Plate tectonics and Magnetic Fields

Physics of the Earth as a Planet, Lecture 2

(Fowler, The Solid Earth, CUP, Chapters 2 & 3, 1st or 2nd edition)

The basic idea of plate tectonics is that the surface layers of the Earth, called the lithosphere, can be considered as rigid caps on a sphere, whose boundaries are outlined by belts of earthquakes. Earthquakes are produced by the relative motion between these caps. Plate tectonics is what is known as a *kinematic* theory, because it is not concerned with the forces that maintain the motions, which are instead taken to be given. In this respect it is unsatisfactory. The other respect in which it is incomplete is that it does not work in continental regions, only in oceanic ones. We will have quite a bit to say about both these issues later on in the course. The second is the reason why the theory was not proposed until 1967, by which time we had a good understanding of the structure and evolution of the sea floor.

The rules

Plate tectonics is concerned with the relative motions between rigid plates. It is not possible to define what you mean by absolute motions. There are three types of plate boundary:

Ridges are where two plates move apart, and are drawn as two parallel lines. The slip vector between A and B need not be at right angles to the plate boundary, though generally it is to a good approximation. Also ridges generally add to both plates at about the same rate, though again there are some exceptions. So people sometimes give the half spreading rate, rather than the true separation rate.

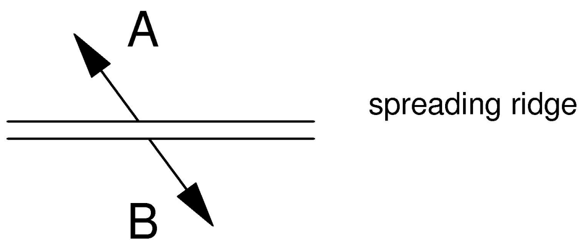

Figure 1: Schematic diagram showing a spreading ridge, with the arrows pointing in the direction of the plate motion. New plate is being created on both A and B.

Trenches are where plates are destroyed, and are also called subduction zones. They are asymmetric, with one plate sliding under the other. Their strike is not related to the slip vector between the plates, and they are drawn with triangles on the plate that is not being destroyed (called the overriding plate).

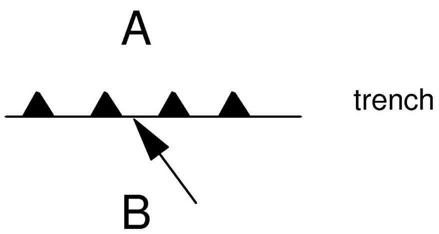

Figure 2: Schematic diagram showing a trench or subduction zone. The arrow denotes the direction of plate motion, with plate B being destroyed by subduction beneath plate A. The triangles are on plate A, which is not being destroyed.

Transform Faults are boundaries of pure slip, and are also known as strike slip faults. They are, by definition, parallel to the slip vector, and are shown by a single solid line. There are two types, left and right handed when viewed from above (sometimes called sinistral and dextral).

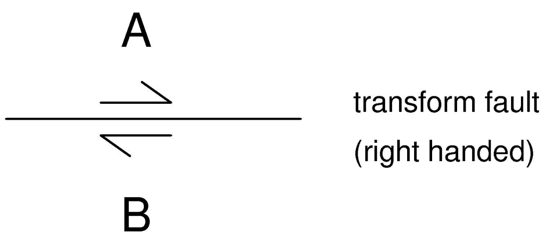

Figure 3: Schematic diagram showing a transform fault, or strike slip fault. The semi arrows define the direction of the plate motion, which is always parallel to the plate boundary. No new plate is being created, nor is any plate being destroyed.

Translation

On a plane the simplest type of motion is a translation. Even this simplest geometry changes with time, because ridges are symmetric and trenches are not.

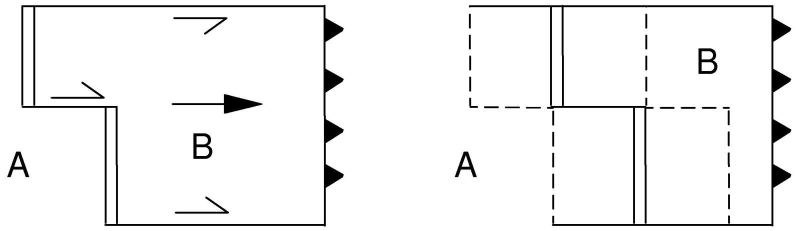

Figure 4: Example of translation motion. The ridges are spreading between plates A and B, which is being accommodated by plate B being destroyed at a subduction zone (or trench). Over time, the ridge moves towards the trench, because the ridges are symmetric and the trench is not.

