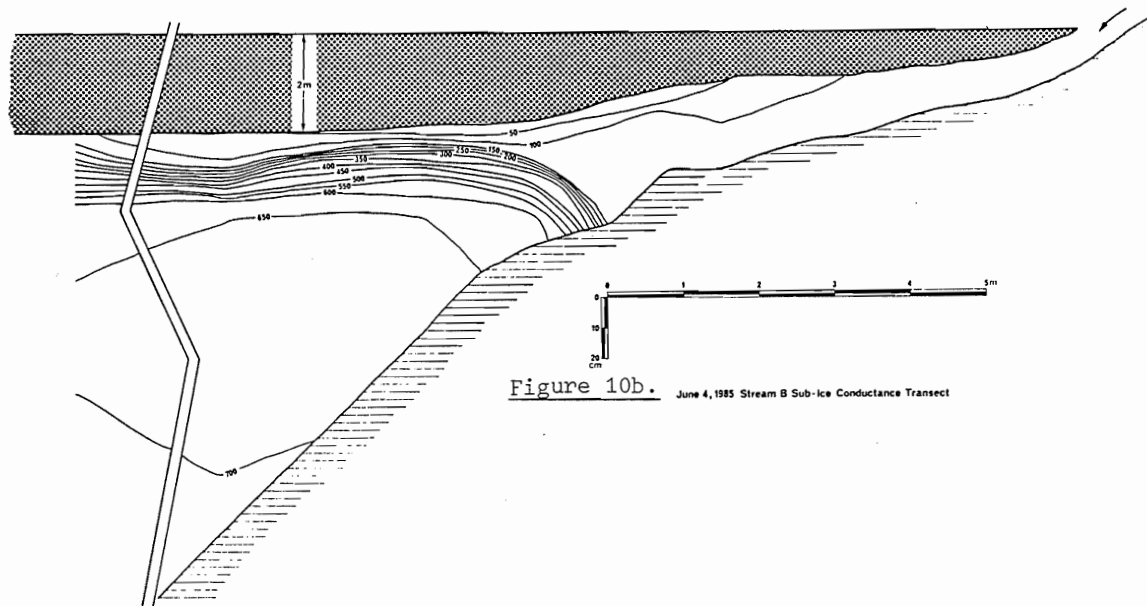
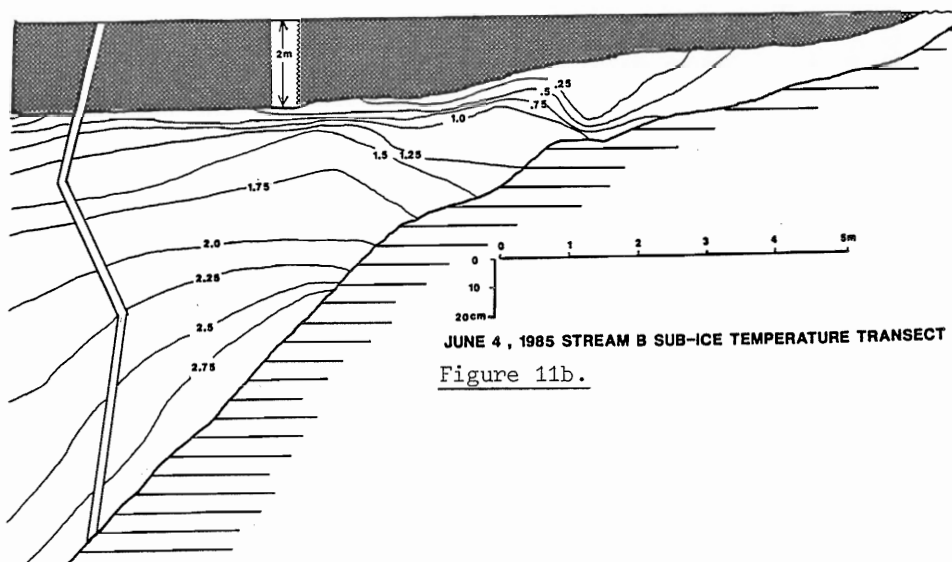
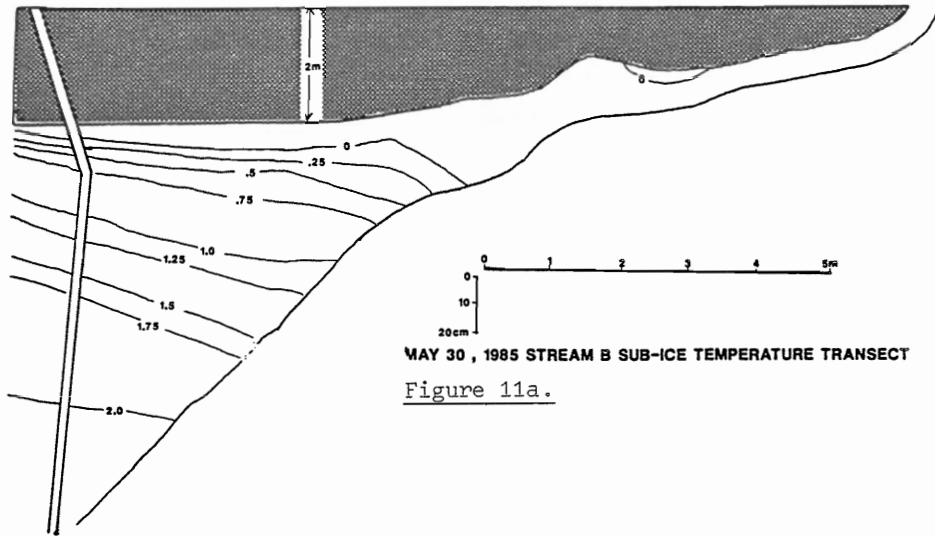


