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Evidence of temperature depression and hydrological variations in the eastern Sierra Nevada during the Younger Dryas stade

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Abstract

Sediment records from two lakes in the east-central Sierra Nevada, California, provide evidence of cooling and hydrological shifts during the Younger Dryas stade (YD; ~12,900–11,500 cal yr BP). A chironomid transfer function suggests that lake-water temperatures were depressed by 2°C to 4°C relative to maximum temperatures during the preceding Bølling–Allerød interstade (BA; ~14,500–12,900 cal yr BP). Diatom and stable isotope records suggest dry conditions during the latter part of the BA interstade and development of relatively moist conditions during the initiation of the YD stade, with a reversion to drier conditions later in the YD. These paleohydrological inferences correlate with similar timed changes detected in the adjacent Great Basin. Vegetation response during the YD stade includes the development of more open and xeric vegetation toward the end of the YD. The new records support linkages between the North Atlantic, the North Pacific, and widespread YD cooling in western North America, but they also suggest complex hydrological influences. Shifting hydrological conditions and relatively muted vegetation changes may explain the previous lack of evidence for the YD stade in the Sierra Nevada and the discordance in some paleohydrological and glacial records of the YD stade from the western United States.

Keywords: Quaternary; Younger Dryas stade; Climate change; Vegetation change; Sierra Nevada; California; Paleoclimatology; Paleohydrology; Paleolimnology

Introduction

Although the cause of the Younger Dryas stade (YD; ~12,900–11,500 cal yr BP) remains controversial (Broecker, 2003; Lowell et al., 2005; Carlson et al., 2007; Firestone et al., 2007), records from the North Atlantic indicate a significant freshwater flux, rapid interruption in Atlantic meridional circulation, a decrease in sea-surface temperatures (SSTs) and a depression of air temperatures in Greenland and the circum-North Atlantic region during that time period (Björck et al., 1996; Alley, 2000; Rutter et al., 2000; McManus et al., 2004; Carlson et al., 2007). Strong North Atlantic cooling should be rapidly teleconnected to the North Pacific region (Mikolajewicz et al., 1997). Indeed,

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paleooceanographic records from the northeastern Pacific indicate a sharp drop in SSTs of 3°C to 8°C and abrupt changes in circulation during the YD stade (Hendy et al., 2002; Barron et al., 2003).

Terrestrial records from California and adjacent areas are mixed with respect to the effects of the YD stade. Whereas pollen stratigraphies from marine cores off northern California and lake sediment cores from Oregon provide evidence of vegetation changes consistent with YD stade cooling (Grigg and Whitlock, 1998; Barron et al., 2003), pollen records from the Sierra Nevada have not shown evidence of a significant climatic reversal (e.g., Adam, 1985; Anderson, 1990; Smith and Anderson, 1992; Koehler and Anderson, 1994). Although there is evidence for linkages between North Atlantic cooling during Heinrich events and glacial advances in the Sierra Nevada (Phillips et al., 1996), evidence for a glacial advance during the YD stade is lacking (Clark and Gillespie, 1997; Licciardi et al., 2004; James et al., 2002; Gillespie and Zehfuss, 2004). Clark and Gillespie (1997) concluded that by 15,000–14,000 cal yr BP glaciers were restricted to the high Sierra Nevada and the somewhat limited Recess Peak readvance during the late glacial ended by about 13,100±85 cal yr BP. Whereas some lake-level records from eastern California and Oregon suggest dry conditions during the YD stade (Benson et al., 1997; Licciardi, 2001; Bacon et al., 2006), other records from the Great Basin indicate moist conditions (Thompson, 1992; Phillips et al., 1994; Quade et al., 1998; Liu et al., 2000; Huckleberry et al., 2001; Briggs et al., 2005; Oviatt et al., 2005).

We present new paleoenvironmental evidence regarding the YD stade from two small lakes in the east-central Sierra Nevada. Although previously published pollen records from these lakes do not provide indications of vegetation changes during the YD stade (Anderson, 1990), we reasoned that multi-proxy records with higher temporal resolution might present a fuller picture of environmental changes in this portion of the Sierra Nevada. In addition to pollen, we analyzed sediment organic content, $\delta^{13}C$ and $\delta^{18}O$ of sediment organics, chironomids, diatoms and charcoal. The resulting records provide evidence of depressed summer temperatures and a series of hydrological shifts coincidental with the YD stade. The timing and nature of the hydrological shifts appear to correspond generally with shifts in lake level recorded at Owens Lake, which lies south-southeast of the study area in the Great Basin.

Study sites and methods

Lake Barrett (37.596°N, 119.007°W) is located at 2816 m a.s.l., east of the Sierra Nevada crest (Figs. 1 and 2). The 1.8-ha lake consists of two basins with a maximum water depth in the south basin of 6.1 m. Starkweather Lake (37.663°N, 119.074°W) is located just west of the crest at an elevation of 2424 m a.s.l. (Figs. 1 and 2). The 1.1-ha lake has a maximum depth of 11.0 m. Both lakes have potential surface outflow channels, but water was not observed flowing out of either lake during the summer field seasons. We did not observe obvious higher shorelines or other geomorphological evidence of significantly higher sustained lake levels.

