[4]
Magnetic properties of natural materials

Walking around this little island (of Chaul) . . . a wonderful thing happened to me . . . placing the needle on top of a big boulder . . . the rose turned . . . it occurred to me that such a strange fact was due to the quality and nature of the rock.

Extract from log book of Joao de Castro 1538

4.1 Introduction

The main impetus for the study of the magnetic properties of rocks and sediments came with the realisation that many natural materials record valuable information about the Earth's ancient field by retaining a magnetic remanence acquired close to the time of their formation. The majority of magnetic studies of natural samples have thus been carried out by palaeomagnetists investigating natural remanent magnetisations. These studies have involved many rock types spanning all ages, and collected from all parts of the world. In association with these researchers into the Earth's fossil magnetism, studies of synthetic materials have also been carried out. These are often referred to as rock magnetic studies (Nagata 1953). They have concentrated on providing background information of use to the palaeomagnetist, and have dealt mainly with natural iron minerals commonly capable of carrying a stable remanence, such as haematite and the titanomagnetites (e.g. Day *et al.* 1977, Bailey & Dunlop 1983). A major endeavour of rock magnetists has been to elucidate the intricacies of thermoremanent magnetisation through laboratory studies of synthetic samples.

When dealing with natural samples we shall inevitably be involved with mixtures of magnetic materials. This chapter is largely concerned with setting a framework within which the magnetic properties of such mixtures can be interpreted by drawing on and summarising the results of a wide variety of rock magnetic studies.

In magnetic studies of an environmental nature, as described in Chapters 7–12 and 16, we have found mineral magnetic investigations to be invaluable in allowing unprecedentedly large quantities of material to be analysed and so we have concentrated on simple, flexible, rapid magnetic measurements which enhance this aspect of the work. Such reconnaissance mineral magnetic studies can then be followed up by more detailed and more diagnostic measurements on selected, or key, samples. The environmental mineral magnetic properties of greatest use have turned out to be magnetic susceptibility and isothermal remanence. Our interest has largely lain with magnetic properties at room temperature, despite the intrinsic connection between temperature and magnetic properties long known to the physicist. This is because in many of our environmental samples chemical changes readily take place at elevated temperatures.

21

Table 4.1 Units in magnetism and their relationships.

Quantity	SI	CGS (emu)	Relationship
induction in free space (field)	B_0 tesla (T)	B gauss (G)	$1\,T = 10^4 G$
magnetic force (field)	$H\,A\,m^{-1}$	H oersted (Oe)	$1\,A\,m^{-1} = 4\pi \times 10^{-3}\,Oe$
permeability of a vacuum	$\mu_0 = 4\pi \times 10^{-7}\,H\,m^{-1}$	$\mu_0 = 1$	$1\,H\,m^{-1}$ equivalent to $10^7/4\pi\,G\,Oe^{-1}$
induction in free space (field)	$B_0 = \mu_0 H$	$B = H$	1 T equivalent to 10^4 Oe
induction in medium	$B = B_0 + \mu_0 M$	$B = H + 4\pi I$	$1\,T = 10^4\,G$
magnetisation per unit volume	$M\,A\,m^{-1}$	I G	$1\,A\,m^{-1} = 10^{-3}\,G$
magnetisation per unit mass	$\sigma = M/\rho\,A\,m^2\,kg^{-1}$ ρ = density	$\sigma = I/\rho\,G\,cm^3\,g^{-1}$	$1\,A\,m^2\,kg^{-1} = 1\,G\,cm^3\,g^{-1}$
susceptibility per unit volume	$\kappa = M/H$	$\kappa = I/H$	1 (SI unit) = $4\pi\,G\,Oe^{-1}$
susceptibility per unit mass	$\chi = \kappa/\rho\,m^3\,kg^{-1}$	$\chi = \kappa/\rho\,G\,Oe^{-1}\,cm^3\,g^{-1}$	$1\,m^3\,kg^{-1} = 4\pi \times 10^{-3}\,G\,Oe^{-1}\,cm^3\,g^{-1}$

4.2 Units

A diversity of magnetic unit systems are currently in use. The system we are following is an SI system based on Crangle (1975). Table 4.1 summarises the SI units in this system and their relationship with the older CGS units. Table 4.2 is a practical table illustrating the differences between the two systems. It tabulates the magnetic properties of single- and multidomain magnetite and haematite using both SI and CGS units.

4.3 Magnetic remanence

4.3.1 Natural magnetic remanences

Rocks, sediments and soils can acquire remanent magnetisations by natural processes. Such natural remanences are not as intense as those which can be artificially imparted in strong laboratory fields but they can be just as stable. Somewhat surprisingly very stable remanences can be acquired in weak magnetic fields such as the Earth's field. This can happen when a suitable natural event occurs to change the co-ercivity of the magnetic minerals from very low to very high values. For example, in igneous rocks, the event is simply the cooling of the magnetic minerals through their Curie and **blocking temperatures**; in red sediments it is the chemical growth of secondary magnetic minerals through their critical **blocking volumes**; and in deep-sea sediments the event is consolidation, which locks the tiny detrital, micron-sized magnetic minerals firmly into the bulk of the sediment.

THERMOREMANENT MAGNETISATION
A thermoremanent magnetisation is acquired simply by a magnetic mineral cooling from above its Curie temperature in a magnetic field. If the sample is isotropic the magnetisation is parallel to the applied field and, for small fields, the **intensity** of the remanence is proportional to the field. Furthermore, the remanence is remarkably stable and this means that thermoremanences can survive with little change through geological time. Thermoremanent mag-netisations can thus accurately record the direction and intensity of weak magnetic fields at remote times in the past as described in Chapter 13.

The intensity of the thermoremanence of multi-domain grains is much lower than that of single-domain grains (Fig. 4.1). Many rocks have magnetic grains, which, although they can be seen under the microscope to be of multidomain size ($> 20\ \mu$m), behave magnetically as though they contain a remanence carried by single-domain grains. There are a number of possible explanations for this behaviour, including the microscopic subdivision of large grains to single-domain size, inclusions of submicroscopic magnetic particles, and pseudo single-domain behaviour. The remanence of mixtures of multi-domain, single-domain and superparamagnetic grains is overwhelmingly influenced by the single-domain fraction. Consequently, the thermoremanence of stable single-domain grains plays an important rôle in many palaeomagnetic studies.

CHEMICAL REMANENT MAGNETISATION
When a magnetic mineral is produced by chemical changes at temperatures below its Curie temperature, it can acquire a remanence in the direction of the ambient field. A **chemical remanent magnetisation** is locked into a magnetic grain when it grows larger than a critical size called its blocking volume. If chemical growth were to continue well above the blocking volume, a grain could eventually become multidomain, passing through a pseudo single-domain stage.

Table 4.2 Magnetic properties of magnetite and haematite in SI and CGS units.

Quantity		Single-domain magnetite		multidomain magnetite		Haematite	
saturation magnetisation	M_S	480 kAm^{-1}	(480 G)	480 kAm^{-1}	(480 G)	2.5 kAm^{-1}	(2.5 G)
saturation magnetisation per per unit mass	σ_S	92 A m^2 kg^{-1}	(92 G cm^3 g^{-1})	92 A m^2 kg^{-1})	(92 G cm^3 g^{-1})	0.5 A m^2 kg^{-1}	(0.5 G cm^3 g^{-1})
saturation remanence	M_{RS}	50 kA m^{-1}	(50 G)	5 kA m^{-1}	(5 G)	1 kA m^{-1}	(1 G)
saturation remanence per unit mass	σ_{RS}	10 A m^2 kg^{-1}	(10 G cm^3 g^{-1})	1 A m^2 kg^{-1}	(1 G cm^3 g^{-1})	0.2 A m^2 kg^{-1}	(0.2 G cm^3 g^{-1})
susceptibility of ARM per unit mass	χ_{ARM}	800 µm^3 kg^{-1}	(0.07 cm^3 g^{-1})	100 µm^3 kg^{-1}	(0.01 cm^3 g^{-1})	10 µm^3 kg^{-1}	(0.001 cm^3 g^{-1})
susceptibility	κ	2.4	(0.19)	2.8	(0.22)	125 × 10^{-6}	(10 × 10^{-6})
susceptibility per unit mass	χ	450 µm^3 kg^{-1}	(0.04 cm^3 g^{-1})	530 µm^3 kg^{-1}	(0.04 cm^3 g^{-1})	6 µm^3 kg^{-1}	(50 µcm^3 g^{-1})
coercive force	$(B_0)_C$	10 mT	(100 Oe)	2 mT	(20 Oe)	0.4 T	(4000 Oe)
coercivity of remanence	$(B_0)_{CR}$	33 mT	(33) Oe)	15 mT	(150 Oe)	0.7 T	(7000 Oe)
density	ρ	5250 kg m^{-3}	(5.25 g cm^{-3})	5250 kg cm^{-3}	(5.25 g cm^{-3})	5000 kg cm^{-3}	(5 g cm^{-3})

Chemical remanence is of a lower magnitude than thermoremanence because the saturation magnetisation and anisotropy energy are lower at temperatures well below the Curie temperature. The stabilities of the two remanence types are however similar. Much less laboratory work has been carried out on the magnetic properties of chemical remanences although they are common in nature and have been widely used for investigating ancient field directions, particularly in sedimentary rocks.

DETRITAL REMANENT MAGNETISATION

Detrital magnetic particles (with a previously acquired thermal or chemical remanence) can align themselves with an applied magnetic field while falling through water, and lead to a **detrital remanence.** However, it is more likely that the detrital remanence of most sediments is formed after the particles have come to rest on the substrate, by the rotation of the magnetic grains in water-filled interstices. This so-called post-depositional remanent magnetisation is locked into the sediment by consolidation due to either compaction or to the growth of authigenic minerals or organic gels. In some slowly deposited sediments it may be many tens of thousands of years after deposition before the detrital magnetic particles become consolidated or buried and record the ambient field. In other sediments the post-depositional remanence can be locked into the material in only a few days or weeks after the time of deposition. The origin of natural remanent magnetisation in sediments is described further in Chapter 13 along with related laboratory redeposition experiments.
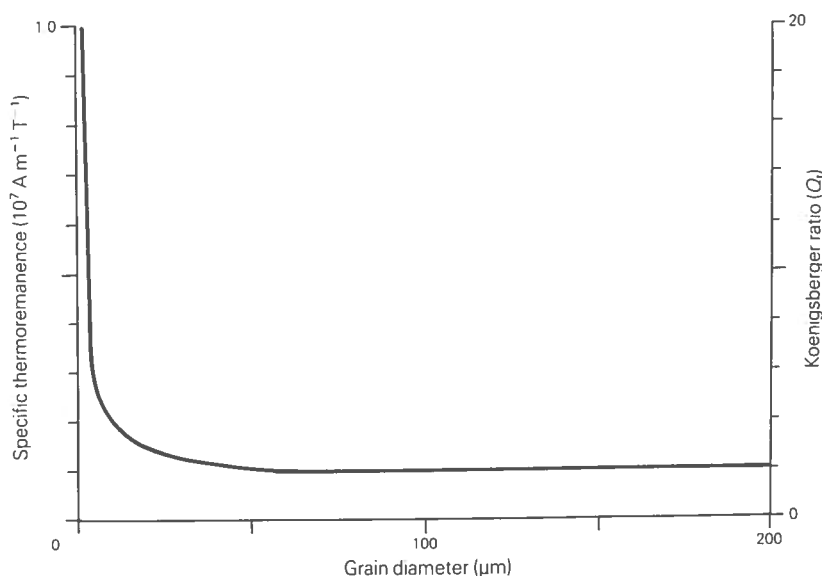

Figure 4.1 Dependency of low field thermoremanence intensity, per tesla, with magnetic grain size. The Koenigsberger ratio, Q_t, expresses the strength of the thermoremanence. It is defined as the ratio of the thermoremanence acquired in a low field to the magnetisation induced by the same low field at room temperature.
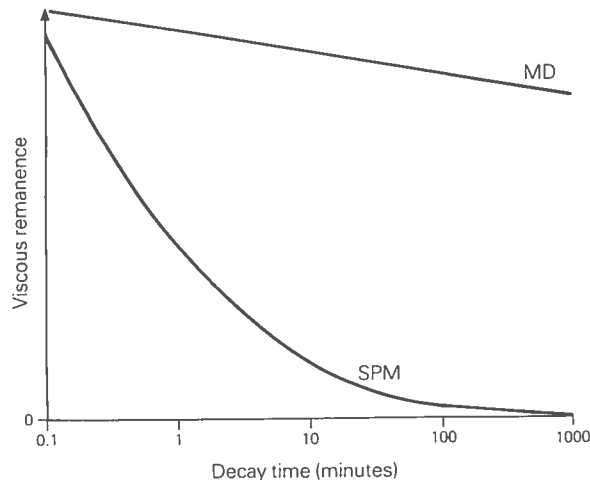
Figure 4.2 Decay of viscous remanence of multidomain (MD) and superparamagnetic (SPM) grain assemblages in zero field.

4.3.2 Laboratory-imparted remanences

VISCOUS REMANENT MAGNETISATION

A viscous remanence can be acquired when a sample is exposed to a new magnetic field. The remanence is, by definition, time dependent (2.5). In the nineteenth century Ewing (1900) observed that viscous remanence had a logarithmic time dependence.

The logarithmic time dependence can be predicted theoretically for both multidomain and interacting assemblages of single-domain grains (Dunlop 1973b). The stable single-domain/superparamagnetic boundary, however, truncates the logarithmic relationship and leads to characteristic time–remanence curves (Fig. 4.2) which contrast with those of multidomain materials. Magnetic grains near the stable single-domain/superparamagnetic boundary with low relaxation times are probably the most important contributors to viscous remanences in natural materials (Mullins & Tite 1973, Dunlop 1983b).

Decay of natural remanence with time during zero field storage indicates the presence of 'soft' components of viscous remanence. Such components can often be removed by waiting a few days or weeks for the natural remanence to stabilise. Such soft remanences are commonly found in recent sediments which have been held in a core store for some time and especially in those which have partially dried. These storage remanences are generally due to a small portion of the magnetic particles gradually realigning themselves in the new field direction of the core store,

rather than to superparamagnetic or domain wall thermal activation effects.

