# DEVELOPMENT AND VALIDATION OF AN ISOTOPIC METHOD FOR ESTIMATING LAKE EVAPORATION

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## ABSTRACT

A study designed to test the validity of an isotopic method for estimating evaporation was conducted within a small, tundra lake situated in the continental Arctic of Canada. Evaporation was determined using an isotope mass balance approach which accounted quantitatively for isotopic fractionation, progressive evaporative enrichment of $\delta^{18}O$ and $\delta^2H$ in lake water, and attenuation of enrichment by liquid inputs and atmospheric moisture. Concurrent determinations made using standard mass balance, energy balance, aerodynamic profile and class A pans permitted rigorous comparisons between methods. Results are presented for two summers which had distinct weather conditions and hydrological balances. Overall, the $\delta^{18}O$ balance was found to be in good agreement with standard methods during both years over time intervals greater than about one week. Owing to a less systematic response of $\delta^2H$ over short time periods, it use is not recommended for quantitative mass balance determinations over time intervals of less than about 50 days in this setting.

KEY WORDS arctic; evaporation; lakes; northern hydrology; stable isotopes, tracers

## INTRODUCTION

Open water evaporation can be an important water loss mechanism in regions with abundant lakes, streams or artificial reservoirs (US Geological Survey, 1954). It is probably very important over vast regions of northern Canada where lakes are numerous, and surface water availability is generally enhanced by low relief, regional occurrence of permafrost and widespread distribution of shallow, crystalline bedrock (Prowse, 1990). The magnitude of evaporation in the region, its climatological significance and relationship to regional hydrology is not well understood, however, owing to limitations in the existing hydroclimatic monitoring network, and the small number of detailed experimental studies that have been conducted (Prowse, 1990).

Use of isotopic tracers has been suggested as a potential method for increasing the scale of experimental studies, for calibration of climate models and for quantifying evaporation loss as required by water development projects in remote, unmonitored locations (Gibson *et al.*, 1993, 1994). A major concern, however, is that the method has not been subject to rigorous comparison with standard and accepted experimental methods.

This study presents a comparison between four standard methods and an isotope mass balance technique, applied concurrently to determine evaporation from a small lake in the continental Arctic. The

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isotopic method was developed specifically for small lakes in this setting where potential exists for highly variable atmospheric conditions and rapid lake volume fluctuations. Such lakes are typical features of seasonally arid Arctic regions (Woo, 1980; FitzGibbon and Dunne, 1981; Woo, 1986; Bigras, 1990). The results demonstrate the sensitivity of small lakes to evaporation during the ice-free period and provide a basis for application of isotopic methods in a wider range of continental environments.

## Background

The stable isotopes of oxygen ($^{16}O$, $^{18}O$) and hydrogen ($^{1}H$, $^{2}H$) occur naturally within the water molecule (i.e. $^{1}H^{1}H^{16}O$, $^{1}H^{1}H^{18}O$, $^{2}H^{1}H^{16}O$) and are commonly used as hydrological tracers owing to their properties of mass conservation and often systematic, measurable variability in water budget components. Isotopic tracing of evaporation relies on the principle that surface waters exposed to the atmosphere, under ordinary conditions, will become enriched in the rare, heavy, water species ($^{1}H^{1}H^{18}O$, $^{2}H^{1}H^{16}O$) relative to the common, light, water species ($^{1}H^{1}H^{16}O$), owing to slight differences in rates of species molecular diffusion through the atmospheric boundary layer (e.g. Merlivat and Coantic, 1975; Gat, 1981a). Isotopic enrichment in surface waters exposed to evaporation has been widely reported, and the degree of enrichment is known to depend on both atmospheric conditions, including relative humidity, temperature and the isotopic composition of atmospheric moisture, and lake water balance parameters, including inflow and outflow (Gat and Bowser, 1991). Several studies have utilized the mean, long-term isotopic enrichment signature of lakes to estimate evaporation rates in the context of steady-state isotope mass balance analysis (i.e. Dinçer, 1968; Gat 1981b; Zuber, 1983; Gonfiantini, 1986; Gibson et al., 1993). Studies of large-scale evaporative recycling have also employed isotopic methods but have relied, conversely, on tracing contributions of isotopically 'light' evaporate to the atmosphere (Leguy et al., 1983; Gat and Matsui, 1991; Ingraham and Taylor, 1991; Gat et al., 1994). This study adopts the former approach, but uses a non-steady-state method in order to account for time variations in water balance and isotopic enrichment observed for a shallow arctic lake. In the following discussion, isotopic composition is expressed as $^{18}O/^{16}O$ and $^{2}H/^{1}H$ ratios, represented by $\delta$ values signifying deviation in parts per thousand (‰) from a designated standard. Values here are cited with respect to SMOW (standard mean ocean water), such that $\delta_{sample} = (R_{sample}/R_{SMOW} - 1)\ 1000$, where $R = {}^{2}H/^{1}H$ or $^{18}O/^{16}O$ ratio measured in sample and standard (Fritz and Fontes, 1980).

## Model development

Steady-state models often poorly describe the isotopic behaviour in small northern lakes because of extreme seasonal changes that occur in the water and isotope balances of these reservoirs (Gibson et al., 1993). Typically, the isotopic composition of a small lake is 'reset' to more depleted values each spring by substantial input of isotopically light snowmelt water. While input of snowmelt water briefly enhances throughflow, it also promotes evaporative enrichment in the subsequent summer period, as the isotopic composition of the lake readjusts to conditions typical of the arid summer climate and evaporation-dominated water balance. Typical summer enrichment in small lakes, shown herein for a lake in the continental Arctic, can be as high as 4·5‰ in $\delta^{18}O$ and 30‰ in $\delta^{2}H$, and comparable in magnitude to changes observed following rare, natural or artificial flooding events in low-latitude regions (Fontes and Gonfiantini, 1967; Zimmerman, 1979). Isotopic changes following such events can provide a high sensitivity record of evaporation, groundwater exchange and outflow conditions as shown for ephemeral lakes in the Sahara (Fontes and Gonfiantini, 1967) and for newly formed artificial reservoirs in Germany (Zimmerman, 1979).

