Holocene lake-level trends in the Rocky Mountains, USA

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1. Introduction

Studies of drought in the western U.S. during the past few centuries and millennia (e.g., Stahle et al., 2000; Woodhouse, 2003; Cook et al., 2004; MacDonald, 2007; Meko et al., 2007) provide insights on patterns of drought generated by “unforced” climate phenomena such as El Nino and decadal variability in Pacific climates (e.g., Swetnam and Betancourt, 1998; McCabe et al., 2004, 2007; Stone and Fritz, 2006; Stevens and Dean, 2008). Future regional changes will, instead, be forced by rising greenhouse-gas levels (e.g., Stewart et al., 2004; Diffenbaugh et al., 2005) and may be analogous to large externally-forced changes in the past, even if such changes developed over long-time scales (Overpeck et al., 1991; Webb et al., 1993). Both past and future boundary-condition changes have the potential to push climatic conditions beyond the envelope of natural variability experienced during the past millennium (Diffenbaugh et al., 2005; Williams et al., 2007). Here, we examine such potential by reconstructing Rocky Mountain lake-level trends during the Holocene, including intervals when insolation and atmospheric composition differed significantly from today (Berger and Loutre, 1991; Monnin et al., 2001). Climate model experiments have simulated large changes in precipitation and annual moisture balance across the western

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Abstract

The availability of water shapes life in the western United States, and much of the water in the region originates in the Rocky Mountains. Few studies, however, have explicitly examined the history of water levels in the Rocky Mountains during the Holocene. Here, we examine the past levels of three lakes near the Continental Divide in Montana and Colorado to reconstruct Holocene moisture trends. Using transects of sediment cores and sub-surface geophysical profiles from each lake, we find that mid-Holocene shorelines in the small lakes (4–110 ha) were as much as ~10 m below the modern lake surfaces. Our results are consistent with existing evidence from other lakes and show that a wide range of settings in the region were much drier than today before 3000–2000 years ago. We also discuss evidence for millennial-scale moisture variation, including an abruptly-initiated and -terminated wet period in Colorado from 4400 to 3700 cal yr BP, and find only limited evidence for low-lake stands during the past millennium. The extent of low-water levels during the mid-Holocene, which were most severe and widespread ca. 7000–4500 cal yr BP, is consistent with the extent of insolation-induced aridity in previously published regional climate model simulations. Like the simulations, the lake data provide no evidence for enhanced zonal flow during the mid-Holocene, which has been invoked to explain enhanced mid-continent aridity at the time. The data, including widespread evidence for large changes on orbital time scales and for more limited changes during the last millennium, confirm the ability of large boundary-condition changes to push western water supplies beyond the range of recent natural variability.
Evidence of past lake levels can reveal the range of moisture variation over multiple time scales (Benson et al., 1990; Harrison and Digerfeldt, 1993; Stine, 1994), but few studies have documented Holocene lake-level changes in the Rocky Mountains (Fritz et al., 2000). A growing number of diatom, geochemical, and geomorphic records from the Rocky Mountains indicate both long- and short-term variability in lake levels (Cumming et al., 2002; Langford, 2003; Stone and Fritz, 2006; Stevens et al., 2006; Bracht et al., 2008; Stevens and Dean, 2008). Here we synthesize new and existing sedimentological evidence for long-term water-level trends, and compare the evidence for these trends with the limited sedimentary evidence of shorter-term variations. Three key questions motivate this study:

1. What long-term hydrologic trends are evident in the Rocky Mountains during the Holocene?
What spatial patterns characterize the long-term trends in moisture variation?

How similar were long-term moisture variations in the Rocky Mountains and mid-continent?

To address these three questions, we studied the lake-level history of a lake in western Montana and two lakes in Colorado, and compiled existing sedimentary evidence for shoreline shifts or desiccation events in an additional 14 lakes or wetlands (Figures 1a and 2). The small size of our new study lakes (5 ha to 1.1 km$^2$) ensures that past lake-level fluctuations capture local hydrologic conditions (Harrison and Digerfeldt, 1993). Each new study lake was chosen to represent a different region defined by Cayan (1996) for having broadly-coherent snowpack variation at interannual scales (Figure 1a), because lake levels depend heavily on surplus moisture derived primarily from winter precipitation (Figure 2), and long-term forcing may have generated broad-scale moisture anomalies similar to those at interannual time scales (Mock and Brunelle-Daines, 1999; Harrison et al., 2003; Whitlock and Bartlein, 2004; Shin et al., 2006). Indeed, some of Cayan’s (1996) snowpack patterns (i.e., the “Idaho” pattern; Figure 1a) are associated with atmospheric and oceanic conditions like those that have been simulated and inferred for the mid-Holocene, including a deep Aleutian low (e.g., Bartlein et al., 1998; Anderson et al., 2005) and a cool north Pacific (Kim et al., 2004). Likewise, Licciardi (2001) compared the histories of four lakes (Chewaucan, Lahontan, Bonneville, and Owens) in the Great Basin and showed important synoptic differences like those that exist at short time scales today.

Figure 2. Study site bathymetry, air photos, core and sub-surface profile locations, and modern climatology. Dashed lines on photos mark bathymetry (note different m intervals among sites); solid lines on photos mark sub-surface profiles; dots indicate core locations. Monthly temperature ($T$), precipitation ($P$), and $P$ minus potential evapotranspiration (PET) are shown at left, and are based on the AD 1971–2000 climatologies for the U.S. climate divisions that contain the lakes (NCDC, 1994) and the Thornthwaite (1948) method for calculating PET. Dashed lines represent $P$; dark gray indicates periods of potential lake and groundwater recharge when PET $>$ $P$. 

cause lake levels depend heavily on surplus moisture derived primarily from winter precipitation (Figure 2), and long-term forcing may have generated broad-scale moisture anomalies similar to those at interannual time scales (Mock and Brunelle-Daines, 1999; Harrison et al., 2003; Whitlock and Bartlein, 2004; Shin et al., 2006). Indeed, some of Cayan’s (1996) snowpack patterns (i.e., the “Idaho” pattern; Figure 1a) are associated with atmospheric and oceanic conditions like those that have been simulated and inferred for the mid-Holocene, including a deep Aleutian low (e.g., Bartlein et al., 1998; Anderson et al., 2005) and a cool north Pacific (Kim et al., 2004). Likewise, Licciardi (2001) compared the histories of four lakes (Chewaucan, Lahontan, Bonneville, and Owens) in the Great Basin and showed important synoptic differences like those that exist at short time scales today.
We present detailed data for our three study sites, and then review all of the other direct evidence of lake desiccation or shoreline changes in the Rocky Mountains (Figure 1a). We discuss the long-term patterns in these data, some evidence for millennial and shorter changes in climate, and comparisons among the lake-level data, other paleoenvironmental records, and climate model simulations.

2. Study sites

We targeted small lakes with small watersheds and minimal stream inputs to obtain records that (a) were minimally affected by local stream processes (e.g., down cutting of lake outlets; variable contributions of stream-derived sediments to sediment sequences), and (b) were controlled by groundwater levels, which integrate multiple years of precipitation. Such lakes have been shown to be climatically sensitive and to have regionally-synchronous trends (Harrison and Digerfeldt, 1993; Shuman and Donnelly, 2006). Two lakes are diamictom-dammed lakes fed by groundwater in unconsolidated sediments near the forest-steppe boundaries of the Flathead Valley of Montana and the North Park Valley of Colorado. The third lake is a small glacially-scoured bedrock basin in subalpine forest in the San Juan Mountains, Colorado.

