east–west trending linear features over a region in the planet’s southern hemisphere. These sub-parallel linear features are attributed to alternating bands of strongly magnetized crust. It is tempting to interpret their origin in the same way as the generation of oceanic magnetic linear features on Earth by a plate tectonic process in the presence of a reversing field. However, as yet the origin of the magnetic anomalies on Mars is not understood.

**Jupiter** has been known since 1955 to possess a strong magnetic field, because of the polarized radio emissions associated with it. The spacecraft *Pioneer 10* and *Pioneer 11* in 1973–4 and *Voyager 1* and *2* in 1979 established that the planet has a bow shock and magnetopause. From 1995 to 2003 the spacecraft *Galileo* made extensive surveys of Jupiter’s magnetosphere. The huge magnetosphere encounters the solar wind about 5,000,000 km “upwind” from the planet; its magnetotail may extend all the way to Saturn. Two reasons account for the great size of the magnetosphere compared to that of Earth. First, the solar wind pressure on the Jovian atmosphere is weaker due to the greater distance from the Sun; secondly, Jupiter’s magnetic field is much stronger than that of the Earth. The dipole moment is almost 20,000$\mu$E which gives a powerful equatorial magnetic field of more than 400,000 nT at Jupiter’s surface. The quadrupole and octupole parts of the non-dipole magnetic field have been found to be proportionately much larger relative to the dipole field than on Earth. The dipole axis is tilted at 9.7° to the rotation axis, and is also displaced from it by 10% of Jupiter’s equatorial radius. The magnetic field of Jupiter results from an active dynamo in the metallic hydrogen core of the planet. The core is probably very large, with a radius up to 75% of the planet’s radius. This would explain the high harmonic content of the magnetic field near the planet.

**Saturn** was reached by *Pioneer 11* in 1979 and the *Voyager 1* and 2 spacecraft in 1980 and 1981, respectively. The on-board magnetometers detected a bow shock and a magnetopause. In 2004 the *Cassini–Huygens* spacecraft entered into orbit around Saturn. In 2005 the Huygens lander descended to the surface of Saturn’s largest moon *Titan* while Cassini continued to orbit and measure the parent planet’s properties. Saturn’s dipole magnetic moment is smaller than expected, but is estimated to be around 500$\mu$E. This gives an equatorial field of 55,000 nT, almost double that of Earth. The magnetic field has a purer dipole character (i.e., the non-dipole components are weaker) than the fields of Jupiter or Earth. The simplest explanation for this is that the field is generated by an active dynamo in a conducting core that is smaller relative to the size of the planet. The axis of the dipole magnetic field lies only about 1° away from the rotation axis, in contrast to 11.4° on Earth and 9.7° on Jupiter.

**Uranus** is unusual in that its spin axis has an obliquity of 97.9°. This means that the rotation axis lies very close to the ecliptic plane, and the orbital planes of its satellites are almost orthogonal to the ecliptic plane. Uranus was visited by *Voyager 2* in January 1986. The spacecraft encountered a bow shock and magnetopause and subsequently entered the magnetosphere of the planet, which extends for 18 planetary radii (460,000 km) towards the Sun. Uranus has a dipole moment about 50 times stronger than Earth’s, giving a surface field of 24,000 nT, comparable to that on Earth. Intriguingly, the axis of the dipole field has a large tilt of about 60° to the rotation axis; there is no explanation for this tilt. Another oddity is that the magnetic field is not centered on the center of the planet, but is displaced by 30% of the planet’s radius along the tilted rotation axis. The quadrupole component of the field is relatively large compared to the dipole. It is therefore supposed that the magnetic field of Uranus is generated at shallow depths within the planet.

**Neptune**, visited by *Voyager 2* in 1989, has a magnetic field with similar characteristics to that of Uranus. It is tilted at 49° to the rotation axis and is offset from the planetary center by 55% of Neptune’s radius. As in the case of Uranus, the magnetic field has a large quadrupole term compared to the dipole and thus probably originates in the outer layers of the planet, rather than in the deep interior.

It is not yet known whether **Pluto** has a magnetic field. The planet is probably too small to have a magnetic field sustained by dynamo action.

There is reasonable confidence that Mercury, Earth, Jupiter, Saturn, and Uranus have active planetary dynamos today. The magnetic data for Mars and Neptune are inconclusive. All available data indicate that Venus and the Moon do not have active dynamos now, but possibly each might have had one earlier in its history.

### 5.5 MAGNETIC SURVEYING

#### 5.5.1 The magnetization of the Earth’s crust

The high-order terms in the energy density spectrum of the geomagnetic field (Fig. 5.33) are related to the magnetization of crustal rocks. Magnetic investigations can therefore yield important data about geological structures. By analogy with gravity anomalies we define a magnetic anomaly as the difference between the measured (and suitably corrected) magnetic field of the Earth and that which would be expected from the International Geomagnetic Reference Field (Section 5.4.4). The magnetic anomaly results from the contrast in magnetization when rocks with different magnetic properties are adjacent to each other, as, for example, when a strongly magnetic basaltic dike intrudes a less magnetic host rock. The stray magnetic fields surrounding the dike disturb the geomagnetic field locally and can be measured with sensitive instruments called magnetometers.

As discussed in Section 5.3.1, each grain of mineral in a rock can be classified as having diamagnetic, paramagnetic or ferromagnetic properties. When the rock is in a magnetic field, the alignment of magnetic moments by the field produces an induced magnetization ($M_i$)
Magnetic field in the geological past. Its direction is usually related to the Earth’s magnetic field, but is related to the Earth’s magnetic field in the rock. Its direction is usually related to the Earth’s magnetic field, but is related to the Earth’s magnetic field in the rock.

Each rock usually contains a tiny quantity of ferromagnetic minerals. As we have seen, these grains can become magnetized permanently during the formation of the rock or by a later mechanism. The remanent magnetization ($M_r$) of the rock is not related to the present-day geomagnetic field, but is related to the Earth’s magnetic field in the geological past. Its direction is usually different from that of the present-day field. A result the directions of $M_r$ and $M_i$ are generally not parallel. The direction of $M_i$ is the same as that of the present field but the direction of $M_r$ is often not known unless it can be measured in rock samples.

The total magnetization of a rock is the sum of the remanent and induced magnetizations. As these have different directions they must be combined as vectors (Fig. 5.40a). The direction of the resultant magnetization of the rock is not parallel to the geomagnetic field. If the intensities of $M_r$ and $M_i$ are similar, it is difficult to interpret the total magnetization. Fortunately, in many important situations $M_r$ and $M_i$ are sufficiently different to permit some simplifying assumptions. The relative importance of the remanent and induced parts of the magnetization is expressed in the Königsberger ratio ($Q_n$), defined as the ratio of the intensity of the remanent magnetization to that of the induced magnetization (i.e., $Q_n = M_r / M_i$).

