

Water Isotopic and Hydrochemical Evolution of a Lake Chain in the Northern Great Plains and its Paleoclimatic Implications

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Abstract

Oxygen isotopes and geochemistry from lake sediments are commonly used as proxies of past hydrologic and climatic conditions, but the importance of present-day hydrologic processes in controlling these proxies are sometimes not well established and understood. Here we use present-day hydrochemical data from 13 lakes in a hydrologically connected lake chain in the northern Great Plains (NGP) to investigate isotopic and solute evolution along a hydrologic gradient. The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ of water from the chain of lakes, when plotted in $\delta^2\text{H} - \delta^{18}\text{O}$ space, form a line with a slope of 5.9, indicating that these waters fall on an evaporation trend. However, 10 of the 13 lakes are isotopically similar ($\delta^{18}\text{O} = 6 \pm 1\text{‰}$ VSMOW) and show no correlation with salinity (which ranges from 1 to 65‰). The lack of correlation implies that the isotopic composition of various source waters rather than in-lake evaporation is the main control of the $\delta^{18}\text{O}$ of the lakes. Groundwater, an important input in the water budget of this chain of lakes, has a lower $\delta^{18}\text{O}$ value (-16.7‰ in 1998) than that of mean annual precipitation (-11‰) owing to selective recharge from snow melt. For the lakes in this chain with salinity <15‰, the water Mg/Ca ratios are strongly correlated with salinity, whereas Sr/Ca is not. The poor correlation between Sr/Ca and salinity results from uptake of Sr by endogenic aragonite. These new results indicate that $\delta^{18}\text{O}$ records may not be interpreted simply in term of climate in the NGP, and that local hydrology needs to be adequately investigated before a meaningful interpretation of sedimentary records can be reached.

Introduction

The northern Great Plains (NGP) of North America are dotted with innumerable glacially-formed lakes. Most are small, shallow, semi-permanent potholes and wetlands, especially those found in the extensive till plains and dead-ice moraines (Sloan, 1972; Winter, 1989). Some lakes situated in outwash sand and gravel tend to be deeper and may preserve continuous sedimentary sequences covering the last several thousand years. These sediments and especially their geochemistry have been used as indicators of lake salinity and effective moistures (precipitation – evapotranspiration, P-E) (e.g., Engstrom &

Nelson, 1991; Xia et al., 1997b; Yu & Ito, 1999; Fritz et al., 2000).

In arid regions lakes in topographically closed basins are generally regarded as hydrologically closed (no surface outflow and limited groundwater interaction) and are considered near-ideal for studying past changes in hydrochemistry resulting from climatic forcing. The expected behavior of lake hydrochemistry and thus of endogenic carbonates (such as ostracodes) that form in the lake is as follows. (1) The increased evaporation of water causes Ca-carbonate supersaturation, so that Ca is continually removed from the water, while more conservative solutes such as Mg and SO_4 accumulate, resulting in higher total dissolved solids (TDS). The

resulting increase in Mg/Ca of the lake is recorded in Mg/Ca of ostracode-shell calcite. (2) The evaporation process allows for selective escape of H_2^{16}O so that $\delta^{18}\text{O}$ of the lake water increases. The higher $\delta^{18}\text{O}$ of the water is recorded in the $\delta^{18}\text{O}$ of ostracode shells and other endogenic carbonates. However, there is mounting evidence that such ideal behavior may be an oversimplification. Recent results from the NGP suggest that complex local hydrology may frequently obscure the climatic signals in these paleo-records and prevent their straightforward climatic interpretation (e.g., Last et al., 1994; Smith et al., 1997). Based on oxygen and hydrogen isotope data of surface water system in the Great Basin, Benson (1994) also stressed the importance of understanding modern processes and controls on water isotopes in interpreting paleo-records.

Here we use present-day water isotopic ($\delta^{18}\text{O}$ and $\delta^2\text{H}$) and hydrochemical data from a chain of 13 hydrologically connected lakes in the NGP to investigate their evolution along a topographic and hydrologic flow path. The spatial pattern of hydrochemistry from these lakes provides a modern analogue of what may occur in an individual lake over time (space-for-time substitution). We thus explore the climatic and hydrologic influences on the geochemical evolution of these hydrologically connected lakes and evaluate the usefulness and reliability of different sedimentary proxies for paleoclimatic studies.

Study region

The glaciated plains are part of the northern Great Plains in central North America (Figure 1A). The region has a subhumid to semi-arid climate, with precipitation significantly less than evapotranspiration. With a continental location and the variable influence of three major air masses (Bryson & Hare, 1974), the NGP has large inter-annual climatic variation. The moist Gulf of Mexico air mass provides most moisture to the region, as Pacific air has been highly modified and has become dry by the time it crosses the Rocky Mountains and the cold continental Arctic air is dry with little moisture (Bryson & Hare, 1974).

The mean annual precipitation ranges from about 520 mm in the northeast to 350 mm in southwest, with peak precipitation in June. The temperature ranges for January and July are from -18 to -10°C and from 20 to 24°C , respectively.

The chain of Rice–Carlson lakes is situated on the Missouri Coteau in north-central North Dakota (Figure 1B). The Coteau is an area of stagnation (dead-ice) moraine composed primarily of till derived from Cretaceous and Tertiary sedimentary rocks with limited areas of glacial outwash in topographically low areas (Sloan, 1972). The chain of lakes span a distance of about 12 km along a glacial melt-water channel filled with outwash gravel and sand and surrounded by till uplands (Figures 1B and 2). There is no surface water connection among these lakes (Figure 2), but they are linked by groundwater flow. The general direction of groundwater flow is towards the center of the chain from both ends, based on the surface elevation of the lakes. Eisenlohr et al. (1972) reported hydrochemical measurements from some of these lakes, which indicated highest conductivity in the central lakes (our lakes D and E with 65,000 and 35,000 $\mu\text{S cm}^{-1}$ conductivity, respectively). Rice Lake, the deepest in the chain, may have been the only surviving lake during extreme droughts in the distant past. Side-valley seepage zones marked by dense colonies of hard-stem bulrush and cattail indicate at least limited groundwater inflow from the uplands. In the southern part of the chain, test holes show that the outwash sand and gravel deposits are about 24 m in depth above the underlying glacial till (Sloan, 1972).

Methods

The 13 lakes and one spring discharging into Lake D in the Rice – Carlson chain (Figure 2) were sampled in late May, 1998. There was no precipitation event during the two-day sampling period, so the samples represent a single-time snapshot of chemical conditions of these lakes. Water from other NGP lakes were sampled for previous geochemical studies by Eisenlohr et al. (1972), Swanson et al. (1988), Engstrom & Nelson (1991), and Xia et al. (1997a).

