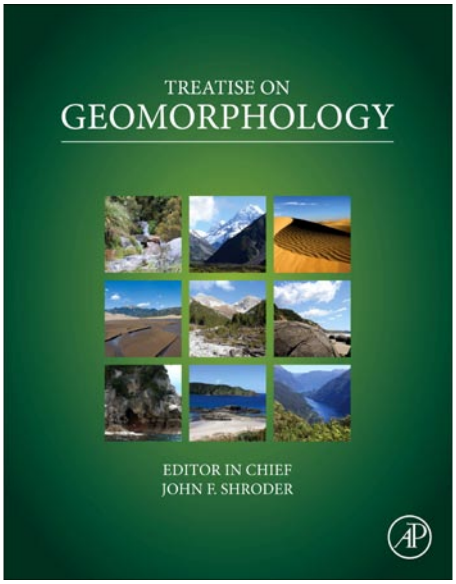

Blenkinsop T., and Moore A. (2013) Tectonic Geomorphology of Passive Margins and Continental Hinterlands. In: John F. Shroder (ed.) Treatise on Geomorphology, Volume 5, pp. 71-92. San Diego: Academic Press.

5.5 Tectonic Geomorphology of Passive Margins and Continental Hinterlands

T Blenkinsop, School of Earth and Environmental Sciences James Cook University, Townsville, QLD, Australia
A Moore, African Queen Mines Ltd., Botswana, South Africa, and Rhodes University, Grahamstown, South Africa

Abstract

Passive margins, created at the margins of rifted continents, are affected by thermal, isostatic, flexural, buckling (plate tectonic), and dynamic (mantle) stresses. Variations among these give rise to diverse topographic expressions. The geometry of rifting also has a major effect on topography. Thus, many low-relief margins lacking a fringing escarpment occur at the failed arms of triple junctions. Variations in lithospheric flexural rigidity influence the response of passive margins following rifting. Postrifting evolution of the continental hinterland is more readily explained by stresses related to plate tectonic processes than by the dynamic uplift over plumes.

5.5.1 Introduction

Passive margins form where continental rifting results in plate divergence and the formation of new oceanic crust. They constitute ~50% of continental margins today (**Figure 1**; Gallagher and Brown, 1999), and they are highly significant to humankind, both as dwelling places for a large proportion of the world's population and as repositories of vital resources – most notably hydrocarbons. Passive margins separate oceanic crust from a continental hinterland, which has a history that is closely linked to that of the adjacent margin (e.g., Summerfield, 2000). Plate tectonics provides a first-order framework for understanding the formation of these features (Bishop, 2007), and yet, the evolution of their tectonic geomorphology can only be linked indirectly to plate tectonic processes because they are commonly remote from plate boundaries.

Some characteristic features of passive margins associated with extrusive igneous rocks are illustrated in **Figure 2(a)**. The shelf break, or slope, approximately at the continent–ocean boundary, appears to mark the maximum Neogene sea lowstand linked to Plio–Pleistocene glacial advances, and joins the continental shelf to the abyssal depths of the ocean. The continent–ocean boundary is not always very well defined (Brown et al., 2003), but at least conceptually, it corresponds to the point at which continuous vertical dikes of oceanic crust are joined to some continental crust. The shelf break is separated from the exposed continent by the continental shelf, which consists of a sedimentary pile of variable thickness underlain by continental crust. Ocean-dipping normal faults have syn-tectonic sediments in their hanging walls; the tops of tilted fault blocks may be eroded. The faults may be listric, and rotate the sediments in their hanging walls (**Figure 2**). Postrift sediments can overlie these features, and these sequences may themselves be tilted and eroded. Continental margin sediments form seaward-dipping seismic reflectors characterized by low P-wave velocities (<7 km s^{-1}) (**Figure 2**). Sedimentary features of the offshore passive margin consist of fans, channels, canyons, and carbonate mounds, with evidence of slumps, debris flows, and regional unconformities (Bagguley and Prosser, 1999; O'Grady et al., 2000; Stoker et al., 2010).

Rifted passive margins develop where disrupted continental fragments diverge essentially perpendicular to the

Blenkinsopand, T., Moore A., 2013. Tectonic geomorphology of passive margins and continental hinterlands. In: Shroder, J. (Editor in Chief), Owen, L.A. (Ed.), Treatise on Geomorphology. Academic Press, San Diego, CA, vol. 5, Tectonic Geomorphology, pp. 71–92.

