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Evaporation from a Small Lake in the Continental Arctic using Multiple Methods

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Daily evaporation from a small lake in the continental Low Arctic of Canada was examined using three independent experimental methods and a simplified combination model. Mean daily lake evaporation (± variability between methods) was estimated to be $3.2^{+0.3}_{-0.3}$ mm d^{-1} and 2.5 $^{+0.6}_{-0.3}$ mmd^{-1} over fifty-day periods during two consecutive summers. Based on these results and additional class-A pan data, total thaw-season evaporation estimates of 220 mm to 320 mm were obtained, equivalent to 70% to 100% of annual precipitation. These values are 15 to 70% higher than predicted by standard evaporation maps of Canada. Our results indicate that the Priestley-Taylor model provides a good approximation of the Bowen ratio energy balance model in this setting. As expected, estimates based on mass balance are highly sensitive to uncertainty in measurement of lake inflow and outflow.

Introduction

Vast areas of the continental Arctic of Canada are characterized by low topographic relief, poorly integrated drainage, continuous permafrost and abundant shallow lakes of less than one hundred hectares. The hydrological regime of this terrain has been described as low-arctic nival, distinguished by long, cold winters when processes are relatively dormant, brief spring snowmelt periods when high streamflow and widespread flooding are accentuated by the presence of frozen ground, and short summers with low streamflow punctuated by infrequent rainstorm-generated peaks (Woo 1990). Although hydrology of the region remains poorly known, field observations suggest that lakes typically undergo large seasonal volume fluctuations due to the combined influence of their shallow depth, extreme spring flooding and intense evaporation during arid summers. Inter-annual lake volume fluctuations may also be considerable since thaw season evaporation is similar in magnitude to annual precipitation, as shown for other northern environments (Latham 1988; Bigras 1990). Recent expansion of gold, base metals and diamond mining activities in the region, and associated development of artificially-altered lakes for storage and treatment of tailings effluent, has prompted recent interest in lake hydrology. Specific practical applications, such as prediction and control of tailings pond levels, consideration of dam stability, and use of shallow water caps to isolate potentally acid-generating tailings requires a sophisticated understanding of lake-water balance and evaporative processes.

This study focuses on the Lupin mine site (65° 45'N, 111° 15'W), situated on the west shore of Contwoyto Lake, Northwest Territories, Canada approximately 375 km NNE of Yellowknife, and 125 km beyond northern treeline, where evaporation is estimated from a small lake (Fig. 1). The site is centred within an area of several thousand km^2 where no lake evaporation studies have been conducted and where use of basic climatological techniques such as the CRWE model (complementary relationship wet-surface evaporation model, Morton 1983) has not been possible due to the short instrumental period-of-record and lack of basic observations such as sunshine duration. In addition to data requirements for management of tailings ponds, evaporation data are also being compiled to assess the validity of isotope-mass balance methods for estimating evaporation which were applied concurrently but are beyond the scope of this article (Gibson 1996; Gibson *et al.*1993, 1994).

Application of several methods has provided a unique opportunity to examine and intercompare standard experimental methods, including mass balance, energy balance and class-A pans to the Priestley-Taylor model and the aerodynamic profile method. A critical concern with experimental observations in this setting is the potential for horizontal advection of latent and sensible heat due to the small surface area of lakes and the stark contrast between the dry tundra landscape and adjacent lake surfaces.

Study Area

The study site is typical of the region southwest of Contwoyto Lake and east of treeline (Fig. 1 b). Mean annual precipitation based on records from 1982 to 1993 at the Lupin Airport weather station is 289 mm about 45% as snow. Mean annual temperature is -12°C and the range in monthly temperature is 45°C. Ice break-up on lakes begins as daytime temperatures rise above 0°C in late-May. The ice-free season on small lakes typically extends from mid-June to late-September. Spring surface drainage from tributary catchments occurs long before ice break-up on Contwoyto

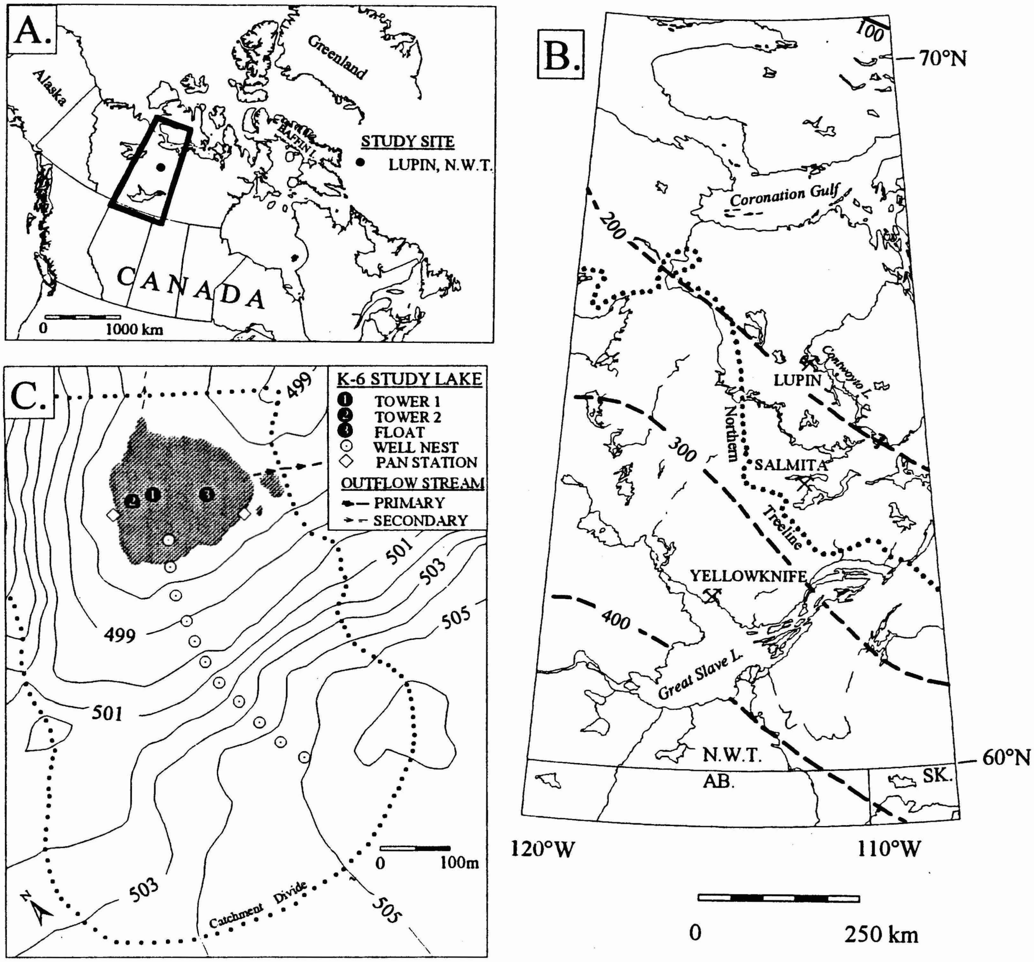

Fig. 1. (a) Map of Canada showing the study area. Bold outline denotes area enlarged in (b). Map of central Northwest Territories showing the Lupin Mine, Yellowknife (nearest climate station with sunshine record), northern treeline (dotted line) and mean small lake evaporation contours (redrawn from DEN Hartog and Ferguson 1978); and, (c) 1 m contour map of the study lake and catchment area showing instrument locations and outflow streams.