Foy Lake in western Montana (48.17°N, 114.35°W, 1006 m elevation, 110 ha surface area) fills a depression in till that formed as glaciers retreated from the Flathead Valley before 13,100 cal yr BP (Smith, 2004). Proterozoic sediments comprise the bedrock near Foy Lake, and contain some carbonate units (Harrison et al., 1986). The lake is primarily a groundwater outcrop with only minor stream inputs, and has a historically-widened outlet. The immediate watershed of the lake covers ~300 ha, but groundwater may be derived from an additional ~1500 ha. Stevens et al. (2006) and Stone and Fritz (2006) studied the diatom assemblages and geochemical composition of a core collected from the center of Foy Lake in 39.9 m of water. Power et al. (2006) and Power (2006) described the local fire and vegetation history based on an examination of charcoal and pollen. Foy Lake lies within Cayan’s (1996) “Idaho” pattern of snowpack variation (Figure 1a).

Hidden Lake (40.51°N, 106.61°W, 2710 m elevation, 6.5 ha surface area) is located behind ridges of unconsolidated sediments on the eastern edge of the Park Range of northern Colorado. This surficially-closed basin is fed by groundwater from a 94-ha watershed. The ridges, which dam the lake, contain boulders composed of Tertiary volcanics and Cretaceous sandstones, which outcrop near the lake (Snyder, 1980). These features have been interpreted as block-slip deposits (Snyder, 1980), but field inspection indicates that they may be moraines, which would date to at least 17,300–16,300 cal yr BP (Benson et al., 2005). The lake has a maximum depth of 9 m, and sits within Cayan’s (1996) “Colorado” region (Figure 1a).

Little Molas Lake (37.74°N, 107.71°W, 3330 m elevation, 5 ha surface area) is located at Molas Pass, in Colorado’s San Juan Mountains, within the region of Cayan’s (1996) “New Mexico” pattern (Figure 1a). The small lake regularly overflows and has one small inflowing stream, which drains a ~275-ha watershed including a near-by cirque. The lake lies in a glacially-scoured karst feature of the Mississippian Leadville Limestone, which is draped by the red silts of the Mississippian Molas Formation (Blair et al., 1996). The bedrock forms a submerged ridge that partially divides the lake basin (Figure 2). Tertiary volcanics form the upper elevations of the Little Molas Lake watershed. The maximum depth of the lake is 5 m. Toney and Anderson (2006) analyzed pollen and sedimentary charcoal in a deep-water core from the lake. Maher (1961) analyzed fossil pollen from Molas Lake located 2.3 km to the east.

3. Methods

Our approach follows the methods of Digerfeldt (1986) and Shuman et al. (2001, 2005) and uses transects of sediment cores within each lake, in combination with sub-surface profiles collected by seismic and ground-penetrating radar (GPR) geophysical techniques, to track shifts in the position of near-shore sediments. The method depends on the assertion that wave energy prohibits the accumulation of sediment in the shallowest areas of the lake. Littoral sediment types (e.g., sands) thus accumulate closer to the center of the basin when water levels are low than when lake levels are high. A rise (fall) in lake level provides one of the few mechanisms for raising (lowering) the elevation of the energy-defined limit of fine sediment accumulation (the “sediment limit” of Digerfeldt, 1986; Dearing, 1997; or “depositional boundary depth” of Rowan et al., 1992). Even as the lake fills with lacustrine sediment, fine-grained muds can only accumulate where the energy of the environment is low; infilling alone cannot cause shoreward expansion of sediment accumulation. Therefore, near-shore sediment accumulation can be used to infer both high and low-lake stands; and the presence of young sediments near shore can be taken to confirm recent high water levels (see Davis and Ford, 1982; Shuman et al., 2005 for contrast). Previous studies have also demonstrated the potential to find stratigraphic evidence of century-scale lake-level changes (e.g., Abbott et al., 1997). Sediment descriptions used here follow Schnurrenberger et al. (2003); all lake sediments are referred to as lacustrine, but sediments deposited in shallow water above wave base are described as littoral.

To estimate past lake levels, we reconstructed the past elevations of wave base (i.e., the physical control on the boundary between shallow- and deep-water sediment facies, see e.g., Shuman, 2003; Anderson et al., 2005). To do so, we find the highest elevation location of fine-grained lacustrine sediments at a particular time. Low-lake stands are identified in sub-surface profiles by the truncation of sedimentary units near shore and in near-shore cores as intervals of low sedimentation rates (<10 cm/ka; Webb and Webb, 1988), coarse-grained sediment lags, and other littoral sedimentary features (Dearing, 1997). High stands appear evident as basin-wide units of fine-grained sediment that extend to high elevations in the lake basin.

3.1. Sub-surface profiles

We collected over ten sub-surface profiles in each study lake to identify the major stratigraphic features. Representative profiles are shown here (Figures 3–5). Evidence of overlapping sediments (increasing shoreward extent of overlying sediment layers) was inferred to indicate time-transgressive increases in water level. Strong near-shore reflectors indicate deposits of coarse-grained littoral sediments (“paleoshorelines”) that are disconformable with underlying sediment units. Paleoshorelines were differentiated from local slumps, because they were found around the entire lake. At Foy Lake, where water depths exceeded 40 m and where water conductivity was high (1120 μS), we used an EdgeTech 424 CHIRP seismic-reflection system. At Hidden and Little Molas Lakes, where water depths were less than 10 m and water conductivity was low (20 and 120 μS respectively), we used a Geophysical Survey Systems, Inc. (CSSI) SIR-3000 GPR with a 200 mHz antenna floated in a raft on the water surface.

3.2. Collection and analysis of sediment cores

Sediment cores were collected with a piston corer along transects from the center to the shore of each lake (Figures 2 & 3). At Foy Lake, sediments were collected in 2002 with a 5-
1865

cm diameter modified Livingstone corer in water depths of 4.0, 7.9, 15.9, and 20.0 m. At Hidden Lake, sediments were collected in 2005 using 2–3 m long, 10-cm diameter polycarbonate tubes with a piston at three deep locations (north, central, and southern) and along two transects toward shore. A central transect of cores collected in 6.6, 6.0, 5.5, 5.0, 4.5, and 3.5 m of water is considered in this paper. A core of the complete sediment sequence from the deepest portion of the lake was collected in 2007. At Little Molas Lake, we used a modified Livingstone corer with 10-cm diameter polycarbonate barrels to collect a single transect of cores from 3.0, 2.0, 1.5, 1.0, and 0.69 m of water in 2005.

Cores were logged and imaged using a Geotek Multi-Sensor Core Logger and DMZ Core Scanner at the National Lacustrine Core Repository Core Lab (LacCore) at the University of Minnesota. The logs provided magnetic-susceptibility measurements (Nowaczyk et al., 2002) for all three lakes and gamma-ray attenuation bulk density measurements (Zolitschka et al., 2002) for the polycarbonate-lined cores from Hidden and Little Molas Lakes. A coulometer was used to measure the total carbonate and total organic content of the carbonate-rich sediments from Foy Lake; loss-on-ignition analyses provided information on the water, organic, and carbonate content (Dean, 1974) of Hidden Lake cores. Sediments were also visually inspected to classify facies based on sediment structure and grain-size and to identify potential unconformity surfaces.

3.3. Chronology

AMS radiocarbon dates were obtained from selected cores to provide a chronology for each basin. Dating targeted key stratigraphic changes, but the chronologies are limited by the
availability of datable material. For this reason, we emphasize
long-term features of the lake histories here, which appear ro-

bust regardless of the specific ages of individual stratigraphic
changes. All dates were calibrated using CALIB 5.0 (Reimer et

al., 2004), and rounded to the nearest 5 years.

AMS radiocarbon ages were obtained on macrofossil ma-
terial, charcoal, and bulk sediment at Hidden and Little Mo-
las lakes (Table 1). Four dates from Foy Lake were obtained on
5 cm$^3$ of core material that was processed to concentrate pollen
using a heavy-liquid separation method (Brown et al., 1989).
The use of bulk sediment for dating at Hidden Lake is sup-
ported by comparable ages and sedimentation rates obtained
on a sub-set of macrofossil and charcoal samples.