ISOTHERMAL REMANENT MAGNETISATION

The remanent magnetisation acquired by deliberate exposure of a material to a steady field at a given temperature (most commonly room temperature) is called an **isothermal remanence**. The magnitude of the remanence depends on the strength of the steady field applied. This dependence can be readily demonstrated in the laboratory by placing a specimen in stronger and stronger fields and measuring the remanence after each field exposure. Figure 4.3 illustrates the increase in isothermal induced *magnetisation* with field, while Figure 2.3 illustrates changes in isothermal *remanence* with field. In later chapters when we are describing experiments which involve subjecting specimens to fields of different strengths almost all our measurements will be of the isothermal remanent magnetisation remaining in the specimens after they have been withdrawn from the influence of the applied fields, rather than their magnetisations while in the fields.
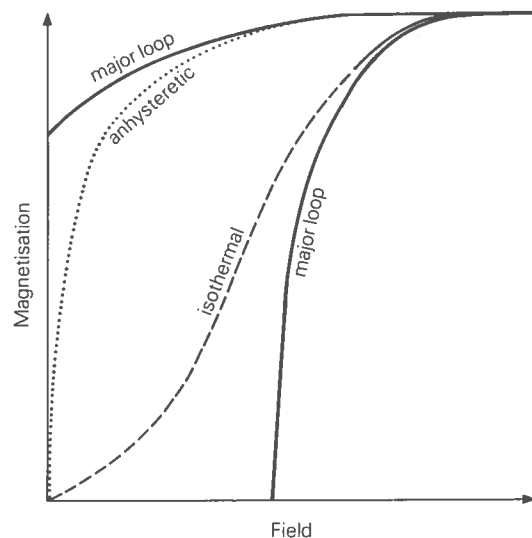

Figure 4.3 Schematic diagram illustrating acquisition curves for anhysteretic (dotted) and isothermal (dashed) magnetisations. The isothermal magnetisation acquisition curve increases slowly in low fields, then more rapidly before saturating. The anhysteretic magnetisation curve follows a simpler path, which closely parallels the right-hand side of the major hysteresis loop, before saturating. On the anhysteretic, or ideal magnetism curve, the domain walls achieve positions of true equilibrium. So the anhysteretic curve always lies above the initial magnetisation curve.

The maximum remanence which can be produced is called the saturation isothermal remanent magnetisation (expressed by the alternative symbols M_{RS}, σ_{RS}, J_{RS} and SIRM). The field at which saturation is reached depends on composition and microstructure (e.g. grain size). The highest isothermal remanence which it is practical to produce using laboratory equipment may not actually be the true saturation remanence but may instead only be an isothermal remanence produced on the approach to saturation. This circumstance arises, for example, when haematite is a constituent mineral. We denote our laboratory measurements by the symbol SIRM rather than σ_{RS} (the physicists' symbol for saturation remanence) in order to indicate that our measurements are of the highest isothermal remanence we can produce with our equipment but recognising that our SIRM will, in certain circumstances, fall somewhat short of the true saturation remanence.

Isothermal remanence is found naturally in materials which have been struck by lightning, as large magnetic fields associated with lightning strikes can induce strong, although generally low stability, isothermal remanences.

ANHYSTERETIC REMANENT MAGNETISATION

Anhysteretic remanent magnetisation is generally imparted by subjecting a sample to a strong alternating field which is smoothly decreased to zero in the presence of a small steady field. Anhysteretic (free from hysteresis) remanence is sometimes referred to as an ideal remanence.

Anhysteretic remanence increases in strength with the application of either a stronger steady field or a stronger alternating field until saturation is reached (Fig. 4.3). A linear change in remanence with field strength is found for steady fields with magnitudes similar to the Earth's field (Patton & Fitch 1962). This linear rate of change is referred to as the susceptibility of anhysteretic remanence. In practice anhysteretic remanences can be easily produced in palaeomagnetic laboratories by using the same equipment as that used for alternating field demagnetisation (6.5.1). The sample can be magnetised rather than demagnetised simply be ensuring that a steady direct field acts on the sample throughout the experiment.

4.4 Magnetic susceptibility

Susceptibility is a measure of the ease with which a material can be magnetised. Volume susceptibility is defined by the relation $\kappa = M/H$, where M is the volume magnetisation induced in a material of susceptibility, κ, by an applied field, H. By using this definition we have volume susceptibility as a dimensionless quantity. SI values of volume susceptibility are 4π times larger than CGS values (Table 2.1). **Specific susceptibility**, χ, is defined as volume susceptibility divided by density $\chi = \kappa/\rho$ (Table 4.1) and has units of $m^3\ kg^{-1}$. Susceptibility is generally measured in small fields, of strength less than 1 mT. At these low fields it is found that susceptibility is reasonably independent of applied field intensity.

Magnetic susceptibility measured by the usual methods (6.3) is an apparent value because of the self-demagnetising effect described in Section 2.4 in connection with anisotropy. When a substance is magnetised its internal magnetic field is less than the externally applied field. κ_i, the intrinsic susceptibility, relates the induced magnetisation to the internal magnetic field, whereas κ_e, the extrinsic susceptibility which we actually observe, relates the induced magnetisation to the externally applied field. The relationship between the two susceptibilities can be easily derived and shown to be

$$\kappa_e = \kappa_i/(1 + N\kappa_i)$$

where N is the demagnetisation factor.

For a strongly magnetic mineral such as pure magnetite, $N\kappa_i > 1$. Hence we have $\kappa_e \approx 1/N$ and, if N is known, then there is a very simple relationship between the susceptibility we measure and the concentration of ferrimagnetic grains in a sample. This is the situation for natural samples where the concentration of ferrimagnetic grains is generally small, being a few percent or less. In terms of a volume fraction of ferrimagnetic grains, f, the measured susceptibility, κ, is given by $\kappa = f\kappa_e$ and so to a good approximation $\kappa = f/N$. In practice for natural samples it is found that N is reasonably constant with a value close to $\frac{1}{3}$, i.e. that expected for a sphere. So if the magnetic grain shapes are roughly spherical and the dominant magnetic mineral is magnetite, as can be expected in most natural samples, the volume fraction ($f \ll 1$) can be estimated directly by dividing the volume susceptibility by three. This simple relationship between susceptibility and magnetite concentration is extensively used in subsequent chapters.

A list of susceptibilities of common natural ferrimagnetic and antiferromagnetic minerals and of some other common minerals is given in Table 3.4. A
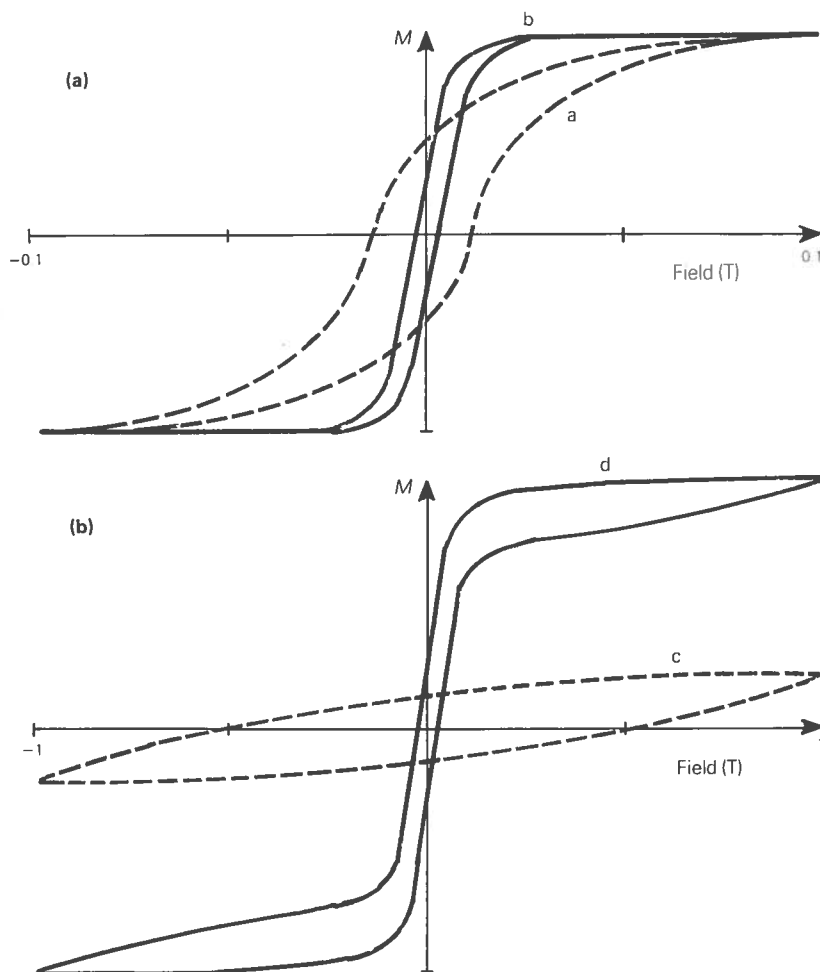
25

Figure 4.4 Magnetic hysteresis loops. (a) Loop a: assemblage of single-domain magnetite grains (dashed); loop b: multidomain magnetite (solid). The heights of the single-domain and multidomain magnetite loops are the same because saturation magnetisation does not vary with grain size. (b) Loop c: haematite (dashed); loop d: mixture of hard and soft minerals (solid).

number of studies have established that the magnetic susceptibility of natural materials mainly depends on their magnetite content (Mooney & Bleifuss 1953, Puranen 1977, Currie & Bornhold 1983). The inset of Figure 4.9 illustrates this simple linear relationship between susceptibility and magnetite content. When ferrimagnetic minerals such as magnetite are scarce and the susceptibility is consequently very weak, paramagnetic and even diamagnetic minerals can make substantial contributions to it (Prasad & Ghildyal 1975). **Frequency dependent susceptibility** can help identify samples with grains spanning the superparamagnetic/stable single-domain boundary (Stephenson 1971, Mullins & Tite 1973) and is discussed in Section 6.3 along with susceptibility instrumentation.

4.5 Anisotropy of susceptibility

The relative ease with which a material can be magnetised in various directions is expressed by its **anisotropy of susceptibility**. Such anisotropy (Section 2.4) in natural samples can arise through external stresses, such as natural tectonic stresses, or be intrinsic due to the shape or crystalline structure of the ferromagnetic minerals in the samples. The first type of anisotropy is of practical interest because of its contribution to changes in the local geomagnetic field as precursors of volcanic eruptions (Johnston & Stacey 1969) or earthquakes (Undzendor & Shapiro 1967, Johnston *et al.* 1976). The second type has application in fabric analyses (Hamilton & Rees 1970).

26

Table 4.3 General hysteresis properties.

Rock type	χ ($\mu m^3 kg^{-1}$)	SIRM ($mA m^2 kg^{-1}$)	SIRM/χ ($kA m^{-1}$)	$(B_0)_{CR}$ (mT)	$\dfrac{IRM_{-100\,mT}}{SIRM}$	NRM ($\mu A m^2 kg^{-1}$)
ultrabasic	0.2	0.4	2	15	0.9	200
gabbro	1.0	2.0	2	15	0.9	500
basalt	1.8	70	40	35	0.8	5000
diorite	0.5	1	2	20	0.7	10
granite	0.2	0.3	1.5	35	0.6	1
limestone	<0.001	0.005	>10	70	0.4	0.5
sandstone	0.1	0.5	7	30	0.8	5
siltstone	0.2	1.4	8	30	0.9	10
clay	0.1	1	10	40	0.9	50
gneiss	0.05	7.5	150	50	0.7	5
schist	0.01	0.01	1	20	0.9	1
slate	0.01	0.007	0.5	50	0.6	1

For natural samples to be anisotropic the shape or crystalline axes of the assemblage of magnetic grains must be aligned to form a fabric. The effect of grain shape alignment will dominate when the magnetic anisotropy is due to ferrimagnetic grains such as magnetite. The shape effect (Section 2.4) of ferrimagnetic grains results from the demagnetising field being less along the long axes of magnetic grains, so that the susceptibility is higher parallel to the long or major axes.

Crystalline anisotropy is more important than shape anisotropy in rhombohedral and hexagonal iron oxide (Section 3.2.2) and sulphide (Section 3.3) minerals as their spontaneous magnetisation is low. In these minerals the direction of minimum susceptibility is perpendicular to the basal crystallographic plane. In some metamorphic rocks the observed magnetic anisotropy results from alignment of the basal planes of pyrrhotite grains (Fuller 1963).

4.6 Magnetic hysteresis

As briefly described in Section 2.3 the magnetisation of a ferromagnetic material lags behind the external inducing magnetic field. This important magnetic phenomenon is known as hysteresis. It can be conveniently displayed for a specimen by plotting a graph of the magnetisation produced by a cyclic field. The graph takes the form of a symmetric loop (Figs 2.2 & 4.4). The area inside the loop is a measure of the hysteresis.

The hysteretic properties of natural materials of different composition vary widely (e.g. Cisowski 1980). For example magnetite produces tall, thin hysteresis loops (e.g. loop b in Fig. 4.4a), whereas haematite produces flat, fat loops (e.g. loop c in Fig.

4.4b). Variation in grain size also plays an equally important rôle in determining the shape of a crystal's hysteresis loop. Multidomain and superparamagnetic grains produce much thinner loops than single-domain grains. Paramagnetic and diamagnetic minerals do not exhibit hysteresis loops; instead they simply show a reversible linear relationship between magnetisation and applied field.

Cycling the magnetic field to values well below saturation produces quite different shaped hysteresis loops from those of saturation. Such minor hysteresis loops characteristically exhibit elliptical shapes. Laboratory hysteresis measurements of rhombohedral ferromagnets (e.g. haematite) tend to be limited to minor loops (loop c in Fig. 4.4b) as haematite saturation fields cannot be produced with conventional equipment. Most natural ferrimagnets saturate in lower fields, well within the capabilities of laboratory magnets so that in practice we observe their saturation or major hysteresis loops.

