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Normalised natural remanent magnetisation intensity during the last 240 000 years in piston cores from the central North Atlantic Ocean: geomagnetic field intensity or environmental signal?

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Abstract

We have examined in detail the stratigraphic variations in magnetic parameters of four cores from the North Atlantic Ocean in areas where the depositional environment has varied with climatic changes. Our objective is to assess whether normalisation of the natural remanent magnetisation (NRM) intensity can cancel the effect of climatically induced variations in magnetic mineral content and grain size and whether a reliable record of relative changes in geomagnetic field intensity can be obtained. After selecting the core sections which meet published criteria for paleointensity normalisation, we have attempted to obtain a record of the geomagnetic field intensity variations over the past 240 kyear, using both ARM and IRM as normalising factors for the NRM. The two methods yield overall similar results, except for the interval 20–30 kyear, where IRM normalisation yields a record more consistent with previous sedimentary and volcanic results, than does the more frequently employed ARM normalisation. The final intensity record documents a picture of the dipole field moment which presents large similarities with profiles obtained from sediments deposited in different environmental conditions in various regions around the world. We observe a major low at about 42 kyear, which could correspond to the Laschamp event, a broad low in the interval 90–130 kyear, possibly connected to the Blake event, and another low at about 190 kyear, which could reflect the Biwa I event. The processes of acquisition of depositional and postdepositional remanent magnetisation and the physics of the normalising methods are still insufficiently understood. However, these results, obtained from a region characterised by complex environmental dynamics, confirm the potential of sedimentary deep sea cores for relative paleointensity determinations.

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1. Introduction

Observatory records and archeomagnetism have provided information on the short period (10–10^3 years), recent variations of the geomagnetic field, while the polarity time-scale gives information on aspects of long period geomagnetic changes (10^5–10^8 years). Records of the variations of the geomagnetic field on intermediate time-scales (10^4–10^5 years) would provide a significant contribution to our understanding of the geodynamo. Such records, if preserved in sediments, would present the attraction of providing continuous reconstruction of the geomagnetic field and would also possibly provide a high resolution magnetostratigraphic tool that would be most useful in paleoceanographic studies.

There have therefore been a number of recent efforts to produce relative paleointensity records from ocean cores (Kent and Opdyke, 1977; Constable and Tauxe, 1987; Tauxe and Valet, 1989; Tauxe and Wu, 1990; Yamazaki and Ioka, 1992; Meynadier et al., 1992; Tric et al., 1992a). Tric et al. (1992a) presented a record from the Mediterranean for the last 80 kyear, which is in agreement with absolute paleointensity data from lavas that date back to 40 kyear. This record has recently been extended to 140 kyear using cores from the equatorial Indian Ocean (Meynadier et al., 1992). Other results, covering longer time periods have been reported from the Western Pacific (Kent and Opdyke, 1977; Tauxe and Wu, 1990; Yamazaki and Ioka, 1992). Very recently a record from the Pacific spanning the last 4 Myear, suggests a link between non-reversing 'stable' polarity behavior and the reversing dynamo (Valet and Meynadier, 1993).

The procedure for obtaining a record of relative paleointensity from a sediment involves a normalisation of the natural remanent magnetisation (NRM) intensity to remove the gross effect of variation of concentration and grain size of the magnetic minerals. An excellent review of all the approaches, methods and results on this topic has recently been given by Tauxe (1993). Several normalisation parameters have been proposed: the anhysteretic remanent magnetisation (ARM), the low-field magnetic susceptibility (χ), and the isothermal remanent magnetisation (IRM). However, the process of acquisition of depositional and postdepositional remanent magnetisation (DRM and PDRM) is imperfectly understood, as is the physics of the normalising parameters such as ARM, and how it relates to grain size variation. So the problem of retrieving geomagnetic paleointensity records from sediments is a rather complex issue.

There are at least two approaches to this problem: the most straightforward and powerful of these is to compare relative paleointensity records from regions which experience quite different variations in sedimentary characteristics with climate. A common signal would then preclude climatic variation as a major factor in the relative intensity signal. A second approach is to characterise a core, both magnetically and sedimentologically, as completely as possible and then attempt to make direct observation as to whether variation in certain parameters coincides with variation in normalised intensity. An attempt to approach the problem along these lines is presented here, with the study of four marine cores from the North Atlantic.

2. Core location and sampling

Optimal areas for relative paleointensity studies of ocean sediment cores are difficult to define a priori. The sedimentation rate should be at least of the order of 4–5 cm kyear^{-1} to allow sufficient temporal resolution. High sedimentation rates usually occur in the deep sea on continental slopes where there are large fluxes of continental detritus in areas of high biogenic productivity. In such areas, problems may arise from changes in mineral magnetic characteristics (magnetite dissolution and iron sulphide formation) (Roberts and Turner, 1993). In this study, we have chosen a low productivity area with high fluxes of detrital minerals: the North Atlantic Ocean. The cores were obtained during the PALEOCINAT cruise of the R/V "Le Suroît" of IFREMER. Their location, length and the water depth are given in Fig. 1. It was not known until very recently (Grousset et al., 1993) that large

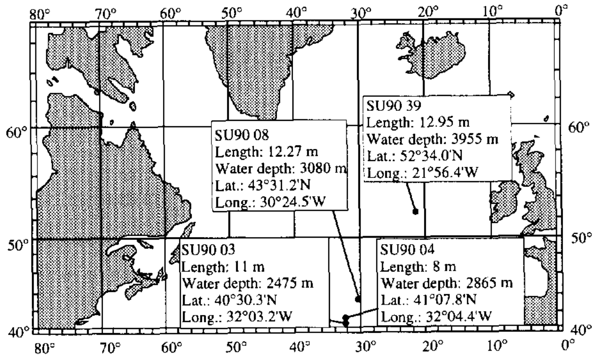

Fig. 1. Location map of the four sites studied in the central North Atlantic Ocean. The characteristics of the sites are also reported close to the corresponding points.

changes in sedimentation during glacial periods between 40° and 55° north occurred about each 7 kyear, associated with the so-called Heinrich events, which arise from the deposition of ice rafted material (Heinrich, 1988; Bond et al., 1992; Broecker et al., 1992). There is strong evidence that the magnetic properties of deep sea sediments from these regions, including the nature and concentration of the magnetic minerals and their grain sizes, are largely correlated with climatic parameters (Robinson, 1986; Mead et al., 1986; Bloemendal et al., 1988, 1992). The sites of cores SU9008 and SU9039 are in the middle of the zone affected by deposition of ice rafted material, while cores SU9003 and SU9004 are from sites close to the southern limit of the area affected by this phenomenon.

The cores were cut into azimuthally oriented sections which in turn were cut longitudinally into half cores along a reference line directly on board. In this work we made major use of the pass-through measurement technique in connection with a small access high-resolution cryogenic magnetometer now installed in our laboratory. Improvements in the method and its extension to measurement of a wider range of rock magnetic parameters, are reported elsewhere (Weeks et al., 1993). The u-channels developed and used in our laboratory (Weeks et al., 1993) are long open-sided 2 × 2 cm square sectional area, stiff plastic liners made of transparent plastic, so recovery of the sediment can be visually controlled. These liners are of the same length as the core sections (usually 1.5 m) and were pushed into the core half. An airtight cover, made of the same plastic, clips over the u-channel to hold the sediment in place and prevent it from drying. This sampling method produces less sediment disturbance than the conventional method of sampling cubic plastic boxes. The large diameter (10 cm) of the cores obtained during this cruise allows parallel sampling of either two or even three u-channels, or of u-channels and cubic (8 cm^3) boxes. On two of the cores, cubes were then taken at regular (10 cm) spacing so that results from the two methods could be compared. We also sampled cubes on the other cores with a larger spacing for the study of the anisotropy of the low field magnetic susceptibility. Some of these were also used for extraction of magnetic minerals for Curie balance and X-ray diffraction experiments.

