INTRODUCTION TO GEOMAGNETISM

The primary objective of paleomagnetic research is to obtain a record of past configurations of the geomagnetic field. Thus, understanding paleomagnetism demands some basic knowledge of the geomagnetic field. In this chapter, we begin by defining common terms used in geomagnetism and paleomagnetism. With this foothold, we describe spatial variations of the present geomagnetic field over the globe and time variations of the recent geomagnetic field. Even this elementary treatment of geomagnetism provides the essential information required for discussing magnetic properties of rocks, as we will do in the succeeding chapters. This chapter includes an appendix dealing with systems of units used in geomagnetism and paleomagnetism and describing the system of units used in this book.

SOME BASIC DEFINITIONS

New subjects always require basic definitions. Initially, we need to define magnetic moment, $\boldsymbol{M}$; magnetization, $\boldsymbol{J}$; magnetic field, $\boldsymbol{H}$; and magnetic susceptibility, χ. Generally, students find developing an intuitive feel for magnetism and magnetic fields more difficult than for electrical phenomena. Perhaps this is due to the fundamental observation that isolated magnetic charges (*monopoles*) do not exist, at least for anything more than a fraction of a second. The smallest unit of magnetic charge is the magnetic dipole, and even this multipole combination of magnetic charges is more a mathematical convenience than a physical reality.

The *magnetic dipole moment* or more simply the *magnetic moment*, $\boldsymbol{M}$, can be defined by referring either to a pair of magnetic charges (Figure 1.1a) or to a loop of electrical current (Figure 1.1b). For the pair of magnetic charges, the magnitude of charge is m, and an infinitesimal distance vector, $\boldsymbol{l}$, separates the plus charge from the minus charge. The magnetic moment, $\boldsymbol{M}$, is

$$\boldsymbol{M} = m\,\boldsymbol{l} \tag{1.1}$$

For a loop with area A carrying electrical current I, the magnetic moment is

$$\boldsymbol{M} = I\,A\,\boldsymbol{n} \tag{1.2}$$

where $\boldsymbol{n}$ is the vector of unit length perpendicular to the plane of the loop. The proper direction of $\boldsymbol{n}$ (and therefore $\boldsymbol{M}$) is given by the *right-hand rule*. (Curl the fingers of your right hand in the direction of current flow and your right thumb points in the proper direction of the unit normal, $\boldsymbol{n}$.) The current loop definition of magnetic moment is basic in that all magnetic moments are caused by electrical currents. However, in some instances, it is convenient to imagine magnetic moments constructed from pairs of magnetic charges.

Magnetic force field or *magnetic field*, $\boldsymbol{H}$, in a region is defined as the force experienced by a unit positive magnetic charge placed in that region. However, this definition implies an experiment that cannot actually be performed. An experiment that you can perform (and probably have) is to observe the aligning torque on a magnetic dipole moment placed in a magnetic field (Figure 1.1c). The aligning torque, $\boldsymbol{\Gamma}$, is given by the vector cross product:

$$\boldsymbol{\Gamma} = \boldsymbol{M} \times \boldsymbol{H} = MH \sin\theta \hat{\boldsymbol{\Gamma}} \tag{1.3}$$

where θ is the angle between $\boldsymbol{M}$ and $\boldsymbol{H}$ as in Figure 1.1c and $\hat{\boldsymbol{\Gamma}}$ is the unit vector parallel to $\boldsymbol{\Gamma}$ in Figure 1.1c.

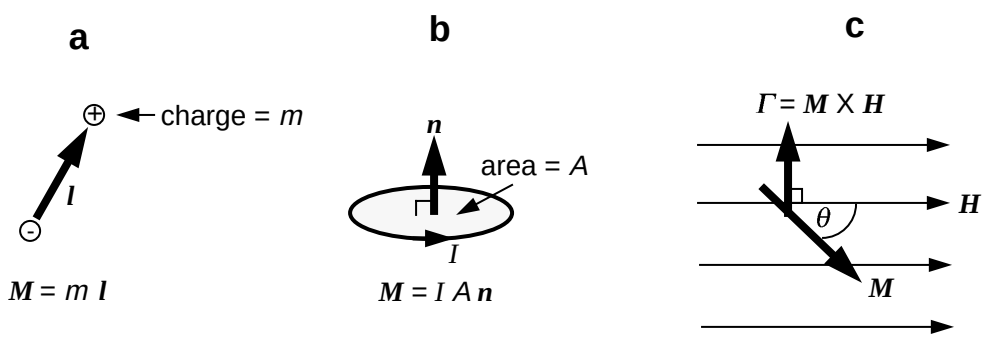

Figure 1.1 (**a**) A magnetic dipole constructed from a pair of magnetic charges. The magnetic charge of the plus charge is m; the magnetic charge of the minus charge is $-m$; the distance vector from the minus charge to the plus charge is $\boldsymbol{l}$. (**b**) A magnetic dipole constructed from a circular loop of electrical current. The electrical current in the circular loop is I; the area of the loop is A; the unit normal vector $\boldsymbol{n}$ is perpendicular to the plane of the loop. (**c**) Diagram illustrating the torque $\boldsymbol{\Gamma}$ on magnetic moment $\boldsymbol{M}$, which is placed within magnetic field $\boldsymbol{H}$. The angle between $\boldsymbol{M}$ and $\boldsymbol{H}$ is θ; $\boldsymbol{\Gamma}$ is perpendicular to the plane containing $\boldsymbol{M}$ and $\boldsymbol{H}$.

A magnetic moment that is free to rotate will align with the magnetic field. A compass needle has such a magnetic moment that aligns with the horizontal component of the geomagnetic field, yielding determination of magnetic azimuth.

The energy of alignment of magnetic moments with magnetic fields will be encountered often in the development of rock magnetism. This potential energy can be expressed by the vector dot product

$$E = -\boldsymbol{M} \cdot \boldsymbol{H} = -MH\cos\theta \tag{1.4}$$

The negative sign in this expression is required so that the minimum energy configuration is achieved when $\boldsymbol{M}$ is parallel to $\boldsymbol{H}$.

The magnetic intensity, or *magnetization*, $\boldsymbol{J}$, of a material is the net magnetic dipole moment per unit volume. To compute the magnetization of a particular volume, the vector sum of magnetic moments is divided by the volume enclosing those magnetic moments:

$$\boldsymbol{J} = \frac{\sum_i \boldsymbol{M}_i}{\text{volume}} \tag{1.5}$$

where $\boldsymbol{M}_i$ is the constituent magnetic moment.

There are basically two types of magnetization: induced magnetization and remanent magnetization. When a material is exposed to a magnetic field $\boldsymbol{H}$, it acquires an *induced magnetization*, $\boldsymbol{J}_i$. These quantities are related through the magnetic susceptibility, χ:

$$\boldsymbol{J}_i = \chi\, \boldsymbol{H} \tag{1.6}$$

Thus, *magnetic susceptibility*, χ, can be regarded as the *magnetizability* of a substance. The above expression uses a scalar for susceptibility, implying that $\boldsymbol{J}_i$ is parallel $\boldsymbol{H}$. However, some materials display *magnetic anisotropy,* wherein $\boldsymbol{J}_i$ is not parallel to $\boldsymbol{H}$. For an anisotropic substance, a magnetic field applied in a direction x will in general induce a magnetization not only in direction x, but also in directions y and z. For anisotropic substances, magnetic susceptibility is expressed as a tensor, $\boldsymbol{\chi}$, requiring a 3×3 matrix for full description.

