Chapter 3

Magnetic Methods

3.1. INTRODUCTION

3.1.1. General

Magnetic and gravity methods have much in common, but magnetics is generally more complex and variations in the magnetic field are more erratic and localized. This is partly due to the difference between the dipolar magnetic field and the monopolar gravity field, partly due to the variable direction of the magnetic field, whereas the gravity field is always in the vertical direction, and partly due to the time-dependence of the magnetic field, whereas the gravity field is time-invariant (ignoring small tidal variations). Whereas a gravity map usually is dominated by regional effects, a magnetic map generally shows a multitude of local anomalies. Magnetic measurements are made more easily and cheaply than most geophysical measurements and corrections are practically unnecessary. Magnetic field variations are often diagnostic of mineral structures as well as regional structures, and the magnetic method is the most versatile of geophysical prospecting techniques. However, like all potential methods, magnetic methods lack uniqueness of interpretation.

3.1.2. History of Magnetic Methods

The study of the earth's magnetism is the oldest branch of geophysics. It has been known for more than three centuries that the Earth behaves as a large and somewhat irregular magnet. Sir William Gilbert (1540–1603) made the first scientific investigation of terrestrial magnetism. He recorded in *de Magnete* that knowledge of the north-seeking property of a magnetite splinter (a *lodestone* or leading stone) was brought to Europe from China by Marco Polo. Gilbert showed that the Earth's magnetic field was roughly equivalent to that of a permanent magnet lying in a general north–south direction near the Earth's rotational axis.

Karl Frederick Gauss made extensive studies of the Earth's magnetic field from about 1830 to 1842, and most of his conclusions are still valid. He concluded from mathematical analysis that the magnetic field was entirely due to a source within the Earth, rather than outside of it, and he noted a probable connection to the Earth's rotation because the axis of the dipole that accounts for most of the field is not far from the Earth's rotational axis.

The terrestrial magnetic field has been studied almost continuously since Gilbert's time, but it was not until 1843 that von Wrede first used variations in the field to locate deposits of magnetic ore. The publication, in 1879, of *The Examination of Iron Ore Deposits by Magnetic Measurements* by Thalén marked the first use of the magnetic method.

Until the late 1940s, magnetic field measurements mostly were made with a magnetic balance, which measured one component of the earth's field, usually the vertical component. This limited measurements mainly to the land surface. The fluxgate magnetometer was developed during World War II for detecting submarines from an aircraft. After the war, the fluxgate magnetometer (and radar navigation, another war development) made aeromagnetic measurements possible. Proton-precession magnetometers, developed in the mid-1950s, are very reliable and their operation is simple and rapid. They are the most commonly used instruments today. Optical-pump alkali-vapor magnetometers, which began to be used in 1962, are so accurate that instrumentation no longer limits the accuracy of magnetic measurements. However, proton-precession and optical-pump magnetometers measure only the magnitude, not the direction, of the magnetic field. Airborne gradiometer measurements began in the late 1960s, although ground measurements were made much earlier. The gradiometer often consists of two magnetometers vertically spaced 1 to 30 m apart. The difference in readings not only gives the vertical gradient, but also, to a large extent, removes the effects of temporal field variations, which are often the limiting factor on accuracy.

Digital recording and processing of magnetic data removed much of the tedium involved in reducing measurements to magnetic maps. Interpretation algorithms now make it possible to produce computer-drawn profiles showing possible distributions of magnetization.

The history of magnetic surveying is discussed by Reford (1980) and the state of the art is discussed by Paterson and Reeves (1985).

3.2. PRINCIPLES AND ELEMENTARY THEORY

3.2.1. Classical versus Electromagnetic Concepts

Modern and classical magnetic theory differ in basic concepts. Classical magnetic theory is similar to electrical and gravity theory; its basic concept is that point magnetic poles are analogous to point electrical charges and point masses, with a similar inverse-square law for the forces between the poles, charges, or masses. Magnetic units in the centimeter-gram-second and electromagnetic units (cgs and emu) system are based on this concept. Système International (SI) units are based on the fact that a magnetic field is electrical in origin. Its basic unit is the dipole, which is created by a circular electrical current, rather than the fictitious isolated monopole of the cgs–emu system. Both emu and SI units are in current use.

The cgs–emu system begins with the concept of magnetic force **F** given by Coulomb's law:

$$\mathbf{F} = (p_1 p_2/\mu r^2)\mathbf{r}_1 \tag{3.1}$$

where **F** is the force on p_2, in dynes, the poles of strength p_1 and p_2 are r centimeters apart, μ is the *magnetic permeability* [a property of the medium; see Eq. (3.7)], and $\mathbf{r}_1$ is a unit vector directed from p_1 toward p_2. As in the electrical case (but unlike the gravity case, in which the force is always attractive), the magnetostatic force is attractive for poles of opposite sign and repulsive for poles of like sign. The sign convention is that a *positive pole* is attracted toward the Earth's north magnetic pole; the term *north-seeking* is also used.

The *magnetizing field* **H** (also called *magnetic field strength*) is defined as the force on a unit pole:

$$\mathbf{H}' = \mathbf{F}/p_2 = (p_1/\mu r^2)\mathbf{r}_1 \tag{3.2}$$

(we use a prime to indicate that H is in cgs–em units); **H′** is measured in oersteds (equivalent to dynes per unit pole).

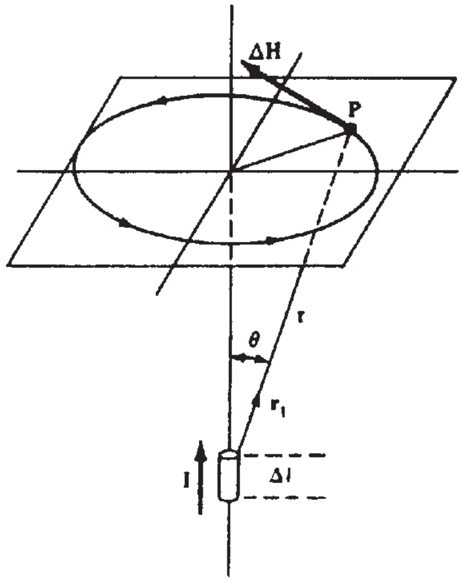

Figure 3.1. Ampère's law. A current **I** *through a length of conductor* Δl creates a magnetizing field $\Delta\mathbf{H}$ *at a point P:*

$$\Delta\mathbf{H} = (\mathbf{I}\,\Delta l) \times \mathbf{r}_1/4\pi r^2$$

where $\Delta\mathbf{H}$ *is in amperes per meter when* **I** *is in amperes and r and* Δl *are in meters.*

A *magnetic dipole* is envisioned as two poles of strength $+p$ and $-p$ separated by a distance $2l$. The *magnetic dipole moment* is defined as

$$\mathbf{m} = 2lp\mathbf{r}_1 \tag{3.3}$$

m is a vector in the direction of the unit vector $\mathbf{r}_1$ that extends from the negative pole toward the positive pole.

