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Spatial and temporal variability of prairie lake hydrology as revealed using stable isotopes of hydrogen and oxygen

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Abstract

Evaporation and groundwater fluxes are thought to regulate hydrologic variability in lakes of the northern Great Plains, but little is known of how the relative importance of these processes may vary in time or space. To address this issue, we measured the isotopic composition of water (δ^{18}O, δ^{2}H) from 70 closed-basin lakes in southern Saskatchewan, Canada. All lakes occupied endorheic basins along a long gradient of salinity (0.2–115 g total dissolved solids L^{-1}). Lakes exhibited synchronous seasonal changes in salinity (synchrony, $S = 0.78$) and δ^{18}O ($S = 0.84$) during the dry summer of 2003 (~195 mm rain), whereas coherence was reduced to 0.56 and 0.22, respectively, during the wet summer of 2004 (~295 mm rain). However, despite evaporative enrichment of isotopic ratios during dry summers, hydrologic balances were regulated mainly by changes in water inflow (I) rather than evaporation (E) in both wet and dry years, with particularly strong influence of inflow (lowest E : I ratio) in dry southwestern regions. Analysis of isotopic composition also identified winter precipitation or groundwater as the most influential source of water to most lakes, despite only ~30% of annual precipitation being delivered during winter. Therefore, although seasonal variability in lake chemistry was influenced by evaporation during summer, long-term mean chemical characteristics of prairie lakes were regulated mainly by changes in winter precipitation or groundwater influx.

Lakes are abundant in the northern Great Plains (Last 1992) despite intense evaporation (Pham et al. 2008) and net precipitation deficits of 40 cm yr^{-1} to 60 cm yr^{-1} (Laird et al. 1996). In general, inter-annual variability of meteorological conditions that affect the persistence of lakes (temperature, seasonal precipitation) is regulated by the interplay between air masses arising over the Arctic, Pacific Ocean, and Gulf of Mexico (Bryson and Hare 1974), as well as global atmosphere–ocean systems such as the El Niño-Southern Oscillation (ENSO; Trenberth and Hurrell 1994), North Atlantic Oscillation (NAO; Hurrell 1995), and the Pacific Decadal Oscillation (PDO; Mantua et al. 1997). In addition, at a regional scale, the spatial and temporal variability of prairie lake hydrology is also affected by groundwater fluxes (van der Kamp and Hayashi 1998; Fritz et al. 2000) and by an unusually high supply of runoff from large catchments, most of which is derived from winter precipitation (Steppuhn 1981; Akinremi et al. 1999). However, despite these generalities, little is known of the specific conditions under which evaporation or water influx may control changes in lake chemistry and persistence, nor of the relative importance of summer and winter precipitation in the hydrologic budget of local and regional lakes. Consequently, an improved understanding of the basic hydrology of prairie lakes is needed to both forecast the effects of future climate change on lakes of the northern Great Plains, and to better interpret paleolimnological records of past climate variability in the region.

Prairie lake hydrology and chemical composition are regulated by transfers of energy and mass to lake basins (Pham et al. 2008). Inputs of energy tend to synchronize the

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seasonal variability in physical (ice melt, thermal stratification) and chemical (salinity, ionic composition) properties of lakes (Magnuson et al. 2000; Bonsal et al. 2006), whereas inputs of mass (water, dissolved substances) reduce temporal coherence among lakes (Pham et al. 2008), likely because of differences among sites in catchment characteristics that regulate mass flow (soil, geology, vegetation, topography, etc.; Blenckner 2005) and because the distribution of precipitation is more spatially variable than that of temperature (Walker et al. 1995; Whitfield et al. 2002; Quiring and Papakyriakou 2005). In the Canadian Prairies, >70% of total annual precipitation falls as summer rain; however, in many basins 80% of runoff is derived from the melting of winter snow packs during spring (Steppuhn 1981; Akinremi et al. 1999). In contrast, the influence of groundwater discharge and recharge to lakes can be difficult to predict because basins are variably connected to groundwater flow systems (Shaw and Prepas 1990; Birks and Remenda 1999), and because of differences in the chemical and hydraulic characteristics of different aquifers. For example, inputs of saline groundwater can increase ionic strength of some lakes (Keller and van der Kamp 1988; Last 1999), whereas differences in water density in lake and soil environments can also lead to diffusive losses of solutes (Sanford and Wood 1991). Furthermore, most research to date suggests that surface and intermediate groundwater systems are recharged mainly by melting snow rather than summer precipitation (van der Kamp and Maathuis 1991; Maule et al. 1994). Thus, although evaporation can have a strong effect on seasonal variations in lake chemistry (Pham et al. 2008), the persistence and overall chemical characteristics of prairie lakes may depend more strongly on the relative influence of individual water sources (summer, winter, groundwater).

In principle, analysis of stable isotopes of hydrogen (δ^2H) and oxygen (δ^{18}O) can be used to quantify the main fluxes of water to and from lake basins (Gibson et al. 1993; Gibson and Edwards 2002; Wolfe et al. 2007). In this approach, ratios of light (^{1}H, ^{16}O) and heavy isotopes (^{2}H, ^{18}O) are measured in regional precipitation (summer, winter), groundwater, and surface waters (rivers, lakes, wetlands) to determine the relative importance of source waters to the hydrology of individual basins, as well as the relative influence of evaporation and water inflow in maintaining a steady-state mass-balance of water in closed basins (reviewed in Gibson et al. 2005). Changes in isotopic ratios occur predictably during phase changes from liquid to gaseous states, and again when precipitation condenses within the atmosphere. These phase changes result in a strong, constant, linear relationship between δ^2H and δ^{18}O in global and regional precipitation (meteoric water lines; Craig 1961), which can be compared to lake waters to evaluate the importance of summer and winter precipitation to water mass-balances. In addition, evaporation of water enriches the heavy isotope content of lake water (Craig and Gordon 1965; Gonfiantini 1986), and produces a systematic deviation in δ^2H and δ^{18}O signatures (local evaporation line [LEL]) which can be used to quantify the relative importance of evaporation and water inflow to individual lake basins (Gat 1996; Gibson and Edwards 2002). However, although mass-balance studies have been conducted for many regions of North America (Gibson et al. 2005) and for prairie rivers (Ferguson et al. 2007), isotopic analyses have rarely been applied to closed-basin lakes of the northern Great Plains (Yu et al. 2002).

In this paper, we used stable isotopes of H and O in surface waters from 70 lakes to quantify the relative importance of evaporation (E) and water inflow (I) as controls of prairie lake hydrology, and to identify the predominant sources of water (summer precipitation, winter precipitation, groundwater) in each lake basin. First, isotope ratios were measured in spring, summer, and autumn in 21 lakes during years with wet or dry summers to quantify how δ^2H and δ^{18}O signatures evolved during the ice-free season, and to test the hypothesis that lakes should exhibit synchronous enrichment with heavy isotopes as a result of evaporative concentration and fractionation (Gat 1996). Second, steady-state water-balance models were used to calculate the E : I ratio for each of 70 lakes. Because most lakes lacked surface inflow, yet were located in a region with a substantial precipitation deficit, we hypothesized that the relative influence of evaporation should be greatest in relatively dry areas. Finally, we compared the δ^2H and δ^{18}O signatures of surface water to the mean amount-weighted isotopic signature of rain and snow to identify the main source of water to each lake basin. In this region, summer rainfall accounts for ~70% of total precipitation (Steppuhn 1981) but only 20% of runoff into lake basins; therefore, we hypothesized that winter precipitation regulated the overall isotopic composition and water balance of prairie lakes.

Methods

Site description—Seventy lakes of the northern Great Plains were selected to quantify the controls of spatial and temporal variability of lake hydrology using stable isotope signatures of lake water (δ^2H, δ^{18}O; Fig. 1). Study lakes occur in a subhumid (northeast) to semiarid (southwest) area of the Canadian Prairies, a region in which mean potential evaporation commonly exceeds total annual precipitation by 40–60 cm yr^{-1}, despite mean annual summer precipitation of 30–40 cm (Laird et al. 1996). Lakes were selected to lie both parallel and perpendicular to the main climate isopleths representing net precipitation deficit (i.e., total precipitation – potential evaporation).

Climate data for the study region was obtained from Environment Canada (http://www.climate.weatheroffice.ec.gc.ca) for five meteorological stations, including Saskatoon, Humboldt, Muenster, Swift Current, and Maple Creek. Calculation of mean (±SD) climate normals for these sites (1971–2000) reveals that monthly daily temperatures during the ice-free season range from 3.7 ± 1.2°C (Apr) to 15.0 ± 0.8°C (Jul), whereas monthly total precipitation ranges from 24.6 ± 2.3 mm in April to 66.7 ± 6.0 mm during June. Mean daily relative humidity based on Saskatoon and Swift Current stations is low and relatively constant among months, ranging from 43.2 ± 1.4% during May to 47.0 ± 0.6% in both June and July.

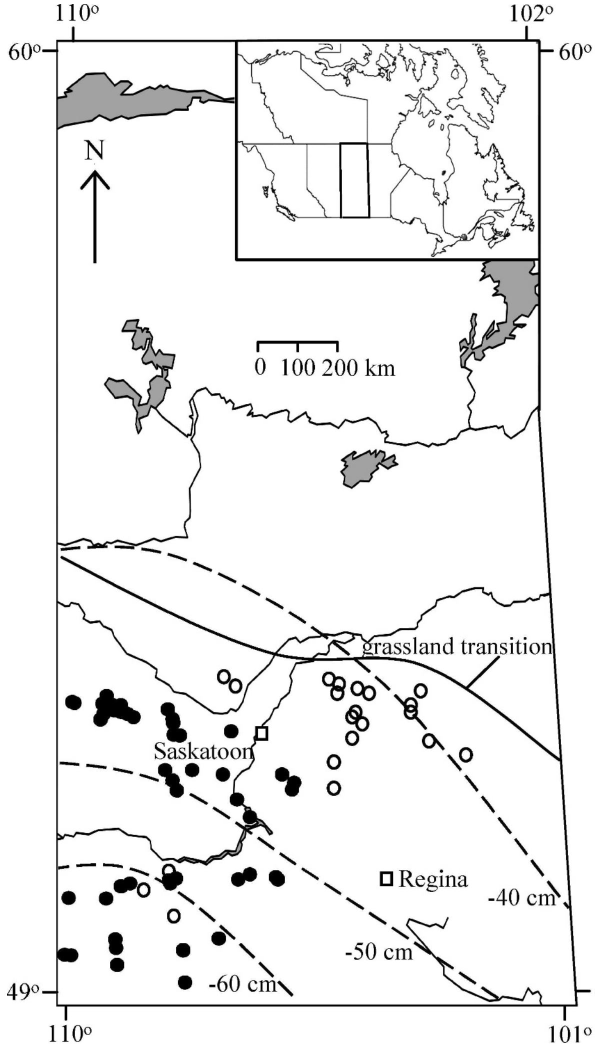

Fig. 1. Location of sampling sites in Saskatchewan, Canada (open circles = seasonal sites; filled circles = spatial survey 2004) and major cities (squares). Dashed isoclines indicate mean annual precipitation deficit (cm yr^{-1}), calculated as total precipitation minus potential evaporation (Laird et al. 1996).

