7. USE OF WATER ISOTOPE TRACERS IN HIGH-LATITUDE HYDROLOGY AND PALEOHYDROLOGY

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Introduction

The stable isotopes ^2H and ^{18}O are highly valuable tracers for investigating present and past hydrology and hydroclimatology because of the existence of robust physically-based understanding of isotopic partitioning in the global water cycle (e.g., see Rozanski et al. (1993), Araguás-Araguás et al. (2000)). Although applicable in any

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geographic location, water isotope data are particularly effective for characterizing hydrology and hydroclimatology in remote regions, especially at high latitudes, providing information about runoff generation processes and surface water balance that may be difficult or impossible to acquire using other methods (e.g., see Burgman et al. (1987), Wolfe and Edwards (1997), Maric (2003), Gibson (2001, 2002), Gibson and Edwards (2002), Gibson et al. (1993, 2002), Leng and Anderson (2003)). Stratigraphic records of water isotope data in lake sediments can also be used to reconstruct or constrain past hydrology and hydroclimatology at high latitudes, relying on the same mechanistic principles and benefiting from the framework provided by knowledge and inference about global paleo-isotope fields (e.g., see MacDonald et al. (1993, this volume), Edwards et al. (1996), Wolfe et al. (1996, 2000, 2003), Sauer et al. (2001b), Shemesh et al. (2001), Anderson et al. (2001), Hammarlund et al. (2002), Hu and Shemesh (2003)).

This chapter briefly reviews key aspects of the use of stable water isotope tracers in hydrological and paleohydrological investigations applicable to the study of long-term environmental change in Arctic and Antarctic lakes, illustrated by selected examples from recent investigations in northern Canada and Sweden (see Figure 1 for locations of sites mentioned in the text). The aim is to provide the reader with basic working knowledge of the major controls on the isotopic partitioning of water in high-latitude catchments and the nature of the information that can be gleaned from records of the isotopic composition of paleowaters obtained from lake sediment archives in such environments.

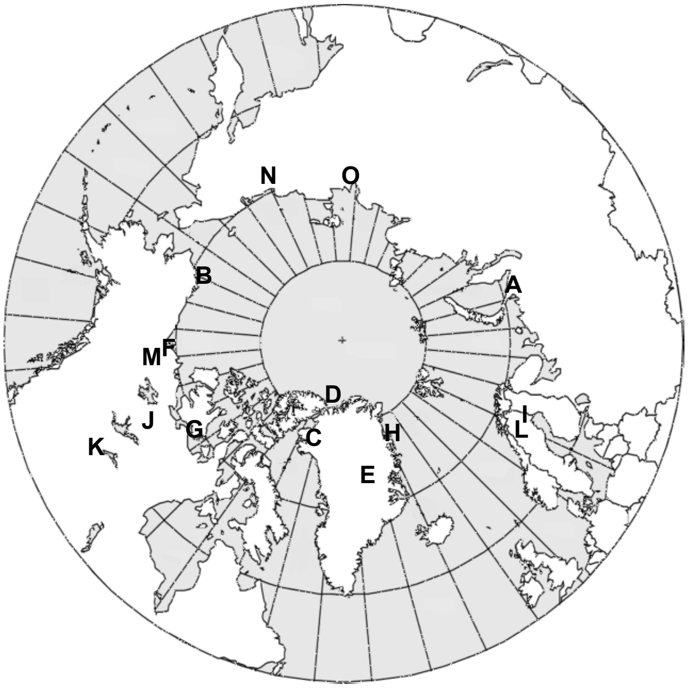

Figure 1. Map of the circumpolar North, showing locations noted in subsequent figures and text. Amderma (**A**), Barrow (**B**), Thule (**C**), Alert (**D**), GRIP (**E**), Inuvik (**F**), Cambridge Bay (**G**), Danmarkshavn (**H**), Naimakka (**I**), Exeter Lake (**J**), Peace-Athabasca Delta (**K**), Lake Tibetanus (**L**), Mackenzie River Delta (**M**), Kolyma River Delta (**N**), Lena River Delta (**O**).

Isotopic labelling in the hydrological cycle

Background

Isotopic partitioning (fractionation) of water in the hydrological cycle arises because of differences in the behaviour of water molecules containing various combinations of the naturally occurring stable isotopes of hydrogen (^{1}H, ^{2}H) and oxygen (^{16}O, ^{17}O, ^{18}O). Most isotope hydrology studies are based on variations in the relative abundances of the two rare heavy isotopic species (isotopomers) containing a single ^{2}H or ^{18}O atom ($^{1}H^{2}H^{16}O$ and $^{1}H^{1}H^{18}O$) with respect to common light water ($^{1}H^{1}H^{16}O$). Isotopic compositions are expressed conventionally as δ-values, representing deviation in per mil (‰) from the isotopic composition of a specified standard, such that $\delta^{2}H$ or $\delta^{18}O$ = $1000((R_{sample}/R_{standard}) - 1)$, where R refers to the $^{2}H/^{1}H$ or $^{18}O/^{16}O$ ratios in sample and standard, respectively. The most widely used standard in hydrological applications is Vienna Standard Mean Ocean Water (VSMOW), which approximates the bulk isotopic composition of the present-day global ocean reservoir, and hence has $\delta^{2}H$ and $\delta^{18}O$ values both defined to be exactly 0‰. This is a logical datum for hydrological studies since evaporation from the oceans is the fundamental source of global atmospheric moisture, which provides the precipitation input for continental water cycling, and the isotopic composition of the oceans is essentially invariant at human time scales. Studies using limnic carbonates (e.g., sedimentary calcite, mollusc shells, ostracode valves) as oxygen-isotope archives also make use of the Vienna Pee Dee Belemnite (VPDB) standard, which approximates the isotopic composition of modern marine carbonate. $\delta^{18}O$ values expressed relative to the VSMOW and VPDB standards are related such that $(1000 + \delta^{18}O_{sample/VSMOW})/(1000 + \delta^{18}O_{sample/VPDB}) = 1.03086$ (Fritz and Fontes (1980) or Clark and Fritz (1997) are recommended for further information about isotopic standards and nomenclature).

