crustaceans, and bryozoans); and a small number of vagile taxa, or strollers, largely gastropods, that roamed the surfaces of the reefs. It is likely that this latter guild is under-represented. These studies suggest that the reefs built up after initially unstable oolitic shoals had been stabilized by both infaunal and epifaunal bivalves, the shells of which then provided the firm seafloor onto which the reef-forming organisms were recruited (**Figure 5**).

See Also

Fossil Invertebrates: Trilobites; Bryozoans; Corals and Other Cnidaria; Graptolites; Molluscs Overview; Bivalves; Gastropods. **Fossil Vertebrates:** Dinosaurs. **Lagerstätten**. **Microfossils:** Foraminifera. **Palaeopathology**.

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PALAEOMAGNETISM

T H Torsvik, Geological Survey of Norway, Trondheim, Norway

Introduction

Palaeomagnetism is the study of the Earth's magnetic field preserved in rocks. The discovery that some minerals, at the time of their formation, can become magnetized parallel to the Earth's magnetic field was made in the nineteenth century. Early in the twentieth century, Bernard Brunhes made the startling discovery that some rocks are magnetized in the opposite orientation to the Earth's present-day magnetic field. This led him to propose that the Earth's magnetic field had reversed its polarity in the past. These reversals have subsequently been shown to be non-periodic and the Earth's magnetic field reversal history is now well known for the past 175 million years and more sketchingly understood to the beginning of the Palaeozoic (*ca*. 545 Ma).

Palaeomagnetism has a range of application potential (*see* **Magnetostratigraphy, Analytical Methods:** Geochronological Techniques), and the focus here is on understanding the importance of palaeomagnetism as an investigative tool in assembling palaeogeographical reconstructions.

Fundamentals

The Earth's magnetic field is believed to originate from the outer fluid core and, at the surface, the field is described by its inclination (angle with respect to the local horizontal plane), declination (angle with respect to the Greenwich meridian), and field strength (Figure 1). The inclination of the Earth's field varies systematically with latitude, which is of prime importance for palaeomagnetic reconstructions. At the north magnetic pole, the inclination of the field is $+90^\circ$ (straight down), at the equator the field inclination is zero (horizontal) pointing north, and at the south magnetic pole the inclination is -90° (straight up; Figure 2). The magnetic north and south poles currently differ from the geographical north and south poles by 11.5° because the magnetic axis is inclined from the geographical (=rotation) axis. The

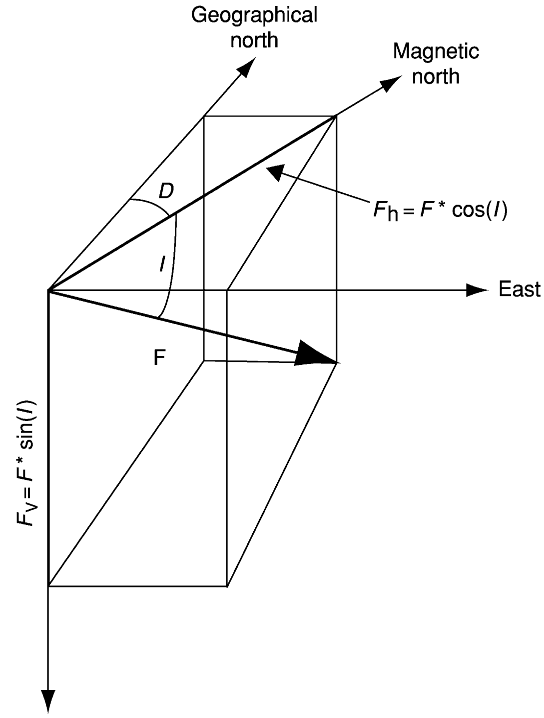

Figure 1 The direction and intensity of the total field vector (**F**) decomposed into declination from geographical north (D) and inclination from the horizontal (I). The equations relate the horizontal (F_h) and vertical (F_v) components of the total field (**F**) with I and D.

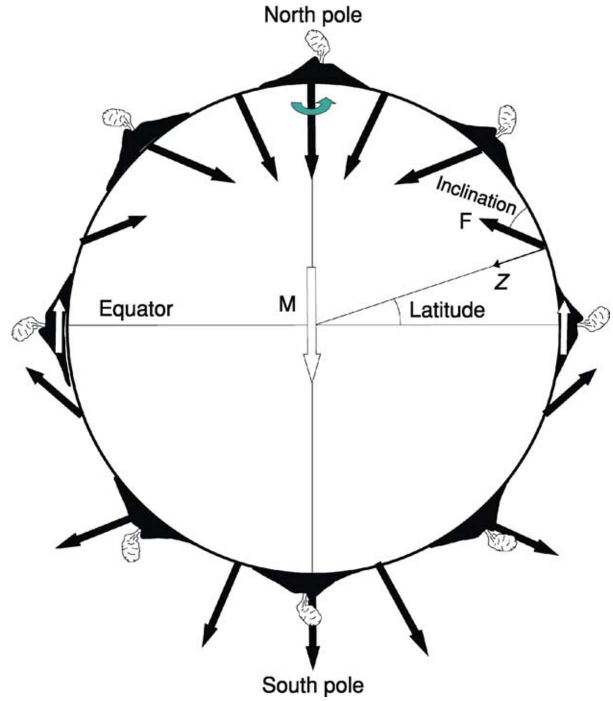

Figure 2 Field lines at the Earth's surface for a geocentric axial dipole. At the equator, the inclination is flat (zero) and, at the north and south poles, the inclination is vertical (+90° and −90°, respectively). The inclination recorded in volcanoes formed on the Earth's surface is dependent on the latitude (see also Figure 3). Declinations in a normal polarity field, such as today's, should point to the north.

magnetic axis, however, is slowly rotating/precessing around the geographical axis (known as secular variation) and, over a period of a few thousand years, it is hypothesized that the averaged magnetic poles correspond reasonably well with the geographical poles. This is known as the geocentric axial dipole (GAD) hypothesis. We can therefore imagine that a magnetic dipole is placed at the centre of the Earth and aligned with the Earth's rotation axis (Figure 2). In palaeomagnetic studies, it is therefore important to sample rocks whose ages range over more than a few thousand years; a study of a single dyke or basalt flow, for example, represents an instantaneous reading of the Earth's magnetic field (cools within a scale of days to weeks) and will not accurately record the position of the Earth's rotation axis.

