Short-term evaporation and water budget comparisons in shallow Arctic lakes using non-steady isotope mass balance

J.J. Gibson[*]

Department of Earth Sciences, University of Waterloo, Waterloo, Ont. N2L 3G1, Canada

Abstract

Shallow lakes in a continental, low Arctic setting are found to undergo substantial fluctuations in heavy-isotope content during the annual cycle due to extreme seasonality in water balance processes. Progressive isotope enrichment during the ice-free period occurs as a consequence of isotope exchange during evaporation under seasonally arid conditions, while enhanced input and flushing by heavy-isotope depleted precipitation and snowmelt during late fall and spring tend to deplete the lakes in heavy isotopes once again. Time-series sampling surveys in a group of nearby lakes, ranging in volume from 41,000 to 2,250,000 m^3, was carried out for two consecutive ice-free periods, in conjunction with a comprehensive hydrological measurement program in a detailed study lake, to assess models that describe isotopic enrichment in lakes and to test their suitability for comparing evaporation rates and water balance. From a non-steady isotope balance analysis, it is found that isotope enrichment rates in lakes during the ice-free period are determined primarily by evaporation rates and volume of the lakes, and are less sensitive to water balance variations for short time intervals when evaporation is less than 50 mm or so. A basic assessment of best-fit and step-wise models is presented which suggests that the former are useful only for predicting evaporation and water balance during month-long periods with relatively stable atmospheric and hydrologic conditions. A step-wise isotope balance approach is presented which demonstrates how isotope-based estimates of evaporation rates can be applied to effectively compare short-term (weekly) water balance in nearby lakes. Practical applications in the region include water balance assessments to assist in design and maintenance of tailings ponds for gold and diamond mining operations. © 2002 Elsevier Science B.V. All rights reserved.

Keywords: Lakes; Evaporation; Stable isotopes; Oxygen-18; Deuterium; Water balance

1. Introduction

Balance equations of the isotopic tracers ^{18}O and ^{2}H provide independent hydrological information that may be useful for estimating evaporation and other

[*] Present address: Isotope Hydrology Section, International Atomic Energy Agency, Wagramer Strasse 5, P.O. Box 100, A-1400 Vienna, Austria..

E-mail address: j.gibson@iaea.org (J.J. Gibson).

water balance parameters, as demonstrated in previous studies of open-water bodies (Dinçer, 1968; Gat, 1970, 1981; Zuber, 1983; Gibson et al., 1993; Krabbenhoft et al., 1990). Isotopic studies to date have relied extensively on tracing of average, long-term water balance conditions through the use of basic models assuming isotopic and hydrologic steady state. Although such models may be effective for describing the annual water balance of large volume lakes in temperate climates, which typically undergo only
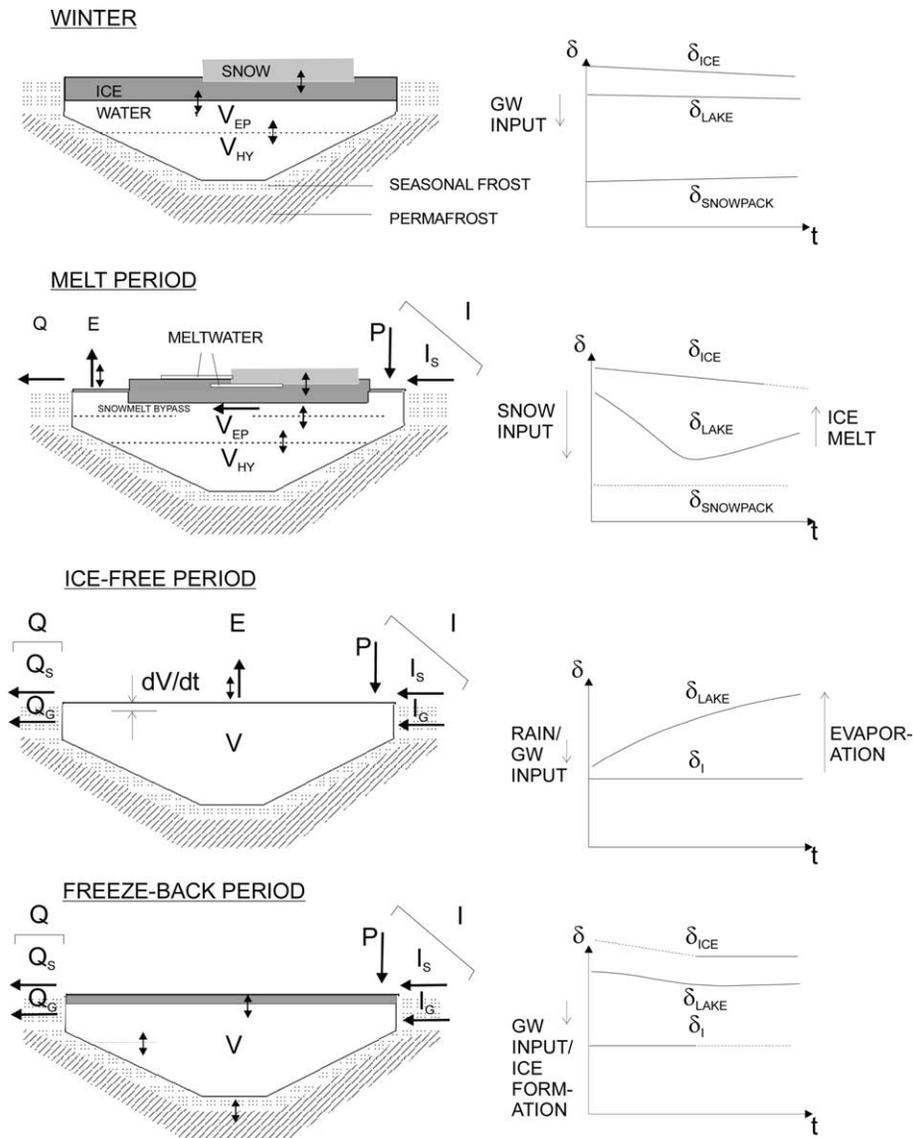
Fig. 1. Left panels: schematic of a lake in a cold climate region showing important hydrologic storages and fluxes for various seasons. Note that lake volume $V = V_{EP} + V_{HY}$, where V_{EP} is the volume of the epilimnion and V_{HY} is the volume of hypolimnion when the lake is not well mixed; dV/dt is the change in lake volume; E is the evaporation; I is the inflow, where $I = I_S + P$, I_S is subsurface outflow, and P is precipitation; Q is the outflow, where $Q = Q_S + Q_G$, Q_S is the surface outflow and Q_G is the subsurface outflow. Bidirectional arrows indicate isotope exchange between compartments. Right panel: Time series of isotopic composition of lake ice, water, snowpack, and input showing general direction of hydrological forcings (arrows). Arrows show major forcings on isotopic composition (δ).

moderate seasonal changes in lake volume and isotopic composition, they may be less appropriate for describing shallow lakes in cold continental climates. In the latter case, this is due to transient isotopic and water balance conditions which arise from the seasonality of atmospheric and hydrologic processes.

In seasonally arid Arctic and Subarctic nival regimes, the isotopic composition of shallow lakes evolves throughout the open-water period, as influenced by abundant throughflow of snowmelt which is depleted in heavy isotopic species in the early melt period, by evaporative isotopic enrichment of heavy

isotopes during mid-summer, and by enhanced active layer input and reduced evaporation rates during the late summer and fall. Lake isotope composition begins the evaporation season substantially depleted in the heavy isotopic species relative to values representative of isotopic steady state with the summer atmosphere which further enhances the overall enrichment. Transient conditions are also perpetuated by the extreme brevity of the open-water period, so that the limiting isotopic composition is often never attained. Previous studies (Gibson et al., 1993; Gibson, 2001) have successfully adapted steady-state models to estimate long-term (or annual) evaporation and outflow in a wide range of lakes in Arctic and Boreal regions, but such models are not well suited to short-term studies conducted over periods of weeks or months.

Fontes and Gonfiantini (1967) and Zimmerman (1979) have demonstrated the application of non-steady methods for estimating evaporation and groundwater recharge in some transient but unidirectional systems, namely desiccating ephemeral ponds in the Sahara, and newly constructed artificial reservoirs in Germany. These also represent systems far from isotopic steady state due to events or changes in the ambient conditions, although not due to extreme seasonality as is demonstrated herein for Arctic lakes. This study presents and compares the isotopic response in several shallow Arctic lakes situated within a distance of 6 km, utilizing supplementary micrometeorological records (humidity and air temperature), and the isotopic composition of atmospheric moisture (δ_A) estimated from mass balance of two class-A pans (Gibson et al., 1999). Important considerations such as characterization of active layer input and its isotopic composition, lake volume, and horizontal and vertical mixing are also discussed. This paper begins with a generalized overview of the relevant theory used for isotope mass balance in a variety of non-steady-state water balance scenarios, intended to clarify and categorize the range of possible non-steady formulations that have been applied or postulated in previous studies. From these broader theoretical discussions, models are selected and applied to a dataset from the field study in northern Canada. These examples illustrate field-based application of the models for characterizing evaporation rates and water balance from Arctic lakes using both best-fit and step-wise approaches. Importantly, the study demonstrates some of the major benefits and limitations of using the non-

steady-state approach for characterizing weekly lake-to-lake variations in evaporation rates.

1.1. Theory

The isotope balance relationships described in the following section are relevant for describing changes in δ^{18}O and δ^2H occurring during the ice-free period in response to evaporative enrichment when the lake is well mixed (Fig. 1c). The models presented do not depict isotopic changes during the winter when the lake may be stratified, or during the melt or freeze-back periods when isotope balance is complicated by incomplete mixing, ice formation, or snow and ice melting (Fig. 1a, b, d).

