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Paleomagnetism and environmental magnetism of GLAD800 sediment cores from Bear Lake, Utah and Idaho

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ABSTRACT

A ~220,000-year record recovered in a 120-m-long sediment core from Bear Lake, Utah and Idaho, provides an opportunity to reconstruct climate change in the Great Basin and compare it with global climate records. Paleomagnetic data exhibit a geomagnetic feature that possibly occurred during the Laschamp excursion (ca. 40 ka). Although the feature does not exhibit excursional behavior (≥40° departure from the expected value), it might provide an additional age constraint for the sequence. Temporal changes in salinity, which are likely related to changes in freshwater input (mainly through the Bear River) or evaporation, are indicated by variations in mineral magnetic properties. These changes are represented by intervals with preserved detrital Fe-oxide minerals and with varying degrees of diagenetic alteration, including sulfidization. On the basis of these changes, the Bear Lake sequence is divided into seven mineral magnetic zones. The differing magnetic mineralogies among these zones reflect changes in deposition, preservation, and formation of magnetic phases related to factors such as lake level, river input, and water chemistry. The occurrence of greigite and pyrite in the lake sediments corresponds to periods of higher salinity. Pyrite is most abundant in intervals of highest salinity, suggesting that the extent of sulfidization is limited by the availability of SO_4^{2-}. During MIS 2 (zone II), Bear Lake transgressed to capture the Bear River, resulting in deposition of glacially derived hematite-rich detritus from the Uinta Mountains. Millennial-scale variations in the hematite content of Bear Lake sediments during the last glacial maximum (zone II) resemble Dansgaard-Oeschger (D-O) oscillations and Heinrich events (within dating uncertainties), suggesting that the influence of millennial-scale climate oscillations

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can extend beyond the North Atlantic and influence climate of the Great Basin. The magnetic mineralogy of zones IV–VII (MIS 5, 6, and 7) indicates varying degrees of post-depositional alteration between cold and warm substages, with greigite forming in fresher conditions and pyrite in the more saline conditions.

INTRODUCTION

Lakes occupying tectonic depressions in the Great Basin have accumulated nearly continuous Quaternary sediment records following their late Tertiary subsidence. These lakes, and the sedimentary processes within them, are sensitive to changes in temperature and precipitation (Benson et al., 1990; Morrison, 1991; Oviatt et al., 1992; Grayson, 1993). As a result, our understanding of Quaternary climate change and its influence on the western United States can be improved by studying the sediments within them. A key requirement for interpreting and correlating any sedimentary record in the context of climate change is a robust age model. Accurate dating is challenging because many of the dating techniques have varying reliability (Colman et al., 2006) and the magnitude of ^{14}C reservoir effects for these lakes is not well known (Benson, 1999). Studies of the past direction and strength of Earth's magnetic field have the potential to provide another method to address some of the questions about timing of climatic events in the Great Basin, with respect to records elsewhere, including the North Atlantic. For sedimentary records shorter than 0.78 m.y. (like our Bear Lake record), there are three types of geomagnetic behavior that can be used to help constrain the chronology. The first type, excursions, are brief (i.e., 10^3 years) but significant departures from the geocentric axial dipole, which describes the behavior of Earth's magnetic field as a "bar magnet" centered on the spin axis of Earth. During these departures, the magnetic field approaches (and commonly attains) a polarity reversal and then returns to its previous state (King and Peck, 2001). The second type of geomagnetic behavior is secular variation. Paleomagnetic secular variation (PSV) is described as short-term (10^2–10^4 years) changes in the non-dipole component of Earth's magnetic field. This secular variability can result in as much as a 30–40° drift of the geomagnetic pole away from the geographic pole and have a regional influence of 3000–5000 km at Earth's surface (King, 1983; Lund, 1996; King and Peck, 2001). The third form of geomagnetic behavior commonly used in constructing an age model is based on changes in the strength (intensity) of Earth's magnetic field over time. The intensity of Earth's magnetic field has both a non-dipole and a dipole component and has changed through time, making it useful for regional correlations on secular variation (SV) time scales and global correlations on longer time scales (10^3-10^6 years) (King, 1983; King and Peck, 2001). Sedimentary records characterizing changes in all three of these geomagnetic behaviors have been used successfully for correlating and dating lacustrine sequences (Liddicoat and Coe, 1979; King, 1983; King et al., 1983; Negrini et al., 1984; Lund et al., 1988; Liddicoat, 1992; Lund, 1996; Brachfeld and Banerjee, 2000; Lewis et al., 2007).

The reliability of paleomagnetic results may be compromised by physical deformation of sediments during or after coring. Tests for sediment deformation can be made by multiple sampling of the sections (Thompson, 1984) and by measurement of magnetic fabric (Rosenbaum et al., 2000). Paleomagnetic reliability can also be compromised by post-depositional chemical alteration of original detrital magnetic minerals (Snowball and Torii, 1999; Roberts et al., 2005; Sagnotti et al., 2005). The degree and effects of such alteration can usually be recognized by a combination of mineral magnetic, petrographic, and geochemical analyses (Reynolds et al., 1994). Although some post-depositional authigenic minerals (e.g., greigite) can record the directional characteristics of Earth's magnetic field at the time of mineral formation, it is often impossible to determine when such minerals formed relative to deposition of the surrounding sediments (Roberts et al., 2005; Sagnotti et al., 2005).

Many of the approaches used to evaluate paleomagnetic reliability also form the backbone of environmental magnetism applied to interpreting the depositional and post-depositional processes (physical, chemical, and biological) responsible for magnetic signals (Thompson and Oldfield, 1986; Reynolds and King, 1995; Verosub and Roberts, 1995). Such magnetic records may provide valuable information about a range of paleoenvironmental factors, such as (1) conditions of weathering, erosion, and sediment transport in the catchment; (2) conditions of sediment transport and deposition in the lake; and (3) chemical conditions in lake water and sediments. The goal of this study is to examine these factors to assess the paleomagnetic directional signal recorded in the sediments of Bear Lake and to interpret changes in the depositional and post-depositional conditions as they relate to glacial and interglacial conditions within the Bear Lake catchment during the past ~220,000 years.

GEOGRAPHIC AND HYDROLOGIC SETTING

Bear Lake is located in northeastern Utah and southeastern Idaho (Fig. 1). The lake is contained within the half-graben Bear Lake Valley (McCalpin, 1993) and has a maximum depth of 63 m, with a mean depth of 28 m (Birdsey, 1989). The present lake surface is 1805 m above sea level. The Bear River Range (Fig. 1) captures most of the moisture from the prevailing westerlies, yielding average annual precipitation of 125 cm yr^{-1} in the western portion of the catchment (http://wcc.nrcs.usda.gov/snow/), mostly occurring in the winter. In contrast, the eastern portion of the catchment receives an average of 30.5 cm yr^{-1}, slightly more during the winter (http://wrcc.sage.dri.edu/summary/climsmid.html). Prior to ca. 1918, when a series of canals was constructed to divert the Bear River into the lake, Bear Lake was topographically closed and evaporation dominated the hydrologic system throughout the Holocene (Dean, this volume). However, geomorphic and stratigraphic evidence indicates that the level of Bear Lake exceeded its present level several times during the Quaternary (Laabs and Kaufman, 2003). When the lake expanded, it may have captured the natural channel of the Bear River whose headwaters are in the Uinta Mountains located ~150 km south of Bear Lake.

METHODS

Two holes, separated by a few tens of meters, were continuously cored to depths of 100 m (BL00-1D) and 120 m (BL00-1E). The holes were located at 41°57'06"N, 111°18'30"W, at a water depth of 54.8 m. The core sampling tools used were developed by DOSECC (Drilling, Observation and Sampling of the Earth's Continental Crust), based on ODP and commercial designs, to recover-high quality core samples. They include a hydraulic piston corer (HPC), a non-rotating, extended shoe corer, and a rotating, extended core bit corer (the Alien). Both the HPC and extended shoe corer recover 6.6 cm diameter cores, whereas the Alien corer recovers 6.1 cm diameter cores. The HPC was used to recover the uppermost 45 m of sediment in both holes. In hole BL00-1D, the Alien corer was used from 45 to 69 m depth. The extended shoe corer was used for the remaining 31 m of hole BL00-1D (69–100 m) and for the lower 75 m of hole BL00-1E (45–120 m). During the initial core description process, the magnetic susceptibility (K) was measured at 2 cm intervals with a Bartington Instruments loop sensor (80-mm diameter) attached to a GEOTEK® multi-sensor core logger, and ~250 unoriented samples were collected in 3.2 cm^3 non-magnetic plastic boxes. After the initial core description phase, u-channel samples were taken from the split core surface.

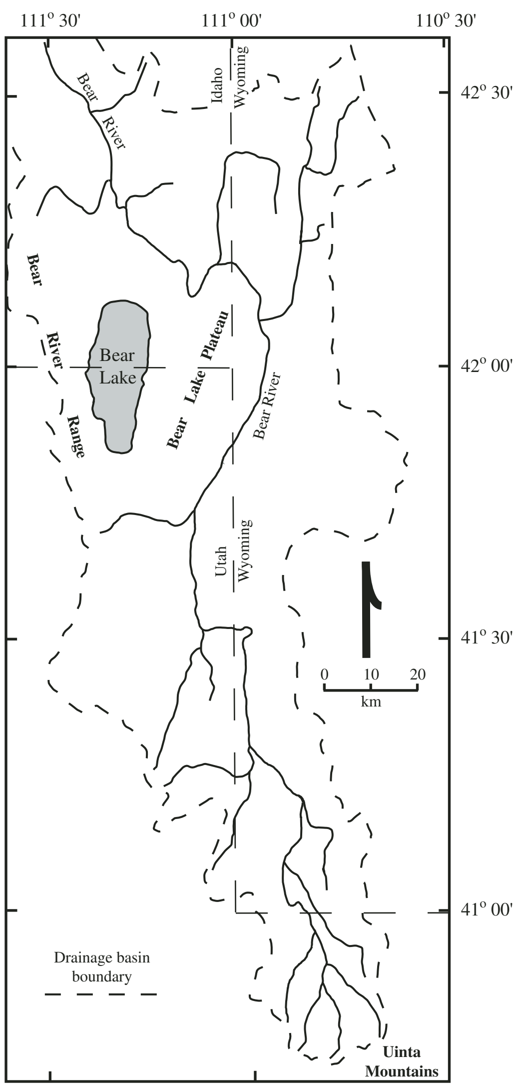

Figure 1. Location map of Bear Lake on the Utah and Idaho border. The curved dashed line delineates the drainage basin of the Bear River (solid black line). The Uinta Mountains are located in the southeastern corner of the map.

Natural remanent magnetization (NRM) was measured at 2 cm intervals along the u-channel samples using a 2-G® Enterprises small-access cryogenic magnetometer (model 755–1.65 UC) at the paleomagnetic laboratory of the Graduate School of Oceanography at the University of Rhode Island. The samples were subjected to stepwise alternating field demagnetization at seven fields (0, 5, 10, 15, 20, 30, and 40 mT). After the 40 mT step, the magnetization was typically less than 40% of its initial value and no further demagnetization was performed. Following demagnetization, the characteristic remanent magnetization (ChRM) and maximum angular deviation (MAD) values were calculated using the software created and described by Mazaud (2005).

