question requires investigation of the way stressors impact the community, and whether it is the dominant or the rare species that are most sensitive, and therefore most rewarding for study in detecting impacts.

Interpretation of impacts also has to proceed against a background of natural changes in benthic communities caused by little-understood, year-to-year differences in annual recruitment. In establishing a baseline there is a need also to take into account the little-understood effects of bottom trawling on coastal benthos. Such disturbance in parts of the North Sea may date back at least 100 years, and now means that virtually every square meter of bottom is trawled over at least once a year. Such monitoring has in the past entailed costly benthic survey and tedious analysis of samples to species level. Consequently, there has been effort to see whether the effects of stress can be detected at higher taxonomic levels, such as families. Higher taxonomic levels may more closely reflect gradients in contamination than they do abundance of individual species because of the statistical noise generated from natural recruitment variability and from seasonal cycles such as reproduction. This hierarchical structure of macrobenthic response means that, as stress increases, the adaptability of first individual animals, then the species, and then genus, family, and so on, is exceeded so that the stress is manifest at progressively higher taxonomic level.

Such new approaches, along with the nascent awareness of conservation of the rich benthic diversity, and with a need for improved environmental impact assessment on the deep continental margin, should ensure a continued active scientific interest in macrobenthos in the years to come.

See also

Benthic Boundary Layer Effects. Benthic Foraminifera. Benthic Organisms Overview. Coral Reefs. Deep-sea Fauna. Demersal Fishes. Fiordic Ecosystems. Grabs for Shelf Benthic Sampling. Meiobenthos. Microphytobenthos. Phytobenthos. Pollution: Effects on Marine Communities. Rocky Shores. Sandy Beaches, Biology of.

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MAGNETICS

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Introduction

Since World War II it has been possible to measure the variations in the intensity of the Earth's magnetic field over the oceans from aircraft or ships. In the 1950s the first detailed magnetic survey of an oceanic area, in the north-east Pacific, revealed a remarkable 'grain' of linear magnetic anomalies, quite unlike the anomaly pattern observed over the continents. In the 1960s it was realized that these linear anomalies result from a combination of seafloor spreading and reversals of the Earth's magnetic field. Hence they provide a detailed record of both the evolution of the ocean basins, and the timing of reversals of the Earth's magnetic field, during the past 160 million years. In addition, because of the dipolar nature of the field and the dominance of 'fossil' magnetization in the oceanic crust, the linear anomalies formed at midocean ridge crests, and the anomalies developed over isolated submarine volcanoes, can sometimes yield paleomagnetic information, such as the latitude at which these features were formed.

Units

In the SI system the unit of magnetic induction, or flux density (which geophysicists refer to as 'field intensity'), is the tesla (T). Because the magnitude of the Earth's magnetic field and magnetic anomalies is very small compared to 1 T, they are usually specified in nanoteslas (nT) ($1\,\mathrm{nT} = 10^{-9}\,\mathrm{T}$). Fortunately, an old geophysical unit called the gamma is equivalent to 1 nT.

History of Measurement

William Gilbert, one-time physician to Elizabeth I of England, is thought to have been the first person to realize that the form of the Earth's magnetic field is essentially the same as that about a uniformly magnetized sphere. This is also equivalent to the field about a bar magnet (or magnetic dipole) placed at the center of the Earth, and aligned along the rotational axis (**Figure 1**). Certainly in terms of the written historical record he was the first person to propose this, in his Latin text *De Magnete*, published in 1600. Presumably, with the extension of European exploration to more southerly latitudes in the late fifteenth and the sixteenth centuries, mariners had problems with their compasses that Gilbert realized could be explained if the vertical component of the Earth's magnetic field varies with latitude. Accurate measurements of the direction of the Earth's magnetic field at London date from Gilbert's time. Measurement of the strength or intensity of the field, however, was not possible until the early part of the nineteenth century. The equipment used then, and for the following one hundred years or so, included a delicate suspended magnet system and required accurate orientation and leveling before a measurement could be made. Measurements on a moving platform such as a ship were extremely difficult, therefore, and with the advent of iron and steel ships became impossible because of the magnetic fields associated with the ships themselves.

Meanwhile, measurements on land had revealed that although the Earth's magnetic field is, to a first approximation, equivalent to that about an axial dipole as envisaged by Gilbert, there is a considerable nondipole component, on average 20% of the dipole field. Moreover, the field is changing in time, albeit slightly and slowly, in both intensity and direction. As rather more than 70% of the Earth's surface is covered by water and there were few magnetic measurements in these areas, detailed mapping of the field at the surface worldwide was

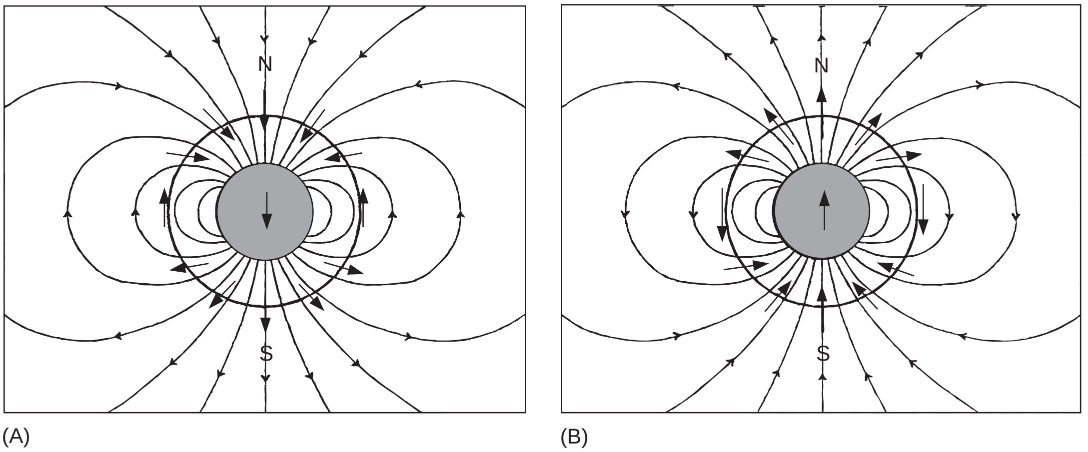

Figure 1 The (A) normal (present-day) and (B) reversed states of the dipolar magnetic field of the Earth. Shaded area shows the Earth's core; heavy arrows indicate directions of the field at the Earth's surface.

seriously hampered. In 1929 the Carnegie Institution of Washington went to the length of building a wooden research ship, the *Carnegie*, to map the Earth's magnetic field in oceanic areas. Similarly the then USSR commissioned a wooden research ship, the *Zarya*, in 1956. However, by this time new electronic instruments had been developed that were capable of making continuous measurements of the total intensity of the Earth's magnetic field from aircraft and ships.

