

## A 2.14-Myr astronomically tuned record of relative geomagnetic paleointensity from the western Philippine Sea

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[1] We present a 2.14-Myr astronomically tuned relative geomagnetic paleointensity record from the western Philippine Sea. Pseudosingle-domain titanomagnetite is the only magnetic mineral identified and variations in titanomagnetite concentration fall well within the accepted limits for relative paleointensity variations. No significant temporally persistent periodicities are observed in wavelet analyses of the paleointensity time series or in the rock magnetic parameters used for relative paleointensity normalization. This suggests that our paleointensity record is largely free of rock magnetic or lithological artefacts and that it represents a reliable record of geomagnetic behavior with no evidence for modulation of the field at Earth orbital periods. The paleointensity record is highly coherent with the Sint-800 global paleointensity stack for the last 800 kyr and with a coeval record from the West Caroline Basin. Our record confirms that it is normal for the geomagnetic field to undergo dynamic changes within polarity intervals, with relatively frequent collapses of the field to low paleointensities and concomitant deviations away from the stable field direction. We do not observe an asymmetrical sawtooth form in our paleointensity record, which might suggest that previously observed asymmetrical sawtooth paleointensities result from rock magnetic artefacts. Also, we do not observe a persistent 100-kyr inclination periodicity, in contrast to the suggestion that geomagnetic field directions are modulated by orbital eccentricity. Good agreement between our paleointensity record and the coeval West Caroline Basin record provides the beginning of a detailed view of geomagnetic field behavior between 0.8 and 2.14 Ma. **INDEX TERMS:** 1520 Geomagnetism and Paleomagnetism: Magnetostratigraphy; 1521 Geomagnetism and Paleomagnetism: Paleointensity; 1560 Geomagnetism and Paleomagnetism: Time variations—secular and long term; 9355 Information Related to Geographic Region: Pacific Ocean; **KEYWORDS:** geomagnetic paleointensity, magnetostratigraphy, spectral analysis, wavelet analysis, Philippine Sea

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### 1. Introduction

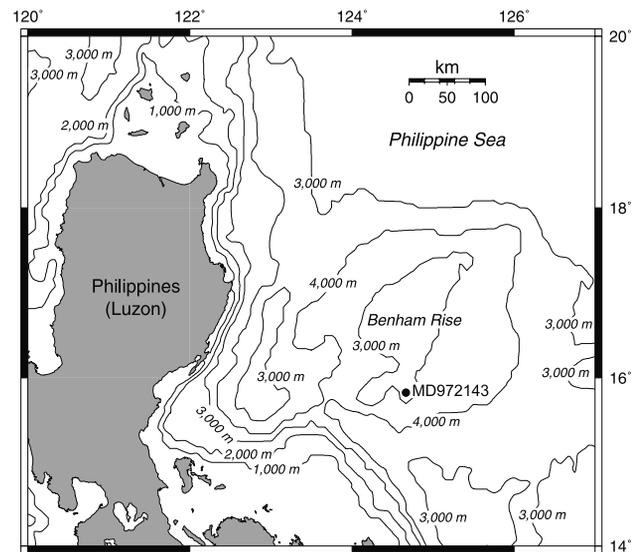
[2] Vector records of the geomagnetic field that span long periods of geological time are important for developing an understanding of the long-term behavior of the dynamo that generates the geomagnetic field. Deep-sea sedimentary sequences represent suitable targets for such studies because they are often deposited more continuously than other geological archives of geomagnetic information. Despite the ease with which the direction of magnetization can be determined for long sedimentary sequences, it has been,

until relatively recently, problematical to obtain full vector time series because the ancient field intensity is more difficult to accurately extract from sediments. In the last decade, there has been a great deal of activity directed toward determining the relative paleointensity of marine sedimentary sequences. Coupled with high-quality dating constraints usually provided by oxygen isotope analyses of foraminiferal calcite, it has been demonstrated that, on a global scale, there is considerable coherency among relative paleointensity records from geographically widely distributed sites [Tauxe and Wu, 1990; Tric et al., 1992; Meynadier et al., 1992; Schneider, 1993; Tauxe and Shackleton, 1994; Yamazaki and Ioka, 1994; Weeks et al., 1995; Stoner et al., 1995, 1998; Yamazaki et al., 1995; Lehman et al., 1996; Schneider and

Mello, 1996; Schwartz *et al.*, 1996; Guyodo and Valet, 1996, 1999; Roberts *et al.*, 1997; Channell *et al.*, 2000; Laj *et al.*, 2000; Guyodo *et al.*, 2001]. The similarity among these globally distributed paleointensity records suggests that the signal is dominated by the dipole component of the geomagnetic field. In addition to providing constraints on geodynamo behavior, such work has been pursued in order to provide a high-resolution magnetostratigraphic dating tool for the Brunhes Chron [e.g., Guyodo and Valet, 1996, 1999; Stoner *et al.*, 1998; Channell *et al.*, 2000; Laj *et al.*, 2000; Kiefer *et al.*, 2001] and to understand how geomagnetically modulated temporal variation of cosmogenic isotope production, such as  $^{14}\text{C}$  [Mazaud *et al.*, 1991; Laj *et al.*, 1996a] and  $^{10}\text{Be}$  [Frank *et al.*, 1997], can affect the use of dating techniques that employ cosmogenic isotopes. Time series analyses of relative paleointensity records indicate that periodicities associated with the Earth's orbit are present in some records [Channell *et al.*, 1998; Yamazaki, 1999] and that other significant periodicities are also present in some records [Tauxe and Wu, 1990; Tauxe and Shackleton, 1994]. However, spectral analysis using a more sophisticated wavelet technique suggests that the orbital periodicities observed for the last 1.1 Myr by Channell *et al.* [1998] are an expression of lithological variations and are not characteristic of the geodynamo [Guyodo *et al.*, 2000]. Spectral analysis of long geomagnetic time series is clearly important for testing such interpretations and for understanding the mechanisms that generate the geomagnetic field.

[3] Fluctuations in geomagnetic paleointensity are now well documented for the last 800 kyr, for which a stacked global paleointensity curve (Sint-800) has been produced [Guyodo and Valet, 1999]. A global paleointensity stack has also been constructed for the interval around the Jaramillo Subchron (0.95–1.10 Ma) [Guyodo *et al.*, 2001]. In order to understand the longer-term behavior of the geomagnetic field, it is necessary to obtain long records with high-resolution age control, but few well-constrained records are available beyond the Jaramillo Subchron. The records of Valet and Meynadier [1993] (hereinafter referred to as VM93) and Meynadier *et al.* [1994], from the equatorial Pacific and Indian Oceans, respectively, span the last 4 Myr. However, the reliability of their observed asymmetrical sawtooth paleointensity pattern has been called into question by numerous workers [e.g., Kok and Tauxe, 1996a, 1996b; Mazaud, 1996; McFadden and Merrill, 1997] and needs to be tested by obtaining further coeval paleointensity records. Several other longer records have been published, including high-resolution records from the California margin for the last 1.4 Myr [Guyodo *et al.*, 1999; Hayashida *et al.*, 1999], low-resolution records for the last 1.8 Myr from the central equatorial Pacific [Laj *et al.*, 1996b; Verosub *et al.*, 1996], and a low-resolution Ontong-Java Plateau paleointensity stack for the interval between 0.78 and 2.58 Ma [Kok and Tauxe, 1999]. Some of these records lack precise age control, with dating constrained only at the respective reversal boundaries. There is clearly a paucity of paleointensity records with high-resolution age control older than 1.1 Ma.

[4] In this paper, we present a 2.14-Myr paleointensity record from the western Philippine Sea, with age control based on astronomical tuning of high-quality  $\delta^{18}\text{O}$  data. The length of the record makes it suitable for developing a detailed view of long-term paleointensity variations which can



**Figure 1.** Location map of core MD972143 from Benham Rise, western Philippine Sea.

also be tested for orbital influences and for the presence or absence of asymmetrical sawtooth paleointensity behavior.

## 2. Geological Setting and Core Description

[5] Core MD972143 (length 38 m) was recovered from Benham Rise (15.87°N, 124.65°E) in the western Philippine Sea (Figure 1; water depth of 2989 m) during the IMAGES III cruise of the French R/V *Marion Dufresne* in 1997. The sediments consist mainly of hemipelagic calcareous oozes with intercalated tephra layers. The core top (0–0.12 m) is void, and cracks occur between subbottom depths of 0.78 and 0.87 m. A white foraminiferal sand layer, dominated by reworked fossils, exists between 32.91 and 33.95 m. The top of this sand layer marks the base of the paleointensity record presented in this paper.

