7.1 GEOMAGNETISM

Fifty years after Wegener's proposal of continental drift failed to gain acceptance, it was revived in the 1950s, when a new type of data showed that continents had moved. The motions are measured with respect to two different references. Wegener argued that continents had moved both relative to each other, as shown by the fit of continents, and relative to the north and south poles, as shown by the glacial deposits. Confirmation of these ideas came from *paleomagnetism,* the study of the history of Earth's magnetic field. To understand these, we first consider some general principles of magnetism and then *geomagnetism,* the study of Earth's present magnetic field.

7.1.1 Magnetic concepts (some may move to chapter 2)

Magnetic fields are similar to, but in some crucial ways different, from the more familiar electric fields. Electric fields arise from charged particles, which are either positively or negatively charged. The electric force between two particles with charges q and Q a distance r apart is

$$\mathbf{F} = \frac{qQ}{4\pi\varepsilon_0 r^2}$$
(7.1.1)

where $\hat{\mathbf{r}}$ is a unit vector in the direction from one charge to the other and ε_0 is the permittivity of free space. The electric field, $\mathbf{E}$, is the electric force acting on a unit positive charge

$$\mathbf{E} = \frac{\mathbf{F}}{q} = \frac{Q}{4\pi\varepsilon_0 r^2} \ .$$
(7.1.2)

Often, we consider electric dipoles, two opposite charges a distance apart. For example, some chemical properties result from molecules acting as electric dipoles.

Although magnetic fields have a similar form to electric fields, they differ because there are no single magnetic "charges," known as magnetic poles. Hence the fundamental entity is the magnetic dipole arising from an electric current I circulating in a conducting loop, such as a wire, with area A (Fig. 7.1-1). The field is described as resulting from a *magnetic dipole* characterized by a *dipole moment*

$$\mathbf{m} = m\,\hat{\mathbf{n}} \qquad m = IA.$$
(7.1.3)

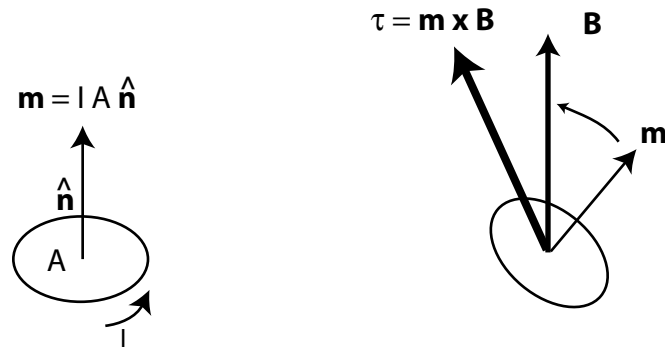Figure 7.1-1: *Left:* A magnetic dipole with moment **m** arising from an electric current I circulating in a conducting loop with area A and normal vector $\hat{\mathbf{n}}$. *Right:* Torque on a magnetic dipole due to magnetic field **B.**
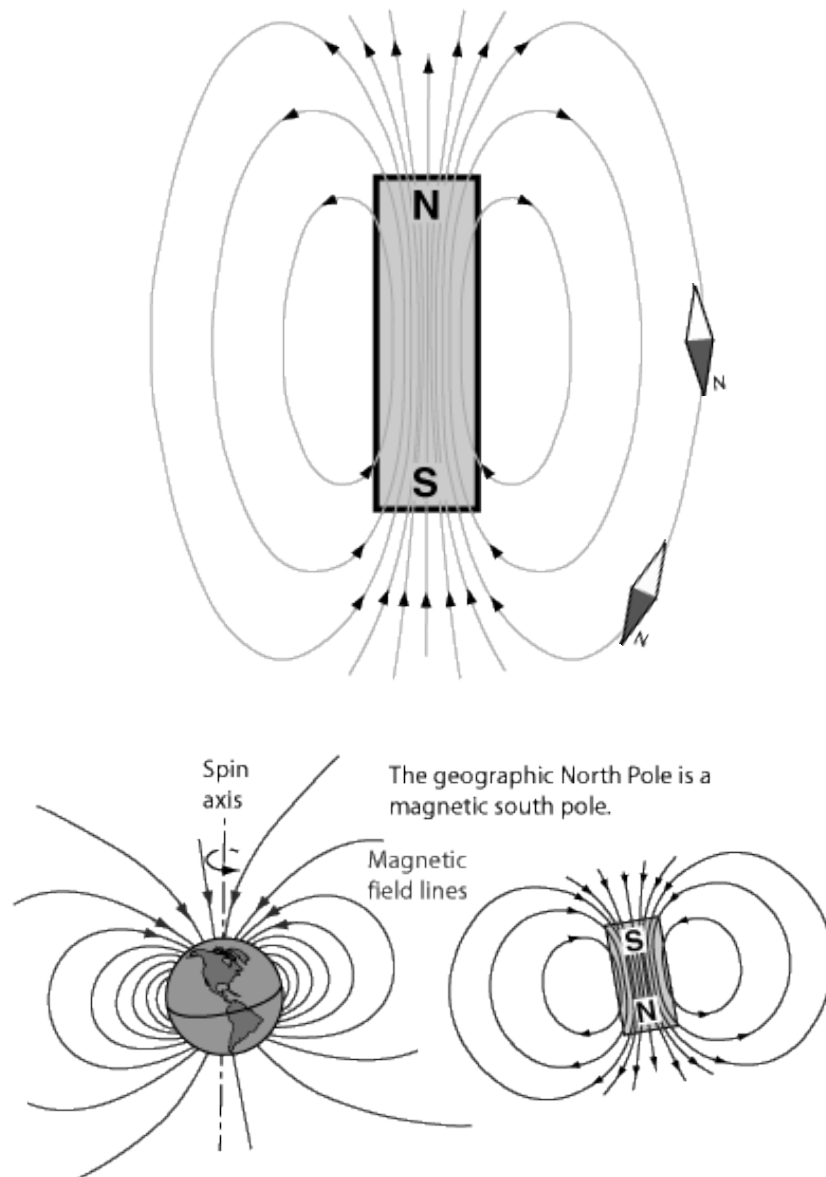

Figure 7.1-2: *Top:* Magnetic field lines for a bar magnet. *Bottom:* Field lines for the earth, treating the field as a dipole. (http://hyperphysics.phy-astr.gsu.edu)

This is a vector with magnitude *IA* pointing in the direction of the normal vector n̂ that is perpendicular to the loop and defined as positive from the direction of the current, using a right hand rule. Both the vector and its magnitude are called the dipole moment. Magnetic dipoles can arise from electric currents - which are moving electric charges - on scales ranging from wire loops to the hot fluid moving in the core that generates the earth's magnetic field. They also arise at the atomic level, where they are intrinsic properties of charged particles like protons and electrons. As a result, rocks can be magnetized, much like familiar bar magnets. Although the magnetism of a bar magnet arises from the electrons within it, it can be viewed as a magnetic dipole, with north and south magnetic poles at opposite ends.

The magnetic field can be defined in terms of the torque on a magnetic dipole in a way similar to defining the electric field from the force on a charge (Eqn 7.1.2). An object with magnetic dipole moment **m** experiences a torque (Fig. 7.1-1)

$$\tau = \mathbf{m} \times \mathbf{B} \tag{7.1.4}$$

that is the vector cross product of **m** with the magnetic field **B.** Because torque equals the rate of change of angular momentum **L** and the object's moment of inertia **I** is unchanged, its angular velocity ω increases

$$\tau = \frac{d\mathbf{L}}{dt} = \mathbf{I}\frac{d\omega}{dt} \ . \tag{7.1.5}$$

Thus the object rotates in a right handed way about an axis in the direction of the torque, which is perpendicular to both **m** and **B.** This rotation aligns the object's dipole moment with the magnetic field. Once these are aligned, the angle between them is zero, so the cross product is zero and the magnetic field exerts no torque.

Using this idea, we visualize the magnetic field of a dipole in terms of *magnetic field lines* pointing outward from the north pole of a bar magnet and in toward the south pole (Fig. 7.1-2). The lines point in the direction another bar magnet, such as a compass needle, would point. At any point, the north pole of the compass needle would point along the field line, toward the south pole of the bar magnet. The earth's magnetic field, as discussed shortly, looks essentially like a dipole field. A tricky

point is that the earth's field is that of a dipole pointing south along the earth's axis, so the magnetic

pole near the geographic North pole is a south magnetic pole. We call this the North magnetic pole.

because the north pole of a compass needle, which is a north magnetic pole, points toward it. (Figure

7.1-2).

