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Orbital- and millennial-scale vegetation and climate changes of the past 225 ka from Bear Lake, Utah–Idaho (USA)

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Abstract

Continuous high-resolution pollen data for the past 225 ka from sediments in Bear Lake, Utah–Idaho reflect changes in vegetation and climate that correlate well with variations in summer insolation and global ice-volume during MIS 1 through 7. Spectral analysis of the pollen data identified peaks at 21–22 and 100 ka corresponding to periodicities in Earth's precession and eccentricity orbital cycles. Suborbital climatic fluctuations recorded in the pollen data, denoted by 6 and 5 ka cyclicities, are similar to Greenland atmospheric temperatures and North Atlantic ice-rafting Heinrich events. Our results show that millennial-scale climate variability is also evident during MIS 5, 6 and 7, including the occurrence of Heinrich-like events in MIS 6, showing the long-term feature of such climate variability. This study provides clear evidence of a highly interconnected ocean–atmosphere system during the last two glacial/interglacial cycles that extended its influence as far as continental western North America. Our study also contributes to a greater understanding of the impact of long-term climate change on vegetation of western North America. Such high-resolution studies are particularly important in efforts of the scientific community to predict the consequences of future climate change.

1. Introduction

Numerous paleoclimate proxy records from Greenland, the North Atlantic and the Pacific Ocean have documented the existence of millennial-scale climate variability during the last glacial period (Bond et al., 1993; Dansgaard et al., 1993; Grootes et al., 1993; Bond and Lotti, 1995; Behl and Kennett, 1996). An increasing number of studies on marine sediments document the occurrence of similar millennial-scale variability during the penultimate glacial period and previous interglacials (Oppo and Lehman, 1995; Maslin and Tzedakis, 1996; Bond et al., 1997; Raymo et al., 1998; Peterson et al., 2000; Tzedakis et al., 2004; Siddall et al., 2006). High-resolution pollen records from continuous sedimentary sequences provide an opportunity to examine the regional response of vegetation to these climatic changes. Although the number of paleovegetation records spanning the last glacial–interglacial cycle in North America has increased in recent years (i.e. Clear Lake (Adam et al., 1981), Humptulips (Heusser and Heusser, 1990), Santa Barbara Basin (Heusser, 1995, 1998, 2000), Carp Lake (Whitlock and Bartlein, 1997), Searles Lake (Litwin et al., 1999) and Owens Lake (Woolfenden, 2003)), only a few pollen studies show suborbital-scale climate variability in continental North America (Grimm et al., 1993, 2006; Whitlock and Grigg, 1999) due to the very fragmentary sedimentary record and the lack of temporal resolution and/or sensitivity necessary to reveal the extent and effects of these environmental fluctuations in vegetation. Documenting natural modes of climate variability during warm intervals of the past that resemble the present interglacial is important in understanding and assessing the nature of postulated future climatic changes. Some pollen studies show that the last two interglacials (MIS 5e and 7c) were warmer than the Holocene (Tzedakis and Bennett, 1995; Whitlock and Bartlein, 1997; Heusser, 2000; Woolfenden, 2003). A detailed study of the vegetation during

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past warm intervals is very important for the prediction of future climate change on the vegetation as temperatures are increasing today (Brohan et al., 2006).

Models of the relationship between lake size and pollen catchment area suggest that large lakes with big drainage basins will primarily record the regional pollen rain (Jacobson and Bradshaw, 1981). We present here high-resolution (ca 600 yr) pollen data from a sediment core from Bear Lake, on the Utah–Idaho border, for the past 225 ka in order to (1) determine the regional vegetation and climate history of this area and (2) compare these data with other high-resolution records of global ice volume, Greenland atmospheric temperature variations and vegetation changes in continental North America.

2. Bear Lake and core BL00-1E

Bear Lake is a large (surface area of 280 km^2), deep (to 63 m depth) lake located directly along the Idaho–Utah border in the northern Great Basin (Bright et al., 2006) and is one of the few large lakes that survived Holocene desiccation in western North America (Bright et al., 2006). The lake is located at the southern end of the larger Bear Lake Valley and is surrounded by the Bear River Range on the west (maximum elevation of approximately 3030 m); on the east is the Bear Lake Plateau with a maximum elevation of approximately 2270 m (Fig. 1). Prior to its diversion into Bear Lake between 1909 and 1918 (Birdsey, 1989), the Bear River, which heads up in the Uinta

