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A 1000-year record of dry conditions in the eastern Canadian prairies reconstructed from oxygen and carbon isotope measurements on Lake Winnipeg sediment organics

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ABSTRACT: A short sediment core (162 cm), covering the period AD 920–1999, was sampled from the south basin of Lake Winnipeg for a suite of multi-proxy analyses leading towards a detailed characterisation of the recent millennial lake environment and hydroclimate of southern Manitoba, Canada. Information on the frequency and duration of major dry periods in southern Manitoba, in light of the changes that are likely to occur as a result of an increasingly warming atmosphere, is of specific interest in this study. Intervals of relatively enriched lake sediment cellulose oxygen isotope values ($\delta^{18}O_{cellulose}$) were found to occur from AD 1180 to 1230 (error range: AD 1104–1231 to 1160–1280), 1610–1640 (error range: AD 1571–1634 to 1603–1662), 1670–1720 (error range: AD 1643–1697 to 1692–1738) and 1750–1780 (error range: AD 1724–1766 to 1756–1794). Regional water balance, inferred from calculated Lake Winnipeg water oxygen isotope values ($\delta^{18}O_{inf\text{-}lw}$), suggest that the ratio of lake evaporation to catchment input may have been 25–40% higher during these isotopically distinct periods. Associated with the enriched $\delta^{18}O_{cellulose}$ intervals are some depleted carbon isotope values associated with more abundantly preserved sediment organic matter ($\delta^{13}C_{OM}$). These suggest reduced microbial oxidation of terrestrially derived organic matter and/or subdued lake productivity during periods of minimised input of nutrients from the catchment area. With reference to other corroborating evidence, it is suggested that the AD 1180–1230, 1610–1640, 1670–1720 and 1750–1780 intervals represent four distinctly drier periods (droughts) in southern Manitoba, Canada. Additionally, lower-magnitude and duration dry periods may have also occurred from 1320 to 1340 (error range: AD 1257–1363), 1530–1540 (error range: AD 1490–1565 to 1498–1572) and 1570–1580 (error range: AD 1531–1599 to 1539–1606). Copyright © 2009 John Wiley & Sons, Ltd.

KEYWORDS: oxygen and carbon isotopes; sediment cellulose; Lake Winnipeg; droughts.

Introduction

The frequency and severity of distinct dry periods (i.e. evaporative water loss *E*, exceeding precipitation *P*) affecting the runoff of the major river basins in Canada's western interior represents a data source that is an essential part of understanding future droughts and dry periods in this sensitive region. For this study a 162 cm core (99–900 Core 8) was retrieved from the south basin of Lake Winnipeg for the purpose of reconstructing recent millennial dry periods in southern Manitoba in addition to conditions conducive to flooding within the Red River Basin (Simpson *et al.*, 2009).

Lake Winnipeg has a surface area of 24 000 km^2, making it the 11th largest lake in the world. Its broad catchment area

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(983 000 km^2) extends between Lake Superior to the east and the Rocky Mountains to the west. Lake Winnipeg consists of two north and south basins separated by a narrow passage (Fig. 1). The shallow (average depth of 9 m; maximum depth of 12 m) and much smaller south basin has a surface area of 3600 km^2. Lake Winnipeg south basin sediments consist of a lower glaciolacustrine unit deposited by glacial Lake Agassiz, overlying till or bedrock. Lake Winnipeg sediments (varying between 1 and 7 m) unconformably overly the Lake Agassiz sediments, representing postglacial sedimentation that began approximately 4000 a BP (Lewis *et al.*, 2001).

Two major rivers, the Red and Winnipeg, enter the south basin. The Red River discharges nearest the Core 8 site and is therefore assumed to have the most direct influence on the short core (the Winnipeg River discharges into the adjacent and more remote Traverse Bay) (Fig. 1). This discussion centres upon the occurrences of major dry periods in southern Manitoba during the last 1000 a as evidenced by the down-core (Core 8) record of cellulose oxygen and organic carbon isotopic variations in direct comparison with other substantiating multi-proxy analyses. These reconstructed dry periods are further quantified through an estimation of the 1000 a hydrological variability (evaporation/input ratio) in southern Manitoba.

Oxygen isotopic analyses on lake sediment cellulose ($\delta^{18}O_{cellulose}$) have proven to be a very effective tool for reconstructing palaeoclimatic and hydrologic conditions in

Figure 1 Map of the south basin of Lake Winnipeg, Manitoba, Canada showing the site location of Core 8 (50°34.343′ N, 96° 50.004′ W; 8.4 m water depth)

lake watersheds (MacDonald *et al.*, 1993; Duthie *et al.*, 1996; Wolfe *et al.*, 1996, 2001a; Abbott *et al.*, 2000). Success is directly related to the fact that the oxygen isotopic composition of cellulose derived from lake sediment organics is generally controlled by only two factors: (1) the isotopic composition of input waters that are determined by precipitation, surface runoff and groundwater; and (2) hydrologic processes such as evaporation which enriches the isotopic composition of the input water in ^{18}O. The isotopic composition of precipitation, surface runoff and groundwater can vary with climate as regional temperatures, precipitation seasonality and air mass distributions change over time. The magnitude of the (secondary) enrichment in the lake water is proportional to the amount of evaporation from the lake water reservoir. Dynamic evaporation from lake water reservoirs is related to low atmospheric humidity. Consequently, lake sediment cellulose, systematically offset from the resident lake water, tends to be relatively enriched in ^{18}O during drier than normal periods. Furthermore, lake sediment cellulose records a consistent, non-temperature-dependent 27–28‰ ^{18}O enrichment, allowing for uncomplicated inferences of lake water isotopic compositions (DeNiro and Epstein, 1981; Sternberg, 1989; Yakir, 1992). Therefore, since distinctly drier periods in southern Manitoba should have detectable effects on the overall water balance of Lake Winnipeg, it is possible to further quantify these effects by assessing the oxygen isotope–mass balance between waters input into the Lake Winnipeg catchment area and water vapour evaporated from the lake's surface (Edwards *et al.*, 2004). In fact, estimation of the evaporation to inflow ratio (*E*/*I*) for the purpose of reconstructing the past hydrologic variability from isotopic palaeorecords have been successfully accomplished through isotope–mass balance considerations in a number of cases (McDonald *et al.*, 1993; Wolfe *et al.*, 2001a; Hammarlund *et al.*, 2003).

