Eddy covariance measurements of evaporation from Great Slave Lake, Northwest Territories, Canada

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Abstract. The first direct measurements of evaporation from a large high-latitude lake, Great Slave Lake, Northwest Territories, Canada, were made using eddy covariance between July 24 and September 10, 1997, and June 22 and September 26, 1998. The main body of the lake was ice-free between June 20 and December 13, 1997, and June 1, 1998, and January 8, 1999, with the extended ice-free season in 1997–1998 coinciding with 4°C above normal air temperatures and an abnormally strong El Niño. Measurements extending roughly 5.0 to 8.5 km across the lake were made from a small rock outcrop located near the main body of the lake. The lake was thermally stratified between mid-July and September, with the thermocline extending down to approximately 15 m. High winds were effective in mixing warm surface waters downward and, when accompanied by cold fronts, resulted in large, episodic evaporation events typically lasting 45 hours. The daily total evaporation was best described as a function of the product of the horizontal wind speed and vapor pressure difference between the water surface and atmosphere. Seasonally, the latent heat flux was initially negative (directed toward the surface) followed by a steady increase to positive values (directed away from the surface) shortly after ice breakup. The latent heat flux then remained positive for the remainder of the ice-free period, decreasing midsummer and then steadily increasing until freeze-up. The sensible heat flux was small and often negative most of the spring and summer yet switched to positive and began to increase in the early fall. Extrapolation of evaporation measurements for the entire ice-free periods gave totals of 386 and 485 mm in 1997 and 1998–1999, respectively.

1. Introduction

It is generally accepted that high latitudes are sensitive to climate change [Houghton et al., 1996] and that the hydrologic cycle is an integral part of climate [Chahine, 1992]. Global warming through increasing atmospheric CO2 concentrations is expected to result in an increase in evaporation (E) thus promoting warming via the water vapor–temperature feedback. There are, however, few if any direct observations of such an increase in E at high latitudes, which would have ecological and climatological significance at regional and global scales. Changes in E from high-latitude lakes should be detectable with climatic perturbations since open water E is uncomplicated by terrestrial vegetation’s adaptive physiological responses. Large, deep lakes would provide a robust indicator of climate sensitivity given the stable integration of the annual climate through the large surface area and volume of water. Since no long-term E measurements exist and as an alternative to using predictive models [e.g., Small et al., 1999], we directly measured lake E from Great Slave Lake, Northwest Territories, Canada, the deepest (614 m) and fifth largest (in terms of surface area, 27,200 km2) lake in North America. Measurements were made before and after the occurrence of the 1997–1998 El Niño, allowing the effects of the coinciding abnormally high air temperatures to be examined.

Several studies have found evidence of climate warming or variability on lakes located at lower latitudes. Schindler et al. [1990] found boreal air and lake temperatures increased by 2°C between 1969 and 1989, which lengthened the ice-free period by 3 weeks, decreased water renewal rates, and increased chemical concentrations overall, resulting in a summer loss of habitat for cold, stenothermic organisms. Anderson et al. [1996] found that the ice phenology records for 20 Wisconsin lakes from 1968 to 1988 exhibited a trend toward earlier breakup. They also found average breakup dates occurred significantly earlier during El Niño events. Global patterns in lake ice phenology have been used to decipher past climate variations [e.g., Palecki and Barry, 1986; Vavrus et al., 1996], and dates of freeze-up and thaw have been successfully numerically modeled [e.g., Walsh et al., 1998]. Modeling of lake response to climatic forcing has been performed usually using the model of Hostetler and Barlein [1990] either coupled to a regional climate model [e.g., Hostetler et al., 1993] for use in, for example, assessing the ability to simulate the hydrology of the Aral Sea...
Small et al., 1999] or to investigate the hydrologic response of central North American Lakes [Hostetler and Giorgi, 1995] or the Southern (Laurentian) Great Lakes to doubling atmospheric CO$_2$ [Croley et al., 1998].

This research represents the first of its kind ever conducted on Great Slave Lake or any high-latitude lake of this size, and forms part of the Global Energy and Water Cycle Experiment (GEWEX), specifically in the Mackenzie River Basin Study [Stewart et al., 1998]. To our knowledge, the only physical limnological research on this lake was reported in a thorough monograph by Rawson [1950], which includes instantaneous water temperature profiles at various locations around the lake. Measurements (air temperature and humidity and wind speed and direction) to improve weather forecasting and aid navigation on the lake have been made at the island research site by Environment Canada and also from a moored buoy located near Hay River operated during the ice-free period by the Canadian Coast Guard [see Hudson, 1991].