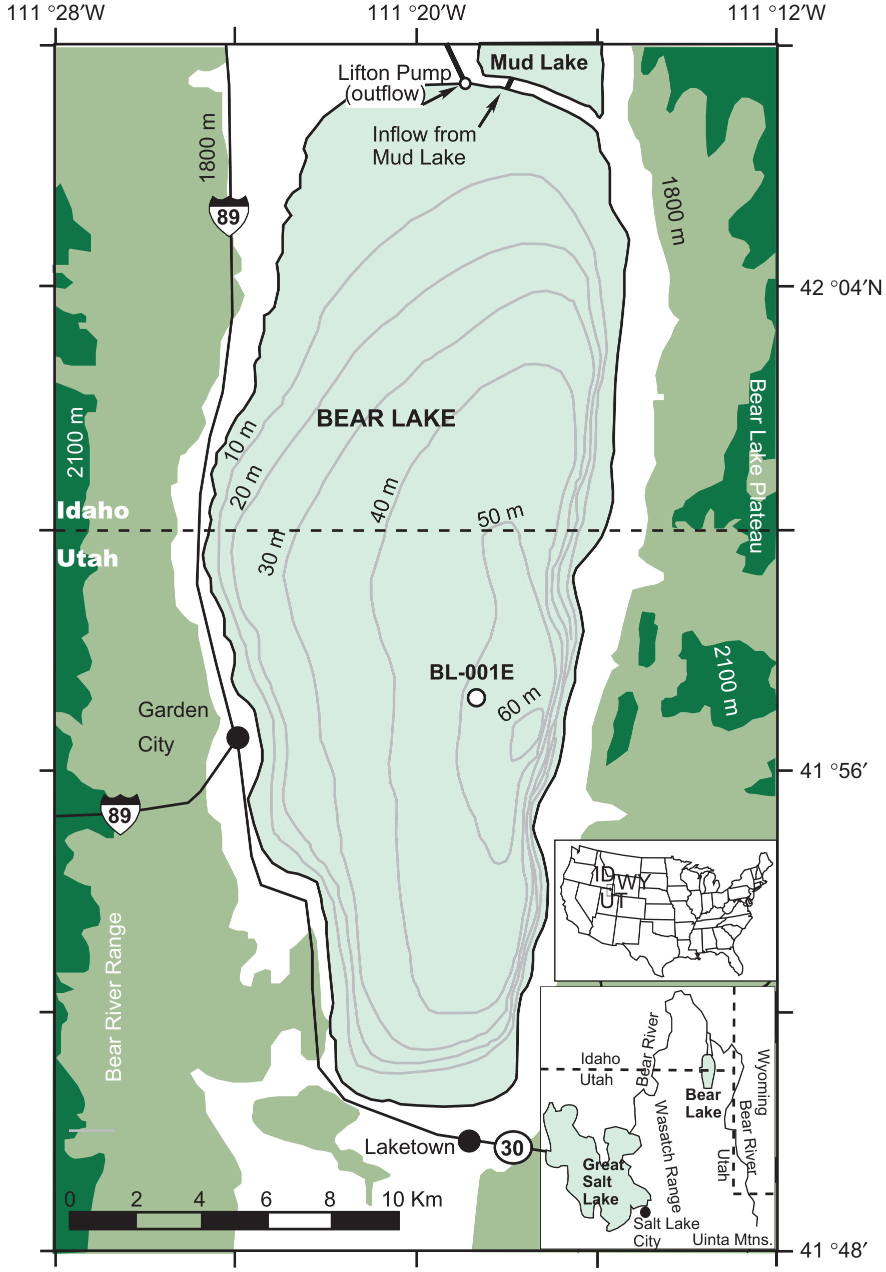

Fig. 1. Location of Bear Lake and borehole BL00-1E (marked with circle). Modified from Dean et al. (2006) and Bright et al. (2006).

Mountains ca 150 km to the south, passed ca 13 km to the north of the lake (Bright et al., 2006).

Climate data for Laketown, Utah (1823 m), immediately south of the Lake, show maximum average July and minimum average January temperatures of 26.4 and −11.8 °C, respectively. Precipitation is received throughout the year, with an annual average of 29.8 cm, with 132 cm of snow depth (http://www.wrcc.dri.edu/Climsum.html).

The Uinta Range and adjacent foothills, to the east of Bear Lake, span a broad vegetational gradient, with upper slopes and summits vegetated by alpine tundra (>3350 m) through subalpine Engelmann spruce (*Picea engelmannii*)/ subalpine fir (*Abies lasiocarpa*) forest (3200–3500 m) and montane lodgepole pine (*Pinus contorta*) forest (2400–3200 m) (Jackson et al., 2005). The southern and eastern flanks of the Uintas, which have relatively high summer precipitation, have extensive ponderosa pine (*P. ponderosa*) forests (1900–2500 m) and Rocky Mountain piñon (*P. edulis*)—Utah juniper (*Juniperus osteosperma*) woodlands (1675–2200 m) grading downward into big sagebrush (*Artemisia*) steppe and desert shrubland (Jackson et al., 2005). *Pinus ponderosa* forests, and woodlands of *P. edulis* and *J. osteosperma* also occur on the north slope of the range, south and west of the study site. On the western flanks, *P. contorta* forests grade downslope into mixed forests dominated by Douglas-fir (*Pseudotsuga menziesii*) and Rocky Mountain aspen (*Populus tremuloides*), which in turn give way to *Artemisia* steppe and desert shrubland below 2000 m.

Many areas directly surrounding Bear Lake today have been converted to agricultural and developed land. Natural vegetation to the east, south and north is largely *Artemisia* steppe and shrubland with *Pinus edulis*—*Juniperus* woodland on scarps to the east. To the west, with increasing elevations in the Bear River Range, are *Populus tremuloides* forest and woodland, *Picea engelmannii*–*Abies lasiocarpa* forest and woodland (Southwest Regional Gap Analysis, 2003). This diversity makes the Bear Lake record a very sensitive environment for recording variations in plant distribution and climate.

The composition of sediments in the lake indicates that the Bear River flowed naturally into Bear Lake several times in the past, mostly during glaciations (here generally considered cold/wet periods), carrying sediments and pollen grains from the higher elevation Uinta Mountains and the more immediate surroundings of Bear Lake (Dean et al., 2006; Bright et al., 2006; this study). The lake was mainly closed during interglaciations (warm/dry periods), when the lake retracted into a topographically closed basin (Bright et al., 2006). The modern Bear Lake is topographically separated from Bear River, except for regulated flow via the canals. When Bear Lake rises and captures flow from the Bear River, the lake then becomes a topographically open basin (Laabs and Kaufman, 2003).

Sediments for this study come from a 120-m-long core (BL00-1E; 41°57′06″N, 111°18′30″W; 51.12 m water depth; altitude 1805 m asl; Fig. 1) recovered in 2000 using the Global Lake Drilling to 800 m (GLAD800) drill rig (Dean et al., 2002; http://esp.cr.usgs.gov/info/lacs/glad.htm). Preliminary age control for the BL00-1E core was provided by ^{14}C dates correlated from cores BL96-1, 2, and 3 with the upper portion of core BL00-1E, one magnetic excursion (correlated to a standard sequence) and by a U/Th date (127.7 ± 3.9 ka) at 66.4 m on fine-grained aragonite layer (Colman et al., 2006) (Fig. 2). Because the initial set of dates was restricted to the upper half of the BL00-1E core, additional ages were inferred by direct correlation of the carbonate and isotopic data to the independently dated paleoclimate record from Devils Hole, Nevada, using maxima and minima of the isotopic records as the tie points (Colman et al., 2006). The age assessments suggest that the core extends back to about 225 ka (Fig. 2). Because the age model contains assumptions about the correlation of Bear Lake to other climate records, this model should be interpreted with caution (Colman et al., 2006). We also provide an alternative, linear age model (Fig. 2), that includes correlated radiocarbon dates plus the U/Th date only.

