

A paleoclimate record for the past 250,000 years from Summer Lake, Oregon, USA: I. Chronology and magnetic proxies for lake level

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Abstract

This study presents the age control and environmental magnetism components of a new, late Pleistocene paleoclimate record for the Great Basin of western North America. Two new cores from the Summer Lake sub-basin of pluvial Lake Chewaucan, Oregon, USA are correlated to basin margin outcrops on the basis of tephrochronology, lithostratigraphy, sediment magnetism and paleomagnetic secular variation. Eleven tephra layers were found in the cores that correlate to tephra identified previously in the outcrop. The Olema ash was also found in one of the cores; its stratigraphic position, relative to 3 dated tephra layers, indicates that its age is 50–55 ka, somewhat younger than has been previously reported. The Summer Lake sediments are divided into deep and shallow lake lithosomes based on sedimentary features. The stratigraphic position of these lithosomes support the tephra-based correlations between the outcrop and the cores. These sediments contain a well resolved record of the Mono Lake Excursion (MLE) and an earlier paleomagnetic excursion as well as a high quality replication of the paleosecular variation immediately above the MLE.

Relative sedimentation rates increased dramatically toward the depocenter during intervals of low-lake level. In contrast, during intervals of high-lake level, relative sedimentation rates were comparable along the basin axis from the basin margin to the depocenter. The magnetic mineralogy of the Summer Lake sediments is dominated by pseudo-single domain (titano)magnetite and intervals of high/low magnetite concentration coincide with lithosomes that indicate high/low lake levels. Magnetic grain size also varies in accord with bulk sediment grain size as indicated by the silt/clay ratio. To a first order, variations in magnetic parameters, especially those attributable to the concentration of magnetic minerals, correlate well with global glacial/interglacial oscillations as indicated by marine oxygen isotope stages. This relationship can be explained by increased dissolution of (titano)magnetite minerals as lake level dropped and the lake became more productive biologically. This inference is supported by a correspondence between lower concentrations of magnetite with higher levels of total organic carbon and vice-versa.

Introduction

The Great Basin hydrological province is a valuable source of paleoclimate data due, in part, to the fact that the sizes of internally-drained lakes within the Great Basin are especially sensitive to changes in precipitation and temperature (Benson et al., 1990; Morrison, 1991; Oviatt et al., 1992; Grayson, 1993). Furthermore, because most of the basins have been actively subsiding along marginal faults for much of the late Tertiary, their deposits potentially carry paleoclimate records throughout the Quaternary Period and earlier (e.g., Adam et al., 1989).

Lake sediments from the northwestern part of the Great Basin may also contain tephra derived from the volcanoes of the Cascade Range which lie to the west.

The geochemical signatures of these tephra layers allow excellent intra-basinal correlations between outcrop and core, as well as extra-basinal correlations with other well-dated sequences of lake sediments or volcanic deposits. The tephra beds can also be dated directly using either thermoluminescence (TL) or radiometric methods, thus extending the age control of these sediments beyond the range of radiocarbon dating (e.g., Sarna-Wojcicki & Davis, 1991).

Summer Lake lies within one of four sub-basins of the Pleistocene Lake Chewaucan pluvial lake system in the northwestern Great Basin (Figure 1). The Summer Lake sub-basin was formed as a north-south trending half-graben which is tilted toward the west (Donath, 1962; Walker, 1969; Baldwin, 1981). Accordingly, the axis

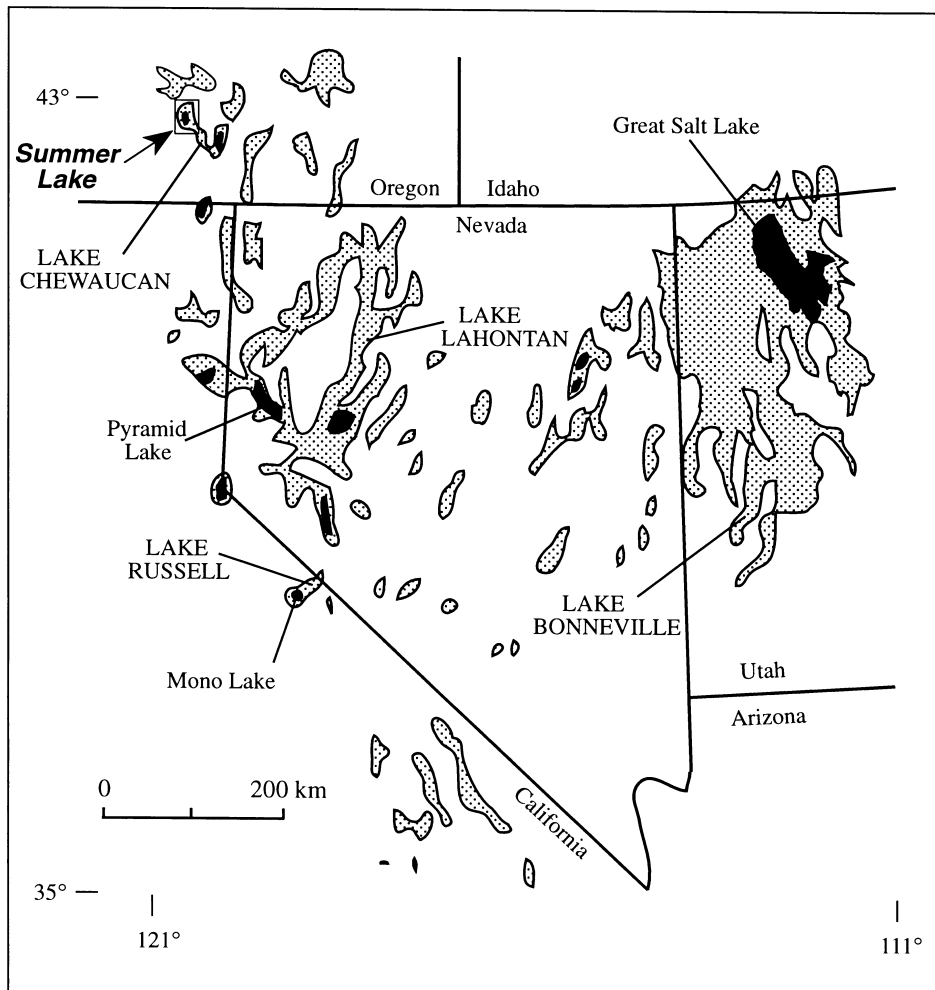


Figure 1. Map of Great Basin lakes with shaded areas indicating Pleistocene highstands (after Spaulding et al., 1983). Summer Lake is in the northwestern corner of the map.

of the basin lies near the western boundary of the basin where the modern remnant lake is presently located (Figure 2). Previous studies of this basin have shown that, in addition to having all of the above-mentioned attributes that are common to pluvial lakes in the northwestern Great Basin (Allison, 1982; Davis, 1985), the deposits of Summer Lake are also useful as repositories of paleoclimate data because of the distinct response of lithology to lake level change (Erbes, 1996), the abundance and diversity of biota (Palacios-Fest et al., 1993; Wigand, et al., 1995; Cohen et al., this volume), and the strong and reliable record of sediment magnetism (e.g., Negrini et al., 1994; Roberts et al., 1994).

In this article we present the age control of a composite sedimentary record including basin-margin outcrops and 2 cores. One core was taken from near the depocenter and the other from a transitional location between the depocenter and the basin margin. We also present environmental magnetic data from these sediments and show that, for this sub-basin, magnetic parameters can be used to indicate relative lake depth.

Age control and correlation

Age control of basin-margin outcrop

Late Pleistocene deposits are exposed at the northern margin of the Summer Lake sub-basin in the canyon of the spring-fed Ana River (Figure 2) near the town of Summer Lake, Oregon. The elevation at the top of these

outcrops is 1,277 m, 15 m above the modern playa surface and 101 m below the highest paleoshoreline feature (Allison, 1982). These deposits contain more than 50 tephra layers, several of which have been dated (Figure 3; Table 1). All age constraints were previously discussed by Negrini et al. (1994), with the exception of new radiocarbon ages for the Wono and Trego Hot Springs (THS) tephra (Benson et al., 1997) and 2) new radiometric and correlation ages associated with the Summer Lake GG tephra layer and the Pringle Falls geomagnetic excursion (Herrero-Bervera et al., 1994; Henyey et al., 1995; McWilliams, 1995).

The new ages for the Wono (27.3 ± 0.3 ^{14}C ka B.P.) and THS (23.2 ± 0.3 ^{14}C ka B.P.) tephra layers were determined by interpolation between 11 precise AMS radiocarbon dates on organic carbon samples and the stratigraphic positions of the 2 tephra layers (Benson et al., 1997). The radiocarbon and tephra samples are from a sediment core that was extracted from Pyramid Lake, Nevada (Figure 1). In part due to the precision added by these new dates, the chronology shown in Figure 3 is remarkably consistent above a depth of 5 m especially considering the variety of methods used to determine age (the ^{14}C and K-Ar radiometric methods, the correlation of paleomagnetic secular variation features, tephrochronology, and TL dating).

In contrast, the chronology below 5 m is characterized by a considerable amount of uncertainty. This uncertainty is exemplified by the range of ages (~ 180 – 228 ka) assigned to the Pringle Falls excursion, a geomagnetic phenomenon that has been documented

Table 1. Age control for Ana River outcrop composite section

Dated interval or horizon	AR depth (m)	Age (ka)	Dating method	Reference
Mono Lake Excursion to top of section	0–2.35	29.5–18.1	paleomagnetic correlation	Negrini & Davis [1992]
Tephra D (Mt St Helens Mp)	1.13	18.6–20.4	radiocarbon	Davis [1985]
Tephra 18 (Trego Hot Springs Tephra)	1.47	23.5 ± 2.5	thermoluminescence	Berger [1991]
Trego Hot Springs Tephra	1.47	23.2 ± 0.3	radiocarbon	Benson et al. [1997]
Tephra F (Wono Tephra)	1.90	27.3 ± 0.3	radiocarbon	Benson et al. [1997]
Tephra 12 (Mt St Helens Cy)	2.61	47.0 ± 2.0	thermoluminescence	Berger [1991]
Tephra 6 (Pumice Castle)	3.62	72 ± 6	K-Ar	Davis [1985] & Bacon[1983]
Tephra 2	4.02	67.3 ± 7.2	thermoluminescence	Berger [1991]
Tephra N	7.00	102.3 ± 11	thermoluminescence	Berger [1991]
Tephra R	7.56	165 ± 19	thermoluminescence	Berger [1991]
Tephra GG	11.76	190 ± 10	global correlation of paleomagnetic excursion	this paper
Tephra GG	11.76	218 ± 10	Ar-Ar	Herrero-Bervera et al. [1994]
Tephra KK	13.46	171 ± 43	Ar -Ar	Herrero-Bervera et al. [1994]
Tephra KK	13.46	200 ± 27	thermoluminescence	Berger [1991]
Tephra LL	13.92	160 ± 35	thermoluminescence	Berger [1991]

