Regional water balance trends and evaporation-transpiration partitioning from a stable isotope survey of lakes in northern Canada

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[1] Regional variations in evaporation losses and water budget are interpreted from systematic isotopic patterns in surface waters across a 275,000 km^2 region of northern Canada. Differential heavy isotope enrichment in a set of >255 nonheadwater lakes sampled by floatplane during 1993 and 1994 is strongly correlated to varying hydroclimatic conditions across the region. Calculated catchment-weighted evaporation losses typically range from ~10–15% in tundra areas draining into the Arctic Ocean to as high as 60% in forested subarctic areas draining to the Mackenzie River via Great Bear or Great Slave Lakes. Because of the diversity in drainage order and the ratio of catchment to surface area, lakes in the region may inherit as little as 30% to as much as 99% of their isotopic enrichment signal from upstream water bodies. Open-water evaporation generally decreases with increasing latitude and accounts for 5–50% of total evapotranspiration. Coupling of meteorological and isotopic data permits a novel assessment of regional evaporation-transpiration flux partitioning in the three major ecoclimatic zones (high-boreal forest, subarctic forest-tundra, and low-arctic shrub tundra), while the differing frequency distributions of lake water balance in these zones provides a new index of landscape-scale hydroclimatology that may have significant potential for investigating ongoing (or past) changes in response to high-latitude climate change. *INDEX TERMS:* 1655 Global Change: Water cycles (1836); 1836 Hydrology: Hydrologic budget (1655); 1833 Hydrology: Hydroclimatology; 1818 Hydrology: Evapotranspiration; 4870 Oceanography: Biological and Chemical: Stable isotopes; *KEYWORDS:* stable isotopes, oxygen-18, deuterium, lakes, regional water balance, spatial variability

1. Introduction

[2] Heavy isotope enrichment is widely observed in lakes and other surface waters undergoing evaporation because of mass-dependent differences in the equilibrium vapor pressures and gas-phase molecular diffusivities among the naturally occurring water isotopomers, including common light water containing only ^{16}O and ^{1}H ($^{1}H^{1}H^{16}O$) and two of the rare heavy isotopomers containing either ^{18}O ($^{1}H^{1}H^{18}O$) or ^{2}H ($^{1}H^{2}H^{16}O$). An evaporating moisture flux is preferentially depleted in ^{18}O and ^{2}H, leading to progressive enrichment of these heavy isotopes in the residual liquid. In contrast, transpiration does not generally produce a fractionation between the moisture flux and residual soil water, since the transfer of moisture to the atmosphere by plants is essentially quantitative, and heavy-isotope enrichment is restricted to internal plant waters [*Jacob and Sonntag*, 1991; *Wang and Yakir*, 2000]. The absolute isotopic enrichment of a surface water body is controlled by both atmospheric and water balance processes [*Gat*, 1995]. The principal atmospheric factors are humidity, temperature, and the isotopic composition of ambient atmospheric moisture, which collectively control the relative rates and ultimate limits of heavy-isotope enrichment, whereas water balance (i.e., the evaporation/inflow ratio of the reservoir, or E/I) determines the degree to which isotopic enrichment is attenuated by dilution. Evaporation processes and the results of experimental studies to characterize turbulent and diffusive mass transfer mechanisms from an isotopic perspective have been extensively addressed in a number of key publications, including those by *Gat* [1996], *Gonfiantini* [1986], *Merlivat and Jouzel* [1979], and *Stewart* [1975].

[3] Valuable insight into the isotopic behavior of lakes in the strongly seasonal, continental climate of northern Canada has been gained through field-based investigations of lake evaporation and water balance over the past decade in the Northwest Territories, Nunavut, and northern Alberta [*Gibson et al.*, 1993, 1996a, 1996b, 1998, 1999, 2002; *Gibson*, 2001, 2002]. Isotopic labeling of waters at catchment to regional scale is commonly manifested by the existence of two linear trends on a conventional ^{2}H-^{18}O crossplot

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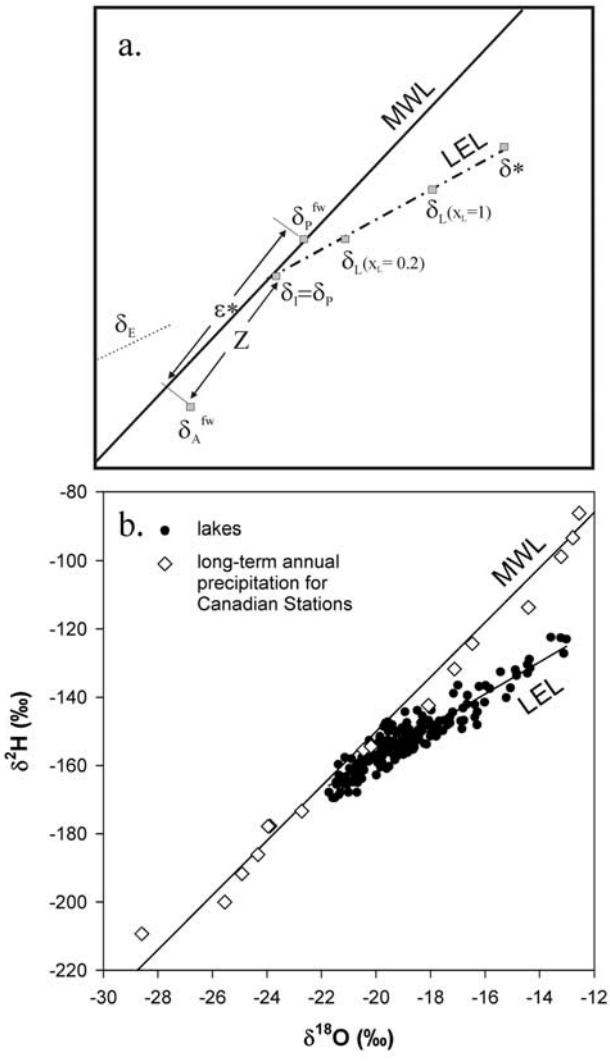


Figure 1. Plots of ^{2}H-^{18}O (a) showing schematic relationships between isotopic compositions in surface water systems including limiting isotopic enrichment (δ^*), mean annual input ($\delta_I = \delta_P$), lake water δ_L under two water balance scenarios where $x = 1$ and $x = 0.2$, and atmospheric moisture (δ_A^{fw}) estimated assuming equilibrium with flux-weighted precipitation (δ_P^{fw}) and noting that $Z = \delta_I - \delta_A^{fw}$, which is apparent inflow-atmospheric moisture separation, and (b) showing lakes sampled in central Arctic and precipitation from Canadian stations. Note that lakes plot along an evaporation line (LEL) with a slope of 4.7 ($r^2 = 0.88$) and precipitation plots close to meteoric water line (MWL) of $\delta^2H = 8\delta^{18}O + 10$. Isotopic composition of lakes is controlled predominantly by variations in evaporation loss from lake and upstream lakes.

(Figures 1a and 1b) (the δ notation expresses the relative abundances of ^{18}O and ^{2}H as deviations in per mil (‰) from a given standard, normally Vienna Standard Mean Ocean Water, such that $\delta_{sample} = 1000(R_{sample}/R_{VSMOW}) - 1$, where R is $^{18}O/^{16}O$ or $^2H/^1H$), allowing differentiation of "meteoric" waters that retain their original isotopic composition (i.e., local precipitation, including that stored in the snowpack, and derived groundwaters) from waters that have undergone subsequent heavy-isotope enrichment due to evaporation (i.e., surface waters and derived stream and groundwaters). The former typically cluster along a local meteoric water line (MWL) having a slope of ~8, reflecting the pervasive influence of mass-dependent isotope exchange processes during the transport and progressive distillation of atmospheric moisture, whereas evaporatively enriched waters generally plot to the right of the MWL, as a consequence of additional kinetic effects during the evaporation process related to variations in both overall mass and its distribution within the water molecules [*Dansgaard*, 1964; *Merlivat and Jouzel*, 1979]. As shown in Figure 1b, the isotopic composition of precipitation in northern Canada generally plots close to the global meteoric water line (MWL) of *Craig* [1961], defined by $\delta^2H = 8\delta^{18}O + 10$, which is the locus of weighted monthly precipitation sampled at several hundred stations worldwide [see *Rozanski et al.*, 1993].

[4] Evaporated waters within a given region often cluster along a local evaporation line (LEL) having a slope in the range 4–6, variably well defined as a function of the range of *E*/*I* ratios and degree of hydraulic connectivity within a system. The intersection point of an LEL with the local MWL commonly affords an excellent approximation of the weighted mean isotopic composition of local precipitation (δ_P), which is a key datum for isotope-mass balance investigations, since offset from the MWL along the LEL increases in proportion to the cumulative fraction of water lost by evaporation upstream of a given sampling station. Basic knowledge or assumptions about the hydrologic status of a lake and its catchment can be coupled with this isotopic information to quantify or constrain both oxygen and hydrogen isotope-mass budgets.

