The impact of a loss of hydrologic connectivity on boreal lake thermal and evaporative regimes

Christopher Spence,1* Ken Beaty,2 Paul J. Blanchfield,2 Lee Hrenchuk,3 Murray D. MacKay4
1Environment and Climate Change Canada, Saskatoon, Canada
2Fisheries and Oceans Canada, Winnipeg, Canada
3IISD – Experimental Lakes Area, Winnipeg, Canada
4Environment and Climate Change Canada, Downsview, Canada

Abstract

This paper summarizes the initial results (first 4 yr) of a whole-catchment manipulation to evaluate the impact of a loss of hydrologic connectivity of a boreal lake with its watershed on lake thermal and evaporative regimes. We diverted the upstream flow to a fourth order lake resulting in a marked reduction (81%) in watershed area, and compared the response to a comparable control lake in a headwater position. The manipulation reduced runoff into the experimental lake from 2400 mm to 90 mm, greatly increased theoretical residence time (from 2.3 yr to 18.3 yr), and reduced average dissolved organic carbon concentration from 5.8 ± 0.3 mg L⁻¹ to 5.5 ± 0.3 mg L⁻¹. Average Secchi depth was increased by 0.6 m, resulting in a 0.25 m deeper epilimnion and cooler surface water temperatures (8% of the time) than those predicted by pre-manipulation relationships. There was some evidence that the response of evaporation to episodic events was altered, but this was not extensive enough to alter annual evaporation or median evaporation rates. Our findings show that the impacts of hydrologic connectivity on lake chemistry can cascade to alter the energy budgets of boreal lakes. These results can inform how differences in hydrological connectivity across the landscape due to lake-watershed topology and climate will impact boreal lakes. Of note is that very clear lakes in the boreal region are highly sensitive to impacts from a change in clarity. Information on their spatial distribution will be necessary to assess impacts of hydrological connectivity on the boreal lake complex.

Lakes are ubiquitous across the boreal regions of the northern hemisphere. Up to 30% of boreal regions can be comprised of lakes and ponds (Molot and Dillon 1996; Bhatti et al. 2003; Rouse et al. 2005), the vast majority of them likely smaller than 0.1 km² (Downing et al. 2006; Cael and Seekell 2016). These water bodies occur across a diversity of climates (Rouse et al. 1997) and play a crucial role in the regional exchanges of heat, mass, and energy (Rouse et al. 2005). While the response of boreal lake biophysical properties and processes to stressors is complicated, literature suggests the importance of watershed processes vs. autochthonsous processes to lake response (Algesen et al. 2004; Benoy et al. 2007; Kellerman et al. 2013) can depend on the degree of interaction between a lake and its catchment. This implies that the dynamics of connectivity, defined as the ability to transfer mass or energy from one point to another (Pringle 2001; Bracken and Croke 2007), acting both within a watershed and between a lake and its watershed, could play a crucial role in the physical and hydrometeorological regimes of boreal lakes and their vulnerability to change. In the context of hydrology, connectivity has been used to explain and describe system response across a diversity of scales from hillslopes, headwaters and higher-order watersheds (Grayson et al. 1997; James and Roulet 2007; Mayor et al. 2008; Jencso et al. 2009). In a study of aquatic-terrestrial biogeochemistry, connectivity is implicit in the ideas of “hot spots” and “hot moments” (McClain et al. 2003) as the movement of solutes across the landscape is disproportionately controlled by processes within a source area (i.e., the hot spot) and for specific periods (i.e., the hot moment).

There are wide seasonal and annual variations in hydrological connectivity among boreal lakes (Oswald et al. 2011; Phillips et al. 2011). The non-linear relationship between precipitation and connectivity in boreal regions is primarily due to heterogeneity in land cover and topography (Devito et al. 2005; Buffam et al. 2007; Lauden et al. 2012). This heterogeneity creates a threshold-mediated streamflow generation process (Allan and Roulet 1994; Peters et al. 1995; Spence and Woo 2003) that results in a dichotomy of either
connection or non-connection depending on antecedent conditions and water inputs. This characteristic of streamflow generation in the boreal region is important because it controls the nature of biogeochemical connectivity and transfer of material from catchments into lakes (Allan et al. 1993). For instance, while the fraction of wetlands within a watershed has consistently been correlated with long term dissolved organic carbon (DOC) export (Gergel et al. 1999; Stasko et al. 2012), it is not only the fraction of wetlands in a watershed, but more specifically, the wetness of these wetlands that controls DOC export. During dry, hydrologically disconnected periods, dissolved organic matter, which includes DOC, is allowed to accumulate in the watershed. When conditions become wet and the water table rises, wetlands become hydrologically connected to downstream lakes; transporting DOC (Hinton et al. 1997; Schindler et al. 1997; Buffam et al. 2007).

DOC has an important role in several limnological processes (Solomon et al. 2015), one of which is the extinction of radiation in the water column (Morris et al. 1995; Stasko et al. 2012). A change in DOC loading to a lake, or the quality of the DOC, is known to influence radiation penetration, thermocline depth and epilimnion temperature (Imberger and Patterson 1990; Schindler et al. 1990, 1997; Snucins and Gunn 2000; Gunn et al. 2002; Houser 2006), but the net result for surface water temperatures has not been evaluated. There are various definitions of “surface” temperature in the literature, such as 1 m below the surface (Snucins and Gunn 2000; Gunn et al. 2002) or epilimnion averages (Houser 2006), but for the purposes of this study, “surface temperature” was defined as the skin temperature because this is what controls energy exchange with the atmosphere. This literature implies surface temperatures should be enhanced with a decrease in DOC, but it is possible they may also become cooler if the same energy inputs are distributed through a clearer and more voluminous epilimnion. The thermal structure of boreal lakes has a profound impact on their hydrometeorology. Short term variations in turbulent, latent, and sensible heat fluxes from lakes are not governed by net radiation, because short wave radiation penetrates the water surface. Instead, temperature and water vapor density gradients between the surface and lower atmosphere are among the primary controls (Granger and Hedstrom 2011). Therefore, chemical loads to a lake could be a second order control on evaporation, as these govern lake clarity and in turn, thermal structure, as illustrated in Fig. 1. How changes in chemical loads to lakes might impact evaporation rates across the boreal lake complex at a variety of time scales (weekly, monthly, seasonally) remains uncertain. However any increase in evaporation could result in a positive feedback, whereby higher evaporation could reduce lake levels further, thus decreasing streamflow and DOC transport, increasing water clarity and evaporation in higher order lakes, and augmenting low water conditions.