It was important to consider the unique features of the hydrological balances of lakes in subarctic to high arctic environments in order to develop a realistic and conservative model for this setting. Specifically: (1) lakes are frozen and hydrologically inactive for 6–10 months or more during winter; (2) lakes undergo significant flooding and high rates of outflow each spring during a brief, 10-day to two-week snowmelt period (Woo, 1980); (3) lakes are commonly poorly mixed prior to ablation of the lake ice cover (Bergmann and Welch, 1985), but owing to their small volume and shallow depth, often less than 1·5 m deep, tend to be thoroughly mixed with uniform temperature distributions during the open water period (Stewart and Rouse, 1976; Bellow and Smith, 1990); (4) the midsummer period is typically arid, and lake volume

commonly declines in response to high evaporation rates, low to negligible outflow rates and low to moderate rates of lateral inflow from the shallow active layer (Gibson et al., 1996); (5) deep groundwater interaction is negligible in most areas owing to the presence of thick permafrost; (6) fluctuations in meteorological and hydrological conditions are common during the open water period as a result of the effect of cold air intrusions, and summer rainfall or snowfall events; and (7) the freeze-back period is characterized by a reversal in the overall hydrological balance as evaporation wanes and lateral inflows and precipitation increase up to the time of permanent winter ice cover development.

For the purpose of this comparative analysis only the ice free period is considered. During this time, lakes are typically well mixed and evaporative enrichment is pronounced, but there is significant potential for small fluctuations in rates of isotopic enrichment associated with short-term meteorological and hydrological events. For this reason, a stepwise approach was adopted as suggested by Gonfiantini (1986) for lakes showing fast volume variations with rates changing in time, whereby each time interval represents a short period where inflow–outflow–evaporation conditions can be assumed to be relatively steady. Although this approach uses fewer assumptions and hence requires more detailed characterization of hydrological and isotopic parameters than other approaches, it represents a significant simplification relative to standard approaches for examining evaporation from many lakes in a given location. For the purposes of this study, it also enables a detailed comparison with standard evaporation methods over short time intervals. The basis of the method is presented below.

*Theory*

The water mass and isotope-mass balance (for either $\delta^{18}O$ or $\delta^2H$) of a well-mixed lake, assuming constant density of water, are given respectively as

$$\frac{dV}{dt} = I - O - E \tag{1}$$

$$\delta_L \frac{dV}{dt} + V \frac{d\delta_L}{dt} = \delta_I I - \delta_Q O - \delta_E E \tag{2}$$

where $V$ is the lake volume at time $t$; $dV$ is the lake volume change over time interval $dt$; $I$ is the mean inflow, comprised of precipitation, surface inflow and subsurface inflow; $O$ is the mean outflow, comprised of surface and subsurface outflow; $E$ is the mean lake evaporation; and $\delta_L$, $\delta_I$, $\delta_Q$ and $\delta_E$ are the isotopic composition of lake water, inflow, outflow and evaporating moisture, respectively. Note that $\delta_Q \approx \delta_L$ for well-mixed conditions.

The isotopic composition of evaporating moisture $\delta_E$ is estimated, assuming zero resistance in the liquid phase, from the linear resistance model given by Craig and Gordon (1965) as

$$\delta_E = \frac{\delta_L - h\delta_A - \varepsilon}{1 - h + \epsilon_k} \tag{3}$$

where $h$ is the ambient (free air) humidity normalized to conditions at the air–water interface, $\delta_L$ is the isotopic composition of the evaporating liquid and $\delta_A$ is the isotopic composition of ambient atmospheric moisture. The total isotopic enrichment factor $\varepsilon$ is derived from $\varepsilon = \varepsilon^* + \varepsilon_K$ where $\varepsilon^*$ is the equilibrium isotopic enrichment between liquid and vapour, $\varepsilon_K = C_K(1 - h)$ is the kinetic isotopic enrichment resulting from diffusion of water molecules through the boundary layer and $C_K$ is a parameter describing relative transport resistance of the heavy and common isotopic species. Typical $C_K$ values of 14·2 and 12·5‰ for oxygen and hydrogen, respectively, were adopted from previous studies and are representative of conditions of evaporation into a fully turbulent atmosphere (Gonfiantini, 1986; Merlivat, 1978). Values of $\varepsilon^*$, which vary only slightly over the range of temperatures observed in the study, were calculated from time-series records of lake and pan temperatures using an equation presented by Gonfiantini (1986) based on experimental data from Majoube (1971).

A constant volume form of the isotope-mass balance Equation (2) was applied to estimate evaporation over discrete time steps defined by the water sampling interval of the lake. Combining Equations (2) and (3), assuming $dV/dt = 0$ for a given time step, we obtain an expression for the rate of change of $\delta^{18}O$

and $\delta^2H$ of lake water, $\delta_L$, given by (Gonfiantini, 1986) as

$$\frac{d\delta_L}{dt} = -[(1 + mx)\delta_L - \delta_I - x\delta^*]\frac{It}{V}$$

(4)

where $x = E/I$ is the fraction of lake water lost by evaporation, $1 - x = O/I$ is the fraction of water lost to liquid outflows, $m = (h - \varepsilon)/(1 - h + \varepsilon_K)$ as defined by Welhan and Fritz (1977) and Allison and Leaney (1982), and $\delta^* = (h\delta_A + \varepsilon)/(h - \varepsilon)$ is the limiting isotopic composition under local meteorological conditions as defined by Gat (1981b).

Integrating with respect to $\delta_L$ over the limits $t_0 \rightarrow t$ using mean values of $\delta_I$, $\delta_A$, $\varepsilon^*$, $\varepsilon_K$, $V$, $h$, $I$, $O$ and $E$ then yields:

$$\delta_L(t) = \delta_S - (\delta_S - \delta_0)\exp\left[-(1 + mx)\frac{It}{V}\right]$$

(5)

where $\delta_0$ is the initial isotopic composition of lake water, $\delta_L$ is the isotopic composition of lake water at time $t$ and $\delta_S = (\delta_I + mx\delta^*)/(1 + mx)$ is the steady-state isotopic composition the lake would reach $t \rightarrow \infty$.

