Chapter 1

Magnetism in nature

Magnetism has fascinated mankind since the invention of compasses that could track invisible magnetic field lines over the earth's surface. Much later came the discovery that rocks can fossilize a record of ancient magnetic fields. Unravelling this record – the 'archeology' of magnetism – is the science of paleomagnetism, and understanding how the microscopic fossil 'compasses' in rocks behave has come to be known as rock magnetism.

Rock magnetism is both a basic and an applied science. Its fundamentals concern ferromagnetism and magnetic domains and were developed most authoritatively by Néel. Its applications continue to expand, giving impetus to new research into the mechanisms and fidelity of rock magnetic recording. Some of the history and applications are described in this chapter.

1.1 A brief history

1.1.1 Earth magnetism

Compasses were used in China and the Arab world for centuries before Petrus Peregrinus in 1269 gave the first European description of a working compass. The earliest compasses were lodestones, naturally occurring ores of magnetite (Fe_3O_4). Particular areas, or poles, of one lodestone would attract or repel the poles of another lodestone. This magnetic polarization is the key to their use as compasses in navigation. A suspended lodestone will rotate until its axis of magnetization or polarization, joining north and south poles of the lodestone, lines

up with imaginary field lines joining the north and south geomagnetic poles. In modern terminology, the magnetization ***M*** aligns with the field ***H***.

Although the analogy between the polarization of a magnet and the polarization of the earth's magnetic field had been known at least since the time of Petrus Peregrinus (and probably much earlier), it was not until 1600 that William Gilbert in his celebrated treatise *De Magnete* documented the similarity between the field lines around a lodestone cut in the form of a sphere (a terella) and the field lines around the earth. He concluded: the earth is a great magnet. At the 'north' pole of the terella (actually a south magnetic pole by modern definition), marked A in Fig. 1.1, a compass needle points down, and at the 'south' pole B, it points up, just as for the earth's field. At the equator, the field is horizontal. At intermediate latitudes on the terella, the field is shown as flat-lying, whereas on the earth the field is inclined 60° or more to the horizontal. Nevertheless, Gilbert correctly recognized the essential similarity between the field of a uniformly magnetized sphere and the earth's field. Both are dipole fields (Chapter 2).

In the early nineteenth century, geomagnetic observatories were established around the world, although the coverage was rather uneven in both latitude and longitude (the same remains true today). The accumulated data on the inclination, declination (deviation from north) and intensity of the earth's field

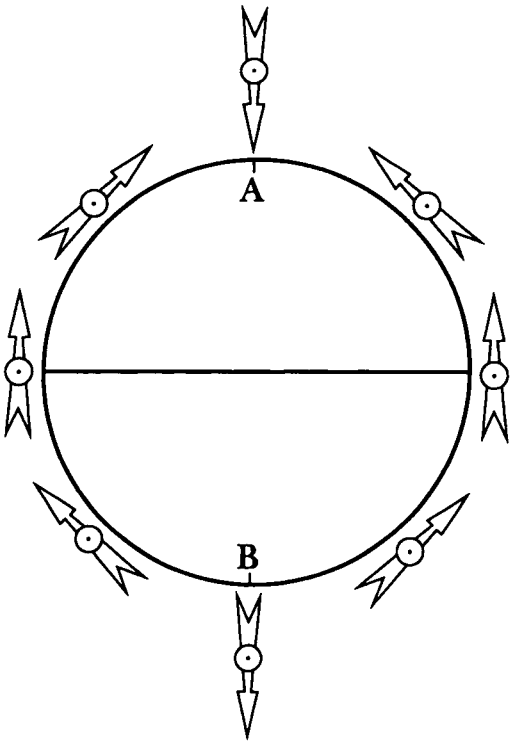

Figure 1.1 Sketch of a terella (a sphere of magnetite) as shown by Gilbert (1600) in *De Magnete*. The inclinations of a compass needle around the terella approximately match those around the earth.

were sufficient for Gauss (1839) to make a pioneering spherical harmonic analysis. He found that the bulk of the field originates within the earth, as Gilbert had surmised, and that it is predominantly a dipole field, again confirming Gilbert's conjecture. Gauss was able to calculate several higher multipole terms in the field with sufficient accuracy to show that the best-fitting dipole is inclined 11.5° to the earth's rotation axis. That is, the magnetic north and south poles do not coincide with the geographic poles.