The hydrogen isotope composition of lake water was determined on 2-5 μL samples that were reduced with 120 mg zinc (Zn) in Pyrex reaction vessels at 450°C for one hour (Kendall & Coplen, 1985). The evolved H_2 was measured on a Finnigan MAT delta E mass spectrometer. The oxygen isotope composition of lake water was measured on 5 mL samples equilibrated overnight with instrument-grade CO_2 (99.99%) at 25°C (Epstein & Mayeda, 1953). The $^2\text{H}/^1\text{H}$ and $^{18}\text{O}/^{16}\text{O}$ ratios of water samples are presented in delta (δ) notation as $\delta = (R_{\text{sample}}/R_{\text{standard}} - 1) \times 1000$, where R is absolute $^2\text{H}/^1\text{H}$ or $^{18}\text{O}/^{16}\text{O}$ ratios and the standard is Vienna mean standard ocean water (V-SMOW). The analytical precision was 0.05‰ for $\delta^{18}\text{O}$ and 1‰ for $\delta^2\text{H}$.

The cation measurements of lake water samples, after treatments with 6N ultrapure HCl, were carried out on a Perkin Elmer/Sciex Elan 5000 inductively-coupled plasma mass spectrometer (ICP-MS). More saline waters were diluted by adding triply distilled de-ionized water. The analytical precision is within 5% on concentration measurements of individual elements. The anion concentrations were measured by a Dionex DX 500 ion chromatography, using dual AS4A analytical columns, a GP40 gradient pump, an AS40 autosampler and a CD20 conductivity detector. The analytical precision is about 1-2% on concentrations that are at least 10 times the detection limits and about 5% on concentrations that are at least 5 times the detection limits. The detection limits for the measured anions are mostly 0.005 – 0.02 ppm.

Results

All lakes had a pH value above 8, ranging from 8.15 to 8.76, while the spring had a pH of 7.31 (Table 1). Water temperatures were 20 ± 2 °C (18.0 to 22.1°C for 12 lakes), except the spring, which was significantly cooler (11.9 °C). The lakes in the center of the lake chain tend to have higher total dissolved solids (TDS; likely equivalent to salinity in our samples) than other lakes (Fig. 3).

The $\delta^{18}\text{O}$ values from the 13 lakes ranged from -8.5 ‰ (Lake A) to -0.2‰ (Lake I) (Table 1), although all but three had a narrow range of $\delta^{18}\text{O}$ values (-5.4 to -6.7‰) (Figure 4A). At the same time salinity ranged from dilute to hypersaline (TDS = 0.6‰ to 86‰). The spring that runs into Lake D had a much lower $\delta^{18}\text{O}$ value of -16.7‰, compared to the lakes. All lakes fit a tight line on the $\delta^{18}\text{O}$ - $\delta^2\text{H}$ plot, with a slope of 5.9 and intercept of -23.5‰ (Figure 4B).

Sodium (Na) and sulfate (SO_4), of the dominant solutes in all the lakes (Table 1), were closely correlated with TDS (Figures 5A and 5B), as were K and Cl (not shown). These are conservative solutes, which usually precipitate as minerals from lake waters only at very high salinities (Donovan and Rose, 1994). Mg also behaves conservatively at low salinity (<15‰), and Mg/Ca ratios show good correlation with TDS (Figure 5C). At higher salinity and/or with suitable anion chemistry Mg begins incorporating into precipitated minerals (such as high Mg-calcite), lowering Mg/Ca ratios in lake water. However, even at low salinity Sr/Ca ratios are poorly correlated with TDS (Figure 5D) because of preferential uptake of Sr by aragonite precipitation (Eugster and Jones, 1979; Haskell et al., 1996; Xia et al., 1997a).

Discussion

Water isotopic evolution of lake chain

The available isotopic data of atmospheric precipitation, groundwater, and lake waters from the NGP are summarized to evaluate lake isotopic behaviors and evolution of the chain of study lakes (Figure 6). There are two official IAEA-GNIP (Global Network of Isotopes in Precipitation) precipitation stations at the northern limit of the NGP (Wynyard, Saskatchewan and Gimli, Manitoba; Figure 1A). They are approximately 450 km north of the Rice–Carlson lake chain and have isotope data for the period of 1975 - 1982. The precipitation season at Wynyard is from May to August and peaks in July with an average of monthly total of 72 mm and a $\delta^{18}\text{O}$ value of -13‰. The cold season has limited amounts of precipitation but with a very low average $\delta^{18}\text{O}$

value of -25‰. The weighted annual mean for $\delta^{18}\text{O}$ is -15.6‰, closer to summer than winter precipitation values. The precipitation season at Gimli is from May to September and also peaks in July (95 mm) with an average $\delta^{18}\text{O}$ of -10‰. Cold season precipitation has an average $\delta^{18}\text{O}$ value of -25‰, and the weighted mean annual $\delta^{18}\text{O}$ value is -14.2‰. Limited isotope data for precipitation are also available from Grand Forks, North Dakota, about 300 km east of the Rice–Carlson chain, for the period of March 1990 – May 1991 (Matheney & Gerla, 1996; R. K. Matheney, 1997, pers. comm.). The weighted mean annual $\delta^{18}\text{O}$ is estimated to be about -11‰. Thus the mean precipitation $\delta^{18}\text{O}$ in the vicinity of the Rice–Carlson chain is estimated to be about -12‰, with seasonal variation in the range between Grand Forks and the Canadian stations. The $\delta^2\text{H} - \delta^{18}\text{O}$ relations from these three stations have similar slopes but different intercepts (7.2 and 0.0 for Grand Forks; 7.7 and 3.0 for Gimli; and 7.6 and -0.3 for Wynyard). Matheney & Gerla (1996) reported a slope of 7.3 and intercept of 1 for their Grand Forks data. Combined, these data sets define a regional meteoric water line (MWL) of $\delta^2\text{H} = 7.6 * \delta^{18}\text{O} + 2.0$ (Figure 5) with a similar slope but lower intercept compared to the global meteoric water line of $\delta^2\text{H} = 8 * \delta^{18}\text{O} + 10$ (Craig, 1961). Harvey & Welker (2000) reported a similar slope of 7.66 from a site in Nebraska.

Scattered groundwater isotopic data are available from the region. Lent (1993) reported isotopic measurements of three buried-valley aquifers from New Town, New Rockford, and Spiritwood in North Dakota. The $\delta^{18}\text{O}$ values of these groundwaters average about -16‰, which is lower than that of mean annual precipitation. Buried-valley groundwater shows little or no tritium, suggesting no modern water recharge. Remenda et al. (1994) found the $\delta^{18}\text{O}$ value of -25‰ for old groundwater derived from glacial Lake Agassiz at depths of 20-30 m in clay deposits in eastern North Dakota and surrounding areas. Shallow groundwater in the same region has a $\delta^{18}\text{O}$ value of -14‰, close to the mean annual value for precipitation. The Dakota (bedrock) aquifer of lower Cretaceous sandstone and sandy shale in the Grand Forks region lies 30-50 m

beneath low-permeability glacio-lacustrine till and silty clays. Its water is Pleistocene melt-water in origin, brackish, and has a $\delta^{18}\text{O}$ of -18‰ (Matheney & Gerla, 1996). In the same region, the groundwater at ~10 m depth has $\delta^{18}\text{O}$ value of -15‰, while shallower groundwater and soil pore-water at 0.5 - 4 m have a $\delta^{18}\text{O}$ of -12‰. In summary, all regional groundwater tends to have lower $\delta^{18}\text{O}$ compared to mean annual precipitation, possibly because of selective recharge from spring snow melt and mixing with deep fossil groundwater of glacial origin. Thus groundwater contribution to the water budgets could have a significant influence on the isotopic mass balance of small shallow lakes.

