Geophys. J. R. astr. Soc. (1967) **12**, 321–362.

The Intensity of the Ancient Geomagnetic Field: A Review and Analysis

Peter J. Smith*

(Received 1966 December 29)†

Summary

In this review all palaeomagnetic determinations of the intensity of the ancient geomagnetic field known to the author at 1966 June are brought together and analysed. The lists of data cover determinations from both historic (archaeomagnetic) and geological specimens. The problems encountered in the determination of ancient field intensities from rocks, especially that of rock alteration during the formation of artificial remanence and that of the presence of secondary components of magnetization, are discussed, and methods used to overcome them are described. Experimental techniques used hitherto are briefly summarized. Methods of treating data are reviewed; and suggestions for comparing data from different latitudes are made.

The main conclusions drawn from an analysis of both the field intensity data and the problems involved in obtaining this data are:

(i) that the problem of rock alteration during the heating necessary to induce an artificial TRM may be overcome by heating selected naturally highly-oxidized specimens in air;

(ii) that the effects of secondary components of magnetization may be avoided by utilizing that part of the primary moment which remains after the removal of secondary moments by thermal or a.c. demagnetization;

(iii) that the geomagnetic dipole is not constant within any given polarity but fluctuates in strength, possibly with a period of the order of 10^4 years;

(iv) that because of these fluctuations and non-dipole field variations no conclusions regarding the mean strength of the geomagnetic dipole at any given time may be made if only a few rock samples are measured, and hence that future workers should use sufficient samples to average out these effects;

(v) that the limited data from transition zones suggest that during a field reversal the dipole moment reduces to zero but the non-dipole field remains.

* Present address: US Geological Survey, Menlo Park, California 94025.

† Received in original form 1966 September 2.

1. Introduction.
2. Basis of field intensity determination.
3. Experimental techniques.
4. The recent geomagnetic field.
5. Treatment of data.
6. Historic and archeomagnetic field intensity data.
7. Geological field intensity data.
8. Discussion.
   8.1. Ancient geomagnetic field intensities: historic and archeological time.
   8.2. Ancient geomagnetic field intensities: geological time.
   8.3. The spread of virtual dipole moments caused by non-dipole fields.
   8.4. Fluctuations of the Earth's dipole moment.
   8.5. Ancient geomagnetic field intensities: transition zones.
9. Conclusions.
10. Acknowledgments.
11. References.

1. Introduction

Since the beginning of this century, and especially during the past decade, considerable effort has been devoted to the determination of the direction of the Earth's main magnetic field at various localities and over the past 600 m.y. or more by utilizing palaeomagnetic measurements on rocks of various types [for the most recent and complete summaries see Irving (1964) and Irving & Stott (1960–1965)]. Comparatively little work, however, has been carried out on similar studies of the ancient intensity of the Earth's field. Undoubtedly the experimental difficulties encountered in the determination of ancient field intensities from rocks and the uncertainties in the interpretation of the results obtained are much greater than in the corresponding measurements of direction only. Nevertheless, in spite of these barriers, several attempts have been made to rectify the lack of data since the pioneering work of Thellier (1938) and Koenigsberger (1938). The momentum of this branch of palaeomagnetism is gradually increasing; and the results are likely to be of no less importance than their directional counterparts.

Many of the studies carried out hitherto, covering both historic and geological time, have been published in theses, conference proceedings, internal publications and other forms, some of which are not readily accessible (including several in foreign languages). It is thought, therefore, that a comprehensive review and analysis of ancient field intensity determinations may be useful at the present time for the following reasons: to bring together in one place and in a fairly detailed form, all field intensity data so far published; to compare results from the various authors; to assess the reliability of the results; to make suggestions on the treatment of data; to summarize and critically examine experimental techniques used; and to interpret the data in the light of physical processes occurring within the Earth, wherever possible.

With regard to the last point, palaeomagnetic measurements of direction have been invoked in large part to explain surface phenomena, or conversely, the existence of surface phenomena (for example, continental drift) has been inferred from the results obtained. In the case of intensity data, on the other hand, emphasis is likely to be placed on the interpretation of processes within the Earth. Dynamo theories of the origin of the Earth's magnetism place the source of the main field within the core; and hence intensity determinations might be expected to lead to an increase in our knowledge of processes in the core. But the first objective must simply be to determine the magnitude of the Earth's field as a function of time. Since this is yet far from realization it may be premature to formulate more detailed questions; but three basic problems may be mentioned here:

(i) From an adequate supply of basic data it should be possible to detect differences between the normal and reversed intensity fields, if such differences exist.

(ii) Similarly it should be possible to determine whether field reversal takes place through a 'flip-over' of the main dipole, or, as the dynamo theory implies, by a decay of the dipole moment to zero followed by a build-up in the opposite sense.

(iii) It should also be possible to detect any 'fine structure' of the intensity field either due to secular variation or actual fluctuations of the dipole intensity.

A previous review of ancient field intensities has been published by Thellier & Thellier (1959). In addition, chapters on this subject by Doell & Cox (1961) and Irving (1964) have been published as sections of longer treatises on palaeomagnetism. Owing to the increase in the amount of data made available during the past few years, these are now incomplete.

2. Basis of field intensity determination

The principle upon which the determination of ancient field intensities from rock magnetic data is based is extremely simple. Thermoremanent magnetization (TRM) is the process most commonly studied in actual determinations; and it may be shown that for a wide variety of rock types, the intensity of magnetization induced in a specimen by cooling it through its Curie point in a magnetic field is proportional to that field as long as it does not exceed about one oersted (Nagata 1943). Johnson *et al.* (1948) have also shown that up to field strengths of about twice that of the Earth's present field, the same proportionality is valid for detrital remanent magnetization (DRM). Thus if M_n is the natural remanence acquired by either TRM or DRM in an ancient field F, and M_a is the artificial remanence acquired by the same specimen under the same magnetization process in a field H:

$$F = H.\frac{M_n}{M_a}. \tag{1}$$

All field intensity results published so far have indicated the Earth's field to have been less than one oersted, and so the validity of the above formula appears not to be in doubt for the case of the Earth's field. In practice, however, its straightforward application is rarely possible for other reasons. Firstly, the value of M_n measured may bear little relation to the original intensity acquired in the field F. Secondary components of magnetization, such as partial TRMs due to slight heating or viscous components acquired in the Earth's field (Rimbert 1958), may have been superimposed on the primary component since the rock was laid down. Secondly, during the process of inducing the artificial remanence the nature and amount of magnetic material present may be altered, so that the original and artificial magnetizations are induced into essentially different substances. This is particularly likely to arise during the heating and cooling necessary to produce an artificial TRM. Finally, there may be difficulty in determining the type of natural magnetization possessed by the rock—a situation especially likely to arise in the case of sediments where it may be difficult to differentiate between DRM and chemical remanent magnetization (CRM). These factors drastically limit the final number of field intensity values which may be regarded as reliable.

The problems of alteration and secondary components will be considered in turn.

(i) *Alteration.* The direction of magnetization in a rock is measured relative to external standards (the horizontal plane and the geographic pole) which do not change and, in particular, are not functions of the rock itself. In order to determine the ancient field intensity, however, it is necessary to use equation 1. In this case the comparison standard is internal (in effect it is the constant of proportionality between the moment and the magnetizing field). This standard is a function of the rock and hence may change.

Because of the necessity of using an internal comparison standard a rock used for field intensity determination must be stable against change during the remagnetizing process. The best approach to the problem of alteration during heating is therefore to select only those specimens which possess small alteration potential. The main type of change liable to take place during heating is oxidation (Akimoto *et al.* 1957, Akimoto *et al.* 1959, Ade-Hall *et al.* 1965, Havard *et al.* 1965) although reduction, ionic ordering, annealing and phase homogenization may also play a part in some cases. Low Curie point rocks tend to oxidize on heating with a consequent undesirable increase in Curie point, and so an initially high Curie point is usually a necessary property. Naturally baked sediments and humanly baked objects have therefore found wide application in field intensity work because not only are their Curie points usually high, but the coercive forces involved are usually high also. They also frequently contain hematite, the most highly oxidized form of iron oxide. Further, since baking has already occurred before collection, many of the physical and chemical changes likely to take place during the laboratory heating will already have occurred before the initial magnetizing process. The Class 5 (highly-oxidized) lavas described by Wilson *et al.* (1967) and utilized for field intensity determination by Smith (1967) are a particularly useful type of igneous rock because, as with baked sediments, they usually possess high Curie points and are very stable against alteration during heating in air.

Many of these points apply to unbaked sediments also; but there are two properties which make this class of rock less suitable. Firstly, the intensity of magnetization is usually about a hundred times smaller than in baked sediments so that unless very sensitive equipment is available the accuracy of measurement will be low. Secondly, not only is it often difficult to differentiate between CRM and DRM in such cases, but these processes are more difficult to reproduce in the laboratory than TRM.

Despite careful selection of rocks, some degree of magnetic change almost always occurs during heating and hence it is necessary to apply a series of tests supplementary to the main field determination, and from the results of these decide whether to reject the intensity value as unreliable or to quote it with a specified error. Supplementary tests will be described below.

(ii) *Secondary components.* Magnetic moment is a vector quantity, that is, it may be represented by a magnitude and a direction. It is possible to increase or reduce the moment, or superimpose a second component at any angle to the original moment, without necessarily affecting the direction of that part of the original moment which remains. Thus as long as the primary component is not completely erased, some of the original directional properties will remain. In practice, in the case of rock magnetization, it frequently occurs that part of the original component remains; and the secondary components may then usually be removed by thermal or a.c. demagnetization. However, the mere acquisition of secondary components usually means that part of the primary has been erased. Even if this is not so it is not, in practice, possible to remove the secondary components without some of the primary being demagnetized simultaneously. In this sense, therefore, direction is more independent of secondary effects than is magnitude.

The problem of secondary components has undoubtedly held back the development of field intensity determination. Greater confidence may be placed in field intensity results obtained if any secondary components present are small; and this explains the emphasis which has been placed on historic and archaeological specimens. Small secondary components may not be a necessary condition, however. Some workers (Wilson 1961, Van Zijl 1961, Irving *et al.* 1965, Smith 1967) have utilized the magnetic moment which remains in the rock after the secondary components have been removed by thermal or a.c. demagnetization. It has been mentioned above that these processes will also remove some of the primary magnetization; but the assumption is that if both the natural and artificial remanences are compared (in equation 1) at the same temperature or at the same a.c. demagnetizing field, the proportion of the primary component remaining is the same in each case. This in turn is based on the assumption that secondary components (whether they be PTRMs or viscous magnetizations) are limited to the lower range of blocking temperatures (or coercive forces, where appropriate) and that the upper limit of blocking temperatures (coercive forces) affected is represented by the point on the vector heating (field) diagrams above which the direction of magnetization is constant. The possibility remains that so-called 'spontaneous decay' of the NRM has occurred above the limits of the secondary moments. There appears to be no mechanism whereby this can happen, and such decay is considered to be unlikely. It is not possible to prove its absence, however.

A further point which must be considered is whether the proportionality between the magnetic moment induced in the rock and the magnetizing field is valid under the conditions in which the rock is laid down and under the conditions in which the artificial remanence is induced in the laboratory. Nagata (1943), for example, showed that for a particular rock from Japan, although the TRM acquired during cooling in a small field was independent of cooling rate to a first approximation, at cooling rates greater than about 0·04°/s there was some diminution of the remanence. In the field, cooling rates are normally much lower than this, but this is not necessarily so in the laboratory. It is thus advisable to induce artificial remanences over as long a time as possible, and to avoid any rocks whose natural moment may have been produced during rapid quenching.

3. Experimental techniques

Several different empirical approaches to the problem of field intensity determination have been made, but by far the most commonly used technique (especially for historic specimens) is that developed by Thellier *et al.* (1938, 1941a, 1941b, 1942, 1946, 1959). Because of its wide usage, the various stages of the method will be summarized in some detail. It is applicable to any type of material which may be considered to be potentially useful for intensity work as long as the original magnetization is TRM; but for reasons described above baked rocks and other baked objects are most frequently used. It is a feature of the method that the reliability of the intensity value obtained is assessed on the basis of internal self-consistency of the results.

The following are the main steps in the method:

(i) The sample is stored for two weeks in the laboratory in the same orientation as at the collection site in order to restore any soft components which may have altered since collection.

(ii) After measurement of the remanence the specimen is rotated through 180° about a horizontal axis normal to the local magnetic meridian, again left for two weeks, and remeasured. The vector difference between the two measurements is calculated; and the size of the probable viscous component acquired since the initial magnetization is estimated on the assumption that the growth of viscous magnetization is proportional to log (time) (Rimbert 1958). If this estimate is greater than a few percent of the total magnetization, the rock is discarded as being too unstable magnetically.

(iii) A rock passing the above test is then heated to 100° C, cooled in the Earth's field at a known orientation and again measured. This is then repeated after another 180° rotation as described above. The vector sum of the two latter measurements represents twice the PTRM acquired in the initial geomagnetic field between 100° C and the Curie point. 100° C is considered by Thellier to be above the effect of any viscous component.

(iv) The rock is then heated to above its Curie point, cooled in the Earth's field, remeasured, rotated through 180° as before, heated to 100° C, cooled and again remeasured. The vector sum of these two latest measurements represents twice the PTRM acquired in the present geomagnetic field between the Curie point and 100° C.

(v) The two quantities from (iii) and (iv) are then used in conjunction with the known value of the present field and equation (1) to calculate F.

The process described above represents the Thelliers' method in its simplest form which utilizes only one temperature interval (100° C–Curie point). The range between 100° C and the Curie point may be sub-divided experimentally in order to determine values of F over several different but smaller temperature intervals, with the advantage that any chemical or phase change which takes place due to heating will produce different values of F over the different temperature ranges and hence expose itself. Any changes in the nature or amount of magnetic material present immediately become obvious and hence lead to rejection of the data as unreliable. Assuming no field change during the intial cooling and no physical or chemical change in the material during the subsequent heating and cooling, it follows from the additive property of PTRMs that:

$$F = H\frac{M_n(T_1-T_2)}{M_a(T_1-T_2)} = H\frac{M_n(T_2-T_3)}{M_a(T_2-T_3)} = \ldots H\frac{\Delta M_n}{\Delta M_a} = \ldots H\frac{M_n(T_i-T_j)}{M_a(T_i-T_j)}, \qquad (2)$$

where T_i, $T_j < T_c$ and F, H, $M_n(T_i-T_j)$ and $M_a(T_i-T_j)$ represent the quantities expressed in equation (1) over the temperature range (T_i–T_j). The experimental validity of equation (2) is thus automatically a test of reliability.

It is frequently the custom to plot ΔM_n against ΔM_a for any rock. Linearity indicates reliability of F values, though deviations at low temperatures due to viscous magnetizations are not usually a cause of rejection. Two examples of successful plots are shown in Fig 1, after Nagata *et al.* (1963). As an example of the errors involved Irving (1964) has calculated from the Thelliers' data in one case that for five different specimens from the same locality measured over three different temperature intervals, the mean standard deviations are: (i) 3% for same specimens over different temperature intervals, (ii) 8% for different specimens over the same temperature intervals.

A further test is to repeat the whole experiment, but this time to maintain the specimen at its Curie temperature for some considerable time. If the intensity values obtained from the second experiment differ from those obtained in the first, then some change must have taken place, and again the data are invalid.

It is clear that even in its simplest form the Thelliers' method is laborious and slow, but that any values finally emerging will almost certainly be free from systematic errors. It is, in fact, the most reliable method available; but it would be convenient to have a quicker and more robust method. Over the historic time scale, where single field intensity determinations are useful, the Thelliers' method is indispensable; but for geological specimens it is important to obtain a large number of field intensity values for any given period in order to average out the effects of secular variation and possible fluctuations in the dipole moment itself.

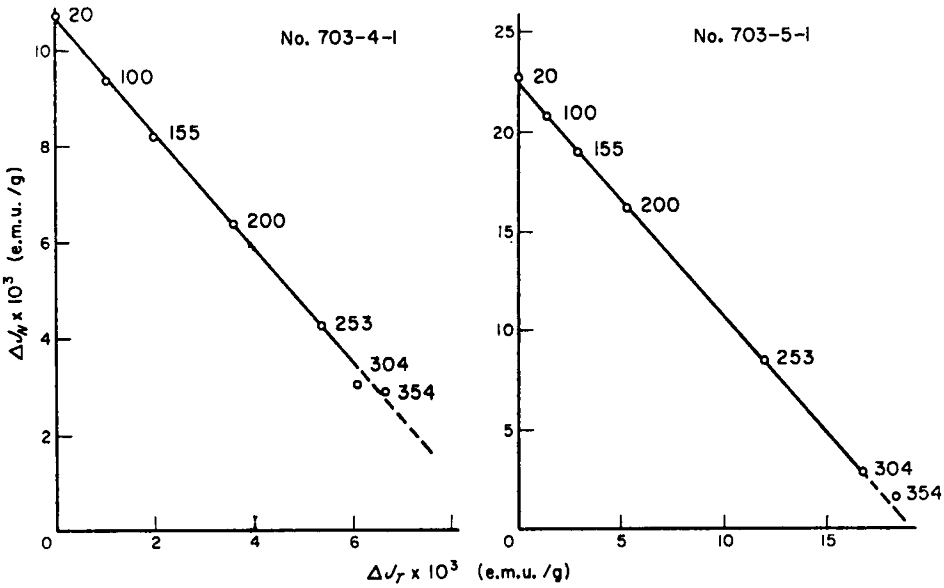

FIG. 1. Examples of plots of the change in natural moment against change in artificial TRM during heating of Japanese rocks (after Nagata, Arai & Momose 1963).

