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A 2200-year record of hydrologic variability from Foy Lake, Montana, USA, inferred from diatom and geochemical data

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Abstract

A 2200-yr long, high-resolution (~5 yr) record of drought variability in northwest Montana is inferred from diatoms and $\delta^{18}\text{O}$ values of bio-induced carbonate preserved in a varved lacustrine core from Foy Lake. A previously developed model of the diatom response to lake-level fluctuations is used to constrain estimates of paleolake levels derived from the diatom data. High-frequency (decadal) fluctuations in the de-trended $\delta^{18}\text{O}$ record mirror variations in wet/dry cycles inferred from Banff tree-rings, demonstrating the sensitivity of the oxygen-isotope values to changes in regional moisture balance. Low frequency (multi-centennial) isotopic changes may be associated with shifts in the seasonal distribution of precipitation. From 200 B.C. to A.D. 800, both diatom and isotope records indicate that climate was dry and lake level low, with poor diatom preservation and high organic carbon:nitrogen ratios. Subsequently, lake level rose slightly, although the climate was drier and more stable than modern conditions. At A.D. 1200, lake level increased to approximately 6 m below present elevation, after which the lake fluctuated between this elevation and full stage, with particularly cool and/or wetter conditions after 1700. The hydrologic balance of the lake shifted abruptly at 1894 because of the establishment of a lumber mill at the lake's outlet. Spectral analysis of the $\delta^{18}\text{O}$ data indicates that severe droughts occurred with multi-decadal (50 to 70 yr) frequency.

Keywords: varves, $\delta^{18}\text{O}$, diatoms, carbon:nitrogen, drought, PDO, Montana

Introduction

Water availability is crucial for the stability of any society, necessitating careful planning to offset the effects of drought. Such mitigation may be refined with improved understanding of climatic controls and the potential recurrence of severe drought. In this paper, we define drought as a marked and relatively short (~2–200 yr) decrease from long-term average values of effective moisture (precipitation minus evapotranspiration). We differentiate drought events from overall aridity, which is a measure of baseline climate over hundreds to thousands of years.

In North America, one of the most important areas for drought research is the Northern Rocky Mountains, because of the expected sensitivity of western moisture balance to the effects of global warming (Chagnon *et al.*, 1991). Recent model-

ing efforts suggest that “even with a conservative climate model *current* demands on water resources in many parts of the West will not be met under plausible future climate conditions...” (Barnett *et al.*, 2004). Furthermore, the Rocky Mountains are an important zone of recharge for major rivers (e.g., Columbia, Missouri) and aquifers (e.g., Northern Great Plains and Columbia Plateau) (Downey and Dinwiddie, 1988), making an understanding of drought in this region critical to water management in the Northwest and Midwest US. We present here a detailed 2200-yr reconstruction of hydrologic balance from Foy Lake in northwestern Montana (Figure 1a) that is based on diatom and geochemical proxies as a means to understand drought severity and periodicity in this region. Although an optimal reconstruction would include seasonal estimates of drought, none of the proxy records proved suitable for parsing shifts in effective moisture into summer and winter sources.

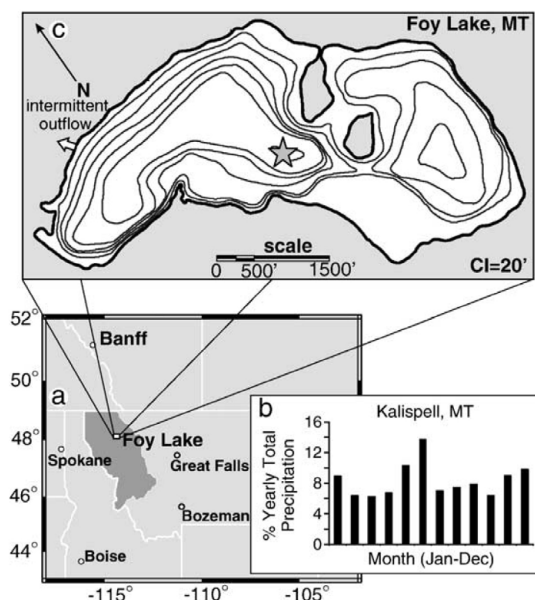


Figure 1. (a) Regional map showing the location of Foy Lake, Montana (MT), and Banff, Canada. Dark shading represents Climatic Division 1 for Montana; (b) monthly percentages of long-term annual precipitation at the Kalispell, MT, climate station; (c) bathymetric map of Foy Lake. A star marks the coring location.

Climatic setting

The principle moisture source for northwest Montana is the Pacific Ocean, with precipitation following westerly storm tracks (Bryson and Hare, 1974). A sizable fraction of winter climate variability in this region has been related to large-scale synoptic features, specifically El Niño-Southern Oscillation (ENSO), Pacific-North America (PNA) pattern, and the Pacific Decadal Oscillation (PDO). Both regional temperature and precipitation are correlated with ENSO, with strong El Niños producing warmer/drier winters and strong La Niñas producing cooler/wetter winters, with an average lag of 4 months (Redmond and Koch, 1991). A positive PNA pattern corresponds to a stronger than average Aleutian low and a reinforced ridge over western North America, resulting in drier conditions from October through March. Alternations between warm (positive) and cool (negative) phases of the PDO correspond with drier/wetter winters, respectively (Mantua *et al.*, 1997). These indices are linked primarily to climatic conditions from October through March and therefore are associated with winter precipitation (especially snow-pack), which is the primary source of stream and ground water recharge in much of western North America. However, local topographic variability superimposed upon broad-scale controls results in a spatially heterogeneous pattern in the seasonal distribution of precipitation in the mountainous West (Mock, 1996).

Meteorological data from Kalispell, Montana (3 km northeast of Foy Lake) suggest a local climatology somewhat different than the general trend described above. Based on a 97-yr series, annual average precipitation is 381 mm/yr and average annual temperature is 6.1°C. Over the last century, less than half (46%) of the precipitation has occurred during the win-

ter months (Oct–Mar). Typically the two wettest months are May and June (Figure 1c). The prominence of late-spring precipitation could indicate a detachment of the local hydrologic framework from the described winter-oriented synoptic atmospheric patterns. On the other hand, the amount of precipitation in late spring may be dependent on the instability of the atmosphere caused by soil moisture that accrued during the preceding winter months.