In order to calculate $E = xI$ for each time step, $dt$, the values of $\bar{h}$, $\bar{T}$, $\bar{\varepsilon}^*$, $\bar{\varepsilon}_K$, $\bar{m}$, $\bar{V}$ and $I$ were assembled, $\delta^*$ was calculated from an independent $\delta_A$ record derived from isotope–mass balance of constant volume class A pans (Gibson, 1996; see also Allison and Leaney, 1982), $\delta_I$ was estimated from weighted sampling of relevant components and $\delta_0$ and $\delta_L$ were measured directly from lake water samples. Using Equation (5) this permitted determination of $x$ and $\delta_S(x)$, yielding total $E$ for both $\delta^{18}O$ and $\delta^2H$ for individual times steps. Outflow was also calculated but is not presented here.

For comparative purposes, actual volumes, volumetric changes and physically gauged inflow are used for each time step. In addition, mean lake depth $\bar{D} = \bar{V}/\bar{A}$ is substituted for $\bar{V}$ in order to estimate $E$ as an equivalent depth. Simplified models requiring no physical gauging of inflow and outflow have also been evaluated but are beyond the scope of this paper.

*Study area*

The study was conducted near Lupin, NWT (Figure 1a), a gold mining operation situated in the continental Arctic of Canada, approximately 375 km north of Yellowknife and 125 km beyond northern tree-line (Figure 1b). The region is characterized by low relief, poorly integrated drainage and abundant shallow lakes. The mean annual temperature, based on records from 1982 to 1993 at the Lupin Airport weather station, is $-12°C$ with mean annual precipitation of 289 mm, about half of which falls as snow. Mean daytime temperatures above 0°C are generally observed from late-May to early-September and the ice free season on small lakes extends from mid-June to late-September. From climatological maps, lake evaporation at the site is estimated at between $200\,\text{mm}\,\text{a}^{-1}$ (den Hartog and Ferguson, 1978) and $325\,\text{mm}\,\text{a}^{-1}$ (Morton, 1979). Strong regional evaporation gradients of about $50\,\text{mm}\,\text{a}^{-1}$ per 100 km are predicted in the region with annual totals decreasing towards the north-east (Figure 1b).

A small lake with a surface area of 6 ha, a catchment area of 27 ha and an average depth of 0·65 m was selected for detailed investigations ('K-6 L', Figure 1c). The lake has only intermittent surface outflow during the summer, occurring through a small drainage stream and a secondary streamlet which were active during periods of high water level. The catchment was chosen based on its apparent representativeness of local lakes, proximity to a road, manageable size and undisturbed character. Dwarf tundra vegetation is common on well-drained slopes, while grass and sedge dominate low areas and the lakeshore. Active-layer development within surficial glacial till deposits is typically restricted to 0·5 to 2 m depth, based on detailed frost-probing surveys (Gibson, 1996). Continuous permafrost is known to extend to 510 m depth based on observations in the nearby mine workings.

*Conventional methods*

Automated sensors were mounted on two towers and a float apparatus within the lake (see Figure 1c), and were configured to provide the necessary parameters for estimation of hourly evaporation using a
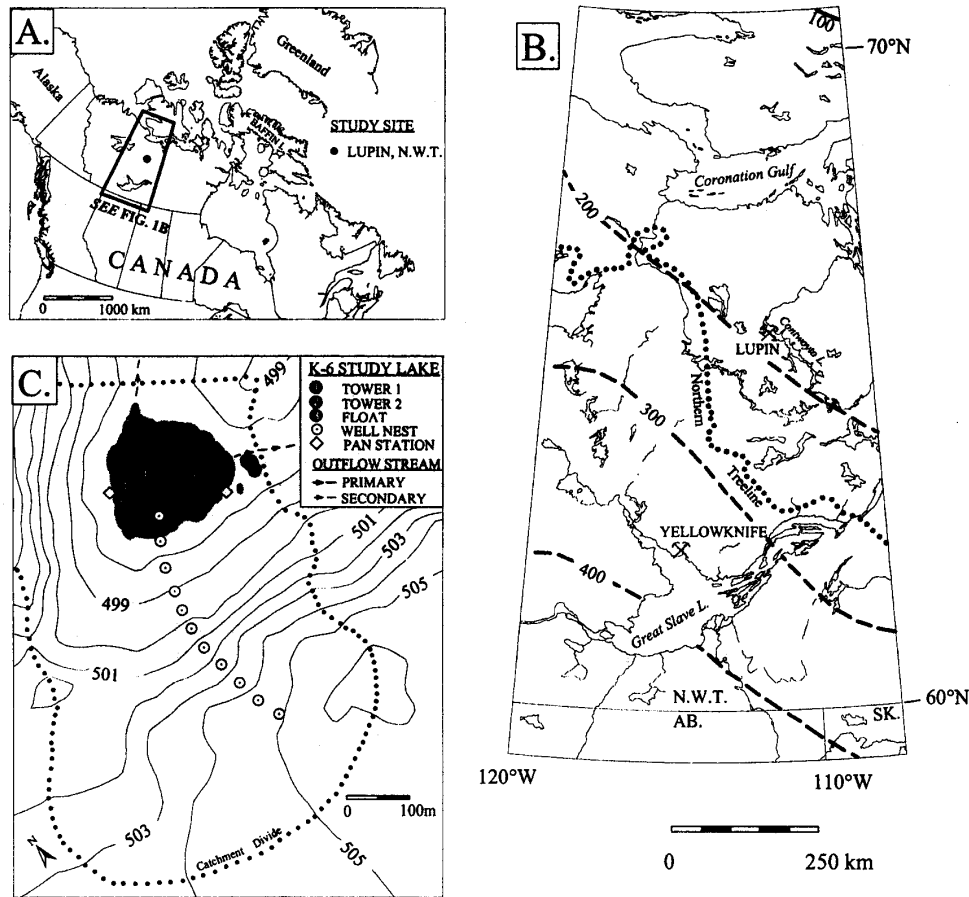
Figure 1. (a) Map of Canada showing study site locations. (b) Enlargement of central Arctic showing Lupin and Yellowknife, approximate position of northern treeline and estimated mean annual lake evaporation contours in $mm\,a^{-1}$ (after den Hartog and Ferguson, 1978); (c) 'K-6' study lake and catchment area showing location of instrumentation and outflow channels