For geological specimens Wilson (1961), Momose (1963) and Smith (1967) have used a method based on the total TRM curves rather than on the PTRMs over various temperature intervals. The specimen under investigation is completely demagnetized thermally whilst its NRM is measured as a function of temperature. The rock is then cooled to room temperature in a known magnetic field, its TRM being measured during cooling as a function of temperature. (Heatings and coolings are carried out in air on specimens which are highly stable against oxidation.) It is then a simple matter to apply equation (1) at any particular temperature, T, or more correctly over any particular temperature interval, T_c-T, where T_c is the Curie point. Momose did this for the temperature interval between room temperature and the Curie point for rocks with small secondary moments, whereas Wilson and Smith calcuated F for various temperature intervals above the effects of secondary magnetizations.

The main disadvantage of this method lies in the fact that it is necessary to heat the specimen above its Curie point in order to give it a TRM and before any intensity values may be calculated. This means that internal consistency tests will not be able to show up any systematic change due to alteration of the magnetic material. It is necessary, therefore, to carry out supplementary tests to detect any such changes. These usually take the form of saturation magnetization, Curie point and low-field moment curve shape comparisons before and after heating. It must be realized, however, that there may be no correlation between high-field and low-field properties. Thus, for example, indication of no change in the saturation magnetization during heating does not necessarily give any information on the corresponding ability of the rock to acquire low-field magnetizations.

Van Zijl (1961) and Smith (1967) have also used the above simplified technique but with a.c. rather than thermal demagnetization. The specimen is a.c. demagnetized and its NRM measured as a function of a.c. demagnetizing field. It is then heated to above its Curie point and cooled in a known field. (Smith heated highly-oxidized specimens in air, Van Zijl heated specimens of unknown oxidation state in nitrogen.) The TRM is then a.c. demagnetized in the same way as the NRM; and equation (1) is applied at particular fields above the effects of secondary magnetizations. This method is especially suitable for igneous rocks which may normally be a.c. demagnetized in fields of only a few hundred oersteds. Supplementary tests must be carried out to estimate the degree of possible alteration taking place during the heating.

All the above techniques have been developed for use with rocks whose initial magnetism was due to TRM. In fact, TRM has been used most frequently for intensity determinations because of its greater stability generally; but one of the earliest determinations of palaeointensity was carried out by Johnson *et al.* (1948) on glacial varves and Pacific ocean bottom sediments whose magnetism was due to DRM. In the first case, after measurement the clays were ground and redeposited under water in a known field. F was then calculated using equation (1). In the second case, the natural magnetization was measured, after which each specimen was placed in a field of 2000 gauss. for a few seconds. On the assumption that all the magnetic grains involved were of the same size, the resultant moment was taken to be proportional to the quantity of magnetic material present. The ancient field was then calculated relative to the present field through the medium of a modern sample of known NRM and making allowances for different amounts of magnetic material in each sample. This method is based on assumptions which must be regarded as doubtful since no correlation between low-field and high-field behaviour can be guaranteed; but the results are consistent with more recent determinations.

These are the main techniques which have been used, although some others have been developed in specific cases. For example, Weaving (1960) has utilized the interaction between two components of magnetization in the same rock; but this method is not open to universal application.

In this section little attention has been paid to supplementary reliability tests; but these are detailed in the main tables of results and are generally self-explanatory.

4. The recent geomagnetic field

The most detailed knowledge we possess of the past geomagnetic field is derived from direct observation. However, magnetic field records have only been made for the past few hundred years; and the earlier measurements are neither complete nor accurate (Bullard *et al.* 1950). The Earth's field and its changes during the present century have been fairly comprehensively documented by Vestine *et al.* (1947). These authors have used all measurements made between 1905 and 1945 by themselves and others to produce in table and chart form a description of the secular variation for the epochs 1912.5, 1922.5, 1932.5 and 1942.5 and a complete description of the Earth's main field for 1945. Using this data, Bullard *et al.* (1950) have computed the non-dipole field for the epochs 1907.5 and 1945 by subtracting from the total field that due to the best-fitting dipole (the best-fitting dipole is that which fits the observed field best from the point of view of least squares). Details of the field for 1955 have been published by Finch *et al.* (1957); and measurements made between 1945 and 1964 have been analyzed by Leaton *et al.* (1965). The United States Air Force Geophysics Research Directorate publishes Earth magnetic data at five-yearly intervals. Fig. 2 shows the total field intensity (F) contours for the year 1960, taken from this source.

The relative ease with which palaeomagnetic directional measurements have been made is paralleled in the development of direct field observation. The south-seeking property of magnets was known to the Chinese during the eleventh century AD and possibly even as long as 4500 years ago. Magnetic declination had been discovered by at least 1300 AD; and Henry Gellibrand (1635) was the first to notice that it changed with time. In 1701 Edmund Halley produced the first isogonic chart for sea areas; and the first isoclinic chart for the whole Earth was published by Johann Carl Wilcke in 1768. Another 36 years were to elapse, however, before field intensity measurements were made, and these by Alexander von Humboldt (1804). By observing the number

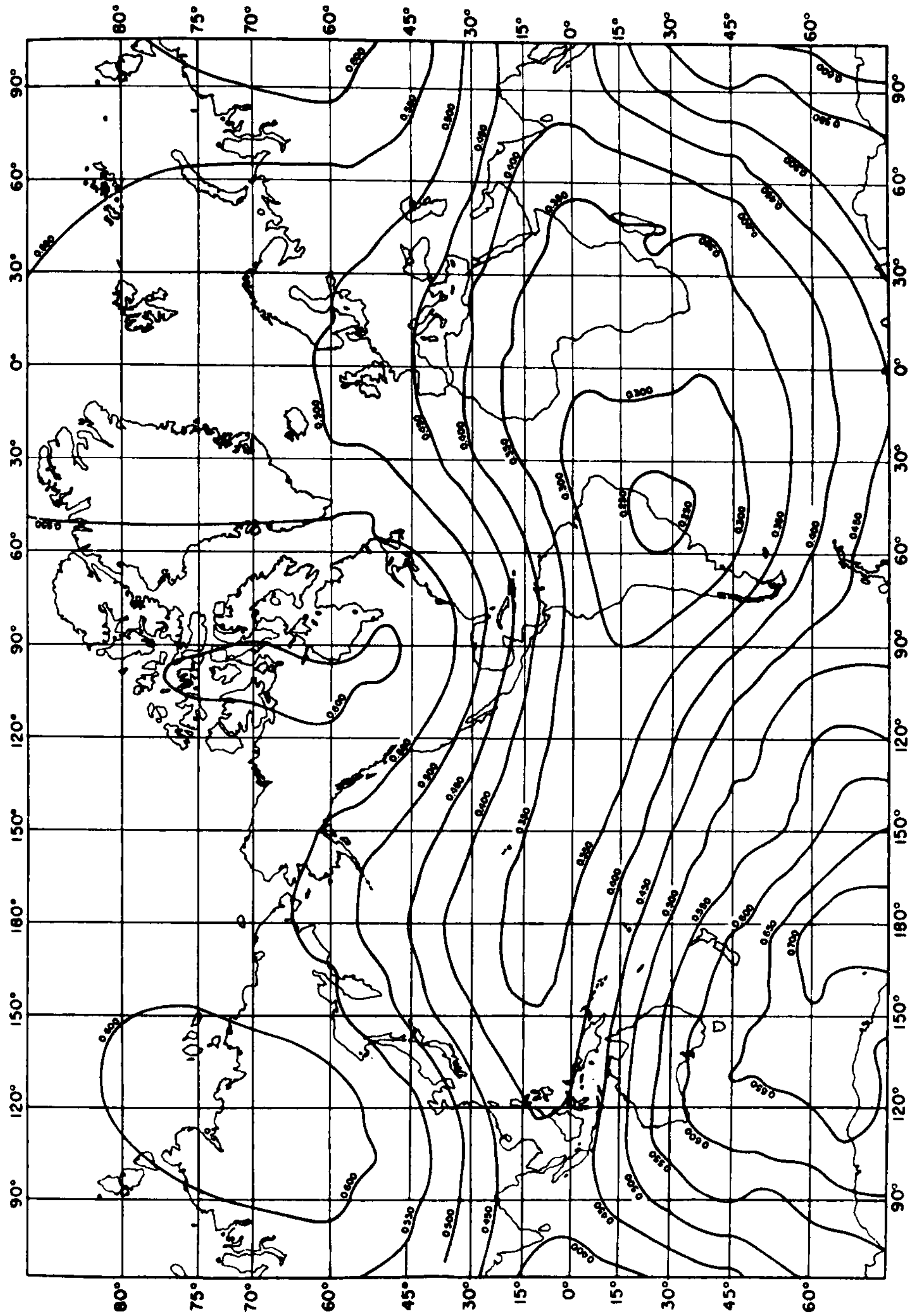

FIG. 2. The total intensity of the Earth's magnetic field (in gauss). United States Air Force Geophysics Research Directorate (1960). Reproduced by kind permission of Macmillan Co.

of swings of a dip-needle in the meridional plane for ten-minute periods and in four northern hemisphere zones and one southern hemisphere zone, von Humboldt was able to discover the increase in the total magnetic intensity from the equator to the poles. Three years later he extended his measurements to horizontal intensity with the aid of a compass needle oscillating in the horizontal plane. These measurements were made before the current scientific unit of field intensity had been introduced, and hence results were expressed in relation to a standard station at Micuipampa in Peru on the magnetic equator where the intensity was arbitrarily taken as unity. Later Gauss was to estimate von Humboldt's unit as 0·3494Γ.

The Earth's present field may be described completely in terms of three components:

(i) A dipole component which is by far the most important part of the field and which is of internal origin.

(ii) A smaller non-dipole component which is also of internal origin.

(iii) A still smaller component originating outside the Earth due to electric currents in the ionosphere, and amounting at most to a few percent of the total field. Despite their proportionately small magnitudes these external fields possess periods of variation ranging from 11 years to a matter of hours due to solar and lunar influences, so that direct field measurements are hampered by irregularities whose amplitude is of the order of 40γ in low and middle latitudes (Vestine *et al.* 1947). From the palaeomagnetic point of view, however, the external fields are not significant and will not be considered further.

It is useful to express the geomagnetic field in terms of spherical harmonics for, by so doing, each component may be shown to possess a different mathematical form. This was first carried out by Gauss (1838) who was thus able to show that both the dipole and non-dipole components were of internal origin. Owing to the imperfect magnetic data available at the time, Gauss inferred that no external source of magnetic field existed; but more sophisiticated measurements have since shown that a small external component is present.

The basis of the analysis is to express the Earth's magnetic potential as a series of the form:

$$V = a \sum_{n=0}^{\infty} \sum_{m=0}^{n} \left(\frac{a}{r}\right)^{n+1} [g_n^m \cos m\phi + h_n^m \sin m\phi]\, P_n^m(\cos\theta)$$

where a is the radius of the Earth,

r is the radial distance of the field point,

θ is the geographic colatitude,

ϕ is the east longitude,

$P_n^m(\cos\theta)$ are associated Legendre polynomials of degree n and order m,

g_n^m and h_n^m are constants (now known as Gauss coefficients) having the units of magnetic intensity.

The field components (X, Y, Z) are then given by the derivatives:

$$X = \frac{1}{r}\frac{\partial V}{\partial \theta}, \quad Y = \frac{-1}{r \sin\theta}\frac{\partial V}{\partial \phi}, \quad Z = \frac{\partial V}{\partial r},$$

whence at $r = a$, X, Y and Z may be expressed solely in terms of g_n^m, h_n^m, $\sin m\phi$, $\cos m\phi$ and $P_n^m(\cos\theta)$. The first and most important coefficient, g_1^0, is that associated with the axial dipole field. It is, in fact, the equatorial intensity of the axial dipole field. The smaller coefficients g_1^1 and h_1^1 are associated with the equatorial components

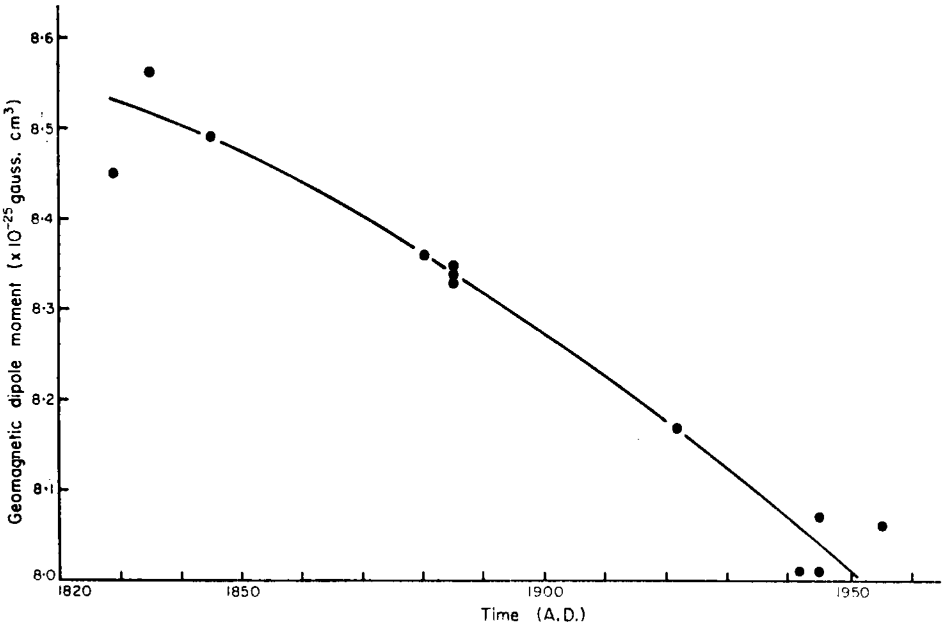

FIG. 3. Variation of the geomagnetic dipole moment with time since direct direct field observations were first made.

of the dipole field. It may be shown that the total geomagnetic dipole moment, M, is given by:

$$M = a^3[(g_1{}^0)^2+(g_1{}^1)^2+(h_1{}^1)^2]^{\frac{1}{2}}.$$

The values of M computed from Gaussian coefficients determined by various authors are shown in Table 1 and are plotted against time in Fig. 3. Despite the scatter it is clear that the dipole moment has been decreasing since direct field measurements were first made. The figures indicate a rate of decrease of about 5% per century since 1835.

Table 1

Geomagnetic dipole moment computed from the data of various authors for different epochs

Author	Epoch	Moment $\times 10^{-25}$ gauss . cm^3
Erman & Petersen	1829	8·45
Gauss	1835	8·56
Adams	1845	8·49
Adams	1880	8·36
Schmidt	1885	8·35
Fritsche	1885	8·34
Neumeyer & Petersen	1885	8·33
Dyson & Furner	1922	8·17
Jones & Melotte	1942	8·01
Afanasieva	1945	8·01
Vestine *et al.*	1945	8·07
Finch & Leaton	1955	8·06

One of the first questions that palaeointensity determination might be expected to answer is therefore: for how long has this decrease been going on and at what rate? This problem has been solved for several centuries preceding direct field measurements; and this will be discussed in Section 8.1.

5. Treatment of data

It is clear from spherical harmonic analysis that the dipole term is by far the predominant one in the Earth's magnetic field; and one consequence of this is that to a first approximation the Earth's field is identical with that of a uniformly magnetized sphere. This fact was first recognized by William Gilbert in his book *De Magnete* (1600). Thus the basic formula relating magnetic inclination, I, to latitude, λ, is $\tan I = 2\tan\lambda$. In palaeomagnetism, I is usually measured and the ancient latitude of the site is calculated using this formula. Thus strictly speaking the latitude obtained is the magnetic latitude. In practice, however, it is commonly assumed that, averaged over long periods ($> 10^4$ years), the Earth's dipole field has been axial, so that the calculated latitude is taken as the geographic latitude and the effects of non-dipole fields or wobble of the main dipole are treated as 'noise' superimposed on the main axial dipole 'signal'. Since this 'noise' is random it may be averaged out by using a large number of specimens representing a time interval of over 10^4 years.

The validity of the axial dipole assumption has been widely discussed previously (Runcorn 1959, Creer 1962, Irving 1964 and others) and so the evidence will not be reviewed here. It is worth mentioning, however, that the best-fitting dipole is at present inclined at about $11\frac{1}{2}°$ to the Earth's rotational axis. The values which have been published since 1829 for the position of the north geomagnetic pole are shown in Table 2. Although the pole appears to have moved by about 5° of longitude towards the west, Bullard *et al.* (1950) consider, on the basis of possible inaccuracy in the early information, that during the period covered by direct observation it cannot be considered to have moved. However, despite the lack of observable movement the mean axial dipole is used as a working hypothesis in palaeomagnetism.