Among the many projects instigated by the Allies in World War II, to counter the submarine menace, was the Magnetic Airborne Detector (MAD) project. The outcome was the development of the fluxgate magnetometer. To increase the sensitivity of the instrument, much of the Earth's magnetic field is 'backed off' by a solenoid producing a biasing field. Initially it was not possible to produce a constant biasing field and the instruments tended to 'drift,' which meant that they were not ideal for scientific purposes. After the war these instruments were redeployed for use in conducting aeromagnetic surveys over land areas, in connection with oil and mineral exploration, and then modified for use from ships. In both instances the detector was housed in a 'fish' that was towed, so as to remove it from the magnetic fields associated with the ship or aircraft.

In the 1950s the proton-precession magnetometer was developed, which had the advantage of achieving the same or somewhat better sensitivity (about 1 part in 50 000) without the problem of drift. Since 1970 even more sensitive magnetometers have been developed – the optical absorption magnetometers. Although in some ways superseded by these magnetometers based on proton and electron precession, fluxgates are still widely used because they have the advantage of measuring the component of the field directed along the axis of the detector rather than the total ambient field. This property makes them particularly useful in satellites, for example, where they are often used as components in orientation systems, at the same time yielding measurements of the Earth's magnetic field.

Measurements of the Earth's magnetic field in oceanic areas are, therefore, now relatively routine, whether they be from satellites, aircraft, ships, submersibles, or remotely operated or deeply towed vehicles.

Nature of the Earth's Magnetic Field

The differences between the measured magnetic field about the Earth and that predicted for a central and axially aligned dipole are considerable. Best-known is the difference between the magnetic poles – where the field is directed vertically – and the rotational, geographic, poles of the Earth. As a result, for most points on the Earth's surface there is an angular difference between the directions to true north and to magnetic north. Mariners refer to this as the magnetic variation; scientists refer to it as the magnetic declination. The centered dipole that best fits the observed field predicts a field strength of 30 000 nT around the equator and a maximum value of 60 000 nT at both poles. The actual field departs considerably from this, as can be seen in **Figure 2**. The intensity and direction of the field, or any of their components (such as magnetic variation) at any one point, vary with time, typically by tens of nanoteslas and a few minutes of arc per year. This is known as the secular variation of the field. The form of the Earth's magnetic field and its secular variation is thought to derive from the fact that it is generated in the outer, fluid core of the Earth, which is metallic (largely iron and nickel) and hence a good electrical conductor. Convective motions of this fluid conductor carrying electrical currents and interacting with magnetic fields produce a dynamo-like effect and an external magnetic field. The essential axial symmetry of this field is probably determined by the influence of the Coriolis force on the precise nature of the convective motions. Historical, archeomagnetic and paleomagnetic data suggest that although the field at any one time departs considerably from that predicted by a geocentric axial dipole, when averaged over several thousand years, the mean field is very close to that about such a dipole. On even longer, geological, timescales the field intermittently reverses its polarity completely (**Figure 1**), probably within a period of approximately 5000 years. The length of intervals of a particular polarity varies widely from a few tens of thousands of years to a few tens of millions of years.

The secular variation and changes in the polarity of the field result from dynamic processes deep in the Earth's interior. There are also shorter-period variations of the field with time that are of external origin, essentially a result of the interaction of the solar wind with the Earth's magnetic field. The most relevant of these in the present context is the daily or diurnal variation of the field. This is a smooth variation that is greatest during daylight hours and typically has an amplitude of a few tens of nanoteslas. It can be greater, however, near the magnetic equator and poles. Increased solar activity can produce higher-amplitude and more irregular variations, and intense sunspot activity produces global magnetic storms; high-amplitude, short-period

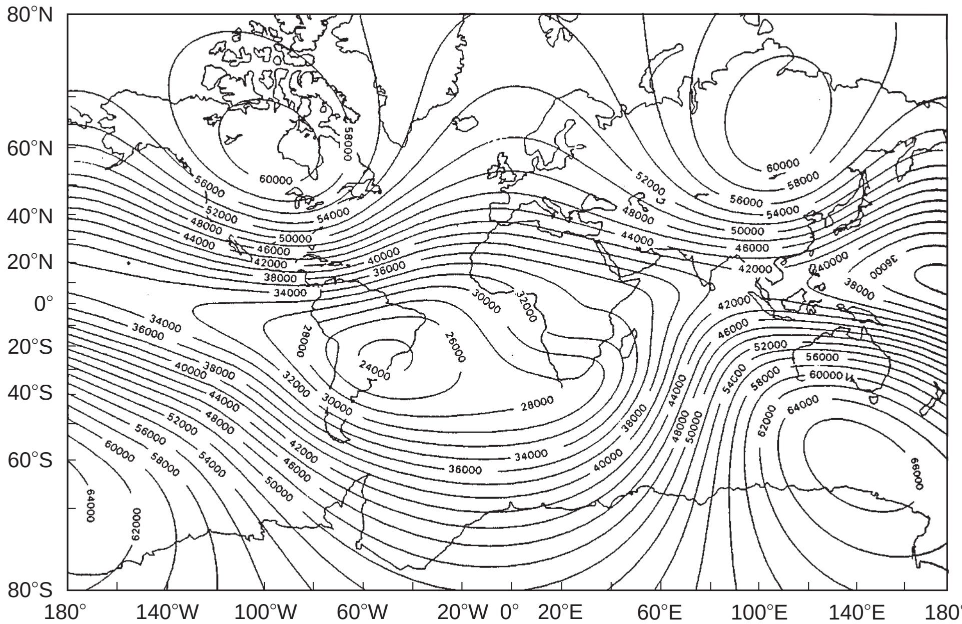

Figure 2 Total intensity of the Earth's magnetic field (in nT) at the Earth's surface (for epoch 1980). (Reproduced from Langel RA, in *Geomagnetism* Vol. 1, JA Jacobs (ed.) © 1987 Academic Press, by permission of the publisher.).

variations during which magnetic surveying has to be discontinued.