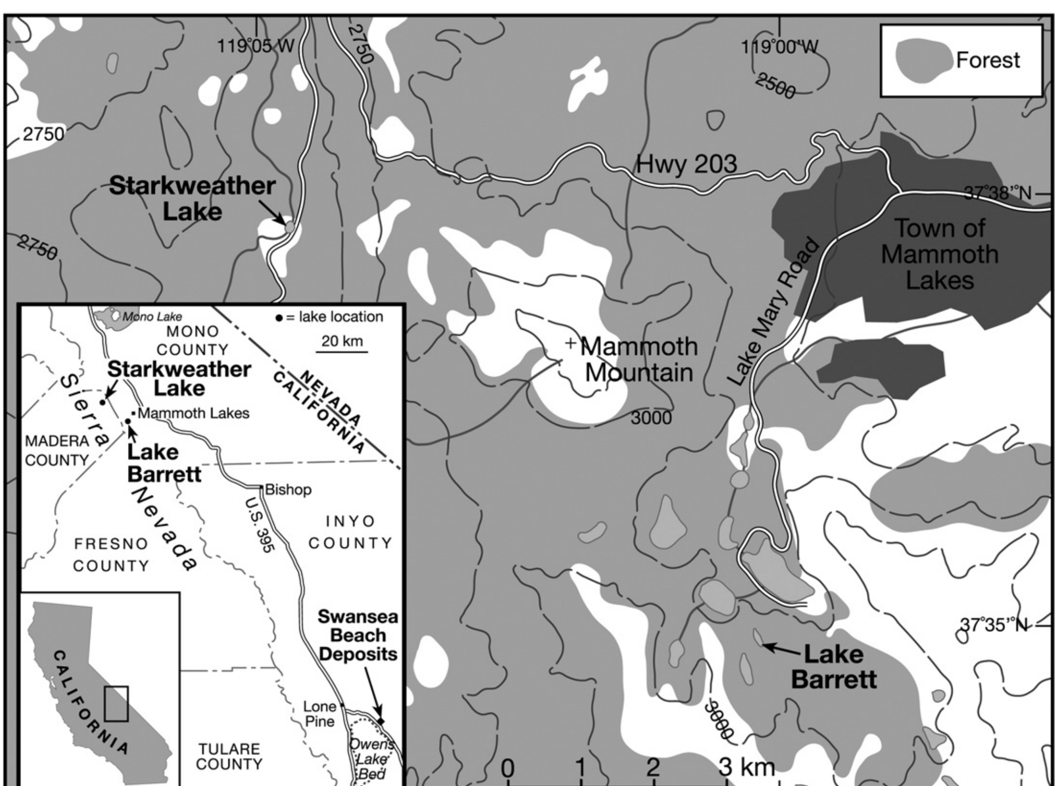

Figure 1. The eastern Sierra Nevada—southwest Great Basin study area with the locations of Lake Barrett, Starkweather Lake and the Swansea shoreline at Owens Lake.

