Journal of Hydrology (2008) **350**, 1–13

available at www.sciencedirect.com

ScienceDirect

journal homepage: www.elsevier.com/locate/jhydrol

A coupled isotope tracer method to characterize input water to lakes

Yi Yi [a,*], Bronwyn E. Brock [a], Matthew D. Falcone [a], Brent B. Wolfe [a,b], Thomas W.D. Edwards [a]

[a] *Department of Earth and Environmental Sciences, University of Waterloo, Waterloo, ON, N2L 3G1 Canada*
[b] *Department of Geography and Environmental Studies, Wilfrid Laurier University, Waterloo, ON, N2L 3C5 Canada*

Received 31 July 2007; received in revised form 7 November 2007; accepted 9 November 2007

KEYWORDS
Water isotope tracers;
Isotope hydrology;
Craig and Gordon model;
Isotope-mass balance;
Peace-Athabasca Delta

Summary We develop a new coupled isotope tracer method for characterizing the isotopic composition of input water to lakes, and apply it in the context of ongoing hydrological process studies in the Peace-Athabasca Delta, a large, remote, riparian ecosystem in the boreal region of western Canada. The region has a highly seasonal climate, with floodplain lakes typically receiving input only during the 4–6 month open-water season from varying proportions of spring snowmelt, summer rains and river flooding. These possible input sources have distinct ranges of isotopic compositions that are strongly constrained to a well-defined local meteoric water line, thus affording the opportunity to derive lake-specific estimates of the integrated isotopic composition of input waters after accounting for the effects of secondary evaporative isotopic enrichment. As shown by comparison of the results of isotopic surveys of delta lakes prior to freeze-up in 2000 and 2005, this isotopic characterization of input waters can be combined with other data and field observations to provide new insight into spatial and temporal variability in delta lake recharge processes. This includes evidence that summer rainfall in 2000 played an important role in replenishing shallow basins delta-wide, especially in the central low-lying region, compensating for below-average snow accumulation during the previous winter. In contrast, 2005 was marked by greater relative contributions from both snowmelt and river flooding because of high winter snow accumulation and a spring ice-jam that caused river floodwaters to enter some basins in the southern part of the delta. The method is readily transferable to investigations in other remote regions that are sparsely monitored by conventional hydrometric networks.

---

[*] Corresponding author. Tel.: +1 519 888 4567x33966; fax: +1 519 746 7484.
E-mail address: y2yi@uwaterloo.ca (Y. Yi).

0022-1694/$ - see front matter
doi:10.1016/j.jhydrol.2007.11.008

Introduction

Understanding the relative roles of hydrological processes on lake water balances in large and remote freshwater ecosystems is important for their management but is challenging because approaches that depend upon conventional instrumentation may be impractical. Installation of hydrometric devices in ecosystems with numerous aquatic basins requires substantial investment and assumptions are frequently required to, for instance, close lake water balances. For basins located in deltas and river floodplains that receive multiple sources of water, lake level recorders may provide useful knowledge of water level increases but other independent hydroclimatic information is needed to identify the cause of the observed change. Furthermore, results derived from only a few instrumented sites may be difficult or inappropriate to extrapolate over complex landscapes where the relative importance of hydrological processes is expected to vary spatially.

In recent years, water isotope tracers have become increasingly utilized as an alternative approach for providing assessments of lake water balances in remote regions. For example, Gibson and Edwards (2002) conducted a systematic survey of lake water isotope compositions to understand regional variations in evaporation losses and water budgets associated with the climate gradient across the northern boreal treeline in Canada. More recently, Mayr et al. (2007) used water isotope tracers in the southern Patagonia of Argentina to characterize contemporary lake water balances with the purpose of informing paleolimnological investigations. In these and other studies, quantitative estimates of lake water balances are derived using variations of the Craig and Gordon (1965) linear resistance model that describes isotopic evaporative enrichment. Applications commonly utilize lake water oxygen and hydrogen isotope compositions separately, which frequently generates small differences in water balance estimates using the individual tracers. Although these differences are often attributed to analytical or model uncertainties, mass conservation dictates that lake water balances calculated from lake water oxygen and hydrogen isotope compositions must agree.

Here we develop a new approach to the application of water isotope tracers that preserves the fundamental assumption of mass conservation allowing additional hydrological information to be derived regarding the nature of source waters to lakes in the Peace-Athabasca Delta (PAD), Canada. This large freshwater ecosystem contains hundreds of shallow (most are <2 m) basins where rainfall, snowmelt and river water are important to sustain aquatic habitat but their relative roles over space and time are not well characterized. These investigations build upon quantitative assessment of lake water balances across the PAD, which identified distinct landscape sectors of hydrolimnological conditions based on integration of isotope and chemistry analyses of lake water samples collected in October 2000 (Wolfe et al., 2007). In this previous study, as elsewhere, comparison of oxygen- and hydrogen-isotope estimated evaporation-to-inflow ratios revealed a small but systematic departure from the 1:1 line that was thought to emanate from uncertainties in model input values. We re-visit this dataset and utilize another from September 2005 to evaluate input water isotope compositions during two years characterized by very different hydrological and meteorological conditions. Results show marked spatial and temporal variability in input waters to lakes using our new coupled isotope tracer method, one that is readily transferable to other large, freshwater ecosystems.

Study area

The Peace-Athabasca Delta (PAD) is a large (~3900 km^2) wetland complex located at the convergence of the Peace, Athabasca and Birch rivers at the western end of Lake Athabasca, northern Alberta, Canada (Fig. 1). The PAD can be subdivided into three deltaic sectors: the Athabasca sector to the south (~1970 km^2), the Peace sector to the north (~1680 km^2) and the much smaller Birch sector to the west (~170 km^2) (PADPG, 1973). Several large shallow lakes (Claire, Baril and Mamawi lakes) are located in the center of the PAD, where the three sectors coalesce (Fig. 1). The Peace sector lies to the north of these lakes and is a relict fluviodeltaic landscape that is covered by mature forests with bedrock inliers in the northeast. This sector is flooded only during major ice-jams that periodically develop on the Peace River. The southern Athabasca sector is an active delta of extremely low relief that frequently receives river floodwaters during both the spring thaw and open-water seasons. Ice-jam flooding of portions of the Athabasca sector occurred in the spring of 2005, preceding one of our lake water sample collections (September 2005).

There are numerous shallow lakes in the PAD, which span a broad hydrological spectrum. Based on previous studies incorporating water isotope tracers and limnological characteristics, basins have been categorized into four drainage types (Wolfe et al., 2007). Open-drainage basins are located mainly in the low-lying central portion of the delta where they frequently receive discharge from many of the rivers and creeks that constitute the complex channel network of the PAD. Closed-drainage basins are generally found in the Peace sector and receive widespread river water only during periodic ice-jam flood events on the Peace River, and thus, input from precipitation is an important source of water to these lakes. Restricted-drainage basins are located mainly in the Athabasca sector, where the input of river water is the primary hydrological process that controls lake water balances. Rainfall-influenced basins are found mainly in the central portion of the delta adjacent to the large open-drainage basins and occupy shallow depressions in the landscape (depth <50 cm). Their water balances are similar to those of restricted-drainage basins but their source waters are dominated by summer precipitation.