Reduction of Magnetic Data

Measurements of the Earth's magnetic field over oceanic areas are corrected for the present-day spatial and time variations of the field described above in order to obtain residual 'anomalies' in the field. These are caused by magnetization contrasts at or near the Earth's surface, i.e., in the upper lithosphere. Such anomalies should therefore yield information on the magnetization and structure of the oceanic crust.

For measurements made at or near the Earth's surface, i.e. from ships, aircraft, or submersibles, the secular and diurnal variation of the field can often be ignored because these effects are very small compared to the amplitudes of the anomalies being mapped. However, should a very accurate survey be required, secular variation has to be taken into account if surveys made at different times are being combined and the observations should be corrected for diurnal variation using records from nearby land stations or moored buoys. If there are sufficient 'cross-overs' during the survey (i.e. repeat measurements at the same point), it may also be possible, indeed preferable, to use these to correct for diurnal variation. It may also be necessary to correct for any magnetic effect of the moving platform itself. If present, this effect will vary according to the direction of travel.

For many purposes, however, the above corrections are so small that they can be ignored. The final correction, the removal of the main or 'regional' field of the Earth, that is, the field generated in the Earth's core, must always be applied. In theory this should be simple. The depth to the core–mantle boundary, 2900 km, means that the field originating in the core should have a smooth, long-wavelength variation at or above the Earth's surface. The magnetization contrasts within a few tens of kilometers of the Earth's surface will produce anomalies of much shorter wavelength. Between these two source regions any magnetic minerals in the mantle are at a temperature above their Curie temperature and are effectively nonmagnetic. In practice, because of its complexity and because it is changing with time, it has proved difficult to accurately define the main field of the Earth, i.e. that originating in the core, and the way in which it is changing with time. The first attempt to define a global 'International Geomagnetic Reference Field' was made in the 1960s and, although revised and greatly improved at five-year intervals since then, its level, if not its gradients, can still be seen to be slightly incorrect for certain oceanic areas. As a result, particularly in the past, the regional field for particular profiles or surveys has been obtained by fitting a smooth long-wavelength curve or surface to the observed data.

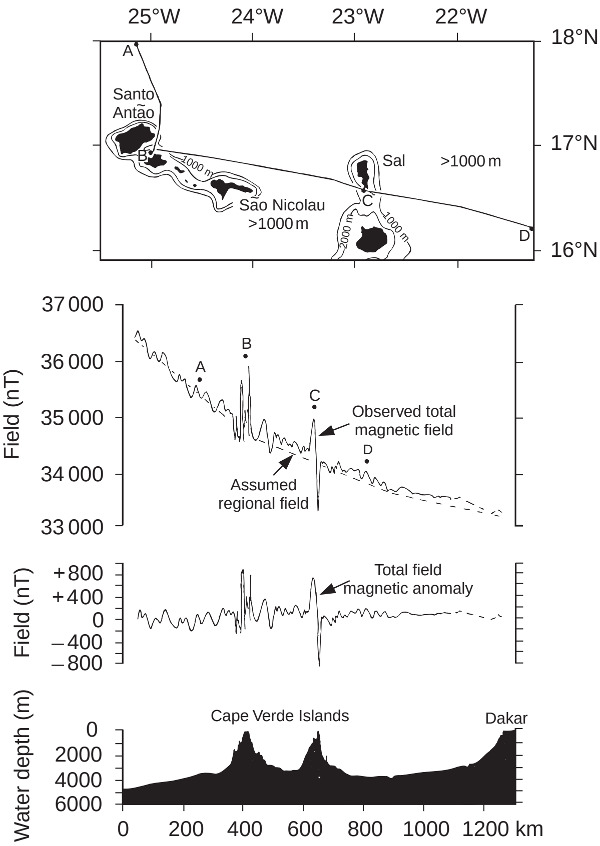

Figure 3 Magnetic profile recorded by a fluxgate magnetometer between the Cape Verde islands and Dakar, Senegal. The ship's track close to the islands is shown in the upper part of the diagram. The dashed line indicates the regional field used to calculate total field magnetic anomalies. The high-amplitude, short-wavelength anomalies close to the Cape Verde Islands reflect the presence of highly magnetic volcanic rocks at shallow depth. (Reproduced from Heezen BC, *et al.*, in *Deep-Sea Research,* Vol. 1 © 1953 Elsevier Science, by permission of the publisher.)

Once the long-wavelength 'regional field' has been removed from magnetic data, the resulting 'residual' or 'total-field' anomalies are assumed to result from magnetization contrasts within the upper, magnetic, part of the Earth's lithosphere (**Figure 3**).

Magnetization of Ocean Floor Rocks

All minerals, and hence all rocks, exhibit magnetic properties. However, the magnetization that most rock types acquire in the relatively weak magnetic field of the Earth is insufficient to produce significant anomalies in the Earth's magnetic field, particularly when this is measured at some distance from the rocks, as is typically the case in oceanic areas. For rocks to be capable of producing appreciable anomalies they must contain more than a few percent by volume of 'ferromagnetic' minerals, that is, certain oxides and sulfides containing iron, notably magnetite (Fe_3O_4). Most sediments and 'acid' (silica-rich) igneous rocks, such as granite, do not meet this criterion. Basic (silica-deficient) igneous rocks such as basalts, and the coarser-grained but chemically equivalent gabbros, and ultrabasic rocks such as peridotite do contain a higher proportion of iron oxides and are capable of producing anomalies. Metamorphic rocks, formed when preexisting rocks are subjected to high temperatures and/or pressures, are typically weakly magnetized except for some formed from basic or ultrabasic igneous rocks.

