1
The Earth's magnetic field

1.1 Introduction

In geomagnetism we are measuring extremely small magnetic fields – at its strongest near the poles, the Earth's magnetic field is several hundred times weaker than that between the poles of a toy horseshoe magnet. In a magnetic compass, the needle is weighted so that it will swing in a horizontal plane, its deviation from geographical north being called the declination, D. A non-magnetic needle which is balanced horizontally on a pivot becomes inclined to the vertical when magnetized. Over most of the northern hemisphere the north-seeking end of the needle will dip downwards, the angle it makes with the horizontal being called the magnetic dip or inclination, I. The total intensity F, the declination D and the inclination I completely define the magnetic field at any point. The horizontal and vertical components of F are denoted by H and Z. H may be further resolved into two components X and Y, X being the component along the geographical meridian (northward) and Y the orthogonal component (eastward). Figure 1.1 illustrates these different magnetic elements. They are simply related to one another by the following equations:

$$H = F \cos I \qquad Z = F \sin I \qquad \tan I = Z/H \tag{1.1}$$

$$X = H \cos D \qquad Y = H \sin D \qquad \tan D = Y/X \tag{1.2}$$

$$F^2 = H^2 + Z^2 = X^2 + Y^2 + Z^2 \tag{1.3}$$

The variation of the magnetic field over the Earth's surface is best illustrated by isomagnetic charts, i.e. maps on which lines are drawn through points at which a given magnetic element has the same value. Contours of equal intensity in any of the elements X, Y, Z, H or F are called isodynamics. Figures 1.2–1.4 are world maps showing contours of equal declination (isogonics), equal inclination (isoclinics) and total intensity for the year 1990. Palaeomagnetists have traditionally used the oersted as the unit of magnetic field strength and the gauss (Γ) as the unit of magnetic induction. The distinction is somewhat pedantic in geophysical applications, since the permeability of air is virtually unity in cgs units. In SI units, which will be used throughout this book,

$$1\ \Gamma = 10^{-4}\ \text{Wb m}^{-2}\ (\text{weber/m}^2) = 10^{-4}\ \text{T (tesla)}$$