Climate in the PAD is strongly seasonal. According to 1971–2000 climate normals at Fort Chipewyan, Alberta, (Weather Station ID 3072658; Environment Canada, 2004), mean annual air temperature is −1.9 °C, mean January air temperature is −23.3 °C and mean July air temperature is 16.7 °C. Precipitation averages 391.7 mm annually, with about 59% falling as rain during the May–September period. Meteorological conditions preceding the two sample collections (October 2000 and September 2005) were different (Fig. 2). The winter of 1999/2000 was warm and dry

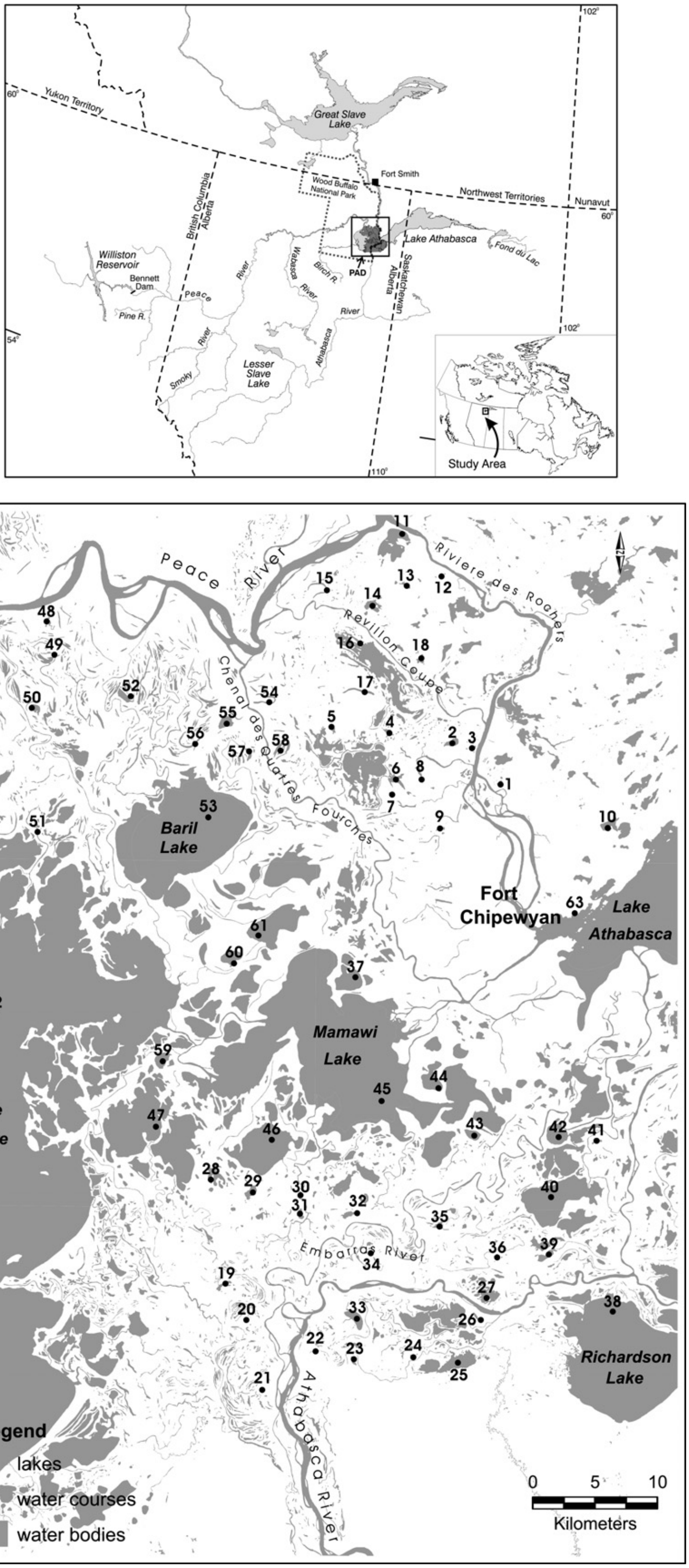

Figure 1 Location of the Peace-Athabasca Delta, Alberta, Canada and lake water sampling sites.

(especially from January to April); rainfall was evenly distributed throughout the following summer. In contrast, the winter of 2004/2005 had much greater snowfall throughout the latter months of the season and greater rainfall in August.

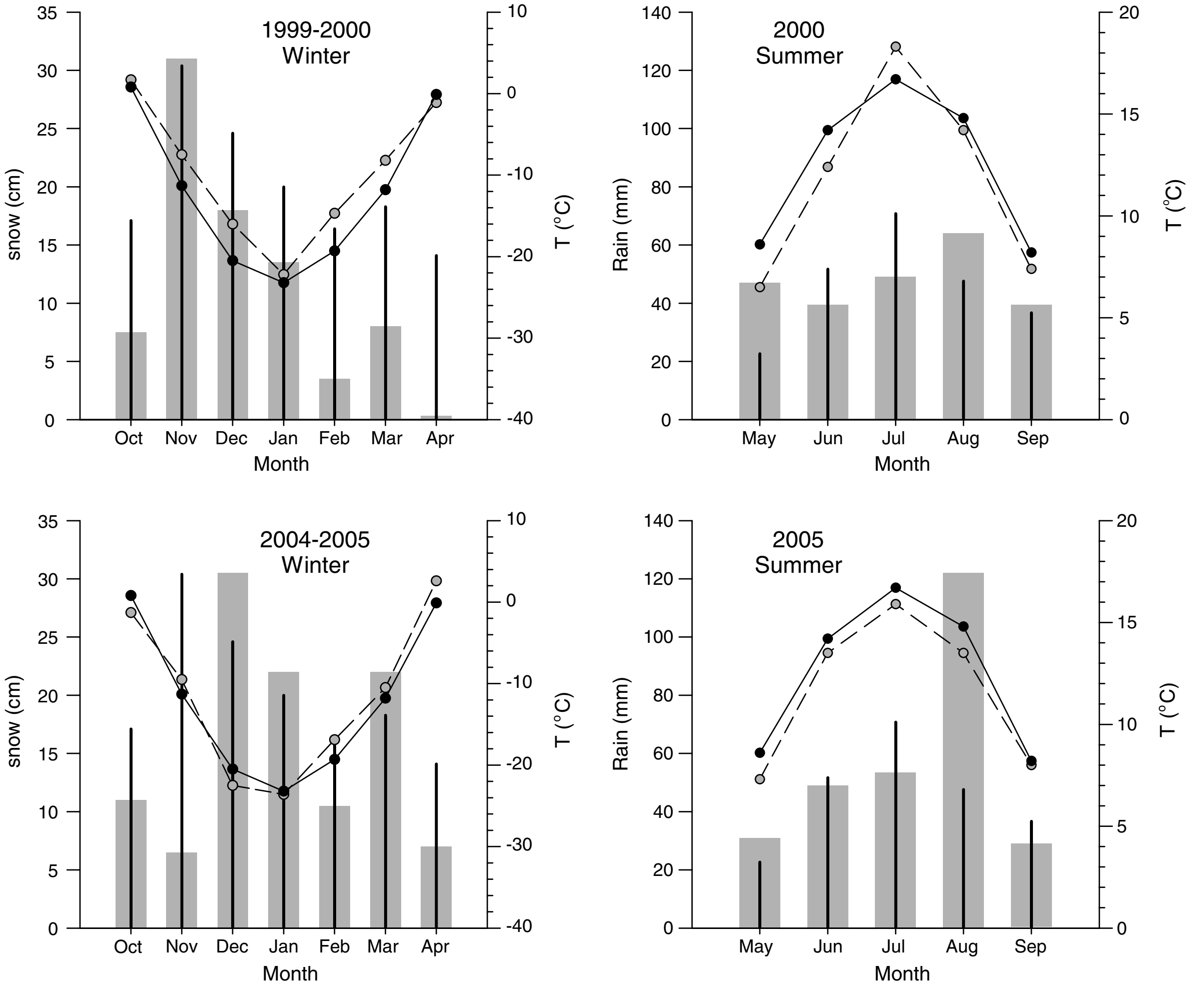

Figure 2 Winter and summer monthly mean temperature and accumulated precipitation for 1999/2000 and 2004/2005 compared to 1971—2000 climate normals. Climate normals are shown as solid circles (temperature) and bars (precipitation).

Sampling and analysis

Water samples for oxygen and hydrogen isotope analysis were collected from 62 lakes on 25 October 2000 and most of these same lakes ($n = 54$) were re-sampled on 14 September 2005 (Fig. 1). Samples were collected from approximately 10 cm below the surface at the center of each lake with the aid of a helicopter. All samples were sealed in 30 ml high-density polyethylene bottles and were analyzed at the University of Waterloo — Environmental Isotope Laboratory, where $^{18}O/^{16}O$ and $^{2}H/^{1}H$ ratios were measured using standard methods (Coleman et al., 1982; Drimmie and Heemskerk, 1993; Epstein and Mayeda, 1953). Results are reported in δ values, representing deviations in per mil (‰) from VSMOW on a scale normalized to values of Standard Light Antarctic Precipitation (−55.5‰ for $\delta^{18}O$; −428‰ for $\delta^{2}H$; Coplen, 1996). Analytical uncertainties are ±0.2‰ for $\delta^{18}O$ and ±2.0‰ for $\delta^{2}H$. To demonstrate the potential variability in input water isotope composition to lakes in the PAD, results from water samples collected over six years from two basins located in different hydrological settings (PAD 37 and PAD 54; Fig. 1) are also reported.

