in terms of domain theory. After reaching saturation, the applied field H is reduced to zero at which point the intensity of magnetisation is that attributed to the remanent magnetisation. To eliminate this magnetisation, a negative field, $-H_c$, the *coercive force*, has to be applied. The *coercivity*, H_c, is an indication as to the 'hardness' or permanence of the magnetisation. Consequently, larger magnetic grains, which thus contain more magnetic domains, are easier to magnetise (and therefore have a higher susceptibility), than fine grains which are magnetically hard as indicated by a relatively high coercivity and low susceptibility. On increasing the applied magnetic field to full saturation, the hysteresis loop is completed. It follows that, for minerals that exhibit a nonlinear behaviour, no unique value of susceptibility exists. Values cited for such materials are usually for weak values of H and prior to saturation ever having been reached. For much more detailed discussion of rock magnetism, see the monographs by Nagata (1961), Stacey and Banerjee (1973) and O'Reilly (1984).

3.3 MAGNETIC PROPERTIES OF ROCKS

3.3.1 Susceptibility of rocks and minerals

Magnetic susceptibility is an extremely important property of rocks and is to magnetic exploration methods what density is to gravity surveys. Rocks that have a significant concentration of ferro- and/or ferri-magnetic minerals tend to have the highest susceptibilities. Consequently, basic and ultrabasic rocks have the highest susceptibilities, acid igneous and metamorphic rocks have intermediate to low values, and sedimentary rocks have very small susceptibilities in general (Table 3.2 and Figure 3.7). In this compilation of data, specific details of rock types are not available and so the values cited should be taken only as a guide. Metamorphic rocks are dependent upon their parent material and metapsammites are likely to have different susceptibilities compared with metapelites, for example.

Whole rock susceptibilities can vary considerably owing to a number of factors in addition to mineralogical composition. Susceptibilities depend upon the alignment and shape of the magnetic grains dispersed throughout the rock. If there is a marked orientation of particles, such as in some sedimentary and metamorphic rocks, a strong physical anisotropy may exist. The variation of magnetic properties as a function of orientation and shape of mineral grains is known as the *magnetic fabric*. Magnetic fabric analysis provides a very sensitive indication as to the physical composition of a rock or sediment, which in turn can be important in interpreting physical processes affecting that rock. For example, it is possible to correlate magnetic fabric variation in estuarine sediments with sonograph

Table 3.2 Susceptibilities of rocks and minerals (rationalised SI units)

Mineral or rock type	Susceptibility*
Sedimentary	
Dolomite (pure)	-12.5 to $+44$
Dolomite (impure)	20 000
Limestone	10 to 25 000
Sandstone	0 to 21 000
Shales	60 to 18 600
Average for various	**0 to 360**
Metamorphic	
Schist	315 to 3000
Slate	0 to 38 000
Gneiss	125 to 25 000
Serpentenite	3100 to 75 000
Average for various	**0 to 73 000**
Igneous	
Granite	10 to 65
Granite (m)	20 to 50 000
Rhyolite	250 to 37 700
Pegmatite	3000 to 75 000
Gabbro	800 to 76 000
Basalts	500 to 182 000
Oceanic basalts	300 to 36 000
Peridotite	95 500 to 196 000
Average for acid igneous	**40 to 82 000**
Average for basic igneous	**550 to 122 000**
Minerals	
Ice (d)	-9
Rocksalt (d)	-10
Gypsum (d)	-13
Quartz (d)	-15
Graphite (d)	-80 to -200
Chalcopyrite	400
Pyrite (o)	50 to 5000
Hematite (o)	420 to 38 000
Pyrrhotite (o)	1250 to 6.3×10^6
Ilmenite (o)	314 000 to 3.8×10^6
Magnetite (o)	70 000 to 2×10^7

(d) = diamagnetic material; (o) = ore; (m) = with magnetic
*$\kappa \times 10^6$ rationalised SI units; to convert to the unrationalised c.g.s. units, divide by 4π
Data from Parasnis (1986), Sharma (1986), Telford *et al.* (1990)

images of the estuary floor. In conjunction with Thematic Mapper images obtained from low-flying aircraft and simultaneous water sampling from boats, it is possible to establish a detailed model of estuarine sediment dynamic processes, as has been achieved for Plymouth Sound in south-west England (Fitzpatrick 1991). For further details of the magnetic fabric method, see the discussions by Lowrie (1990) and Tarling (1983), for example.