When rocks form on an ideal planet, they acquire a remanent (permanent) magnetization parallel to the Earth's magnetic field at that location (Figures 2 and 3). There are a number of ways in which a rock can acquire a remanent magnetization, but most rocks are magnetized by one of the following processes: (1) as magma solidifies and cools below the Curie temperature (T_C, i.e., temperature above which a magnetic material loses its magnetism because of thermal agitation), magnetic minerals acquire a thermoremanent magnetization aligned with the Earth's magnetic field at the time of cooling; (2) during the deposition of sediments, magnetic mineral grains settle statistically in the direction of the Earth's magnetic field and a detrital remanent magnetization is acquired; and (3) when magnetic minerals are formed during chemical processes (diagenesis or metamorphism), the magnetic minerals grow as magnetized crystals with their magnetization in the direction of the external magnetic field; this creates a chemical remanent magnetization. Most magnetic minerals are iron–titanium oxides that belong to two solid solution series (Figure 4): the titanomagnetites (e.g., the end-member magnetite) and the titanohaematites (e.g., haematite). In the titanomagnetite series ($Fe_{3-x}Ti_xO_4$), there is an approximately linear variation of spontaneous magnetization ($\mathbf{M}_s$) and T_C with composition (x). Other common magnetic minerals include goethite and pyrrhotite (Table 1).

As an example, we can consider the remanence acquisition of a basaltic lava flow during cooling. The most important magnetic mineral in basaltic rocks is titanomagnetite. Low-titanium phases (e.g., pure

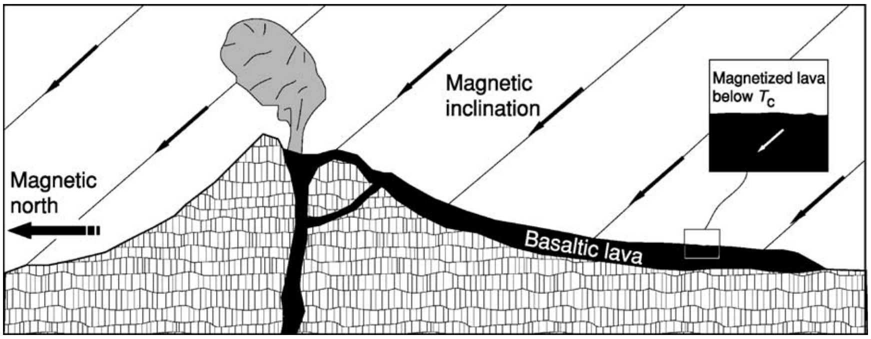

Figure 3 Example of acquisition of a thermoremanent magnetization (TRM) at intermediate northerly latitudes (acquired in a normal polarity field similar to today's). A lava will acquire a TRM upon cooling below the Curie temperature (see text), and the inclination will parallel the inclination of the external field and have declinations due north.

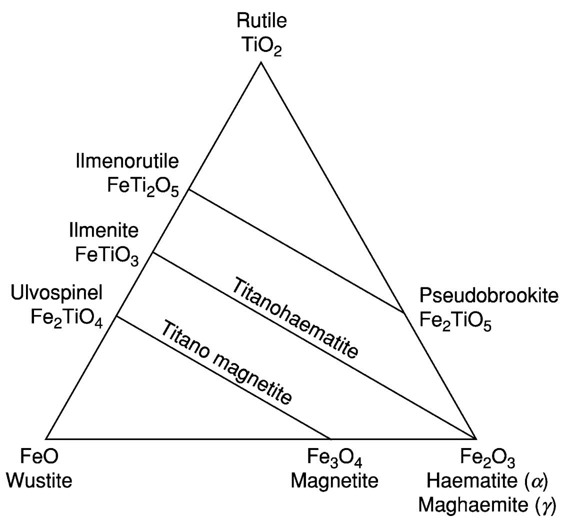

Figure 4 Ternary diagram showing the magnetically important iron oxide minerals in the titanomagnetite and titanohaematite solid solution series.

Table 1 Magnetic properties of some common minerals in rocks

Mineral (composition)	M_s*(kA m^{-1})*	T_C *(°C)*
Titanomagnetite ($Fe_{2.4}Ti_{0.6}O_4$)	125	150
Magnetite (Fe_3O_4)	480	580
Maghaemite (γ-Fe_2O_3)	380	590–675
Haematite (α-Fe_2O_3)	~2.5	675
Goethite (α-FeOOH)	~2	120
Pyrrhotite (Fe_7S_8)	~80	320

magnetite; Table 1) have T_C values close to 580°C, whereas the presence of titanium lowers the T_C value. During a basaltic volcanic eruption, the temperature of a lava is approximately 1200°C; when a lava flow cools below T_C, the Earth's magnetic field is recorded within the lava flow (Figure 3). The declination, inclination, and magnetization intensity, which are proportional to the strength of the field, can today be measured in the laboratory. The magnetization intensity can vary by several orders of magnitude between different rock types, and thus different laboratory instruments are required to measure the magnetization precisely. Volcanic rocks normally have high intensities and the magnetization can be measured on standard spinner magnetometers. Conversely, sedimentary rocks can be extremely weakly magnetized, and highly sensitive superconducting magnetometers (superconducting quantum interference devices, SQUIDs) are required to measure and unravel their magnetization history.

