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Palaeomagnetism

Roy Thompson

In this review the broad outline of the study of palaeomagnetism, recent advances in the subject, and its interrelationship with other branches of science are described in simple terms. The palaeomagnetic methods and techniques which are now widely used are presented in the early sections of the review. In the latter sections palaeomagnetic results, now in rapidly increasing quantities and of greater quality, and their implications for several aspects of the earth sciences, are discussed.

Preamble

The study of palaeomagnetism has (by analogy with a chess game) progressed through the opening stages, where the main methods and techniques were developed, into the transition to the middle game. This is an interesting and absorbing stage of advancement where the fruits of the opening labours are seen more clearly. During the opening moves new ground was traversed which paved the way for the establishment of such concepts as sea floor spreading and plate tectonics. The subject is now advancing strategically on a broad front and expanding onto new flanks, with individual research workers providing the tactical details!

Introduction

Palaeomagnetism is the study of the geomagnetic field throughout geological time as recorded in the permanent magnetization of rocks. The history of palaeomagnetism is the oldest of any branch of geophysics, as the magnetic properties of lodestone (a magnetite-rich rock) were known to the Chinese several centuries B.C. Most probably by the second century B.C. the Chinese were using lodestone as a magnetic compass. Further developments came between A.D. 1100 and 1600 when the properties of the geomagnetic field were studied and observed in Europe, particularly by Petrus Peregrinus and William Gilbert. The next large step was when systematic studies of the magnetization of recent lava flows were made in the nineteenth and early twentieth centuries. However, it was not until the development in the early 1950s of highly sensitive instruments for measuring the remanent magnetizations of rocks that palaeo-

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magnetic studies began in earnest. Since then palaeomagnetic work has continuously increased both in the number of investigations being conducted, and the standard of results attainable. Palaeomagnetic studies now have importance in several branches of the earth sciences including stratigraphy of marine and continental sequences, evolution rates, palaeoclimates, continental drift and sea floor spreading, constitution of the earth's core and long-term variations of the geomagnetic field. The various applications of palaeomagnetism to these subjects are described in the latter half of this article after the methods and techniques commonly used by palaeomagnetists have been discussed.

Origin of the remanent magnetization of rocks

Iron oxide minerals hold the permanent magnetization, which may have been formed by a variety of processes. In igneous rocks the magnetization is of a thermoremanent (TRM) origin. It grows when the rocks cool down, from their molten state, through the Curie point of each iron oxide mineral they contain, to ambient temperatures. Below the Curie points, the thermal vibrations in the iron oxide lattices are too weak to randomize the magnetic exchange interactions of the unpaired electrons of the iron atoms, and a permanent magnetization is able to form. The most common carriers of remanence in igneous rocks are titanomagnetites which are ferrimagnetic and have Curie points of between 400 and 600°C, well below the melting point of rocks. During cooling the titanomagnetites often oxidize and exsolve to some extent, but are still capable of carrying a stable thermoremanence.

Another type of magnetic remanence, chemical remanent magnetization (CRM), results from the formation at low temperatures of magnetic minerals, and is often associated with diagenesis or metamorphism. For example, in red sediments the haematite pigment often carries a stable remanence which has grown diagenetically. Once the haematite grains have grown through a critical size (diameter 0·1 μm) the magnetization becomes 'locked in' and is then stable over long intervals of geological time. DRM, detrital remanent magnetization, is a third possible type of remanence. It results simply from the alignment of sediment grains, which already hold a remanence, in the ambient magnetic field direction as they settle down through water or in the water filled interstices of a consolidating sediment.

A. Methods and techniques

Collection of samples

The establishment of a continuous geomagnetic record is hampered by the inherently discontinuous nature of the geological record and the variable accessibility of suitable sampling localities. Thus the palaeomagnetic sampling method which has evolved is one of collecting samples from a number of separate exposures (sites) which are part of a geological unit. Palaeomagnetic results from a site (e.g. a lava flow) relate to a single moment in time. Samples are commonly cut into a number of specimens which are measured and investigated individually. Results from the specimens are combined to give sample mean directions. Results from several sites (obtained by combining sample means) are averaged to yield a formation or rock unit mean which describes the average direction of the geomagnetic field over the extended period of time represented by that rock unit. If results of formation means of the same age (from one continent) can be shown to be consistent, they can be combined to define finally a palaeomagnetic pole position for the continent and for the time interval involved.

Two methods are widely used to collect samples. One method is simply to collect orientated blocks of rock which are then drilled and sliced into convenient sized specimens (e.g. 2·5 cm high, 2·5 cm diameter cylinders) in the laboratory. The other method is to drill cores from an exposure using light weight, field, drilling equipment. In both methods orientation of the sample is usually made with a sun compass (a portable sun dial) and a clinometer, and an accuracy of a

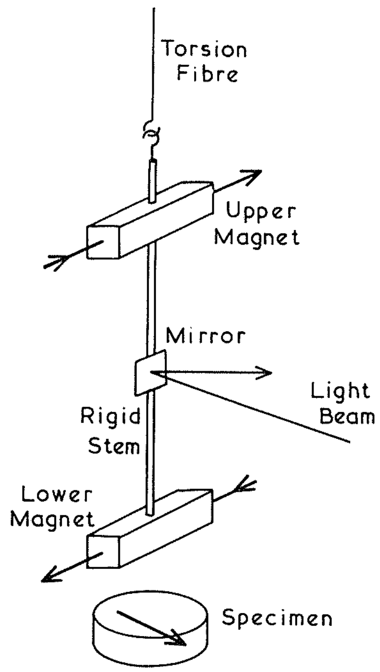

Fig. 1. Simplified diagram of the magnet system of an astatic magnetometer.

few degrees is normally aimed for. However, for certain more detailed studies, e.g. palaeosecular variation studies, an accuracy of one degree or less is desirable.

Measurements of specimens

Magnetic remanence measurements are mainly made with either astatic or spinner magnetometers.[1] The principle of the astatic magnetometer is based on the behaviour of a suspended magnet system which is insensitive to uniform magnetic fields but which responds to the field gradient produced by the remanence of a specimen positioned below the system. High sensitivity is achieved by aligning the magnets antiparallel (Fig. 1) and by making the controlling torque on the system very weak. When a specimen is placed below the magnetometer the astatic pair rotates slightly. This rotation is measured by reflecting a light beam from an attached mirror (Fig. 1). Deflections of the light beam are measured for a number of different orientations of the specimen (commonly sixteen) and combined to calculate the direction and intensity of magnetization of the specimen.

Spinner magnetometers operate on the principle that the magnetic remanence of a specimen when spun inside a coil of wire, about an axis in the plane of the coil, produces an alternating e.m.f. The amplitude and phase of the e.m.f. are measured about three orthogonal axes of the specimen, relative to a fiducial mark on the rotating head. From these measurements the intensity and direction of magnetization can be calculated directly. Recently fluxgate spinner magneto-

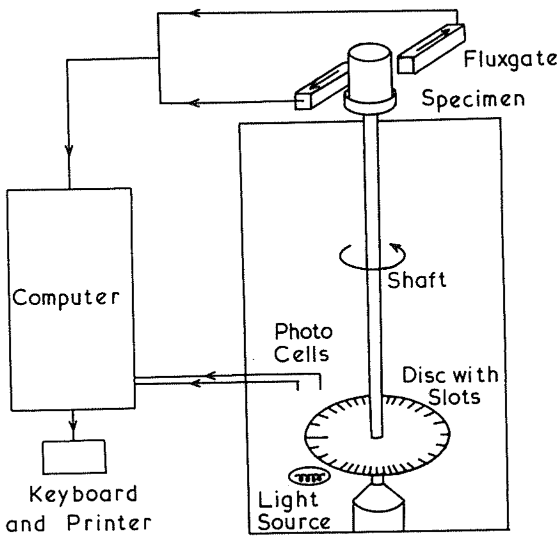

Fig. 2. Block diagram of a computerized, fluxgate slow speed spinner magnetometer

meters have been developed which have several advantages over earlier methods.[2] In the fluxgate system two flux-sensitive probes are positioned in opposition near the rotating specimen (Fig. 2). As the sample rotates the fluxgate probes are subjected to a fluctuating magnetic field. A varying output results which is analysed by either a lock-in amplifier or by a digital computer with a reference signal generated by the rotating shaft. In the computer system, measurements are taken 128 times each cycle, timed by the illumination of photocells through the radial slots of a rotating disc, and are integrated over several cycles (Fig. 2). The measurements are then subjected to Fourier analysis to determine the intensity and direction of magnetization of the specimen. The advantage of the computerized fluxgate system is that a wide range of intensities of remanence can be measured rapidly in a magnetically and vibrationally noisy environment.

