1.4.3 Simulation

An alternative, but also complementary, approach to determining the nature of landscape change through time is through the use of **simulation models** of which there are essentially three types: hardware models, analogue models and mathematical models.

Hardware models, of which laboratory flumes are perhaps the best known example, are frequently used to study the operation of specific geomorphic processes, such as the transport of sediment by wind or water; but they can also be used to explore the sequence of landform adjustments that arise as a result of changes in external variables influencing these processes. An excellent example of this kind of application of hardware models is the assessment of the effects of a lowering of base level on an artificial channel created in a flume.

The major problem with hardware models is that of scale. Such models are almost invariably constructed at a size much smaller than reality and usually the aim is to accelerate specific geomorphic effects under controlled conditions so that substantial changes in form can be observed over a relatively short period of time. Unfortunately, though, changes in scale affect the relationships between the various properties of hardware models and the real world in different ways. Scale ratios for length are different for those involving velocities and acceleration, while properties such as mass and inertia which are critical in processes such as sediment transport are also not simply a linear function of length. A partial solution to this problem is to maintain the correct scaling for a particular variable of interest by distorting the scale relationships of other variables which are not of specific interest.

A second type of simulation is carried out using **analogue models**. These involve the replacement of materials which occur in the real world with other materials which enable the effects of processes to be more readily observed. Examples include the use of clay to synthesize the effects of glacier flow, or the development of fault patterns. Such models have now been largely replaced by mathematical models which represent geomorphic phenomena and the relationships between them in terms of mathematical expressions.

These mathematical models constitute the third kind of simulation model. **Deterministic mathematical models** are based on exact relationships between independent (causal) variables and dependent (response) variables, such as those that predict the way in which a slope profile is transformed over time from some initial form. By contrast **stochastic mathematical models** incorporate a random component which allows different possible outcomes to arise from a given set of initial conditions. One of the most widely applied types of stochastic model is that used to simulate the development of stream networks where the growth of individual channel links in nature is affected by numerous minor factors, the net effects of which can be adequately treated as random (Fig. 1.12).

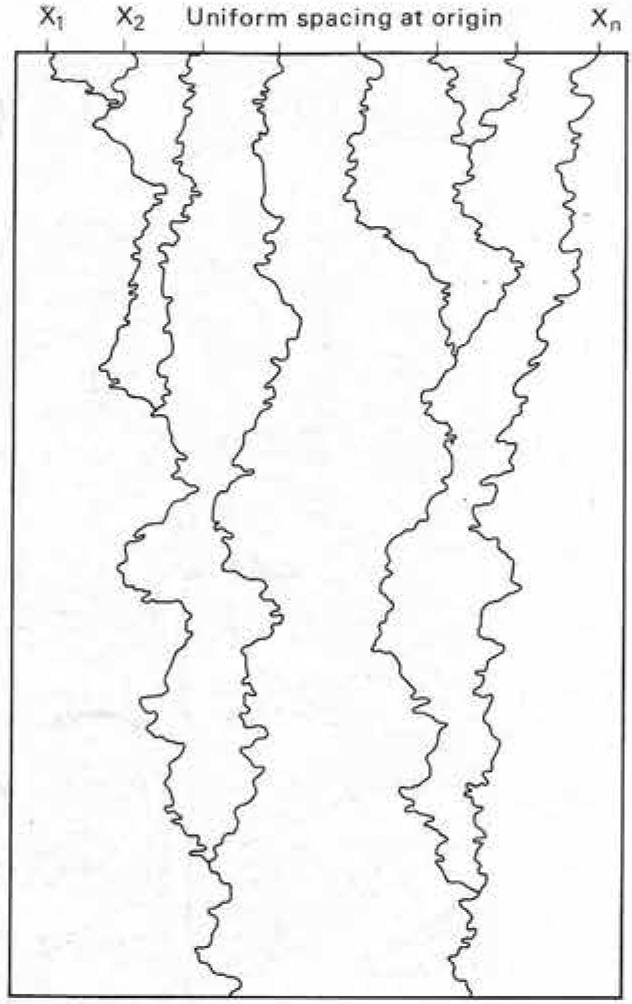

Fig. 1.12 The development of rills on a hillslope simulated using a stochastic (random) model. The model specified that in each time increment the 'random walk' could proceed forward at any angle, but could not go backward. Note the equal spacing between the points of origin (X_1–X_n) for each rill. (After L. B. Leopold, M. G. Wolman and J. P. Miller (1964) Fluvial Processes in Geomorphology. *W. H. Freeman, San Francisco, Fig. 10–4, p. 416.)*

Given the complexity of most geomorphic processes, the construction of mathematical models necessitates a good deal of abstraction and selection, and the geomorphic phenomena considered are almost invariably represented in a highly idealized fashion. Landform properties and processes must be strictly defined if they are to be represented mathematically. Mathematical models simplify the complexity of form existing in the real world and in so doing enable us to predict how landforms will change through time. How successful such predictions are depends on how good a representation of the real world our model is. The most accurate simulations of landform change are generally those which treat relatively simple geomorphic systems, and which are based on a detailed knowledge of the geomorphic processes involved.

There are two major problems in using mathematical models to simulate landform change over time. One is **equifinality** – different models may generate similar results and unless the geomorphic processes and landforms involved are very well understood it may not be possible to decide which version most adequately describes the processes that are operating. The second problem is that we need to have a detailed knowledge of the actual changes in form that occur in the landscape in order that mathematical models predicting particular changes may be tested. This is an often neglected but crucially important issue.

1.5 Endogenic and exogenic factors

1.5.1 Sources of energy

The processes that shape the world's landscapes are powered by two major sources of energy (Fig. 1.13). The energy for endogenic mechanisms comes primarily from geothermal heat, although small contributions are also made by tidal energy generated by the gravitational attraction of the Sun and the Moon and by rotational energy derived from the momentum of the Earth's rotation. The ultimate sources of energy for exogenic processes are the potential energy arising from the height of material above base level, and that proportion of solar radiation received by the Earth.

1.5.1.1 Internal energy

Evidence of the Earth's internal energy is provided by the **geothermal heat flow** which can be measured, and in some cases observed, at the surface. Volcanoes, of course, represent local areas of higher than average heat flow, but the Earth's internal heat is also the ultimate source of energy for virtually all tectonic processes and the associated horizontal and vertical movements of the crust. Although the energy involved in the various endogenic mechanisms is not known, the present rate of heat flow to the surface is fairly

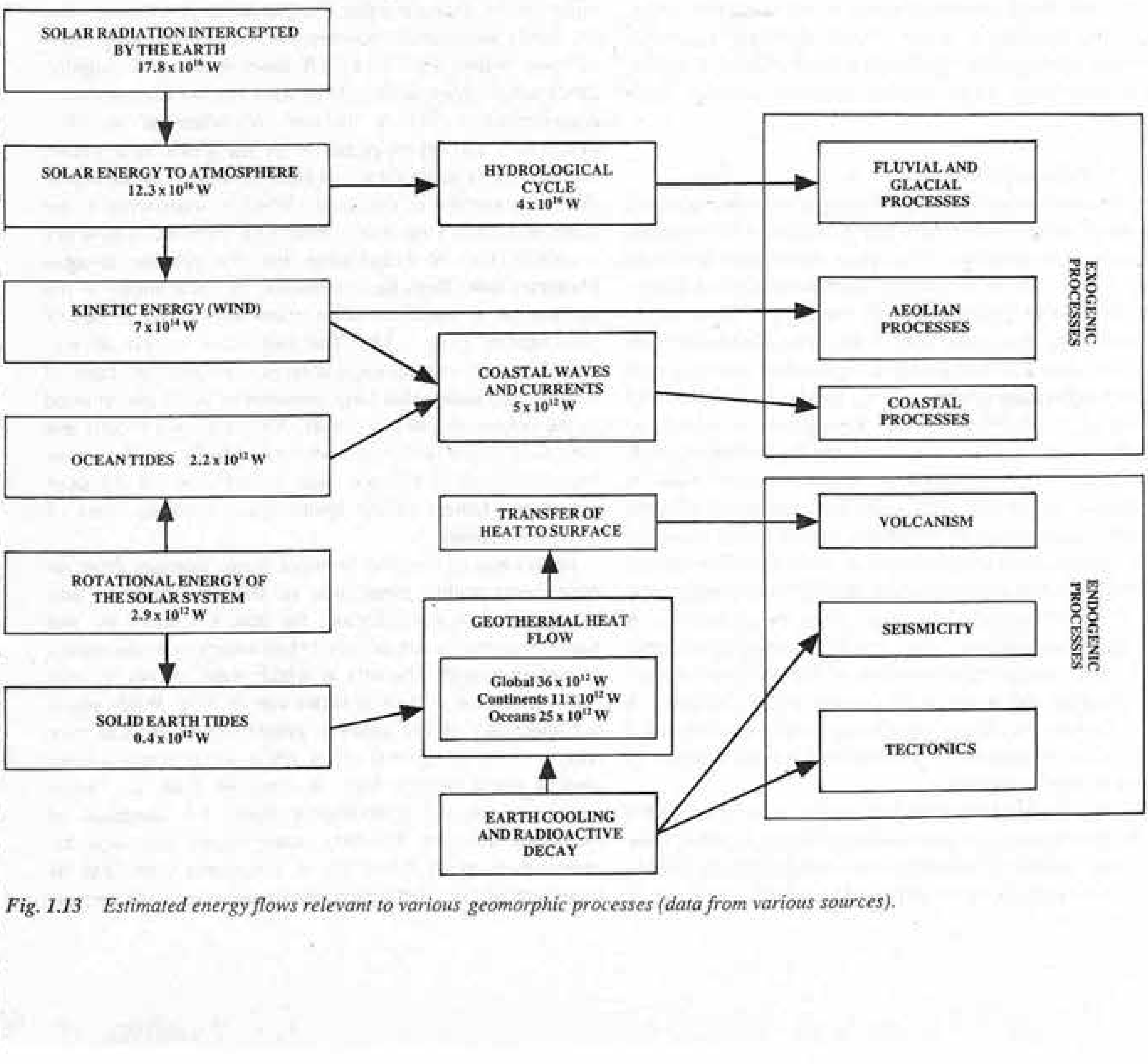

Fig. 1.13 Estimated energy flows relevant to various geomorphic processes (data from various sources).

well established. It can be measured from the increase in temperature with depth (averaging around 20–30 °C km^{-1}), although allowances must be made for the thermal conductivity of the rock. Heat flow rates within the continents average around 56 mW m^{-2} and range up to 200 mW m^{-2} in areas of active volcanic activity. In the ocean basins the average is about 78 mW m^{-2}, but rates as high as 250 mW m^{-2} have been recorded. These variations in heat flow provide important clues as to the nature of the Earth's interior and the operation of endogenic processes.

The major source of geothermal heat is the radioactive decay of the long-lived isotopes of uranium, thorium and potassium. About 83 per cent of the heat flow to the Earth's surface is attributable to this process, the remainder being provided by the continued cooling of the Earth which has been proceeding since its formation some 4.6 Ga ago. Because the half-lives of the major heat-producing isotopes are in the range 10^9–10^{10} a the supply of heat from radioactive decay has been more or less constant over the past several hundred million years. Eventually, though, this heat supply will gradually diminish and the endogenic processes arising from it will become less active and ultimately cease. Here, then, we have a system which, although in a steady state for a geologically significant period of time, is in fact, in the very long term, moving towards a decay equilibrium.

1.5.1.2 Solar radiation

Solar radiation provides an enormous source of energy, but only a very small proportion of this is utilized in the operation of geomorphic processes. The upper atmosphere intercepts about 17.8×10^{16} W of largely short-wave radiant energy from the Sun of which about 30 per cent is immediately reflected back into space (Fig. 1.13). The remainder heats the atmosphere and the surface and generates a mean global surface temperature of about 15 °C. Of the 12.3×10^{16} W of solar energy received by the atmosphere a significant amount (about 33 per cent) drives the **hydrological cycle** (see Section 1.5.2), the continuous movement of water in its gaseous, liquid and solid states between and within the atmosphere, oceans and landsurface. Latent heat is absorbed by the vaporization (evaporation) of water from the surface of the oceans and continents and through transpiration (the loss of water vapour from plant cells) by vegetation. A small proportion (about 1 per cent) is converted into kinetic energy and powers the circulation of the air in the atmosphere (winds) and water in the oceans (ocean currents). A further minute, but highly significant, proportion (about 0.1 per cent) is consumed in photosynthesis (the fixation of radiant energy by plants).

The receipt of solar radiation varies over the Earth's surface both temporally and spatially. Taking a global view there is an excess of incoming over outgoing radiation between about latitude 40° N and S and a deficit polewards of these latitudes. The resulting latitudinal temperature gradient gives rise to the general circulation of the atmosphere which contributes, along with ocean currents, to the redistribution of heat from the equator towards the poles and is the major factor in the climatic zonation of the Earth. At the regional scale marked temperature differences are generated by the contrasting thermal properties of the continents and oceans, with the higher heat capacity of the oceans helping to moderate in coastal areas the extremes of temperature characteristic of continental interiors. At a local scale altitude becomes a primary factor affecting the heat budget. With increasing elevation there is a progressively greater heat loss through long-wave radiation from the Earth's surface. This leads to an overall decrease in mean temperature but an increase in diurnal range.

1.5.2 The hydrological cycle

The hydrological cycle can be conceived as a system of storages between which water is transferred. The oceans represent by a considerable margin the largest storage, but ice sheets and glaciers account for a significant proportion of fresh water (Fig. 1.14). Of more geomorphic significance are the magnitudes of the transfers between storages. Approximately 517×10^3 km^3 of water is annually evaporated and reprecipitated over the globe as a whole; this is roughly equivalent to a layer of water 1 m thick over the entire surface of the Earth. Water is transferred to the atmosphere as a vapour by evaporation from the oceans and a combination of evaporation and transpiration (**evapotranspiration**) from the continents, and is returned to the surface in a liquid or solid state as various forms of precipitation (Fig. 1.14). The significant excess of precipitation over evapotranspiration on the continents (around 40 per cent) means that large amounts of water are returned to the oceans as surface runoff. Although only 0.0001 per cent of all water (and 0.004 per cent of fresh water) is to be found in rivers at any one time, runoff is by far the most important element of the hydrological cycle in terms of landform genesis.

In addition to transfers between major storages there are movements within them (that is, transfers between substorages). Most precipitation, for instance enters the soil water or ground water storage before returning to the oceans as surface runoff. The rate at which water moves through the hydrological cycle also varies significantly. Water which becomes part of the shallow ground water storage may remain there for several years while water entering deep ground water storage may be removed from the active circulation of the hydrological cycle for hundreds of thousands of years. Similarly, water frozen into large ice sheets may travel hundreds of kilometres over tens of thousands of years before eventually returning to the ocean.

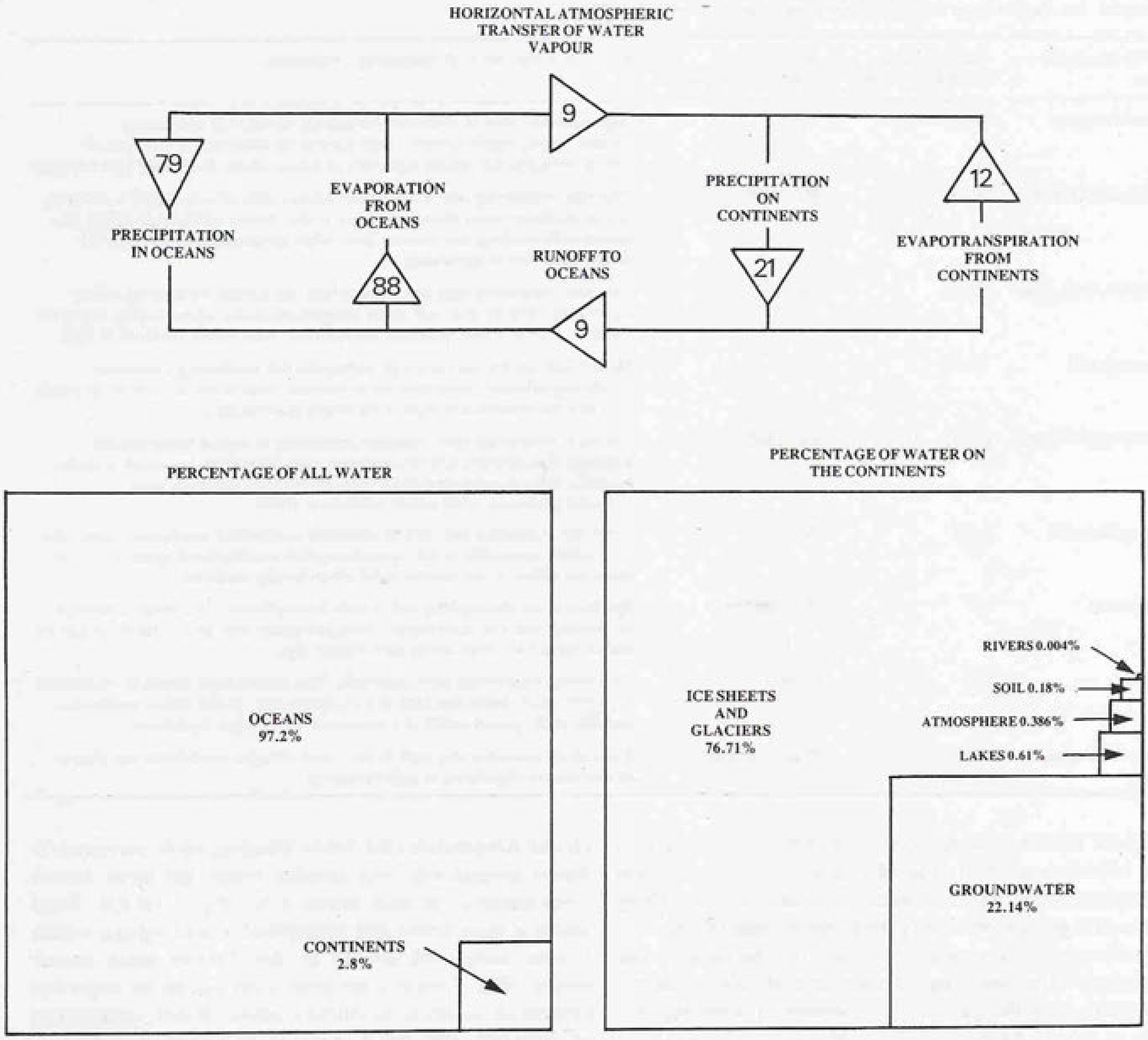

Fig. 1.14 The hydrological cycle. The figures representing transfers are percentages of the mean annual global precipitation of approximately 1000 mm. The proportion of water in the various storages is also illustrated. (Data from various sources.)