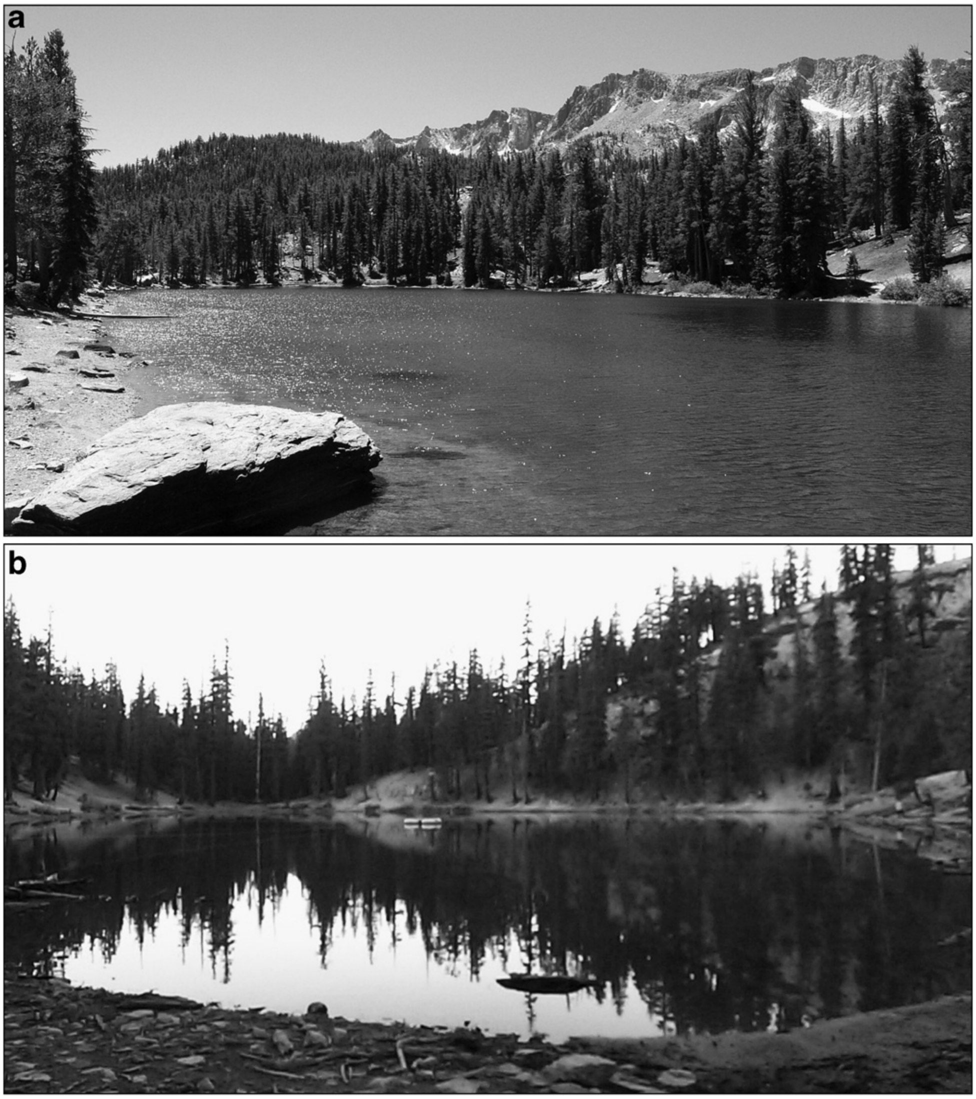

Figure 2. Photographs of Lake Barrett (a) and Starkweather Lake (b).

Maximum elevations along Sierra Nevada crest in the vicinity of the lakes reach over 3300 m at the summit of Mammoth Mountain (Fig. 1). The bedrock of the region includes Mesozoic metavolcanic and granitic rocks, and extensive late Cenozoic rhyolite, basalt and quartz latite (Huber and Rinehart, 1965; Hill et al., 1985; Bailey, 1989). Surficial deposits of Pleistocene glacial sediments and Holocene volcanic ash are present within the lake catchments. Although Mammoth Mountain is a lava-dome complex, the upper ash layers in the region were largely deposited by the eruption of the Inyo and Mono Craters to the north (reviewed in Anderson, 1990; Hildreth, 2004). The lake basins are likely glacial in origin and related to the late Wisconsin-age Tioga glaciation. The lakes lie beyond the ice limits of the subsequent late glacial Recess Peak glaciation and the Holocene-aged Matthes glaciation (Bailey, 1989; Burbank, 1991; Clark and Gillespie, 1997; Gillespie and Zehfuss, 2004). Cirques are present at upper elevations in the surrounding area and within 2 km of Lake Barrett (Fig. 1). The local areas of the lakes and their watersheds do not contain glaciers at present.

The vegetation in the vicinity includes open sagebrush typical of the Great Basin at lower elevations to the east of Lake Barrett, conifer forest at the elevations of the lakes, and areas of open alpine tundra at elevations greater than 3000 m (Figs. 1 and 2). Barrett Lake is surrounded by mature but relatively open forest composed of *Pinus contorta* (lodgepole pine) and *P. monticola* (western white pine) with *Tsuga mertensiana* (mountain hemlock) interspersed. Anderson (1990) reported finding *P. flexilis* (limber pine) near the site as well. Conifer forest surrounding Starkweather Lake includes *Abies magnifica* (red fir), *P. contorta*, *P. monticola*, scattered *T. mertensiana* and some *Juniperus occidentalis* (sierra juniper). Average January and July temperatures reported at the USFS station in nearby Mammoth Lakes (Station 045280: 1993–2007) are −2°C and 17°C, respectively. The mean annual precipitation over the period of available record is 64 cm.

Sediment was recovered from Lake Barrett (July 16, 2000) and Starkweather Lake (October 13, 2001) using a modified Livingston piston corer (Wright, 1991). A 201-cm-long core was obtained from Lake Barrett and a 371-cm-long core was obtained from Starkweather Lake. At both lakes, coring ceased when

Table 1
Radiocarbon dating results and calibrations

Lake	Core depth (cm)	Material	Lab number	Age (^{14}C yr BP)	Calibration y-intercept (cal yr BP)	2σ age range (cal yr BP)	Relative area under distribution
Lake Barrett							
	146.0	Twig fragment	Beta-153338	7820±40	8597	8514–8722	0.986
	163.5	Pine needle	Beta-153339	9170±40	10,330	10,237–10,429	0.969
	180.5–181.0	Bulk sediment	Beta-164174	11,180±60	13,083	12,946–13,202	1.000
	196.0–197.5	Bulk sediment	NSRL-12025	11,900±65	13,765	13,616–13,931	1.000
Starkweather Lake							
	290.5–291.5	Bulk sediment	Beta-164175	9410±50	10,640	10,511–10,756	1.000
	301.0	Twig fragment	Beta-160790	10,300±70	12,108	11,805–12,392	0.993
	312.0–313.0	Bulk sediment	Beta-161881	11,490±50	13,334	13,240–13,432	1.000

