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Middle/Late Pleistocene relative palaeointensity of the geomagnetic field from lacustrine sediments, Lake Chewaucan, western United States

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SUMMARY

Detailed palaeomagnetic and rock magnetic studies of a 15 m succession of Middle/Late Pleistocene lacustrine sediments from ancient Lake Chewaucan, southern Oregon, western United States, indicate that the remanence-bearing grains are sufficiently uniform to be applicable to relative palaeointensity studies. We have used ARM, SIRM and χ for normalization of the NRM. All three parameters give essentially identical results in their relative stratigraphic variations, which indicates that the normalizations efficiently remove the effects of variation in magnetic mineral concentration. Patterns in grain-size variation, as indicated by small-scale quasi-cyclic fluctuations in hysteresis parameters, may be due to environmental changes such as lake-level variation. However, these fluctuations are within the acceptable range of grain sizes for palaeointensity studies and cannot be correlated with any of the features of the normalized remanence record. We therefore conclude that the large-scale variations in the normalized remanence record are due to geomagnetic palaeointensity fluctuations. Parts of the normalized remanence record, where firm chronological constraints exist, may correlate with features of relative palaeointensity records from deep-sea sediments. Our results also confirm the observation that low geomagnetic field intensities dominate during geomagnetic excursions. Further studies of relative palaeointensity of the geomagnetic field may enable the development of an independent time-scale which would make possible the direct correlation of palaeoclimate records from deep-sea and continental environments.

Key words: lacustrine, Oregon, palaeointensity, palaeomagnetism, Pleistocene, rock magnetism, USA.

1 INTRODUCTION

Most palaeomagnetic studies concentrate on changes in direction of the geomagnetic field. Although such studies have provided much information about the geomagnetic field, they are fundamentally incomplete in that they do not describe the vector behaviour of the field. Measurements of both direction and intensity of the geomagnetic field are available from observatory and archaeomagnetic records, but they do not extend far back in time, and they do not provide much evidence concerning long-term variations. Studies of volcanic sequences can potentially yield long-term records, but high-precision dating of such successions is difficult and the temporal resolution of the resulting palaeointensity record is dependent on the frequency of volcanic eruption. Long-lived volcanoes that erupt frequently are rare, thus extensive records are difficult to obtain.

Sedimentary records, particularly from the deep-sea, are usually deposited continuously and can have high-resolution time-scales (e.g. $\delta^{18}O$ stratigraphy). Sediments therefore represent an excellent opportunity to investigate the direction and intensity of the geomagnetic field. Although palaeomagnetic directions can be measured relatively easily, palaeointensity determination is less straightforward. The intensity of magnetization of a sediment (the natural remanent magnetization, or NRM) is a function of both the palaeointensity of the geomagnetic field and the mineralogy, concentration and grain size of the magnetic grains in the sediment. The effects of these factors can be separated by using different laboratory-induced magnetizations to understand the NRM behaviour (Opdyke, Kent & Lowrie 1973;

Banerjee & Mellema 1974; Levi & Banerjee 1976; Tucker 1981). The three most common types of magnetization are anhysteretic remanent magnetization (ARM), saturation isothermal remanent magnetization (SIRM) and low field magnetic susceptibility (χ). If a laboratory-induced magnetization is always produced under the same conditions, the intensity should be proportional to the concentration of the magnetic carriers. Therefore, if one takes the ratio of the NRM to the intensity of the laboratory-induced magnetization, the effects of concentration should cancel, leaving an approximation to the relative palaeointensity of the geomagnetic field. King, Banerjee & Marvin (1983) have suggested that this approach is most reliable when the sediment under investigation is 'uniform' with respect to mineralogy, grain size and concentration of remanence carriers. Their criteria are: (1) the dominant magnetic mineral must be magnetite; (2) the particle size range for the magnetite must be pseudo-single domain (PSD), between 1 and 15 μm; and (3) the maximum magnetite concentration must be less than 20–30 times greater than the minimum concentration. The last criterion has been suggested as being too liberal (Tauxe 1993) because a factor of 30 change in concentration can result in a 10 per cent difference in ARM acquired in a given field. Tauxe (1993) therefore suggests a limit of a factor of 10 or less for changes in concentration when ARM is used for normalization.

Several recent studies have attempted to obtain relative palaeointensity records from deep-sea sediments (Constable & Tauxe 1987; Tauxe & Valet 1989; Tauxe & Wu 1990; Meynadier *et al.* 1992; Tric *et al.* 1992). Tric *et al.* (1992) produced a record from the Mediterranean Sea that extends back 80 kyr, which is in agreement with, and has been calibrated against, palaeointensity data from lavas covering the period 0–40 kyr. This record shows significant agreement with earlier studies from the western equatorial Pacific by Constable & Tauxe (1987) and Tauxe & Wu (1990). Meynadier *et al.* (1992) extended the record back to 140 kyr in the Somali Basin where they were able to reproduce much of the detailed record of Tric *et al.* (1992). All of these records have significant intensity lows at about 40 and 60 kyr and a high at 80 kyr (Tauxe 1993). The coherence of the existing records from around the world, over distances of several thousands of kilometres, and over time periods of tens of kyr to hundreds of kyr, is interpreted to reflect changes in the Earth's dipole field (Meynadier *et al.* 1992; Tric *et al.* 1992). These results mark a significant step towards the establishment of a palaeointensity reference curve for the last several 100 kyr, and based on data covering intervals including several geomagnetic reversals over the past several million years, Valet & Meynadier (1992) suggest that palaeointensity of the geomagnetic field may provide the basis for a new global stratigraphy.

A further test of the validity of the relative palaeointensity method is provided by the study of records from entirely different depositional environments. We have been studying lacustrine sequences from the western United States in order to determine whether they contain the same relative palaeointensity records as deep-sea sediments. The western United States and north-eastern Pacific Ocean are ideal localities for such a study because these lacustrine and marine sediments often contain volcanic ash layers from major eruptions of Cascade and other volcanoes (Sarna-Wojcicki *et al.* 1987). The trace-element geochemistry of each eruption is usually quite distinct, and extensive catalogues of common tephra from the western United States now exist (Sarna-Wojcicki *et al.* 1991). A sedimentary sequence can therefore be placed within the chronological framework provided by the tephrostratigraphy. Additionally, many of the major lakes in the western United States are at low enough elevations and at low enough latitudes that they were not covered with ice during glacial intervals and did not dry out completely during the warmer interglacial intervals. As a result, these lakes often have basins where a nearly complete history of lake level and palaeoclimate fluctuations has been preserved (e.g. Smith 1984; Adam *et al.* 1989, 1990; Benson *et al.* 1990).

