PALEOECOLOGICAL AND ISOTOPIC RECORDS OF CLIMATE CHANGE AND VARIABILITY FROM LAKES AND SPELEOTHEMS, BEAR RIVER RANGE, SOUTHEASTERN IDAHO

by

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STATEMENT OF DISSERTATION APPROVAL

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ABSTRACT

A paleoenvironmental reconstruction is presented from the Bear River Range, southeast Idaho. Reconstructed environmental conditions are inferred from paleoecological proxies, including pollen, macrofossils, and macroscopic charcoal from lacustrine sediments taken from Plan B Pond. A second, complementary paleoenvironmental reconstruction is presented based on stable carbon and oxygen isotope measurements of speleothem calcite collected from Minnetonka Cave, located just 8 km from Plan B Pond. The paleoenvironmental reconstructions from the two Bear River Range records suggest that each record contains significant seasonal bias, with the speleothem and sedimentary charcoal records primarily recording winter-season variability over time, and the pollen and macrofossil data recording summer conditions. Together, the Bear River Range data show a comprehensive picture of Holocene hydroclimatic conditions, and refines our understanding of controls on water resource variability in the region.

Based on the Bear River Range records, the study area experienced cool, wet winters, and hot, dry summers in the early Holocene (prior to ~7500 BP), as compared to modern. The middle Holocene (7500-4000 BP) was characterized by dry and consistently warming winters, with dry, warm summers. The Late Holocene (4500 BP-Modern) has been characterized by cooler summers, with generally wetter, but increasingly variable, winter conditions.

The results of this study indicate that previously published paleoclimate reconstructions from the Western United States must be considered carefully when using them to understand large-scale ocean atmosphere teleconnection patterns such as El-Niño Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO). Because these teleconnection patterns are associated with spatially consistent seasonal precipitation and temperature anomalies, paleoclimate records must be evaluated for seasonal bias, if they are to be used for reconstructing past strength or presence of the teleconnection anomalies.

For Kari, Grayson, and Allan.

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CHAPTER 1

WESTERN HYDROCLIMATOLOGY AND THE NEED FOR LONG-TERM DATA: THE BASIS FOR PALEO-ENVIRONMENTAL INVESTIGATIONS IN THE BEAR RIVER RANGE

Introduction

Recent research into the effects of global climate change on spatiotemporal precipitation patterns and river hydrology in the arid Western United States has shown that the magnitude and seasonality of precipitation and peak runoff in many watersheds, especially those that are snowmelt dominated, have been altered over the last several decades and are likely to continue changing for the foreseeable future (e.g. Barnett et al., 2004; Regonda et al., 2005; Carson, 2007; Bedford and Douglass, 2008). Problems generated by the altered hydrology include conflicts between agricultural irrigation needs and aquatic wildlife habitat conservation, increased wildland fire danger, decreased hydroelectric production, and impacts on the economic health of the winter recreation industry, among others (Barnett et al., 2004; BRAC, 2007).

Given the importance of water resources in the West, and the potential impacts of altered hydroclimatology in the future, developing a better understanding of long-term variability and trends, the mechanisms driving variability, and the environmental effects of past changes in hydroclimate are important research goals. This project was developed with such goals in mind. The project focuses specifically on the Bear River Basin, located at the intersection of Utah, Idaho, and Wyoming. As discussed below, the Bear River Basin is an interesting watershed due to both human and environmental factors, making it an ideal research location.

Study Objectives and Contributions

The primary objective of this study was to reconstruct the long-term record of climate variability in Bear River Basin by analyzing geochemical and paleoecological indicators of hydroclimatic conditions over the last 10,000-15,000 years. The data collected provide a long-term baseline against which historic conditions can be compared to evaluate the significance of recent changes to hydroclimatic variability, especially related to seasonality and magnitude of snowpack. The data also provide information about changes in vegetation composition as a function of climate forcings, and establish a locally relevant fire/vegetation/climate relationship through the Holocene. Finally, due to the unique geologic history of the Bear River Range, this study evaluates the comparability of two different types of terrestrial paleoclimate archives, and utilizes the biases of the different records to better understand the reasons for disparate paleoenvironmental records from around the west.

Study Site

The Bear River Range (BRR) in northern Utah and southeast Idaho is the northernmost extension of the Wasatch Mountains. The range lies in the approximate center of the Bear River drainage basin, and rises to a maximum elevation of approximately 3000m. The mountain range is a primary hydrologic recharge area for the Bear River watershed (Dion 1969; UDWR, 2004). As a primary recharge area, the hydrologic record of the Bear River Range provides a local signal that is applicable over a broad area. Furthermore, the range provides several different climate archives, including glacial lakes and speleothems. Thus, a new comprehensive multi-archive, multiproxy paleoenvironmental investigation was possible within a relatively small area in the Bear River Range. Lake sediment proxies analyzed for the study include pollen, macroscopic charcoal, and plant macrofossils. Climate proxies extracted from the speleothem include stable carbon and oxygen isotopes. The investigation provides cross-calibration and verification of proxy data from each of the archives, allowing for more accurate interpretation of data at a variety of temporal resolutions.

Research Questions

Four primary research questions have been developed for this study.

1. How does the magnitude and frequency of climate variability in the Bear River region during the Late-Glacial and Holocene compare to historic records, especially with respect to precipitation?
    - Hypothesis #1: The historic record is an adequate representation of hydroclimatic variability in the Bear River Basin for long-term water resource planning.

2. How sensitive is vegetation in the Bear River Range to changes in climate, and is there a primary climate variable (temperature vs. precipitation, seasonality of precipitation, etc.) responsible for vegetation shifts?
    - Hypothesis #2: Bulk vegetation composition is highly responsive to changes in climate, with significant sensitivity to multiple climate variables.
3. Do speleothem and lake sediment proxies from essentially the same mid-latitude location record climate change differently from one another? (i.e., Are they directly comparable?)
    - Hypothesis #3: Co-located speleothem and lake sediment based paleoclimate reconstructions do reflect similar paleoenvironmental histories, and can be directly compared with one another.
4. How does the localized climate record of the Bear River Range compare with the Bear Lake climate record generated by the USGS and collaborators? Is the Bear Lake record an accurate indicator of climate in the Bear River Range over the last several thousand years?
    - Hypothesis #4: The records extracted from the Bear River Range do correlate well with the Bear Lake paleoenvironment reconstructions, suggesting Bear Lake does provide an accurate record of local paleoenvironmental conditions.

Study Motivation/ Background

To fully understand the significance of recent observations and predictions of future hydroclimatic change in the West, a long-term baseline is needed to place the new information into perspective. The length of record necessary to provide “long-term” data is debatable; however, the Colorado River Compact exemplifies how historical data alone are often not sufficient. Water allocations in the Colorado River Compact of 1922 were based upon a 20-year instrumental dataset. We now know that the 20-year record represented one of the wettest periods in several centuries (Woodhouse et al., 2006; Meko et al., 2007). Thus, the actual range of variability in the available water resources of the Colorado River was underestimated, and the long-term mean was overestimated. As a result, a new agreement had to be negotiated in 2007 to provide guidelines for allocating Colorado River water when inevitable shortages occur (U.S. Bureau of Reclamation, 2007).

The Bear River, located in the northeastern corner of the Great Basin, flows across state lines in an arid environment where water resources are limited (Figure 1.1), much like the Colorado River. The Bear River follows an approximately 500-mile path, originating on the north flank of the Uinta Mountains in eastern Utah, flowing northward into Wyoming and Idaho before reversing directions and flowing southward back into Utah and ending at Great Salt Lake (Figure 1.1). The river is the largest contributor to Great Salt Lake, supplying slightly more than half of the total riverine input. The Bear River Compact of 1958 and Amended Bear River Compact of 1980 established the framework under which the waters of the Bear River are allocated among the states of Utah, Wyoming, and Idaho.

Investigations of stream hydrology in the Uinta Mountains (Carson, 2007) show that records from the last century display a trend toward having larger relative contributions of spring and summer precipitation to the total annual precipitation record. Additionally, research on snowpack in the Great Salt Lake Basin has shown that peak snow-water equivalent occurs approximately 15 days earlier today than it did 20 years ago (Bedford and Douglass, 2008). These studies indicate that the local region surrounding the Bear River Basin is experiencing similar hydroclimatic changes to those reported in other Western watersheds (e.g., Barnett et al., 2004), but they do not provide extended records of past variability to determine how notable these changes are in a longer context.

Recently, political and economic factors have brought Bear River water resources into the public's eye. The State of Utah has identified the Bear River as one of the last developable water resources in the state, and laid out a multiyear plan for increasing the storage capacity of reservoirs and building new diversions of Bear River water to accommodate growing populations in the Wasatch front region (UDWR, 2001, 2004). Nearly synchronous with the release of Utah's plans for the Bear River Project, the State of Idaho declared the Idaho portion of the Bear River basin as an official ground water management area (GWMA). The State of Idaho defines a GWMA as all or part of a ground water basin that may be approaching critical conditions of not having sufficient ground water to provide a reasonably safe supply for irrigation or other uses at the current or projected rates of withdrawal (IDWR, 2001). Although Idaho's declaration pertains specifically to groundwater, it affects the Bear River surface discharge as well, due to the fact that the Bear River has been classified as an effluent (hydrologically gaining) stream throughout most of the Idaho portion of the river basin (Dion, 1969). This means that over-use of groundwater resources in the Idaho portion of the basin could diminish the in-stream flow of the Bear River as it passes through Idaho and into Utah by artificially drawing down aquifers and preventing groundwater from adding to in stream flows.

In addition to being a potential political and economic point of contention, the Bear River is also located at an interesting geographic boundary in terms of both observed climate and projected climate change. The Bear River watershed represents a location in the latitudinal transition zone between areas to the north that are predicted to get more precipitation as climate warms, and areas to the south that are predicted to get less precipitation (IPCC, 2007).

The Bear River Basin also lies along a major latitudinal transition zone (~ 40-42°N) defined by spatial winter season precipitation anomalies associated with the large-scale ocean-atmosphere climate teleconnection patterns known as the El-Niño Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997; Wise, 2010), with areas to the north (south) of the transition zone typically wetter (drier) than average in La Niña phases of ENSO and/or cool phases of the PDO, and drier (wetter) in El Niño phases of ENSO and/or warm phases of PDO. This contrasting north-south precipitation anomaly pattern is commonly referred to as the Western precipitation dipole (e.g., Wise, 2010). The Bear River Basin's location along these boundaries make it an interesting location for evaluating long-term climatic and hydrologic variability in order to better understand future impacts of a warming climate on a major regional water resource.

Manuscripts

The next three chapters are each written to be independent manuscripts, but as a compilation they present a comprehensive paleoenvironmental record of the Bear River Range that further refines our understanding of the paleoenvironmental history of the Great Basin and Western United States. The three manuscripts presented are: (1) A 14,000-Year Record of Fire and Vegetation from the Bear River Range, Southeast Idaho (2) A Speleothem Record of Holocene Paleoclimate from the Northern Wasatch Mountains, Southeast Idaho, USA (3) Complementary seasonal bias in two paleoclimate records from Southeast Idaho: Refining the regional hydroclimatic history of the Northeast Great Basin.

Manuscript 1, "A 14,000-Year Record of Fire and Vegetation from the Bear River Range, Southeast Idaho", was written with either *The Holocene* or *Quaternary Research* as its intended target journal. The manuscript presents pollen and macrofossil evidence of vegetation changes over time, and a macroscopic charcoal record as and indicator of past fire activity. Mechanisms controlling vegetation and fire regime changes are also discussed.

Manuscript 2, "A Speleothem Record of Holocene Paleoclimate of the Northern Wasatch Mountains, Southeast Idaho, USA", was written with *Quaternary International* as its intended target journal. Specifically, it is intended to be included as part of a proceedings volume from the 25th Pacific Climate Workshop (2011) held in Monterrey, California. The paper presents isotope data from a Minnetonka Cave speleothem, discusses the systematics thought to be controlling the isotope data, and discusses the climatic states responsible for precipitation variability at Minnetonka Cave.

Manuscript 3, "Complementary Seasonal Bias in Two Paleoclimate Records from Southeast Idaho: Refining the Regional Hydroclimatic History of the Northeast Great Basin", was written with *Quaternary Science Reviews* as its intended target journal The manuscript discusses the need for understanding seasonal bias in paleoenvironmental records, and presents a season-specific paleoclimatic history from the Bear River Range. Then, by correlating different season-specific proxies with other paleclimate records from around the region the paper addresses likely seasonal bias in other records to provide a clearer understanding of changing spatiotemporal climate patterns across the Great Basin and other nearby locations.

The final chapter of this dissertation summarizes the findings of the research project, presents implications and significance of findings in a larger context, and discusses some of the outstanding questions and future research that might be developed from this work.

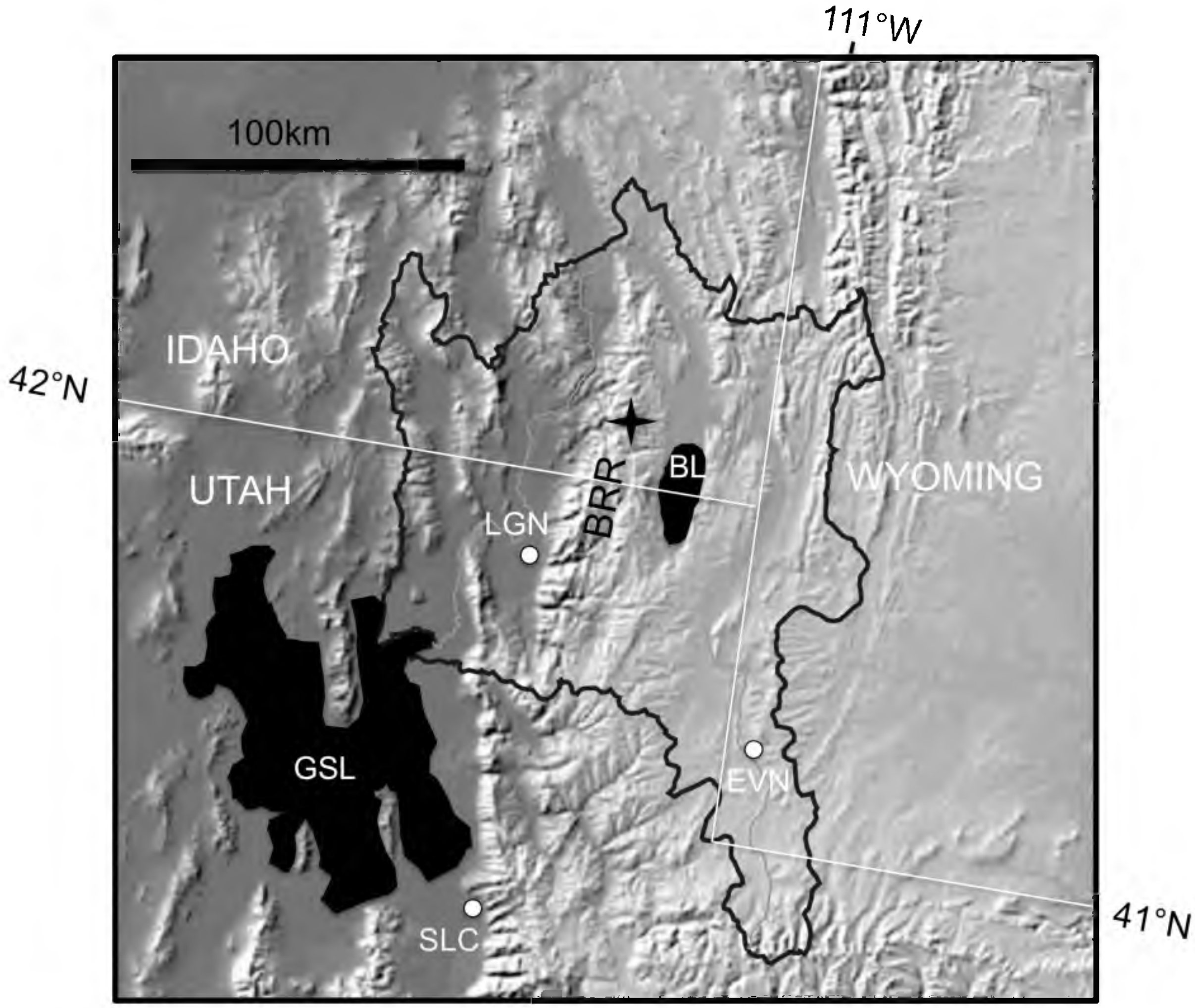

Figure 1.1. Study Area. BRR=Bear River Range, GSL=Great Salt Lake, BL=Bear Lake, LGN=Logan, ENV=Evanston, SLC=Salt Lake City. Black Line indicates boundary of Bear River Basin. Star indicates approximate location of study area.

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CHAPTER 2

A 14,000-YEAR RECORD OF FIRE AND VEGETATION FROM THE BEAR RIVER RANGE, SOUTHEAST IDAHO

Introduction

Research conducted over the past few decades has identified a complex relationship between climate, vegetation, and fire regime in terrestrial ecosystems. In most cases, climate has been interpreted to concurrently affect both fire regime and vegetation composition through time, with bilateral feedbacks between fire regime and forest composition (e.g., Long et al., 1998; Brunelle and Anderson, 2003; Prichard et al., 2009; Long et al. 2011). However, in some cases, climate is identified as the primary control on fire regime, independent of changes in vegetation composition (e.g., Millspaugh et al., 2000; Carcaillet et al., 2001).

Understanding climate-fire-vegetation interaction is important given that changes in forest fire activity have been observed in the Western U.S. (hereafter West) over the past several decades in response to changes in seasonal hydroclimatology (Westerling et al., 2006). Changes to forest composition and health have also been observed or are predicted to occur in response to global climate changes (Williams et al., 2010). Predicting and adapting management strategies to ongoing and future changes to vegetation composition and fire regimes across the West will be an important challenge for resource and conservation managers.

Paleoenvironmental reconstructions provide data that can be used to inform land management professionals about ranges of natural variability, rates of change, ecological resiliency, and ecosystem sensitivity to climatalogical forcings and natural disturbance agents (e.g., Froyd and Willis, 2008). In addition to aiding in forest management decisions, paleoenvironmental records can be used to identify and understand the mechanisms driving changes in the amount and seasonality of precipitation (Woodhouse, 2004), as has been observed across the West in recent decades (Barnett et al., 2004; Regonda et al., 2005; Carson, 2007). With growing water resource demands in many parts of the West, improved accuracy in long-term and year-to-year hydroclimate predictive capability is of vital need.

Here we present a 14,000-year record of fire and vegetation history from Plan B Pond, Bear River Range, southeast Idaho, based on pollen, macrofossils, and macroscopic charcoal from a lacustrine sediment core. The primary objectives of this research are to provide insights into the response of forest communities and fire regimes to past changes in climate, as well as add a spatial data point to further refine our understanding of spatiotemporal changes to hydroclimate in the West over the period of record.

Study Site

Plan B Pond (42.148°N, 111.579°W) is a small (~2.5 ha) closed-basin pond in the Bear River Range (BRR). The pond is located at 2500m elevation at the head of

Bloomington Canyon, approximately 30km southwest of Montpelier, Idaho (Figure 2.1). The pond has a drainage basin to lake surface ratio of approximately 9:1, and had a maximum depth of ~1.1 m at the time cores were collected in July, 2007. The hydrology of the pond is primarily groundwater-controlled. Plan B Pond is situated within a compound cirque basin, and surrounded by glacial till mapped as Holocene or Latest-Pleistocene in age (Reheis et al., 2009). Bedrock within and around the drainage basin is Paleozoic quartzite and dolomite (Oriel and Platt, 1980).

Modern climate data from the Franklin Basin SNOTEL station, located at 2450m elevation, approximately 10 km southwest of the study site, show a high degree of seasonality. Over the period 1983-2010, July temperatures on average ranged from 8° to 24°C, with a mean of 15.5°C. January temperatures, on average, ranged from -11° to + 1°C, with a mean of -6.5°C. Total precipitation was approximately 1200mm annually, with just more than 75% of the annual precipitation occurring from October-April, primarily falling as snow. July and August were the driest months, contributing less than 6% of the annual precipitation, combined.

Modern vegetation at the study site varies significantly with slope and aspect. Much of the open canyon bottom and steep south facing slopes are composed of open meadows and shrub communities with scattered occurrence of *Pseudotsuga menziesii* (Douglas fir), *Pinus flexilis* (limber pine), *Abies lasiocarpa* (subalpine fir), and *Picea engelmanii* (Engelmann's spruce). North facing slopes are generally composed of more closed *Psuedotsuga* or mixed *Abies/Picea* canopies. Lower in the valley *Pinus contorta* (lodgepole pine) has a significant presence, as well as both intermixed and dense patches of *Populus tremuliodes* (quaking aspen). Low-growing dwarf *Populus tremuloides* is also present in the area directly surrounding Plan B Pond. Examples of other plant species identified from the study site in 2007 include *Artemisia tridentata* (sagebrush), *Thalictrum occidentale* (western meadowrue), *Symphoricarpos oreophilus* (mountain snowberry), *Penstamon cyananthus* (Wasatch beard tongue), *Geranium richardsonii* (Richardson's geranium), *Oxyria dignya* (alpine mountan sorrel), *Agastache urticifolia* (nettleleaf giant hyssop), and *Lonicera involucrata* (twinberry honeysuckle).

Materials and Methods

A 3.98 m sediment core (PBP0701) was collected from the approximate center of Plan B Bond in ~1.1 m water using a modified Livingstone corer. The uppermost 43cm of sediment were not recoverable with the Livingstone corer due high water content and lack of consolidation. Sediments were recovered in 1 m increments and extruded in the field. Extruded core sections were described, wrapped in plastic wrap and aluminum foil, and stored in wood storage boxes for transport.

An 80 cm short-core (PBP0702) was collected to recover the uppermost, unconsolidated sediments that were lost from the Livingstone core. The short core was extruded in 1cm intervals into Whirlpak bags in the field prior to transporting back to the lab. All samples were immediately stored under refrigeration upon returning from the field site.

Sediments from Plan B Pond consisted of brown to greenish-brown organic rich gyttja in the upper 3.5 m of sediment. Occasional sand layers of 1-2 mm thickness were also observed. The lower 43 cm of the Livingstone core consisted of dense inorganic grayish-brown sandy clay, consistent with glacially derived sediment inputs of rock flour.

Sample preparation for charcoal analysis followed procedures and recommendations described by Brunelle and Anderson (2003). A 5 cc sediment sample was collected from each 1cm depth interval. The sample aliquots were soaked in a solution of hexametaphosphate (50 g/L) to aid in disaggregation. Due to the presence of abundant clay and unidentified mucilaginous material, the sodium hexametaphosphate solution was only marginally effective. The partially disaggregated samples were then rinsed through nested 250 μm and 125 μm sieves. Remaining samples from each size fraction were rinsed into gridded petri dishes and scanned under a 10X-70X magnification microscope for macrofossils. Macrofossil identification was verified using known samples from an index collection. The two size fractions were then recombined into a glass beaker and soaked in a 10% solution of chlorine bleach, heated to 50-60 °C for approximately 10 minutes to eliminate mucilaginous material, and then rinsed through nested sieves a second time. Once again, the two size fractions were placed in gridded petri dishes and charcoal particles of each size fraction were counted. For plotting, counts from the two size fractions of charcoal were combined.

Charcoal data analysis procedures generally follow those presented by Long et al. (1998), wherein charcoal counts are deconvolved into a low-frequency background component and a high-frequency peaks component with fire events identified by peaks that exceed a statistically significant threshold above the background. The deconvolution and identification of fire events from charcoal data was done using CharAnalysis software (Higuera et al., 2009). Specific input parameters to the CharAnalysis software are shown in Table 1. Based on the on the CharAnalysis results a running fire frequency (fire events/1000 years) and mean fire recurrence intervals (average years between fire events) were calculated.

Pollen sample preparation followed a method modified from Faegri et al. (1989), using a *Lycopodium* spore tracer. Samples were concentrated from 1cc of sediment, and collected from the core at intervals providing approximately equal age distribution, generally at multiples of 4 cm within the core. Actual sampling interval ranged from 4-32 cm (mean ≈ 10 cm), providing a mean temporal resolution of 362 years. Concentrated pollen samples were preserved in silicone oil. Each sample was counted at 500X magnification. A minimum of 300 terrestrial grains was counted, except in cases where the amount of recovered pollen sample was insufficient to do so. Pollen data are presented in relative percentages of the terrestrial taxa sum. CONISS (Grimm, 1987) was used to perform stratigraphically constrained cluster analysis. Pollen zones are numbered in stratigraphic order, oldest first.

Chronology

Chronological constraints are provided by four radiocarbon dates and a ^{210}Pb series measured from the upper 20 cm of core sediments (Table 2). An age model (Figure 2.2) was created from the ^{14}C and ^{210}Pb data by fitting a fourth-order polynomial to the data. For the upper 20 cm, actual ^{210}Pb ages were used in place of model results. Radiocarbon ages were calibrated to calendar year before present (Cal Yr BP), using the web version of the CALIB 6.0 radiocarbon calibration application utilizing the IntCal09 calibration dataset (Stuiver and Reimer, 1993). All ages reported as before present (BP) are in calibrated Calendar Years BP unless specifically noted otherwise.

Results

Charcoal

Smoothed background charcoal accumulation rates (Figure 2.3b) are consistently low from the beginning of the record to approximately 7000 years ago. Periods of extremely low charcoal accumulation occur at ~12,500 BP and again between 10,000-8,000 BP. After 7000 BP, background charcoal accumulation shows a steady increase until around 1800 BP. From 1800 BP to 800 BP charcoal background shows a steady decline to values similar to those seen in the early Holocene. After ~800 BP charcoal background abruptly returns to higher values and remains high up to modern.

Based on the analyses of the peaks component of the charcoal data (Figure 2.3a), long-term fire frequency averages 3.47 fires/1000 years, giving a mean fire return interval of 288 years. Fire frequency is lower than the long-term average over the intervals 12800-11900 BP, 11100-7500 BP, 4500-1800 BP, and 1150-800 BP. Fire frequency is higher than average over the intervals 13900-12800 BP, 11800-11100 BP, 7500-4500 BP, 1800-1200 BP, and from 600 BP to modern. The magnitude of charcoal peaks is generally lower in the early half of the record than the latter half, with nine of the ten largest peaks occurring after 6500 BP. The largest peak in the record coincides with a historically documented fire that occurred in Bloomington Canyon in 1934 (Wilde, 1980), coincident with widespread fire activity throughout the region that year (Barrett, 1994).

Pollen and Macrofossils

Pollen results from Plan B Pond (Figure 2.4), for the most part, show gradual trends in taxa abundance over time, rather than discrete periods bracketed by abrupt composition changes. However, as is common practice with pollen data, the pollen spectra into discreet pollen zones to simplify discussion of the data. Five pollen zones were defined using CONISS (Grimm, 1987).

Pollen Zone 1 (13800-11300 BP) shows high relative abundances of *Artemisia*, *Pinus*, *Picea*, and *Asteraceae* pollen. The last few hundred years included in Zone 1 begin to show a modest, but notable increase in the abundance of *Ambrosia*, *Pseudotsuga*, and *Amaranthaceae*. The overall ratio of arboreal to nonarboreal pollen (AP/NAP) is quite low over this interval, and abundance of the algal palynomorph *Botryococcus* is the lowest at any time in the record. Macrofossil data confirms the local presence of *Abies*, *Picea,* and *Pinus flexilis* by 13000 BP.