Palaeomagnetic Analysis

In the early days of palaeomagnetic studies, it was common to measure the magnetization in a rock and assume that this magnetization, referred to as the natural remanent magnetization (NRM), represented a primary magnetization that had survived magnetic resetting from subsequent thermal or chemical activity. However, during the 1970s, it became more and more evident that rocks can undergo magnetic resetting, and it is therefore now standard procedure to test the stability of the NRM by thermal, alternating field, or chemical (rare) demagnetization. With the former method, a sample is measured following heating to higher and higher temperatures in a 'zero' field oven. From the 1980s, it became standard procedure to display the demagnetization data in orthogonal vector plots, also referred to as Zijderveld diagrams (Figure 5). These diagrams portray directional and intensity changes on a single diagram – magnetization components are identified as linear segments in both the horizontal ('declination') and vertical ('inclination') planes. Components and the degree of linearity can be computed using least-squares algorithms. A single-component

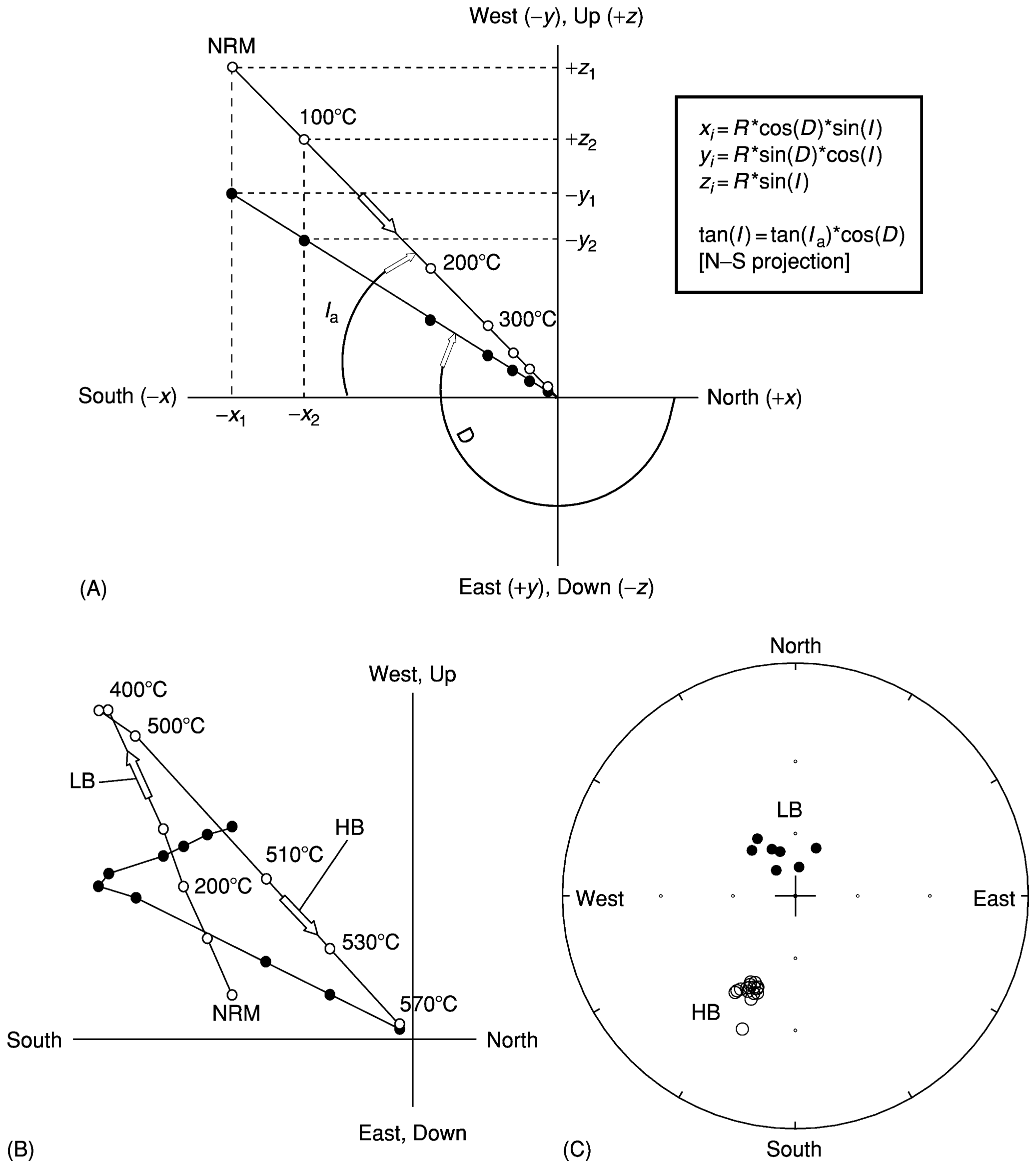

Figure 5 A Zijderveld plot illustrating vector components of magnetization during progressive thermal demagnetization, projected onto orthogonal horizontal (declination; filled symbols) and vertical (inclination; open symbols) planes. For each demagnetization step, the measured declination (D), inclination (I), and magnetization intensity (R) are decomposed to cartesian coordinates, x_i, y_i, and z_i, using the formulae in the box in (a). N–S (+x,−x) is selected as the projection plane as the declination is closer to N–S than E–W. The horizontal component is plotted as x_i,y_i, whereas the vertical plane is plotted as x_i,z_i. This procedure is repeated for each demagnetization step (in the example, natural remanent magnetization (NRM): 100°C, 200°C, etc.). Using this procedure, a magnetic component is recognized as a linear vector segment. The declination for a component can be read directly from the diagram, normally numerically computed by least-squares analysis, whereas the inclination is apparent (I_a) and always larger than the real I; the 'distortion' of I_a depends on the projection plane. (A) Single-component magnetization decaying towards the centre of the diagram. (B) Two-component magnetization with a low unblocking component (LB) identified below 400°C and a high unblocking component (HB) decaying towards the centre of the diagram; the stability up to 570°C suggests pure magnetite as the remanence carrier. (C) Mean site compilation of HB and LB components shown in a stereoplot. LB is a recent overprint, whereas HB should be considered to be a Late Permian magnetization from the Oslo area (Baltica). The mean declination/inclination in this hypothetical study is 205°/−41° (95% confidence circle around the mean, α_{95}, is 2.3°). The calculated palaeomagnetic pole for the site (60° N, 10° E) is 49.3° N and 152.3° E. In the stereoplot, open (filled) symbols denote negative (positive) inclinations.

