
DOI: http://doi.org/doi:10.1111/j.1365-246X.2006.03263.x
Palaeomagnetism of greigite bearing sediments from the Dead Sea, Israel

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Accepted 2006 October 12. Received 2006 October 12; in original form 2005 September 27

SUMMARY
Detailed magnetostratigraphic investigations were carried out on five sediment cores obtained from three sites on the northwestern shore of the Dead Sea, Israel. The sediments are heterogeneous, composed of fine clastic layers alternating with authigenic aragonite. The magnetic remanence is carried by Ti-magnetite and greigite, reflecting a detrital as well as an authigenic phase. Directions of the characteristic remanent magnetization were obtained from stepwise AF-demagnetization of the natural remanent magnetization and subsequent principle component analysis. The acquisition of a gyroremanent magnetization in a noticeable amount of the samples could be linked to core disturbances. The depth–age models, based on AMS 14C datings on plant remains, yielded an age of 10.04 ka cal. BP for the base of the longest profile from site Ein Gedi, a continuous lacustrine sediment sequence. The profiles from the two other sites representing near shore sedimentation are interrupted by a hiatus in sedimentation. Comparison of the inclination and declination records from all Dead Sea cores and nearby sited lakes suggest that low frequency variations documented in the Dead Sea sediments are likely of geomagnetic origin.

Key words: Dead Sea, Israel, greigite, lacustrine sediments, Late Holocene, palaeomagnetism.

INTRODUCTION
During the last decades palaeomagnetic investigations became a common part of lake sediment studies throughout the world and now geomagnetic palaeosecular variation (PSV) records are known as a powerful tool for correlation and dating of sediment sequences (e.g. Mothersill 1996; Saarinen 1998; Ali et al. 1999; Brachfeld et al. 2000; Gogorza et al. 2000; Kotilainen et al. 2000; Nourgaliev et al. 2000a). Since the directions of the geomagnetic field vector undergo not only temporal but also spatial variations, this tool, sensu strictu, can be used accurately only within a limited region. One of these regions is Europe, where the synchronicity of PSV records from Great Britain in the West, Poland in the East, Finland to the North and the Mediterranean Sea in the South, have been proven in several studies (e.g. Thompson & Turner 1979; Creer 1985; Mothersill 1996; Saarinen 1998; Ali et al. 1999; Brandt et al. 1999; Gogorza et al. 2000; Kotilainen et al. 2000; Nourgaliev et al. 2000b; Snowball & Sandgren 2002). However there is a lack of reference data for North Africa, the Near East and the Middle East. A well-established PSV mastercurve for these areas would not only allow for palaeomagnetic dating but could contribute to a detailed analysis of the variations of the geomagnetic field vector with latitude and longitude (Korte et al. 2005). Investigations in the Near East are in progress and yielded archaemagnetic PSV records from Iraq (Hammo-Yass 1987/1988) and Israel (Marion et al. 1994; Katari et al. 1995; Sternberg et al. 1999), as well as lacustrine records from Lake Kinneret, Israel (Thompson et al. 1985) and Birkat Ram, Golan Heights, Israel (Frank et al. 2002) so far. Further magnetostratigraphic investigations were carried out on sediment profiles from three different sites on the western shore of the northern basin of the Dead Sea, Israel.

SITE DESCRIPTION AND MATERIAL
The Dead Sea is located within a NNE–SSW stretching pull-apart basin, a part of the Red Sea-rift zone. The formation of the basin started in the Upper Miocene about 15 Ma ago and is still active with ongoing seismic activity (Garfункel & Ben-Avraham 1996). The Dead Sea basin is flanked by graben walls on both sides build up of carbonatic rocks. Most of the clastic material brought into the lake is transported by temporarily active rivers at the western shore of the lake during spring time and by the Jordan River in the North (Fig. 1). Both fluviatile systems are forming fan delta structures at their front. The detrital magnetic material mainly originates from basaltic rocks of quaternary age and the sediments of Lake Lisan (the palaeo-Dead Sea) that were passed through by the Jordan river on its way to the Dead Sea.

During a drilling campain in 1997 a total of nine cores were recovered from four different sites at the western shore of the Northern Basin of the Dead Sea with an Usinger piston corer (a modified Livingston piston corer) (Usinger 1991). The individual core
sections are up to 2 m long with a diameter of 80 mm in the upper part and 55 mm below. They are not overlapping and material might be missing at the core boundaries (Migowski 2002). Therefore, the recovery of two parallel cores with different starting depth is requested to obtain a complete profile. Two of the sites chosen for drilling, Ze’elim and Hever, are located within fan deltas whereas Ein Feshkha and Ein Gedi represent near shore and lacustrine environments, respectively (Fig. 1). Although all coring sites were near the recent shore line in 1997, the top of the cores have different heights below mean sea level (mbsl) due to topographical variations of the Dead Sea shore.

