4

Landforms and tectonics of plate interiors

4.1 Landscapes of plate interiors

We saw in Chapter 3 how a number of major morphological features, notably orogenic mountain belts and island arcs, coincide with convergent plate boundaries; a superficial glance at the Earth's morphology might even suggest that all significant sub-aerial topographic features are confined to such boundaries. But the fact that the effects of plate interactions can extend for thousands of kilometres away from plate boundaries, as we have seen in the case of the Tibetan Plateau and central Asia (see Section 3.4.4.2), shows that this is certainly not the case. Moreover, a closer inspection of global morphology reveals a range of major landform features which are located in the interior of plates and which are unrelated to even distant plate convergence events (Fig. 4.1).

Some continents, most notably Africa, have extensive plateaus which lie at elevations (2000 m or more) which far exceed the average height of the continental platforms. **Basins** represent another major type of landform of plate interiors. These are commonly 1000 km or more across and are either entirely enclosed and therefore drained internally, such as the Lake Eyre Basin of Australia and the Chad and Kalahari Basins of Africa, or they are breached by one or more major river system, such as the region drained by the Zaire (Congo) river system (Fig. 4.1). Although volcanoes and associated volcanic landforms are clearly concentrated along plate margins, significant occurrences are also to be found in plate interiors. On the continents some of this intra-plate volcanic activity is closely related to rift valleys which themselves represent sites of **continental rifting** where the crust has been stretched and faulted.

The break-up of the supercontinent of Pangaea, which began about 180 Ma ago (Fig. 2.16), created many new continental margins. This event was of great significance for the development of landscapes on a continental scale because it led to the establishment of new base levels for continental erosion. These margins exhibit a much lower level of tectonic activity than the active margins located along convergent plate boundaries and are consequently termed **passive continental margins** (or simply **passive margins**). Where formed

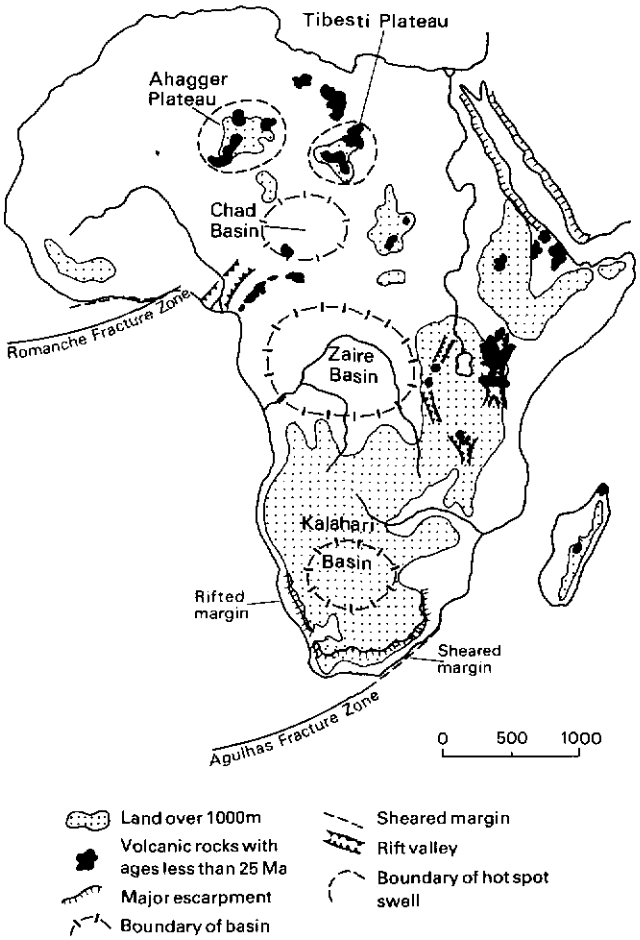


Fig. 4.1 Examples of landforms related to tectonic processes operating in plate interiors on the African continent. Note that for clarity only some of the basins and hot-spot swells are shown.

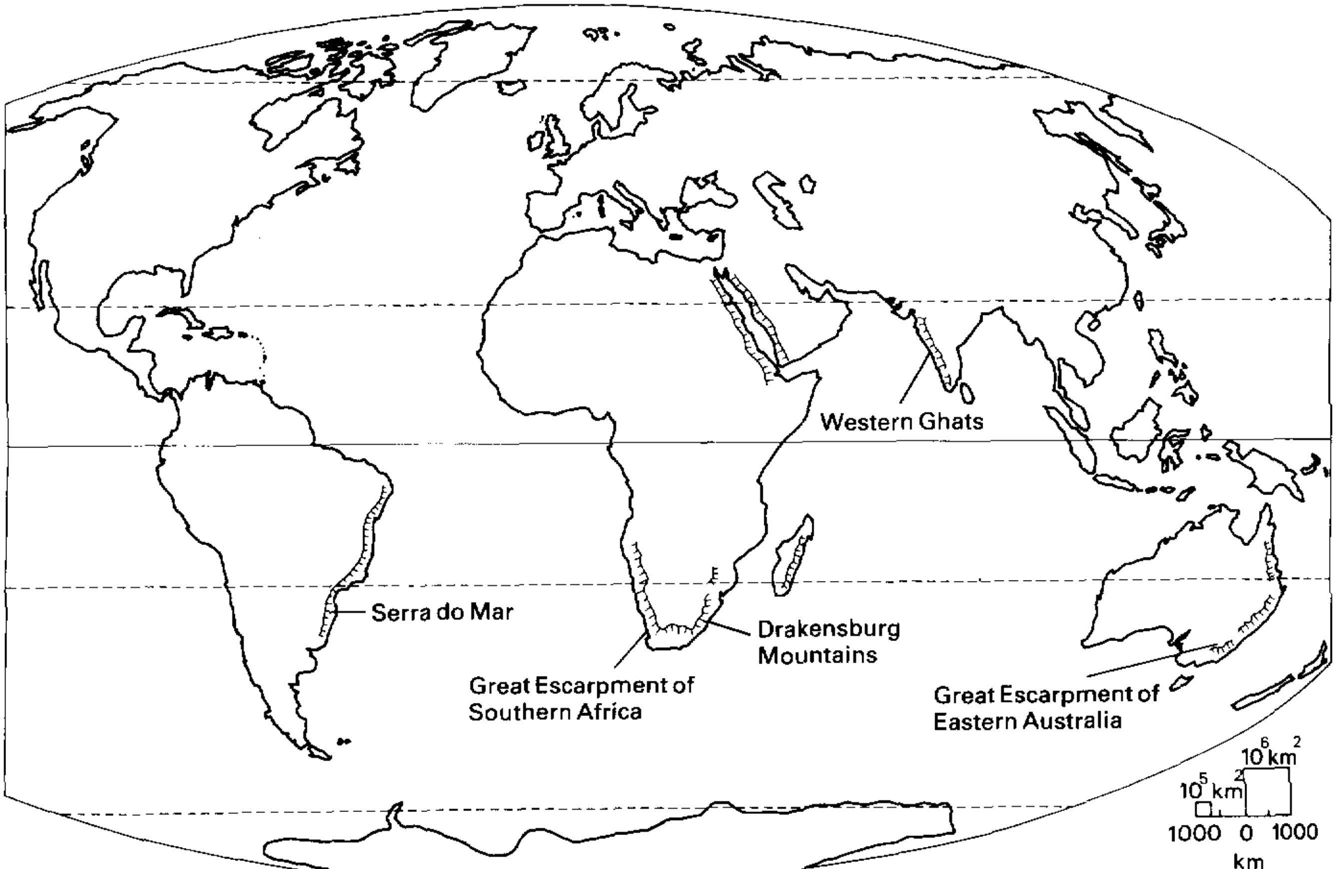


Fig. 4.2 Great escarpments on passive margins. Although lacking major escarpments, a number of other passive margins possess low amplitude marginal upwarps flanked by a significant break of slope. An example is the Fall Line of the eastern seaboard of North America which marks a significant increase in stream gradients and in places this develops into a distinct escarpment.

by divergent plate movement they are described as **rifted margins,** but where the motion between two adjacent continental blocks has been transform they are called **sheared margins** (but note that the term rifted margin is often applied rather loosely to passive margins of any type). Many, though certainly not all, passive continental margins are characterized by major escarpments (sometimes called **great escarpments**) (Fig. 4.2). By any standards these are significant morphological features – the Great Escarpment in southern Africa has a relief exceeding 1000 m in places – and they typically separate distinct geomorphic environments. Below great escarpments the topography is usually highly dissected, but inland on the plateau surface the relief is generally subdued.

Along plate margins the horizontal motion of plates is the primary force driving the uplift which occurs in orogens (although as we saw in Section 3.3 vertical movement resulting from thermal effects is of great importance in continental-margin orogens). In plate interiors horizontal plate movements play a less direct role in creating landforms and it is the processes which generate the broad warping and vertical movements of the crust, that is, the mechanisms of epeirogeny, that are of most concern.

4.2 Mechanisms of epeirogeny

While we now have a reasonably good understanding of how interactions at plate boundaries give rise to major morphological features on the continents our knowledge of the tectonic processes operating in plate interiors is by comparison rather poor. Numerous mechanisms have been suggested to explain the often significant uplift that has occurred in plate interiors but none seems to be applicable to all cases. Although the various models defy any simple classification, for convenience we will consider them under the headings of thermal, phase change and mechanical models.

4.2.1 Thermal models

There are many occurrences of volcanic activity in areas remote from plate margins; the volcanoes of the East Africa

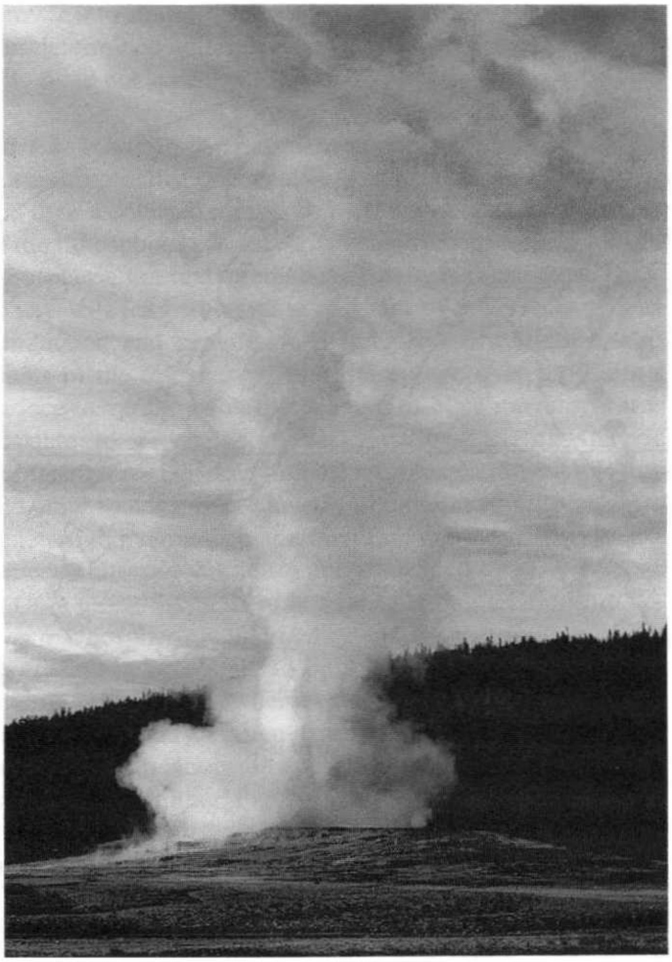


Fig. 4.3 Old Faithful geyser, Yellowstone National Park, Wyoming, USA, representing hydrothermal activity associated with a hot spot beneath the North American continent. The surrounding region is characterized by active volcanism and uplift.

Rift system, such as Kilimanjaro and Mount Kenya, the volcanic and associated hydrothermal activity of the Yellowstone region in Wyoming, U S A (Fig. 4.3), and the chain of volcanic peaks making up the Hawaiian Islands are examples from both continental and oceanic regions. There are also anomalous areas of volcanism on, or near, mid-oceanic ridges; Iceland, for instance, is formed of unusually thick oceanic crust produced by very high localized rates of volcanic activity on the Mid-Atlantic Ridge. Such an area of anomalous volcanism, which is not related to processes of subduction or normal mid-oceanic ridge spreading, is termed a **hot spot**. In addition to volcanic activity hot spots are in many cases associated with significant crustal uplift. The resulting hot-spot swells are typically several hundred kilometres across and may rise 1 km or more above the surrounding terrain.