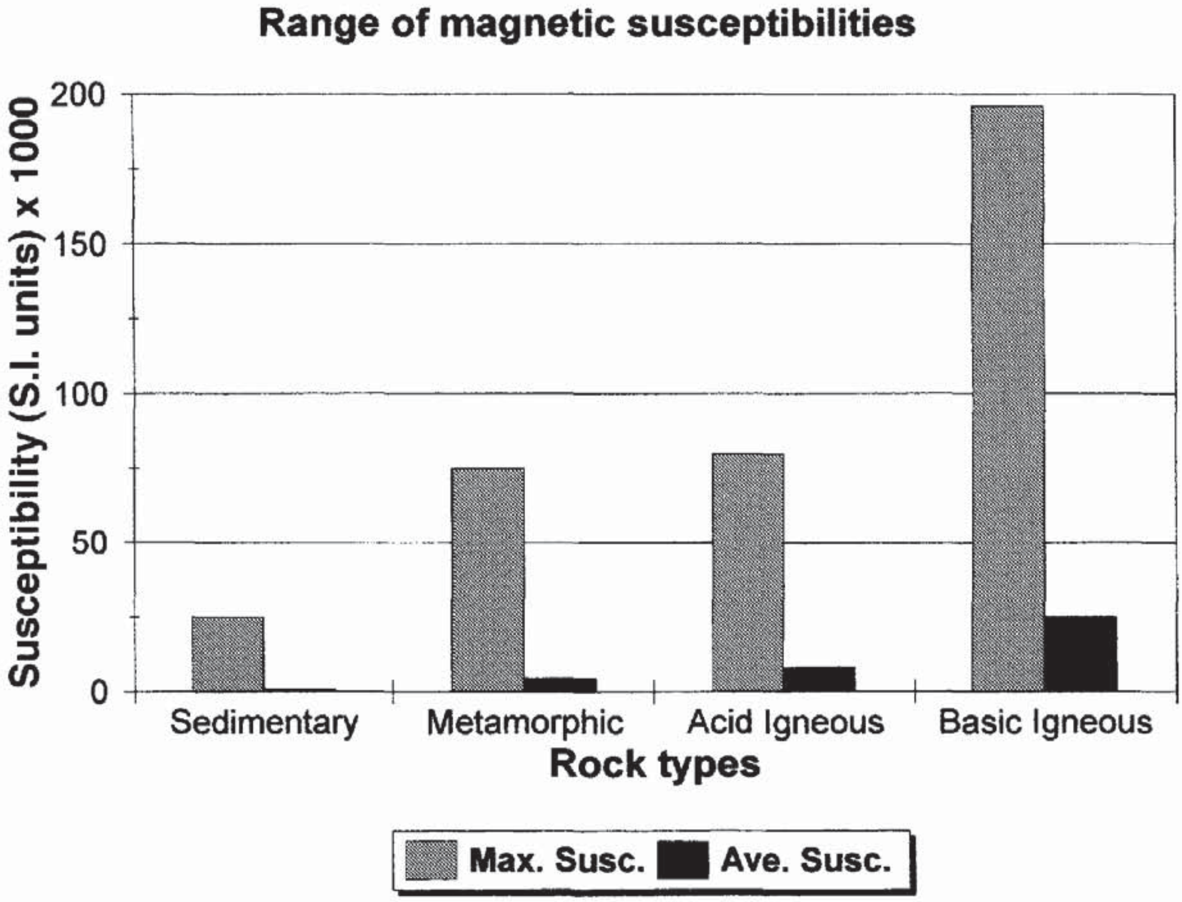

Figure 3.7 Susceptibilities of major rock types

For magnetic ore bodies with extremely high susceptibilities ($\kappa \geqslant 10^6$ SI), the measured susceptibility – more correctly referred to as the *apparent susceptibility* κ_a – can be reduced substantially by a shape demagnetisation effect (see Box 3.4). This involves a *demagnetisation factor* N_α which depends on a direction α. For a sphere, $N_\alpha = 1/3$ in all directions. In the case of a thin sheetlike body with a high true susceptibility ($\kappa \sim 10^6$ SI): $N_\alpha \sim 1$ in the transverse direction, giving a susceptibility $\kappa_a \approx 0.5K$; and $N_\alpha \sim 0$ in the longitudinal direction, so that $\kappa_a \approx \kappa$. Demagnetisation factors are discussed further by Parasnis (1986) and Sharma (1986).

Box 3.4 Apparent susceptibility κ_a and the demagnetisation factor N_α

$$\kappa_a = \kappa/(1 + N_\alpha \kappa)$$

Susceptibilities can be measured either in the field using a hand-held susceptibility meter such as the kappameter, or on samples returned to a laboratory where they can be analysed more accurately.

3.3.2 Remanent magnetisation and Königsberger ratios

In addition to the induced magnetisation, many rocks and minerals exhibit a permanent or *natural remanent magnetisation* (NRM) of intensity $\boldsymbol{J}_r$ when the applied field $\boldsymbol{H}$ is zero. The various processes by which rocks can acquire a remanent magnetisation are listed in Table 3.3 and discussed in more detail by Merrill (1990), Sharma (1986) and Tarling (1983).

Table 3.3 Types of remanent magnetisation (RM). After Merrill (1990), by permission

Type of RM	Process
Natural (NRM)	Acquired by a rock or mineral under natural conditions
Thermal (TRM)	Acquired by a material during cooling from a temperature greater than the Curie temperature to room temperature (e.g. molten lava cooling after a volcanic eruption)
Isothermal (IRM)	Acquired over a short time (of the order of seconds) in a strong magnetic field at a constant temperature (e.g. such as by a lightning strike)
Chemical (CRM)	Also crystallisation RM; acquired at the time of nucleation and growth or crystallisation of fine magnetic grains far below the Curie point in an ambient field
Thermal–chemical (TCRM)	Acquired during chemical alteration and cooling
Detrital (DRM)	Also depositional RM; acquired by the settling out of previously magnetised particles to form ultimately consolidated sediments which then have a weak net magnetisation, but prior to any chemical alteration through diagenetic processes
Post-depositional (PDRM)	Acquired by a sediment by physical processes acting upon it after deposition (e.g. bioturbation and compaction)
Viscous VMR	Acquired after a lengthy exposure to an ambient field with all other factors being constant (e.g. chemistry and temperature)
Anhysteretic (ARM)	Acquired when a peak amplitude of an alternating magnetic field is decreased from a large value to zero in the presence of a weak but constant magnetic field

