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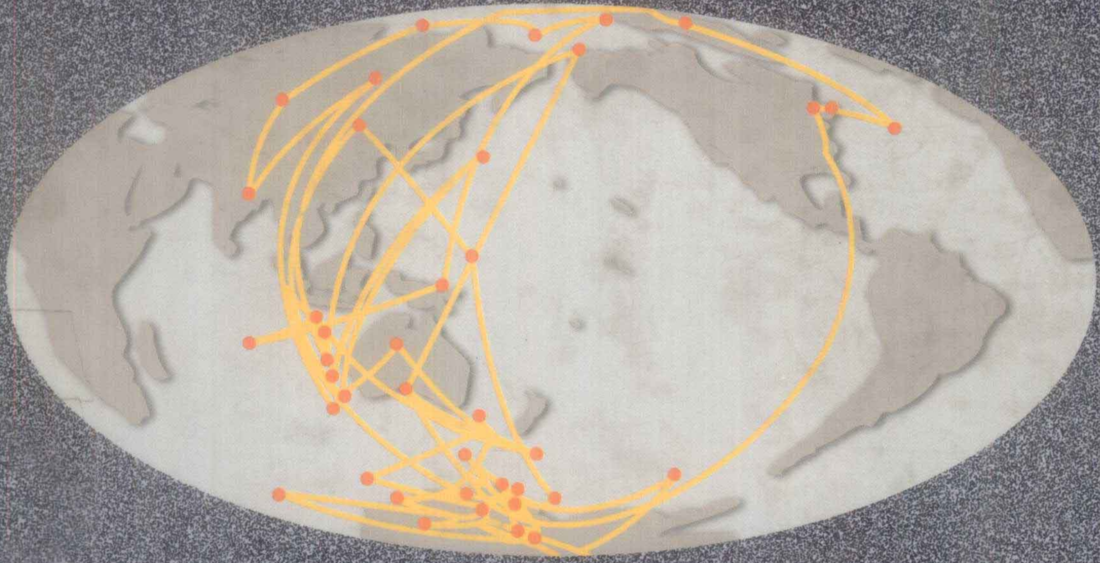

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STUDIES ON PALEOMAGNETISM AND REVERSALS OF GEOMAGNETIC FIELD IN CHINA

A Brief Introduction

This book is an attempt to summarize the most important advances in paleomagnetism that have been made in China. It includes 10 chapters: Chapters 1—3 present general surveys and some topic remarks. Chapters 4—6 describe studies on magnetostratigraphy in Quaternary. Chapters 7—8 are a comprehensive treatment of the reversals/excursions of the Earth's magnetic field with special attention to the structure and morphology of the transition field, which is significant for us to understand the mechanism of the reversal process. Chapters 9—10 provide a paleomagnetic constraints on the plate tectonic motions of the major Chinese blocks over the Phanerozoic.

This book will be of value to graduate students and researchers in geophysics, particularly in geomagnetic polarity reversals, loess magnetism and tectonophysics, as well as geologists and environmental scientists interested in magnetic signals recorded in rocks.

Contents

Chapter 1 The Earth Magnetic Field

1.1 Geomagnetic elements

The earth is a great magnet, similar to a uniformly magnetized sphere; at the strongest near the poles, the geomagnetic field is several hundred times weaker than that between the poles of a toy horseshoe magnet. The main elements of the geomagnetic field are shown in Fig. 1.1. The intensity of the whole magnetic force is denoted by F, which is called the total intensity or the total force. The magnetic force at any point O can be specified by means of the rectangular components X, Y, Z and defined as follows: X is the component along the horizontal direction in the geographical meridian; and it is reckoned as positive if northward, but negative if southward; Y is the horizontal component transverse to the geographical meridian, and it is reckoned as positive if eastward, but negative if westward; Z is the vertical component, often called the vertical intensity or the vertical force, reckoned as positive if downward and negative if upward. Another common way of specifying the magnetic force is by means of H, D, I: H is the magnitude of the horizontal component, considered positive whatever its direction and called horizontal intensity or horizontal force; D is the azimuth of the horizontal component and is reckoned as positive from the geographical north towards the east, ranged from 0° to 360°, but negative from the geographical north to the west; it is called the magnetic declination; I is the angle made by the direction of the whole magnetic force with the horizontal force (H), and it is reckoned as positive if the whole magnetic force inclines downwards, but negative if upwards; it is called the magnetic dip or magnetic inclination. D and I are measured in degrees and minutes of arc.

