PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Spectral Analysis of Palaeomagnetic Time Series and the Geomagnetic Spectrum [and Discussion]

C. E. Barton and F. J. Lowes

Phil. Trans. R. Soc. Lond. A 1982 **306**, 203-209
doi: 10.1098/rsta.1982.0080

Phil. Trans. R. Soc. Lond. A **306**, 203–209 (1982) [203]
Printed in Great Britain

Spectral analysis of palaeomagnetic time series and the geomagnetic spectrum

By C. E. Barton†
Geophysics Department, University of Edinburgh,
Mayfield Road, Edinburgh EH9 3JZ, U.K.

There are now sufficient archaeomagnetic data from rapidly deposited sediments and baked clays to start bridging the gap in the geomagnetic spectrum between the frequency ranges covered by observatory records and polarity reversals. The form of the continuum spectrum of internal origin can be only loosely constrained but is broadly consistent with earlier speculations. The power spectral density function appears to increase rapidly with period up to periods of about 60 years, then more slowly up to a plateau in the region of 10^4 to 10^5 years, and thereafter starts to fall. There is somewhat inconclusive evidence for a drop in power density at periods around 10^2 years. Prospects for refining the spectrum are excellent.

1. Introduction

Any satisfactory explanation of the origin of the geomagnetic secular variation must predict the form of the continuous power spectrum (Braginskiy 1970). Observatory data have been used to obtain spectra for the variation in vertical and horizontal components of the field in the periodicity range 1–100 years (Currie 1968; Alldredge 1977). The form of the spectrum at longer periods has so far been left to the realm of speculation (see, for example, Doell & Cox 1972). This paper describes an initial attempt to use archaeomagnetic data from baked clays and recently deposited sediments to provide observational bounds for the continuous spectrum up to periods of about 10^5 years.

The frequency domain falls roughly into five poorly defined bands characterized by the types of observation that are possible (figure 1). Except at high frequencies, the source mechanism given is largely speculative. 'Internal' dipole variations refers to the intrinsic magnetohydrodynamic (m.h.d.) behaviour of the core, as opposed to variations excited externally, e.g. by changes in global ice volumes (Rampino 1979) or orbital parameters of the Earth (see, for example, Wollin *et al.* 1978; Chave & Denham 1979). The classification of source mechanisms in figure 1 follows the pattern of previous suggestions (Jacobs 1975, p. 122).

2. Choice of variable

No single parameter adequately represents the variation in both magnitude and direction of the geomagnetic field at a site. For observatory records we can compute power spectra for individual Cartesian components: H (horizontal), X (north), Y (east) and Z (down). There are sufficient worldwide archaeomagnetic data (Cox 1968; Bucha 1969; Smith 1970; Burlatskaya & Nachisova 1977) to describe the gross variation of the Earth's dipole moment during the last

† Present address: Graduate School of Oceanography, University of Rhode Island, Narragansett, Rhode Island 02882, U.S.A.