Nineteen extra samples were taken directly from the u-channels where a distinct low in inclination was identified. These samples were placed in plastic boxes and used for measuring anisotropy of magnetic susceptibility (AMS) to determine whether the directional data were the result of core deformation or whether these recorded a geomagnetic feature. The AMS measurements were made with a KLY-2 Kappa bridge.

Following NRM measurements, anhysteretic remanent magnetization (ARM) was imparted in a 100 mT alternating field with a bias field of 100 µT (79.6 Am^{-1}). The u-channels were then demagnetized at the same levels as the NRM in order to generate NRM/ARM ratios for paleointensity reconstructions (Meynadier et al., 1992; Yamazaki and Ioka, 1994; Lehman et al., 1996). Also, the susceptibility of anhysteretic remanent magnetization (K_{ARM}) was determined by dividing the initial ARM intensity by the bias field (79.6 Am^{-1}).

Isothermal remanent magnetizations (IRM) were imparted to the u-channel samples, following measurement of ARM. Samples were first subjected to a steady field of 1.2 T using a direct-field CENCO electromagnet, from which the resulting magnetization is referred to as the saturation IRM (SIRM). Samples were then given a backfield IRM (BIRM) in an oppositely directed steady field of 0.3 T. The S parameter and "hard" IRM

(HIRM) were calculated using the SIRM and BIRM as described in King and Channell (1991). Measurement of SIRM was problematic for u-channel samples from depths greater than 30 m, because the imparted magnetization approached the upper limits of the system's sensitivity. For this reason, we adopted the method described by Roberts (2006), who suggested imparting an IRM at 1.2 T and then demagnetizing the resulting IRM at 100 mT before measurement. The measured value is here called IRM_{100mT}. In addition to the u-channel measurements, the mineral magnetic properties of the ~250 discrete magnetic samples were characterized. For the discrete samples, IRM was measured with a AGICO JR-5 high-speed spinner magnetometer.

The magnetic properties reflect types, amounts, and magnetic grain sizes of magnetic minerals in the sediments. For sediments lacking in ferrimagnetic minerals (e.g., titanomagnetite and greigite) K reflects the content of paramagnetic minerals and of diamagnetic materials (e.g., calcite, quartz, and water). Magnetic Susceptibility (K), ARM, and IRM commonly reflect the content of ferrimagnetic minerals (for example, magnetite, titanomagnetite, titanohematite, and greigite) but differ in their sensitivity to variations in magnetic grain size (Dunlop and Özdemir, 1997). The ferrimagnetic minerals are characteristically strongly magnetic and have relatively low coercivities. The HIRM parameter is a measure of the concentration of high-coercivity minerals (e.g., hematite and goethite). For ferrimagnetic grains large enough to carry remanent magnetization (≥0.03 µm), K has a weak dependence on magnetic grain size, increasing somewhat as grains increase from single domain to multi-domain. For such remanence-carrying grains, both IRM and ARM decrease with increasing grain size so that large multi-domain grains contribute relatively little in comparison to smaller grains. In comparison to IRM, ARM varies more strongly for very small grains (<1 µm), with ARM being particularly strong for magnetite grains with diameters on the order of 0.1 µm or smaller. Because of the differing dependence of these magnetic properties on the size of ferrimagnetic grains, the ratios ARM/K, IRM/K, and ARM/IRM provide convenient qualitative measures of "magnetic grain size" (with higher values indicating finer sizes). Grain-size interpretations based on these ratios assume that K, ARM, and IRM reflect the same ferrimagnetic minerals. This assumption is commonly violated when K is significantly affected by paramagnetic and or diamagnetic material, when IRM contains a large hematite component, and when there are major changes in ferrimagnetic minerals.

Both low- and high-coercivity minerals (e.g., magnetite and hematite, respectively) acquire significant magnetization at 1.2 T. Demagnetization at 100 mT usually removes most of the magnetization from low-coercivity, ferrimagnetic phases, but has little effect on high-coercivity phases. When a significant amount of single-domain ferrimagnetic material (e.g., greigite) is present, however, the ferrimagnetic component dominates IRM_{100mT} because the 100 mT field is not large enough to demagnetize these extremely stable, highly magnetic grains. Greigite, an authigenic ferrimagnetic Fe-sulfide mineral, typically occurs as dominantly single-domain size grains (Roberts, 1995; Snowball, 1997). Therefore, IRM_{100mT} either provides a measure of the content of high-coercivity minerals such as hematite or is an indicator for the presence of greigite. The ambiguity in the interpretation of IRM_{100mT} can be largely avoided by using an additional screen for the presence of greigite described by Reynolds et al. (1994, 1998). Single-domain greigite acquires a large IRM in a 1.2 T field but is largely unaffected by the 100 mT alternating field used to impart ARM. Therefore, samples characterized by high values of SIRM/K and relatively low values of K_{ARM}/K are likely to contain significant amounts of greigite.

First-order reversal curves (FORCs) were generated to further characterize some of the magnetic grain size and mineralogical changes. Following the methods of Roberts et al. (2000) and Pike et al. (2001), a set of 200 partial hysteresis curves was generated from 39 samples from the Bear Lake section using a Princeton Measurements Corporation, Alternating Gradient Magnetometer. Samples were chosen on the basis of magnetic property variations and petrographic observations (Reynolds and Rosenbaum, 2005). FORC diagrams were generated using the FORCOBELLO program of Winklhofer and Zimanyi (2005) with a smoothing factor of five. In sediments from Bear Lake, with a complex mixture of magnetic minerals, the usual hysteresis parameters can be ambiguous because of the complexity of grain interactions. FORC diagrams can help identify the presence or absence of magnetostatic interactions and characterize the contributions of different magnetic grain sizes (Roberts et al., 2000; Pike et al., 2001). Magnetostatic grain interactions cause spreading of contours parallel to the H_b axis. The shape of the contours is indicative of magnetic grain size, with single-domain grains producing closed contours and multi-domain grains producing divergent contours (Roberts et al., 2000). The median switching field, a measure of the coercivities of the different magnetic grain sizes and mineralogies in the sample, is plotted on the H_c axis (Roberts et al., 2000; Pike et al., 2001). The FORC diagrams and hysteresis data were used in conjunction with the other magnetic properties and petrographic observations to characterize variations in magnetic mineralogy.

Identification of magnetic minerals can be unambiguously made through direct reflected-light petrographic observations when the mineral grains exceed ~5 µm (Petersen et al., 1986). Less precise identification (e.g., distinguishing Fe-oxides from Fe-sulfides) can be made for grain sizes down to ~1 µm. Reflected-light petrographic analyses were made on magnetic minerals that had been concentrated from bulk sediment with a pumped-slurry separator (Reynolds et al., 2001) and then mounted in epoxy and polished.

RESULTS

Magnetic Properties

Mineral magnetic parameters for the Bear Lake sequence show intervals of distinctive variations related to the input, formation, and preservation of magnetic minerals. Values of K range from near zero in the upper few meters to greater than 60×10^{-5} (SI) in several intervals down-core (Fig. 2). Values of K_{ARM}/K, SIRM/K, and ARM/SIRM display large-magnitude variations with some intervals having consistently low values of one or more ratios, and other zones having higher, more variable values. Similarly, IRM_{100mT} displays large-scale variations with intervals of low values separated by intervals of higher, more variable values. Based on the variations in these magnetic properties, the Bear Lake record was divided into seven zones (Table 1, Fig. 2). Dean (this volume) and Kaufman et al. (this volume) describe the sedimentology of the GLAD800 cores from Bear Lake and identify seven zones that roughly coincide with our mineral magnetic zones (Fig. 2), illustrating the need to consider the bulk sediment variations when interpreting the mineral magnetic signal. Although each peak and trough of the magnetic properties do not correspond exactly to sedimentologic variations, the relationship between carbonate content and the amount of magnetic material is obvious. Intervals containing aragonitic marl have the lowest K values due to dilution of detrital material, dissolution of magnetic Fe-oxides, or reduced input of terrestrially derived silt. Conversely, the intervals with higher silt content (e.g., Zone II) and lower carbonate have the highest K values. The mineral magnetic zones and a brief description of the bulk sedimentology are given below.

Zone I

Zone I spans the uppermost 9 m. This interval has low values of K, IRM_{100mT}, and SIRM/K, and high values of K_{ARM}/K and ARM/SIRM. The low values of K and IRM_{100mT} are due in part to dilution of detrital material by endogenic carbonate and in part to post-depositional destruction of magnetite and hematite (Rosenbaum and Heil, this volume). In addition to dilution and post-depositional alteration, the low silt content of this interval suggests a diminished input of detrital, Fe-oxide-bearing material. The diamagnetic properties of the abundant carbonate contribute to the extremely low values of K. High values of ARM/SIRM and FORC data (Fig. 3A), however, indicate that the small quantity of ferrimagnetic material in this interval has a fine magnetic grain size, single domain (SD) or pseudo–single domain (PSD). We speculate that this fraction is protected from alteration within larger detrital silicate grains and rock fragments, and that the quantity of these protected grains is so small that they affect the grain-size indicators only in the absence of other ferrimagnetic minerals. The upward decrease in ARM/SIRM associated with extremely low values of K may indicate that even these protected grains have been destroyed in the uppermost sediments. Petrographic observations of magnetic separates from Bear Lake cores BL96-1, -2, and -3, located a short distance from the BL00 drill site, show that sediments correlative with zone I contain small quantities of magnetite and hematite, and that pyrite and greigite are either absent or present in minor quantities (Reynolds and Rosenbaum, 2005). Many of the Fe-oxides observed in these samples, including both magnetite and hematite, occur in rock fragments, lending support to the speculation that such inclusions are protected from alteration. It should be noted, however, that grains large enough to be optically identified are too large to explain the fine magnetic grain size discussed above.

Zone II

This zone, which extends from 9 to 18 m (Fig. 2), differs markedly from overlying and underlying sediments. This zone consists of massive, reddish-gray silty clay and contains less than 20% $CaCO_3$ (Dean, this volume; Kaufman et al., this volume). High values of both K and IRM_{100mT} indicate an abundance of ferrimagnetic material. The grain-size-sensitive ratios and FORC data (Fig. 3B) indicate the presence of coarse (multi-domain, MD) magnetic grains, and the FORC data also indicate the subordinate presence of SD, high-coercivity minerals as well. The contribution of the high-coercivity minerals is overwhelmed by the more magnetic and coarser low-coercivity minerals in the grain-size-sensitive ratios. Petrographic observations in correlative sediments from other cores (Reynolds and Rosenbaum, 2005) indicate that magnetite, titanomagnetite, and hematite control magnetic properties. The magnetic minerals represent a wide variety of compositions, textures, and sizes including homogeneous magnetite grains tens of micrometers in diameter. Little dilution of detrital material by endogenic carbonate (Fig. 2) and good preservation of magnetite account for high values of K. These factors also contribute to high hematite content (indicated by the high IRM_{100mT} values) having multiple origins, including input of hematite-rich rock fragments from the Uinta Mountains (Reynolds and Rosenbaum, 2005; Rosenbaum and Heil, this volume).