Triple Junctions

The other structure we need to understand is called a triple junction, where three plates meet at a point. The simplest of these is where three ridges meet. If we choose some (arbitrary) frame, and define the plate velocities in this frame to be $\mathbf{v}_A$, $\mathbf{v}_B$ and $\mathbf{v}_C$, then the relative velocity of B with respect to A, ${}_A\mathbf{v}_B$ is

$${}_A\mathbf{v}_B = \mathbf{v}_B - \mathbf{v}_A$$

Similarly

$${}_B\mathbf{v}_C = \mathbf{v}_C - \mathbf{v}_B$$

$${}_C\mathbf{v}_A = \mathbf{v}_A - \mathbf{v}_C$$

and hence

$${}_A\mathbf{v}_B + {}_B\mathbf{v}_C + {}_C\mathbf{v}_A = 0 \tag{1}$$

We can represent these velocities in velocity space by a triangle whose vertices are the plates. The sides of the triangle represent the velocities between the pairs of plates. The dashed lines *ab*, *bc* and *ac* in the velocity triangle, show frames in which the geometry of the associated plate boundaries do not change. If these lines meet at a point there is a frame in which the geometry of the entire triple junction does not change, and it is known as a *stable* triple junction. If this condition is not satisfied it is called an *unstable* junction.

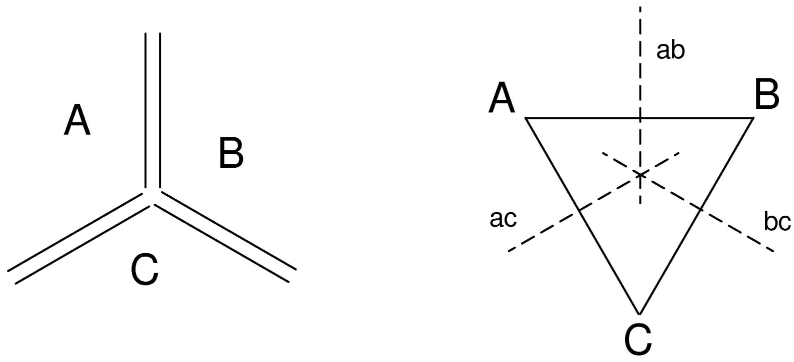

Figure 5: Example of a triple junction consisting of three ridges.

The definition of the line which defines the frame in which the geometry of the associated plate boundary does not change, depends on the type of plate boundary involved. The line *ab* is always parallel to, i.e. in the direction of, the plate boundary between the two plates. For a spreading ridge between two plates A and B, the line *ab* is given by the bisector of ${}_A\mathbf{v}_B$ (Fig. 6). For a trench between two plates A and B, the line *ab* goes through the vertex of ${}_A\mathbf{v}_B$ representing the plate which is not being destroyed (Fig. 7). Finally, for a transform fault, the lines *ab* goes through the two vertices of the plates involved (Fig. 8).

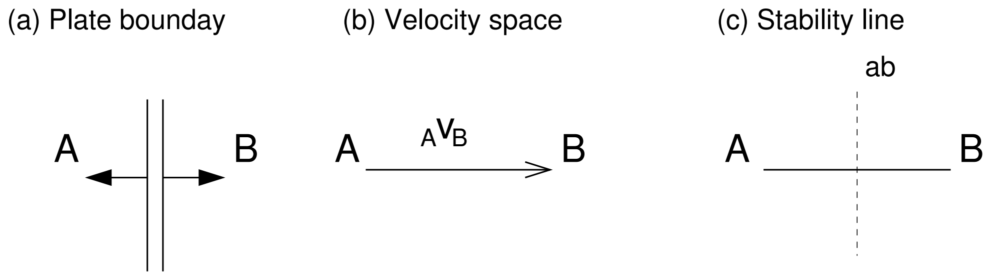

Figure 6: Determining the line *ab* which defines the frame in which a spreading ridge is stable.