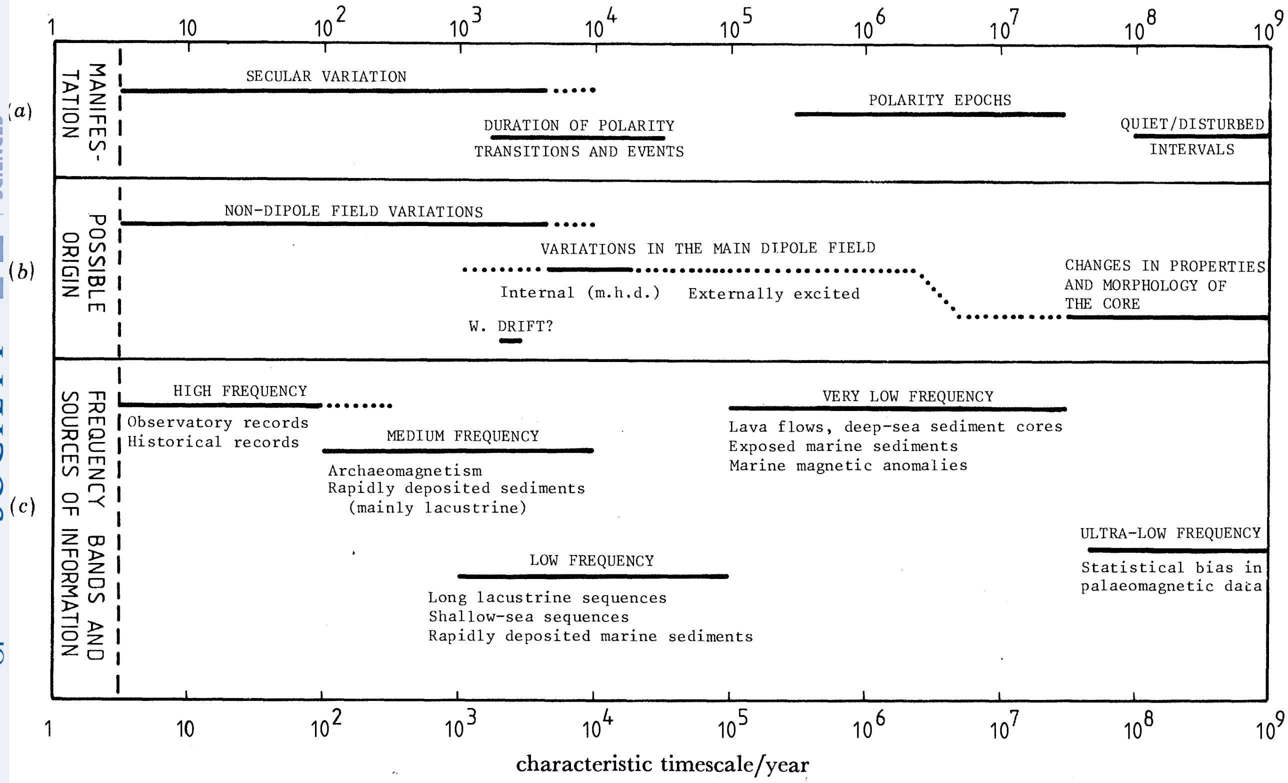

FIGURE 1. Principal frequency bands of the geomagnetic spectrum of internal origin, showing (*a*) how the variations are manifested, (*b*) hypothesized origins, and (*c*) frequency band classification and sources of observational data.

10000 years, but the discontinuous nature of the records at all but a few sites makes them generally unsuitable for the type of analysis that can be applied to observatory records. Before about 10000 years ago our information about the field is essentially directional only.

By virtue of the geometry of different types of source mechanism (e.g. dipole wobble, drifting and oscillating radial dipoles and current loops), the spectra of declination and inclination time series may be somewhat different. Declination is particularly ill-suited as a variable for spectral analysis owing to its instability in the region of a dip pole. Inclination is more satisfactory as it is more easily determined (being independent of the azimuthal orientation of samples), and a correction can be applied to allow for the latitudinal dependence of inclination change with respect to dipole movement. Disadvantages are that the inclination spectrum is not directly related to spectra of Z and H, which are commonly computed for observatory data, and that in sediments inclinations are subject to bedding errors (Johnson *et al.* 1948). The compromise used herein is to sum contributions from each component to the spectral power at a given frequency:

$$SS(f) = S_X(f) + S_Y(f) + S_Z(f).$$

If spectra for H and Z only are available, S_H is used instead of $S_X + S_Y$. For series of directional data the magnetic vector is assigned a constant magnitude, equal to that due to an axial geocentric dipole of present-day moment (8.0×10^{22} A m^2). This value is chosen as being fairly typical for both the recent (Holocene) and the remote past. During the past 10000 years the dipole moment, as calculated essentially from European data, has displayed a strong sinusoidal

Table 1. Observatory annual mean data analysed

code	observatory	latitude (°N)	longitude (°E)	interval analysed/year
MAUR	Mauritius (Royal Alfred)	−20.10	57.55	74
TANA	Tananarive	−18.92	47.55	72
AGIN	Agincourt	43.78	280.73	70
ALIB	Alibag	18.63	72.87	74
VALE	Valentia	51.93	349.75	82
COIM	Coimbra	40.22	351.58	110
STON	Stonyhurst	53.85	357.53	75
SFER	San Fernando	36.47	353.80	86
ESKD	Eskdalemuir	55.32	356.80	72
RUDE	Rude Skov	55.84	12.46	72
GREE	Greenwich	51.48	0.00	80
APIA	Apia	−13.80	188.22	74
PILA	Pilar	−31.67	296.12	70

Observations are listed in order of descending spectral power density at the low frequency cutoff in figure 2.