In addition to the induced magnetization resulting from the action of present magnetic fields, a material may also possess a *remanent magnetization*, $\boldsymbol{J}_r$. This remanent magnetization is a recording of past magnetic fields that have acted on the material. Much of the coming chapters involves understanding how rocks

can acquire and retain a remanent magnetization that records the geomagnetic field direction at the time of rock formation.

In paleomagnetism, the direction of a vector such as the surface geomagnetic field is usually defined by the angles shown in Figure 1.2. The vertical component, H_v, of the surface geomagnetic field, ***H***, is defined as positive downwards and is given by

$$H_v = H \sin I \tag{1.7}$$

where H is the magnitude of ***H*** and I is the *inclination* of ***H*** from horizontal, ranging from –90° to +90° and defined as positive downward. The horizontal component, H_h, is given by

$$H_h = H \cos I \tag{1.8}$$

and geographic north and east components are respectively,

$$H_N = H \cos I \cos D \tag{1.9}$$

$$H_E = H \cos I \sin D \tag{1.10}$$

where D is *declination*, the angle from geographic north to horizontal component, ranging from 0° to 360°, positive clockwise. Determination of I and D completely describes the direction of the geomagnetic field. If the components are known, the total intensity of the field is given by

$$H = \sqrt{H_N^2 + H_E^2 + H_V^2} \tag{1.11}$$

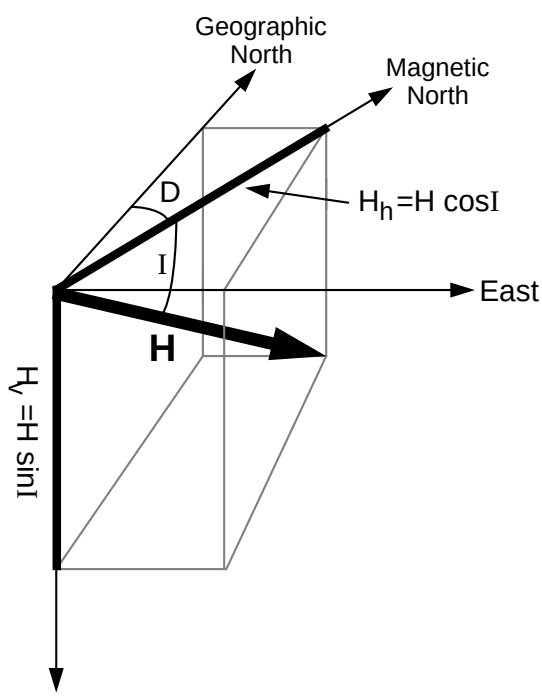

Figure 1.2 Description of the direction of the magnetic field. The total magnetic field vector ***H*** can be broken into (1) a vertical component, $H_v = H \sin I$ and (2) a horizontal component, $H_h = H \cos I$; inclination, I, is the vertical angle (= dip) between the horizontal and ***H***; declination, D, is the azimuthal angle between the horizontal component of ***H*** (= H_h) and geographic north; the component of the magnetic field in the geographic north direction is $H \cos I \cos D$; the east component is $H \cos I \sin D$. Redrawn after McElhinny (1973).

GEOCENTRIC AXIAL DIPOLE MODEL

A concept that is central to many principles of paleomagnetism is that of the *geocentric axial dipole* (GAD), shown in Figure 1.3. In this model, the magnetic field produced by a single magnetic dipole at the center of the Earth and aligned with the rotation axis is considered. The GAD field has the following properties, which are derived in detail in the appendix on derivations:

$$H_h = \frac{M \cos \lambda}{r_e^3} \tag{1.12}$$

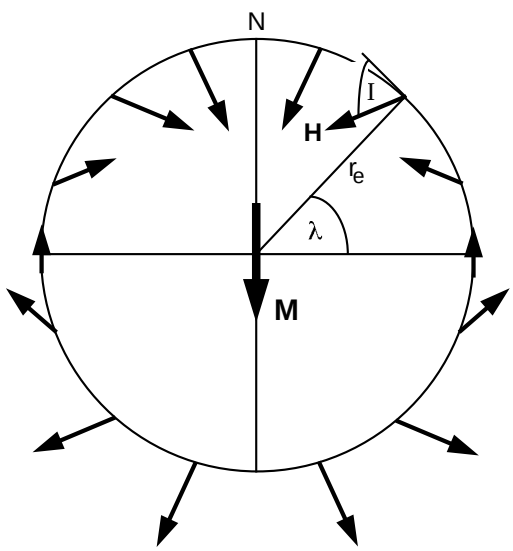

Figure 1.3 Geocentric axial dipole model. Magnetic dipole ***M*** is placed at the center of the Earth and aligned with the rotation axis; the geographic latitude is λ; the mean Earth radius is r_e; the magnetic field directions at the Earth's surface produced by the geocentric axial dipole are schematically shown; inclination, *I*, is shown for one location; N is the north geographic pole. Redrawn after McElhinny (1973).

$$H_v = \frac{2M\sin\lambda}{r_e^3} \tag{1.13}$$

$$H = \frac{M}{r_e^3}\sqrt{1+3\sin^2\lambda} \tag{1.14}$$

where *M* is the dipole moment of the geocentric axial dipole; λ is the geographic latitude, ranging from –90° at the south geographic pole to +90° at the the north geographic pole; and r_e is the mean Earth radius.

The lengths of the arrows in Figure 1.3 schematically show the factor of 2 increase in magnetic field strength from equator to poles. The inclination of the field can be determined by

$$\tan I = \left(\frac{H_v}{H_h}\right) = \left(\frac{2\sin\lambda}{\cos\lambda}\right) = 2\tan\lambda \tag{1.15}$$

and *I* increases from –90° at the geographic south pole to +90° at the geographic north pole. Lines of equal *I* are parallel to lines of latitude and are simply related through Equation (1.15), which is a cornerstone of many paleomagnetic methods and is often referred to as "the dipole equation." This relationship between *I* and λ will be essential to understanding many paleogeographic and tectonic applications of paleomagnetism. For a GAD, *D* = 0° everywhere.

THE PRESENT GEOMAGNETIC FIELD

The morphology of the present geomagnetic field is best illustrated with *isomagnetic charts*, which show some chosen property of the field on a world map. Figure 1.4 is an *isoclinic chart* showing contours of equal inclination of the surface geomagnetic field. The *geomagnetic equator* (line of *I* = 0°) is close to the geographic equator, and inclinations are positive in the northern hemisphere and negative in the southern hemisphere. This is roughly the morphology of a geocentric axial dipole field, but there are obvious departures from that simplest configuration. The *magnetic poles* (locations where *I* = ±90°; also called *dip poles*) are not at the geographic poles as expected for a GAD field, and the magnetic equator wavers about the geographic equator. The present geomagnetic field is obviously more complex than a GAD field, and the GAD model must be modified to better describe the field.