Isotopic labelling of precipitation – Global and Local Meteoric Water Lines

The distribution of isotopes in precipitation and surface waters is characterized by the existence of strongly linear relations between $\delta^{2}H$ and $\delta^{18}O$ values over a broad range of spatial and temporal scales, reflecting systematic mass-dependent partitioning of water molecules in the hydrologic cycle. The most striking expression of the coupling between hydrogen- and oxygen-isotope labelling of water is the Global Meteoric Water Line (Craig 1961), described by $\delta^{2}H = 8\delta^{18}O + 10$, which closely approximates the observed relation between mean annual amount-weighted precipitation $\delta^{2}H$ and $\delta^{18}O$ world-wide (Figure 2a). The existence of this fundamental baseline is consistent with the notion that global atmospheric moisture arises primarily from a well-mixed source (i.e., the subtropic ocean surface) and undergoes progressive rain-out of mass and heavy isotopes during subsequent poleward atmospheric transport. The linearity and slope of the GMWL reflect the pervasive influence of temperature-dependent equilibrium partitioning of the two heavy isotopic species between atmospheric vapour and condensing precipitation. This can be modelled conceptually as a multi-step Rayleigh-type open-system distillation process, leading to the combination of low amount and depleted heavy-isotope content (low δ values) of precipitation at high latitudes. This important coupling of mass and isotope depletion is well-illustrated by comparison of the global fields of precipitable moisture and precipitation δ values (e.g., see Figure 2 of Birks et al. (2002)), and is also reflected indirectly in the spatial relations that are observed between precipitation δ values and air temperatures at mid- to high latitudes, often used as the basis for paleotemperature reconstruction from the isotopic composition of paleoprecipitation stored in glacier ice or ground water (e.g., see Rozanski et al. (1993, 1997), Schotterer et al. (1996)).

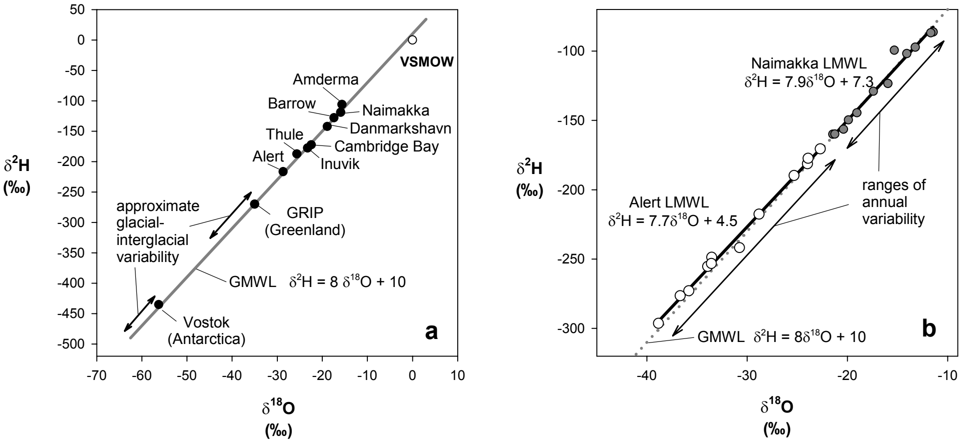

Figure 2. a) Isotopic labelling of weighted mean annual precipitation (δ_P) collected at selected high-latitude stations, based on data compiled by Birks et al. (2002) from the Global Network for Isotopes in Precipitation (GNIP), operated jointly by the International Atomic Energy Agency and the World Meteorological Organization (see IAEA/WMO (2001)). Modern mean annual composition of precipitation for ice-core sites from Greenland and Antarctica are also indicated, with arrows showing the approximate ranges of variability between average glacial and interglacial conditions (ice-core data archived at the World Data Centre for Paleoclimatology, Boulder, CO, USA: http://www.ngdc.noaa.gov/paleo/data.html). Vienna Standard Mean Ocean Water (VSMOW) is also shown, offset slightly below the Global Meteoric Water Line (GMWL). *b)* Isotopic labelling of mean composite monthly precipitation at Alert (Canada) and Naimakka (Sweden), based on data compiled by Birks et al. (2002). Data from both stations are strongly localized along the GMWL (dotted line), defining individual local meteoric water lines (LMWLs) having slopes slightly less than 8 and large ranges of monthly variability. Maximum and minimum mean composite monthly δ^2H and δ^{18}O values at both stations occur, respectively, in July and January. Note that the GMWL commonly provides an excellent proxy for the LMWL in the absence of direct isotopic characterization of local precipitation, especially at high latitudes.

Amount-weighted δ^2H and δ^{18}O values of precipitation received over the year at individual sites also commonly plot in strongly linear clusters in δ^2H-δ^{18}O space close to the GMWL, and best-fit Local Meteoric Water Lines (LMWLs) drawn through these clusters can provide highly useful characterization of local isotopic input functions for hydrological studies (Figure 2b). Substantial seasonal variability is typical along LMWLs at mid- to high latitudes, especially in cold regions, with winter precipitation generally strongly depleted and more variable in heavy-isotope content compared to that received during the summer season.