A magnetic field is a consequence of the flow of an electrical current. As expressed by Ampère's law (also called the Biot–Savart law), a current **I** in a conductor of length Δl creates, at a point P (Fig. 3.1), a magnetizing field $\Delta\mathbf{H}$ given by

$$\Delta\mathbf{H} = (\mathbf{I}\,\Delta l) \times \mathbf{r}_1/4\pi r^2 \tag{3.4}$$

where **H** has the SI dimension amperes per meter [$= 4\pi \times 10^{-3}$ oersted], r and Δl are in meters, **I** is in amperes, and $\Delta\mathbf{H}$, $\mathbf{r}_1$, and $\mathbf{I}\,\Delta l$ have the directions indicated in Figure 3.1.

A current flowing in a circular loop acts as a magnetic dipole located at the center of the loop and oriented in the direction in which a right-handed screw would advance if turned in the direction of the current. Its dipole moment is measured in ampere-meter^2 ($= 10^{10}$ pole-cm). The orbital motions of electrons around an atomic nucleus constitute circular currents and cause atoms to have magnetic moments. Molecules also have spin, which gives them magnetic moments.

A magnetizable body placed in an external magnetic field becomes magnetized by induction; the magnetization is due to the reorientation of atoms and molecules so that their spins line up. The magnetization is measured by the *magnetic polarization* **M** (also called *magnetization intensity* or *dipole moment per unit volume*). The lineup of internal dipoles produces a field **M**, which, within the body, is added to the magnetizing field **H**. If **M** is constant and has the same direction throughout, a body is said to be *uniformly magnetized*. The SI unit for magnetization is ampere-meter^2 per meter^3 [= ampere per meter (A/m)].

For low magnetic fields, **M** is proportional to **H** and is in the direction of **H**. The degree to which a body is magnetized is determined by its *magnetic susceptibility k*, which is defined by

$$\mathbf{M} = k\mathbf{H} \tag{3.5}$$

Magnetic susceptibility in emu differs from that in SI units by the factor 4π, that is,

$$k_{SI} = 4\pi k'_{emu} \tag{3.6}$$

Susceptibility is the fundamental rock parameter in magnetic prospecting. The magnetic response of rocks and minerals is determined by the amounts and susceptibilities of magnetic materials in them. The susceptibilities of various materials are listed in Table 3.1, Section 3.3.7.

The *magnetic induction* **B** is the total field, including the effect of magnetization. It can be written

$$\mathbf{B} = \mu_0(\mathbf{H} + \mathbf{M}) = \mu_0(1 + k)\mathbf{H} = \mu\mu_0\mathbf{H} \tag{3.7a}$$

$$\mathbf{B}' = \mathbf{H}' + 4\pi\mathbf{M}' = (1 + 4\pi k')\mathbf{H}' = \mu\mathbf{H}' \tag{3.7b}$$

when **H** and **M** (**H′** and **M′**) are in the same direction, as is usually the case. The SI unit for **B** is the tesla = 1 newton/ampere-meter = 1 weber/meter^2 (Wb/m^2). The electromagnetic unit for **B′** is the gauss [= 10^{-4} tesla (T)]. The permeability of free space μ_0 has the value $4\pi \times 10^{-7}$ Wb/A-m. In vacuum $\mu = 1$ and in air $\mu \approx 1$. Confusion sometimes results between **H′** and **B′** because the em units gauss and oersted are numerically equal and dimensionally the same, although conceptually different; both **H′** and **B′** are sometimes called the "magnetic field strength." In magnetic prospecting, we measure **B** to about 10^{-4} of the Earth's main field (which is about 50 μT). The unit of magnetic induction generally used for geophysical work is the nanotesla (also

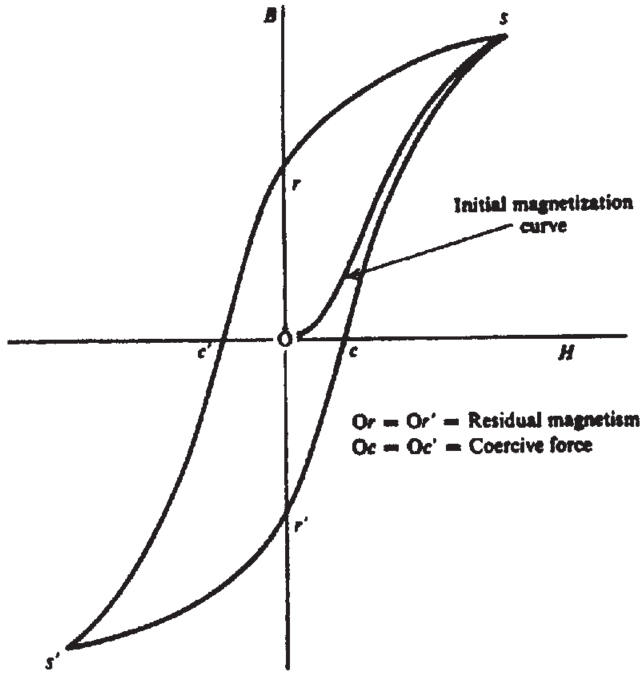

Figure 3.2. Hysteresis loop. s, s′ = saturation, r and r′ = remanent magnetism, c and c′ = coercive force.

called the gamma, γ):

$$1\gamma = 10^{-9}\ \mathrm{T} = 1\ \mathrm{nT}$$

There is often confusion as to whether the quantity involved in magnetic exploration is **B** or **H**. Although we measure $\mathbf{B}_e$, we are interested in the Earth's field $\mathbf{H}_e$. However, because **B** and **H** are linearly related [Eq. (3.7)] and usually $\mu \approx 1$, we can (and do) treat a map of $\mathbf{B}_e$ as if it were a map of $\mathbf{H}_e$.

We also speak of *magnetic flux* or *magnetic lines of force* ϕ:

$$\phi = \mathbf{B} \cdot \mathbf{A} \tag{3.8}$$

where **A** is a vector area (§A.3.2). Thus $|\mathbf{B}| = \phi/|\mathbf{A}|$ when **A** and **B** are parallel, that is, **B** is the density of magnetic flux. The SI unit for magnetic flux is the weber (= T-m^2) and the em unit is the maxwell (= 10^{-8} Wb).

