# 4

# LARGE-SCALE TECTONIC AND STRUCTURAL LANDFORMS

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Deep-seated geological processes and structures stamp their mark on many large landforms. This chapter looks at:

- Plate tectonic, diastrophic, and volcanic and plutonic processes
- How tectonic plates bear characteristic large-scale landforms at their active and passive margins and in their interiors
- The connections between tectonic geomorphology and large-scale landforms

## Splitting a continent

On 14 September 2005, a 4.7-magnitude earthquake in Dabbahu, 400 km north-east of Addis Ababa, Ethiopia, was followed by moderate tremors. Between 14 September and 4 October 2005, 163 earthquakes greater than magnitude 3.9 and a small volcanic eruption (on 26 September) occurred along the 60-km Dabbahu segment of the Afar rift (Figure 4.1). This volcano-seismic event marked a sudden sundering of the African and Arabian tectonic plates (Wright *et al.* 2006). It created an 8-m rift in just three weeks (Plate 4.1), a thin column of which filled up with magma forming a dyke between 2 and 9 km deep, with 2.5 km$^3$ of magma injection. The sudden rifting added to the long-term split that is currently tearing the north-east of Ethiopia and Eritrea from the rest of Africa and could eventually create a huge new sea. The earth movements of September 2005 are a small step in the creation of a new whole ocean that will take million of years to complete. However, this event is unparalleled in geological investigation and it has given geologists a rare opportunity to monitor the rupture process first-hand.

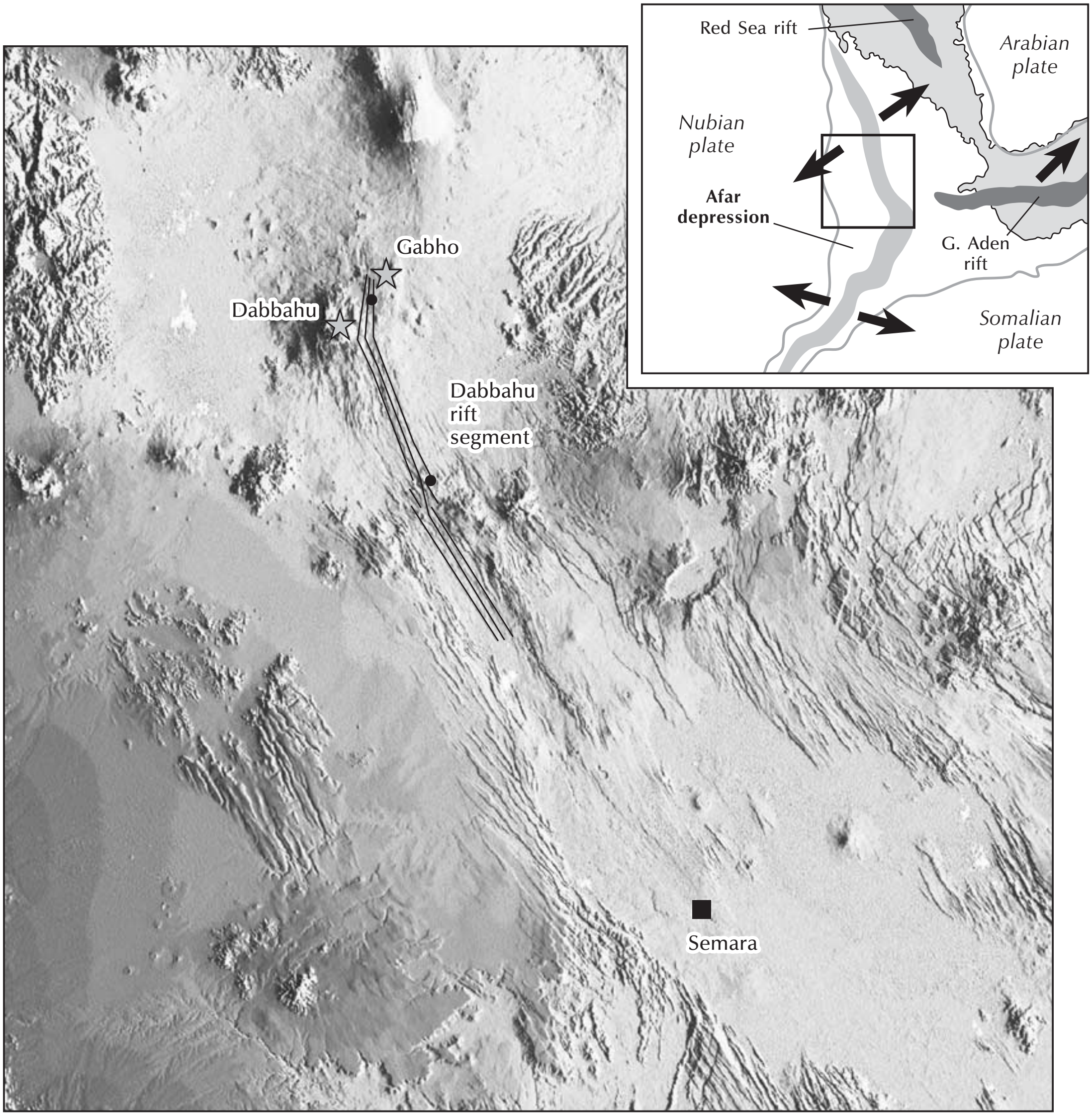


*Figure 4.1* Topographic relief of the 60-km-long Dabbahu rift segment within the Afar Depression. The inset shows directions of plate divergence between the stable African (Nubian), Arabian, and Somalian plates.
*Source:* Adapted from Cynthia Ebinger, Royal Holloway University of London.

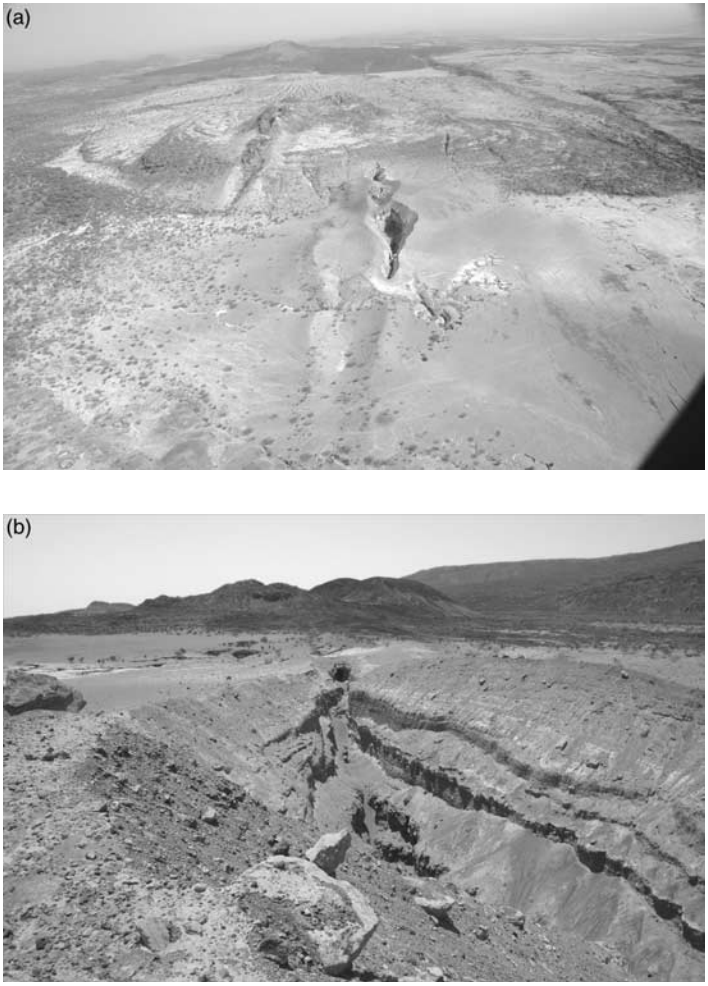


*Plate 4.1* The explosive volcanic vent that opened on 26 September 2005 after two days of nearly continuous seismic activity. (a) The 500-m-long, 60-m-wide vent looking north. To the right lies a 200-m-wide, 4-km-long zone of open fissures and normal faults that may mark the subsurface location of the dyke. *Photograph by Elizabeth Baker, Royal Holloway, University of London*. (b) View to the south from the north end of the vent. Notice the tunnel at the southern end. Notice the layers of ash that built up over a periods of days around the vent. The rhyolitic rocks in the foreground were blown out of the vent. (*Photograph by Julie Rowland, University of Auckland*)

The ascent of internal energy originating in the Earth's core impels a complicated set of geological processes. Deep-seated lithospheric, and ultimately baryspheric, processes and structures influence the shape and dynamics of the toposphere. The primary surface features of the globe are in very large measure the product of geological processes. This primary tectonic influence is manifest in the structure of mountain chains, volcanoes, island arcs, and other large-scale structures exposed at the Earth's surface, as well as in smaller features such as fault scarps.