Two situations are of particular interest. The first is when $Q_n$ is very large (i.e., $Q_n \gg 1$). In this case (Fig. 5.40b), the total magnetization is dominated by the remanent component and its direction is essentially parallel to $M_r$. Oceanic basalts, combined with rapid underwater cooling at oceanic ridges, are an example of rocks with high $Q_n$ ratios. Due to the rapid quenching of the molten lava, titanomagnetite grains form with skeletal structures and very fine grain sizes. The oceanic basalts carry a strong thermoremanent magnetization and often have $Q_n$ values of 100 or greater. This facilitates the interpretation of oceanic magnetic anomalies, because in many cases the induced component can be neglected and the crustal magnetization can be interpreted as if it were entirely remanent.

The other important situation is when $Q_n$ is very small (i.e., $Q_n \ll 1$). This requires the remanent magnetization to be negligible in comparison to the induced magnetization. For example, coarse grained magnetite grains carry multidomain magnetizations (Section 5.3.5.3). The domain walls are easily moved around by a magnetic field. The susceptibility is high and the Earth’s magnetic field can induce a strong magnetization. Any remanent magnetization is usually weak, because it has been subdivided into antiparallel domains. These two factors yield a low value for $Q_n$.

Magnetic investigations of continental crustal rocks for commercial exploitation (e.g., in ancient shield areas) can often be interpreted as cases with $Q_n \ll 1$. The magnetization can then be assumed to be entirely induced (Fig. 5.40c) and oriented parallel to the direction of the present-day geomagnetic field at the measurement site, which is usually known. This makes it easier to design a model to interpret the feature responsible for the magnetic anomaly.

5.5.2 Magnetometers

The instrument used to measure magnetic fields is called a magnetometer. Until the 1940s magnetometers were mechanical instruments that balanced the torque of the magnetic field on a finely balanced compass needle against a restoring force provided by gravity or by the torsion in a suspension fiber. The balance types were cumbersome, delicate and slow to operate. For optimum sensitivity they were designed to measure changes in a selected component of the magnetic field, most commonly the vertical field. This type of magnetometer has now been superseded by more sensitive, robust electronic instruments. The most important of these are the flux-gate, proton-precession and optically pumped magnetometers.
### 5.5.2.1 The flux-gate magnetometer

Some special nickel–iron alloys have very high magnetic susceptibility and very low remanent magnetization. Common examples are Permalloy (78.5% Ni, 21.5% Fe) and Mumetal (77% Ni, 16% Fe, 5% Cu, 2% Cr). The preparation of these alloys involves annealing at very high temperature (1100–1200 °C) to remove lattice defects around which internal stress could produce magnetostrictive energy. After this treatment the coercivity of the alloy is very low (i.e., its magnetization can be changed by a very weak field) and its susceptibility is so high that the Earth's field can induce a magnetization in it that is a considerable proportion of the saturation value.

The sensor of a flux-gate magnetometer consists of two parallel strips of the special alloy (Fig. 5.41a). They are wound in opposite directions with primary energizing coils. When a current flows in the primary coils, the parallel strips become magnetized in opposite directions. A secondary coil wound about the primary pair detects the change in magnetic flux in the cores (Fig. 5.41b), which is zero as soon as the cores saturate. While the primary current is rising or falling, the magnetic flux in each strip changes and a voltage is induced in the secondary coil. If there is no external magnetic field, the signals due to the changing flux are equal and opposite and no output signal is recorded. When the axis of the sensor is aligned with the Earth's magnetic field, the latter is added to the primary field in one strip and subtracted from it in the other. The phases of the magnetic flux in the alloy strips are now different; one saturates before the other. The flux changes in the two alloy strips are no longer equal and opposite. An output voltage is produced in the secondary coil that is proportional to the strength of the component of the Earth's magnetic field along the axis of the sensor.

The flux-gate magnetometer is a vector magnetometer, because it measures the strength of the magnetic field in a particular direction, namely along the axis of the sensor. This requires that the sensor be accurately oriented along the direction of the field component to be measured. For total field measurements three sensors are employed. These are fixed at right angles to each other and connected with a feedback system which rotates the entire unit so that two of the sensors detect zero field. The magnetic field to be measured is then aligned with the axis of the third sensor.

The flux-gate magnetometer does not yield absolute field values. The output is a voltage, which must be calibrated in terms of magnetic field. However, the instrument provides a continuous record of field strength. Its sensitivity of about 1 nT makes it capable of measuring most magnetic anomalies of geophysical interest. It is robust and adaptable to being mounted in an airplane, or towed behind it. The instrument was developed during World War II as a submarine detector. After the war it was used extensively in airborne magnetic surveying.

### 5.5.2.2 The proton-precession magnetometer

Since World War II sensitive magnetometers have been designed around quantum-mechanical properties. The proton-precession magnetometer depends on the fact that the nucleus of the hydrogen atom, a proton, has a magnetic moment proportional to the angular momentum of its spin. Because the angular momentum is quantized, the proton magnetic moment can only have specified values, which are multiples of a fundamental unit called the nuclear magneton. The situation is analogous to the quantization of magnetic moment associated with electron spin, for which the fundamental unit is the Bohr magneton. The ratio of the magnetic moment to the spin angular momentum is called the gyromagnetic ratio \( (\gamma_p) \) of the proton. It is an accurately known fundamental constant with the value \( \gamma_p = 2.675 \times 10^8 \text{s}^{-1} \text{T}^{-1} \).

The proton-precession magnetometer is simple and robust in design. The sensor of the instrument consists of a flask containing a proton-rich liquid, such as water. Around the flask are wound a magnetizing solenoid and a detector coil (Fig. 5.42); some designs use the same solenoid alternately for magnetizing and detection. When the current in the magnetizing solenoid is switched on, it creates a magnetic field of the order of 10 mT, which is about 200 times stronger than the Earth’s field. The
magnetizing field, $B_t$ (= 0.03–0.06 mT)

Fig. 5.42 (a) The elements of a proton-precession magnetometer. (b) Current in the magnetizing coil produces a strong field $F$ that aligns the magnetic moments (“spins”) of the protons. (c) When the field $F$ is switched off, the proton spins precess about the geomagnetic field $B_t$, inducing an alternating current in the coil with the Larmor precessional frequency $f$.

The intensity of the Earth’s magnetic field is in the range 30,000–60,000 nT. The corresponding precessional frequency is approximately 1250–2500 Hz, which is in the audio-frequency range. Accurate measurement of the signal frequency gives an instrumental sensitivity of about 1 nT, but requires a few seconds of observation. Although it gives an absolute value of the field, the proton-precession magnetometer does not give a continuous record. Its portability and simplicity give it advantages for field use.

The flux-gate and proton-precession magnetometers are widely used in magnetic surveying. The two instruments have comparable sensitivities of 0.1–1 nT. In contrast to the flux-gate instrument, which measures the component of the field along its axis, the proton-precession magnetometer cannot measure field components; it is a total-field magnetometer. The total field $B_t$ is the vector sum of the Earth’s magnetic field $B_E$ and the stray magnetic field $\Delta B$ of, say, an orebody. Generally, $\Delta B \ll B_E$, so that the direction of the total field does not deviate far from the Earth’s field. In some applications it is often adequate to regard the measured total field anomaly as the projection of $\Delta B$ along the Earth’s field direction.