Table 2

Positions of the north geomagnetic pole computed by various authors for different epochs

Author	Epoch	Latitude °N	Longitude °W
Erman & Petersen	1829	78·3	64·6
Gauss	1835	77·7	63·6
Adams	1845	78·7	64·3
Adams	1880	78·3	68·1
Schmidt	1885	78·6	69·6
Fritsche	1885	78·6	67·9
Neumeyer & Petersen	1885	78·3	67·6
Dyson & Furner	1922	78·4	69·1
Jones & Melotte	1942	78·8	68·6
Afanasieva	1945	78·4	68·8
Vestine *et al.*	1945	78·6	69·9
Finch & Leaton	1955	78·3	69·0

Field intensity determinations have been presented in several different ways. Thellier *et al.* (1959 review) (who were first to publish historic intensity determinations) were able to compare their values directly by correcting for the differences in latitude of the sites from which the samples came. They used the palaeomagnetically determined inclination to reduce the intensity value of any given specimen to that it would have had had it been magnetized at the 65° isocline. 65° was chosen as being typical of the sampling region as a whole. For a dipole field of moment P, it can be shown that at radial distance r and latitude λ, or inclination I, the field intensity is given by:

$$F = \frac{P}{r^3}(4-3\cos^2\lambda)^{\frac{1}{2}} = \frac{2P}{r^3}(1+3\cos^2 I)^{\frac{1}{2}}. \tag{3}$$

Thus assuming the Earth's field to be essentially dipolar, the Thelliers used the following reduction formula derived from (3):

$$F_{65} = F\,\frac{(1+3\cos^2 I_m)^{\frac{1}{2}}}{(1+3\cos^2 65°)^{\frac{1}{2}}}, \tag{4}$$

where F is the determined ancient field intensity, I_m is the measured palaeomagnetic inclination and F_{65} is the corresponding intensity at the 65° isocline. Since this equation assumes the field to be essentially dipolar, the non-dipole variations will manifest themselves as scatter or 'noise' superimposed on the main moment.

Doell *et al.* (1961) have suggested an alternative method of treatment by which all intensity values are reduced to the same geographic position. They have recalculated the Thelliers' data for the reference position 50° N, 5° E according to the conversion formula:

$$F^1 = F(1+3\cos^2 p_0)^{\frac{1}{2}}(1+3\cos^2 p)^{-\frac{1}{2}}, \tag{5}$$

where p is the angular distance from the sampling site to the magnetic pole, p_0 is the distance from the reference position to the magnetic pole, and F^1 is the reduced intensity. Again the field is assumed to be dipolar, and further, the geomagnetic pole is assumed to be fixed in its present position. Thus variations in F^1 are only proportional to changes in the main dipole if this latter assumption is correct. By this method also, non-dipole variations will appear as superimposed scatter.

Yet a third method of presentation has been used by many workers (Burlatskaya 1961, 1962, Nagata *et al.* 1963, Nagata *et al.* 1964, Sasajima *et al.* 1964), for historic intensity values. In these cases F/F_0 ratios have been given, where F is the measured ancient field intensity and F_0 is the present intensity at the sampling site. Similarly, geological field intensity values have been presented by Nagata (1943), Van Zijl (1961) and Momose (1963) as J_n/J_T ratios, where J_n is the natural moment and J_T is an artificial TRM acquired in the Earth's present field.

It would be convenient to have some method of treating intensity results which would enable values from sampling sites at different latitudes to be compared directly, with the minimum of assumptions regarding the past behaviour of the field. It was suggested by Smith (1967), therefore, that it would be preferable to calculate the equivalent dipole moment which would have produced the measured intensity at the magnetic palaeolatitude of the sample. This is analogous to the calculation of virtual geomagnetic pole positions (VGP) from palaeomagnetic directional data; and hence the calculated dipole moment may be called the *virtual dipole moment* (VDM).

One advantage of this method is that no scatter is introduced by wobble of the main dipole since the palaeolatitude determined is independent of the orientation of the dipole relative to the Earth's rotational axis. This is also true of the reduction method used by the Thelliers; but they had recourse to a somewhat arbitrarily chosen isocline of 65° which may not be convenient for all rocks examined in the future. Calculation of the VDM is equally appropriate to historic or geological determinations as long as the palaeomagnetic latitude is known. Apart from experimental errors, the non-dipole field will also produce scatter.

In all examples of geological intensity values published so far, and for the Thelliers' historic values, the associated magnetic palaeolatitudes have also been presented. It is unfortunate that in all cases where historic F/F_0 ratios have been determined no

Table 3

Historic and archaeological reduced dipole moments.
N = number of samples used

No.	Reference	Site	Rock type	Age	RDM $\times 10^{-25}$ gauss . cm^3	N
1	B	Tbilisi, Russia	Brick	1900	8·6[(1)]	3–4
2	SM	Kagoshima, Japan	Andesite	1780±5	11·2	1
3	NAM	Japan	Basalt	1778	11·3	5
4	TT	Versailles, France	Brick	1750	>7·4	?
5	B	Tbilisi, Russia	Brick	1725	8·4*	6–8
6	B	Tbilisi, Russia	Brick	1625	9·2*	6–8
7	NAM	Japan	Basalt	1552	11·3	4
8	SM	Kagoshima, Japan	Andesite	1470±5	12·5	1
9	TT	Lille, France	Brick	1460	8·5	7
10	NAM	Japan	Basalt	1421	12·5	6
11	NKS	Peru	Pottery	1400	10·6	2
12	NAM	Japan	Basalt	1300±30	10·3	4
13	SM	Toyota, Japan	Baked earth	1250±80	12·3	4
14	B	Tbilisi, Russia	Brick	1250	10·9*	9–12
15	SM	Seto, Japan	Baked earth	1120±20	12·8	4
16	B	Tbilisi, Russia	Brick	1100	11·1	3–4
17	NAM	Japan	Basalt	1070±70	9·6	4
18	B	Tbilisi, Russia	Brick	1050	11·6*	6–8
19	SM	Seto, Japan	Baked earth	1000±20	14·5	4
20	SM	Sakai/Seto, Japan	Baked earth	935	14·0*	2
21	B	Tbilisi, Russia	Brick	900	11·8*	6–8
22	B	Tbilisi, Russia	Brick	850	12·5	3–4
23	NAM	Japan	Tile	750	10·1*	7
24	B	Tbilisi, Russia	Brick	750	12·5	3–4
25	SM	Sakai/Seto, Japan	Baked earth, tile	730	13·4*	9
26	SM	Sakai, Japan	Baked earth	650±50	14·9	2
27	NKS	Bolivia	Pottery	600±100	15·1	3
28	SM	Sakai, Japan	Baked earth	545	14·7*	5
29	NAM	Japan	Pottery	545	12·4*	11
30	NKS	Bolivia; Peru	Pottery	530	14·2	3
31	SM	Tsu, Japan	Baked earth	480±20	14·5*	3
32	NKS	Bolivia	Pottery	400±100	12·5	1
33	SM	Ueno, Japan	Clay idol	380±20	13·6	2
34	NAM	Japan	Basalt	300±100	14·4	5
35	SM	Hiraoka, Japan	Pottery	220±20	14·3	2
36	TT	Paris, France	Brick	200	10·7	12
37	TT	Basle, Switzerland	Tile	175	11·4	5
38	B	Tbilisi, Russia	Brick	150	12·3*	6–8
39	SM	Fukuoka, Japan	Pottery	100±50	14·3	1
40	TT	Fréjus, France	Brick	25	10·6	11
41	NKS	Mexico; Peru	Pottery, rock [(2)]	0	10·5	11
42	SM	Daito, Japan	Pottery	−100±50	16·4	1
43	TT	Carthage	Kiln	−146	12·6	6
44	NAM	Japan	Basalt	−150±150	14·5	2
45	NAM	Japan	Pottery	−300±250	11·5	4
46	TT	Carthage	?	−600	13·5	1
47	SM	Shiga, Japan	Pottery	−900±100	9·3	2
48	NAM	Japan	Pottery	−1000±400	9·4	1
49	NAM	Japan	Pottery	−2000±500	8·9	1
50	NAM	Japan	Pottery	−2500±300	7·9	3
51	B	Tbilisi, Russia	Brick (?)	−2600	9·9	3–4
52	NAM	Japan	Pottery	−3100±300	7·4	1
53	JMT	South Newbury, Vermont	Glacial clay	*ca.* −8000	~12·0	?[(3)]

[(1)] All B values scaled from Fig. 3 of Burlatskaya (1961).

[(2)] Lavas and pyroclastic rocks.

[(3)] Many samples covering a depositional period of 200 years.

* Century-means (Section 6).

directional data have been given, so that in order to calculate dipole moments a further assumption must be made. Although the best-fitting dipole has not been observed to move, there is no guarantee that it has not done so over the past few thousand years, and so some assumption has to be made about its orientation at the time that any particular rock was laid down. Since it probably takes at least 10000 years for the dipole to mean out axially (Cox *et al.* 1964) the axial dipole cannot be regarded as valid in this case. The only course, therefore, is to assume that the geomagnetic pole has remained fixed in its present position and hence to calculate dipole moments using present geomagnetic latitudes. Dipoles calculated in this way have been termed *reduced dipole moments* (RDMs). Some error will be introduced if the geomagnetic dipole has moved; but this is not likely to be much greater than that introduced by a latitude change of about $11\frac{1}{2}°$ since this represents the standard deviation due to dipole wobble (Cox *et al.* 1964).

In this review all ancient field intensity data will be converted to VDMs or RDMs, as appropriate, irrespective of the source. Results from all workers have been converted to this common system; and details of the calculations are given where appropriate.

6. Historic and archaeomagnetic field intensity data

In Table 3 reduced dipole moments covering historic and archaeological time are listed in order of increasing age. The list includes all results known to the author at 1966 June. Details of the sampling locality and the type of rock used are given together with the number (N) of specimens used in the determination of the RDMs. Dates AD are expressed as positive numbers and dates BC as negative numbers.

Where an author has made several determinations on material from the same century, the resultant moments and ages have been meaned. In certain cases this has already been done by the author concerned; but where this is not so, the present author has made the relevant calculation for the sake of uniformity. Such century-means are denoted by an asterisk. All errors are those quoted by the original authors except where the present author has carried out the averaging operation described above.

For clarity, code letters have been used under the heading 'References' in Table 3 and denote the relevant sources as follows:

JMT	Johnson, Murphy & Torreson 1948.
TT	Thellier & Thellier 1959 (see Note 1).
B	Burlatskaya 1961, 1962.
NAM	Nagata, Arai & Momose 1963.
NKS	Nagata, Kobayashi & Schwarz 1964 (see Note 2).
SM	Sasajima & Maenaka 1964 (see Note 3).

Note 1. This is a review of all field intensity work carried out by the Thelliers up to 1959.

Note 2. Also published by Nagata, Kobayashi & Schwarz 1965.

Note 3. Also published by Sasajima 1965.

A summary of methods and reliability tests is given in Table 4.

In all cases reduced dipole moments have been calculated from F/F_0 ratios on the assumption that the geomagnetic dipole has remained fixed in its present position. Thus the geomagnetic latitude of the sampling site (or the laboratory, where appropriate) was used in equation (3) to calulate P. The geomagnetic latitudes used are shown in Table 5, which also gives the values of the present field F_0 at the sites (or laboratories). F_0 values were usually quoted by the original authors.

Table 4

Summary of methods and reliability tests for historic and archaeological field intensity determinations

'Internal consistency' refers to the constancy of field intensity values obtained by the Thelliers' method as described in Section 3

1. **Johnson, Murphy & Torreson (1948)**
   Method: Artificial deposition under water in the laboratory (DRM).
   Reliability: Stability inferred from uniformity of directional measurements and high coercive forces.

2. **Thellier & Thellier (1959)**
   Method: Thelliers' (Section 3) (TRM).
   Reliability: Internal consistency.

3. **Burlatskaya (1961, 1962)**
   Method: Thelliers' (TRM).
   Reliability:
   (i) Internal consistency.
   (ii) Stability of directions during heating.
   (iii) Experiments carried out on bricks magnetized during the period of direct field measurement gave intensity values correct to $<5\%$.

4. **Nagata, Arai & Momose (1963)**
   Method: Thelliers' (TRM).
   Reliability:
   (i) Internal consistency.
   (ii) Lack of linearity of $\Delta M_n - \Delta M_a$ curves at high temperature interpreted as chemical change or initial baking below the Curie point and such specimens discarded.
   (iii) Chemical and X-ray analyses carried out on certain specimens.

5. **Nagata, Kobayashi & Schwarz (1964)**
   Method: Thelliers' (TRM).
   Reliability:
   (i) Internal consistency.
   (ii) Temperatures above 200°C used to avoid viscous magnetizations.
   (iii) Basalts, pottery and pyroclastic rocks of same age gave consistent values even though from widely separated sites.
   (iv) No Curie point changes.
   (v) J_s change $<\pm 10\%$ for pottery and $<\pm 12\%$ for rocks (with one exception at $+58\%$).

6. **Sasajima & Maenaka (1964)**
   Method: Thelliers' (TRM).
   Reliability:
   (i) Internal consistency.
   (ii) High stability under thermal and a.c. demagnetization.

Table 5

Data used for the conversion of F/F_0 ratios and F values to RDMs

$\lambda_m{}^0$ = *mean geomagnetic latitude of site.*

F_0 = *present field at sampling or laboratory site*

* = *read from Vestine et al.* (1947)

Site	Author	$\lambda_m{}^0$	F_0 (oe)
Western Japan	SM	22°N	0·465
Tokyo, Japan	NAM	26°N	0·461
Tbilisi, Russia	B	37°N	0·480*
Mexico	NKS	28°N	0·448
Peru	NKS	1°N	0·291
Bolivia	NKS	5°S	0·291
Versailles	TT	52°N	—
Lille	TT	53°N	—
Paris	TT	52°N	—
Basle	TT	49°N	—
Fréjus	TT	45°N	—
Carthage	TT	38°N	—

7. Geological field intensity data

For igneous rocks the natural remanent magnetization (J_n) is usually greater than the magnetization induced isothermally in the Earth's present field (H_0), so that $J_n > \kappa H_0$, where κ is the susceptibility. Koenigsberger (1938) expressed J_n, κ and H_0 in terms of the Q_n-ratio, defined as:

$$Q_n = \frac{J_n}{\kappa H_0}.$$

Similarly Q_T may be defined in terms of the TRM (J_T) acquired in field H_0; and combining the two Q-ratios, we may write:

$$Q_{n/T} = \frac{Q_n}{Q_T} = \frac{J_n}{J_T}.$$

Koenigsberger was also among the first to consider field intensity in any systematic way. His well-known diagram of $Q_{n/T}$ values plotted against time is shown in Fig. 4. The twenty-five points here represent basalts, andesites, granites, diabases, a granodiorite, a porphyrite and an amphribolapite from various parts of the world, but mainly Germany. Koenigsberger considered the low $Q_{n/T}$ ratios for the older rocks to be due to (i) movement of the lavas during solidification and at temperatures below the Curie point of pure magnetite (585° C), and (ii) spontaneous decrease of the NRM over a period of time due mainly to seismic shocks, chemical changes and

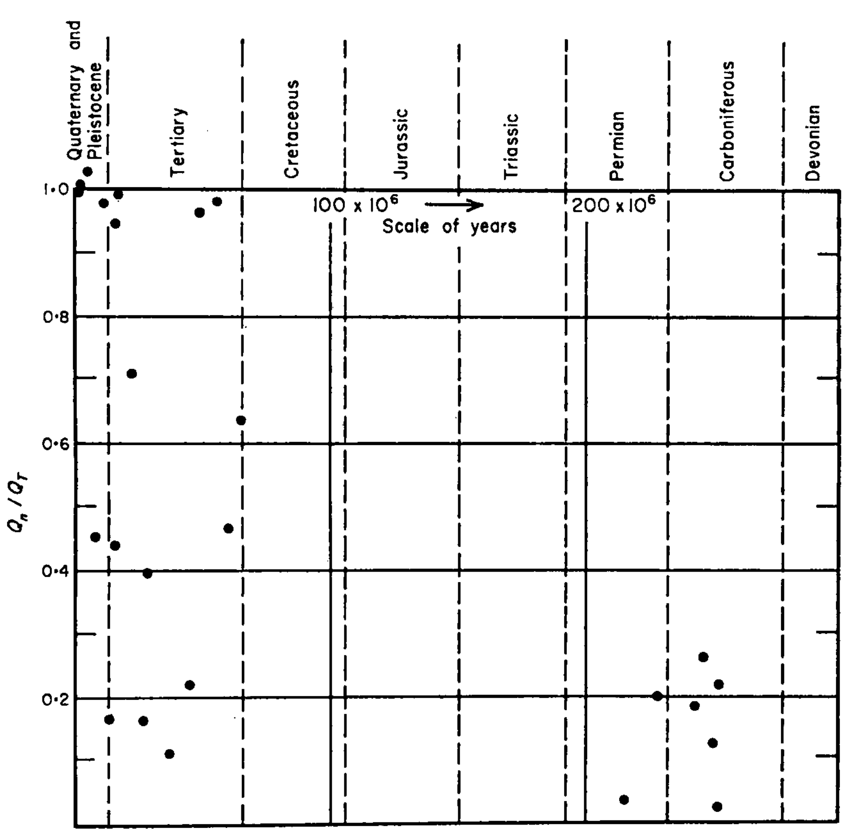

FIG 4. Variation of Q_n/Q_T with geological age for igneous rocks, mainly from Germany (after Koenigsberger 1938).

the effects of changes in the magnetic field direction (including reversals). Movements of the material during the acquisition of NRM will certainly reduce the resultant intensity of magnetization, but it is doubtful whether such movements can have taken place below 600° C. The question of a time-dependent decrease of magnetization is an important one for all field intensity results. The extent of secondary magnetization in Koenigsberger's rocks is not known. Further, Koenigsberger made no allowance for change of latitude between the acquisition of the NRM and the artificial TRM due to continental drift; and hence his Permo-carboniferous $Q_{n/T}$ will be under-estimated by a factor of about two. The Tertiary values will only be affected by a few percent on account of continental drift. It will be shown later that virtual dipole moments calculated approximately from Koenigsberger's data are in remarkable agreement with more modern data.