Endogenic landforms may be tectonic or structural in origin (Twidale 1971, 1). **Tectonic landforms** are productions of the Earth's interior processes without the intervention of the forces of denudation. They include volcanic cones and craters, fault scarps, and mountain ranges. The influence of tectonic processes on landforms, particularly at continental and large regional scales, is the subject matter of **morphotectonics**. **Tectonic geomorphology** investigates the effects of active tectonic processes – faulting, tilting, folding, uplift, and subsidence – upon landforms. A recent and prolific development in geomorphology is the idea of '**tectonic predesign**'. Several landscape features, patently of exogenic origin, have tectonic or endogenic features stamped on them (or, literally speaking, stamped under them). Tectonic predesign arises from the tendency of erosion and other exogenic processes to follow stress patterns in the lithosphere (Hantke and Scheidegger 1999). The resulting landscape features are not fashioned directly by the stress fields. Rather, the exogenic processes act preferentially in conformity with the lithospheric stress (see p. 138). The conformity is either with the direction of a shear or, where there is a free surface, in the direction of a principal stress.

Few landforms are purely tectonic in origin: exogenous forces – weathering, gravity, running water, glaciers, waves, or wind – act on tectonic landforms, picking out less resistant rocks or lines of weakness, to produce **structural landforms**. An example is a volcanic plug, which is created when one part of a volcano is weathered and eroded more than another. A breached anticline is another example. Most textbooks on geomorphology abound with examples of structural landforms. Even in the Scottish Highlands, many present landscape features, which resulted from Tertiary etching, are closely adjusted to underlying rock types and structures (Hall 1991). Such passive influences of geological structures upon landforms are called **structural geomorphology**.

## PLATE TECTONICS AND VOLCANISM

The outer shell of the solid Earth – the **lithosphere** – is not a single, unbroken shell of rock; it is a set of snugly tailored **plates** (Figure 4.2). At present there are seven large plates, all with an area over 100 million km$^2$. They are the African, North American, South American, Antarctic, Australian–Indian, Eurasian, and Pacific plates. Two dozen or so smaller plates have areas in the range 1–10 million km$^2$. They include the Nazca, Cocos, Philippine, Caribbean, Arabian, Somali, Juan de Fuca, Caroline, Bismarck, and Scotia plates, and a host of **microplates** or **platelets**. In places, as along the western edge of the American continents, continental margins coincide with plate boundaries and are **active margins**. Where continental margins lie inside plates, they are **passive margins**. The break-up of Pangaea created many passive margins, including the east coast of South America and the west coast of Africa. Passive margins are sometimes designated rifted margins where plate motion has been divergent, and sheared margins where plate motion has been transformed, that is, where adjacent crustal blocks have moved in opposite directions. The distinction between active and passive margins is crucial to interpreting some large-scale features of the toposphere.

Earth's tectonic plates are continuously created at mid-ocean ridges and destroyed at subduction sites, and are ever on the move. Their motions explain virtually all tectonic forces that affect the lithosphere and thus the Earth's surface. Indeed, plate tectonics provides a good explanation for the primary topographic features of the Earth: the division between continents and oceans, the disposition of mountain ranges, and the placement of sedimentary basins at plate boundaries.

### Plate tectonic processes

The plate tectonic model currently explains changes in the Earth's crust. This model is thought satisfactorily to explain geological structures, the distribution and

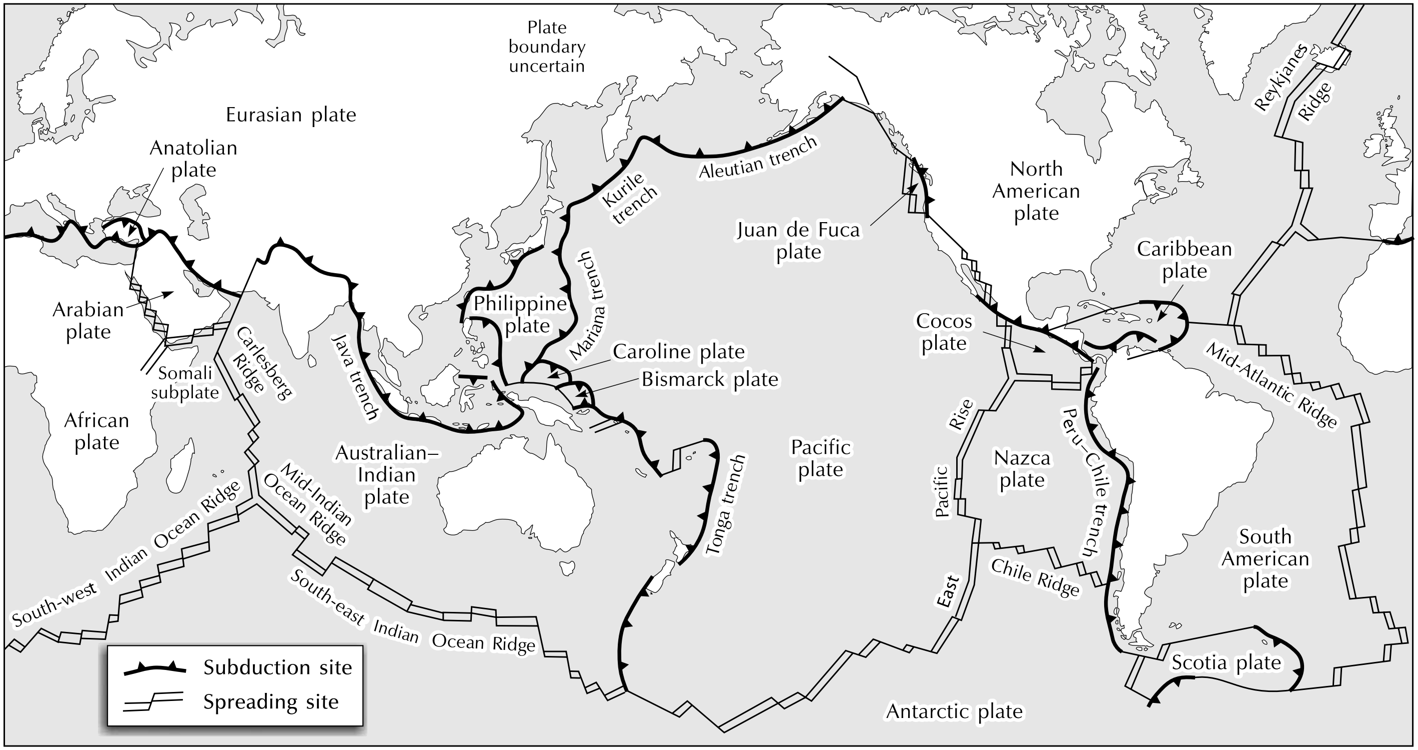


*Figure 4.2* Tectonic plates, spreading sites, and subduction sites.
*Source:* Adapted from Ollier (1996)

variation of igneous and metamorphic activity, and sedimentary facies. In fact, it explains all major aspects of the Earth's long-term tectonic evolution (e.g. Kearey and Vine 1990). The plate tectonic model comprises two tectonic 'styles'. The first involves the oceanic plates and the second involves the continental plates.

## Oceanic plate tectonics

The **oceanic plates** are linked into the cooling and recycling system comprising the mesosphere, asthenosphere, and lithosphere beneath the ocean floors. The chief cooling mechanism is subduction. New oceanic lithosphere is formed by volcanic eruptions along mid-ocean ridges. The newly formed material moves away from the ridges. In doing so, it cools, contracts, and thickens. Eventually, the oceanic lithosphere becomes denser than the underlying mantle and sinks. The sinking takes place along **subduction zones**. These are associated with earthquakes and volcanicity. Cold oceanic slabs may sink well into the mesosphere, perhaps as much as 670 km or below the surface. Indeed, subducted material may accumulate to form 'lithospheric graveyards' (Engebretson *et al.* 1992).