Zone III

This zone, which extends from 18 to 41 m, is characterized by K and ARM/SIRM values intermediate to those of zones I and II (Fig. 2), and low values of IRM_{100mT}. The sediment is calcareous clay with centimeter-scale gray to greenish-gray bands that are massive and show signs of bioturbation. In addition, there is silt-sized quartz, diatoms, and >3% organic carbon (Dean, this volume; Kaufman et al., this volume). The most abundant mineral in a magnetic separate from a depth of 35.07 m is titanohematite (Reynolds and Rosenbaum, 2005). The sample contains a small amount of magnetite, but most magnetite has been dissolved or replaced by pyrite. It is difficult to compare the degree of alteration between zones I and III using the petrographic data, but the difference in carbonate content between the two zones largely accounts for the difference in K, suggesting that the degree of alteration is similar. FORC data (Fig. 3C) indicate the presence of coarser (MD), low-coercivity minerals with a contribution of finer (SD), slightly higher coercivity minerals (indicated by the less divergent contours extending to higher fields). These "higher coercivity," SD minerals should not be confused with the high-coercivity, coarser minerals identified in zone II (e.g., hematite). They are low-coercivity minerals but have a slightly higher coercivity than the other low-coercivity minerals present (note: for the remainder of this paper, the phrase "higher coercivity" is used in reference to these minerals). It is likely that this mixture of SD,

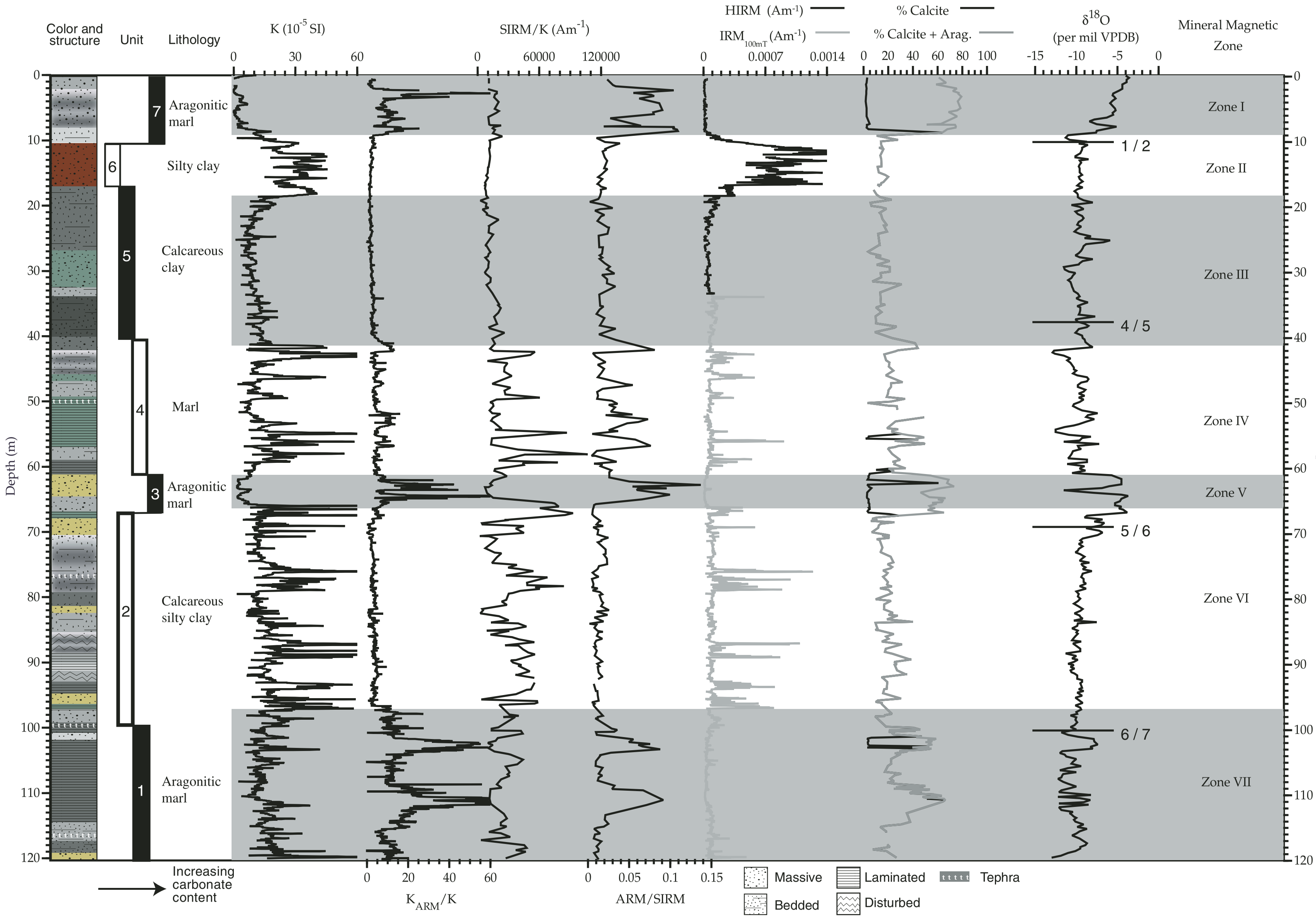

Figure 2. Magnetic properties, calcite and calcite + aragonite content from Dean (this volume), and $\delta^{18}O$ data with interpreted stage boundaries from Bright et al. (2006). The SIRM/K and ARM/SIRM data are from discrete samples and are lower resolution than the u-channel and core data (K, K_{ARM}/K, HIRM, and IRM_{100mT}). The sedimentologic description and summary (sediment color, structure, unit, and lithology) are taken from Kaufman et al. (this volume) and are based on the initial core descriptions conducted at the National Lacustrine Core Repository (LaCore) at the University of Minnesota, Minneapolis.

TABLE 1. GENERALIZED CHARACTERISTICS OF MAGNETIC MINERAL ZONES FROM BEAR LAKE

Magnetic zones	Magnetic concentration (K, M_{rs}, M_s)	Magnetic grain size (FORC, M_{rs}/M_s, H_{cr}/H_c)	Magnetic mineralogy (H_{cr}, petrography, IRM_{100mT})	Lake conditions
I: (0–9 m, 0–17 ka)	Low	SD and MD (0–2 m MD only)	High proportion of low-coercivity Mt and Ti-ht as well as Py	High salinity and reducing conditions during present interglacial
II: (9–18 m, 17–33 ka)	High	SD and MD	Mixture of low-coercivity, MD Mt and Ti-ht and high-coercivity, SD Ht	Lake transgression captures Bear River, which carries Fe-oxide minerals from Uinta Mts. during the LGM
III: (18–41 m, 33–76 ka)	Moderate	MD (occasional SD)	High proportion of low-coercivity, MD Ti-ht, as well as Py with very sparse higher-coercivity,* SD Gt	Saline and reducing conditions, but perhaps less than zone I because greigite is still sparsely present
IV: (41–61 m, 76–113 ka)	Variable	SD and MD	Mixture of low-coercivity MD Ti-ht and higher-coercivity, SD Gt	Fluctuating salinity; greigite formation during fresher intervals
V: (61–66 m, 113–122 ka)	Low	MD (and much less SD)	Mostly low-coercivity, MD Ti-ht and pyrite with sparse higher-coercivity, SD Gt	Similar to present interglacial conditions; high salinity and reducing conditions
VI: (66–97 m, 122–180 ka)	Variable	SD (and MD to a much lesser extent)	High proportion of higher-coercivity, SD Gt and sparse low-coercivity, MD Ti-ht and sparse Py	Salinity fluctuations similar to zone IV, with greigite more prominent in fresher intervals
VII: (97–120 m, 180–222 ka)	Variable	SD and MD	Alternates between higher-coercivity, SD Gt and low-coercivity, MD Ti-ht	High salinity and reducing conditions, perhaps slightly less than zones I and V but more than zones III, IV, and VI

Note: Magnetic grain size: SD—single domain; MD—multi-domain. Magnetic mineralogy: Mt—magnetite; Ht—hematite; Ti-ht—titanohematite; Py—pyrite; Gt—greigite. LGM—last glacial maximum. See text for explanation of other abbreviations.

*Although greigite is a low-coercivity mineral, single-domain greigite tends to have a higher coercivity than magnetite and titanohematite (Roberts, 1995). Here (and in the text) the phrase "higher coercivity" is used in regard to greigite, but is not to be confused with "high coercivity" as used in regard to hematite.