Figure 10b. June 4, 1985 Stream B Sub-Ice Conductance Transect

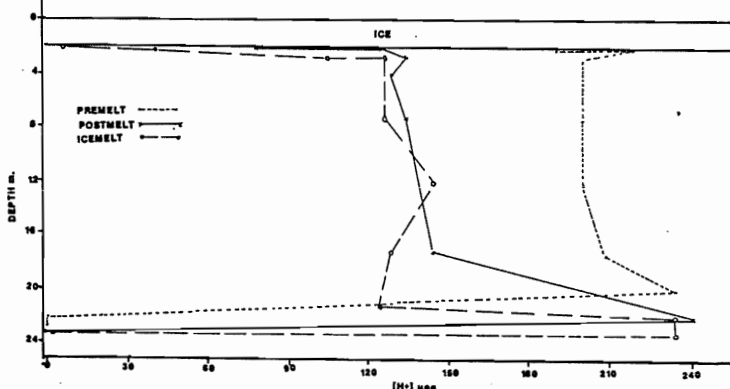


Mass calculations do not show a significant change (-0.3 to +2.4%) over the 1984 and 1985 snowmelt periods using the conductance data (Tables 4 and 5). Most parameters show a slight dilution in 1984 while in 1985 there is little or no change during terrestrial snowmelt, possibly because of the greater snowmelt inputs in 1984. Any dilution is largely a result of the establishment of the near-ice meltwater layer (Tables 4 and 5). Total unfiltered phosphorus increased during snowmelt as a result of terrestrial runoff having higher concentrations of phosphorus than resident lake water.

Hydrogen ion concentration shows a 32% decrease over the 1985 spring snowmelt period (Tables 4 and 5 and Fig. 12). Alkalinity measurements indicate a substantial (up to 329 ueq/l) buffering capacity in the inflows (Schiff unpublished data), particularly in the West and East inflows, the two largest. Unlike the other ionic parameters there is a substantial decrease of hydrogen ion below the near-ice meltwater layer (Fig. 12). Approximately one third of the hydrogen ion decrease can be attributed to the neutralization of the H^+ in the meltwater layer adjacent to the icecover. The remaining change in acidity occurs below this meltwater layer and is somewhat more difficult to explain. A possible neutralization mechanism could be the buffering of the lake by

carbonitic particulate matter which settles through the density gradient supporting the meltwater layer, thus affecting the whole water column. Some support for this hypothesis is presented in Table 5, where the decrease in H^+ is roughly balanced, on an equivalent basis, by an increase in magnesium and potassium. Unlike more southerly lakes which may experience an "acid-shock" as polluted snow melts, Colour Lake appears to undergo an alkaline titration each spring.

Figure 12. 1985 COLOUR LAKE [H+] PROFILES



Outflow chemistry shows some interesting temporal and spatial trends. Mean concentrations of Outflow 1 are lower and more alkaline than those of Outflow 2, indicating a meltwater source for its water (Tables 1, 2 and 3). Outflow 2, which contained 93% of the water which left the lake has a moderately higher conductance and acidity than Outflow 1, indicating a partial contribution of lake water to this outlet (Tables 2 and 3).