Primary remanent magnetisations are acquired by the cooling and solidification of an igneous rock from above the Curie temperature (of the constituent magnetic minerals) to normal surface temperature (TRM) or by detrital remanent magnetisation (DRM). Secondary remanent magnetisations, such as chemical, viscous or post-depositional remanent magnetisations, may be acquired later on in the rock's history. This is especially true of igneous rocks which have later undergone one or more periods of metamorphism, particularly thermal metamorphism.

The intensity of the remanent magnetisation J_r may swamp that of the induced magnetisation J_i, particularly in igneous and thermally metamorphosed rocks. The ratio of the two intensities (J_r/J_i) is called the *Königsberger ratio*, Q, which can be expressed in terms of the Earth's magnetic field at a given locality and the susceptibility of the rocks (Box 3.5). Just as susceptibility can vary within a single rock type, so too can the Königsberger ratio. However, similar rock types have characteristic values of Q, some of which are listed in Table 3.4. Nagata (1961) has made four broad generalizations on the basis of Q:

- $Q \sim 1$ for slowly crystallised igneous and thermally metamorphosed rocks in continental areas;
- $Q \sim 10$ for volcanic rocks;

Table 3.4 Examples of values of the Königsberger ratio

Rock type	Location	Q
Basalt	Mihare volcano, Japan	99–118
Oceanic basalts	Northeast Pacific	15–105
Oceanic basalts	Mid-Atlantic Ridge	1–160
Sea-mount basalts	North Pacific	8–57
Cainozoic basalts	Victoria, Australia	5
Early tertiary basalts	Disko, West Greenland	1–39
Tholeiite dykes	England	0.6–1.6
Dolerite sills	North England	2–3.5
Dolerite	Sutherland, Scotland	0.48–0.51
Quartz dolerite	Whin Sill, England	2–2.9
Gabbro	Småland, Sweden	9.5
Gabbro	Minnesota, USA	1–8
Gabbro	Cuillin Hills, Scotland	29
Andesite	Taga, Japan	4.9
Granite	Madagascar	0.3–10
Granite plutons	California, USA	0.2–0.9
Granodiorite	Nevada, USA	0.1–0.2
Diabase	Astano Ticino, Switzerland	1.5
Diabase dykes	Canadian Shield	0.2–4
Magnetite ore	Sweden	1–10
Magnetite ore	South India	1–5

- $Q \sim 30–50$ for many rapidly quenched basaltic rocks;
- $Q < 1$ in sedimentary and metamorphic rocks, except when iron ore is involved.

Box 3.5 Königsberger ratio, Q

$$Q = \boldsymbol{J}_r/\kappa(\boldsymbol{F}/\mu_0)$$

where $\boldsymbol{J}_r$ is the intensity of remanent (NRM) magnetisation, κ is the susceptibility, μ_0 is the permeability of free space and $\boldsymbol{F}$ is the magnitude of the Earth's magnetic field (in tesla) at a given location in the same sense as the B-field (flux density).

It is also very important to consider that not only may $\boldsymbol{J}_r$ exceed $\boldsymbol{J}_i$, but the direction of remanent magnetisation may be quite different from that of the ambient induced field at a location. Consequently, the resultant magnetisation (i.e. the vectorial sum of the remanent and induced magnetisations) will give rise to characteristic magnetic anomalies (refer back to Figure 3.3) when reversely magnetised rocks are present.

3.4 THE EARTH'S MAGNETIC FIELD

3.4.1 Components of the Earth's magnetic field

The geomagnetic field at or near the surface of the Earth originates largely from within and around the Earth's core. Currents external to the Earth in the ionosphere and magnetosphere associated with the Van Allen radiation belts (Figure 3.8), currents induced in the Earth by external field variations and the permanent (remanent) and steady-state induced magnetisations of crustal rocks, also contribute to the overall geomagnetic field. The magnetosphere is vital for the survival of life on Earth as it forms the primary force field which protects the planet from harmful radiation from the Sun. The various components of the geomagnetic field affect exploration surveys in a variety of ways which will be discussed in turn.

3.4.1.1 The main dipole field

The main component of the geomagnetic field is called the *dipolar field* as it behaves, to a first-order approximation, like a dipolar electromagnet located at the centre of the Earth but inclined at 11.5° to the rotational axis (Figure 3.9).