Theory

Isotopic labelling in the hydrological cycle

The distribution of water isotopes on the earth's surface (i.e., in precipitation and surface waters) is characterized by the existence of strong linear relations between $\delta^{2}H$ and $\delta^{18}O$ over a broad range of spatial and temporal scales, reflecting systematic mass-dependent isotope fractionation in the hydrological cycle (Rozanski et al., 1993; Gat, 1996; Gibson and Edwards, 2002). The most salient feature of the covariant behaviour between hydrogen and oxygen isotopes in the global water cycle is the global meteoric water line (GMWL), which is expressed by the linear function $\delta^{2}H = 8\delta^{18}O + 10$ (Craig, 1961). The slope reflects the dominant influence of temperature-dependent equilibrium fractionation of heavy isotope species between atmospheric vapour and condensing precipitation, while the linearity is consistent with the notion that, at the global scale, atmospheric moisture primarily arises from one large water source (i.e., the subtropic ocean surface) and undergoes progressive distillation during poleward atmospheric transport. The GMWL has proven to be an especially useful reference line for understanding spatial patterns in the variability of the isotopic composition of mean annual precipitation (Rozanski et al., 1993). Analogous labelling of precipitation occurs at local scale in the form of local meteoric water lines (LMWLs), reflecting temporal variability in the isotopic composition of local precipitation. LMWLs commonly lie close to the GMWL, usually with a lower slope because of raindrop re-evaporation and kinetic effects during snow formation (Rozanski et al., 1993).

The isotopic composition of water that has undergone evaporation diverges from the GMWL (or LMWL) and the variation of hydrogen- and oxygen-isotope signals in surface waters is systematic because of mass-dependent fractionation that occurs during isotopic enrichment. Typically, the isotopic compositions of neighbouring water bodies receiving input water similar in isotopic composition cluster along a well-defined linear trajectory termed a local evaporation line (LEL). In δ^{18}O-δ^{2}H space, the slope of the LEL usually ranges between 4 and 6 and is primarily controlled by local atmospheric conditions including relative humidity (h), temperature (T) and the isotopic composition of atmospheric moisture (δ_A) (Gibson et al., in press). Moreover, the relative position of a given lake along the LEL is strongly associated with the water balance of the lake (Gonfiantini, 1986; Gat, 1996; Gibson and Edwards, 2002). Overall, the existence of two linear trends in δ^{18}O-δ^{2}H space allows differentiation of meteoric water from surface water that has undergone secondary isotopic enrichment due to evaporation.

Isotope-mass balance modelling

Quantitative investigations of lake water balances using water isotope tracers have been based on a conceptual framework incorporating individual water budget components in δ^{18}O-δ^{2}H space (Fig. 3). For any given lake, δ_L, δ_E and δ_I represent the isotopic composition of the lake water, evaporative flux from the lake and the input water to the lake, respectively. δ_I is commonly estimated by the intersection of the LMWL with the best-fit line through a time-series of lake water isotope compositions or a spatial distribution of lake water isotope compositions in a watershed (e.g., Gibson et al., 1993). According to mass conservation, $_E$ should lie on the extension of the LEL to the left of the LMWL. Also shown in Fig. 3 is the isotopic composition of atmospheric moisture δ_A and the limiting isotope composition δ^*, which is the maximum isotopic enrichment attainable under local atmospheric conditions when a water body approaches desiccation. δ^* is independent of hydrological conditions (i.e., δ_L and δ_I) but related to atmospheric conditions (i.e., h, T and δ_A; Welhan and Fritz, 1977; Allison and Leaney, 1982). Theoretically, regardless of the hydrological complexity of a landscape, the isotopic composition of all lakes in an area with similar atmospheric conditions will converge towards δ^*.

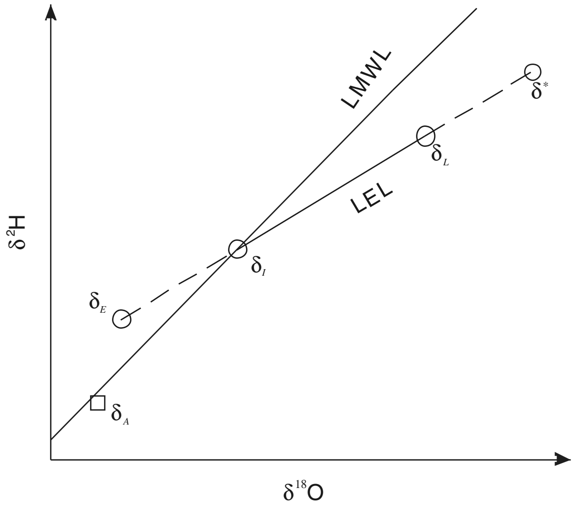

Figure 3 Schematic δ^{18}O-δ^{2}H diagram identifying key isotopic parameters used in isotope-mass balance studies. These include lake water isotope composition (δ_L), input water isotope composition (δ_I), isotopic composition of evaporated vapour from the lake (δ_E) and the limiting isotopic composition (δ^*). The atmospheric moisture isotope composition (δ_A) is also shown.

Isotopic composition of evaporated vapour (δ_E)

Evaporated flux is a critical component for isotope mass-balance calculations but it is difficult to obtain samples directly for isotopic measurements. Studies have conventionally quantified δ_E using the Craig and Gordon (1965) linear resistance model, which includes a molecular diffusion layer bounded below by a ''virtually saturated'' layer and above by a turbulent transportation layer. The fluxes through the molecular diffusion layer are linearly proportional to the vapour concentration difference between the upper and lower boundaries. As such, the vapour flux of $^{1}H_2{}^{16}O$ (i.e., the common ''light'' water isotopologue) across the liquid—vapour interface is given by

$$E = (C_s - C_a)/\rho = C_s(1-h)/\rho \tag{1}$$

where C_s is the vapour concentration in the ''virtually saturated'' layer, C_a is the vapour concentration in the turbulent transportation layer, h is the relative humidity of the turbulent layer normalized to the water surface temperature and ρ is a resistance coefficient (Gonfiantini, 1986; Gat, 1996). Similarly, the vapour flux of heavy water isotopologues (i.e., $^{1}H^{2}H^{16}O$ or $^{1}H_2{}^{18}O$) across the same molecular diffusion layer is given by

$$E_i = (C_s R_s - C_a R_a)/\rho_i = C_s(R_s - hR_a)/\rho_i \tag{2}$$

Here, the subscript i designates water molecules bearing the heavy isotope species. R_s and R_a are the isotopic ratios of the water vapour in the saturated and turbulent layers, respectively. Since water vapour in the saturated layer should be in equilibrium with surface water at the saturated condition, R_s can be quantitatively related to the isotopic ratio of lake water (R_L) by the definition of equilibrium fractionation:

$$R_L/R_s = \alpha^* \tag{3}$$

where α^* is the liquid—vapour equilibrium fractionation factor (i.e., $\alpha^* > 1$). Isotopic fractionation is also commonly expressed by a separation term (ε^*), where $\varepsilon^* = \alpha^* - 1$ (for ε^* in decimal notation). The ε^* and α^* parameters are dependent on temperature and can be calculated using empirical equations (Horita and Wesolowski, 1994).

According to the definition of the isotopic ratio, R_E can be expressed as $R_E = E_i/E$. By combining Eqs. (1)—(3), R_E can be expressed as

$$R_E = \frac{\rho}{\rho_i} \cdot \frac{R_L/\alpha^* - hR_a}{1-h} \tag{4}$$

Introducing ε_k, the kinetic separation term according to Craig and Gordon (1965) and Gonfiantini (1986):

$$\varepsilon_k = (1-h)\left(\frac{\rho_i}{\rho} - 1\right) \tag{5}$$

Eq. (4) can then be rewritten in δ-notation:

$$\delta_E = \frac{(\delta_L - \varepsilon^*)/\alpha^* - h\delta_A - \varepsilon_k}{1 - h + \varepsilon_k} \tag{6}$$

where δ_L and δ_A are the isotopic compositions of lake water and atmospheric moisture, and ε^* and ε_k are equilibrium and kinetic separation terms, as defined above. ε_k is estimated as a function of the relative humidity deficit, described elsewhere (Gonfiantini, 1986; Gibson and Edwards, 2002; Edwards et al., 2004). Note that Eq. (6) is formulated for δ, ε and h values in decimal notation. Eq. (6) can be further simplified on the basis of various assumptions (Welhan and Fritz, 1977; Allison and Leaney, 1982; Ferhi et al., 1983; Gat, 1996; Yakir and Sternberg, 2000; Gibson and Edwards, 2002; Vallet-Coulomb et al., 2006) but we apply Eq. (6) as it was presented by Gonfiantini (1986) and as shown here.