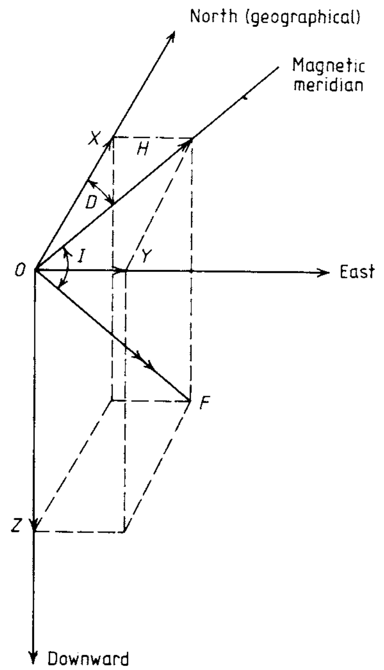

Figure 1.1

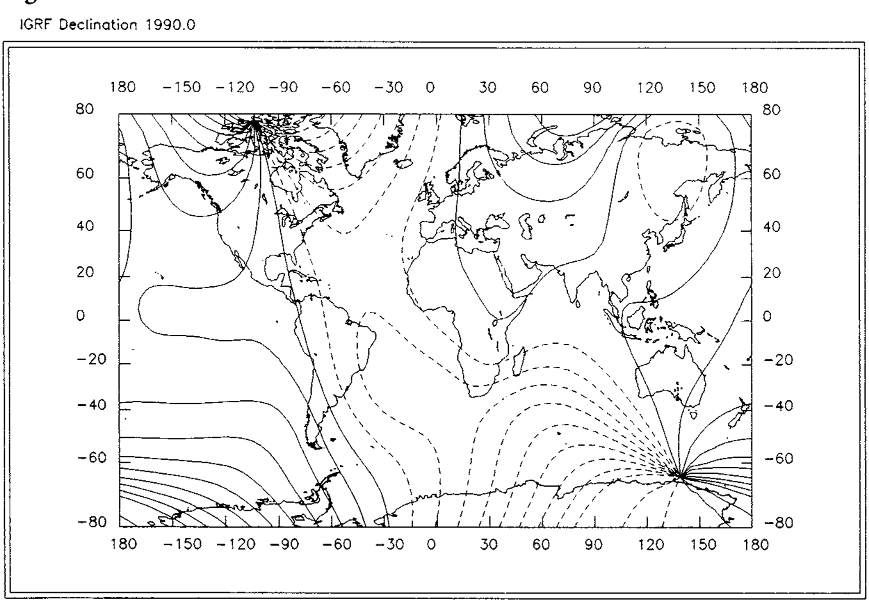

Figure 1.2 World map showing contours of equal declination D for 1990. The zero and positive contours are shown as solid lines and the negative contours as dashed lines. Contours are shown for the following values: $-150°$, $-120°$, $-90°$ to $+90°$ with a contour interval of $10°$, $+120°$ and $+150°$. Provided by D. R. Barraclough. Courtesy of the Geomagnetism Unit, British Geological Survey.

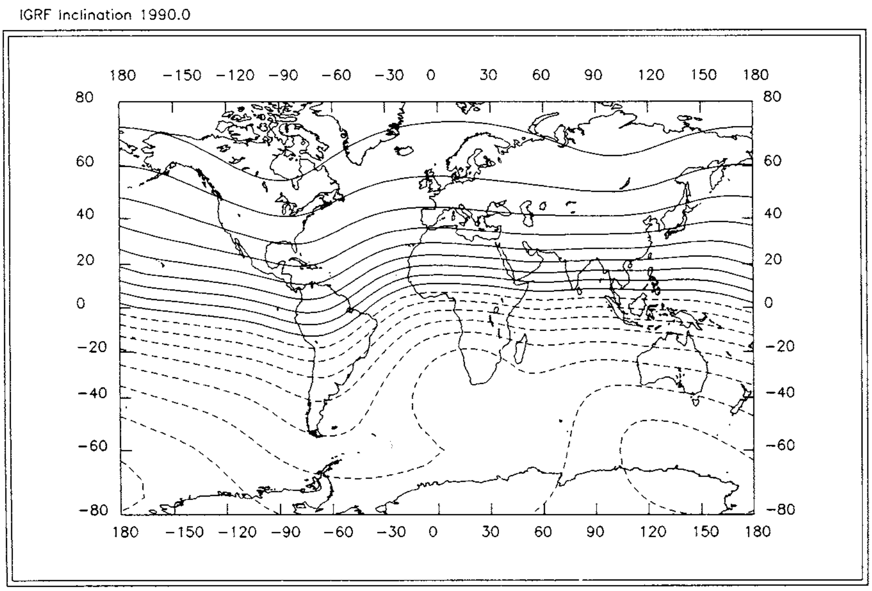

Figure 1.3 World map showing contours of equal inclination *I* for 1990. The zero and positive contours are shown as solid lines and the negative contours as dashed lines. The contours are from −80° to +80° with a contour interval of 10°. Provided by D. R. Barraclough. Courtesy of the Geomagnetism Unit, British Geological Survey.

