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^{14}C DATING AND MAGNETOSTRATIGRAPHY

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ABSTRACT. The history of the earth's magnetic field is preserved in the fossil magnetism of archaeologic specimens, natural rocks and sediments. Samples such as lava flows and baked sherds that acquired a thermoremanent magnetization on cooling can be used to estimate ancient geomagnetic field intensities and directions. Paleofield directions can also be obtained from fine-grained sediments that acquired detrital magnetic remanence when deposited. Study of the earth's magnetic field over the last few tens of thousands of years yields information on geomagnetic dynamo theories, causes of fluctuations in cosmic-ray activity, and the formulation of a new regional chronologic tool.

INTRODUCTION

Both the intensity and shape of the geomagnetic field change with time. Over 1000 paleo-intensity determinations have been made on materials covering the last 10,000 years. These determinations show that the global mean dipole moment during this time was 9×10^{22} Am^2 and that it varied from a broad minimum of 6.5×10^{22} Am^2 ca 6500 years BP to a maximum of 11.5×10^{22} Am^2 near 2700 years BP (McElhinny and Senanayake, in press). Superimposed on this overall trend are more rapid intensity fluctuations, such as those observed in the last century. Since AD 1850 the axial dipole has been decreasing at an average rate of 5% per century, the rate dropping for a few years ca AD 1950 to almost zero. Variations of even higher frequency have recently been observed. In AD 1970 the earth's magnetic field underwent a jerk (Malin and Hodder, 1982) which noticeably altered the rate of secular variation at many localities of the world. Sustained intensity changes causing the local field to vary by a factor of 2 or 3 in 100 years were suggested on the basis of a sequence of archaeomagnetic results from Greece (Walton, 1979). The general pattern of field intensity changes summarized above largely confirms that discussed by Smith (1966) and Bucha (1970) and reinforces the possibility of important global intensity changes occurring in only a few hundred years.

Detailed information about the evolution of the shape of the geomagnetic field over the last 10,000 years recently became available through paleomagnetic studies of lake sediment cores (Mackereth, 1971) from different regions of the world. Two key advances made the acquisition of global paleolimnomagnetic data possible. The first advance was the development of piston corers, such as Mackereth's (1958) pneumatic 6m corer which allowed multiple, long, undisturbed sediment cores to be collected from a lake during one day's fieldwork. The second advance was the computerization of fluxgate and cryogenic magnetometers which permitted the magnetic remanence of hundreds of fragile, weakly magnetized samples to be measured during one day's laboratory work. Following these technical developments, much of the paleomagnetic data synthesized below were collected from lake deposits. In Europe, e g, well over 50,000 samples from tens of lakes have been subjected to paleomagnetic measurements.

DATA QUALITY

PALEOMAGNETIC DATA. As most natural materials do not hold a reliable paleomagnetic record of ancient geomagnetic field changes, it is very important to demonstrate the credibility of the data before accepting them for magnetostratigraphic or geophysical work. The standard approach for such an assessment is to use minimum reliability criteria. In this approach practical quantitative criteria are used in order to reject results that are unlikely to be reliable indicators of the paleomagnetic field. In paleolimnomagnetic studies the most important criteria concern 1) reproducibility between samples, cores and lakes, and 2) independence from sediment lithology. Unusual paleomagnetic results from only one sediment sampling site should be viewed with great caution.

^{14}C DATA. Most paleolimnomagnetic records have been dated directly by the ^{14}C method. A limited number of key magnetic records have also been dated using a pollen biostratigraphy or tephra chronology, tied to the ^{14}C time scale. Sediments dated by yearly laminations are now beginning to yield reliable paleomagnetic records.

Contamination of lake sediments by old carbon (Olsson, 1979) is a recurrent problem in dating lake sediments. Natural carbon contamination due to either erosion of materials such as graphite, coal, or old soils in the catchment, or to the redeposition of old sediments cannot, at present, be dealt with satisfactorily by either laboratory pretreatment or modelling procedures. Particularly pronounced natural contamination can lead to out-of-sequence dates. Such highly

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aberrant ages can of course be easily discarded. More commonly, however, natural contamination leads to subtle systematic errors that can be difficult to detect unless the ^{14}C dates from lake sediments can be compared with dates from materials such as peat and charcoal.