3. Dating, time resolution and intercorrelation of the cores

Several independent parameters have been used to date and intercorrelate the cores. In order to get a reference time-scale, the oxygen isotopic ratios of the planktonic foraminifera *Globigerina bulloides* and *Neogloboquadrina pachyderma* were analysed in cores SU9008 and SU9039, with a resolution of 5–10 cm. The isotopic records are typical of the North Atlantic Ocean for the last two climatic cycles and all stages and substages may be correlated to the SPECMAP isotopic record which we use as our reference time-scale (Pisias et al., 1984) (Fig. 2).

The other cores were linked to the isotopic stratigraphy of these cores following the method defined by Bond et al. (1991), by a continuous correlation of the light reflectance signal ('gray scale'), which is a first order proxy of the carbonate content in the North Atlantic. The gray-scale reflectance signal was constructed by digitising reflected light along each segment of the cores with colour photography. This signal was averaged on a width of about 1 cm to smooth the effect of bioturbation. In addition, the carbonate content was measured at low resolution (each 10 cm) to support the gray-scale records.

The dating of the cores was thus performed in two steps. First, we have considered the carbonate and the gray-scale records by selecting well-identified common characteristic events for each core. The depth scales were adjusted linearly between these tie points. The best intercorrelation was reached by minimising the differences between the four cores for the gray-scale and the carbonate signals. These correlations were performed using the software 'Analyserie' developed at our institute (Paillard and Labeyrie, 1993).

The low field magnetic susceptibility, χ, of each core was measured continuously using a vertical system, made entirely of rigid plastic. A Bartington sensing coil, with 4 cm diameter, was displaced along a u-channel section placed in a vertical position. Measurements were made every 1 cm with the sensing coil moving continuously at a rate consistent with the sensitivity range used (Weeks et al., 1993). The resolution of the coil is about 3 cm. This signal is very sensitive to the pseudo-periodic inputs of ice rafted detritus which characterise the glacial North Atlantic (the 'Heinrich layers' (Heinrich, 1988; Bond et al., 1992; Broecker et al., 1992; Grousset et al., 1993)). The χ records were used for testing the quality of the stratigraphic correlation described above, but no tie points were defined by correlation of the χ records. Finally, the resulting uncertainty in the correlation is of the order of 3 cm ($\approx$ 0.5 kyear) during climatic transitions, but may reach $\approx$ 10 cm ($\approx$ 2–4 kyear) within climatic substages when the gray reflectance is constant.

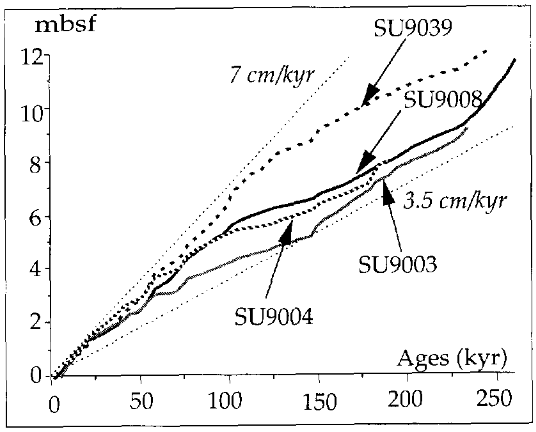

Fig. 3. Depth vs. age diagrams for the four studied cores. The average sedimentation rates are comprised between 3.5 and 7 cm per 1000 years.

The depth scales of the four cores were thus matched together and this common depth scale was converted into an absolute time-scale by reference to the SPECMAP $\delta^{18}O$ stack (Pisias et al., 1984). The sedimentation rates observed for the four cores are shown in Fig. 3; they average around 5 cm kyear^{-1} with minima around 3 and maxima around 10 cm kyear^{-1}.

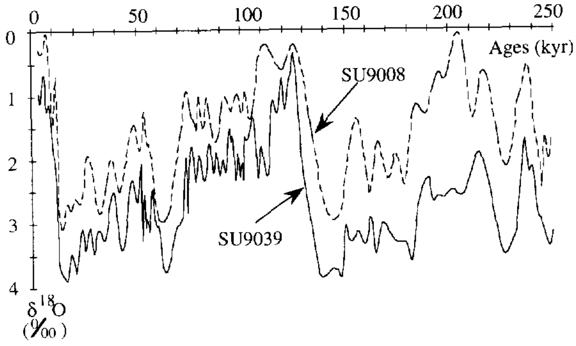

Fig. 2. Variations of the oxygen isotopic ratio of two planktonic foraminifera with time in cores SU9008 and SU9039. The curves were first obtained as a function of depth and were then transferred to ages using the SPECMAP isotopic record (Pisias et al., 1984).

4. Rock magnetic properties and measurements of the NRM

Stepwise acquisition of isothermal remanent magnetisation (IRM) was carried out on both u-channels and standard cubic samples. The u-channels were passed through the poles of an electromagnet and measured after each step. Cubic samples were magnetised using a 2G pulse magnetiser. The results obtained by the two methods are very consistent. Complete saturation of the IRM (SIRM) was observed in fields of 0.1–0.3 T. The u-channels were then stepwise demagnetised using alternating field (AF) methods in order to determine the median destructive field (MDF) of the SIRM. After a new acquisition of the SIRM, the u-channels were again passed through the magnet and increasing fields were applied stepwise in the opposite direction, to measure the coercivity of the remanence, Hcr, (defined as the backfield which reduces the SIRM to zero). Values of the MDF were always lower than 25 mT, and those of Hcr lower than 55 mT. These values are consistent with the presence of magnetite but cannot exclude the possible presence of iron sulfides which are characterised by similar values of these parameters.

Magnetic extracts from different horizons of the cores were obtained by circulating a slurry (usually the contents of five or six cubic samples) between the poles of a high gradient electromagnet. Thermomagnetic analysis of the extract, using a horizontal Curie balance, revealed the presence of a single magnetic phase with Curie temperatures varying between 550°C and 580°C, characteristic of low Ti-content magnetites (Fig. 4(a)). The non-reversible character of the curve during cooling is probably a result of the oxidation of magnetite. Oxygen appears to be always present in small amounts in the extracts, and possibly also in the measurement cell despite the use of a nitrogen atmosphere. Magnetite was also unambiguously identified by X-ray diffraction analyses which revealed characteristic peaks at 2.53, 2.97, 2.09, 1.61 and 1.48 Å (Fig. 4(b)).