[5] The main objective of this study is to explore the potential application of isotope tracers as indicators of water balance systematics and variability at the large scale and to gain a better understanding of the role of lakes in the regional runoff regime of a boreal-arctic transition zone. Here we present and discuss a previously unreported stable isotope data set acquired from a regional survey of lake water quality across a remote 275,000 km^2 region of the continental Arctic and subarctic encompassing northern tree line (Figure 2). Water samples were collected by the Department of Indian and Northern Affairs Canada as part of a study to investigate baseline controls on water chemistry and potential impacts related to mining and recreation [*Puznicki*, 1996]. Archived samples were separately analyzed for $\delta^{18}O$ and δ^2H by the authors at the University of Waterloo (Figure 1b), revealing systematic spatial variability in evaporative isotopic enrichment in the lakes consistent with previous point observations at localities within the region and affording the opportunity to gain a better understanding of regional variations in water budgets. As outlined in section 1.1.1, we apply a steady state isotope mass balance approach to the data set to quantify regional trends in evaporation loss as a fraction of total outflows, evaporation losses from lakes in proportion to that from upstream reservoirs, and evaporation losses as a fraction of the total evapotranspiration flux. Supporting

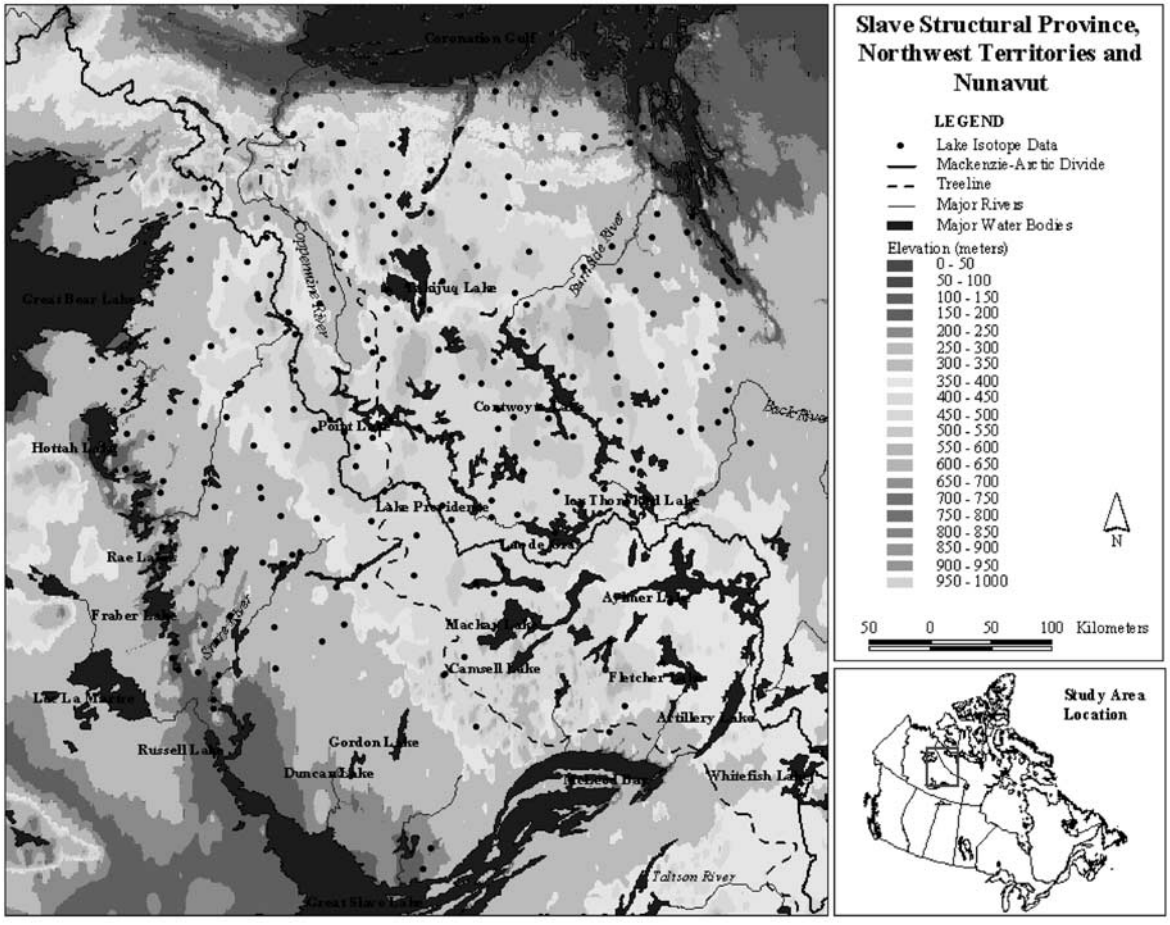


Figure 2. Relief map of central Arctic of Canada showing sampling locations for this study. Note that solid line indicates Arctic-Mackenzie drainage divide and dotted line shows approximate position of northern tree line. See color version of this figure at back of this issue.

information for the calculations is obtained from a geographic information system (GIS) and map-based analysis of isotopic and hydroclimatic parameters interpolated for each sampling site and basin. Use of the steady state approach is justified by the large volume of the lakes, most of which are larger than 10^9 m^3, with average depths of >14 m. Such lakes are known from previous studies to have subdued seasonal isotopic cycles [see *Gibson*, 2001]. The analysis offers a unique, broad-scale perspective of regional water balance using models that have been intensively evaluated and refined in field-based studies and provides valuable insight into the role of lakes in the runoff regime of Precambrian Shield terrain underlain by permafrost.

1.1. Theory

1.1.1. General lake balance

[6] The water and isotope mass balances of a well-mixed lake undergoing evaporation while maintaining a long-term constant volume (assuming constant density of water) are

$$I_L = Q_L + E_L \tag{1}$$

$$I_L\delta_I = Q_L\delta_Q + E_L\delta_E, \tag{2}$$

where I_L is combined surface and subsurface inflow, Q_L is combined surface and subsurface outflow, E_L is lake evaporation, and δ_I, δ_Q, and δ_E are the isotopic compositions of inflow, outflow, and evaporative flux, respectively. Substituting $Q_L = I_L - E_L$ from (1) and $\delta_Q = \delta_L$ (acknowledging that average outflow will be isotopically similar to the isotopic composition of lake water δ_L), equation (2) can be rearranged as

$$\frac{E_L}{I_L} = \frac{\delta_I - \delta_L}{\delta_E - \delta_L}, \tag{3}$$

which assumes no long-term storage changes in the reservoir. As noted earlier, the evaporation flux δ_E is typically depleted in the heavy isotopes relative to lake water δ_L. Although impossible to measure directly, the magnitude of isotopic separation between lake water and the isotopic composition of the evaporation flux has been shown to be dependent on evaporation temperature, boundary layer state (i.e., laminar, turbulent, or static), and ambient atmospheric conditions (relative humidity and isotopic composition of atmospheric moisture). The stan-

dard approach for estimating δ_E from these boundary characteristics is by the *Craig and Gordon* [1965] model assuming negligible resistance to mixing in the liquid phase [see *Gat*, 1995] (note that the *Craig and Gordon* [1965] equation is modified to directly utilize isotopic data in per mil rather than as a decimal fraction, i.e., −15‰ rather than −0.015 should be used),

$$\delta_E = \frac{\alpha^*\delta_L - h\delta_A - \varepsilon}{1 - h + 10^{-3}\varepsilon_K}, \tag{4}$$

in parts per mil, where α^* is the equilibrium liquid-vapor isotope fractionation ($\alpha^* = 1 + \varepsilon^*$), h is the atmospheric relative humidity (ranging from 0 to 1) normalized to the saturation vapor pressure at the temperature of the air-water interface, δ_A is the isotopic composition of ambient moisture, and

$$\varepsilon = \varepsilon^* + \varepsilon_K, \tag{5}$$

in parts per mil, where ε is the total isotopic separation factor including both equilibrium ε^* and kinetic ε_K components. The equilibrium separations can be evaluated using the empirical equations determined experimentally by *Horita and Wesolowski* [1994] given by

$$\begin{aligned}\varepsilon^* \approx 10^3 \ln\alpha\left({}^{18}\mathrm{O}\right) = &-7.685 + 6.7123\left(10^3/T\right)\\ &- 1.6664\left(10^6/T^2\right)\\ &+ 0.35041\left(10^9/T^3\right)\end{aligned} \tag{6}$$

$$\begin{aligned}\varepsilon^* \approx 10^3 \ln\ \alpha\left({}^{2}H\right) = &\ 1158.8\left(T^3/10^9\right) - 1620.1\left(T^2/10^6\right)\\ &+ 794.84\left(T/10^3\right) - 161.04\\ &+ 2.9992\left(10^9/T^3\right),\end{aligned} \tag{7}$$

in parts per mil, where T is the interface temperature (K). These relatively new equations yield values only slightly different than those reported by *Majoube* [1971] and *Bottinga and Craig* [1969] in the range of temperatures expected for average evaporation conditions (0°–25°C).

[7] Kinetic enrichment factors ε_K are dependent on both the boundary layer conditions and the humidity deficit evaluated according to

$$\varepsilon_K = C_K(1 - h), \tag{8}$$

in parts per mil, where constant, experimentally-determined C_K values of 14.2‰ for oxygen and 12.5‰ for hydrogen are used as representative of typical lake evaporation conditions [*Gonfiantini*, 1986; *Araguás-Araguás et al.*, 2000].

[8] Introducing the term X_L, which is the fraction of water loss by evaporation (E/I ratio),

$$x_L = \frac{E_L}{I_L} = \frac{E_L}{E_L + Q_L}, \tag{9}$$

and substituting equation (4) into equation (3) then yields

$$x_L = \frac{(\delta_L - \delta_I)}{m(\delta^* - \delta_L)}, \tag{10}$$

where δ_L is the steady state isotopic composition of the lake, m is the enrichment slope

$$m = \frac{h - 10^{-3}\varepsilon}{1 - h + 10^{-3}\varepsilon_K} \tag{11}$$

as defined in previous studies [*Welhan and Fritz*, 1977; *Allison and Leaney*, 1982], and

$$\delta^* = \frac{h\delta_A + \varepsilon}{h - \varepsilon \cdot 10^{-3}}, \tag{12}$$

in parts per mil, is the limiting isotopic composition under local climatological conditions [*Gat and Levy*, 1978; *Gat*, 1981].

[9] Residence time of a reservoir in long-term isotopic and hydrologic steady state is given by

$$\tau = \frac{1}{x_L}\frac{E}{V}, \tag{13}$$

in years, which assumes that the evaporative enrichment signal that defines X_L is derived entirely from the lake itself. If appreciable isotopic enrichment also occurs in upstream reservoirs, then this must be viewed as an upper limit.