An *inclined geocentric dipole* is inclined to the rotation axis, as shown in Figure 1.5. The inclined geocentric dipole that best describes the present geomagnetic field has an angle of ~11.5° with the rotation axis. The poles of the best-fitting inclined geocentric dipole are the *geomagnetic poles*, which are points on

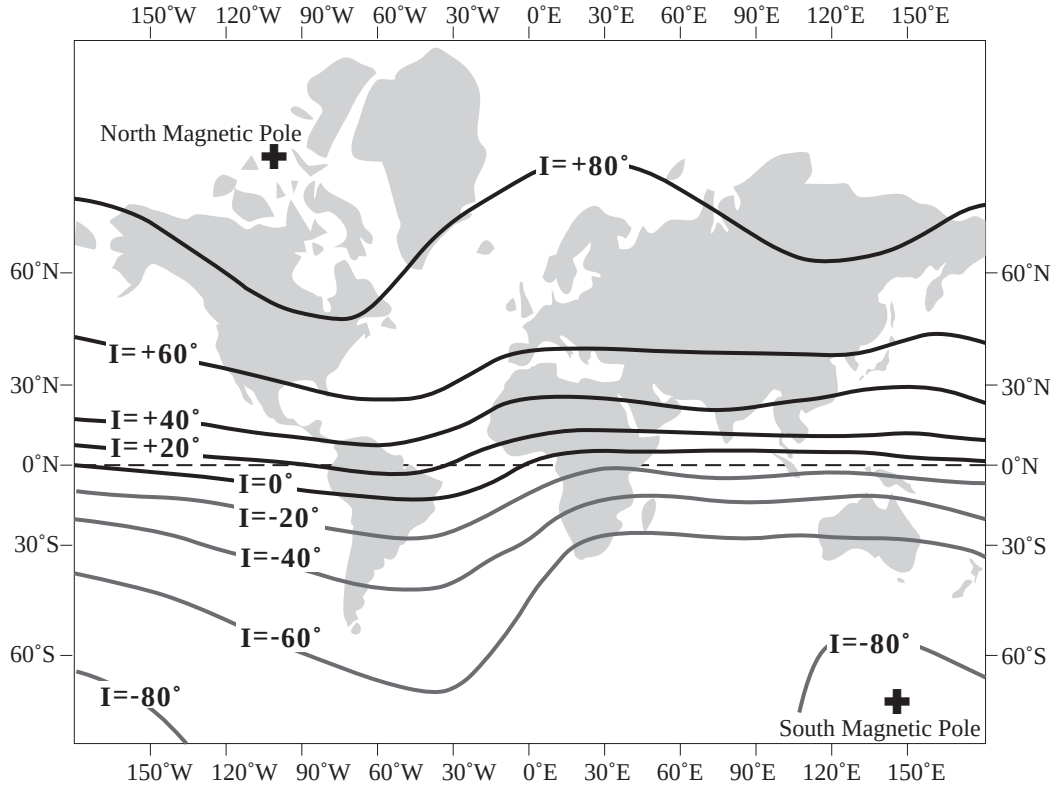

Figure 1.4 Isoclinic chart of the Earth's magnetic field for 1945. Contours are lines of equal inclination of the geomagnetic field; the locations of the magnetic poles are indicated by plus signs; Mercator map projection. Redrawn after McElhinny (1973).

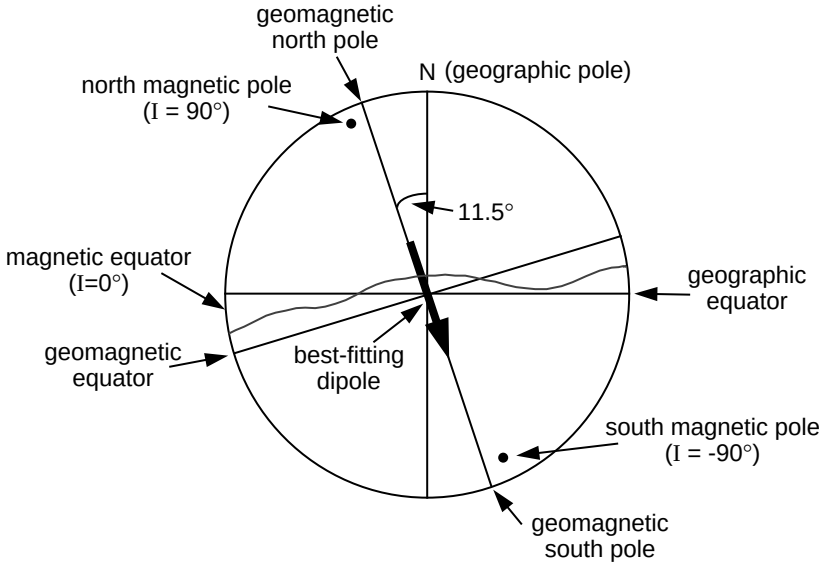

Figure 1.5 Inclined geocentric dipole model. The best-fitting inclined geocentric dipole is shown in meridional cross section through the Earth in the plane of the geocentric dipole; distinctions between magnetic poles and geomagnetic poles are illustrated; a schematic comparison of geomagnetic equator and magnetic equator is also shown. Redrawn after McElhinny (1973).

the surface where extensions of the inclined dipole intersect the Earth's surface. If the geomagnetic field were exactly that of an inclined geocentric dipole, then the geomagnetic poles would exactly coincide with the dip poles. The fact that these poles do not coincide indicates that the geomagnetic field is more complicated than can be explained by a dipole at the Earth's center. Although the inclined geocentric dipole accounts for ~90% of the surface field, the amount remaining is significant.

It is possible to further refine the fit of a single dipole to the geomagnetic field by relaxing the geocentric constraint, allowing the dipole to be positioned to best fit the field. This best-fitting dipole is the *eccentric dipole*, which describes the field only marginally better than the inclined geocentric dipole. For the present geomagnetic field, the best-fitting eccentric dipole is positioned about 500 km (~8% of Earth radius) from the geocenter, toward the northwestern portion of the Pacific Basin.

The ability of the best-fitting eccentric dipole to describe the geomagnetic field depends on location on the Earth's surface. At some locations, the best-fitting eccentric dipole perfectly describes the geomagnetic field. But at other locations, up to 20% of the surface geomagnetic field cannot be described by even the best-fitting dipole. This discrepancy indicates the presence of a higher-order portion of the geomagnetic field, which is called the *nondipole field*. This nondipole field is determined by subtracting the best-fitting dipolar field from the observed geomagnetic field. A plot of the nondipole field (for the year 1945) is shown in Figure 1.6, where the contours give the vertical component of the nondipole field and the arrows show the magnitude and direction of the horizontal component of the nondipole field.