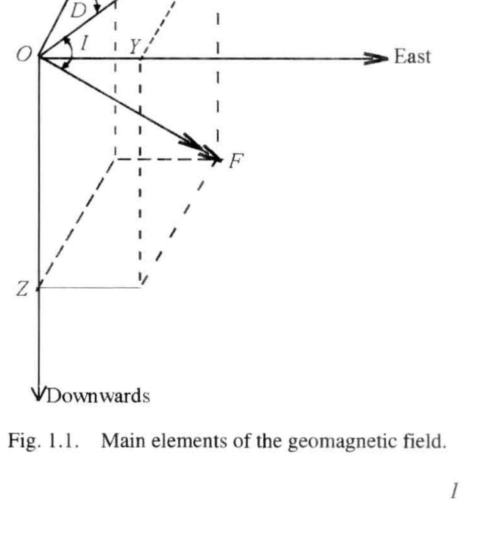

Fig. 1.1. Main elements of the geomagnetic field.

As shown in Fig. 1.1, these magnetic elements are evidently connected by the following equations:

$$H = F\cos I, \quad Z = F\sin I, \quad \tan I = Z/H, \tag{1.1}$$

$$X = H\cos D, \quad Y = H\sin D, \quad \tan D = Y/X, \tag{1.2}$$

$$F = H^2 + Z^2 = X^2 + Y^2 + Z^2. \tag{1.3}$$

Three independent elements are required in specifying F, and when the three elements are given, any of the others can be determined from the above relations.

The intensity of the earth's field is commonly expressed in gauss (Γ) using the c.g.s. electromagnetic system of units, whilst in the M.K.S.A. system the tesla (T) can be used :

$$1\Gamma = 10^{-4}\ \text{Wb m}^{-2} = 10^{-4}\ \text{T}. \tag{1.4}$$

Since geomagnetic fields are extremely small, a more convenient unit, gamma (γ) is defined as

$$1\ \text{gamma}\ (\gamma) = 10^{-5}\ \text{gauss}\ (\Gamma) = 10^{-9}\ \text{tesla (T)} = 1\ \text{nT}. \tag{1.5}$$

Geomagnetic data are derived in general surveys by both land, sea and air (airplane or satellite), in local surveys for prospecting, and in the geomagnetic observatories and in this field geomagnetic data provide the most accurate, continuous and complete information. The maximum value of the Earth's magnetic field at the surface is currently a little more than 0.7Γ and occurs in the region of the south magnetic pole.

1.2 Isomagnetic charts

The results of magnetic surveys by land and sea reduced to a common epoch can be conveniently represented graphically by charts of various kinds. One method is to draw lines having the property that a given magnetic element has a constant value along each line. The lines are called isomagnetic lines, and a chart in which the distribution of a magnetic element is thus indicated, whether for the whole or for a part of the earth, is called an isomagnetic chart. Special names are assigned to the lines and charts for certain of the elements: those referring to the magnetic elements *D*, *I* and *X*, *Y*, *Z*, *H* or *F* respectively are called isogonic (*D*) (agonic for $D = 0°$), isocline (*I*), and isodynamic. The charts may be drawn on a globe or on a plane projection of any kind.

The magnetic field observed at the surface of the earth could be produced by sources inside the earth, by sources outside the earth's surface or by electric currents crossing the earth. Early in 1839 Gauss concluded that the field was solely of internal origin; in practice the external field is not totally absent. The earth's field *F*, measured near the surface, can be described roughly as resembling that of a dipole, or that of a uniformly magnetized sphere. Consider therefore the field of a uniformly magnetized sphere whose magnetic axis runs north-south, and let *P* be any external point with distance *r* from the center *O* and θ the angle *NOR*, i.e. θ is the magnetic co-latitude (Fig. 1.2).

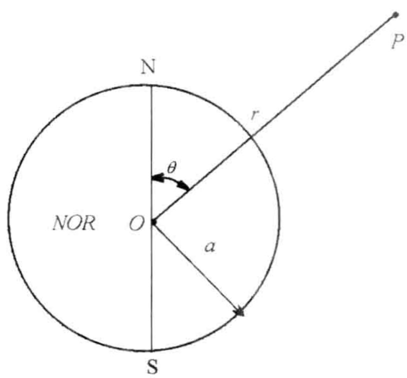

Fig. 1.2. Sketch of axial dipole.

