Physics of the Earth and Planetary Interiors 110 (1999) 115–128

PHYSICS OF THE EARTH AND PLANETARY INTERIORS

Paleointensity variations of Earth's magnetic field and their relationship with polarity reversals

Jon D. Pelletier

Division of Geological and Planetary Sciences, California Institute of Technology, Mail Stop 150-21 Pasadena, CA 91125, USA

Received 7 August 1998; accepted 9 October 1998

Abstract

Power-spectral analyses of the intensity of Earth's magnetic field inferred from ocean sediment cores and archeomagnetic data from time scales of 100 yr to 10 Myr have been carried out. The power spectrum is proportional to $1/f$ where f is the frequency. These analyses compliment previous work which has established a $1/f^2$ spectrum for variations at time scales less than 100 yr. We test the hypothesis that reversals are the result of variations in field intensity with a $1/f$ spectrum which occasionally are large enough to cross the zero intensity value. Synthetic binormal time series with a $1/f$ power spectrum representing variations in Earth's dipole moment are constructed. Synthetic reversals from these time series exhibit statistics in good agreement with the reversal record, suggesting that polarity reversals may be the end result of autocyclic intensity variations with a $1/f$ power spectrum.

Keywords: Magnetic field; Frequency; Intensity

1. Introduction

Earth's magnetic field has exhibited significant variability over a wide range of time scales. On time scales less than a couple of hundred years historical data are available for variations in the intensity and orientation of the geomagnetic field. Archeomagnetic data can be used to infer the intensity of the field from time scales of centuries to millenia. Sediment cores provide the widest range of time scales of variations in the geomagnetic field with internal origin: 1 kyr to 10 Myr. Techniques of time series analysis can be used to characterize this variability. The power spectrum is the square of the coefficients of the Fourier transform of the time series. It quantifies the average variability of the series at different time scales. Barton (1982) has performed spectral analysis of paleointensity using historical observations and sediment cores. He identified a broad, continuous power spectrum with a steep dependence on time scale for time scales less than a century and a flatter spectrum at longer time scales.

The geomagnetic field also exhibits reversals with a complex history including variations over a wide range of time scales. The reversal history can be characterized by the polarity interval distribution and the reversal rate. Polarity intervals vary from those short enough to be barely resolved in the magnetic anomalies of the seafloor to the 35 Myr Cretaceous superchron. Reversals are also clustered in time such that short polarity intervals tend to be followed by short polarity intervals and long intervals by long intervals. This clustering has been quantified with the reversal rate which gradually decreases going

PII: S0031-9201(98)00140-X

back to 100 Ma and then increases going back further in time before the Cretaceous superchron.

Due to the availability of many new time series data sets for paleointensity of Earth's magnetic field since the work of Barton (1982), it would be useful to perform power-spectral analyses of some of these recent data sets to further characterize the temporal variability of the geomagnetic field. In this paper, we perform power-spectral analyses of time series data for the intensity of Earth's magnetic field inferred from sediment core and archeomagnetic data. We find that the power spectrum of the intensity of the geomagnetic field from time scales of 100 yr to 10 Myr is well approximated by a $1/f$ dependence, where f is the frequency. Variations in the intensity of the geomagnetic field in one polarity exhibit a normal distribution. When a fluctuation crosses the zero intensity value a reversal occurs. We test the hypothesis that reversals are the result of intensity variations with a $1/f$ power spectrum which occasionally are large enough to cross the zero intensity value, driving the geodynamo into the opposite polarity state. Synthetic time series with a $1/f$ power spectrum and a binormal distribution are used to generate reversal statistics. These are found to be in good agreement with those of the real reversal history. The reversal statistics are sensitive to the form of the power spectrum of intensity variations. We conclude that the agreement between the synthetic reversal record produced with $1/f$ noise variations in the intensity of Earth's magnetic field is strong support for $1/f$ behavior over the length of the reversal record. This suggests that processes internal to the core (autocyclic) may determine geomagnetic variability up to very long time scales. This contrasts with the hypothesis that internal core processes dominate secular variations while changes in conditions at the core–mantle boundary determine variations on longer time scales (McFadden and Merrill, 1995).

2. Power-spectral analyses of the virtual axial dipole moment

Paleomagnetic studies clearly show that the polarity of the magnetic field has been subject to reversals. Kono (1971) and Tanaka et al. (1995) have compiled paleointensity measurements of the magnetic field from volcanic lavas for 0–10 Ma. They concluded that the distribution of paleointensity is well approximated by a symmetric binormal distribution with mean 8.9×10^{22} Am^2 and standard deviation 3.4×10^{22} Am^2. One normal distribution is applicable to the field when it is in its normal polarity and the other when it is in its reversed polarity.

We have utilized four datasets for computing the power spectrum of the intensity of Earth's magnetic field. They are archeomagnetic data from time scales of 100 yr to 1.6 kyr from Kovacheva (1980), marine sediment data from the Somali basin from time scales of 1.6 kyr to 25 kyr from Meynadier et al. (1992), marine sediment data from the Pacific and Indian Oceans from 25 kyr to 4 Myr from Meynadier et al. (1994), and marine sediment data from 10 kyr to 10 Myr from Tauxe and Hartl (1997). The data were published in table form in Kovacheva (1980), obtained from L. Meynadier (Meynadier, 1995) for the marine sediment data in Meynadier et al. (1992) and Meynadier et al. (1994), and obtained from L. Tauxe (Tauxe, 1998) for the marine sediment data in Tauxe and Hartl (1997). The archeomagnetic data from Kovacheva (1980) is a master curve for southeastern Europe. The intensity is expressed in terms of the Virtual Axial Dipole Moment (VADM) where the effects of latitude of the measurement locations were taken into account in estimating the strength of Earth's dipole moment. This calibration is important for comparison with marine sediment data on longer time scales which are calibrated in units of VADM by comparison with the average paleointensity estimated from volcanic lavas from the same time range as that of each sediment core. Thus, although each core represents variations in the geomagnetic field at different locations, they are calibrated with respect to the estimated dipole intensity so that they may be compared and compiled to obtain information of the variation of the geomagnetic field over a continuous range of frequencies.