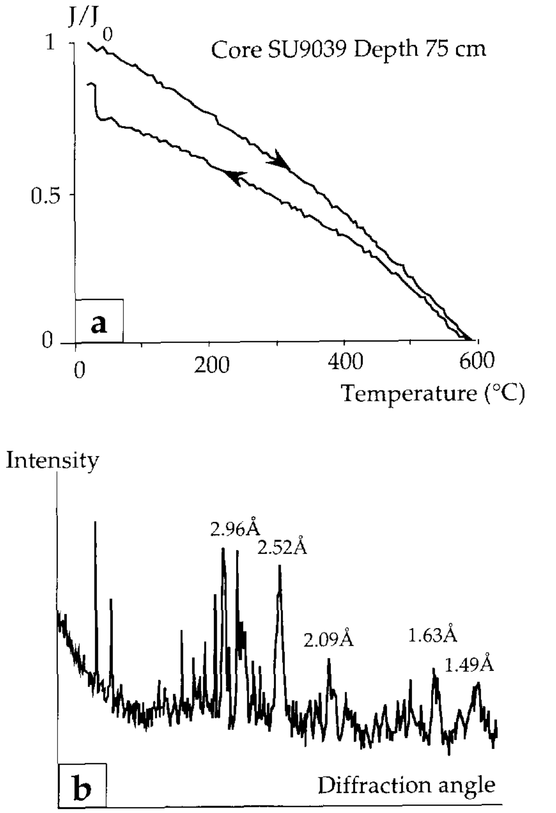

Fig. 4. (a) Representative thermomagnetic curve obtained from magnetic extracts with a horizontal Curie balance in a nitrogen atmosphere, and (b) characteristic X-ray diffraction spectrum of a magnetic extract. Both analyses are consistent with a magnetic mineralogy entirely dominated by magnetite.

In each core, the anisotropy of the low field susceptibility (AMS) has been analysed on cubic specimens sampled about every 50 cm with a Kappabridge KLY2. In most cases, the susceptibility tensor is accurately defined with a mean percentage of anisotropy of about 2.5%. The observed magnetic fabrics are oblate with minimum susceptibility axes ($K3$) perpendicular to the bedding plane and with $K1$ and $K2$ axes randomly oriented within this plane with no clear magnetic lineation. The typical orientation of the three principal axes of the anisotropy ellipsoid is shown in Fig. 5 for core SU9039. This fabric is typical of low-energy sedimentary environments.

The measurements of the NRM were made in a μ-metal shielded room. The u-channels were stepped through the pick-up coil system and readings were taken at 1 cm intervals. Using the deconvolution method of Constable and Parker (1991), one can attain a spatial resolution of ≈ 2.5 cm, comparable with that obtained from conventional cubic samples (Weeks et al., 1993). However, the intercorrelation between the different cores could not be obtained to better than 5–10 cm, depending on the particular climatic stages, as explained above. So the resolution of 2.5 cm is not attained in the final stack. We have thus used the data prior to deconvolution which provide a resolution of about 4 cm and do not contain high frequency components inevitably introduced by the deconvolution process.

In-line demagnetising coils were used for AF demagnetisation. A routine, computer-controlled,

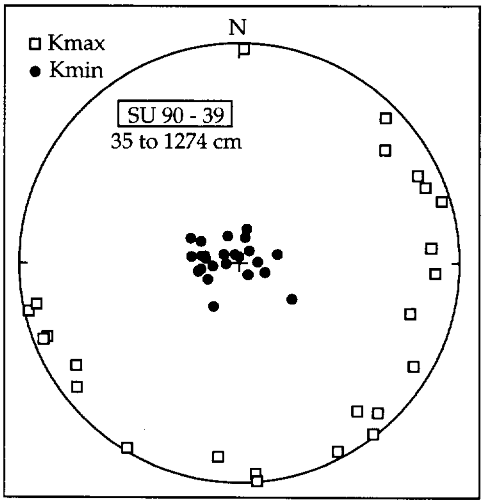

Fig. 5. Equal area projection of the magnetic fabric observed in core SU9039. This fabric, with the $K1$ and $K2$ axes randomly oriented in the horizontal plane is typical of low energy environment.

stepwise technique was used throughout, with at least six steps of demagnetisation up to 40–50 mT. The directional stability during demagnetisation was assessed by generating demagnetisation plots every 5 cm. This was possible because of the high reproducibility of the core position at each demagnetisation step (better than 1 mm). Some of the demagnetisation plots are shown in Fig. 6, together with plots from discrete 8 cm^3 samples from the same stratigraphic height. In general these plots are of very good quality. Apart from a small viscous component which is removed at about 15 mT, in most cases a single stable component of magnetisation is recognised with a median destructive field in the range 20–40 mT.

In the early stages of this study (for core SU9003), this routine involved four steps of demagnetisation up to 21 mT, which was the maximum amplitude of the AF-field available at the time. Technical improvements have increased the maximum available field and the routine has been changed to six steps of demagnetisation up to 40 mT. In one case (SU9004) pilot cubic samples taken every 15 cm were first demagnetised with at least seven steps of demagnetisation in order to select the steps of demagnetisation for the continuous measurements. U-channels from the same core were then demagnetised using five steps up to 25 mT. Although even 40 mT was sometimes insufficient to completely remove the NRM, it was possible in all cases to check whether the intensity decreases linearly to the origin, accurately defining the direction of the stable component of magnetisation. At certain depths, AF demagnetisation did not isolate a high coercivity component and the magnetisation did not decrease linearly to the origin. The last three or four points of the demagnetisation plots were then used as independent entries to calculate α_{95} values, which reflect the directional stability of the NRM demagnetisation. Values of α_{95} were calculated every 5 cm. The values are usually very low, of the order of 2–3°. For certain depths, however, sharp increases in the α_{95} values documented instabilities of the remanent magnetisation or the presence of high coercivity minerals.

The declination, inclination, and NRM (after demagnetisation at 250 mT) of the four cores are plotted against age in Fig. 7. In core SU9008, a clear change of the directions is observed at about 180–190 kyear. The demagnetisation diagrams obtained from this zone isolate a reverse component during the last two or three steps of the demagnetisation, after removal of a normal component at about 25 mT. Because this change is observed both in declination and in inclination, it may reflect a geomagnetic excursion (Biwa I event?) (Champion et al., 1988). This observation needs, however, to be tested by further investigations.

The ratio of the susceptibility of anhysteretic magnetisation (ARM) vs. χ was used to investigate downcore changes of the grain size of the magnetic minerals (King et al., 1982, 1983) directly using u-channels (Weeks et al., 1993). ARM was produced in the u-channels using a Schonstedt AF demagnetising coil within a large set of Ruben's coils (1 m each side). These coils provide a region of constant field, approximately 40 cm long, along the same axis as, but towards one end of the AF coils. The u-channels were pushed through the AF coils, which were kept at a constant alternating field, into, and through, the region of constant field. The ARM is therefore acquired during the movement of the u-channel through the system: as the AF field decreases with distance from the AF coils the u-channels experience the constant field throughout the major decrease in AF field intensity (Weeks et al., 1993). The constant biasing field was approximately 0.05 mT and maximum AF fields of 70 mT were used. Although higher AF fields are generally desirable, we do not think that our results would be significantly altered by inclusion of the highest coercivity fraction. The plots of the ARM, of χ and of the ARM/χ ratios as a function of age are shown in Fig. 8 for the four cores. The plots of ARM vs. χ are shown in Fig. 9. In these plots points that lie along a straight line through the origin indicate variation in concentration of grain with constant size, while changes in slope of the line indicate variation in grain size, if the magnetic mineralogy is homogeneous. For each core, we show in the upper case the results observed when the entire length of the core is considered, in the lower case those observed after rejection of the intervals in which the ARM/χ ratios exceed a factor of 5 (see below).