The single-domain loop of Figure 4.4a has been drawn for an assemblage of grains. An individual single-domain grain has a quite differently shaped hysteresis loop from that of the average loop of Figure 4.4a. As shown in Figure 2.8 the hysteresis loop of an individual grain depends on the orientation of the grain with respect to the applied field. Calculations of compound magnetisation curves for collections of prolate ellipsoids orientated at random (Stoner & Wohlfarth 1948) readily explain the observed hysteresis loops of natural assemblages of single-domain ferrimagnetic grains. Superparamagnetic ferrimagnetic grains do not display hysteresis but show a steep magnetisation versus field curve which saturates at low fields. The saturation magnetisation of superparamagnetic grains is the same as that of multidomain or single-domain grains.
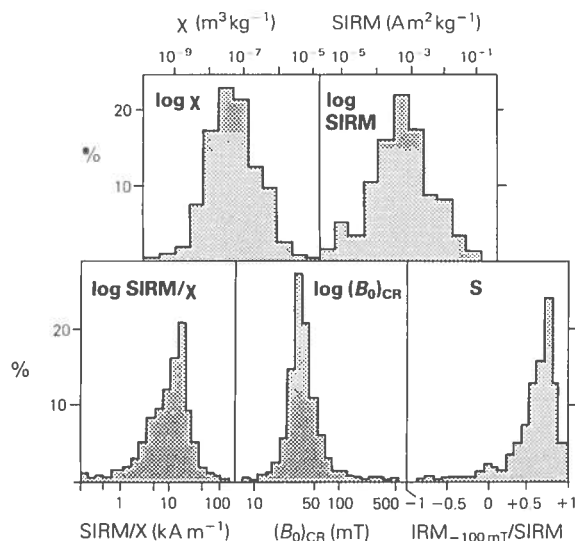
Figure 4.5 Histograms showing the distribution of values of five mineral magnetic parameters in 1000 natural samples. Susceptibility (χ) and saturation remanence (SIRM) are closely log normally distributed.

Investigations of natural materials nearly always have to take into account mineral mixing effects. A number of experimental (Wasilewski 1973) and theoretical studies (Kneller & Luborsky 1963) have been carried out on the systematics of hysteresis properties of mixtures. The most striking effect of extreme mixing is the production of constricted or wasp waisted loops (loop d in Fig. 4.4b). A related effect is that a small addition of a high coercivity material can produce a large increase in coercivity and in coercivity of remanence (Kneller & Luborsky 1963).

4.7 General magnetic properties of natural materials

The main magnetic parameters that are used in later sections for quickly characterising the magnetic mineralogy and granulometry of natural samples are susceptibility, saturation remanence, moderate field ($\sim$ 0.1 T) isothermal or anhysteretic remanences, remanence coercivity and saturation magnetisation. The main characteristics of these magnetic properties are described below and the typical values and ranges of five of those room temperature magnetic parameters are illustrated by measurements on a set of 1000 natural samples selected as being representative

of the wide assortment of rocks, soils, minerals and recent sediments encountered during our environmental studies (Table 4.3).

SUSCEPTIBILITY

As described in Section 4.4, the susceptibility of natural materials is mainly a consequence of their magnetite content so that susceptibility can often be used as a rapid, surrogate measure of magnetite concentration. Susceptibility values of rocks are commonly found to be log normally distributed (Irving *et al.* 1966, Yanak & Uman 1967). Such a distribution can be seen for both susceptibility and saturation remanence (Fig. 4.5) in our 1000 sample collection. The form of the distribution reflects the log normal heavy mineral content distribution of natural materials. Susceptibility ranges from weak negative values ($\sim$ -1 nm^3 kg^{-1}) for diamagnetic rocks such as quartzites to high values ($\sim$ 10 000 nm^3 kg^{-1}) in iron
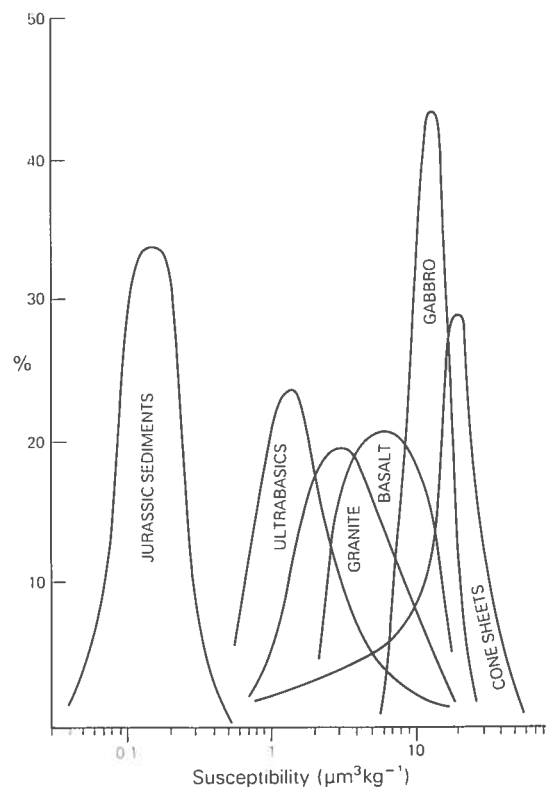

Figure 4.6 Variability of susceptibility for various rock types illustrated using 288 *in situ* measurements made on the Isle of Skye, Scotland.

28

Table 4.4 Average hysteresis properties of 1000 natural samples.

Parameter	Median value	Range	
		minimum	maximum
χ	43 nm^{3}kg^{-1}	-1	10000
SIRM	0.58 mAm^{2}kg^{-1}	0.002	100
SIRM/χ	13 kAm^{-1}	0.11	320
$(B_0)_{CR}$	40 mT	8	570
IRM$_{-100\,mT}$/SIRM	0.7	-0.92	1.00

ores (Table 4.4). An idea of the variation of susceptibility within various rock types is given in Figure 4.6 by the results of a survey of *in situ* susceptibility measurements on six rock types on the Isle of Skye.

ARTIFICIALLY IMPARTED REMANENCES

The simplest artificial remanence to grow is iso-thermal, room temperature, remanent magnetisation. Some workers prefer to use anhysteretic rather than isothermal remanences (Banerjee *et al.* 1981), although the range of magnetising fields available tends to be more restricted. In detail, anhysteretic remanences are more difficult to interpret than isothermal remanences (Dunlop 1983a) as they display somewhat more complicated variations with grain size and they are reduced more by grain interactions (Jaep, 1971). The most useful single artificially imparted remanence is probably the saturation isothermal remanence (Cisowski 1980), which although mainly a measure of magnetite content, is also dependent on grain size and can be strongly influenced by other magnetic minerals such as haematite.

Figure 4.7a illustrates the variation of saturation remanence with grain size for magnetite as deduced
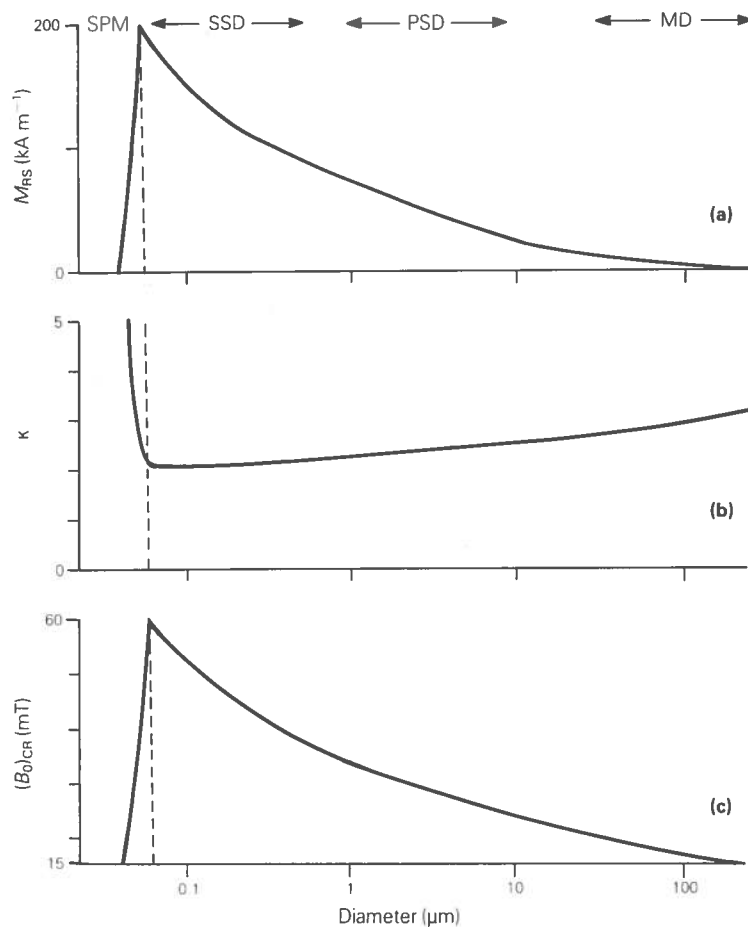

Figure 4.7 Summary diagram of the variation of (a) saturation remanence, (b) susceptibility and (c) coercivity of remanence with magnetite grain size. The dashed lines highlight the changes in trend of mineral magnetic properties at the superparamagnetic/stable single-domain boundary.

from experiments on dispersed magnetite powders (Parry 1965). Notice how saturation remanence is much higher in single-domain grains of around 0.1 μm diameter than in large multidomain grains and how the range of variation through the multidomain and stable single-domain regions is much greater for saturation remanence than for susceptibility (Figs 4.7a & b). Saturation remanence values range from virtually zero, in materials devoid of iron or manganese minerals, to over 100 mA m² kg⁻¹ in finely disseminated iron oxide ores or heavy mineral bands

(Table 4.2). The histogram of saturation remanence values of our 1000 samples (Fig. 4.5) has a median value of 0.58 mA m² kg⁻¹ and the usual log normal distribution of a magnetic parameter that mainly depends on concentration.

SATURATION REMANENCE TO SUSCEPTIBILITY RATIO

The dependence of the two parameters saturation remanence and susceptibility upon each other is displayed in the scattergram of Figure 4.8. The clear
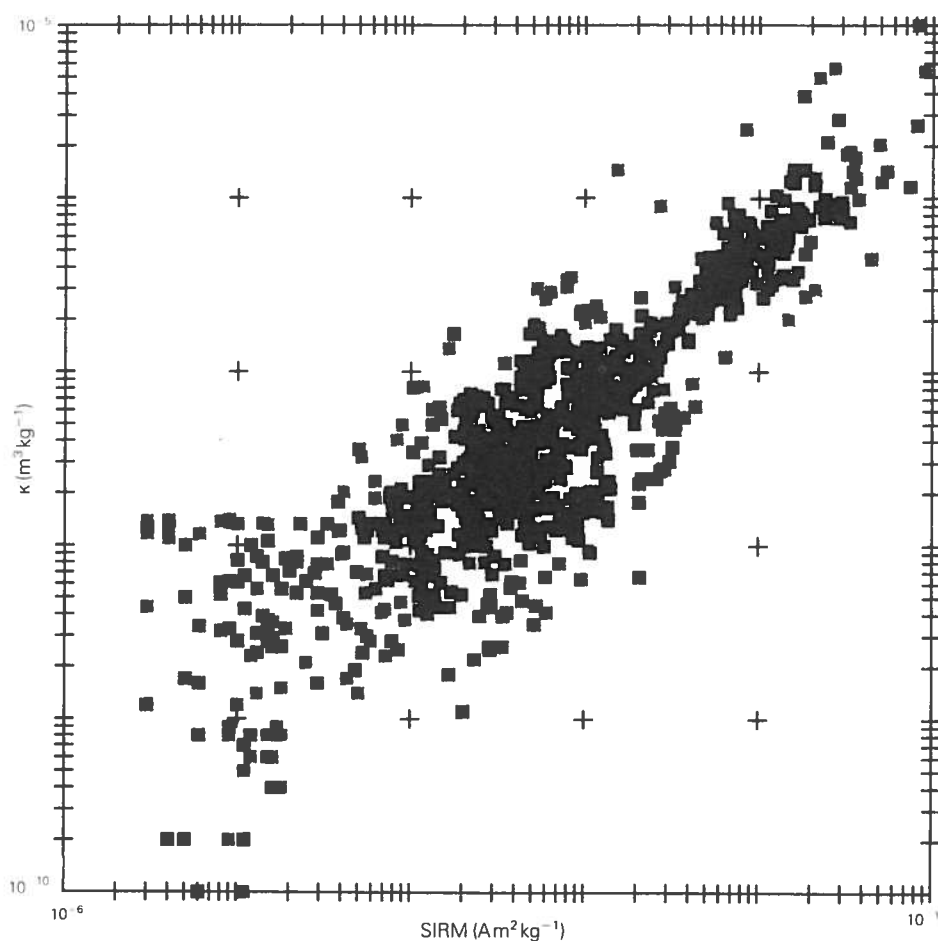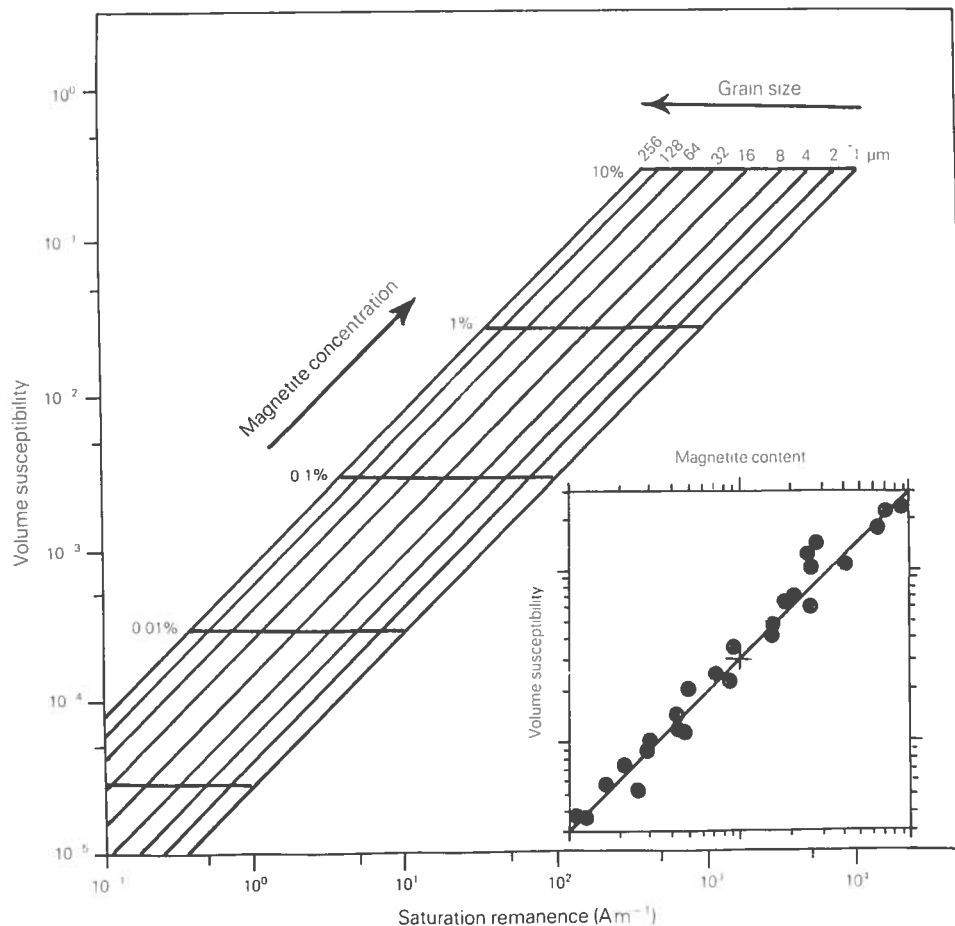

Figure 4.8 Bilogarithmic saturation remanence versus susceptibility plot for 1000 natural samples. Squares denote results from individual specimens. High magnetite concentrations plot towards the top right and low concentrations towards the bottom left corners of the diagram. Specimens lying below and to the right of the main group tend to have high haematite to magnetite ratios. Low SIRM, but moderate κ specimens, plotting at the far left have relatively high susceptibilities on account of paramagnetic contributions. Multidomain magnetites and assemblages with high concentrations of superparamagnetic magnetite plot on the upper side of the main group.