1.1.2. Headwater lakes, nonheadwater lakes, and catchments

[10] Definition of the input signal is the most important distinction between headwater and nonheadwater basins. For headwater lakes, the assumption that inflow is close to the isotopic composition of precipitation ($\delta_I = \delta_P$) is a reasonable first approximation, as demonstrated in previous studies [*Gibson et al.*, 2002]. Because of the so-called "string-of-lakes" effect (adopting the terminology of *Gat and Bowser* [1991]), inflow to nonheadwater lakes is expected to be variably enriched in the heavy isotopes by evaporation from upstream lakes. This makes it impossible to distinguish isotopically the evaporation occurring directly from a lake from that occurring upstream, without additional information on the isotopic composition of inflows. Nevertheless, evaporation losses from the entire catchment can be evaluated by substitution of $\delta_I = \delta_P$ into equation (10), acknowledging that input integrated over the catchment as a whole must be isotopically similar to the isotopic composition of precipitation, as demonstrated by *Gibson* [2001]. This provides an index of evaporation loss for a catchment area (or headwater lake) given by

$$x_C \text{ or } x_{\mathrm{HW}} = \frac{(\delta_L - \delta_P)}{m(\delta^* - \delta_L)}. \tag{14}$$

[11] One important application, as suggested by *Gibson et al.* [2002] is for tracing of long-term runoff from the catchment area. Assuming that runoff from the catchment is equal to inflow to the lake minus precipitation on the lake, then the runoff can be calculated as

$$Q_{\mathrm{DBA}} = \left(\frac{E_L}{x_{\mathrm{HW}}} - P_L\right)\beta, \tag{15}$$

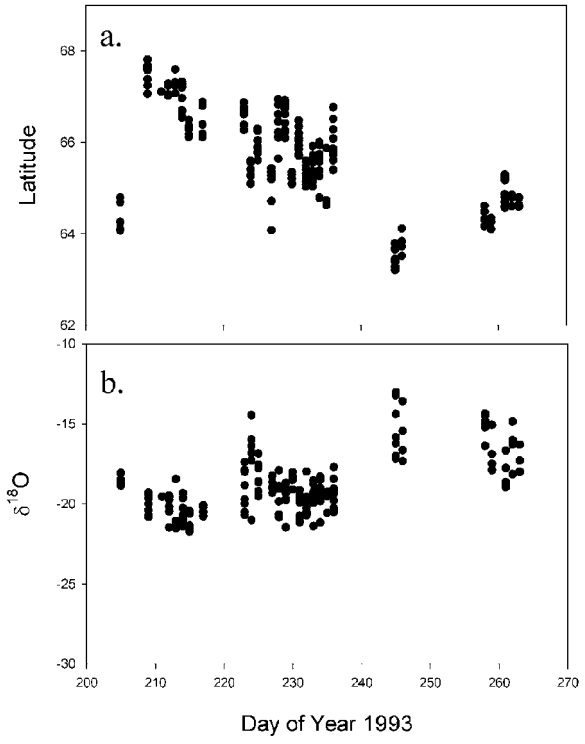


Figure 3. 1993 time series of (a) water sampling latitude and (b) $\delta^{18}O$, showing that temporal changes in isotopic composition are related to latitude rather than due to substantial seasonal differences in isotopic enrichment.

where

$$\beta = \frac{\mathrm{LA}}{\mathrm{DBA}} \tag{16}$$

and LA is the lake area, DBA is the drainage basin area, and CA is the catchment area such that CA = DBA + LA.

[12] For nonheadwater systems, it is possible to distinguish isotopically the evaporation losses from successive reservoirs [*Gat and Bowser*, 1991], although the data for inflows are not available for the present study. The evaporation losses in a nonheadwater lake consist of evaporation occurring from the lake and evaporation from upstream lakes,

$$x_C = \frac{E_L + \sum E_L^U}{P_L + Q_{\mathrm{DBA}}\beta^{-1}}, \tag{17}$$

where E_L is evaporation from the lake, $\sum E_L^U$ is the sum of evaporation from upstream lakes and reservoirs, P_L is the precipitation on the lake, and Q_{DBA} is the runoff from the catchment area. Partitioning of evaporation losses from upstream lakes can therefore be evaluated by incorporating hydrologic data from other sources. In this case,

$$\frac{E_L}{E_{\mathrm{TOTAL}}} = \frac{E_L}{E_L + \sum E_L^U} = \frac{e_L}{x_c(q\beta^1 + p)}, \tag{18}$$

where e_L, q, and p are, respectively, mean annual lake evaporation, runoff, and precipitation interpolated for each site from the Hydrological Atlas of Canada [*denHartog and Ferguson*, 1978a, 1978b].

[13] The contribution of evaporation (E) to the total evapotranspiration flux (E_T) in the catchment is likewise evaluated using

$$\frac{E}{E_T} = \frac{E_L + \sum E_L^U}{E_T} = \frac{x\left(q\beta^{-1} + p\right)}{x\left(q\beta^{-1} + p\right) + e_t}, \tag{19}$$

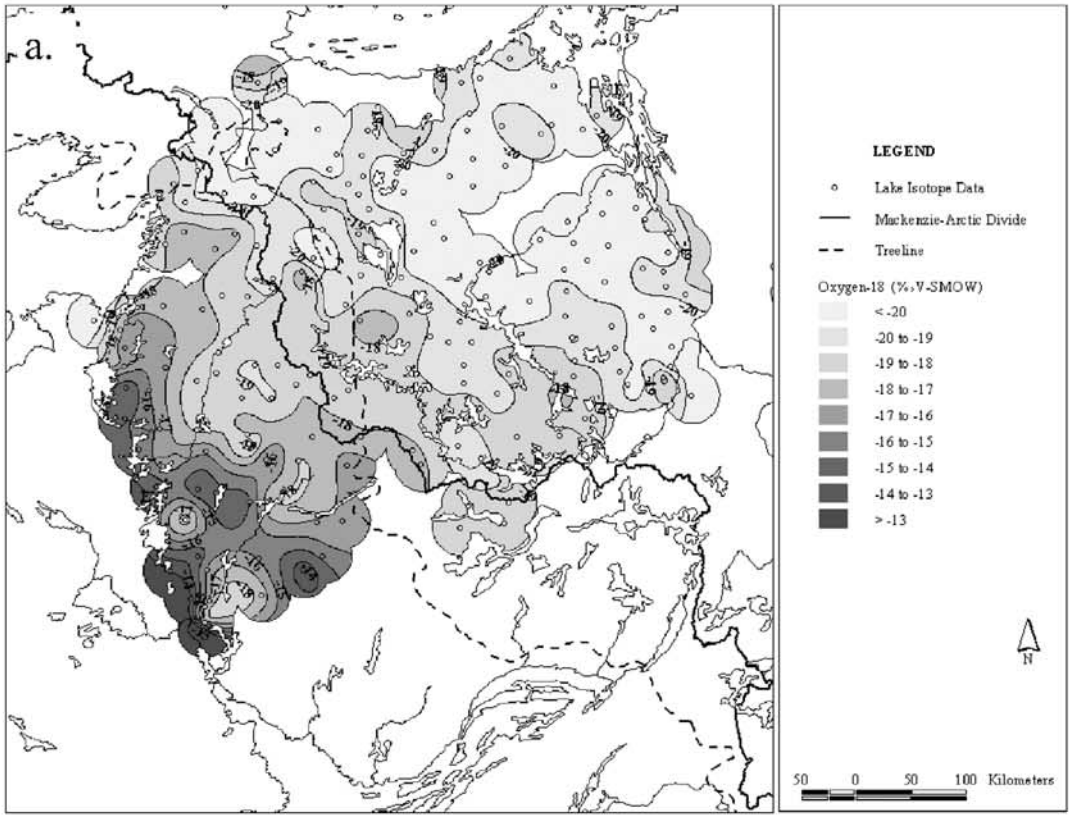
where e_t is derived evapotranspiration from the land surface, also interpolated from the Hydrological Atlas of Canada [*denHartog and Ferguson*, 1978b]. Although the latter two indices rely heavily on nonisotopic information, the isotope data assist in constraining the mechanisms of water transfer and offer a new perspective on the role of lakes in the regional hydrologic regime.

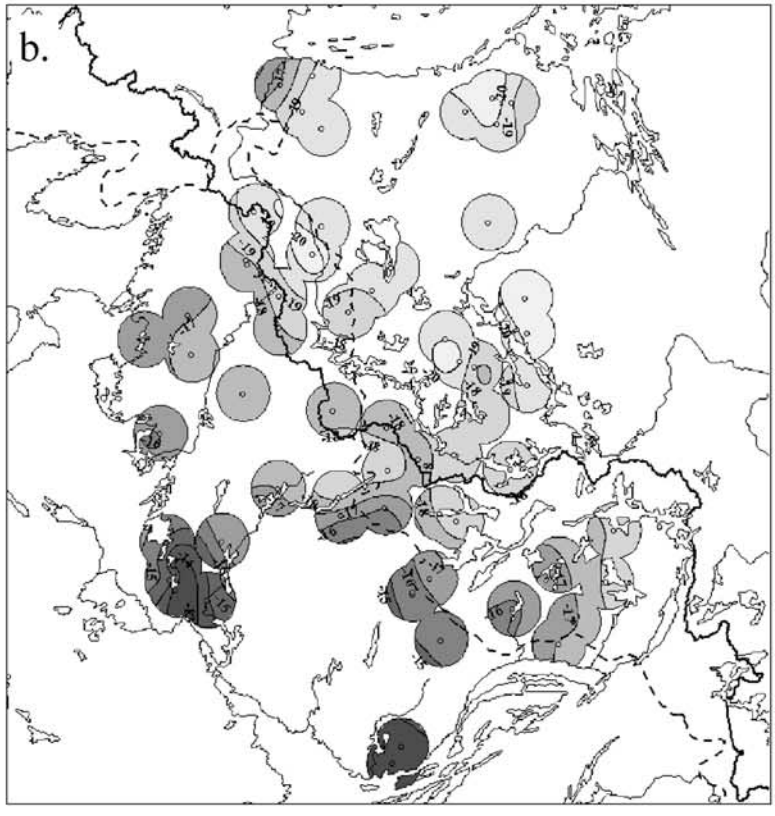
1.2. Study Area and Methods

[14] The study was conducted in the continental Arctic and subarctic of the Northwest Territories and Nunavut (Figure 2), within the ranges of 62°N–68°N and 107°W–118°W, in an area characterized by low relief and myriad lakes. Bedrock consists mainly of fractured Canadian Shield rocks of Precambrian age with a thin to discontinuous overburden of glacial till, moraine, and lesser organic soils. Permafrost extent ranges from thin, discontinuous occurrences in bogs and peaty soils near Yellowknife to thick, continuously frozen ground up to several hundred meters deep in the north and northeast. Apart from the extreme northwestern region along the Coppermine River, where river channels may be deeply incised, drainage is disorganized, and large areas of up to 20,000 km^2 may be drained through individual lakes. Lakes in this setting can be considered as nodes in a regional string-of-lakes drainage network. The region straddles the divide between the Mackenzie River basin and the Arctic coastal drainage, producing a prominent southwest-northeast bidirectional drainage pattern. Steep southwest-to-northeast hydroclimatic gradients also exist in the area, with mean annual temperatures ranging from −6° to −14°C. Precipitation and evaporation range from about 350 and 400 mm yr^{-1}, respectively, in the southwest to <200 mm yr^{-1} for both parameters in the northeast and along the coast of Coronation Gulf, according to mapped compilations presented in the Hydrological Atlas of Canada [*denHartog and Ferguson*, 1978a, 1978b].