5. The records of normalised field intensity

Following the initial suggestions by Levi and Banerjee (1976) and King et al. (1982), the ARM normalising method has been widely used in this kind of study. The initial criteria given by these

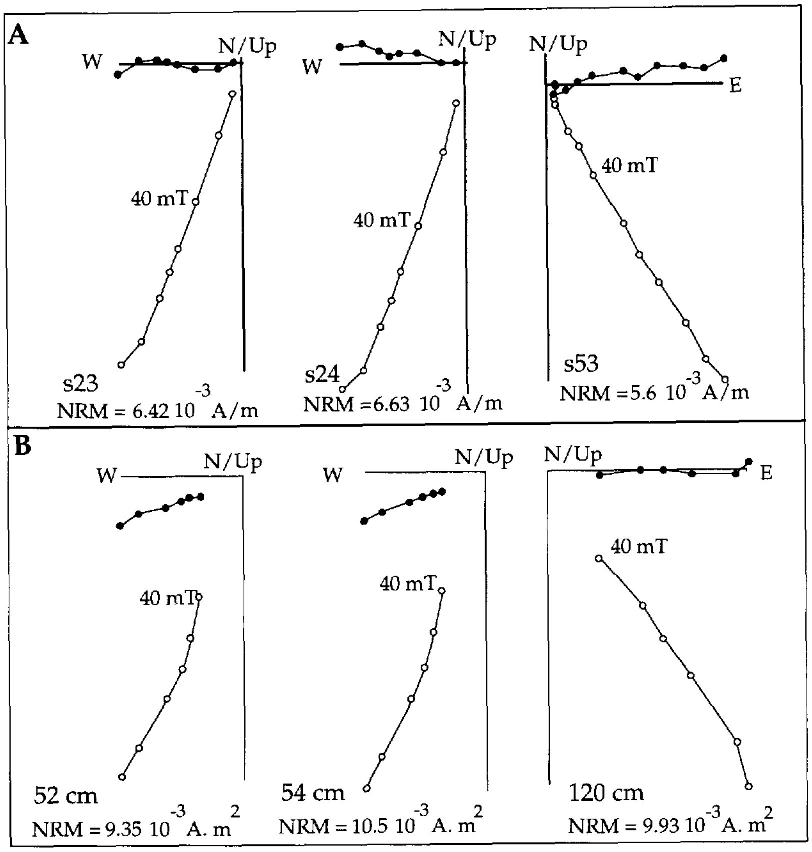

Fig. 6. Demagnetisation diagrams obtained from core SU9008: (A) from cubic samples; (B) from the equivalent depths in the u-channels (non-deconvoluted signal). The black and white dots refer to the projection on to horizontal and vertical planes, respectively. The results obtained with the much faster u-channel technique are quite consistent with those obtained from single samples, with virtually identical spatial resolution.

authors for magnetic uniformity to which the sediments should conform for the normalising procedure to be safely applied, are: (1) magnetite should be identified as the main carrier of magnetic remanence; (2) magnetite grain size should be between 1–15 μm; (3) magnetite concentration should not vary by more than a factor of 20–30. These criteria have been subsequently improved by different authors (King et al., 1983; Tauxe and Valet, 1989; Tauxe, 1993). Tauxe and Wu (1990), Tric et al. (1992a) and Tauxe (1993) have employed a more restrictive approach on point (3) and suggested that changes in concentration of the magnetic mineral should not exceed a factor of 10. Meynadier et al. (1992) have considered that ARM/χ should not vary of more than a factor of ± 2.5 with respect to the mean value. In this work we consider these criteria and use both ARM and saturation IRM (SIRM) as normalising parameters to attempt to obtain semicontinuous records of normalised field intensity.

The rock-magnetic analyses discussed above show that these criteria are not all met simultaneously over the entire length of the four cores. However, they are met over a significant fraction of the total length of each core. For instance, values of ARM and of ARM/χ do not change by more than a factor of ± 2.5 (which is the selection criteria of Meynadier et al. (1992)) over 87%, 85%, 75%, and 83% of the total length of cores SU9003, SU9004, SU9008, and SU9039, respectively. We have then attempted to obtain a semicontinuous record of the field intensity using only those portions of the cores where these criteria were met. In addition, we excluded those sections where the α_{95} values defined above exceeded a factor of 10. The rejected sections are shown by the shaded areas on the ARM and ARM/χ curves vs. age in Fig. 8. These sections include, in

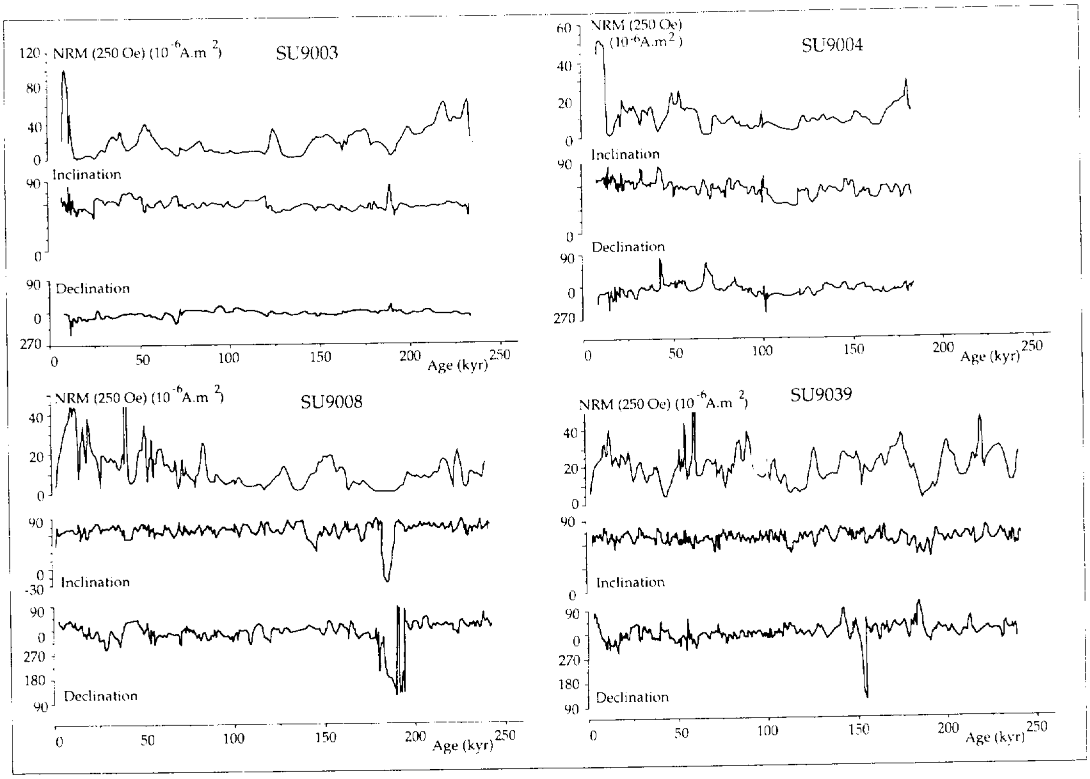

Fig. 7. Declination, inclination and NRM (after demagnetisation at 250 mT) values observed in the four cores, plotted vs. age.

particular, the sediment associated with the Heinrich peaks (Heinrich, 1988; Bond et al., 1992; Broecker et al., 1992; Grousset et al., 1993), where most of the high α_{95} values were documented from the directional data.