The water mass and isotope mass balance for a shallow, well-mixed lake during the ice-free period (Fig. 1c), assuming constant density of water, may be written respectively as

$$\mathrm{d}V/\mathrm{d}t = I - Q - E \qquad (1)$$

$$V\frac{\mathrm{d}\delta_L}{\mathrm{d}t} + \delta_L\frac{\mathrm{d}V}{\mathrm{d}t} = I\delta_I - Q\delta_Q - E\delta_E \qquad (2)$$

where V is the volume of the lake, t the time, $\mathrm{d}V$ the change in volume over time interval $\mathrm{d}t$, I the instantaneous inflow where $I = I_S + I_G + P$ (I_S is the surface inflow, I_G the groundwater inflow, and P the precipitation on the lake surface), Q the instantaneous outflow where $Q = Q_S + Q_G$ (Q_S is the surface outflow and Q_G the groundwater outflow), E the evaporation, and δ_L, δ_I, δ_Q and δ_E the isotopic compositions of the lake, inflow, outflow, and evaporative flux, respectively.[1] Note that instantaneously $\delta_I = (\delta_S I_S + \delta_G I_G + P\delta_P)/I$ and $\delta_Q = (\delta_S Q_S + \delta_G Q_G)/Q$, the latter of which, δ_Q and its subcomponents, δ_S and δ_G will be identical and approximately equal to δ_L in well mixed lakes to maintain the isotope balance. While characterization of the isotopic composition of most components is possible through weighted sampling, δ_E is impossible to measure directly. Based on a linear resistance

---

[1] δ values express isotopic ratios as deviations in per mil (‰) from the Vienna-SMOW (standard mean ocean water), such that $\delta_{SAMPLE} = 1000((R_{SAMPLE}/R_{SMOW}) - 1)$, where R is ^{18}O/^{16}O or ^{2}H/^{1}H. Values cited herein are normalized on the SMOW-SLAP scale so that SLAP (standard light arctic precipitation) has a value of -55.5‰ in δ^{18}O and -428‰ in δ^2H (see Coplen, 1996).

Table 1
Summary of mass balance and isotope balance equations. Note that n/a indicates not applicable or useful

Mass	Isotope balance	
Balance	Fraction dependent	Time dependent
$dV/dt = I - Q - E$ (1)	'f-model': $$\delta_L = \delta_S - (\delta_S - \delta_0)f^{\left[\frac{-(1+mx)}{1-x-y}\right]} \quad (6)$$	By i steps using $I_i = Q_i + E_i$ and measured water levels: $$\delta_L = \delta_S - (\delta_S - \delta_0)\exp\left[-(1+mx)\frac{It}{V}\right] \quad (10a)$$
$I = Q + E$ (1a)	f = constant	't-model': $$\delta_L = \delta_S - (\delta_S - \delta_0)\exp\left[-(1+mx)\frac{It}{V}\right] \quad (9)$$
$dV/dt = -Q - E$ (1b)	$$\delta_L = \delta^* - (\delta^* - \delta_0)f^{\left[\frac{mE}{E+O}\right]} \quad (7a)$$	n/a
$dV/dt = I - E$ (1c)	$$\delta_L = \delta_S - (\delta_S - \delta_0)f^{\left[\frac{-(1+mx)}{1-x}\right]} \quad (7b)$$	By i steps using $I_i = E_i$, $V_i = V_{1-i} - E_{1-i}$: $$\delta_L = \delta_S - (\delta_S - \delta_0)\exp\left[-(1+m)\frac{Et}{V}\right] \quad (10b)$$
$I = E$ (1d)	f = constant	't-pan model': $$\delta_L = \delta_S - (\delta_S - \delta_0)\exp\left[-(1+m)\frac{Et}{V}\right] \quad (10c)$$
$dV/dt = -E$ (1e)	$$\delta_L = \delta^* - (\delta^* - \delta_0)f^m \quad (7c)$$	n/a

model as described by Craig and Gordon (1965) (but modified to utilize isotope data in per mil rather than as a decimal fraction, i.e. use $-15\%o$ rather than -0.015) δ_E can be estimated by

$$\delta_E = (\delta_L - h\delta_A - \varepsilon)/(1 - h + \varepsilon_K/1000) \quad (3)$$

where h is the ambient humidity normalized to the saturation vapor pressure at the temperature of the air–water interface, δ_A the isotopic composition of ambient moisture, and $\varepsilon = \varepsilon^* + \varepsilon_K$ the total isotopic fractionation comprised of both equilibrium ε^* and kinetic ε_K components. Note that ε^* is a function of temperature

[2] ε values represent instantaneous isotopic separations in per mil between liquid and vapor, such that $\varepsilon_{LIQUID-VAPOR} = 1000((R_{LIQUID}/R_{VAPOR}) - 1) \approx (\delta_{LIQUID} - \delta_{VAPOR})$. ε^* represents conditions where liquid and vapor are in equilibrium. Typical values range from 9 to 12‰ in $\delta^{18}O$ and 80–110‰ in δ^2H for T (temperatures) of 0–25 °C; ε_K represents the additional isotope separation due to diffusive kinetic effects. Typical values range from 2.9 to 6.4 in $\delta^{18}O$ and 2.5 to 5.6 in δ^2H for relative humidity in the range of 80–55‰ ($h = 0.8$–0.55).

(T), whereas ε_K is controlled by the turbulent/diffusion mass transfer mechanisms and humidity.[2] Under constant atmospheric and hydrologic conditions (i.e. hydrologic steady-state, such that $dV/dt = 0$ in Eq. (1)), air–water isotopic exchange will proceed and the lake water will enrich (or deplete) to an isotopic steady-state reflective of the isotopic and hydrologic characteristics of the system. Perturbations from steady state are commonly observed due to normal variability in atmospheric and hydrologic parameters and their isotopic composition.

Substitution of Eq. (3) into Eq. (2) assuming well-mixed conditions yields

$$V\frac{d\delta_L}{dt} + \delta_L\frac{dV}{dt} = I\delta_I - Q\delta_L$$
$$-\frac{E}{1 - h + \varepsilon_K/1000}(\delta_L - h\delta_A - \varepsilon) \quad (4)$$

which can be further simplified given various water balance scenarios as discussed below.

1.1.1. Fraction-dependent models

Fraction- (or volume-) dependent models have been shown to be most applicable in the case of lakes with isotopic composition that is evolving in response to volume reduction (i.e. desiccating ponds or evaporation pans) or volume increase (i.e. flooded ponds). If we define the remaining fraction of lake water as $f = V/V_o$, where V is the residual volume and V_o the original volume (and noting that $df = dV/V_o = (I - Q - E)dt/V_o$), then Eq. (4) can be simplified by substitution of the relations $V = V_o f$, $V_o = (I - Q - E)dt/df$, and from the basic theorem of calculus $f/df = 1/d \ln f$, to obtain (Gonfiantini, 1986)

$$d\delta_L/d \ln f = 1/(I - Q - E)[I(\delta_I - \delta_L)$$

$$+ mE(\delta^* - \delta_L)] \tag{5}$$

where $m = (h - \varepsilon/1000)/(1 - h + \varepsilon_K/1000)$ as defined by Welhan and Fritz (1977) and Allison and Leaney (1982), $\delta^* = (h\delta_A + \varepsilon)/(h - \varepsilon/1000)$ is the limiting isotopic composition under local meteorological conditions as defined by Gat and Levy (1978) and Gat (1981).

Assuming constant values for δ_I, δ_A, ε^*, ε_K, h, I, Q and E, and integrating Eq. (5) yields (Gonfiantini, 1986)

$$\delta_L = \delta_S - (\delta_S - \delta_0)f^{\left[\frac{-(1+mx)}{1-x-y}\right]} \tag{6}$$

where $\delta_L(f)$ describes the change in the isotopic composition of the lake with f, δ_0 the initial isotopic composition of lake water, $\delta_S = (\delta_I + mx\delta^*)/(1 + mx)$ is the steady-state isotopic composition the lake will attain as $f \to 0$, $x = E/I$ the fraction of lake water lost by evaporation, and $y = Q/I$ the fraction of lake water lost by liquid or isotopically non-fractionating outflows (Gonfiantini, 1986). Various integrated forms of Eq. (6) for simplified mass balance situations are given in Table 1.

The equations have the general form

$$\delta_L = \underbrace{\delta_K}_{\text{steady-state term}} - \underbrace{(\delta_K - \delta_0)}_{\text{transient fractional term}} f^\psi \tag{7}$$

where δ_K is given by δ^* for situations where inflow is negligible, or by δ_S when the lake has appreciable inflow. For a given interval, the transient term is defined by the incremental isotopic enrichment

$\delta_K - \delta_0$, f, and a rate constant ψ describing the enrichment slope. Note that $\delta_L \to \delta_K$ as $f \to 0$, where the transient term vanishes and the equations revert to steady-state conditions, and that $\delta_S = \delta_0$ when $f = 1$.

1.1.2. Time-dependent models

In situations where volumetric changes are negligible and lake volume can be assumed constant (i.e. $I = Q + E$), it is possible to describe isotopic changes in lake water over time t by assuming $dV/dt = 0$ in Eq. (4) which yields (Gonfiantini, 1986)

$$d\delta_L/dt = -[(1 + mx)\delta_L - \delta_I - x\delta^*](I/V) \tag{8}$$

where $x = E/I$ is again the fraction of lake water lost by evaporation, and in this case $1 - x = Q/I$ is the fraction of water lost to liquid outflows.

Integrating using mean values of δ_I, δ_A, ε^*, ε_K, h, V, I, Q, and E over the given integration period yields (Gonfiantini, 1986)

$$\delta_L = \delta_S - (\delta_S - \delta_0)\exp[-(1 + mx)(It/V)] \tag{9}$$

where $\delta_L(t)$ describes the change in isotopic composition of the lake with time t.

Various integrated forms of Eq. (8) for more simplified mass balance situations are given in Table 1. In general, the equations have the form

$$\delta_L = \underbrace{\delta_S}_{\text{steady-state term}} - \underbrace{(\delta_S - \delta_0)\exp \chi}_{\text{transient exchange term}} \tag{10}$$

For a given interval, the transient term is defined by the incremental isotopic enrichment $(\delta_S - \delta_0)$, and a time-constant χ describing the enrichment slope. Note that $\delta_L \to \delta_S$ as $t \to \infty$, where the transient term vanishes and the equations revert to steady-state conditions (see time-dependent equations, Table 1). These isotopic enrichment equations, in conjunction with time-series isotope sampling and meteorological measurements can be used to estimate water balance parameters by best-fit or step-wise approaches (Gonfiantini, 1986). Two operationally useful step-wise approximations are also included in Table 1 for situations where dV/dt may be small and only minor errors are introduced by assuming constant volume over short time intervals dt. In this case, V must then be modified for subsequent time steps in order to minimize approximation errors.