An interesting complexity of magnetic fields is that because they can magnetize objects, the

total field we observe is the sum of both the external magnetic field and the field it produced in the

object. Hence the total magnetic field B is defined as

$$\mathbf{B} = \mu_o(\mathbf{H} + \mathbf{M}), \tag{7.1.6}$$

the sum of the *magnetizing field* **H** and the resulting *magnetization* **M**, times μ_o, a constant called the

magnetic permeability of free space.[1]

The magnetization M is the dipole moment m per unit volume. Its magnitude depends on how

easily the solid is magnetized, which is described by its *magnetic susceptibility,* χ, with

$$\mathbf{M} = \chi\mathbf{H}. \tag{7.1.7}$$

Because **M** and **H** are vectors, this equation states that rocks or other materials placed in a external

magnetizing field acquire a magnetization parallel to the magnetizing field with a strength propor-

tional to the susceptibility.

Using this definition, Eqn 7.1.5 can be written

$$\mathbf{B} = \mu_o(1 + \chi)\mathbf{H} = \mu_o\mu\mathbf{H} \tag{7.1.8}$$

where $\mu = (1 + \chi)$ is defined as the *magnetic permeability* of the material. Thus **B** and **H** are pro-

portional. Hence although measurements of the earth's magnetic field measure **B**, a map of **H** would

look like a map of **B**, with different units.

If the susceptibility is small, the permeability is about 1. This is the case in a vacuum, where χ

= 0, and in air where χ is small. In contrast, some rocks and minerals have significant susceptibility,

and so can be magnetized by the earth's magnetic field (Table 1).

---

[1] The distinction between the fields **B** and **H** is often confusing, especially because both are sometimes called
the magnetic field. **B** is also called the magnetic induction, and **H** is also called the magnetic field strength or
external field.

The choice of units to describe magnetic fields and their effects can also be confusing, because the form of equations like Eqn 7.1.6 depends on the units chosen. We use SI units, in keeping with (for better or worse) recent trends. In addition to the familiar meter, kilogram, and second, some unit must be chosen to describe electrical and magnetic quantities. In SI, this is the *ampere,* denoted A, which is a unit of electric current.[2] Substituting in Eqn 7.1.3 shows that the dipole moment has units of current times area, or A-m^2. Eqn 7.1.4 shows that the magnetic field **B** has the units of torque, force times distance or N-m, divided by those of the dipole moment, A-m^2. Thus **B** has units defined as teslas, T, where

$$1 \text{ T} = \text{N/A} - \text{m} = \text{kg/A} - \text{s}^2. \qquad (7.1.9)$$

The earth's magnetic field is about 50 μT, or 50×10^{-6} T. A common smaller unit is a *gamma* or nanotesla, with

$$1 \gamma = 10^{-9} \text{ T} = 1 \text{ nT}. \qquad (7.1.10)$$

For comparison, the cgs unit is the Gauss, equal to 10^{-4} T.

The units of other quantities are given by their definitions. The magnetization **M** is magnetic moment per unit volume, so its units are A-m^2/m^3 or A/m. **H** has the same units, and χ is dimensionless. μ_o, the permeability of free space, is $4\pi \times 10^{-7}$ N-A^2.

7.1.2 Earth's magnetic field

At a point on the earth's surface, the observed magnetic field **B** is described by its magnitude B and two angles giving its direction (Fig. 7.1-3). The vertical component, B_v, is defined as positive downward with

$$B_\text{v} = B \sin I, \qquad (7.1.11)$$

where I is the *inclination,* the field's angle from the horizontal, defined as positive downward. The horizontal component, B_h, is

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[2] Typical home electric circuits allow currents up to 20 A, or 20 "amps." For comparison, currents over 0.1 A can kill people.
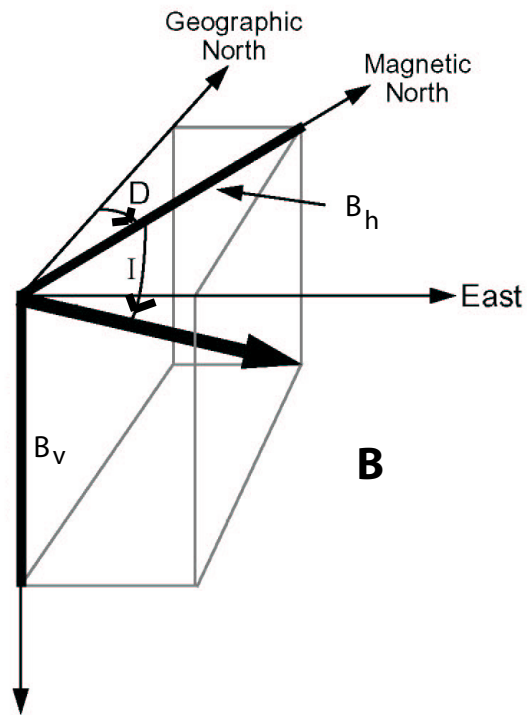
Figure 7.1-3: Geometry of the earth's magnetic field observed at a point on the surface. (Butler, 1992)
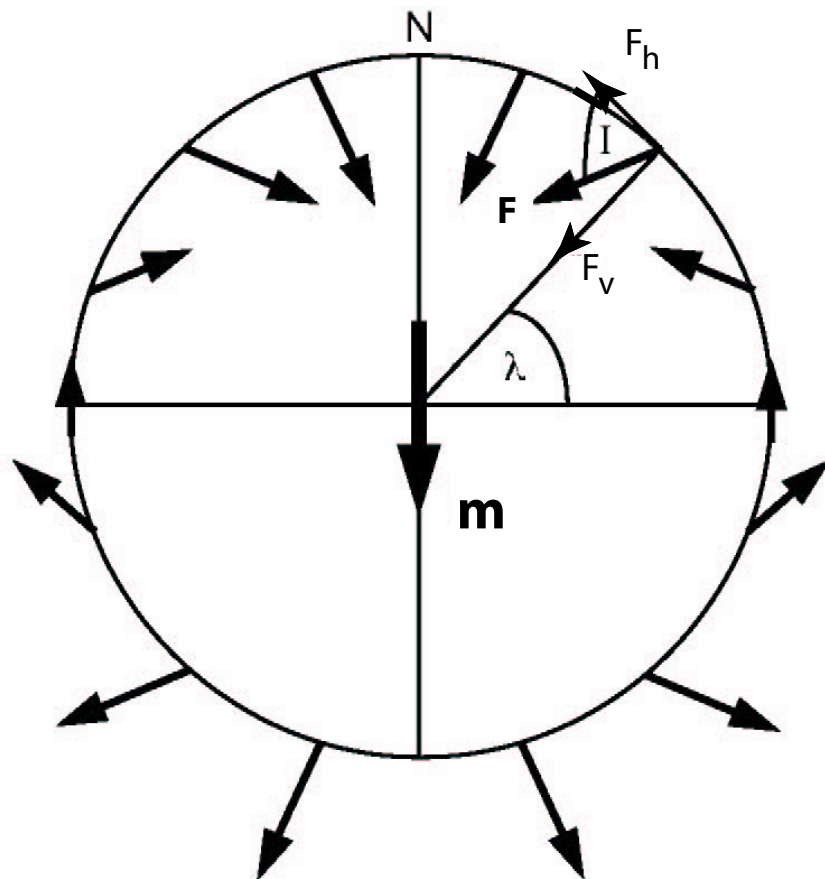

Figure 7.1-4: Geometry of the dipole magnetic field. (Butler, 1992)

$$B_h = B \cos I, \tag{7.1.12}$$

and can be decomposed into north and east components

$$B_N = B \cos I \cos D \quad \text{and} \quad B_E = B \cos I \sin D. \tag{7.1.13}$$

Here D is the field's *declination* angle, measured clockwise from north.

