12.710 Marine Geology and Geophysics
Geomagnetism, Paleomagnetism and Magnetostratigraphy

Fowler: Chapter 3
Stacey: Chapter 7
Kennett: Chapter 7

Introduction

It is possible to use magnetization of rocks in Earth's magnetic field to find the age of the seafloor. This lecture discusses the characterization of the main and secular magnetic fields and the concept of magnetostratigraphy.

The Main Field and Deviations

Systematic mapping of the magnetic field began in the 1400's by early explorers who traversed the Atlantic Ocean in search of the East Indies. The nature of the magnetic field was of significant practical interest because it was recognized that magnetic observations could be used to assist in navigation.

Evidence that the Earth behaved as a giant magnet was first presented by **William Gilbert**, who was, during the early 1600's, the physician to Queen Elizabeth I. Gilbert's analysis was the first documented quantitative study of **geomagnetism**, and for his contribution his name has been assigned to one of the paleomagnetic polarity epochs. The mathematical explanation that the field was generated internally was not published until 1838 by **C.F. Gauss**. Of course the notion that the field was maintained by electrical currents in the liquid outer core occurred much later than that.

To a first approximation the geomagnetic field is a **dipole field**, with field lines that emanate from the south geomagnetic pole and enter at the north geomagnetic pole. (Note that the north pole corresponds to the south end of magnet and vice versa -- because opposites attract.)

The magnetic field lines are normal to lines of equipotential. The pattern of field lines is consistent with a magnetic dipole at approximately the center of the Earth. But a perfect dipole would align itself with the rotation axis, and that isn't observed, so the Earth's field is not a perfect dipole, but it is close to one. The dominant dipole nature of the Earth's magnetic field raised the obvious thought that the interior of the Earth contained a gigantic bar magnet. But there are several characteristics that are inconsistent with the Earth's interior containing a solid magnet:

(1) there are **non-dipole components of the field** (though small compared to the dipole). These non-dipole contributions vary in amplitude by ~10 nT yr^{-1}; and drift westward at a rate of 0.2° yr^{-1};
(2) it has been known for 400 years that there are secular variations and in particular a **westward drift** of the dipole field of 0.05° yr^{-1}; and
(3) the field switches **polarity** over geologic time. There is no obvious pattern to these polarity reversals, and the period varies.

All of these observations are consistent with temporal variations in magnetization and any mechanism for magnetic field generation must allow for time variations. As seismology began to show evidence for a fluid outer core within the Earth, theoretical work converged to indicate that a **self-sustaining dynamo**, generated by motions in the fluid core, was the most plausible mechanism of magnetic field generation.

The Geomagnetic Dynamo

The differential motion of electrically conductive fluid in the Earth's outer core couples to the rotation to give rise to a self-sustaining magnetic field, a process called the **magnetic dynamo** process, which is described by principles of **magnetohydrodynamics**. After decades of research, satisfactory models of this phenomenon are only now beginning to be achieved.

As we shall discuss, we do know that Earth's magnetic field is not simply the result of permanent magnetism in its rocks, because the **paleomagnetic record** shows that field has reversed its polarity many times, on timescales of 10^3-10^6 years. Prolonged periods of predominantly one polarity are called **epochs**, with the most recent ones named after prominent geomagnetic researchers; shorter periods of reversals within epochs are called **events** and are named after places that contain rocks that led to their discovery. The longest epoch occurred during the Permian period and lasted for 50 Ma.

The shortest events are so short that they it is sometimes argued that the field does not completely reverse everywhere on Earth (at least for some reversals). It is possible that the field reversed in some areas and only weakened in others. Clearly the process of reversals is complex.

The mechanism for the generation and sustanence of the geomagnetic field has its basis in the **frozen flux principle**. The essential element of this principle is that if a combination of the conductivity, magnitude and velocity of fluid motion is "sufficient", then a magnetic field can be transported and deformed with the fluid. In this mechanism **the kinetic energy of fluid motion is converted into magnetic field energy**.

There are **two kinds of motions**. The first is differential rotation within the core that is a consequence of the tendency of a convecting fluid to conserve angular momentum as it is transported radially in response to buoyancy forces. Thus the inner part of the core rotates with a higher angular velocity than the outer part and it draws out the lines of a **poloidal** field to produce an additional **toroidal** field. The two different kinds of motion re-enforce each other, leading the dynamo to be self-sustaining.