4. Site-specific analyses

4.1. Foy Lake, Montana

4.1.1. CHIRP data

Seismic profiles from Foy Lake show widespread stratifi-
ced sediments across deep areas of the lake (Figure 3). The
Mazama Ash (MZ) layer, dating to ca. 7600 cal yr BP (Zdanow-
icz et al., 1999), appears evident as a dark band in the profiles.
Biogenic gas obscures the lowermost sediments in many parts
of the lake, but appears constrained below specific layers. Less
gas is evident in areas where homogenized cones of sediment
extend upward from the MZ as though these features formed
during the release of gas trapped below the MZ.

A dark reflector in the seismic profiles (Figure 5) extends
outward from shore and truncates underlying features includ-
ing the MZ and the gas-rich layers, but ends before reaching
areas of about 20 m water depth. The feature appears in mul-

Figure 4. Interpretation of key features in sub-surface profiles. Dashed line
on profile of Little Molas Lake shows elevation of top of bedrock outcrop
in comparison to lower extent of near-shore unconformity.

Figure 5. Details from the near-shore segments of two seismic-reflection profiles across Foy Lake. Eastern segments are shown in the two columns on the left
with a schematic interpretation of the major onlapping and truncated reflectors, and unconformities, to the left. Western segments are shown in the two col-
umns to the right with the schematic interpretation to the right. Arrows labeled “PS” indicate possible paleoshoreline features.
Six major sediment facies were found in the Foy Lake cores (Figure 6). The facies represent a gradient from deep to shallow water with a transition from fine laminations to coarse sand and gastropods (Figure 6a). The deepest water facies (0) includes a series of thin organic-and-carbonate laminations, which are likely varves similar to those observed in the deeper, central core study by Stevens et al. (2006) and Power et al. (2006). Correlative sediments in shallower portions of the lake contain thick carbonate-rich laminations (facies 1). Cores A and C from 15.9 and 7.9 m of water respectively also contain a dense lag of gastropod shells (facies 5). The sandy facies (3–5) and back (Figure 6b). Laminated and massive marls (facies 2), fine carbonate sands (facies 3), and coarse-grained carbonate sands (facies 4). Carbonate fragments form the sand-sized grains of facies 3 and 4 and appear to have been precipitated as crusts around littoral organisms (Charophyte casts). The lowermost sediments in shallow-water core C contain a dense lag of gastropod shells (facies 5). The sandy facies were not found in the deep-water core from 39.9 m of water that was the focus of previous work (Power, 2006; Power et al., 2006; Stevens et al., 2006; Stone and Fritz, 2006). Core B in 20.0 m of water is similar to A, and is not discussed.

The cores contain five stratigraphic units, which represent repeated transitions from fine-grained facies (0–2) to coarse-grained facies (3–5) and back (Figure 6b). Laminated and massive marls (facies 1, 2) form Unit I at the base of cores A and C. Massive carbonate sands comprised of Charophyte casts (facies 3–4) then overlie Unit I and form Unit 2. The MZ tephra marks a sharp transition (Figures 6b and 7c) back to massive marl (facies 2), which forms Unit 3 in both cores. Carbonate sands (facies 4) then form Unit 4 in cores C and A, but transition abruptly to the fine laminated sediments (facies 0–1) of Unit 5 in the upper 125–140 cm of all of the cores (Figure 6b).

Several of the stratigraphic units are only well represented in the deepest core A (Figure 6b). Unit 1 in cores A and C contains the Glacier Peak (GP) tephra, which dates to 13,100 cal yr BP (Sarna-Wojcicki et al., 1983), but the thickness of sediment between the GP and MZ tephras (Figure 6b) decreases from core A (151 cm) to core C (74.5 cm). In fact, a distinct unconformity surface in core C at the top of Unit 1 is marked by gastropod shells and a sharp transition from the submarine laminations (facies 1, 2) of Unit 1 to the carbonate sands (facies 4) of Unit 2 (Figure 7d). The unconformity lies 240 cm below the top of core C, and is 1030 cm below the modern lake surface (bmls). In core A, a dark, high-magnetic-susceptibility sand layer marks the transition from Unit 1 to Unit 2 at 347–350 cm below the surface of the core (1937–1940 cm bmls).

Likewise, unit 4 is 90 cm thick in core A, but only 15 cm thick or less in cores C and D (Figure 6b). An unconformity at the base of Unit 4 is evident 140 cm below the top of core C (930 cm bmls; Figure 7b). This greenish contact between the marl of Unit 3 and the sands of Unit 4 (Figure 7b) appears visually to contain abundant glauconite, which would be consistent with a non-accumulation surface inhabited by gastropods and other littoral organisms. In core D, a gastropod lag (Figure 6a, facies 5) overlies the unconformity. The unconformity appears to correlate to the prominent unconformity and paleoshoreline in the CHIRP profiles (Figures 3–5). Only Unit 5 is well defined in all cores, and coincides with the first rise in the organic carbon content of the cores above 5% in D and C and above 10% in core A (Figure 8).

### Table 1. Radiocarbon samples and their calibrated ages based on CALIB 5.0.2 (Reimer et al., 2004) by core.

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* Collected from the location I in 2007; all other cores from Hidden Lake were collected in 2005.
The two other AMS radiocarbon dates on pollen from 237 to 238 cm and 245 to 246 cm in core C bracket the lower unconformity between Units 1 and 2. The ages for these two dates (6750–6670 and 7790–7700 cal yr BP; Table 1) are inconsistent with their position relative to the MZ and GP tephras and were not used in our age model. The upper date was obtained 55 cm below the MZ but was 850–920 yrs younger than the MZ; the lower date was 12 cm above the GP but was 5310–5400 yrs younger than the GP.

### 4.1.4. Sedimentation rates

The chronology indicates hiatuses in sediment accumulation (Figure 8) consistent with the two apparent unconformities in core C (Figure 7b, d; note that a break in the plot of core C at ca. 2000 yrs BP in Figure 8 is due to a lack of data not an unconformity). The MZ and the two calibrated radiocarbon ages of ca. 2780 cal yr BP bracket the upper unconformity surface in core C. A linear sedimentation rate between the MZ and the mean age of these dates would equal 11.1 cm/ka and is close to Webb and Webb’s (1988) threshold for nonconstant accumulation. The unconformity dates to 6175–2780 cal yr BP, if a sedimentation rate of 31.2 cm/ka is applied to the 23-cm massive marl in core C between MZ and the unconformity. The rate equals that between MZ and the varve-like laminations in core A, and may be a reasonable assumption for core C given the similar spacing of CHIRP reflectors below the unconformity at the locations of cores C and A (Figures 3–5).

The MZ and GP ashes bracket the lower unconformity in core C; a linear sedimentation rate between tephras would equal 23.0 cm/ka. The unconformity is 14 cm above the GP tephra, and thus its lower age is slightly younger than 13,100 cal yr BP. The unconformity would last from 12,600 to 9760 cal yr BP if a sedimentation rate of 36.8 cm/ka is applied to the sediments above and below—based on the assumption that the continuous sedimentation rate from core A between the MZ and GP ashes applies to core C. Alternatively, the unconformity represents a period of limited accumulation over 1040 yrs based on age difference between the median calibrated ages of the bracketing but anomalously young radiocarbon dates. If so, the age range of the unconformity is

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**Figure 6.** Major sediment facies (A) and core stratigraphies (B) from Foy Lake, Montana. Magnetic-susceptibility data are presented as dashed lines; graphs of visually-identified facies (values equal numbers associated with photos in A) are plotted as solid lines. Calibrated ages of two radiocarbon dates in core C and the Mazama (MZ) and Glacier Peak (GP) tephras are shown.
Holocene lake-level trend in the Rocky Mountains

12,600–11,560 cal yr BP, which is similar to the age of a low-water-level period from ca. 12,500 to ca. 11,000 cal yr BP inferred from diatoms (Stone and Fritz, 2006).