resistant material, presumably bedrock or glacial debris, was encountered at the base of the soft lake sediments. Chronologies for the basal sections of the cores were developed by AMS radiocarbon dating of terrestrial plant macrofossils and bulk organic sediment (Table 1). Because these lakes occupy basins formed in areas with overwhelmingly silica-rich volcanics and aluminum- and silica-rich metavolcanic rock with negligible carbon (Dodge, 1971; Metz and Mahood, 1991), we assumed that the radiocarbon-dating uncertainties in bulk-sediment that are typically associated with carbonate rock dissolution and hard-water effects (MacDonald et al., 1991) would be negligible. Radiocarbon ages were converted to calibrated ages using CALIB 5.0.2 and the IntCal04 calibration dataset. For comparative purposes with earlier studies, CALIB 4.4 and the IntCalL98 calibration dataset were also used (Stuiver and Reimer, 1993; Stuiver et al., 1998; Reimer et al., 2004; Stuvier et al., 2005). The radiocarbon dates were converted to calibrated ages before AD 1950 (cal yr BP) following the approach of Bacon et al. (2006) and Adams (2007) to calculate a midpoint calibrated age based on the 2-sigma highest probability around the y-intercept between the conventional uncalibrated ^{14}C date and the calibration curve. Age–depth models were produced by linear interpolations between dated samples (Fig. 3). The chronologies applied to the cores are taken directly from these conservative age–depth models and do not incorporate any additional manipulation.

Loss-on-ignition (LOI) analysis was performed to examine changes in the organic content of the sediments (Heiri et al., 2001). The δ^{13}C of sediment organics and the δ^{18}O of sediment cellulose were determined on the Starkweather core using standard techniques (Wolfe et al., 2001), with additional refinements (Wolfe et al., 2007). There was insufficient sediment left from other analyses to conduct δ^{13}C and δ^{18}O analyses on the Barrett core. Lake water δ^{18}O values were reconstructed using a cellulose-water oxygen isotope fractionation factor of 1.028 (Wolfe et al., 2001; 2007). Results are expressed in standard δ-notation in per mil (‰) values relative to Vienna PeeDee Belemnite (VPDB) for δ^{13}C and Vienna Standard Mean Ocean Water (VSMOW) for δ^{18}O.

Chironomid and diatom analysis followed standard procedures (Battarbee et al., 2001; Walker, 2001) that have been previously applied in the Sierra Nevada (Porinchu et al., 2002, 2003; Bloom et al., 2003; Potito et al., 2007). A minimum of 40 chironomid head capsules and 600 diatom valves were identified in each sample. Reconstructions of summer surface water temperatures were developed using a chironomid-based inference model based on analysis of 57 lakes and 44 taxa from the Sierra Nevada (details in Porinchu et al., 2002). A one-component weighted-averaging partial least-squares model ($r^2=0.72$, $p<0.05$; root mean squared error, RMSE=1.1 °C) was applied using WA-PLS version 1.1 (Juggins and ter Braak, 1993).

Following Bloom et al. (2003), a diatom-based lake-depth inference model and an updated salinity-inference model were developed for the Sierra Nevada using C^2 version 1.3 (Juggins, 2003). The diatom-inference model for lake depth is based on weighted averaging with inverse de-shrinking and incorporates 49 lakes and 309 diatom taxa ($r^2=0.84$, $p<0.05$; RMSE=2.7 m), whereas the updated diatom-inference model for salinity is based on weighted averaging with inverse de-shrinking and incorporates

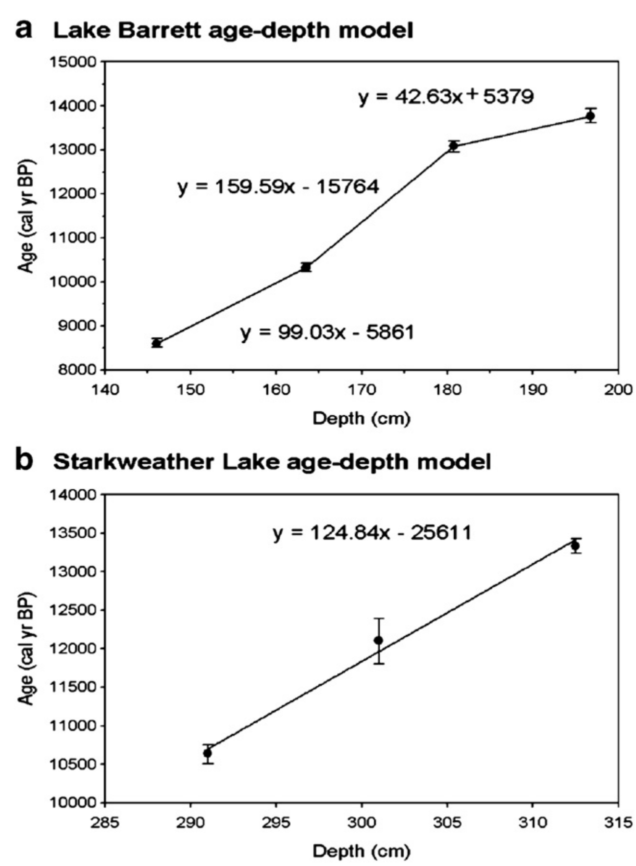

Figure 3. The age–depth models for Lake Barrett (a) and Starkweather Lake (b) with the 2σ error bars for calibrated dates.

