Summary. Four box cores collected from the Ontong–Java plateau during the Eurydice expedition have been used to make relative geomagnetic palaeointensity measurements. Rock magnetic measurements on the sediments show that they are characterized by a uniform magnetic mineralogy, and that they are suitable for relative intensity estimates. These are obtained by normalizing the NRM by an ARM imparted in a low DC bias field. The palaeoceanographic event known as the preservation spike is used to establish a crude time-scale for the record so that it may be compared with other data from the same region, and also with global palaeointensity estimates. The marine sediment data are quite similar to Australian intensity data from lake sediments and archaeomagnetic sources, but as might be expected exhibit some obvious differences from the global record.

Key words: Palaeomagnetic field intensity, palaeosecular variation, relative palaeointensity

Introduction

Over the past few decades, there has been a tremendous increase in our understanding of palaeomagnetic secular variation over a range of time-scales [see Merrill & McElhinny (1983) for an excellent review]. However, advances in our knowledge of palaeomagnetic intensity variations have lagged somewhat behind those for directional variations owing to the stricter requirements necessary for obtaining reliable palaeointensity data. The ideal record for palaeomagnetic intensity information is a stable total thermal remanent magnetization (TRM) that is acquired when material cools through the Curie temperature of its magnetic phase and does not suffer subsequent remagnetization. Palaeointensity information can therefore be derived from archaeological material (e.g. baked hearth stones and fired clays or...
pottery) as well igneous rocks such as lava flows. Such data provide spot readings of the geomagnetic field at the time of cooling; unfortunately, the nature of the material required and the extensive effort necessary to obtain reliable laboratory results mean that the record is not only discontinuous but also very sparse prior to about 12,000 yr BP (McElhinny & Senanayake 1982).

Until comparatively recently, archaeomagnetic and lava flow data were the only reliable palaeointensity information available. However, sedimentary sequences can provide high resolution, continuous records of palaeomagnetic field information (e.g. Clement & Kent 1984), and under certain circumstances these records can be used to provide accurate relative palaeointensity estimates (King, Banerjee & Marvin 1983; Constable 1985). The method for obtaining these estimates relies on the assumption that the intensity of magnetization retained by a sediment is linearly proportional to the field causing it. Variations in concentration of magnetic minerals down the length of the stratigraphic section can therefore be compensated for by normalizing the NRM using a suitable magnetic parameter, ideally one which varies with concentration in the same way as the NRM. This normalized NRM will then reflect relative changes in field intensity along the length of the section. Investigations into magnetic mineralogy suggest that low field ARM is usually the most suitable normalizing parameter (Johnson, Kinoshita & Merrill 1975; Levi & Banerjee 1976). This method for obtaining relative palaeointensities will obviously only be successful in situations where the magnetic mineralogy is uniform throughout the section and where the linearity constraint mentioned above is valid. Based on a detailed rock magnetic examination of some lake sediments, King et al. (1983) have outlined some characteristics which are desirable for sediments before they might be considered as reliable recorders of relative palaeointensity. These are (1) the carrier of NRM must be magnetite or some similar mineral such as titanomagnetite, (2) the magnetite must be in the pseudo-single domain grain size range, and (3) the concentration of magnetic material may not vary by more than a factor of 20 or 30.

Lake sediments which satisfy the above criteria have been exploited for relative palaeointensity information with some degree of success (Constable 1985; King et al. 1983). However, rapidly deposited deep-sea sediments, which also have the potential to provide high-quality records of geomagnetic field variation (Kent & Opdyke 1977; Clement & Kent 1984; Kent 1973; Harrison 1974) have been largely ignored for relative intensity work. These deep sea records have the advantage of providing long continuous records over time scales which are not normally available from lake sediments. The lower sedimentation rate in deep sea sediments compared with lake sediments means that the resolution of these records will not be as high as in lake sediments, however this is more than compensated for by the additional length of time available. The work of Karlin & Levi (1983, 1985) has aroused some scepticism about the value of marine sediments for high-resolution palaeomagnetic studies. They found substantial down-core changes in the magnetic mineralogy of several hemi-pelagic cores, arising from the reduction of organic matter. Such diagenetic processes lead to the dissolution of magnetite and result in a reduction in palaeomagnetic stability and overall quality of the sedimentary record. Changes in grain size, magnetic stability and/or mineralogy with depth should therefore be viewed with extreme caution. It should, however, be noted that the data of Karlin & Levi (1983, 1985) come from hemi-pelagic sediments with high organic carbon contents and are not typical of pelagic high-carbonate oozes.