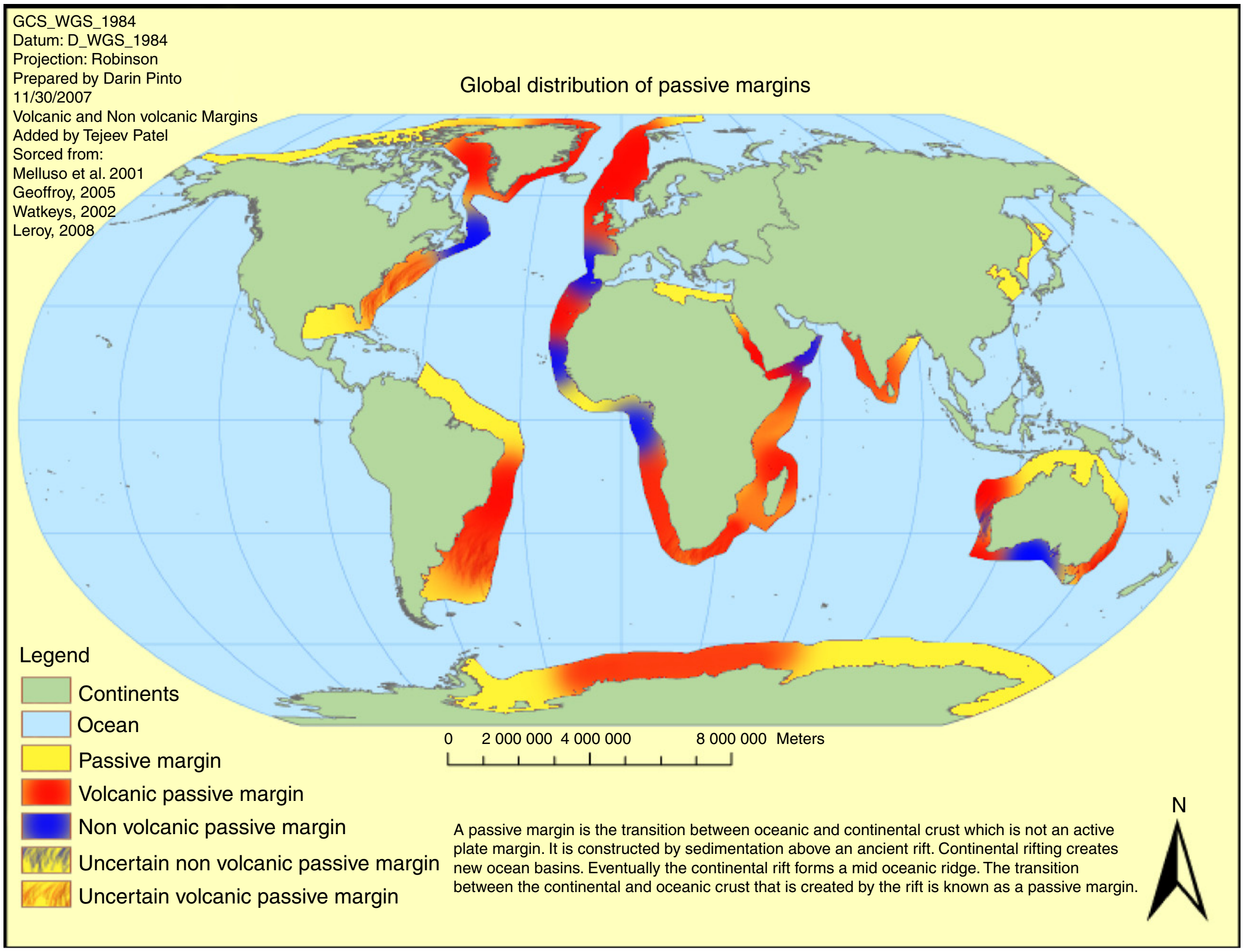

Figure 1 Global distribution and classification of passive margins. http://commons.wikimedia.org/wiki/File:Globald.png

coastline, as for example, the Atlantic coasts of Africa and South America. Sheared passive margins form where separation is along strike–slip faults, as in the case of the separation of the Agulhas Bank from the Falkland plateau along a dextral strike–slip fault (**Figure 3(a)**). These types of margins have high continental shelves and deep sedimentary basins (Brown et al., 2003). Passive margins can also be divided into Nonvolcanic and Volcanic margins (**Figure 1**). The former, also referred to as sedimentary passive margins, develop where rifting and associated crustal stretching and thinning is not accompanied by volcanic activity; they are generally broader than volcanic passive margins (Geoffroy, 2005). Passive margins can indeed be classified by their width: Narrow passive margins are less than 100 km wide, compared with Wide margins with a broad continental shelf (e.g., Brown et al., 2003). These types are well illustrated by contrasts around the Australian continent, where the southeast and northwest Australian margins are typical examples of the narrow margin and the wide margin, respectively.