A second problem concerns hard-water corrections. In regions of carbonate-rich bedrock or glacial deposits, ages can again be systematically old. The usual method of correcting hard-water errors is to extrapolate the ^{14}C age/depth relationship to the mud/water interface in order to estimate the present day hard-water error. This error is then assumed constant and all the ^{14}C ages are then reduced by the amount of the present hard-water error. Constant hard-water correction errors of several 100 years have had to be applied to the South Australian, North American, and Near Eastern master sediments. The need for duplicate dating techniques such as tephra or pollen zonation combined with ^{14}C dates is apparent. In this paper all ^{14}C dates were calibrated to calendar dates using Clark's (1975) tree-ring curve. Dates older than 6500 years (the limit of Clark's curve) were modified using an extension to Clark's calibration curve that decreases smoothly to zero correction at 10,000 years BP.

REGIONAL GEOMAGNETIC MASTER CURVES

Regional master curves (fig 1) were derived by analyzing high quality paleomagnetic results from lake sediments. The curves plot the variation of declination and inclination with calibrated ^{14}C age. All the directional records were confirmed by stacking results from more than one lake or from multiple samples. Stacking, however, was not used in constructing the master curves because of signal attenuation. Each curve consists of a cubic spline fitted to the sequential directional data of a type section. The type lakes and cores selected for the six regions are: South Australia, Bullenmerri core BC extended by Keilambete core KF (Barton and McElhinny, in press); North America, St Croix core 75 (Banerjee, Lund and Levi, 1979); western Europe, Lomond core LLRPI extended by Windermere core W3 (Thompson and Turner, 1979); eastern Europe, Paajarvi core P4 (Huttonen and Stober, 1980) extended by Lovojarvi core D (Tolonen, Siiriainen and Thompson, 1975); Near East, Kinneret core K8 (Thompson and Stiller, ms in preparation); East Asia, Kizaki core K3 (Horie et al, 1981). The North Pacific curve of fig 1 is based on the paleomagnetic results from ^{14}C dated lava flows of McWilliams, Holcomb and Champion (1982).

SECULAR VARIATION MAGNETOSTRATIGRAPHY

The regional master curves form a basis for dating Holo-

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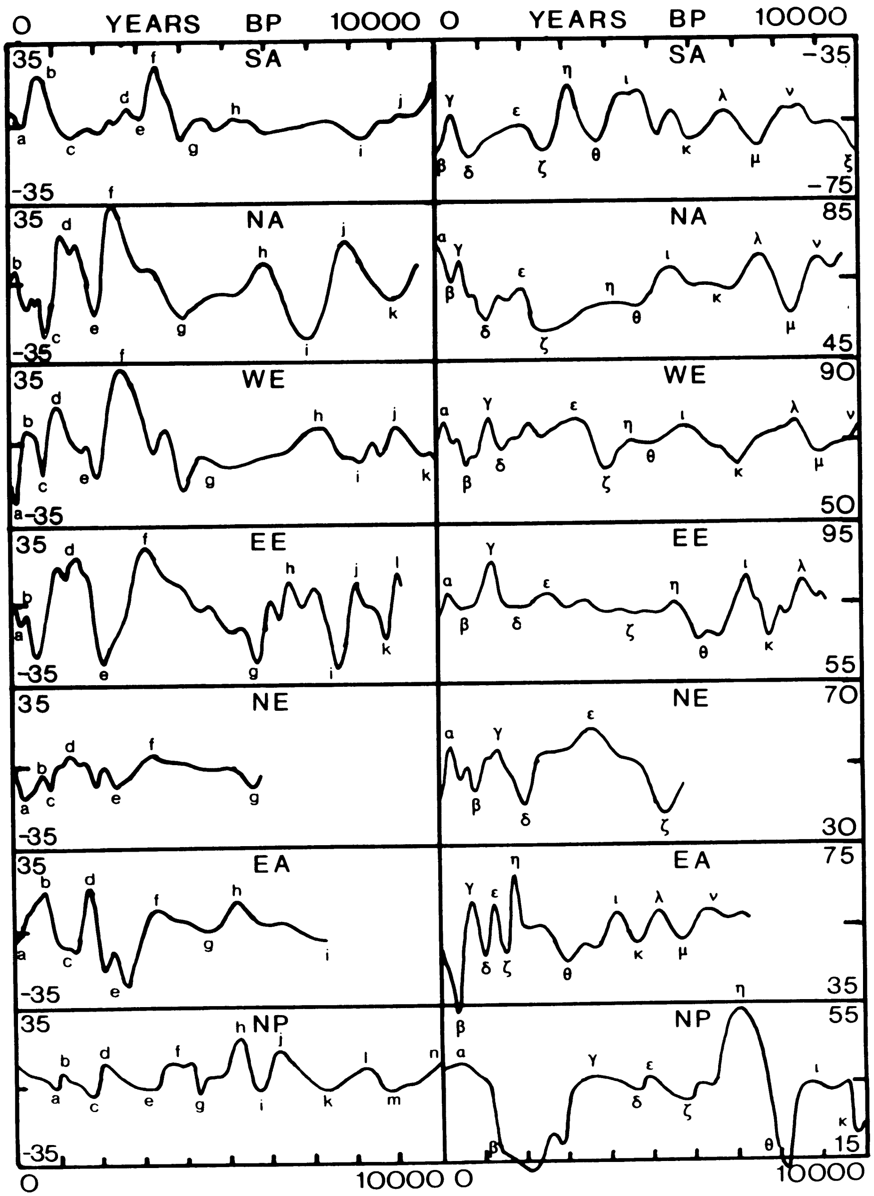