There are various geometric and topological situations across the boreal landscape that allow for a range of possible hydrological connectivity conditions for individual lakes. Headwater lakes, for instance, may have no concentrated input of water or materials from a stream, and often only provide intermittent outflow. The perennial streamflow conditions experienced by higher order lakes would be typically associated with greater and more sustained hydrological connectivity, but depending on climate conditions, may be subject to the same intermittent flows. With regards to geometry, when lakes are small relative to the size of their watershed, their capacity to influence hydrological connectivity between upslope and downstream areas may be diminished. Before it can be determined how this variety of hydrological and biogeochemical connectivity conditions across the boreal region may control hydrometeorological regimes, it is important to evaluate how watershed area (or position) influences the physical properties of lakes. Furthermore, connectivity varies with climatic inputs, so how connectivity controls thermal and energy regimes will not be static for an individual lake. There remains much uncertainty over the future direction and magnitude of change in precipitation over the boreal region (Benoy et al. 2007; Sucker and Krause 2010; IPCC 2014; DiDato et al. 2016). Understanding how DOC, water clarity, thermal structure, and lake hydrometeorology may react to changes in hydrological connectivity, and whether it is due to landscape or climatic influence, is critical to determine lake vulnerability to change.

We conducted a watershed-scale manipulation to test these hydrometeorological hypotheses as a way to improve predictions of how boreal lakes react to changes in hydrological connectivity which can be caused by climate and resource development stressors. While there exists a long history of deductive methodologies in aquatic ecology (e.g., Blanchfield et al. 2009), and hydrometeorology (e.g., Rouse and Kershaw 1971), no aquatic manipulation studies have set out to directly examine these hydrometeorological questions via a landscape manipulation. Physical manipulations with large submerged propellers in Scandinavian lakes have deepened the thermocline and increased heat content, but these experiments did not pursue impacts on lake hydrometeorology (Lydersen et al. 2008; Forsius et al. 2010; Verta et al. 2010). Therefore, conducting a whole-catchment experiment would represent a first attempt at a deductive approach to study the influence of a catchment on lake hydrometeorology within the context of hydrological connectivity. To this end, we conducted a large-scale manipulation by diverting all upstream flow around a boreal lake. The objective was to quantify the impacts of a loss of hydrological connectivity on surface water temperatures and evaporation rates of a boreal lake. We interpret the results and discuss how structural landscape controls such as the distribution of lakes across a watershed, and variable climate controls such as precipitation amount, frequency and intensity, may alter spatial and temporal hydrological connectivity, as
well as how landscape alterations may cascade through boreal lake chemistry, and thermal and evaporative regimes.

Methods

Study site

The study lakes are located at IISD Experimental Lakes Area (IISD-ELA) southeast of Kenora, Ontario, Canada. The IISD-ELA has served as a natural laboratory for over 40 yr in the study of various human impacts through whole-lake manipulations, which permit investigations to occur at ecosystem-level spatial scales and multi-year temporal scales (Blanchfield et al. 2009). Each experimental lake is compared with one or a suite of unmanipulated lakes that are monitored over the long term to provide a reference envelope for experimental manipulations. In this study, Lake 626 was manipulated and Lake 373, a nearby ELA long-term reference lake, served as the control (Fig. 2; Table 1). Lake 626 is 25.8 ha in size, has a maximum depth of ~12 m, and naturally receives runoff from three upstream lakes and their catchments (Fig. 2). The advantages of using Lake 373 as a control include its similar surface area, its similar watershed : surface area ratio as Lake 626 following manipulation, and its proximity (300 m) to Lake 626 means both receive similar meteorological inputs throughout the study (Table 1). The disadvantage is that Lake 373 is deeper ($Z_{\text{max}} = 22$ m) and its volume is approximately double that of Lake 626 (Table 1), resulting in a colder water temperature regime than Lake 626, a higher heat storage capacity, and lower annual evaporation volume. These differences were detected during the pre-manipulation phase.

Experimental design

The field experimental methodology included two phases. The first phase, from April 2009 through October 2010, established baseline conditions and documented differences in thermal and evaporative regimes between the two lakes. The second phase, from April 2011 to November 2014, involved a catchment-level manipulation of water input; a novel approach not traditionally attempted in hydrometeorological studies to quantify the impacts of climatic variation on boreal lakes. Usually, lakes are observed for long periods of time and the influence of climatic conditions on processes and fluxes are deduced from the lake reaction to such conditions within the observation record (e.g., Schindler et al. 1990; Parker et al. 2009). The experimental design proposed here takes advantage of the long history of ecosystem-level experimentation at the Experimental Lakes Area, including various water level manipulations that comprise both drawdown and flooding studies (e.g., Hesslein et al. 2005; Turner et al. 2005). In November 2010, hydraulic structures were added to Lake 627, immediately upstream of Lake 626, to permit the diversion of runoff from Lake 627 around Lake 626 to the wetland immediately below it. These structures permitted the simulation of a loss of hydrological connectivity by reducing the effective watershed area of Lake 626 to 19% of its pre-manipulation size (Fig. 2; Table 1), essentially turning it into a headwater lake for the remainder of the

Fig. 1. Schematic illustrating the hypothesized interactions between upstream inflow, dissolved organic carbon (DOC) loading, residence time ($R_l$), light extinction ($K_d$), lake thermal structure ($T_s$) and lake evaporation ($E$) for two scenarios. The left depicts a situation with sustained hydrological connectivity between the upper basin and a lake. The right depicts a situation where that hydrological connectivity is severed.
study. The theoretical residence time of Lake 626, calculated as mean annual yield (225 mm yr\(^{-1}\)) divided by lake volume, increased from approximately 2 yr to 18 yr.