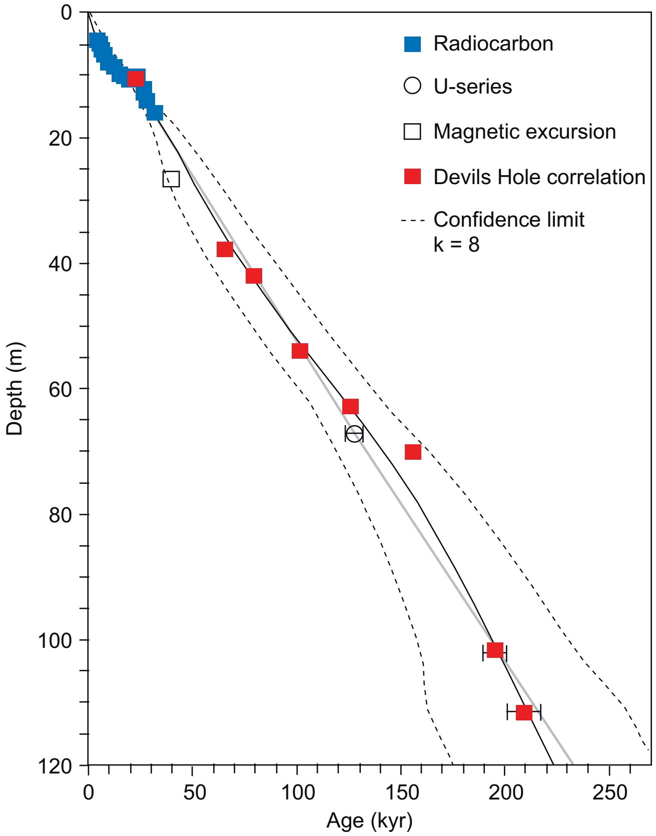

Fig. 2. Age model and 95% confidence limits for core BL00-1E, from Colman et al. (2006). The age model (black) was constructed using the radiocarbon, magnetic excursion and uranium-series age estimates, as well as the climate correlations with Devils Hole (Colman et al., 2006). A second, linear age model (grey) includes correlated radiocarbon dates plus the U/Th date only. K is the number of spline functions used in the fit, which controls the degree of smoothness (see Colman et al., 2006, for more details).

Three additional radiocarbon-dated cores of 4–5 m in length have provided a record of the last glacial–interglacial cycle from Bear Lake (Dean et al., 2006). Sediments deposited during the last glacial (MIS 2) include reddish silty clays that were transported to the lake by the Bear River and contain glacial flour derived from the Uinta Mountains to the east. These siliceous sediments have relatively high magnetic susceptibility, and contain an ostracode (*Cytherissa lacustris*) indicative of cold climates and dilute water (Dean et al., 2006). Toward the end of MIS 2, Bear Lake disconnected from the Bear River (became a topographically closed basin) and sedimentation became progressively dominated by carbonate precipitation, which characterizes the majority of the Holocene (MIS 1) sediments. Previous BL00-1E core studies show that the large-scale trends in $\delta^{18}O$ in the lake broadly parallel glacial–interglacial fluctuations known from other proxy climate records (Bright et al., 2006). Three long intervals with abundant endogenic aragonite and enriched $\delta^{18}O$ and $\delta^{13}C$ values have been interpreted as warm and dry periods and are correlated with the interglacials MIS 1, 5e and 7a (Bright et al., 2006).

3. Methods

Sediment samples were obtained from the joint University of Minnesota-NSF's LacCore facility, where the cores are curated permanently. Our core BL00-1E samples for pollen analysis were taken at an average interval of 30-cm (ca 600-yr) for the 120-m (ca 225 ka) record. Two *Lycopodium* tracer tablets were added to each sample ($2\,cm^3$) for pollen concentration calculation. Pollen extraction methods followed a modified Faegri and Iversen (1989) procedure, including treatment with KOH, washing through a 180-μm screen, and treatment with dilute HCl, HF and acetolysis solution. Samples were then sieved through an 8-μm Nitex screen to remove additional clay and fine organics. Sample residues were stained with safranin-O and mounted in silicone oil. Counting was performed at 400 × magnification on a compound light microscope. Pollen identification was accomplished by comparing the fossils with their present-day relatives using published keys and the modern pollen reference collection at Northern Arizona University. At least 300 terrestrial pollen grains per slide were identified. The percentages of each pollen taxa were calculated, and the results were plotted in a detailed simplified diagram (Fig. 3). To highlight basic patterns, taxa with similar relative temperature and moisture requirements (i.e., warm and arid) were grouped together and plotted in Fig. 4.

A cyclostratigraphic analysis was performed on the BL00-1E pollen time series on the relative abundance of taxa that grow in relatively warm and/or arid conditions. We used the program REDFIT (Schultz and Mudelsee, 2002) with the objective of characterizing the different periodicities present in the unevenly spaced pollen time series and estimating its red-noise spectra. We ran two scenarios, using (1) the age model of Colman et al. (2006) and (2) a linear age model that was independent of assumptions of the climate correlations that are built into the Colman et al. model (Fig. 2). This simple age model is based only on the U/Th age of 128 ka, plus the correlated ^{14}C ages from the BL96 cores. The spectral analysis assisted in identifying recurrent features or periodicities through spectral peaks registered at differing frequencies throughout the studied core.

4. Results and interpretation

Based upon the chronology established by Colman et al. (2006), our average sampling interval corresponds to ca 600 years which has allowed us to produce a high-resolution paleovegetation record covering the last three interglacials. Consequently, our pollen zones are coeval with Oxygen Isotope Stages 1 through 7 (Fig. 3), and we do not define new pollen zones.

Interglacials (Zones 1, 5e and 7c) and interstadials (Zones 3, 5a, 5c, 7a and 7e) have higher percentages of Chenopodiaceae—*Amaranthus*, *Sarcobatus*, and *Ambrosia*, as well as *Juniperus* and *Quercus* pollen (Fig. 3). These taxa today are characteristic of arid, intermontane basins of Utah and Idaho. We also use the term "warm" in a relative sense to describe these taxa, since they are most abundant during the Holocene sediment record from nearby Dingle Swamp (L. Doner, http://esp.cr.usgs.gov/info/lacs/polperc.htm). Chenopodiaceae—*Amaranthus* pollen, primarily from flats of *Atriplex* spp., as well as *Sarcobatus*, is commonly found today on basin floors throughout the Great Basin (Welsh et al., 1987). High *Ambrosia* concentrations may result from species like *A. psilostachya* or *A. acanthicarpa*, which can be common in similar environments (Welsh et al., 1987). Our confidence in including these five taxa together comes not only from similar occurrence in OIS 1 from Dingle Swamp but also from OIS 1 and 5e sediments in the Indian Cove Well core from a Great Salt Lake (Davis and Moutoux, 1998). We included *Juniperus* in this group as well for similar reasons (Davis and Moutoux, 1998).