Pollen Zone 2 (11300- 9400 BP) continues to be dominated by high abundances of *Pinus* and *Artemisia*, although *Artemisia* declines steadily throughout the interval. *Picea* declines to its lowest level in the record at the beginning of the zone, and remains uniformly low for the duration of the interval. *Abies* also reaches its lowest levels in this zone. In contrast, *Ambrosia* reaches its highest abundance in Zone 2. Amaranthaceae shows a small but steady increase over the interval, and the overall AP/NAP ratio remains low and steady. *Botryococcus* increases through the zone, reaching a localized peak just after 10,000 BP. Macrofossils of *Picea*, *Abies*, and 5-needle *Pinus* all continue to be present in Zone 2, although the occurrence frequency of *Picea* needles is lower than in Zone 1, while the occurrence frequency of *Pinus* needles is higher, at least prior to

~10300 BP. *Pseudotsuga* macrofossils are also present regularly in Zone 2, in contrast to their absence in Zone 1.

Pollen Zone 3 (9400-6200 BP) shows a decline in *Pinus* abundance compared to the previous two zones. *Artemisia* remains high with a slight increasing trend. *Ambrosia* remains near the high levels seen in Zone 2, with a very slight declining trend. *Abies* and *Picea* both show a gradual increase over the duration of Zone 3. Amaranthaceae is consistently high in Zone 3, reaching its highest abundance during this time period, along with *Sarcobatus*. Despite the first appearance or more consistent presence of a few arboreal taxa such as *Quercus*, *Populus*, *Acer*, and *Juniperus*, the AP/NAP declines to its lowest value in Zone 3. *Bottryoccoccus* initially declines in abundance in Zone 3, reaching a local minimum just before 8000 BP. After 8000 BP, *Botryococcus* abundance increases sharply until 7000 BP, and then levels off. Macrofossils show continued steady occurrence of *Abies* needles, and regular but less frequent occurrences of *Picea* and *Pseudotsuga*. *Pinus flexilis* needles also occur on a more sporadic basis. Macrofossil evidence for *Pinus* and *Pseudotsuga* is absent after 7000 BP.

Pollen Zone 4 (6200-2500 BP) is a period of marked transition. *Abies* shows a steady increase along with a more stepwise increase in *Picea*. *Artemisia* shows a steady decline, as does *Ambrosia*, although the *Ambrosia* decline is not as pronounced as that of *Artemisia*. Amaranthaceae remains near its highest levels in the record. The AP/NAP ratio shows an increasing trend, especially after 4000 BP. *Botryococcus* abundance hits its highest level in the Holocene at the beginning of Zone 4, just after 6000 BP, then decreases sharply until 4700 BP. After 4700 BP *Botryococcus* continues its decline, but at a much slower rate up to 2500 BP. *Picea* and *Abies* macrofossils are present throughout Zone 4, but no macrofossil evidence for *Pseudotsuga* or *Pinus flexilis* is present in the zone. One needle from a 2-needle pine (likely *Pinus contorta*) occurs at 2800 BP, and represents the only macrofossil from a 2-needle pine in the entire record.

In Pollen Zone 5 (2500 BP-Modern), *Pinus* abundance returns to high levels similar to those seen in Zone 1 and Zone 2. *Artemisia* continues its decline, while *Abies* and *Picea* continue to increase in abundance. *Psuedotsuga* is slightly higher than in the previous two pollen zones, while *Ambrosia* and *Amaranthaceae* are lower in this period. A very noticeable increase in *Cyperaceae* also occurs in Pollen Zone 5. The AP/NAP ratio increases to its highest value in the record in Pollen Zone 5. *Botryococcus* abundance in Zone 5 initially increases to a local maximum at 1700 BP, before dropping back nearly to levels seen at the 2500 BP minimum. Macrofossil evidence is dominated by *Abies* needles in Zone 5. Occurrences of *Picea*, *Pinus*, and *Pseudotsuga* macrofossils are limited to one or two samples each, with the *Pinus* and *Pseudotsuga* occurring early in the zone and the *Picea* occurring late.

Discussion

Vegetation and Insolation

Overall, vegetation changes throughout the Plan B Pond record show a close relationship with mid-to late summer (July-August) insolation. This relationship is clearly evident when the data are plotted as the ratio of cool/moist to warm/dry (CM:WD) indicator taxa (Figure 2.5). Here we use *Picea* +*Pseudotsuga* +*Abies* + other Asteraceae as the cool/moist taxa and *Juniperus* +*Quercus* +*Ambrosia* +Amaranthaceae +*Sarcobatus* as the warm/dry taxa. These taxa are an amalgamation of those used by

Davis (1998) to define “glacial” vs. “interglacial” taxa in Great Salt Lake pollen records and those used by Jimenez-Moreno et al. (2007) to define “cold” vs. “warm+dry” in Bear Lake pollen records.

With the exception of a notable and distinct cool/moist excursion coincident with the Younger Dryas period (12,900-11,500 BP) and increased variability after 2500 BP, the pollen ratio in Figure 5 generally follows mid- to late summer (July + August) insolation changes, and suggests that summer temperatures are the dominant variable controlling vegetation composition for most of the Holocene, either directly or indirectly by affecting evapotranspiration rates and effective moisture. This is similar to the conclusions of Power et al. (2011), which also found that vegetation in high elevation sites in the Northern Rocky Mountains was most sensitive to summer temperature, especially in the early to middle Holocene when summer insolation was highest.

Summer Temperatures

The direct importance of summer temperature is particularly well demonstrated by the taxa, *Picea* and *Ambrosia*. Throughout the record, relative abundances of *Picea* and *Ambrosia* generally behave in an offsetting manner, with one increasing at the apparent expense of the other. An examination of published relationships between climate variables and relative pollen abundance in modern sediments (Williams et al., 2006) shows that mean July temperature is the one climate variable where *Picea* and *Ambrosia* show little overlap in their distribution, each with abrupt thresholds at approximately 17.5 °C (Figure 2.6a). *Picea* shows a sharp increase in abundance when mean July temperatures are below this threshold and *Ambrosia* increases sharply when temperatures are above the threshold. Currently, mean annual temperature at Plan B Pond is ~ 15.5 °C, supporting a high ratio of *Picea* to *Ambrosia* (Figure 2.6b).

The decline in *Picea* to minimal levels at 10500 BP and lasting until 9500 BP, as *Ambrosia* was reaching its peak Holocene abundance, may suggest that summer temperatures at Plan B Pond peaked near or above the 17.5 °C threshold, equal to or greater than 2 °C warmer than modern. This is consistent with midge-based summer temperature reconstructions from SE British Columbia that suggest peak temperatures 3-4 °C warmer than modern from 10500-9000 BP (Chase et al., 2008).

Based on the CM:WD ratio, near-peak or at least warmer than modern summer temperatures were sustained throughout Pollen Zones 2 and 3 (11400-6200 BP). This is also consistent with midge inferred temperature record of Chase et al. (2008) which shows continued warmer than modern temperatures by ~2 °C until ~6500 BP. However, the bulk of this period of maximum summer warmth at Plan B Pond (from ~11100-7500BP) is characterized by lower than average fire frequency. This suggests that while summer temperatures have been the dominant driver of vegetation change, other variables may have been more important for determining fire activity, despite the large effects that such warm summer temperatures would have had on evapotranspiration and fuel/soil-moisture.

Background Charcoal and Fuels

Marlon et al. (2006) showed background charcoal accumulation rate is largely controlled by the availability of woody fuel (proportion of woody taxa on landscape), and is generally unaffected by changes in fire frequency or sedimentation rate. A comparison of charcoal accumulation rate and AP:NAP from Plan B Pond confirms this relationship (Figure 2.7) at our site, given that arboreal species represent the bulk of woody fuel. Background charcoal trends correspond well with the fuel available on the landscape, with the exception of departures at ~12500 BP, 9000 BP, and 1000 BP when fuel availability increases as charcoal background decreases. These three time intervals may represent periods of time when climatic controls limiting fire activity allowed woody fuels to accumulate. Fire frequency is in decline during each of the three anomalous intervals and suggests a more or less synchronous response between fire activity and vegetation composition, or a perhaps a causal response with fire causing a vegetative change. However, the relationship at these three discrete time periods are not representative of the record as a whole.

Charcoal Peaks/ Fire Frequency and Winter Snowpack

Today, fire frequency in the West is strongly linked to the length of fire season, a function of the timing of spring snowmelt (Westerling et al., 2006). The timing of spring snowmelt is affected by both spring temperatures and winter precipitation, but Westerling et al. (2006) found that when comparing early and late spring snowmelt years in the West there was a strong association between warm spring temperatures and reduced winter precipitation. This association was also noticed by Cayan (1996) when exploring the interannual variability in snowpack across the West. Thus, the two variables, winter precipitation and spring temperature, do not tend to behave independently. Rather, one can infer that periods of higher than average fire activity likely exhibit both low snowpack and warmer than average spring temperatures, while low fire activity is associated with higher than average snowpacks and cooler spring temperatures.

Based on this relationship between fire activity and hydroclimatic conditions we hypothesize that the time periods 12800-11900 BP, 11100-7500 BP, 4500-1800 BP, and 1150-800 BP were characterized by higher than average snowpacks and cool springs, while the periods 11800-11100 BP, 7500-4500 BP, 1800-1200 BP, and from 600 BP to modern were characterized by lower than average snowpacks and warm springs. The frequency of charcoal peaks from 13900-12800 BP is also higher than average, but because sediments over most of this interval are inorganic glacial flour we are not confident that charcoal in these earliest sediments is not reworked material. Thus, we are hesitant to provide any definite interpretation of the charcoal content.

The accuracy of our overall fire frequency reconstruction is bolstered by an obvious correlation between *Botryococcus* abundance and our reconstructed fire frequency curve (Figure 2.8). The *Botryococcus* increases during periods of high fire activity (dry winters/warm springs) and declines during periods of low fire activity (wet winters/cool springs). The correspondence of fire activity and *Botryoccoccus* abundance is most likely explained by a common dependence of each variable on winter snowpack. Groundwater hydrology in the BRR is snowmelt dominated (Spangler, 2001), and lake level in Plan B Pond is primarily groundwater controlled, so it is expected that during periods of higher than average snowpacks lake level would remain higher. In addition, water temperatures would likely be cooler, and the ice-free period would be shorter. In lacustrine deposits, especially from small lakes, *Botryococcus* is often an inverse indicator of water level (Clausing, 1999), and has been shown to be weakly, but positively correlated with temperature (Rull et al., 2008). Thus, higher than average snowpacks at Plan B Pond would likely lead to reduced *Botryococcus* abundance. Decreased fire activity associated with high snowpacks could also limit nutrient delivery to the pond as compared to periods of high fire activity, so some of the correlation between fire activity and *Botryococcus* may in fact be causal.

Summer Precipitation and Fire

One might argue that an alternative explanation for the low fire frequency in the BRR during the Early Holocene was increased summer precipitation due to enhanced convective precipitation ("summer monsoon") resulting from higher than modern summer insolation. This mechanism has been widely applied to explain climate, fire, and vegetation reconstructions across much of the Western U.S. (e.g., Whitlock and Bartlein, 1993; Whitlock et al., 1995; Brunelle et al., 2005). However, in the course of said investigations, it has consistently been shown that that the effective spatial footprint of summer monsoonal precipitation has remained fixed throughout the Holocene, such that sites which today lack significant summer precipitation ("summer dry/winter wet" sites in Whitlock and Bartlein, 1993), were not significantly affected by summer precipitation in the Early Holocene either. Rather than a change in spatial footprint, the summer precipitation gradient increased between the "summer-wet" and "summer-dry" sites (Whitlock et al., 1995; Brunelle et al., 2005). As Plan B Pond is a definitive summer-dry site today, we do not consider intensified summer convective/ "monsoon" activity in the early Holocene a likely cause of reduced fire activity from 11100-7500 BP.

Climate/Fire/Vegetation Relationships

Overall, the data from Plan B Pond suggests that pollen assemblages and macroscopic charcoal data are each primarily explained by one dominant climatic variable, but not the same dominant climatic variable. Vegetation composition as indicated by pollen assemblages is most responsive to mid to late summer temperature, a function of summer insolation. In some cases, temperature itself may have been limiting, as discussed previously with the taxa, *Picea* and *Ambrosia*. In other cases, the effect of temperature on evapotranspiration and effective growing season moisture may have been more important. Variability in fire frequency at Plan B Pond is independent of vegetation change, and most likely has been controlled by the length of the fire season throughout the Holocene.

Holocene Paleoenvironments

Since fire season is a function of winter precipitation and spring/early summer temperature, low fire frequency during most of the early Holocene (11100-7500 BP) suggests that winter precipitation was higher than the long-term average during that time, offsetting the enhanced evapotranspiration that would be expected during the peak of summer insolation.

With higher than average winter snowpack partially or completely offsetting higher summer temperatures, we speculate that peak hydrologic drought conditions would have occurred during the interval of time when summer temperatures were still near their peak (before ~6200 BP), but after winter precipitation decreased (after ~7500 BP) from the high levels of the early Holocene. This 1300-year window coincides with many regional paleoenvironmental indicators of intense drought conditions. For example, the Great Salt Lake is thought to have nearly desiccated between approximately 6800-7500 BP (6700-6000 ^{14}C yrs BP) (Murchison, 1989), Lundeen (2001) documented an interval of lake desiccation in the Sawtooth Mountains of Central Idaho from ~7500-7000 BP (6800-6200 ^{14}C yrs BP), and both palynological and isotopic indicators from Pyramid Lake, Nevada suggest the dominance of intensely dry conditions from 7500-6300 BP (Benson et al., 2002; Mensing et al., 2004). Model results also show minimum Holocene moisture index values, a measure of annual effective precipitation for plant growth, occurred from ~7800-5700 BP (7000-5000 ^{14}C yrs BP) in the northeastern Great Basin (Broughton et al., 2008).

Beginning in Pollen Zone 4 (~6200BP), pollen data suggest the initiation of a consistent trend toward cooler summer conditions and increased effective moisture, consistent with declining summer insolation. This can be seen as an inflection point in the AP:NAP and the CM:WD ratios, and as a dramatic shift in the *Picea*:*Ambrosia* ratio. Although the effective moisture was generally increasing over the entirety of period Pollen Zone 4 (6200-2500 BP), higher than average fire frequency continued until 4500 BP suggesting winters were still relatively dry up to that point. A stepwise increase in *Picea* abundance, a drop in *Botryococcus*, and a slight increase in herbaceous pollen also occur at ~4500 BP, all indicating a shift to wetter conditions.

Based on the fire frequency reconstruction, wetter than average winters persisted from ~4500 BP, through the remainder of the Pollen Zone 4 chronozone, and into Pollen Zone 5, finally returning to drier conditions ~ 2000 BP. Pollen data suggests a continued cooling trend after 4500 BP, and an overall increase in effective moisture. However, the transition from the Pollen Zone 4 to Pollen Zone 5 at ~2500 BP marks a shift to a climatic regime with much higher frequency and higher magnitude variability than seen previously in the record. This shift is most evident in the CM:WD pollen ratio. In addition to the increased variability in the pollen ratio, a sharp increase in aquatic taxa pollen (mostly Cyperaceae) can be seen after 2500 BP. We interpret the increased Cyperaceae after 2500 BP to be indicative of a more variable lake level that would consistently create new lake-margin habitat for Cyperaceae in both transgressive and regressive phases.

Increased climatic variability in the late Holocene is consistent with the results of several studies, especially as it relates to variability in El Niño-Southern Oscillation (ENSO) (e.g., Moy et al., 2002; Conroy et al., 2008; Barron and Anderson, 2011). The primary effects of ENSO variability in the study region are changes to the amount of winter precipitation. Specifically, winters in negative ENSO phases (La Niña) tend to be wetter than average and winters in positive ENSO phases (El Niño) tend to be drier than average (Cayan, 1996; Jain and Lall, 2000). Reconstruction of ENSO activity by Moy et al. (2002) and Conroy et al. (2008) both suggest that the frequency of El Niño events increased in the late Holocene, with the highest frequency at ~2000 BP (Conroy et al., 2008) or shortly thereafter (Moy et al., 2002). Based on a compilation of records, Barron and Anderson (2011) suggests that that ENSO variability in the eastern North Pacific began to intensify after ~4000 BP in Southern California, followed by a northerly migration of the ENSO teleconnection pattern over the next couple of millennia.

Underlying the intensified climate variability, likely caused by ENSO intensification, Pollen Zone 5 is reflective of effectively wet conditions, as compared to the three previous zones. This is reflected by the highest AP:NAP values in the record, driven in part by high contributions of pine in conjunction with declining *Artemisia* (Figure 2.9). The ratio of *Pinus:Artemisia* is often interpreted as being an indicator of lower treeline in montane systems, primarily controlled by effective moisture (Fall, 1997; Jiménez-Moreno et al., 2010). A declining trend in summer insolation explains part of the effective moisture increase; however, the changes in winter precipitation due to increased ENSO variability significantly impacted fire regime, lake level, and vegetation at shorter timescales, making paleoclimatic interpretations more complex in Pollen Zone 5 than earlier portions of the record.

Medieval Climate Anomaly

One sub-interval in Pollen Zone 5 (~1800-750 BP) has a particularly interesting suite of proxy indicators, suggesting unique environmental conditions at that time. This sub-interval at Plan B Pond includes the time period widely recognized as the medieval climate anomaly (MCA), occurring ~1000-700 BP (Mann et al., 2009). Anomalous conditions at Plan B Pond begin to occur considerably earlier than the traditionally accepted onset of the MCA, but the peak temperature and moisture anomalies do occur during the MCA, as discussed below.

Background charcoal accumulation at Plan B Pond decline from 1800-750 BP, despite AP:NAP ratio (indicating fuel availability) reaching its highest values between 1300 BP and 700 BP (Figure 7). The disconnect between fuel availability and background charcoal production suggests that fires were limited to small ground fires rather than larger magnitude crown fires, which would have helped bolster the AP:NAP increase. The ratio of *Pinus:Artemisia* also peaks increases to peak values between 1300 BP and 700 BP suggesting effectively wetter conditions. However, peaks in fire frequency and *Botryococcus* suggest reduced winter snowpack and/or warmer spring/early summer temperatures, at least during the first half of the sub-interval (~1800-1300 BP). The *Picea:Ambrosia* ratio also show an increase at this time indicating warmer summer temperatures.

Together these data suggest that from 1800-700 BP, but especially from 1300-700 BP, climate was warmer and effectively wet overall. Because at least part of this interval shows evidence of reduced snowpack, we infer that the time period must have experienced anomalously high summer precipitation. This interpretation is supported by the isotope record from carbonates in Bison Lake, Colorado that also suggests a shift in seasonal precipitation regimes, with a greater relative importance of rain from ~1600-600 BP (Anderson, 2011). The possibility of significantly increased summer precipitation in the Late Holocene is at odds with our earlier dismissal of intensified monsoonal precipitation in the Early Holocene. However, the combination of proxy indicators in the Early Holocene do not indicate the same conditions seen from 1800-700 BP with the possible exception of a few centuries ~9000 BP. Thus, if summer precipitation did increase in the Early Holocene, it does not appear that it sufficiently offset the higher rates of evapotranspiration associated with higher insolation and summer temperatures, and did not greatly affect moisture availability for vegetation.

Effective moisture likely increased after ~1300 BP when a drop in fire frequency and *Botryococcus* suggest snowpacks returned to normal or above normal until ~700 BP. Together with increased summer precipitation, one could infer that the time period from

1300-700 BP would have been exceptionally wet, with long moist summers ideal for plant productivity. It is not a coincidence that this interval corresponds with the peak occupation period of the Fremont Culture in the northern Great Basin and adjacent areas (Coltrain and Leavitt, 2002).

After ~700 BP, the declining *Pinus:Artemisia* ratio suggests more xeric conditions than in the preceding millennia. This is corroborated by an increase in fire frequency, and is synchronous with the abandonment of many Fremont sites throughout the region.

Paleoenvironmental Summary

Late Glacial/Younger Dryas

Dominant sedimentation of inorganic glacial clays in Plan B Pond prior to ~13000 BP suggests that small cirque glaciers were present in Bloomington Canyon well into the late-glacial. By 13000 BP, macrofossil evidence shows local presence of *Picea*, *Abies*, and *Pinus flexilis*, and pollen ratios indicate that conditions were ameliorating. Just after 13000 BP, coincident with the onset of the Younger Dryas period, conditions became much cooler and effectively wetter. By 12000-11500 BP climate in the BRR had mostly returned to its pre-Younger Dryas trend of warming temperatures.

Early Holocene

Summer temperatures continued to climb into the Early Holocene, driven by increases in summer insolation. Peak summer temperatures were reached 10500-9000 BP, and may have been >2°C warmer than modern. Summer temperatures remained close to 2°C warmer than modern until ~6200 BP. Despite warm summers and the expected effects of increased evapotranspiration, annual moisture balance was partially offset by larger than average snowpacks that kept fire activity low until ~7500 BP. Peak hydrologic drought in the BRR would have occurred from 7500-6200 BP when summer temperatures were still much higher than modern, and no longer offset by higher than average winter precipitation.

Late Holocene

After ~6200 BP declining summer temperatures drove effective moisture up, initiating a steady increase in the relative proportion of arboreal vegetation; this, despite lower than average snowpacks until ~4200 BP. From 4200-1800 BP, increased winter precipitation kept fire frequency lower than average, and effective moisture continued to climb. After ~2500 the overall climate variability increased significantly at the study site, consistent with an intensification of ENSO activity. Continued declines in summer insolation, driving a trend toward effectively wetter conditions, were overlain by higher frequency oscillations in winter precipitation.

In addition to, or possibly resulting from the increased ENSO variability, an anomalous warm, but wet period began ~1800 BP, and climaxed between 1300-700 BP, coincident with the MCA. This warm, wet interval was unique in that the extra moisture was not from excess snowpack, but rather summer precipitation. No other interval in the record has the same combination of proxy indicators requiring a “summer-wet” interpretation. After 700 BP, data suggest cooler but drier conditions at Plan B Pond than during the MCA.

Conclusions

The primary objectives of this study were to provide insights into the response of forest communities and fire regimes to past changes in climate, as well as add a spatial data point to further refine our understanding of spatiotemporal changes to hydroclimate in the West over the period of record. The paleoenvironmental record from Plan B Pond, Bear River Range, Southeast Idaho, offers a 14000 record of vegetation, fire, and climate relationships. Throughout the record, the dominant vegetation changes are driven primarily by summer insolation, either directly through phytological temperature limits, or indirectly by affecting rates of evapotranspiration. In a select few instances at ~1000 BP, 9000BP, and 12500 BP, climatically controlled reductions in fire activity may have influenced vegetation composition. Otherwise, fire frequency and vegetation change have largely been independent of one another in the BRR. Fire frequency is primarily driven by the magnitude of winter snowpack and spring/ early summer temperatures, which both affect the timing of spring snowmelt and the length of the fire season.

In the Late Holocene, higher frequency and amplitude variability, indicative of a strengthening ENSO system are overlain on the longer summer insolation driven climate trend. In addition seeing the intensification of ENSO variability at Plan B Pond, definitive Younger Dryas cooling and a distinctive warm, wet MCA are apparent in the data. The MCA anomaly appears to have began several hundred years earlier in the BRR than the global average, but shows peak temperature and effective moisture anomalies in synch with the accepted MCA time window.

An important contribution of this record is the recognition of seasonally biased proxies. Pollen data from Plan B Pond are clearly most sensitive to summer temperature and effective growing season moisture, while charcoal data are relatively insensitive to effective growing season moisture conditions, responsive to winter precipitation patterns instead. Recognition of these seasonal biases will help us have a better understanding of the hydrological variability we infer from paleoclimate records.

Table 2.1. Input parameters input to CharAnalysis software for analysis of Plan B pond charcoal record.

	Parameter	Value Used	Unit
Pretreatment	Years Interpolated	10	year
	Log Transform ?	None	
Smoothing	Method	Lowess smoother	
	Background Smoothing	300	year
Peak Analysis	Peak identification (residuals/ratios)	Ratios	
	Threshold Type (Local/Global)	Local	
	Threshold Method	Base threshold values on a percentile cut-off of a noise distribution determined by a Gaussian mixture model.	
	Threshold Values	0.990	Percentile of noise distribution
Peak Analysis Results	Smoothing for fire frequency and return interval	1000	year

Table 2.2. Chronological data used to construct age-depth model for Plan B Pond sediment cores.

DEPTH (CM)	Type	^{14}C Yrs BP (1950)	1-sigma ^{14}C error (^{14}C Yrs)	Calendar Yrs BP (1950)
1	^{210}Pb	-	-	-54.7
3	^{210}Pb	-	-	-49.2
5	^{210}Pb	-	-	-42.8
7	^{210}Pb	-	-	-35.1
9	^{210}Pb	-	-	-26.0
11	^{210}Pb	-	-	-15.1
13	^{210}Pb	-	-	-3.5
15	^{210}Pb	-	-	6.9
17	^{210}Pb	-	-	15.8
19	^{210}Pb	-	-	51.1
49	^{14}C	910	24	840
125	^{14}C	3540	30	3800
248	^{14}C	8400	40	9450
358	^{14}C	10990	70	12900

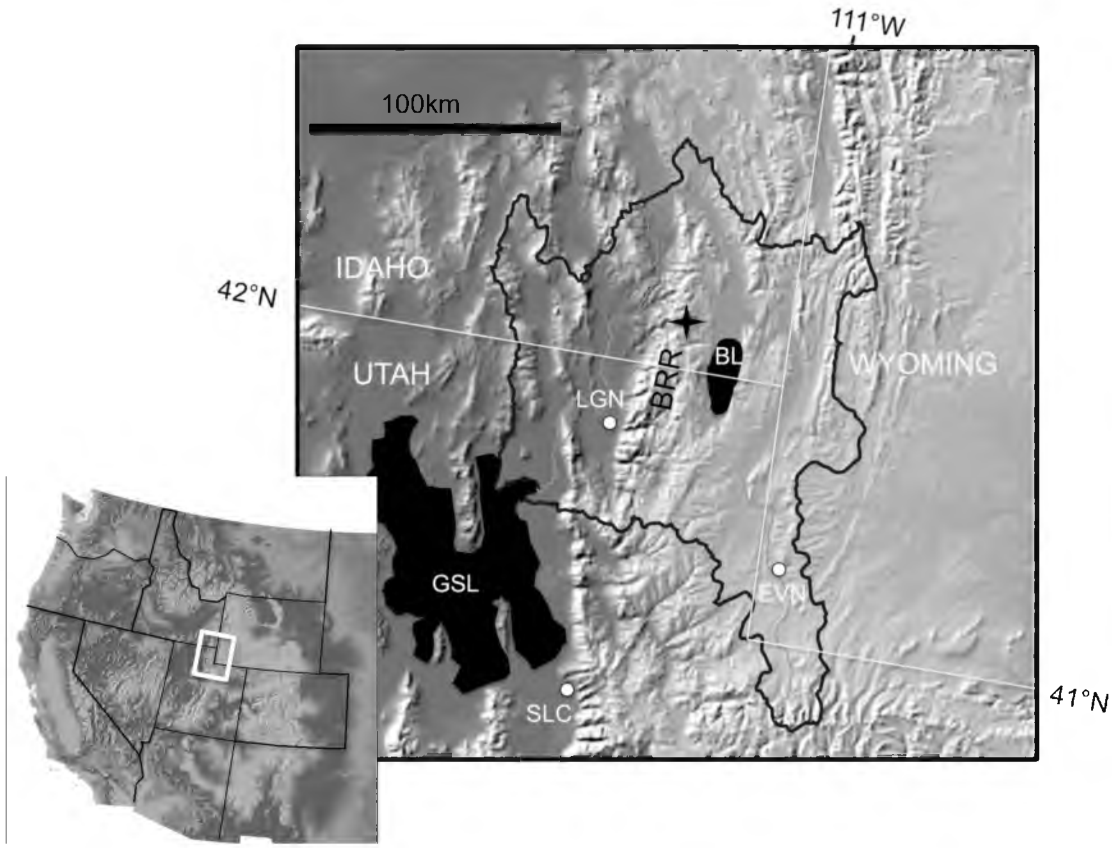

Figure 2.1. Plan B Pond Location Map. BL=Bear Lake; BRR= Bear River Range; EVN=Evanston, GSL= Great Salt Lake; LGN= Logan. Star indicates approximate location of Plan B Pond.