2 LAKE CHEWAUCAN

We have studied a 15 m sequence of Middle/Late Pleistocene lacustrine sediments from Lake Chewaucan, southern Oregon (Fig. 1). Several studies of palaeomagnetic directions have been reported from the Lake Chewaucan sediments (Negrini, Davis & Verosub 1984; Negrini, Verosub & Davis 1988; Negrini & Davis 1992). The sequence contains over 50 tephra layers. Some of the tephra layers have been directly dated (Berger 1991) and others have been correlated, by trace-element geochemistry, to dated tephra layers from Cascade volcanoes (Davis 1985; Sarna–Wojcicki *et al.* 1991). The presence of numerous readily recognizable tephra layers facilitates correlation with other localities where relative palaeointensity work is currently ongoing.

Age constraints for the Lake Chewaucan sediments are summarized in Table 1 and Fig. 2. These constraints are largely provided by thermoluminescence (TL) dates on tephra layers (Berger 1991) and are consistent with ages based on K/Ar and radiocarbon dating of stratigraphic sections from other localities.

The youngest part of the Lake Chewaucan chronology is well constrained by radiocarbon and TL dating. Correlation of palaeo-secular variation features between Lake Chewaucan and a well-dated (^{14}C) Pleistocene lacustrine succession from Lake Russell near Mono Lake, California, supports this chronology (Negrini & Davis 1992). The only samples from this part of the Lake Chewaucan succession analysed in the present study occur in the vicinity of the Mono Lake excursion (29–27 kyr), as reported by Negrini *et al.* (1984).

TL dates have also been obtained lower in the Lake Chewaucan succession from tephra 12 to tephra N (Berger 1991; Table 1). Dates of 165 ± 19 kyr for tephra R and 102 ± 11 kyr for tephra N differ significantly, even though these tephra occur stratigraphically close together. This result is consistent with the presence of a major unconformity between tephra N and tephra R. The presence of a prominent ostracode lag deposit (Davis 1985), the coated nature of the ostracodes (Cohen, A., University of Arizona, private communication, 1991), and the absence of a prominent tephra (N1), which is preserved above this deposit towards the depositional centre of the lake, all suggest that this interval represents a significant sedimentary hiatus. This is the only prominent unconformity in the

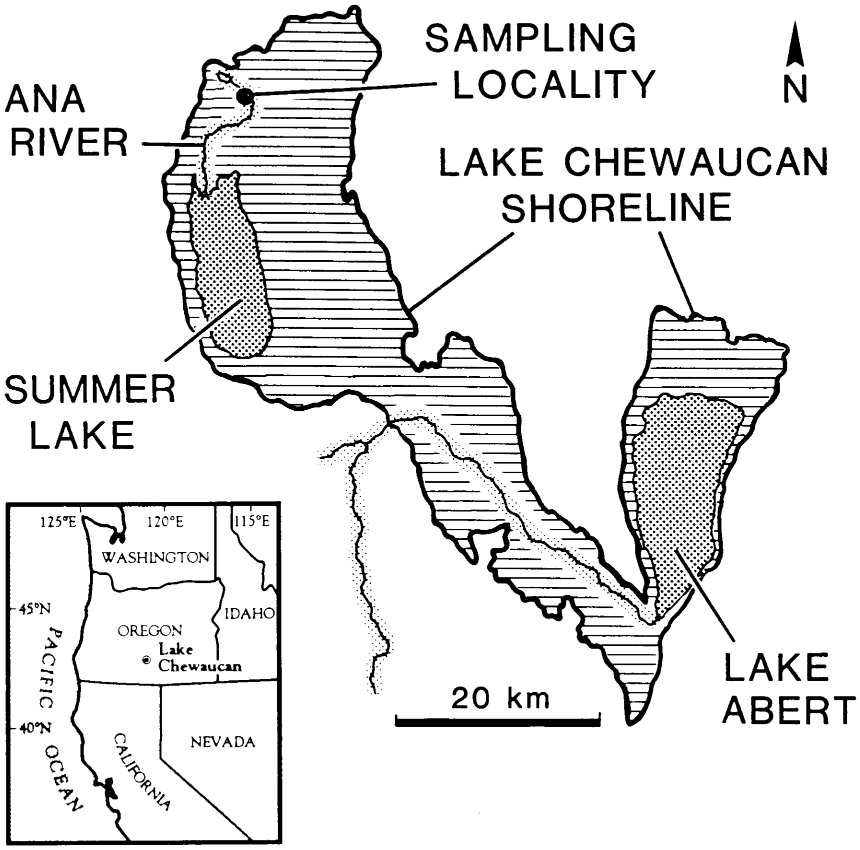

Figure 1. Shoreline of the Pleistocene lake and its Holocene remnants, Summer Lake and Lake Abert. Inset: map of western United States showing location of Lake Chewaucan.

succession studied here. Its age is consistent with that of the oxygen isotope stage 6/5 boundary (130 kyr; Martinson *et al.* 1987).

The TL dates from tephra R and below have large standard errors, and we can only effectively constrain the age of the entire lower part of the sequence to be between 200 kyr and perhaps 130 kyr. We are currently undertaking to resolve the uncertainties in the chronology in this part of the sequence using ^{40}Ar/^{39}Ar dating. We have attempted to obtain a relative palaeointensity record for the entire sequence at Lake Chewaucan. However, due to the unresolved chronological problems, we will make correlations only with the upper part of the record, above the prominent unconformity. We are confident that the age control in the upper part of the succession is accurate to within 10 per cent.

3 METHODS

All remanence measurements were made either on a horizontal three-axis 2G Enterprises cryogenic magnetometer or on a Schonstedt spinner magnetometer. All

Table 1. Summary of age constraints for Lake Chewaucan sediments.