Glacial (Zones 2, 4 and 6) and stadial (Zones 5b, 5d, 7b and 7d) periods are characterized at Bear Lake by higher percentages of *Picea*, other Asteraceae, and *Eriogonum*. The *Picea* pollen could represent either Engelmann (*P. engelmannii*) or Colorado blue (*P. pungens*), both of which grow today in the region (Welsh et al., 1987). The occurrence of *Picea* clearly suggests colder and perhaps wetter conditions (greater effective moisture), since each has distributions today at higher elevations. It is unclear which species contributed to higher percentages of other Asteraceae and *Eriogonum* pollen. However, both pollen types have higher percentages in Lateglacial sediments from Dingle Swamp (L. Doner, http://esp.cr.usgs.gov/info/lacs/polperc.htm), while Other Asteraceae is more abundant in glacial period records from the Great Salt Lake (Davis and Moutoux, 1998), ca 150 km to the southwest.

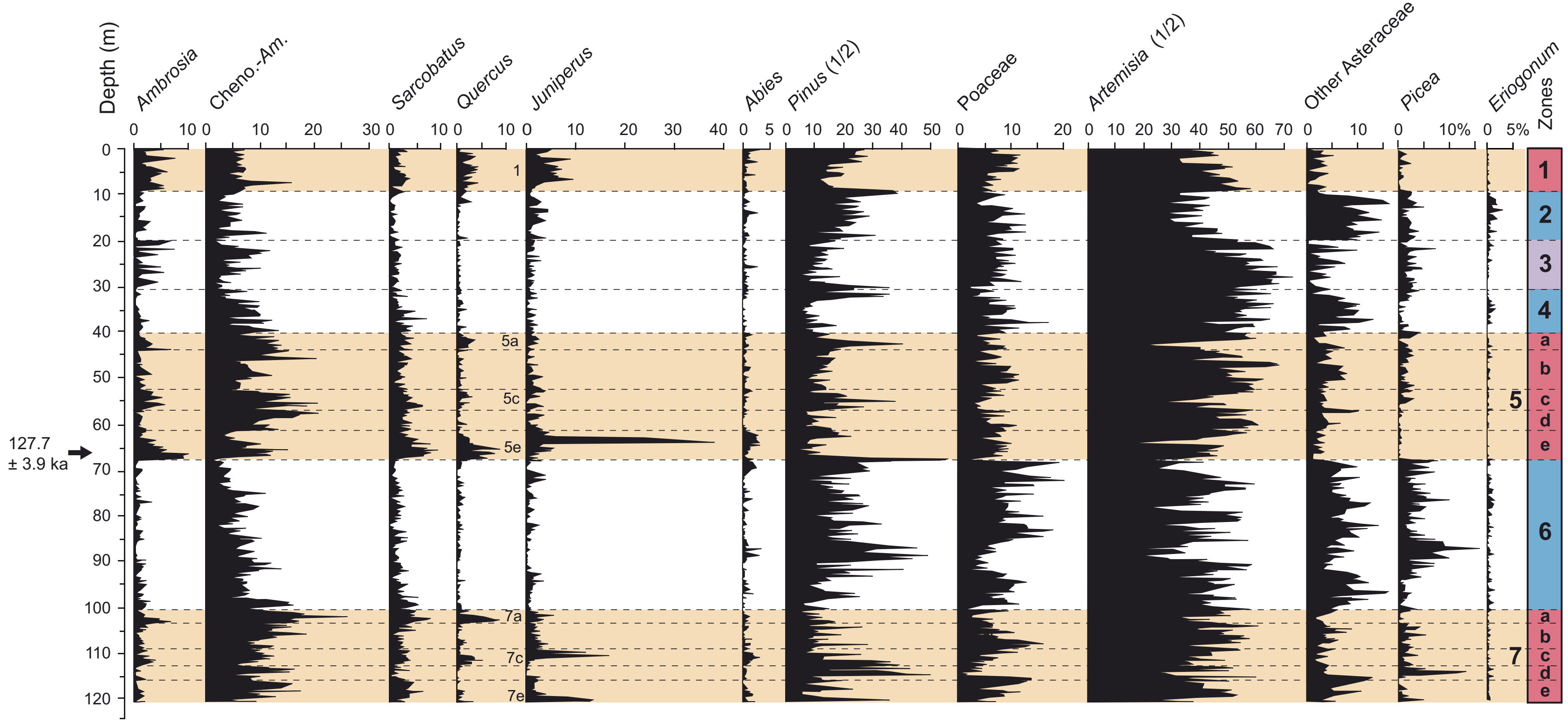

Fig. 3. Summary terrestrial pollen diagram of the BL00-1E core showing percentages of selected taxa. A minimum of 300 grains of terrestrial pollen was counted in every sample. Note change of scale for both *Pinus* and *Artemisia*. The U/Th date of 127.7 ± 3.9 ka at 66.4 m on fine-grained aragonite layer (Colman et al., 2006) is shown on the left. On the right, the main pollen zonation on the core.

