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Quantitative comparison of lake throughflow, residency, and catchment runoff using stable isotopes: modelling and results from a regional survey of Boreal lakes

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Abstract

An isotope-based approach for water balance assessment is presented and applied to estimate throughflow, residence time and catchment runoff to 70 headwater lakes on the Boreal plain and uplands of northern and north-central Alberta, Canada. The survey reveals a complex hydrologic regime with systematic variability in water balance due to local site characteristics. On average, runoff to lakes in wetland-dominated catchments is found to be significantly higher than runoff to upland-dominated lakes, with generally higher contributions from catchments with low bog/fen ratios. The isotope method, which relies primarily on water sampling and isotopic analysis, can be easily integrated in routine water quality surveys and is shown to be a practical alternative to conventional hydrological modelling for comparative analysis of water balance controls on hydrochemistry and aquatic ecology of lakes, particularly in low-relief wetland-rich terrain.

Keywords: Stable isotopes; Oxygen-18; Deuterium; Lakes; Regional water balance; Spatial variability; Disturbance hydrology; Regional limnology

1. Introduction

Hydrologic modelling is commonly employed to empirically or statistically evaluate potential water balance and landscape controls on lake water quality (Thierfelder, 1999). The rationale for this approach is that substances are being carried to the lake via surface and subsurface runoff from the catchment slopes, and therefore, may inherit distinct hydrochemical signatures associated with differing pathways, transport mechanisms, or variations in landscape element distributions. Although this approach has been successfully applied to assess water quality impacts from disturbance in many areas including the Boreal Shield of Quebec (Carignan et al., 2000), difficulties have been encountered in basic application and validation of such models in low-relief wetland-rich terrain (Pietroniro et al., 1996). The primary limitation in this setting has been that landscape hydrological models are topographically driven, and low-relief can result in problems with defining catchment areas, and poor representation of internal drainage structure, especially closed depressions (Martz and Garbrecht, 1997; Garbrecht and Martz, 1999). Similar problems

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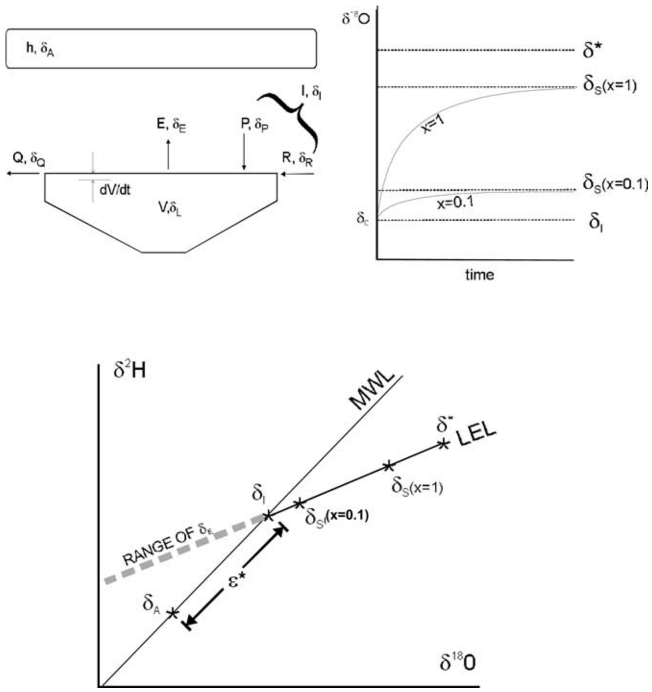

Fig. 1. (a) Generalised lake balance schematic showing storage and hydrologic fluxes and their isotopic compositions for a typical lake. Note that I is total inflow, where $I = R + P$, P is precipitation and R is surface and subsurface runoff; E is evaporation, Q is the isotopic composition of outflow, V is lake volume, dV/dt is change in lake storage, and h is ambient atmospheric humidity. Note that δ values refer to the isotopic compositions of the respective components. δ_A and δ_L denote the isotopic compositions of ambient atmospheric moisture and lake water, respectively. (b) Time series plot of δ^{18}O enrichment for lakes with differing values of x (evaporation/inflow). Note that δ^* is the limiting isotopic enrichment under local atmospheric conditions, δ_S ($x = 1$) is the steady-state isotopic composition for a terminal lake, δ_S ($x = 0.1$) is the steady-state isotopic composition for a throughflow lake with $x = 0.1$, δ_I is the isotopic composition of inflow and δ_0 is the initial isotopic composition of the lakes. (c) Plot of δ^{18}O versus δ^2H depicting isotopic compositions and typical isotopic separation between various components depicted in Fig. 1(a) and (b). Note the position of the MWL of Craig (1961) and LEL. This scenario depicts the case of isotopic equilibrium between precipitation and atmospheric moisture, i.e. $\delta_A = \delta_P - \varepsilon^*$, ε^* being the equilibrium isotopic separation for oxygen and hydrogen.

are also encountered due to lack of high-resolution topographic data for small-scale applications in remote or developing regions (e.g. parts of northern Canada).

Herein, we present an alternative approach for estimating lake water balance and landscape runoff contributions in headwater systems based mainly on water sampling and laboratory analysis of the oxygen-18 and deuterium composition of the lake itself, and utilising supplementary physical and climatic data. The rationale for this approach is that the degree of natural evaporative isotopic enrichment in lakes is a sensitive indicator of lake water balance parameters, specifically the partitioning of water losses by evaporation versus liquid outflow (Gat, 1995; Gibson et al., 1993; Gibson, 2001), and can be used to trace lake-to-lake or temporal changes in lake balance, including runoff from the catchment. The current study was motivated by the need for hydrological control in an investigation of forest fire and harvesting impacts on aquatic chemistry and ecology of headwater lakes in low-relief catchments of northern Alberta. The required field data, which includes time-series collection of water samples during the thaw season and bathymetric surveys, can easily obtained during routine water quality sampling surveys. Supplementary meteorological information and precipitation isotope data can be obtained in most areas by interpolation from existing climate station databases.

The isotope mass balance approach for estimating water balance parameters has been demonstrated in previous studies of open water bodies (Dinçer, 1968; Gat, 1970, 1981; Zuber, 1983; Krabbenhoft et al., 1990) and was recently reviewed by Gat (1995). In addition, a series of recent isotope-based studies on water budget assessment, spatial water balance variability, and regional hydrology provide important background information on isotopic variations in seasonal climates of northern Canada (Gibson et al., 1993, 1996a,b, 1999; Gibson, 2001). This is the first analysis of water balance variability and controls inferred from isotope-based analysis of a large number of lakes. Results are presented along with a brief summary of observed morphologic controls on lake water balance. A comprehensive review of water balance controls on hydrochemistry and aquatic biota, including the inferred effects of natural forest fire and clear-cut harvesting are presented elsewhere (Prepas et al., 2001; McEachern et al., 2000). The theoretical basis of the approach, models and practical sampling and parametric weighting issues are discussed later.