Bowen ratio energy balance and a two-point aerodynamic profile method. A detailed account of field methods and results has been presented elsewhere (Gibson et al., 1996). A standard Bowen ratio approach was adopted from other northern lake studies (Stewart and Rouse, 1976; Bellow and Smith, 1990) with one notable exception. Namely, lake heat storage and ground heat flux were evaluated at two locations within the lake to account for potential differences in bed characteristics and heat flux in shallow and deep areas. A two-point aerodynamic profile method was applied similar to that of Ohmura (1982), with stability corrections determined by the method outlined in Monteith and Unsworth (1990).

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

$$E = I - O - \mathrm{d}V/\mathrm{d}t \qquad (6)$$

where $I = P + LI$, and $P$ is the daily volume of precipitation falling on the lake surface, $LI$ is the volume of lateral surface and subsurface inflow, $O$ is the volume of surface and subsurface outflow, and $\mathrm{d}V/\mathrm{d}t$ is the volumetric change in lake storage. $P$ was determined based on the daily depth of recorded rainfall

and daily mean values of lake surface area, and $dV/dt$ was derived from lake water level records converted to volumes using a rating curve. *LI* and *O* were estimated from records of hillslope storage surplus and lake stage records, respectively, calibrated through manual flow gauging. Summaries of mass balance components for 1992 and 1993 are shown in Figures 2 and 3, respectively.

### Water sampling and analysis

Water samples for isotopic analysis were collected manually at regular intervals from the study lake, lateral inflow sources and two class A evaporation pans situated near the lake shore. A custom-designed automatic sampler, which pumped water from the lake and maintained an air-tight seal on collected samples, was used during periods when a regular visit could not be made. Rainfall was sampled as required from a roof drainage system at the nearby mine site, and snowfall, when it formed substantial accumulations, was collected on plywood boxes at ground level. Active-layer water was characterized based on weighted sampling of shallow wells (Figure 1c) and sampling of overland flow when it occurred. In all cases, sampling sites were carefully chosen to be more or less representative of each water balance component. For the lake, one-litre samples were normally collected several metres from shore at mid-depth and then transferred to smaller sampling bottles.

Additional samples were also collected over the course of the study to monitor potential isotopic variations resulting from incomplete mixing of the lake. Overall, depth and spatial variations in isotopic composition within the lake were found to be minor during the ice free period (Gibson, 1996). Daily temperature distributions, continuously monitored within the lake using 12 thermistors and a tower-mounted infrared radiometer, were found to be uniform within 1 or 2°C, supporting the assumption that the lake was well mixed. All samples were collected in 30-ml HDPE bottles and returned to the Environmental Isotope Laboratory, University of Waterloo for analysis of oxygen and hydrogen stable isotopes. Analytical uncertainties are approximately $\pm 0.1‰$ for $\delta^{18}O$ and $\pm 2‰$ for $\delta^2H$.
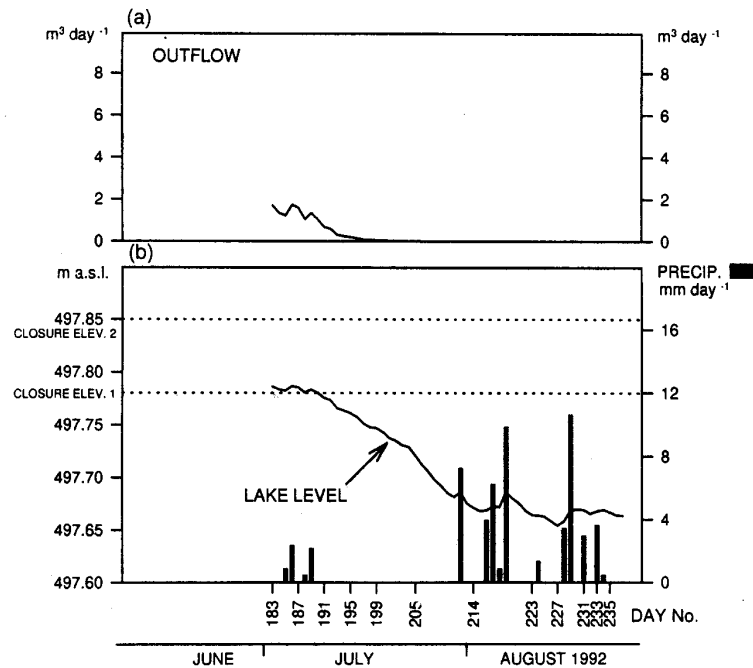


Figure 2. 1992 daily mass balance summary, (a) outflow, and (b) lake level and daily precipitation. Note that 'closure elevation 1 and 2' are the minimum lake levels required for surface outflow in the main drainage stream and secondary streamlet, respectively (Figure 1c). Also, 'Day No.' indicates day of year of water sampling for isotopic analysis
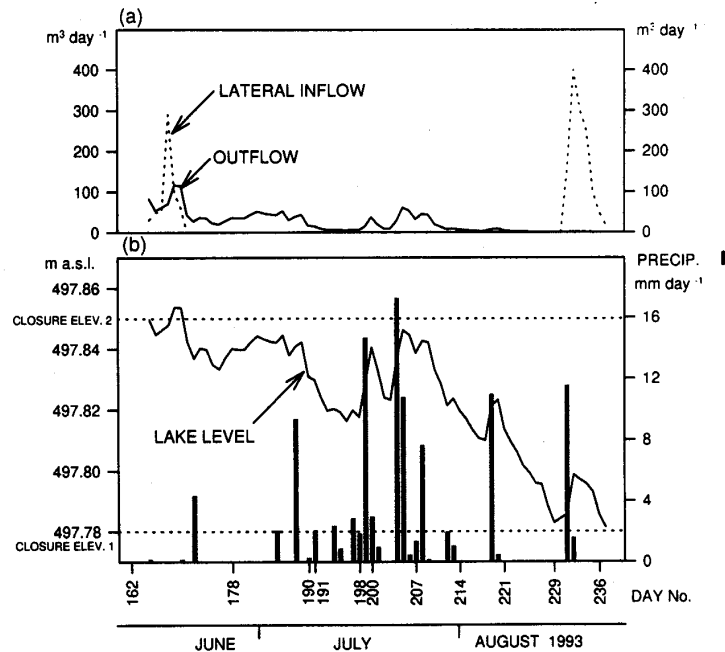
Figure 3. 1993 daily mass balance summary, (a) outflow and lateral inflow, and (b) lake level and daily precipitation. Note that 'closure elevation 1 and 2' are the minimum lake levels required for surface outflow in the main drainage stream and secondary streamlet, respectively (Figure 1c). Also, 'Day No.' indicates day of year of water sampling for isotopic analysis