All study lakes are located in topographically closed basins, with little lotic inflow and no visible outflow as streams or rivers (Table 1). Maximum lake depth varies from ~0.1 m to 30 m, whereas average summer salinity measured as g total dissolved substances (TDS) L^{-1} varies from fresh (0.2 g L^{-1}) to hyper-saline (115 g L^{-1}). In addition, lakes exhibit high nutrient concentrations characteristic of other prairie ecosystems (Patoine et al. 2006) and were composed, when known, of predominantly sulfate or carbonate brines rather than chloride solutions. Overall, most lakes were polymictic, although six sites appeared to be meromictic based on the presence of a pronounced chemocline in each sampling season (Table 1).

Physical and chemical analyses—Seasonal water balances were determined in 21 lakes by measuring lake water salinity (g TDS L^{-1}) and surface-water stable isotope composition (δ^2H, $\delta^{18}O$) three times (spring, summer, autumn) during 2003 and 2004 (Pham et al. 2008). All lakes were sampled within a 2-week interval in any given season, with collections centered on calendar day-of-year (DOY) 120 (spring), 200 (summer), and 280 (autumn). Due to missing samples, some lakes were excluded from individual analyses.

Salinity, conductivity, and power of hydrogen (pH) were determined at 1-m-depth intervals for the entire water column using a YSI model 85 probe (YSI Inc., Yellow Springs, Ohio, USA) following Pham et al. (2008). In addition, the uppermost 5 m of all lakes were surveyed for water chemistry by deploying an integrating tube sampler within the water column. Water collected was filtered through 0.45-μm-pore sterile membrane filters and analyzed for major ions sodium (Na^+), chloride (Cl^-), calcium (Ca^{++}), magnesium (Mg^{++}), sulfate ($SO_4^{=}$), bicarbonate (HCO_3^{-}), and carbonate ($CO_3^{=}$). Salinity determined by probe measurement was highly correlated with the sum of the measured ions ($r^2 = 0.884$, $p < 0.0001$; Pham et al. 2008). Other chemical determinations included dissolved organic carbon (DOC), total Kjeldahl nitrogen (TKN), and total dissolved phosphorus (TDP). All analyses followed the standard procedures of Pfaff (1993) and Stainton et al. (1977) and were conducted at the University of Alberta water chemistry laboratory.

Stable isotopes of H and O were analyzed in water collected from the uppermost 5 m of each lake. Water was filtered through a 0.45-μm-pore membrane filter immediately following sampling. Stable isotope analyses were conducted by Environment Canada, Saskatoon, Canada; $\delta^{18}O$ was measured by standard CO_2–H_2O equilibration techniques (Koehler et al. 2000), whereas δ^2H was determined by the chromium reduction decomposition method of Morrison et al. (2001). Stable isotope results were expressed in standard delta (δ) notation (‰) relative to Vienna Standard Mean Ocean Water (SMOW), where $\delta_{\mathrm{SAMPLE}} = 1000 \times (R_{\mathrm{SAMPLE}} \times R^{-1}_{\mathrm{SMOW}}) - 1$, and R is $^{18}O : ^{16}O$ or $^2H : ^1H$. As many basins had high concentrations of dissolved salts (Table 1), it was necessary to correct for the effects of salt on determinations of $\delta^{18}O$ following Sofer and Gat (1972). No correction was made for determinations of δ^2H. Salt corrections did not elicit a large change in $\delta^{18}O$ values (corrected vs. uncorrected $r^2 = 0.99$, $p < 0.0001$); therefore, salinity-corrected $\delta^{18}O$ values were used in all mass-balance equations (*see* below). Isotopic values were not corrected for the effects of elevation or latitude (Dansgaard 1964) because preliminary regression analysis revealed no significant relationship of these variables with either $\delta^{18}O$ ($r^2_{\mathrm{elevation}} = 0.037$, $p > 0.11$; $r^2_{\mathrm{evaporation}} = 0.004$, $p > 0.6$) or δ^2H ($r^2_{\mathrm{elevation}} = 0.047$, $p > 0.07$; $r^2_{\mathrm{evaporation}} = 0.019$, $p > 0.25$), respectively.

Temporal and spatial variation in lake hydrology—Previous chemical analysis of these lakes revealed that seasonal changes in lake-water salinity were synchronous among sites during dry summers, but that temporal coherence declined as a function of summer precipitation (Pham et al. 2008). In this paper, we compared these seasonal changes in salinity with those of $\delta^{18}O$ and δ^2H to evaluate how evaporative forcing and precipitation may alter the isotopic composition of lake water. In this procedure, sampling dates were first standardized to DOY 120 (spring), DOY

Table 1. Summary characteristics of study lakes. Abbreviations include Elev (elevation), cond. (conductivity), temp. (temperature), DOC (dissolved organic carbon), TKN (total Kjeldahl nitrogen), TDP (total dissolved phosphorus), and ND (not determined). Surface-water temperature measured at 1-m depth. Stratification (strat.) is designated as T (thermal), N (no stratification), or meromictic (M). Isotope values from 2004 spatial survey. $\delta^{18}O$ corrected for salinity. First 20 lakes (Antelope–Wolverine) were also sampled in each season (data not shown).

Lake	Latitude (°N)	Longitude (°W)	Vol. ($m^3\times10^6$)	Surface area (km^2)	Elev. (m)	Zmax (m)	Salinity ($g\ L^{-1}$)	Cond. ($\mu S\ cm^{-1}$)	Temp. (°C)	Strat.	pH
Antelope	50 17	108 24	62.9	13.8	700.9	5.2	10.7	17060	21.4	N	9.1
Arthur	52 34	105 26	11.5	2.9	540.5	4.0	16	19560	17.6	T, M	8.2
Charron	52 24	104 20	13.9	4.0	555.5	5.0	7	10640	18.5	N	9.0
Clair	51 59	104 03	2.0	1.2	523.8	3.0	2.2	4033	18.5	N	8.9
Deadmoose	52 19	105 10	74.6	10.9	538.5	29.9	19	26250	18.6	T, M	9.2
Edouard	52 23	104 20	2.8	1.0	580.4	5.0	0.3	625	18.7	N	8.9
Fishing	51 50	103 30	175.5	32.1	529.4	19.5	2.2	3589	19.1	N	8.6
Humboldt	52 09	105 06	70.5	19.1	543.5	6.0	1.5	2383	19.4	N	8.5
Kipabiskau	52 34	104 12	19.4	5.2	521.9	8.0	0.4	672	19.7	N	8.7
Lenore, N. basin	52 30	104 59	1.4	0.5	536.8	10.0	3	1056	19.8	N	8.4
Lewis Crk, S. basin	51 26	105 31	1.7	2.0	525.5	1.3	1.7	3062	20.3	N	8.5
Little Manitou	51 45	105 30	40.1	12.8	492.9	4.3	50.7	66800	20.2	T	8.5
Middle	52 34	105 10	8.2	5.9	533.9	4.0	33	38960	18.7	T	8.8
Rabbit	52 36	107 00	9.1	4.6	504.3	4.9	5.6	8530	19.3	N	9.4
Redberry	52 43	107 09	531.5	60.7	501.5	15.5	12.2	17480	19.2	T	8.7
Shannon	52 38	105 26	2.6	1.0	548.9	7.3	2.4	3986	18.9	N	8.9
Success	50 29	108 01	3.3	0.7	715.4	20.0	28	26130	19	T, M	9.9
Wakaw	52 40	105 35	36.0	10.7	510.5	10.0	2.2	3488	19.5	N	9.5
Waldsea	52 17	105 12	37.2	4.7	533.1	12.9	17.5	22650	18.3	T, M	8.7
Wolverine	52 01	105 14	3.4	1.4	566.3	1.5	4.4	8670	18	N	8.4
Alioth	52 30	109 33	ND	ND	642	0.09	8.6	16300	30	N	8.6
Aroma	52 15	108 33	ND	ND	630	0.42	41.6	60600	23.9	N	8.6
Bitter	50 06	109 47	ND	ND	733	0.12	25.5	35570	19.4	N	8.9
Blackstrap	51 44	106 28	ND	ND	570	7.4	0.3	471	19.2	N	8.5
Brightwater	51 36	106 32	ND	ND	549	5.9	0.2	390	20.2	N	8.2
Coldspring	52 20	108 34	ND	ND	642	0.36	20.6	31030	25.1	N	8.7
Cutbank	50 32	106 59	ND	ND	700	2.12	2.5	415	19.2	N	10.0
Cypress	49 28	109 28	ND	ND	969	2.8	0.5	861	17.3	N	8.5
Dewey	50 35	107 03	ND	ND	712	1.6	12.3	17920	18.2	N	9.5
Eaglehill	52 22	109 00	ND	ND	636	0.17	11.7	20510	27.3	N	9.5
Elkwater	49 40	110 17	ND	ND	1228	8.5	0.2	451	18.8	N	8.8
Farewell	52 28	109 22	ND	ND	580	0.07	2.2	4340	28.5	N	9.1
Freefight	50 24	109 06	ND	ND	1224	16	32.3	45890	21.4	T, M	8.8
Gull	50 06	108 29	ND	ND	660	5	1.5	2595	19	N	8.5
Handsome	50 29	107 23	ND	ND	754	1.3	2.6	4107	17.9	N	9.0
Harris	49 48	109 30	ND	ND	851	4.1	0.3	499	19.1	N	8.9
Hazlet	50 24	108 38	ND	ND	717	3	1.4	2354	18.2	N	8.6
Ingebrigt	50 21	109 21	ND	ND	710	4.8	3.9	6610	21.8	ND	9.0
Island	49 07	108 13	ND	ND	874	2.6	0.2	320	19.9	ND	10.2
Junction	49 58	109 31	ND	ND	860	8.6	0.8	1495	19.8	N	8.8
L. Manito	52 36	109 35	ND	ND	601	1.68	10.4	19140	29.2	ND	9.6
Lonetree	50 29	106 56	ND	ND	576	3	3.3	5260	18.2	N	9.6
Luck	51 03	107 03	ND	ND	577	0.43	17.1	24950	19.9	N	9.1
Macklin	52 19	109 56	ND	ND	639	1.54	1	1877	22.5	N	9.1
Manito	52 47	109 46	ND	ND	591	5.71	24.9	40420	26.6	T	8.6
Mason	50 21	109 22	ND	ND	689	0.96	7.7	12390	21.4	N	8.2
McLean	52 36	109 22	ND	ND	613	0.09	26.3	47050	29.6	N	9.1
Milden	51 24	107 25	ND	ND	590	0.25	8.4	14760	25.7	N	8.6
Notokeu	49 45	108 16	ND	ND	711	1.8	2.7	4287	17.4	N	9.2
Opuntia	51 47	108 34	ND	ND	606	0.24	8.3	15370	29.5	N	9.2
Pellitier	49 59	107 56	9.7	2.9	825.0	8.5	0.4	654	18.6	N	8.8
Pike	51 54	106 49	ND	ND	478	2.2	0.2	403	22	N	8.6
Reesor	49 39	110 06	ND	ND	756	6.3	0.2	285	20.1	N	8.7
Reflex	52 39	110 00	ND	ND	589	1.9	115.8	137800	25	N	9.3
Richmond	51 59	108 01	ND	ND	662	0.43	37.7	54800	23.3	N	8.8
Shrimp	51 35	108 37	ND	ND	680	2.4	0.5	851	21.1	N	8.8
Sink	52 26	109 11	ND	ND	624	1.86	4.5	8430	26.7	N	8.8
Snakehold	50 30	108 28	ND	ND	858	3	52.7	70600	21.4	ND	5.2

Table 1. Extended.