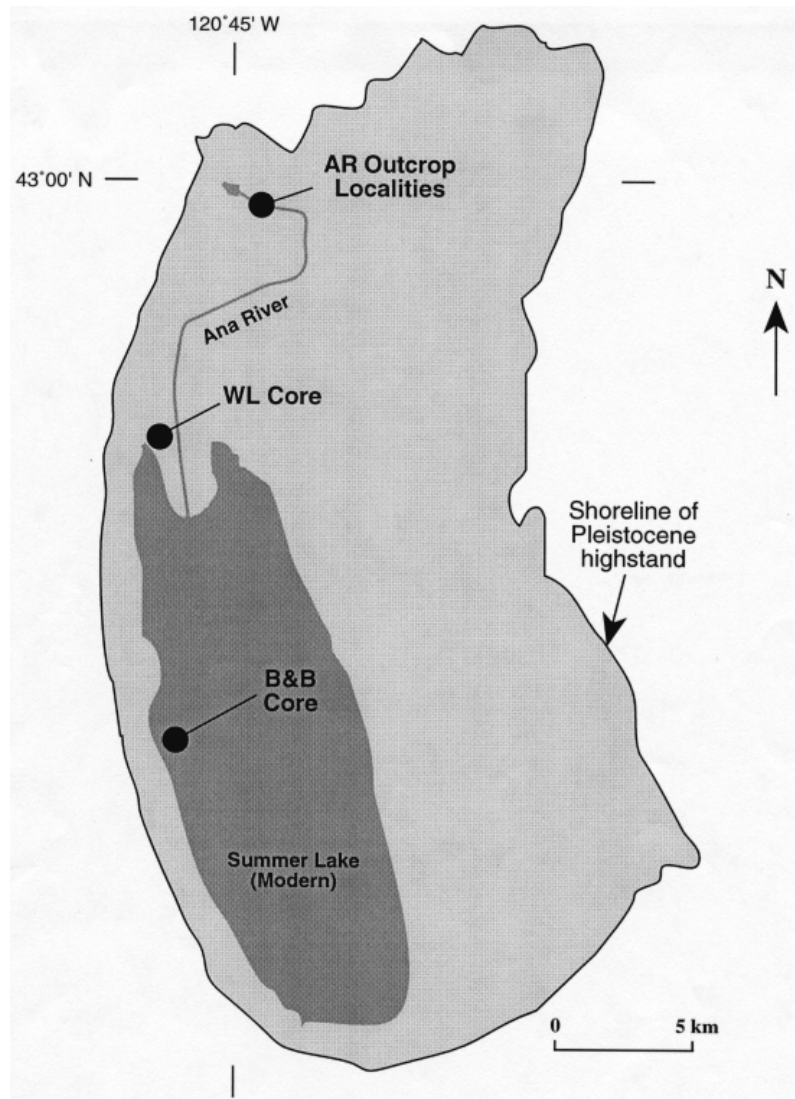


Figure 2. Map of Summer Lake sub-basin showing locations of outcrop and core.

in several sedimentary records including two records containing tephra layer GG at the Ana River locality (Negrini et al., 1994) and 2 records that contain a correlative tephra layer D at the Pringle Falls, Oregon locality which lies 170 km to the northwest (Herrero-Bervera et al., 1994). McWilliams (1995) has suggested that the true age of the excursion is toward the older part of this range, based on the similarity of its transitional virtual geomagnetic pole (VGP) path with that of an excursion found in volcanic rocks from New Zealand that have been dated at 223 ± 4 ka (Shane et al., 1994) and on a radiometric age on tephra layer D (Herrero-Bervera et al., 1994). In contrast, a proposed correlation of the Pringle Falls event with an excursion

found in a marine sediment record from the north-western Atlantic Ocean (Heney et al., 1995) suggests that the Pringle Falls excursion correlates with a 190 ± 10 ka geomagnetic excursion/paleointensity low found throughout the world in marine sediments (e.g., Yamazaki & Ioka, 1994; Heney et al., 1995; Weeks et al., 1995; Lehman et al., 1996; Roberts et al., 1997; Channell, in press). Because the VGP path of the Pringle Falls excursion is similar to several different excursions and polarity transitions (e.g., Clement, 1991; Laj et al., 1991; Tric et al., 1991), the correlation between the Pringle Falls and New Zealand excursion is unconvincing. We favor the correlation of Heney et al. (1995) because, in contrast, it is based on the correlation of several

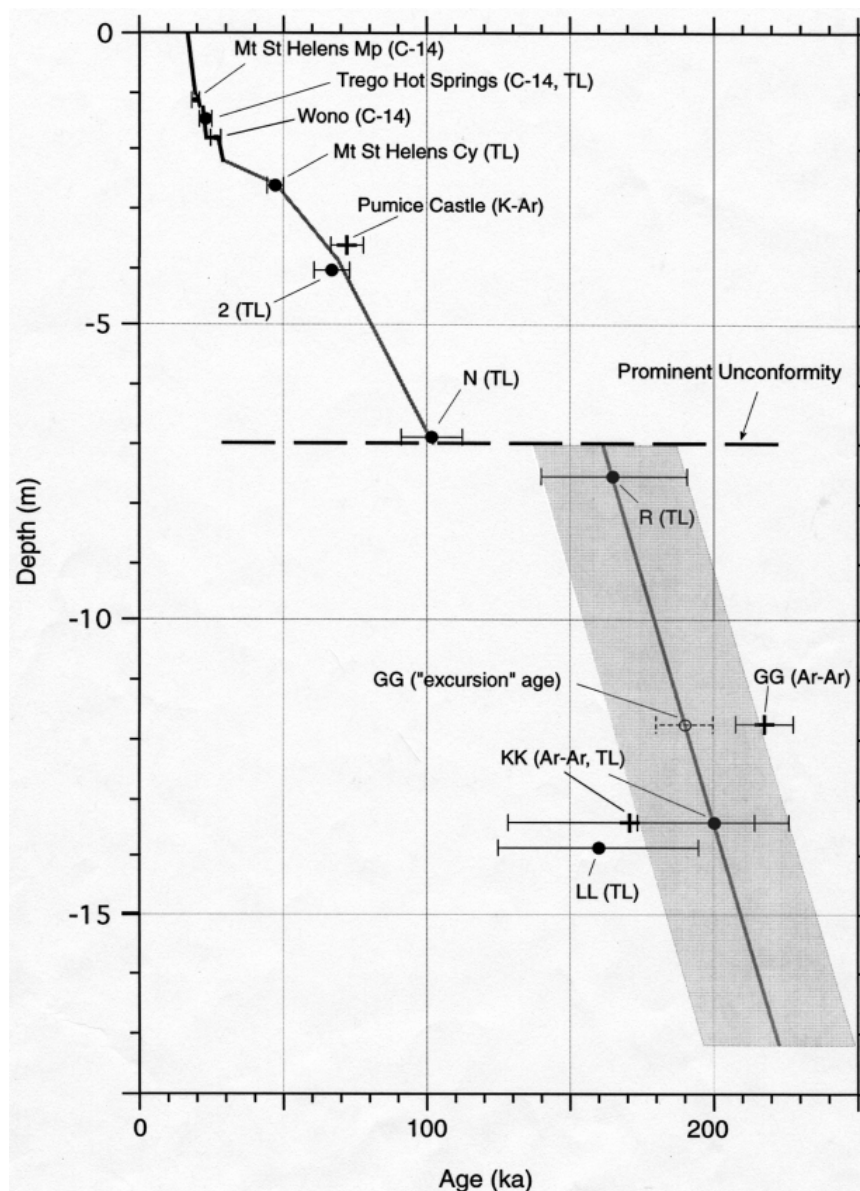


Figure 3. Depth vs. age for Ana River outcrop sediments (see Table 1 and discussion in text). Solid circles represent thermoluminescence dates. Crosses represent radiometric dates. Solid black line represents age control from correlation of paleomagnetic secular variation features. Heavy dashed line represents a major unconformity associated with marine oxygen isotope stage 6/5 transition (Davis, 1985; Berger, 1991). Lack of deposition at this basin-margin outcrop during extremely low lake levels and/or erosion of the underlying sediments deposited during the Stage 6 high lake interval are probably responsible for the missing stratigraphic interval corresponding to the time interval surrounding the unconformity.

distinct paleomagnetic directional and intensity features found in both the excursion and in the subsequent secular variation (Negrini et al., 1994; Henyey et al., 1995; Negrini, in press).

Based on the above discussion, we assign an 'excursion' age of 190 ± 10 ka for tephra layer GG. This constrains our preferred chronology represented by the

thick shaded line in Figure 3. Our preferred chronology for the older part of the sequence is constrained at the upper end by a TL date on tephra layer R. This date provides the youngest age constraint below a prominent unconformity, which is defined sedimentologically by a lag deposit of carbonate-coated ostracodes (Figure 3). The unconformity corresponds to a major

lake-level regression that has been associated with marine oxygen isotope stage 5e (Davis, 1985; Berger, 1991; Erbes, 1996). We have also enclosed our preferred chronology in a shaded field that corresponds to the limits of the error associated with the TL dates on tephra layers KK and R. This field intersects acceptable ages associated with every age constraint shown in Figure 3 within the bounds of their experimental precision. Therefore, the actual chronology should lie somewhere within this field.

The preferred chronology given in Figure 3 was used to assign ages to the Ana River outcrop sediments in this paper and in the companion paper that follows (Cohen et al., 2000). This chronology also provides the framework for the chronologies of the 2 cores described below. That is, age-depth relationships will be developed for each core after they are correlated to the outcrop by a variety of methods, including tephrochronology, lithostratigraphy, and the correlation of paleomagnetic and sediment magnetic features. Any interpretation based on this preferred chronology is, of course, subject to the imprecision reflected in the error bars and scatter of data shown in Figure 3.

Correlation of outcrop with cores

In the summer of 1992 sediment cores were collected from two localities in the Summer Lake sub-basin (Figure 2). The sites were chosen so that the cores, in conjunction with the Ana River outcrops, would represent a transect along the N-S axis of the basin from one margin to the depocenter. The 'Wetlands Levee' (WL) core was taken near the northernmost margin of the modern playa at an elevation of 1265 m (latitude 42° 54' 40.2" N, longitude 120° 42' 39.6" W). The 'Bed and Breakfast' (B&B) core was taken from near the depocenter of the sub-basin in the modern Summer Lake playa at an elevation of 1264 m (latitude 42° 48' 27.0" longitude 120° 46' 56.4").

The WL site was accessed via a gravel road on top of a dike. A relatively long core was obtained with a truck-mounted, Central Mining Equipment (CME) 75 drill rig using a 5-foot-long (1.52 m), 3-inch (7.62 cm) internal diameter, core-barrel sampler within 8 inch (20.3 cm) hollow stem augers. The entire length of the WL core is 30.5 m and the recovery was 90–95%. The core was encased in cellophane and PVC pipe and was subsequently transported to a cold storage room at California State University, Bakersfield.

Because the B&B site was accessible only by foot, we were only able to collect a relatively short, 11.9 m

core (100% recovery) using a modified Livingston piston coring device. Drive segments were collected and preserved in 4-inch (10.2 cm) diameter PVC casing and were stored cold at the Desert Research Institute in Reno, Nevada.

Correlation methods

Tephrochronology

The geochemical signatures of volcanic glass from tephra layers sampled from core and outcrop were analyzed using the electron microprobe in the Geo-analytical Laboratory at Washington State University, Pullman. The experimental details are given in Foit et al. (1993). These analyses and their statistical comparison to tephra from the outcrop sediments and from other western North American localities are summarized in Table 2. The similarity coefficients were calculated using unit weighting (1.0) of the oxide concentrations of Si, Al, Ca, Fe, Na, and K and 0.25 weighting of the oxide concentrations of Ti and Mg. The latter 2 oxides were given a lower weighting because of their low concentrations and consequently high relative error of measurement. The stratigraphic positions of these tephra layers are shown in Figure 4a along with their correlatives from the outcrop and other dated tephra layers within this sequence.

Lithostratigraphy

Based on lithological features, the sediments of the Summer Lake sub-basin can be divided into intervals which correspond to deposition under relatively uniform physical and chemical conditions (i.e., lithosomes). Shallow lake lithosomes are characterized by the preponderance of well-sorted lenses of coarse silt and sand as well as carbonate breccias (Erbes, 1996). In contrast, deep lake lithosomes seldom contain such features and, instead, are characterized by a dominance of thinly-bedded to massive clay and muddy silt. More detailed descriptions of the methods used in the analysis and interpretation of the lithological features are given by Erbes (1996) and in the companion paper by Cohen et al. (2000). The stratigraphic positions of lithosome intervals associated with shallow or deep lake levels are plotted in Figure 4a.