30

Figure 4.9 Bilogarithmic saturation remanence versus susceptibility plot. The concentration/grain size grid is for pure magnetite. Inset illustrates the linear relationship commonly observed, in sets of natural samples, between susceptibility and magnetite concentration.

grouping along the diagonal corresponding to a SIRM/κ ratio of 10 kAm^{-1} in Figure 4.8 results from magnetite having an average effective grain size of 5 μm in natural samples, and from both susceptibility and saturation remanence dominantly reflecting magnetite concentration. The Pearson correlation coefficient of this correlation between the logarithms of susceptibility and saturation remanence for the 1000 natural samples is 0.86. Samples with unusually high haematite to magnetite ratios plot to the right of the general trends as they have SIRM/κ ratios of around 1000 kAm^{-1}.

The ratio of saturation remanence to susceptibility can be used as a rough estimate of magnetic grain sizes in crystals larger than a few hundred angstroms. The grid, based on the data of Parry (1965) and

Dankers (1978) and unpublished results of the authors, on the bilogarithmic plot of Figure 4.9 illustrates how the concentration and grain size of magnetite crystals in a rock can be estimated from measurements of susceptibility and saturation remanence. The magnitude of the magnetic measurements gives the concentration and the ratio gives the grain size.

An important factor in many soils and soil-derived sediments is the presence of ultrafine iron oxides in the superparamagnetic size range (Mullins 1977 and Ch. 8). Superparamagnetic particles have distinctively high susceptibilities and low SIRM/κ ratios. Mixtures of superparamagnetic and single-domain grains thus have similar SIRM/κ ratios to those of multidomain grains and plot in the larger grain size region of Figure
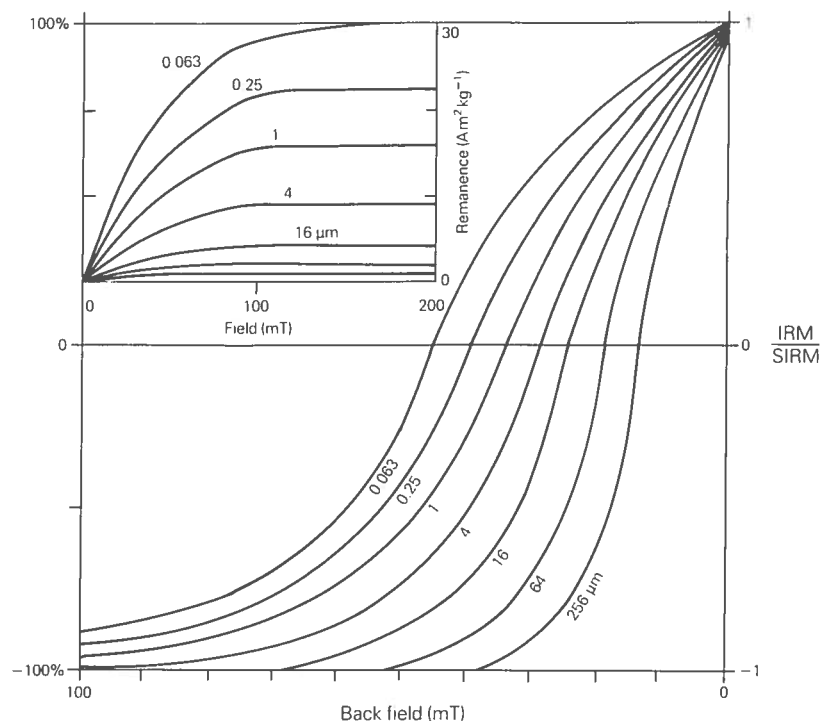
31

Figure 4.10 Coercivity plots of isothermal remanence versus back applied field for magnetite grains ranging in diameter from 0.063 to 256 µm. The back coercivity curves are normalised with respect to saturation remanence SIRM in order to display the variation in stability with grain size. Inset shows un-normalised acquisition of isothermal remanence curves for the same range of magnetite grain sizes, emphasising the dependence of saturation remanence on grain size.

4.9. Further hysteresis measurements, as described below, are needed to distinguish between large multi-domain magnetite grains and samples containing mixtures which include very small superparamagnetic grains. The increase in scatter in the lower left-hand part of Figure 4.8 reflects the importance of superparamagnetic and paramagnetic effects in samples with very low magnetite concentrations.

REMANENCE COERCIVITY

Coercivity of remanence $(B_0)_{CR}$, is a relatively straightforward, very useful hysteresis parameter which can be used in determining magnetic mineralogy and grain size and in helping to characterise magnetic mixtures. It is the field which reduces the saturation isothermal remanence to zero (Fig. 2.2). For crystals of one mineralogy coercivity of remanence is a sensitive indicator of grain size, varying for magnetite from less than 10 mT for multi-domain grains to almost 100 mT for small elongated grains (Figs 4.7c & 4.10).

Examples of remanence hysteresis curves for basalts in various states of oxidation are shown in Figure 4.11. The basalts range from low oxidation states (type I) containing large homogeneous titano-magnetites with remanence coercivity of around 10 mT, through moderately oxidised lavas bearing exsolved spinels containing small magnetite crystals (type III) with higher coercivities of about 40 mT, to highly oxidised lavas in which the main magnetic crystals are haematites (type VI) characterised by very high remanence coercivities of over 300 mT.

REMANENCE RATIOS

The ratios of remanences produced in different laboratory fields can also be profitably used for characterising samples. One ratio that we have found particularly helpful is the ratio of a moderate (100mT) back field isothermal remanence to the saturation remanence. This ratio (often referred to as S) has been found to be expedient for recognising samples with unusual haematite to magnetite ratios (Stober & Thompson 1979). The basic idea behind the S ratio is that in practice the magnetisation of most ferrimagnets will have saturated in fields below 0.1 T and that the major differences in high field remanence (i.e. $SIRM - IRM_{100mT}$) will be due to the imperfect antiferromagnets such as haematite and goethite.

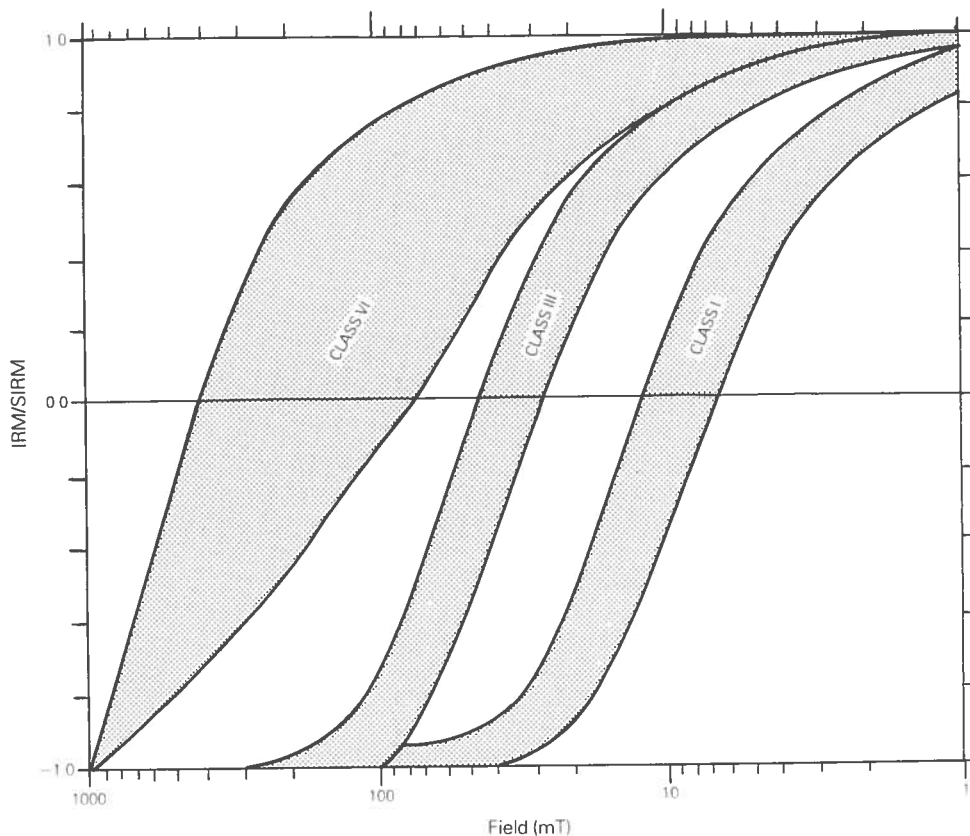Some instruments (e.g. most portable discharge induction coils (Section 6.6.2)) do not produce fields

Figure 4.11 Envelopes of coercivity plots for basalt samples of low (I), medium (III) and high (VI) oxidation states.

as high as 1 T and so isothermal remanence ratios similar to S but employing fields other than 1 T and 0.1 T have been used in later chapters.

The ratio of saturation remanence to anhysteretic remanence, SIRM/ARM, has been found to be particularly useful for detecting grains slightly larger than superparamagnetic in size on account of their low SIRM/ARM ratios. Gillingham and Stacey (1971) and Kneller (1980) have discussed theoretical relationships between SIRM and ARM for multi-domain and single-domain grains respectively.

SATURATION MAGNETISATION

Saturation magnetisation is unaffected by grain size, pure magnetite and haematite having saturation magnetisations of 480 kA m^{-1} and 2.5 kA m^{-1} respectively (Table 2.2). Saturation magnetisation is thus theoretically a most attractive mineral magnetic parameter. Unfortunately absolute measurements of the saturation magnetisation of the low magnetic mineral concentrations with which we have largely been concerned have not proved straightforward and so examples of absolute measures of saturation magnetisation have tended to be restricted to strongly magnetic igneous rocks (e.g. Nagata 1953) or synthetic minerals (e.g. Parry 1965, Dunlop 1973a). Saturation magnetisation can however be usefully used in ratio with saturation remanence or susceptibility (Fuller 1974).

MAGNETISATION AND COERCIVITY RATIOS

Two ratios that are frequently employed to characterise magnetic materials are the magnetisation ratio and the coercivity ratio. Both can be obtained from instruments that plot out hysteresis loops (Sections 6.4.3 & 6.4.4).

The magnetisation ratio of saturation remanence to saturation magnetisation, M_{RS}/M_S, is a sensitive

33

indicator of magnetisation state. Theoretical computations show that ideally M_{RS}/M_S is exactly 0.5 for an assemblage of randomly orientated non-interacting uniaxial magnetic monodomain grains (Stoner & Wohlfarth 1948). In contrast the magnetisation ratio M_{RS}/M_S is less than 0.1 for multidomain grains and even lower for superparamagnetic grains (Bean & Livingston 1959).

The coercivity ratio $(B_0)_{CR}/(B_0)_C$ relates the coercivity of remanence to the saturation coercivity. The lower limit of this coercivity ratio is 1.0. Stoner and Wohlfarth's random assemblage model predicts a coercivity ratio of 1.09 for uniaxial single-domain grains no matter what value their demagnetising co-efficients take. Multidomain grains are expected to have coercivity ratios in excess of 4.0, while super-paramagnetic grains will have ratios in excess of 10.0 (Wasilewski 1973). Rhombohedral ferromagnetic components can have ratios of 3.0 or more.

Magnetic materials can be neatly classified by determining their magnetisation and coercivity ratios (Day *et al.* 1977). Figure 4.12 illustrates how single-domain, pseudo-single-domain and multidomain grains can be recognised through their magnetisation and coercivity ratios. Indeed such hysteresis ratios
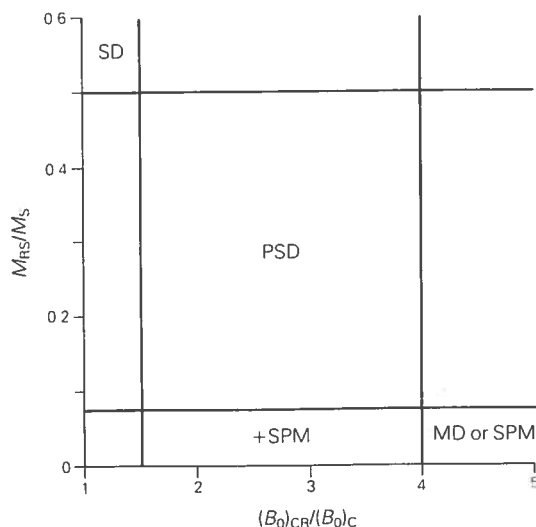
and parameters have been widely used in the magnetic tape, permanent magnet and transformer core manu-facturing industries for product grading.