In terms of carbon isotopes, photosynthesising organisms (phytoplankton) preferentially incorporate ^{12}C from dissolved inorganic carbon (DIC) pool, which is usually assumed to be in isotopic equilibrium with atmospheric carbon (Schelske and Hodell, 1991, 1995; Helie *et al.*, 2002). While the carbon isotopic composition of the DIC reservoir can be influenced by lake water temperature, CO_2 exchange between the atmosphere and lake water (isotopic equilibrium is assumed in this case), and the influx of CO_2 from both the decomposition of ^{13}C depleted organic matter and the chemical weathering of ^{13}C enriched carbonate rocks in the catchment (both assumed to be insignificant in this case) it is primarily regulated by seasonal changes in photosynthesis and respiration (Stuiver, 1975; McKenzie, 1985; Buchardt and Fritz, 1980; Hakansson, 1985; Talbot and Johannessen, 1992). Therefore, the stable isotopic composition of organic carbon in lake-bottom sediments can be used as a proxy for palaeoproductivity (palaeoactivity) in lakes since enhanced productivity during certain times can

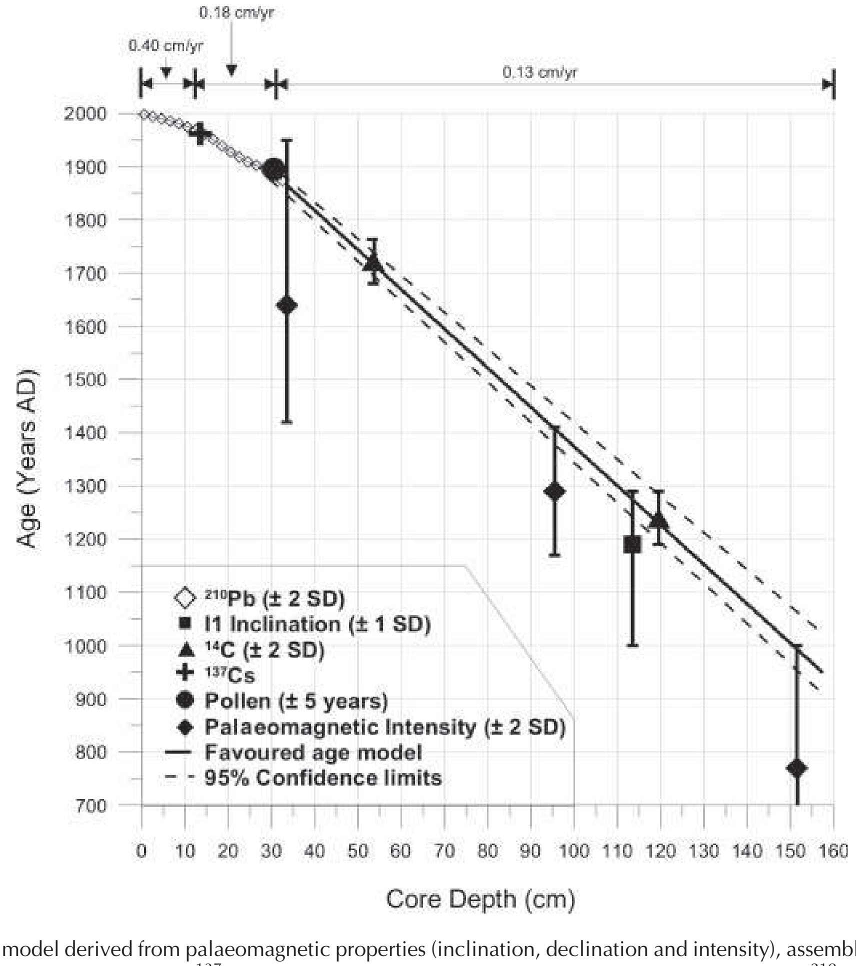

Figure 2 The Core 8 age model derived from palaeomagnetic properties (inclination, declination and intensity), assemblages, ^{210}Pb, ^{137}Cs, ^{14}C and pollen analyses (SD, standard deviation). Peak ^{137}Cs activity occurred at the 14 cm core level corresponding to a ^{210}Pb age of AD 1965. The first appearance of Salsola (Russian thistle) implies a marker date of AD 1895 at the 32 cm core level. The linear relationship Age (AD) = 211.19 – 8.0981 (depth cm) was employed between core depths 33 and 158 cm based on the oldest ^{210}Pb age (33 cm) and the ^{14}C date (wood fragment) at 119 cm with an age of AD 1240 ± 50 a. A palaeomagnetic inclination feature I1 at 114 cm with an age of AD 1190 validates the ^{14}C date. Upper measurement uncertainty ranges for the three palaeomagnetic intensity ages (34, 95 and 151 cm) overlap with the age model age estimates (Simpson *et al.*, 2003)

result in ^{13}C enrichment of the lake's DIC reservoir, which causes progressive enrichment of the ^{13}C composition of the autochthonous organic matter (Schelske and Hodell, 1991, 1995).

It is very likely that sustained dry periods would generally result in lower water input into lakes and increased water residence time due in large part to reduced outflow rates. Increased evaporation from the surface of the lake would elevate the evaporation/input ratio, resulting in subsequent ^{18}O enrichments of both the lake water and resident cellulose producers. Reduced nutrient levels in response to anomalously lower water levels during dry conditions force resident organisms into reduced activity and productivity.

Materials and methods

Core 8 is one of 15 10 cm diameter gravity cores (varying in length between 107 and 170 cm) sampled in the south basin of Lake Winnipeg (Fig. 1) in August 1999. All the cores retrieved were sealed with silicone and eventually transported upright by a refrigerated truck to the Geological Survey of Canada core storage facility in Dartmouth, Nova Scotia. Core 8 (162 cm in length) was split longitudinally, digitally imaged and subsampled at 1 cm intervals for a total of 158 individual samples. The core age model established (Fig. 2) relies upon palaeomagnetic properties (inclination, declination and intensity), assemblages, ^{210}Pb, ^{137}Cs, ^{14}C and pollen analyses.

^{210}Pb ages were calculated from unsupported ^{210}Pb activity for the uppermost 33 intervals of Core 8, using the constant rate of supply (CRS) method (Oldfield and Appleby, 1984; Wilkinson and Simpson, 2003). The ^{210}Pb dates obtained are considered reliable since the peak ^{137}Cs activity occurred during the AD 1965 ^{210}Pb age (Fig. 2). Palynological analysis revealed the first appearance of Salsola (Russian thistle) at the 32 cm level of Core 8 (Anderson, 2003). This implies a marker date of AD 1895 at the 32 cm level corresponding to the first appearance of this exotic species in the Red River Basin (Dewey, 1985; Jacobson and Engstrom, 1989). Between the 33 and 158 cm core depths the linear relationship Age (AD) = 211.19 – 8.0981 (depth cm) was employed based on the oldest ^{210}Pb age (33 cm) and the ^{14}C dates (wood fragments) at 54 and119 cm with ages of AD 1710 ± 40 and 1240 ± 50 a, respectively. A palaeomagnetic inclination feature I1 at 114 cm with an age of AD 1190 (assuming 1σ uncertainty ranging from AD 1000 to 1290) (Lund, 1996) serves to validate the older ^{14}C date. Additionally, the three palaeomagnetic intensity estimates (34, 95 and 151 cm levels) that suggest ages older than the age model estimates all overlap with the age model within their respective upper measured uncertainty ranges (Fig. 2) (Simpson *et al.*, 2003).