3.2.2. B-H Relations: The Hysteresis Loop

The relation between **B** and **H** can be complex in ferromagnetic materials (§3.3.5). This is illustrated by hysteresis (Fig. 3.2) in a cycle of magnetization. If a demagnetized sample is subjected to an increasing magnetizing field **H**, we obtain the first portion of the curve in which **B** increases with **H** until it flattens off as we approach the maximum value that *B* can have for the sample (*saturation*). When **H** is decreased, the curve does not retrace the same path, but it does show a positive value of **B** when **H** = 0;

this is called *residual* (*remanent*) *magnetism*. When **H** is reversed, **B** finally becomes zero at some negative value of **H** known as the *coercive force*. The other half of the hysteresis loop is obtained by making **H** still more negative until reverse saturation is reached and then returning **H** to the original positive saturation value. The area inside the curve represents the energy loss per cycle per unit volume as a result of hysteresis (see Kip, 1962, pp. 235–7). Residual effects in magnetic materials will be discussed in more detail in Section 3.3.6. In some magnetic materials, **B** may be quite large as a result of previous magnetization having no relation to the present value of **H**.

3.2.3. Magnetostatic Potential for a Dipole Field

Conceptually the magnetic scalar potential A at the point P is the work done on a unit positive pole in bringing it from infinity by any path against a magnetic field $\mathbf{F}(r)$ [compare Eq. (2.4)]. (Henceforth in this chapter $\mathbf{F}$, F indicate magnetic field rather than force and we assume $\mu = 1$.) When $\mathbf{F}(r)$ is due to a positive pole at a distance r from P,

$$A(r) = -\int_{-\infty}^{r} \mathbf{F}(r) \cdot d\mathbf{r} = p/r \qquad (3.9)$$

However, since a magnetic pole cannot exist, we consider a magnetic dipole to get a realistic entity. Referring to Figure 3.3, we calculate A at an external point:

$$A = \left(\frac{p}{r_1} - \frac{p}{r_2}\right)$$

$$= p\left\{\frac{1}{\left(r^2 + l^2 - 2lr\cos\theta\right)^{1/2}} - \frac{1}{\left(r^2 + l^2 + 2lr\cos\theta\right)^{1/2}}\right\} \qquad (3.10)$$

We can derive the vector $\mathbf{F}$ by taking the gradient of A [Eq. (A.17)]:

$$\mathbf{F}(r) = -\nabla A(r) \qquad (3.11)$$

Its radial component is $F_r = -\partial A/\partial r$ and its angular component is $F_\theta = -\partial A/r\partial\theta$; these are

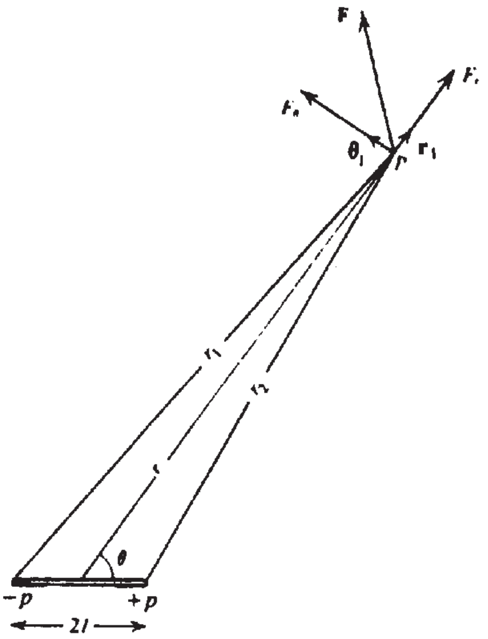

Figure 3.3. Calculating the field of a magnetic dipole.

$$F_r = -p\left\{\frac{r + l\cos\theta}{\left(r^2 + l^2 + 2rl\cos\theta\right)^{3/2}} - \frac{r - l\cos\theta}{\left(r^2 + l^2 - 2rl\cos\theta\right)^{3/2}}\right\} \qquad (3.12a)$$

$$F_\theta = p\left\{\frac{l\sin\theta}{\left(r^2 + l^2 + 2rl\cos\theta\right)^{3/2}} + \frac{l\sin\theta}{\left(r^2 + l^2 - 2rl\cos\theta\right)^{3/2}}\right\} \qquad (3.12b)$$

When $r \gg l$, Equation (3.10) becomes

$$A \approx |\mathbf{m}|\cos\theta/r^2 \qquad (3.13)$$

where $\mathbf{m}$ is the dipole moment of magnitude $m = 2lp$. Equations (3.11) and (3.13) give [§A.4 and Equation (A.33)]

$$\mathbf{F} \approx (m/r^3)(2\cos\theta\,\mathbf{r}_1 + \sin\theta\,\boldsymbol{\theta}_1) \qquad (3.14a)$$

where unit vectors $\mathbf{r}_1$ and $\boldsymbol{\theta}_1$ are in the direction of increasing r and θ (counterclockwise in Fig. 3.3). The resultant magnitude is

$$F = |\mathbf{F}| \approx (m/r^3)(1 + 3\cos^2\theta)^{1/2} \qquad (3.14b)$$

and the direction with respect to the dipole axis is

$$\tan\alpha = F_\theta/F_r = (1/2)\tan\theta \qquad (3.14c)$$

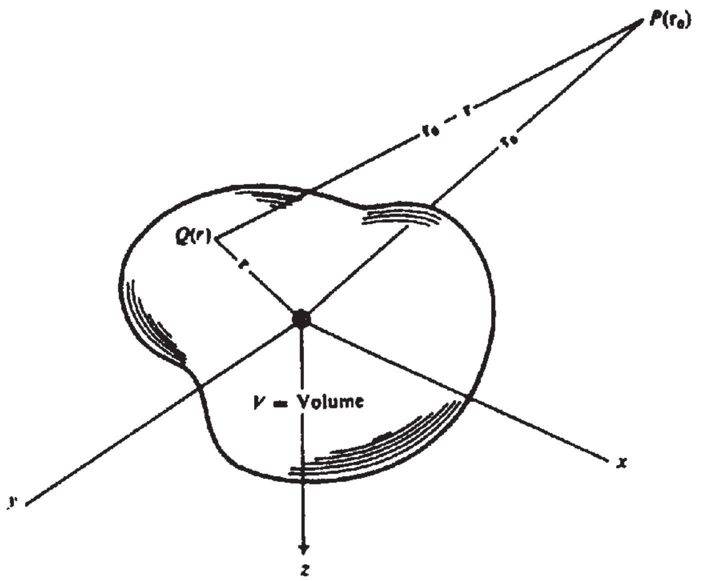

Figure 3.4. General magnetic anomaly.