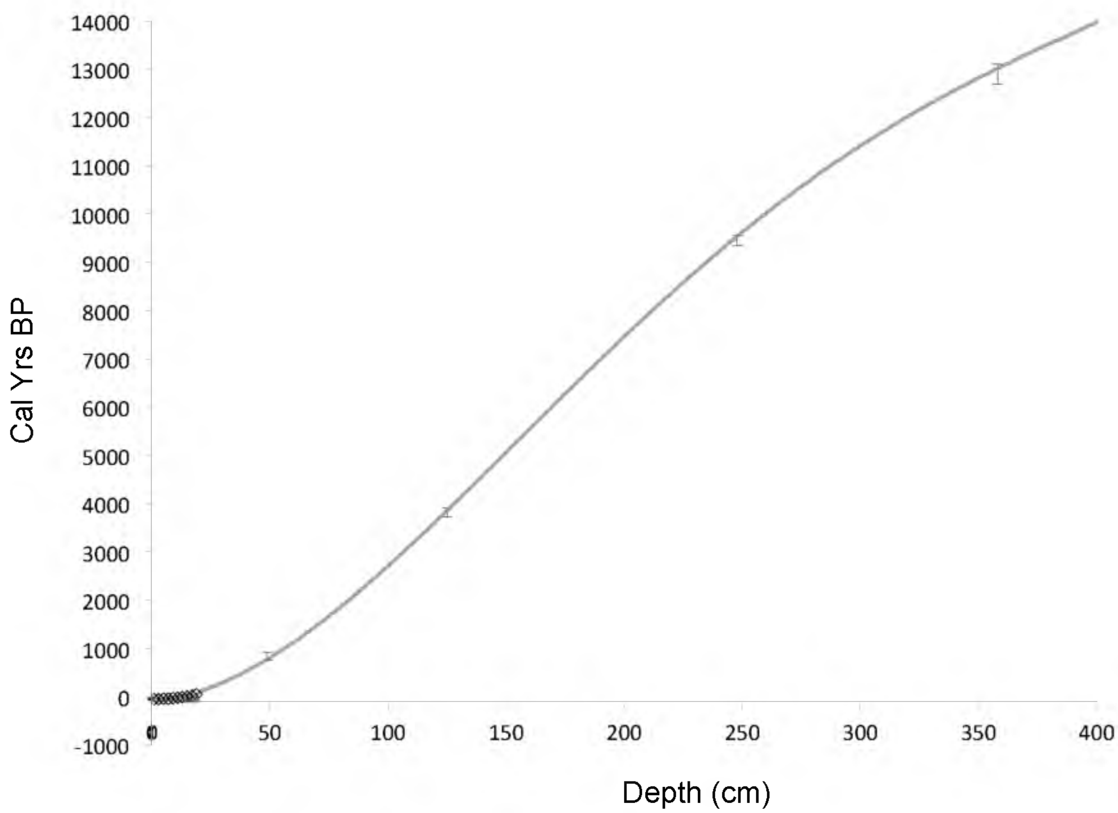

Figure 2.2. Plan B Pond Age-Depth Model. Vertical bars indicate calibrated ^{14}C ages. Diamonds indicate ^{210}Pb ages. Gray line indicates polynomial model used.

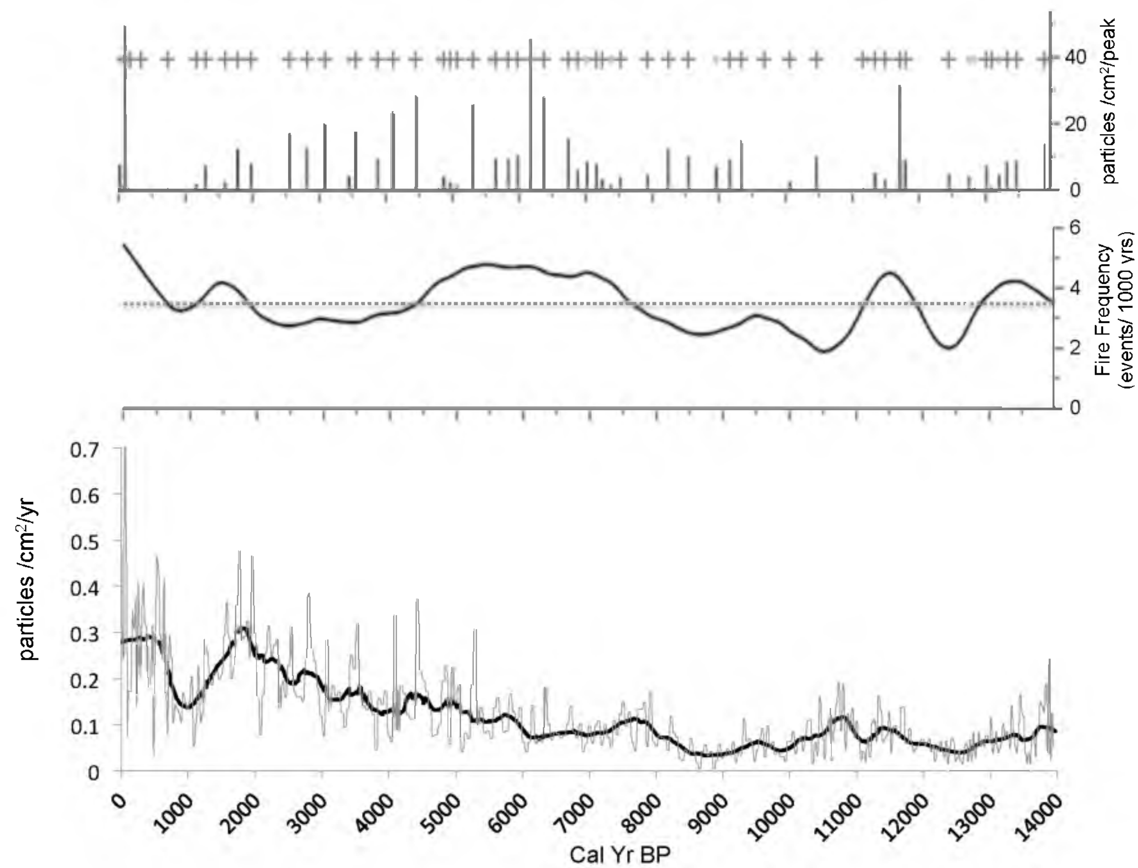

Figure 2.3. Plan B Pond Charcoal Data. Top panel- peak occurrence and magnitude. Middle Panel- 1000 year smoothed fire frequency. Bottom Panel- charcoal accumulation rate (light gray line) and 300-year smoothed background (black line).

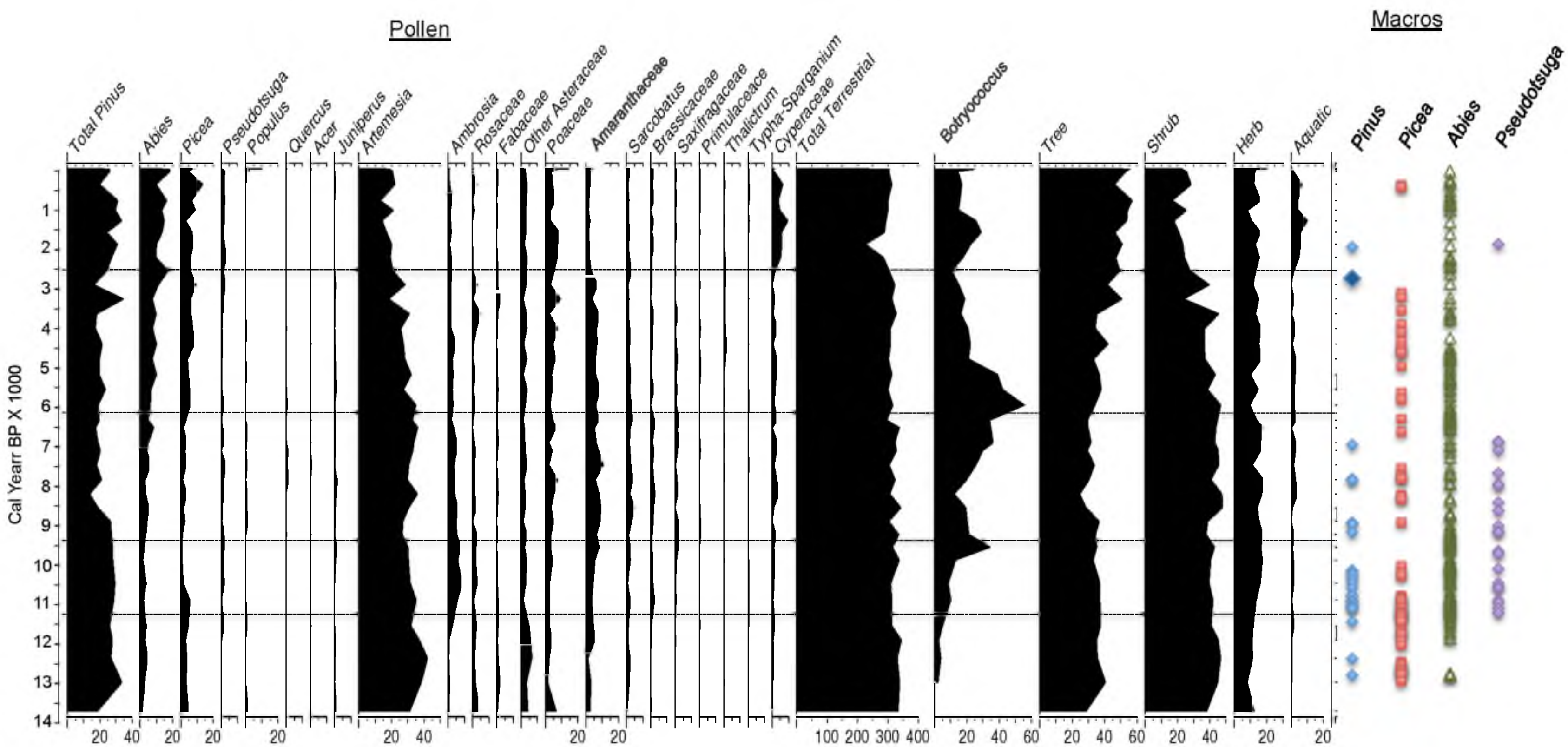

Figure 2.4. Plan B Pond Pollen and Macrofossil Data. Pollen data presented as relative percent of total terrestrial taxa pollen sum. Horizontal lines indicate pollen zones based on stratigraphically constrained cluster analysis. Macrofossil data indicates presence/absence in given sample.

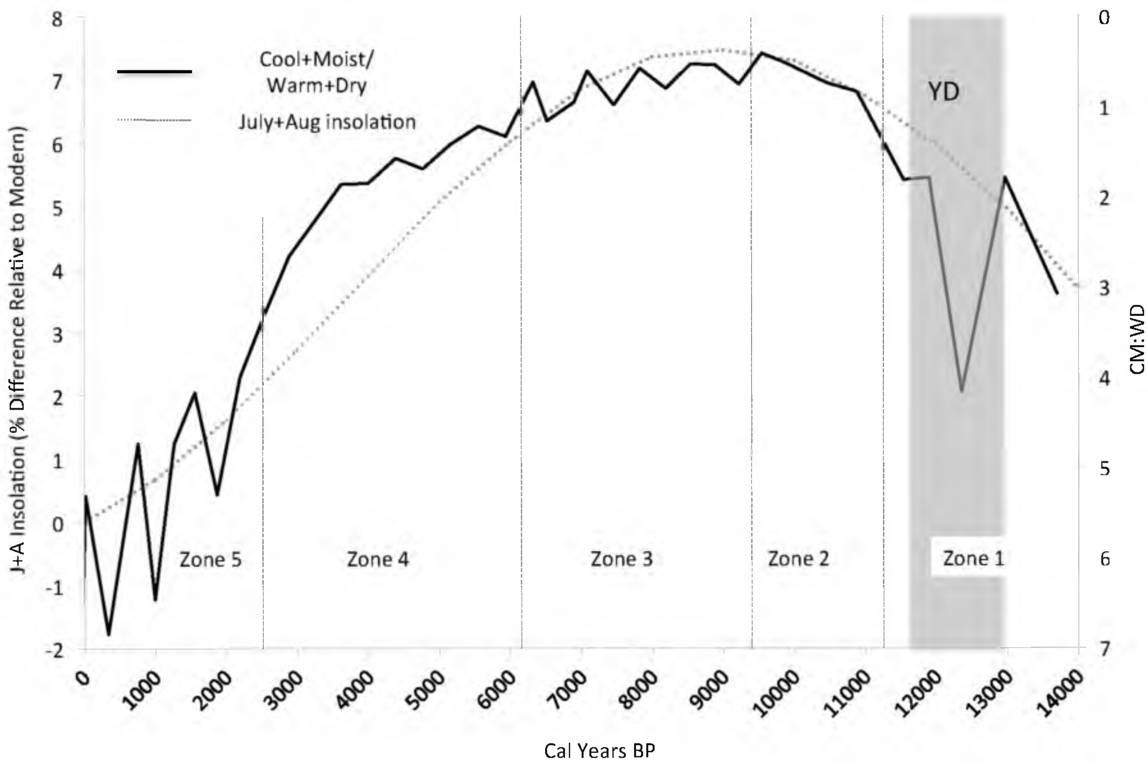

Figure 2.5. Plan B Pond Cool+Moist Taxa : Warm+Dry Taxa Pollen Ratio. Cool+Moist=*Picea* +*Pseudotsuga* +*Abies* + other Asteraceae. Warm+Dry= *Juniperus* +*Quercus* +*Ambrosia* +Amaranthaceae +*Sarcobatus*. Pollen zones from Figure 4 are shown by vertical lines. Shaded bar indicates Younger Dryas period.

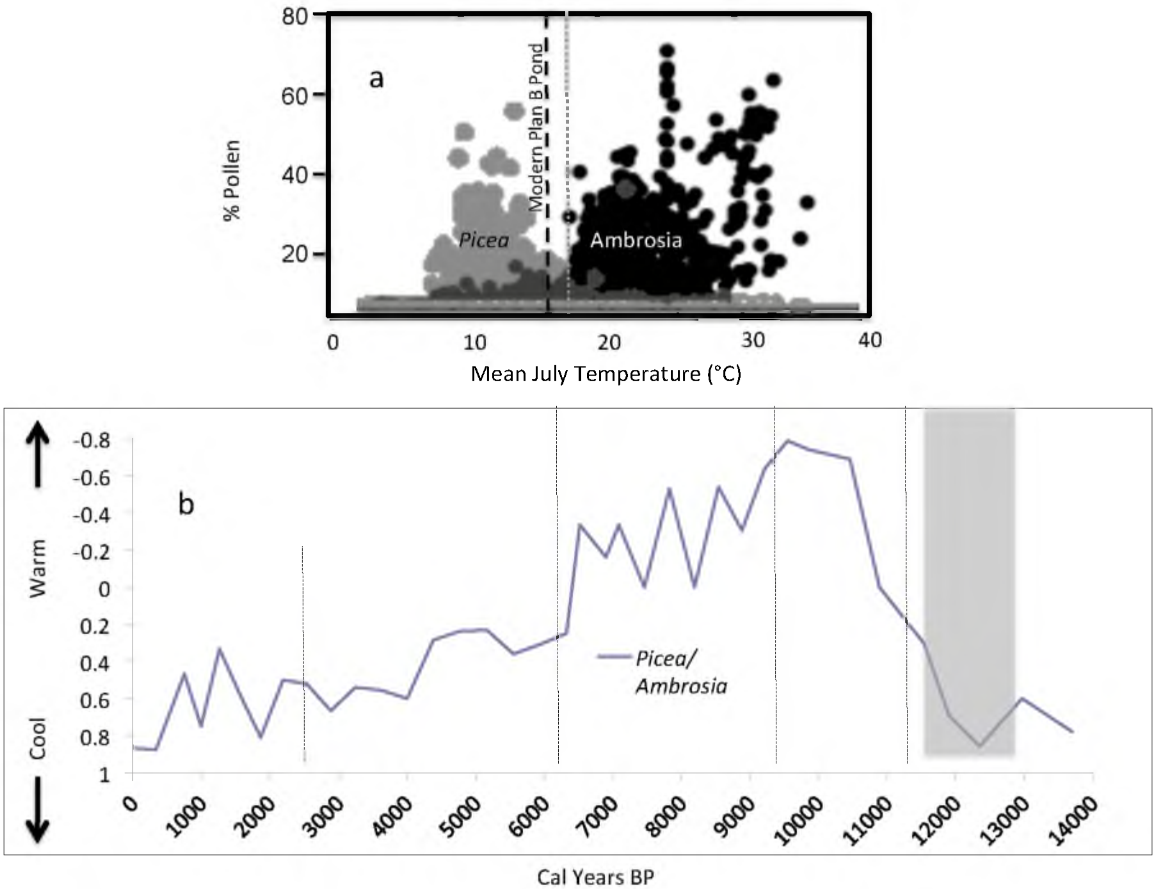

Figure 2.6. *Picea and Ambrosia* Temperature Distributions and Abundance Ratio. A) Scatter plot of pollen abundance vs. mean July temperature from the North American Pollen Atlas (Williams et al., 2006). Black dashed line indicates modern mean July temperature at Plan B Pond. Gray dotted line indicates threshold temperature at approximately 17.5°C. B) Ratio of *Picea:Ambrosia* from Plan B Pond.

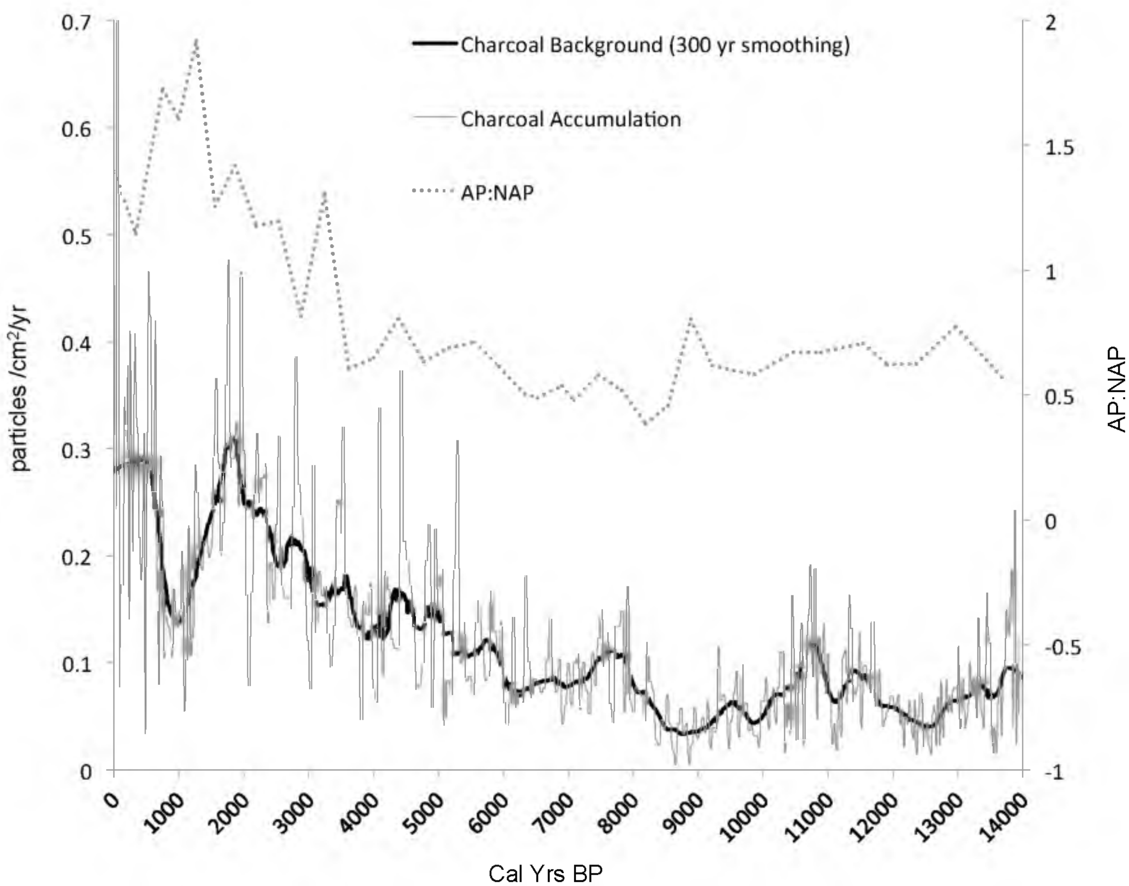

Figure 2.7. Plan B Pond AP:NAP Ratio and Charcoal Data. Top panel ratio of arboreal terrestrial pollen to non-arboreal pollen (AP:NAP). Bottom panel Plan B Pond charcoal data- charcoal accumulation rate (light gray line) and 300-year smoothed background (black line).

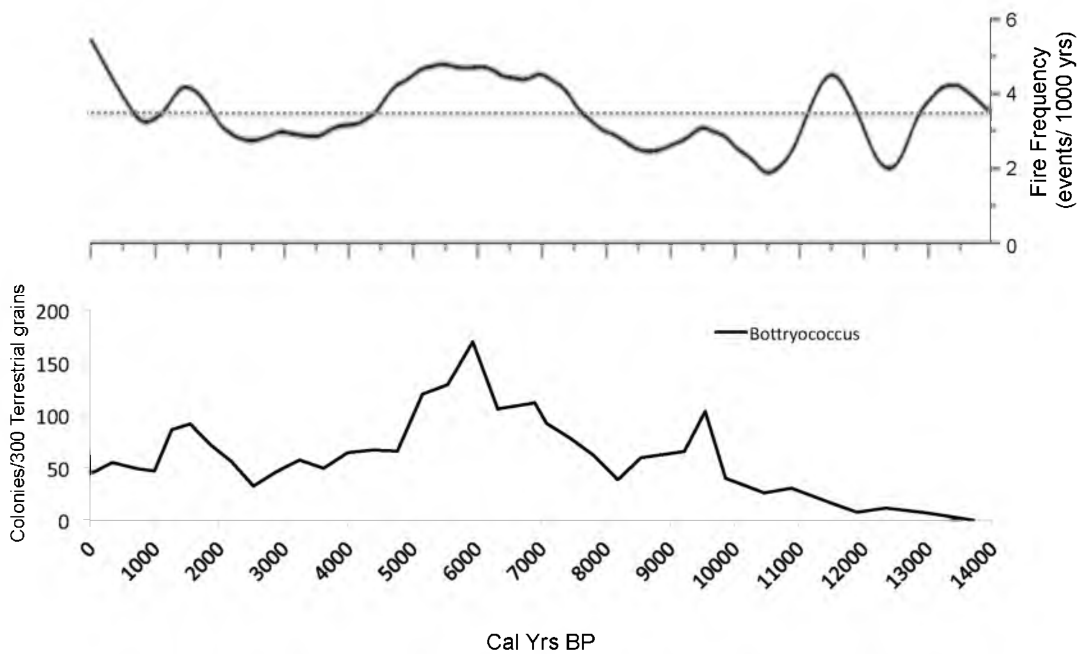

Figure 2.8. Fire Frequency and *Bottryococcus* Abundance. Top panel- reconstructed fire frequency from Figure 3. Bottom panel- *Botryococcus* abundance. High fire frequency and Bottryococcus abundance are interpreted to reflect periods of lower winter snowpack.

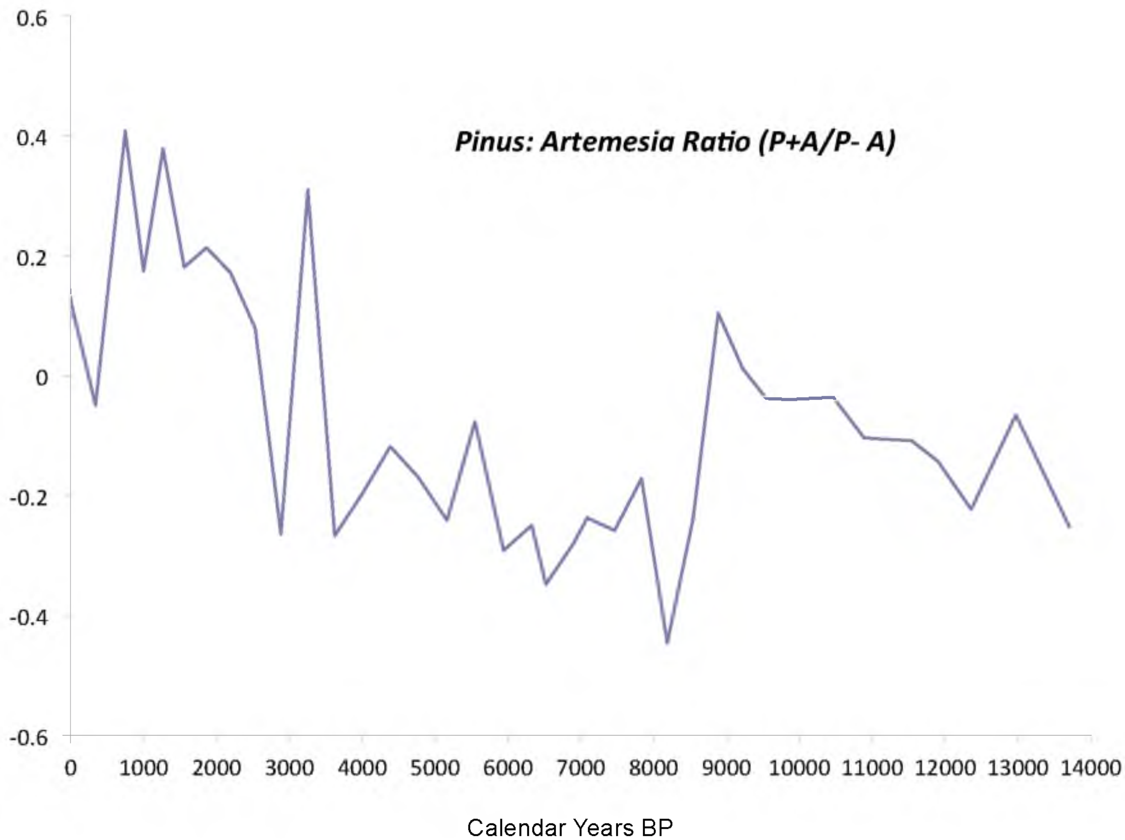

Figure 2.9. *Pinus/Artemisia* Ratio. Ratio is calculated as *Pinus-Artemisia/Pinus +Artemisia.* Higher values are interpreted to indicate lowering of lower treeline and effectively wetter conditions.