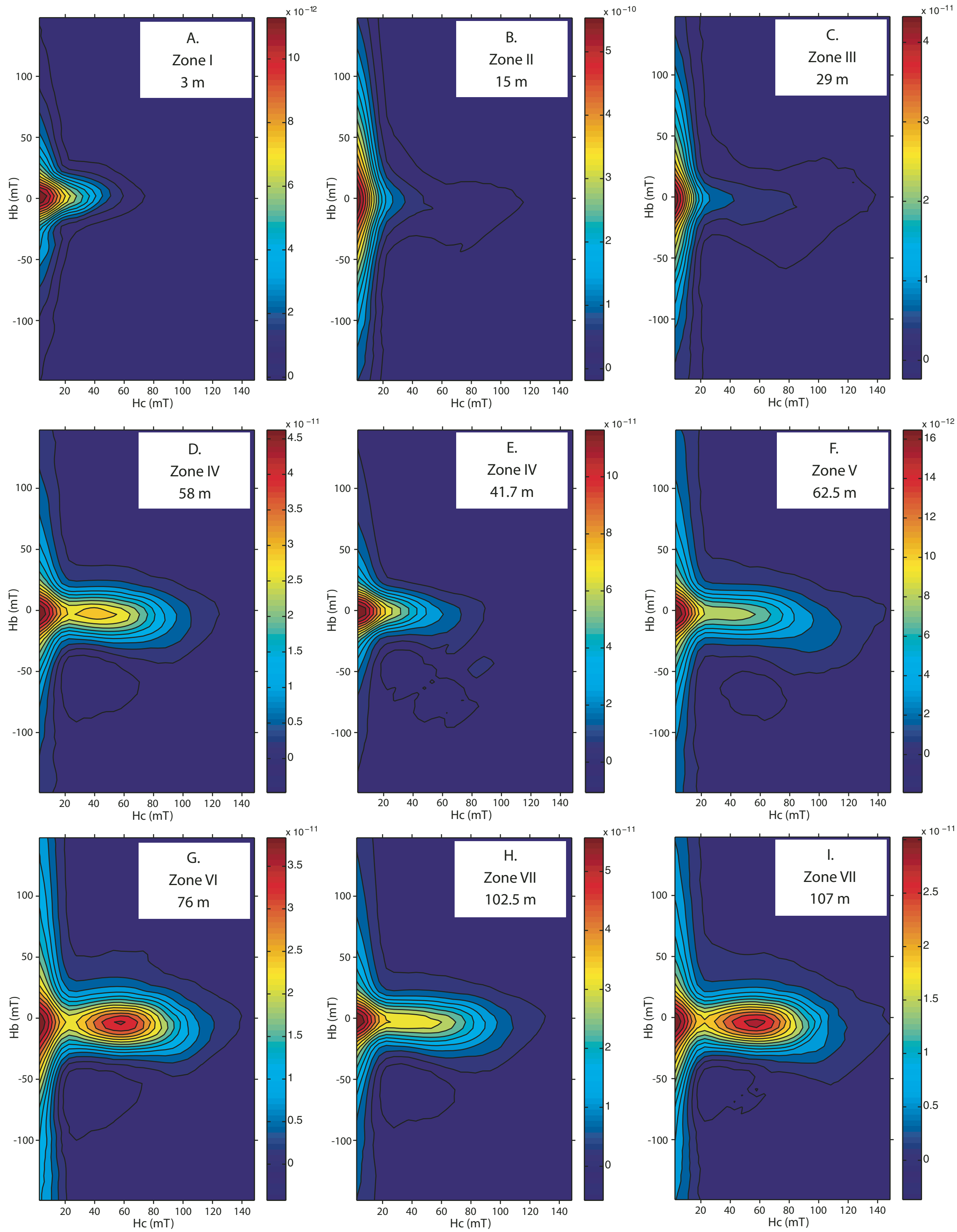

Figure 3. Representative first-order reversal curve (FORC) diagrams from each of the magnetic zones. The H_b axis shows the degree of grain-to-grain interaction by the amount of vertical spread in the contours. For instance, there will be little or no vertical spread in the contours for a sample consisting of non-interacting single-domain magnetic minerals. Also, the presence of a mean stabilizing field (such as that resulting in the alignment of the magnetic minerals in the sample) would be represented by a displacement of the distribution below the H_b axis. The H_c axis indicates the range of coercive fields (in mT) identified for the different minerals present in the sample, which is a function of magnetic grain size and mineralogy. The scale bar next to each FORC diagram indicates the proportion (or concentration) of the different coercive fields identified in a particular sample, with warmer colors indicating higher concentrations.

higher-coercivity minerals and MD, low-coercivity minerals is the cause of the intermediate values of ARM/SIRM.

Zone IV

This zone extends from 41 to 61 m. The zone is characterized by highly variable values of all magnetic parameters, indicating large variations in the types and quantities of magnetic minerals. The zone is a diatomaceous marl with centimeter-scale gray banding, variable bioturbation, and variations in $CaCO_3$ that often exceeds 20% (Dean, this volume; Kaufman et al., this volume). Magnetic separates were examined from two samples within this interval. Titanohematite appears to be the most abundant phase in a magnetic separate from a sample (41.59 m) (Reynolds and Rosenbaum, 2005) that has a low value of IRM_{100mT}. A small quantity of magnetite in this sample is preserved inside silicate grains. The other sample (43.00 m) has a high value of IRM_{100mT}, and petrographic observations show that greigite is the most abundant mineral in the magnetic separate. High values of IRM_{100mT} within this zone are therefore interpreted as an indicator of greigite and not hematite. Titanohematite is a minor component of the magnetic separate from this sample, and there is good evidence of magnetite dissolution. The FORC data indicate variations in the proportions of SD, higher-coercivity magnetic minerals and MD, low-coercivity magnetic minerals (Fig. 3D and E). Most samples with high K also have high IRM_{100mT} and a higher proportion of SD, higher-coercivity minerals (Fig. 3D); conversely, samples with lower K have low IRM_{100mT} values and a lower proportion of SD, higher-coercivity minerals (Fig. 3E). Based on similar magnetic properties and petrographic observations of the two samples, the samples in this zone with lower values of K are thought to have magnetic mineralogy similar to sediment in zone III. The variable magnetic properties in zone IV appear to reflect variable amounts of greigite.

Zone V

This thin zone, extending from 61 to 66 m, is characterized by magnetic properties like those of zone I including extremely low values of K. Petrography indicates that detrital magnetite was destroyed and that the most abundant magnetic mineral is titanohematite. Like zone I, the low content of magnetic minerals is due to the combination of destruction of Fe-oxide grains and dilution of the detrital component by a high carbonate content (>50%) composed largely of aragonite. The FORC diagram from this zone (Fig. 3F) indicates a slightly higher contribution of the SD, higher-coercivity minerals than in zone I (Fig. 3A).

Zone VI

This zone extends from 66 to 97 m. Magnetic properties are highly variable with similar characteristics to those in zone IV. Sedimentologically, this interval is made up of calcareous silty clay, which are gray to greenish-gray and massive with "whispy" black staining suggesting the presence of sulfides (Dean, this volume; Kaufman et al., this volume). Greigite is the most abundant mineral in magnetic separates from three samples with high values of IRM_{100mT} (67.10, 78.33, and 96.2 m), whereas titanohematite is the most abundant mineral in separates from intervals with low values of IRM_{100mT} (83.2 and 97.5 m). As in zone IV, the FORC data indicate variations in the proportions of SD, higher-coercivity minerals (high K and IRM_{100mT} values) and MD, low-coercivity minerals (low K and IRM_{100mT} values; Figs. 2 and 3G) that are apparently due to variations in the amount of greigite.

Zone VII

This zone spans the lowermost portion of the section from 97 m to the bottom of the core and is delineated by the transition to lower, less variable IRM_{100mT}. Although there is a thin interval of aragonitic marl (101.6–103 m), most of the unit is a calcitic marl that is diatomaceous and variably laminated and massive (Dean, this volume; Kaufman et al., this volume). Relative to zones IV and VI, the K values are intermediate and less variable, whereas the K_{ARM}/K values alternate between ~10 and 60 (Fig. 2). Reynolds and Rosenbaum (2005) determined from magnetic separates that greigite is the most abundant mineral in the higher-SIRM/K intervals and titanohematite is the dominant mineral in the lower-SIRM/K intervals. Of the five FORC samples taken within this zone, three were taken from high-SIRM/K intervals (99, 107, and 115.5 m) and two were taken from low-SIRM/K intervals (102.5 and 111 m). The FORC data for the greigite-bearing intervals show a higher proportion of SD and minerals having a higher coercivity, as well as a smaller fraction of MD, low-coercivity minerals (Fig. 3I). The FORC data from the low SIRM/K intervals also show SD, higher-coercivity minerals mixed with MD, low-coercivity minerals, but for these intervals the coarse fraction constitutes a higher proportion (contours are not closed in the higher coercivity range) (Fig. 3H). Despite having somewhat higher K values, the titanohematite intervals of this zone have FORC diagrams and mineral magnetic properties similar to zone V, in which titanohematite is the dominant magnetic mineral (Fig. 3F).

Paleomagnetism

Variable magnetic mineralogies influence the validity of the paleomagnetic directional data (Fig. 4). For instance, the sediments from intervals with a relative abundance of Fe-oxide minerals (zones II and III) record changes in field direction of ±30° around the geocentric axial dipole value, which is typical of secular variation behavior of Earth's magnetic field (Fig. 4; King, 1983; King and Peck, 2001; Lund, 1996). We interpret this behavior to reflect primary magnetization (a depositional remanence). On the other hand, changes in directional properties from sediments with abundant greigite (zones IV, VI, and VII) vary by ±15° (Fig. 4), casting doubt on the paleomagnetic integrity of these zones. The demagnetization behavior of Bear Lake sediments indicates that the magnetic minerals record a stable remanence (Fig. 5). Following the removal of a weak overprint at the 5 mT demagnetization step, the sediments generally demagnetized linearly toward the origin of the vector endpoint diagrams, with <40% of their initial values remaining after the 40 mT level.

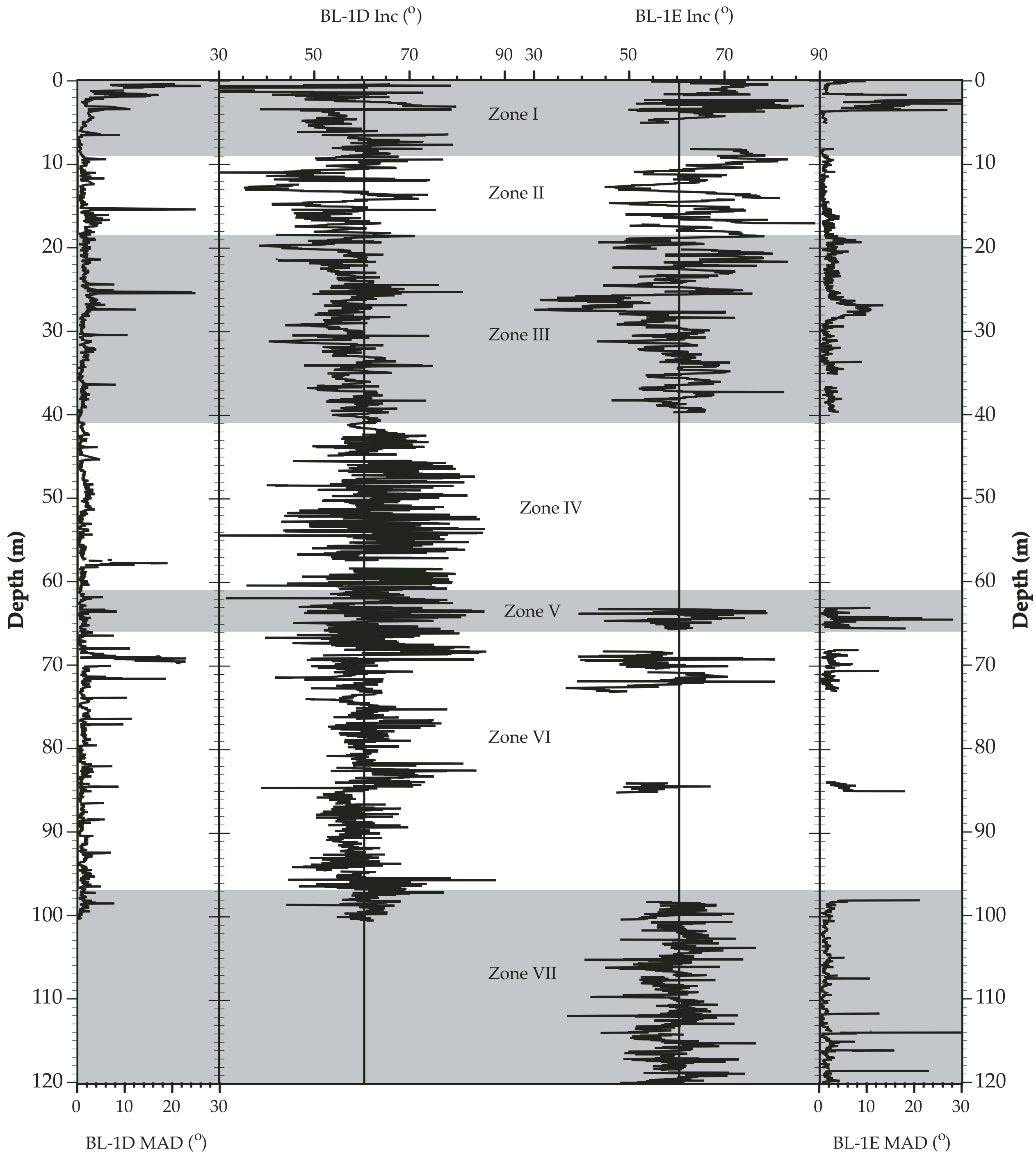

Figure 4. Paleomagnetic inclination data from cores BL00-1D and -1E and mineral magnetic zones from Figure 1. Hole 1E was measured to complete the lowermost 20 m of the Bear Lake section that was not recovered in hole 1D. In addition, the upper ~40 m were measured to illustrate the reproducibility and fidelity of the paleomagnetic record from that interval. The brief intervals of data shown for hole 1E between 40 and 100 m were generated in an attempt to duplicate potential geomagnetic features from hole 1D, which were later determined to reflect sediment disturbance. The maximum angular deviation (MAD) values are shown for each record. MAD values were calculated using the software created by Mazaud (2005).