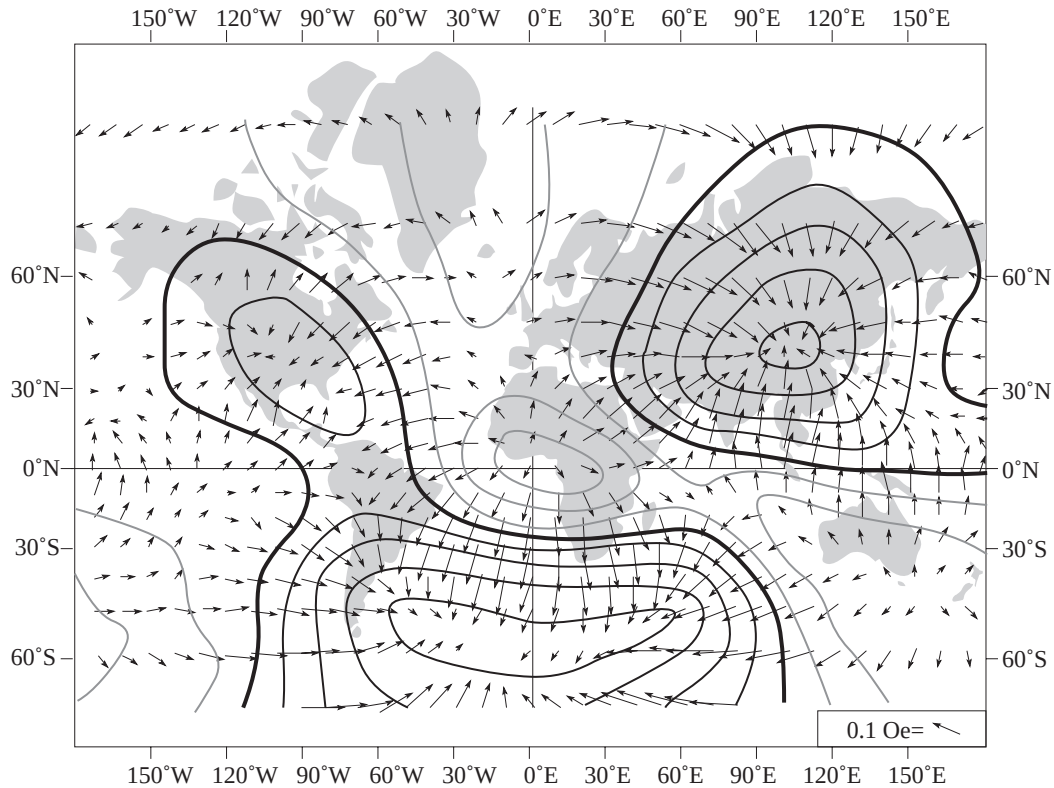

Figure 1.6 The nondipole geomagnetic field for 1945. Arrows indicate the magnitude and direction of the horizontal component on the nondipole field; the scale for the arrows is shown at the lower right corner of the diagram; contours indicate lines of equal vertical intensity of the nondipole field; heavy black lines are contours of zero vertical component; thin black lines are contours of positive (downward) vertical component, while gray lines are contours of negative vertical component; the contour interval is 0.02 Oe. Notice the clown-face appearance with the nondipole magnetic field going into the eyes and mouth and being blown out the nose. Redrawn from Bullard et al. (*Phil. Trans. Roy. Soc. London*, v. A243, 67–92, 1950).

Note that in Figure 1.6 there are six or seven continental-scale features that dominate the nondipole field. Some of these features have upward-pointing vertical field and horizontal components that point away from the center of the feature. Magnetic field lines are emerging from the Earth and radiating away from these features. Other nondipole features show the opposite pattern, with magnetic field lines pointing downward and toward the center of the feature. These patterns of the nondipole field can be modeled (at least mathematically) by placing radially pointing magnetic dipoles under each nondipole feature. (However, be advised that the physical interpretation of nondipole features is a matter of debate among geomagnetists.) These radial dipoles are (by best-fit mathematics) placed within the fluid outer core near the boundary with the overlying mantle. Opposite signs of these radial dipoles can account for the opposing field patterns of the nondipole features. This morphology and modeling of the nondipole field suggest an origin in fluid eddy currents in the outer core near the interface with the overlying solid mantle. Indeed, nondipole features are dynamic and exhibit growth, decay, and motions similar to eddy currents in turbulent fluid flow. These time variations have been measured historically and can be determined prehistorically through various paleomagnetic methods.

GEOMAGNETIC SECULAR VARIATION

The direction and magnitude of the surface geomagnetic field change with time. Changes with periods dominantly between 1 yr and 10^5 yr constitute *geomagnetic secular variation*. Even over the time of historic geomagnetic field records, directional changes are substantial. Figure 1.7 shows historic records of geomagnetic field direction in London since reliable recordings were initiated just prior to 1600 A.D. The range of inclination is 66° to 75°, and the range of declination is –25° to +10°, so the directional changes are indeed substantial.

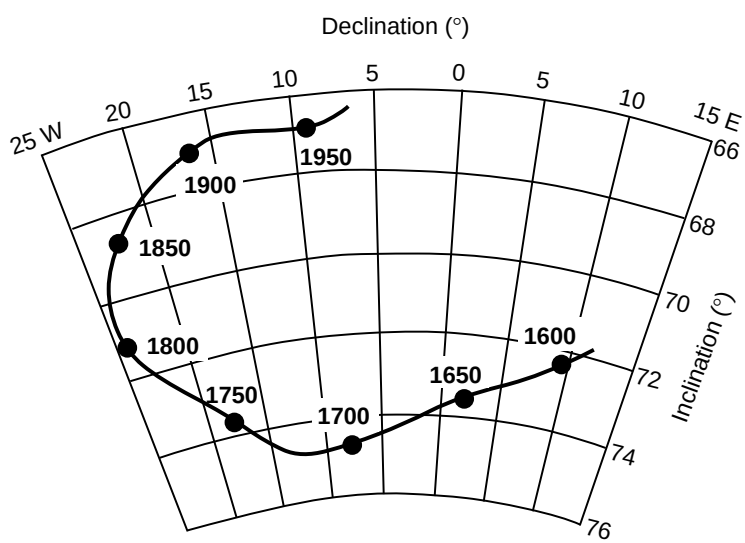

Figure 1.7 Historic record of geomagnetic field direction at Greenwich, England. Declination and inclination are shown; data points are labeled in years A.D.; azimuthal equidistant projection. Redrawn after Malin and Bullard (*Phil. Trans. Roy. Soc. London*, v. A299, 357–423, 1981.)

Patterns of secular variation are similar over subcontinental regions. For example, the pattern of secular variation observed in Paris is similar to that in London. However, from one continent to another, patterns of secular variation are very different. This observation probably reflects the size of the nondipole sources of geomagnetic field within the Earth's core.

The dominant period of the secular variation is longer than the London record, and this sometimes leads to the incorrect impression that secular variation is cyclic and predictable. One of the early objectives of

paleomagnetic investigations (and an area of active research now) was to obtain records of geomagnetic secular variation. Paleomagnetism of archeological artifacts (*archeomagnetism*), Holocene volcanic rocks, and postglacial lake sediments have provided information about secular variation.

A record of geomagnetic secular variation recorded by sediments in Fish Lake in southern Oregon is shown in Figure 1.8. Most directions are within 20° of the mean, but short-term deviations of larger amplitude are present. The observed directional changes are not cyclic. Instead, the directional change is better characterized as a *random walk* about the mean direction. There is a range of periodicities dominantly within 10^2–10^4 yr. Spectral analysis indicates a broad band of energy with periods in the 3000- to 9000-yr interval and maximum energy with periods in the 2500- to 3000-yr range.