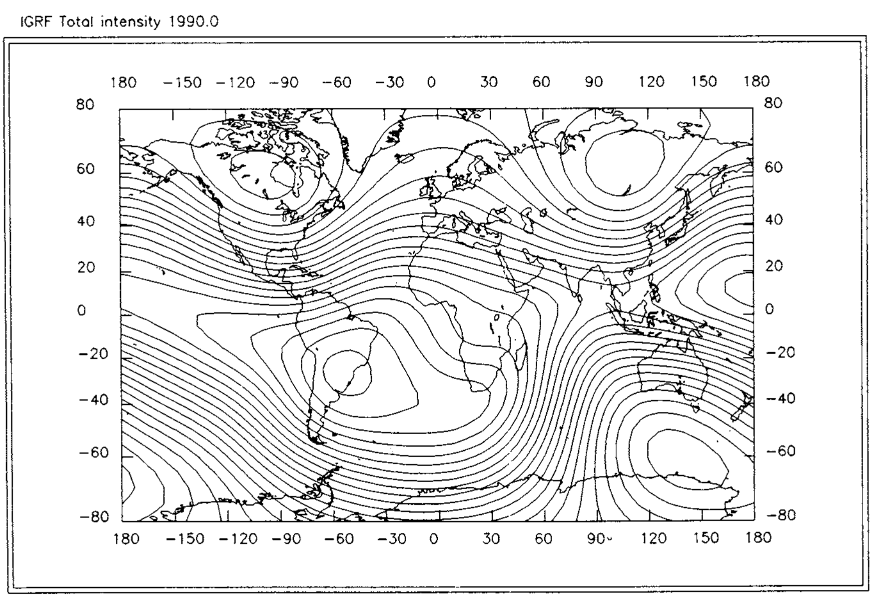

Figure 1.4 World map showing contours of equal total intensity *F* for 1990. The contour interval for *F* is 200 nT and the contours range from 24 000 nT, centred over southern Brazil, to 66 000 nT, centred approximately on the southern dip-pole south of Australia. Provided by D. R. Barraclough. Courtesy of the Geomagnetism Unit, British Geological Survey.

Since in geomagnetism we are measuring extremely small magnetic fields, a more convenient unit is the gamma (γ), defined as

$$1\ \gamma = 10^{-9}\ \mathrm{T} = 1\ \mathrm{nT}$$

Apart from its spatial variation, the Earth's magnetic field also shows temporal changes ranging from variations on a timescale of seconds to secular variations on a timescale of hundreds of years and on an even longer timescale to complete reversals of polarity. The short-period, transient, variations are due to external influences and have no lasting effect on the Earth's main magnetic field, which is of internal origin. They will not be discussed at all in this book. Variations over $10{-}10^4$ a may be determined from archaeomagnetic and palaeomagnetic studies of the secular variation. This time range is probably characteristic of core fluid motions. If successive annual mean values of a magnetic element are obtained from a particular station, it is found that these secular changes are in the same sense over a long period of time, although the rate of change does not usually remain constant. Figure 1.5 shows the changes in declination and inclination at London, Boston and Baltimore. The declination at London was 11½° E in 1580 and 24½° W in 1819, a change of almost 36° in 240 years. Lines of equal secular change (isopors) in an element form sets of ovals centring on points of local maximum change (isoporic foci). Figures 1.6 and 1.7 show the secular change in Z for the years 1922.5 and 1942.5. It can be seen that the secular variation is a regional rather than a planetary phenomenon and that considerable changes can take place in the general distribution of isopors even within 20 years. The secular variation is anomalously large and complicated over and around Antarctica; on the other hand, it is markedly smaller in the Pacific hemisphere (between about 120° E and 80° W). There is a strong secular change focus in the Atlantic, where the vertical intensity is changing non-linearly. Z was approximately -50 nT a^{-1} in the 1960s and -150 nT in 1978. The isophoric foci drift westward at a fraction of a degree per year.

One of the most interesting results of palaeomagnetic studies is that many igneous rocks show a permanent magnetization approximately opposite in direction to that of the present field. Reverse magnetization was first discovered by David (1904) and by Brunhes (1906) in a lava from the Massif Central mountain range in France – since then examples have been found in almost every part of the world. About one-half of all rocks measured are found to be normally magnetized and one-half reversely. Dagley *et al.* (1967) carried out an extensive palaeomagnetic survey of Eastern Iceland sampling some 900 separate lava flows lying on top of each other. The direction of magnetization of more than 2000 samples representative of individual lava flows was determined covering a time interval of 20 Ma. At least 61 polarity zones, or 60 complete changes of polarity, were found giving an average rate of at least 3 inversions/Ma. The same pattern of reversals observed in igneous rocks was also found in deep-sea sediments (see, e.g., Opdyke *et al.*, 1966).