Apart from its sedimentary veneer, the upper part of the oceanic lithosphere consists almost entirely of basic and ultrabasic rocks, i.e. basalts, gabbros, and peridotites. This is a consequence of the way in which it is formed by the process of seafloor spreading. At midocean ridge crests the ultrabasic peridotite of the Earth's mantle undergoes partial melting, producing basic magma that rises and collects as a magma chamber within oceanic crust. Solidification of such magma chambers ultimately forms the main crustal layer of gabbro, but not before some ultrabasic rocks have formed at the base of the chamber from the accumulation of first formed crystals, and magma has been extruded through near vertical fissures to form pillow basalts on the seafloor. Solidification of the magma in these fissures forms a layer consisting of dikes between the gabbro and the basalts.

Thus, because of the rock types present, the oceanic crust and upper mantle are relatively strongly magnetized and capable of producing large-amplitude anomalies in the Earth's magnetic field, even when measured at sea level. The thickness of the magnetic layer is determined by the depth to the Curie point isotherm. Because of the way in which the oceanic lithosphere is formed, by spreading about ridge crests, this varies from a few kilometers depth within the crust at ridge crests to a depth of approximately 40 km in oceanic lithosphere that is 100 million years, or more, in age. In places the thermal regime associated with seafloor spreading is modified by mantle 'hot spots.' As a result there is an enhanced degree of partial melting of the mantle and the larger volumes of magma produced extrude on to the seafloor to form seamounts, oceanic islands and, exceptionally, oceanic plateaux. These all involve an appreciable thickening of the oceanic crust, but the rock types involved are all essentially basic and potentially strongly magnetic (**Figure 3**).

Observed Anomalies

The marked contrast in the way in which continental and oceanic crust are formed, and hence in the predominant rock types in each setting, gives rise to a striking difference in the character of the total field anomalies developed over the two types of lithosphere. Within the continents the variety of rock types in mountain belts and their juxtaposition by folding and faulting produces magnetization contrasts and anomalies that delineate the general trend of the belt. Areas of igneous activity that include basic igneous rocks are characterized by very large-amplitude and typically short-wavelength anomalies, and sedimentary basins and extensive areas of granite are quiet magnetically. This pattern of anomalies is characteristic of the continental shelves out to the continental slope, the true geological boundary between the continents and the oceans, although, in that the shelves are typically underlain by a great thickness of sediments, they are often magnetically 'quiet.'

Deep sea areas, that is, those underlain by oceanic lithosphere, are characterized by remarkably linear and parallel anomalies that extend for hundreds of kilometers and are truncated and offset by fracture zones (**Figure 4**). The fracture zones are formed by

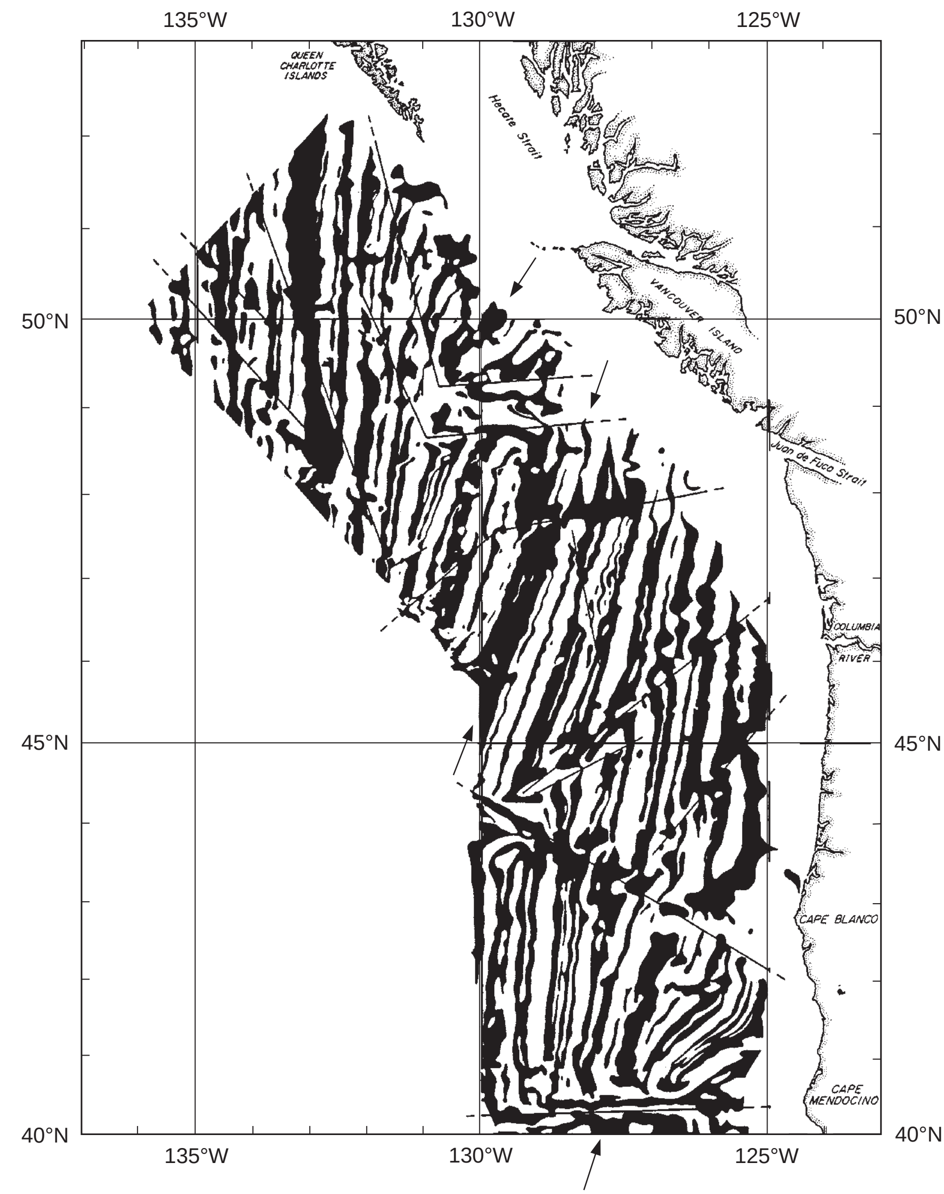

Figure 4 Linear magnetic anomalies in the north–east Pacific. Areas of positive anomaly are shown in black. Straight lines indicate faults offsetting the anomaly pattern; arrows, the axes of three short ridge lengths in the area – from north to south, the Explorer, Juan de Fuca and Gorda ridges. (Based on Figure 1 of Raff AD and Mason RG in *Bull. Geol. Soc. Amer.*, Vol 72. © 1961 *Geological Society of America.* Reproduced by permission of the publisher.)