Hot spots appear to be rather irregularly distributed over the Earth's surface (Fig. 4.4) but the real pattern is difficult to establish because of the problems of identifying specific hot spots and incomplete data in some regions. Along mid-oceanic ridges and subduction zones it is often a problem distinguishing between 'normal' and anomalous volcanism. Since it is also difficult in some cases to determine whether neighbouring centres of active volcanism are associated with one or more hot spots, estimates of the total number of hot spots vary significantly from around 40 to well over 100.

A hot spot arises from the presence of unusually hot mantle at the base of the lithosphere (Fig. 4.5(A)). Such a **sub-lithospheric thermal anomaly** may be generated by a **mantle plume**, an upwelling of hot mantle originating deep in the Earth's interior, possibly at the core-mantle boundary; alternatively, it may be caused by localized heating at shallower depths in the mantle. In either case the presence of hot mantle leads to melting at the base of the lithosphere and the lithosphere as a whole becomes thinner (Fig. 4.5(B)). Since part of the lithosphere is replaced by hot mantle of lower density from the asthenosphere, the lithosphere as a whole experiences isostatic uplift. As the change in density is the result of a change in temperature this kind of isostatic adjustment is termed **thermal isostasy**. Eventually uplift may be sufficient to stretch the brittle overlying crust to the point where it fractures and a rift valley forms (Fig. 4.5(C)).

The time scale over which uplift takes place depends upon the way in which heat is transferred through the lithosphere. Conduction of heat occurs only very slowly through rock, and if this is the only mechanism operating uplift will occur over a very long period – something of the order of 100 Ma. In contrast **penetrative magmatism**, whereby hot and partially molten material from the asthenosphere forces its way up towards the surface, is a much more rapid mechanism of heat transfer and is capable of thinning 100 km thick lithosphere and producing uplift at the surface in only about 20 Ma.

As several million years are required for the transfer of a significant amount of heat to the surface of continental lithosphere, it has been suggested that hot spots are most likely to develop where the lithosphere is more or less stationary with respect to sub-lithospheric thermal anomalies as this would allow time for sustained heating to occur. This idea has been applied to the African Plate which, according to some analyses of plate motions, appears to have been almost stationary with respect to the underlying asthenosphere for the past 25–35 Ma. This period coincides with a well-established increase in volcanic activity in Africa and, less certainly, with a phase of uplift.

Although the idea of linking volcanism and uplift to sub-lithospheric thermal anomalies is attractive there are problems with this hypothesized relationship. One difficulty is that there is evidence indicating that hot spots can migrate with respect to each other at velocities of up to 20 mm a^{-1}. This means that both hot spots can move under 'stationary' lithosphere and that the lack of motion of the African Plate is

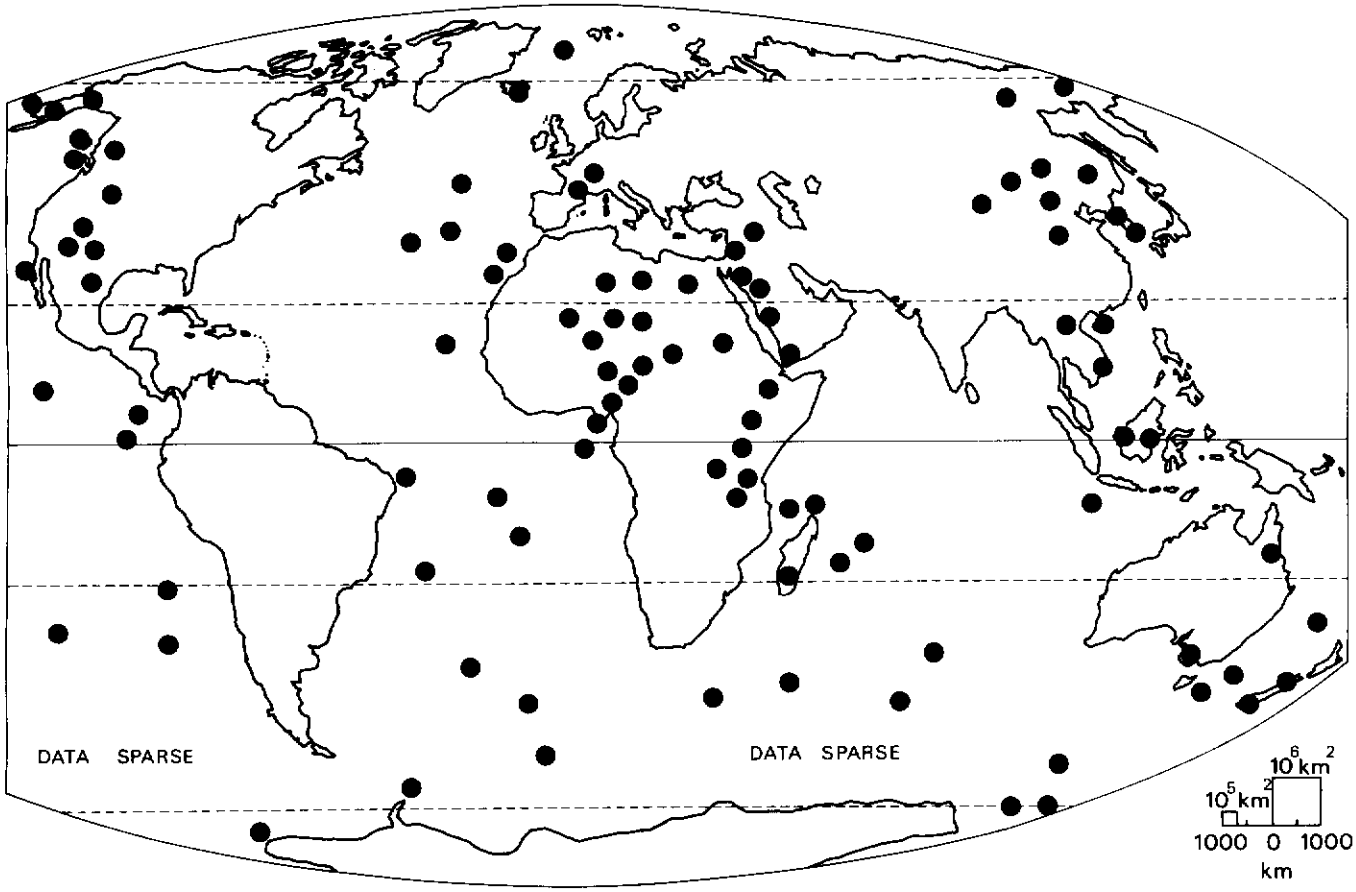


Fig. 4.4 Global distribution of hot spots. (Based on P. R. Vogt, (1981) Journal of Geophysical Research *86, Fig. 1, p. 951. Published by The American Geophysical Union.)*

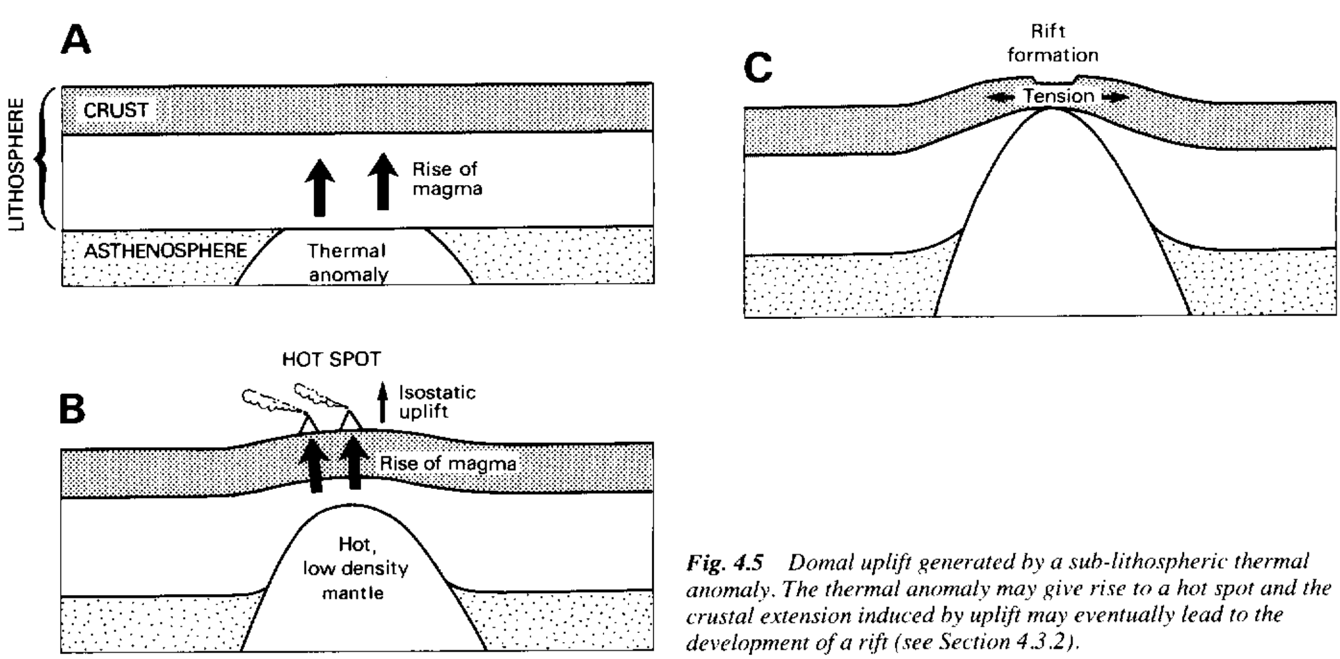


Fig. 4.5 Domal uplift generated by a sub-lithospheric thermal anomaly. The thermal anomaly may give rise to a hot spot and the crustal extension induced by uplift may eventually lead to the development of a rift (see Section 4.3.2).

thrown into doubt because its past positions have been in part determined with respect to a reference frame of supposedly fixed hot spots.

4.2.1.1 Lithospheric vulnerability

For a more realistic analysis of the occurrence of hot spots and their associated uplifts we need to consider the thickness of the lithosphere as well as its velocity with respect to sub-lithospheric thermal anomalies. This is because thick lithosphere will tend to be more resistant to the effects of heat conduction and penetrative magmatism. By relating variations in the thickness of the lithosphere over the Earth's surface to variations in the velocity of plate movement it is possible to determine global variations in **lithospheric vulnerability** (Fig. 4.6; Box 4.1). Lithosphere which is thin and slow moving is predicted to be the most vulnerable to penetrative magmatism and therefore the most likely to experience hot-spot volcanism and associated uplift.

On a global basis there is a fairly strong relationship between lithospheric vulnerability and the distribution of hot spots; few hot spots are to be found in areas of low lithospheric vulnerability whereas they are common in areas of high vulnerability. When specific regions are examined, however, this relationship does not always seem to apply. By carefully examining Figure 4.6 you can see that on the African Plate the relationship is, in fact, the opposite of that predicted with several hot spots located in areas of low lithospheric vulnerability. There are several possible reasons for this including inaccuracies in the estimates of lithospheric thickness and the over-zealous identification of hot spots on the African Plate. Nevertheless it is probable that factors other than lithospheric vulnerability, such as the actual global distribution of sub-lithospheric thermal anomalies, are important in influencing the occurrence of hot-spot volcanism and associated uplift.

Box 4.1 Lithospheric vulnerability

The lithospheric vulnerability index (v) is defined in terms of lithospheric thickness (l) and plate velocity (u) as

$$v = \frac{K}{lu^{1/2}}$$

where K is a constant for a property of the lithosphere known as thermal diffusivity which expresses the rate of propagation of a temperature change through time. This expression demonstrates that vulnerability is the same for 200 km thick lithosphere moving very slowly at 10 mm a^{-1} as it is for lithosphere which is only 63 km thick but which is moving rapidly at 100 mm a^{-1}.