The majority of classified rifted passive margins are volcanic (**Figure 1**), for example, the margins of southern and eastern Africa and Greenland, and the western margin of India. These passive margins are generally linked to large volumes of basaltic lavas, sometimes with associated acid extrusives, which form Large Igneous Provinces (LIPs). The volcanism may precede continental rifting, as in the case of the Drakensberg in southern Africa (Marsh et al., 1997), or may be broadly coeval, as exemplified by the Deccan Traps (de Wit, 2003), and volcanic activity may persist after breakup. Seaward-dipping reflectors in some volcanic passive margins are now considered to be due to rollover above listric normal faults that dip toward the continent (e.g., Geoffroy, 2005; **Figure 2(b)**). The lower crust of these margins typically has an anomalously high seismic P-wave velocity (7.1–7.8 km s^{-1}), which is interpreted to reflect basic igneous rocks termed Lower Crustal Bodies (LCBs), linked to the volcanic activity, which have been welded to the lower crust by a process termed 'underplating' (McKenzie, 1984). There is evidence that volcanic rifted margins are divided along strike into 50–70-km-long segments, which may have originally represented separate magma chambers (Geoffroy, 2005).

Onshore, passive margins are characterized by diverse topographic expressions. Thus, extensive low-relief coastal plains (Low Elevation passive margins; e.g., Gilchrist and Summerfield, 1990) are a feature of central south Australia, the Argentinean coast of South America, and the west coast of central Africa (Fullard and Darby, 1974). Such topography contrasts markedly with 'High Elevation' passive margins

