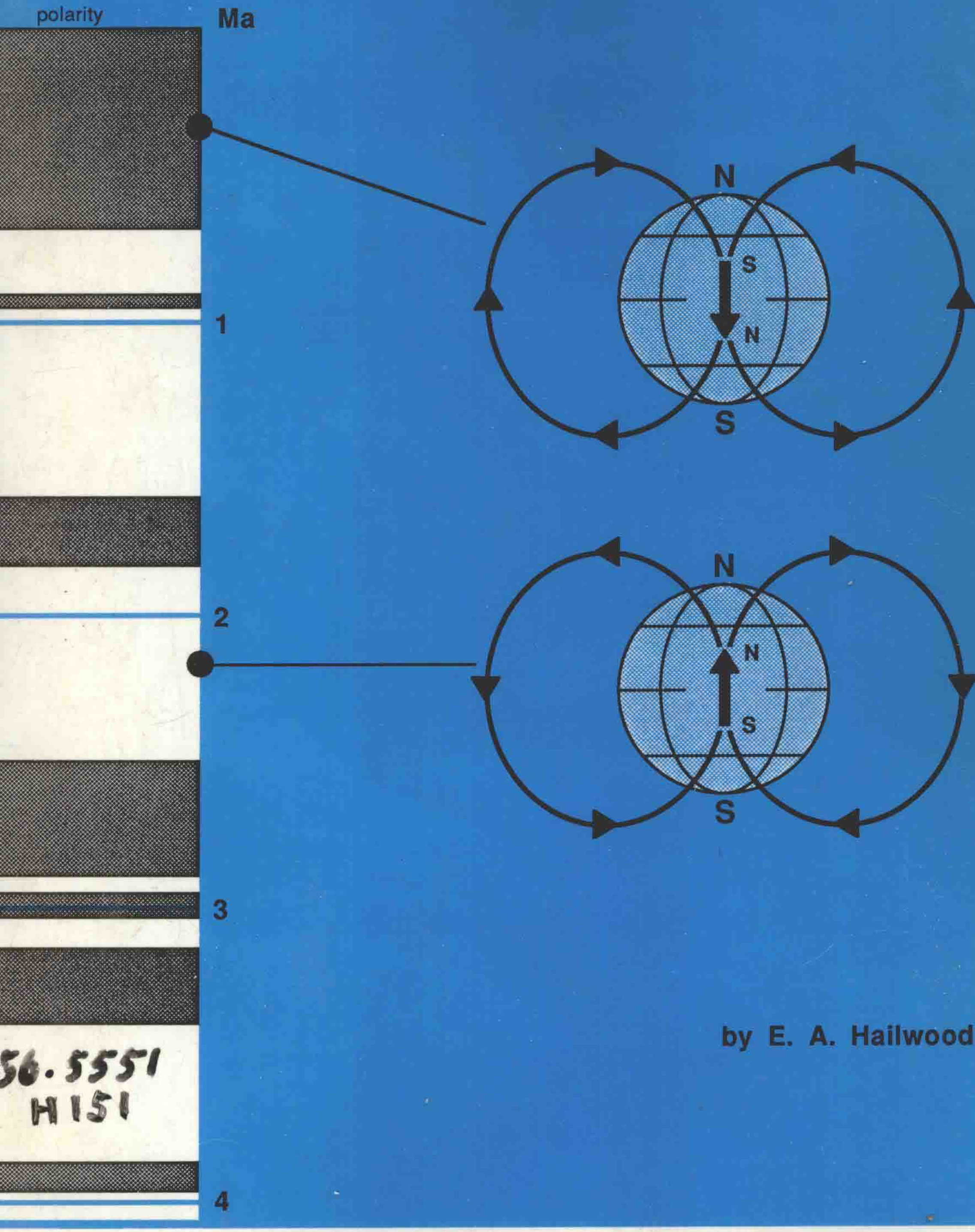
MAGNETOSTRATIGRAPHY



by E. A. Hailwood

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Magnetostratigraphy

E. A. HAILWOOD

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Contents

SUMMARY

The fact that rocks carry a weak but measureable permanent magnetism that provides a record of the nature of the geomagnetic field in the past has been known since the end of the last century. The potential stratigraphical value of this record was soon recognized, but attempts to put it to geological use were inhibited by the difficulties of measuring the weak magnetism of sedimentary rocks and a lack of understanding of the variation of the geomagnetic field with time.

In the early 1960s it was first demonstrated clearly that the Earth has reversed the polarity of its field at irregular intervals throughout geological history, and that a reliable record of the succession of polarity changes is provided in the permanent magnetism of sequences of sedimentary and igneous rocks and also in the patterns of magnetic anomalies produced in the ocean basins by sea floor spreading. This discovery paved the way for development of the magnetic polarity time scale, which has provided an important global time framework. Developments in instrumentation over the past two decades have permitted the application of magnetostratigraphic studies to virtually all rock types. This subject has now taken its place alongside biostratigraphy and radiometric dating as one of the standard branches of stratigraphy.

1. THE FUNDAMENTALS OF PALAEOMAGNETISM

The science of biostratigraphy is founded upon the identification of evolutionary changes in fossil organisms to provide a basis for subdividing geological time. In a similar way, magnetostratigraphy depends upon the study of the fossil (or remanent) magnetism of rocks to identify patterns of changes in the earth's magnetic field, which also can be used to subdivide the geological past. In general the biostratigraphical record is more definitive than the magnetostratigraphic record, since particular assemblages of organisms are normally characteristic of a restricted interval in geological history, whereas particular geomagnetic field configurations may have recurred a great many times in the past. However, magnetostratigraphy has the particular attribute that the geomagnetic field is a global phenomenon, and although short-term variations (the *secular variation*) may occur on a geographically localized scale, the longer-term changes, particularly polarity reversals, occur essentially simultaneously over the whole surface of the earth. Therefore they provide a potential basis for identifying true 'time planes' in geological sequences. In contrast, the problems of possible facies control or palaeoenvironmental dependence of biostratigraphical zonal boundaries are well known.