magnetization is identified by single vector decay towards the origin of the diagram as the sample is progressively demagnetized (Figure 5A). Multi-component magnetizations, in which a primary component has been partly overprinted by younger components, can be recognized by the presence of two or more linear segments (Figure 5B). In the latter example, a hypothetical Permian dyke magnetization from the Oslo region (Norway), with SSW declinations and negative (upward-pointing) inclinations (high unblocking component), is partially overprinted by a younger magnetization of Holocene origin, with NNW declinations and positive (downward-pointing) inclinations (low unblocking component). When the blocking temperatures overlap, curved segments are observed.

In a typical palaeomagnetic study, 5–10 samples are analysed from each site. Magnetization components are computed by least-squares analysis, a mean direction is computed for each site using Fisher statistics, and, finally, a mean direction is calculated from all the sites. In our example, we can imagine that the stereoplot in Figure 5C represents the mean directions from numerous individual dykes in the Oslo region.

Palaeomagnetic Stability Tests

Magnetic overprinting or resetting presents a problem. However, there are four fundamental tests used to check the stability and the potentially primary character of magnetizations: (1) the fold test; (2) the reversal test; (3) the conglomerate test; and (4) the contact test (Figure 6). The fold test determines whether a magnetization was formed prior to or after folding of a rock, and is therefore a relative test that does not prove a primary origin of the remanence. If a remanence is pre-fold, the site vectors should be dispersed after folding (Figure 6A); conversely, if a magnetization is post-fold, the site vectors should be similar throughout the fold structure (Figure 6B). The presence of antipodal stratigraphically linked reversals in a sedimentary or basaltic sequence is the best evidence for primary remanence (Figure 6A). The conglomerate test is also a powerful test. If magnetizations are random between individual boulders (Figure 6A), this is an indication for a primary magnetization in the host rock. Conversely, if boulder magnetizations concur with those in the surrounding rocks, magnetic overprinting is indicated (Figure 6B). A contact test is employed to check whether an intrusion or a dyke carries a primary magnetization. A dyke and its baked margin should coincide, whilst non-baked samples should differ if significantly older

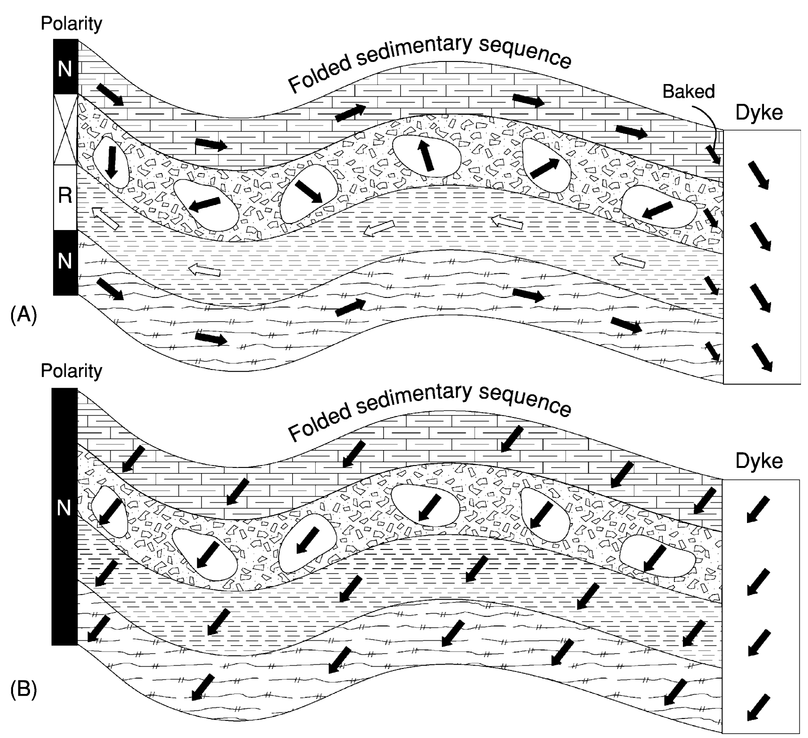

Figure 6 (A) Positive field tests. We assume that a sedimentary sequence was deposited at the equator (flat inclinations) and later folded. Polarity is indicated by the arrows. As the inclinations follow the bedding (except in the conglomerate), this is a pre-fold magnetization. Evidence for a primary magnetization is witnessed by layers with antipodal polarity, and we can therefore establish a magnetostratigraphy (alternating normal and reversed magnetic fields). Further evidence for a primary magnetization is observed from a layer with boulders that show random magnetization vectors. A dyke intruded the folded sequence later, and steeper inclinations (indicating that the continent must have moved to higher latitudes) from the dyke and its baked/chilled margin indicate a primary dyke magnetization. (B) Negative field tests. All layers, conglomerate boulders, and the dyke have similar magnetizations and no reversals are observed. This indicates a regional secondary overprint after the folding and dyke intrusion.

than the dyke (Figure 6A). Conversely, if the dyke and all the surrounding rocks have a similar magnetization, the dyke records a younger overprint (Figure 6B).