2. Site Description

The research site was a small rock outcrop located in the main body of the lake (Inner Whalebacks, 61.92°N, 113.73°W) 70 km southeast of Yellowknife, 150 km northeast of Hay River (Figure 1). Despite significant logistical problems the site was chosen based on its location near the main body of the lake (i.e., west of 113°W) with surrounding waters' depths matching the lake's main body mean depth of 41 m [Rawson, 1950]. Maximum depth in the main lake is 163 m and is 614 m in the East Arm, which is considered a physically distinct unit from the main lake [Rawson, 1950]. Fetch exceeded 12 km in all directions. The island's approximate height above the mean water surface, width, and length are 10, 100, and 180 m, respectively. Eddy covariance, radiation, and water temperature profile measurements were made between July 24 and September 10, 1997, and June 22 and September 26, 1998 (August 11, 1998, for radiation and water temperature measurements). Ancillary meteorological measurements were made continuously.

Rawson [1950] reported the following general thermal and mixing properties of the main body of the lake. Thermal stratification occurred between mid-July and late August, with the mid-August thermocline between 10- and 18-m depths. Summer-time water input from the Slave River entering at the south shore was equivalent to roughly 0.6% of the volume of the main lake, with enough heat to raise the temperature of the main lake by only 0.6°C. Vernal circulation occurred around July 1, partial circulation occurred around September 1, and complete circulation probably occurred between October 1 and freeze-up in late December. The total summer heat income (amount of heat taken into the lake) was between 0.63 and 0.80 GJ m$^{-2}$.

Coinciding with an abnormally strong El Niño event [McPhaden, 1999], the mean annual 1998 air temperature measured in Yellowknife was 4°C above the 1961-1990 mean (1.5°C in 1997), equal in 1998 to the anticipated CO$_2$-induced air tem-

Figure 1. Location of the Inner Whalebacks Island study site on Great Slave Lake, Northwest Territories, Canada, 70 km southeast of the City of Yellowknife (YK), 150 km northeast of Hay River (HR).
temperature increase at this latitude [Washington and Mehl, 1996] (Figure 2). Using special sensor passive microwave imager (SSM/I) satellite data, it was determined that the main body of the lake was ice-free between June 20 and December 13, 1997, and June 1, 1998, and January 8, 1999, a 45-day extension (A. Walker, personal communication, 1999). The main body of the lake is typically ice-free sometime between June 1–15 and December 15 to January 1 [Rawson, 1950].

3. Materials and Methods

The latent heat flux (\( \lambda E \)) (divided by the latent heat of vaporization \( \lambda \) to give \( E \) the equivalent millimeters of evaporated water) and sensible heat flux (\( H \)) were directly measured with the Mk2 Hydra eddy covariance system. Turbulent fluxes away from the surface are represented as positive values. A brief description of the system is provided here since a full description is provided by Shuttleworth et al. [1988]. A pair of vertically oriented ultrasonic transducers with 200-mm separation measured the vertical wind speed. Absolute humidity was measured horizontally and centered between the ultrasonic transducers. A fast-response chromal-constantan 38 \( ^\circ \)C thermocouple mounted horizontally and centered between the ultrasonic transducers measured air temperature, and a fast-response cup anemometer mounted above the ultrasonic transducers measured horizontal wind speed. All Hydra measurements were made at a frequency of 20 Hz, with hourly average fluxes corrected for sensor separation and frequency response [Moore, 1986]. The Hydra was mounted on a telescopic pneumatic mast positioned near the center of the island, yet away from a meteorological tower, at a height of 6.9 m above the ground surface, with 150-mA power supplied by two 12-V batteries charged by two solar panels. The Hydra ran successfully throughout both field seasons; it was visited on a monthly basis to change the data storage module and to ensure that it was operating satisfactorily. Prior to field use, the Hydra was operated over a short-grass surface for a period of 10 days. During this period the cumulative daily sum of \( \lambda E \) plus \( H \) plotted against the cumulative daily sum of net radiation (\( R_n \))