1.5.3 Climatic controls

The wide range of climatic conditions over the Earth's surface, together with the role of climatic factors in influencing the nature and rate of operation of geomorphic processes, has led some geomorphologists to suggest that different climates are associated with characteristic landform assemblages. More specifically, it has been argued that particular climatic conditions can have an effect on geomorphic processes sufficient to outweigh the influence on landform development of differences in tectonic setting, rock type and relief, and over a period of time can generate a distinctive association of landforms of regional extent. This is the essential assumption of climatic geomorphology which seeks to establish the nature of landforms associated with distinctive climatic regimes and to identify the specific combinations of geomorphic processes which give rise to them. Areas characterized by landforms associated with a particular climatic environment are called **morphoclimatic zones** (or regions). The term morphogenetic region is also used, but this is arguably less appropriate as it implies that climatic factors alone control landform genesis.

The validity of the assumptions behind climatic geomorphology have been severely challenged by many geomorphologists who see little evidence of a close relationship between climate and landform morphology except under

Table 1.4 The Earth's major morphoclimatic zones (based on various sources)

MORPHOCLIMATIC ZONE	MEAN ANNUAL TEMPERATURE (°C)	MEAN ANNUAL PRECIPITATION (mm)	RELATIVE IMPORTANCE OF GEOMORPHIC PROCESSESS
Humid tropical	20–30	>1500	High potential rates of chemical weathering; mechanical weathering limited; active, highly episodic mass movement; moderate to low rates of stream corrasion but locally high rates of dissolved and suspended load transport
Tropical wet–dry	20–30	600–1500	Chemical weathering active during wet season; rates of mechanical weathering low to moderate; mass movement fairly active; fluvial action high during wet season with overland and channel flow; wind action generally minimal but locally moderate in dry season.
Tropical semi–arid	10–30	300–600	Chemical weathering rates moderate to low; mechanical weathering locally active especially on drier and cooler margins; mass movement locally active but sporadic; fluvial action rates high but episodic; wind action moderate to high
Tropical arid	10–30	0–300	Mechanical weathering rates high (especially salt weathering); chemical weathering minimal; mass movement minimal; rates of fluvial activity generally very low but sporadically high; wind action at a maximum
Humid mid–latitude	0–20	400–1800	Chemical weathering rates moderate, increasing to high at lower latitudes; mechanical weathering activity moderate with frost action important at higher latitudes; mass movement activity moderate to high; moderate rates of fluvial processes: wind action confined to coasts
Dry continental	0–10	100–400	Chemical weathering rates low to moderate; mechanical weathering, especially frost action, seasonally active; mass movement moderate and episodic; fluvial processes active in wet season; wind action locally moderate
Periglacial	<0	100–1000	Mechanical weathering very active with frost action at a maximum; chemical weathering rates low to moderate; mass movement very active; fluvial processes seasonally active; wind action rates locally high
Glacial	<0	0–1000	Mechanical weathering rates (especially frost action) high; chemical weathering rates low; mass movement rates low except locally; fluvial action confined to seasonal melt; glacial action at a maximum; wind action significant
Azonal mountain zone	Highly variable	Highly variable	Rates of all processes vary significantly with altitude; mechanical and glacial action become significant at high elevations

the most extreme climatic contrasts. Suffice it to say here that with the exception of landforms which have very short relaxation times, most elements of the landscape are likely to be, to a greater or lesser extent, out of equilibrium with prevailing climatic conditions because of the rapidity and magnitude of global climatic changes that have occurred, especially over the past 2–3 Ma. Indeed in some regions, such as central Australia and central southern Africa, there is generally a very low rate of geomorphic activity and the landscape is dominated by **relict landforms** developed under climates quite different from those prevailing now.

Notwithstanding uncertainties about the precise relationship between climate and landform genesis, it is clear that there are major contrasts in the kind of geomorphic processes active under certain climatic regimes and that all climatically related processes vary in the intensity with which they operate from one climatic region to another (Table 1.4). Indeed a number of attempts have been made to produce global maps of morphoclimatic zones, and Fig. 1.15 shows a modified version of the widely cited map by the French geomorphologists J. Tricart and A. Cailleux.

The boundaries between morphoclimatic zones are somewhat arbitrary, but some climatic parameters are rather specifically related to the operation of particular geomorphic processes. Frost action can obviously occur only where ground temperatures fall below freezing, while permanently frozen ground will only develop where the mean annual temperature is at least below 0 °C (Fig. 1.16(A)). Wind action is most active and widespread in arid regions within a zone fairly well defined by the 200 mm mean annual isohyet (Fig. 1.16(B)), although wind can be an important process on coasts in all climatic zones. Global variations in precipitation undoubtedly have some influence on rates of weathering and fluvial activity, but the frequency of high-intensity rainfall events is probably far more significant in affecting rates of erosion than overall rainfall amounts (Fig. 1.16(B)).

1.5.4 Human agency

Landforms, and the processes that create them, impinge on human activity in many ways. High magnitude geomorphic events, such as large landslides or major floods, become natural hazards if they affect people. Whether this happens depends on the distribution of people and geomorphic events in time and space. By examining the landforms themselves it is often possible to tell which areas are likely to be at risk from high-magnitude–low-frequency geomorphic events, while a knowledge of the conditions that occur immediately before a major event, such as high-intensity

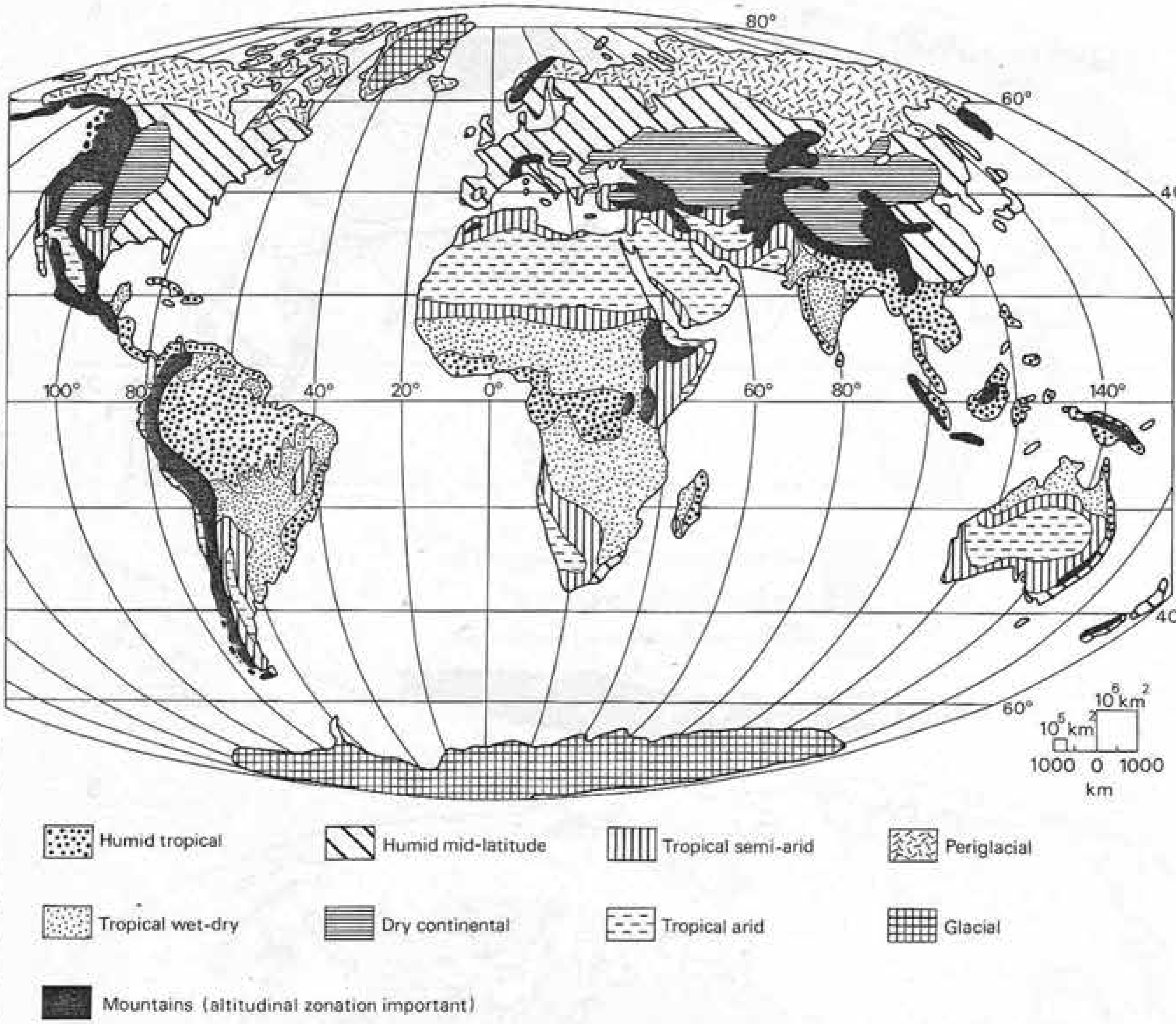

Fig. 1.15 Global distribution of morphoclimatic zones (see Table 1.4 for details of climatic parameters and geomorphic processes.) (Modified from J. Tricart and A. Cailleux (1972) Introduction to Climatic Geomorphology. *Longman, London, Map III.)*

rainfall preceding a flood, can provide the opportunity to issue warnings for evacuation. A knowledge of geomorphology can also contribute to environmental management in regions where special account has to be taken of particular geomorphic phenomena; examples include the impact of pipeline and road construction on permanently frozen ground in high-latitude environments, and the encroachment of sand dunes on human settlements in arid regions. In some cases an understanding of the factors controlling geomorphic processes can be directly relevant to engineering problems, an example being the application of geomorphic research on weathering to the problem of the deterioration of buildings in arid regions due to the presence of salts. Finally, a knowledge of landform genesis can assist in prospecting for certain kinds of mineral resources.

This book does not specifically consider these topics which together constitute the subdiscipline of applied geomorphology. None the less, it is vital to take account of the way human activity itself has affected the natural landscape. Human agency as a factor in landform genesis began with the first fires lit by our ancestors which burnt the vegetation and temporarily left the soil more vulnerable to erosion. The impact of human activity on the landscape has been increasing at an accelerating rate, especially since the Industrial Revolution and the introduction of modern farming techniques.

Human activity can affect the landscape in two ways; by the direct creation of new landforms, and by an alteration in the rate at which geomorphic processes operate. Examples of the creation of new features in the landscape include reservoirs formed by the construction of dams (Fig. 1.17), and the effects of large-scale strip mining. In the

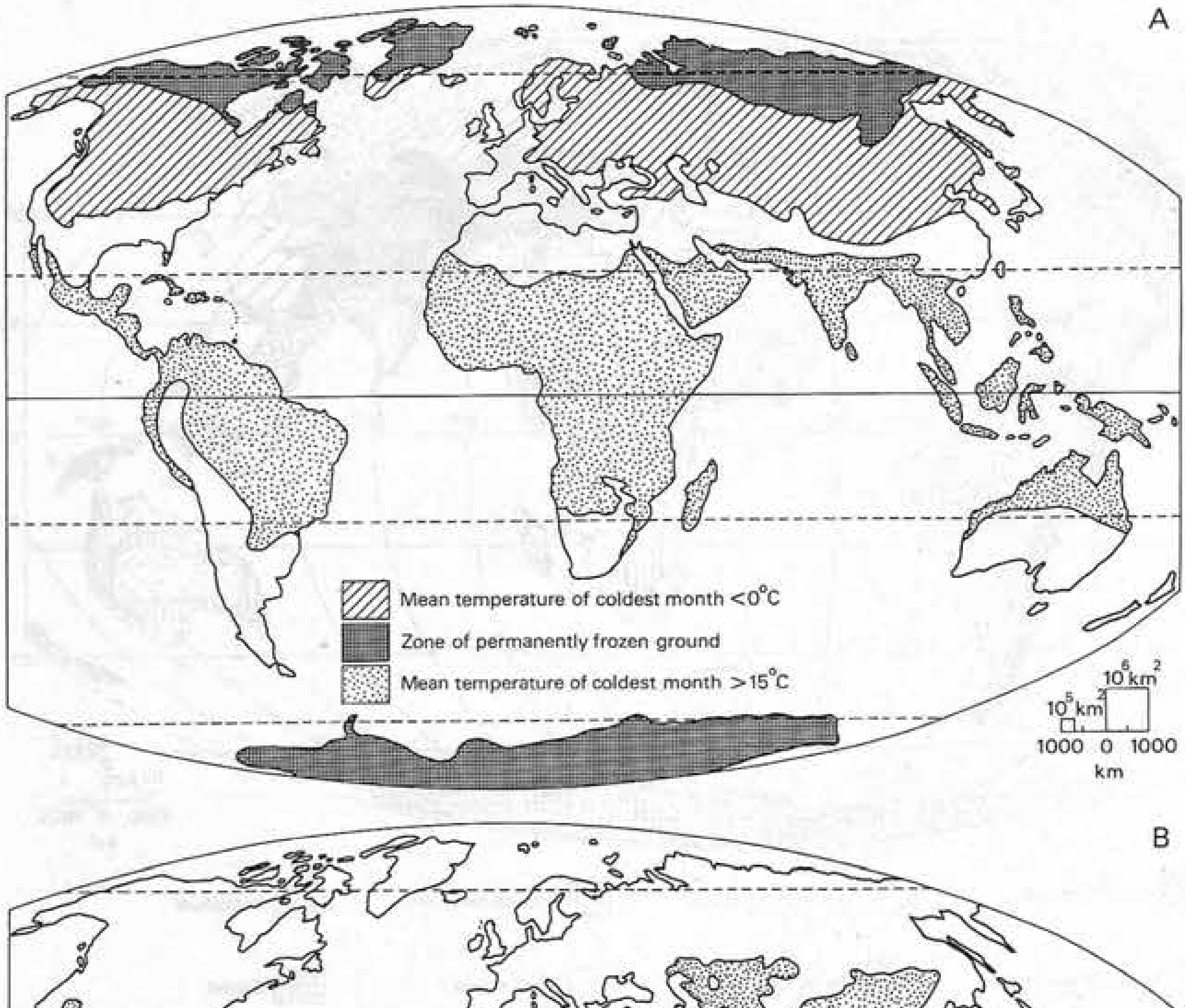

***Fig. 1.16** Global distribution of some key climatic variables: (A) important temperature limits; (B) areas experiencing high precipitation intensities and regions with low mean annual precipitation. (Based on data in R. Common (1966) in: G. H. Dury (ed.)* Essays in Geomorphology. *Heinemann, London 53–81; and B. M. Reich, 1963,* Journal of Hydrology *1, 3–28).*

Fig. 1.17 Oblique aerial view of the western end of Lake Mead on the Arizona–Nevada border, USA, a large reservoir formed through the impoundment of the Colorado River by the Hoover Dam (visible in the centre of the photograph).

Appalachian region of the U S A strip mining has produced benches on hillsides formed of rock waste which, it has been estimated, extend for a total distance of around 30 000 km. The modification of hillslopes is not, however, only a recent phenomenon, as a visit to many areas in the Mediterranean region would demonstrate (Fig. 1.18).

Of greater global significance than the direct creation of landforms is the effect of human activities on the rates of geomorphic processes. Agriculture, lumbering and the construction of settlements can all cause dramatic changes in land use which in turn may greatly change the susceptibility of the landsurface to erosion. Not surprisingly, the extension of activities such as agriculture into new regions over the past 100 years or so has led to some dramatic changes in the quantity of sediment being transported by rivers in the areas affected. Fluvial systems have, in many instances, undergone dramatic and complex adjustments as a result. If we are to understand how the natural landscape behaves it is vital that these effects are taken into account, otherwise we are in danger of interpreting the past in terms of an atypical present.

Fig. 1.18 Terraced hillsides in southern Cyprus dating back to at least the period of Roman occupation.

Further reading

A wide range of material is available for those interested in the history of geomorphology. An excellent recent survey is that by Tinkler (1985) while Chorley *et al.* (1964) provide a detailed assessment of developments up to the turn of the century. The second volume of this work (Chorley *et al.*, 1973) is devoted to W. M. Davis but two further planned volumes have yet to appear. A valuable study of the early years of geomorphology in Britain is that by Davies (1969) who has also recently presented a sober re-evaluation of Hutton's original contribution (Davies, 1985). Gould (1987) contains brilliant and highly readable essays exploring the conceptions of time embodied in the models of Earth history presented by Hutton and Lyell; this is required reading for those who really wish to grasp the philosophical framework of landform analysis established in the late eighteenth and early nineteenth centuries. Rudwick (1985) is (in spite of its title) also worth consulting for its discussion of the various meanings of uniformitarianism.