TABLE 1
AXEL HEIBERG COLOUR LAKE OUTFLOW 1984

Date	Time	pH	Cond (umho/cm)	Calcium (mg/L)	Magnesium (mg/L)	Sodium (mg/L)	Potassium (mg/L)	Chloride (mg/L)	Sulphate (mg/L)	Silica (mg/L)	Iron (mg/L)	Doc (mg/L)	Total P (mg/L)	Nitrate (mg/L)	Sum Cations (Eq/L)	Sum Anions (Eq/L)	Cations- Anions (Eq/L)	Percent Imbalance
7 Jun	2220	6.15*	297.9	28.80	4.34	0.76	1.14	1.03	81.92	1.15	0.135	12.8	0.019	1857.2	1734.6	122.6	6.6	
8 Jun	1815	6.15*	227.2	32.00	4.48	0.80	1.06	0.95	79.01	0.78	0.070	2.1	0.007	2028.1	1671.7	356.3	17.6	
9 Jun	1446	4.80	365.1	36.90	8.30	2.26	2.32	0.63	143.90	1.40	0.290	0.8	0.011	2697.8	3013.7	-315.9	-11.7	
10 Jun	1533	3.75	531.7	48.70	13.20	3.50	3.56	0.58	207.60	2.09	0.465	0.7	0.003	3937.5	4338.5	-401.0	-10.2	
11 Jun	1750	3.75	409.8	34.10	8.76	2.40	2.56	0.43	137.55	1.63	0.385	0.6	0.012	2770.1	2879.8	-109.7	-3.8	
12 Jun	1440	4.60	248.6	22.20	4.82	1.40	1.56	0.34	88.45	1.05	0.340	0.8	0.003	1630.3	1851.1	-220.7	-13.5	
Mean			346.7	33.78	7.32	1.85	2.03	0.66	123.07	1.35	0.281	3.0	0.009	2486.8	2580.9	-94.1	-2.5	
Standard Dev.			113.9	8.88	3.49	1.07	0.97	0.28	50.24	0.47	0.151	4.8	0.006	844.9	1043.0	286.2	12.3	
Maximum Value			531.7	48.70	13.20	3.50	3.56	1.03	207.60	2.09	0.465	12.8	0.019	3937.5	4338.5	-401.0	17.6	
Minimum Value			227.2	22.20	4.34	0.76	1.06	0.34	79.01	0.78	0.070	0.6	0.003	1630.3	1671.7	-401.0	-13.5	
# of Samples			6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6