As discussed in Chapter 2, the earth's magnetic field is essentially that of an *axial geocentric dipole,* an imaginary bar magnet at the center of the earth and aligned along the rotation axis. In this approximation, the field points toward the north magnetic pole, which is at the geographic north pole. It thus has no east-west component, so the declination is zero. The dipole field **F** as a function of latitude λ (Fig. 7.1-4) can be resolved into a horizontal component

$$F_h = B_0 \cos \lambda \tag{7.1.14}$$

and a vertical component

$$F_v = 2B_0 \sin \lambda. \tag{7.1.15}$$

Thus the magnitude of the field is

$$F = \sqrt{F_h^2 + F_v^2} = B_0 \sqrt{1 + 3 \sin^2 \lambda} \tag{7.1.16}$$

where the factor

$$B_0 = \mu_0 m/4\pi a^3 \tag{7.1.17}$$

is the strength of the horizontal component at the magnetic equator ($\lambda = 0$). Here μ_0 is the magnetic permeability of free space, m is the earth's dipole moment, and a is the earth's radius.

The dipole field's inclination, or angle with the horizontal, is related to the latitude by

$$\tan I = F_v/F_h = 2 \tan \lambda, \tag{7.1.18}$$

which is called the *dipole equation.* Thus at the magnetic equator $\lambda = 0°$ so $I = 0°$, because the field is horizontal. At the north magnetic pole $\lambda = 90°$ and $I = 90°$, because the field points downward. Conversely, at the south magnetic pole $\lambda = -90°$ and $I = -90°$, because the field points upward. At intermediate latitudes the field has intermediate values with I positive in the northern hemisphere and

negative in the southern hemisphere. For example, at 45°N $\lambda = 45$ and $I = 63°$, whereas at 45° S $\lambda =$ -45° and $I = -63°$.

The real magnetic field is more complex than an axial geocentric dipole. Fig. 7.1-5 shows the magnetic field interpolated from data collected at magnetic observatories, and compiled as the International Geomagnetic Reference Field, or IGRF. The IGRF is revised regularly, because the field changes with time. Contrary to the predictions of Eqn 7.1.15, the field's magnitude (intensity) does not vary only with latitude, and the highest intensities are not at the north and south poles. Similarly, the declination - the angle relative to geographic north at which a compass needle points - varies significantly rather than being zero everywhere. This phenomenon is familiar to hikers, who correct their compass readings using the declination direction shown on topographic maps for the area they are in (Fig. 7.1-6).

A better approximation to the field uses the dipole that best describes the real magnetic field, which is currently inclined about 11.5° to the spin axis (Fig. 7.1-7). The positions where this dipole's axis intersects the Earth's surface, known as the *geomagnetic north and south poles,* are at 80°N, 72°W and 80°S, 108°E for the 2005 version of the IGRF. The equator of this dipole is the *geomagnetic equator.* This field is described by Eqns. 7.1.14-16, using the latitude relative to the geomagnetic poles.

Although the inclined dipole is a better description of the field, neither it nor any other dipole fully describes the field. The remaining part of the field, the *non-dipole field,* is about 10% of the total field. This portion causes several complexities. The actual locations where the field is horizontal define the *magnetic equator,* which differs from the geomagnetic equator. Similarly, the actual points where the field is vertical are the *magnetic north and south poles,* which are currently near 83°N, 114°W and 65°S, 139°E. These points move and so are determined by magnetic measurements made every few years.

The magnetic field changes with time, presumably because of changing fluid motions in the core. Changes on time scales less than 100,000 years are called *secular variation.* Fig. 7.1-8 (top) shows the changes in the declination and inclination observed over the past 300 years in Greenwich,
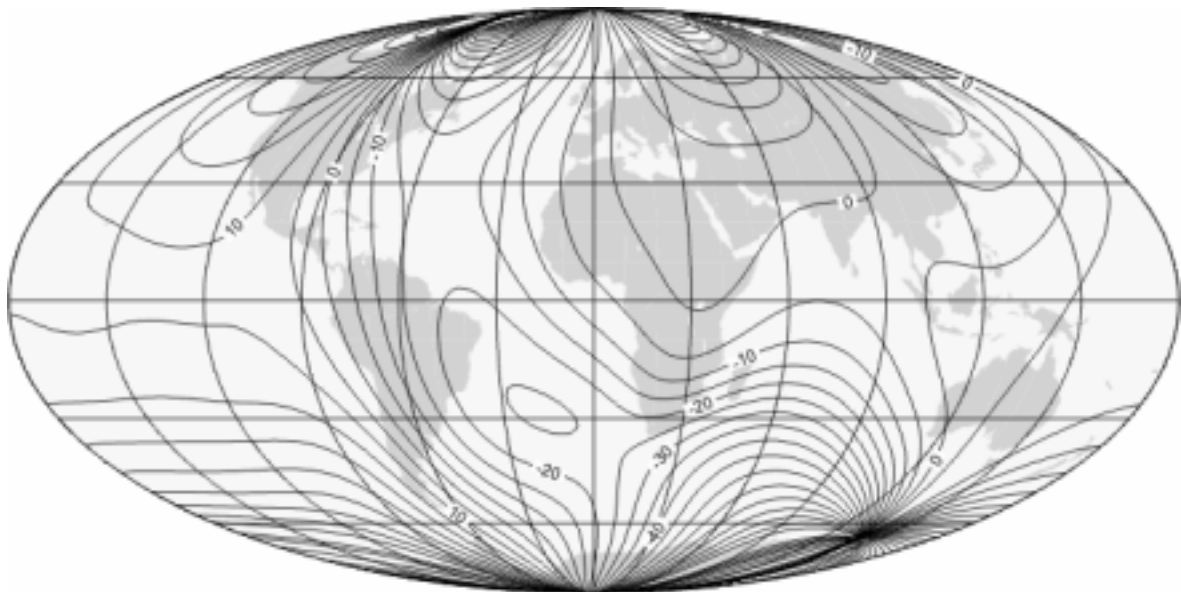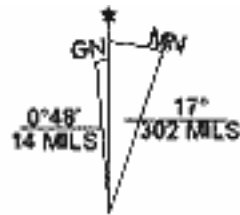
INTENSITY
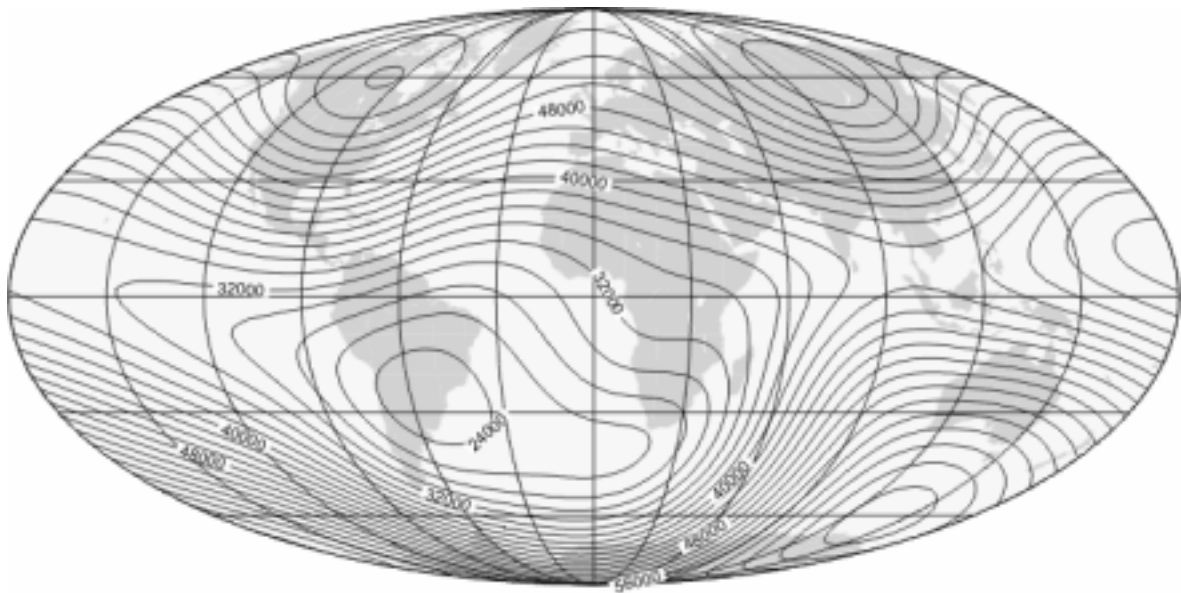
DECLINATION