Irving (1964) has pointed out that the ratio $Q_n{}^1 = J_n/\kappa$ depends more strongly on the ancient field than the NRM alone does because the use of κ will, to a certain extent, correct for the quantity and type of magnetic material present. He suggests, therefore, that for a sequence of rocks magnetized by a common process, the determination of $Q_n{}^1$ would give a relative measure of the Earth's field over the period covered by the rock sequence assuming no viscous decay. On the other hand, $Q_n{}^1$ ratios have been used to indicate the stability of rocks with regard to such decay (Powell 1963, Irving 1964). Nagata (1943), assuming the Earth's field intensity to have remained constant, also used the $Q_{n/T}$ ratio as a measure of stability. He showed, for example, that whereas for ten recent Japanese lavas $Q_{n/T} = (1{\cdot}02 \pm 0{\cdot}04)$, for eleven Japanese Pleistocene ejecta $Q_{n/T} = (0{\cdot}68 \pm 0{\cdot}05)$. Nagata *et al.* (1954) suggested the rejection of rock data in cases where $J_n/J_T < 0{\cdot}1$ on the ground of possible instability, but at the same time admitted that such low values do not necessarily imply viscous decay. A perfectly stable, but low, NRM could arise, for example, during a reversal of the Earth's field if such a reversal occurred by the decay of the moment followed by a build-up in the opposite sense. On the other hand, Akimoto *et al.* (1960) showed that for two Japanese Tertiary dolerite sheets, low values of J_n/J_T lying between 0·023 and 0·099 were due to oxidation of the original titanomagnetite content to titanomaghemite, thereby obliterating the NRM with an IRM component. It is clearly not possible to generalize on the causes of low $Q_{n/T}$ values. Particular cases must be treated on their merits.

Thellier *et al.* (1959) criticized the use of the $Q_{n/T}$ ratio exclusively as a test of stability by pointing out that low ratios could indicate a lower value of the Earth's field intensity in the past. Momose (1963) has argued similarly; and all workers presenting ancient field intensity determinations necessarily make this assumption. Thellier *et al.* have obtained one precise geological field intensity value for the Middle Quaternary using metamorphic clays baked by the lavas of the Gravenoire Plateau in France. Four specimens gave consistent values of field intensity at 0·18 oersted, which at the measured magnetic palaeolatitude corresponds to a virtual dipole moment of $3{\cdot}13 \times 10^{25}$ gauss.cm^3. An attempt was also made to obtain a field intensity value from an andesite extruded by the Nugere volcano in France (Upper Quaternary), but values from six specimens ranged from 0·20 oersted to 0·46 oersted which the Thelliers considered to be an indication of unreliability.

Using the one-interval TRM technique (Section 3), Wilson (1961) obtained field intensity values of $(0{\cdot}42 \pm 0{\cdot}03)$ oersted and $(0{\cdot}42 \pm 0{\cdot}01)$ oersted from two specimens of laterite baked by a lava in Northern Ireland during the Tertiary. The quoted values represent means of several one-interval determinations at various temperatures between 400° C and 600° C, that is, above the effects of secondary magnetizations. The corresponding mean virtual dipole moment is $7{\cdot}00 \times 10^{25}$ gauss.cm^3. Weaving (1960) was able to deduce two spot intensity values by utilizing a particular property of lavas from the Linga area of Central India (Deccan Traps). Two specimens each possessed two magnetic components having widely separated Curie points.

Since in each case the low Curie point component must have originally acquired its magnetization in the resultant field of the Earth and the high Curie point component, by heating and cooling the low Curie point component in various fields Weaving was able to estimate the initial ambient cooling fields to be 0·37 oersted and 0·40 oersted. The correponding mean virtual dipole moment is $7{\cdot}55 \times 10^{25}$ gauss.cm^3.

DRM has also been utilized in field intensity determination. In 1954 Clegg *et al.* obtained a rough value of the Triassic field intensity using Cheshire sandstones by redepositing the ground specimens under water. These experiments gave $J_n/J_d \sim \frac{1}{3}$ whence, taking into account latitude changes, the VDM is $\sim 3{\cdot}2 \times 10^{25}$ gauss.cm^3. In 1955, King, in similar experiments obtained field intensity values leading to a VDM of $\sim 36 \times 10^{25}$ gauss.cm^3 for a silt of about 1000 years old from Angermanland in Sweden, and to a VDM of $\sim 24 \times 10^{25}$ gauss.cm^3 for a somewhat older specimen. The latter two values are inexplicably at least a factor of two higher than any others reported.

Most of the above values, however great the experimental accuracy and however reliably they are thought to reflect the past field, are of only limited usefulness because of the small number of samples in each set. As in the case of directional measurements, a large number of specimens spread over a period of at least 10^4 years are required in order to mean out the effects of secular variation. Of the above results only those of Nagata (1943), Koenigsberger (1938) and Clegg *et al.* (1954) satisfy this criterion, and these results may only be regarded as approximate. There are, however, four further sets of results which obey the criterion, namely, Van Zijl (1961), Momose (1963), Briden (1966) and Smith (1967). For analysis later in this paper field intensity determinations on *sets* of rocks have been grouped as follows:

Class 1. Accurate determinations; stability tests carried out; at least ten specimens measured.

Van Zijl (1961) — Briden (1966)
Momose (1963) — Smith (1967).

Class 2. Approximate determinations; method of doubtful validity or no stability tests carried out or fewer than ten specimens measured or any combination of these.

Koenigsberger (1938) — Clegg *et al.* (1954)
Nagata (1943) — Irving *et al.* (1965).
Johnson *et al.* (1948)

Details of Class 1 virtual dipole moments are given in Table 6 and details of Class 2 virtual dipole moments are given in Table 7. The ages stated are those according to Holmes' *Revised Time Scale* of 1959 unless it is otherwise stated that radioactive age determinations have been carried out. Details of the VDM calculations are given in the relevant Table.

When further accurate (Class 1) data becomes available, approximate (Class 2) data may perhaps be discarded, but at present the approximate VDMs lend useful support to the more accurate VDMs with which they agree very well.

8. Discussion

8.1. *Ancient geomagnetic field intensities: historic and archaeological time.* The division of intensity determinations into those for historic and archaeological time (defined here as $< 10^4$ years before present) and those for geological time is in some ways an arbitrary distinction, but a useful one, for by concentrating on a fairly well-dated period of a few thousand years it may be possible to resolve short period (10^3 years or less) variations in field intensity. On the geological time scale such short period variations are usually meaned out as in palaeodirectional studies.

Table 6

Class 1 (*accurate*) *geological virtual dipole moments in order of increasing age.*

VDMs have been given in the form: VDM, standard deviation (%), *standard error of mean* (%).

Units of VDM are 10^{25} *gauss.cm*^{3}. *Approximate ages obtained from Holmes*' (1959) *Revised Time Scale unless radioactive ages are available.* F=*field intensity.* J_s=*saturation magnetization.* N=*normal polarity,* R=*reversed polarity,* I=*intermediate polarity*

1. **Smith (1967)**

Age: 2·1–18·1 m.y. (K–A dating).
Rocks: Baked laterites and Class 5 lavas from Iceland.
Number and polarity: 14*N*, 15*R*, 2*I*.
Type of NRM: TRM.
Method: One-interval heating and cooling in 0·60 oersted field (laterites); a.c. demagnetization of NRM and artificial TRM (lavas).
Reliability:
(i) No high-field (5000 oersteds) Curie point change on heating.
(ii) No zero-field Curie point change on heating.
(iii) NRM and TRM curves possess same shape.
(iv) Heating and cooling saturation magnetization curves possess same shape.
(v) Mean J_s change$=-(10{\cdot}7\pm4{\cdot}5)$% for laterites and $-(6{\cdot}4\pm4{\cdot}3)$% for lavas.
(vi) F calculated above the effects of secondary components.
(vii) Mean standard deviation in F at 270°C, 320°C and 370°C$=\pm5{\cdot}8$% (laterites), at 100, 150, 200, 250 and 300 oersted fields$=\pm4{\cdot}5$% (lavas).
(viii) Mean standard deviation in F for pairs of specimens from same contact $=\pm10{\cdot}6$% (laterites).
VDM: Calculated using mean palaeomagnetic latitudes (50°N for laterites, 47°N for lavas).
VDM=5·34, 2·52 (47·2%), 0·45 (8·4%).
Notes (i) Mean site latitude=65°N.

2. **Momose (1963)**

Age: Upper Miocene–Upper Pliocene (~11 m.y.).
Rocks: Welded tuffs, andesitic tuff, basalts, andesites, porphyrite, dacite and biparite from (mainly) Komoro, Shigarami and Enrei formations of Japan.
Number and polarity: 14*N*, 23*R*.
Type of NRM: TRM.
Method: One-interval heating and cooling in Earth's present field.
Reliability:
(i) No Curie point change on heating even after one hour at Curie point.
(ii) NRM and TRM curves possess same shape.
(iii) J_s–T curves repeatable; 75% specimens gave J_s (heating)/J_s (cooling) ~1, 25% specimens gave J_s change <5%.
(iv) Directions unchanged after storage for over one year in Earth's field.
(v) Stable under a.c. demagnetization.
VDM:
(a) Calculated using palaeomagnetic latitudes for individual specimens.
Normal: VDM=5·26, 3·13 (59·5%), 0·84 (16·0%).
Reversed: VDM=6·07, 2·30 (37·9%), 0·48 (7·9%).
All: VDM=5·77, 2·68 (46·4%), 0·44 (7·6%).
(b) Calculated using mean palaeomagnetic latitudes (computed by reviewer).
Normal: VDM=5·29, 2·38 (45·0%), 0·67 (12·7%). λ_m=43·5°N.
Reversed: VDM=6·12, 2·43 (39·7%), 0·48 (7·8%). λ_m=21·0°N.
All: VDM=5·95, 2·52 (42·3%), 0·45 (7·6%). λ_m=28·5°N.
Notes:
(i) Mean site latitude=36°N.
(ii) There is no significant difference between VDMs calculated using mean latitudes and VDMs calculated using individual latitudes.
(iii) *N* and *R* sequences separated by a sequence of twelve transition zone specimens (not included in above analysis).

Table 6—*continued*

3. **Smith (1967)**
   Age: 33·9–57·1 m.y. (K–A dating).
   Rocks: Baked contacts (red sandstones, conglomerates, dykes) and dykes from Scotland.
   Number and polarity: 1*N*, 17*R*.
   Type of NRM: TRM.
   Method: One-interval heating and cooling in 0·60 oersted field.
   Reliability:
   (i) No high-field (5000 oersteds) Curie point change on heating.
   (ii) No zero-field Curie point change on heating.
   (iii) NRM and TRM curves possess same shape.
   (iv) Heating and cooling J_s curves possess same shape.
   (v) Mean J_s change $= -(13{\cdot}0 \pm 9{\cdot}0)\%$.
   (vi) F calculated above the effects of secondary cononents.
   (vii) Mean standard deviation in F at 270°C, 320°C and 370°C $= \pm 10{\cdot}2\%$.
   (viii) Mean standard deviation in F for pairs of specimens from same contact $= \pm 10{\cdot}6\%$.
   (ix) F values from one dyke and corresponding baked contact agreed within $\pm 7{\cdot}5\%$.
VDM: Calculated using mean palaeomagnetic latitude (40°N).
   VDM=4·94, 2·58 (52·3%), 0·61 (12·4%).
   Note: Mean site latitude=56°N.

4. **Van Zijl (1961)** (also Van Zijl, Graham & Hales 1962)
   Age: Triassic–Jurassic (~180 m.y.).
   Rocks: Basalts from Stormberg (Maseru and Sani Pass) area of Basutoland.
   Number and polarity: 52*N*, 16*R*.
   Type of NRM: TRM.
   Method: A.c. demagnetization at 219 oersteds (peak) of NRM and artificial TRM.
   Reliability:
   (i) Heatings carried out in nitrogen in attempt to avoid chemical changes, and specimens left for one hour at 650°C.
   (ii) Close correspondence of NRM and TRM curves.
   (iii) No large directional changes during heating.
   (iv) Specimens with VRM or lightning NRM rejected.
   (v) Ore microscopy and X-ray analysis carried out on specimens before and after heating.
   VDM: Calculated using mean palaeomagnetic latitudes (33°S for normal specimens, 41°S for reversed specimens).
   Normal: VDM=3·24, 2·00 (61·6%), 0·28 (8·6%).
   Reversed: VDM=2·77, 1·36 (49·1%), 0·34 (12·3%).
   All: VDM=3·12, 1·89 (60·5%), 0·23 (7·4%).
   Notes:
   (i) F data scaled from Van Zijl (1961), Fig. 5 . 1.
   (ii) Mean site latitude=29·6°S (Sani Pass), 29·4°S (Maseru).
   (iii) Normal and reversed sequences are separated by thirty-eight transition zone specimens (not included in above analysis).
   (iv) The Karroo dolerites of South Africa (which according to Van Zijl were the feeders for the Stormberg lavas) have been K–A dated in the range 154–190 m.y. (McDougall 1963).

5. **Briden (1966)**
   Age: Late Silurian or early Devonian (~400 m.y.).
   Rocks: Mugga Mugga Porphyry instrusion near Canberra, Australia.
   Number and polarity: Twenty-nine specimens from four samples (N).
   Type of NRM: TRM.
   Method: The Thelliers' method with detail differences.
   Reliability:
   (i) A.c. and thermal demagnetization curves similar.
   (ii) Specimens heated in nitrogen atmosphere.
   (iii) F values internally consistent.
   (iv) Susceptibility did not change on heating.
   (v) Directions diverge consistently from present field direction.
   (vi) No systematic difference in F values from discs and cylinders cut from same sample, so superficial alteration discounted.
   VDM: Calculated using mean palaeomagnetic latitude (23°S) and by giving unit weight to each specimen.
   VDM=2·25, 0·97 (43·1%), 0·18 (8·0%).
   Note: Mean site latitude=35°S.

Table 7

Class 2 (approximate) virtual dipole moments in order of increasing age. Explanation of symbols as for Table 6

1. **Johnson, Murphy & Torreson (1948)**

Age:	Recent (~0·6 m.y.).
Rocks:	Pacific ocean bottom sediments (red clay).
Number and polarity:	Eight specimens from single vertical core (*N*).
Type of NRM:	DRM.
Method:	See Section 3 . 2.
VDM:	Determined relative to Earth's present field by comparison with specimen of present age. VDM=5·28, 0·71 (13·4%), 0·25 (4·7%).
Notes:	(i) Specimens spread at roughly equal time intervals from 35 000 to 1 165 000 years. (ii) Mean site latitude=32·4°S. (iii) Palaeomagnetic latitude=8·2°S.

2. **Nagata (1943)**

Age:	Pleistocene (~1 m.y.).
Rocks:	Basalts and andesites from Kantô area of Japan.
Number and polarity:	11, ?
Type of NRM:	TRM.
Method:	One-interval heating and cooling in Earth's present field.
VDM:	Calculated on the assumption that the effective latitude of the site in Pleistocene times was the same as the present geographic latitude (40°N). VDM:=5·50, 1·70 (30·9%), 0·49 (8·9%).

3. **Irving, Stephenson & Major (1965)**

Age:	Mainly Pleistocene to Recent.
Rocks:	Basalt dykes, sills and lavas plus intrusion in Lower Tertiary limestone, from Heard Island.
Number and polarity:	11, *N*, *R* and discordant.
Type of NRM:	TRM.
Method:	A.c. demagnetization at 150 oe of NRM and TRM (obtained through heating and cooling in nitrogen atmosphere).
VDM:	Calculated from mean field intensity (0·40 oe) and mean inclination (−66°). VDM=6·30 (range 1·73–13·50).

4. **Koenigsberger (1938)**

Age:	Tertiary (~35 m.y.).
Rocks:	Various igneous, mainly from Germany.
Number and polarity:	18, ?
Type of NRM:	TRM.
Method:	One-interval heating and cooling in Earth's present field.
VDM:	Calculated approximately on the following basis: Data gives J_n/J_T=0·65 (mean). Taking coordinates (51°N, 12°E) for Germany, present *F*=0·47 oe (Vestine *et al.* 1947). Assume 5° continental drift (Blackett *et al.* 1960, Fig. 2). VDM=5·02, 2·31 (47·1%), 0·54 (10·7%).