Sediment magnetism

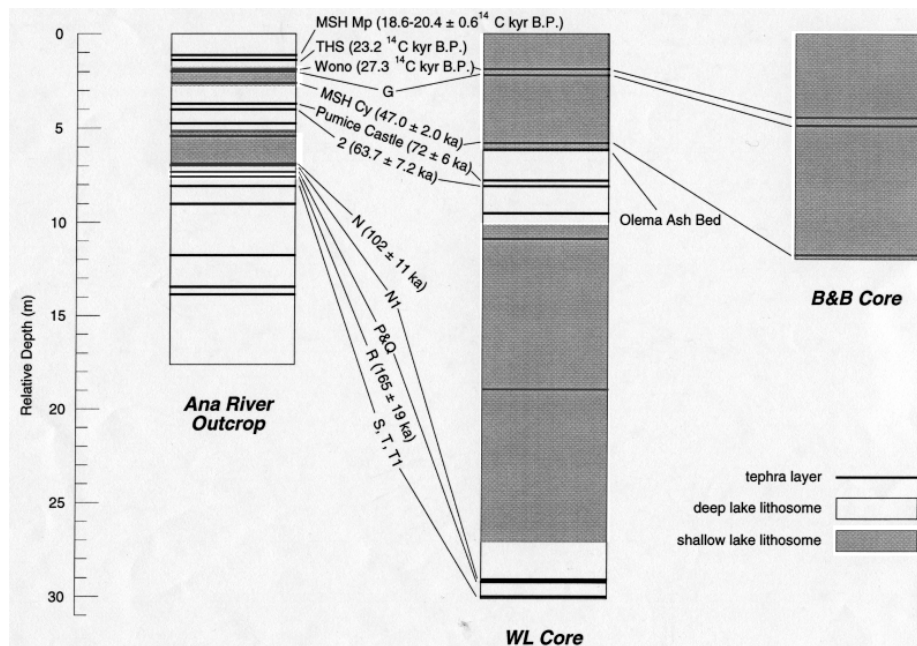
Mineral magnetic and paleomagnetic records are commonly used to assist in correlating sedimentary records (e.g., Thompson & Oldfield, 1986; Peck et al., 1994; Verosub & Roberts, 1995). Three such measure-

Table 2. Compositions and similarity coefficients and correlations of glasses in the Ana River outcrop, Wetland Levee, and Bread and Breakfast tephra

Location & Depth ¹	Sample	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MgO	MnO	CaO	BaO	TiO ₂	Na ₂ O	K ₂ O	Cl	SC ⁸
Ana River 1.13m	Tephra D	76.3	13.9	1.23	0.30	0.06	1.51	0.06	0.16	4.1	2.4	0.11	0.97
	MSHM Std ²	76.45(2.50)	13.81(20)	1.39(20)	0.32(4)	n.d.	1.48(8)	n.d.	0.16(3)	3.96(30)	2.36(10)	0.10(2)	
Ana River 1.47m	Tephra 18	75.2(2)	13.7(2)	1.65(1)	0.20(1)	0.04(2)	1.00(3)	0.10(1)	0.25(3)	4.6(1)	3.2(1)	0.12(1)	0.99
	Trego H.S. Std ³	75.12	13.94	1.67	0.20	0.05	1.05	n.d.	0.25	4.51	3.21	n.d.	
Ana River 1.90m Wetlands Levee 1.84m Bed & Breakfast ⁴ .42m	Tephra F	73.7	14.1	2.23	0.28	0.02	1.30	0.10	0.31	4.7	3.1	0.10	0.96
	WL-2-2	74.82(24)	13.87(12)	1.95(8)	0.25(4)	n.d.	1.28(6)	n.d.	0.19(3)	4.45(10)	3.10(8)	0.09(2)	0.97
	Tephra IB	74.86(24)	13.87(9)	1.95(11)	0.25(4)	n.d.	1.29(5)	n.d.	0.20(3)	4.38(12)	3.11(10)	0.09(3)	0.97
	Wono Std ⁴	74.57(83)	13.77(26)	2.00(30)	0.29(5)	n.d.	1.33(16)	n.d.	0.27(7)	4.54(17)	3.12(12)	0.11(2)	
Ana River 1.95m Wetlands Levee 2.2m Bed & Breakfast	Tephra G	77.0	13.3	1.03	0.10	0.04	0.75	0.10	0.10	4.2	3.4	0.09	0.95
	WL-2-4	77.39(17)	12.91(10)	1.16(13)	0.12(2)	n.d.	0.81(4)	n.d.	0.09(6)	4.07(8)	3.35(8)	0.10(4)	0.99
	Tephra II	77.51(19)	12.86(11)	1.12(11)	0.12(2)	n.d.	0.80(7)	n.d.	0.09(5)	4.00(7)	3.42(11)	0.08(3)	
Ana River 2.61m Wetlands Levee 5.7m Bed & Breakfast	Tephra 12, glass 1	76.5	14.0	0.99	0.24	0.04	1.50	0.07	0.10	4.1	2.4	0.07	0.97
	WL-7-1	76.54(31)	13.71(13)	1.15(16)	0.29(4)	n.d.	1.60(7)	n.d.	0.12(6)	4.04(9)	2.48(11)	0.07(2)	0.96
	Tephra V	76.58(18)	13.77(9)	1.12(8)	0.26(2)	n.d.	1.63(7)	n.d.	0.10(3)	4.03(9)	2.41(7)	0.10(1)	0.97
	MSH Cy Std ⁵	76.68(16)	13.93(20)	1.05(9)	0.27(3)	n.d.	1.55(5)	n.d.	0.11(5)	3.90(12)	2.43(11)	0.07(4)	
	WL-7-2	75.24(29)	13.25(12)	1.98(12)	0.11(3)	n.d.	0.50(6)	n.d.	0.17(3)	4.53(23)	4.06(10)	0.16(4)	0.98
Olema ⁷	75.01	13.63	1.76	0.11	0.06	0.50	0.50	n.d.	0.17	4.48	4.05	n.d.	
Ana River 3.62m Wetlands Levee 7.66m	Tephra 6	72.2	14.7	2.47	0.49	0.03	1.77	0.10	0.49	4.7	3.0	0.14	0.96
	WL-9-2	72.04(39)	14.43(14)	2.81(13)	0.49(4)	n.d.	1.83(12)	n.d.	0.45(13)	4.65(12)	3.15(8)	0.15(2)	0.96
Pumice Castle ⁶	72.64(60)	14.39(10)	2.41(10)	0.53(3)	n.d.	1.85(6)	n.d.	0.46(4)	4.23(20)	3.12(10)	0.24(22)		
Ana River 4.02m Wetlands Levee 8.06m	Tephra 2	72.7	14.5	2.43	0.26	0.04	1.07	0.10	0.33	5.0	3.4	0.08	0.97
	WL-9-4	72.73(23)	14.36(11)	2.71(14)	0.26(5)	n.d.	1.10(7)	n.d.	0.32(4)	4.88(9)	3.54(11)	0.10(2)	
Ana River 7.05m Wetlands Levee 8.06m Wetlands Levee 8.06m	Tephra NI	69.7	15.0	3.59	0.51	0.09	1.65	0.09	0.49	5.5	3.0	0.12	0.97
	WL-37-I	70.54(27)	14.86(11)	3.87(10)	0.53(3)	n.d.	1.72(6)	n.d.	0.48(3)	4.92(22)	2.98(4)	0.10(2)	0.99
	WL-37-II	70.40(21)	14.85(7)	3.89(14)	0.53(4)	n.d.	1.73(7)	n.d.	0.48(6)	5.03(17)	2.99(8)	0.10(2)	
Ana River 7.31m Wetlands Levee 29.16m	Tephra P	71.6	14.4	3.12	0.33	0.07	1.15	0.09	0.34	5.3	3.5	0.11	0.96
	WL-37-III	72.14(17)	14.39(12)	3.26(14)	0.31(4)	n.d.	1.23(4)	n.d.	0.32(7)	4.90(19)	3.34(7)	0.11(2)	
Ana River 7.36m Wetlands Levee 29.19m	Tephra Q	71.8	14.3	3.10	0.32	0.07	1.15	0.07	0.33	5.2	3.5	0.11	0.97
	WL-37-IV	72.31(24)	14.34(12)	3.22(14)	0.32(3)	n.d.	1.19(4)	n.d.	0.31(4)	4.78(19)	3.42(7)	0.11(1)	
Ana River 8.01m Wetland Levee 30.02m	Tephra T	74.7	13.5	1.86	0.12	0.06	0.52	0.09	0.18	4.9	4.1	0.13	0.97
	WL-38-3	74.97(27)	13.57(26)	1.88(19)	0.10(7)	n.d.	0.56(16)	n.d.	0.17(2)	4.62(17)	4.02(16)	0.12(3)	

1. Compositions of the glasses in Ana River tephra from Davis (1985). 2. Mount St Helens Mm Standard glass compositional data normalized to 100% from Busacca et al., (1992). 3. Trego Hot Springs Standard G551A from Benson et al., 1997. 4. Re-analysis by the author (FFF) of the medium sand fraction of Wono tephra sample DR-29 (Davis, 1985). 5. Mount St Helens Cy standard collected by D. Mullineux. 6. Mount Mazama Pumice Castle Standard MA-79C138 (Busacca et al., 1992). 7. Olema or Ash Bed 1 (AH-4(3)), Clear lake, CA (Sarna-Wojcicki et al., 1992). 8. The similarity coefficient (Borchardt et al., 1972) corresponds to the match between the given glass composition and the composition immediately following it in the table.

a



b

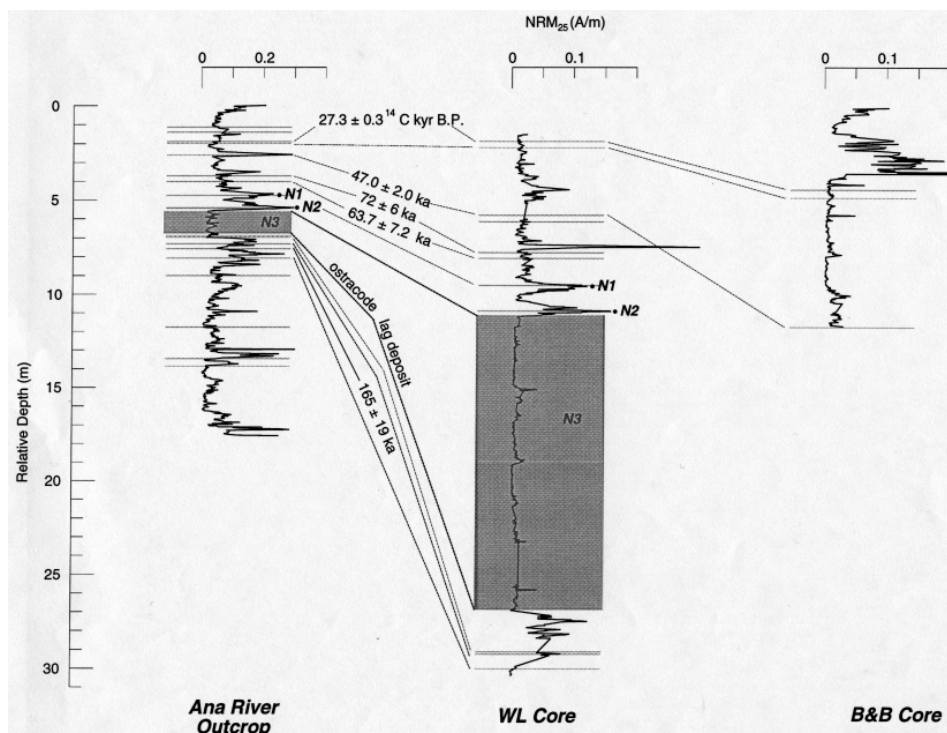
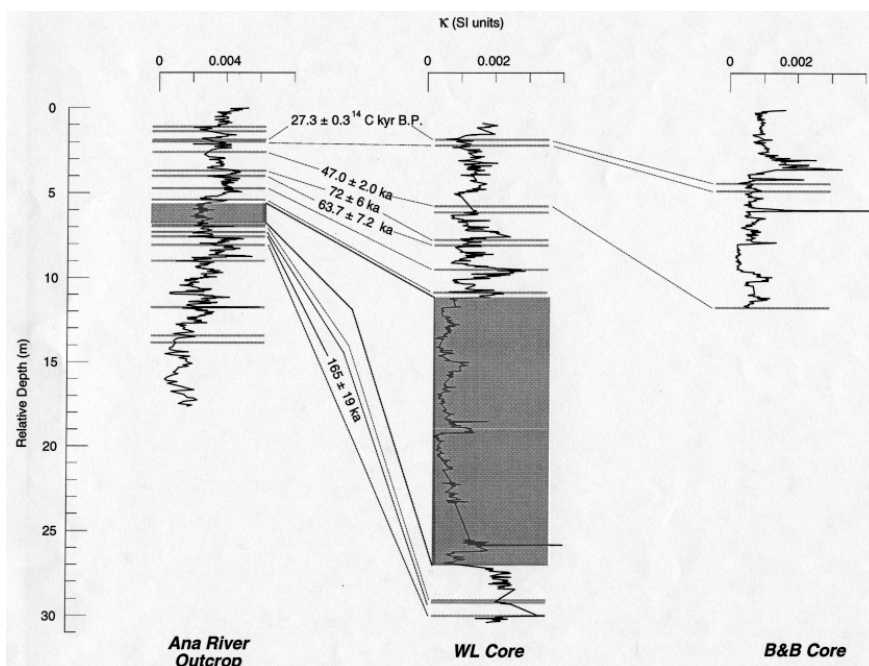
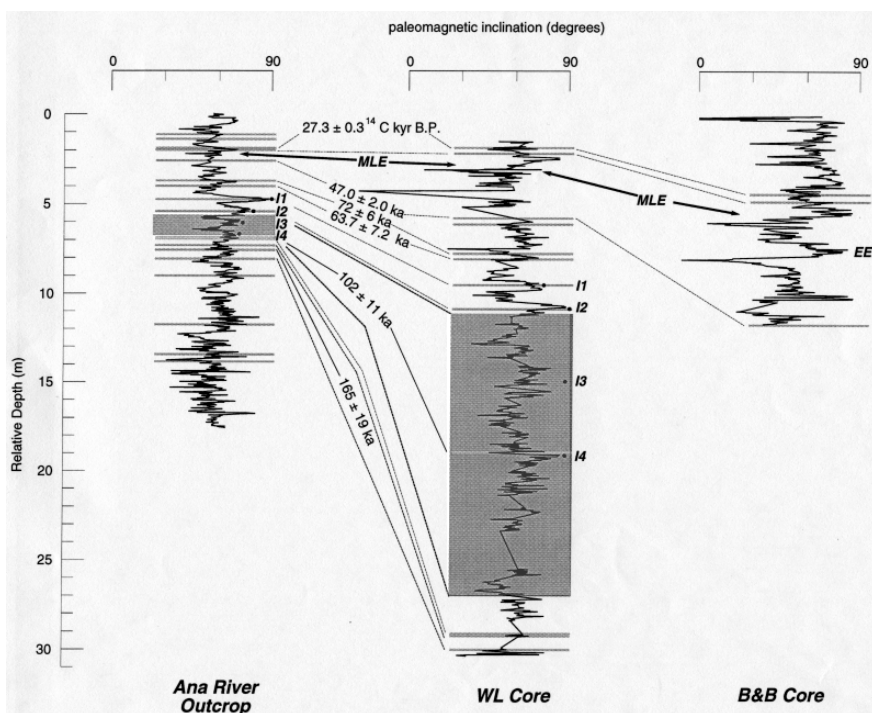