Ancient field intensities

If the remanence measured is stable and of a thermoremanent origin (e.g. in lavas or pottery) the ancient magnetic field intensity may, in certain cases, be deduced by reheating the specimen and cooling it in a known field (H_a). The ancient field intensity (H_o) is given by the relationship

$$H_o = H_a \frac{J_o}{J_a}$$

where J_o is the natural remanence and J_a is the remanence acquired in the known field H_a.

Statistical analyses

Palaeomagnetic measurements require statistical analysis because the palaeomagnetic directions they give are scattered due to secular variation and random errors introduced during sampling, measurement, etc. Sir Ronald Fisher[3] invented a statistical method appropriate to the palaeomagnetic problem in which the density of points on a sphere at an angle θ with the mean direction is taken to be proportional to exp (K cos θ) where K is the precision. This frequency distribution in the spherical problem corresponds to the Gaussian error function of the one dimensional theory. K, the precision, corresponds to the reciprocal of the variance in the Gaussian distribution. If $K = 0$ the points are randomly distributed: if K is large the points are closely grouped about the mean direction. A further estimate of precision which is commonly used in palaeomagnetic studies is α_{95}. This is the semi angle of the circular cone about the calculated mean direction within which the true mean lies at the 95% probability level. For small α the relationship

$$\alpha_{95} = \frac{140}{\sqrt{KN}}$$

holds approximately, where N is the number of measurements. Using Fisher's distribution statistical tests may be rigorously constructed to compare various mean directions and test if they are significantly distinct, or to test if mean directions are equivalent to a known direction, e.g. the earth's magnetic field direction.

Stability tests

A variety of stability tests can be used to help in confirming that the magnetic remanence was acquired close to the time of formation of the rock, rather than

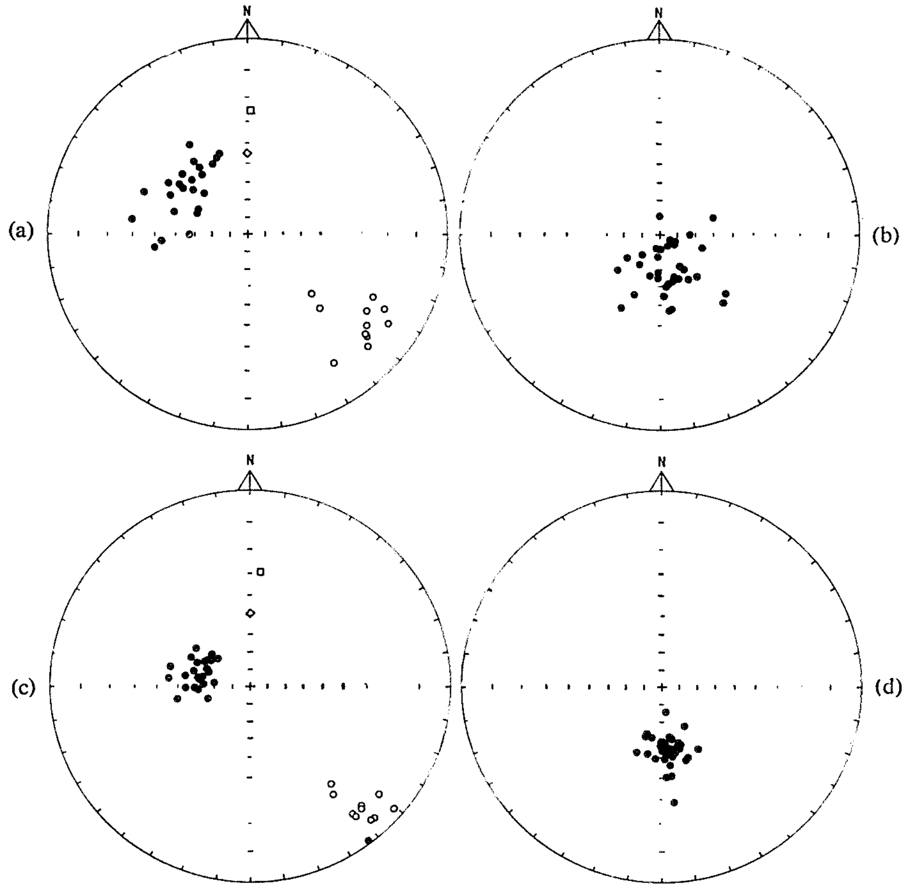

Fig. 3. Example of stability tests. Data plotted on stereographic projections where each circle represents the mean direction of the remanence of three specimens from a single hand sample. Closed circles, directions down; open circles, directions up; open diamond, present axial field direction; open square, present magnetic field direction. (a) and (b) NRM; (c) and (d) thermally cleaned; (a) and (c) referred to the present horizontal; (b) and (d) referred to the palaeo-horizontal.

at some later time.[1] Stability tests are thus extremely useful and are used extensively in palaeomagnetic studies. The tests can be divided into two main groups: field and laboratory. An example of the field test is the 'fold test'. It can be used when the rocks under investigation have been tilted or folded. The dispersion (e.g. α_{95}) of the remanence vectors is calculated for the rock samples in their present orientation (i.e. the position in which they were collected) and also their palaeo orientation (i.e. the position in which they were originally formed). If the grouping of the magnetic vectors is tighter when referred to the palaeohorizontal, then the remanence must have been acquired before the folding or tilting of the rocks. A good example of the fold test is shown by a collection made by the author[4] of South American Permian sandstones from two limbs of an anticline (Figs. 3a and b). In Fig. 3(a) the directions of magnetic remanence are plotted on a stereographic (Wulff) projection and are referred to the present horizontal. The two separate groups of directions (one pointing to the north-west and inclined downwards at about 50°, the other pointing south-east and upwards at about 20°) correspond to samples collected from the two limbs of the anticline. When the directions are referred to the palaeohorizontal by rotating the bedding planes (and the magnetic directions with them) back to the ancient horizontal, the results plot in a single group with a steep inclination (Fig. 3b) clearly illustrating the stable nature of the remanence.

Partial thermal demagnetization or thermal cleaning is an example of a laboratory test. The specimens are heated in stages to successively higher temperatures, being cooled in field free space and remeasured between each heating stage. Less stable components, that is those most likely to be of secondary origin, are removed first by this heat treatment. The most stable, and most probably primary components thus remain at the higher temperatures. The removal of the weaker components (which are often in the direction of the earth's present magnetic field) can be traced and the stable high temperature component used alone to calculate the past direction of the ancient field. This type of test is also illustrated in Fig. 3. The two lower diagrams show the high temperature (primary) components and the upper diagrams the natural remanence (which is composed of the primary plus a secondary remanence). The secondary remanence is of a variable intensity, in the direction of the earth's present magnetic field, and causes a scattering and streaking of the results towards the present magnetic field position (Fig. 3a). The considerably reduced scatter and lack of streaking in Figs. 3(c) and 3(d) indicate that thermal demagnetization has successfully isolated a stable primary remanence, which, when referred to the palaeohorizontal, can be used to calculate a Permian palaeomagnetic pole for South America.

