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Forest-tundra water balance signals traced by isotopic enrichment in lakes

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Abstract

Enrichment of oxygen-18 and deuterium in surface waters is shown to be a useful indicator of water balance variations in remote, permafrost regions of northern Canada where hydroclimate monitoring networks are limited. Necessarily, such indicators must be applied with care as isotopic signals in each lake trace water balance and atmospheric conditions integrated over different time-periods and spatial areas, as determined by lake water-residence time and catchment drainage area. Isotopic enrichment in large lakes is found to be relatively stable in time, and, as interpreted in the context of steady-state models, will yield information more representative of regional climatological-scale processes. In contrast, seasonal isotopic enrichment in shallow lakes, which occurs due to extreme seasonality of the northern climate and short water residence times, may be useful for estimating short-term 'point' evaporation rates. © 2001 Elsevier Science B.V. All rights reserved.

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1. Introduction

Physically-based hydrologic research, including the study of climate change and associated impacts on water resources, is severely hampered by poor spatial distribution and short length-of-record of hydrometric monitoring stations in northern Canada (Prowse, 1990). Hydrometric networks north of 60°N fall below standards recommended by the World Meteorological Organization (1981) mainly due to lack of coverage of inland areas. Such inadequacies have also complicated evaluation of more localized but rapidly expanding water development impacts associated with mining in the region (Latham, 1988). Increased awareness of the hydro-ecological sensitivity of permafrost environments to climate change and pollution has prompted ongoing development of water balance approaches to supplement hydrometric monitoring, including the use of isotopic tracers.

This study explores the use of oxygen and hydrogen stable isotopes to characterize changes in hydrologic regime across northern treeline. Research has focused on selected mine sites within subarctic forest, forest-tundra transition and tundra zones. An understanding of the isotopic response of lakes, rivers and watershed compartments to changing hydrologic conditions, which has been gained through parallel application of isotopic and non-isotopic methods in site-specific water balance studies (Gibson et al., 1993a,b, 1996a,b, 1999), is being used to evaluate systematic variations in isotopic enrichment of lakes at the sites. In addition to contributing baseline information on hydroclimatic variability, which has immediate practical value for design and management of mine-tailings ponds in the

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region, the study provides a better understanding of variations in water balance and related isotopic behaviour of primary interest to lake paleohydrological studies (Edwards, 1993).

A fundamental isotope mass balance approach is applied that utilizes measured isotopic enrichment in lakes as a quantitative index of partitioning between evaporation and non-fractionating outflow pathways. The theoretical basis of the method is outlined below.

1.1. Theoretical background

The water-mass and isotope-mass balance for a well-mixed reservoir, assuming constant density of water, may be written respectively as

$$\mathrm{d}V/\mathrm{d}t = I - Q - E \tag{1}$$

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

where V is the volume of the reservoir, t is time, $\mathrm{d}V$ is the change in volume over time interval $\mathrm{d}t$, I is inflow, Q is outflow, E is evaporation, and δ_L, δ_I, δ_Q and δ_E are the isotopic compositions of the reservoir, inflow, outflow, and evaporative flux, respectively. Providing that isotopic compositions of components in Eq. (2) can be measured or estimated, and given that systematic isotopic enrichment occurs during exposure to evaporation under normal climate conditions (Gat and Bowser, 1991; Gonfiantini, 1986), it is possible to combine Eqs. (1) and (2) to solve for two unknown water balance components. While characterization of the isotopic composition of most components is possible through weighted sampling, δ_E is difficult to measure directly. Using the linear resistance model of Craig and Gordon (1965), assuming zero resistance to transport in the liquid phase, δ_E can be estimated by

$$\delta_E = (\alpha^*\delta_L - h\delta_A - \epsilon)/(1 - h + \epsilon_K/1000) \tag{3}$$

where α^* is the equilibrium isotopic fractionation, h is the atmospheric humidity normalized to the saturation vapour pressure-temperature of the air–water interface, δ_A is the isotopic composition of ambient moisture, and $\epsilon = \epsilon^* + \epsilon_K$ is the total isotopic separation comprised of both equilibrium ϵ^* and kinetic ϵ_K components.[1] Equilibrium separation factors ϵ^* for both oxygen and hydrogen are well-defined as a function of temperature from laboratory experiments by Majoube (1971) and others. Kinetic separation factors ϵ_K, although the subject of considerable controversy in the past, are fairly well-described from theoretical and experimental studies. For time-scales relevant to water balance applications, ϵ_K and can be approximated by $C_K(1 - h)\cdot 10^3$ where $C_K = (D/D_i)^n - 1$, D and D_i are the molecular diffusion coefficients of the common ($^1H^1H^{16}O$) and heavy isotopic species ($^1H^1H^{18}O$ or $^1H^2H^{16}O$) respectively, and n is a turbulence parameter such that $n = 1/2$ for mean turbulent flow, $n = 2/3$ for laminar flow and $n = 1$ for static transport (Gonfiantini, 1986; Vogt, 1976; Merlivat, 1978a,b; Merlivat and Coantic, 1975; Brutsaert, 1975). As noted by Gonfiantini (1986), $n = 1/2$, which corresponds to C_K values of 14.3 and 12.5‰ for oxygen and hydrogen, respectively, appears to reasonably represent the conditions observed most frequently in nature.