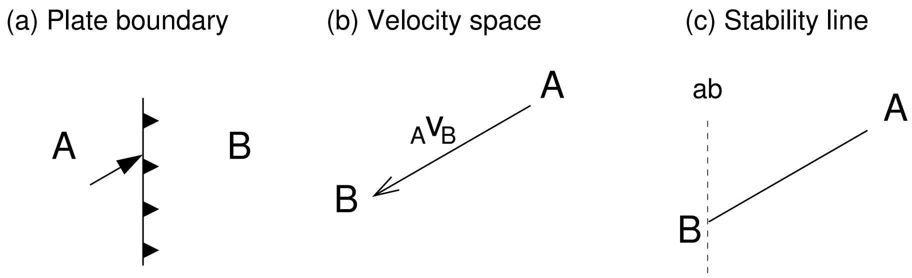

Figure 7: Determining the line *ab* which defines the frame in which a trench or subduction zone is stable.

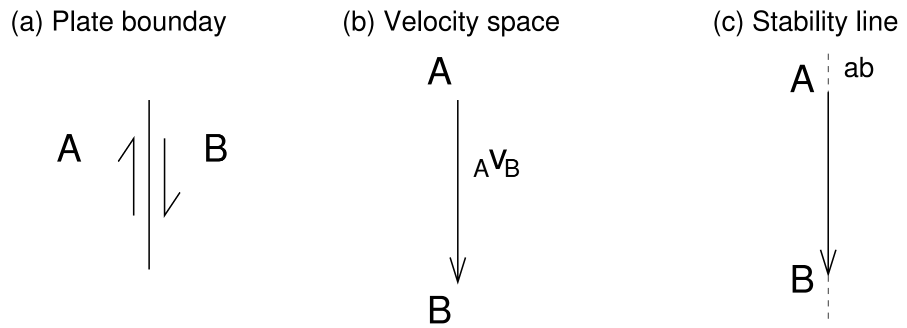

Figure 8: Determining the line ab which defines the frame in which a transform fault is stable.

Motions on a sphere

The Earth is a sphere, and so we must generalise these geometric ideas to motions on a spherical surface. The key idea here is a theorem due to Euler, which states that any motion on the surface of a sphere can be described by a rotation about an axis through the centre of the sphere. Any movement of a rigid object can be described in terms of a translation and a rotation. If spherical caps are to remain on the surface of the Earth, their rotation axes must pass through its centre. In exactly the same way as before

$$ {}_A\boldsymbol{\omega}_B + {}_B\boldsymbol{\omega}_C + {}_C\boldsymbol{\omega}_A = 0 \qquad (2) $$

where ${}_A\boldsymbol{\omega}_B$ is the relative angular velocity vector of plate B with respect to plate A. However, the plane described by these three angular velocity vectors does not lie in the tangent plane at the triple junction. Taking the cross product with $\mathbf{r}$, the radius vector to the triple junction, gives equation (1).

Finite rotations on a sphere are harder to deal with. Clearly we can move any plate from one position to another by two rotations. The first takes A to A', and could, for instance, be a rotation about the pole of the great circle through A and A', the second is about A' and brings B'' to B' (Fig. 9).

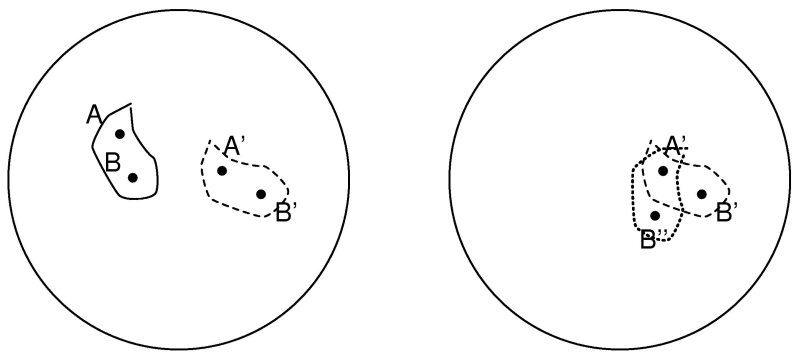

Figure 9: Diagram illustrating combining two finite rotations.

Two rotations are easily shown to be equivalent to a single rotation about a different axis. Consider two rotations, θ about B and ϕ about A, and construct the two spherical triangles shown. Clearly a rotation first about B then about A carries P_2 to P_1, and then back to P_2, whereas A then B carries P_1 to P_2, then back to P_1. P_2 and P_1 are therefore the two points that do not move when these rotations are carried out. These points are therefore, by definition, the two poles of finite rotation. They only coincide as $\theta,\ \phi \to 0$, when they become infinitesimal rotations. Because these two poles are different, finite rotations do not in general commute with each other, because the order in which they are carried out makes a difference to the final result.