All palaeomagnetic samples are nowadays subjected to a variety of statistical, field, and laboratory tests which they must satisfy before they are used for drawing inferences about the past behaviour of the earth's magnetic field.

B. Geomagnetic and tectonic inferences

Secular variation

Secular changes of the geomagnetic field span periods from tens of years to hundreds of thousands of years. Magnetic observatories all over the world record changes in direction and intensity of the field. For some observatories, e.g. London and Paris, records extend back over about 400 years (Fig. 4). Secular changes can be very rapid: for example, at London magnetic declination swung through 35° in 240 years and at Cape Town the horizontal intensity decreased by 30% in 100 years. One branch of palaeomagnetism, named

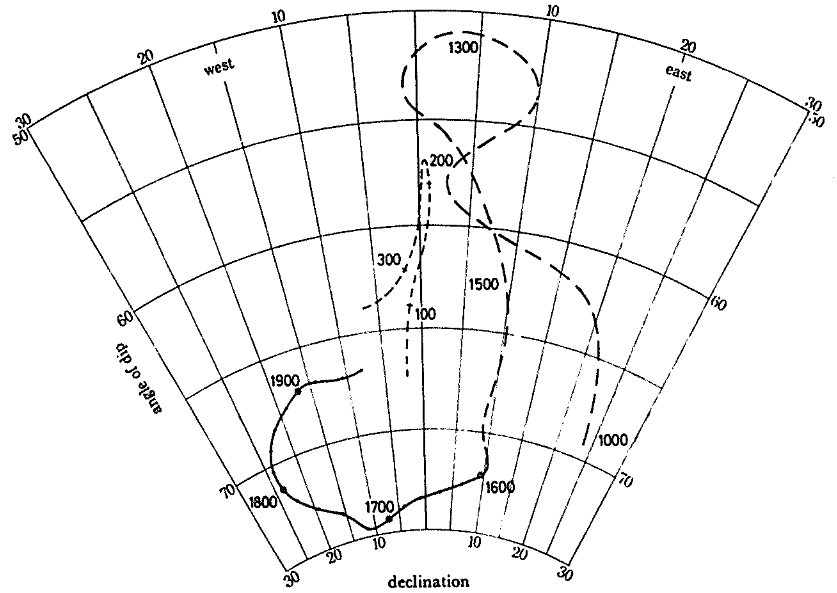

Fig. 4. Secular variation of declination and inclination at London from observatory and archaeomagnetic data from Aitken (1970).[5]

archaeomagnetism, extends the secular variation record through historical times. Orientated samples of baked brick walls and floors of pottery kilns hold a thermoremanent record of the direction, and sometimes intensity, of the geomagnetic field at the time of their last firing. By dating the hearths (and hence the magnetizations) archaeologically, or by radiocarbon methods on associated material, palaeomagnetists have built up a pattern of variations for Roman times and from A.D. 1000 to the present day for Britain[5] (Fig. 4). It has been suggested, on the basis of the observatory records alone, that secular variation is a cyclic phenomenon, but the archaeomagnetic results show that secular variation is much more complicated. Furthermore the clockwise motion of the magnetic field vector in the observatory records (Fig. 4), linked with the present westward drift of the geomagnetic field, is not seen in the archaeomagnetic data where an anticlockwise motion predominates.

The pattern of secular changes can be extended even further back in time by the results from a recent development in palaeomagnetism, the investigation of lake sediments.[6, 7] Deposition rates of a few millimetres per year

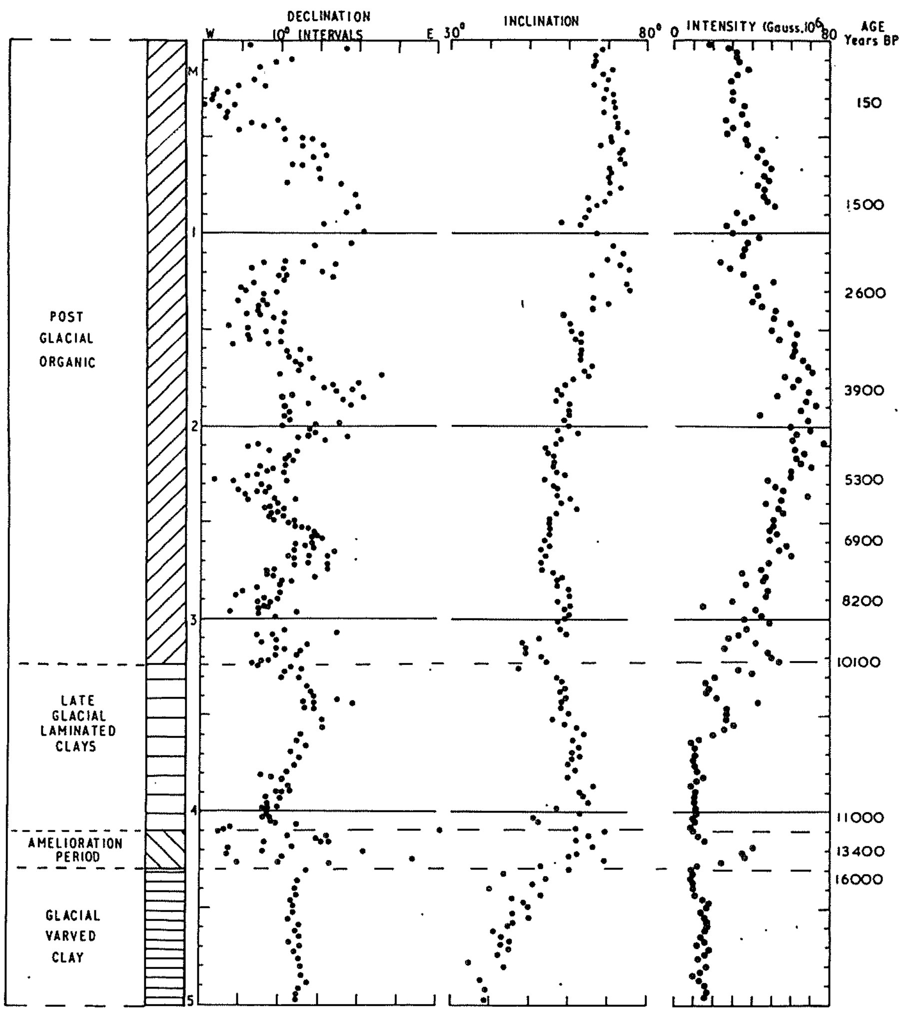

Fig. 5. Variation of declination, inclination and intensity for the last 15,000 years in 6 m cores of sediment from Lake Windermere. The age of the sediment is known from ^{14}C studies.

are common in lakes and a continuous record spanning about 15,000 years can be recovered. The sediments are very soft and are collected in plastic core tubes 6 m long. A strong, stable magnetic remanence is carried by haematite, and is probably of a chemical origin having grown close to the sediment/water interface in oxidizing conditions. Radiocarbon dating and pollen analyses provide a time scale for dating the changes in declination and inclination (Fig. 5). These results show that secular changes have continued over the past 15,000 years and have possibly been periodic, and also confirm that the mean inclination has been very close to that predicted by a geocentric axial dipole field model, i.e. the pattern of the geomagnetic field is the same as that produced by a bar magnet (dipolar) aligned along the earth's rotation axis (axial) and is centred at the middle of the earth (geocentric). A useful application of these variations is that once the secular changes have been accurately dated they in turn can be used to date sediments from other lakes, simply by measurement of their magnetic remanence. This can be accomplished very rapidly and nondestructively by spinning the whole core tube and using the computerized fluxgate magnetometer described above.