Figure 4. (a) Stratigraphic positions of tephra found in the cores and outcrop. The ages and geochemical analyses of the tephra are summarized in Tables 1 and 2, respectively; (b) Natural remanent magnetization after alternating field demagnetization at 25 mT (NRM_{25}) vs. depth for the outcrop and cores. Tephra-based correlations are supported by the correlation of two sharp peaks (N1 and N2) and a pronounced, broad interval of consistently low NRM_{25} (N3). The broad interval defining N3 is shaded. Tephra layers are plotted as horizontal, lightly-shaded lines. Correlations between the tephra layers are indicated by the light dashed lines; (c) Volume magnetic susceptibility (k) vs. depth for the outcrop and cores. A broad region of anomalously-low susceptibilities occupies the same general area of low NRM_{25} intensities defined as N3 in Figure 4b. The shaded region corresponds to the position of N3; (d)

c



d



Paleomagnetic inclination vs. depth for the outcrop and cores. The distinctive signature of the Mono Lake Excursion (MLE) as indicated by anomalously-low inclinations followed immediately by anomalously-high inclinations, is observed in all three records. An earlier excursion (EE) is observed in the record from the B&B core. Four distinctive peaks in inclination (I1–I4) are correlated between the Ana River outcrop record and the WL core. The broad nature of the I3 and I4 peaks in the WL core record relative to the outcrop record is consistent with the higher sedimentation rate in that region of the WL record, as suggested by the tephra correlations.

ments were used in this paper to correlate the outcrop with the two cores: (1) the intensity of natural remanent magnetization (NRM); (2) low-field, volume susceptibility (κ); and (3) paleomagnetic inclination. The NRM intensity of sediments is dependent principally on the concentration of magnetic minerals as well as the intensity of the Earth's magnetic field at the time the magnetization was acquired (Tauxe, 1993). Inclination is dominantly influenced by the Earth's magnetic field direction at the time the magnetization was acquired, although it can be distorted by sedimentary processes such as differential compaction and it can be reset to produce a post-depositional magnetization (Verosub, 1977; Anson & Kodama, 1987; Celaya & Clement, 1988; Arason & Levi, 1990a, 1990b; Deamer & Kodama, 1990; Levi & Banerjee, 1990; Tauxe, 1993). In sediments dominated by 1 magnetic mineral type like those of the Summer Lake sub-basin (see below and Roberts et al., 1994), the κ of fine-grained sediments is primarily a function of the concentration of magnetic minerals (Thompson & Oldfield, 1986; Verosub & Roberts, 1995; Heider et al., 1996).

Samples for all magnetic analyses were taken from outcrop according to the procedure outlined in Negrini et al. (1994). Samples were taken from core after each core segment was split parallel to its vertical axis with a

fine wire. The split was offset slightly so that the samples could be taken from as close to the axis of the core as possible in order to minimize the effects of sediment deformation at the core margins. In almost all cases, 2 samples were taken per horizon and averaged values are reported associated with each sample pair.

NRM intensity and inclination were measured in a Molspin spinner magnetometer after alternating field (AF) demagnetization in a Molspin alternating field demagnetizer at 25 mT in order to remove a viscous remanent magnetization. Because the demagnetized remanence is no longer strictly a 'natural' magnetization, this parameter will henceforth be referred to as NRM_{25} . Susceptibility was measured with a Bartington Instruments Model MS1 susceptibility bridge. The three measurements are plotted versus depth for the outcrop and two cores in Figure 4, b–d.

The results of previous studies have shown that, after the removal of weak viscous remanence, the magnetization of the Lake Chewaucan outcrop sediments is carried by pseudo-single domain magnetite and reliably records the Earth's magnetic field at or near the time of sediment deposition (Negrini et al., 1984; 1988; 1994; Roberts et al., 1994). The sediments from the B&B and WL cores have a similar 2-component magnetization as shown by the typical demagnetization

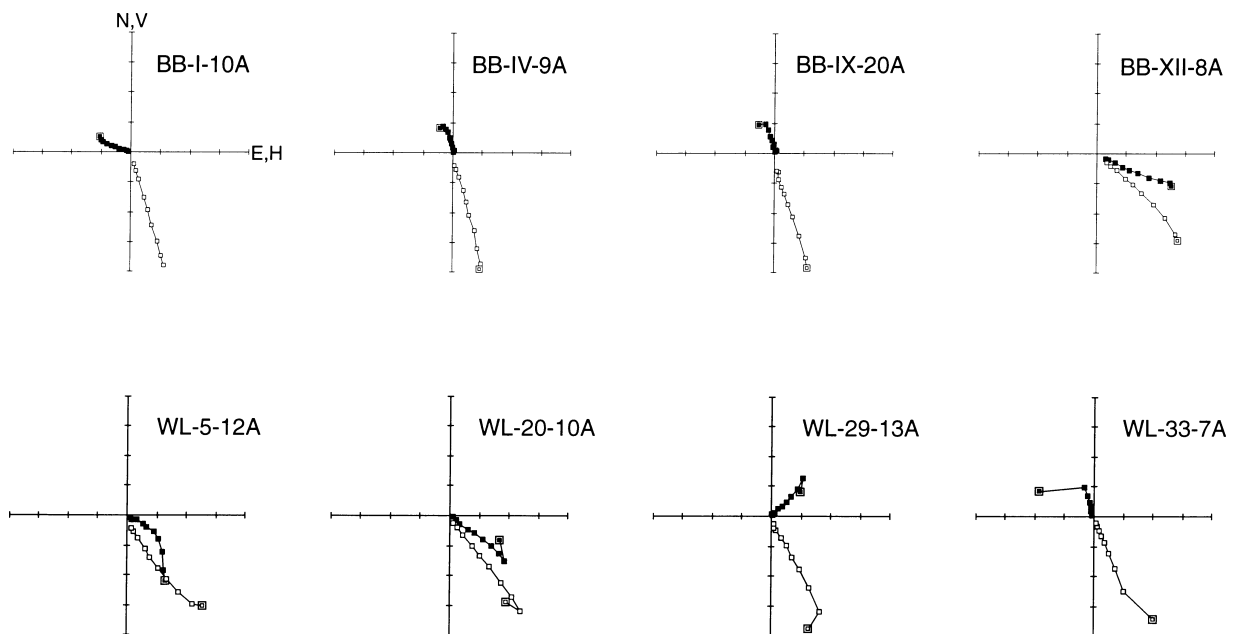


Figure 5. Vector component plots for representative samples from the WL and B&B cores. The filled (unfilled) squares depict the horizontal (vertical) components of the paleomagnetic vector after alternating field demagnetization levels at 0, 5, 10, 15, 20, 25, 30, 40, 50, and 60 mT.

behavior of pilot samples (Figure 5); the magnetizations exhibit a stable, uni-vectorial direction component after AF demagnetization at 10–20 mT. The median destructive fields of the samples from both cores range from 10–25 mT. The average NRM_{25} intensities are 19.9 and 30.9 mA/m for the WL and B&B samples, respectively. These NRM_{25} intensities are a factor of 2–3 lower than those of the outcrop samples (54.8 mA/m), as are the average magnetic susceptibility values (Figure 4b). Upon dessication, samples from the cores and outcrop typically decreased in volume by factors of 3 and 0–1.5, respectively. Thus, we attribute the relatively lower values of κ and NRM_{25} in the core samples to elevated water content in the core sediments relative to those from the outcrop.

The average inclination of the WL core sediments ($59.5^\circ \pm 10.0^\circ$) is statistically indistinguishable from the expected inclination for a geocentric axial dipole at this locality (61.7°). This observation supports the supposition that the paleomagnetic directions in the WL core provide a reliable record of the geomagnetic field at or near the time of deposition and are thus suitable for use in correlation with the outcrop. Further support for this supposition and for the

reliability of the inclination record from the B&B and WL cores is provided by comparison of these 2 records with coeval records of paleomagnetic inclination from western North America (Figure 6). The correspondence between the record from the B&B core and the Pluvial Lake Russell record (Lund et al., 1988) is particularly impressive. Both the Mono Lake Geomagnetic Excursion, which is a distinct, geomagnetic chronostratigraphic marker horizon in western North America (Conway et al., 1994; Negrini & Davis, 1992), and a subsequent paleosecular variation waveform (Lund et al., 1988) are reproduced in detail.

Correlation results

The age of the core sediments was tied into the outcrop chronology by correlating tephra and several distinct features in the lithologic and magnetic records. Additional constraints were placed on the age of the cores, especially the B&B core, via direct correlation of paleomagnetic secular variation (PSV) features to PSV records from elsewhere in western North America. These correlations are summarized below and the relevant data are presented in Figures 4 and 6 and in Tables 2–4.