5. **Clegg, Almond & Stubbs (1954)**

Age:	Triassic (~180 m.y.).
Rocks:	Sandstone from Cheshire, England.
Number and polarity:	?
Method:	Artificial deposition in the laboratory.
VDM:	Unknown number of experiments gave $J_n/J_D \sim \frac{1}{3}$ and palaeomagnetic latitude 32·6°N, whence: VDM~3·24.
Note:	Site latitude~53°N.

Table 7—*continued*

6. Koenigsberger (1938)	
Age:	Permo-Carboniferous (~300 m.y.).
Rocks:	Various igneous, mainly from Germany.
Number and polarity:	7, ?
Type of NRM:	TRM.
Method:	One-interval heating and cooling in Earth's present field.
VDM:	Calculated approximately on following basis: Data gives $J_n/J_T = 0{\cdot}15$ (mean). Taking coordinates (51°N, 12°E) for Germany, present $F = 0{\cdot}47$ oe (Vestine *et al.* 1947). Assume 51° continental drift (Blackett *et al.* 1960, Fig. 2). VDM=1·99, 0·66 (33·2%), 0·25 (12·6%).

The study of magnetic *directions* in materials that acquired their magnetizations during the past few thousand years has been directed in the main towards the elucidation of the nature of secular variation of the non-dipole fields. This has been made possible by the large number of man-made objects and structures (for example, pottery and kilns) produced during this period as well as historic lavas. These materials are not only magnetic but are frequently datable to a high degree of accuracy by modern radiometric methods (especially carbon-14 analysis) and in relation to historic events. Archaeomagnetic field direction data have been combined with direct measurements of the Earth's field to give some idea of the amplitudes and periods of secular changes in direction.

In the case of direct field direction measurements, since the main geomagnetic dipole has not been observed to move since about 1800 (Bullard *et al.* 1950) any changes in direction may be attributed to changes in the non-dipole field. But during this period the *intensity* of the main dipole has been decreasing at the rate of about 5% per century (Section 4), and hence in the analysis of recent field intensities obtained

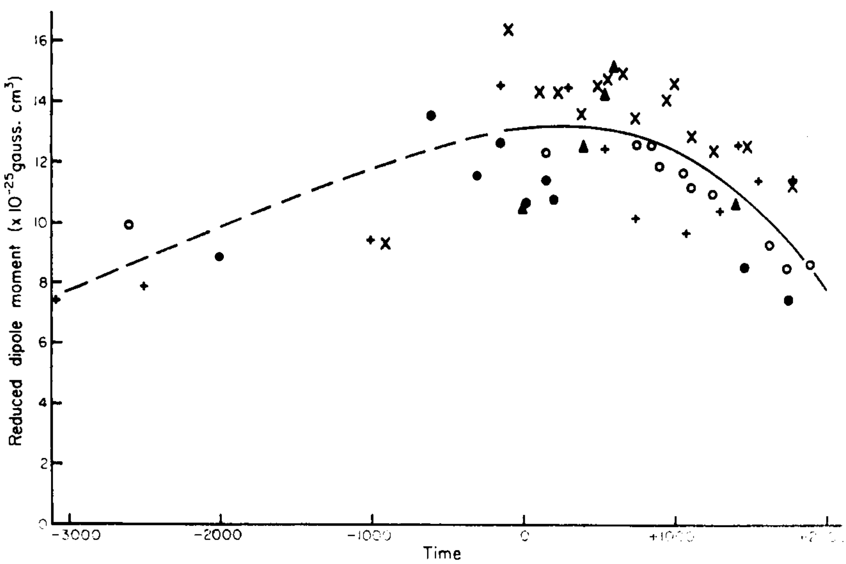

Fig. 5. Historic and archaeological reduced dipole moments plotted against time. Times A.D. are expressed as positive numbers; times B.C. are expressed as negative numbers. ● Thellier & Thellier; ○ Burlatskaya; × Sasajima & Maenaka; + Nagata, Arai & Momose; ▲ Nagata, Kobayashi & Schwarz.

by archaeomagnetic methods both variations in dipole and non-dipole field intensity must be considered.

In all, fifty-three field intensity values covering the period from -8000 to $+1900$ have been presented in various forms. For the present review, reduced dipole moments have been calculated from the published data assuming the main dipole to have remained fixed in its present inclination ($11\frac{1}{2}°$ to the rotational axis) (Section 6). Fifty-two of the RDMs are plotted against time in Fig. 5. The RDM at -8000 due to Johnson *et al.* (1948) has been omitted because of the scale of the diagram.

The following points may be noted:

(a) During the past 2000 years the geomagnetic dipole moment has apparently decreased by about one-third of its peak value to the present directly observed value of $8{\cdot}0 \times 10^{25}$ gauss.cm^3 (Fig. 5). Prior to the year 0 the dipole moment was increasing. The scatter of the points in Fig. 5 is large; and hence in order to show the main trend of the result more clearly the RDMs have been meaned in groups of ten commencing with the youngest. The results of this are shown in Fig. 6.

In Fig. 7 the results from individual workers are plotted separately. In each case there is a general decrease in dipole moment with time since about the year 0, although there are variations within the decrease. (A possible exception is NKS, but in this case the data are too few to judge properly.) These variations will be discussed below.

(b) The question arises as to whether the results indicate a real variation in the dipole moment or whether they merely represent the effect of secular variation of the non-dipole field or wobble of the main dipole. The three possibilities will be discussed in turn.

(i) *Secular variation.* The quasi-period deduced by extrapolation of the curve in Fig. 6 is of the order of 10^4 years, whereas non-dipole field variations generally possess period in the range 10^2–10^3 years (Cox *et al.* 1964). In particular, curves of the archaeomagnetically determined inclination for Sicily (Chevallier 1925), Rome (Chevallier 1953), Paris (Thellier 1953), Japan (Watanabe 1958, Yukutake 1961) and Russia (Burlatskaya 1962) exhibit quasi-periodicities of 400–1000 years. This is in good agreement with the well-known directly observed secular variation curves for

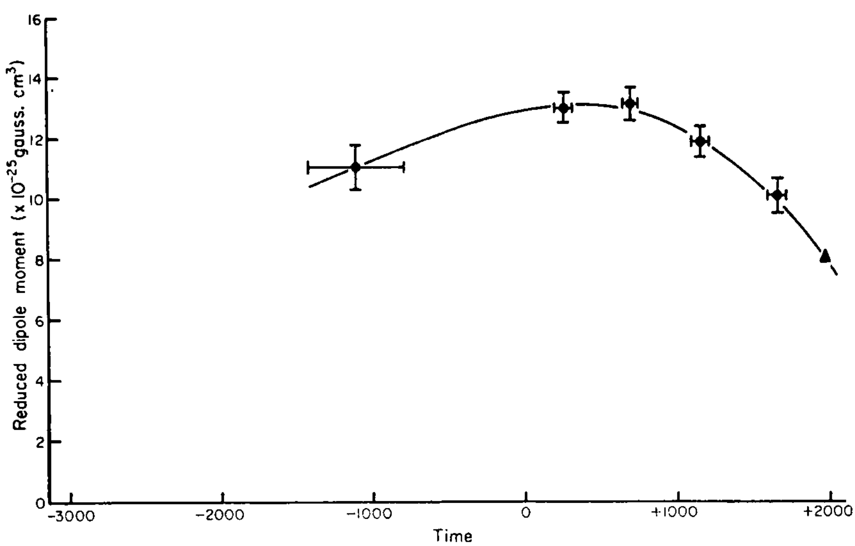

FIG. 6. Historic and archaeological reduced dipole moments (meaned in groups of ten commencing with the youngest) plotted against time. Error bars represent standard errors of the means. Times A.D. are expressed as positive numbers; times B.C. are expressed as negative numbers.

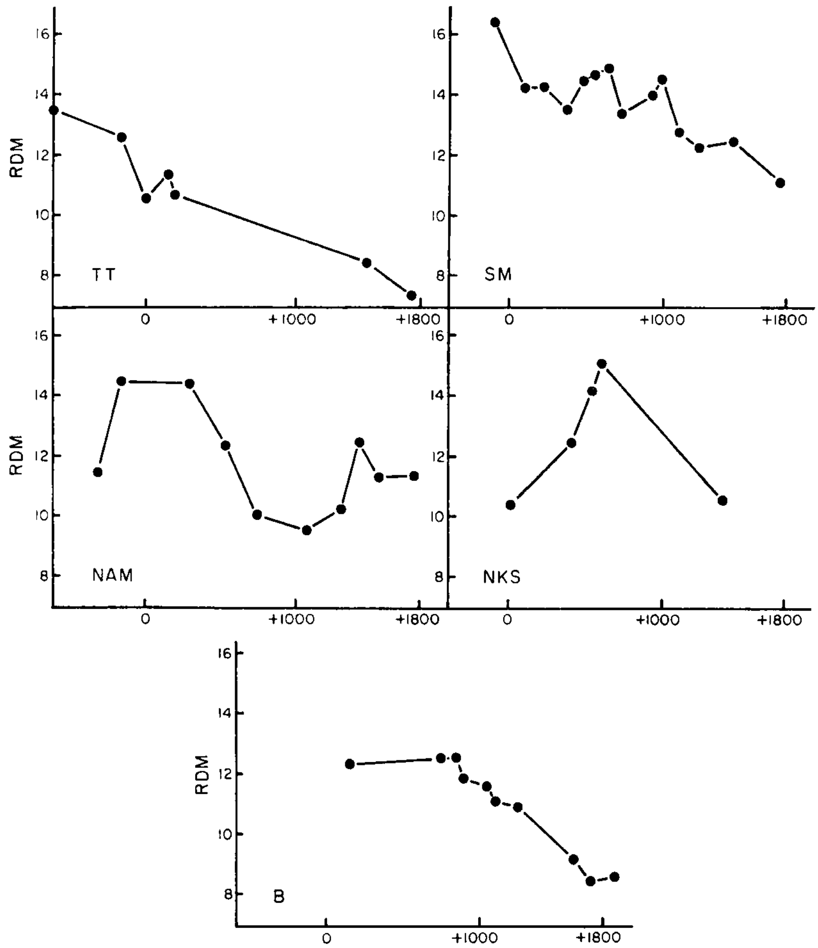

FIG. 7. Histoiic and aechaeological reduced dipole moments (computed from the data of individual authors) plotted against time. RDMs are in units of 10^{25} gauss.cm³. TT=Thellier & Thellier; SM=Sasajima & Maenaka; NAM=Nagata, Arai & Momose; NKS=Nagata, Kobayashi & Schwarz; B=Burlatskaya.

London and Paris which appear to complete a cycle in about 500 years. Bullard *et al.* (1950) have shown by analysing direct field data obtained between 1907 and 1945 that the main features of the non-dipole field are drifting westward at an average rate of ($0{\cdot}18° \pm 0{\cdot}015°$) per year, which implies a period of about 1800 years were any given feature to persist for so long. Yukutake (1962), using a different analytic method, arrived at an average rate of 0·2° a year for the westward drift. Not only does the non-dipole field drift westward, but due to the growth and decay of eddies in the Earth's core the intensity of the field varies with periods of up to 10^3 years (Cox *et al.* 1964).

The curve in Fig. 6 thus exhibits a quasi-periodicity of about one order of magnitude higher than that of known non-dipole field effects. Further, the dipole moments have been obtained from four widely separated areas in three different continents, so that it is unlikely that the cause of the RDM variation lies in any phenomenon as localized as non-dipole features.

(ii) ***Dipole wobble.*** The wobble of the main dipole has a period of the order of 10^4 years (Cox *et al.* 1964), and hence it is tempting to ascribe the curve in Fig. 6 to the effects of such a wobble. There are arguments against this idea, however. Firstly, if dipole wobble were responsible for the observed effects there would be a correlation between magnetic inclination and field intensity (the greater the inclination the greater should be the field intensity). Thellier *et al.* (1959) presented combined directional and field intensity data, and no such correlation was evident. Secondly, since the angular standard deviation of dipole wobble is only about 11° (Cox *et al.* 1964) a field intensity change of no more than 15% could be attributed to this cause in middle latitudes, compared to the 50% indicated by Fig. 6.

(iii) *Variation of dipole moment.* The above arguments suggest that the curve in Fig. 6 is due neither to secular variation of the non-dipole field nor wobble of the main dipole; and hence it is concluded that there has been a real variation in the dipole moment. Fig. 6 indicates that over the past 150 years the geomagnetic dipole moment has been decreasing at the rate of about 7% per century. This is in good agreement with the value of 5% per century deduced from direct field measurement (Section 4).

(c) If the dipole moment variation is real, having a quasi-period an order of magnitude higher than the period of secular variation of the non-dipole field, it might be expected that secular variations would be superimposed on the main dipole variation. There is a little evidence for this. The curves in Fig. 7 due to Nagata *et al.* (1963) and Sasajina *et al.* (1964) show apparent variations having periods of about 1000 years and 400 years respectively, superimposed on the main decrease of dipole moment since the year 0. However, since both of these curves arise from Japanese rocks and yet possess different characteristics it must be concluded that the data are insufficient to accurately resolve secular variation.

Nevertheless it seems likely that a large part of the scatter in Fig. 5 is due to secular variation in field intensity. However, it will be recalled (Section 6) that the RDMs plotted in Fig. 5 were calculated assuming the main dipole to have maintained its present inclination over the whole period covered by the results and hence some of the scatter could be caused by any dipole wobble that might have occurred. Kawai *et al.* (1965) have suggested, on the basis of an analysis of published secular variations directional data, that for the past 1200 years the Earth's dipole has been rotating about the Earth's geographic axis in an anticlockwise direction. Sasajina *et al.* (1965) have shown that if this hypothesis is used to correct all field intensity results to a common latitude the scatter of points is only slightly reduced. Most of the scatter in Fig. 5 must therefore be due to non-dipole field variations or experimental error.

To summarize, the evidence suggests that the curve in Figs. 5 and 6 indicates a real variation in the Earth's dipole moment over the past few thousand years. The scatter of points in Fig. 5 is probably due mainly to non-dipole field variations and experimental error. The rate of decrease of the dipole moment over the past 150 years (7%) as deduced from archaeomagnetic data agrees well with the directly observed rate (5%).

8.2. *Ancient geomagnetic field intensities: geological time.* The five accurate mean virtual dipole moments detailed in Table 6 and the six approximate mean virtual dipole moments detailed in Table 7 are plotted as a function of time (m.y.) in Fig. 8. Individual VDMs from the five accurate determinations are also plotted as histograms as follows:

Fig. 9.—Momose; Japanese; normal and reversed specimens plotted together and separately.

Fig. 10.—Smith; Icelandic; normal and reversed specimens plotted together and separately.

Fig. 11.—Smith; Scottish; all specimens plotted together (all seventeen specimens except one were reversed).

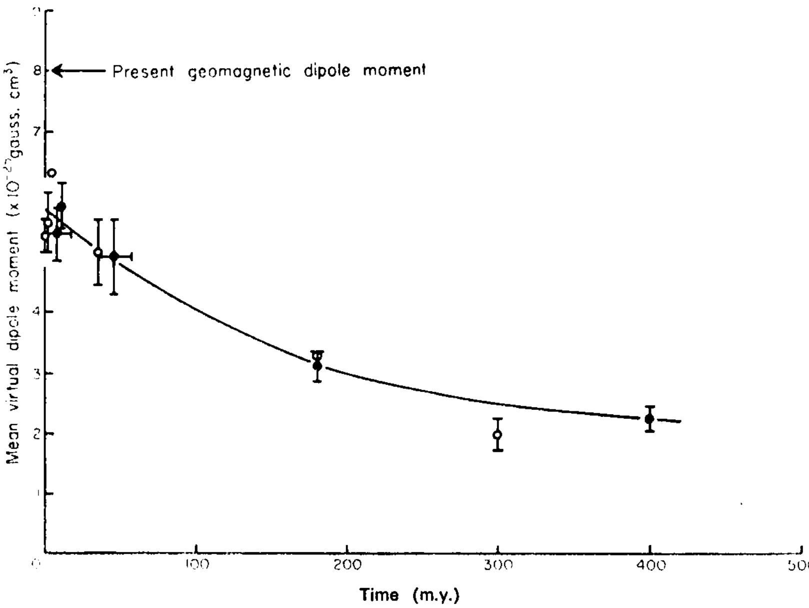

FIG. 8. Mean geological virtual dipole moments plotted against time (m.y.). ● accurate (Class 1) values. ○ approximate (Class 2) values. Error bars in VDM represent standard errors of the means; error bars in time represent range of potassium-argon ages where appropriate.

Fig. 12.—Van Zijl; South African; normal and reversed specimens plotted together and separately.

Fig. 13.—Briden; Australian; all specimens plotted together (all normal).

The following points may be noted from Figs. 8–13:

(i) Fig. 8 indicates that for at least the past 400 m.y. the mean geomagnetic dipole moment was smaller than the Earth's present dipole moment ($8{\cdot}0 \times 10^{25}$ gauss.cm^3). The highest accurate mean VDM ($5{\cdot}77 \times 10^{25}$ gauss.cm^3 due to Momose) is 28% below the present moment, the lowest accurate mean VDM ($2{\cdot}25 \times 10^{25}$ gauss.cm^3 due to Briden) is 72% below the present moment, and all other accurate mean VDMs lie between these limits. However, Figs. 9–13 indicate that VDMs from some individual specimens were higher than the present dipole moment. For example, of the thirty-seven individual VDMs obtained by Momose (Fig. 9), six were higher than $8{\cdot}0 \times 10^{25}$ gauss.cm^3. The highest was $14{\cdot}60 \times 10^{25}$ gauss.cm^3; the highest accurate VDM yet obtained from a geological specimen.