Figure 7.1-5: The observed magnetic field shows deviations from a pure dipole field. (http://gsc.nrcan.gc.ca/geomag/field/index_e.php)
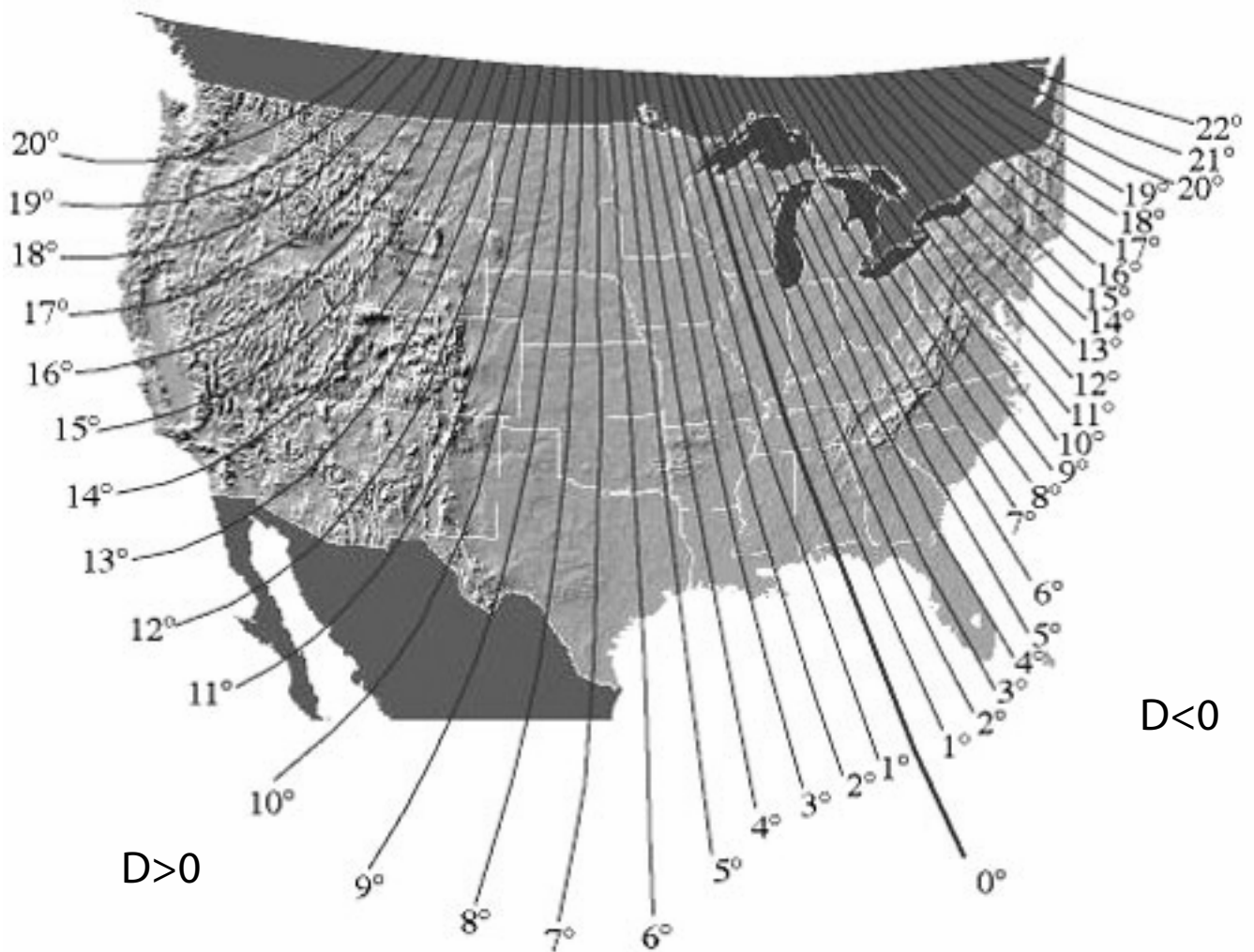
Figure 7.1-6: *Top:* Declination is shown on topographic maps by arrows showing the orientation of a compass needle (magnetic north) relative to geographic north. This example shows declination 17°. *Bottom:* Declination in the United States. Sites in the eastern part of the country have negative declinations. (http://geology.isu.edu/geostac)
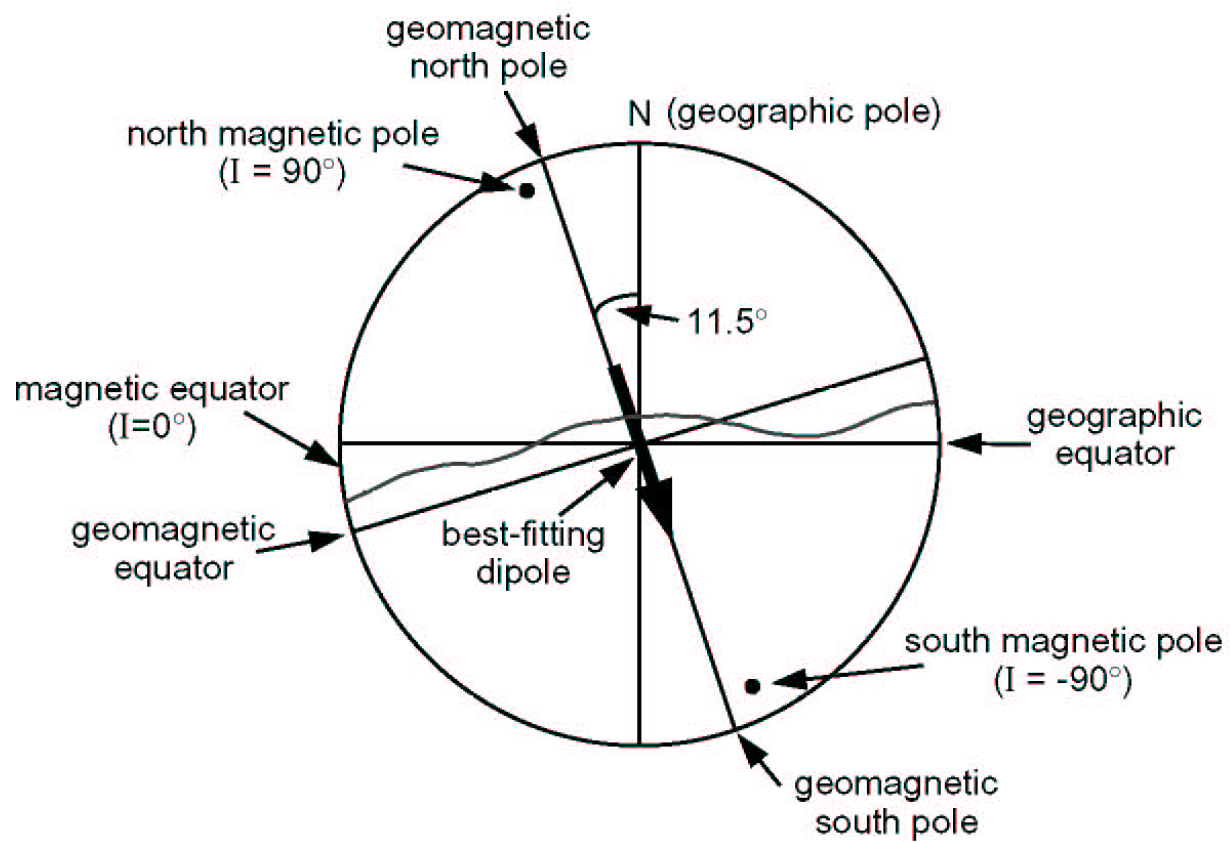
Figure 7.1-7: An inclined dipole model for the magnetic field. (McElhinny, 1973)
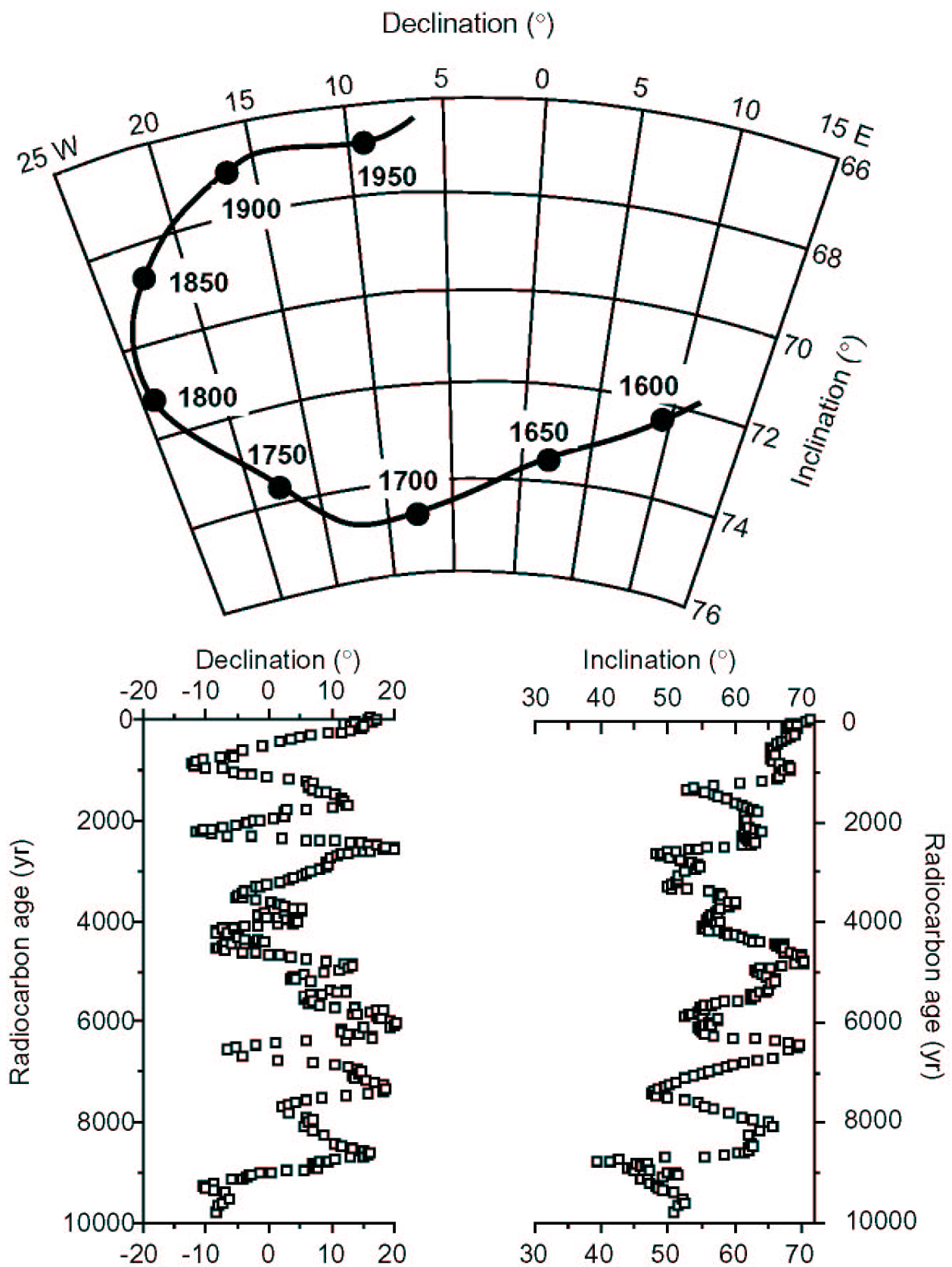
Figure 7.1-8: Secular variation of the magnetic field. *Top:* Short timescale observations in Greenwich, England. *Bottom:* Long timescale results from paleomagnetic data in Oregon. (Butler, 1992)