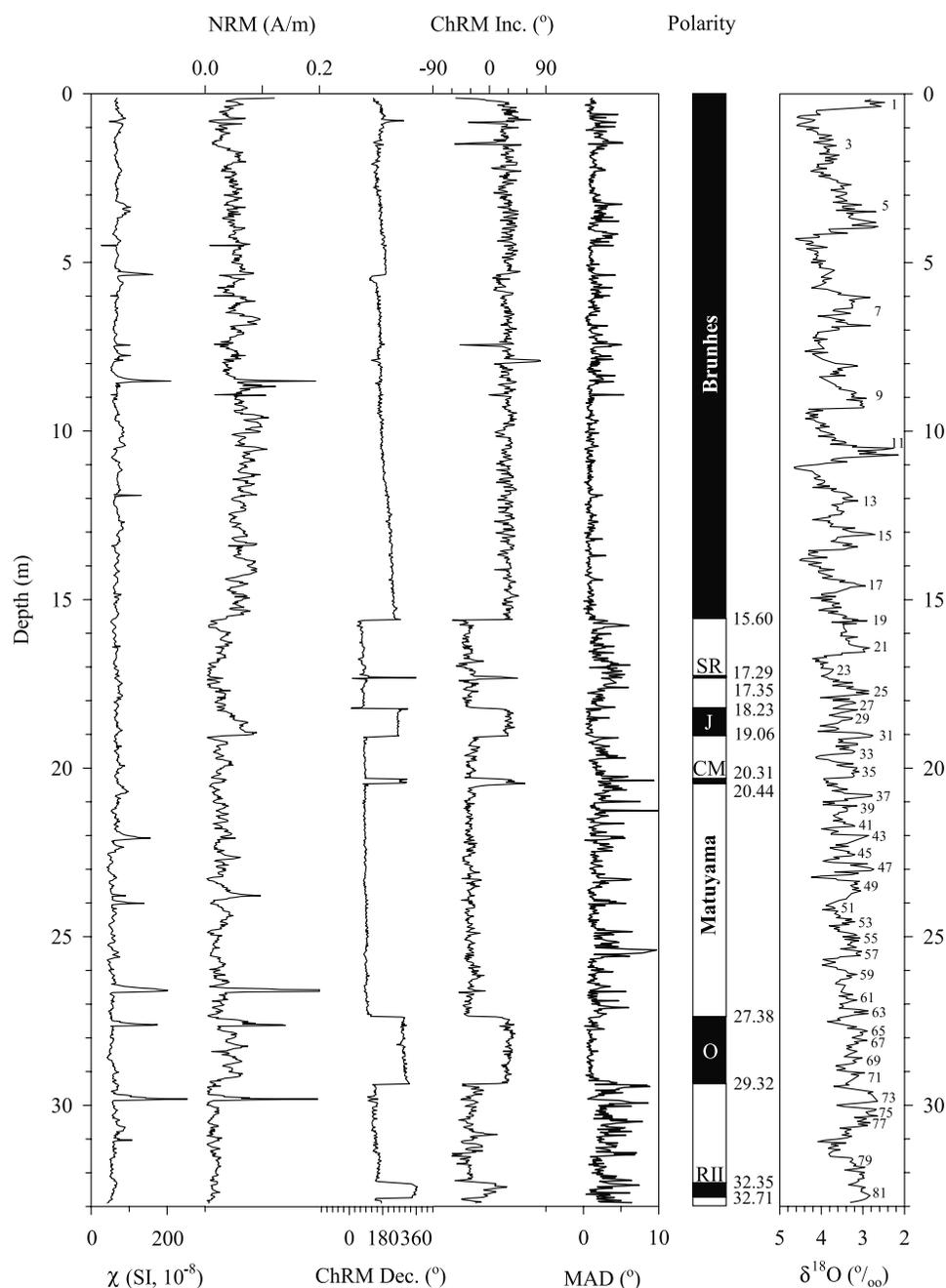
## 3. Methods

### 3.1. Paleomagnetic Measurements

[6] Paleomagnetic samples were obtained by pressing plastic cubes (7 cm<sup>3</sup>) into the sediments. A total of 1441 samples were continuously collected, with minimal gaps between samples down to the foraminiferal sand layer at 32.91 m. The natural remanent magnetization (NRM) was measured on all samples with a 2G Enterprises superconducting rock magnetometer. To examine the stability of the NRM, stepwise alternating field (AF) demagnetization was carried out on each sample using an in-line three-axis demagnetizer linked with the rock magnetometer. After measurement of the NRM, 12 demagnetization steps were measured between 5 and 80 mT, with a 5–10 mT increment at each step. Characteristic remanent magnetization (ChRM) directions were determined by performing a linear regression through the data for multiple demagnetization steps [Kirschvink, 1980].

### 3.2. Mineral Magnetic Measurements

[7] Several magnetic properties were measured to check whether the MD972143 sediments are suitable for relative



**Figure 2.** Magnetic properties and chronostratigraphic framework for core MD972143. From left to right, versus depth: low-field magnetic susceptibility ( $\chi$ ), intensity of the natural remanent magnetization (NRM) prior to demagnetization, declination and inclination, respectively, of the characteristic remanent magnetization (ChRM), which was calculated using data from multiple demagnetization steps, maximum angular deviation (MAD) values associated with the ChRM determinations, polarity (black, normal; white, reversed), with depths indicated for polarity boundaries (SR, Santa Rosa polarity interval; J, Jaramillo Subchron; CM, Cobb Mountain Subchron; O, Olduvai Subchron; and RII, Réunion II Subchron), and  $\delta^{18}\text{O}$  variations, with numbers indicating  $\delta^{18}\text{O}$  stages after Shackleton and Pisias [1985], Shackleton *et al.* [1990], and Shackleton *et al.* [1995]. Peaks in magnetic susceptibility and NRM represent volcanic ash layers.

paleointensity determinations. Low-field magnetic susceptibility ( $\chi$ ) was measured for all samples using a Bartington Instruments MS2 magnetic susceptibility meter. Anhyseretic remanent magnetizations (ARMs) were imparted to the z axis of all samples using an 80 mT AF and a 0.1 mT bias field.

Technical difficulties prevented acquisition of high-quality ARM data, which are therefore not used in this paper. Saturation isothermal remanent magnetizations (SIRMs) were imparted using a DC field of 1 T. The SIRMs were then subjected to AF demagnetization at peak fields of 10, 15 and

20 mT. Hysteresis parameters, including saturation magnetization ( $M_s$ ), saturation remanence ( $M_r$ ), coercive force ( $B_c$ ) and coercivity of remanence ( $B_{cr}$ ), were measured on 313 subsamples ( $\sim 30$  mg) up to maximum fields of 1 T using a Princeton Measurements Corporation Micromag Alternating Gradient Magnetometer (sensitivity of  $10^{-11}$  A m<sup>2</sup>).

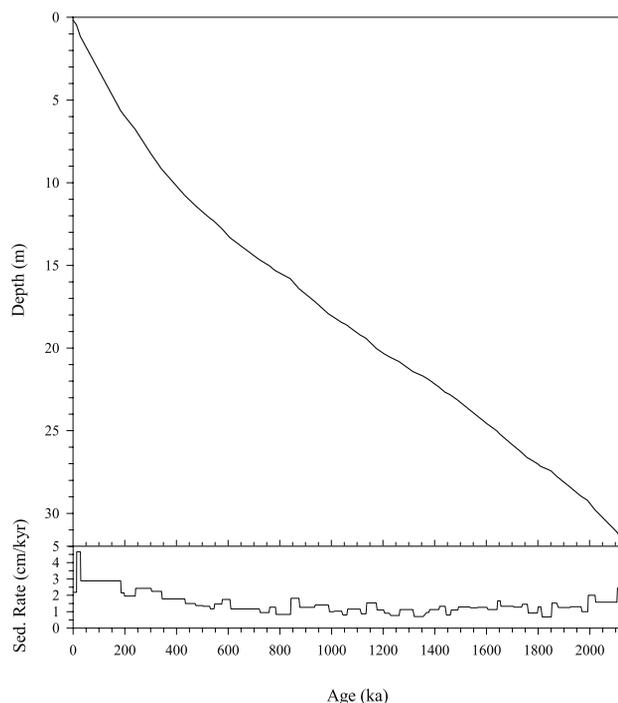
### 3.3. Spectral Analysis

[8] Wavelet analysis was used to test for the presence of significant spectral power within the relative paleointensity time series and within the rock magnetic parameters used for paleointensity normalization. For nonstationary time series, wavelet analysis is particularly useful for identifying periodicities that might be present at different times. In order to conduct wavelet analysis on a time series with irregular spacings between points, it is necessary to resample the time series with uniform spacing. This was achieved by calculating the average time spacing for the original data points and by resampling the record at this average time spacing. A red-noise background spectrum was assumed and was estimated from the best fit with the global spectrum. A Morlet wavelet was used as the basis function. Details of the wavelet analysis technique are given by *Torrence and Compo* [1998]. In addition, the software package “AnalySeries” [*Paillard et al.*, 1996] was used to perform cross-spectral analysis between our relative paleointensity record and the Sint-800 and VM93 records to delineate their spectral power, coherency and phase relationships. Before conducting cross-spectral analysis, each investigated time series was normalized to unit variance and resampled at a constant 2-kyr interval. The Blackman-Tukey technique with a Bartlett window type was used for cross-spectral analysis.

## 4. Results

### 4.1. Magnetic Polarity Stratigraphy and Age Control

[9] *Hornig et al.* [2002] developed a magnetic polarity stratigraphy and detailed  $\delta^{18}\text{O}$  stratigraphy for the MD972143 core in order to provide new astronomically calibrated age estimates for short polarity events in the Matuyama Chron. In this paper, we focus on paleointensity determinations for the MD972143 core. We briefly discuss relevant magnetostratigraphic and  $\delta^{18}\text{O}$  results, as presented by *Hornig et al.* [2002], in order to establish the suitability of the studied sediments for relative paleointensity determinations. The sediments recovered in core MD972143 generally have stable magnetizations, with 95% of samples showing clear univectorial decay to the origin of vector component diagrams. Secondary remanence components are generally small and a characteristic remanent magnetization (ChRM) was almost always identified after demagnetization at 10–20 mT. Using the ChRM data from the stepwise-demagnetized samples, it was possible to construct a clear magnetic polarity stratigraphy (Figure 2). The average ChRM inclinations in the normal and reversed polarity intervals are  $30.3^\circ$  and  $-30.7^\circ$ , respectively, which are not significantly different from the inclination expected for a geocentric axial dipole field at the site latitude ( $\pm 29.6^\circ$ ). The ChRM declinations clearly undergo progressive down-core changes (Figure 2), which probably indicates that the sediments twisted during core recovery.