54 lakes and 311 diatom taxa (r^2 = 0.91, p < 0.05; RMSE = 5.87 mgL^{-1}). The relatively high RMSE of the depth model suggests that these results should be considered evidence of the direction of water-level changes rather than precise depths.

Total dissolved solids in Sierra Nevada lakes and runoff are typically low (Melack et al., 1985) and provide a small range of variability for salinity reconstructions. Our reconstructed values are quite low during the YD stade and several negative values arose in the diatom-inferred salinity records due to de-shrinking (Birks, 1998). In addition, the Starkweather lake-depth and salinity reconstructions performed poorly during reliability assessment following the analogue approach described by Bradshaw et al. (2000). Therefore, we restrict our application of the diatom-based lake-depth and salinity estimates to Lake Barrett and caution that the reconstructions are best considered as indicators of general trends in hydrologic balance.

Fossil pollen samples were processed and analyzed using standard techniques (Faegri and Iversen, 1989). Pollen identifications were based on keys (Kapp et al., 2000) and UC Berkeley and UCLA reference collections. A terrestrial pollen sum of ~500 grains was counted for each sample. In addition, microscopic charcoal particles >12.5 µm in diameter were counted for each sample. The zonation of the pollen stratigraphies into pollen-assemblage zones was conducted using the CONISS program (Grimm, 1987). The program produces a stratigraphically constrained cluster analysis of pollen stratigraphic data (Grimm, 1987).

Results and interpretation

Radiocarbon dating indicates that the lower portions of the Lake Barrett and Starkweather Lake cores extend through the