### 5.5.2.3 The absorption-cell magnetometer

The absorption-cell magnetometer is also referred to as the alkali-vapor or optically pumped magnetometer. The principle of its operation is based on the quantum-mechanical model of the atom. According to their quantum numbers the electrons of an atom occupy concentric shells about the nucleus with different energy levels. The lowest energy level of an electron is its ground state. The magnetic moment associated with the spin of an electron can be either parallel or antiparallel to an external magnetic field. The energy of the electron is different in each case. This results in the ground state splitting into two sublevels with slightly different energies. The energy difference is proportional to the strength of the magnetic field. The splitting of energy levels in the presence of a magnetic field is called the Zeeman effect.

Absorption-cell magnetometers utilize the Zeeman effect in vapors of alkali elements such as rubidium or cesium, which have only a single valence electron in the outermost energy shell. Consider the schematic representation of an alkali-vapor magnetometer in Fig. 5.43. A polarized light-beam is passed through an absorption cell containing rubidium or cesium vapor and falls on a photodetector cell, which measures the intensity of the light-beam. In the presence of a magnetic field the ground state of the rubidium or cesium is split into two sublevels, $G_1$ and $G_2$. If the exact amount of energy is added to the vapor, the electrons may be raised from their ground state to a higher-energy level, $H$. Suppose that we irradiate the cell with light from which we have filtered out the spectral line corresponding to the energy needed for the transition $G_2H$. The energy for the transition $G_1H$ has not been removed, so the electrons in ground state $G_1$ will receive energy that excites them to level $H$, whereas those in ground state $G_2$ will remain in this state. The energy for these transitions comes from the incident light-beam, which is absorbed in the cell. In due
course, the excited electrons will fall back to one of the more stable ground states. If an electron in excited state H falls back to sublevel G₁ it will be re-excited into level H; but if it falls back to sublevel G₂ it will remain there. In time, this process – called “optical pumping” – will empty sublevel G₁ and fill level G₂. At this stage no more energy can be absorbed from the polarized light-beam and the absorption cell becomes transparent. If we now supply electromagnetic energy to the system in the form of a radio-frequency signal with just the right amount of energy to permit transitions between the populated G₂ and unpopulated G₁ ground sublevels, the balance will be disturbed. The optical pumping will start up again and will continue until the electrons have been expelled from the G₁ level. During this time energy is absorbed from the light-beam and it ceases to be transparent.

In the rubidium-vapor and cesium-vapor magnetometers a polarized light-beam is shone at approximately 45° to the magnetic field direction. In the presence of the Earth’s magnetic field the electrons precess about the field direction at the Larmor precessional frequency. At one part of the precessional cycle an electron spin is almost parallel to the field direction, and one half-cycle later it is nearly antiparallel. The varying absorption causes a fluctuation of intensity of the light-beam at the Larmor frequency. This is detected by the photocell and converted to an alternating current. By means of a feedback circuit the signal is supplied to a coil around the container of rubidium gas and a radio-frequency resonant circuit is created. The ambient geomagnetic field $B_t$ that causes the splitting of the ground state is proportional to the Larmor frequency, and is given by

$$B_t = \frac{2\pi}{\gamma_e} f$$  \hspace{1cm} (5.44)

Here, $\gamma_e$ is the gyromagnetic ratio of the electron, which is known with an accuracy of about 1 part in 10⁷. It is about 1800 times larger than $\gamma_p$, the gyromagnetic ratio of the proton. The precessional frequency is correspondingly higher and easier to measure precisely. The sensitivity of an optically pumped magnetometer is very high, about 0.01 nT, which is an order of magnitude more sensitive than the flux-gate or proton-precession magnetometer.

### 5.5.3 Magnetic surveying

The purpose of magnetic surveying is to identify and describe regions of the Earth’s crust that have unusual (anomalous) magnetizations. In the realm of applied geophysics the anomalous magnetizations might be associated with local mineralization that is potentially of commercial interest, or they could be due to subsurface structures that have a bearing on the location of oil deposits. In global geophysics, magnetic surveying over oceanic ridges provided vital clues that led to the theory of plate tectonics and revealed the polarity history of the Earth’s magnetic field since the Early Jurassic.

Magnetic surveying consists of (1) measuring the terrestrial magnetic field at predetermined points, (2) correcting the measurements for known changes, and (3) comparing the resultant value of the field with the expected value at each measurement station. The expected value of the field at any place is taken to be that of the International Geomagnetic Reference Field (IGRF), described in Section 5.4.4. The difference between the observed and expected values is a magnetic anomaly.

#### 5.5.3.1 Measurement methods

The surveying of magnetic anomalies can be carried out on land, at sea and in the air. In a simple land survey an operator might use a portable magnetometer to measure the field at the surface of the Earth at selected points that
form a grid over a suspected geological structure. This method is slow but it yields a detailed pattern of the magnetic field anomaly over the structure, because the measurements are made close to the source of the anomaly.

In practice, the surveying of magnetic anomalies is most efficiently carried out from an aircraft. The magnetometer must be removed as far as possible from the magnetic environment of the aircraft. This may be achieved by mounting the instrument on a fixed boom, A, several meters long (Fig. 5.44a). Alternatively, the device may be towed behind the aircraft in an aerodynamic housing, B, at the end of a cable 30–150 m long. The “bird” containing the magnetometer then flies behind and below the aircraft. The flight environment is comparatively stable. Airborne magnetometers generally have higher sensitivity ($\approx 0.01 \text{ nT}$) than those used in ground-based surveying (sensitivity $\approx 1 \text{ nT}$). This compensates for the loss in resolution due to the increased distance between the magnetometer and the source of the anomaly. Airborne magnetic surveying is an economical way to reconnoitre a large territory in a short time. It has become a routine part of the initial phases of the geophysical exploration of an uncharted territory.

The magnetic field over the oceans may also be surveyed from the air. However, most of the marine magnetic record has been obtained by shipborne surveying. In the marine application a proton-precession magnetometer mounted in a waterproof “fish” is towed behind the ship at the end of a long cable (Fig. 5.44b). Considering that most research vessels consist of several hundred to several thousand tons of steel, the ship causes a large magnetic disturbance. For example, a research ship of about 1000 tons deadweight causes an anomaly of about 10 nT at a distance of 150 m. To minimize the disturbance of the ship the tow-cable must be about 100–300 m in length. At this distance the “fish” in fact “swims” well below the water surface. Its depth is dependent on the length of the tow-cable and the speed of the ship. At a typical survey speed of 10 km h$^{-1}$ its operational depth is about 10–20 m.

### 5.5.3.2 Magnetic gradiometers

The magnetic gradiometer consists of a pair of alkali-vapor magnetometers maintained at a fixed distance from each other. In ground-based surveying the instruments are mounted at opposite ends of a rigid vertical bar. In airborne usage two magnetometers are flown at a vertical spacing of about 30 m (Fig. 5.44c). The difference in outputs of the two instruments is recorded. If no anomalous body is present, both magnetometers register the Earth’s field equally strongly and the difference in output signals is zero. If a magnetic contrast is present in the subsurface rocks, the magnetometer closest to the structure will detect a stronger signal than the more remote instrument, and there will be a difference in the combined output signals.