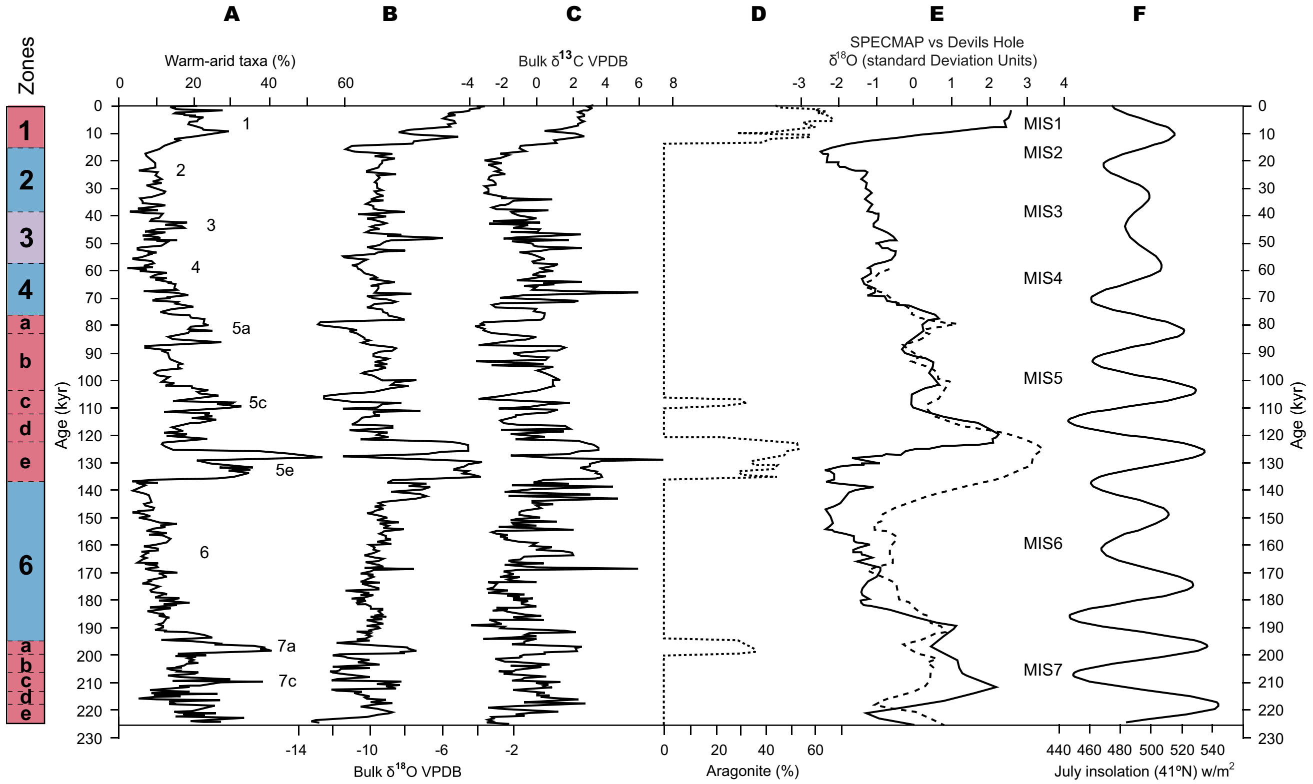

Fig. 4. Comparison of palaeoenvironmental records of the last 225,000 yr. (A) Relative abundance of warm-arid taxa (*Ambrosia*, Chenopodiaceae—*Amaranthus*, *Sarcobatus*, *Quercus* and *Juniperus*) in the pollen spectra and main pollen zones discussed in the text. (B) Bulk sediment δ^{18}O‰ VPDB record for the BL00-1E core (Bright et al., 2006). (C) Bulk sediment δ^{13}C‰ VPDB record for the BL00-1E core (Bright et al., 2006). (D) Aragonite content in the sediments from BL00-1E core. Percentage is determined by X-ray diffraction (intensity of the mean aragonite peak divided by the sum of the main peak intensities) (Bright et al., 2006). (E) SPECMAP vs. Devils Hole (dashed line) δ^{18}O record in standard deviation units (Martinson et al., 1987; Winograd et al., 1997) and the standard Marine Isotope Stage zonation. (F) July insolation at 41°N (Berger, 1978).

We consider these occurrences to be important support for including these taxa within the cold-wet indicators.

Artemisia and Poaceae are abundant in most of the samples, and even though they oscillate widely throughout the core, neither taxa showed a very clear association with either interglacial/interstadial or glacial/stadial assemblages. For *Artemisia*, this could reflect its relatively wide elevational range (*A. tridentata* alone today grows from ca 1200 to over 3050 m elevation; Welsh et al., 1987), while Poaceae includes many species of different ecological requirements. *Pinus*, also consisting of many species, appears to be very abundant during transitions between the different stages.

4.1. Orbital-scale changes

High-resolution analysis of the Bear Lake sediments shows close agreement between the pollen zones identified here (Fig. 3) and the Marine Isotope Stage (MIS) and Devils Hole δ^{18}O stratigraphy (Fig. 4). Our interpretations of the pollen record correlate well with those of (1) abundance in carbonates (calcite and aragonite) (Bright et al., 2006; Dean et al., 2006) and (2) δ^{18}O and δ^{13}C in the core (Bright et al., 2006), as well as with (3) the insolation curve for the same time-span (Berger, 1978) (Fig. 4). During periods of summer insolation maxima (Fig. 4F) and low Northern Hemisphere ice volume (Fig. 4E) the warm-arid taxa were very abundant (Fig. 4A) and the lake sedimentation was generally characterized by abundant aragonite deposition (Fig. 4D) and enriched δ^{18}O and δ^{13}C values (Figs. 4B and 4C; Bright et al., 2006). This correlation is particularly striking during the last three interglacials in the record (MIS 1, 5e and 7c) (Fig. 4). During times of insolation minima (Fig. 4F) and extensive Northern Hemisphere ice volume (Fig. 4E), higher elevation tree (mainly *Picea*) and herb (mainly Other Asteraceae) pollen was more abundant (Fig. 3), sediments were poor in aragonite (Fig. 4D) and the δ^{18}O and δ^{13}C values were usually more negative (Bright et al., 2006) (Figs. 4B and 4C). This correlation clearly links global climate variations, regional vegetation changes and the main chemical–mineralogical sedimentary fluctuations in Bear Lake (Bright et al., 2006). We interpret these changes as demonstrating a strong orbital signal in the BL00-1E record which is obvious in the pollen spectra as discussed above (Figs. 3 and 4).

Even though generally excellent correlation exists between the different proxies studied from Bear Lake, some discrepancies have also been observed. These differences occur mainly during glaciations (i.e. MIS 6), and were previously observed by Bright et al. (2006) in the isotopic data, with a low $\delta^{18}O_{bulk}$–$\delta^{13}C_{bulk}$ correlation coefficient. They are probably related to non-linear responses in the isotopic evolution of both $\delta^{18}O$ and $\delta^{13}C$ due to differences in biologic productivity and geomorphic changes, independently of climate, but the differences in $\delta^{18}O$ may also reflect differences in isotopic fractionation between calcite and aragonite (Bright et al., 2006). Interpretations from the pollen data are also occasionally at odds with the isotopic record, which can be explained as variations in the response (sensitivity, intensity, etc.) of the different proxies to environmental or climate change as well as different factors that could change the water chemistry independent of climate, such as the effect of tectonic and geomorphic processes (Laabs and Kaufman, 2003; Bright et al., 2006).

Our data support the hypothesis that the previous interglacial (MIS 5e) was the warmest period of the last 225 ka in western North America, with the highest percentages of warm-arid indicators (i.e., *Ambrosia*, Chenopodiaceae—*Amaranthus*, *Sarcobatus*, with *Quercus* and *Juniperus*; Figs. 3 and 4). Pollen grains, together with the aragonite content and $\delta^{18}O$, seem to indicate that stage 7a was warmer-drier than stage 5a. The early Holocene was also characterized by the high abundances of warm taxa coinciding with the summer insolation maxima (Figs. 3 and 4). Insolation maxima at these times undoubtedly led to increased summer temperatures, decreased effective precipitation, and indirect strengthening of the eastern Pacific subtropical high-pressure system, which intensified drought in this area (Whitlock and Bartlein, 1997).