It is uncertain why plates should move. Several **driving mechanisms** are plausible. Basaltic lava upwelling at a mid-ocean ridge may push adjacent lithospheric plates to either side. Or, as elevation tends to decrease and slab thickness to increase away from construction sites, the plate may move by gravity sliding. Another possibility, currently thought to be the primary driving mechanism, is that the cold, sinking slab at subduction sites pulls the rest of the plate behind it. In this scenario, mid-ocean ridges stem from passive spreading – the oceanic lithosphere is stretched and thinned by the tectonic pull of older and denser lithosphere sinking into the mantle at a subduction site; this would explain why sea-floor tends to spread more rapidly in plates attached to long subduction zones. As well as these three mechanisms, or perhaps instead of them, mantle convection may be the number one motive force, though this now seems unlikely as many spreading sites do not sit over upwelling mantle convection cells. If the mantle-convection model were correct, mid-ocean ridges should display a consistent pattern of gravity anomalies, which they do not, and would probably not develop fractures (transform faults). But, although convection is perhaps not the master driver of plate motions, it does occur. There is some disagreement about the depth of the convective cell. It could be confined to the asthenosphere, the upper mantle, or the entire mantle (upper and lower). Whole mantle convection (Davies 1977, 1992) has gained much support, although it now seems that whole mantle convection and a shallower circulation may both operate.

The lithosphere may be regarded as the cool surface layer of the **Earth's convective system** (Park 1988, 5). As part of a convective system, it cannot be considered in isolation (Figure 4.3). It gains material from the asthenosphere, which in turn is fed by uprising material from the underlying mesosphere, at constructive plate boundaries. It migrates laterally from mid-ocean ridge axes as cool, relatively rigid, rock. Then, at destructive plate boundaries, it loses material to the asthenosphere and mesosphere. The fate of the subducted material is not clear. It meets with resistance in penetrating the lower mantle, but is driven on by its thermal inertia and continues to sink, though more slowly than in the upper mantle, causing accumulations of slab material (Fukao *et al.* 1994). Some slab material may eventually be recycled to create new lithosphere. However, the basalt erupted at mid-ocean ridges shows a few signs of being new material that has not passed through a rock cycle before (Francis 1993, 49). First, it has a remarkably consistent composition, which is difficult to account for by recycling. Second, it emits gases, such as helium, that seem to be arriving at the surface for the first time. Equally, it is not 'primitive' and formed in a single step by melting of mantle materials – its manufacture requires several stages. It is worth noting that the transformation of rock from mesosphere, through the asthenosphere, to the lithosphere chiefly entails temperature and viscosity (rheidity) changes. Material changes do occur: partial melting in the asthenosphere generates magmas that rise into the lithosphere, and volatiles enter and leave the system.

## Continental plate tectonics

The **continental lithosphere** does not take part in the mantle-convection process. It is 150 km thick and

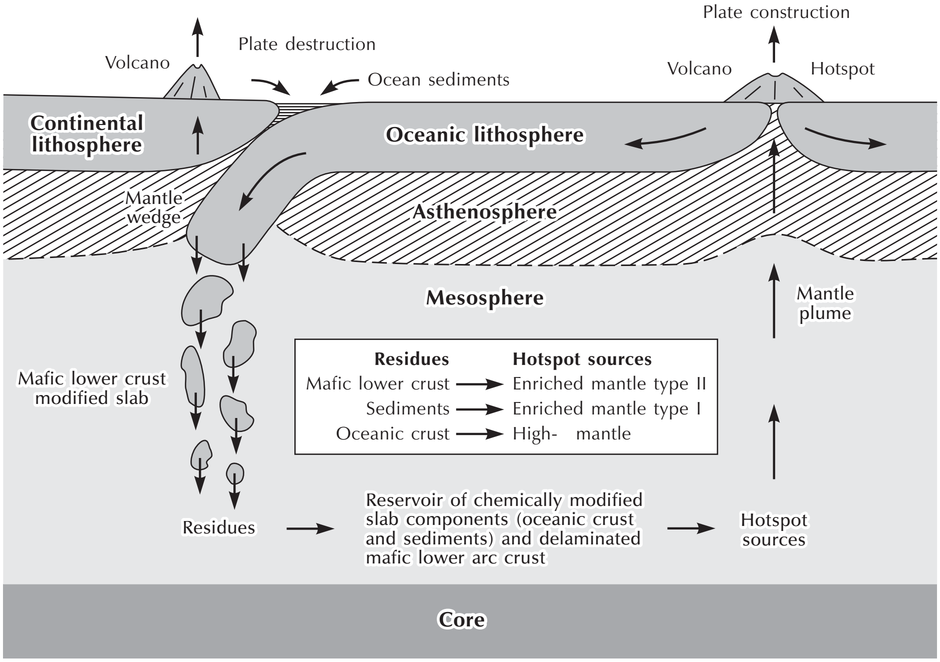


*Figure 4.3* Interactions between the asthenosphere, lithosphere, and mesosphere. The oceanic lithosphere gains material from the mesosphere (via the asthenosphere) at constructive plate boundaries and hotspots and loses material to the mesosphere at destructive plate boundaries. Subduction feeds slab material (oceanic sediments derived from the denudation of continents and oceanic crust), mantle lithosphere, and mantle wedge materials to the deep mantle. These materials undergo chemical alteration and accumulate in the deep mantle until mantle plumes bear them to the surface where they form new oceanic lithosphere.
*Source:* Adapted from Tatsumi (2005)

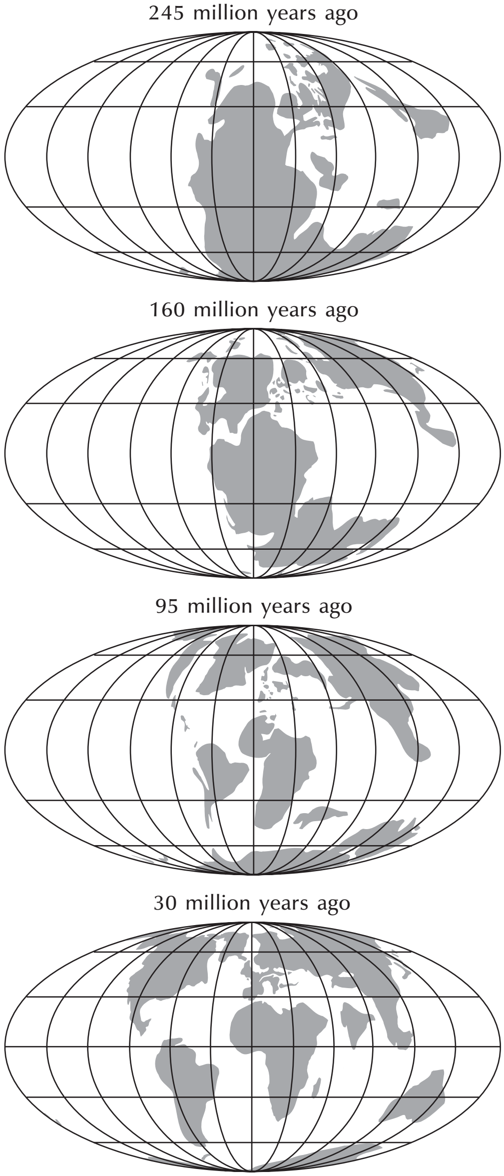
consists of buoyant low-density crust (the tectosphere) and relatively buoyant upper mantle. It therefore floats on the underlying asthenosphere. Continents break up and reassemble, but they remain floating at the surface. They move in response to lateral mantle movements, gliding serenely over the Earth's surface. In breaking up, small fragments of continent sometimes shear off; these are called **terranes**. They drift around until they meet another continent, to which they become attached (rather than being subducted) or possibly are sheared along it. As they may come from a different continent from the one they are attached to, they are called **exotic** or **suspect terranes** (p. 113). Most of the western seaboard of North America appears to consist of these exotic terranes. In moving, continents have a tendency to drift away from mantle hot zones, some of which they may have produced: stationary continents insulate the underlying mantle, causing it to warm. This warming may eventually lead to a large continent breaking into several smaller ones. Most continents are now sitting on, or moving towards, cold parts of the mantle. An exception is Africa, which was the core of Pangaea. Continental drift leads to collisions between continental blocks and to the overriding of oceanic lithosphere by continental lithosphere along subduction zones.

Continents are affected by, and affect, underlying mantle and adjacent plates. They are maintained against erosion (rejuvenated in a sense) by the welding of sedimentary prisms to continental margins through metamorphism, by the stacking of thrust sheets, by the sweeping up of microcontinents and island arcs at their leading edges, and by the addition of magma through intrusions and extrusions (Condie 1989). Geologists have established the relative movement of continents over the Phanerozoic aeon with a high degree of confidence, although pre-Pangaean reconstructions are less reliable than post-Pangaean reconstructions. Figure 4.4 charts the probable break-up of Pangaea.