Two special cases, $\theta = 0$ and $\pi/2$ in Equation (3.12), are called the *Gauss-A* (end-on) and *Gauss-B* (side-on) positions. From Equations (3.12) they are given by

$$F_r = 2mr/(r^2 - l^2)^2 \quad F_\theta = 0 \quad \theta = 0 \tag{3.15a}$$

$$F_r = 0 \quad F_\theta = m/(r^2 + l^2)^{3/2} \quad \theta = \pi/2 \tag{3.15b}$$

If $r \gg l$, these simplify to

$$\left.\begin{aligned} F_r &\approx 2m/r^3 \quad & \theta = 0 \\ F_\theta &\approx m/r^3 \quad & \theta = \pi/2 \end{aligned}\right\} \tag{3.15c}$$

3.2.4. The General Magnetic Anomaly

A volume of magnetic material can be considered as an assortment of magnetic dipoles that results from the magnetic moments of individual atoms and dipoles. Whether they initially are aligned so that a body exhibits residual magnetism depends on its previous magnetic history. They will, however, be aligned by induction in the presence of a magnetizing field. In any case, we may regard the body as a continuous distribution of dipoles resulting in a vector dipole moment per unit volume, **M**, of magnitude M. The scalar potential at P [see Fig. 3.3 and Eq. (3.13)] some distance away from a dipole M ($r \gg l$) is

$$A = M(r)\cos\theta/r^2 = -\mathbf{M}(r) \cdot \nabla(1/r) \tag{3.16}$$

The potential for the whole body at a point outside the body (Fig. 3.4) is

$$A = -\int_V \mathbf{M}(\mathbf{r}) \cdot \nabla\left(\frac{1}{|\mathbf{r}_0 - \mathbf{r}|}\right) dv \tag{3.17}$$

The resultant magnetic field can be obtained by employing Equation (3.11) with Equation (3.17). This gives

$$\mathbf{F}(\mathbf{r}_0) = \nabla \int_V \mathbf{M}(\mathbf{r}) \cdot \nabla\left(\frac{1}{|\mathbf{r}_0 - \mathbf{r}|}\right) dv \tag{3.18}$$

If **M** is a constant vector with direction $\boldsymbol{\alpha} = \ell\mathbf{i} + m\mathbf{j} + n\mathbf{k}$, then the operation

$$\mathbf{M} \cdot \nabla = M\frac{\partial}{\partial \alpha} = M\left(\ell\frac{\partial}{\partial x} + m\frac{\partial}{\partial y} + n\frac{\partial}{\partial z}\right) \tag{3.19}$$

[Eq. (A.18)] and

$$A = -M\frac{\partial}{\partial \alpha}\int_V\left(\frac{dv}{|\mathbf{r}_0 - \mathbf{r}|}\right) \tag{3.20}$$

The magnetic field in Equation (3.20) exists in the presence of the Earth's field $\mathbf{F}_e$, that is, the total field **F** is given by

$$\mathbf{F} = \mathbf{F}_e + \mathbf{F}(r_0)$$

where the directions of $\mathbf{F}_e$ and $\mathbf{F}(r_0)$ are not necessarily the same. If $\mathbf{F}(r_0)$ is much smaller than $\mathbf{F}_e$ or if the body has no residual magnetism, **F** and $\mathbf{F}_e$ will be in approximately the same direction. Where $\mathbf{F}(r_0)$ is an appreciable fraction (say, 25% or more) of $\mathbf{F}_e$ and

A= magnetic potential
U= gravitational potential
M= magnetic moment

has a different direction, the component of $\mathbf{F}(r_0)$ in the direction of $\mathbf{F}_e$, F_D, becomes [Eq. (3.20)]

$$F_D = -\mathbf{f}_1 \cdot \nabla A = -\frac{\partial A}{\partial f} = M\frac{\partial^2}{\partial \alpha\, \partial f}\int_V \frac{dv}{|\mathbf{r}_0 - \mathbf{r}|} \tag{3.21a}$$

where $\mathbf{f}_1$ is a unit vector in the direction of $\mathbf{F}_e$ (§3.3.2a). If the magnetization is mainly induced by $\mathbf{F}_e$, then

$$F_D(\mathbf{r}_0) = M\frac{\partial^2}{\partial f^2}\int_V \frac{dv}{|\mathbf{r}_0 - \mathbf{r}|} = kF_e\frac{\partial^2}{\partial f^2}\int_V \frac{dv}{|\mathbf{r}_0 - \mathbf{r}|} \tag{3.21b}$$

The magnetic interpretation problem is clearly more complex than the gravity problem because of the dipolar field (compare §2.2.3).

The magnetic potential A, like the gravitational potential U, satisfies Laplace's and Poisson's equations. Following the method used to derive Equations (2.12) and (2.13), we get

$$\nabla \cdot \mathbf{F} = -\nabla^2 A = 4\pi\mu p$$

p is the net positive pole strength per unit volume at a point. We recall that a field $\mathbf{F}$ produces a partial reorientation along the field direction of the previously randomly oriented elementary dipoles. This causes, in effect, a separation of positive and negative poles. For example, the x component of $\mathbf{F}$ separates pole strengths $+q$ and $-q$ by a distance ζ along the x axis and causes a net positive pole strength $(q\zeta)\,dy\,dz = M_x\,dy\,dz$ to enter the rear face in Figure A.2a. Because the pole strength leaving through the opposite face is $\{M_x + (\partial M_x/dx)\,dx\}\,dy\,dz$, the net positive pole strength per unit volume (p) created at a point by the field $\mathbf{F}$ is $-\nabla \cdot \mathbf{M}$. Thus,

$$\nabla^2 A = 4\pi\mu\nabla \cdot \mathbf{M}(r) \tag{3.22}$$

In a nonmagnetic medium, $\mathbf{M} = 0$ and

$$\nabla^2 A = 0 \tag{3.23}$$

3.2.5. Poisson's Relation

If we have an infinitesimal unit volume with magnetic moment $\mathbf{M} = M\boldsymbol{\alpha}_1$ and density ρ, then at a distant point we have, from Equation (3.16),

$$A = -\mathbf{M} \cdot \nabla(1/r) = -M\,\nabla(1/r) \cdot \boldsymbol{\alpha}_1 \tag{3.24}$$

From Equations (2.3a), (2.5), and (A.18), the component of $\mathbf{g}$ in the direction $\boldsymbol{\alpha}_1$ is

ρ= density

$$g_\alpha = -dU/d\alpha = -\nabla U \cdot \boldsymbol{\alpha}_1 = -\gamma\rho\nabla(1/r) \cdot \boldsymbol{\alpha}_1 \tag{3.25}$$

Thus,

$$A = (M/\gamma\rho)\,g_\alpha \tag{3.26}$$

If we apply this result to an extended body, we must sum contributions for each element of volume. Provided that M and ρ do not change throughout the body, the potentials A and U will be those for the extended body. Therefore, Equations (3.24) to (3.26) are valid for an extended body with constant density and uniform magnetization.