The quantitative evaluation of δ_E, as expressed in Eq. (6), requires values for the isotopic composition of atmospheric moisture (δ_A), the isotopic composition of lake water (δ_L), relative humidity (h) and temperature (T). Of these parameters, δ_L, h and T can be measured routinely. Direct measurement of δ_A, on the other hand, is difficult because sample collection is logistically challenging and can introduce unintended isotopic fractionation. Given the dynamic nature of the atmosphere, single sample of atmospheric moisture may not accurately represent an integrated isotopic mean value, especially in regions with strong seasonality (e.g., Jacob and Sonntag, 1991). Alternatively, several indirect methods have been used to estimate δ_A. The most straightforward and frequently applied method is to assume isotopic equilibrium between evaporation-flux-weighted local precipitation and atmospheric moisture (Gibson, 2002). Although this method has been applied extensively in hydrological studies (Zuber, 1983; Gibson et al., 1993; Edwards et al., 2004), the assumption of equilibrium between precipitation and atmospheric moisture may not be valid if terrestrial recycling of evapotranspirative vapour is significant, as is the case in the Great Lakes region (Gat et al., 1994) and the Amazon basin (Gat and Matsui, 1991). A constant-volume pan can be used to derive δ_A (Welhan and Fritz, 1977; Allison and Leaney, 1982; Gibson et al., 1999) but this requires careful maintenance of the pan over the investigation period, which may not be practical in remote regions. Where available, studies have used an index lake (e.g., Dinçer, 1968) as a natural analogue of a constant-volume pan to generate a time-integrated estimate of δ_A.

Fortunately, our studies have identified a deep (>8 m) closed-drainage lake (PAD 18; see Fig. 1) that is an excellent index lake for the PAD region. The lake is perched well beyond the possible influence of river flooding, and hence is fed solely by precipitation. Several years (2000—2006) of field observation have revealed minimal interannual variation in water level (±0.2 m) and near-constant isotopic composition ($-9.1 \pm 0.5‰$ for $\delta^{18}O$; $-104 \pm 2‰$ for δ^2H), as expected if input is closely compensated by evaporation (i.e., $E/I \approx 1$) (Hall et al., 2004; Falcone, 2007; Wolfe et al., 2007). Based on catchment area and meteorological records, the lake water has an estimated mean residence time on the order of 10 years, which suggests that the system integrates precipitation inputs over multi-annual time-scales. The mean annual isotopic composition of precipitation based on the nearest monitoring station (Fort Smith, Northwest Territories, 60 °N 112 °W; $-18.8‰$ for $\delta^{18}O$; $-147‰$ for δ^2H, Birks et al., 2004) agrees very closely with shallow groundwaters sampled directly in the catchment of PAD 18 in 2003 ($-18.5‰$ for $\delta^{18}O$; $-146‰$ for δ^2H). By solving Eq. (6) for both isotopes on the basis that the isotopic composition and mass of annual liquid input and thaw-season vapour output from PAD 18 are identical ($\delta_E = \delta_I$), we obtain an estimate of δ_A ($-26.6‰$; $-205‰$). Notably, this is very close to δ_A ($-25.4‰$; $-211‰$) estimated previously by Wolfe et al. (2007) assuming equilibrium with thaw-season precipitation (Gibson and Edwards, 2002), and in good agreement with various ''snapshot'' estimates of δ_A that we have obtained from field experiments using constant-volume pans (Falcone, 2007). As a result, we apply the index lake method to infer δ_A in this paper.

General lake water balance

The water-mass and isotope-mass balances of a well-mixed lake at hydrological and isotopic steady-state are (Gonfiantini, 1986; Gibson and Edwards, 2002):

$$I = Q + E \tag{7}$$

$$I\delta_I = Q\delta_Q + E\delta_E \tag{8}$$

where I is the inflow rate ($m^3\,s^{-1}$), Q is the outflow rate ($m^3\,s^{-1}$) and E is the evaporation rate ($m^3\,s^{-1}$). δ_I, δ_Q and δ_E represent the isotopic compositions of their corresponding hydrological components. Physical outflow does not cause isotopic fractionation, therefore outflow is isotopically equal to lake water (i.e., $\delta_Q = \delta_L$). By combining Eqs. (7) and (8), the evaporation-to-inflow (E/I) ratio can be expressed using the isotopic compositions of each water budget component:

$$E/I = \frac{\delta_I - \delta_L}{\delta_E - \delta_L} \tag{9}$$

Eq. (9) has been used extensively to assess lake water balances (e.g., Dinçer, 1968; Gat and Levy, 1978; Gonfiantini, 1986; Gibson and Edwards, 2002; Vallet-Coulomb et al., 2006; Mayr et al., 2007).

Results and discussion

Developing a coupled isotope tracer method to characterize lake-specific input water

Isotopic composition of water samples collected from two lakes, PAD 37 and PAD 54, over a six-year period (2000—2005) show characteristic linear patterns when plotted in $\delta^{18}O$-δ^2H space (Fig. 4a). Although all samples plot below the LMWL ($\delta^2H = 6.7\delta^{18}O - 19.2$; Birks et al., 2004), indicating varying degrees of the importance of evaporation to the

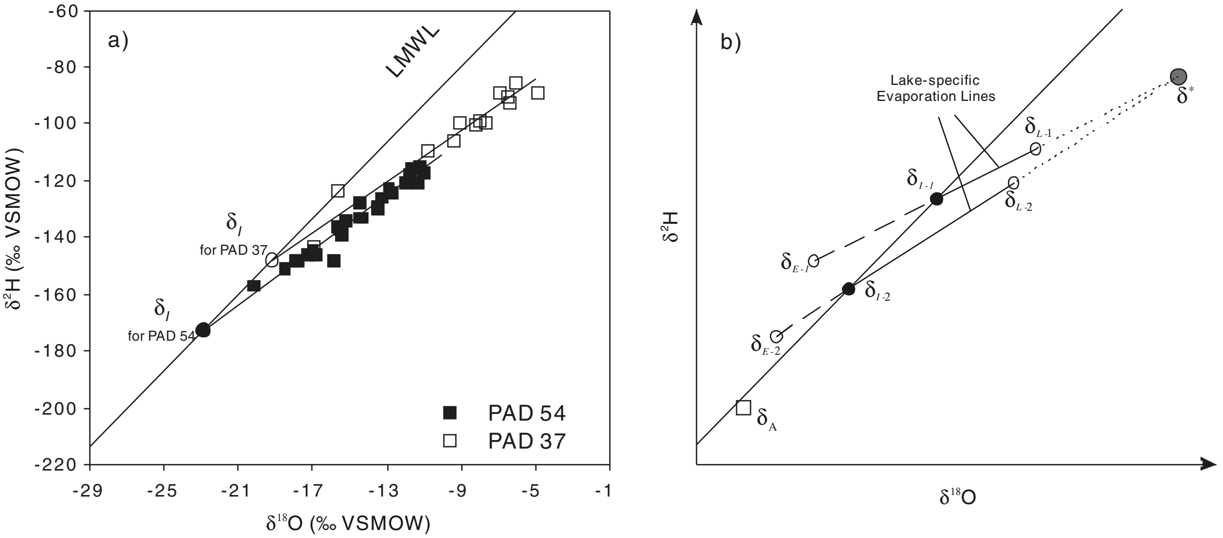

Figure 4 (a) Lake water isotope compositions (2000—2005) for PAD 54 and PAD 37. Best-fit lines through these datasets intersect the LMWL at different points, suggesting that the lakes are fed by water of differing isotopic compositions. (b) Schematic δ^{18}O-δ^2H diagram illustrating that the isotopic composition in a region that experiences similar atmospheric conditions will converge to the limiting isotopic composition (δ^*) with increasing evaporation, independent of δ_I.

lake water balances, their respective trajectories are subtly offset, suggesting these two lakes are fed by input waters with slightly different time-integrated isotopic compositions (δ_I). For PAD 54, the estimated δ_I ($-22.9‰$ for δ^{18}O and $-173‰$ for δ^2H) is more depleted than that for PAD 37 ($\delta^{18}O = -19.2‰$; $\delta^2H = -148‰$), which is consistent with their different hydrological settings. PAD 54 is a deep (~4 m) oxbow lake near the Peace River, which is known to receive isotopically-depleted input from both occasional river floodwater and snowmelt runoff generated from its large contributing area. On the other hand, PAD 37 is a shallow basin (<1 m deep) in the central part of the PAD having minimal contributing catchment, and is thus more dependent on the input from rainfall, which is typically isotopically-enriched relative to snowmelt and river water. Indeed, at least one sampling episode appears to have captured almost complete replenishment of the lake by recent rainfall, judging by the sample point lying on the LMWL (Fig. 4a).