## Additional monitoring

For the isotope-mass balance determinations, normalized humidity, lake temperature and class A pan temperatures for each sampling interval were calculated from hourly meteorological records at the lake site. The computer program SURFER$^{TM}$ (Golden Software Inc.) was used to estimate mean lake volume $\bar{V}$, mean area $\bar{A}$ and mean depth $\bar{D} = \bar{V}/\bar{A}$ for each sampling interval based on a high-resolution theodolite survey of the lake bed and vicinity, and from water level records. Inflow was measured as described in 'Conventional methods'.

## RESULTS AND DISCUSSION

### Isotopic variations

The study lake enriched by 4·6‰ in $\delta^{18}$O over July to mid-September 1992 (day 183 to 255) from a minimum of $-18\cdot1$‰ to a maximum of $-14\cdot3$‰ (Figure 4), reflecting relatively high rates of evaporation, low precipitation and negligible lateral inflow during this period (Figure 2). Values of $\delta^{18}$O for precipitation ranged from greater than $-16$‰ to less than $-24$‰ with a mean weighted value close to $-20$‰, similar to mean active-layer water. A midsummer snowfall event, contributing a water equivalent of 11 mm ($\delta^{18}$O $= -21\cdot1$‰), briefly interrupted the lake enrichment trend. From the time of peak enrichment to freeze-up in late-September, $\delta^{18}$O declined by 0·8‰ to $-15\cdot2$‰ in response to declining rates of evaporation and reactivation of lateral inflow. Similar, although less regular, enrichment trends are observed for $\delta^2$H (Figure 5), with a net evaporative enrichment of 24‰ over the season. Over the comparative interval when all methods were applied (day 183 to 235), net enrichment was 3·4‰ for $\delta^{18}$O and 18‰ for $\delta^2$H.

Sampling during 1993 extended form the late-winter dry snow period to the time of freeze-up on the lake, and included a 250-point depth-integrated snow survey in the catchment area, and a three-point lake ice
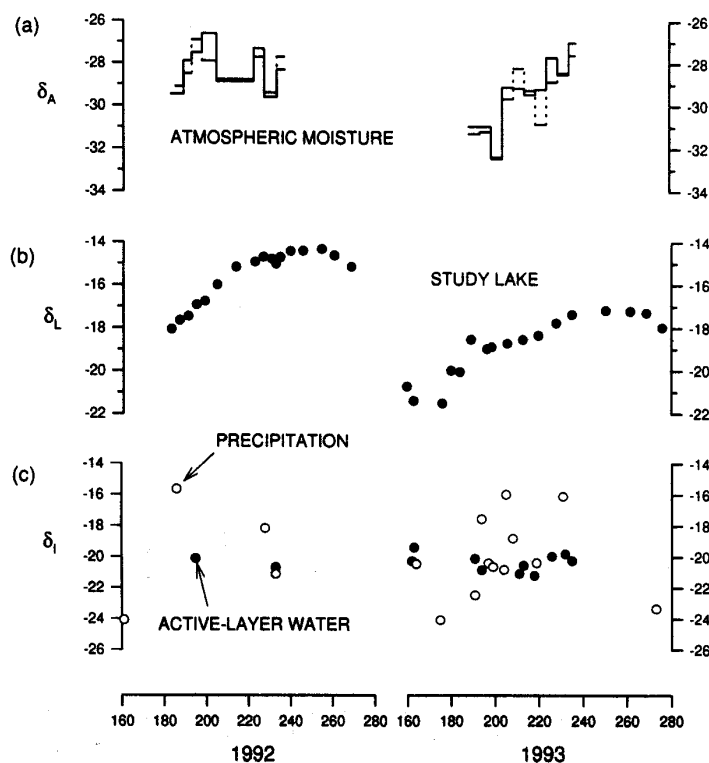
Figure 4. Time series of $\delta^{18}O$ for thaw season mass balance components sampled at Lupin, NWT. Components include: (a) atmospheric moisture derived from isotope balance of two class-A evaporation pans, (b) 'K-6' lake water, and (c) input sources, namely precipitation and active-layer water (lateral inflow). Note that all points represent analysis of single water samples with the exception of active-layer water points, which are based on integrated sampling of all wells (Figure 1c) (after Gibson *et al.*, 1995)

profile survey (Gibson, 1996). The water equivalent of snow in the catchment was 39 mm and its mean isotopic composition was $-24 \cdot 1 \pm 2 \cdot 4‰$ in $\delta^{18}O$ and $-175 \pm 19‰$ in $\delta^2H$. Lake ice, which extended to the lake bottom in two out of three profiles, was slightly enriched in $\delta^{18}O$ and $\delta^2H$ relative to the lake at the time of freeze-up ($\delta^{18}O = -13 \cdot 8‰$, $\delta^2H = 0127‰$). Taking into account the isotopic composition of the subice lake water ($\delta^{18}O = -21 \cdot 7‰$, $\delta^2H = -173‰$), and an estimate of the ice : water ratio (roughly 5 : 1), we obtain an estimate of the mean isotopic composition of the lake in late-winter which is consistent with that observed at the time of freeze-up the previous year.