Dated interval or horizon	Depth (relative to Tephra 12)	Age (kyr)	Dating method	Reference
Mono Lake Excursion to top of section	16 - 261 cm above	28.9 - 16.7	Palaeomagnetic correlation	Negrini and Davis (1992)
Tephra D	147 cm above	18.1 - 20.8	Radiocarbon	Davis (1985)
Trego Hot Springs Tephra	113 cm above	24.3 ± 2.7	Thermoluminenscence	Berger (1991)
Tephra 12 (Mt St Helens Cy)	0	35 - 50	Radiocarbon	Various dates summarised by Sarna-Wojcicki *et al.* (1991), Crandell (1987) and Davis (1985)
Tephra 12 (Mt St Helens Cy)	0	50.2 ± 3.4	Thermoluminescence	Berger (1991)
Tephra 6 (Pumice Castle)	91 below	72 ± 6	K/Ar	
Tephra 2	113 below	67.3 ± 7.5	Thermoluminescence	Berger (1991)
Tephra N	393 below	102 ± 11	Thermoluminescence	Berger (1991)
Tephra R	460 below	165 ± 19	Thermoluminescence	Berger (1991)
Tephra V	607 below	100 - 200	Tephrochronology and K/Ar	Sarna-Wojcicki *et al.* (1991)
Tephra KK	1042 below	100 - 200	Tephrochronology and K/Ar	Sarna-Wojcicki *et al.* (1991)
Tephra KK	1042 below	200 ± 27	Thermoluminescence	Berger (1991)
Tephra LL	1093 below	162 ± 35	Thermoluminescence	Berger (1991)

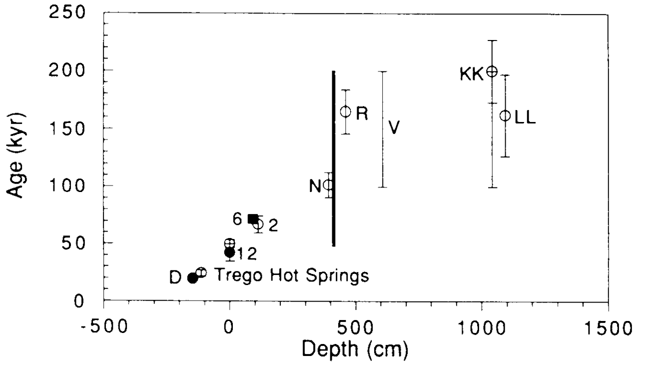

Figure 2. Age versus depth for the Lake Chewaucan sedimentary sequence. Names of tephra layers are given beside each date. Tephra 12 is taken as the zero depth for this study: all stratigraphic thicknesses cited are depths relative to tephra 12 at Ana River section C (Davis 1985). Open circles represent TL dates (Berger 1991); solid circles represent radiocarbon dates; solid square represents K/Ar date tied into section by tephrochronologic correlation; lines without symbols represent approximate age constraints from tephrochonologic correlation (Table 1); heavy line represents major unconformity at oxygen isotope stage 6/5 boundary.

alternating field (AF) demagnetizations were made using a Schonstedt AF demagnetizer. ARMs were imparted on one sample axis in an alternating field of 100 mT with a coaxial bias field of 0.05 mT. IRMs were induced on one sample axis up to maximum fields of 500 mT with a 4 inches (1 inch = 25.4 mm) electromagnet. Bulk magnetic susceptibilities were measured with a Bartington Instruments M.S.2 magnetic susceptibility meter. Hysteresis parameters were measured on small subsamples (20–40 mg) taken from larger palaeomagnetic samples after all other analyses had been completed. A Princeton Measurements Corporation Micromag Alternating Gradient Magnetometer was used for hysteresis measurements, up to maximum fields of 1 T. Because of the small sample size, several subsamples from the same sample were measured in some cases in order to test the consistency of the results. In each case, the replicate analyses did not differ significantly from the initial hysteresis measurement, indicating magnetic homogeneity of the sediment at least at the scale of single palaeomagnetic samples.

4 ROCK MAGNETISM

Our initial approach to determining the suitability of the Lake Chewaucan record for relative palaeointensity studies has been to examine the rock magnetic properties of the sediments in detail in order to ascertain whether the sediments satisfy the criteria for magnetic uniformity, as suggested by King *et al.* (1983).

For any sample, the presence of magnetite can be readily determined from the high field behaviour and thermomagnetic properties. SIRM acquisition experiments indicate that the magnetic mineralogy at Lake Chewaucan is dominated by a strongly ferrimagnetic mineral with maximum coercivities of around 200–300 mT (Fig. 3a). The SIRM acquisition behaviour of the subsamples subjected to hysteresis analysis confirms this conclusion. Stepwise thermal demagnetization of the SIRM indicates that >98 per cent of the remanence is unblocked up to temperatures of 550–600 °C (Fig. 3b). Both of these observations are indicative of a mineralogy dominated by magnetite.