England. These values vary by 10-15°, which is significant. Long records from paleomagnetic data (Fig. 7.1-8, bottom) show similar changes over thousands of years. Much of the changes on timescales shorter than about 3000 years are due to variations in the non-dipole field, whereas changes on longer timescales are due to variations in the dipole field.

Fig. 7.1-9 shows that over the past 2000 years the location of the geomagnetic pole has varied about the geographic north pole. Each geomagnetic pole in this figure represents an average over 100 years. On such short timescales, the geomagnetic pole varies about Earth's spin axis and thus the North pole. The average over 2000 years, in contrast, is nearly coincident with the spin axis. In general when paleomagnetic data are averaged over tens or hundreds of thousands of years, the location of the resulting mean paleomagnetic pole corresponds to the axis of a dipole that is located at Earth's center and aligned with the spin axis. This observation is known as the *geocentric axial dipole hypothesis* and is the basis for using paleomagnetic data to determine the past positions of continentants and plates. This result is fortunate for geophysics, because it means that averaging pole positions determined over tens or hundreds of thousands of years gives a reasonable position consistent with an axial dipole, so the paleomagnetic pole can be treated as a geographic pole. If this were not the case, paleomagnetism could not reliably identify the effects of plate motions that occur over much longer times. Interestingly, although secular variation is a problem in studying past plate motions, it gives important constraints on how the field is generated within the core. This illustrates the principle that one scientist's signal is another's noise.
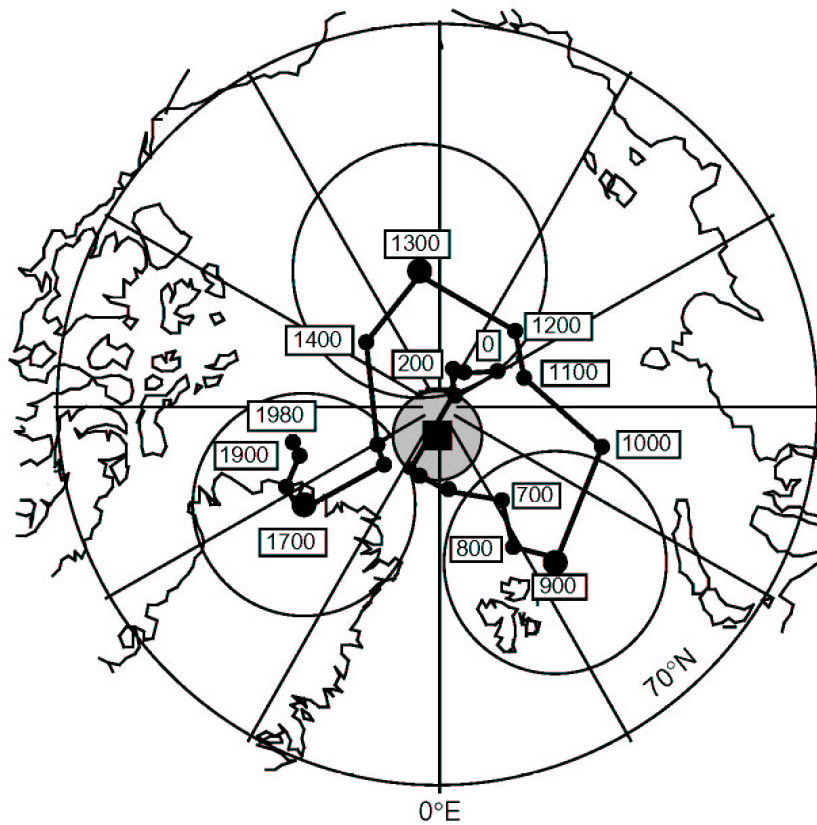
Figure 7.1-9: Positions of the north geomagnetic pole from paleomagnetic data averaged over 100 years, spanning the past 2000 years. The circles give uncertainties for each time interval, and the square gives the average over 2000 years. (Butler, 1992)

7.2 PALEOMAGNETISM

7.2.1 Rock magnetism

As we have seen in section 7.1, rocks or other materials placed in an magnetic field acquire a magnetic field parallel to and proportional to the external magnetic field. The strength of this magnetization is described by the magnetic susceptibility (Eqn 7.1.6).

Rocks can record the earth's magnetic field in several ways. The most important, called *Thermal Remnent Magnetism (TRM),* results when molten volcanic rock cools. Basalt and other volcanic rocks contain iron and titanium bearing minerals like magnetite (Fe_3O_4), hematite (Fe_2O_3), and ilmenite ($FeTiO_3$). These minerals are examples of materials known as *ferromagnetic,* whose atomic structure causes them to retain magnetization they acquired, even after an external field is removed. Such materials, typically containing iron, nickel, or cobalt, form the magnets we use on refrigerators and in other familiar applications. Ferromagnetism arises because electrons have both spins and magnetic moments. When electrons in atomic orbital shells are paired - have opposite spins - they have no net magnetic moment. In contrast, unpaired electrons have magnetic moments, so strongly ferromagnetic materials have many unpaired electrons. The resulting regions of parallel dipoles are called *magnetic domains.*

Normally, the domains are randomly oriented, so the material has no net magnetization. However, applying an external magnetic field aligns the domains and reorganizes them, giving the material a net magnetization (Fig. 7.2-1). This ordering persists after the external field is removed. However, the ordering is destroyed if the material is heated above its *Curie temperature,* because thermal oscillations overcome the domain alignments.