If characteristic sequences of polarity reversals are observed in different sections through a particular geological formation, then, under favourable circumstances, the matching of these sequences can provide a basis for geological correlation between the sections, even in the absence of biostratigraphical data. More commonly, magnetostratigraphy is used to provide an extension to conventional biostratigraphical techniques, rather than a substitute for them. For example, if a given sedimentary succession is to be correlated with a 'standard' geological time scale using magnetostratigraphy, then some biostratigraphical control is normally necessary to decide which parts of the polarity reversal sequence observed in this succession should be matched with which parts of the polarity reversal time scale. Once this approximate matching based on biostratigraphy has been carried out, a refined correlation of specific points in the succession (the individual polarity reversal horizons) with the polarity reversal time scale can be obtained, and a numerical age then assigned to these points.

The earth's magnetic field

Since magnetostratigraphy is based upon the identification of records of past changes in the geomagnetic field it is important to understand the basic nature of the field and its fluctuations with time.

The earth's present magnetic field approximates to that of a uniformly magnetized sphere, or a magnetic *dipole* located at the centre of the earth. The dipole axis is inclined at an angle of about 11.5° to the geographical (rotational) axis (Fig. 1a). This model accounts for approximately 80% of the main field, the remaining 20% having

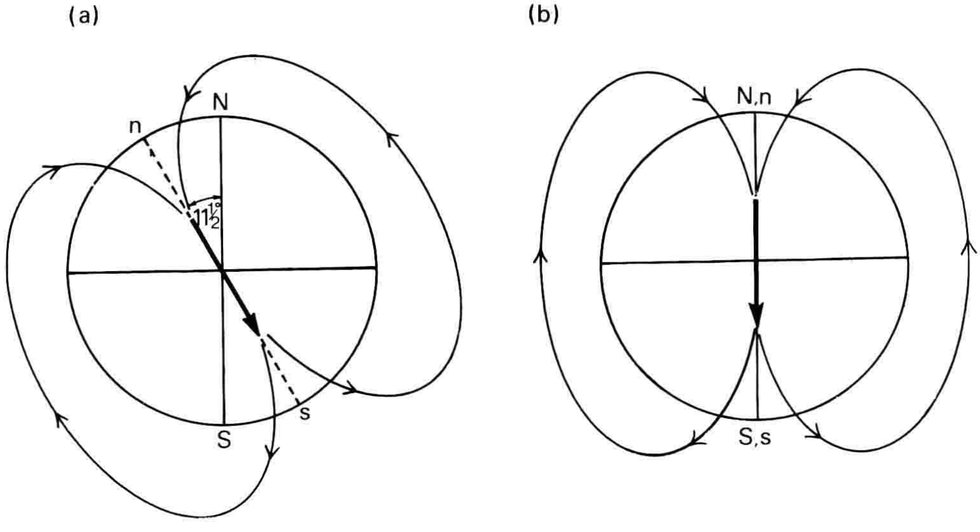

FIG. 1. (a) Model for present day geomagnetic field. Best fitting magnetic dipole axis is inclined at about 11.5° to the geographical axis. (b) Model for time averaged geomagnetic field (over $c.10^4$ years). Best fitting magnetic dipole axis coincides with the geographical axis. N and S are the geographical poles; n and s are the magnetic poles.

a more complex non-dipole nature. It is known that the orientation of this best-fitting dipole axis varies with time. Palaeomagnetic data indicate that its position corresponds closely with that of the geographical axis when averaged over periods of a few thousand years (Fig. 1b). This is referred to as the *axial geocentric dipole model*.

The source of the main part of the geomagnetic field is known to be the earth's core, but predicted core temperatures preclude any possibility of materials within this region of the earth carrying a permanent (or 'remanent') magnetism. Therefore the geomagnetic field must be maintained continuously by processes operating within the core, and it is believed that this involves a '*self-exciting dynamo*' mechanism which arises from convective motions of the hot, electrically conductive fluid within the outer core. As a result of the *dynamic* processes involved in the creation of the geomagnetic field, it is to be expected that the latter will vary with time, and it is the record of these variations throughout geological history, preserved in the permanent magnetism of rocks formed at the earth's surface, that form the basis of magnetic stratigraphy.

Variations of the field with time

The *direction* of the geomagnetic field vector, $\bar{F}$, at any point on the earth's surface is defined by two angles (Fig. 2), the *declination* (D, specified relative to geographical north), and the *inclination* (I, specified relative to the horizontal, and regarded as positive if the field is downward-directed and negative if upward-directed).

Observatory records, and studies of the remanent magnetism of rocks, soft sediments and archaeological materials, indicate that the geomagnetic field fluctuates in both direction and intensity on a number of different time scales. Determinations of ancient geomagnetic field *intensity* values are difficult, and often of uncertain

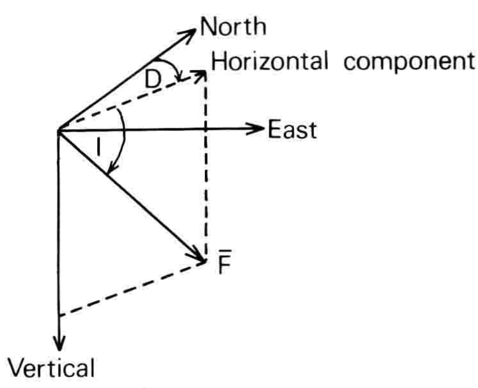

FIG. 2. Specification of direction of geomagnetic field vector, $\bar{F}$, in terms of angles of declination (D) and inclination (I).

reliability, whereas determinations of ancient field *directions* normally can be made with a much higher degree of precision and confidence. In consequence, this publication is concerned largely with the stratigraphical applications of *directional* changes of the field. However, certain other magnetic parameters which may be useful in lithostratigraphical correlations (e.g. intensity of remanent magnetism, and magnetic susceptibility) are also considered briefly.