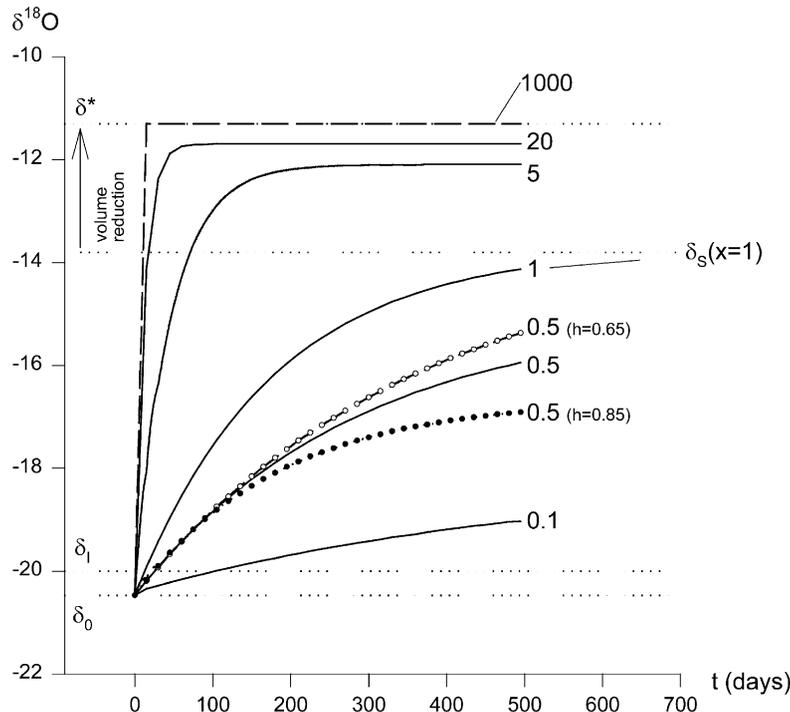An increase in the fraction of water lost by

Fig. 2. Schematic time-series of $\delta^{18}O$ enrichment predicted for a lake for a range of x (E/I) simulated by Eqs. (9), (10b) and (10c). Note that δ_S $(x = 1)$ corresponds to $E = I$ which is the maximum potential enrichment for a water body without volume reduction. Lakes with $E > I$ will evaporate toward dryness at which time their composition will range between δ_S $(x = 1)$ and the limiting isotopic enrichment δ^*. δ_I refers to inflow and δ_0 refers to the initial isotopic composition of the lake. Base case scenario adopts $h = 0.75$ and $\delta_A = -30‰$, but effect of varying humidity on lake with $x = 0.5$ is also shown.

evaporation x (E/I) will increase both the rate of isotopic enrichment and the ultimate δ_S that the water body will achieve, as shown schematically in Fig. 2 for a water body described by time-dependant formulae which illustrates the principle responsible for water balance scenarios identified in Table 1. As required, the isotope mass balance approach can be adapted or simplified to accommodate any given natural water balance scenario. Choice of a suitable isotope balance equation is governed by the mass balance situation described in Eqs. (1) and (1a)–(1e). Note that the equations summarized herein require slight modification in cases where salinity is significantly greater than seawater, such that the thermodynamic activity of water is substantially less than 1 (see Horita, 1990).

1.2. Field site, methods and observations

Field studies were conducted near Lupin, Nunavut,

Canada (Fig. 3), a gold mining operation situated on the south shore of Contwoyto Lake within the continental Arctic. The mine site is located about 90 km south of the Arctic circle, 375 km north of Yellowknife, and 125 km above northern treeline (Fig. 3b). The study area is typical of the low Arctic regions, with less than 300 mm of annual precipitation, mean annual temperatures of $-12\,°C$, and extreme annual variation in monthly temperature of $45\,°C$. 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. Local topographic relief is subdued and runoff is disorganized creating an abundance of shallow lakes ranging from 4 to 100 ha. From climatological maps, small lake evaporation at the site is estimated between 200 mm a^{-1} (den Hartog and Ferguson, 1978) and 325 mm a^{-1} (Morton, 1979). Strong evaporation gradients of about 50 mm a^{-1} per 100 km are predicted in the region
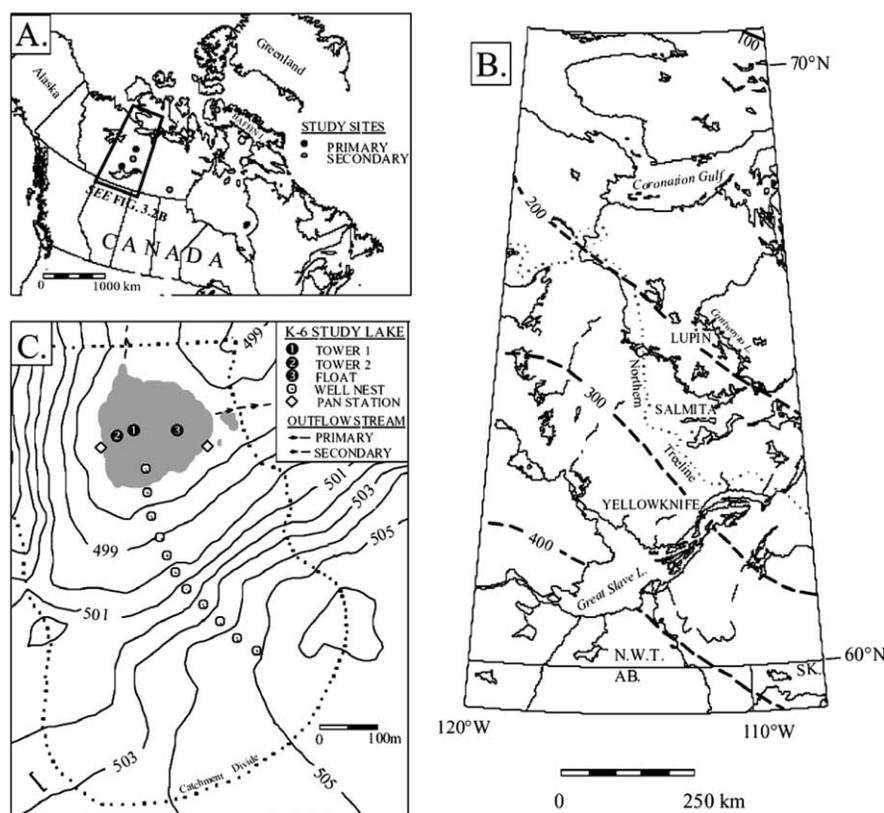
Fig. 3. (a) Study site locations, (b) enlargement showing Lupin, Yellowknife and estimated mean annual lake evaporation contours in millimeter (after den Hartog and Ferguson, 1978), and (c) K-6 study lake, catchment area and instrumentation.

with annual totals decreasing towards the northeast (Fig. 3b).

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 intensive study ('K-6 L', Fig. 3c). 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. Five nearby lakes (4.4–30.2 ha) and two mine-tailings ponds (20.5 and 79.3 ha) were selected for ancillary study. Dwarf tundra vegetation is common on all well-drained catchment slopes, while grass and sedge dominate low-lying areas and lakeshores. Continuous permafrost is known to extend to 510 m depth based on observations in the Lupin mine workings.

1.2.1. Water sampling and instrumentation

Water samples for isotopic analysis were manually collected at regular intervals during June, July and August 1992 and 1993 from all lakes, two evaporation pans and input sources, including lateral inflow and precipitation. In all cases, sampling sites were carefully chosen to be more-or-less representative of each water balance component. For lakes, 1-L samples were normally collected several meters from shore at mid-depth and then transferred to smaller sampling bottles. 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. Details of active layer water sampling and gauging are presented later on. Two class-A pans situated near the lakeshore were monitored by standard procedures and sampled at 5–7 days intervals in order to derive δ_A using the method of Gibson et al. (1999). The
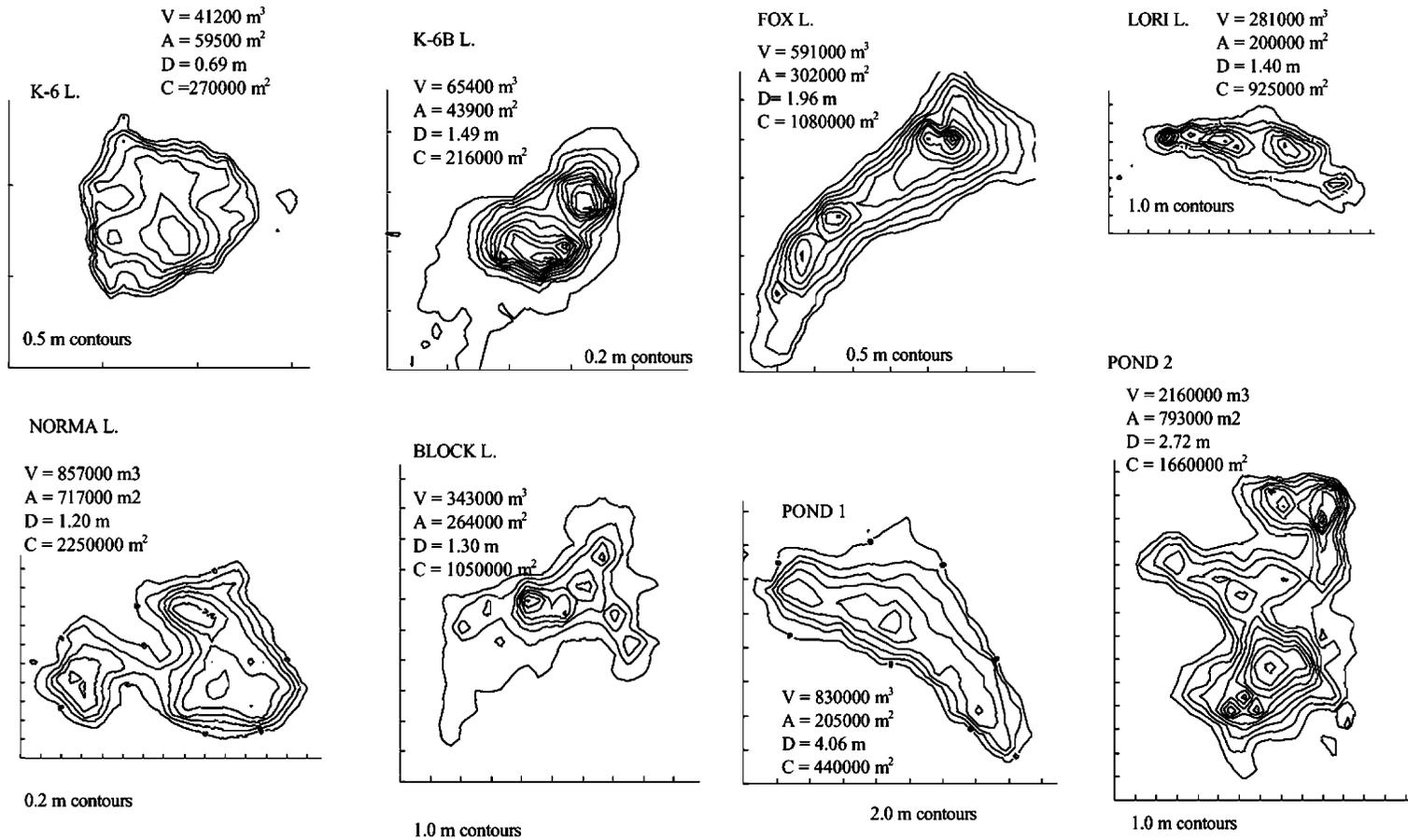
K-6 L.