## Diastrophic processes

Traditionally, tectonic (or geotectonic) forces are divided into two groups: (1) diastrophic forces and (2) volcanic and plutonic forces. **Diastrophic forces** lead to the folding, faulting, uplift, and subsidence of the lithosphere. **Volcanic forces** lead to the extrusion of magma on to the Earth's surface as lava and to minor intrusions (e.g. dykes and sills) into other rocks. **Plutonic forces**, which originate deep in the Earth, produce major intrusions (plutons) and associated veins.

Diastrophic forces may deform the lithosphere through folding, faulting, uplift, and subsidence. They are responsible for some of the major features of the physical toposphere. Two categories of **diastrophism** are recognized: orogeny and epeirogeny, but these terms are a source of much confusion (Ollier and Pain 2000, 4–8). **Orogeny** literally means the genesis of mountains, and when first used it meant just that. Later, it became associated with the idea of folding, and eventually it came to mean the folding of rocks in fold belts. As mountain building is not associated with the folding of rocks, it cannot be synonymous with orogeny (Ollier 2003). **Epeirogeny** is the upheaval or depression of large areas of cratons without significant folding or fracture. The only folding associated with epeirogeny is the broadest of undulations. Epeirogeny includes **isostatic movements**, such as the rebound of land after an ice sheet has melted, and cymatogeny, which is the arching, and sometimes doming, of rocks with little deformation over 10–1,000 km. Some geomorphologists believe that mountains result from the erosion of areas uplifted epeirogenically (e.g. Ollier and Pain 2000, 8; Ollier 2003; see Huggett 2006, 29–30).



245 million years ago

160 million years ago

95 million years ago

30 million years ago

*Figure 4.4* Changing arrangement of continents over the last 245 million years, showing the break-up of Pangaea, during the Early Triassic period; during the Callovian age (Middle Jurassic); during the Cenomanian age (Late Cretaceous); and during the Oligocene epoch. All maps use Mollweide's equal-area projection.
*Source:* Adapted from maps in Smith *et al.* (1994)

The relative motion of adjacent plates primarily creates the many tectonic forces in the lithosphere. Indeed, relative plate motions underlie almost all surface tectonic processes. Plate boundaries are particularly important for understanding geotectonics. They are sites of strain and associated with faulting, earthquakes, and, in some instances, mountain building (Figure 4.5). Most boundaries sit between two adjacent plates, but, in places, three plates come into contact. This happens where the North American, South American, and Eurasian plates meet (Figure 4.2). Such Y-shaped boundaries are known as **triple junctions**. Three plate-boundary types produce distinctive tectonic regimes:

1 **Divergent plate boundaries** at construction sites, which lie along mid-ocean ridges, are associated with divergent tectonic regimes involving shallow, low-magnitude earthquakes. The ridge height depends primarily on the spreading rate. Incipient divergence occurs within continents, including Africa, and creates rift valleys, which are linear fault systems and, like mid-ocean ridges, are prone to shallow earthquakes and volcanism (p. 143). Volcanoes at divergent boundaries produce basalt.

2 **Convergent plate boundaries** vary according to the nature of the converging plates. Convergent tectonic regimes are equally varied; they normally lead to partial melting and the production of granite and the eruption of andesite and rhyolite. An oceanic trench, a volcanic island arc, and a dipping planar region of seismic activity (a Benioff zone) with earthquakes of varying magnitude mark a collision between two slabs of oceanic lithosphere. An example is the Scotia arc, lying at the junctions of the Scotia and South American plates. Subduction of oceanic lithosphere beneath continental lithosphere produces two chief features. First, it forms an oceanic trench, a dipping zone of seismic activity, and volcanicity in an **orogenic mountain belt** (or **orogen**) lying on the continental lithosphere next to the oceanic trench (as in western South America). Second, it creates **intra-oceanic arcs** of volcanic islands (as in parts of the western Pacific Ocean). In a few cases of continent–ocean collision, a slab of ocean floor has overridden rather than underridden the continent. This process, called **obduction**, has produced the Troödos Mountain region of Cyprus. Collisions of continental lithosphere result in crustal thickening and the production of a mountain belt, but little subduction. A fine example is the Himalaya, produced by India's colliding with Asia. Divergence and convergence may occur obliquely. Oblique divergence is normally accommodated by transform offsets along a mid-oceanic ridge crest, and oblique convergence by the complex microplate adjustments along plate boundaries. An example is found in the Betic cordillera, Spain, where the African and Iberian plates slipped by one another from the Jurassic to Tertiary periods.

3 **Conservative** or **transform plate boundaries** occur where adjoining plates move sideways past each other along a transform fault without any convergent or divergent motion. They are associated with strike-slip tectonic regimes and with shallow earthquakes of variable magnitude. They occur as fracture zones along mid-ocean ridges and as **strike-slip fault zones** within continental lithosphere. A prime example of the latter is the San Andreas fault system in California.

Tectonic activity also occurs within lithospheric plates, and not just at plate edges. This is called **within-plate tectonics** to distinguish it from plate-boundary tectonics.

## Volcanic and plutonic processes

Volcanic forces are either intrusive or extrusive forces. **Intrusive forces** are found within the lithosphere and produce such features as batholiths, dykes, and sills. The deep-seated, major intrusions – batholiths and stocks – result from plutonic processes, while the minor, nearer-surface intrusions such as dykes and sills, which occur as independent bodies or as offshoots from plutonic intrusions, result from hypabyssal processes. **Extrusive forces** occur at the very top of the lithosphere and lead to exhalations, eruptions, and explosions of materials through volcanic vents, all of which are the result of volcanic processes.

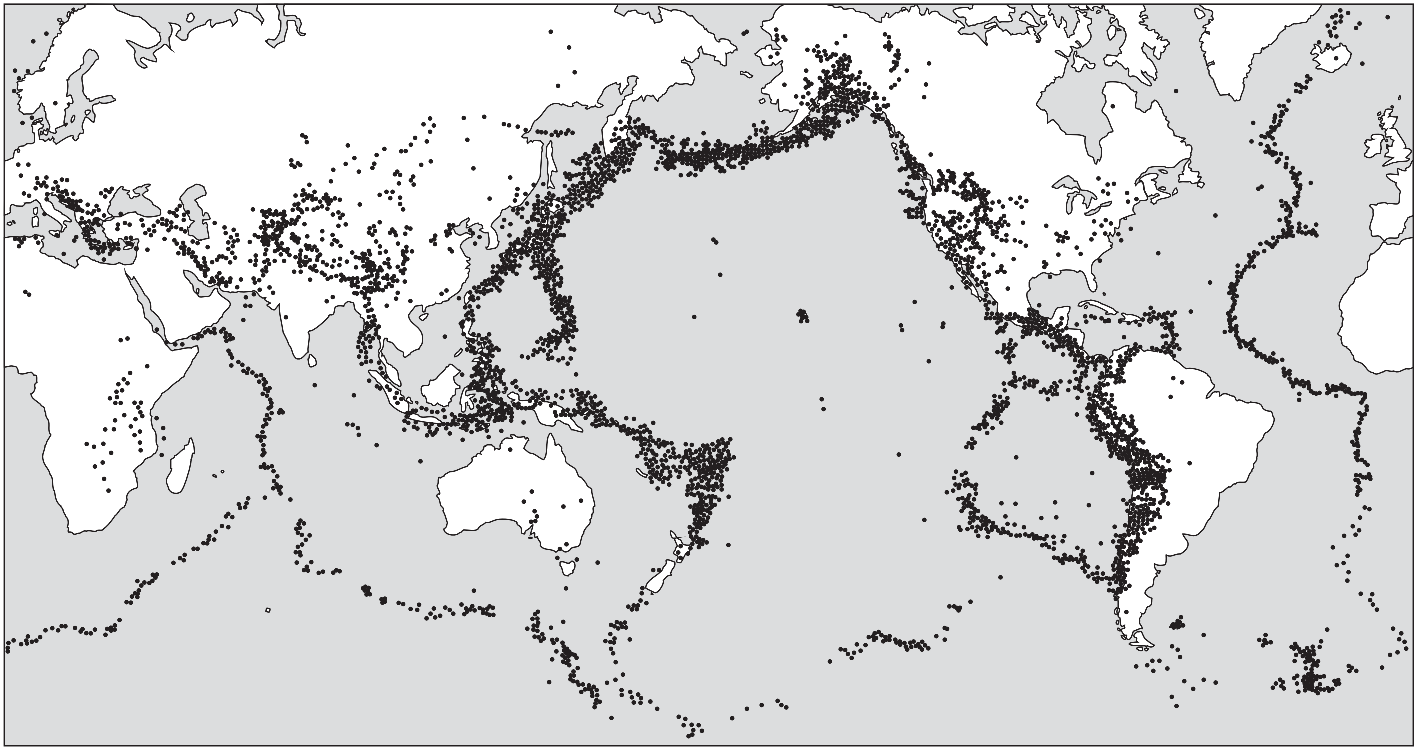


*Figure 4.5* Global distribution of earthquakes.
*Source:* Adapted from Ollier (1996)