Variability in δ_I can also affect the accuracy of isotope-based water balance assessments (see Eq. (9)). In this environment, individual lakes (such as PAD 37 and PAD 54) possess their own evaporation lines anchored by lake-specific input water isotope compositions. Moreover, in a region with similar, well-mixed atmospheric conditions (i.e., h, T and δ_A), the limiting isotope compositions (δ^*) for different lakes, regardless of lake water balance status (including δ_I), are predicted to be the same. As a result, all lake-specific LELs of varying δ_I and slope will converge towards a common δ^* (Fig. 4b).

In previous isotope-mass balance studies of lakes in the PAD (Wolfe et al., 2007), variability in δ_I was recognized and estimated from the intersection of lake-specific LELs, drawn parallel to a regional predicted LEL, and the LMWL. This resulted in conservative estimates of δ_I with respect to variability about the mean annual isotope composition of precipitation and generated reasonable water balance estimates largely supported by limnological parameters. However, this approach is not consistent with the notion that all lakes in a region should tend towards the same δ^*. Below, we present a more rigorous mathematical solution for δ_I by assuming equivalent E/I ratios derived from both δ^{18}O and δ^2H signatures, thereby conserving both mass and isotopes.

As illustrated in Fig. 4b, δ_I is expected to lie on the line between δ_L and δ_E. δ_I can be further constrained to the LMWL, since the input water is primarily meteoric water that has not undergone evaporation. Rain and snow samples collected during 2000—2005 showed good agreement with this LMWL (Falcone, 2007). This leads to the expression below that directly couples oxygen and hydrogen isotope signatures from Eq. (9):

$$\frac{\delta^2H_I - \delta^2H_L}{\delta^2H_E - \delta^2H_L} = \frac{\delta^{18}O_I - \delta^{18}O_L}{\delta^{18}O_E - \delta^{18}O_L} \tag{10}$$

With calculation of δ_E from Eq. (6) and measurement of δ_L, Eq. (10) can be used to solve for a unique pair of $\delta^{18}O_I$ and δ^2H_I values by incorporating the LMWL. Herein, we refer to this approach as a coupled isotope tracer method for estimating input water isotope composition to lakes. In the following section, we test this approach by characterizing input waters to lakes in the PAD sampled at the end of two thaw seasons having contrasting hydrological and meteorological conditions.

Lake water input composition in the Peace-Athabasca Delta for 2000 and 2005

Isotopic compositions of lake water sampled in October 2000 and September 2005 show similar linear trends offset from the LMWL in δ^{18}O-δ^2H space but different distribution patterns (Fig. 5a and b). For lake waters sampled in 2000 (Fig. 5a), the isotopic compositions range from $-13.9‰$ to

-6.7‰ in δ^{18}O (-124‰ to -89‰ in δ^{2}H) with the exception of PAD 45, which has a low isotopic composition (-17.3‰ and -139‰ for δ^{18}O and δ^{2}H, respectively) and lies close to the LMWL. For 2005 (Fig. 5b), lake water isotope compositions range from -16.9‰ to -9.6‰ in δ^{18}O (-137‰ to -105‰ in δ^{2}H). In general, lake water isotope compositions in 2005 are more depleted than for 2000 and exhibit less scatter (R^2 is 0.88 for 2000 and 0.96 for 2005).

Applying the coupled isotope tracer method (Eq. (10)) to the 2000 and 2005 lake water isotope composition datasets yields lake-specific δ_I values, integrated over the thaw season, for each corresponding δ_L (Fig. 5a and b). Calculations revealed that δ_I values for the two years span discrete (though overlapping) spectra along the LMWL. In 2000, δ_I values vary from -22.7‰ to -12.0‰ in δ^{18}O (-171‰ to -100‰ in δ^{2}H), while δ_I values in 2005 range from -26.2‰ to -16.1‰ in δ^{18}O (-195‰ to -127‰ in δ^{2}H).

Snowmelt, rainfall and river water are the three main sources of water to lakes in the PAD (Peters, 2003; Peters et al., 2006). It is expected that snowmelt and rainfall will affect all delta lakes to some extent, while influence from river water occurs perennially for lakes in direct connection with the channel network (i.e., open-drainage basins) and only during flood events for other lakes. These different types of source waters also bear distinct isotopic signatures based on extensive sampling of snow, rain and river water from 2000-2005. The isotopic composition of rain ranges from -20.5‰ to -13.9‰ for δ^{18}O (-168‰ to -112‰ for δ^{2}H), whereas the isotopic composition of snow ranges from -30.6‰ to -19.4‰ for δ^{18}O (-233‰ to -141‰ for δ^{2}H). While a wide range of isotopic compositions have been obtained for rain and snow, the isotopic composition of river water spans a much narrower range. The isotopic composition of the Athabasca River ranges from -19.3‰ to -17.5‰ for δ^{18}O (-154‰ to -142‰ for δ^{2}H). The Peace River is isotopically depleted compared to the Athabasca River but also brackets a narrow range of values (-20.1‰ to -19.0‰ for δ^{18}O; -159‰ to -150‰ for δ^{2}H). For the convenience of illustration and discussion, we present the isotopic ranges of potential source waters to lakes (i.e., rain, snow and river water) as gray lines in Fig. 5.

The relative influence of the three main source waters to lakes can be evaluated by the position of δ_I along the defined segments of the LMWL. Based on mixing, input water with a high proportion of snowmelt will tend to be more isotopically depleted than input water composed of a high proportion of rainfall. Therefore, lakes with low δ_I values, within the isotopic range of snow, would suggest that snowmelt plays an important role in the lake's water budget for that particular year, while higher δ_I values would suggest a greater proportion of rainfall contributing to these lake basins. Because of the narrow isotopic variability of river water, δ_I values for flooded lakes are expected to be similar to that of river water. This concept is readily demonstrated in the δ_I results for lakes with well-understood and relatively simple water budgets. For instance, several lakes ($n = 10$) that have previously been identified as rainfall-influenced basins based on their hydrolimnological characteristics and others, such as PAD 45 (Mamawi Lake; Fig. 1), are open-drainage basins that have permanent channel connections with major rivers (see Appendix D in Wolfe et al., 2007). The input to these basins is dominated by a single source of water (i.e., summer rainfall for the rainfall-influenced lakes and river water for PAD 45) and they have straightforward water budgets. δ_I results for rainfall-influenced basins are positioned at the enriched end of the isotopic spectrum of input waters for both 2000 and 2005, whereas δ_I for PAD 45 averages -18.7‰ in δ^{18}O (-144‰ in δ^{2}H) over the two sample collection years, which is well within the range of its primary source of water, the Athabasca River (-19.3‰ to -17.5‰ for δ^{18}O; -154‰ to -142‰ for δ^{2}H) (Fig. 5a and b).

