[12]

Mineral magnetism in marine sediments

> We have *first* the matters derived from the wear of coasts, and those brought to the sea by rivers . . . In oceans affected with floating ice we have land debris . . . Second – we have dust of deserts . . . In the trade wind region of the N. Atlantic we have a very red-coloured clay . . . which is largely made up of dust from the Sahara . . . Third – we have the loose volcanic materials, which have been . . . universally distributed as pumice or as ashes carried by the wind . . . While examining the deposits during the cruise I frequently observed among the magnetic particles from our deep-sea clays small round black-coloured particles which were attracted by the magnet, and I found it difficult to account for the origin of these. On our return home I entered into a more careful examination of the magnetic particles. Some of the particles are little [iron] spherules . . . that appear to have a cosmic origin.
>
> John Murray Esq., 1876
> *Proc. R. Soc. Edinburgh, 247, 261*

12.1 Introduction

Of all the environmental systems considered in this book, the sea is, from a mineral magnetic point of view, by far the most complex. Potentially significant sources of magnetic minerals are at their most varied and include not only all those of relevance to lacustrine and atmospheric studies but also submarine and extraterrestrial sources. Supply pathways are often more complex and extended than anywhere else. Depositional environments range from extreme high to extreme low energy, and also span a bewildering range of chemically and biologically modulated variation. Moreover marine environments include some of the least accessible on earth. In consequence it is not possible to attempt a comprehensive and integrated review. The approach adopted here is selective, thematic and empirical, focusing on a limited number of quite independent studies in contrasted marine contexts. These range from pelagic environments where our major concern is the palaeoclimatic record, to coastal situations dominated by recent human impact.

12.2 The origin and flux of marine magnetic minerals

Figure 12.1 summarises in schematic form the major sources and pathways of the magnetic minerals found in marine sediments. The present section attempts to evaluate these sources and pathways in the light of the concern with possible palaeoclimatic linkages and implications developed in 12.4 below.

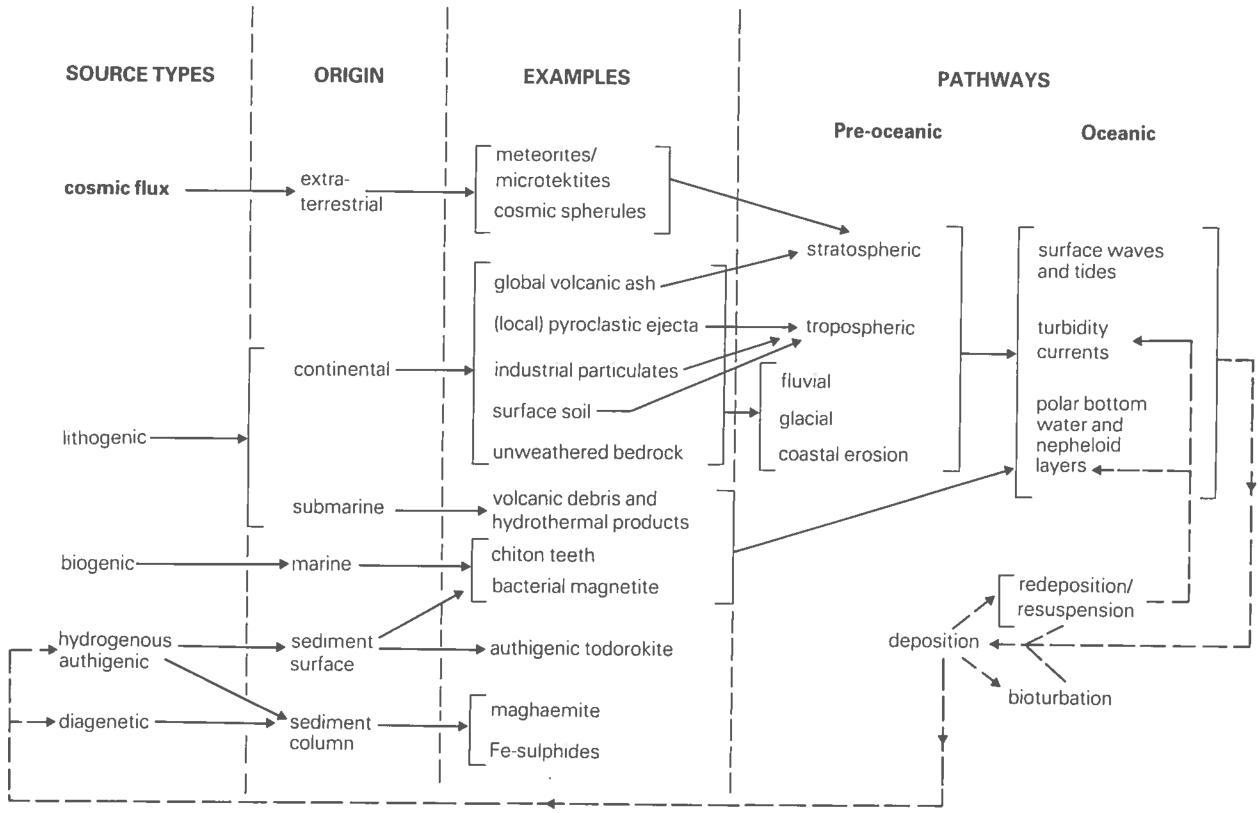

Figure 12.1 The sources and supply pathways of magnetic minerals encountered in marine sediments.

COSMIC

Due to magnetite formation during the entry of extra-terrestrial material into the Earth's atmosphere, most of the cosmic spherule component of marine sediments is strongly ferrimagnetic (Brownlee 1981). Three conditions must be met before cosmic particles can be of any real relative significance in the marine sediment record from a given area:

(a) The sediments should predate the Industrial Revolution, since as shown in Chapter 11 industrial and fossil-fuel combustion processes have, over the last century, greatly increased the atmospheric concentration of magnetic spherules, even in areas remote from urban and industrial activity.
(b) Even in pre-industrial times, the sediments must come from those areas of the sea bed most remote from any kind of particulate terrigenous input.
(c) Sediment accumulation rates must be extremely slow.

In practice these conditions are met in some abyssal sediments from open oceans, especially the Pacific. It is possible to extract cosmic spherules from this type of sediment using simple magnetic techniques and the concentrations (up to several mg kg^{-1}) may be significant in the mineral magnetic record. However, none of the cores considered in the succeeding sections fulfills criteria (a) and (b) and it is unlikely that cosmic spherules comprise more than a very minor 'background' component in the mineral magnetic record. Meteorites and microtektites are of much more limited significance spatially and temporally and can be ignored in the context of the present account.

LITHOGENIC

We can infer from preceding chapters that three types of continental source will be important in the flux of magnetic minerals to marine sediments on the timescale of interest in this section. These are volcanic sources, soil and bedrock. Volcanic inputs to marine sediments will range from highly local pyroclastic debris through widely distributed tephra layers, to deposition from global volcanic dust veils resulting from the injection of fine ash particles into the stratosphere. Haggerty (1970) regards the volcanic input of magnetic minerals to deep-sea sediments as a significant component in natural remanence studies.

Soil and bedrock sources contribute particulates to the oceans as a result of wind, fluvial, glacial and shoreline erosion. The figures summarised by Prospero (1981) and reproduced in Table 11.1 suggest that most authorities regard land surfaces exposed to wind erosion as much greater contributors of atmospheric dust than are volcanoes at the present day, though this may be in part a function of recent human activity. If we bear in mind the often high magnetic concentration in such surface material and add the effects of fluvial inputs (cf. Currie & Bornhold 1983), as well as glacial rafting (Mullen *et al.* 1972, Ruddiman *et al.* 1971), we must conclude that continental erosion is a major source of detrital marine magnetic minerals even in many pelagic environments. The processes of weathering and erosion which control the nature and the release of magnetic minerals in soils, as well as the atmospheric and hydrospheric pathways through which magnetic minerals will pass from source to sediment, are all strongly controlled by climatic change.

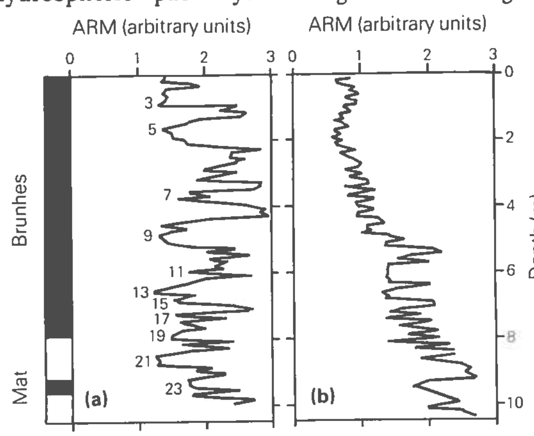

Figure 12.2 Down-core variation in intensity of anhysteretic remanent magnetisation for two North Pacific cores. (a) A fossiliferous core with a 1 m year polarity timescale; (b) a non-fossiliferous core with authigenic iron oxides. (After Johnson *et al.* 1975.) The sequence of numbers 1 to 22 in (a) indicates a possible correlation of the mineral magnetic fluctuations with Pleistocene climatic change and the oxygen isotope stages.

BIOGENIC

It is rather difficult to determine the relative importance of biogenic magnetic minerals in marine sediments. In nearshore sediments, as in lakes, a wealth of circumstantial evidence points to a detrital origin for the bulk of the magnetic minerals present. However, Kirschvink (1982a) has recently inferred the presence of bacterial and possibly algal (Lins de Barros *et al.* 1981) magnetite in marine clays of Miocene age in Crete and transmission electron microscopy has revealed the presence in a variety of deep-sea sediments of several morphologically distinct magnetite crystals, which closely resemble those formed by magnetotactic bacteria (Kirschvink & Chang 1984). There is also the possibility that in the sediments of the open oceans, magnetite derived not only from magnetotactic bacteria but also from chiton teeth (cf. Lowenstam 1981) makes a significant contribution.

AUTHIGENIC AND DIAGENETIC

Recent reviews by Henshaw and Merrill (1980) and by Burns and Burns (1981) show that both authigenic and diagenetic magnetic phases are relatively common in marine sediments. Moreover, their formation is often related to variables such as accumulation rate and organic content, which are in turn often controlled by climatic variations. However, the conditions under which these phases form are becoming increasingly well understood and defined.

Authigenic maghaemite is found in 'red clay' deep-sea sediments from the north and central north Pacific, where it carries an unstable, low coercivity magnetisation (Kent & Lowrie 1974, Johnson *et al.* 1975). The maghaemite is most probably formed *in situ* by the low temperature oxidation of magnetite. Kent and Lowrie noted its widespread occurrence in sediments formed before three million years ago, possibly on account of the warmer upper Cainozoic climate and associated slower sediment deposition rates. Such low sedimentation rates increase the exposure time of sedimenting particles to oxidising bottom waters and permit oxidation near the sediment/water interface before burial occurs. Johnson *et al.* found that the most heavily oxidised magnetic minerals occurred in the non-fossiliferous cores, which displayed a slow erratic decline of magnetic mineral content towards the sediment surface (Fig. 12.2b) and poor palaeomagnetic records. In contrast clear palaeomagnetic polarity zones and distinctive fifty thousand year variations in magnetic mineral concentration (Fig. 12.2a) were found for the fossiliferous cores. As discussed below in Section 12.4 such distinctive mineral magnetic variations can be matched to the oxygen isotope stratigraphy.

Authigenic magnetic iron sulphide minerals produced in highly reducing conditions have been identified by Kobayashi and Nomura (1972, 1974) in deep-sea sediment cores from the Sea of Japan. Unusually strongly reducing conditions are needed to grow pyrrhotite. Perhaps times of stagnant bottom waters and abundant organic matter, associated with low sea levels when the Sea of Japan may have been entirely surrounded by land, could have produced suitable conditions for the authigenic *in situ* growth of pyrrhotite and pyrite.

Selective dissolution of magnetite grains in strongly reducing hemipelagic muds on the Oregon continental slope has been demonstrated by Karlin and Levi (1983) to lead to downcore coarsening of the magnetic grain size as the smallest grains are removed. This change in the top 30 to 80 cms is accompanied by a reduction in magnetic concentration. Further downcore the magnetic grain size begins to decrease while concentration continues to decline as the remaining grains dissolve.

12.3 Core correlation in marine sediments

Several published studies (e.g. Radhakrishnamurty *et al.* 1968) point to the link between magnetic susceptibility variations and tephra layers in marine sediments, and in each case, the opportunity for core correlation using magnetic measurements is at least implicit. In the case of lake sediments (Ch. 10) detailed core correlation is often possible even where tephra layers are absent and, by analogy, we may hope to find that marine sediments offer similar opportunities despite the increases in temporal and spatial scale involved.