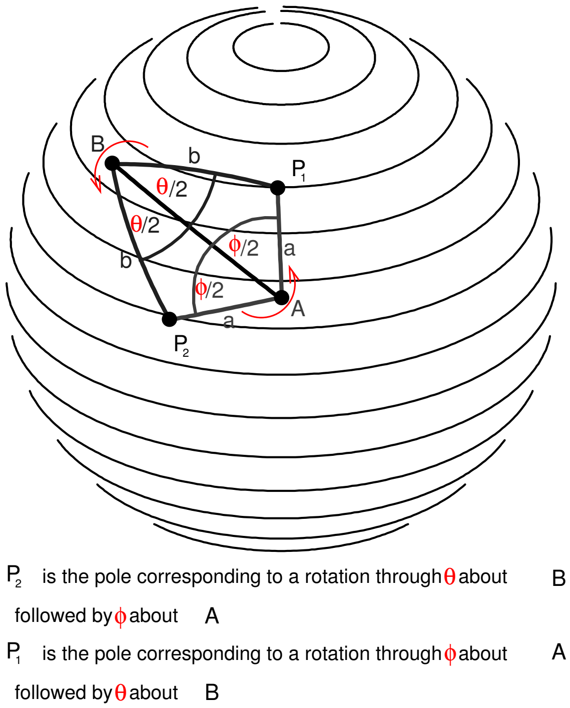

Figure 10: Diagram illustrating the two poles of finite rotation.

Plate motions

The plate boundaries are defined by earthquake locations. Most earthquakes result from relative motion between plates, whose interiors are aseismic because they do not deform. This idea works well in oceanic regions, but in continental regions seismicity is distributed over broad bands and the motion is not taken up on a single fault. We cannot therefore expect simple geometric ideas to describe continental deformation, though the relative motion of the aseismic regions on either side can be described by a rotation about a pole.

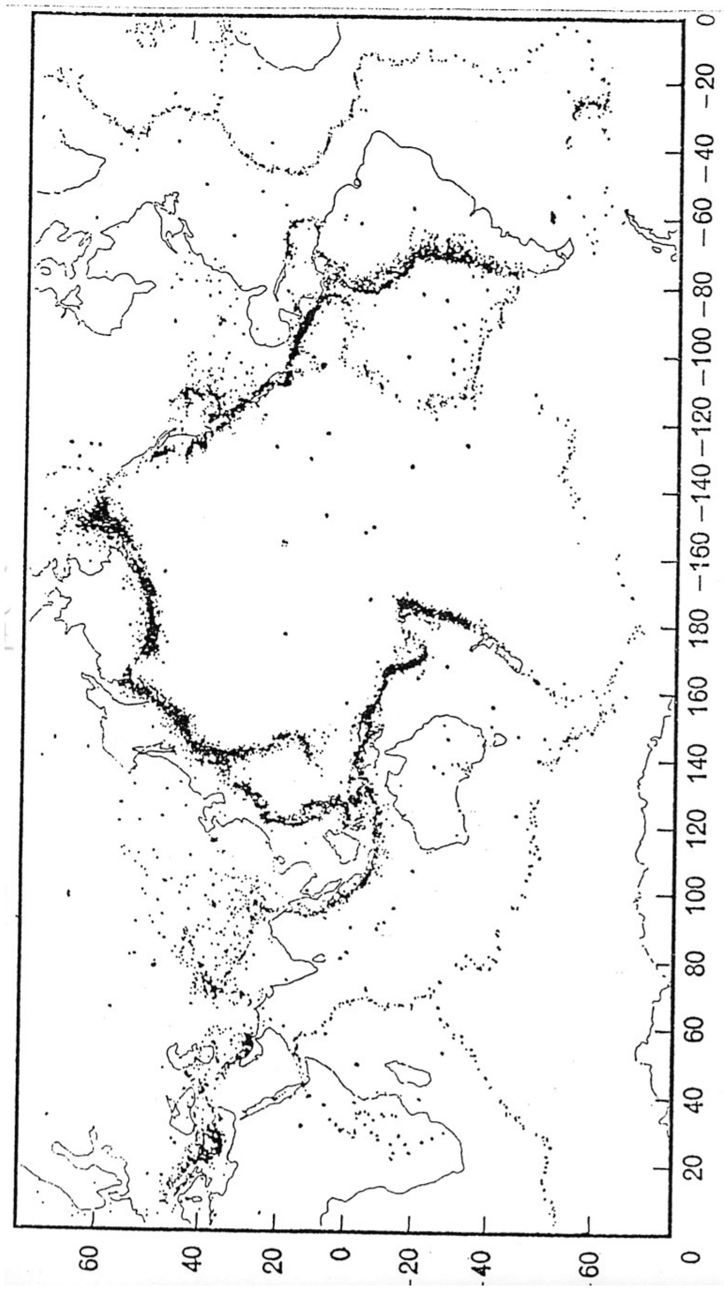

Figure 11: Earthquake locations, from Barazangi & Dorman (1969). A daily updated version can obtained from the IRIS webpage (http://www.iris.edu).