1.1. Theory

The water-mass and isotope-mass balance for a well-mixed lake (Fig. 1), assuming constant density of water, may be written, respectively, as

$$dV/dt = I - Q - E \tag{1}$$

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

where V is the volume of the reservoir, t is time, dV is the change in volume over time interval dt, I is combined surface and subsurface inflow, Q is combined surface and subsurface 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. Due to differences in the saturation vapour pressure and molecular diffusivities in air of the rare, heavy isotopic species of water ($^1H_2{}^{18}O$,$^1H^2H^{16}O$) relative to the common, light species ($^1H_2{}^{16}O$), the evaporation flux δ_E is typically depleted in the heavy isotopes relative to lake water δ_L (Gat, 1995). The magnitude of isotopic separation between lake water and the isotopic composition of the evaporation flux has been shown to be dependent on evaporation temperature, the details of boundary layer mass transfer processes (i.e. laminar, turbulent or static), and ambient atmospheric conditions (humidity and isotopic composition of atmospheric moisture). To account for these effects, a boundary-layer flux model developed by Craig and Gordon (1965) can be applied to estimate the isotopic composition of the evaporation flux (neglecting resistance to mixing in the liquid phase[1]) as a function of other more readily determined parameters. The evaporating moisture is given by

$$\delta_E = (\alpha^*\delta_L - h\delta_A - \varepsilon)/(1 - h + 10^{-3}\varepsilon_K) \tag{3}$$

where α^* is the equilibrium liquid–vapour isotope fractionation, h is the atmospheric relative humidity (expressed as a decimal fraction ranging from 0 to 1, and normalised to the saturation vapour pressure–temperature of the air–water interface), δ_A is the isotopic composition of ambient moisture, and

$$\varepsilon = \varepsilon^* + \varepsilon_K \tag{4}$$

where ε is the total isotopic separation factor comprised of both equilibrium ε^* and kinetic ε_K components. Equilibrium separation factors ε^* for oxygen and hydrogen are adequately known as a function of temperature from laboratory experiments by Majoube (1971) and others as summarised in Gonfiantini (1986). Likewise, kinetic separation factors ε_K are narrowly constrained both from theoretical and experimental studies. For water balance applications, ε_K can be approximated by

$$\varepsilon_K = C_K(1 - h) \tag{5}$$

where $C_K = (D/D_i)^n - 1$ and D is the molecular diffusion coefficient of $^1H_2{}^{16}O$, D_i is the molecular diffusion coefficient of $^1H_2{}^{18}O$ or $^1H^2H^{16}O$, 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; Merlivat, 1978a,b; Merlivat and Coantic, 1975; Brutsaert, 1975). Use of $n = 1/2$, which corresponds to C_K values of approximately 14.3 and 12.5‰ for oxygen and hydrogen, respectively, is a suitable approximation for natural evaporation from lakes (Gonfiantini,

[1] Isotopic gradients in the liquid phase may occur near the air–water interface. This effect is considered negligible on water balance time scales of days to years (Gat, 1995).

1986). Additional corrections are sometimes required to account for build-up of evaporate in the overlying boundary layer, although this effect is not generally significant for small area lakes (Gat, 1995).

1.2. Steady-state models

The most common application of isotope tracers has been in the context of simple balance models assuming invariant hydrologic and climatic conditions, although models of more complex systems have also been developed and applied (Gonfiantini, 1986). Conceptually, there are three main hydrologic settings for lakes (Horita, 1990):

(i) dessicating water bodies, where inflow takes place once or sporadically (i.e. $dV/dt = -Q - E$),
(ii) terminal (or zero throughflow) lakes, where inflow is continuous, and where long-term evaporation balances inflow, so that no liquid outflow occurs (i.e. $I = E$), and
(iii) throughflow lakes, where inflow is continuous and balanced by a combination of outflow via evaporation and liquid outflow (i.e. $I = Q + E$).

Desiccating lakes are transient systems, and therefore, are not well represented by steady-state models, nor are such lakes common in Boreal regions. In the case of both terminal lakes and throughflow lakes, where hydrologic fluxes remain constant and volumetric changes are minor (i.e. $dV/dt \approx 0$), the lake can be considered as close to hydrologic steady state. In a scenario, where lake water begins with an initial isotopic composition close to that of input water (i.e. $\delta_0 \approx \delta_I$), it will undergo progressive enrichment in the heavy isotopes as evaporation proceeds during residency of water in the lake (Fig. 1). In terminal and throughflow situations, an integrated expression for temporal changes in lake water δ_L is obtained by combining Eqs. (2) and (3) according to Gonfiantini (1986) assuming other parameters are invariant as

$$\delta_L = \delta_S - (\delta_S - \delta_0)\exp[-(1 + mx)(It/V)] \tag{6}$$

where δ_0 is the initial isotopic composition of the reservoir,

$$\delta_S = (\delta_I + mx\delta^*)/(1 + mx) \tag{7}$$

where δ_S is the steady-state isotopic composition that the reservoir will attain over time as $t \to \infty$, and

$$\delta^* = (h\delta_A + \varepsilon)/(h - 10^{-3}\varepsilon) \tag{8}$$

where δ^* is the limiting isotopic enrichment for a reservoir evaporating to dryness under local climate conditions (Gat and Levy, 1978; Gat, 1981), and

$$m = (h - 10^{-3}\varepsilon)/(1 - h + 10^{-3}\varepsilon_K) \tag{9}$$

where m is the enrichment slope defined in previous studies (Welhan and Fritz, 1977; Allison and Leaney, 1982), and

$$x = \frac{E}{I} = \frac{E}{E + Q} \tag{10}$$

where x is a throughflow index, i.e. the fraction of total water inputs lost by evaporation.

Rearranging Eq. (7) to obtain an expression for the throughflow index (x) yields

$$x = \frac{(\delta_L - \delta_I)}{m(\delta^* - \delta_L)} \tag{11}$$

In the case of the non-transient reservoirs (terminal lakes and throughflow lakes), heavy isotope enrichment in lake water will approach an isotopic steady state δ_S (see Fig. 1(b)), so that deviation from δ_I along the local evaporation line (LEL) is constrained to be less than $\delta_{S(x=1)} - \delta_I$. Enrichment beyond $\delta_{S(x=1)}$ is restricted to the desiccating reservoirs.

On plots of $\delta^{18}O$ versus δ^2H (Fig. 1(c)), the kinetic fractionation of oxygen and hydrogen isotopes by evaporation produces enrichment along LEL with slopes ranging from 4 to 7, in contrast to shifts associated with pure equilibrium isotopic exchange, which lie close to the MWL of Craig (1961) and have a slope close to 8 as fixed by the ratio of $\varepsilon_2^*/\varepsilon_{18}^*$ (Fig. 1(c)). Note that degree of displacement along the LEL is proportionate to the fraction of water lost by evaporation (x). In non-seasonal systems, the isotopic composition of atmospheric moisture, which fixes the potential enrichment via control on the δ^* and δ_S values is normally close to equilibrium with mean annual precipitation ($\delta_A \approx \delta_P - \varepsilon^*$).