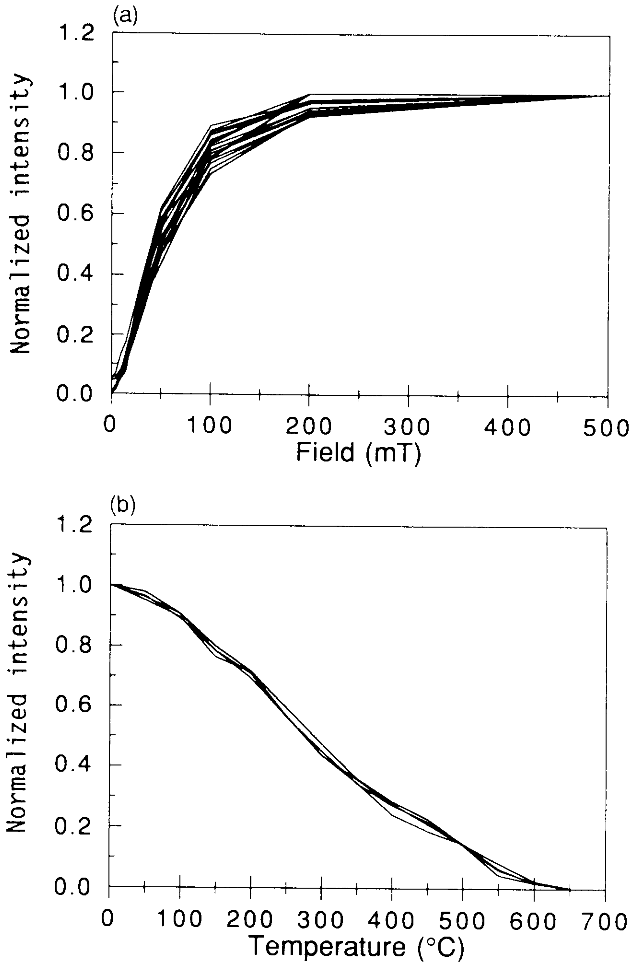

Figure 3. (a) Representative normalized SIRM acquisition curves; (b) typical intensity decay curves during stepwise thermal demagnetization for samples from Lake Chewaucan.

Magnetite grain size in bulk samples of sediment can be estimated by two approaches. The first uses plots of susceptibility to ARM acquisition (χ_{ARM}) versus magnetic susceptibility (χ). As outlined by King *et al.* (1983), changes in slope on this type of plot indicate changes in magnetite grain size, while changes along a line of constant slope are indicative of variations in the concentration of magnetite. For the Lake Chewaucan samples, the variation in magnetite grain size is small (Fig. 4a). The second approach involves measurement of four high field hysteresis parameters, M_{rs}, M_s, B_c and B_{cr}, as outlined by Day, Fuller & Schmidt (1977). M_{rs} is the saturation remanence, M_s is the saturation magnetization, B_c is the coercive force, or the field required to reduce M_s to zero, and B_{cr} is the coercivity of remanence, or the back-field required to reduce M_{rs} to zero. The method of Day *et al.* (1977) enables a more quantitative estimate of grain size. For the Chewaucan samples (Fig. 4b), the values of M_{rs}/M_s and B_{cr}/B_c fall within the PSD field, well within the acceptable range of

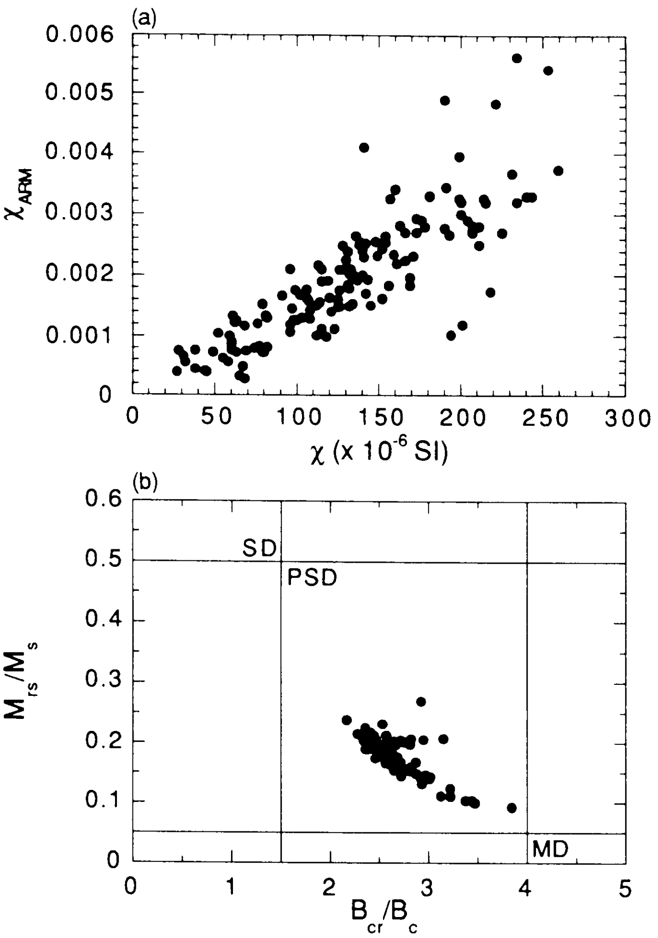

Figure 4. Variation of magnetite grain size from Lake Chewaucan sediments. (a) χ_{ARM} versus χ (*cf.* King *et al.* 1983). Variation in magnetite grain size appears to be small (see text); (b) M_{rs}/M_s versus B_{cr}/B_c (*cf.* Day *et al.* 1977). Values of M_{rs}/M_s versus B_{cr}/B_c fall within the PSD size range.

values cited by King *et al.* (1983) for 1–15 μm magnetite grains.

Stratigraphic variation in magnetite grain size at Lake Chewaucan can be estimated by plotting various proxies for grain size, such as χ_{ARM}/χ, SIRM/χ, M_{rs}/M_s, and B_{cr}/B_c versus depth (Fig. 5). In Fig. 5, SIRM/χ is much less variable in the Lake Chewaucan sediments than χ_{ARM}/χ. Because SIRM is a saturation parameter, it is relatively insensitive to changes in inducing field and it therefore varies as a simple function of magnetic mineral concentration. The efficiency of ARM acquisition varies significantly with concentration. This can give rise to non-linear ARM acquisition even with relatively small changes in magnetic mineral concentration (Tauxe 1993). We therefore believe that the SIRM/χ curve is a more accurate proxy for grain size in the Lake Chewaucan sequence. It is evident in Fig. 5 that the overall variation in grain size is small. Grain-size variations are evident as small-scale, quasi-cycle fluctuations in hysteresis parameters (Fig. 5c), which may be due to environmental changes such as lake level variation. However, these fluctuations are all within the acceptable range of grain sizes suggested by King *et al.* (1983).

Variations in χ_{ARM}/χ and M_{rs} indicate that the maximum magnetite concentration is approximately 12 times the minimum concentration although concentration variation for the majority of the samples is within a factor of six. Our detailed rock magnetic studies therefore indicate that the remanence-bearing grains in the Lake Chewaucan succession meet the criteria proposed by King *et al.* (1983), as well as the more restrictive criteria of Tauxe (1993), and are sufficiently uniform to be applicable to relative palaeointensity studies.