Palaeomagnetic Poles and Reconstruction of a Continent

Based on the measurement of the remanent inclination, we can calculate the ancient latitude for a continent when the rock formed from the formula: $\tan(I) = 2 \times \tan(\text{latitude})$. In addition, the remanent declination, which deviates from 0° or 180° (depending on the polarity of the Earth's magnetic field), provides information about the rotation of a continent.

The inclination and declination change with the position of the sampled rock on the globe (Figure 2), but the position of the magnetic pole of a geocentric axial dipole is independent of the locality at which the rock acquired its magnetization. Thus, it is practical to calculate pole positions in order to compare results from various sites or to perform plate tectonic reconstructions.

Ideally, as a time average, a palaeomagnetic pole (calculated from the declination, inclination, and the geographical site location) for a newly formed rock will correspond with the geographical north or south pole. If a continent moves later, the palaeomagnetic pole must move with the continent. To perform a reconstruction with palaeomagnetic poles, we therefore have to calculate the rotation (Euler) pole and angle which will bring the palaeomagnetic pole back to the geographical north or south pole, and then rotate the continent by the same amount (Figure 7A). In our example, a palaeomagnetic pole (latitude, 49.3° N; longitude, 152.3° E), calculated from the situation depicted in Figure 5C, will position the Baltica continent (most of northern Europe eastward to the Urals) at latitudes between 15 and 50° N, causing the city of Oslo to have been located

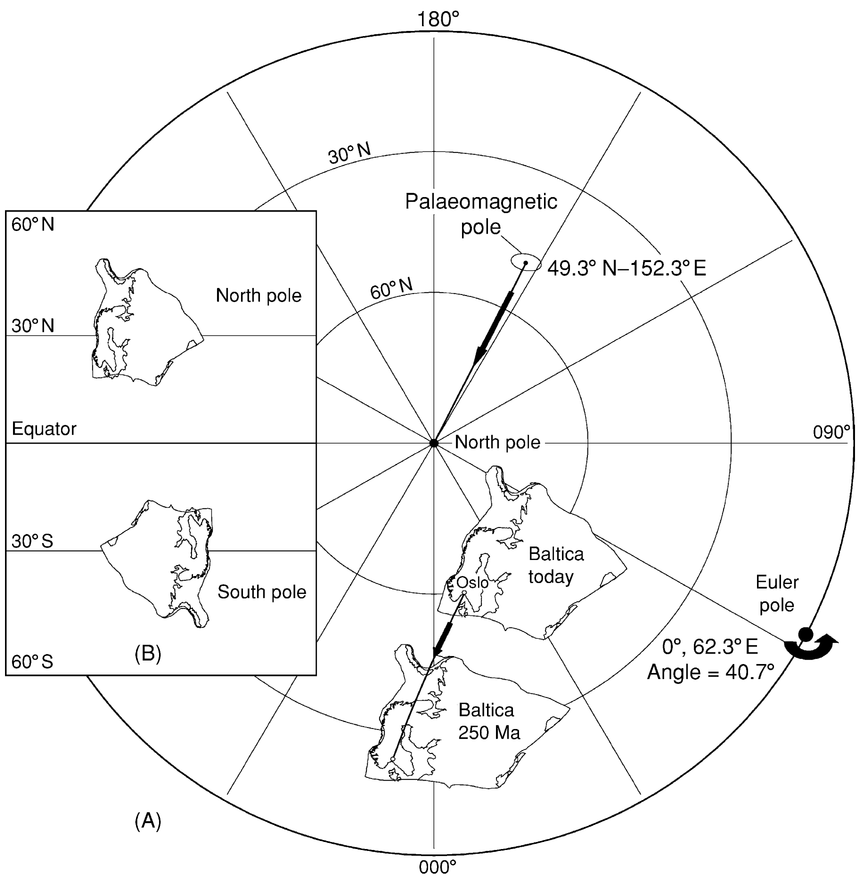

Figure 7 (A) The reconstruction of a continent, for example Baltica, is performed as follows. Determine the Euler pole needed to rotate a palaeomagnetic pole (in our case 49.3° N, 152.3° E) to the geographical north pole (we calculate 0°, 62.3° E and a rotation of 40.7°). This Euler pole is then used to rotate the continent by the same amount. Thus, Baltica today is rotated back about this pole to the position it occupied in Permian times. (B) In (A), we assumed that the palaeomagnetic pole was a north pole. If we assume a south pole, then the continent will be placed in the opposite hemisphere and geographically inverted.

at 24° N in the Late Permian (Figure 7A). Because the current latitude of Oslo is 60° N, Baltica must have drifted northwards since the Permian.

Palaeomagnetic data can only constrain latitude (based on inclination) and the amount of angular rotation (based on declination). Because the palaeolongitude is unknown, we can position Baltica at any longitude we wish, subject to other geological constraints. In addition to that uncertainty, we cannot tell in old rocks whether a palaeomagnetic pole is a south or north pole. In Figure 7A, we assumed that the pole was a north pole, but if we used a south pole, Baltica would plot in the southern hemisphere but in a geographically inverted orientation (Figure 7B). Hence, there is freedom to select north or south poles when producing reconstructions, placing the continent in an opposite hemisphere and rotated by 180°.