The records of normalised NRM intensity were obtained using values of NRM, ARM and SIRM after AF demagnetisation at 25 mT, i.e. after removal of the small secondary component present in the NRM. The results are shown in Fig. 10, which gives the normalised intensity signal NRM/ARM and NRM/SIRM as a function of age for the four cores, obtained after rejection of the portion of the cores where the selection criteria are not met.

Before these normalised records can be considered representative of the changes of the geomagnetic field, one has to verify, as a basic requirement, that the normalising process has removed the influence of environmental factors. It

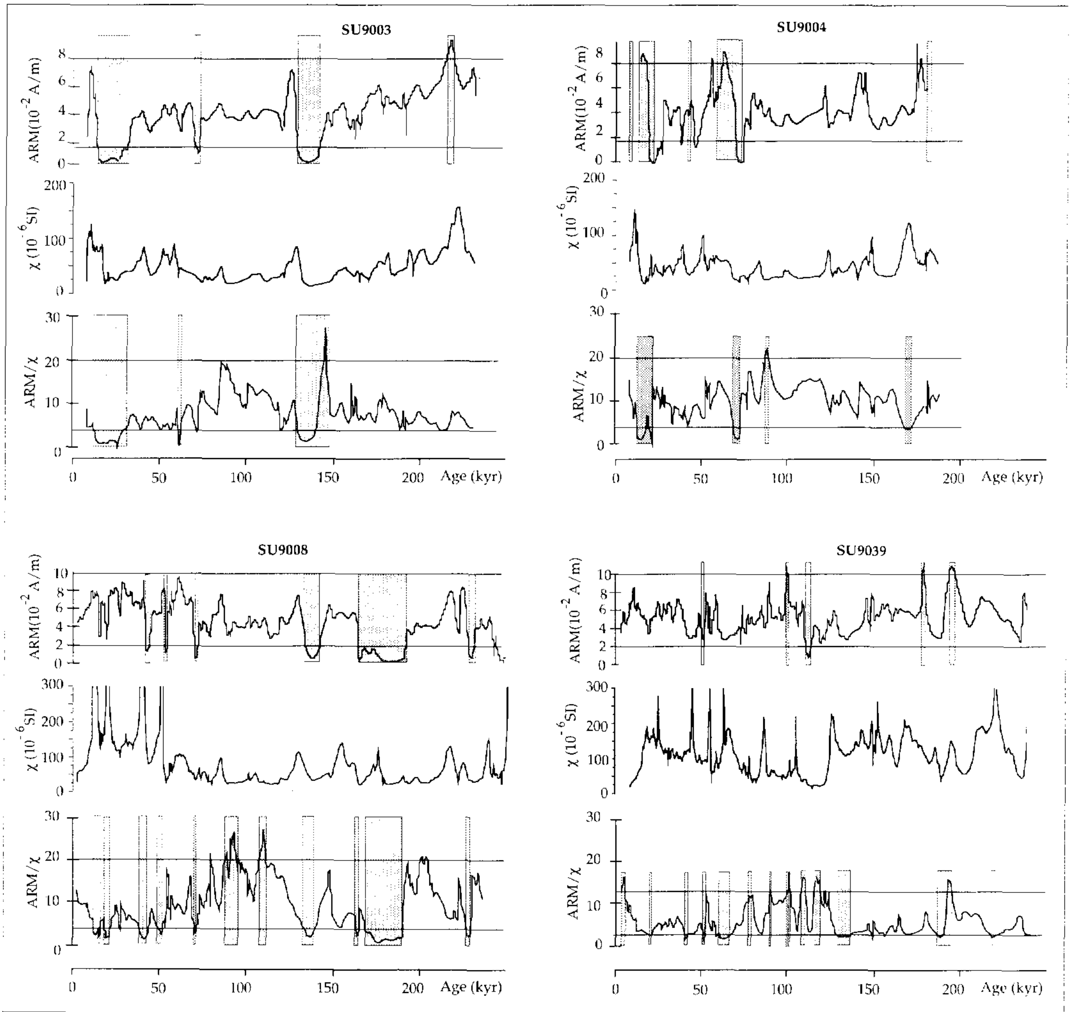

Fig. 8. Records of the variations of the ARM, of the magnetic susceptibility χ, and of the ratio ARM/χ for the four studied cores as a function of time. In the ARM/χ and ARM curves vs. age, horizontal lines mark the limits within which the values do not vary by more than a factor of ± 2.5 with respect to the mean value. The zones outside these lines are clearly identified in gray. These zones correspond to core intervals which are not suitable for paleointensity determinations.

is known that χ and NRM may largely reflect depositional conditions. Therefore, if the normalised intensity and the low field susceptibility records are largely correlated, then it is reasonable to conclude that normalisation has not successfully removed climatically controlled factors from the NRM (Tauxe and Wu, 1990). Conversely, the lack of correlation is an indication that the normalised intensity changes arise mainly from changes of the magnetic field (Tauxe and Wu, 1990).

Following Tauxe and Wu (1990), we have therefore calculated for each core the squared coherence function vs. frequency between the normalised intensity signals NRM/ARM and NRM/SIRM and the low field susceptibility, χ. The squared coherence functions vs. frequency between the normalised intensity signal NRM/