Figures 12.3 to 12.5 illustrate susceptibility-based core correlation in a variety of marine contexts. The cores for which whole core volume susceptibility is plotted in Figure 12.3 are part of a large set of piston and gravity cores from the eastern Mediterranean between Cyprus and southern Turkey. Even without a detailed timescale the correlation scheme highlights major spatial changes in deposition rates and patterns. The basis for the correlations is thought to be rather complex. The total time interval involved includes the later part of the last glacial interval and much of the Holocene, and part of the variation reflects fluctuations in terrigenous influx resulting from climatic change and human activity. The period of minimum susceptibility coincides with the well

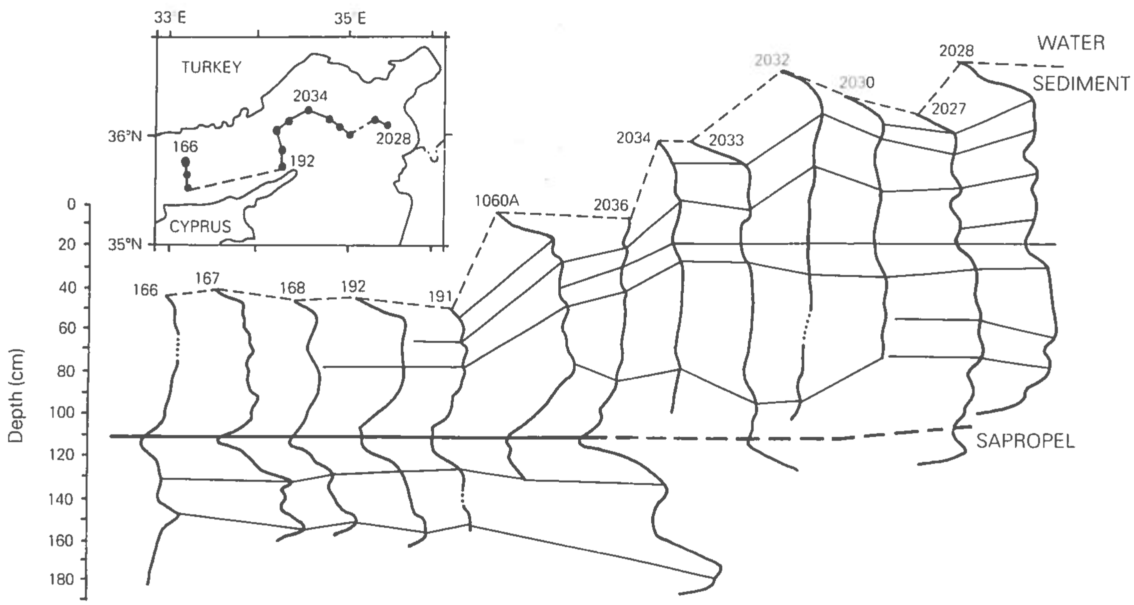

Figure 12.3 Whole core volume susceptibility logs from late-Pleistocene piston and gravity cores from the eastern Mediterranean. (All cores were scanned in the Department of Geological Sciences, Imperial College, London, UK.)

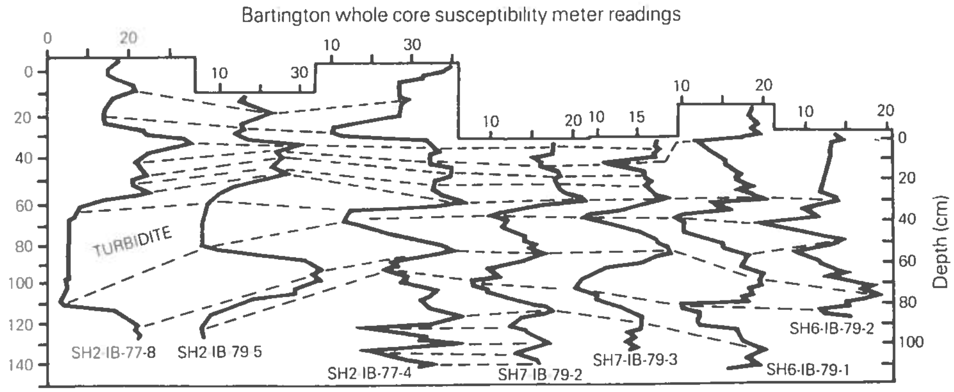

Figure 12.4 Single-sample mass susceptibility logs for cores from the Azores region of the eastern Atlantic.

known sapropel layer dated to *c.* 5000–7000 BP and may reflect a period of minimum terrigenous input between the end of the last glaciation and the beginnings of prehistoric farming and accelerated soil erosion (cf. Section 10.6). In addition, tephra layers are certainly present though not always in a recognisably continuous form from core to core, and these are responsible for some of the main susceptibility peaks.

The core traces plotted in Figure 12.4 are from a topographically varied area of the eastern North Atlantic in the region of the Azores. They include tephra layers of local origin reflecting the proximity of a region of active vulcanicity (Robinson 1982). Several cores (e.g. 79–5) show, in addition, an interval of uniformly low susceptibility corresponding with a microturbidite layer. Robinson (1982) has reported additional mineral magnetic data from cores in which the occurrence of the microturbidites has been inferred. In each core a major change in the $IRM_{-100\,mT}/SIRM$ ratio is associated with the microturbidite layer. The saturation remanence is clearly much 'harder' in the layer than in the samples on either side, probably reflecting some diagenetic change from paramagnetic and/or ferrimagnetic forms to antiferromagnetic as a result of post-depositional bacterially mediated organic oxidation. The parallel between the $IRM_{-100}/SIRM$ changes and the distribution of uranium and the redox-sensitive metals in these cores (Colley *et al.* 1984) suggests that the identification of this type of mineral magnetic feature has significant geochemical implications.

Figure 12.5 plots a series of single-sample susceptibility traces, compiled by Robinson (1982) from a set of cores obtained by the Institute of Oceanographic Sciences, for an area close to the King's Trough area of the North Atlantic. Here the correlations are extremely clear and largely independent of such tephra and turbidite layers as have been recognised. These cores are much more representative of relatively undisturbed pelagic environments and the degree of correlation is likely to be typical of many deep-sea sediment cores obtained from comparable environments in all the worlds major ocean floors. The basis of the correlation is believed to be climatic modulation of the magnetic mineral types and concentrations and is discussed fully in the following section.

12.4 Mineral magnetism and palaeoclimate in deep-sea sediments

One of the most striking observations to emerge from studies of deep-sea sediments has been the parallelism between palaeomagnetic parameters and palaeoclimatic indicators noted by many authors and encountered in all the major ocean basins of the world. Frequently, variations in the intensity of natural remanent magnetisation (NRM) and sometimes also changes in inclination closely parallel the palaeoclimatic signature whether this is derived from ^{18}O ratio determination, foraminiferal, coccolith or calcium carbonate analysis. Figure 12.6 illustrates this parallelism. Many authors have taken these and similar results as confirmation of some type of fundamental linkage between the behaviour of the geomagnetic field and changes in climate. Clearly, the

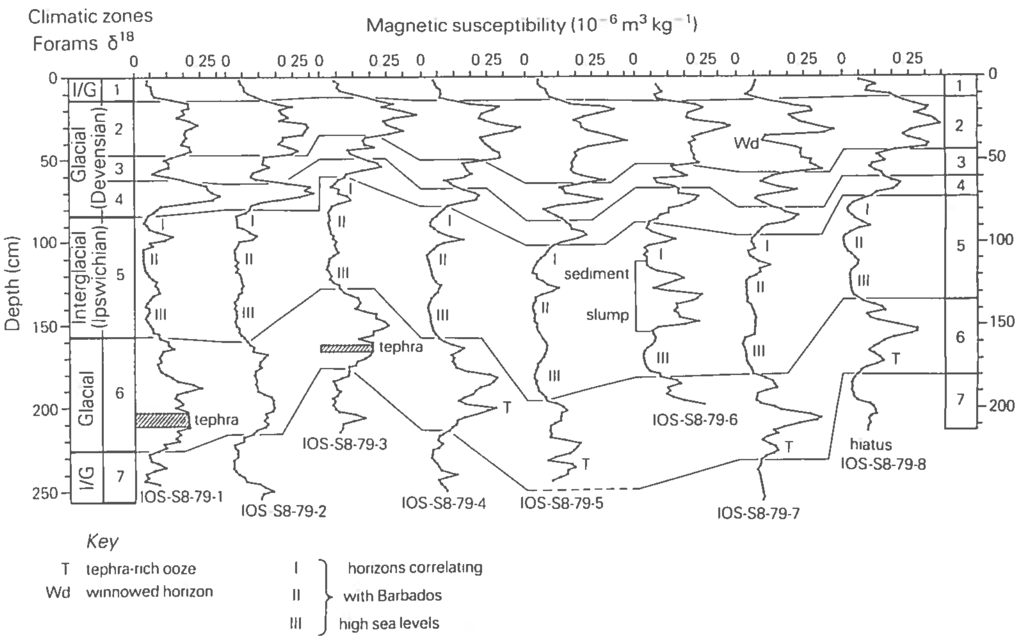

Figure 12.5 Single-sample mass susceptibility logs for eight cores from the North Atlantic. The correlations shown are related to the palaeoclimatic inferences developed in Section 12.4 of the text and illustrated in Figure 12.7.

linkage, if real, must exist on a global scale. The theories proposed have invoked a range of possible mechanisms. Harrison and Prospero (1977) proposed a direct effect of the magnetic field on the atmosphere by means of a causal link between low geomagnetic field intensities, higher effective cosmic radiative flux and warmer climates. Doake (1977) suggested that the distribution of ice may have influenced the geomagnetic field by modifying the fluid motions within the core. Wollin *et al.* (1977) and Opdyke and Kent (1977) speculated that astronomical variables, for example changes in the eccentricity of the Earth's orbit, may control both climatic and geomagnetic intensity variations. The problem of accounting for the palaeoclimatically related inclination variations has also attracted some attention. Stuiver (1972) suggested that long-term changes in planetary configuration may be responsible for these apparent correlations.

More recently several authors have either challenged the reality of a geomagnetic–palaeoclimatic link (Kent 1982) or put forward a completely different kind of explanation for the observed parallelism (Bloemendal 1980; Oldfield & Robinson 1985; Robinson 1986). Kent (1982) uses data from the southern Indian Ocean and from the equatorial Pacific to establish two important points. He shows that not only does NRM intensity parallel calcium carbonate concentration and, where available, the ^{18}O record, but so do the mineral magnetic parameters, susceptibility and IRM. Whereas changes in NRM intensity can, in theory, reflect geomagnetic variations, changes in susceptibility and IRM cannot. Moreover, in the equatorial Pacific, the peak $CaCO_3$ concentrations and the NRM, IRM and susceptibility minima are associated with maximum ice volume and minimum temperature during glacial intervals, whereas in the southern Indian Ocean the converse applies. The parallelism between *mineral* magnetic parameters and palaeoclimatic indices observed in these cores as well as in those studied by Bloemendal from the South Atlantic and Robinson from the North Atlantic makes a global geomagnetic–palaeoclimatic linkage difficult to sustain as the primary cause of the magnetic–climatic relationship on the timescales resolved by marine sediment. This difficulty is further compounded by the opposite nature of the magnetic–palaeoclimatic relationship at Kent's two sites. Clearly an alternative interpretation is required.