1.2.1. Volume, residence time, and seasonality effects

While lake volume is not a primary control on the isotopic composition of lake water in systems which are close to isotopic and hydrologic steady state, the

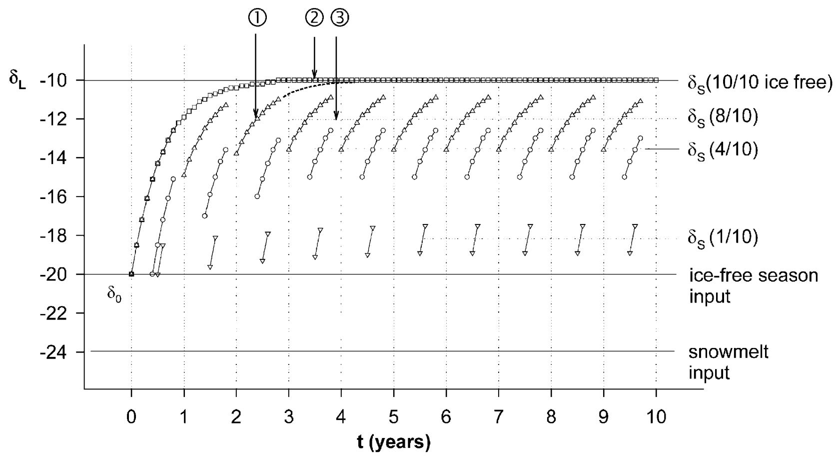

Fig. 2. Time-series of isotopic composition of lake water (δ_L) for hypothetical lakes with ice-free periods ranging from 1/10 to 10/10 year. Note that the scenario assumes initial δ_0 of −20‰, $\delta_S = -10$‰, constant steady-state values for $I = E = 1.4$ mm/d during the ice-free (evaporation) season, $V = 1000$ mm, $h = 0.7$, $T = 10$°C during the ice-free period, an instantaneous snowmelt inflow of 250 mm at the beginning of each ice-free season. The lake is assumed to remain inactive during the ice-on period (i.e. no isotopic exchange and no water inflow or outflow or evaporation). In the example, the various systems only reach pseudo-steady state (dotted horizontal lines) after 3–5 years, depending on the water balance scenario. Note that the appropriate pseudo-steady-state values are close to average δ_L during the thaw season. The symbol ① identifies the δ_L record obtained from time-series sampling during a single ice-free period for the 8/10 scenario; ② identifies the δ_S value obtained from best-fit of δ_L data from ①, which yields estimates representative of water balance conditions during the ice-free period only; and ③ identifies the δ_S value obtained by averaging ice-free period δ_L data, which yields estimates representative of mean annual water balance conditions. Note that the latter approach was employed to obtain mean annual water balance parameters in the present study.

temporal footprint of δ_L depends on the residence time of water in a lake given by

$$\tau = V/I \tag{12}$$

which can be calculated from the annual dimensionless throughflow index (x), mean lake volume (in mm) and lake evaporation (in mm) as

$$\tau = xV/E(\text{years}) \tag{13}$$

Lakes with smaller volume (or mean depth) will tend to be more affected by short-term perturbations related to flood-drought cycles or other events that may be buffered by larger lakes. In situations, where residence time is longer than 1 water year, steady-state approximations are expected to yield estimates close to annual values. In situations, where residence time is shorter than 1 water year, lakes are expected to systematically fluctuate on a seasonal basis.

The steady-state model is a straightforward and reasonable approximation for most lakes with $\mathrm{d}V/\mathrm{d}t < \sim 5\%$ in temperate climate zones although it does not account for short-term isotopic shifts, which often occur in seasonal climates. Use of steady-state approximations in seasonal climates is not necessarily problematic, although parameters in Eq. (11) must be carefully defined. Overall, the atmospheric parameters (δ_A, δ^*, h, m) should be flux-weighted according to the seasonal timing of evaporation and variations in evaporation rates. In strongly, seasonal climates lakes will not attain isotopic steady state (unchanging isotopic composition) during the ice-free season as shown in Fig. 2. Ice cover serves to isolate a water body from evaporation and atmospheric exchange during the winter and portions of the transitional months, and is therefore strongly linked to the seasonality. Duration of ice cover and seasonality also influence the flux-weighted atmospheric parameters, which may differ significantly from the mean annual values in extreme climates. Fig. 2 is a time-series plot of isotopic enrichment in a series of hypothetical lakes with different ice-free durations ranging from 1/10 to 10/10 year. In the scenario, ice cover is allowed to vary in order to simulate lakes with varying seasonality and water balance. The scenario assumes that: (i) the lakes behave like terminal lakes ($x = 1$) during the ice-free season (i.e.

$I = E = 1.4$ mm/d) and $\delta_S = -10‰$ and $\delta_I = -20‰$, (ii) the lake is 1000 mm deep and is initially filled with water of $\delta_0 = -20‰$, and (iii) at the beginning of each year, the lake water is assumed to be reset to lower δ values by instantaneous input of 250 mm of snowmelt water of $\delta_I = -24‰$. Progressive, monthly time-steps in isotopic enrichment are shown for each ice-free period as predicted by Eq. (6). Note that differences in the ice-free period also imply differences in the annual evaporation and water balance.

In seasonal settings, it is important to note that temporal isotope sampling of lakewater during the ice-free period can provide an estimate of δ_S and hence x, that is, representative of the ice-free period if a best-fit approach is applied whereas an estimate of δ_S representative of long-term (annual) conditions is obtained from the average of measured δ_L values during the ice-free period (Fig. 2). In principle, as seasonality decreases so does the difference between the mean annual and ice-free only δ_S values. Redistribution of isotopes in ice covers due to freezing effects may also complicate isotope distributions in lake water in the very early thaw season, although this effect is largely restricted to a short interval following melting of the ice cover.

1.2.2. Incomplete mixing and stratification

In stratified lakes or lakes with pronounced horizontal inhomogeneities, it can be necessary to account separately for epilimnion and hypolimnion volumes and exchanges, provided these have distinct isotopic compositions (Gat, 1995). Neglecting stratification can lead to overestimation of the importance of evaporation loss, if sampling is conducted during dry, stratified periods and underestimation of evaporation loss, if sampling is conducted during wet, stratified periods. Inflow bypass or short-circuiting of the system may also reduce the effective volume of the lake during wet periods or reduce the effective input in the opposite situation. Incomplete mixing within the lake itself is also a potential source of error, when applying Eq. (5) to large lakes. The effect of stratification is minimised in Boreal environments as reservoirs typically turnover annually. In principle, incomplete mixing can be characterised by spatial and temporal sampling to bracket potential errors to any desired level of precision, although this is not always practical. A simple approximation for systems with similar epilimnion and hypolimnion compositions is to use an average value to represent the undifferentiated lake volume. Potential errors related to stratification are discussed later.