In terms of fields,

$$\begin{aligned} \mathbf{F} &= -\nabla A = -(M/\gamma\rho)\nabla g_\alpha \\ &= (M/\gamma\rho)\nabla(\nabla U \cdot \boldsymbol{\alpha}_1) \\ &= (M/\gamma\rho)\nabla U_\alpha \end{aligned} \tag{3.27a}$$

where $U_\alpha = dU/d\alpha$. For a component of $\mathbf{F}$ in the direction β_1, this becomes

$$F_\beta = (M/\gamma\rho)\,U_{\alpha\beta} \tag{3.27b}$$

In particular, if $\mathbf{M}$ is vertical, the vertical component of $\mathbf{F}$ is

$$Z = (M/\gamma\rho)\,U_{zz} = (M/\gamma\rho)(\partial g_z/\partial z) \tag{3.28}$$

These relations are used to make pseudogravity maps from magnetic data.

3.3. MAGNETISM OF THE EARTH

3.3.1. Nature of the Geomagnetic Field

As far as exploration geophyics is concerned, the geomagnetic field of the Earth is composed of three parts:

1. The main field, which varies relatively slowly and is of internal origin.
2. A small field (compared to the main field), which varies rather rapidly and originates outside the Earth.
3. Spatial variations of the main field, which are usually smaller than the main field, are nearly constant in time and place, and are caused by local magnetic anomalies in the near-surface crust of the Earth. These are the targets in magnetic prospecting.

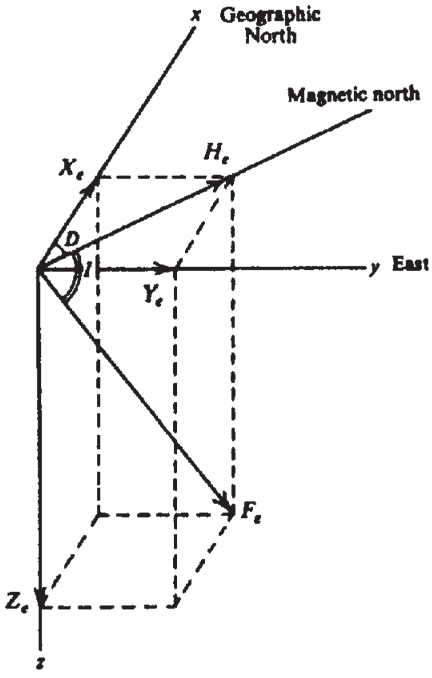

Figure 3.5. Elements of the Earth's magnetic field.

3.3.2. The Main Field

(a) The Earth's magnetic field. If an unmagnetized steel needle could be hung at its center of gravity, so that it is free to orient itself in any direction, and if other magnetic fields are absent, it would assume the direction of the Earth's total magnetic field, a direction that is usually neither horizontal nor in-line with the geographic meridian. The magnitude of this field, F_e, the *inclination* (or *dip*) of the needle from the horizontal, I, and the angle it makes with geographic north (the *declination*), D, completely define the main magnetic field.

The *magnetic elements* (Whitham, 1960) are illustrated in Figure 3.5. The field can also be described in terms of the vertical component, Z_e, reckoned positive downward, and the horizontal component, H_e, which is always positive. X_e and Y_e are the components of H_e, which are considered positive to the north and east, respectively. These elements are related as follows:

$$\left.\begin{aligned} F_e^2 &= H_e^2 + Z_e^2 = X_e^2 + Y_e^2 + Z_e^2 \\ H_e &= F_e \cos I \qquad Z_e = F_e \sin I \\ X_e &= H_e \cos D \qquad Y_e = H_e \sin D \\ \tan D &= Y_e/X_e \qquad \tan I = Z_e/H_e \\ \mathbf{F}_e &= F_e \mathbf{1}_1 = F_e(\cos D \cos I\mathbf{i} \\ &\qquad + \sin D \cos I\mathbf{j} + \sin I\mathbf{k}) \end{aligned}\right\} \quad (3.29)$$

As stated earlier, the end of the needle that dips downward in northern latitudes is the north-seeking or positive pole; the end that dips downward in southern latitudes is the south-seeking or negative pole.

Maps showing lines of equal declination, inclination, horizontal intensity, and so on, are called *isomagnetic maps* (Fig. 3.6). *Isogonic, isoclinic,* and *isodynamic maps* show, respectively, lines of equal declination D, inclination I, and equal values of F_e, H_e, or Z_e. Note that the inclination is large (that is, $Z_e > H_e$) for most of the Earth's land masses, and hence corrections do not have to be made for latitude variations of F_e or Z_e ($\approx$ 4 nT/km) except for surveys covering extensive areas. The overall magnetic field does not reflect variations in surface geology, such as mountain ranges, mid-ocean ridges or earthquake belts, so the source of the main field lies deep within the Earth. The geomagnetic field resembles that of a dipole whose north and south magnetic poles are located approximately at 75°N, 101°W and 69°S, 145°E. The dipole is displaced about 300 km from the Earth's center toward Indonesia and is inclined some 11.5° to the Earth's axis. However, the geomagnetic field is more complicated than the field of a simple dipole. The points where a dip needle is vertical, the *dip poles*, are at 75°N, 101°W and 67°S, 143°E.

The magnitudes of F_e at the north and south magnetic poles are 60 and 70 μT, respectively. The minimum value, ~ 25 μT, occurs in southern Brazil —South Atlantic. In a few locations, F_e is larger than 300 μT because of near-surface magnetic features. The line of zero inclination (*magnetic equator*, where $Z = 0$) is never more than 15° from the Earth's equator. The largest deviations are in South America and the eastern Pacific. In Africa and Asia it is slightly north of the equator.

(b) Origin of the main field. Spherical harmonic analysis of the observed magnetic field shows that over 99% is due to sources inside the Earth. The present theory is that the main field is caused by convection currents of conducting material circulating in the liquid outer core (which extends from depths of 2,800 to 5,000 km). The Earth's core is assumed to be a mixture of iron and nickel, both good electrical conductors. The magnetic source is thought to be a self-excited dynamo in which highly conductive fluid moves in a complex manner caused by convection. Paleomagnetic data show that the magnetic field has always been roughly along the Earth's spin axis, implying that the convective motion is coupled to the Earth's spin. Recent exploration of the magnetic fields of other planets and their satellites provide fascinating comparisons with the Earth's field.