The cores from the sites Ein Feshkha, Ein Gedi and Ze’elim were subjected to detailed palaeomagnetic investigations. An overview about the core recovery is given in Table 1. The sediments are composed of alternating fine layers of detrital material and authigenic aragonite, probably reflecting seasonal variations (Fig. 2). Superimposed on this layering is the change between sequences with a comparably higher content of clastic material and those which are dominated by aragonite (Migowski 2002). The laminated sequences are interrupted by intercalated silt, sand or salt layers with thickness up to 2 m. Detailed rock magnetic investigations performed on three of the cores presented here, DSF-B, DSEn-C/A and DSZ-A, revealed that Ti-magnetite and greigite are the carriers of the magnetic remanence (Frank et al. 2007.). The greigite is present in form of thin dark brownish green layers intercalated into the sediment sequences throughout most of the profiles, indicating synsedimentary formation of greigite.

**METHODS**

Continuous high resolution logs of magnetic susceptibility were measured with a Bartington MS2E spot reading sensor in steps of 1 mm on the split halves of the cores directly after opening of the core tubes at the GeoForschungsZentrum Potsdam (GFZ). The sensor is integrated in an automated core logging system build at the Laboratory for Palaeo- and Rockmagnetism in Potsdam. The core halves were than sealed in polyethene sheets and cold stored until subsampling started 2–5 yr later. Due to a high content of salt in the sediments the cores were still wet and not visibly altered except for an oxidation of the sediment surface. Subsampling was performed

**Table 1.** Overview about the sediment cores available from the four Dead sea sites including site coordinates, core length investigated and number of paleomagnetic samples taken.

<table>
<thead>
<tr>
<th>Site</th>
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</thead>
<tbody>
<tr>
<td>Ein Feshkha</td>
<td>31°42’N, 35°27’E</td>
<td>DSF-A</td>
<td>2.62</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td></td>
<td></td>
<td>DSF-A1</td>
<td>0.65</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td></td>
<td></td>
<td>DSF-B</td>
<td>16.59</td>
<td>16.59</td>
<td>671</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Ein Gedi</td>
<td>31°30’N, 35°24’E</td>
<td>DSEn-C/A</td>
<td>20.05</td>
<td>19.88</td>
<td>741</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td></td>
<td></td>
<td>DSEn-B</td>
<td>18.96</td>
<td>18.96</td>
<td>724</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Ze’elim</td>
<td>31°21’N, 35°24’E</td>
<td>DSZ-A</td>
<td>11.73</td>
<td>11.02</td>
<td>310</td>
<td>Intercalated sand layers</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>DSZ-B</td>
<td>12.29</td>
<td>9.84</td>
<td>319</td>
<td>Intercalated sand layers</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>DSZ-C</td>
<td>6.83</td>
<td>–</td>
<td>–</td>
<td>Mostly sand layers</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hever</td>
<td>31°24’N, 35°23’E</td>
<td>DSH-A</td>
<td>4.47</td>
<td>–</td>
<td>–</td>
<td>–</td>
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</table>

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with cubic plastic boxes (20 × 20 × 15 mm) that were pushed into the cleaned surface of the core halves in intervals of 25 mm. Although there are massive disturbances in the sediment structure of the cores DSEn-C/A and DSEn-B from site Ein Gedi below 12.4 and 10.4 m, respectively, affecting more than 2 m of the sediment sequence, both cores were completely subsampled. For the determination of the magnetic bulk susceptibility $\kappa_{LF}$ of the subsamples, a Kappabridge KLY-3S (AGICO Brno) with an operating frequency of 875 Hz was used.

Directions and intensities of the natural remanent magnetization (NRM) were measured with a fully automated long-core 2G-755-SRM (Superconducting Rock Magnetometer). All samples were subjected to stepwise alternating field (AF) demagnetization, using the integrated in-line three-axis AF demagnetizer. Demagnetization was carried out in 10 steps up to 100 mT. The directions of the characteristic remanent magnetization (ChRM) were derived from principle component analysis (Kirschvink 1980) of the results of successive demagnetization steps. The demagnetization steps from 0 (NRM) to 10 mT as well as those from 65 to 100 mT were always excluded from the vector analysis. All cores were obtained without azimuthal orientation, therefore, the mean ChRM-declination value of the top core section was set to zero. Afterwards the remaining core sections were also rotated, fitting the ends of the individual declination logs together.