Lake which is delayed due to its considerable volume. Ice cover on Contwoyto Lake commonly persists until mid-August and freeze-up begins again in late-October. Drainage from Contwoyto Lake occurs through the Burnside River and Back River, which discharge into the Arctic Ocean.

The lake selected for detailed study has an approximate surface area of 4 ha and a catchment area of 27 ha (Fig. 1c). The site was chosen based on its representativeness, manageable size, minimal outflow, and proximity to a road. Dwarf tundra vegetation is common on well-drained slopes and uplands, with grass and sedge dominating low areas and the lakeshore. Active-layer development is restricted to depths ranging from 0.5 to 2 m above continuous permafrost extending to a depth of several hundred metres. Bedrock outcrop is common above 500 m.a.s.l. within the catchment and consists mainly of low-grade phyllite, argillite and slate. The lake has a mean depth of 0.65m and is characterized by a peripheral shelf where depth of water is less than 1.2 m, and a deep, central hole where depth reaches a maximum of 2.5 m. The peripheral shelf accounts for about two-thirds of the lake area and is marked by rib and trough features (Shilts and Dean 1975), and rounded cobbles and boulders mantled by relatively minor accumulations of organic sediments. The deep, central portion of the lake is dominated by relatively thick deposits of organic sediments. Surface outflow is intermittent.

Evaporation monitoring was limited to the months of July and August during the ice-free season. Temperature, rainfall and snowfall during the study months were typical of average conditions (Table 1). Rainfall, however, was lower than average in July 1992 and higher in July 1993.

Table 1 – Climatic conditions during the study compared to monthly normals†

	Mean Daily Temperature (°C)			Total Rainfall (mm)			Total Snowfall (mm)		
	1992	1993	Normal	1992	1993	Normal	1992	1993	Normal
July	11.0	10.4	10.8 (1.5)	14.8	87.0	46.7 (24.6)	0	0	0.6 (1.3)
August	8.2	9.2	8.5 (2.4)	41.6	25.8	49.4 (25.5)	5.4	3.0	3.7 (6.7)

† Based on monthly observations for 1982 to 1993 at the Lupin airport AES weather station. Values in brackets denote 1 standard deviation from the normal.

Field Methods

Hillslope Monitoring

Ten two-point well nests, installed by the method outlined in Lee and Cherry (1978) were used to monitor water tables along a representative transect of the catchment hillslope (Well Nest, Fig. 1c). Depth to frozen ground was determined along the well transect at 30 locations every 7 to 14 days by frost-probing (*i.e.* by manually hammering 10 mm-diameter re-bar vertically into the ground until impenetrable frost is encountered). Depth profiles of the active-layer till were sampled, stored in air-tight plastic bags and returned for laboratory analysis of grain size and moisture content by standard methods (Freeze and Cherry 1979, p. 335). Mean porosity and unfrozen hydraulic conductivity were estimated to be 0.23 and 1.8×10^{-6} m s^{-1} respectively.

Bail testing of the wells (Freeze and Cherry 1979, p. 339) suggested a slightly lower *in situ* hydraulic conductivity of 7.8×10^{-7} m s^{-1} likely related to cementation of the till by liquefaction processes. An inventory of mean active-layer storage capacity and mean water storage was compiled from frost table measurements and well measurements to provide a record of subsurface runoff potential. Surface runoff, when it occurred, was measured periodically from a representative hillslope plot using a plastic bag or bucket and volumetric cylinder.

Outflow Monitoring

Ten water table wells were installed along the main drainage pathway to estimate potential for subsurface outflow. Frost probing at 30 points was also carried out at 14 day intervals along this transect. The high specific retention of the dense, compacted peat material, combined with a shallow active layer averaging 0.5 m suggested that subsurface flow was small. Surface outflow, when it occurred was manually gauged using the method outlined for hillslope runoff.

Automated Instrumentation

Automated sensors were mounted on two towers and a float apparatus within the lake (Fig. 1c) and configured to provide necessary parameters for estimation of evaporation using mass balance, Bowen ratio energy balance, the Priestley-Taylor method, and a two-point aerodynamic profile method. Campbell Scientific 21X data loggers were programmed to read sensor outputs at 10 second intervals and record hourly mean values.

Lake water level was measured using a Druck model PDCR 950 pressure transducer verified with the independent record of a UDG01 ultrasonic depth gauge and manual leveling surveys. Initial lake volume and surface area was estimated from a computerized theodolite survey of the lake bathymetry and surrounding topography. Precipitation was measured using a TE 525M tipping bucket rain gauge.

Energy balance sensors mounted on Tower 2 consisted of a Middleton net radiometer aspirated with nitrogen gas and an infrared radiometer recording lake surface temperature. Water temperatures at 0.30 m intervals throughout the water column were measured using Model 107 thermistors suspended from a separate float apparatus anchored near the centre of the lake. Bed thermistors were manually installed by a dry-suit diver at the water-bed interface and at a depth of 0.30 m below the interface to provide representative temperature profiles in both deep and shallow areas of the lake. Daily morning inspections were conducted to ensure that the net radiometer was positioned normal to the lake surface and free from obstruction by dew and hoar frost. Isopropyl alcohol was used to clean polypropylene radiometer domes as required.

Aerodynamic profile sensors were mounted on Tower 1 near the mid-point between the centre and edge of the lake. Air temperature and relative humidity were measured at two heights above the water surface using Vaisala HMP35CF probes housed in Gill radiation shields. Wind speed at identical heights was measured using 3-cup micro-response contact anemometers. These sensors were initially set at 0.45 m and 1.8 m above the surface and periodically lowered as water levels declined. Interim heights were determined by compensating for water level change. Wind vector and barometric pressure were measured using an R.M. Young wind monitor and a Sentra SBP270 sensor respectively. Class-A pans were positioned near the lakeshore and maintained daily using standard methods.

Evaporation Calculations

Mass Balance

The daily volume of lake evaporation E was determined as the residual of the water mass balance using the relationship

$$E = P + I - O - \Delta S \tag{1}$$

where P is daily total volume of precipitation falling on the lake surface, I is volume of lateral surface and subsurface inflow, O is volume of surface and subsurface outflow, and ΔS is the volumetric change in lake storage. P is determined based on the daily depth of recorded rainfall multiplied by mean lake surface area and ΔS is derived from lake water level records converted to volumes using a rating curve. I and O are estimated from records of hillslope storage surplus and lake stage respectively, calibrated through manual flow gauging.