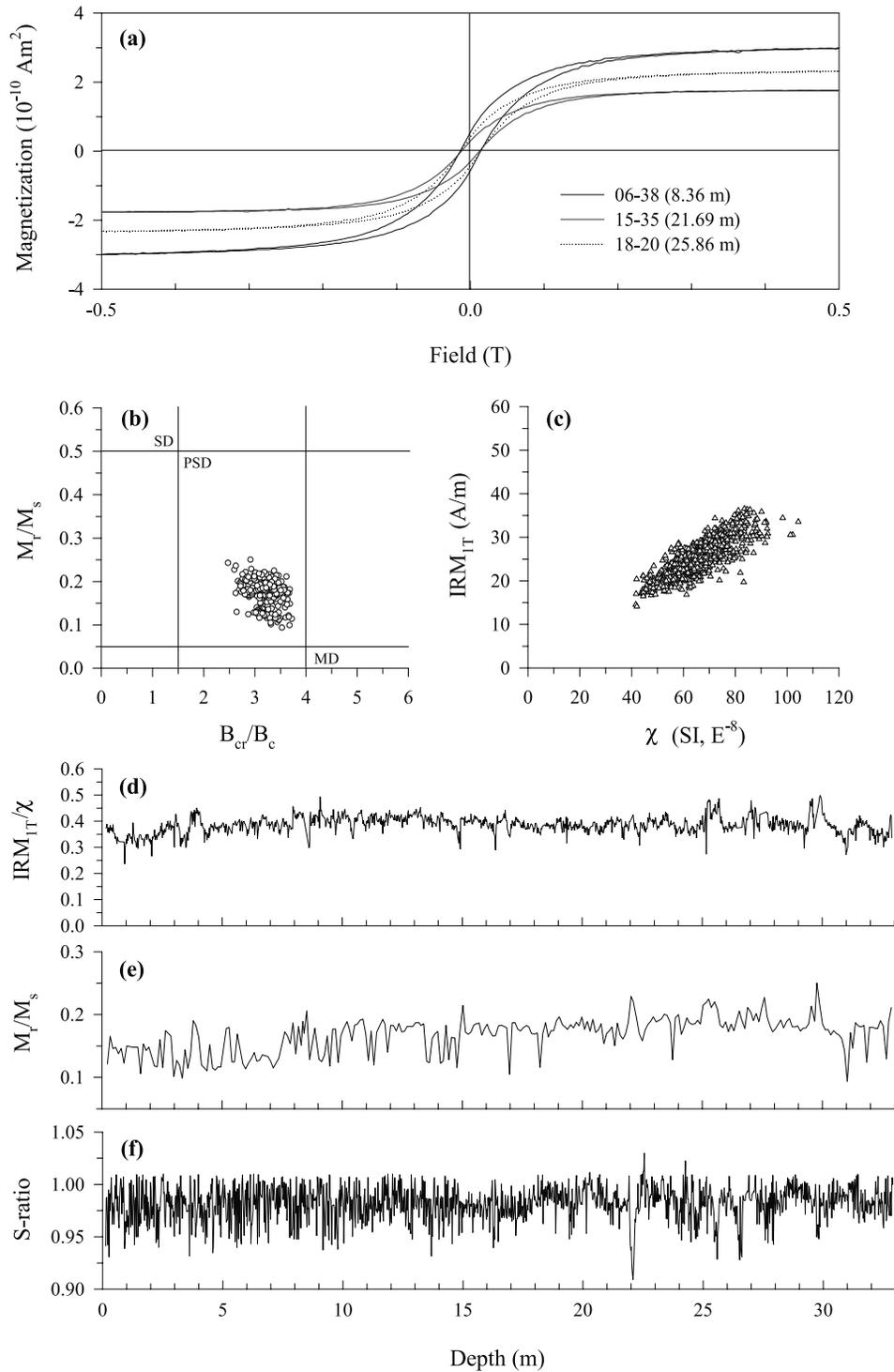


**Figure 3.** Depth versus age plot for core MD972143, using the astronomically tuned age model of *Hornig et al.* [2002], which was constructed after subtraction of 0.93 m of sediment represented by volcanic ash layers. Variations in sedimentation rate throughout the studied age interval are shown in the bottom panel.

Regardless, it is evident that the declinations change by  $180^\circ$  at each polarity reversal and that the paleomagnetic directions reliably record the geomagnetic signal. Values of the maximum angular deviation [*Kirschvink*, 1980] are generally below  $5^\circ$ , which indicates that ChRM directions are well defined. In addition to the Jaramillo and Olduvai subchrons, it is possible to recognize the Santa Rosa [following *Singer and Brown*, 2002], Cobb Mountain and Réunion II polarity intervals within the Matuyama Chron in core MD972143 (Figure 2). The  $\delta^{18}\text{O}$  stratigraphy of *Hornig et al.* [2002] is clearly delineated back to stage 81 (Figure 2) by correlation of the oxygen isotope record from core MD972143 with the composite  $\delta^{18}\text{O}$  record of *Shackleton and Pisias* [1985], *Shackleton et al.* [1990], and *Shackleton et al.* [1995]. Stage 81 is identified immediately above the foraminiferal sand layer [*Hornig et al.*, 2002], which provides an age constraint of 2.14 Ma for the base of the paleointensity record presented here.

[10] Astronomical calibration of the MD972143 record was achieved [*Hornig et al.*, 2002] by tuning the  $\delta^{18}\text{O}$  record to a target curve constructed using the astronomical solution of *Laskar et al.* [1993]. The orbital components of the  $\delta^{18}\text{O}$  data from core MD972143 clearly match the target curve and provide confidence in the results of the astronomical tuning procedure. We use the age model produced by *Hornig et al.* [2002] for presenting our paleointensity record.

[11] In Figure 3, we show a plot of sedimentation rate for the interval above the white foraminiferal sand layer (32.91–33.95 m) after removing 0.93 m of the sediment record represented by volcanic ash layers which are treated as



**Figure 4.** Mineral magnetic properties for core MD972143. (a) Representative hysteresis loops. (b) Plot of  $M_r/M_s$  versus  $B_{cr}/B_c$  for 313 samples [Day *et al.*, 1977], which indicates that the magnetic mineral assemblage is dominated by pseudosingle-domain (PSD) titanomagnetite (SD, single domain, and MD, multidomain). (c) Plot of  $IRM_{1T}$  versus  $\chi$  for all samples except those from volcanic ash horizons which are not considered for paleointensity determinations. The clustering of the data indicates that the titanomagnetite particles have a relatively narrow range of concentrations and the near linearity of the cluster indicates that the range of grain size variation is also narrow. If a line of best fit is fitted through the data, it would have a positive intercept on the horizontal axis, which indicates that paramagnetic minerals contribute to the susceptibility. (d) and (e) Down-core variations of grain-size-dependent magnetic parameters  $IRM_{1T}/\chi$  and  $M_r/M_s$ , respectively. (f) Down-core variations of the  $S$  ratio indicate that a ferrimagnetic mineral dominates the magnetic properties. This is consistent with the identification of titanomagnetite [Horng *et al.*, 2002].

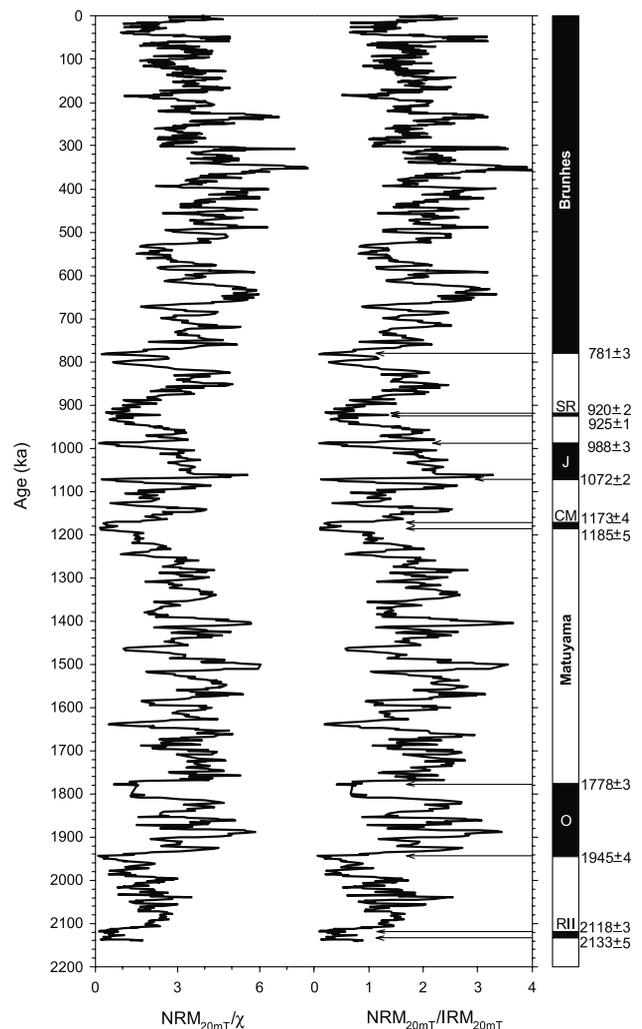
instantaneous deposits that are not representative of longer-term sedimentation. These volcanic ash layers are generally evident as peaks in down-core profiles of  $\chi$  and NRM intensity (Figure 2). Below the foraminiferal sand layer, the chronology is less secure, and this older portion of the core is therefore not used in the present study. Temporal variations in sedimentation rate were determined (lower panel of Figure 3) using 56 age control points from the  $\delta^{18}\text{O}$  stratigraphy [Hornig *et al.*, 2002]. Sedimentation rates varied between 1 and 2 cm/kyr for the studied time interval, except for the last  $\sim 340$  kyr where sedimentation rates were generally higher than 2 cm/kyr. This higher apparent sedimentation rate for the uppermost part of the core might be partially attributable to lower compaction of the surficial sediments.