Techniques for assessing the paleointensity of Earth's magnetic field are reviewed in Tauxe (1993). King et al. (1983) has established minimum quality standards for establishing the reliability of a sediment core to determine accurate relative paleointensity measurements. All of the cores we analyze satisfy these criteria (Meynadier et al., 1992; Meynadier et al., 1994; Tauxe and Hartl, 1997). The cores were tested for the homogeneity of rock magnetic properties and for the stationarity of sedimentation rate. Meynadier et al. (1992, 1994) noted that except for the topmost sections, their cores had homogeneous magnetic properties within the standards of King et al. (1983). The cores of Tauxe and Hartl (1997) were described as 'remarkably consistent' in terms of their magnetic properties. All cores feature steady sedimentation rates. The effect of variable sedimentation rate has been investigated by Guyodo and Valet (1996). They found that variable sedimentation rate can act as a high-pass filter removing some of the variance at the highest frequencies of the core. However, this is not expected to be the case in these cores due to their uniform sedimentation rates. The other potential uncertainty in the data we analyze is associated with the time scale calibration of the core. In our spectral analysis we are interested in obtaining the exponent of the power-law characterizing the power spectrum. The exponent quantifies how strongly the power-spectral density depends on frequency. If the calibration of the core is in error, this does not affect the power-spectral exponent at all since the relative power at different frequencies is unchanged. This can also be seen by considering the units of power-spectral density: field intensity squared multiplied by time. If the time scale changes by a constant factor, the power-spectral density at all frequencies simply goes up or down by the same factor and the exponent is unchanged.

Marine sediment data can be accurate measures of relative paleointensity but give no information on absolute intensity. In order to calibrate marine sediment data, the data must be compared to absolute paleointensity measurements from volcanic lavas sampled from the same time period as the sediment record. Meynadier et al. (1994) have done this for the composite Pacific and Indian Ocean dataset. They have calibrated the mean paleointensity in terms of the virtual axial dipole moment for 0–4 Ma as 9×10^{22} Am^2 (Valet and Meynadier, 1993). This value is consistent with that obtained by Kono (1971) and Tanaka et al. (1995) for the longer time interval up to 10 Ma. Using this calibration, we calibrated the Somali data with the time interval 0–140 ka from the composite Pacific and Indian Ocean dataset. The data from Meynadier et al. (1994) are plotted in Fig. 1 as a function of age in Ma. The last reversal at

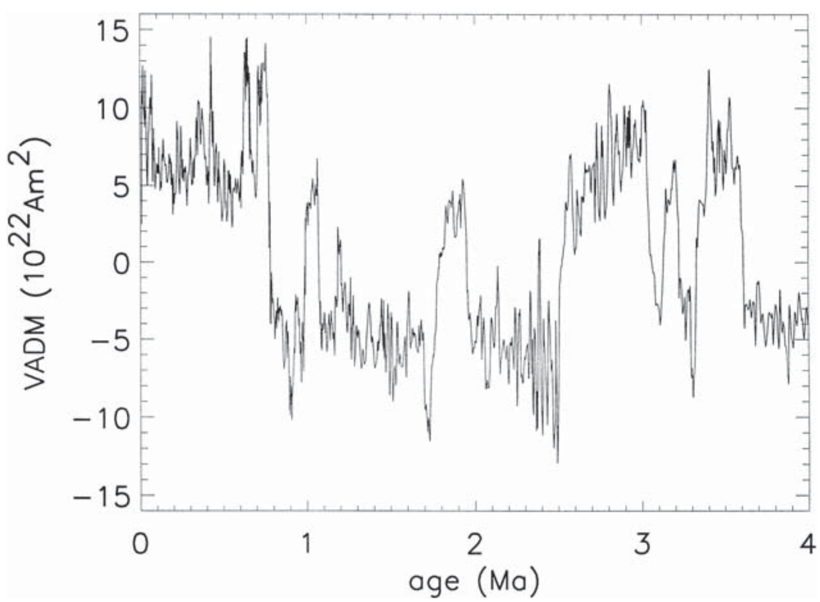

Fig. 1. Paleointensity of the intensity of Earth's magnetic field as inferred from sediment cores for the past 4 Ma from Meynadier et al. (1994). The data are expressed as the virtual axial dipole moment (VADM) with reversed polarity data given by negative values.

approximately 780 ka is clearly shown. We computed the power spectrum of each of the time series with the Lomb periodogram (Press et al., 1992). This technique is used for data which are nonuniformly sampled in time. The compiled spectra are given in Fig. 2 with the frequencies where one spectrum ends and the other begins given by a solid dot. The composite sediment record from the Pacific and Indian Oceans are plotted up to a frequency corresponding to a period of 25 kyr. Above this time scale good synchroneity is observed in the Pacific and Indian Ocean datasets (Meynadier et al., 1994). This suggests that non-geomagnetic effects such as variable sedimentation rate are not significant in these cores above this time scale. From frequencies corresponding to time scales of 25 kyr down to 1.6 kyr we plot the power spectrum of the Somali data. Although the Somali data has a uniformly high sedimentation rate (all sections of the core have rates greater than 5 cm/kyr), the highest frequencies in this section of the spectrum may not be reliable since sediment cores are usually thought to be reliable only down to time scales of several thousand years. From time scales of 1.6 kyr to the highest frequency we plot the power spectrum of the data of Kovacheva (1980). A least-squares linear regression to the data yields a slope of -1.09 over 4.5 orders of magnitude. This indicates that the power spectrum is well approximated as $1/f$ on these time scales.

We have performed similar analyses on sediment cores covering the time scales of 10 kyr to 10 Myr obtained by Tauxe and Hartl (1997). This core enabled us to cover the same frequency range, 25 kyr to 4 Myr, with cores from two different time periods. The core of Meynadier et al. (1994) cover 0–4 Ma and the cores of Tauxe and Hartl (1997) cover 11 Ma during the Oligocene. The power spectrum of the Tauxe and Hartl (1997) core, estimated with the Lomb periodogram, is presented in Fig. 3. Although the variations in time in this core are very different from that of Meynadier et al. (1994), the power spectrum is nearly proportional to $1/f$ just as for the Meynadier et al. (1994) data. This suggests that the $1/f$ spectrum is a universal feature of geomagnetic variations at these time scales.

Many analyses of variations in paleointensity of Earth's magnetic field concentrate on identifying characteristic time scales of variation. In contrast,

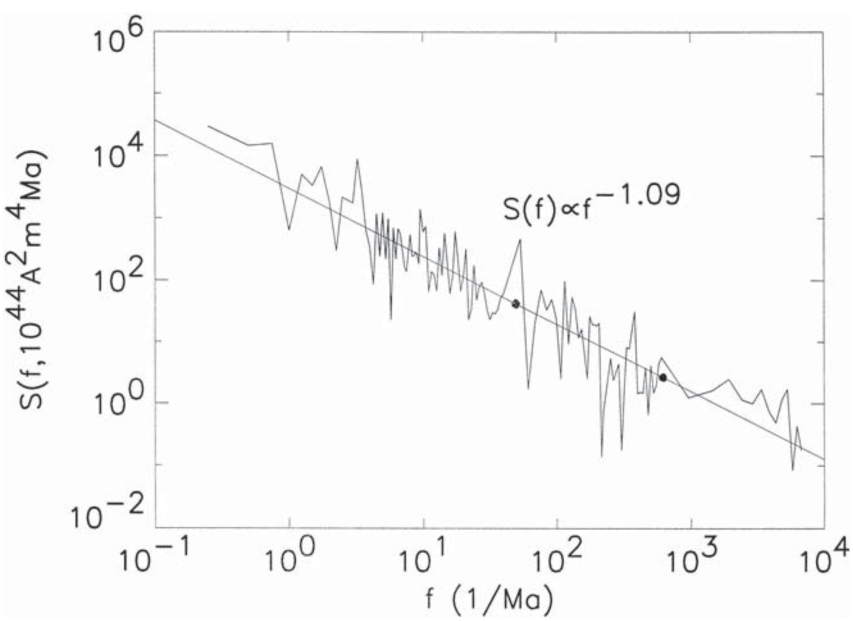

Fig. 2. Power spectrum of the geomagnetic field intensity variations estimated with the use of the Lomb periodogram from sediment cores of Meynadier et al. (1992) and Meynadier et al. (1994) and archeomagnetic data from Kovacheva (1980). Solid dots indicate the two frequencies where the three spectra meet. The power spectrum S is given as a function of frequency f for time scales of 100 a to 4 Ma.