1.2.3. Catchment effects

To maintain a relatively stable long-term water level, the inflows to a lake must roughly balance the discharges by both evaporation and liquid outflow. Although short-term signals may be hidden in the volumetric change of the reservoir, a knowledge of the long-term or annual lake water balance should provide information on the mean long-term catchment runoff. Inflow to a lake is comprised of both precipitation and lateral inflow from the catchment area. The general water and isotope balance of such inflow is given by

$$I = P + R \tag{14}$$

$$\delta_I I = \delta_P P + \delta_R R \tag{15}$$

where P is the precipitation falling on the lake surface, R is the combined surface and subsurface runoff from the catchment to the lake and δ values represent the isotopic composition of the various components. Owing to the fact that catchment runoff is derived directly or indirectly from precipitation in headwater systems, the inflow to a lake is to a first approximation given by $\delta_I \approx \delta_P \approx \delta_R$ although exposure to evaporation (not transpiration) during recharge and surface runoff may lead to alteration of the evaporative signal in groundwater, wetlands, dry soils, and other watershed compartments (Gibson, 2001). Transpiration can effectively remove substantial amounts of potential runoff, although it does not significantly alter the isotopic composition of the soil water reservoir except in extremely dry soils. A contribution from upstream lakes has also been referred to as the chain-of-lakes effect (Gat and Bowser, 1991). Catchment effects are also be minimised in the case, where the drainage basin area is small relative to the surface area of the lake, so that R is a minor component of I. A tortuous surficial drainage configuration may promote evaporative modification of water during its residency in the catchment. In addition, Eqs. (14) and (15) may not apply to lakes with deep-seated groundwater contributions derived from distant recharge outside the catchment area (i.e. from a non-local precipitation source).

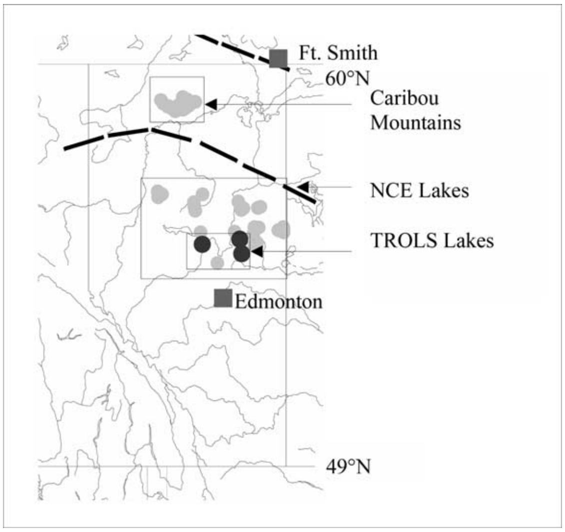

Fig. 3. Map of Alberta, Canada and vicinity showing the approximate location of lakes and lake groups included in the study, and location of nearest IAEA/WMO Global Network of Isotopes in Precipitation stations at Edmonton and Ft. Smith. A more extensive network of climate stations used to interpolate meteorological parameters is not shown.

From the assumption that the volume of runoff from the catchment is equal to the volume of lake inflow minus the precipitation to the lake, the catchment runoff to a lake of area LA can also be evaluated from the lake isotopic data using

$$R = \left(\frac{E}{x} - P\right)\frac{\mathrm{LA}}{\mathrm{DBA}} \tag{16}$$

where DBA is the drainage basin area such that (LA + DBA = CA), where CA is the total catchment area, R is the catchment runoff, P is precipitation, and DBA is the drainage basin area. This also permits estimation of the runoff/precipitation ratio (R/P) or the effective drainage basin area (eDBA) as

$$\mathrm{eDBA} \approx \frac{R}{P}\mathrm{DBA} \tag{17}$$

1.3. Study area and methods

In 1996 and 1997, water samples were collected in 70 lakes (3 times per year for most lakes) during the ice-free season, which extends from May to October. The lakes, belonging to three study groups, namely the Caribou Mountains Lakes, NCE Lakes and the Road-Accessible lakes (TROLS), are located in northern and north-central Alberta (Fig. 3). Some lakes were accessible by highways, logging roads or off-road trails, while others were only visited by fixed-wing Cessna float plane. Details on site selection have been discussed elsewhere (McEachern et al., 2000; Prepas et al., 2001). Vertically integrated samples were collected from the euphotic zone (depths receiving $\geq 1\%$ of ambient surface light) at two sites (including the deepest) along the longitudinal axis of each lake. In some of TROLS, grab samples of the epilimnion and hypolimnion were also obtained, along with supplementary sampling of groundwaters, surface waters, precipitation, snowpack, and evaporation pans. Water samples were collected in tightly sealed 30 ml high-density polyethylene (HDPE) bottles and returned to the University of Waterloo for standard analysis of $\delta^{18}O$ and $\delta^{2}H$ by mass spectrometry within 6 months of collection. δ values are reported as deviations in permil (‰) from the Vienna-standard mean ocean water (SMOW), such that $\delta_{\mathrm{sample}} = 1000((R_{\mathrm{sample}}/R_{\mathrm{smow}}) - 1)$, where R is $^{18}O/^{16}O$ or $^{2}H/^{1}H$. Values cited herein are normalised on the SMOW-SLAP scale, so that standard light arctic precipitation (SLAP) has a value of -55.5‰ in $\delta^{18}O$ and -428‰ in $\delta^{2}H$ (Coplen, 1996). Analytical uncertainty is estimated to be ± 0.1‰ for $\delta^{18}O$ and ± 2‰ for $\delta^{2}H$.

Lake volume, surface area, mean lake depth, and maximum lake depth were estimated from bathymetric maps, and drainage basin area and percent coverage by wetlands, uplands, bogs and fens were estimated using 1:20,000 or 1:15,000 aerial photographs.

The annual precipitation was estimated for each lake-site along with relevant ice-free exchange parameters (lake evaporation, weighted air temperature and humidity[2]) by spatial interpolation (kriging) from all climate stations in the Mackenzie Basin.

For the NCE and TROLS lakes, isotopic composition of input δ_I for each individual lake was estimated based spatial interpolation (kriging) of long-term amount-weighted precipitation data from 15 Canadian climate stations in the global network for isotopes in precipitation (GNIP) database (nearest neighbours are

[2] Ice-free season estimates are based on monthly data for May–September weighted according to the monthly evaporation flux.