#### 4.2. Mineral Magnetic Properties

[12] In order for a sediment to be suitable for relative paleointensity studies, the magnetization of the sediment must be linearly related to the Earth's magnetic field strength [e.g., King *et al.*, 1983; Tauxe, 1993]. In order to meet this condition, empirical studies indicate that the magnetization of the sediment should be carried by magnetite, preferably in the pseudosingle-domain (PSD) grain size range (1–15  $\mu\text{m}$ ), and that the concentration of the magnetite should not vary by more than a factor of 10 [Tauxe, 1993]. Mineral magnetic properties are shown for core MD972143 in Figure 4 (excluding samples with high volcanic ash contents). Hornig *et al.* [2002] carried out X-ray diffraction analysis and electron probe microanalysis on magnetic extracts. Their results indicate that a surficially oxidized low-titanium magnetite is the only magnetic mineral present. Hysteresis loops have regular shapes (Figure 4a) and are not wasp-waisted [e.g., Roberts *et al.*, 1995], which is consistent with the presence of a single magnetic mineral without a bimodal mixture of grain sizes. Hysteresis parameters, as shown in a “Day plot” [Day *et al.*, 1977] (Figure 4b), are consistent with the dominance of particles in the PSD size range. A plot of  $\text{IRM}_{1\text{T}}$  versus  $\chi$  has relatively uniform slope, which suggests that the titanomagnetite grains in core MD972143 have a narrow range of grain sizes (Figure 4c). Down-core plots of grain-size-dependent parameters such as  $\text{IRM}_{1\text{T}}/\chi$  and  $M_r/M_s$  (Figures 4d and 4e) have no resemblance to  $\delta^{18}\text{O}$  variations (Figure 2), which indicates that the observed minor grain size variations are not climatically controlled. Maximum and minimum values of  $\chi$  and  $\text{IRM}_{1\text{T}}$  vary by a factor of  $\sim 2.5$  (Figure 4c). The titanomagnetite concentration therefore varies by much less than a factor of 10, as suggested by Tauxe [1993] as a requirement for relative paleointensity studies. The interpretation that titanomagnetite is the dominant magnetic mineral in core MD972143 is supported by  $S$  ratio data ( $S$  ratio =  $\text{IRM}_{-0.3\text{T}}/\text{IRM}_{1\text{T}}$ ; see Verosub and Roberts [1995]), which generally vary between 0.95 and 1.00, with an average of 0.98 (Figure 4f). Thus, in summary, it appears that the sediments of core MD972143 meet the mineralogical requirements for relative paleointensity determinations because the magnetization is dominated by titanomagnetite with a narrow range of concentrations in the PSD size range.

#### 4.3. Relative Paleointensity Determinations

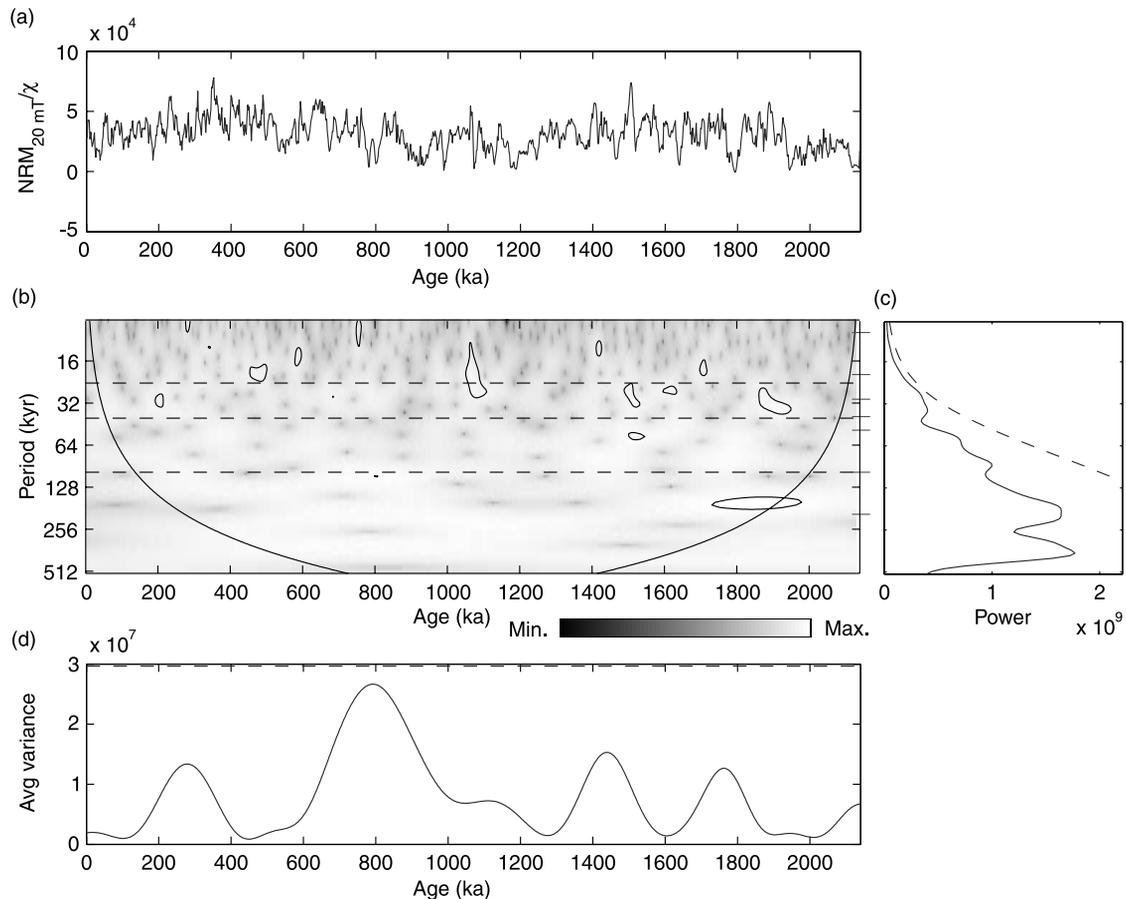
[13] We have normalized the NRM with  $\chi$  and IRM in order to estimate relative geomagnetic paleointensity varia-



**Figure 5.** Two estimates of relative geomagnetic paleointensity for the last 2.14 Myr from core MD972143:  $\text{NRM}_{20\text{mT}}/\chi$  and  $\text{NRM}_{20\text{mT}}/\text{IRM}_{20\text{mT}}$ . Ages are plotted with respect to the astronomically tuned age model of Hornig *et al.* [2002]. The magnetic polarity zonation for the core is summarized on the right-hand side of the figure, along with the ages derived for the polarity boundaries by Hornig *et al.* [2002]. Abbreviations for polarity intervals are the same as in Figure 2.

tions for the MD972143 core (after demagnetization of the NRM and IRM at 10, 15, and 20 mT). Comparison of  $\text{NRM}/\chi$  and  $\text{NRM}/\text{IRM}$  data indicates that both parameters provide equivalent estimates of the geomagnetic relative paleointensity (Figure 5). Indistinguishable results were obtained for normalizations at each of the above demagnetization levels; therefore, for the sake of simplicity, we have only shown the normalization at 20 mT in Figure 5. The details of the paleointensity curves, including variations in frequency and amplitude of the signal, are reproduced for both normalization parameters. We therefore use the  $\text{NRM}_{20\text{mT}}/\chi$  curve for comparison with other paleointensity records for the remainder of this paper.