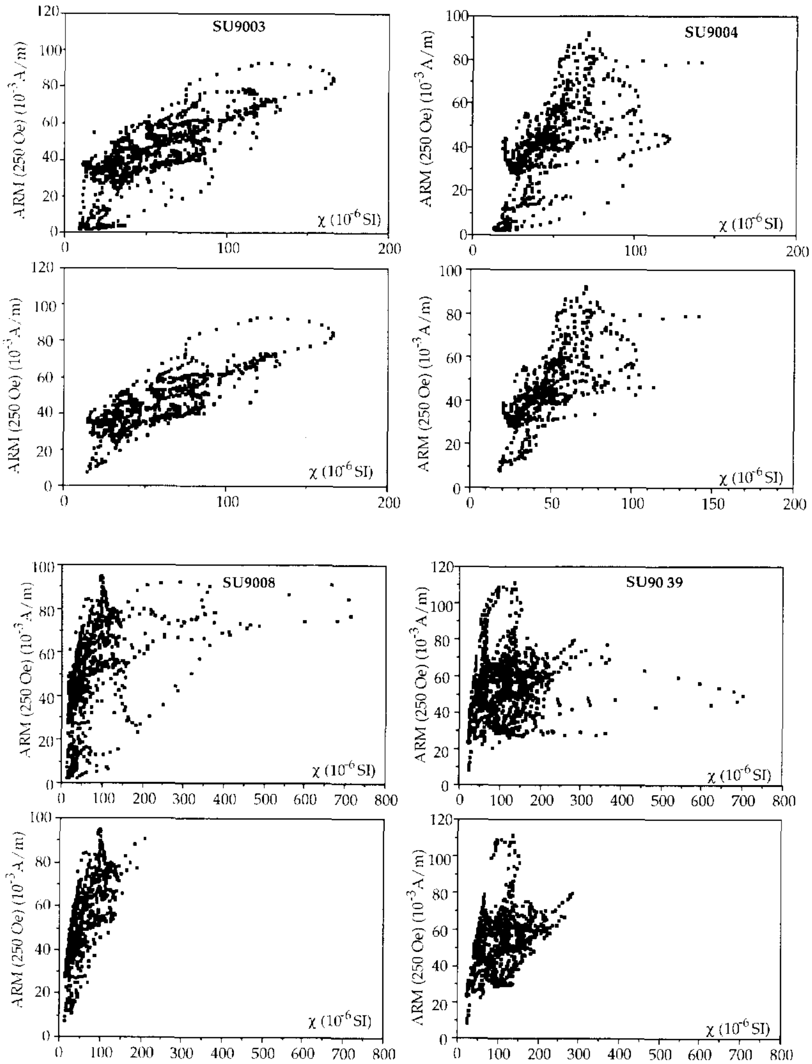

Fig. 9. variations of ARM as a function of χ showing the evolution of the grain size. For each core we show in upper case the results observed when the entire core is considered, in lower case the effect of rejecting intervals characterised by ARM/χ > ± 2.5 the mean value.

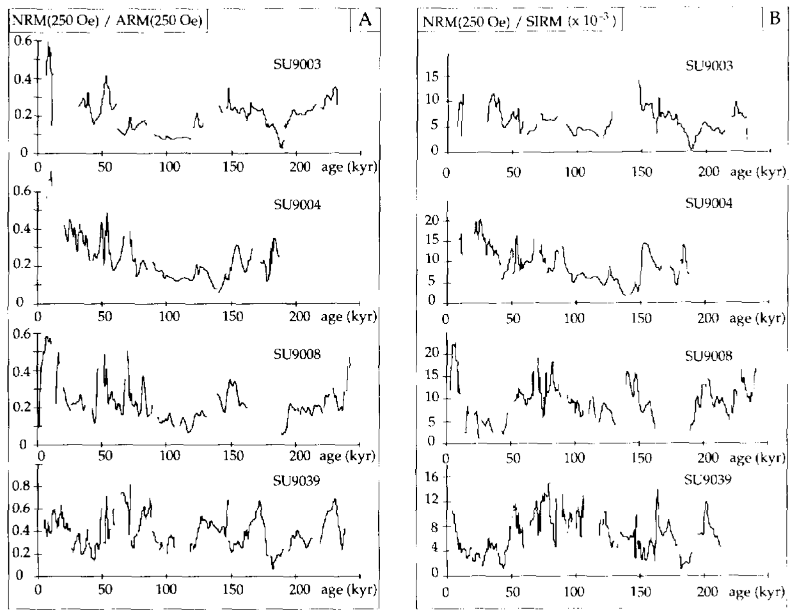

Fig. 10. Individual records of the normalised intensity vs. age, obtained from the selected intervals of the four cores (see text). In (A) and (B), ARM and SIRM are used as normalising factors of the NRM, respectively.

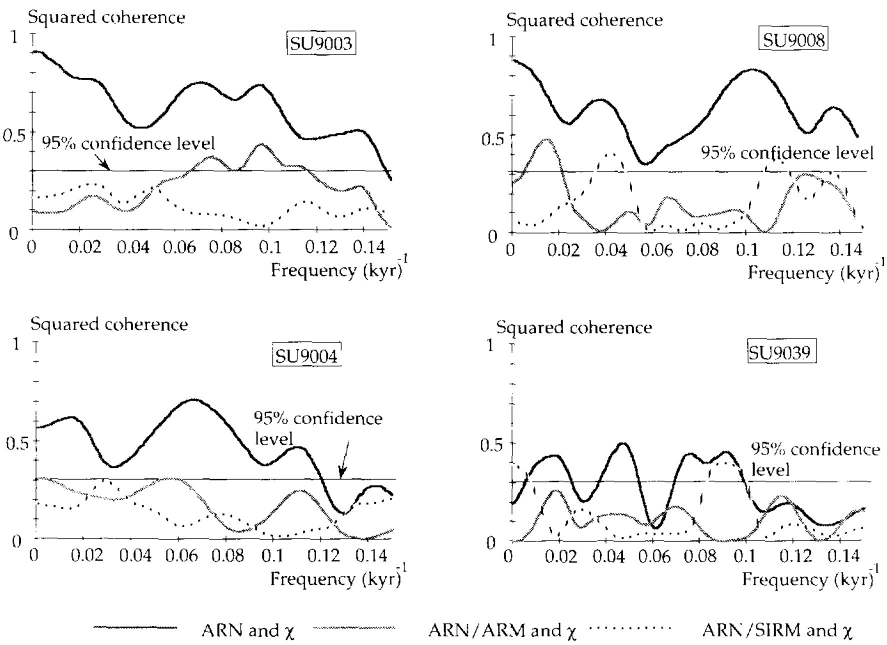

Fig. 11. Squared coherence functions vs. frequency between the NRM and the low field susceptibility and between the normalised intensity signal (NRM/ARM) and (NRM/SIRM) and the low field susceptibility. The NRM and the low field susceptibility appear significantly correlated over the entire frequency range considered. On the other hand, normalisation has lowered the coherence below the 95% significance level over most of the range. This is particularly true when IRM is used as a normalising factor. Overall the climatic induced contribution has been largely removed from the signal.