To make radiation measurements directly over the water
surface avoiding contamination from the island, a 14-m-long retractable horizontal boom, 2 m above the mean water surface, was constructed on the eastern side of the island. Lake \( R_w \) was measured with a net radiometer (model NR Lite, Kipp and Zonen, Delft, Netherlands) at the end of the horizontal boom. Immediately prior to installation, the net radiometer was factory calibrated and was found to underestimate \( R_w \) measured by a Middleton model CN1 over a short-grass surface by 3%. Calculation of the net radiometer’s view factor indicated that 95% of the instrument’s signal received from the water surface originated from a circle with a radius of 8.7 m, well beyond the shore of the island. Acoustic bathymetric measurements made by boat around the vicinity of the net radiometer revealed a mean water depth of 9 m. Water surface or “skin” (0–10 \( \mu \)m depth) temperature \( (T_0) \) was measured with an infrared thermometer (model 4000.GL, Everest Interscience, Tucson, Arizona) with a 15° field of view, mounted 45° from vertical, also at the end of the horizontal boom. This configuration resulted in a water surface viewing area of 4.6 m². This instrument was also factory calibrated immediately prior to field use and is reported to have an accuracy of ±0.5°C. Vapor pressure at the water surface \( (e_v) \) was calculated as the saturation vapor pressure \( [\text{Buck}, 1981] \) at infrared-determined \( T_0 \). Incident \( (R_{i, i}) \) and reflected \( (R_{r, i}) \) solar radiation were measured with separate pyranometers (models PSP and Black and White, respectively, Eppley Laboratories, Newport, Rhode Island), the latter mounted inverted at the end of the horizontal boom. Signals from these instruments were sampled every 5 s and stored as 15-min means, standard deviations, minimums, and maximums, using a data logger (model CR10X, Campbell Scientific Inc. (CSI), Logan, Utah) and a data storage module (model SM192, CSI). Two 12-V batteries connected in parallel charged by a solar panel (model MSX10R, Solarex, Frederick, Maryland) supplied power. Unfortunately, high waves generated by persistent, strong winds destroyed the horizontal boom on August 11, 1998.

Near the center of the island, additional instruments were positioned 8.5 m above the ground (18 m above water surface) supported by a 9-m vertical meteorological tower. Wind speed \( (u_{18}) \) and direction (model 5310, R. M. Young, Traverse City, Michigan) and relative humidity (calculated vapor pressure \( e_{v, 18} \) and temperature \( (T_{18}) \) (model HMP-35D, Vaisala, Helsinki, Finland, with a R. M. Young model 41002 radiation shield) were all measured on this tower. A 12-V lithium battery, with power supplied by a solar panel (model MSX10R, Solarex, Frederick, Maryland) maintained the power requirements of the data logger (model 21X, CSI) which scanned the instruments every 5 s and stored 15-min means, standard deviations, minimums, and maximums on a data storage module (model SM716, CSI). Rainfall was measured with both a tipping bucket rain gauge (model TR-525M, Texas Electronics, Dallas, Texas) and an Environment Canada standard collecting rain gauge, both located 5 m from the meteorological tower. At the end of the 1997 experiment both rain gauges totals agreed within 1 mm.

A water temperature profile \( (T_{w,i}) \) 1 km southwest of the island at depths of 0.7, 5, 10, 15, 20, and 25 m was measured with thermistors and data loggers housed in watertight containers (model WT05-35, Onset Computer Corp., Bourne, Massachusetts). Water temperatures at 35 and 55 m were measured with sensors capable of handling the greater pressures at these depths (model TidBit, Onset Computer Corp.). All thermistors were factory calibrated immediately prior to use, and predeployment and postdeployment intercomparisons revealed a maximum discrepancy of 0.23°C between thermistors spanning a 0°–23°C temperature range. All thermistors were synchronized to sample \( T_w \) every 15 min. The thermistors were cable tied to a marine-grade rope, anchored with a 20-kg anchor, and suspended by two fishing net floats. Global satellite positioning of the profile’s positions both at deployment and retrieval in 1997 indicated that the sensors did not drift. The same persistent high waves that destroyed the horizontal boom in August 1998 also destroyed the water temperature profile; hence all data collected in 1998 were lost.