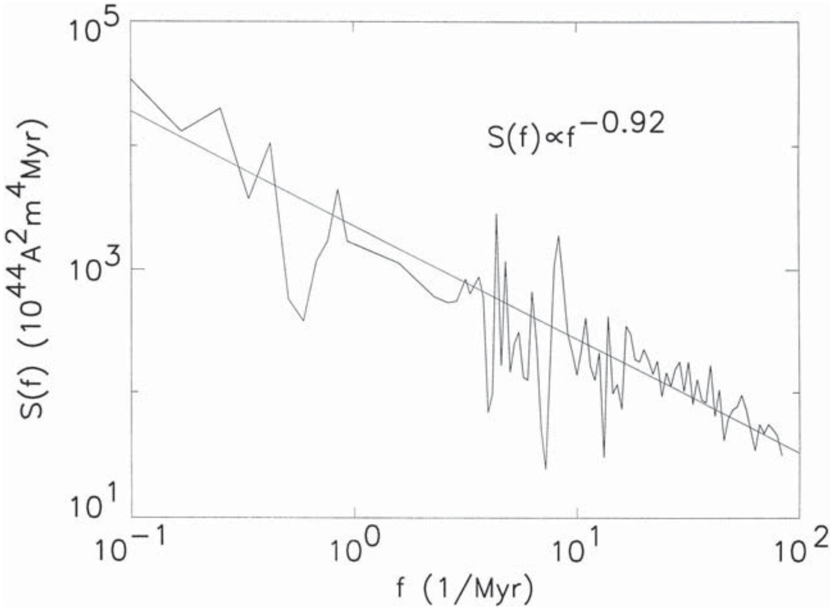

Fig. 3. Power spectrum of the geomagnetic field intensity variations estimated with the use of the Lomb periodogram from 11 Myr of sediment cores from the Oligocene obtained by Tauxe and Hartl (1997). The power spectrum S is given as a function of frequency f for time scales of 10 ka to 10 Ma.

our work shows that a continuous spectrum described by $1/f$ is most appropriate. Many such characteristic time scales have been identified. Valet and Meynadier (1993) suggested, based on the same sediment core data analyzed in this paper, that Earth's magnetic field regenerates following a reversal on a time scale of a few thousand years and then decays slowly on a time scale of 0.5 Ma before the next reversal. They termed this an 'asymmetric saw-tooth' pattern. More recent data have shown that the 'asymmetric saw-tooth' is not a robust pattern. Longer cores show a slow decay preceding a reversal to be rare (Tauxe and Hartl, 1997). These authors observed most intensity patterns between reversals to be characterized by broad, symmetric arches. Moreover, Lehman et al. (1996) has shown that the magnetic field does not always regenerate quickly after a reversal. McFadden and Merrill (1997) have tested the sawtooth-variation hypothesis against the record of reversals and found it to be unsupported. Thibal et al. (1995) have quantified the rate of decrease in field intensity preceding a reversal and found it to be inversely proportional to the length of the polarity interval. The authors concluded from this that the length of the reversal was predetermined. Such behavior is not indicative of a predetermined polarity length. This can be concluded by considering the null hypothesis that variations in the field are characterized by any stationary random process. By definition, a stationary time series has a variance which is independent of the length of the series. The average rate of change of the time series over a time interval will then be a constant value divided by the interval of time, i.e., inversely proportional to time interval. Therefore, any stationary random function satisfies the relationship that Thibal et al. (1995) observed.

In the power-spectral analyses of geomagnetic variations inferred from sediment cores by Lund et al. (1988), Meynadier et al. (1992), Lehman et al. (1996), Tauxe and Shackleton (1994), and Tauxe and Hartl (1997) dominant periodicities in the record were identified and proposed as characteristic time scales of geodynamo behavior. However, it must be emphasized that any finite length record will exhibit peaks in its power spectrum, even if the underlying process is random, such as a $1/f$ noise. Periodicity tests such as those developed by Lees and Park (1995) need to be applied to data in order to assess the probability that a peak in a spectrum is statistically significant. The periodicity tests developed by Lees and Park (1995) are especially valuable because they do not depend on a particular model of the stochastic portion of the spectrum. Some of the periodicity tests that have been used in the geomagnetism literature assume forms for the stochastic portion of the spectrum that are not compatible with the $1/f$ process we have identified. See Mann and Lees (1996) for an application of these techniques to climatic time series.

The power spectrum of secular geomagnetic intensity variations has been determined to have a $1/f^2$ power spectrum between time scales of one and one hundred years (Currie, 1968; Barton, 1982; Courtillot and Le Mouel, 1988). This is consistent with the analysis of McLeod (1992) who found that the first difference of annual means of geomagnetic field intensity is a white noise since the first differences of a random process with power spectrum $1/f^2$ is a white noise. Our observation of $1/f$ power-spectral behavior above time scales of approximately 100 yr together with the results of Currie (1968) and Barton (1982) suggests that there is a crossover from $1/f$ to $1/f^2$ spectral behavior at a time scale of approximately one hundred years.