Oldfield and Robinson (1985) draw together evidence from both lake and marine sediments to propose a sedimentological/mineralogical as against

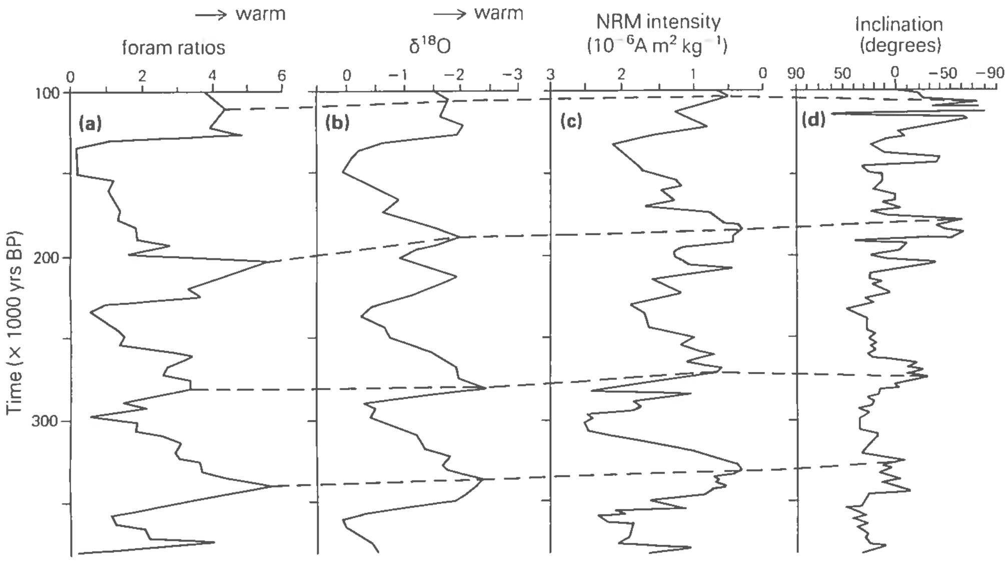

Figure 12.6 Correlation of (a) the *Globorotalia menardii* climate curve, (b) the oxygen isotope curve of *Globigerinoides ruba*, (c) magnetic intensity and (d) magnetic inclination in deep-sea cores V12-122 from the Caribbean from 380 000 to 100 000 years BP. (After Wollin *et al.* 1977.)

a geomagnetic basis for the palaeomagnetic–palaeoclimatic relationships. The essential components of the explanation are as follows:

(a) The fundamental palaeoclimatic linkage is with mineral magnetic and not palaeomagnetic properties. The linkage is reflected in variations in the intensity of NRM because these are strongly dependent on the changes in magnetic concentrations, grain size and mineralogy which are reflected in the mineral magnetic parameters (see Ch. 4). Confusion has arisen in the past from the failure of many workers to distinguish clearly between NRM intensity and geomagnetic palaeointensity, and from the fact that interest in sedimentary *palaeo*magnetism preceded any detailed *mineral* magnetic studies of deep-sea cores.

(b) Where a linkage is apparent between inclination and palaeoclimate it reflects the relationship between sediment structure and 'inclination error' noted more often in glacial varve than marine sediment studies (cf. Sections 13.2.2 & 14.5).

(c) All aspects of the magnetic–climatic linkages so far identified in deep-sea sediments are a function of the control exercised by climate over the complex of variables determining shifts in magnetic mineral sources and types, in the flux of magnetic minerals to the sediments, and in the concentration of non-magnetic sediment components such as calcium carbonate.

Figure 12.7 shows results from IOS Core S8-79-4 in the North Atlantic. It spans most of oxygen isotope stages 1 to 7 and thus represents the last *c.* 250 000 years of the Pleistocene. The foraminiferal, calcium carbonate and ^{18}O records are set alongside plots of χ, SIRM, ARM and $IRM_{-100\,mT}$/SIRM. The non-magnetic indices provide a mutually consistent picture of palaeoclimatic variation, which is in turn reflected in each of the curves of mineral magnetic variation. Furthermore, the speed and relative ease with which detailed magnetic characterisation can be accomplished makes it feasible to resolve variations much more closely than is normally done using non-magnetic techniques. At the detailed level of resolution achieved here the minor variations in all magnetic parameters during ^{18}O zone 5 appear to reflect the sequence of subdivisions previously established by Shackleton (1977) and the reality of the palaeoclimatic signature in the mineral magnetic record can hardly be doubted.

It is especially significant that here as in Bloemendal's mid-Pliocene record from South Atlantic DSDP Core 514 (Fig. 12.8), not only do

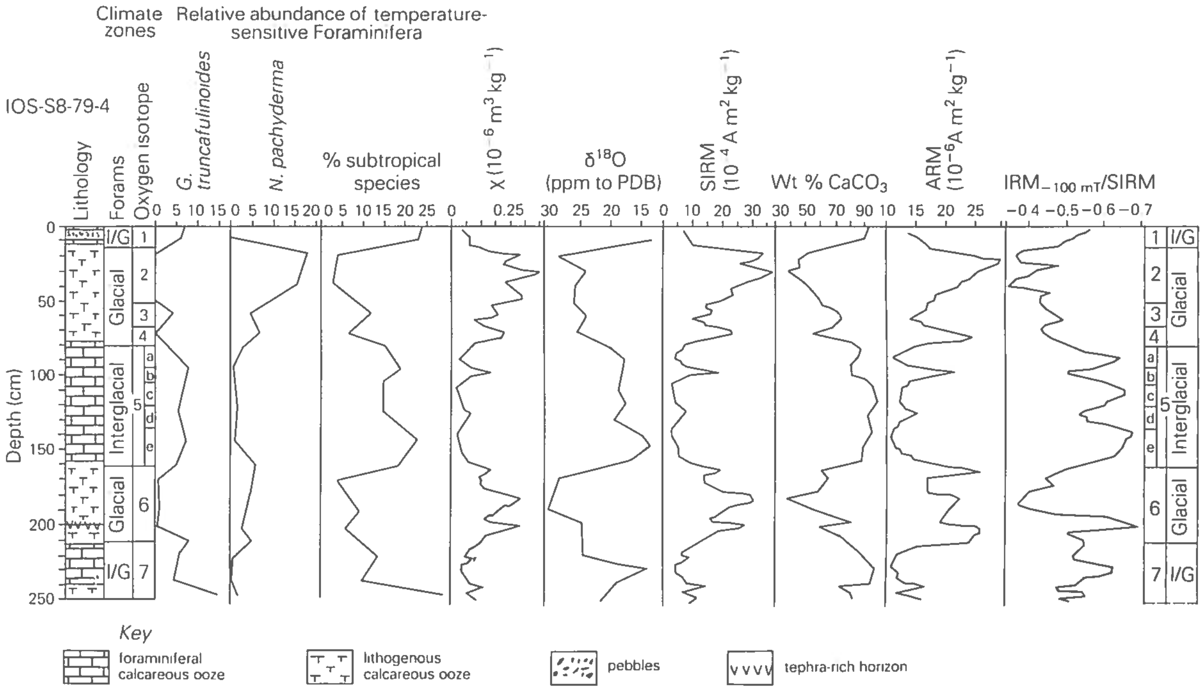

Figure 12.7 Mineral magnetic and palaeoclimatic signal in Core IOS-S8-79-4 from the North Atlantic (see Oldfield & Robinson 1982 and text).

concentration-dependent parameters such as χ, SIRM and ARM reflect palaeoclimate control but so also does the normalised, hence essentially qualitative $IRM_{-100\,mT}$/SIRM parameter. This confirms that the mineral magnetic signature, though in part an expression of varying degrees of dilution by the non-magnetic sediment components such as $CaCO_3$, reflects also changes in magnetic grain size and/or mineralogy. In the IOS core, $IRM_{-100\,mT}$/SIRM varies from −0.34 to −0.68, and in the 'hardest' samples up to 15% of the total SIRM remains unsaturated in a reverse field of −0.3 T. In the DSDP, core variations in the stability of the remanence are even greater and the occasional extreme sample has a coercivity of SIRM close to 0.1 T. Although recent studies by King *et al.* (1982) have stressed the importance of stable single-domain magnetite in samples with relatively high remanent coercivities, the wide range of variation found at these sites, the high coercivities of the 'hardest' samples and the presence of a significantly varying and at times highly unsaturated component at 0.3 T strongly suggest important mineralogical variations including shifts in the ratio of ferrimagnetic to canted antiferromagnetic grains. The glacial parts of the record are marked by high magnetic concentrations and a relatively 'hard' mineral assemblage and the interglacials by the converse. Clearly the record implies changes in magnetic mineral sources and pathways as well as changes in the flux of both magnetic and non-magnetic components to the sediment. Thus in order to find an explanation to account for the full range of climatically modulated mineral magnetic variations represented here and in other deep-sea cores we need to consider the implications of Section 12.2 on the origin and flux of marine magnetic minerals.

Although tephra layers as such will give rise to changes in magnetic mineral concentration unrelated to climatic change, it is nonetheless possible that variations in the volcanic dust veil, discussed by Lamb (1970) and Kennett (1981) for example, have been of major significance in controlling both climatic change and the flux of magnetic minerals to marine sediments. Kennett, summarises many studies supporting a link between volcanic activity and climate and concludes that 'the general synchronism of the volcanic explosive episode and global climatic evolution is probably not coincidental'. However, in

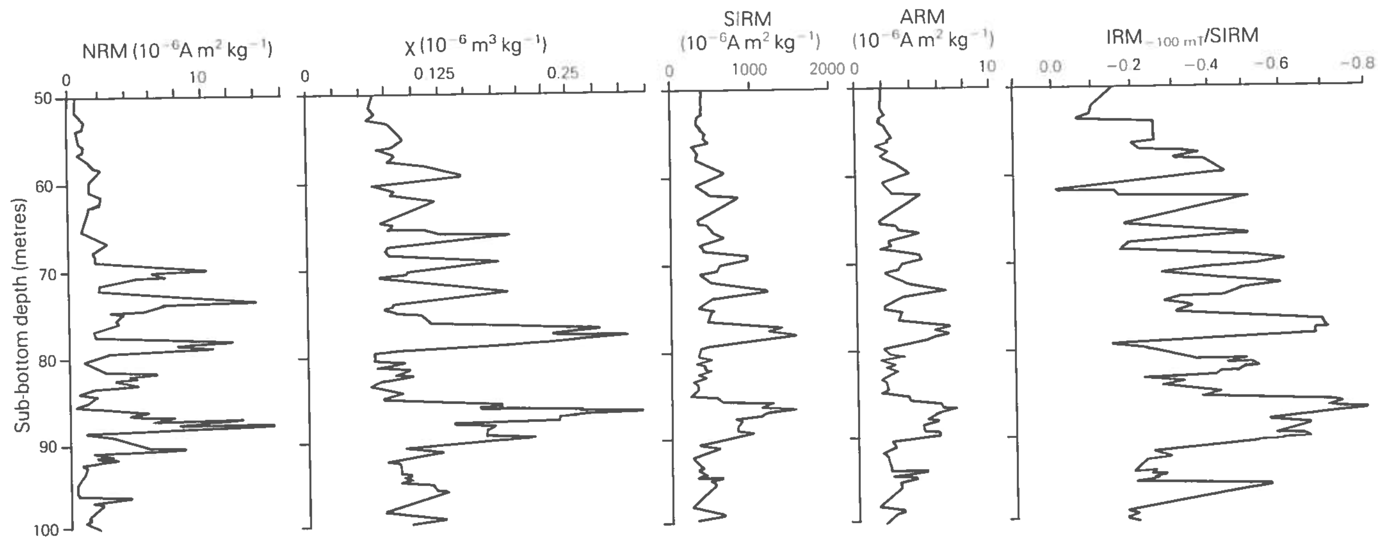

Figure 12.8 Mineral magnetic measurements from part of DSDP Core 514 from the South Atlantic (Bloemendal 1980).

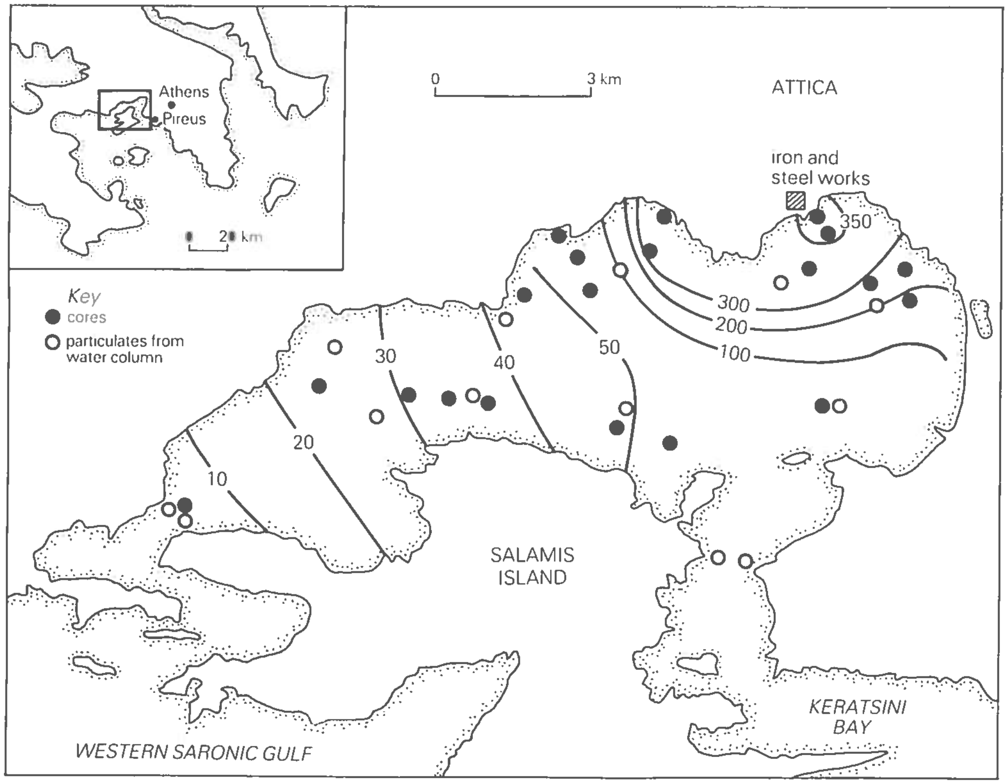

Figure 12.9 The Elefsis Gulf, Greece. The map locates both sediment coring sites and contemporary particulate sampling stations. The contours plot the decline in sediment surface mass susceptibility away from the iron and steel works on the north-eastern shore.

the light of the foregoing chapters, the magnetic evidence for the association between recurrent changes in mineralogy and climatic change is more readily interpretable as the result of shifts in weathering regimes, dust source areas and both atmospheric and oceanic pathways, than as the result of varying intensities of volcanic activity, though the evidence is far from conclusive. Moreover the importance of non-atmospheric inputs to the sediments considered here remains an open question.