Palaeointensity measurements, made mainly on archaeological material, show that the decrease of the earth's dipole moment, recorded directly during the past 130 years (Fig. 6) began about 1500 years ago. Then the dipole moment

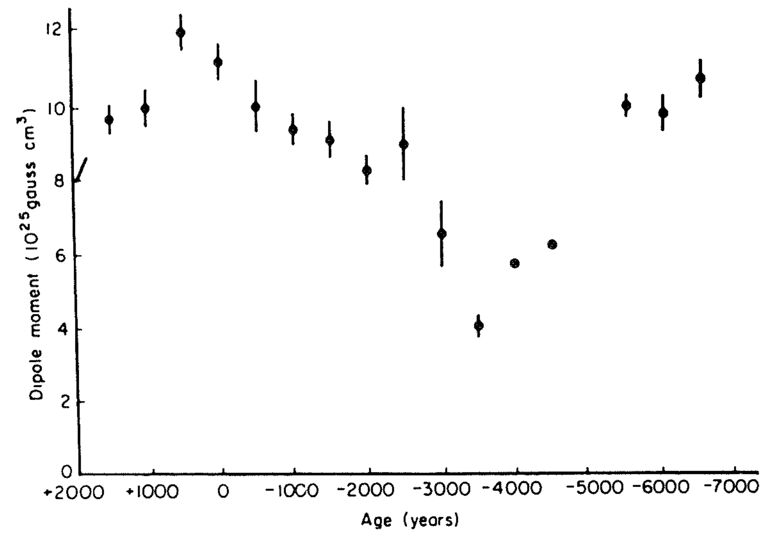

Fig. 6. Variation in the geomagnetic intensity over the past few thousand years. Palaeointensity determinations averaged over 500 year intervals with standard errors shown by vertical lines. Changes in intensity over the last 130 years as deduced from direct observations are shown by the diagonal line. From Smith (1970).[8]

had a peak value 50 per cent higher than the present moment of $8{\cdot}0 \times 10^{25}$ gauss cm^3. Before this decline the dipole moment had fluctuated quite markedly with a time scale of about 4000 years (Fig. 6).

Axial dipole field

The best fitting dipole to the earth's magnetic field lies along an axis inclined at about 11° from the axis of rotation and accounts for about 90 per cent of the main field. The residual field is known as the non-dipole field. Both the dipole and non-dipole fields vary with time, so producing secular variation. Now palaeomagnetic pole positions are calculated on the hypothesis that the time-average magnetic dipole field of the earth closely approximates a geocentric axial dipole. This means that the non-dipole field components should average out to zero and that the main field is aligned with the axis of rotation rather than an axis which is ten or more degrees out of alignment, as at present. It is important to establish if this has been the predominant effect over geological time.

Palaeomagnetic pole positions from the Quaternary (Fig. 7), although scattered, group strikingly around the rotation axis. Also the mean magnetic inclinations of deep sea cores, taken from all the world's oceans, show no significant departures from those expected using an axial geocentric dipole model. Both these results suggest that over periods of the order of 10^5 to 10^6 years the dipole field has indeed been central and aligned with the axis of rotation. As will be described below, palaeoclimatic information from several geological periods agrees with palaeolatitudes based on palaeomagnetic results, and hence lends further support to the hypothesis of a geocentric axial dipole existing throughout geological time.

Reversals of the earth's magnetic field

In 1906 Brunhes described some lavas from France which were magnetized in a direction exactly opposite to that of the present geomagnetic field, and he suggested that the earth's magnetic field had been reversed in the past. Mercanton in 1926 extended the records of rocks with a reverse magnetization to the southern hemisphere, and Matuyama in 1929 constructed the first magnetic stratigraphy by showing that there was a relationship between the magnetic polarity of Japanese lavas and the time of their extrusion. The concept of magnetic field reversals, however, was not widely accepted until more detailed studies were carried out in the 1950s and 1960s.

An alternative explanation to field reversal is that rocks magnetized in a direction opposite to that of the present magnetic field have distinct chemical or mineralogical properties which cause their magnetization to self-reverse. Several theoretical ways in which self-reversal can occur have been described.[9] For example, a rock which has two distinct, but intergrown, magnetic mineral phases with different Curie points would when cooling firstly obtain a magnetization, in the direction of the ambient field, in the phase with the higher Curie point. The magnetic field in the region around this high Curie point mineral phase would then be dominated by its thermoremanent magnetization. The direction of this magnetic field would be largely opposite to that of the ambient magnetic field (as in the case of the magnetic field close to a bar magnet). Thus the second intergrowth phase, still above its Curie point, would be subjected to the field of the first phase and on cooling through its own, lower Curie point would acquire a magnetization antiparallel to the external applied field. If on further cooling the magnetization of this lower Curie point phase was stronger than that of the

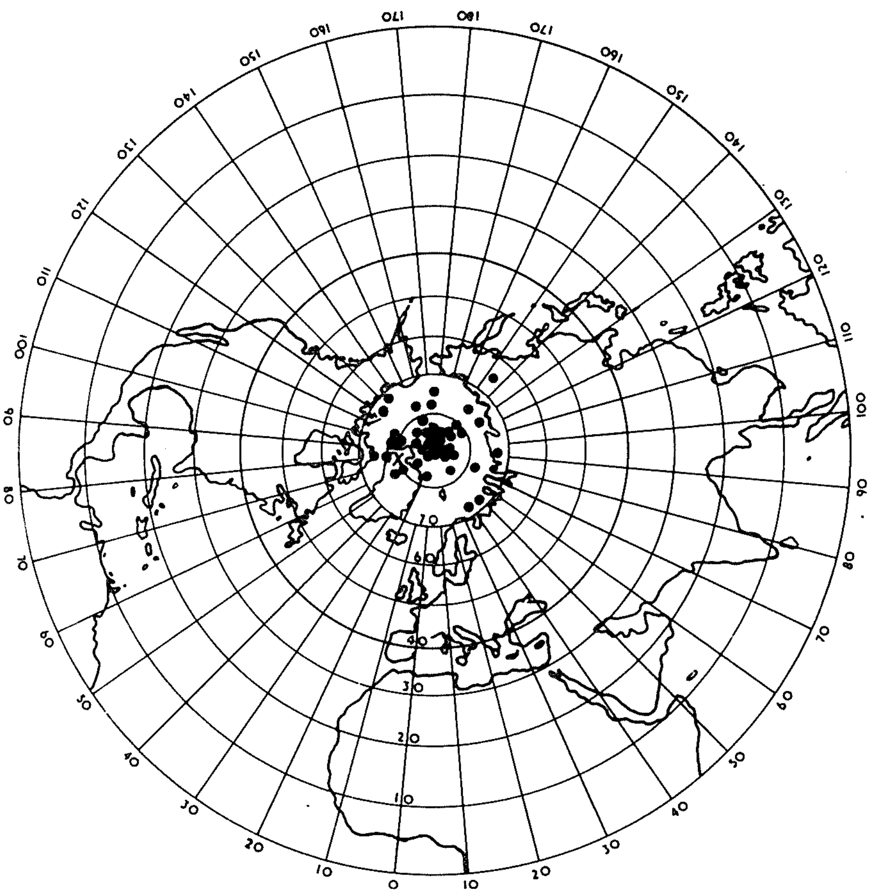

Fig. 7. Polar stereographic projection of palaeomagnetic poles (circles) for the last 5 million years. Cross is the present geomagnetic pole. Note how the palaeomagnetic poles cluster around the geographic pole rather than the geomagnetic pole.

higher Curie point phase, a self reversal would occur. Natural rocks which have self-reversed, and can be demonstrated to do so on cooling in an applied magnetic field in the laboratory, have been discovered. However, less than 1 per cent of rocks tested in the laboratory exhibit the property of self-reversal, and furthermore the phenomenon is restricted to minerals of a very limited range of chemical composition. Many palaeomagnetic results, some of which are described below, now provide convincing evidence that the main explanation for rocks acquiring a reverse magnetization is that of magnetic field reversal rather than self reversal.