4.1.5. Inferred changes in lake level

Based on the elevations and ages of massive marls (facies 2), thickly laminated marls (facies 1), and varve-like laminations (facies 0), we infer that the primary long-term signal at Foy Lake was a rise in the position of wave base, and thus the lake level, after early- and mid-Holocene (Figure 8). Given the depth of the uppermost unconformity in core C and absence of units 1–4 in core D (Figure 6b), wave base in Foy Lake was at or below 930 cm bmls for an extended period between 7600 and 2780 cal yr BP, and then rose to <525 cm bmls to allow for the initiation of deposition in core D (Figure 8). The accumulation of sandy sediments (Units 2 and 4) during much of the Holocene in cores A and C, and as deep as 1940 cm bmls, appears consistent with a lack of earlier high stands. The inference fits with a period of low or fluctuating lake levels at Foy Lake inferred from diatom assemblages between 7000 and 2400 cal yr BP (Stone and Fritz, 2006). Likewise, Stevens et al. (2006) estimate that the water level rose ~12 m within the past 2200 years on the basis of diatom and geochemical evidence. Our data confirm that the total lake-level rise since before ca. 2800 cal yr BP was >5.3 m based on the depth of the unconformity in core C, and could have been >16 m based on the deposition of sands (Unit 4) in core A (Figures 6b & 7a).

Two additional fluctuations in wave base—and therefore water level—appear superimposed on the persistently lower-than-modern levels prior to ca. 2800 cal yr BP (Figure 8). First, the lowest unconformity in core C, and an associated sand layer in core A, mark a period when wave base was likely at least 1030 cm bmls and as much as 1940 cm bmls between the deposition of the GP and MZ tephras. The unconformity ages suggest that the low stand has a Younger Dryas age (12,900–11,600 cal yr BP), which is consistent with low-water levels at ca. 12,500–11,000 cal yr BP inferred from diatoms by Stone and Fritz (2006). Second, the transition from sands to marl associated with the MZ is consistent with a relatively high wave base for a few centuries to a millennium after the eruption.

Figure 7. Four important facies transitions in cores from Foy Lake. (A) Transition from carbonate sands to fine varve-like laminations is shown from core A. (B) The upper unconformity surface, indicated by white dashed lines, and the transition from carbonate sands to broad laminations are shown from core C. (C) The Mazama ash layer and associated transition from sands to massive marl is shown for core A, and is representative of a similar transition in core C. (D) The lower unconformity surface between laminated marl and carbonate sand, and including gastropod shells, is shown from core C. Depth intervals of individual images are listed; “Dr.” indicates drive or core sub-section. Breaks in core surfaces occurred because cores dried before imaging.
Stone and Fritz (2006) observe an abrupt change in diatom assemblages following the deposition of MZ tepha and infer high water levels from 7600 to 7100 cal yr BP; Power (2006) observed a sharp change in pollen assemblages and fire regimes at the same time.

Our core stratigraphies and sub-surface profiles contain little evidence for lake-level changes on century time scales in the past 2000 years, except for subtle facies changes (massive to banded marls) in core D (Figure 6b). Geochemical and diatom data indicate, however, that the lake level continued to change with the two basins of Foy Lake merging only at ca. 800 cal yr BP (Stevens et al., 2006; Stone and Fritz, 2006), and therefore, near the end of the dry Medieval period recorded by tree-ring data (Cook et al., 2004).

4.2. Hidden Lake, Northern Colorado

4.2.1. GPR data

GPR profiles from Hidden Lake show well-stratified sediments extending across the lake basin in water depths of >3 m, and show that only the uppermost strata extend above water depths of about 6 m (Figure 3). The profiles also show evidence of three near-shore unconformities, which truncate the underlying sediments and form the base of onlapping sequences (Figure 9). Two strong reflectors (labeled “PS” in Figure 9) are associated with the unconformities and are evident near shore in multiple profiles throughout the basin. We infer that the reflectors represent the extent of the littoral zone during low stands of the lake. The older and larger of the two features is associated with two unconformities and may, therefore, represent a feature formed during two consecutive low stands.

4.2.2. Core data

Cores C, D, and E in water depths of 6.0, 5.5, and 5.0 m respectively penetrate the older and most prominent low-stand reflector in the GPR profiles (Figures 2 & 3). Cores F and H in 4.5 and 3.5 m of water penetrate only the uppermost strata that extend nearest to shore. Core F penetrates the uppermost unconformity, which is not evident at the location of core H (Figure 3). The shallow cores (C–H) decrease in thickness shoreward with 91 cm of organic-rich silts in core C and only 6 cm of organic-rich silts in core H (Figure 10).

The cores contain two facies: (1) a dark-banded silt with >20% loss-on-ignition at 550 °C and (2) a light gray inorganic silt with <10% loss-on-ignition at 550 °C. Twenty centimeters of dark-banded silt (facies 1) form the base of cores C, D, and E. These basal sediments are denser and contain less water than sediments above, except for a prominent gray layer of facies 2 that directly overlies the lowermost sediments (Figure 10). Radiocarbon ages from cores E and C indicate that the basal unit of facies 1 also has the same age (ca. 4400–3700 cal yr BP) in both cores, and is much younger than a calibrated basal radiocarbon date of 8410 (8425–8400) cal yr BP from a core collected in the depocenter of the lake (location I in Figure 2b).

Facies 2 is similar to the fine silt- and clay-rich matrix of the till surrounding the lake, and forms a distinct 8–19-cm thick layer separating the lower unit of facies 1 from subsequent units in cores E, D, and C (Figure 10). The facies 2 layer dates from 3750 to 2950 cal yr BP in core C, and from 3720 to 2025 cal yr BP in core E. It has a sharp lower contact in cores E and D at 563 and 610 cm bmls respectively, and becomes progressively thicker toward shore where the bases of cores F and H contain an upward fining sequence of silts, sands, and pebbles. The layer is denser than other sediments in the cores (Figure 10) and likely correlates with the older of the two strong reflectors in the GPR profiles (Figure 9). In a second transect of cores at the southern end of the lake (not shown), the layer overlies and fills a mud crack preserved in one core.

Two additional light-colored bands of facies 2 are evident amid the sediments of facies 1, which comprise the uppermost sediments in all of the cores above the prominent gray layer (Figure 10). Peaks in density and decreases in sediment water content mark the first of these bands at 39–42 and 29–31 cm depth in cores D and E, where the band is associated with a calibrated radiocarbon age of 1220 cal yr BP (Figure 10). The second band at 10–20 cm depth in cores F, E, and D, becomes increasingly light colored and dense toward shore, and is bracketed by ages of 850 and 490 cal yr BP (Figure 10). These bands may correlate with the upper two unconformities in the GPR profiles (Figure 9). Neither band is evident amid the 6 cm of facies 1 at the top of core H.

4.2.3. Sedimentation rates

The basal ages from cores C and E of 4395 and 4230 cal yr BP were obtained on material 689–690 cm bmls and 568–569 cm bmls respectively, and indicate a lack of sediment accumulation >390 cm below the modern “sediment limit” until ca. 4400 cal yr BP (Figure 11). By contrast, sediment had been accumulating at the location of core I since ca. 8400 cal yr BP (Table 1).

Net sedimentation rates in core C reached 22.6–34.7 cm/ka from 4370 to 3750 cal yr BP, and were 29.8–56.0 cm/ka in core E from 4230 to 3720 cal yr BP (Figure 11). The rates slowed, however, after statistically similar ages of 3820–3690 cal yr BP in
Holocene lake-level trends in the Rocky Mountains

In core C, the prominent gray layer of facies 2 accumulated between 3750 and 2950 cal yr BP at a net sedimentation rate of 12.3–16.5 cm/ka, which is close to Webb and Webb’s (1988) threshold for constant accumulation. In core E, the net sedimentation rate was below the threshold: 7.7–8.7 cm/ka. The rate in core E remained low (3.5–4.1 cm/ka) until an age of 1260–1180 cal yr BP. The gray layer, therefore, probably coincides with an unconformity that spans from ca. 3700 to 2990 cal yr BP at 673–674 cm bmls in core C, and from ca. 3700 to 1200 cal yr BP at 549–536 cm bmls in core E. The ages therefore indicate a 930–2690 yr period of limited sedimentation >320 cm below the modern “sediment limit” (Figure 11). The different time ranges of slow sedimentation in cores E and C match with onlapping above the lowest paleoshoreline in the GPR data (Figure 9).