A fast–finite Fourier transformation of \( T_w \) at each depth revealed spectral peaks (periods) decreased with depth. For example, a period of 24 hours at the 0.7- and 5-m depths decreased to 13.6 hours at the 10- through 25-m depths, then further decreased to 11.1 hours at the 35- and 55-m depths. Clearly, the surface waters (0–5 m) responded to radiative forcing corresponding to the rotational period of the Earth, whereas the 13.6-hour period at the 10- through 25-m depths exactly matched inertia period or half a “pendulum day” at this latitude (the time required for the Foucault pendulum to rotate through 360° = 23.93 hours/sin(61.92°) = 27.12 hours/2 = 13.6 hours). The different physical processes controlling the timescales of energy exchange in the atmosphere compared to water lead to large variations in short-term calculations of the total energy stored in the water therefore requiring long-term means of \( T_w \) to calculate the energy storage.

The total energy stored in the water over a given time interval \( (J_w) \) (W m⁻²) was determined by first calculating a mean water temperature,

\[
\overline{T_w} = \frac{1}{z} \sum \frac{T_w \Delta z_i}{\Delta z},
\]

where \( T_{w,i} \) is the water temperature at sensor \( i \), \( \Delta z_i \) is the depth segment assignment to \( i \) (midpoint between successive levels), and \( z \) is the total water depth at the profile measurement (55 m). Energy storage in the water was then calculated as

\[
J_w = \rho_v c_p \Delta \overline{T_w} \Delta z,
\]

where \( \rho_v \) and \( c_p \) are the density and specific heat of water, respectively, and \( \Delta \overline{T_w} \) is the (mean) water temperature change over the period \( t_2 - t_1 \) (Δt). The accuracy of the turbulent flux measurements was assessed by plotting the cumulative 24-hour mean \( \lambda E \) plus \( H \) against the cumulative 24-hour mean available energy, \( R_{n,i} - J_{w,i} \). A linear regression forced through the origin had a slope of 0.96, indicating the turbulent flux measurements underestimated the available energy by 4% over the 49-day 1997 period when \( R_{n,i} - J_{w,i} \) measurements were available. Logistics forced the \( R_{n,i} \) and \( J_{w,i} \) measurements to be at different locations, the former measured over water 9 m deep and the latter measured over water 55 m deep. The impact on the available energy measurement, however, was negligible, first, because the thorough mixing of surface waters resulted in minimal discrepancy between the skin and 0.7 m depth water temperatures at the two locations (linear regression slope, \( y \) intercept, and \( r \) were 0.96, 0.88°C, and 0.96, respectively, based on 0.5-hour mean \( T_{w,i} \) (independent) and \( T_w \) at 0.7-m depth (dependent)) and, second, because typically only 1% of the incident light penetrates to the 9-m depth [Rawson, 1950].
4. Results and Discussion

Water temperature isotherms (Plate 1) show that the 1997 thermal stratification period was short, beginning mid-July and starting to end by September. Despite this short period the thermocline, as represented by the position of the 10°C isotherm, extended down to approximately 15 m. Periods of high winds, common on such a large lake, effectively mixed warmer surface waters downward (Figure 4). For example, the quiescent period in early July 1998 allowed the water surface to warm to 19°C (July 11, 1998). Increasing winds, from an average of 2 (July 12, 1998) to 18 m s⁻¹ (July 13, 1998), mixed the waters, decreasing \( T_0 \) from 19°C to 5°C in just over 3 days (77.5 hours). Another example of such a wind-driven mixing event occurred in late August 1997, when \( T_0 \) dropped from 17°C (August 29) to 9°C (August 31) over 39 hours, corresponding to an increase in \( u_{18} \) from 3 (August 29) to 15 m s⁻¹, gusting to 18 m s⁻¹ (August 31) over 16 hours.

The impact of these high-wind events and their associated synoptic meteorological conditions on \( E \) is shown in Figure 5, representing a typical episodic \( E \) event. Under clear skies a cold front advanced from the north on August 6, 1997. The wind direction completely rotated through 360° clockwise from north and back to north again, and \( u_{18} \) steadily increased. The front passed over the site midday on August 7 as indicated by cloudy skies, strong winds, an 8°C drop in \( T_{18} \), and a 1-kPa drop in \( e_{18} \). Evaporation at this time increased dramatically, propelled by the sudden increase in \( \Delta e (\Delta e = e_0 - e_{18}) \) and the high \( u_{18} \). Evaporation remained significant but steadily decreased over the 36 hours after the passage of the cold, dry air as \( T_{18} \) and \( e_{18} \) gradually recovered and \( u_{18} \) decreased. Note the small \( E \) before the passage of the cold front and large \( E \) after, even when comparing cloud-free days. The dominance