Energy Balance

Evaporation was estimated using the Bowen ratio energy balance (BREB) method based on the energy balance of the lake surface layer given by

$$Q^* = Q_H + Q_E + Q_G \tag{2}$$

where Q^* is net radiation, Q_H is the sensible heat flux, Q_E is the latent heat flux, and Q_G is the subsurface heat flux. Rearranging Eq. (2), and substituting $\rho_W L_V E$ for Q_E, the evaporation flux is given by

$$E = \frac{Q^* - Q_G}{(1+\beta)\,\rho_W L_V} \tag{3}$$

where E is evaporation, L_V is latent heat of vapourization of water, ρ_W is the density of water, $\beta = C_p \Delta T / L_V \Delta q$ is the Bowen ratio, C_p is the heat capacity of air at a given temperature, ΔT and Δq are the vertical temperature and specific humidity gradients respectively, and Q_G is the subsurface heat flux given by

$$Q_G = Q_{GL} + Q_{GB} \tag{4}$$

The lake heat storage Q_{GL} was evaluated over time t for each horizontal water slice of volume Δv and mean surface area Δa and summed according to

$$Q_{GL} = C_W \sum \frac{(\Delta T/\Delta t)\Delta v}{\Delta a} \tag{5}$$

where C_W is the heat capacity of water, and ΔT is the temperature change over time step Δt. Heat conduction into the lake bed Q_{GB} is determined for each time step from

$$Q_{GB} = K_B(T_2 - T_1) \tag{6}$$

where K_B is the bed thermal conductivity estimated from measurements of bed grain size (Farouki 1981), T_1 is the temperature at the water-bed interface and T_2 is the bed temperature at a depth of 0.3 m below the interface. Q_{GB} is evaluated separately for shallow and deep portions of the lake bed using separate temperature profiles at representative points, and areally weighted.

Hourly Bowen ratio data were evaluated using the rejection criteria outlined in Ohmura (1982a). Of 2400 hours-of-record only 119 were excluded, with resulting gaps appearing for up to several hours during evening and early morning hours when evaporation is small. Mean daily BREB estimates were therefore computed from available hourly data within each 0000-2400 hour MLT averaging period using simple interpolation to account for missing values.

Priestley-Taylor Model

Evaporation was also estimated using the Priestley-Taylor combination model, a technique similar to the BREB method but commonly used in northern Canada for operational simplicity to eliminate the need for boundary layer profiles of temperature and specific humidity. It is expressed as

$$E = \alpha\frac{S}{S+\gamma}\;\frac{Q^*-Q_G}{\rho_W L_V} \tag{7}$$

where α is an empirically-derived evaporability factor, S is the slope of the saturation vapour pressure-air temperature curve, and γ is the psychrometric constant. For the model calculations presented in this paper α is taken as 1.26 as suggested by Priestley and Taylor (1972) for describing mean evaporation from large saturated land surfaces and 'advection-free' water surfaces. Although these conditions are certainly not met in the present study due to the small size of the study lake, several previous studies in northern Canada have found that the use of $\alpha \approx 1.26 \pm 0.1$ is satisfactory for examining evaporation from saturated surfaces including small lakes over periods of weeks or months (Rouse 1990). In studies where BREB estimates have also been available this simplification has been tested by combining Eqs. (3) and (7) to obtain an expression for α given by

$$\alpha = \frac{S+\gamma}{S(1+\beta)} \tag{8}$$

This expression is used in the present study to determine mean daily values for α and thereby examine and assess the validity of $\alpha = 1.26$.

Aerodynamic Profile

As demonstrated in a northern setting by Ohmura (1982b) and others, evaporation may be calculated by measurement of humidity, wind speed and temperature gradients in the logarithmic sublayer. For a two-point profile, the evaporation rate is determined as

$$E = -\frac{\rho_A \kappa^2}{\phi_M \phi_W} \frac{(u_2 - u_1)(q_2 - q_1)}{(\ln z_2/z_1)^2} \tag{9}$$

where ρ_A is the density of air, κ is the VON Kármán constant 0.4, u is hourly-mean wind speed, q is the hourly-mean specific humidity, z is the measurement height above the water surface, subscripts 1 and 2 referring to upper and lower levels respectively, and ϕ_M and ϕ_W are stability corrections for momentum and water vapour respectively. Stability corrections were determined by analysis of gradient Richardson numbers using the method outlined in Monteith and Unsworth (1990, p. 238).

It has been shown that closely spaced measurements near the surface are required over small northern lakes to ensure that readings are taken within the fully adjusted boundary layer (Bello and Smith 1990). In the present study instruments were situated to optimize fetch given the lake geometry, prevailing wind direction, and logistical constraints (Tower 1, Fig 1c). Although fetch to Tower 1 was limited for winds originating from the near-shore direction (220-270° designated as Group A), such conditions occurred only 15% of the time or 351 of 2400 hours-of-record. For the remaining 85% of the time, fetch to Tower 1 exceeded 90 m wherein the upper sensors (≤ 1.8 m in height) fell within a boundary layer defined by a height-to-fetch ratio of about 0.02 (Group B). For estimation of daily evaporation rates Group A results were therefore not included. Instead, evaporation was estimated from available Group B hours within each 24-hour averaging period, once again interpolating for missing values.

Results

Mass Balance

Evaporation from the study lake was calculated using Eq. (1) for 54 days during 1992 and 71 days during 1993. Mass balance components are summarized in Table 2. Evaporation resulted in a steady decline in lake level during 1992 with only brief rainfall-generated peaks (Fig. 2). Evaporation for the observation period, which extended for about 50% of the ice-free season, totalled 181 mm (3.4 mm d^{-1}). Only minor surface outflow from the lake was observed during early July when lake levels were above the zero discharge elevation of the main drainage stream (closure elevation 1, Fig. 2). Outflow however was likely sustained at increased rates throughout the preceding snowmelt period prior to initiation of field monitoring. Lateral inflow