Fig 1. Regional declination (left) and inclination (right) master curves for South Australia, SA; North America, NA; West Europe, WE; East Europe, EE; Near East, NE; East Asia, EA; North Pacific, NP. Tree ring calibrated time scale in calendar years BP (0 BP = AD 1950).

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cene sediments. By matching paleomagnetic features from new sedimentary sequences with those of the curves, dates can be transferred to the new materials under investigation. Turning points of the direction variations prove to be the clearest features to recognize; thus, they are labelled (Fig 1). Some turning points are poorly developed; others, eg, WEe and WEg, have double peaks. Both of these are difficult to label. Bearing in mind difficulties discussed above, the typical uncertainty in the estimated ages of the secular variation features (Table 1) is probably ca 5%. Dominant errors relate to natural carbon contamination problems and not to paleomagnetic or isotopic laboratory experimental precision limitations. Geographic areas over which the regional magnetostratigraphy master curves will apply can best be judged from a study of the historic magnetic field.

Table 2 summarizes global Holocene geomagnetic field changes. The main geomagnetic variation is the fluctuation of the axial dipole intensity (g_1^o). The magnitude and timing of these variations can be used to account for a large proportion of the changes in ^{14}C production through a shielding modulation of the flux of galactic cosmic rays. Spherical harmonic analyses of the paleomagnetic data plotted in Figure 1 reveal slow changes in the axis of the geomagnetic dipole about the earth's spin axis (Table 2). The analyses indicate that the dipole axis was markedly tilted away from the rotation axis around 7500 and 2500 years ago as well as during the last few hundred years. An estimate of the relative importance of the non-dipole field has been made through the residual errors (RSS) to the best fitting dipole (Table 2). The non-dipole field appears to have been relatively strong around 7500 years BP and relatively weak around 5000 years BP.

THE HISTORICAL FIELD

Spherical harmonic analyses of historic and archaeomagnetic data provide a consistent set of harmonic coefficients at 50-year intervals since AD 1600 (Thompson and Barraclough, in press). By taking differences between these coefficients, global maps of magnetic field changes can be constructed. Two maps are shown in Figure 2A and B for the North Atlantic region. These illustrate the change in position of the westerly declination maxima WEa (Fig. 1) and the inclination maxima WEα (Fig 1) between AD 1625 and 1975. The maps were prepared by calculating and contouring the zero isopores of the rate of change of declination (D) and inclination (I). The zero isopores, in common with many features of the non-dipole field, exhibit a general westward drift. Their drift rate varied markedly from place to place, eg, compare the AD 1675 and 1875

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Table 1. Ages of magnetostratigraphic features

Declination

Turning points	SA	NA	WE	EE	NE	EA	NP
a	300	–	140	160	220	0	900
b	680	100	450	300	700	700	1100
c	1300	750	600	600	850	1200	1800
d	2000	1200	1000	1400	1300	1650	2150
e	2800	2000	2000	2200	1900	2200	3200
f	3500	2400	2600	3100	2100	3100	3900
g	4500	4000	4900	5700	2400	4400	4400
h	5500	5900	7100	6500	3200	5100	5300
i	8300	7000	8300	7600	5600	7300	5600
j	9000	7900	9100	8000	–	–	6000
k	–	9000	10000	8700	–	–	7300
l	–	–	–	9000	–	–	8350
m	–	–	–	–	–	–	8900