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CHAPTER 3

A SPELEOTHEM RECORD OF HOLOCENE PALEOCLIMATE OF THE NORTHERN WASATCH MOUNTAINS, SOUTHEAST IDAHO, USA

Introduction

Modern precipitation regimes across the Western United States (hereafter "West") exhibit a high degree of heterogeneity with regards to both amount and seasonality. This heterogeneity is attributed to the interplay between large-scale ocean/atmosphere patterns, mesoscale regional circulation patterns, and local topographic effects on climate (Mock, 1996; Shinker, 2010). Investigations into the relative effects of modern ocean/atmosphere teleconnection patterns on precipitation regimes across the West have demonstrated spatiotemporal correlations that explain much of the large-scale variability. For example, the association of different phases of El-Niño Southern Oscillation (ENSO) with predictable but opposite winter precipitation anomalies in the American Pacific Northwest and desert Southwest are well known (e.g., Cayan, 1996; Wise 2010). Similar precipitation anomaly patterns have been attributed to the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997; McCabe et al., 2004), and the magnitude of both ENSO and PDO related precipitation anomalies has been linked to the phase of yet another teleconnection pattern, the Atlantic Multidecadal Oscillation (AMO) (McCabe et al., 2004; Wise, 2010).

Studies of modern climate controls in the West show that the large-scale ocean/atmosphere teleconnections are associated with anomalous precipitation primarily in winter (e.g., Cayan et al., 2010; Wise, 2010). Elucidating the relative significance or role of these large-scale teleconnection patterns over time from paleoenvironmental records is often delimited by the lack of season-specific climate proxies, or having data that are sensitive to the wrong season for evaluating the influence of the winter-dominant teleconnection patterns. As a result, interpretations of paleoclimate data are often forced to use generalities such as "effectively wet" or "effectively dry" to describe mean annual hydroclimatic conditions, and any links with large-scale teleconnections patterns are more speculative than they might be if, for example, "effectively wet" could instead be stated definitively as "more winter precipitation".

In this context, we present results from a speleothem record from Minnetonka Cave in the Bear River Range (BRR), the northernmost extension of the Wasatch Mountains, Southeast Idaho, USA (Figure 3.1). The Minnetonka Cave speleothem data represent a record not of annual climate or moisture balance, but primarily of winter precipitation amount and temperature. As a result, the Minnetonka Cave record provides an opportunity to more accurately assess the long-term variability of Pacific influences on climate in the West through the Holocene.

Location and Site Characteristics

Minnetonka Cave is a public tour cave located in the eastern BRR (42.0875 °N, 111.519 °W), southeast Idaho. The entrance to the cave is at approximately 2320 m in elevation, and the cave network extends below surface terrain that reaches ~2600 m in elevation. Temperature inside the cave is monitored by the U.S. Forest Service and/or tour concessionaires and remains constant at ~4 °C year-round, approximating the mean annual air temperature outside the cave. Modern vegetation near the cave is a mix of open and closed canopy subalpine conifer forest, consisting mainly of *Abies lasiocarpa* (subalpine fir), *Picea engelmannii* (Engelmann spruce), *Pseudotsuga menziesii* (Douglas fir) and other subalpine taxa.

Data from a Natural Resources Conservation Service SNOTEL station located 8km southwest of the cave at 2460 m elevation (site #484- Franklin Basin) shows mean annual precipitation of ~1200mm/year. Precipitation occurs mainly as winter snow with more than 75% of the annual precipitation occurring from October through April, and only 15% of the annual precipitation occurring during the period of June through September.

Hydrogeologic studies in the area show that groundwater recharge in the BRR occurs primarily via snowmelt, with summer precipitation generally being inadequate to meet the needs of vegetation and to saturate the soils (Dion, 1969). Isotope data from springs and creeks in the eastern BRR show highly negative $\delta^{18}O$ and $\delta^{2}H$ values (average = -17.4‰ and -129.7‰, respectively) that fall on or very near the global meteoric water line (GMWL), regardless of late-spring or late-summer sampling dates (Bright, 2009). Similar isotopic results are reported for the Central Wasatch Mountains

(Mayo and Loucks, 1995), and across much of the Great Basin (Smith et al., 2002), demonstrating the dominance of high elevation winter precipitation in groundwater recharge in the region and a lack of significant evaporative effects prior to infiltration.

Groundwater in the eastern BRR is modern in age based on tritium measurements (Bright, 2009). Data from a nearby site in the northern Wasatch Mountains with similar geology shows mean groundwater residence times of 3-13 years (Rice and Spangler, 1999). Groundwater recharge in the BRR occurs via two modes, the first being relatively slow diffuse flow with a residence time of months to years, and the second mode being conduit flow with much higher flow rates and residence times of days or weeks (Rice and Spangler, 1999). Dye-pack test results from areas in the northern Wasatch Mountains indicate average groundwater flow rates up to 2700 feet/day during periods of high snowmelt when conduit flow is likely to be a dominant component of spring discharge (Rice and Spangler, 1999; Spangler, 2001).

Materials and Methods

Stalagmite MC08-1 is approximately 27 cm tall and was collected from Minnetonka Cave in August of 2008. Slow but active dripping on the sample was observed at the time of collection. The stalagmite was split vertically, and sampled along the growth axis for stable carbon and oxygen isotope analyses. Stable isotope samples were collected every 0.2 mm by incrementally advancing the stalagmite on a moveable stage under a 0.5 mm dental burr. The speleothem surface and drill bit were cleaned with ethanol prior to sampling, and brushed clean between samples. Samples were analyzed in the Department of Geosciences at the University of Massachusetts Amherst on a

Finnegan Delta XL mass spectrometer coupled to an automated carbonate preparation system. Stable isotope data are presented in standard δ-notation. Carbon and oxygen isotope data are reported relative to the Vienna Pee Dee Belemnite (VPDB) standard for carbonates, and Vienna Standard Mean Ocean Water (VSMOW) standard for water.

Seventeen $^{234}U/^{230}Th$ dates (Table 3.1) measured adjacent to the stable isotope sampling transect were used to build an age-depth model (Figure 3.2). All ages are reported as calendar years before present (BP), where present is defined as 1950 AD, for ease of comparison with radiocarbon dated paleoenvironmental records. Ages postdating 1950 AD are reported as negative. Ages reported from other records are also reported in calendar years BP, unless explicitly stated otherwise.

Dates show that the upper 152mm of MC08-1 are Holocene in age. Within the Holocene record, three growth hiatuses, or periods of extremely slow growth are inferred from sudden changes in age-depth relationships. Narrow (<1mm) white opaque layers can be seen in the speleothem at the depth of each of these steps in the age-depth plot. These layers are easily differentiated from the main stalagmite matrix and are being treated as hiatus layers. As such, our age model fits multiple polynomials to sections of the speleothem with uninterrupted growth, and extrapolates those growth curves up to the hiatus layers, rather than fitting curves across the steps in the age-depth plot.

Results

Speleothem Growth and Hiatuses

Growth hiatuses in MC08-1 occur from approximately 7400-9500 BP, 3800-1850 BP, and 1750-700 BP. The period of growth that occurred from 1850-1750 BP has only one date constraining it, so the rate of growth and the exact duration of growth are unknown. A constant growth rate of 0.0286 mm/yr was applied to this interval for the purposes of developing the age model. The applied growth rate approximates the average rate of growth over continuous growth intervals.

$\delta^{13}C$ Results

Minnetonka Cave $\delta^{13}C$ values mainly fluctuate in the range of -7 to -3‰ (Figure 3.3). Modern values are ~ -5.25‰, close to the mean Holocene value of ~5.1‰. Compared to modern values, the earliest part of the Minnetonka Cave record shows relatively high $\delta^{13}C$ values of -5 to -4‰. Beginning at ~10500 BP, $\delta^{13}C$ increases to the highest Early Holocene values of ~ -3‰ at ~9800 BP, then fall back to between -4‰ and -5‰ shortly thereafter. Values generally remain above modern, in the -5‰ to -4‰ range, when growth resumes following the first hiatus (9500-7400 BP), then drop to lower than modern values by just after 7200 BP. Values remain consistently less than -5.5‰ until ~ 6200 BP, except for two brief pulses at 6550 BP and 6450 BP when they rise to near modern. Between 6000 BP and 4000 BP $\delta^{13}C$ values are more variable, fluctuating around modern values, except for a prolonged period of lower values between ~4350 BP and 4050BP. Between 4000 BP and ~3800 BP $\delta^{13}C$ values peak in the record, just before the second growth hiatus (3800-1850 BP). When growth resumes briefly from ~1850 BP to 1750 BP, $\delta^{13}C$ values are for the most part higher than modern, in the range -5 to -4‰. From the time growth resumes after the final hiatus (1800-730 BP), values fluctuate around modern with three prolonged periods of lower values centered at 650 BP, 550 BP, and ~10 BP (1940 AD). There are also two prolonged periods of higher than modern $\delta^{13}C$ values from 500-450 BP, and 150-70 BP.

$\delta^{18}O$ Results

Compared to the $\delta^{13}C$ time series from Minnetonka Cave, the $\delta^{18}O$ data show a smaller range of variability overall; however, whereas $\delta^{13}C$ values from modern or near-modern samples approximate the long-term mean, the oxygen isotope results show that modern materials with $\delta^{18}O$ values ~ -14.5 ‰, are more than two standard deviations ($1\sigma = 0.418$) higher than the Holocene mean value of -15.45‰, not taking into consideration growth hiatus periods. Thus, although modern $\delta^{18}O$ values are not unprecedented in the record, they are not representative of Holocene conditions on the whole.

The earliest part of the Minnetonka record shows $\delta^{18}O$ values that generally fluctuate around a mean value of -15.4‰. The early Holocene data shows significant variability, but no obvious trend. The mid-Holocene, shows a consistent trend toward higher $\delta^{18}O$ values, starting out at values averaging ~ -16 ‰ at 7400 BP, and ending at values ~ -14.75‰ at 3800 BP. The period from 7400-6000 BP is the most extended period with values consistently lower than the long-term mean of -15.45‰. The represented portion of the Late Holocene record (<3800 BP) has $\delta^{18}O$ values that average -15.25‰. From 700-600 BP, 400-300 BP, and 50 BP-modern, $\delta^{18}O$ values remain consistently at or above that average, while the periods 600-400 BP and 300-100 BP are consistently below.

Discussion

Controls on Speleothem Growth

We postulate four possible explanations for the intermittent cessation of speleothem growth. The first possibility is a change in the drip water flow path that temporarily cut off flow to the drip site of our sample. This mechanism could have little or no relationship to climatic conditions. However, based on the temporal correlation of hiatus intervals with regional climate episodes recorded in other paleoclimate archives (discussion follows) it seems likely that hiatuses are related to changes in climatic conditions.

The second possible growth-halting mechanism is a lack of drip water due to significant reduction in winter precipitation. Cutting off cave drip water due to lack of adequate precipitation seems unlikely, as it would require snowpacks to be essentially nonexistent at elevations above 2300 m during hiatus intervals.

A more likely cause of halted growth might be significant reductions in temperature. If mean annual temperatures dropped enough to bring cave temperatures or below freezing, or to maintain permafrost, growth may have been halted by a lack of liquid water in the cave. A drop in mean annual temperature $\geq$ 4°C during the Holocene, the magnitude of change required under this scenario, seems unlikely given that Laabs et al. (2006) estimate that temperatures were probably only ~ 6-7°C cooler than modern in the Wasatch Mountains during full glacial conditions. However, early Holocene temperatures ~3°C cooler are suggested from paleoecological data in the GSL basin (Madsen et al., 2001). Moreover, model results from caves in Oregon suggest that a more modest drop in temperature, insufficient to form permafrost but cool enough to maintain frozen soils through the period of maximum snowmelt, could halt or slow speleothem growth by significantly increasing the ratio of surface runoff to infiltrating meltwater (Ersek et al., 2009).

Finally, speleothem growth may be halted due to inadequate saturation of $CaCO_3$ in the cave dripwater as a result of short residence time of water during wetter than average periods. If the spring meltwater pulse was consistently higher than average, groundwater and cave dripwater may be dominated by relatively dilute solutions derived primarily from conduit flow through the epikarst. Very short residence times from the point of infiltration to the drip site could prohibit the water from absorbing sufficient respired CO_2 to promote dissolution of the bedrock. This mechanism seems highly likely to be the cause for speleothem hiatuses since modern stream and spring water in the BRR already has low concentrations of Ca^{2+} and most other ions (Dean et al., 2007). A definitive conclusion about the cause of hiatuses in speleothem MC08-1 is not possible; however, it seems reasonable to conclude that hiatus intervals most likely correspond to anomalously wet conditions, or somewhat less likely, anomalously cold conditions, or both.

Controls on $\delta^{13}C$

When in isotopic equilibrium, speleothem $\delta^{13}C$ values are affected by open vs. closed system behavior (Hendy, 1971), the ratio of C_3 and C_4 plants overlying the site (Cerling, 1984; McDermott, 2004), vegetation density and soil/plant respiration rate (Quade et al. 1989, Baldini et al., 2005), drip interval (Mühlinghaus et al., 2007), and temperature (Emrich et al., 1970). At Minnetonka Cave we can rule out changes in C_4

plant abundance as a significant variable due to the high elevation and associated low summer minimum temperature (Teeri and Stowe, 1976), but the relative influence of other variables must be considered.

To evaluate primary controls on $\delta^{13}C$ at Minnetonka Cave, we started by estimating the expected $\delta^{13}C$ value for modern samples, assuming isotopic equilibrium. Studies from a similarly vegetated high elevation site in Colorado have reported $\delta^{13}C_{respired\ CO_2}$ values in the range of -26 to -24‰ (Bowling et al., 2009). If we assume a $\delta^{13}C$ enrichment of ~4.4‰ due to differential diffusion through soils (Cerling et al., 1991) and an equilibrium fractionation ($10^3 \ln \alpha_{Calcite-CO2gas}$) of ~ 11.5‰ at 4°C (Emrich et al., 1970), the speleothem $\delta^{13}C_{Calcite}$ values should be approximately -10 to -8‰. Thus, modern $\delta^{13}C$ values from MC08-1 are higher than expected for an open system in complete isotopic equilibrium.

Part of the enrichment may be a result of closed, rather than open system behavior (Hendy, 1971). In a closed system, dissolution of the limestone bedrock occurs after the groundwater solution is no longer in contact with soil gas having elevated CO_2 concentrations. This scenario results in a significant fraction of the carbon in the stalagmite coming from the limestone rather than just the soil CO_2 (Hendy, 1971).

However, higher than expected $\delta^{13}C$ values are common in temperate-zone cave systems and have been attributed to several different mechanisms (McDermott, 2004). The most prominent of these mechanisms include early degassing of drip waters and precipitation of carbonate in the epikarst or cave ceiling prior to dripping (Johnson et al., 2006), low soil respiration rates that allow a residual atmospheric isotope signal to be retained in the soil water dissolved inorganic carbon (DIC) (Cerling, 1984; Quade et al.

1989; Baldini et al., 2005), or short drip water residence times in the epikarst that prevents the water from reaching isotopic equilibrium with the soil CO_2 (Baker, 1997).

Any and all of these mechanisms are possible at Minnetonka Cave, but available evidence suggests that residence time is the most likely the primary explanation for the higher than expected $\delta^{13}C$ values. This conclusion is largely based on the relationship observed between historic GSL surface elevation data and Minnetonka $\delta^{13}C$ values (Figure 3.4), where high lake levels (wet conditions) correspond with higher $\delta^{13}C$ values and low lake levels (dry conditions) correspond with lower $\delta^{13}C$ values. McDermott (2004) suggests that such a relationship is likely indicative of short groundwater residence time in the zone of elevated soil CO_2.

The correlation of elevated speleothem $\delta^{13}C$ values with anomalously wet conditions is somewhat fortuitous in that there are few mechanisms that can lead to such a relationship. Most controls on speleothem $\delta^{13}C$ lead to more negative $\delta^{13}C$ values under wet conditions. For example, wetter conditions at the study site should lead to 1) higher overall soil respiration rates due to higher soil moisture conditions (Scott-Denton et al., 2003); 2) increased soil respiration in winter with increased snow cover (Monson et al., 2006; Bowling et al., 2009); 3) increased isotope discrimination by plants during photosynthesis (Bowling et al., 2002); 4) less early groundwater degassing and precipitation of calcite in the epikarst (Johnson, 2006); and 5) reduced exchange of soil CO_2 with the atmosphere because of increased snow cover (Solomon and Cerling, 1987). These phenomena would all have the effect of increasing soil CO_2 concentrations, or lowering the $\delta^{13}C$ value of the soil CO_2. In either case, the net effect should be more negative $\delta^{13}C$ values in cave calcite. Higher drip rates should also be associated with wet conditions, and should also lead to more negative values (Mühlinghaus et al., 2007). The fact that less (more) negative $\delta^{13}C$ values in MC08-1 correlate with well-documented wet (dry) intervals, suggests that none of these mechanism is the primary control on $\delta^{13}C$ variability at Minnetonka Cave.

Wetter than average conditions at Minnetonka Cave likely equate to higher than average snowpacks, as it is only the winter precipitation that infiltrates and recharges the groundwater today. On average, higher snowpacks should generate higher magnitude and more prolonged pulses of rapid conduit-type groundwater movement in the fractured karst bedrock overlying Minnetonka Cave. This fast moving water would have a very short residence time in the zone of active microbial and root respiration, and would likely draw atmosphere into the soils as it infiltrated (Solomon and Cerling, 1987). Together, these mechanisms would limit the amount of ^{13}C-depleted soil CO_2 that could be dissolved into the infiltrating water, and dilute it with isotopically heavy atmospheric CO_2.

We interpret the $\delta^{13}C$ signal from Minnetonka Cave as indicating relatively dry (wet) winters when values are more (less) negative. This interpretation is corroborated by the fact that $\delta^{234}U$ values from MC08-1 show considerable variability over time with a negative correlation to $\delta^{13}C$ values (r= 0.57), as would be expected if residence time was a common driver for both $\delta^{234}U$ and $\delta^{13}C$ values.

Controls on $\delta^{18}O$

In the absence of kinetic effects, the oxygen isotope composition of the stalagmite is a function of temperature inside the cave and the isotope composition of drip water

(Hendy, 1971). The temperature inside the cave is a reflection of mean annual temperature outside the cave. The isotopic composition of drip water is a function of condensation temperature of precipitation, distance travelled by the precipitating airmass, original vapor source composition, and any evaporative enrichment that may occur between condensation of water and the time when it drips on the stalagmite (Lachniet, 2009). The evaporative enrichment can occur within the atmosphere, in the soils, or on the cave ceiling. We have ignored the "amount effect" on precipitation $\delta^{18}O$, given that it typically is only a factor in low-latitude systems (Lachniet, 2009).

Using a modern average $\delta^{18}O$ value of -17.4±0.5‰ for groundwater in the BRR (Bright, 2009), a cave temperature of 4°C, and a fractionation ($10^3 \ln \alpha_{H_20\text{-}CaCO_3}$) of 32.67‰ (Kim and O'Neil, 1997), the $\delta^{18}O_{calcite}$ value that would be expected from calcite precipitated in isotopic equilibrium with dripwater is ~ -14.2‰. Thus, the measured $\delta^{18}O_{calcite}$ value (-14.3‰) of the most modern sample from MC08-1 suggests that precipitation of calcite is occurring in isotopic equilibrium with drip water and that kinetic fractionation resulting from evaporation or rapid degassing of drip waters inside the cave is negligible. This conclusion is supported by the fact that the correlation between $\delta^{18}O$ and $\delta^{13}C$ values along the growth axis of the sample is very weak (r^2=0.077). A high correlation between the two variables has been suggested as evidence of kinetic fractionation effects (Hendy, 1971).

Since streams and springs within the BRR have isotopic compositions that plot very near the GMWL ($\delta^2H=8.2*\delta^{18}O+13.2$; Bright, 2009), evaporative enrichment of precipitation in soils and the atmosphere is considered negligible, at least for the winter precipitation that is responsible for groundwater recharge. This conclusion is consistent with the findings in companion papers by Friedman et al. (2002) and Smith et al. (2002) for isotope compositions of precipitation and groundwater across the entire Great Basin.

It can be concluded that the dominant controls on speleothem $\delta^{18}O$ at Minnetonka Cave are a combination of temperature, storm-track, and original vapor source composition. Air mass trajectories that produce precipitation in the Great Basin are highly variable (Friedmann et al., 2002); however, due to the overwhelming dominance of winter precipitation in the dripwater composition at Minnetonka Cave, the variation in airmass trajectories and vapor source waters affecting precipitation in the BRR is more limited. We therefore suggest that temperature is likely the most dominant variable in the $\delta^{18}O$ signal, but probably not the only source of variability.

With the assumption that $\delta^{18}O$ variability during the Holocene was entirely due to temperature, we can estimate the range of temperature change. For mid- to high-latitudes, Dansgaard (1964) provides a precipitation $\delta^{18}O$ temperature dependence of 0.7‰/°C, based on the spatial relationship of mean annual temperature and precipitation $\delta^{18}O$, where cooler temperatures correspond to lower $\delta^{18}O$. Dansgaard's value has since been revised for most midlatitude locations to ~0.55‰/°C (Fricke and O'Neil, 1997; Kohn and Welker, 2005). However, data from two Great Basin sites (Winnemucca, NV and Cedar City, UT) show slopes of 0.67‰/°C and 0.62‰/°C, respectively, when the isotopic composition of precipitation is plotted against mean temperature during precipitation events, a more accurate measure of the $\delta^{18}O$/temperature relationship (Kohn and Welker, 2005). Therefore, it seems prudent to use a value ~ 0.65‰/°C for temperature estimates at Minnetonka Cave.

In addition to the temperature effect on $\delta^{18}O$ of precipitation, the temperature effect on the fractionation that occurs during the precipitation of $CaCO_3$ from drip waters must also be considered. This effect is ~ -0.23‰/°C at near-modern cave temperatures (Kim and O'Neil, 1997). Adding the two temperature effects together results in a total stalagmite $\delta^{18}O$ temperature effect of ~ 0.41‰/°C, or a total Holocene temperature range of ~ 6°C. The calculated temperature range is heavily biased toward winter temperatures because that is when the majority of precipitation events that generate the dripwater occur. However, the temperature inside the cave is a reflection of mean annual temperature. Since both the cave temperature and the temperature during precipitation events affects the $\delta^{18}O$ signal, the temperature trends inferred from Minnetonka Cave $\delta^{18}O$ are most reflective of winter temperatures, but partially influenced by mean annual temperatures.

Paleoclimate Interpretation

Data from the earliest Holocene (11200-9500 BP) part of the Minnetonka Cave record suggest cooler and significantly wetter conditions than present in the BRR, as indicated by relatively low $\delta^{18}O_{calcite}$ and high $\delta^{13}C_{calcite}$ values. On average, early Holocene $\delta^{18}O$ values are ~1‰ more negative than modern. By applying the 0.41‰/°C gradient from above, we can infer that the period was approximately 2.5°C cooler than modern. The interpretation of cooler than modern conditions is consistent with the paleoecological record from the GSL Basin which suggests conditions were wet and ~3°C cooler than present during the early Holocene (Madsen et al., 2001).

The interval represented by the earliest hiatus (9500-7400BP) in MC08-1 is interpreted to be much wetter and/or much colder than modern. The 9500-7400 BP interval corresponds very well with a period interpreted to be very wet at Bear Lake, as evidenced by a host of sedimentary, paleoecological, and geochemical data (Dean, 2006; Moser and Kimball, 2009; Smoot and Rosenbaum, 2009). Dean (2009) suggests that the period from ~9300-8300 BP was much wetter than present near Bear Lake due to increased delivery of winter Pacific-derived moisture.

The middle Holocene (7400-3800 BP) record from Minnetonka Cave shows a persistent warming trend from 7200 BP to just after 4000 BP. Based on our $\delta^{18}O$ data, an estimated warming of ~3°C is inferred from the linear trend. This magnitude of warming is consistent with model results that predict winter temperature anomalies of -1°to -2°C relative to pre-industrial temperatures over the Great Basin at 6000 BP, part way through our inferred warming trend (Diffenbaugh and Sloan, 2004). However, both model results and proxy data suggest that mean annual temperatures were warmer at ~ 6000 BP because warm summer temperature anomalies outweighed cooler winter temperatures (Diffenbaugh and Sloan, 2004).

The middle Holocene started out slightly wetter than modern, but underwent a rapid and sustained reduction in precipitation at ~7200 BP that persisted for almost a millennium, as indicated by a prolonged interval of more negative $\delta^{13}C$ values. A brief reprieve from the prolonged dry conditions occurred at 6550 and 6450 BP, when carbon isotope data suggests precipitation approached modern levels, briefly. Mensing et al. (2004) interpret the period 7530-6300 BP as being the driest period on record in the

Western Great Basin, and Great Salt Lake reconstructions suggest that the lake level was at its Holocene minima at approximately this time interval (Murchison, 1989).

From 6200 BP to 4400 BP, precipitation levels fluctuated at around modern values in the BRR. Then, from 4350 to just after 4100 BP, dry conditions again persisted at Minnetonka Cave. The ~ 4200 BP drought was not as severe as the 7200-6200 BP drought, either in duration or intensity, but it stands out as a significant and prolonged dry period. The 4200 BP drought has been recognized as an extreme dry period across much of the central U.S (Booth et al., 2005), and has been shown as a period of high fire activity in central Idaho (Nelson and Pierce, 2010).

Immediately following the period of prolonged drought at ~4200 BP, conditions at Minnetonka Cave appear to have rapidly transitioned to very wet winters by just after 4000 BP. The two Late-Holocene hiatus intervals (3800-1850 BP and 1750-700 BP) are again thought to represent wetter and/or colder conditions in the BRR. The hiatus intervals are consistent with the onset of neoglacial cooling documented in Western North America. For example, chironomid-based temperature reconstructions from Stella Lake, Great Basin National Park, suggest that July temperatures in the Great Basin were on average 1-1.5°C cooler than modern between approximately 3800 and 1500 BP, an then gradually warmed toward modern (Reinemann et al., 2009). GSL also began a major transgressive phase during this period of time, reaching its maximum level at approximately 2000 ^{14}C years BP (Murchison, 1989). The salinity of GSL is thought to have dropped during this transgressive phase to levels low enough to support large populations of Utah chub (*Gilia atraria*) at ~ 3400 ^{14}C yr BP, and again at ~1000 ^{14}C yr BP (Broughton et al., 2000). The two separate pulses of inferred low salinity correspond well with the two Late Holocene hiatus periods in MC08-1, that are also separated by a brief growth interval (i.e., relatively dry) from ~1850-1750 BP.

In general, the last 700 years of the Minnetonka Cave record are inferred to be drier than the preceding millennia. For most of the interval 700-520 BP, precipitation was much below average, and temperatures from 700-600 were probably slightly warmer than the Holocene average. Just after 600 BP temperatures cooled abruptly and oscillated at cooler than average temperatures until 400 BP. From ~500-450 BP precipitation increased, which in association with cooler temps may have led to another GSL highstand (Murchison, 1989). Temperatures warmed slightly from 400 to just before 200 BP when another abrupt cooling event occurred. Precipitation between 400 and 150 BP was about average compared to mean Holocene conditions, and then increased between ~150 and 70 BP. Precipitation was lower than average from 70 BP to the most recent part of the record, and temperatures show a consistent warming trend after ~130 BP that results in ~ 2.4°C of total winter season warming. Very similar to these results, the 10-year running mean of winter (Nov-April) seasonal temperature at Minnetonka Cave from PRISM climate data (Daly et al., 2008) shows ~2.2°C warming over the period 1900-2000 AD.