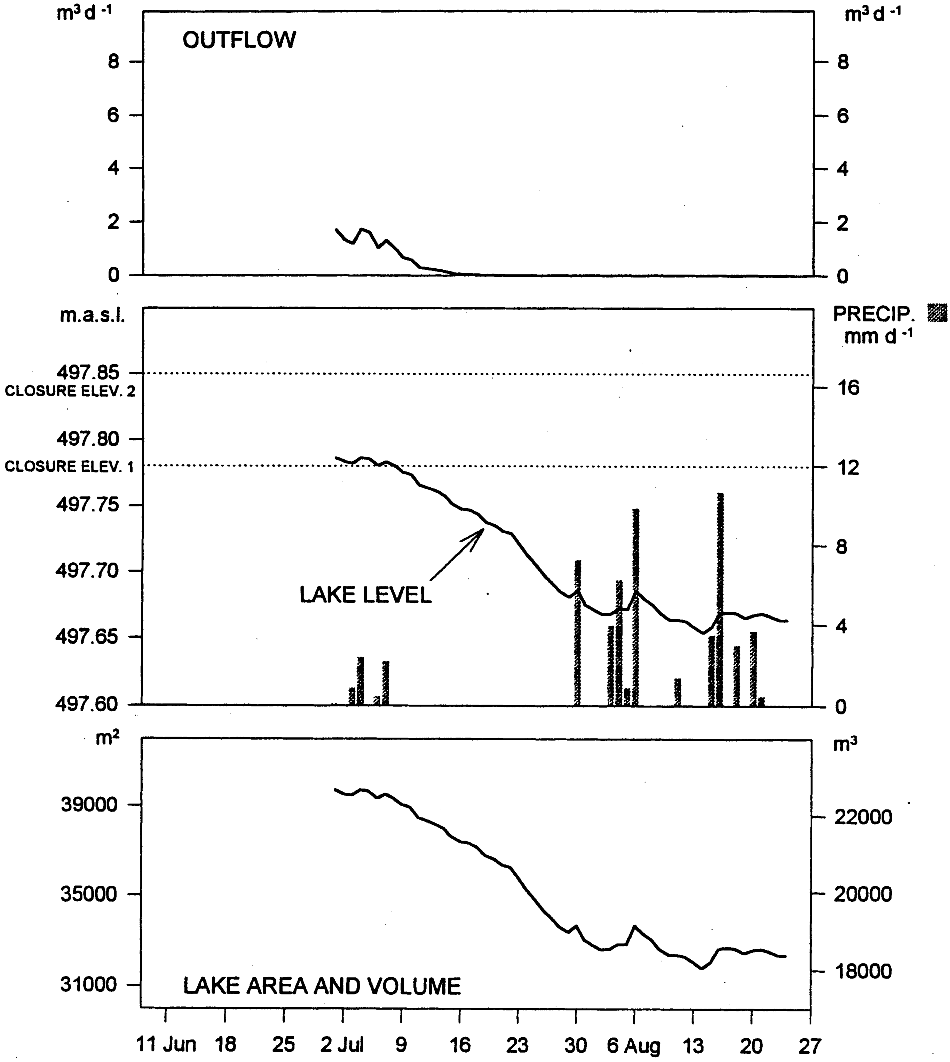

Fig. 2. 1992 Daily Mass Balance Summary. i) outflow; ii) mean lake level and total precipitation; and iii) mean lake area and volume. Note that closure elevation 1 and 2 refer to maximum lake level for zero discharge in the main and secondary surface outflow streams respectively (Fig. 1c).

from the catchment area was negligible as a result of unsaturated conditions in the hillslope active layer, which allowed recharge of incident precipitation. Due to the high roughness of the frost table, water derived from active-layer melting was largely detained in isolated pockets on the hillslope or released by evaporation through the unsaturated zone. Measured hillslope storage capacity exceeded water storage

Table 2 – Mass balance summary for the study lake. ΔS is change in lake storage, I is lateral inflow, P is precipitation on the lake surface, O is surface outflow, and E is evaporation.

		Total			INPUT		OUTPUT	
Year	Period	No. Days	Ice-free Days†	ΔS mm	I mm	P mm	O mm	E mm
1992	2 July-24 August	54	110	124	0	57	0.3	181
1993	22 June-24 August	71	105	68	39	112	40	180
Total		125	215	192	39	169	40	361

† based on observations made by mine personnel.

throughout the observation period (Fig. 3). Negligible input except via precipitation directly on the lake surface provided exceptional control for determination of evaporation rates. The magnitude and rapid response time of the lake hydrograph during precipitation events confirms that input is occurring only in the proximity of the lake.

Instrumentation was installed in the study lake shortly after ice break-up in 1993 and prior to complete ablation of the snowpack. Lake levels exceeded the observed 1992 maximum throughout the 71 day observation period such that surface outflow occurred continuously through the main drainage stream, and in addition, via a secondary streamlet at the time of peak snowmelt (closure elevation 2, Fig. 4). Although lateral groundwater inflow was small and the hillslope active-layer was un-

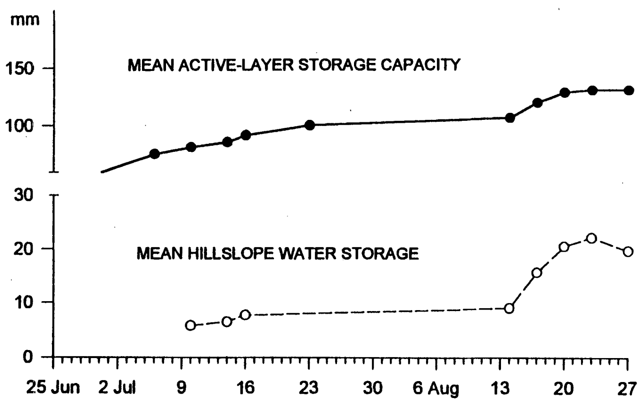

Fig. 3. Mean hillslope water storage and storage capacity curves for 1992 based on monitoring of the frost table (30 locations) and water table (10 locations). Note that active-layer storage capacity = mean porosity × frost depth. Hillslope water storage = mean porosity × (mean frost depth – mean water table depth) or depth of precipitation when active-layer storage is exceeded. As storage capacity was not exceeded no surface runoff occurred.

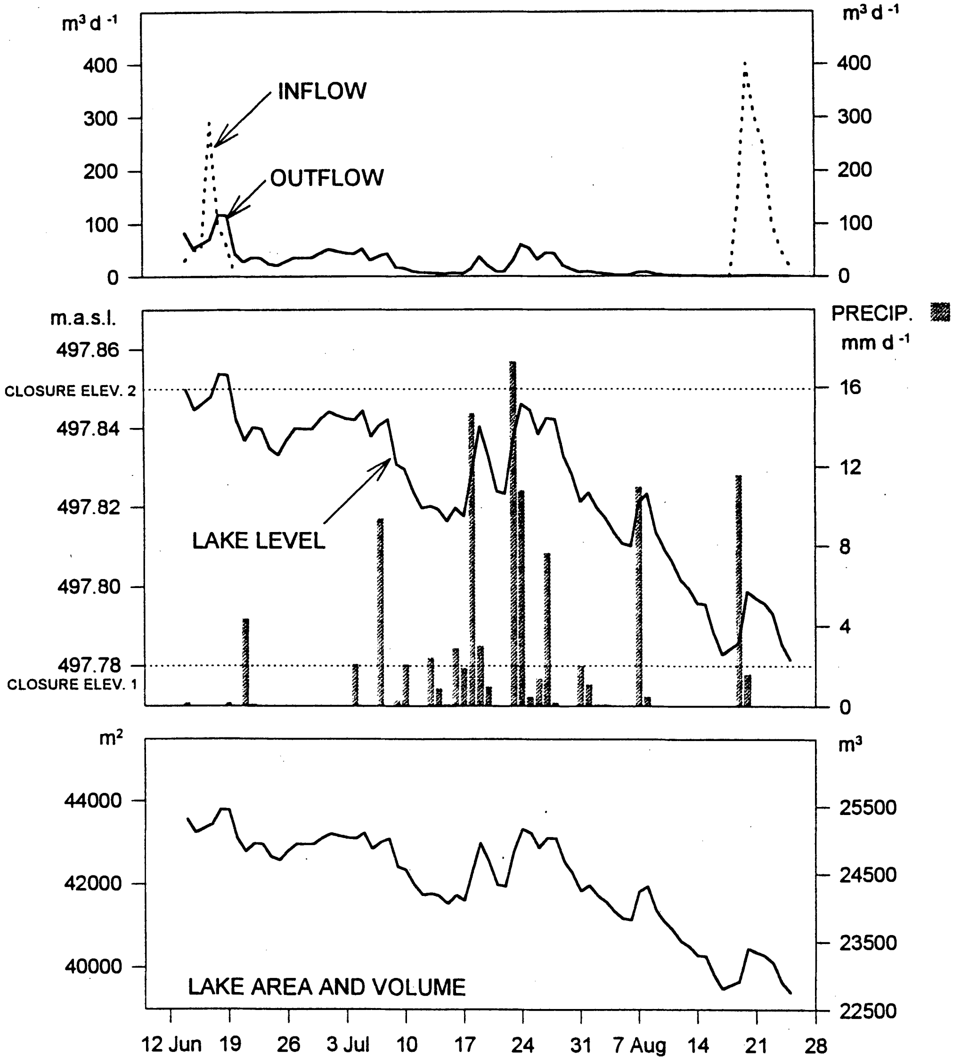

Fig. 4. 1993 Daily Mass Balance Summary. i) lateral inflow and outflow, ii) mean lake level and total precipitation, iii) mean lake area and volume. Closure elevation 1 and 2 are defined in Fig. 2.