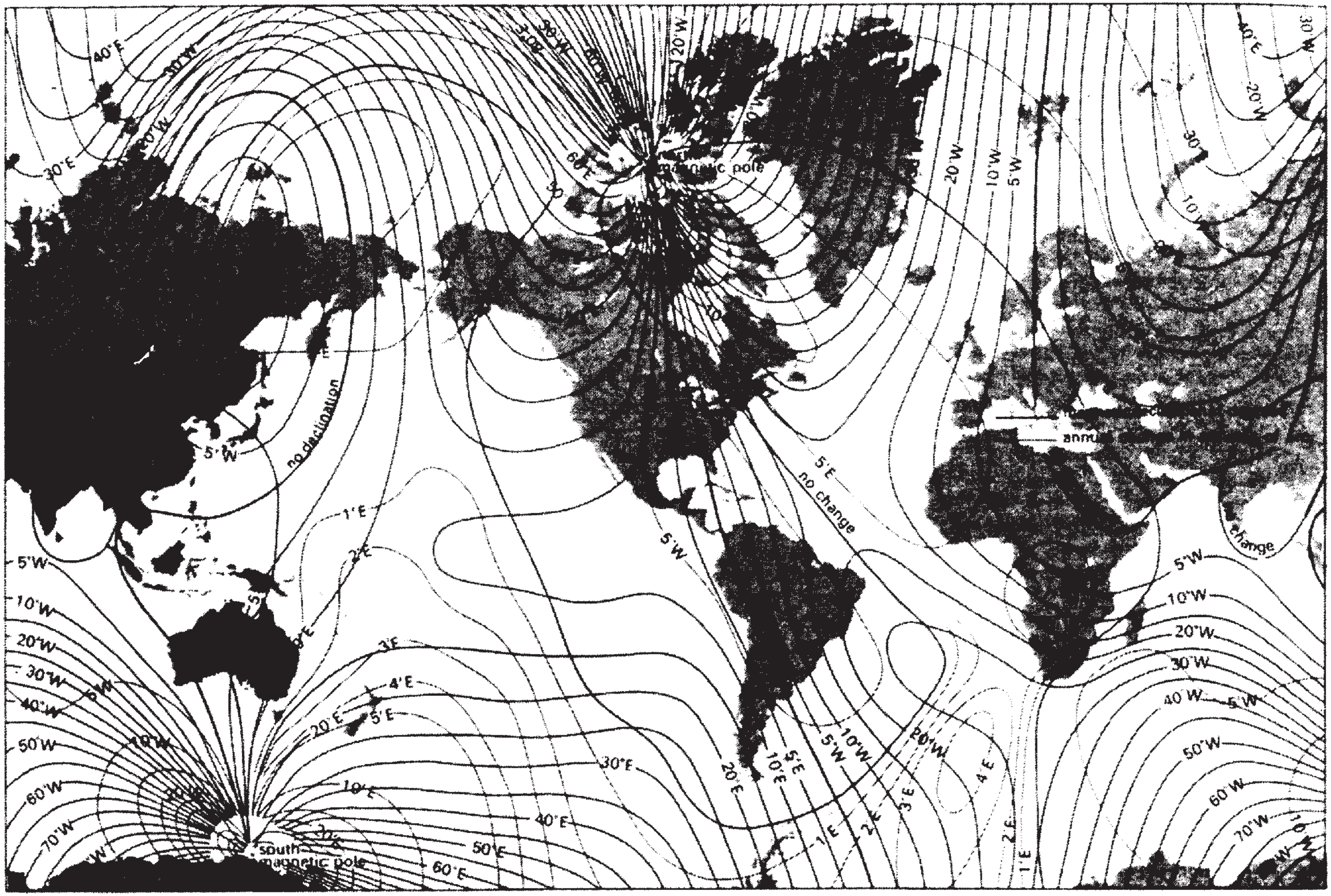

Figure 3.6. The Earth's magnetic field in 1975. (From Smith, 1982). (a) Declination (heavy lines) and annual rate of change in minutes/year (light lines)