The $\delta^{18}\text{O}$ values of 27 additional lakes from the NGP (covering latitudes of 43.5-49.2°N and longitudes of 95.0-105.9°W) range from -4.6 to -0.5‰ (Table 2; Figure 6 inset), with significant seasonal variations (e.g., Xia et al., 1997a). Given the much lower $\delta^{18}\text{O}$ of precipitation, evaporative enrichment has a large influence on the isotopic values of lake water. However, because of the small volume of these lakes, changes in the seasonality of precipitation ($\delta^{18}\text{O} = -10‰$ for summer months, but -25‰ for winter months), the proportion of different groundwater (-15 to -25‰) contribution, and atmospheric circulation paths would also have significant influences on the $\delta^{18}\text{O}$ values of the lake water. The 27 lake samples in $\delta^2\text{H} - \delta^{18}\text{O}$ space define a "lake line" ($\delta^2\text{H} = 7.4 * \delta^{18}\text{O} - 18$; $R = 0.74$), with an almost indistinguishable slope from the regional meteoric water line (7.6). Together with the 13 lakes in the Rice–Carlson chain, these 40 NGP lakes define a lake line with a slope of 6.7. These lake lines, rather than being simple evaporation lines, result from differences among lakes in the isotopic composition of input water (δ_{in}). An evaporation line for an individual lake over time (or for several lakes in a region) assumes that there is a constant δ_{in} , and that the slope of the evaporation line is mostly determined by atmospheric parameters such as relative humidity and isotopic composition of precipitation (Gat & Bowser, 1991; Gat, 1995). The evaporative evolution, or the distance from δ_{in} along the evaporation line is mostly determined by hydrologic factors such as the ratio of inflow rate

to evaporation rate (F_{in}/E). However, if δ_{in} is not constant over time, there won't be a fixed evaporation line, but rather a series of evaporation lines starting from different δ_{in} and ending at lake water isotopic values.

Our study region has a relative humidity of about 0.53 (for Williston, North Dakota), which would generate an evaporation line with a theoretically-predicted slope of <5 , given the region's isotopic composition of atmospheric moisture (Gat, 1995; his Figure 2). The lakes in the Rice–Carlson chain have a slope of 5.9 compared to 7.4 for other NGP lakes, which suggests that there are some hydrologic constraints acting in this setting. The significant evaporation effect ($\Delta^{18}O = 10$ to 15‰) is most likely responsible for increasing $\delta^{18}O$ of lake water to -1 to -6‰ from the original input values of -11 to -25‰. But within the $\delta^{18}O$ range of lake water, the slight variation in relative groundwater contributions and different degrees of modifications could easily cause a 2-3‰ fluctuation of lake water $\delta^{18}O$ as observed.

Hydrochemistry of the chain of lakes

The variable chemistry of evaporative groundwater-fed lakes in the NGP results from the chemical variability of source groundwater, which in turn reflects aquifer chemistry and mineralogy. Donovan & Ross (1994) maintain that lake chemistry in the NGP is governed by both groundwater flow path/length and subsequent evaporative concentration at the surface in which other process (geochemical, climatic, and lake-aquifer hydraulics) are dominant controls. Accordingly, evaporative lakes derive much of their chemical character from the local groundwater feeding them and this is strongly influenced by position in the flow system at which a lake intercepts groundwater. Owing to hydraulic factors, shallow groundwater might be expected to supply the largest number of lakes and highest groundwater flux in a uniform, low relief topography. However, deep depressions might attract flow from deeper aquifer(s).

Groundwater in the NGP has high Ca concentrations, but with an apparent upper limit of

about 500 ppm (Figure 7A), while lakes tend to have lower Ca concentrations, owing to precipitation of calcium carbonate. Lake I is a very small lake at the southern end of the chain (Figure 2) and has the highest $\delta^{18}O$ value (Figure 4). However, it is perplexed that Lake I has a much higher Ca concentration than other lakes in the chain (Figure 7A), apparently suggesting a large proportion of groundwater in its hydrologic budget. The local groundwater tends to have lower Mg/Ca ratios than the lakes (due to high Ca concentrations in groundwater), but slightly higher salinity/conductivity compared to lake water (Figure 7B). Outwash lakes, such as in the Rice–Carlson chain, are strongly influenced by groundwater supplies that have high Mg/Ca ratios and high salinity (Figure 7B). Evaporation has two aspects of influence on hydrochemistry of lakes. Groundwater supplies solutes to lakes and increases lake salinity (elevating lake-water concentrations of both Ca and Mg), but has a limited role in altering Mg/Ca ratios.

The strong correlation between Mg/Ca ratios and salinity at low and intermediate salinity level, as observed in the chain of lakes, also appears in other NGP lakes (Figure 7B), as suggested by clustered data points below 10,000 $\mu S\ cm^{-1}$ conductivity. At Devils Lake, North Dakota, Engstrom & Nelson (1991) found that the relationship between Mg/Ca ratios and salinity had a break in slope around 9‰ salinity owing to non-conservative behavior of Mg at higher ionic strengths.

Paleoclimatic implications

Oxygen isotopes from sedimentary carbonates have frequently been used as proxies of temperatures in humid northern and mid-latitudes and of effective moisture (precipitation - evapotranspiration) in arid and semi-arid regions. Trace-element ratios (Mg/Ca, Sr/Ca) of endogenic carbonates and ostracode shells have also been used as proxies of lake salinity in arid regions. However, complex local hydrology in the NGP (and elsewhere) may obscure the climatic signals in these records and limit their utility as simple aridity proxies.

An ideal proxy for salinity is one that would record the values of a conservative solute, i.e., any solute that shows a linear relationship with salinity. Although solutes such as Na and SO₄ show such behavior (Figures 5A and 5B), there are no sedimentary proxies for these solutes except at very high salinities when Na and SO₄ evaporites precipitate. Mg/Ca ratios and salinity of lake waters in the Rice–Carlson chain show a good linear correlation at salinities lower than about 15 ‰ (Figure 5C). Lake waters with higher salinities (>15‰) show greater scatter (Figures 5C and 7B), suggesting precipitation of high-Mg calcite or some other Mg-mineral that removes Mg from water. Salinity increases in the direction of the shallow groundwater flow (Figure 5), suggesting that each lake acts as a source of evaporated water to the next lake down the chain. Sr/Ca ratios show a poorer correlation with salinity, presumably because higher Mg/Ca leads to inorganic precipitation of aragonite, which preferentially incorporates Sr (as opposed to calcite, which excludes Sr) (Eugster & Jones, 1979; Engstrom & Nelson, 1991; Haskell et al., 1996).