the transform faults that offset the crest of the midocean ridge system. Thus the linear magnetic anomalies parallel the ridge crests. The anomalies are remarkable for their linearity, their high amplitude and the steep magnetic gradients that separate highs from lows. Any explanation of them in terms of linear structures and/or lateral variations in rock type within the oceanic crust is extremely improbable. It transpires that they result from a combination of sea floor spreading and reversals of the Earth's magnetic field. As new oceanic crust and upper mantle form at a ridge crest they acquire a permanent (remanent) magnetization which parallels the ambient direction of the field. If, as spreading occurs, the Earth's magnetic field reverses, then the ribbon of newly formed oceanic lithosphere along the whole length of the spreading ridge system acquires a remanent magnetisation in the opposite direction. It is these contrasts between normally and reversely magnetized material which produce the high amplitude linear anomalies and the steep gradients between them.

Rates of seafloor spreading vary greatly for different ridges and the interval between reversals of the Earth's magnetic field is also very variable throughout geological time. However, rates of spreading are typically a few tens of millimeters per year and the average polarity interval is about 0.5 million years. Thus typical linear anomalies are 10–20 km in width. Initially, spreading rates could only be reliably determined for the past 3.5 million years. For this period the reversal timescale had been independently determined from measurements of the age and polarity of remanent magnetization of both subaerial lava flows and deep-sea sediments, and it is clearly reproduced in the anomalies recorded across midocean ridge crests (**Figure 5**). With the dating of older oceanic crust by the international Deep Sea Drilling Program it became possible to deduce spreading rates at earlier times and to calibrate the timescale of reversals of the Earth's magnetic field implied by the older linear anomalies. In this way the geomagnetic reversal timescale for the past 160 million years has been deduced (**Figure 6**).

In recording the times at which the Earth's magnetic field has reversed its polarity, the linear magnetic anomalies also serve as time or growth lines that reveal the evolution of the ocean basins in terms of seafloor spreading (**Figure 7**). Thus it is possible to accurately reconstruct the most recent phase of continental drift (during the past 180 million years) when a former supercontinent was split up to form the present-day continents and the Atlantic and Indian Oceans. The Pacific Ocean was formed during the same period, and in the process

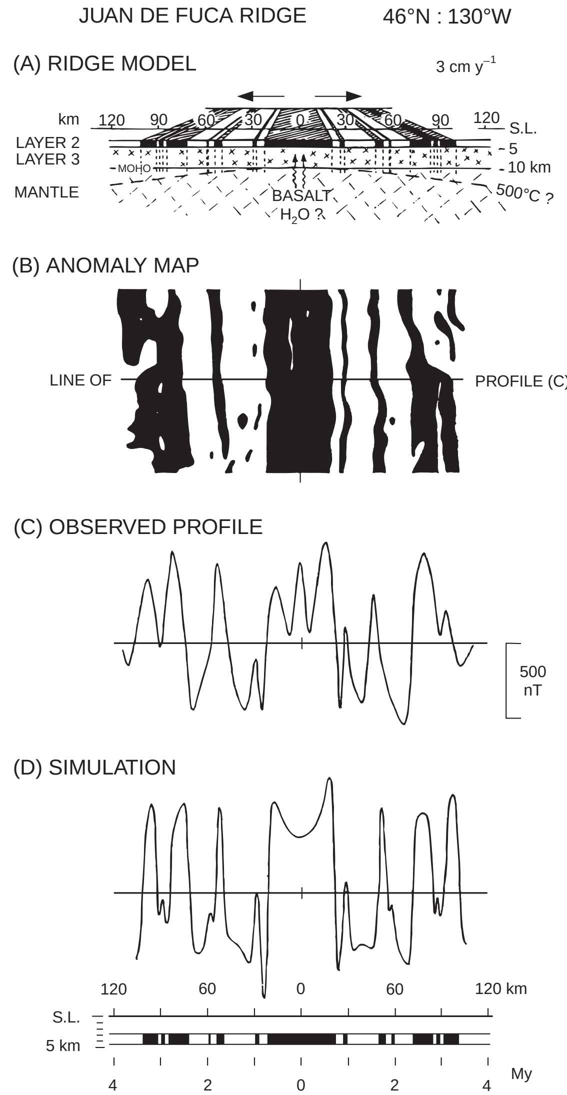

Figure 5 (A) Schematic crustal model for the Juan de Fuca Ridge, south-west of Vancouver Island. Shaded material in layer 2 is normally magnetized; unshaded material is reversely magnetized. SL = sea level. (B) Part of the summary map of magnetic anomalies recorded over the Juan de Fuca Ridge (**Figure 4**). (C) Total field magnetic anomaly profile along the line indicated in (B). (D) Computed profile assuming the model and reversal timescale for the past 3.5 million years. (Reproduced from Vine FJ in *The History of the Earth's Crust.* RA Phinney (ed.). © 1968 Princeton University Press, by permission of the publisher.)

dispersed marginal fragments of the supercontinent around its rim where they are now recognized as 'suspect terranes.' In contrast to this very detailed record of spreading and drift for the past 180 million years there is no such record for the previous 96% of geological time, and earlier phases of drift and mountain building have to be deduced from the more complex and fragmentary geological