## The location of volcanoes

Most volcanoes sit at plate boundaries. A few, including the Cape Verde volcano group in the southern Atlantic Ocean and the Tibesti Mountains in Saharan Africa, occur within plates. These **'hot-spot' volcanoes** are surface expressions of thermal mantle plumes. Hot-spots are characterized by topographic bumps (typically 500–1,200 m high and 1,000–1,500 km wide), volcanoes, high gravity anomalies, and high heat flow. Commonly, a mantle plume stays in the same position while a plate slowly slips over it. In the ocean, this produces a chain of volcanic islands, or a **hot-spot trace**, as in the Hawaiian Islands. On continents, it produces a string of volcanoes. Such a volcanic string is found in the Snake River Plain province of North America, where a hot-spot currently sitting below Yellowstone National Park, Wyoming, has created an 80-km-wide band across 450 km of continental crust, producing prodigious quantities of basalt in the process. Even more voluminous are **continental flood basalts**. These occupy large tracts of land in far-flung places. The Siberian province covers more than 340,000 km$^2$. India's Deccan Traps once covered about 1,500,000 km$^2$; erosion has left about 500,000 km$^2$.

## Mantle plumes

**Mantle plumes** appear to play a major role in plate tectonics. They may start growing at the core–mantle boundary, but the mechanisms by which they form and grow are undecided. They may involve rising plumes of liquid metal and light elements pumping latent heat outwards from the inner-core boundary by compositional convection, the outer core then supplying heat to the core–mantle boundary, whence giant silicate magma chambers pump it into the mantle, so providing a plume source. Mantle plumes may be hundreds of kilometres in diameter and rise towards the Earth's surface. A plume consists of a leading 'glob' of hot material that is followed by a 'stalk'. On approaching the lithosphere, the plume head is forced to mushroom beneath the lithosphere, spreading sideways and downwards a little. The plume temperature is 250–300°C hotter than the surrounding upper mantle, so that 10–20 per cent of the surrounding rock is melted. This melted rock may then run on to the Earth's surface as flood basalt, as occurred in India during the Cretaceous period when the Deccan Traps were formed.

**Superplumes** may form. One appears to have done so beneath the Pacific Ocean during the middle of the Cretaceous period (Larson 1991). It rose rapidly from the core–mantle boundary about 125 million years ago. Production tailed off by 80 million years ago, but it did not stop until 50 million years later. It is possible that superplumes are caused by cold, subducted oceanic crust on both edges of a tectonic plate accumulating at the top of the lower mantle. These two cold pools of rock then sink to the hot layer just above the core, and a giant plume is squeezed out between them. **Plume tectonics** may be the dominant style of convection in the major part of the mantle. Two super-upwellings (the South Pacific and African superplumes) and one super-downwelling (the Asian cold plume) appear to prevail (Figure 4.6).

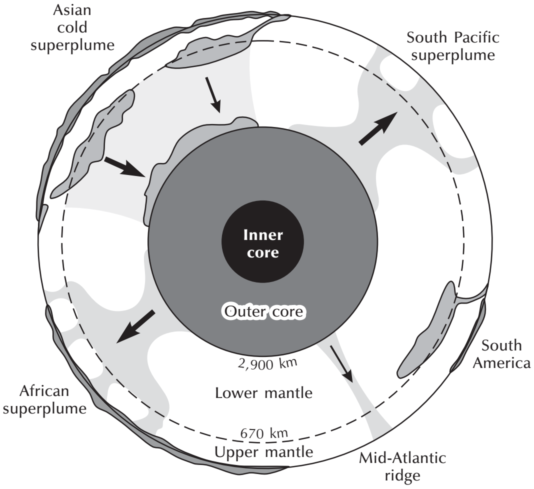


*Figure 4.6* A possible grand circulation of Earth materials. Oceanic lithosphere, created at mid-ocean ridges, is subducted into the deeper mantle. It stagnates at around 670 km and accumulates for 100–400 million years. Eventually, gravitational collapse forms a cold downwelling on to the outer core, as in the Asian cold superplume, which leads to mantle upwelling occurring elsewhere, as in the South Pacific and African hot superplumes.
*Source:* Adapted from Fukao *et al.* (1994)

It should be mentioned that a minority of geologists have always spoken out against plumes. However, since about the turn of the millennium the number of voices has swollen, and the validity of the plume model has emerged as a key debate in Earth science (see Foulger *et al.* 2005; Huggett 2006, 21–5).

## LANDFORMS RELATED TO TECTONIC PLATES

Tectonic processes primarily determine large-scale landforms, though water, wind, and ice partly shape their detailed surface form. Geomorphologists classify large-scale landforms in many ways. One scheme rests on crustal types: continental shields, continental platforms, rift systems, and orogenic belts. It is convenient to discuss these large units under three headings – plate interiors, passive plate margins, and active plate margins.

### Plate-interior landforms

**Cratons** are the broad, central parts of continents. They are somewhat stable continental shield areas with a basement of Precambrian rocks that are largely unaffected by orogenic forces but are subject to epeirogeny. The main large-scale landforms associated with these areas are basins, plateaux (upwarps and swells), rift valleys, and intracontinental volcanoes. Equally important landforms lie along passive continental margins, that is, margins of continents created when formerly single landmasses split in two, as happened to Africa and South America when the supercontinent Pangaea broke apart.

**Intra-cratonic basins** may be 1,000 km or more across. Some, such as the Lake Eyre basin of Australia and the Chad and Kalahari basins of Africa, are enclosed and internally drained. Others, such as the region drained by the Congo river systems, are breached by one or more major rivers.

Some continents, and particularly Africa, possess extensive **plateaux** sitting well above the average height of continental platforms. The Ahaggar Plateau and Tibesti Plateau in North Africa are examples. These plateaux appear to have been uplifted without rifting occurring but with some volcanic activity.

**Continental rifting** occurs at sites where the continental crust is stretched and faulted. The rift valley running north to south along much of East Africa is probably the most famous example (p. 98), and its formation is linked with domal uplift. Volcanic activity is often associated with continental rifting. It is also associated with hot-spots.

### Passive-margin landforms

Figure 4.7 shows the basic geomorphic features of **passive** or **Atlantic-type margins** with mountains (see Battiau-Queney 1991; Ollier 2004b). It seems likely that these features start as an old plain (palaeoplain) of a continental interior that breaks along a rift valley (Ollier and Pain 1997). The palaeoplain at the new continental edge, which is created by the rifting, experiences downwarping. Sea-floor spreading then favours the growth of a new ocean in which post-rift sediments accumulate as a wedge on the submerged palaeoplain to form a seawards-sloping basal unconformity. This is the **breakup unconformity** owing to its association with the fragmenting of a supercontinent (Ollier 2004). Inland the palaeoplain survives as **plateaux**. Some plateaux may be depositional but most are erosion surfaces formed of uplifted palaeoplains. In areas where the sedimentary strata form folds, the uplands are bevelled cuestas and accordant, level strike ridges. The plateaux may extend over large areas or they may have suffered dissection and survive as fragments on the hardest rocks. They often retain the ancient drainage lines. **Marginal swells** are widespread asymmetrical bulges along continental edges that fall directly into the sea with steeper (2°) slopes towards the coast. They develop after the formation of plateaux and major valleys. **Great escarpments** are highly distinctive landforms of many passive margins. They are extraordinary topographic features formed in a variety of rocks (folded sedimentary rocks, granites, basalts, and metamorphic rocks) and separate the high plateaux from coastal plains. The great escarpment in southern Africa in places stands more than 1,000 m high. Great escarpments often separate soft relief on inland plateaux from highly dissected relief beyond the escarpment foot. Not all passive margins bear great escarpments, but many do (Figure 4.8). A great escarpment has even been identified in Norway,