**Correlation and Chronology**

The intrasite correlation between cores is based on the high resolution logs of magnetic susceptibility (Fig. 3) as well as on macroscopic identification of marker layers. Continuous composite profiles were obtained for sites Ein Gedi and Ze’elim combining cores DSEn-C/A and DSE-B and DSZ-A and DSZ-B, respectively. The results of the single cores will be presented versus composite depth. Correlation between the cores obtained from different sites is not possible by sedimentological means (Figs 2 and 3).

The chronologies for the profiles from Ein Gedi, Ein Feskha and Ze’elim are based on Accelerator Mass Spectrometry (AMS) $^{14}$C datings on terrestrial plant remains (Migowski et al. 2004) (Fig. 4).
In contrast to the depth/age model by Migowski et al. (2004) where mean sedimentation rates were calculated from all $^{14}$C ages available, the chronologies presented here shows variable sedimentation rates incorporating all reliable $^{14}$C ages. Data from sediment sequences that are either deformed or could not be assigned to a composite depth unambiguously were rejected. Additional age information for the uppermost 650 cm of the Ze’elim cores is available by correlation with the neighbouring outcrop in the Ze’elim gully (Ken-Tor et al. 2001; Migowski 2002). Based on the $^{14}$C-ages a 3 ka long hiatus was identified within the sediment sequence of Ein Feshkha at about 620 cm depth (Fig. 4). The sediment sequence from Ze’elim is interrupted by one hiatus in the time interval from 3 to 8 ka. The hiatuses are interpreted as a result of erosion and reworking during low stands of the Dead Sea water level, associated with dry periods (Frumkin & Elitzur 2002). The lacustrine sediments from the Ein Gedi profile are not affected, but show a continuous record of sedimentation for the last 10 ka with a mean sedimentation rate of 2 mm a$^{-1}$. The profiles from Ein Feshkha and Ze’elim span the last 7.5 and 9 ka, respectively (Fig. 4).

RESULTS

Stability of the NRM and occurrence of a gyroremanent magnetization (GRM)

The results from progressive AF demagnetization are shown by some typical demagnetization curves in Fig. 5. Except for a viscous overprint, acquired during the cold storage of the core halves, there

Figure 3. Intrasite correlation of cores DSZ-A and DSZ-B from Ze’elim and between DSEn-C/A and DSEn-B from Ein Gedi based on the logs of high-resolution magnetic susceptibility measurements. Some correlation levels are marked by dashed lines. The record from core DSF-B from Ein Feshkha is shown at the right side. Correlation between the cores from the different sites is not possible by sedimentological means.
is no evidence for a secondary component masking the primary direction as it is shown in the orthogonal demagnetization (Zijderveld) diagrams (Figs 5a–f). Further steps revealed a high stability of the magnetization directions, until some samples acquired a gyroremanent magnetization (GRM) from fields of 65 mT on (Figs 5e and f). The vector end points did not move towards the origin of the coordinate system. The acquisition of a GRM during demagnetization is a behaviour typical for sediments containing greigite (Snowball 1997; Sagnotti & Winkler 1999). However, sediments that are disturbed by the piston’s suction during coring, acquire also a GRM during demagnetization (Fig. 5f). This conclusion was derived from a comparison of stratigraphic position of the samples that acquired a GRM with the core photographs. The deformed sediment sequences could also be identified by calculating the difference between $J_{NRM}$ after demagnetization at 65 mT ($J_{NRM65}$) and 100 mT ($J_{NRM100}$) (Fig. 6). Here the rate of GRM acquisition is in the range of more than 100 per cent. In extremely deformed sequences such as between 1240 and 1320 cm depth in core DSEn-A, the GRM acquisition reaches up to more than 600 per cent relative to the NRM intensity after AF-demagnetization at 65 mT (Fig. 6). Unfortunately, it is not possible to distinguish between Greigite bearing samples and those from deformed sections just by the calculated GRM acquisition rate. This can be seen in the comparison of the records of $J_{SBM}/k_{LF}$, a greigite indicative rock magnetic parameter (see Frank et al. 2007) and the GRM acquisition rate (Fig. 6). Samples with a high amount of greigite ($J_{SBM}/k_{LF} > 30 \text{ kA m}^{-1}$) did not necessarily have high GRM acquisition rates and they did not necessarily have to catch a GRM at all. The magnetic carriers in the sample shown in Fig. 5(d) show a distinct decrease in magnetization between 300 and 400 °C (Frank et al. 2007), indicative for greigite (Snowball & Thompson 1990; Snowball 1991; Jelinowska et al. 1995; Roberts 1995) but there is obviously no GRM acquired during demagnetization (Fig. 6).

Sagnotti et al. (2005) indicated that sediments from the Ross Sea, Antarctica with a ratio of $J_{SBM}/k_{LF} < 20 \text{ kA m}^{-1}$ will not show significant gyro magnetic effects during AF demagnetization. This limit is obviously higher, around 40 kA m$^{-1}$ in the sediments from the Dead Sea, probably due to the more complex composition of the sediment. Analyzing the anisotropy of the magnetic susceptibility would probably allow for some additional answers to this problem, but this data is not available.