HCO_3 (mg L^{-1})	DOC (mg L^{-1})	Cl (mg L^{-1})	SO_4 (mg L^{-1})	Na (mg L^{-1})	K (mg L^{-1})	Ca (mg L^{-1})	Mg (mg L^{-1})	TKN (μg L^{-1})	TDP (μg L^{-1})	δ^2H (‰)	δ^{18}O (‰)
914.4	68	524	12478	3670	265	14	1650	4062	250	−79	−7.3
190.0	61	467	33265	3240	370	49	4360	3681	253	−85.9	−5.3
802.3	39	424	8689	1490	210	46	1630	2491	131	−66.6	−5.3
ND	ND	79	2511	584	63	126	285	1722	103	−99.6	−8.3
134.6	41	7136	14673	5950	303	48	1900	2607	62	−74.6	−6.7
ND	ND	6	209	30	22	35	59	1983	20	−84.3	−6.5
461.0	19	64	2366	344	76	98	406	1187	16	−67.8	−6.8
368.2	21	176	1271	199	64	104	240	1474	111	ND	−8.0
ND	ND	8	201	26	11	47	50	742	9	−95.8	−8.8
251.2	35	215	4670	748	169	40	900	2233	28	−84.6	−7.8
242.5	32	71	1889	307	61	142	264	1907	610	−93.5	−8.5
167.5	159	19009	49690	16500	905	48	10950	6337	437	−85.7	−4.5
1365.7	143	2760	51587	9990	768	25	9450	8009	324	−77.8	−4.6
108.2	36	157	7296	1250	134	33	1230	2214	57	−80.9	−6.5
412.1	39	273	17584	2580	230	18	3020	1973	35	−65.3	−5.7
323.3	23	51	2936	271	110	59	640	1465	40	−66.6	−6.3
1771.4	53	1286	20182	6900	406	5	2400	2951	84	−82.7	−6.1
3001.7	16	79	1261	336	36	220	356	934	15	−83.3	−7.6
402.3	31	4973	20214	3860	317	325	3360	2027	45	−100	−6.8
1158.9	44	292	20214	739	173	189	900	2864	68	−82	−4.7
412.3	60	1467	27277	7780	244	50	1730	5123	681	−97.7	−9.3
1259.6	268	7071	64670	14700	1140	26	10080	8933	555	−85.5	−6.5
540.2	59	3724	25455	10825	390	5	1420	3010	713	−73.4	−5.3
495.4	10	11	129	39	6	37	26	707	17	−114.8	−12.8
158.6	8	9	90	30	4	35	19	406	12	−126.2	−14.8
791.0	219	1848	24069	7530	372	14	3020	9564	900	−91.5	−8.0
117.8	36	55	2265	930	53	25	242	2287	63	−63.4	−5.7
311.6	19	9	204	106	19	44	61	1604	567	−86.4	−7.7
166.8	50	482	26364	3570	283	23	2330	3255	37	−90.7	−7.7
729.4	158	2967	7573	4830	314	19	298	9786	409	−82.9	−5.4
312.5	12	28	31	23	5	38	38	658	9	−91.3	−9.7
750.1	43	136	830	871	24	14	17	6338	1991	−95.4	−9.0
812.4	212	9363	92923	9580	2010	16	7840	5317	329	−81.9	−5.1
227.4	23	88	1261	466	35	97	212	1887	433	−107.5	−11.2
513.7	45	34	2163	930	73	72	182	3138	937	−87.2	−7.0
402.7	14	6	145	54	7	29	27	865	52	−125.6	−14.2
271.6	23	32	1098	336	21	90	202	1821	464	−116.3	−10.8
892.5	40	295	2891	1530	18	21	272	2343	260	−100.9	−9.6
39.7	19	3	15	8	19	26	16	1504	103	−88.9	−10.3
262.6	17	42	618	200	16	72	78	1243	389	−113.7	−11.3
2494.7	341	976	7122	5090	187	1	232	3907	960	−75.7	−3.1
387.1	41	76	2689	1230	80	12	238	2823	67	−77.7	−6.0
227.6	91	865	32568	5190	400	33	3400	6345	204	−80.3	−6.8
692.4	48	69	377	416	39	25	52	3538	944	−98.7	−8.0
657.8	130	660	32883	13500	318	21	1050	9081	381	−85.8	−3.3
181.9	54	3304	19153	12350	326	2	495	3142	354	−108.6	−7.1
1131.7	155	651	6589	3020	42	23	532	4140	346	−82.4	−7.0
342.4	74	289	18837	1850	255	263	2100	4939	205	−87.6	−6.0
984.2	30	42	1704	1210	41	21	83	ND	1182	−88.2	−7.9
571.3	50	693	7013	3500	120	41	416	3948	214	−64.4	−3.6
353.6	12	21	181	88	17	22	76	852	16	−85.2	−9.0
180.7	12	19	50	38	5	21	28	854	10	−97.5	−8.8
491.5	6	1	17	4	2	43	14	410	10	−98.8	−10.7
5888.8	818	64038	6788	47400	273	1	151	2286	18	−87.1	−2.8
690.0	293	1204	61961	17300	840	17	5390	6174	287	−90.1	−6.6
206.6	27	23	350	66	40	46	71	1965	52	−94	−8.9
1291.9	68	708	2465	2020	72	36	254	6352	4454	−88	−6.5
136.3	402	14043	77307	18400	934	87	12400	2067	1408	−86.4	−6.0

Table 1. Continued.

Lake	Latitude (°N)	Longitude (°W)	Vol. ($m^3 \times 10^6$)	Surface area (km^2)	Elev. (m)	Zmax (m)	Salinity ($g\ L^{-1}$)	Cond. ($\mu S\ cm^{-1}$)	Temp. (°C)	Strat.	pH
Springwater	52 00	108 22	ND	ND	699	0.96	16.7	25990	22.6	N	9.1
Sufferin	52 38	109 53	ND	ND	625	5.9	0.2	441	23.4	N	8.6
Sunny	52 01	108 32	ND	ND	673	15	10	15630	20.9	T, M	9.0
Thackeray	52 31	108 42	ND	ND	644	0.37	2.3	4198	23.7	N	8.6
Tramping	51 59	108 47	ND	ND	649	0.4	41.3	54900	19.6	N	9.3
Vincent	50 13	108 57	ND	ND	736	0.125	106.6	108200	24.7	N	8.3
X1_AB	50 07	110 1	ND	ND	789	0.135	52	77300	26.5	N	8.9
X2	52 05	107 35	ND	ND	559	0.65	5.6	9670	23	N	8.2
X3	51 54	107 39	ND	ND	614	0.19	52	80600	29.4	N	8.4
X4	52 26	109 51	ND	ND	636	0.09	5.2	10250	29.1	N	9.2
X5	52 40	110 04	ND	ND	606	4.2	13	21370	24.5	ND	9.2
X6	52 34	109 47	ND	ND	645	0.26	72.9	102400	26.1	N	8.3

200 (summer), and DOY 280 (autumn), before Z-transforming annual time series for each lake, and calculating lake synchrony (S) as the mean pair-wise Pearson correlation coefficient among all lake pairs (Patoine and Leavitt 2006). This coefficient S has a value of 1.0 if changes in lake chemistry are perfectly synchronous, and 0.0 if lakes vary in an independent fashion. Strong temporal coherence of seasonal changes in lake chemistry suggests that lakes vary in response to large-scale phenomena such as climate or land use (Patoine and Leavitt 2006), and that detailed analysis of individual sentinel ecosystems (e.g., paleoclimate reconstructions) can provide insight into broad-scale regulatory processes.

We extended our seasonal lake survey to a total of 70 lakes during the summer of 2004 to gain a better understanding of the spatial variability of water sources to prairie lakes (Table 1). Surface waters of lakes were sampled during 2 weeks of summer (DOY ~190) for stable isotopes and water chemistry as described above. The isotopic values from these 70 lakes were also used in an isotope mass-balance model to estimate ratios of evaporation to inflow (E : I) and to identify the main source of water (summer, winter) to each lake based on the perpendicular distance to the LEL (Wolfe et al. 2007; *see* below). This spatial survey encompassed substantial variability in lake water salinity (~0.2 to 115 g TDS L^{-1}) and stable isotopes of $\delta^{18}O$ (−15 to −2‰) and $\delta^{2}H$ (−127 to −61‰) and allowed us to place our seasonal surveys into a broad geographic context (>100,000 km^2).

Isotope hydrology—Stable isotope signatures of surface waters were compared to known local meteoric inputs and local evaporative effects for the Canadian Prairies to evaluate controls of the hydrology of individual lakes (Fig. 2a). Prior analysis of isotopes in regional precipitation collected during 1990–2005 (Fig. 2a; Environment Canada unpubl.) revealed that precipitation samples were generally clustered along a Local Meteoric Water Line (LMWL; $\delta^{2}H = 7.7 \times \delta^{18}O - 1.2$), which was very similar to the Global Meteoric Water Line (GMWL; $\delta^{2}H = 8 \times \delta^{18}O + 10$) described by Craig (1961). The LMWL is defined by the amount-weighted estimates of the isotopic composition ($\delta^{2}H$, $\delta^{18}O$) of summer ($\delta_{\mathrm{Psummer}}$) and winter precipitation ($\delta_{\mathrm{Pwinter}}$), where summer precipitation is enriched in heavy isotopes relative to winter samples. Similarly, isotope analysis of groundwater from 129 shallow (<100-m deep) wells within the survey area was obtained as part of the 2005 Saskatchewan Research Council monitoring well network and was used to demonstrate that the mean (±SD) regional signature of groundwater ($\delta^{18}O_{\mathrm{groundwater}} = -18.3 \pm 2.0‰$, $\delta^{2}H_{\mathrm{groundwater}} = -146 \pm 14‰$) was similar to that observed in other studies ($\delta^{18}O_{\mathrm{groundwater}} = -15$ to $-20‰$, $\delta^{2}H_{\mathrm{groundwater}} = -135$ to $-175‰$) of the Canadian Prairies (Kelley and Holmden 2001). Both precipitation and groundwater data are available from Environment Canada on request.