Additional isotopic characterization of local precipitation is provided by the derived "deuterium excess" parameter (*d*-excess = $\delta^2H - 8\delta^{18}O$), which gives an indication of location in δ^2H-$\delta^{18}O$ space relative to the GMWL reference (*d*-excess$_{GMWL}$ = +10 ‰). This characteristic feature of the distribution of isotopes in precipitation reflects the combined influence of equilibrium and kinetic isotopic partitioning as atmospheric moisture is generated by evaporation from the sea surface, which tends to be conserved during subsequent transport and distillation. Local precipitation typically has a seasonal cycle marked by lower *d*-excess values in summer, often attributable to slight evaporative isotopic enrichment of falling raindrops, and higher *d*-excess in winter due to kinetic effects associated with ice formation in clouds (Merlivat and Jouzel 1979, Jouzel and Merlivat 1984).

Isotopic enrichment of surface waters - Local Evaporation Lines

Strong coupling of hydrogen- and oxygen-isotope variations is also evident from the development of characteristic heavy-isotope buildup in surface waters undergoing evaporation. The isotopic signatures of neighbouring water bodies receiving input of similar isotopic composition typically lie along more-or-less well-defined linear arrays in δ^2H-$\delta^{18}O$ space, termed Local Evaporation Lines (LELs; Figure 3), which deviate from the LMWL along slopes that usually range between 4 and 6, depending on local atmospheric conditions during the thaw season (primarily relative humidity, temperature, and the isotopic composition of ambient moisture). Intersection of the LEL with the LMWL often provides an excellent approximation of the weighted-mean isotopic composition of input waters to a catchment (δ_I), while displacement of a given lake water (δ_L) along the LEL provides an index of water balance, which can be quantified in terms of evaporation:inflow ratio (E/I) via isotope-mass balance considerations, as described for hydrogen or oxygen isotope data by:

$$E/I = (\delta_I - \delta_L)/(\delta_E - \delta_L) \qquad (1)$$

where δ_E represents the isotopic composition of the evaporative flux. δ_E can be evaluated using the well-validated linear resistance model of Craig and Gordon (1965), which accounts for the differing volatilities of the water isotopomers as a combination of mass-dependent variations in equilibrium vapour pressures ("equilibrium effects") plus variations in molecular diffusivities arising from the combination of differing absolute mass and its distribution within water molecules ("kinetic effects").

Assuming negligible resistance to liquid-phase mixing,

$$\delta_E = ((1 + 10^{-3}\varepsilon^*)\,\delta_L - h\delta_A - \varepsilon)/(1 - h + 10^{-3}\varepsilon_K) \qquad (2)$$

where ε^* and ε_K represent the respective equilibrium and kinetic effects, expressed as per mil (‰) separations between the liquid and vapour phases, $\varepsilon = \varepsilon^* + \varepsilon_K$, and h is the atmospheric relative humidity normalized to the saturation vapour pressure at the temperature of the air-water interface. ε^* values for both hydrogen and oxygen can be estimated from the empirical relations derived by Horita and Wesolowski (1994). For δ^2H, $\varepsilon^* = 1158.8(T^3/10^9) - 1620.1(T^2/10^6) + 794.84(T/10^3) - 161.04 + 2.9992(10^9/T^3)$ and for $\delta^{18}O$, $\varepsilon^* = -7.685 + 6.7123(10^3/T) - 1.6664(10^6/T^2) + 0.3504(10^9/T^3)$, where T is the interface temperature (in K) and ε_K values for typical natural conditions can be accurately approximated as a function of relative humidity deficit (Gonfiantini 1986, Araguás-Araguás et al. 2000). For δ^2H, $\varepsilon_K = 12.5(1 - h)$ and for $\delta^{18}O$, $\varepsilon_K = 14.2(1 - h)$.

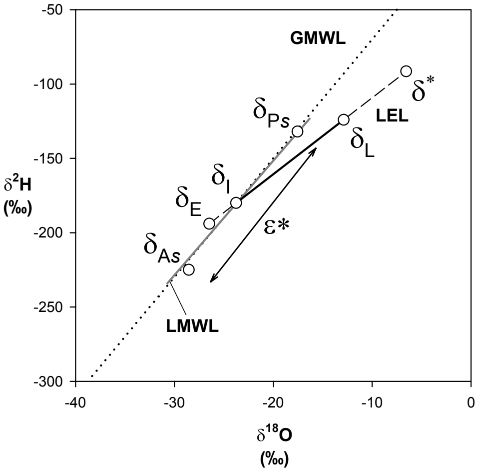

Figure 3. Isotope labelling of water balance components, consistent with a steady-state E/I ratio of about 0.8 for a hypothetical high-latitude lake, as described by equations (1)-(4) in the text (based on mean thaw season relative humidity and air temperature, respectively, of about 0.7 and 10˚C). The isotopic composition of the lake water at the time of sampling (δ_L) is offset from the input water composition (δ_I) along a local evaporation line (LEL) having a slope of about 5. Although impossible to sample directly, mass conservation dictates that the evaporative flux from the lake (δ_E) lies on the extension of the LEL to the left of the LMWL. Evaporation and exchange occur in association with ambient atmospheric moisture during the summer evaporation season (δ_{As}), weighted for evaporation flux, which is in approximate isotopic equilibrium (ε^*) with the mean isotopic composition of summer precipitation (δ_{Ps}). Also shown is the limiting isotopic composition (δ^*), which is the maximum enrichment attainable under local atmospheric conditions as a water body approaches dryness, providing an indication of potential extreme non-steady-state behaviour. See Gibson and Edwards (2002) for additional discussion.

Equations (1) and (2) can be applied readily to assess instantaneous lake water balances in systems approximating hydrologic steady-state (e.g., see Gibson (2001), Gibson and Edwards (2002)) and a similar approach has been used for reconstructing or constraining past hydrologic variability from isotopic paleorecords (e.g., see Edwards and Fritz (1988), Edwards and McAndrews (1989), MacDonald et al. (1993), Edwards et al. (1996), Wolfe et al. (2001a), Hammarlund et al. (2003)). Substitution and rearrangement of equations (1) and (2) also yields:

$$E/I = [(1 - h + 10^{-3}\varepsilon_K)/(h - 10^{-3}\varepsilon)][(\delta_L - \delta_I)/(\delta^* - \delta_L)] \tag{3}$$

and

$$\delta^* = (h\delta_A + \varepsilon)/(h - 10^{-3}\varepsilon) \tag{4}$$

where δ^* is the limiting isotopic enrichment attainable when a water body evaporates to near-zero volume, and has hence progressed well beyond hydrologic steady-state (see Figure 3). Modelling of the non-steady-state isotopic enrichment leading to δ^* is also possible, affording information about mass and isotope fluxes under transient conditions (e.g., see Gibson (2002)).