The gradiometer emphasizes anomalies from local shallow sources at the expense of large-scale regional variation due to deep-seated sources. Moreover, because the gradiometer registers the difference in signals from the individual magnetometers, there is no need to compensate the measurements for diurnal variation, which affects each individual magnetometer equally. Proton-precession magnetometers are most commonly used in ground-based magnetic gradiometers, while optically pumped magnetometers are favored in airborne gradiometers.

### 5.5.3.3 The survey pattern

In a systematic regional airborne (or marine) magnetic survey the measurements are usually made according to a predetermined pattern. In surveys made with fixed-wing aircraft the survey is usually flown at a constant flight elevation above sea-level (Fig. 5.45a). This is the procedure favored for regional or national surveys, or for the investigation of areas with dramatic topographic relief. The survey focuses on the depth to the magnetic basement, which often underlies less magnetic sedimentary surface
rocks at considerable depth. In regions that are flat or that do not have dramatic topography, it may be possible to fly a survey at low altitude, as close as possible to the magnetic sources. This method would be suitable over ancient shield areas, where the goal of the survey is to detect local mineralizations with potential commercial value. If a helicopter is being employed, the distance from the magnetic sources may be kept as small as possible by flying at a constant height above the ground surface (Fig. 5.45b). The usual method is to survey a region along parallel flight-lines (Fig. 5.45c), which may be spaced anywhere from 100 m to a few kilometers apart, depending on the flight elevation used, the intensity of coverage, and the quality of detail desired. The orientation of the flight-lines is selected to be more or less normal to the trend of suspected or known subsurface features. Additional tie-lines are flown at right angles to the main pattern. Their separation is about 5–6 times that of the main flight-lines. The repeatability of the measurements at the intersections of the tie-lines and the main flight-lines provides a check on the reliability of the survey. If the differences (called closure errors) are large, an area may need to be resurveyed. Alternatively, the differences may be distributed mathematically among all the observations until the closure errors are minimum.

5.5.4 Reduction of magnetic field measurements

In comparison to the reduction of gravity data, magnetic survey data require very few corrections. One effect that must be compensated is the variation in intensity of the geomagnetic field at the Earth’s surface during the course of a day. As explained in more detail in Section 5.4.3.3 this diurnal variation is due to the part of the Earth’s magnetic field that originates in the ionosphere. At any point on the Earth’s surface the external field varies during the day as the Earth rotates beneath different parts of the ionosphere. The effect is much greater than the precision with which the field can be measured. The diurnal variation may be corrected by installing a constantly recording magnetometer at a fixed base station within the survey area. Alternatively, the records from a geomagnetic observatory may be used, provided it is not too far from the survey area. The time is noted at which each field measurement is made during the actual survey and the appropriate correction is made from the control record.

The variations of magnetic field with altitude, latitude and longitude are dominated by the vertical and horizontal variations of the dipole field. The total intensity $B_1$ of the field is obtained by computing the resultant of the radial component $B_r$ (Eq. (5.38)) and the tangential component $B_t$ (Eq. (5.39)): 

$$B_1 = \sqrt{B_r^2 + B_t^2} = \frac{\mu_0 m}{4\pi} \sqrt{1 + 3 \cos^2 \theta} \frac{1}{r^3}$$ (5.45)

The altitude correction is given by the vertical gradient of the magnetic field, obtained by differentiating the intensity $B_1$ with respect to radius, $r$. This gives

$$\frac{\partial B_1}{\partial r} = -\frac{3}{4\pi} \frac{\mu_0 m}{r^4} \sqrt{1 + 3 \cos^2 \theta} = -\frac{3}{r} B_1$$ (5.46)

The vertical gradient of the field is found by substituting $r = R = 6371$ km and an appropriate value for $B_1$. It clearly depends on the latitude of the measurement site. At the magnetic equator ($B_1 \approx 30,000$ nT) the altitude correction is about 0.015 nT m$^{-1}$; near the magnetic poles ($B_1 \approx 60,000$ nT) it is about 0.030 nT m$^{-1}$. The correction is so small that it is often ignored.

In regional studies the corrections for latitude and longitude are inherent in the reference field that is subtracted. In a survey of a small region, the latitude correction is given by the north–south horizontal gradient of the magnetic field, obtained by differentiating $B_r$ with respect to polar angle, $\theta$. This gives for the northward increase in $B_r$ (i.e., with increasing latitude)

$$\frac{\partial B_r}{\partial \theta} = \frac{\mu_0 m}{4\pi} \frac{1}{r^3} \frac{1}{\sqrt{1 + 3 \cos^2 \theta}} \frac{3 B_r \sin \theta \cos \theta}{r(1 + 3 \cos^2 \theta)}$$ (5.47)

The latitude correction is zero at the magnetic pole ($\theta = 0^\circ$) and magnetic equator ($\theta = 90^\circ$) and reaches a maximum value of about 5 nT per kilometer (0.005 nT m$^{-1}$) at intermediate latitudes. It is insignificant in small-scale surveys.

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**Fig. 5.45** In airborne magnetic surveying the flight-lines may be flown at (a) constant altitude above sea-level, or (b) constant height above ground-level. The flight pattern (c) includes parallel measurement lines and orthogonal cross-tie lines.
In some land-based surveys of highly magnetic terrains (e.g., over lava flows or mineralized intrusions), the disturbing effect of the magnetized topography may be serious enough to require additional topographic corrections.

5.5 Magnetic anomalies

The gravity anomaly of a body is caused by the density contrast ($\Delta \rho$) between the body and its surroundings. The shape of the anomaly is determined by the shape of the body and its depth of burial. Similarly, a magnetic anomaly originates in the magnetization contrast ($\Delta M$) between rocks with different magnetic properties. However, the shape of the anomaly depends not only on the shape and depth of the source object but also on its orientation to the profile and to the inducing magnetic field, which itself varies in intensity and direction with geographical location. In oceanic magnetic surveying the magnetization contrast results from differences in the remanent magnetizations of crustal rocks, for which the Königsberger ratio is much greater than unity (i.e., $Q_n \gg 1$). Commercial geophysical prospecting is carried out largely in continental crustal rocks, for which the Königsberger ratio is much less than unity (i.e., $Q_n \ll 1$) and the magnetization may be assumed to be induced by the present geomagnetic field. The magnetization contrast is then due to susceptibility contrast in the crustal rocks. If $k$ represents the susceptibility of an orebody, $k_0$ the susceptibility of the host rocks and $F$ the strength of the inducing magnetic field, Eq. (5.17) allows us to write the magnetization contrast as

$$\Delta M = (k - k_0)F$$

Some insight into the physical processes that give rise to a magnetic anomaly can be obtained from the case of a vertically sided body that is magnetized by a vertical magnetic field. This is a simplified situation because in practice both the body and the field will be inclined, probably at different angles. However, it allows us to make a few observations that are generally applicable. Two scenarios are of particular interest. The first is when the body has a large vertical extent, such that its bottom surface is at a great depth; the other is when the body has a limited vertical extent. In both cases the vertical field magnetizes the body parallel to its vertical sides, but the resulting anomalies have different shapes. To understand the anomaly shapes we will use the concept of magnetic pole distributions.