Palaeomagnetic measurements on rocks formed during the last few million years show that the remanence lies generally within 30 degrees of the normal or reverse direction of the earth's magnetic field, and only rarely at an intermediate direction. Also about half the rocks measured hold a magnetic remanence parallel to the present magnetic field direction and the rest hold a remanence in the opposite direction. The combination of potassium–argon dating studies with palaeomagnetic polarity determinations has made it possible to answer the question of whether there have been distinct intervals of geological time within each of which all newly-forming rocks acquired the same magnetic polarity. The investigations have shown clearly that there have been intervals when rocks from all over the world had one and the same polarity of magnetization and alternating with these, other intervals when all rocks acquired a magnetization of the opposite polarity.[10] In the early 1960s a reversal time scale was first established and throughout the 1960s the scale was extended and defined more accurately to produce the picture shown in Fig. 8. Over 150 radiometric ages and polarity determinations were used to construct the sequence of reversals for the last 5 million years. The clear grouping of normal and reverse data in a distinct sequence thus conclusively confirmed the reality of field reversals. Four major normal and reverse polarity intervals were defined and named epochs. Reversals of a shorter duration of the magnetic field were discovered and named events (Fig. 8). Now that the reversal time scale has become more accurately determined it has been found that there is no natural division between epochs and events, but there is a continuous distribution of the lengths of polarity intervals from periods of 10^4 to 10^7 years.

Although potassium–argon dating is considerably better than any other dating technique for lavas with an age of the order of a few million years, it is limited to an accuracy of about 3 per cent. This means that beyond about 5 million years ago the resolution of potassium–argon dating is such that the possibility of recognizing individual events is very limited. However, exceptionally long periods of constant polarity can be distinguished much further back in geological time. For example, the Kiaman reversed polarity interval has been shown by potassium–argon dating to have begun about 300 million years ago and to have ended about 240 million years ago.

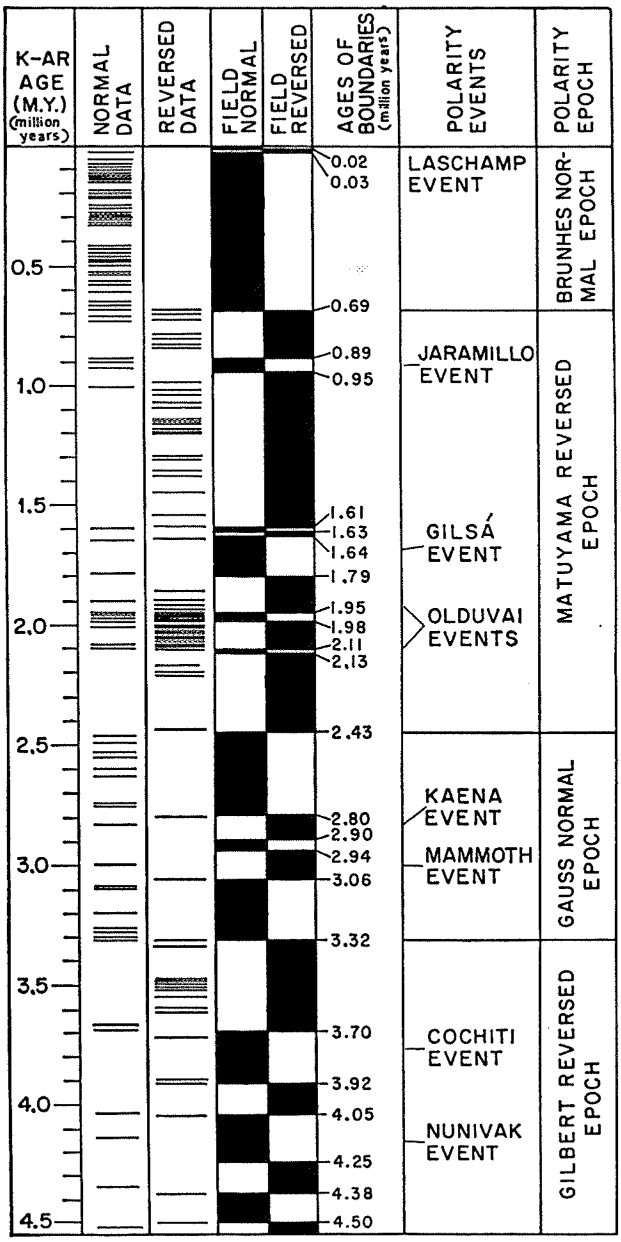

Fig. 8. Time scale for geomagnetic reversals. Each short horizontal line shows the age as determined by potassium–argon dating and the magnetic polarity of one volcanic cooling unit. Normal polarity intervals as shown by the solid portions of the 'field normal' column, and reversed-polarity intervals, by the solid portions of the 'field reversed' column. [From Cox (1969) *Science* **163**, 237–245. © 1969 by the American Association for the Advancement of Science.]

Deep sea sediments

The reversal time scale as shown in Fig. 8 can be used to date sequences of strata which carry a stable magnetic remanence. An important application of this kind is in the dating of oceanic sediments.[11] Deep sea sediments have been shown to record the sequence of epochs and events of magnetic reversals as dated from the continental lava sequences, and hence the rate of deposition of the oceanic sediments can be calculated. This magnetic stratigraphy has proved invaluable for long distance core to core correlations between different facies and faunal provinces. Palaeomagnetic measurements from oceanic sediment cores over 25 m long provide a possibility of extending the reversal time scale to about 15 million years ago by extrapolating the result using the rate of sedimentation deduced for the topmost sediments. Intensity measurements from the cores show that at field reversals the magnetic field intensity decreases and rebuilds over a period of about 20,000 years, whereas the polarity change is completed in about a tenth of that time. These estimates are in good agreement with ones based on results from continental lava flows derived from a comparison of the number of flows which were extruded during a time when the earth's magnetic field was in the process of reversing, with the number extruded during a time when the main field was stable, over a set number of reversal periods. Also magnetic intensity measurements from deep sea sediments and lavas show that when the magnetic field is reversing the field intensity does not fall to zero but only to about one quarter of its usual intensity.

Another interesting result from the palaeomagnetic research on deep-sea cores is that a number of independent investigations have revealed a close correlation between reversals and faunal extinctions.[12, 13] For example, during the last 2·5 million years, eight species of widely distributed Radiolaria became extinct. Six of the eight species died out at a time of polarity change. An example of the extinction of Radiolaria close to the Brunhes-Matuyama boundary as revealed in Antarctic cores is shown in Fig. 9. Three main explanations have been proposed to account for these correlations. (i) During reversals the earth's magnetic shielding against cosmic rays would be reduced so that the amount of radiation reaching the earth's surface would increase and might accelerate mutation rates and affect the course of evolution. (ii) The solar wind might produce changes in the upper atmosphere during reversals which might influence the climate. (iii) Geomagnetic reversals might directly influence organisms. However, some workers feel that all the mechanisms proposed to date are inadequate, and some have suggested that the correlations between extinctions and reversals are simply due to chance. Much additional information will have to be obtained before these problems are solved.