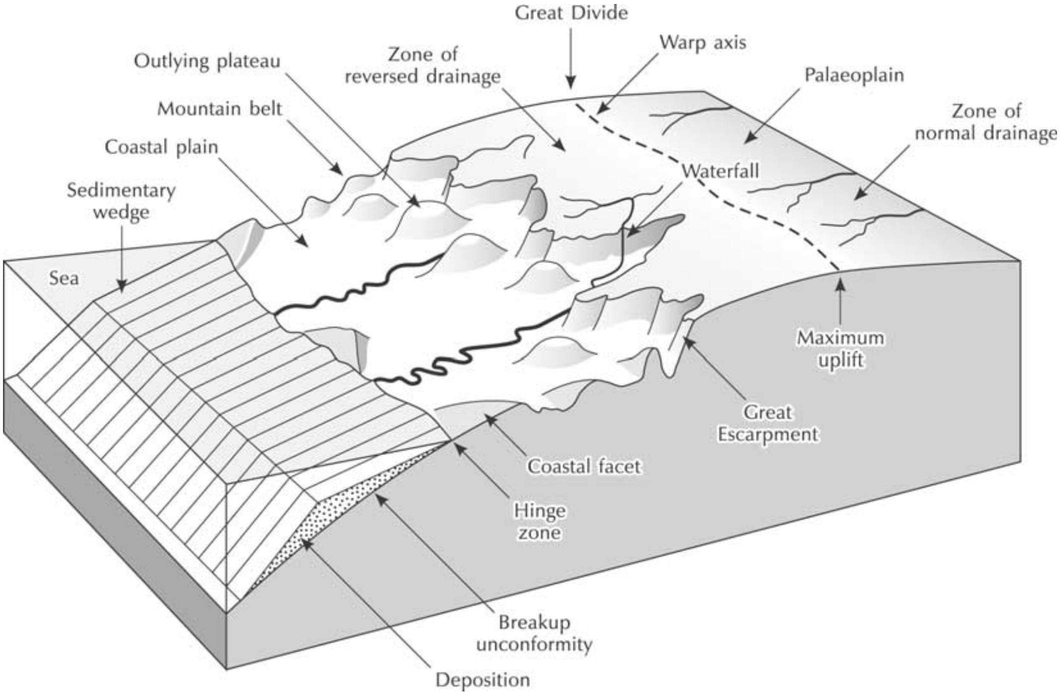


*Figure 4.7* The chief morphotectonic features of a passive continent margin with mountains.
*Source:* Adapted from Ollier and Pain (1997)

where the valleys deeply incised into the escarpment, although modified by glaciers, are still recognizable (Lidmar-Bergström *et al.* 2000). Some passive margins that lack great escarpments do possess low marginal upwarps flanked by a significant break of slope. The **Fall Line** on the eastern seaboard of North America marks an increase in stream gradient and in places forms a distinct escarpment. Below great escarpments, **rugged mountainous areas** form through the deep dissection of old plateaux surfaces. Many of the world's large waterfalls lie where a river crosses a great escarpment, as in the Wollomombi Falls, Australia. **Lowland** or **coastal plains** lie seawards of great escarpments. They are largely the products of erosion. Offshore from the coastal plain is a wedge of sediments, at the base of which is an unconformity, sloping seawards.

Interesting questions about passive-margin landforms are starting to be answered. The Western Ghats, which fringe the west coast of peninsular India, are a great escarpment bordering the Deccan Plateau. The ridge crests stand 500–1,900 m tall and display a remarkable continuity for 1,500 km, despite structural variations. The continuity suggests a single, post-Cretaceous process of scarp recession and shoulder uplift (Gunnell and Fleitout 2000). A possible explanation involves denudation and backwearing of the margin, which promotes flexural upwarp and shoulder uplift (Figure 4.9). Shoulder uplift could also be effected by tectonic processes driven by forces inside the Earth.

## Active-margin landforms

Where tectonic plates converge or slide past each other, the continental margins are said to be **active**. They may be called **Pacific-type margins** as they are common around the Pacific Ocean's rim.

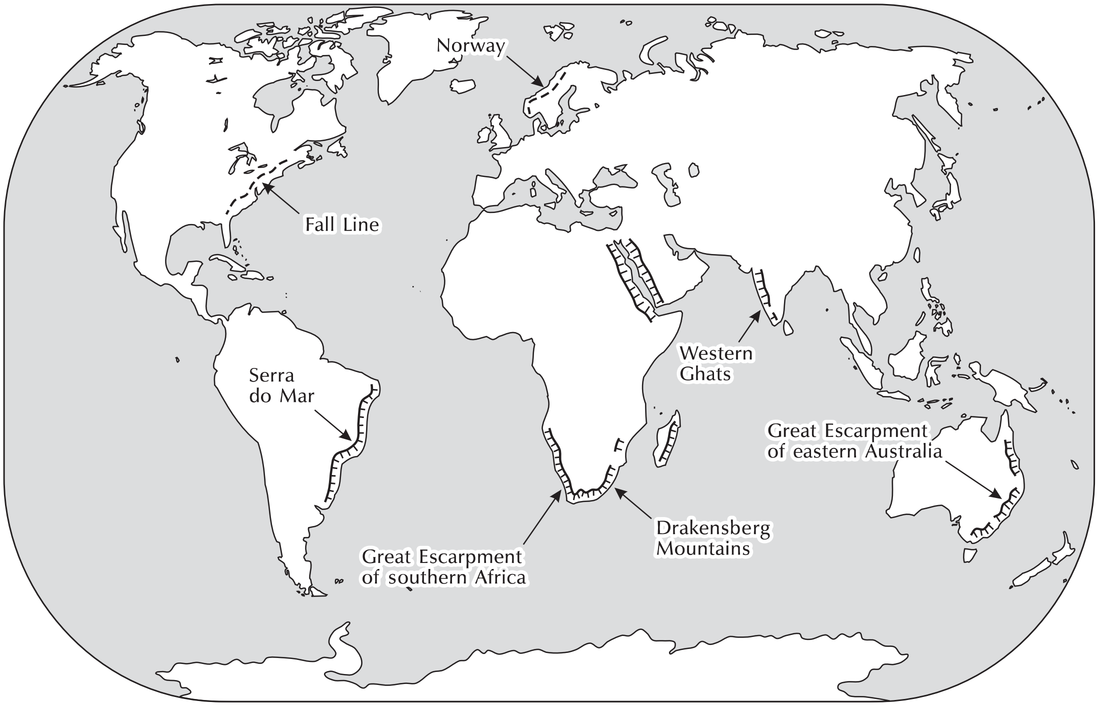


*Figure 4.8* Great escarpments on passive margins.
*Source:* Adapted from Summerfield (1991, 86)

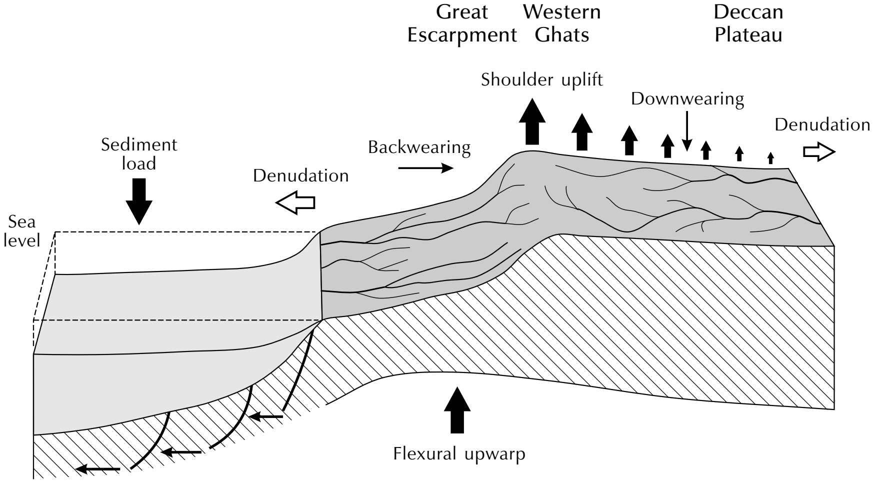


*Figure 4.9* Conceptual model of passive margin denudation and shoulder uplift by flexural rebound based on the Western Ghats, India.
*Source:* Adapted from Gunnell and Fleitout (2000)

The basic landforms connected with convergent margins are island arcs and orogens. Their specific form depends upon (1) what it is that is doing the converging – two continents, a continent and an island arc, or two island arcs; and (2) whether subduction of oceanic crust occurs or a collision occurs. **Subduction** is deemed to create steady-state margins in the sense that oceanic crust is subducted indefinitely while a continent or island arc resists subduction. **Collisions** are deemed to occur when the continents or island arcs crash into one other but tend to resist subduction.

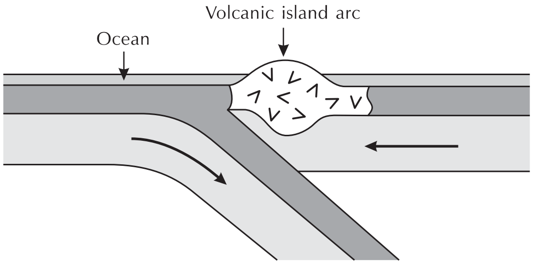
## Steady-state margins

Steady-state margins produce two major landforms – intra-oceanic island arcs and continental-margin orogens (Figure 4.10).