A further magnetic ratio, that of SIRM/κ to $(B_0)_{CR}$, can be helpful in distinguishing mixtures containing superparamagnetic grains (Bradshaw & Thompson 1985). The reason for this is that superparamagnetic grains contribute to susceptibility, but not to satura-tion remanence and furthermore they do not affect coercivity of remanence. Consequently, samples with a large superparamagnetic component have unusually high susceptibilities which lead to low SIRM/κ to $(B_0)_{CR}$ ratios. Figure 4.13 sketches the main types of magnetic mixtures found for a wide range of SIRM/κ and $(B_0)_{CR}$ values.

It will be apparent that hysteresis parameters can be combined in a large number of ways and that specific situations or special investigations may justify particular combinations. The hysteresis parameters and ratios used in classifying natural samples will to some extent have to depend on the availability of local equipment. In later chapters we have endeavoured to use ratios that can be measured on simple, readily available instruments and that appear to differentiate assemblages of soils and sediment samples very sensitively although they may permit no more than rather general qualitative estimates of the basis of the mineral magnetic varia-tion. In the final prospects chapter, an approach to quantifying the interpretation of general mineral magnetics is discussed.

4.8 Temperature dependence of magnetic properties

Competition between magnetic ordering energy and thermal randomising energy leads to the basic thermomagnetic properties observed in ferromagnets.

SATURATION MAGNETISATION
The thermal change of spontaneous magnetisation is theoretically a fundamental magnetic property since it depends only on a ferromagnet's composition and crystal structure. It is independent of variables such as crystal size, shape and internal stress. Consequently M_S–T curves have been widely used for mineral identification in rock magnetic studies. Spontaneous magnetisation, which at most temperatures can be distinguished from saturation magnetisation only with difficulty, has its largest value at absolute zero
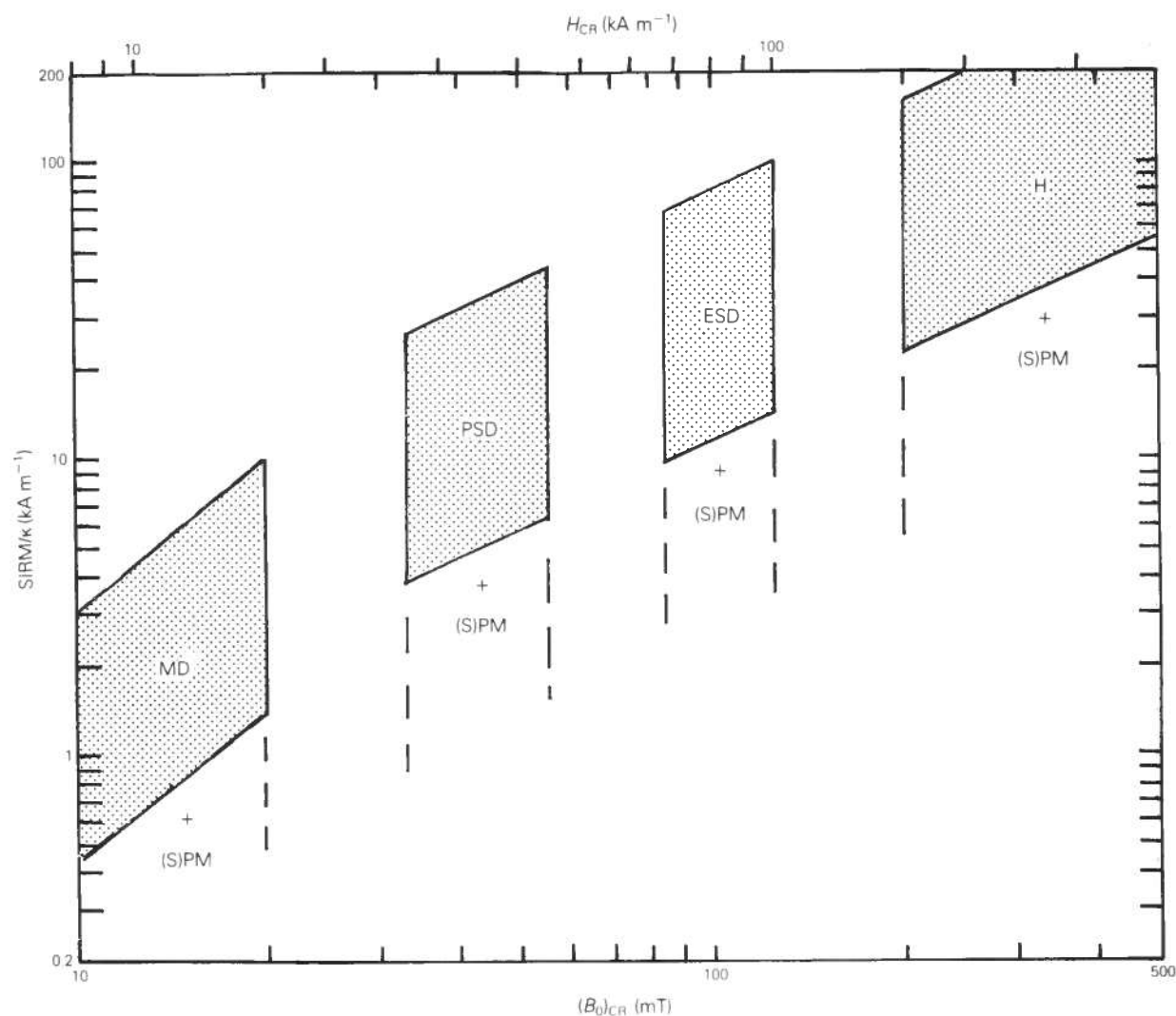

Figure 4.12 Classification of magnetic minerals in terms of magnetisation and coercivity ratios. Single-domain (SD) grains plot in the upper left corner, multidomain (MD) in the lower right while pseudo-single-domain (PSD) lie in between. Superparamagnetic grains (SPM) plot in the lower part of the diagram (after Day *et al.* 1977).

34

Figure 4.13 Magnetic mixture SIRM/χ versus $(B_0)_{CR}$ diagram. The grid schematically divides the diagram into magnetic mineralogies and magnetisation states. Multidomain (MD), pseudo single-domain (PSD) and elongated single-domain (ESD) magnetites fall in the upper left to centre of the diagram. Haematite (H) lies in the upper right corner; mixtures containing (super)paramagnetic grains lie further towards the lower right.
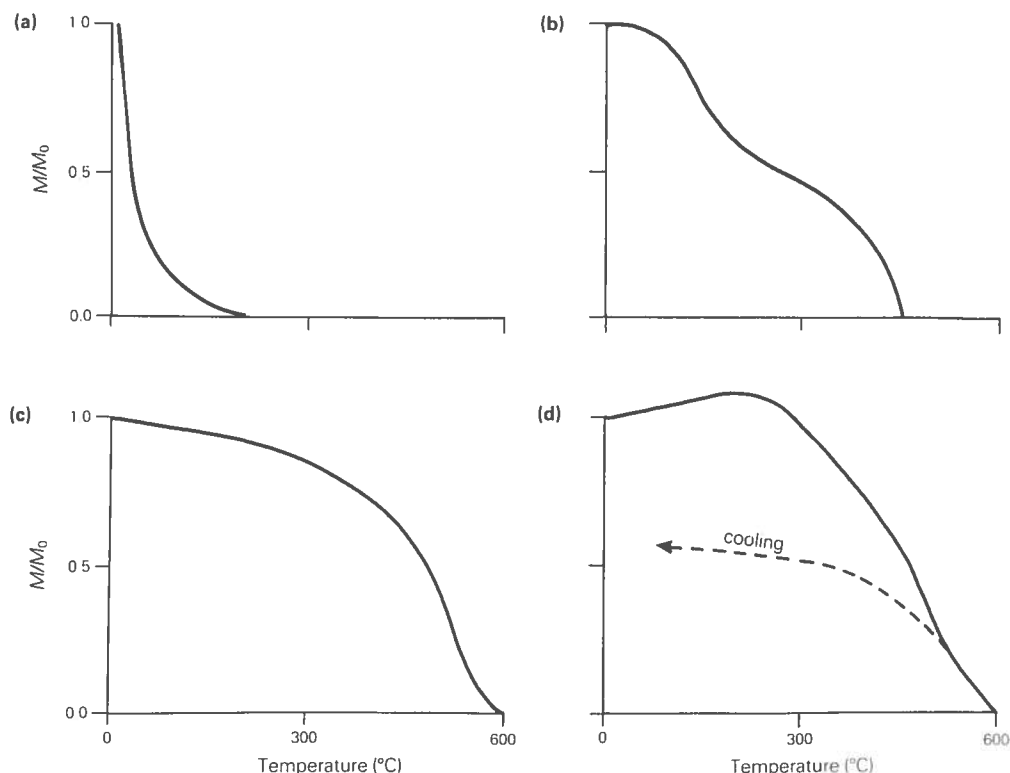
Figure 4.14 Strong field thermomagnetic curves for basalts with spinel minerals in various states of oxidation and alteration. (a) Fresh homogeneous low Curie temperature titanomagnetites; (b) exsolving titanomagnetites with two magnetic minerals; (c) magnetites produced by deuteric oxidation; (d) maghaemite in deep-sea basalt.

and falls at an increasing rate with increasing temperature to become zero at the Curie temperature.

Examples of the variation of saturation magnetisation with temperature of basalts containing different magnetic minerals are given in Figure 4.14. Abrahamsen *et al.* (1984) describe several of these types of variation of saturation magnetisation with temperature from one deep drill core through Tertiary basalts on the Faeroe Islands. Equipment for producing such thermomagnetic curves is described in Section 6.4.

SUSCEPTIBILITY

Susceptibility, κ, varies with temperature in a more complicated manner than saturation magnetisation. The reason for this difference in behaviour is that susceptibility is sensitive to various parameters, such as internal stress and crystalline anisotropy, which change with temperature.

Figure 4.15 plots typical changes of susceptibility with temperature for 'magnetite' crystals of various grain sizes and various compositions. All the curves have been normalised so that the value of susceptibility at 0 °C is plotted as unity. The susceptibility of single-domain magnetite tends to vary with temperature in a manner similar to the dashed curve of Figure 4.15 with only small susceptibility changes, except at the Curie temperature. For multidomain magnetites (solid curve) more pronounced susceptibility changes are observed especially on cooling below room temperature. Multidomain titanium-rich magnetites (dotted curve) have characteristic κ–T curves with low Curie temperatures and with a steady fall in susceptibility on cooling. The susceptibility of paramagnetic minerals follows the Curie–Weiss law of $\kappa = C/T$, i.e. that paramagnetic susceptibility is inversely proportional to absolute temperature, C being a constant. Superparamagnetic crystals follow a similar law at temperatures above their blocking temperature when their susceptibility falls at a rate slightly greater than $1/T$. This κ–T behaviour is illustrated by the curve of
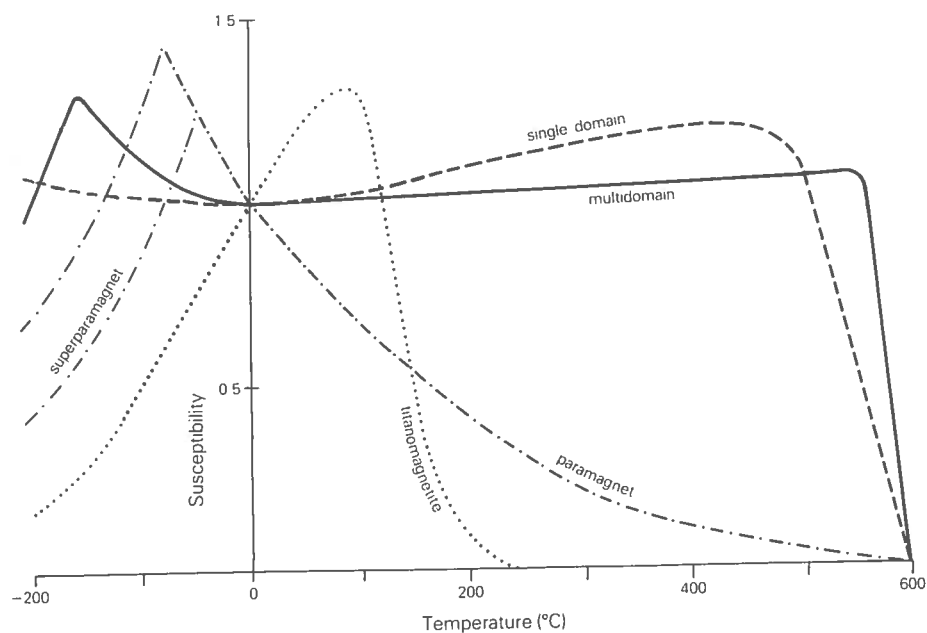
36

Figure 4.15 Low field susceptibility versus temperature curves for various 'magnetite' domain states and compositions. Schematic diagram with curves arbitrarily normalised to equivalent susceptibilities at 0°C. Titanomagnetite (dotted), single-domain magnetite (dashed), multidomain grains (solid), superparamagnetic grains (dashes and dots). (The temperature of the peak susceptibility of superparamagnetic grains depends on their volume.)

dots and dashes in Figure 4.15. On cooling below the blocking temperature the susceptibility of superparamagnets drops sharply as a stable magnetisation is blocked in and the coercivity increases. This blocking effect is diagrammatically illustrated in Figure 4.15 by the additional curves of dots and dashes falling away at low temperatures.

Measurements of susceptibility at low temperature are easy to make (Radhakrishnamurty *et al.* 1978) and can provide valuable additional magnetic information, particularly about the likely relative importance of the contribution of paramagnetic or superparamagnetic crystals to the bulk susceptibility.

REMANENCE

In palaeomagnetic studies of natural remanent magnetisation, blocking temperatures can be of more interest than Curie temperatures as they control long-term magnetic stability at elevated temperature.

Blocking temperatures are determined by heating the remanence in a zero field and observing the temperature at which the remanence is destroyed. A remanence *v.* temperature plot derived from such an experiment is shown in Figure 4.16. The plot compares the thermal demagnetisation results of a natural remanence and a laboratory thermoremanence. We can see that the natural remanence has lost part of its primary magnetisation. The missing remanence (shaded region of Fig. 4.16) is that of the low stability, low blocking temperature grains. The high stability, high blocking temperature magnetisation is seen to be the same in both the natural remanence and the artificial laboratory-produced thermoremanence.