Pacific Influences

Interannual snowpack anomalies in the West have been shown to exhibit a consistent relationship with ENSO and PDO (Cayan, 1996), as well as the strength of the winter Pacific North American (PNA) pattern, characterized in part by the strength of the Pacific Aleutian Low (AL) pressure cell (Wallace and Gutzler, 1981). However, Minnetonka Cave lies very close to the ENSO/PDO precipitation dipole boundary (Wise, 2010), and some regional studies have suggested that the Pacific teleconnection influence may not be so clearly defined in this transitional area (e.g., Tingstad et al., 2010).

To explore the dynamics of our specific study site we used the NCEP/NCAR reanalysis dataset and historical APRIL 1 snow water equivalent (SWE) derived from snow course data to generate composite 700mb geopotential height anomalies (Figure 3.5), and 700mb mean wind vectors (Figure 3.6) for October-March of the ten highest and ten lowest April 1 SWE years between 1949 and 2010. We found that the winter 700mb geopotential height composite anomaly from high snowpack years in BRR indicated a consistently weakened AL/PNA pattern, whereas low snowpack years in the BRR showed a 700mb geopotential height composite anomaly indicative of a deepened or stronger AL/PNA pattern.

The 700mb mean wind vectors for low snowpack years show that the steering level winds have a more meridional flow over Western North America, with the effect of diverting Pacific storms to the north of the study site. In high snowpack years, a more zonal flow is maintained by the steering level winds. The observed 700mb wind vector anomalies are consistent with descriptions by Wallace and Gutzler (1981) of conditions associated with a strong vs. weak cold-season PNA pattern, and a correspondingly strong vs. weak AL.

Two independent paleorecords of the strength/position of the AL have been derived from $\delta^{18}O$ data from the Mt. Logan Ice Core (Fisher et al., 2008) and authogenic carbonate sediments in Jellybean Lake (Anderson et al., 2005), both located in the southwestern Yukon Territory, Canada. These two records are both interpreted primarily as a signal of meridional vs. zonal moisture trajectories, but are said to be reflective of the strength of the AL. If our interpretation of the Minnetonka Cave $\delta^{13}C$ record is correct, we should expect that periods inferred from the two Yukon records to be dominated by meridional flow and a strengthened AL would correspond to dry conditions at Minnetonka Cave (less negative $\delta^{13}C$). Conversely, periods dominated by zonal flow and a weaker AL would correspond to wetter conditions at Minnetonka Cave (more negative $\delta^{13}C$). A comparison of our Minnetonka Cave $\delta^{13}C$ data with the Jellybean Lake and Mt. Logan Ice Core records (Figure 3.7) generally supports our interpretation of the Minnetonka Cave $\delta^{13}C$ data and hiatus intervals. Hiatuses in the Minnetonka Record correspond to periods when the Jellybean Lake and/or the Mt. Logan records suggest a prolonged weakening of the AL. Even smaller scale fluctuations in the Yukon records generally correspond well with the Minnetonka $\delta^{13}C$ data.

The correlation between Jellybean Lake and Minnetonka Cave appears more consistent through time than the Mt.Logan/Minnetonka Cave correlation. This may be an artifact of the Mt. Logan Age model which shows significant compression of the ice core record prior to ~1000 years BP, and that has few constraints, especially prior to ~ 4000 BP (Fisher et al., 2008). However, local rather than regional climate variables at the two Yukon sites have also been invoked to explain periods of apparent disagreement between them, such as the period from 3000-2000 BP (Barron and Anderson, 2011). Considering this localized variability and the distance between the Yukon sites and Minnetonka Cave, we are not surprised by some inconsistencies between our record and the two Yukon records, even though the Yukon records generally do support our interpretation of the Minnetonka record and the dominant large-scale climatologies indicated by our interpretation.

Conclusions

The Minnetonka Cave isotope record is a winter season-specific recorder of precipitation and heavily winter-biased recorder of temperature. As such, correlations to the dominant Pacific teleconnection patterns and associated precipitation anomalies are likely more appropriate than many paleoclimate records in the West. This will help provide a better understanding of past variability in the frequency, spatial fingerprint, and intensity of Pacific teleconnection patterns.

Overall, the Minnetonka Cave isotope record shows a pattern of cooler and wetter than average early Holocene winters, and more variable mid-Holocene winter precipitation with prolonged dry periods from 7200-6200BP and ~ 4200BP. Winter temperatures over the mid-Holocene (7400-4000 BP) show a consistent warming trend of ~ 3°C. A period of consistently wetter conditions began ~4000 BP and generally persisted until ~700 BP, with the exception of a brief period of dryness at ~1850-1750 BP. Much of this wet interval corresponds to a period of cooler summer temperatures in the Great Basin. The last 700 years has been variable with respect to both precipitation and temperature. Notable climate excursions occurred at 700-500 BP when conditions were largely drier than average, followed by a wet pulse at ~ 475BP, and a consistent warming trend over the last ~130 years equivalent to > 2°C winter season temperature change.

Climatology composites for wet and dry winters in the BRR show that positive and negative precipitation anomalies in the BRR are largely associated with the relative strength of the AL/PNA pattern. The Minnetonka Cave $\delta^{13}C$ record, interpreted to be a good indicator of winter precipitation in the BRR, correlates reasonably well with two independent isotope-based reconstructions of the AL from the southern Yukon. This supports our interpretation of the isotope signal at Minnetonka Cave.

Table 3.1 . Uranium-series Data for Minnetonka Cave Stalagmite.

sample	^{238}U (ng/g)	^{232}Th (pg/g)	$^{230}Th/^{232}Th$ activity ratio	$^{230}Th/^{238}U$ activity ratio	measured $\delta^{234}U$ (‰)	initial $\delta^{234}U$ (‰)	uncorrected age (yrs BP)	corrected age (yrs BP)
MINN-2 mm	157.5 ± 0.4	499 ± 65	22 ± 3	0.0223 ± 0.0012	7921 ± 8	7923 ± 8	273 ± 15	**85 ± 98**
MINN-4 mm	154.6 ± 0.4	185 ± 35	50 ± 10	0.0197 ± 0.0008	8155 ± 7	7728 ± 6	235 ± 10	**166 ± 38**
MINN-9 mm	152.6 ± 0.4	91 ± 33	145 ± 52	0.0281 ± 0.0007	7721 ± 6	7728 ± 6	353 ± 9	**317 ± 24**
MINN-14.5 mm	154.4 ± 0.4	60 ± 35	330 ± 191	0.0417 ± 0.0008	8104 ± 6	8115 ± 6	501 ± 10	**479 ± 20**
MINN-18.5 mm	143.2 ± 0.5	81 ± 36	289 ± 129	0.0531 ± 0.0014	7991 ± 17	8005 ± 17	647 ± 18	**614 ± 28**
MINN-24 mm	155.6 ± 0.4	100 ± 32	290 ± 94	0.0609 ± 0.0008	8174 ± 8	8190 ± 8	727 ± 10	**690 ± 24**
MINN-29 mm	150.9 ± 0.4	110 ± 39	698 ± 247	0.1670 ± 0.0015	8790 ± 8	8836 ± 9	1875 ± 17	**1835 ± 30**
MINN-33.5 mm	152.6 ± 0.4	225 ± 34	648 ± 97	0.3119 ± 0.0018	7513 ± 6	7598 ± 7	4055 ± 24	**3965 ± 53**
MINN-44 mm	144.6 ± 0.4	163 ± 35	930 ± 201	0.3440 ± 0.0020	7114 ± 7	7207 ± 7	4702 ± 205	**4629 ± 49**
MINN-53 mm	149.0 ± 0.4	126 ± 34	1411 ± 377	0.3908 ± 0.0026	6916 ± 7	7023 ± 7	5491 ± 37	**5435 ± 49**
MINN-70 mm	107.8 ± 0.3	192 ± 46	920 ± 219	0.5366 ± 0.0033	8645 ± 14	8796 ± 15	6201 ± 41	**6104 ± 67**
MINN-83 mm	143.2 ± 0.5	159 ± 38	1377 ± 331	0.5009 ± 0.0034	7402 ± 13	7541 ± 13	6655 ± 47	**6586 ± 60**
MINN-103 mm	146.2 ± 0.4	185 ± 44	1305 ± 314	0.5405 ± 0.0032	7431 ± 10	7582 ± 10	7169 ± 44	**7091 ± 62**
MINN-113 mm	137.5 ± 0.4	453 ± 37	633 ± 52	0.6825 ± 0.0035	6771 ± 8	6959 ± 8	9911 ± 54	**9692 ± 123**
MINN-128 mm	131.8 ± 0.4	933 ± 35	301 ± 12	0.6980 ± 0.0053	6561 ± 9	6748 ± 10	10439 ± 84	**9954 ± 256**
MINN-140 mm	128.3 ± 0.4	1421 ± 38	221 ± 6	0.8007 ± 0.0044	7116 ± 13	7330 ± 15	11181 ± 66	**10476 ± 359**
MINN-151 mm	121.0 ± 0.3	808 ± 40	381 ± 19	0.8331 ± 0.0047	7115 ± 9	7345 ± 10	11652 ± 69	**11228 ± 223**
MINN-156 mm	182.7 ± 0.6	90 ± 51	35405 ± 20025	5.7177 ± 0.0259	9196 ± 17	11477 ± 27	78423 ± 496	**78402 ± 496**

All errors are absolute 2σ. Subsample powder sizes range from 70 to 140 mg. Initial $^{230}Th/^{232}Th$ atomic ratio used to correct ages is 0.00008 (activity ratio = 15) ± 50%. Yrs BP = years before present, where present = *AD* 2009.

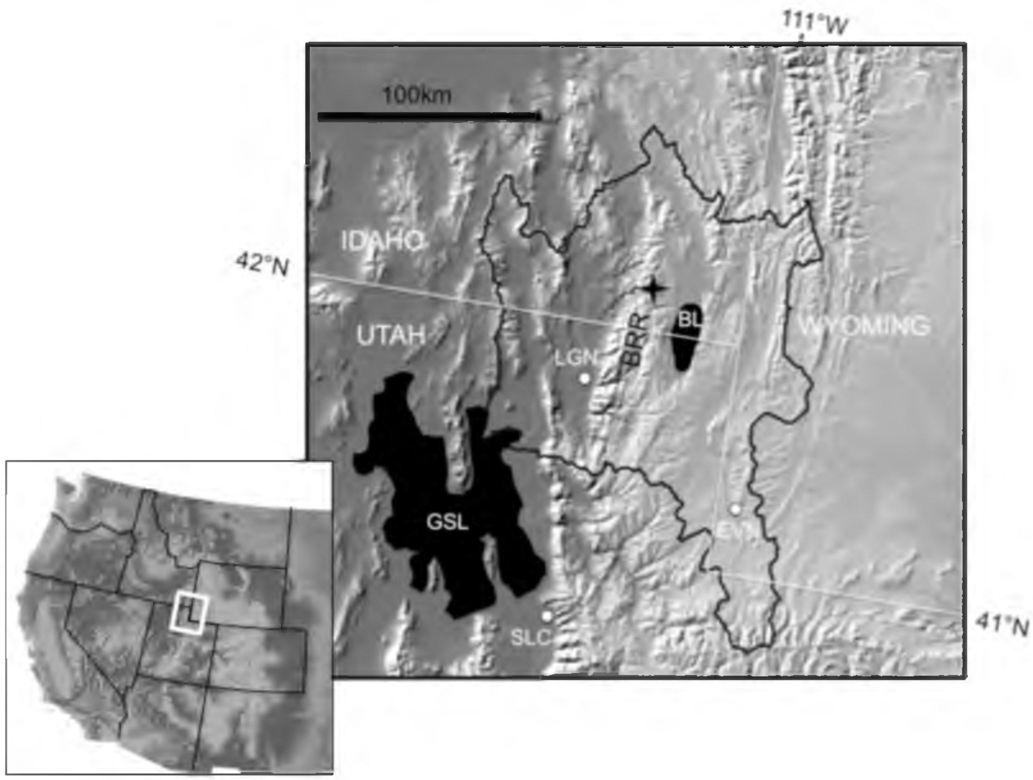

Figure 3.1. Minnetonka Cave Location Map. BRR=Bear River Range, GSL=Great Salt Lake, BL=Bear Lake, LGN=Logan, ENV=Evanston, SLC=Salt Lake City. Black Line indicates boundary of Bear River Basin. Star indicates approximate location of Minnetonka Cave.

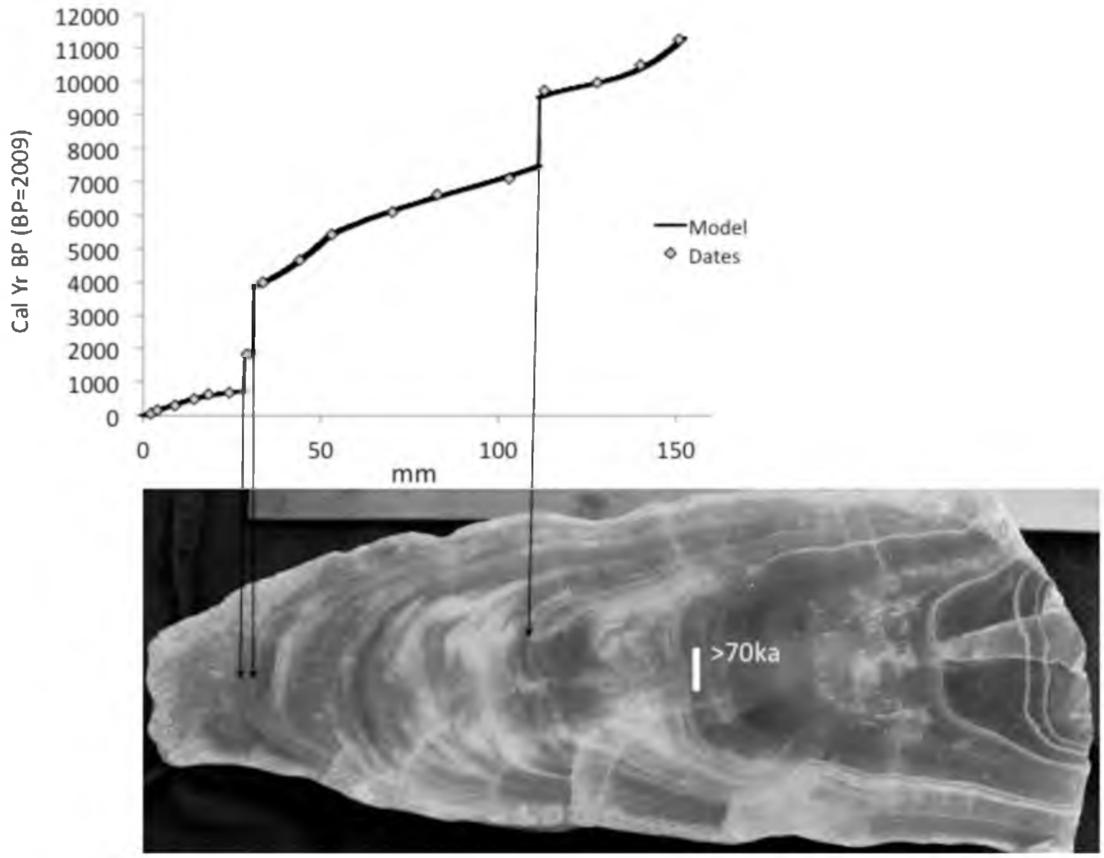

Figure 3.2. Speleothem U-Th Age Model. Black arrows show correspondence of hiatuses in age-depth model with opaque layers in speleothem.

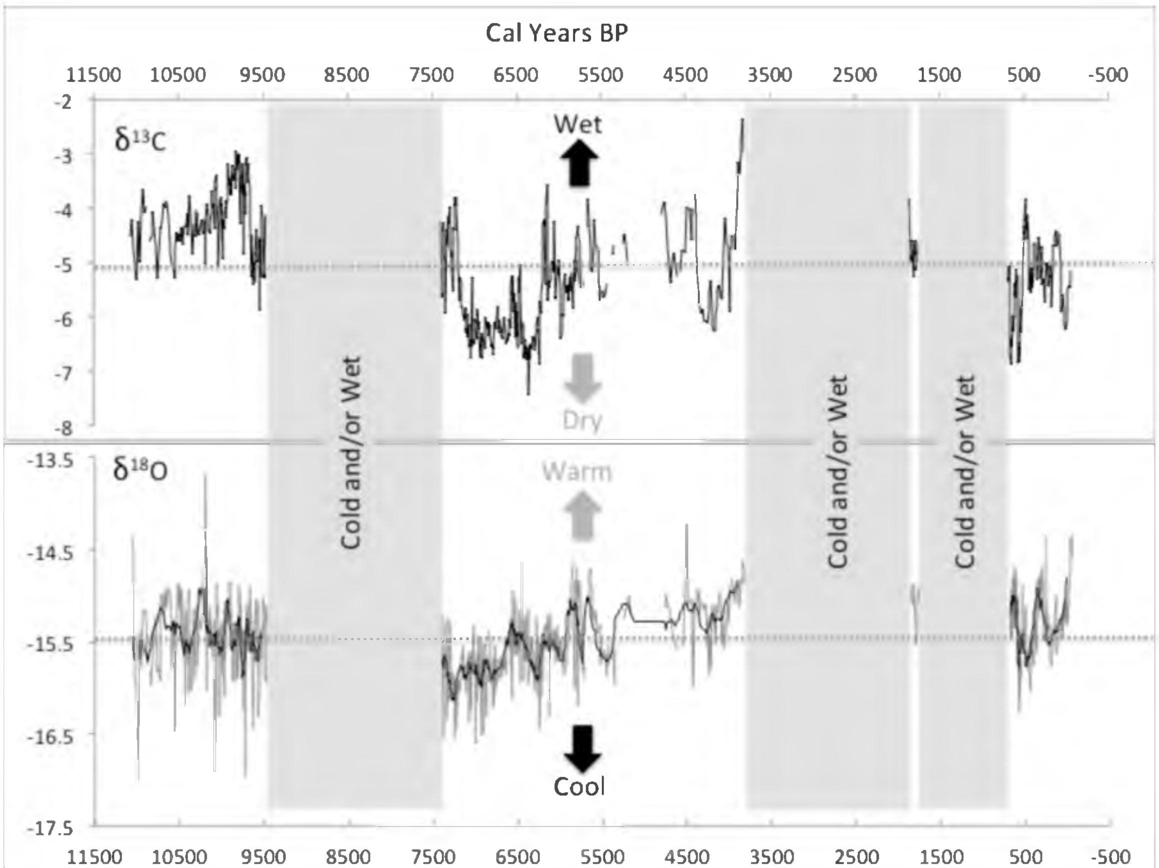

Figure 3.3. Stable Isotope Results. Horizontal dashed lines represent mean value for entire record.

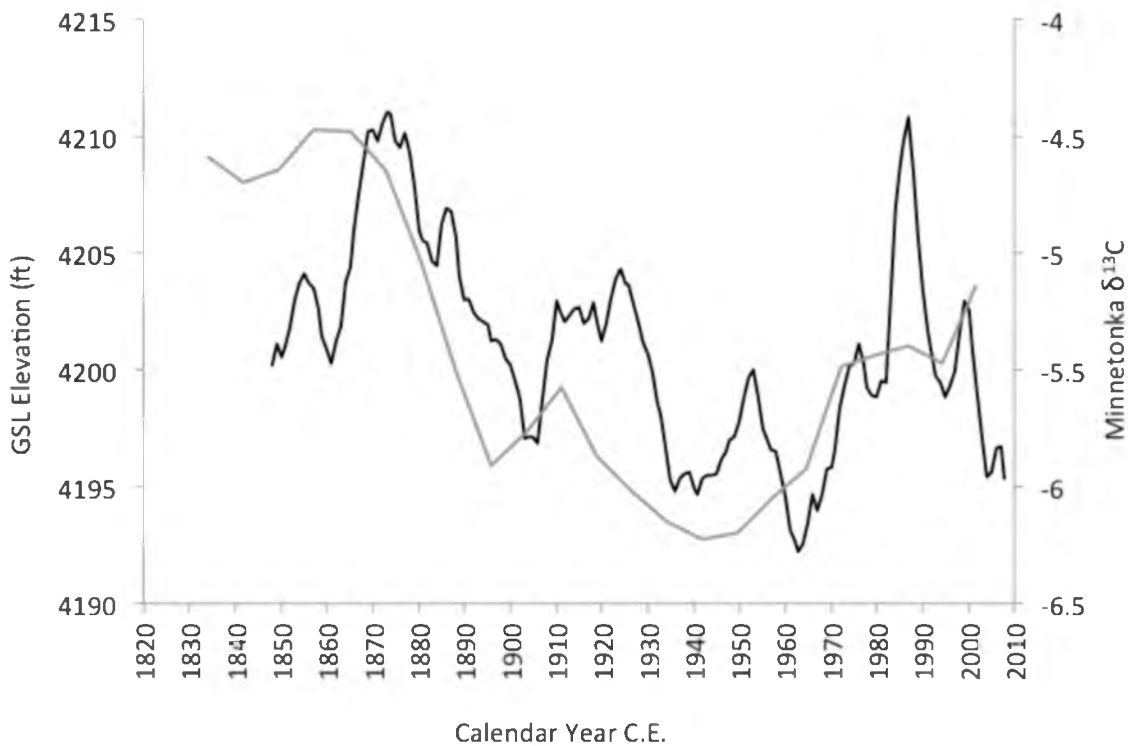

Figure 3.4. Minnetonka Cave $\delta^{13}C$ Data (gray) with Lake Surface Elevation Data From Great Salt Lake (black). Note that low lake levels correspond with more negative speleothem $\delta^{13}C$ values, and high lake levels correspond with less negative $\delta^{13}C$ values.

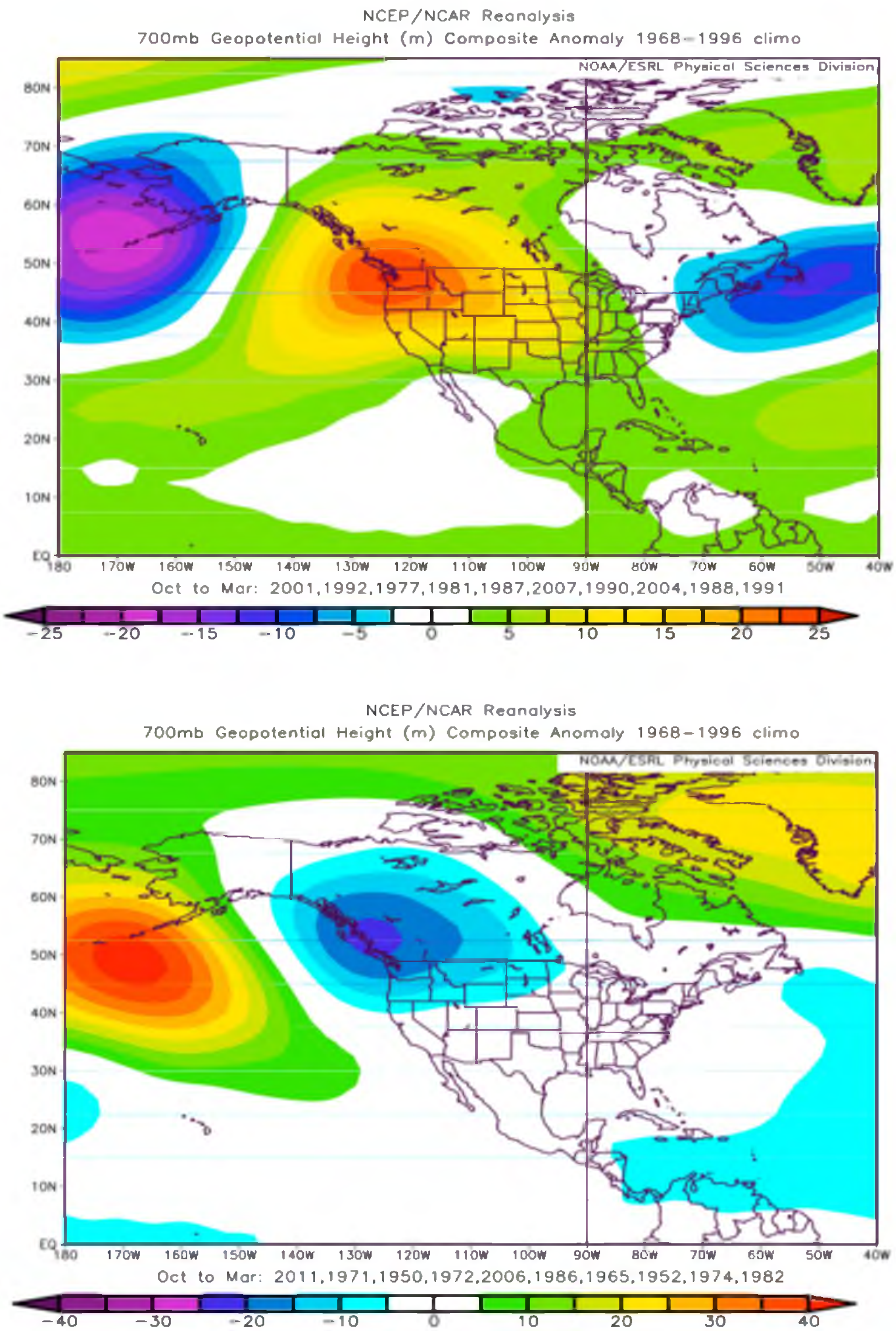

Figure 3.5. Composite October-April 700-mb Geopotential Height Anomalies. Top) low April 1 SWE (dry) years; Bottom) high April 1 SWE (wet) years.