Although it is not possible to identify the greigite bearing samples by means of the calculated GRM acquisition rate, which would be a rather quick method, the samples from deformed sediment sequences could easily be detected: they are (1) often linked to core boundaries and (2) core deformation normally affects a whole series of samples not only single ones (Fig. 6). Because of the clear lamination of the Dead Sea sediments the deformed sequences could also be identified by eye very easily, so there is no real need for this method in here, but it could be a useful tool analyzing homogeneous sediments.

NRM-intensity

The variations in the NRM-intensity measured for the five cores from the Dead Sea are similar to those in the magnetic susceptibility and thus reflect the heterogeneous composition of the sediments. The lowest intensities were obtained from sediment sequences with a high content of aragonite, building up the uppermost 10 m of the two Ein-Gedi cores (Figs 7b and c). This profile reflects the lacustrine sediment sequences deposited nearly continuously during the last 10 ka whereas the cores from Ein Feshkha (Fig. 7a) and Ze’elim (Figs 7d and e) reflect near shore sedimentation, the latter in alternation with fan deposits (Fig. 1). The highest NRM-intensities were measured for samples from DSZ-A and DSZ-B in the depth interval from 250 to 400 cm, corresponding to the time interval...
Figure 5. Alternating field demagnetization results of six samples, Fig. 5(a)–(f), from Core DSF-B, with different sediment composition. Sampling depth, intensity of the natural remanent magnetization ($J_{\text{NRM}}$), and median destructive field of the NRM (MDF$_{\text{NRM}}$) of each subsample is shown in the plots of the demagnetization curves. The related orthogonal projections (Zijderveld diagrams) of the stepwise demagnetized samples are always shown below the demagnetization curves. Demagnetization steps run from 0 to 100 mT. Units are $10^{-3}$ Am$^{-1}$. The NRM measurement in each Zijderveld diagram is marked with an asterisk. Closed symbols denote X plotted versus Y, and open symbols denote Z plotted versus H ($H^2 = X^2 + Y^2$).
1.5–2 ka (Figs 2, 7d and e). In the profile from Ein Feshkha there is also a distinct increase in intensity in this time interval (200–400 cm) as well as between 1200 and 1500 cm depth, that is 6.5–7.5 ka (Fig. 7a). The latter is also visible in core DSEn-C/A from Ein Gedi (Figs 2 and 7b), indicating a distinct change either in the level of the Dead Sea, in the minerogenic components transported into the lake, or in the geochemical composition of the waterbody. The rock magnetic investigations revealed, that the sediment sequences with high magnetic susceptibilities are those with highest content of greigite (Frank et al. 2007). This could also be seen in the records of the median destructive fields of the NRM (MDFNRM) which varies between 10 and 50 mT in all cores (Figs 7a–e). The lowest values (10–24 mT), that are typical for (Ti-)magnetite in the MD- to PSD range (Hartstra 1982), are found in the sediment sequences with low magnetic susceptibilities and low NRM intensities, the higher MDFs, typical for the greigite bearing samples (Frank et al. 2007), are linked to intervals with increased NRM intensities.

A quite more interesting aspect are the similarities between the records of MDFNRM and the ratio of NRM-intensity after demagnetization at 20 mT (\(J_{\text{NRM}}(20 \text{ mT})\)) normalized by \(\kappa_{\text{LF}}\)
Figure 7. Records of $\kappa_{LF}$ and $J_{NRM}$, ratio of NRM-intensity after demagnetization at 20 mT ($J_{NRM}(20 \text{ mT})$) normalized by $\kappa_{LF}$, median destructive field of the NRM (MD$F_{NRM}$), inclination and declination of the characteristic remanent magnetization ($I_{ChRM}$, $D_{ChRM}$) as well as the GRM acquisition rate for (a) core DSF-B from site Ein Feshkha, (b) core DSEn-C/A and (c) core DSEn-B from site Ein Gedi and (d) core DSZ-A and (e) Core DSZ-B from Ze’elim. The grey shaded areas mark core sections with major sediment deformations.