The directional variations of the geomagnetic field with time are summarized in Fig. 3. The shortest-term fluctuations occur on time scales which range from seconds to a few years, and these are largely due to solar effects. They account for only a small proportion (typically *c.* 1%) of the total field, and have little, if any, geological significance.

Near-continuous records of the somewhat longer-term *secular variations* are available for the past few hundred years (at the most) from magnetic observatories (Fig. 4a) and for post-glacial times from studies of the remanent magnetism of lake sediments (Fig. 4b). Discontinuous records for the past few thousand years are available from palaeomagnetic studies of archaeological materials, and for earlier periods from the palaeomagnetism of older sedimentary sequences and volcanic lavas. These records indicate that both the declination and the inclination of the field undergo smooth quasi-cyclical fluctuations with 'periodicities' that range from 10^2 to 10^4 years. Under favourable circumstances such secular variations can provide a basis for subdividing geological sequences into stratigraphical units which can be used for correlation purposes, but apparently on a local or regional, rather than a worldwide scale.

Perhaps the most useful geomagnetic field changes from a stratigraphical point of view are polarity reversals. It is now well established that the geomagnetic field inverts its polarity (so that the magnetic north and south poles are interchanged) at irregular intervals. The present state, with the magnetic north pole lying close to the geographical north pole, is referred to as the *normal* polarity state. The opposite state, when the magnetic north pole lies close to the geographical south pole, is referred to as the *reverse* polarity state (Fig. 5). The time spent in one or the other state varies from about 10^4 to 10^7 years, and the transition from one to the other appears to take between 10^3 and 10^4 years (see, for example, Clement *et al.* 1982). In view of the relatively short duration of the transition period, geomagnetic polarity reversals can be regarded as effectively 'instantaneous' events on a geological time scale. Their globally synchronous nature makes them particularly suitable features for geological correlation purposes, especially if sets of normal and reversed polarity intervals with distinctive 'fingerprints' can be identified. Furthermore, if an unambiguous correlation of these with the appropriate part of a geochronometrically calibrated polarity reversal time scale can be achieved, then it is possible to *date* the sedimentary sequence.

Palaeomagnetic units

The magnetic units referred to in this publication are summarized in Fig. 6. SI units are used throughout the text, but the equivalent cgs units (together with conversion factors) are listed in Fig. 6 since these have been used widely in the palaeomagnetic literature in the past.

Principal carriers of magnetism in rocks

The remanent magnetism of rocks and unconsolidated sediments normally resides in iron oxides and occasionally in other iron compounds

TIME SCALE OF FLUCTUATIONS, OR RECURRENCE INTERVAL	NATURE OF VARIATION	SOURCE
< 5 years	Transient field fluctuations	External (solar activity)
$10^2 - 10^4$ years	Secular variations and geomagnetic 'excursions'	Internal (core processes)
$10^4 - 10^7$ years	Polarity reversals	
$10^7 - 10^8$ years	Polarity bias and reversal frequency.	

FIG. 3. Variations of the geomagnetic field with time.

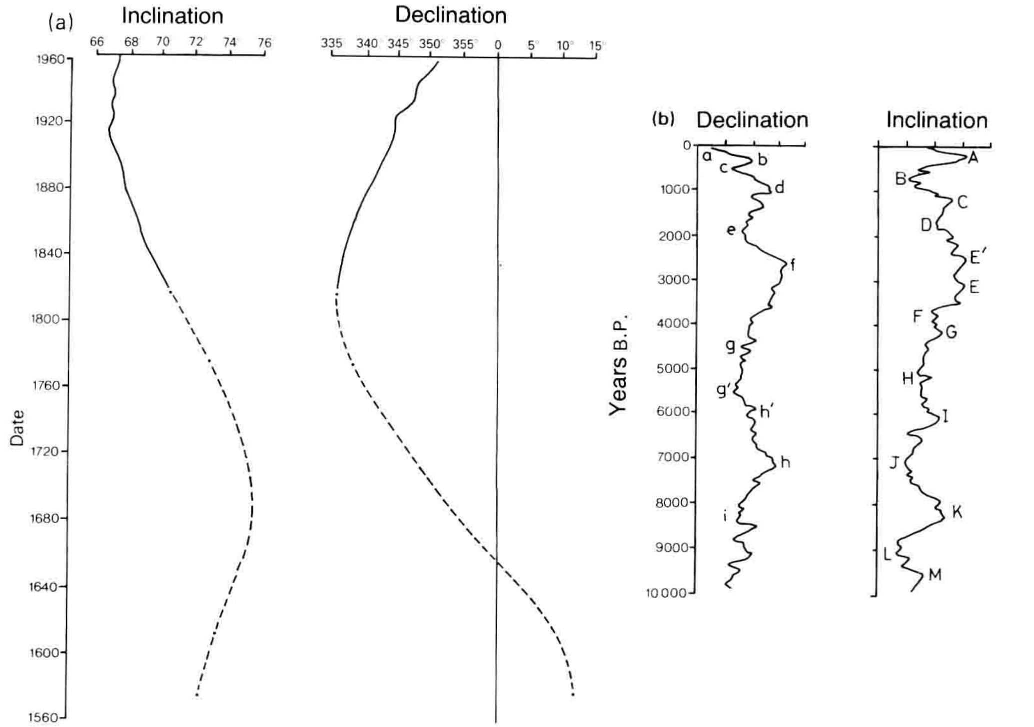

FIG. 4. (a) Variation of geomagnetic inclination and declination at the Greenwich magnetic observatory since 1580. (b) Variations of geomagnetic declination and inclination for the past 10 000 years recorded in the remanent magnetism of UK lake sediments (from Turner & Thompson 1981). Scale divisions are at 5° intervals on both plots. Features of these records that have proved useful for stratigraphical correlations between cores are lettered.