Pollen indicators of colder conditions during pollen Zones 2 (the Last Glacial Maximum) and 6 (the end of MIS 6) coincide with insolation minima (Fig. 4). During glacial periods, lower insolation values, along with the extent of the Laurentide ice sheet and low atmospheric CO_2 (Weaver et al., 1998) greatly influenced temperature, effective precipitation and atmospheric circulation with a displacement of polar jet stream to the south of its present position, causing cooling and an increase in Pacific moisture in the Bear Lake area (Bartlein et al., 1998).

The BL00-1E pollen record also shows the typical structure of the glacial cycles, with long glaciations and short deglaciations (saw-tooth structure) (Fig. 4). However, in this record pollen changes are more pronounced during the Termination II transition than during Termination I. It is tempting to ascribe this to the greater summer solar insolation during MIS 5e than MIS 1.

Vegetation records spanning the last 125 ka in North America are rare but do reflect changes mainly related to the orbital-scale controls of climate (Adam et al., 1981; Heusser and Heusser, 1990; Whitlock and Bartlein, 1997; Heusser, 1998; Litwin et al., 1999; Heusser, 2000; Woolfenden, 2003). Independent of the different chronologies used in these studies, a broad correspondence between the patterns of western US (Santa Barbara Basin, Clear Lake, Carp Lake, Owens Lake and Bear Lake) pollen records since MIS 5 is apparent. For instance, similar to the Bear Lake record, xerothermic indicators are highest at Carp Lake (*Pseudotsuga* + *Quercus* + Cupressaceae; Whitlock and Bartlein, 1997), at Clear Lake, California (*Quercus* over *Pinus*; Adam et al., 1981) and in Santa Barbara Basin sediments (*Quercus*; Heusser, 2000) during MIS 5e, 5c, 5a and 1 (Fig. 5). Differences between these records are probably due to the complex response of different species to climate change, regional peculiarities or errors in chronology. For example, the Owens Lake pollen record (Woolfenden, 2003) differs from the other records in that no clear subdivision exists between substages within MIS5. The transition from stages 5a to 4 is more gradual in the Bear Lake record, especially compared to the coastal California sites (Santa Barbara Basin and Clear Lake; Fig. 5).

These patterns of environmental change noted in the Bear Lake and other western North American records resemble sequences reconstructed from long European pollen records, such as La Grande Pile (France) (Woillard, 1978), Ioannina (Greece) (Tzedakis and Bennett, 1995; Tzedakis et al., 2002), MD95-2042 (Iberian Peninsula) (Sánchez-Goñi et al., 1999, 2000) and MD95-2043 (western Mediterranean Sea) (Sánchez-Goñi et al., 2002). In these studies interglacials and interstadials are rich in temperate trees while glacials and stadials are characterized by higher percentages of steppe indicators, mainly *Artemisia*.

4.2. Millennial-scale changes

Millennial-scale changes are also noted in the pollen spectra from Bear Lake. Colder periods here, characterized by higher percentages of *Picea*, other Asteraceae, and *Eriogonum*, are followed by relatively abrupt warmings, indicated by higher percentages of Chenopodiaceae-*Amaranthus*, *Sarcobatus*, and *Ambrosia*, as well as *Juniperus* and *Quercus* pollen. During the last glaciation (MIS 2, 3, and 4) these fluctuations show patterns similar in age and duration with millennial-scale climate variability as recognized in marine records from the Santa Barbara (Hendy and Kennett, 1999) and Cariaco (Peterson et al., 2000) basins, and the North Atlantic, including Heinrich events (Heinrich, 1988; Bond et al., 1993; Bond and Lotti, 1995) and Greenland surface temperatures (Grootes et al., 1993; Dansgaard et al., 1993) (the latter is particularly well shown for interstadials 19, 20 and 21, Fig. 6). In particular, events similar in timing to Heinrich events H1 through H6 are recorded in our pollen data (well shown for Heinrich events 4, 5 and 6, Fig. 6). However, evidence of the Younger Dryas does not show up in this core, either in the pollen data, or in other proxies (Bright et al., 2006), nor are there reversals in proxies analyzed from more closely spaced samples in core BL96-2 (Dean et al., 2006). This needs further investigation.

Long pollen records that demonstrate more than just a few individual Heinrich events over the last glacial–interglacial cycle are very rare in continental North America