Oldfield and Robinson (1985) tentatively conclude that the linkage between mineral magnetic variations and climatic change in deep-sea sediments probably arises from variations in continental sources, both volcanigenic and more especially erosive. The sedimentary expression of these variations must also be controlled by the effect of climate on the pathways by which the lithogenic minerals reach the sea bed, especially through changes in atmospheric circulation and ocean currents. Both Ellwood (1980) and Bloemendal (1980) suggest that changing bottom current velocities and directions have been responsible for mineral magnetic variations in their cores from the South Atlantic.

It is clear that we are some way from understanding all aspects of the mineral magnetic–climatic linkage mechanisms. Moreover these are likely to vary with latitude and with core location in relation to extensive deflated arid land surfaces as well as to patterns of movement by deep benthic water masses. Even so there are strong incentives for studying the linkages since a co-ordinated view could provide valuable new insights into the nature and expression of climatic change. At the very least, the results obtained so far confirm that mineral magnetic measurements in deep-sea sediments can provide not only a convenient and informative method of logging but one which is of major palaeoenvironmental significance.

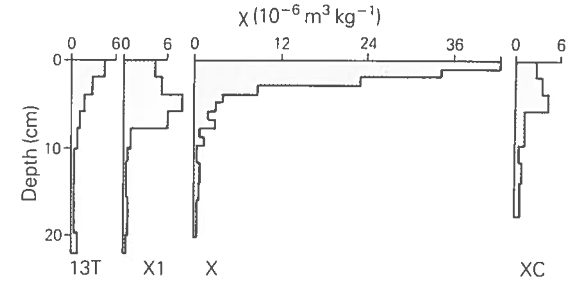

Figure 12.10 Down-core variations in mass susceptibility at sites close to the iron and steel works located in Figure 12.9.

12.5 Particulate pollution monitoring in coastal waters

A recent study by Scoullos *et al.* (1979) illustrates the potential value of magnetic measurements in coastal pollution monitoring where major sources are discharging high particulate concentrations which include ferrimagnetic oxides. The Elefsis Gulf is a shallow, almost enclosed embayment of the eastern Mediterranean close to the main areas of industrial development associated with Athens. The major point source of particulates is an iron and steel works near the northeastern corner of the Gulf, established from 1925 onwards and first brought into production in 1948 using scrap iron in electric arc furnaces. A first blast furnace became operational in 1963 and a second in 1972. The factory complex also includes a cokery, steel making and oxygen plants, a rolling mill and port facilities.

Figure 12.9 plots the location of sixteen sample stations from which monthly samples of particulates were taken on Millipore membrane filters of 0.45 μm at depths of 0, 10, 20 and 30 m during the period March 1977 to February 1978 inclusive. Also plotted are seventeen localities from which sediment cores up to 1 m long were taken using a pneumatic Mackereth (1969) minicorer. Susceptibility and SIRM were measured on all cores and whole core volume susceptibility scans were carried out on additional paired cores taken from Station 2 in 1980. Frequency-dependent (cf. quadrature) susceptibility was measured on suites of samples from Cores 2A and 3. The same figure shows the spatial distribution of sediment surface values of susceptibility in contour form and confirms the dominance of the point source at the iron and steel works.

Figure 12.10 shows down-core variations in susceptibility in several cores taken close to the industrial site. Core X was taken only 0.3 km from the artificial lagoon which receives the plant's effluents, the others at distances varying from 1 to 3 km. The recent nature of the peak in values is clearly indicated and has been subsequently confirmed by ^{137}Cs (Baxter, pers. comm.).

Figure 12.11 plots SIRM *v.* total particulate iron for all filter paper samples taken during January 1978. This was shown by means of chlorophyll-a analysis to be the period of minimum biomass. A strong linear relationship emerges suggesting that careful calibration and use of SIRM in the Elefsis Gulf may provide a rapid index of total particulate iron concentrations. The whole core volume susceptibility plots show that rapid scanning could be used as effectively as more time-consuming single-sample measurements to monitor variations in the magnetic content of near-surface sediments in the Gulf.

The plots of frequency-dependent (cf. quadrature) susceptibility (Fig. 12.12) show that whereas coercivity curves fail to differentiate clearly between the 'magnetite' deposited prior to the major effect of the iron and steel works from that postdating its

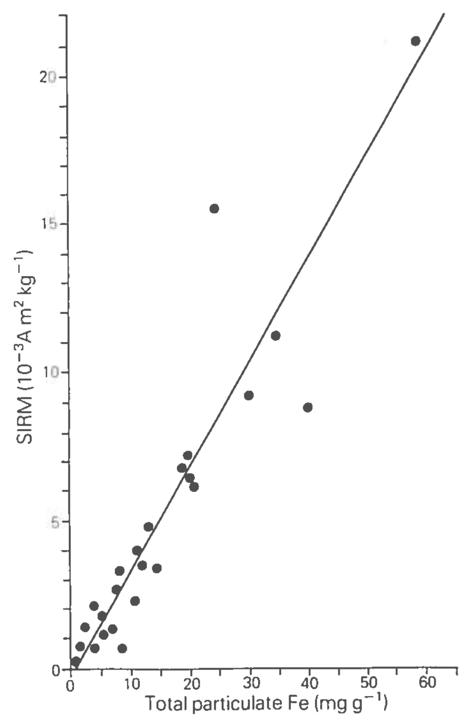

Figure 12.11 SIRM values versus total particulate iron concentration for filter samples from the water column taken from the Elefsis Gulf in January 1978.

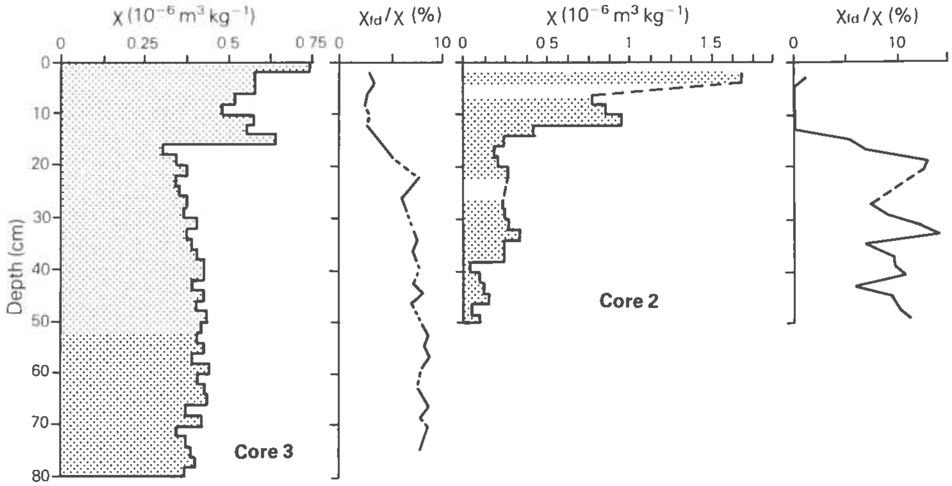

Figure 12.12 Mass susceptibility and percentage frequency-dependent (χ_{fd}/χ) susceptibility in two Elefsis Gulf cores.

development (Scoullos *et al.* 1979), χ_{fd}, expressed as a percentage of total susceptibility, declines sharply at the depth where the rapid rise in total susceptibility records the onset of the industrial development. This is consistent with a secondary soil-derived origin for the 'magnetite' in the lower part of the cores (cf. Ch. 8) and an industrial origin for the most recent material. The results as a whole, point to the value of this approach, especially in preliminary appraisals of both historical trends and current dispersion patterns. Not only is total particulate iron related to the magnetic measurements, but as in the case study noted in Chapters 9 and 11, so are other heavy metals including zinc. In considering the possible extension of this approach to other areas it is worth noting that Doyle *et al.* (1976) infer particulate pollution of the sediments of the eastern Gulf of Mexico from the identification of magnetic spherules over areas hundreds of kilometres from potential industrial sources.

12.6 Summary and conclusions

Mineral magnetic studies of marine sediments can already provide a basis for core correlation, for characterising diagenetic change, for establishing tephrochronologies, for reconstructing the course of climatic change and for near-shore particulate pollution monitoring. However, they are in their infancy and the range of illustrations and applications outlined here is unlikely to encompass anything approaching the full potential. Moreover, in each of the aspects touched on, e.g. palaeoclimatic implications and diagenetic effects, a great deal more critical work remains to be done before the crucial processes and mechanisms can be understood. This work calls for the *routine* association of mineral magnetic studies with oxygen isotope, palaeoecological, sedimentological and geochemical studies, both as precursors and complements to the traditional lines of enquiry. The speedy non-destructive nature of the techniques, the growing portability of the equipment and the development of sensors capable of scanning not only enclosed cores of any diameter but also free faces, makes the methodology ideally suited for this rôle. In addition, a broadening of the marine palaeomagnetist's concerns beyond the reconstruction and validation of palaeomagnetic directions and palaeointensities could lead to major advances in the quality and the spatial and temporal coverage of the mineral magnetic record. Only in this way will it be possible to build up a picture of the global patterns of variation upon which more general inferences of palaeoclimatic and geochemical value can be based.

[13]
Reversal magnetostratigraphy

"Footprints on the sands of time"

Longfellow
A psalm for life

13.1 Introduction

The geomagnetic field has been likened to a chronometer (Cox 1973). The magnetic clock, driven relentlessly by electric currents in the Earth's liquid core, oscillates back and forth switching between its two stable modes of a 'normal' state in which the Earth's magnetic field points north and a 'reverse' state in which the field points south. The geomagnetic chronometer lacks a regulator so it runs unevenly with switching intervals varying from a few thousand years to several tens of millions of years. The magnetic timing system of north–south flips, tags rocks as they form with its binary reversal code. Palaeomagnetists have unravelled the magnetic coding scheme so that palaeomagnetic remanence signatures can now be used to date sequences of lava flows and sediments.

Nature in her usual mischievous way has, in addition to providing such an elegant magnetic clock, laid down a rich assortment of false leads, red herrings, decoys and tricks which can bamboozle and deceive the impetuous palaeomagnetic investigator. The mechanics and workings of the magnetic chronometer are outlined in some detail below in order to alert the reader to the types of pitfalls to be encountered in magnetostratigraphy.

13.2 Geomagnetic signatures

The fidelity and accuracy with which geomagnetic polarity reversals have been recorded in rocks have varied greatly. This wide range in quality has resulted from the diversity of the geological processes which can record changes of the ancient field.

The process of remanence acquisition is generally simpler in igneous rocks than in sedimentary deposits. In some sequences of lava flows polarity reversals have been recorded in such great detail that individual reversals can be recognised by the characteristic pattern of their polarity transitions. Some sediments which have steadily accumulated, such as deep-sea sediments, have also recorded polarity changes in detail. Other palaeomagnetic records can however be rather misleading and so an understanding of the processes and mechanisms involved in palaeomagnetic remanence acquisition can be very helpful in assessing the usefulness of a palaeomagnetic record in a magnetostratigraphic study.