5.5.5.1 Magnetic anomaly of a surface distribution of magnetic poles

Although magnetic poles are a fictive concept (see Section 5.2.2.1), they provide a simple and convenient way to understand the origin of magnetic field anomalies. If a slice is made through a uniformly magnetized object, simple logic tells us that there will be as many south poles per unit of area on one side of the slice as north poles on the opposite side; these will cancel each other and the net sum of poles per unit area of the surface of the slice is zero. This is no longer the case if the magnetization changes across the interface. On each unit area of the surface there will be more poles of the stronger magnetization than poles of the weaker one. A quantitative derivation shows that the resultant number of poles per unit area $\sigma$ (called the surface density of poles) is proportional to the magnetization contrast $\Delta M$.

The concept of the solid angle subtended by a surface element (Box 5.4) provides a qualitative understanding of the magnetic anomaly of a surface distribution of magnetic poles. Consider the distribution of poles on the upper surface with area $A$ of a vertical prism with magnetization $M$ induced by a vertical field $B_z$, as illustrated in Fig. 5.46a. At the surface of the Earth, distant $r$ from the distribution of poles, the strength of their anomalous magnetic field is proportional to the total number of poles on the surface, which is the product of $A$ and the surface density $\sigma$ of poles. Equation (5.2) shows that the intensity of the field of a pole decreases as the inverse square of distance $r$. If the direction of $r$ makes an angle $\theta$ with the vertical magnetization $M$, the vertical component of the anomalous field at $P$ is found by multiplying by $\cos \theta$. The vertical magnetic anomaly $\Delta B_z$ of the surface distribution of poles is

$$\Delta B_z \approx \frac{(\sigma A) \cos \theta}{r^2} \propto \frac{(\Delta M) \Omega}{r^2}$$

A more rigorous derivation leads to essentially the same result. At any point on a measurement profile, the magnetic anomaly $\Delta B_z$ of a distribution of poles is proportional to the solid angle $\Omega$ subtended by the distribution at the point. The solid angle changes progressively along a profile (Fig. 5.46b). At the extreme left and right ends, the radius from the observation point is very oblique to the surface distribution of poles and the subtended angles $\Omega_1$ and $\Omega_4$ are very small; the anomaly distant from the body is nearly zero. Over the center of the distribution, the subtended angle reaches its largest value $\Omega_2$ and the anomaly reaches a maximum. The anomaly falls smoothly on each side of its crest corresponding to the values of the subtended angles $\Omega_1$ and $\Omega_4$ at the intermediate positions. A measurement profile across an equal distribution of “north” poles would be exactly inverted. The north poles create a field of repulsion that acts everywhere to oppose the Earth’s magnetic field, so the combined field is less than it would be if the “north” poles were not there. The magnetic anomaly over “north” poles is negative.

5.5.5.2 Magnetic anomaly of a vertical dike

We can now apply these ideas to the magnetic anomaly of a vertical dike. In this and all following examples we will assume a two-dimensional situation, where the horizontal length of the dike (imagined to be into the page) is infinite. This avoids possible complications related to “end effects.” Let us first assume that the dike extends to

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Box 5.4: Solid angles

A solid angle is defined by the ratio between the area of an element of the surface of a sphere and the radius of the sphere. Let the area of a surface element be \( A \) and the radius of the sphere be \( r \), as in Fig. B5.4. The solid angle \( \Omega \) subtended by the area \( A \) at the center of the sphere is defined as

\[
\Omega = \frac{A}{r^2}
\]  

(1)

The angle subtended by any surface can be determined by projecting the surface onto a sphere. The shape of the area is immaterial. If an element of area \( A \) is inclined at angle \( \alpha \) to the radius \( r \) through a point on the surface, its projection normal to the radius (i.e., onto a sphere passing through the point) is \( A \cos \alpha \), and the solid angle it subtends at the center of the sphere is given by

\[
\Omega = \frac{A \cos \alpha}{r^2}
\]  

(2)

A solid angle is measured in units of steradians, which are analogous to radians in planar geometry. The minimum value of a solid angle is zero, when the surface element is infinitesimally small. The maximum value of a solid angle is when the surface completely surrounds the center of the sphere. The surface area of a sphere of radius \( r \) is \( A = 4\pi r^2 \) and the solid angle at its center has the maximum possible value, which is \( 4\pi \).

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Fig. B5.4 Definition of the solid angle \( \Omega \) subtended by an area \( A \) on the surface of a sphere with radius \( r \).

---

very great depths (Fig. 5.47a), so that we can ignore the small effects associated with its remote lower end. The vertical sides of the dike are parallel to the magnetization and no magnetic poles are distributed on these faces. However, the horizontal top face is normal to the magnetization and a distribution of magnetic poles can be imagined on this surface. The direction of magnetization is parallel to the field, so the pole distribution will consist of “south” poles. The magnetized dike behaves like a magnetic monopole. At any point above the dike we measure both the inducing field and the anomalous “stray field” of the dike, which is directed toward its top. The anomalous field has a component parallel to the Earth’s field and so the total magnetic field will be everywhere stronger than if the dike were not present. The magnetic anomaly is everywhere positive, increasing from zero far from the dike to a maximum value directly over it (Fig. 5.46b).

If the vertical extent of the dike is finite, the distribution of north poles on the bottom of the dike may be close enough to the ground surface to produce a measurable stray field. The upper distribution of south poles causes a positive magnetic anomaly, as in the previous example. The lower distribution of north poles causes a negative anomaly (Fig. 5.47b). The north poles are further from the magnetometer than the south poles, so their negative anomaly over the dike is weaker. However, farther along...
the profile the deeper distribution of poles subtends a larger angle than the upper one does. As a result, the strength of the weaker negative anomaly does not fall off as rapidly along the profile as the positive anomaly does. Beyond a certain lateral distance from the dike (to the left of L and to the right of R in Fig. 5.47b) the negative anomaly of the lower pole distribution is stronger than the positive anomaly of the upper one. This causes the magnetic anomaly to have negative side lobes, which asymptotically approach zero with increasing distance from the dike.

The magnetized dike in this example resembles a bar magnet and can be modelled crudely by a dipole. Far from the dike, along a lateral profile, the dipole field lines have a component opposed to the inducing field, which results in the weak negative side lobes of the anomaly. Closer to the dike, the dipole field has a component that reinforces the inducing field, causing a positive central anomaly.