Apparent Polar Wander Paths

Apparent polar wander (APW) paths represent a convenient way of summarizing palaeomagnetic data for a continent or terrane, instead of producing palaeogeographical maps at each geological period. APW paths represent the apparent motion of the rotation axis relative to the continent, depending on whether one plots the movement of the north or south pole. APW paths can therefore be constructed as north or south paths. To construct an APW path, a set of palaeomagnetic poles of varying geological age

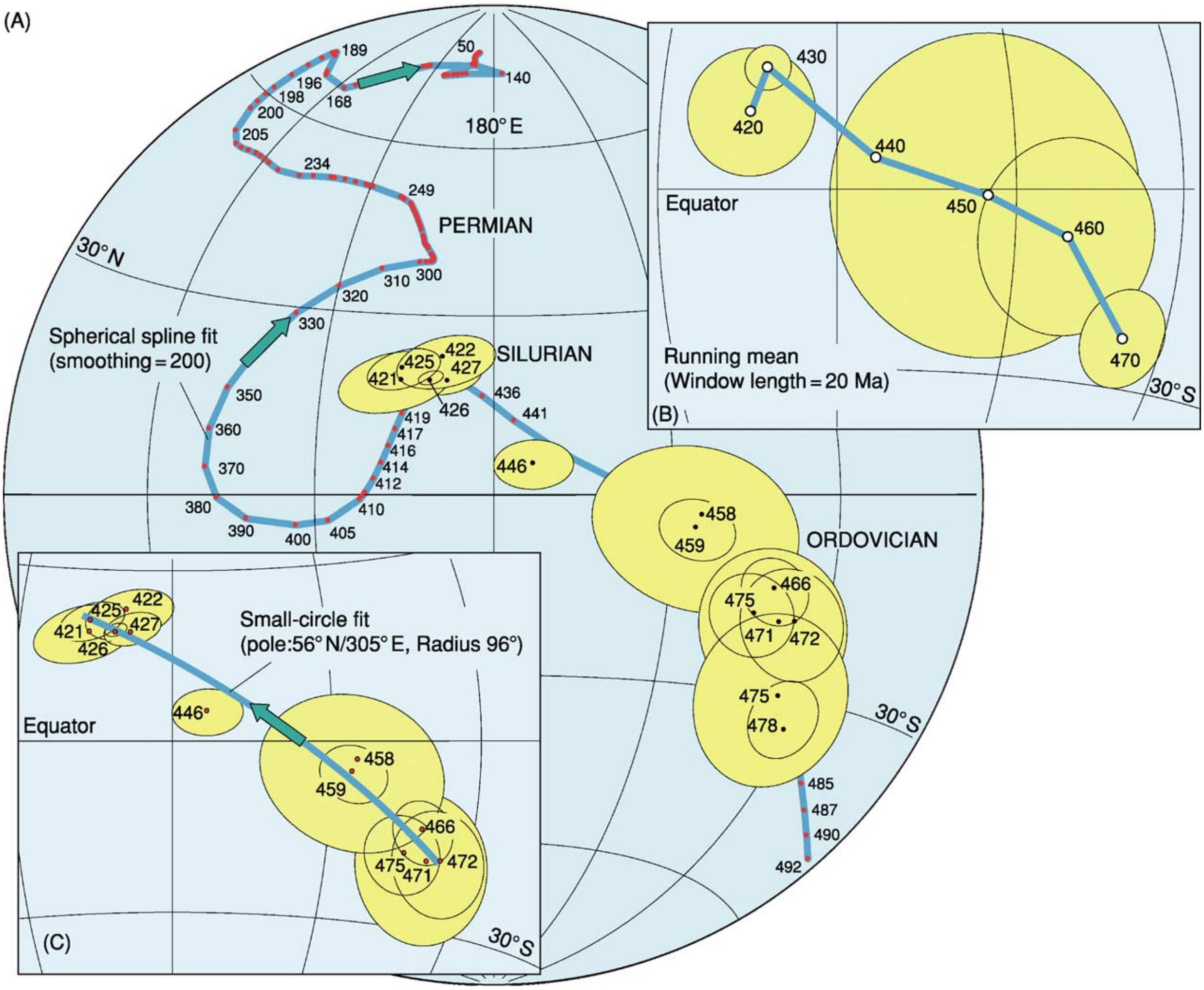

Figure 8 Examples of apparent polar wander (APW) paths for Baltica (include stable European data from Permo-Carboniferous times). (A) Moderately smoothed spherical spline APW path from Early Ordovician to Early Tertiary times. Only Ordovician through Silurian input poles are shown. (B) Running mean path using the same input poles as in (A) and a 20 million year window. Mean poles are shown with 95% confidence ellipses (A95), except for the 440 Ma mean pole for which there was only one pole entry. The running mean path is only shown for the Ordovician–Silurian section of the APW path. (C) Small circle path fitted to Ordovician–Silurian poles as in (A). Input poles in (A) and (C) are shown with 95% confidence ovals (known as dp/dm).

are presented in a single diagram, and a synthetic path is fitted to the incrementing poles (Figure 8A). There are three common methods for generating APW paths: (1) spherical splines; (2) running mean (sliding time window); and (3) the small circle method.

The spherical spline method of modelling APW paths has been employed since the late 1980s. In brief, a spline constrained to lie on the surface of a sphere is fitted to the palaeomagnetic poles (Figure 8A), themselves weighted according to the precisions of the input palaeopoles. In the running mean method, palaeomagnetic poles from a continent are assigned absolute ages, a time window is selected (e.g., 20 million years), and all palaeomagnetic poles with ages falling within the time window are averaged. Using Fisher statistics, 95% confidence ellipses (known as A95 when averaging poles) can be calculated for each mean pole (Figure 8B). Both the spline method and the running mean technique are effective in averaging out random noise and allowing the basic pattern of APW paths to be determined.

The small circle method is based on the fact that movements of continents, APW paths, hotspot trails, ocean fracture zones, etc., must describe small circular paths if the Euler pole is kept constant. It is reasonable to assume that continents may drift around Euler poles that are kept constant for, say, some tens of millions of years. One can therefore fit APW segments along an APW path. This is demonstrated in Figure 8C where we can fit a small circle to Baltica poles from 475 to 421 Ma. However, after 421 Ma, the path changed direction markedly and this resulted from the collision of Baltica with Laurentia (North America, Greenland, and the British Isles north of the Iapetus Suture), which radically changed the plate tectonic boundary conditions and the APW path for Baltica.