Inclination

Turning points	SA	NA	WE	EE	NE	EA	NP
α	–	50	240	300	300	–	200
β	0	420	650	600	550	400	2150
γ	400	750	1150	1300	700	760	3500
δ	900	1200	1650	1900	900	1000	4700
ε	1900	2300	3100	2600	1400	1300	5100
ζ	2600	2900	3800	4600	2000	1550	5800
η	3200	3700	4300	5500	3600	1750	7000
θ	3600	4400	5000	6400	5300	2800	8200
ι	4600	5300	6000	7200	–	4100	8950
κ	6000	6600	7100	7800	–	4600	9800
λ	6800	7700	8300	8600	–	5100	–
μ	7900	8400	8800	–	–	5600	–
ν	8600	9600	9700	–	–	6600	–
ξ	10000	–	–	–	–	–	–

SA South Australia (35°S 140°E)
NA North America (45°N 90°W)
WE Western Europe (55°N 05°W)
EE Eastern Europe (60°N 30°E)
NE Near East (30°N 35°E)
EA Eastern Asia (35°N 140°E)
NP North Pacific (20°N 155°W)

Ages tabulated in calibrated ^{14}C years BP. The pre-2000 BP EA magnetostratigraphic features are taken from Horie *et al* (1981). The EA ages are rather poorly known, based here on a linear interpolation between the basal tephra layer and the archaeomagnetic features in the upper sediments. The NP turning points are based on the ^{14}C dated lava flow palaeomagnetic data of McWilliams *et al* (1982).

Table 2 Geomagnetic coefficients and parameters

Epoch BP	g_1^o μT	λ_c	ϕ_c	g_2^o/g_1^o	RSS	Regions
Present	-30.5	79°N	070°W	+0.05	-	>100
0	-30.5	81	087 W	+0.02	0.011	7
500	-37.6	83	086 W	-0.02	0.009	7
1000	-40.9	89	041 W	+0.04	0.014	7
1500	-41.8	85	163 E	+0.03	0.008	7
2000	-42.0	87	165 W	+0.07	0.018	7
2500	-44.0	80	155 E	+0.04	0.019	7
3000	-40.2	85	121 E	+0.02	0.011	7
3500	-33.9	89	110 W	-0.04	0.023	7
4000	-32.5	88	001 E	+0.00	0.015	7
4500	-32.6	88	008 E	-0.01	0.004	7
5000	-30.0	86	020 W	-0.02	0.003	7
5500	-26.9	84	080 W	-0.02	0.004	7
6000	-25.4	86	090 W	-0.01	0.008	6
6500	-25.2	86	015 W	+0.02	0.005	6
7000	-25.5	82	004 W	-0.06	0.019	6
7500	-27.1	80	054 W	+0.13	0.036	5
8000	-30.2	87	155 W	-0.01	0.004	5
8500	-32.8	88	113 E	-0.09	0.011	5
9000	-33.0	88	023 W	-0.04	0.011	5

g_1^o is the strength of the axial dipole. This is the dominant term in the spherical harmonic expansion of the earth's internal magnetic field and is the main parameter to be considered in assessing the influence of geomagnetic field variations on ^{14}C production. λ_c is the latitude of the north magnetic pole. ϕ_c is the longitude of the north magnetic pole. g_2^o/g_1^o is the ratio of the axial quadrupole to the axial dipole and is an important geomagnetic parameter in assessing asymmetry between the northern and southern hemispheres. RSS is the mean residual sum of squares, here presented as a proportion of the axial dipole (i e, with $g_1^o = 1.0$), to the palaeolimnomagnetic direction data when fitted by a 'tilted' geocentric dipole. RSS is a measure of the non-dipole field. Large values indicate times of high non-dipole to dipole field ratio. The right hand column notes the number of regions used in the calculations. The data used are shown in Fig 1. The effect of data distribution can be judged by comparing the 0 BP parameters with those of the present field. The small number of regions available for analysis and the lack of absolute orientation of the palaeomagnetic records can lead to systematic errors in spherical harmonic analyses.