G. K. Gilbert is the subject of an excellent biographical study by Pyne (1980) emphasizing his methodological approach; briefer assessments of Gilbert's work are provided by Pyne (1975) and Yochelson (1980) while his key monographs are Gilbert (1877, 1914). Other original works of particular importance include: Hutton (1788) (contained in Eyles, 1970); Playfair (1802), which is available as a facsimile edition (White, 1956); Lyell (1830–33); Agassiz (1840), available in an English translation (Carozzi, 1967); Davis (1909); and Penck (1924), which is also available in translation (Penck, 1953). More recent developments, with a particular emphasis on fluvial geomorphology are reviewed by Gregory (1985), while Horton's seminal paper is well worth consulting (Horton, 1945).

Recommended reading on the various topics considered in the discussion of future directions in geomorphology is indicated in the various chapters which develop these themes, but it is worth mentioning here the thought-provoking essays by Baker (1986) and Hayden *et al.* (1986), the assessment of impact cratering on the Earth by Grieve (1987), the review of comparative planetary geomorphology by Sharp (1980) and the excellent case study of interpreting planetary landforms in terms of terrestrial analogues contained in Howard *et al.* (1988)

Thorn (1988) provides a useful introduction to concepts in geomorphology (although I feel he severely underplays the importance of endogenic factors in landscape development) while Baker (1988), Chorley (1978), Ritter (1988) and Scheidegger (1987) discuss from various perspectives key principles and methodological approaches. On specific concepts in geomorphology, Scheidegger (1979) considers the 'antagonism' between endogenic and exogenic processes and Summerfield (1981) briefly discusses their relative significance as the spatial scale is altered. Magnitude–frequency relationships are treated in a classic paper by Wolman and Miller (1960) and reconsidered by Wolman and Gerson (1978). The application of systems analysis to landform studies is discussed by Chorley (1962), Chorley and Kennedy (1971) and Huggett (1985). The problem of temporal scale in geomorphology is treated from various perspectives by Cullingford *et al.* (1980), Montgomery (1989), Schumm and Lichty (1965), Thorn (1982) and Thornes and Brunsden (1977). The related issue of appropriate conceptions of equilibrium in geomorphic research is considered by Chorley and Kennedy (1971) and Schumm (1977) (especially Chapter 1), although it is important to be aware of the different uses of the terms dynamic equilibrium and steady state in the latter in comparison with some earlier work. The modern usage of the term uniformitarianism is critically evaluated by Gould (1965) and Simpson (1970) provides a detailed discussion which introduces the notion of the immanent and configurational aspects of reality. There are relatively few general treatments of the problem of landform change through time, but Thornes and Brunsden (1977) cover a broad range of issues.

On methods of analysis Gardiner and Dackombe (1981) and Goudie (1981) between them comprehensively cover field and laboratory techniques in geomorphology. Schumm's detailed investigation of the evolution of microscale badland topography (Schumm, 1956) and the study by Schumm and Chorley (1964) of the fall of Threatening Rock provide classic examples of the value of direct observations. The general principles of space–time substitution are thoroughly reviewed by Paine (1985) while Brunsden and Kesel (1973) provide an example of the application of this approach. A useful introduction to simulation modelling is provided by Chorley and Kennedy (1971) and examples of the application of simulation modelling to fluvial and coastal landforms are presented by Howard (1971) and King and McCullagh (1971). Mosley and Zimpfer (1978) and Schumm *et al.* (1987) provide examples of the use of hardware models and Anderson (1988) contains detailed presentations of mathematical models.

The Earth's internal energy and the present global heat flow are treated at an introductory level by Pollack and Chapman (1977) while Williams (1982) outlines the surface energy budget (see especially Chapters 1 and 3). Barry (1969) provides a useful summary of the global hydrological cycle, although some of the data quoted have subsequently been revised. The principles of climatic geomorphology are critically summarized by Stoddart (1969) and presented in detail by Büdel (1982) and Tricart and Cailleux (1972). There is a large literature on applied geomorphology and the impact of human activities on the landscape with useful starting points being provided by Coates (1981), Cooke and Doornkamp (1974), Craig and Craft (1980), Gregory and Walling (1987) and Hails (1977).

Finally, it is appropriate here to mention some of the more important general sources of research and reference information in geomorphology. The four major journals covering geomorphic research are *Catena*, *Earth Surface Processes and Landforms*, *Geomorphology* and *Zeitschrift für Geomorphologie*, although articles on geomorphology are to be found in a wide range of geography and earth science journals. The reviews of recent developments in various fields within geomorphology contained in *Progress in Physical Geography* are particularly useful. Fairbridge (1968) is still a valuable reference work, but this has recently been supplemented by the less detailed but broader Goudie (1985). Snead (1981) is a useful geomorphic atlas while Short and Blair (1986) is an invaluable source of images of landforms taken from space with detailed accompanying commentaries.

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2

Global morphology and tectonics

2.1 Global morphology

2.1.1 The geoid

The Earth is only approximately a sphere. As a result of the centrifugal force of rotation it bulges at the equator and its polar radius (6378 km) is 21 km shorter than its equatorial radius (6397 km); thus the Earth is more accurately described as an oblate spheroid. Even this description, however, is not completely accurate as inhomogeneities in the distribution of mass in the Earth's interior produce further small but measurable irregularities on its surface. These irregularities have been determined with great precision over recent years through the very accurate measurement of deviations in the orbits of artificial satellites.

The shape of the Earth determined in this way is known as the **geoid** and is represented by the surface defined by mean sea level over the oceans and the extension of sea level along imaginary canals across the continents. As is evident in Figure 2.1 this surface shows many irregularities compared with a simple oblate spheroid; there is, for instance, a bulge of 76 m near New Guinea and a depression of some 104 m to the south of India. The significance of the geoid for the operation of geomorphic processes is that it defines the ultimate base level for denudation; any change in the geoid would therefore cause consequential changes in base levels. This is an important issue in studies of sea-level change and is examined further in Chapter 17.

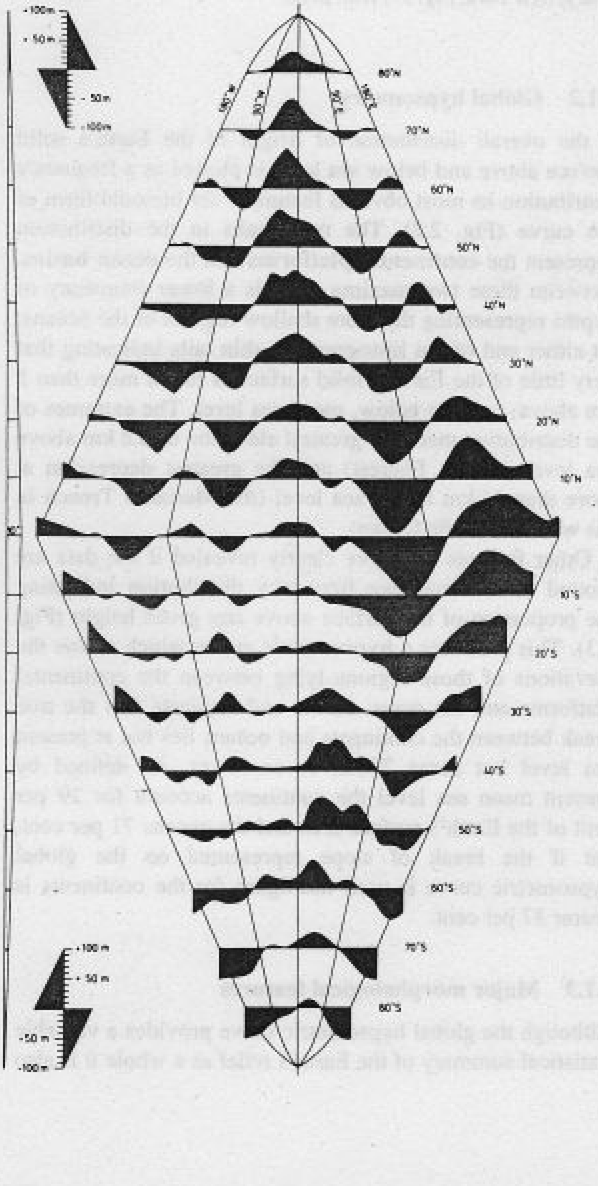


Fig. 2.1 Geodetic sea level profiles at intervals of 10° latitude. The horizontal latitudinal lines separate positive geoid values (above) from negative geoid values (below). (From N. –A. Mörner (1976), Journal of Geology *84, Fig. 1, p. 124.)*

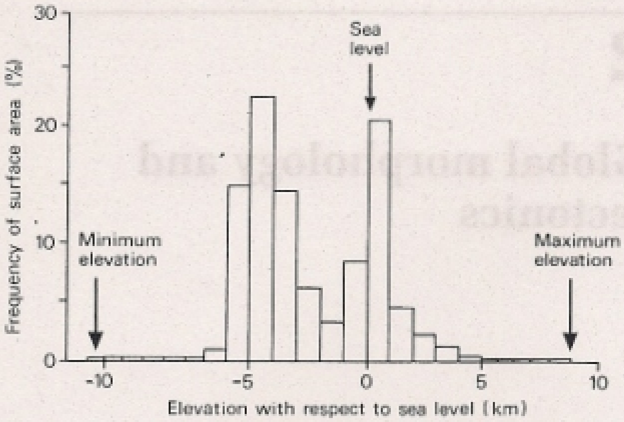


Fig. 2.2 Proportionate areal distribution of the solid surface of the Earth between successive elevations. Note the two peaks representing the ocean basins and the continental platforms. (Modified from P. J. Wyllie 1976, The Way the Earth Works. *Wiley, New York, Fig. 3–11(a), p. 38.*

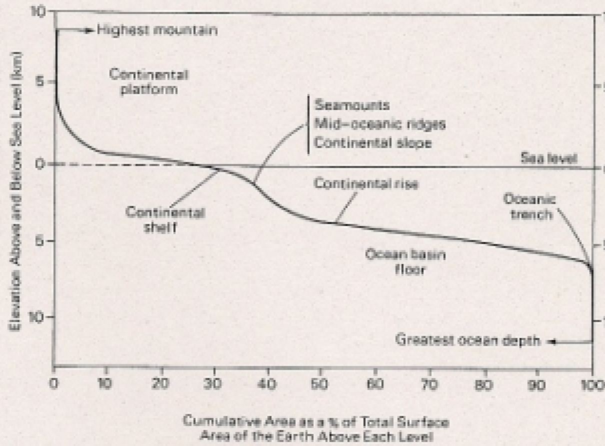


Fig. 2.3 Global hypsometric curve indicating the main morphological features characterizing the Earth's surface at different elevations. (Modified from P. J. Wullie (1976) The Way the Earth Works. *Wiley, New York, Fig. 3–11(b), p. 38.).*

2.1.2 Global hypsometry

If the overall distribution of height of the Earth's solid surface above and below sea level is plotted as a frequency distribution its most obvious feature is the bimodal form of the curve (Fig. 2.2). The two peaks in the distribution represent the **continental platforms** and the **ocean basins.** Between these two maxima there is a lower frequency of depths representing the more shallow regions of the oceans. At either end of the histogram are thin tails indicating that very little of the Earth's solid surface is found more than 5 km above, or 6 km below, mean sea level. The extremes of the distribution mark the greatest elevation of 8.8 km above sea level (Mount Everest) and the greatest depression at more than 11 km below sea level (the Marianas Trench in the western Pacific Ocean).

Other features are more clearly revealed if the data are plotted as a cumulative frequency distribution indicating the proportion of the surface above any given height (Fig. 2.3). This produces a **hypsometric curve** which shows the elevations of those regions lying between the continental platforms and the ocean basins and suggests that the true break between the continents and oceans lies not at present sea level but some 200 m or so lower. As defined by present mean sea level the continents account for 29 per cent of the Earth's surface area and the oceans 71 per cent, but if the break of slope represented on the global hypsometric curve is used the figure for the continents is nearer 37 per cent.

2.1.3 Major morphological features

Although the global hypsometric curve provides a valuable statistical summary of the Earth's relief as a whole it is also necessary to identify the major components of global morphology that give rise to the curve (Fig. 2.4). Beginning with the more familiar form of the continents we can distinguish between the continental platforms formed by plateaus and lowlands, and the major linear mountain systems known as **orogenic mountain belts**, or simply **orogens**. One major mountain system cuts across central and southern Eurasia extending from the Alps in the west, through the Himalayas to the mountains of western China. Another, formed by the Andes and the North American Cordillera, runs along the entire western margin of the Americas. Other older and more subdued ranges include the Appalachians in eastern North America and the Urals in western Eurasia. Extensive areas within continental platforms are formed of **basement**, a complex of metamorphic and igneous rocks of Palaeozoic or Precambrian age. In some localities, notably East Africa, continental platforms are traversed by **rift valleys** consisting of linear troughs formed by the subsidence of crust between parallel systems of faults.

The submarine extension of a continent is called the **continental shelf**. This is bordered by a **continental slope** which inclines at an angle of around 3–6° towards the ocean basin and which is separated from it by a **continental rise** (Fig. 2.5). The ocean basins themselves are traversed by a vast system of **mid-oceanic ridges** up to 1000 km in width and tens of thousands of kilometres in length. On average they rise some 2 km above the surrounding ocean floor but in a few localities, notably at Iceland, they reach above sea level. Although the surface of the ocean basins is relatively uniform it is punctuated in places by volcanoes. Some of these rise above sea level, but the vast majority are submarine features known as **seamounts**, or if they are flat-topped, **guyots**. The greatest ocean depths are

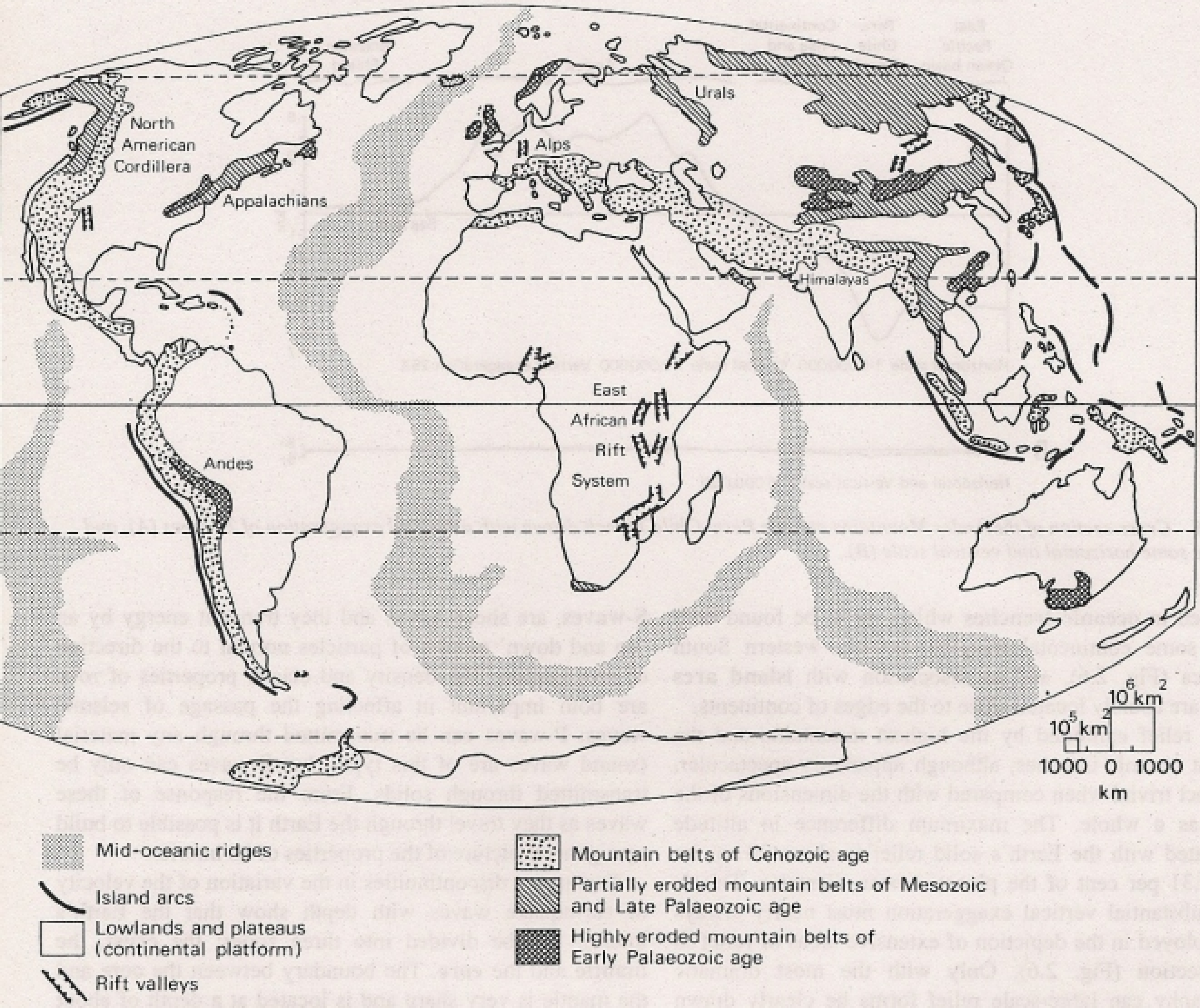


Fig. 2.4 Major morphological features of the Earth.