[14] In order to assess the robustness of the paleointensity proxies shown in Figure 5, it is useful to conduct spectral

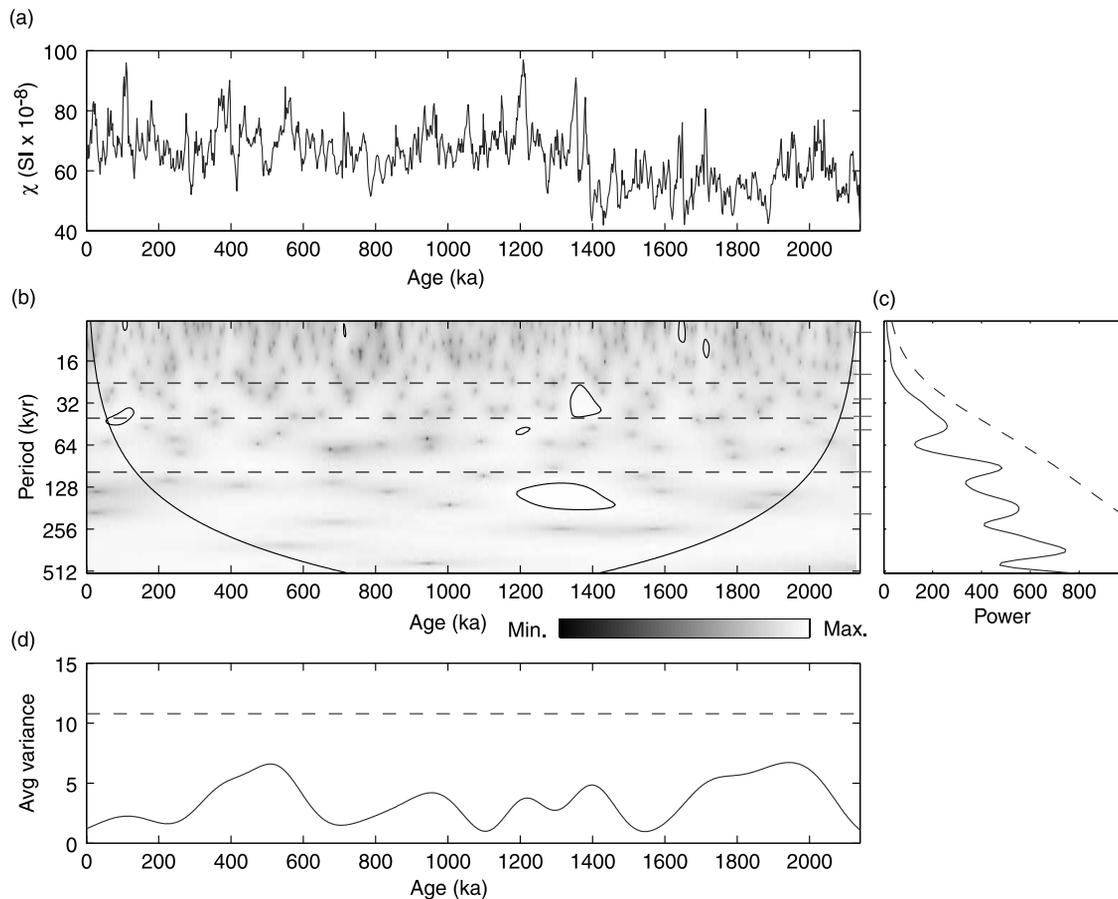


**Figure 6.** Wavelet analysis of the relative paleointensity proxy ( $\text{NRM}_{20\text{mT}}/\chi$ ) for core MD972143. (a)  $\text{NRM}_{20\text{mT}}/\chi$  for the studied 2.14 Myr time series. (b) Wavelet power spectrum for the paleointensity time series. Orbital periods of precession (23 kyr), obliquity (41 kyr), and eccentricity (100 kyr) are shown as dashed lines. Other ticks on the right-hand scale indicate periods of 10, 20, 30, 40, 50, and 200 kyr. Contour lines surrounding intervals with light shading indicate statistically significant power at the 95% confidence level for red noise with a lag 1 autoregressive process [see *Torrence and Campo*, 1998]. Curved lines on either side of the figure indicate the “cone of confidence” where edge effects become important. (c) Global wavelet spectrum for the entire paleointensity time series. The dashed line indicates the 95% confidence level (i.e., none of the spectral peaks are globally significant). (d) The scale-averaged variance of the wavelet power spectrum, which shows the localization of spectral power for periods ranging from 90 to 110 kyr.

analysis of the paleointensity time series as well as for the normalizing parameters to check for nongeomagnetic contamination of the signal. In paleointensity studies, it is normal to calculate a power spectrum for the entire time series to test for significant periodicities in the signal. However, *Guyodo et al.* [2000] showed that wavelet analysis [see *Torrence and Campo*, 1998] is a more powerful technique for paleomagnetic data sets because it enables identification of temporal variations in observed periodicities. Wavelet analyses of the  $\text{NRM}_{20\text{mT}}/\chi$  paleointensity proxy and of the normalizing parameter ( $\chi$ ) for the last 2.14 Myr are shown in Figures 6 and 7, respectively. In both cases, the global wavelet spectrum (Figures 6c and 7c) contains no significant periodicities for the entire time series, although some temporally localized periods are significant in the wavelet power spectra (Figures 6b and 7b). These localized peaks in the wavelet power spectrum are not coincident for the  $\text{NRM}_{20\text{mT}}/\chi$  and  $\chi$  time series,

which suggests that the paleointensity proxy is not contaminated by nongeomagnetic lithological signals. We therefore conclude that the  $\text{NRM}_{20\text{mT}}/\chi$  paleointensity proxy for the MD972143 core provides a useful geomagnetic signal.

[15] In Figure 8, we compare the relative paleointensity record from core MD972143 with the Sint-800 global stack of *Guyodo and Valet* [1999] for the last 800 kyr and with the West Caroline Basin record of *Yamazaki and Oda* [2002] and the equatorial Pacific record of *Valet and Meynadier* [1993] for the older interval of the core. Agreement between the MD972143 and Sint-800 data sets is generally good, with close agreement between the form of the curves, and generally in the timing and amplitude of maxima and minima. The only significant discrepancies occur at 310, 430–450, 490, 590–620, and 730 ka. The similarity between the Sint-800 and MD972143 paleointensity records can be quantitatively tested by cross-spectral



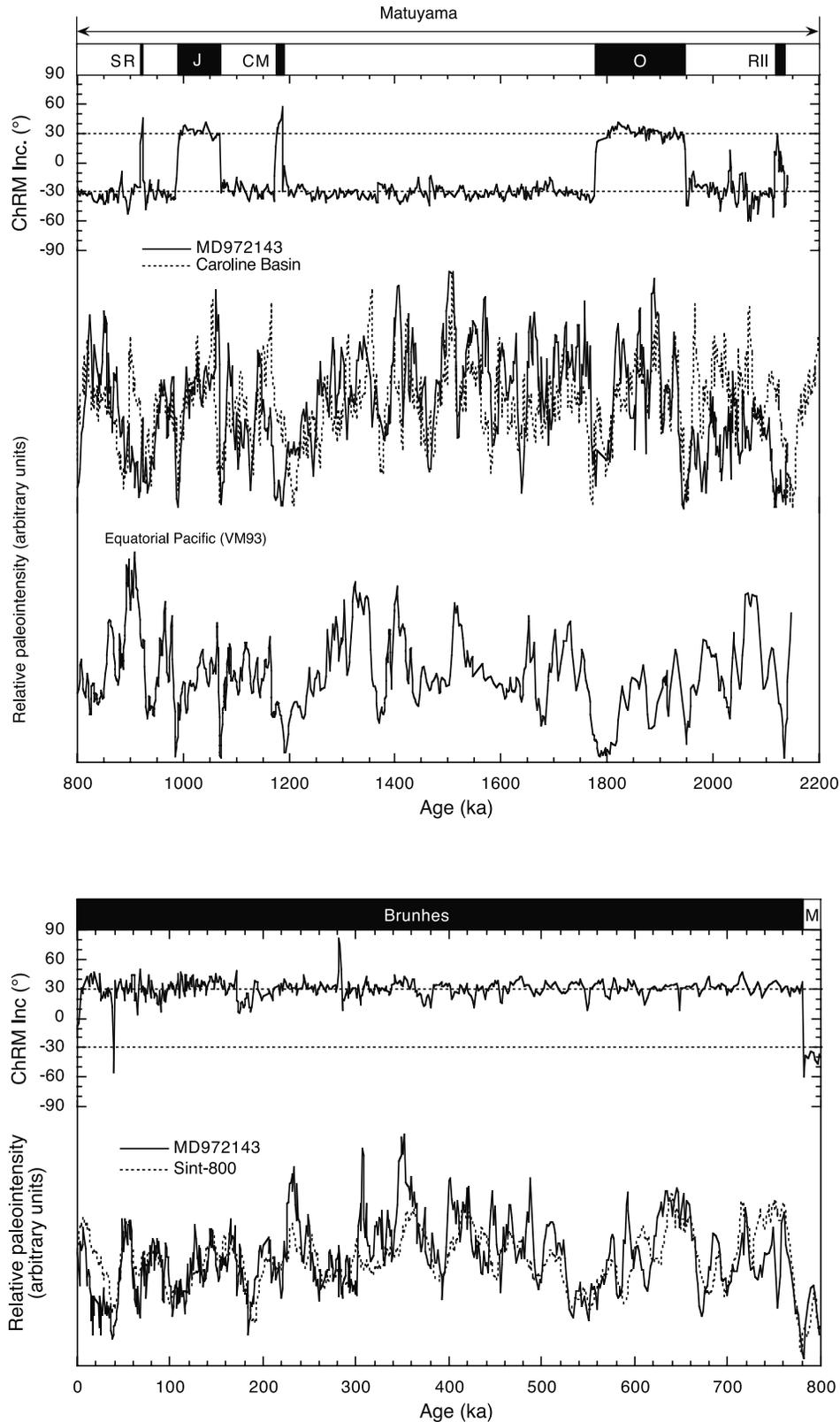
**Figure 7.** Wavelet analysis of the relative paleointensity normalizing parameter ( $\chi$ ) for core MD972143. (a) Magnetic susceptibility ( $\chi$ ) for the studied 2.14 Myr time series. (b) Wavelet power spectrum for the  $\chi$  time series (compare Figure 6). (c) Global wavelet spectrum for the entire  $\chi$  time series. The dashed line indicates the 95% confidence interval (i.e., none of the spectral peaks are globally significant). (d) The scale-averaged variance of the wavelet power spectrum, which shows the localization of spectral power for periods ranging from 90 to 110 kyr.

analysis (Figure 9). Significant coherency is evident across a broad range of frequencies, which suggests that the records are quantitatively similar and that the MD972143 core has recorded dominantly dipolar geomagnetic intensity fluctuations at least over the last 800 kyr.