The sediment samples were first treated with 10% HCl to remove carbonate material. The samples were then rinsed repeatedly with distilled water and freeze-dried. Carbon and nitrogen elemental analysis (% C, % N) were completed on Core 8 at a 1 cm resolution. The sediment samples were sieved (<500 μm) to eliminate any macrofossil plant debris of possible terrestrial origin. Additional sample treatment involving solvent extraction (2:1 chloroform:methanol and ethanol), bleaching

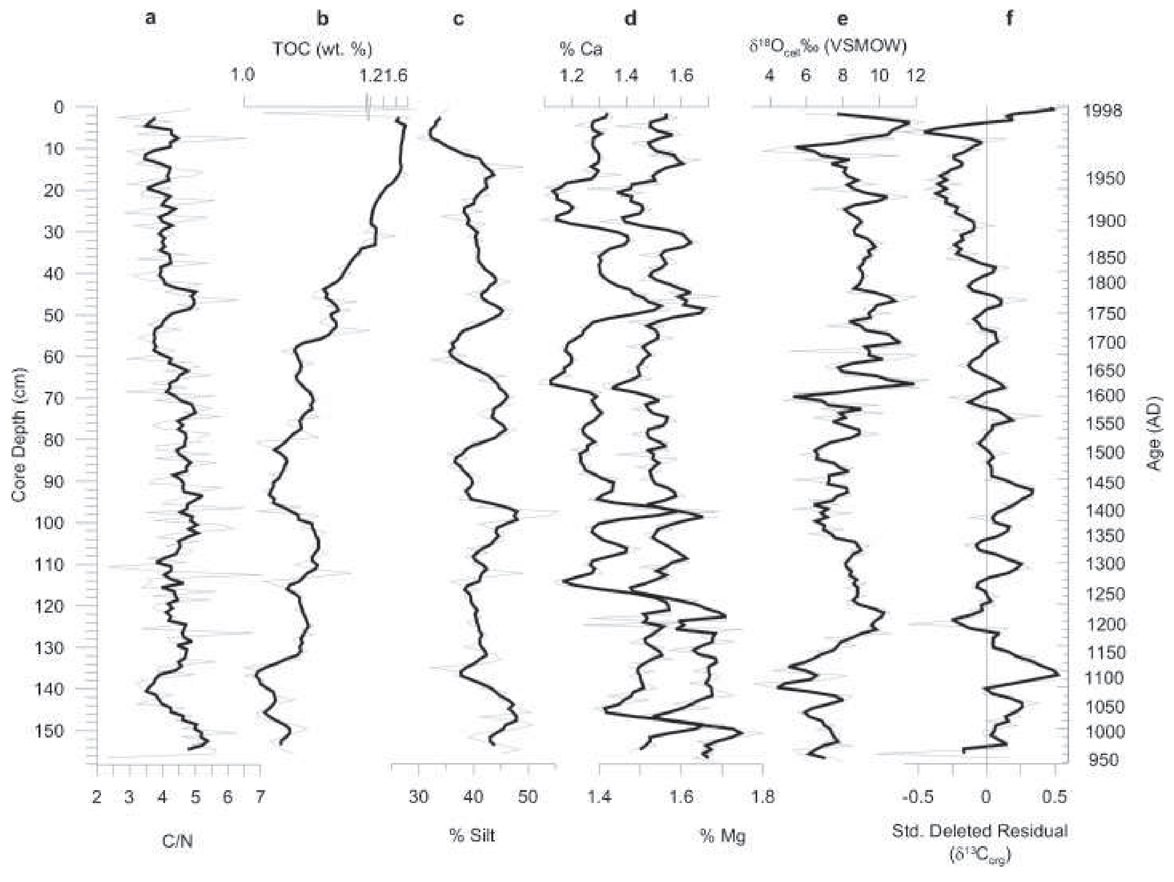

Figure 3 Core 8 depth series (dark lines are three-point running averages) of sediment (a) carbon to nitrogen (C/N) ratios, (b) wt% of total organic carbon (TOC), (c) silt percents, (d) calcium (Ca), left, and magnesium (Mg), right, %, (e) extracted cellulose oxygen isotopic compositions ($\delta^{18}O_{cell}$) with respect to VSMOW and (f) organic matter standardised deleted residual (SDR) carbon isotopic compositions (SDR$\delta^{13}C_{OM}$). Note: the ^{13}C Suess effect (Keeling *et al.*, 1979) was removed from the measured $\delta^{13}C_{OM}$ values (post AD 1850) prior to SDR analysis according to the Saurer *et al.* (1997) method

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and alkaline hydrolysis removed non-cellulose organic constituents, oxyhydroxide leaching removed possible inorganic oxygen sources and sodium polytungstate heavy liquid separation isolated the sediment cellulose (Green, 1963; Wolfe *et al.*, 2001b).

The organic matter sediment samples were combusted in an attached elemental analyser (EuroVector®) and the carbon dioxide gas (for carbon isotopes) produced was carried in a stream of helium (carrier gas) to the University of Winnipeg Isotope Laboratory (UWIL) continuous-flow isotope ratio mass spectrometer (GV Instruments IsoPrime®). The isolated sediment cellulose samples were pyrolysed to produce carbon monoxide for oxygen isotope analysis (Wolfe *et al.*, 2001b). The water oxygen and hydrogen isotope analyses were performed by standard CO_2 equilibration and manganese reduction methods, respectively, at the Environmental Isotope Laboratory (EIL), University of Waterloo (VG Instruments 903 and 602).

The isotope results are expressed as δ values that represent deviations per mil (‰) from the VPDB standard for carbon and the VSMOW standard for oxygen such that $\delta_{sample} = [(R_{sample}/R_{standard}) - 1] \times 10^3$, where R is the $^{13}C/^{12}C$ or $^{18}O/^{16}O$ ratio in the sample and standard. The $\delta^{13}C_{OM}$ and $\delta^{18}O_{cellulose}$ values have analytical uncertainties of ±0.1‰ and ±0.2‰, respectively. The water $\delta^{18}O$ and $\delta^{2}H$ values have analytical uncertainties of ±0.1‰ and ±2.0‰, respectively. Accuracy was obtained through the analysis of laboratory standards used for calibration of results.

Results and discussion

In the case of Lake Winnipeg it is possible to establish a link between the carbon isotope values for the dominantly autochthonous or lacustrine sediment organic matter and lake biological activity. Initially, for some large lakes there is a direct correlation between total phosphorus (TP) loading, the dominant control on lacustrine productivity, and $\delta^{13}C_{OM}$ values for organic matter in their sediments (Schelske and Hodell, 1995; O'Reilly *et al.*, 2003). As the input of nutrients and lake productivity increased, enrichment of $\delta^{13}C_{OM}$ values of organic matter occurred. Suitably, Mayer *et al.* (2003) report a positive correlation between an increase in Core 8 TP loading and a progressive $\delta^{13}C_{OM}$ enrichment starting in the late 19th century and continuing throughout the 20th century, suggesting a progressive increase in Lake Winnipeg productivity.