Combining Eqs. (2) and (3) and integrating between the limits t_0 and t for time intervals where water balance fluxes and their isotopic compositions can be assumed constant and $\mathrm{d}V/\mathrm{d}t \approx 0$ (i.e. hydrologic steady-state) yields (Gonfiantini, 1986)

$$\delta_L = \delta_S - (\delta_S - \delta_0)\exp[-(1 + mx)(It/V)] \tag{4}$$

where δ_0 is the initial isotopic composition of the reservoir, $\delta_S = (\delta_I + mx\delta^*)/(1 + mx)$ is the steady-state isotopic composition the reservoir will attain as $t \to \infty$, $x = E/I$ is the fraction of reservoir water lost by evaporation, $m = (h - \epsilon)/(1 - h + \epsilon_K/1000)$ as defined in previous studies (Welhan and Fritz, 1977; Allison and Leaney, 1982) and $\delta^* = (h\delta_A + \epsilon)/(h - \epsilon/1000)$ is the limiting isotopic composition under local climate conditions (Gat and Levy, 1978; Gat, 1981).

In the special case where reservoirs are large enough to buffer transient isotopic variations related to seasonality of hydroclimate conditions, and when long time intervals are considered, it can be assumed that the lake is also close to isotopic steady-state

[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 $^2H/^1H$. $\delta^{18}O$ and δ^2H values cited herein are normalized to -55.5 and -428‰, respectively, for SMOW and SLAP (Standard Light Arctic Precipitation). See Coplen (1996). ϵ values represent instantaneous isotopic separations in per mil between co-existing liquid and vapour, such that $\epsilon_{LIQUID-VAPOUR} = 1000((R_{LIQUID}/R_{VAPOUR}) - 1) \sim ((\delta_{LIQUID} - \delta_{VAPOUR})$.

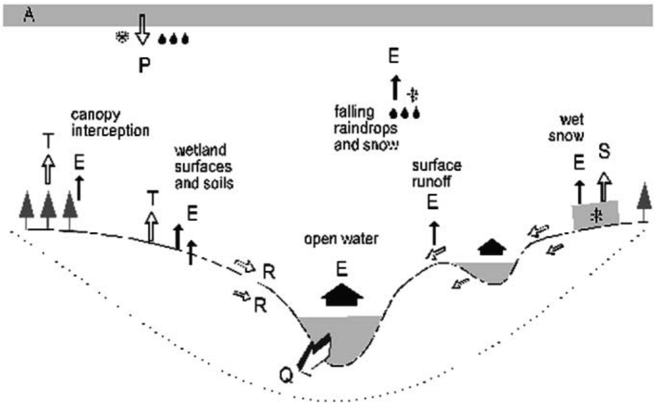

Fig. 1. Water balance schematic illustrating evaporative (solid arrows) vs non-fractionating processes (hollow arrows). Note that large lakes will inherit the isotopic enrichment signal of all evaporation (*E*) from water-contributing sources across the catchment as integrated over the residence time of lake water. Non-fractionating processes include transpiration (*T*), sublimation from snow (*S*), and water transport as surface/subsurface runoff (*R*) and outflow (*Q*). Precipitation (*P*) derived from the atmosphere (*A*) is also subject to fractionation during descent through unsaturated air or when a portion of canopy intercepted moisture is evaporated.

($\delta_L \approx \delta_S$) and Eq. (4) can be simplified to yield

$$x = E/I = (\delta_L - \delta_I)/[m(\delta^* - \delta_L)] \quad (5)$$

which is a key relationship describing the dependency between the water balance of a large reservoir x and its isotopic enrichment by evaporation $\delta_L - \delta_I$. Atmospheric controls on this enrichment include h and δ_A through their influence on m and δ^*.