### Canada Small Lake Model (CSLM) application and hypothesis derivation

To derive hypotheses, lake thermal response to variations in transparency was explored through numerical experimentation. The Canadian Small Lake Model (MacKay 2012; MacKay et al. 2017) is a dynamic mixed layer scheme developed for use in Environment and Climate Change Canada’s land surface modelling programme. It consists of a simple one-dimensional thermal model coupled with a turbulent mixing parameterization based on the integrated turbulent kinetic energy (TKE) approach proposed by Rayner (1980) and Imberger (1985), and developed by Spigel et al. (1986). Briefly, a fully nonlinear energy balance is computed in a surface skin layer of arbitrary thickness. Turbulent mixing in the epilimnion is achieved through surface stirring and buoyancy production as well as shear production along the (diurnal) thermocline. In this instance it was configured to emulate Lake 626, and forced with meteorological conditions and incoming shortwave and longwave radiation observed at Lake 239, 10 km to the south. Shortwave radiation extinction was user manipulated within the model to develop hypotheses of how a change in upstream inflow would influence thermal conditions, particularly surface temperatures and latent heat fluxes from the lake. These manipulations included a standard model run with the observed baseline extinction condition of 0.5 m\(^{-1}\), and simulations with extinctions of 0.4 m\(^{-1}\) and 0.3 m\(^{-1}\).

**Fig. 2.** Location of the study (Lake 626) and reference (Lake 373) lakes. Bold outlines are watersheds, with the hatched portion of the Lake 626 watershed denoting the area producing streamflow that was diverted around Lake 626 post-manipulation (2011–2014). The upper photograph shows the dyke at the Lake 627 outlet and the lower photograph the new diversion outlet stream with lake level gauge. White circles show approximate locations of meteorological rafts and lake sampling stations. The arrow denotes the direction of the re-routed streamflow.
Table 1. Physical characteristics of experimental Lake 626 during baseline conditions (2009–2010) and the period of watershed manipulation (2011–2014) compared to headwater reference Lake 373 (2009–2014). Land cover data are from the Ontario Land Cover Data Base, Second Edition (https://www.ontario.ca/data/provincial-land-cover).

| Characteristic                      | Lake 626 | Lake 373 |
|------------------------------------|----------|----------|
| Area (ha)                          | 25.8     | –        |
| Watershed area at outlet (ha)      | 372.4    | 80.6     |
| Watershed lake fraction            | 0.3      | 0.4      |
| Watershed bedrock fraction         | 0.03     | 0.02     |
| Watershed forest fraction          | 0.61     | 0.48     |
| Watershed wetland fraction         | 0       | 0.02     |
| Mean depth (m)                     | 6.4      | 10.7     |
| Maximum depth (m)                  | 12       | 20.8     |
| Volume (m$^3$)                     | 1,772,000| 2,991,000|
| Theoretical residence time (yr)    | 2.3      | 25.1     |
| Mean Secchi depth (m)              | 4.9      | 5.5      |
| Mean light extinction coefficient  | 0.51     | 0.46     |

Field instrumentation

To estimate lake evaporation rates and observe the thermal regime, a meteorological tower on a floating platform was deployed in the center of each lake (Fig. 2) from May through October each year. Open water fetches for the predominant wind direction to the rafts on Lakes 626 and 373 were 300 m and 600 m, respectively. Each raft was equipped with a Kipp and Zonen NR-lite radiometer placed 1.1 m above the water surface to measure net radiation (Q$_n$; W m$^{-2}$). Three Vaisala HMP35CF thermohygrograms were located between the water surface and a height of 2 m on each raft to measure air temperature (T$_a$; °C) and vapor pressure (e$_v$; kPa). Anemometers were set at the same heights as the thermohygrograms to measure wind speed (u; m s$^{-1}$). These sensor heights resulted in maximum height : fetch ratios of 1 : 150 and 1 : 300 for the predominant wind direction. The water surface temperature (T$_w$; °C) was measured continuously with an Apogee infrared thermometer with an accuracy of 0.2°C. The surface vapor pressure (e$_v$; kPa) was calculated from this temperature assuming 100% relative humidity. These meteorological terms were measured every 10 s (Campbell Scientific CR3000 data loggers) with averages output half hourly from which daily average values were calculated to estimate evaporation using the equations described below.

Daily change in heat storage ($J_w$, W m$^{-2}$) was calculated from half hourly water temperatures measured with Onset Pendant thermistors attached to a cable anchored to the lake bottom adjacent to each tower. Thermistors were located immediately below the surface and every meter to the bottom of the lake. For each day, the mean water temperature $T_w$(°C) was calculated as:

$$T_w = \frac{1}{Z} \sum_{i=1}^{n} T_{wi} \cdot \Delta z$$

(1)

which allows the calculation of $J_w$ as:

$$J_w = \rho \cdot c_p \frac{\Delta T_w}{\Delta t} \cdot \Delta z$$

(2)

where $z$ is depth (m), $T_{wi}$ is the water temperature (°C) at the individual thermistor, $\Delta z$ is the depth segment assigned to that thermistor, $\rho$ is the density of water (kg m$^{-3}$), $c_p$ is the specific heat of water (J kg$^{-1}$ °C$^{-1}$) and $\Delta t$ is the duration of the averaging period (s).