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(Figs 7a–c), an estimation of the relative palaeointensity. High values of $MDF_{NRM}$ correspond to high relative palaeointensities. This relation is not an uncommon observation in sediments with a monomineralic composition of the magnetic carrier fraction and normally interpreted as an influence of the magnetic grain size (Brachfeld & Banerjee 2000), or of the degree of alignment of the magnetic particles (Frank et al. 2003). Both explanations are invalid in case of the Dead Sea sediments with its mixture of magnetic minerals.
Here it must be interpreted as a parameter also linked to the ratio of the amount of the different magnetic carrier minerals, Ti-magnetite and greigite. The latter carry a chemical remanent magnetization (CRM), the alignment of the magnetic particles is much more effective compared to the detrital remanent magnetization of the Ti-magnetite. Thus the MDF\textsubscript{NRM} and the NRM-intensities are generally higher for greigite bearing samples, masking the variations in the intensity of the geomagnetic field as recorded by the Ti-magnetites.

Directions

The ChRM-inclination and -declination records obtained for the five cores investigated, are shown in Figs 7(a)–(e), together with the GRM-acquisition rates calculated for each sample. The directions of samples taken from sections with high GRM acquisition rates, mostly identified as distorted sediment sequences, deviate clearly from the mean inclination and declination values measured above and below. The only exceptions from this rule are the samples from the cores DSZ-A (Fig. 7d) and DSZ-B (Fig. 7e). The rate of GRM-acquisition is not higher than 380 per cent and below 200 per cent, respectively, and the deformation of the sediment, clearly visible in the core photographs, did not affect the directional records significantly. In the three other cores, DSF-B (Fig. 7a), DSE\textsubscript{n}-C/A (Fig. 7b) and DSE\textsubscript{n}-B (Fig. 7c) at least one directional parameter is affected in case of core disturbances. The reason for this different behaviour are probably based on the followings:

(i) On the composition of the sediment.
(ii) On the rate and form of the deformation.
(iii) In how the core was split along the \(z\)-axis with respect to the \(x\)- and \(y\)-axis of the field vector.

All the samples from these disturbed core sections, except for those from the Ze'elim cores, as well as all outliers linked to core boundaries, lithogenic fragments etc., were excluded from further investigations and the cleaned inclination and declination records are shown in Figs 8(b) and (c). There is no hint, if and to which extent, the recorded variations in ChRM-inclination and -declination are affected by the secondary formation of sulphidic ferromagnetic minerals. The amount in greigite is highest in the sections with the highest magnetic concentration as for example between 200 and 400 cm in core DSF-B (Fig. 7a), as was revealed from the detailed rock magnetic investigations carried out on cores DSF-B, DSE\textsubscript{n}-C/A and DSZ-A (Frank et al. 2007). However, there is nothing within the directional records that could be attributed to a time-delayed magnetic overprint such as a smoothing of the records. So, it must be assumed that it is not possible to separate the magnetic signal carried by Ti-magnetite on one hand and greigite on the other hand, a mixed signal throughout the whole profiles must be expected.

The amplitudes in the variations of the ChRM inclination records obtained from DSF-B, DSE\textsubscript{n}-C/A, DSE\textsubscript{n}-B, DSZ-A and DSZ-B are in the range of up to \(\pm20^\circ\) around a mean value of \(42^\circ\) (Figs 8a and b). The expected dipole inclinations are \(50.6^\circ\) for Ze'elim, \(50.8^\circ\) for Ein Gedi and \(51^\circ\) for Ein Feshkha. The range and morphology of the inclination variations in all five records corresponds to known secular variations of the Earth’s magnetic field although the mean value is 8\(^\circ\) lower than the expected dipole inclination for the Dead Sea region (Figs 8a to c). This so-called inclination error is a common observation in sediments; the reasons for this are manifold. In the sediments from the Dead Sea it could be assumed, that the high sedimentation rate, the composition and grain size of the sediment matrix as well as the laminated structure contribute to the inclination shallowing. There is no indication that the samples containing

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higher amounts of greigite relative to Ti-magnetite are not affected by the inclination shallowing as would be expected for a CRM.

There is an overall agreement in the shape and the succession of the inclination variations obtained from both Ein Gedi core (Fig. 8b). There is less agreement between the ChRM-inclination records from DSZ-A and DSZ-B (Fig. 8c), although the general trend is the same. Both records seem to be smoothed compared to those from Ein Feshkha and Ein Gedi which is a result of the different sedimentation rates. The time-interval covered by the uppermost 600 cm in DSZ-A and DSZ-B corresponds to only 350 cm in DSEn-C/A (Figs 8c and b). The ChRM-declination record shows variations with amplitudes of ±40° and a pronounced westerly swing between 540 and 440 cm depth in the record from DSF-B (Fig. 8a).

Intralake comparison

In order to obtain a better overview about the long term variations in the ChRM-declination as well as in the ChRM-inclination records all 10 curves were transformed into time series following the age models presented in Fig. 4. The low frequency variations in the records are marked by a polynomial fit of degree 10, making the identification of similar trends easier, but slightly masking the real variations. While the lowermost part of the DSF-B record is long enough to correlate the inclination and declination feature to those in the Ein Gedi records, the time interval covered by the lowermost 2 m of the two Ze’elim cores is too short for any kind of correlation despite the independent chronology. Therefore, this section will not be shown in the following figures.