The uppermost band of facies 2 in core E dates to 850–490 cal yr BP (Figure 10), and may correlate with the uppermost unconformity evident in GPR profiles near the top of the section (Figure 9). The sedimentation rate between the bracketing dates is 23.3–35.3 cm/ka, which is higher than Webb and Webb’s (1988) threshold. The rate, however, is slower than that for the immediately underlying sediments in core E (25.5–44.0 cm/ka) and for the sediments of equal age in the central core, I (47.1–54.2 cm/ka). The relatively low sedimentation rates, and the high near-shore density of the layer, could be consistent with a low-lake stand during the Medieval period. The 6 cm of organic silts (facies 1) in core H only date to 55 cal yr BP (Figure 11; Table 1).

Figure 9. Details from the western near-shore segments of three GPR profiles of Hidden Lake. Radar data are shown in the top row of panels with a schematic interpretation of the major onlapping and truncated reflectors, and unconformities, below. Arrows labeled “PS” indicate strong (dark) reflectors that are possible paleoshoreline features.

Figure 10. Core stratigraphies and images from Hidden Lake in northern Colorado. Thin lines show the core water content, and dark lines show bulk density based on gamma-ray attenuation. Dots indicate the location of AMS radiocarbon samples. Italics denotes dates based on macrofossil material or sedimentary charcoal; median calibrated ages are listed. Gray bars marks the position of the layer of low-organic content that correlates with the lower (more prominent) of the two potential paleoshorelines in the radar profiles (“PS” in Figure 9).
4.2.4. Inferred change in lake level

The Hidden Lake data track a progressive rise in lake levels since ca. 4400 cal yr BP, which was interrupted by an abrupt, sub-millennial high stand centered at about 4000 cal yr BP and a brief Medieval-age low stand after ca. 850 cal yr BP (Figure 11). Sediments dating to the 19th and 20th centuries extend higher in the lake basin than any earlier sediment and may indicate that Hidden Lake is currently higher than any time in the Holocene. The basal ages of cores C, E, and I indicate that the sediment limit was also much higher in the past 4400 years than during earlier portions of the Holocene. Indeed, the basin may have been desiccated before ca. 8400 cal yr BP based on the basal age from core I (Table 1). The depth difference between the base of core C and modern sediment deposition at the top of core H indicates that wave base, and by association the water surface, have risen by >340 cm since before 4400 cal yr BP.

Based on cores C and E, and the lack of onlapping at the base of the GPR profiles (Figure 9), water levels rose rapidly at first: the sediment limit rose >120 cm between 4400 and 4200 cal yr BP. The rise does not appear to have reached above the elevation of core F (450 cm bmls) given the absence of facies 1 below the gray layer (facades 2) at the base of that core (Figure 10). The rise marks the beginning of a ~700-yr-long high stand that lasted until ca. 3700 cal yr BP. The overlapping ages below the gray layer in cores E and C are consistent with a rapid decline in water level of >120 cm at the end of the high stand. The truncation of the lowermost reflectors in the GPR profiles (Figure 9), increases in core sediment densities (Figure 10), and declines in sedimentation rates (Figure 11) show that wave base fell below the location of core C by 3700 cal yr BP. After the end of the low stand, wave base rose by >125 cm to allow organic deposition (a band of facies 1) in core E by 2020 cal yr BP, but probably declined again by 1220 cal yr BP based on slow sedimentation rates and a second band of facies 2.

Similar features may indicate that the lake experienced at least one additional low stand during Medieval times between ca. 850 and 490 cal yr BP. The sedimentary evidence for a Medieval low stand is less pronounced than the low stand after ca. 3700 cal yr BP, but seems consistent with a drop in wave base of >50 cm given the differences in the depths of the unconformity (as recorded by density peaks) in cores F and E (Figure 11). The timing of the low stand is consistent with low-water levels at Mono Lake, California, after ca. 840 cal yr BP (Stine, 1994), and with tree-ring evidence for extensive drought in the western U.S. between ca. 1150 and 650 cal yr BP (Cook et al., 2004).

4.3. Little Molas Lake, Southern Colorado

4.3.1. GPR data

Four major sedimentary units appear in the GPR profiles from Little Molas Lake (Figures 3 & 4). The lowermost unit (Unit 4, Figure 4) contains few obvious reflectors and does not extend as close to shore as more recent sediments. Unit 4 also laps onto basement material both near shore and near the top of a mid-lake outcrop (at right in Figure 3). Unit 3 is well stratified with multiple reflective bands. It appears to conformably overlie unit 4, except near shore and near the mid-lake basement outcrop, where the base of unit 3 onlaps over the top of unit 4. Unit 2 has three sub-units: a section of highly reflective strata between two sections without strong reflectors. The reflectors of unit 2 lap over the top of unit 3.

Unit 1 (‘Late-Holocene sediments’ in Figure 4) extends higher along the lake margin and over the mid-lake outcrop than any of the older units. Units 2–4 do not cover the top of the mid-lake outcrop, and an unconformity truncates units 2 and 3 near shore. Unit 1 overlies the near-shore unconformity and the top of the bedrock outcrop. An erosional unconformity does not, however, appear to truncate units 2 and 3 near the mid-lake outcrop. The highest extent of units 2 and 3 just below the top of the outcrop is approximately the same elevation as the lowest elevation of the near-shore unconformity (Figure 4).

4.3.2. Core data

Cores B, C, and D contain similar and complete sequences from water depths of 3–1.5 m and appear to include continuous accumulation since the formation of Little Molas Lake. Core E, however, penetrates through the near-shore unconformity in 1 m of water, and core F in 69 cm of water contains only unit 1 where it extends beyond the older units.

Cores E, D, and C contain highly similar magnetic susceptibility and bulk density stratigraphies (Figure 12). The lowermost sediments in all four cores are light pink silts and clays that are dense and have high-magnetic susceptibility. These sediments likely correspond to unit 4 in the GPR profiles (Figure 4).

Organic-rich banded silts with multiple fibrous intervals comprise the remainder of the four cores. A rise in bulk density at a depth of 228–224 cm in cores C and D and at a depth of 195 cm in core E can be tracked among the cores, as can four small susceptibility peaks (gray lines, Figure 12). Above these features (at 85 cm in core C, 80 cm in core D, and 50 cm in core E), susceptibility declines conspicuously. The low susceptibility sediments have similar thicknesses of about 50 cm in cores D and C but are reduced in thickness to 25 cm in core E (Figure 12). No unconformity surface is apparent in core E, but the transition from uniformly thick features among cores to the thin layer in core E (Figure 12) matches with the truncation of unit 2 in the GPR data (Figure 4).

Figure 11. Lake-level interpretation and summary of core data from Hidden Lake. Symbol locations indicate linearly-interpolated ages between age control points and depth below the modern lake surface. Symbol size indicates percent loss-on-ignition at 550 °C. The arrow labeled “MD” marks the lake low-stand associated with a Medieval drought. Gray shaded area represents the area inferred to have been in the zone of sediment accumulation below the wave base, based on the presence of sediment in the cores. Ages of core F sediments were inferred by stratigraphic correlation to core E. Dashed lines at base of core A indicate continued but uncollected sedimentation.
A rise in susceptibility marks the uppermost sediments in cores E, D, and C (Figure 12), and probably corresponds to unit 1 in the GPR profiles (Figure 4). Notably, the thickness of the uppermost sediments (25–35 cm) is again similar in cores E, D, and C (Figure 12).

Core F is short (25 cm long) and contains dark silts that are denser and have higher susceptibility than the sediments of the other cores (Figure 12). The bulk density declines throughout the core, but susceptibility increases to a peak at 20 cm and then declines until a second rise begins in the upper 10 cm.