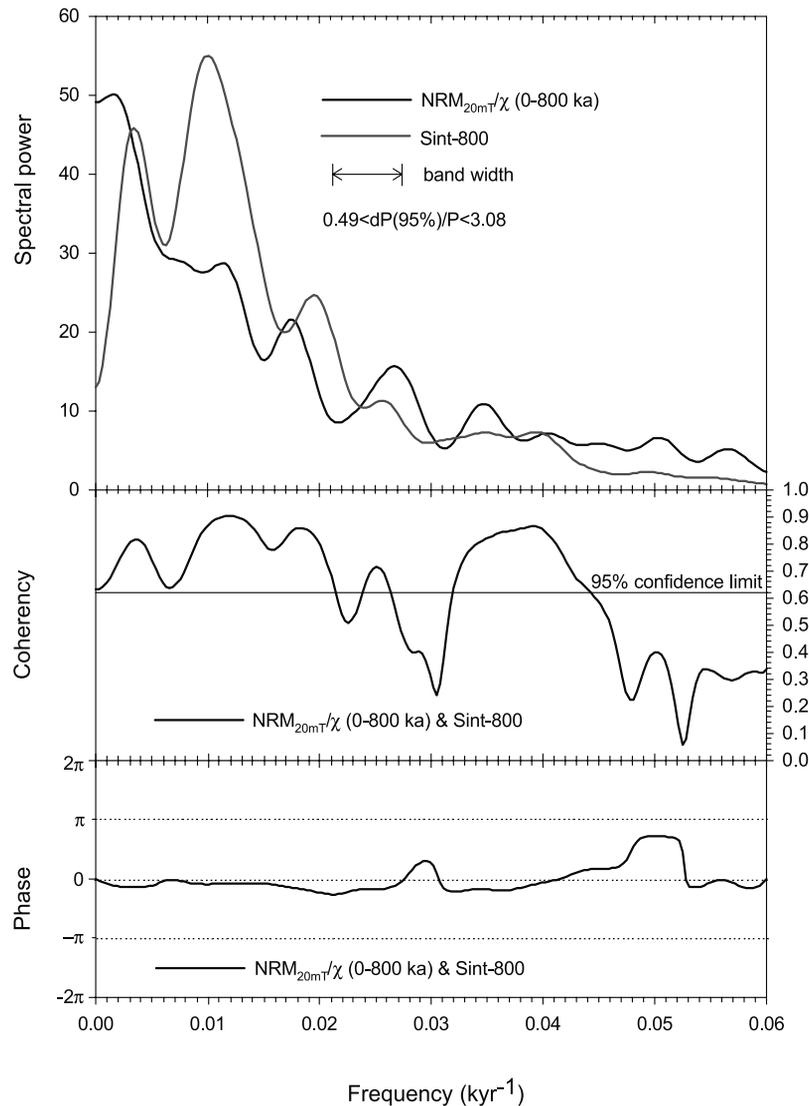
[16] In contrast to the general agreement between the MD972143 paleointensity record and Sint-800, visual correlation between our record and VM93 for the interval between 0.80 and 2.14 Ma is less convincing (Figure 8). In particular, the records show no similarity between 800 and 930 ka, 1430–1630 ka, and within the Olduvai Subchron (1770–1950 ka). This generally poor agreement is demonstrated in Figure 10 where we plot results of cross-spectral analysis between the two paleointensity records for the interval from 0.80 to 2.14 Ma. In contrast to the comparison with Sint-800 (Figure 9), the coherency with VM93 is significant at only a narrow range of frequencies (Figure 10). Despite the relatively poor agreement with VM93, visual correlation between the MD972143 paleointensity record and the West Caroline Basin record of *Yamazaki and Oda* [2002] is good. Apart from discrepancies at 900 ka and 1970–2070 ka, there is close agreement

between the two records in terms of frequency and amplitude of the signal. In some cases, there are temporal offsets between correlative paleointensity features, which simply results from differences in the age models for the two time series. These offsets are particularly obvious around the Cobb Mountain and Réunion II polarity intervals (Figure 8). The generally impressive agreement between these two records provides clear evidence that both cores contain a robust dominantly dipolar signal that will enable development of a detailed view of geomagnetic field behavior between 0.8 and 2.14 Ma.

[17] One of the intervals where there is a discrepancy between the MD972143 and West Caroline Basin records is at around 900 ka, between the Jaramillo Subchron and the Matuyama/Brunhes (M/B) boundary. Several published paleointensity records have good chronological control for the interval between 0.73 and 1.10 Ma, which makes it possible to examine this apparent discrepancy in more detail (Figure 11). The additional paleointensity records shown in Figure 11 include the Mediterranean LC07 record of *Dinarès-Turell et al.* [2002], the North Atlantic ODP Site 983 record of *Channell and Kleiven* [2000], and the



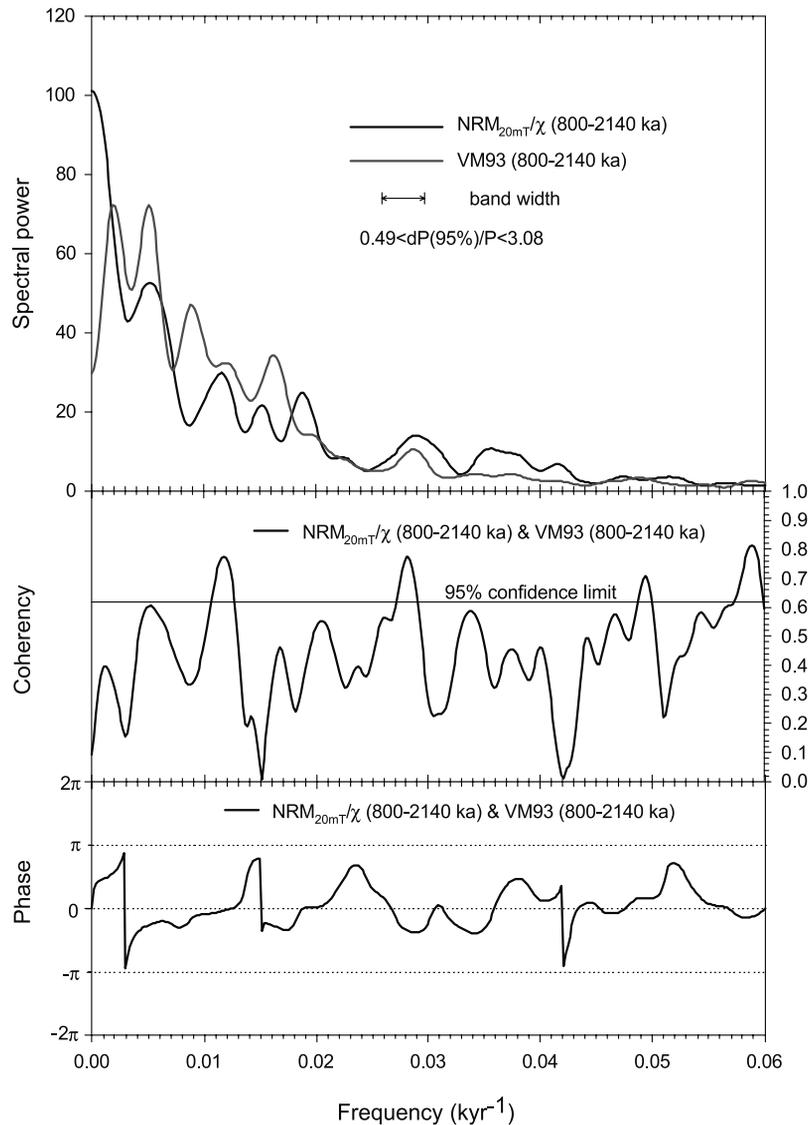
**Figure 8.** The relative paleointensity record for core MD972143 compared to the Sint-800 global paleointensity stack [Guyodo and Valet, 1999] for the last 800 kyr and compared to the West Caroline Basin record of Yamazaki and Oda [2002] and the equatorial Pacific paleointensity record of Valet and Meynadier [1993] (VM93) for the interval from 800 ka to 2.14 Ma. ChRM inclinations are also shown to indicate the positions of polarity reversals documented in core MD972143 (dotted lines indicate the inclinations expected for a geocentric axial dipole field at the site latitude). Black, normal polarity, and white, reversed polarity on the polarity log, which has the same labels as in Figure 2.



**Figure 9.** Spectral power, coherency and phase resulting from cross-spectral analysis of  $\text{NRM}_{20\text{mT}}/\chi$  for core MD972143 and the Sint-800 paleointensity stack of *Guyodo and Valet* [1999] for the last 800 kyr. There is significant coherency between both signals across a broad range of frequencies, which suggests that the records are quantitatively similar.