**Intra-oceanic island arcs** result from oceanic lithosphere being subducted beneath another oceanic plate. The heating of the plate that is subducted produces volcanoes and other thermal effects that build the island arc. Currently, about twenty intra-oceanic island arcs sit at subduction zones. Most of these lie in the western Pacific Ocean and include the Aleutian Arc, the Marianas Arc, the Celebes Arc, the Solomon Arc, and the Tonga Arc. The arcs build relief through the large-scale intrusion of igneous rocks and volcanic activity. A deep trench often forms ahead of the arc at the point where the oceanic lithosphere starts plunging into the mantle. The Marianas Trench, at −11,033 m the deepest known place on the Earth's surface, is an example.

**Continental-margin orogens** form when oceanic lithosphere is subducted beneath continental lithosphere. The Andes of South America are probably the finest example of this type of orogen. Indeed, the orogen is sometimes called an Andean-type orogen, as well as a Cordilleran-type orogen. Continental-margin island arcs form if the continental crust is below sea level. An example is the Sumatra–Java section of the Sunda Arc in the East Indies.

(*a*) Intra-oceanic island arc



(*b*) Continental-margin orogen

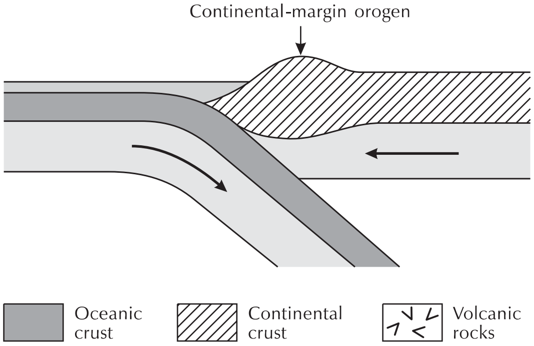


*Figure 4.10* The two kinds of steady-state margins. (a) Intra-ocean island arc formed where an oceanic plate is subducted beneath another oceanic plate. These are common in the western Pacific Ocean. (b) A continental margin orogen formed where an oceanic plate is subducted beneath a continental plate. An example is the Andes.
*Source:* Adapted from Summerfield (1991, 58)

## Collision margins

Landforms of collision margins vary according to the properties of the colliding plate boundaries. Four types of collision are possible: a continent colliding with another continent; an island arc colliding with a continent; a continent colliding with an island arc; and an island arc colliding with an island arc (Figure 4.11):

1 Continent–continent collisions create **intercontinental collision orogens**. A splendid example is the Himalaya. The collision of India with Asia produced an orogen running over 2,500 km.

2 Island arc–continent collisions occur where an island arc moves towards a subduction zone adjacent to

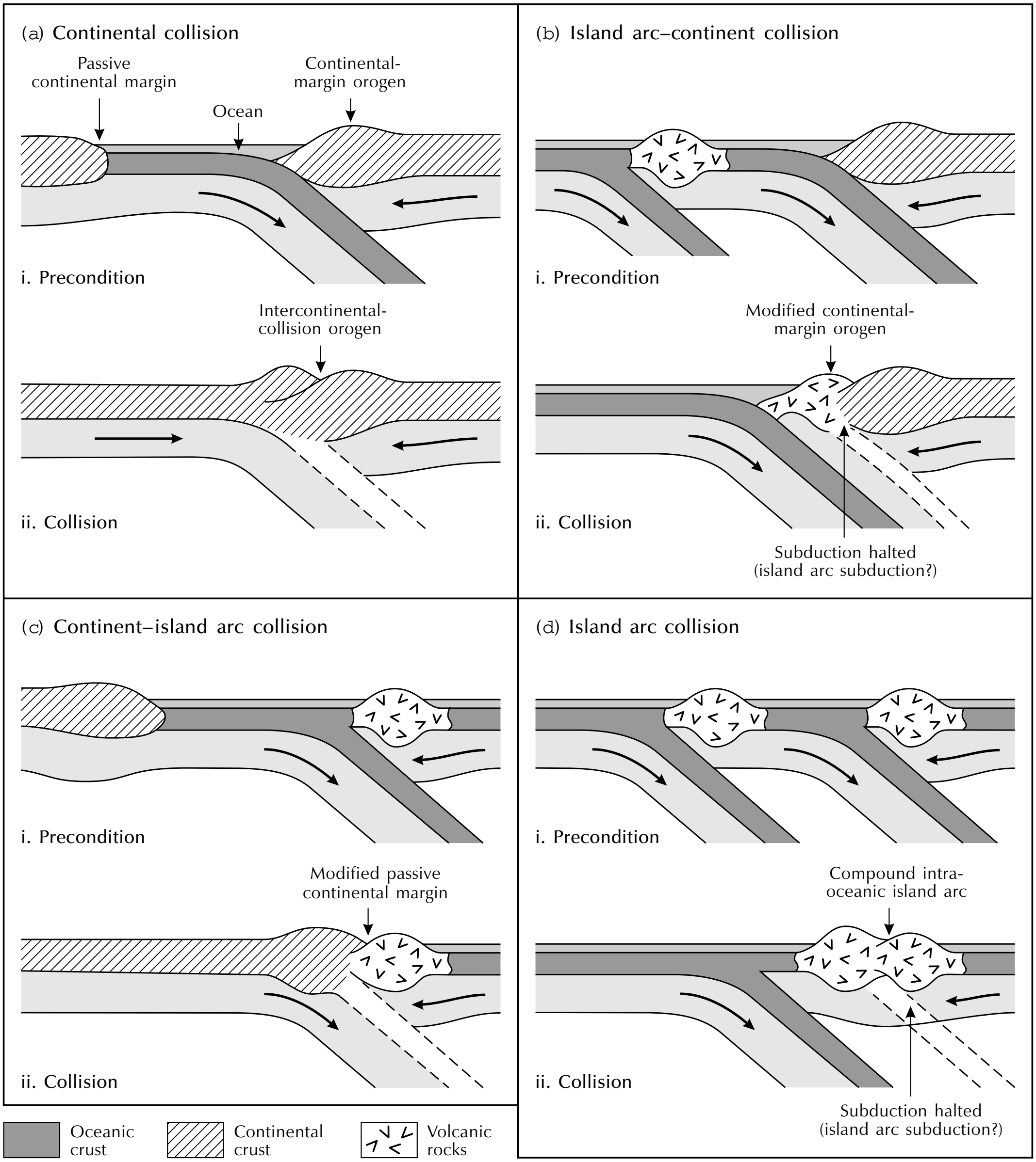


*Figure 4.11* Four kinds of collisional margins. (a) Intercontinental collision orogen formed where two continental plates collide. An example is the Himalaya. (b) Modified continental-margin orogen formed where an intra-oceanic island arc moves into a subduction zone bounded by continental crust. (c) Modified passive continental margin formed where a continent moves towards a subduction zone associated with an intra-oceanic island arc. (d) Compound intra-oceanic island arc formed by the collision of two intra-oceanic island arcs.
*Source:* Adapted from Summerfield (1991, 59–60)

a continent. The result is a **modified continental-margin orogen**.

3 Continent–island arc collisions occur when continents drift towards subduction zones connected with intra-oceanic island arcs. The continent resists significant subduction and a **modified passive continental margin** results. Northern New Guinea may be an example.

4 Island arc–island arc collisions are poorly understood because there are no present examples from which to work out the processes involved. However, the outcome would probably be a **compound intra-oceanic island arc**.

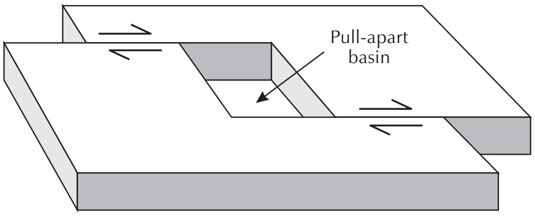
## Transform margins

Rather than colliding, some plates slip by one another along **transform** or **oblique-slip faults**. Convergent and divergent forces occur at transform margins. Divergent or transtensional forces may lead to **pull-apart basins**, of which the Salton Sea trough in the southern San Andreas Fault system, California, USA, is a good example (Figure 4.12a). Convergent or transpressional forces may produce **transverse orogens**, of which the 3,000-m San Gabriel and San Bernardino Mountains (collectively called the Transverse Ranges) in California are examples (Figure 4.12b). As transform faults are often sinuous, pull-apart basins and transverse orogens may occur near to each other. The bending of originally straight faults also leads to spays and wedges of crust. Along **anastomosing faults**, movement may produce upthrust blocks and down-sagging ponds (Figure 4.13). A change in the dominant direction of stress may render all these transform margin features more complex. A classic area of transform margin complexity is the southern section of the San Andreas fault system. Some 1,000 km of movement has occurred along the fault over the last 25 million years. The individual faults branch, join, and sidestep each other, producing many areas of uplift and subsidence.