The Penman–Monteith equation was used to calculate the evaporation rate ($E$, m s$^{-1}$) following:

$$E = \frac{1}{\Delta} \left[ \frac{\Delta \cdot (Q_n - J_w) + \rho_a \cdot c_p \cdot (c_v - e_v) / \rho_v}{\Delta + \gamma \cdot (1 + \rho_v / \rho_a)} \right]$$

(3)

where $\lambda$ is the latent heat of vaporization (J kg$^{-1}$), $\Delta$ is the slope of the saturated vapor pressure-temperature curve (kPa °C$^{-1}$), $\rho_a$ is air density (kg m$^{-3}$), $c_p$ is specific heat of moist air (J kg$^{-1}$ °C$^{-1}$), $\gamma$ is the psychrometric constant (kPa °C$^{-1}$), $\rho_v$ is the aerodynamic resistance calculated assuming a surface roughness length of 1 mm (s m$^{-1}$), and $e_v$ is surface resistance (s m$^{-1}$), assumed to be zero. Values of $E$ were converted to units of mm d$^{-1}$. Estimated uncertainty in evaporation estimates using these methods is ± 20% (Spence et al. 2003). To protect the rafts from ice damage, they were removed prior to freeze-up. This resulted in a loss of approximately 1–2 weeks of evaporation data at the end of the season. In order to open water season cumulative evaporation totals to be calculated with a standardized season extending from 2 weeks after ice-off until freeze-up, a bulk evaporation depth for the period after raft removal ($E_2$, mm) was added to the cumulative evaporation rates calculated with (3). $E_2$ was estimated by assuming the remaining seasonal lake heat storage dominated the available energy, and the latent heat flux was calculated using the average Bowen Ratio from the 2 weeks prior to raft removal:

$$E_2 = \frac{1}{\Delta} \left[ \frac{\Sigma T_w}{1 + \beta} \right]$$

(4)

To demonstrate the magnitude of the manipulation on the water budget of Lake 626 and compare it to that of Lake 373, V-notch weirs (120° at Lake 373 and 150° at Lake 626) were installed immediately downstream of each lake outlet. Discharge ($Q$, mm d$^{-1}$) from each lake was estimated using standard stage-discharge equations with stage at the weirs.
measured every 10 min with Ott Thalamedes float potentiometer data loggers. V-notch weirs result in accurate estimates of velocity, so the ±5% measurement error was primarily associated with the stage measurement from the data loggers. Catchment inflow \( i; \text{mm d}^{-1} \) was calculated by pro-rating measured streamflow by basin area using runoff data from the headwater catchment “Northwest Inflow to Lake 239” collected in a control structure in the same manner; a standard method of estimating streamflow from ungauged basins at the IISD-ELA (Schindler et al. 1976; Newbury and Beatty 1977).

Each open water season, the epilimnions of Lakes 626 and 373 were sampled bi-weekly for water chemistry, including DOC concentration, as well as limnological parameters.

The data were continuous except for a break in 2013 (Hering et al. 2012; Hoag 2012). To determine epilimnion depth, water temperatures were measured at 1 m depth increments with a multi-function probe (model RBR XRX620, RBR Ltd.). Near the bottom of the epilimnion, temperatures were measured at 0.25 m depth increments so as to pinpoint the depth of this layer more precisely. The bottom of the epilimnion was defined as the deepest measured depth where there was less than 0.25°C change in water temperature from the previous (shallower) quarter meter depth. To collect water samples for DOC analysis, an integrated sampler (Shearer 1978) was moved vertically between the surface of the lake and 0.5 m above the bottom of the epilimnion until the sampling bottle (2 L opaque amber polypropylene copolymer, Nalgene) was full. The sample was refrigerated at 4°C for up to 4 months prior to DOC analysis. DOC analysis was carried out using an automated instrument (Shimadzu TOC-VCPH + TNN-1 Total Organic Carbon and Total Nitrogen Analyzer with Gas Purification Kit) following the method described by Stainton et al. (1977). The method detection limit was 1.40 μmol L\(^{-1}\) (0.017 mg L\(^{-1}\)), the reporting detection limit to be 2.80 μmol L\(^{-1}\) (0.034 mg L\(^{-1}\)), and the level of quantification to be 4.21 μmol L\(^{-1}\) (0.050 mg L\(^{-1}\)).

Coincident Secchi depth and light extinction measurements were collected biweekly and used to assess lake transparency. Secchi depth, \( z_S \), was taken by lowering a Secchi disk into the water on the shaded side of the boat until the observer could no longer see the disk. The disk was then slowly brought back up in the water column until it was just visible, and the depth was noted and recorded to the nearest 0.25 m increment. Photosynthetically Active Radiation (PAR) was measured with an underwater PAR sensor (LI-192 Underwater Quantum Sensor attached to a LI-1400 logger, LI-COR) at 1 m depth increments on the sunny side of the boat. Readings were taken just above the surface of the water at the start and end of the profile, and if they differed by more than 10%, the profile was redone. Light extinction coefficients, \( K_d \), were calculated from PAR profile data as:

\[
K_d = \frac{n \cdot \left( \sum z \cdot \ln(\%PAR) - \left( \sum z \cdot \sum \ln(\%PAR) \right) \right)}{n \cdot \sum z^2 - \left( \sum z \right)^2}
\]

where \( \ln(\%PAR) \) is the natural log of the percent of surface PAR at each depth, and \( n \) is the number of data points (i.e., depths) used in the calculation. The manufacturer reports typical error as 3% and 7.4% as the maximum total error.

**Hypothesis testing**

A common approach in paired catchment studies (Watson et al. 2001; Moore et al. 2005; Gomi et al. 2006; Som et al. 2012) to detecting change due to a manipulation includes three steps. The first step is to derive a relationship between the entire dataset of control and pre-manipulation responses, which in this study could be of the form:

\[
x_{626,t} = m_x \cdot x_{373,t} + b_x + e_x
\]

where \( x_{626,t} \) is a predicted value of \( x_{626,t} \) using \( x_{373,t} \), where \( x \) is a hydrometeorological characteristic of the lake (i.e., DOC concentration, light extinction, surface temperature, evaporation rate) on day \( t \). The terms \( m_x \), \( b_x \), and \( e_x \) are slope and intercept coefficients estimated by regression and an error term, respectively. The second step is to estimate the manipulation effect by calculating the differences between the post-manipulation observations and the predicted values from Eq. 6 using:

\[
M_e = x_{626,t} - x'_{626,t}
\]

where \( x_{626,t} \) is the post-manipulation observed value.