The relative velocity between two plates can be measured directly by various methods. The most widely used is satellite laser ranging, SLR, where lasers at two stations are used to measure the distance to a passive satellite as it passes overhead. The distance between them can be calculated. One method depends on the satellite being visible over a considerable part of the sky from both stations at the same time. When this condition is not satisfied, the ranging measurements must be used to obtain the orbit of the satellite, and this in turn used to measure the distance between the two stations. The satellite widely used for such studies is called LAGEOS, and consists of a tungsten ball covered with corner reflectors. The plot shows a comparison of the relative motions of a number of stations determined by SLR with geological estimates from magnetic anomalies. We now know that the (small) difference between the two methods is real, and results from errors in the radiometric dates used to calibrate the magnetic time scale. When these errors are corrected, the two measurements agree to within about 1%, which is about the likely error. The geological rates over millions of years therefore are the same as those measured over a few years.

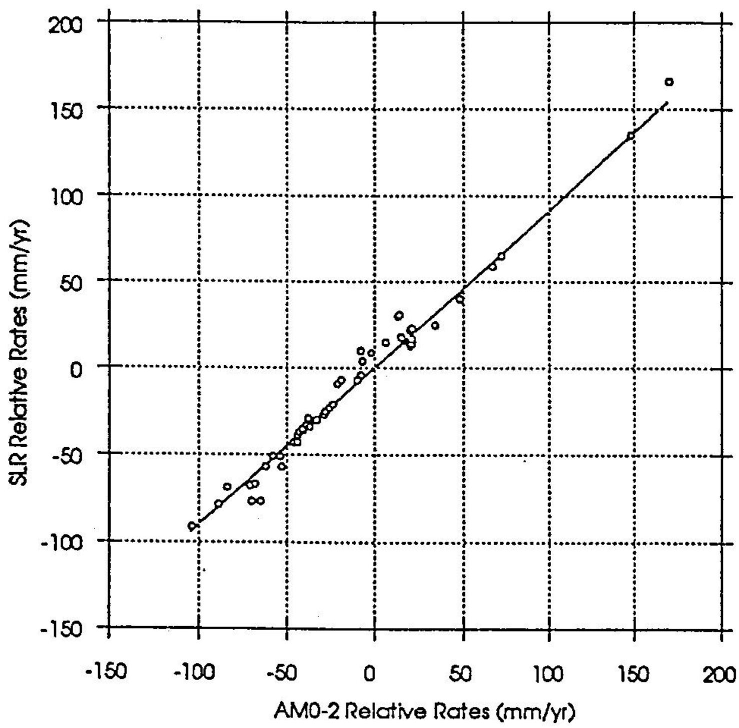

Figure 12: Comparison of SLR determined geodesic rates with those implied by the AMO-2 geologic plate motion model for 54 lines connecting stations on five plates that are well within plate interiors and crossing at least one plate boundary. The slope of the line is 0.914 ± 0.017.

Finite rotations of plates cannot in general occur about the instantaneous poles, because the instantaneous poles cannot remain at the same position on each pair of plates as all plates move. But there is always a single finite rotational pole, called the Euler pole, that describes any finite movement between two plates. Figure 13 shows this pole for the South Atlantic, with small circles drawn about it. In this case the finite and instantaneous poles approximately coincide. The small circles about the finite pole are therefore parallel to the transform faults, and are lines of pure slip. The two continents are equidistant from the spreading ridge, because the ridge has added to both plates at the same rate. When the two continents are rotated towards each other through equal angles in a frame in which the ridge is fixed, it is clear that the present geometry of the ridge is inherited from that of the initial break, and that the transform faults are produced when the break is not normal to the slip vector.