13.2.1 Origin of remanent magnetisation in igneous rocks

The natural remanent magnetisation of igneous rocks is generally the residual thermoremanent (Section 4.3.1) magnetisation acquired by the rocks on cooling from their molten state. This thermoremanence is locked into the igneous material at Curie and blocking temperatures of around 500 °C, well below the temperatures of between 1000 and 1250 °C at which the material crystallised out of the igneous melt. Many igneous rocks have been found to record accurately the direction and intensity of the geomagnetic field with uniform directions within a single igneous unit. Fine-grained igneous rocks with a thermoremanence held by small iron oxide crystals have been found to be particularly suitable for palaeomagnetic work.

LABORATORY EXPERIMENTS

The stable remanence of igneous rocks can be recreated in the laboratory by heating a sample of rock to above its Curie temperature and then allowing it to cool down again to room temperature (Koenigsberger 1938). Such artificial remanences are found to duplicate both the stability and intensity of the natural remanence. These types of experiment lend strong support to the weak field cooling origin of the natural remanence of igneous material and demonstrate that a thermoremanence direction is generally coincident with the field applied during cooling. Experiments also show that the rate of cooling is not important in affecting the remanence intensity at rates slower than 0.1 °C per second (Nagata 1953, Dodson & McClelland-Brown 1980).

Experiments on remanence acquisition over restricted temperature ranges have established that the addition law (Thellier 1946) of thermoremanence applies to natural as well as synthetic samples. In essence the addition law states that the partial thermoremanence of one temperature interval is independent of the remanence produced by field cooling in other temperature intervals. This basic property of thermoremanence is exploited in two important laboratory techniques. These are first, the method of isolating more than one component of remanence by partial demagnetisation (Section 6.5.2), and secondly the method of obtaining more than one estimate of ancient field intensity by the double heating palaeointensity technique (Thellier & Thellier 1959). Both methods involve cooling the sample from some chosen temperature, below the Curie temperature, to room temperature in a well controlled zero field (Section 6.6.3) environment. This cooling procedure results in isolating the remanence of magnetic grains with blocking temperatures above the chosen temperature, as grains with blocking temperature below the chosen temperature lose their remanence when cooled in zero field. The practical benefits of these types of experiment are that igneous rocks with quite complicated multicomponent magnetic histories, which may involve partial remagnetisation by geological events after the time of their formation, can still be used in geomagnetic and magnetostratigraphic studies.

One disturbing phenomenon observed in the laboratory is that a small percentage ($<1\%$) of natural igneous rocks acquire a thermoremanence antiparallel to that of the applied field (Nagata *et al.* 1952). Néel (1955) has described a number of physical mechanisms which could cause such a reverse thermoremanence. The two major mechanisms in natural materials are first the magnetic interaction of two different magnetic constituents in a rock, and secondly the interaction of spin moments of neighbouring sites within a crystal lattice. One might expect that such reversal behaviour in a natural remanence could easily be detected by simply testing the rock in the laboratory. However, nature manages to arrange events, such as later chemical change, which can mislead or confound the investigator and in practice self-reversals of some key magnetostratigraphic igneous rocks have gone undetected for many years (Heller 1980). On the positive side the annoying complexities of self-reversals were a major challenge to early magnetostratigraphic workers and the phenomenon spurred them on to improve field collections and laboratory methods. Their endeavours resulted in the establishment of the polarity timescale (Cox *et al.* 1963) which played a major rôle in the emergence of the global plate tectonic hypothesis which was to revolutionise much of the Earth sciences.

BASALTS

Basalts display a wide range of oxidation and alteration states and a corresponding range of magnetic mineralogies (Section 3.6.1). Distinctive changes of magnetic mineralogy with oxidation have been documented in numerous suites of basalts from igneous provinces around the world (Ade-Hall *et al.* 1971). Progressive high temperature (deuteric) oxidation of natural iron oxides has been classified into six stages from no oxidation (class I) to maximum oxidation (class VI). The oxidation series begins with uniform titanomagnetite grains, with well formed outlines, which crystallised directly out of the basaltic melt. As oxidation gets under way the titanomagnetite grains begin to show exsolution of magnetite and thin ilmenite lamellae. With further oxidation titanohaematite and sphene appear. Rutile and titanohaematite become more common as the final stages of oxidation are approached and finally, at the highest oxidation state, titanohaematite and pseudobrookite or ferrorutile completely replace the original titanomagnetite. The above oxidation succession charts mineralogical alterations which can occur as a basalt cools through temperatures of between about 800 and 500 °C. The varied magnetic assemblages produced by such high temperature oxidation can still pick up a geomagnetic signal, as they cool through their blocking temperatures, and can end up carrying excellent palaeomagnetic records. Class III rocks from the middle of the range of oxidation states often display a particularly strong and stable palaeomagnetic signal on account of the elongated shapes, small sizes and pure magnetite composition of their iron oxide grains.

Low temperature alterations, such as may arise from later hydrothermal events, can further complicate the high temperature sequence outlined above. Oceanic basalts extruded under water tend to be quenched and consequently have different bulk mineral magnetic properties from those of continental basalts. Many oceanic basalts display signs of extensive low temperature maghaematisation (Ade-Hall 1964, Ozima & Ozima 1971). Although the remanence of ocean basalts has been found to be predominantly stable a significant proportion also have appreciable soft magnetic components.

MAGNETISATION OF OCEAN CRUST

Following the invention of the proton precession magnetometer (6.6.1), oceanographic ships and aircraft exploring the ocean basins charted the intensity of the Earth's magnetic field by towing magnetometers behind them. Typical cruise profiles reveal magnetic anomalies with amplitudes of up to a few hundred nanotesla and widths of tens of kilometres. The anomalies have been found to have lengths of several thousands of kilometres and to be broken only when intersected by large oceanic fracture zones. The anomalies thus form a series of remarkably regular and continuous magnetic stripes which are found in all the major oceans, some stripes having been traced half way around the world (Sclater & Parsons 1981).

The explanation for the origin of the neat, regular pattern of the magnetic stripes came from combining the ideas of continental drift, sea-floor spreading and geomagnetic reversals. The idea is that the sea floor, as it forms continuously at ocean ridge crests becomes magnetised in the Earth's field, and the steady spreading of the ocean crust away from the ridges leads, as the Earth's field reverses polarity, to a regular magnetic pattern (Vine & Matthews 1963). The positive magnetic stripes formed when the field was normal, as it is today, and the negative stripes formed when the field was reversed. This link of oceanic anomalies with geomagnetic reversals allows ocean crust to be dated by magnetostratigraphic methods and conversely it allows continuous oceanographic magnetic field profiles to form the basis for a polarity reversal sequence (Heirtzler *et al.* 1968, Larson & Pitman 1972). The remarkable simplicity and symmetry of the movement of rigid plates on the Earth's spherical surface (Morgan 1968) has facilitated the magnetostratigraphic interpretation of oceanic magnetic anomalies, allowing the age of all of the oceanic crust to be deduced, together with surprisingly accurate and detailed reconstructions of the past positions of the world's major plates.

Curiously, despite the simplicity and beauty of the tectonic insights derived from oceanic magnetic stripes, the actual physical source of the magnetic anomalies remains in some doubt. The magnitude and shape of the anomalies most commonly lead oceanographers to model the magnetic source as being the uppermost 0.5 km of basaltic lavas and dykes with an average remanent magnetisation intensity of 10–20 A m^{-1}, although early block models tended towards a 1–2 km thick layer with an intensity of 5 A m^{-1}. Reasonably high Koenigsberger ratio (see Fig. 4.1) measurements from oceanic dredge and drill samples confirm that the magnetic anomalies result from a remanent rather than an induced magnetisation. However, the remanence intensity of recovered oceanic basalts as measured in the laboratory tends to be only around 3 A m^{-1}, that is roughly a half to a quarter of the modelled remanence intensity (Kent *et al.* 1978, Lowrie 1979). The dispersion of the remanence directions and the presumed restriction of even those moderate magnetisations to the pillow basalts (defined seismically as the upper 0.5 km (layer 2A) of the oceanic crust) have led to suggestions that the source of the stripe anomalies may lie deeper in the ocean crust. Rock magnetic and palaeomagnetic investigations indicate that the gabbroic rocks of layer 3 form possible important source regions (Dunlop & Prevot

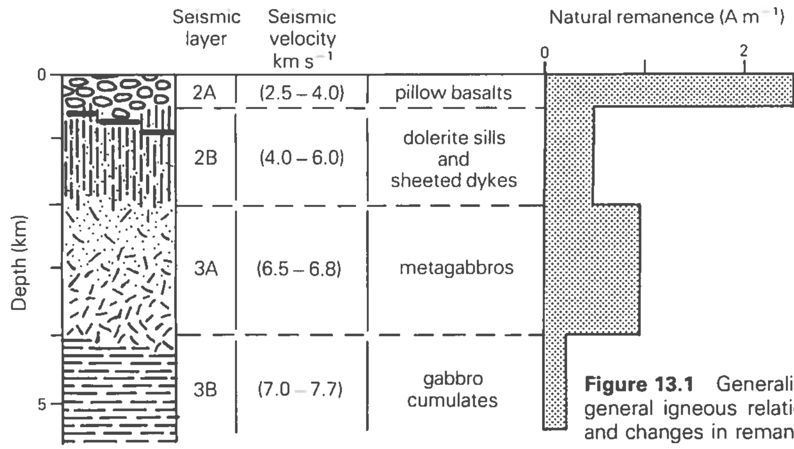

Figure 13.1 Generalised section through oceanic crust depicting general igneous relationships, seismic layers, seismic velocities and changes in remanent magnetisation with depth.

1982). Figure 13.1 illustrates the general magnetic zonation of the ocean crust envisaged as a result of these rock magnetic and palaeomagnetic investigations.

13.2.2 Origin of magnetisation in sediments

Many types of sediment have been found to contain excellent records of the past behaviour of the geomagnetic field. The exact mechanisms by which sediments acquire their natural remanent magnetism are, however, still poorly understood. This lack of understanding has arisen partly because of natural sedimentological complexities but also partly because of difficulties in interpreting the natural remanence acquisition in terms of laboratory studies.

Laboratory studies suggest that there are two basic mechanisms by which sediments can acquire a natural remanence (Section 4.3.1). One of these mechanisms leads to a remanence called a detrital remanent magnetisation. The other mechanism by which sediments can acquire a remanence is of a chemical nature and produces a remanence called a chemical remanent magnetisation. Two forms of chemical remanent magnetisation have been recognised. In one of these forms, named authigenic chemical remanent magnetisation, the remanence is produced by the growth of new minerals from solution. In the other form, named diagenetic chemical remanent magnetisation, the remanence results from the chemical alteration of pre-existing minerals.

LABORATORY EXPERIMENTS

Numerous experiments have been carried out in order to determine the properties of detrital remanent magnetisations, but very little laboratory work has been performed in connection with chemical remanent magnetisations (Verosub 1977).

Experiments using a range of sedimentary materials, such as crushed basalt fragments (Nagata *et al.* 1943), dispersed glacial clays (Johnson *et al.* 1948) and synthetic sediment (Irving & Major 1964), have shown that depositional detrital remanent magnetisation correctly records the declination of the applied field. However, the inclination recorded by this depositional process is consistently too shallow. This biasing effect to low inclinations has been named the 'inclination error' (King 1955). It can be explained in terms of the interaction of the settling magnetic particles as they touch the substrate. For example, the actions of nearly spherical particles rolling into depressions or disc-shaped particles rotating into the horizontal plane as they come to rest at the sediment/water interface are two processes in which the action of gravity can overcome the geomagnetic field alignment and so cause an inclination error (Griffiths *et al.* 1960). A further biasing effect has been observed in the laboratory to occur on dipping beds. Particles rolling down slope at the time of deposition distort the remanent inclination in an effect called the 'bedding error' (King 1955, King & Rees 1966). Finally, laboratory deposition in flowing water has been observed to cause errors in both declination and inclination. This 'current rotation effect' (Rees 1961) has been studied in experiments using flume tanks.

Other laboratory experiments such as stirring reconstituted sediments have been performed in order to study the properties of post-depositional detrital remanent magnetisation (e.g. Kent 1973). In contrast to the experiments on depositional detrital remanences these experiments have revealed no systematic deviations of the remanence directions from the laboratory field directions. In addition to this encouraging result concerning the ability of sediments to record field directions, post-depositional detrital remanent magnetisation intensities have been found to be linearly proportional to the intensity of the applied magnetic field (Kent 1973, Tucker 1980). So these laboratory experiments, coupled with theoretical calculations based on the alignment of magnetic particles perturbed by Brownian motion, indicate that both the direction and intensity of the ancient geomagnetic field can be accurately recorded by the process of post-depositional detrital remanent magnetisation.