At the other end of the input water isotope spectrum, some lakes are primarily influenced by snowmelt runoff during the spring thaw. Based on the δ_I results, PAD 15 and PAD 49 appear to be strongly influenced by snowmelt in 2000 and 2005, respectively (Fig. 5a and b). PAD 15 is a deep (~4 m) oxbow lake possessing catchment attributes similar to PAD 54 described above. Its large contributing area likely also provides substantial isotopically-depleted snowmelt runoff to the lake. PAD 49 is a small, shallow (~0.6 m), closed-drainage basin. Both of these lakes are surrounded by mature forests, which tend to accumulate snow during the winter, while the forest canopy intercepts direct rainfall during the thaw season. As a result, locally accumulated winter snowpack recharges these lakes rapidly during spring snowmelt and more gradually by sustained shallow groundwater flow over the thaw season. Examination of the spatial distribution pattern of δ_I

Table 1 Input parameters used in Eqs. (6) and (10) to calculate lake-specific input water isotope composition

	2000	2005	Reference
h(%)	65.7	69.7	Environment Canada (2004)
T (°C)	12.0	11.4	
α^* (^{18}O, ^{2}H)	1.0105, 1.0943	1.0106,1.0951	Horita and Wesolowski (1994)
ε^* (^{18}O, ^{2}H) ‰	10.5, 94.3	10.6, 95.1	
ε_K (^{18}O, ^{2}H) ‰	4.9, 4.3	4.3, 3.7	Gonfiantini (1986)
δ_A (^{18}O, ^{2}H) ‰	-26.6, -205	-26.6, -205	Index lake method (based on average h and T of 67.7% and 11.7 °C)

Thaw season temperature (T) and relative humidity (h) are based on evaporation flux-weighted mean daily values measured at Fort Chipewyan, Alberta. Note, all parameters are in decimal expression when applied in Eq. (6). For example, 60% = 0.6 for h values; -18‰ = -0.018 for δ values and 10‰ = 0.01 for ε values.

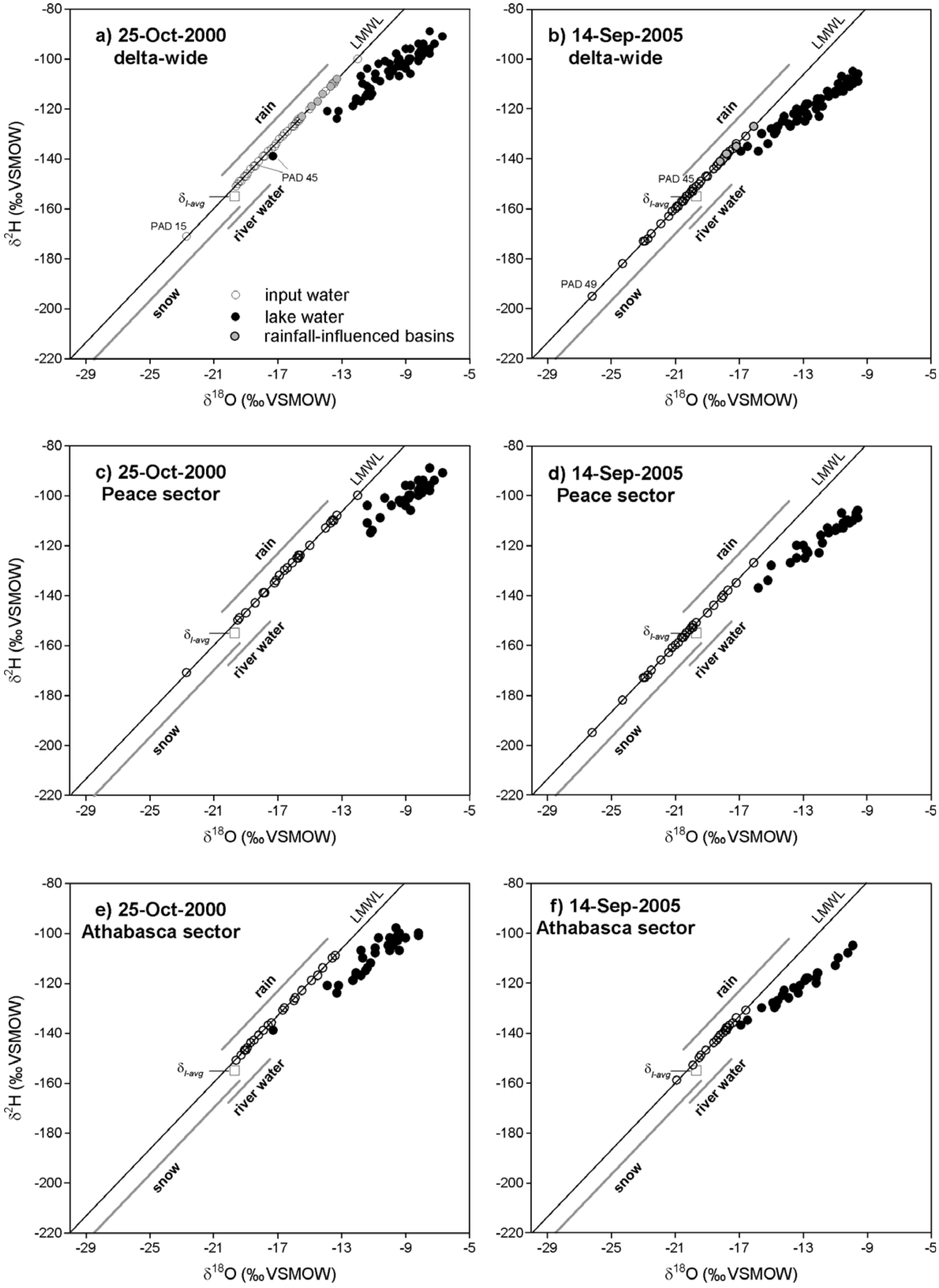

Figure 5 δ^{18}O-δ^{2}H diagrams with measured lake water isotope compositions and calculated lake-specific input water isotope compositions using Eqs. (6) and (10) (see Table 1 for input parameters). (a) and (b) Delta-wide results for 25-October-2000 and 14-September-2005. The gray circles are for previously identified rainfall-influenced basins (Wolfe et al., 2007). Labelled sites are mentioned in the text. (c) and (d) Peace sector results for 25-October-2000 and 14-September-2005. (e) and (f) Athabasca sector results for 25-October-2000 and 14-September-2005. Note that drier and more variable conditions in 2000 led to greater overall enrichment and scatter in isotopic composition of lake waters. Isotopic ranges for snow, rain and river water are also shown (gray lines), as well as the average isotope composition ($\delta_{\text{I-avg}}$) for the three water sources (Falcone, 2007). Note that these ranges lie on or are close to the LMWL, but are offset for graphic purposes only.

reveals that lakes with the lowest δ_I values for the two sample collection years are all located in the northern Peace sector where closed-drainage basins are dominant and mature forest covers the landscape (Fig. 5c and d). Topographic relief from outcropping bedrock knolls in the northeast likely also captures snowdrifts. In comparison, δ_I results indicate that no basins from the southern Athabasca sector were strongly influenced by snowmelt in either 2000 or 2005 (Fig. 5e and f), consistent with the wetland-covered, active floodplain landscape containing mainly restricted-drainage basins in this portion of the PAD. Instead, a small number of lakes in the Athabasca sector may have received river floodwaters from an ice-jam on the Athabasca River during the spring of 2005, which likely also accounts for the narrower range of δ_I for these basins (of composition similar to the Athabasca River) compared to 2000 (Fig. 5e and f).

There are also notable differences in the distribution patterns of δ_I for Peace sector basins between collection years 2000 and 2005 (Fig. 5c and d). The average isotopic composition of the three main sources of water (i.e., rain, snow and river water) is $-19.7‰$ for $\delta^{18}O$ and $-155‰$ for δ^2H (Falcone, 2007; also shown in Fig. 5). For lakes sampled in 2000, 30 of 31 (97%) have δ_I values isotopically-enriched relative to this reference point (Fig. 5c). In contrast, only 9 of 30 (30%) lakes in 2005 have δ_I values that are higher compared to this reference point (Fig. 5d). As shown in Fig. 2, the winter of 1999–2000 was relatively warm with less-than-average snowfall (especially from January to April) and summer 2000 rainfall was near-normal. In contrast, snowfall in 2004–2005 was much higher than 1999–2000, although subsequent total summer rainfall was similar. Consistent with these meteorological conditions, our δ_I results indicate far greater relative importance of snowmelt to Peace sector lake water budgets in 2005 compared to 2000 (Fig. 5c and d).

Conclusions

A coupled isotope tracer method is described to characterize the dual-isotope composition of input water (δ_I) to individual lake basins. Estimation of δ_I is based on the assumption that input waters fall along a definable LMWL in $\delta^{18}O$-δ^2H space and that mass and isotopes are conserved during evaporation, which is described by the linear resistance model of Craig and Gordon (1965). We demonstrate the viability of this technique by applying it in the Peace-Athabasca Delta to show that lakes receive input derived from varying proportions of rainfall, snowmelt and river water, which can be systematically related to lake catchment characteristics and hydrological settings, as well as antecedent meteorological conditions.