Somewhat surprisingly, magnetic remanence can be destroyed by cooling as well as by heating. The magnetisation of pure haematite crystals undergoes a transition at −10 °C (Section 3.2.2), while magnetite shows a thermomagnetic effect at −150 °C. For both of these minerals these low temperature effects are caused by a change in magnetocrystalline anisotropy (Section 2.4.1). At the transition temperature the anisotropy disappears as it changes sign and there is an associated loss of remanence which can be diagnostic of mineral type.

Irreversible thermomagnetic remanence effects can even be employed in the recognition of magnetic minerals (Dunlop 1972). Maghaemite, for example, is
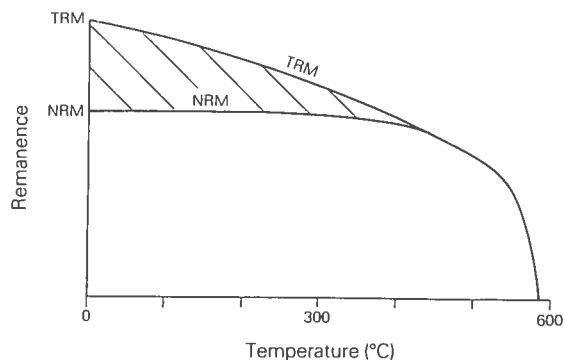
Figure 4.16 Remanence versus temperature demagnetisation plot for a laboratory thermoremanence (TRM) and a natural remanence (NRM) in the same specimen. Low blocking temperature magnetisations have been lost from the NRM.

metastable (Section 3.2.1) and on heating to 300 °C inverts to haematite. Its saturation magnetisation, susceptibility and remanence all show decreases around 300 °C which are not recovered on subsequent cooling. This irreversible behaviour is often used to infer the occurrence of maghaemite in natural samples (Fig. 4.14d). Pyrrhotite (Section 3.3) characteristically shows a distinctive rise in magnetisation on heating to 200 °C, due to a reordering of lattice vacancies, followed by a fall at the Curie temperature of 320 °C.

4.9 Summary

The magnetic properties of natural materials can be used as a petrological tool and as a method of characterising samples. Artificially imparted remanences and magnetisations can be used in order to distinguish between various types of magnetic minerals and their grain sizes. Low field magnetisation, or susceptibility, is a particularly easy property to measure and is a quick first guide to magnetite concentration. Hysteresis and thermomagnetic properties can provide further information about the types of magnetic minerals and magnetic mixtures in natural materials.

Geological materials can acquire natural remanence magnetisations through natural processes. Some such naturally grown remanences are stable and can record information about the ancient magnetic field, with little alteration, for millions of years.

Further reading

General journal paper

Clark 1983. Comments on magnetic petrophysics.

Advanced journal papers

Stacey 1963. The physical theory of rock magnetism.
O'Reilly 1976. Magnetic minerals in the crust of the Earth.

Advanced books

Nagata 1953. *Rock magnetism.*
O'Reilly 1984. *Rock and mineral magnetism.*

[5]

The Earth's magnetic field

Magnus magnes ipse est globus terrestris.

W. Gilbert 1600

5.1 Geomagnetism

Scientists have been measuring the Earth's magnetic field for several centuries. They have built up an increasingly more detailed worldwide picture of the field and its variation with time. Field measurements are now made from ships, aeroplanes and satellites as well as on the ground. For older records of the geomagnetic field and its changes with time we must turn away from historical documents to the study of the fossil magnetism in the Earth's rocks and sediments. The palaeomagnetic signature of the geomagnetic field, recorded by the remanent magnetisation of rocks and sediments, naturally produces a sparser and less detailed picture than the one we can build up for the present field, but the strength of this record is that it extends the time of observation of the field by seven orders of magnitude to cover 3000 million years. As the field originates in the Earth's molten core, the palaeomagnetic record is crucial to the understanding of both the origin and evolution of the geomagnetic field and the Earth's deep interior.

5.1.1 Description of the geomagnetic field

In a rough way the field outside the Earth resembles that outside a uniformly magnetised sphere as illustrated in Figure 5.1. This approximation of the geomagnetic field to that of a simple dipole was first recognised by William Gilbert, physician to Queen Elizabeth I, and published in his treatise *De Magnete* in 1600. Gellibrand in 1635 was the first to demonstrate that the geomagnetic field varied not only regionally but also with time.

The direction and strength of the geomagnetic field can be most simply determined using a magnetic needle. Freely pivoted it will swing into a position parallel to the local lines of force of the geomagnetic field, pointing approximately to the north and, if free to tilt, taking up an inclined position. In the northern hemisphere the magnetised needle, in its equilibrium position will point downwards towards the north, whereas in the southern hemisphere it will point upwards towards the north. The field intensity can be
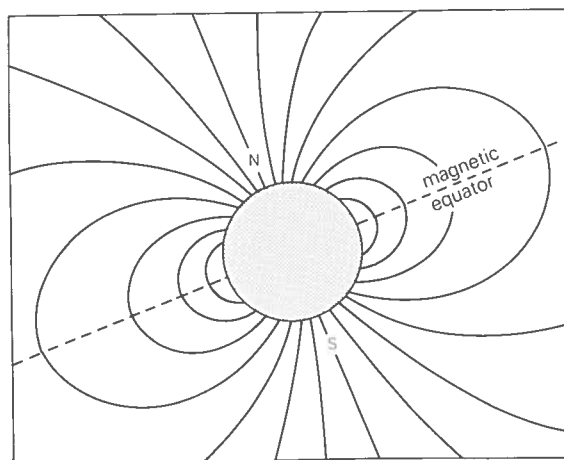

Figure 5.1 The magnetic lines of force of a geocentric dipole field. The Earth's magnetic axis is presently tilted about 10° from its spin axis.

determined by turning the needle through an angle of 90° from its rest position. The torque required to restrain the needle from returning to its equilibrium position is then a measure of the field intensity.

To describe completely the magnetic field, which is a vector, we need three numbers. These numbers may be chosen in different ways (Fig. 5.2). It is generally convenient to take the intensity of the field (F), the angle of dip of the field below the horizontal plane, i.e. the **inclination** (I), and the angle between the horizontal component of the field and the true, geographical north, i.e. the **declination** (D). The total intensity (F), declination (D) and inclination (I) then completely define the magnetic field at any point. Another method of specifying the field is to use the field components along the northward (x), eastward (y) and downward (z) directions (Fig. 5.2). These

three components also completely define the field. These various field components along with the horizontal field intensity (H) are related by the following equations:

$$H = F \cos I \qquad z = F \sin I \qquad \tan I = z/H$$

$$x = H \cos D \qquad y = H \sin D \qquad \tan D = y/x$$

$$F^2 = H^2 + z^2 = x^2 + y^2 + z^2$$

Variations of the declination and inclination of the geo-magnetic field over the surface of the Earth are illustrated in Figure 5.3 for the year 1980. The line along which the inclination is zero in Figure 5.3b is called the magnetic equator. The points where the inclination is +90° and −90° are the north and south magnetic poles respectively. The field intensity varies from roughly 7×10^{-5} T at the magnetic poles to 2.5×10^{-5} T at the magnetic equator.

The dipole field that best fits the actual field of the Earth today has its poles $11\frac{1}{2}°$ from the geographic poles. It is only an approximation accounting for 90% of the Earth's field. In some places the difference between the actual field and that of the best fitting dipolar field reaches 20% of the real field. The intersections of the best fitting dipole axis with the Earth's surface are called the geomagnetic poles. The north and south geomagnetic poles lie at $78\frac{1}{2}°$N, 70°W and $78\frac{1}{2}°$S, 110°E respectively and should not be confused with the magnetic poles in Figure 5.3a. Subtraction of the best fitting **dipole field** from the actual field leaves the non-dipole field. We find that the present magnetic field differs substantially from the dipole field over half a dozen areas a few thousand kilometres across. These anomalous regions show no obvious relationships with geographical or geological features. Indeed the features of the non-dipole field have changed markedly over the last 400 years (Yukutake & Tachinaka 1968). In contrast the dipole has hardly changed its orientation during this period (Barraclough 1974) although its strength has decreased, on average, at a rate of 0.05% per year (Yukutake 1979).

Figure 5.2 Relationship of the magnetic elements. The magnetic field at any point on the Earth's surface can be fully described by three parameters. Geomagnetists often use the orthogonal north (x), east (y) and down (z) components. Palaeomagnetists, involved mainly with angular data, tend to use declination (D), inclination (I) and intensity (F).

5.1.2 Secular variation

The importance of changes in direction of the geo-magnetic field over the time of a few hundred years is illustrated in Figure 5.4 which shows the changes in field direction at London. The first measurement of
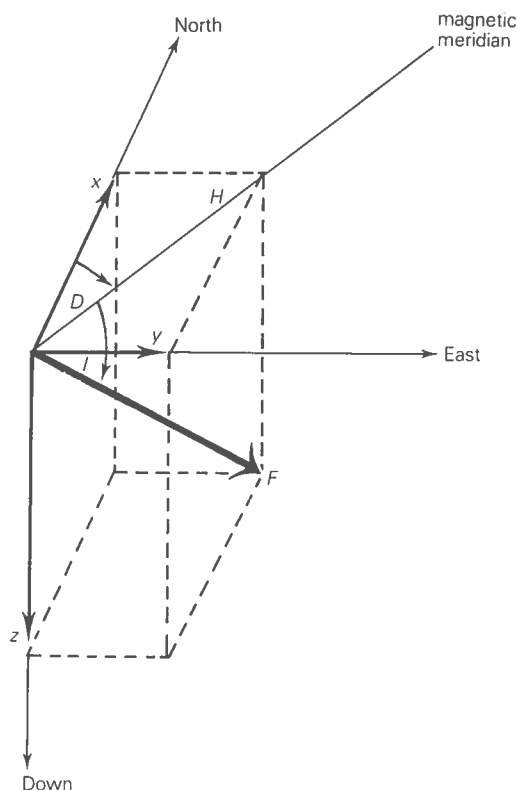
(a) IGRF 1980   Declination (*D*)   (degrees)

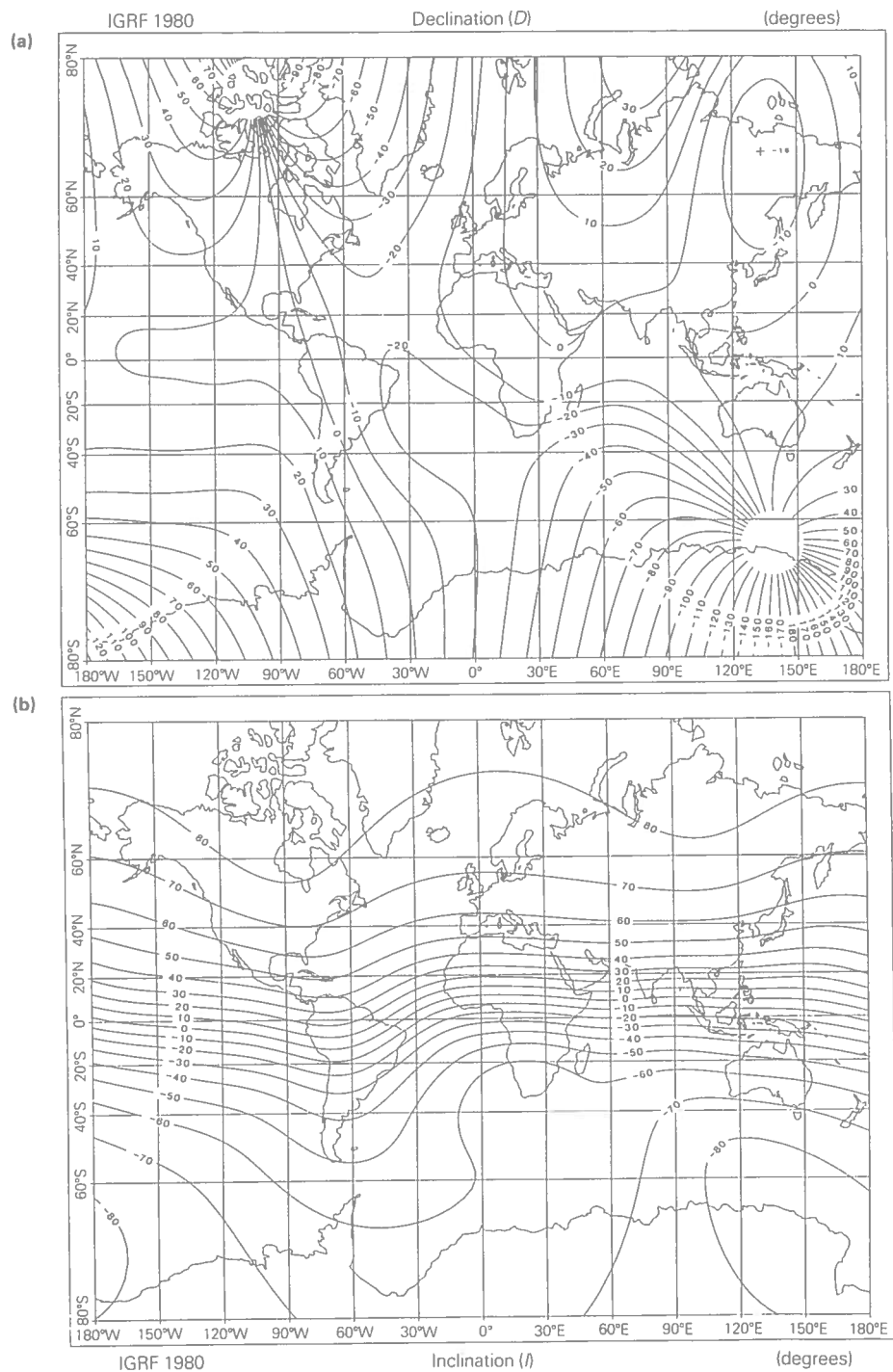**(b)** IGRF 1980   Inclination (*I*)   (degrees)