In principle, analysis of the isotopic composition of lake water samples relative to the LMWL can be used to quantify both the relative importance of evaporation and water inflow to individual lake hydrology, and the relative importance of summer and winter precipitation to mass-balances of lake water (Gat and Levy 1978; Gibson 2001; Gibson and Edwards 2002). In both cases, this analysis assumes that an isolated lake basin is in steady state, and that the main hydrologic exchanges arise from evaporation and precipitation. The procedure then requires the calculation of the theoretical LEL, defined as the trajectory that an individual lake would exhibit through $\delta^{18}O - \delta^{2}H$ space if the lake experienced evaporative loss of water (Fig. 2a). The theoretical LEL is defined by three parameters: δ_P, the mean amount-weighted isotopic composition of precipitation; δ_{SS}, the isotopic composition of lake waters in a closed-basin in which evaporation equals inflow (E : I = 1) from precipitation δ_P under isotopic steady state, and; δ^*, the theoretical maximum (limiting) isotopic enrichment of water in a desiccating basin. The theoretical LEL is plotted as a straight line from δ_P through δ_{SS} to δ^* and represents the isotopic evolution of water in a closed basin lake with inputs of average amount-weighted precipitation (δ_P) evaporating towards δ^* (Fig. 2a). In the northern Great Plains, deviations from LEL arise both from disproportionate inputs of seasonal precipitation (Wolfe et al. 2007; Yi et al. 2008), and from groundwater influx which reduces the intercept of LMWL and LEL from δ_P towards $\delta_{\mathrm{groundwater}}$, and increases the slope of the observed relationship between $\delta^{2}H$ and $\delta^{18}O$ relative to the theoretical LEL (Yu et al. 2002; Gibson et al. 2005).

Table 1. Extended continued.

HCO_3 (mg L^{-1})	DOC (mg L^{-1})	Cl (mg L^{-1})	SO_4 (mg L^{-1})	Na (mg L^{-1})	K (mg L^{-1})	Ca (mg L^{-1})	Mg (mg L^{-1})	TKN (μg L^{-1})	TDP (μg L^{-1})	δ^2H (‰)	δ^{18}O (‰)
370.0	188	584	4040	5090	444	12	3500	7099	131	−79.8	−5.6
276.6	18	5	32	39	5	22	40	1149	12	−77.8	−6.3
891.6	54	353	22289	2840	236	13	2010	3046	103	−81.8	−5.8
584.7	47	56	1903	784	52	73	199	3982	344	−98.7	−9.2
2172.7	604	3783	35438	17400	971	3	1480	61	865	−77.6	−3.4
940.3	185	16187	53524	24000	532	658	4220	10292	1455	−79.2	−5.3
426.5	69	5010	84601	28000	544	39	1640	4880	181	−76.6	−4.5
345.9	50	85	7166	1120	135	381	1140	3215	51	−87.6	−7.1
666.1	337	1358	89644	18100	642	398	7580	3816	375	−83.3	−5.8
800.6	348	1764	6229	5500	558	1	27	5227	2463	−92.6	−4.9
1301.6	156	1260	12195	4920	329	3	1280	5132	211	−75.9	−4.2
657.7	241	3026	121724	43100	814	46	6460	10071	220	−86.6	−5.9

Although highly seasonal patterns of evaporation can also increase observed slopes of LEL relative to theoretical expectations (Gibson et al. 2008), the relative humidity at our sites during April to September was too low (monthly mean <48%) to greatly bias LEL determinations.

In our analysis, δ_P was derived directly from measurements of precipitation during 1990–2005 (Environment Canada unpubl.), whereas δ^* was derived using the equation of Gat and Levy (1978; Gat 1981),

$$\delta^*_{(\delta18O\text{ or }\delta2H)} = [h(\delta_A) + \varepsilon] \times \left[h - \left(10^{-3} \times \varepsilon\right)\right]^{-1} \quad (1)$$

in which h is evaporation flux-weighted mean annual humidity derived from long-term climate normals (Environment Canada, http://www.climate.weatheroffice.ec.gc.ca) of humidity and evaporation (R. Warren, Agriculture and Agri-Food Canada, Regina, Canada pers. comm.), δ_A is a measure of atmospheric moisture that is assumed to be in equilibrium with δ_P (Gat and Bowser 1991) as

$$\delta_A = \delta_P - \varepsilon^* \quad (2)$$

and ε is the total isotopic separation factor calculated as

$$\varepsilon_{(\delta18O\text{ or }\delta2H)} = \varepsilon^* + \varepsilon k \quad (3)$$

In both Eq. 2 and 3, ε* is the equilibrium fractionation effect that describes mass-dependant partitioning between ${}^1H_2^{16}O$, ${}^1H\,{}^2H^{16}O$, and ${}^1H_2^{18}O$ (Horita and Wesolowski 1994) and is a value that is not significantly different from those reported by Majoube (1971) for temperatures ranging from 0°C to 25°C. Thus ε* is estimated empirically from

$$\begin{aligned}\varepsilon^*_{(\delta18O)} = 10^3\ln\alpha = {} & -7.685 + \left[6.7123 \times \left(10^3 \times T^{-1}\right)\right] \\ & - \left[1.6664 \times \left(10^6 \times T^{-2}\right)\right] \\ & + 0.3504 \times \left(10^9 \times T^{-3}\right)\end{aligned} \quad (4)$$

and,

$$\begin{aligned}\varepsilon^*_{(\delta2H)} = 10^3\ln\alpha = {} & 1158.8 \times \left(T^3 \times 10^{-9}\right) \\ & - 1620.1 \times \left(T^2 \times 10^{-6}\right) + 794.84 \\ & \times \left(T \times 10^{-3}\right) - 161.04 + 2.9992 \\ & \times \left(10^9 \times T^{-3}\right)\end{aligned} \quad (5)$$

where T is evaporation flux-weighted temperature in Kelvin (K) obtained from Environment Canada long-term climate data (http://www.climate.weatheroffice.ec.gc.ca). In contrast, εk in Eq. 3 is the equilibrium kinetic effect that describes the differences in molecular diffusivities of isotopes, and which varies as a function of the relative humidity deficit of typical lake evaporation. We estimated εk using the equations derived by Gonfiantini (1986) and Araguas-Araguas (2000);

$$\varepsilon k_{(\delta2H)} = 12.5 \times (1 - h) \quad (6)$$

and

$$\varepsilon k_{(\delta18O)} = 14.2 \times (1 - h) \quad (7)$$

Finally, the parameter δ_{SS}, which represents the expected isotopic composition of a closed basin fed by δ_P in isotopic steady state (e.g., E : I = 1) was determined by

$$\delta_{SS(\delta18O\text{ or }\delta2H)} = [(\delta_P + (m \times \delta^*)] \times (1 + m)^{-1} \quad (8)$$

where

$$\begin{aligned}m_{(\delta18O\text{ or }\delta2H)} = {} & \left[h - 10^{-3}(\varepsilon)\right] \\ & \times \left\{1 - h + \left[10^{-3}(\varepsilon k)\right]\right\}^{-1}\end{aligned} \quad (9)$$

In this calculation, m is the enrichment slope defined by Welhan and Fritz (1977) and Allison and Leaney (1982).

Isotope mass-balance—To determine the relative influence of evaporation and water inflows on the water balance of individual lakes, we calculated E : I ratios using the surface-water δ^2H and δ^{18}O values observed in the 2004 spatial survey, local climatic parameters (humidity, temperature, evaporation), and experimentally determined fractionation factors described in the previous section. In this calculation, we assumed that all lakes receive water only from precipitation (closed basins), that each lake is well mixed, and that all sites are in isotopic steady state (Gibson and Edwards 2002). With these assumptions, the E : I ratio was calculated for each lake using both δ^{18}O and δ^2H as,

$$E : I_{(\delta18O\text{ or }\delta2H)} = (\delta_I - \delta_L) \times (\delta_E - \delta_L)^{-1} \quad (10)$$

where δ_I represents the isotopic value of input to each lake basin and was estimated as δ_P, δ_L is the isotopic value of surface waters at each site, and δ_E is the isotopic composition of evaporation which cannot be directly measured. Fortunately, δ_E can be estimated using a robust linear resistance model of Craig and Gordon (1965) which quantifies the isotopic behavior of surface waters evaporating under normal conditions. This linear resistance model takes into account both equilibrium fractionation effects and kinetic fractionation effects, and is most appropriate for regions with either low relative humidity during summer (<50%), or low seasonal variation in evaporation (Gibson et al. 2008). The equation for δ_E is reformatted from Craig and Gordon (1965) by Gibson and Edwards (2002) and Gibson et al. (2002) as:

$$\delta_E = \left\{\left[1 + 10^{-3}(\varepsilon^*)\right] \times \left[\delta_L - h\delta_A - \varepsilon\right]\right\} \times \left[1 - h + 10^{-3}(\varepsilon k)\right]^{-1} \tag{11}$$

and, with substitution of Eq. 10 and 11, yields

$$E : I_{(\delta 18O \text{ or } \delta 2H)} = \left\{\left[1 - h + 10^{-3}(\varepsilon k)\right] \times \left[h - 10^{-3}(\varepsilon)\right]^{-1}\right\} \times \left[(\delta_L - \delta_I) \times (\delta^* - \delta_L)^{-1}\right] \tag{12}$$

When calculated as above, E : I is unity when a lake exhibits steady-state evaporation (E = I), whereas a ratio greater than one indicates a basin desiccating towards its theoretical limiting isotopic enrichment, and E : I < 1.0 indicates a basin with inputs that exceed evaporative losses (Fig. 2b). Although it is possible to derive E : I estimates using either δ^{18}O or δ^2H, we used δ^{18}O-based estimates because they are more commonly used by other researchers (Gibson et al. 2005), and because there was good correlation between δ^{18}O- and δ^2H-derived estimates of E : I ($r = 0.7$, $p < 0.001$). In prairie lakes, addition of isotopically depleted groundwater will lead to an underestimate of E : I when estimated relative to the theoretical LEL, but will not alter the general conclusions concerning the relative importance of evaporation and water influx as controls of overall lake hydrology.

We approximated the relative influence of summer precipitation vs. winter precipitation by determining the perpendicular distance of each basin to the theoretical LEL using the 2004 spatial survey data and following the procedures of Wolfe et al. (2007). In lakes without groundwater, positive values represent conditions in which surface waters are more similar to the isotopic values of summer rainfall ($\delta_{Psummer}$) than would be expected from the amount-weighed input of precipitation, whereas negative values suggest that lake water is more characteristic of winter precipitation ($\delta_{Pwinter}$; Fig. 2b). In instances where