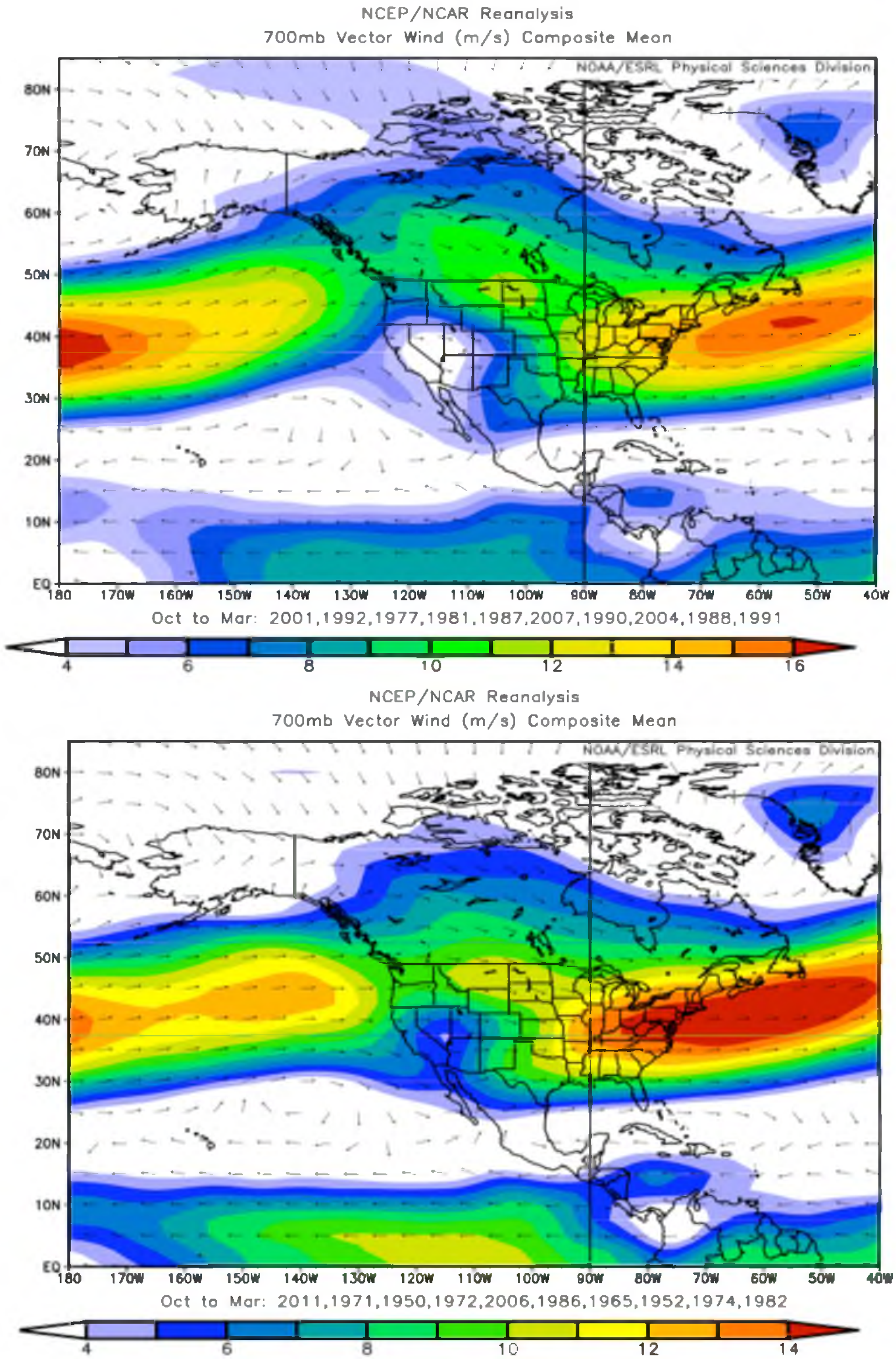

Figure 3.6. Composite October-April 700-mb Mean Vector Wind. Top) low April 1 SWE (dry) years; Bottom) high April 1 SWE (wet) years.

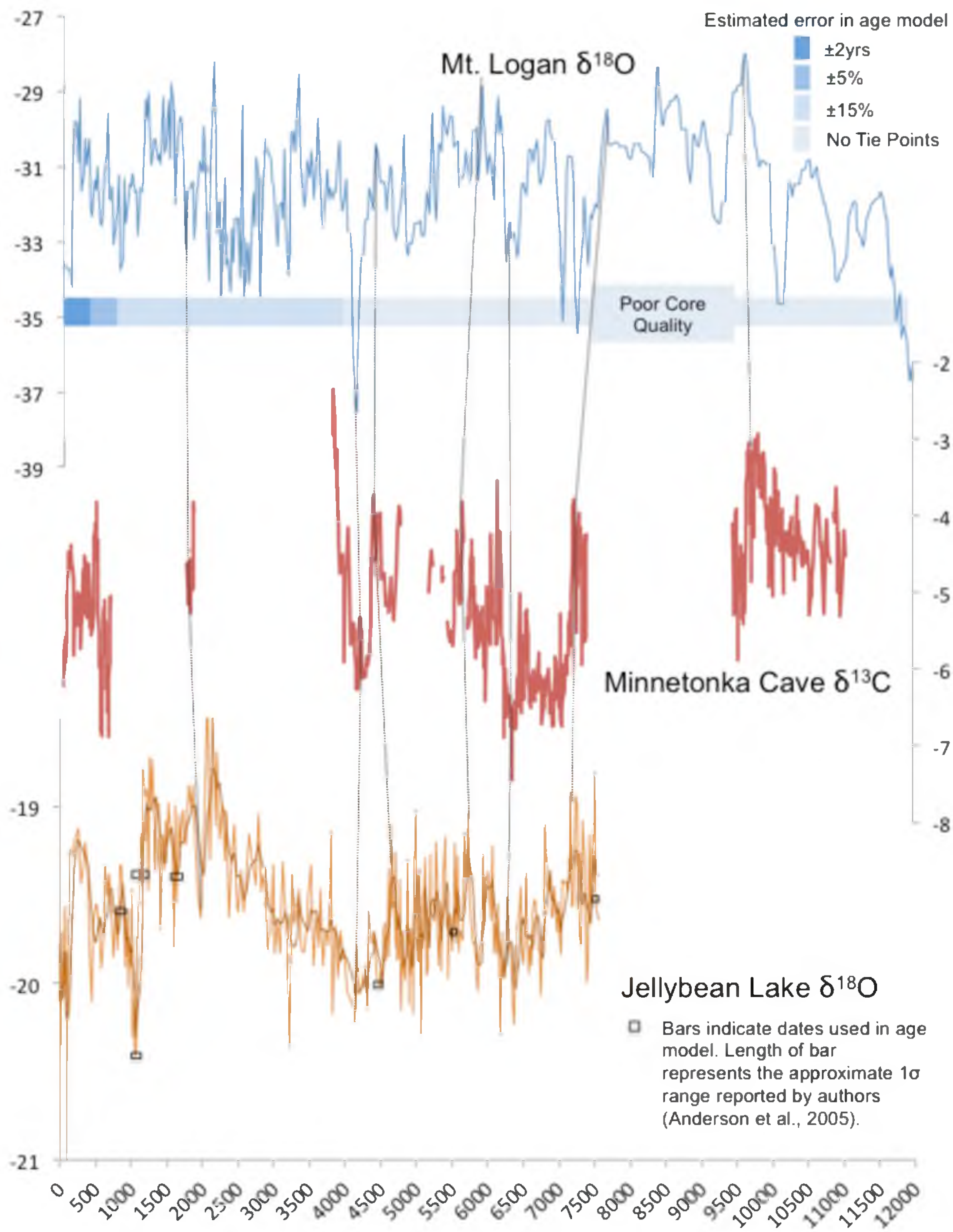

Figure 3.7. Minnetonka Cave $\delta^{13}C$ Record (Middle) Shown with Mt. Logan (top) and Jellybean Lake (bottom) $\delta^{18}O$ Records. Minnetonka Cave record interpreted to be dry (wet) when $\delta^{13}C$ values are more (less) negative. Mt. Logan (Fisher et al., 2008) and Jellybean Lake (Anderson et al., 2005) records are interpreted to indicate stronger (weaker) AL when $\delta^{18}O$ values are more (less) negative. Common X-Axis is in years BP. Estimates of age model error depicted by shaded bar below Mt. Logan record are given in Fisher et al., 2008.

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CHAPTER 4

COMPLEMENTARY SEASONAL BIAS IN TWO PALEOCLIMATE RECORDS FROM SOUTHEAST IDAHO: REFINING THE REGIONAL HYDROCLIMATIC HISTORY OF THE NORTHEAST GREAT BASIN

Introduction

The Great Basin of the Western United States is an expansive internally drained province extending from the southwest corner of Wyoming to the Sierra Nevada Mountains, and from central Oregon to southern California (Figure 4.1). The Great Basin has been the focus of many paleoclimate studies due to the presence of numerous pluvial lakes in the region intermittently throughout the Pleistocene (e.g., Benson et al., 1990; Currey, 1990; Oviatt, 1997; Davis, 1998). The greatest of these pluvial lakes, Lake Bonneville, occupied much of the northeast corner of the Great Basin until as recently as 12,000 ^{14}C years BP (Godsey, 2005). The changes in annual hydrologic balance indicated by the appearance and disappearance of the Great Basin pluvial lakes are obvious, and have been explained in large part by changes to atmospheric circulation caused by the extent of the North American ice sheets (Benson et al., 1995; Bartlein et al., 1998).

Changes in the hydrologic budget since the disappearance of the Great Basin Pluvial lakes have been both more subtle and more geographically heterogeneous than the glacial to interglacial pluvial cycles. In addition to large hemispheric or continental scale controls on the hydrologic budget, more regional ocean-atmosphere teleconnection patterns such as the El Niño-Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO), and Atlantic Multidecadal Oscillation (AMO), in conjunction with local topography, interact to create diverse patterns of hydrologic variability in time and space (Mock, 1996; McCabe et al., 2004; Wise, 2010; Shinker, 2011).

Understanding past changes in the hydrologic budget of the Western U.S. and the various geographic footprints of hydrologic variability is often complicated by the fact that many paleoclimate records have an inherent seasonal bias, especially when evaluating precipitation and drought (e.g., St. George et al., 2010; Power et al., 2011). Thus, inferences about mean annual hydrologic conditions must be made cautiously. An alternative, but equally problematic limitation to paleoclimate data, can actually be the lack of seasonal sensitivity in proxy data when local climate, especially precipitation, is strongly seasonal. In this case, proxy-inferred mean annual hydrologic budgets or effective moisture may mask significant seasonal anomalies, especially if multiple seasonal anomalies have offsetting hydrologic effects.

In addition to recognizing the relative importance of seasonal climate contributions to paleoclimate records, accounting for the magnitude of seasonal variability can also be critical to understanding differences in paleoenvironment across regions. The significance of seasonality has been shown to be particularly relevant in the early Holocene when summer insolation was much higher than modern and winter insolation was lower than modern (e.g., Whitlock and Bartlein; 1993; Davis, 1999; Broughton et al., 2000; Anderson, 2011).

One final complication for elucidating paleoclimate conditions on a seasonal basis, and therefore developing a more rigorous understanding of hydrologic budgets in the West, is that topography limits the geographic range of extrapolation from a single record. In some cases topography is a simple barrier between moist and dry air masses (Whitlock and Bartlein, 1993; Brunelle et al., 2005), in other cases the sensitivity of proxy indicators to different variables may change with elevation (Power et al., 2011). Thus, paleoclimate records in fairly close proximity may be sensitive to vastly different forcing mechanisms.

In the context of understanding the effects of enhanced Holocene seasonality and changing states of dominant ocean-atmosphere teleconnection patterns on hydrologic budgets in the northeast Great Basin, we review the regional paleoclimate literature and discuss the significance of two recently published paleoclimate archives from the northeast corner of the Great Basin. This review is presented in order to identify and understand seasonal bias in regional paleoclimate records so that it will not be mistakenly confused for spatial climate heterogeneity. The two new records are essentially co-located, providing control conditions for topography and elevation, allowing for the opportunity to utilize their complementary seasonal biases to better evaluate hydroclimatic conditions in the eastern Great Basin during the Holocene on a season specific basis. As will be discussed, isolation of seasonal variability is critical in an area that relies heavily on winter precipitation and has a highly negative summer moisture balance. The two new records include a speleothem stable isotope record from Minnetonka Cave (Chapter 3, this volume), and a lacustrine pollen and charcoal stratigraphy from Plan B Pond (Chapter 2, this volume).

The study sites lies in the southeast corner of Idaho, between the Great Salt Lake (GSL) and Bear Lake, and are situated in a primary hydrologic recharge area common to both lake basins (Figure 4.2). The GSL/Bonneville basin and Bear Lake have been well studied as paleoclimate archives (e.g., Murchison, 1989; Oviatt, 1997; Madsen et al., 2001; Rosenbaum and Kaufman, 2009), but the postglacial paleoclimate records from Bear Lake and GSL do not correlate well with one another despite their close proximity. The lack of correlation may just reflect the general hydroclimatic heterogeneity characteristic of the West. However, misinterpretations from both records may exist due to tectonic, geomorphic, or isostatic effects that have nothing to do with climate (e.g., Currey, 1990; Laabs and Kaufman, 2003). The Minnetonka Cave and Plan B Pond records refine the Bear Lake and GSL paleoclimate records by helping to isolate elements of the lake basins' water budgets, in addition to informing about broader regional paleoclimate patterns.

Setting and Significance

The Bear River Range (BRR), Utah and Idaho, is the northernmost extension of the Wasatch Mountains. The range lies centrally within the Bear River watershed (Figure 4.2) and acts as a primary hydrologic recharge area for the Bear River, the most significant river, by volume, flowing into the Great Salt Lake. For most of the Holocene when the Bear River was excluded from Bear Lake, runoff from the BRR is also thought to have accounted for ~99% of the inflow to Bear Lake (Dean et al., 2005 & 2007).

Based on SNOTEL data over the period 1983-2009, precipitation in the Bear River Range averaged more than 1200 mm/year, falling primarily as snow in the months

October through April. Summer precipitation is minimal with average July+August precipitation contributing less than 6% of the annual total. Mean July temperature is approximately 15.5°C, and mean January temperature is -6.5 °C.

Several large, once-glaciated canyons drain the east side of the Bear River Range, including (from north to south) Paris Canyon, Bloomington Canyon, and St. Charles Canyon. Plan B Pond is located at 2500 m elevation at the head of Bloomington Canyon. Minnetonka Cave is located at 2320 m elevation in St Charles Canyon. The linear distance between Plan B Pond and Minnetonka Cave is just over 8 km.

Explanation of Minnetonka Cave and Plan B Pond Records

Detailed analysis and discussion of the Minnetonka Cave stalagmite record is presented in Chapter 3 of this volume, and details of the Plan B Pond record can be found in Chapter 2 of this volume. Here we provide a brief narrative summarizing the records for general reference.

Minnetonka Cave

The Minnetonka Cave record includes stable carbon and oxygen isotope analyses from a single stalagmite. The record spans just over 11000 years, but includes three significant hiatuses (Figure 4.3). Hiatuses have been interpreted to indicate anomalously wet and/or cold conditions at Minnetonka Cave. Wet conditions are thought to lead to short groundwater water residence times in the active soil respiration zone, reducing the amount of dissolved CO_2 in groundwater, and preventing dissolution of carbonate bedrock. Thus, the calcium carbonate saturation index would be too low for calcite precipitation in the cave environment. Cold conditions could either lead to freezing conditions in the cave, thus inhibiting stalagmite growth, or could help maintain frozen soils long enough into the snowmelt season that much of the snowmelt would runoff the surface rather than infiltrating into the epikarst, thus starving the cave system of drip water.

Stable carbon isotope ratio variability in the Minnetonka Cave speleothem is interpreted to primarily reflect winter precipitation amounts. More negative $\delta^{13}C$ values reflect dry conditions or time periods when infiltrating water had a long residence time in zones of active soil respiration, allowing significant respired CO_2 from soils to be absorbed. More positive $\delta^{13}C$ values reflect wet conditions when respired CO_2 contributed less carbon to the dissolved inorganic carbon pool in cave drip waters due to limited residence time. Since groundwater recharge in the study region is snowmelt dominated, the Minnetonka $\delta^{13}C$ record is strongly winter precipitation biased.

Stable oxygen isotope ratios in the Minnetonka Cave record are interpreted to be primarily a recorder of temperature, also with a significant winter season bias. Other factors, including storm trajectory and vapor source changes over time are also likely to have had some effect on $\delta^{18}O$ values, but they are considered secondary controls compared to temperature.

Plan B Pond

The paleoenvironmental record from Plan B Pond is approximately 14000 years long. The record is based on stratigraphic changes in pollen, plant macrofossils, and macroscopic charcoal from a 3.97m-long lacustrine sediment core. The PBP record is also seasonally biased, but the season of bias depends on the proxy being evaluated. The pollen record is primarily sensitive to summer temperature with certain taxa being directly limited by temperature, and other taxa being limited by the effects of summer temperature on evapotranspiration. A plot of the ratio of select cool+moist to warm+dry (CM:WD) indicator taxa (warm/dry taxa = *Ambrosia*+ *Juniperus*+ Amaranthaceae +*Sarcobatus* +*Quercus*; cool/moist taxa= Asteraceae+ *Picea* +*Abies* +*Pseudotsuga*) shows that summer insolation is the dominant control on the vegetative composition through the Holocene (Figure 4.4), especially prior to 2.5ka BP (ka BP= 10^3 calendar years before present.

The PBP charcoal record (Figure 4.5) (methods in Chapter 2, this volume) also shows sensitivity to summer insolation in the background component, while the peaks component (fire frequency) is sensitive to winter precipitation. Background charcoal is primarily a function of fuel availability (Marlon et al., 2006), so it makes sense that the background component of charcoal would be correlated with the composition of vegetation at the site, specifically with the abundance of woody taxa.

Fire frequency across the Western U.S. is strongly linked to the length of the fire season. The length of the fire season is controlled by the timing of spring snowmelt in mountainous areas of the West, itself a function of winter precipitation and spring temperature (Westerling et al. 2006). Since spring temperature and preceding winter precipitation are generally not independent of one another (Cayan, 1996; Westerling et al., 2006) it is inferred that periods of low fire frequency indicate wet winters and cool springs, while periods with high fire frequency indicate dry winters and warm springs.

Bear River Range Records in a Regional Context

Late Glacial/Younger Dryas (14ka-11.5ka BP)

The earliest period of record from the BRR does not benefit from having overlapping speleothem and lake core records, as the Minnetonka Cave record does not begin until just before 11ka BP. Based on the Plan B Pond record, we infer that climate was warming from 14-13ka BP. Deposition of inorganic glacial flour in Plan B Pond from 14-13ka BP suggests small cirque glaciers were maintained in the upper reaches of the BRR well into the Late Glacial. However, *Picea, Abies*, and *Pinus flexilis* were locally present around Plan B pond by 13ka BP, based on the presence of macrofossils. Charcoal data indicate higher than average fire frequency from 14-13ka BP, indicative of dry winters. Between 13ka BP and 11.9ka BP, the CM:WD pollen ratio (Figure 4.4) suggests a brief but dramatic shift to cooler and effectively wetter conditions synchronous with the Younger Dryas cooling event. A high *Picea/Ambrosia* ratio, a more specific indicator of summer temperature (Lundeen and Brunelle, this volume), suggests that summers in particular were colder during this time. Charcoal data suggest a lower fire frequency, indicative of wetter than average Holocene winters from just after 13ka BP until 11.9ka BP, followed by a period of increasing fire frequency to 11.1ka BP.

Paleoecological records from the Bonneville basin indicate cooler summers during the Younger Dryas interval (~13-11.5ka BP) (Madsen et al., 2001) in agreement with BRR pollen assemblages. The BRR records generally concur with hydrologic indicators in the Bonneville Basin over the Bølling/Allerød and Younger Dryas chronozones, as well. Isotopic data and carbonate sedimentology/mineralogy from Blue Lake, near Wendover, Utah, indicate very low lake levels in the Bonneville basin during the Bølling/Allerød (14.7-13.3ka BP) with levels increasing to the Younger Dryas-age Gilbert highstand by 13.1ka BP. Water levels remained high until 11.6 ka BP, before falling sharply at 11.5ka BP (Benson et al., 2011). Similarly, Thompson (1992) reports that the Ruby Marshes in northeast Nevada may have desiccated sometime between approximately 17-13ka BP (15000-11000 ^{14}C yrs BP), but by ~12.8ka BP (10760 ^{14}C yrs BP) a deeper than modern freshwater lake was established, reaching the least saline conditions in the record shortly thereafter. The Ruby Marshes maintained deeper than modern water through the early Holocene, but a minor regressive period is suggested ~ 12.2 ka BP (~10400 ^{14}C yrs BP) (Thompson, 1992).

Hydrologic indicators from Bear Lake are not as consistently correlative with the BRR records as the Bonneville basin records during the Bølling/Allerød and Younger Dryas chronozones. For example, Bear Lake surface elevation reconstructions suggest lower than modern lake levels persisted at Bear Lake after 15ka BP, but lake levels fell further during the Younger Dryas to some of the lowest levels in the last 20,000 years, indicating very dry conditions (Smoot and Rosenbaum, 2009). On the other hand, hydrologic inputs based on diatom assemblages suggest that river inputs to Bear Lake were consistently higher from 13.8-7.6ka BP, than after 7.6ka BP (Moser and Kimball, 2009), and low $\delta^{18}O$ values of sedimentary calcite deposited in Bear Lake between 14.5-12ka BP are interpreted to indicate the dominance of abundant winter precipitation (Dean, 2009).

The discrepancy between reconstructed Bear Lake surface elevations and the BRR record may be explained by tectonic and geomorphic controls on Bear Lake's surface elevation and surface inflow from the BRR (Laabs and Kaufman, 2003). For example, reduced surface flow from the BRR to Bear Lake during the Younger Dryas and early Holocene may have occurred due to the presence of fault controlled northward-flowing fluvial channels at the eastern foot of the BRR. The channels are thought to have been occupied from ~13ka BP until at least 7.4ka BP (Reheis et al., 2009). Bright (2009) states that surface flow and shallow groundwater in the BRR was responsible for maintaining the hydrologic balance in Bear Lake prior to the Bear River diversion in the early 20th century, and that lake-marginal or sublacustral springs were likely of little significance. Thus, diversion of significant portions of the BRR surface flow away from Bear Lake, as implied by the existence of northward running channels, may have reduced river input and lowered lake level despite relatively wet conditions. Reheis et al. (2009) also suggest that renewed threshold incision by the Bear River ~12ka BP may have caused the lake regression.

Early Holocene (11.5-7.5ka BP)

Pollen data from Plan B Pond suggests rapidly warming summer temperatures after 11.5ka BP, peaking between 10.5-9.5ka BP, but remaining considerably warmer than modern throughout the early Holocene. Winter temperatures, as indicated by oxygen isotope ratios from Minnetonka Cave, suggest that winter temperatures from 11-9.5 ka BP were slightly cooler than modern, but about the same as the long-term Holocene average. The period 9.5-7.5ka BP is represented by a hiatus in the Minnetonka Cave record so definitive winter temperature inferences are difficult. However, colder conditions are one of the mechanisms thought to possibly be the cause of hiatuses, so a shift to cooler winter temperatures may be indicated from 9.5-7.5ka BP.

Charcoal data indicate wetter than average winters from 11.1-7.5ka BP, with the period from 10-9ka BP being slightly less wet, but still wetter than the long-term average. Carbon isotopes from Minnetonka Cave also indicate wet winter conditions from 10.5-9.7ka BP, followed by a brief dry excursion before the growth hiatus begins at just after 9.5ka BP. The hiatus from ~9.5-7.5ka BP is interpreted to indicate very wet winter conditions. Thus, both winter precipitation indicators from our BRR records suggest wet winter conditions for much of the early Holocene (Figure 4.6).

Warmer than modern early Holocene summers have been widely reported across the West, often based on evidence for upward movement of the upper treeline. For example, Munroe (2003) suggests treeline reached its highest elevation between 9.4ka BP and 8.1ka BP in the Uinta Mountains, and infer that July temperatures had to have been a minimum of 1°C warmer than modern. In the Medicine Bow Mountains of south central Wyoming, Mensing et al. (2011) report that treeline began moving upslope rapidly after 11.5ka BP, and remained above modern to ~9-6ka BP. Similar paleoecological results suggesting warm early Holocene summer temperatures are reported from the Colorado Rockies (Fall, 1997; Jiménez-Moreno et al., 2010), the White Mountains in central California (LaMarche, 1973); Yellowstone National Park (Huerta et al., 2009), and the Bitterroot Range in the Northern Rockies of Idaho and Montana (Brunelle et al., 2005).

Wet early Holocene winter conditions, as suggested by the BRR data, are also consistent with many records across the west. A significant $\delta^{18}O$ excursion to more negative values in Bear Lake carbonates is reported from 9ka-7ka BP, interpreted to be the indirect result of increased snowpack in the region (Dean et al., 2006). Bear Lake diatom assemblages also indicate generally higher river inputs prior to 7.6ka BP than any time after (Moser and Kimball, 2009). Higher than modern lake levels were maintained in the Ruby Marshes, Nevada until ~7.6ka BP (Thompson, 1992). Davis (1999) also reports high lake levels at Mono Lake from ~13.5ka -7.8ka BP (11600-7000 ^{14}C yrs BP), and reconstructed levels for Owens Lake indicate a consistent transgressive phase beginning ~12ka BP, reaching peak Holocene levels just before 7.5ka BP (Bacon et al., 2006). High levels of winter precipitation are also inferred from pollen data from Blacktail Pond, Yellowstone National Park (Huerta et al., 2009).

Also indicative of wet early Holocene winters in the West is evidence for early Holocene glacial advances in several locations, particularly in the region around the American Pacific Northwest. For example, early Holocene advances have been reported in southwest British Columbia (Menounos et al., 2004), on Mt. Baker, Washington (Thomas et al., 2000), and elsewhere in the Northern Cascades (Waitt et al., 1982). Some of these early Holocene advances are not uniformly accepted, and remain controversial (e.g., Reasoner et al., 2001; Davis et al., 2009).

In contrast to the records indicating wet early Holocene conditions, many records, particularly paleobotanical records, suggest that the early Holocene was a period of peak aridity in the West, at least in areas that are strongly dominated by winter precipitation, and that lack significant moisture inputs from summer convective storms (Whitlock and Bartlein, 1993; Brunelle et al., 2005). For example, in the West Range of west-central Idaho, Doerner and Carrera (1999) report that the interval ~10.5ka-7.5ka BP (9300-6700 14C yr BP) was the hottest/driest in the Holocene. Beiswenger (1991) reports xeric conditions at Greys Lake, eastern Idaho from ~11.5ka-8ka BP (10000-7100 ^{14}C yr BP), with a peak in xeric conditions ~9.2ka BP (8200 ^{14}C yr BP). At Blue Lake, in the

Bonneville basin, the early Holocene is described as a period of warming and drying, with lake desiccation by ~8ka BP, associated with peak abundances of xeric taxa such as Amaranthaceae and minimum abundances of conifers between 8ka-9ka BP (Louderback and Rhode, 2009).

The Early Holocene represents a period when enhanced seasonality significantly complicates the paleoenvironmental record. Much work has been published investigating the role and spatiotemporal footprint of enhanced summer convective precipitation in the early Holocene (e.g., Whitlock and Bartlein, 1993, Brunelle et al., 2005), with the conclusion that sites that have a unimodal winter peak in precipitation were warmer and drier in the early Holocene, whereas sites that experience significant summer convective precipitation were warmer and wetter at that time. However, this conclusion reflects mean annual or growing season effective moisture, and does not specifically address winter precipitation, which is the dominant climatic variable affected by major Pacific ocean/atmosphere teleconnection patterns.

Based on pollen data alone, the traditional interpretation for early Holocene conditions in the BRR would also be hot and dry. The relatively high abundance of xeric taxa (e.g., Amaranthaceae, *Ambrosia*, *Sarcobatus*) and low abundance of mesic taxa (e.g., *Picea* and *Abies*) do suggest effectively dry conditions. Significant presence of *Pinus flexilis* and Pse*udotsuga menziesii* at Plan B Pond during the early Holocene, as indicated by macrofossils, is also indicative of effectively drier conditions. However, charcoal data from Plan B pond and isotope data from Minnetonka Cave suggest that snowpacks in the early Holocene were above average.

Collectively, data from BRR, suggest that wet winter conditions dominated the early Holocene, but were more than offset by very warm and dry summers, in terms of the relative effects on effective moisture. Broughton et al. (2008) reached similar conclusions in the Bonneville Basin, based on model results and faunal assemblages. As a result of the offsetting seasonal anomalies, effectively dry conditions are suggested by paleoecological indicators in the BRR, much the same as other paleoecological records from "summer-dry" sites (á la Whitlock and Bartlein, 1993) across the Intermountain West. However, to call the early Holocene "dry" in the context of understanding long term changes in the hydroclimatology of the Great Basin and the west is misleading, given that that winter precipitation was above average. Recognition of this distinction is critical for long-term reconstructions of Pacific teleconnection patterns and their effects over time, as their effects are most pronounced in winter precipitation anomalies (Wallace and Gutzler, 1981; Cayan, 1996; Wise, 2010).