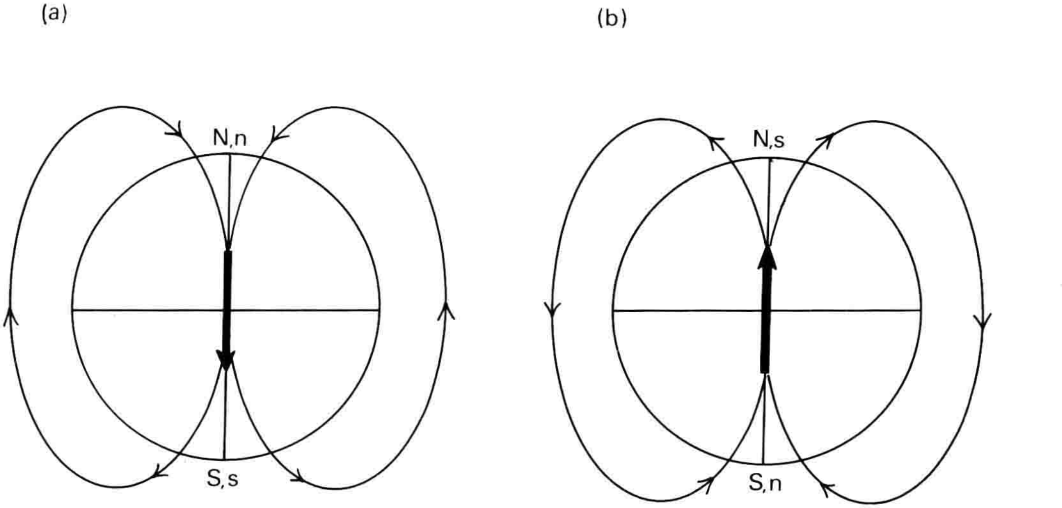

FIG. 5. Configuration of the axial dipole geomagnetic field during (a) a normal polarity interval: and (b) a reverse polarity interval. During a normal polarity interval the magnetic north pole (n) lies close to the geographical north pole (N). Consequently, during a normal polarity interval the lines of force are directed towards the geographical north pole. The magnetic declination is northerly, and the inclination is positive (downward-directed) in the northern hemisphere and negative (upward-directed) in the southern hemisphere. Conversely, during a reverse polarity interval the magnetic declination is southerly, and the inclination is negative in the northern and positive in the southern hemisphere.

PROPERTY	SI		COMMONLY USED cgs UNITS	CONVERSION FACTOR
	UNIT	COMMONLY USED SUB-UNIT		
Intensity of remanent magnetization of rock	Amperes per metre (Am^{-1})	Milliamperes per metre ($mA\ m^{-1}$)	Gauss (G)	$1\ mA\ m^{-1} = 10^{-6} G$
Magnetic moment of rock	Ampere metre^2 (Am^2)	Milliampere metre^2 ($mA\ m^2$)	Gauss cm^3 ($G\ cm^3$)	$1\ mA\ m^2 = 1 G\ cm^3$
Magnetic field	Tesla (T)	Millitesla (mT) Nanotesla (nT)	Oersted (Oe) Gamma (γ)	$1\ mT = 10\ Oe$ $1\ nT = 1\gamma = 10^{-5}\ Oe$
Magnetic susceptibility (per unit volume)	Dimensionless		Gauss/oersted ($G.Oe^{-1}$)	1 SI unit $= 4\pi$ $G.Oe^{-1}$

FIG. 6. Magnetic units.

such as iron hydroxides and sulphides. Other rock forming minerals such as phyllosillicates, amphiboles, and pyroxenes may contribute significantly to the *induced* magnetism of the rock in the present geomagnetic field (and therefore to the magnetic susceptibility of the rock), but these minerals do not have the capacity to carry a remanent magnetism. After the initial formation of the rock, subsequent weathering, diagenetic or metamorphic processes may produce iron oxides from these minerals, and such oxides may themselves carry a component of remanent magnetism which dates from the time of alteration. It is clear that a full understanding of the history of magnetization of a rock unit requires knowledge of the composition of the magnetic minerals present, as well as the processes of formation of these minerals. This information cannot always be readily obtained, but even a limited knowledge of the magnetic mineralogy can assist considerably in the interpretation of the magnetic properties observed in the rock.

The most important carriers of remanent magnetism in rocks are members of the magnetite-ulvospinel (titanomagnetite), and hematite-ilmenite (ilmenohematite) solid solution series. Deuteric exsolution in igneous rocks commonly leads to preferential formation of the end members of these series (for example, intergrowths of magnetite and ilmenite). Low temperature oxidation may result in the formation of intermediate oxidation products such as maghemite (γFe_2O_3) which carries a relatively strong remanent magnetism. More extensive alteration, particularly in sedimentary rocks, ultimately tends to lead to formation of hematite (αFe_2O_3), either by direct oxidation from magnetite, or through the formation of iron hydroxides such as goethite ($\alpha FeOOH$), which may later be reduced to hematite. The spontaneous magnetization of magnetite (92 $Am^2\ kg^{-1}$) is considerably greater than that of hematite (0.5 $Am^2\ kg^{-1}$), so that magnetite-bearing rocks (notably basic igneous rocks) normally have much stronger remanent magnetizations than rocks whose principal magnetic carriers are hematite (for example, many red bed formations).

Processes of magnetization in rocks

An appreciation of the reliability of magnetostratigraphic data requires an understanding of the processes by which a magnetic record is produced in rocks. Iron oxides commonly comprise only a few per cent by weight of a rock, and usually only a small proportion of these oxide grains possess a remanent magnetization. Thus the magnetism of the rock as a whole is normally carried by a fraction of 1% of the total minerals present. The nature of this magnetism will depend both on the magnetic properties of the individual magnetic mineral grains, and on the mechanisms by which the magnetic moments of these grains become aligned within the rock. A number of different types of naturally occurring remanent magnetization may be identified, according to the mode of alignment.