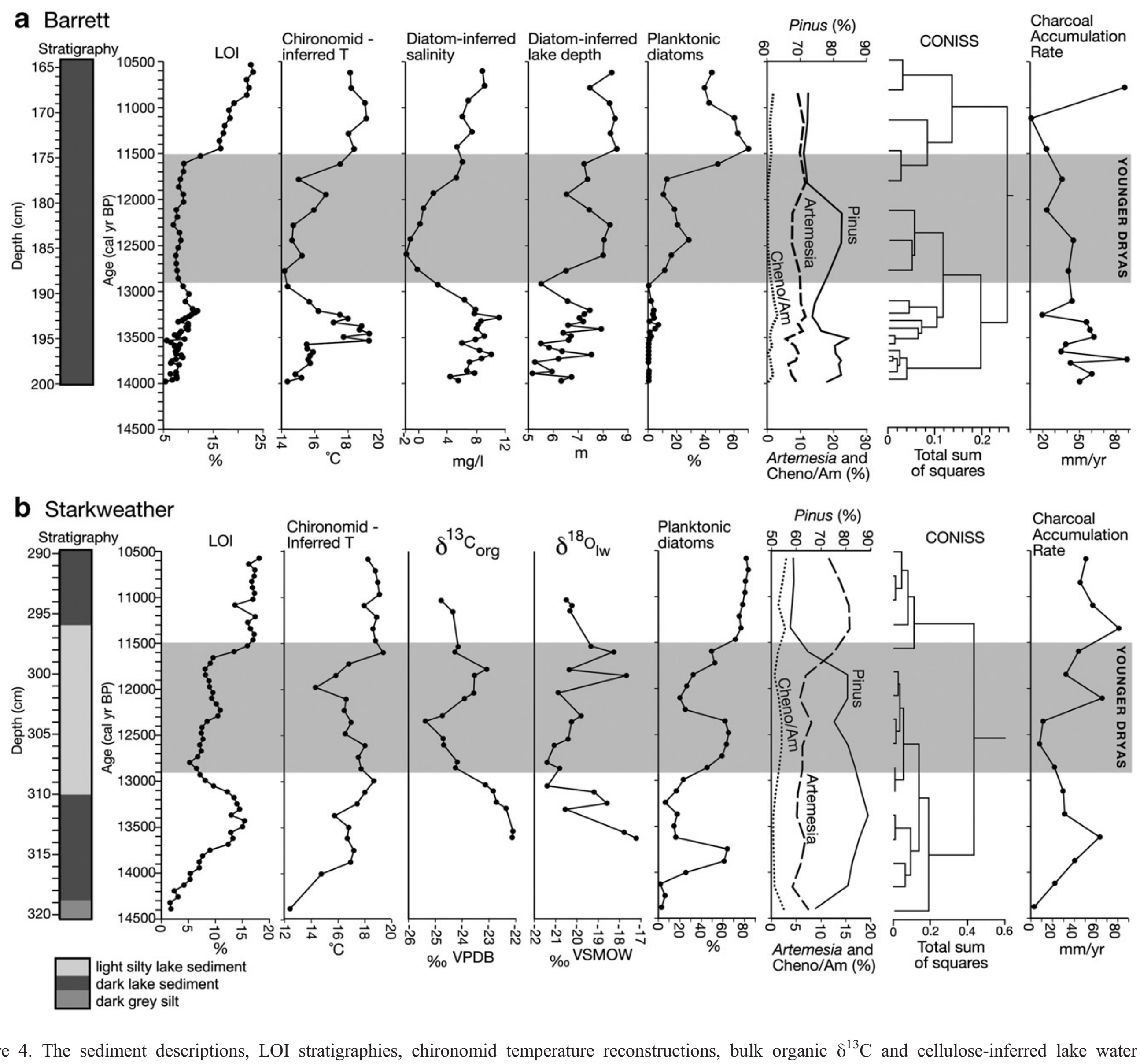

Figure 4. The sediment descriptions, LOI stratigraphies, chironomid temperature reconstructions, bulk organic $\delta^{13}C$ and cellulose-inferred lake water $\delta^{18}O$ stratigraphies, diatom salinity and depth reconstructions, planktonic diatom percentages, *Pinus, Artemesia* and Chenopodiaceae-Amaranthaceae (Cheno/Am) pollen percentages, CONISS pollen zonation (based upon all terrestrial pollen), and charcoal stratigraphies from Lake Barrett (a) and Starkweather Lake (b).

YD stade (Table 1; Fig. 3). Sedimentation rates for this time period at Lake Barrett are slightly higher than at Starkweather (Fig. 3). The ages from the bulk sediment dates appear consistent with the ages provided by the terrestrial macrofossil dates. However, because of the limited number of dates for each core, the analytical error associated with the radiocarbon dates, and the fact that dates from terrestrial macrofossils and bulk organic sediment can be prone to different sources of taphonomic uncertainty, the chronologies presented here are subject to uncertainties of a few centuries. While recognizing this uncertainty, the records from the two lakes suggest that the observed environmental changes are generally contemporaneous and consistent with the YD stade (Figs. 4 and 5).

The chironomid-based temperature estimates from the two lakes indicate a ~2°C to 4°C depression of summer water temperatures compared to the preceding maximum of the Bølling–Allerød interstade (BA; ~14,500–12,900 cal yr BP, Fig. 4). The depression in temperature is most strongly expressed in the Lake Barrett reconstruction. The LOI stratigraphies (Fig. 4) from both Barrett and Starkweather lakes indicate decreased organic content during the YD stade, consistent with cooling and a decline in lake productivity. Although the Lake Barrett sediments remain dark and relatively organic throughout the YD interval, at Starkweather the sediment deposited during the YD stade is slightly lighter than the underlying or overlying sediment.

The diatom-estimated water depths at Lake Barrett suggest an increase in lake levels at the start of the YD stade and then a decrease in the latter half of the interval (Fig. 4). The salinity reconstruction suggests a decrease in salinity in the early YD stade and a shift back towards higher salinities in the latter part of the stade (Fig. 4). The planktonic diatom percentages fluctuate in concert with the lake depth estimates at Lake Barrett and display a similarly timed pattern at Starkweather Lake (Fig. 4). The coincidental fluctuations in planktonic diatoms at both lakes suggest low water levels and relatively dry conditions during the late BA interstade, increasing moisture in the early YD stade, and a decrease in moisture in the middle to late YD stade. However, although relatively high salinity persists into the early Holocene, water levels appear to have increased again at the termination of the YD stade. The change in the relationship between lake level and salinity may be related to the seasonal distribution of insolation in the early Holocene. Davis (1999) has suggested that greater summer insolation in the early Holocene may have produced increased evaporation and summer drought, but decreased winter insolation may have produced a larger snowpack. The rapid spring melt of this snowpack may have produced greater runoff resulting in higher lake levels.

There is a decline in cellulose-inferred lake water $\delta^{18}O$ ($\delta^{18}O_{lw}$) at Starkweather Lake during the early part of the YD stade (Fig. 3). Modern runoff and unevaporated lake waters in the Sierra Nevada typically have $\delta^{18}O$ values of −15 to −17‰ (Space et al., 1991). The low values $\delta^{18}O$ of −20 to −21‰ observed at Starkweather Lake during the early part of the YD stade could reflect a decline in the temperature of condensation and/or a decline in evaporation rates in the lake itself. The general covariance between $\delta^{18}O_{lw}$ and bulk organic $\delta^{13}C$ ($\delta^{13}C_{org}$) suggests that changes in hydrologic balance played at least some role in the declining $\delta^{18}O_{lw}$ values during the YD stade. It is likely that there was a rapid flushing consistent with the lowered organic matter content, lower salinity and indications of higher lake levels during the early YD stade. The subsequent increase in $\delta^{18}O_{lw}$ and $\delta^{13}C_{org}$ values in the middle to late YD stade likely represents a shift to increased evaporative ^{18}O-enrichment and reduced flushing, consistent