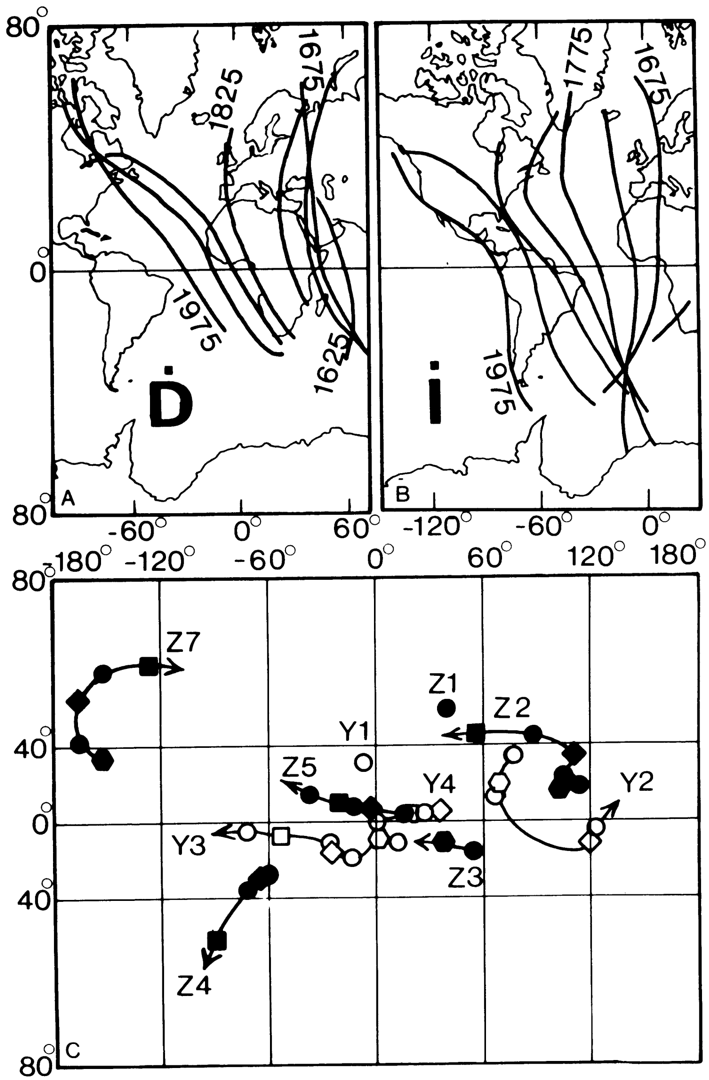

Fig 2. Examples of geomagnetic changes between AD 1625 and 1975. A. Location of the isopore D = 0. B. Location of the isopore I = 0. C. Major foci of changes in Y (open symbols) and Z (closed symbols) at 50-year intervals. Positive anomalies have even codes (eg, Y2), negative anomalies have odd codes. Hexagons, diamonds, and squares correspond to epochs AD 1725, 1825 and 1925, respectively.

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positions of the zero D isopore at the equator and at the latitude of Britain in Figure 2A. These differences and the overall changes can be understood in terms of the growth, movement, and decay of local centers of secular change. Figure 2C shows the location of isoporic foci of the field elements Y and Z calculated from Thompson and Barraclough's coefficients. Changes in the easterly horizontal element (Y) and the vertical element (Z) are closely related to changes in declination and inclination, respectively. The 17th century pattern of North Atlantic declination changes was dominated by the foci Y1 and Y2. The change in location of the westerly declination turning point WEa (ie zero D isopore) has been controlled partly by the establishment of the new foci Y3 and Y4 and partly by the movement of all four foci. The change in position of the zero I isopore (WEα) was controlled by the evolution and movement of the minimum focus Z5 gradually replacing the foci Z1 and Z3, (Fig 2C).

SUMMARY

Holocene field intensities have varied by a factor of 2 or 3 due to fluctuations in the geomagnetic dipole moment, whereas field direction changes have been dominated by the turbulent evolution of the non-dipole field with short-lived foci of secular change. The foci have combined, largely at random, to produce the field variations of Figure 1. Analysis of historic field variations reveals regions of similar field changes several thousand kilometers in extent. The boundaries between such regions undoubtedly alter along with the field changes; nevertheless, the regional master curves (Fig 1) are likely to be useful over distances of several hundred kilometers and possibly a few thousand kilometers. The age differences of turning points, due to source drift, across such distances is only ca 100 years. When dating older Holocene sediments, such differences are not of critical importance, and for younger historic materials, the turning point chronology has now been mapped (Fig 2A, B).

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