Thermoremanent magnetization (TRM)

Igneous rocks are formed by the cooling of magma from temperatures well in excess of

1000°C accompanied by the crystallization of various mineral phases; including iron oxides. As the iron oxides continue to cool after crystallization, an exchange interaction between atoms begins to dominate thermal disordering at a particular temperature called the *Curie point*. This results in a strong spontaneous magnetization setting in within these crystals, and this magnetism becomes closely aligned with the ambient geomagnetic field at that locality. The Curie points for the most important iron oxides, magnetite and hematite, are approximately 575°C and 675°C respectively.

Just below the Curie temperature the *relaxation time* (a measure of the response time of the magnetic moments of the grains to changes in the applied magnetic field) is very short, in the order of seconds. However, the relaxation time increases exponentially with increasing grain volume and with decreasing temperature. On cooling through a particular temperature called the *blocking temperature* the relaxation time rapidly exceeds 100s, and the magnetism of the grain then becomes effectively fixed in the direction of the ambient geomagnetic field. As the grain continues to cool the relaxation time may reach 10^9 years or more, so that a record of the geomagnetic field at the time of cooling is effectively 'frozen' into the rock for the rest of geological time. This type of magnetism is called a *thermoremanent magnetization* (TRM).

If the rock is reheated at a later stage in its history (for example, through deep burial, intrusion of a nearby igneous body, or a regional heating event associated with a subsequent orogenic episode) then a new component of TRM will be acquired. The direction of this new component will be parallel with the geomagnetic field at the time of reheating. This component will be carried by those grains that have blocking temperatures up to the maximum reheating temperature. However, provided that the latter is less than the Curie temperature of the magnetic constituents, a component of the original magnetization, dating from the initial cooling, will remain in those grains whose blocking temperatures lie between the reheating temperature and the Curie temperature. It is clear, therefore, that a given rock unit may carry several components of TRM, each dating from a different heating episode in the rock's history, and each carried by populations of magnetic mineral grains with different blocking temperature. This *principle of superposition of magnetization* components applies equally to other types of magnetism.

Chemical remanent magnetization (CRM)

Various physicochemical changes may take place in rocks subsequent to their initial formation, and commonly these involve crystallization of new magnetic mineral phases. Following nucleation, when the grains are extremely small, their relaxation times will be very short. As they grow through a particular volume, the *blocking volume* (analogous to the blocking temperature), so their magnetism becomes effectively fixed in direction, parallel with the ambient geomagnetic field, and the rock acquires a component of magnetism dating from this time. This is called a *chemical remanent magnetization* (CRM). CRM may be particularly important in sedimentary rocks, where it may arise from various processes occurring during lithification and diagenesis, and also during chemical weathering.

Depositional and post-depositional remanent magnetization (DRM and PDRM)

This type of magnetism is acquired by clastic sediments as a result of the physical alignment of magnetic mineral grains during the depositional process, prior to lithification. The individual magnetic mineral particles normally will have been eroded from pre-existing rock formations, and the majority of these grains will have been derived at some stage from igneous or metamorphic bodies. These grains will carry a magnetization that dates from their time of cooling. In consequence they will behave as small magnets, and, when free to rotate during suspension in the water column, they will take up an orientation which, in the absence of other significant aligning forces, will be parallel with the ambient geomagnetic field. Under certain circumstances this orientation may be preserved during the depositional process, so that the resultant sediment acquires a remanent magnetism which is parallel with the field at the time of deposition. This is called a *depositional remanent magnetization* (DRM). However, laboratory redeposition experiments indicate that various systematic processes may cause departures of the direction of the remanent magnetism (particularly the inclination) from that of the ambient field. These include the effects of the gravitational couple and hydrodynamic shear due to bottom currents on the orientation of the grains. In laboratory experiments, inclination 'errors' have been observed to be as great as 25° (see, for example, King 1955; Rees 1961) but in most natural sediments they appear to be much smaller, and more often are completely absent (see, for example, Thompson & Turner 1979). This is believed to be due to the ability of the magnetic mineral grains to rotate within the water-filled interstitial cavities between the (generally larger) non-magnetic grains, subsequent to deposition (see, for example, Irving & Major 1964; Tucker 1980). This process allows the orientation of the magnetic grains to be realigned parallel with the geomagnetic field, so that any departures of this alignment arising during deposition are corrected. This process of reorientation may occur for a relatively limited time after deposition until compaction or cementation finally locks the grain in place. The resultant magnetism is called a *post-depositional remanent magnetization* (PDRM).

Viscous remanent magnetization (VRM)

This is a time-dependent magnetization acquired by rocks as a result of their being subjected to a weak magnetic field for long periods. Thus most rocks have been situated in the earth's magnetic field for long periods after their initial formation, and during this time the magnetic moments of grains with short relaxation times will gradually become aligned parallel with the ambient field. This results in the development of a component of *viscous magnetism* within the rock, which is usually parallel with the recent field direction. Fortunately this type of magnetism can normally be erased by the standard laboratory procedures discussed below.

Anhysteretic remanent magnetization (ARM)

This is a relatively rare type of magnetism in natural rocks, which may occur, for example, as a result of lightning strikes. It is due to the simultaneous application of an alternating and a direct magnetic field.

In addition to the five types of natural magnetization outlined above, other types of magnetism may be induced during field and laboratory treatment of palaeomagnetic samples. These include *isothermal remanent magnetization* (IRM), produced by the application of a strong magnetic field for a short interval of time, and rotational remanent magnetization (RRM) which is produced under some circumstances when rock samples are rotated in the presence of an alternating magnetic field (see, for example, Wilson & Lomax 1972; Stephenson 1981).