Ten of 13 lakes in the Rice–Carlson chain have $\delta^{18}\text{O}$ values of $6 \pm 1\text{‰}$ VSMOW, while at the same time span a large range in salinity (1 to 65‰). This pattern implies that evaporation does not necessarily control the $\delta^{18}\text{O}$ of the lake water (which is the basis to infer P-E from $\delta^{18}\text{O}$), but rather suggests that each lake has a different hydrologic budget with variable contribution and isotopic composition of source waters, or spatial variation in groundwater chemistry. Groundwater, an important component of lake water budget in this region, usually has lower $\delta^{18}\text{O}$ values than that of mean annual precipitation owing to selective recharge from snow melt. So interaction between lakes and groundwater plays an important role in determining lake isotopic composition.

Closed basin lakes from the same geographic region are typically thought to fall on a single evaporation line which intersects the global or regional MWL at a point similar to the isotopic value of the mean annual precipitation for the region (Gat, 1995). This is not the case for 13 lakes in the Rice–Carlson chain. In particular, the low isotopic values for Spring D suggest that

groundwater is preferentially recharged by snowmelt. Such seasonally “biased” recharge may be common in the NGP (D. Rosenberry, pers. comm., 1999). The isotope values of the lakes are bracketed by the evaporation line coming off the groundwater recharge and another evaporation line coming off the mean annual precipitation. A particular lake’s water will fall within this field, depending on the ratio between the surface input and groundwater inflow. This ratio will vary depending on the annual amount of precipitation and its seasonality. NGP lakes are known to undergo considerable isotopic evolution during the summer months (Xia et al., 1997a), but the lakes do not evolve from the same isotopic starting point after every ice-out, nor do they evolve in a similar manner from year to year. Rather, the starting point depends on the amount of snowmelt runoff as well as summer temperature and precipitation. The $\delta^{18}\text{O}$ value of lake water, and thus of endogenic carbonate and ostracode shells, reflect the sum total of all processes that control the hydrologic budget of the lake.

A number of published studies document the complex interaction of lakes and groundwater in the NGP (e.g., LaBaugh et al., 1987; Winter & Rosenberry, 1995). Smith et al. (1997) suggest that during the mid-Holocene ostracode Mg/Ca and $\delta^{18}\text{O}$ records were contradictory because of changes in the isotopic composition of groundwater discharge to the lake. In fact, the divergent behavior hinted at the severity of the drought. Although valuable insight can still be gleaned from oxygen isotopes, reliable interpretations require additional independent proxies as well as an understanding of local hydrology.

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Table 1. Water isotopes and hydrochemistry of the 13 lakes and one spring in the Rice– Carlson chain in the northern Great Plains of North America.

Lake*	pH	Temp. (°C)	d ¹⁸ O (‰)	d ² H (‰)	Na (ppm)	K (ppm)	Cl (ppm)	SO ₄ (ppm)	Ca (ppm)	Mg (ppm)	Sr (ppm)	Mg/Ca (molar)	Sr/Ca (molar)	TDS (ppm)
Lake #1	8.70	19.2	-6.1	-61	342	83.9	72.1	1120	16.1	281.6	0.049	28.9	0.0014	2857
Rice Lake	8.42	18.0	-6.1	-59	123	40.8	29.0	588	23.7	198.9	0.057	14.0	0.0011	1728
Lake A	8.15	21.8	-8.5	-75	30	8.4	7.3	157	29.9	57.4	0.077	3.2	0.0012	587
Lake B	8.56	20.7	-6.0	-60	449	75.7	57.0	1007	20.6	209.3	0.067	16.8	0.0015	2893
Lake B2	8.42	19.9	-5.4	-56	3549	129.6	182.0	8682	42.6	482.6	0.149	18.7	0.0016	13890
Lake C	8.53	22.0	-5.4	-56	17117	487.3	659.8	38055	36.6	862.4	0.200	38.8	0.0025	58354
Lake D	8.76	19.8	-6.0	-60	12787	390.8	642.8	28210	12.4	737.7	0.166	97.9	0.0061	44095
Lake E	8.76	14.9	-3.7	-44	24231	572.0	568.9	54907	29.6	1798.6	1.205	100.3	0.0187	85610
Lake F	8.73	18.5	-5.4	-56	6514	228.5	187.9	15292	12.0	875.7	0.117	120.3	0.0045	25403
Lake G	8.60	18.3	-6.5	-64	240	23.0	9.9	579	19.7	114.9	0.080	9.6	0.0019	1657
Lake H	8.63	19.1	-5.9	-56	2709	143.0	108.6	7372	13.3	627.8	0.096	77.7	0.0033	12034
Lake I	8.30	22.1	-0.2	-25	5416	320.0	223.2	18352	346.7	1660.9	4.925	7.9	0.0065	26900
Lake J	8.20	20.5	-6.7	-62	141	13.8	5.6	342	30.3	88.3	0.254	4.8	0.0038	1185
Lake D	7.31	11.9	-16.7	-129	1359	8.4	21.3	1875	43.9	33.6	1.440	1.3	0.0150	4947
Spring														

* Some lakes have official names on USGS topographic maps: Lake A = Tangedahl Lake; Lake B = Lake Vernon; Lake G = Nelson Lake; and Lake J = Carlson Lake.

Table 2. Water isotopes ($\delta^2\text{H}$ and $\delta^{18}\text{O}$) of 27 additional lakes in the northern Great Plains.

Lake	Latitude (°N)	Longitude (°W)	$\delta^{18}\text{O}$ (‰)	$\delta^2\text{H}$ (‰)
Albert, SD	44.53	97.15	-1.9	-23
Bitter, SD	45.28	97.32	-1.5	-24
Clear, MN	43.63	94.97	-2.7	-52
Cochrane, SD	44.72	96.48	-1.8	-25
Coldwater, ND	46.02	99.07	-2.1	-30
Dead Coon, MN	44.37	96.23	-4.6	-41
Fife, Sask.	49.23	105.88	-2.9	-48
George, ND	46.75	99.50	-3.4	-46
Godwit, ND	47.95	101.52	-2.9	-43
Hazelsen, SD	45.52	97.47	-0.5	-26
Herman, SD	44.00	97.17	-0.6	-17
Horseshoe, ND	47.88	98.80	-3.7	-43
Isabel, ND	46.82	99.73	-2.1	-41
Madison, SD	43.95	97.00	-1.8	-28
Medicine, SD	44.82	97.35	-1.9	-31
Minnewaska, MN	45.62	95.42	-2.5	-36
Moon, ND	46.85	98.17	-3.0	-39
Nelson, ND	47.92	101.47	-3.9	-49
Norden, SD	43.58	97.20	-2.3	-41
Oakwood, SD	44.43	96.97	-1.5	-22
Pickerel, SD	45.50	97.38	-4.1	-38
Poinsett, SD	43.53	97.08	-1.8	-28
Roslyn, SD	45.52	97.43	-2.3	-28
Round, ND	48.03	100.00	-2.7	-46
Roy, SD	45.72	97.45	-1.1	-29
Skoyal, ND	47.92	101.47	-2.5	-44
Waubay, SD	45.42	97.40	-1.8	-29

Figure Captions:

Figure 1. Location and geologic setting of the Rice–Carlson chain of lakes in the northern Great Plains. **A.** Location of the study site (Rice Lake, North Dakota) and climate stations (Wynyard, Saskatchewan; Gimli, Manitoba; and Grand Forks, North Dakota) with isotope measurements of precipitation in central North America. **B.** Glacial deposits around the Rice–Carlson chain of lakes (from Bluemle, 1991). The lake chain is situated in a meltwater channel filled with outwash sand and gravel and surrounded by till uplands with low hydraulic conductivity.