![Figure 4](image-url)

**Figure 4.** Water surface temperature \( (T_0) \) integrated over a depth of approximately 10 μm measured with an infrared thermometer and horizontal wind speed at a height of 18 m above water surface \( (u_{18}) \), both plotted as 15-min means. Dashed and solid curves represent 1997 and 1998 measurements, respectively.
of these episodic $E$ events throughout the ice-free seasons was revealed by a spectral analysis (fast-finite Fourier transformation) of the hourly $E$. For both years a prominent spectral peak was revealed at a period of 45 hours, a measure of the dominant timescale for $E$. In fact, 50% of the total measured $E$ occurred over only 25% and 20% of the days (1997 and 1998, respectively).

The importance of the variables shown in Figure 5 on the magnitude of $E$ is further explored in Figure 6. The daily mean $R_n$ was negatively and poorly correlated with $E$, as overcast days were often associated with a large $\Delta e$ and a high $u_{18}$. The relationships between $E$ versus $u_{18}$ and $E$ versus $\Delta e$ were both positive and statistically significant (student’s $t$ test, $\alpha = 0.05$), with a greater proportion of the variance in $E$ explained by $\Delta e$ than by $u_{18}$. The variables $\Delta e$ and $u_{18}$, however, were not independent as periods of high winds were usually accompanied by cold, dry air. Therefore most of the variance in $E$ was explained by the product $u_{18}\Delta e$.

The observed seasonal patterns in $E$ and the associated energy balance components are shown in Figure 7. Net radiation (Figure 7a) decreased away from the summer solstice, and overcast days often were those with the highest $E$ (Figure 7b), as those days were often associated with the passage of cold, dry air over the lake. The energy required to fuel $E$ likely came from the release of stored energy in the lake. In June 1998, fog and condensation ($\lambda E$, $0$) occurred when the vapor pressure in the warmer atmosphere exceeded that of the cold (near 4°C) lake surface. Only after $T_0$ warmed above 4°C (approximately July 1, 1998, Figure 4), which occurred roughly 30 days after the June 1 date of complete ice breakup, did $\lambda E$ become

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**Figure 5.** Time series of 15-min mean (a) incident solar ($R_s$) (bold curve) and net radiation ($R_n$) (fine curve), (b) Horizontal wind speed at 18 m above water surface ($u_{18}$) (bold curve), wind direction (fine curve), (c) Air temperature ($T_{18}$) (bold curve), water surface temperature ($T_0$) (fine curve), (d) vapor pressure at 18 m ($e_{18}$) (bold curve), vapor pressure at water surface ($e_0$) (fine curve), and hourly mean latent heat flux ($\lambda E$) or equivalent evaporation rate ($E$) during August 5–9, 1997. This period represents a typical episodic (36–72 hours) evaporation event following the passage of a cold front (midday August 7).

**Figure 6.** Relationships between daily (24 hours UT-7 hours) 1997 (solid circles) and 1998 (open circles) total evaporation ($E$) and mean (a) net radiation ($R_n$), (b) horizontal wind speed at 18 m above water surface ($u_{18}$), (c) vapor pressure difference between the water surface and 18 m ($\Delta e = e_0 - e_{18}$), and (d) the product $u_{18}\Delta e$. Lines are linear regressions given by the equations on each graph.

**Figure 7.** Daily (24 hours UT-7 hours) mean (fine curves) (a) net radiation ($R_n$), (b) latent heat flux ($\lambda E$) and equivalent $E$, (c) sensible heat flux ($H$), and (d) the residual of the energy balance ($e = R_n - \lambda E - H$) indicative of energy storage in the water. To allow direct comparisons between years, time scales are the same for both years. Seasonal trends are represented by a 10-day running mean (bold curves).
positive. Between July 1 and mid-July 1998, \( E \) steadily increased as relatively calm winds allowed strong surface warming (hence a large \( e_o \)) to occur. The atmosphere at this time was still relatively cool and dry; hence \( \Delta e \) was large. Episodic surface cooling due to wind-driven mixing of surface water coupled with a warmer, more humid atmosphere acted to weaken the vapor pressure difference and therefore suppressed \( E \) between mid-July and mid-August 1998. Throughout much of the spring and summer, \( H \) was often directed toward the lake surface (Figure 7c), as the air at 18 m was warmer than that near the cold water surface. This pattern reversed in the fall as the air temperature gradient reversed; thus the atmosphere warmed at the expense of the surface. The residual of the energy balance (\( e = R_n - \lambda E - H \)), indicative of the energy stored in the water, indicated the lake was an energy sink (positive \( e \)) throughout most of the spring and summer, likely switching to an energy source (negative \( e \)) in the fall and early winter (Figure 7d).