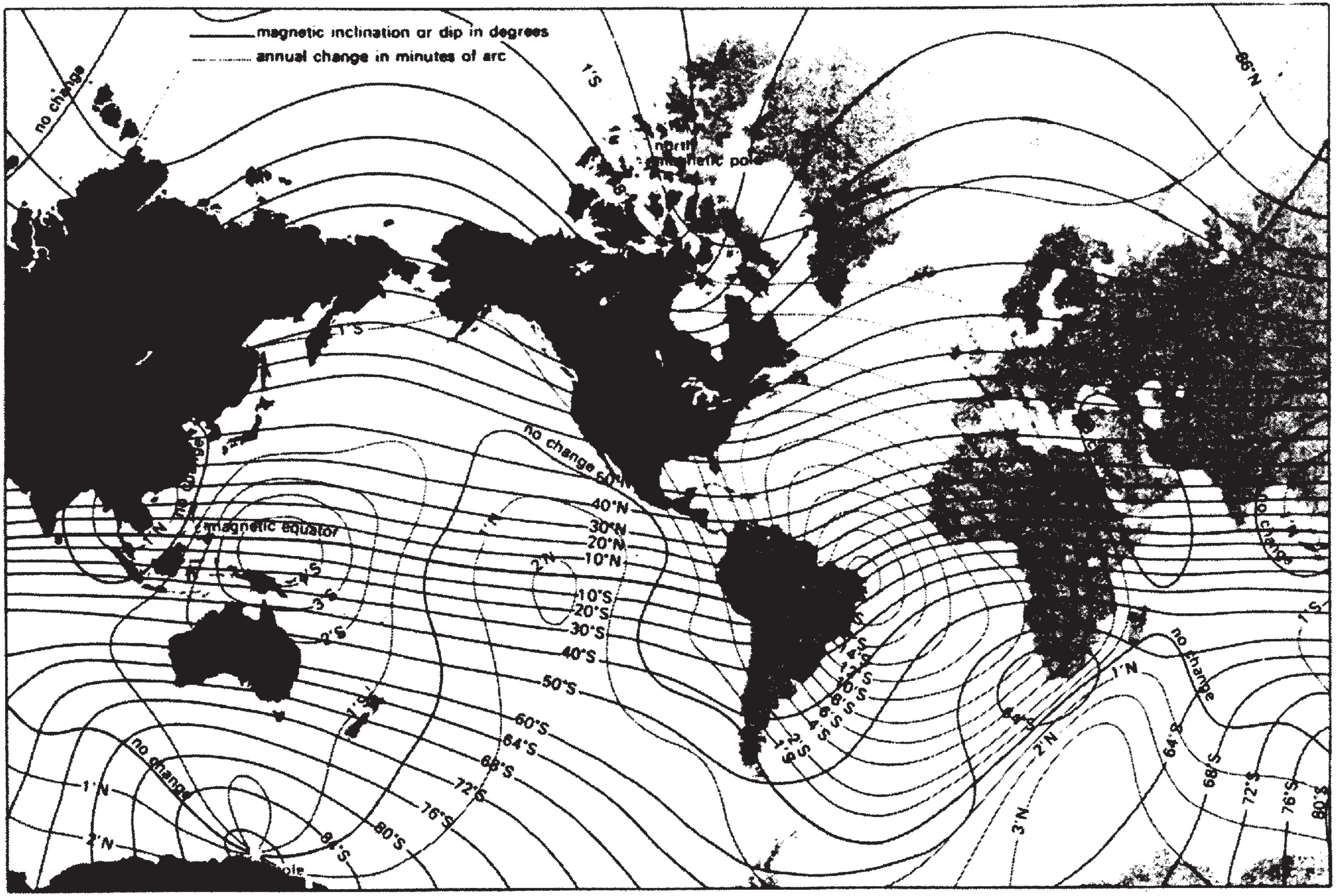

Figure 3.6. (Continued) (b) geomagnetic latitude (heavy lines) and annual rate of change in minutes/year (light lines)

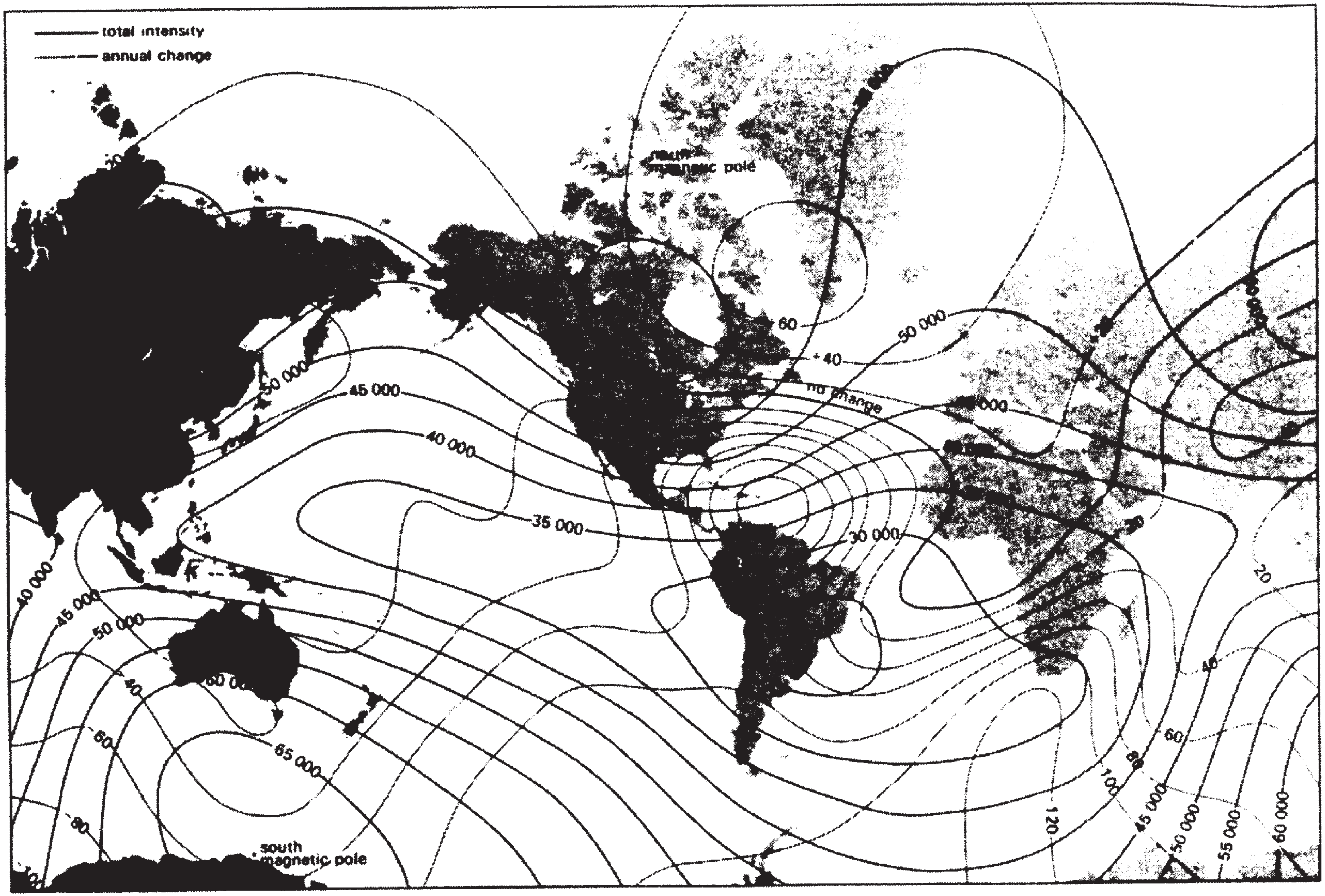

Figure 3.6. (Continued) (c) total field strength in nanotesla (heavy lines) and rate of change in nanotesla/year (light lines).

(c) Secular variations of the main field. Four hundred years of continuous study of the Earth's field has established that it changes slowly. The inclination has changed some 10° (75° to 65°) and the declination about 35° (10°E to 25°W and back to 10°W) during this period. The source of this wandering is thought to be changes in convection currents in the core.

The Earth's magnetic field has also reversed direction a number of times. The times of many of the periodic field reversals have been ascertained and provide a *magnetochronographic time scale*.

3.3.3. The External Magnetic Field

Most of the remaining small portion of the geomagnetic field appears to be associated with electric currents in the ionized layers of the upper atmosphere. Time variations of this portion are much more rapid than for the main "permanent" field. Some effects are:

1. A cycle of 11 years duration that correlates with sunspot activity.
2. Solar diurnal variations with a period of 24 h and a range of 30 nT that vary with latitude and season, and are probably controlled by action of the solar wind on ionospheric currents.
3. Lunar variations with a 25 h period and an amplitude 2 nT that vary cyclically throughout the month and seem to be associated with a Moon–ionosphere interaction.
4. Magnetic storms that are transient disturbances with amplitudes up to 1,000 nT at most latitudes and even larger in polar regions, where they are associated with aurora. Although erratic, they often occur at 27 day intervals and correlate with sunspot activity. At the height of a magnetic storm (which may last for several days), long-range radio reception is affected and magnetic prospecting may be impractical.

These time and space variations of the Earth's main field do not significantly affect magnetic prospecting except for the occasional magnetic storm. Diurnal variations can be corrected for by use of a base-station magnetometer. Latitude variations ($\approx$ 4 nT/km) require corrections only for high-resolution, high-latitude, or large-scale surveys.