If *m* is the magnetic moment of a geocentric dipole along the axis, the potential at *P* is

$$V = \frac{m}{4\pi}\frac{\cos\theta}{r^2}. \tag{1.6}$$

The inward radial component of force corresponding to the magnetic component *Z* is given by

$$Z = -\mu_0 \frac{\partial V}{\partial r} = \frac{\mu_0 m}{2\pi}\frac{\cos\theta}{r^3} \tag{1.7}$$

and the component at right angle to *OP* in the direction of decreasing θ, corresponding to the magnetic component *H*, is given by

$$H = -\mu_0 \frac{1}{r}\frac{\partial V}{\partial \theta} = \frac{\mu_0 m}{2\pi}\frac{\cos\theta}{r^3}, \tag{1.8}$$

where μ_0 is the permeability of free space.

The inclination *I* is given by

$$\tan I = Z/H = 2\cot\theta \tag{1.9}$$

and the magnitude of the total force by

$$F = (H^2 + Z^2)^{\frac{1}{2}} = \frac{\mu_0 m}{4\pi r^3}(1 + 3\cos^2\theta)^{\frac{1}{2}}. \tag{1.10}$$

Thus intensity measurements are a function of latitude.

Fig. 1.3 illustrates the distinction between the magnetic, geomagnetic and geographical poles and equators. The line along which the inclination is zero is called the magnetic equator, whilst the magnetic poles (or dip poles) are points where the inclination is ± 90°. The north magnetic pole is situated where $I = +90°$, and the south magnetic pole where $I = -90°$. The geomagnetic poles, i.e. the points where the axis of the geocentric dipole which best approximates the earth's field meets the surface of the earth are situated at approximately 78.8°N, 289.1°E (in northwest Greenland) and 78.8°S, 109.1°E (in Antarctic) for epoch 1980. The geomagnetic axis is thus inclined at about 11° to the geographical axis. The circle on the earth's surface coaxial with the dipole axis and midway between the geographical poles is called the geomagnetic equator, and is different from the magnetic equator which is not a circle in any case.

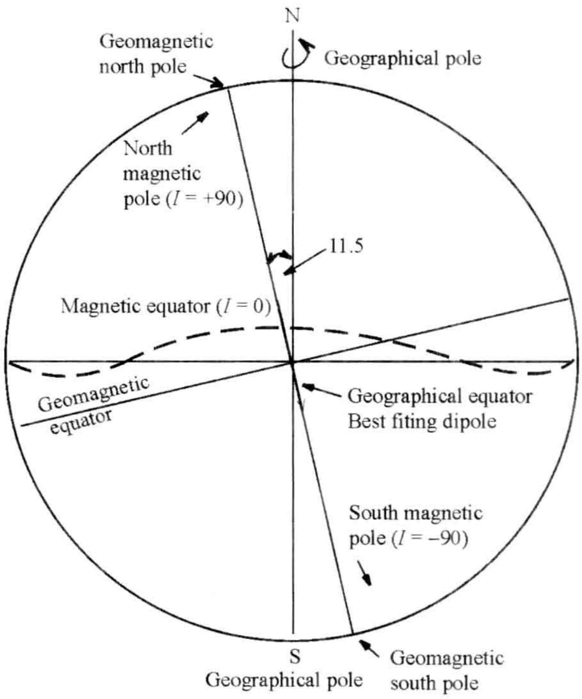

Fig. 1.3. The distinction between the magnetic, geomagnetic and geographic poles and equators.

In addition to the isomagnetic chart for each element, the rate of the secular variation in each element at any epoch (reckoned as the change per year, positive when the change is an increase) can be represented by isovariational charts or isoporic charts. It may show that (i) isopors tend to form closed ovals around certain foci of rapid annual change; (ii) the areas of rapid change are not permanent but may appear, or undergo radical changes in the form or the position, in so brief a time as one or two decades.

