Geophysics 420: Outline 12

Electromagnetism

Electricity and magnetism are expressions of electromagnetic fields which introduces a separate set of fundamental forces that we can use in the exploration of the Earth's structure and dynamics. EM is used in exploration (magnetic or electrically conducting ores), studies of the present day Earth's magnetic field and the geodynamo, and in paleomagnetism to help unravel past plate motions.

Electrostatic forces follow Coulomb's law with magnitude

$$F_q = C\,\frac{q_1 q_2}{r^2}$$

where the force is repulsive if $q_1 q_2 > 0$. Magnetic fields are generated when electrical charges move. The force on a charge q in a magnetic field $\mathbf{B}$ is given by

$$\mathbf{F} = q\mathbf{v} \times \mathbf{B}$$

with magnitude

$$F = qvB \sin\phi$$

where ϕ is the angle between the velocity and the direction of the magnetic field. The magnetic field is measured in Tesla, where $1\ \mathrm{T} = 1\ \mathrm{N/Am} = 10^4$ Gauss. The Gauss is a convenient, but non-SI unit since the magnitude of the Earth's magnetic field at the Earth's surface is approximately 1 Gauss.

The Earth's magnetic field

The Earth's magnetic field can be approximated by that of a magnetic dipole (think: bar magnet) near the center of the Earth at an angle of 11.5° with the Earth's rotation axis. Non-dipole components include those of the quadrupole, octopole, etc.

At the surface of the Earth, the magnetic field lines at most of the northern hemisphere point into the Earth. The direction of the field lines can be described by the *declination* δ (angle of horizontal component with respect to geographic north) and *inclination* i (angle of magnetic field with respect to the horizontal plane. For example, the magnetic field at Tuscon AZ (to pick a spot) in 1964 had $\delta = 13°\mathrm{E}$, $i = 59°$, $B = 56\mu\mathrm{T} = 0.56$ gauss. The magnetic field shows large variability, induced by the interaction with the solar wind, atmospheric phenomena (thunderstorms) and secular variations in the internal magnetic field (which indicates a dynamic origin for the generation of the field).

Rocks can respond to and/or record the Earth's magnetic field through the magnetization of its minerals. In general the magnetization is related to the magnetic field by

$$\mu_0 \mathbf{M} = \chi \mathbf{B}$$

where μ_0 is the magnetic permeability of space and χ is the (dimensionless) magnetic susceptibility. Some elements are intrinsically magnetic due to the spin of electrons which generates a mini-dipole. Materials that lack intrinsic magnetization can still become temporarily magnetic by magnetic induction. These materials are *diamagnetic* and have small and negative susceptibilities. In a strong enough magnetic field the induced magnetic field can nevertheless be sufficiently strong to offset other forces (think: levitating frogs). *Paramagnetic* materials are intrinsically magnetic but the orientation of the electron dipoles can be random in particular due to thermal agitation. The effective susceptibility is temperature dependent following Curie's law

$$\mu_0 \mathbf{M} = \chi \frac{\mathbf{B}}{T}$$

where $\chi > 0$ but generally small. The most efficient recorders of the Earth's magnetic field are *ferromagnetic*, with pure iron and magnetite Fe_3O_4 as important examples. The magnetization is strong due to alignment of the electron dipoles by a quantum physical effect called exchange coupling. This effect disappears at the material-dependent Curie temperature, above which the material is paramagnetic. The Curie temperatures of Fe ($770°C$) and magnetite ($580°C$) indicate that surface rocks are at sufficiently low temperature to retain a record of the magnetic field. A related property of some materials is *anti-ferromagnetic* behavior, where the electron dipoles alternate. Although these materials appear non-magnetic, they record the magnetic field in a subtle but very stable manner. An important example of this type of material is hematite (Fe_2O_3) which has a Curie temperature of $680°C$.

An important cause of permanent rock magnetization is thermoremanent magnetization (TRM) which is seen in igneous rock which record the Earth's magnetic field when it cools down through the Curie temperature of the minerals. Other forms include depositional or detrital remanent magnetization (DRM) due to the alignment of magnetic minerals during the deposition of sediments, or chemical remanent magnetization (CRM) which occurs when minerals record the magnetic field as the crystals grow. The field caused by the magnetized rock is always weaker than the Earth's magnetic field, and is generally not larger than about 1% for basalts and 0.01% for sediments.

The french scientist Brunhes made the important discovery that the magnetic field of the Earth occassionally reverses it self. Although the field can be approximately described by a dipole that is aligned with the Earth's rotation axis, the South and

North pole can change position. The magnetic reversal record forms the basis of the paleomagnetic time scale which can be used for example to map out the age of the ocean floor. In the last 4 million years there are 4 major time periods: Brunhes normal, Matuyama reverse, Gauss normal and Gilbert reverse. The reversals occur episodically with a period on the order of 1 Myr in the Tertiary and Quaternary. Most of the Cretaceous is characterized by an unusually long normal period (the ’superchron’) which may be related to the extensive volcanism and fast spreading of this time period.

The record of reversals can be mapped on the ocean floor, which led to the establishment of sea floor spreading (first demonstrated by Vine and Matthews in 1963). The record of declination and inclination also allows to determine the relative orientation of the rock to the magnetic field when it was formed. For example, the measured inclination i is related to the paleolatitude λ by

$$\tan i = 2 \tan \lambda$$

The paleolatitude can be used to distinguish latitudinal movement which provides minim estimates for the speed of plate tectonics and continental drift. The location of the paleopole can also be determined. If the present day latitude of a rock sample is given by λ_x and the difference in present day and recorded declination is D then the paleopole latitude λ_p is given by

$$\sin \lambda_p = \sin \lambda_x \sin \lambda + \cos \lambda_x \cos \lambda \cos D$$

The two coordinates λ and λ_p allow us to establish the apparent polar wander (APW) path, which indicates motion of the pole relative to the continent. Keith Runcorn established in 1956 that the different APW paths of north-America and Europe can be matched if it is assumed that the continents moved together and formed a supercontinent some 250 million years ago.