The average $\delta^{13}C_{OM}$ value is −26.3 ‰ (VPDB) with a narrow standard deviation of 0.26. To highlight the carbon isotope variations present in Core 8 sediment organics for each core depth, least squares regression procedures were applied to independent data from the previous 10 core depths to obtain a linear fit of carbon isotope ratio (Y) as a function of year (X) (Friedman *et al.*, 2007). The observed carbon isotope ratio for a specified core depth (observed Y) was compared to the predicted carbon isotope ratio (predicted Y) obtained from the least-squares regression line for the previous 10 core depths and the standardised deleted residual was computed using the

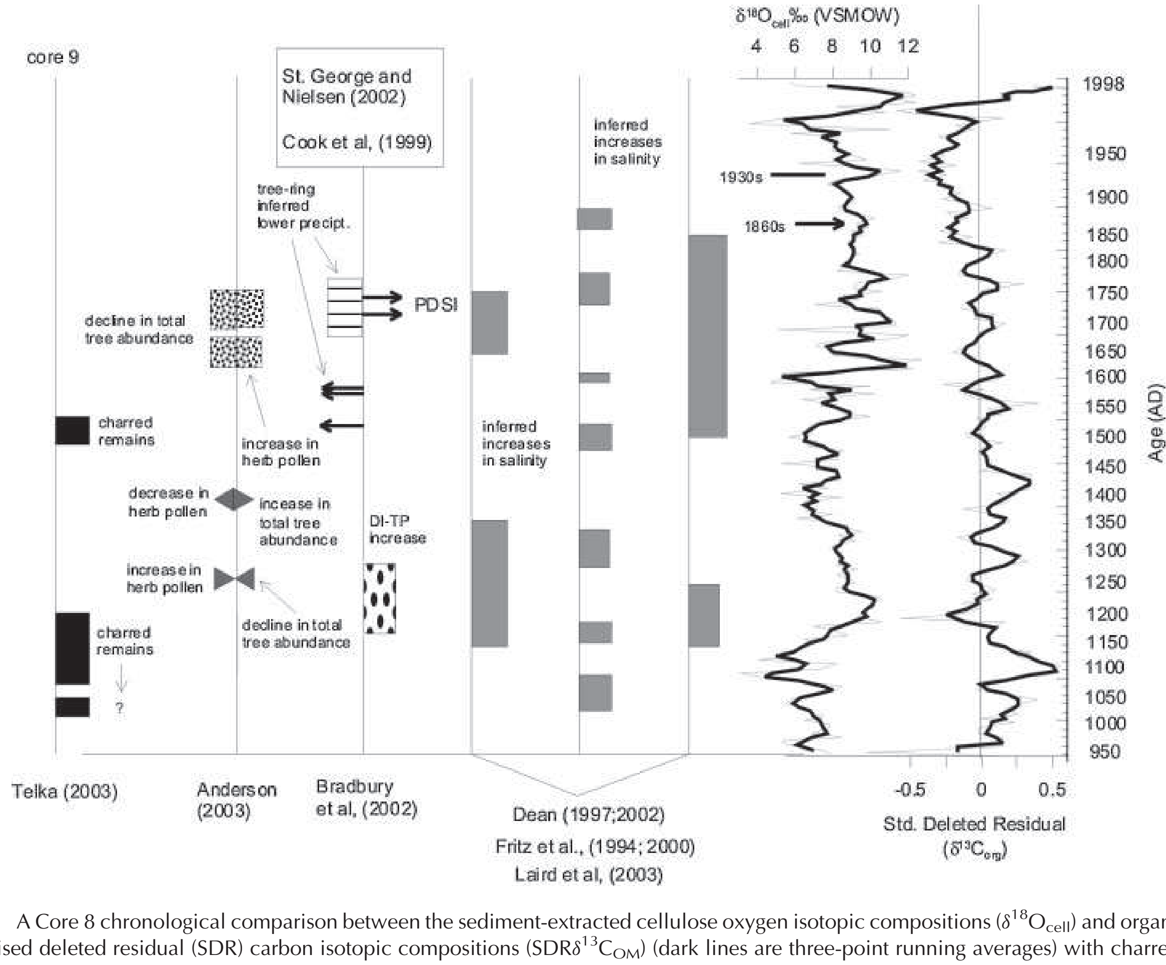

Figure 4 A Core 8 chronological comparison between the sediment-extracted cellulose oxygen isotopic compositions ($\delta^{18}O_{cell}$) and organic matter-standardised deleted residual (SDR) carbon isotopic compositions (SDR$\delta^{13}C_{OM}$) (dark lines are three-point running averages) with charred remains from Lake Winnipeg core 9 (Telka, 2003), Core 8 herb and total tree pollen (Anderson, 2003), diatom-inferred total phosphorus (DI-TP) from Elk Lake, Minnesota (Bradbury *et al.*, 2002), inferred lake salinities from Elk Lake, Minnesota (Dean, 1997, 2002), Moon, Coldwater and Rice Lakes, North Dakota (Fritz *et al.*, 2000), Devils Lake, North Dakota (Fritz *et al.*, 1994), Nora Lake, Manitoba (Laird *et al.*, 2003), upper Red River basin tree ring-inferred precipitation (St George and Nielsen, 2002), North Dakota and Minnesota tree ring-inferred palmer drought severity index (PDSI) (Cook *et al.*, 1999)

formula: standardised deleted residual = (observed *Y* predicted *Y*)/RMSE, (equation 1), where RMSE denotes the standard error of the residuals obtained from the linear fit to the 10 earlier core depths. The values of standardised residuals were used to scan for outlying data points. A plot of the standardised deleted residuals ((SDR$\delta^{13}C_{OM}$)), *versus* the corresponding age model years for the various core depths is depicted in Figs 3(f) and 4. Secondly, a purely lacustrine origin of the sediment organic matter is immediately suggested by C/N values <10 for Core 8 in Fig. 3(a) (C/N range: 2.3–7.1; average: 4.3) (Meyers *et al.*, 1984; Meyers, 1994; Meyers and Lallier-Vergès, 1999). However, Snowdon and Simpson (2003) report a contradiction between Core 8 Rock-Eval (an evaluation of the properties of sediment organic matter) results suggesting that most of the organic matter is derived from terrestrial sources and the low C/N results, implying a dominantly autochthonous organic matter source (Fig. 3(a)). Additionally, Snowden and Simpson (2003) show a strong inverse relationship between hydrogen and oxygen indices. These combined results strongly suggest that significant alteration of the lacustrine organic matter has occurred.