Figure 2. Topographic map of the Rice–Carlson chain of lakes and surrounding region. The solid-filled lakes with labels were sampled for this study. Based on surface topography, groundwater appears to flow from both ends toward the middle of the chain.

Figure 3. Spatial variation of the total dissolved solids (TDS; parts per thousand) in lakes of the Rice–Carlson chain, ND.

Figure 4. Isotopic composition of 13 lakes in the Rice–Carlson chain. **A.** Correlation of $\delta^{18}\text{O}$ and total dissolved solids (TDS) from 13 lake and one spring samples. **B.** Plot of isotopic data from 13 lake and one spring in $\delta^2\text{H} - \delta^{18}\text{O}$ space.

Figure 5. Hydrochemistry of 13 lakes in the Rice–Carlson chain. **A.** Na vs. TDS (total dissolved solids). **B.** SO_4 vs. TDS. **C.** Mg/Ca vs. TDS. **D.** Sr/Ca vs. TDS.

Figure 6. A plot of $\delta^2\text{H}$ vs. $\delta^{18}\text{O}$ in precipitation, groundwater and lake waters from the northern Great Plains (Gimli and Wynyard precipitation data from the IAEA-GNIP database; Grand Forks precipitation and groundwater isotope data from Matheney & Gerla, 1996; NGP lake-water isotopes from this study). The regional meteoric water line (MWL) combined from the three precipitation-isotope stations has the form of $\delta^2\text{H} = 7.6 \cdot \delta^{18}\text{O} + 2.0$ ($R = 0.99$). It has very similar slope but lower intercept compared to the global MWL ($\delta^2\text{H} = 8 \cdot \delta^{18}\text{O} + 10$; Craig, 1961). Inset shows detail of isotopic values of NGP lake samples which fit a “lake line” with a slope of 6.7. Alone, the Rice–Carlson lakes fit a line with a slope of 5.9 (Figure 3B), whereas other NGP lakes have a slope of 7.4. The main plot shows all individual isotope measurements from the NGP lakes, while the inset shows averages for each lake. See text for additional discussion.

Figure 7. Hydrochemistry of lakes and groundwater in the northern Great Plains. **A.** Relationship between Mg and Ca concentrations. This plot shows that groundwater has high Ca concentration with an apparent upper limit about 500 ppm. **B.** Correlation of Mg/Ca ratios and conductivity. In most cases, conductivity (in $\mu\text{S cm}^{-1}$) is equal to total dissolved solids (TDS or salinity). Groundwater and other NGP lake data are from Eisenlohr et al. (1972), Swanson et al. (1988), Engstrom & Nelson (1991), and J.J. Donovan (unpublished data). See text for details.

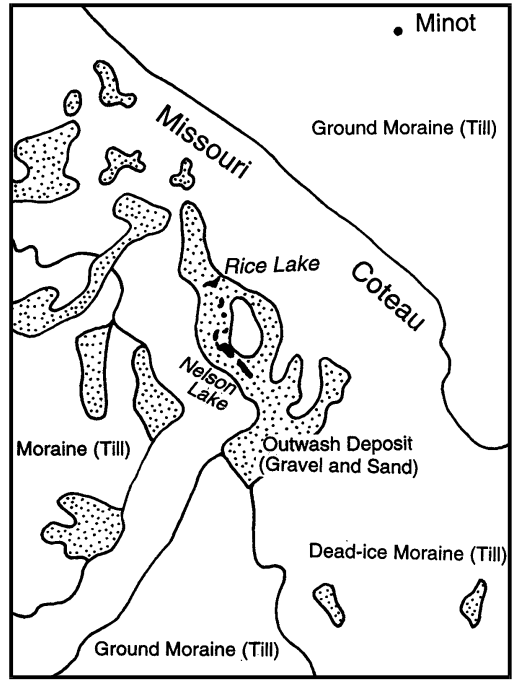
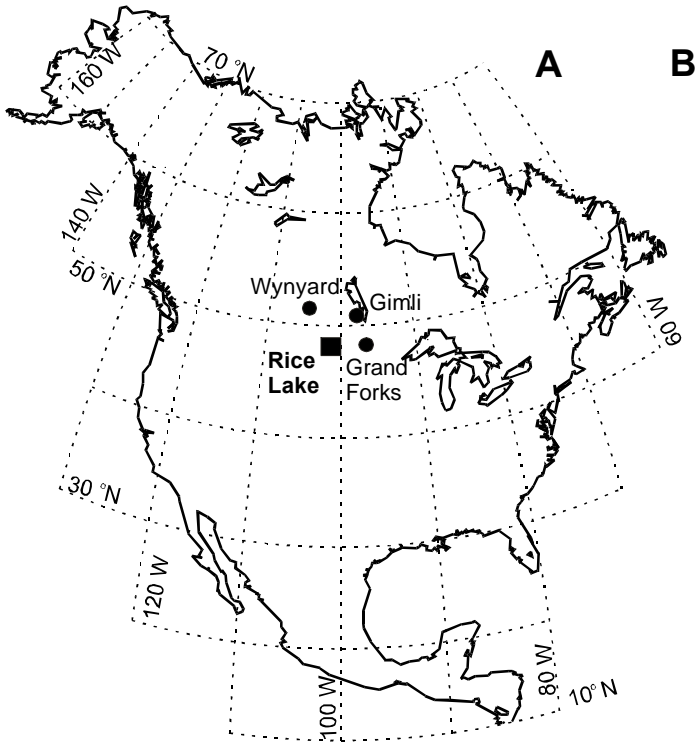


Fig. 1 (Yu et al.)