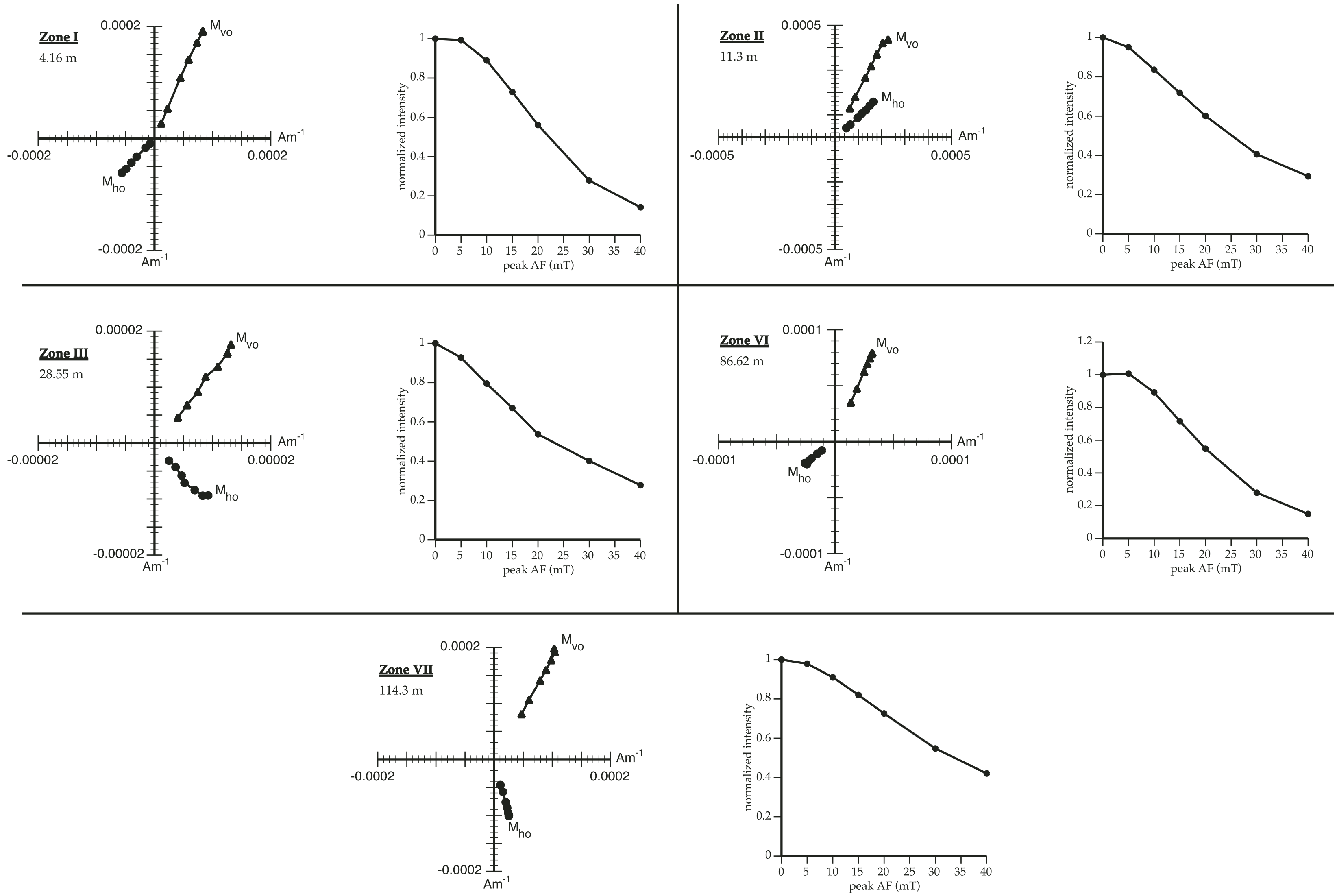

Figure 5. Typical Zijderveld plots from the zones I–III, VI, and VII. Zones IV and V are not shown because directional data from this interval are highly variable due to coring disturbance. M_{vo}—the initial magnetization in the vertical plane, M_{ho}—the initial magnetization in the horizontal plane, Am^{-1}—amps per meter, and AF—alternating field.

A visual comparison of the inclination data from hole BL00-1D and -1E for the interval spanning 0–40 m indicates a reasonably close correspondence between the two data sets, particularly in the 8.5–20.5 m and 29–40 m intervals (Fig. 6). Maximum angular deviation values (MAD; Fig. 4) are generally low (~3°) except for portions of zone I and the interval between ~25–28 m. The high MAD values and the erratic inclination values in the top 4 m of zone I are the result of post-depositional destruction of the remanent carrying detrital Fe-oxides. Between 20.5 and 29 m, the records show poor correspondence. In particular, BL00-1D does not have the same high amplitude variations that are present in hole 1E (Fig. 6). Between 25.5 and 27 m in hole BL00-1E, there is a distinct shallowing of the inclination (Figs. 4 and 6); however, the same feature is not present in the BL00-1D hole. The MAD values are generally higher in this interval for both holes; however, values in hole 1D are more variable, with maxima exceeding 25°, whereas hole 1E values are generally around 10° with one peak value of 13° (Fig. 4). Minimum susceptibility axes from the 20–40 m interval are mostly near vertical, but several samples between 24 and 27 m from hole 1D yielded shallower inclinations of susceptibility axes (Fig. 6). Although the sediments are not visually disturbed, the lack of sedimentologic evidence for rapid deposition (such as turbidites) suggests that these shallowed susceptibility axes result from deformation of the sedimentary

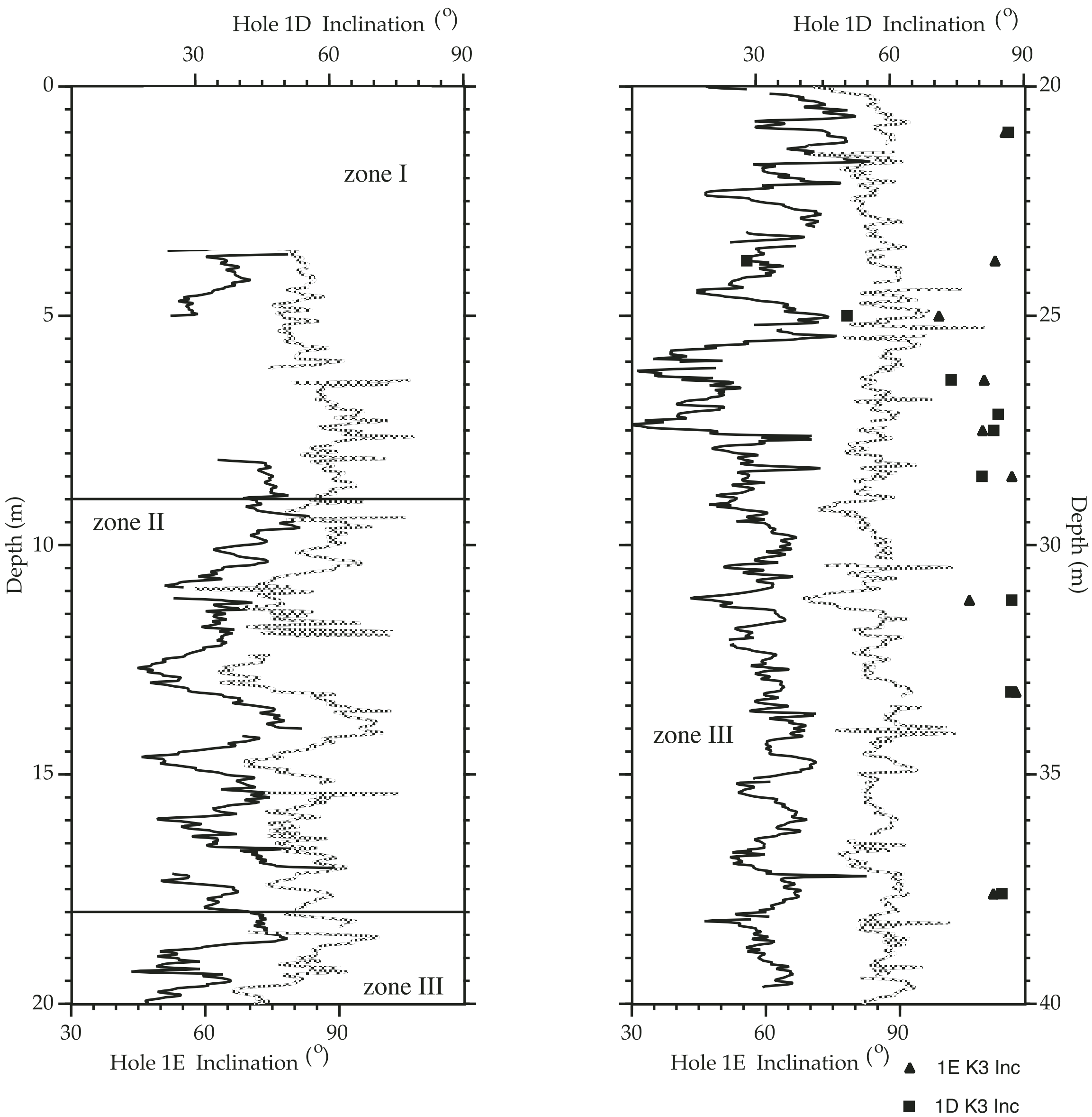

Figure 6. The uppermost 40 m of paleomagnetic inclination data and the inclinations of minimum magnetic susceptibility axes (K_3). Note: (1) There is generally good correlation of paleomagnetic data between the cores in zones II and III. (2) Low inclinations of K_3 axes occur between ~23.5 and 27 m in hole 1D.

fabric (Rosenbaum et al., 2000). From ~40–70 m, the inclination data are highly variable. This behavior is likely an artifact of the change from the hydraulic piston corer to the rotary corer that can cause the sediment to fracture, rotate, or tilt (Poag and Foss, 1985). The lowermost part of the sequence (70–120 m) was cored using the extended nose corer and, as a result, contains less variable paleomagnetic directions with lower amplitude peaks and troughs (Fig. 4). The MAD values below 40 m are generally low except for zone V, in which the environmental magnetic evidence suggests post-depositional destruction of Fe-oxides (Fig. 2) resulting in the loss of remanent-carrying minerals.