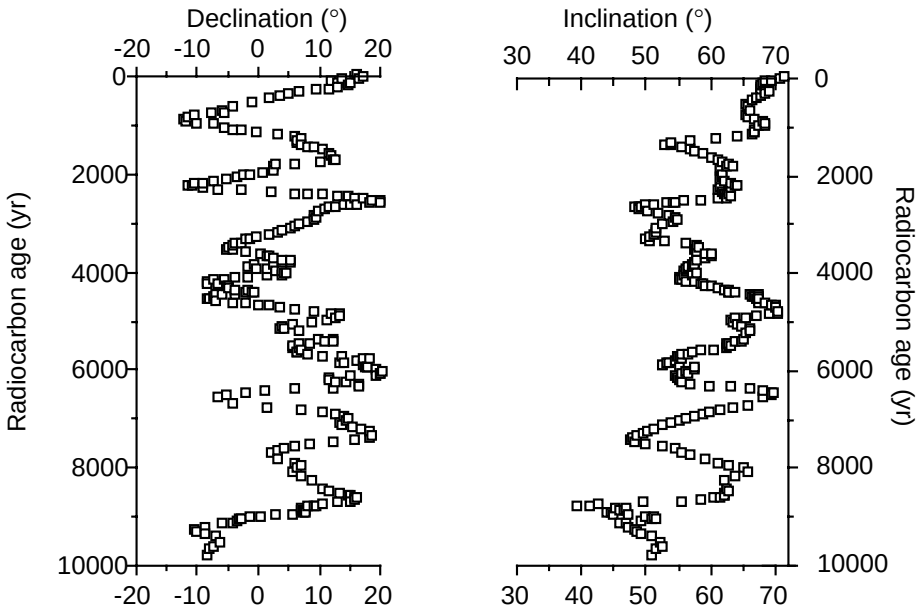

Figure 1.8 Record of Holocene geomagnetic secular variation recorded by sediments in Fish Lake in southeastern Oregon. Declination and inclination are shown against radiocarbon age. Data kindly provided by K. Verosub.

The origins of geomagnetic secular variation can be crudely subdivided into two contributions with overlapping periodicities: (1) nondipole changes dominating the shorter periods and (2) changes of the dipolar field with longer periods. Changes in the nondipole field dominate periodicities less than 3000 yr. Nondipole features appear to grow, decay, and deform with lifetimes of ~10^3 yr. Over historic time, there has been a tendency for some features of the nondipole field to undergo *westward drift*, a longitudinal shift toward the west at a rate of about 0.4° longitude per year. Other nondipole features appear to be stationary.

The dipole portion of the geomagnetic field (90% of the surface field) also changes direction and amplitude. To separate changes of the dipole and nondipole fields, historic records as well as archeomagnetic records and paleomagnetic records from Holocene volcanic rocks have been analyzed. Eight regions of the globe were defined within which mean directions of the geomagnetic field were determined at 100-yr intervals. Magnetic pole positions determined from these regional mean directions were then averaged to yield a global average geomagnetic pole for each 100-yr interval over the past 2000 yr. Results are shown in Figure 1.9.

Because this procedure has provided a global spatial average, effects of the nondipole field have been averaged out, and the secular variation evident in Figure 1.9 is that of the dipole field. The record shows the geomagnetic pole performing a random walk about the north geographic pole (the analogy is a drunk staggering around a light pole). The average position of the geomagnetic pole is indistinguishable from the

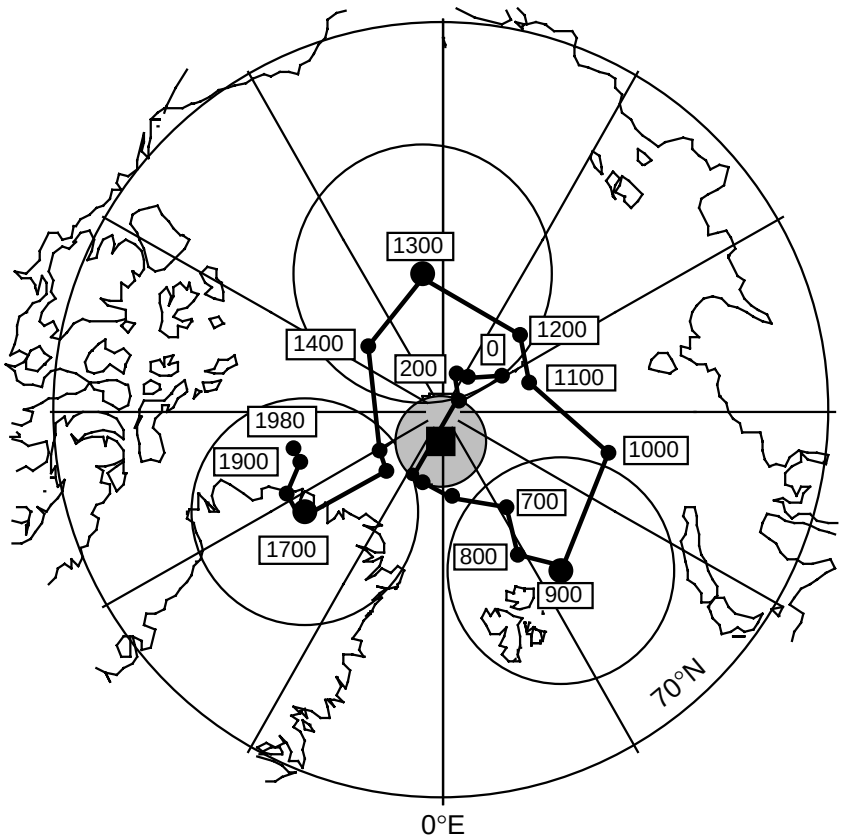

Figure 1.9 Positions of the north geomagnetic pole over the past 2000 yr. Each data point is the mean geomagnetic pole at 100-yr intervals; numbers indicate date in years A.D.; circles about geomagnetic poles at 900, 1300, and 1700 A.D. are 95% confidence limits on those geomagnetic poles; the mean geomagnetic pole position over the past 2000 yr is shown by the square with stippled region of 95% confidence. Data compiled by Merrill and McElhinny (1983).

rotation axis, indicating that the geocentric axial dipole model describes the time-averaged geomagnetic field when averaged over the past 2000 yr. This supports a crucial hypothesis about the geomagnetic field known as the *geocentric axial dipole hypothesis*. This hypothesis simply states that the time-averaged geomagnetic field is a geocentric axial dipolar field. Because this hypothesis is central to many applications of paleomagnetism, it will be explored in considerable detail later.

In addition to changes in orientation of the best-fitting dipole (depicted by changes in geomagnetic pole position shown in Figure 1.9), the amplitude of the geomagnetic dipole also changes with time. A compilation of results is shown in Figure 1.10, which shows variations in the magnitude of the dipole moment. Over the past 10^4 yr, the average dipole moment is 8.75×10^{25} G cm^3 (8.75×10^{22} A m^2). Changes in dipole moment appear to have a period of roughly 10^4 yr, with oscillations of up to ±50% of the mean value.

The picture of the geomagnetic field that emerges from examination of secular variation is one of directional and amplitude changes that are quite rapid for a geological phenomenon. Although short-term deviations of the geomagnetic field direction from the long-term mean direction can exceed 30° or so, the time-averaged field is strikingly close to that of the elegantly simple geocentric axial dipole.