1.3 Composition and time spectrum of the geomagnetic field

It may be convenient to divide the earth's field F into two parts: the mean field F^0 and the variation field $\delta F(t)$, the latter being only 2.4% to less 1% of the total field. Both parts are still separated into those due to internal or external origin as expressed symbolically in the following:

$$F = F^0 + \delta F(t) = F^i + F^e + \delta F^i(t) + \delta F^e(t), \tag{1.11}$$

where F^i and F^e are 24% and 6% of F^0 in the magnitude, whilst δF^i and δF^e are 1/3 and 2/3 of $\delta F(t)$ respectively. In addition, F can be described as

$$F = F_M + F_m + F_a + F^e + \delta F(t), \tag{1.12}$$

where F_M denotes the field of a magnetic dipole, or that of a uniformly magnetized sphere; F_m is called the continental (residue) field; F_a is the field of regional and local anomalies; and F^e and $\delta F(t)$ are indicated as above. Again, if F_n is called the main field (normal or basic field), we have

$$F_n \approx F_M + F_m + F^e \approx F_M + F_m, \tag{1.13}$$

$$F^i \approx F_M + F_m + F_a \approx F_M + F_m = F_n. \tag{1.14}$$

The difference between F_n (or F) and F_M is usually called the nondipole field. With regard to F^e and F^e(t), they originate from electric currents in the ionosphere and magnetosphere. Through intensive study after the International Geophysical Year (IGY) during the last three decades our knowledge of the overhead current-systems has been considerably improved and advanced. The following current systems may be listed and evaluated: (i) Quiet-day solar diurnal current (S_q) and small lunar diurnal current (L); (ii) magnetopause (Chapman-Ferraro boundary) current; (iii) ring current the symmetric component); (iv) solar wind-magnetosphere (S_M) dynamo-generated currents: i) the cross-tail current; ii) the substorm current; iii) the polar cap current; iv) the cusp current; v) the pulsation current.

The temporal variation of the earth's field is very broad; it may be summarized in Table 1.1. Generally speaking, three variations, S (solar), L (lunar), and D (disturbed), together with slower-changed secular variation, comprise the chief changes in the earth's magnetic field. Since S, L, and D, unlike the secular variation, produce no really long-enduring change in the field, they will be called the transient magnetic variations. Of S the mean taken over extremely quiet days is referred to as S_q; on normal days or days with only minor disturbance, there is in addition the solar disturbance daily variation, S_D. The variation S_D is part of the D field generally and only in the absence of storms and substorms there will have $D \approx S_D$. In general terms, we have

$$D = \Delta F - S_q - L \text{ (on normal days)}, \tag{1.15}$$

$$D = D_{st} + S_D + D_i \text{ (during magnetic storms)}. \tag{1.16}$$

Magnetic storms typically can be divided into three phases: the initial, main and recovery phases. The initial phase may be gradual or represented by an abrupt change (sudden commencement). In addition to the S_D component related to local time, there is a component related in time to the beginning of the storm, termed the storm-time variation D_{st}. The irregular part D_i only becomes significant at high latitudes.

As shown in Table 1.1, the study of all temporal variations of geomagnetic field must be an extensive, complicated and laborious task.

Table 1.1 Time spectrum of geomagnetic field

Time variation of external sources	Period	Time variation of internal source	Period (year)
Micropulsations	Several seconds to minutes	Min. detectable period	3.7
Substorms & bays	Several minutes to hours	Non-dipole field	10^3—10^4
Magnetic storms	Several hours to days	Polarity transition	10^3—10^4
Daily variations	1 day	Main field	10^3—10^4
27-day recurrence	26—28 days	Osi. of dipole	$\geqslant 10^4$
Annual variation	1 year	Polarity interval	10^6—10^7
Sunspot-cycle	11 & 22 years	One polarity duration	10^8—10^9

1.4 Spherical harmonic analysis description of the geomagnetic field

Assuming that there is no magnetic material near the ground, the earth's magnetic field can be derived from a potential function V which satisfies Laplace's equation and can thus be represented as a series of spherical harmonics

$$\dot{V} = \frac{a}{\mu_0} \sum_{n=1}^{\infty} \sum_{m=0}^{n} P_n^m(\cos\theta) \left\{ \left[C_n^m \left(\frac{r}{a} \right)^n + (1 - C_n^m \left(\frac{a}{r} \right)^{n+1} \right] A_n^m \cos m\phi + \left[S_n^m \left(\frac{r}{a} \right)^n + (1 - S_n^m \left(\frac{a}{r} \right)^{n+1} \right] B_n^m \sin m\phi \right\}, \tag{1.17}$$