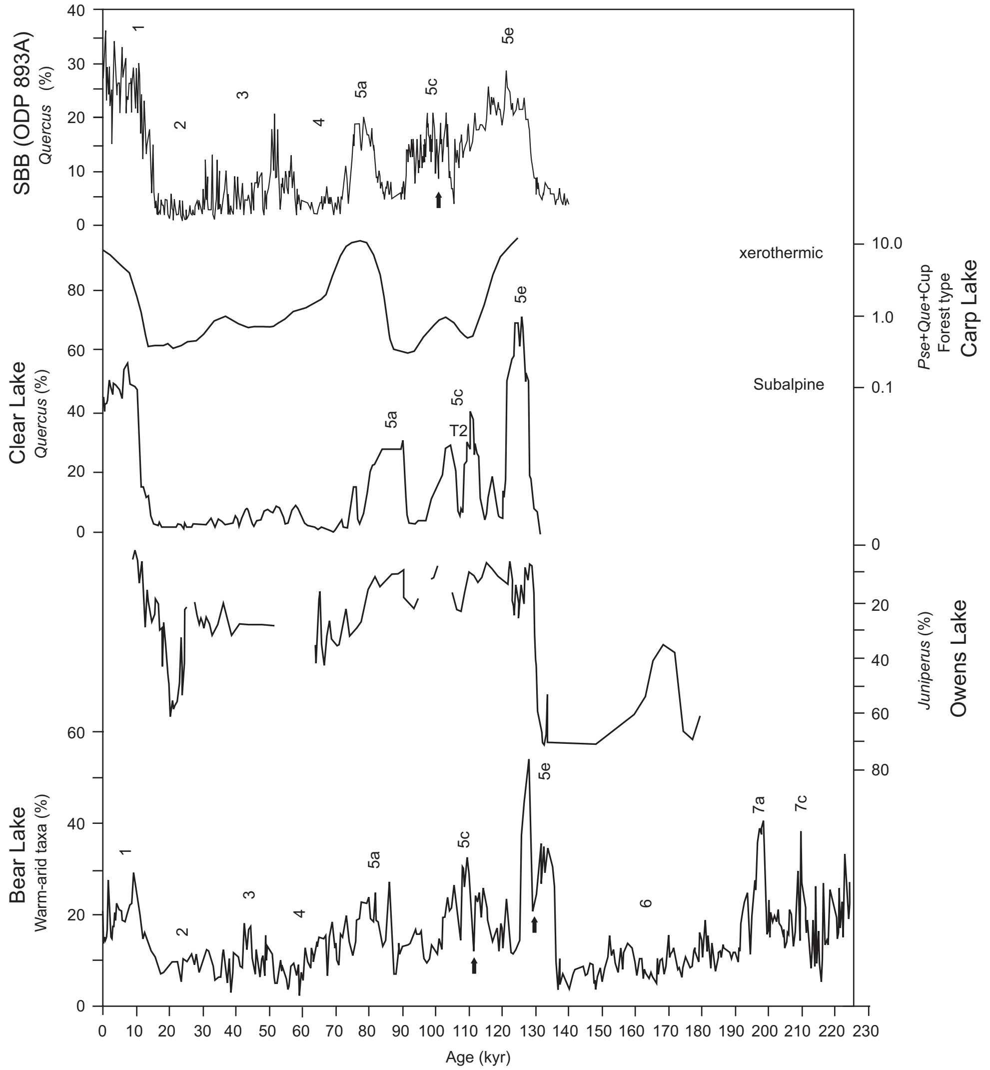

Fig. 5. Comparison of long pollen records from western US for the last 225 ka. Pollen data are from BL00-1E, Bear Lake, Utah–Idaho (this study); Owens Lake, California (Woolfenden, 2003); Clear Lake, California (Adam et al., 1981); Carp Lake, Washington (Whitlock and Bartlein, 1997); and ODP 893A, Santa Barbara Basin, California (Heusser, 1998, 2000). Arrows pointing to a cooling during stage 5c in the SBB and Bear Lake records correlate with Clear Lake event Tsiwi 2, identified as T2. Arrow in Bear Lake record pointing to a cooling during stage 5e can be interpreted as the "Eemian cooling event" (Maslin and Tzedakis, 1996). Note that each of the time series is based on independent age models. The line plotted for Carp Lake is a locally weighted mean and not actual points as in the other records.

(Grimm et al., 2006; Huang et al., 2006), mostly due to poor dating control (Kershaw and Whitlock, 2000) and lack of high-resolution pollen analyses (Anderson et al., 2000). In most of the northern hemisphere, Heinrich events are associated with wetter and/or cooler temperatures (Heinrich, 1988; Andrews, 1998; Hemming, 2004). Our pollen data suggest that warm and arid pollen types decline during these events, and are replaced by higher percentages of *Picea* and Other Asteraceae, which are more characteristic of cooler and wetter conditions (see discussion above).

Millennial-scale climatic variability is also apparent in sediments from the interglacials MIS 1, 5 and 7 (Fig. 4) and the glacial MIS 6 (Fig. 7) in the BL00-1E record. An example of this is the record in the BL00-1E pollen record of global cooling events as the Montaigu substage (Woillard, 1978) during MIS 5c, or the "intra-Eemian cooling event" (Maslin and Tzedakis, 1996) during MIS 5e as in Clear Lake (Adam et al., 1981), Santa Barbara Basin (Heusser, 2000) in North America, (Fig. 5), as well as in some European records (e.g. Tzedakis et al., 1997). During MIS 6 millennial-scale variations in the pollen and isotopic record (Bright et al., 2006; Fig. 7) from Bear Lake bear close temporal similarities to those observed during the most recent glaciation (MIS 2–4; Fig. 6). We have identified eight significant Heinrich-like cooling events and seven subsequent warming periods between stages 7a and 5e (Fig. 7). A very clear correlation exists between the variability observed in the pollen data and $\delta^{13}C$ values, indicating a strong connection between climate variability and changes in water chemistry. Where discrepancies exist between these two paleoclimate proxies and the $\delta^{18}O$ (Fig. 7), we ascribe these to variations in the response

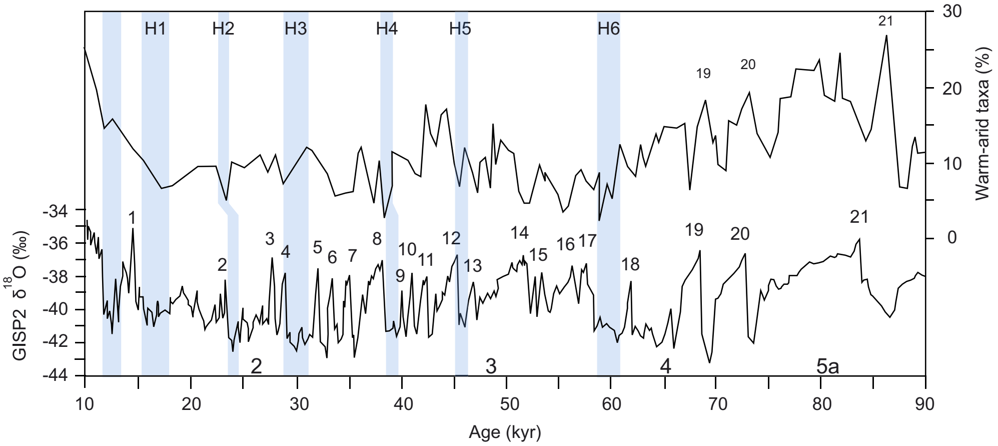

Fig. 6. Millennial-scale vegetation and climate change for the 10–90 ka interval at Bear Lake. Comparison of the relative abundance of warm-arid taxa (*Ambrosia*, Chenopodiaceae-*Amaranthus*, *Sarcobatus*, *Quercus* and *Juniperus*) from BL00-1E pollen record with the variations in $\delta^{18}O$ composition in ice in the GISP2 ice record, Greenland (Grootes et al., 1993). H1 to H6 represent positions of Heinrich events (Bond et al., 1993; Bond and Lotti, 1995). Proposed interstadial (Dansgaard et al., 1993; Grootes et al., 1993) correlation with the pollen record (numbers 19–21).