On longer time scales than those considered above, the dipolar geomagnetic field has been observed to switch polarity. The present configuration of the dipole field (pointing toward geographic south) is referred to as *normal polarity*; the opposite configuration is defined as *reversed polarity*. Reversal of the polarity of the dipole produces a 180° change in surface geomagnetic field direction at all points. We shall investigate this phenomenon (especially the geomagnetic polarity time scale) in a later chapter. For now, the essential

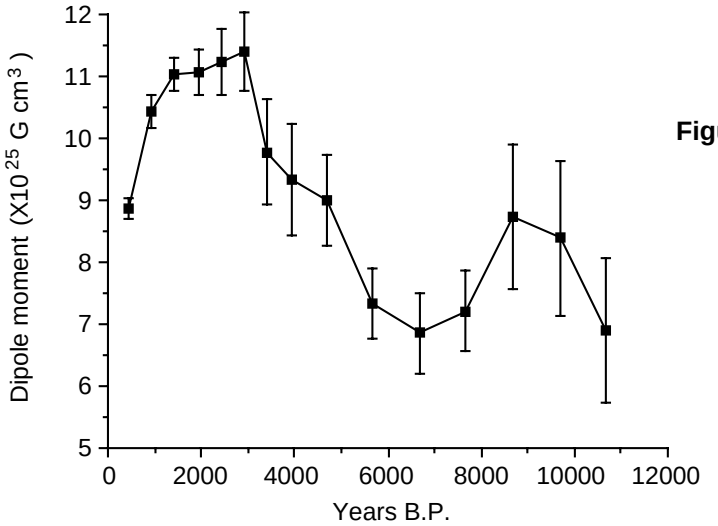

Figure 1.10 Geomagnetic dipole moment over the past 10,000 years. Means for 500-yr intervals are shown to 4000 yr B.P.; 1000-yr means are shown from 4000 to 10,000 yr B.P.; error bars are 95% confidence limits. Redrawn after Merrill and McElhinny (1983).

feature is that the geocentric axial dipole model describes the time-averaged geomagnetic field during either normal-polarity or reversed-polarity intervals.

ORIGIN OF THE GEOMAGNETIC FIELD

Measurement and description of the geomagnetic field and its spatial and temporal variations comprise one of the oldest geophysical disciplines. However, our ability to describe the field far exceeds our understanding of its origin. All plausible theories involve generation of the geomagnetic field within the fluid outer core of the Earth by some form of *magnetohydrodynamic dynamo*. Attempts to solve the full mathematical complexities of magnetohydrodynamics have driven some budding geomagnetists into useful but nonscientific lines of work. In fact, complete dynamical models have not been accomplished, although the plausibility of the magnetohydrodynamic origin of the geomagnetic field is well established.

Quantitative treatment of magnetohydrodynamics is (mercifully) beyond the scope of this book, but we can provide a qualitative explanation. The first step is to gain some appreciation for what is meant by *self-exciting dynamo*. A simple electromechanical disk-dynamo model such as that shown in Figure 1.11 contains the essential elements of a self-exciting dynamo. The model is constructed of a copper disk rotating on an electrically conducting axle. An initial magnetic induction field, $\boldsymbol{B}$ (see Appendix 1.1 for definition), is present in an upward direction perpendicular to the copper disk. Electrons in the copper disk experience a *Lorenz force*, $\boldsymbol{F}_L$, when they pass through this field. The Lorenz force is given by:

$$\boldsymbol{F}_L = q\,\boldsymbol{v} \times \boldsymbol{B} \tag{1.16}$$

where q is the electrical charge of the electrons, and $\boldsymbol{v}$ is the velocity of electrons. This Lorenz force on the electrons is directed toward the axle of the disk and the resulting electrical current flow is toward the outside of the disk (Figure 1.11).

Brush connectors are used to tap the electrical current from the disk, and the current passes through a coil under the disk. This coil is wound so that the electrical current produces a magnetic induction field in the same direction as the original field. The electrical circuit is a *positive feedback system* that reinforces the original magnetic induction field. The entire disk-dynamo model is a self-exciting dynamo. As long as the disk is kept rotating, the electrical current will flow, and the magnetic field will be sustained.

With this simple model we encounter the essential elements of any self-exciting dynamo:

1. A moving electrical conductor is required and is represented by the rotating copper disk.
2. An initial magnetic field is required.

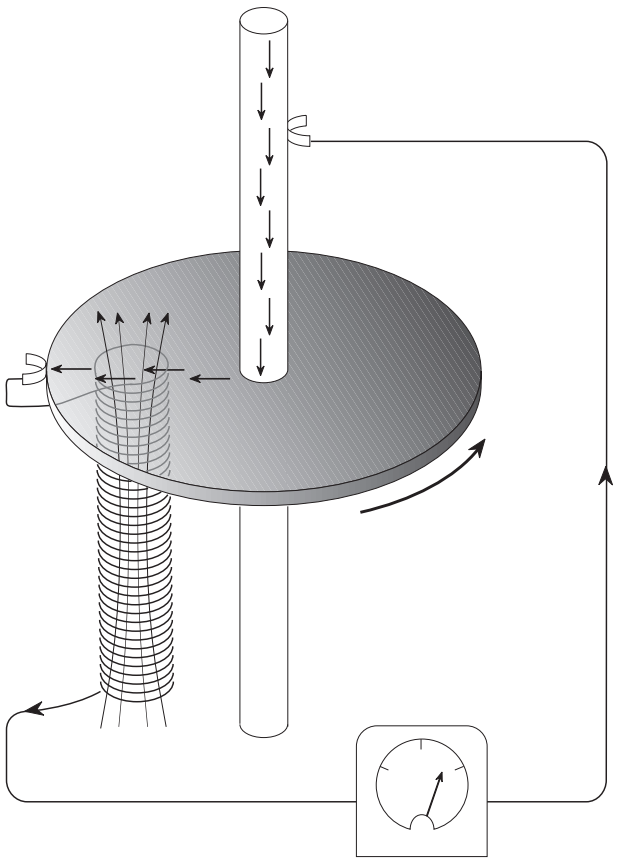

Figure 1.11 Self-exciting disk dynamo. The copper disk rotates on an electrically conducting axle; electrical current is shown by bold arrows; the magnetic field generated by the coil under the disk is shown by the fine arrows. (Adapted from *The Earth as a Dynamo*, W. Elsasser, Copyright© 1958 by Scientific American, Inc. All rights reserved.)

3. An interaction between the magnetic field and the conductor must take place to provide reinforcement of the original magnetic field. In the model, this interaction is the Lorenz force with the coil acting as a positive feedback (self-exciting) circuit.
4. Energy must be supplied to overcome electrical resistivity losses. In the model, energy must be supplied to keep the disk rotating.

Certainly no one proposes that systems of disks and feedback coils exist in the Earth's core. But interaction between the magnetic field and the electrically conducting iron-nickel alloy in the outer core can produce positive feedback and allow the Earth's core to operate as a self-exciting magnetohydrodynamic dynamo. For reasonable electrical conductivities, fluid viscosity, and plausible convective fluid motions in the Earth's outer core, the fluid motions can regenerate the magnetic field that is lost through electrical resistivity. There is a balance between fluid motions regenerating the magnetic field and loss of magnetic field because of electrical resistivity.

Apparently, fluid motions in the Earth's core are sufficient to regenerate the field, but there is enough leakage to keep the shape of the geomagnetic field fairly simple. Thus, the dominant portion of the geomagnetic field is the (simplest possible) dipolar shape with subsidiary nondipolar features probably resulting from fluid eddy currents within the core near the boundary with the overlying mantle.