5.5.5.3 Magnetic anomaly of an inclined magnetization

When an infinitely long dike is magnetized obliquely rather than vertically, its anomaly can be modelled either by an inclined dipole or by pole distributions (Fig. 5.48). The magnetization has both horizontal and vertical components, which produce magnetic pole distributions on the vertical sides of the dike as well as on its top and bottom. The symmetry of the anomaly is changed so that the negative lobe of the anomaly is enhanced on the side towards which the horizontal component of magnetization points; the other negative lobe decreases and may disappear.

The shape of a magnetic anomaly also depends on the angle at which the measurement profile crosses the dike, and on the strike and dip of the dike. The geometry, magnetization and orientation of a body may be taken into account in forward-modelling of an anomaly. However, as in other potential field methods, the inverse problem of determining these factors from the measured anomaly is not unique.
can be compensated by the method of interpretation of the oceanic crustal magnetizations responsible. The procedure has proved to be important for detailed inter-margins of the disturbing body. Among other applications, an anomaly (Fig. 5.49) and allows a better location of the center of the prism (after Lindner et al., 1984).

The asymmetry (or skewness) of a magnetic anomaly can be compensated by the method of reduction to the pole. This consists of recalculating the observed anomaly for the case that the magnetization is vertical. The method involves sophisticated data-processing beyond the scope of this text. The observed anomaly map is first converted to a matrix of values at the intersections of a rectangular grid overlying the map. The Fourier transform of the matrix is then computed and convolved with a filter function to correct for the orientations of the body and its magnetization. The reduction to the pole removes the asymmetry of an anomaly (Fig. 5.49) and allows a better location of the margins of the disturbing body. Among other applications, the procedure has proved to be important for detailed interpretation of the oceanic crustal magnetizations responsible for lineated oceanic magnetic anomalies.

5.5.5.4 Magnetic anomalies of simple geometric bodies

The computation of magnetic anomalies is generally more complicated than the computation of gravity anomalies. In practice, iterative numerical procedures are used. However, the Poisson relation (Box 5.5) enables the computation of magnetic anomalies for bodies for which the gravity anomaly is known. This is most easily illustrated for vertically magnetized bodies, such as the following examples.

(1) Sphere. The gravity anomaly \( \Delta g_z \) over a sphere of radius \( R \) with density contrast \( \Delta \rho \) and center at depth \( z \) (representing a diapir or intrusion) is given by Eq. (2.83), repeated here:

\[
\Delta g_z = \frac{4}{3} \pi G \Delta \rho R^3 \left( \frac{z}{R^2 + z^2} \right)^{3/2}
\]

Assuming the same dimensions and a magnetization contrast \( \Delta M_z \), the potential of the magnetic anomaly over a vertically magnetized sphere according to the Poisson relation is

\[
W = \frac{\mu_0}{4\pi} \left( \frac{\Delta M_z}{G \Delta \rho} \right) \Delta g_z = \frac{1}{3} \mu_0 R^3 \Delta M_z \left( \frac{z}{R^2 + z^2} \right)^{3/2} \quad (5.50)
\]

By differentiating with respect to \( x \) or \( z \), we get the horizontal or vertical field anomaly, respectively. The vertical field magnetic anomaly \( \Delta B_z \) of the sphere is

\[
\Delta B_z = -\frac{\partial W}{\partial z} = -\frac{1}{3} \mu_0 R^2 \Delta M_z \frac{\partial}{\partial z} \left( \frac{z}{(R^2 + z^2)^{3/2}} \right)
= -\frac{1}{3} \mu_0 R^2 \Delta M_z \frac{(z^2 + x^2)^{3/2} - z(3/2)(2z)(z^2 + x^2)^{1/2}}{(z^2 + x^2)^{3/2}} \quad (5.51)
\]

\[
\Delta B_z = \frac{1}{3} \mu_0 R^2 \Delta M_z \frac{2z^2 - x^2}{(z^2 + x^2)^{3/2}} \quad (5.52)
\]

(2) Horizontal cylinder. The gravity anomaly \( \Delta g_z \) over a cylinder of radius \( R \) with horizontal axis centered at depth \( z \) and with density contrast \( \Delta \rho \) (representing an anticline or syncline) is given by Eq. (2.93). If the structure is vertically magnetized with magnetization contrast \( \Delta M_z \), Poisson’s relation gives for the magnetic potential

\[
W = \frac{\mu_0}{4\pi} \left( \frac{\Delta M_z}{G \Delta \rho} \right) \Delta g_z = \frac{1}{3} \mu_0 R^2 \Delta M_z \frac{z}{z^2 + R^2} \quad (5.53)
\]

The vertical magnetic field anomaly \( \Delta B_z \) over the horizontal cylinder is

\[
\Delta B_z = \frac{1}{3} \mu_0 R^2 \Delta M_z \frac{2z^2 - x^2}{(z^2 + x^2)^{3/2}} \quad (5.54)
\]

(3) Horizontal crustal block. The gravity anomaly for a thin horizontal sheet of thickness \( t \) at depth \( \delta \) between horizontal positions \( x_1 \) and \( x_2 \) (Fig. 2.54 b), extending to infinity normal to the plane of observation, is given by Eq. (2.96). Let the width of the block be \( 2m \), and let the horizontal position be measured from the midpoint of the block, so that \( x_1 = x - m \) and \( x_2 = x + m \). Applying Poisson’s relation, we get the magnetic potential for a semi-infinite horizontal thin sheet of vertically magnetized dipoles, of thickness \( t \) at depth \( z \)

\[
W = \frac{\mu_0}{4\pi} \left( \frac{\Delta M_z}{G \Delta \rho} \right) \Delta g_z = \frac{\mu_0 \Delta M_z}{2\pi} t \times \left[ \tan^{-1} \left( \frac{x + m}{z} \right) - \tan^{-1} \left( \frac{x - m}{z} \right) \right] \quad (5.55)
\]
Box 5.5: Poisson’s relation

Poisson (1781–1840) observed a relationship between the gravitational and magnetic potentials of a body, which allows a simple method of computing magnetic field anomalies if the gravity anomaly of the body is known. Consider an arbitrary volume \( V \) with homogeneous density and vertical magnetization (Fig. B5.5). If the density of the body is \( \Delta \rho \) a small element with volume \( dV \) has mass \( (\Delta \rho V) \). The gravitational potential \( U \) at a point on the surface at a distance \( r \) from the element is

\[
U = -\frac{G \Delta \rho dV}{r} \quad (1)
\]

The vertical gravity anomaly \( \Delta g_z \) of the volume element is found by differentiating \( U \) with respect to \( z \)

\[
g_z = -\frac{\partial}{\partial z} \left( -\frac{G \Delta \rho dV}{r} \right) = G \Delta \rho \frac{dV}{r} \frac{\partial}{\partial z} \left( \frac{1}{r} \right) \quad (2)
\]