Gauss and his contemporaries observed that the earth's field is not static. The most striking change in the field between Gauss' time and our own is a >5% decrease in the earth's dipole moment. Similar changes in the paleofield intensity are well documented in rocks. Sometimes they presage a reversal of the field, in which the magnetic north pole becomes the new south pole and vice versa. The question of whether the entire field reverses or only the main dipole is still hotly debated, but the reality of polarity reversals is not in question. Earlier this century, however, it was not clear whether the field had actually reversed or whether some rocks could become magnetized antiparallel to the field ('self-reversal').

Brunhes (1906) and David (1904) were the first to measure natural magnetizations that were antiparallel to the local geomagnetic field. They observed that both lavas and the clays they had baked were reversely magnetized. Since the two materials have very different mineralogies, it is unlikely that both would exhibit self-reversal. Matuyama (1929) showed that younger Quaternary lavas, now known to be <0.78 Ma in age, were all magnetized normally (parallel to the present geomagnetic field) whereas older lavas were reversely magnetized. The association of normal and reverse polarities with different geological times leaves no doubt that the earth's field has reversed. It also forms the basis for the geomagnetic polarity time scale. The most recent polarity epochs, covering the past 4 Ma of earth history, are named in honour of Brunhes (and David), Matuyama, Gauss and Gilbert (see Fig. 1.4).

1.1.2 Ferromagnetism and magnetic domains

Weiss (1907) was intrigued by the fact that iron and other 'soft' ferromagnetic materials with small permanent magnetizations become strongly magnetized when exposed to quite weak magnetic fields. Weiss proposed that external fields play a minor role compared to a hypothesized internal 'molecular field' which aligns the magnetic moments of individual atoms, producing a spontaneous magnetization, $\boldsymbol{M}_s$. In the absence of any external field, the magnetic moments of regions with different directions of $\boldsymbol{M}_s$ (now called domains) cancel almost perfectly, but even a small applied field will either rotate domains or enlarge some at the expense of others. Weiss' theory is the starting point for modern ideas about ferromagnetism (Chapter 2) and ferromagnetic domains (Chapter 5).

The origin of Weiss' molecular field was explained by Heisenberg (1928). Electrostatic 'exchange coupling' of 3d orbitals in neighbouring atoms results in the alignment of the magnetic moments of their electrons. Néel (1948) returned to Weiss' molecular field description in his Nobel-prize-winning treatment of exchange coupling between magnetic sublattices in antiferromagnetic and ferrimagnetic substances.

It was not until the 1930's that F. Bitter developed a method of observing magnetic domains. In the Bitter technique, one observes a smooth (usually polished) surface of a ferromagnetic grain or crystal under the microscope after it has been coated with a thin colloidal suspension of ultrafine ($\ll 1$ μm) magnetite particles. The magnetic colloid gathers at domain boundaries, where flux leaks out of the crystal, and moves with the boundaries in response to applied fields. A photograph of simple domain boundaries on a {110} crystal face of magnetite appears in Fig. 1.2.

Without the benefit of any substantial data base, Landau and Lifschitz (1935) correctly predicted the form and dimensions of magnetic domains and domain walls (the boundaries between domains). Reasoning that domain boundaries would be positioned so as to maximize internal flux closure and eliminate 'magnetic charges' on both internal boundaries and the crystal surface, they postulated exactly the structure illustrated in Fig. 1.2. The plate-like main or body domains (seen here edge on) are terminated by wedge-shaped surface or closure domains which provide a continuous flux path between neighbouring body domains.