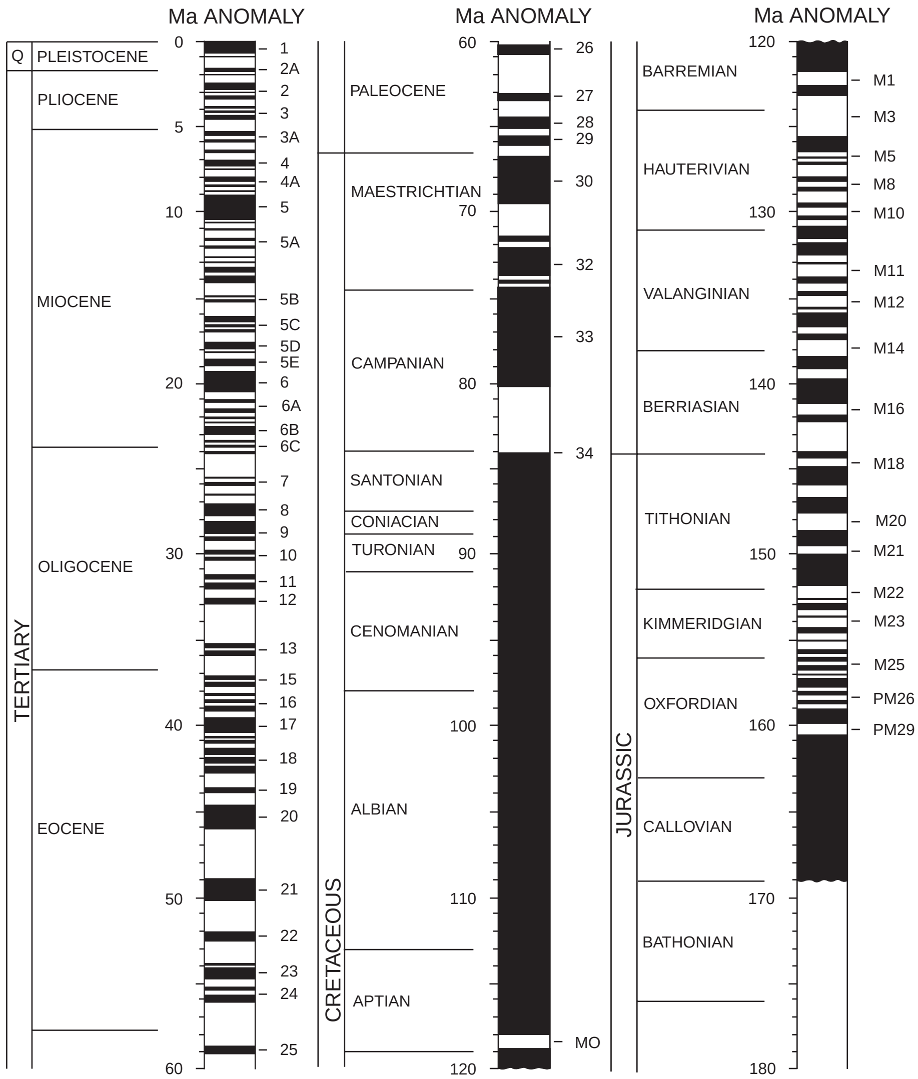

Figure 6 Geomagnetic polarity timescale for the past 160 million years. (Reproduced from Jones EJW, *Marine Geophysics*. © 1999 John Wiley and Sons, by permission of the publisher.)

record within the remaining 40% of the Earth's surface that is covered by continental crust.

Paleomagnetic Information Contained in Oceanic Magnetic Anomalies

As a result of the dipolar nature of the Earth's magnetic field, whereby its inclination to the horizontal varies systematically from the equator to the poles (**Figure 1**), anomalies in the total field over a relatively simple and symmetrical feature such as the central, normally magnetized ribbon of crust at a midocean ridge crest typically have an asymmetry (**Figure 8**). The exceptions occur at the poles and across ridges trending exactly north–south, where the anomaly is a symmetrical high, and over an east–west trending ridge at the Equator, where the anomaly is a symmetrical low. For all other latitudes, and all orientations other than north–south, the degree of asymmetry, or the 'phase-shift' of the anomaly, is a function of the latitude and orientation. As a result of spreading, all older linear anomalies, unless formed about a north–south trending, east–west spreading ridge, will now be at a different latitude from that at which they were formed, and the direction of their remanent magnetization will be different from the direction of the ambient magnetic field. As a consequence, the asymmetry of the anomaly is different from what one would predict for similarly directed remanence and ambient field. This difference