## Terranes

Slivers of continental crust that somehow become detached and then travel independently of their parent body, sometimes over great distances, may eventually attach to another body of continental crust. Such wandering slivers go by several names: allochthonous terranes, displaced terranes, exotic terranes, native terranes, and suspect terranes. **Exotic** or **allochthonous terranes** originate from a continent different from that against which they now rest. **Suspect terranes** may be exotic, but their exoticism is not confirmable. **Native terranes** manifestly relate to the continental margin against which they presently sit. Over 70 per cent of the North American Cordillera is composed of displaced terranes, most of which travelled thousands of kilometres and joined the margin of the North American craton during the Mesozoic and Cenozoic eras (Coney *et al.* 1980). Many displaced terranes also occur in the Alps and the Himalayas, including Adria and Sicily in Italy (Nur and Ben-Avraham 1982).

(*a*) Transtension



(*b*) Transpression

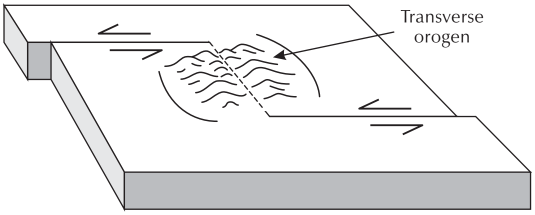


*Figure 4.12* Landforms associated with oblique-slip faults. (a) Pull-apart basin formed by transtension. (b) Transverse orogen formed by transpression.

## TECTONIC GEOMORPHOLOGY AND CONTINENTAL LANDFORMS

Important interactions between endogenic factors and exogenic processes produce macroscale and megascale

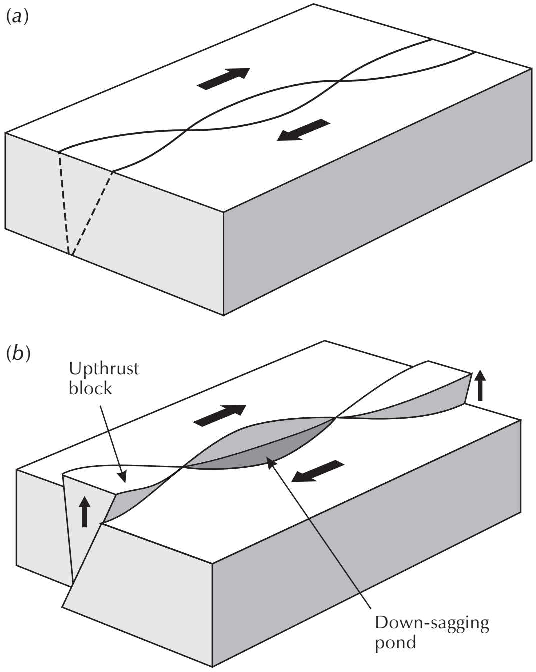


*Figure 4.13* Landforms produced by anastomosing faults. (a) Anastomosing faults before movement. (b) Anastomosing faults after movement with upthrust blocks and down-sagging ponds.
*Source:* Adapted from Kingma (1958)

landforms (Figure 1.1). Plate tectonics explains some major features of the Earth's topography. An example is the striking connection between mountain belts and processes of tectonic plate convergence. However, the nature of the relationship between mountain belts (orogens) and plate tectonics is far from clear, with several questions remaining unsettled (Summerfield 2007). What factors, for example, control the elevation of orogens? Why do the world's two highest orogens – the Himalaya–Tibetan Plateau and the Andes – include large plateaux with extensive areas of internal drainage? Does denudation shape mountain belts at the large scale, and are its effects more fundamental than the minor modification of landforms that are essentially a product of tectonic processes? Since the 1990s, researchers have addressed such questions as these by treating orogens, and landscapes more generally, as products of a coupled tectonic–climatic system with the potential for feedbacks between climatically influenced surface processes and crustal deformation (Beaumont *et al.* 2000; Pinter and Brandon 1997; Willett 1999).

The **elevation of orogens** appears crucially to depend upon the crustal strength of rocks. Where crustal convergence rates are high, surface uplift soon creates (in geological terms) an elevation of around 6 to 7 km that the crustal strength of rocks cannot sustain, although individual mountain peaks may stand higher where the strength of the surrounding crust supports them. However, in most mountain belts, the effects of denudation prevent elevations from attaining this upper ceiling. As tectonic uplift occurs and elevation increases, river gradients become steeper, so raising denudation rates. The growth of topography is also likely to increase precipitation (through the orographic effect) and therefore runoff, which will also tend to enhance denudation (Summerfield and Hulton 1994). In parts of such highly active mountain ranges as the Southern Alps of New Zealand, rivers actively incise and maintain, through frequent landslides, the adjacent valley-side slopes at their threshold angle of stability. In consequence, an increase in the tectonic uplift rate produces a speedy response in denudation rate as river channels cut down and trigger landslides on adjacent slopes (Montgomery and Brandon 2002). Where changes in tectonic uplift rate are (geologically speaking) rapidly matched by adjustments in denudation rates, orogens seem to maintain a roughly **steady-state topography** (Summerfield 2007). The actual steady-state elevation is a function of climatic and lithological factors, higher overall elevations being attained where rocks are resistant and where dry climates produce little runoff. Such orogens never achieve a perfect steady state because there is always a delay in the response of topography to changing controlling variables such as climate, and especially to changing tectonic uplift rates because the resulting fall in baselevel must be propagated along drainage systems to the axis of the range. Work with simulation models suggests that variations in denudation rates across orogens appear to affect patterns of crustal deformation (Beaumont *et al.* 2000; Willett 1999). For relatively simple orogens, the prevailing direction of rain-bearing winds seems significant.

On the windward side of the orogen, higher runoff generated by higher precipitation totals leads to higher rates of denudation than on the drier, leeward side. As a result, crustal rocks rise more rapidly on the windward flank than on the leeward flank, so creating a patent asymmetry in depths of denudation across the orogen and producing a characteristic pattern of crustal deformation. Such modelling studies indicate that a reversal of prevailing rain-bearing winds will produce a change in topography, spatial patterns of denudation, and the form of crustal deformation (Summerfield 2007). In addition, they show that the topographic and deformational evolution of orogens results from a complex interplay between tectonic processes and geomorphic processes driven by climate.

## SUMMARY

Geological processes and geological structures stamp their marks on, or in many cases under, landforms of all sizes. Plate tectonic processes dictate the gross landforms of the Earth – continents, oceans, mountain ranges, large plateaux, and so on – and many smaller landforms. Diastrophic forces fold, fault, lift up, and cast down rocks. Orogeny is a diastrophic process that builds mountains. Epeirogeny is a diastrophic process that upheaves or depresses large areas of continental cores without causing much folding or faulting. The boundaries of tectonic plates are crucial to understanding many large-scale landforms: divergent boundaries, convergent boundaries, and transform boundaries are associated with characteristic topographic features. Incipient divergent boundaries may produce rift valleys. Mature divergent boundaries on continents are associated with passive margins and great escarpments. Convergent boundaries produce volcanic arcs, oceanic trenches, and mountain belts (orogens). Transform boundaries produce fracture zones with accompanying strike-slip faults and other features. Plate tectonic processes exert an important influence upon such continental-scale landforms as mountain belts, but there is an important interplay between uplift, climate, and denudation.

## ESSAY QUESTIONS

1. **Explain the landforms associated with active margins.**
2. **Explain the landforms associated with passive margins.**
3. **Examine the factors that determine the major relief features of the Earth's surface.**

## FURTHER READING

Burbank, D. W. and Anderson, R. S. (2001) *Tectonic Geomorphology: A Frontier in Earth Science*. Malden, Mass.: Blackwell Science.
A detailed and insightful discussion of one of geomorphology's latest developments, but not easy for trainee geomorphologists.

Godard, A., Lagasquie, J.-J., and Lageat, Y. (2001) *Basement Regions*. Translated by Yanni Gunnell. Heidelberg: Springer.
An insight into modern French geomorphology.

Huggett, R. J. (1997) *Environmental Change: The Evolving Ecosphere*. London: Routledge.
You may find some of the material in here of use. I did!

Summerfield, M. A. (ed.) (2000) *Geomorphology and Global Tectonics*, pp. 321–38. Chichester: John Wiley & Sons.
Not easy for the beginner, but a dip into this volume will reward the student with an enticing peep at one of geomorphology's fast-growing fields.