Site description

Foy Lake (N 48° 10', W 114° 21', 1006 masl) is a deep freshwater lake, consisting of two basins with a combined surface area of 1.1 km² (Figure 1c). The lake is currently mesotrophic with a summer secchi disk depth of 7 m. The lake is located at the eastern edge of the Salish Mountains, approximately 120 m above the Kalispell valley floor. The Salish Mountains are composed primarily of argillites, metacarbonates, and quartzites from the Proterozoic-age Belt Supergroup (Harrison *et al.*, 1992) and were heavily glaciated by the Flathead sub-lobe of the Cordilleran ice sheet (Alden, 1953).

A shallow outlet on the northern end of the lake operates intermittently during spring and is believed to have been modified in the late 1800s to power a lumber mill. The precise influx of ground water to the lake is not known, although residents have described numerous springs along the southeast shore. Although a comprehensive study of surface and ground water chemistry is lacking, scattered wells in the valley and along the foothills indicate that ground water is Ca–Mg–HCO₃ dominated, with relatively low conductivity and pH ranging spatially from 7.3 to 8.3 (LaFave, 2000). Water from a shallow well, up-gradient from Foy Lake has a chemical composition similar to the regional pattern and is saturated with respect to calcite and dolomite and only slightly undersaturated with respect to opaline silica and aragonite (<http://mbmggwic.mtech.edu>).

Methods

In February 2000, we cored the deepest basin of Foy L. ($z_{\max} = 42$ m) with a Livingstone piston-corer, retrieving a mid- to late-Holocene sequence, 6 m in length, and a freeze-core of the upper 57 cm of sediment. The piston core was split lengthwise and photographed in 10-cm intervals. Sediment samples, macrofossil locations, and lamination counts were recorded on enlarged photographs from each interval. Contiguous samples, 2 to 5 mm in length, were collected along distinct bedding planes. Laminations were counted visually in each sample three times prior to collection. Samples were homogenized before being sub-sampled for diatom and isotopic analyses. The remaining sediment was freeze-dried and used for geochemical analyses.

Diatom samples were prepared following the techniques outlined in Battarbee (1986). A minimum of 300 valves was identified and counted at each level. Interpretation of planktic to benthic diatom ratios is based on a model discussed in detail in Stone and Fritz (2004). Oxygen and carbon-iso-

topes were measured at 70°C on a Finnigan MAT 252 coupled with a Kiel II carbonate device after the sediment had been bleached to remove organic matter and sieved at 63 µm to remove ostracodes. Data are reported as ‰ relative to Vienna PeeDee Belemnite (VPDB). X-ray diffraction (XRD) was performed on 60 samples with a Rigaku Miniflex® using a sealed CuKα tube with a diffracted beam monochromator to reduce β-rays. Total carbon and nitrogen values for every other sample (~10-yr intervals) were measured with a Costech® ECS4101. Total inorganic carbon content was measured with a UIC® coulometer. Contiguous samples from the upper 53 cm of freeze core were analyzed for ²¹⁰Pb following the technique outlined in Eakins and Morrison (1978). Power and evolutive spectral analyses of de-trended oxygen isotopic data were run with a red noise assumption. Evolutive spectral analyses were run with a 200-yr moving window with a 2-data-point offset (Mann and Lees, 1996 and Mann and Park, 1999).

Results

Stratigraphy and age model

The upper 1.7 m of sediment consists of red and green gyttja with fine laminations composed of carbonate/organic couplets (Figure 2a). Packets of very distinct white laminations occur from 0 to 1.6 m. Below 1.6 m, the sediment is banded, with grey non-laminated intervals, 1–3 cm thick, becoming increasingly frequent down-core. Laminations disappear completely below 2.5 m. A carbonate-rich interval (pink zone), approximately 13 cm thick, occurs between 18 and 31 cm. Based on rip-clasts and mottling preserved in this section of

the freeze-core, we have designated three cm as slump and excluded this sediment from our analyses and age model.

A chronology was constructed by counting the laminations with the aid of magnification. In sections where laminations were indistinct or absent, the number of years assigned to a sample was calculated from the averaged sedimentation rate of the samples immediately above and below it. Several lines of evidence suggest that the laminations are varves, although direct confirmation with ²¹⁰Pb dating was problematic due to the “pink zone.” An abrupt drop in unsupported ²¹⁰Pb activity at this boundary suggests either a hiatus in sedimentation or dilution by a rapid influx of carbonate. The latter is more likely given (1) the depth of the lake basin, (2) a modern age on a radiocarbon date at 31.2 cm (CURL-5141), and (3) the evidence of slumping. When the ²¹⁰Pb data are modeled to account for the rapid sedimentation, a comparison of the ²¹⁰Pb and varve chronologies has a correlation coefficient of 0.99. The coincidence of a discrete, large charcoal peak in the sample dated A.D. 1910–1914 with the only large local fire during the last 200 years in A.D. 1910 (M. Power, personal communication) provides further support for the chronology.

Although there is no a priori reason to believe that the laminations do not continue to be annual, radiocarbon dates were used to verify varve counts. Wood collected at 87.3 cm yields a date of 925 ± 35 ¹⁴C yr B.P. (CURL-5142) or A.D. 1106 when calibrated (Stuiver and Reimer, 1993) (Figure 2a). The median varve age for that sample is A.D. 1103, which is within the analytical error of the ¹⁴C date. In addition, we can assign a basal date for the varved sequence from a ¹⁴C age model covering the entire Holocene sequence (not reported in this paper). The resultant model is based on ten calibrated ¹⁴C