One of the shortcomings of whole-ecosystem manipulation studies, such as the one described here, are the small number of values available for assessing effect, if an annual estimate is used. As such, it was necessary to evaluate more frequent data; specifically, we compared daily hydrometeorological data and bi-weekly limnological data.

It is recognized that these higher frequency data may be subject to autocorrelation, violating the assumptions of the tests listed above. When necessary, autocorrelation of residuals was removed to provide an estimate of the random disturbance on day \( t \), \( M_t \):

\[
M_t = (x_{626,t} - x'_{626,t}) - \rho_1 \cdot (x_{626,t-1} - x'_{626,t-1}) - \rho_2 \cdot (x_{626,t-2} - x'_{626,t-2}) - \ldots - \rho_k \cdot (x_{626,k-1} - x'_{626,k-1})
\]

where \( \rho_1 \ldots k \) is an estimate of the lag \( t \) autocorrelation coefficient.

The third step is to test the statistical significance of the differences between the observed and predicted characteristics. Since these random disturbances are independent, 95% confidence limits could be estimated as \( \pm 1.96 \sigma(M_t) \).
suggested by Watson et al. (2001) and Som et al. (2012), if more than 5% of the values of $M_r$ exceeded the 95% confidence limits, a statistically significant manipulation of a hydrometeorological characteristic of the lake was implied. When the data of any parameter did not adhere to the assumptions of this method, analysis of variance (ANOVA) and a paired $t$-test were used to evaluate if there was a difference in observations and predicted post-manipulation values. Last, the Mann–Whitney test was applied in those instances when the distributions did not pass normality or equal variance assumptions. There were some lake characteristics that were not assessed statistically because they were not associated with specific hypotheses, but provide context nonetheless. This includes maximum cumulative change in heat storage and epilimnion depth; the latter was measured using the July–August average, as in Schindler et al. (1996).

**Results**

**Study hypotheses**

The premise of this study was that disrupting hydrological connectivity from the upper basin to Lake 626 would diminish DOC concentrations in Lake 626, making it more transparent. The hydrometeorological impact of reducing shortwave extinction was explored with the CSLM (Fig. 3). Surface temperatures for all three simulations were similar to observed, though reducing extinction tended to systematically reduce surface temperature until fall turnover (27 September; DOY 270) and then systematically increase surface temperatures (Fig. 3a). Cumulative evaporation in the model was suppressed early in the season by reducing extinction, enough that seasonal evaporation totals were reduced from the standard scenario (Fig. 3b). By 27 October (DOY 300), after 152 d of accumulation the difference between the

![Fig. 3. Simulations of the CSLM of Lake 239 at the Experimental Lakes Area for 07 May 2009–27 October 2009 for three different shortwave extinction coefficients (0.5 blue, 0.4 green, 0.3 red) compared with observations (black): (a) surface temperature; (b) cumulative evaporation (accumulation begins DOY 148); (c) depth of 15°C isotherm. Also indicated in (c) are observed depths for dissolved oxygen at 6 mg L$^{-1}$ threshold (*).](image-url)
standard (0.5 m$^{-1}$) and clearest (0.3 m$^{-1}$) simulation was 22 mm (~6%).

We examined change in total heat content over the open water season as another way to determine the impacts of shortwave extinction on the thermal properties of a lake. For our simulations, the increase in heat content between 10 May and 22 September (135 d) was 406 MJ m$^{-2}$, 462 MJ m$^{-2}$, and 514 MJ m$^{-2}$ for extinctions of 0.5 m$^{-1}$, 0.4 m$^{-1}$, and 0.3 m$^{-1}$, respectively. These results suggest with a loss of hydrological connectivity, decline in DOC concentrations and reduction in light extinction will come two cascading effects; (1) reduced light extinction will result in a warmer water column with lower surface water temperatures earlier in the year and enhanced surface water temperatures after fall turnover; and, (2) resulting changes in surface water temperatures will suppress evaporation at the beginning of the open water season, but cumulative seasonal evaporation will not be affected as increases late in the season will offset these declines.

**Streamflows**

The strength of hydrological connectivity between a lake and its catchment can be expressed by the amount of streamflow entering the lake. Pre-manipulation, there was an order of magnitude difference in the streamflow experienced by the experimental and reference lakes (Table 2). Post-manipulation, without the inputs from the upstream watershed, the water level of Lake 626 dropped approximately 0.2 m (Fig. 4). Lake 626 inflows and outflows of an average 2400 ± 120 mm and 3100 ± 157 mm, respectively, before the manipulation decreased to 90 ± 5 mm and 130 ± 6 mm, respectively, after the manipulation. These are comparable to those estimated for Lake 373 throughout the experiment (Fig. 4; Table 2) and reflect that Lake 626 post-manipulation and Lake 373 had a similar degree of hydrological connectivity to their watersheds. It also reflects Lake 626’s new status as a headwater lake.

**Lake chemistry and light extinction**

Pre-manipulation, average DOC concentration was 3.9 ± 0.3 mg L$^{-1}$ in Lake 373 and 5.8 ± 0.3 mg L$^{-1}$ in Lake 626 (Fig. 5). The lower pre-manipulation DOC concentrations in Lake 373 relative to Lake 626 are assumed to be due to the combined lower rates of loading associated with the lower rates of inflow, and the longer water residence time for this headwater lake (Table 2). The lower DOC in Lake 373 contributed to less light extinction (0.41 ± 0.05 m$^{-1}$) than in Lake 626 (0.51 ± 0.07 m$^{-1}$) pre-manipulation (Fig. 5; Table 1). This was also reflected in the mean Secchi depths of each lake, which were 0.5 m deeper in Lake 373 (5.4 m) than Lake 626 (4.9 m) (Table 1).