There are some similarities between the fitted ChRM-inclination and declination records with a minimum in inclination between 2 and 1.8 ka and a subsequent relative inclination low at 1.2 ka visible in all five cores (Fig. 9a). Unfortunately, the profiles of Ein Feshkha and Ze’elim are interrupted by sand layers which did not yield reliable directions, and there are hiatuses in sedimentation, limiting the comparability of the records (Fig. 2). The comparison of the declination records (Fig. 9b) reveals that the variations stored in the cores from one site are similar, but with different amplitudes. For Ein Gedi at least four eastward swings in declination starting around 8.8, 6.5, 4.6 and 1.8 ka cal. BP as well as the westward movement of the geomagnetic vector at 2.8, 5, 7.8 and 9.6 ka, respectively, could be identified in both cores. In the short time interval covered by the Ze’elim cores two westerly swings are (partly) observed between 2.4 and 2.2 ka and between 1.4 and 1 ka cal. BP, respectively (Fig. 9b). The latter is also present in the record from DSF-B occurring from of 2.6–2 ka cal. BP.

In order to obtain representative ChRM-inclination and declination curves for each site, the records derived from the investigated cores were stacked, filtered with a weighted three-point-running-average and again, transformed into time series (Figs 10a and b). The cores from Ein Gedi and Ze’elim were stacked after numerical resampling of the ChRM-inclination and declination records in 2.5 cm intervals and calculating the mean inclination and declination for each horizon of cores DSEn-C/A and DSEn-B, and DSZ-A and DSZ-B, respectively. Most of the high frequency-variations in the inclination records obviously do not represent the secular variations in the local geomagnetic field; they are not correlatable between different sites.

Interlake comparison

The inclination and declination records obtained from the three Dead Sea sites were subjected to a regional comparison in order to test the significance of the reconstructed signal, especially regarding the mixing of different magnetic carriers (Figs 10a and b). The intralake comparison revealed no hint, whether and to which extent the recorded variations are affected by the secondary formation of sulphidic ferromagnetic minerals. The records chosen for comparison are from the nearby sites Lake Kinneret (Thompson et al. 1985) and Birkat Ram (Frank et al. 2002) (Fig. 1). The chronology for Birkat Ram was improved on the base of new radiocarbon datings published by Schwab et al. (2004). It was corrected for the reservoir effect, which is in the range of 500–700 yr, as it was estimated by the results of pollen analysis (Schwab et al. 2004). Because the age shift caused by the reservoir effect is not linear throughout the profile, a constant value of +600 yr was applied. The Kinneret record is that of core Kind4, plotted versus calibrated C14-ages as published by Thompson et al. (1985) although this calibration was revised in the meantime by Stiller et al. (2001). The most prominent correlatable features are marked by dashed lines. These are an inclination minimum dated between 2.1 and 1.8 ka (Fig. 10a) and the already mentioned westward swing in declination between 3 and 1.6 ka (Fig. 10b). Both features are well known from European PSV records (e.g. Thompson & Turner 1979; Saarinen 1998; Stockhausen 1998). The eastward deflection at 3 ka is also known from Japan and...
North America and is interpreted as a feature of the non-dipole field drifting westwards with a rate of about 0.13 °a−1, reaching Bulgaria around 2.8 ka ago (Hyodo et al. 1993). However, the age difference of 0.4 ka for this feature, between the Ein Gedi record and the archaeomagnetic record of Bulgaria, (Kovacheva 1997) is too long to be attributed to the westward drift alone. The given drift rate would allow for a difference in age of only 80 yr. An explanation is given, assuming a dating error of ±5 per cent in both chronologies. Thus the inclination low at 1.9 ka in the Dead Sea records corresponds to a pronounced inclination low in Europe at 1.7 ka BP, as can be achieved from the archaeomagnetic records of Bulgaria (Kovacheva 1997, Fig. 11) and Hungary (Márton 2003). Summarizing the preceding discussion it can be concluded, that during the last 3 ka the sediments from the Dead Sea recorded variations that are similar to known directional variations of the Earth magnetic field.

For the sequences older than 3 ka, there are fewer similarities between the records from the different lakes from Israel. The inclination maximum at 6 ka in Birkat Ram seems to have an equivalent in the Ein Feshkha and Ein Gedi record at around 5.4 ka (Fig. 10a) and the general trends in the declination records are similar (Fig. 10b). The inclination record from Lake Kinneret show an opposite trend compared to Ein Gedi and Birkat Ram, indicating that the given ages for this profile might be too old by up to 1.0 ka.