[15] The study area spans the northern limit of trees, and regional vegetation patterns are strongly influenced by the steep climatic gradients, leading to a steady transition from high-boreal forest to subarctic forest-tundra to low-arctic shrub tundra toward the northeast [*Ritchie*, 1993; see also *Ruhland and Smol*, 1998].

[16] Water samples were collected in 1993 and 1994 as part of a water quality survey of the Slave Structural Province of the Canadian Shield [see *Puznicki*, 1996]. Random and thorough coverage of the study area was ensured by selecting lakes closest to intersections of a 25-km grid on 1:250,000 topographic map sheets that were large enough to accommodate the landing and takeoff of a single-engine float plane. Additional lakes were also


a.
LEGEND
Lake Isotope Data
Mackenzie-Arctic Divide
Treeline
Oxygen-18 (‰ V-SMOW)
< -20
-20 to -19
-19 to -18
-18 to -17
-17 to -16
-16 to -15
-15 to -14
-14 to -13
> -13
N
50 0 50 100 Kilometers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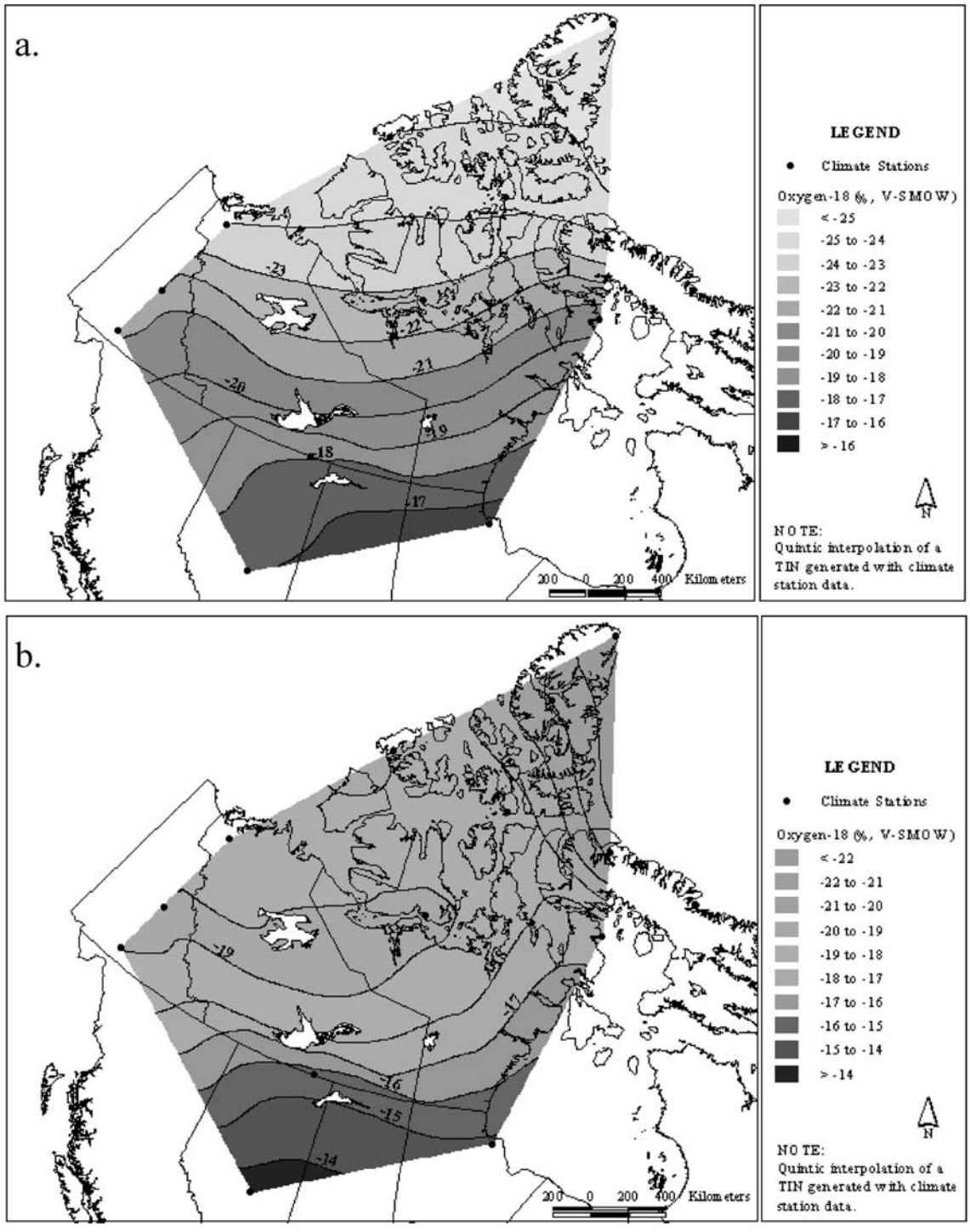

Figure 5. Map showing $\delta^{18}O$ composition of (a) mean annual precipitation and (b) evaporation-flux-weighted (thaw season) precipitation, interpolated from nearby Global Network for Isotopes in Precipitation (GNIP) stations. Similar trends are noted for $\delta^{2}H$. See color version of this figure at back of this issue.

Figure 4. (opposite) Map of $\delta^{18}O$ in lake water sampled during (a) 1993 and (b) 1994. A 25-km contouring radius is maintained to show only areas with high spatial sampling density. Similar trends were noted for $\delta^{2}H$ (not shown). Note that solid line indicates Arctic-Mackenzie drainage divide and dotted line shows approximate position of northern tree line. See color version of this figure at back of this issue.

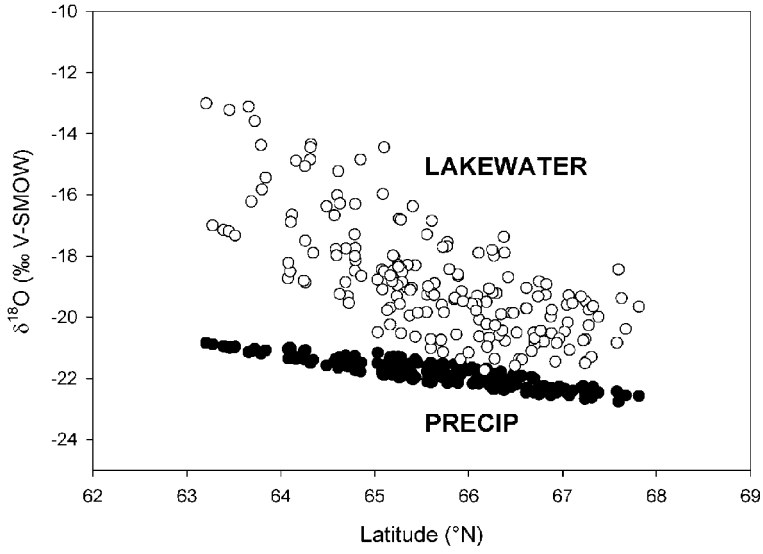

Figure 6. Plot of latitudinal variations of lake water (open circles) sampled during 1993 (from Figure 2) and GNIP mean annual precipitation (solid circles) for each lake site (interpolated from Figure 5a). Note that latitudinal trends in isotopic composition of lake water are larger than variations in precipitation, which is attributed to higher evaporation losses at lower latitudes.

included in specific catchments prone to potential impact from mining or recreational development. Lakes were sampled during the open water season (July to September 1993 and 1994), with sampling conducted as close as possible to the lake center. Irregularly shaped lakes were sampled more than once to ensure that a representative set of samples was obtained. All samples were collected using a 3-L horizontal Van Dorne water sampler, with collection typically at 4 m depth (or middepth in the case of shallower lakes). Measurements were made of lake and sample depths, water temperature, pH, and conductivity on site and were reported with results of major ion and trace element analyses performed in the laboratory by *Puznicki* [1996]. Limnological characteristics of a subset of lakes were also evaluated by *Ruhland and Smol* [1998]. The 182 filtered, but otherwise untreated, samples from the 1993 survey were analyzed for $\delta^{18}O$ and $\delta^{2}H$. Sampling in 1994 was carried out on a smaller set of different lakes overlapping the same area. Seventy of the latter samples were also analyzed for stable isotopes, and the results are presented here to test the reproducibility of the regional trends discerned from the larger sample set.