The general term *natural remanent magnetization* (NRM) is used to describe the magnetization of rocks in their natural state, before they have been subjected to any laboratory demagnetization procedures. It is clear from the above discussion that the NRM of a rock may include a number of different components of magnetization, perhaps of different type (for example, a VRM and a TRM superimposed on an original PDRM), each acquired at different stages in the rock's history, and each residing in different populations of magnetic mineral grains within the rock. It is customary to use the term *primary magnetization* to describe the component that is acquired at, or very close to, the original time of formation of the rock, and *secondary magnetization* to describe components acquired at later stages in its history. In magnetostratigraphic studies we are normally concerned with the primary component, but under some circumstances secondary components can also provide useful information, provided that their time of acquisition by the rock can be established reliably. Very often a particular component of stable magnetism dominates a geological unit, and yet there is no clear evidence for or against this component representing the primary magnetization. Often the assumption is made that this component is primary, but without proof usage of the term 'primary magnetization' is inappropriate. A better term is *characteristic* magnetization, and this term is used throughout this publication, except in cases where the primary nature of the magnetism is established unequivocably.

Field and laboratory procedures

The principal aims of palaeomagnetic laboratory procedures are to examine the stability of the magnetism of the rock, and to attempt to isolate the different components of magnetization present. A full interpretation of these data then involves establishing the ages of the different components, and in particular attempting to identify the primary magnetization. Quite often palaeomagnetic studies fall short of these targets; for example, because the primary component may have been overprinted completely by secondary components, or because the ages of one or more components cannot be established reliably, or simply because the magnetism of the rock as a whole is too weak or too unstable for any reliable palaeomagnetic determinations to be made.

Sampling

The first stage in any magnetostratigraphic study is the collection of orientated samples. In studies of natural exposures it is commonplace to use a portable coring drill to take 25 mm diameter cores, up to 100 mm long, from which 25 mm right-cylindrical specimens can be cut subsequently. These cores are orientated with respect to north and the vertical before separation from the outcrop. For conventional palaeomagnetic studies a minimum of six separately orientated cores are normally taken at each stratigraphical level (or site), in order to average out small orientation errors and other sources of local variability, and to improve the precision of the mean site remanence directions. A precision of $1-2°$ is desirable in the final palaeomagnetic determination if the results are to be used for plate motion studies. However, in studies of polarity reversal stratigraphy a somewhat lower degree of precision (*c*.5°) is usually acceptable, but sampling at close stratigraphical intervals is essential in order to maximize the stratigraphic resolution. As a compromise, the number of independently orientated samples per stratigraphic level can be reduced to two or three, the principal purpose of replicate sampling in this case being to check the internal consistency of the palaeomagnetic data.

As an alternative to field drilling, orientated blocks (hand samples) may be taken. The numbers of samples involved in a detailed magnetostratigraphic study may be very large, and handling and transport of these in the field, and the subsequent preparation of specimens from them in the laboratory, can pose particular problems. Furthermore, the orientation is usually less precise than for cores drilled in the field. Specialized techniques have been developed for sampling poorly consolidated sediments such as clays and loose sands and silts in the field (see, for example, Townsend & Hailwood 1985).

Magnetostratigraphic investigations have been carried out widely on unconsolidated sediments taken from freshwater lakes, continental shelves and deep ocean basins by hydraulic, gravity, piston, and rotary corers. These sediment cores are usually sampled by inserting small, non-magnetic sample boxes (commonly 8 ml cubes) at regular intervals into a longitudinal section of the core. Although the direction of the vertical in these cores is known, it is seldom possible to orientate them with respect to geographical north. Thus, although the inclination of remanent magnetism can be specified reliably it is often only possible to specify relative declination changes with depth, rather than absolute declination values. Since the inclination changes sign, and the declination changes by approximately 180° when the geomagnetic field reverses, this information is normally sufficient to define a reliable magnetic polarity stratigraphy. However, in the case of cores taken by standard rotary drilling techniques, such as those used by the Deep Sea Drilling Project and in drilling for hydrocarbons, relative rotation occurs between successive fragments of core. Relative declination values for successive core pieces cannot then be defined for the remanent magnetism. In such cases the magnetic polarity sequence must be deduced from the remanent inclination values alone. Thus in the northern hemisphere positive (downward-directed) inclinations represent a normal polarity, and negative (upward-directed) inclinations a reverse polarity. The opposite applies for sites at which the magnetism was acquired in the southern hemisphere. Such determinations are normally quite reliable for sites at intermediate and high latitudes, but for those within a few degrees of the equator the inclination values are so small that it is often difficult or impossible to discriminate between normal and reverse polarity values within the experimental uncertainties of the palaeomagnetic determinations. An important advance in magnetostratigraphic studies of cores obtained by the Deep Sea Drilling Project has been the introduction of the hydraulic piston corer, which permits recovery of unconsolidated sediments to a depth of about 250 m below the sea floor without rotating the drill pipe. This has enabled relative declination values to be defined, and reliable polarity reversal sequences to be established in cores from low latitude sites.

Palaeomagnetic measurements

The remanent magnetism of rock specimens is measured by means of laboratory magnetometers. Two types of instrument that are widely used at present are *balanced fluxgate* and *cryogenic* magnetometers. In the former, the sensor operates at room temperature and is driven by a high frequency alternating magnetic field, which is modulated by the weak direct field due to the rock specimen. The output is a very small voltage which is proportional to the component of magnetism being measured, and which can be processed readily by computer. The specimen is normally rotated at a low frequency (about 10 Hz) during the measurement, to allow filtering of external noise and to permit specific components of magnetization to be detected. Normally three orthogonal components are measured in sequence, and these are combined by the computer to give the direction and intensity of remanent magnetism of the specimen. The weakest magnetic intensities that can be measured reliably with this instrument are normally about 0.05 to 0.1 mA m^{-1}. In the cryogenic magnetometer the sensing element operates at liquid helium temperature (−183°C), and the presence of a weakly magnetic rock specimen (in a chamber at room temperature) within this sensor causes a persistent current to flow, whose value is proportional to the component of magnetism being measured. In principle the cryogenic magnetometer is perhaps two orders of magnitude more sensitive than the best balanced fluxgate instruments, but in practice instrumental noise often limits the effective sensitivity to about 0.01−0.05 mA m^{-1}. However, a major advantage of the cryogenic magnetometer is its more-or-less '*instantaneous*' response (*c*.1 s), compared with the significant times required to integrate the signal on a balanced fluxgate instrument to the required level for reliable measurement (between 2 and 20 min., depending on the intensity). Thus, in magnetostratigraphic investigations of weakly magnetic sediments such as carbonate sequences, the cryogenic magnetometer has clear advantages over other instruments.