In order to get additional information about the variations in inclination that can be expected for the time interval from 3 to 10 ka, the archaeomagnetic record from Bulgaria (Kovacheva 1997; Kovacheva et al. 1998) was taken for comparison with the Ein Gedi curve (Fig. 11a). Although the time resolution in the archaeomagnetic record is much lower than in the lacustrine record, there is a good agreement in the succession of the inclination variations between both records. The only major feature missing in the inclination curve from the Dead Sea is the pronounced inclination low at 0.7 ka (Fig. 11a). This must be interpreted as a different behaviour of the geomagnetic vector in the Near East compared to Southeast Europe during the last 1000 yr, because this feature, well known as feature β in the British mastercurve, (Turner & Thompson 1981) is also missing in the record from Birkat Ram (Fig. 10a) and in the archaeomagnetic data from Egypt (Hussain 1983, 1987). In contrast to the comparison of the inclination curves, similarities between the declination curves from Bulgaria and Ein Gedi are rare and restricted to the time interval from 0 to 3 ka cal. BP. (Fig. 11b). This might
Figure 8. (Continued.)

Figure 9. Comparison of (a) the ChRM-inclinations and (b) the ChRM-declinations obtained from cores DSF-B, DSEn-C/A, DSEn-B, DSZ-A and DSZ-B. All records were transformed into time series using the age models presented in Fig. 4. The low frequency variations in the inclination and declination records are highlighted by a 10 point polynomial fit (black line).
Figure 10. Comparison of the stacked lacustrine (a) inclination and (b) declination records from sites Ein Feshkha, Ein Gedi and Ze’elim with those from Lake Kinneret, Israel (Thompson et al. 1985) and Birkat Ram, Golan Heights, Israel (Frank et al. 2002). The dating of the Birkat Ram record was modified following the new age model by Schwab et al. (2004). The correlatable features are marked by dashed lines.

Figure 11. Comparison of the stacked (a) inclination and (b) declination record from site Ein Gedi with the archaeomagnetic record from Bulgaria (Kovacheva 1997; Kovacheva et al. 1998). The latter was not corrected for latitude.

Discussion and Conclusion

Considering the results of the rock magnetic investigations, it has to be discussed, whether sediments with such a composition of the magnetic mineral fraction like those from the Dead Sea, can be suitable recorders of geomagnetic field variations. As was shown by the interlake comparison of the inclination and declination records, the directional variations obtained from sites Ein Feshkha, Ein Gedi and Ze’elim on the western shore of the Dead Sea (Figs 10a and b) correspond to PSV expected for this region for the last 3 ka cal. PB., at least. The agreement between the inclination and, to a much lesser extent, the declination records indicates, that the sediment magnetization in all three profiles partly reflect the directional variations of the geomagnetic field. Because the amplitudes in the directional records are in the same range as those from Birkat Ram, Lake Kinneret and Bulgaria it could further be assumed, that the signal was only minimally smoothed during recording. A high degree of smoothing would be expected if the sedimentation rate is low, such as in marine records (Lund & Keigwin 1994) or if a primary signal carried by a detrital phase is superimposed by a secondary signal recorded with a time delay longer than the expected lock-in-time and/or over a longer time interval. In the latter case, two components with different directions should be separated during stepwise demagnetization of the NRM, on the condition that the coercitivity spectra or Curie-temperatures of the two magnetic mineral fractions are clearly different.

In the sediments from the Dead Sea Ti-magnetite and greigite carry the magnetic remanence. The coercitivity of greigite is in the range of SD-magnetite with a coercitive force ($B_C$) between 30 and 50 mT and the coercitivity of remanence ($B_{CR}$) varying between 60 and 80 mT (Snowball 1991; Roberts 1995; Dekkers & Schoonen...
These values are more than two times higher compared to the mean values of 10 mT (B_C) and 33 mT(B_CR) obtained from detailed rock magnetic investigations on Dead Sea samples carrying only detrital Ti-magnetite (Frank et al. 2007). However, the primary and the secondary magnetization could not be separated during AF-demagnetization as it is presented in the Zijderveld plots (Figs 5a–f). This indicates, that either the time delay between the alignment of the detrital magnetic particles in the sediments and the growth of Fe-sulphides is in the range of the lock-in time for the detrital particles so both components are parallel, or that the AF-demagnetization can not effectively separate both components. As was revealed from the rock magnetic investigation (Frank et al. 2007), the coercivity spectra for the samples containing mostly greigite and those dominated by Ti-magnetite are overlapping. Thus a successful separation by AF demagnetization is not possible. A rock magnetic study on sediments from the outcrop Ze’elim (Ron et al. 2006) yielded similar results during AF demagnetization. These findings corresponds to the observations made by Snowball & Thompson (1990) on sediments from Loch Lomond, Great Britain, also realizing that the magnetic characteristics of magnetites and greigites embedded in sediments are too similar to be distinguished from their contribution to the magnetic remanence during AF-demagnetization. A thermal demagnetization of the Dead Sea samples presented in this study was not possible due to the sample methods used, that is plastic boxes. However, thermal demagnetization of the Ze’elim samples taken by Ron et al. (2006) yielded a two-component magnetization with greigite as the dominant magnetic carrier and Ti-magnetite, as would be expected. Unfortunately, it is not shown if and to what extend the directions recorded by the two magnetic carriers differ.