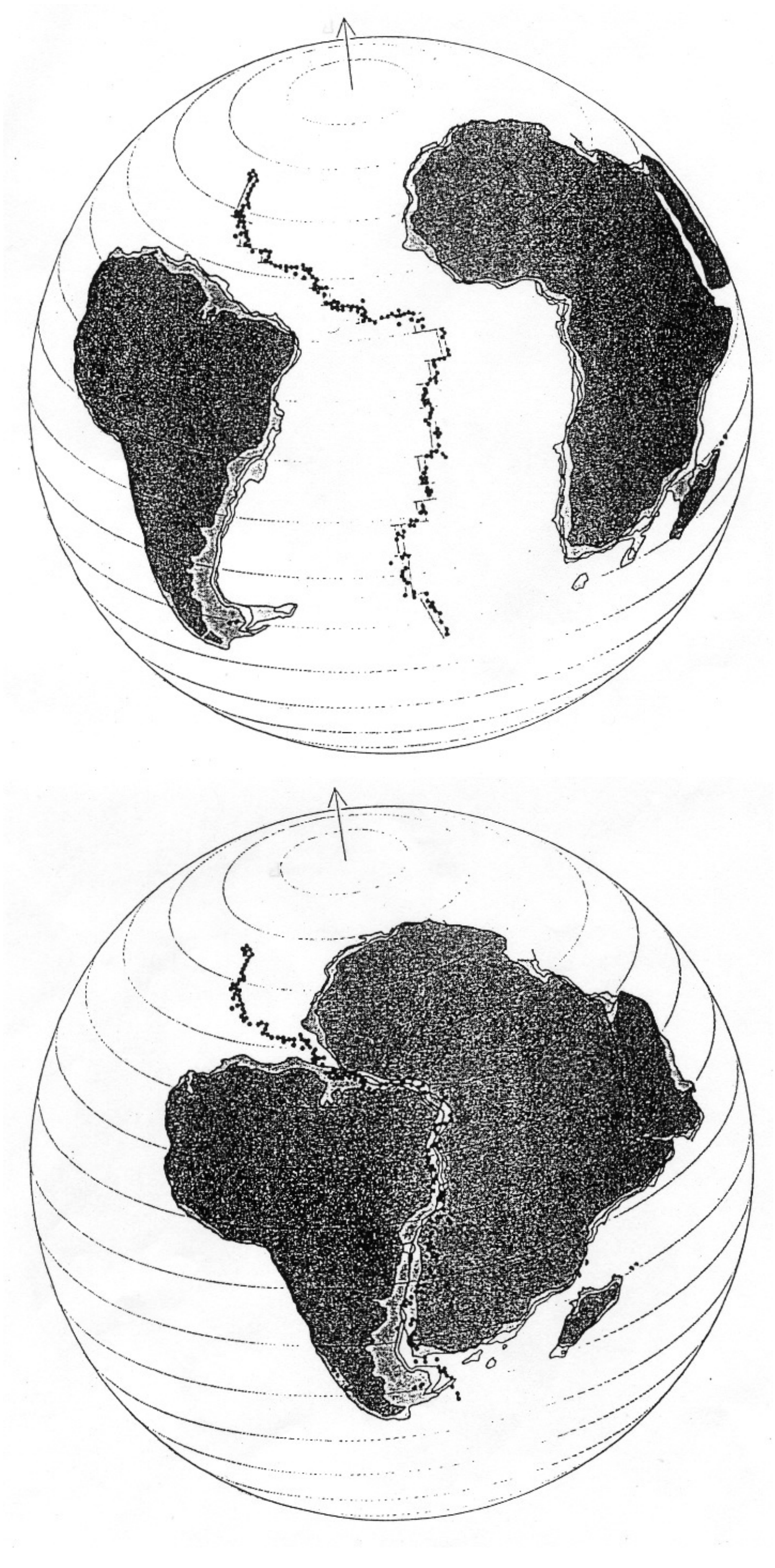

Figure 13: Euler pole for the South Atlantic.

Magnetic anomalies

The principal method we use for reconstructing past movement of plates depends on the magnetic anomalies that are formed on the sea floor when two plates separate. The main field of the Earth's core can be approximated by a dipole at the Earth's surface, largely because the thickness of the Earth's mantle (≈ 3000 km) is sufficient to attenuate the higher harmonics.