Palaeolatitudes and Drift Rates – Links to Facies

Based on APW paths, we can calculate palaeolatitudes and plate velocities for a specific geographical location. Plate velocities are minimum velocities as the longitude is unconstrained; we only calculate latitudinal velocities. Figure 9 shows an example of such calculations based on the APW path in Figure 8A. In this diagram, we have also separated the northward and southward drift of Baltica. Drift velocities are typically below 8 cm per year, but peak velocities of around 14 cm per year are seen after collision of Baltica with Laurentia in Late Silurian–Early Devonian times.

The calculation of latitudinal velocities is important in order to check whether drift rates are compatible with ‘modern’ plate tectonic velocities. A rate of 18 cm per year (India) is the highest reliable value reported for the last 65 million years. When values appear unrealistically high (e.g., more than 20–30 cm per year), some authors have appealed to true polar wander (TPW) as a plausible explanation. TPW is a highly controversial subject that implies rapid tilting of the Earth’s rotation axis, and is not generally accepted.

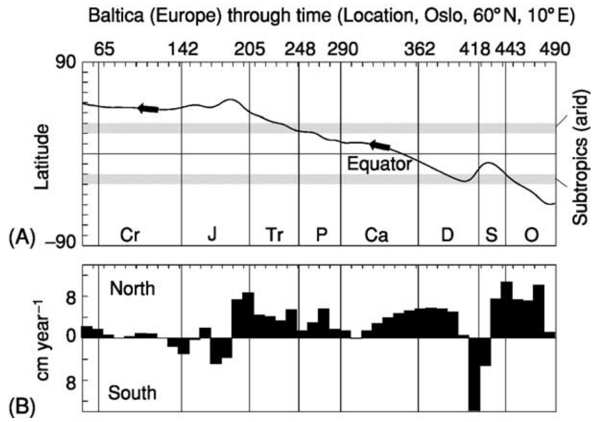

Figure 9 Latitude motion (A) and velocities (B) for Baltica (city of Oslo) from Ordovician to present times based on palaeomagnetic data (Figure 8A). Baltica was in the southern hemisphere during Ordovician through Devonian times, crossed into the northern hemisphere in the Early Carboniferous, and continued a general northward drift throughout the Mesozoic and Cenozoic. Northward and southward latitudinal translations throughout the Early Palaeozoic were accompanied by velocity peaks in the Early Silurian (northward) and the earliest Devonian (southward). Cr, Cretaceous; J, Jurassic; Tr, Triassic; P, Permian; Ca, Carboniferous; D, Devonian; S, Silurian; O, Ordovician.

The distribution of climatically sensitive sediments, such as glacial deposits, coal, carbonates, and evaporites, is useful to check the palaeolatitudes derived from palaeomagnetic data. Glacial deposits are usually confined to polar latitudes and, except during the recent ice ages, there is no evidence for such deposits in Southern Baltica, as predicted by the palaeomagnetic data (maximum 60° S in the Early Ordovician). Carbonates, particularly in massive build-ups, such as reefs, are more common in lower latitudes. During Ordovician and Silurian times, Baltica drifted to subtropical and tropical latitudes, as witnessed by the presence of Bahamian-type reefs in Southern Baltica. Evaporites typically record dry climates within the subtropics (20–30°). During the Late Permian, Baltica was located at subtropical northerly latitudes, and the Late Permian coincides with large evaporite deposits in the North Sea area that subsequently became important in hydrocarbon trap development.

Palaeomagnetism and Palaeogeography: the Big Picture

Quantitative reconstructions are most commonly derived from hotspot traces (Cretaceous–Tertiary times) and ocean-floor magnetic anomalies, but prior to the earliest *in situ* ocean floor preserved today (approximately 175 Ma), the positioning of continents can only be quantitatively recognized by palaeomagnetism. As longitude is unknown from palaeomagnetic data, the identification and discrimination of faunas and floras can indicate that continents with similar faunas were in proximity to one another; conversely, different faunas of the same age can indicate the separation of the continents.

In order to construct a global palaeogeographical map, palaeomagnetic data from individual continents or terranes must be compiled and evaluated in terms of reliability. At any given time, palaeomagnetic data may not exist for all continents and additional criteria, such as fauna, flora, facies, and tectonic and