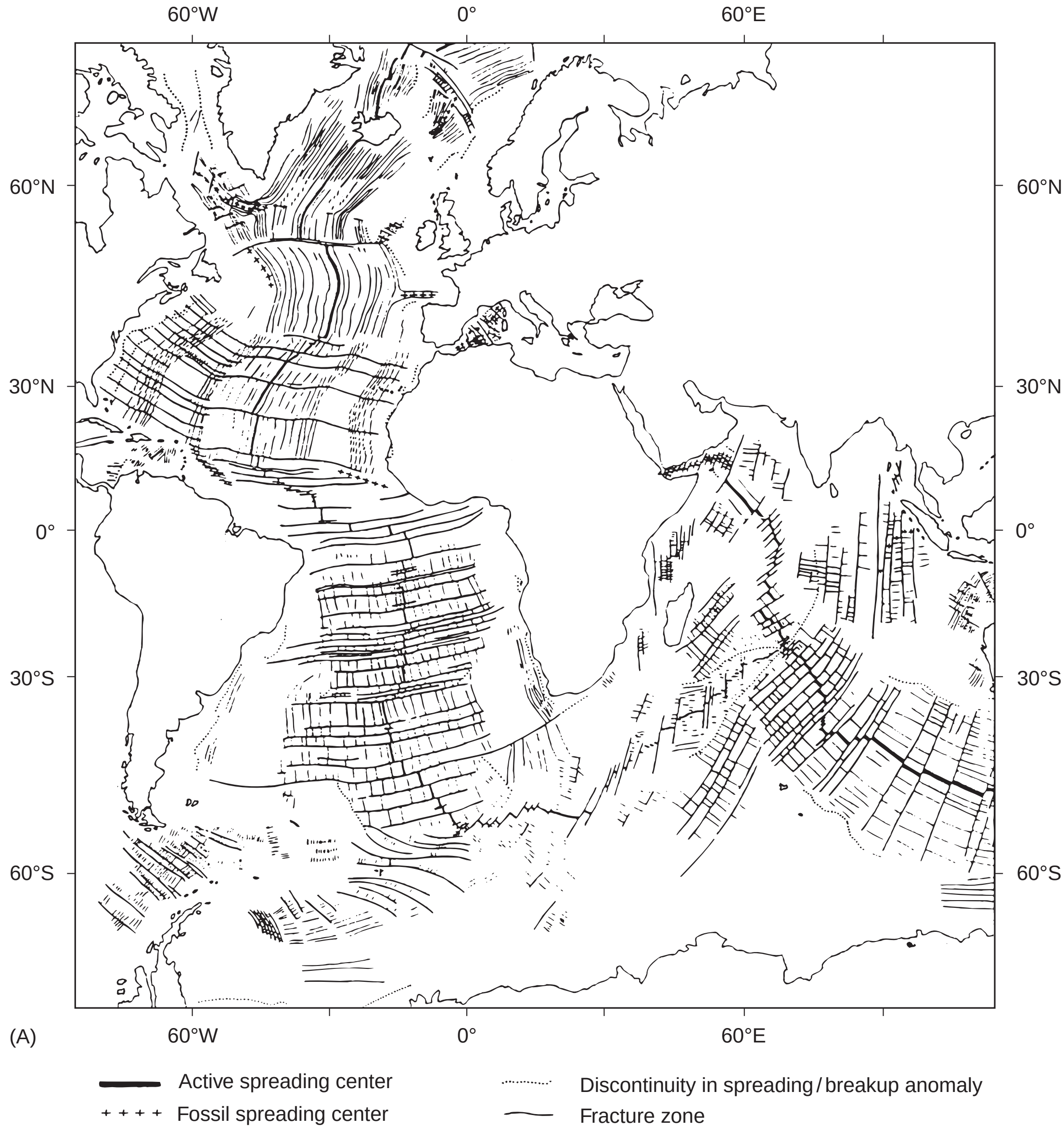

Figure 7 A compilation of the trends of linear magnetic anomalies in (A) the Atlantic and Indian Oceans and (B) the Pacific Ocean. (Reproduced from Jones EJW, *Marine Geophysics.* © 1999 John Wiley and Sons, by permission of the publisher.)

between the observed and predicted phase shift reflects the latitudinal change. Although the interpretation of such data is somewhat ambiguous if the orientation of the anomaly is thought to have changed since the time of formation, it does provide paleolatitude, i.e. paleomagnetic, information for oceanic areas, which is otherwise rather sparse. As with all paleomagnetic data, they can be used to test independently derived models for the 'absolute' motion of plates and plate boundaries, such as ridge crests, across the face of the Earth.

In theory the ambiguity of paleolatitude determination mentioned above can be removed by carrying out the analysis on anomalies of the same age on either side of a particular ridge. An intriguing result of such studies, however, is that in some cases different latitudes of formation are deduced for the same anomaly on either side of the ridge. In that they were formed at the same time and at the same ridge crest, this cannot be so, and the result is giving us yet more information on the geometry or magnetization of the source region. Such a result could be produced by a decay in the intensity of the Earth's magnetic field or an increase in the number of magnetic excursions or very short-lived reversals as a particular polarity interval progresses, and/or a more complex geometry for the boundaries between normally and reversely magnet-