3.3.4. Local Magnetic Anomalies

Local changes in the main field result from variations in the magnetic mineral content of near-surface rocks. These anomalies occasionally are large enough to double the main field. They usually do not persist over great distances; thus magnetic maps generally do not exhibit large-scale regional features (although the Canadian Shield, for example, shows a magnetic contrast to the Western Plains). Many large, erratic variations often make magnetic maps extremely complex. The sources of local magnetic anomalies cannot be very deep, because temperatures below ~ 40 km should be above the *Curie point*, the temperature ($\approx$ 550°C) at which rocks lose their magnetic properties. Thus, local anomalies must be associated with features in the upper crust.

3.3.5. Magnetism of Rocks and Minerals

Magnetic anomalies are caused by magnetic minerals (mainly magnetite and pyrrhotite) contained in the rocks. Magnetically important minerals are surprisingly few in number.

Substances can be divided on the basis of their behavior when placed in an external field. A substance is *diamagnetic* if its field is dominated by atoms with orbital electrons oriented to oppose the external field, that is, if it exhibits negative susceptibility. Diamagnetism will prevail only if the net magnetic moment of all atoms is zero when H is zero, a situation characteristic of atoms with completely filled electron shells. The most common diamagnetic earth materials are graphite, marble, quartz, and salt. When the magnetic moment is not zero when H is zero, the susceptibility is positive and the substance is *paramagnetic*. The effects of diamagnetism and most paramagnetism are weak.

Certain paramagnetic elements, namely iron, cobalt, and nickel, have such strong magnetic interaction that the moments align within fairly large regions called *domains*. This effect is called *ferromagnetism* and it is $\sim 10^6$ times the effects of diamagnetism and paramagnetism. Ferromagnetism decreases with increasing temperature and disappears entirely at the Curie temperature. Apparently ferromagnetic minerals do not exist in nature.

The domains in some materials are subdivided into subdomains that align in opposite directions so that their moments nearly cancel; although they would otherwise be considered ferromagnetic, the susceptibility is comparatively low. Such a substance is *antiferromagnetic*. The only common example is hematite.

In some materials, the magnetic subdomains align in opposition but their net moment is not zero, either because one set of subdomains has a stronger magnetic alignment than the other or because there are more subdomains of one type than of the other. These substances are *ferrimagnetic*. Examples of the first type are magnetite and titanomagnetite, oxides of iron and of iron and titanium. Pyrrhotite is a magnetic mineral of the second type. Practically all magnetic minerals are ferrimagnetic.

3.3.6. Remanent Magnetism

In many cases, the magnetization of rocks depends mainly on the present geomagnetic field and the magnetic mineral content. Residual magnetism (called *natural remanent magnetization*, NRM) often contributes to the total magnetization, both in amplitude and direction. The effect is complicated because NRM depends on the magnetic history of the rock. Natural remanent magnetization may be due to several causes. The principal ones are:

1. *Thermoremanent magnetization* (TRM), which results when magnetic material is cooled below the Curie point in the presence of an external field (usually the Earth's field). Its direction depends on the direction of the field at the time and place where the rock cooled. Remanence acquired in this fashion is particularly stable. This is the main mechanism for the residual magnetization of igneous rocks.
2. *Detrital magnetization* (DRM), which occurs during the slow settling of fine-grained particles in the presence of an external field. Varied clays exhibit this type of remanence.
3. *Chemical remanent magnetization* (CRM), which takes place when magnetic grains increase in size or are changed from one form to another as a result of chemical action at moderate temperatures, that is, below the Curie point. This process may be significant in sedimentary and metamorphic rocks.
4. *Isothermal remanent magnetization* (IRM), which is the residual left following the removal of an external field (see Fig. 3.2). Lightning strikes produce IRM over very small areas.
5. *Viscous remanent magnetization* (VRM), which is produced by long exposure to an external field; the buildup of remanence is a logarithmic function of time. VRM is probably more characteristic of fine-grained than coarse-grained rocks. This remanence is quite stable.

Studies of the magnetic history of the Earth (*paleomagnetism*) indicate that the Earth's field has varied in magnitude and has reversed its polarity a number of times (Strangway, 1970). Furthermore, it appears that the reversals took place rapidly in geologic time, because there is no evidence that the Earth existed without a magnetic field for any significant period. Model studies of a self-excited dynamo show such a rapid turnover. Many rocks have remanent magnetism that is oriented neither in the direction of, nor opposite to, the present Earth field. Such results support the plate tectonics theory. Paleomagnetism helps age-date rocks and determine past movements, such as plate rotations. Paleomagnetic laboratory methods separate residual from induced magnetization, something that cannot be done in the field.

3.3.7. Magnetic Susceptibilities of Rocks and Minerals

Magnetic susceptibility is the significant variable in magnetics. It plays the same role as density does in gravity interpretation. Although instruments are available for measuring susceptibility in the field, they can only be used on outcrops or on rock samples, and such measurements do not necessarily give the bulk susceptibility of the formation.

From Figure 3.2, it is obvious that k (hence μ also) is not constant for a magnetic substance; as H increases, k increases rapidly at first, reaches a maximum, and then decreases to zero. Furthermore, although magnetization curves have the same general shape, the value of H for saturation varies greatly with the type of magnetic mineral. Thus it is important in making susceptibility determinations to use a value of H about the same as that of the Earth's field.

Since the ferrimagnetic minerals, particularly magnetite, are the main source of local magnetic anomalies, there have been numerous attempts to establish a quantitative relation between rock susceptibility and Fe_3O_4 concentration. A rough linear dependence (k ranging from 10^{-3} to 1 SI unit as the volume percent of Fe_3O_4 increases from 0.05% to 35%) is shown in one report, but the scatter is large, and results from other areas differ.

Table 3.1 lists magnetic susceptibilities for a variety of rocks. Although there is great variation, even for a particular rock, and wide overlap between different types, sedimentary rocks have the lowest average susceptibility and basic igneous rocks have the highest. In every case, the susceptibility depends only on the amount of ferrimagnetic minerals present, mainly magnetite, sometimes titano-magnetite or pyrrhotite. The values of chalcopyrite and pyrite are typical of many sulfide minerals that are basically nonmagnetic. It is possible to locate minerals of negative susceptibility, although the negative values are very small, by means of detailed magnetic surveys. It is also worth noting that many iron minerals are only slightly magnetic.

3.3.8. Magnetic Susceptibility Measurements

(a) Measurement of k. Most measurements of k involve a comparison of the sample with a standard. The simplest laboratory method is to compare the deflection produced on a tangent magnetometer by a