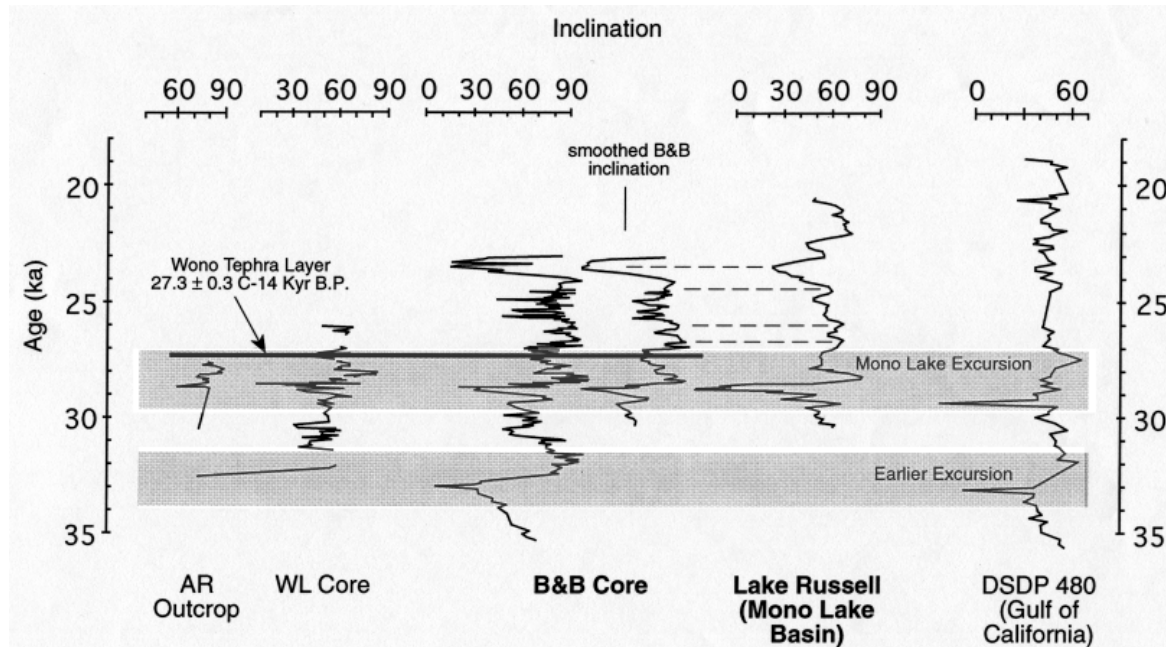


Figure 6. High-resolution correlation of paleomagnetic inclination features between the B&B record from the Lake Chewaucan sediments and the Wilson Creek Record from the sediments of Pluvial Lake Russell. A portion of the B&B record was replotted after smoothing with a seven-point Gaussian algorithm. Detailed correlations between the smoothed version of the B&B record and the Wilson Creek record are indicated by dashed lines. Also shown are the record from the sediments of DSDP 480 (Gulf of California) and correlation of a distinctive excursion that precedes the Mono Lake Excursion at ~34–32 ka.

Table 3. Age control for B&B core

Dated interval or horizon	B&B depth (m)	Age (ka)	Dating method	Reference
Inclination low above Mono Lake excursion (MLE)	0.25	23.5	age of this feature in Wilson Creek record corrected for new age of Wono Tephra	this paper
Inclination high above (MLE)	0.56	24.5	ditto	Davis [1985]
Tephra I (Wono Tephra)	4.42	27.3 ± 0.3	radiocarbon	Benson et al. [1997]
Inclination high within MLE	5.38	28.4	age of this feature in Wilson Creek Record corrected for new age of WonoTephra	this paper
Inclination low within MLE	6.10	28.8	ditto	this paper
Inclination low within earlier excursion	8.12	33.0	age of this feature in DSDP 480 core	Levi & Karlin [1989]
Tephra V (Mt St Helens Cy)	11.81	47.0 ± 2.0	thermoluminescence	Berger [1991]

Table 4. Age control for WL core

Dated interval or horizon	WL Depth (m)	Age (ka)	Dating method	Reference
Tephra WL-2-2 (Wono)	1.84	27.3 ± 0.3	radiocarbon	Benson et al. [1997]
Tephra WL-2-4 (Tephra G)	2.20	27.8	age of Tephra G in B&B core (Tephra II)	this paper
Inclination high within MLE	2.47	28.4	age of this feature in Wilson Creek record corrected for new age of Wono Tephra	this paper
Inclination low within MLE	3.11	28.8	ditto	this paper
Inclination low within earlier excursion	4.27	33.0	age of this feature in DSDP 480 core	Levi & Karlin [1989]
Tephra WL-7-1 (Mt St Helens Cy)	5.70	47.0 ± 2.0	thermoluminescence	Berger [1991]
Tephra WL-9-2 (Pumice Castle Tephra)	7.66	72 ± 6	K-Ar	Davis [1985] & Bacon [1983]
Tephra WL-9-4 (Tephra 2)	8.06	67.3 ± 7.2	thermoluminescence	Berger [1991]
14 inclination feature associated with Tephra N in Ana River outcrop	19.10	102.3 ± 11	thermoluminescence	Berger [1991]
ostracod lag deposit/bottom of N3 NRM feature	27.10	118.2 above unconformity 151.0 below unconformity	extrapolation from nearest dated horizon to the unconformity based on assumption of constant sedimentation rate	this paper
Tephra WL-38-1 (Tephra R)	29.57	165 ± 19	thermoluminescence	Berger [1991]
Tephra WL-38-3 (Tephra T)	29.98	167.9	age of 'T' in outcrop	this paper

Tephrochronology

Eleven tephra from the WL core were correlated with tephra from the Ana River outcrop based on their geochemical signature. Using the outcrop tephra terminology defined by Davis (1985), tephra T, Q, P, N1, 2, the Pumice Castle set of 3 ashes (4, 6, and 8), tephra layer 12 (the Mt. St. Helens Cy ash), tephra G and F (the

Wono tephra layer) were all identified in the WL core (Figure 4a and Table 2). Three tephra layers in the B&B core correlate with the Mt. St. Helens Cy ash, tephra layer G, and the Wono tephra layer as well as with their correlatives in the WL core (Figure 4a and Table 2). In all cases, the similarity coefficients (Borchardt et al., 1972) associated with these matches were 0.95 or greater.

Furthermore, each of the above tephra correlations was the best match found among all other tephra in the Summer Lake sequence as well as among the > 1,000 tephra analyses from Western North America that are included in the database of the WSU Geoanalytical Laboratory.

A tephra layer from the WL core was correlated with the Olema tephra bed. This ash layer was previously identified in two lacustrine cores, one taken from Clear Lake near the town of the same name in northern California (Sarna-Wojcicki et al., 1988) and 1 from near Tulelake near Klamath Falls, Oregon 125 km SSW of Summer Lake (Reick et al., 1992). The Olema tephra layer was previously estimated to have an age of 55–75 ka based on extrapolation of sedimentation rates determined from radiocarbon dates and palynological correlation with oceanic isotopic records (Rieck et al., 1992). The stratigraphic position of this tephra layer in the WL core in relation to the 45–50 ka Mt St Helens Cy ash (Berger, 1991; Berger & Busacca, 1995) and the 72.0 ± 6 ka Pumice Castle tephra layer (Bacon, 1983; Davis, 1985) suggests a slightly younger age (50–55 ka) for the Olema ash (Figure 4a).

Lithostratigraphy

The lithosome boundaries presented in Figure 4a support the tephra-based correlations between the cores and outcrop discussed above. Specifically, the Ana River outcrop and the WL core contain an interval of sediments that are consistent with shallow lake conditions beginning above tephra layer N1 and ending below tephra layer 2. Deeper lake conditions can be inferred from about 2 meters below tephra layer 2 upward to just below the Mt. St. Helens Cy tephra layer. Relatively shallow lake conditions prevailed from this point upward to above the Wono tephra layer.

Sediment magnetism

Three distinct NRM_{25} features (N1–N3) can be correlated between the Ana River outcrop and the WL core records (Figure 4b). These correlations are consistent with, and supplement, the correlations based on tephrochronology and lithostratigraphy.

N1 and N2 are sharp, high amplitude peaks in NRM_{25} intensity that are found in both the AR and WL records, a meter or 2 below tephra layer 2. N3 lies immediately below these features in both records; it is a prominent, relatively broad interval of extremely low and constant NRM_{25} intensity. Below N3, in both cases, are the N1, P, Q and T tephra layers. Notably, a prominent ostracode lag deposit is also found at the base of N3 in both the

outcrop exposures and in the WL core (Figure 4b). This lag deposit, which is bounded in the outcrop by tephra layers R (165 ± 19 ka) and N (102 ± 11 ka), defines the unconformity associated with marine oxygen isotope stage 5e as noted previously by Davis (1985) and Berger (1991).

A broad low in κ , that is consistent with N3 and the shallow lake lithosome, is evident in the outcrop and the WL core (Figure 4, b & c). This is not surprising given that both NRM intensity and κ are largely dependent on the concentration of magnetic minerals.

Four prominent and sharp peaks in paleomagnetic inclination (I1–I4) appear in the Ana River outcrop record below tephra layer 2 and above the unconformity at the base of N3 (Figure 4d). These features can be correlated with similar features in the WL core inclination record. The correlation of features I1 and I2 is straightforward because they coincide with tephra layers and because their morphology is very similar between core and outcrop where they reside in intervals of similar sedimentation rate. Correlation of I3 and I4 is less straightforward because these features lie within an interval of dramatically higher sedimentation rate in the WL core. Thus they appear much broader in the WL core record relative to the outcrop record. Because Tephra layer N (102 ± 11 ka) lies within the interval of Ana River outcrop sediments that corresponds to feature I4, we use I4 as a correlation point with which to project this date onto the WL core (Figure 4d).

The inclination signature of the Mono Lake Excursion (MLE) has been previously identified in the outcrop sediments immediately below the Wono and G tephra layers (Negrini et al., 1984; Negrini & Davis, 1992). As mentioned above, the MLE is also evident in the same relative position in both core records (Figure 4d). When compared to the high quality record of the MLE from the sediments of Pluvial Lake Russell (Lund et al., 1988), it is apparent that the inclination record from the B&B core contains a particularly well-resolved record of the excursion, as well as a high quality replication of the paleosecular variation found immediately above the MLE (Figure 6). An earlier excursion (EE) is found in the B&B core approximately 2 m below the MLE and 4 m above the 47 ± 2.0 ka Mt St Helens Cy tephra layer (Figure 4d). We correlate this feature with the 34–32 ka excursion also found below the MLE in DSDP core 480 (Levi & Karlin, 1989). These paleomagnetic correlations, in conjunction with the radiometric and TL dates on the Wono and Mt St Helens Cy tephra (Table 1), serve as the basis for dating the B&B core and the uppermost part of the WL core. We note that the recently revised

age (27.3 ± 0.3 ^{14}C kyr B.P.) of the Wono tephra layer (Benson et al., 1997) is consistent with the age of its position on the Lake Russell PSV curve as projected by the PSV correlation shown in Figure 6. The chronology of the Lake Russell sedimentary sequence is summarized in Benson et al. (1990).

Relative sedimentation rates and completeness of the records

The age versus depth curves for the Summer Lake outcrop and core records (Figure 7) are based on the outcrop chronology (Figure 3 and Table 1), as well as on the previously published chronology of the Lake Russell sediments which are exposed in the banks of Wilson Creek (Benson et al., 1990), and the chronology of the DSDP Hole 480 (Levi & Karlin, 1989). The core chronologies were tied to these established chronologies via the correlations presented above and via the data summarized in Tables 2–4. An age was

calculated for every sample from each record by linear interpolation between correlated age points, based on the relative depths of the samples and the age points.