When \( E \) was not directly measured (i.e., after September 10 and 26 in 1997 and 1998, respectively), it was estimated as a function of the measured \( u_1 \), \( \Delta e \) (see Figure 6d). The standard error of the estimated \( E \) from this linear regression was 0.7041 mm d\(^{-1}\), and the critical value of the two-tailed \( t \) distribution with \( \alpha = 0.05 \) and \( n - 2 = 105 \) days is 1.983. Hence the error associated with estimating \( E \) from the linear regression shown in Figure 6d is \( \pm(1.983)(0.7041) = \pm 1.396 \) mm d\(^{-1}\) [Zar, 1984]. When \( T_o \) was unavailable (i.e., after September 10, 1997, and August 11, 1998), the \( e_o \) required to calculate \( \Delta e \) was estimated from a linear regression of the daily mean \( e_o \) against the daily mean \( e_{18} \) corresponding to a range in \( T_o \) from 4.1° to 20.7°C (Figure 8). This introduces additional uncertainty in the estimation of \( E \) since this relationship may not hold late in the season when vapor pressures and \( T_o \) fall below the range of the linear regression. Late-season measurements, however (e.g., October 3, 1999, \( e_{18} = 0.29 \) kPa, \( T_{18} = -2°C, T_o = 4°C, \) and \( e_o = 0.82 \) kPa), initially confirmed a predicted \( e_o \) of 0.82 kPa calculated from the linear regression shown in Figure 8, giving some confidence to late-season estimates. The cumulative \( E \) calculated using the regressions shown in Figures 6d and 8 was 5% above (1997) and 8% below (1998) the measured \( E \) during the same periods. This gives an indication of the minimum error involved in using this estimation procedure as the estimation parameters were derived from the same data to which the comparisons were made.

This method of estimating \( E \) applies Fick’s diffusion law in finite difference form (\( E = K_E u_{18} (e_o - e_j) \), where \( K_E \) is eddy diffusivity for water vapor) and was empirical only in so far as that \( K_E \) (a function of instrument heights \( z \) and location, size, and shape of the lake) was determined statistically for this experiment. This method is often referred to as the “mass transfer equation/method” and has been used extensively for many lakes and reservoirs [see Harbeck, 1962; Quinn and den Hartog, 1981; Spahrt and Ruddy, 1983; Winter et al., 1995]. More physically based methods of calculating open water \( E \) such as the Penman [1948] equation, however, all require \( J_w \), which is problematic to calculate at a daily time step in such a large lake [see Stannard and Rosenberry, 1991].

The annual course of the daily total \( E \) for both years (Figure 9) shows a secondary peak shortly after ice breakup, followed by steady decrease through the late summer, followed by a steady increase until ice cover was complete. The reduction of \( E \) following the ice breakup maximum is not observed over the Southern (Laurentian) Great Lakes [Scherzer, 1997] and may stem from faster humidifying of a shallower convective boundary layer at higher latitudes, acting to decrease \( E \) through the negative feedback between \( E \) and the saturation deficit.

The annual cumulative \( E \) for both years extending over the entire ice-free season (Figure 10) shows a total of 386 and 485 mm in 1997 and 1998–1999, respectively. The maximum probable (\( \alpha = 0.05 \)) ranges associated with these estimates based on the linear regression shown in Figure 6d were determined by adding, then subtracting, the maximum error (1.396 mm d\(^{-1}\)) to the daily estimated total \( E \), then cumulating these results.
Figure 10. Cumulative evaporation (E) for 1997 (fine curve) and 1998–1999 (bold curve). Extrapolation to annual E totals of 386 ± 127 mm (1997) and 485 ± 144 mm (1998–1999) beyond the measurement periods (as indicated by circles with vertical lines) was accomplished using the linear regression shown in Figure 6d based on measured vapor pressure and wind speed 18 m above water level. Vapor pressure at the water surface (eₚ) was estimated using the linear regression shown in Figure 8, because (except for the initial four days of 1997) water surface temperature was not available when this estimate was being used.

individually. This calculation gave a range of 386 mm ± 127 mm (33%) for 1997 and 485 mm ± 144 mm (30%) for 1998–1999. The 1997 total E falls at the upper end of the 300–400 mm yr⁻¹ estimated for lakes at this latitude by Phillips [1990], whereas the 1998–1999 total approached the 500–750 mm yr⁻¹ (long-term estimated annual total determined using the mass transfer equation) for the Southern (Laurentian) Great Lakes [Scherertz, 1997].