Fig. 4.6 Global distribution of hot spots (triangles) and contours of the dimensionless lithospheric vulnerability index; lower values of the index indicate areas more vulnerable to penetrative magmatism. Major plate boundaries are also shown. Note that the set of hot-spot locations shown is slightly different from that illustrated in Figure 4.4. (After H. N. Pollack et al., *(1981)* Journal of Geophysical Research *86, Fig. 3, p. 965. Copyright by the American Geophysical Union.)*

4.2.1.2 *Effects of hot-spot migration*

One intriguing aspect of the effects of sub-lithospheric thermal anomalies on continental topography arises from the consequences of plates moving over such regions of unusually hot asthenosphere. As yet there has been no detailed attempt to assess the possible effects of such hot-spot 'migration' on landscape development, but there is some evidence from the stratigraphic record of the movement of topographic swells across continents which may be related to the passage of hot spots or similar thermal phenomena. For example, from about 160 to 100 Ma BP, North America drifted over a hot spot which is now located near the Great Meteor Seamount in the eastern Atlantic Ocean. Patterns of erosion, sedimentation and igneous activity recorded over the area during this period seem to be consistent with a migrating region of uplift related to the track of this hot spot.

On the basis of an estimated 10 per cent of the Earth's surface currently being part of a hot-spot swell it has been calculated that on average hot-spot epeirogeny should affect a particular area of crust about every 600 Ma. We would expect the slow migration of hot-spot swells across continental interiors to have a significant impact on landscape development through, for instance, the disruption and diversion of drainage systems. Geomorphologists have yet to assess this possibility, although the relationship between migrating hot-spot swells and landscape development provides an intriguing subject for future research.

4.2.2 Phase changes

A major problem with hot spots as a general explanation for uplift in plate interiors is that many elevated continental regions in such areas show no evidence of volcanism. Large parts of southern and eastern Africa stand at over 1500 m yet recent volcanism has been very localized. This might be explained by sub-lithospheric thermal anomalies causing uplift without penetrative magmatism developing to a point where volcanic activity occurs at the surface. Another difficulty with the hot-spot model is that unrealistically large increases in heat flow are apparently required to explain the magnitude of uplift recorded in some regions.

An alternative explanation of such uplifts involves the effects of density changes in minerals in the upper mantle. Such **phase changes** occur when minerals adopt different atomic configurations as temperatures and pressures change. The temperatures experienced by a mineral in the crust or mantle can be affected by heat flow from below while the pressure can be altered by loading and unloading of the crust above. Consequently the depth at which particular phase changes (and therefore density changes) occur can migrate up and down within the lithosphere. The surface above will rise or fall in response to the changes in volume brought about by these alterations in mineral density. A number of possibly significant phase changes have been suggested but our lack of detailed knowledge of the abundance of specific minerals in the lower crust and upper mantle makes their evaluation rather uncertain.

A phase change, triggered by a small amount of heating at the base of the lithosphere, has been proposed to account for the uplift evident in a number of regions in plate interiors, including southern and eastern Africa, south-east Australia and the Transantarctic Mountains of Antarctica. Each of these areas of continental lithosphere are thought to have overridden regions of hot asthenosphere associated with former mid-oceanic spreading ridges. It is thought that this would have increased the flow of heat to the base of the lithosphere sufficiently to cause a phase change in the uppermost continental mantle after a delay of up to 70 Ma (the time taken for heat to be conducted from the base of the lithosphere to just below the Moho). The conversion of eclogite, a rock with a density of 3400 kg m^{-3}, to basalt (density 3000 kg m^{-3}) is the kind of phase change envisaged. If this is so a 11.25 km thick layer of eclogite would need to be converted to a thickness of 12.75 km of basalt to produce an uplift of 1500 m.

The attractiveness of this model is that it requires only a modest increase of temperature of around 100 °C at the base of the lithosphere. This seems much more reasonable than the very large increases of up to 1000 °C which would be required to produce an uplift of around 1500 m by purely thermal effects. A major problem with the model, however, is that it predicts a rather uniform and extensive pattern of uplift which does not accord with the typical morphology of continental plate interiors. Africa, for instance, is characterized by areas of domal uplift, from 500 to 2000 km across and by a prevalence of high elevations along its continental margins.

4.2.3 Mechanical models

The most straightforward mechanical models of uplift are those that involve the isostatic adjustment that takes place when a load is removed from the crust (see Section 2.2.4). The term epeirogeny was in fact first coined by G. K. Gilbert in 1890 to describe the isostatic uplift which occurred as a result of the evaporation of Lake Bonneville in Utah in the western U S A (leaving the much smaller Great Salt Lake of the present day). During a period of high lake levels 15 000–25 000 a BP the load provided by the average water depth of about 145 m depressed the crust. After unloading by evaporation the crust rose and shoreline features are now elevated up to 64 m above the present water level.

Far more spectacular examples of isostatic uplift are to be found in the areas of the northern hemisphere covered by great ice sheets during the Pleistocene. Detailed records of post-glacial uplift have been established for such areas as the Fenno-Scandian Shield and the Canadian Shield from both raised and warped shorelines and gravity data. Uplift of the central region of the Fenno-Scandian Shield began about

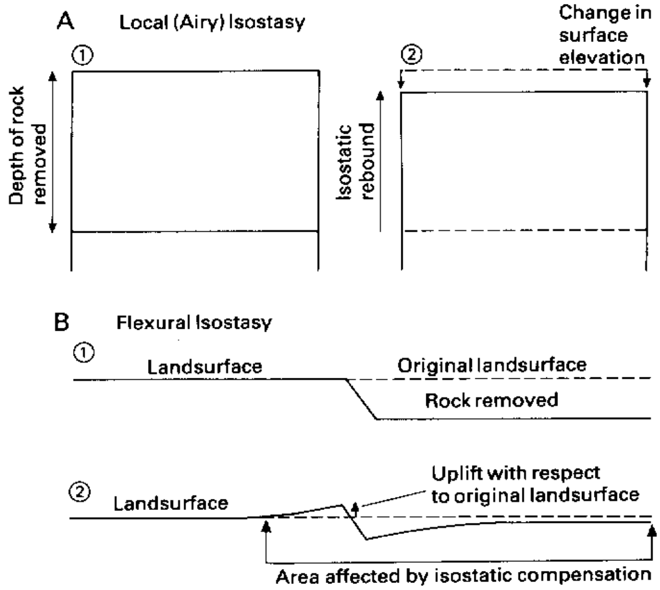


Fig. 4.7 *Models of isostasy showing the contrasting modes of vertical adjustment between local (Airy) isostasy (A), and flexural isostasy (B).*

13 000 a BP and the total uplift since then has exceeded 100 m.

Though involving less spectacular rates of uplift, crustal unloading by denudation is a far more pervasive process which extends over long periods of time, in contrast to the rather transient effects of deglaciation and evaporation of lakes. The latter two phenomena essentially restore the crust to its approximate elevation before loading by ice or water, but denudation can lead to continued uplift of the crust for as long as continental crust is sub-aerially exposed. If isostatic compensation is local, that is if vertical adjustments of the lithosphere in response to the removal of a load are confined just to the area covered by the original load, then isostatic uplift simply reduces the rate at which a landsurface is lowered by denudation (Fig. 4.7(A)). Such local compensation, however, cannot occur at the small scale because the lithosphere has a finite strength and so changes in load result in regional isostatic adjustments over a greater area than that actually affected by the change in load (Fig. 4.7(B)).

Such flexural isostasy (see Section 2.2.4) is a familiar phenomenon to geomorphologists who study the uplift and warping of shorelines as a result of crustal loading by nearby ice sheets. In this case flexing of the lithosphere leads to the development of a **forebulge** at some distance from the ice margin which experiences an increase in surface elevation. It is possible that a similar flexural effect is associated with great escarpments along passive continental margins. In this case there is active erosion (and therefore unloading) along the escarpment edge but relatively little erosion on the plateau surface immediately inland of the escarpment. Consequently it is possible that the slowing eroding plateau summit could actually increase in elevation as a result of the erosional unloading along the escarpment (Fig. 4.7(B)). A similar effect can lead to the summits of mountains in orogenic belts increasing in elevation as the incision of deep valleys between peaks causes unloading of the crust. The extent to which isostatic compensation occurs regionally through flexure, rather than locally, depends on the **flexural rigidity** of the lithosphere, a property which indicates its resistance to bending in response to a change in load.

Another proposed mechanism of epeirogenesis rests on the observation that continental lithosphere is in an unstable mechanical equilibrium because its mantle is denser than the underlying asthenosphere. Consequently if some process, such as cracking or 'erosion' by mantle plumes, punctures the continental lithosphere as far as the crust the relatively dense mantle layer of the lithosphere could become detached and sink (Fig. 4.8). This process has been termed **delamination**

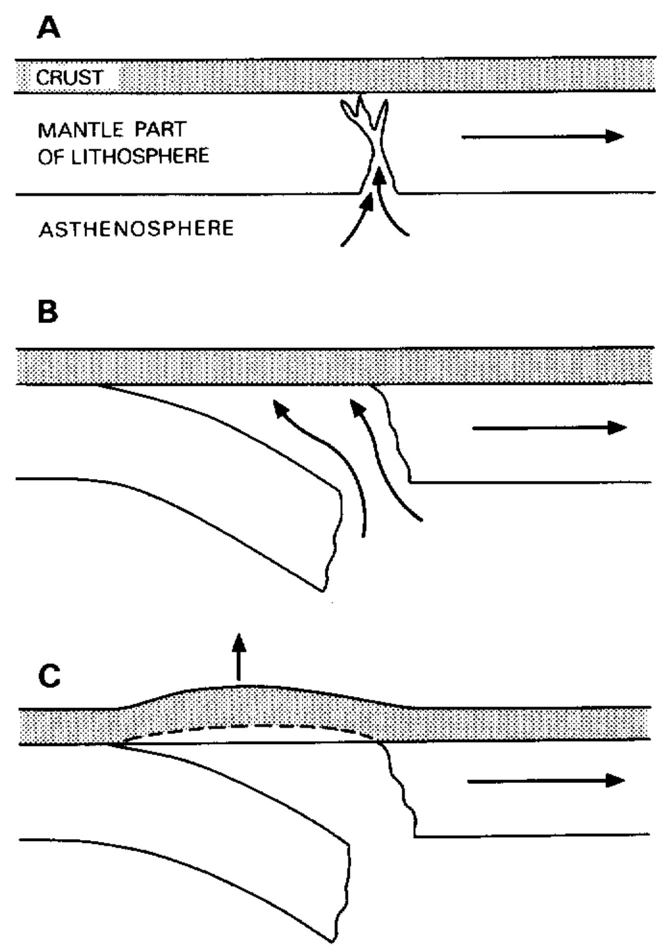


Fig. 4.8 *Schematic representation of delamination showing: (A) initial cracking of the mantle part of the lithosphere by penetrative magmatism; (B) the sinking of this portion of the lithosphere into the less dense asthenosphere; (C) crustal uplift associated with the replacement of lithosphere by warmer and less dense sub-lithospheric mantle. Plate motion is to the right in relation to the underlying asthenosphere.*

and may be capable of causing epeirogenic uplift through the exposure of the cold base of the crust to hot mantle from the asthenosphere and the replacement of relatively dense lithospheric mantle by less dense mantle from the asthenosphere. Two such delamination 'events' may account for the episodes of uplift of the Colorado Plateau in the south-west U S A. around 30 Ma and 5 Ma BP.