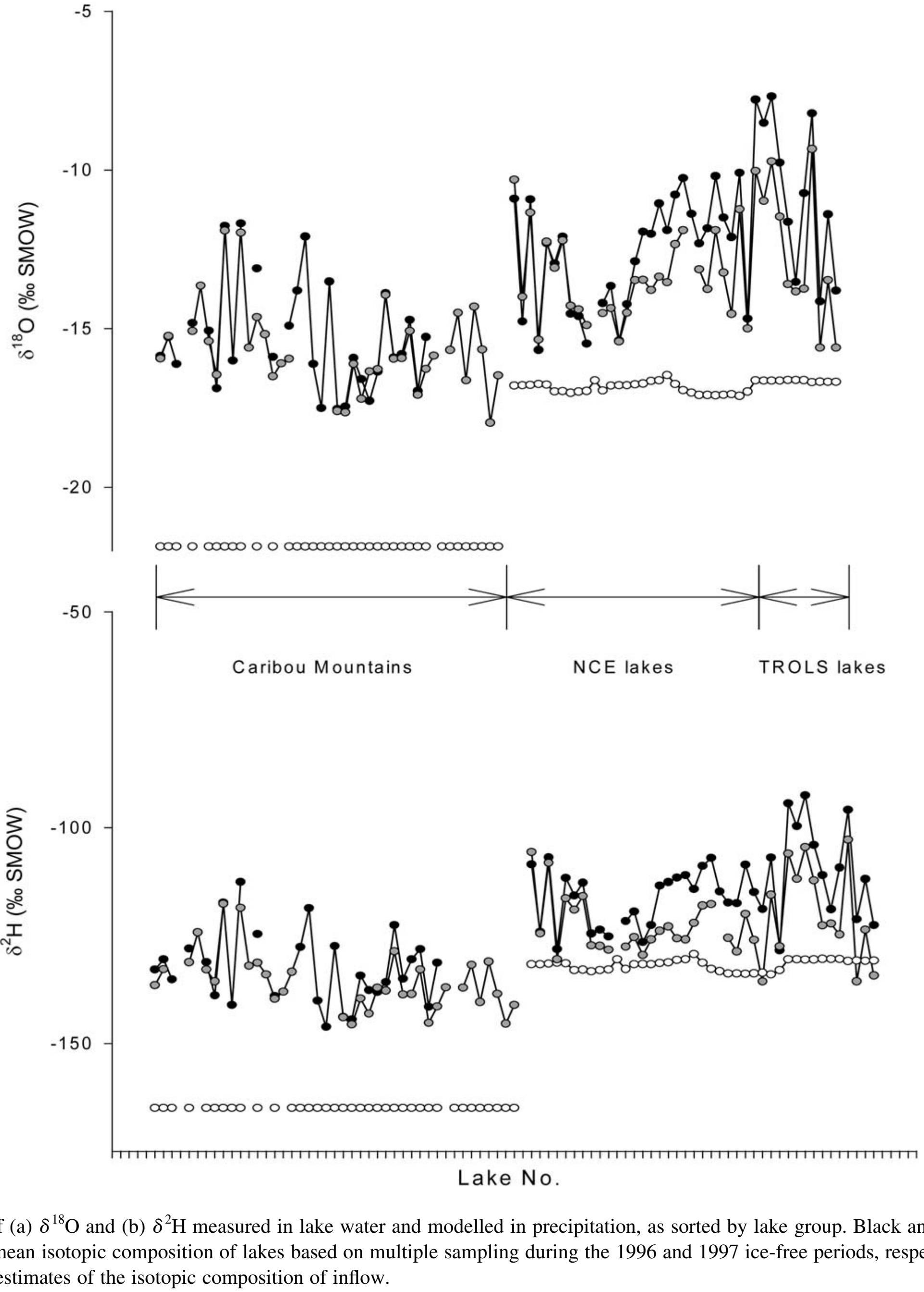

Fig. 4. Plots of (a) δ^{18}O and (b) δ^{2}H measured in lake water and modelled in precipitation, as sorted by lake group. Black and grey circles represent the mean isotopic composition of lakes based on multiple sampling during the 1996 and 1997 ice-free periods, respectively. Open circles depict estimates of the isotopic composition of inflow.

shown in Fig. 3). These values were found to be in close agreement with measured isotopic composition of precipitation and groundwater collected at selected field stations except in the Caribou Mountains. Due to the distinct upland setting of the Caribou Mountains Lakes, δ_I was estimated as the intersection of the MWL and LEL in δ^{18}O versus δ^{2}H space (Gibson et al., 1993). This method produced input estimates that were in better agreement with measured precipitation and groundwater in the area. The isotopic

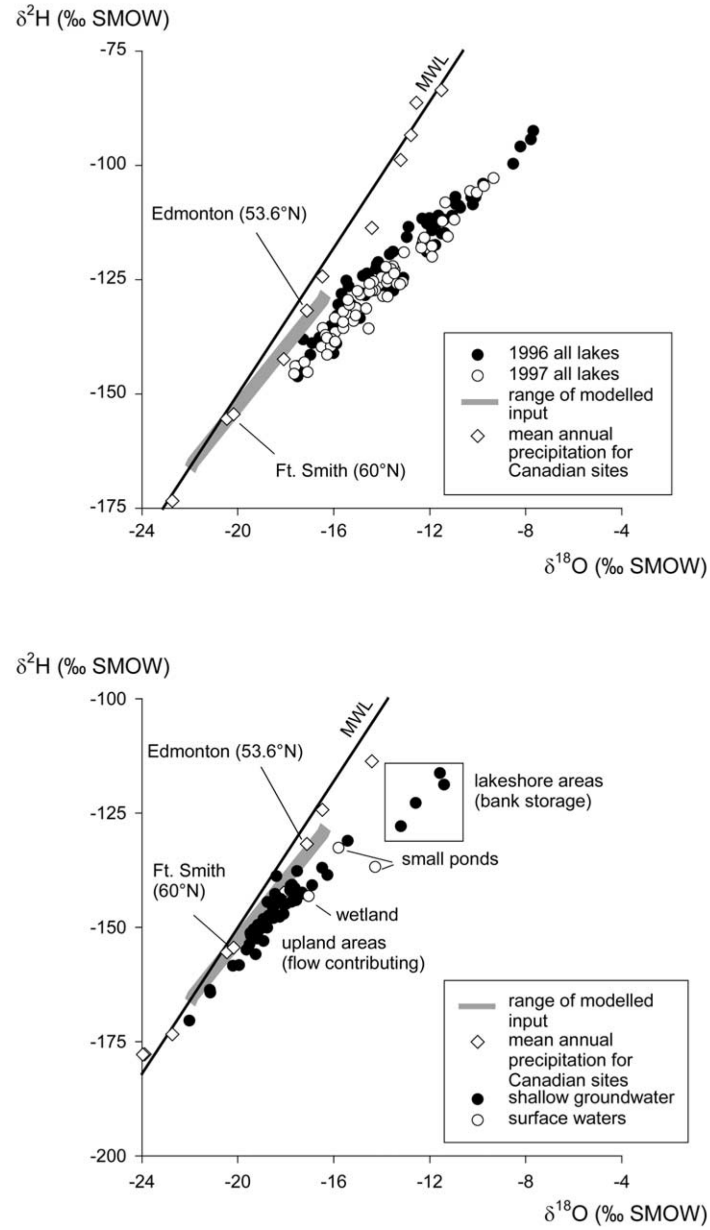

Fig. 5. (a) Plot of $\delta^{18}O$ versus $\delta^{2}H$ showing lake water, range of modelled precipitation for lakes in this study, and measured precipitation at GNIP stations in Canada. Note that MWL denotes the MWL of Craig (1961) with a slope of 8. Lake samples cluster to form a local evaporation trend (LEL) with a slope of about 5. Note that the intersection of MWL and LEL corresponds closely to the precipitation input composition expected for the study area. Enhanced evaporation loss (lower throughflow) corresponds to enhanced displacement of lake water from the MWL along the LEL; (b) Plot of $\delta^{18}O$ versus $\delta^{2}H$ showing measured and modelled input sources.

Table 1
Summary of water balance parameters by lake group calculated by (1) $\delta^{18}O$ balance and (2) δ^2H balance († and ‡ denote maximum and minimum representative values, respectively (4 extreme values removed))

Lake group		x (%)		τ (years)		R/P (%)	
		1	2	1	2	1	2
Caribou Mountains ($n = 27$)	Mean	32	42	1.8	2.6	33	24
	Max†	75	96	9.3	13.6	175	186
	Min‡	16	18	0.2	0.2	3	1
	1 std	13	18	2.1	3.2	33	35
NCE ($n = 27$)	Mean	25	25	0.6	0.6	75	75
	Max†	51	61	2.3	2.3	186	187
	Min‡	6	5	0.1	0.1	19	19
	1 std	13	13	0.5	0.5	44	44
TROLS ($n = 11$)	Mean	45	45	2.1	2.2	20	20
	Max†	81	94	7.2	8.3	77	76
	Min‡	12	10	0.2	0.1	9	5
	1 std	26	30	1.9	2.2	19	19

composition of atmospheric moisture was calculated based on the precipitation equilibrium assumption ($\delta_A = \delta_P - \varepsilon^*$), where δ_P and ε^* were weighted according to the monthly evaporation flux. It should be noted that, as expected, the interpolation approach yielded $\delta^{18}O$ estimates that were less enriched than measured lake water isotope compositions for all lakes. For three locations, however, δ^2H of input was estimated to be slightly more enriched than observed lake water, reflecting greater uncertainty with this tracer or slight problems with the kriging routine used in the present setting.