The cumulative E compare well for both years between mid-August and mid-November, implying that the difference in the cumulative totals stems from the longer 1998–1999 ice-free period. With the 19-day earlier ice breakup in 1998, 29 mm of water already evaporated by the time E was positive in 1997. The start of the evaporation season was captured in both years, yet the end of the season extended beyond December 12 in 1998 as aerial inspection of the site and surrounding area revealed that the main body of the lake was still ice-free, with shore ice present in bays and around islands. The SSM/I passive microwave satellite data confirmed complete freeze-up did not occur until January 8, 1999 (A. Walker, personal communication, 1999). Thus the 26-day later freeze-up in 1998–1999 compared to 1997 allowed for an additional 70 mm of evaporative water loss. The dates of ice formation were more significant than the dates of ice breakup, as the daily E rates late in the year were double those at the beginning of the year.

5. Summary and Conclusions

We described the first direct eddy covariance measurements of evaporation from Great Slave Lake during segments of the 1997 and 1998 ice-free period. Within the footprint of these measurements, which extended roughly 5.0 to 8.5 km across the lake, the mid-July to September thermal stratification period and thermocline, which extended down to approximately 15-m depth, were sometimes disturbed by periods of high winds. These periods of high winds, when accompanied by cold, dry air (e.g., passage of a cold front), resulted in episodic evaporative water losses. These periods usually lasted 45 hours, and 50% of the evaporative water loss occurred over only 25% and 20% of the days (1997 and 1998, respectively). The daily (24 hours) total evaporation was best described as a function of the product of the horizontal wind speed and the difference between the vapor pressure at the water surface and atmosphere. As measurement of the short-term energy storage on the lake was problematic, the mass transfer equation was used to estimate the total evaporative water loss during both the 1997 and 1998–1999 ice-free periods. A 4°C above normal air temperature and a 45-day extended ice-free period, coincidental with the 1997–1998 El Niño, resulted in an estimated 485 mm total evaporative water loss in 1998–1999 compared to 386 mm in 1997.

The adjacent terrestrial environment may be affected by changes in the lake's energy and water cycles. Although the distance inland that may be influenced by the lake is unknown, even boreal lakes much smaller than Great Slave Lake have a significant effect on the adjacent terrestrial areas [Onley et al., 1997; Sun et al., 1997]. Areas around large lakes tend to experience fewer clouds, lake breezes, and reduced sensible heat fluxes in the summer. Although more moisture is advedt into the Mackenzie basin from the east (Pacific air) than leaves the basin to the east, summertime precipitation is convective-driven; hence local evaporation is an important atmospheric moisture source [Gyakum, 1997; Strong et al., 1997]. In the fall and early winter the lake acts as an important moisture and heat source at a time when terrestrial sources for both are minimal. One may speculate that late-fall moisture release from this lake, if it results in lake-effect snow, would result in wetter terrestrial springtime conditions.

Changes in the magnitude and seasonality of evaporation from the lake, driven by changes in interannual meteorological conditions, could also have anthropogenic impacts in the Mackenzie basin. The Mackenzie River (twelfth longest river in the world), with Great Slave Lake as its origin, serves as the primary transportation route for supplying northern communities with commodities from the south. Changes in the lake's discharge, either by changes in ice phenology or increased evaporative losses not matched by increased precipitation could have an economic impact on those who rely on the river as a transportation route. Changes and variability in ice phenology may impact downstream flooding, as Great Slave Lake may be thawed while the north flowing Mackenzie River is still frozen farther north. The local fishing and tourist industries may be adversely affected if the aquatic ecosystem is adversely affected by a decrease in water renewal rates. This may occur if an increase in lake evaporation is not matched by precipitation and runoff inputs into the lake.

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