The time involved in the acquisition of a post-depositional detrital remanent magnetisation and the effect of sediment consolidation have been investigated in the laboratory by the gradual redeposition of long columns of sediment. Experiments have been carried out involving steady day by day redeposition, by continuous addition of sediment, for periods of up to half a year. In one experiment using organic lake muds and a redeposition rate of 2 m a^{-1} (Barton & McElhinny 1979) an artificial reversal of the ambient magnetic field was quickly locked into the accumulating sediment. The remanent acquisition process certainly took place within a period of less than 2 days. In a similar experiment using deep-sea foraminiferal clay and a redeposition rate of 3 m a^{-1} (Lovlie 1974) the field reversal was not recorded in the redeposited sediment until 10 days after the reversal occurred at a burial depth of 10 cm.

Although it must be remembered that these experiments involve deposition rates some tens to thousands of times more rapid than those in nature, they indicate that in non-organic sediments consolidation of the sediment matrix is important in fixing the magnetic particles in position. In organic-rich sediments, on the other hand, it seems that the mobility of the magnetic particles can be restricted soon after deposition, possibly within months in some sediments, by the action of organic gels (Stober & Thompson 1977, 1979).

NATURAL REMANENCE OF DEEP-SEA SEDIMENTS

The widespread correlation of faunal zonations with palaeomagnetic reversal sequences in deep-sea sediment cores demonstrates that the acquisition of remanence by many deep-sea sediments is more or less contemporaneous, at least in terms of the resolution of dating, with deposition (Opdyke 1972). However, as many deep-sea sediment cores which exhibit a coherent pattern of magnetisation are bioturbated near the surface, their remanence has probably resulted from a post-depositional mechanism with its accompanying magnetisation delay. The similarity of the mean inclination of the remanent magnetisation of recent marine sediments to the local geomagnetic field inclination further suggests that the predominant remanence in oceanic materials is a post-depositional remanence rather than a depositional detrital remanence.

The depth of active bioturbation has been estimated in deep-sea sediments by studying the redistribution of ash layers and microtektite horizons and by investigating the mixing of radioactive contaminants in surface samples (Ruddiman *et al.* 1980). These various techniques show that surface material can be mixed with subsurface materials to depths ranging from 10 to 60 cm. Possible effects of variable bioturbation on a polarity subzone are illustrated in Figure 13.2. These effects can range from boundary displacements to the elimination or spurious addition of short polarity subzones.

The range of magnetic iron and manganese oxides, hydroxides and sulphides found in marine sediments has been described by Haggerty (1970) and by Henshaw and Merrill (1980), and Chapter 12 outlines their sources and significance in marine sediment

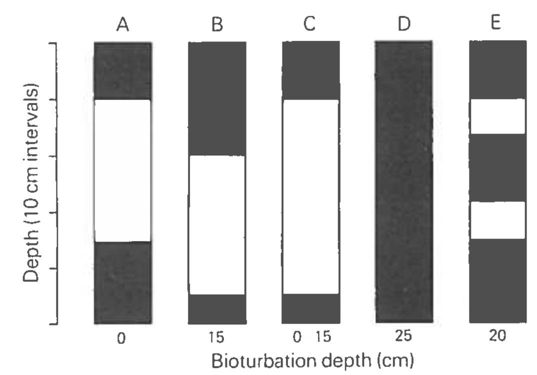

Figure 13.2 Possible effects of bioturbation on a single polarity subzone. A perfect palaeomagnetic recording of the subzone is shown in sediment column A. Distorted recordings of the same subzone are shown in columns B to E. The minimum depths of faunal activity required to create each effect are listed beneath each column. (A) no bioturbation; (B) constant bioturbation; (C) intermittent surface bioturbation; (D) constant bioturbation after a period of inactivity; (F) intermittent subsurface bioturbation. (After Watkins 1968.)

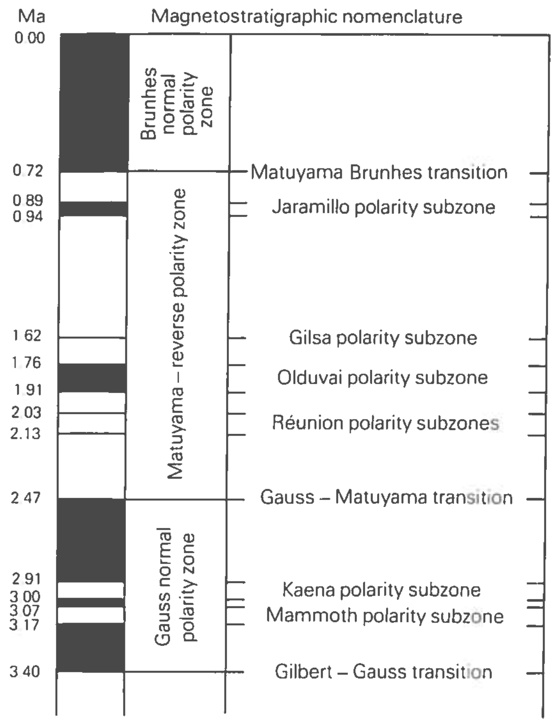

Figure 13.3 Magnetostratigraphic nomenclature as applied to the past 3.5 million years. Ages of boundaries derived by McDougall (1979) by combining K–Ar age determinations from many localities around the world using the direct timescale approach. Some debate still remains over the existence and age of short subchrons such as the Gilsa subchron and possible reversed polarity subchrons in the Brunhes zone. Normal polarities are depicted by solid shading.

stratigraphy. The loss of quality and resolution of the magnetic stratigraphy at depth in some cores, particularly from the north central Pacific, is attributed to a chemical remanence replacing the primary post-depositional detrital remanence (Kent & Lowrie 1974).

PARTICLE REALIGNMENT

The original palaeomagnetic record of a sediment can be distorted by later physical movements of the micron-sized particles which carry the natural remanence. Particle realignment can take place rapidly if the sediment is disturbed or more slowly over hundreds or thousands of years. On the thousand year timescale, it can gradually lead to the resetting of reversal patterns or the reduction of secular variation features while on a timescale of years, prolonged core storage can lead to marked changes in remanence direction. Magnetic particle realignments occur within minutes when the sediment is disturbed, for example by drying (Granar, 1958). Such physical rotations of magnetic particles and their associated remanence realignments can be difficult to detect because the magnetic coercivity and character of the new remanence is often very similar to that of the original remanence.

The effect of drying is surprisingly important and rather complicated (Johnson *et al.* 1975). Drying inorganic sediments with an original high water content generally fixes their remanence (Henshaw & Merrill 1979). In contrast drying organic-rich lake sediments can lead to a reduction of up to 50% of their remanence (Stober & Thompson 1977, 1979). In further contrast, drying of some sediments can lead to the acquisition of a new remanence. These three distinct effects are all thought to be related to the physical mobility and realignment of magnetic carriers. The first effect, of remanence stabilisation, can occur naturally due to either evaporation or compaction processes. Some laboratory experiments (Verosub *et al.* 1979) suggest that stabilisation can be produced by a remarkably small change in water content through a critical limit. The precise critical water content appears to depend both on the size of the magnetic particles and the sediment matrix (Payne & Verosub 1982), but to be about 75% for many deep-sea sediments with compaction being the controlling factor. The second effect, of remanence loss, has been found in some lake sediments steadily to reduce the natural remanent intensity as the water content is decreased. A likely mechanism is that during dehydration the small magnetic carriers are physically pulled towards the larger sediment particles by surface tension effects. The net remanence is thus

Table 13.1 Recommended terminology for magnetostratigraphic polarity units.

Magnetostratigraphic polarity units	Approximate duration in years	Geochronological equivalent	Chronostratigraphical equivalent
polarity subzone	$10^4 - 10^5$	subchron	subchronozone
polarity zone	$10^5 - 10^6$	chron	chronozone
polarity superzone	$10^6 - 10^7$	superchron	superchronozone

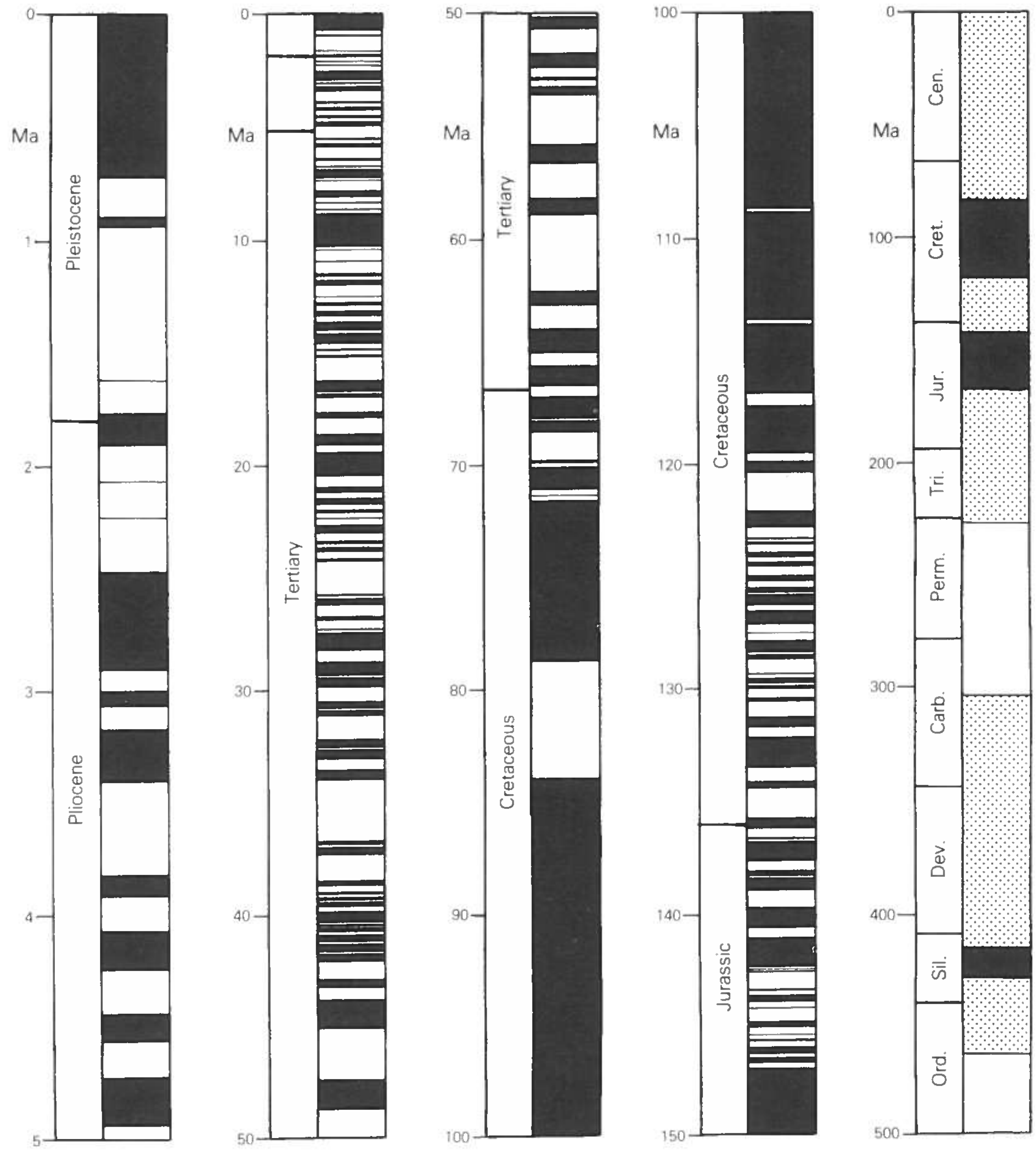

Figure 13.4 Polarity sequence of the geomagnetic field extended to 150 million years by inference from marine magnetic anomalies (modified from Lowrie & Alvarez 1981, Channell *et al.* 1982). The right-hand column divides the last 500 million years of geological time into normal and reverse polarity superchrons and disturbed (mixed) superchrons (light shading). Normal polarities are depicted by solid shading.

reduced in intensity, but the random surface tension forces do not lead to any change in remanence direction. A similar effect occurs in freezing lake sediments when disturbance of the magnetic carriers reduces the remanence intensity while the remanence direction remains largely unaltered. The third phenomenon, of the production of a new stable remanence by drying, has been attributed to sediment fabric disturbance which allows the magnetic particles to continue to rotate, at unusually low water contents, into the new field direction.

Two additional effects of drying sediments are that chemical or biochemical change associated with slow drying can reduce magnetic susceptibility (e.g. Bloemendal, 1982) and that the drying of small samples can distort their shape. This drying distortion may also change the direction of their remanence. The distortion of large blocks of sediment during drying or compression has, however, very little effect on the net remanence, because large blocks tend to crack along fracture sets which leave the remanence direction unaltered (Kodoma & Cox 1978).