The process occurs in reverse as molten igneous rocks cool, first becoming solid at about 800-1100°C (Chapter 4), and eventually cooling through the Curie temperature of the magnetic minerals, about 600°C. Below the Curie temperature the minerals retain a magnetization that records the earth's field when and where they cooled, even when the earth's field changes and the rock is transported on a moving plate. Igneous rocks magnetized at midocean ridges and at volcanos on land
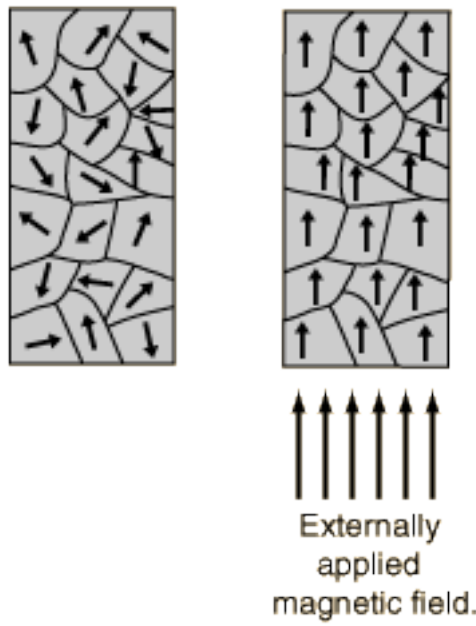
Externally
applied
magnetic field.

Figure 7.2.1: Magnetic domains in a ferromagnetic material are aligned and reor ganized by an exter-

nal magnetic field, causing a net magnetization. (http://hyperphysics.phy-astr.gsu.edu)

provide the primary data used to establish the history of the magnetic field and of plate motions.

Sediments can acquire magnetization in a different way, called *Detrital Remnant Magnetism (DRM).* This occurs when sediment is deposited in water still enough for magnetized grains to be aligned by the earth's field. Conceptually, this occurs because the earth's field exerts a torque that rotates grains' magnetic moments toward the direction of the magnetic field (Fig 7.1-1). How this actually occurs in the real world during the complex process of sediment deposition - that involves a variety of grains with different sizes, only some of which are magnetic, subject to a number of processes (Fig 7.2-2) including biological perturbations - remains an area of research. Paleomagnetic data from deep sea sediments provides important data about the evolution of the ocean basins.

Rocks can also be magnetized as a result of the growth of magnetic minerals, a process called *Chemical Remnant Magnetism (CRM).* The most common case involves the growth of hematite by oxidation reactions - rusting - in iron- rich sediments like sandstones. For example, iron oxidizes and combines with water to produce the mineral goethite, $FeO(OH)$, which then dehydrates

$$2FeO(OH) \rightarrow Fe_2O_3 + 2H_2O \tag{7.2.1}$$

to hematite. Once hematite grains reach a critical size, they are magnetized and record the earth's field. Because these "red bed" sediments are common in the continental rock record, their paleomagnetism is valuable for understanding the history of continents.

7.2.2 Paleomagnetic measurements

Paleomagnetic results require collecting rock samples of known age and determining the magnetic field within them. For sites on land, this is done by using a rock drill to collect a core sample, whose orientation is recorded (Fig. 7.2-3). Measurements are also made on samples drilled from the sea floor. A number of samples are collected at each site, in order to sample the paleomagnetic field over about 100,000 years and thus average out the effects of secular variation.

The sample's magnetic field is measured using a magnetometer. A complication is that samples often have secondary magnetization due to effects like lightening strikes on the rock, which can
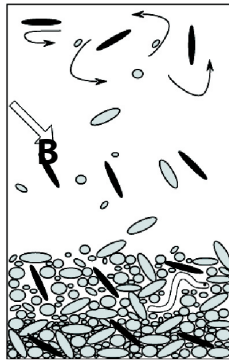
Figure 7.2-2: Schematic process causing detrital remnant magnetism: magnetized sediment grains are aligned by the earth's field. (http://www.science.siu.edu/geology/courses/geol535/geol535b.html)
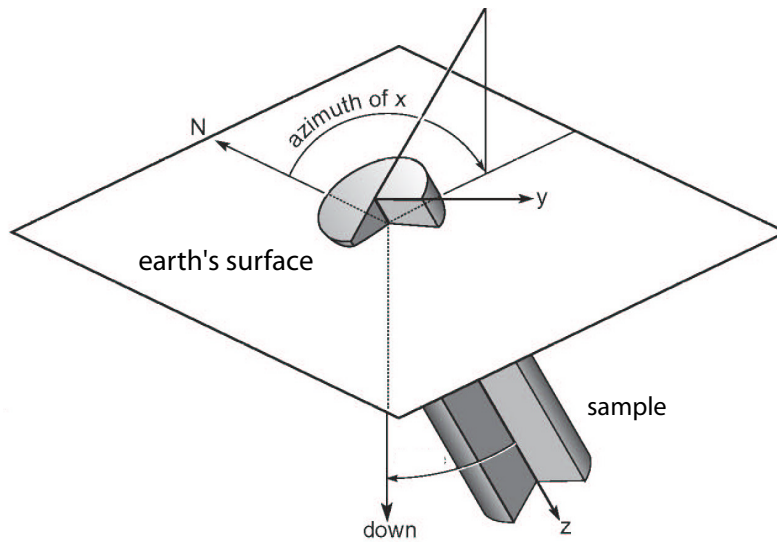

Figure 7.2-3: *Top:* Pioneering paleomagnetist Allan Cox using a portable drill to collect rock samples from the Galapagos islands in the early 1960's. (Cox, 1973) *Bottom:* The orientation of a paleomagnetic sample is recorded so laboratory measurements can be used to find the paleomagnetic field. (Butler, 1992).

obscure the primary magnetization.[1] This is removed by "cleaning" the samples by using techniques including heating them and subjecting them to alternating magnetic fields.

Various types of magnetometers are used to measure the magnetization of rock samples. The easiest to understand is the classic spinner magnetometer. This relies on Faraday's law that the voltage, V, generated in a wire coil is proportional to the rate of change of the magnetic flux through the coil

$$V = - \frac{d \; flux \; \mathbf{B}}{dt} \; .$$

(7.2.2)

The magnetic flux is the product of the coil's area and the component of the magnetic field perpendicular to that area, which can be written using the usual formulation for the flux of a vector field through a surface with area dA and normal vector $\mathbf{\hat{n}}$ (Fig. 7.2-4)

$$flux \; \mathbf{B} = \mathbf{B} \cdot \mathbf{\hat{n}} \; dA.$$

(7.2.3)

The rock sample is placed in a holder on a shaft that rotates, producing a changing magnetic flux through coils in the magnetometer, and thus a voltage proportional to the component of the rock's magnetization in the plane perpendicular to the shaft. Spinning the sample in different orientations gives all three components of magnetization.

Far more sensitive measurements are made using superconducting magnetometers that rely on the properties of superconducting materials, which are very different from those we encounter in daily life. Electrical currents in conventional materials follow Ohm's law

$$V = IR$$

(7.2.4)

in which the current I is conducted by electrons moving in response to a voltage V. The resistance R arises from the thermal vibrations of the material's crystal lattice, and so increases with temperature. In contrast, *superconductivity* is a phenomenon that occurs in some metals at extremely low temperatures, generally below 25°K (-248°C or -414°F), where electricity is conducted with no resistance. These supercurrents are due to the motion of Cooper pairs, pairs of electrons that are coupled due to distortions of the material's crystal lattice (Fig. 7.2-5).