Isotope hydrology at high latitudes

Runoff generation and evaporation

Hydrological processes at high latitudes are almost universally characterized by intense climatic and isotopic seasonality. Winter is typically marked by significant reduction or even total suspension of runoff generation because of deep frost penetration into soils and the storage of highly ^{2}H- and ^{18}O-depleted seasonal precipitation in the frozen state within the overlying snowpack. In spite of dry atmospheric conditions, open-water evaporation is strongly suppressed by development of ice-cover on lakes and all but the most rapidly flowing streams, though significant quantities of moisture may be recycled to the atmosphere via sublimation directly from the snow.

Warming in spring results in rapid melting of the snowpack, introducing pulses of isotopically-depleted runoff into streams and lakes, while still-frozen soils inhibit infiltration. Lake-ice begins to thin, with early break-up on small, shallow water bodies leading to immediate onset of seasonal evaporation, while persistence of ice-cover for days or weeks on large, deep lakes may suppress this process well into the thaw season. The isotopic expression of these two primary hydrological processes, snowmelt dilution and evaporative enrichment, is often strongly evident at both lake and catchment scales, as shown for data from a selection of small upland lakes in the continental Low Arctic of northern Canada (Figure 4).

The isotopic evolution of a given lake during any particular thaw season is highly variable, depending on the fluxes and isotopic compositions of inputs (surface and subsurface runoff, inflow from upstream water bodies, precipitation falling directly upon the lake) and outputs (surface and subsurface outflow, evaporative flux), as well as the storage capacity within the lake itself, which acts to buffer seasonal variation (note data for Yamba Lake in Figure 4b). Annual thawing of the active layer in permafrost terrain also serves to suppress seasonal heavy-isotope buildup in small lakes by sustaining input of unevaporated precipitation recharged during the latter part of the prior thaw season (e.g., see Bursey et al. 1991).

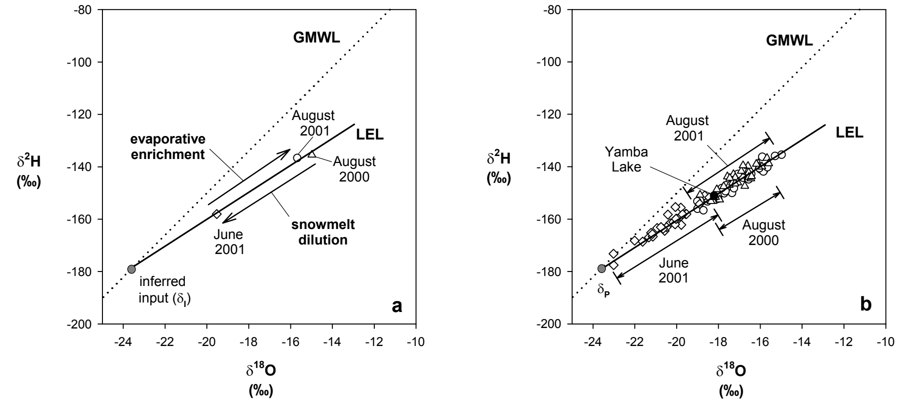

Figure 4. a). Changing isotopic composition of a small upland tundra lake in the west Exeter Lake catchment (64.7˚N; 111.1˚W), a headwater tributary of the Coppermine River, Canada, sampled on three occasions, immediately prior to freeze-up (late August 2000 and 2001) and during spring melt (mid-June 2001), reported by Maric (2003). The plot clearly shows the seasonal cycle produced by the opposing effects of snowmelt dilution and evaporative enrichment. Pronounced enrichment attained during summer 2000, conserved under winter ice-cover, was strongly offset by isotopically depleted snowmelt runoff in spring 2001. Subsequent accumulated evaporative enrichment during summer 2001 almost exactly countered this effect, returning the lake to a composition very similar to that of the previous year. A best-fit line through these data points provides a good approximation of the LEL and the weighted mean isotopic composition of input waters (δ_I). *b)* Isotopic data from 30 small tundra lakes in the west Exeter Lake catchment (including the lake in Figure 3a), sampled on the same three occasions, showing pronounced clustering along the inferred LEL and the equivalent signals of snowmelt dilution and evaporative enrichment at catchment scale. The existence of limited scatter about the LEL reflects the occurrence of slightly differing evaporation lines for each lake, arising because of minor variations in snowmelt isotopic composition and varying contributions from rainfall and ground water runoff, but generally also confirms that the LEL-GMWL intersection is a reasonable proxy for the isotopic composition of local weighted mean annual precipitation (i.e., $\delta_I \approx \delta_P$). Also shown for comparison is the isotopic composition of outflow from Yamba Lake, a large lake downstream, which exhibited no measureable variation between the three sampling episodes, reflecting the damping of seasonal fluctuations because of its large volume (and hence long hydraulic residence time).

Although most upland lakes exhibit systematic seasonal cycles of snowmelt depletion and progressive evaporative enrichment, much less predictable seasonal isotopic behaviour may occur in lakes in fluvial and deltaic settings, as in the myriad water bodies along the Mackenzie, Kolyma, Lena and other major northern river systems. As illustrated in Figure 5, episodic high-water events (often induced by high spring runoff, local ice-jam flooding or excess precipitation in upstream reaches of the river) can significantly influence the seasonal isotopic cycling in a lake in such environments by abruptly introducing water of strongly differing composition originating outside of the lake's catchment.