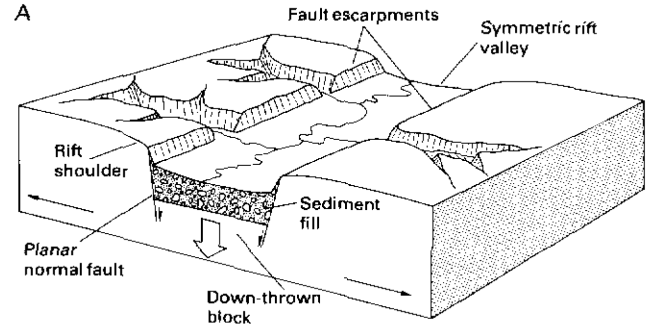
4.3 Continental rifts

4.3.1 Rift structure and location

Rift valleys are important landforms on continental crust in plate interiors where tensional stresses predominate in the *lithosphere*. They typically have widths of 30–60 km, but some are up to 100 km across. Pioneers of rift valley geology such as J. W. Gregory were strongly influenced in their interpretations of rift structure by the apparently simple rift morphology that they encountered in areas such as East Africa. The conventional image of a rift valley is of a broad, flat-floored and symmetric trough flanked by steep escarpments. The most obvious structural interpretation of this morphology is that of a graben with the rift floor being formed on a downthrown block bounded by normal faults which create the steep escarpments (Fig. 4.9(A)). But a closer examination of both the morphological and structural evidence shows that this model is inadequate. In particular the application of **seismic stratigraphy** to the study of rift valleys during the 1980s has led to a considerable revision of ideas about their structure.

Seismic stratigraphy is a technique whereby seismic waves generated by small explosions set off at the surface are reflected or refracted from discontinuities in the underlying sediments which represent changes in sediment properties. Such changes thus enable discrete sedimentary units to be identified and subsurface faults to be mapped (see Section 17.2.2.2. for a more detailed explanation). Seismic data which show the deep structure of rift systems have revealed a much more complex pattern of faulting than was suspected in the classic rift model, with the presence of listric as well as normal faults. Most significantly, the structure of many rifts examined by seismic methods has been shown to be asymmetric with most of the downthrow occurring along a major boundary listric fault on one side of the rift. Such rifts are said, therefore, to have a **half-graben** structure with the main boundary listric fault forming a **footwall,** and the opposing downwarped crust forming a **hanging wall** leading down from a **roll-over** zone (Fig. 4.9(B)). The presence of an **antithetic fault** on the hanging wall margin can give the impression of a symmetric rift valley if it is exposed and forms an escarpment, even though the overall structure is asymmetric.

Further evidence contradicting the traditional symmetric rift valley model comes from observations of their morphology and surface structure. Careful examination of fault



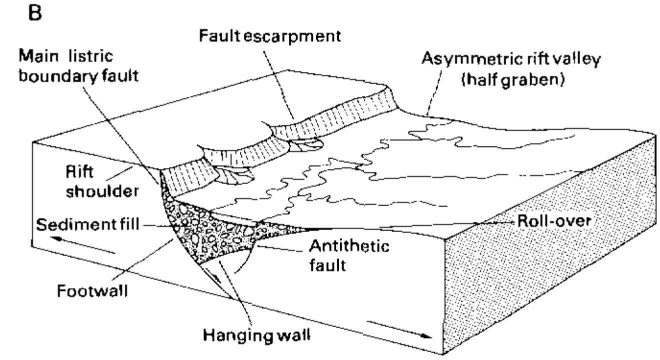


Fig. 4.9 Schematic representation of contrasting rift structures: (A) classic symmetric graben structure with a downthrown block bounded by normal faults; (B) asymmetric, half-graben structure. In both cases the number of faults in real rifts and the complexity of their structure is much greater than is indicated here.

patterns, for instance, reveals that major faults tend to occur on only one side of a rift along any one section of the valley. Moreover, these major faults tend to be discontinuous and curved in plan and it seems that main boundary listric faults alternate along rift valleys and are separated by **transfer faults** (Fig. 4.10). Some rift valley lakes, such as Lake Tanganyika, Lake Malawi and Lake Turkana in East Africa

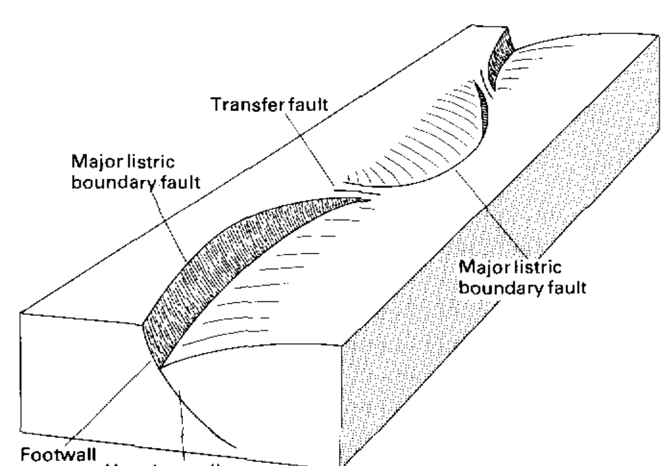


Fig. 4.10 Highly schematic illustration of alternating half-graben along a rift valley.

have a slightly sinuous plan form which may reflect the structural control exerted by these alternating curved boundary faults.

Even a cursory examination of global structure and morphology shows the generally close relationship between continental rifts and areas of mid-plate volcanism and crustal doming and uplift. The location of rift systems also seems to be closely related to the distribution of ancient structures representing zones of weakness between stable cratons. The East African Rift System, for instance, parallels the trend of old mobile belts (regions of highly deformed rocks formed during early phases of folding and orogenesis) and tends to follow the margins of Precambrian cratons (Fig. 4.11). Such ancient lines of weakness are apparently more readily activated than adjacent crust and consequently are favoured sites for later faulting and rift development. Some recently active rifts, however, do cut across ancient Precambrian structures, suggesting that the stress pattern initiating rifting, as well as the position of zones of crustal weakness, is important in determining the location and orientation of continental rifts.

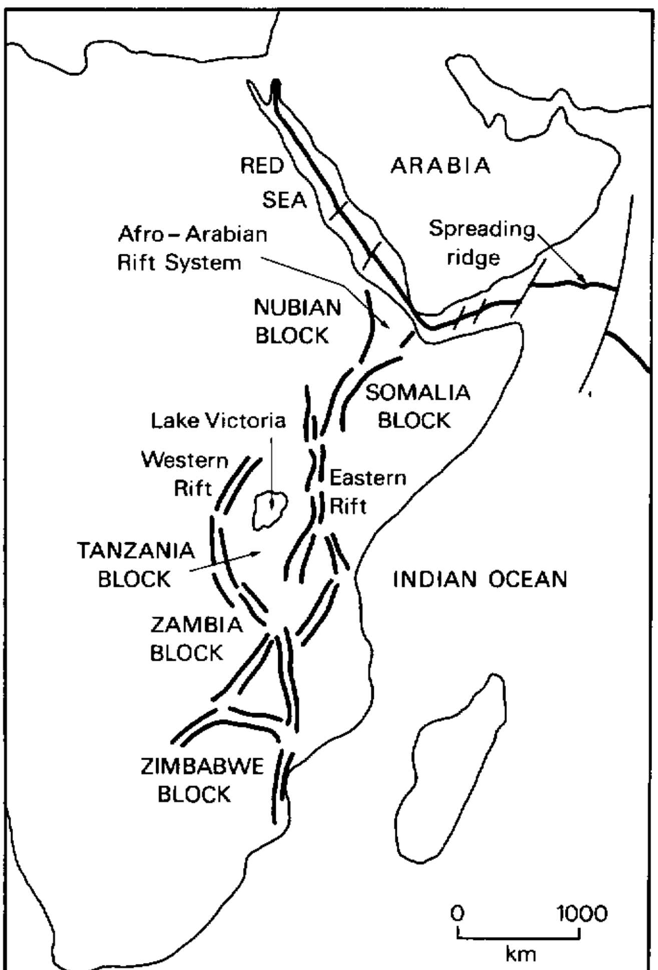


Fig. 4.11 Simplified map of the main components of the East African Rift System. Note the continuity with the spreading ridge leading from the Red Sea through the Gulf of Aden and into the Indian Ocean. (Based on various sources.)

The major differences in rift characteristics relate to their position with respect to plate boundaries and the intensity of volcanic activity associated with them. Some rifts appear to be clearly related to divergent crustal movements with lateral extension being normal to the rift axis. The East African Rift System is of this type and its interpretation as an incipient site of plate rupture is further supported by its continuity with the divergent plate boundaries of the Red Sea and Gulf of Aden. In other cases, such as the Dead Sea Rift, there is a significant transform component in the extensional movement and this feature can in fact be regarded as a large pull-apart basin (see Section 3.5). The other main rift type is to be found orientated approximately at right angles to the strike of intercontinental collision orogens. The development of this type of rift appears to be related to the crustal extension that occurs, as a result of continental collision, beyond the immediate collision zone itself. Examples of this type are the Rhine Rift in the foreland zone of the Alps, and the Baikal Rift System which lies some 3000 km to the north of the Himalayas (see Section 3.4.4.2).

There are significant differences in the morphology and degree of volcanic activity associated with these two types of rift. True intra-plate rifts, such as the East African Rift System, are characterized by prolonged volcanism and typically exhibit 1–2 km of down-faulting along the crest of substantial crustal upwarps. In contrast orogen-related rifts, such as the Rhine and Baikal structures, experience generally less intense volcanism and much more marked down-faulting, amounting to 5–6 km in the case of the Baikal Rift. In this kind of rift, however, the marginal uplift is either narrow or lacking altogether.

4.3.2 Rift formation

The commonly close relationship between rifting and crustal doming in rifts not associated with orogens suggests a causal relationship. There are two possibilities: one involves **active rifting** in which rifting develops as a response to the tensional stresses induced in the crust by uplift resulting from upwelling of the asthenosphere (Fig. 4.5(C)); the other involves **passive rifting** in which rifting is initiated by extensional stresses in the lithosphere, and this permits the subsequent upwelling of hot mantle which in turn induces thermal uplift (Fig. 4.12).

Clearly the timing of uplift, volcanism and rifting is crucial in the assessment of the validity of the active and passive rifting models. The typical sequence for passive rifting is rifting followed by volcanism, which may in turn be succeeded by uplift. Active rifting, on the other hand, would be expected to begin with volcanism which is then followed by uplift before rifting itself is finally initiated; but in some

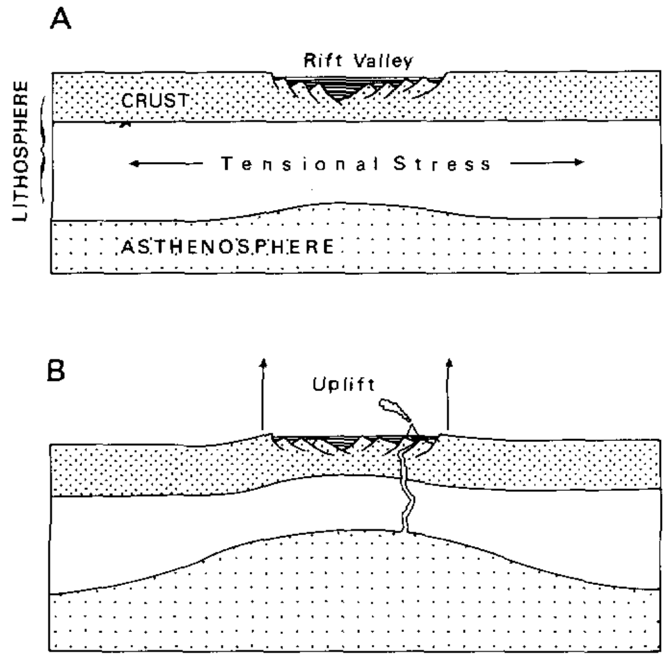


Fig. 4.12 Formation of a rift through extension of the lithosphere. In contrast to the active rifting model, in this passive rifting mechanism extension and thinning of the lithosphere (A) precedes uplift of the rift flanks (B).

cases uplift seems to precede volcanism. In practice some rifts seem to be of the active type whereas others seem to have been initiated by passive rifting, although in many cases there is no consensus as to whether specific rifts are active or passive.

4.3.3 Rifting and uplift in Africa

The domal uplifts and rift structures of Africa are at various stages of development and thus provide a useful basis for examining the evolution of rift systems. Two major periods of uplift and rift formation have been recognized in Africa; a Jurassic-Cretaceous rifting phase between 180 and 130 Ma BP associated with the break-up of Gondwana and a more recent period initiated some 35 to 25 Ma BP and continuing up to the present.