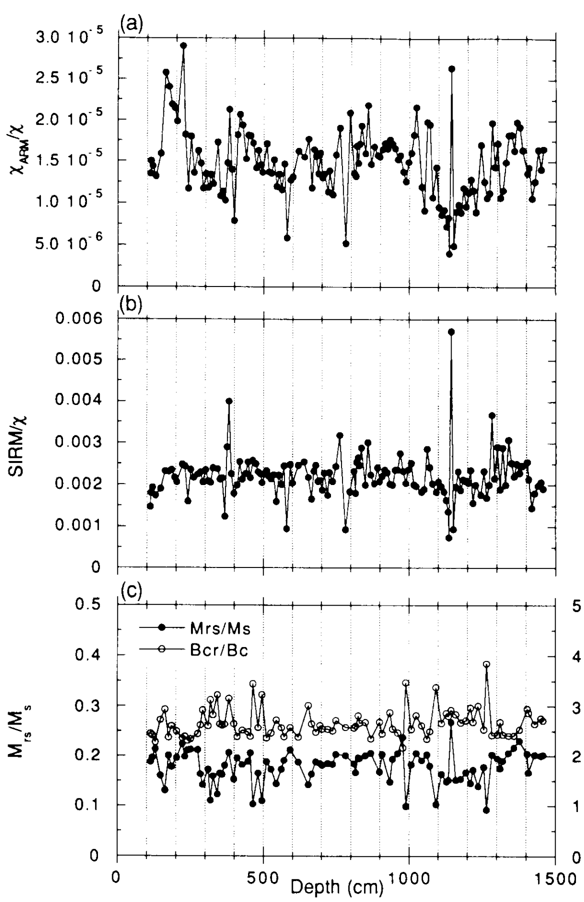

Figure 5. Stratigraphic variation of magnetite grain size in the Lake Chewaucan succession, as indicated by (a) χ_{ARM}/χ, (b) SIRM/χ (c) M_{rs}/M_s and B_{cr}/B_c.

5 NORMALIZED REMANENCE

In order to derive relative palaeointensity records from sediments, it is important that the laboratory-induced remanent magnetization be a close analogue of the NRM. Typical examples of the NRM, ARM and SIRM behaviour of Lake Chewaucan samples during AF demagnetization are shown in Fig. 6. For these samples, it is clear that both the ARM and SIRM closely mimic the behaviour of the NRM; however, the SIRM curve is consistently more nearly parallel to the NRM curve. SIRM may therefore be the more appropriate laboratory-induced magnetization for palaeointensity normalization of these samples. This point is further illustrated by plotting SIRM versus NRM and ARM versus NRM for various samples (Fig. 7). It is clear that the

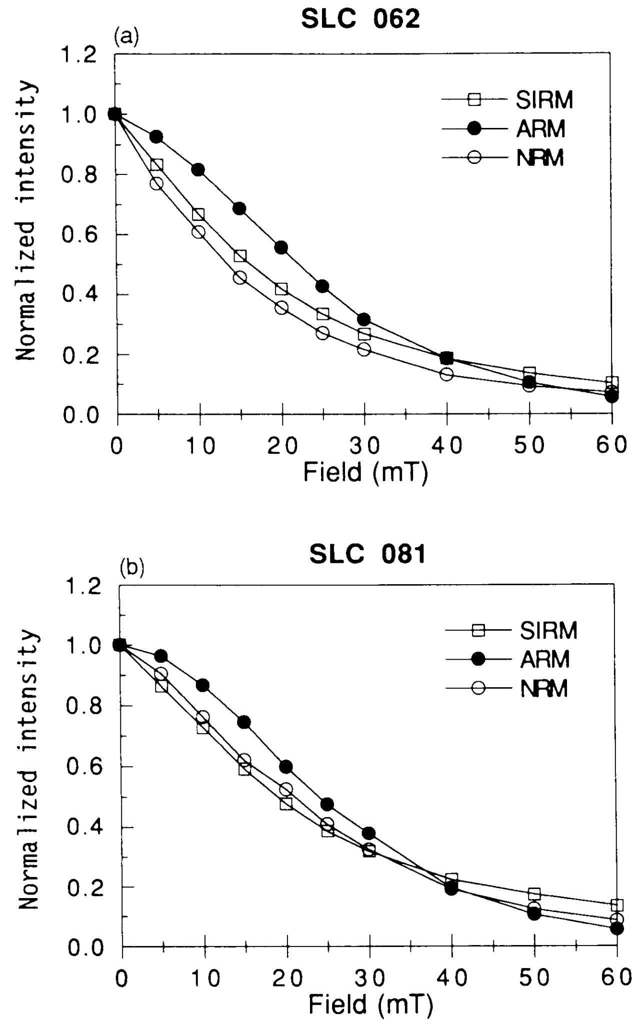

Figure 6. AF demagnetization of NRM, ARM and SIRM for two representative samples (SLC 062 and 081) from Lake Chewaucan. The correspondence between the SIRM and NRM curves indicates that SIRM may be the most appropriate laboratory-induced magnetization for relative palaeointensity normalization for these samples.