In 1839 Gauss showed that the field of a uniformly magnetized sphere, which is the same as that of a dipole at its centre, is an excellent first approximation to the Earth's magnetic field. Gauss further analysed the irregular part of the Earth's

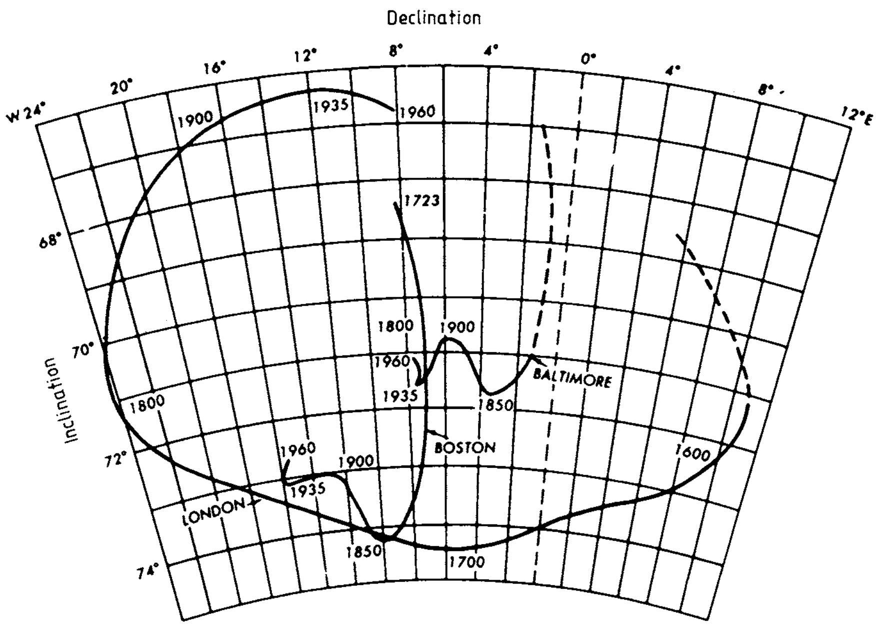

Figure 1.5 Secular change of declination and inclination at London, Boston and Baltimore. After Nelson *et al.* (1962).

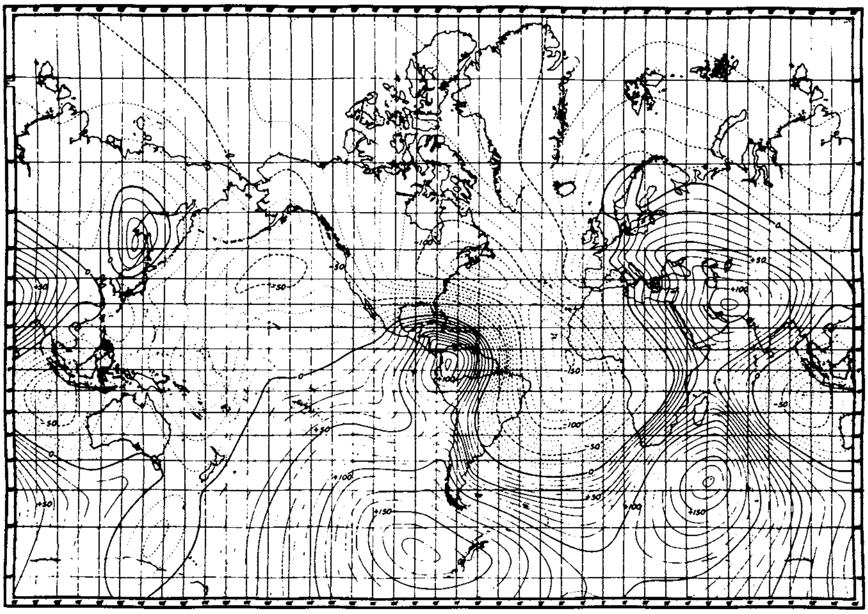

Figure 1.6 World map showing the geomagnetic secular variation of the vertical component *Z*. Epoch 1922.5. After Vestine *et al.* (1947a).