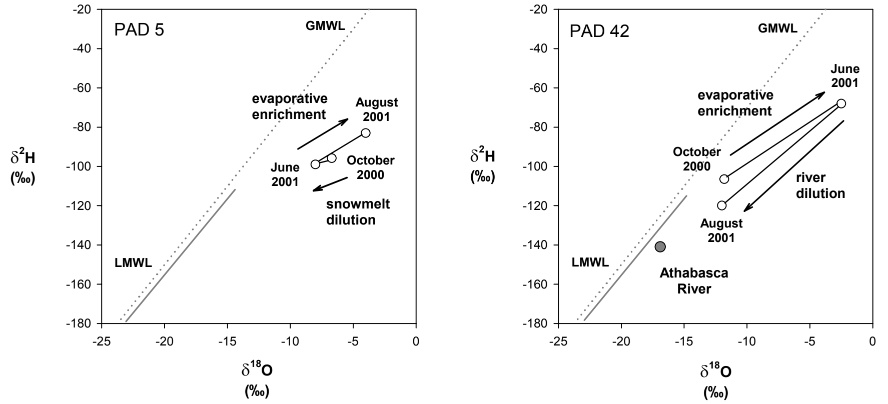

Figure 5. Strongly differing isotopic evolution of two lakes in the Peace-Athabasca Delta, Canada (55.8°N; 111.5°W), sampled immediately prior to freeze-up in October 2000 and again in early June and late August 2001. The relative depletion of PAD 5 between October 2000 and June 2001 reflects the over-riding influence of snowmelt dilution, likely offset somewhat by evaporative enrichment. In contrast, evaporation completely dominated the early-season evolution of PAD 42, at that time a much shallower, desiccating water body, as shown by extreme non-steady-state heavy-isotope enrichment. The subsequent evolution of PAD 5 between June and August 2001 was dominated by evaporative enrichment, yet PAD 42 became prominently depleted as a result of an influx of Athabasca River water during a summer high-water event (after Wolfe et al. (2002)).

Various other factors can also potentially affect the regularity of the seasonal cycle, including variation in the timing and abundance of precipitation, which may force a lake across key hydrologic thresholds. Particularly strong shifts accompany the development of seasonal or perennial closure of a lake, for example, because of the rapid onset of non-steady-state evaporative enrichment (e.g., see Wolfe et al. (1996)). Seasonal variations in water levels can also influence important processes like snowmelt bypassing, which can occur when high water levels lead to rapid flushing of snowmelt under or around ice-cover with minimal mixing, potentially leading to under-representation of isotopically-depleted winter precipitation in a lake's water budget (e.g., see Edwards and McAndrews (1989)). Perhaps most fundamentally, climate variability and change can be expected to affect the isotopic baseline provided by the LMWL through shifts in the range and weighted mean isotopic composition of annual precipitation, possibly in company with temperature change and/or shifting seasonality of precipitation, as well as through alteration of the LEL due to changes in relative humidity and the isotopic composition of atmospheric moisture.

Water isotope tracers in paleolimnology

Development of water isotope records from lake sediments is typically undertaken with the goal of reconstructing the isotopic history of the overlying lake water, as the basis for inferences about paleoclimate and paleohydrology. The majority of such investigations have focused on the use of specific carbonate or organic sediment constituents as oxygen-isotope archives, building upon pioneering studies by Stuiver (1968, 1970) on lacustrine carbonates and Edwards and McAndrews (1989) on algal cellulose. Promising results have also been achieved using other substrates, including biogenic silica as an alternative oxygen-isotope archive (see Shemesh et al. (2001)) and kerogen and lipids as hydrogen-isotope archives (see Krishnamurthy et al. (1995); Sauer et al. (2001a); Huang et al. (2002)). Although advances in the development of hydrogen-isotope archives may eventually afford the opportunity for simultaneous reconstruction of both $\delta^{18}O$ and $\delta^{2}H$ histories, only single-isotope records have yet been produced. As a result, the interpretation of lake water isotopic history is fundamentally dependent on the ability to partition shifts in lake water $\delta^{18}O$ (or $\delta^{2}H$) into MWL- and LEL-parallel components, as shown schematically in Figure 6. Various strategies can be employed to deconvolute lake water $\delta^{18}O$ histories, including use of independently derived records of paleoprecipitation $\delta^{18}O$ to account for variations in input water composition, thus allowing direct resolution of lake-specific changes in water balance (e.g., see Edwards and Fritz (1988); Wolfe et al. (2000, 2001a)). The latter can also be achieved to varying degrees of confidence by comparing inferred lake water $\delta^{18}O$ records from multiple lakes in the same region spanning a range of hydrologic sensitivities, on the assumption that they share a common input water $\delta^{18}O$ history controlled by the isotopic composition of regional precipitation (e.g., see Edwards et al. (1996)). Clues from other indicators in lake sediment records (e.g., $\delta^{13}C$ or $\delta^{15}N$ as nutrient isotope tracers or other biological proxies like diatoms) can also inform or constrain hydrologic interpretations (e.g., see Wolfe et al. (1996), Hammarlund et al. (1997)).