Jaramillo paleointensity stack of *Guyodo et al.* [2001]. Paleointensity minima are evident at each polarity transition in each record. In addition to the paleointensity minimum at the M/B boundary, all of the records contain evidence of a minimum (DIP 1) that preceded the M/B boundary by  $\sim 15$  kyr [Kent and Schneider, 1995]. Comparison of the positions of these paleointensity minima indicates that there are discrepancies between the ages used for the polarity boundaries. The principal reason for the discrepancy is that *Hornig et al.* [2002] used the most recent astronomical target curve of *Laskar et al.* [1993] for their astronomical calibration of the MD972143 record, whereas some of the other records are based on different astronomical target curves. These discrepancies are of the order of only a few thousand years, which does not compromise comparison of the paleointensity records. In particular, it should be noted that agreement among the published records is good in the vicinity of the Jaramillo

Subchron. For other parts of the paleointensity record between the Jaramillo Subchron and the M/B boundary, serial correlation between coeval features is less clear (Figure 11). *Dinarès-Turell et al.* [2002] showed that, for the interval between 0.78 and 0.88 Ma, the records of *Valet and Meynadier* [1993] and *Meynadier et al.* [1994] compare less favorably with other records from the Pacific Ocean [*Guyodo et al.*, 1999], the Ontong Java Plateau [*Kok and Tauxe*, 1999], and the Mediterranean Sea. *Dinarès-Turell et al.* [2002] suggested that there are reasonable matches between these records when one considers the uncertainties in age control for some of the records. Modification of the age models for these records and those in Figure 11, within allowable age constraints and with tuning to a common astronomical target curve, will help to reduce the apparent discrepancies among the records. Nevertheless, on the basis of published records, paleointensity variations in this age interval are less clearly coherent



**Figure 10.** Spectral power, coherency and phase resulting from cross-spectral analysis of  $\text{NRM}_{20\text{mT}}/\chi$  for core MD972143 and the equatorial Pacific paleointensity record of *Valet and Meynadier* [1993] (VM93) for the interval from 800 ka to 2.14 Ma. Coherency between these records is poor compared to that between core MD972143 and the Sint-800 paleointensity stack, as shown in Figure 9 (see text for discussion).

than for the Jaramillo Subchron [*Guyodo et al.*, 2001] and older intervals (Figure 8).

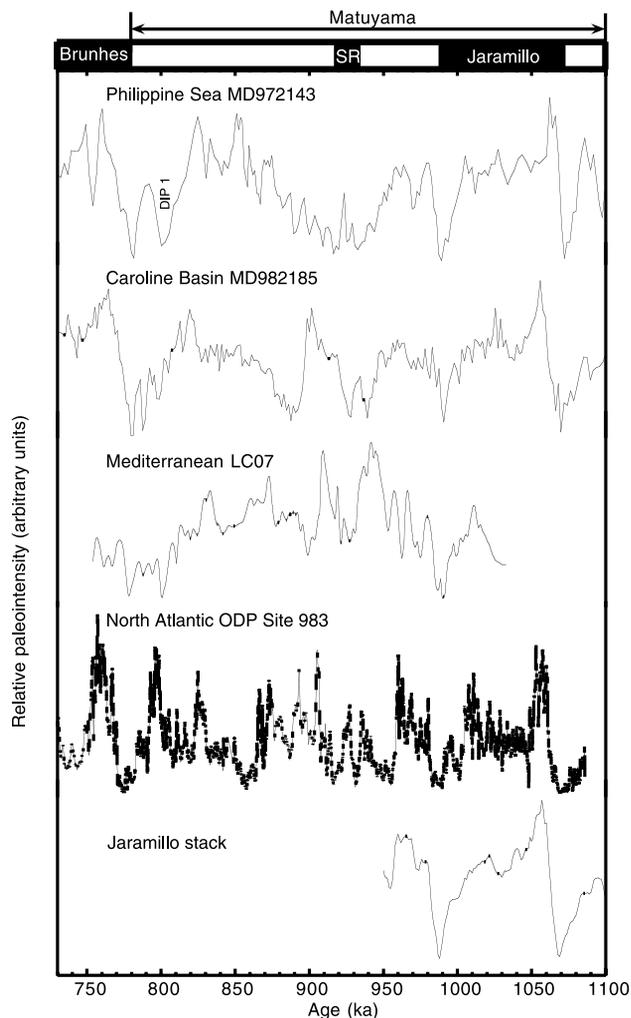
## 5. Discussion

[18] The relative paleointensity data presented in this paper provide a record with high-quality age control; such records are still rare for the time interval beyond 1.1 Ma. The record from core MD972143 provides the opportunity to address some important questions concerning long-term behavior of the geomagnetic field. First, are any significant (orbital) periodicities present in the relative paleointensity data from core MD972143? Second, the MD972143 paleointensity record shows significant coherency with the Sint-800 global paleointensity stack for the past 800 kyr, and with the West Caroline Basin record of *Yamazaki and Oda* [2002] back to 2.14 Ma, but it shows poor coherency with

the VM93 record for the age interval between 0.8 and 2.14 Ma. Does this observation have any implications concerning the reliability of VM93 and are asymmetrical sawtooth paleointensity variations observed in the MD972143 record? Third, is there any evidence in the paleomagnetic record from core MD972143 that suggests the presence of a 100-kyr periodicity in inclination, as recently suggested by *Yamazaki and Oda* [2002]?

### 5.1. Orbital Influence on the Intensity of the Geomagnetic Field?

[19] Statistically significant power at the orbital eccentricity period (100 kyr) and at other periods has been reported in several paleointensity studies of sedimentary sequences [*Tauxe and Wu*, 1990; *Tauxe and Shackleton*, 1994; *Channell et al.*, 1998; *Yamazaki*, 1999]. The magnetic properties of core MD972143 are highly uniform and



**Figure 11.** Comparison of paleointensity records between 0.73 and 1.10 Ma, including the records from core MD972143, West Caroline Basin core MD982185 [Yamazaki and Oda, 2002], Mediterranean core LC07 [Dinarès-Turell et al., 2002], North Atlantic ODP Site 983 [Channell and Kleiven, 2000], and the Jaramillo paleointensity stack of Guyodo et al. [2001].

clearly satisfy the criteria of Tauxe [1993] for relative paleointensity investigations. It is therefore useful to investigate the spectral content of the paleointensity record from core MD972143 to test the claim that the geodynamo is in some way energized by orbital forcing. Wavelet analysis of the MD972143 paleointensity record indicates that there is no statistically significant power at the orbital eccentricity period (Figure 6b). The scale-averaged variance contains a peak for the 100-kyr eccentricity period at 800 ka; however, this peak is not statistically significant at the 95% confidence level (Figure 6d). Although there are also peaks in the global power spectrum (Figure 6c), the 2.14-Myr record from core MD972143 is notable for its lack of significant temporally persistent power at any period. This suggests, in contrast to the findings of Channell et al. [1998] and Yamazaki [1999], that orbital eccentricity or other orbital components are not geomagnetically significant over the last 2 Myr. Our results are consistent with those of Guyodo

et al. [2000], who used wavelet analysis to demonstrate that the inferred eccentricity modulation of geomagnetic field intensity suggested by Channell et al. [1998] represents a subtle expression of lithological variations rather than being a characteristic of the geodynamo. Wavelet analysis therefore appears to be a particularly powerful tool for testing the origin of persistent periodicities in records of geomagnetic relative paleointensity.

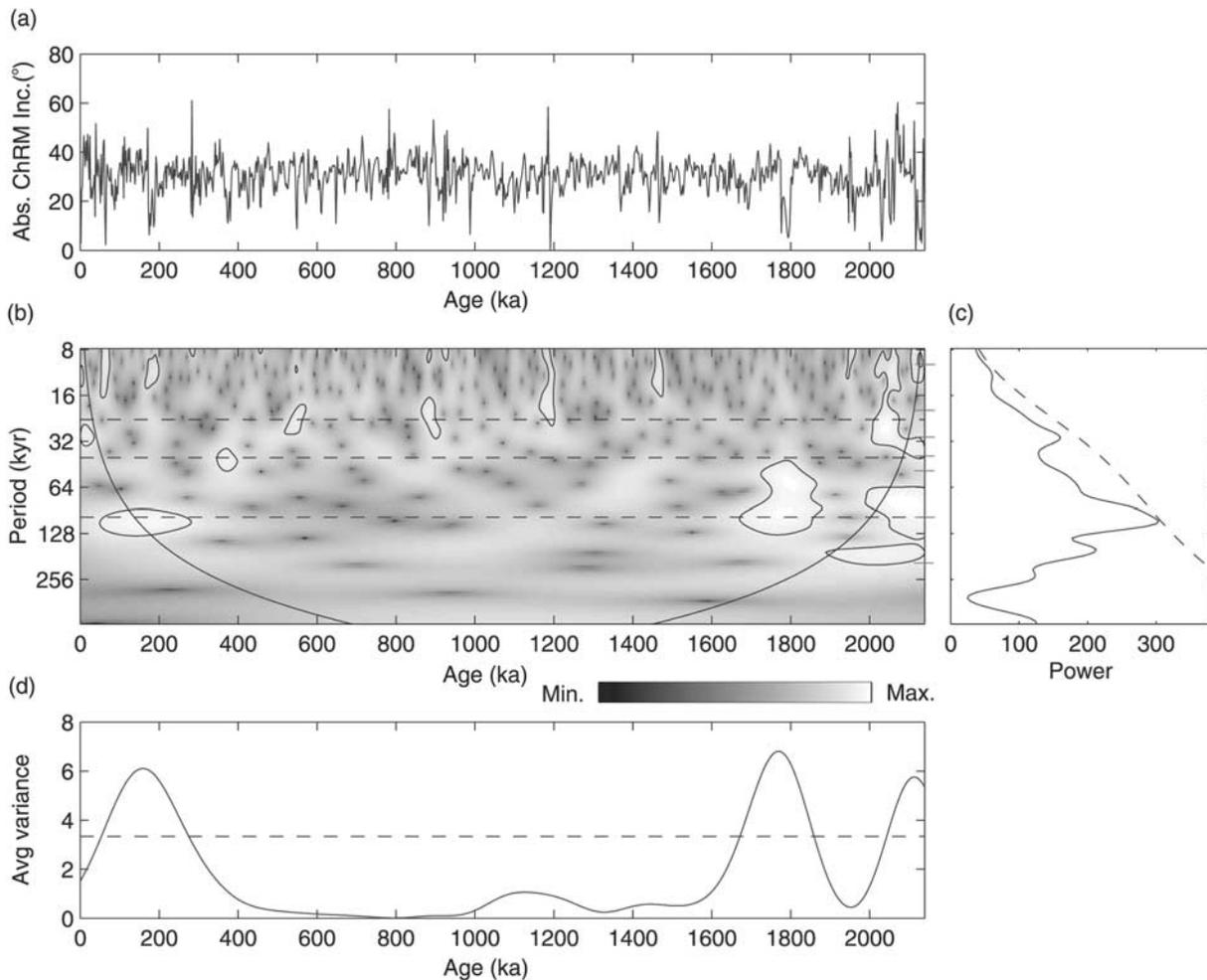
## 5.2. Asymmetrical Sawtooth Paleointensity Behavior?