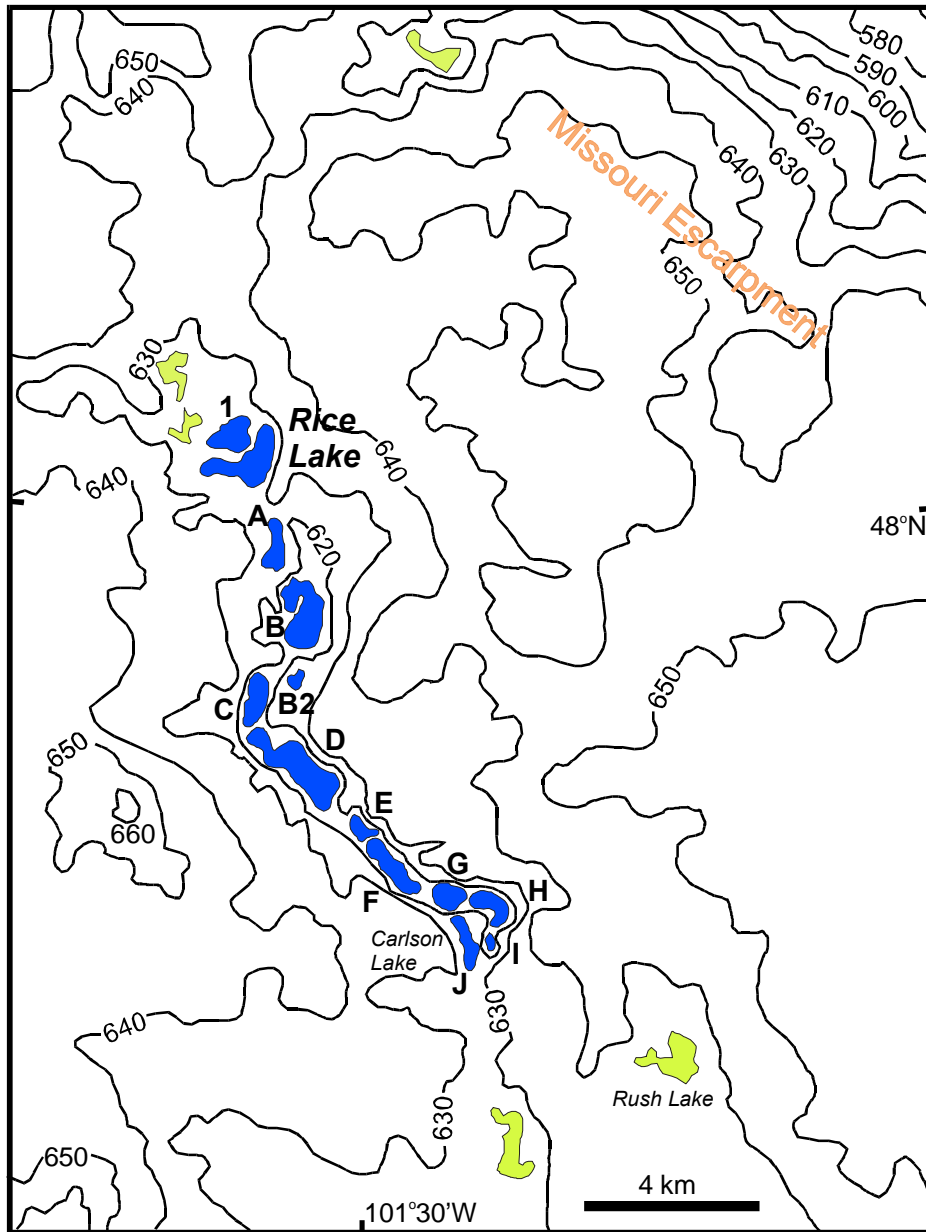


Fig. 2 (Yu et al.)

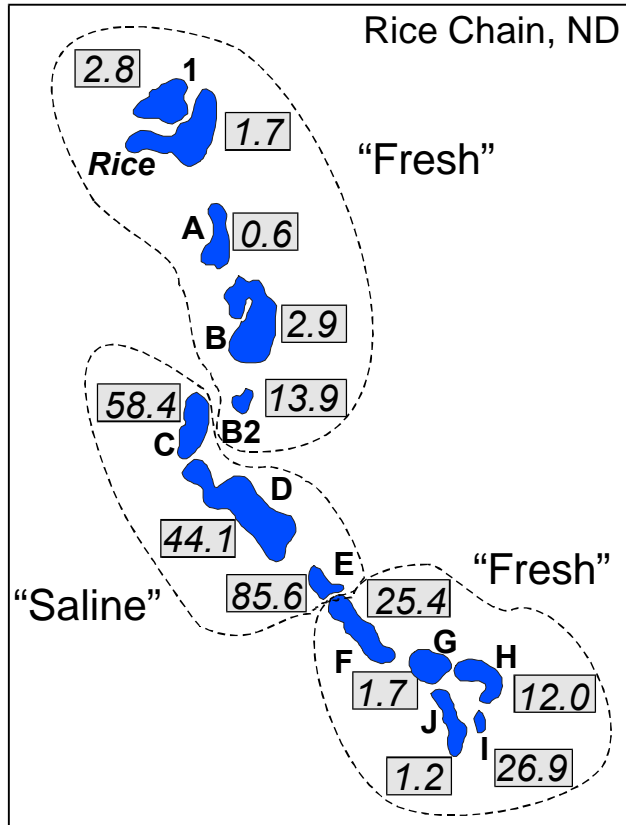


Fig. 3 (Yu et al.)

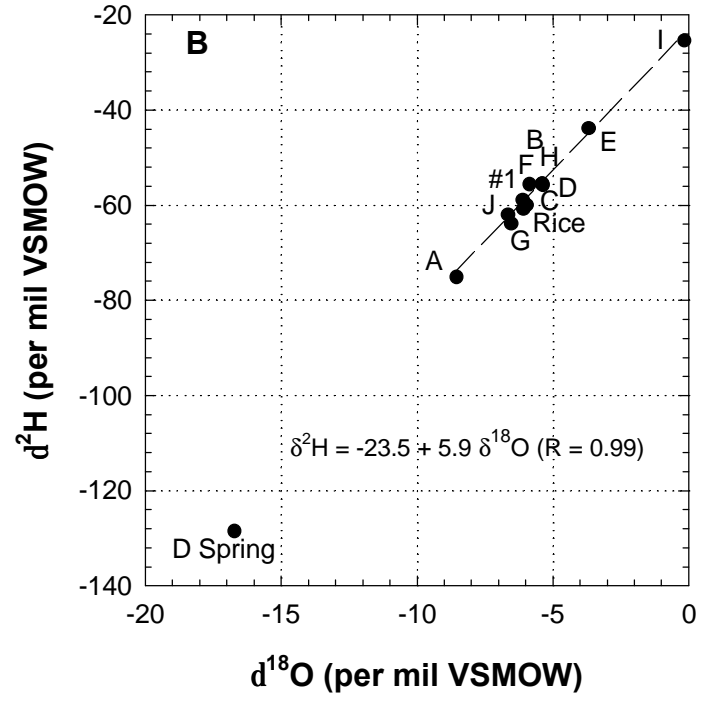
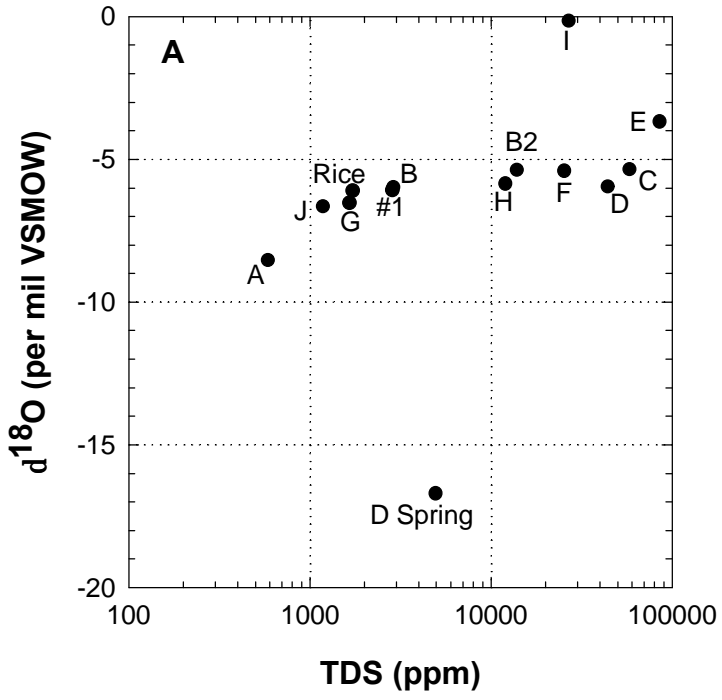


Fig. 4 (Yu et al.)