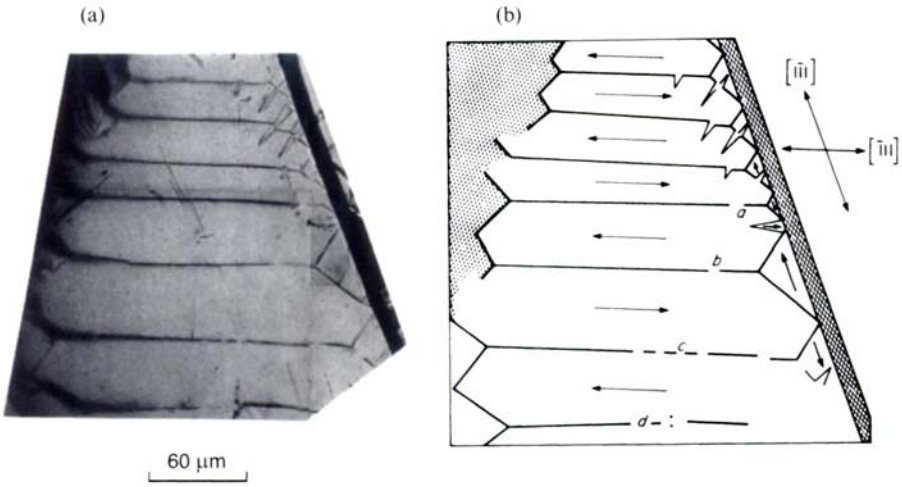

Figure 1.2 Body and surface (closure) domains observed on a polished {110} surface of a large natural magnetite crystal. The arrows are inferred directions of M_s within each domain. The black lines are domain walls, made visible by means of a magnetic colloid. [After Özdemir and Dunlop (1993a) © American Geophysical Union, with the permission of the authors and the publisher.]

The final development of classic domain theory came with the work of Kittel (1949), whose calculations of wall and domain widths extended and refined Landau and Lifschitz's estimates, and Rhodes and Rowlands (1954) and Amar (1958), who first calculated the magnetic self-energy of a rectangular block, representing a domain wall. Although the self-energy (or demagnetizing energy) had long been known to control the size and shape of domains, its importance in determining domain wall energy and size had not been recognized. In small (< 1 μm) magnetite crystals, walls become broad and their demagnetizing energy increases accordingly. Rhodes and Rowlands' results are also the starting point for most micromagnetic calculations of fine-scale magnetization structure (Chapter 7).

1.1.3 Rock magnetism

Earth magnetism and ferromagnetism developed in mutual isolation until Koenigsberger (1938), Thellier (1938) and Nagata (1943) brought earth magnetism into the laboratory. They attempted to reproduce and understand the process by which igneous rocks are magnetized in nature. The new science was given a name with the publication in 1953 of Nagata's classic book, *Rock Magnetism*.

Koenigsberger, Thellier and Nagata gave new magnetizations to their rocks (lavas and archeological materials like bricks and baked clays) by heating them to high temperatures and cooling them in a weak field. This **thermoremanent magnetization** (TRM for short) was always parallel to the field in which it was acquired and its intensity was proportional to the strength of the field. TRM was therefore an accurate recorder of magnetic field directions, as geomagnetists had assumed, and was also potentially a means of determining paleofield intensities.

TRM in the laboratory had a series of remarkable properties which were most clearly described by Thellier. When a sample was given a **partial TRM**, by exposing it to a field only in a narrow cooling interval (between T_1 and T_2 say), and was subsequently reheated to high temperatures in zero field, the TRM remained unchanged below T_2, but completely disappeared between T_2 and T_1.

Néel (1949) explained this observation as a consequence of **blocking** of TRM during cooling at a single blocking temperature T_B determined by the size and shape of a particular magnetic grain. The TRM is unblocked when reheated through T_B. The wide spectrum of grain sizes and shapes in a rock leads to a continuous distribution of blocking temperatures of partial TRM, but each individual grain has one and only one value of T_B.

This individuality of blocking temperatures illuminates other experimental TRM 'laws'. Partial TRM's acquired in different temperature intervals are mutually independent. For example, if two partial TRM's are added vectorially in a single cooling run, by rotating the field by 90° after the first partial TRM is produced, the two partial TRM's are observed to demagnetize independently of

each other during zero-field reheating. Each partial TRM disappears over its own blocking temperature interval and the total magnetization vector retraces the exact pattern it followed during cooling.

The 'additivity law' is another consequence of the one grain/one blocking temperature principle. In this case, we combine results of separate experiments in which partial TRM's are produced over different temperature intervals. The intervals, taken together, cover the entire temperature range from the Curie point (the ferromagnetic ordering temperature, usually several hundred °C) to room temperature. The sum of all partial TRM intensities is found to be equal to the intensity of total TRM produced in a single cooling. Since each partial TRM contains a unique part of the T_B spectrum of the total TRM, this law is easy to understand also. Figure 1.3 illustrates how total TRM is built up from a spectrum of partial TRM's.