Other types of instrument that have been used widely in the past, and which are still in use in a number of laboratories, are the *astatic* and the *spinner* magnetometer. In the former, the rock specimen is brought close to a sensitively balanced suspended magnet system and the deflection of this system is monitored as the specimen is moved into a set of different orientations. These instruments are very sensitive to magnetic and mechanical noise, but when operated in a low noise environment, intensities of magnetization as low as 0.1 mA m^{-1} can be measured reliably. In the spinnner magnetometer the specimen is rotated at a high frequency within a pick-up coil. The magnetic moment of the specimen induces an alternating voltage in this coil, and the direction and intensity of the component of magnetization within the plane of measurement can be determined from the phase and amplitude of this signal. As with most other laboratory magnetometers, three orthogonal components are measured and then combined to give the resultant magnetization.

Progressive demagnetization

As noted above (p. 7), most rocks carry a number of different components of magnetization, which have been acquired at different times in the rock's history. These components will reside in different populations of magnetic mineral grain within the rock, which commonly will have different magnetic *stabilities*. The stability of the magnetism in a grain may be specified in terms of its blocking temperature or its *coercivity* (the remanent magnetism of the grain may be destroyed by the application of a particular magnetic field in the opposite direction to the magnetism, and the coercivity is a measure of this field).

Separation of the different components of magnetism in the rock is normally achieved by progressive demagnetization procedures, the lower stability components being removed preferentially during the early stages of this treatment, and the higher stability components during the later stages. Three types of progressive demagnetization methods are in use, namely alternating field, thermal and chemical demagnetization.

(i) *Alternating field (af) demagnetization.* In this technique the specimen is placed within a demagnetizing coil, which generates an alternating field with a very pure sinusoidal wave form. The coil is situated inside a magnetic shield or Helmholtz cage, to remove the effect of any other magnetic fields present in the laboratory. The applied alternating field is ramped up to some predetermined maximum value, H_1, and then smoothly reduced to zero. During each half-cycle of the applied field the magnetism of all grains in the rock with coercivities up to H_1 will follow the applied field, and alternate in direction along the demagnetizing coil axis. As a result of the progressive reduction of the alternating field to zero, equal numbers of grains will be left with their magnetic moments pointing in opposite directions along this axis, so that no net magnetism results from these grains. Thus the magnetism of all grains with coercivities up to H_1 is effectively randomized. This process must be carried out sequentially along three orthogonal axes of the specimen, or alternatively the specimen must be 'tumbled' during application of the alternating field, in order for it to be demagnetized uniformly throughout. This treatment is then repeated at progressively higher applied fields, H_2, H_3, etc., the magnetization of the specimen being measured after each step. From the changes in the resultant magnetization vector during this treatment it is possible to identify the different components of magnetization present in the rock. In routine palaeomagnetic investigations it is common to increment the applied alternating field at 2.5 or 5 mT intervals up to a maximum of about 50 mT, but occasionally fields in excess of 100 mT may be necessary to successfully decompose the remanent magnetism. This technique is used widely for rocks in which the remanence is carried by titanomagnetites, but because the coercivities of hematite grains are often greater than 300 mT, and therefore beyond the maximum limit of most af demagnetizers, it is usually necessary to use thermal (or chemical) demagnetization for hematite-bearing rocks.

(ii) *Thermal demagnetization*. In this technique the specimen is heated in a non-magnetic oven to a precisely determined temperature and then cooled back to room temperature in a magnetic field-free space. The thermal fluctuations due to heating will randomize the magnetization of all grains with blocking temperatures up to this applied temperature. The magnetization of the specimen is then measured at room temperature and the procedure is repeated at a slightly higher temperature. This process is continued until the magnetism of the specimen has been destroyed completely. It is common to use 50°C or 100°C temperature increments up to about 500°C, but after this, as the Curie temperatures of the magnetic constituents are approached, the increments may be reduced to 5°C or 10°C, in order to reliably identify the high blocking temperature components. Although the thermal demagnetization technique often allows the isolation of components of magnetization carried by very high coercivity grains, which are not amenable to standard af demagnetization procedures, it suffers from the disadvantage that heating of the rock may lead to undesirable chemical or mineralogical changes, in particular the generation of new magnetic phases.

(iii) *Chemical demagnetization*. In some cases the magnetic stabilities of different components of magnetism within the rock may be closely similar, so that two (or more) components are removed simultaneously during af or thermal demagnetization, and these components cannot then be identified separately. In such cases, if the different components are carried by grains with different mineralogical compositions (for example, different phases of hematite), then it may be possible to remove one or more of these phases preferentially by selective dissolution of the appropriate grains in a suitable acid (see, for example, Collinson 1967; Henshaw & Merrill 1980). Because of the difficulties of identifying appropriate leaching agents, ensuring complete penetration of these through the specimens, and also because of the long duration of the experiments (typically >800 h) this technique has not been used very widely, but it appears to have particular value in attempting to understand the complicated magnetization of certain red bed formations that have been the subject of magnetostratigraphic investigation in recent years.