Elevation is also an important factor when considering early Holocene records and inferring climatic conditions (Power et al., 2011). The magnitude of the difference between effects of increased winter precipitation and increased summer temperatures on effective moisture would be greater at lower elevations than high elevations due to orographic effects on winter precipitation. Thus, lower elevation sites such as Blue Lake, Nevada (Louderback and Rhode, 2008) may be severely summer temperature-biased in their effective moisture reconstructions in the early Holocene.

Middle Holocene (7.5ka-4.5ka BP)

Based on pollen PBP pollen ratios, the BRR continued to experience warmer than modern summer temperatures in the middle Holocene, with summer cooling taking place over the entire period, but particularly after ~6.2ka BP. In contrast, Minnetonka Cave $\delta^{18}O$ data indicate winter temperatures were much cooler than either modern conditions or long-term averages at 7.5 ka BP, but exhibited a steady warming trend that continued until after 4ka BP. From 7.5ka-4.5ka BP fire frequency was higher than average suggesting dry winter conditions. Dry middle Holocene winter conditions are also indicated by low $\delta^{13}C$ values from Minnetonka Cave, especially between 7.2ka-6.1ka BP.

With prolonged intervals of dry winters, and summer temperatures still well above modern, especially before ~6ka BP, the middle Holocene is inferred to be the driest period in the BRR, hydrologically. An overall peak in *Botryococcus* abundance just after 6ka BP, an indication of consistent low lake level, indicates the relative magnitude of mid-Holocene hydrologic drought (Chapter 2 this volume). However, relative pollen abundances indicate a consistent trend toward effectively wetter conditions, especially as compared to the early Holocene.

Warmer than modern mid-Holocene summer temperatures inferred from BRR data are consistent with records from the nearby Uinta Mountains, where Carson et al. (2007) suggest larger than average floods from 8ka-5ka BP were driven, in part, by warmer than modern summer temperatures, and Munroe (2003) reports warmer than modern summer temperatures from 9.5ka BP until at least 5.4ka BP. Similar findings are reported by Madsen and Curry (1979) from pollen ratios from Snowbird Bog, Wasatch Mountains where relatively warm temperatures are inferred up to ~6ka BP. Relatively warm (comparable to modern) mid-Holocene summer temperatures are also reported from Great Basin National Park, east central Nevada, based on a midge-based reconstruction (Reinemann, 2009).

In addition to being warm, much of the Intermountain West is interpreted to have been quite dry during the mid-Holocene, especially between 7ka-4ka BP (e.g., Shuman et al., 2009). Lake level reconstructions from Owens Lake, California indicate a prolonged lowstand during this period (Bacon et al., 2006). Lake levels in Pyramid Lake, Western NV, were also low, particularly ~7.6ka-6.5ka BP (Benson et al., 2002). Ruby Marshes in eastern Nevada had low water levels from ~7.5ka-5.3ka BP (Thompson, 1992), and Bear Lake diatom data indicate dry conditions from 7.6ka-5.8ka BP. Lake surface elevations in Bear Lake were also consistently 15m or more below modern from ~7ka-4.5ka BP (Smoot and Rosenbaum, 2009).

Lake level reconstructions for Great Salt Lake show that the lake was near desiccation from ~7.5ka-6.8 ka BP (6700-6000 ^{14}C yrs BP) rebounded briefly just after 6.8ka BP, and then fell back to lower than modern levels until ~4ka BP (Murchison, 1989). Model simulated moisture index in the Bonneville Basin also reaches its lowest Holocene values from ~8ka-4.5ka BP(Broughton et al., 2008)

Whereas the early Holocene was dominated by wet winters, but often interpreted to be effectively dry by paleoecological indicators, the middle Holocene is often interpreted to be effectively moister than the early Holocene, but actually had less winter precipitation. This disparity is likely attributable to declining summer insolation and slightly cooler summer temperatures than were seen in the early Holocene, which offset precipitation deficits and allowed growing season moisture to remain higher than during the early Holocene.

Late Holocene (4.5ka BP-Modern)

After 4.5ka-4ka BP, climate conditions in the BRR began shifting to a regime cooler and wetter than the mid-Holocene. Pollen data from Plan B Pond show markedly increased contributions from conifers, and declining inputs from *Artemisia* and Amaranthaceae. The AP:NAP ratio increases, indicating a shift to more closed-canopy conditions, and *Botryococcus* abundance declines from its mid-Holocene peak, especially prior to ~2.5ka BP. The CM:WD pollen ratio suggest rapidly cooling summers after ~4ka BP and significantly increased climate variability after ~2.5ka BP.

Charcoal data from Plan B Pond suggest generally wet winter conditions in the late Holocene, with below average fire frequency from ~4.5ka-1.8 ka BP and from 1.2ka-0.6ka BP. Minnetonka Cave $\delta^{13}C$ data show a relatively intense dry period from 4.5ka-4 ka BP to start the late Holocene, but $\delta^{13}C$ values indicate a shift toward very wet winters just after 4ka BP, before initiation of a growth hiatus that lasts until ~1.9ka BP, interpreted to reflect prolonged wet and or cold winter conditions. Growth resumes for a brief duration at ~1.9ka BP, indicating a temporary shift to drier conditions, then ceases again from ~1.8ka-0.8ka BP. Continuous growth from 0.8ka BP-modern suggest that most of the last millennium has been the driest period in the late Holocene, in terms of winter precipitation.

Oxygen isotope data from Minnetonka Cave suggest, for the most part, warmer winter temperatures in the late Holocene than in the mid Holocene. There is continuation of the mid Holocene warming trend from 4.5 ka BP up to the hiatus at 3.8 ka BP. The $\delta^{18}O$ values remain relatively high (warm temperatures) during the brief period of growth at 1.9ka-1.8ka BP, and then oscillate at mostly above average values for the last 0.8ka, with the exception of a century-long cool period ~0.5ka BP. The $\delta^{18}O$ values over the last 130 years show a steady warming trend, consistent with instrumental records.

The late Holocene BRR records are generally quite consistent with other regional paleoclimate reconstructions. Hydrologically, the Great Basin and much of the West in general have been shown to be quite wet in the late Holocene, especially from ~4ka-2ka BP. The GSL was in a transgressive phase from ~4ka-2ka BP (Murchison, 1989), and faunal evidence from Homestead Cave in the GSL basin indicates that salinity in the GSL was low enough from ~3.7ka-3.6ka BPand ~1.3ka-0.9ka BP to support populations of Utah chub (*Gilia atraria*) (Broughton et al., 2000). This correlates well with reconstructions from Owens Lake, California that indicate rapid lake transgression from ~4.0ka-3.5ka BP, and maintenance of a lake levels above historical high levels until ~1ka BP (Bacon et al., 2006). Isotopic data from Pyramid Lake, Nevada show that following a period of mid-Holocene aridity, Lake Tahoe rose to its sill depth and began spilling into Pyramid Lake by 3.1ka BP, supporting a return to wetter conditions in the Late Holocene (Benson et al., 2002). Increased precipitation in the northern Sierra Nevada, the headwaters of Lake Tahoe and Pyramid Lake, is also supported by data from the San Francisco Bay estuary where increased river inputs from the Sacramento and San Joaquin Rivers are inferred from 3.8ka-2ka BP based on paleosalinity indicators (Goman and Wells, 2000).

Bear Lake provides some mixed signals in the late Holocene. Although there is significant variability, the lake level reconstruction from Bear Lake is consistently higher in the late Holocene than the mid Holocene (Smoot and Rosenbaum, 2009), suggesting wetter conditions, overall. Pollen data from Bear Lake also suggest cooler, more mesic conditions after ~3.8ka-3.6ka BP (Doner, 2009). However, diatom data are interpreted to indicate dry and increasingly evaporative conditions from 4.3ka-3.8ka BP and after 2.9ka BP (Moser and Kimball, 2009). These conflicting interpretations are difficult to reconcile, but may be indicative that non-climatic factors are at play, or may be related to limited sampling resolutions in a more variable climatic period.

Other evidence from the region supports a shift to cooler conditions after ~4.5ka BP. The midge-based July temperature reconstruction from Great Basin National Park, Nevada, suggests summer temperatures were consistently cooler from ~4.0ka -1.5 ka BP than the two and a half millennia prior, or any time since (Reinemann, 2009). After ~4.5ka BP, cooler summers and wetter winters than during the mid-Holoceneare also inferred from pollen records in northern Colorado (Jiménez-Moreno et al., 2010), southern Wyoming (Mensing et al., 2011), Western Idaho (Doerner and Carerra, 2001), and northwestern Nevada (Thompson, 1992).

Discussion

Based on the correlation of different elements of the two BRR records with paleoenvironmental records from around the region, we can make some generalizations about the relative seasonal biases of different types of proxy records. These seasonal biases in many cases have been recognized and specified by other authors (e.g., Davis, 1999), but the importance of recognizing which types of records are comparable, and which are not, justifies repeating the conversation.

For the most part, lake level reconstructions seem to correlate with BRR records of winter precipitation throughout the Holocene. For example, the reconstructed lake levels from Owens Lake, California show remarkable correlation with the BRR $\delta^{13}C$ and fire frequency data (Figure 4.7). Obviously, there are some cases where non-climatic controls such as tectonics and fluvial dynamics may alter threshold elevations, change river inputs, or otherwise affect lake level reconstructions. Based on the intermittent correspondence between Bear Lake and the BRR records, it is apparent that such complications hinder climate inferences from the Bear Lake record over certain time intervals, a point recognized and discussed at length by Laabs and Kaufman (2003).

In the case of GSL, rather than tectonic or geomorphic complications, surface area/volume may be a significant factor explaining discordance between reconstructed lake level and the BRR paleoclimate record. The GSL has approximately 233 km^2 of surface area for every km^3 of volume. By comparison, Bear Lake and Lake Tahoe have approximately 35 km^2 and 3.3 km^2 surface area per km^3 of volume, respectively. Thus, evaporative fluxes from GSL would be expected to be much higher.

For much of the Holocene, especially the latter half, the reconstructed GSL surface elevation matches well with the winter precipitation records from the BRR (Figure 4.7). However, during the period of peak insolation in the early Holocene, when many other lake records in the West, including Bear Lake, record high water and BRR records indicate large snowpacks, GSL shows a delayed transgression of modest magnitude between ~8.8ka-7.8ka BP, based on the most accepted lake level curve (e.g., Murchison, 1989). Patrickson et al. (2010), have proposed a possible highstand during this time period of much greater magnitude, but this highstand is not widely accepted.

Given the bathymetry of the GSL basin, high surface area/volume ratios would be maintained for the entire range of Holocene hydrologic variability, making the lake particularly susceptible to evaporative losses during the early Holocene summer insolation peak, but also limiting surface elevation change, even with large changes in hydrologic budget. The limitations of using lake surface elevation as an indicator of hydrologic balance are discussed by Benson and Paillet (1989), for the Lahontan Basin, Nevada. The authors suggest that for comparative purposes between lake basins of different bathymetry, lake surface area is a much better metric to evaluate climatic effects on water budgets. However, it is not common that complete shorelines can be mapped to reconstruct past changes in surface area; therefore, recognition and consideration of the limits of lake surface elevation data and lake-specific bathymetry may have to suffice.

Instead of a winter precipitation bias, it is clear from Plan B Pond pollen spectra that BRR paleobotanical records are more sensitive to summer insolation and summer temperatures than precipitation variability through most of the Holocene. Based on the number of pollen records that characterize the early Holocene as extremely xeric, despite widespread evidence for high lake levels and abundant winter snow, it seems likely that many (most?) pollen records carry the same bias. This is consistent with the findings of Power et al. (2011) who state that they also found vegetation change at high elevation sites in the Northern Rockies to be responsive to summer temperatures. This summer-biased sensitivity makes it difficult to infer information about past precipitation regimes from the pollen record, especially in regions that are heavily dominated by winter precipitation. As such, pollen records may be a poor data set for elucidating temporal changes in winter precipitation patterns associated with Pacific teleconnection patterns such as ENSO and PDO.

At Plan B Pond, the peaks component of the charcoal data is a much better indicator of winter precipitation than the pollen/macrofossil record, largely due to the significance of winter snowpacks in determining the length of the fire season (Westerling et al., 2006). Our findings from Plan B pond suggest that charcoal data, in conjunction with pollen data may be useful in extracting information about periods of intense summer precipitation as well. As stated previously, the background component of the charcoal dataset is related to the amount of woody fuel, and is largely independent of fire frequency (Marlon et al., 2006). As such, the ratio of arboreal:non-arboreal (AP:NAP) pollen tracks the background charcoal quite well. During select time periods, this relationship breaks down and the fuel load (as indicated by AP:NAP) increases as charcoal background decreases. We suggest this is reflective of reduced fire severity due to wet summer conditions, and may be used to identify periods with anomalously wet summers.

To test this idea, we compare the Plan B Pond background charcoal and AP:NAP data with the isotope record from Bison Lake, Colorado (Anderson, 2011) (Figure 4.8). The Bison Lake data are interpreted to be a measure of the partitioning of rain and snow precipitation. We note that at ~9ka BP and again ~1ka BP when Plan B Pond background charcoal declines as the AP:NAP ratio increases, the Bison Lake record shows distinct excursions toward a more rain dominated precipitation balance. This suggests the paired charcoal and pollen datasets may indeed have some potential for extracting a summer precipitation signal.

In addition to the noted excursions at ~9ka BP and 1ka BP, it is worth noting that the Bison Lake record shows a very similar trend as the Plan B Pond background charcoal record. Anderson (2011) interprets the Bison Lake trend to be reflective of diminishing summer monsoonal precipitation inputs and increased winter moisture over time, an explanation we find unlikely to be applicable at Plan B Pond, given the complete lack of summer monsoonal precipitation today and the long-term stability of the summer monsoon footprint (Brunelle et al., 2005). An alternative explanation of the Bison Lake data, which Anderson (2011) mentions, then largely dismisses, is reduced evaporation over time as a result of declining summer insolation. Given the correspondence between our data and the Bison Lake data, we wonder if this alternative does not merit more consideration. We suggest the high frequency variability may be changes in seasonal rain/snow partitioning of annual precipitation, while the long-term trend is an evaporative signal related to changing summer insolation values.

Summary and Conclusions

Paleoenvironmental reconstructions from the Great Basin and elsewhere in the West suggest significant spatial heterogeneity in climate conditions through the Holocene. On one hand this is to be expected, given the combination of large scale, mesoscale, and local controls affecting climate across the topographically complex region (Mock, 1996; Shinker, 2011). However, the datasets from the BRR show that even in collocated records such as Minnetonka Cave and Plan B Pond, the climate record extracted from different archives, or even different proxies from the same archive, can provide very different climate histories. In the BRR records, the differences are attributed to the presence of a strong seasonal bias in each proxy dataset. This determination is possible because of the collocated records and the ability to control for localized effects related to topography and other fine scale climate variables.

With season specific paleoclimate information, we were able to assess potential biases in other paleoenvironmental records in the region to refine our understanding of hydroclimatic conditions through the Holocene. Once seasonal/forcing biases are accounted for a clearer and more homogeneous climate history can be seen. In general, paleobotanical records from the early Holocene tend to be most responsive to summer temperature and its effects on effective moisture in the growing season. Hydrologic indicators such as paleoshorelines are more responsive to winter precipitation, unless the lakes have a very high surface area/volume ratio, in which case higher summer evaporation rates in the early Holocene largely offset the winter precipitation signal.

The early Holocene history shows broad agreement across the West for much warmer summer conditions. However, winters were likely cooler and definitely much wetter, at least at higher elevations. Summer precipitation in the early Holocene, likely varied in intensity depending on location (Whitlock and Bartlein, 1993; Brunelle et al., 2005). Summer precipitation is not a climate component that any one BRR record is particularly sensitive to, but paired charcoal background accumulation and pollen data show promising potential for identifying periods of intensified summer precipitation.

The mid Holocene was characterized by dry winters and summers that were cooler than the early Holocene, but still warmer than the modern. Winter temperatures were steadily warming through the mid Holocene. In terms of hydrologic inputs, the mid Holocene was the driest interval of the Holocene. Low lake levels and periods of desiccation are widespread at this time (e.g., Shuman et al., 2009). However, because of declining summer insolation and reduced evapotranspiration compared to the early Holocene, many pollen-based environmental reconstructions indicate a shift to more mesic conditions.

Records from the late Holocene show broad agreement that summer temperatures were cooler than the early or mid Holocene, winters were wetter than the mid Holocene, and winter temperatures were slightly warmer than the long-term Holocene average. The disparity between hydrologic indicators and paleobotanical indicators is not as great in the late Holocene due to reduced magnitude of seasonality.

The reduced seasonality in the mid to late Holocene allowed GSL surface elevation data to respond more cohesively with the BRR winter precipitation record. The winter precipitation signal in the GSL surface elevation during the early Holocene was likely largely or completely offset by the increased evaporative losses associated with peak summer insolation values.

The Bear Lake paleoenvironmental record is only intermittently correlative with the BRR record, despite close proximity and a near-complete reliance on BRR hydrologic inputs. Disagreement between different Bear Lake climate proxies interpreted to be sensitive to the same subset of climate variables, suggests that nonclimatic factors have complicated the Bear Lake record.

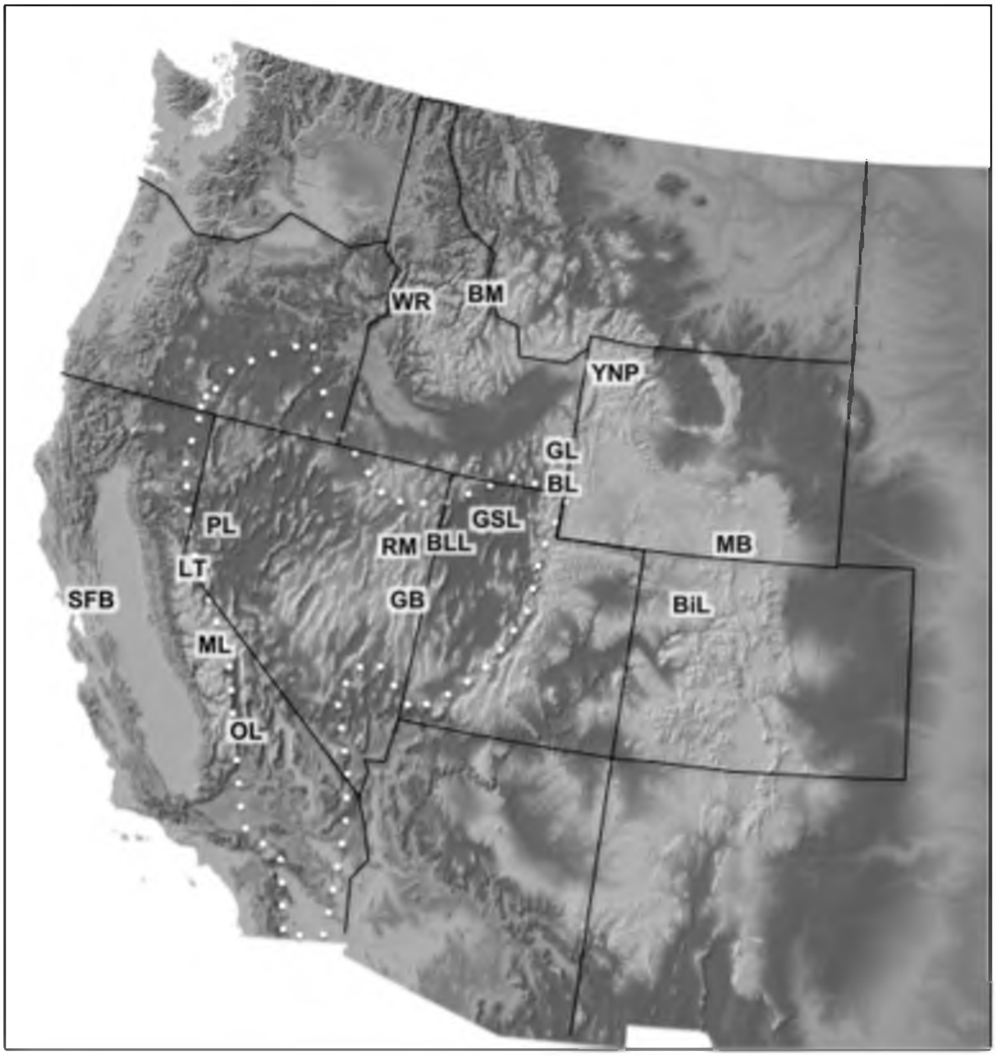

Figure 4.1. Map of Western U.S. Showing Locations of Some Paleoclimate Records Referred to in Text. SFB= San Francisco Bay; PL= Pyramid Lake, LT=Lake Tahoe, ML=Mono Lake, OL= Owens Lake, RM=Ruby Marshes, WR=West Range, BM=Bitterroot Mountains; YNP=Yellowstone National Park; GL=Gray's Lake; BL=Bear Lake; GSL=Great Salt Lake; BLL= Blue Lake; GB=Great Basin National Park; MB=Medicine Bow Mountains; BiL= Bison Lake. Dotted white line indicates approximate boundary of Great Basin.

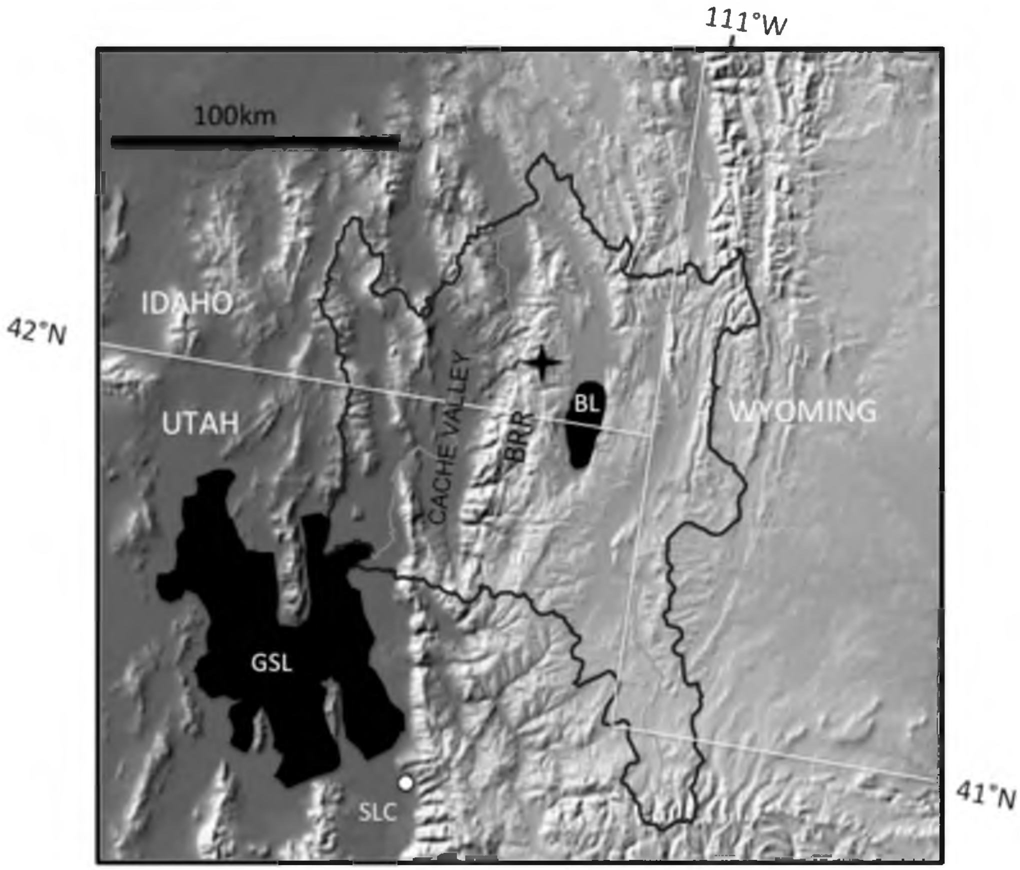

Figure 4.2. Map of Bear River Drainage Basin. BL=Bear Lake; BRR= Bear River Range; GSL=Great Salt Lake; SLC= Salt Lake City. Black line indicates drainage basin boundary.

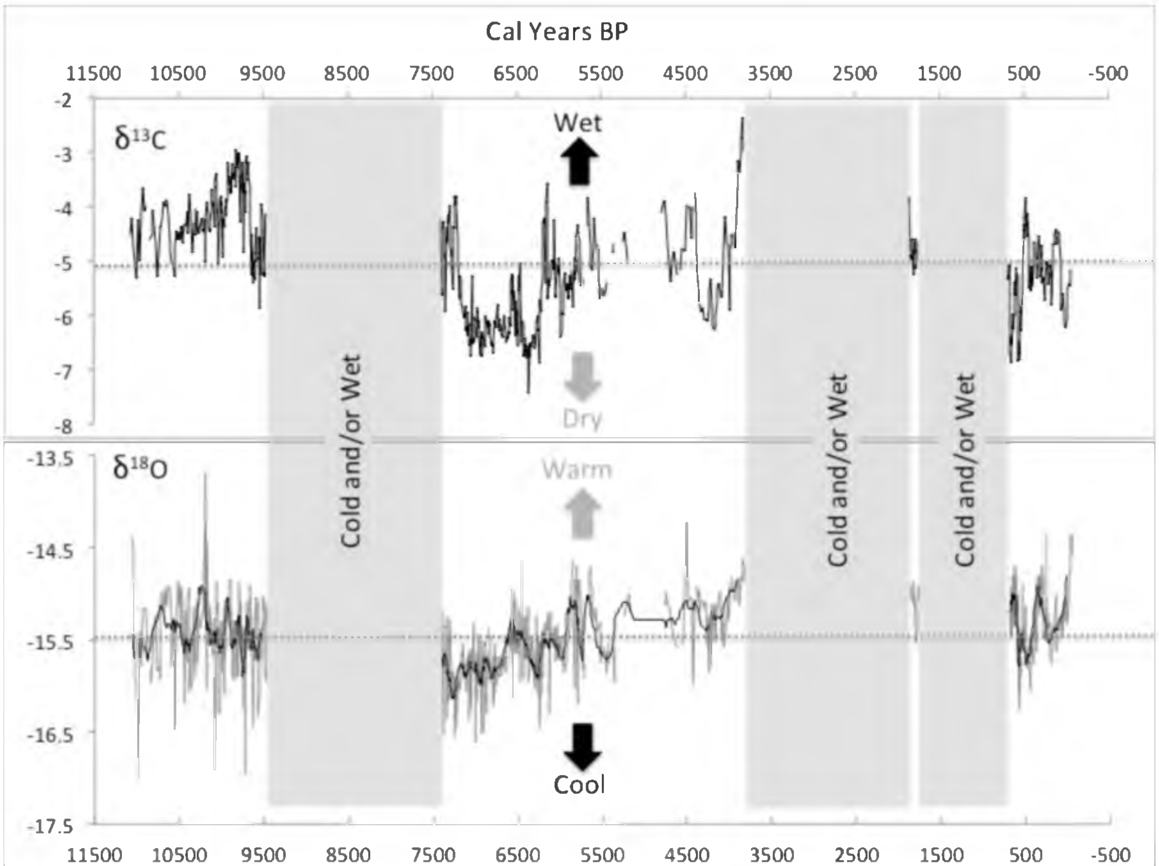

Figure 4.3. Minnetonka Cave Stable Isotope Data and Interpretation.