4.3.3. Chronology
A basal AMS date from core F on a wood fragment from 25 cm calibrates to 1695–1620 cal yr BP. A second AMS date on macrofossil material at 221 cm near the base of the organic silts in core E calibrates to 10,510–10,300 cal yr BP (Table 1; Figure 12). Based on these two dates, the average net sedimentation rates for cores F and E are 15.1 and 21.1 cm/ka respectively. Therefore, the top ~35 cm of core E, which contain the uppermost susceptibility rise, accumulated during the time represented by core F.

4.3.4. Inferred change in lake level
The sediments in core F represent a significant rise in lake level. Wave base likely rose by >30 cm before ca. 1650 cal yr BP because the base of core F is ~30 cm shallower than the correlative sediments in core E (Figure 12b). As the sediment limit shifted higher than before, sediment began to accumulate at the location of core F and the sediments became more organic and fine grained, and thus less dense and magnetic, over time. The near-shore extent of unit 1 in the GPR profiles beyond the uppermost extent of older units (Figure 4) matches with this interpretation. The uppermost rise in susceptibility in cores E, C, and D may indicate a correlative increase in streamflow into the lake and an increase in the deposition of volcanically-derived minerals in the past 1650 years. The evidence for increased runoff and high water levels during this period are consistent with a trend toward mesic vegetation recorded in fossil pollen data (Toney and Anderson, 2006).

The extent of the sediment units on the mid-lake outcrop (Figure 3) is inconsistent with previous periods of when wave-base elevations equaled or exceeded those of the past 1650 yrs. GPR units 2 and 3 did not accumulate on top of the outcrop and show no evidence of having been eroded from there; these units lap onto the outcrop with their uppermost reflectors extending to just below the top of the basement material. Therefore, wave base was probably at or below the top of the outcrop before ca. 1650 cal yr BP.

The unconformity near shore below Unit 1 (Figure 4) may have resulted from sediment infilling near shore or a period of extremely low-water levels before ca. 1650 cal yr BP, but the later would have also produced an unconformity over the mid-lake outcrop (which is not evident). However, the onlapping at the base of units 2 and 3 (Figures 3 & 4) may indicate earlier declines in wave-base elevation. The limited shoreward extent of unit 4 (Figure 4) shows that wave base was proba-
bly several meters lower than today before ca. 10,500 cal yr BP. The core and GPR data show no evidence for water-level changes after ca. 1650 cal yr BP, despite evidence for dry Medieval conditions at other near-by lakes (Davis, 1994). The small amplitude of the changes at Little Molas Lake likely resulted from the ability of the lake to overflow; the observed change in wave base may indicate changes from seasonal or intermittent overflow to constant overflow.

4.4. Other direct evidence of Holocene lake-level change in the Rocky Mountains

Foy, Hidden, and Little Molas lakes show stratigraphic evidence for high water levels in the past few millennia. Sediment cores and sub-surface profiles from other lakes throughout the Rocky Mountain region also contain similar evidence (Figure 13). The data appear consistent among lakes both within Cayan’s (1996) regions (Figure 1), indicating that our study sites may be regionally representative, and across the Rocky Mountains as a whole. Drier-than-modern conditions appear evident at all sites between ca. 7000 and 4500 cal yr BP (Figure 13).

Near Foy Lake, Hofmann et al. (2006) used seismic profiles to infer a late-Holocene (<3000 cal yr BP) high stand of Flathead Lake. Like Foy Lake, low levels during the mid-Holocene began after the deposition of the MZ ash. In central Montana, geochemical analysis of carbonates at Jones Lake shows a transition from aragonite-dominated sediments to calcite-dominated sediments beginning at ca. 3100 cal yr BP and likely indicates a late-Holocene rise in lake level (Shapley, 2005). The last occurrence of aragonite at Jones Lake follows a multi-step decline that ended by ca. 1400 cal yr BP, and thus parallels the sedimentary evidence for lake-level rise at Foy by 2800 cal yr BP and the diatom evidence for the highest levels at Foy Lake after ca. 1250 cal yr BP (Stevens et al., 2006).

In Wyoming, but also within Cayan’s (1996) “Idaho” pattern of snow variability (Figure 1), three kettle lakes in the Wind River Range show evidence of low-water levels during the mid-Holocene. Two contain sedimentary hiatuses at 8000-3100 and 7500–5700 cal yr BP respectively, and the third lake did not begin to accumulate sediment until after 5700 cal yr BP (Lynch, 1998). Seismic-reflection profiles and paleobiotic data from Yellowstone Lake, 100 km northwest of the Wind River kettle lakes, may also indicate low-water levels during the mid-Holocene (Otis et al., 1977; Ariztegui et al., 2000; Johnson et al., 2003; Theriot et al., 2006; Morgan et al., 2007). The profiles show onlapping of recent sediments and submerged shoreline features (Johnson et al., 2003; Morgan et al., 2007), which cores confirm date after ca. 7000 cal yr BP (Ariztegui et al., 2000). Pierce et al. (2002) obtained a radiocarbon age of 3835 cal yr BP near the top of a submerged beach sand 5 m below the modern lake surface. Unfortunately, the volcanic and tectonic history of the Yellowstone Caldera complicates the climatic interpretation of these data.

In central Colorado within the region of snow variability that includes Hidden Lake, Cottonwood Pass Lake contains peat layers that date to ca. 6200–5000 cal yr BP suggesting shallow or wetland conditions. Near-by wetlands also accumulated sediment only in the past 5400 and 2900 years as though they had been dry until the later half of the Holocene (Fall, 1997). In northern Arizona, and within the region of snow variability that includes Little Molas Lake, Weng and Jackson (1999) report hiatuses attributable to low-lake levels at Bear Lake from 7400 and 4230 cal yr BP and at Fracas Lake from 8500 and 2000 cal yr BP. Cores from Potato and Walker Lakes in Arizona also contain evidence of desiccation between 11,500 and 3000 and at ca. 6000 cal yr BP respectively (Hevly, 1985; Anderson, 1993). Similarly, the core from Montezuma’s Well, Arizona, contains intervals of gravel and shallow-water macrofossils that date from >8000 to 1500 cal yr BP and interrupt otherwise fine-grained organic sediment accumulation (Davis and Shafer, 1992).

Figure 13. Summary of regional lake-level data over the past 13,000 years.

Gray lines show the inferred changes in wave-base elevation (WBE), and by association, lake level at Foy, Hidden, and Little Molas Lakes; dashed lines indicate when lower-than-modern levels are inferred but the specifics of any additional fluctuations have not been constrained. Black dots indicate the ages of radiocarbon dates at the base of late-Holocene high-stand sediments. Numbered bars indicate the timing of low-lake stands at additional study sites, which are labeled as in Figure 1a: 1, Flathead Lake, MT (Hofmann et al., 2006); 2, Jones Lake, MT (Shapley, 2005); 4, Park Pond 2, WY (Lynch, 1998); 5, Forest Pond 2, WY (Lynch, 1998); 6, Park Pond 1, WY (Lynch, 1998); 7, Cottonwood Pass Pond, CO (Fall, 1997); 8, Red Lady Fen, CO (Fall, 1997); 9, Red Well Fen, CO (Fall, 1997); 10, Bear Lake, AZ (Weng and Jackson, 1999); 11, Fracas Lake, AZ (Weng and Jackson, 1999); 13, Montezuma Well, AZ (Davis and Shafer, 1992); 14, Potato Lake, AZ (Anderson, 1993). The timing of low stands at Yellowstone Lake, WY (Otis et al., 1977; Figure 1a site 3), and Walker Lake, AZ (Hevly, 1985; Figure 1a site 12) is not shown because it is poorly constrained.
5. Discussion

5.1. What long-term hydrologic trends are evident during the Holocene?

5.1.1. Orbital-scale trends

We find evidence that mid-Holocene lake levels were lower than today at a wide range of settings (Figure 13). Differences among our records indicate that moisture trends were spatially patterned and modulated by other local factors (i.e., hydrology) and short-term variability, but important long-term trends existed throughout the region. The trends are consistent with the effects of orbital forcing, especially given that mid-latitude lake-level responses to the seasonal progression of orbital forcing can result in large water-level rises in the last few millennia (Shuman and Donnelly, 2006). In particular, the regional rise in water levels by ca. 2000 cal yr BP is consistent with the persistence of >20 W m⁻² insolation anomalies in early autumn until after 3000 yrs BP (Berger, 1978). Summer insolation anomalies declined steadily over the course of the Holocene, but September and October insolation anomalies increased until 6000–4000 cal yr BP. The autumn insolation anomalies are directly relevant to the lake-level record because they affect a portion of the year when the balance of precipitation and evapotranspiration shifts from negative to positive today (Figure 2). The insolation anomalies were also important for generating atmospheric dynamics that probably limited precipitation rates during much of the Holocene (e.g., Harrison et al., 2003; Diffenbaugh et al., 2006) and for affecting temperatures in a manner that may have reduced recharge from spring snowmelt (e.g., Bartlein et al., 1998; Stewart et al., 2004); see further discussion below.