* Outflow1, all other samples are from Outflow2

TABLE 2
AXEL HEIBERG OUTFLOW 1 CHEMISTRY 1985

Date	Time	Discharge (l/sec)	pH	Cond (umho/cm)	Calcium (mg/L)	Magnesium (mg/L)	Sodium (mg/L)	Potassium (mg/L)	Chloride (mg/L)	Fluoride (mg/L)	Sulphate (mg/L)	Nitrate (mg/L)	Total Unfilt P(mg/L)	Total Filt P(mg/L)	Total Kjeldahl (mg/L)	Sum Cations (Eq/L)	Sum Anions (Eq/L)	Cations- Anions (Eq/L)	Percent Imbalance
27 May	727		6.08	501.8	71.18	7.63	1.114	1.832	2.17	0.05	188.20	0.25	0.025	0.006	0.300	4275.9	3982.1	293.9	6.9
27 May	1156	0.0029	7.08	327.5	42.70	6.21	0.933	1.387	2.01	0.23	188.20	0.25	0.016	0.005	0.380	2717.7	2717.7		
28 May	1500	0.013	6.20	247.4	29.21	3.98	0.685	0.863	1.90	0.05	79.60	0.25	0.021	0.005	0.190	1837.8	1713.5	124.3	6.8
29 May	416		6.28	167.6	20.97	2.02	0.535	0.724	0.64	0.08	49.50	0.05		0.005		1255.1	1052.8	202.2	16.1
29 May	1730		6.07	310.5	35.96	5.05	1.254	1.431								2301.8	2301.8		
30 May	2245	0.002	6.37	251.3	33.71	3.58	0.759	1.096								2038.2	2038.2		
31 May	1605	0.003	6.07	122.3	13.74	1.26	0.406	0.661	0.42	0.01	35.90	0.05	0.016	0.005	0.200	1100.5	995.5	105.0	9.5
1 Jun	2147	0.0062	6.13	157.0	18.72	2.07	0.535	0.767	0.52	0.11	35.60	0.05	0.010	0.005	0.150	859.5	761.6	97.9	11.4
2 Jun	1652	0.007	6.08	121.1	12.94	1.58	0.442	0.649	0.47	0.11	46.90	0.05	0.024	0.004	0.170	1100.5	995.5	105.0	9.5
3 Jun	1535	0.0013	6.04	147.9	13.17	3.94	0.664	1.053	0.55	0.12	45.00	0.05	0.010	0.006	0.150	859.5	761.6	97.9	11.4
4 Jun	2130	0.0096	6.07	123.9	11.58	3.27	0.737	0.920	0.46	0.07	35.90	0.05	0.008	0.007	0.130	903.2	764.1	139.1	15.4
Mean				225.31	27.63	3.69	0.733	1.035	1.02	0.09	64.58	0.12	0.016	0.005	0.206	1741.1	1373.5	367.6	10.2
Stand Dev.				118.77	17.98	2.00	0.271	0.377	0.76	0.06	51.98	0.10	0.007	0.001	0.089	1060.7	1099.4	-38.7	3.8
Maximum Value				501.82	71.18	7.63	1.254	1.832	2.17	0.23	188.20	0.25	0.025	0.007	0.380	4275.9	3982.1	293.9	6.9
Minimum Value				121.08	11.58	1.26	0.406	0.649	0.42	0.01	35.60	0.05	0.008	0.004	0.130	825.3	759.8	65.5	6.8
Number				11	11	11	11	11	9	9	8	9	8	9	8	11	8	11	8