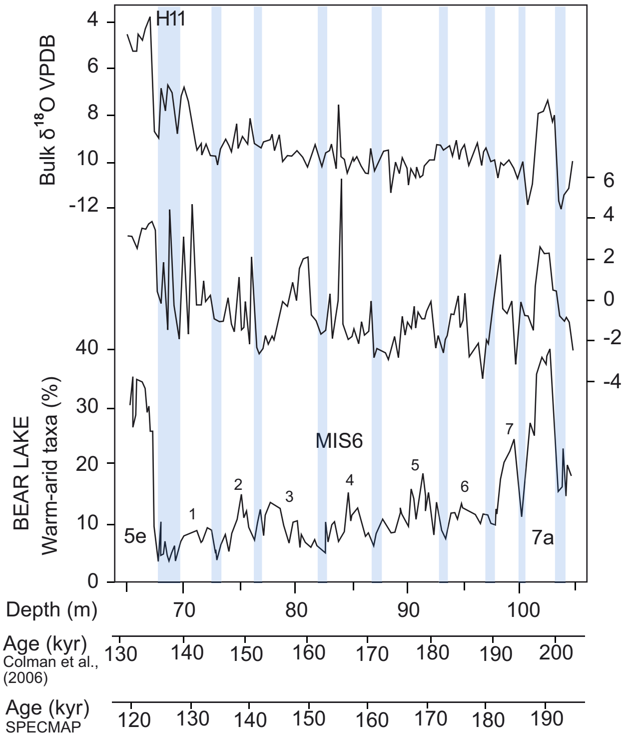

Fig. 7. Millennial-scale vegetation and climate change during MIS 6 in the BL00-1E core. Comparison of the relative abundance of warm-arid taxa (*Ambrosia*, Chenopodiaceae-*Amaranthus*, *Sarcobatus*, *Quercus* and *Juniperus*) with the variations in bulk sediment $\delta^{18}O$‰ VPDB and bulk sediment $\delta^{13}C$‰ VPDB from the BL00-1E core (Bright et al., 2006). Blue shadings highlight Heinrich-like cold periods. Numbers 1–7 indicate warm stadials within MIS 6. Ages from age model for the BL00-1E core (Colman et al., 2006) and SPECMAP (Martinson et al., 1987).

and sensitivity of the different proxies to environmental or climate change (see Section 4.1 above).

The two spectral analysis runs (either the Colman et al. (2006) or linear age models) on the BL00-1E pollen time series using the warm-arid taxa both show prominent peaks at 100 ka, with a 22 ka (Colman et al. model) or a 21 ka (linear model) peak (Fig. 8), correlating precisely with the periodicities of the Earth's precession and eccentricity cycles, respectively. Although we recognize a distinct 100 ka periodicity in our analysis, the fact that only a little over two full eccentricity cycles (225 ka) are represented in the Bear Lake record precludes a definitive assessment of this cycle. However, the 21–22 ka peak correlating to the precession cycle is very strong in this record. Two smaller and less significant cycles are also present, centered on 6–7 and 5 ka periodicities (Fig. 8), confirming additional suborbital-scale climatic variability in the pollen record of Bear Lake. Though variations at these periodicities are evident throughout the pollen time series (Figs. 6 and 7), the ca 6 ka cycle is most obvious in the pollen and isotope record during the penultimate glaciation (Fig. 7).

The suborbital component of climatic variability observed in this core appears to have parallels in other paleoclimatic time series. The variability at the 6 ka frequency in the Bear Lake pollen is comparable to the (a) ca 6.7 ka periodicity recorded in abundance cycles of herb pollen in South China Sea (Luo et al., 2005); (b) 7 ka periodicity in loess records from 70–120 ka in China (An and Porter, 1997); (c) 7.5 ka periodicity in sediment-lightness record within MIS 5 sediments in the North Atlantic (Chapman and Shackleton, 1999); and (d) the ca 8.4 ka periodicity recorded in abundance cycles of the marine alga *Florisphaera profunda* in equatorial Atlantic (McIntyre and Molfino, 1996). Periodicities very similar to the 5 ka have also been identified in North Atlantic records from the last glacial cycle (Chapman and Shackleton, 1998, 1999; An and Porter, 1997) and in a pollen record from South China Sea (Luo et al., 2005). It is noteworthy that 7 and 5 ka periodicities are the 3rd and 4th harmonics of the precession cycle suggesting that this band of millennial-scale variability is related to precession (e.g., Berger et al., 2006).

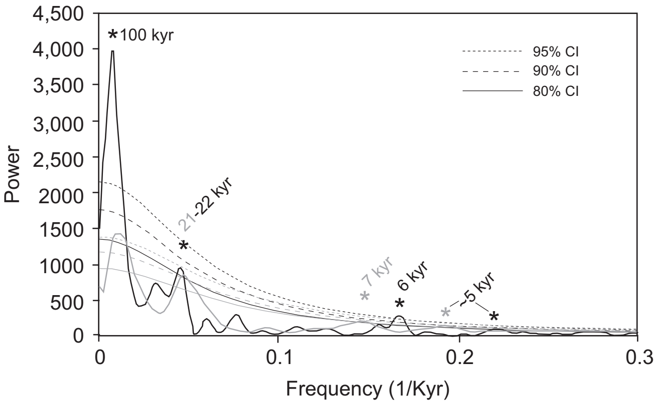

Fig. 8. Spectral analysis of BL00-1E pollen time series on warm-arid taxa (*Ambrosia*, Chenopodiaceae-*Amaranthus*, *Sarcobatus*, *Quercus* and *Juniperus*), using the Colman et al. (2006) age model (black) and linear age model (grey). The confidence intervals (80%, 90% and 95%) are shown by different lines. The spectral analysis was done using REDFIT (Schultz and Mudelsee, 2002).

5. Conclusions

The pollen record from Bear Lake is arguably the most detailed and continuous record of vegetation change for the last 225 ka in North America. Although the Bear Lake sedimentary record is similar in many regards to patterns of paleoecologic and paleoclimatic change from other long records in western North America, it differs substantially in that it extends the paleoenvironmental record back to MIS 7 with sampling spacing of ca 600-year intervals.

Though potential problems exist in the dating of this record (Colman et al., 2006), the original Colman et al. age model and the linear model show very similar results. Our record reflects changes in vegetation and climate that correlate well with orbital-scale variations in summer insolation and global ice-volume. This confirms a tight coupling between global climate and vegetation change from sites beyond the limits of Pleistocene glaciation. The Bear Lake paleovegetation record is particularly sensitive to, and undoubtedly influenced by, climatic cycles related to Earth's precession, obliquity and eccentricity parameters. Suborbital climatic fluctuations reflected in the pollen data, with cyclicities of 6–7 and 5 ka, have also been identified. These are characteristic of the entire pollen record and bear similarities to the records of Greenland and Antarctic atmospheric temperatures, as well as to North Atlantic ice-rafting events. We believe that our analysis here is one of the first continental records to show millennial-scale variability and Heinrich-like events in stage MIS 6, demonstrating the importance of high-resolution pollen analyses over long timescales, and the high sensitivity of the regional vegetation not only to orbital-scale but also to millennial-scale climatic variability. The concordance of terrestrial, oceanic and atmospheric paleoclimatic records supports the conclusion that the high-frequency climatic variability represents widespread changes in the global climate system. This, in turn, is particularly important in the efforts of the scientific community to predict the consequences of future climate change.

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