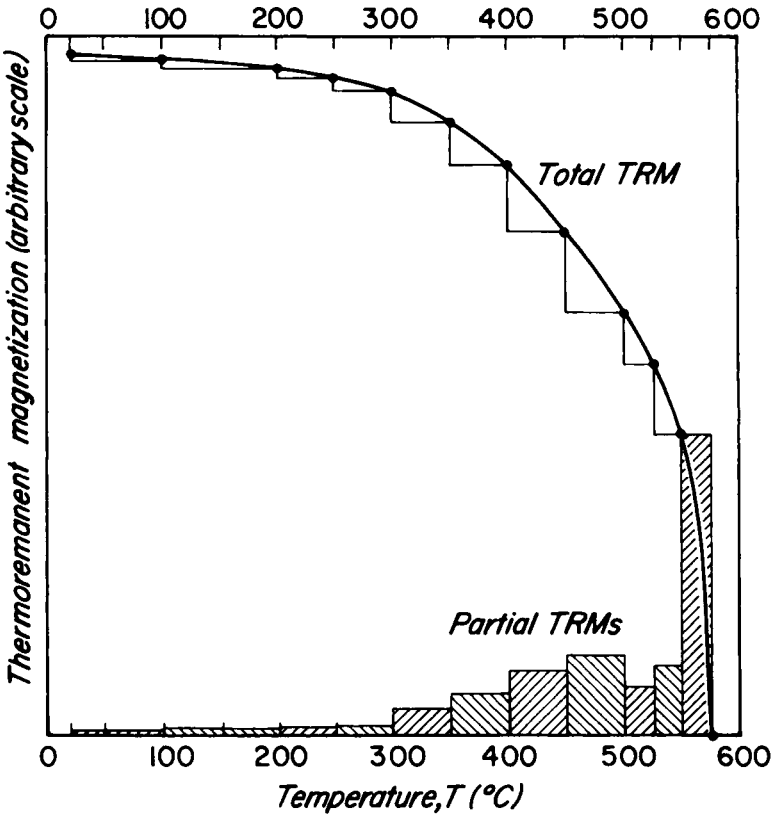

Figure 1.3 Relationship between the unblocking temperature spectrum of total TRM (solid curve, with steps indicating increments over 25–100 °C heating intervals) and the blocking temperature spectrum of partial TRM's (hatched). The partial TRM gained over a particular temperature interval equals the incremental TRM lost over the same interval (the reciprocity law) and the sum of all partial TRM's equals the intensity of total TRM (additivity law).

Koenigsberger, Thellier and Nagata were fortunate in their choice of fine-grained rocks containing an abundance of single-domain magnetite and hematite (αFe_2O_3) grains. **Single-domain** grains are too small to accommodate a domain wall and must change their magnetizations by rotation. Only single-domain grains exhibit a unique TRM blocking temperature and obey the Thellier laws (see Chapter 8). The TRM of larger **multidomain** grains (Chapter 9) is complicated by the mobility of domain walls and is less obviously suitable for recording the paleomagnetic field. Stacey (1962, 1963) proposed a new view: multidomain grains that have aspects of single-domain behaviour. Pseudo-single-domain (PSD) models have inspired and shaped rock magnetic research ever since, and form the subject of Chapter 12.

1.2 How rock magnetism is applied

1.2.1 Magnetic lineations: the seafloor record of reversals

Lithospheric plate motions result in a smooth spreading of newly erupted seafloor lavas away from mid-ocean ridges. The lavas contain fine-grained titanomagnetite ($Fe_{2.4}Ti_{0.6}O_4$) which acquires an intense TRM on cooling below its Curie temperature. The geomagnetic field which produces the TRM reverses at irregular intervals. The overall result (Fig. 1.4) is a set of normally and reversely magnetized bands of seafloor, symmetrical about mid-ocean ridges and increasing in age with distance from the ridge.

The magnetized seafloor produces a magnetic field (the 'anomaly' field) which is added to or subtracted from the local geomagnetic field. The polarized strips of seafloor are typically 10 km (Atlantic) to 80 km (Pacific) wide. Usually the field is measured at or just above the sea surface, 1–2 km above the seafloor. Because of this fortunate source geometry, fine details of the seafloor field are filtered out and the anomaly field is quite flat except over the boundaries between magnetized bands of seafloor, where it reverses sign. When contoured over the ocean, the anomaly field reproduces the lineated pattern of seafloor TRM that underlies it. These lineated anomalies are often called **magnetic stripes**. The boundaries between stripes mark times of field reversals.