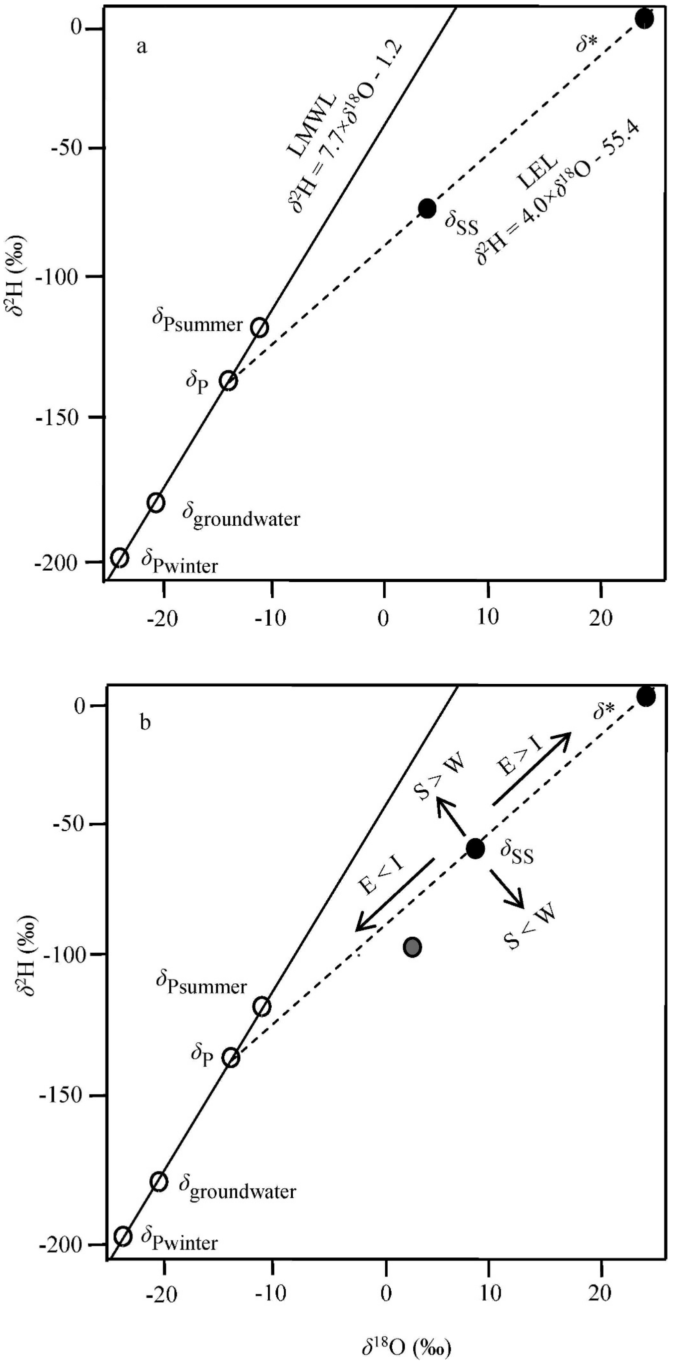

Fig. 2. (a) Isotopic water-balance model identifying a Saskatoon, Saskatchewan, Canada–based Local Meteoric Water Line (solid line; LMWL) and theoretical Local Evaporation Line (dashed line; LEL). Data for LEL were based on Saskatoon long-term (1971–2000) climatic parameters. LMWL is defined based on mean annual amount-weighted isotopic signatures of total (δ_P), summer ($\delta_{Psummer}$), and winter precipitation ($\delta_{Pwinter}$), as well as the mean isotopic value groundwater ($\delta_{groundwater}$). δ_{SS} is the isotopic value of a basin in isotopic steady state (evaporation = inflow) and δ^* is the theoretical limiting isotopic enrichment (both closed circles) of a desiccating basin if fed by amount-weighted average precipitation. (b) The position of an individual lake (grey circle) along the LEL can be used to determine the relative importance of evaporation (E) and water inflow (I), whereas

←

offsets of individual lakes from the LEL can help identify the relative influence of summer (S) and winter (W) precipitation on the isotopic signature of lake waters (Wolfe et al. 2007). *See* Methods for further details.

groundwater influx is important, we also expected LEL distances to be influenced by $\delta_{\text{groundwater}}$, a value similar to that of winter precipitation in this study (*see* above).

Spatial patterns of isotopic mass-balance (E : I) and water source determinations (LEL distance) were estimated from the 2004 spatial survey data using SURFER 8 (Golden Software, Inc. 2003) to evaluate how lake hydrology varied in relation to long-term patterns of precipitation and evaporation. Interpolations among lakes were calculated using a radial-basis gridding method and a multiquadratic kernel function with equidistant X and Y grid spacing, where X and Y dimensions are lake latitude and longitude in decimal degrees, and Z was defined as either E : I or distance to LEL. We predicted that E : I would follow regional patterns of precipitation deficit (Fig. 1) and should be greatest in lakes of southwest Saskatchewan where semiarid conditions often occur. Similarly, we expected the importance of summer precipitation to be greatest in the northeast sector of our study area where rainfall is greatest (Pham et al. 2008).

Multivariate statistics were used to identify the main correlates of E : I ratio and LEL distance derived from the 2004 lake survey. Preliminary analysis of data using detrended correspondence analysis revealed the presence of short environmental gradients; therefore, redundancy analysis (RDA) was used to determine whether the among-lake variability in E : I and LEL distances could be explained by linear combinations of measured environmental predictor variables (Hall et al. 1999). All variables were centered and standardized prior to analysis. Environmental variables included those parameters measured at all lakes at the time of isotope sampling (Table 1) and were included only if they were not collinear with other predictor variables (i.e., variance inflation factor <20) and individually explained a significant ($p < 0.05$) amount of variation in E : I or distance to LEL (Hall et al. 1999). Monte Carlo permutation tests with 999 iterations were used to determine significance of the first RDA axis. As a result of standardization, length of arrows in RDA are not indicative of the absolute strength of correlations among predictor and response variables, although ordinations can still be interpreted with respect to basic correlations (positive, negative) and factor loading on the ordination axes. All calculations were performed using CANOCO version 4 (ter Braak and Šmilauer 1998). Further statistical details are provided in Hall et al. (1999). We hypothesized that E : I ratios should increase as a function of lake water salinity due to evaporative concentration of dissolved salts and enrichment of heavy isotopes during summer. In contrast, we hypothesized that the distance to LEL should be most negative in the southeast regions where mean snow pack was greatest, but recognized that this analysis might be biased because of groundwater influx (*see* above).

Results

Chemical synchrony—As detailed in Pham et al. (2008), seasonal changes in standardized salinity were more coherent among lakes in dry 2003 than in wet 2004 (Fig. 3a). For example, 13 of 16 lakes exhibited similar increases in standardized salinity, leading to an elevated mean (±SD) lake-pair synchrony (S) of 0.78 ± 0.19 during 2003, a year with low total summer precipitation (182 mm), more snow (101.3 cm), and higher mean maximum temperatures (20.1°C ± 6.5). In addition, all but one site showed decreasing lake levels during summer (data not shown), consistent with results of salt mass-balance models which demonstrate that changes in salinity are controlled mainly by evaporation (Pham et al. 2008). In contrast, 2004 was less coherent ($S = 0.56 \pm 0.34$), wetter in summer (295.3 mm) but not winter (81.8 cm), almost 4° cooler (16.9°C ± 5.0), and exhibited six lakes with increased water levels.

Intra-annual changes in $\delta^{18}O$ (Fig. 3b) also showed greater seasonal synchrony among lakes in 2003 ($S = 0.84 \pm 0.15$) than in 2004 ($S = 0.22 \pm 0.37$), a result which was not affected by corrections for salinity (analysis not shown). In general, most sites revealed isotopic enrichment from spring to autumn during 2003 (mean $\delta^{18}O$ enrichment 1.7 ± 1.2‰), consistent with salinity increases and lake volume declines, whereas little net isotopic enrichment was recorded during 2004 (0.20 ± 0.14‰). In contrast, seasonal changes in Z-transformed $\delta^{2}H$ were asynchronous during both 2003 (0.04 ± 0.34) and 2004 (−0.01 ± 0.25; Fig. 3c), although nine sites showed seasonal enrichment of $\delta^{2}H$ during 2003 (mean enrichment 5.9 ± 4.2‰), and seven lakes exhibited seasonal enrichment of $\delta^{2}H$ during 2004 (0.0 ± 0.01‰).

Seasonal changes in Z-transformed salinity and $\delta^{18}O$ within lakes were significantly correlated in 2003 ($r = 0.858$, $p < 0.001$) and 2004 ($r = 0.443$, $p < 0.001$), either with or without correction for the effects of salinity on $\delta^{18}O$. In contrast, changes in standardized salinity were less correlated with $\delta^{2}H$ both in 2003 ($r = 0.336$, $p = 0.01$) and 2004 ($r = 0.151$, $p = 0.25$). Significant correlations also existed between seasonal changes in $\delta^{18}O$ and those of $\delta^{2}H$ both in 2003 ($r = 0.384$, $p = 0.004$) and in 2004 ($r = 0.250$, $p = 0.05$), particularly if $\delta^{18}O$ and $\delta^{2}H$ values were not Z-transformed ($r = 0.745$, $p < 0.001$ and $r = 0.553$, $p < 0.001$, respectively). However, unlike analysis of seasonal chemical changes, the absolute $\delta^{18}O$ and $\delta^{2}H$ values were only correlated weakly with lake water salinity (e.g., in 2004, $r = 0.048$, $p = 0.847$).

Isotope hydrology—Isotopic values of lakes sampled during the spatial survey of 2004 were clustered around the predicted LEL (Fig. 4). Regressions of $\delta^{2}H$ vs. $\delta^{18}O$ exhibited a slope (4.27 ± 0.40) similar to that of the theoretical LEL (4.0), although its intercept with the LMWL was depleted isotopically relative to δ_P. Unexpectedly, observed slopes of regressions of $\delta^{2}H$ vs. $\delta^{18}O$ were somewhat greater than the LEL slope in analyses based on the mean values of lakes sampled seasonally in 2003 (4.77 ± 0.95) and 2004 (5.14 ± 1.00), possibly reflecting a more substantial influence of groundwater in this narrower range of lakes (analysis not shown).

Calculated E : I ratios indicated that hydrologic balances of the 70 survey lakes were regulated mainly by the inflow of water rather than by evaporation during 2004 (Fig. 5a). Overall, E : I ranged from 0.0 to 0.84, with a mean (±SD) value that was substantially lower (0.44 ± 0.18) than the point at which lakes are in steady state (E : I = 1.0).