Figure 5.3 The Earth's magnetic field (1980). (a) The declination (variation) from the geographical meridian (10° intervals). (b) The inclination (dip) from the horizontal (10° intervals). (From Fabiano *et al.* 1983)
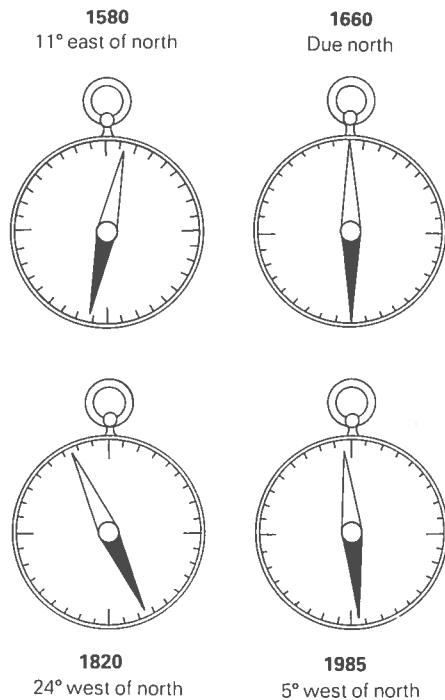
1580
11° east of north

1660
Due north

1820
24° west of north

1985
5° west of north

Figure 5.4 The variation in direction of a compass needle at London since 1580 AD due to slow secular changes of the Earth's magnetic field (after Press & Siever 1974).

the declination of a compass needle at London was made around 1570 AD when it was found that the needle pointed 11° to the east, in 1660 AD it pointed due north and by 1820 AD it had swung round to 24° to the west (Fig. 5.4 and Bauer 1896). Since then the declination has steadily decreased and now, at London, is 5°W and is decreasing at the rate of 9 minutes per year. The inclination of a dip needle at London has also varied with time: in 1700 AD it reached a maximum of over 74° and is now close to 66°. These secular or gradual variations of field direction at London are plotted in Figure 5.5 along with the historical variations measured at Rome and at Boston. Notice how the secular variations at London and Rome, some 1500 km apart, have been relatively similar whereas the variations at Boston, some 5000 km from Europe, have been very different. Geomagnetic field intensities as well as directions have been found to have changed appreciably over the past few hundred years. For example, at Cape Town the horizontal component of the field decreased by 45% in

140 years. The worldwide average (root mean square) secular intensity change amounts to some 50 nT per year.

Sudden worldwide changes in the rate of secular variation, termed jerks or impulses (Courtillot *et al.* 1978, Malin & Hodder 1982) occur at intervals. During this century globally synchros jerks occurred in 1912–13 and 1969–70, each had a duration of around one year.

World **secular variation** maps of the rate of change of components of the geomagnetic field show a general resemblance in character to maps of the non-dipole field and it is generally felt that the secular variation reflects changes which are taking place in the non-dipole field. Halley in 1692 noticed that a large proportion of the features of the secular variation could be explained by a westward movement. This general westward drift averages about 0.2° per year, for non-dipole features, and in places has been as high as 0.6° per year. Both the non-dipole foci and the secular variation foci tend to move westwards. In addition these two kinds of foci form, deform and decay, rather like eddies in a stream of water, implying that the non-dipole field is temporary, always changing its features over lifetimes of a few hundred years (Fig. 5.6). Secular variation then is a complex regional phenomenon with some local centres drifting rapidly, some apparently remaining stationary, some being short lived and others being more persistent (Thompson 1984).

5.1.3. Origin of the geomagnetic field

Gauss (1839) first developed the mathematical method of analysing the geomagnetic field in terms of a potential and representing it as an infinite series of spherical harmonic functions. He used the method with observations of the geomagnetic field made on the surface of the Earth and was able to demonstrate that the field was predominantly of internal origin. We now know that there is a very small net external contribution, of about 0.1%, to the geomagnetic field at the Earth's surface. The external contributions which rise somewhat during magnetic storms are due to currents circulating in the upper atmosphere. They tend to vary rapidly with periods ranging from a few seconds to years. Our main concern in this and later chapters about the geomagnetic field is with the Earth's internally generated field.
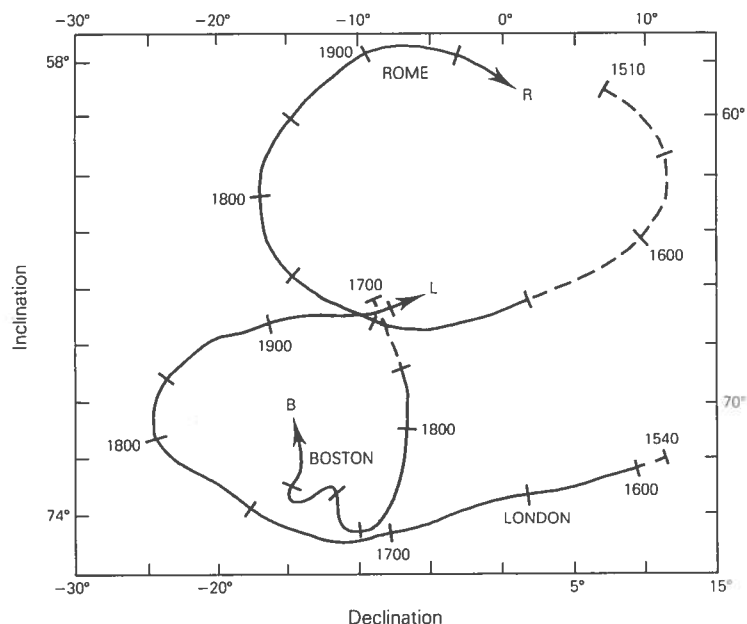
The geomagnetic field pattern bears little relation

Figure 5.5 Secular changes in magnetic declination and inclination as observed in London, Rome and Boston (solid lines with tick marks every 50 years). The dashed curves begin at the time of the first declination measurements at each locality. The early inclination changes are based on archaeomagnetic data. Secular direction changes largely result from fluctuations in the Earth's non-dipole field which vary from locality to locality.

to geological features and varies much more rapidly than geological processes of uplift or continental drift and so we must look beneath the Earth's crust and mantle to motions in the fluid core of the Earth for the origin of the geomagnetic field. The most promising theory for the origin of the Earth's field is the dynamo theory which was developed by Elsasser (1946) and Bullard (1948) in the 1940s and 1950s.

The basic idea of the dynamo theory is that the Earth's magnetic field arises in the Earth's dense metallic liquid core, owing to the circulation of electric currents (Hide & Roberts 1956). A system of currents once generated in the Earth's core could be expected to run for some time because of the effects of self-induction. The electric energy would be gradually dissipated through Joule heating and after about 10^5 years any current system would have virtually disappeared. We know from palaeomagnetic studies that the Earth has had a magnetic field comparable in form and strength to that of the present field for at least 3×10^9 years. As this length of time is much longer than the 10^5 year free decay time there must be a mechanism controlling and maintaining the electric currents in the Earth's core. The most plausible mechanism is the interaction of magnetic fields with the flow of electric currents arising from the fluid motion of the Earth's core, i.e. a self-exciting dynamo. Theoretical calculations suggest that in the Earth's

core, fluid motions across an existing magnetic field will produce their own magnetic fields and induced electric currents and can be very important hydrodynamically. Fluid motions could thus reinforce and maintain the geomagnetic field through a self-exciting dynamo mechanism (Elsasser 1946, Bullard & Gellman 1954). Although the basic mathematical formulation of such motions and interactions (which form the subject of magnetohydrodynamics) involves no more than the laws of classical mechanics, thermodynamics and electromagnetism, the subject is extremely complex because the mathematical problems are mostly non-linear and there are many physical unknowns and few observational constraints. So only a few solutions have been obtained to the mathematical problems and these relate to particularly simple solutions, but the general consensus is that any sufficiently complicated and vigorous motion in the Earth's core will operate as a dynamo. The primary energy source of the geomagnetic dynamo is thought to be either the radioactive decay of elements in the Earth's core (e.g. Verhoogen 1973) or the gravitational energy released by the sinking of heavy material in the outer core (Braginsky 1963). The resulting thermal or compositional instabilities lead to the formation of convection currents which in turn, through their magnetohydrodynamic actions, drive the dynamo (Lowes 1984).

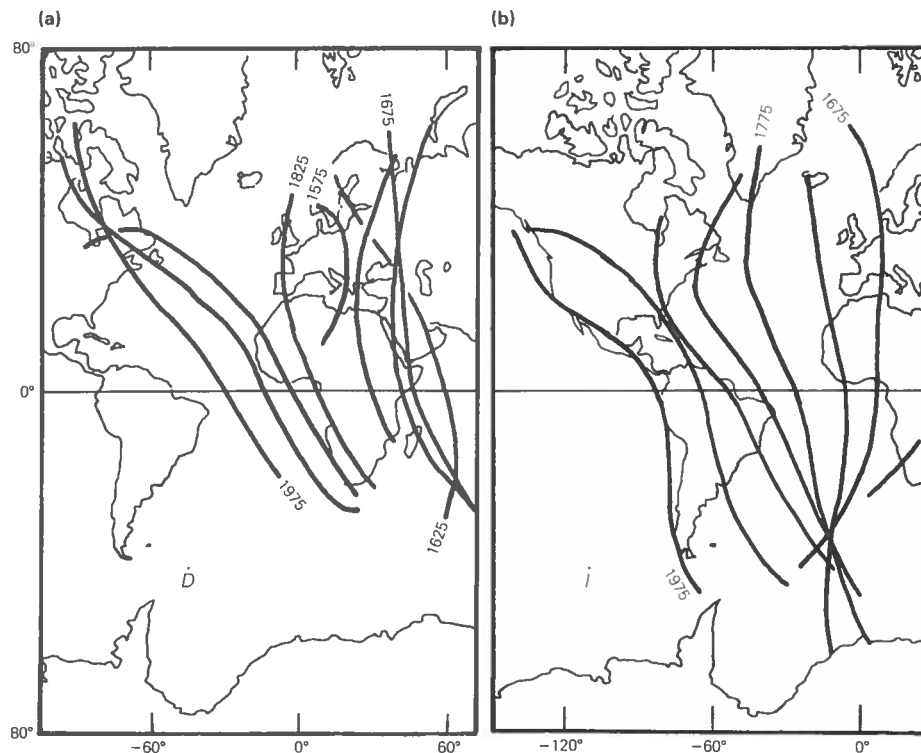The fluid motions of the Earth's core may also

43

Figure 5.6 Examples of geomagnetic change between 1575 and 1975 AD illustrated by the locations of the (a) $D = 0$ and (b) $I = 0$ isopores (i.e. the locations where declination and inclination are passing through maxima or minima). The zero secular change isopores, in common with many features of the geomagnetic field, change position with time. For example, in central Europe in 1575 AD declination is reaching an easterly maximum (Fig. 5.5). The position of the easterly maximum moves slowly eastwards with time. By 1675 AD inclination in central Europe passes through a maximum (Fig. 5.5), the position of the maximum drifting steadily westwards to cross the Atlantic by 1825 AD. Around 1800 AD a westerly declination maximum grows up and crosses central Europe and Africa as marked by the successive positions of the Zero declination change isopore.

account for the temporal fluctuations in the Earth's magnetic field in addition to the overall origin of the field. The maintenance of the geomagnetic field involves a range of electromagnetic interactions, feedback systems and transfers of kinetic, magnetic and electrical energies. So eddying or turbulent motion would result in complicated and varying disturbances of the magnetic field and furthermore local irregularities in fluid flow or current systems near the surface of the core could be expected to produce magnetic disturbances on the scale of the non-dipole field. The magnetic field of a shallow electric current loop seated in the top layer of the core would resemble that of a radial magnetic dipole buried in the core. The non-dipole field has often been modelled in terms of radial dipoles rather than in terms of spherical harmonic components. The number, location and magnitude of the dipoles used in the models vary considerably as no unique model can be derived from the observational data on the Earth's surface. Models with as few as eight radial dipoles can be constructed which produce a magnetic pattern at the Earth's surface which closely resembles that of the geomagnetic field. A further problem in constructing dipole models is that the amount of magnetic shielding or screening in the conducting fluid of the Earth's core is difficult to estimate. The modern geomagnetic approach to the question of secular magnetic variations is in terms of the dynamics of the redistribution of flux lines crossing the core/mantle boundary.

5.2 Palaeomagnetism

The history of the geomagnetic field can be extended into the geological past through the study of fossil magnetisation since natural materials can acquire a remanence (fossil magnetisation) in the Earth's magnetic field and so record its ancient direction and intensity (Section 4.3.1.). Many natural materials are able to retain such a signature of the geomagnetic field through later field changes or geological events. If we can find such materials, reconstruct their palaeo-orientation, measure their remanent magnetisation and date the time of origin of the magnetisation, we have a measure of the geomagnetic field in the past. Palaeomagnetists have assembled a remarkable picture of the ancient geomagnetic field using such fossil remanent magnetisations from many parts of the world and for ages ranging back to those of the very oldest rocks found on Earth.

The youngest materials studied, such as archaeological pottery shards or lake sediments, produce a palaeomagnetic picture which ties in well with the historically documented observations of the geomagnetic field. Palaeomagnetic results from rocks and sediments a few million years old show that the ancient field, although of similar form and intensity to the present field, was often completely reversed. This surprising phenomenon has been particularly well documented. It can be shown that the field has **reversed polarity** many hundreds of times and that the field has been in a 'reversed' state, when a compass needle anywhere on the Earth's surface would have pointed south, as often as it has been in the present, 'normal' state. Even older rocks with ages of tens and hundreds of million years reveal a further palaeomagnetic feature called polar wander. The palaeomagnetic pole positions of such rocks, in which the effects of secular variation have been averaged out, do not coincide with the present geographic pole, but are found to lie on a path which gradually leads away from the present pole as older and older rocks are considered. Such paths are called apparent polar wander paths as the movement of the palaeomagnetic **pole position** reflects the drift of continental blocks rather than the movement of the geomagnetic pole.