The *geomagnetic poles*, the positions on the Earth's surface through which the axis of the best-fitting dipole passes – which are located in Hayes Peninsula in northern Greenland and near the Russian Vostok research station in Greater Antarctica – are not the same as the

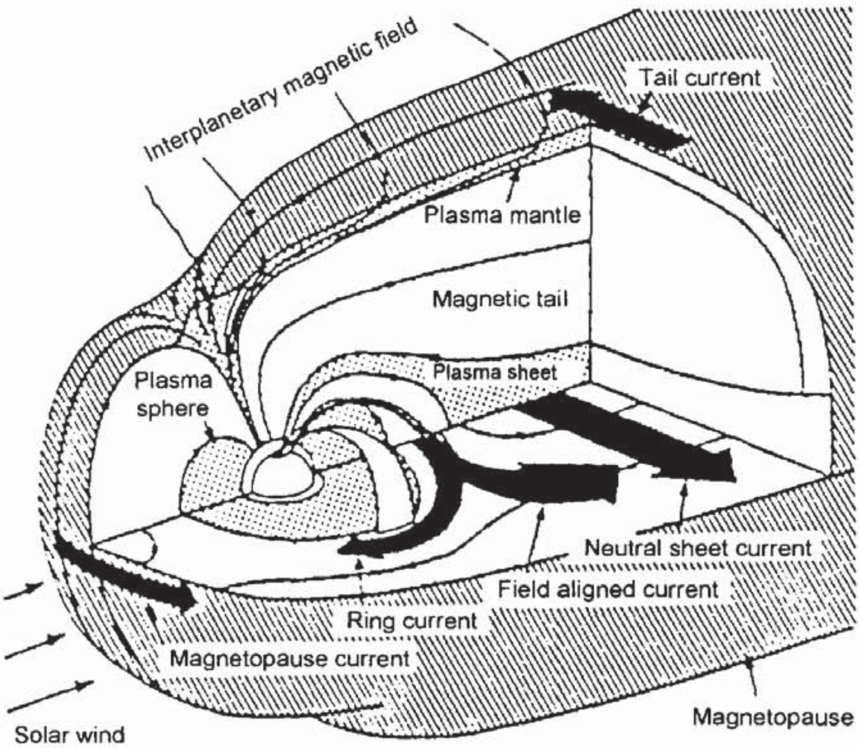

Figure 3.8 The geomagnetic field showing the magnetosphere, magnetopause and Van Allen radiation belts. From James (1990), by permission

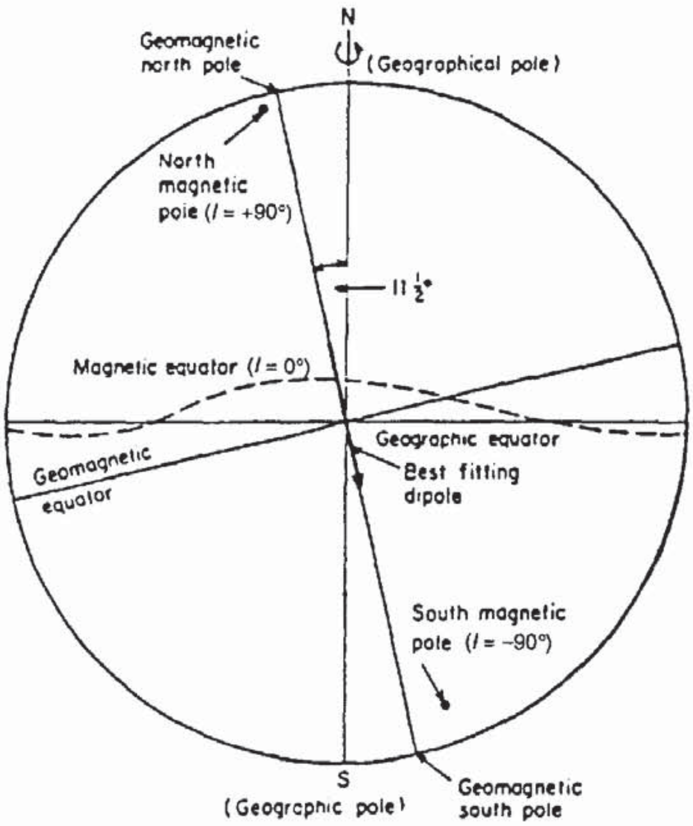

Figure 3.9 The field due to an inclined geocentric dipole. From McElhinny (1973), by permission