Even this qualitative view of magnetohydrodynamics provides an explanation for the time-averaged geocentric axial dipolar nature of the geomagnetic field. Rotation of the Earth must be a controlling factor on the time-averaged fluid motions in the outer core. Therefore, the time-averaged magnetic field generated by these fluid motions is quite logically symmetric about the axis of rotation. The simplest such field is a geocentric axial dipolar field.

It should also be pointed out that the magnetohydrodynamic dynamo can operate in either polarity of the dipole. All the physics and mathematics of magnetohydrodynamic generation are invariant with polarity of the dipolar field. Thus, there is no contradiction between the observation of reversals of the geomagnetic

dipole and magnetohydrodynamic generation of the geomagnetic field. However, understanding the special interactions of fluid motions and magnetic field that produce geomagnetic reversals is a major challenge.

As wise economists have long observed, there is no free lunch. The geomagnetic field is no exception. Because of ohmic dissipation of energy, there is a requirement for energy input to drive the magnetohydrodynamic fluid motions and thereby sustain the geomagnetic field. Estimates of the power (energy per unit time) required to generate the geomagnetic field are about 10^{13} W (roughly the output of 10^4 nuclear power plants). This is about one fourth of the total geothermal flux, so the energy involved in generation of the geomagnetic field is a substantial part of the Earth's heat budget.

Many sources of this energy have been proposed, and ideas on this topic have changed over the years. The energy source that is currently thought to be most reasonable is gradual cooling of the Earth's core with attendant freezing of the outer core and growth of the solid inner core. This energy source is plausible in terms of the energy available from growth of the inner core and is efficient in converting energy to fluid motions of the outer core required to generate the geomagnetic field.

APPENDIX 1.1: ABOUT UNITS

Any system of units is basically an arbitrary set of names created to facilitate communication about measured or calculated quantities. These units can be broken down into fundamental quantities: mass, length, time, and electric charge. Before about 1980, most geophysical literature used the cgs system, for which fundamental units were gram (gm), centimeter (cm), seconds (s), and coulomb (C). In an effort to obtain uniformity across various disciplines of physical sciences, international committees have lately recommended usage of the Système Internationale (SI). The SI fundamental units are the meter (m), kilogram (kg), second (s), and coulomb (C). For basic quantities (e.g., force), both the cgs and SI systems are simple and conversions from one system to the other are by integral powers of 10.

However, things are not simple for magnetism, and for various reasons, conversion from cgs to SI has led to confusion rather than clarity. Obviously, we must have a system to follow in this book, and so we must confront the potentially confusing issue of units. In doing so, I adhere to our objective of making the paleomagnetic literature accessible and so provide a basic guide to units as they are actually used by paleomagnetists. First the cgs and SI governing equations and units are explained and a table of the units and conversions is provided. Then the current usage of units in paleomagnetism and the (we hope) simplified system used in this book are explained.

In dealing with units of magnetism, the cgs system is sometimes known as the Gaussian system or emu (electromagnetic) system. In the cgs system, the basic quantities are

$\boldsymbol{B}$ = magnetic induction
$\boldsymbol{H}$ = magnetic field
$\boldsymbol{J}$ = magnetic moment per unit volume, or magnetization

These quantities in cgs are related by

$$\boldsymbol{B} = \mu_0 \boldsymbol{H} + 4\pi \boldsymbol{J} \tag{A1.1}$$

where

$$\boldsymbol{J} = \chi \boldsymbol{H} \tag{A1.2}$$

χ = magnetic susceptibility

and: μ_0 = magnetic permeability of free space = 1.0

$\boldsymbol{B}$, $\boldsymbol{H}$, and $\boldsymbol{J}$ all have the same fundamental units. However, common practice has been to refer to units of $\boldsymbol{B}$ as gauss (G), units of $\boldsymbol{H}$ as oersteds (Oe), and units of $\boldsymbol{J}$ as either gauss or emu/cm^3. Susceptibility, χ, is dimensionless.

In the SI system, $\boldsymbol{B}$, $\boldsymbol{H}$, and $\boldsymbol{J}$ are also used, but an additional quantity, $\boldsymbol{M}_v$, is introduced as the magnetic moment per unit volume. (The symbol $\boldsymbol{M}_v$ is used for volume density of magnetic moment in an attempt to avoid confusion with $\boldsymbol{M}$, which is used for magnetic moment.) These quantities in SI are related by

$$\boldsymbol{B} = \mu_0 \boldsymbol{H} + \boldsymbol{J} \tag{A1.3}$$

where $\mu_0 = 4\pi \times 10^{-7}$ henries/m = permeability of free space and

$$\mathbf{J} = \frac{\chi \mathbf{H}}{\mu_0} \tag{A1.4}$$

In SI, $\boldsymbol{B}$ and $\boldsymbol{J}$ have the same fundamental unit, given the name tesla (T), and $\boldsymbol{M}_v$ and $\boldsymbol{H}$ have the same fundamental unit, amperes/meter (A/m). Again, χ is dimensionless (although this is not so obvious as it was for cgs). Table 1.1 summarizes the fundamental dimensions, units, and conversions for basic quantities in cgs and SI.

Those advocating strict usage of SI would force us to use SI units throughout this book and convert all previous paleomagnetic literature according to Table 1.1. I am not going to do that, not only because I happen to be a little stubborn, but because the current paleomagnetic literature does not strictly conform to SI. I could write this book to conform strictly to SI (honest I could), but the reader would then have unnecessary difficulties in following units in past and current paleomagnetic literature.

The current usage of units in paleomagnetism has developed in the following way. Paleomagnetism and rock magnetism developed when cgs (emu) was the prevailing system. Early literature employs cgs units, and almost all instruments are calibrated in cgs. In addition, for some considerations (like energetics of interactions of magnetic dipole moments with magnetic fields), the cgs system is simply easier to deal with. However, because adherence to SI is now required by most Earth science journals, most paleomagnetists currently do their laboratory work (and thinking?) in cgs, then convert to SI at the last moment to conform with requirements for publication. The conversions used in doing so are really a perversion of the proper SI usage.

For example, let us say that a paleomagnetist does laboratory work on a suite of rocks that have intensity of magnetization, J, of 10^{-4} G. Almost invariably, this observation will get converted to SI by reporting intensity of magnetization as 10^{-1} A/m. Strict adherence to SI would require converting the observed 10^{-4} G magnetization to proper SI units of J which would yield $4\pi \times 10^{-8}$ T. But that procedure requires the dreaded 4π factor and is almost never done. To convert by simple integral powers of ten, the observed intensity of magnetization in cgs is converted (perhaps knowingly but maybe not) to an equivalent SI value of magnetic moment/unit volume, M_v, thus yielding 10^{-1} A/m.

In converting intensities of magnetic fields, H, from cgs units of Oe to SI units, a similar trick is employed. Again, strict adherence to SI would require converting an observed 100 Oe magnetic field to proper SI units of H, yielding $(1/4\pi) \times 10^5$ A/m. Once again to avoid the undesirable 4π factor, the observed magnetic field in oersteds is converted to the equivalent magnetic induction, B = 100 G. Then this value is converted to SI to yield a "magnetic field" (really magnetic induction) value of 10^{-2} tesla or 10 millitesla (mT). This commonly employed scheme of conversion from cgs (emu) to SI is summarized at the bottom of Table 1.1. Clearly, the confusion introduced by these conversions is considerable.