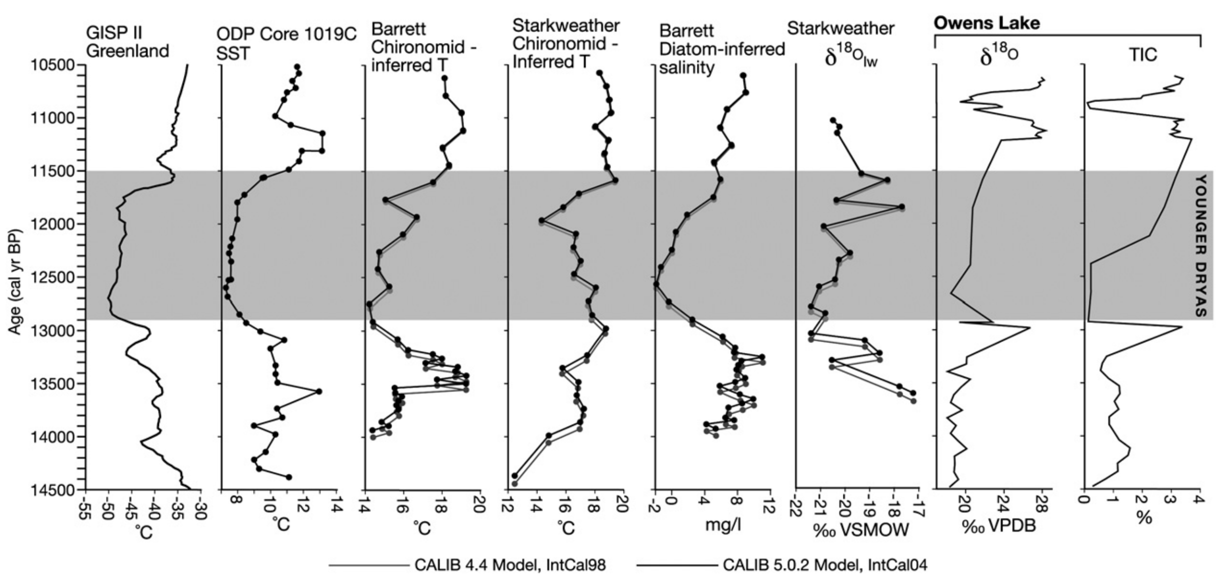

Figure 5. Comparison of the GISP II Greenland temperatures (data and chronology from Alley, 2000), the alkenone-based SST record for ODP Site 1019C off the coast of northern California (data and chronology from Barron et al., 2003), Lake Barrett and Starkweather Lake chironomid-based temperature estimates, diatom-inferred salinity, cellulose-inferred $\delta^{18}O_{lw}$ record, and the Owens Lake $\delta^{18}O$ and total inorganic carbon (TIC) stratigraphies (Benson et al., 1997, 2002; Owens Lake chronology from Benson et al., 2002 assuming minimal reservoir effect). The gray lines on the Barrett and Starkweather series depict the timing of changes using a chronological model based upon ^{14}C date calibration using CALIB4 and IntCal98 so the series can be compared directly to earlier calibrated series.

with the increasing salinity, generally lowered lake levels, and the modest increase in LOI. Since the chironomid records show continued low temperatures during the latter part of the YD, this apparent change in hydrology to drier conditions during the second half of the YD stade may have largely reflected a decline in precipitation rather than an increase in evaporation.

At Lake Barrett there is a decline in *Artemisia* and Chenopodiaceae-Amaranthacea (Cheno/Am) pollen percentages and an increase in *Pinus* at the initiation of the YD stade. These changes suggest increased tree cover at that time (Fig. 4). Vegetation at Lake Barrett, which lies on the eastern slope of the Sierra and closer to the sagebrush-forest ecotone, may have been more climatically sensitive than at Starkweather Lake. At both sites the greatest split in the CONISS zonation dendrogram occurs near the end of the YD stade, when increasing *Artemisia* and declining *Pinus* suggest the initiation of drier and more open vegetation conditions at both sites. The relatively subtle changes in the vegetation during the YD stade appear to reflect changes in moisture and evaporative stress. The charcoal records, although noisy, may suggest a decline in regional fire activity during the YD stade.

Discussion

The Barrett and Starkweather records indicate decreased summer water temperatures of ~2°C–4°C during the YD stade. The timing of the YD stade as presented in our records generally corresponds to the timing of temperature depression evident in the North Atlantic sector and in the North Pacific (Fig. 5). Uncertainties inherent in our radiocarbon-based chronologies and the potential of some sediment mixing make it impossible to infer confidently if the start of temperature depression and initiation of other changes at the eastern Sierra sites led, coincided with, or followed the decline in temperatures evident in the North Atlantic sector or the northeastern Pacific.

Although the precise relationship between lake-water temperature and air temperature is not known, the decline in water temperatures is comparable with YD stade temperature depressions reported from the northeast Pacific region and elsewhere in the western United States. Our reconstructions of the timing and magnitude of temperature depression in the eastern Sierra are generally similar to that reported from alkenone-based SST reconstructions for the northeast Pacific Ocean (Barron et al., 2003). Palynological analysis of Owens Lake (Mensing 2001), packrat middens from the Grand Canyon (Cole and Arundel, 2005), and a δ^{18}O speleothem record from Oregon (Vacco et al., 2005) provide evidence of similar temperature depression at these land-surface sites. The Barrett and Starkweather lake records suggest that cooling during the YD, similar in magnitude to these other regions, also occurred in the east-central Sierra Nevada and further support a linkage between North Atlantic sector cooling, northeastern Pacific cooling and cooling in southwestern North America.