2. Results and discussion

2.1. Isotopic variations

Lakes were found to be systematically, but variably enriched in both heavy isotopes relative to measured and modelled inputs at each site (Fig. 4), similar to distributions observed elsewhere in surveys of small lakes in Canada (Gibson et al., 1993). The isotopic separation between lake water and modelled input is different for each lake, although comparable patterns are found for both tracers and similar (although not identical) patterns are obtained for 1996 and 1997. As shown in plots of $\delta^{18}O$ versus δ^2H (Fig. 5(a)), this separation is clearly the result of evaporative enrichment of lake water, which produces a LEL, the degree of offset of individual lakes from the MWL reflecting proportion of water loss by evaporation. The consistent LEL slopes for both years, and high correlation between $\delta^{18}O$ and δ^2H suggests an evaporation regime governed by robust and consistent exchange parameters.

The isotopic composition of modelled input used in the water balance calculations (Fig. 5(b)) is intermediate between observed δ values in groundwaters from upland areas and the meteoric water line (MWL), reflecting the mean global trend of $\delta^{18}O$ and δ^2H in precipitation. Also shown (Fig. 5(b)) are δ values for some lakeshore areas, which are evidently affected by subsurface outflow from the lake (in these cases likely due to bank storage), and wetlands and ponds, which are also evaporatively enriched. Although enrichment does occur in waters residing in surface storage for significant periods, the overall flux-weighted inputs (accounting for the small contributions of many of these highly evaporated sources) are not likely significantly different than predicted in precipitation. For example, in the upland catchment, where the detailed sampling was conducted, the arithmetic mean isotopic composition of groundwater ($n = 59$) is -18.1 and -146‰ for $\delta^{18}O$ and δ^2H, respectively, depleted by 0.3 and 2.5‰ compared to precipitation at the site, and close to the limit of analytical uncertainty. This effect is attributed to contributions of snowmelt,

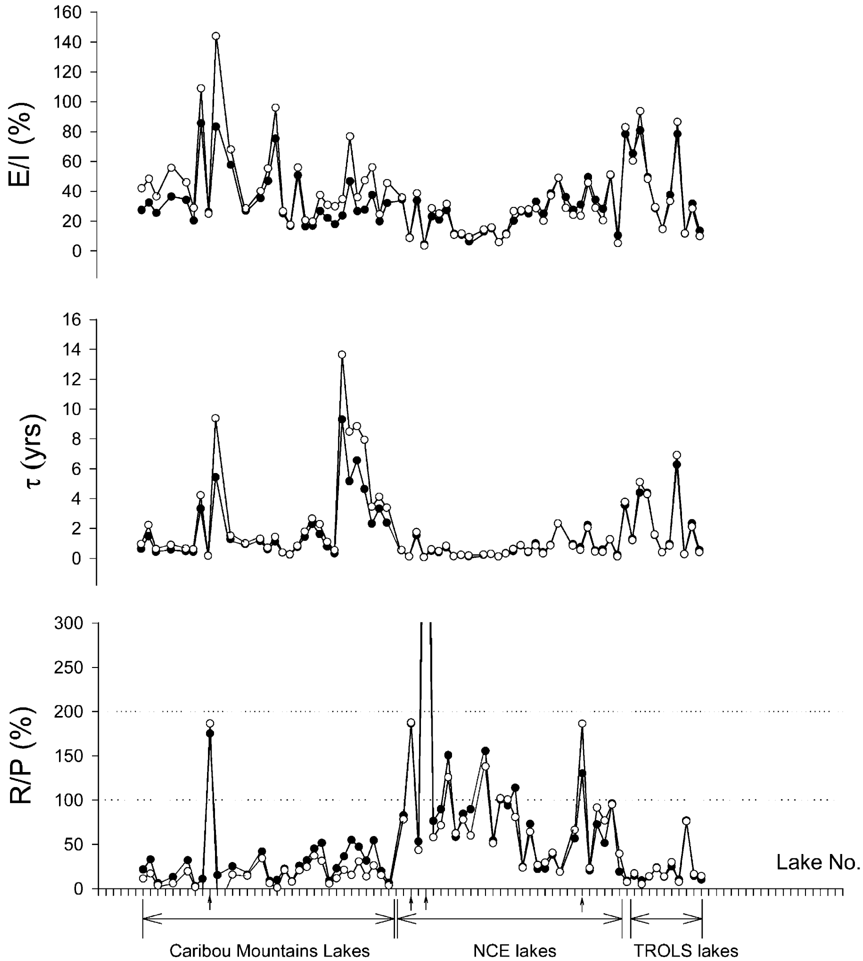

Fig. 6. Isotope-based estimates of throughflow index (x or E/I), residence time (τ), and runoff/precipitation ratios (R/P) for lakes sampled in 1996, sorted by lake group. Note that good agreement is obtained between estimates using oxygen-18 ($\delta^{18}O$) and deuterium ($\delta^{2}H$). E/I in excess of 100% may suggest model calibration problems or effect of volumetric drawdown. R/P values are found to be higher than the expected maximum of about 50% (see text for explanation). Small vertical arrows indicate lakes excluded from the summary in Table 1.

which tends to be depleted in the heavy isotopic species, rather than by evaporative enrichment. Based on the available isotopic information, the evaporative enrichment during runoff to the lakes is expected to be minor in upland-dominated lakes. In the case of wetland-dominated lakes, there is certainly more potential for some interference of the isotopic signal by evaporation from wetlands, which would lead to overestimation of x and the residence time, and underestimation of the catchment runoff. However, wetland catchments (>50% wetland cover) generally had higher average depth-equivalent runoff than non-wetland catchments suggesting that this effect is either small or it is overwhelmed by problems with underestimation of catchment area in wetland areas.