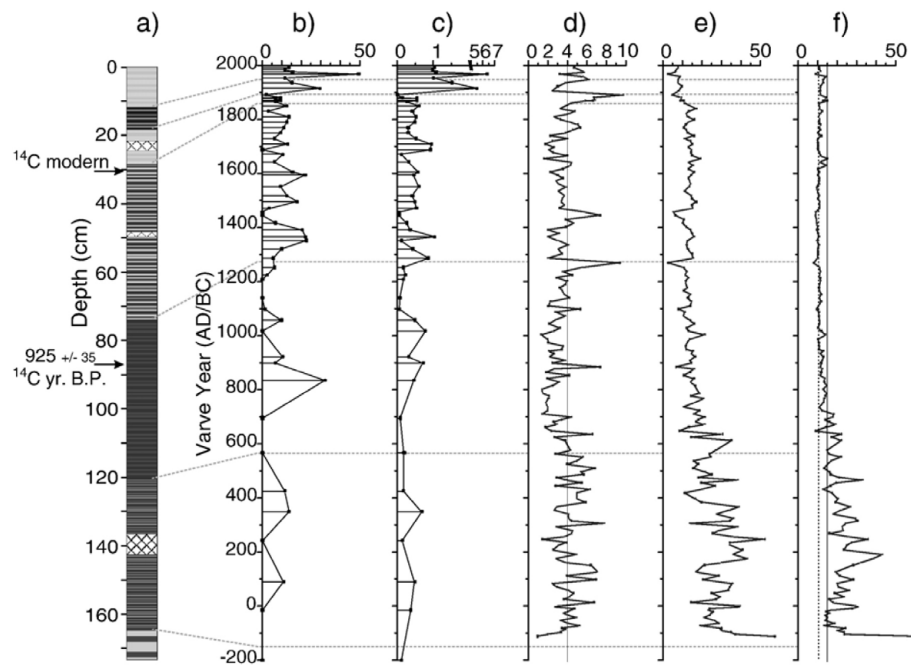


Figure 2. Sediment chemistry and structure for the upper 2200 yr (1.73 m). Solid lines in (d) and (f) indicate the long-term mean values. Dashed line in ‘f’ indicates ratio below which all carbon is of aquatic origin. (a) Sediment stratigraphy (X pattern indicates slumps.); (b) percentage carbonate that is calcite; (c) ratio of quartz peak intensities in XRD to aragonite peak intensities; (d) % TIC; (e) % TOC; (f) OC:N.

dates on macrofossils and two tephtras, Mazama and Glacier Peak I. The assigned calibrated date of 150 B.C. falls within 125 years of the varve age, for an age offset of 6%. We conclude that it is reasonable to accept that the laminations are annual and have plotted all data against the median varve ages for their respective samples.

Lithology

The carbonate mineralogy in many samples is a mixture of calcite and aragonite. In most cases, the amount of calcite estimated from peak intensities of the diffractograms is $\leq 15\%$ (Figure 2b). The average amount of calcite is 10%. In general, greater proportions of calcite occur in the younger part of the record. Given that $\delta^{18}\text{O}$ values of aragonite are 0.6‰ higher than calcite formed in the same water at the same temperature (Tarutani *et al.*, 1969), variations in carbonate mineralogy can affect the $\delta^{18}\text{O}$ record. Although several calcite-rich samples, such as the one at A.D. 1967, coincide with exceptionally low $\delta^{18}\text{O}$ values, no systematic correlation exists between $\delta^{18}\text{O}$ values and % calcite. Variations in the $\delta^{18}\text{O}$ profile caused solely by differing ratios of calcite and aragonite are insufficient ($<0.3\%$) to change the overall isotopic profile, and the record is, therefore, not corrected. The only other major mineral is quartz, which is significant especially in the post-settlement phase (A.D. 1890 to present) (Figure 2c).

Carbon/nitrogen

Values for total inorganic carbon (TIC), total organic carbon (TOC), and the ratio of organic carbon to nitrogen (OC:N) are shown in Figures 2d–f. TIC ranges from 1 to 10%, with an average of 4%. In general TIC values are relatively high in the interval from 200 B.C. to A.D. 600, then drop below av-

erage from A.D. 600 to 1700, with two peaks at \sim A.D. 1260 and 1450. TIC values begin to increase at A.D. 1700 with high values sustained during the latter part of the 1800s, concomitant with the ‘pink zone’. The low values in the 20th century span the first 40 years.

The TOC and OC:N curves are very similar with high values from A.D. 0 to 700 and low relatively invariant values from A.D. 700 to present. All OC:N data younger than A.D. 720 are below the long-term average of 15.5. The early part of the OC:N record shows a large protracted peak (centuries in duration) beginning at 50 B.C. and centered on A.D. 190. A shorter, 75-yr peak is centered on 200 B.C. The high OC:N values near the base of the sequence are consistent with the change in lithology to frequent packets of non-laminated sediment.

Diatoms

Fossil diatom assemblages throughout most of the last 1400 years show a dominance of a single planktic taxon (*Cyclotella bodanica* var. aff. *lemanica*). Benthic diatom taxa, including *Craticula halophila*, *Cymbella cymbiformis*, *Anomoeoneis costata*, *Encyonopsis descripta*, *Encyonopsis angustata*, *Amphora libyca*, and *Navicula* cf. *oblonga*, increase dramatically in relative abundance in several discrete multi-decadal intervals (Figure 3). Peaks in the relative abundance of benthic taxa exceed 60% during the intervals A.D. 1925 to 1935, 1660 to 1680, and 1350 to 1380. Several lower amplitude peaks range from 15 to 30%. Between A.D. 800 and 1200, the relative abundance of benthic taxa is distinctly less than periods before and after, with relative abundance ranging from 0 to 10%. From A.D. 1400–1440, 1580–1620, 1700–1750, fossil diatom populations are nearly monotypic ($>95\%$ *C. bodanica* var. aff. *lemanica*) and are represented as ‘planktic highs’

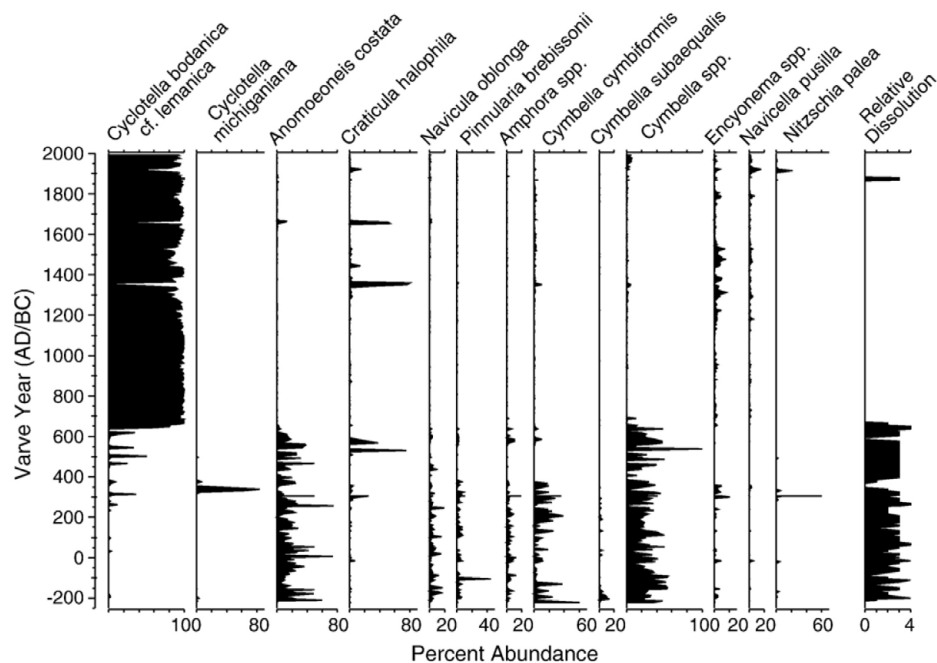


Figure 3. Time-series of select diatom species and a histogram of relative dissolution in diatom samples where higher values indicate increasing dissolution.

on relative abundance plots (Figure 4). The precise abundance of benthic diatoms prior to A.D. 600 is unknown due to dissolution. However, several levels with pristine preservation clearly suggest benthic dominance in this older interval.