where θ, ϕ are the co-ordinates of the magnetic colatitude and longitude, r is the distance from the center of the earth, a is the earth's mean radius (6371.2 km), and $P_n^m(\cos\theta)$ is the spherical harmonic function of Schmidt of degree n and order m. Written in this form the coefficients A_n^m and B_n^m have the dimensions of magnetic field, C_n^m and S_n^m are numbers lying between 0 and 1, and represent the fraction of the potential associated with sources of external origin ($r>a$). The coefficients $(1-C_n^m)$ and $(1-S_n^m)$ indicate the fraction of the potential associated with sources of internal origin ($r<a$). The Gauss coefficients g_n^m and h_n^m are to be evaluated (see eq. (1.23) below) and in SI units, these are traditionally given in nanotesla. Hence in eq. (1.17) the factor μ_0 (permeability of free space) is included, so that the coefficients in SI units have the same numerical values as in the c.g.s electromagnetic system of units. Note that the surface harmonics $P_n^m(\cos\theta)\sin m\phi$ (or $\cos m\phi$) vanish along (n−m) circles of latitudes, corresponding to the zero of P_n^m; and also because of the factor $\cos m\phi$ or $\sin m\phi$, along 2 m meridians at equal intervals π/m. These zero-lines divide the surface of the sphere into four-cornered (or, at the poles, three-cornered) regions. Consequently, for $n>m>0$, the surface harmonics are called tesseral or non-zonal surface harmonics. When $m=0$, the functions are called zonal surface harmonics; and when $m=n$, they are called sectorial surface harmonics. There is no term with $n=0$, which would correspond to a magnetic monopole within the earth.

The potential V cannot be measured directly; what can be determined is the three components of force $X = (\mu_0/r)(\partial V/\partial\theta)$ (horizontal, northward), $Y = (-\mu_0/r\sin\theta)(\partial V/\partial\phi)$ (horizontal, eastward) and $Z = \mu_0(\partial V/\partial r)$ (vertical, downward) at the earth's surface, $r = a$. Z (at $r=a$) may be expanded as a series of spherical hamonics:

$$Z = \frac{\mu_0 \partial V}{\partial r} = \sum_{n=1}^{\infty}\sum_{m=0}^{n} P_n^m(\cos\theta)(\alpha_n^m \cos m\phi + \beta_n^m \sin m\phi) \tag{1.18}$$

and the coefficients α_n^m, β_n^m determined from the observed values of Z.

By differentiating eq. (1.17) with respect to r and then writing $r=a$, we have

$$Z = \frac{\mu_0 \partial V}{\partial r} = \sum_{n=1}^{\infty}\sum_{m=0}^{n} P_n^m(\cos\theta)\left\{\left[nc_n^m (n+1)(1-c_n^m)\right]A_n^m \cos m\phi \right.$$
$$\left. + \left[nS_n^m - (n+1)(1-S_n^m)\right]B_n^m \sin m\phi\right\}. \tag{1.19}$$

The coefficients of each separate hamonic term for each n and m must be equal in the two expressions of Z given by eqs. (1.18) and (1.19). Hence

$$\alpha_n^m = \left[nc_n^m - (n+1)(1-c_n^m)A_n^m\right],$$

$$\beta_n^m = \left[ns_n^m - (n+1)(1-s_n^m)B_n^m\right]. \tag{1.20}$$

Again from an analysis of the observed values of X and Y, the coefficients in the following two expansions derived from eq. (1.17) may be obtained:

$$Y_{r=\alpha} = \left(\frac{-\mu_0}{r\sin\theta}\frac{\partial V}{\partial\phi}\right)_{r=\alpha}$$
$$= \frac{1}{\sin\theta}\sum_{n=1}^{\infty}\sum_{m=0}^{n} P_n^m(\cos\theta)(mA_n^m \sin m\phi - mB_n^m \cos m\phi), \tag{1.21}$$

$$Y_{r=\alpha} = \left(\frac{-\mu_0}{r}\frac{\partial V}{\partial\phi}\right)_{r=\alpha}$$
$$= \sum_{n=1}^{\infty}\sum_{m=0}^{n} \frac{\partial}{\partial\theta} P_n^m(\cos\theta)(A_n^m \sin m\phi - mB_n^m \cos m\phi). \tag{1.22}$$