V = 41200 m³
A = 59500 m²
D = 0.69 m
C = 270000 m²

0.5 m contours

K-6B L.

V = 65400 m³
A = 43900 m²
D = 1.49 m
C = 216000 m²

0.2 m contours

FOX L.

V = 591000 m³
A = 302000 m²
D= 1.96 m
C = 1080000 m²

0.5 m contours

LORI L.   V = 281000 m³
A = 200000 m²
D = 1.40 m
C = 925000 m²

1.0 m contours

NORMA L.

V = 857000 m3
A = 717000 m2
D = 1.20 m
C = 2250000 m²

0.2 m contours

BLOCK L.

V = 343000 m³
A = 264000 m²
D = 1.30 m
C = 1050000 m²

1.0 m contours

POND 1

V = 830000 m³
A = 205000 m²
D = 4.06 m
C = 440000 m²

2.0 m contours

POND 2

V = 2160000 m3
A = 793000 m2
D = 2.72 m
C = 1660000 m²

1.0 m contours

Fig. 4. Bathymetric contour maps of lakes near Lupin mine, Nunavut. Grid intervals are 100 m. Note that scale and contour intervals vary (see each map). *V*: mean volume, *A*: mean area, *D*: mean depth and *C*: catchment area.
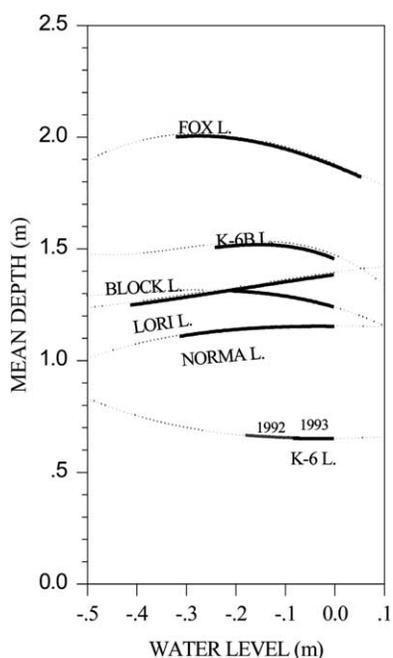
Fig. 5. Plot of mean depth versus mean water level for Lupin study lakes. Water level is relative to that observed on 11 June 1993 (day 162). Solid lines indicate range observed for 1993. Dashed line indicates 1992 range for K-6 L. Dotted lines indicate predicted values from catchment and bathymetric survey data. Note that mean depth changes by less than about 7% for all lakes.

method utilizes the isotope balance of the evaporation pan, as described by (10c) to determine the δ_A. All samples were collected in HDPE (high density polyethylene) bottles and returned to the Environmental Isotope Laboratory, University of Waterloo and analyzed within several months for oxygen and hydrogen stable isotope ratios. Analytical uncertainties are approximately $\pm 0.1\permil$ for $\delta^{18}O$ and $\pm 2\permil$ for δ^2H. An extensive micrometeorological observation program was also undertaken within the K-6 study lake in support of parallel evaporation studies using aerodynamic profile, Bowen ratio energy balance, and physical water balance methods (Gibson et al., 1996a,b). Humidity, air temperatures and water temperatures were monitored at 30 s intervals and recorded hourly at the K-6 study site. Water temperatures were also monitored at the time of sampling for other lakes and tailings ponds.

1.2.2. Lake characteristics

To define the lake volume V, multiple transects of lake-bottom topography were mapped for each lake using a sonic transmitter–receiver with a continuous chart recorder operated from an inflatable motorboat. Surveys of the lakeshore area of each catchment and periodic measurements of lake level were also conducted using a computerized theodolite. To define dV/dt and f, water levels within the K-6 Lake were also monitored continuously using both an ultrasonic depth gauge and a pressure transducer. From digitized topographic and water level records, the computer program SURFER™ (Golden Software Inc.) was used to construct bathymetic contour maps of each lake (Fig. 4), as well as estimates of lake volume, lake area, mean depth and variations of all parameters with lake level. A summary of mean depth and lake level is shown in Fig. 5. Catchment areas, with the exception of the K-6 catchment which was surveyed in detail by computer theodolite (650 points), were estimated from NTS 1:50,000 scale topographic maps in addition to high-resolution maps compiled by Lupin mine personnel.

Although shallow lakes in northern Canada are often well-mixed during the ice-free period (Stewart and Rouse, 1976a,b; Bello and Smith, 1990), additional water temperature data and a limited number of water samples were also collected over the course of the study to monitor the potential for incomplete mixing of selected lakes. Depth and spatial variations in isotopic composition within K-6 Lake (6 ha) were found to be minor during the ice-free period. Daily temperature distributions were found to be uniform within 1 or 2 °C, also supporting the assumption that the lake was well mixed. Temperature and isotopic profiles of the larger Pond 2 tailings reservoir (79 ha) also suggested that it was ostensibly well mixed during the ice-free period, despite periodic artificial inputs of water decanted from Pond 1 (see Gibson, 1996).

1.2.3. Active layer inflow

Lateral inflow to the study lake was monitored and sampled along a transect of shallow well nests within the study catchment (Well Nests, Fig. 3c). Frost tables were also monitored at 50 locations along the well transect at regular intervals throughout the thaw season. During 1993, a shallow irregular frost table combined with abundant snowmelt produced overland flow in rills on catchment slopes from early to mid-June. The active layer remained unsaturated for most of the thaw season. Pockets of isolated groundwater were observed in the
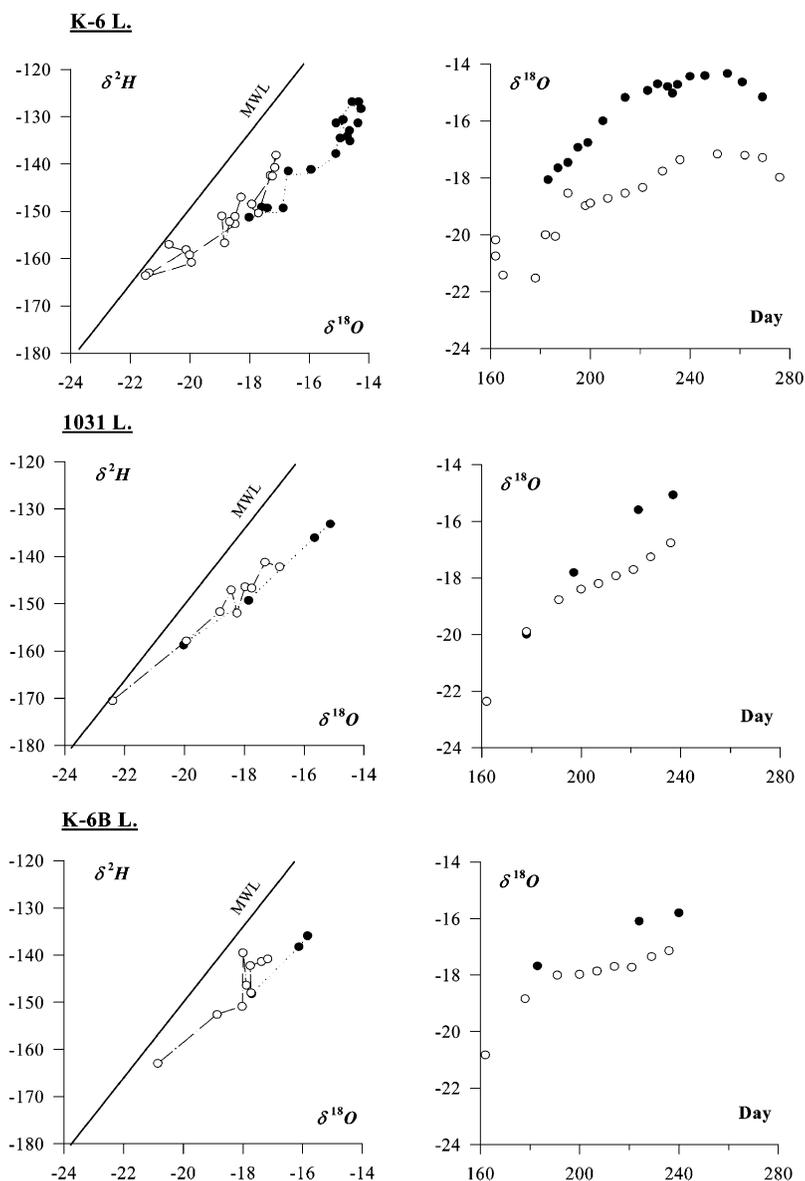
Fig. 6. Plots of (i) ^{2}H–^{18}O and (ii) time-series $\delta\,^{18}$O of lake water for the Lupin study lakes (solid dots: 1992 samples, hollow dots: 1993 samples, lines joining points depict temporal trends). Note that global meteoric water line (MWL) of 5^2H $= 8\delta\,^{18}$O $+ 10$ (Craig, 1961) defines the approximate trend for precipitation input to the lakes.