Palaeomagnetic phenomena, in addition to extending our knowledge of the geomagnetic field, can have important stratigraphic, tectonic and climatic implications. Interpretation of palaeomagnetic records can however be problematical and subjective although the collection and measurement of palaeo-magnetic samples is generally straightforward. Many old rocks have undergone complex geological histories so that their original magnetic records may have been modified by later magnetisations and geological events. In consequence, a number of geological and laboratory tests must be applied in order to extract useful geomagnetic information. The equipment used in such palaeomagnetic measurements and testing is described in Chapter 6. The palaeomagnetic method is examined in more detail in Chapters 13 and 14, along with some applications and implications of palaeomagnetic results. The remainder of this chapter summarises the main geomagnetic information which has been won from studies of fossil magnetism, starting with results from recent materials and then proceeding deeper into the geological past.

5.2.1 Archaeomagnetism

Archaeomagnetism is the study of the magnetic remanence of archaeological artifacts and structures. Most archaeomagnetic results have been derived from baked materials such as pottery or tiles, which acquired a thermoremanent magnetisation on cooling. Such baked materials may also be used to find the
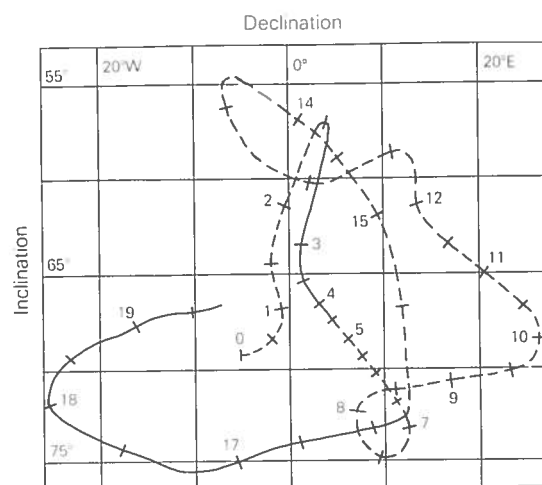

Figure 5.7 Archaeomagnetic variation in declination and inclination at London since 0 AD. The dashed line shows the secular direction changes deduced from the thermo-remanence of bricks, tiles and pottery. The solid curve shows the direct historically documented variation of Fig. 5.5. Although the motion of the magnetic field vector is complex, showing both clockwise and anticlockwise looping, the average direction is close to that which would be produced by a geocentric axial dipole (data after Aitken 1974, Clark 1980).

45

ancient intensity of the geomagnetic field. Figure 5.7 illustrates the way in which archaeomagnetic records from France (Thellier 1981) and Britain (Aitken 1974, 1978) have extended the historical pattern of field direction changes back to the times of the Roman empire. Historical records of declination and inclination generally performed a clockwise loop during the period of direct observations (Fig. 5.5), a motion that is attributed to westward drift of the non-dipole field (Runcorn 1955). The clockwise sense of motion is also seen in the European archaeomagnetic record from 1300 AD to the present day (Fig. 5.7), but at earlier times an anticlockwise motion is found suggesting that in the past the non-dipole field has at times drifted eastwards (Aitken 1974, Kovacheva 1982).

5.2.2   Palaeosecular variation

Variations of the geomagnetic field, with characteristic timescales of 100 or 1000 years, are found in the palaeomagnetic records of rapidly deposited sediments. Organic-rich muds at the bottom of lakes have been found to carry continuous records of magnetic field changes spanning tens of thousands of years (Mackereth 1971). Figure 5.8 shows the last 5000 years of the palaeomagnetic record recovered from the bed of Loch Lomond. The familiar declination and inclination features of the historical and archaeomagnetic records can be seen in the uppermost sediments along with magnetic variations of similar amplitude and character in the lower sediments. The clockwise looping of the historical declination and inclination vector and the eastward looping of the archaeomagnetic vector are preserved in the uppermost Lomond record. The deeper sediments show that before AD 0 looping was once again mainly clockwise. Figure 5.8 also demonstrates that the secular variation of the geomagnetic field has been following a complex rather than a regular pattern. No simple periodicity is revealed although the characteristic timescale of the variations has a length of between 2000 and 3000 years (the spectral energy peaks at a period of 2700 years). Lake sediment records from other parts of the world show variations with similar characteristic timescales but different declination and inclination patterns. The potential of these limnomagnetic records as a magnetostratigraphic tool is discussed in Chapter 14.

Palaeointensity investigations of lava and very old archaeological samples have been used to estimate the ancient intensity of the geomagnetic field. Figure 5.7 changes in geomagnetic dipole moment over the last 11 000 years (Fig. 5.9). Variations from different parts of the world and rapid local fluctuations have been averaged out to reveal pronounced changes in dipole intensity with durations of order 10^4 years. The dipole field intensity, for instance, varied by a factor of two between a minimum at 6500 years BP and a maximum at 2800 years BP (Fig. 5.9).

5.2.3   Polarity timescale

Palaeomagnetic results show that the ancient geomagnetic field has reversed itself at frequent, but highly irregular, intervals in the geological past. Detailed radiometric dating studies have also shown that the polarity reversals take place simultaneously all around the world (Cox *et al.* 1963, Cox 1969). Averaged over a long sequence of polarity reversals the total amount of time spent by the geomagnetic field in the reversed state is equal to that spent in the normal state. Furthermore, the average field intensity of the two states is also similar. The energy levels of the geomagnetic dynamo in its normal and reverse states thus appears to be very similar.
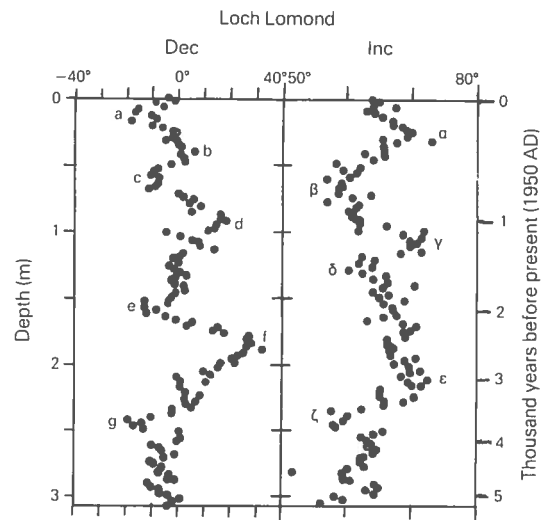

Figure 5.8 Palaeomagnetic directions recorded in the top 3 m of sediment of Loch Lomond. The declination measurements are centred on zero. ^{14}C age determinations provide the timescale at the right. Dots mark the natural remanence directions of individual 10 ml volume palaeomagnetic subsamples. Major palaeomagnetic turning points labelled following the convention of Mackereth (1971) (data from Turner & Thompson 1979).

Figure 5.9 Variation in dipole moment of the Earth's magnetic field during the past 11 000 years as deduced from archaeomagnetic palaeointensity studies. The archaeomagnetic results, with their standard deviations shown by vertical bars, are world averages over 1000 year intervals (500 year intervals after 3000 years BP) (data after McElhinny & Senanayake 1982).
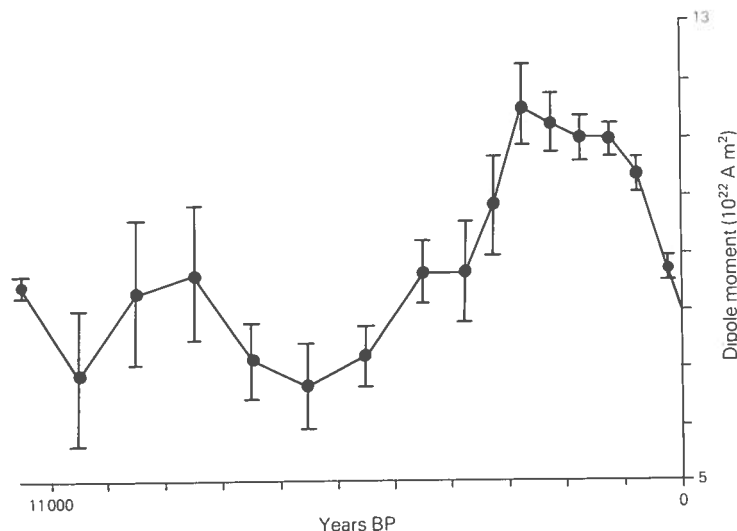
Polarity reversals have been found preserved in the thermoremanence of lava flows, in the detrital remanence of deep-sea sediments and the magnetic anomalies charted over the mid-ocean ridges. All these diverse recordings can be combined into one consistent polarity timescale. Marine anomaly profiles provide a particularly detailed and nearly continuous record of polarity reversals over the past 170 million years as changes of polarity of the geomagnetic field are preserved in the ocean crust, as it spreads away from the mid-oceanic ridge crests as part of the plate tectonic process (Vine & Matthews 1963, Larson & Pitman 1972).

Over the past fifty million years the geomagnetic reversal process appears to have been almost completely random. Characteristic periods associated with individual reversals have been about two hundred thousand years while the time to complete a transition between polarity states is estimated to be between one and ten thousand years. Over 50 million years ago the dynamo process was less symmetrical with respect to polarity and long periods of single polarity are found, the longest being a period of reversed polarity which persisted for 85 million years. Changes in the average frequency of reversals have been found to occur at intervals of 50 million years or longer. Such geomagnetic periods of tens and hundreds of millions of years have a geological ring about them which suggests that their physical origin is to be sought in the Earth's solid mantle rather than the fluid core. Mantle convection currents could alter the physical properties of the core/mantle interface by producing hot or cold spots or even bumps over time-scales of hundreds of millions of years. These slow temperature or topographic changes could affect the operating characteristics of the dynamo and govern the long period timetable of geomagnetic polarity changes, such as the changes in average frequency of geomagnetic reversals.

5.2.4 Apparent polar wandering

When averaged over some hundred thousand years the geomagnetic dipole axis has been found to coincide with the Earth's spin axis. Indeed palaeo-magnetic evidence suggests that throughout the Earth's history the geomagnetic field has been pre-dominantly dipolar and axial. This simple relationship between the geomagnetic axis and the spin axis has allowed palaeomagnetic measurements to be used to make continental reconstructions and to chart the drift of continents over the surface of the Earth (Runcorn 1956, Irving 1977).

The directions of fossil magnetisation in rocks from different parts of the world can be most easily compared by plotting the magnetic results in terms of palaeomagnetic pole positions. Figure 5.10 shows the change in palaeomagnetic pole positions of rocks from Europe and North America over the past 500 million years. The shapes of the two polar wander paths can be explained very well by the drift of the continents associated with formation of the Atlantic ocean.
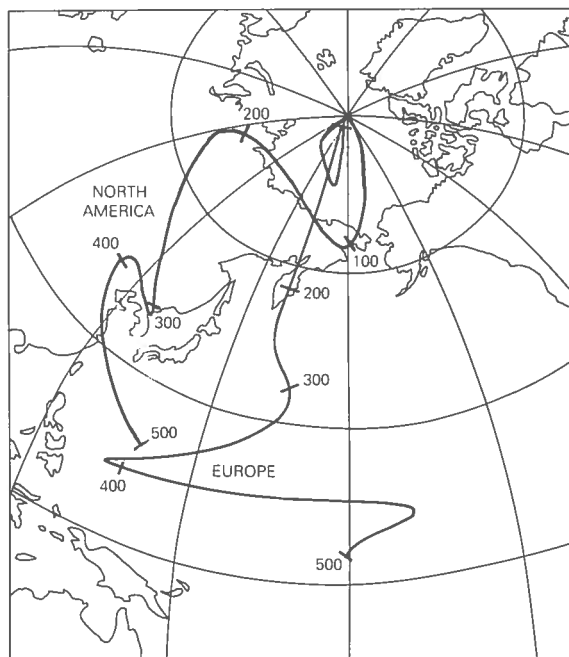
47

Figure 5.10 Apparent polar wander paths for North America and Europe (tick marks every 100 million years). The general trend of the paths from equatorial regions to the north pole reflects the northward drift of both continents during the past few hundred million years. The longitudinal difference between the two paths is a measure of the separation of the two continents which has taken place with the formation of the North Atlantic ocean.

During the past 100 million years the Atlantic ocean has been gradually opening as Europe and America have moved apart. The continents and the floor of the Atlantic ocean have also been slowly drifting northwards. These plate motions combined with a time-averaged geocentric axial dipole field produce the polar wander paths of Figure 5.10. It is possible to use palaeomagnetic data to make detailed, quantitative reconstructions of the past positions of the continents. The palaeolatitude, λ, is given by the palaeomagnetic inclination, I, through the dipole field formula $\tan I = 2 \tan \lambda$. The palaeomeridian is given directly by the palaeomagnetic declination. The palaeolongitude cannot be found from palaeomagnetic data, although relative palaeolongitudes can be derived by comparing sections of apparent polar wander paths.

5.3 Summary

Stars and large planets generate their own magnetic fields through dynamo actions deep in their interiors. The Earth has been generating a field for at least the last three billion years. On account of Coriolis forces planetary magnetic fields average out to be symmetrical about their rotation axis. Throughout the Earth's history its field has closely corresponded to that of a dipole with its axis aligned with the spin axis.

The geomagnetic field can be generated with either its north magnetic pole or its south magnetic pole in the northern hemisphere and it has frequently switched between these two stable polarity states. The present polarity of the Earth's field is referred to as normal. Saturn currently has a reversed field while Jupiter's field is also reversed. In addition to the major reorganisations of the Earth's field lines which take place when the whole field switches polarity, minor field reorganisations are continuously taking place and are referred to as secular changes. These long-lasting fluctuations affect both the intensity and the direction of the field and have been documented over the past few hundred years by observations of changes in the declination of magnetic compasses and the inclination of dip needles.

Palaeomagnetic studies of the fossil magnetism of rocks and sediments extend our knowledge of the geomagnetic field into the ancient past and can be used for investigating geomagnetic dynamo behaviour, for geological dating purposes and for continental drift and microplate tectonic studies.

Further reading

General books

McElhinny 1973. *Palaeomagnetism and plate tectonics.*
Strangway 1970. *History of the Earth's magnetic field.*
Irving 1964. *Palaeomagnetism and its application to geological and geophysical problems.*
Tarling 1983. *Palaeomagnetism.*
Merrill and McElhinny 1983. *The Earth's magnetic field.*
Aitken 1974. *Physics and archaeology.*

Advanced book

Jacobs 1975. *The Earth's core.*