By using measurements of the three components of the earth's magnetic field and by truncating the series of some value n, the relative importance of the internal (S_n^m and C_n^m) versus external ($1-S_n^m$, $1-C_n^m$) sources can be determined. Also, the coefficients determined by this way are not invariant in time. As seen in eq. (1.20), C_n^m and S_n^m can be determined from a knowledge of the coefficients A_n^m, B_n^m, α_n^m and β_n^m. Gauss found from data available at that time (1839) that $C_n^m = S_n^m = 0$, i.e. the source of the earth's magnetic field is entirely internal. The coefficients of the field of internal origin are

$$g_n^m = (1-c_n^m)A_n^m \quad h_n^m = (1-s_n^m)B_n^m, \tag{1.23}$$

and are known as Gauss coefficients. If the external field is negligible, eq. (1.23) is reduced to

$g_n^m = A_n^m$, and $h_n^m = B_n^m$.

Thus, in considering the main field, the external sources may be ignored and the potential is written as

$$V = \frac{\alpha}{\mu_0}\sum_{n=1}^{\infty}\sum_{m=0}^{n}\left(\frac{a}{r}\right)^{n+1} P_n^m(\cos\theta)(g_n^m \cos m\phi + h_n^m \sin m\phi). \tag{1.24}$$

Note that g_n^m is zero because there is no monopole contribution to the potential. The g_1^0 term is given by

$$\frac{g_1^0}{\mu_0} P_1^0 \frac{a^3}{r^2} = \left(\frac{g_1^0 4\pi\, a^3}{\mu_0}\right)\frac{\cos\theta}{4\pi\, r^2}. \tag{1.25}$$

This term is the potential associated with a geocentric dipole with strength $\left(\frac{g_1^0 4\pi\, a^3}{\mu_0}\right)$ oriented along the z axis. The g_1^1 term is $\left(\frac{g_1^0 4\pi\, a^3}{\mu_0}\right)\left(\frac{\cos\phi\sin\theta}{4\pi\, r^2}\right)$; if γ is the angle between x and r axes,

$$\cos\gamma = \sin\theta\cos\phi,$$

so that the g_1^1 term corresponds to a geocentric dipole oriented along the x direction. Similarly, it is found that the h_1^1 term $\left(h_1^1 \frac{4\pi\, a^3}{\mu_0}\right)\left(\frac{\cos\phi\sin\theta}{4\pi\, r^2}\right)$ corresponds to a geocentric dipole oriented in the y direction. The magnitude and direction of the geocentric dipole, P can be determined by the usual vector addition:

$$P = \frac{4\pi\, a^3}{\mu_0}(g_1^{0^2} + g_1^{1^2} + h_1^{1^2})^{1/2} \tag{1.26}$$

and it is found to be tilted at roughly 11.5° to the rotation axis. The terms involving $n = 2$ (r^3 in the potential) represent a geocentric quadrupole, the terms $n = 3$ (r^4 in the potential) represent the geocentric octrapole and so forth. The g_1^1, g_1^1 and h_1^1 terms collectively represent the dipole field, while the remaining terms collectively represent the non-dipole field. Again, it is sometimes convenient to represent the earth's field as a dipole displaced from the center with a distance d along the z axis. The potential of such an offset dipole is proportional to

$$\cos\theta(r^2 + d^2 - 2rd\cos\theta)^{-1} = \frac{\cos\theta}{r^2}\left(1 - \frac{2d}{r}\cos\theta + \frac{d^2}{r^2}\right)^{-1}. \tag{1.27}$$

Expanding in a Taylor series for $d<r$, it can be seen that the offset dipole is equivalent to a dipole plus a series of higher degree multipole fields at the earth's center. Further, it may be shown that

$$\frac{g_2^0}{2g_1^0} = \frac{d}{a}, \quad \frac{g_3^0}{g_2^0} = \frac{3d}{2a}. \tag{1.28}$$

In practice the series expansion of magnetic potential V in eq. (1.24) is truncated at a maximum N of m and n, usually in the range of 8—15. The total number of coefficients to degree N is $(N+1)^2 - 1$, i.e. at $N = 8$, then 80 coefficients; at $N = 10$, then 120 coefficients. In 1968 the International Association of Geomagnetism and Aeronomy (IAGA) adopted an International Geomagnetic Reference Field (IGRF) describing the main geomagnetic field in 1965 by means of 80 spherical harmonic coefficients ($N = 8$). An additional set of 80 coefficients describing the secular variation (SV) was included for use in extending the main-field model in time both backward and forward. Since then the main-field models are defined for successive 5-year epochs by IAGA Working Group V-8.The IGRF now consists of a new set of IGRF models at 5-year epochs from 1900.0 to 1940.0, the existing DGRF (Definite GRF) models at 5-year epochs from 1945.0 to 1985.0, a new DGRF 1990 model that replaces IGRF 1990, and a new IGRF 1995 model that includes SV terms for forward continuation of the 1995 field to the year 2000. Coefficients for dates between the 5-year epochs are obtained by linear interpolation between the corresponding coefficients for the neighboring 5-year epochs.