3. Analysis of the reversal history

In this section, we will test the hypothesis that reversals are the result of intensity variations in one polarity state becoming large enough to cross the zero intensity value into the opposite polarity state. We will show that the statistics of the reversal record are consistent those of a binormal, $1/f$ noise paleointensity record which reverses when the intensity crosses the zero value. We will compare the polarity length distribution and the clustering of reversals between synthetic reversals produced with $1/f$ noise intensity variations and the reversal history according to Harland et al. (1990) and Cande and Kent (1992a, 1995). The sequence of polarity intervals according to the time scale of Harland et al. (1990) is presented in Fig. 4. The polarity interval length is plotted as a function of the order of occurrence. This

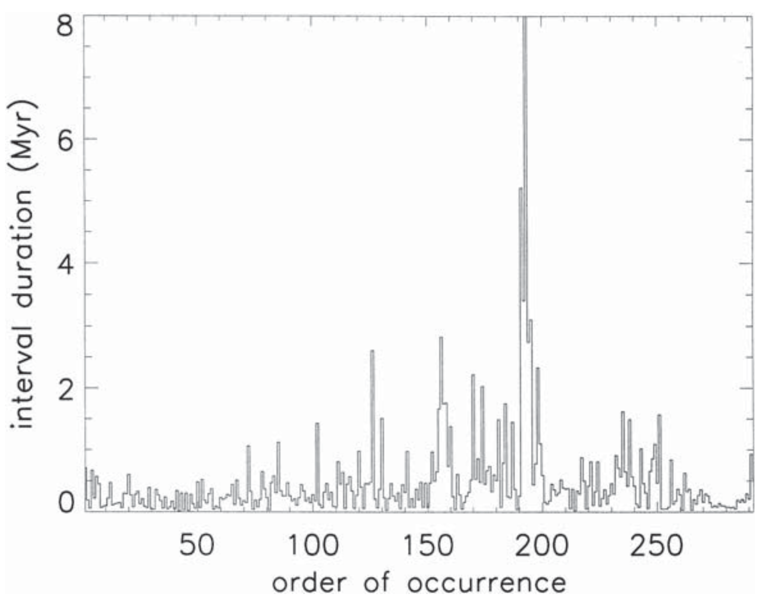

Fig. 4. Polarity interval duration as a function of order of occurrence from the magnetic time scale of Harland et al. (1990). The series of intervals shows a wide distribution and correlation of interval duration such that long intervals tend to be clustered with long intervals and short ones with short ones.

figure illustrates the broad distribution of polarity interval lengths that characterizes the reversal history (with both very short and very long polarity intervals) and the clustering of clustering of reversals with long polarity intervals clustered with long polarity intervals and short polarity intervals clustered with short ones. Note that in Fig. 4, the Cretaceous superchron is far off scale.

First we consider the polarity length distribution of the real reversal history. The polarity length distribution calculated from the chronology of Harland et al. (1990) is given as the solid line in Fig. 5. The polarity length distribution is the number of interval lengths longer than the length plotted on the horizontal axis. A reassessment of the magnetic anomaly data has been performed by Cande and Kent (1992a, 1995) to obtain an alternative magnetic time scale. Their chronology, normalized to the same length as the Harland et al. (1990) time scale, is presented as the dashed curve. The two distributions are nearly identical. These plots suggest that the polarity length distribution is better fit by a power law for large polarity lengths than by an exponential distribution, as first suggested by Cox (1968). The same conclusion has been reached by Gaffin (1989) and Seki and Ito (1993). The entire distribution of polarity lengths is also well fit by a Gamma process (McFadden and Merrill, 1986). It should be emphasized that the polarity length distribution is a very different analysis than that of the spectral analysis of Section 2. There is no simple relationship between the power spectrum of intensity variations and the polarity length distribution of reversals. The observation of a power-law power spectrum in Section 2 does not necessarily imply a power-law polarity length distribution.

The third curve, plotted with a dashed and dotted line, represents the polarity length distribution estimated from the magnetic time scale between C1 and C13 with 'cryptochrons' included and scaled to the length of the Harland et al. (1990) time scale. Cryptochrons are small variations recorded in the magnetic anomaly data that may either represent variations in paleomagnetic intensity or short reversals (Blakely, 1974; Cande and Kent, 1992b). Cryptochons occur with a time scale at the limit of temporal resolution of the reversal record from magnetic anomalies of the sea floor. The form of the polarity

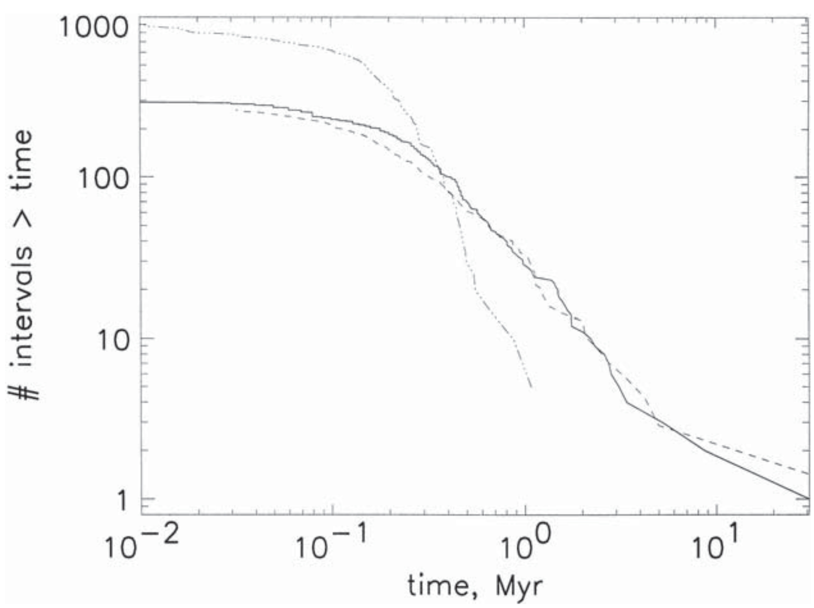

Fig. 5. Cumulative frequency-length distribution of the lengths of polarity intervals during the last 170 Ma from the time scale of Harland et al. (1990) (solid curve), Cande and Kent (1992a, 1995) (dashed curve), and the Cande and Kent (1992a, 1995) time scale from C1 to C13 with cryptochrons included (dashed and dotted line).

length distribution estimated from the record between C1 and C13 including cryptochrons is not representative of the entire reversal history because of the variable reversal rate which concentrates many short polarity intervals in this time period. However, this distribution enables us to estimate the temporal resolution of the reversal record history. The distribution estimated from C1 to C13 has many more short polarity intervals than those of the full reversal history starting at a reversal length of 0.3 Myr.