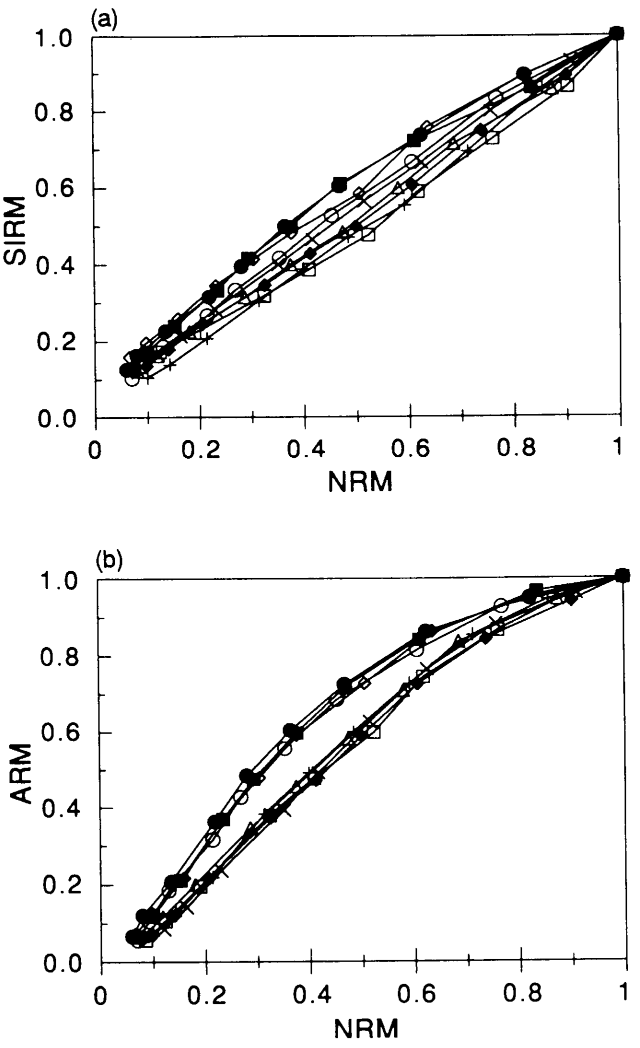

Figure 7. (a) SIRM versus NRM for representative samples from Lake Chewaucan after stepwise AF demagnetization; (b) ARM versus NRM for representative samples from Lake Chewaucan after stepwise AF demagnetization. Measurements were made at 0, 5, 10, 15, 20, 25, 30, 40, 50 and 60 mT for all samples. Both plots are normalized to the initial values for NRM, SIRM and ARM, respectively.

SIRM and NRM vary linearly with increasing demagnetization field, whereas ARM is distinctly non-linear with respect to NRM. This approach can be a useful tool for evaluating the most suitable normalization parameter in the early stages of such a study.

Some workers favour ARM over SIRM normalization because SIRM is a much stronger magnetization than NRM or ARM, and different parts of the grain-size distribution may be activated by SIRM compared to NRM or ARM (e.g. Levi & Banerjee 1976; Levi & Merrill 1978). Susceptibility normalization can also be problematic because paramagnetic minerals contribute to the susceptibility but not to the remanence.

Instead of relying solely on one normalization parameter, we have used three (ARM, SIRM and χ) in order to examine their relative efficiency. We have used the 25 mT demagnetization step for normalization of remanence intensity for each type of magnetization because, at this level, all secondary components are removed from the NRM, the AF decay curves are closely parallel (Fig. 6), and the remanence intensities are still relatively strong and well above the noise levels of the magnetometers used. All three parameters give remarkably similar results, with a one-to-one correspondence existing between almost all of the features on each curve (Fig. 8).

The greatest difference between the curves, and the greatest amount of variation in the relative amplitude of the peaks, occurs with the χ normalization (Fig. 8c). As noted above and as indicated by our results, variations in paramagnetic mineral content, and hence variable contribution to magnetic susceptibility, make χ the least useful normalization parameter.

Stratigraphic variations in magnetite particle size are within the acceptable range for relative palaeointensity studies, and these variations do not appear to affect the normalized intensity signal because major changes in

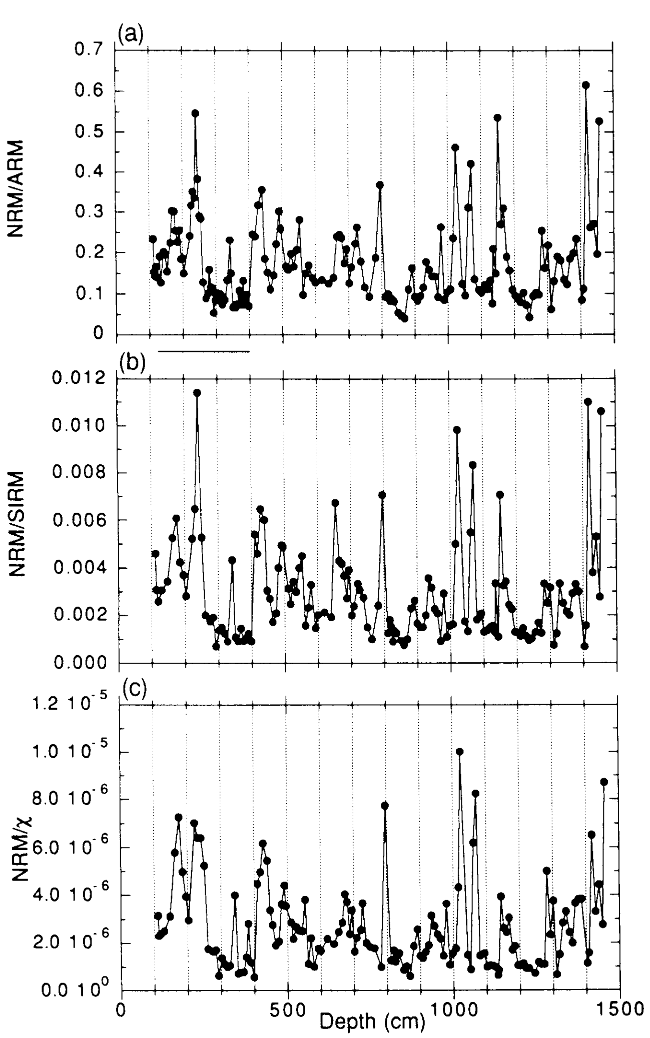

Figure 8. Results of three types of relative palaeointensity normalization from Lake Chewaucan sediments: (a) NRM/ARM; (b) NRM/SIRM; (c) NRM/χ. In each case the NRM is normalized by a laboratory-induced magnetization. The similarities between the curves suggest that each laboratory-induced magnetization provides a good approximation to the relative palaeointensity record. The solid line in (a) indicates the depth range shown in Fig. 9(b).

intensity (Fig. 8) do not correspond to any areas where grain-size variation is evident (Fig. 5). We therefore conclude that the large-scale variations in our normalized remanence record are due to relative geomagnetic palaeointensity fluctuations.

6 CORRELATION WITH OTHER RELATIVE PALAEOINTENSITY RECORDS

Because the Lake Chewaucan sediments appear to provide a record of relative palaeointensity, a comparison can be made with existing palaeointensity records. The records of Tric *et al.* (1992) and Meynadier *et al.* (1992) are currently the most extensive Late Pleistocene records of relative palaeointensity. We therefore use the records from these two studies as the basis for our comparison.