1.5 Origin of the earth's magnetic field

Early at the beginning of this century Einstein described the origin of the earth's magnetic field as one of the five important problems in physics. Since then there has been much speculation on it and its secular variation and many possible sources have been suggested, most of which have proved to be inadequate. These hypotheses include permanent magnetization, residual electric currents set up in the earth's interior early in its history, thermoelectric effect, geomagnetic effect, Hall effect, galvanomagnetic effect, electromagnetic induction by magnetic storms, and differential rotation effect. It is doubtful that they are responsible for the earth's magnetic field, although some may perhaps contribute a little bit to it. The only possible means seems to be some forms of electromagnetic induction, electric currents flowing in the earth's fluid, electrically conducting outer core, i.e. the so-called dynamo problem. In the development of models for a self-sustaining dynamo they include several stages: (i) the homopolar dynamo (disk dynamo) approach; (ii) kinematics dynamo models; (iii) hydrodynamic approach; (iv) magnetohydromagnetic (MHD) instability models; and (v) turbulent dynamo models or mean-field electromagnetic models. Here we only describe briefly some physics involved, together with mean field electrodynamics.

The dynamo problem involves the solution of a highly complicated system of coupled partial differential equations (electrodynamics, hydrodynamics and thermodynamics), along with the appropriate boundary and initial conditions. In geophysical and astrophysical problems the displacement current and all purely electrostatic effects are negligible, i.e. $\partial D/\partial t = 0$. Thus the electromagnetic field equations are the usual Maxwell equations:

$$\nabla \times E = -\partial B / \partial t, \tag{1.29}$$

$$\nabla \times B = \mu_0 j, \tag{1.30}$$

$$\nabla B = 0, \tag{1.31}$$

where B and E are the magnetic and electric fields respectively, μ_0 the magnetic permeability (in earth's core) and j the electric current density. The electromotive forces which give rise to j are due both to electric charge and to motional induction so that the total current j is given by

$$j = \sigma(E + U \times B), \tag{1.32}$$

where σ is the electrical conductivity and U the fluid velocity. Then taking the curl of both sides of eq. (1.30) and using eqs. (1.29)—(1.32), E can be eliminated, and we finally obtain the magnetic induction equation:

$$\partial B / \partial t = \nu_m \nabla^2 B + \nabla \times (U \times B), \tag{1.33}$$

where $L_m = (\sigma\mu_0)^{-1}$ is the magnetic diffusivity or magnetic viscosity. When $U = 0$, this equation is reduced to the vector diffusion equation for B. The term $\nu_m \nabla^2 B$ represents the tendency for the field to decay through ohmic dissipation by the electric current supporting the field. Dimensional arguments indicate a decay time of the order L^2/ν_m, where L is a length representation of the dimensions of the region in which current flows. As an alternative limiting case, suppose that the material is in motion but has negligible electrical resistance, then eq. (1.33) becomes

$$\partial B / \partial t = \nabla \times (U \times B). \tag{1.34}$$

This equation is identical to that satisfied by the vorticity in the hydrodynamic theory of the flow of a non-viscous fluid. It is shown that vortex lines move with the fluid. Thus eq. (1.34) implies that the field changes are the same as if the magnetic lines of force were "frozen" into the material. The term $\nabla \times (U \times B)$ in eq. (1.33) gives the interaction of the velocity field with the magnetic field. This interaction can cause breakdown of the magnetic field or build-up of the magnetic field, depending on the nature of the velocity field. Again, when neither term on the right-hand side of eq. (1.33) is negligible, both the above effects are observed, i.e. the lines of force tend to be carried about with the moving fluid and at the same time leak through it. If L, T, V represent the order of magnitude of a length, time and velocity respectively, transport dominates leak if $LV >> \nu_m$, or by analogy, when a magnetic Reynolds number $R_m = LV/\nu_m >> 1$. In the case of a dynamo operating in the earth's core, it has been estimated that $R_m \geqslant 10$.