There is some evidence in Africa for the idea that uplift may precede volcanism since the Adamawa dome in the Cameroon region, which rises to elevations in excess of 2000 m, lacks evidence of recent volcanic activity. Volcanism and uplift without rifting is exemplified in North Africa where Neogene volcanic rocks are to be found on the Tibesti and Ahagger Plateaus (Fig. 4.1). Domal uplifts with rifts cutting across them are particularly common in East Africa where the East African Rift System comprises a whole series of such structures (Fig. 4.11). The crests of the domal uplifts typically have three rifts meeting at angles of about 120°; this geometrical arrangement appears to arise from the way in which the tensional stresses within the lithosphere are most readily accommodated. Eleven such triple junctions, each possibly associated with a separate hot spot, have been identified on the continental part of the African Plate and an intriguing problem is the way in which they may subsequently promote continental rupture and develop into spreading centres.

Spreading can apparently only occur when continental separation can be accommodated into the motions of the global plate system. Where this is not possible rift development is halted, rifts become filled with sediment and any associated uplift subsides. Although the triple junctions identified in Africa, including those along the continental margin as well as those in the interior, have variously evolved by spreading along one, two or all three rift arms, the most common sequence has been for one arm to remain inactive and form an **aulacogen,** with spreading occurring along the other two (Fig. 4.13). Of the 29 rifts identified in Africa, five have led to continental separation and a further three have been associated with the movement of Madagascar away from the African mainland. The two rifts in the Benue Trough in Nigeria went through a phase of opening followed by later closing, and of the remaining rifts ten became inactive before the spreading stage and nine are currently active and may spread in the future.

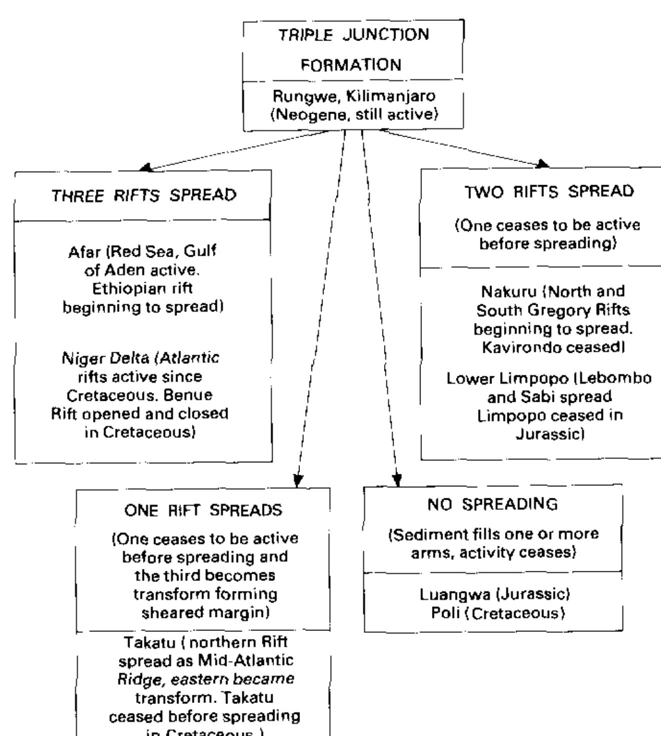


Fig. 4.13 Mode of evolution of triple junctions illustrated from rift structures in Africa. (Adapted from K. Burke and A. J. Whiteman (1973) in: D. H. Tarling and S. K. Runcorn (eds) Implications of Continental Drift for the Earth Sciences, *2 vols. Academic Press, London, Table 3, p. 749.)*

4.4 Continental basins

4.4.1 Basin morphology

In addition to upward vertical movements of the crust which give rise to broad upwarps or swells, downward movements also occur creating basins. Basins located in continental interiors are sometimes referred to as **intra-cratonic basins** while those located along continental margins in intra-plate settings are called **passive margin basins.** The latter are primarily submarine features and will not be discussed here although they are certainly an important component of passive margin development (see Section 4.5).

Areas of crustal subsidence usually form topographic features with minimal local relief because they become partly or wholly filled by sediment transported from surrounding highlands; examples include the Lake Eyre Basin in southern Australia, the Kalahari Basin in southern Africa and the Ob Basin in the USSR. The volume of sediment that accumulates in a basin depends on the rates of subsidence, sediment supply and sediment removal. Evacuation of basin sediments is primarily by rivers for basins with an outlet to the ocean (such as the Zaire Basin) but wind action is important in enclosed basins (such as the Kalahari Basin) (Fig. 4.1). Basins in continental interiors are a few hundred to a thousand or more kilometres across and the sedimentary units, with which they are partially or completely filled, generally thicken towards the centre. This sediment thickening indicates the role of sediment loading in sustaining basin subsidence but it does not explain how basins originate. Continental basins generally seem to be long-lived features with a subsidence history often extending over a period of 100 Ma or more.

4.4.2 Mechanisms of subsidence

Basins form as a result of downwarping of the crust, as a consequence of uplift of the surrounding region, or through a combination of both of these effects. Epeirogenic uplift and subsidence are closely related phenomena and a number of the mechanisms of uplift discussed in Section 4.2 may be relevant to crustal subsidence. For example, while heating of the lithosphere may account for uplift, subsidence may result from lithospheric cooling. In this case the connection between uplift and subsidence is especially close as there must first be a heating event for the subsequent cooling to give rise to a basin. On its own heating and cooling of the lithosphere will not create a basin as the crust will simply return to its original elevation. What is required is a mechanism for thinning the lithosphere during the heating phase so that when it cools and subsides a region of negative relief is formed.

One possible way of accomplishing this is through the erosional thinning of the crust when it is elevated during the period of uplift. The amount of thinning and consequently the amount of eventual subsidence will depend on the depth of denudation that has occurred. In most cases, however, this appears to be insufficient to account for the amount of subsidence observed.

If denudation is insufficient as a thinning mechanism then sub-crustal processes must be considered. Phase changes provide a possible mechanism which could account for subsidence as well as uplift, depending on whether the crystalline transformations involved give rise to more, or less, compact mineral structures. A further potential thinning mechanism which has received considerable support since the late 1970s is the stretching of the lithosphere as a result of sustained tensional stress (Fig. 4.12). Lithospheric extension will promote faulting and an initial phase of subsidence associated with upwelling of hot mantle from the asthenosphere into the base of the lithosphere. This is followed by a slower phase of subsidence promoted by cooling of this mantle material as heat is conducted up to the surface.

Irrespective of the mechanism initiating basin formation the amount of subsidence will be amplified by the weight of sediments which accumulate. This is undoubtedly a highly significant process where sediment thicknesses of several kilometres occur. It is also a major reason for the thickening of sedimentary units towards the centre of basins. The initial depression will be the first to accumulate sediment and this will lead to a positive feedback cycle of further subsidence and sediment accumulation extending outwards from the initial focus of subsidence. Because of the rigidity of the lithosphere, down-flexure may extend for 150 km or more beyond the area of sediment loading. While subsidence is maintained by sediment loading, the surrounding upland areas supplying sediment will rise isostatically through unloading. Once initiated such a complementary system of uplift and subsidence is probably self-sustaining for long periods of time in the absence of major tectonic disturbance. In enclosed basins water-loading by lakes may further contribute to subsidence, though their effect will be far more ephemeral than sediment loading as they are likely to undergo repeated phases of growth and desiccation due to changes in climate.

4.5 Passive continental margins

The sequence of uplift, rifting and continental rupture which leads to the formation of passive continental margins is of considerable importance in any attempt to understand the large-scale geomorphology of the present-day continents. Moreover, passive margins have assumed great economic significance since the 1960s as their hydrocarbon potential has begun to be extensively exploited by the oil industry. The break-up of Pangaea (Fig. 2.16) generated a considerable length of passive continental margins and analysis of the palaeomagnetic record of the continents and the subsequently created ocean floor has given us a fairly good idea of the timing of most of these rifting episodes.

A feature common to many, but certainly not all, passive continental margins is a broad upwarp which separates the coast from interior basins. Such upwarps consist of broad swells running parallel to the coast and are in most cases flanked on their oceanward side by a great escarpment. When rifting leads to the rupture of a continent the original elevation of the landsurface above sea level prior to rifting inevitably creates some relief once a new base level is established. But a number of passive margins appear to have experienced uplift either during or soon after rifting which has led to the development of continental-margin upwarps which stand higher than landsurfaces further inland. Clearly models of passive margin evolution must be able to account for the development of these upwarps.

Our understanding of the vertical (as opposed to the horizontal) movements of the lithosphere during continental rupture is largely derived from the interpretation of the sediments laid down in passive margin basins. Knowledge of these sedimentary sequences, which are now almost entirely located below sea level, has been greatly enhanced by the application of seismic stratigraphy (see Section 4.3.1). Together with evidence from boreholes drilled through continental-margin deposits, investigations using seismic stratigraphy have shown that most passive margins support a wedge of sediment which becomes gradually thinner as it passes seawards from continental to oceanic lithosphere (Fig. 4.14). Deposits laid down during the rifting phase are termed **synrift sediments** while those laid down on the margin once continental separation has occurred are termed **postrift sediments.** Interpretations of these sedimentary sequences in conjunction with morphological and structural evidence has led to the development of numerous models of the tectonic evolution of passive margins.

4.5.1 Active rifting

In the active rifting model (see Section 4.3.2) rifts propagate between a series of domal uplifts formed above sub-lithospheric thermal anomalies (Fig. 4.15(A)). This gives rise to a sequence of domes and intervening saddles which, as the break-up of the continent proceeds and a new ocean is created, are split to form a sequence of ruptured domes and troughs along the newly formed passive margins. At this early stage of development these new continental margins are termed **nascent passive margins** (Fig. 4.15(B)).

As sea-floor spreading continues the margin ages and subsidence begins to predominate over uplift, at least along its oceanward flank (Fig. 4.15(C)). Such subsidence eventually leads to the submergence of the oceanward part of the margin and the formation of a continental shelf. Subsidence during this **mature passive margin** stage is driven by both **thermal subsidence** arising from the cooling of the margin as it moves away from the region of mantle upwelling

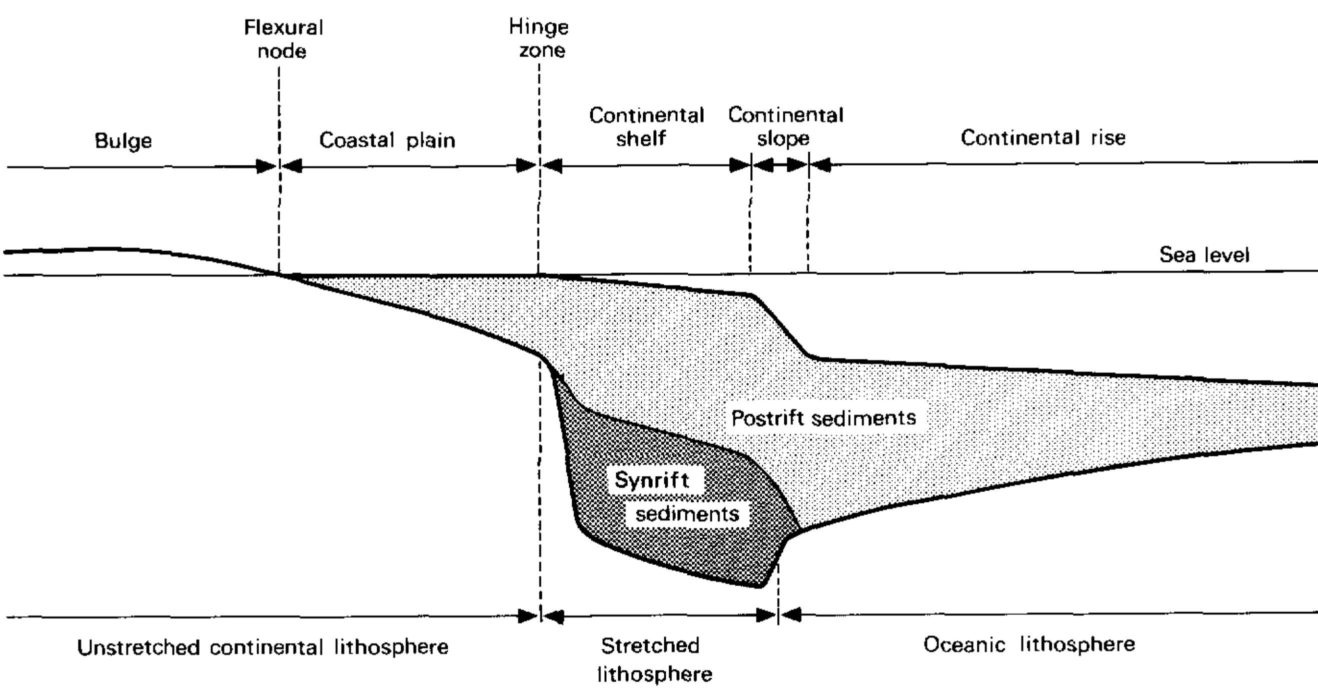


Fig. 4.14 The major structural and morphological elements of passive (rifted) continental margins. Synrift sediments are those deposited in a rift prior to continental rupture, whereas postrift sediments date from the initiation of sea-floor spreading. The ***hinge zone*** *marks the line along the margin dividing continental lithosphere of normal thickness from lithosphere which has been thinned and stretched during rifting and subsequently subsided under the load of overlying sediments. The* ***flexural node*** *separates the (oceanward) portion of the margin subject to subsidence from that experiencing uplift. (Modified from A. B. Watts (1982)* Nature ***297****, Fig. 2, p. 470.)*