A basinward trend of increasing sedimentation rate is clearly evident in the relative chronologies of the Summer Lake outcrop and cores (Figure 7). This relation is most prominent during intervals of low lake level. For example, the average sedimentation rate of the WL core during shallow lake conditions is 4.5 times that of the Ana River marginal outcrop, whereas, during deep lake conditions, the sedimentation rates are more similar. We infer from this observation that, when lake levels were high, the 2 core locations and the outcrop were all well below lake level and fluvially transported detritus was distributed relatively evenly along the basin axis. On the other hand, during intervals of low lake level, the higher elevation sites (e.g., the basin margin outcrop) were often at or above lake level. Thus these sites were not receiving sediment and may have even lost sediment through erosion by wave

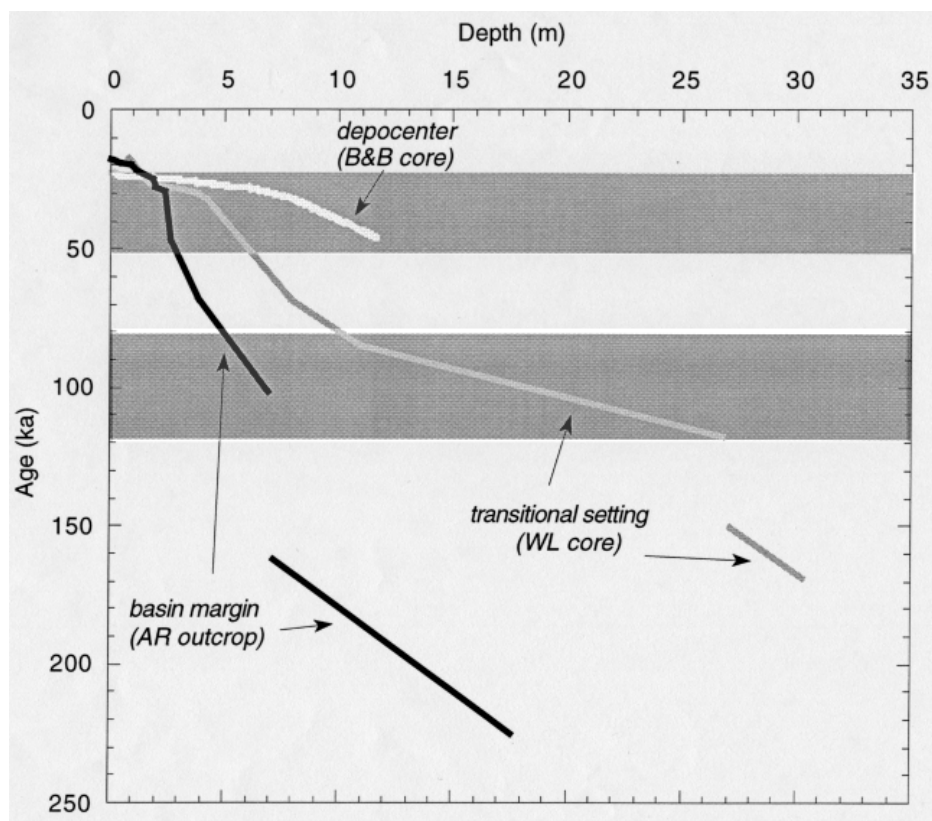


Figure 7. Age vs. depth plots for the three Summer Lake records. Note that the sedimentation rates increased dramatically toward the basin depocenter during intervals that correspond to shallow lake lithosomes (shaded regions). In contrast, sedimentation rates were similar during deep lake intervals (unshaded regions).

action, by meandering of streams, and/or by deflation, which are active processes in the modern basin. The resultant reworked detritus was then transported into the remaining shallow lake thus exacerbating the contrast in sedimentation rates between basin margin and depocenter sites.

As a result of these processes, the basin margin outcrop is missing ~ 60 ka of record from ~ 160 to ~ 100 ka. The WL core site is at a lower elevation and was probably under water more often than the basin margin; this core therefore is missing less of the record, perhaps as little as 35–40 ka based on extrapolation of sedimentation rates above and below the unconformity (Figure 7). Because the missing parts of these records are probably due to both nondeposition and erosion of previously deposited sediments, the period of low-lake level responsible for the stratigraphic hiatus probably occurred somewhere near the middle of the missing time period rather than near the beginning or end. Davis (1985) and Berger (1991) attributed this gap in the outcrop record to low-lake levels associated with the interglacial period corresponding to marine oxygen isotope stage (MIS) 5. The transition from the maximum glacial conditions of MIS 6 (~ 135 ka) to the maximum interglacial conditions of MIS 5 (~ 125 ka) took place within several thousand yrs or less (Martinson et al., 1987). Thus, if lake levels at Summer Lake are directly linked to the global glacial/interglacial cycles, then ~ 30 ka or more of lake sediments were eroded from the Ana River outcrop and sedimentation did not effectively resume until ~ 30 ka after the onset of MIS 5.

Negrini & Davis (1992) noted a 4,000 yr gap in the basin margin outcrop record from ~27 to ~23 ka. They attributed this hiatus to low water levels corresponding to the same arid conditions presumably responsible for the 'Wizard's Beach Recession' at Pyramid Lake, Nevada (Dansie et al., 1988). Because this interval of time is clearly represented in the B&B core (Figure 6), it is evident that the Summer Lake basin did not dry up completely during this period. Sedimentological, geochemical, and paleontological indicators from these depocenter sediments support the existence of a warm shallow lake at this time (Cohen et al., 2000).

The lowering of lake level can also be caused by local processes such as stream diversion. In fact, even though Summer Lake is the sub-basin with the lowest elevation in the Lake Chewaucan pluvial system, it currently receives no surface recharge from the diverted Chewaucan River, which is the only major stream that feeds the Lake Chewaucan pluvial basin (Allison, 1982; Davis, 1985). Though the coincidental timing of several

lake level fluctuations observed both here (see below) and in the companion paper (Cohen et al., 2000) favor global climate change as the primary cause of lake level change in the Summer Lake sub-basin, stream diversion cannot be ruled out as a contributing factor.

Environmental magnetism

In environmental magnetism, magnetic properties are used to infer changes in concentration, grain-size, and mineralogy of magnetic minerals in sediments. Such quantitative petrologic information can, in turn, implicate changes in weathering and erosion, sediment transport and lacustrine depositional environments (e.g., lake level) caused by either regional or even global climate or by local geomorphic events (Thompson & Morton, 1979; Thompson & Oldfield, 1986; Snowball, 1993; Thouveny et al., 1994; Peck et al., 1994; Jelinowska et al., 1995; Verosub & Roberts, 1995; Rosenbaum et al., 1996). Many magnetic parameters, especially magnetic susceptibility (κ), can be measured rapidly, thereby allowing unprecedented resolution, even for continuous sampling of sedimentary sequences. Furthermore, these measurements are generally non-destructive thereby saving the samples for additional analysis. In the remainder of this paper we apply environmental magnetic methods to assist in constructing a model of depositional environments for Lake Chewaucan over the last 250,000 yrs.

Methods

The sediment magnetism measurements used are low-field volume susceptibility (κ), susceptibility of anhysteretic remanent magnetization (κ_{ARM}), isothermal remanent magnetization (IRM), the S parameter, and 4 hysteresis parameters: Ms (saturation magnetization), Mrs (saturation remanent magnetization), Bc (coercive force), and Bcr (coercivity of remanence). κ was measured as described in the previous section. κ_{ARM} is the susceptibility of samples to acquiring magnetizations in a small bias field that is superimposed on a stronger, decaying alternating field (AF). In this study, all samples were given ARMs with a 0.05 mT bias field with a 100 mT AF in a Molspin AF demagnetizer. κ_{ARM} was calculated by dividing the resultant ARM by the applied bias field. IRM is the intensity of remanent magnetization acquired during magnetization in a strong static field. IRMs were given to all of the samples in a 100 mT static field with a model 6002

electromagnet manufactured by Alpha Scientific. The S parameter (see Versoub & Roberts, 1995) is the absolute value of the ratio of the IRM acquired by the samples in a relatively small reversed magnetic field (0.3 T) to the IRM previously acquired by the samples in a larger forward magnetic field (1.2 T). The S parameter essentially determines the relative amount of high susceptibility ferrimagnets (i.e., magnetite-like minerals) and low susceptibility canted antiferromagnets (i.e., hematite-like) in the sediments. If the sample is composed entirely of ferrimagnets, then this ratio will simply be equal to unity because these minerals saturate completely in applied fields of 0.3 T or less. If the sample contains an appreciable amount of canted antiferromagnets, then this ratio will take on a value between zero and 1. The hysteresis parameters (M_s , M_{rs} , B_c , and B_{cr}) were determined as part of a previous study (Roberts et al., 1994).

Magnetic susceptibility as a proxy for lake level and global climate change

As noted above, κ values for the Summer Lake sediments are high in high-lake lithosome intervals and vice-versa (Figure 4, a & c). This relation is strengthened in the companion paper by Cohen et al. (2000; Figure 17) where major trends in κ are shown to be consistent with several sedimentological, geochemical and paleontological indicators of lake depth (e.g., ostracode paleosalinity indices). Furthermore, when κ is plotted against age rather than depth, it becomes clear that κ , and hence, lake level, correlate temporally to first order with established proxies of global climate change (Figure 8). Specifically, given the uncertainty in the age control, the Summer Lake κ signal varies in concert with the major global climate changes associated with marine $\delta^{18}\text{O}$ stages 3–7. There may even be a response to the more prominent substages (e.g., 5a–e) though this relationship is less convincing (Figure 8a). In contrast, though cycles with periods of approximately 2–3 ka are apparent in the κ record from the B&B core (Figure 8b), these features are difficult to correlate with climate variations of similar periods exhibited in records from high latitude ice cores such as the GISP2 temperature record shown also in Figure 8b (Grootes et al., 1993).

Origin of magnetic susceptibility changes in the Summer Lake sediments

The above observations suggest that κ can be used, with caution, as a paleoenvironmental and chronological tool

in studies of the Summer Lake pluvial sub-basin. To better understand the linkage between κ and lake level (and, therefore, to increase our confidence in the use of κ as a lake level proxy), we first attempt to determine which of the 2 major environmental magnetic influences (magnetic mineralogy or concentration) is responsible for the connection between lake levels and κ .

Variations in magnetic mineralogy have been shown to contribute to changes in κ due to important events within lake catchments, such as through the nearly complete replacement of high- κ magnetite-like minerals by low- κ hematite-like minerals (e.g., Thompson & Oldfield, 1986; Rosenbaum et al., 1996). In order to test for mineralogical variations, the S parameter was measured for 17 samples from the Ana River outcrop and 15 samples from the WL core. At least 2 samples were chosen from each interval of high and low susceptibility for each location. Thus, if changes in the dominant magnetic mineral are responsible for major decreases in susceptibility, such changes should be detectable in this sample set. As shown in Figure 9, all of the measured samples from both the Ana River outcrop and the WL core had S values of at least 0.97, with most values between 0.99 and 1.00. This variation lies within the range of repeated measurements on individual samples and is consistent with the dominance of low coercivity minerals such as magnetite within both the low and high κ intervals. These results are consistent with an initial petrographic study of polished grain mounts from 10 magnetic separates from Summer Lake sediment (R.L. Reynolds, U.S.G.S., Denver, written communication, 1999). The petrographic observations indicate that magnetite and titanomagnetite dominate the magnetic mineral assembly in the Summer Lake sediments, as was previously concluded by Roberts et al. (1994). Based on the above evidence, we conclude that changes in the concentration of magnetite-like minerals are responsible for high amplitude changes in κ rather than changes in the magnetic mineralogy.