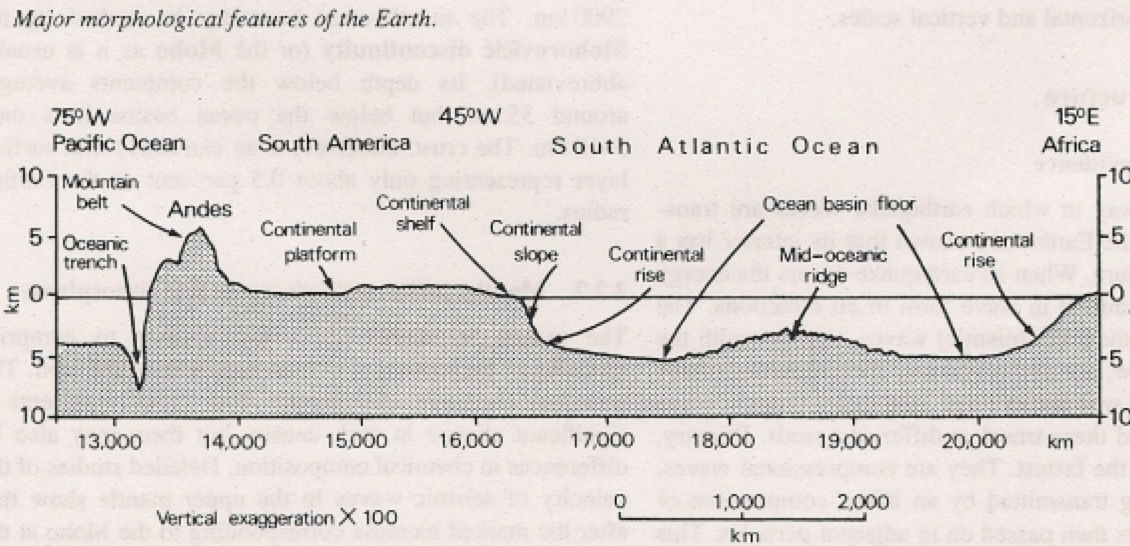


Fig. 2.5 Schematic cross-section of major morphological features from western South America to the west coast of Africa at the latitude of the Tropic of Cancer. Note that the curvature of the Earth is not shown. (Modified from P. J. Wyllie (1976) The Way the Earth Works. *Wiley, New York, Fig. 3–10, pp. 36–7.)*

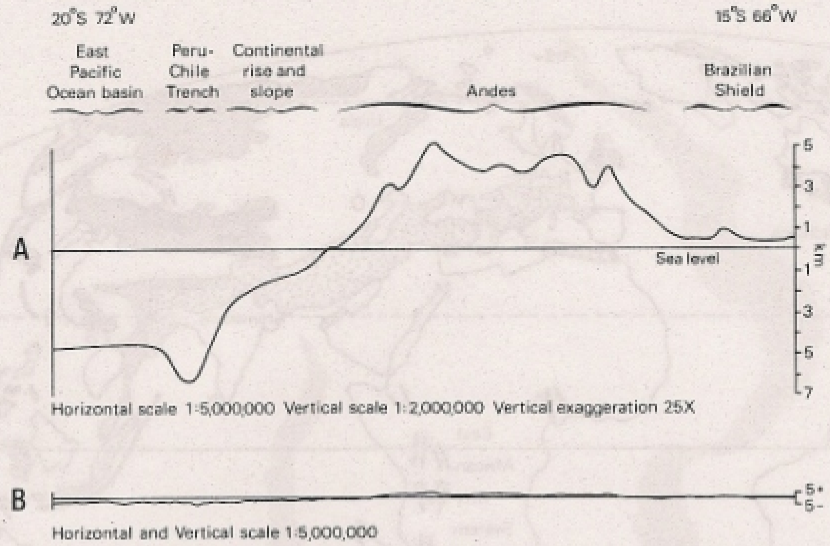


Fig. 2.6 Cross-section of the Andes Mountains and the Peru–Chile Trench drawn with a vertical exaggeration of 25 times (A), and with the same horizontal and vertical scale (B).

achieved in **oceanic trenches** which are to be found both along some continental margins, notably western South America (Fig. 2.6), and in association with **island arcs** which are usually located close to the edges of continents.

The relief exhibited by the highest mountains and the deepest oceanic trenches, although apparently spectacular, is in fact trivial when compared with the dimensions of the Earth as a whole. The maximum difference in altitude associated with the Earth's solid relief is about 20 km, or only 0.31 per cent of the planet's mean diameter. This is why substantial vertical exaggeration must nearly always be employed in the depiction of extensive areas of relief in cross-section (Fig. 2.6). Only with the most dramatic topography can large-scale relief forms be clearly drawn with identical horizontal and vertical scales.

2.2 Earth structure

2.2.1 Seismic evidence

Studies of the way in which earthquake waves are transmitted through the Earth have shown that its interior has a concentric structure. When an earthquake occurs the energy released is transmitted in wave form in all directions. The velocities of earthquake (seismic) waves, together with the paths they follow, are determined by the properties of the material through which they pass. Two major types of wave are produced and these travel at different speeds. Primary, or **P-waves**, are the fastest. They are compressional waves, the energy being transmitted by an initial compression of particles which is then passed on to adjacent particles. This produces a sequence of zones of compression and expansion which travel away from the source. Secondary, or **S-waves**, are shear waves and they transmit energy by an 'up and down' motion of particles normal to the direction of propagation. The density and elastic properties of rock are both important in affecting the passage of seismic waves. P-waves can be transmitted through any material (sound waves are of this type), but S-waves can only be transmitted through solids. From the response of these waves as they travel through the Earth it is possible to build up a general picture of the properties of its interior.

Two major discontinuities in the variation of the velocity of earthquake waves with depth show that the Earth's interior can be divided into three zones; the **crust**, the **mantle** and the **core**. The boundary between the core and the mantle is very sharp and is located at a depth of about 2900 km. The mantle-crust boundary is marked by the **Mohorovičić discontinuity** (or the **Moho** as it is usually abbreviated). Its depth below the continents averages around 35 km, but below the ocean basins it is only 5–10 km. The crust, therefore, is an extremely thin surface layer representing only about 0.5 per cent of the Earth's radius.

2.2.2 Mantle, asthenosphere, crust and lithosphere

The mantle is mainly solid and appears to comprise minerals of high density, rich in magnesium and iron. The boundary between the mantle and crust represents a significant change in rock density but there may also be differences in chemical composition. Detailed studies of the velocity of seismic waves in the upper mantle show that after the marked increase corresponding to the Moho at the crust–mantle boundary their velocity gradually becomes greater to a depth of around 100 km, where a small, but

Table 2.1 Summary of the properties of the crust, mantle, lithosphere and asthenosphere

	THICKNESS	VELOCITIES OF P-AND S-WAVES (km s^{-1})	MEAN DENSITY (kg m^{-3})	BEHAVIOUR	
Crust	Continental crust Mean: 35 km Min: <30 km Max: 70 km Oceanic crust: 5–10 km	Increase with depth to 6.6 P-waves 3.8 S-waves but considerable spatial variation	Continental crust: 2700 Oceanic crust: 3000	Solid to top of low velocity zone. Elastic deformation under vertical crustal loading.	**Lithosphere**
	Moho				
Mantle	Base of crust to depth of 2900 km	Immediately below Moho: 8.5 P-waves 4.8 S-waves	3320 immediately below Moho to 5600 at base of mantle		**Lithosphere**
		Decrease to minimum of 7.8 P-waves 4.2 S-waves in low velocity zone between depth of 100 and 300 km		Very weak in low velocity zone (depth range 100–300 km). Inelastic deformation or 'flow'.	**Asthenosphere**
		Below 300 km increase to 14.0 P-waves 7.5 S-waves at base of mantle		Below low velocity zone progressively less weak, but convection probably possible deep into the mantle.	

significant, decrease is found which continues to a depth of about 300 km (Table 2.1).

This layer of attenuated velocity is known as the **low velocity zone** and the reduction in seismic wave velocities is considered to be due to partial melting in this region of the mantle. It partly corresponds to the **asthenosphere**, the plastic-like properties of which permits slow 'flow' of material in response to forces applied over long periods of time (Fig. 2.7). Lubrication may be provided by melting between mineral boundaries, but molten material must only form a small proportion of the asthenosphere since it is capable of transmitting S-waves.

Compared with the mantle, the crust forms a very thin layer. Beneath the oceans it is thought to extend to a depth of around 10–15 km, that is to only 5–10 km below the ocean floor. Below the continents the crust is much thicker, averaging about 35 km but reaching up to 70 km beneath some mountain ranges.

Of greatest significance to the creation of macroscale landforms is the boundary of the asthenosphere with the solid mantle above. The crust and the immediately underlying mantle located above the asthenosphere in fact appear to behave as a coherent semi-rigid layer called the **lithosphere**. Two types can be distinguished: **oceanic lithosphere** is capped by thin oceanic crust, while **continental lithosphere** is capped by much thicker continental crust. The thickness of the lithosphere itself shows considerable variability and the boundary with the underlying asthenosphere is gradational and difficult to define precisely. In addition to the analysis of seismic data, heat

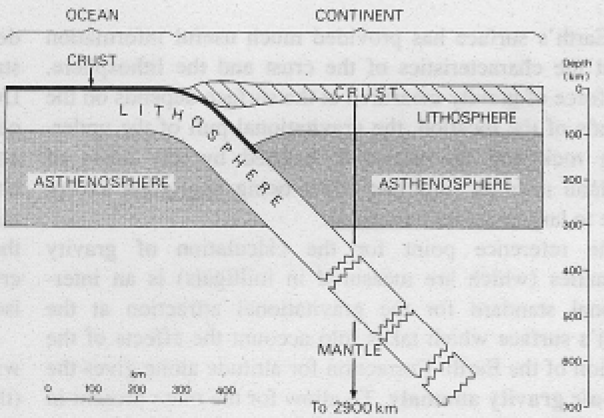


Fig. 2.7 Schematic representation of the relationship between crust, mantle, lithosphere and asthenosphere. Reincorporation of the lithosphere into the underlying mantle is also shown (see Section 2.4.)

flow measurements have been employed to map variations in lithospheric thickness on a global basis (Fig. 2.8). These indicate generally thin lithosphere beneath the oceans, especially in the vicinity of mid-oceanic ridges, and a marked thickening in certain continental areas, including Antarctica, North Africa, Brazil and north-eastern North America.

2.2.3 Gravity anomalies

Measurement of variations in the gravitational attraction at

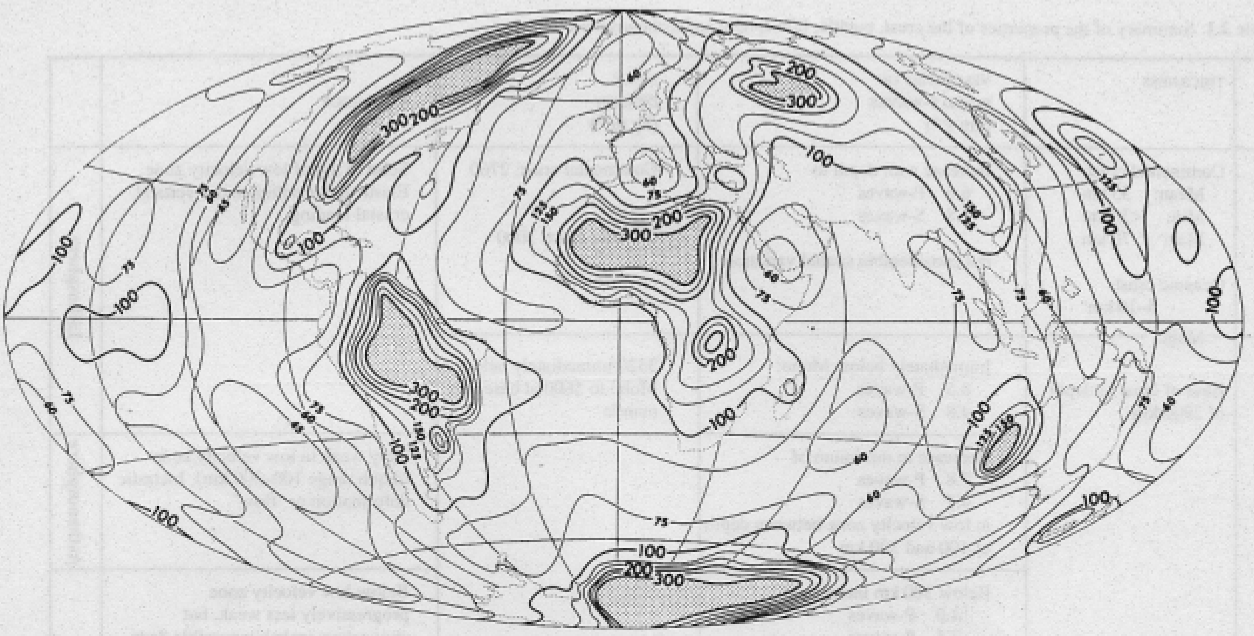


Fig. 2.8 Global variation in the thickness of the lithosphere estimated from heat flow measurements. (From D. S. Chapman and H. N. Pollack (1977) Geology *5, Fig. 4, p. 268.)*

the Earth's surface has provided much useful information about the characteristics of the crust and the lithosphere. The force of gravity measured at the surface depends on the altitude of the location, the gravitational pull of the underlying rock and the attraction exerted by any areas of highland near by (this last effect being negligible except close to large mountain masses).

The reference point for the calculation of gravity anomalies (which are measured in milligals) is an international standard for the gravitational attraction at the Earth's surface which takes into account the effects of the rotation of the Earth. Correction for altitude alone gives the **free-air gravity anomaly**. To allow for the mass present in a mountain range above the reference surface the theoretical pull of the rock, based on an assumed average rock density, is subtracted from the free-air anomaly (or a correction is made for the presence of sea water if the point is over the ocean). This gives the **Bouguer anomaly** which is the most widely used measure of gravitational deviations over the Earth's surface.

2.2.4 Isostasy

As all but the uppermost part of the mantle appears capable of viscous flow when subject to prolonged stress, the semi-rigid lithosphere, capped by continental or oceanic crust, can be viewed as 'floating' on the underlying asthenosphere (the most easily deformable viscous part of the mantle). To attain hydrostatic equilibrium the position of the lithosphere adjusts vertically in accordance with its density and thickness. The term **isostasy** (meaning 'equal standing') was introduced by the American geologist Dutton in 1889 to describe this state of equilibrium. Until quite recently it was considered that it was the crust that attained equilibrium with respect to the mantle, but it is now known that isostatic adjustments also involve the rigid mantle forming the lower part of the lithosphere. Nevertheless it is differences in the density and thickness of the crust that are largely responsible for variations in the isostatic adjustment of the lithosphere as a whole.

If it is in isostatic equilibrium one section of lithosphere will stand higher than another because it is of lower density (the crustal density or **Pratt model**), of the same density but thicker (the crustal thickness or **Airy model**), or through a combination of both a lower density and greater thickness (Fig. 2.9). Continental lithosphere stands higher than oceanic lithosphere because continental crust is both of greater thickness and lower density than oceanic crust. The great differences in elevation within the continents are, in most cases, related to variations in crustal thickness with areas at high elevations generally being underlain by deep roots of buoyant crustal rock (Box 2.1).

Bouguer anomalies are generally negative on land and positive over the oceans. This would be expected from differences in the density and thickness of continental and oceanic crust. Bouguer anomalies include a correction for altitude but only on the basis of an average crustal density. The continents, particularly where elevations are high, exhibit lower than predicted (negative) gravity values because of the greater thickness of underlying crust of

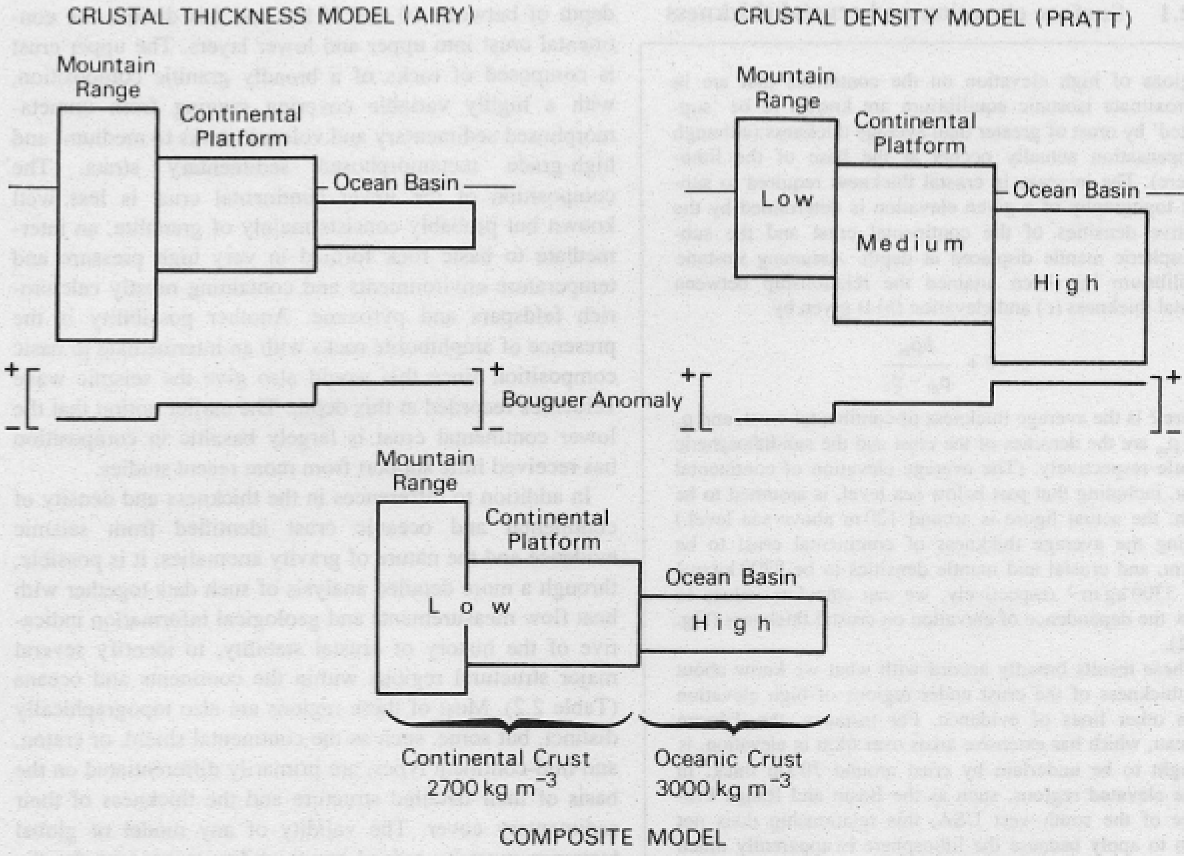


Fig. 2.9 A schematic representation of three models of isostatic equilibrium. Although differences in the density and thickness of the crust are illustrated, the lithosphere as a whole is involved in isostatic compensation.

relatively low density. Higher than predicted (positive) gravity values occur over the oceans because they are underlain by thin and relatively dense crust.