Table 2. Palaeomagnetic and archaeomagnetic time series analysed

code	location	latitude (°N)	F_{ad}† / μT	source material	interval analysed / year B.P.	data spacing / year	reference
BI	Lake Biwa, Japan	35.2	43.8	single 50 m core	2800–63000	200	1
BL	Black Sea, station 1474	42.3	47.6	single core	7250–23500	250	2
BU	Lake Bullenmerri, SE Australia	−38.3	45.5	eight 6 m cores	50–9750	100	3, 4
CH	Lake Charzykowski, N. Poland	53.7	53.3	two 6 m cores lake sediments	50–5550	100	5, 6
GN	Lake Gnotuk, SE Australia	−38.3	45.5	three 4½ m cores	500–10600	100	3, 4
JA	SW Japan	35	43.7	archaeomagnetic samples	200–1940	20	7
KE	Lake Keilambete, SE Australia	−38.2	45.4	six 4½ m cores lake sediments	0–11100	100	3, 4
LWG	Loch Lomond, Lake Windermere and Llyn Geirionydd, U.K.	54.5	53.6	sets of 6 m cores	40–6960	40	8
P43	NW Pacific Ocean, core V12/36–43P	33.8	43.0	12.2 m piston core	5000–605000	5000	9
P45	NW Pacific Ocean, core V12/36–45P	31.1	41.8	12.25 m piston core	20000–650000	5000	9

References: 1, Yaskawa *et al.* (1973); 2, Creer (1974); 3, Barton & Polach (1980); 4, Barton & McElhinny (1981); 5, Creer *et al.* (1979); 6, Tucholka (1980); 7, Hirooka (1971); 8, Turner & Thompson (1981); 9, C. Barton, unpublished data.

† The field strength at the site corresponding to a geocentric axial dipole of moment 8.0×10^{22} A m^2.

trend (present phase ≈ π, period ≈ 7500 years) which would appear as a sharp peak in the power spectra of X, Y and Z. However, this trend is less clear in results from other parts of the world and, furthermore, data for the last 50000 years indicate that this periodicity is not a persistent characteristic of the field (McElhinny & Senanayake 1981). Although $SS(f)$ is not a measure of the variance in any single, measurable property of the field, it is sensitive to any type of field variation. In this respect it is superior as a parameter for monitoring the spectral character of the field to either $S_Z(f)$ or $S_H(f)$, or to the powers inthe declination and inclination time series.

3. Results

A pilot study has been completed by using the data set listed in tables 1 and 2. Although many analyses of observatory annual mean data have already been published (see, for example, Currie 1968; Alldredge 1977), a new analysis was done to ensure that all records are processed in the same manner. Most of the palaeomagnetic and archaeomagnetic data sets were smoothed by taking inter-quartile means (i.e. the mean after discarding the upper and lower quartiles)