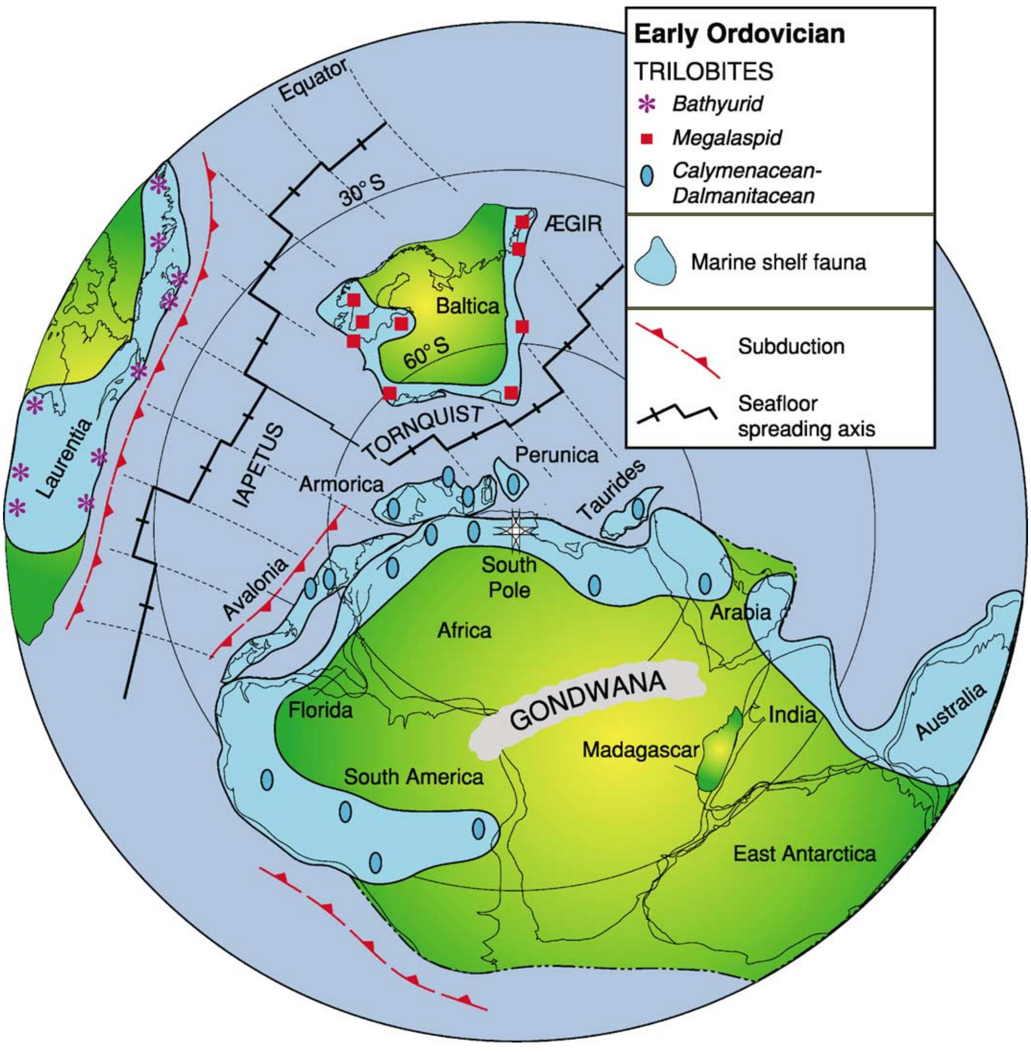

Figure 10 Reconstruction for Early Ordovician times (490–470 Ma) with the major terranes and some key Arenig–Llanvirn trilobite faunas. Laurentia (located at the equator) includes North America, Greenland, and the British Isles north of the Iapetus Suture. Baltica (intermediate southerly latitudes) includes most of Scandinavia and Russia eastwards to the Urals. The core of Gondwana consists of Africa, Arabia, Madagascar, Greater India, most of Antarctica, most of Australia, Florida, and most of South America. Gondwana formed at around 550 Ma and covered more than 90° of latitude in the Early Ordovician. Gondwanan dispersal history commenced with the rifting off of Avalonia at *ca.* 465 Ma. Avalonia includes the British Isles and north-west Europe south of the Iapetus Suture, eastern Newfoundland, most of the Maritime Provinces of Canada, and parts of the eastern USA. Armorica includes the Armorican Massif of Normandy and Brittany, the Massif Central, and the Montagne Noire areas of France, together with parts of the Iberian Peninsula. Perunica comprises the area north of the Barrandian basin of Bohemia. The Taurides comprises most of central and southern Turkey.

magmatic history, must be incorporated in order to construct rational maps. Figure 10 illustrates an Early Ordovician (490–470 Ma) reconstruction; it does not show all the continents or terranes that existed at this time, but some major players, such as Gondwana, Baltica, Laurentia, and some selected peri-Gondwanan terranes with palaeomagnetic and/or faunal data (Avalonia, Armorica, Perunica, and Taurides). Not all the Gondwanan continents have reliable palaeomagnetic data for this time. However, Gondwana was amalgamated at around 550 Ma and continental elements, such as South America and India (no palaeomagnetic data), remained attached to Africa until the breakup of the Pangea supercontinent during the Mesozoic.

An integrated approach of palaeomagnetic and faunal analysis is applicable for the entire Phanerozoic, but works best for the Early Palaeozoic and, notably, the Early Ordovician (Figure 10). At this time, Gondwana stretched from the south pole (Africa) to the equator (Australia and East Antarctica), Baltica occupied intermediate southerly latitudes, separated by the Tornquist Sea, whereas Laurentia straddled the equator. Laurentia was separated from both Baltica and Gondwana by the Iapetus Ocean that had opened in the Late Precambrian. The Iapetus Ocean (approximately 5000 km across the British sector) and the Tornquist Sea (approximately 1100 km between southern Baltica and Armorica–Perunica) were at their widest. This is probably why benthic trilobites from Laurentia (bathyurid) and north-west Gondwana (calymenacean–dalmanitacean) are so markedly different from those of Baltica (megalaspid).

In summary, palaeomagnetism can be seen to be the best and only quantitative method to establish the positions of old terranes and continents as they drifted across the globe over geological time.

See Also

Analytical Methods: Geochronological Techniques. **Gondwanaland and Gondwana**. **Magnetostratigraphy**. **Mantle Plumes and Hot Spots**. **Palaeozoic:** Ordovician. **Plate Tectonics**.

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PALAEONTOLOGY

L R M Cocks, The Natural History Museum, London, UK

Introduction

The word 'fossil' has no exact definition, but it is loosely taken to mean any organism whose remains or traces of remains are preserved in some kind of sediment. The term is derived from the Latin *fossare*, meaning 'to dig'. The definition implies nothing about age: fossils range from those that are about three billion years old to those that are preserved in lime-saturated water deposited only a few days or even hours ago. The study of fossils is termed palaeontology, which is a word derived from Greek that literally means 'knowledge of ancient things'; however, for more than two centuries the word 'palaeontology' has been restricted to the study of formerly living (organic) not inorganic remains or traces. The word palaeobiology is sometimes used as an alternative. In this encyclopaedia there follow numerous