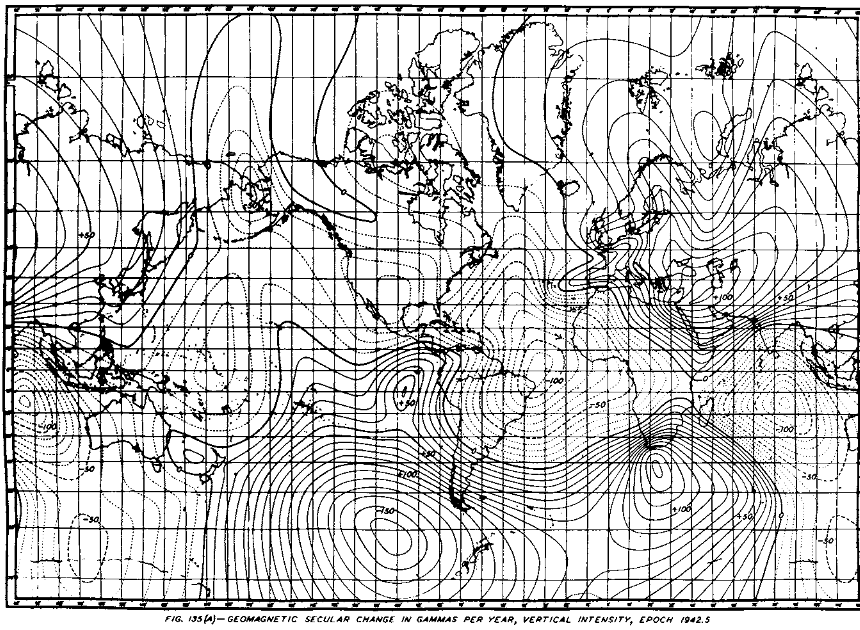

Figure 1.7 World map showing the geomagnetic secular variation of the vertical component *Z*. Epoch 1942.5. After Vestine *et al.* (1947a).

field, i.e. the difference between the actual observed field and that due to a uniformly magnetized sphere, and showed that both the regular and irregular components of the Earth's field are of internal origin.

Since the north-seeking end of a compass needle is attracted towards the northern regions of the Earth, those regions must have opposite polarity. Consider therefore the field of a uniformly magnetized sphere whose magnetic axis runs north–south, and let P be any external point distant r from the centre O and θ the angle NOP, i.e. θ is the magnetic co-latitude (see Figure 1.8). If m is the magnetic moment of a geocentric dipole directed along the axis, the potential at P is

$$V = \frac{m}{4\pi}\frac{\cos\theta}{r^2} \tag{1.4}$$

The inward radial component of force corresponding to the magnetic component Z is given by

$$Z = \mu_0 \frac{\partial V}{\partial r} = \frac{-\mu_0 m}{2\pi}\frac{\cos\theta}{r^3} \tag{1.5}$$

and the component at right angles to OP in the direction of decreasing θ, corresponding to the magnetic component H, by

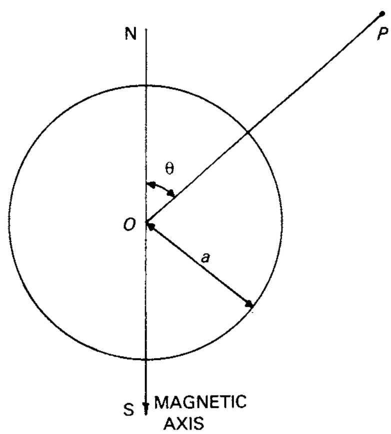

Figure 1.8

$$H = \mu_0 \frac{1}{r}\frac{\partial V}{\partial \theta} = \frac{-\mu_0 m}{4\pi}\frac{\sin\theta}{r^3} \tag{1.6}$$

where μ_0 is the permeability of free space.