By substituting $\delta_I = \delta_P$ into Eq. (5), this relationship may also be applied to estimate x from lake (or river) drainage basins, as characterized by the isotopic separation between the lake (or river) δ_L and precipitation δ_P. Providing lakes with long residence times are considered, it is probably sufficient to assume that storage changes in various compartments of the watershed are minor. In this case, isotopic enrichment at lowest level of the drainage hierarchy reflects partitioning between evaporation losses (over land and water) vs liquid outflow/transpiration losses integrated over the catchment area (Fig. 1). From a regional perspective, this separation ($\delta_L - \delta_P$) is less difficult to quantify as it can be determined by measuring or estimating the isotopic composition of precipitation and lake (or river) water without detailed sampling of sub-watershed compartments.

In applying Eq. (5) to watershed reservoirs as described above, it is important to note that some processes such as soil evaporation impart an isotopic enrichment signal which is distinct from that of open-water evaporation due to the dominance of static transport through the soil zone (Allison et al., 1983). In many cases however, as in environments with shallow permafrost, these processes occur in land areas with very low moisture contents which function essentially as closed reservoirs, and so contribute only negligible volumes of water to local lakes. Analogous 'reservoir' isotope enrichment effects are found in the case of water within the leaves of transpiring plants, where isotopic enrichment is not generally passed on to the transpiration flux under steady-state conditions (Wang and Yakir, 2000). Other high-closure compartments such as vegetation canopies where precipitation is intercepted, or high-closure processes such as evaporation from snow will not be isotopically traceable in the liquid phase if residual volumes completely volatize. This raises an important consideration with use of isotopic tracers, as discrepancies between catchment areas defined by hydrometric networks (i.e. based on topographic relief) and isotope-mass balance methods (i.e. based on water-contributing areas) therefore exist. Although a detailed understand of such high-closure processes is extremely important for standard hydrometric approaches and for coupled atmosphere-surface modelling, their net influence (which may be small) is intrinsically weighted in the isotope composition of runoff.