In this paper we assess the potential of sediments from the Ontong Java Plateau for obtaining relative palaeointensity information. We have studied several box cores and found that they meet the minimum necessary requirements for retaining a palaeointensity record.
We compare the relative intensity estimates obtained from these cores with global and Australian palaeointensity records.

**Palaeomagnetic measurements**

We have chosen four box cores from the Ontong Java Plateau (ERDC 83Bx, 92Bx, 102Bx and 120Bx) that were collected in 1975 during the Eurydice expedition. Fig. 1 shows the location of the cores and its relationship to the nearest other palaeointensity site. The core locations and water depths are listed in Table 1. Extensive sedimentologic and geochemical data are available for these cores (Johnson, Hamilton & Berger 1977; Berger, Johnson & Hamilton 1977; Berger & Killingley 1982; Berger 1982; and SIO reference series, in preparation). These cores have been kept wet and refrigerated since they were taken, thus minimizing potential damage to their magnetic properties from dehydration.

Palaeomagnetic specimens were prepared in a manner inspired by Clement, Kent & Opdyke (1982). Using an Exacto knife, the centre slabs (2 cm thick) of the box cores were sliced into wafers from which two to four specimens from each stratigraphic level could be cut. The interval from 28 cm to 38 cm depth in core ERDC 120Bx was sliced into 0.5 cm wafers but the data from a single stratigraphic level were scattered owing to the softness of the sediment. All the other wafers were 1 cm thick. Individual specimens were weighed in order to normalize palaeo- and rock magnetic measurements. The specimens were refrigerated wherever possible to minimize desiccation.

![Figure 1. Sites of the Eurydice cores (squares) and the Australian lake sediment data (triangle).](image-url)
Table 1. Eurydice box cores used in this study.

<table>
<thead>
<tr>
<th>Core no.</th>
<th>Water depth (m)</th>
<th>Length (cm)</th>
<th>Latitude °</th>
<th>Longitude ° E</th>
<th>Preservation spike depth (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>83</td>
<td>2342</td>
<td>40</td>
<td>1° 24.1' N</td>
<td>157° 18.6' E</td>
<td>23.5</td>
</tr>
<tr>
<td>92</td>
<td>1598</td>
<td>32</td>
<td>2° 13.5'S</td>
<td>156° 59.9' E</td>
<td>26.5</td>
</tr>
<tr>
<td>102</td>
<td>2306</td>
<td>32</td>
<td>3° 36.3'S</td>
<td>161° 19.1' E</td>
<td>17.5</td>
</tr>
<tr>
<td>120</td>
<td>2247</td>
<td>38</td>
<td>0° 1.0'S</td>
<td>158° 41.6' E</td>
<td>26.0</td>
</tr>
</tbody>
</table>

All the palaeomagnetic measurements described here were performed in the Scripps Palaeomagnetic Laboratory. Remanence measurements were made on a CTF 3-axis cryogenic magnetometer housed in a magnetically shielded room (ambient field less than 200 nT). Laboratory procedures are described by Tauxe et al. (1985).

In Fig. 2 we show representative examples of vector end-point diagrams during alternating field and thermal demagnetization. The specimens tend to exhibit a very slight change in direction when demagnetized to 10 mT, after which they decay univectorially to the origin. The median destructive field for all the box core specimens was between 20 mT and 30 mT, indicating high stability for the NRM. The maximum blocking temperature of between 550°C and 600°C suggests that the remanence is carried by pure or nearly pure magnetite, fulfilling the first requirement of King et al. (1982) for palaeointensity studies.

We used a technique developed by Banerjee, King & Marvin (1981) for investigating magnetic granulometry using measurements of anhysteretic remanent magnetization (ARM).