When corrections are made to take account of these differences in crustal density the magnitude of gravity anomalies is significantly reduced. Remaining discrepancies are termed **isostatic anomalies**, but why should such anomalies exist when the lithosphere is apparently free to attain isostatic equilibrium with respect to the asthenosphere?

One reason is that the lithosphere is not divided into small discrete blocks able to move freely up and down with respect to each other. The lithosphere possesses a certain degree of rigidity so that the mass of any load (such as an ice sheet) placed on it is supported over a greater area than that covered by the load itself. In other words the lithosphere experiences flexure, just as a springboard does when a diver walks along it. This kind of behaviour, known as **flexural isostasy**, is especially important for oceanic lithosphere and for thick continental lithosphere with a high rigidity (see Section 4.2.3).

A second reason for isostatic anomalies is that the lithosphere is not capable of adjusting instantaneously to a change in load. Although the great ice sheets which covered much of North America and northern Europe in the recent past had largely disappeared by 10 000 a BP, the landsurface in these regions is still rising rapidly in response to the removal of this load.

A third reason for isostatic anomalies is that there are dynamic forces present in the sub-lithospheric mantle which are capable of actively dragging down or pushing up the lithosphere. These forces also play a key role in the large horizontal movements experienced by the lithosphere which we will be examining in detail later in this chapter.

2.2.5 Crustal structure

Seismic and gravity data, together with direct evidence from the rocks themselves, allow us to identify the structural and compositional differences between oceanic and continental crust. Oceanic crust has a mean density of about 3000 kg m^{-3} and is composed of layers of basic rocks, broadly basaltic and gabbroic in composition, with a thin veneer of sediments. Over most of the ocean floor this sedimentary cover is only 1–2 km thick and in the vicinity of mid-oceanic ridges it becomes very thin indeed. One of the most remarkable discoveries arising from oceanographic research

Box 2.1 Surface elevation and crustal thickness

Regions of high elevation on the continents that are in approximate isostatic equilibrium are known to be 'supported' by crust of greater than average thickness (although compensation actually occurs at the base of the lithosphere). The increase in crustal thickness required to support topography of a given elevation is determined by the relative densities of the continental crust and the sub-lithospheric mantle displaced at depth. Assuming isostatic equilibrium has been attained the relationship between crustal thickness (c) and elevation (h) is given by

$$c = \bar{c} + \frac{h\rho_m}{\rho_m - \rho_c}$$

where $\bar{c}$ is the average thickness of continental crust, and ρ_c and ρ_m are the densities of the crust and the sub-lithospheric mantle respectively. (The average elevation of continental crust, including that part below sea level, is assumed to be 0 km; the actual figure is around 120 m above sea level.) Taking the average thickness of continental crust to be 35 km, and crustal and mantle densities to be 2700 kg m^{-3} and 3300 kg m^{-3} respectively, we can calculate values to show the dependence of elevation on crustal thickness (Fig. B2.1).

These results broadly accord with what we know about the thickness of the crust under regions of high elevation from other lines of evidence. For instance, the Tibetan Plateau, which has extensive areas over 6km in elevation, is thought to be underlain by crust around 70 km thick. In some elevated regions, such as the Basin and Range Province of the south-west USA, this relationship does not seem to apply because the lithosphere is apparently much thinner than average and is underlain by unusually hot asthenosphere of lower than normal density.

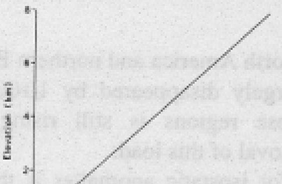


Fig. B2.1 Elevation of continental areas above sea level as a function of crustal thickness.

over the past two decades or so has been the youth of oceanic sediments and the underlying basaltic crust. The oldest known rocks from the ocean floor come from the western Pacific, but these are only of Jurassic age. This can be contrasted with the antiquity of rocks exposed over large areas of the continents which in some cases are more than 3000 Ma old.

Like the oceanic crust, continental crust has a layered structure, but one which is much more complex and less clearly defined. A seismic boundary, known as the **Conrad discontinuity**, underlies some continental regions at a depth of between 10 and 30 km and this divides the continental crust into upper and lower layers. The upper crust is composed of rocks of a broadly granitic composition, with a highly variable covering ranging from unmetamorphosed sedimentary and volcanic rocks to medium- and high-grade metamorphosed sedimentary strata. The composition of the lower continental crust is less well known but probably consists mainly of granulite, an intermediate to basic rock formed in very high pressure and temperature environments and containing mostly calcium-rich feldspars and pyroxene. Another possibility is the presence of amphibolite rocks with an intermediate to basic composition since this would also give the seismic wave velocities recorded at this depth. The earlier notion that the lower continental crust is largely basaltic in composition has received little support from more recent studies.

In addition to differences in the thickness and density of continental and oceanic crust identified from seismic evidence and the nature of gravity anomalies, it is possible, through a more detailed analysis of such data together with heat flow measurements and geological information indicative of the history of crustal stability, to identify several major structural regions within the continents and oceans (Table 2.2). Most of these regions are also topographically distinct, but some, such as the continental shield, or craton, and mid-continent types, are primarily differentiated on the basis of their detailed structure and the thickness of their sedimentary cover. The validity of any model of global tectonics must be judged by its ability to account for the distribution and characteristics of these major structural regions.

2.3 Development of ideas on global tectonics

Serious scientific attempts to explain the major structural and relief features of the Earth began only in the nineteenth century. In 1829 Elie de Beaumont put forward the idea that the Earth is contracting and argued that compressional stresses set up in the crust as a result of the cooling of the Earth's interior would give rise to faulting, folding and thickening of the crust, and eventually to the formation of mountain ranges. This proposal found support in the work of Lord Kelvin, the pioneer Victorian geophysicist, who attempted to calculate the age of the Earth from its probable rate of cooling on the assumption that it had formed as a molten offshoot of the Sun. However, as the role of radioactive decay in generating heat in the Earth's interior (and thus drastically reducing the previously hypothezised rate of cooling) became appreciated the notion of a contracting Earth was rejected. Various global tectonic models were proposed during the nineteenth and the first part of the twentieth century, with most attention being focused on the origin of the major mountain systems of the Earth's surface. The majority of the models assumed that the

Table 2.2 Structural classification of crustal types

CRUSTAL TYPE	TYPICAL CRUSTAL THICKNESS (km)	HEAT FLOW ($mW\ m^{-2}$)	BOUGUER ANOMALY (mgal)	DEGREE OF STABILITY	CHARACTERISTICS
Continental shield, or craton	35	29–38	–10 to –30	Very stable	Low to moderate elevation. Composed of highly deformed Precambrian metamorphic and plutonic rocks, unaffected by post-Precambrian tectonism. No covering of post-Precambrian sediments
Mid-continent	38	33–50	–10 to –40	Stable	Generally similar to continental shields, the major difference being the development since the Precambrian of broad undulations which have led to the accumulation of thick sedimentary sequences in extensive basins, particularly near continental margins. The mid-continent structural type commonly occurs adjacent to continental shields and together they account for the majority of the area of continental platforms
Basin and range	30	71–105	–200 to –250	Very unstable	Named after the type area of the Basin and Range Province of the western U S A, this structural type is characterized by great instability associated with significant extension of the crust giving rise to a series of basins and intervening, usually significantly eroded, upland areas. Notable are the very thin crust (for continental regions), the high mean elevation, high heat flow, strongly negative gravity anomalies and marked volcanic and seismic activity
Young mountain belt or active orogen	55	29–84	–200 to –300	Very unstable	Comparatively narrow, elongated regions attaining elevations in excess of 3 km which have experienced relatively recent and often rapid uplift, in many cases preceded by intense folding and thrusting of thick sedimentary sequences as a result of crustal compression. Crustal thicknessess are highly variable, ranging up to a maximum of about 70 km. There is also a wide range of heat flow rates with higher rates being characteristic of younger mountain belts
High plateau	35?	84?	–150 to –250	Very unstable	This crustal type lacks a distinct geophysical character, but is typified by high elevations resulting from uplift lacking associated folding or thrust faulting. The Colorado Plateau provides an example
Island arc	30	29–167	–50 to +100	Very unstable	This type includes a wide range of structural forms, including significant 'continental' fragments such as Japan and New Zealand, as well as arcs formed of numerous individual volcanic peaks of predominantly andesitic composition. Crustal thickness, heat flow rates and Bouguer anomalies are all highly variable both between and within island arc systems. Volcanic and seismic activity is intense
Oceanic trench	?	Low	Strongly negative	Very unstable	Closely associated with island arcs but are best considered separately as they may be found adjacent to young mountain belts. Earthquake activity is marked, particularly towards the adjacent island arc or mountain belt
Ocean basin	11	54	+250 to +350	Very stable	Covers extensive areas of the ocean. Ocean basins are broken by long linear fractures or faults and are punctuated in places by volcanoes of basaltic composition
Mid-oceanic	10	42–335	+200 to +250	Unstable	Crustal type typified by shallow earthquakes and abundant volcanic activity and composed of basaltic lava. Some ridges have a central rift valley and at a limited number of points they break the surface of the ocean to form islands

Source: Based mainly on J. N. Brune (1969) *American Geophysical Union Monograph* 13, 230–42; and E. W. Spencer (1977) *Introduction to the Structure of the Earth* (2nd edn) McGraw-Hill, New York.

positions of the continents were fixed and that the ocean basins were ancient features.

2.3.1 Continental drift

The notion of **continental drift** is not new. It was canvassed on a number of occasions in the nineteenth century, but its most influential advocate was Alfred Wegener who intially presented his ideas in 1912. He outlined three main lines of supporting evidence. First, he pointed out the 'fit' of coastlines now separated by thousands of kilometres of ocean, in particular those of South America and Africa. He reconstructed the present land masses into a single supercontinent called Pangaea and suggested that this had initially split into two continents, Gondwanaland (Gondwana) to the south and Laurasia to the north, before further rupture and drift resulted in the familiar shape and location of the continents today.

Secondly, Wegener, a climatologist by training, referred to the global distribution of rocks characteristic of particular climatic environments, particularly ancient glacial deposits (**tillites**). He maintained that the distribution of tillites and patterns of glacial striations produced by ice sheets during the Late Palaeozoic glaciation (now termed the Gondwana Ice Age) found in the now widely dispersed continental areas of southern Africa, Australia, South America, India and Antarctica indicated that these land masses were contiguous at that time and probably located fairly close to the South Pole. He also pointed out, though somewhat less convincingly, that some geological structures could be traced from one continent to another across what are now wide stretches of ocean.

Wegener's third main category of evidence was palaeontological. He argued that there were marked similarities between fossil terrestrial fauna and flora from Palaeozoic strata in the various southern continents and between North America and Europe, suggesting that during this era there was free movement over a single large land mass. Fossils from more recent strata showed progressively less similarities with time in the different continents; this he attributed to the contrasting evolutionary paths of groups of animals and plants separated by continental drift.

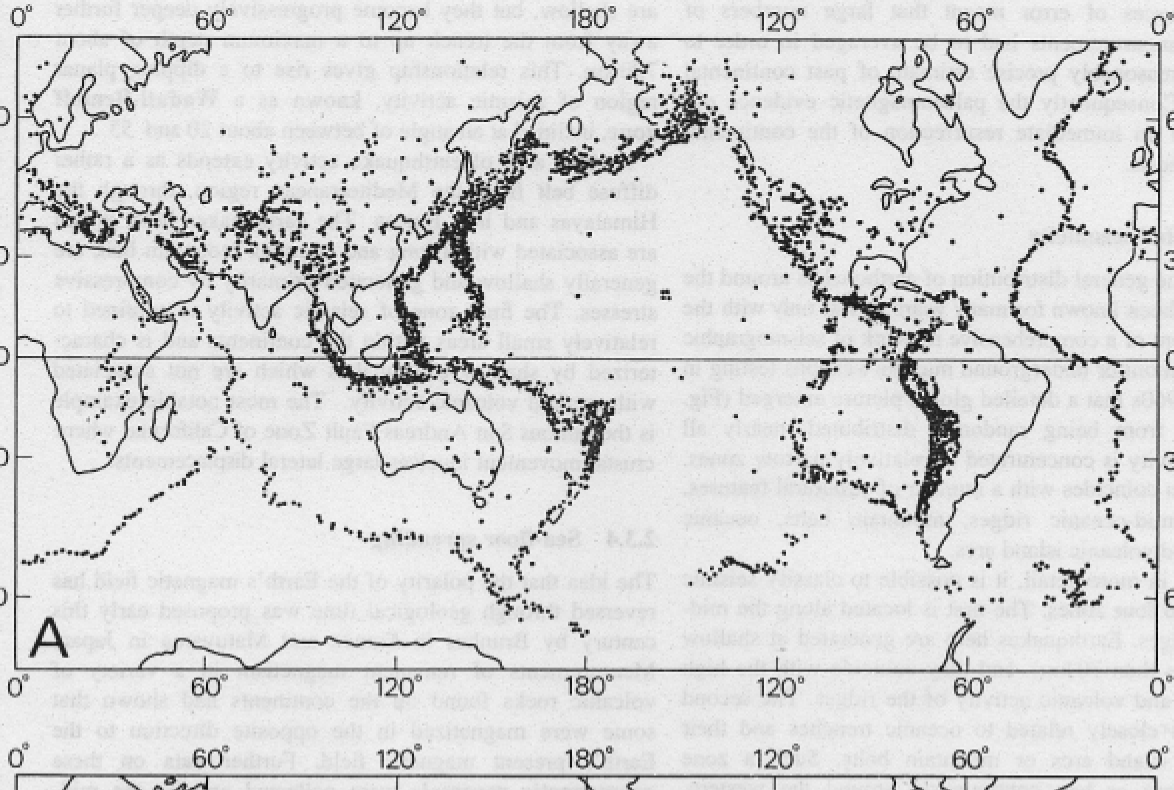
The greatest obstacle to the acceptance of the drift hypothesis was not so much the nature of this supporting evidence (although alternative explanations were readily proffered) but rather the failure to find a convincing mechanism by which it could occur. From an examination of global hypsometry Wegener had noted the concentration of large areas of the Earth's surface around two levels representing the continental platforms and the ocean basins. He saw this as being compatible with the crust being made up of two layers, the upper one of relatively low density, the lower layer of higher density. He suggested that tidal forces resulted in the lighter continental crust 'ploughing' through the substratum of denser crust underlaying the oceans. This proposal was readily dismissed by geophysicists who were easily able to demonstrate that the Earth is far too strong to be deformed by such tidal forces. Suggestions made by Arthur Holmes and others that continents could be moved by convection currents within the mantle were ignored rather than countered by opponents of drift. While information about the Earth's interior and the ocean floors was so sparse, most geologists preferred the safety of the established doctrine of stationary continents.

2.3.2 Palaeomagnetic evidence

During the mid-1950s, at a time when continental drift was not seriously considered by most earth scientists, new evidence in the form of palaeomagnetic data from rocks again began to bring into question the notion of stationary continents. S.K. Runcorn and his associates, working in Britain, conducted an intensive programme of data collection involving the measurement of **remanent magnetism** in rocks of various ages from around the world. Earlier studies of such palaeomagnetism in France and Japan had shown that iron-rich volcanic rocks, such as basalt, record the magnetic field prevailing at the time they are formed. As basaltic lavas cool through the temperature interval 500 to 450 °C (the **Curie point**) the atomic groups within the iron minerals they contain become aligned parallel to the magnetic lines of force acting upon them. Once the temperature of such rocks falls below 450 °C this magnetic orientation becomes 'frozen' into the individual minerals, only being subsequently disturbed by marked heating. It was found that the palaeomagnetism of young rocks tended to be close to that expected from the present magnetic field, but older rocks showed marked deviations.

As a rough approximation the Earth's present magnetic field can be represented by a regular dipolar pattern (similar to that produced when a bar magnet is held beneath iron filings scattered on a sheet of paper). If it is assumed that the Earth's magnetic field was of this form in the past, then it is possible to estimate the position of the magnetic pole for rocks of known ages (dated by radiometric methods) by measuring their palaeomagnetism. On the basis of such measurements on rocks from Europe, Runcorn and his colleagues demonstrated an apparent movement of the magnetic pole over the past 500 Ma. This phenomenon was termed **polar wandering** and is now thought to arise from the movements of the continents themselves rather than from any significant shift in the location of the magnetic pole itself.