saturated for most of the thaw season, storage capacity was exceeded both during the snowmelt period and in late summer resulting in inflow via overland pathways (Fig. 5). Evaporation for the total observation period, which extended for about 70% of the ice-free season, was estimated to be 180 mm (2.5 mm d^{-1}). For the shorter comparative interval of 46 days estimated evaporation was 141 mm (3.1 mm d^{-1}). This interval was approximately 44% of the ice-free season for 1993.

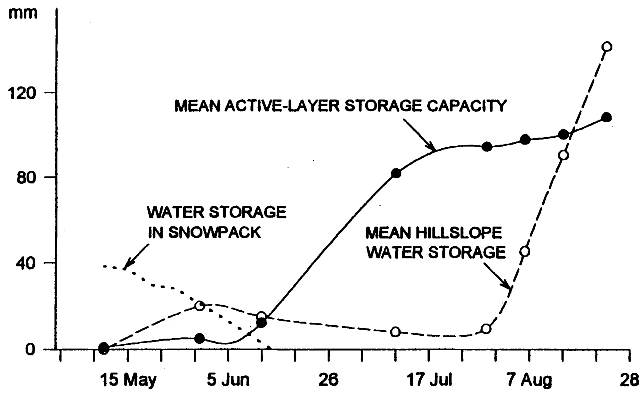

Fig. 5. Mean hillslope water storage and storage capacity curves for 1993 based on monitoring of the frost table (30 locations) and water table (10 locations). Hillslope water storage and active-layer storage capacity are defined in Fig. 3. Surface runoff occurred when active-layer storage capacity was exceeded during snowmelt and in the late summer.

Energy Balance

Evaporation was calculated using the BREB method (Eq. (3)) over 54 days during 1992 and 46 days during 1993. Estimated evaporation for the two respective years is 154 mm (2.9 mm d^{-1}) and 111 mm (2.4 mm d^{-1}). From analysis of the energy partitioning for each year (Table 3), 58 to 64% of total net radiation energy was directed to latent heating, 28 to 30% was directed to sensible heating and 3% to 9% was dissipated through heating of the lake bed. Lake heat storage, which fluctuated as the lake warmed and cooled in response to air temperatures, approached zero over 2 to 4 day periods. Overall, Q_{GL} was a net energy sink during the observation interval for both years, accounting for roughly 6% of Q^*. E for 1992 was approximately 70% of E for 1993 due mainly to proportionately lower Q^*. Daily evaporation varied from

Table 3 – Bowen Ratio Energy Balance (BREB) summary for the study lake. Q^* is net radiation, Q_{GL} is lake heat storage, Q_{GB} is heat flux through the lake bed, Q_H is sensible heat flux, and Q_E is latent heat flux.

Year	Period	No. Days	Q^* MJ × m^{-2}	% of Q^*: Q_{GL}	Q_{GB}	Q_H	Q_E	Evaporation mm
1992	3 July-24 August	54	+599	6	9	28	58	154
1993	10 July-25 August	46	+391	6	3	30	64	111
Total		100	+990	-	-	-	-	265
Mean		-	-	6	6	29	61	-

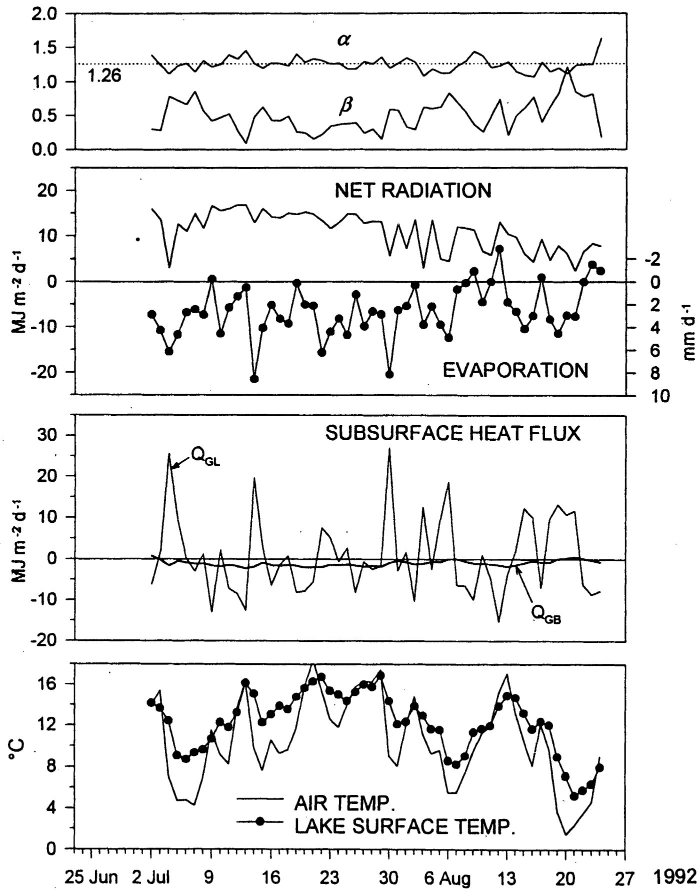

Fig. 6. 1992 Energy Partitioning Summary. i) Mean daily Bowen ratio β determined from boundary layer profiles of temperature and specific humidity, and mean daily evaporability factor α derived using Eq. (9), *i.e.* as the ratio of BREB and equilibrium latent heat fluxes. Note that α = 1.26 is shown for reference. ii) Daily total net radiation measured over the lake surface and mean daily BREB evaporation; iii) mean daily subsurface heat fluxes, namely the heat storage in the lake Q_{GL} and heat conduction into the lake bed Q_{GB}; and iv) mean daily air temperature and lake surface temperature.

about -2 to 8 mm with no strong seasonal trends (Figs. 6b ,7b). Daily peaks in E are commonly associated with positive peaks in Q_{GL} (Figs. 6c, 7c) and often lag about a day or so behind peaks in Q^* (Figs. 6b, 7b). Such a cycle reflects solar heating of the lake followed by release of stored heat through Q_E and Q_H. Q_{GB}, in contrast, is relatively insensitive to changes in Q^* and remains mainly small and negative throughout the summer. As expected, less-negative Q_{GB} values are observed when lake heat storage declines as both lake and air temperatures drop (Figs. 6d ,7d).