Figure 2. Vector end-point diagrams of pilot specimens from the Eurydice box core collection. The cores are unoriented in the horizontal plane. The squares are vector projections onto the horizontal plane, and the circles are projected onto the vertical plane. The units are in 10^{-4} mA m^{-2} (total moment).
and bulk low-field susceptibility ($\chi$). King et al. (1982) elaborated on the technique and illustrated how these parameters may be used to determine variations in magnetic grain size and concentration. Changes in the ARM/$\chi$ ratio imply changes in grain size, with higher ratios indicating smaller grain sizes. Changes in ARM and $\chi$ that maintain a constant ratio imply changes in concentration of magnetic material only.

We have measured ARM and $\chi$ for all four box cores; however, as the magnetic properties are remarkably similar among the cores we will discuss only the results from ERDC 92Bx. One specimen from each level of 92Bx was given an ARM in an AF peak field of 100 mT using a DC bias field of 0.05 mT. The bulk low field susceptibilities for these specimens were measured on a Sapphire Instruments SI-1 susceptibility bridge and were quite low, averaging $4.5 \times 10^{-7}$ SI. The susceptibility values reported here are the mean of at least three measurements. A calibration sample composed of 20 mg of MgO was measured between specimens to correct for instrument drift.

We plot ARM versus $\chi$ for ERDC 92Bx in Fig. 3, where the squares are data points from the upper 12 cm of the core and the octagons come from levels below 12 cm. These data cannot be compared directly with those of King et al., because the susceptibilities have not been corrected for the effect of carbonate matrix. However, relative changes in concentration and grain size will still be apparent. With the exception of a single outlier (probably a measuring error), the data lie along a line, suggesting that there are only minor changes in the grain size and that changes in magnetic mineral concentration are the dominant effect. It is probably not significant that the octagons all lie slightly below the squares, suggesting either a very slight coarsening of grain size down-core or a trend in the matrix contribution to $\chi$, as this effect is at the limit of the resolution of the susceptibility measurements. Changes in the magnetic mineral concentration can be assessed using the ratio of the maximum susceptibility to the minimum; this is about two in the case shown in Fig. 3, which is far less than the fact of 20 considered a maximum by King et al. (1983).

![Figure 3. Plot of ARM intensity versus bulk low field susceptibility for core ERDC 92Bx.](image-url)
The absolute grain size of the magnetic material cannot be determined from Fig. 3 because of the contribution of diamagnetic material to the susceptibility measurements. However, based on the behaviour observed during demagnetization we argue that these sediments retain a stable remanence, and the magnetic material is probably well within the pseudo single domain grain size. From the foregoing evidence, we tentatively conclude that the sediments from the Eurydice box core collection are suitable for relative palaeointensity studies. In the following section we present our relative palaeointensity data and then compare them with independent absolute and relative palaeointensity data.

Results

Weight normalized NRMs for all the cores are plotted as a function of depth in Fig. 4. This material is very strongly magnetized for a marine sediment, with NRMs averaging more than $10^{-6}$ Am$^2$ kg$^{-1}$. Several measurements were made per horizon for each core and these are plotted individually in the figure. The results from core ERDC 83Bx are highly reproducible between specimens at each horizon. Some of the other cores are more scattered; in particular, the data near the bottom of core ERDC 120Bx, where we attempted to obtain very closely spaced samples, is of distinctly poorer quality. We attribute this to the difficulty inherent in taking such small samples. Cores ERDC 83Bx and ERDC 102Bx exhibit the most similarity between their NRM profiles; however, we do not necessarily expect the profiles to be identical because the cores come from sufficiently different locations and water depths (see Table 1) that differences in mineralogy and magnetic mineral concentrations are quite probable. We would naturally expect these differences to be reflected in the NRM profiles.

In Fig. 5 we plot the ARM intensities determined for each of two specimens per level for all the box cores. As with the NRM data, there are distinct differences in behaviour among

![Figure 4](image-url)
and within the cores, reflecting the variable concentrations in fine grained magnetite. Here again we may expect differences among the cores, reflecting the different water depths and sedimentary environments of the core locations.