13.3 The geomagnetic polarity timescale

TERMINOLOGY

The polarity timescale, although it has evolved to an extent where it is undoubtedly capable of being the foundation of an understanding of diverse global

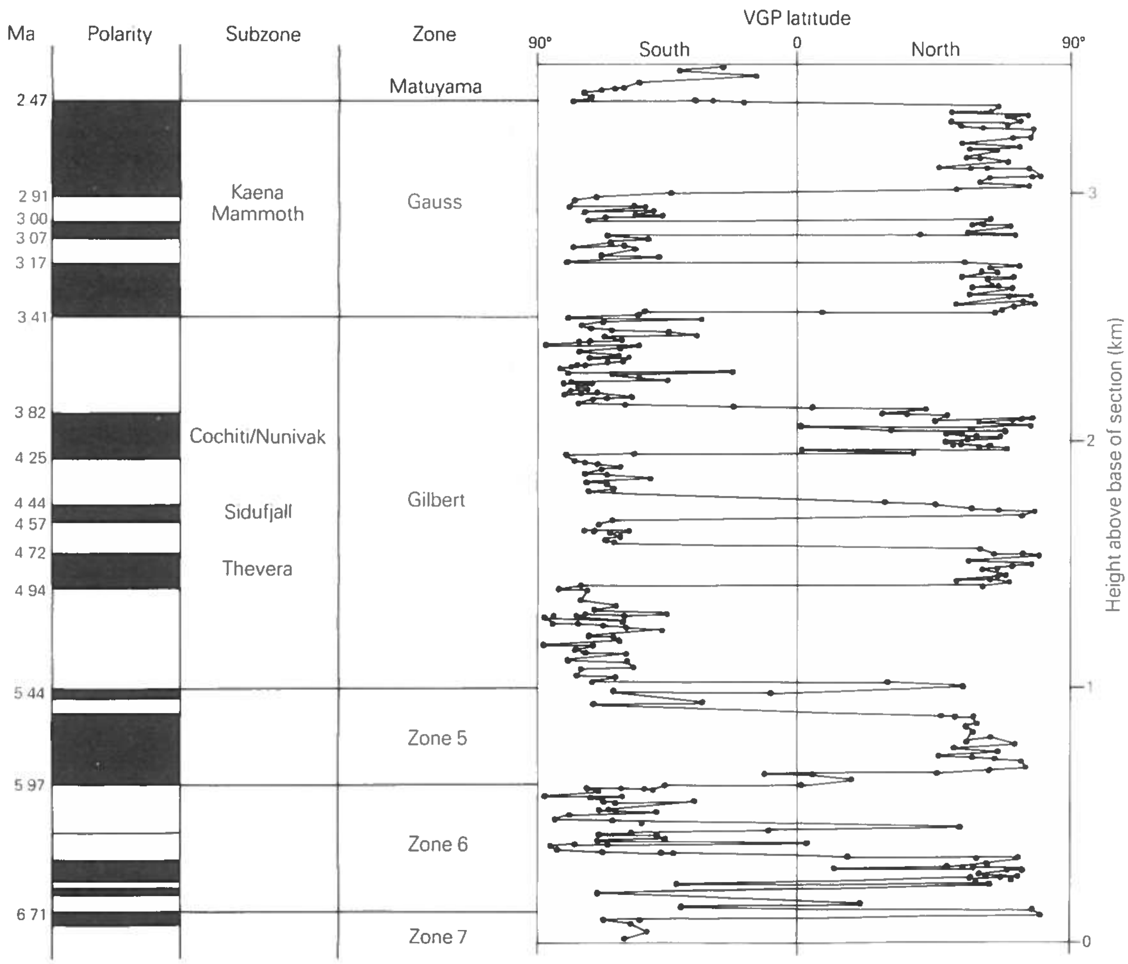

Figure 13.5 Palaeomagnetic results and polarity timescale recorded in a 3400 m sequence of lava flows in Iceland. Normal polarities are depicted by solid shading (after McDougall *et al.* 1977). The probability of a deep sea sediment sequence being datable by palaeomagnetic study is assessed at 20% by Stupavsky and Gravenor (1984).

phenomena, is still being developed and refined (Watkins 1972, Irving & Pullaiah 1976). A polarity nomenclature system is thus needed which incorporates the early magnetostratigraphic conventions and usages but which can also be developed in a coherent manner, incorporating modifications and resolving inconsistencies, confusion and controversies. A sub-commission of the International Commission on Stratigraphy has accordingly recommended a system of terms and hierarchies for magnetostratigraphic use (Table 13.1). The sub-commission discourages the use of the previously applied magnetostratigraphic terms of epoch, event and interval on account of conflicts with established stratigraphic terminology, and because the terms are not strictly physically appropriate. Furthermore, it is

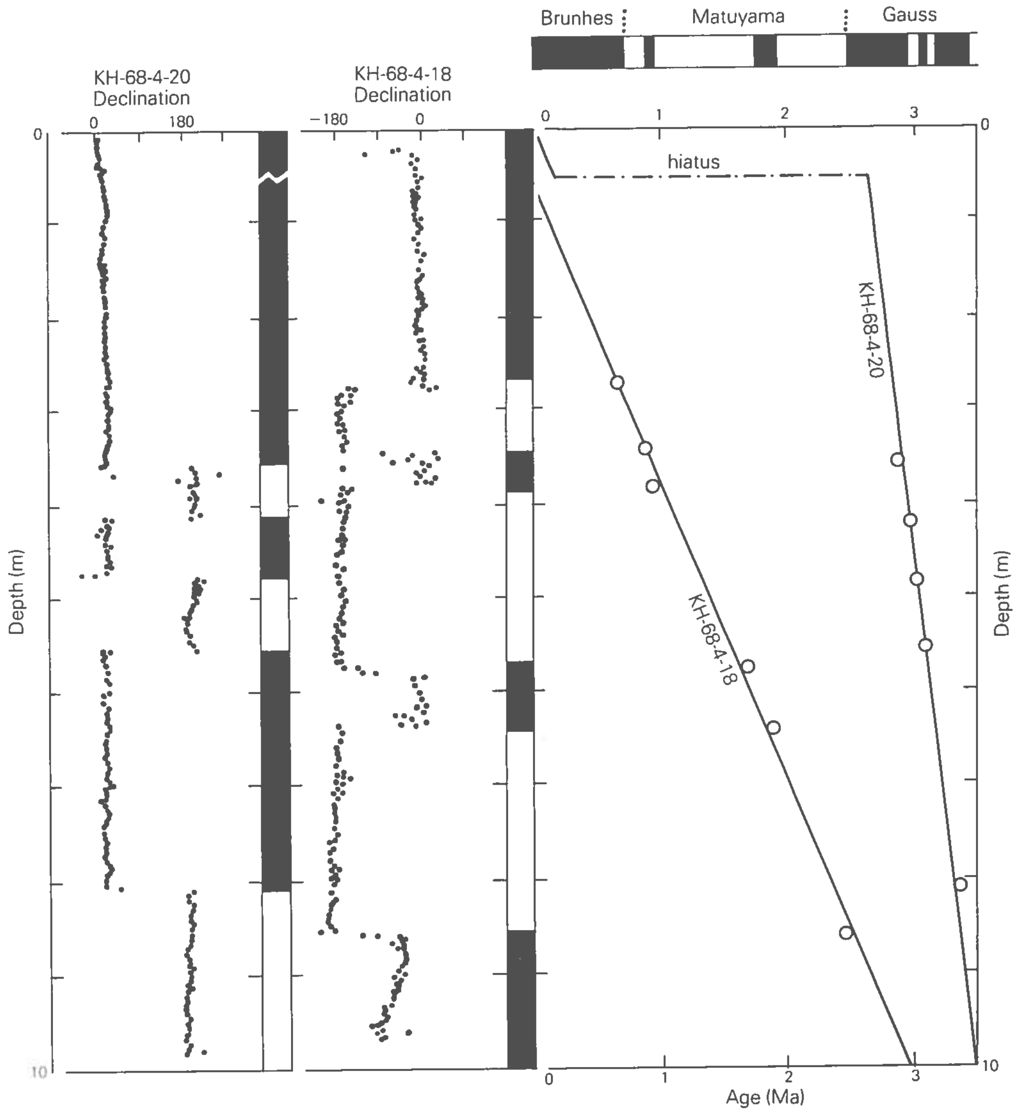

Figure 13.6 Palaeomagnetic NRM declination directions in two west-central equatorial Pacific sediment cores. The polarity zonation of each declination record is shown and correlated (open circles at right) with the polarity timescale of Figure 13.3 to give sedimentation rate estimates. A long hiatus covering the whole of the Matuyama reversed polarity chron and a large proportion of the Brunhes normal polarity chron is inferred for Core KH-68-4-20 (data from Kobayashi *et al.* 1971).

now realised that there is no fundamental geophysical difference between a polarity event and a short polarity epoch. The preferred terms for magnetic intervals are polarity zone, subzone and superzone (e.g. Harland *et al.* 1982). Figure 13.3 illustrates the use of this revised magnetostratigraphic nomenclature for the past 3.4 Ma. Geographically derived names are preferred for major units although names, in use, derived from distinguished contributors to the science of geomagnetism, are also acceptable.

Magnetic polarity units have been established in two quite different ways: (a) by combining laboratory determinations of the remanent magnetisation direction of rock samples or cores with radiometric or biostratigraphic age determinations, and (b) through the shipboard magnetometer profiles of the intensity of oceanic magnetic field anomalies (e.g. Irving and Pullaiah, 1976). The first approach fits neatly into conventional stratigraphic classification procedures such as the use of type localities and type sections. The second ocean survey approach is not handled so easily. The practical oceanic anomaly numbering system, which has proved so useful for reconstructing plate motions and the development and history of ocean basins, although falling outside conventional stratigraphic practice, is likely to remain as an extremely useful quasi-stratigraphic method. Figure 13.4 illustrates a polarity timescale for the last 150 Ma erected by combining the two approaches.

0–5 Ma DIRECT RADIOMETRIC APPROACH

This is the original and most elegant method of establishing a polarity timescale. It uses radiometric ages and magnetic polarities of rock samples from any part of the world. The ages and polarities are simply plotted next to each other in order to form the polarity scale. The polarity timescale was developed and refined through the 1960s principally by the work of two research groups based in California and Australia. Cox (1973) described the timescale developments as 'a rather lively competition, resembling a long-distance chess game in which the two sides communicated via letters to the journals *Nature* and *Science*'.

The direct radiometric method is limited by the

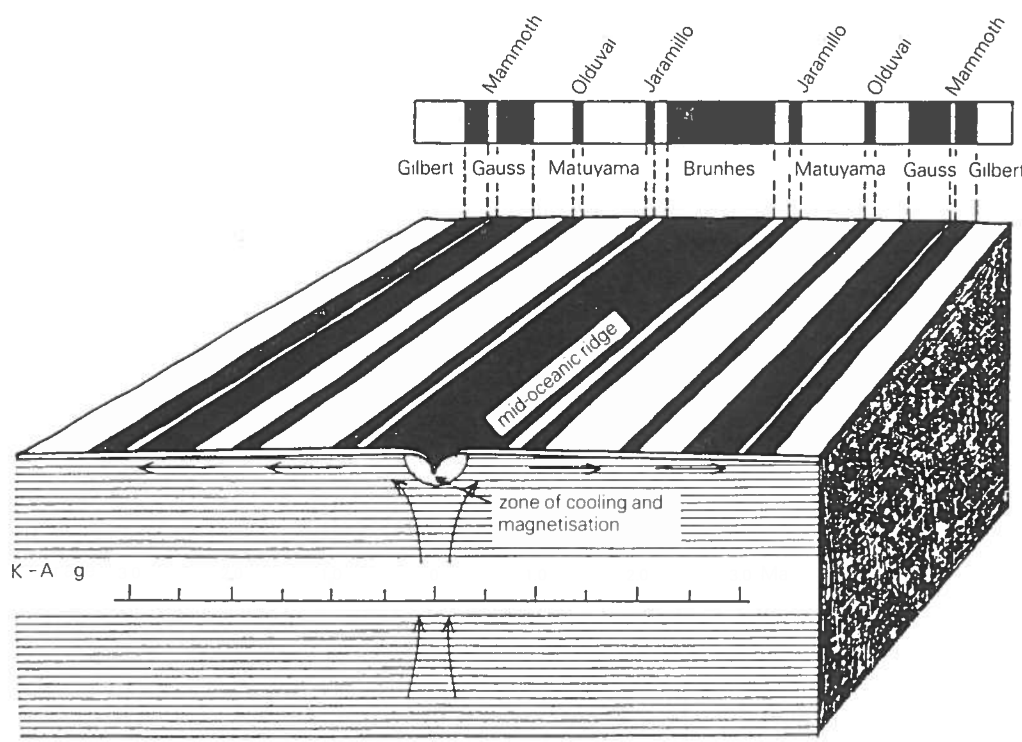

Figure 13.7 Schematic representation of the principle of sea-floor spreading and reversals of the Earth's magnetic field as proposed by Vine and Matthews (1963) (after Allan 1969).

accuracy of dating techniques. Currently the K–Ar dating precision is around 3% for rocks a few million years in age. As the average length of time between polarity inversions, during the last 5 million years, has been about 200 000 years it works out that the possibilities of recognising individual polarity subzones earlier than four or five million years ago by K–Ar dating are extremely limited, even with prodigious amounts of data (Dalrymple 1972) because the resolution of the dating is inadequate.