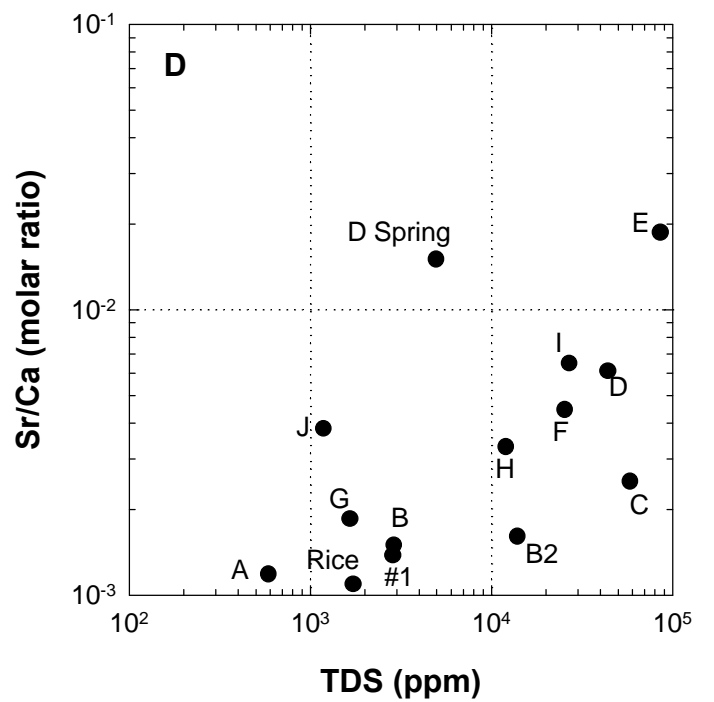
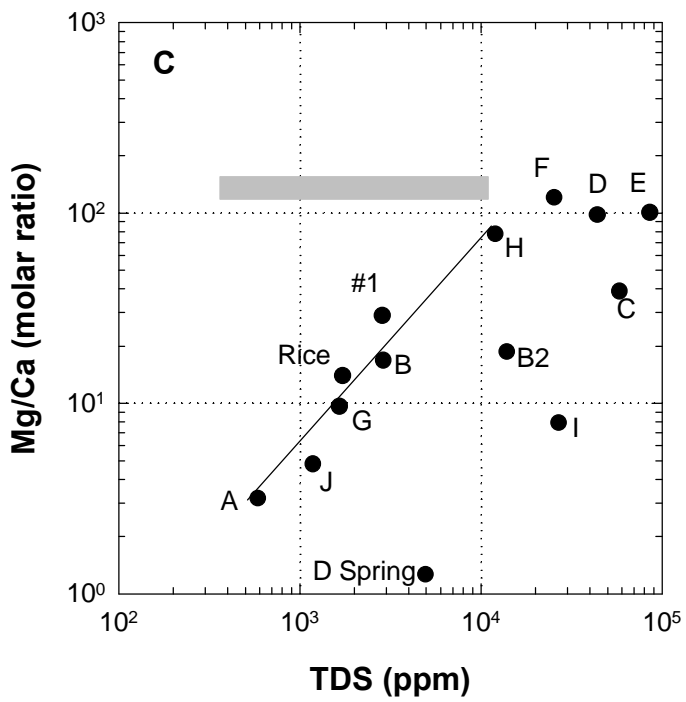
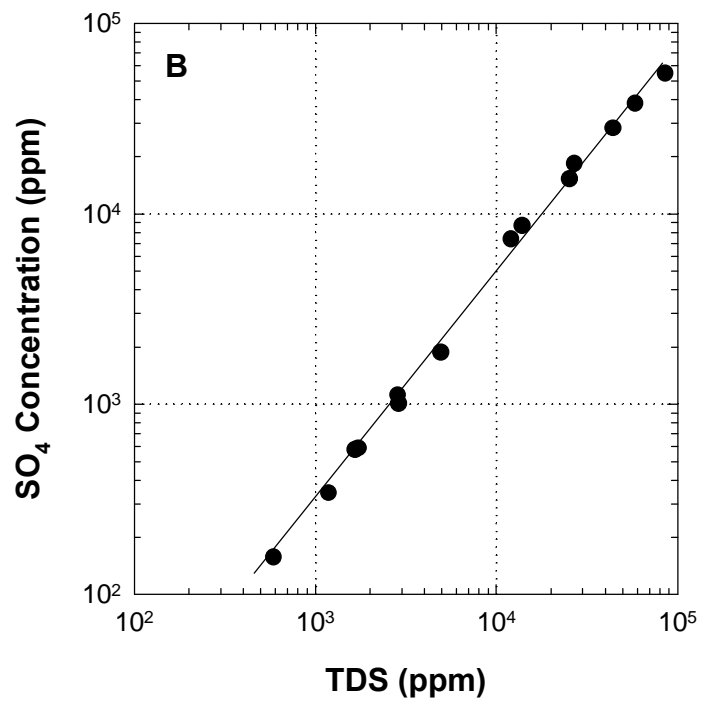
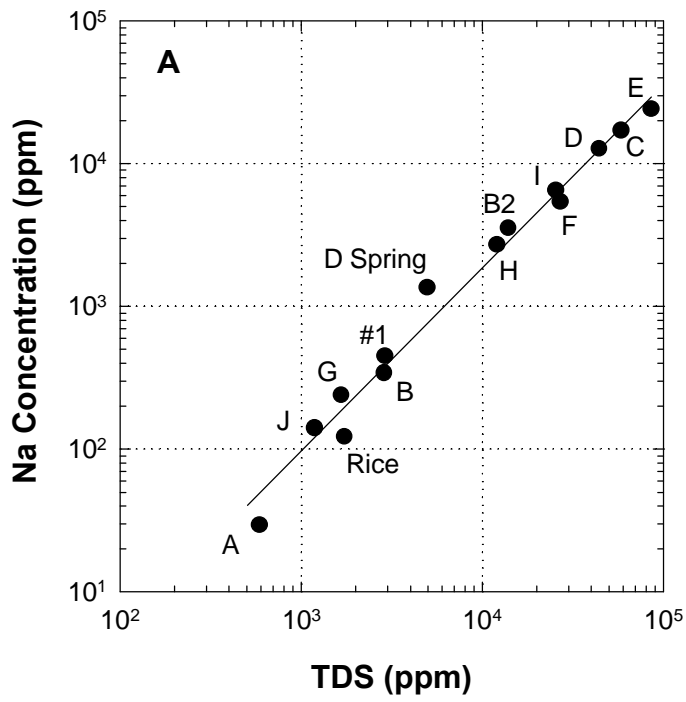


Fig. 5 (Yu et al.)