After extending palaeomagnetic investigations to rocks from North America it was found that while these too showed an apparent movement of the magnetic pole the path of polar wandering appeared to differ systematically from that determined for Europe. This discrepancy could be explained if North America and Europe had been moving with respect to each other as well as with respect to the magnetic pole. Further work began to suggest that palaeomagnetic data supported the model of continental drift proposed by Wegener although there were initial uncertainties over the accuracy of the technique. Palaeomagnetic data could indicate north–south movements (palaeolatitudes) but they could not give past longitudinal positions. Moreover, the weakness of the remanent magnetism contained within volcanic rocks together with



ig. 2.10 Location of approximately 30 000 earthquakes recorded by the US Coast and Geodetic Survey between 1961 and 1967. In) all the registered earthquakes are shown, whereas (B) records only those earthquakes generated at depths in excess of 100 km, with a separate indication of those regions experiencing deep focus seismic events (300–700 km). (Modified from M. Barazangi and J. orman (1969) Bulletin of the Seismological Society of America *59, 370–6.)*

various sources of error meant that large numbers of individual measurements had to be averaged in order to provide a reasonably precise estimate of past continental positions. Consequently the palaeomagnetic evidence did not lead to an immediate resurrection of the continental drift hypothesis.

2.3.3 Global seismicity

Although the general distribution of earthquakes around the world had been known for many years, it was only with the establishment of a comprehensive network of seismographic stations to monitor underground nuclear weapons testing in the early 1960s that a detailed global picture emerged (Fig. 2.10). Far from being randomly distributed, nearly all seismic activity is concentrated in relatively narrow zones. This pattern coincides with a number of structural features, including mid-oceanic ridges, mountain belts, oceanic trenches and volcanic island arcs.

Looking in more detail, it is possible to classify seismic activity into four zones. The first is located along the mid-oceanic ridges. Earthquakes here are generated at shallow depths (less than 70 km), and they coincide with the high heat flows and volcanic activity of the ridges. The second category is closely related to oceanic trenches and their associated island arcs or mountain belts. Such a zone extends more or less continuously around the western, northern and eastern margins of the Pacific Ocean. The most notable feature of this seismic zone is the relationship between the depths of earthquake generation and their location with respect to the adjacent island arc or mountain belt (Fig. 2.11). On their oceanic margin earthquake foci are shallow, but they become progressively deeper further away from the trench up to a maximum depth of about 700 km. This relationship gives rise to a dipping planar region of seismic activity, known as a **Wadati-Benioff zone**, inclined at an angle of between about 20 and 55°.

A third area of earthquake activity extends as a rather diffuse belt from the Mediterranean region, through the Himalayas and into Burma. The earthquakes here, which are associated with a long and complex mountain belt, are generally shallow and generated primarily by compressive stresses. The final zone of seismic activity is confined to relatively small areas within the continents and is characterized by shallow earthquakes which are not associated with marked volcanic activity. The most notable example is the famous San Andreas Fault Zone of California, where crustal movement involves large lateral displacements.

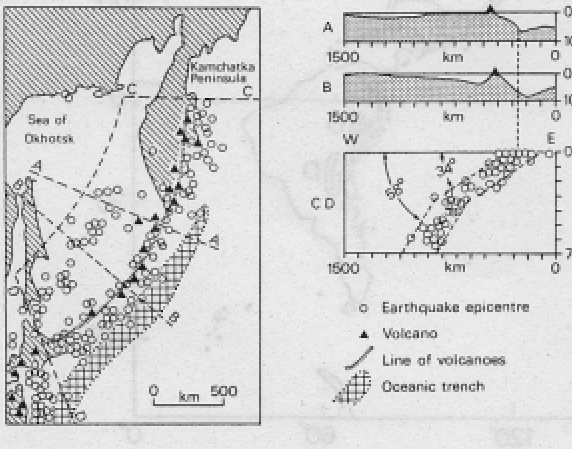


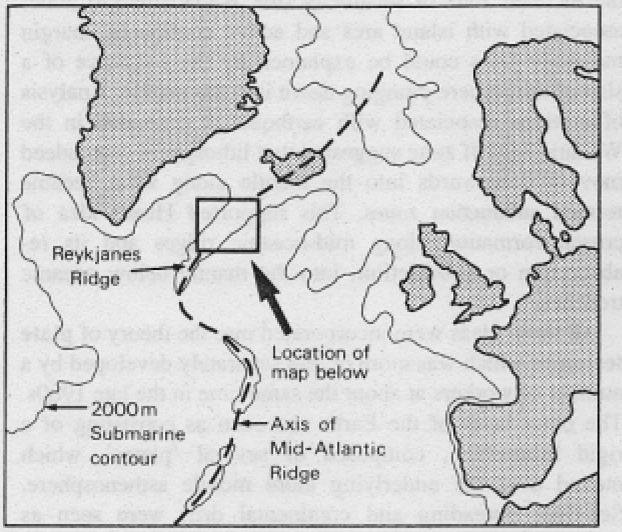
Fig. 2.11 Map of earthquake epicentres in the northern Japan–Kurile–Kamchatka region of the north-western Pacific. Earthquakes with a magnitude of 6 or more on the Richter scale are shown. The composite profile plots the depth of all the earthquakes on the map in relation to their distance from the oceanic trench. (After H. Benioff, 1954, Bulletin of the Geological Society of America *65, Fig. 2, p. 388.)*

2.3.4 Sea-floor spreading

The idea that the polarity of the Earth's magnetic field has reversed through geological time was proposed early this century by Brunhes in France and Matuyama in Japan. Measurements of remanent magnetism in a variety of volcanic rocks found on the continents had shown that some were magnetized in the opposite direction to the Earth's present magnetic field. Further data on these **geomagnetic reversals** were collected and by the mid-1960s a chronology of reversals had been established through the radiometric dating of a large number of samples from all over the world. The resulting detailed palaeomagnetic record showed that during the past 4 Ma the average duration of 'normal' fields (like the present one) had been about 420 ka and that of 'reversed' fields a little longer (about 480 ka). Periods of a particular polarity are termed **epochs**, with shorter phases of opposite magnetization within these being referred to as **events**.

Geophysical surveys during the late 1950s involving the measurement of the magnetic characteristics of the crust of the ocean floor indicated the presence of anomalous linear patterns (Fig. 2.12). These **magnetic anomalies** were found to be hundreds of kilometres long and typically 20–30 km across, and it initially seemed that they were caused by alternating bands of weakly and strongly magnetized rocks. No convincing explanation was initially produced to explain these features.

In 1962 H. H. Hess, of Princeton University in the USA, published a radical explanation for much of the geological data that had been accumulating over the previous decade or so. This proposed that mid-oceanic ridges represent regions where new oceanic crust is being generated by the upwelling of hot mantle material. He suggested that this new crust spreads laterally away from these ridges until it reaches an island arc or mountain belt where it descends into the mantle along the adjacent oceanic trench. This



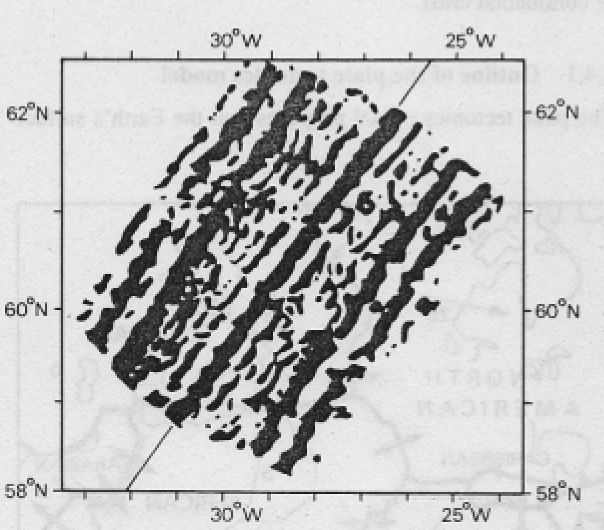


Fig. 2.12 Magnetic anomalies across the Reykjanes Ridge in the North Atlantic. This is a summary diagram of several traverses across the ridge, the axis of which is indicated by the diagonal line. Areas of positive anomalies are shown in black. (Magnetic anomaly map after J. R. Heirtzler et al. *(1966)* Deep Sea Research *13, Fig. 7, p. 435.)*

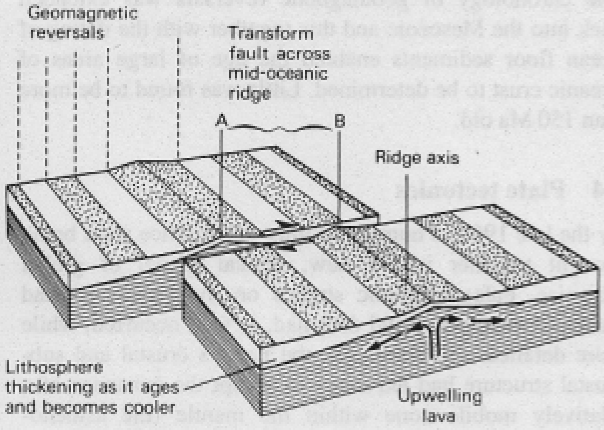


Fig. 2.13 Schematic illustration of the sea-floor spreading hypothesis of Hess as related to the record of palaeomagnetic reversals by Vine and Matthews. A displacement across the ridge representing a transform fault is also shown.

'conveyor belt' view of the oceanic crust, which owed something to the earlier ideas of Holmes, was termed **sea-floor spreading** by Dietz, another American geologist.

In 1963 Vine and Matthews, two Cambridge geophysicists, took up the idea of sea-floor spreading and linked it with data on the palaeomagnetic anomalies observed along mid-oceanic ridges (Fig. 2.12). They proposed that these anomalies, which seemed to be arranged symmetrically on either side of, and roughly parallel with, the ridge crest, were not a result of variations in the *intensity* of magnetization, as earlier suggested, but rather a consequence of the *direction* of magnetization. Positive anomalies represented underlying rocks with normal polarity, negative anomalies those with reversed polarity. As new crust emerged at mid-oceanic ridges, they suggested, it became magnetized with the polarity of the prevailing magnetic field (Fig. 2.13).

Subsequently the widths of the linear magnetic zones were related to the duration of each successive epoch of normal and reversed polarity on the assumption that the ocean floor spreads at a constant rate away from mid-oceanic ridges. The close relationship observed between the width of each magnetic anomaly and the corresponding time scale of magnetic reversals, which by now had been accurately established by the radiometric dating of terrestrial lavas, convinced many earth scientists of the reality of sea-floor spreading. Further corroborative evidence was provided by the palaeomagnetic record of ocean sediments containing particles of iron-rich minerals which lay on the basaltic oceanic crust.

When the patterns of magnetic anomalies across mid-oceanic ridges were mapped out it was noticed that in several cases the parallel bands of similarly magnetized crust were broken by numerous offsets or displacements which seemed to be aligned approximately at right angles to the axis of the ridge. Some of these displacements were very large, amounting to 1000 km or more in places. It appeared that these offsets marked some kind of lateral movement between adjacent sections of oceanic crust. In 1965 J. T. Wilson of the University of Toronto proposed that these were not simple strike-slip faults, as had been previously suggested, but were an entirely new variety, which he termed a **transform fault** (Fig. 2.13).

During the late 1960s considerable effort was invested in establishing the history of the ocean floor, much of this involving the activities of the Deep Sea Drilling Project.

The chronology of geomagnetic reversals was extended back into the Mesozoic and this together with the dating of ocean floor sediments enabled the age of large areas of oceanic crust to be determined. Little was found to be more than 150 Ma old.

2.4 Plate tectonics

By the late 1960s a number of lines of evidence were being brought together into a new, radical model of global tectonics. Palaeomagnetic studies on the continents had indicated that continental drift had, in fact, occurred, while more detailed information on the Earth's crustal and sub-crustal structure had led many to accept the existence of a relatively mobile zone within the mantle (the asthenosphere). Detailed mapping of the mid-oceanic ridges, together with magnetic and seismic surveys, had indicated that these were zones of crustal tension where new crust was being extruded from the mantle below. The history of polarity reversals of the Earth's magnetic field recorded in this newly generated crust as it cooled indicated that, once formed, it moved away from ridges towards oceanic trenches. On the basis of the much more complete seismic data available by the late 1960s it was being suggested that the inclined zone of seismicity (the Wadati-Benioff zone) associated with island arcs and active continental margin mountain belts could be explained by the existence of a slab of lithosphere plunging down into the mantle. Analysis of motions associated with earthquakes generated in the Wadati–Benioff zone suggested that lithosphere was indeed moving downwards into the mantle along what became termed **subduction zones**. This supported Hess's idea of crustal formation along mid-oceanic ridges and its re-absorption or **subduction**, into the mantle below oceanic trenches.

All these ideas were incorporated into the theory of **plate tectonics** which was more or less separately developed by a number of workers at about the same time in the late 1960s. The outer layer of the Earth was seen as consisting of a rigid lithosphere, composed of several 'plates', which moved over the underlying more mobile asthenosphere. Sea-floor spreading and continental drift were seen as involving these lithospheric **plates** rather than just oceanic or continental crust.

2.4.1 Outline of the plate tectonics model

The plate tectonics model proposes that the Earth's surface

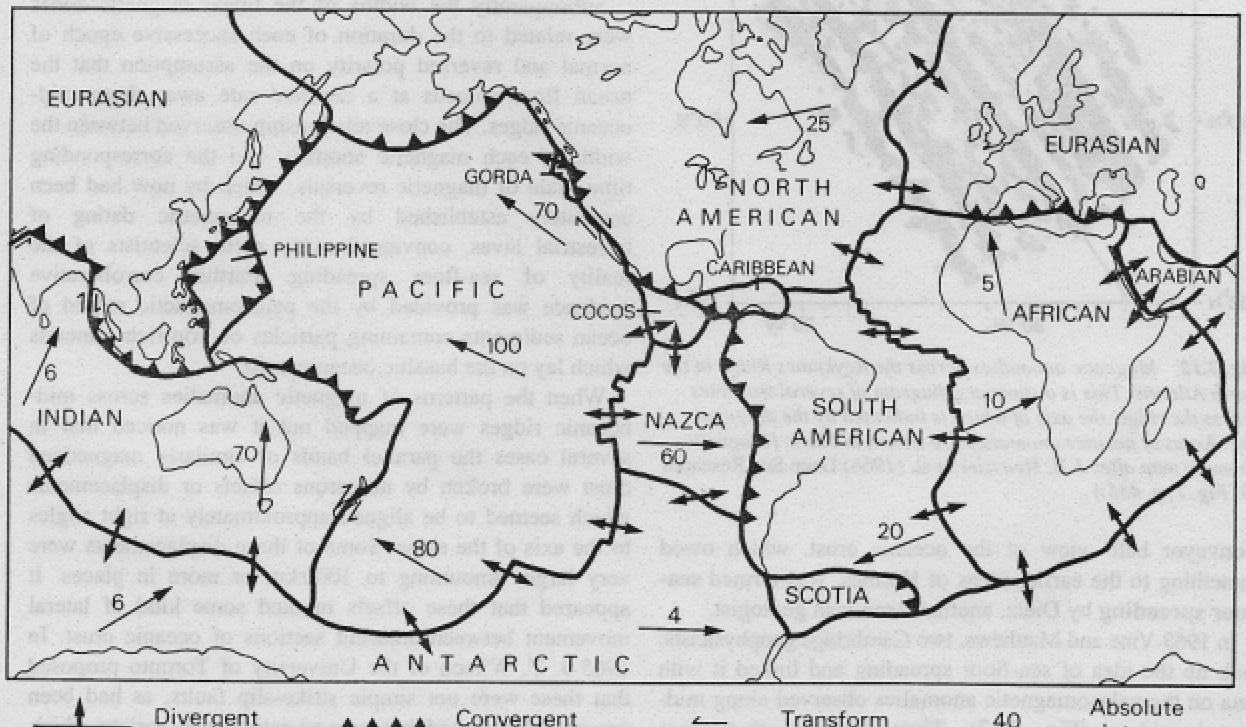


Fig. 2.14 Map of the major lithospheric plates. The various types of plate boundary are shown and the estimated current rates and directions of plate movement are indicated by arrows (rates in mm a^{-1}.)

comprises seven major, and at least a dozen minor, lithospheric plates composed of the crust and the upper, more rigid, part of the mantle (Fig. 2.14). These plates are constantly in motion with respect to one another and to the Earth's axis of rotation, and the motion of one plate influences the movement of others. Rates of movement range up to 100 mm a^{-1} (100 km Ma^{-1}) and average around 70 mm a^{-1} (70 km Ma^{-1}). Much, though not all, of the Earth's tectonic, volcanic and seismic activity is directly or indirectly associated with movement between neighbouring plates. The narrow zone marking the relative movement between two plates, which in most cases is fairly clearly demarcated by seismic activity, is termed a **plate boundary**, with the peripheral region adjacent to this boundary being referred to as a **plate margin**.

There are three types of plate boundary. At a **divergent boundary**, such as the Mid-Atlantic Ridge, new crust is formed and attached to the upper part of adjoining lithospheric plates while new upper mantle is accreted to the lower part. Plate movement is laterally away from the **spreading ridge**. At a **convergent boundary**, such as the western coast of South America, two plates are in motion towards each other, with one plate slipping down below the other along a subduction zone. The surface area of a plate is reduced at a subduction zone whereas it is increased along a spreading ridge. Along a **transform boundary**, two plates simply move laterally past each other along a transform fault without any major element of divergence or convergence. The most famous example of this kind of boundary is the San Andreas Fault System in California, USA. At some localities three plates may come into contact. Such a boundary is known as a **triple junction**, an example being the junction of the Pacific, Nazca and Cocos Plates (Fig. 2.14).

The upper layer of a plate is composed of either oceanic or continental crust or both. There is a fundamental difference between the behaviour of lithosphere capped with oceanic crust (oceanic lithosphere) and that covered with continental crust (continental lithosphere) since only oceanic lithosphere can be generated at mid-oceanic ridges and subducted at oceanic trenches.