Time-lag of water movement from watershed

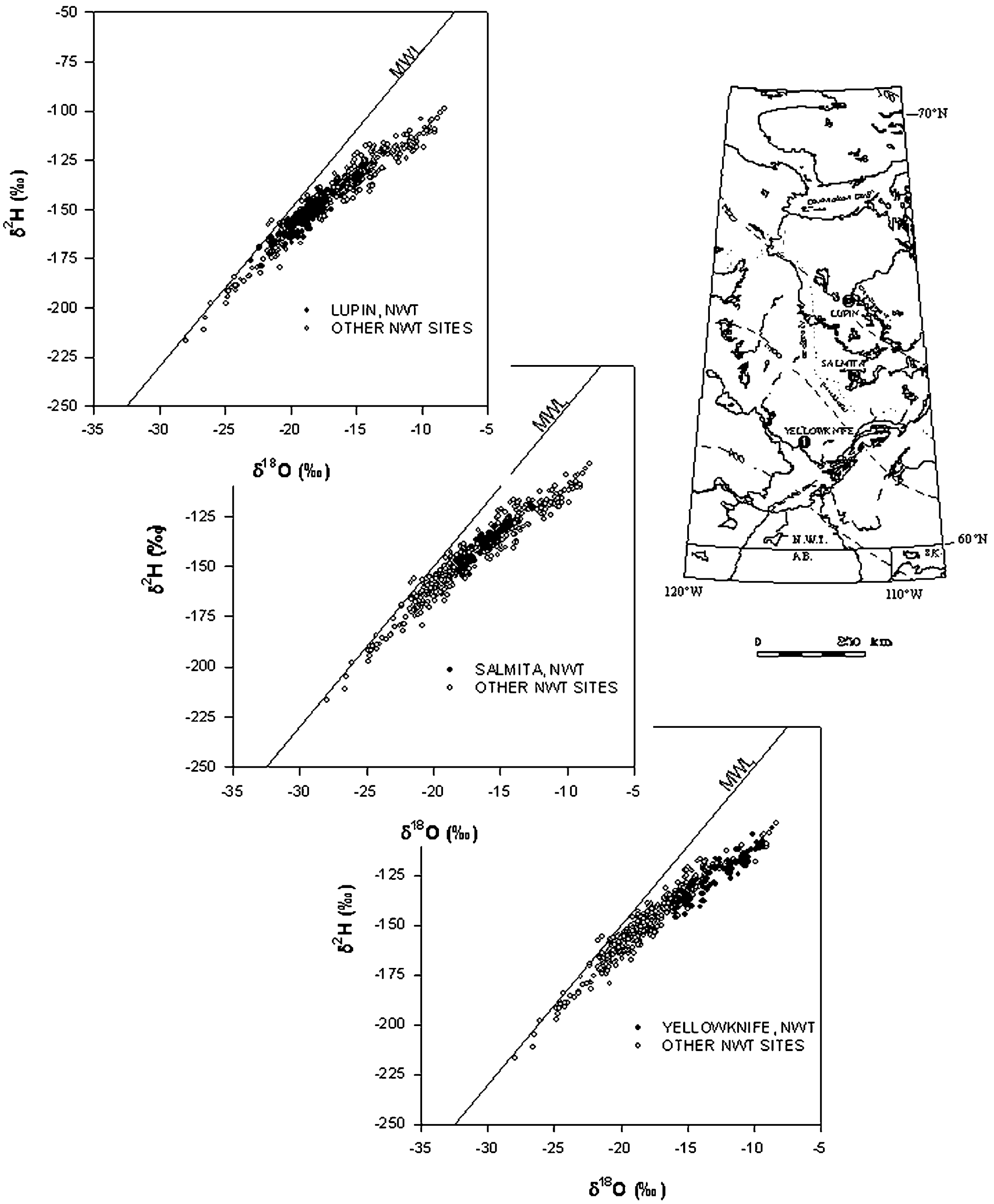

Fig. 2. Plots of δ^{18} O vs δ^{2}H for surface waters sampled at sites across northern Canada. Samples from Lupin (top), Salmita (centre), and Yellowknife (bottom) are highlighted (solid circles) against samples collected at all NWT stations (hollow circles). The plots illustrate the distinct isotopic enrichment signatures at each site reflecting differences in water balance regime. Note that time-series water samples were collected from numerous reservoirs at each site during the thaw season. Map contours are mean annual small lake evaporation (mm/a) based on limited class-A pan and climate data (den Hartog and Ferguson, 1978).

compartments to the lake may also be an important consideration with application of Eq. (5) to watershed reservoirs. However, this effect can be viewed simply as a 'resistance to mixing' of the reservoir that may be characterized from an understanding of local runoff processes. In permafrost environments, where water residence times in shallow lakes, wetlands and the active layer are relatively short, watershed compartments are generally flushed every one or two years. In comparison, large lakes can be identified with residence times of a decade or more. As such, time-lag of water delivery from the catchment may not be a substantial concern as watershed 'mixing' occurs on time-scales an order of magnitude shorter than the residence time of water in such lakes. In systems with deep groundwater contributions or widespread degradation of permafrost, or for river drainage systems without the regulating and integrating influence of lakes, this approach may be more problematic.

Incomplete mixing within the lake itself is also a potential source of error when applying Eq. (5) to large lakes. Although reservoirs typically turnover annually in permafrost environments this is not always the case (Hathersly et al., 1970) and such lakes may be laterally inhomogeneous. In principle, incomplete mixing can be characterized by sampling over time and space to bracket potential errors to any desired level of precision.

Such determinations may also be sensitive to possible shifts in the mean isotopic composition of precipitation over the residence time of lake water if isotopic records or suitable proxy data (groundwater and permafrost) are unavailable over equivalent periods.

In summary, characterization of isotopic separation between large lakes and precipitation input *via* Eq. (5) provides a practical basis for examining regional water balance trends as shown below for areas both north and south of the treeline. The quantitative analysis of such trends, while improving the understanding of regional water balance variations in a remote region of northern Canada, is based on available data and may be improved through further research.

1.2. Methodology

A field program has been underway since 1989 in the Northwest Territories, Canada to improve the understanding of hydrologic processes in cold regions through characterization and modelling of isotopic distribution between water balance components. These activities have included evaporation studies using standard water balance, energy balance, and aerodynamic profile methods (Gibson et al., 1996a), comparisons between standard evaporation methods and transient isotope mass balance methods over weekly time steps for small lakes (Gibson et al., 1996b, 1998), development and testing of new evaporation pan approaches for characterizing temporal variations in δ_A (Gibson et al., 1999), use of steady-state models for evaluating evaporation losses from lakes and wetlands (Gibson et al., 1993a), and runoff generation (Gibson et al., 1993b). A total of over 2000 water samples have been analysed for $\delta^{18}O$ and $\delta^{2}H$ including atmospheric moisture, precipitation, shallow groundwater, permafrost, lakes, wetlands, and tailings ponds. These activities have provided a broad basis for the current analysis of regional trends in water balance regime.

At each site, time-series sampling was conducted at a variety of local lakes to characterize isotopic enrichment at various levels of the drainage hierarchy. Lakes ranged from small, shallow reservoirs of less than 10 ha, averaging less than 1 m-depth, to large lakes on the order of 10–500 km^2 and 10–50 m-depth. Details on sampling protocols, instrumentation, and other field-based activities have been adequately discussed in preceding studies as outlined above.