(ii) The mean geomagnetic dipole moment appears to have been increasing for the past 400 m.y. (Fig. 8). In view of the paucity of data beyond the Tertiary it would be unwise to consider the increase of VDM with time as more than a possibility until many more data are available.

(iii) Much more convincing is the close agreement of the seven Tertiary mean virtual dipole moments (Fig. 8), notwithstanding the approximate nature of four of them. All six of the mean VDMs lie within $\pm 15{\cdot}6$% of their mean value ($5{\cdot}45 \times 10^{25}$ gauss.cm^3). All three of the accurate mean VDMs lie within $\pm 7{\cdot}9$% of their mean value ($5{\cdot}35 \times 10^{25}$ gauss.cm^3). Such close agreement gives confidence in the validity of the mean VDMs. The three accurate mean VDMs were obtained from eight different types of rock collected from three different areas of the world. Although the mean of these accurate determinations is 33% lower than the Earth's present dipole

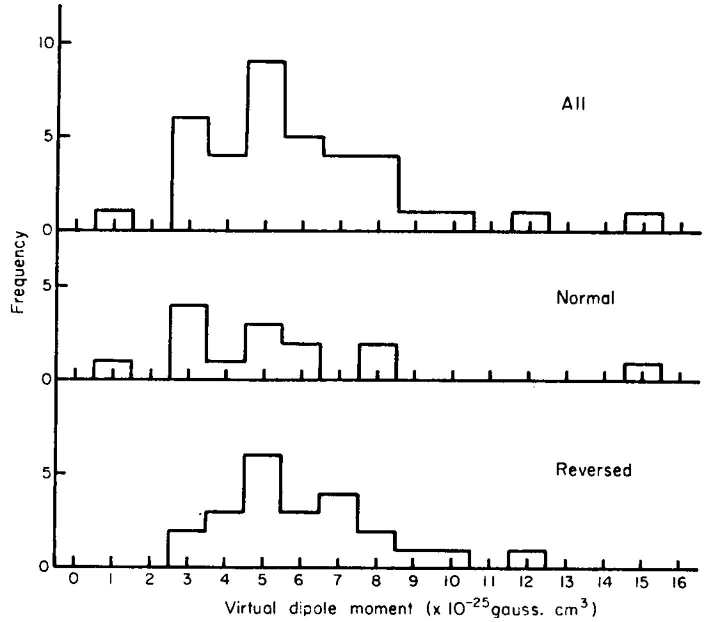

FIG. 9. Histograms of the Japanese Tertiary virtual dipole moments computed from the data of Momose (1963).

moment, it is unlikely that, for example, magnetic decay would have had such a similar effect on so many different rock types as to produce mean VDMs in such close agreement with each other.

(iv) Fig. 8 indicates that on the geological time scale the expected present dipole moment should be $(5{\cdot}6 \pm 0{\cdot}5) \times 10^{25}$ gauss.cm^3. This is 30% below the presently observed dipole moment of $8{\cdot}0 \times 10^{25}$ gauss.cm^3.

(v) No obvious differences may be discerned between the normal and reversed intensity fields as represented by the VDMs plotted in Figs. 9, 10 and 12. The data

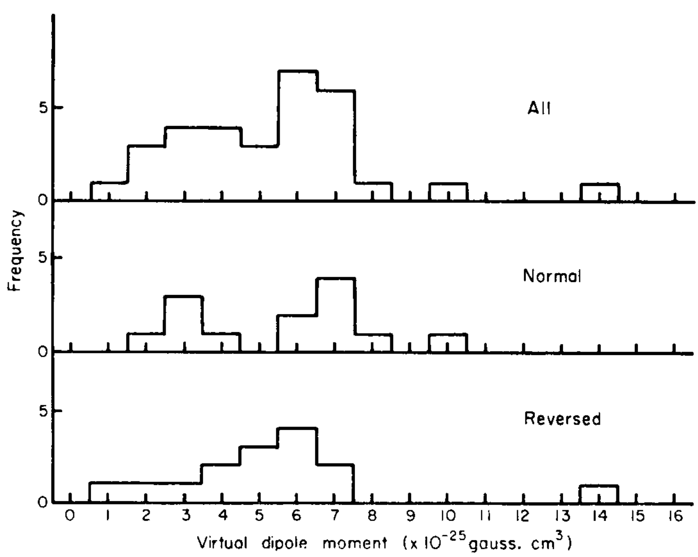

FIG. 10. Histograms of the Icelandic Tertiary virtual dipole moments obtained by Smith (1967).

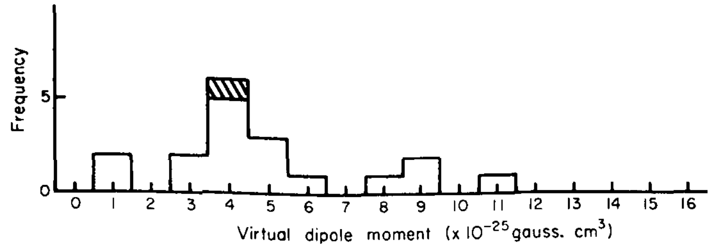

FIG. 11. Histogram of the Scottish Tertiary virtual dipole moments obtained by Smith (1967). All specimens were reversed except one (shaded).

are too few at present to rule out the possibility that differences in the normal and reversed fields are non-existent.

(vi) Although the histograms of the VDMs (Figs. 9–13) are peaked in the region of their respective mean VDMs, in each case there is a large spread of individual VDMs giving rise to the large standard deviations listed in Table 6. The five standard deviations of the VDMs lie in the range $\pm 43{\cdot}1\%$ to $\pm 60{\cdot}5\%$.

There are three possible causes for the large spreads of individual VDMs, namely, experimental error, secular variation of the non-dipole fields and variations in the dipole moment itself. None of the spread will be due to dipole wobble because all

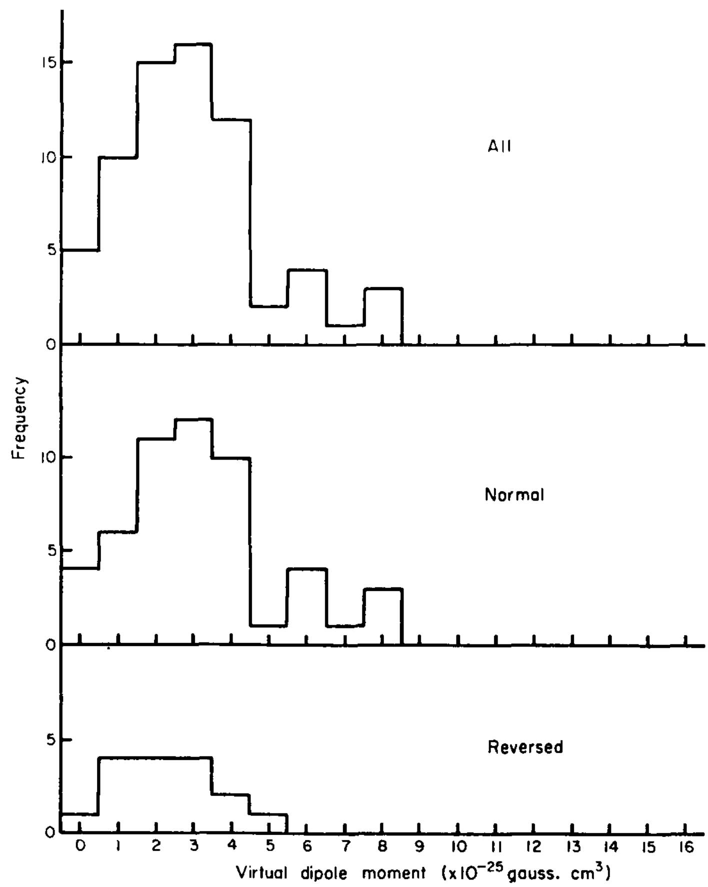

FIG. 12. Histograms of the South African Triassic–Jurassic virtual dipole moments computed from the data of Van Zijl (1961).

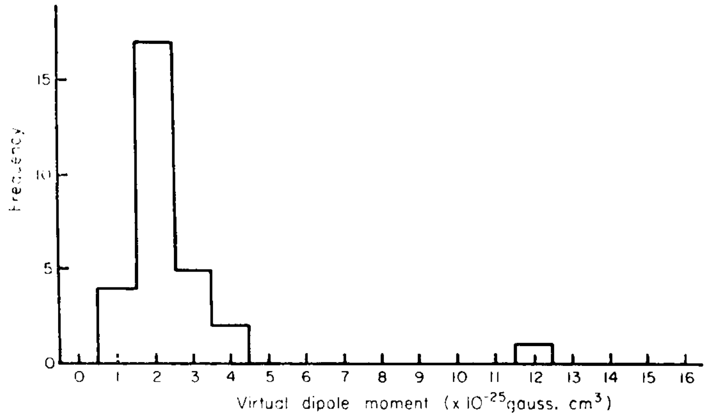

Fig. 13. Histogram of the Australian Silurian–Devonian virtual dipole moments computed from the data of Briden (1966). All specimens were normal.

VDMs were calculated using the relevant measured palaeomagnetic latitudes (Section 5).

The large spreads are not likely to arise through experimental error alone ('experimental error' is here taken to include both errors of measurement and variations introduced through slight alteration in the rock during the laboratory remagnetizing process). It is difficult to estimate the spread expected from this source but, for example, Smith (1967) has estimated that for the VDMs deduced from Scottish baked contacts and dykes the maximum standard deviation due to experimental error should be $\pm 17\%$ (compared with $\pm 52{\cdot}3\%$ obtained from the actual VDMs) and that for the VDMs deduced from Icelandic lavas and baked laterites the maximum standard deviation due to experimental error should be $\pm 13\%$ (compared with $\pm 47{\cdot}2\%$ obtained from the actual VDMs).

The possibility that the large standard deviations in the measured VDMs represent experimental error together with non-dipole field variations will be examined in the following section.

8.3. *The spread of virtual dipole moments caused by non-dipole fields.* In this section will be investigated the possibility that the large spreads of virtual dipole moments (as exemplified by Figs. 9–13) are due mainly to secular variation of the non-dipole field. It will be shown that such secular variation, together with experimental error, is not sufficient to account for the large spreads and hence that actual variations in the dipole moment must be invoked.

Cox *et al.* (1960) suggested that the variations in the Earth's present or recent magnetic field might be used to estimate the 'signal to noise ratio' for palaeomagnetic data. Creer (1962) has computed the angular dispersions in the directions of magnetization introduced by secular variation using three models based on the 1945 field of the Earth; and Cox (1962) has done the same by somewhat different methods. Since both these authors were interested mainly in palaeodirectional studies no attempts were made to analyse field intensity variations. This has now been carried out here in order to determine the fluctuations likely to be produced by secular variation in a set of VDMs from a given latitude.

The present field of the Earth deviates from an axial dipole field in two ways. Firstly, the best fitting dipole (whose strength is at present $8{\cdot}0 \times 10^{25}$ gauss.cm^3) is inclined at $11\frac{1}{2}°$ to the Earth's rotational axis; and secondly, in addition to the dipole there are non-dipole components which are believed to originate in the region of

Table 8

Intensity statistics for the 1945.0 geomagnetic field computed from the data of Vestine et al. (1947)

λ° = *latitude*
F = *field intensity*
σ = *standard deviation about the λ-mean*
ε = *standard error of the λ-mean*

λ°	(i) F for axial dipole* (oe)	(ii) Mean F around λ_n (oe)	(iii) Difference (ii−i) (oe)	(iv) σ (oe)	(v) ϵ (oe)	(vi) σ (%)	(vii) ϵ (%)	(viii) Mean F around λ (oe)	(ix) σ (oe)	(x) σ (%)
N80	0·613	0·561	−0·052	0·019	0·003	3·4	0·5	0·584	0·042	7·2
70	0·592	0·564	−0·028	0·030	0·005	5·3	0·9	0·569	0·061	10·7
60	0·559	0·558	−0·001	0·039	0·007	7·0	1·3	0·543	0·079	14·6
50	0·515	0·532	+0·017	0·047	0·008	8·8	1·5	0·504	0·097	19·3
40	0·464	0·492	+0·028	0·050	0·008	10·2	1·6	0·464	0·098	21·1
30	0·410	0·444	+0·034	0·047	0·008	10·6	1·8	0·424	0·086	20·3
20	0·360	0·400	+0·040	0·039	0·007	9·8	1·8	0·389	0·069	17·8
10	0·324	0·363	+0·039	0·028	0·005	7·7	1·4	0·360	0·047	13·0
00	0·310	0·349	+0·039	0·034	0·006	9·7	1·7	0·349	0·034	9·7
10	0·324	0·356	+0·032	0·059	0·010	16·1	2·8			
20	0·360	0·377	+0·017	0·088	0·015	23·4	4·0			
30	0·410	0·403	−0·007	0·109	0·018	27·0	4·5			
40	0·464	0·436	−0·028	0·121	0·020	27·8	4·6			
50	0·515	0·476	−0·039	0·126	0·021	26·5	4·4			
60	0·559	0·527	−0·032	0·112	0·019	21·2	3·6			
70	0·592	0·574	−0·018	0·081	0·014	14·1	2·4			
S80	0·613	0·607	−0·006	0·045	0·008	7·4	1·3			

Columns (vi) and (vii) are expressed as percentages of the λ-mean (column (ii)).
Column (x) is expressed as a percentage of the λ-mean (column (viii)).
* Calculated for a dipole moment of $8{\cdot}0\times10^{25}$ gauss.cm^3 and an Earth radius of $6{\cdot}37\times10^8$ cm.

the core-mantle boundary (Lowes *et al.* 1951). In his first model, Creer (1962) hypothesized that the scatter of geomagnetic field directions at a particular place in the course of time is the same as that observed for a particular time around the line of latitude passing through that place. The calculated dispersion factors and their latitude dependence were in good agreement with palaeomagnetic data derived from rock formation of different ages, whence Creer suggested that the intensity of the secular variation field relative to the main geomagnetic field has always been much the same as at present.

In the present field intensity analysis it has been assumed that Creer's original hypothesis holds for field intensity as well as direction. The mean field intensity (hereafter termed the λ-mean) was calculated at 10° intervals of latitude by meaning the intensity values at 10° intervals of longitude for the 1945 field. The data for the 1945 field were taken from Vestine *et al.* (1947). For each latitude, the standard deviation (σ) of the intensity values about the λ-mean was calculated ($n = 36$) together with the standard error of the λ-mean (ε). Geographic latitudes were used and hence σ and ε included the effects of the non-dipole field and of the best-fitting dipole inclined at $11\frac{1}{2}°$ to the rotational axis.

All computed data are shown in Table 8 which gives, for each latitude used:

(i) field intensity for an axial dipole of $8{\cdot}0\times10^{25}$ gauss.cm^3;

(ii) λ-mean of field intensity;

(iii) difference between (i) and (ii);

(iv) standard deviation (σ) about the λ-mean;
(v) standard error of the λ-mean (ε);
(vi) σ expressed as a percentage of the λ-mean;
(vii) ε expressed as a percentage of the λ-mean;
(viii) λ-mean for both hemispheres taken together;
(ix) standard deviation (σ) about the λ-mean;
(x) σ expressed as a percentage of the λ-mean.

The standard deviations of the 1945 field strength, expressed as percentages of the respective λ-means, are plotted against latitude in Fig. 14. There is a clear latitude dependence which is double-valued in each hemisphere and different from the single-valued latitude dependence of the angular dispersion as deduced by Creer (1962). The five accurate standard deviations (%) listed in Table 6 are plotted as points in Fig. 14.

The rock data are too few to enable any definite conclusions to be drawn with regard to latitude dependence. What is clear is that the rock VDMs have much higher standard deviations than predicted by the scatter of the 1945 field (by a factor greater than 4 in the northern hemisphere). If the standard deviations of the rock VDMs were due only to the variations of the non-dipole field, the points in Fig. 14 should lie below the curves since the curves include both non-dipole field and inclined dipole effects whereas the rock standard deviations conceal no inclined dipole effect.