[17] Isotopic ratios were determined by conventional mass spectrometric techniques in the University of Waterloo Environmental Isotope Laboratory, with $\delta^{18}O$ and $\delta^{2}H$ values reported with respect to Vienna standard mean ocean water (VSMOW) on a scale normalized such that Standard Light Arctic Precipitation (SLAP) has values of −55.5 and −428‰, respectively [see *Coplen*, 1996]. Maximum analytical uncertainties are ±0.1‰ for $\delta^{18}O$ and ±2‰ for $\delta^{2}H$.

1.3. Potential Incomplete Mixing or Temporal Isotopic Enrichment

[18] Prior to analyzing spatial variations the isotopic data were scrutinized to test whether potential "noise" from within-lake heterogeneity or seasonal isotopic enrichment of lake water over the 50-day sampling interval strongly influenced primary isotopic water balance signals. To assess the former, isotopic analyses were obtained on an additional

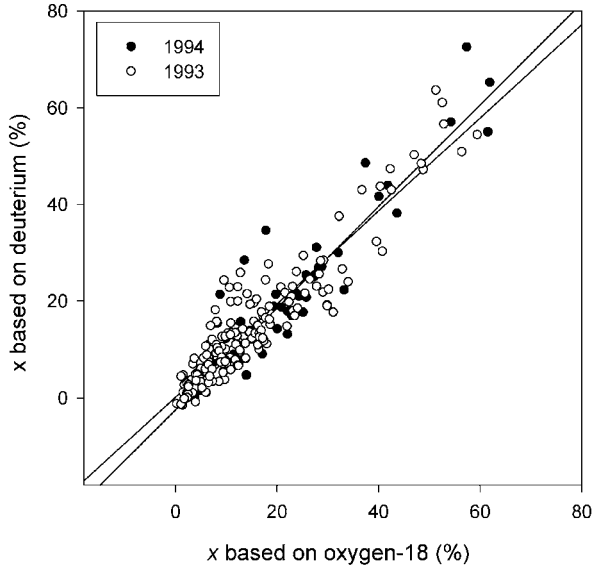

Figure 7. Evaporation/inflow *x* based on deuterium (%) versus *x* based on oxygen-18 (%) for lakes sampled in 1993 and 1994. Good agreement between oxygen-18 and deuterium suggests that basic model exchange parameters are adequately defined. Note that 1994 survey area included several lakes near Great Slave Lake that were not sampled in 1993. These lakes have the highest *x* values observed.

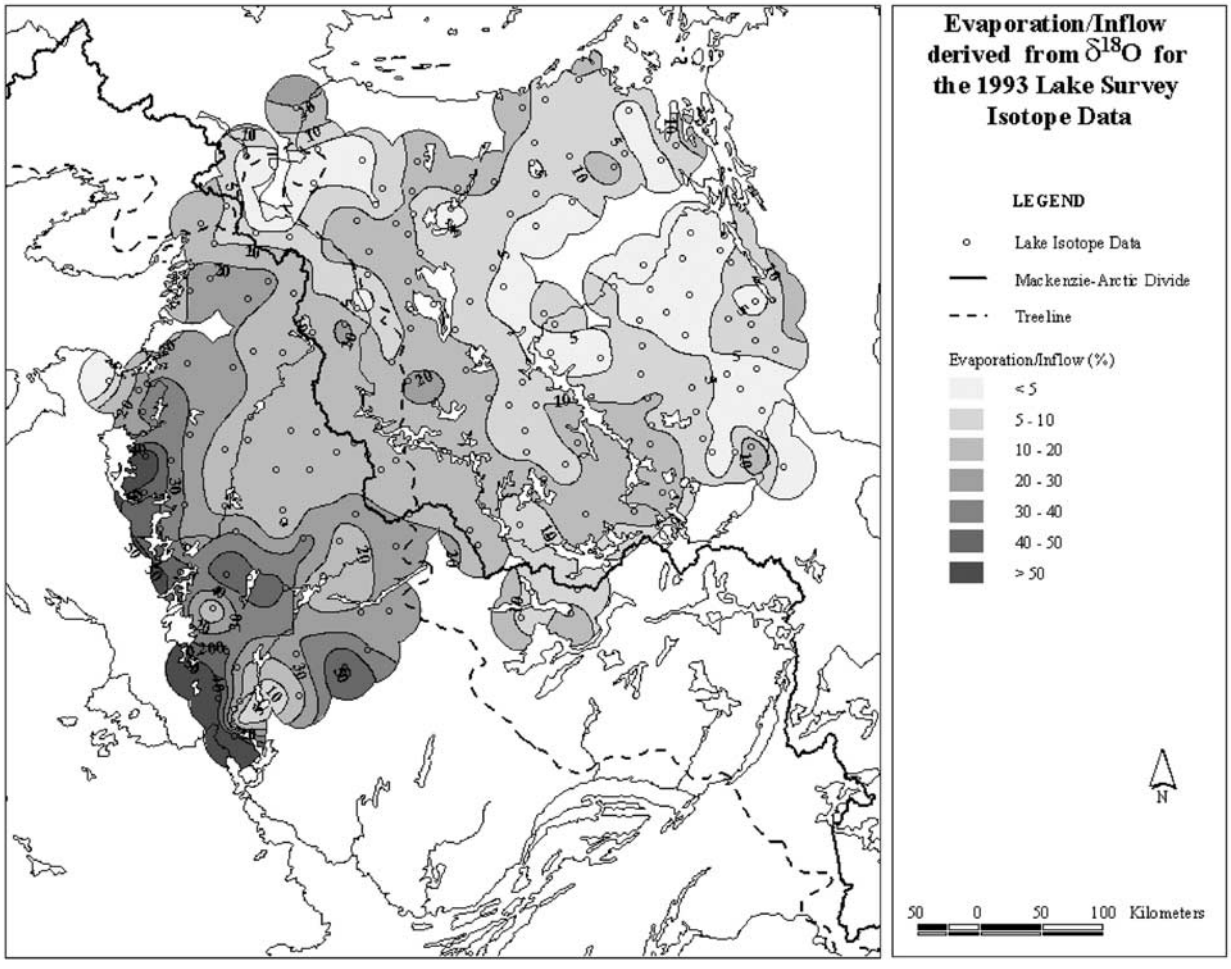

Figure 8. Map showing *E*/*I* ratios (%) for lakes based on 1993 sampling. See color version of this figure at back of this issue.

29 samples from 12 lakes, with two or three samples collected per lake, either in a vertical profile at a given site or at bottom or middepth at several different locations within the same lake. On average, the standard deviation of the $\delta^{18}O$ and $\delta^{2}H$ values for the 12 lakes was found to be 0.14 and 1.2‰, respectively, which is similar to the analytical uncertainty for each tracer. This shows that lakes are generally well mixed, consistent with results from a survey of lakes in northern Alberta involving extensive comparative sampling of epilimnion, hypolimnion, and euphotic zone waters [*Gibson et al.*, 2002]. The only significant within-lake variability (0.5 and 3.8‰ for $\delta^{18}O$ and $\delta^{2}H$, respectively) was apparent for an unusually large and irregular lake having several inflows, which was atypical for water bodies included in the present survey. Full consideration of water balance in such lakes would require much more detailed basin-specific assessment than the current regional-scale analysis permits.

[19] On the basis of extensive prior observations and modeling studies in the region [*Gibson*, 2001], seasonal evaporative isotopic enrichment is expected to be subdued in this set of lakes, affording strong potential to preserve longer-term spatial water balance information. Although analysis of isotope data over the course of the 50-day sampling campaign does reveal an apparent overall shift with time (Figure 3), the systematic relation with latitude suggests that this is primarily an artifact of the sampling strategy, imposed by logistical constraints, rather than a reflection of progressive seasonal enrichment. Nevertheless, this could be an important factor in surveys of highly responsive water bodies (especially small headwater lakes), possibly necessitating explicit compensation or detrending if sampling extended over a protracted time period.

2. Results and Discussion

2.1. Spatial Isotopic Variations

[20] Contour maps of isotopic data for lake waters and precipitation are shown in Figures 4 and 5. The most striking feature in the spatial variability of lake water

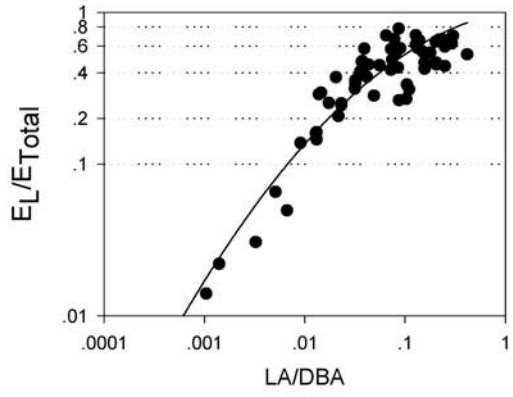

Figure 9. Relationship between lake area/drainage area ratios and E_L/E_{Total} (or fraction of water loss from sampled lake versus all lakes in catchment area) for a subset of 60 lakes. Note that LA refers to area of sampled lake only (not including upstream lakes) and all lakes can be classified as nonheadwater lakes (that is, contributing area contains appreciable open water). Note that lakes in large basins have minor influence on evaporation loss signal. Where lake area exceeds 10% of basin area, lake evaporation becomes dominant signal.

isotopic composition is the prominent northeast-to-southwest progression of increasing $\delta^{18}O$ in the 1993 data set (Figure 4a), consistent with regional ecoclimatic gradients [see *Prowse*, 1990; *Ecoregions Working Group*, 1989]. Indeed, this regional isotopic signal is remarkably distinct, in spite of localized anomalies that are likely related to lake-specific variability in depth, volume, and other factors. Nearly identical patterns over large and small scales are observed for δ^2H (not shown), as expected given the strongly linear relation between $\delta^{18}O$ and δ^2H highlighted in Figure 1. Importantly, similar regional $\delta^{18}O$ (and δ^2H) patterns are also evident in the smaller set of samples from a different selection of lakes surveyed in 1994 (Figure 4b), indicating that the observed trends are generally representative and persistent, in this case even between years characterized by an oscillation between unusually wet (1993) and rather dry (1994) conditions.