The isotopic composition of the lake is shown in Figures 4 and 5 from the time that meltwater was present on the lake ice until the time of freeze-up, and is described below for $\delta^{18}O$. The $\delta^{18}O$ of the earliest meltwater in the lake was intermediate between snow and lake ice, and reflects a mixture of these sources. $\delta^{18}O$ values slowly increased from about $-20 \cdot 8‰$ to a maximum of $-18 \cdot 5‰$ as ice ablation proceeded. Over the ice free period (day 186 to 276), total evaporative enrichment in $\delta^{18}O$ was $2 \cdot 9‰$ ($25‰$ for $\delta^2H$). Over the comparative interval, when all methods were applied (day 186 to 236), net enrichment was $1 \cdot 6‰$ for $\delta^{18}O$ and $13‰$ for $\delta^2H$, or 47 and 72%, respectively, of that observed for 1992.

Net enrichment in $\delta^{18}O$ and $\delta^2H$ in constant volume class A pans situated near the lake shore was about 7 and $35‰$, respectively, for 1992, and $5 \cdot 3$ and $31‰$, respectively, for 1993 (not shown). Pan-derived atmospheric moisture values (assuming $C_K$ values for fully turbulent conditions) varied between $-26 \cdot 5$ and $-32 \cdot 5‰$ in $\delta^{18}O$ and $-200$ to $-250‰$ in $\delta^2H$, and were consistent between pans (Figures 4 and 5). Comparisons with values derived from direct vapour sampling using cold traps is presented elsewhere (Gibson, 1996).
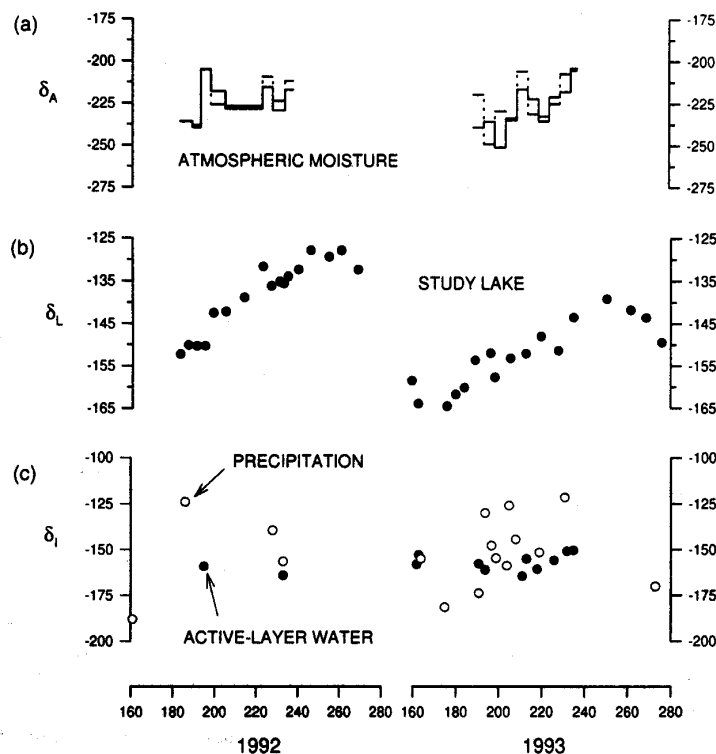
Figure 5. Time series of $\delta^2$H for thaw season mass balance components sampled at Lupin, NWT. Components include: (a) atmospheric moisture derived from isotope balance of two class-A evaporation pans, (b) 'K-6' lake water, and (c) input sources, namely precipitation and active-layer water (lateral inflow). Note that all points represent analysis of single water samples with the exception of active-layer water points which are based on integrated sampling of all wells (Figure 1c)

Evaporation was calculated using Equation (5) and the isotopic records shown in Figures 4 and 5 for 11 intervals totalling 52 days during 1992 and eight intervals totalling 50 days during 1993.

*Evaporation: estimates and comparisons*

*Cumulative totals.* Cumulative evaporation curves for each method are presented in Figures 6 and 7 and totals are summarized in Table 1.

Table I. Lake and pan evaporation totals and daily rates for July and August, 'K-6' Lake, Lupin, NWT

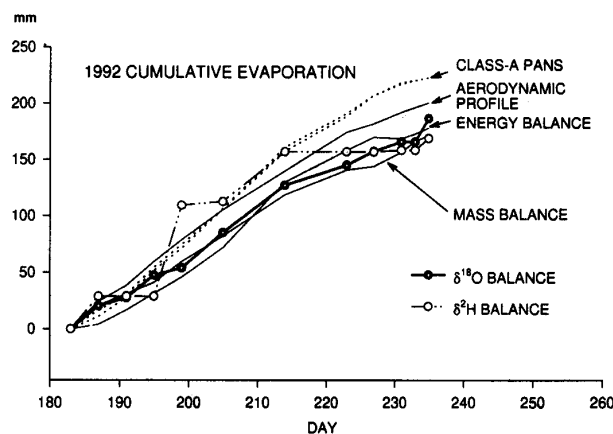| Year no. days | | Lake evaporation methods | | | | | |
|---|---|---|---|---|---|---|---|
| | | $\delta^{18}$O | $\delta^2$H | Profile | Energy balance | Mass balance | Class A pans |
| 1992 | Total (mm) | 187 | 169 | 201 | 171 | 178 | 223 |
| 52 d | Mean (mm day$^{-1}$) | 3·6 | 3·3 | 3·9 | 3·3 | 3·4 | 4·3 |
| | (% Diff.)* | (0) | (−10) | (+7) | (−8) | (−5) | (+20) |
| 1993 | Total (mm) | 91 | 107 | 101 | 97 | 156 | 139 |
| 50 d | Mean (mm day$^{-1}$) | 1·8 | 2·1 | 2·0 | 1·9 | 3·1 | 2·8 |
| | (% Diff.)* | (0) | (+18) | (+11) | (+7) | (+71) | (+53) |

* %Difference from $\delta^{18}$O balance estimate

Figure 6. 1992 cumulative evaporation curves of 'K-6' study lake based on the indicated methods. Circles denote times of water sampling for isotopic analysis. Also shown are results from two class A pans (uncorrected) situated on the lake shore

Several obvious trends for 1992 (Figure 6) are noted, as follows. (1) Cumulative evaporation totals are very similar for all methods. (2) Cumulative evaporation for lake evaporation methods is lower in all cases than class A pans. (3) The slope of the $\delta^{18}$O balance curve is very consistent with that of the aerodynamic profile method, the energy balance method and the standard mass balance method. (4) the slope of the $\delta^2$H balance curve is more erratic than other methods, with very poor agreement over some intervals.