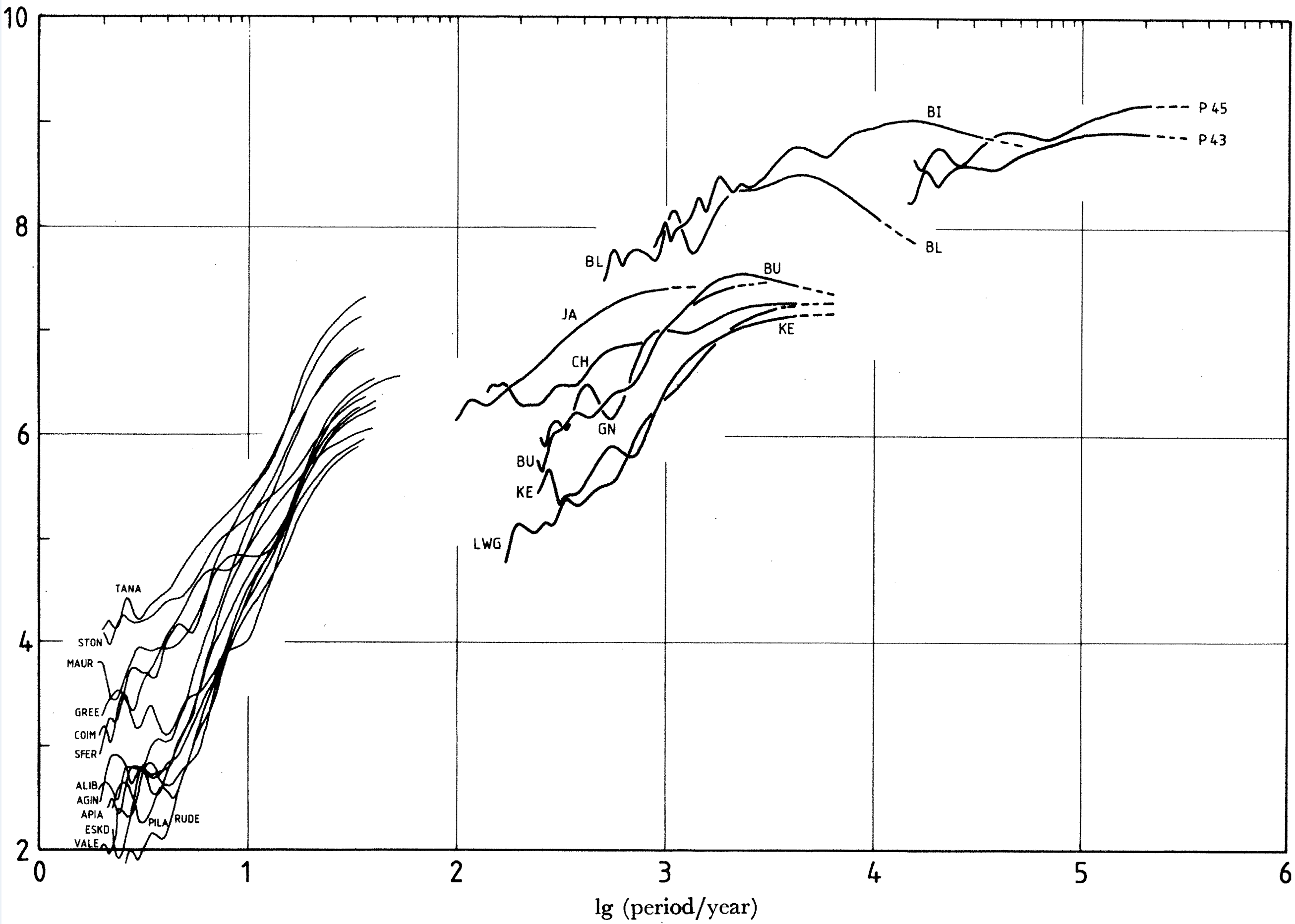

Figure 2. Compilation of summed spectral power density estimates. Spectra are shown dashed for periods longer than half the extended record lengths, and are not plotted beyond the 40th harmonic of the record lengths. Abbreviations are listed in tables 1 and 2.

over equal time intervals after transformation from depth to time scales by using linear interpretation between dated horizons. Radiocarbon ages were corrected to calendar years (after Clark (1975), with a constant correction of 800 years before to 6700 years B.P.).

Periodogram analysis has been used following the scheme described by Bloomfield (1976). Series were de-trended with a first-order polynomial, tapered with a split cosine belt covering 10 % of the data at each end, then padded with zeros to the next radix-2 number of points. Square moduli of the Fourier transforms of X, Y and Z series were summed before computing the periodograms. Spectral power density estimates were obtained by smoothing periodograms with three passes of a modified Daniell filter of half-width 2 (bandwidth = 13 periodogram ordinates (see Bloomfield 1976, p. 169)).

Results of this analysis (figure 2) demonstrate that only weak constraints can be placed on the form of the continuous geomagnetic spectrum beyond periods of about 10^2 years. Spectral density estimates from lacustrine (and to a lesser extent archaeomagnetic) records will be too low owing to a loss of resolution during stacking and averaging, irregularities in timescales, and imperfections in the palaeomagnetic recording process. These effects will be particularly apparent at high frequencies. For the two deep-sea core records, however, there is little doubt that the power levels observed are due to noise, and should therefore be treated as upper bounds to the true spectrum. Some measure of the true uncertainty in the spectral estimates may be guaged from a comparison between the three curves from SE Australia (KE, BU, GN) in figure 2. These records are each based on unusually large data sets and each have comprehensive radiocarbon chronologies. Spectral estimates at low frequencies are sensitive to the form of detrend applied to the data, particularly for observatory records that display strong trends (Courtillot *et al.* 1977) Thus the absolute scaling at the low-frequency end of the observatory spectra in figure 2 is not very reliable. Despite the above limitations the results are of considerable interest.