Oceanic lithosphere has a mean density rather close to that of the immediately underlying asthenosphere. When newly formed along mid-oceanic ridges it is hot and thin and it probably has a slightly lower density. However, as oceanic lithosphere ages, cools and thickens it becomes more dense than the asthenosphere and rests upon it in an unstable state. Along zones of plate convergence, especially where these occur along the margins of continents (such as along the west coast of South America), buckling of the oceanic lithosphere causes it to founder and sink into the underlying asthenosphere, thereby forming a subduction zone. Alternatively, it seems possible that old, cold and thick oceanic lithosphere can attain a density sufficient for it to subside spontaneously into the asthenosphere. This may have occurred in the western Pacific Ocean where the age of the lithosphere presently being subducted indicates that cooling over a period of about 180–200 Ma is required for this process of spontaneous subduction to occur. By contrast continental lithosphere has a significantly lower mean density than oceanic lithosphere and its buoyancy with respect to the asthenosphere prevents all but limited subduction.

The two fundamental assumptions underlying the plate tectonics model are that the surface area of the Earth has not changed significantly with respect to the rate of generation of new oceanic crust, and that there is a lack of internal deformation within plates compared to the relative motion between them. If the Earth's surface area has increased significantly, sea-floor spreading could occur without associated plate subduction, the rate of crustal formation at mid-oceanic ridges matching the increase in surface area. There are, however, good reasons to believe that the Earth's radius has not increased significantly during the past 500 Ma or so.

Other evidence supports the idea of a lack of major deformation within plates, at least over plates composed of oceanic lithosphere. Oceanic crust has a layered structure which shows remarkably little disturbance and this would not be the case if significant deformation had occurred. In order for plates to behave in the manner proposed the lithosphere of which they are composed must be sufficiently rigid compared with the underlying asthenosphere for stress to be transmitted from one side to the other. If this were not the case significant deformation would occur within plates rather than being concentrated along plate margins. The movements of plates would be greatly complicated if they were to undergo significant internal deformation rather than acting as thin but rigid caps in motion around a sphere. It is quite clear, however, that marked deformation does in fact occur within some continental regions which are thousands of kilometres away from the nearest plate boundary.

In the remainder of this book we will be using plate tectonics as a framework for the interpretation of the Earth's large-scale topographic features. As will become apparent, however, there are several instances where it will be necessary to modify the basic model outlined in this chapter.

2.4.2 Classification of plate boundaries

In Table 2.3 each form of plate boundary is classified according to the type of interacting lithosphere involved. The most well-defined divergent boundaries are associated with mid-oceanic ridges, but they may also occur in an incipient form within continents as rift valleys comprising linear systems of faults associated with tension in the crust. Some rift valleys are analogous to mid-oceanic ridges in

Table 2.3 Classification and primary characteristics of plate boundary types

		MORPHOLOGICAL AND STRUCTURAL FEATURES		
Boundary type	Stress	Oceanic-oceanic lithosphere	Oceanic-continental lithosphere	Continental-continental lithosphere
Divergent	Tensional	Mid-oceanic ridge Volcanic activity	—	Rift valley Volcanoes
Convergent	Compressional	Oceanic trench and volcanic island arc	Oceanic trench and continental-margin mountain belt and volcanicity	—
		Complex island arc collision zone	Modified continental-margin mountain belt	Mountain belt, limited volcanic activity
Transform	Shear	Ridges and valleys normal to ridge axis	—	Fault zone, no volcanicity

experiencing shallow earthquakes and volcanicity.

Convergent boundaries between oceanic lithosphere are marked by an oceanic trench, a volcanic island arc and a Wadati–Benioff zone. Subduction of oceanic lithosphere beneath continental lithosphere is also associated with a trench and an inclined zone of seismicity, but in this case volcanic activity occurs within the active mountain belt located on the continental lithosphere fringing the trench. Where two continents collide neither experiences significant subduction but some crustal thickening occurs and a mountain belt is formed.

Transform boundaries, marked by transform faults, occur both as fracture zones along mid-oceanic ridges and as strike-slip fault zones within continental lithosphere. Plate movements can be oblique as well as simply divergent, convergent or transform. Oblique divergence is most commonly accommodated by transform offsets along a mid-oceanic ridge crest, while oblique convergence is resolved by the complex adjustment of lithospheric fragments along the plate boundary.

2.4.3 Plate motion

Plates vary greatly in area (Fig.2.14). The Pacific Plate, for instance, underlies almost the whole Pacific Ocean whereas a number of minor plates within the continents have an area of less than 1×10^6 km^2. Except where subduction zones lie adjacent to mountain belts on continental margins, plate boundaries do not coincide with continental coastlines. Many continental margins are not separated by subduction zones from the divergent boundaries marked by mid-oceanic ridges. For example, apart from its northern margin along the Mediterranean Sea, the African Plate is surrounded by divergent or transform boundaries. This means that it must be growing in area and that, consequently, the total area of all the other plates must be declining if the 'constant area' assumption of plate tectonics is to be satisfied. Moreover, since sea-floor spreading is taking place from both the Mid-Atlantic Ridge and the Carlsberg Ridge (in the Indian Ocean), one or both of these ridges must be experiencing movement with respect to the Earth's rotational axis.

If plates are regarded as rigid caps involved in relative motion around a sphere their movement can be analysed in terms of specific geometric controls. In Figure 2.15A we can imagine a segment of the Earth's surface (ABE) which has split into two, the two halves (ABC and ADE) remaining in contact only at point A. Point A is the **pole of rotation** about which movement occurs, and the separation of originally adjacent points on each segment occurs along small circles about this point. If we consider irregularly shaped areas on the sphere (X and Y), rather than the regular segments, these can be seen to move apart in such a way that their motion can similarly be described with respect to the same pole of rotation at A. The effect is clearly illustrated in the alignment of transform faults which mark the off-sets along either side of divergent plate boundaries (Fig. 2.15B). These represent segments of small circles concentric about the pole of rotation with respect to which the diverging plates are moving. A consequence of this is that the rate of spreading along a ridge will be a function of the sine of the angular distance from the pole of rotation; the greater the distance from the pole of rotation: the greater the rate of spreading (up to an angle of 90°). At convergent boundaries the rate of plate subduction is determined in a similar way.

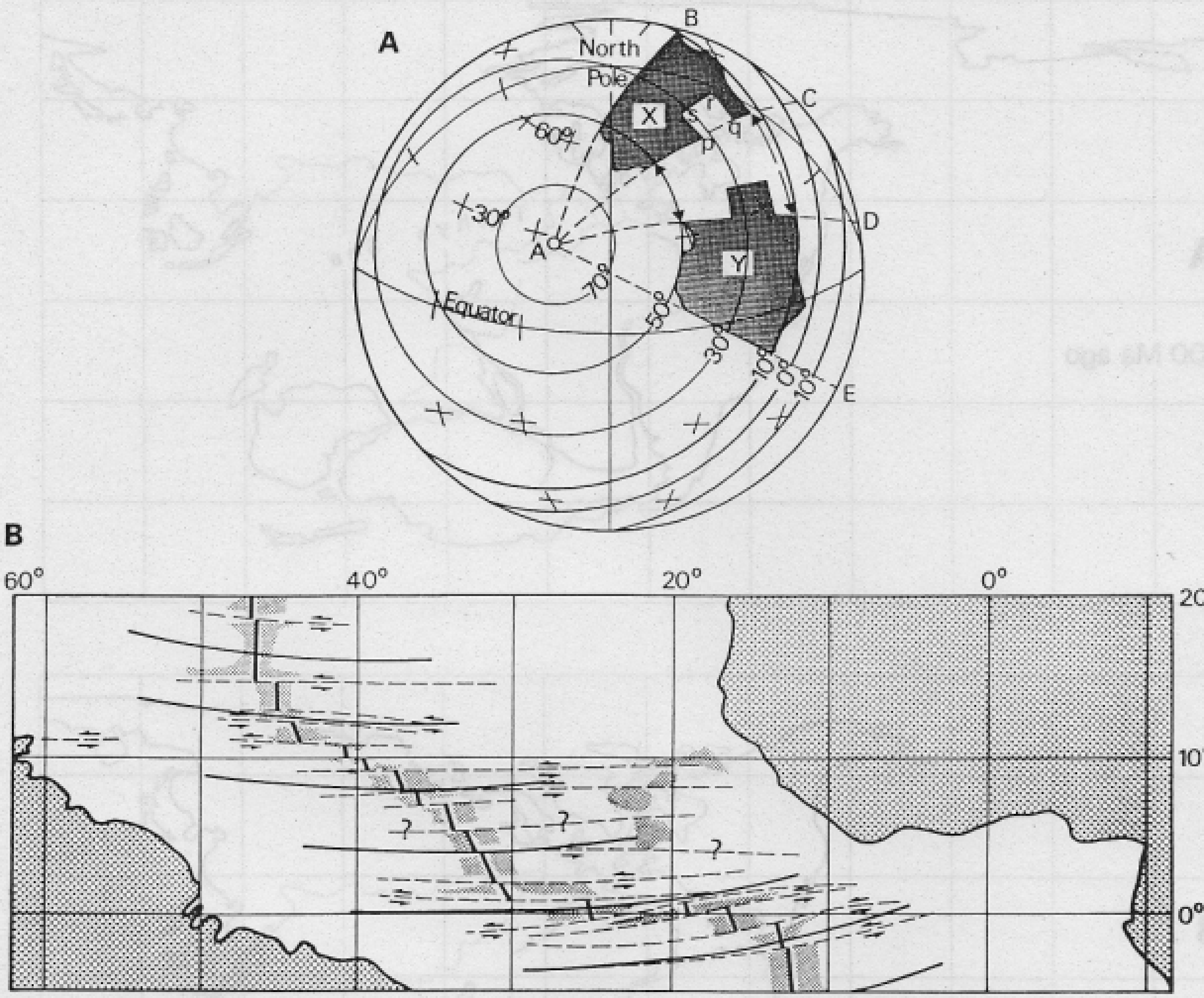
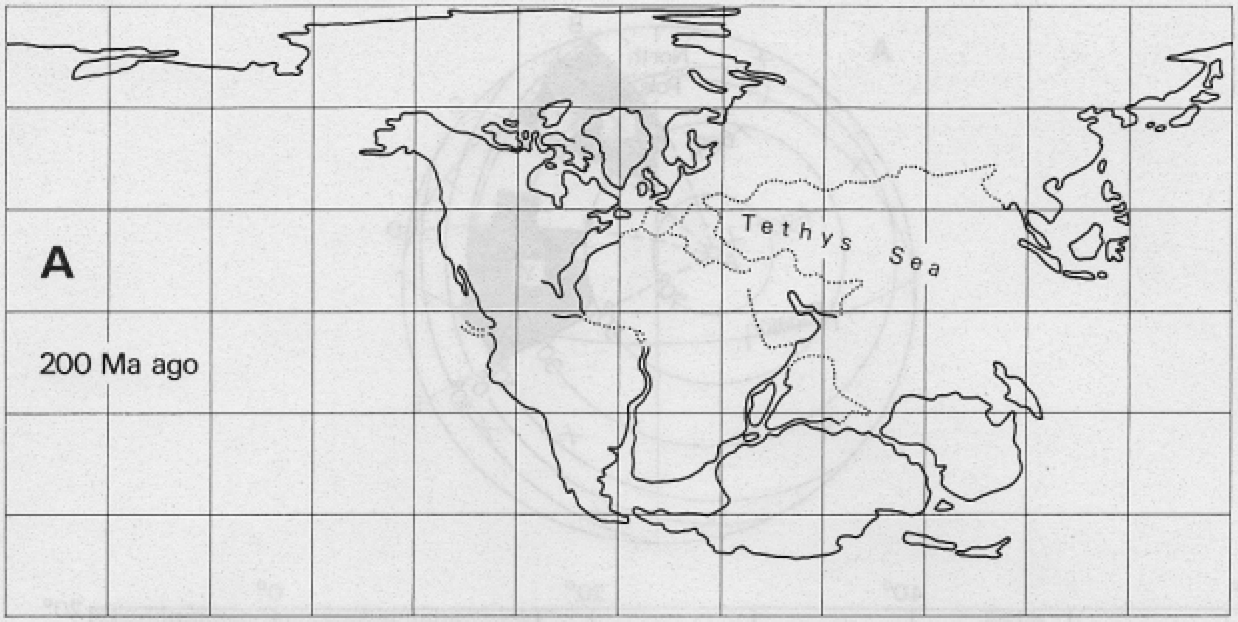


Fig. 2.15 The geometry of plate motion: (A) the movement of two segments of the Earth's surface with respect to a common pole of rotation; (B) the alignment of transform faults (dashed lines) in the equatorial Atlantic compared with small circles (solid lines) concentric about a pole located at 58°N, 36°W. (After E. R. Oxburgh (1971) in: I. G. Gass, P. J. Smith and R. C. L. Wilson (eds) Understanding the Earth. *Open University Press, Milton Keynes, Figs. 19.2 and 19.3, p. 266.)*

By comparing the alignments of transform faults, the poles of rotation of the major plates have now been established. Moreover, by relating this information to data on spreading rates, the directions of relative plate motions derived from the analysis of earthquakes, and palaeomagnetic evidence of continental palaeolatitudes, it has been possible to establish the direction and speed of present-day plate motions and what is probably a fairly accurate history of movements over the past 200 Ma or so (Fig. 2.16).

Since the mid-1980s a direct means of measuring present rates and directions of plate motion has become available. This technique, known as **very-long-baseline interferometry (VLBI)**, can determine with great accuracy the distance between two points on the Earth's surface by using radio telescopes. The measurement is derived from the very small difference in the time of arrival of signals received at different telescopes from a distant radio source (usually a quasar). The accuracy of this technique has been steadily improving and measurements can now be made over thousands of kilometres to an accuracy of a few millimetres. Preliminary results have confirmed that plates are moving in the directions, and at the rates, indicated by already acquired geophysical data. These results are important for plate tectonic theory because they show that the Earth is not currently expanding as has been suggested by a small minority of earth scientists, at least not at a rate which is significant compared with rates of plate motion. The 1990s will see further refinements to these measurements as the **Global Positioning System** becomes fully



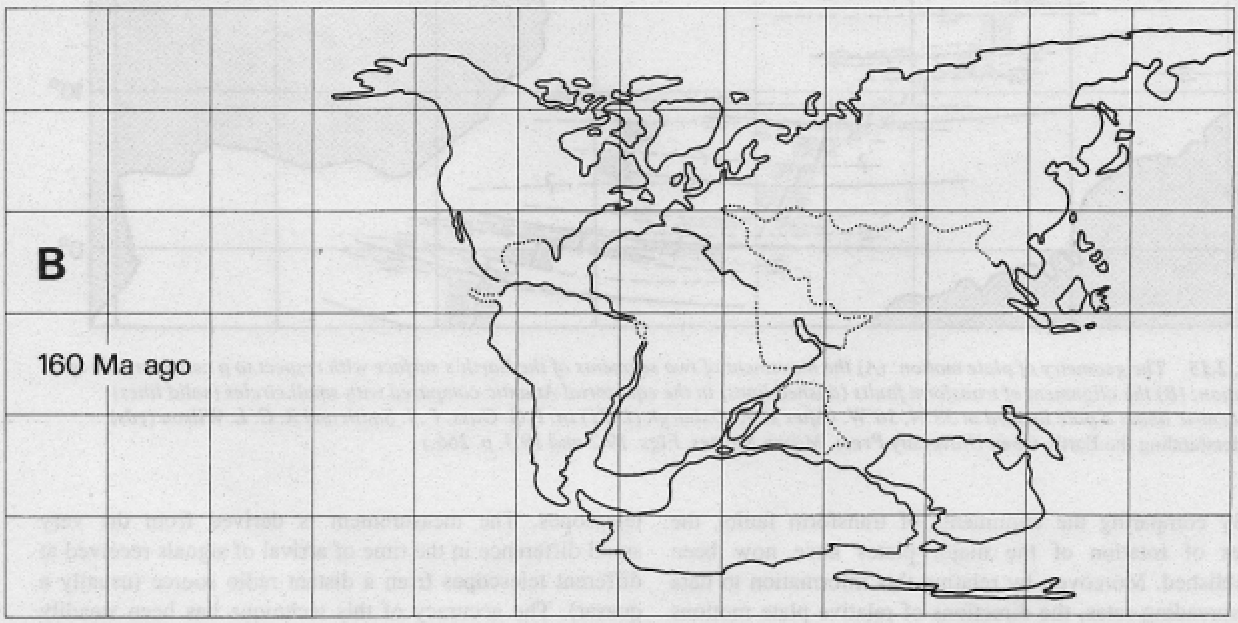
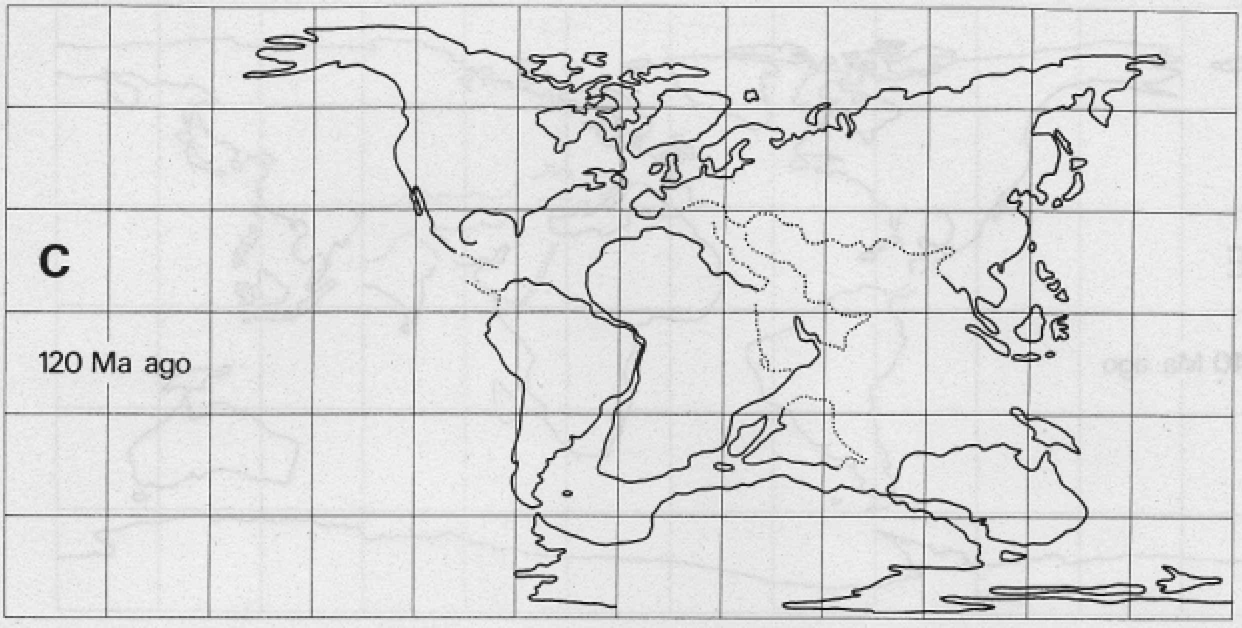
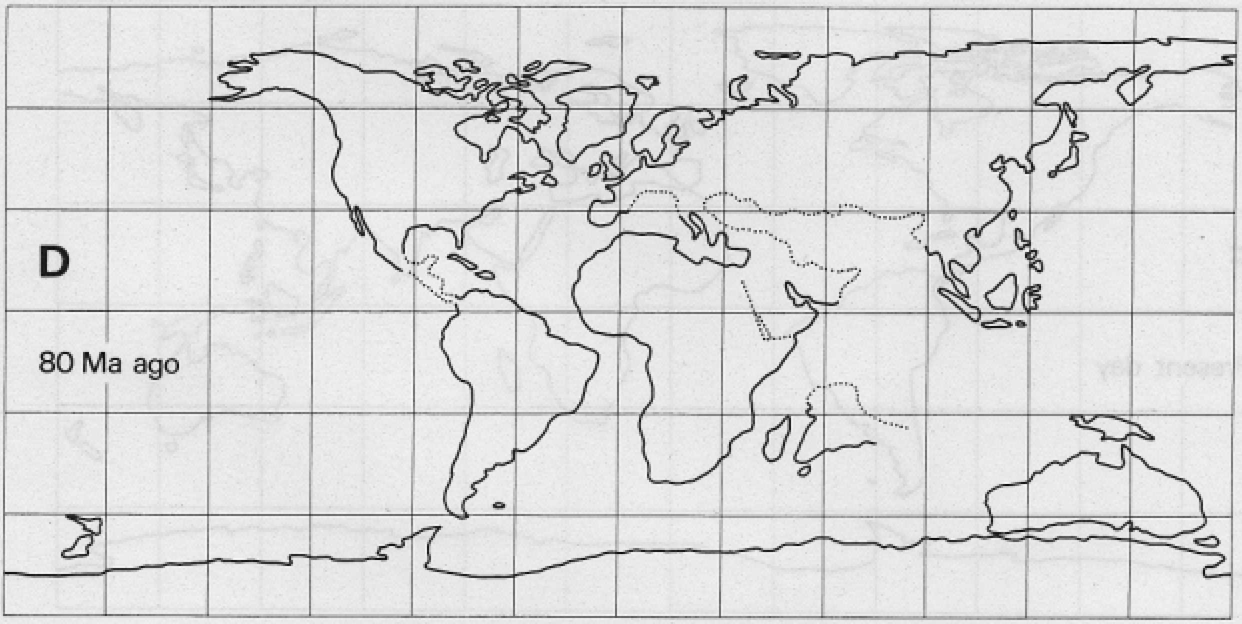