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[1] In the same way, credit cards and other magnetic media can lose their magnetization.
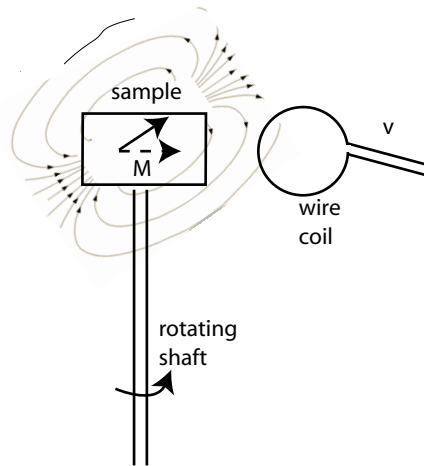
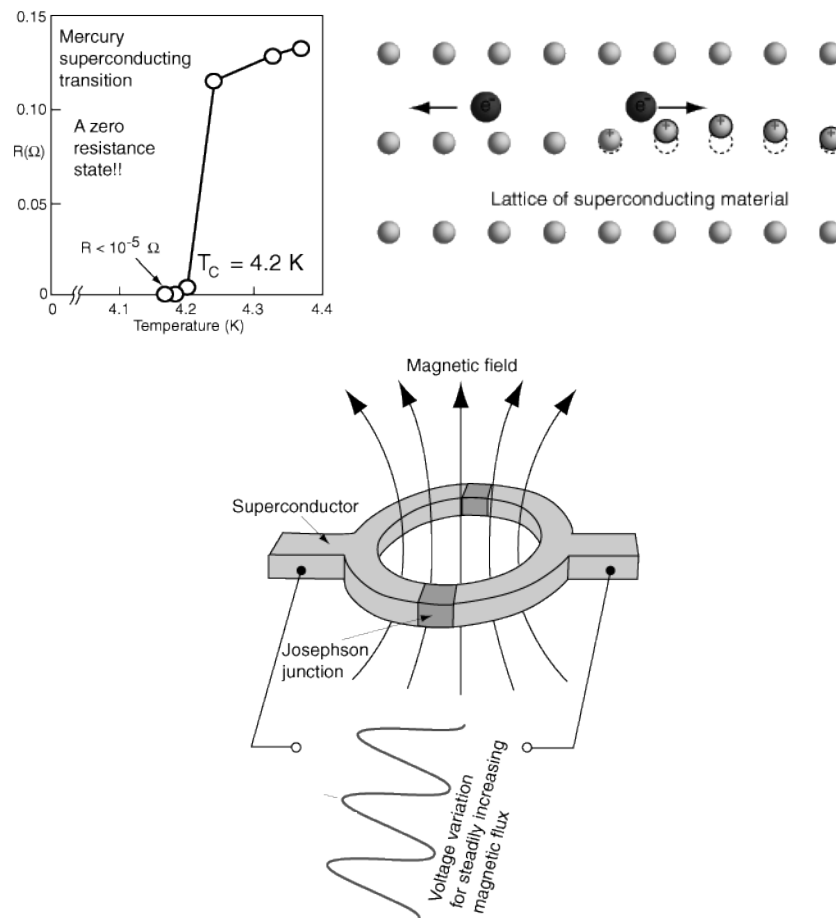Figure 7.2-4: Schematic sho wing workings of a spinner magnetometer.

Figure 7.2-5: Schematic principles of a superconducting magnetometer . *Top left:* At extremely low temperatures electricity is conducted with no resistance. *Top right:* Current in superconductor due to the motion of Cooper pairs, pairs of electrons that are coupled due to distortions of the materialís crystal lattice *Bottom:* Superconducting quantum interference device (SQUID) consisting of two superconductors separated by thin insulating layers to form two parallel Josephson junctions. (http://hyperphysics.phy-astr.gsu.edu)

Superconducting magnetometers are based on the properties of Josephson junctions, which consist of two superconductors separated by a thin insulating layer. Cooper pairs can "tunnel" across the junction due to their quantum mechanical probability of being on the other side, causing a current across the junction. The magnetometer is based on a superconducting quantum interference device (SQUID) consisting of two superconductors separated by thin insulating layers to form two parallel Josephson junctions. The voltage across the SQUID oscillates with the changes in phase at the two junctions, which depend upon the change in the magnetic flux and can thus be used to measure it. Superconducting magnetometers use three SQUIDS to measure all the components of the sample's magnetization. Although they are costly, because they require cooling to 4°K using liquid helium, their high sensitivities and fast measurements make them the systems of choice in paleomagnetic laboratories.

Magnetometers yield the three components of the sample's magnetization in a coordinate system aligned with the sample (Figure 7.2-3). In this coordinate system, the declination is

$$D = \tan^{-1}(M_y/M_x) \tag{7.2.5}$$

and the inclination is

$$D = \tan^{-1}\left(M_z/\sqrt{M_x^2 + M_y^2} \right) \tag{7.2.6}$$

where $\tan^{-1}$ is the inverse tangent function. These are then converted to geographic declination and inclination using the orientation of the core.

7.2.3 Apparent polar wander paths

The fact that rocks record the magnetic field where and when they were magnetized makes it possible to determine how they moved since they were magnetized, as illustrated in Fig. 7.2-6. Imagine a rock formed at the equator, which thus has a horizontal magnetic field. If it is moved to another place - for example 45°N - its magnetization will not be parallel to the field there. Thus the rock's magnetization shows that it is now 45° away from where it was magnetized. This can be phrased by saying that the paleomagnetic pole, or "apparent pole", inferred from the rock is 45° from the current
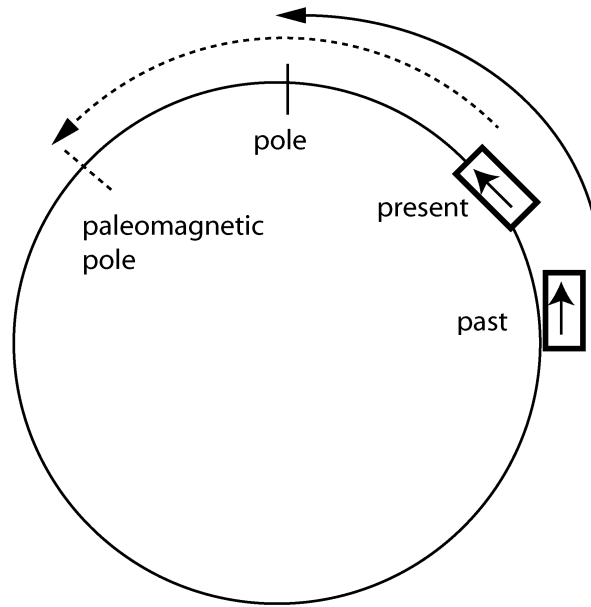
Figure 7.2-6: Motion of a rock magnetized at the equator mak es its paleomagnetic pole differ from the north pole.
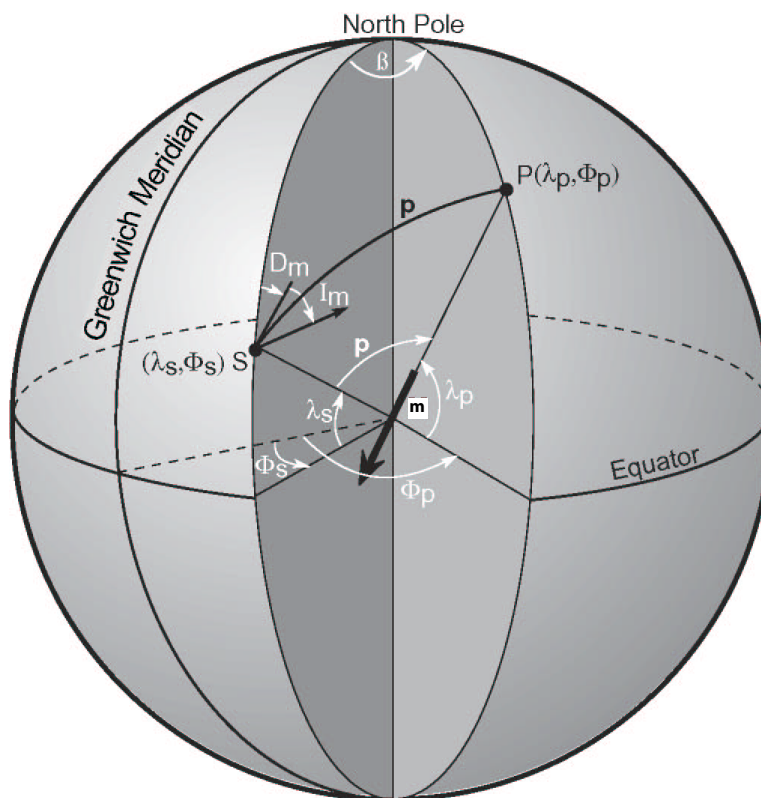

Figure 7.2-7: Geometry in volved in finding a paleomagnetic pole - in the present geographic coordinate system - from paleomagnetic data. (Butler, 1992).