4. Discussion and conclusions

If we assume that spectral power densities for the lacustrine records are too low by about one order of magnitude (amplitudes attenuated by approximately threefold), and that power densities for P43 and P45 are too high at very low frequencies, then the complete spectrum can be divided into four zones of distinctive character:

(*a*) periods less than about 5 years where the gradient in figure 2 is low and there is considerable variability between observatories;

(*b*) periods between about 5 and 30 years, characterized by high gradients and little variation between observatories;

(*c*) periods between about 30 years and 10^4 years where the gradient is intermediate, possibly with a dip near 10^2 years, and

(*d*) periods beyond about 10^4 years where the power density levels off at about 10^9 nT^2 year and possibly start to fall.

It was expected that a change in power level and character of the spectrum might delineate the different frequency domains of non-dipole and dipole source mechanisms. This does not appear to be so in any obvious manner. Verosub & Cox (1971) have observed that during this century the total magnetic energy of the field external to the core has remained fairly constant, and the decay in energy of the dipole field has been largely (though not completely) offset by an increase in energy of the non-dipole field. The present results suggest that this might be a permanent characteristic of the field, and that no definable spectral boundary exists between the frequency ranges of dipole and non-dipole field effects other than a broad maximum in the continuum around periods of 10^4 years. This period coincides with that of Braginskiy's (1970) fundamental oscillation of the magnetohydrodynamic dynamo.

Within zones (*a*) and (*c*) variations in the geomagnetic field are presumably controlled by external and internal source mechanisms respectively. The shape of the spectrum suggests that there is no clear-cut boundary separating their periodicity ranges. Alldredge (1977) considers that variations with periods greater than 13 years are of internal origin.

If the spectrum does dip at periods around 10^2 years, a peak must occur in the region of 50–60 years. A concentration of power in the geomagnetic spectrum at about 60 years has been noted by many authors (e.g. Vestine & Kahle 1968; Yukutake 1973; Currie 1973; Braginskiy & Fishman 1977). However, the apparent dip is defined by the high-frequency information in the palaeomagnetic records, which is generally the most poorly resolved part of the geomagnetic signal. Furthermore, any phase irregularities in high-frequency components, for example due to timescale imperfections, will be manifested as a reduction in power density, particularly at high frequencies. An analysis of high-quality archaeomagnetic data covering the last 1000 years will, it is hoped, resolve this uncertainty.

There is considerable scope for refining the final spectrum. Some of the problems affecting the high-frequency end of palaeomagnetic spectra can be avoided by stacking 'dynamic' spectra computed for sliding sequences of a limited length of each record. This would help to resolve the uncertainty about the spectrum between periods of 50 and 500 years. A more comprehensive data set is required with emphasis on archaeomagnetic and high-resolution deep-sea records. Notable omissions from the present set are the large bodies of archaeomagnetic data from Europe and the U.S.S.R., and the high-quality lacustrine records from North America (see, for example, Lund & Banerjee 1979, 1981; King *et al.* 1982). A strong limitation on what can be achieved with currently available palaeomagnetic records is the poor quality of most of the associated chronologies.

I thank K. M. Creer, G. Ingham and P. Tucholka for providing listings of their data, F. J. Lowes and C. Green for helpful discussions, M. W. McElhinny for first suggesting this topic, and the Institute of Geological Sciences, Edinburgh, for providing listings of observatory data. The work was supported by the N.E.R.C., grant no. GR3/3552.

References (Barton)