In addition to paleomagnetic inclination data, the ratios of NRM intensity to K and IRM and ARM intensities (Fig. 7B) were calculated as possible measures of relative paleointensity. Generally, the mineral magnetic properties in the Bear Lake section do not meet the criteria for geomagnetic paleointensity studies defined by King et al. (1982) that include (1) a uniform mechanism of detrital remanent magnetization, (2) uniform magnetic mineralogy (preferably magnetite), and (3) uniform magnetic grain size. The exception is zone III (18–33 m) in which the magnetic concentration (K), grain size (ARM/SIRM and FORC), and mineralogy (FORC, IRM_{100mT}, and petrography) are uniform, and in which low-coercivity, multi-domain minerals (probably ferrimagnetic titanohematite) are the dominant remanence carriers (FORC diagrams, Fig. 3). Within this zone, the NRM/ARM, NRM/IRM, and NRM/K data show distinctly low values between 25 and 27 m, coincident with low-inclination values (Fig. 7). Representative demagnetization curves for NRM/ARM from the interval indicate that the low values are present regardless of demagnetization level (Fig. 7C).

DISCUSSION

Magnetic Properties

Magnetic properties of the Bear Lake section reflect varying degrees of post-depositional alteration of detrital Fe-oxide minerals and varying quantities of the magnetic sulfide mineral, greigite. Zone II (9–18 m) is the only interval in which detrital Fe-oxide minerals are largely preserved. Sediments within this interval contain large amounts of glacial flour derived from the Uinta Mountains and carried to Bear Lake by the Bear River (Rosenbaum and Heil, this volume). Preservation of Fe-oxides reflects some combination of fresh water, rapid deposition (sulfidization may cease when sediment has been buried to the point that diffusion can no longer replenish sulfate in the pore waters, see Canfield and Berner, 1987), and perhaps large amounts of Fe-oxide minerals.

The relatively uniform magnetic properties through zone III (18–41 m) probably reflect post-depositional conditions. In this zone, detrital magnetite and hematite have been largely destroyed and the most important magnetic mineral is ferrimagnetic titanohematite. Titanohematite, which is relatively resistant to post-depositional alteration (Canfield et al., 1992; Reynolds et al., 1994), probably occurs in small amounts throughout the entire Bear Lake section and becomes important phase when other detrital magnetic minerals have been destroyed.

Zones I and V (0–9 m and 61–66 m, respectively) have similar magnetic mineralogy to zone III but lower concentrations of detrital material due to high carbonate content. The upward coarsening of magnetic grain size in the uppermost 2 m of zone I (indicated by decreasing values of ARM/SIRM) coincides with increasing values of $\delta^{18}O$ (Fig. 2). Bright et al. (2006) suggest that this is an interval of increased salinity. Higher concentrations of sulfate in this interval may have caused more complete destruction of Fe-oxide grains including small magnetite grains enclosed in detrital rock particles. Zones IV (41–61 m) and VI (66–97 m) are characterized by intervals dominated by greigite and intervals with characteristics like those of zone III.

Zone VII (97–120 m) alternates between intervals with greigite (higher SIRM/K values) and intervals with titanohematite (lower SIRM/K values). Dean (this volume) identifies two intervals in this zone that are enriched in calcite + aragonite and interprets this enrichment (particularly the increased aragonite) as more saline lake conditions. The correspondence of the calcite + aragonite peaks with the relative dominance of titanohematite in zone VII suggests that the increased availability of sulfate contributed to sulfidization of the less-resistant Fe-oxides. The salinity of this zone is likely to have been similar, or just slightly less than that of zones I and V (based on the calcite + aragonite data), which are also dominated by the more-resistant titanohematite. The dominance of titanohematite in zone III is not as easily explained by higher salinity conditions since the calcite + aragonite content is not as high as zones I, V, and VII. In this case, the dominance of titanohematite may result from a combination of the reduction of the less-resistant Fe-oxides and a decrease in the amount of Fe-oxides deposited in the lake at that time, i.e., lower salinities (lower sulfate concentrations) are required for sulfidization of fewer Fe-oxides.

Variations in magnetic properties reflect differences in lake conditions during deposition. Zone II records freshwater conditions during a period when the Bear River delivered large volumes of glacial flour from the Uinta Mountains to Bear Lake (Rosenbaum and Heil, this volume). Zones I and V correspond closely to intervals previously correlated with marine isotope stages (MIS) 1 and 5e (Fig. 2; Bright et al., 2006). During these periods, the lake is thought to have been topographically closed, highly evaporative, and productive. Based on high aragonite content (Fig. 2), Dean (this volume) interprets these intervals to represent the most saline conditions in the cored section. Differences in magnetic properties in the other four zones largely reflect the presence or absence of greigite. The concentration of sulfate may be an important factor in determining the fate of Fe-oxide minerals and production of Fe sulfides (Snowball and Torii, 1999).

In a study of Owens Lake, Reynolds et al. (1998) suggest that the formation of greigite was controlled by sulfate availability. If reducing conditions exist and sulfate is not available, Fe may go into solution and be lost from the sediment. If sulfate is abundant,

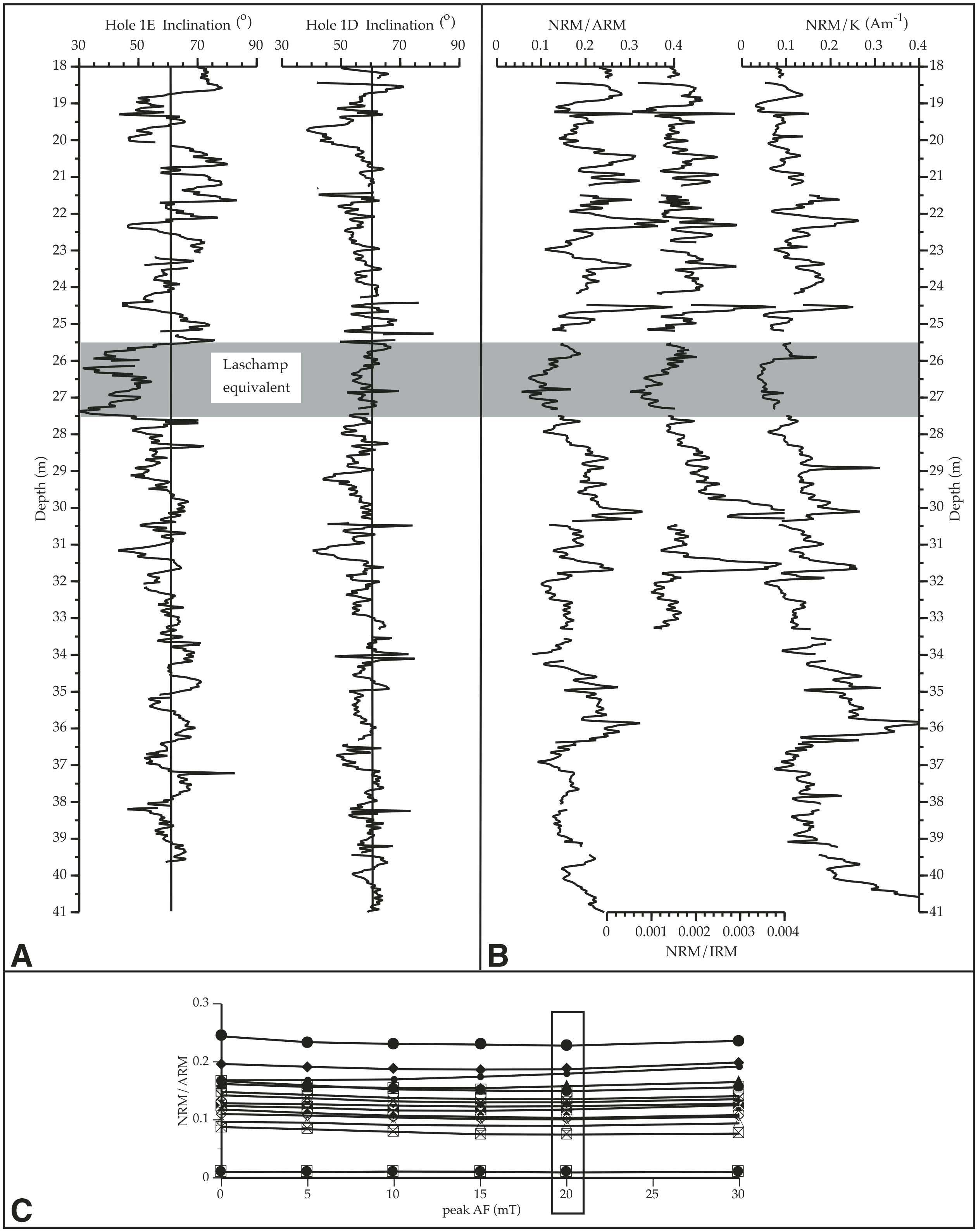

Figure 7. (A) Paleomagnetic inclination data from zone III for BL00-1E and -1D showing the geomagnetic feature interpreted as the time-equivalent of the Laschamp excursion found in higher-fidelity records. The expected geocentric axial dipole value (~61°) for Bear Lake at 42°57'06"N is indicated with solid vertical lines. (B) Relative paleointensity proxy data (NRM/ARM, NRM/IRM, and NRM/K) from BL00-1E showing low values at the time of the interpreted Laschamp equivalent. (C) Representative NRM/ARM data versus demagnetization level from the interval of low relative paleointensity. Constant ratio values regardless of demagnetization level illustrate the ability of the sediments within this interval to reliably record relative paleointensity variations (King et al., 1982). ARM, IRM, NRM—anhysteretic, isothermal, and natural remanent magnetization, respectively.

and therefore not a limiting factor, pyrite is likely to form at the expense of Fe-oxide minerals. Greigite may form under intermediate conditions, when sulfate is present in quantities less than needed to form pyrite. The absence of greigite in portions of the Bear Lake record could indicate more saline conditions (i.e., more sulfate leading to more pyrite upon reduction) or less saline (i.e., less sulfate and Fe-oxide preservation) than conditions for the greigite-bearing intervals. These possibilities cannot be distinguished on the basis of magnetic properties alone. We interpret greigite-free intervals (other than zone II) to indicate more saline conditions because (1) some pyrite occurs in these sediments (Reynolds and Rosenbaum, 2005), (2) intervals deposited under the most saline conditions, as indicated by high concentrations of aragonite, lack greigite, and (3) all other high-carbonate intervals, which reflect elevated salinities, also lack greigite.

Paleomagnetism

Although variable amounts of the detrital remanence-carrying minerals have been removed by post-depositional alteration, detrital Fe-oxide minerals are present in most of the sequence, even if only in small quantities as resistant titanohematite minerals (Reynolds and Rosenbaum, 2005). In addition, the inclination data are centered on the geocentric axial dipole value of 61° (Fig. 4). Excluding zone II, which has a relative abundance of magnetite and hematite, it is likely that a combination of titanohematite minerals and greigite contributes to the paleomagnetic record from Bear Lake. Greigite produces a strong and stable chemical remanence (CRM; Roberts, 1995) acquired post-depositionally. The difficulty associated with relying on greigite as a remanence carrier is the time it takes for the acquisition of the CRM. If greigite is formed either syndepositionally, or soon after deposition, then the CRM may be an accurate recorder of geomagnetic field behavior. If there is a significant lag time associated with the acquisition of the CRM, the paleomagnetic directional signal recorded in the sediments is not representative of the geomagnetic field behavior at the time of deposition.