mid-slope areas by early August and saturated conditions and overland flow occurred once again by late-August as precipitation input exceeded the storage capacity of the active layer. During 1992, similar frost conditions were observed, however widespread saturation of the active layer did not occur up until the end of August due to the limited amount of total rainfall. Overall, saturated groundwater flow was never a dominant source of input to the lake, but was likely a minor source of inflow during periods of overland flow (Gibson, 1996). Estimates of inflow were calculated from records of precipitation, mean active layer water
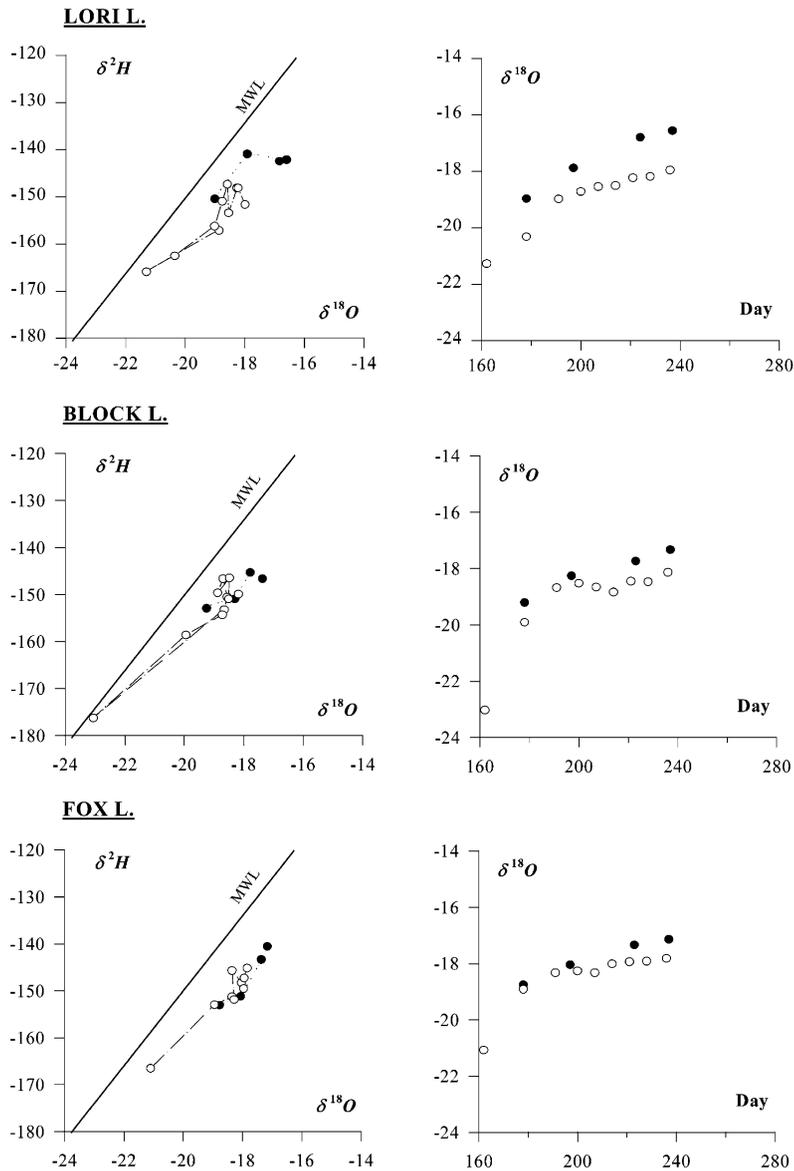
Fig. 6 (*continued*)

storage and storage capacity (Gibson et al., 1996a). Due to the presence of thick regional permafrost, deep groundwater movement was presumed to be negligible. Similar responses were noted for hill slopes in the other lake catchments.

Water samples were collected from the snowpack, overland flow, depression storage, the unsaturated zone, and from wells in order to characterize the isotopic composition of lateral inflow sources and runoff. Lateral inflow during 1992 and 1993 is estimated to be approximately -19.0 and $-20.0‰$, respectively for $\delta^{18}O$ (Gibson, 1996).

1.2.4. Evaporative enrichment

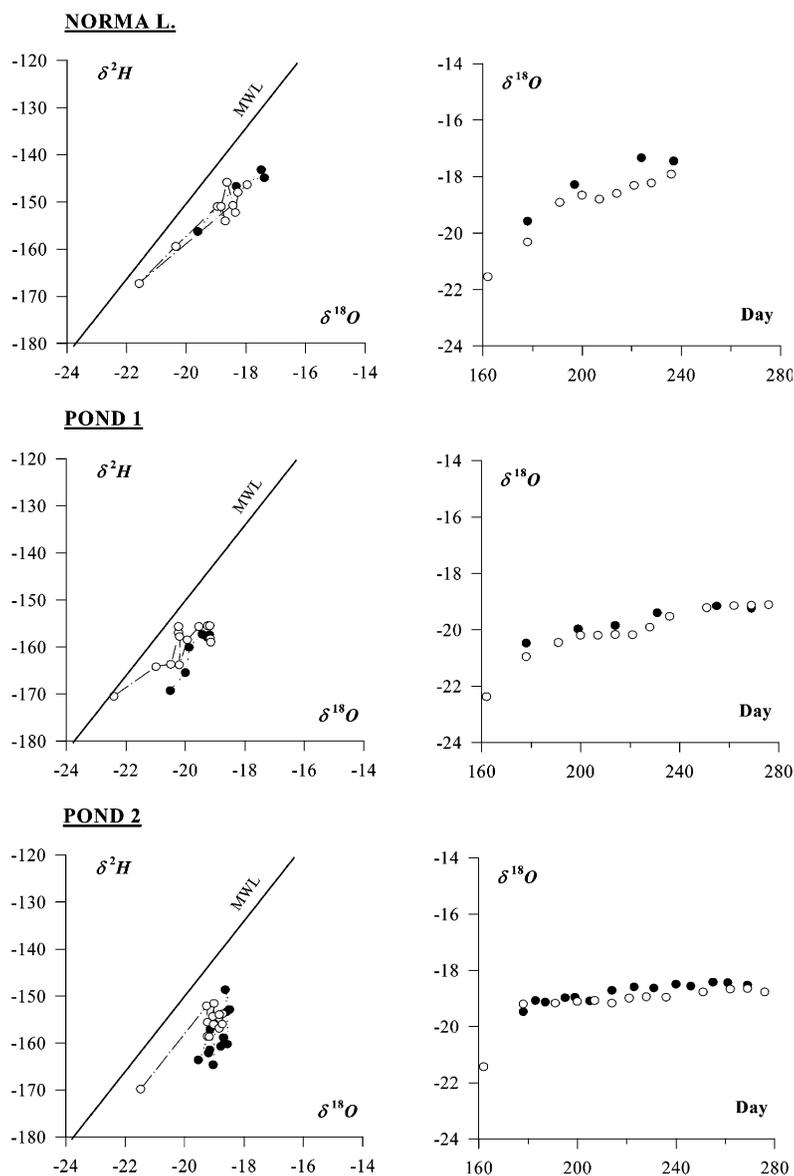Systematic isotopic variations were observed in lake waters during the course of the thaw season

Fig. 6 (*continued*)

period in response to changing water balance conditions. As shown in Fig. 6, lake waters plot below the MWL of Craig (1961) in ^{2}H–^{18}O space and seasonal fluctuations generally occur along trends with slopes of about five, characteristic of isotopic changes in open-water bodies subject to evaporation (Craig and Gordon, 1965). Distinct temporal isotopic enrichment patterns are noted

for 1992 and 1993 as illustrated in time-series plots of δ^{18}O (Fig. 6).

During 1992, water sampling was initiated shortly after open-water conditions were observed in late-June (day 178) and samples were collected periodically throughout July and August. The detailed study lake, K-6 L, and the two mine-tailings reservoirs, Pond 1 and Pond 2 were sampled more frequently

Table 2
Summary of best-fit model parameters and results for lakes during 26 June–27 August 1992 (day 178–240), Lupin, Nunavut

Reservoir[a]	Input parameters						Regression output			Volume-calibrated depth equivalents			
	h	T_a	δ_I	δ_A	δ^*	δ_S	$\Delta \ln(\delta_S - \delta_L)/\Delta t$	r^2	x	E (mm/day)	E (mm)	I (mm)	Q (mm)
K-6 L	0.64	10.7	−19.0	−28.3	−7.6	−13.3	−0.025	0.962	0.446	3.8	238	532	295
1031 L	0.64	10.7	−19.0	−28.3	−7.6	−13.6	−0.026	0.962	0.402	–	–	–	–
K-6B L	0.64	10.7	−19.0	−28.3	−7.6	−15.2	−0.025	0.999	0.164	4.5	277	1682	1405
Lori L	0.64	10.7	−19.0	−28.3	−7.6	−15.8	−0.025	0.998	0.174	4.3	269	1546	1276
Block L	0.64	10.7	−22.0	−28.3	−7.6	−16.1	−0.015	0.983	0.308	3.6	221	716	495
Fox L	0.64	10.7	−19.0	−28.3	−7.6	−16.5	−0.022	0.999	0.125	4.1	257	2052	1794
Norma L	0.64	10.7	−22.0	−28.3	−7.6	−16.3	−0.019	0.916	0.292	4.1	251	859	608
Pond 1	0.64	10.7	−20.0	−28.3	−7.6	−18.2	−0.015	0.899	0.073	2.4	221	3030	2810
Pond 2	0.64	10.7	−20.0	−28.3	−7.6	−18.0	−0.025	0.962	0.058	2.3	171	3882	3711
Contwoyto L	0.64	10.7	−22.0	−28.3	−7.6	−19.2	−0.009	0.997		–	–	–	–

[a] Arranged from smallest to largest volume (see Fig. 2).

over a protracted period extending to the time of freeze-up in late-September. High evaporation rates, low precipitation and minor lateral inflow during 1992 led to substantial isotopic enrichment in lakes by late-August. Isotopic enrichment was found to be rapid and more-or-less unidirectional. Reversals in the isotopic enrichment trend were observed during periods of reduced evaporation rates and higher inflow, as evident in the case of K-6 L during a period of cold-air intrusion on day 233 and in the several weeks prior to freeze up (K-6 L, Fig. 6).