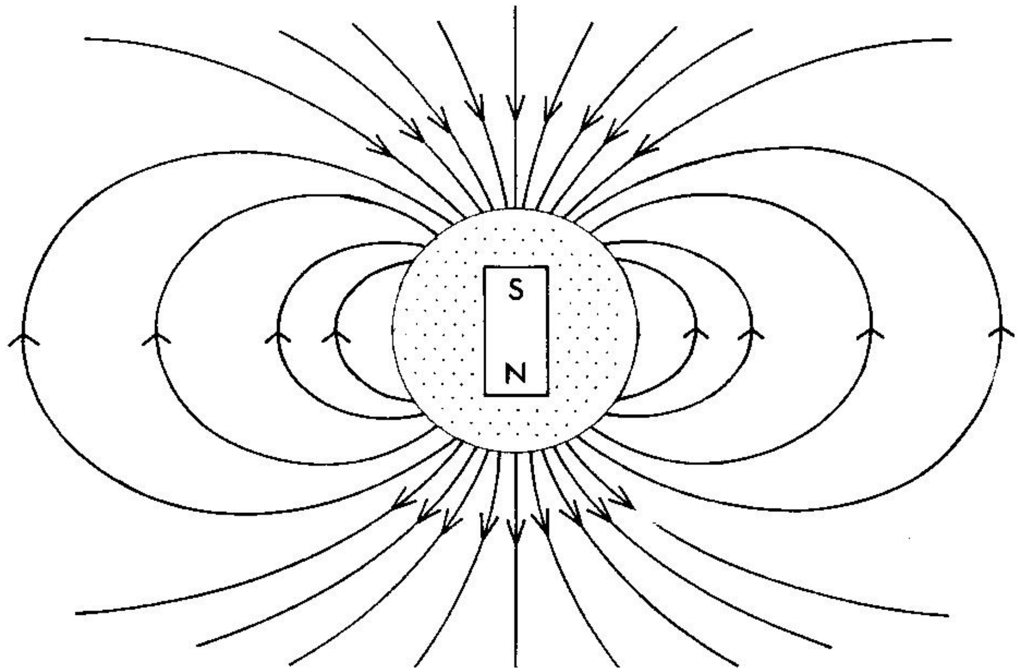

Figure 14: A magnetic dipole at the centre of the Earth is a good approximation of the Earth's magnetic field.

Outside the Earth's core, with the absence of any free currents, the magnetic field strength $\mathbf{H}^c$ satisfies the following Maxwell equation (Ampère's law)

$$\nabla \times \mathbf{H}^c = \mathbf{0}, \tag{3}$$

and as a result it can be written in terms of a magnetic scalar potential Φ^c as

$$\mathbf{H}^c = -\nabla\Phi^c. \tag{4}$$

To a good approximation, the region outside of the core can be treated as an electromagnetic vacuum with constitutive law for the magnetic field $\mathbf{B}^c$ as

$$\mathbf{B}^c = \mu_0 \mathbf{H}^c, \tag{5}$$

where μ_0 is the magnetic permeability of free space. Another of Maxwell's equations (Gauss's law for magnetism) is

$$\nabla \cdot \mathbf{B}^c = 0 \tag{6}$$

from which it follows that the magnetic scalar potential satisfies Laplace's equation,

$$\nabla^2\Phi^c = 0. \tag{7}$$

Solutions to Laplace's equation can be written in terms of a multipole expansion (or equivalently, as a sum of spherical harmonics). On the Earth's surface, the magnetic scalar potential is dominated by the leading order term, which represents a magnetic dipole. The potential Φ^c from a dipole at the origin is

$$\Phi^c = \frac{\mathbf{m} \cdot \mathbf{r}}{4\pi r^3} = -\frac{m \cos\theta}{4\pi r^2} \tag{8}$$

where the dipole moment $\mathbf{m}$ points towards the south pole, and $\mathbf{r}$ is the position vector. The magnetic field is obtained by taking the gradient (from (4) and (5)),

$$\mathbf{B}^c = -\mu_0 \nabla \Phi^c = \frac{\mu_0}{4\pi}\left(\frac{3\hat{\mathbf{r}}(\hat{\mathbf{r}} \cdot \mathbf{m}) - \mathbf{m}}{r^3}\right) \tag{9}$$

where $\hat{\mathbf{r}} = \mathbf{r}/r$. The magnitude of the magnetic field is

$$|\mathbf{B}^c| = \frac{\mu_0 m}{4\pi r^3}\left(3\cos^2\theta + 1\right)^{1/2}. \tag{10}$$