Oxygen and carbon isotope records

The $\delta^{18}\text{O}$ record has a maximum range of $\sim 6\%$, although most of the data vary between -2.5 and -6.5% (Figure 4). The record can be separated into pre- and post-settlement phases. Pre-settlement (200 B.C. to A.D. 1890) values have an average of -4.4% . Values drop to an average of -6.2% during post-settlement (A.D. 1890 to present). High-frequency (decadal) fluctuations dominate the record, with longer-term variations becoming more apparent near the base. In the earliest part of the record (A.D. 85–160), $\delta^{18}\text{O}$ values are nearly as low as the post-settlement phase. Beginning at A.D. 150, century-scale fluctuations are superimposed on an overall 4% increase that peaks at A.D. 550. From A.D. 700 to 860, $\delta^{18}\text{O}$ values are consistently lower than the long-term average. Although a broad, low-amplitude peak is centered at A.D. 1140, the high frequency variability that characterizes the earlier part of the record is greatly diminished between A.D. 880 and 1250. A trend toward generally higher values begins at A.D. 1350 and ends at A.D. 1620 with a pronounced 1% decrease. A period with lower than average $\delta^{18}\text{O}$ values occurs in the 1700s and is followed by a several decades of high, uniform values in the 1800s. An abrupt drop in $\delta^{18}\text{O}$ signals the post-settlement phase of the lake beginning at A.D. 1890. In

addition to the high frequency variability of the record, several abrupt and transient decreases in $\delta^{18}\text{O}$ occur at A.D. 510, 830, 1280, 1588, 1740, 1896, and 1967.

The difference between pre- and post-settlement is less pronounced in the $\delta^{13}\text{C}$ record. As with the $\delta^{18}\text{O}$ values, $\delta^{13}\text{C}$ values are nearly as low in the interval A.D. 75–15 as at present. In contrast to the more gradual increase of the $\delta^{18}\text{O}$ values, the carbon isotopes increase abruptly by $\sim 4\%$ at A.D. 150 before gradually decreasing to steady, near-average values from A.D. 600 to 1180. $\delta^{13}\text{C}$ values are generally lower than average in the period 1180 to 1790, with the exception of the interval 1550 to 1620. Values in the 1800s are uniform.

Discussion

Hydrology and chemistry of the lake

How well the modern hydrology and water chemistry of the lake reflect prehistoric conditions is uncertain. The abrupt drop in $\delta^{18}\text{O}$ values at the end of the 19th century suggests a decrease in the residence time of the water that may be a consequence of the construction of a lumber mill at the outlet around 1894.

The dominant carbonate phase in the lake sediments is aragonite. Because incoming ground water has a Mg/Ca molar ratio of 0.9, significant removal of Ca (perhaps by macrophytes in the littoral zone) must occur in order for the aragonite to precipitate (Müller *et al.*, 1972). The historical dominance of aragonite suggests that the residence time of water has always been sufficiently long to achieve this Ca removal, even though the modern lake is seasonally open. The mixture of calcite with the aragonite can be produced by two possible mechanisms, dependent on whether the calcite is detrital or authigenic. If the calcite is detrital, higher values likely correspond with greater surface run-off and wetter conditions in the catchment. If the calcite is authigenic, it most likely precipitates during years with wet springs, which cause the lake to overflow, thereby reducing the residence time of both water and dissolved solids. The latter scenario is supported by the absence of a systematic correlation between $\delta^{13}\text{C}$ values and % calcite, which might be expected if the $\delta^{13}\text{C}$ profile resulted from the mixing of two discrete (one detrital, one lacustrine) carbon pools. In either case, the occurrence of calcite points to a more positive hydrologic balance of the lake.

The modern pH range in the lake (9–10) is higher than the ground water and places the lake near a threshold for silica dissolution. Dissolution of diatoms prior to A.D. 650 and from A.D. 1870 to 1890 (concurrent with the “pink zone”) is attributed to high pH, associated with enhanced rates of primary productivity. Near saturation of ground water with respect to opal suggests that low silica concentrations are not the cause of dissolution.

Lake-level variations and droughts

Estimates of lake levels are based on a site-specific model of Stone and Fritz (2004), which compares the benthic/planktic diatom ratio to potential changes in benthic habitat based

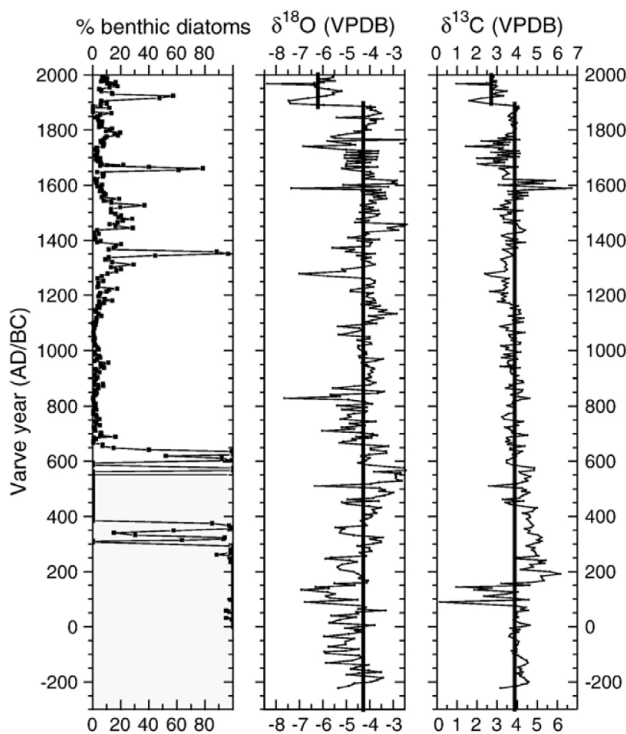


Figure 4. Time-series of (a) percent benthic diatoms; (b) $\delta^{18}\text{O}$ values; and (c) $\delta^{13}\text{C}$ values. Gray box on diatom plot indicates zone of dissolution discussed in text. Thick black lines on the isotope plots show average values for pre- and post-settlement periods.

on basin morphometry. The non-linear relationship between lake depth and the extent of benthic habitat within the photic zone indicates that increases in the abundance of benthic diatoms do not necessarily imply dropping lake levels (e.g., severe drought conditions). According to the model (Figure 5), a 3- to 6-m drop from the present water elevation results in a nearly 4-fold reduction of planktic diatom habitat—a zone referred to as the ‘benthic threshold zone’. At lake elevations below the benthic threshold zone, the two sub-basins become isolated from each other, and the relative proportion of planktic habitat increases abruptly.