Changes in magnetic grain size due to changes in lake level

To first order, κ_{ARM} and IRM should, like κ , depend on the concentration of magnetic minerals. Thus, it comes as no surprise that all three of these parameters display similar variations for both the Ana River outcrop and the WL core (Figure 9). Small amplitude changes in these parameters may nevertheless be present as a result of variations in the size of the magnetic minerals. For example, because IRM decreases with increasing grain

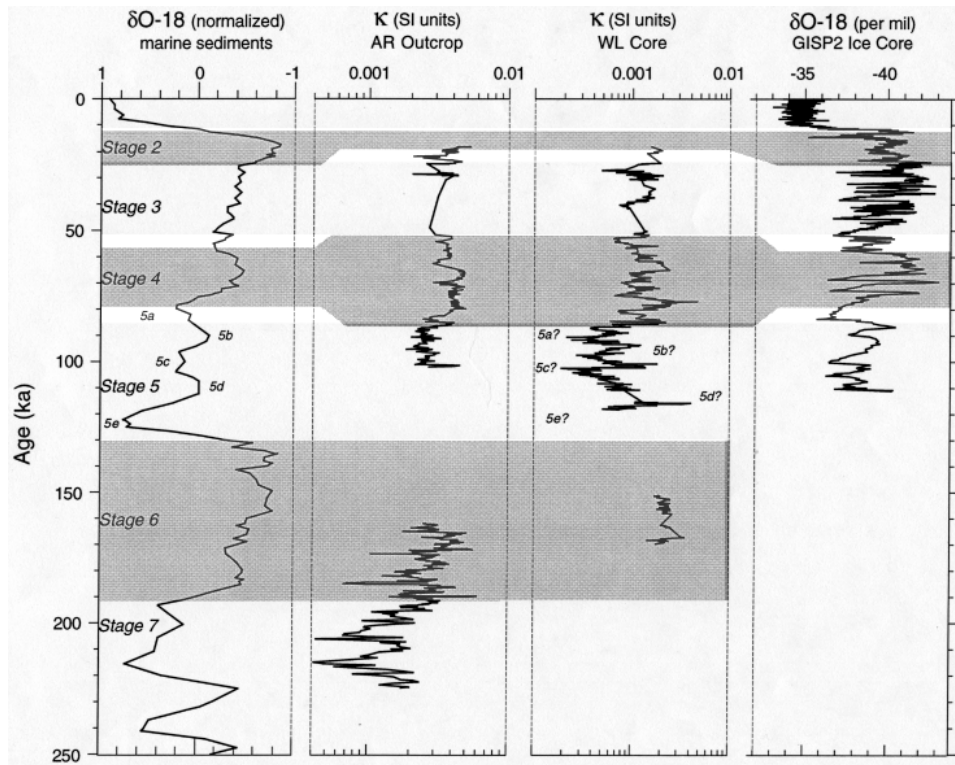
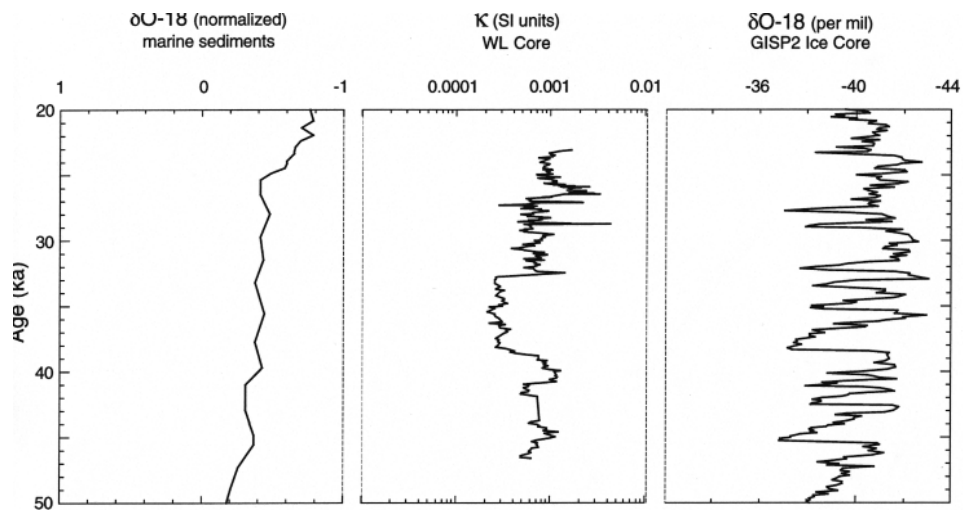
a**b**

Figure 8. (a) Volume susceptibility (κ) records from the Summer Lake sub basin (AR outcrop and WL core) compared to $\delta^{18}\text{O}$ proxies for global climate change. Summer Lake ages are based on the chronology shown in Figure 3. The marine $\delta^{18}\text{O}$ record is from Martinson et al. (1987). The GISP2 ice core $\delta^{18}\text{O}$ record is from Grootes et al. (1993). Major changes in global climate are clearly represented in the susceptibility records from Summer Lake sub basin. Tentative correlations with the substages of Stage 5 are indicated on the WL core record. Note that relatively high κ values in the Summer Lake sediments occur during glacial intervals and subintervals; (b) High resolution κ records from the Summer Lake sub-basin (B&B core) compared with the same indicators of global climate change plotted in Figure 8a. Though a slow progression towards more glacial conditions is reflected in a gradual overall increase in κ over time, high frequency variations cannot be correlated between the κ record from the B&B core and the high resolution $\delta^{18}\text{O}$ record from the GISP2 ice core.

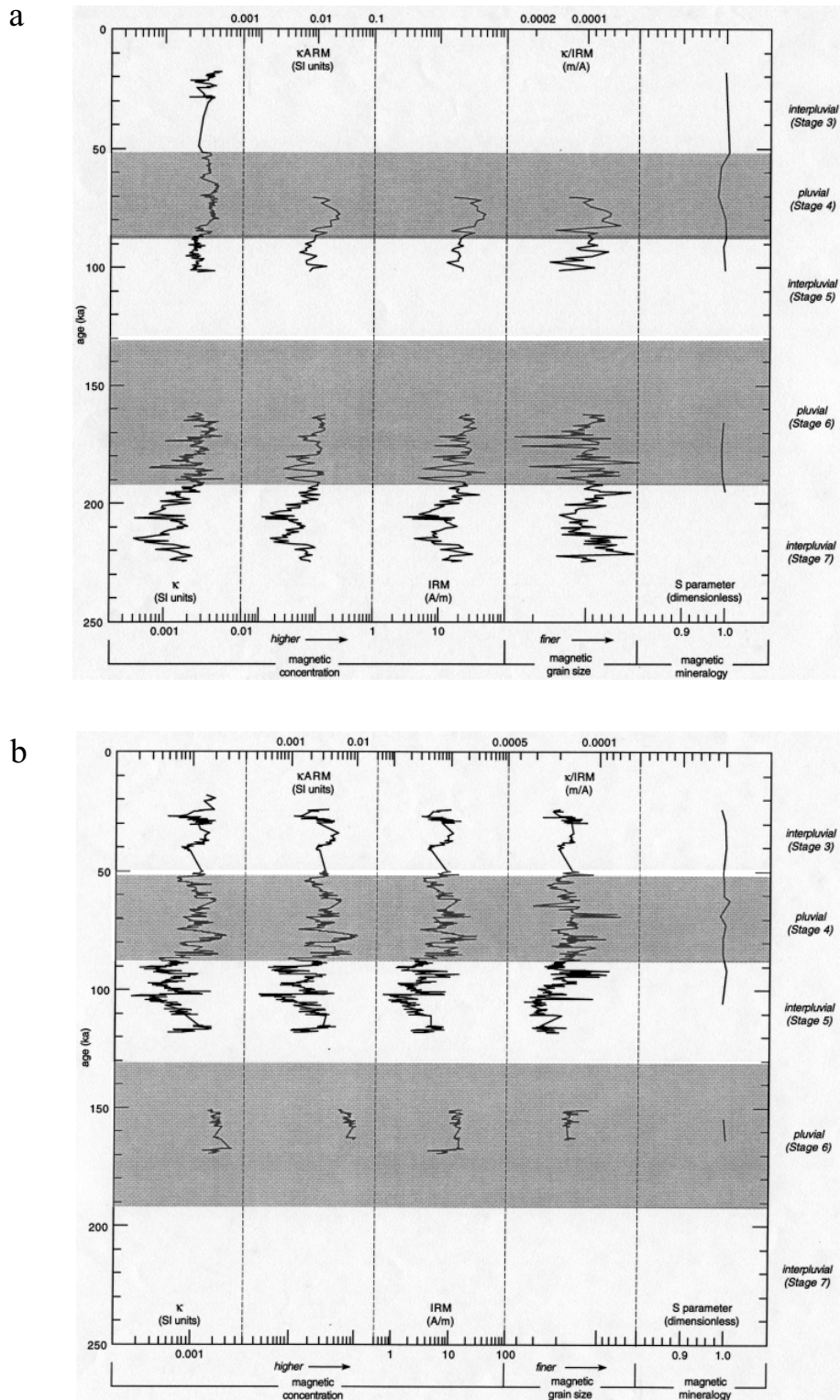


Figure 9. (a) Environmental magnetism of the Ana River outcrop sediments; (b) Environmental magnetism of the WL core sediments.

size over the range of sizes relevant to this study (Parry, 1965; Thompson & Oldfield, 1986) and because κ is relatively insensitive to changes in grain size (Heider et al., 1996), the ratio of κ to IRM can be used as an indicator of magnetic grain size. This estimate of magnetic grain-size is plotted in both Figure 9, a & b with scales reversed so that finer grains plot toward the right of the diagram. In both the outcrop and core sediments, the variations in magnetic grain-size usually vary in opposition to magnetic concentration. That is, when the concentration of magnetic minerals is high, the magnetic grain-size is relatively fine and vice-versa. Thus, pluvial intervals are characterized by smaller magnetic grains and interpluvials are characterized by larger magnetic grains.

The above hypothesis can be tested by comparing the hysteresis parameters of the 'low-lake' and 'high-lake' samples. That is, samples with smaller magnetic grain sizes should have higher ratios of M_{rs}/M_s and lower ratios of B_{cr}/B_c and vice-versa (Day et al., 1977). Hysteresis parameters were measured as part of an earlier study of the Ana River outcrop sediments (Roberts et al., 1994) in which all of the data were plotted together in order to demonstrate that the magnetization of these sediments was carried entirely by pseudo-single domain magnetite grains. In the present study, the data are separated into 4 subsets according to pluvial or interpluvial intervals (Figure 10). The slight but discernible bias of 'high lake' samples toward the upper left of these diagrams and 'low lake' samples toward the lower right support our earlier conclusion that smaller magnetic grains are dominant during deposition in deeper lakes and vice-versa.

Comparison of magnetic and non-magnetic grain sizes

In order to test whether the size variations of the (titano)magnetic grains are representative of variations in the size of non-magnetic grains, we determined the bulk grain size of sediments from the WL core using both laser granulometry and pipette/sieve analysis (Erbes, 1996). Because 95% by weight of the Lake Chewaucan sediments consists of silt and clay-sized particles, we use the ratio of silt to clay to represent the overall granulometry of the WL core sediments.

The silt/clay data shown in Figure 11 generally demonstrate the same trends as the magnetic grain size indicators. That is, smaller grains characterized the sediments from high lake/pluvial intervals (MIS 6 and 4) and larger grains were characteristic during low lake/interpluvial intervals (MIS 5). The silt/clay ratios

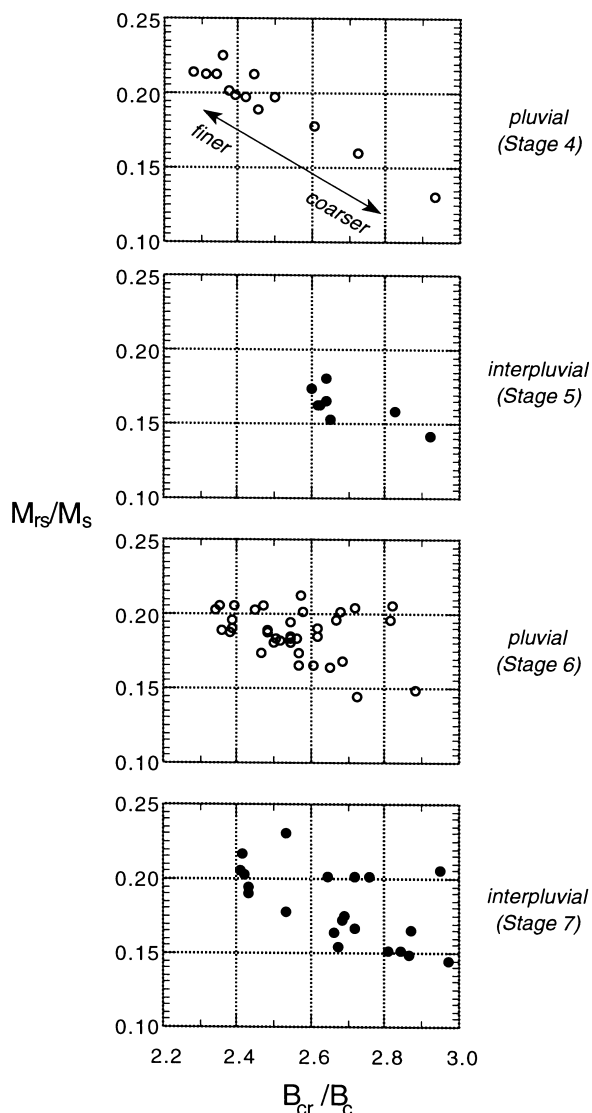


Figure 10. Magnetic grain size of Lake Chewaucan pluvial and interpluvial sediments from hysteresis parameters (after Day et al., 1977). Pluvial (interpluvial) sediments exhibit a clearly discernible bias toward smaller (larger) magnetic grain-size, interpluvial sediments toward larger grain-size. Note: all samples lie well within the field characteristic of pseudo-single domain grains (Roberts et al., 1994).