The field $\mathbf{B}^c$ is strongest at the poles

$$|\mathbf{B}^c(0)| = 2|\mathbf{B}^c(\pi/2)|. \tag{11}$$

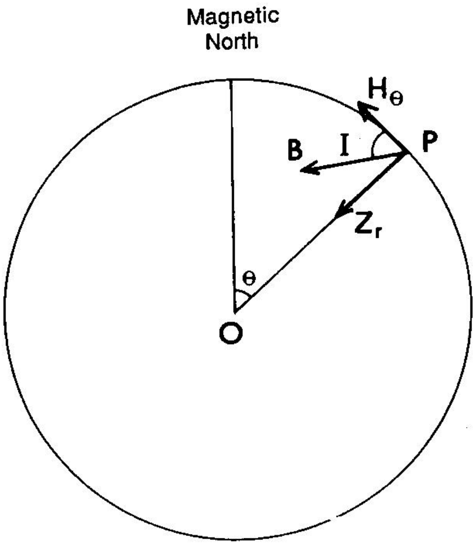

Figure 15: Diagram illustrating I the angle of inclination. H_θ represents the horizontal component of the magnetic field $\mathbf{B}^c$, and Z_r represents the inward radial component (i.e. the downward vertical).

If the inclination of the field to the horizontal is I

$$\tan I = 2\cot\theta = 2\tan\lambda \tag{12}$$

where λ is the latitude (try deriving this result!). Equation (12) is very useful. See pages 36-38 (1st ed.) or 48-50 (2nd ed.) of 'The Solid Earth' for further details on the derivation of these equations.

Paleomagnetism

When sediments are deposited, they can acquire remnant magnetisation. This is much less than the magnetisation of igneous rocks (e.g. basalt flows from volcanic eruptions), but it can still be measured. We can measure the inclination in igneous and sedimentary rocks and use it to compute the paleomagnetic latitude. If we manage to obtain paleomagnetic positions from rocks of different ages on the same continent, then we can track the path of the continent on a map (Figure 16).

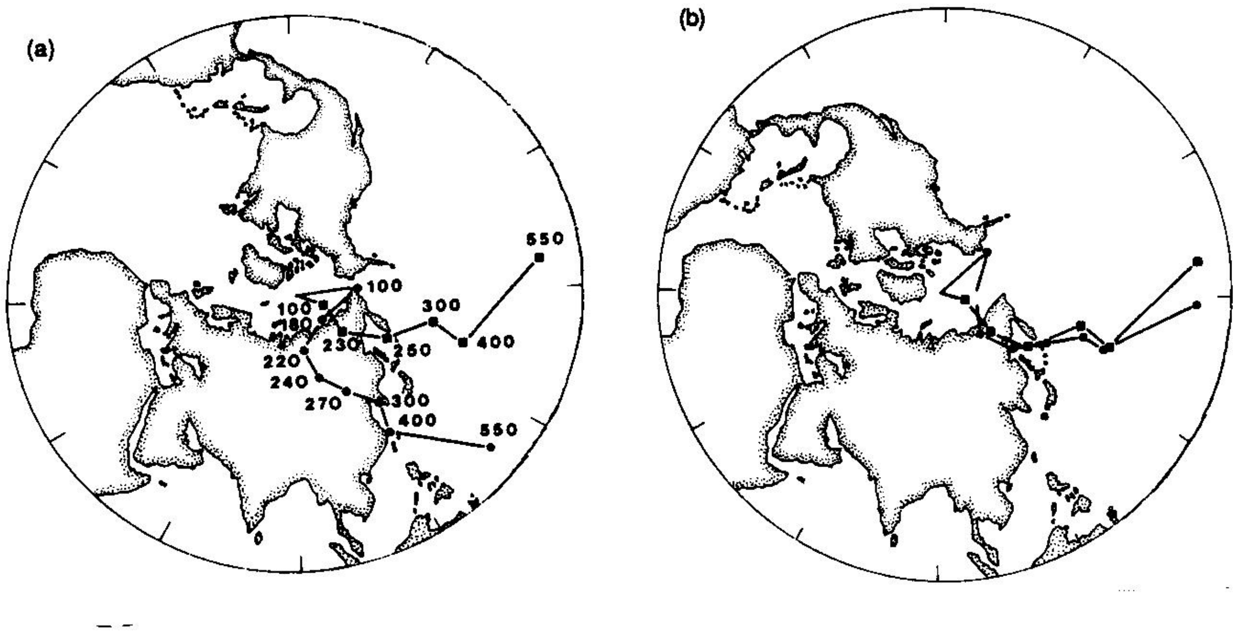

Figure 16: (a) Polar wander curves for North America (circles) and Europe (squares). (b) Polar wander curves for North America and Europe when allowance has been made for the opening of the Atlantic Ocean. The two curves are now almost coincident. (After McElhinney 1973.)