Although many factors, such as temperature, hydrology, and timing of carbonate precipitation (cf. Shanley *et al.*, 1998) can influence the $\delta^{18}\text{O}$ values of lacustrine carbonate, we argue that the $\delta^{18}\text{O}$ record from Foy mainly tracks changes in effective moisture (and hence droughts). This assertion is based on the positive correlations between the $\delta^{18}\text{O}$ record and the Palmer Hydrologic Drought Index (PHDI) (Figure 6a) and between the de-trended $\delta^{18}\text{O}$ record and the 700-yr record of estimated precipitation derived from Banff tree-ring data (Watson and Luckman, 2004) (Figure 6b). (We selected the Banff record for comparison because it is significantly longer than more proximal records.) Although the overall patterns in the isotope and tree-ring data are similar, the magnitudes of wet/dry fluctuations in the two records are not equivalent. A notable example of this is a protracted wet interval in the early 1500s recorded by the tree-rings that is not evident in the isotope record. Some of the discrepancies may reflect local

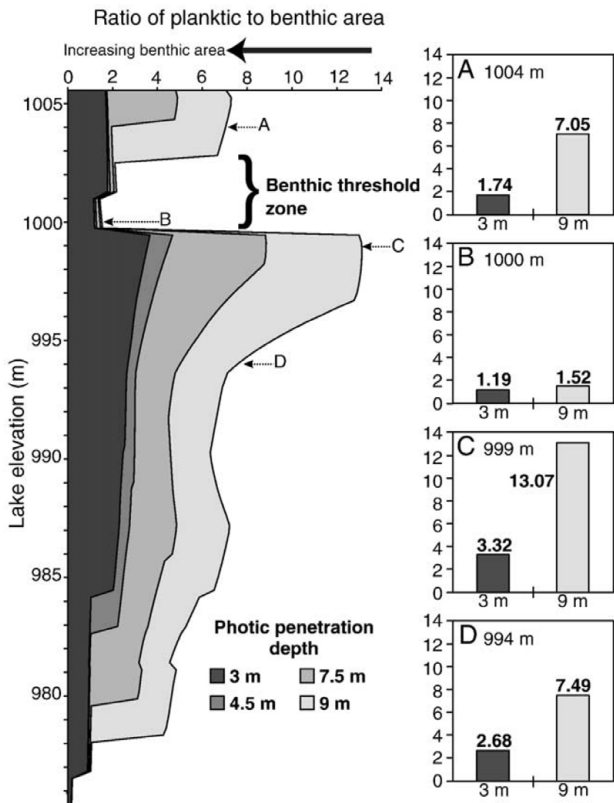


Figure 5. Model showing changes in planktic to benthic habitat ratios with lake-level change from Stone and Fritz (2004). Boxes on the right show the effects of different photic depths on the ratios at specific lake elevations.

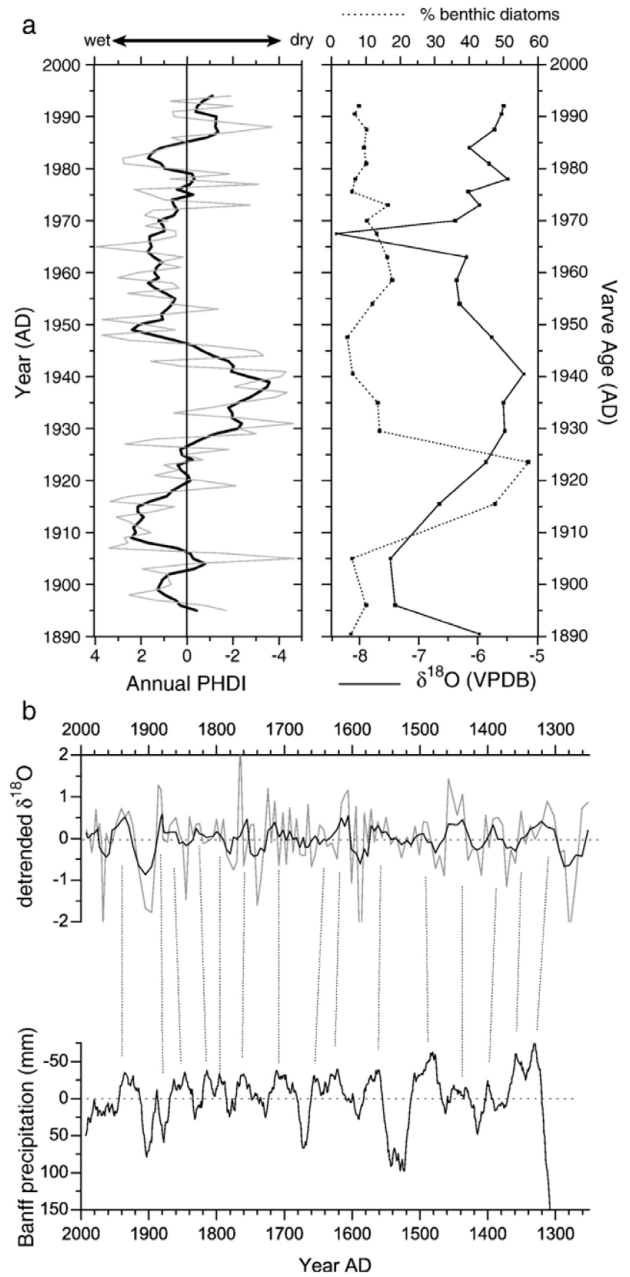


Figure 6. (a) Comparison of the annual (gray line) and 5-yr smoothed (black line) PHDI record for Division 1, Montana (left panel) with the $\delta^{18}\text{O}$ record (black line, right panel) and % planktic diatom record (dashed line, right panel) from Foy Lake for the last century. Negative values of the PHDI correspond with drought years. (b) Comparison of the Foy Lake de-trended $\delta^{18}\text{O}$ record from A.D. 1250 to 2000 with the reconstruction of absolute precipitation from Banff tree-rings (Watson and Luckman, 2004). Black line in top figure is a 5-pt running average of the de-trended isotope values. Dashed lines indicate tentative correspondence of droughts recorded in the lake and in the trees.

climate conditions. Minor offsets between the two records are likely due to errors in the varve chronology. The tree-ring chronology is considered more robust because it incorporates hundreds of trees rather than counts on a single core.