During 1993, similar isotopic responses were observed for the ice-free period (day 191–271), however isotopic enrichment was relatively subdued due to colder, wetter conditions which resulted in lower rates of evaporation, higher precipitation and increased lateral inflow. Such findings are consistent with the predicted response of lakes for changing water balance conditions as characterized by x (E/I) (see Fig. 2).

It is important to note that isotopic changes observed prior to complete ablation of lake-ice during the early melt period (day 160–191, 1993) are not related primarily to evaporative enrichment processes but rather arise from mixing between variable contributions of snowmelt, lake-ice melt and sub-ice lake water (see Fig. 1b). As application of theory developed in Section 1.1 should be restricted to ice-free periods when the lake is well mixed, it was very important to carefully document ice-cover conditions.

This point was also noted in an ongoing study of isotopic variations in lakes on Baffin Island (Sauer, 1997) and at additional mine sites in the Northwest Territories and Nunavut, Canada (Gibson, 1996).

2. Results and discussion

Several isotope mass balance models were applied to estimate evaporation and outflow from lakes at Lupin, Nunavut based on the data described in Section 1.2 and temporal records of δ_A reported previously (Gibson et al., 1999). The isotope data in Fig. 6 clearly show the different enrichment patterns in wet and dry years, with enrichment levels varying in each lake. To predict the evaporation and water balance using the isotope data, a best-fit model and three step-wise models were applied as discussed below.

2.1. Application of isotope balance models

2.1.1. Best-fit model

Eq. (9) can be rearranged to obtain

$$\ln\left(\frac{\delta_S - \delta_L}{\delta_S - \delta_0}\right) = -(1 + mx)\frac{It}{V} \quad (11)$$

which permits estimation of both δ_S and x from linear regression of $\ln(\delta_S - \delta_L)$ provided that h, T, δ_I and δ_A are known and volumetric changes are negligible. Similar methods have been used by Fontes and
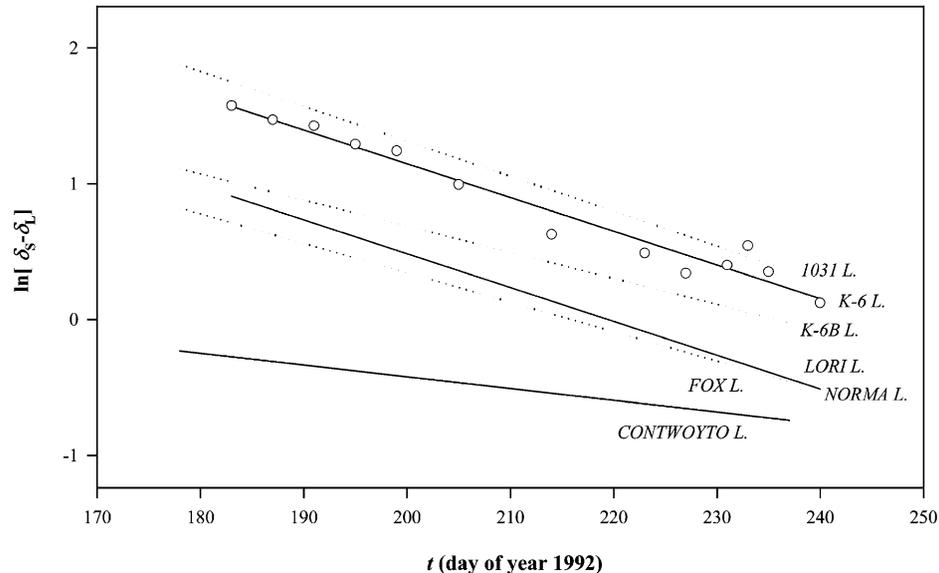
Fig. 7. Plot of $\ln(\delta_S - \delta_L)$ versus t for selected lakes at Lupin, NWT illustrating best-fit regression lines. For each reservoir, derived isotopic steady-state value δ_S provides the mean evaporation/inflow ratio x for the period. Derived slopes are used to solve explicitly for I/V, and lake volume and area estimates are used to calibrate depth equivalents of E, I, and O. Negative slopes indicate that lakes are tending to isotopic steady-state. Individual water samples for K-6 L are also plotted.

Gonfiantini (1967) to derive m and δ_S from ephemeral ponds in the Sahara using Eq. (7a), by Welhan and Fritz (1977) and Allison and Leaney (1982) to derive m and δ_S from evaporation pans using Eqs. (7c) and (10c), respectively, and by Zimmerman (1979) to derive δ_S and x using Eq. (9).

The best-fit model was applied only to time-series records of $\delta^{18}O$ for mid-summer, ice-free periods which were typified by progressive evaporative enrichment. Results for 1992 are summarized in Table 2, and $\ln(\delta_S - \delta_L)$ versus t regressions are illustrated in Fig. 7.

The regression results are reminiscent of plots presented by Zimmerman (1979) describing isotopic evolution in newly formed artificial lakes as they readjusted to contemporaneous conditions. In the present setting, transient isotopic variations are perpetuated mainly by the counteractive effects of spring snowmelt mixing and high rates of evaporation over the subsequent mid-summer period. Shallow lakes are distinguished from Contwoyto Lake, as illustrated in Fig. 6, by more negative regression slopes, reflecting higher I/V ratios and more rapid temporal isotopic changes. Minor volumetric decrease which occurred in most lakes, although restricted to

less than about 7%, caused only slight downward concavity in some of the plots, but this effect was generally subdued.

Weaker and less-regular isotopic enrichment prevented use of the best-fit method entirely during the 1993 mid-summer period. Application of under these conditions, when relatively little information is preserved in the slope $\Delta \ln(\delta_S - \delta_L)/\Delta t$ resulted in lack of convergence on δ_S. (Although it may have been possible at least for K-6 Lake, to apply the method over two separate periods, i.e. day 200–225 and 225–260, see trends in Fig. 5, but this was not done).

Volume-calibrated depth equivalents of evaporation, inflow and outflow are presented in Table 2. Note that δ_I was taken as $-19.0‰$ for headwater lakes fed only by active layer sources, but proportionately adjusted values were used in the case of the Pond 1 and Pond 2 reservoirs in order to account for artificial inputs, and in the case of Block L, Norma L and Contwoyto L in order to account for input from late-lying snowbanks. Although reasonable results were obtained for 1992, the best-fit model was neither universally applicable nor did it adequately describe short-term isotopic fluctuations that were systematic
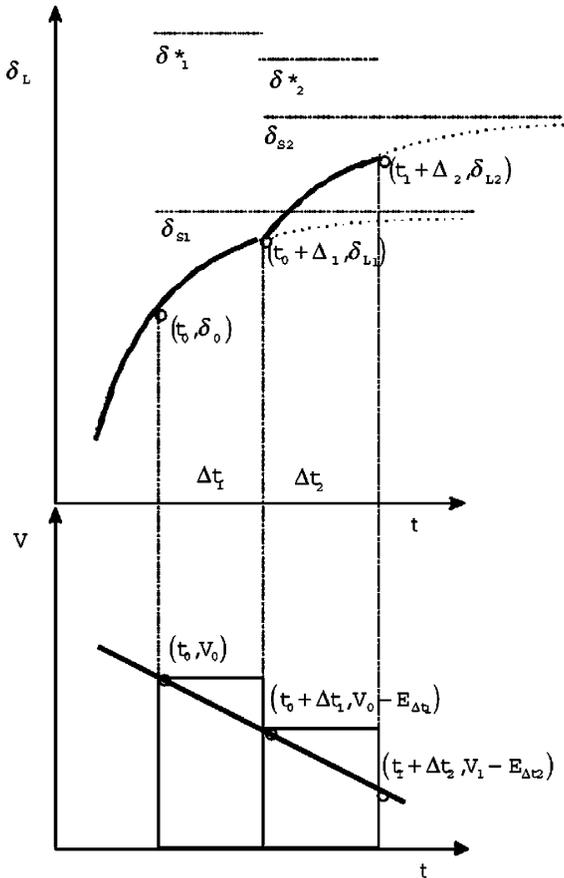
Fig. 8. (a) Schematic time-series plot of δ_L and V showing step-wise isotope balance scenario according to Eqs. (1) and (10a). For each time step Δt, δ^* is the limiting isotopic enrichment (i.e. the mean atmospheric forcing parameter), and δ_S is the steady-state enrichment (i.e. the mean atmospheric/hydrologic forcing parameter) which control isotopic enrichment of lake water (solid line). Subscripts refer to step numbers. Note that δ_I and δ_A are held constant during each time step such that $\delta_L \rightarrow \delta_S$ as $t \rightarrow \infty$. (b) Also shown is a useful volume approximation for times of zero outflow as described by Eqs. (1c) and (10b). Volume V is held constant during each time step (i.e. $I = E$) and subsequently adjusted for calculated evaporation.

and representative of changing water balance conditions. Overall, uncertainty in the assessments using best-fit models, as previously discussed in some detail by Zimmerman (1979) and Gonfiantini (1986) is usually no better than $\pm 30\%$ under most situations. Potential error is somewhat reduced in the current application ($\pm 20\%$) due to excellent temporal records available for h, T, δ_I and δ_A (see Gibson et al., 1999). Despite good distinction between water balance of

nearby lakes, the method is restrictive in terms of when it can be applied and is therefore not a universal tool for comparing nearby lakes in the Arctic and other highly seasonal systems. Best-fit models benefit from using all or a substantial number of lake water data to constrain estimates of water balance parameters.