If the body is vertically magnetized with uniform magnetization \( \Delta M_z \), the magnetic moment of the volume element is \( (\Delta M_z dV) \). The magnetic moment is directed downward as in Fig. B5.5. The radius vector from the element to a point on the surface makes an angle \((\pi - \theta)\) with the orientation of the magnetization. The magnetic potential \( W \) at the point \((r, \theta)\) is

\[
W = \frac{\mu_0}{4\pi} \frac{\Delta M_z dV \cos(\pi - \theta)}{r^2}
= -\frac{\mu_0}{4\pi} \frac{\Delta M_z dV \cos \theta}{r^2} = -\frac{\mu_0}{4\pi} \frac{\Delta M_z dV}{r^2} \left( \frac{z}{r} \right) \quad (3)
\]

Note the following relationship

\[
\frac{\partial}{\partial z} \left( \frac{1}{r} \right) = -\frac{1}{r^2} \frac{\partial r}{\partial z} = -\frac{1}{r^2} \frac{\partial}{\partial z} \sqrt{x^2 + z^2} = -\frac{1}{r^2} \frac{z}{r} \quad (4)
\]

Comparing Eq. (2) and Eq. (5) and eliminating the volume \( \Delta V \) we get Poisson’s relation:

\[
W = \frac{\mu_0}{4\pi} \frac{(\Delta M_z)}{(\partial \Delta \rho) / \partial z} g_z \quad (5)
\]

Substituting this result in (3) gives

\[
W = \frac{\mu_0}{4\pi} \frac{\Delta M_z dV \frac{\partial}{\partial z} \left( \frac{1}{r} \right)}{(\partial \Delta \rho) / \partial z}
\]

This derivation for a small element is also valid for an extended body as long as the density and magnetization are both uniform.

Assume that the horizontal crustal block is made up of layers of thickness \( t = dz \). If the top of the block is at depth \( z_1 \) and its base at depth \( z_2 \), the magnetic potential of the block is found by integrating Eq. (5.55) between \( z_2 \) and \( z_1 \),

\[
W = \frac{\mu_0}{2\pi} \frac{\Delta M_z}{2} \int_{z_1}^{z_2} \left[ \tan^{-1} \left( \frac{x + m}{z} \right) - \tan^{-1} \left( \frac{x - m}{z} \right) \right] dz \quad (5.56)
\]

Differentiating with respect to \( z \) gives the vertical magnetic field anomaly over the block:

\[
\Delta B_z = -\frac{\partial}{\partial z} W
= -\frac{\mu_0}{2\pi} \frac{\Delta M_z}{2} \int_{z_1}^{z_2} \frac{\partial}{\partial z} \left[ \tan^{-1} \left( \frac{x + m}{z} \right) - \tan^{-1} \left( \frac{x - m}{z} \right) \right] dz
\]

where the angles \( \alpha_1, \alpha_2, \alpha_3 \) and \( \alpha_4 \) are defined in Fig. 5.50a. Note that the angles \( (\alpha_1 - \alpha_2) \) and \( (\alpha_3 - \alpha_4) \) are the planar angles subtended at the point of measurement by the top and bottom edges of the vertically magnetized crustal block respectively. This is similar to the dependence of magnetic anomalies of
5.5.5 Effect of block width on anomaly shape

The effect of the width of a crustal block on anomaly shape is illustrated by use of Eq. (5.59) to model the vertical field magnetic anomaly of a vertically magnetized block with its top at depth 2.5 km and base at depth 3 km. The block is effectively a thin magnetized layer, similar to the source of oceanic magnetic anomalies. Three cases are considered here: a narrow block of width \( w = 2m = 5 \) km for which \( mlz_1 = 1 \), a block of width 10 km \( (mlz_1 = 2) \), and a wide block of width 40 km \( (mlz_1 = 8) \).

The narrowest block gives a sharp, positive central anomaly with negative side lobes (Fig. 5.50b), as explained in Section 5.5.5.2. As the block widens with respect to its depth, the top of the central anomaly flattens (Fig. 5.50c), its amplitude over the middle of the block decreases, and the negative side lobes grow. When the block is much wider than the depth to its top (Fig. 5.50d), a dip develops over the center of the block. The positive anomalies are steep sided and are maximum just within the edges of the block, while the null values occur close to the edges of the block. The negative side anomalies are almost as large as the positive anomalies.

The pronounced central dip in the anomaly is due to the limited vertical thickness of the layer. If the layer is very wide relative to its thickness, the central anomaly may diminish almost to zero. This is because the angle \((\alpha_1 - \alpha_2)\) subtended by the magnetized base of the layer is almost (but not quite) as large as the angle \((\alpha_3 - \alpha_4)\) subtended by the top of the layer. For a very large width-to-thickness ratio, the central anomaly is zero, the edge anomalies separate and become equivalent to separate anomalies over the edges of the block.

Examination of Fig. 5.50a shows that the subtended angles \((\alpha_1 - \alpha_2)\) and \((\alpha_3 - \alpha_4)\), and thus the anomaly shape, depend also on the height of the measurement profile above the surface of the block. A low-altitude profile over the block will show a large central dip, while a high-altitude profile over the same block will show a smaller dip or none at all.

In contrast to the example of a thin layer described above, if the crustal block is very thick, extending to great depth, the angle \((\alpha_3 - \alpha_4)\) is zero and the effects of the magnetization discontinuity (or pole distribution) on its base are absent. The shape of the anomaly is then determined by \((\alpha_1 - \alpha_2)\) and is flat topped over a wide block.

5.5.6 Oceanic magnetic anomalies

In the late 1950s marine geophysicists conducting magnetic surveys of the Pacific ocean basin off the west coast of North America discovered that large areas of oceanic crust are characterized by long stripes of alternating positive and negative magnetic anomalies. The striped pattern is best known from studies carried out across oceanic ridge systems (see Fig. 1.13). The striped anomalies are hundreds of kilometers in length parallel to the ridge axis, 10–50 km in width, and their amplitudes amount to several hundreds of nanotesla. On magnetic profiles perpendicular to a ridge axis the anomaly pattern is found to exhibit a remarkable symmetry about the axis of the ridge. The origin of the symmetric lineated anomaly pattern cannot be explained by conventional methods of interpretation based on susceptibility contrast.

Seismic studies indicate a layered structure for the oceanic crust. The floor of the ocean lies at water depths
of 2–5 km, and is underlain by a layer of sediment of variable thickness, called seismic Layer 1. Under the sediments lie a complex of basaltic extrusions and shallow intrusions, about 0.5 km thick, forming seismic Layer 2A, under which are found the deeper layers of the oceanic crust consisting of a complex of sheeted dikes (Layer 2B) and gabbro (Layer 3). The magnetic properties of these rocks were first obtained by studying samples dredged from exposed crests and ridges of submarine topography. The rocks of Layers 2B and 3 are much less magnetic than those of Layer 2A. Samples of pillow basalt dredged near to oceanic ridges have been found to have moderate susceptibilities for igneous rocks, but their remanent magnetizations are intense. Their Königsberger ratios are commonly in the range 5–50 and frequently exceed 100. Recognition of these properties provided the key to understanding the origin of the lineated magnetic anomalies. In 1963 the English geophysicists F. J. Vine and D. H. Matthews proposed that the remanent magnetizations (and not the susceptibility contrast) of oceanic basaltic Layer 2 were responsible for the striking lineated anomaly pattern. This hypothesis soon became integrated into a working model for understanding the mechanism of sea-floor spreading (see Section 1.2.5 and Fig. 1.14).