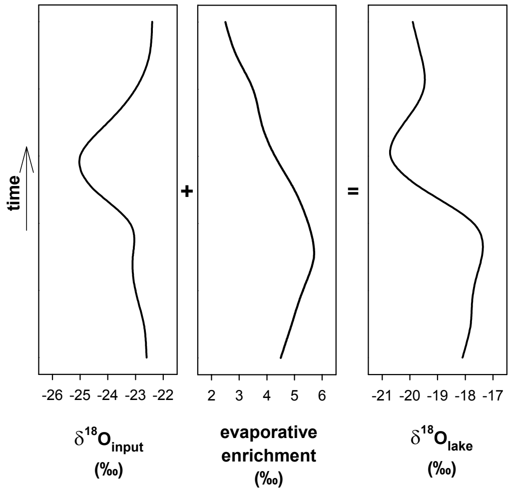

Figure 6. Schematic diagram showing that changes in lake water $\delta^{18}O$ over time are controlled by a combination of changes in the isotopic composition of input waters, which should occur parallel to the GMWL (or LMWL) in $\delta^{18}O$-$\delta^{2}H$ space, modified by changes in the extent of secondary evaporative enrichment, which occur along a shallower slope parallel to the LEL. The equivalent hydrogen-isotope input and evaporative enrichment records would obviously display similar trends, although the MWL-LEL geometry would result in a differing lake water $\delta^{2}H$ history.

Comprehensive discussions of archive-specific methods and interpretation of lake sediment oxygen isotope records have been presented by Edwards and McAndrews (1989), Talbot (1990), Talbot and Kelts (1990), Edwards (1993), Holmes (1996), Li and Ku (1997), Ito (2001), Wolfe et al. (2001b), Shemesh et al. (2001) and others. Lake sediment oxygen-isotope studies at high latitudes have mainly been conducted as part of multidisciplinary investigations aimed at documenting regional Late Glacial and Holocene climatic and hydrological history (e.g., Anderson et al. (2001); Shemesh et al. (2001); Hammarlund et al. (2002); Hu and Shemesh (2003)), and paleohydrological conditions associated with the advance and retreat of circumpolar boreal treeline (e.g. Wolfe et al. (1996); MacDonald et al. (this volume)).

Below we highlight two examples from recent investigations demonstrating, respectively, the derivation and interpretation of millennial-scale MWL-parallel fluctuations in lake water $\delta^{18}O$ inherited by sedimentary carbonates in a ground water-fed lake in northern Sweden, and the analysis of decadal-scale LEL-parallel changes in the oxygen-isotope composition of a lake in northern Canada recorded by algal cellulose.

Lake Tibetanus, Sweden

Abisko National Park, northern Sweden, is situated east of the main drainage divide of the Scandes mountain range, close to the Norwegian border. Following the establishment of the Abisko Scientific Research Station 100 years ago, numerous ecological research projects have been carried out in this subarctic area, many of which have focused on the distinct climatic and vegetational gradients, particularly the altitudinal forest-tundra ecotone. Holocene vegetation dynamics are relatively well known through paleoecological investigations of peat sequences and lake sediments (Sonesson 1968, 1974; Barnekow 1999, 2000), and the climatic forcing of past environmental change in the area has received increased attention as part of multi-disciplinary research based at the Climate Impacts Research Centre at Abisko since the late 1990s (e.g., see Bigler et al. (2002)).

Multi-proxy investigations of the sediments underlying Lake Tibetanus, a small hard-water lake within the upper mountain-birch zone close to Abisko, have yielded particularly crucial information, including the development of a paleoprecipitation $\delta^{18}O$ history, derived through coupling of paleo-isotope data obtained from sedimentary carbonate with pollen-inferred paleotemperature estimates (Hammarlund et al. 2002). As revealed in Figure 7, the modern isotope hydrology of Lake Tibetanus is characterized by complete absence of evaporative enrichment in the lake water, attributable to rapid throughflow of local ground water. Independent evidence suggests that this hydrologic situation has persisted throughout the lake's existence (Hammarlund et al. 1997) and thus the lake water $\delta^{18}O$ history should be characterized by purely MWL-parallel variability.

$\delta^{18}O$ records were obtained on three different carbonate components from a sediment sequence spanning the last 10,000 years (Figure 8). The longest record was developed from fine-grained sedimentary calcite, originating primarily from photosynthesis in submerged *Chara* algae during early summer, while slightly shorter records were gained from analyses of aragonitic shells of *Pisidium* sp. molluscs and calcitic valves of adult *Candona candida* ostracodes precipitating, respectively, throughout the summer and in late autumn. Consistent oxygen-isotope offsets between the three records can be attributed to a combination of mineralogical, biological and temperature-dependent effects (the last related to the differing annual periods of calcification).

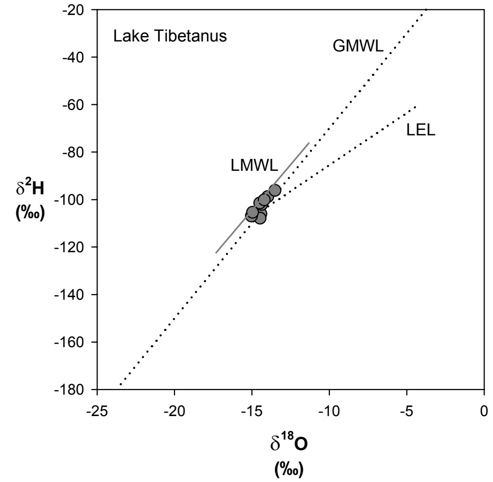

Figure 7. Plot of δ^{18}O versus δ^2H values for samples of lake water collected at intervals over a three-year period (1995-98) from Lake Tibetanus, northern Sweden (68.3°N; 18.7°E), reported by Hammarlund et al. (2002). The LMWL was estimated from isotopic data obtained on samples of local precipitation and ground water and the LEL was calculated based on local climate normals and data from the Global Network for Isotopes in Precipitation (IAEA/WMO 2001). Owing to the short water residence time, the lake water isotopic composition is strongly controlled by that of inflowing ground water with negligible evaporative enrichment, even during the summer months. Independent evidence suggests that rapid flushing has characterized the lake throughout its history and hence that reconstructed lake water δ^{18}O should be a good proxy for the δ^{18}O of local paleoprecipitation.