A simple mechanical conceptualization for the self-sustaining dynamo has been demonstrated experimentally. It is given by two interconnected disk dynamos. Each disk rotates in an axial field produced by a coil carrying a current driven by the other disk. Rotation in the field generates an electromotive force between the perimeter of the disk and the axis, providing the current that drives the other disk. The **system is self-maintaining when current generation exceeds the ohmic dissipation via an appropriate combination of velocity, size and loop conductance**. The model is symmetrical with respect to the polarities of the currents and fields, and that the system works equally well if the polarities are reversed. Instabilities in the system, including spontaneous current reversals, have been demonstrated in simple mechanical systems and computer models of more complex systems. This effect may be an analog for the record of magnetic field reversals that is preserved in the geologic record.

Mathematical Representation of the Internal Field

The potential of an arbitrary geomagnetic field is usually represented as an infinite series of **spherical harmonic** functions. Spherical harmonics are analagous to fourier transforms on a sphere and they represent the solution of **Laplace's equation**

$$\nabla^2 V_m = 0 \qquad (7)$$

which, in spherical polar coordinates is expressed

$$\nabla^2 V_m = \frac{1}{r^2}\frac{\partial}{\partial r}\left(r^2 \frac{\partial V_m}{\partial r}\right) + \frac{1}{r^2 \sin\theta}\frac{\partial}{\partial \theta}\left(\sin\theta \frac{\partial V_m}{\partial \theta}\right) + \frac{1}{r^2 \sin^2\theta}\frac{\partial^2 V_m}{\partial \lambda^2} = 0. \qquad (8)$$

Spherical harmonic solutions to (8) effectively separate out the dipole and non-dipole components of the magnetic field. A general solution of (8) for the internal field can be written

$$V_m = \frac{a}{\mu_o}\sum_{l=1}^{\infty}\left(\frac{a}{r}\right)^{l+1}\sum_{m=0}^{l}\left(g_{lm}\cos m\lambda + h_{lm}\sin m\lambda\right)P_{lm}(\cos\theta) \qquad (9)$$

where g_{lm} and h_{lm} are the spherical harmonic coefficients or "Gauss coefficients" of the field. These coefficients have the dimensions of the field and usually have amplitudes of nanotesla = 10^{-9} T= 10^{-5} Gauss. Note that (9) has no *l=0* term, which would correspond to a magnetic monopole. The *l=1* term is the longest wavelength component of the field and it corresponds to the dipole. All other terms represent non-dipole contributions to the field.

Strength and Pattern of the Main Field

The **magnetic or dipole moment** *(m)* is a product of pole strength and separation. The best-fitting dipole has a moment m_{Earth}=7.856x10^{22} A m^2 and an orientation of 11.5° with respect to the Earth's spin axis.

The field of a dipole is typically represented in terms of the **scalar magnetic potential** V_m, where

$$V_m = -\frac{\bar{m}\cdot\bar{r}}{4\pi r^3} = -\frac{m\cos\theta}{4\pi r^2} \qquad (1)$$

where $\bar{r}$ is the radius vector and θ is the angle between the dipole axis and $\bar{r}$ to the point considered (*i.e.* the co-latitude with respect to the dipole axis), and the negative sign indicates that *m* is aligned along the -*z* axis. The **magnetic field** is

$$B(r) = -\mu_o \nabla V_m(r) \qquad (2)$$

where μ_o (=4πx10^{-7} kg m A^{-2} s^{-2}) is the **magnetic permeability of free space** and

$$\nabla = \hat{e}_r \frac{\partial}{\partial r} + \hat{e}_\theta \frac{1}{r}\frac{\partial}{\partial \theta} + \hat{e}_\lambda \frac{1}{r\sin\theta}\frac{\partial}{\partial \lambda} \qquad (3)$$

is the gradient in spherical coordinates where $\hat{e}_i$ are unit vectors. At the Earth's surface the **inclination** *(I)* is the angle between the magnetic field and the horizontal, and can be expressed

$$\tan I = \frac{Z}{H} = 2\tan\phi \qquad (4)$$

where $\phi=\pi/2-\theta$ is the **magnetic latitude at the time the rock cooled through the Curie temperature**,

$$H = -B_o \sin\theta$$
$$Z = -2B_o \cos\theta \qquad (5)$$

which are the **horizontal and radial components** of the field at the Earth's surface,

$$B_o = \frac{\mu_o}{4\pi}\frac{m_{Earth}}{a^3} = 3.03x10^{-5}T = 0.30G \qquad (6)$$

and *a* is the radius of the surface. Equation (4) can be used to tell how far from the pole a rock was. Equation (6) is the strength at the equator of the best-fit dipole field. The SI unit of field intensity is the **Tesla (T)**, where **1 T= 10^4 Gauss**.