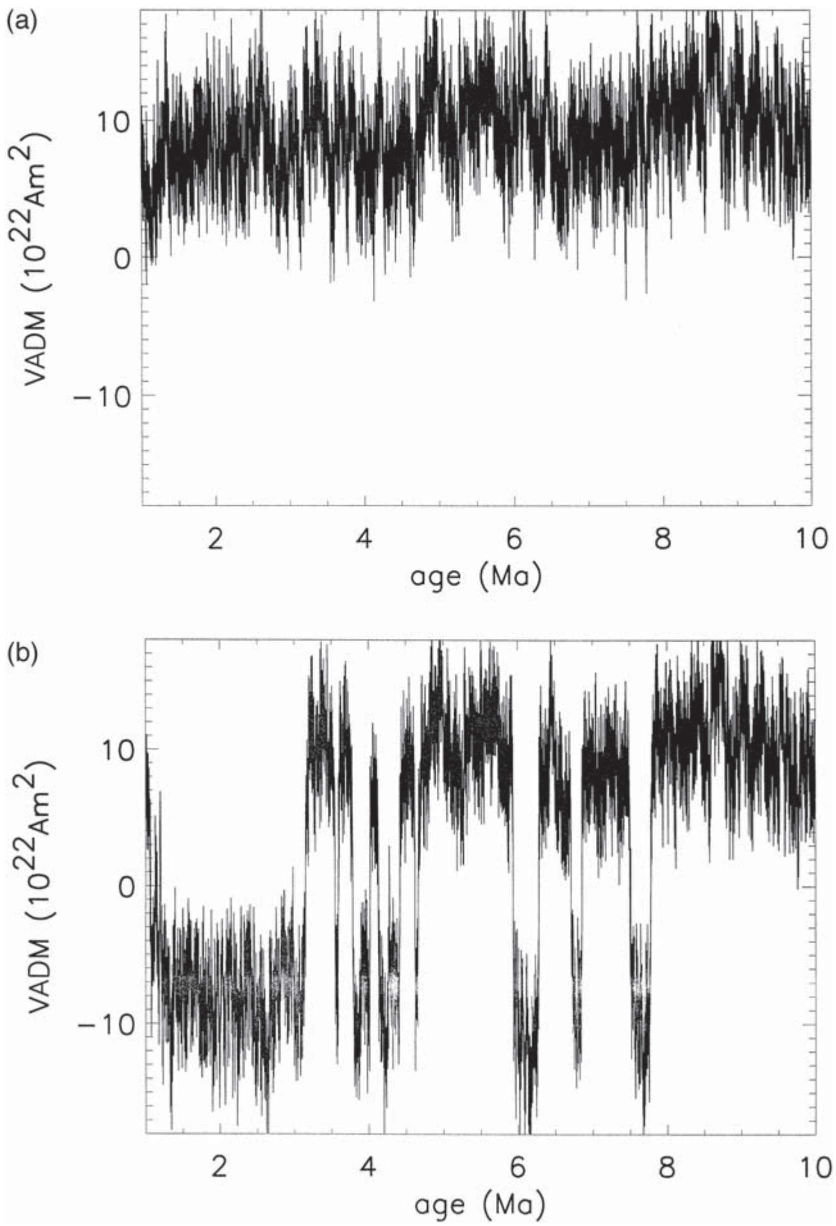

Fig. 6. (a) A $1/f$ noise with a normal distribution with mean of 8.9×10^{22} Am^2 and standard deviation of 3.4×10^{22} Am^2 representing the geomagnetic field intensity (VADM) in one polarity state. (b) Binormal $1/f$ noise constructed from the normal $1/f$ noise of (a) as described in the text.

Above a time scale of 0.3 Myr the magnetic time scale is nearly complete. Below it many short polarity intervals may be unrecorded.

To show that the polarity length distribution of the real reversal record is consistent with that produced by binormal, $1/f$ noise variations in Earth's dipole moment, we have generated synthetic Gaussian noises with a power spectrum proportional to $1/f$, a mean value of 8.9×10^{22} Am^2 and a standard deviation of 3.4×10^{22} Am^2. These synthetic noises represent the field intensity in one polarity state. The synthetic noises were generated using the Fourier-domain filtering technique described in Turcotte (1992). An example is shown in Fig. 6a. In order to construct a binormal intensity distribution from the synthetic normal distribution, we inverted every other positive polarity interval to the opposite polarity. This 'cut-and-paste procedure' is the best approximation of the statistical hypothesis we wish to test. We wish to test the hypothesis that Earth's magnetic field undergoes Gaussian $1/f$ noise variations in one polarity state until the intensity crosses the zero intensity value and the geodynamo is pushed into another basin of attraction characterized by the same $1/f$ variations with the same variance but now with a negative mean value. One way to construct such a stochastic process is to generate a single Gaussian $1/f$ noise with positive mean and, when the intensity crosses the zero axis, take the next continuous data segment above zero and invert it below the zero intensity axis to simulate the geodynamo flipping into another basin of attraction characterized by a negative mean value. The result of this procedure on the Gaussian, $1/f$ noise of Fig. 6a is presented in Fig. 6b. We note that there are several instances in Fig. 6 where it appears that a single zero crossing from Fig. 6a does not lead to a reversal in Fig. 6b. However, in these cases the zero crossings of Fig. 6a contain two zero crossings so close together that they appear to be one. The irregular polarity lengths of Fig. 6b are similar to those in the marine sediment data of Fig. 1. The synthetic polarity lengths produced by this model as a function of the order of occurrence are presented in Fig. 7. This figure can be compared with that of the real reversal history of Fig. 4. Although the model does not predict when

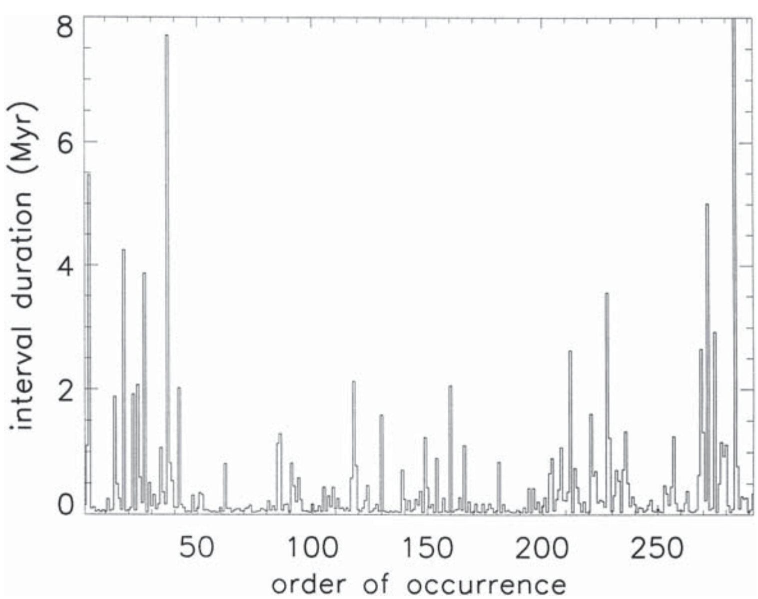

Fig. 7. Polarity interval duration as a function of order of occurrence for the synthetic reversals produced with binormal $1/f$ noise intensity fluctuations. As with the corresponding plot for the real magnetic time scale in Fig. 4, the series of intervals shows a wide distribution and correlation of interval duration such that long intervals tend to be clustered with long intervals and short ones with short ones. The distribution of intervals for the synthetic distribution are presented in Fig. 8 and their correlation is analyzed with the pair-correlation function presented in Fig. 10.

the polarity intervals will be longest or shortest (since no stochastic model can predict behavior in time), there is a wide distribution of polarity intervals and the polarity intervals tend to be autocorrelated as in the real reversal history.