Several steps are necessary to convert the NRM versus depth data of Fig. 4 into relative palaeointensity as a function of time. First we normalize the NRM data by the ARM. Then isochronous tie points must be found in the various records and an absolute time scale established for the combined data set. Isochronous tie points that are independent of the magnetic relative intensity data could be provided by a variety of methods, e.g. stable and radioisotope stratigraphies, regionally consistent lithological criteria such as percentage of CaCO₃ or its complement, magnetic susceptibility. The δ¹⁸O data are very scattered and the distinctive δ¹⁸O maximum (which would be an obvious tie point) occurs at about 18 000 yr which is too old for these sediments, thus reducing the utility of stable isotope data for providing independent high precision tie points in these young sediments. Also, although there are many radiocarbon dates available for these cores (Berger & Kilingley 1982), differential dissolution (Keir 1984) and bioturbation (Peng, Broecker & Berger 1977) complicate the record so that the ¹⁴C stratigraphies are not sufficiently accurate for precise correlation between cores. We have therefore chosen to use a peak in carbonate preservation (the preservation spike) as a tie point. The preservation spike is a readily identifiable regional palaeoceanographic event associated with changes in carbonate preservation (Broecker & Broecker 1974; Shackleton 1977; Berger et al. 1977). It is expected to be isochronous on the Ontong–Java Plateau and its position has been identified in all our box cores (Berger et al. 1977, see also Table 1).

A common depth-scale can now be derived for the NRM/ARM data by setting the tops of the cores and the preservation spike levels at the same age in all the cores. Fig. 6 shows the
ratios of NRM to ARM for the individual cores with their depth scales adjusted so that the tie points coincide. We are now in a position to compare the relative intensity data from the different Eurydice cores. We assume in what follows that sediment accumulation rates were either linear or covaried in all the records presented here. This is obviously an approximation; how valid it is will be apparent in the degree of mismatch which occurs among the combined records. Non-linear sediment accumulation will also result in discrepancies between the combined Eurydice data and data from elsewhere.
Two specimens per horizon were used, to assess the consistency of the record within a single core, and the error bars in Fig. 6 represent one standard deviation based on data from these two specimens. It is immediately obvious from the size of these error bars that the data from cores ERDC 83Bx and ERDC 102Bx are more reproducible than those from ERDC 92Bx and ERDC 120Bx. It is also of interest to note that the ratios NRM/ARM are uniformly about a factor of 2 or 3 lower in the latter cores. We attribute this to sedimentological differences between the sites, although we were unable to determine any systematic differences between the two sets of cores. ERDC 92Bx which has the least reproducible results is the core which comes from the shallowest water depth, and there is some indication that winnowing might have taken place (Peng et al. 1977). Also, the specimens from ERDC 92Bx were inadvertently allowed to dry out before measurement, which resulted in reduced NRM intensities. Whatever the reason, we do not need to be unduly concerned by these scale differences between cores, as minor differences in water content or magnetic grain size between core could readily account for them. We should rather be encouraged by the fact that despite the somewhat different environments in which these sediments appear to have been deposited (reflected in the differing NRM and ARM data among the cores), the NRM/ARM profiles all exhibit a similar type of behaviour. The relative intensities all increase gradually with depth in core to a sub-bottom depth of between 5 and 10 cm and then fall off slowly with continued increasing depth. We believe that these profiles could well represent relative intensity variations in the geomagnetic field, because there are no indications of chemical or magnetic mineralogical changes occurring which could account for the variations down core. This viewpoint is further substantiated by the comparison with data from other sites made in the next section.

Before the data from different cores can be superposed and combined, it is necessary to compensate for the scale differences in relative intensities. Since it is an implicit assumption of the method used here, that there is a linear relationship between the applied palaeofield and the relative intensity, it is only necessary to estimate one parameter; the zeroes of all the relative intensity curves should be the same. As the intensity scale is completely arbitrary we simply choose the core with the smoothest, most consistent looking profile and scale the other data to match it. This was achieved by interpolating the data for the master core to provide values at the same depths as in the core to be scaled, and then finding the multiplicative factor between the cores which resulted in the minimum possible misfit in a least-squares sense between the two data sets.