It is suggested by Snowden and Simpson (2003) that the majority of the organic matter is oxidised within the water column prior to sediment deposition and that post-depositional organic matter alteration is much less prevalent. In the case of Core 8, microbial oxidation (mineralisation) of the most easily degraded forms of organic matter would serve to enrich the ^{13}C content of the preserved sediment bulk organic matter. However, during times of reduced microbial activity, such as could be expected during extended dry periods or droughts, less microbial recycling activity in the water column would result in preservation of more organic matter (as evidenced by an increase in the total organic carbon (TOC) content of the sediments) coupled with values which are relatively more ^{13}C depleted than at other core levels with lower TOC amounts.

Figure 3(b) indicates that slightly higher TOC wt% values (1–1.5%) occur in Core 8 during the period AD 1150–1350, corresponding to primarily negative SDR$\delta^{13}C_{OM}$ values (Fig. 3(f)) and enriched $\delta^{18}O_{cellulose}$ values (Fig. 3(e)). As suggested above, dry climate periods that promote more water surface evaporation will likely provide a legacy of anomalously enriched ^{18}O signatures in the sediment cellulose extracted from Core 8. Therefore, the more enriched $\delta^{18}O_{cellulose}$ (1.5–2.0 ‰), particularly between AD 1180 and 1240, suggests increased evaporitic enrichment of the south basin, possibly due to relatively drier conditions. Additionally, marked negative SRD$\delta^{13}C_{OM}$ values centred at AD 1200 could be due to reduced organic matter degradation in the water column in relation to dry period hampered microbial activity and lower phytoplankton productivity, both as a result of reduced terrestrial 'nutrient flushing' during minor terrestrial runoff.

As indicated in Fig. 4, a mid-13th-century dry period may be evidenced by a 10–15% decline in total tree abundance (mostly *Pinus*) accompanied by a 25–30% increase in herbaceous plant pollen abundance in Core 8 sediments centred at AD 1250 (Anderson, 2003). Additionally, total tree pollen abundance increases (25%) accompanied by herbaceous plant decline (25–30%) at AD 1356 could mark the termination of the proposed dry period. Figure 3(c) reveals a steady (up to) 10% decrease in the Core 8 percentage silt (4–63 μm) between AD 1150 and 1260, suggestive of reduced Red River flow rates

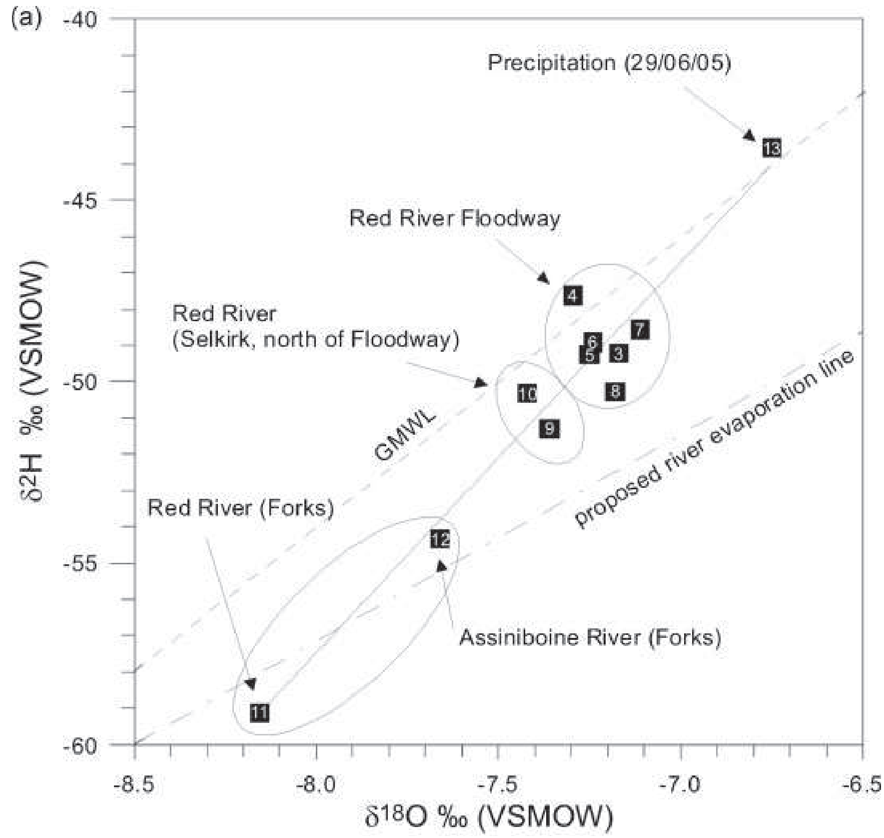

Figure 5 (a) Oxygen *versus* hydrogen isotopic compositions for Red River, Assiniboine River and Red River Floodway water samples collected on 7 July 2005. The Winnipeg precipitation sample was collected in Winnipeg on 29 June 2005 ($\delta^{18}O = -6.8$, $\delta^{2}H = -43.6$). The global meteoric water line (GMWL) is represented by the dashed line. A proposed average evaporation line for river samples 11 and 12 (located right of the GMWL) is indicated by the dash-dash-dot line. The proposed isotopic mixing line between the precipitation, river and floodway samples is indicated by the solid line. (b) Sample locations of Red River (9, 10, 11), Assiniboine River (12) and Red River Floodway (3, 4, 5, 6, 7, 8) 7 July 2005 water samples