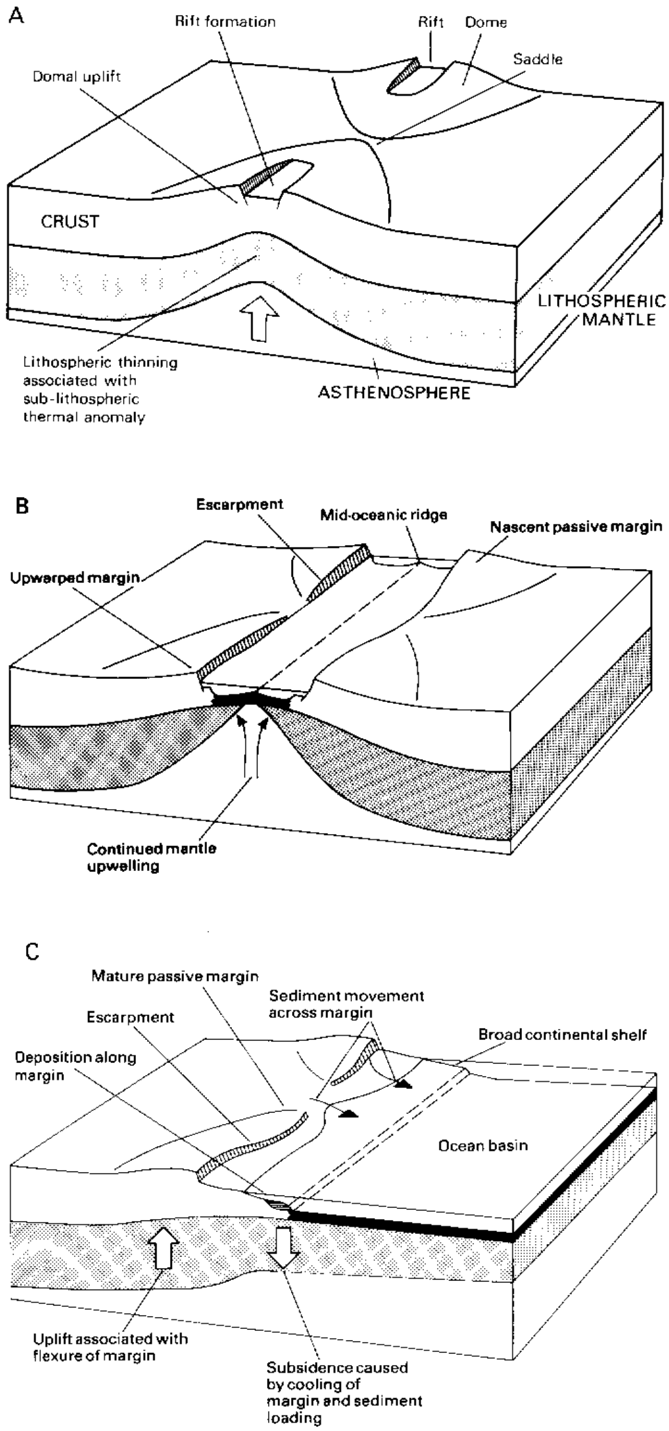


Fig. 4.15 Schematic representation of passive margin evolution through active rifting showing the initial uplift and rifting (A), nascent passive margin (B) and mature passive margin (C) stages.

located at the site of rifting, and by the isostatic loading caused by the growing wedge of sediment accumulating offshore. This sediment loading is coupled with isostatic unloading on the landward part of the margin as material is transported offshore.

Some models of passive margin tectonics assume that the isostatic compensation in response to these changes in load is local rather than regional, in which case no net uplift of the landsurface occurs (see Section 4.2.3). But if the lithosphere along the margin has sufficient rigidity the resulting imbalance in load will be accommodated through flexural isostasy. This will lead to a net uplift of landsurfaces along the least eroded part of the margin, that is, inland of major escarpments. The magnitude of this effect is difficult to estimate, but it will depend in part on the flexural rigidity of the lithosphere. Nevertheless, flexural isostasy provides a possible uplift mechanism for passive margins during their mature stage of development when the effects of the initial thermal uplift have diminished.

Unless the zones of sub-lithospheric mantle upwelling which precipitate uplift and rifting are in a line, sheared margins will develop along zones of transform plate motion linking offset domal upwarps (Fig. 4.16(A)). Modelling by geophysicists indicates that a 200 km diameter convective upwelling could produce a domal uplift about 1000 km across with a maximum elevation of around 1000 m. Rift arms at triple junctions which fail to spread will form aulacogens striking at a high angle into the continental margin (Fig. 4.16(B), (C)).

The extensive series of domal uplifts traversed by the East African Rift System has been widely regarded as a clear example of active rifting with uplift preceding the development of rift structures. Some recent work, however, has cast doubt on this interpretation and suggests instead that rifting may have been initiated before uplift. The evolution of the nascent rifted margins of the Red Sea has also been interpreted both in terms of active and passive rifting mechanisms.

4.5.2 Passive rifting

The idea that simple extension of the lithosphere can produce the structural features observed along passive margins was first presented by D. McKenzie in 1978. In this initial presentation of the passive rifting model it was assumed that the amount of extension was uniform throughout the lithosphere (Fig. 4.17(A)). Calculations based on this assumption, however, indicated that any uplift would be confined to the region over which stretching occurs and that no uplift at all would occur if the thickness of the crust were less than about 20 per cent of the thickness of the lithosphere. In fact this **uniform extension** model predicted that in most geologically reasonable situations subsidence caused by thinning of the lithosphere through stretching would outweigh any thermal isostatic uplift arising from the partial replacement of cool lithosphere by hot asthenosphere below the zone of extension.

Since this simple model was unable to explain the kind of uplift observed along rifted margins there were immediate attempts to modify it. One way the model was altered was to assume that the the mantle part of the lithosphere stretches

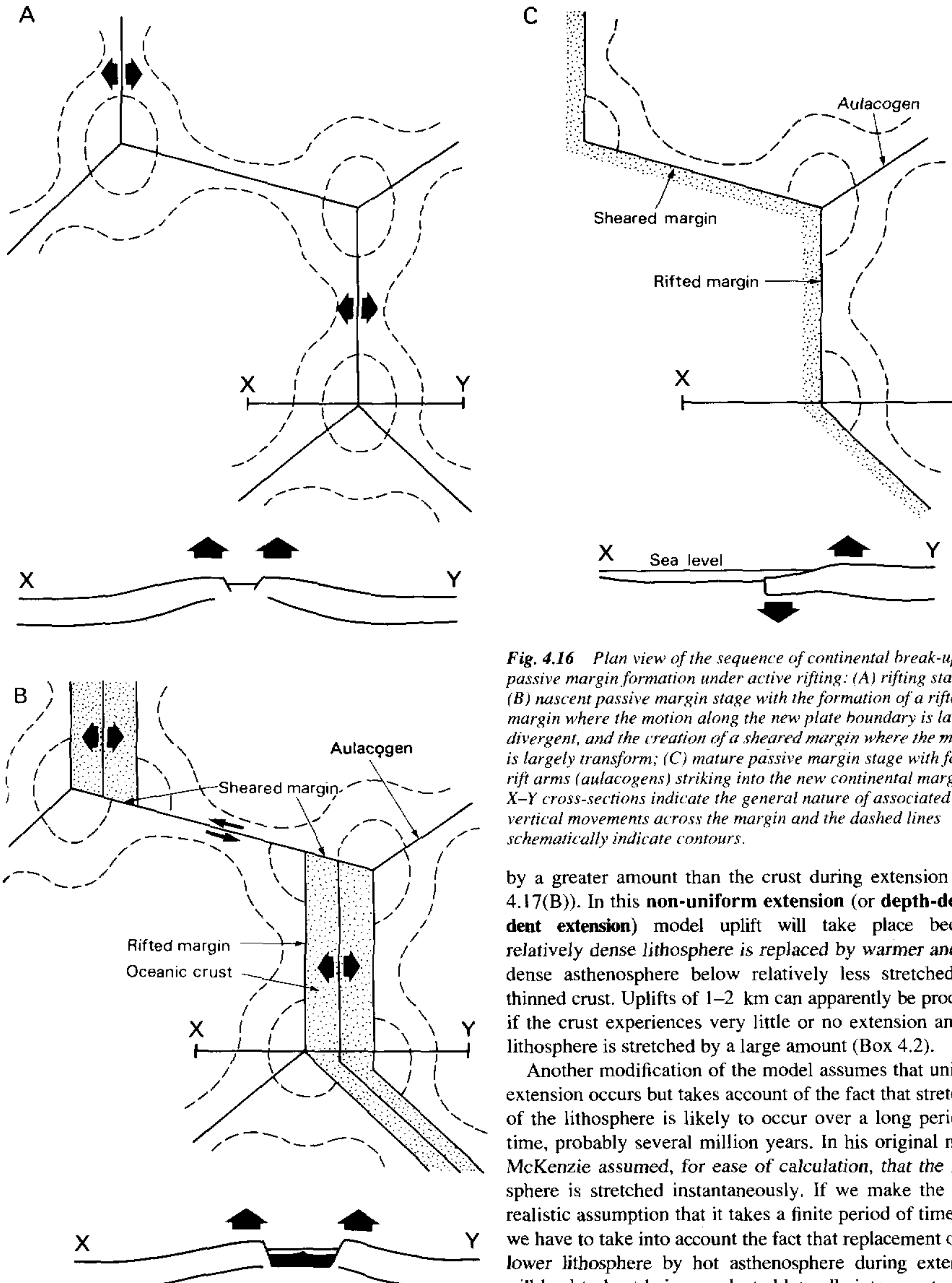


Fig. 4.16 Plan view of the sequence of continental break-up and passive margin formation under active rifting: (A) rifting stage; (B) nascent passive margin stage with the formation of a rifted margin where the motion along the new plate boundary is largely divergent, and the creation of a sheared margin where the motion is largely transform; (C) mature passive margin stage with failed rift arms (aulacogens) striking into the new continental margin. X–Y cross-sections indicate the general nature of associated vertical movements across the margin and the dashed lines schematically indicate contours.

by a greater amount than the crust during extension (Fig. 4.17(B)). In this **non-uniform extension** (or **depth-dependent extension**) model uplift will take place because relatively dense lithosphere is replaced by warmer and less dense asthenosphere below relatively less stretched and thinned crust. Uplifts of 1–2 km can apparently be produced if the crust experiences very little or no extension and the lithosphere is stretched by a large amount (Box 4.2).