Finally, we must convert the depth scale to time in order to compare our data with that from other localities. This is accomplished by assuming that the top of the core coincides with the present day, an assumption which is probably accurate to within 500 years, since generally less than 1 cm has been lost from the top of the core in the coring process (Peng et al. 1977). The preservation spike has been dated by $^{14}$C at about 11 000 BP (Berger 1982). When the age of the surface water reservoir is taken into account, this age is corrected to about 10 500 BP (Berger, private communication). This is the date which we have used in the establishment of the time scale for these sediments.

In Fig. 7 we plot the data from Fig. 6 after scaling the palaeointensities and converting the depth scale to time. The upper portion of the figure is the data from the two best cores, ERDC 83Bx and ERDC 102Bx, for which the data are really of very high quality. The lower diagram shows the data from all the cores superposed. Despite the much larger error bars associated with the other two cores, it is satisfying to note that the general trends in the magnetic field variation are reflected in all the data. Undoubtedly the most detailed information will come from cores ERDC 83Bx and ERDC 102Bx, but the other data do not contradict this information, they merely lack the same resolution.
Lock-in depth of the magnetization

The foregoing discussion established a time scale for the sedimentation process in the Eurydice box cores. However, this does not necessarily correspond to the time at which the magnetic remanence is locked into the sediment. It is well known that if sediments have a sufficiently low shear strength, they may be disturbed (e.g. by bioturbation) and become remagnetized in the direction of the prevailing magnetic field, with an intensity of
Marine sediment palaeointensity

magnetization which depends linearly on the applied field strength (Irving & Major 1964; Kent 1973). The maximum depth within the sediment at which this occurs is not known (and will depend on the constitution of the sediment) but is generally assumed to lie between 10 cm and 60 cm (Verosub 1977). The sediments in the Eurydice box cores are thoroughly bioturbated (Ekdale & Berger 1978) and extensive efforts have been made to model the effects of bioturbation in these cores (e.g. Berger & Killingley 1982; Peng et al. 1977).

Berger & Heath (1968) envisioned bioturbation involving complete mixing to a certain depth below which no mixing takes place. Guinasso & Shenk (1975) modified this simple concept by modelling bioturbation as a diffusion process, which is assumed to occur at a fixed rate to a specified maximum depth. High diffusion rates or low sedimentation rates reduce this model to that of Berger & Heath (1968). Peng et al. (1977) successfully applied the Guinasso–Shenk approach to explain lead and carbon isotope concentrations in ERDC 92Bx by assuming a mixing depth of 8 cm. However, they also point out that the data can be satisfied equally well by an entirely different method which assumes complete mixing to a depth of 3 cm below which the degree of mixing is assumed to fall off exponentially (leaving the sediment relatively undisturbed below about 4 cm). Although the two processes are equally successful in accounting for the isotopic data in the Eurydice box cores, they would not produce the same magnetic records.

Jones & Ruddiman (1982) provide a plausible physical basis for the exponential model of Peng et al. (1977). They point out that there are two kinds of mixing; homogeneous and heterogeneous. Homogeneous mixing is produced by continuous foraging and movement through the uppermost sediments [as described by the Berger & Heath (1968) model]. Heterogeneous mixing, however, involves the random forays of large burrowing organisms, occurring sporadically below the well-mixed layer. The probability of such heterogeneous mixing is likely to decrease exponentially with depth. Magnetic records produced by homogeneous mixing would be uniform within the mixed layer and retain a record of the palaeofield below that. Modification by subsequent heterogeneous mixing would result in variability in parallel records with different heterogeneous mixing histories.