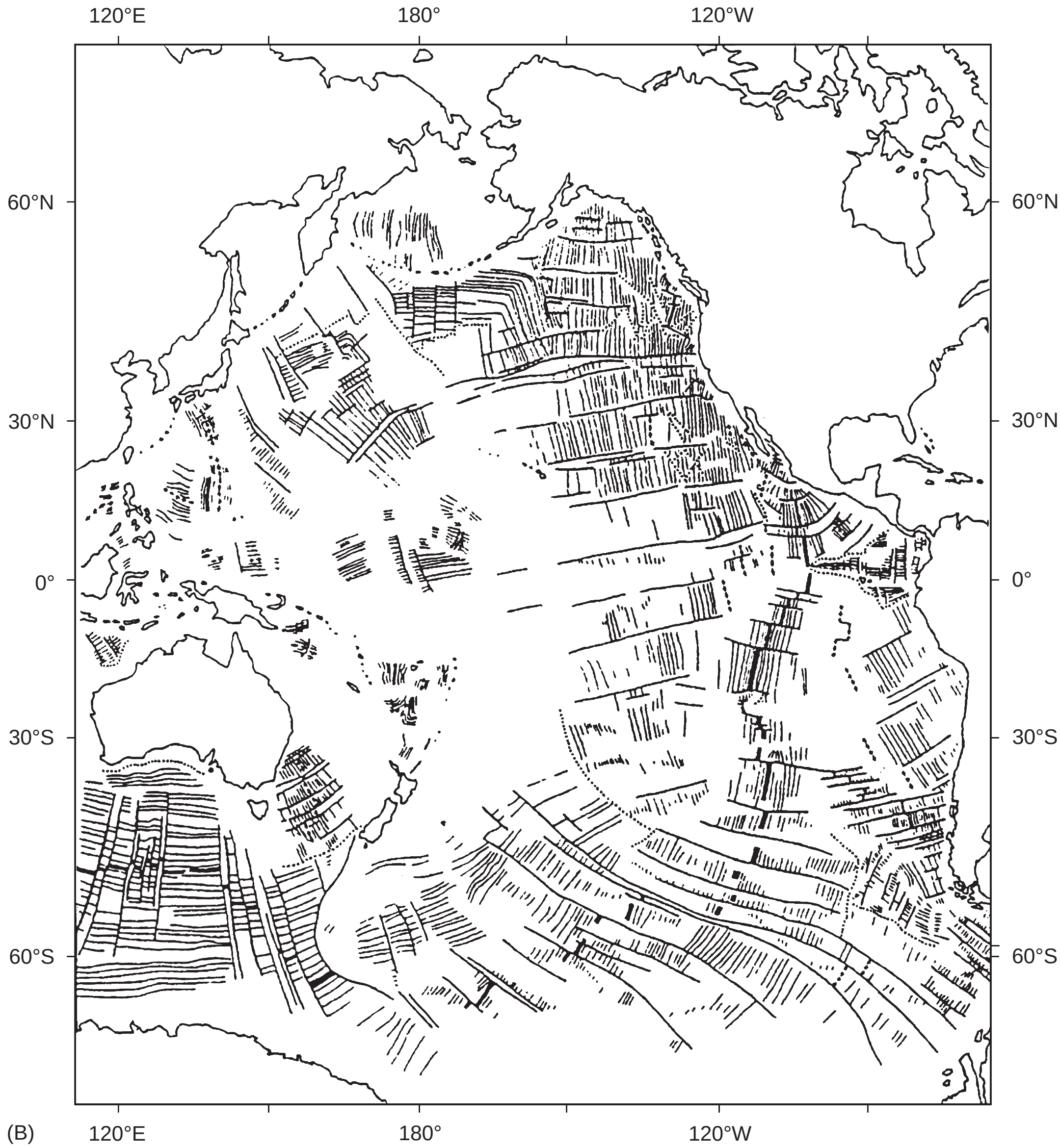

Figure 7 *Continued.*

ized material in the oceanic lithosphere (**Figure 9**). One would predict inwardly sloping boundaries in the upper crust as a result of the lateral extrusion and vertical accretion of lava flows, and outwardly sloping boundaries in the lower crust reflecting the location of the Curie point isotherm at a ridge crest. Increasingly this latter interpretation seems to be the more likely.

The isolated magnetic anomalies that are associated with submarine volcanoes and oceanic islands can also, sometimes, yield useful paleomagnetic information. If it can be assumed that the edifice is uniformly magnetized throughout, then, knowing the topographic shape of the feature, the magnetic anomaly over it can be analyzed to yield a direction and intensity of magnetization. Strictly, this is the sum of the induced and remanent magnetization of the feature, but the former is almost certainly small compared to the latter and the direction of magnetization deduced can be considered to be a good approximation to the direction of the remanent magnetization and hence treated as a paleomagnetic result. It seems probable that many of these features, particularly the large edifices, were formed during more than one polarity interval, in which case the method is inapplicable. However, many seamounts, particularly those formed during the long interval of normal polarity during mid-Cretaceous time, do yield satisfactory results. Examples are the New England seamounts of the

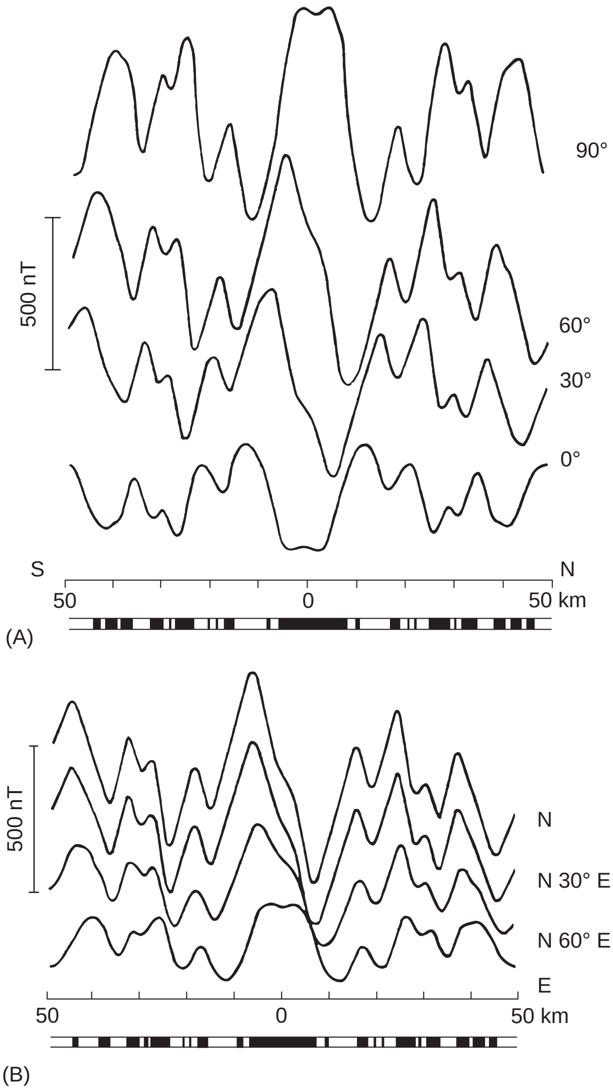

Figure 8 Variation of the magnetic anomaly pattern (A) with geomagnetic latitude (all profiles are N–S; angles refer to magnetic inclination; no vertical exaggeration) and (B) with direction of the profile at fixed latitude (magnetic inclination is 45° in all cases; no vertical exaggeration). (Reproduced from Kearey P and Vine FJ. *Global Tectonics*. © 1996 Blackwell Science, by permission of the publisher.)

north–west Atlantic, and the Musician seamounts of the central Pacific. These yield paleomagnetic pole positions that are consistent with other results for the mid-Cretaceous for the North American and Pacific plates, respectively.

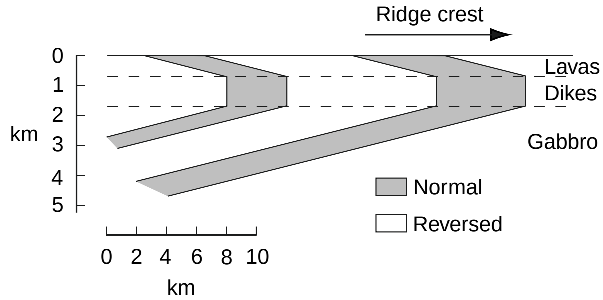

Figure 9 Postulated geometry of normal/reverse magnetization contrasts in the oceanic crust, which could explain the anomalous phase shift of certain linear anomalies.

Conclusion

Thus, because of the dominance of remanent magnetization in the oceanic crust, and the relative simplicity of the way in which it is formed by seafloor spreading and volcanic activity, it records both the history of reversals of the Earth's magnetic field, and the lateral and latitudinal displacement of the crust during the past 160 million years. This record can be played back by measuring the anomalies in the intensity of the Earth's magnetic field over the oceans at the present day.

See also

Deep Sea Drilling Results. Geomagnetic Polarity Time Scale. Mid-ocean Ridge Geochemistry and Petrology. Mid-Ocean Ridge Seismic Structure. Mid-Ocean Ridge Tectonics, Volcanism and Geomorphology. Propagating Rifts and Micro-plates. Seamounts and Off-ridge Volcanism. Seismic Structure.

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MALVINAS CURRENT

See **BRAZIL AND FALKLANDS (MALVINAS) CURRENTS**