1.3. Study sites

This study focuses on a sub-set of three sites located along a 400 km transect across northern treeline in the continental Arctic and Subarctic of northern Canada (see map, Fig. 2). The continental Arctic is distinguished from coastal arctic regions by more extreme annual variations in temperature and humidity, and by increased solar radiation input due to lower latitude, reduced cloudiness and absence of sea fog. Mean annual temperature ranges from −5°C at Yellowknife to −12°C at Lupin, with a range in monthly temperature at all sites of about 45°C (Environment Canada, 1994). Monthly temperatures above 0°C are generally observed from June to mid-September, although length of the thaw season decreases rapidly from south to north. While permafrost is discontinuous in

Table 1
Site location, climatic and hydrologic characteristics

Site	Latitude/longitude	Mean no. days lakes ice-free[a]	Climate normals[b] (for months mean T above °C)			Field observation period 1 July–31 August 1993		
			Precip. (mm/a)	Surface air T (°C)	h	h^{b}	$h_{norm}{}^{c}$	$h_{measured}{}^{d}$
Lupin	65°45′N/111°15′W	90–110	287/251	6.8/5.9	0.75/0.81	0.72/discontinued	0.72	0.74
Salmita	64°03′N/111°11′W	100–110	194	8.8	–	–	–	0.77
Yellowknife	62°30′N/114°24′W	130–140	268	12.6	0.65	0.62	0.70	0.78

[a] Estimated based on field observations during 1992–1994; see also Reid (1994).
[b] Data from Lupin A/Contwoyto L., Tundra, and Yellowknife A (June–September) (Environment Canada, 1994).
[c] Humidity normalized to lake surface water temperature at field micrometeorological station.
[d] Humidity measured within 0.5 m above lake water surface at field micrometeorological station.

Table 2
Measured and approximated model input parameters, calculations, output, and validation by site (1. δ^{18}O, 2. δ^{2}H)

Site/reservoir (basin area)[a]		Input					Calculated		Output		Validation	
		δ_I	δ_A	Mean $\delta_{L-\min}^{+\max}$ (no. samples)	T	h	m	δ^*	E/I (%)		Slope LEL	
									Oxygen	Hydrogen	Observed	Predicted[b]
Lupin/Contowoyto L.	1.	−22.1	−29.4	$-19.8_{-0.2}^{+0.2}$ (10)	6.4	0.74	2.75	−10.0	9_{-1}^{+1}		5.4 ($r^2$0.88; $n = 302$)	6.2
(15,000 km^2)	2.	−167	−226	-158_{-6}^{+4} (10)	6.4	0.74	2.43	−104		7_{-5}^{+3}		
Salmita/Matthews L.	1.	−21.4[c]	−27.8[c]	$-17.7_{-0.7}^{+0.4}$ (4)	8.8	0.77	3.24	−10.1	15_{-4}^{+2}		5.8 ($r^2$0.88; $n = 97$)	5.7
(∼300 km^2)	2.	−161[c]	−220[c]	-151_{-4}^{+5} (4)	8.8	0.77	2.89	−108		-8_{-4}^{+5}		
Yellowknife/Prosperous L.	1.	−20.7[d]	−26.2	$-15.3_{-0.1}^{+0.2}$ (7)	12.6	0.78	3.44	−9.3	-26_{-1}^{+2}		4.8 ($r^2$0.87; $n = 104$)	5.1
(16,300 km^2)	2.	−156[d]	−214	-136_{-4}^{+3} (7)	12.6	0.78	3.09	-108		-23_{-7}^{+7}		

[a] Reservoir depths are estimated at 45, 15 and 25 m, respectively.
[b] See Gat and Bowser (1991), their Eq. (7).
[c] For Salmita, these values were interpolated from measurements at the other two sites.
[d] See GNIP (1999).