2.2. Quantitative estimates

The throughflow index (x or E/I), was estimated based on Eq. (11) and interpolated humidity, temperature and isotope exchange parameters for each site. Note that isotope values for each lake were averages obtained from three samples taken during July to

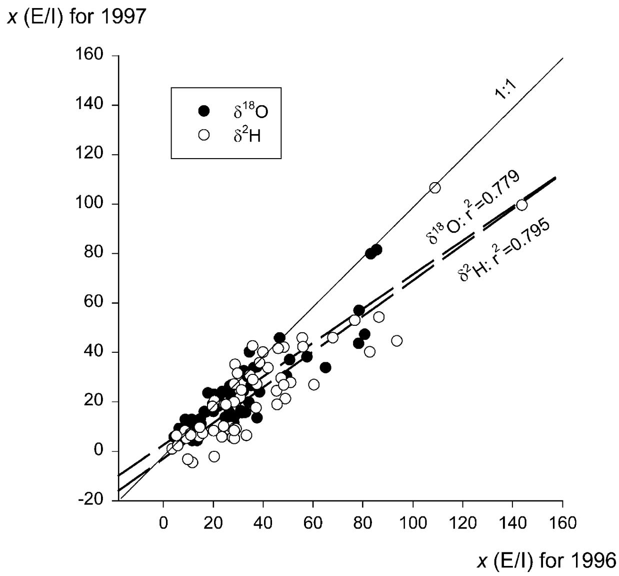

Fig. 7. Plot of x (E/I) for 1996 versus 1997 based on $\delta^{18}O$ and $\delta^{2}H$. Note that 1997 evaporation losses are estimated to be ~80% on average of 1996 water losses, based on both tracers, which is attributed mainly to inter-annual climate variations.

September. By introducing estimates of lake volume (mean depth), interpolated annual evaporation, and catchment morphological parameters, the residence time (τ) for each lake was estimated from Eq. (13) and the runoff/precipitation ratio was estimated using Eq. (16). Results are summarised in Table 1 and illustrated in Fig. 6.

Overall, the results show a wide range in lake water balance in the headwater lakes. Although there are regional consistencies (i.e. each lake group shows similar properties), the high intra-group variability verifies that local morphological setting may be equally important or more important than climatic controls on the water balance. This effect would be expected to diminish at lower levels of the drainage hierarchy as amalgamation of headwater basins are integrated and converge on a regional signal.

The calculated throughflow index (x) ranged from about 5 to 100% in individual reservoirs suggesting that lakes in the survey spanned the full range from terminal reservoirs to high-throughflow lakes. Note that several lakes (denoted by arrows in Fig. 6) were excluded from the summary table due to extreme conditions attributed to poor headwater situation (e.g. they were receiving water from upstream lakes, which was not detected in air photos). Overall, lakes in the upland Caribou Mountains and TROLS groups had mean x of about 30–50% (i.e. 30–50% of water loss by evaporation, 50–70% by surface/subsurface runoff). Lakes of NCE group, which tended to have a higher percentage of wetland cover had about 25% water loss by evaporation on average. A cross plot of x for 1996 versus 1997 (Fig. 7) reveals a close agreement between years with systematically reduced evaporation losses during 1997.

Comparison between estimates based on oxygen-18 and deuterium was used as a guideline to judge the uncertainty of the model. In general, estimates based on both tracers are in agreement to within ±10%.

Lakewater residence times (Table 1 and Fig. 6) are predicted to range from 0.1 to 14 years, and are consistent with values expected for lakes of the Boreal Plain (Prepas et al., 2001). Note that 'years' in this context refer to water years rather than calendar years. To calculate actual years, it is necessary to account for the inactive winter season, which would result in calendar year estimates of about 40% longer in the present setting.

Runoff/precipitation ratios are estimated to range from 1 to 185%, with average values of between 20 and 35% in both the Caribou Mountains and TROLS groups and close to 75% in the NCE group. Calculated

R/P values are high given that the runoff ratio is not expected to exceed about 50% in most situations (i.e. 50% of precipitation falling on the catchment reaches the lake). This may in part reflect problems with defining isotope exchange parameters, such as the limiting isotopic enrichment δ^*, but more likely reflects poor definition of catchment areas (required to estimate *R*/*P*) in very low-relief, large wetland basins. Despite this problem, isotope-based *R*/*P* ratios are still very useful for comparative analysis of the runoff regime.

As the throughflow index (x) is only dependent on the isotopic enrichment, the humidity, and the air temperature, and is not affected by potential errors related to catchment morphological characteristics or precipitation amounts, it is likely the most reliable quantitative isotope indicator. Residence time is also quite robust as it relies on x, the mean annual evaporation rate and mean lake depth, but absolute accuracy of *R*/*P* (and eDBA) is limited by quality of the catchment parameters. Analysis using isotope-based information could avoid this by comparing volumetric inflow rather than depth-equivalent parameters although this provides less information on the effect of various catchment characteristics and terrain configurations. Nevertheless, the isotope method provides a quantitative basis for comparison and classification of the hydrological regime.

2.3. Errors and sensitivity

Previous studies have provided a detailed overview of potential errors associated with application of isotope-based water balance models (Gibson et al., 1993, 1998). Although absolute errors are difficult to evaluate, the estimates are likely comparable to standard physically based water balance (Carignan et al., 2000), which has also been confirmed in detailed evaluation studies (Gibson et al., 1996a,b, 1998). Overall, accuracy of annual water balance estimates for the current application is likely not better than ±20–30%, although it is expected to be half an order of magnitude more precise for comparative applications. The principal advantage of the current approach is that accuracy is significantly improved by use of multiple time-series samples to characterise each lake water value. The principal drawback is that interpolation rather than direct measurement of isotope composition is used to characterise the input signatures. However, a good indication that the method is robust is that it was repeatable from isotope surveys conducted in 1996 and 1997 (Fig. 7). For regional applications, the field-based assessment of water balance parameters using isotope techniques is likely comparable or better than state-of-the-art hydrological modelling techniques that are not field-based, but this remains to be tested. One primary source of potential error is stratification effects, which are discussed later.

2.4. Effects of stratification

In general, integrated euphotic zone samples were collected and analysed from two locations in each lake, and the average value was used to estimate the isotopic composition of lake water. Integrated sampling ensured that values used for δ_L were representative. In some cases, the euphotic zone constituted the whole water column, but more often excluded the deepest portion of the lake. To verify the potential errors in water balance estimates arising from possible stratification effects, an additional 45 pairs of grab samples of epilimnion and hypolimnion water were collected and analysed from lakes in the TROLS group. Overall the pairs had mean differences in δ^{18}O (and δ^2H) of 0.25‰ (1.5‰), with maximum difference of 1.7‰ (11‰) and standard deviation of 0.46‰ (3.7‰). As the euphotic zone samples were integrated it is expected that differences between epilimnion and hypolimnion pairs represent an extreme scenario of potential errors arising from incomplete mixing. One standard deviation in δ_L of 0.46 and 3.7‰ in δ^{18}O and δ^2H, respectively, translates to potential errors in x and τ of not more than ±8% for x values in the range of 45 ± 28%, characteristic of the TROLS group, where the sampling was conducted (Table 1). Due to generally shallower depth of lakes in the NCE and Caribou Mountains groups, potential errors associated with stratification effects are likely to be less than a few percent, and do not overwhelm the primary water balance signals.