Alldredge, L. R. 1977 *J. Geomagn. Geoelect., Kyoto* **29**, 123–135.
Barton, C. E. & McElhinny, M. W. 1981 *Geophys. Jl R. astr. Soc.* **67**, 465–485.
Barton, C. E. & Polach, H. A. 1980 *Radiocarbon* **22**, 728–739.
Bloomfield, P. 1976 *Fourier analysis of time series: an introduction* (258 pages.) New York: Wiley.
Braginskiy, S. I. 1970 *Geomagn. Aeron.* (*Engl. Transl.*) **10**, 172–181.
Braginskiy, S. I. & Fishman, V. M. 1977 *Geomagn. Aeron.* **17**, 607–613.
Bucha, V. 1969 *Nature, Lond.* **224**, 681–683.
Burlatskaya, S. P. & Nachisova, I. Ye. 1977 *Archaeomagnetic determinations of geomagnetic field elements*, *Mater. Mirov. Tsentra Dannykh B.* (112 pages.) Moscow: Sov. Geophys. Comm. Acad. Sci. U.S.S.R.
Chave, A. D. & Denham, C. R. 1979 *Earth planet. Sci. Lett.* **44**, 150–152.
Clark, R. M. 1975 *Antiquity* **49**, 251–266 (and comments by Suess and reply, *Antiquity* **50**, 61–63).
Courtillot. V., Le Mouël, J. L. & Mayaud, P. N. 1977 *J. geophys. Res.* **82**, 2641–2649.
Cox, A. 1968 *J. geophys. Res.* **73**, 3247–3260.
Creer, K. M. 1974 *Earth planet. Sci. Lett.* **23**, 34–42.
Creer, K. M., Hogg, E., Malkowski, Z., Mojski, J. E., Niedziolka-Krol, E., Readman, P. W. & Tucholka, P. 1979 *Geophys. Jl R. astr. Soc.* **59**, 287–313.
Currie, R. G. 1968 *J. geophys. Res.* **73**, 2779–2786.
Currie, R. G. 1973 *S. African J. Sci.* **69**, 379–383.
Doell, R. R. & Cox, A. 1972 In *The nature of the solid Earth* (ed. E. C. Robertson), pp. 245–284. McGraw-Hill.
Hirooka, K. 1971 *Mem. Fac. Sci. Kyoto Univ., Ser. of Geol. Miner.* **38**, 167–207.
Jacobs, J. A. 1975 *The Earth's core.* (258 pages.) London: Academic Press.
Johnson, E. A., Murphy, T. & Torreson, O. W. 1948 *Terr. Magn. atmos. Elect.* **53**, 349–372.
King, J., Banerjee, S. K., Marvin, J. & Lund, S. 1982 *Palaeogeog. Palaeoclim. Palaeoecol.* (Submitted.)
Lund, S. P. & Banerjee, S. K. 1979 *Rev. Geophys. Space Phys.* **17**, 244–249.
Lund, S. P. & Banerjee, S. K. 1981 In *4th IAGA Assembly, Edinburgh, Abstracts*, p. 175.

McElhinny, M. W. & Senanayake, W. 1981 *4th I.A.G.A. Assembly, Edinburgh, Abstracts*, p. 176.
Rampino, M. R. 1979 *Geology* **7**, 584–587.
Smith, P. J. 1970 In *Palaeogeophysics* (ed. S. K. Runcorn), pp. 79–80. London: Academic Press.
Tucholka, P. 1980 *Earth planet. Sci. Lett.* **48**, 379–384.
Turner, G. M. & Thompson, R. 1981 *Geophys. Jl R. astr. Soc.* **65**, 703–725.
Verosub, K. L. & Cox, A. 1971 *J. Geomagn. Geoelect., Kyoto* **23**, 235–242.
Vestine, E. H. & Kahle, A. B. 1968 *Geophys. Jl R. astr. Soc.* **15**, 29–37.
Wollin, G., Ryan, W. B. F. & Ericson, D. B. 1978 *Earth planet. Sci. Lett.* **41**, 395–397.
Yaskawa, K., Nakajima, T., Kawai, N., Torii, M., Natsuhara, N. & Horie, S. 1973 *J. Geomagn. Geoelect., Kyoto* **25**, 447–474.
Yukutake, T. 1973 *J. Geomagn. Geoelect., Kyoto* **25**, 195–212.

Discussion

F. J. Lowes (*School of Physics, The University, Newcastle upon Tyne, U.K.*). Dr Barton has suggested that the power spectrum drops at low frequency. However, if main-field reversals are modelled by a random rectangular waveform having amplitude A and average frequency of reversals $2f_0$, their power spectrum would be $(2A^2/f_0)\,\mathrm{d}f/\{1+(\pi f/f_0)^2\}$. For $A = 45000$ nT and $f_0 = 2\times 10^{-6}$ years this gives 2×10^{25} nT^2 year at low frequency.

C. E. Barton. My analysis does not extend into the periodicity range of geomagnetic reversals where the spectral power density sum $S_X(f)+S_Y(f)+S_Z(f)$ will of course be large. The question is whether there is a significant geomagnetic secular variation signal at periodicities beyond 10^5 years within constant polarity epochs. I am suggesting that the low-frequency ends of curves P43 and P45 in figure 2 provide an approximate upper bound to the spectral power density in this region.