While average DOC concentrations in Lake 373 remained stable at 4.1 ± 0.3 mg L$^{-1}$, they decreased from 5.8 ± 0.3 mg L$^{-1}$ pre-manipulation to 5.5 ± 0.3 mg L$^{-1}$ post-manipulation in Lake 626 (Table 2; Fig. 5). Thus, there was a significant difference between the pre- and post-manipulation DOC concentrations in Lake 626 (Mann–Whitney $p < 0.01$) (Table 3). Thirty two percent of the values of $M_p$ exceeded 1.96$\sigma(M_p)$, also indicating a significant change in DOC concentrations with the reduction of inflows to Lake 626. Moreover, the manipulation resulted only in decreases to DOC concentrations in Lake 626 relative to those in Lake 373 (Fig. 6). Similarly, light extinction coefficients in Lake 373 increased slightly from 0.41 m$^{-1}$ to 0.43 m$^{-1}$ during the post-manipulation period, whereas light extinction in Lake 626 decreased from 0.51 m$^{-1}$ to 0.46 m$^{-1}$ (Table 3). The regression between the pre-manipulation Lake 626 and Lake 373 light extinction was poor ($r^2 = 0.06$), which implies the Watson et al. test statistic is not robust enough to believe the 0.09 fraction of observed values exceeding 1.96$\sigma(M_p)$. The disturbance histogram, however, as with the Mann–Whitney results, demonstrates that the change in Lake 626 was consistently toward a reduction in light extinction (Fig. 6; Table 3). Lake 626 became more transparent following the loss of hydrological connectivity and the reduction in size of its watershed area.

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**Table 2.** Estimates of inflow ($I$) and outflow ($Q$) from Lakes 626 and 373 for each open water season of the study period demonstrating the impact of the watershed manipulation on the lake water budget of Lake 626. Totals are for 2 weeks after break-up until 31 October. Streamflow values are in mm over the area of each lake. Samples refer to the number of DOC samples and light extinction and Secchi depth measurements per year. The presented values for these terms are open season averages plus or minus one standard deviation. DOC is presented in mg/L, $K_d$ in m$^{-1}$, and $z_S$ in m. Baseline years are 2009 and 2010, and manipulation years are 2011–2014.

| Period  | Samples | Lake 626 | Lake 373 |
|---------|---------|----------|----------|
|         | Year    | $n$      | $I$      | $Q$ | DOC | $K_d$ | $z_S$ | $I$ | $Q$ | DOC | $K_d$ | $z_S$ |
| Baseline| 2009    | 13       | 1200 ± 60| 1770 ± 90| 5.7 ± 0.3 | 0.5 ± 0.05 | 5.1 ± 0.5 | 170 ± 9 | 320 ± 16| 3.8 ± 0.2 | 0.4 ± 0.04 | 5.3 ± 0.9 |
|         | 2010    | 13       | 3600 ± 180| 1500 ± 220| 5.9 ± 0.3 | 0.5 ± 0.08 | 4.6 ± 0.8 | 510 ± 26 | 820 ± 41| 4.1 ± 0.4 | 0.4 ± 0.06 | 5.5 ± 0.7 |
| Manipulation| 2011 | 14       | 110 ± 6 | 130 ± 7 | 5.8 ± 0.3 | 0.5 ± 0.04 | 5.7 ± 0.9 | 130 ± 7 | 250 ± 13| 4.2 ± 0.2 | 0.4 ± 0.09 | 5.6 ± 1.1 |
|         | 2012    | 14       | 40 ± 2 | 130 ± 7 | 5.3 ± 0.3 | 0.47 ± 0.05 | 5.7 ± 0.9 | 40 ± 2 | 220 ± 11| 3.9 ± 0.2 | 0.4 ± 0.07 | 5.1 ± 0.9 |
|         | 2013    | 6        | 60 ± 3 | 230 ± 12 | 5.1 ± 0.3 | 0.5 ± 0.1 | 4.8 ± 1.3 | 60 ± 3 | 260 ± 13| 4.1 ± 0.7 | 0.5 ± 0.07 | 5.1 ± 1.0 |
|         | 2014    | 13       | 160 ± 8 | 320 ± 16 | 5.6 ± 0.3 | 0.4 ± 0.06 | 5.4 ± 1.3 | 190 ± 10 | 520 ± 26| 4.4 ± 0.3 | 0.4 ± 0.07 | 5.1 ± 0.5 |
Lake temperatures

The increase in water transparency resulted in greater changes to heat storage in Lake 626 relative to that in Lake 373. Prior to the manipulation, the maximum cumulative change in heat stored from 2 weeks after spring break-up in Lake 373 during the baseline years (2009–2010) was 530 MJ...
and 615 MJ m$^{-2}$, respectively. In comparison, these values in Lake 626 were 365 MJ m$^{-2}$ and 482 MJ m$^{-2}$, respectively (Table 4). The smaller heat storage in Lake 626 can be attributed to the shallower depth. Post-manipulation, maximum cumulative change in heat storage became more similar between the two lakes, with the difference decreasing from an average of 149–49 MJ m$^{-2}$ (Table 4). Much of the additional energy was stored in the epilimnion, but the average temperature of this layer did not change. Rather, the thermocline in Lake 626 tended to be deeper post-manipulation. Pre-manipulation, the differences in this depth between the two lakes averaged 0.5 m, but post-manipulation this was reduced to 0.25 m (Table 4). The Lake 626 July–August epilimnion depth thickened by an average 1.6–2.5 m, and in 2013 and 2014 was deeper than that in Lake 373.