Similar trends are observed for 1993, although cumulative evaporation based on the standard mass balance is significantly higher than other methods, including class A pan estimates. Given that class A pans normally overestimate lake evaporation by 10–60% (Kohler, 1954), and considering the poor agreement with all other methods, this result is viewed as anomalous. Severe overestimation of evaporation in this case is likely to be related principally to underestimation of stream discharge, which occurred only during 1993 and was difficult to measure accurately given the disorganized channel morphology. This result demonstrates the limitations of the mass balance methodology employed in this study. Whereas the consistent results obtained for 1992 suggest that the method is appropriate for times when the
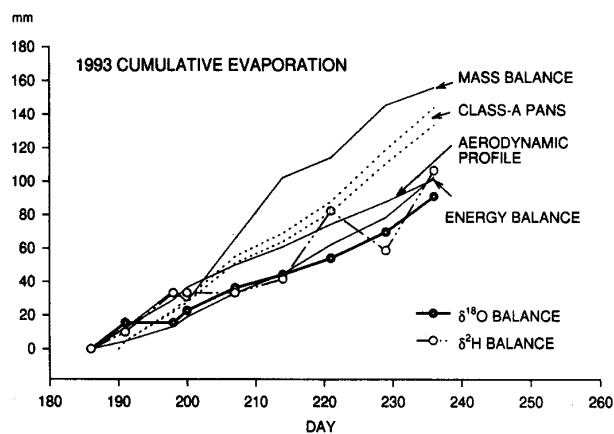


Figure 7. 1993 cumulative evaporation curves for 'K-6' study lake based on the indicated methods. Circles denote times of water sampling for isotopic analysis. Also shown are results from two class A pans (uncorrected) situated on the lake shore

mass balance is approximated by $E = P - dV/dt$, as shown by Marsh and Bigras (1988) for other northern lakes.

Overall, the $\delta^{18}O$ balance method provided satisfactory results over the 50-day intervals and was consistent to within ±10% of the energy balance and aerodynamic profile methods (Table I). In general, the consistency between methods is excellent considering that under ideal field conditions it is difficult to measure evaporation accurately to within 10 or 15% for short periods (Winter, 1981). In contrast, the $\delta^{2}H$ balance method yielded less satisfactory results, and predicted neither the timing nor the magnitude of the evaporative flux as accurately as the $\delta^{18}O$ balance.

*Limits of temporal resolution.* Experimental results for individual time steps, as defined by the water sampling interval of the lake, are displayed in Figure 8.

Mean evaporation rates based on $\delta^{18}O$ balance are shown to be in broad agreement with the other methods over most intervals, and variability is not dissimilar to that observed between the standard methods. Although more consistent results are usually obtained for longer intervals with high enrichment (greater than about 0·4‰ in $\delta^{18}O$ this is not always the case. It is interesting to note that net depletion in $\delta^{18}O$ occurred in the lake during 1992 over interval nos. 8 and 9 (day 223 to 231), yet modelled evaporation is positive and consistent with the standard methods. This is not surprising, however, if we consider the influence of a cold air intrusion that dominated the region at this time. Owing to depleted atmospheric moisture compositions (Figure 4) and input of isotopically depleted snowfall associated with this weather system, the evaporative enrichment in the lake was completely attenuated and the lake underwent a net depletion in $\delta^{18}O$. The consistent evaporation estimates obtained for the period, despite changing hydrological and atmospheric conditions, suggests that the fundamental controls on the $\delta^{18}O$ behaviour of the lake are well described by the model. In contrast, the poor correlation between $\delta^{2}H$ balance estimates and standard estimates, even for six- or nine-day intervals, during periods when the water balance was
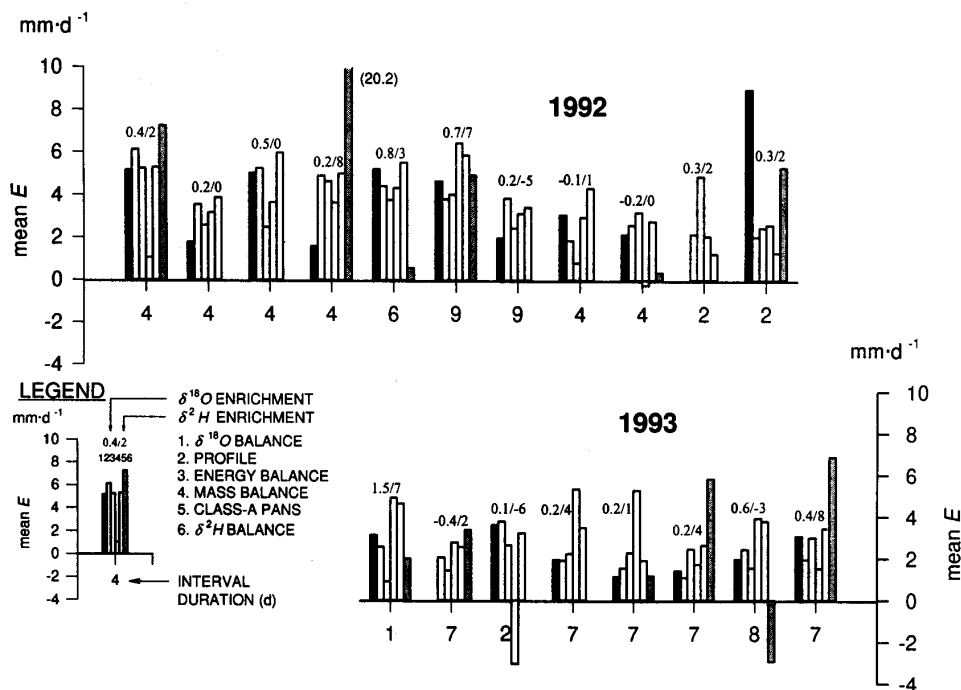


Figure 8. Mean evaporation rates for the specified methods. Note that non-isotopic methods were calculated daily and compiled over water sampling intervals to permit comparison with isotopic methods

dominated by evaporation, suggests that it may be less appropriate for high time-resolution applications. This may in part be related to higher analytical uncertainty compared with $\delta^{18}O$.