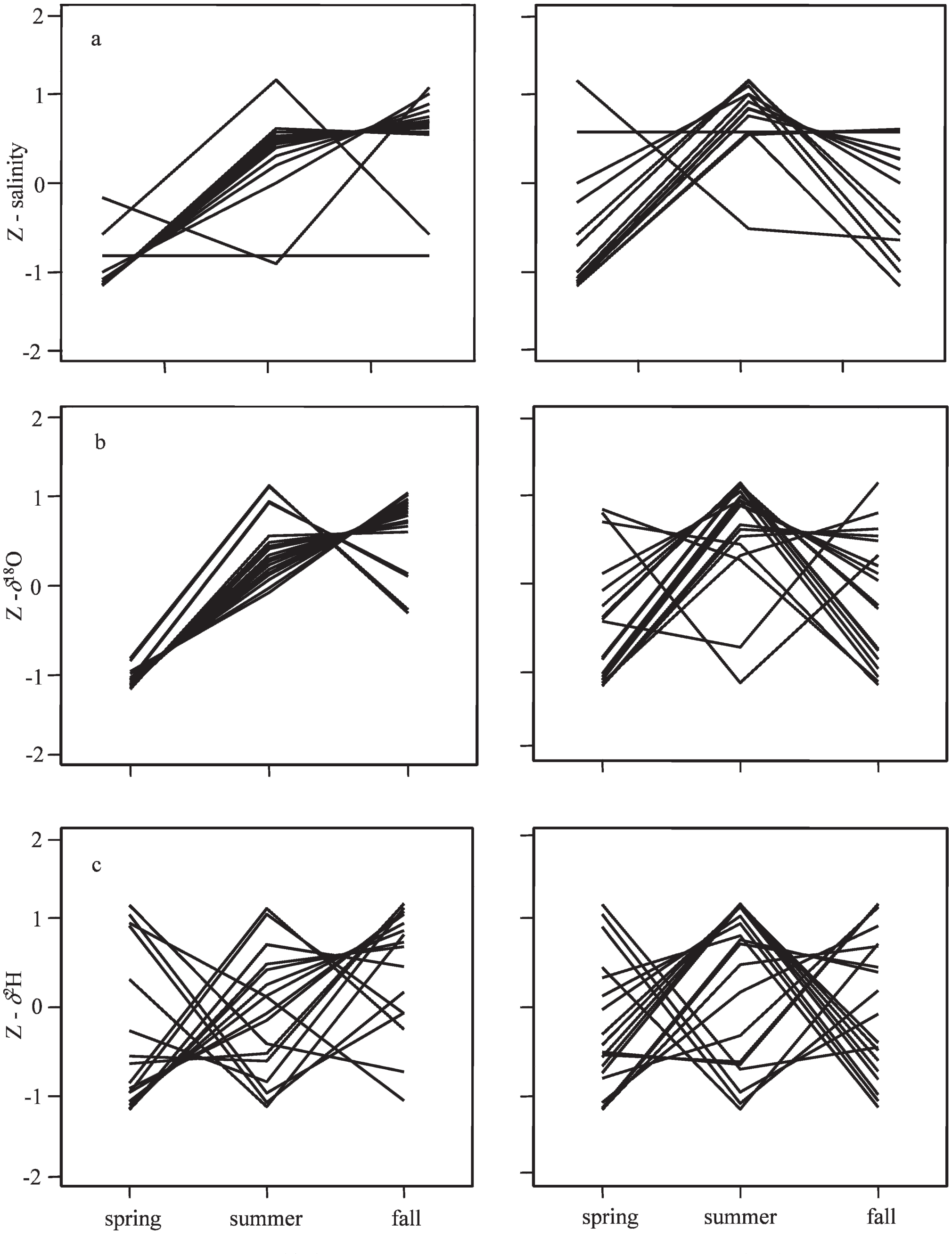

Fig. 3. Seasonal changes in Z-transformed time series of (a) lake water salinity (g TDS L^{-1}), (b) $\delta^{18}O$ signatures, and (c) $\delta^{2}H$ during a year with a dry (2003) and wet (2004) summer. Seasonal samples were collected during a 2-week interval centered around calendar day-of-year (DOY) 120 (spring), 200 (summer) and 280 (autumn).

Because isotopic enrichment was greater in mid-summer (DOY ~200) than during autumn 2004 (DOY ~280; Fig. 3), calculated E : I ratios based on DOY 200 likely overestimated the importance of evaporation to annual water balances. As expected, the isotope signatures of lakes were more similar to that of winter precipitation or groundwater than to that of summer precipitation during 2004 (Fig. 5b). Specifically, 72% of lakes had isotopic values that fell below the theoretical LEL during 2004, with a mean (±SD) distance to LEL of -4.22 ± 6.22. Unexpectedly, seven lakes exhibited positive offsets from the LEL that were in excess of that expected from mean summer precipitation during the past 15 yr.

Ratios of E : I were spatially structured in 2004, with greater influence of evaporation in the northwest, and greater effects of water inflow in the southwest and northeast (Fig. 6a). In particular, the strong north–south gradient of E : I observed in western lakes (longitude >107°W) contrasted sharply with estimates of mean annual precipitation deficit based on long-term meteorological

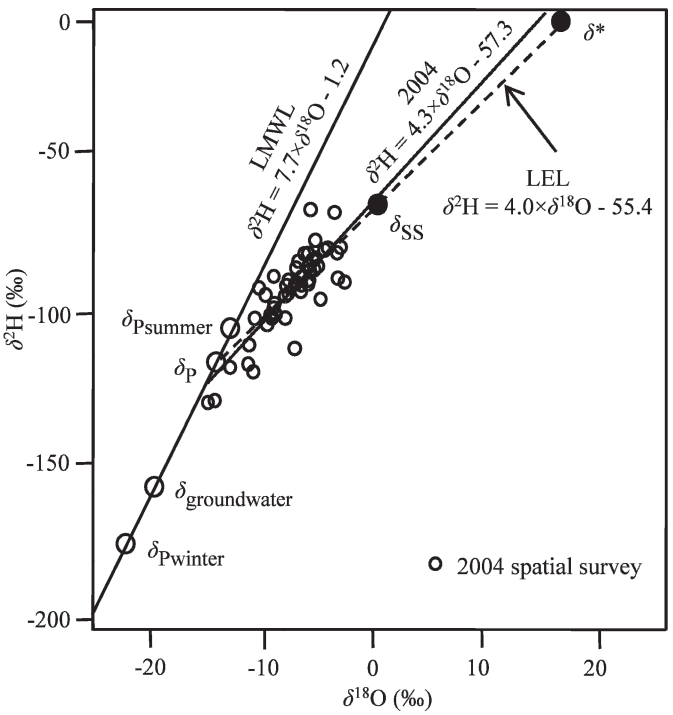

Fig. 4. Relationship between lake surface water (0–5 m) signatures of δ^2H and $\delta^{18}O$ for 70 lakes sampled ~DOY 200 in 2004 relative to the local meteoric water line (LMWL). Note that the best-fit least-squared regression (solid line) lies near the theoretically defined local evaporation line (LEL) based on sampling of summer and winter precipitation (dashed line), but that lakes are more depleted than a typical lake in steady state between evaporation losses and precipitation influx (δ_{SS}; E = I; solid circle).

records for the region (dashed lines in Fig. 6). Similarly, the distance to LEL was also geographically structured, with groundwater or winter precipitation exhibiting the strongest influence on lake water isotopic composition in the western region of the survey, while summer precipitation predominated in the south-central and northeastern regions (Fig. 6b). Again, spatial patterns of water sources appeared to have little relationship to long-term gradients of precipitation deficit.

Redundancy analysis of isotopic data from the spatial survey of 70 lakes was significant ($p < 0.001$) and explained 31.8% of lake variation on the first axis, and 10.8% on the second axis (Fig. 7). Overall, E : I ratios were correlated strongly with RDA axis 1 ($r = 0.770$, $p < 0.001$), as well as with longitude position ($r = 0.401$, $p < 0.001$), salinity ($r = 0.524$, $p < 0.001$), DOC content ($r = 0.572$, $p < 0.001$), and concentrations of Cl ($r = 0.310$, $p < 0.01$), K ($r = 0.540$, $p < 0.001$), SO_4 ($r = -0.285$, $p < 0.05$), and TKN ($r = -0.458$, $p < 0.001$). In contrast, distance to LEL distance was correlated with RDA axis 2 ($r = -0.45$, $p < 0.001$), as well as with maximum lake depth ($r = 0.301$, $p < 0.01$).

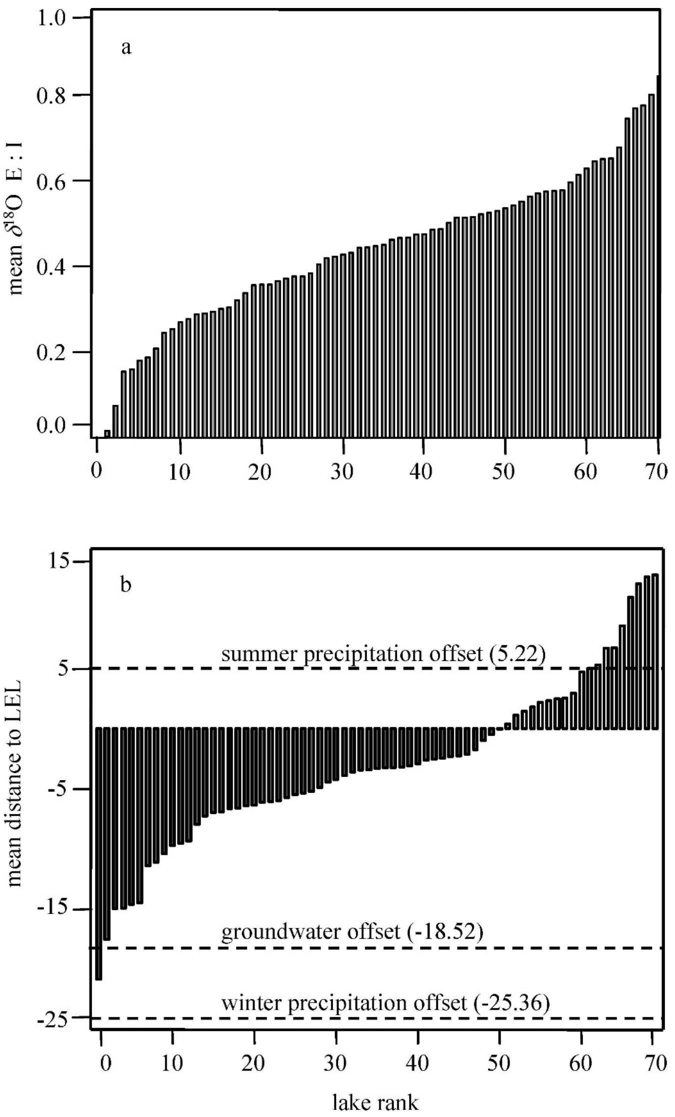

Fig. 5. Distribution of (a) calculated ratios of evaporation: inflow (E : I) and (b) perpendicular distance to the local evaporative line (LEL) based on $\delta^{18}O$ analysis of 70 lakes sampled ~DOY 200 during 2004. Horizontal dashed lines represent these expected maximum offsets arising from the influence of summer (+5.22) and winter precipitation (−25.36), or groundwater (−18.52).