Our experience with the geomagnetic signal is that it appears to be locked in on a much faster time-scale than the Guinasso–Shenk model for Eurydice isotope data suggests. The consistency of the magnetic record near the top of the core precludes a homogeneous mixed layer of more than a few centimetres (and it could be less than this). Diffusion of water or specific isotopes may be taking place to a deeper level but it does not appear to affect the magnetic record. It thus seems that the favourite mixing models used for isotopes are not applicable to the magnetic record. The exponential model of Peng et al. (1977), however, is more compatible with the data presented here. The acquisition of remanence in sediments is still a relatively poorly understood process, in the sense that it is highly dependent on the nature of the sediment. The effects of diffusion of water and isotopes will depend on the relative grain sizes of the magnetic particles and the sediment matrix (Tucker 1980), as well as other characteristics of the sediment. In the absence of a better model we have simply assumed that the top of the core represents the present-day record of the geomagnetic field; we do not expect that this will be in error by more than about 1000 years.

Comparison with other sites

Perhaps one of the trickiest problems with relative geomagnetic intensity data has been establishing whether or not they provide a truthful description of palaeointensity field
variations. The paucity of long palaeointensity records at a single site is due to the stringent requirements for obtaining absolute palaeointensity measurements, i.e. there are very few sites where suitable materials are available covering a long time span in anything like a continuous fashion. Indeed, it is only comparatively recently that sufficient data have become available for reliable estimates to be made of average global palaeointensity variations (McElhinny \& Senanayake 1982). The closest data which are available for comparison with the Eurydice data come from Eastern Australia and are relative intensity data obtained from lake sediments (Constable 1985). The site location is indicated by the triangle in Fig. 1. Despite the fact that these are relative palaeointensity data, we are confident that they do represent an accurate record of the geomagnetic field over at least the past 6000 calendar years because there are independent corroborative archaeointensity data to support them (Barbetti 1983).

The rather large distance between the Eurydice (latitude \(-0^\circ\)) and the Australian lake sediment sites (latitude \(-17^\circ\)) raises the question of whether it is valid to make comparisons between raw magnetic field intensities or whether we should actually compare virtual dipole moments (VDMs). The computation of VDMs will remove scatter in the records caused by dipole wobble. As is well known (see e.g. Merrill \& McElhinny 1983) the geocentric dipole moment \(p\) is related to the field intensity \(F\), of the Earth radius \(R\) at a site with inclination \(I\) by

\[
p = \frac{4\pi R^3}{2\mu_0} (1 + 3 \cos^2 I)^{1/2}.
\]

Directional data are available for the Australian lake sediment data (Constable \& McElhinny 1985), however, since the inclination variations at the site in general appear to be small (\(-10-15^\circ\)), and the latitude is quite low, the effect on the record of computing VDMs is quite small (less than 10 per cent changes in amplitude). The effect at the equatorial Eurydice site would be smaller still. We may therefore make direct comparisons between the relative intensity data, with the knowledge that the picture will change very little when VDMs are compared. It also facilitates a direct comparison between our data and the global palaeointensity record compiled by McElhinny \& Senanayake (1982). Since there is an arbitrary scale factor in our data this is equivalent to treating the relative intensity data as virtual axial dipole moments (VADMs).

The lake sediment records come from two volcanic crater lakes, and when combined, they provide an intensity record extending from about 1500 BP to 16 000 BP, approximately the same time interval as the Eurydice data. Comparison between these data and the Eurydice data is rendered somewhat difficult by the differing frequency content of the two records. The Australian lake sediments have an order of magnitude higher sedimentation rate than the Eurydice cores; a typical Eurydice sample spans about 500 yr, compared with about 40 yr for the lake sediments, so that much higher frequency variations may be detected in the Australian record. We attempted to compensate for this by taking 500 yr block medians of the lake sediment data. Medians were used rather than means because they are more robust to the presence of outliers in the data, and there were small numbers of extreme outliers contaminating the Australian record. The result is shown in Fig. 8, where the triangles represent data from Lake Eacham and the squares data from Lake Barrine. The error bars represent one standard deviation in the estimation of the median. It may be seen from this figure that the two lake records are in excellent agreement with one another over their coeval time span. Constable (1985) also shows that they agree with Barbetti’s (1983) archaeomagnetic data which cover the past 6000 yr.