Fig. 2.16 The pattern of continental drift over the past 200 Ma. The six maps show the estimated past positions of the continents at 40 Ma intervals. Note the assemblage of the continents into one major landmass (Pangaea) 200 Ma ago with the Tethys Sea as a large embayment on its eastern flank (A). By 160 Ma BP this landmass had begun to rupture between North Africa and North America (B) and 40 Ma later the separation of Gondwana (the southern continent) and Laurasia (the northern continent) was complete (C). Over the past 120 Ma the Atlantic Ocean has opened and the various continents making up Gondwana have drifted apart (D–F). Particularly noticeable are the rapid northward movement of India and Australia away from Antarctica. For purposes of comparison all the maps have been produced on a cylindrical equidistant projection. Such a projection, however, grossly distorts areas and shapes in high latitudes. (Modified from A. G. Smith, A. M. Hurley and J. C. Briden (1981) Phanerozoic Paleocontinental World Maps. *Cambridge University Press, Cambridge, Maps 1 and 2, pp. 8–9, Maps 13 and 14, pp. 20–1, Maps 21 and 22, pp. 28–9, Maps 29 and 30, pp. 36–7, Maps 37 and 38, pp. 44–5 and Maps 45 and 46, pp. 52–3.)*





operational. This method of measuring distances between points on the Earth's surface is similar to VLBI but uses signals from Earth-orbiting satellites rather than extra-terrestrial radio sources.

2.4.4 Mechanisms of plate movement

In spite of the wide range of supporting evidence for the existence of plate tectonics the question of what causes plates to move has yet to be fully resolved. One idea is that lateral flow in the asthenosphere, arising from **convection** in the mantle, drags along the overlying lithosphere (Fig. 2.17(A–C). Convection currents rise and diverge below mid-oceanic ridges and converge and descend along subduction zones. Convection occurs when too much heat is present at depth to be conveyed upwards solely by

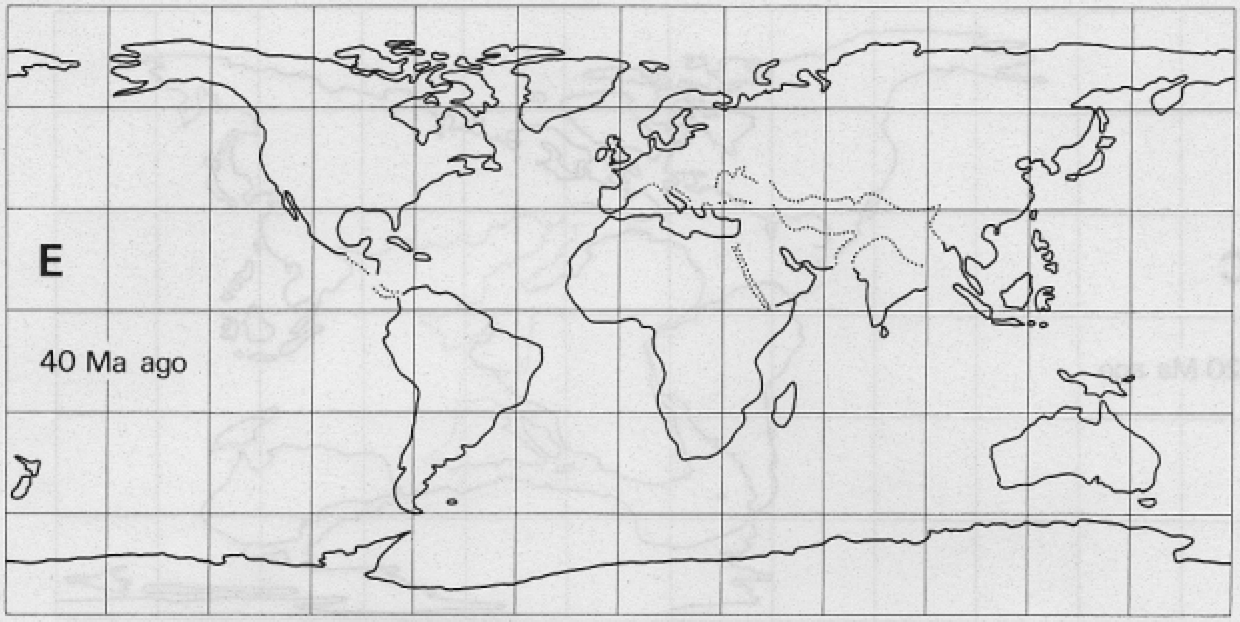


Fig. 2.16 *cont.*

thermal conduction. Temperatures increase and material begins to rise since it is less dense than the surrounding medium; cooler, denser material moves laterally at depth to take its place. If, however, the medium is too viscous convection will be inhibited, or even prevented altogether. Although our knowledge of the physical properties of the Earth's interior is at present far from complete, most geophysicists now support the idea that convection does in fact occur in the mantle; whether this convection is solely, or even primarily, responsible for plate movement is another matter.

Most of the dispute about the nature of convection within the Earth revolves around whether the circulation pattern involves all, or only part, of the mantle. On the basis of certain assumptions about the chemical heterogeneity of the mantle and a postulated marked increase in viscosity with

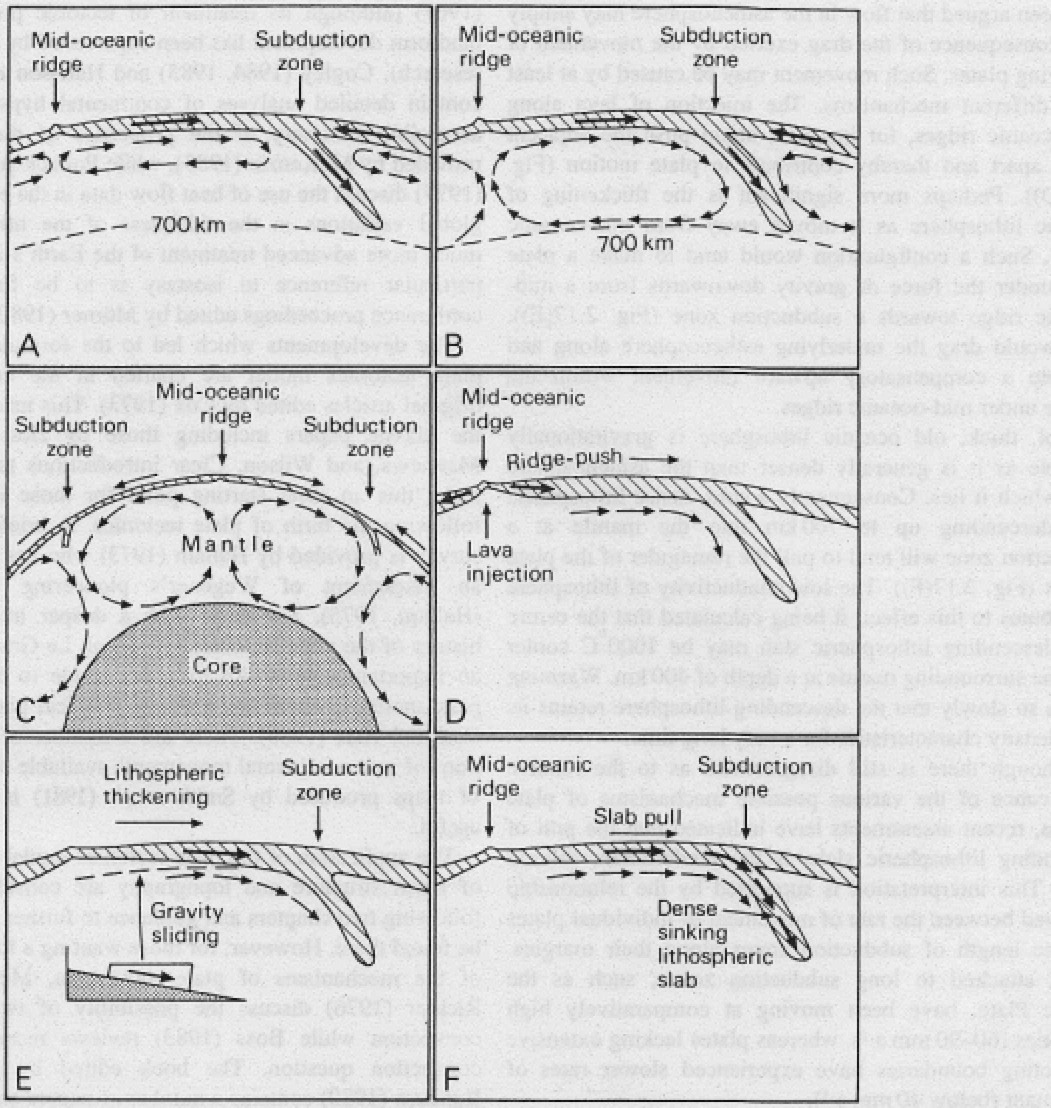


Fig. 2.17 Possible mechanisms of plate movement: (A) shallow convection confined to the asthenosphere (maximum depth about 300 km); (B) convection in the upper mantle extending to a depth of about 700 km; (C) whole mantle convection; (D) ridge push; (E) gravity sliding; (F) slab pull.

depth, it has been argued that convection would be confined to the asthenosphere, that is to a maximum depth of about 300 km (Fig. 2.17(A)). Some researchers, however, have suggested that 700 km may represent the lower limit of convection (Fig. 2.17(B)).

Recently the idea of **whole-mantle convection** has gained support, largely as a result of new estimates of the viscosity of the mantle below the asthenosphere. It has been suggested that a hot layer at the core–mantle boundary, generated by convection in the mobile outer core, produces a circulating system extending up through the entire mantle (Fig. 2.17(C)). An important point is that these large-scale convection cells fit in with the dimensions of plates. Most convection models predict that convection cells have similar horizontal and vertical dimensions, although arguments have been presented against this idea. As some plates have horizontal dimensions of thousands of kilometres it has been argued that convection cells of equivalent horizontal and vertical dimensions would be required to move them. An alternative view is that convection involves a two-scale pattern consisting of large-scale convection in the upper mantle responsible for plate movement superimposed on smaller-scale convection cells confined to the topmost 650 km of the mantle.

Although convection in the mantle has attracted considerable support as a likely cause of plate motion, it has also been argued that flow in the asthenosphere may simply be a consequence of the drag exerted by the movement of overlying plates. Such movement may be caused by at least three different mechanisms. The injection of lava along mid-oceanic ridges, for instance, could push the adjacent plates apart and thereby contribute to plate motion (Fig. 2.17(D)). Perhaps more significant is the thickening of oceanic lithosphere as it moves away from mid-oceanic ridges. Such a configuration would tend to make a plate slide under the force of gravity downwards from a mid-oceanic ridge towards a subduction zone (Fig. 2.17(E)). This would drag the underlying asthenosphere along and promote a compensatory upward movement within the mantle under mid-oceanic ridges.

Cool, thick, old oceanic lithosphere is gravitationally unstable as it is generally denser than the asthenosphere over which it lies. Consequently a cold, dense lithospheric slab descending up to 700 km into the mantle at a subduction zone will tend to pull the remainder of the plate with it (Fig. 2.17(F)). The low conductivity of lithosphere contributes to this effect, it being calculated that the centre of a descending lithospheric slab may be 1000°C cooler than the surrounding mantle at a depth of 400 km. Warming occurs so slowly that the descending lithosphere retains its high density characteristics for a very long time.

Although there is still disagreement as to the relative significance of the various possible mechanisms of plate motion, recent assessments have indicated that the pull of descending lithospheric slabs is the predominant driving force. This interpretation is supported by the relationship observed between the rate of movement of individual plates and the length of subduction zones along their margins. Those attached to long subduction zones, such as the Pacific Plate, have been moving at comparatively high velocities (60–90 mm a^{-1}), whereas plates lacking extensive subducting boundaries have experienced slower rates of movement (below 40 mm a^{-1}).

Further reading

Since the intention of this chapter has been to provide an introduction to global tectonics and the major elements of the Earth's morphology and structure, the suggested reading is mainly limited to texts which will be useful in providing more details on these topics. For those readers with little or no background in geology there are a number of excellent introductory texts. Two of the best are those by Press and Siever (1986) and Skinner and Porter (1987). A more advanced coverage of the Earth's internal structure is provided by Brown and Mussett (1981) while Turcotte and Schubert (1982) give a thorough quantitative grounding in tectonic processes.

A continent-by-continent survey of the Earth's major morphological features is to be found in the book by King (1967) (although its treatment of tectonic processes and landform development has been superseded by more recent research). Cogley (1984, 1985) and Harrison *et al.* (1983) contain detailed analyses of continental hypsometry. An accessible summary of the properties of the mantle is provided by McKenzie (1983), while Pollack and Chapman (1977) discuss the use of heat flow data in the estimation of global variations in the thickness of the lithosphere. A much more advanced treatment of the Earth's interior with particular reference to isostasy is to be found in the conference proceedings edited by Mörner (1980).

The developments which led to the formulation of the plate tectonics model are charted in the collection of original articles edited by Cox (1973). This incorporates all the classic papers including those by Hess, Vine and Matthews, and Wilson. Clear introductions to each topic make this an ideal starting point for those interested in following the birth of plate tectonics. A briefer historical survey is provided by Hallam (1973) who has also written an assessment of Wegener's pioneering contribution (Hallam, 1975). For those with a deeper interest in the history of the plate tectonics revolution Le Grand (1988) is an important source. An excellent guide to the study of plate motions, which has a strong practical bias, is that by Cox and Hart (1986). There are a number of reconstructions of past continental movements available but the series of maps produced by Smith *et al.* (1981) is particularly useful.

The application of the plate tectonics model to problems of Earth structure and topography are considered in the following two chapters and guidance to further reading will be found there. However, for those wanting a fuller account of the mechanisms of plate movement, McKenzie and Richter (1976) discuss the possibility of two scales of convection while Boss (1983) reviews research on the convection question. The book edited by Davies and Runcorn (1980) contains a number of papers on the various suggested mechanisms. Finally, Van Andel (1984) and Molnar (1988) present brief but thought-provoking assessments of the current status of plate tectonics theory and highlight the remaining problems of the model particularly when applied to the continents.

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