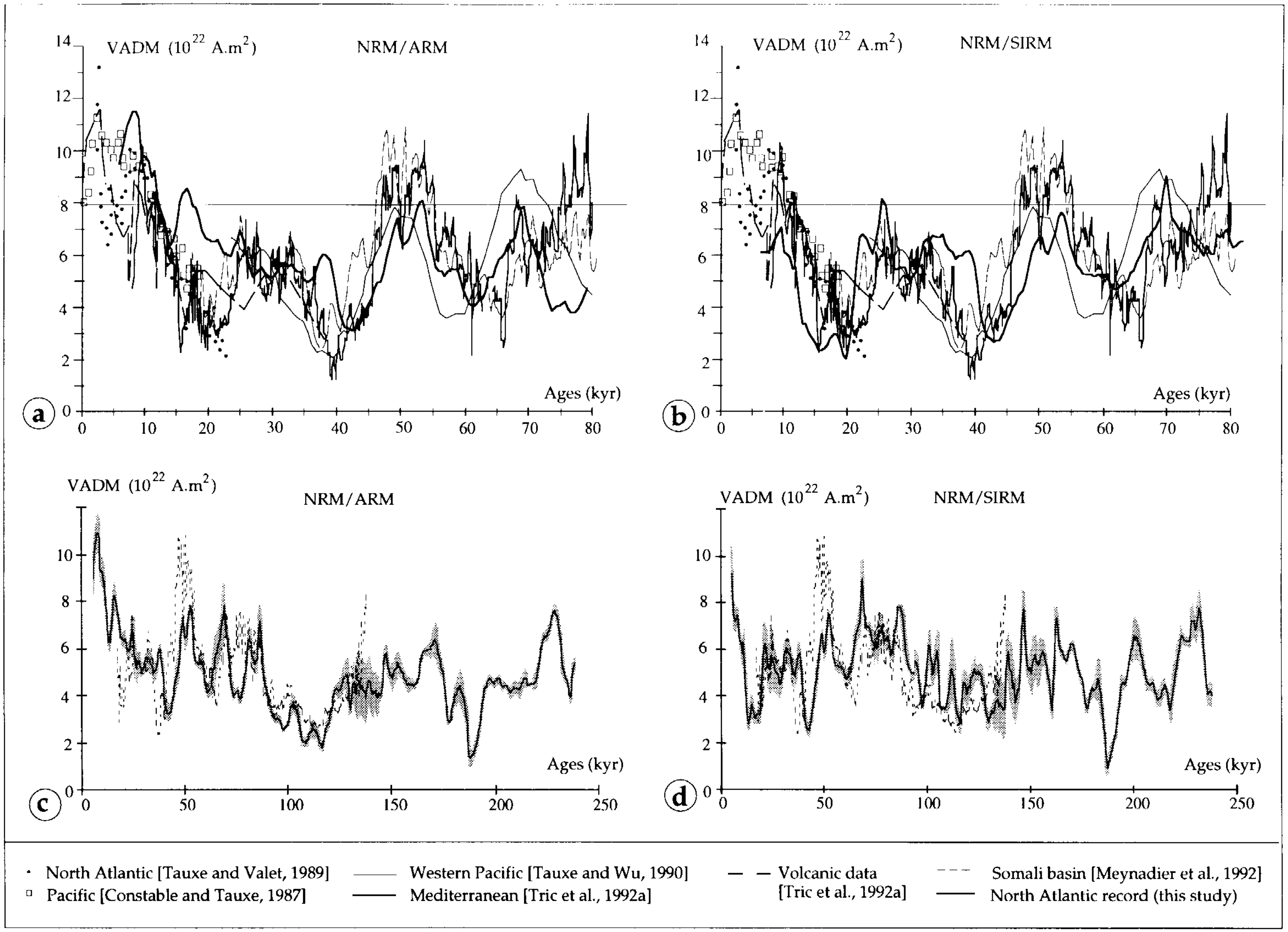

Fig. 12. Variations of the Virtual Axial Dipole Moment values vs. age for both normalisations (NRM/ARM and NRM/SIRM). (a, b) Enlargement of the curves between 0 and 80 kyear, together with those already reported for the same period from different geographic areas. (c, d) Records obtained for the past 240 kyear from this study, together with the curve obtained for the past 140 kyear by Meynadier et al. (1992).

ARM and NRM/SIRM and the low field susceptibility, together with the squared coherence function between the NRM and the low field susceptibility are shown in Fig. 11. In order to reduce any bias resulting from spectral leakage, the data series were multiplied by a taper — here a Bartlett window — before spectral calculations were performed (Jenkins and Watts, 1968; Constable and Parker, 1991). For the four cores, almost all values of squared coherence between NRM/ARM or NRM/SIRM and χ lie below the 95% confidence level. Only a few values lie slightly over the confidence level for core SU9003 — at a frequency of $\approx$ 10 kyear, and for the core SU9008 — at a frequency of $\approx$ 60 kyear. Values of the squared coherence between NRM/ARM or NRM/SIRM and χ, all lie largely below the values of calculated squared coherence between χ and NRM. This is a clear indication that normalisation has been largely successful in removing the paleoenvironmental influence from the NRM of the selected segments of the four cores.

We have then stacked the individual NRM/ARM and NRM/SIRM records obtained from each core to obtain a more accurate documentation of the relative changes of the geomagnetic field intensity over the past 240 kyear. The following procedure has been followed. First, to give equal weight to each record in the final stack, we have multiplied each individual NRM/ARM or NRM/SIRM record by a slightly different factor so that the same mean value of the normalised intensity calculated over its entire length is observed for each core. The main differences in this mean value are observed between the three cores SU9003, SU9004, and SU9008 and core SU9039. This last core has a higher value of the mean normalised intensity. It was sampled from a zone further north than the three other cores, so this difference probably reflects different environmental conditions at deposition. Fundamentally, it illustrates the fact that only relative changes of field intensity can be obtained from marine sediments. The records of the four cores were stacked into a single signal, using a 2 kyear sliding window which was moved in steps of 0.5 kyear. In the stratigraphic intervals where the magnetic properties of a particular core were not suitable for normalisation, we combined data from three and occasionally only two cores.

The two records of normalised intensity obtained with ARM and SIRM normalisation were then separately adjusted to the profiles of the Earth's magnetic field intensity from the Mediterranean (Tric et al., 1992a), the Somali Basin (Meynadier et al., 1992), and the Pacific (core ERDC 113p) (Tauxe and Wu, 1990). The record of Tric et al. (1992a) has been calibrated using absolute determinations of the geomagnetic field intensity obtained from volcanic rocks over the period 10–40 kyear, and both the Somali Basin and the Pacific records are largely consistent with it. This is an indication that these three records yield a rather precise picture of the changes of the field intensity over at least the last 80 kyear. We have multiplied the two stacked records from the North Atlantic by a constant factor, chosen so that the obtained profiles have the same mean value as the Mediterranean, the Somali Basin and the Pacific records in the interval 40–80 kyear (see below why the interval 0–40 kyear was not considered).

The final results, which represent our best approximation of the relative variations of the geomagnetic field intensity over the last 240 kyear using the two normalisation parameters, are shown in Fig. 12. In this figure, for each method of normalisation we have considered first the last 80 kyear (Figs. 12(a) and (b)), then the total interval 0–240 kyear (Figs. 12(c) and (d)). The results from the Mediterranean, the Somali Basin and the Pacific are also reported in this figure for comparison.

6. Discussion

We have critically examined the results obtained with the two methods of normalisation employed for the North Atlantic cores. The stacked records of Fig. 12 show that, apart from some high frequency fluctuations present in the SIRM normalised record, the two methods of normalisation yield quite consistent results except in the 0–30 kyear and in the 90–145 kyear intervals.

In the first (0–30 kyear) interval (Figs. 12(a) and (b)) when using SIRM, we observe a slightly higher scatter of the data (around 25 kyear) in the final stack, arising from some differences between the individual normalised records of the four cores. More significantly, the stacked record obtained with SIRM lies below the record obtained with ARM normalisation. A clear drop in intensity is observed around 15–20 kyear in the first case, while a maximum is observed in the ARM normalised record. In this interval the SIRM normalised record from the North Atlantic appears more consistent with the absolute field intensity determination and the record from the Mediterranean, than the record obtained using ARM normalisation (Figs. 12(a) and (b)).

An intensity maximum very similar and coeval to the one observed in the North Atlantic with ARM normalisation, is observed in the lacustrine record from Lac du Bouchet (Thouveny et al., 1993) while other lacustrine or maar records in the Mediterranean realm yield a marked low (N. Thouveny and D. Williamson, personal communication, 1993). These continental records were all obtained using ARM normalisation. Given the proximity of some of the sites from which conflicting results are obtained, the differences observed in the different records cannot arise from local difference in the magnetic field intensity. Most probably, they reflect differences in the rock magnetic properties of the sediments, which react differently to one or the other method of normalisation. Changes in the ARM of the four cores from the North Atlantic are observed in this interval, probably related to changes in the amount of magnetite. It is possible that the normalisation by the ARM did not entirely remove this effect, because ARM acquisition is not linear with the applied field. Basically, however, and despite the detailed rock magnetic analysis of these North Atlantic cores, we have no reasonable physical idea of why such differences are observed between the two methods of normalisation in this particular stratigraphic interval. This interval, which corresponds to the last glacial maximum and the beginning of deglaciation and thus to drastically changing environmental conditions, is the object of more detailed studies.