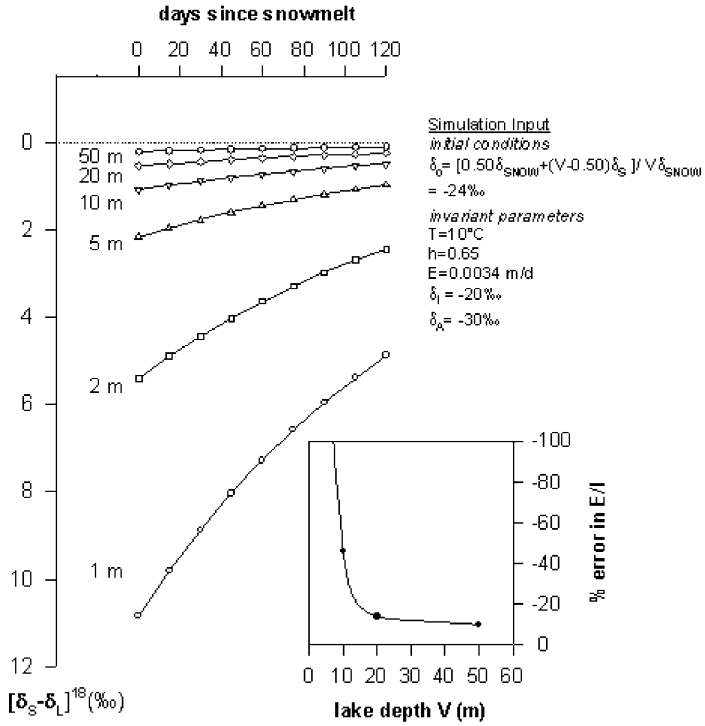

Fig. 3. Time-series plot of deviation from steady-state ($\delta_S - \delta_L$) for reservoirs of varying depth following a 0.5 m-depth snowmelt event based on Eq. (4) assuming parameters as indicated. Inset shows the relationship between lake depth and error in *E*/*I* (*x*) arising from use of mean $\delta_L$and the steady-state model [Eq. (5)].

the Yellowknife area, it is known from observations in the mine workings at Salmita and Lupin to be continuous and several hundred metres thick. Active-layer depths generally range from about 1 to 2 m-depth where permafrost is present.

Lakes are a dominant feature of the landscape in the region despite low annual precipitation of 200–300 mm (Environment Canada, 1996). Winter accumulated snow, which accounts for about 50% of annual precipitation, is rapidly released during a 1- to 2-week snowmelt period, and is the dominant annual hydrologic event both in terms of peak discharge and total volume of runoff (Marsh, 1990). During the ensuing thaw season, moisture availability near the ground surface, and hence evaporation, is enhanced by shallow permafrost, and by low relief and disorganized drainage typical of Precambrian Shield terrain. Annual small lake evaporation predicted from limited class A pan and climate data ranges from about 350 mm near Yellowknife to 200 mm near Lupin (den Hartog and Ferguson, 1978). Fifteen to 70% higher rates were obtained from field studies using multiple methods (Gibson et al., 1996a,b, 1998).

Ice-free days, annual precipitation, surface air temperature, and relative humidity during the thaw season (June–September) are given in Table 1. Comparisons between ambient relative humidity measured at climate stations, and normalized and experimentally measured values within 0.5 m above detailed study lakes, suggest very small humidity gradients across the region despite obvious differences in vegetation characteristics.

2. Results and discussion

2.1. $\delta^{18}O$ vs δ^2H relationships

Plots of δ^{18}O vs δ^2H (Fig.2) reveal distinct ranges of evaporative isotopic enrichment in surface waters at each site. Overall, lakes and other surface waters plot below the Meteoric Water Line (MWL) (Craig, 1961) along well-defined Local Evaporation Lines (LELs) with slopes ranging from 4.8 to 5.4 (see Table 2). By comparison, similar LEL trends were observed at seven additional study sites distributed across the Northwest Territories (NWT) over more than 10° of latitude and 37° degrees of longitude, in environments ranging from boreal forest to arctic desert. The LEL for all NWT surface waters collected during 1989–1995 is given by $\delta^2H = 5.4\delta^{18}O - 55.1$ ($r^2 = 0.94$; $n = 727$). Consistent slopes of the LEL at each site likely reflect consistent and robust relationships between controlling parameters such as δ_A, δ_I, h, C_K^{18} and C_K^2 (see Gat and Bowser, 1991).

2.2. Volumetric controls on isotopic variability

Mean offset along the LEL from the MWL due to evaporative enrichment clearly decreases from south-to-north from Yellowknife to Lupin (Fig. 2). Distinct ranges of enrichment variability are also evident, with extremes defined by seasonal cycles of snowmelt input and transient evaporative enrichment in the shallowest lakes (Fig. 2). Such seasonal oscillations systematically diminish with increasing lake volume as illustrated in a typical thaw season scenario preceded by a 0.5 m-depth input of snowmelt water (Fig. 3). As depicted, amplitude of isotopic variability in a small lake averaging 1 m-depth may be on the order of 7‰ in $\delta^{18}O$ compared to less than 1‰ for a large lake averaging 10–50 m-depth.