TABLE 3 AXEL HEIBERG COLOUR LAKE OUTFLOW 2 1985

Date	Time	Discharge (l/sec)	pH	Cond (unho/cm)	Calcium (mg/L)	Magnesium (mg/L)	Sodium (mg/L)	Potassium (mg/L)	Chloride (mg/L)	Fluoride (mg/L)	Sulphate (mg/L)	Nitrate (mg/L)	Total Unfilt P(mg/L)	Total Filt P(mg/L)	Total Kjeldahl (mg/L)	Sum Cations (Eq/L)	Sum Anions (Eq/L)	Cations- Anions (Eq/L)	Percent Imbalance
28 May	1440	0.0084	4.59	637.0	50.20	39.44	5.719	2.636	2.03	0.38	261.20		0.006	0.006	0.150	6092.3	5515.3	577.0	9.5
29 May	413		4.63	594.7	44.20	33.33	5.169	2.531	1.97	0.17	238.70	0.250		0.004		5261.2	5034.1	227.1	4.3
29 May	1735		4.64	509.3	35.21	27.16	4.177	2.447	1.98	0.17	186.60	0.250		0.005		4259.1	3949.7	309.4	7.3
30 May	2230	0.0015	4.58	496.4	41.95	28.20	4.484	3.040								4712.9		4712.9	
1 Jun	2150	0.0127	4.38	466.4	30.71	21.74	4.884	3.404	3.26	0.13	163.00	0.250		0.007		3662.6	3492.4	170.2	4.6
2 Jun	1652	0.0321	4.77	188.3	13.48	7.25	1.301	1.502	0.50	0.10	58.20	0.250	0.010	0.004	0.140	1381.2	1231.1	150.2	10.9
3 Jun	1530	0.1050	5.88	140.1	11.58	3.95	0.879	1.035	0.50	0.10	38.80	0.250	0.011	0.015	0.120	968.9	827.2	141.7	14.6
4 Jun	1622	0.0770	5.83	124.1	11.35	3.46	0.737	0.894	0.50	0.10	35.00	0.250	0.011	0.007	0.160	907.5	748.0	159.4	17.6
5 Jun	1419	0.0612	5.48	104.4	9.31	2.62	0.588	0.790	0.50	0.10	28.50	0.250	0.013		0.130	729.3	612.7	116.5	16.0
6 Jun	2330	0.0377	5.48	109.4	10.22	2.81	0.643	1.650	0.44	0.10	33.40	0.050	0.006		0.090	814.7	713.0	101.6	12.5
9 Jun	1500	0.0184	5.71	272.5	6.47	2.03	0.444	0.655	0.42	0.26	22.00	0.050	0.004		0.070	527.9	483.6	44.4	8.4
10 Jun	1700	0.0260	6.20	166.0	4.20	1.19	0.292	0.412	0.34	0.01	13.80	0.050	0.007		0.100	331.4	297.2	34.2	10.3
12 Jun	2200	0.0196	5.50	51.2	4.43	1.39	0.301	0.461	0.13	0.01	15.60	0.050	0.005	0.007	0.080	363.5	328.6	34.9	9.6
13 Jun	1955	0.0223	4.88	135.5	13.62	3.80	0.788	0.950	0.25	0.10	54.60	0.125	0.005	0.005	0.060	1064.1	1149.1	- 85.0	- 8.0
14 Jun	2230	0.0168	4.50	255.9	23.22	6.91	1.615	0.804	0.25	0.10	92.20	0.250	0.011	0.002	0.120	1849.7	1931.9	- 82.1	- 4.4
15 Jun	1600	0.0170	4.55	235.1	21.72	6.22	1.500	1.567	0.25	0.10	85.50	0.250	0.002	0.002	0.050	1729.2	1792.4	- 63.2	- 3.7
18 Jun	2146	0.0164	4.23	234.0	18.72	5.97	1.373	1.457	0.25	0.10	78.10	0.250	0.001	0.001	0.040	1581.3	1638.3	- 57.1	- 3.6
20 Jun	2217	0.0190	4.42	364.1	34.46	10.06	2.328	2.367	0.25	0.10	133.60	0.250	0.002		0.080	2747.2	2793.8	- 46.6	- 1.7
Mean				281.4	21.39	11.53	2.068	1.589	0.81	0.12	90.52	2.757	0.007	0.006	0.099	2165.8	1914.0	358.1	6.1
Standard Deviation				182.1	14.61	12.41	1.891	0.931	0.91	0.09	78.66	10.577	0.004	0.004	0.038	1832.9	1660.3	1099.0	7.8
Maximum Value				637.0	50.20	39.44	5.719	3.404	3.26	0.38	261.20	43.800	0.013	0.015	0.160	6092.3	5515.3	4712.9	17.6
Minimum Value				51.2	4.20	1.19	0.292	0.412	0.13	0.01	13.80	0.050	0.001	0.002	0.040	331.4	297.2	- 85.0	- 8.0
Number				18	18	18	18	18	17	17	17	17	14	11	14	331.4	297.2	18	17

Temporal trends indicate an initial displacement of lake water (high conductance and low pH) at low initial discharges followed by a period of dilute, less acidic water during the peak discharge and recession period (Figs. 13a and 13b). The broad shallow nature of the outflows leads us to believe that water exiting the outflow after the initial lake pulse is mainly near-ice meltwater.

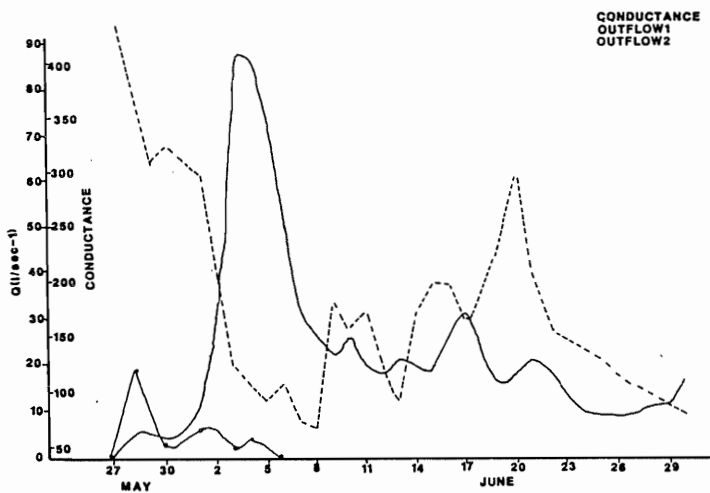


Figure 13a Colour Lake discharge and Conductance over time.

Figure 13b. Colour Lake discharge and pH over time.