Discussion

Analysis of stable isotopes revealed that while evaporation during the dry summer of 2003 caused regionally synchronous increases in both dissolved salt and heavy isotope contents of lake water (Fig. 3), the hydrological balance of these prairie lakes was maintained by inflow of water (Fig. 5a), either from groundwater (van der Kamp and Hayashi 1998; Fritz et al. 2000; Yu et al. 2002) or from usually high supply of runoff from catchments, most of which is derived from winter precipitation (Steppuhn 1981; Akinremi et al. 1999). These findings are consistent with the observation that permanent lakes are common on the northern Great Plains (Last 1992) despite precipitation deficits that exceed 40 cm yr^{-1} (Fig. 1), and with hydrologic balances from individual prairie lakes which demonstrate strong effects of both evaporation (Fritz 1990) and groundwater (van der Kamp and Hayashi 1998; Yu et al. 2002). However, although the relative importance of

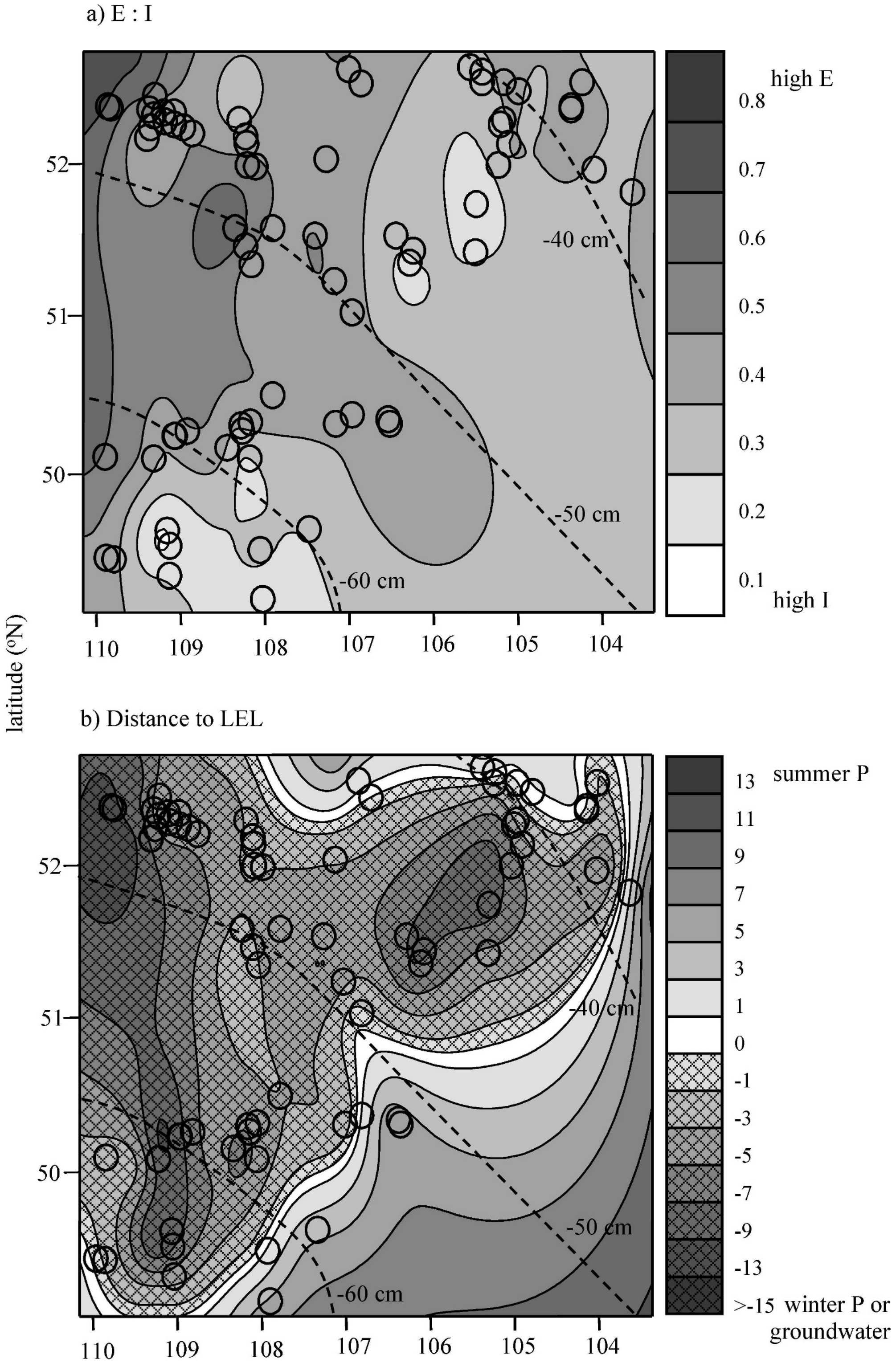

Fig. 6. Spatial patterns of isotopically derived estimates of (a) E : I and (b) perpendicular distance to local evaporation line derived from a survey of 70 lakes in 2004. Surface interpolations were determined using radial basis method with a multiquadratic kernel function using SURFER version 8.0. Due to extrapolation beyond the data range, spatial patterns should not be interpreted in the southeast sector of the survey.

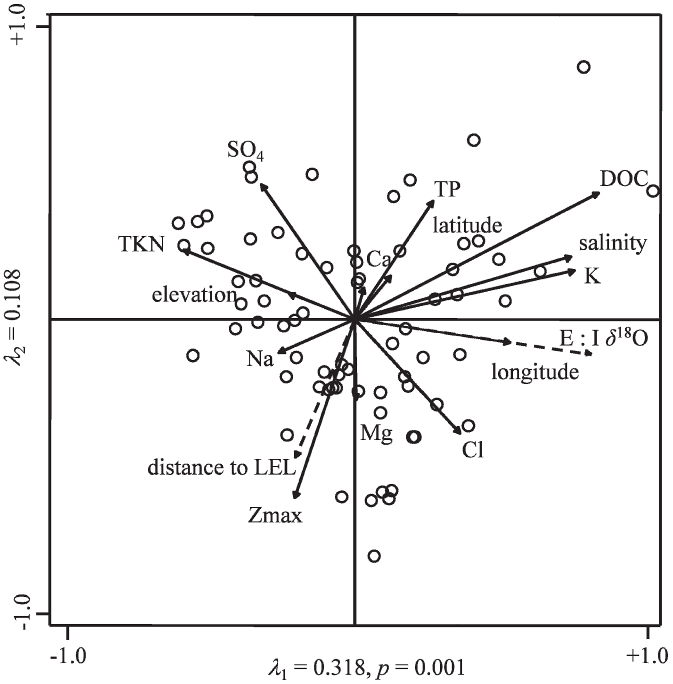

Fig. 7. Redundancy analysis tri-plot illustrating the relationship between response variables (distance to LEL, E : I based on δ^{18}O; dashed arrows) and predictor variables (solid arrow; Table 1) for 70 survey lakes (open circles) sampled ~DOY 200 during 2004. Analysis reveals that E : I is positively correlated with lake-water salinity and longitude, and negatively correlated with total Kjehldahl nitrogen (TKN), whereas distance to LEL is positively correlated with lake depth.

evaporation and inflow was spatially structured (Fig. 6) and increased with lake water salinity (Figs. 3, 7), isotopically derived E : I ratios were unrelated to long-term meteorological records of annual precipitation deficit (Fig. 6). Taken together, these findings suggest that winter climate plays an unexpectedly important role in sustaining prairie lake ecosystems and, as a consequence, that paleoclimate records in this region may mainly reflect temporal variation in climate systems that alter the supply (snowfall), persistence (winter temperature), and runoff (spring temperature) of winter precipitation (ENSO, NAO).

Seasonal effects of evaporation on water chemistry—Previous whole-lake water budgets demonstrate that evaporation of surface water causes regionally synchronous increases in the salinity of these prairie lakes during dry years, but that temporal coherence among sites declines as a linear function of summer precipitation (Pham et al. 2008). Consistent with this view, our isotopic analyses demonstrate that both the extent (Fig. 3) and synchrony ($S = 0.84 \pm 0.15$) of seasonal enrichment of δ^{18}O was greater during the dry summer of 2003 than during wet 2004 ($S = 0.22 \pm 0.37$). Similarly, seasonal changes in Z-transformed δ^{18}O were highly correlated with those of lake-water salinity in both years ($r = 0.858$ and 0.443, respectively; $p < 0.001$), suggesting that the same mechanism underlies intra-annual variability in both water chemistry and isotope ratios. Consistent with the findings of Pham et al. (2008) that evaporation caused synchronous changes in lakes during 2003, we note that mean seasonal enrichment of isotopes from spring to autumn during 2003 fell along a line with a slope (3.47) similar to that expected to arise from evaporation (LEL = 4.0), whereas the isotopic composition of lakes exhibited little net change during 2004 (enrichment slope = 0.00). Together, these findings support the observation that the transfer of energy (evaporation) is a key physical factor inducing temporal coherence of aquatic ecosystems, whereas the transfer of mass (precipitation) reduces synchrony among lakes.

Despite clear evidence of the effects of evaporation on lake chemistry, absolute changes in δ^{18}O and δ^{2}H were slight relative to those documented in other desiccating basins in Canada (Gibson et al. 2005; Brock et al. 2007). For example, δ^{18}O values can vary 5 to >20‰ from spring to autumn in isolated boreal lakes (Brock et al. 2007), whereas mean seasonal enrichment of δ^{18}O was only 1.7 and 0.2‰ in 2003 and 2004, respectively. Similarly, unpublished analysis of lakes of the Qu'Appelle Valley, Saskatchewan (50.80°N, 103.39°W), reveals that annual variability of δ^{18}O during 2007 exceeds 5‰ even in strongly river-influenced lakes (B. Wissel, S. Pham, and P. R. Leavitt unpubl.). Analysis of E : I ratios (Fig. 5a) suggests that the low degree of isotopic enrichment likely arises because of the uniformly strong influence of water inflow from snowmelt or groundwater that buffers all lakes against short-term inter-annual variation in chemical composition, either by dilution of enriched water with depleted groundwater (Birks and Remenda 1999; Yu et al. 2002), or by export of enriched surface waters into shallow aquifers (van der Kamp and Hayashi 1998). Similarly, the relatively low sensitivity of lake isotopic values to inter-annual variations in precipitation is consistent with the long residence time of these lakes (>10 yr when known) and the relatively low amounts of total precipitation in this subhumid to semiarid region (Walker et al. 1995; Quiring and Papakyriakou 2005).

Unexpectedly, seasonal changes in δ^{2}H signatures of surface waters were substantially less synchronous ($S = 0.04 \pm 0.34$) than those of δ^{18}O (0.84 ± 0.15) in 2003 despite clear evidence of isotopic enrichment from spring to autumn (mean 5.9‰) and significant correlations of seasonal changes in both Z-transformed ($r = 0.384$, $p = 0.004$) and raw time series ($r = 0.745$, $p < 0.001$). Examination of individual lake plots (Fig. 3) shows that low temporal coherence arose mainly from the effects of five lakes that exhibited unusually enriched values of δ^{2}H, but not δ^{18}O, in spring 2003. Analysis of the untransformed data (not shown) demonstrates that depletion of δ^{2}H from spring to summer ranged from 3‰ to >20‰ among these outlier lakes, suggesting that lack of synchrony did not arise from consistently greater effects of nonequilibrium kinetic fractionation of H compared to that of O (Kendall and Caldwell 1998), or technical errors in estimation of small changes in ^{2}H content of surface waters. Instead, we note that during spring 2003, all outlier lakes were highly saline, deep (Fig. 7), and exhibited a pronounced surface lens of freshwater that had dissipated by midsummer.

Although many prairie lakes are polymictic due to high area : depth ratios and high regional wind speed (Leavitt et al. 2006; Pham et al. 2008), strongly stratified meromictic lakes can also occur (Michels et al. 2007; Table 1). We hypothesize that this subset of lakes may have exhibited highly complicated mixing of internal waters, which bias estimates of surface water isotope signatures during spring, but not at later times of the year. Tests of this hypothesis are ongoing, but require highly resolved temporal and vertical time series of isotope and water chemistry, which are beyond the scope of this paper. Regardless, we note that despite seasonal differences in enrichment of heavy isotopes of H and O among lakes (Fig. 3), estimates of E : I ratios derived from δ^2H were highly correlated ($r = 0.70$, $p < 0.001$) to those calculated from δ^{18}O, suggesting that the lack of synchrony of δ^2H did not have large effects on our determination of overall lake hydrology.