It is evident that the fidelity of the paleomagnetic record is related to the magnetic mineralogy. The uppermost 4 m of the sequence (top of zone I) has highly variable inclination values that are not correlative between holes 1D and 1E, possibly reflecting a very low abundance of detrital Fe-oxide minerals. In contrast, zone II, which has abundant detrital Fe-oxide minerals, yields a record that resembles typical secular variation (King, 1983; Lund, 1996; King and Peck, 2001), and the inclination data are highly correlative between the two cores (Fig. 6). A close comparison of the 18–40 m interval (zone III) shows that, although the amplitude of the peaks and troughs is damped, the data can be correlated between the two holes (Fig. 6). From 40 to 120 m (zones IV through VII), the sediments have been variably affected by post-depositional alteration, and greigite is relatively abundant. The directional signal throughout this interval seems "muted" compared to zone II. This muted effect is likely the result of the CRM carried by greigite overprinting the DRM carried by the titanohematite.

Paleointensity interpretations from normalized NRM intensity data are problematic due to changes in mineralogy and grain size (Fig. 2). Because of post-depositional changes in magnetic minerals (King et al., 1982), Bear Lake is not a good location to obtain a paleointensity record. However, the sediments of zone III (18–41 m) meet the criteria of King et al. (1982) in that they have (1) a uniform mechanism of detrital magnetization, (2) uniform magnetic mineralogy, and (3) uniform magnetic grain size (Fig. 2). Although magnetite is the preferred magnetic mineral (King et al., 1982), the reproducibility of the inclination data (Fig. 6) and normalized NRM ratios (Fig. 7B) between the two holes suggests that titanohematite provides a sufficient detrital magnetization. In addition, the demagnetization behavior of the NRM/ARM data does not change with demagnetization level (Fig. 7C; see King et al., 1982). On the basis of this evidence, we interpret the NRM/ARM, NRM/IRM, and NRM/K ratios as relative paleointensity records for this interval in the Bear Lake sequence.

Despite post-depositional changes in magnetic mineralogy, the low-inclination values between 25.5 and 27.5 m in hole 1E appear to record a geomagnetic feature (Fig. 7). The absence of this low-inclination feature in hole 1D is probably due to core disturbance, as suggested by the AMS data (Fig. 6) and higher MAD values of hole 1D (Fig. 4). The stratigraphic position of this feature relative to the $\delta^{18}O$ data of Bright et al. (2006) suggests that it occurred during MIS 3. Guillou et al. (2004) identified a geomagnetic excursion in MIS 3, the Laschamp, with an $^{40}Ar/^{39}Ar$ age of 40.4 ± 2.0 ka. Although the low-inclination values from hole 1E do not deviate from the GAD value by 40°, as required for excursion classification (Barbetti and McElhinny, 1972), we interpret this feature to be synchronous with the Laschamp excursion of higher-fidelity paleomagnetic records (e.g., Lund et al., 2001; Guillou et al., 2004). The low paleointensity values from the same interval in Bear Lake (Fig. 7B) support our interpretation, because published paleointensity records (Guyodo and Valet, 1996; Channell et al., 1997; 2000; Stoner et al., 1998, 2000, 2002) indicate a paleointensity low ca. 40 ka.

The linear sedimentation rate (0.54 mm yr^{-1}) used by Kaufman et al. (this volume) indicates that the stratigraphic position of this geomagnetic feature corresponds to an age of 50 ka, ~7 k.y. older than the reported age of the Laschamp excursion (40.4 ± 2.0 ka; Fig. 8; Guillou et al., 2004). Alternatively, extrapolating the Colman et al. (this volume) ^{14}C age model through the U-series age at 67 m depth (127.7 ka; Colman et al., 2006) suggests an age for this geomagnetic feature that is somewhat closer to the reported age of the Laschamp excursion (Fig. 8). Although the geochronologic control is insufficient to assign an age to this geomagnetic feature, consideration of its stratigraphic position suggests that this feature is equivalent to the Laschamp excursion of other records.

Mineral Magnetic Zone Ages

In order to assign ages to our mineral magnetic zones (Table 1) we use the linear sedimentation rate age model

(0.54 mm yr^{-1}) used by Kaufman et al. (this volume). This sedimentation rate was constructed from a simple linear fit of four control points: 0 m = 0 ka, 11.3 m = 18.5 ka, 26.5 m = 41 ka, and 67 m = 127.7 ka. Although there are some discrepancies between this model and the ^{14}C-based age model for the uppermost 20 m (Colman et al., this volume), a comparison of our magnetic data (K_{ARM}/K) and the $\delta^{18}O$ data of Bright et al. (2006) (both plotted using the linear sedimentation rate) with the $\delta^{18}O$ stack of Lisiecki and Raymo (2005) shows a relatively good correlation (Fig. 9). Rather than tuning the ages of our record to match the timing of global MIS stratigraphy, however, we use the linear-sedimentation-rate model to interpret our record in the context of regional and global climate without the circularity associated with a climatically tuned age model, as discussed by Kaufman et al. (this volume).

A linear sedimentation rate indicates that mineral magnetic zone I (0–9 m) represents the Holocene and latest Pleistocene (9 m = 17 ka). In zone II (9–18 m), high HIRM values (Fig. 10)

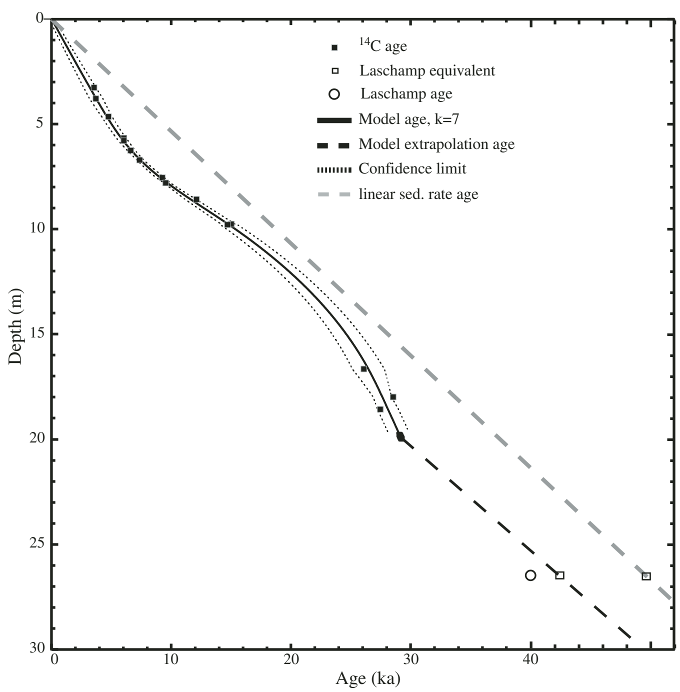

Figure 8. Age models for the uppermost 30 m of Bear Lake. The solid dark line is the ^{14}C-based spline function of Colman et al. (this volume). The "k" value associated with the confidence limit refers to the number of spline functions used and controls the degree of smoothness. The dark dashed line is the extrapolation of the ^{14}C age model through a U-series date at 67 m (127.7 ka from Colman et al., 2006). The light dashed line is the linear-sedimentation-rate model (0.54 mm yr^{-1}) used by Kaufman et al. (this volume). The open squares indicate the inferred age of our Laschamp equivalent feature for each age model. The open circle indicates the age of the Laschamp excursion (40.4 ± 2.0 ka) from Guillou et al. (2004). Considering the stratigraphic position of our Laschamp equivalent feature and the uncertainties associated with the different age models, it is likely that our feature formed around the time of the Laschamp excursion of other records.

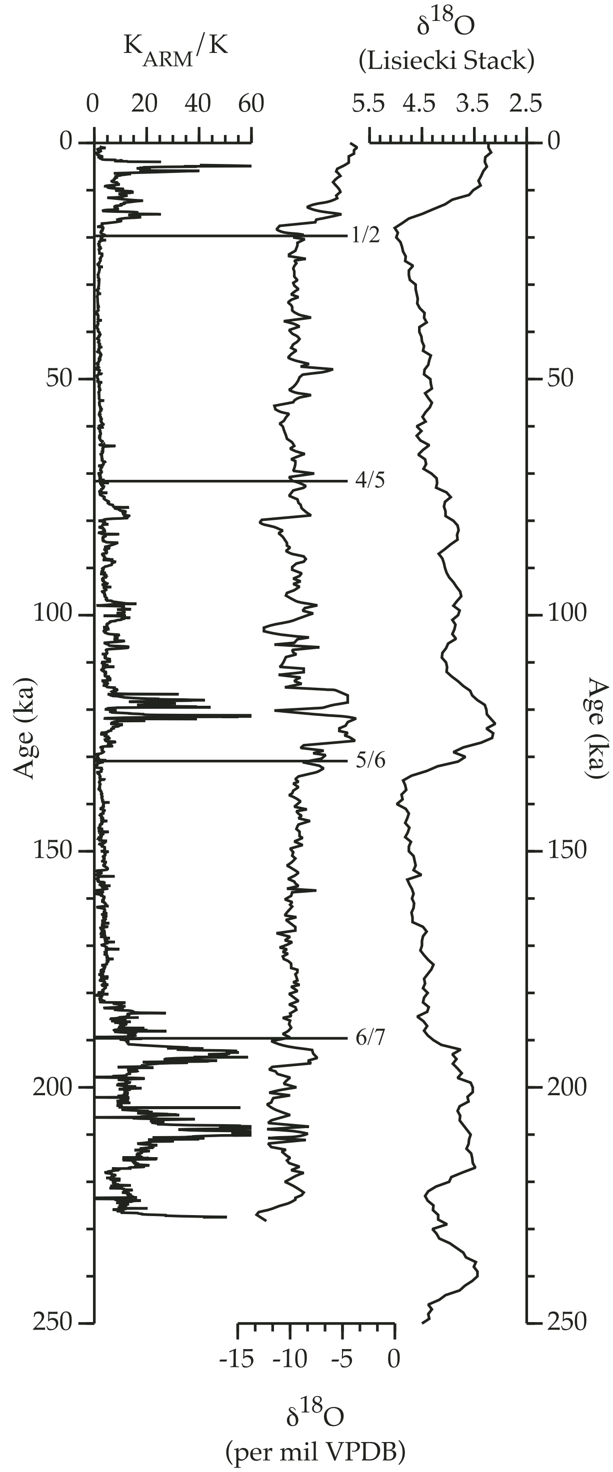

Figure 9. Comparison of K_{ARM}/K and $\delta^{18}O$ (Bright et al., 2006) from the Bear Lake record (using the linear sedimentation rate age model of Kaufman et al. [this volume]) to the $\delta^{18}O$ global stack of Lisiecki and Raymo (2005). The interpreted marine isotope stage (MIS) boundaries for the Bear Lake record (from Bright et al., 2006) are indicated. The comparison shows a relatively good correlation between the Bear Lake record and the global stack and suggests that the linear-sedimentation-rate age model is an appropriate first-order approximation for the age of Bear Lake sediments. VPDB—Vienna Pee Dee Belemnite.