Presentation and analysis of palaeomagnetic data

Demagnetization data

The results of progressive demagnetization investigations can be plotted in either of two ways: (a) as stereographic projections of the direction of remanent magnetism, with accompanying graphs showing the changes in intensity of magnetization (Fig. 7a); or (b) as 'vector end-point' (VEP) diagrams (sometimes called As-Zijderveld diagrams) (Fig. 7b). In the latter case the end point of the magnetization vector after each demagnetizing step is projected on to the horizontal plane and also on to a vertical plane which is normally aligned either north–south or east–west. VEP diagrams have the advantage that they combine both the directional and the intensity changes on to a single plot, rather than treating these parameters separately. If, during a particular part of the demagnetization procedure a single component of magnetism is being removed, then the corresponding points will lie on straight line segments in both orthogonal projections of the vector diagram, and the precise direction of the component can be determined from the gradients of these lines. Thus in the example shown in Fig. 7b, the specimen possesses two distinct components of magnetization, with the lower stability component (A) being isolated during af treatment up to 25 mT, and the higher stability (approximately antiparallel) component (B) by treatment at higher fields. When only a single component of magnetism remains in the specimen (in this case component B) the straight line segments for both the horizontal and vertical projections are directed through the origin. In this example component B can be identified from the stereographic projection (Fig. 7a) as the point at which no further directional changes take place (this is often referred to as the demagnetization '*stable end point*'). The direction of this magnetization

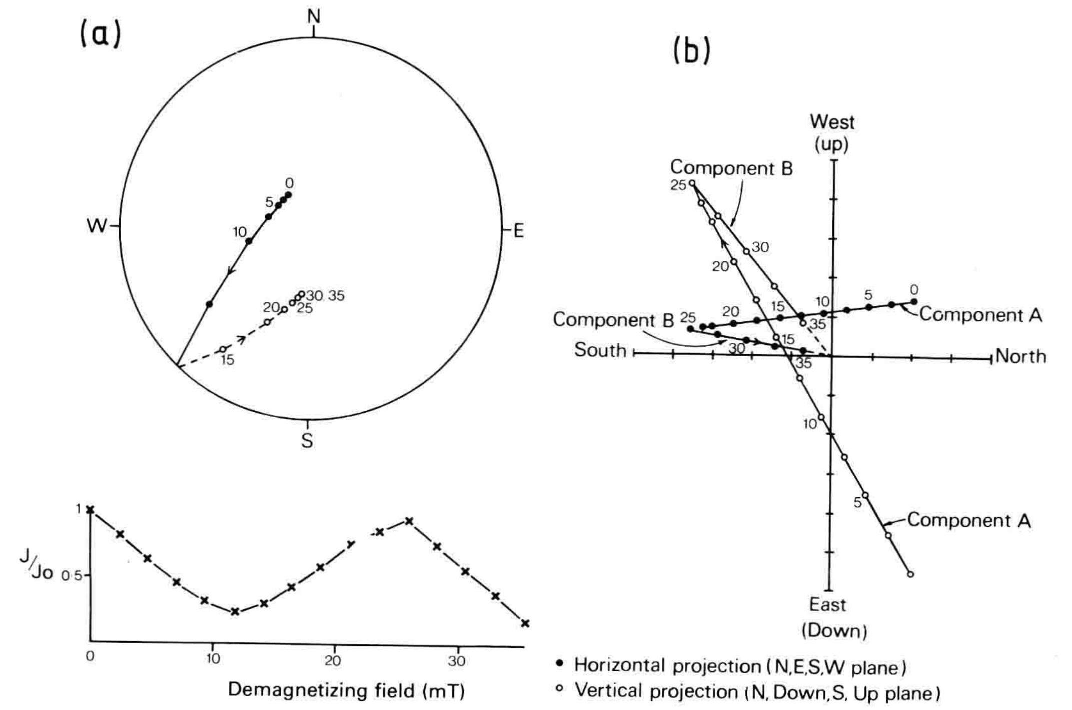

FIG. 7. Alternative methods used for presenting demagnetization data. (a) The successive directions of magnetization may be plotted on a stereographic projection (solid symbols lower hemisphere, open symbols upper hemisphere), with an accompanying graph showing the variation of intensity of magnetization (J) (normalized against initial value, Jo) with demagnetizing field. (b) The end point of the magnetization vector after each demagnetization step may be plotted on to the horizontal plane (solid symbols) and on to a vertical plane, which in this example has a north–south orientation (open symbols). These two plots are usually combined into a single vector end-point diagram, as shown, with the azimuth of the vertical plane (in this case north–south) forming the common axis. For details of interpretation see text.

component can be read directly from the projection, but that of component A cannot be deduced simply from this diagram alone. This illustrates the need to present the demagnetization data in both ways. In some cases, due to overlapping stabilities of two or more components, a stable end point may not be reached before the resultant magnetism becomes too weak for further reliable measurement (in this case the direction of resultant magnetization of the specimen continues to change systematically up to the highest demagnetization step). This will be reflected on the VEP diagram by the appropriate points lying on a curved, rather than a linear, segment. In this case the direction of the high stability component cannot be derived from the VEP diagram. However, on the stereographic projection these points will lie on a great circle segment which will pass through the required stable end point. Provided that the lower stability components have slightly different orientations in different specimens, then by plotting the great circle trends for a number of different specimens the true stable end point direction can be found from the point of intersection of these great circles (see, for example, Halls 1978; Townsend & Hailwood 1985).

Magnetostratigraphic data

Basic magnetostratigraphic data are normally presented in graphical form as stratigraphical plots of the declination and inclination of the stable characteristic magnetization. These are usually translated into diagrammatic form by denoting intervals of normal polarity in black, and reverse polarity in white. Occasionally basic data are plotted as 'virtual geomagnetic pole' (VGP) latitudes. The VGP represents the position of the effective north geomagnetic pole calculated from the magnetization vector on the assumption of an axial geocentric dipole field model. These plots have some advantage in that both the declination