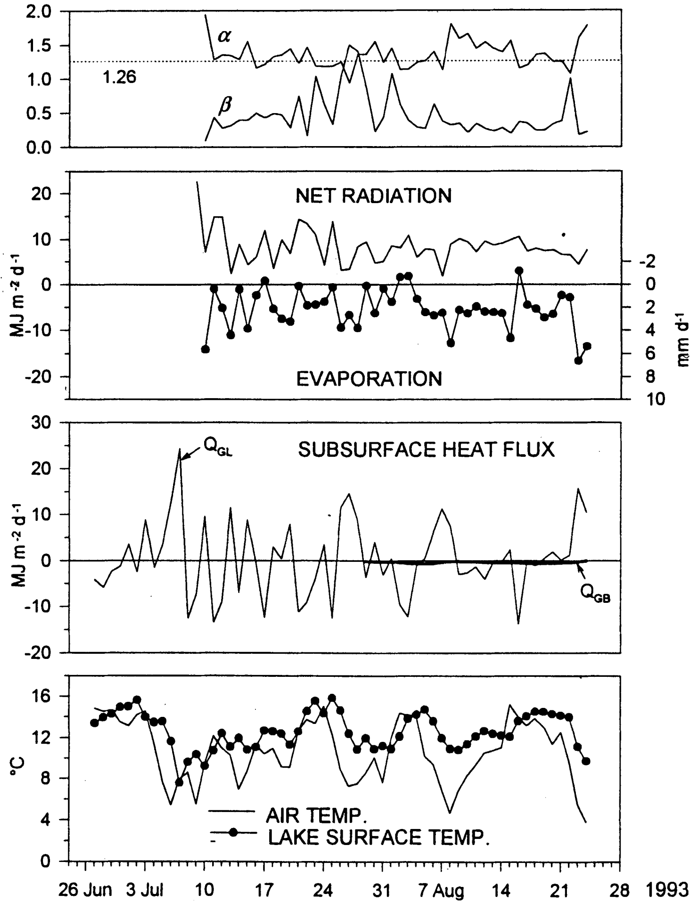

Fig. 7. 1993 Energy Partitioning Summary. i) Mean daily Bowen ratio β determined from boundary layer profiles of temperature and specific humidity, and mean daily evaporability factor α derived using Eq. (9), *i.e.* as the ratio of BREB and equilibrium latent heat fluxes. Note that α = 1.26 is shown for reference. ii) Daily total net radiation measured over the lake surface and mean daily BREB evaporation; iii) mean daily subsurface heat fluxes, namely the heat storage in the lake Q_{GL} and heat conduction into the lake bed Q_{GB}; and iv) mean daily air temperature and lake surface temperature.

Daily Bowen ratios ranged from 0.1 to 1.5 with very similar mean values of 0.49 and 0.47 derived for 1992 and 1993 (Figs. 6a, 7a). Mean values are slightly higher than β = 0.35 and 0.38 measured during mid-summer over a small lake and wet sedge tundra in the Hudson Bay Lowlands (Rouse *et al.* 1977), and daily ranges are slightly larger than β = 0.14 to 0.86 measured for a saturated wetland near Baker Lake, N.W.T. (64°27'N, 97°47'W) (Roulet and Woo 1986a). As anticipated for a saturated surface with no resistance to evaporation, daily variability in β is fully

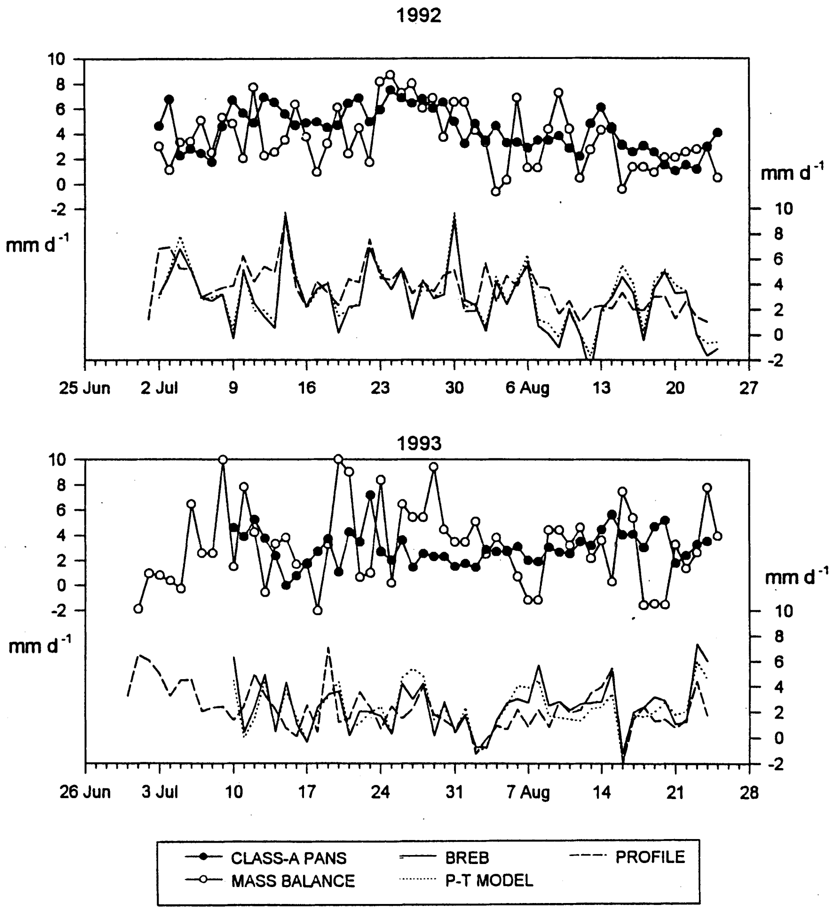

Fig. 8. Time-series of daily lake evaporation based on the specified methods.

linked to variations in the moisture and air temperature regime. A good positive correlation exists between both precipitation occurrence and amount (Figs. 2 and 4) and β (Figs. 6 and 7), primarily as a result of a reduction in the atmospheric moisture deficit Δq, but also due to increases in air temperature gradients ΔT during rainy and cool periods. Q_H is commonly directed towards the lake surface during short periods at night, causing negative hourly values of β. This occurred on all but 28 of 100 days of record.

For the comparative intervals, evaporation calculated as a residual of the mass balance was 18% and 27% greater than that calculated from the BREB method (Tables 4 and 5). Daily evaporation trends are broadly consistent (Fig. 8), however correlation is somewhat diminished for 1993 evidently due to errors in measurement of inflow and outflow during dynamic periods.