[21] The strength of the regional spatial evaporative enrichment signal is also apparent from comparison of the large-scale patterns of variability in lake water $\delta^{18}O$ and long-term amount-weighted precipitation $\delta^{18}O$ across the region (Figure 5a). Although this reveals broad resemblance in the sense of a general northward decrease in both parameters, the greater overall range and steeper latitudinal gradient of lake water $\delta^{18}O$ clearly demonstrates the overwhelming influence of evaporative isotopic enrichment, enhancing and overprinting the precipitation source signal (Figure 6). This is also clearly evident in the strong delineation of the evaporation trend in ^2H-^{18}O space, reflecting the much greater range of variability in $\delta^{18}O$ (and δ^2H) values of lake waters than of regional precipitation (Figure 1b). In the following discussions we translate these systematic variations in isotopic enrichment into quantitative measures of the fraction of water loss by evaporation and combine this information with other hydrologic data to place constraints on the principle vapor transfer mechanisms.

2.2. Model Calculations

[22] The fraction of water loss by evaporation (including all upstream contributions) was estimated based on equation (14) for each lake basin using measured lake water values to characterize δ_L and using input values taken as mean annual amount-weighted precipitation interpolated from the Global Network of Isotopes in Precipitation (GNIP) database (Figure 5a). Consistent with the model presented in Figure 1a, δ^* was evaluated using evaporation-flux-weighted (fw) exchange parameters (h^{fw}, T^{fw}, and δ_A^{fw}) where δ_A^{fw} is estimated assuming equilibrium with flux-weighted precipitation (Figure 5b) via $\delta_A^{fw} = \alpha^*(T^{fw})\delta_P^{fw} - \varepsilon^*(T^{fw})$ and $\varepsilon_K = C_K(1 - h^{fw})$. Use of the evaporation-flux-weighted approach takes into account that isotopic signals develop under conditions prevailing during times when evaporation takes place. The exactness of the model is demonstrated by good agreement between E/I (X_L) derived using both tracers for 1993 and 1994 (Figure 7), and hence good reproducibility of the evaporation line slopes is acquired without any fitting of exchange parameters, contrary to some previous studies [e.g., see *Zuber*, 1983; *Gibson et al.*, 1993].

[23] A map of E/I ratios based on oxygen-18 (Figure 8) illustrates a systematic regime of northeast-southwest gradients, with evaporation accounting for up to 60% of water losses from catchments in the extreme southwest, as opposed to only 5–15% in the northeast. This also implies that water losses by liquid outflows may vary from as little as 40% in forested areas to 85–95% in tundra catchments.

2.3. Origin of Isotopic Enrichment Signals

[24] To better understand the significance of these enrichment signals, equation (18) is applied to partition potential evaporation losses occurring in the lake from those inherited from upstream sources. A subset of 60 lakes is used for which the lake surface and drainage basin areas could be readily defined from topographic maps. As shown in Figure 9, the fraction of evaporation losses occurring from a lake itself, as compared to that lost from upstream reservoirs, ranges from ~70% in basins with moderately low LA/DBA to <1% in lakes with very low LA/DBA. Note that a high degree of correlation between changing E_L/E_{TOTAL} and LA/DBA is anticipated, given the tendency for lake area to vary with total drainage basin area, but the distribution of data in Figure 9 clearly demonstrates that the isotopic signals of evaporation are being consistently integrated and transmitted through drainage networks. Understanding where enrichment signals are being generated within a catchment gains increasing importance when evaluating isotope mass balance in complex systems, though such information is obviously less critical in basins having very low LA/DBA, such that E_L/E_{TOTAL} approaches zero [see *Gibson*, 2001].

2.4. Evaporation and Transpiration Partitioning

[25] Equation (19) is applied to estimate the fraction of water loss by evaporation as a component of the total

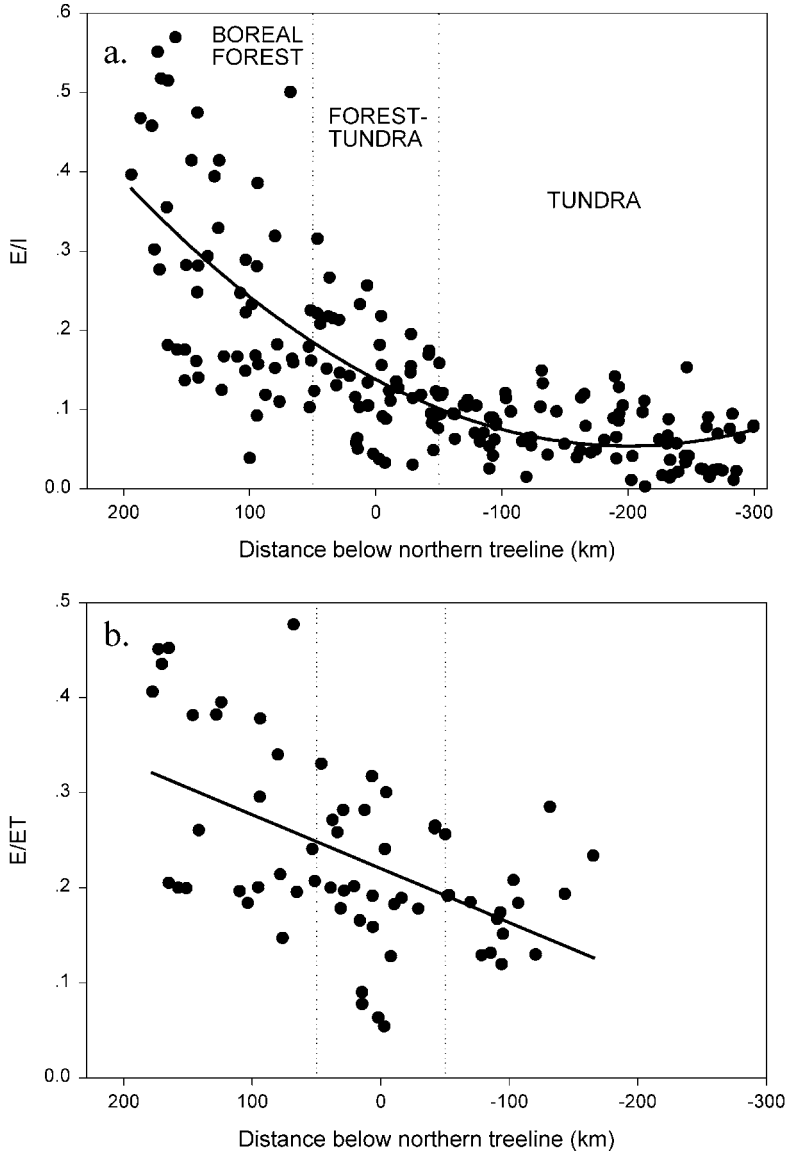

Figure 10. Southwest-northeast (boreal-tundra) variations in water balance ratios estimated by oxygen-18 mass balance. (a) Fraction of water loss by evaporation. (b) Fraction of total evapotranspiration loss by evaporation. Boreal forest, forest-tundra transition, and tundra regions are defined approximately based on distance below northern tree line.

evapotranspiration. Southwest-to-northeast trends in E/E_T for the 60-lake subset along with E/I for all lakes analyzed in 1993 are shown in Figure 10. Overall, evaporation loss from open water bodies, both as a fraction of the total outflow and as a fraction of total vapor loss, is shown to vary systematically from southwest to northeast along the ecoclimatic gradient. E/E_T ranges from >45% in forested areas of the southwest to <15% in areas of shrub tundra. One significant source of uncertainty in use of equation (19), and one that is encountered in all studies of water balance in northern regions, is the systematic underestimation of precipitation due to gauge undercatch and trace precipitation events. *Metcalfe et al.* [1994] estimate that precipitation is underestimated by roughly 30% in the region of interest, which also provides a useful indication of the range of potential error in the precipitation term in equation (19). A sensitivity analysis was conducted to examine the potential errors in E/E_T arising from a ±30% error in precipitation estimates. Uncertainty in E/E_T is found to be only half that of potential error in precipitation (approximately ±15%), which suggests that the computed ratios are meaningful and relatively robust.

[26] These results can be placed into perspective within the context of regional hydroclimatologic variability by considering Figure 11, which translates representative (modal) values for partitioning of evaporation and transpiration for catchments in the three major ecoclimatic zones in the study area into quantitative estimates of evaporation

Region Vegetation	Input	Outflows	
		Isotope Fration-ating	Isotope Non-fractionating
Northeast Arctic Tundra	P 310	E 25	T 105	Q 180
Central Near Treeline	P 325	E 40	T 160	Q 125
Southwest Boreal Forest	P 340	E 60	T 240	Q 40

Figure 11. Catchment-weighted, modal flux-partitioning summary across survey region based on isotopic tracing of E/I and E/E_T ratios. Absolute quantities are approximate and are based on climatological estimates of precipitation input, assuming no storage changes. Note that evaporation is not normally distinguished from transpiration in routine hydrological analysis, but is distinguishable using isotopes due to fractionation. Note that individual basins may vary significantly from modal water balance due to local terrain heterogeneities, especially in high evaporation regimes (see Figure 12). Modal water balance is expected to represent the integrated signature of a representative area for each terrain type.

and transpiration (and derived discharge) obtained from the calculated catchment-weighted flux partitioning, on the basis of estimated representative precipitation inputs. This summary clearly shows the steady decreases in precipitation, evaporation, and transpiration and the increase in discharge, yielding progressively decreasing E/E_T and increasing Q/P from high-boreal forest to subarctic forest-tundra to low-arctic shrub tundra. Note that these modal values are considered to be representative of the regional water balance signals, as they reflect conditions in the larger catchments that have acquired the integrated signals of smaller heterogeneous contributing areas as discussed in section 2.5. Water balance trends are broadly consistent with those summarized by *denHartog and Ferguson* [1978a, 1978b] and *Wedel* [1990], although evaporation-transpiration partitioning has not previously been analyzed.