Sea floor spreading

One of the most important applications of palaeomagnetism has been in

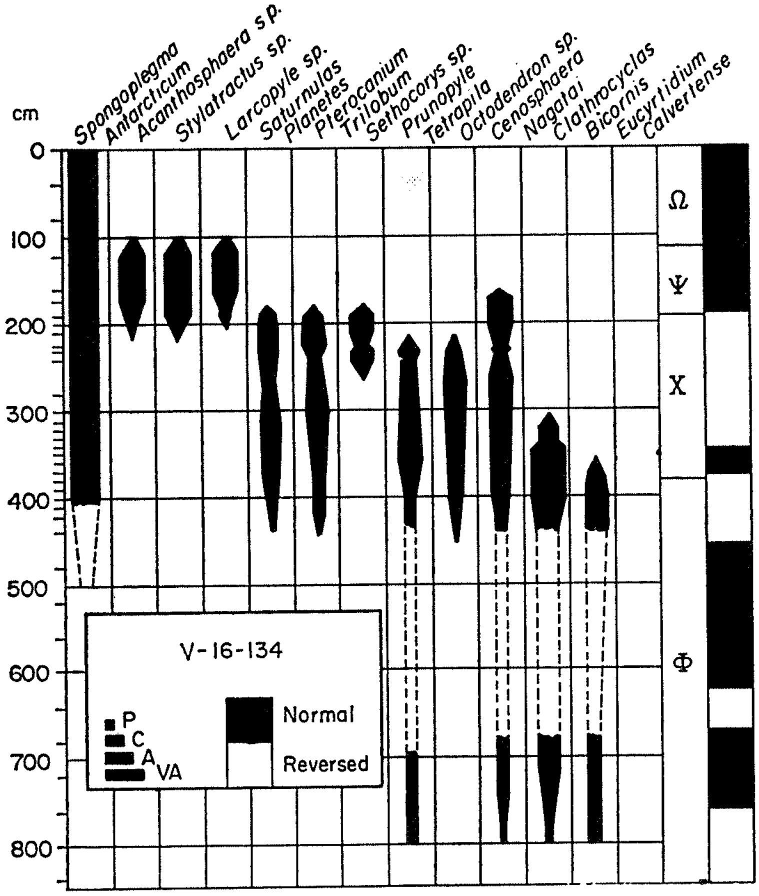

Fig. 9. Extinction and appearance of Radiolaria near the time of the Brunhes-Matuyama boundary in Antarctic deep-sea cores. Abundance of species illustrated by width of columns: P, present; C, common; A, abundant; VA, very abundant. Magnetic stratigraphy shown in right hand column. [From Opdyke *et al.* (1966) *Science* **154,** 349–357. © 1966 by the American Association for the Advancement of Science.]

providing a time scale for sea floor spreading. The concept of sea floor spreading is briefly that oceanic crustal material is continuously being moved laterally by convection currents in the mantle. Basaltic ocean floor is constantly being regenerated at, and spreading outwards from, mid-oceanic ridges which are located above upwellings of the convection cells, and is being destroyed by sinking beneath island arcs. Continental blocks 'drift' along with the oceanic material on the backs of the convection cells and may collide with other blocks to form mountain chains as in the case of the Himalayas, where India collided with Asia. Vine and Matthews in 1963[15] suggested that reversals of the earth's magnetic field might be frozen into the cooling material at mid-oceanic ridges. As sea floor spread outwards from a ridge crest, strips of alternately normal and reverse magnetic polarity, lying parallel to the mid-oceanic ridge, would be formed (Fig. 10). Recordings from fluxgate and proton-precession magnetometers towed at sea level by oceanographic research ships have revealed linear magnetic anomalies which run for thousands of kilometres parallel to the oceanic ridges, and which are alternately high and low, with steep gradients in between (Fig. 11). On the Vine-Matthews hypothesis these strips of high and

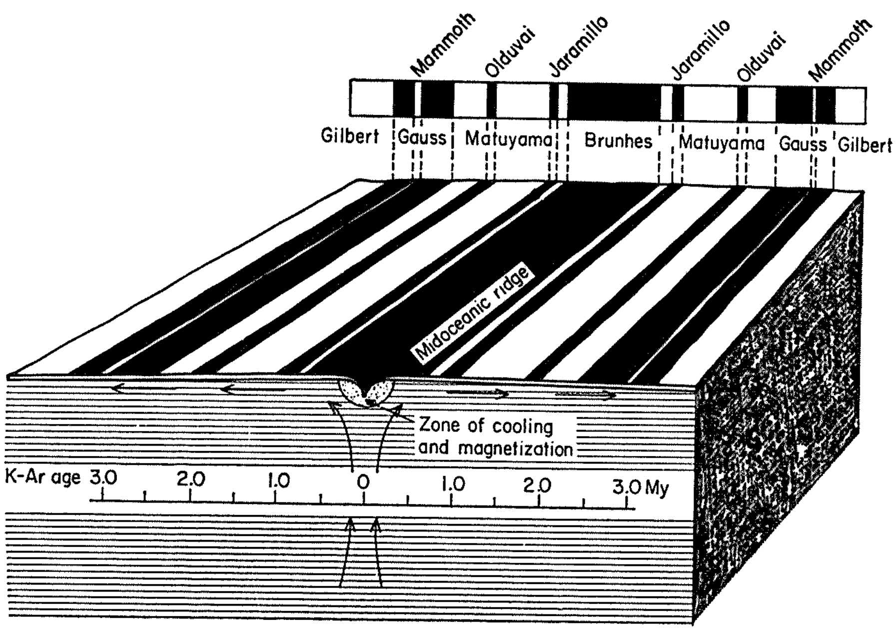

Fig. 10. Schematic representation of the principle of sea-floor spreading and reversals of the earth's magnetic field. Normal polarity zones are shown shaded. [Figure as drawn by McElhinny (1973)[14].]

low magnetic anomalies result directly from the strips of ocean floor with alternately normal and reverse polarity thermoremanent magnetization. In the mid 1960s[15, 16, 17] it was shown that the reversal time scale as deduced from continental lava flows for the last 3·5 million years matched extremely closely with the sequence of anomalies observed near ocean ridges. So the sea-floor spreading hypothesis was vindicated and could then be quantified. A spreading rate of a few centimetres per year was found to fit the results. By adding the assumption that spreading rates have been constant for periods longer than

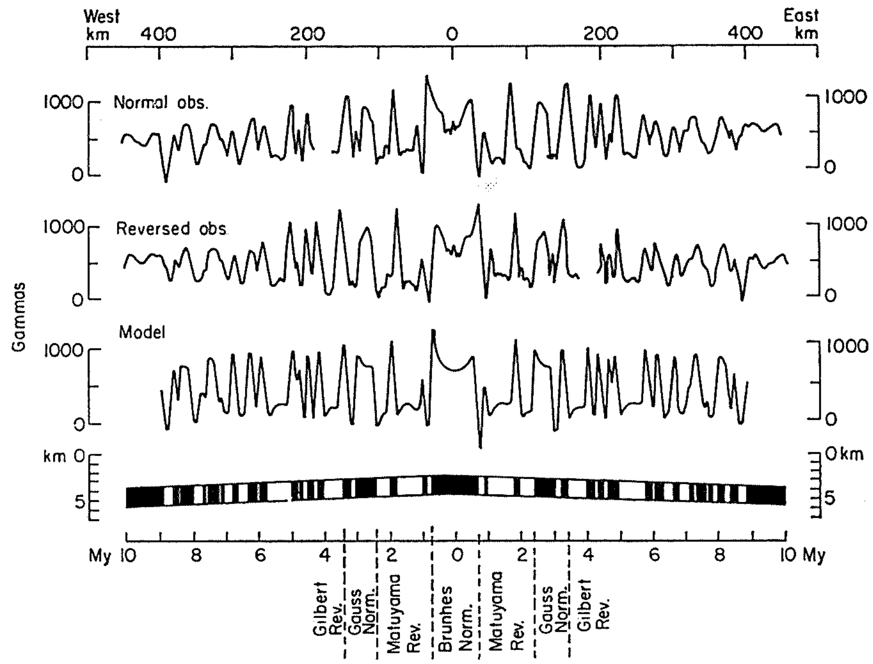

Fig. 11. Observed and reversed Eltanin-19 magnetic anomaly profiles across the Pacific–Antarctic ridge. The time scale is calculated using a constant spreading rate of 4·5 cm/year deduced from the central anomalies, and the reversed sequence before the Gilbert Epoch is inferred from the observed magnetic anomalies. [Figure as redrawn by McElhinny (1973),[14] after Pitman & Heirtzler (1966) *Science* **154,** 1164–1171. © 1966 by the American Association for the Advancement of Science.]