The evidence for wetter conditions during the early YD stade that is recorded in the Barrett and Starkweather lake sediments is consistent with a number of paleolimnological records from the neighboring Great Basin. A paleoshoreline record from Owens Lake (Fig. 1) during the Pleistocene–Holocene transition is preserved in two major transgressive beach complexes at Swansea (36.60°N, 117.90°W). The earlier, higher beach reaches 1128 m a.s.l; the later, lower beach 1118–1119 m a.s.l. Calibrated radiocarbon ages of ~12,880 cal yr BP and ~12,820 cal yr BP are reported by Orme and Orme (1993, 2008) for fragments of the clam *Anodonta californiensis* from the lower beach, which suggest that the later transgression culminated shortly after the onset of the YD stade. Farther south along the paleo-Owens River system, Searles Lake also contains indications of a brief highstand during the YD stade (Phillips et al., 1994). The Searles Lake record for the past ~40,000 yr indicates a positive relationship between lake highstands and earlier late Pleistocene stadials (Phillips et al., 1994). Beach ridges related to a late Pleistocene highstand at Pyramid Lake, Nevada, date to 12,840 cal yr BP (Briggs et al., 2005). High water levels are reported for the Ruby Marshes of central Nevada at ~12,830 cal yr BP (Thompson, 1992). Shallow lakes also developed in the Long Valley region of north-central Nevada (Huckleberry et al., 2001). In western Utah, the Gilbert Shoreline formed during a Lake Bonneville transgression starting ~12,900 cal yr BP (Oviatt et al., 2005; Orme and Orme, 2008).

The shift to drier conditions during the mid- to late YD stade that is apparent in our records is supported by evidence for similar hydrological variability from Owens Lake (Benson et al., 1997, 2002; Mensing, 2001; Bacon et al., 2006). Sedimentological, stable-isotope and pollen evidence from Owens Lake (Benson et al., 1997, 2002; Mensing, 2001) indicate a dry late BA interstade and an abrupt increase of moisture in the early part of the YD stade. Evidence for this shift includes sharp decreases in δ^{18}O and total inorganic carbon (TIC). These Owens Lake records indicate a sharp reversal in these trends in the middle to late YD stade (Fig. 5). A more recent paleohydrological reconstruction of Owens Lake water levels, based upon stratigraphic exposures of fluvio-deltaic and lacustrine sediments, suggests high levels at the very start of the YD stade and a decline to low levels in the middle of the YD (Bacon et al., 2006).

Similar hydrological shifts during the YD stade appear to have been widespread and are present in records from as far away as Asia and Africa (Zhou et al., 2001). The causes of pronounced hydrological variability during the YD stade may relate to low-latitude ocean−atmosphere processes and their influence on precipitation in the mid- to high latitudes (Zhou et al., 2001). In addition, large variations in the flux of freshwater through the Gulf of St. Lawrence into the North Atlantic may have generated climatic variability within the YD stade (Carlson et al., 2007).

Our records from the eastern Sierra suggest temperature depression throughout the YD stade and increased moisture in the early part of the YD. As reviewed above, evidence for cooler conditions during the YD stade and moist conditions early in the YD are now available from a number of other studies in southwestern North America. One question that remains to be resolved is the lack of evidence for a glacial readvance in the eastern Sierra during the YD stade. The available data indicate that Recess Peak ice had retreated by 13,100 cal yr BP with no apparent readvance with the onset of the YD stade (Clark and Gillespie, 1997; James et al., 2002). Interestingly, there is evidence of a glacial readvance in the San Bernardino Mountains of southern California (~500 km south of our sites) during the YD stade (Owen et al., 2003). Our evidence for warm temperatures during the late BA is consistent with retreat of Recess Peak ice by 13,100 cal yr BP. The Barrett and Starkweather sediments do not contain any evidence of direct glacial input, such as dropstones or glacial flour which might directly signal a large readvance during the YD stade. However, it is likely that their elevation is too low to expect such direct glacial influence even if there had been a minor advance in higher elevation cirques.

Seeming inconsistencies in evidence for changes in climate, lake level and glacial advances in the western United States during the YD stade have been attributed to the different sensitivities of various systems and regional differences in climatic responses (Licciardi, 2001, Licciardi et al., 2004; Oviatt et al., 2005). Perhaps insufficient temperature depression and/or the relatively short span of moist conditions during the early YD stade followed by dry conditions may have precluded significant glacial readvance in the eastern Sierra Nevada during the YD stade.

Conclusion

The paleoenvironmental records from Lake Barrett and Starkweather Lake indicate a marked cooling of summer water temperatures during the YD stade. It appears that there was an initial increase in effective moisture and then the onset of drier conditions during the middle period of the YD. The combination of thermal and hydrologic changes during the YD stade evident in our records and other studies from the region support a close linkage between conditions in the North Atlantic, the North Pacific and California climate. The linkage underscores a clear challenge that future climate change may present to California and the West. Changes in the North Atlantic sector that impact thermohaline circulation appear to have been consistently reflected in changes in northeastern Pacific SSTs and hydrological shifts in eastern California over the late Pleistocene (Phillips et al., 1994; Benson et al., 1997, 2002; Hendy et al., 2002; Bacon et al., 2006). The possibility that the North Atlantic thermohaline circulation regime and/or northeastern Pacific SSTs might be significantly affected by future global climate warming (Broecker, 1997; MacDonald et al., 2007) makes these challenges particularly pressing.

Acknowledgments

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