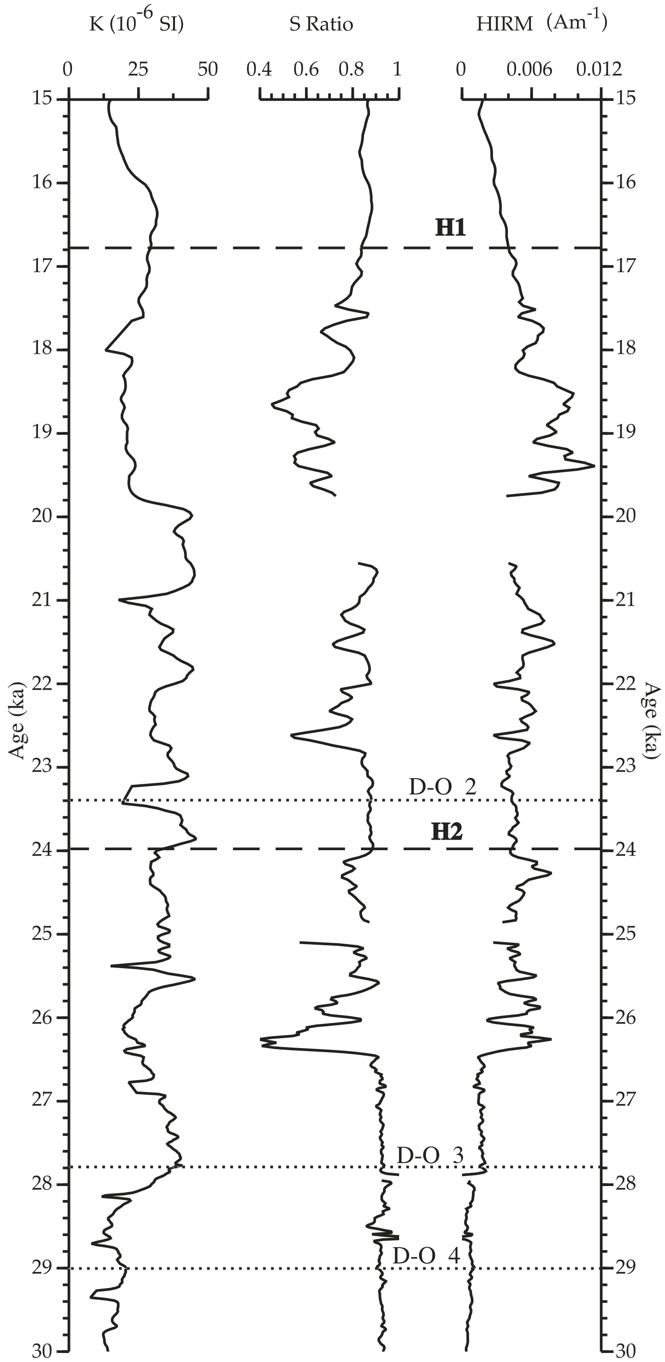

Figure 10. Magnetic properties (K, HIRM, and S ratio) versus age (age model from Colman et al., this volume), showing changes in the abundance of Fe-oxide minerals (Rosenbaum and Heil, this volume). The ages of Heinrich events 1–2 (Hemming, 2004, and references therein) and Dansgaard-Oeschger (D-O) oscillations 2–4 (Rahmstorf, 2003) are indicated with dashed lines and dotted lines, respectively. HIRM—hard isothermal remanent magnetization.

are interpreted by Rosenbaum and Heil (this volume) as glacial flour derived from the Uinta Mountains. The highest values occur around 10.8 m and are assigned an age of 20 ka, which is consistent with an age of 21 ± 3 ka assigned to the Pinedale glacial maximum in the Wind River Range (Pierce, 2004). The upper and lower boundaries for zone III (18–41 m) correspond to 33 ka and 76 ka, respectively. As discussed previously in this paper, the presence of a geomagnetic feature synchronous with the Laschamp excursion (ca. 40 ka) between 25.5 and 27.5 m in this zone suggests that the linear-sedimentation-rate age model does not provide the best fit for this interval (Fig. 8). Zone IV (41–61 m) represents most of MIS 5 (Bright et al., 2006) spanning the time interval of 76–113 ka. MIS 5e (ca. 117–131 ka; Lisiecki and Raymo, 2005) is represented by zone V (61–66 m) and spans the interval 113–122 ka. Zones VI and VII span MIS 6 and most of MIS 7, respectively (Bright et al., 2006). The upper boundary of zone VI (66 m) corresponds to 122 ka and the upper boundary of zone VII (at 97 m) corresponds to 180 ka. The extrapolated basal age of the GLAD800 core from Bear Lake is ca. 222 ka.

Millennial-Scale Variability during the Last Glacial Maximum

Hematite content (as indicated by HIRM and the S parameter) shows well-defined millennial-scale variability between 15 and 30 ka (Fig. 10). Here we use the ^{14}C-based age model of Colman et al. (this volume) because this model is more accurate in this interval than the linear-sedimentation-rate model used for the entire Bear Lake GLAD800 core. These variations during the last glacial maximum (LGM) are interpreted to reflect changes in the content of hematite-rich glacial flour derived from the Uinta Mountains (Rosenbaum and Heil, this volume). Changes in the amount of glacial flour are interpreted to reflect changes in the extent of glacier size in the Uinta Mountains. Using this interpretation, we compared our data from Bear Lake to the timing of North Atlantic Heinrich events 1–2 and Dansgaard-Oeschger (D-O) cycles, well-known millennial-scale climate phenomena observed during the LGM (Fig. 10). A link between alpine glaciation in the western United States and Heinrich events and D-O cycles has been made on the basis of evidence of episodic growth and retreat of alpine glaciers in the Rocky Mountains, the Cascade Range, and the Sierra Nevada (Clark and Bartlein, 1995; Benson et al., 2003). Given existing age constraints, these episodes are in phase with the growth and collapse of the Laurentide ice sheet. In addition, studies from the Great Basin show evidence of Heinrich events and D-O oscillations in the sediment records from several lakes (Benson et al., 1997, 1998, 2003; Benson, 1999; Zic et al., 2002).

North Atlantic Heinrich events are associated with periods of increased iceberg discharge (Heinrich, 1988) initiated by ice sheet instabilities following short-term growth of the Laurentide ice sheet (MacAyeal, 1993). Dansgaard-Oeschger oscillations are the result of either internal oscillations in Earth's climate system (i.e., the formation of North Atlantic Deep Water) or external forcings (i.e., changes in solar irradiance) (Broecker et al., 1990). Although the mechanism is not entirely understood, Benson et al. (2003) showed that the mean position of the polar jet stream (PJS) shifted between 35°N during D-O stadials (cold phase) and ≥43°N during interstadials (warm phase). Thompson et al. (1993) indicated a similar displacement of the PJS during the last glacial maximum. The position of the polar jet stream affects temperature and moisture, with cold, dry air north of the polar jet and cool, moist air south of the polar jet (Thompson et al., 1993). The central location of Bear Lake within the path of the polar jet stream may have made the lake sensitive to changes in the size of the Laurentide ice sheet. We speculate that changes in the position of the PJS, like those associated with D-O oscillations and Heinrich events, affected glaciation in the Uinta Mountains by altering both air temperature and the amount of moisture available for growth of the alpine glaciers. Although uncertainties in the age model of Colman et al. (this volume) are too large to make firm correlations, we note that, if the model is taken at face value, the major peaks in hematite content do not coincide with the timing of any Heinrich events or D-O cycles (Fig. 9). However, it appears that changes do occur on the order of 1–2 k.y., which is similar in duration to D-O cycles (Grootes and Stuiver, 1997). We emphasize that millennial-scale oscillations in the magnetic mineralogy of Bear Lake sediment during the LGM may reflect only regional climate changes within the Great Basin and not hemisphere-wide climatic events.

According to Bright et al. (2006), zone VI corresponds to MIS 6, the penultimate glacial period. Climatic correlations are not suggested for this zone because sediments from this interval do not have magnetite and hematite concentrations like those of zone II. In large part, this condition is related to preservation of Fe-oxide minerals in zone II and destruction of these minerals elsewhere. The high degree of preservation in zone II may reflect the absence of strongly reducing conditions due to fresh water (i.e., low sulfate content) and low productivity, or limited reduction related to rapid sedimentation linked to the influx of glacial flour. Perhaps the Bear River did not flow into the lake during MIS 6, thus preventing glacial flour from entering the lake, or other conditions that would have allowed the preservation of Fe-oxide minerals did not exist.

CONCLUSIONS

The sediments of Bear Lake show a complex record of environmental change. Variations in lake salinity caused by changing freshwater inputs and evaporative conditions associated with glacial and interglacial periods have affected the preservation and deposition of magnetic minerals. Preservation of Fe-oxide minerals during the LGM was likely related to rapid burial of glacial flour input from the Bear River. At other times, changes in the hydrologic balance of Bear Lake near marine isotope stage and substage boundaries led to significant changes in the sulfate concentrations of the lake waters, forming greigite during the fresher, colder stages, and pyrite in the more saline, warmer stages.

Preservation of Fe-oxide minerals in some intervals of the Bear Lake deposits, and varying degrees of post-depositional alteration in others, makes it impossible to establish a continuously reliable paleomagnetic record. Despite these complications, directional variations within an interval of relatively constant magnetic mineralogy (zone III) can be interpreted as a geomagnetic feature equivalent to the Laschamp excursion, thereby providing an important age constraint. Although the sediments of Bear Lake do not generally lend themselves to paleointensity studies because of the complex variations in magnetic minerals, normalized NRM intensities within zone III provide paleointensity information. Lows in NRM/K, NRM/ARM, and NRM/IRM values coincident with the directional variations support the interpretation of the Laschamp equivalent geomagnetic feature.

The sediment record from Bear Lake for the past 220,000 years is complex in both a paleomagnetic and environmental magnetic sense, but some intervals provide important information about climate change in the Great Basin and its relation to both regional and global temporal and spatial scales. Variations in the amount of glacially derived hematite (Fig. 10) delivered to Bear Lake from the Uinta Mountains (Rosenbaum and Heil, this volume) during the LGM occurred on millennial time scales, suggesting a potential link to millennial-scale internal oscillations of the Laurentide ice sheet and the resulting displacement of the polar jet stream. However, uncertainties in the age model prevent us from unequivocally correlating these magnetic features to Heinrich events or Dansgaard-Oeschger oscillations. Although the exact mechanisms of D-O oscillations are not entirely understood, they have a periodicity of ~1500 years (Grootes and Stuiver, 1997), which is strikingly similar to the duration of the hematite peaks in zone II from Bear Lake. At this point these links are speculative and are suggested only as feasible possibilities based on evidence from other lake records and glacial deposits within and around the Great Basin.

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