The operation of reversing the paleomagnetic intensity when it crosses the zero intensity value is consistent with models of the geodynamo as a system with two symmetric attracting states of positive and negative polarity such as the Rikitake disk dynamo (Rikitake, 1958). Between reversals, the geomagnetic field fluctuates until a fluctuation large enough occurs to cross the energy barrier into the other basin of attraction. Kono (1987) has explored the statistical similarity between the Rikitake disk dynamo and the distribution of paleointensity. Our construction of the binormal $1/f$ noise is consistent with this model.

We have computed the distribution of lengths between successive reversals for twenty synthetic noises scaled to 170 Ma, the length of the reversal chronology, and averaged the results in terms of the number of reversals. The results are given in Fig. 8. The dots are the maximum and minimum values obtained in the twenty synthetic reversal chronologies, thus representing 95% confidence intervals. The polarity interval distribution from the Harland et al. (1990) time scale indicated by the dashed curve falls within the 95% confidence intervals of our synthetic data over all time scales plotted except for the Cretaceous superchron, which lies slightly outside of the 95% confidence interval and reversals separated by less than about 0.3 Myr. The overprediction of very short reversals could be a limitation of the model or a result of the incompleteness of the reversal record for short polarity intervals. As mentioned, the temporal resolution of the magnetic time scale inferred from magnetic anomalies is approximately 0.3 Myr. In addition, the model fails to produce a polarity interval as long or longer than the Cretaceous normal superchron in twenty synthetic reversal histories. The superchron cannot, therefore, be reproduced in this model with a high probability. Nevertheless, we conclude that the polarity length distribution produced from binormal $1/f$ intensity variations are consistent with the observed polarity length distribution for nearly all time scales at which the reversal record is complete.

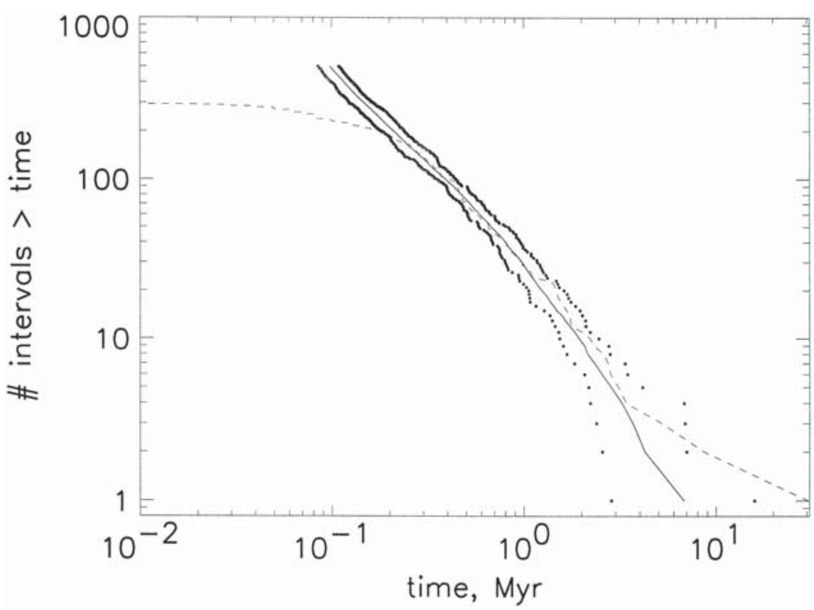

Fig. 8. Cumulative frequency-length polarity interval distributions from the Harland et al. (1990) time scale and that of the binormal, $1/f$ noise model of intensity variations. The distribution from the Harland et al. (1990) time scale (dashed curve) was also given in Fig. 4. The solid line represents the average cumulative distribution from the $1/f$ noise model. The dotted lines represent the minimum and maximum reversal length distributions for 20 numerical experiments, thereby representing 95% confidence intervals.

We next consider whether the agreement illustrated in Fig. 8 is unique to $1/f$ noise. We have computed polarity length distributions using the binormal intensity variations with power spectra $f^{-0.8}$ and $f^{-1.2}$. These results along with the $1/f$ result from Fig. 8 are given in Fig. 9. The shape of the polarity length distribution is very sensitive to the exponent of the power spectrum. A slight increase in the magnitude of the exponent results in many more long polarity intervals than with $1/f$ noise. The analysis of Fig. 2 leads to $1/f^{1.09}$ and that of Fig. 3 leads to $1/f^{0.92}$. The polarity-length distribution produced by intensity variations with these spectra, as with those with $1/f^{1.2}$ and $1/f^{0.8}$ spectra, are not as consistent with that of $1/f$ noise variations, which show a very close match to the polarity-length distribution of real reversals. Here we have implicitly assumed that, to within the uncertainty in the data, the spectrum is $1/f$. Either the exponents 1.09 and 0.92 are exact exponents characterizing geomagnetic processes during the times that each time series covers and the synthetic reversals produced by variations with $1/f^{1.09}$ and $1/f^{0.92}$ spectra do not match the reversal distribution very well, or the spectrum is closer to $1/f$ and the values of the exponents 1.09 and 0.92 are only approximate. We have assumed that the latter is true. The spectral analysis we employ here is probably not capable of determining the exponent to better than 10%. The values of 1.09 and 0.92 are obtained from a least-square fit to a series of unevenly spaced points and neither value should be considered an exact value. We conclude that the agreement in Fig. 8 between the synthetic reversal distribution and the true reversal history is unique to $1/f$ noise and provides strong evidence that the intensity of Earth's magnetic field has $1/f$ behavior up to 170 Ma.

A binormal, $1/f$ noise geomagnetic field variation is consistent with the qualitative results of Pal and Roberts (1988) and Tauxe and Hartl (1997) who found an anticorrelation between reversal frequency and paleointensity. This anticorrelation, first suggested by Cox (1968), is evident in the synthetic $1/f$ noise of Fig. 6b. During the time intervals of greatest average paleointensity the reversal rate is lowest.