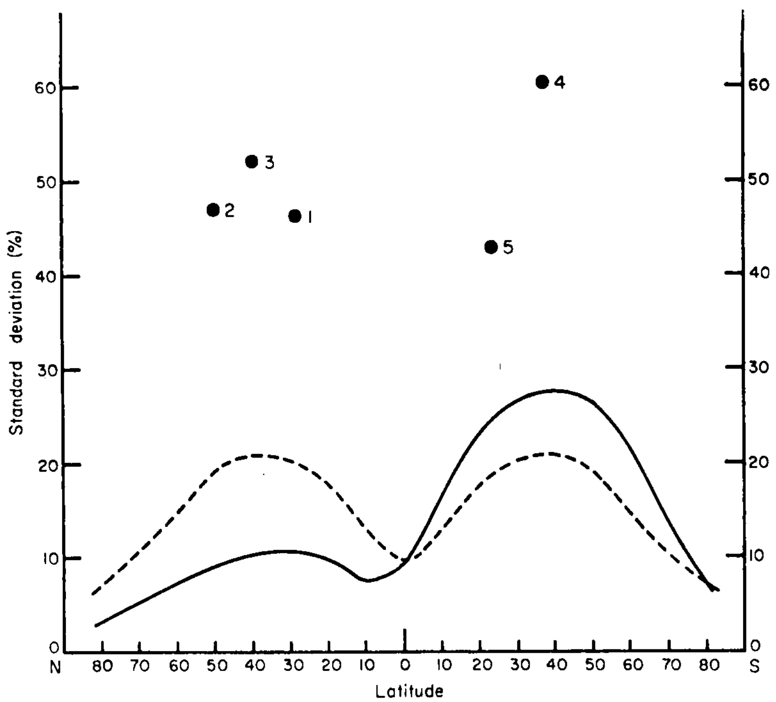

FIG. 14. Percentage standard deviation of the 1945 field intensity around lines of geographic latitude plotted against latitude. Full line is for northern and southern hemispheres computed separately; dotted line is for the two hemispheres computed together. Points represent measured percentage standard deviations of field intensity values from accurate (Class 1) palaeomagnetic data as follows: 1, Momose—Japan; 2, Smith—Iceland; 3, Smith—Scotland; 4, Van Zijl—South Africa; 5, Briden—Australia.

If experimental error is now included we have, for example, for the Scottish and Icelandic VDMs obtained by Smith (1967):

Scottish

Maximum standard deviation due to experimental error $\sim 17\%$. Standard deviation at $\lambda = 40°$ N due to non-dipole field and inclined dipole $\sim 10\%$.

Therefore, combined standard deviation $< 20\%$.

Since the 17% is a maximum value and the 10% includes scatter due to an inclined dipole, the combined standard deviation of the measured VDMs should be less than 20%. This compares with a measured standard deviation of 52·3% as deduced from the spread of the VDMs.

Icelandic

Maximum standard deviation due to experimental error $\sim 13\%$. Standard deviation at $\lambda = 50°$ N due to non-dipole field and inclined dipole $\sim 9\%$.

Therefore, combined standard deviation $< 16\%$.

Again, since the 13% is a maximum value and the 9% includes scatter due to an inclined dipole, the combined standard deviation should be less than 16%. This compares with a measured standard deviation of 47·2% as deduced from the spread of the VDMs.

It is clear, therefore, that if the relative intensities of the dipole and non-dipole fields have always been similar (as suggested by Creer (1962) and Irving *et al.* (1964) on the basis of directional analysis) then the observed standard deviations in the measured VDMs are between two and three times larger than expected from experimental error and non-dipole field variations alone. The simplest and most likely explanation is that actual fluctuations in the dipole moment have taken place throughout the period covered by the rock record (in addition to field reversals). This possibility will be discussed now.

8.4. *Fluctuations of the Earth's dipole moment.* It was shown in the previous section that experimental error and secular variation of the non-dipole field are not sufficient to explain the large standard deviations in the measured VDMs. The most likely explanation is that the geomagnetic dipole moment fluctuates in strength whilst maintaining any given polarity and hence causes part of the spread of the measured VDMs (Figs. 9–13). It is known from direct observation of the dipole moment over the last 150 years that it can vary (Section 4). Further, the historic and archaeomagnetic field intensity data indicate that the dipole moment has been decreasing since about the year 0, prior to which it was increasing (Section 8.1).

In 1955 Bullard introduced the disk dynamo as an analogy to the homogeneous dynamo thought to exist in the Earth's core and thought to be the cause of the geomagnetic field. Rikitake (1958) showed theoretically that the currents in a system of two mutually exciting disk dynamos could reverse. Allan (1958, 1962), using a digital computer, showed that under a wide range of conditions not only that the currents in such a system can reverse, but that within any given polarity the currents fluctuate about their equilibrium values. Mathews *et al.* (1963) have obtained similar solutions using an analogue computer.

The successful simulation of reversals, similar to those observed in palaeomagnetism, by the disk dynamo model leads to the possibility that other predictions of the dynamo model might be applicable to the Earth's magnetic field. In particular, it is suggested here that both palaeomagnetic and archaeomagnetic field intensity results indicate that fluctuations of the geomagnetic dipole moment, in addition to reversals, have occurred.

The following points may be made in relating dipole moment fluctuations to palaeomagnetic and archaeomagnetic field intensity data:

(i) It was suggested above that actual variations of the dipole moment are necessary to explain the large spreads in the measured VDM distributions (Figs. 9–13). It is not possible at present to estimate precisely the standard deviations of the VDMs due to dipole fluctuation, but the following table has been compiled from published data. If σ is the total standard deviation of the VDMs, σ_e is that due to experimental error, σ_s that due to secular variation of the non-dipole field and σ_f that due to dipole moment fluctuations, then:

$$\sigma^2 = \sigma_e{}^2 + \sigma_s{}^2 + \sigma_f{}^2.$$

	Momose Japanese	*Smith Scottish*	*Smith Icelandic*	*Van Zijl South African*	*Briden Australian*
$\sigma\%$	46·4	52·3	47·2	60·5	43·1
$\sigma_e\%$	< 5·0	<17·0	<13·0	?	?
$\sigma_s\%$	<11·0	< 9·0	<10·0	<25·0	<28·0
$\sigma_f\%$	>44·8	>48·7	>44·2	> ~55·1	> ~32·8

The inequality sign appears in the σ_e values because these values are maxima and in the σ_s values because they include scatter due to the inclined main dipole (Section 8.3). The calculated σ_f values are therefore minima.

The table shows that over 90% of the spread of VDMs is due to the supposed dipole moment fluctuations, with the exception of the Australian results (76%). In this case the lower limit of σ_f is much lower than in the other four cases probably because although Briden measured twenty-nine specimens they were cut from only four samples. It is therefore likely that the full range of dipole moment fluctuation was not covered by Briden's samples.

(ii) The variation of dipole moment with time over the past few thousand years as deduced from archaeomagnetic data (Figs. 5 and 6) presumably represents a small part of one of the supposed dipole moment fluctuations.

(iii) It follows from a dipole moment variation that at any instant of time the dipole moment may be different from the mean dipole moment in any given polarity. This would explain why the present mean VDM as deduced from Fig. 8 ($5{\cdot}6 \times 10^{25}$ gauss.cm^3) is 30% below the present directly measured dipole moment ($8{\cdot}0 \times 10^{25}$ gauss.cm^3) (Section 8.1).

It has been shown, therefore, that by invoking fluctuations of the geomagnetic dipole moment in addition to changes in polarity, it is possible to consistently explain the features of the present and ancient geomagnetic dipole field strength as deduced from direct measurement and palaeomagnetic and archaeomagnetic data.

8.5. *Ancient geomagnetic field intensities: transition zones.* Of great interest in palaeomagnetism is the rock record of the behaviour of the geomagnetic field during polarity reversal. The reality of field reversal has been widely discussed previously (Irving 1964, and Wilson 1966, have recently reviewed the evidence) and hence no summary of the relevant arguments will be presented here. Given the occurrence of reversals of the Earth's dipole, however, questions arise as to the nature of the polarity change. The dynamo theory of the Earth's field, which places the source of the main field in motions of the fluid core, implies that during a reversal the dipole reduces to zero followed by an increase in the opposite sense (Elsasser 1956); but there is the possibility that the dipole rotates through 180° with or without change of moment. The further question arises as to what happens to the non-dipole field intensity during a dipole reversal. For example, does the non-dipole field intensity decrease or remain steady during the dipole transition?

Ancient field intensity values from two transition zones have been reported, namely, by Van Zijl (1961) for the Stormberg lavas of Basutoland (Triassic–Jurassic) and by Momose (1963) for Japanese lavas (Tertiary). These data will be analysed here, whence it will be shown that in neither case did the polarity reversal take place by a 'flip-over' of the dipole without decrease in moment, that in neither case did the non-dipole field fall to zero during the transition and that the field intensity results are consistent with constant proportionality of non-dipole to *mean* dipole field strength throughout geological time.

Creer (1962) computed for the 1945 field the angular dispersion due to non-dipole components, and by comparing the dispersion with angular dispersions calculated from palaeomagnetic data, showed that during geological time the intensity of the non-dipole field relative to the *mean* main dipole field has always been approximately the same. A similar conclusion was reached by Irving *et al.* (1964) using a different approach. These authors used a statistical model of the Earth's magnetic field in which the geocentric dipole is axial, producing an equatorial field H_0, and in which the non-dipole fluctuations are simulated by a component h which has constant magnitude but whose direction is randomly distributed in time. For an assumed probability distribution of angles between h and H (where H is the field intensity produced by the geocentric axial dipole at latitude λ) it was shown that $f^2 \sim 3k^{-1}$, where $f = hH^{-1}$ and k is the Fisher (1953) precision parameter estimate. Defining $f_0 = hH_0^{-1}$ it was further shown that:

$$f_0 = f(1+3\sin^2\lambda)^{\frac{1}{2}}. \tag{6}$$

Values of f_0 for the 1945 field and from palaeomagnetic directional data (that is, calculated from k) are shown in Table 9, which is taken from Irving *et al.* (1964). The small standard deviation in f_0 for the 1945 field indicates that the field may be accurately simulated by the model for which $f_0 = 0{\cdot}4$. Further, since the 1945 field is known to be mainly dipolar and since the mean value of f_0 calculated from palaeomagnetic data is also about 0·4, the results are consistent with a constant ratio of non-dipole to *mean* dipole field strength since the Pre-Cambrian (the age of the earliest results included in the palaeomagnetic f_0 calculations).

The methods of Creer and Irving *et al.* both used palaeomagnetic measurements of *direction* to conclude something about the relative *intensities* of the dipole and non-dipole fields. Yet a third method is possible, utilizing field intensity values directly. This approach was suggested by Irving *et al.* but no detailed numerical results were presented. The idea is that if field reversal has occurred by the decrease of the dipole moment to zero followed by a build-up in the opposite sense, then rocks acquiring their magnetization at the time when the dipole moment was zero will have become magnetized only in the non-dipole field which will not necessarily have been zero when the

Table 9

Mean estimates of $f_0 = h/H_0$ derived from the 1945 field and the palaeomagnetic observations using the Model A of Irving and Ward. N = number of observations; SD = standard deviation.

The paleomagnetic results are grouped as follows: (1) dispersion values obtained from a Fisher analysis of site means; (2) between-site dispersions obtained by the two-tier method of Watson & Irving (1957). Data from Irving & Ward (1964).

Data from:	N	Mean f_0	SD
Analysis of 1945 field, unweighted	9	0·396	0·013
Analysis of 1945 field, weighted	9	0·395	—
Igneous rocks (1)	19	0·42	0·12
Igneous rocks (2)	10	0·31	0·12
Sedimentary rocks (1)	5	0·41	0·12
Sedimentary rocks (2)	1	0·31	—
Average of palaeomagnetic values	35	0·38	0·13

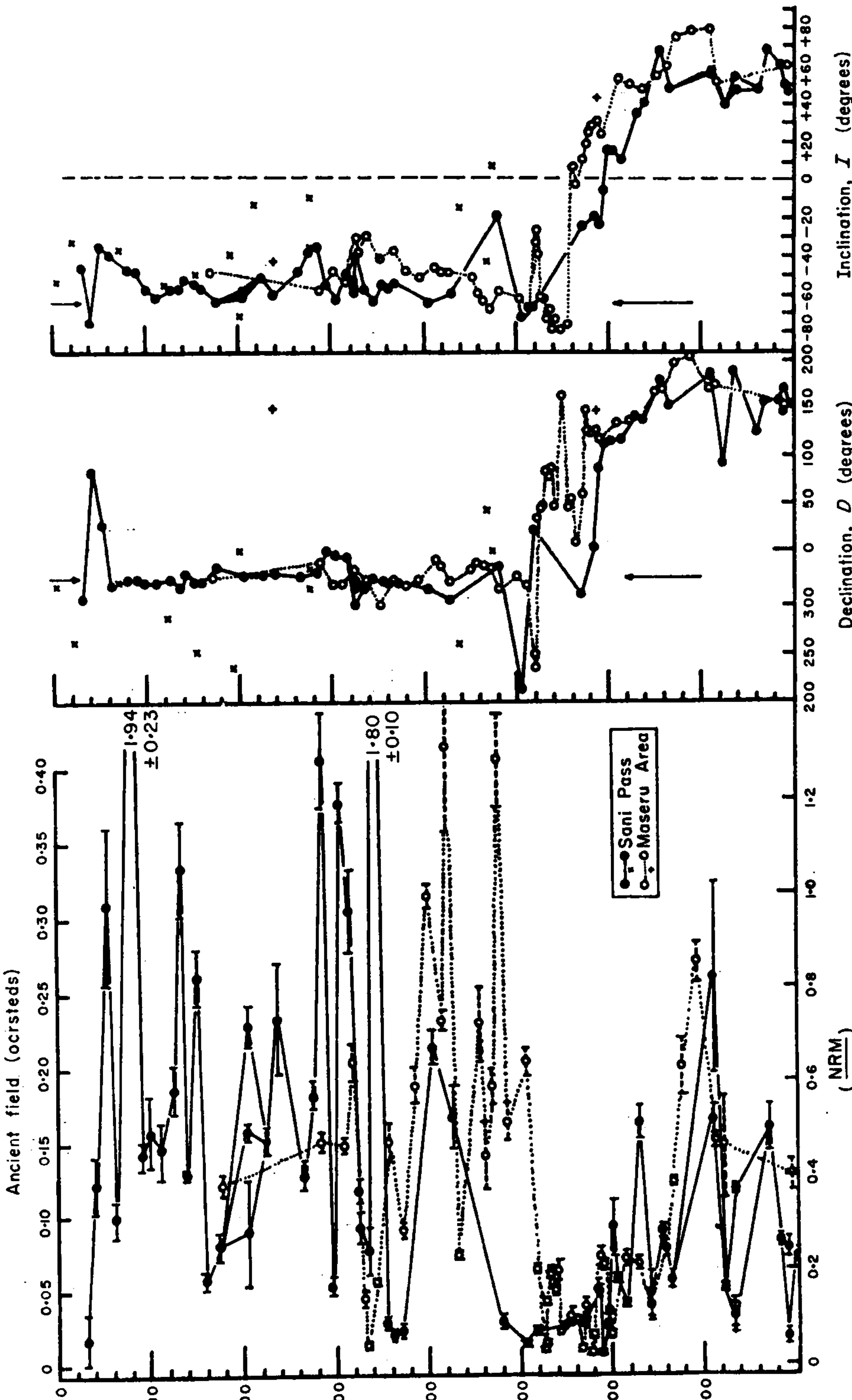

FIG. 15. Ancient field intensity and directional data for the Stormberg lavas plotted against the height in feet above the base of the lavas. Normal and reversed zones are separated by a transition zone (after Van Zijl, Graham & Hales 1962).

dipole field was zero. Field intensities deduced from such rocks may then be compared with field intensities deduced from rocks magnetized at approximately the same time but in the fully normal or reversed dipole field. In this way it would be possible to calculate the approximate ancient f_0 ratio.

The two transition zones will be considered in turn.

(i) *Stormberg lavas* (*Van Zijl* 1961). The field intensity data obtained by Van Zijl are shown in Fig. 15 (taken from Van Zijl *et al.* 1962). In the transition zone the field intensity fell rapidly, and so thirty-eight values scaled from the 'trough' of the intensity curve have been utilized here.

The mean transition zone field intensity was $(0{\cdot}05 \pm 0{\cdot}03)$ oersted. The mean equatorial field intensity for the fully normal and reversed fields was $(0{\cdot}13 \pm 0{\cdot}07)$ oersted. Now any given dipole produces its minimum field intensity in the equatorial plane. It follows, therefore, that if the dipole rotates through 180° with no change in moment during a reversal the transition zone field intensity cannot be less than the equatorial field. In this transition zone the mean field intensity is only 38% of the equatorial dipole field and hence it may be concluded that simple rotation of a constant strength dipole has not occurred.

From the above figures, $f_0 = (0{\cdot}39 \pm 0{\cdot}35)$ where the error is the standard deviation. Since this f_0 value agrees with those from the 1945 field and palaeomagnetic directional data (Table 9) to within less than 3%, it seems likely that the field remaining in the transition zone is the non-dipole field. Further, Fig. 15 indicates that this field does not reach zero at any point in the transition.