0–20 Ma STRATIGRAPHIC APPROACH

The polarity timescale of the past 5 million years can be extended further back in time by studying stratigraphic sections rather than isolated rocks. Figure 13.5 displays the palaeomagnetic results from a thick sequence of lava flows on Iceland. The reversal pattern can easily be seen to match and extend the older part of the polarity timescale of Figure 13.3. Although radiometric dating of the older lavas presents some difficulties, analyses from the whole sequence allow an overall age scale to be assigned to the polarity transitions (McDougall *et al.* 1977). Deep-sea sediments present a further geological sequence which can be used in the stratigraphic approach to erecting an extended polarity timescale, although there are some difficulties in obtaining long, old, unbroken records (Hammond *et al.* 1974, Opdyke *et al.* 1974). Figure 13.6 exhibits palaeomagnetic results from Pacific cores illustrating the value and quality of the magnetostratigraphic information which can be found in ocean sediments (Kobayashi *et al.* 1971). A very clear reversal magnetostratigraphy has also been found preserved in Chinese loess (Heller and Liu, 1982).

0–170 Ma MARINE ANOMALIES APPROACH

By dating magnetised ocean crust a polarity timescale can be generated from the oceanic linear magnetic anomalies (Fig. 13.7) back to the time of formation of the oldest ocean crust of around 170 million years. The age of the magnetised ocean crust layer 2A can be assessed by: (a) assuming that sea-floor spreading has occurred at the same steady rate as during the past few million years (Heirtzler *et al.* 1968), (b) equating the age of the magnetised crust with that of the biostratigraphic age of the immediately overlying sediments (Larson & Pitman 1972, Winterer 1973) and (c) matching the magnetic anomaly sequence to dated magnetostratigraphic-type sections on land (La Brecque *et al.* 1977, Lowrie & Alvarez 1981, Channel *et al.* 1982). By using these three dating methods marine magnetic anomalies have been employed in constructing the polarity reversal sequence of Figure 13.4 which stretches back continuously from the present day to 150 million years ago. The marine magnetic anomalies have been invaluable in reconstructing elements of the geological history of the oceans, such as the evolution of the Atlantic ocean depicted in Figure 13.8 (Phillips & Forsyth 1972).

0–600 Ma QUIET ZONE APPROACH

The key to magnetostratigraphy before 170 Ma lies in the long, quiet polarity superchrons. Old periods of frequent field reversal are difficult to use in magnetostratigraphy because the geomagnetic details are difficult to resolve reliably through either the direct radiometric or the stratigraphic approaches. The pattern of quiet and disturbed magnetic superchrons of Figure 13.4, however, has the potential for magnetostratigraphic correlation on an intercontinental scale. Correlations between the rocks of six continents have been proposed on the basis of the most spectacular of these quiet zones, the Permo-Carboniferous reversed superchron (previously known as the Kiaman reversal).

13.4 Polarity transitions

Occasionally palaeomagnetists have managed to obtain records of the magnetic field when it was actually in the act of changing polarity (e.g. Fig. 13.9). Some of these polarity transitions have been studied in sufficient detail to permit mathematical models of the behaviour of the field during the polarity switch to be developed (Fuller *et al.* 1979). Estimating the time taken for the field to change polarity is extremely difficult, but it is generally taken, on the basis of the thickness of deep-sea sediment layers recording intermediate palaeomagnetic directions across reversal boundaries and on the proportion of intermediate directions found in lavas, to be around 5000 years. Clement and Kent (1984) in a very detailed study of the Matuyama-Brunhes polarity transition in seven deep-sea cores from the Pacific Ocean have found evidence that the duration of this transition at mid-latitudes was twice that at equatorial latitudes. During a polarity transition the field intensity typically drops to one-fifth of its pretransition value. On the basis of relative field intensity measurements

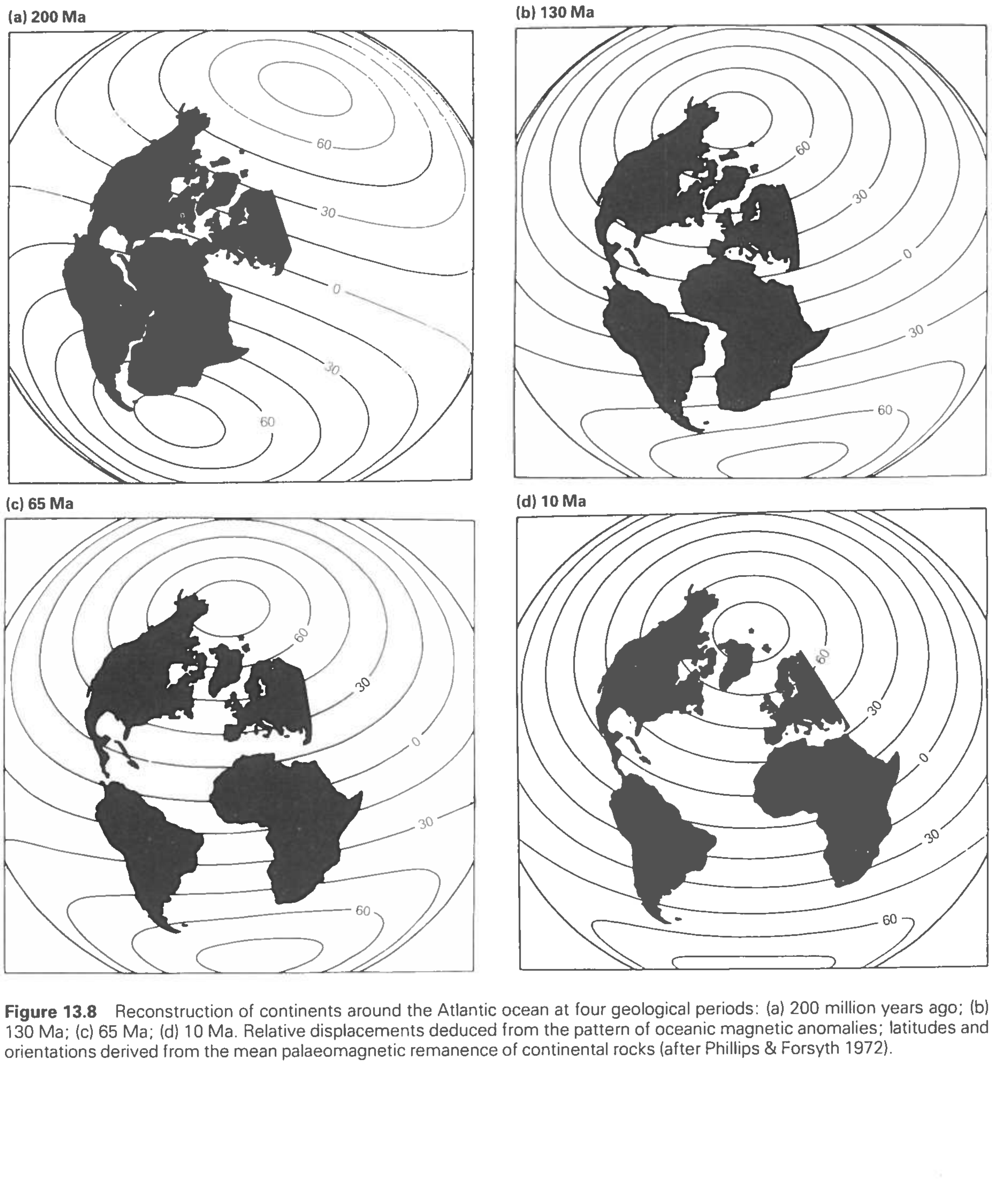

Figure 13.8 Reconstruction of continents around the Atlantic ocean at four geological periods: (a) 200 million years ago; (b) 130 Ma; (c) 65 Ma; (d) 10 Ma. Relative displacements deduced from the pattern of oceanic magnetic anomalies; latitudes and orientations derived from the mean palaeomagnetic remanence of continental rocks (after Phillips & Forsyth 1972).

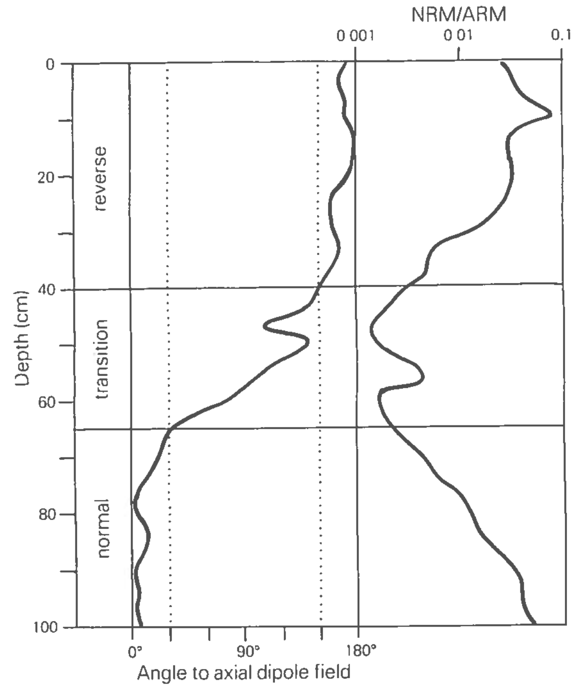

Figure 13.9 Diagrammatic representation of a polarity transition captured by the remanence of steadily accumulating sediments. Transitional directions are defined as vectors which deviate from the local axial dipole field directions by more than 30°. Low NRM/ARM ratios largely indicate times of low geomagnetic field intensity, but may also reflect low NRM intensities caused by unresolved dual component remanences.

in sediments it has often been stated that such field intensity changes persist for a longer time than the field direction changes. More recent palaeointensity data on lavas, however, suggest that the direction and intensity change more or less synchronously (Coe *et al.* 1983) and the early drop and late recovery of intensity commonly observed in sediment transition records is more likely to be a remagnetisation effect. Palaeomagnetic records of certain polarity transitions have now been obtained from different parts of the world. These records show that the predominantly dipole nature of the stable geomagnetic field is not preserved during the transitions. Two approaches to modelling the more complicated field structure during polarity changes are (a) by axial quadrupole and octupole field configurations and (b) by transitions through standing non-dipole fields. More observations of polarity transitions are needed to evaluate the applicability of each model and in particular more transition records from the southern hemisphere are needed to distinguish between transition fields dominated by quadrupole terms and those dominated by higher order axisymmetric terms (Fuller *et al.* 1979).

13.5 Summary

Since the first polarity timescales were constructed in the early 1960s, reversal magnetostratigraphy has developed into an important surrogate dating technique. Use of the polarity timescale played a key rôle in the discovery of sea-floor spreading and the development of plate tectonics. Probably all polarity intervals younger than 150 million years which lasted for longer than 0.1 million years have been discovered. Additional shorter polarity intervals are continually being found as more detailed geological records of particular time periods are discovered and investigated. The most recent firmly established global polarity reversal was the Matuyama to Brunhes transition which took place 700 000 years ago.

Geomagnetic reversal sequences are recorded in the thermal remanent magnetisation of igneous rocks, the chemical remanence of red beds and loess, and the detrital remanence of fine-grained sediments such as clays and limestones. This broad range of remanence and rock types makes reversal magnetostratigraphy a widespread chronological and correlation tool.

Details of geomagnetic behaviour during the reversal process are beginning to be unravelled. The characteristics of individual reversal transitions can be seen to be varied. Some reversal transitions display exceedingly complex changes with pronounced non-dipole, non-axisymmetric behaviour, while other transitions have performed simpler more axisymmetric switches between polarity states.