Table 4 – Lake evaporation for the 54-day period 3 July-25 August 1992

Method	Evaporation mm	mm d^{-1}	% rel. to BREB
Mass Balance	181	3.4	118
BREB	154	2.9	100
Priestley-Taylor	171	3.2	111
Profile	190	3.5	123
Mean	174	3.2	113

Table 5 – Lake evaporation for the 46-day period 10 July-24 August 1993

Method	Evaporation mm	mm d^{-1}	% rel. to BREB
Mass Balance	141	3.1	127
BREB	111	2.4	100
Priestley-Taylor	103	2.2	93
Profile	112	2.4	100
Mean	117	2.5	105

Priestley-Taylor Method

Evaporation was also derived using Eq. (7) over equivalent periods. Results obtained were very similar to the BREB method due to mutual dependence on the parameters Q^*, Q_{GL}, and Q_{GB}. Using Eq. (8) the evaporability factor α is calculated by comparison with BREB estimates (Figs. 6a and 7a). Mean values (± one standard deviation) of α were found to be 1.25±0.10 for 1992 with values for individual days ranging from 1.06 to 1.63 (Fig. 6a) and individual hours ranging from 1.0 to 3.5. Similar variability was observed for 1993 (Fig. 7a) with a mean value of 1.36±0.20 calculated for the entire period. A positive correlation is evident between daily Q^* and α (Figs. 6 and 7). Roulet and Woo (1986a) and others have discussed the dependence of α on atmospheric moisture deficit, aerodynamic resistance, and available energy (Q^*-Q_G). As moisture deficit and aerodynamic resistance were similar throughout the study, higher mean values of α during the second year are attributed primarily to lower available energy.

Relative to the BREB method, use of the simplified Priestley-Taylor model led to an 11% overestimation of evaporation for 1992 and a 7% underestimation for 1993 (Tables 4 and 5). Daily time series of the two methods reveals a nearly identical pattern with only minor differences in the magnitude of some day-to-day peaks (Fig. 8). However, as noted for example on 27 July 1993, and the period from 9 August to 15 August 1993, differences are substantial only when extreme values of the Bowen ratio are observed.

Aerodynamic Profile

Mean daily evaporation was also calculated using Eq. (9) based on hourly mean profiles of specific humidity, wind speed, and temperature. Daily evaporation rates ranged from -2 to 8 mm d^{-1}, with mean rates of 3.5 mm d^{-1} for 1992 and 2.4 mm d^{-1} for 1993. Total estimated evaporation for the comparative intervals is 190 mm and 112 mm respectively (Tables 4 and 5). An additional 15 days of data were collected early in 1993 before necessary sensors for some other methods were operational. Over this interval from 30 June to 10 July 1993 profile E averaged 3.0 mm d^{-1}, similar in magnitude but with poor daily correlation to mass balance E. Based on analysis of gradient Richardson numbers the mean flow regime for both years was near-neutral tending slightly to stable atmospheric conditions. The assumption of neutral stability would have caused an overestimation of evaporation by about 4%. Inclusion of Group-A wind data where fetch was limited would have resulted in additional overestimation of evaporation by 7 to 10%. Of interest, an analysis of profile data by the method outlined in Marciano and Harbeck (1954) reveals that the lake is aerodynamically rough most of the time despite its small size. Mean values of upper wind speed, surface roughness (z_0), friction velocity(u_*), and roughness Reynolds number ($u_* z_0/v$) were determined to be 3.5±1.8 m s^{-1}, 0.0005 m, 0.07 m s^{-1}, and 2.5 respectively.

Results compare acceptably with the BREB method. Similar evaporation trends are evident with excellent correspondence between positive day-to-day peaks (Fig. 8). It is apparent however that differences in daily rates occur frequently during days when lake heat storage is strongly negative (*i.e.* when the lake is rapidly warming). For 1992, profile E is 23% greater than BREB E, 5% greater than mass balance E, and 11% greater than Priestley-Taylor E. For 1992, profile and Bowen ratio estimates are less than 1% different, and profile E is 27% less than mass balance E and 7% greater than Priestley-Taylor E.

Uncertainty and Potential Errors

A root-mean-square (RMS) analysis was conducted in order to estimate potential errors in the independent methods for average conditions. Potential error in E calculated from the mass-balance method is likely better than ±7% for 1992 but close to ±30% for 1993 based on conservative estimates of uncertainty in measurement of precipitation (±5%), lake volume (±5%), inflow (±50%) and outflow volumes (±50%). Substantially larger potential error for the latter year solely reflects the high uncertainties associated with estimation of inflow and outflow. Potential error in E calculated from the BREB method is about ±13% based on uncertainty in measurement of Q^* (±10%), Q_G (±15%) and β (±15%). For average conditions, where β is moderate, error was found to be limited mainly by the uncertainty in Q^*. Potential error in E calculated from the aerodynamic method is about ±18% based on estimated uncertainty in measurement of gradients (±15%) and sensor heights (±1%). Similar values have been cited for comparable methods in northern Canada (Weick and Rouse 1991) and elsewhere (Winter 1981).

Pan and Lake Evaporation

Comparison of lake evaporation results with concurrently measured Class-A pan evaporation (Table 6) suggests that use of a constant pan-to-lake coefficient of 0.81 or 0.82 provides a good approximation of study lake evaporation for both years. This differs from the value of 0.70 commonly used in the Northwest Territories to approximate lake evaporation (Reid 1995). For periods of less than a month, however, use of a constant pan-to-lake coefficient does not accurately predict lake evaporation as measured by all other methods, and so should be regarded only as a qualitative indicator of evaporation for days or weeks.

Class-A pan data were also collected at the Lupin Airport weather station during the 1983 and 1984 ice-free seasons. Mean daily rates were similar to the data presented in this study (Table 6). Assuming a pan-to-lake coefficient of 0.81, estimated lake evaporation for 1983 is 260 mm over a period extending the entire ice-free sea-

Table 6 – Class-A Pan Evaporation

Year		Period	No. Days	Pan Evaporation mm	mm d^{-1}
1983†		18 June-15- Sept.	90	324	3.6
1984†		14 June-31 August	79	280	3.5
1992	1.	4 July-25 August	53	223	4.2
	2.	2 July-25 August	55	237	4.3
1993	1.	8 July-24 August	48	137	2.8
	2.	8 July-24 August	48	147	3.1
Total			373	1348	-
Mean					3.6

† data from AES observations at the Lipun airport weather station.

Table 7 – Estimated lake evaporation, precipitation and evaporation to precipitation ratios (E/P) for the indicated years. Precipitation is based on Lupin airport weather records for each water year totalled from October to September.

Year	Total Ice-Free Days	Small Lake Evaporation mm	Precipitation mm	*E/P*
1983	90	260	372	0.70
1984	90	320	341	0.94
1992	110	300	299	1.00
1993	105	220	306	0.72

son on small lakes. For 1984, estimated lake evaporation is 230 mm over a period extending for about 80% of the ice-free season on small lakes. As these observations extend beyond the months of July and August considered in this study, they are also a useful indicator of evaporation in the early and late ice-free season. Incorporating these additional data, total estimated small lake evaporation for 1983, 1984, 1992, and 1993 is 260 mm , 320 mm, 300 mm and 220 mm respectively. These values range in magnitude from 15 to 70% greater than predicted from a standard evaporation map of Canada (Fig. 1b) and range from 70 to 100% of the total annual precipitation recorded in each water year (Table 7). Also of interest, mean annual precipitation is 20% larger than predicted by standard maps compiled before the Lupin weather station was established (Prowse 1990; Fig. 1.5, p. 10) and roughly 22 to 52% larger over the years outlined in Table 7.