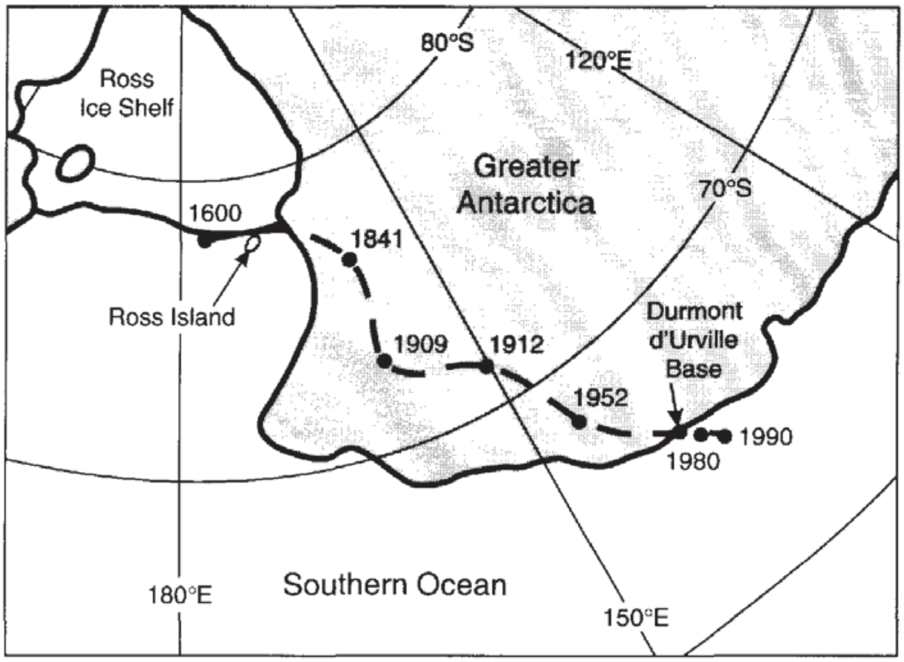

Figure 3.10 Location of the south magnetic pole and its drift since the year 1600

magnetic or *dip poles*. These are located where the magnetic field is directed vertically. The north magnetic pole is currently just north of Bathurst Island in the Canadian Arctic Archipelago and was discovered on 1 June 1831 by James Clark Ross and his uncle Sir John Ross. The south magnetic pole is currently about 150 km offshore from the French Research Station Durmont d'Urville on the Adélie Coast of Greater Antarctica. James Ross came extremely close to locating the south magnetic pole in 1841 but as it was then inland (Figure 3.10) he was thwarted by icebergs and the land ice. It was only on 15 January 1909 that Alistair Mackay, Edgeworth David and Douglas Mawson reached the south magnetic pole after an epic sledge journey from their base on Ross Island.

The geomagnetic field is produced by electric currents induced within the conductive liquid outer core as a result of slow convective movements within it (Figure 3.11). It is for this reason that the analogy of the Earth's field to that induced by an electromagnet is preferred to that of a permanently magnetised bar magnet. The liquid core behaves as a geodynamo but the precise nature of the processes involved has yet to be resolved. Models to explain the disposition of the magnetic field must also account for the slow but progressive change in field intensity and westward drift in direction known as the *secular variation*. Furthermore, the model must also explain how the Earth's magnetic field goes through reversals of magnetic polarity. The study of how the Earth's magnetic field has changed through geological time is known as *palaeomagnetism*. The use of magnetic reversals to provide global chronometric calibration of geological events is known as *magnetostratigraphy* (Tauxe, 1990).

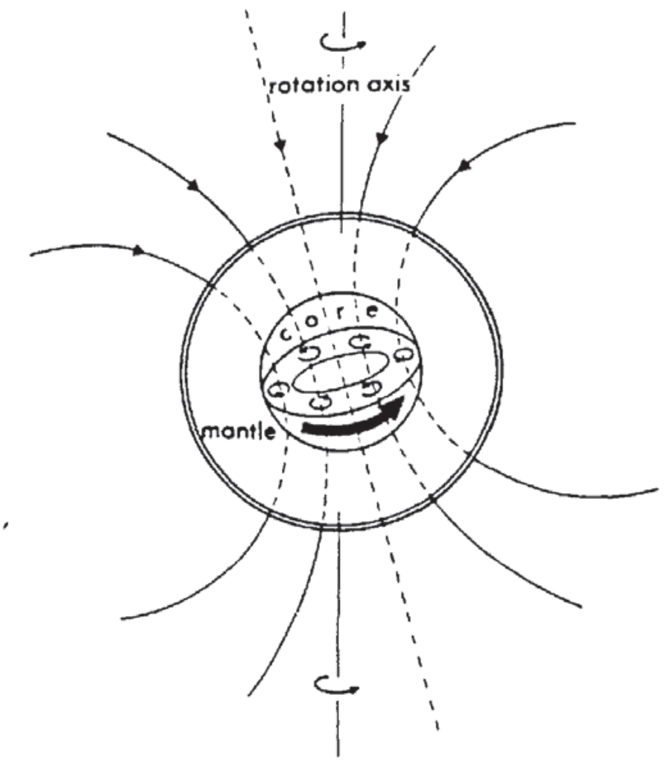

Figure 3.11 Schematic of the cause of the majority of the Earth's magnetic field. From Sharma (1986), by permission

The geomagnetic field can be described in terms of the declination, D, inclination, I, and the total force vector $\boldsymbol{F}$ (Figure 3.12). A freely suspended magnetised needle will align itself along the $\boldsymbol{F}$ vector so that at the magnetic (dip) north, the inclination is 90°; i.e. the needle will point vertically downwards. At the south magnetic (dip) pole, the needle will point vertically upwards. At the magnetic equator, the needle will lie horizontally (Figure 3.13). Furthermore, the vertical component of the magnetic intensity of the Earth's magnetic field varies with latitude, from a minimum of around 30 000 nT at the magnetic equator to 60 000 nT at the magnetic poles.