Fig. 3 also confirms that application of steady-state models as given by Eq. (5) are only realistic approximations in the present setting for deeper lakes. Fig. 3 (inset) shows error in *E/I* (x) arising from use of average δ_L values to characterize δ_S. Although errors will vary in relation to the magnitude and isotopic composition of snowmelt input, the scenario demonstrates a basic principle that errors are minimized when steady-state models are applied to large lakes with diminished seasonal isotopic variability. In the current scenario which reflects typical conditions in the study area, deep lakes can be considered to be greater than about 15 m-deep, using a threshold uncertainty of 1‰ in $\delta^{18}O$ or 20% in *E/I* in the steady state calculation. Potential errors related to uncertainty in estimates of h and δ_A, and minor seasonal volumetric changes have been discussed elsewhere (Gibson et al., 1993b).

2.3. Application of steady-state model

Individual lakes were selected based on their isotopic stability in time from the available data at each site. Although logistical difficulties did not permit rigorous time-series sampling or volumetric estimates to be determined, the lakes with the most constant isotopic composition both seasonally and interannually were also the deepest, as predicted from the concepts presented in the preceding scenario (Fig. 3). These reservoirs also represented the lowest level of the drainage hierarchy at each site and in effect, are probably the best indicators of regional long-term evaporative enrichment in each watershed. As shown in Table 2, the amplitude of variability in the reservoirs was less than 0.2‰ in $\delta^{18}O$ and 6‰ in δ^2H for Lupin and Yellowknife, where catchment area is on the order of 15,000 km^2, and 0.7‰ in $\delta^{18}O$ and 5‰ in δ^2H for Salmita, where catchment area was approximately 300 km^2.

Eq. (5) was applied to estimate the fraction of water lost by evaporation over each watershed reservoir. Input parameters, which were in part based on detailed field studies by Gibson et al. (1999), and results are given in Table 2. Importantly, δ_I at Lupin and Salmita was estimated from multi-year sampling of groundwater, precipitation, and permafrost (Gibson, 1996) and δ_I at Yellowknife was estimated from GNIP (1999) data. Values of ϵ^* were calculated based on surface air temperature during the thaw season using Majoube's (1971) formula and ϵ_K was applied assuming turbulent transfer through the boundary layer (Gonfiantini, 1986). As shown in Table 2, similar results obtained for both oxygen and hydrogen and good agreement between observed and predicted slopes of LELs suggests that basic controls on enrichment are well-described by the model. Slight improvement in predictability of the LEL slope is obtained through use of non-normalized humidities as suggested from previous studies in Canada (Allison et al., 1979), although detailed experimental analysis at the sites does not support this adjustment. Moreover, slight adjustments to the isotopic composition of atmospheric moisture and

Table 3
Comparisons with available runoff and precipitation data

Site/area	Hydrometric/climatic data		This study	
	Runoff (mm)	1. Runoff/precipitation[a]	2. Q/I $(1-x)$	Residual column 1–2
Lupin	212[b]	0.78	0.92	0.13
Salmita			0.88	
Yellowknife	81[c]	0.30	0.75	0.45

[a] Soulis et al. (1994) p. 203, Back R. and Rivers to the Arctic.
[b] Environment Canada (1996) Yellowknife R. at outlet from Prosperous L.
[c] Environment Canada (1994).

precipitation produce analogous effects. Nevertheless, the estimates, as presented, represent a preliminary basis for quantitative interpretation of the observed patterns of isotopic enrichment.

2.4. Water balance trends

Derived watershed *E*/*I* estimates for Lupin, Salmita and Yellowknife of 8, 12 and 25% based on average results from oxygen and hydrogen, suggest a steep, positive north-to-south evaporation gradient in the region which is broadly consistent with trends predicted from class A pan and climate-based estimates of small lake evaporation (den Hartog and Ferguson, 1978). Unlike pan or climate-based estimates, catchment-weighted signals of this kind explicitly relate the importance of evaporation loss to the overall hydrologic regime, and can in principle be compared more directly to hydrometric data. As evaporation losses are evaluated independently of transpiration, the method also provides a unique perspective from which to evaluate the role of open-water evaporation in the regional hydrologic regime. In this regard, the results indicate that open-water evaporation is a significant sink for precipitation input across the region despite the fact that reservoirs are typically frozen for 225–275 days per year. Using available precipitation data to characterize *I*, the results yield estimates of catchment-weighted evaporation rates ranging from about 22 mm/a near Lupin to 70 mm/a near Yellowknife. Similar rates were obtained based on isotopic sampling at the lowest level of the drainage hierarchy for a tundra site near Baker Lake NWT (64°41′N; 97°03′W) and a boreal forest site near Fort Simpson NWT (61°52′N; 121°35′W), respectively (Gibson et al., 1993b). In addition, these studies also attempted to apply the steady-state method to small shallow lakes based on reconnaissance-level isotopic sampling. However, in view of temporal variability in isotopic composition, such models are now known to be an inexact representation of such reservoirs as described previously.