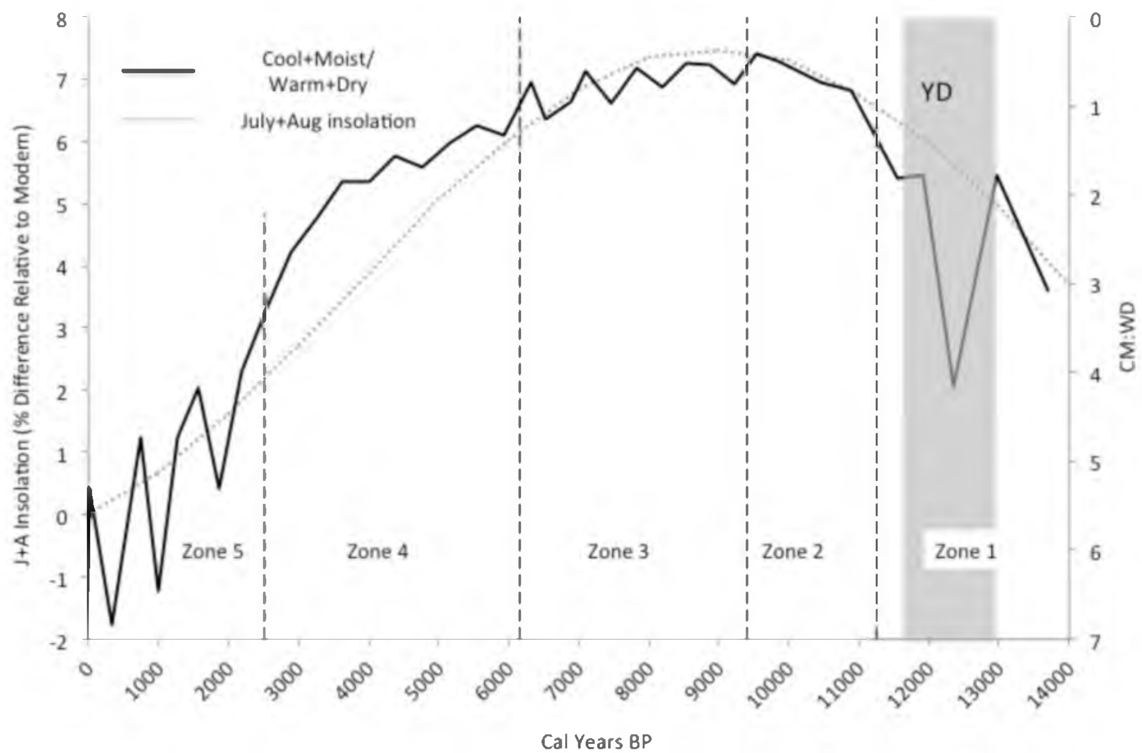

Figure 4.4. Plan B Pond Cool+Moist to Warm+Dry Taxa Pollen Ratio. Cool+Moist=Picea +*Pseudotsuga* +*Abies* + other Asteraceae. Warm+Dry= *Juniperus* +*Quercus* +*Ambrosia* +Amaranthaceae +Sarcobatus. Pollen zones based on stratigraphically constrained cluster analysis shown as vertical lines. Shaded bar indicates Younger Dryas period. Dashed line is July+August insolation in % difference from modern.

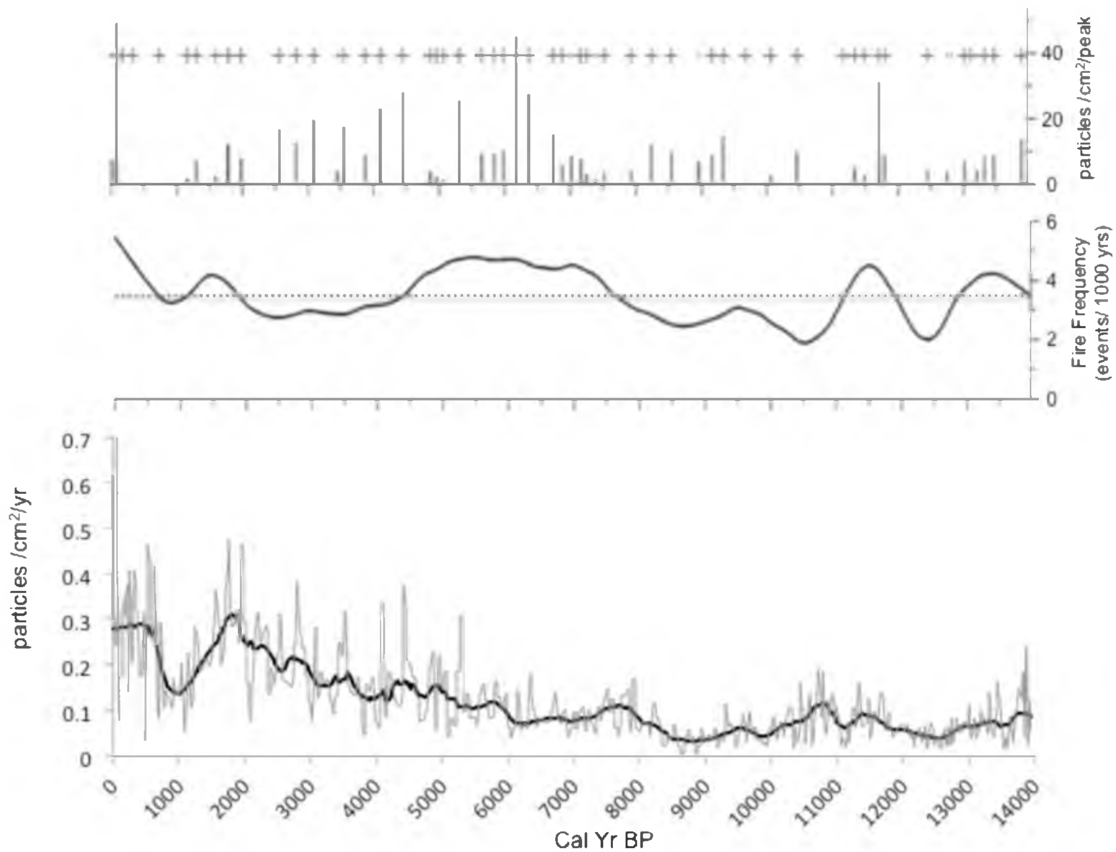

Figure 4.5. Plan B Pond Charcoal Data. Top panel--peak occurrence and magnitude. Middle Panel--1000 year smoothed fire frequency. Bottom Panel--charcoal accumulation rate (light gray line) and 300--year smoothed background (black line).

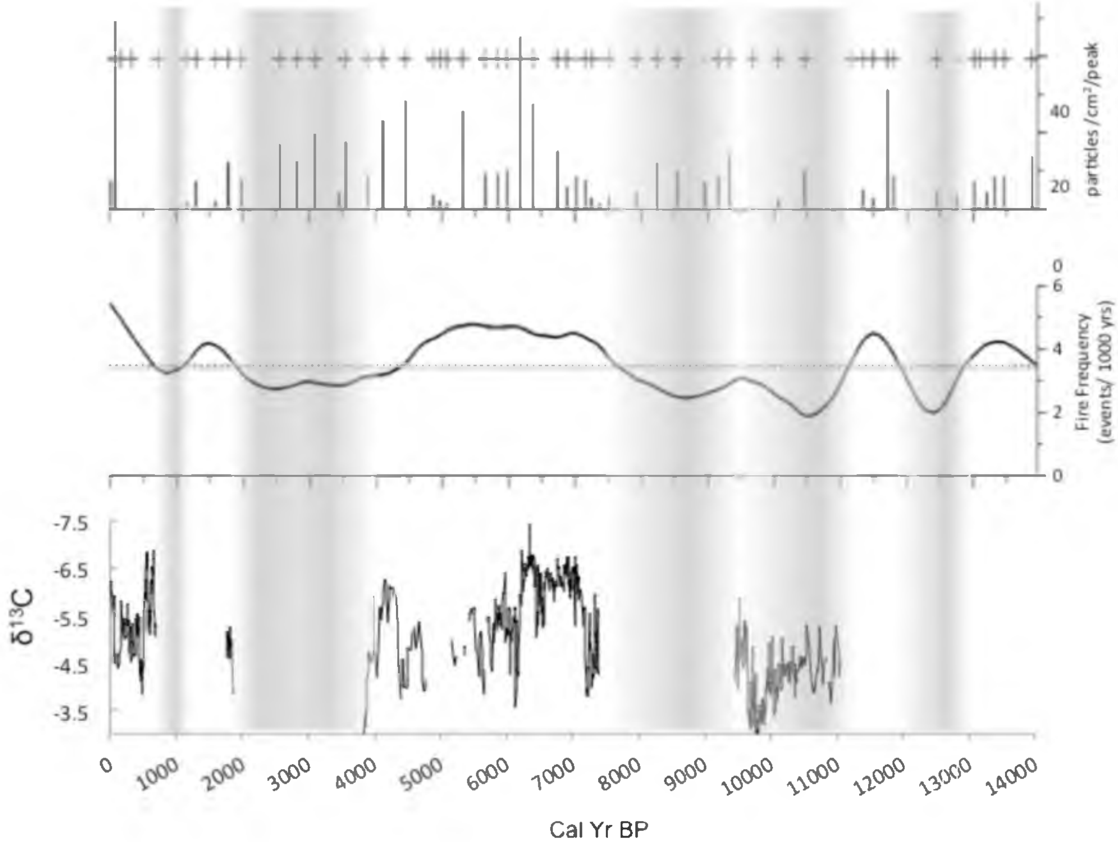

Figure 4.6. Winter Precipitation Sensitive Proxies. Top panel--fire event occurrence and peak magnitude. Middle Panel--smoothed (1000yr) reconstructed fire frequency. Bottom Panel--Minnetonka Cave δ13C data. Shaded vertical bars represent periods interpreted be anomalously wet in winter (large snowpacks).

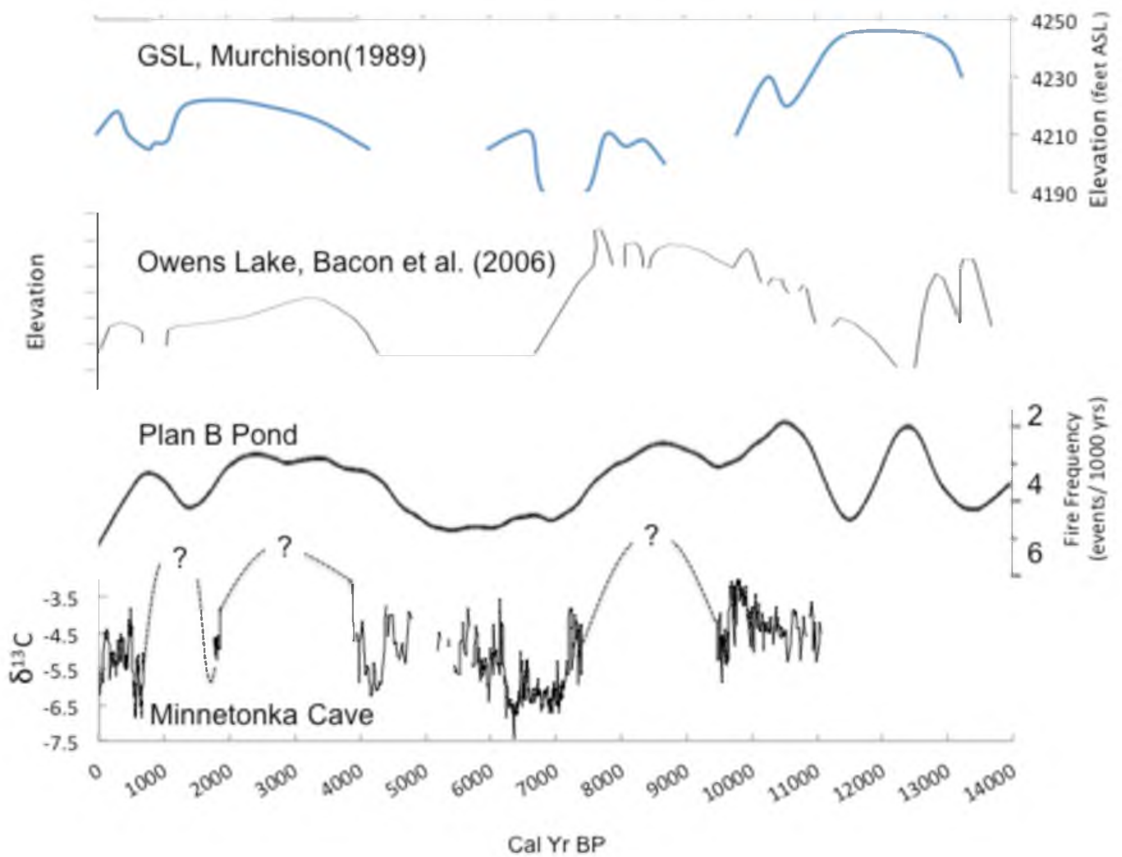

Figure 4.7. Great Salt Lake and Owens Lake Surface Elevation Reconstructions Compared to Bear River Range Winter Precipitation Sensitive Proxies. Dashed lines with question marks intended to show the direction of isotopic excursion expected or inferred from hiatus in growth of Minnetonka Cave speleothem.

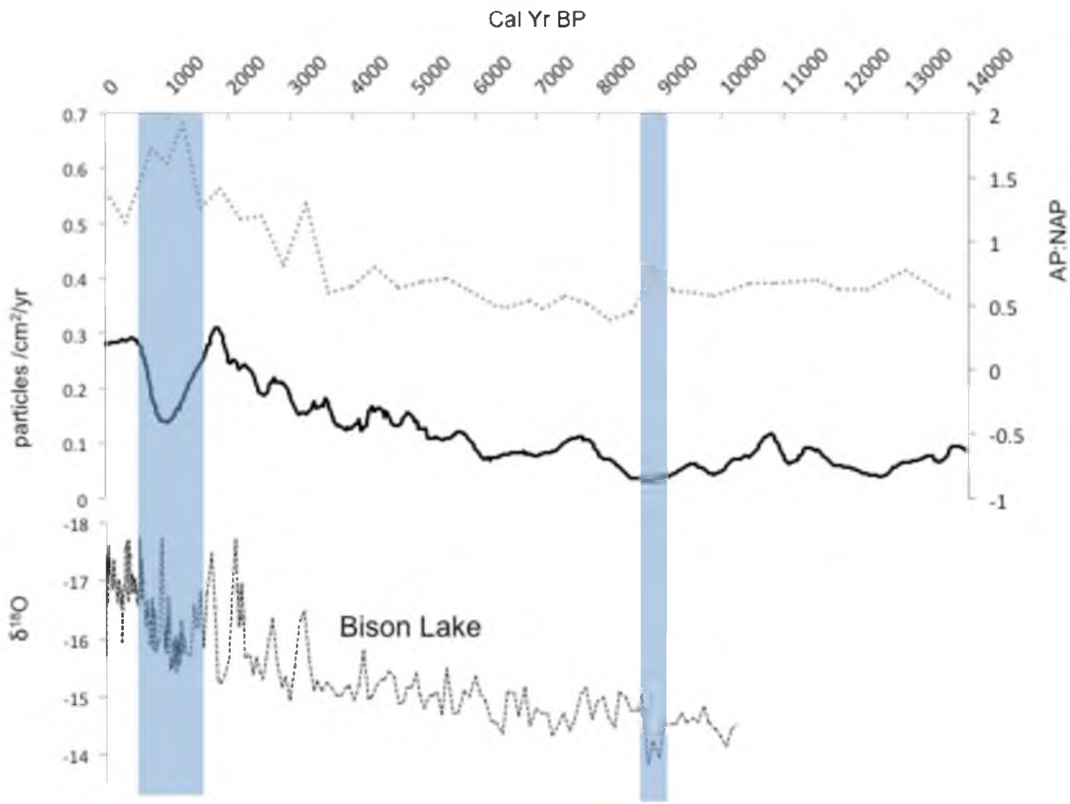

Figure 4.8. Plan B Pond AP:NAP Ratio and Charcoal Background Compared to Bison Lake Isotope Record. Top panel--ratio of arboreal terrestrial pollen to non-arboreal pollen (AP:NAP). Middle panel--Plan B Pond 300-year smoothed charcoal accumulation background (black line). Bottom panel--Bison Lake $\delta^{18}O$ record (Anderson, 2011).

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CHAPTER 5

SUMMARY BEAR RIVER RANGE PALEOENVIRONMENTAL HISTORY, IMPLICATIONS, FUTURE DIRECTIONS, AND LESSONS LEARNED

Summary and Conclusions

From this investigation we have learned much about the paleoenvironmental history of the Bear River Range, especially with respect to the long-term changes in winter precipitation, summer temperatures, and the dominant controls on vegetation composition and fire frequency. While the paleoenvironmental history of the Bear River Range is consistent with other records from the region, this study is notable do to its ability to control for localized climate variability between sites, thus allowing for identification and utilization of complementary seasonal bias for a more comprehensive paleoenvironmental reconstruction.

Results of our research allow us to address our original research questions and hypotheses.

1. How does the magnitude and frequency of climate variability in the Bear River region during the Late-Glacial and Holocene compare to historic records, especially with respect to precipitation?

- Hypothesis #1: The historic record is an adequate representation of hydroclimatic variability in the Bear River Basin for long-term water resource planning.

Our data has shown that the Great Basin and surroundings generally experienced much wetter winters in the early and late Holocene than the mid-Holocene, while summer temperatures peaked in the early Holocene and then cooled steadily through the mid and late Holocene. Winter temperatures were generally cooler in the early Holocene, warmed through the mid-Holocene, and have been more variable, but generally warmer than the long-term average in the late Holocene. Thus, we found that the both the magnitude of variability and the frequency of variability have changed over time, with the latest Holocene generally showing smaller magnitude but higher frequency variability than earlier time periods.

As a result of our findings, we can say that Hypothesis #1 is false. Durations of extreme drought or successive high snowpack years have been much longer in the past than the historic record represents, and the magnitude of extreme events is also not well represented in the historic record.

2. How sensitive is vegetation in the Bear River Range to changes in climate, and is there a primary climate variable (temperature vs. precipitation, seasonality of precipitation, etc.) responsible for vegetation shifts?

- Hypothesis #2: Bulk vegetation composition is highly responsive to changes in climate, with significant sensitivity to multiple climate variables.

Vegetation in the Bear River Range was largely unresponsive to high amounts of early Holocene winter precipitation, instead showing a much more dominant response to high summer temperatures and their associated effects on available growing season moisture. Fire, on the other hand, was responsive to both insolation and winter precipitation. The fire frequency was closely linked to winter precipitation, but fire severity and magnitude was linked with available fuel, aside from a few anomalous periods, with increasing fuel availability in the mid to late Holocene, as compared to the early Holocene, as a result of declining summer insolation and temperatures.

Based on our results, Hypothesis #2 is false, but may depend on the resolution and timespan of the study. In our study, bulk vegetation composition was not particularly sensitive to the full suite of climatic variables, but rather dominated by one climate variable, that variable being summer temperature. Due to the overwhelming direct and indirect effects of summer temperature, sensitivity to changing precipitation regimes was fairly limited until the last few thousand years when insolation has been close to modern. This suggests that sensitivity to factors other than summer temperature may exist, but can only be inferred from samples that are close in age. Thus, a study that focused on a short window of time would be less affected by large differences in insolation, making the response of vegetation to other climatic variables more visible.

3. Do speleothem and lake sediment proxies from the same mid-latitude location record climate change differently from one another? (i.e., Are they directly comparable?)

- Hypothesis #3: Collocated speleothem and lake sediment based paleoclimate reconstructions do reflect similar paleoenvironmental histories, and can be directly compared with one another.

Our results show that the Minnetonka Cave speleothem record and the Plan B Pond record both show definite seasonal biases, with the speleothem recording primarily winter conditions, and the lake sediment proxies showing different seasonal sensitivities, depending on the proxy being evaluated. Pollen, macrofossil data, and charcoal background from Plan B Pond are all mot sensitive to summer conditions, and do not correlate well with the Minnetonka Cave isotope data. However, the charcoal peaks data (fire frequency) does correlate well with the carbon isotope signal from Minnetonka Cave because they are both most sensitive to winter precipitation. Oxygen isotope data from Minnetonka Cave does not correlate well with any of the Plan B Pond proxies, as it records winter temperature, a variable that is not very influential on any of the lake core proxies.

To simply say that the Minnetonka Cave or Plan B Pond data indicate amounts of effective moisture or temperature without specifying their seasonal bias would definitely lead us to conclude that Hypothesis #3 is false, especially when comparing vegetation records with speleothem isotope data. These two data sets would indicate very different paleoenvironmental histories if they were both interpreted as indicating mean annual conditions. Even knowing that the records are strongly biased to one season or another, we must conclude that Hypothesis #3 is false, because the data sets really are not directly comparable. We would argue they are complementary rather than correlative because together they provide a clearer understanding of seasonal dynamics through time.

4. How does the localized climate record of the Bear River Range compare with the Bear Lake climate record generated by the USGS and collaborators? Is the Bear Lake

record an accurate indicator of climate in the Bear River Range over the last several thousand years?

- Hypothesis #4: The records extracted from the Bear River Range do correlate well with the Bear Lake paleoenvironment reconstructions, suggesting Bear Lake does provide an accurate record of local paleoenvironmental conditions.

The Bear River Range Records show periods of both agreement and disagreement with the Bear Lake paleoclimate records. Since the Bear River Range supplied more than 95% of the inflow to Bear Lake for most of the Holocene it is hard to imagine that Bear Lake would not show a consistent relationship with the Bear River Range Records. However, intervals of definite disagreement do exist between the two locations. This investigation has concluded that nonclimatic variables such as tectonic and/or geomorphic controls on lake hydrology have significantly affected the Bear Lake records making them less reliable over the timescale of this investigation. The magnitude of climate changes on glacial-interglacial timescales may have been large enough to outweigh the nonclimatic effects on lake hydrology. However, based on the results of this study and the late glacial-Holocene time period, Hypothesis #4 is false.

Implications and Outstanding Questions

Part of the motivation for our study in the Bear River Range was tied to the question of water resource variability and contextualizing historic hydroclimatic observations with a long-term baseline. From our data we can say that the range of snowpack variability experienced over the Holocene is not well represented by the historic record. Extended periods of much higher snowpack (e.g., 4000-2000 BP, BP= calendar years before present) and periods of much lower snowpack (e.g., 7200-6200 BP) have occurred in the region in the past, and the frequency of variability has been much lower in the past (i.e., prior to 2500 BP). Thus, extended periods of drought and extreme flood episodes have occurred far exceeding the duration and magnitude of those experienced in the historic time period.

As stated in the first chapter of this volume, precipitation in the Western U.S. is greatly affected by the changing states of large-scale ocean/atmosphere teleconnection patterns, including PNA, ENSO, and PDO. Today, the study site in the Bear River Range, tends to be wet when one or more of the following occur: the PNA is weak, ENSO is in the La Niña phase, and PDO is in the cold phase. However, the study site is located on the edge of the Western North American precipitation dipole transition zone, so it is plausible that a northward shift in the dipole boundary could cause the study site to become anomalously dry when the teleconnection patterns are in phases that today lead to locally wet conditions.

The potential for this shift in response to teleconnection phases is perhaps most significant in the early to mid-Holocene. Marine paleoenvironmental records suggest that sea surface temperature patterns from ~10000-4500 BP were indicative of sustained La Niña- like conditions (Marchitto et al., 2010). Moreover, the inter-tropical convergence zone was shifted northward from ~ 9000-4500 BP, which facilitated sustained eastern trade winds and maintained the La Niña conditions (Kaoutavas et al., 2006).

Based on modern ENSO precipitation anomaly associations in the Bear River Range, one might expect that the site would have experienced consistently wet winters for much of the early Holocene and all of the mid-Holocene. The early Holocene was indeed very wet at our study site in the Bear River Range, especially between ~9500-7500 BP, but the mid-Holocene was relatively dry, especially from ~ 7200-6200 BP. During the interval 9500-7500 BP, a speleothem record from southern New Mexico, a locale that is definitely in the La Niña-dry side of the modern precipitation dipole, shows prolonged and extreme dry conditions (Asmerom et al. 2007), suggesting that modern dipole precipitation anomalies associated with ENSO were similar to modern. However, during the interval from 7300-6300 BP, the New Mexico record continues to show dry conditions, as would be expected if SST patterns and ITCZ position continued to favor La-Niña conditions, while our Bear River Range records also show very dry conditions.

This change in phase relationship between the Bear River Range records and the New Mexico record at ~7500 BP suggests that the dipole either broke down, or that the dipole boundary shifted northward. Wise (2010) showed that the position dipole transition zone has been quite stable over the historic period, but our findings indicate further examination of that stationarity over longer periods is needed.

Another question that merits additional investigation is in regards to the conditions leading to significantly increased summer precipitation in the northeastern Great Basin, as we have inferred during the Medieval Climate Anomaly (MCA), ~ 1000 BP. Many paleoenvironmental studies have suggested that intensified convective storms or "monsoon" precipitation affected portions of the West during the early Holocene during the summer insolation maxima (e.g., Whitlock and Bartlein, 1993), but studies also show that the intensified summer precipitation was relegated to sites that have a modern summer "monsoon" precipitation signal (Brunelle et al., 2005).

Since our study site does not have significant modern summer precipitation, we would not expect to see high levels of summer precipitation in the early Holocene, and indeed we do not find such evidence in our data, with the possible exception of a very brief (few centuries) interval ~9000 BP. However, we did find evidence to suggest that summers during the time of the MCA were anomalously wet for several centuries. Unlike the 9000 BP episode, which could be related to an insolation forcing, the MCA summer precipitation anomaly does not have any clear candidates for an explanatory mechanism. Further investigations into regional climate patterns during this time period are needed to verify the summer precipitation anomaly (although the archaeological record is very supportive [e.g., Coltrain and Leavitt, 2002]) and to understand the spatial pattern of the precipitation anomaly. This may provide insights to its ultimate cause.

Final Thoughts and Lessons Learned

This study has demonstrated that even collocated paleoclimate records may record significantly different climate histories, depending the mechanisms controlling the proxy climate variable being measured. Thus, in reconstructing regional and larger-scale paleoclimate histories, significant efforts must be made to ensure spatial heterogeneity is not being inferred from what is actually variable seasonal bias in different records. Because modern climate teleconnection patterns do have distinct spatial anomaly footprints, accurate identification of paleo-footprints is needed to assess long-term climate change and to understand associated forcing mechanisms. A primary contribution of this dissertation is to aid in the assessment and identification of seasonal biases in order to better define these paleo-footprints.

Even with site specific biases identified, challenges to our understanding of regional climate variability remain. One of the most difficult challenges in a regional assessment of changing paleoenvironmental conditions is chronological control. At lower resolutions, error ranges in dating are not too limiting for understanding site-to-site correlations and relative phase relationships, but even at millennial resolutions, let alone century or decadal scales, chronological errors make broad comparisons difficult. Advances in dating methods have improved greatly in the last several decades, but much of the published literature presents poor dating precision, or limited dating control. Revisiting some of the previously studied sites and improving dating precision of these records may be a valuable exercise for paleoclimate researchers.

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