3.4.1.2 The non-dipolar field

While the single dipole field approximates to the Earth's observed magnetic field, there is a significant difference between them, which is known as the *non-dipole field*. The total intensity for the non-dipole field is shown in Figure 3.14, from which several large-scale features can be seen with dimensions of the order of several thousand kilometres and with amplitudes up to 20 000 nT, about one-third of the Earth's total field. Using the method of spherical harmonic analysis, it can be demonstrated that the non-dipole field and the associated large-scale features can be represented by a fictitious set of 8–12 small dipoles radially located close to the liquid core. These dipoles serve to simulate the eddy currents associated with the processes within the liquid core.

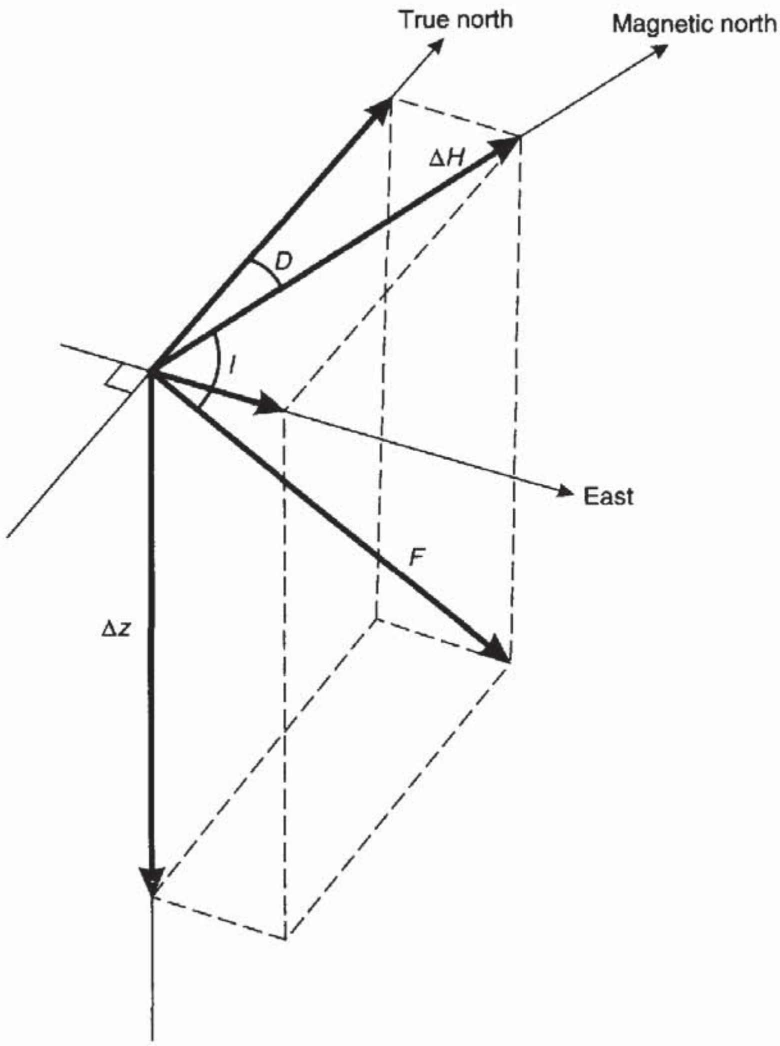

Figure 3.12 Elements of the magnetic field: inclination ***I***, declination ***D***, and total magnetic force ***F***