At this point is must also be taken into consideration that there is no information on the moment in time the digenetic formation of greigite started after deposition of the sediments. Investigations on sediments from White Rock Lake, USA showed that greigite is present already in sediments not older than 25 yr (Reynolds et al. 1999). The sediments from this reservoir lake are not comparable to the sediments from the Dead Sea, with regard to water chemistry and environmental conditions, but these results indicate that the time delay between the deposition of the sediments and the formations of greigite can be very short compared to the time resolution provided by sedimentation rate, sampling intervals and sample size. Additional information is available from geochemical investigations on the iron content in the water body of the Dead Sea by Nishri & Stiller (1984). They found that Fe-sulphides are even present in the anoxic lower water body (LWB) below a water depth of 80–100 m. Further investigations by Gavrieli et al. (2001) yielded that bacterial sulphate reduction also occurs in anoxic subsurface brines that emerge from thermal springs at the shores from the Dead Sea. One of this springs, Enot Zuzim, is situated near Ein Feshkha. Both observations indicate that the authigenic formation of Fe-sulphides in the Dead Sea sediments not only starts shortly after the deposition but that Fe-sulphides are formed already within the water body. So they are a portion of the detrital magnetic fraction, too. This assumption is corroborated by the observation that the directions recorded in samples with greigites as the dominant magnetic carrier minerals did not differ from those carried by Ti-magnetite in neighbouring samples (Figs 7a–c). Thus it can be concluded that in the sediments from site Ein Feshkha the detrital as well as the CRM have formed contemporaneously, considering the time-resolution given by the sample intervals and the lock-in depth, as a primary magnetization. Than the question arises if and to what extend this primary magnetic signal was altered after recovery of the cores, that is oxidation of the greigite exposed to air.

Although the subsampling of the cores started some years after opening of the cores, there was no loss in the magnetization as it was shown by Frank et al. (2007), comparing the results of susceptibility measurements on the split halves of the cores and on the freshly taken subsamples. The hypersaline water was still in the sediments and prevented oxidation of the greigites except for the surface of the core halves. Therefore, it could be assumed that the directions obtained from the Dead Sea sediments are the recordings of the geomagnetic signal at time of or shortly after deposition. However, this says nothing about the quality of this stored signal itself, which might be very low in the coarse grained sediment sequences. Additionally, the greigite will be oxidized in the sediments that were exposed to the air during low stands of the Dead Sea leaving only the coarse grained Ti-magnetite behind. It has also to be considered that the lacustrine sediment sequences from the Dead Sea presented here, are not only tens of km apart but that they also reflect different sedimentation enviroments. So the differences between the records of the five cores from the Dead Sea (Figs 9a and b) are rather due to differences in sedimentology, sedimentation rate and weathering of exposed sediments as to the presence of diagenetically formed greigite.

Despite all the objections on the recording fidelity of the Dead Sea sediments there is an agreement in the age of the correlatable features as obtained from the interlake comparison of the inclination records in Figs 9(a), 10(a) and 11(a) with the Ein-Gedi and Ein Feshkha record at least. The divergences are in the range of a dating error of ±5 per cent in the age models for the sediment profiles from Birkat Ram, where a hard water effect is documented, the Dead Sea and the archaeomagnetic record from Bulgaria. This implies, that sediments with a heterogeneous composition of the mineral magnetic fraction are not a priori useless for the reconstruction of geomagnetic field variations, even if it is not possible to separate the remanence components carried by different magnetic minerals during AF treatment.

Nevertheless, there is a need for further detailed investigations on this kind of sediments, in order to gain an improved understanding on how the magnetic signal is recorded in laminated sediments affected by reductive diagenesis and authigenic Fe-sulphide formation.

ACKNOWLEDGMENTS

The authors wish to thank J. Mingram, D. Berger, M. Köhler and M. Prena for coring, and H. Lippitz for help during laboratory work. M. Stein contributed to the organization of the drilling campaign. M. Kovacheva and an anonymous reviewer are acknowledged for their comments on a former version of this paper. This study was funded by the German Research Foundation (DFG), grant Ne 154/37-1.

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