2.5. Frequency Distribution of Water Balance Types

[27] As a corollary of these observations, an analysis of the frequency distribution of lake E/I ratios also reveals intriguing expressions of the current hydrologic regimes in the three ecoclimatic zones in the study area. As shown in Figure 12, the distribution in each zone is prominently skewed to higher values of evaporation/inflow ratio, with the modal values progressively shifting to lower E/I with higher latitude. The skewness clearly reflects, not unexpectedly, that downstream lakes tend to accumulate more water from catchments having low E/I and low variability, whereas upstream lakes can be expected to have higher E/I and higher variability; that is, as lake order increases, evaporation loss signals converge toward the modal value for that ecoclimatic zone.

[28] Perhaps less expected, but equally obvious from Figure 12, are the pronounced and highly systematic differences between the E/I frequency distributions in the three zones, which have captured distinctive ''fingerprints'' that

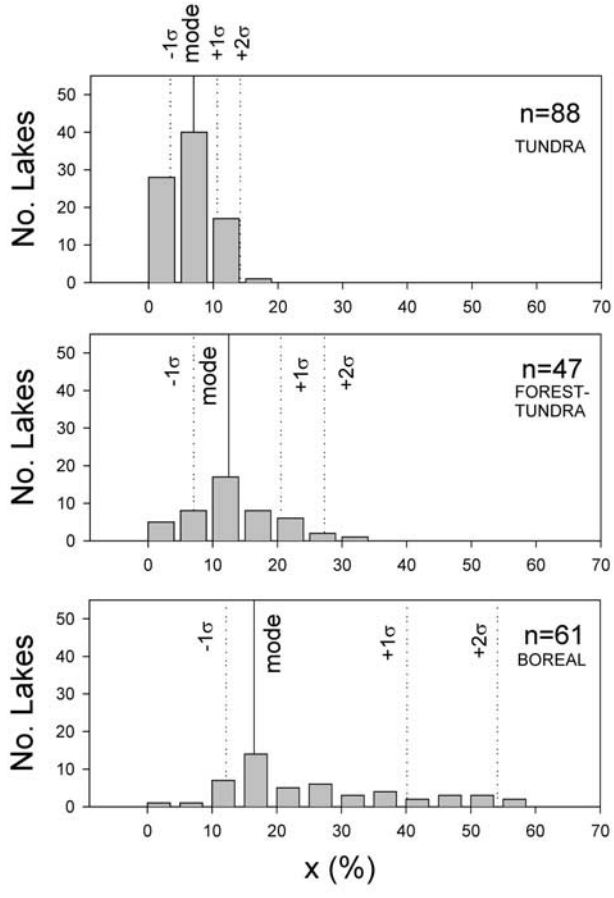

Figure 12. Frequency of lakes with specified fraction of water loss by evaporation, shown separately for each major terrain unit. Note systematic decrease in mode of x from boreal to tundra environments. Distributions are all right-skewed, and variability in evaporation loss (x) decreases progressively from boreal forest to tundra.

reflect the intrinsic structure and contemporary water balance of these areas. This ability to characterize spatial variations in landscape-scale hydroclimate may have highly significant implications, especially in light of the demonstrated sensitivity of tree line environments in this region to past climate change [see *MacDonald et al.*, 1993; *Wolfe et al.*, 1996] and its suspected sensitivity to ongoing and future changes.

3. Concluding Remarks

[29] These investigations have demonstrated that valuable additional hydroclimatic information can be garnered by coupling results from isotopic surveys of surface waters with basic conventional (i.e., nonisotopic) hydrometeorological data. This demonstration is even more convincing in light of the realization that the sampling strategy for this study was not designed with this purpose in mind and that conventional data in the region are notably sparse. On one hand, this emphasizes the potential to substantially enhance the value of existing data and monitoring activities using basic sampling methods and, on the other, provides a compelling argument for fuller integration of water isotope tracers as primary parameters in hydroclimatology field studies from their initial conception.

[30] These results particularly highlight the ability to use isotope methods for improving representation of vapor transfer processes, providing quantitative information that is difficult to produce or constrain using conventional approaches. Although our estimates of E/I and E/E_T partitioning represent only a first-order approximation at this stage, the capability of the methodology is clear.

[31] Finally, this work has led to the identification of a promising new method for characterizing the intrinsic hydroclimatic properties of landscapes within major ecoclimatic zones. Further exploration of this feature is certainly warranted to discern whether observed spatial variations along the steep hydroclimatic gradients in the study area can also serve as a guide to temporal changes in response to future climate forcing.

[32] **Acknowledgments.** Portions of this study were funded by research grants from Water Resources Division, Indian and Northern Affairs Canada, Yellowknife (Coppermine River Basin Study), the Mackenzie GEWEX Study, and NSERC research grants to J. Gibson and T. Edwards. We thank A. Hebb and S. Mrozewski for GIS support and land cover definitions, respectively, and B. Latham and W. Puznicki for graciously providing the water samples. We wish to express our gratitude to T. Prowse and J. Price for in-kind support provided during the course of this study, and to D. Milburn for informative discussions and ongoing support for isotope-based studies as part of the "full house" of hydrologic research in northern Canada. This manuscript benefited from the incisive comments of two anonymous reviewers.

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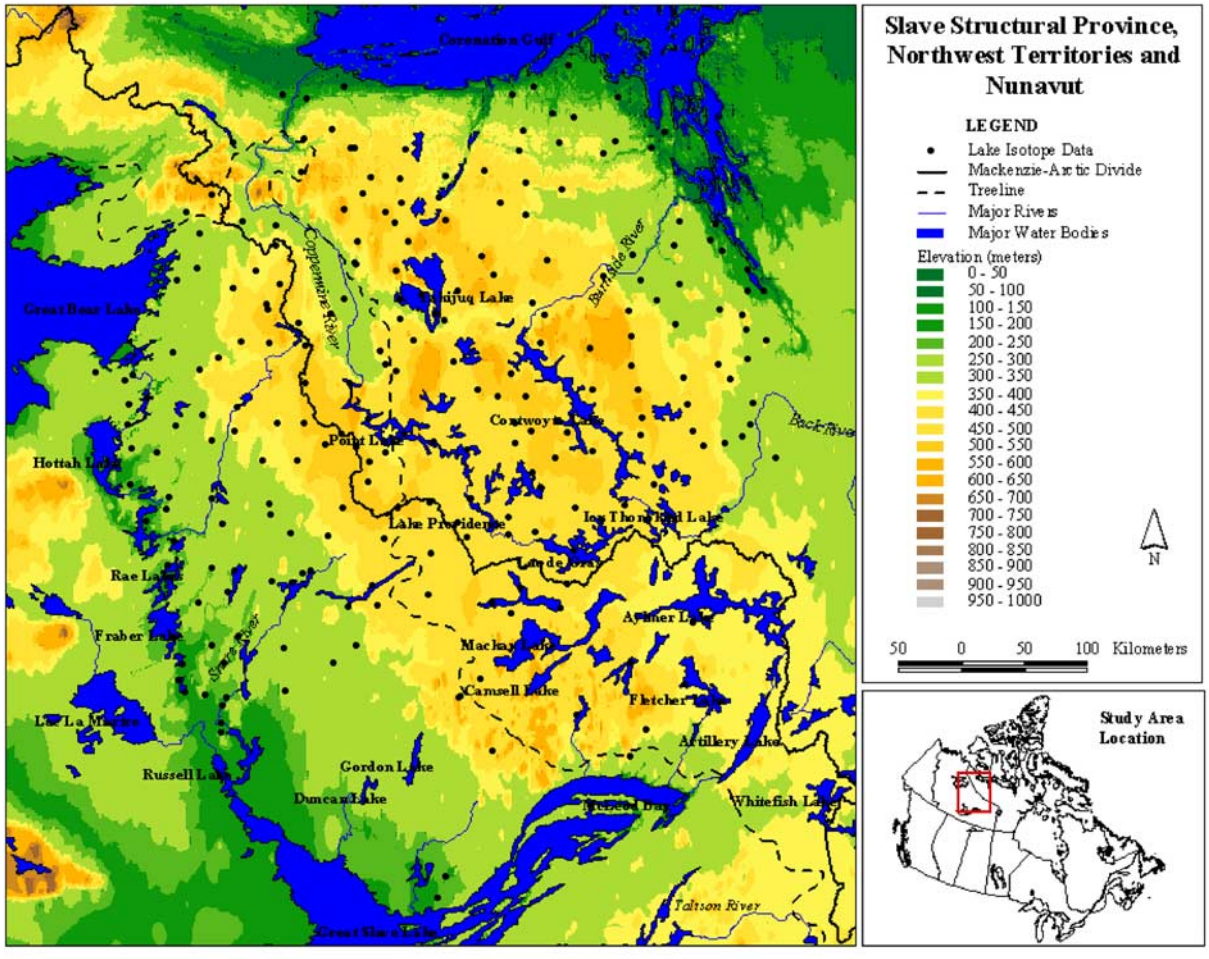
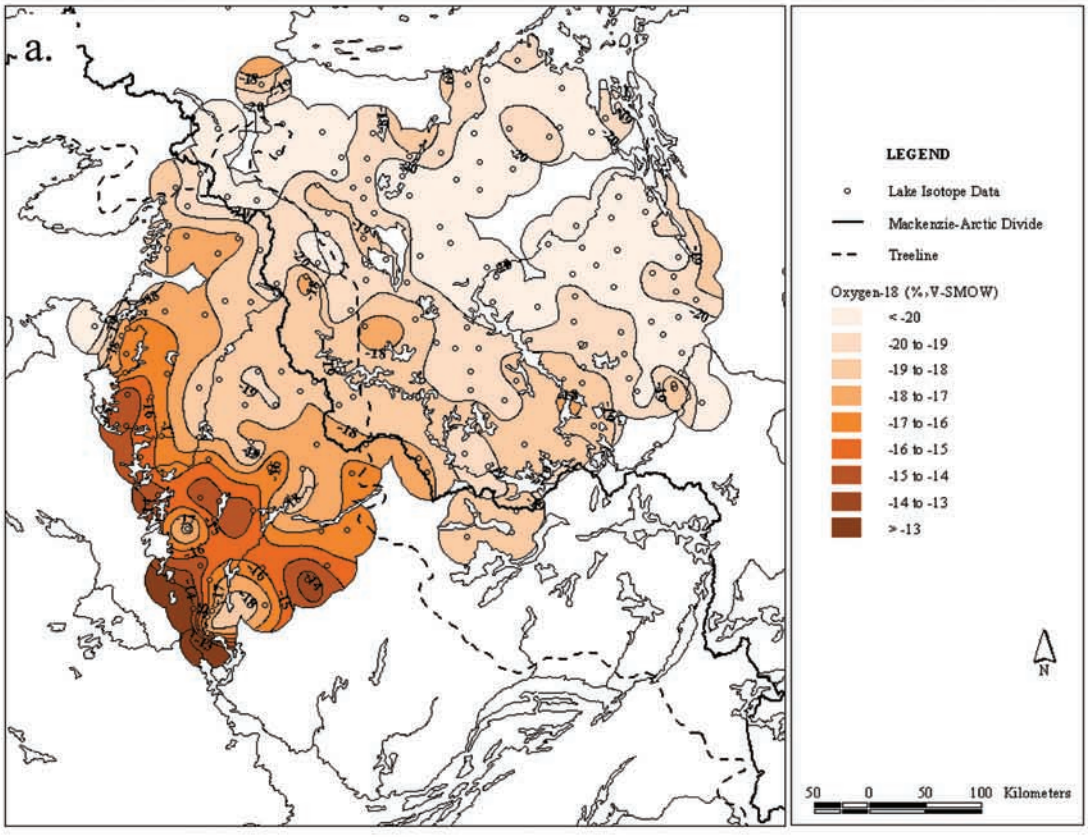
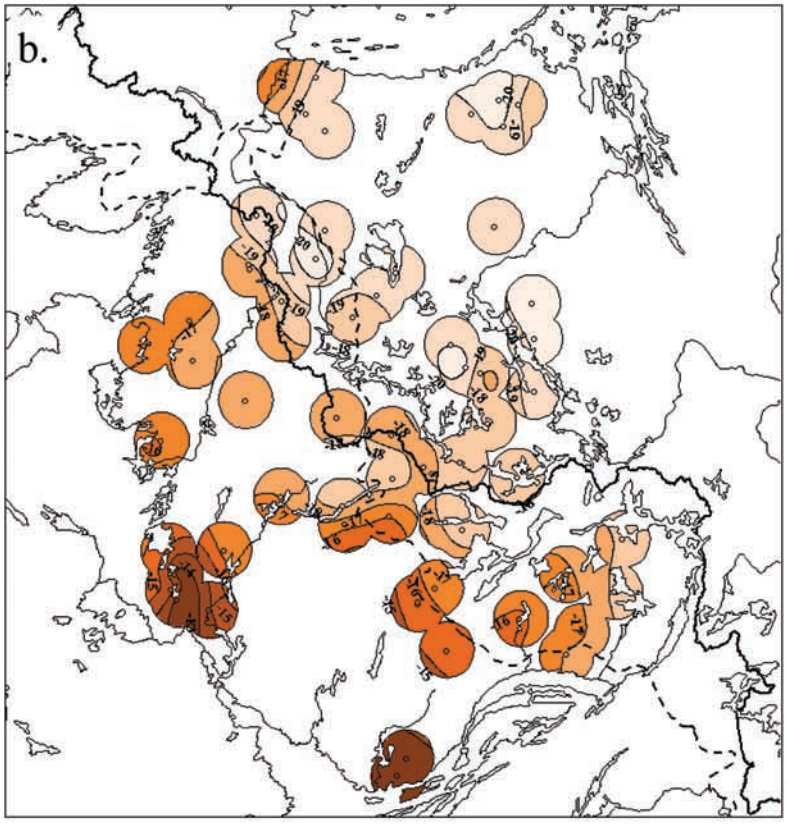