North pole. This apparent polar wander results primarily from the rock, not the pole, having moved.

Using this idea, the apparent pole positions from rocks formed at different times, but on the same

plate, can be used to construct an *apparent polar wander (APW)* path.

Fig. 7.2-7 illustrates how a paleomagnetic pole is found from paleomagnetic data.[2] If rocks

now at site with latitude λ_s and longitude ϕ_s have mean (average) declination D_m and inclination I_m,

then if they were magnetized in an axial dipole field, that field's north pole in the present coordinate

system would be at latitude λ_p and longitude ϕ_p. As usual, we define latitude as increasing from -90°

at the south pole to 90° at the north pole, and longitudes east of the Greenwich meridian increase

from 0° to 360° (or equivalently, west longitudes are negative).

Using Eqn 7.1.18 gives the paleomagnetic colatitude, p, which is the great circle distance from

the site to the pole

$$p = \cot^{-1}\left(\frac{\tan I_m}{2}\right) = \tan^{-1}\left(\frac{2}{\tan I_m}\right) \tag{7.2.7}$$

and the present latitude of the paleomagnetic pole is

$$p = \sin^{-1}(\sin \lambda_s \cos p + \cos \lambda_s \sin p \cos D_m) . \tag{7.2.8}$$

The difference in longitudes between the pole and the site is

$$\beta = \sin^{-1}\left(\frac{\sin p \sin D_m}{\cos \lambda_p}\right) \tag{7.2.9}$$

and the pole longitude is

$$\phi_p = \phi_s + \beta \quad \text{if} \quad \cos p \geq \sin \lambda_s \sin \lambda_p \tag{7.2.10}$$

or

$$\phi_p = \phi_s + 180° - \beta \quad \text{if} \quad \cos p < \sin \lambda_s \sin \lambda_p. \tag{7.2.11}$$

Paleomagnetists have determined APW paths from rocks from different areas. One of the most

famous results (Fig. 7.2-8) showed that APW paths for Europe and North America over time are dif-

ferent. Because the magnetic pole can only be in one place at any given time, this shows that the two
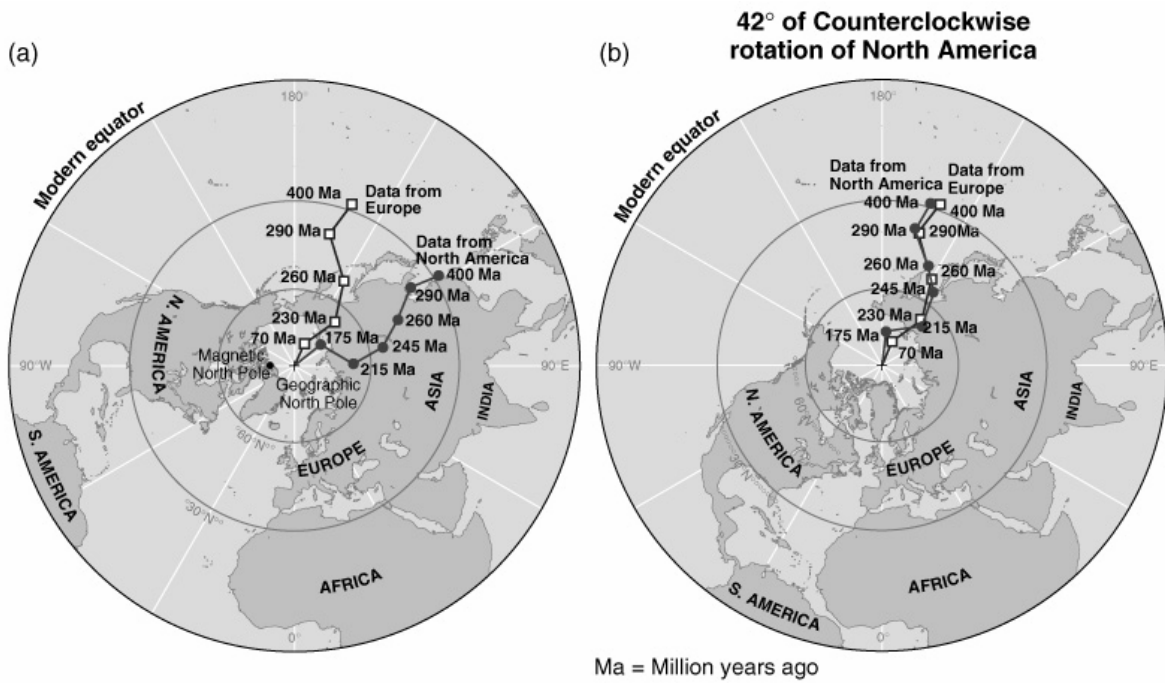
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[2] for this derivation, see Butler (1992).

Figure 7.2-8: The different APW paths for Europe and North America over time can be reconciled by correcting for the opening of the Atlantic. (Davidson et al., 2002)
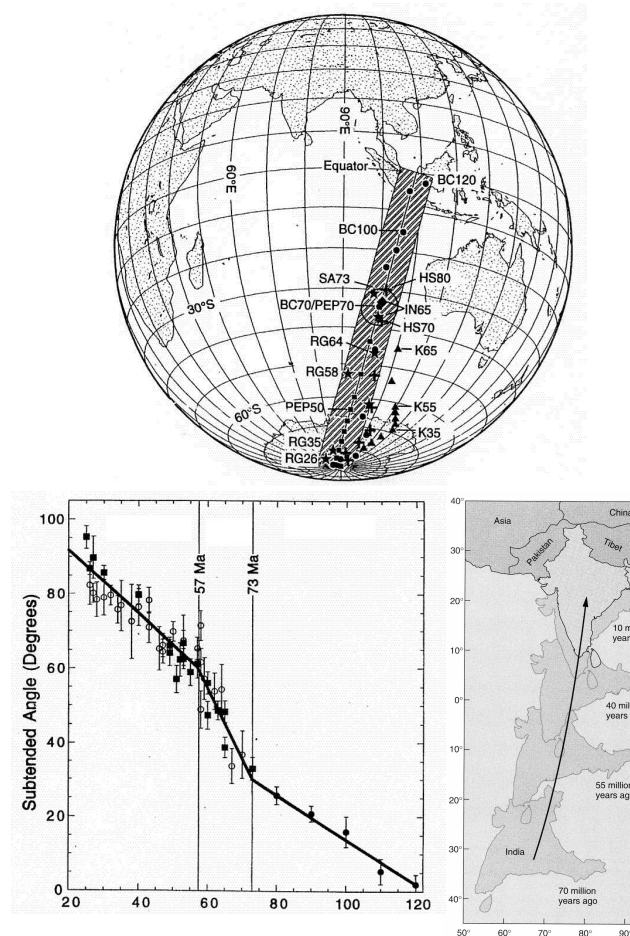

Figure 7.2-9: Apparent polar wander path for the Indian plate, showing the northward motion of India (Acton, 1999).

continents must have moved relative to each other. To test this, paleomagnetists reanalyzed their data assuming that Europe and North America had drifted apart as the Atlantic Ocean opened. Correcting for this drift, the APW paths are the same, confirming the existence of continental drift. This result agreed with the fit of the coastlines (Fig. 6.1-1), which became even more compelling when the continental shelves - the geological edges of the continents - were used rather than the coastlines.[5] Moreover, fitting the coastlines brought rocks of similar ages together.

Figure 7.2-9 shows another striking example, the APW path for the Indian plate. The apparent position of the south pole has moved steadily southward since 120 Ma, showing that India moved northward. The rate of northward motion can be found by plotting the angle between the apparent latitude of the pole and the true pole, as a function of age. The slope of this plot shows that the rate of motion slowed about 57 Ma, as India began colliding with Eurasia, but that northward motion continued as the collision continued to the present.

Although most of the apparent polar wander is due to plate motions, some appears to due to *true polar wander (TPW)* in which the magnetic pole has moved relative to the earth's spin axis. We will return to this issue later.

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[5]The difference is that the coastlines depend on sea level and thus the volume of water in the oceans, and so differ between ice ages - when some of the water is stored in glaciers - and warm periods like today.