Based on the assumption of stable open-drainage conditions resembling the modern hydrological situation throughout the lake's history, long-term changes in carbonate δ^{18}O were interpreted as reflecting variations in lake water δ^{18}O inherited from precipitation δ^{18}O, modified slightly due to temperature-dependent variations in carbonate-water equilibrium isotope fractionation. The pollen-based reconstruction of mean July air temperature was used to account for the effect of changing water temperature on the sedimentary carbonate δ^{18}O record, assuming temperature sensitivity of -0.25 ‰/K (Friedman and O'Neil 1977). Since inferred temperatures were similar to present at the beginning and end of the record, but warmer during the intervening period, this temperature correction resulted in a modest increase in inferred lake water δ^{18}O throughout the mid-part of the record (see Figure 8).

Comparison of the derived paleoprecipitation δ^{18}O record and the pollen-based temperature reconstruction from Lake Tibetanus reveals clear evidence for changing temporal isotope-temperature relations over the Holocene at this site. As addressed in detail by Hammarlund et al. (2002), use of the traditional "isotope paleothermometer", based on the observed modern spatial relation between δ^{18}O of local amount-weighted mean annual precipitation and mean annual temperature of about +0.7‰/K (Dansgaard 1964), is only applicable over the past c. 6000 years. Evidently, significantly different and variable δ^{18}O-MAT relations prevailed 10,000-6000 years ago, presumably reflecting substantially different atmospheric and oceanic circulation patterns than those of the present. As suggested by Hammarlund et al. (2002), these differences likely included enhanced westerly air-mass circulation leading to relatively moist maritime conditions in response to high summer insolation, elevated sea-surface temperatures in the Norwegian Sea and steep meridional pressure gradients. The subsequent transition to drier conditions, which is clearly reflected in the pollen-based reconstruction of mean annual precipitation shown in Figure 8, was accompanied by major changes in forest extent and composition, recorded unequivocally by pollen and plant macrofossil data (Barnekow 1999).

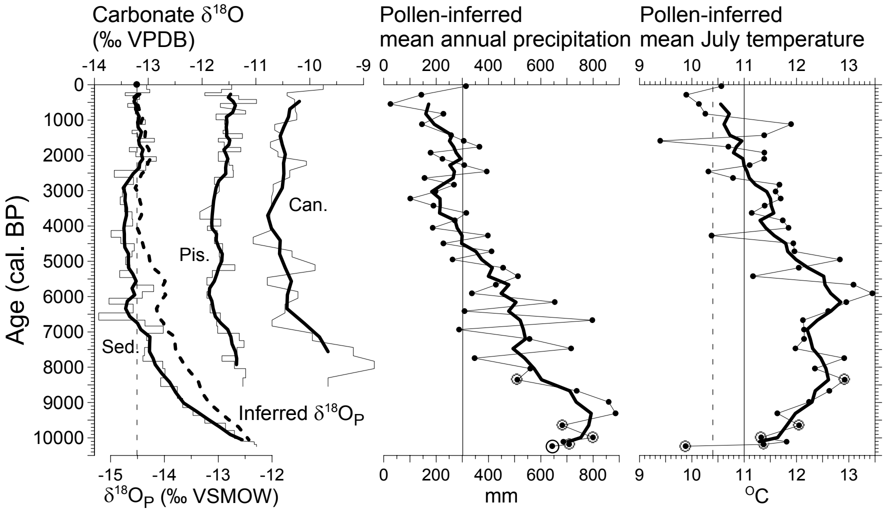

Figure 8. Oxygen-isotope records for a core from Lake Tibetanus plotted against calibrated radiocarbon age, along with pollen-inferred precipitation and temperature, based on data from Hammarlund et al. (2002). Samples with poor modern pollen analogues are circled. Thick lines represent five-point running averages. Modern temperature and precipitation values for Abisko are shown by vertical lines (dashed line represents calculated modern mean July air-temperature at Lake Tibetanus). Sed = fine-grained sedimentary carbonate (primarily *Chara* calcite); Pis = *Pisidium* sp. mollusc aragonite; Can = adult *Candona candida* ostracode calcite. The heavy dashed line represents the inferred δ^{18}O of local paleoprecipitation ($\delta^{18}O_P$) after accounting for temperature-dependent fractionation between sedimentary carbonate and lake water using pollen-inferred mean July temperature.

Jemis Lake, Canada

The Peace-Athabasca Delta (PAD), situated at the confluence of the Peace, Athabasca and Fond-du-Lac (Lake Athabasca) drainages in northern Alberta, Canada, is an extensive high-latitude fluvio-deltaic complex of considerable cultural, historical and ecological significance (e.g., see Peace-Athabasca Delta Project Group (1973); Prowse et al. (1996)). The PAD is also a key node in the Mackenzie River basin, the major North American source of fresh water to the Arctic Ocean. Concerns about the combined influence of climate change, regulation of Peace River flow due to hydro-electric power generation, increasing consumptive use of Athabasca River water associated with tar sands development, and changes in land use upstream have led to numerous environmental studies in the PAD over the past 30 years. Many of these investigations have focussed on the frequency and severity of periodic ice-jam floods on the Peace and Athabasca rivers, which are believed to provide crucial fluxes of water and nutrients to lakes in the delta that are perched beyond the reach of normal seasonal high-water flooding from these two rivers. Increasing attention is also being paid to the implications downstream in the Mackenzie River system.