We believe that the Lake Chewaucan chronology is sufficiently reliable for comparison only for the upper part of the succession, that is, from approximately 65 to 105 kyr. As the basis for our chronology, we have used the available TL dates for the directly dated tephra layers as well as K/Ar dates from correlative tephra layers elsewhere (Table 1). Depths have been converted to ages by linear interpolation between these dates, assuming a constant rate of sedimentation. Our relative palaeointensity record, from 65 to 105 kyr, is shown in Fig. 9(b) below the detailed records (Fig. 9a) of Tric *et al.* (1992) and Meynadier *et al.* (1992). Although we believe that SIRM is the best normalizing parameter for the Lake Chewaucan sediments, we use the NRM/ARM record in Fig. 9(b) because we have more ARM data and because all of the normalization parameters yield comparable curves (Fig. 8).

The correspondence between our curve and those of Tric *et al.* (1992) and Meynadier *et al.* (1992) is reasonably good, with all records showing a short intensity maximum at about 70 kyr and a broader, stronger maximum at about 80 kyr, which is preceded by a rapid rise in intensity. There is, however, a temporal offset of about 5 kyr between features in our record and those of Tric *et al.* (1992; Fig. 9). This offset could easily result from the relative imprecision and large standard errors of our chronology (generally 10 per cent; *cf.* Fig. 2). The record of Meynadier *et al.* (1992) has a broad intensity low from 90 to 120 kyr, in agreement with our record. A prominent unconformity, which we correlate with the oxygen isotope stage 6/5 boundary (130 kyr; Martinson *et al.* 1987), truncates the Lake Chewaucan

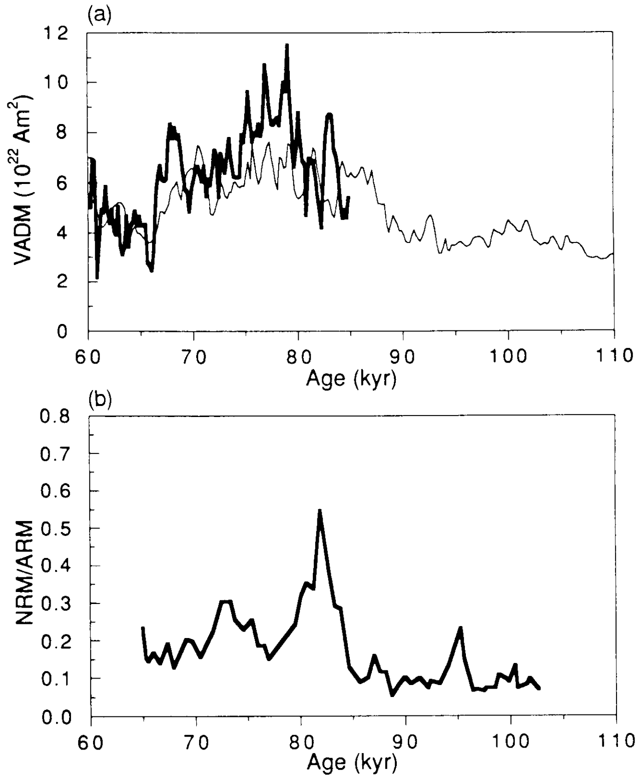

Figure 9. (a) Sedimentary palaeointensity records from 60 to 110 kyr from the Mediterranean Sea (Tric *et al.* 1992; bold line) and from the Somali Basin (Meynadier *et al.* 1992; thin line). Both records have been converted into virtual axial dipole moments by the original authors. (b) The NRM/ARM record from the same interval for Lake Chewaucan.

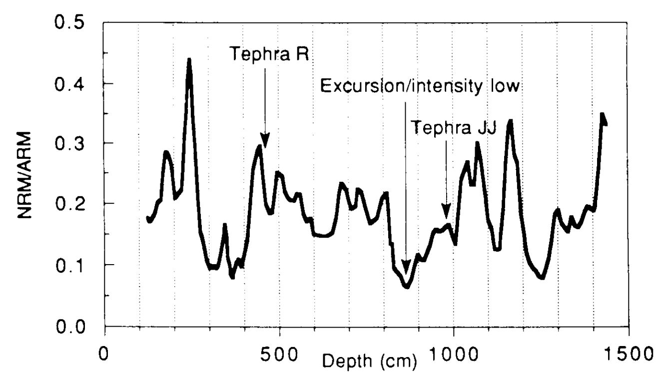

Figure 10. NRM/ARM for the Lake Chewaucan record, smoothed with a seven-point Gaussian filter (coefficients: 0.037, 0.23, 0.693, 1.0, 0.693, 0.23, 0.037). The position of the geomagnetic excursion of Negrini *et al.* (1988) and the palaeointensity low of this study is shown, along with the positions of tephra R and tephra JJ.

record just below tephra N which has a TL date of 102 kyr (Table 1; Berger 1991). If this prominent unconformity is correlative with the stage 6/5 boundary, then the marked drop in palaeointensity from 140 to 130 kyr observed by Meynadier *et al.* (1992) may also be recorded above tephra R at Lake Chewaucan (Fig. 10). Given the uncertainty in the dating of the curves in Fig. 9, we find the correspondence between features of our record and the deep-sea records encouraging.

Negrini *et al.* (1988) reported a geomagnetic excursion at a depth of ~8.5 m in the lower part of the Lake Chewaucan succession studied here. In Fig. 10 we show our relative palaeointensity record which is smoothed in order to show only the major features of the curve. In this record, the lowest palaeointensity occurs at the same stratigraphic position as the geomagnetic excursion reported by Negrini *et al.* (1988; Fig. 10). Liddicoat & Bailey (1986) reported a similar geomagnetic excursion, in the Long Valley region of California, with the same stratigraphic relationship to tephra JJ. Further tephrostratigraphic evidence, and strikingly similar directional signatures (Negrini *et al.*, submitted; Herrero-Bervera *et al.*, in preparation), suggest that these excursion records are correlative to the Pringle Falls, Oregon, excursion reported by Herrero-Bervera *et al.* (1989). The large standard errors associated with the TL dates of Berger (1991) make it difficult to accurately constrain the age of the excursion with existing data. However, the TL date on tephra R (165 ± 19 kyr; Table 1), which lies above this excursion at Lake Chewaucan (Fig. 10), provides strong evidence that it is certainly older than the Blake event (115–120 kyr; *cf.* Tric *et al.* 1991).