Figure 2. Relief map of central Arctic of Canada showing sampling locations for this study. Note that solid line indicates Arctic-Mackenzie drainage divide and dotted line shows approximate position of northern tree line.


a.
LEGEND
Lake Isotope Data
Mackenzie-Arctic Divide
Treeline
Oxygen-18 (‰,V-SMOW)
< -20
-20 to -19
-19 to -18
-18 to -17
-17 to -16
-16 to -15
-15 to -14
-14 to -13
> -13
N
50 0 50 100 Kilometers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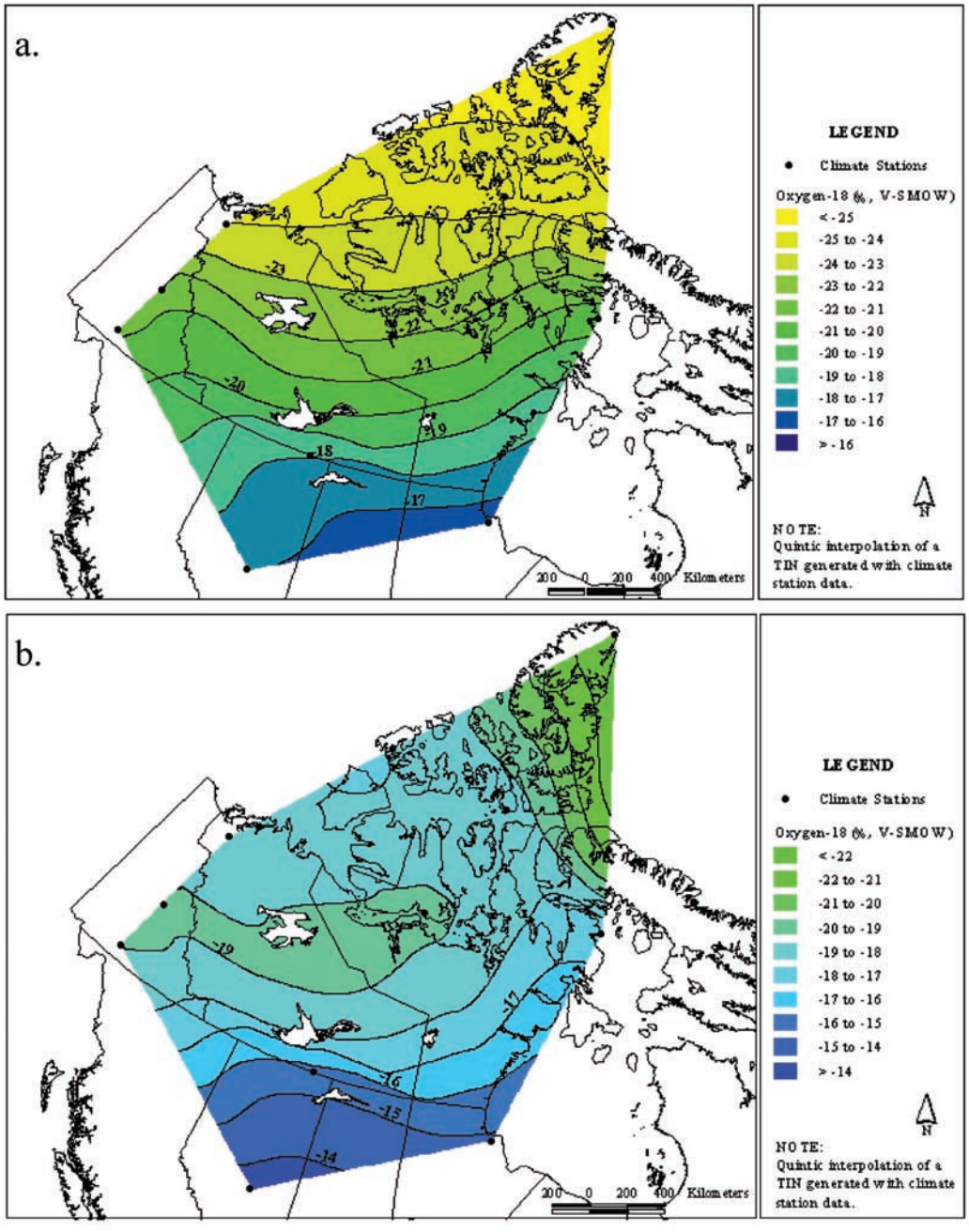

Figure 5. Map showing $\delta^{18}O$ composition of (a) mean annual precipitation and (b) evaporation-flux-weighted (thaw season) precipitation, interpolated from nearby Global Network for Isotopes in Precipitation (GNIP) stations. Similar trends are noted for $\delta^{2}H$.

Figure 4. (opposite) Map of $\delta^{18}O$ in lake water sampled during (a) 1993 and (b) 1994. A 25-km contouring radius is maintained to show only areas with high spatial sampling density. Similar trends were noted for $\delta^{2}H$ (not shown). Note that solid line indicates Arctic-Mackenzie drainage divide and dotted line shows approximate position of northern tree line.

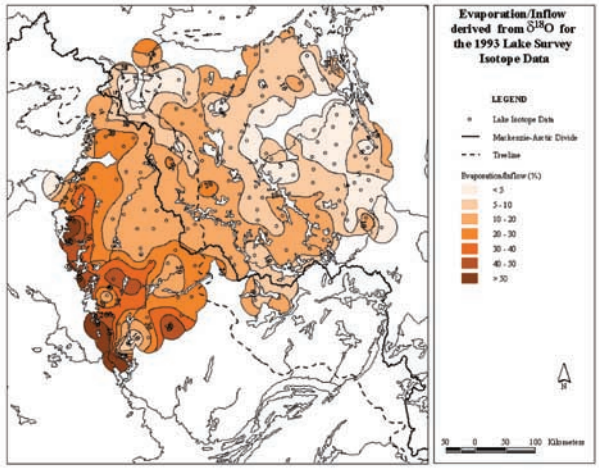

Figure 8. Map showing *E*/*I* ratios (%) for lakes based on 1993 sampling.