Over the entire comparative intervals, potential errors in $E$ related to analytical uncertainty of $\delta^{18}O$ and $\delta^2H$ alone were likely to be less than $\pm 3$ and $\pm 11\%$, respectively, for 1992, and $\pm 6$ and $\pm 15\%$, respectively, for 1993. In order to maintain potential errors in $E$ related to analytical uncertainty to within $\pm 15\%$ for individual time intervals for the study lake, sampling intervals evidently should have been spaced to allow enrichment of at least 0·6‰ or 37 mm $E$ (ca. 14 days) of evaporation for $\delta^{18}O$ and 13‰ or 120 mm $E$ (ca. 44 days) of evaporation for $\delta^2H$. For larger lakes, where the rate of isotopic enrichment is expected to be lower, evaporation should probably be calculated over intervals of a month or more, whereby several samples are used to characterize the enrichment trend. Certainly, based on such a procedure, and also using basic judgement through comparison between $\delta^{18}O$ and $\delta^2H$, it is possible to exclude anomalous individual analyses and minimize resulting errors.

*Comparisons based on amalgamated intervals.* The isotopic methods and standard methods were also compared over amalgamated intervals, comprising multiple, concurrent time steps, in order to assess the consistency between methods over longer periods. Plots of energy balance $E$ versus $\delta^{18}O$ balance $E$ and $\delta^2H$ balance $E$ are shown in Figure 9 for all individual time steps and amalgamated intervals.

Evidently, the $\delta^{18}O$ balance estimates are in good agreement with energy balance estimates and only slightly overestimated evaporation during 1992 and moderately underestimated evaporation during 1993 (Figure 9). Systematic differences between 1992 and 1993 estimates may suggest systematic errors in the characterization of one or more parameters in Equation (5). A sensitivity analysis reveals that the most likely explanation for systematic underestimation during 1993 is underestimated inflow. If higher inflow did occur during this time, reservoir enrichment would have been attenuated beyond what is accounted for in the model, leading to an apparent reduction in lake evaporation rates, and a cumulative error in total $E$.

The $\delta^2H$ balance estimates are once again less consistent than $\delta^{18}O$ (Figure 9) but importantly, the method appears to underestimate evaporation during 1993. This is consistent with the hypothesis that inflow during 1993 was underestimated. A better understanding of the behaviour of $\delta^2H$, including reassessment of reported analytical uncertainties, is certainly required for routine application using this methodology.

## SUMMARY AND CONCLUSIONS

While, in principle, the isotope mass balance method can be applied with any desired level of precision by detailed sampling and proper characterization of the components (Gat, 1970), it is often difficult or
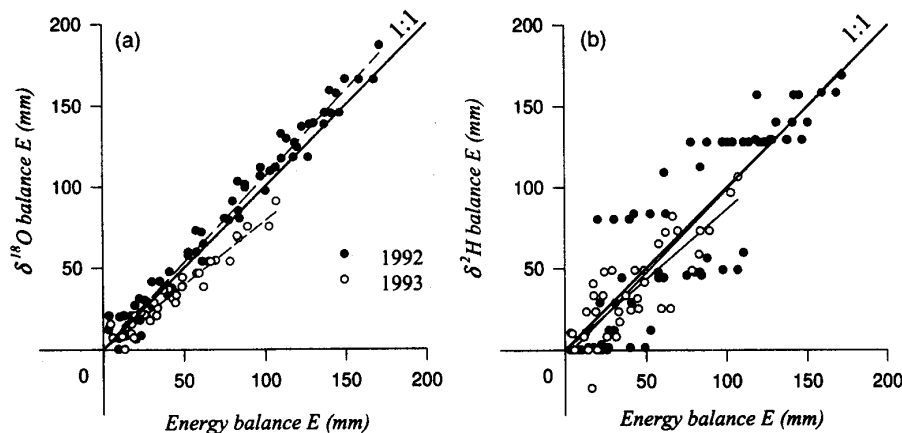


Figure 9. (a) Energy balance $E$ versus $\delta^{18}O$ balance $E$, and (b) energy balance $E$ versus $\delta^2H$ balance $E$, for all time intervals and amalgamated intervals evaluated for the 'K-6' study lake, Lupin, NWT. See text for discussion

impossible to estimate the precision for a given approach. Previous isotopic studies have cited errors ranging from as high as 30 to 50% or more for some common applications (Gonfiantini, 1986). Through a detailed comparative study we have demonstrated that the accuracy of a selected approach can be comparable to aerodynamic profile or energy balance methods, even over short time periods. The very good agreement observed, particularly for $\delta^{18}O$, illustrates the high sensitivity of lakes to evaporation and associated isotopic changes in the present setting. It also suggests that the basic model assumptions, including the use of laboratory values for $C_K$ and pan-derived $\delta_A$ are a suitable approximation for $\delta^{18}O$ in this setting. For $\delta^2H$, we have demonstrated that proper characterization of the isotope balance would require a more detailed sampling programme, use of longer integration periods or reassessment of reported analytical uncertainties.

Although it is perhaps simpler to use standard hydrological approaches for monitoring evaporation from individual small lakes in northern Canada, the potential of the isotopic approach lies in the examination of variability, both from lake to lake within a given locality, and regionally. This is a primary objective of ongoing research. This study has established an initial basis for comparison of results so that both conventional and isotopic research can be applied to achieve a broader understanding of hydrological processes in northern Canada and elsewhere.

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