Several practical difficulties become evident when the isotope-mass balance results are compared with available estimates of runoff/precipitation predicted from hydrometric and climatic data (Table 3). As isotopic fractionation traces only evaporative losses, the estimated ratio of non-fractionating outflow to input (Q/I) includes both liquid outflow (runoff) and transpiration loss, which prohibits direct validation of the results. As shown in Table 3 (residual), the difference between runoff/precipitation and Q/I may suggest that transpiration losses (which is assumed to be the dominant non-fractionating vapour flux) vary from 13% in the Lupin area to 45% near Yellowknife. Although this is reasonable considering differences expected in tundra vs forested systems, the residual also incorporates errors in measurement of snowfall, which are typically on the order of 50% in northern areas due to high frequency of trace events and blowing snow (Metcalfe et al., 1994), and errors of 30% in estimation of runoff due to the effects of river ice and extreme variations in annual discharge (Wedel, 1990).

In view of the fact that *E*/*I* is not sensitive to errors in estimating the absolute quantities of precipitation and runoff, the method does offer considerable promise for under-monitored regions provided basic atmospheric conditions such as humidity and thaw season temperature are known or can be estimated.

3. Summary

This research has provided insight into isotopic distribution in the water cycle in northern permafrost environments and its use in deriving water balance information. In particular, the study has improved the understanding of volumetric and residence time controls on isotopic signatures in lakes and their dependency on water balance and atmospheric conditions. As demonstrated, steady-state isotope-mass balance models are particularly useful for examining large-scale patterns in regions with limited hydroclimate data. One fundamental conclusion of the overall research program is that transient isotopic enrichment in small lakes can be a reliable indicator of short-term 'point' estimates of evaporation as validated using standard evaporation techniques (Gibson et al., 1996b). Such techniques can be practically applied in water balance studies of shallow lakes or tailings ponds without reliance on extensive instrumentation or climate-data extrapolated over large distances.

The research has also contributed to a better understanding of the role of surface reservoirs in the regional runoff in subarctic and arctic environments. In addition to their regulatory role in runoff, lakes and wetlands serve to enhance the residence time of runoff and thereby promote evaporation losses. Despite ice-cover during 60–75% of the year, evaporation losses evidently account for about one-tenth and one-quarter of total watershed water losses from tundra and forested regions, respectively in the continental Northwest Territories, Canada.

Although problems are encountered in validating or comparing quantitative estimates at the watershed-scale, isotope mass balance methods are certainly a valuable tool for data-sparse regions considering that errors in estimation of rainfall and runoff in permafrost regions are large and extrapolated to regions with little or no monitoring.

This study also has important implications for paleohydrological studies which rely on the $\delta^{18}O$ signature of aquatic cellulose incorporated within lake sediments and precipitation proxy data to characterize past changes in the term $(\delta_L - \delta_P)$ from Eq. (5). It is evident that deep lakes with relatively stable isotopic compositions on decadal scales should be selected to examine the significance of water balance and climate changes on the scale of millennia. For quantitative analysis, such records are also more closely approximated by steady-state models.

4. Future studies

Large-scale, multi-disciplinary hydroclimate research has emerged as an important objective of world-wide initiatives such as GEWEX (Global Energy and Water Cycle Experiment) including the cold-regions component MAGS (MAckenzie Gewex Study) in Canada and BALTEX (Baltic Sea Experiment) in northern Europe. In view of limitations in conventional monitoring networks in Canada, particularly in northern regions, use of 'non-conventional' methods is being fully explored within MAGS. These initiatives have supported an enhanced research program in the continental arctic using isotopic methods to examine water balance variability across a 250,000 km^2 area based on isotopic data collected from several hundred lakes. The activities will extend the application of this research to yield a continuous regional perspective of water balance variability from Great Slave Lake to the Arctic Ocean. Ongoing studies within MAGS will emphasize application of isotopic techniques in parallel with other hydrologic methods, as required to gain a complimentary understanding of hydrological processes at the large scale. Isotopic tracing of source-areas, flow components, and catchment-weighted evaporation losses is therefore being undertaken at selected nodes within the Mackenzie Basin to aid in meso- to macro-scale hydrograph modelling where hydrologic regime response is poorly described or unknown. Analogous activities are being considered by an international group of scientists in support of the BALTEX project.

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