Although it would be ideal to identify changes in summertime versus wintertime ‘‘effective moisture,’’ neither the $\delta^{18}\text{O}$ nor diatom data are capable of resolving seasonal shifts. Even though the precipitation of carbonates is induced by seasonal

algal productivity (most commonly in spring/summer and early fall), the $\delta^{18}\text{O}$ signal inherited from the lake water is an integrated value because a hydrologic residence time greater than 1 year creates an “isotopic” memory of previous years. Similarly, diatoms cannot resolve the seasonal timing at which less (or more) precipitation occurred to change lake level prior to their bloom. Thus, our interpretations are constrained to annual changes in effective moisture. To facilitate discussion of the history of the Foy Lake region, we have divided the record into four time intervals, discussed below.

20th century

Analysis of the diatoms and isotopes during the 1930s drought in North America (Dust-Bowl Drought) illustrates the behavior of both proxies during a severe drought and drop in lake level. (Note: the proxy data represent integrated 5-yr samples and thus lack the resolution of the annual PHDI record). In this case, the peak in benthic abundance precedes the timing of maximum drought conditions (Figure 6a)—a predictable product of the morphometry of the lake basin. As drought developed, the lake fell to ~ 2 m below modern level creating a rapid increase in the area for benthic diatoms. In contrast, the $\delta^{18}\text{O}$ value of the lake water increased gradually through the cumulative effects of evaporative concentration. As drought progressed, continued evaporation further increased the $\delta^{18}\text{O}$ values, but once the lake dropped below the -6 m level, the decrease in available benthic habitat caused an abrupt rise in the relative abundance of planktic diatoms. For this reason, the $\delta^{18}\text{O}$ values appear to be decoupled from the diatom record—a feature of the two records that occurs in the past as well. In addition the pronounced shifts in the $\delta^{18}\text{O}$ record at \sim A.D. 1922, 1950, and 1976 roughly follow the regime changes in the PDO at 1923, 1947/1948, and 1976 (cf. Minobe, 1999).

A.D. 1200 to present

We suggest that for most of the last 800 yr, lake level fluctuated around the benthic notch (-3 to -6 m below modern elevation). Three prominent benthic spikes in the diatom record during this interval (Figure 4) are almost certainly related to lake-level fluctuations that maintained the lake at this level (Figure 5), as no other lake depths could produce this particular sedimentary signal (Stone and Fritz, 2004). Annual laminations in this zone preclude slumping as the cause of these ‘spikes’. Furthermore the duration, sharp transition, rapid recovery, and magnitude of the assemblage changes indicate that these are threshold events and not the result of the long-term climate changes that would be required to reduce the lake depth more than 6 m. As the diatom model indicates, these increases in benthic abundance do not necessarily imply dropping lake levels (e.g., severe drought conditions). Antecedent lake levels may have been either above or below the -6 m mark.

The “planktic highs” during intervals A.D. 1400–1440, 1580–1620, and 1700–1750, as in the 1930s, likely represent periods where lake level actually dropped below 6 m and the two sub-basins became isolated. The ‘highs’ are less likely

to result from lake level rising above the “benthic threshold zone” because the benthic values exceed the levels of the past 60 yr in which the lake (at least seasonally) is at maximum level (Stone and Fritz, 2004).

The long-term trend of increasing $\delta^{18}\text{O}$ values from A.D. 1270 to 1620 is tentatively interpreted as an increase in the relative amount of summer moisture, which would increase the annually averaged $\delta^{18}\text{O}$ value of the lake water. The relatively high percent of calcite and lake levels inferred from the diatom data suggest that this was not an exceptionally arid interval, although it had distinct drought events. The most impressive of these occurred around A.D. 1435–1460. The gradual increases in benthic diatoms, with percent abundance values in excess of 20%, likely represent periods where lake level was lowered by moderate drought. The benthic spike at \sim A.D. 1350 may actually represent a modest rise in lake level, as does the spike at \sim A.D. 1650.

The decrease in the $\delta^{18}\text{O}$ values between A.D. 1720 and 1780 suggests a significant cooling and/or greater contribution to the lake from ^{18}O -depleted snowfall. Either of these conditions would have contributed to the expansion of Rocky Mountain alpine glaciers in the 18th and 19th centuries (Luckman, 2000). The isotopic decrease is not likely to have resulted from overall wetter conditions, based on precipitation estimates at Banff (Watson and Luckman, 2001). The low values are generally consistent with negative PDO values reconstructed from tree rings (D’Arrigo *et al.*, 2001) but do not agree with estimates of low snow-pack and summer drought in Glacier National Park (Pederson *et al.*, 2004). Furthermore, the high, uniform $\delta^{18}\text{O}$ values during most of the 1800s are consistent with neither the large shifts in reconstructed PDO (D’Arrigo *et al.*, 2001) nor the high snow-pack and cool/wet summers reconstructed in Glacier (Pederson *et al.*, 2004). Part of the discrepancy may be related to uncertainty in dating around the “pink zone.” Further data are required to address this issue.

A.D. 800–1200

Planktic percentages are greater during this interval than at any other time during the Late Holocene, including the last century when the lake seasonally overflows, suggesting lake levels ~ 7 to 10 m below present. Thus, we argue that in general lake levels prior to A.D. 1200 were lower than during the last 800 yr and that overall climate was dry.

The relatively uniform $\delta^{18}\text{O}$ values during the Medieval Period (A.D. 900–1250) (Lamb, 1965) indicate a quiescent climate and/or decreased sensitivity of the lake to climatic change. Superimposed on this generally dry climate is a severe, protracted drought centered at A.D. 1140 following a wetter period from A.D. 1050 to 1100, indicated by both diatom and isotope data. The interpretation of a dry Medieval period is consistent with other records in the western US (Stine, 1994 and Benson *et al.*, 2002), although these are considerably south of Foy Lake. Despite the general aridity, the region was not subject to the intense, multi-decadal droughts that characterize the preceding millennium.