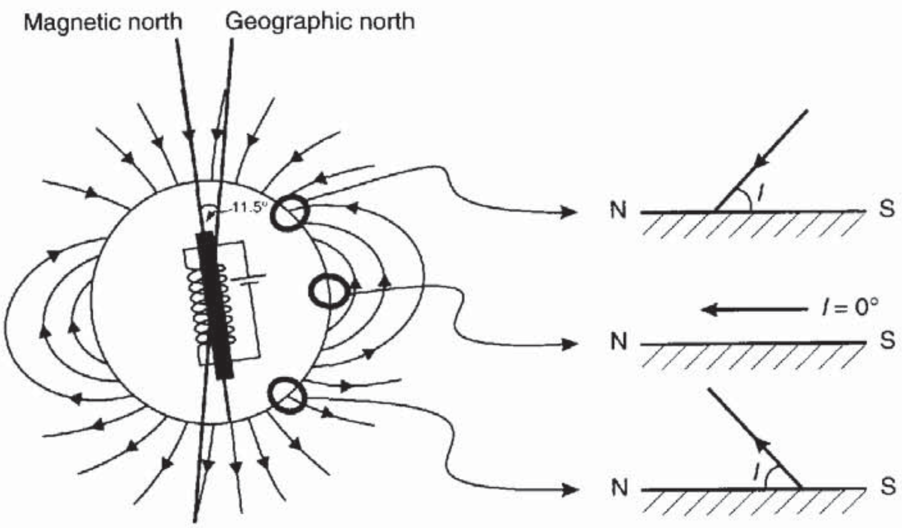

Figure 3.13 Variation of inclination with latitude

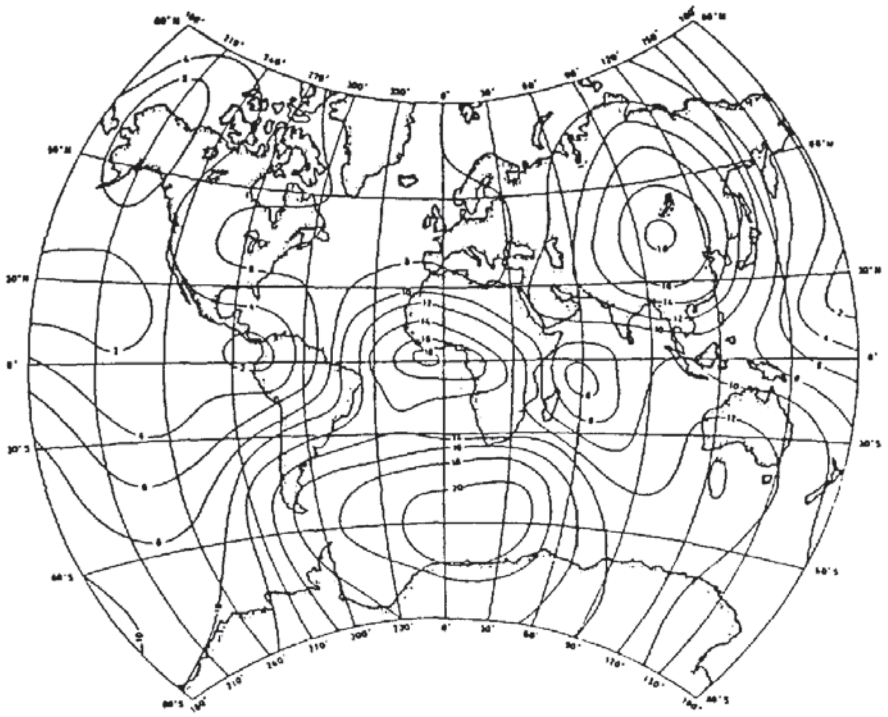

Figure 3.14 Variation in the intensity of the non-dipole field for epoch 1980. From Sharma (1986), by permission

A further use of the spherical harmonic analysis is that it provides a means whereby the spatial distribution and intensity of the total magnetic field can be calculated for the whole globe. The total field, which is calculated every five years, is called the *International Geomagnetic Reference Field* (IGRF) and the year of calculation is known as the *epoch*. It has to be recalculated regularly because of the secular variation (see, for example, IAGA (1987) and Peddie (1982)). Consequently, it is possible to obtain a theoretical value for the field strength of the Earth's magnetic field for any location on Earth (Figure 3.15). It can be seen from this figure that instead of the anticipated two maxima consistent with a truly dipolar field, there are in fact four maxima. The significance of the IGRF in processing magnetic data is discussed in Section 3.6.3.

Data used in the computation of revisions of the International Geomagnetic Reference Field have been obtained by satellite (e.g. during 1965–71, Polar Orbiting Geophysical Observatory series, POGO; October 1979 to June 1980, MAGSAT). However, at satellite orbit ranges, perturbations in the earth's magnetic field caused by magnetic materials in the crust are not resolvable. Surface or airborne measurements can detect considerable high-amplitude small-scale features within the crust down to a depth of 25–30 km where the Curie isotherm is reached. These features may be caused by induction due to the Earth's field or remanent magnetisation or a mixture of both.

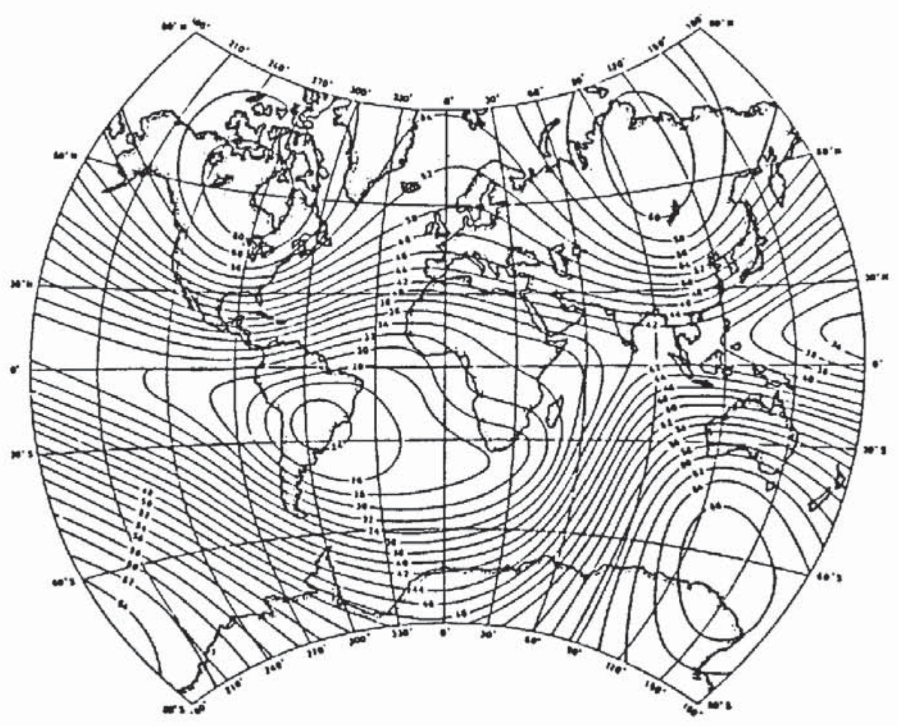

Figure 3.15 Total field intensity derived using the IGRF epoch 1980. From Sharma (1986), by permission

3.4.2 Time variable field

Observations of the Earth's magnetic field have been made for over four centuries at London and Paris. From these data, it is clear that the geomagnetic and magnetic pole positions drift with time, known as the secular variation in the magnetic field (Figure 3.16). In addition, the intensity of the main magnetic field is decreasing at about 5% per century. These rates of change, although very significant on a geological time scale, do not affect data acquisition on a typical exploration survey unless it covers large geographical areas and takes many months to complete, or if such surveys are being used to compare with historical data.