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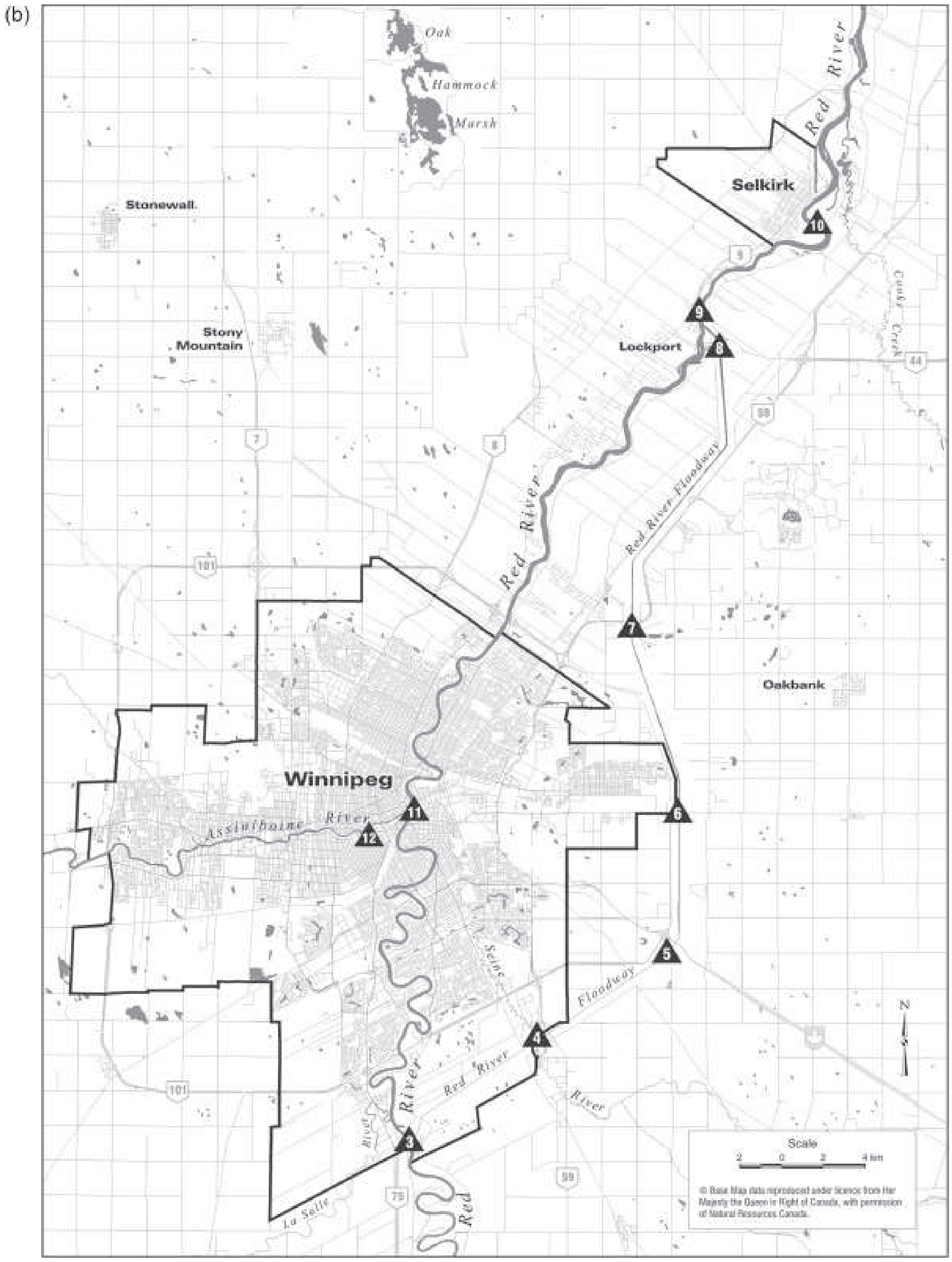

Figure 5 (Continued)

during this period. Additionally, a significant increase in charred remains in Lake Winnipeg 99–900 Core 9 (within 1 m of Core 8), suggestive of an increase in forest fires accompanying a drier climate, between Core 9 depths 125 and 140 cm corresponds to the adjacent Core 8 period AD 1080–1200 or AD 1020–1050 (Telka, 2003) (Fig. 4). Diatom-based salinity reconstructions from Nora Lake in southwestern Manitoba (225 km southwest of the south basin of Lake Winnipeg) signifying changes in effective moisture (evaporation to precipitation; *E*/*P*) place evidence of a large drought between AD 1150 and 1250 (Laird *et al.*, 2003) (Fig. 4). Similar decadal resolution lake water salinities, compared for three lakes in North Dakota (Moon, Coldwater and Rice Lakes; approximately 460, 520, 470 km southwest of Lake Winnipeg's south basin, respectively) suggest that drier than average conditions prevailed in this region of the Great Plains from AD 1040 to 1090; AD 1160–1190; and 1280–1340 (Fritz *et al.*, 2000). Dean (2002) recorded a distinct percentage increase in associated detrital variables (particularly Al %) in Elk Lake, Minnesota (375 km southeast of Lake Winnipeg's south basin) for the period AD 1150–1350 implying drier, windier climate conditions during this period (Fig. 4). Additionally, diatom-inferred total phosphorous (DI-TP) from Elk Lake, Minnesota suggests windier and possibly drier atmospheric conditions promoting water column mixing between AD 1180 and 1280 (Bradbury *et al.*, 2002). Figure 3(d) indicates that notable Core 8 sediment percentage reductions in calcium (Ca) and magnesium (Mg) occur between AD 1180 and 1220, AD 1240–1320 and AD 1340–1390 suggesting reduced terrestrial overland flow (compliant with drier conditions) during these periods (Birks and Edwards, 2004). Note, while the concentrations of several elements in Core 8 do covary with TOC, Ca and Mg do not, suggesting that their reductions are not a result of remobilisation in the sediment column (Snowdon and Simpson, 2003).

Noticeably enriched $\delta^{18}O_{cellulose}$ (1.5–2.0 ‰) intervals are also prevalent between AD 1610 and 1640, 1670–1720 and 1750–1780 (Fig. 3(e)) and are also suggestive of increased

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evaporation within the south basin of Lake Winnipeg due to drier atmospheric conditions. In Fig. 3(f), average to positive SDR$\delta^{13}C_{OM}$ intervals (centred at AD 1630, 1700 and 1770) do arise within the period AD 1610–1780, interspersed with distinct negative SDR$\delta^{13}C_{OM}$ values. These negative SDR$\delta^{13}C_{OM}$ intervals are likely a result of less pre-depositional microbial recycling of the organic matter (as indicated by an average 1.5–2% increase in wt% TOC values between the period AD 1690 and 1780; Fig. 3(b)) and lower lake productivity related to drier than average climate conditions. The Core 8 pollen profiles from Anderson (2003) show marked herbaceous plant abundance increases (20%) between AD 1621–1673 and 1695–1754, accompanied by corresponding declines (5–10%) in total tree abundances (Fig. 4). A 10% reduction in percentage silt (4–63 μm) in Core 8 between AD 1660 and 1730 (Fig. 3(c)) could indicate a period of reduced Red River flow. A record of annual precipitation inferred from a southern Manitoba tree ring chronology suggests that the Red River basin experienced extremely dry conditions between AD 1670 and 1775 (St George and Nielsen, 2002) (Fig. 4). Lower than average sediment percentage Al concentrations from Elk Lake, Minnesota varved sediments are also suggestive of an arid period between AD 1650 and 1750 (Dean, 1997, 2002). A reconstruction of Palmer Drought Severity Indices (PDSI) by Cook *et al*. (1999) suggests that neighbouring American states (including North Dakota and Minnesota) experienced severe drought in AD 1721 and 1743, qualitatively corresponding to the later oxygen isotope-inferred dry intervals (Fig. 4).

In Fig. 4, the Fritz *et al*. (2000) comparison between diatom and ostracod salinity indicators from Moon, Coldwater and Rice lakes indicates above-average salinity levels (indicative of drier conditions) for the periods AD 1475–1505, 1600–1610, 1730–1780 and 1860–1880. Similar diatom and ostracod-inferred salinity reconstructions are suggestive of prevailing arid conditions around Devils Lake, North Dakota ($\sim$210 km southwest from the south basin of Lake Winnipeg) from approximately AD 1500 to 1850 (Fig. 4) (Fritz *et al*., 1994). Figure 3(d) shows low calcium and magnesium concentrations in Core 8 sediments between AD 1620 and 1730, also suggesting reduced terrestrial overland flow during this period (Birks and Edwards, 2004).