200 B.C. to A.D. 800

Poor diatom preservation prior to A.D. 600 and the notably high values in OC:N suggest exceptionally low lake levels, likely the lowest of the entire 2200-yr record. High OC:N values (>10) coincident with the dissolution zone suggest the expansion of emergent macrophytes toward the coring location, which requires significantly lower lake levels than previously described. Increasing OC:N ratios by preferential degradation of algal OM during early diagenesis is considered insufficient to alter the original source signatures (Meyers, 1994), and burial may even stabilize OM from further alteration (Meyers and Ishiwatari, 1993). Elevated $\delta^{13}\text{C}$ values between A.D. 200 and 400 coincide with maximum OC:N values, suggesting greater photosynthetic removal of CO_2 from surface waters that is, at least partly, attributable to aquatic vascular plants. Although the maximum in OC:N values at A.D. 190 does not correspond with peak $\delta^{18}\text{O}$ values, higher OC:N values do correlate with a zone of generally higher $\delta^{18}\text{O}$ values.

The $\delta^{18}\text{O}$ record for the first 1000 yr shows multi-decadal to century-scale droughts superimposed on a steady increase in values until ~A.D. 600. The overall trend toward more positive $\delta^{18}\text{O}$ values suggests a fundamental shift in base climate conditions, such as an increase in general aridity, a shift to warmer temperatures, more summer precipitation relative to winter, or some combination of these factors. A general increase in aridity is considered least likely, given that decreasing OC:N values are interpreted as rising lake levels. The multi-decadal increases at ~A.D. 160, 300, 400, 490 and 550 are considered drought events.

The depletion of ^{18}O after A.D. 600 corresponds with the onset of diatom preservation and suggests rising lake levels and lowered lake pH or alkalinity associated with either increased precipitation and/or cooler summers. Regardless, four multi-decadal peaks in $\delta^{18}\text{O}$ indicate the continued occurrence of frequent drought events.

Drought cycles

Despite the current dominance of late-spring precipitation, regional moisture availability is apparently linked to large-scale climate modes. Evolutive spectral analysis identified three intervals with distinctive periodicities and power (Figure 7). The older part of the record (up to ~A.D. 700) has no defined periodicities that exceed the 90% confidence interval (CI). However, significant variability exists in the mid-frequency (50–100 yr) range, whereas there is little variability in the high-frequency (10–20 yr) range. Between A.D. 700 and 1100, the isotope record shows well-defined periodicities at ~33 and 22 years. The 22-yr cycle may be driven by solar activity (e.g., double-Hale cycle) or may be a manifestation of the bidecadal shift in the strength of the Aleutian low described by Minobe (1999). In addition, for a 200-yr interval, there is a ~14-yr cycle (>95% CI) of unknown origin, although it could be related to oscillatory modes identified in the North Atlantic sector (Moron *et al.*, 1998). In the latter part of the Medieval Period (A.D. 1100–1250), periodic changes once again have

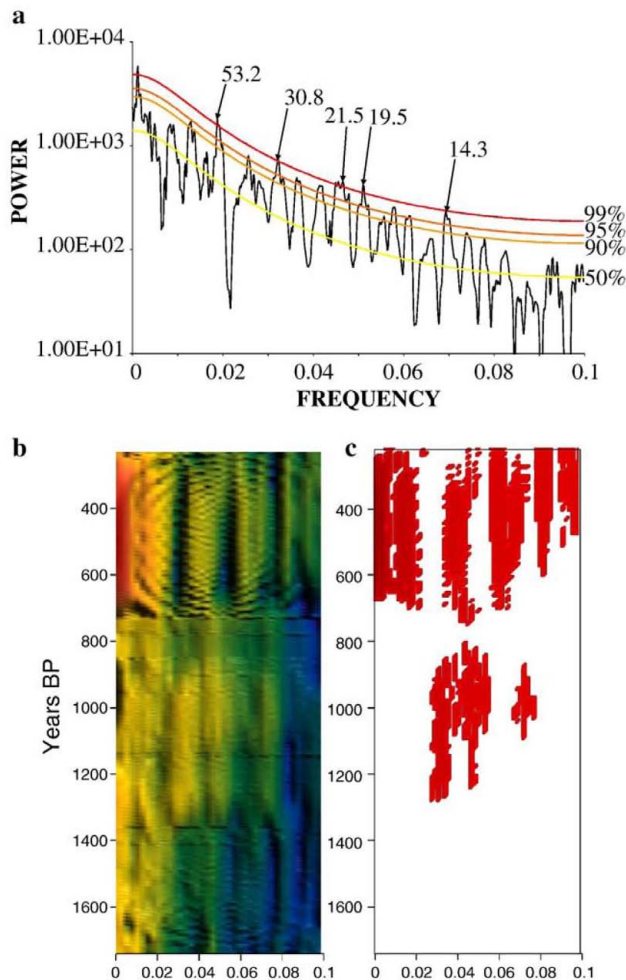


Figure 7. Spectra of the de-trended $\delta^{18}\text{O}$ record for the last 2000 yr. (a) MTM power spectra plotted against a red noise assumption. Red, orange, and gold lines signify the 99, 95, and 90% confidence intervals ($P < 0.01$), respectively; (b) Evolutive spectra employing a 200-yr moving window with a 10-yr step. Frequency plotted against years before 1950. Record is truncated at ~200 and ~1800 yr ago due to window size; (c) 90% confidence plot of the evolutive spectra showing persistent frequencies.

little power. Two possibilities could explain the lack of periodic variability. One, the climate was particularly quiescent and the decadal source of climate variability in the North Pacific was muted or absent. A second possibility is that the lake and its hydrologic framework were insensitive to climate variability at this time. Additional sites, preferably with different hydrologic frameworks, are needed to resolve this question.