In addition to the broad distribution of polarity lengths, the reversal history is also characterized by a clustering of reversals. This behavior has been quan-

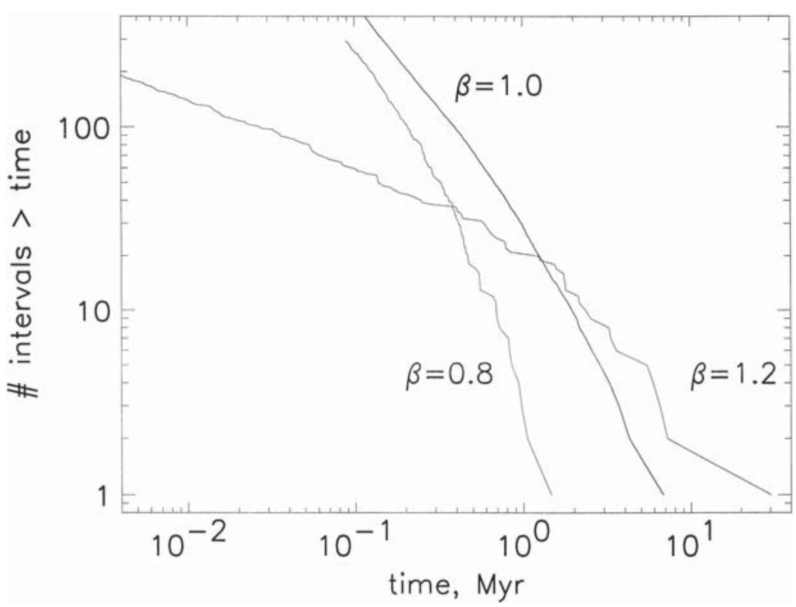

Fig. 9. Cumulative frequency-length polarity interval distributions for the $1/f$ noise model of intensity variations (shown in the middle, also given in Fig. 5) and for intensity variations with power spectra proportional to $f^{-0.8}$ and $f^{-1.2}$. This plot illustrates that the polarity length distribution is very sensitive to the form of the power spectrum, allowing us to conclude that the agreement between the model and the observed distribution in Fig. 5 is unique to $1/f$ noise intensity variations.

tified with the reversal rate. The reversal rate has been relatively high from 0–20 Ma and has decreased gradually going back in history to the Cretaceous superchron. An alternative approach to quantifying the clustering of reversals is with the pair correlation function. The pair correlation function $c(t)$ is the number of pairs of reversals whose separation is between t and $t+\Delta t$, per unit time (Vicsek, 1992) where Δt is a time bin chosen to be small enough to have adequate temporal resolution of $c(t)$ but large enough to include enough pairs of reversals for minimal variance between adjacent bins. The pair correlation function for a set of points can be compared to that for a Poisson process to detect non-random clustering. The pair-correlation function analysis is more appropriate for comparison of the reversal history to the synthetic reversal history generated by a stochastic model such as model based on turbulent processes in the core. This is because stochastic models cannot predict behavior in time, such as when the reversal rate is large or small. However, a stochastic model may accurately reflect the extent to which small polarity intervals are followed by small polarity intervals and long intervals by long intervals as quantified with the pair correlation function.

The pair correlation function of reversals according to the Harland et al. (1990) and Cande and Kent (1992a, 1995) reversal history are shown in Fig. 10 as filled and unfilled circles, respectively. We have used logarithmically-increasing time bins since we wished to have better resolution for small t where $c(t)$ decreases rapidly. Also presented in Fig. 10 is the pair correlation function for a synthetic reversal data set based on binormal $1/f$ noise dipole moment variations (boxes) and the pair correlation function for a Poisson process (triangles). The functions are offset so that they may be placed on the same graph. The Poisson process was constructed with 293 points, the same number of reversals as the Harland et al. (1990) time scale, positioned with uniform probability on the interval between 0 and 170 Ma. The Poisson process yields a correlation function independent of separation. The real and synthetic reversal histories variations exhibit significant clustering with more pairs of points at small separation and fewer at large separations than for a Poisson process. Straight-line fits of the form $c(t) \propto t^{\alpha}$ were obtained. The purpose of this was to show that similar

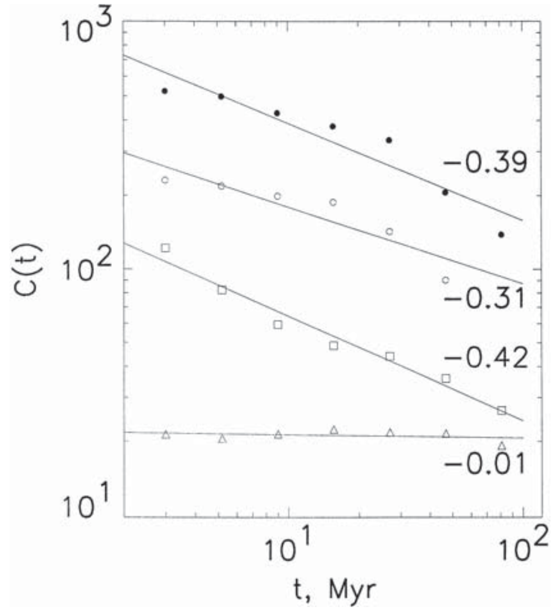

Fig. 10. Pair correlation function of the reversal history according to the Harland et al. (1990) time scale (filled circles), Cande and Kent (1992a, 1995) (unfilled circles), synthetic reversals produced from $1/f$ noise model of intensity variations (boxes), and a Poisson process (triangles). The real and synthetic reversals exhibit similar non-random clustering.

clustering is observed in the real and synthetic reversals. The exponents α of the Harland et al. (1990), Cande and Kent (1992a, 1995), and synthetic reversals are -0.39, -0.31, and -0.42, respectively. Similar non-random clustering is observed in the real and synthetic reversals. We conclude from the consistency between the polarity interval distribution and pair correlation function of real reversals and those generated by fluctuations of a bionormal $1/f$ noise process that there is strong evidence for $1/f$ intensity variations for time scales up to the length of the reversal record.

It is generally believed that secular geomagnetic variations are the result of internal dynamics while longer time scale phenomena such as variations in the reversal rate are controlled by variations in boundary conditions at the core–mantle boundary (CMB) (McFadden and Merrill, 1995). Variations in mantle activity have been proposed as the driving force for reversals (Loper and McCartney, 1986; Gubbins, 1987; Larson and Olson, 1991). However, our observation of continuous $1/f$ spectral behavior from time scales of 100 yr to 170 Myr suggests that internal variability may control variations in geomagnetic intensity over the entire range of time scales analyzed here. Indeed, McFadden and Merrill (1995) state that due to the complex, nonlinear aspects of the dynamo process, it may be that reversals themselves and the long-term variations in their mean rate of occurrence are all manifestations of deterministic chaos.

4. Conclusions

We have presented the results of power-spectral analyses of variations in the intensity of Earth's magnetic field which show the spectrum to be proportional to $1/f$ from time scales of 100 yr to 4 Myr. We have shown that reversals generated by binormal, $1/f$ noise geomagnetic field intensity variations are consistent with the distribution of polarity lengths and clustering of real reversals.

Acknowledgements

I wish to thank Donald L. Turcotte for helpful conversations and Laure Meynadier and Lisa Tauxe for access to the marine sediment data and helpful conversations regarding its interpretation. I also wish to thank two anonymous reviewers for their careful reviews.

References

Barton, C.E., 1982. Spectral analysis of palaeomagnetic time series and the geomagnetic spectrum. Phil. Trans. R. Soc. London A 306, 203–209.