3·5 million years, a reversal sequence can be inferred from the older oceanic magnetic anomalies. A reversal time scale back to the Mesozoic has thus been constructed by extrapolating across the ridge flanks. Although errors associated with such a large extrapolation could be serious, recent results, especially dating of sediments from the JOIDES deep-sea drilling programme, have provided independent confirmation of the magnetic time scale.

Origin of magnetic field reversals

It is now generally accepted that the earth's magnetic field is produced in the earth's molten core by a self exciting dynamo process. Convection currents produce fluid motions in the iron core and under certain circumstances the resulting electric currents produce magnetic fields. Theoretical calculations suggest that if the patterns in the spherical core lack symmetry a dynamo action could exist which may fluctuate to produce secular variation and even reversals of the main dipole field. However, the problems of why reversals are so irregular in length and how they are related to the shorter time variations of dipole fluctuation remain. Cox[18] has proposed a model which accounts for many of the observed features of reversals. His probabilistic model for reversals assumes that the main geomagnetic dynamo is a steady oscillator which reverses when it is triggered by random variations in the much more rapidly fluctuating non-dipole field. On this model the average length of polarity intervals is related to conditions at the core–mantle interface, and the length of single polarity intervals is connected with random processes in the convecting core.

Continental drift

In addition to providing information about processes in the deep interior of the earth, palaeomagnetism provides quantitative information about processes which affect the surface layers of the earth, revealed as large lateral movements of the continents, or 'continental drift'. In the 1950s palaeomagnetic results revived interest in the suggestions that continents had moved appreciably with respect to one another in geological time.[19] Wegener had proposed in 1910 that before the Mesozoic all the continents were grouped in a single land mass or super continent which he called Pangaea. He suggested that this super continent then slowly split up after the Triassic period. However, his considerable body of evidence failed to convince many scientists, and the discussion about continental drift stagnated until it was enlivened by palaeomagnetic results.

The most convenient way to represent palaeomagnetic data related to continental drift studies is in terms of palaeomagnetic pole positions. Palaeomagnetic pole positions are calculated from the measurements of magnetic remanent declination and inclination of geological formations means by using the assumption that the long-term average magnetic field on the earth has been that of an axial geocentric dipole, the axis being that of the earth's rotation. As this assumption has been shown to be valid for the past few million years, extending the model back to earlier times is simply an application of the geological principle of uniformitarianism. A relationship between magnetic inclination and geographical latitude can be readily derived from the axial geocentric dipole model. The relationship is most simply expressed in the form

$$\tan I = 2 \tan L$$

where I is the magnetic inclination and L the latitude. Also on the axial dipole model the palaeomagnetic declination is a direct measurement of the direction of magnetic north. Thus a palaeomagnetic pole position which is defined as the point where the dipole axis meets the earth surface can be calculated from measurements of remanent magnetic declination and inclination alone.

Palaeomagnetic pole positions from consecutive geological periods for a single continent (e.g. South America, Fig. 12) are linked together to produce a polar

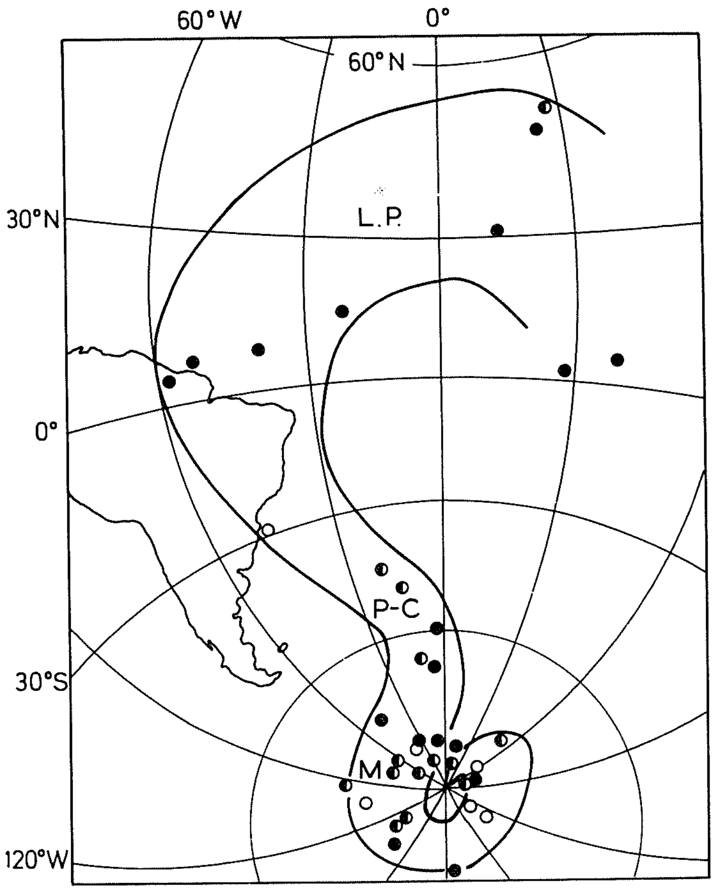

Fig. 12. Phanerozoic South American polar wander path (the width gives an indication of the accuracy of the path). Circles represent individual palaeomagnetic pole positions: solid circles, reverse polarity; open circles, normal polarity; shaded circles, mixed polarity. L.P. Lower Palaeozoic; P-C, Permo-Carboniferous; M, Mesozoic.

wander path. During the 1950s the broad outlines of polar wander curves were constructed for several continents. It was immediately apparent that the paths for all continents had moved through many tens of degrees of arc during the Phanerozoic, i.e. that the continents had moved with respect to the rotation axis. By comparing the polar wander paths from different continents (e.g. Europe and North America) it was possible to show that the continents had moved away from each other since the Triassic. The quantity of palaeomagnetic data has expanded considerably since the 1950s and polar wander paths for all the continents are now known in much greater detail.[20] However, results from some continents for particular periods of geological time are still sparse. For reconstructing the positions of continents during the Mesozoic phase of continental drift, it is clear that sea floor spreading, plate tectonics and continental edge fitting techniques have to a large extent superseded palaeomagnetism. But for older periods of geological time palaeomagnetic data remain invaluable for determining past latitudes and continental reconstructions.

An efficient way to use palaeomagnetic data is to use reconstructions of Pangaea derived from the fitting of the continental edges[21] and to rotate all continents along with their polar wander paths back to a Pangaean reassembly.