In kinematics dynamo models, U is specified in some reasonable way along with some initial magnetic field B. The problem then is whether this U field can support a B field that does not decay to zero as time goes to infinity. In hydrodynamic dynamo problems, both the magnetic induction eq. (1.33) and the Navier-Stokes equation (by neglecting $U \times B$ term) should be simultaneously solved. In the case of MHD dynamo approach, however, the magnetic (Lorentz) forces are assumed to be very strong. This leads to the magnetohydromagnetic instability models, or M.A.C. waves, pioneered by Breginskiy. It is easy to see that dynamos can produce a field in either direction. The induction eq. (1.33) is linear and homogeneous in the field and the Navier-Stokes equation (here not shown) inhomogeneous and quadratic. Thus, if a given velocity field supports either a steady or a varying magnetic field, it will also support the reversed field and the same forces will drive it. This, however, merely shows that the reversed field satisfies the equation, it does not prove that reversal will take place.

Recently, several attempts were made to produce models in which turbulent (i.e. random and small-scale) velocities might act as dynamos. This has been called mean field-electrodynamics or turbulent dynamics. In mean field dynamo models the velocity U and magnetic field B are separately represented as the sum of a statistical average and a fluctuating part i.e. $U = U_0 + U'$, $B = B_0 + B'$, and $\langle U' \rangle = \langle B' \rangle = 0$. The induction equation may then be divided into its mean and fluctuating parts:

$$\frac{\partial B_0}{\partial t} = \nabla \times (U_0 \times B_0) + \nabla \times \varepsilon + \nu_m \nabla^2 B_0, \tag{1.35}$$

$$\frac{\partial B_0}{\partial t} = \nabla \times (U_0 \times B') + \nabla \times (U' \times B_0) + \nabla \times G + \nu_m \nabla^2 B', \tag{1.36}$$

where $\varepsilon = \langle U' \times B' \rangle$ and $G = U' \times B' - \langle U' \times B' \rangle$. As seen from eq. (1.35) ε may be regarded as an extra mean electric force arising from the interaction of the turbulent motion and the magnetic field. If the velocity field is isotropic, it can be shown that

$$\varepsilon = \alpha B_0 - \beta \nabla \times B_0, \tag{1.37}$$

where α and β depend on the local structure of the velocity field. The induction eq. (1.35) satisfied by the mean field is

$$\partial B_0 / \partial t = \nabla \times (\alpha B_0 + U_0 \times B_0) + (\nu_m + \beta) \nabla^2 B_9. \tag{1.38}$$

The term ∂B_0 represents an electric field parallel to B_0. This is the so-called α-effect; α need not be a scalar, but can be a second order tensor. The quantity β is an eddy diffusivity, similar in its effect to the ohmic diffusivity. The α-effect is essentially a mechanism by which turbulent energy is converted to electrical energy. All turbulent models depend to some degree on this effect to magnify the magnetic field. There are two of the most fashionable dynamo types, the α^2 and $\alpha\omega$ dynamos. Briefly, in an α^2 dynamo the effect generates the poloidal field from the toroidal field and generates the toroidal field from the poloidal field. The toroidal field has lines of force that lie on spherical surfaces and has no component external to the core. The poloidal field has a radial component in general and joins continuously with the external observed field. In an $\alpha\omega$ dynamo the α-effect is used in conjunction with a large scale shear flow (the ω effect); an α-effect from cyclone turbulence generates the poloidal field from the toroidal field, and differential rotation creates the toroidal field from the poloidal field thereby completing the cycle.

1.6 Earth's interior and dynamo theory

The properties of the earth's interior, particularly the core, are important for evaluating theories dealing with the earth's magnetic field. Geophysicists have done a remarkable work in obtaining information about the earth's interior, considering the difficulties involved. The inversion of seismic data to obtain the compression and shear-wave velocity distribution throughout the earth's interior has probably provided the best information on the structure and composition of the core. A parameterized earth's model PREM (Preliminary Reference Earth Model) developed by Dziewonski and Anderson (1981) based on astronomical geodetic data, free oscillation and long-period surface wave data, and body wave data is a good first order estimation of the earth's seismic structure. Table 1.2 gives some of the core parameters taken from this model. These parameters are quite adequate for dynamo theories, with the exception that the magnitude of the sharp change in density between the inner and outer core is still not well known. The properties of