2.1.2. Step-wise models

Step-wise models were also applied to investigate the potential for utilizing isotopic information under changing atmospheric and hydrologic conditions. This approach was needed to account for non-uniform and non-unidirectional isotopic changes as observed during 1993. The step-wise approach is discussed by Gonfiantini (1986) and was demonstrated in Gibson et al. (1996b, 1998). A plot of δ_L versus time (Fig. 8) illustrates the basis of the approach and the inter-relationship between the δ_S value for a given time interval and the short-term isotopic enrichment slope which is fixed by the isotope balance situation. Unlike the best-fit approach, one drawback of the step-wise approach is that the calculations are performed on individual isotope pairs (to define δ_0 and δ_L), and the enrichment slope is estimated rather than fitted.

Calculations were initially carried out for each lake and sampling interval given the most general balance situation described in Eqs. (1) and (6), hereafter referred to as the 'f-model'. For comparison, and specifically to test the sensitivity of the various water formulae under the current conditions, two simplified models were also applied, based on Eqs. (1a) and (9), which assumes no volumetric changes in the lakes, and Eqs. (1c) and (10b), which assumes no volumetric changes and no outflow from the lakes. These models are hereafter referred to as the 't-model' and the 't-pan model', respectively (see Table 1). Due to the minimal mean depth fluctuations in the lakes (Fig. 5) the t-model was thought to be a reasonable approximation. The t-pan model, which in addition assumes no outflow was thought to be fundamentally poorer representation of balance conditions except when the lakes were behaving like evaporation pans.

When compared, it was evident that the choice of model did not significantly influence the evaporation results obtained by fitting to the observed enrichment trends. Further analyses also showed that changing water balance parameters within a reasonable range also had little influence on results. Evidently, due to
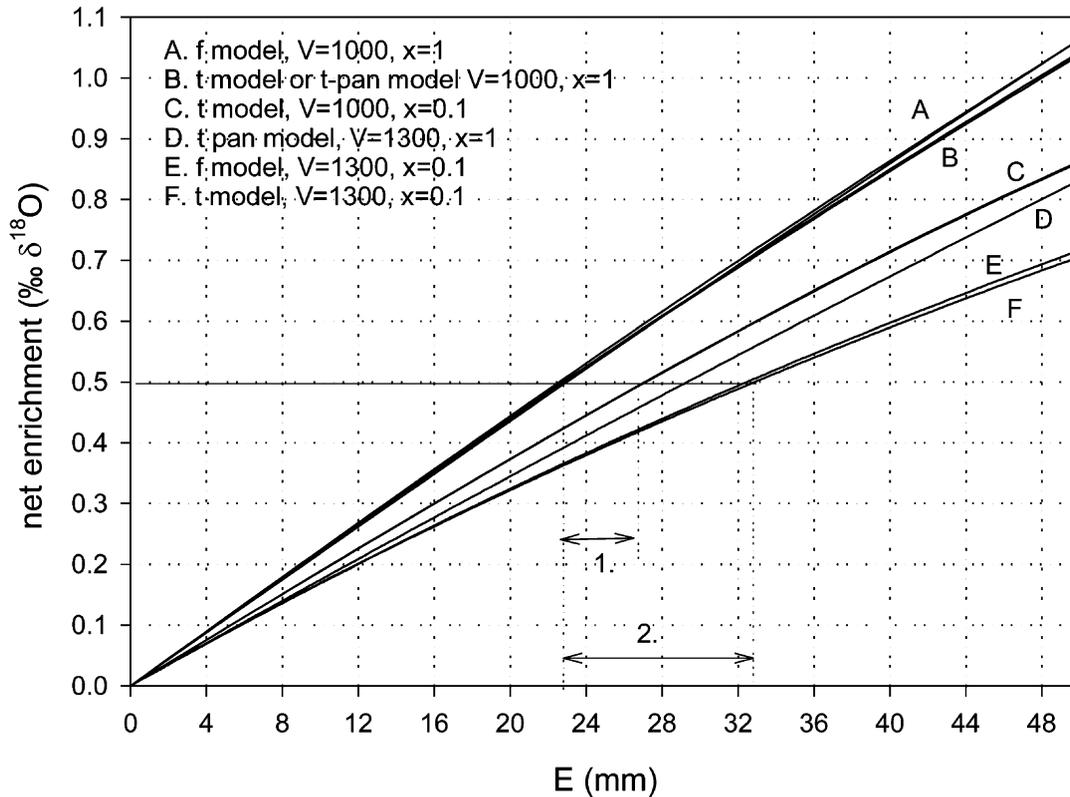
Fig. 9. Plot of enrichment in ^{18}O of lake water versus estimated lake evaporation (E) versus net enrichment in δ^{18}O of lake water (‰) for a lake of V (volume in mm), illustrating differences between f-model, t-model, and t-pan model. Note that A–F refer to various model scenarios, 1 refers to difference in evaporation estimated for scenarios B and C arising from uncertainty in the water balance and 2 refers to difference in evaporation estimated for B and F arising from uncertainty in water balance and a 30% error in estimate of reservoir volume. Note that enrichment slope is virtually identical for the f-model and the t-model, so that A–B and E–F are similar. Note that a change in x from 0.1 to 1 results in a change of only 17% of the evaporative enrichment signal.

extreme deviation from isotopic steady-state in the present setting (up to 6‰ in δ^{18}O or more in the smallest lakes), the rate of isotopic enrichment of lake water over short time intervals of a week or so is primarily controlled by evaporation rates and lake volume, whereas enrichment is only slightly attenuated by the water balance effect (i.e. in high through-flow situations where x is small). A schematic of evaporation versus isotopic enrichment for a shallow water body (1 m depth) based on the three models illustrates this effect (Fig. 9). As shown for an enrichment of 0.5‰ in δ^{18}O (which corresponds to an evaporation loss of 23 mm, a typical weekly total at the site) sensitivity to x variations in the range of 0.1–1 produces uncertainty in total evaporation estimates of about 4 mm or 15% of the enrichment signal (Fig.

9). Although water balance does exert ultimate control on the value of the isotopic steady-state (δ_S) and on the shape of the enrichment slope of the reservoir, the reservoirs are so depleted in heavy isotopes that any outflow effects are overwhelmed by the evaporation signals. Volume is a dominant control, as shown in Fig. 9, but this is accounted for in the models.

The close agreement between the f-model and t-model is not surprising considering that the models are expected to be identical for $f = 1$, and very similar for $f > 0.95$ which was met for all intervals examined. The similarity of the t-pan model and t-model is due to the fact that the enrichment slope given by $(1 + mx)I/V$ is between 3 and 9 times more sensitive to evaporation than to inflow or water balance due to the influence of the m
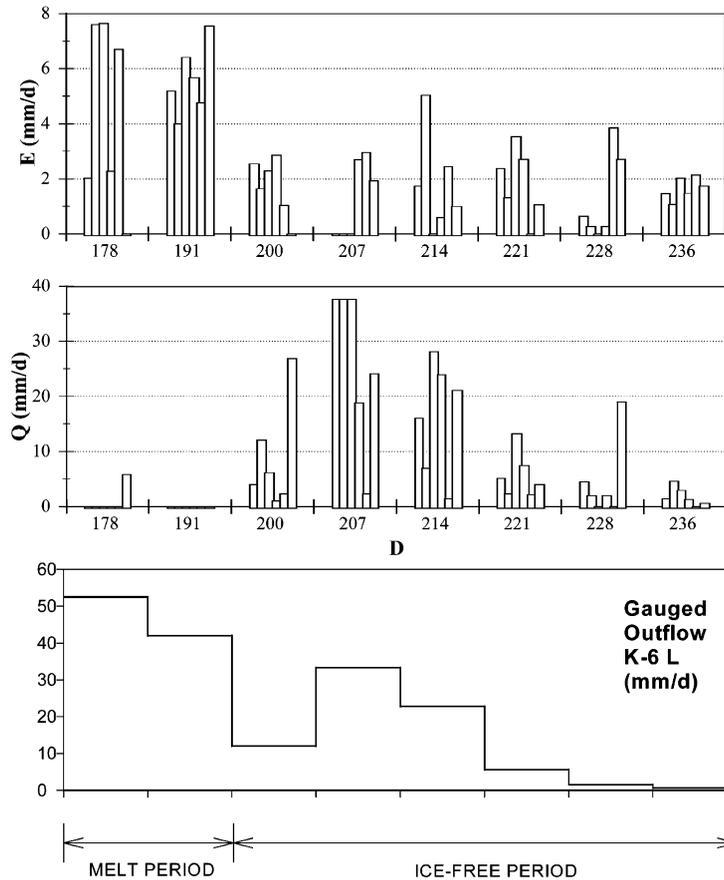
Fig. 10. Lake evaporation E in mm/day for Lupin lakes over 1993 time-intervals ending on the specified day as determined by step-wise method (see text). Note that the first interval began on day 162. Bars (left to right) are: Norma L, Fox L, Block L, Lori L, K-6B L, and K-6 L. Note that evaporation is over-estimated during the first two time intervals due to incomplete mixing during the melt period. Differences between lakes during subsequent time intervals are attributed to variability in evaporation rates (±errors). Errors are expected to be on the order of ±1 mm/day.

term in Eqs. (9) and (10a)–(10c).[3] Water balance information is therefore only strongly imprinted on the enrichment slope when the approach to steady state is established via best-fit approaches. Otherwise evaporation loss dominates the isotope enrichment signal. This realization was found to be quite useful for the application of the step-wise balance approach as discussed below.