The inclination I is then given by

$$\tan I = Z/H = 2\cot\theta \tag{1.7}$$

and the magnitude of the total force F by

$$F = (H^2 + Z^2)^{1/2} = \frac{\mu_0 m}{4\pi r^3}(1 + 3\cos^2\theta)^{1/2} \tag{1.8}$$

Thus intensity measurements are a function of latitude.

The geomagnetic poles, i.e. the points where the axis of the geocentric dipole which best approximates the Earth's field meets the surface of the Earth, are situated at approximately 79° N, 70° W and 79° S, 110° E. The geomagnetic axis is thus inclined at about 11° to the Earth's geographical axis.

Studies of the secular variation in historic times (HSV) show that the dipole intensity has continued to drop over the last 150 a, the rate of decrease increasing since about 1965. This appears to be accompanied by an increase in the non-dipole field near the core–mantle boundary (CMB), implying that there is no observable energy-density loss in the core field. The possible significance of this will be discussed later (Section 5.8) when we consider reversal models. The westward drift of the non-dipole field also appears to have been decreasing over the last 100 a. Newitt and Dawson (1984) find no evidence for it over North America for the last 100–200 a and Kalinin and Rozanova (1983) find significant regional differences in the westward drift rate.

Data on palaeomagnetic secular variation (PSV) can be obtained from archaeological kilns and fireplaces, from lava flows and from sequences of lake sediments. The longest records are from sedimentary sequences, mainly lacustrine, but some from marine cores – see Figure 1.9, which shows type curves of the secular variation obtained from lake sediments from east-central North America. This type of record can in principle provide continuous detailed data, but suffers from low resolution due to smoothing of the signal. Hanna and Verosub (1989) have given a review of lacustrine palaeomagnetic records from western North America covering the past 40 ka and Creer and Tucholka (1983) an overview of type PSV curves suitable for magnetostratigraphic correlation and dating. Particularly interesting is the PSV record from Mono Lake, California (Lund *et al.*, 1988), which shows a distinctive periodic vector waveform that follows an excursion of the magnetic field – this is discussed in detail in Section 4.4. Levi and Karlin (1989) obtained a 60 ka record from sediments in the Gulf of California which show recurring fluctuations about the geocentric axial dipole value. They suggest that these fluctuations may be related to a generally reduced dipole moment between about 20 and 5 ka which may be connected with possible geomagnetic excursions at 51–49 ka and 29–26 ka (see Sections 4.2–4.4).

Comparison of PSV records from Europe and North America show that PSV waveforms with periods longer than a few hundred years can be correlated over ~ 3000 km. Sproul and Banerjee (1989) noted that the Elk Lake Minnesota PSV record shows striking similarity to European archaeomagnetic records (Kovacheva, 1980) if the Elk Lake record is offset by 520 a corresponding to a westward drift rate of 0.23°/a. Distinctive PSV waveforms which reappear about every 2400–3000 a have been seen at some sites (e.g. Lake St Croix, Mono Lake). Lund and Banerjee (1975) have suggested that this is evidence of westward (or eastward) drift of a complex non-dipole waveform which changes very slowly with time compared with the time it takes for the waveform to drift entirely round the Earth (2400–3000 a).