Our case study is drawn from late thaw-season lake water sample collections during two years of contrasting hydrological and meteorological conditions. Lake sat this time are likely to closely approach hydrologic and isotopic steady-state, although a similar approach could be applied under non-steady-state conditions. The method has the added advantage of potentially combining the identification of δ_I with basin evaporation-to-inflow ratio assessments over large, hydrologically complex landscapes like deltas and floodplains from single point-in-time measurements, where widespread time-series isotopic sampling of individual basins is not feasible.

Acknowledgements

We would like to thank the staff of Wood Buffalo National Park for logistical support and Dörte Köster, Trish Stadnyk, Niloshini Sinnatamby and Johan Wiklund for field assistance. We are also grateful to Natalie St. Amour for compiling meteorological data for the PAD and the Environmental Isotope Laboratory — University of Waterloo for conducting stable isotope analysis of water samples. Funding for this research was provided by the British Columbia Hydro and Power Authority, the Polar Continental Shelf Project, the Northern Scientific Training Program of Indian and Northern Affairs Canada, the Canada Foundation for Innovation and Ontario Innovation Trust, and the Northern Research Chair Program of the Natural Sciences and Engineering Research Council of Canada. We especially thank the reviewers and the editor for their helpful comments.

Appendix I

Water isotope composition of PAD lakes from two regional sampling campaigns

Lake	Date	$\delta^{18}O_L$	δ^2H_L	Lake	Date	$\delta^{18}O_L$	δ^2H_L
Athabasca sector							
PAD 19	25-October-2000	−8.2	−100	PAD 19			
PAD 20	25-October-2000	−11.5	−115	PAD 20	14-September-2005	−12.2	−120
PAD 21	25-October-2000	−9.4	−107	PAD 21			
PAD 22	25-October-2000	−9.8	−105	PAD 22	14-September-2005	−13.3	−124
PAD 23	25-October-2000	−10.1	−105	PAD 23	14-September-2005	−11.0	−113
PAD 24	25-October-2000	−10.0	−107	PAD 24	14-September-2005	−13.9	−126
PAD 25	25-October-2000	−13.3	−124	PAD 25	14-September-2005	−14.8	−130
PAD 26	25-October-2000	−12.3	−119	PAD 26	14-September-2005	−14.9	−128
PAD 27	25-October-2000	−9.5	−101	PAD 27	14-September-2005	−9.9	−105

Appendix I (*continued*)

Lake	Date	$\delta^{18}O_L$	$\delta^{2}H_L$	Lake	Date	$\delta^{18}O_L$	$\delta^{2}H_L$
PAD 28	25-October-2000	−9.4	−100	PAD 28	14-September-2005	−12.1	−116
PAD 29	25-October-2000	−12.1	−117	PAD 29	14-September-2005	−12.7	−118
PAD 30	25-October-2000	−11.2	−112	PAD 30	14-September-2005	−14.6	−127
PAD 31	25-October-2000	−11.4	−114	PAD 31	14-September-2005	−15.6	−130
PAD 32	25-October-2000	−9.0	−102	PAD 32	14-September-2005	−12.2	−118
PAD 33	25-October-2000	−12.1	−116	PAD 33	14-September-2005	−14.2	−123
PAD 34	25-October-2000	−11.8	−117	PAD 34	14-September-2005	−12.8	−118
PAD 35	25-October-2000	−9.0	−102	PAD 35	14-September-2005	−10.2	−108
PAD 36	25-October-2000	−9.5	−103	PAD 36			
PAD 37	25-October-2000	−8.2	−101	PAD 37	14-September-2005	−10.8	−110
PAD 38	25-October-2000	−13.2	−121	PAD 38	14-September-2005	−14.3	−125
PAD 39	25-October-2000	−9.6	−98	PAD 39	14-September-2005	−10.8	−110
PAD 40	25-October-2000	−10.9	−108	PAD 40	14-September-2005	−13.6	−122
PAD 41	25-October-2000	−11.7	−110	PAD 41	14-September-2005	−12.9	−119
PAD 42	25-October-2000	−11.8	−107	PAD 42	14-September-2005	−13.2	−121
PAD 43	25-October-2000	−9.4	−100	PAD 43			
PAD 44	25-October-2000	−10.9	−106	PAD 44			
PAD 45	25-October-2000	−17.3	−139	PAD 45	14-September-2005	−16.9	−137
PAD 46	25-October-2000	−11.7	−110	PAD 46	14-September-2005	−16.5	−135
PAD 47	25-October-2000	−10.7	−102	PAD 47			
PAD 59	25-October-2000	−10.0	−102	PAD 59			
PAD 62	25-October-2000	−13.9	−121	PAD 62	14-September-2005	−14.7	−129
Peace sector							
PAD 1	25-October-2000	−8.0	−94	PAD 1	14-September-2005	−12.8	−122
PAD 2	25-October-2000	−7.9	−98	PAD 2	14-September-2005	−11.0	−114
PAD 3	25-October-2000	−8.0	−99	PAD 3	14-September-2005	−10.7	−113
PAD 4	25-October-2000	−7.2	−94	PAD 4	14-September-2005	−10.3	−110
PAD 5	25-October-2000	−6.7	−91	PAD 5	14-September-2005	−10.1	−111
PAD 6	25-October-2000	−7.8	−96	PAD 6	14-September-2005	−10.2	−110
PAD 7	25-October-2000	−8.2	−98	PAD 7	14-September-2005	−10.5	−111
PAD 8	25-October-2000	−11.4	−111	PAD 8	14-September-2005	−13.8	−127
PAD 9	25-October-2000	−7.5	−96	PAD 9	14-September-2005	−10.9	−113
PAD 10	25-October-2000	−9.9	−104	PAD 10			
PAD 11	25-October-2000	−8.7	−100	PAD 11	14-September-2005	−11.5	−113
PAD 12	25-October-2000	−8.1	−97	PAD 12	14-September-2005	−9.9	−110
PAD 13	25-October-2000	−8.8	−100	PAD 13	14-September-2005	−12.7	−123
PAD 14	25-October-2000	−9.1	−102	PAD 14	14-September-2005	−10.9	−113
PAD 15	25-October-2000	−8.7	−106	PAD 15	14-September-2005	−12.9	−125
PAD 16	25-October-2000	−8.8	−101	PAD 16	14-September-2005	−11.9	−116
PAD 17	25-October-2000	−9.4	−102	PAD 17	14-September-2005	−11.8	−119
PAD 18	25-October-2000	−8.2	−100	PAD 18	14-September-2005	−9.6	−106
PAD 48	25-October-2000	−11.1	−114	PAD 48	14-September-2005	−13.4	−125
PAD 49	25-October-2000	−9.0	−104	PAD 49	14-September-2005	−12.0	−123
PAD 50	25-October-2000	−7.5	−98	PAD 50	14-September-2005	−9.6	−109
PAD 51	25-October-2000	−8.7	−96	PAD 51	14-September-2005	−13.0	−120
PAD 52	25-October-2000	−9.3	−103	PAD 52	14-September-2005	−10.5	−113
PAD 53	25-October-2000	−11.4	−104	PAD 53	14-September-2005	−10.6	−107
PAD 54	25-October-2000	−11.2	−115	PAD 54	14-September-2005	−15.2	−134
PAD 55	25-October-2000	−10.6	−109	PAD 55			
PAD 56	25-October-2000	−8.8	−101	PAD 56	14-September-2005	−9.7	−107
PAD 57	25-October-2000	−7.5	−89	PAD 57	14-September-2005	−9.7	−108
PAD 58	25-October-2000	−8.2	−94	PAD 58	14-September-2005	−11.4	−115
PAD 60	25-October-2000	−9.0	−96	PAD 60			
PAD 61	25-October-2000	−10.3	−101	PAD 61	14-September-2005	−13.4	−120
PAD 13 A				PAD 13 A	14-September-2005	−15.8	−137