Additionally proposed arid periods, as indicated by enriched $\delta^{18}O_{cellulose}$ values, include AD 1320–1340, 1530–1540 and 1570–1580 (Fig. 3(e)). Although the Core 8 record corresponding to the early 14th century shows moderate Ca and Mg percentages indicating relatively normal terrestrial runoff during this time, inferred increases in North Dakota and Minnesota lake water salinities (Fritz *et al*., 2000; Dean, 2002) and Core 8 increases in herb pollen (Anderson, 2003) support a period of aridity between AD 1320–1340 (Fig. 4). St George and Nielsen (2002) indicate that AD 1529, 1581, 1582 and 1587, years that correspond closely with the two later oxygen isotope dry intervals, were among the very driest years inferred from tree rings (Fig. 4). It is possible that another interval of concentrated charred remains shown in Fig. 4 between 80 and 85 cm in Core 9 (corresponding to AD 1488–1525) actually correlates with the AD 1530–1540 oxygen isotope dry interval (Telka, 2003). Negative SDR$\delta^{13}C_{OM}$ values correspond to the AD 1530–1540 and 1570–1580 ^{18}O enrichments (Fig. 3(f)). Again, reduced water column recycling of organic matter as evidenced by higher than average wt% TOC values (Fig. 3(b)) during the majority of the 16th century could be primarily responsible for the AD 1530–1540 and 1570–1580 corresponding negative SDR$\delta^{13}C_{OM}$ values.

Finally, the historical AD 1865 (Rannie, 2001) and instrumentally recorded AD1930s southern Manitoba dry periods correspond nicely to enriched $\delta^{18}O_{cellulose}$ and negative SDR$\delta^{13}C_{OM}$ intervals between AD 1850–1870 and AD 1930–1940 (Fig. 4). Note the higher wt% TOC and corresponding broad interval of negative SDR$\delta^{13}C_{OM}$ values throughout the early 20th century (peaking in the 1930s) are a function of both the enhanced preservation of sediment organic material as a result of the drier 1930s in addition to the higher sedimentation rate commencing in the late 19th century (see Fig. 2), which would serve to broaden the related negative SDR$\delta^{13}C_{OM}$ peak. At the top of Core 8, the distinct $\delta^{18}O_{cellulose}$ depletion accompanied by a rapid positive SDR$\delta^{13}C_{OM}$ excursion and reduction in wt% TOC is likely related to sedimentation related to the AD 1996–97 Red River floods. ^{18}O-depleted snow melt, entering the south basin of Lake Winnipeg from the swollen Red River during the spring and early summers of AD 1996 and 1997, would manifest as relatively depleted $\delta^{18}O_{cellulose}$ values in these upper sediments (Fig. 3(e)). Additionally, high productivity and microbial oxidation directly related to the nutrient rich Red River water would also serve to enrich the lake water dissolved inorganic carbon isotopic composition (resulting in relatively positive SDR$\delta^{13}C_{OM}$ values; Fig. 3(f)) and subsequently reduce the TOC preserved in these top sediments (Fig. 3(b)).

In Fig. 3(e), the marked post AD 1970 $\delta^{18}O_{cellulose}$ enrichment is likely a result of the warmer conditions (possibly related to more evaporation) that have persisted in southern Manitoba during the late 20th century in addition to the change in the surface hydrology and runoff characteristics imparted by the Red River Floodway diversion channel operating since AD 1968. While Hesch and Burn (2005) show a 30-year constant (no trend) to downward trend in evaporation in southern Manitoba, temperatures have certainly increased during this period. As shown by Hastings *et al*. (2005), the mean annual temperature increase for southern Manitoba for the period AD 1971–2000 was 1.05°C (0.35°C per decade). However, the majority of the temperature increase during this period occurred between December and January (1.0°C per decade), accounting for more than a 3°C rise in winter temperatures. These temperature increases could be responsible for up to 2‰ enrichment in the oxygen isotopic composition of precipitation in southern Manitoba, accounting for some of the 3–4‰ $\delta^{18}O_{cellulose}$ increase shown post AD 1968 in Fig. 3(e) (Dansgaard, 1964; Yurtsever, 1975). Additionally, the implementation of the Red River Floodway diversion channel (AD 1968) may have proportionally increased the amount of enriched summer precipitation (representing the volume majority of southern Manitoba precipitation) entering the Red River system on route to its temporary repository in the south basin of Lake Winnipeg. In Fig. 5(a), the oxygen and hydrogen isotopic composition of water collected along the Red River Floodway (7 July 2005) is clearly enriched ($\delta^{18}O = +1.5$‰; $\delta^{2}H = +8$‰) with respect to Red and Assiniboine River water collected close to the Forks (the intersection of the two rivers) on the same day. This could be a result of the surface runoff associated with the heavy precipitation event that occurred on 29 June 2005 (Fig. 5(a)). Floodway-diverted Red River water south of Winnipeg subsequently mixed with the precipitation event runoff ($\delta^{18}O = -6.8$, $\delta^{2}H = -43.6$) and was consequently enriched (Fig. 5(a)). The isotopic composition of the Red River water collected north of the reconnection of the Red River Floodway to the Red River (near Lockport and Selkirk, Manitoba, see Fig. 5(b)) is isotopically enriched and lies on a mixing line between the Red/Assiniboine River and the Red River Floodway oxygen and hydrogen isotopic compositions (see Fig. 5(a)). In this particular incidence, the enhanced collection of enriched summer precipitation by the Red River Floodway served to increase the isotopic composition of the