Discussion

The continental arctic is distinguished from coastal arctic regions by more extreme annual variations in temperature and humidity, and by increased solar radiation input due to lower lattitude, reduced cloudiness and absence of sea fog. Situated below the Arctic Circle and within 125 km of northern treeline, Lupin has a climate, vegetation, and hydrology similar to portions of central District of Keewatin, Northwest Territories where moderate or strong maritime influences from Hudson Bay are absent. Hydrologic investigations conducted west of Baker Lake (Roulet and Woo 1986a-b) and south of Ferguson Lake (Bursey *et al.* 1991; Gibson *et al.* 1993) have reported comparable rates of evaporation and runoff for the mid-summer period in catchments draining east to Hudson Bay. Ours is nonetheless the first hydrologic field study of a tundra lake situated in the mainland Northwest Territories draining to the Arctic Ocean.

Evaporation during the open-water period is evidently an important water loss component for a small lake of 4 ha and ranges from about 220 to 320 mm a^{-1} or about 70 to 100% of annual precipitation. Due to the small size of the study lake it is evident that evaporation may be augmented by horizontal advection of sensible heat from the surrounding landscape. The derived evaporation rates are therefore more typical of the local lake regime than of conditions in the centre of any large lake where evaporation would be uninfluenced by the surrounding landscape. Considering that profiles for the Bowen ratio and aerodynamic profile method were measured near the mid-point between the centre and edge of the lake, it is likely that measurements adequately reflect the mean advective influence for the study lake as do mass balance measurements which integrate evaporation over the entire lake surface. We can also expect that lake evaporation varies with lake area. Given that the average lake area A in the vicinity is about 8 times that of the study lake (or about 32 ha) we can suggest, based on the empirically-derived relationship proposed by Harbeck

(1962) in which $E \propto A^{-0.05}$, that average small lake evaporation may range from as low as 200 to 290 mm a^{-1}. These estimates are still higher than predicted from the map of DEN Hartog and Ferguson (1978), based mainly on class-A pan data, and much lower than predicted by the map of Morton (1983), based on the CRWE model. Similar discrepancies have been reported for the Mackenzie Delta region (Marsh and Bigras 1988) suggesting a universal need for better calibration of climatological models.

Due to extreme gradients in lake evaporation predicted along a transect extending from Yellowknife to Lupin (≈ 50 mm per 125 km, Fig. 1b), results are likely representative only for very short distances north and south of the site. This is corroborated by parallel studies conducted at the nearby Salmita mine site (Fig. 1b) where summer evaporation from small ponds was found to more or less balance precipitation inputs over several years.

The use of several methods in this study has provided a comparative basis for examining reliability. In general, good agreement is observed between methods, considering estimated potential errors. The most conservative results were obtained for the BREB method, which is independent of atmospheric stability and less susceptible to imperfect fetch conditions than aerodynamic profile methods (Brutsaert 1982, p. 210). Although the profile method did tend to overestimate evaporation, estimates were usually intermediate between the mass balance and BREB methods suggesting that this effect is only minor. Furthermore, application of the profile method was a worthwhile check on the direction and magnitude of the BREB estimates and errors resulting from time lag of inflow when computing the daily mass balance.

The simplified Priestley-Taylor model was not substantially different from the BREB method when examined on time-scales ranging from individual days to periods of 50 days. Daily predictability of the energy balance is demonstrated in a plot of BREB *E versus* Priestley-Taylor *E* (Fig. 9). The high correlation observed is ex-

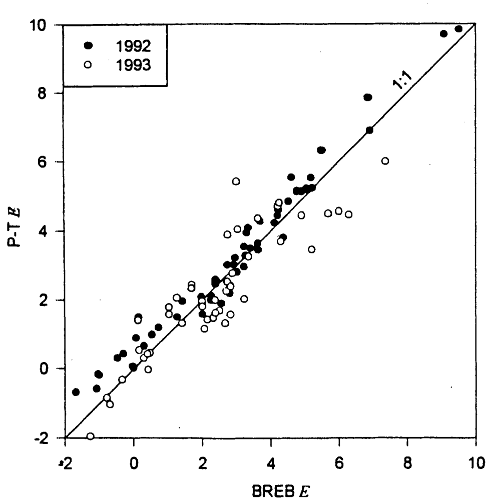

Fig. 9. Plot of mean daily values of BREB *E versus P-T E*.

pected considering that the models are controlled primarily by available energy (Q^*-Q_G). A running mean analysis of the difference between these methods indicates that for averaging periods of 30 days or more there is a 95% probability that estimates are within ±10%. The results show convincingly that the Priestley-Taylor model is a good approximation of the BREB method in the present context. Evaporability factors have been derived using Eq. (8) in a wide variety of studies in northern Canada and results presented herein are not significantly different from results previously reported. It appears that the propensity towards $\alpha \approx 1.26$ for saturated surfaces is mainly a function of the limited range and temperature dependence of β in arctic and subarctic environments. Use of $\alpha \approx 1.26 \pm 0.10$ is equivalent to assuming $\beta \approx 0.98 \pm 0.15$ at 0°C and $\beta \approx 0.32 \pm 0.12$ at 15°C. These values are indeed consistent with other reported observations (Rouse 1990).

The mass-balance method, although prone to substantial errors for even moderate inflow and outflow, may be an appropriate method for monitoring evaporation from tailings ponds with artificially altered catchments where inflow and outflow are controlled.

From the predicted E/P ratios for Lupin (Table 7) it appears that total containment of liquid wastes in closed ponds is not feasible following decommissioning and eventual abandonment of the mine site. One characteristic of the area is that some lakes and depressions selectively catch large amounts of blowing snow during winter causing rapid and unpredictable flooding during spring melt. It may be possible, therefore, to lower input to some degree through careful positioning of snow fences and removal of unfavourable obstacles within the tailings impoundment area.

One fundamental concern with the analysis presented is that evaporation was estimated at a point, though spatial variability in lake evaporation rates over short distances may be significant. Although we suggest and discuss empirical relationships relating lake size and evaporation as derived in temperate climates, this phenomenon remains largely unknown in the Arctic. Direct observations from many lakes, a formidable logistical undertaking using conventional approaches,would be required to fully characterize the range and temporal variability of lake evaporation. This is the primary objective of ongoing investigations using isotope mass balance methods which do not rely on extensive instrumentation.

Summary and Conclusions

Monitoring conducted over a total of one hundred days during two consecutive years is used to characterize the summer mass balance, energy balance, and boundary layer profile of a typical small lake in the continental Arctic. Lake evaporation and mass balance is shown to be very different for a wet year than for a dry year, with evaporation controlled predominantly by net radiation input. Additional class-A pan observations in the early and late summer thaw season are used to predict the total evaporation over the entire ice-free period.

Evaporation is shown to be variable but appears to be consistently less than the annual precipitation total for four typical water years. Consequently, it appears that total containment of liquid wastes in tailings ponds is not a feasible decommissioning strategy for the Lupin operation following cessation of mining activities. Even for an ideal pond with no significant catchment area outflow would likely occur each spring following snowmelt. As demonstrated previously in a wide range of northern environments, the Priestley-Taylor method is shown to be a suitable approximation of the energy balance for a small lake.

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