Another modification of the model assumes that uniform extension occurs but takes account of the fact that stretching of the lithosphere is likely to occur over a long period of time, probably several million years. In his original model McKenzie assumed, for ease of calculation, that the lithosphere is stretched instantaneously. If we make the more realistic assumption that it takes a finite period of time then we have to take into account the fact that replacement of the lower lithosphere by hot asthenosphere during extension will lead to heat being conducted laterally into unextended,

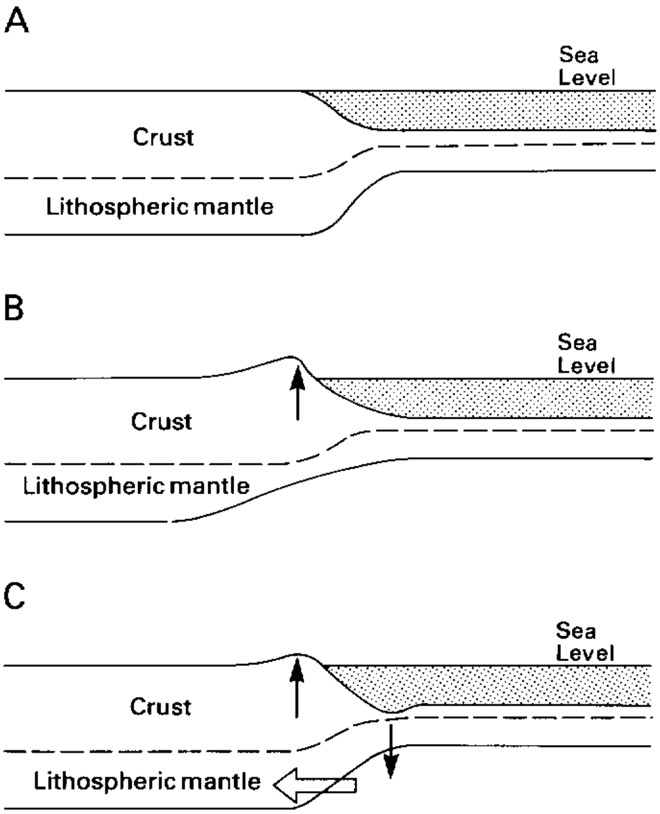


Fig. 4.17 Response of rifted margins to different modes of passive rifting: (A) (instantaneous) uniform extension; (B) non-uniform (depth-dependent) extension in which the sub-crustal lithosphere stretches by a greater amount than the crust and uplift occurs in the zone of relatively unthinned crust; (C) uniform extension over a finite period of time with uplift promoted by lateral heat flow into adjacent unthinned lithosphere. (Modified from C. Beaumont et al. *(1982)* Philosophical Transactions of the Royal Society London ***A305***, Fig. 10, p. 310.)*

Box 4.2 Uplift by non-uniform extension

We can calculate the maximum uplift to be expected during non-uniform, or depth-dependent, extension if we know how much the sub-crustal lithosphere has been stretched in regions where the overlying crust has experienced little or no extension. It is in fact often difficult to determine these stretching factors empirically, but we can make some calculations using geologically reasonable values. The maximum uplift (Δh) is given by

$$\Delta h = \frac{0.5\rho_m \alpha (T_a - T_s)\, l_l}{\rho_c}\left(1 - \frac{1}{\beta_L}\right)$$

where l_l is the initial thickness of the lithosphere, β_L the proportionate amount of stretching experienced by the sub-crustal lithosphere, ρ_m and ρ_c the densities of the mantle (3300 kg m^{-3}) and crust (2700 kg m^{-3}) respectively, $T_a - T_s$ the difference in temperature between the base and the top of the lithosphere (1300 °C) and α the volumetric coefficient of thermal expansion (that is, the proportionate change in volume accompanying a change in temperature) ($\alpha = 3.4 \times 10^{-5}$ °C^{-1}).

If we assume that the initial lithospheric thickness is 150 km and the sub-crustal lithosphere is stretched by a factor (β_L) of 1.1. (that is, its 'length' is increased by 10 per cent) then the maximum uplift in unstretched overlying crust will be about 0.37 km. But if we increase the stretching factor we obtain greater changes in surface elevation; thus for β_L = 1.2 we get 0.68 km and for β_L = 1.5 we get 1.35 km. These higher figures are comparable to the amount of uplift observed on a number of passive margins. The initial lithospheric thickness also affects the ultimate amount of uplift generated in unstretched overlying crust. For example, if the initial lithospheric thickness is only 100 km where the stretching factor (β_L) is 1.5, the uplift is reduced to 0.9 km.

and therefore unthinned, lithosphere, as well as vertically into the overlying upper lithosphere (Fig. 4.17 (C)). Depending on the amount of heat involved this mechanism could lead to a significant uplift through thermal isostasy due to the heating of this unthinned lithosphere. Since this heat transfer will occur only slowly the maximum uplift of the passive margin would probably occur as long as 60 Ma after rifting; this contrasts with the non-uniform extension model in which uplift takes place concurrently with stretching and rifting.

A further approach to explaining the development of significant continental-margin upwarps as a consequence of passive rifting has been to invoke a process known as **secondary convection.** In this model it is proposed that thinning of the lithosphere and its partial replacement by hot asthenosphere might allow convective upwelling to become established below the zone of extension. It is envisaged that this active mantle upwelling would lead to a significant heating of the overlying lithosphere and could generate uplifts of up to around 1 km. The difference between this model and active rifting is that the convective upwelling is a consequence, not a cause, of lithospheric thinning and rifting. The secondary convection model has been applied to the Gulf of Suez at the northern end of the Red Sea. Here significant uplift of rift flanks has occurred although there is convincing evidence (from the deposition of marine sediments at the time of rifting) that rifting preceded uplift.

The sequence of rifted margin evolution to be expected from passive rifting is illustrated schematically in Figure 4.18. This figure is non-committal as to the precise mechanism likely to lead to the generation of passive margin upwarps as it illustrates secondary convection as well as non-uniform (depth-dependent extension) and the effects of lateral heat flow into unthinned lithosphere. As is evident from comparing Figures 4.15 and 4.18, by the mature stage continental margins formed by both active and passive rifting tend to assume a similar mode of evolution, with subsidence, at least in the offshore zone, predominating over uplift as the margin both cools and is loaded by sediment shed from the adjacent land mass.

A further component of passive rifting that needs to be considered is whether a rifting event is accompanied by

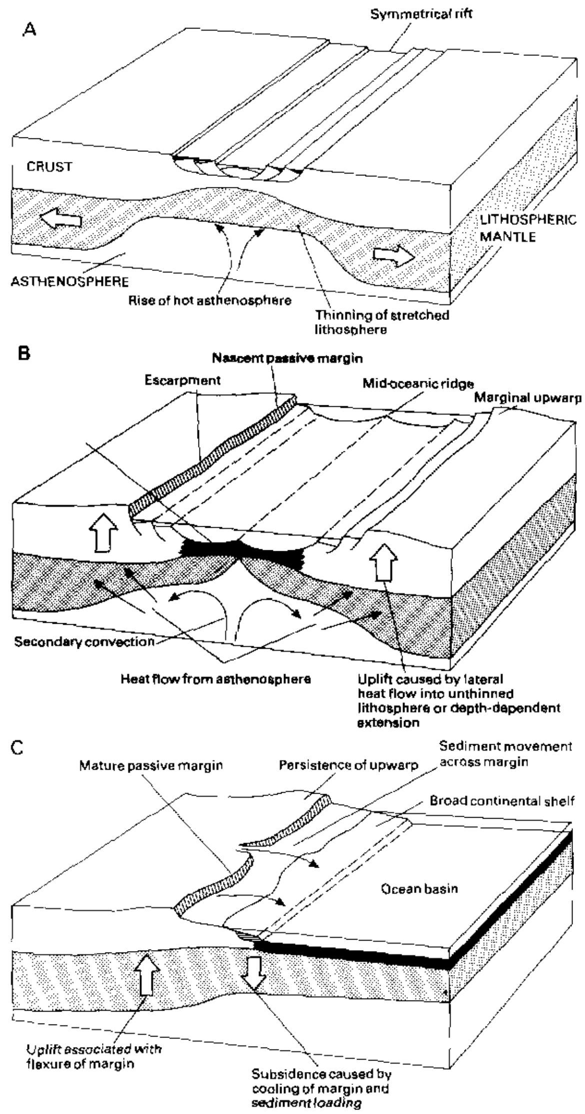


Fig. 4.18 *Schematic representation of passive margin evolution through symmetric passive rifting showing the initial rifting (A), nascent passive margin (B) and mature passive margin (C) stages.*

abundant volcanism. It has been pointed out that enormous thicknesses of basaltic lavas erupted during the early stages of rifting characterize some passive margins, such as those of eastern Brazil, western India and the south-eastern part of southern Africa. A number of other passive margins, however, show no sign of such major volcanism. This observation has led to the suggestion that whether a passive margin is volcanic or non-volcanic depends on whether the site of rifting happens to coincide with the location of a mantle plume and its associated hot spot.

Where rifting occurs over a mantle plume the sub-lithospheric mantle will be up to 200 °C hotter than normal and large quantities of magma will be generated. This occurs because rifting of the lithosphere allows hot mantle rock to move towards the surface. The resulting reduction in pressure, or decompression, experienced by this material causes it to melt. Some of the magma generated is erupted on to the surface to form enormous lava flows (see Chapter 5), but by far the largest proportion is accreted to the base of the crust, a process known (somewhat misleadingly) as **underplating**. According to the underplating model the addition of volcanic rock thickens the crust and the resulting isostatic adjustment leads to the formation of a broad hot-spot swell up to 2000 km across and an increase in surface elevation of up to 2000 m. It is indeed interesting to note that the highest passive margins also seem to be those characterized by rift-related volcanism, just as the model predicts. Although mantle plumes do not drive the rifting in this model, the surface uplift caused by underplating would assist the rifting process as diverging plates would tend to slide down either side of a hot-spot swell.

All the models of passive rifting discussed so far assume that rifting is more or less a symmetric process; that is, we would expect the opposing passive margins formed through continental break-up to have a similar structure and morphology because they have experienced a similar tectonic history. But does symmetric rifting accord with the evidence? We saw in our discussion of continental rifts (see Section 4.3) that asymmetric rather than symmetric rifting appears to be the norm, at least at the scale of individual rift structures. Moreover, if we compare opposing passive margins there seem to be a number of cases where their morphology and structure are very different. For example the western margin of southern Africa has a relatively narrow continental shelf and a well-developed marginal upwarp, whereas the opposing margin of the eastern coast of southern South America to which it was joined prior to the break-up of Gondwana has a very broad continental shelf and no significant marginal upwarp. Another instance of apparent passive margin asymmetry is provided by the eastern margin of Australia, which has a marked upwarp which forms the Great Dividing Range, and the Lord Howe Rise which represents a now submerged fragment of continental crust which rifted away from eastern Australia around 95 Ma BP.

Studies of extensional terranes, such as the Basin and Range Province of the south-western USA, using seismic methods have revealed shallow-dipping faults which appear to extend through the entire lithosphere. These have been interpreted as shear zones representing surfaces of lithospheric detachment. If lithospheric extension occurs primarily along such **detachment faults** rather than by thinning throughout the entire zone of lithospheric extension, then we would expect opposing passive margins to display a marked complementary asymmetry (Fig. 4.19).