Due to the predominant control of evaporation losses on the isotope enrichment signals, it was decided to apply the t-pan model to predict an initial

estimate of E for each time interval. This value, along measurements of dV/dt and estimates of I, was then plugged into the fully defined f-model and used to calculate Q. To account for the slight water balance effects discussed in relation to Fig. 9, this initial solution was then refined slightly by iterative substitution in Eqs. (1) and (6) until the estimates converged. This method was found to yield results for E and Q that were reasonable and consistent with other hydrological measurements carried out during the 1993 season (Fig. 10). In effect, as for the analysis depicted in Fig. 10, the isotopic enrichment was primarily used to estimate the evaporation losses, and outflow was then estimated as a residual of the water

---

[3] In the humidity range of 0.75–0.90 observed most frequently in this study.

balance. Similar to the best-fit approach, uncertainty in the use of the step-wise approximation in the current context is estimated at no better than ± 20%. This limit does however permit monitoring of significant differences in water balance of nearby lakes or artificial reservoirs if they indeed exist. Overall, step-wise models are found to be most useful for highly transient systems but the primary weakness is that the method relies on individual data points.

2.2. Water balance and evaporation estimates

Consistent and reasonable lake-to-lake patterns illustrated in Fig. 10 suggest that the f-model can potentially be utilized under similar seasonal conditions to characterize the evaporation rate and outflow of lakes on short time scales. From examining the estimates for the ice-free period (i.e. day 200–236, Fig. 10), a similar pattern of evaporation and outflow emerges for the nearby study lakes. Mean lake evaporation is estimated to be about 1.6 mm/day and weekly averages for individual reservoirs ranged from 0 to 5 mm/day. Of interest, trends in predicted evaporation and outflow are in agreement with conventional methods applied concurrently (Gibson et al., 1996a,b), and accurately depict the timing of outflow in comparison with physical gauging methods (Fig. 10). Estimates of evaporation and outflow for the melt period (day 162–200) are not in agreement with physical observations, due to complex processes operating during this season, as discussed in relation to Fig. 2.

Mean evaporation for the natural lakes over the mid-summer period was estimated to range from about 1.3 to 1.9 mm/day during 1993 and 3.8 to 4.5 mm/day during 1992. Lake-to-lake variability of this magnitude, on the order of ± 15–40%, is in broad agreement with previous studies of northern lakes (Marsh and Bigras, 1988; Bigras, 1990). Variability is likely somewhat masked by uncertainty in the estimates, which is in the order of ± 20%. The analysis does confirm that the nearby lakes with broadly similar characteristics do not have vastly different evaporation rates or water balances on a weekly time scale. To improve on this in future, it might be beneficial to sample the lakes daily and to apply best-fit methods on weekly time scales. The use of several lakes in such an analysis may also serve to reduce uncertainty in defining mean values of evaporation and water balance for local lakes.

Overall, results do suggest that local lake-to-lake effects are more subdued during a wet season (1993) and more pronounced during a dry season (1992). This may be the result of the advection influence, where sensible heat from the tundra landscape can significantly augment evaporation in smaller water bodies during dry years.

3. Concluding comments

It was found that evaporation rates predominantly control the rates of isotopic enrichment in small shallow lakes in a continental, low arctic setting due to extreme seasonality and extreme departures from isotopic steady-state during the early ice-free period. Non-steady isotope balance methods were applied in conjunction with conventional water balance analysis to constrain and compare short-term (weekly) water balance variations in nearby lakes. Although similar results were obtained for nearby lakes in the present study, the method is expected to be effective for identifying lakes with substantially different water balance characteristics.

This study revisits isotope mass balance concepts that have been developed and applied in various climatic regions over the past four decades, although few detailed studies have been conducted in cold region environments. The enhanced seasonality of the continental Arctic provided an ideal and unprecedented setting for experimental testing of basic field approaches for characterizing isotope mass balance within a highly variable atmospheric and hydrologic system. Current and future studies are focusing more and more on climatological water balance signals which are preserved in large, deep lakes which are more readily compared (Gibson, 2001; Gibson and Edwards, 2002). A regional survey of lakes across the continental Arctic, which are one or two orders of magnitude larger in volume than those examined in this study, reveals systematic trends corresponding to climatological evaporation loss and water balance signals. Nevertheless, better understanding of shallow arctic lakes is still required to design mitigation strategies appropriate for mine-tailings ponds in the region. These reservoirs are essentially modified

shallow lakes and are expected to acquire similar water balances in the post-operational phase of mining activities.

Acknowledgments

References

Allison, G.B., Leaney, F.W., 1982. Estimation of isotopic exchange parameters, using constant-feed pans. J. Hydrol. 55, 151–161.

Bello, R., Smith, J.D., 1990. The effect of weather variability on the energy balance of a lake in the Hudson Bay Lowlands, Canada. Arctic Alpine Res. 22, 98–107.

Bigras, S.C., 1990. Hydrological regime of lakes in the Mackenzie Delta, Northwest Territories, Canada. Arctic Alpine Res. 22, 163–174.

Coplen, T.B., 1996. New guidelines for reporting stable hydrogen, carbon, and oxygen isotope-ratio data. Geochim. Cosmochim. Acta 60, 3359–3360.

Craig, H., 1961. Isotopic variations in meteoric waters. Science 133, 1702–1703.

Craig, H., Gordon, L.I., 1965. Deuterium and oxygen 18 variations in the ocean and marine atmosphere. In: Tongiorgi, E., (Ed.), Stable Isotopes in Oceanographic Studies and Paleotemperatures, Lab. Geologia Nucleare, Pisa, pp. 9–130.

Dinçer, T., 1968. The use of oxygen-18 and deuterium concentrations in the water balance of lakes. Water Resour. Res. 4, 1289–1305.

Fontes, J.Ch., Gonfiantini, R., 1967. Comportement isotopique au cours de l'evaporation de deux bassins Sahariens. Earth Planet. Sci. Lett. 3, 258–266.

Gat, J.R., 1970. Environmental isotope balance of Lake Tiberias, Isotopes in Hydrology, IAEA, Vienna, pp. 151–162.

Gat, J.R., 1981. Lakes. In: Gat, J.R., Gonfiantini, R. (Eds.), Stable Isotope Hydrology—Deuterium and Oxygen-18 in the Water Cycle, IAEA Techn. Rep. Ser. 210, IAEA, Vienna, pp. 203–221.

Gat, J.R., Levy, Y., 1978. Isotope hydrology of inland sabkhas in the Bardawil area, Sinai. Limnol. Oceanog. 23, 841–850.

Gibson, J.J., 1996. Non-steady isotopic methods for estimating lake evaporation: development and validation in Arctic Canada. PhD Thesis. University of Waterloo, Canada, 238 pp.

Gibson, J.J., 2002. Forest-tundra water balance traced by isotopic enrichment in lakes. J. Hydrol. 251, 1–13.

Gibson, J.J., Edwards, T.W.D., 2002. Regional water balance trends and evaporation-transpiration partitioning from a stable isotope survey of lakes in northern Canada. Global Biogeochem. Cycles 16, in press.

Gibson, J.J., Edwards, T.W.D., Bursey, G.G., Prowse, T.D., 1993. Estimating evaporation using stable isotopes: quantitative results and sensitivity analysis for two catchments in northern Canada. Nordic Hydrol. 24, 79–94.

Gibson, J.J., Prowse, T.D., Edwards, T.W.D., 1996a. Evaporation from a small lake in the continental arctic using multiple methods. Nordic Hydrol. 27, 1–24.

Gibson, J.J., Edwards, T.W.D., Prowse, T.D., 1996b. Development and validation of an isotopic method for estimating lake evaporation. Hydrol. Process. 10, 1369–1382.

Gibson, J.J., Reid, R., Spence, C., 1998. A six-year isotopic record of lake evaporation in the Canadian Subarctic. Hydrol. Process. 12, 1779–1992.

Gibson, J.J., Edwards, T.W.D., Prowse, T.D., 1999. Pan-derived isotopic composition of water vapour and its variability in northern Canada. J. Hydrol. 217, 55–74.

Gonfiantini, R., 1986. Environmental isotopes in lake studies. In: Fritz, P., Fontes, J.Ch. (Eds.), Handbook of Environmental Isotope Geochemistry, vol. 3. Elsevier, New York, pp. 113–168.

den Hartog, G., Ferguson, H.L., 1978. Mean annual lake evaporation. Plate 17, Hydrol. Atlas Canada, Ottawa. Department of Fisheries and Environment, Ottawa, Ont., map.

Horita, J., 1990. Stable isotope paleoclimatology of brine inclusions in halite: modeling and applications to Searles Lake, California. Geochim. Cosmochim. Acta 54, 2059–2073.

Krabbenhoft, D.P., Bowser, C.J., Anderson, M.P., Valley, J.W., 1990. Estimating groundwater exchange with lakes 1. The stable isotope mass balance method. Water Resour. Res. 26, 2445–2453.

Marsh, P., Bigras, S.C., 1988. Evaporation from Mackenzie Delta lakes, NWT, Canada. Arctic Alpine Res. 20, 220–229.

Morton, F.I., 1979. Climatological estimates of lake evaporation. Water Resour. Res. 15, 64–76.

Sauer, P., 1997. Records of climate and late quaternary paleoclimate from stable isotopes in lakes and lake sediments, Eastern Canadian Arctic. PhD Thesis. University of Colorado at Boulder, USA, 224 pp.

Stewart, R.B., Rouse, W.R., 1976a. A simple method for determining evaporation from shallow lakes and ponds. Water Resour. Res. 12, 623–628.

Stewart, R.B., Rouse, W.R., 1976b. Simple models for calculating evaporation from wet and dry tundra surfaces. Arctic Alpine Res. 8, 263–274.

Welhan, J.A., Fritz, P., 1977. Evaporation pan isotopic behaviour as an index of isotopic evaporation conditions. Geochim. Cosmochim. Acta 41, 682–686.

Zimmerman, U., 1979. Determination by Stable Isotopes of Underground Inflow and Outflow and Evaporation of Young Artificial Groundwater Lakes, Isotopes in Lake Studies, IAEA, Vienna, pp. 87–94.

Zuber, A., 1983. On the environmental isotope method for determining the water balance of some lakes. J. Hydrol. 61, 409–427.