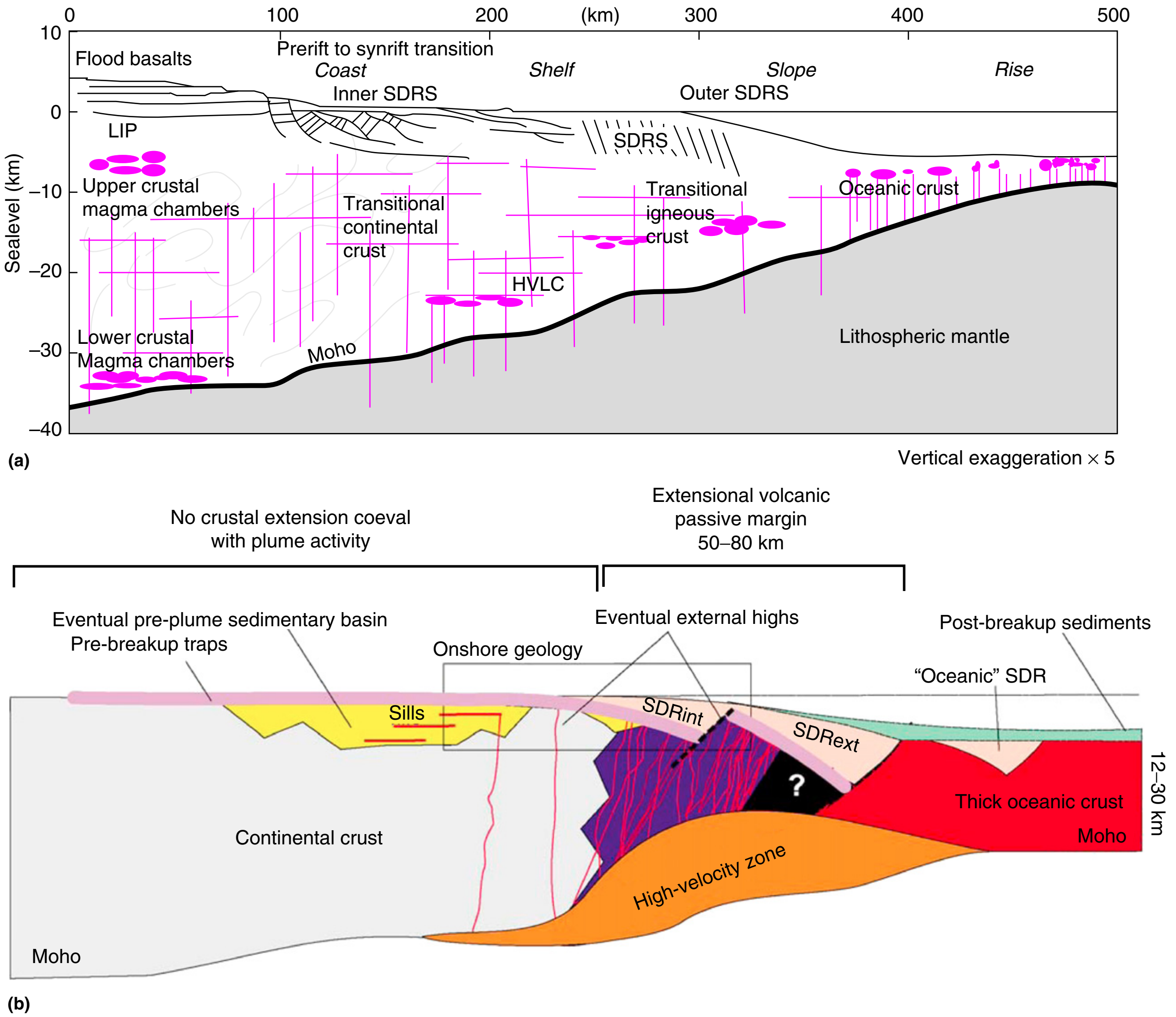

Figure 2 Features of volcanic rifted margins. (a) The SDRS may be due to ocean-dipping listric faults. Reproduced with permission from Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J., 2002. Characteristics of volcanic rifted margins. In: Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J. (Eds.), Volcanic Rifted Margins. Geological Society of America (GSA), Boulder, CO, Special Paper 362, pp. 1–14. HVLC is the high-velocity lower crust, LIP is Large Igneous Province. (b) Alternatively, the SDRS is due to tilting of volcanic rocks and sediments in the hanging walls of continent-dipping listric faults. Reproduced from Geoffroy, L., 2005. Volcanic passive margins. Comptes Rendus Geoscience 337, 1395–1408. Abbreviations: SDRint, internal seaward-dipping reflectors; SDRext, external seaward-dipping reflectors; SDRS, seaward-dipping reflectors series.

with dramatic, commonly continuous, seaward-facing scarps, which are located some 60–200 km inland of the coast. Examples of the latter are India (Ollier and Powar, 1985), the east coast of Australia (Pain, 1985; Ollier, 2004), and the remarkable horseshoe-shaped escarpment that girdles virtually all of southern Africa (Ollier and Marker, 1985) (**Figure 4**), except where breached by the Orange and Limpopo rivers.

It is important to emphasize however, that the nature of the southern African escarpment varies along its length. Thus, the Drakensberg section is characterized by dramatic seaward-facing cliffs (**Figure 4(b)**), whereas elsewhere it may be represented by a marked increase in gradient separating the relatively low-relief coastal plain from the interior plateau (**Figure 4(a)**). In some sections of the escarpment, there is an apparent accordance of summits, which have been inferred to reflect the dissected relics of a Gondwana peneplain, extant before rifting (King, 1976; Lister, 1987). However, this view has been challenged by Partridge and Maud (1987) and Fleming et al. (1999).

The morphology of escarpments can be classified into two groups: Shoulder Type (e.g., the southern African escarpments), in which the escarpment forms a drainage divide between a coastal plain and the continental hinterland, and Arch Type (e.g., southeast Australia), in which the drainage divide is inland of the escarpment (**Figure 5**; Ollier, 1984).

The classification of escarpments by Pazzaglia and Gardner (2000), supplemented by Bishop (2007), distinguishes Type 1 escarpments as those etched into an upland associated with rifting (e.g., southern Africa), Type 2, which are etched into a postrift flexural bulge and are dominated by postrift offshore