External Influences

The magnetic field away from the Earth is strongly influenced by the sun.
Solar wind – the constant stream of ionized particles emitted by sun confines magnetic field to **magnetosphere**; deforms field lines so field lines form a **magnetotail**.
Bow shock – the shock wave produced where solar wind is slowed by interaction with the magnetic field.
Van Allen Belts -- zones of charged particles trapped by the Earth's magnetic field.

Paleomagnetic Poles

If the angles of declination and inclination of rock samples are known, the position of the **paleomagnetic pole** can be calculated. This is the pole at earlier times, different from the present due to the motions of the tectonic plates. One must use the laws of cosines and sines in spherical geometry to do the calculation. We find the **geographic latitude** (ϕ_p) **of the paleomagnetic north pole** *P* to be

$$\cos\left(\frac{\pi}{2}-\phi_p\right) = \cos\left(\frac{\pi}{2}-\phi_S\right)\cos\left(\frac{\pi}{2}-\phi\right) + \sin\left(\frac{\pi}{2}-\phi_S\right)\sin\left(\frac{\pi}{2}-\phi\right)\cos D \qquad (10)$$

where ϕ_s is the geographic latitude of the sample, ϕ is the paleolatitude given by (4), and *D* is the measured **magnetic declination** or the azimuth of the horizontal component of the magnetic field. We may also write the **difference between the longitudes of the paleomagnetic north pole and the sample location** ($\lambda_p - \lambda_s$)

$$\sin(\lambda_p - \lambda_s) = \frac{\sin\left(\frac{\pi}{2}-\phi\right)\sin D}{\sin\left(\frac{\pi}{2}-\phi_p\right)} = \frac{\cos\phi\sin D}{\cos\phi_p} \quad for \sin\phi \geq \sin\phi_p \sin\phi_s$$
$$\sin(\pi + \lambda_s - \lambda_p) = \frac{\cos\phi\sin D}{\cos\phi_p} \quad for \sin\phi < \sin\phi_p \sin\phi_s \qquad (11)$$

Hence, from measurements of rock magnetization and equations (4), (10) and (11), past positions of the magnetic pole can be found.

Polar Motion

Observations of the Earth's rotation axis via astronomical or space geodetic methods (*e.g.* very long baseline interferometry) show that the Earth's pole is moving in the direction 285° E at a rate of approximately 10 cm yr^{-1}. This yields a change in the orientation of the pole with respect to the celestial sphere. (In practice, what is moving are locations on the Earth, rather than the spin axis, that is moving.) This motion is due to changes in the Earth's moments of inertia due to mass re-distributed due to post-glacial rebound associated with unloading past ice caps. There are also various "wobbles" associated with dynamical phenomenon (winds, atmosphere-ocean coupling, motions at the core-mantle boundary) with 12- and 14-month periods that are superposed on the longer term drift. Collectively, the motion of the pole with respect to the celestial sphere is referred to as **polar motion**. Polar motion cause movement of the surface of the earth with respect to the celestial reference frame of a few km. Polar motion is a transient phenomenon compared to the longer term drift of the plates with respect to the geomagnetic pole, but it obscures paleomagnetic motions on short time frames. The typical rate of polar wander is about 0.3° per million years and this dictates that it is difficult to isolate paleopole movements from movements due to polar motion for rocks younger than about 20 Ma.

Rock Magnetization

Many common rocks contain small amounts of iron. Some rocks are inherently magnetic and others can become magnetized by the magnetic field. The latter has permitted the past movements of plates to be tracked.

In a material that has a tendency to become magnetized, **domains** are randomly aligned in the absence of a magnetic field. In such a case there is no magnetic moment because magnetic domains cancel out.

As a volcanic rock cools, it passes through a series of critical temperatures at which iron grains acquire **spontaneous magnetization**. These are called **Curie points** or temperatures, and they are different for different minerals. Here are a couple of common ones:

- magnetite (Fe_3O_4) = 580° C
- hematite (Fe_2O_3) = 680° C.