That the secular variation of the present-day geomagnetic field is lower in the Pacific Ocean region than over the rest of the world has been known for some time (see, e.g., Vestine and Kahle, 1966; Doell and Cox, 1971; Bingham and Stone, 1972). Early palaeomagnetic studies of Brunhes age Hawaiian lava flows ($\lesssim$ 0.7 Ma) suggested that a low non-dipole field in the Pacific region was not just a recent feature (Doell and Cox, 1971; Doell, 1972). Studies of other areas of the Pacific (the Galapagos Islands – Doell and Cox, 1972; Easter Island – Isaacson and Heinrichs, 1976; and the Society Islands – Duncan, 1975) revealed no anomalously low secular variation thus constraining the region of a low non-dipole field. Because the eruption of lava flows is episodic and dating is not very precise, some authors have suggested that the low non-dipole field in Hawaii is either a very recent phenomenon or the result of incomplete sampling (McElhinny and Merrill, 1975; Coe *et al.*, 1978). More recently McWilliams *et al.* (1982), using ^{14}C-dated lava flows in Hawaii, concluded that lower secular variation is a real feature of the Central Pacific, although their data come mainly from younger samples (the oldest is 31 ka). Their conclusion is reinforced by Peng and King (1992), who obtained a palaeomagnetic record for the past 13 ka from sediments in Lake Waiau near the summit of Mauna Kea, Hawaii. Their data are similar to those obtained from lava flows (Holcomb *et al.*, 1986), but are of much higher resolution. It seems that anomalously low secular variation has occurred beneath the central Pacific area on timescales of 10^4 a.

The determination of the palaeointensity of the geomagnetic field is more

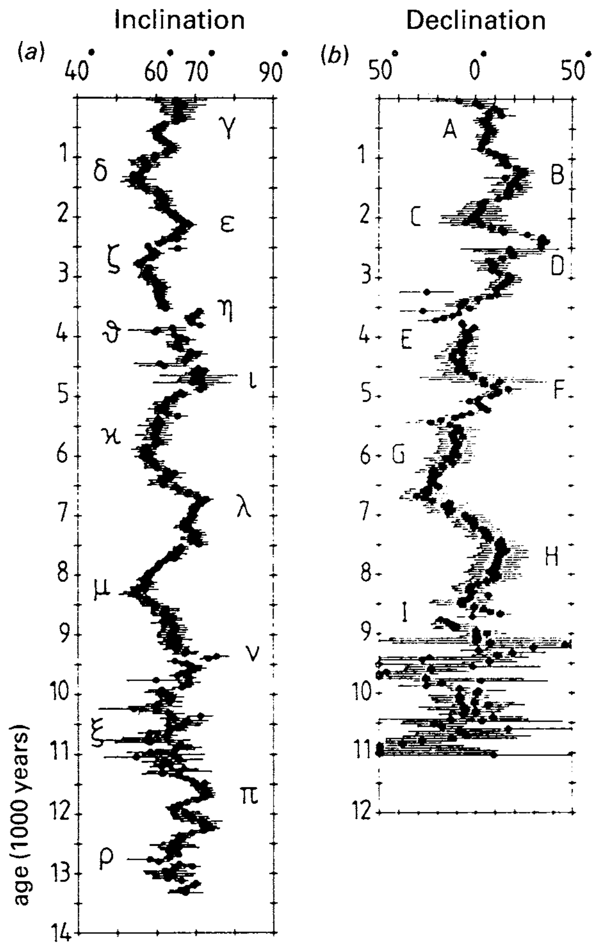

Figure 1.9 Type curves of the secular variation showing (a) inclination and (b) declination for east-central North America (Great Lakes – Minnesota region) obtained by stacking individual records at 40-year increments. The bars represent standard errors stacked at each level. The Greek and English labels identify the principal maxima and minima which were used for between lake correlation. After Creer and Tucholka (1982).

difficult to establish than its palaeodirection, since the intensity of the magnetic remanence is not only dependent on the intensity of the field but is also strongly related to the material. Chauvin *et al.* (1991) have obtained palaeointensity measurements using the Thellier method for two late Quaternary volcanic sequences on Réunion Island in the southern hemisphere (21° S). Values of the strength of the palaeofield varied from 19 to 55 μT. For significantly older rocks (0.6–2 Ma) from Réunion Island, Senanayake *et al.* (1982) had obtained field strengths in the range 19.4–40.2 μT, which is not substantially different from the late Quaternary values. Chauvin *et al.* found that for the youngest sequence

(5–12 ka) VDMs were mostly in the range 7.5×10^{22} to 9.9×10^{22} A m^2, but for the oldest sequence (82–98 ka) the range was from 4.1×10^{22} to 8.8×10^{22} A m^2.