2.5. Water balance and catchment morphology

The relationship between catchment morphology and water balance is not straightforward, but

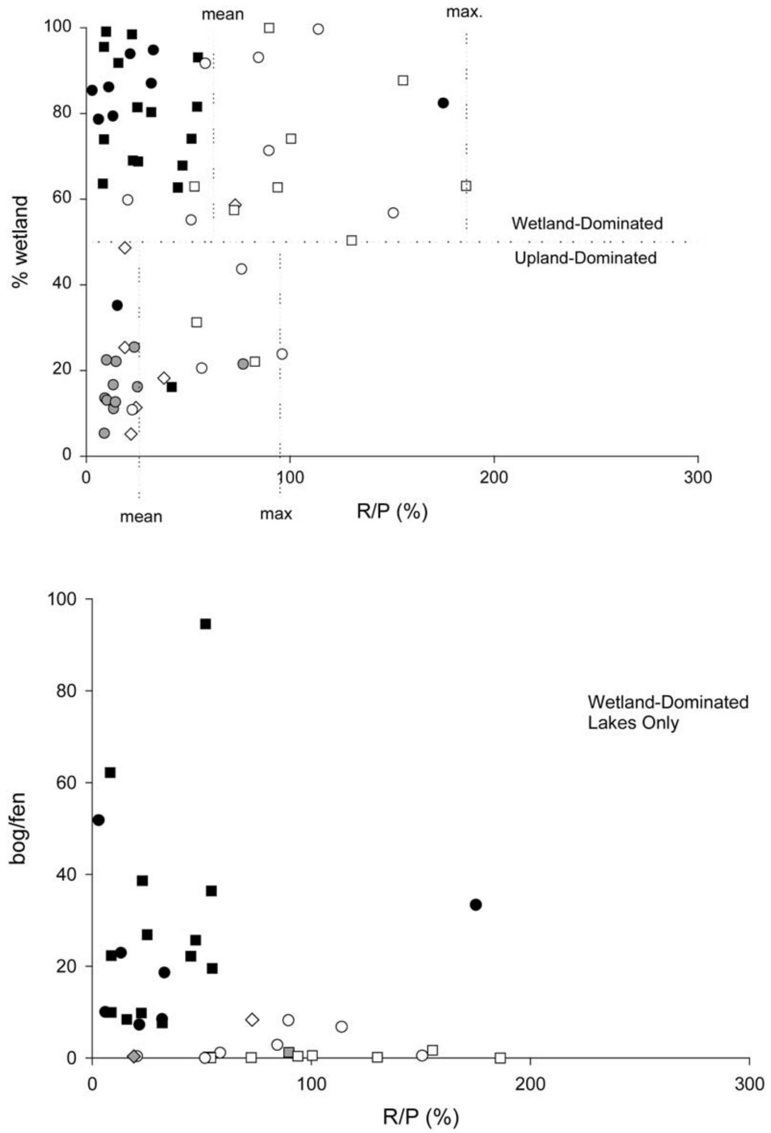

Fig. 8. (a) Plot of *R*/*P* (%) versus %wetland showing mean and maximum values of *R*/*P* for wetland- and upland-dominated basins and (b) *R*/*P* (%) versus bog/fen ratio for wetland-dominated basins. Note that higher *R*/*P* is noted for wetland-dominated basins, and especially for basins with a higher proportion of fens. Note that closed symbols are Caribou Mountains lakes, open symbols are NCE lakes and grey symbols are TROLS lakes; circles being for reference lakes, squares being for lakes in recently burnt catchments and diamonds being for lakes in harvested catchments.

systematic differences are detected in wetland- and upland-dominated catchments. Overall, lakes from all groups and treatment types spanned a wide range of water balance conditions. A plot of *R*/*P* versus %wetland for 1996 (Fig. 8(a)) reveals that lakes with wetland-dominated drainage basins (>50% wetland cover) have higher mean and maximum runoff ratios than upland-dominated basins. Although *R*/*P* should only be taken as a relative index of runoff, it is initially apparent that wetland-dominated basins appear to be hydraulically better connected to lakes than upland basins. Higher *R*/*P*

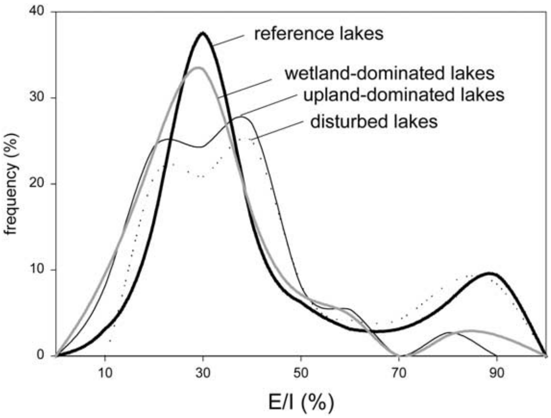

Fig. 9. Frequency (%) of lakes in a given *E*/*I* class. The plot distinguishes subgroups including (i) reference lakes (no disturbance in catchment area), (ii) disturbed lakes (with burnt or harvested catchments), (iii) wetland-dominated lakes, and (iv) upland-dominated lakes. In all cases, skewed distributions were obtained. This is attributed to similar internal structure of all systems, i.e. reduced contributions of water to lakes from high *E*/*I* (evaporative) watershed compartments and increased contributions from low *E*/*I* compartments.

ratios in wetland-dominated basins are not attributed to differences in vapour loss mechanisms in wetland versus upland systems. This would lead to more enriched lake water in wetland basins, higher x, and reduced R/P, as upland basins lose water predominantly via transpiration from soil zone, whereas wetland-dominated basins lose water by both transpiration and evaporation. One plausible explanation for this observation may be that upland-dominated basins are more efficient at detaining and removing water due to higher infiltration capacity and more efficient evapotranspiration. In addition, it is probable that R/P ratios may be somewhat overestimated in wetland-dominated basins due to systematic underestimation of catchment areas in lower-relief areas, where wetlands tend to occur.

Additional insight into controls on the runoff ratios can be gleaned from examining the bog/fen ratios in wetland-dominated catchments (Fig. 8(b)). In general, it is found that R/P tends to be much higher in fen-dominated areas than in bog-dominated areas. This suggests that fens may be more effective at delivering water from catchment areas to the lake.

Systematic patterns in the regional lake hydrology mosaic are particularly evident in frequency distribution plots of the throughflow index (x) (Fig. 9). Note that the regional distribution of x in lakes is skewed to lakes with lower x and lower isotopic enrichment, which is consistent with reduced runoff contributions from highly evaporation sources. This applies to reference lakes, wetland-dominated lakes, upland-dominated lakes and disturbed lakes. Non-headwater lakes and streams at lower levels in the drainage hierarchy, which collect water from large number of lakes might be expected to acquire an isotopic signal reflecting close to 25–30% water loss by evaporation in the present setting or about -16 and -138‰ in $\delta^{18}O$ and $\delta^{2}H$, respectively. This is close to summer baseflow values observed in large local rivers (Hitchon and Krouse, 1972).

3. Concluding comments

Application of the isotope balance approach in the present study provides a framework for quantitative comparison of hydrology and hydrologic changes arising from forest fire and clear-cut harvesting (McEachern et al., 2000; Prepas et al., 2001). As the method is ideally suited to incorporation in water quality surveys, it offers new possibilities for examining hydrologic controls on the aquatic chemistry and ecology of lakes. It is also a potentially valuable tool for study of the role of lakes in regional water balance at nodes in the drainage hierarchy. Ongoing research is focused on developing a more sophisticated understanding of large-scale characteristics of the isotope exchange parameters to permit more extensive use of isotope mass balance in regional surveys. Additional studies are investigating hydrograph and flux-based methods for partitioning of evaporation and transpiration, mainly in Boreal and Arctic areas.

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