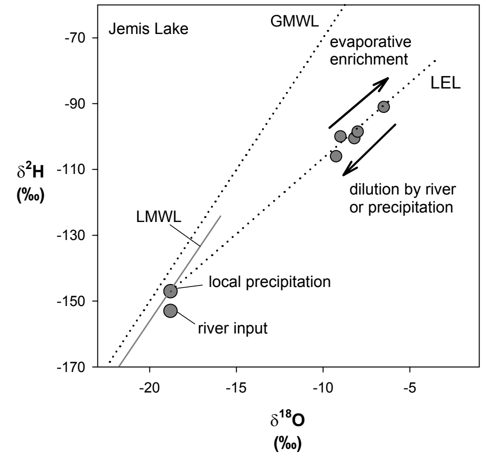

Figure 9. Lake water δ^{18}O and δ^2H data from samples collected over two years (2000-02) from Jemis Lake in the Peace-Athabasca Delta, Canada (58.3°N; 111.5°W). Shown for reference is the calculated local evaporation line (LEL) for lakes fed by local amount-weighted mean annual precipitation, based on local climate normals and data from the Canadian Network for Isotopes in Precipitation (see Birks et al. (2002); Gibson and Edwards (2002)). Jemis Lake receives input from precipitation and river water, which both have similar isotopic compositions, thus leading to LEL-parallel shifts in lake water composition, regardless of source, in response to changing water balance.

Jemis Lake, a small shallow water body located centrally in the PAD, has been targeted by a number of investigators as being potentially representative of broader conditions in the delta over time, as well as being relatively well-known to the residents of Fort Chipewyan, the major local settlement. Seasonal monitoring of the isotopic composition of Jemis Lake as part of longer-term isotope hydroclimatology studies in the PAD suggests that the water balance of this lake (Figure 9), like many others in the delta, is strongly controlled by the opposing effects of evaporation, leading to heavy-isotope enrichment, and periodic inflow of isotopically depleted river water and precipitation (see also Figure 5). Because river water inputs to Jemis Lake and local precipitation have similar isotopic compositions, fluctuations in the lake's water balance are expressed as

LEL-parallel shifts in $\delta^{18}O$-$\delta^{2}H$ space, which can thus be translated directly into shifting E/I ratio using equation (3). Indeed, consideration of reconstructed lake water $\delta^{18}O$ obtained from algal cellulose in a short core of sediment from Jemis Lake suggests that changes in the isotopic composition of lake water throughout at least the past 150 years (and perhaps much longer; see Wolfe et al. (2002)) have also been strongly controlled by LEL-parallel fluctuations, as revealed by strong inverse correspondence between inferred E/I and an independently derived flood history (Figure 10). Such insight and context are critical to understanding the hydro-ecological evolution of the Peace-Athabasca Delta, both because of its substantial natural heritage value and because of overarching concerns about the responsible stewardship of water resources within the Mackenzie River system as a whole.

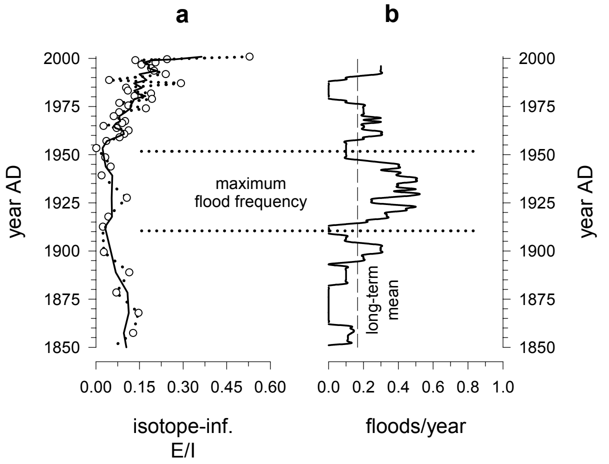

Figure 10. a) Speculative isotope-inferred history of evaporation:inflow ratio (E/I) in Jemis Lake AD 1850-2000, based on lake water $\delta^{18}O$ reconstructed from algal cellulose in a short sediment core (after Wolfe et al. (2002)). E/I was calculated using Equation (3), assuming input of either river water or precipitation approximating the $\delta^{18}O$ of modern local mean annual precipitation (see Figure 9). The chronology was developed from ^{210}Pb and ^{14}C data. The solid line on the E/I plot is a three-point running mean. *b)* Peace River spring ice-jam flood frequency record in the Peace-Athabasca Delta derived from anecdotal and historical sources by Timoney et al. (1997), expressed as a 10-year mean. The water balance of Jemis Lake has apparently fluctuated in concert with flood frequency throughout the past 150 years, as shown clearly by the inverse relation between isotope-inferred E/I and estimated floods/year, with particularly well-marked depiction of the prolonged period of wetter conditions between about 1910 and 1955.

Summary and future perspectives

This chapter has outlined the major features of water-isotope labelling in the hydrological cycle, highlighting the characteristic signals in high-latitude lakes that may provide clues to interpret stratigraphic variations in sediment cores. Although lakes respond individualistically, the strongest signals at seasonal time scales in most water bodies having short hydraulic residence times are likely to derive from snowmelt influx and evaporative enrichment. Samples from lake sediment cores tend to naturally average and integrate seasonal signals (as do the waters in lakes having long residence times), thus often affording the opportunity to resolve longer-term fluctuations in the isotopic composition of lake water into two respective components, one reflecting shifting input water composition (and hence expected to vary parallel to the Global or Local Meteoric Water Line in δ^{18}O-δ^{2}H space) and the other reflecting varying evaporative enrichment along trajectories parallel to the Local Evaporation Line. Deconvolution of MWL-parallel signals and associated estimates of the isotopic composition of paleoprecipitation may provide direct links to global paleoprecipitation isotope fields, analogous to those afforded by high-latitude ice cores (e.g., see Rozanski et al. 1997)), whereas LEL-parallel signals provide the basis for reconstructing local hydroclimate (e.g., MacDonald et al. (1993)).

Future advances in the application of water-isotope tracers in high-latitude lakes are likely to be realized from the development of coupled water-isotope records, using new archive materials to reconstruct lake water isotopic history in δ^{18}O-δ^{2}H space, thus permitting fuller and more sophisticated paleohydrologic and paleoclimatic interpretation.

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