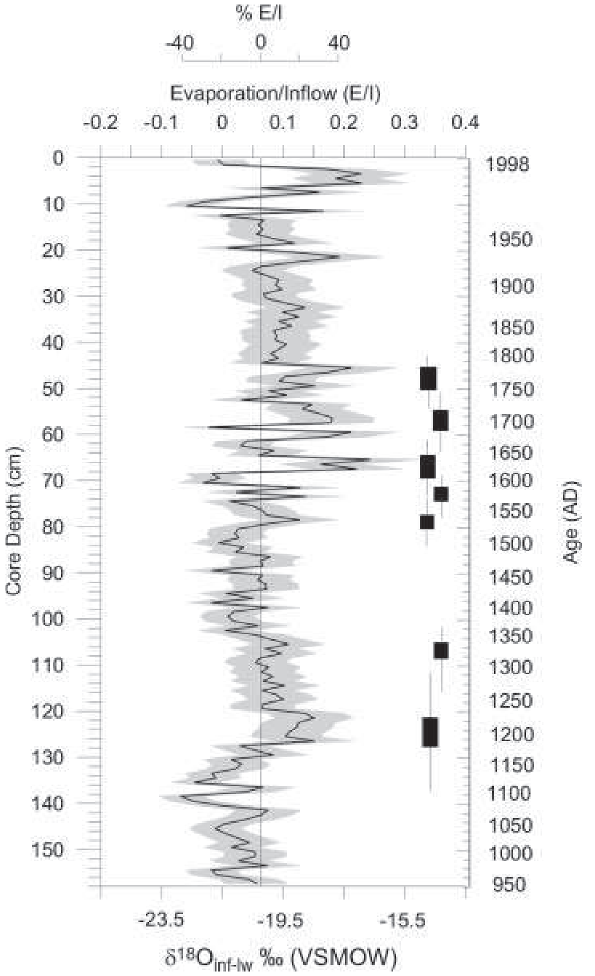

Figure 6 The evaporation to inflow ratio (*E/I*) was calculated using the equation $E/I = [(1 - h + 10^{-3}\ \varepsilon k)/(h - 10^{-3}\ \varepsilon)][(\delta_{inf\text{-}lw} - \delta_I)/(\delta^* - \delta_{inf\text{-}lw})]$, where the atmospheric relative humidity h is considered to range between 0.5 to 0.7 during June to August; the June to August average temperatures for southern Manitoba range between 16.2 and 20.2°C (Environment Canada); $\varepsilon k = 14.2(1 - h)$; $\varepsilon = -7.685 + 6.7123(10^3 T) - 0.6664(10^6 T^2) + (0.3504(10^9 T^3)$ (note: T in K); $\delta_{inf\text{-}lw} = \delta^{18}O_{cellulose} - 27.5‰$; δ_I is assumed to range between −15 and −18‰ and the limiting isotopic enrichment, $\delta^* = (h\delta_a + \varepsilon)/(h - 10^{-3}\ \varepsilon)$, where the atmospheric oxygen isotopic composition of water vapour, $\delta_a = (\delta_I - \varepsilon)$. The dark line represents the median of the calculated *E/I* range (shaded region) assuming the variables discussed above. Percent *E/I* changes were calculated from an average *E/I* value of 0.057. Black blocks with error bars represent the proposed dry intervals based on Core 8 sediment $\delta^{18}O_{cell}$ and SDR$\delta^{13}C_{OM}$ values: AD 1180–1230 (range: AD 1104–1231 to 1160–1280), 1320–1340 (range: AD 1257–1363), 1530–1540 (range: AD 1490–1565 to 1498–1572), 1570–1580 (range: AD 1531–1599 to 1539–1606), 1610–1640 (range: AD 1571–1634 to 1603–1662); 1670–1720 (range: AD 1643–1697 to 1692–1738) and 1750–1780 (range: AD 1724–1766 to 1756–1794). Note that while the elevated $\delta^{18}O_{inf\text{-}lw}$ and *E/I* values corresponding to the period between 1930 to 1940 is probably in response to the very dry conditions existing in the eastern Canadian prairies at this time (1930–40; St George, 2007), the very high values occurring at the top of the core, corresponding to the late 20th century (post 1968), are probably related to the combined influence of both known intervals of dry conditions (1973, 76, 77, 80, 81, 87, 88, 91, 92, 98; St George, 2007) and the oxygen isotopic enrichment of water periodically diverted through the Red River floodway

lower Red River, poised to enter the south basin of Lake Winnipeg, by almost 1‰ and 5‰ in oxygen and hydrogen isotopic composition, respectively. Therefore, the Red River Floodway could also account in part for some of the 3–4‰ $\delta^{18}O_{cellulose}$ increase shown post AD 1968 in Fig. 3(e).

Further, since lake sediment cellulose is consistently enriched in ^{18}O by 27–28‰ compared to lake water (as discussed above), inferring lake water isotopic composition from sediment cellulose oxygen isotope values is routine. Subsequently, inferring a Lake Winnipeg $\delta^{18}O$ water history from the $\delta^{18}O_{cellulose}$ values can provide a watershed hydrology record for southern Manitoba (Fig. 6). Variations in southern Manitoba water balance is best quantified in terms of an evaporation/inflow ratio (*E/I*) through an isotope–mass balance consideration of the inferred lake water oxygen isotope ($\delta^{18}O_{inf\text{-}lw}$) record (Edwards *et al.*, 2004). As expected, Fig. 6 shows distinct increases in the *E/I* ratios associated with the major drought periods identified above. Specifically, the AD 1180–1230 drought interval is associated with a 25–30% increase in the *E/I* ratio with respect to the millennial average, while the AD 1610–1640, 1670–1720 and 1750–1780 drought intervals suggest a 35–40% increase in the *E/I* ratio.

Conclusions

In the case of the south basin of Lake Winnipeg, significantly enriched $\delta^{18}O_{celluose}$ values, corroborated by other hydroclimate indicators that can be readily linked to drier environmental conditions, suggest that the time intervals corresponding to the Core 8 model ages AD 1180–1230 (range: AD 1104–1231 to 1160–1280), 1610–1640 (range: AD 1571–1634 to 1603–1662); 1670–1720 (range: AD 1643–1697 to 1692–1738) and 1750–1780 (range: AD 1724–1766 to 1756–1794) represent major dry periods in southern Manitoba, Canada (Fig. 6).

Assuming reasonable atmospheric and isotopic parameters, the average ratio of lake surface evaporation to catchment input into the south basin of Lake Winnipeg (*E/I*) were significantly higher (between 25% and 40%) during the major dry periods designated here, perhaps qualifying them as major periods of drought in southern Manitoba (Fig. 6). Correlated with most of these oxygen isotope-designated major dry periods are intervals of relatively depleted carbon isotopes that most likely relate to reduced microbial oxidation of washed-in terrestrial organic matter (as evidenced by reduced sediment TOC) during less productive times. It also seems reasonable to assume that a reduction in the flushed input of terrestrial nutrients, accompanying drier regional conditions, could also partially account for the negative SDR$\delta^{13}C_{OM}$ intervals. Based on similar isotope and corroborating multi-proxy evidence other, less severe and enduring, dry periods may have occurred during the periods AD 1320–1340 (range: AD 1257–1363), 1530–1540 (range: AD 1490–1565 to 1498–1572) and 1570–1580 (range: AD 1531–1599 to 1539–1606)(Fig. 6). Additionally, precipitation and water samples collected from the Red and Assiniboine Rivers and the Red River Floodway suggest that the floodway has served to increase the input of isotopically enriched summer precipitation into the south basin of Lake Winnipeg post AD 1968. Overall, the information provided here helps to augment a number of other previous accounts of anomalously dry periods in the eastern Prairies or Great Plains during the last millennium. It is desirable that the now better-constrained dry periods in this central region of North America will assist various agencies and institutions in the formulation of helpful adaptation strategies to prepare for the predicted future hydroclimate changes. Finally, while the use of oxygen and carbon isotopes from sediment organics is likely to remain controversial in light of the problems associated with the

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possibility of mixed organic sources and cellulose isolation, the success demonstrated in this study suggests that its utility should neither be unconsciously accepted or ubiquitously rejected for all lakes. It seems prudent that the environmental reconstruction value of using sediment organics should always be evaluated carefully when needed.

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