Effects of water influx on lake chemistry—Analysis of E : I ratios revealed that the isotopic composition of all survey lakes was more strongly regulated by water inflow than by evaporation, despite the absence of permanent rivers feeding these lakes (Fig. 5a) and the observation that the entire study region exhibits a net annual precipitation deficit of >40 cm yr^{-1} (Fig. 1). In fact, >30% of sites had E : I ratios characteristic of hydrologically open boreal lakes (<0.4), which are subject to regular flooding or river input (Wolfe et al. 2007; Brock et al. 2007), whereas the remainder of our sites were isotopically characteristic of restricted-drainage lakes which are occasionally fed by surface-water inflow (E : I <0.8). Such a strong influence of inflow despite the absence of permanent rivers is consistent with the findings of Yu et al. (2002) who showed that gross isotopic composition of lakes was regulated by snowmelt-fed influx of groundwater, and not evaporation, in a series of 13 prairie sites in the U.S.A.

Several processes may interact to emphasize the importance of winter precipitation and groundwater to the hydrologic budget of prairie lakes. First, although prairie snowfall contributes only 30% of total annual precipitation during the past 30 yr, very little is lost to the atmosphere during winter (Winter 1989), and intense snow-pack melting during the brief spring is responsible for 80% of annual surface runoff (Akinremi et al. 1999). Second, snowmelt, rather than summer precipitation, accounts for most groundwater recharge in this region, particularly via the numerous wetlands that feed shallow aquifers (van der Kamp and Hayashi 1998; Birks and Remenda 1999). Finally, the very low mean topographic relief (mean change <25 cm km^{-1}) in southern Saskatchewan (Patoine and Leavitt 2006) leads to 10-fold variability among years in the area of land contributing runoff to lakes (i.e., realized drainage area), with particularly high runoff and contributing catchment area during years of maximal snow-pack accumulation (Patoine et al. 2006). In contrast, precipitation from low-intensity summer events is usually returned to the atmosphere directly by evaporation, or through plant evapotranspiration (Ferguson et al. 2007), leading to disproportionately low contribution of summer rainfall to runoff relative to expectations based on seasonal precipitation totals (70% in summer). Taken together, these factors emphasize the importance of winter precipitation on the hydrologic budgets of most prairie lakes, and caution managers, farmers, and limnologists that the 0.95 cm yr^{-1} decline in winter precipitation observed since 1970 (Akinremi et al. 1999) may have much greater effects on the persistence and quality of aquatic resources in the Prairies than will future increases in regional summer temperature (Schindler and Donahue 2006; van der Kamp et al. 2008).

Despite clear evidence of the importance of winter precipitation or groundwater in the hydrologic budgets of prairie lakes, analysis of stable isotopes did not distinguish clearly among these sources of water to individual lake basins (Fig. 5b). Although analysis of the distance to the theoretical LEL has been used successfully to quantify the relative proportion of summer and winter precipitation in lakes of northern boreal plains (Wolfe et al. 2007), several lake features related to groundwater characteristics interact to complicate the use of this metric in prairie ecosystems. For example, the mean isotopic value from 129 regional wells ($\delta_{\text{groundwater}}$) is both consistent with other determinations for the northern Great Plains (Kelley and Holmden 2001; Yu et al. 2002) and is similar to the mean composition of regional snowfall in our study area (Figs. 4, 5b). Thus, without site-specific quantification of the isotopic composition of groundwater and local snowfall, it is difficult to partition the relative importance of snow and subsurface inflow. Second, under conditions in which groundwater contributions are important, we expect that isotopic composition of source waters will be depleted from δ_{P} towards $\delta_{\text{groundwater}}$ (Fig. 2), but that estimates of δ^* will not necessarily change (*see* Eq. 1 above). As a result, slope of regressions describing the relationship between δ^2H and δ^{18}O will be steeper than that of the theoretical LEL, leading to over estimation of groundwater influence in lakes with low E : I ratios, and overestimation of summer contributions in lakes with higher E : I ratios.

As documented by Yu et al. (2002), lakes with strong groundwater contribution can exhibit $\delta^2\text{H} - \delta^{18}\text{O}$ slopes of 5.9 to 7.4, values substantially greater than that expected given regional temperature and humidity (4.0–5.0 in western Canada; Gibson et al. 2005 and this study). This effect is expected to be particularly great in small groups of local lakes which may exhibit similar groundwater influence. Consistent with these ideas, analysis of relationships between δ^2H and δ^{18}O slopes based on our seasonal survey of 21 lakes revealed high slopes in both 2003 (4.77) and 2004 (5.14), while seven of 70 survey lakes exhibited positive deviations from LEL that lay outside the theoretical bounds defined by summer precipitation (δ_{Psummer}). This latter anomaly can be explained if groundwater-fed lakes evolve along a steeper slope than that defined by the theoretical LEL (e.g., Fig. 4), but suggests that further whole-lake mass-balances will be required to determine whether insights from LEL offset determinations can be applied to sites with substantial subsurface inflow. In particular, we recommend that highly resolved (biweekly) time series of isotopes and flux be determined for surface water, groundwater, and precipitation in a series of lakes of contrasting lotic influence (riverine vs. topographically isolated) to better evaluate the role of groundwater and runoff as controls of lake hydrology.

Spatial structure of hydrologic regulation—One of the main findings of this study is that the relative effect of evaporation on surface water isotope chemistry (as E : I) is geographically structured, but that the east–west gradient identified visually (Fig. 6) and by redundancy analysis (Fig. 7) is unrelated to long-term, north–south geographic patterns of annual precipitation deficit (dashed lines in Figs. 1, 6), a metric which is highly influenced by hydrologic processes which occur during summer. Thus, simple meteorological metrics of precipitation and potential evaporation do not appear to provide strong predictions of the effects of evaporative forcing on lake chemistry. In general, this conclusion is similar to that of Pham et al. (2008), who used whole-lake water-balance models to determine that the effects of evaporation in the absence of groundwater were greatest in the relatively humid northeast sector of our study region, rather than the drier southwest. Here, we extend this analysis to demonstrate that the influence of evaporation relative to that of groundwater is generally greatest in western saline lakes (Fig. 7). Together, these findings suggest that the determination of the vulnerability of aquatic systems to future climate change on the northern Great Plains cannot be easily assessed from modern meteorologic data alone, and may depend more on the supply of snow and groundwater to the region, than on the direct effects of global warming.

Caveats to the use of isotope models—Stable isotope mass-balance models are subject to several assumptions that can influence the reliability of interpretations related to lake hydrology (Gibson 2002; Gibson et al. 2005; Wolfe et al. 2007). In our case, we assumed that prairie lakes were influenced mainly by precipitation and evaporation (i.e., steady state) because each lake lacked visible river influence (terminal or isolated basins), remained permanently inundated (no intermittent wetlands), and exhibited residence times >10 yr, when known. Fortunately, although we conclude that this assumption is unlikely to be true due to the potential influence of groundwater, analysis of lake synchrony and the relative influence of evaporation and water inflow were not affected by the presence of subsurface flux. Instead, we feel that estimates of water source based on distance to LEL are of unknown reliability (*see* above) and require further investigation. Second, we assumed that all lakes could be modeled using a single theoretical LEL rather than lake-specific evaporation lines (Gibson et al. 2005). Consistent with this finding, the regression of δ^2H vs. δ^{18}O was not statistically different from that of the theoretical LEL (Fig. 4), although as discussed above, we suggest that groundwater could increase the slope of the observed LEL in small groups of lakes (Yu et al. 2002) and that site-specific estimates of $\delta_{\text{groundwater}}$ may be required in future mass-balance models. Finally, we assumed that precipitation received during 2003 and 2004 was characteristic of the long-term fluxes and isotopic signatures of the region, even though we recognize that the climate and hydrology of the northern Great Plains is characterized by very high inter-annual variability (Walker et al. 1995; Quiring and Papakyriakou 2005). In this regard, we believe that the inter-annual variation in slope of isotope regressions based on seasonal surveys in 2003 (4.77) and 2004 (5.14) may reflect differences in mean temperature and humidity among years, both of which influence determinations of δ^* and LEL slope (Gibson et al. 2005, 2008). However, despite these violations, we feel that the use of stable isotope mass-balances provides an excellent first order approximation of the basic hydrological fluxes and E : I ratios of prairie lakes.

Implications for climate–lake relationships—Our analyses suggest that winter precipitation is the main control of hydrologic variability in lakes of the northern Great Plains, either directly due to its predominant contribution to surface runoff (Steppuhn 1981; Akinremi et al. 1999), or indirectly via replenishment of groundwater aquifers (Hayashi et al. 1998; Birks and Remenda 1999; Yu et al. 2002). Consequently, changes in climate systems that regulate the supply, persistence, or melting of snow are predicted to have the strongest effects on prairie lake chemistry and sustainability. In this regard, it is interesting to note that shifts in the Southern Oscillation Index are correlated to changes in both winter temperature and snowfall in the Canadian Prairies (Trenberth and Hurrell 1994). For example, during El Niño events (e.g., 1997), the winter jetstream splits over the northern Great Plains, with one arm extending north beyond the grassland boundary (Fig. 1), while the southern arm traverses through the northern U.S.A. (Shabbar et al. 1997). In this situation, the Canadian Prairies often experience relatively warm and dry winter conditions, with little snow-pack accumulation and low spring runoff (Shabbar and Skinner 2004). In contrast during La Niña events (1999), the jetstream lies to the south of the Canadian Prairies, leading to cold, relatively wet winters. These patterns can be intensified or diminished by climatic variations associated with the PDO (McCabe et al. 2004), and from outbreaks of cold arctic air, which are more common in the positive phase of the NAO or Arctic Oscillation (Hurrell 1995). Given these observations, we conclude that the mechanisms underlying changes in the sign and intensity of atmosphere–oceanic teleconnections during winter and spring may have a profound effect on aquatic resources in the Canadian Prairies, and that better understanding of the controls of flux, persistence and melting of winter snow may provide important insights on the best means of adaptation to future climate change.

Insights from whole-lake mass-balances of isotopes and chemistry have important implications for the interpretation of paleoclimatic records from lakes on the northern Great Plains. First, our analysis of lake synchrony revealed that sentinel systems used for climatic reconstruction are most likely to represent the broader hydrologic landscape when energy transfer (evaporation) is the predominant mechanism causing chemical variability. In contrast, our analyses suggest that changes in precipitation are more likely to induce variability among basins (Fritz et al. 2000), and that short-term deviations among lakes likely reflect catchment-specific differences in how precipitation is exported to lakes from land (Blenckner 2005). However, because we also conclude that lakes of the Canadian Prairies are most strongly influenced by changes in winter precipitation, we suggest that long-term variations in lake chemistry and synchrony likely reflect major reorganizations in climate systems that regulate the transportation, persistence, and melting of winter precipitation (Michels et al. 2007). Validation of these hypotheses is ongoing and requires first that we calibrate the isotopic signatures of lake sediments with the chemical and isotopic characteristics of overlying waters as has been done elsewhere (Huang et al. 2004), and second, that we compare historical changes in isotope composition of sediments with independent metrics of past lake levels and salinity (e.g., diatoms; Fritz et al. 2000). Rather than varying coherently with each other as is commonly expected in most paleoclimate studies (e.g., comparisons with ice cores, tree-rings, etc.), we anticipate that spatially structured but independent variation among lakes will provide key insights into historical changes in the climate systems which regulate winter precipitation (Michels et al. 2007).

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