Magnetic stripes are the most tangible evidence of seafloor spreading and plate tectonics. Historically they were decisive in converting geologists to a mobilist view of the earth. They are the main means of determining plate velocities, and they provide the most complete record of geomagnetic reversal history during the last 175 Ma. In view of this importance, and the exhaustive research that has been carried out since Vine and Matthews first proposed their model in 1963, it might be thought that very little remains to be learned. Surprisingly, this is not so. Using a tape recorder as an analogy, the tape drive (the spreading

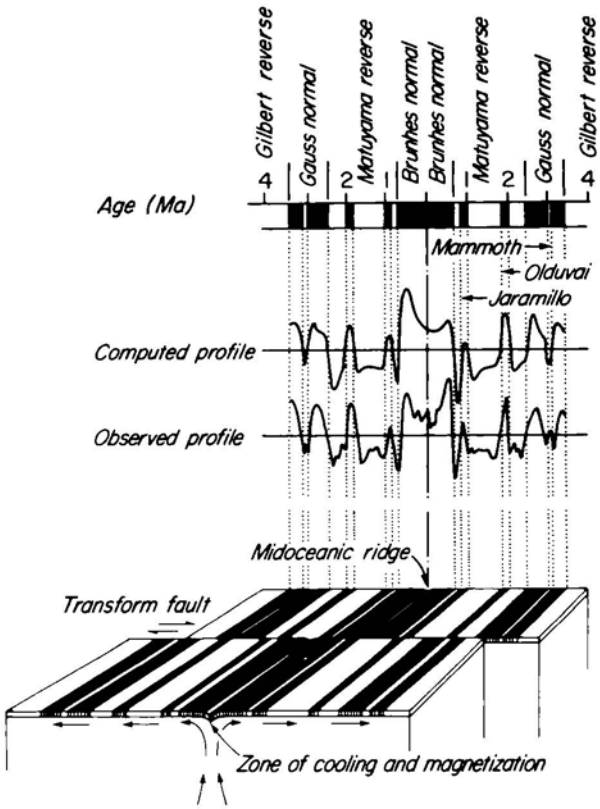

Figure 1.4 The Vine and Matthews' (1963) model of seafloor spreading (bottom) and linear magnetic anomalies over the oceans. The anomaly profile (centre) is a spatial replica of the time sequence of geomagnetic field reversals (top). [After Takeuchi *et al.* (1970) Fig. 8.4, with the permission of the authors.]

seafloor and the tectonic forces that drive it) and the time series of geomagnetic reversals that are recorded have been thoroughly looked into, but the recording medium itself – the seafloor and its magnetic minerals – have not.

The first surprise is that much of the seafloor carries no TRM at all. Before new seafloor has spread more than 20 km from its parent ridge, seawater has penetrated deep along fissures and oxidized the primary titanomagnetites, changing both their chemistry and their magnetization. A second major question is how much of the magnetic stripe signal originates in the surface layer of seafloor lavas and how much has a deeper source in rocks to which we have no direct access. How these problems are being tackled is described in Chapters 13 and 14.

1.2.2 Other magnetic anomalies

Magnetic stripes are not the only interesting anomalies in the geomagnetic field. Anomalies of local and regional extent are used by geologists as tools in mapping lithologies, structures and deformational styles, and delineating metamorphic terranes. They may also have direct economic interest if they pinpoint mineable concentrations of magnetite or hematite or if they trace iron formations outlining

the structure of greenstone belts in which gold and other precious metals are concentrated.

The exact connection between anomaly fields and the magnetic petrology of the source body or region is not always clear, usually because the source cannot be sampled directly. The interpretation of an anomaly also depends on whether the magnetization is mainly **induced** by the present geomagnetic field and is therefore in the present field direction, or is mainly **remanent** (permanent) and parallel to some ancient field, which because of plate movements and field reversals may be at a large angle to the present field direction. The Koenigsberger Q_n ratio of remanent to induced magnetization is therefore an important parameter.

Some major questions have been raised by long-wavelength (regional to global scale) anomalies of the earth and moon revealed by satellite and spacecraft magnetometers. Lunar anomalies are often too large to be explained by typical magnetizations of returned Apollo samples (Chapter 17). Closer to home and just as vexing are broad anomalies over both continents and oceans whose amplitudes demand a substantial source of magnetization in the lower crust or upper mantle. Exotic magnetic minerals seen in crystal xenoliths have been favoured by some (Wasilewski and Mayhew, 1982; Haggerty and Toft, 1985) and an enhancement of magnetization by high temperatures at depth by others (see §10.2.2). The question is far from settled (Toft and Haggerty, 1988).