On the basis of its higher consistency with the volcanic results, we have tentatively chosen the SIRM normalised record as more representative of the geomagnetic field variation in the North Atlantic than the ARM normalised record. Again, we emphasise that a significant difference is observed only around 15–20 kyear (Figs. 12(a) and (b)).

The record from the North Atlantic shows large similarities with the other published records. All the available records, for instance, document a marked intensity low at $\approx$ 38–43 kyear, and a maximum at $\approx$ 50 kyear. This low, which is coeval with the Laschamp event (Chauvin et al., 1989), appears less pronounced, and slightly displaced towards older ages (42 kyear) in the North Atlantic record compared with the other records ($\approx$ 40 kyear) but this difference is well within dating accuracy. The 60 kyear low observed in the record from North Atlantic appears at 58 kyear in the Pacific record (core ERDC113p) (Tauxe and Wu, 1990), and at $\approx$ 65 kyear in the data from the Mediterranean (Tric et al., 1992a) and from the Somali Basin (Meynadier et al., 1992). There is also a maximum at 70 kyear in the North Atlantic and Pacific records, which appears to occur around 78 kyear in the Mediterranean and Somalia records.

For older periods, the low intensity zone documented by SIRM normalisation between 90 and 145 kyear appears broader and noisier than the intensity low obtained from ARM normalisation. The ARM normalisation appears to be more consistent with the Somali Basin record (Figs. 12(c) and (d)). The broad low has already been observed at 110–130 kyear in sediments from the Somali Basin (Meynadier et al., 1992), although the end of the low intensity period is not exactly synchronous in the two records. This period embraces the reported dates for the Blake event. This broad low is followed by a large high from 125 to 180 kyear. Both the 90–110 kyear low and the 125–180 kyear high are consistent with the paleointensity values recently obtained from lava flows (Tric et al., 1992b).

Finally, a marked low is documented at 190 kyear. During this interval the intensity drops to the minimum value observed over the entire time interval explored by the four cores. A similar low at 190 kyear is observed in another record of normalised NRM from the Western Equatorial Pacific (Yamazaki and Ioka, 1992) and has been ascribed to the Biwa I event.

A link between changes of the geomagnetic field intensity and the Earth's orbital parameters has been considered by some authors (Kent and Opdyke, 1977; Negi and Tiwari, 1984; Tauxe and Wu, 1990; Tric et al., 1992a; Meynadier et al., 1992). Visual examination of the record from the North Atlantic apparently does not support any claim for periodicity or stationarity over the observed interval. More quantitatively, we have made a spectral analysis of the records of normalised intensity, using an autocorrelation power spectrum algorithm successfully used for time series analysis in paleoclimatic studies (Pestiaux and Berger, 1982, 1984). The results (Fig. 13) show that while the first 80 kyear undeniably suggest some oscillatory behavior (frequency peak around 0.06, i.e. period of 17 kyear), the rest of the record does not sustain this hypothesis. Fundamentally, there is no evidence that the signal is stationary, and it is certainly premature to conclude from these data that there is a significant periodicity in the pattern of geomagnetic field intensity changes.

7. Conclusions

This paper is an attempt to derive a reliable record of the relative geomagnetic field intensity changes from a region characterised by a low biologic productivity but also by rather drastic

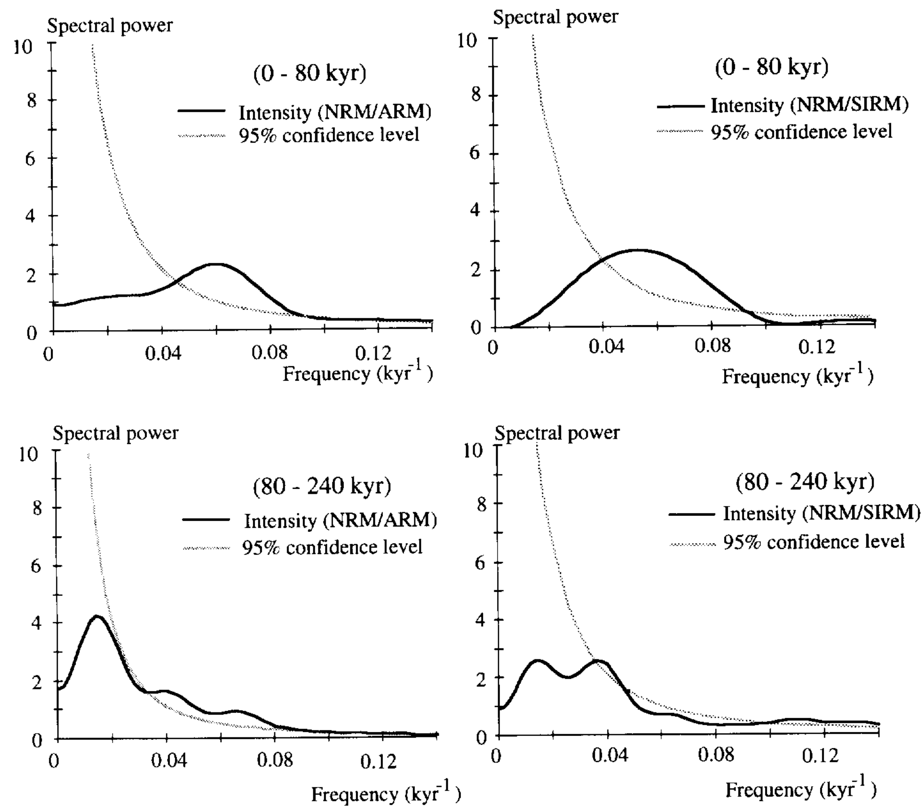

Fig. 13. Power spectra of the NRM/ARM and NRM/SIRM obtained for the two periods 0–80 kyear and 80–240 kyear using the algorithm of Pestiaux and Berger (1982, 1984).

changes in fluxes of detrital minerals. As not all the stratigraphic intervals met the criteria for normalisation, we have restricted our analysis to only those sections which appear suitable for paleointensity determinations using published selection criteria. Although selection of much more homogeneous sediments is clearly desirable, this procedure appears to provide results consistent with those previously obtained by other authors in different geographical areas, and appears encouraging for further studies of geomagnetic paleointensity in regions where homogeneous sediments are not available.

The choice of saturation IRM as a suitable normalising parameter for NRM has been done a posteriori on the basis of the comparison with absolute determinations of paleointensity obtained from well dated volcanic rocks spanning only a fraction of the time interval explored here. The consistency of our results with those obtained by different authors in different geographical area is a good indication that we have successfully extracted a real geomagnetic signal from the North Atlantic sediments. Nevertheless, these results clearly draw attention to the necessity of a better understanding of the processes of sedimentary magnetisation acquisition as well as of the physics of the normalising processes such as ARM and IRM, before any method of extracting an independent record of the paleointensity of the geomagnetic field can be considered as thoroughly reliable. An effort in this sense, based on hysteresis loops characteristics, is now being carried on in our laboratory and will be the object of a future paper.

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