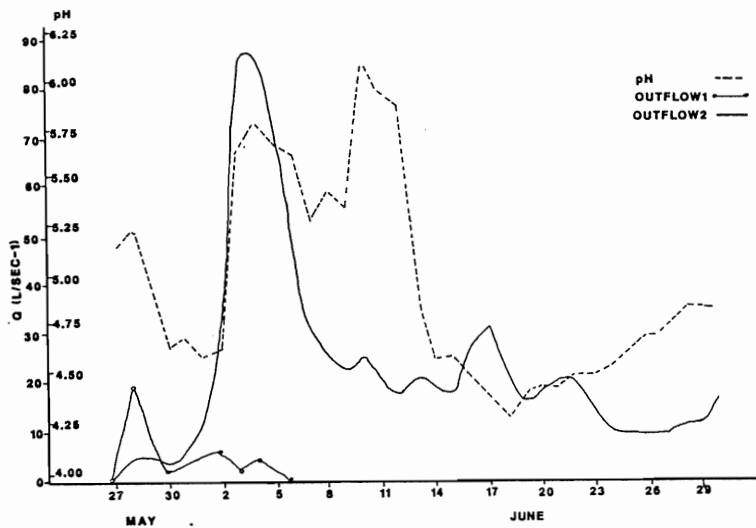
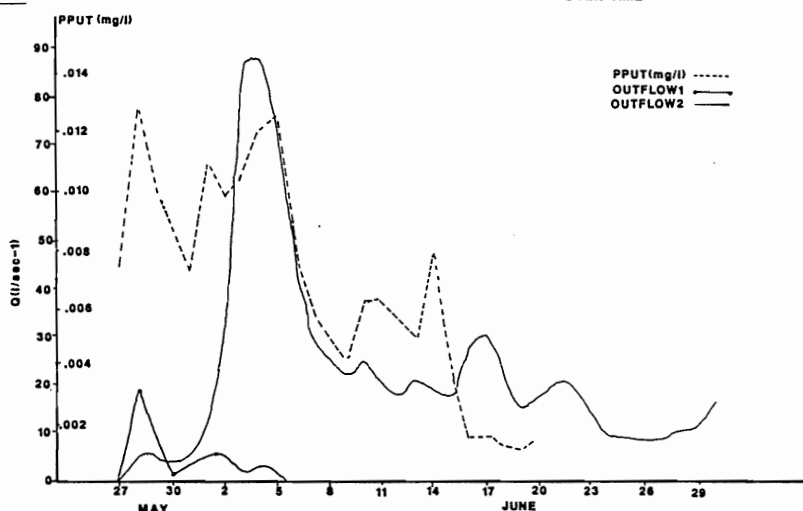


Figure 13c. COLOUR LAKE DISCHARGE AND TOTAL UNFILTERED PHOSPHOROUS OVER TIME



Several interesting characteristics reflect the non-lake origin of the outflow water. Ratios of Ca-Mg and Na-K in the outflows are decreased as a result of the weathering of crustal material and the higher proportion of Mg and K in the runoff as compared to the lake. Total unfiltered phosphorus, which is relatively high in the runoff as compared to the lake, follows a different temporal pattern than conductivity and pH. Phosphorus concentrations are high early in the runoff, including the peak flow period and decline as the proportion of ice meltwater increases in the outflow (Fig. 13c). The initial high outflow phosphorus levels, while the water is acidic, may be a result of the close proximity of Outflow 1 to the base camp kitchen and some runoff from this source contributing to the outflow in addition to the lakewater exiting the system at this time.

Discussion

Several snow and ice meltwater pathways exist in a high arctic lake system. Early in the melt, while the ice surface is flooded, a large portion of the terrestrial runoff can move over the ice surface around the lake in a meltwater moat. The importance of the moat in the transfer of meltwater through an arctic lake system can vary between years, depending on snow accumulation over the previous winter and timing and intensity of the melt that spring.

If the ice surface is breached while its surface is flooded, water can enter the lake in a turbulent vertical vortex which can substantially influence the under-ice water profile to a great depth for at least a short period of time. In the case of Colour Lake this water appears to return to the under-ice surface because of its lower density in relation to the resident lake water.