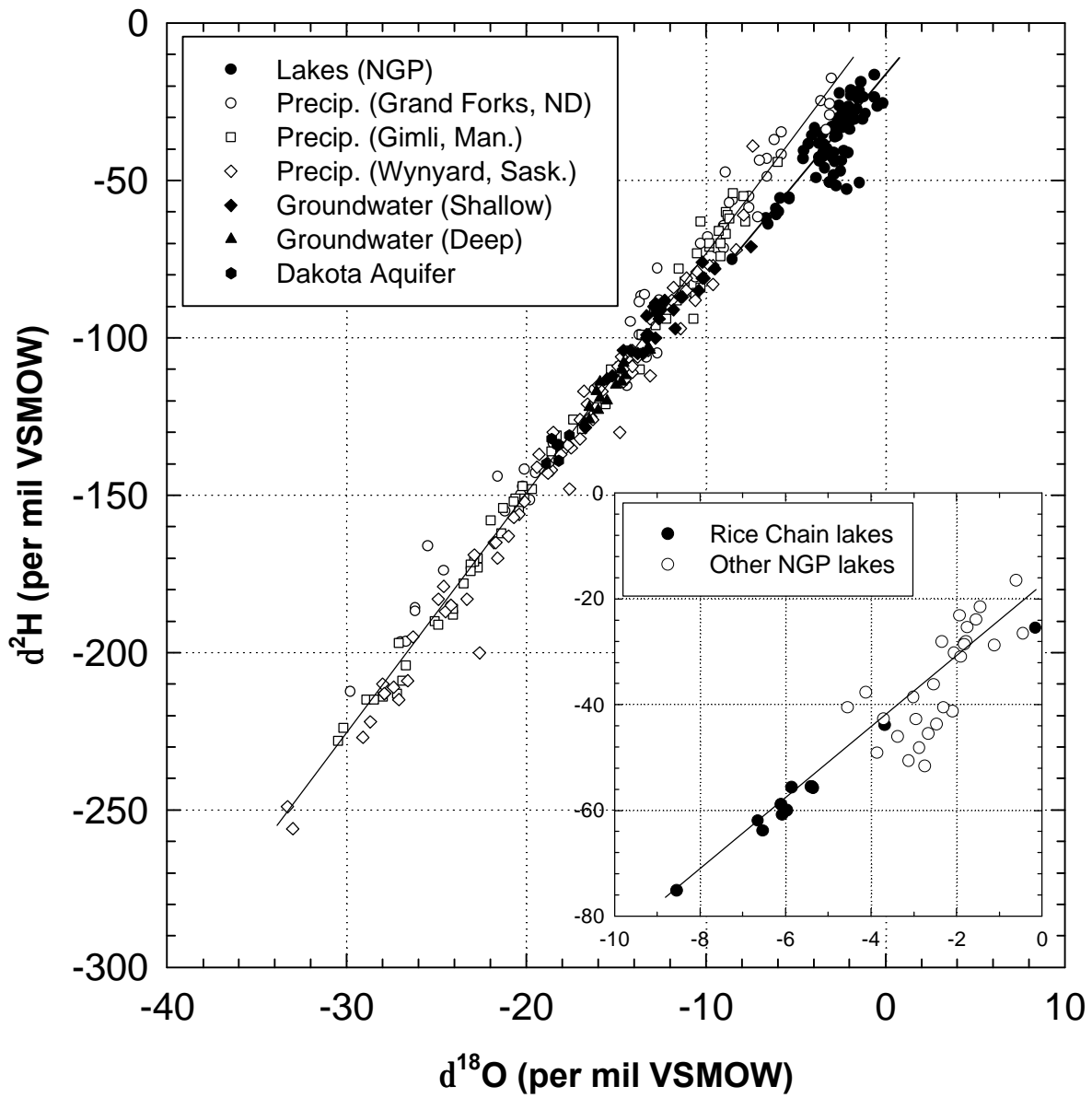


Fig. 6 (Yu et al.)

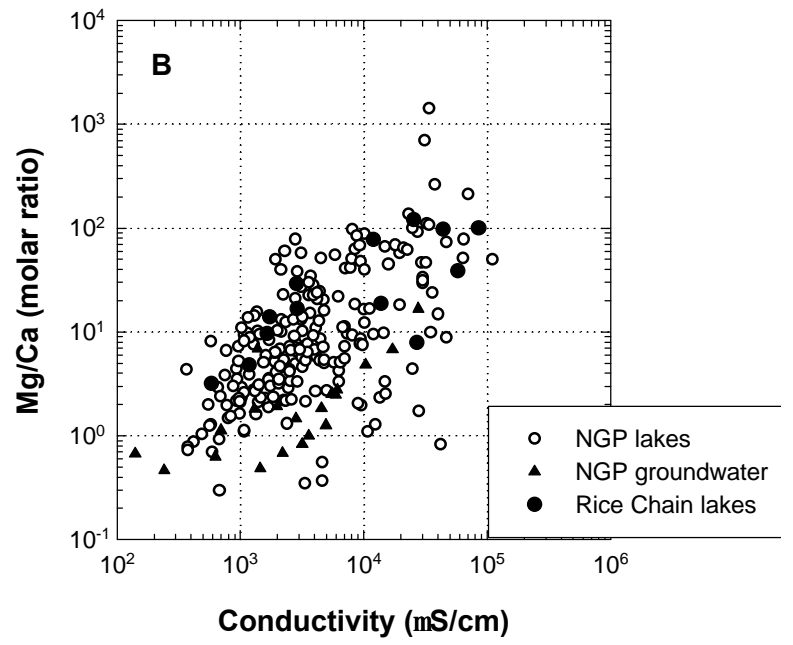
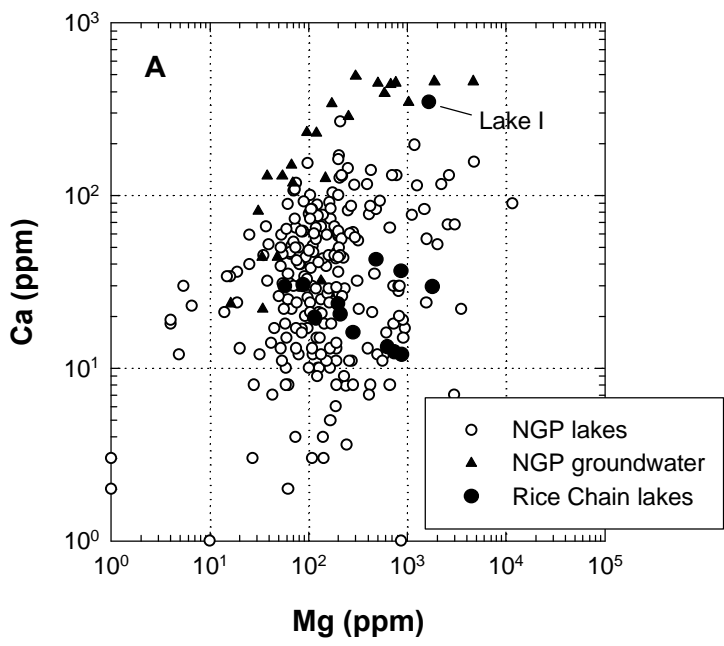


Fig. 7 (Yu et al.)