1.2.3 Records of geomagnetic field variation

Reversals are only one (albeit the most spectacular one) of many geomagnetic field variations. Magnetostratigraphy, the dating and correlation of geological strata based on the magnetic reversals they record, has become a major science. For example, the Cretaceous–Tertiary boundary at which the dinosaurs became extinct has been dated very precisely at 66 Ma using magnetostratiography (Lowrie and Alvarez, 1981). Reversal records are recovered both from sediments cored in the oceans and from ancient sedimentary sections preserved on land. For times before the mid-Jurassic (> 175 Ma) there are no surviving ocean sediments or seafloor and our only record of reversals comes from sedimentary rocks. Most sections include times when sedimentation was slow or non-existent and the quality of the magnetic recording may vary considerably down the section. Deciphering the reversal record requires much time, labour, insight, comparison of sections – and rock magnetic tests (Chapter 11).

The behaviour of the geomagnetic field during a polarity transition and during excursions (which may be aborted transitions) is of vital importance in understanding the core dynamo. A reversal lasts only a few thousand years. The best magnetic records of reversals come from lavas which are closely spaced in time and from rapidly deposited ocean sediments. Figure 1.5 is a high quality record from Ocean Drilling Program cores in the equatorial Pacific. The gradual decline

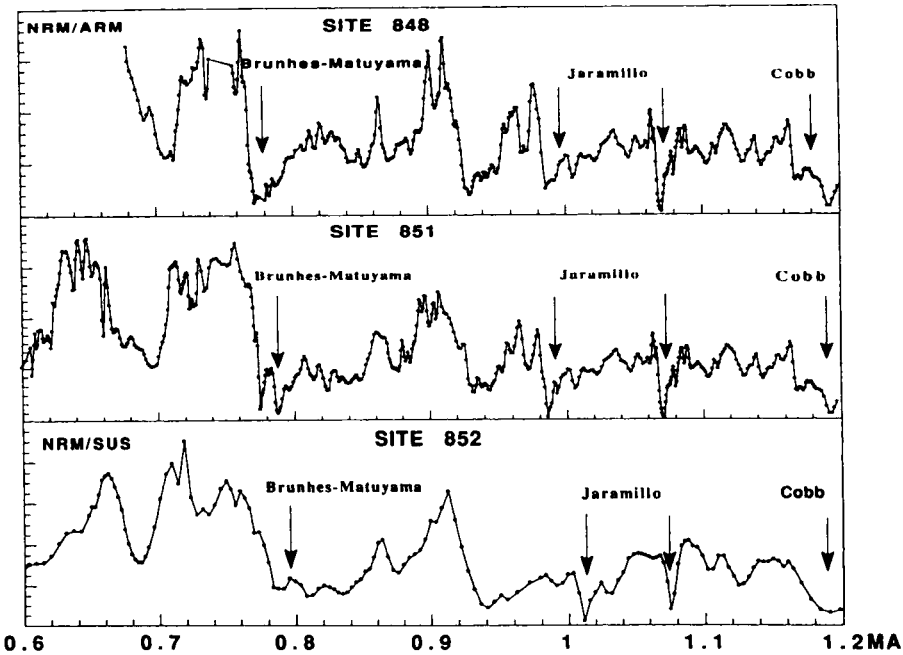

Figure 1.5 A high quality record of geomagnetic intensity variations recorded in oceanic sediment cores. The Brunhes–Matuyama polarity transition (see Fig. 1.4) and the Jaramillo and Cobb events (aborted reversals) are indicated. [After Meynadier *et al.* (1995), with the permission of the authors and the publisher, The Ocean Drilling Program, College Station, TX.]

in field intensity preceding a transition or excursion, the rapid recovery of intensity after the event, and the correlation between the amount of intensity recovery and the time until the next event are clear in the record.

The geomagnetic field changes direction and intensity even at times of constant polarity. The decline in dipole moment in the past 150 years was mentioned earlier. The field declination and inclination also vary. Lake sediments, soils, and loess (windblown sediment) record similar behaviour of the earth's field in the past, known as paleosecular variation (PSV). PSV records, if of high quality, are a powerful tool for dating and correlating sediments and soils over the past few thousand years. The correlation cannot be made over large distances because secular variation records fairly local variations in the field.

The main problem in paleofield variation records is fidelity of the recording. Both sediments and lavas may record a magnetization whose inclination is systematically shallower than the field inclination. Why inclination errors occur and what can be done to recognize and overcome them will be considered in Chapters 14 and 15. Random noise often obscures the details of the signal in sediments. Some of this noise is introduced in the coring process, but much is inherent to the magnetic grains themselves. Effective ways of 'cleaning' the records to