In this book, we use a system of units that is most effective for teaching paleomagnetism and for providing an introduction to the past and current paleomagnetic literature. We use definitions and governing equations for magnetic quantities that are rooted in the cgs system and provide easy conversions to SI. With any system of units, there are some pitfalls, and our system is no exception. Frankly, the primary pitfall is that even the most diligent student is likely to be bored by this discussion of units. Another pitfall is that many presentations employing SI use $\boldsymbol{M}$ as the symbol for dipole moment per unit volume. But the paleomagnetic literature is full of usages of $\boldsymbol{M}$ as magnetic dipole moment. In an effort to be consistent with that common usage, we also use $\boldsymbol{M}$ for magnetic dipole moment. (The only known antidotes to discussions of units are undisturbed silence in a dark room for 15 minutes or a brisk walk in the park. Excess worry about systems of units may cause you to give up the quest of paleomagnetism and take up, say, modern dance.)

TABLE 1.1. Units and Conversions for Common Quantities of Magnetism

	cgs (emu) System		Système Internationale (SI)		
	Fundamental Units	Unit	Fundamental Units	Unit	Conversion
Energy		erg		joule (J)	1 erg = 10^{-7} joule
Force (*F*)	gm cm s^{-2}	dyne	kg m s^{-2}	newton (N)	1 dyne = 10^{-5} newton
Current (*I*)	10 C s^{-1}	abampere	C s^{-1}	ampere (A)	1 abampere = 10 ampere
Magnetic Induction (*B*)	0.1 gm s^{-1} C^{-1}	gauss (G)	kg s^{-1} C^{-1}	tesla (T)	1 gauss = 10^{-4} tesla
Magnetic Field (*H*)	0.1 gm s^{-1} C^{-1}	oersted (Oe)	C s^{-1} m^{-1}	ampere m^{-1} (A/m)	1 Oe = $(1/4\pi) \times 10^3$ A/m
Magnetization (*J*)	0.1 gm s^{-1} C^{-1}	gauss (G) (= emu cm^{-3})	kg s^{-1} C^{-1}	tesla (T)	1 gauss = $4\pi \times 10^{-4}$ tesla
Magnetic Dipole Moment/Unit Volume	0.1 gm s^{-1} C^{-1}	gauss (G) (= emu cm^{-3})	C s^{-1} m^{-1}	A/m	1 gauss = 10^3 A/m
Magnetic Moment (*M*)	0.1 gm s^{-1} C^{-1} cm^3	gauss cm^3 (G cm^3 = emu)	C s^{-1} m^2	A m^2	1 gauss cm^3 = 10^{-3} A m^2
Magnetic Susceptibility (χ)	Dimensionless		Dimensionless		χ (cgs) = $4\pi\,\chi$ (SI)

Conversions commonly employed in paleomagnetism: Magnetization, $J = 10^{-3}$ G converts to "magnetization" = 1 A/m. Magnetic field, H = 1 Oe converts to magnetic "field" = 10^{-4} T = 0.1 mT.

Some Examples: Surface geomagnetic field strength: 0.24–0.66 Oe = 0.024–0.066 mT. Magnetic field generated by laboratory electromagnet: 2000 Oe = 0.2 T = 200 mT. Magnetic dipole moment of the earth: 8×10^{25} G cm^3 = 8×10^{22} A m^2.

Natural remanent magnetization of rocks: basalt: 10^{-3} G = 1 A/m; granite: 10^{-4} G = 0.1 A/m; nonmarine siltstone: 10^{-5} G = 10^{-2} A/m; marine limestone: 10^{-7} G = 10^{-4} A/m.

SUGGESTED READINGS

M. W. McElhinny, *Palaeomagnetism and Plate Tectonics*, Cambridge, London, 356 pp., 1973.
Chapter 1 presents an introduction to the geomagnetic field.

R. T. Merrill and M. W. McElhinny, *The Earth's Magnetic Field*, Academic Press, London, 401 pp., 1983.
An excellent text on geomagnetism. Chapter 2 provides a thorough introduction to the geomagnetic field and historical secular variation.

P. N. Shive, Suggestions for the use of SI units in magnetism, *Eos Trans. AGU*, v. 67, 25, 1986.
Summarizes the problems with units in magnetism.

PROBLEMS

1.1 The pattern of the nondipole geomagnetic field around a major feature of the nondipole field can be modeled by a magnetic dipolar source placed near the core-mantle boundary directly under the center of the feature. Figure 1.12 shows a meridional cross section through the Earth in the plane of the nondipole feature and the magnetic dipole used to model the nondipole feature. At the location directly above the model dipole, the nondipole field is directed vertically downward and has intensity 0.1 Oe. The model dipole is placed at 3480 km from the center of the Earth. Adapt the geometry of Figure 1.3 and the equations describing the magnetic field of a geocentric axial dipole to the model dipole in Figure 1.12. Calculate the magnetic dipole moment of the model dipole and compare your answer to the magnetic dipole moment of the best-fitting dipole for the present geomagnetic field (~8.5×10^{25} G cm^3). Remember to get all required input parameters in cgs units; then your answer will be in cgs units of magnetic moment (G cm^3); mean Earth radius = 6370 km.

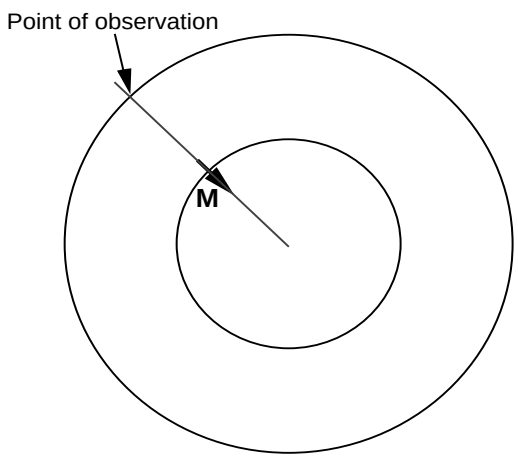

Figure 1.12 Model magnetic dipole for a nondipole feature of the geomagnetic field. The figure is a meridional cross section in the plane containing the middle of the nondipole feature (labeled "point of observation"), the center of the Earth, and the magnetic dipole used to model the nondipole feature.

1.2 The rate of westward drift of the nondipole geomagnetic field is about 0.4° of longitude per year. Features of the nondipole field are generally considered to originate from sources in the outer core near the boundary with the overlying mantle. Imagine a feature of the nondipole field that is centered on the geographic equator. If the source of this nondipole feature is at a distance of 3400 km from the geocenter, what is the linear rate of motion of the source with respect to the lower mantle? Calculate the linear rate in km/yr and in cm/s. (*Note*: On the Earth's surface at the equator, 1° of longitude ≈110 km. 1 yr = 3.16×10^7 s.)

1.3 Convert the following measured quantities in cgs units to SI units using the conversions generally applied in the paleomagnetic literature and described in Appendix 1.1.

a. $J = 3.5 \times 10^{-5}$ G
b. $M = 2.78 \times 10^{-20}$ G cm^3
c. $H = 128$ Oe