The Earth's magnetic field changes over a daily period, the *diurnal variations*. These are caused by changes in the strength and direction of currents in the ionosphere. On a magnetically 'quiet' (Q) day, the changes are smooth and are on average around 50 nT but with maximum amplitudes up to 200 nT at the geomagnetic equator. The changes are least during the night when the background is almost constant, and decrease in amplitude from dawn to midday whereupon they increase to the daily maximum about mid-late afternoon before settling down to the night-time value.

Magnetically disturbed (D) days are marked by a rapid onset of fluctuations of the order of hundreds of nanoteslas followed by slower but still erratic fluctuations with decreasing amplitude. These

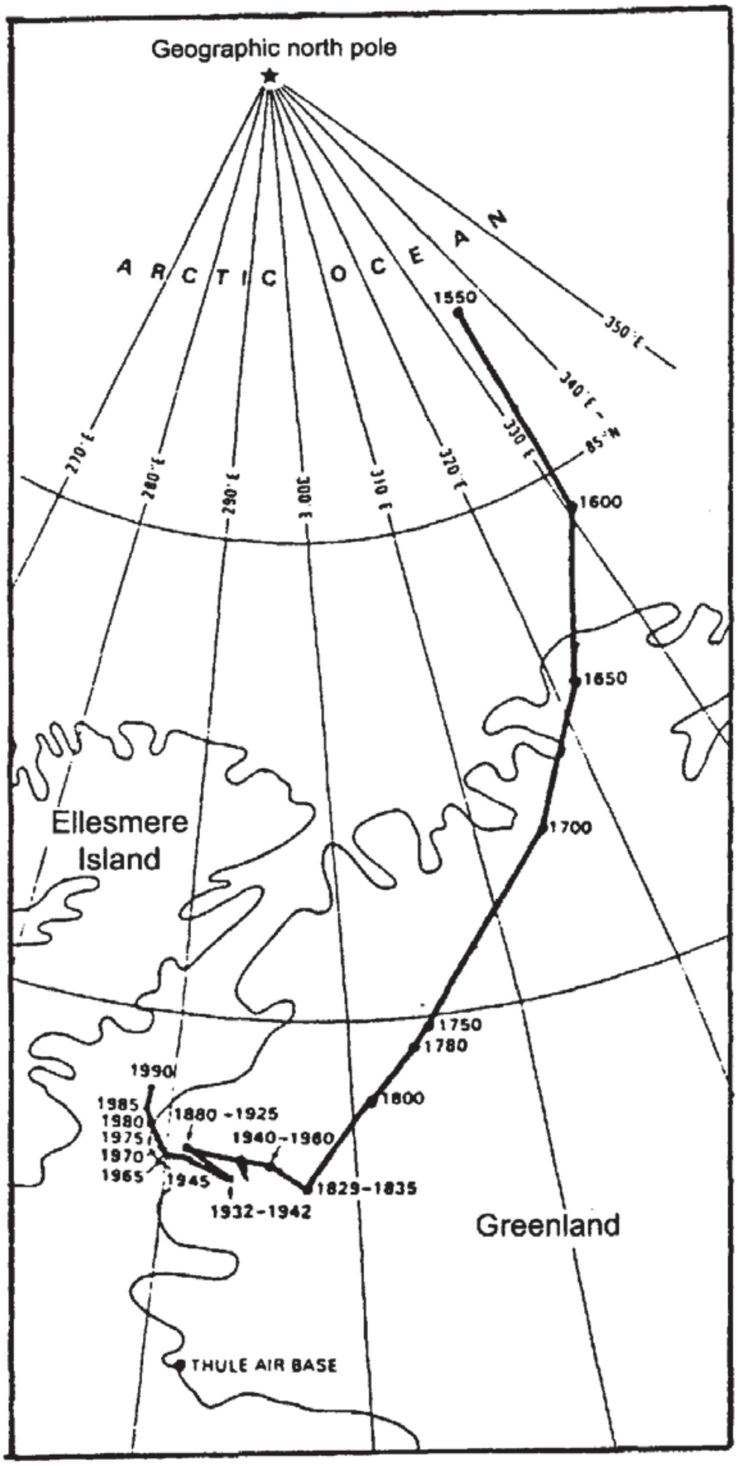

Figure 3.16 Drift of the north magnetic pole position from 1550 to 1990. From James (1990), by permission

disturbances, which are called *magnetic storms*, may persist for several hours or even days. Such frenetic magnetic activity is caused by sunspot and solar activity resulting in solar-charged particles entering the ionosphere. This may happen on fine sunny days and not necessarily in stormy weather. Magnetic observatories around the