remained low (finer grains) during MIS 3. This exception is consistent with subdued Stage 3 responses in the environmental magnetic data (i.e., magnetic concentrations were not as low and grain sizes were not as large in Stage 3 as they were during other low lake intervals). In fact, these 'exceptions' are consistent with the marine oxygen isotope signal. That is, MIS 3 is an interpluvial/interglacial only in a relative sense; its

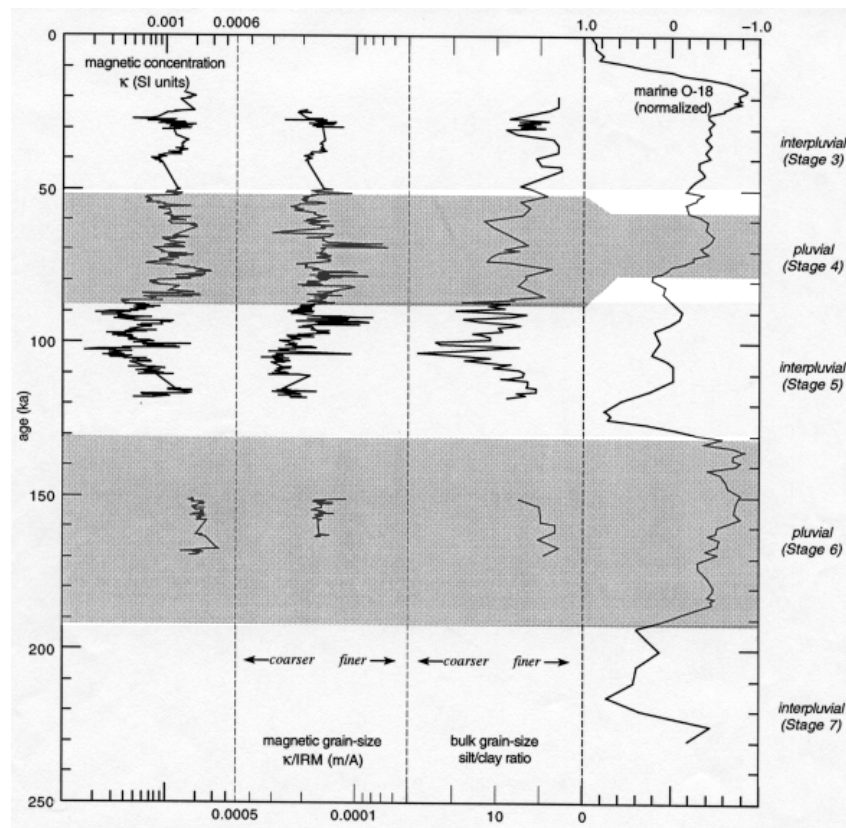


Figure 11. Grain size of magnetic minerals, as determined by the κ /IRM ratio, compared to grain size of bulk sediment (silt/clay ratio) determined by laser granulometer for the WL core. A magnetic concentration parameter (κ) and a global climate proxy (marine $\delta^{18}\text{O}$) are also plotted. As in Figure 9, magnetic concentration increases to the right, magnetic grain-size increases to the left.

isotopic signature is actually closer in value to those of MIS 2, 4 and 6 than those of interpluvial/interglacial MIS 5 or 7.

Discussion and conclusions

The following is a summary of observations that are pertinent to the depositional history of the Summer Lake sub-basin of Pluvial Lake Chewaucan during the last 250 ka. The observations are from this paper and from Roberts et al. (1994), Erbes (1996), and Cohen et al. (2000).

1. The lithostratigraphy of the sediments can be divided into deep lake (pluvial) and shallow lake (interpluvial) lithosomes.
2. The age control of these sediments, based on several independent methods, is consistent and precise, at least for the part of the record younger than 100 ka.
3. Sedimentation rates were relatively constant along

the basin axis during pluvial intervals. They increase dramatically, but only toward the depocenter, during interpluvial intervals.

4. The magnetic mineralogy is dominated by pseudo-single domain, (titano)magnetite throughout the last ~250 ka regardless of lake level and climate.
5. Magnetic and bulk sediment grain sizes were generally finer during deeper lake intervals and vice-versa.
6. The concentration of magnetic minerals was higher during intervals of deeper lake level and vice-versa.
7. To first order, intervals of deeper lakes in this sub-basin, as indicated by many proxies including magnetic susceptibility, correspond temporally to global glacial stages.

A depositional model for the Summer Lake sub-basin in response to global climate change

The above observations are consistent with the following depositional model for the Summer Lake

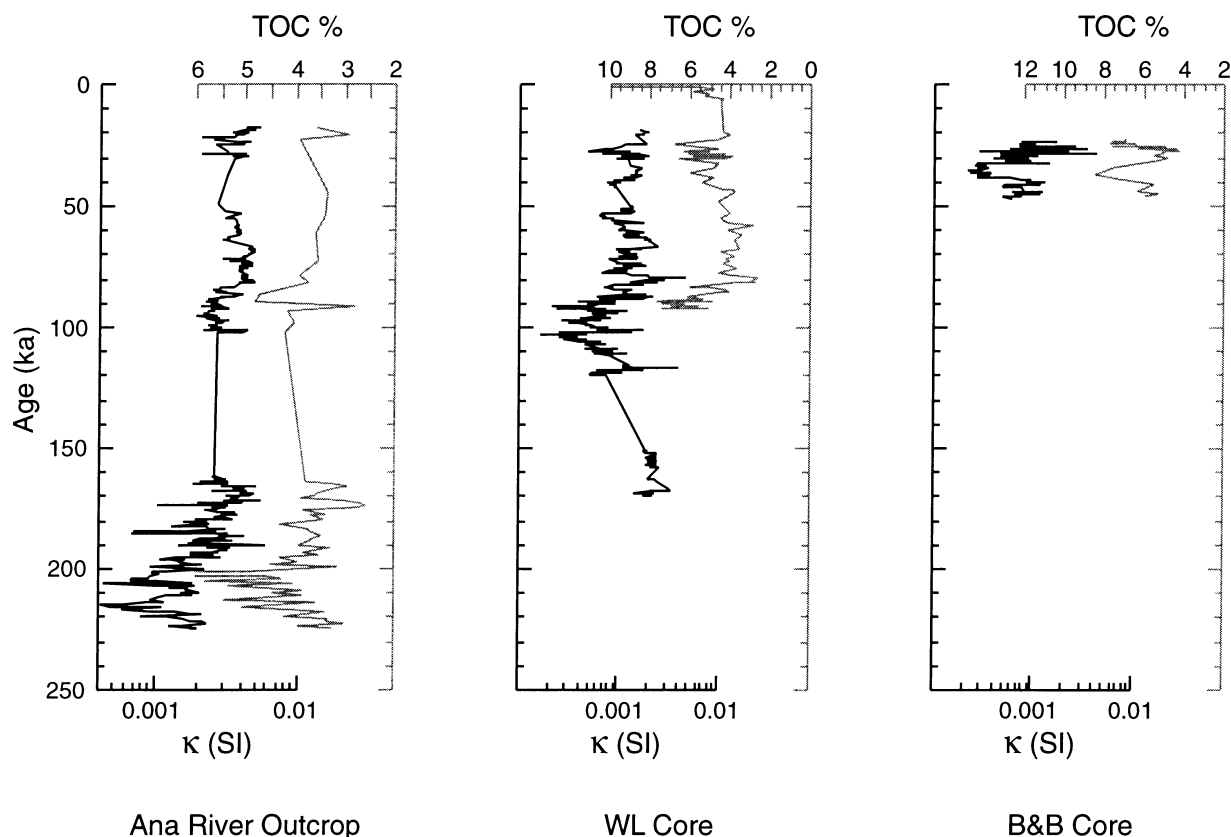


Figure 12. Comparison of total organic carbon (TOC) from Cohen *et al.* (this volume) with magnetite concentration as represented by magnetic susceptibility (κ). TOC increases to the left; κ increases to the right; lake level increases to the right.

sub-basin. First, during low-lake intervals, effective sedimentation rates became much higher toward the depocenter, because (a) the entire load of fluvial detritus is deposited in the relatively small lake located near the depocenter; (b) marginal deposits were eroded and these reworked sediments added to the sediment load at the depocenter. Second, when lake levels were high, grain size decreased throughout the lake because of lower depositional energy related to greater distances from the mouths of streams.

Third, changes in lake-level cause measurable variations in sediment magnetic properties throughout the lake. In particular, the lake-level response of the sediment magnetic properties is dominated by changes in (titano)magnetite concentration, with subsidiary changes in grain-size, but without perceptible changes in magnetic mineralogy. The clear connection between lake levels and sediment magnetic properties strengthens the case for the use of magnetic measurements as the basis for correlation of lake sediments and as paleo-environmental proxies. The exact mechanism by which

magnetic and lacustrine processes are linked remains unclear, however. For example, because (titano)-magnetite is a high density mineral, one would expect that magnetic concentration would have decreased along with magnetic grain size as the lake grew in size and deepened. In contrast, (titano)magnetite concentration consistently increased as the lake deepened. Two possible explanations for this paradox are dilution and/or dissolution of (titano)magnetite related to processes within the lake such as biological productivity which generally increases as the level of the lake drops. Dilution as a result of increased biological productivity during low lake levels probably does not suffice by itself because (titano)magnetite concentration usually decreased by a factor of 2–5 thus requiring increases in the diluting material amounting to 50–80% of the total sediment volume. In contrast, although biological productivity did indeed increase in the Summer Lake sub-basin during low lake level (e.g., Figure 12), changes in the percentages of organic and inorganic carbon combined rarely exceeded 15%

(Cohen et al., 2000). Furthermore, during routine inspection of smear samples under an optical microscope, diatoms were rarely observed in anything but trace amounts; thus biogenic silica was probably never very high in the Summer Lake sediments. Dilution by authigenesis, on the other hand, may contribute significantly to changes in the susceptibility of magnetite in the Summer Lake sub-basin based on the formation of substantive amounts of authigenic minerals such as smectites in the nearby Lake Abert sub-basin of Lake Chewaucan (Banfield et al., 1991). Notably, authigenesis in Lake Abert was suggested to be controlled, at least in part, by lake level.

Even if the increases in sediment volume due to biological productivity and authigenesis are not sufficient to explain decreases in the concentration of magnetite by dilution, small increases in total organic carbon (TOC) can lead to dramatic decreases in magnetite concentrations through targeted dissolution of magnetite due to oxidation reactions mediated by bacteria feeding on organic matter (e.g., Karlin, 1990; Leslie et al., 1990; Snowball, 1993; Rosenbaum et al., 1996). A strong correspondence between TOC and magnetic susceptibility in all three outcrop and core records (Figure 12) supports such a dissolution process as the mechanism for the inverse link between magnetic concentration and lake level in the Summer Lake sub-basin. Because a preferential dissolution of smaller (titano)magnetite grains has been noted in the above-cited studies, this process could also contribute to the increase in magnetic grain size with decreasing lake level observed in this study (Figures 9–11).

Fourth, we propose that major rises and falls of lake level in the Summer Lake sub-basin are caused by changes in climate associated with the growth and recession, respectively, of the North American ice sheet, for which the global marine oxygen isotope signal has been shown to be a nearly synchronous proxy (Mix, 1987; Ruddiman, 1987). The mechanism for this causal relationship is the southward/northward migration of the mean position of the northern jet stream storm track in response to the growth/recession of the North American ice sheet as proposed originally by Antevs (1948). It is important to note that this conclusion is based on an independent chronology for the Summer Lake sediments rather than on assumed correlations of key lacustrine events with key events in the global climate record.

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