The sedimentary signatures of low-lake stands before ca. 2000 yrs BP (Figures 5, 8, & 13) are also more pronounced than stratigraphic features in our study lakes associated with recent decadal-to-centennial droughts (e.g., Stine, 1994; Cook et al., 1999, 2004; MacDonald, 2007; Meko et al., 2007). Lake levels rose >6 m from 6000 yrs BP to today at Foy Lake (Figure 7) and >3 m at Hidden Lake (Figure 9), but likely fluctuated much less during the last millennium. At Hidden Lake, potential stratigraphic evidence for a recent low-stan date to the Medieval period, but the associated lake lowering would have been substantially less than occurred before 4400 cal yr BP and between 3700 and 1200 cal yr BP. At Little Molas and Foy Lakes, water levels may have continued to rise until after ca. 1600 and 800 cal yr BP respectively, but we do not find stratigraphic evidence for pronounced fluctuations in water levels at these sites during the last millennium.

The blunt character of sedimentary records provides a “filter” that is useful for gauging the relative severity of different climatic events. Although short-term droughts were not strong enough to generate an obvious sedimentary record, whereas changes forced by boundary-condition changes over the course of the Holocene passed the sedimentary “filter” of climatic signals and produced obvious stratigraphic evidence. Therefore, although “megadroughts” are part of the natural variability of the late-Holocene (e.g., Cook et al., 2004), changes in the large-scale controls on the climate system in the past (i.e., insolation) and, by analogy, in the future (i.e., greenhouse gases) have the potential to drive regional water levels well beyond the envelope of variability that existed during the past millennium.

5.1.2. Millennial-to-centennial variations

Some moisture changes that took place over millennial-to-centennial time scales were also large enough to pass through the sedimentary “filter” and produce direct stratigraphic evidence. Although the regional patterning of the millennial-to-centennial record is hard to evaluate because of the small number of radiocarbon ages at Foy and Little Molas Lakes, some millennial-scale patterns can be inferred. For example, multiple dates from Hidden Lake reveal that the lake was high at 4400–3800 cal yr BP and that a rapid decline of >120 cm in less than a century followed (Figure 11). The high stand correlates with evidence for a period of high water levels between 4500 and 3500 cal yr BP inferred from diatom assemblages at Foy Lake (Stone and Fritz, 2006). Likewise, Miao et al. (2007) summarize the dated evidence of loess deposition in Nebraska at sites ~450–750 km east of Hidden Lake, and include dates of 4810–4450 and 5060–4480 cal yr BP on paleosol formation during a period of minimal dune activity (Miao et al., 2007). These ages are slightly older than the ages of 4420–4300 and 4285–4160 cal yr BP, which mark the beginning of the high water levels at Hidden Lake, but the subsequent period of maximum loess deposition (above the paleosols) in Nebraska dates from 3890 to 2430 cal yr BP when Hidden Lake was low (Figures 10 & 11). Therefore, Hidden Lake may have been high when paleosols developed in Nebraska, and low during the periods of maximum aeolian activity. Other episodes of dune and loess activity date to 9400–6600 and 770–570 cal yr BP (Miao et al., 2007) when Hidden Lake was also low (Figure 11).

Some dune data contrast with the Hidden Lake record, however. A period of enhanced aridity has been inferred from dune activity at 4300–4100 cal yr BP in the Ferris Dunes of central Wyoming (Stokes and Gaylord, 1993). The Ferris Dunes are located 180 km north of Hidden Lake, and contain ~10 m of aeolian deposition associated with ages of 4340–4040 cal yr BP, and interbedded interdune deposits that date from 8290 to 5650 cal yr BP (Stokes and Gaylord, 1993; Forman et al., 2001). Based on these ages, little dune activity occurred in central Wyoming when Hidden Lake was low, and extensive dune activity took place when the lake was high. The regional contrast between two moisture indicators (lake levels and dunes) may depend, in part, on differences in seasonal moisture influences on lakes and dunes. Dune activity depends on vegetation cover controlled by summer moisture, as well as wind (Forman et al., 1995, 2001), but lakes may rely heavily on winter precipitation (Figure 2).

5.1.3. Comparison with other proxies

Like the dune data, fossil pollen and plant macrofossil records confirm that lake-level trends do not fully represent all aspects of Holocene hydroclimatology. Specifically, the lake-level data from the southern Rocky Mountains differ from other evidence from Colorado and Arizona that indicates wetter-than-modern conditions in the early to mid-Holocene as a result of an insolation-enhanced southwestern monsoon (Beinroth et al., 1990; Davis and Shafer, 1992; Thompson et al., 1993; Fall, 1997; Mock and Brunelle-Daines, 1999; Harrison et al., 2003). In Colorado, forests expanded downslope into new dry valleys in Colorado in response to wet conditions before ca. 6000 yrs BP (Fall, 1997; Markgraf and Scott, 1981; Mayer et al., 2005). Mayer et al. (2005) found evidence of forest parkland during the early Holocene where sagebrush steppe grows today in Middle Park, Colorado, ~45 km from Hidden Lake, but Hidden Lake was low at the time based on the lack of nearshore sediments older than ca. 4400 cal yr BP (Figures 10 & 11). Little Molas Lakes and other lakes in the southwest U.S. (Figure 13) were also lower than today during the same period.

The coincidence between forested valleys, inactive Wyoming dunes (see Section 5.1.2), and low lakes may be explained by a high-relative importance of winter (non-growing season) moisture for the lakes (e.g., Prentice et al., 1992). Forest extent and dune activity, by contrast, may have depended in large part
upon spring and summer precipitation (Forman et al., 1995, 2001; Weltzin and McPherson, 2000). Isotope data from southern Colorado near Little Molas Lake indicate that summer precipitation probably decreased relative to winter precipitation over the course of the Holocene (Friedman et al., 1988). Likewise, climate models simulate lower snowpack than today in the early- and mid-Holocene (Bartlein et al., 1998), and the absence of glacial activity in Colorado between ca. 8000 and 1000 cal yr BP (Benson et al., 2007) indicates long-term periods of low snow accumulation. Therefore, in addition to the direct effects of insolation anomalies on evaporation rates (discussed above in Section 5.1.1), the long-term lake-level trends in Colorado may be related to an increase in winter precipitation, as well as changes in the volume, rate, and seasonal timing of snowmelt.

In other portions of the Rocky Mountains, and even locally within Colorado, vegetation trends track moisture-level changes that are similar to those inferred from the lake-level histories. High-elevation forests near Little Molas Lake became more mesic in the past few millennia when the lake level was high (Toney and Anderson, 2006). Vegetation trends are also consistent with our lake-level reconstruction at Foy Lake because local forests became closed and mesophytic tree taxa, such as fir (Abies) and spruce (Picea), increased in abundance after 2200 cal yr BP (Power et al., 2006). A trend back toward more open vegetation near Foy Lake in the past 700 years could indicate local human activity, low annual moisture balance, or a shift in the seasonal timing of precipitation (Power et al., 2006). The later two possibilities are not recorded in the sedimentary record of lake-level change (including diatom and geochemical data; Stevens et al., 2006). More work is needed to fully understand the seasonal climatic patterns and other factors that explain both spatially-variable fire and vegetation trends (e.g., Thompson and Anderson, 2000; Whitlock and Bartlein, 2004; Brunelle et al., 2005) and the lake-level trends documented here.