Blakely, R.J., 1974. Geomagnetic reversals and crustal spreading rates during the Miocene. J. Geophys. Res. 79, 2979–2985.

Cande, S.C., Kent, D.V., 1992a. A new geomagnetic polarity time scale for the Late Cretaceous and Cenozoic. J. Geophys. Res. 97, 13917–13951.

Cande, S.C., Kent, D.V., 1992b. Ultrahigh resolution marine magnetic anomaly profiles: a record of continuous paleointensity variations?. J. Geophys. Res. 97, 15075–15083.

Cande, S.C., Kent, D.V., 1995. Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic. J. Geophys. Res. 100, 6093–6095.

Courtillot, V., Le Mouel, J.L., 1988. Time variations of the Earth's magnetic field: from daily to secular. Ann. Rev. Earth Planet. Sci. 16, 389–476.

Cox, A., 1968. Lengths of geomagnetic polarity intervals. J. Geophys. Res. 73, 3247–3260.

Currie, R.G., 1968. Geomagnetic spectrum of internal origin and lower mantle conductivity. J. Geophys. Res. 73, 2768–2779.

Gaffin, S., 1989. Analysis of scaling in the geomagnetic polarity reversal record. Phys. Earth Planet. Inter. 57, 284–290.

Gubbins, D., 1987. Mechanism for geomagnetic polarity reversals. Nature 326, 167–169.

Guyodo, Y., Valet, J.-P., 1996. Relative variations in geomagnetic intensity from sedimentary records: the past 200,000 years. Earth Planet. Sci. Lett. 143, 23–36.

Harland, W.B., Cox, A., Llewellyn, P.G., Pickton, C.A.G., Smith, A.G., Walters, R., 1990. A Geologic Time Scale 1989. Cambridge Univ. Press.

King, J.W., Banerjee, S.K., Marvin, J., 1983. A new rock-magnetic approach to selecting sediments for geomagnetic paleointensity studies: Application to paleointensity for the last 4000 years. J. Geophys. Res. 88, 5911–5921.

Kono, M., 1971. Intensity of the Earth's magnetic field during the Pliocene and Pleistocene in relation to the amplitude of mid-ocean ridge magnetic anomalies. Earth and Planet. Sci. Lett. 11, 10–17.

Kono, M., 1987. Rikitake two-disk dynamo and paleomagnetism. Geophys. Res. Lett. 14, 21–24.

Kovacheva, M., 1980. Summarized results of the archeomagnetic investigation of the geomagnetic field variation for the last 8000 yr in south-eastern Europe. Geophys. J. R. Astr. Soc. 61, 57–64.

Larson, R.L., Olson, P., 1991. Mantle plumes control magnetic reversal frequency. Earth and Planet. Sci Lett. 107, 437–447.

Lees, J.M., Park, J., 1995. Multiple-taper spectral analysis: a stand-alone C C-subroutine. Comp. Geol. 21, 199–236.

Lehman, B., Laj, C., Kissel, C., Mazaud, A., Paterne, M., Labeyrie, L., 1996. Relative changes of the geomagnetic field intensity during the last 280 kyear from piston cores in the Acores area. Phys. Earth Planet. Inter. 93, 269–284.

Loper, D.E., McCartney, K., 1986. Mantle plumes and the periodicity of magnetic field reversals. Geophys. Res. Lett. 13, 1525–1528.

Lund, S.P., Liddicoat, J.C., Lajoie, K.R., Henyey, T.L., Robinson, S.W., 1988. Paleomagnetic evidence for long-term (10^4 year) memory and periodic behavior in the Earth's core dynamo process. Geophys. Res. Lett. 15, 1101–1104.

Mann, M.E., Lees, J.M., 1996. Robust estimation of background noise and signal detection in climatic time series. Climatic Change 33, 409–445.

McFadden, P.L., Merrill, R.T., 1986. Geodynamo energy source constraints from paleomagnetic data. Earth Planet. Sci Lett. 42, 22–33.

McFadden, P.L., Merrill, R.T., 1995. History of the Earth's magnetic field and possible connections to core–mantle boundary processes. J. Geophys. Res. 100, 307–316.

McFadden, P.L., Merrill, R.T., 1997. Sawtooth paleointensity and reversals of the geomagnetic field. Phys. Earth Planet. Inter. 103, 247–252.

McLeod, M.G., 1992. Signals and noise in magnetic observatory annual means: mantle conductivity and jerks. J. Geophys. Res. 97, 17261–17290.

Meynadier, L., 1995. Personal communication.

Meynadier, L., Valet, J.-P., Weeks, R., Shackleton, N.J., Hagee, V.L., 1992. Relative geomagnetic intensity of the field during the last 140 ka. Earth Planet. Sci. Lett. 114, 39–57.

Meynadier, L., Valet, J.-P., Bassonot, F.C., Shackleton, N.J., Guyodo, Y., 1994. Asymmetrical sawtooth pattern of the geomagnetic field intensity from equatorial sediments in the Pacific and Indian oceans. Earth Planet. Sci Lett. 126, 109–127.

Pal, P.C., Roberts, P.H., 1988. Long-term polarity stability and strength of the geomagnetic dipole. Nature 331, 702–705.

Press, W.H., Teukolsky, S.A., Vetterling, W.T., Flannery, B.P., 1992. Numerical Recipes in C: The Art of Scientific Computing. Cambridge Univ. Press.

Rikitake, T., 1958. Oscillations of a system of disc dynamos. Proc. Cambridge Phil. Soc. 54, 89–105.

Seki, M., Ito, K., 1993. A phase-transition model for geomagnetic polarity reversals. J. Geomag. Geoelectr. 45, 79–88.

Tanaka, H., Kono, M., Uchimara, H., 1995. Some global features of paleointensity in geological time. Geophys. J. Int. 120, 97–102.

Tauxe, L., 1993. Sedimentary records of relative paleointensity of the geomagnetic field: theory and practice. Rev. Geophys. 31, 319–354.

Tauxe, L., 1998. Personal communication.

Tauxe, L., Hartl, P., 1997. 11 million years of Oligocene geomagnetic field behavior. Geophys. J. Int. 128, 217–229.

Tauxe, L., Shackleton, N.J., 1994. Relative paleointensity records from the Ontong-Java Plateau. Geophys. J. Int. 117, 769–782.

Thibal, J., Pozzi, J.-P., Barthes, V., Dubuisson, G., 1995. Continuous record of geomagnetic field intensity between 4.7 and 2.7 Ma from downhole measurements. Earth Planet. Sci. Lett. 136, 541–550.

Turcotte, D.L., 1992. Fractals and Chaos in Geology and Geophysics. Cambridge Univ. Press.

Valet, J.-P., Meynadier, L., 1993. Geomagnetic field intensity and reversals during the past four million years. Nature 366, 234–238.

Vicsek, T., 1992. Fractal Growth Phenomena, World Scientific.