Also of interest is the **blocking temperature**, which is tens of degrees less than the Curie temperature for most minerals. Below this magnetized grains can't re-orient and on average remain with their magnetic moments parallel to the Earth's magnetic field lines at the time. Both the Curie and blocking temperatures are significantly less than the temperatures at which lavas crystallize (800° -1100° C). Thus magnetization becomes permanent some time after magmas solidify. This process is called **thermoremnant magnetization (TRM).**

Sedimentary rocks can acquire remnant magnetization even though they have never reached temperatures of 500° C:

- **depositional remnant magnetization (DRM)** - occurs during deposition; any previously magnetized grains deposited in still water will align themselves with their magnetic moments parallel to the Earth's field.
- **chemical remnant magnetization (CRM)** - acquired *in situ* after deposition during chemical growth of iron oxide grains; when the grains reach a critical size the magnetic field

required to change their magnetization is stronger than Earth's' field, so the magnetization is essentially permanent.

The degree to which a rock can be magnetized by an external field is determined by its **magnetic susceptibility** (χ). A rock placed in the Earth's magnetic field **B(r)** where **r** is radius follows the relationship

$$\mu_o M(r) = \chi B(r) \qquad (12)$$

where **M(r)** is the **induced magnetization** of the rock when it is exposed to the Earth's magnetic field **B(r)**. Values of χ for basalt range from 10^{-4}-10^{-1}. TRM intensities are somewhat stronger; 10^{-3}-10^{-1}.

Magnetic Anomalies and Seafloor Spreading

To use magnetic field measurements to ascertain the magnetization of the crust, it is necessary to subtract out the regional field; what's left is the **magnetic anomaly.** Magnetic anomalies are variations in the intensity or magnitude of the magnetic field. Over the oceans magnetic field measurements are made by dragging a **magnetometer** behind a ship. This is relatively easy to do so many magnetic observations of the seafloor have been collected. Plots show alternating stripes of positive and negative values. They exhibit mirror symmetry across mid-ocean ridges, are offset by fracture zones, and are a few tens of km in width. Thus magnetic anomaly maps are useful in delineating ridge axes and fracture zones.

The origin of the magnetic stripes first interpreted correctly in 1963 by **F.J. Vine** and **D.H. Matthews**. They realized this pattern of striping to be a consequence of **seafloor spreading**. Oceanic crust formed at mid-ocean ridges as basaltic intrusions, acts as a magnetic tape recorder, preserving past reversals of the magnetic field. Vine and Matthews assumed that the depth of magnetization is related to the Curie Point isotherm but the situation is more complicated than this. Analysis has shown that much of the magnetization resides in oceanic crustal Layer 2, and more specifically in the uppermost 200-500 m of Layer 2, where sampling has shown high magnetic intensities. But the rocks in this region cannot account for the observed intensity of magnetization. It is now thought that a significant fraction of the anomaly amplitude may originate in gabbroic Layer 3. Materials in this region is more weakly magnetized than those in Layer 2 but their significantly greater volume may compensate for this.

It is now known that the **width of a magnetic stripe is determined by the rate of plate motion (half spreading rate) times the length of time between magnetic reversals**. To interpret a magnetic anomaly pattern one needs to know either when reversals occurred or what the half-spreading rate of the ridge was. The interpretation of seafloor striping as due to polarity reversals was facilitated significantly by the fact that the mid-ocean ridge in the Atlantic is aligned normal to the path taken by ships traversing between North America and Europe.

Block models of oceanic crust have been commonly used to determine the plate velocity that best fits the magnetic anomaly pattern. These models are used to determine the magnetic intensity across a block of uniformly magnetized material. In these models the block has finite width but infinite length. Detailed analysis has shown that such a representation is overly simplistic and the magnetic structure of the crust is more complex than would be suggested by this model representation. There is evidence for:

- large systemic variations of inclinations from expected dipole values,

- large variations in intensity, and
- significant lateral heterogeneity.

It is known that the magnetic properties of basalt undergo aqueous alteration, but stable remanence is nonetheless the dominant component of the Layer 2 magnetization.

The correct interpretation of the magnetic stripes was one of the key facts that convinced the geophysical community in the mid 1960's of the reality of **plate tectonics**. By 1966, geophysicists had established a **reversal timescale** dating back 4 Ma. The absolute ages were "anchored" by the **potassium-argon (K-Ar) dating technique**, which is useful in dating melting-cooling events. By 1968 the timescale was extended back to 80 Ma. The timescales have been verified by drilling the seafloor as part of the **Deep Sea Drilling Project (DSDP)**, which initiated in 1968 and continues to the present day. Combinations of magnetic anomaly studies, analysis of fossils in DSDP cores, and absolute radiometric dating by the K-Ar technique determined the timescale of the seafloor.