Despite the uncertainties in the dating, our results are consistent with the observation that low field intensities dominate during geomagnetic reversals (e.g. Opdyke *et al.* 1973; Bogue & Coe 1984; Prévot *et al.* 1985; Valet, Laj & Langereis 1988) and geomagnetic excursions (e.g. Marshall, Chauvin & Bonhommet 1988; Roperch, Bonhommet & Levi 1988).

We have also obtained a normalized intensity record from the vicinity of the Mono Lake excursion in the upper part of the Lake Chewaucan succession. Our record of the Mono Lake excursion (Negrini *et al.* 1984) is shorter than the directional record of Lund *et al.* (1988) due to an unconformity above the excursion (Negrini & Davis 1992). The sediments above the excursion are coarser grained than those that record the excursion, due to low lake levels (*cf.* Negrini & Davis 1992), and are not sufficiently uniform for palaeointensity study. In the original study of the Mono Lake excursion by Negrini *et al.* (1984), relative palaeointensity variations were not given because of the observed differences between the NRM and ARM demagnetization curves (Figs 6 and 7). However, in this study we have shown that even though the ARM may not be as good an analogue for the NRM as the SIRM, it still gives a normalized intensity signal that is comparable with the SIRM normalization. We have normalized the NRM data from the Mono Lake excursion with ARM, SIRM and χ, and find that, as with the older part of the succession, all three normalizations give comparable results.

In Fig. 11(a), we show the χ normalization because we have more χ data. We also show the palaeointensity record of Tric *et al.* (1992) and Meynadier *et al.* (1992) for the period 36–20 kyr (Fig. 11b). In the Tric *et al.* (1992) record, there is a broad high with a low superposed from 30–27 kyr.

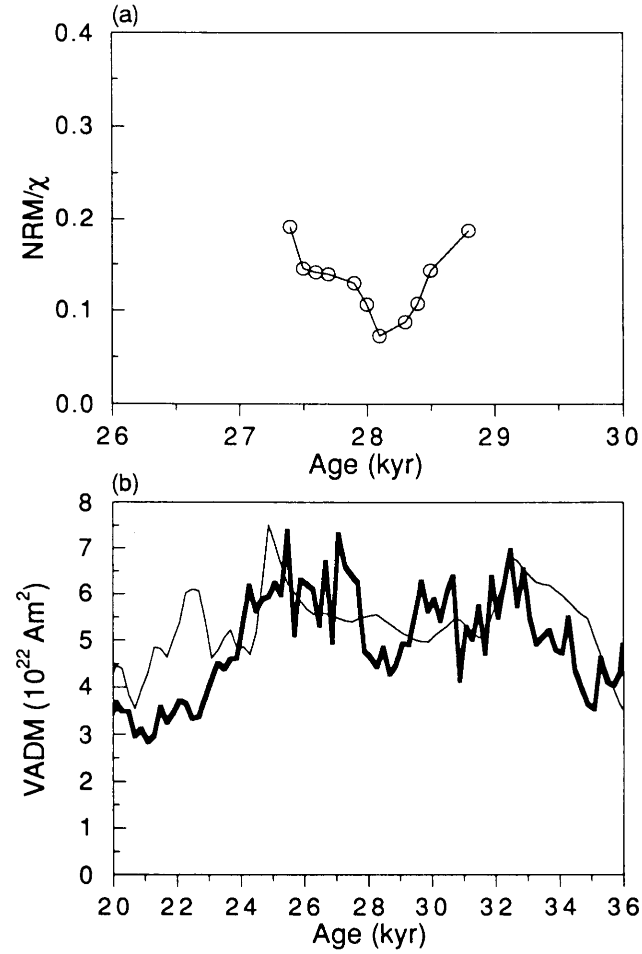

Figure 11. Comparison of relative palaeointensity record from (a) Mono Lake excursion, recorded at Lake Chewaucan (NRM/$\chi \times 10^{-5}$ SI units), with (b) the records of Tric *et al.* (1992) from the Mediterranean Sea (bold line) and Meynadier *et al.* (1992) from the Somali Basin (thin line). The records in (b) were transformed to virtual axial dipole moments by the original authors.

The record of Meynadier *et al.* (1992) is much smoother than that of Tric *et al.* (1992), and it shows a broader low from 33 to 25 kyr superposed on the same high. No anomalous palaeomagnetic directions were reported by Tric *et al.* (1992) or Meynadier *et al.* (1992) in this interval. This result suggests that although the directional manifestation of a geomagnetic excursion may differ spatially, the associated low field intensities may be a useful tool for large-scale, inter-regional correlation. The observation of low field intensities associated with the two geomagnetic excursions in the Lake Chewaucan succession reinforce the importance of obtaining both directional and palaeointensity records of geomagnetic reversals and excursions.

7 CONCLUSIONS

We have found good general agreement in the relative palaeointensity records from a western North American lacustrine record and from two marine records from the Mediterranean Sea and the Somali Basin during the time intervals from ~140 to 65 kyr, and from 29 to 27 kyr. Correlation of relative palaeointensity records from widespread regions, even globally, may be possible as further high-quality records are obtained (such studies *must* be supported by detailed rock magnetic characterization). Valet & Meynadier (1992) have even suggested that it may be possible to develop a new global stratigraphy based on palaeointensity of the geomagnetic field. A new global stratigraphy will be particularly useful in providing a common time-scale for correlating marine and continental records of palaeoclimate, which is presently limited by imprecisions inherent in the age determinations of continental sediments.

Our results also provide further evidence that low geomagnetic field intensities dominate during geomagnetic excursions. Thus, relative palaeointensity records, in conjunction with conventional studies of palaeomagnetic directions, have the potential to lead to a better understanding of the long-term vector behaviour of the geomagnetic field.

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