Table 3. Results of Kolmogrov–Smirnov (K–S) tests ($p$-values) to determine normality, and ANOVA or Mann–Whitney tests ($p$-values) to evaluate differences between means or medians of post-manipulation observed and predicted values. Results of the regression between the control and pre-manipulation conditions and the fraction of $M_r$ values exceeding the 95% prediction interval. Bolded terms are those tests results that are robust and imply a significant change post-manipulation.

|          | ANOVA          | Mann–Whitney | Watson et al. |
|----------|----------------|--------------|---------------|
|          | K-S $p$        | $p$          | $r^2$ SE      | $p$ $M_r>1.96\sigma(M_r)$ |
| DOC      | $<0.01$        | $<0.01$      | 0.43 0.12    | $<0.01$ 0.32 |
| $K_d$    | $<0.01$        | $<0.01$      | 0.06 0.2     | 0.71 0.09 |
| $T_s$    | $<0.01$ 0.23   | 0.01         | 0.93 0.01    | $<0.01$ 0.08 |
| $E$      | $<0.01$ 0.88   | 0.55         | 0.85 0.02    | $<0.01$ 0.06 |

Fig. 6. Disturbance histograms for DOC, light extinction, surface water temperature, and daily evaporation rate in Lake 626 post-manipulation (2011–2014). The dashed vertical lines represent $1.96\sigma(u_0)$. 
Collectively, the hydrometeorological responses in Lake 626 post-manipulation are similar to model predictions in Fig. 3c, which shows how a decrease in light extinction could increase the heat storage and the depth of the epilimnion, and are consistent with the hypothesis that hydrological disconnectivity would result in a warmer water column (Fig. 1). A warmer water column at depth implies that the same amount of energy was distributed through a larger volume of water which resulted in reduced surface temperatures in Lake 626 relative to Lake 373 (0.4°C). There was a significant difference in surface water temperatures detected using the Mann–Whitney test (Table 3). There was autocorrelation at 1-, 2- and 3-d periods detected in the time series, and was subsequently accounted for with the Watson et al. (2001) test statistic. This showed 8% of disturbances exceeded 1.96σ(Mr), indicating a slight change in surface temperatures with the severe reduction in hydrological connectivity to Lake 626 (Fig. 6). The strong kurtosis in the distribution (κ = 4.43) toward negative values, and the lack of disturbance values above + 3.9 (1.96σ(Mr)) for Tt demonstrates that all the changes in surface water temperatures due to the manipulation were associated with cooling.

**Evaporation rates**

The evaporation response to manipulation was characteristically different than the other hydrometeorological traits. Mann–Whitney test results suggest there was not a significant statistical change in observed and predicted post-manipulation evaporation rates. However, autocorrelation was detected within the observed-predicted residual time series at frequencies of 4 d, 10 d, 11 d, and 13 d. The Watson et al. disturbance test statistic, which addressed this autocorrelation, for evaporation equaled 0.06; that is 6% of the values of Mr exceeded 1.96σ(Mr). These results imply there was a subtle change in some daily evaporation rates with the reduction of inflows to Lake 626. Evaporation disturbances were almost normally distributed with both kurtosis and skewness near zero (κ = 0.85; γc = –0.06). Unlike with DOC concentrations, light extinction and surface water temperatures, there were disturbance values both above + 1.96σ(Mr) and below – 1.96σ(Mr) (Fig. 6) suggesting that there were instances when the manipulation enhanced or suppressed daily evaporation rates. The disturbance time series from 2014 (Fig. 7) illustrates that when there is confidence that evaporation was suppressed by the manipulation, it tended to be either early in the open water season or associated with lake heating events (e.g., 21 July 2014 and 28 September 2014). As well, the largest positive disturbances associated with enhanced evaporation were in conjunction with lake cooling events in August and September, implying the response to manipulation was a subtle change in how the lake responded to episodic events through the open-water season. The more transparent Lake 626 permitted more energy into the lake (Table 4) vs. supporting the latent heat flux, but this heat was released later in the season. The manipulation effect on annual evaporation totals was negligible (Table 4).

**Discussion**

**Lake topology and hydrological connectivity**

Extreme climate events and uncertainty regarding future precipitation regimes may result in changes to the frequency and duration of hydrological connectivity between lakes and their stream inputs. This whole-catchment manipulation provides one of the first attempts to explicitly examine the hydrometeorological responses of a lake to a loss of hydrological connectivity. A change in, or loss of, hydrological connectivity across space or time can be in response to either relative location (topology) or climate or physical disturbance. The observed changes in Lake 626 demonstrate the cascading effect of DOC concentrations on lake clarity, surface water temperatures and evaporation rates in response to upstream flow manipulation, and highlight that lake location within the watershed may influence lake hydrometeorology. The results suggest that boreal lakes with less hydrological connectivity with their watersheds, that is, those subject to lower streamflow inputs relative to their regional counterparts, may exhibit lower DOC concentrations, which is supported by research in other boreal regions (Buffam et al. 2007). Pace and Cole (2002) observed that DOC patterns in lakes with and without sustained flows were not correlated, further supporting the idea that topology or relative location along the stream network is an important control on the DOC regime of a boreal lake. However, this cannot necessarily be extrapolated across hydroclimatic regions of the boreal, as both Zhang et al. (2010) and Seekell et al. (2014) found significant inter-regional differences in DOC patterns across the Canadian and Swedish boreal forests and Temnerud et al. (2010) observed in Sweden that lower order boreal streams tended to have higher total organic carbon concentrations than higher order...
streams. If there are lower concentrations in headwaters, these would be relative to a regional average, because DOC concentrations are known to be controlled by land cover (e.g., catchment wetland fraction) and climate (e.g., precipitation) (Gergel et al. 1999; Buffam et al. 2007).