At A.D. 1250, variability in the de-trended isotope record becomes pronounced with a nearly 600-yr interval dominated by cyclical changes at 55–75 yr, 22–31 yr, 19-yr, and 12–12.5 yr. The 55–75 yr periodicity is consistent with ‘pentadecadal’ shifts in Pacific sea surface pressures and temperatures (Minobe, 1997, Minobe, 1999, Mantua *et al.*, 1997 and McCabe *et al.*, 2004), which have been identified in tree-ring records of the central and southern Rocky Mountains (Gray *et al.*, 2003) and lake sediments (Benson *et al.*, 2003). The strength of this signal suggests that decadal variability in the North Pacific and

Aleutian low may be the main driver in decadal climate variability over northwest Montana. The 22- to 31-yr cycle may be a blurred expression of two discrete cycles at 33 and 20 yr, which are dominant in the early part of the Medieval period. The origin of the cycle may be the 'bidecadal' variation in the Aleutian low, which, when combined with the first cycle, results in the North Pacific regime changes (Minobe, 1999 and Mantua *et al.*, 1997). The 19-yr cycle may be an harmonic of the longer cycles. The 12-yr cycle has an unknown origin and may represent 'clusters' of ENSO variability. However, the strong correspondence between modern climate and La Niña

years cannot be tested with our present data set, because the sampling interval (~5 yr) is too long. In any case, the period after A.D. 1250, which we delineate as the onset of the Little Ice Age (*sensu* Porter, 1986), stands out as having very well-defined and strong cyclicity in moisture availability.

Conclusions

The combination of diatoms and stable isotopes allows clearer reconstructions of century to millennial trends in aridity. Figure 8 summarizes the relationship among major cli-

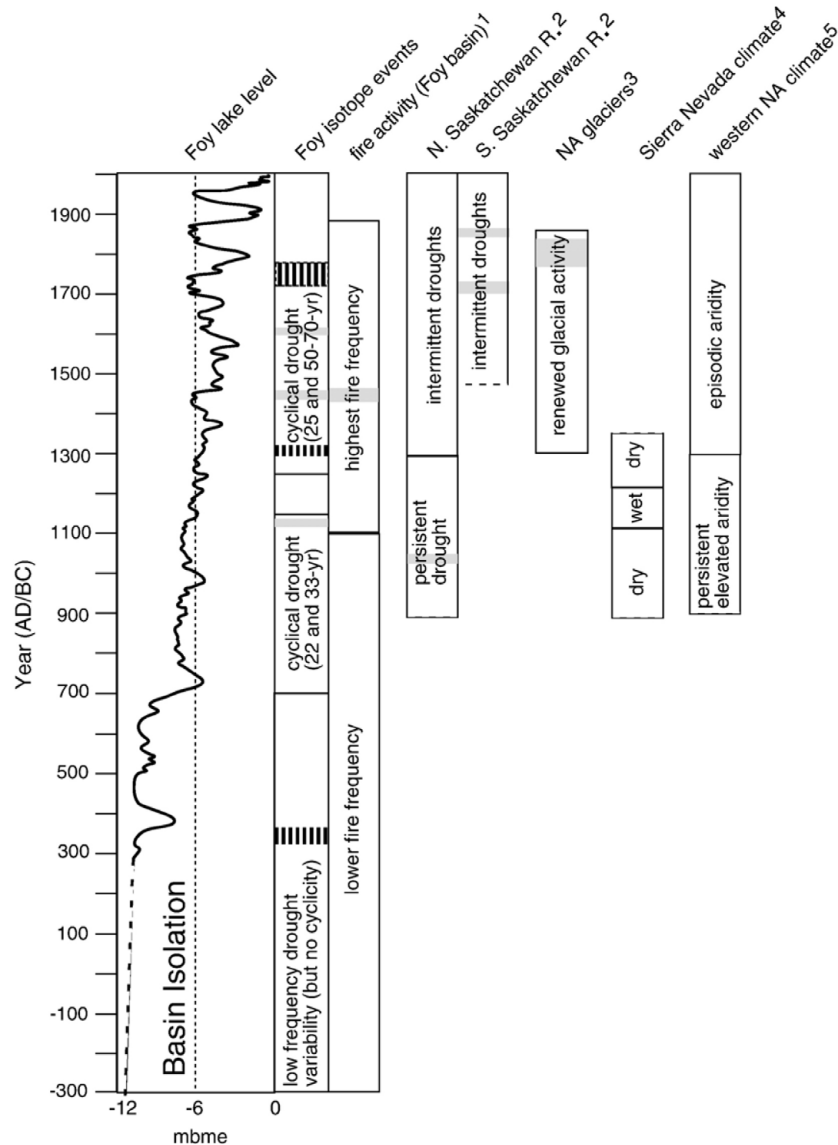


Figure 8. Major climatic events in the Foy L. record and their relationship to regional paleoclimatic records. Dashed edges on the boxes indicate the beginning and/or termination date for each particular record. From left to right: stylized lake-level curve for Foy L. in meters below modern elevation (mbme). Lake-level elevations are not exact (with the exception of modern elevation and the elevation of basin isolation) but are meant to show relative changes inferred from the diatom record. Dashed line indicates unknown levels in diatom dissolution zone. Main boxes of Foy isotope events indicate intervals with defined drought cycles. Gray (black) boxes are significant drought (wet) events in the $\delta^{18}\text{O}$ record defined by peaks of exceptionally high (low) values compared to data on either side of the peak. 1) Fire activity in the Foy drainage basin (from Power *et al.*, in press). Gray box indicates peak fire activity. 2) Drought frequency determined from reconstructed discharge in the North and South Saskatchewan River basins (Case and MacDonald, 2003). Records are 550 and 510 yr long, respectively. Gray boxes indicate the most severe droughts lasting at least 20 yr. 3) North American glacier activity (Porter, 1986). Gray box indicates cool/wet conditions in Glacier Natl. Park inferred from tree rings (Pederson *et al.*, 2004). 4) 360-yr record of hydroclimate for the Sierra Nevada inferred from tree stumps (Stine, 1994). 5) General hydroclimate for the last 1200 yr for all of western NA based on the drought area index (Cook *et al.*, 2004).

matic changes at Foy Lake based on lake-level variations, isotope events, and fire activity (Power *et al.*, in press) and regional estimates of drought and glacier activity. In general, climate was very dry from 200 B.C. to A.D. 600, with the lowest lake-levels of the last 2200 yr. Discrete drought events occurred within the overall dry climate, although these events have no distinct recurrence interval. A slight increase in effective moisture occurred from A.D. 600 to 800, with rising lake levels and fresher lake chemistry permitting diatom preservation. Climate remained drier than present, but more stable, from A.D. 800 to 1250. In the early part of this interval, climate varied at bidecadal and tridecadal time intervals. A.D. 1250 marks the beginning of an entirely different climatic regime. Lake-levels rose within 3 to 6 m of present elevation, and variations in effective moisture occurred with defined frequencies, several of which are consistent with multi-decadal variations in the strength of the Aleutian low and the PDO. An extended oxygen-isotopic reversal in the 1700s offers insight into the climate conditions that led to the advance of Rocky Mountain alpine glaciers in the 1800s.

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