In Fig. 9 we superpose the Australian record and the data of Fig. 7(a), i.e. the best of the
Eurydice data (circles are Australian, squares Eurydice). Here the two Australian records have been combined and the 500 yr medians of the combined record are plotted. The Australian and Eurydice records are quite similar between about 1500 BP and 14000 BP, beyond which point there is an increase in intensity in the Australian record, followed by a decrease right at the end of the record. It is not clear whether this increase truly represents geomagnetic intensity variations, or whether the Eurydice record is more accurate. It is known that the sedimentation rate in Lake Barrine is much higher prior to 11000 or 12000 BP and that the time scale associated with the sediments is very uncertain before about 11000 BP. It is possible that what is plotted between 11000 BP and 16000 BP for the Australian record actually took place on a much more rapid time scale (say 1000–3000 yr).

The data of Fig. 9 show that the Australian and Eurydice records agree between 0 BP and 10000 BP. It is not necessarily obvious that this is to be expected between two sites which are so widely separated (~17° in latitude and 15° in longitude), although a clear correlation is seen between the archaeointensity data of Barbetti (1983) and the lake sediment data of Constable (1985), which are also quite widely separated. To compare the two sites more closely we computed the 500 yr median values for all the Eurydice data (not just those from the two best cores). We then plotted these median values for the two data sets against each
Figure 9. Comparison of Australian record with data from the two best Eurydice cores. The age scale is fixed and the Eurydice data has been scaled vertically using the least squares procedure described in the text. Circles are Australian, and squares Eurydice data.

other and found the scale factor relating the two relative intensity records (by a least squares fit to a line constrained to pass through the origin). In the lower part of Fig. 10 we plot these 500 yr median values for the two records, after applying the appropriate scaling factor. Error bars in this figure are 1 standard deviation in the median, except for the last three 500 yr bins which each contain only a single datum, where we have used the value for the adjacent bin at 16 900 BP. This is almost certainly a conservative estimate for the error. The Australian data are again represented by solid circles and the Eurydice data by open squares. Also plotted in the upper part of this figure (solid triangles) are the global average VDM data of McElhinny & Senanayake (1982). The Eurydice data have been scaled to match the global VDMs using the same least-squares minimization procedure described earlier. For the global data we have plotted the distributional standard deviation since this may be largely attributed to secular variation and provides an idea of the range within which the Eurydice data might be expected to fall. There are a number of systematic differences between the Pacific region records (i.e. Australian and Eurydice) and the global average record. First the Pacific data are systematically lower than the global data from 0 BP to 2000 BP. This feature is also strongly evident in the archaeointensity data of Barbetti (1983). Secondly, the pronounced minimum in the global data around 6000 BP is not as low in the Pacific data,
nor is the low value around 10,000 BP. There are very few global data older than about 11,000 BP, but what are available are consistent with the generally low field-values observed in the Eurydice cores. We should not really expect the global data to look like the Pacific region data, when the averages are taken over such short time spans. 500 years is not enough to average out the secular variation at a single site. In addition, the global data set is heavily
biased towards European sites. The Pacific data do however lie within the statistical variation expected in the global data, based on the distributional standard deviation computed by McElhinny & Senanayake (1982).

Conclusions

We have made relative intensity estimates of the geomagnetic field variations using ARM normalized measurements of the NRM for a site in the western equatorial Pacific ocean. The rock magnetic uniformity of these sediments suggested that they might contain material suitable for intensity estimates. The magnetic remanence appears to have been locked in within the uppermost few centimetres and there is a high degree of reproducibility of the broad trends in relative intensity among the four cores studied. There is good agreement between the Australian lake-sediment data and the Eurydice records over the past 10,000 yr, supporting the contention that these records reflect palaeointensity variations. Using the global data set compiled by McElhinny & Senanayake we were able to scale our relative intensity data to provide VADMs. These show similar trends to the average global data but do not replicate it exactly. This is to be expected for such short time averages at a single site. The data lie well within the expected distributional variation of the global data.

Although the data are inherently very smooth because of the low sedimentation rates, they provide a valuable new source of information, because of the potential for future work on the associated gravity and piston cores. If these associated cores can be shown to possess the same mineralogic uniformity throughout their length as those studied here, then relative intensity estimates extending back hundreds of thousands of years may become available.

Acknowledgments

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References

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