One of the expected outcomes of our catchment manipulation was that DOC concentrations would decrease in the experimental lake, which would in turn result in increased water clarity (Fig. 1). This association has been documented before in studies evaluating long-term observations (Schindler et al. 1996) and synoptic sampling of boreal lakes (Fee et al. 1996). While these studies support the link between DOC, water clarity and epilimnion depth, our study is the first to demonstrate the relatively cooler surface temperature of a less hydrologically connected lake. The impact of hydrological connectivity on the variability of evaporation was subtle in our study, but does not necessarily imply that landscape position is unimportant. Previous studies have highlighted the role of lake size on the seasonality of evaporation (Oswald and Rouse 2004; Rouse et al. 2005) and the role wind can play on the thermal structure and evaporation rates of larger lakes (Blanken et al. 2000; Keller 2007). Our findings demonstrate that the energy budget response of the boreal lake complex to a loss of hydrological connectivity will depend on the distribution of not just lake size but on the relative distribution within the drainage network of those headwater lakes that receive only intermittent inputs to those higher order lakes that receive perennial streamflow.

While the evaporation response to the manipulation was only just measureable, upscaling of that response across headwaters, which are the vast majority of most stream networks (Bishop et al. 2008), may not be inconsequential to the regional response to a shift in moisture conditions.

This interaction between lakes and their watersheds needs to be captured in how lakes are parameterized in land surface schemes; particularly the latter’s influence on light extinction properties. Schemes to simulate small lake interactions with the atmosphere are only just beginning to become incorporated into coupled land-atmosphere models. Research is required to determine if turbulent flux exchanges in these models are sensitive to parameterization of light extinction. The diversity in light extinction among boreal lakes suggests that one value may not be adequate to simulate the variety of responses that may exist. A more appropriate approach may be to apply a probability distribution function to light extinction parameters, which could be informed by information on the diversity of boreal lake DOC properties (Gergel et al. 1999; Guerrero et al. 2017).

Dry conditions and hydrological connectivity
This experiment was designed to determine the impacts and consequences of a loss of hydrological connectivity on lake thermal conditions and energy budgets. Annual evaporation volumes from Lake 626 and Lake 373, as with previously documented boreal lakes, are primarily dictated by incoming energy and the length of the open water period.
Over the past four decades, the ice-free period has increased by \( \Delta D = 19 \) d for lakes at the IISD-ELA (Guzzo and Blanchfield 2017), with regional predictions of 10, 17, and 29 more ice-free days in the 2020s, 2050s, and 2080s, respectively (Keller 2007). Applying the evaporation rates documented in this study, extended ice-free seasons would result in an additional 35 mm, 60 mm, or 100 mm of seasonal evaporation. This impact is far greater at seasonal scales than any changes in daily evaporation rates that may cascade from a loss in hydrological connectivity. The dry conditions caused by the removal of the connection between Lake 626 and its upper basin resulted in no more than slight changes in how lake evaporation responded to episodic events of a time scale of days to weeks. Previous research suggests that the major effects of changes in hydrological connectivity on small boreal lakes will be indirect with water transport and chemistry being the major determinants of transparency (Keller 2007; Benoy et al. 2007; Stasko et al. 2012). The results here support these hypotheses in regards to DOC, transparency, thermocline depth and surface temperatures, but not seasonal evaporation.

As with previous lake thermal structure experiments (Lyderson et al. 2008; Forsius et al. 2010; Verta et al. 2010), the stress associated with this manipulation exceeds any natural short term stress associated with a loss in hydrological connectivity. This manipulation simulated the complete disappearance of upstream inflow for 4 yr. The longest no-flow period from Lake 239, the longest recorded reference lake at the IISD-ELA, is 278 d, with an average no-flow period of 68 d. The chances of a no-flow period lasting 347 d, 300 d, 234 d, 184 d, and 53 d are 1%, 2%, 5%, 10%, and 50%, respectively. This demonstrates that a year-long no-flow period would be rare, but statistically possible under current climate conditions. It can be concluded with some confidence that the influence of a loss of hydrological connectivity on lake evaporation and atmospheric-lake feedbacks is minor. While not the focus of this study, it is apparent from Fig. 3c that increasing transparency will deepen the thermocline enough to encroach on the hypoxic zone, which may have serious consequences for cold-adapted organisms that require well-oxygenated water, such as Mysis and lake trout (Salvelinus namaycush) (e.g., Plumb and Blanchfield 2009; Guzzo and Blanchfield 2017).

Shifts in vegetation, wetland fraction, rates of DOC production, runoff rates relative to that production, and lake residence time control baseline and future DOC loading into boreal lakes. Perhaps the most important characteristic controlling lake response to dynamic hydrological connectivity is the current baseline chemical regime. Observations by Gunn et al. (2002) demonstrate the very non-linear nature of the relationship between DOC concentration and water clarity, with a significant increase in slope at \( \sim 1-2 \) mg L\(^{-1}\). Those lakes with low baseline DOC concentrations will be the most vulnerable to stressors associated with climate. In lakes with low DOC concentrations or concentrations just above this threshold, extreme changes in hydrological

![Fig. 8. Relationship between open water period and annual evaporation for Lakes 626 and 373. Relationships are significant with 95% confidence and exhibit \( r^2 \) values of 0.76.](image-url)
Connectivity and lake energy regimes

Conclusions

Four hypotheses were derived for this study. The first, the loss of hydrological connectivity invoked by the experiment would result in lower DOC concentrations in the boreal lake, was confirmed. The second hypothesis was that this would decrease light extinction, which was also confirmed. This allowed incoming radiation to penetrate to deeper depths, leading to a warmer water column. Our third hypothesis—that reduced light extinction will result in lower DOC concentrations in the boreal lakes—was confirmed. The second hypothesis was that this would result in lower DOC concentrations in the boreal lakes (Gergel et al. 1999; Hudson et al. 2003; Stasko et al. 2012). In that respect, since most current published predictions of future precipitation regimes do not include such characteristics, the information is not available to assess how boreal lakes will respond to this particular climate change stressor.

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