Tric *et al.* (1992) obtained high-resolution records of the palaeointensity of the geomagnetic field for the past 80 ka from five marine cores – three from the Tyrrhenian Sea, one from the eastern Mediterranean and one from the southern Indian Ocean. The results correlate well with archaeomagnetic data before 10 ka and with volcanic data for the last 40 ka (see Figure 1.10). The dipole field moment shows large-scale changes – it fell to 22 and 28 % of its present value 39 and 60 ka ago, respectively. These lows alternated with periods of higher intensity. There is some indication of a periodic nature in these intensity variations, but the record is not long enough to allow of any conclusion being reached.

Meynadier *et al.* (1992) obtained relative magnetic field intensities for the last 140 ka from three marine cores in the Somali basin, Western Indian Ocean (a time versus depth correlation was established from the δ^{18}O record). The remanence intensity was normalized with respect to the anhysteretic remanent magnetization. The quasi-cyclic pattern for the past 80 ka confirms the results obtained in the Mediterranean by Tric *et al.* (1992) – see Figure 1.11. A power spectrum analysis indicated two dominant peaks centred at 100 ka and 22–25 ka and two smaller peaks at 19 and 43 ka. The longer period (100 ka) reflects the large parabolic aspect of the curve and cannot be considered for the limited time interval studied – a longer time series is needed to come to any conclusions on

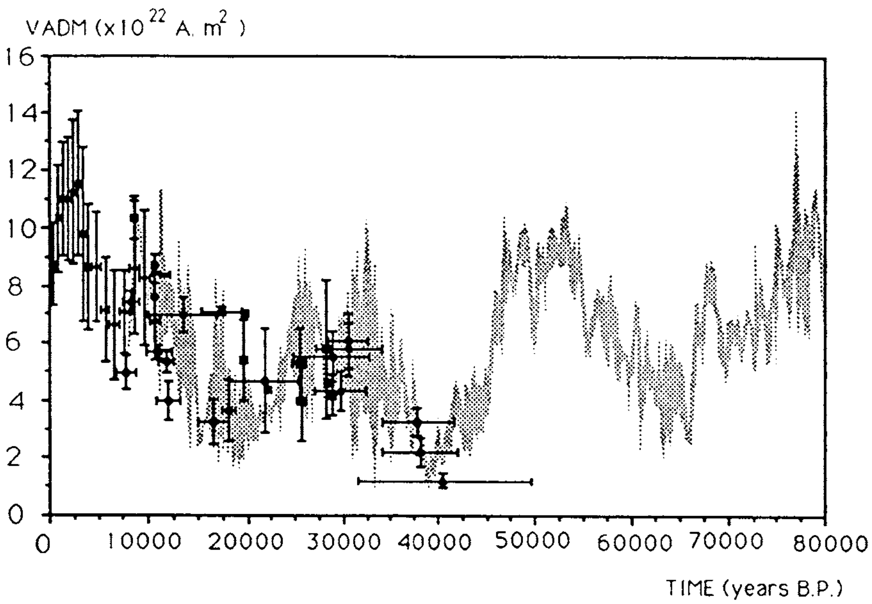

Figure 1.10 Comparison of the intensity of the geomagnetic field between the sedimentary records and archaeomagnetic and volcanic data for the period 0–40 ka. Data from the sedimentary records are shown hachured and extend to 80 ka. After Tric *et al.* (1992).