5.2. What spatial patterns characterize the long-term trends in moisture variation?

5.2.1. Comparison with patterns of interannual-to-multidecadal variability

Many short-term moisture patterns, including snowpack variability (Figure 1), emphasize a north–south contrast in the Rocky Mountain region (e.g., Trenberth et al., 1988; Cayan, 1996; Mock, 1996; Dettinger et al., 1998; McCabe et al., 2004). Likewise, comparison of lake-level histories in the Great Basin has shown variable synoptic responses to climate forcing (Licciardi, 2001). The rise in lake levels at long-time scales documented here, however, is evident throughout the Rocky Mountains; all of the fourteen lakes discussed were lower than today ca. 7000-4500 cal yrs BP (Figure 13). To the west, throughout the Great Basin, both small and large lake systems also record evidence of aridity before ca. 3000 yrs BP (e.g., Quade et al., 1998; Broughton et al., 2000; Benson et al., 2002; Briggs et al., 2005). Therefore, lakes throughout much of the interior of the western U.S. do not show the spatial variability that exists at short time scales, but rather widespread aridity during the mid-Holocene.

This inference is significant because some datasets indicate meaningful changes in Pacific SST patterns during the Holocene that are similar to those associated with the El Nino Southern Oscillation (ENSO) on annual scales (Gagan et al., 1998; Overpeck and Webb, 2000; Koutavas et al., 2002). Likewise, the mechanisms that shift North American precipitation patterns on interannual time scales, such as atmospheric circulation changes associated with shifting Pacific SST gradients have probably also applied over the time scale of the Holocene (e.g., Harrison et al., 2003; Shin et al., 2006). Aridity in the Rocky Mountains, thus, could underscore the potential for large boundary-condition changes to override the influences of the internal modes of climate variability that were important during the past millennium (see e.g., Diffenbaugh et al., 2006). However, the spatial patterns of response to ENSO and other modes of Pacific SST variability may vary through time (Shinker and Bartlein, 2009), and the influences of these patterns may not generate consistently strong ‘spatial fingerprints’ at both short- and long-time scales. Indeed, some historic droughts, such as in AD 1856–1865 and AD 1890–1896, have been attributed to the effects of cool La Nina-like conditions like those inferred for the mid-Holocene (e.g., Overpeck and Webb, 2000), and have also not shown a north–south contrast in the Rocky Mountains and adjoining regions (Herweijer et al., 2006). If so, the lack of a match between the Holocene lake trends and recent patterns of variability (Figure 1) may not preclude a role for the drivers of recent variability (e.g. ENSO) over the course of the Holocene.

5.2.2. Comparison with climate model simulations

Widespread aridity in the Rocky Mountains at 6000 yrs BP is inconsistent with large wet anomalies over large areas of the Rocky Mountains in coupled ocean–atmosphere GCM simulations of this period (Harrison et al., 2003; Shin et al., 2006). The specific inclusion of La Nina-like tropical Pacific SST in one GCM generates enhanced aridity in the central U.S., including in the southern Rocky Mountains, but the sign of the regional moisture anomaly was reversed when realistic greenhouse-gas forcing (i.e., lower than preindustrial mid-Holocene levels of CO2 and CH4) was included in a fully forced and coupled experiment using the same model (Shin et al., 2006). Harrison et al. (2003) provide a conceptual model for producing aridity in the northern Rocky Mountains when ocean–atmosphere feedbacks amplify a monsoon over the southern Rocky Mountains, but their model simulations do not produce the hypothesized pattern and instead generate wet conditions across most of the central U.S. in contrast to datasets there.

The most accurate match between data and model results comes from regional-scale model simulations with more accurate topography than in the GCMs (Diffenbaugh et al., 2006). In the regional model results, seasonal insolation forcing at 6000 years BP drives a robust pattern of enhanced aridity (i.e., not significantly changed by other forcing such as SST variation) in most high-elevation areas of the Rocky Mountains (Diffenbaugh et al., 2006). The regional model’s success in simulating the moisture-balance patterns inferred from the data underscores the importance of fine-scale processes in creating the patterns of response, and indicates that the long-term lake-level trends are probably driven by a hierarchy of local to global responses to orbital change. For example, local responses probably include effects on surface energy-budgets and soil moisture, which mediate the regional-scale (e.g., topographic) modification of synoptic-scale climatic patterns.

5.3. How similar were long-term moisture variations in the Rocky Mountains and mid-continent?

Intense aridity prevailed in the Great Plains and Great Lakes region during the mid-Holocene (Webb et al., 1983; Baker et al., 1992; Forman et al., 1995, 2001; Laird et al., 1996; Grimm, 2002), and may be expected to coincide with an amplified east–west moisture gradient with wet conditions in the Rocky Mountains. A widely-stated hypothesis links dry mid-continent conditions during the mid-Holocene to the westward flow of air masses of Pacific origin (e.g., Bartlein et al., 1984; Bradbury and Dean, 1993; Yu et al., 1997; Lewis et al., 2008). According to the hypothesis, strong zonal atmospheric
flow carried the air masses into the mid-continent, where regional aridity was generated by the loss of moisture from the air masses as they travelled over the mountains in the western U.S. (Figure 1b). Our data do not support this conclusion at orbital time scales because we do not find evidence of enhanced orographic precipitation in the Rocky Mountains.

Both the mid-continent and the Rocky Mountains were drier than today during the mid-Holocene. Therefore, our data, like historic analyses and modeling efforts (e.g., Diffenbaugh et al., 2006; Shinker et al., 2006), are inconsistent with the “zonal-flow” hypothesis of mid-continent aridity. Instead, aridity in both the Rocky Mountains and mid-continent is probably the direct result of seasonal insolation anomalies (i.e., increased evapotranspiration), and a hierarchy of insolation-induced effects, including (a) broad synoptic changes that enhanced atmospheric subsidence and, thus, prohibited precipitation over the mid-continent (Harrison et al., 2003; Diffenbaugh et al., 2005), (b) a reduction of cool-season moisture advection resulting, in part, from low winter insolation, (c) reduced soil moisture (Forman et al., 2001), and (d) surface-atmosphere feedbacks like those linked to snowpack volume and extent in the southwest U.S. today (Zhu et al., 2005; McCabe and Clark, 2006).

6. Conclusions

Well-documented mid-Holocene aridity in the central U.S. extended into the headwater regions of the Rocky Mountains, and demonstrates a high sensitivity of key water supplies to global changes. The stratigraphies of Foy, Hidden, and Little Molas Lakes show consistent evidence of high lake levels during the past two millennia, and low-lake levels throughout the majority of the Holocene. Aside from these long-term trends, however, the records differ significantly with unique millennial-scale fluctuations documented by stratigraphic features at Foy and Hidden Lakes. The Foy Lake stratigraphy indicates a low stand that may date to the Younger Dryas chronozone (specifically 12,600–11,560 cal yr BP); sediments at Hidden Lake indicate an abruptly-initiated and -terminated high stand between 4400 and 3800 cal yr BP. Only Hidden Lake contains evidence of lake-level change resulting from the severe droughts of the last millennium, but the associated stratigraphic evidence is less pronounced than the evidence for earlier low-lake levels.

Long-term changes in the hydroclimate of the Rocky Mountains, as inferred from the lake levels, likely arose from seasonal insolation anomalies and related atmospheric processes. However, the success of regional climate models for simulating these patterns (in contrast to GCMs) indicates that local-to-regional processes were probably also important. The trends indicate that changes in the broad-scale controls on climate can significantly alter the availability of water in the Rocky Mountain region and associated river basins. Large shifts in boundary-condition changes in the future (i.e., greenhouse-gas concentrations) could, therefore, also produce dramatic hydrologic impacts.

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