Meltwater that does enter the lake under the ice cover appears to layer under the ice in a thin (<1m) stratum and is preferentially lost to the outflow. Outflows chemically reflect the near-ice meltwater layer rather than the resident lake water while the lake is ice covered. Any dilution of Colour Lake can largely be attributed to the establishment of the near-ice meltwater layer, however a significant whole lake circulation below the meltwater layer does occur. The chemical density differences between the meltwater layer and the underlying lake water are not sufficiently large to prevent thermal circulation. For example, if meltwater up to one tenth the concentration of Colour Lake water was heated to 4°C, it would be able to sink through the underlying lake water despite its lower concentration. Less heating would be required to circulate more concentrated water. The deep convective circulation process must begin in the nearshore zones of the lake. The presence of the dark nearshore moat with its low albedo allows a differential heating to occur. The nearshore meltwater is

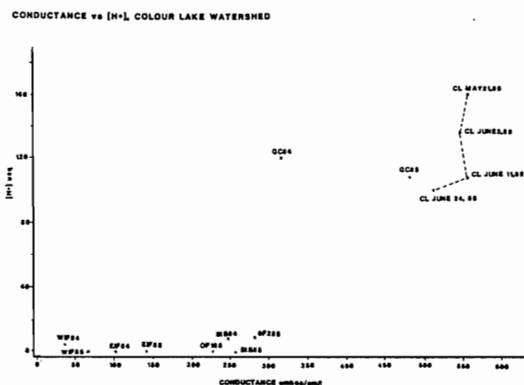
lower

heated and begins to sink along the sediment-water interface. The overwinter lake water was likely at some equilibrium with the sediment prewater. The replacement of this lake water with dilute meltwater probably leads to an ionic flux from the sediments to the overlying meltwater. The material flux from the sediments, combined with the differential solar heating, would allow the meltwater to move downslope beginning the whole under-ice circulation process which brings about the deep heating and aeration observed throughout Colour Lake during the spring snowmelt. Observations of snowmelt water moving across siltstone deposits in a solifluction scar in the Gordon Creek subbasin (Fig. 3) indicate a 30X increase in conductivity and acidity over a distance of less than ten centimetres, giving some indication of the exchange capacity of the sediment deposits under the Colour Lake basin (Pierson et al 1985).

Ice cover has two effects which act to increase the concentration of solutes of a lake over its inflows. The first is cryoconcentration from ice growth. Secondly, in lakes with small diffuse inflows, the ice cover seals the lake from wind-induced mixing, and allows the movement of the dilute meltwater, both inflows and icemelt, through and out of the lake with incomplete mixing with the lake body (Bergmann and Welch 1985). This effect increases the actual renewal time of a lake, the discrepancy being greatest for lakes with low AD:AO, high Zmean : AO or AD, thick, persistent ice covers and low summer rainfall in relation to snowmelt (Welch and Legault in prep).

The results of cryoconcentration, flow-through effects and subbasin geologic differences, are shown in Fig. 14 and Tables 4 and 5. Fig. 14 shows Colour Lake's ionic relationship with that of its inflows and outflows. Colour Lake in the late winter is from 1.1 to 15X more concentrated (conductance data), and 1.31 to 700X more acidic than its inflows and outflows. Gordon Creek most closely resembles Colour Lake water, as mentioned earlier this is a result of the stream channel lying in a deeply incised solifluction scar exposing acidic mineral deposits which are likely the source of Colour Lake's acidity (Pierson et al 1985). There appears to be a slight diluting of most chemical parameters in 1984 which can largely be accounted for by the establishment of the under-ice meltwater layer excluding more concentrated lakewater. In 1985, which is a complete account of the terrestrial snowmelt, there appears to be a slight increase in the mass of most elements in the lake. This indicates that there is a preferred loss of dilute snow and ice meltwater from the outflows. The slight increase in chemical masses may be a result of the replacement of ultrapure lake ice with more concentrated streamwater inputs.

Figure 14.



It appears that a substantial proportion of the yearly spring snowmelt water is exported directly from Colour Lake as a result of incomplete mixing. This allows Colour Lake to maintain its concentrational difference from its outflows and supports the Schindler et al (1974) hypothesis. For the acidity differences between the lake and its inputs to be maintained a reacidification must occur. Two acidification mechanisms which likely act in combination are, summer rainfall percolating through a deeper active layer and coming into the same mineral deposits which acidify Gordon Creek, and the oxidation of pyrite at the sediment water interface which would release mineral acidity in the form of sulphuric acid: $(FeSO_4 + 2H_2O) - (H_2SO_4 + FeOH_x)$. Preliminary sediment porewater chemistry indicates the latter may be the most important process (Schiff unpublished data).

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