(ii) *Japanese lavas* (*Momose* 1963). The basic field intensity data obtained by Momose are shown in Fig. 16, where J_n/J_T ratios are plotted as a function of measured virtual geomagnetic polar latitude. In this case the transition zone is not so well

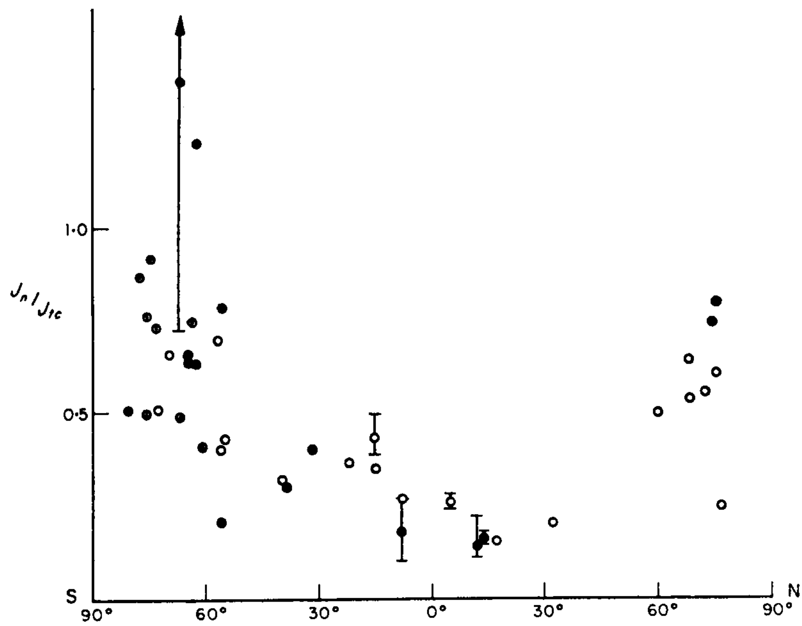

FIG. 16. J_n/J_T ratios for Japanese rocks plotted against corresponding palaeomagnetic pole position latitudes. Normal and reversed zones are separated by transition zone specimens (after Momose 1963). ○, Shigarami; ●, Komoro; ⊙, Enrei.

defined and so two different estimates of the non-dipole field have been made as follows:

(a) using the five points in the 'trough' of the transition for which the equivalent pole positions lie within $\pm 12°$ latitude of the equator ($h = 0{\cdot}095 \pm 0{\cdot}022$ oersted);

(b) using the seven points in the 'trough' whose equivalent pole positions lie within $\pm 17°$ latitude of the equator and for which $J_n/J_T < 0{\cdot}3$ ($h = 0{\cdot}086 \pm 0{\cdot}020$ oersted).

The object was to select only those field intensity values obtained from rocks which were magnetized when the main dipole strength was closest to zero.

The equatorial field intensity deduced from specimens magnetized in the normal and reversed fields was $(0{\cdot}222 \pm 0{\cdot}095)$ oersted, whence:

$$\text{(a) } f_0 = 0{\cdot}43 \pm 0{\cdot}20$$

$$\text{(b) } f_0 = 0{\cdot}39 \pm 0{\cdot}19$$

(errors are standard deviations).

These f_0 values agree with those from the 1945 field and palaeomagnetic directional data (Table 9) to within less than 13% and 3% respectively, again suggesting that the dipole reduces to zero during the transition but that the non-dipole field remains.

Since the transition zone field is less than 43% of the equatorial dipole field, as before the dipole cannot have rotated through 180° without change of moment. Fig. 16 shows further that the non-dipole field never reduces to zero during the transition.

Both sets of transition zone field intensity data are thus consistent with the view that during a transition the dipole moment passes through zero, but that the non-dipole field remains approximately constant. The data are not consistent with the view that during a reversal the dipole rotates through 180° without change of moment. The results are further consistent with constant proportionality of non-dipole to *mean* dipole field strengths over geological time, where $f_0 \sim 0{\cdot}4$.

9. Conclusions

It is not proposed to summarize in detail the numerous conclusions drawn from the field intensity data and presented in Section 8. There are, however, several points which have arisen during the preparation of this review and from the field intensity work carried out by the present author which may be usefully mentioned here.

(i) One of the main technical difficulties in field intensity determination is to control the nature and amount of magnetic material present in the specimen under investigation, during the heating and cooling necessary to produce an artificial TRM. Since specimens in which no such changes take place during heating are rare it is frequently necessary to utilize material in which some alteration takes place and hence difficulties arise in the estimation of the degree of alteration.

It has been shown (Havard 1965) that even when rock specimens are heated in a vacuum, oxidation or reduction may take place in the presence of the rock minerals themselves. Since heating and measuring cores in a vacuum is also a technically difficult operation, the simplest and most reliable technique is to heat in air specimens which are naturally highly oxidized. Baked materials have therefore been widely used in field intensity determination. Highly-oxidized igneous rocks are equally useful when obtainable, the class 5 lavas described by Wilson *et al.* (1967) being especially useful. The discovery of highly-oxidized lavas depends largely on the use of efficient rock polishing techniques and a high-powered microscope for examination of the opaque magnetic minerals present. The selection of specimens by a study of the opaque petrology is thus the most important recent development in the progress of field intensity determination. However, much igneous material does not contain highly-oxidized opaque minerals. At present no perfect technique is available for controlling magnetic mineral alteration during the heating of such specimens.

(ii) The evidence presented in Section 8 led to the suggestion that scatter of the virtual dipole moments is due not only to experimental error and secular variation of the non-dipole fields but mainly to actual fluctuation in the dipole moment. It should be realized therefore that large spreads of VDMs do not necessarily imply invalidity of the results. However, it follows that many specimens must be measured in order to obtain a reliable estimate of the mean VDM.

In reviewing palaeodirectional data, Wilson (1966) considered that *at least* ten independently oriented specimens are necessary to mean out the effects of experimental error and secular variation. In the case of field intensity determination where there is an additional source of scatter the number of specimens measured should be even higher. It is difficult to say exactly how many independently oriented specimens should be utilized; but it is suggested here that in future about twenty should be regarded as a minimum number.

(iii) The full potential of field intensity determinations cannot be realized unless magnetic directional data (especially inclination) are also obtained. In the case of geological baked material, orientation is frequently difficult, but it is usually possible to obtain palaeodirections from the associated igneous bodies. For historic and archaeological specimens whose orientations *in situ* may not be known, it may not be possible to obtain directions of magnetization, although every attempt should be made to do so.

(iv) The over-riding conclusion drawn from the field intensity data obtained hitherto is that the geomagnetic dipole moment fluctuates in strength within any given polarity. Direct field measurements over the past 150 years (Section 4), the variation of the historic reduced dipole moments over the past few thousand years (Fig. 6), the 30% difference between the present dipole moment and that predicted for the present time from the mean geological virtual dipole moments (Fig. 8) and the large scatter of geological VDMs (Section 8.4) all support this view. Evidence of dipole moment fluctuation supports the dynamo theory of the origin of the Earth's dipole field which predicts such fluctuation. Field intensities from transition zones also support the dynamo theory since they indicate that during polarity reversal the dipole moment decreases to zero followed by an increase in the opposite sense (Section 8.5).

10. Acknowledgments

I would like to thank Dr R. L. Wilson for making many valuable suggestions for the improvement of this paper and Dr P. Dagley for writing the computer programs used in the data analyses.

Sub-Department of Geophysics,
University of Liverpool.

1966 *August.*

References

Ade-Hall, J. M., Wilson, R. L. & Smith, P. J., 1965. The petrology, Curie points and natural magnetizations of basic lavas, *Geophys. J. R. astr. Soc.*, **9**, 323.

Akimoto, S. & Katsura, T., 1959. Magnetochemical study of the generalised titano-magnetite in volcanic rocks, *J. Geomagn. Geoelect. Kyoto*, **10**, 69.

Akimoto, S., Katsura, T. & Yoshida, M., 1957. Magnetic properties of Fe_2TiO_4–Fe_3O_4 system and their change with oxidation, *J. Geomagn. Geoelect. Kyoto*, **9,** 165.

Akimoto, S. & Kushiro, I., 1960. Natural occurrence of titanomaghemite and its relevance to the unstable magnetization of rocks, *J. Geomagn. Geoelect. Kyoto*, **11,** 94.

Allan, D. W., 1958. Reversals of the Earth's magnetic field, *Nature, Lond.*, **182,** 469.

Allan, D. W., 1962. On the behaviour of systems of coupled dynamos, *Proc. Camb. phil. Soc. math. phys. Sci.*, **58,** 671.

Blackett, P. M. S., Clegg, J. A. & Stubbs, P. H. S., 1960. An analysis of rock magnetic data, *Proc. R. Soc.*, A, **256,** 291.

Briden, J. C., 1966. Estimates of direction and intensity of the palaeomagnetic field from the Mugga Mugga Porphyry, Australia, *Geophys. J. R. astr. Soc.*, **11,** 267.

Bullard, E. C., 1955. The stability of a homopolar dynamo, *Proc. Camb. phil. Soc. math. phys. Sci.*, **51,** 744.

Bullard, E. C., Freedman, C., Gellman, H. & Nixon, J., 1950. The westward drift of the Earth's magnetic field, *Phil. Trans. R. Soc.*, A, **243,** 67.

Burlatskaya, S. P., 1961. Archaeomagnetic evidence for the Earth's magnetic field near Tiflis during the past, *Geomagn. Aeronom.*, **1,** 707.

Burlatskaya, S. P., 1962. The ancient magnetic field of the Earth, *Izv. Akad. Nauk SSSR*, p. 524.

Chevallier, R., 1925. L'aimantation des laves de l'Etna et l'orientation du champ terrestre en Sicile du XII ou XVII siécle, *Annls Phys.*, **4,** 5.

Chevallier, R., 1953. Variacion secular del campo geomagnetico, Cönsejo superior de investigaciones cientificas, *Memoria*, No. 11, Tarragona.

Clegg, J. A., Almond, M. & Stubbs, P. H. S., 1954. The remanent magnetism of some sedimentary rocks in Britain, *Phil. Mag.*, **45,** 583.

Cox, A., 1962. Analysis of present geomagnetic field for comparison with palaeomagnetic results, *J. Geomagn. Geoelect. Kyoto*, **13,** 101.

Cox, A. & Doell, R. R., 1960. Review of palaeomagnetism, *Bull. geol. Soc. Am.*, **71,** 645.

Cox, A. & Doell, R. R., 1964. Long period variations of the geomagnetic field, *Bull. seism. Soc. Am.*, **54,** 2243.

Creer, K. M., 1962. The dispersion of the geomagnetic field due to secular variation and its determination for remote times from palaeomagnetic data, *J. geophys. Res.*, **67,** 3461.

Doell, R. R. & Cox, A., 1961. Palaeomagnetism, *Adv. Geophys.*, **8,** 221.

Elsasser, W. M., 1956. Hydromagnetism, II: A review, *Am. J. Phys.*, **24,** 85.

Finch, H. F. & Leaton, B. R., 1957. The Earth's main magnetic field—epoch 1955.0, *Mon. Not. R. astr. Soc. geophys. Suppl.*, **7,** 314.

Fisher, R. A., 1953. Dispersion on a sphere, *Proc. R. Soc.*, A, **217,** 295.

Gauss, C. F., 1838. Allgemeine Theorie des Erdmagnetismus, in *Resultate Magnetische Verein.*

Gellibrand, H., 1635. *A Discourse Mathematical on the Variation of the Magneticall Needle.* London.

Gilbert, W., 1600. *De magnete.* Translation: Dover, New York, 1958.

Havard, A. D., 1965. Thesis, University of London.

Havard, A. D. & Lewis, M., 1965. Reversed partial thermo-magnetic remanences in natural and synthetic titanomagnetites, *Geophys. J. R. astr. Soc.*, **10,** 59.

Holmes, A., 1959. A revised geological time-scale, *Trans. Edinb. geol. Soc.*, **17,** 183.

Irving, E., 1960–1965. Palaeomagnetic directions and pole positions, Parts I–VII, *Geophys. J. R. astr. Soc.*, **3**, 96; **3**, 444; **5**, 72; **6**, 263; **7**, 263; (with Stott, P. M.) **8**, 249; **9**, 185.

Irving, E., 1964. *Palaeomagnetism.* Wiley, London, New York.

Irving, E., Stephenson, P. J. & Major A, 1965. Magnetism in Heard Island Rocks, *J. geophys. Res.*, **70**, 3421.

Irving, E. & Ward, M. A., 1964. A statistical model of the geomagnetic field, *Geofis. pura appl.*, **57**, 47.

Johnson, E. A., Murphy, T. & Torreson, O. W., 1948. Pre-history of the earth's magnetic field, *Terr. Magn. atmos. Elect.*, **53**, 349.

Kawai, N., Hirooka, K. & Sasajima, S., 1965. Counter-clockwise rotation of the geomagnetic dipole axis revealed in the world-wide archeo-secular variations, *Proc. Japan Acad.*, **41**, 398.

King, R. F., 1955. The remanent magnetism of artificially deposited sediments, *Mon, Not. R. astr. Soc. geophys. suppl.*, **7**, 115.

Koenigsberger, J. G., 1938. Natural residual magnetism of eruptive rocks, Parts I and 2, *Terr. Magn. atmos. Elect.*, **43**, 119.

Leaton, B. R., Malin, S. R. C. & Evans, M. J., 1965. An analytical representation, of the estimated geomagnetic field and its secular change for the epoch 1965.0, *J. Geomagn. Geoelect. Kyoto*, **17**, 187.

Lowes, F. J. & Runcorn, S. K., 1951. The analysis of the geomagnetic secular variation, *Phil. Trans. R. Soc.*, A, **243**, 525.

Mathews, J. H. & Gardner, W. K., 1963. Field reversals of the palaeomagnetic type in coupled disc dynamos, US Naval Res. Lab., Washington DC, Report 5886.

McDougall, I., 1963. Potassium–argon age measurements on dolerites from Antarctica and South Africa, *J. geophys. Res.*, **68**, 1535.

Momose, K., 1963. Studies on the variations of the Earth's magnetic field during Pliocene time, *Bull. Earthq. Res. Inst.*, **41**, 487.

Nagata, T., 1943. The natural remanent magnetism of volcanic rocks and its relation to geomagnetic phenomena, *Bull. Earthq. Res. Inst.*, **21**, 1.

Nagata, T., Akimoto, S., Uyeda, S., Momose, K. & Asami, E., 1954. Reverse magnetization of rocks and its connection with the geomagnetic field, *J. Geomagn. Geoelect. Kyoto*, **6**, 182.

Nagata, T., Arai, Y. & Momose, K., 1963. Secular variation of the geomagnetic total force during the last 5000 years, *J. geophys. Res.*, **68**, 5277.

Nagata, T., Kobayashi, K. & Schwarz, E. J., 1964. *Archeomagnetic Intensity Studies of South and Central America*, Progress Report to the US–Japan program on the *Studies on Natural Remanent Magnetism of Rocks in the Pacific Basin Area*, Scientific Report No. 3; also, 1965. *J. Geomagn. Goelect. Kyoto*, **17**, 399.

Powell, D. W., 1963. Significance of differences in magnetization along certain dolerite dykes, *Nature, Lond.*, **199**, 674.

Rikitake, T., 1958. Oscillations of a system of disc dynamos, *Proc. Camb. phil. Soc. math. phys. Sci.*, **54**, 89.

Rimbert, F., 1958. Thesis, University of Paris.

Runcorn, S. K., 1959. On the hypothesis that the mean geomagnetic field for parts of geological time has been that of a geocentric axial multipole, *J. atmos. terr. Phys.*, **14**, 167.

Sasajima, S., 1965. Geomagnetic secular variation revealed in baked earths in west Japan, *J. Geomagn. Goelect. Kyoto*, **17**, 413.

Sasajima, S. & Maenaka, K., 1964. *Intensity Studies of the Geomagnetic Field for the Past 2000 Years in West Japan*, p. 44. 1964 Progress Report of the Rock Magnetism Group in Japan, Tokyo.

Smith, P. J., 1967. The intensity of the Tertiary geomagnetic field, *Geophys. J. R. astr. Soc.*, in press.

Thellier, E. & Thellier, O., 1959. Sur l'intensité du champ magnétique terrestre dans le passé historique et géologique, *Annls Géophys.*, **15**, 285. (This review paper contains full references to the Thelliers' individual papers.)

United States Air Force Geophysics Research Directorate, 1960. *Handbook of Geophysics.* Macmillan, New York.

Van Zijl., J. S. V., 1961. Thesis, University of Witwatersrand.

Van Zijl, J. S. V., Graham, K. W. T. & Hales, A. L., 1962. The palaeomagnetism of the Stormberg lavas: II, The behaviour of the magnetic field during a reversal, *Geophys. J., R. astr. Soc.*, **7**, 169.

Vestine, E. H., Laporte, L., Cooper, C., Lange, I. & Hendrix, W. C., 1947. Description of the Earth's main magnetic field and its secular change, 1905–1945, *Publs Carnegie Instn. Washington*, No. 578.

Watanabe, N., 1958. Secular variation in the direction of geomagnetism as the standard scale for geomagnetochronology, *Nature, Lond.*, **182**, 383.

Watson, G. S. & Irving, E., 1957. Statistical methods in rock magnetism, *Mon. Not. R. astr. Soc. geophys. Suppl.*, **7**, 289.

Weaving, B., 1960. Thesis, University of London.

Wilson, R. L., 1961. Palaeomagnetism in Northern Ireland: Part 2, The thermal demagnetization of natural magnetic moments of rocks, *Geophys. J. R. astr. Soc.*, **5**, 45.

Wilson, R. L., 1966. Palaeomagnetism and rock magnetism, *Earth Sci. Rev.*, **1**, 175.

Wilson, R. L. & Watkins, N. D., 1967. Correlation of petrology and natural magnetic polarity in the Columbia Plateau basalts, *Geophys. J. R. astr. Soc.*, in press.

Yukutake, T., 1961. Archeomagnetic study on volcanic rocks in Oshima Island, Japan, *Bull. Earthq. Res. Inst.*, **39**, 467.

Yukutake, T., 1962. The westward drift of the magnetic field of the Earth, *Bull. Earthq. Res. Inst.*, **40**, 1.