The oceanic crust formed at a spreading ridge acquires a thermoremanent magnetization (TRM) in the geomagnetic field. The basalts in Layer 2A are sufficiently strongly magnetized to account for most of the anomaly measured at the ocean surface. For a lengthy period of time (measuring several tens of thousands to millions of years) the polarity of the field remains constant; crust formed during this time carries the same polarity as the field. After a polarity reversal, freshly formed basalts acquire a TRM parallel to the new field direction, i.e., opposite to the previous TRM. Adjacent oceanic crustal blocks of different widths, determined by the variable time between reversals, carry antiparallel remanent magnetizations.

The oceanic crust is magnetized in long blocks parallel to the spreading axis, so the anomaly calculated for a profile perpendicular to the axis is two dimensional, as in the previous examples. Consider the case where the anomalies on a profile have been reduced to the pole, so that their magnetizations can be taken to be vertical. We can apply the concept of magnetic pole distributions to each block individually to determine the shape of its magnetic anomaly (Fig. 5.51a). If the blocks are contiguous, as is the case when they form by a continuous process such as sea-floor spreading, their individual anomalies will overlap (Fig. 5.51b). The spreading process is symmetric with respect to the ridge axis, so a mirror image of the sequence of polarized blocks is formed on the other side of the axis (Fig. 5.51c). If the two sets of crustal blocks are brought together at the spreading axis, a magnetic anomaly sequence ensues that exhibits a symmetric pattern with respect to the ridge axis (Fig. 5.51d).

This description of the origin of oceanic magnetic anomalies is over-simplified, because the crustal magnetization is more complicated than assumed in the block model. For example, the direction of the remanent magnetization, acquired at the time of formation of the ocean crust, is generally not the same as the direction of the magnetization induced by the present-day field. However, the induced magnetization has uniformly the same direction in the magnetized layer, which thus behaves like a uniformly magnetized thin horizontal sheet and does not contribute to the magnetic anomaly. Moreover, oceanic rocks have high Königsberger ratios, and so the induced magnetization component is usually negligible in comparison to the remanent magnetization. An exception is when a magnetic
survey is made close to the magnetized basalt layer, in which case a topographic correction may be needed.

Unless the strike of a ridge is north–south, the magnetization inclination must be taken into account. Skewness is corrected by reducing the magnetic anomaly profile to the pole (Section 5.5.5.3). A possible complication may arise if the oceanic magnetic anomalies have two sources. The strongest anomaly source is doubtless basaltic Layer 2B, but, at least in some cases, an appreciable part of the anomaly may arise in the deeper gabbroic Layer 3. The two contributions are slightly out of phase spatially, because of the curved depth profiles of cooling isotherms in the oceanic crust. This causes a magnetized block in the deeper gabbroic layer to lie slightly further from the ridge than the corresponding block with the same polarity in the basaltic layer above it. The net effect is an asymmetry of inclined magnetization directions on opposite sides of a ridge, so that the magnetic anomalies over blocks of the same age have different skewnesses.

5.6 PALEOMAGNETISM

5.6.1 Introduction

A mountain walker using a compass to find his way in the Swiss Alps above the high mountain valley of the Engadine would notice that in certain regions (for example, south of the Septimer Pass) the compass-needle shows very large deviations from the north direction. The deflection is due to the local presence of strongly magnetized serpentinites and ultramafic rocks. Early compasses were more primitive than modern versions, but the falsification of a compass direction near strongly magnetic outcrops was known by at least the early nineteenth century. In 1797 Alexander von Humboldt proposed that the rocks in these unusual outcrops had been magnetized by lightning strikes. The first systematic observations of rock magnetic properties are usually attributed to A. Delesse (1849) and M. Melloni (1853), who concluded that volcanic rocks acquired a remanent magnetization during cooling. After a more extensive series of studies in 1894 and 1895 of the origin of magnetism in lavas, G. Folgerhaiter reached the same conclusion and suggested that the direction of remanent magnetization was that of the geomagnetic field during cooling. By 1899 he had extended his work to the record of the secular variation of inclination in ancient potteries. Folgerhaiter noted that some rocks have a remanent magnetization opposite to the direction of the present-day field. Reversals of polarity of the geomagnetic field were established decisively early in the twentieth century.

In 1922 Alfred Wegener proposed his concept of continental drift, based on years of study of paleoclimatic indicators such as the geographic distribution of coal deposits. At the time, there was no way of explaining the mechanism by which the continents drifted. Only motions of the crust were considered, and the idea of rigid continents ploughing through rigid oceanic crust was unacceptable to geophysicists. There was as yet no way to reconstruct the positions of the continents in earlier eras or to trace their relative motions. Subsequently, paleomagnetism was to make important contributions to understanding continental drift by providing the means to trace past continental motions quantitatively.

A major impetus to these studies was the invention of a very sensitive astatic magnetometer. The apparatus consists of two identical small magnets mounted horizontally at opposite ends of a short rigid vertical bar so that the magnets are oriented exactly antiparallel to each other. The assembly is suspended on an elastic fiber. In this configuration the Earth’s magnetic field has equal and opposite effects on each magnet. If a magnetized rock is brought close to one magnet, the magnetic field of the rock produces a stronger twisting effect on the closer magnet than on the distant one and the assembly rotates to a new position of equilibrium. The rotation is detected by a light beam reflected off a small mirror mounted on the rigid bar. The device was introduced in 1952 by P. M. S. Blackett to test a theory that related the geomagnetic field to the Earth’s rotation. The experiment did not support the postulated effect. However, the astatic magnetometer became the basic tool of paleomagnetism and fostered its development as a scientific discipline. Hitherto it had only been possible to measure magnetizations of strongly magnetic rocks. The astatic magnetometer enabled the accurate measurement of weak remanent magnetizations in rocks that previously had been unmeasurable.

In the 1950s, several small research groups were engaged in determining and interpreting the directions of magnetization of rocks of different ages in Europe, Africa, North and South America and Australia. In 1956 S. K. Runcorn put forward the first clear geophysical evidence in support of continental drift. Runcorn compared the directions of magnetization of Permian and Triassic rocks from Great Britain and North America. He found that the paleomagnetic results from the different continents could be brought into harmony for the time before 200 Ma ago by closing the Atlantic ocean. The evaluation of the scientific data was statistical and at first was regarded as controversial. However, Mesozoic paleomagnetic data were soon obtained from the southern hemisphere that also argued strongly in favor of the continental drift hypothesis. In 1957 E. Irving showed that paleomagnetic data confirmed better with geological reconstructions of earlier positions of the continents than with their present-day distribution. Subsequently, numerous studies have documented the importance of paleomagnetism as a chronicle of past motions of global plates and as a record of the polarity history of the Earth’s magnetic field.

5.6.2 The time-averaged geomagnetic field

A fundamental assumption of paleomagnetism is that the time-averaged geomagnetic field corresponds to that of