Such detachment models predict that two types of passive margin will be produced by continental rupture. An **upper-plate margin** is formed by crust lying above the detachment

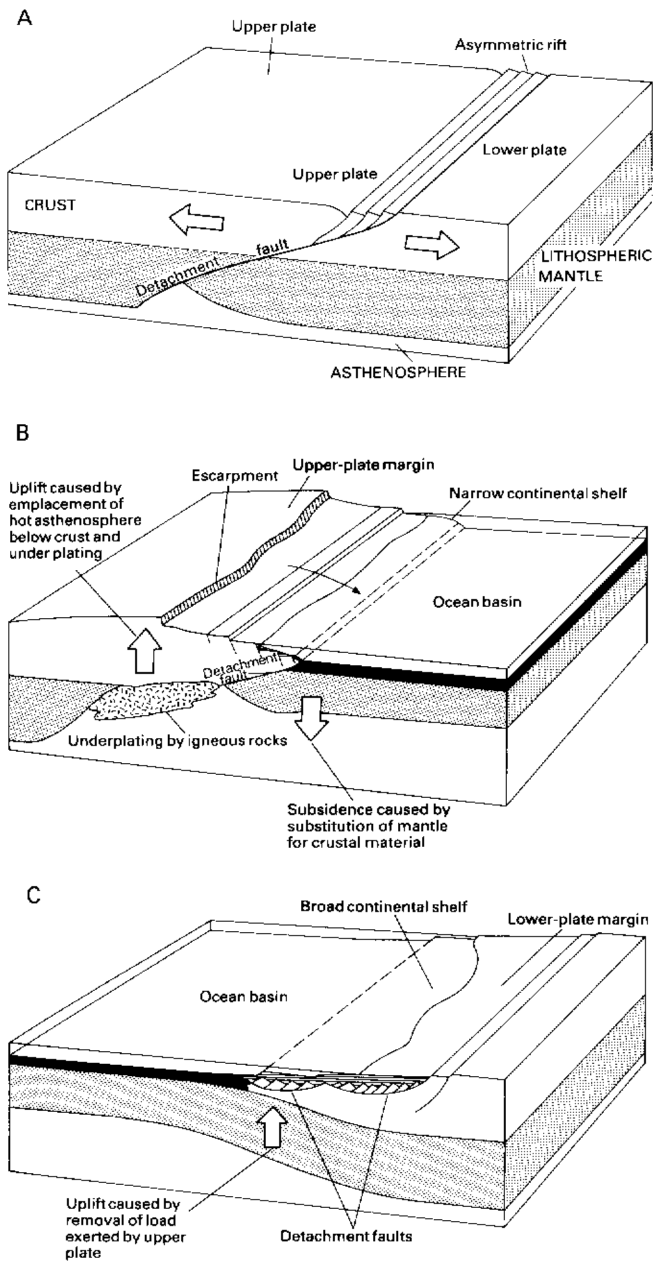


Fig. 4.19 Schematic representation of the detachment model of asymmetric passive margin evolution. Asymmetric rifting (A) leads to the development of morphologically and structurally distinct upper-plate (B) and lower-plate (C) margins.

fault while a **lower-plate margin** is developed on deeper crustal rocks lying below the detachment zone and which may be overlain by faulted fragments of the upper plate. Upper- and lower-plate margins will differ significantly in their uplift and subsidence history.

As displacement occurs along the detachment fault the margin of the lower plate will be upwarped as the load previously exerted by the upper plate is progressively removed (Fig. 4.19(C)). The predominant effect, however, will be subsidence due to the considerable thinning of the lower plate lithosphere. The resulting continental margin morphology is likely to consist of a broad continental shelf and a rather modest amount of uplift inland.

The behaviour of the upper plate is quite different. The pulling away of the lower plate along the detachment fault exposes the base of the lithosphere on the upper plate to hot, rising asthenosphere (Fig. 4.19(B)). This substitution of less dense asthenosphere for denser lithosphere will result in uplift of the landsurface on the upper-plate margin. Oceanward of this uplift zone, however, there will be subsidence since here movement along the detachment fault results in the substitution of mantle for less dense lower crust. The coupling of this offshore subsidence and onshore uplift will induce a flexure of the margin which will accentuate uplift inland. The resulting upper-plate margin morphology will comprise a broad marginal upwarp which drops abruptly down to a relatively narrow continental shelf.

Superimposed on all these models of passive margin rifting is the possible contribution to uplift and subsidence made by flexural isostasy as material is eroded from the margin and transported offshore. As with the active rifting model this effect could generate uplift of the margin, even during its mature phase when other mechanisms are largely inoperative.

4.5.3 Passive margins: a key research focus

The modelling of passive margin evolution is still in its infancy. A wide range of mechanisms have been proposed by geophysicists attempting to model the broad patterns of uplift and subsidence observed but modelling has run well ahead of empirical testing. Figure 4.20 summarizes some of the key processes that appear to be involved. To date relatively little attention has been paid to the detailed morphological evidence that might indicate which, if any, of the proposed models is applicable to any particular passive margin. Nevertheless, it is impossible to underestimate the

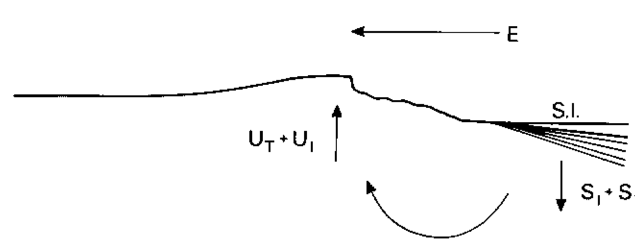


Fig. 4.20 Summary of some of the major tectonic factors controlling the morphological evolution of rifted passive margins: U_T = thermally-driven uplift; U_I = isostatic uplift associated with denudational unloading; S_T = thermally driven subsidence; S_I = isostatic subsidence associated with sediment loading; r = rotation of the margin resulting from U_I and S_I; E = erosion by escarpment retreat; S.l. = sea level.

importance of passive margins to an understanding of long-term landform development, a topic we will be returning to in Chapter 18. It seems probable that many of the models outlined above will be found to be applicable to at least some passive margins, but much work remains to be done in this field. This is especially true of sheared margins which have been neglected by geophysicists. Their attention has been focused on rifted margins largely because of their potential oil and gas reserves.

4.6 The break-up of supercontinents

Discussion about the way continental rifts form and how, in some cases, they develop into sites of continental rupture and sea-floor spreading raises the much broader question of why supercontinents break up. At least part of the answer to this question may lie in the contrasting properties of continental and oceanic crust. Continental crust is only about one-half as efficient at conducting heat as oceanic crust, so if a **supercontinent**, such as Pangaea, covers a significant part of the Earth's surface heat will build up in the mantle below it. After perhaps 80 Ma or so the increase in the temperature in the sub-lithospheric mantle below the supercontinent will lead to uplift, rifting and continental fragmentation. New ocean basins are created which become wider through time but eventually subduction zones will become established, either along previously passive margins, or within ocean basins. This establishment of subduction zones results in a change from continental dispersal to continental aggregation which eventually leads to the reassembly of a large land mass.

This idea of a repetitive sequence of continental fragmentation and aggregation is termed the **Wilson cycle** after its originator J. T. Wilson. The average duration of each cycle seems, on the basis of geological evidence, to be very roughly around 500 Ma. This is well beyond the time scale generally relevant to landform development but it is still important to be aware of how present landscapes, especially in plate interiors, relate to the most recent episode of supercontinent fragmentation. For instance, dispersal patterns of the continents following the break-up of Pangaea show that the now separate continents have been moving away from Africa, which itself appears to have remained relatively stationary. In other words Africa may have remained close to the region of major sub-lithospheric heating which preceded the break-up of Pangaea and this may explain why it has a high mean elevation today in spite of lacking significant orogenic belts. Australia, on the other hand, has moved rapidly away from its original location on the edge of Pangaea and it has the lowest mean elevation of the major continents. There is obviously much uncertainty in these broad generalizations, but geomorphic evidence has a key role to play in relating such global tectonic models to the geological history of the continents.

Further reading

The large-scale relationships between landforms and tectonics in plate interiors have received comparatively little attention from geomorphologists so the literature from this perspective is rather limited. King (1967) is still useful for its descriptions of the continental landscapes of plate interiors although his interpretations of the tectonic processes involved have now been superseded by more recent work. Ollier (1981) provides an introduction to several topics discussed in this chapter, and my own brief reviews in *Progress in Physical Geography* (Summerfield, 1986, 1988, 1989) try to relate some of the work by geophysicists to problems of landscape development. The morphology of passive margins is discussed in general terms by Ollier (1985), and with respect to specific regions by Ollier (1982) (eastern Australia), Summerfield (1985a) (Africa), Ollier and Marker (1985) (southern Africa) and Ollier and Powar (1985) (India).

There is an extensive and rapidly growing literature on the tectonics of plate interiors, but unfortunately much of it is written from a geophysical rather than a geomorphic perspective and it is not readily accessible to readers lacking a good grounding in physics and maths. Two general reviews of the field presented at an introductory level are Bally *et al.* (1980) and Burchfiel *et al.* (1980). Both of these edited volumes contain a number of relevant and clearly written papers on a variety of specific topics. A more detailed treatment of epeirogenic uplift is to be found in McGetchin and Merrill (1979); in spite of its general title it is almost solely concerned with the Colorado Plateau, but it does contain a brief but wide-ranging review of uplift mechanisms. An interesting assessment of the differential uplift of the continents through the Late Cenozoic is provided by Bond (1978).

A clear introduction to hot spots is provided by Burke and Wilson (1976), while a more recent assessment of their global distribution is that by Stefanick and Jurdy (1984). Hot spots as a cause of uplift are discussed by Crough (1983) and Gass *et al.* (1978), while Theissen *et al.* (1979), Summerfield (1985b) and Sahagian (1988) examine the relationship between epeirogeny and hot spot activity in Africa. The important, but neglected, problem of the range of scales of uplift associated with thermal mechanisms is addressed by Le Bas (1980), and the idea of lithospheric vulnerability is presented by Pollack *et al.* (1981). Crough (1979) explores the possible morphological consequences of continents migrating over hot spots and also cites a possible example from North America (Crough, 1981).

Turning to other possible uplift mechanisms, the phase change model is considered by Smith (1982) with particular reference to Africa and Australia, and by Smith and Drewry (1984) with respect to the Transantarctic Mountains. The mechanism of delamination is described by Bird (1979). Flexural isostasy is discussed by Karner and Watts (1982) and Watts *et al.* (1982), and at a more introductory level by Watts

(1981), but unfortunately these treatments focus on offshore subsidence rather than uplift inland.

Rift systems have been studied for many decades but have received renewed attention since the early 1970s because of their association with crustal upwarps and their role in continental rupture. An accessible introduction is provided by Pàlmason (1982); this is an edited compilation containing a brief but stimulating review of continental rifts by Mohr (1982). More recent ideas about the half-graben structure of many rifts are introduced by Frostick and Reid (1987) and examined in depth by Rosendahl (1987) and in Frostick *et al.* (1986). The influence of pre-rift structures on eventual rift form is considered by Versfelt and Rosendahl (1989) and the relationship between rifts and domal uplift in Africa is discussed by Burke and Whiteman (1973).

Much of the research into the tectonic evolution of basins overlaps with work on passive margin development, and although written for students of sedimentology the book by Miall (1984) provides a valuable introduction for geomorphologists on both of these themes. Bally and Snelson (1980) provide an extremely comprehensive review of basin types and origins and Turcotte (1980) presents one of the more accessible reviews of the geophysics of basin subsidence.

A good introduction to the tectonics of passive continental margins is provided by Scrutton (1982) while Courtillot and Vink (1983) and Bonatti (1987) look at the processes of continental rifting and break-up. The extent and magnitude of uplift to be expected from active and passive rifting mechanisms are compared by Keen (1985), while Beaumont *et al.* (1982) and Buck *et al.* (1988) assess the patterns of uplift likely to result from different models of passive rifting. The uniform extension model is presented by McKenzie (1978) and the idea of non-uniform stretching is clearly described by Rowley and Sahagian (1986). Steckler (1985) proposes secondary convection as an explanation of the uplift in the Gulf of Suez region, White and McKenzie (1989) outline the underplating model, and the asymmetric detachment model of continental break-up is presented by Lister *et al.* (1986). The eastern Australian passive margin is one region where there has been a fruitful interchange of ideas and data between geomorphologists and geophysicists, although this has not resulted in a consensus as to either the history of uplift or the mechanisms that have caused it. Veevers (1984) provides a comprehensive background to the history of the eastern Australian margin while Bishop (1988) reviews the various models that have been proposed to account for uplift. Lambeck and Stephenson (1986) and Wellman (1987, 1988) provide contrasting interpretations of the uplift mechanisms that have operated in the region.

Finally, the possible mechanisms underlying the cyclic fragmentation and reassembly of continents are discussed at an introductory level by Nance *et al.* (1988) and in detail by Gurnis (1988). Both these discussions derive ideas from earlier work by Anderson (1982).

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