Appendix II

Water isotope composition of PAD 54 and PAD 37 between 2000 and 2006

PAD 54			PAD 37		
Date	$\delta^{18}O_L$	$\delta^{2}H_L$	Date	$\delta^{18}O_L$	$\delta^{2}H_L$
25-October-2000	−11.2	−115	25-October-2000	−8.2	−101
04-March-2001	−11.3	−121	10-June-2001	−8.0	−99
04-June-2001	−11.0	−117	27-August-2001	−6.5	−91
27-August-2001	−12.0	−121	03-June-2002	−9.4	−106
03-June-2002	−12.9	−123	21-September-2002	−9.0	−100
24-July-2002	−11.8	−118	02-May-2004	−16.9	−143
24-August-2002	−11.6	−116	17-July-2004	−4.9	−89
21-September-2002	−11.7	−116	19-September-2004	−15.7	−124
01-May-2003	−20.2	−157	16-May-2005	−15.5	−139
22-May-2003	−18.5	−151	18-July-2005	−7.7	−99
04-June-2003	−17.9	−148	14-September-2005	−10.8	−110
16-June-2003	−17.8	−148	01-June-2006	−6.4	−92
03-July-2003	−17.3	−146	20-July-2006	−6.1	−85
15-July-2003	−16.8	−146	01-September-2006	−7.0	−89
23-September-2003	−15.6	−137			
02-May-2004	−15.4	−138			
11-June-2004	−15.4	−139			
16-July-2004	−14.5	−128			
28-July-2004	−14.3	−133			
17-September-2004	−13.5	−130			
19-September-2004	−13.5	−129			
12-March-2005	−15.9	−148			
16-May-2005	−17.0	−145			
18-July-2005	−15.7	−136			
28-July-2005	−15.6	−137			
14-September-2005	−15.2	−134			
01-June-2006	−14.5	−133			
20-July-2006	−13.3	−126			
01-September-2006	−12.7	−124			

References

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

Birks, S.J., Edwards, T.W.D, Gibson, J.J., Drimmie, R.J., Michel, F.A., 2004. Canadian Network for Isotopes in Precipitation. (<http://www.science.uwaterloo.ca/~twdedwar/cnip/cniphome.html>) (accessed 06.07.07.).

Coleman, M.L., Shepherd, T.J., Durham, J.J., Rouse, J.E., Moore, G.R., 1982. Reduction of water with zinc for hydrogen isotope analysis. Analytical Chemistry 54, 993–995.

Coplen, T.B., 1996. New guidelines for reporting stable hydrogen, carbon, and oxygen isotope-ratio data. Geochimica et Cosmochimica 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 the marine atmosphere. In: Tongiorgi, E. (Ed.), Stable Isotope in Oceanographic Studies and Paleotemperatures. Laboratorio di Geologia Nucleare, Pisa, Italy, pp. 9–130.

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

Drimmie, R.J., Heemskerk, R.A., 1993. Water ^{18}O by CO_2 equilibration. Technical Procedure 13.0, Rev. 02. Environmental Isotope Laboratory: Department of Earth Sciences, University of Waterloo, 11 p.

Edwards, T.W.D., Wolfe, B.B., Gibson, J.J., Hammarlund, D., 2004. Use of water isotope tracers in high-latitude hydrology and paleohydrology. In: Pienitz, R., Douglas, M.S.V., Smol, J.P. (Eds.), Long-term Environmental Change in Arctic and Antarctic Lakes, Developments in Paleoenvironmental Research, vol. 8. Springer, Dordrecht, pp. 187–207.

Environment Canada, 2004. Climate Data Online (<http://www.climate.weatheroffice.ec.gc.ca/climateData/hourlydata_e.html>) (accessed 08.04.07.).

Epstein, S., Mayeda, T.K., 1953. Variation of O^{18} content of waters from natural sources. Geochimica et Cosmochimica Acta 4, 213–224.

Falcone, M., 2007. Assessing hydrological processes controlling the water balance of lakes in the Peace-Athabasca Delta, Alberta, Canada, using water isotope tracers. MSc thesis, University of Waterloo, Waterloo, 190 p.

Ferhi, A., Bariac, T., Letolle, R., 1983. $^{18}O/^{16}O$ ratios in leaf water and cellulose of aquatic and terrestrial plants. In: IAEA (Ed.), Palaeoclimates and Palaeowaters: a Collection of Environmental Isotope Studies. International Atomic Energy Agency, Vienna, pp. 85–99.

Gat, J.R., 1996. Oxygen and hydrogen isotopes in the hydrological cycle. Annual Reviews in Earth and Planetary Sciences 24, 225–262.

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

Gat, J.R., Matsui, E., 1991. Atmospheric water balance in the Amazon Basin: an isotopic evapotranspiration model. Journal of Geophysical Research 96, 13179–13188.

Gat, J.R., Bowser, C.J., Kendall, C., 1994. The contribution of evaporation from the Great Lakes to the continental atmosphere: estimate based on stable isotope data. Geophysical Research Letters 21, 557–560.

Gibson, J.J., 2002. A new conceptual model for predicting isotope enrichment of lakes in seasonal climates. PAGES News 10, 10–11.

Gibson, J.J., Edwards, T.W.D., 2002. Regional water balance trends and evaporative-transpiration partitioning from a stable isotope survey of lakes in northern Canada. Global Biogeochemical Cycles 16. doi:10.1029/2001GB001839.

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 Hydrology 24, 79–94.

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

Gibson, J.J., Birks, S.J., Edwards, T.W.D., in press. Global prediction of δ_A and $\delta^{18}O$-$\delta^{2}H$ evaporation slopes for lakes and soil water accounting for seasonality. Global Biogeochemical Cycles.

Gonfiantini, R., 1986. Environmental isotopes in lake studies. In: Fritz, P., Fontes, J.-C. (Eds.), Handbook of Environmental Isotope Geochemistry, The Terrestrial Environment, vol. 2. Elsevier, New York, pp. 113–168.

Hall, R.I., Wolfe, B.B., Edwards, T.W.D., 2004. A Multi-century Flood, Climatic, and Ecological History of the Peace-Athabasca Delta, Northern Alberta, Canada. Final Report. Published by BC Hydro, 163 p + Appendices.

Horita, J., Wesolowski, D., 1994. Liquid–vapour fractionation of oxygen and hydrogen isotopes of water from the freezing to the critical temperature. Geochimica et Cosmochimica Acta 58, 3425–3437.

Jacob, H., Sonntag, C., 1991. An 8-year record of the seasonal variation of ^{2}H and ^{18}O in atmospheric water vapour and precipitation at Heidelberg, Germany. Tellus 43B, 291–300.

Mayr, C., Lücke, A., Stichler, W., Trimborn, P., Ercolano, B., Oliva, G., Ohlendorf, C., Soto, J., Fey, M., Haberzettl, T., Janssen, S., Schäbitz, F., Schleser, G.H., Wille, M., Zolitschka, B., 2007. Precipitation origin and evaporation of lakes in semi-arid Patagonia (Argentina) inferred from stable isotopes ($\delta^{18}O$, $\delta^{2}H$). Journal of Hydrology 334, 53–63.

PADPG, 1973. Peace-Athabasca Delta Project, technical report and appendices, vol 1: hydrological investigations, vol 2: ecological investigations. Peace-Athabasca Delta Project Group, Delta Implementation Committee, Governments of Alberta, Saskatchewan and Canada; 176 p.

Peters, D.L., 2003. Controls on the persistence of water on perched basins of the Peace-Athabasca Delta, northern Canada. PhD thesis, Trent University, Peterborough, 194 p.

Peters, D.L., Prowse, T.D., Marsh, P., Lafleur, P.M., Buttle, J.M., 2006. Persistence of water within perched basins of the Peace-Athabasca Delta, northern Canada. Wetlands Ecology and Management 14, 221–243.

Rozanski, K., Araguas-Araguas, L., Gonfiantini, R., 1993. Isotopic patterns in modern global precipitation. In: Swart, P.K., Lohmann, K.C., McKenzie, J., Savin, S. (Eds.), Climate Change in Continental Isotopic Records. American Geophysical Union Geophysical Monograph 78, pp. 1–36.

Vallet-Coulomb, C., Gasse, F., Robinson, L., Ferry, L., 2006. Simulation of the water and isotopic balance of a closed tropical lake at a daily time step (Lake Ihotry, South-West of Madagascar). Journal of Geochemical Exploration 88, 153–156.

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

Wolfe, B.B., Karst-Riddoch, T.L., Hall, R.I., Edwards, T.W.D., English, M.C., Palmini, R., McGowan, S., Vardy, S.R., 2007. Classification of hydrologic regimes of northern floodplain basins (Peace-Athabasca Delta, Canada) from analysis of stable isotopes ($\delta^{18}O$, $\delta^{2}H$) and water chemistry. Hydrological Processes 21, 151–168.

Yakir, D., Sternberg, L.S.L., 2000. The use of stable isotopes to study ecosystem gas exchange. Oecologia 123, 297–311.

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