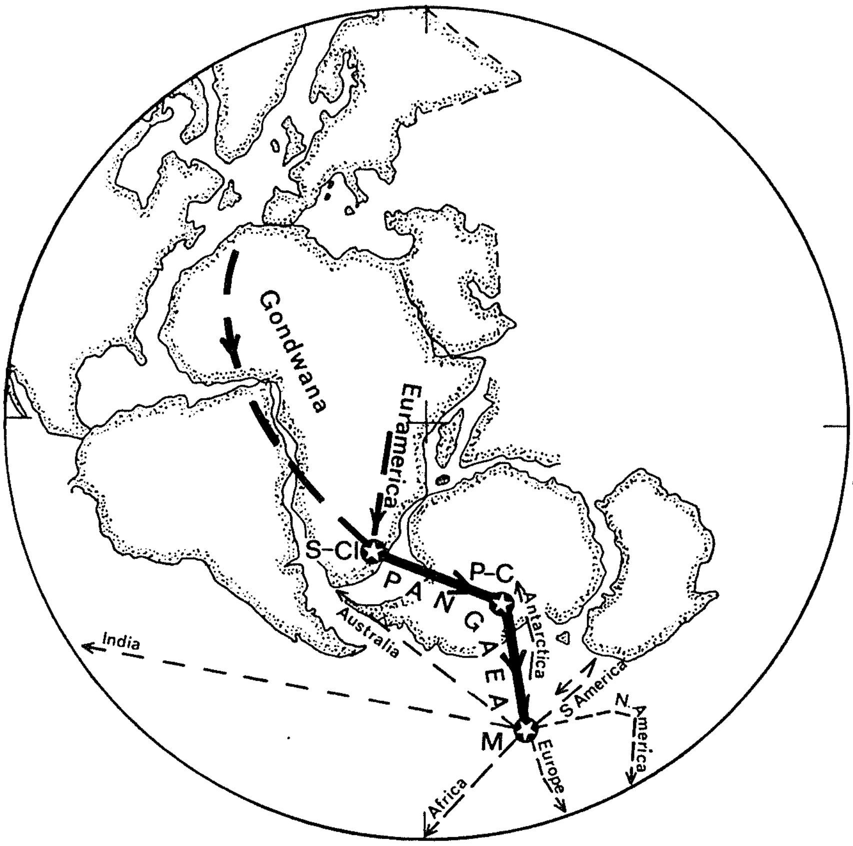

Fig. 13. Generalized diagram showing the path of the south pole during the Phanerozoic. The two continents of Gondwanaland and Euramerica combine into Pangaea during the Silurian and then there exists a common polar-wander path until the Mesozoic (M). The paths from the present continents then diverge as shown. [Modified from McElhinny (1973).[14]]

Phanerozoic palaeomagnetic results are presented diagrammatically in this form in Fig. 13. From Lower Carboniferous to Triassic times polar wander paths from all continents lie on a common curve which passes from South Africa across Antarctica. This curve represents the locus of the positions of the south pole during the time Pangaea was in existence. During the Mesozoic, polar wander paths of the present-day individual continental blocks separate, as shown in Fig. 13, as a result of the fragmentation of Pangaea. The amount of divergence between polar wander paths at any particular time is a measure of the degree of separation of the continents. Before the Carboniferous the polar wander paths do not lie on a single curve but separate paths for various earlier continents can be seen to converge as Pangaea is welded (Fig. 13). Thus the convergence of the Gondwanan and Euramerican polar wander paths before the Lower Carboniferous resulted from the meeting of the Gondwanan and Euramerican plates which culminated in the formation of the Hercynian and Appalachian mountain chains where the two blocks collided. The pattern and extent of continental drift before the existence of Pangaea is thus becoming clarified as new palaeomagnetic results are obtained for the Lower Palaeozoic. The problem of continental drift and continental reconstruction in the Precambrian is much more difficult to decipher because of the vast span of time covered by the Precambrian. Attempts at reconstruction are now being made for certain continents, but many more years of palaeomagnetic research will be required before a reliable picture can be built up.

Palaeoclimates

The large latitudinal movements of continents deduced from palaeomagnetic data naturally suggest the question of whether palaeoclimatic data, which should reflect palaeolatitudes, agree with the proposed reconstructions. A range of types of palaeoclimatic data which includes the distribution of evaporites and red-beds, the variation in diversity of temperature-dependent organisms, and $^{18}O/^{16}O$ palaeotemperature measurements have been used to test palaeolatitudes as deduced from palaeomagnetic measurements. Unfortunately, the palaeoclimatic data are too limited, too scattered or not necessarily latitudinally dependent to produce definite conclusions. In many cases, however, there is often good agreement. Ancient glacial deposits, for example, agree well with palaeomagnetic prediction of polar regions. Widespread Permian glacial deposits are found in South Africa, India, Australia and eastern South America, and Ordovician glacial deposits are found in the Sahara. (Compare these localities with the position of the south palaeomagnetic pole for these periods in Fig. 13.) It is interesting to note that in North America the Ordovician sequences consist of thick limestones and beds characteristic of low latitudes, in confirmation of the Lower Palaeozoic separation of North America and Gondwanaland.

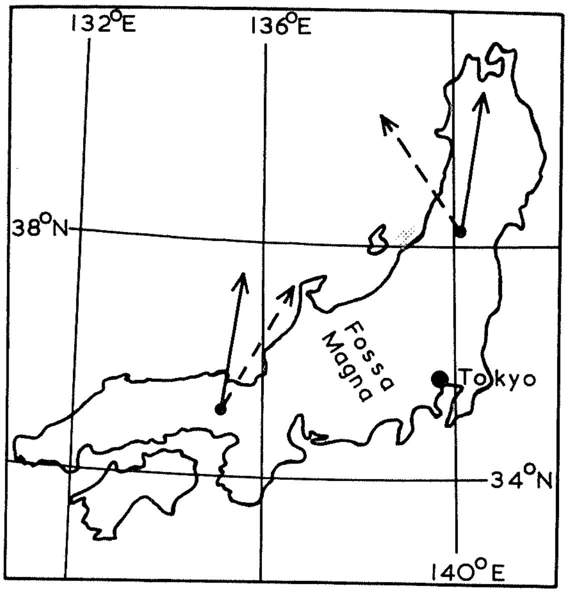

Fig. 14. Example of land mass rotation. Horizontal component of the mean direction of magnetization from Cenozoic rocks (dashed arrow) and pre-Tertiary rocks (solid arrow) from southwestern and northeastern Japan.

Tectonic movement of small continental blocks

Palaeomagnetism has also found application in the small scale movements of continental blocks,[22] particularly where the movement has been one of rotation which is revealed by differences of magnetic declination which was acquired by rocks before the time of rotation of the blocks. A classic example of this type of application is in the bending of Japan. Cretaceous palaeomagnetic declinations from the southwest part of Japan point towards the east whereas declinations from northeast Japan point towards the west. The difference in declination of remanence between the two parts is about 60–70° (Fig. 14). In contrast Tertiary and Quaternary rocks yield declinations from the two parts which agree very closely with each other and with geographical north (Fig. 14). The palaeomagnetic declinations thus strongly suggest that after the Cretaceous Japan was deformed resulting in a relative rotation of the two areas. This is substantiated by the fact that the mean long axes of the southwest and northeast parts of Japan lie at an angle of about 120°, indicative of a bending of about 60°.

The Moon

The magnetic properties of lunar rock samples returned from the Apollo missions have been investigated and provide intriguing information about the history and constitution of the Moon. At present the Moon has no measurable main magnetic field. If it does possess a magnetic field its value must be extremely weak and less than 10 gammas.

Crystalline lunar samples between 3700 and 3400 million years old have been shown to possess stable natural remanent magnetizations by tests using partial thermal, and alternating field, demagnetization techniques. Thermomagnetic and hysteresis experiments have shown that the remanence is carried by native iron particles. Partial thermoremanent magnetization experiments (as described in section 4) on the intensity of the remanence have shown that an ambient field of about 1 oersted (10^6 gammas) must have been present when the stable magnetization was acquired 3500 million years ago. The lunar samples were not in situ when collected but were simply loose rocks on the lunar regolith. Hence the direction of the ancient lunar field is not known.

Because of the lack of lunar magnetic field today, the origin of the magnetization and source of the ancient ambient field are problematic. It is thought that unusual sources of magnetization, e.g. shock, thermal cycling and meteorite impact, can be discounted, and that there must indeed have been a magnetic field of internal or external origin present during the early history of the Moon. The magnetic field must, of course, have since died away. Possible sources of an external magnetic field are the Sun and Earth but neither are satisfactory. The past intensity of the Sun's magnetic field has most probably not been strong enough to produce the required field value at the Moon because stars of the same type as the Sun do not have strong magnetic fields. The past intensity of the geomagnetic field has always been of the same order as at the present day. To produce a strong enough field at the Moon's surface the Moon must have been much closer to the Earth 3000–4000 million years ago. However, it could not have remained close enough to the Earth over the period of hundreds of millions of years during which the lunar material acquired a strong stable remanence. Hence the ambient magnetic field was most probably of internal origin, possibly produced by a dynamo action in a small, molten, iron core which has since solidified.

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