[20] When Valet and Meynadier [1993] provided evidence that geomagnetic paleointensity variations have an asymmetrical sawtooth form between successive reversals, it was suggested that geomagnetic field behavior was deterministic. That is, the primary constraint on the timing of the next reversal is the intensity to which the field rebounded after the previous reversal. This conclusion conflicts with over 30 years of analysis of long-term field behavior [e.g., Cox, 1968; McFadden et al., 1987], where the probability of a polarity transition should not be a function of field intensity. The observation of asymmetrical sawtooth paleointensity behavior has therefore been hotly debated. Criticism has focused on geomagnetic [e.g., McFadden and Merrill, 1997] and rock magnetic [e.g., Kok and Tauxe, 1996a, 1996b; Mazaud, 1996] arguments, with new records [e.g., Laj et al., 1996b; Kok and Tauxe, 1999] often not confirming the observations of Valet and Meynadier [1993]. The number of records that span several reversals is still small; thus new records that can constrain the problem are valuable.

[21] Inspection of the paleointensity record from core MD972143 indicates that it does not have an obvious asymmetrical sawtooth form (Figure 8). Rather, the paleointensity variations seem to occur on a similar scale to those in the Brunhes Chron, with relatively broad maxima and intermittent minima that seem to correspond to geomagnetic excursions or short polarity intervals. There is close agreement between our record and that of Valet and Meynadier [1993] for some intervals (e.g., 950–1400 ka; Figure 8). However, lack of coherency between the two records for other intervals (Figure 10) and the generally excellent agreement between our record and that of Yamazaki and Oda [2002] raises the question of whether VM93 is consistently reliable. The sediments from core MD972143 are magnetically much more uniform than those studied by Valet and Meynadier [1993]. The fact that for the last 800 kyr, the geomagnetic paleointensity appears to be well recorded by sediments with the same magnetic properties as the interval from 800 ka to 2.14 Ma suggests that this earlier interval should record paleointensities as faithfully as the younger interval in core MD972143. The coeval record of Yamazaki and Oda [2002] also contains no evidence for asymmetrical sawtooth paleointensity behavior. Together, these records provide independent evidence that asymmetrical sawtooth paleointensities might result from rock magnetic artefacts, as has been suggested in the literature.

## 5.3. Orbital Influence on Paleomagnetic Inclination?

[22] Yamazaki and Oda [2002] provided evidence for a 100-kyr periodicity in paleomagnetic inclination data as well as in their paleointensity record from the West Caroline Basin. As demonstrated above, our wavelet analysis for



**Figure 12.** Wavelet analysis of the absolute value of the paleomagnetic inclinations for core MD972143. (a) Absolute value of the paleomagnetic inclination for the studied 2.14 Myr time series. (b) Wavelet power spectrum for the inclination time series (compare. Figure 6). (c) Global wavelet spectrum for the entire inclination time series. The dashed line indicates the 95% confidence level. (d) The scale averaged variance of the wavelet power spectrum, which shows the localization of spectral power for periods ranging from 90 to 110 kyr.

the MD972143 relative paleointensity record does not support a long-term modulation of the geomagnetic field by orbital eccentricity. It is therefore worth testing whether our inclination data contain any evidence for a temporally persistent eccentricity-related periodicity. We have performed a wavelet analysis of the absolute value of the paleomagnetic inclinations from the MD972143 core. Absolute values have been used to remove the effect of step shifts in the inclination records at each respective polarity transition. Wavelet analysis of the MD972143 inclination record (Figures 12b and 12c) confirms the finding of *Yamazaki and Oda* [2002] that the greatest concentration of power in the global power spectrum occurs at the 100-kyr eccentricity period (Figure 12c). However, this dominant periodicity is only statistically significant at the 95% confidence level at around 1.8 Ma and near both ends of the record (Figures 12b and 12d), where edge effects become important and where the results are less meaningful. Lack of statistically significant, temporally

persistent power at the 100-kyr eccentricity period in our record suggests that the conclusion concerning orbital eccentricity modulation of the geomagnetic field [*Yamazaki and Oda*, 2002] is questionable. It should be noted that *Yamazaki and Oda* [2002] did not demonstrate that the spectral power at the 100-kyr orbital eccentricity period was statistically significant. We therefore consider it premature to accept the hypothesis of orbital modulation of the geomagnetic field without more robust evidence.

#### 5.4. Long-Term Geomagnetic Field Behavior

[23] As expected, the MD972143 paleointensity record has minima in correspondence with each geomagnetic reversal. In addition, there are minima at times of known geomagnetic excursions during the Brunhes Chron (Figure 8). Directional fluctuations away from the expected axial geocentric dipole inclination at the site latitude are evident in the MD972143 record (Figure 8) in association with the observed paleointensity minima, although the Laschamp

event at  $\sim 40$  ka is the only excursion for which reversed polarity inclinations are observed. Our data therefore confirm the suggestion that collapse of the field to low paleointensities, with associated directional fluctuations, are a normal part of long-term geomagnetic field behavior.

[24] A similar pattern of geomagnetic behavior is evident within the MD972143 record during the Matuyama Chron. Along with the Jaramillo and Olduvai subchrons, three short normal polarity intervals are present in the Matuyama Chron, including the Santa Rosa polarity interval and the Cobb Mountain and Réunion II subchrons (Figure 8). In addition, the presence of numerous paleointensity minima, which are often associated with directional fluctuations, confirms observations from the Brunhes Chron that these features represent a normal aspect of long-term geomagnetic field behavior.

## 6. Conclusions

[25] We have presented a detailed relative paleointensity record spanning the last 2.14 Myr from the western Philippine Sea, with age control based on well-defined  $\delta^{18}\text{O}$  data that enable astronomical tuning of the age model for core MD972143. The MD972143 record confirms that, within stable polarity chrons, it is normal for the geomagnetic field to undergo dynamic changes, with relatively frequent collapses of the field to low paleointensities, and concomitant directional fluctuations. Results of wavelet analyses suggest that there are no significant temporally persistent periodicities in either the paleointensity or rock magnetic time series and that there is no significant coherency between the relative paleointensity proxy and the normalizing parameter used to derive the proxy. Wavelet analysis of our relative paleointensity record for the last 2.14 Myr therefore does not support the suggestion that 100-kyr orbital eccentricity periodicities are characteristic of the geodynamo [cf. Channell *et al.*, 1998; Yamazaki, 1999]. This result supports the conclusion of Guyodo *et al.* [2000] and confirms that wavelet analysis is a powerful tool for testing the origin of periodicities in geomagnetic time series. Furthermore, wavelet analysis of our paleomagnetic inclination record does not support the suggestion of a persistent 100-kyr inclination periodicity [Yamazaki and Oda, 2002].

[26] In contrast to the good comparison between the MD972143, Sint-800 and the West Caroline Basin [Yamazaki and Oda, 2002] paleointensity records, agreement between our record and that of Valet and Meynadier [1993] for much of the interval between 0.8 and 2.14 Ma is less convincing. Furthermore, in agreement with Yamazaki and Oda [2002], we do not observe an asymmetrical sawtooth form in our paleointensity record. These results provide independent evidence that asymmetrical sawtooth paleointensities might result from rock magnetic artefacts, as has been suggested in the literature.

[27] Our results suggest that the MD972143 paleointensity record is largely free of rock magnetic, lithological or climatic artefacts and that it represents a reliable record of geomagnetic field behavior. This conclusion is supported by the observation that the MD972143 paleointensity record is highly coherent with the Sint-800 global paleointensity stack for the last 800 kyr and by good correlation between the MD972143 record and the coeval West Caroline Basin

record of Yamazaki and Oda [2002]. Good coherence between records for the older age interval suggests that we are starting to obtain a reliable and detailed view of geomagnetic field behavior for the last 2 Myr.

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