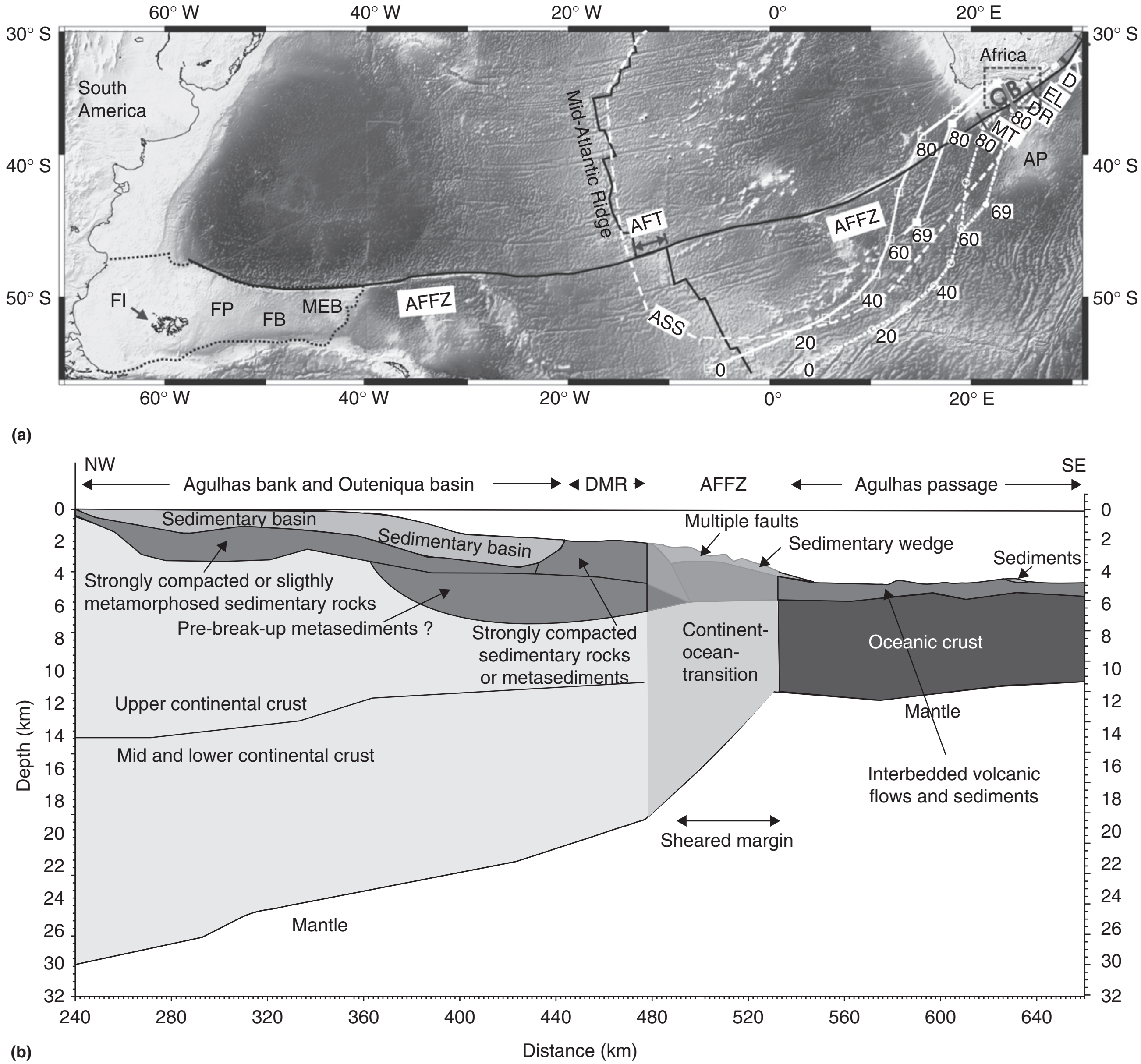

Figure 3 Rifted and Sheared passive margins. (a) The facing margins of South America (north of 47 °S) and Africa (north of 35 °S) are rifted. The margins to the north of the Falkland Plateau (FP) and along the SE coast of Africa are Sheared margins. Reproduced with permission from Parsiegla, N., Gohl, K., Uenzelmann-Neben, G., 2007. Deep crustal structure of the sheared South African continental margin: first results of the Agulhas-Karoo Geoscience Transect. South African Journal of Geology 110, 393–406. FB = Falkland Basin, FI = Falkland Islands, MEB = Maurice Ewing Bank, AFT = Agulhas-Falkland Transform, MT = Mallory Trough segment, DR = Diaz Ridge segment, EL = East London segment, D = Durban segment. (b) Cross-section along the bold gray line (between O and B of the Outeniqua Basin in **Figure 3(a)**) to show the crustal structure of a sheared margin.

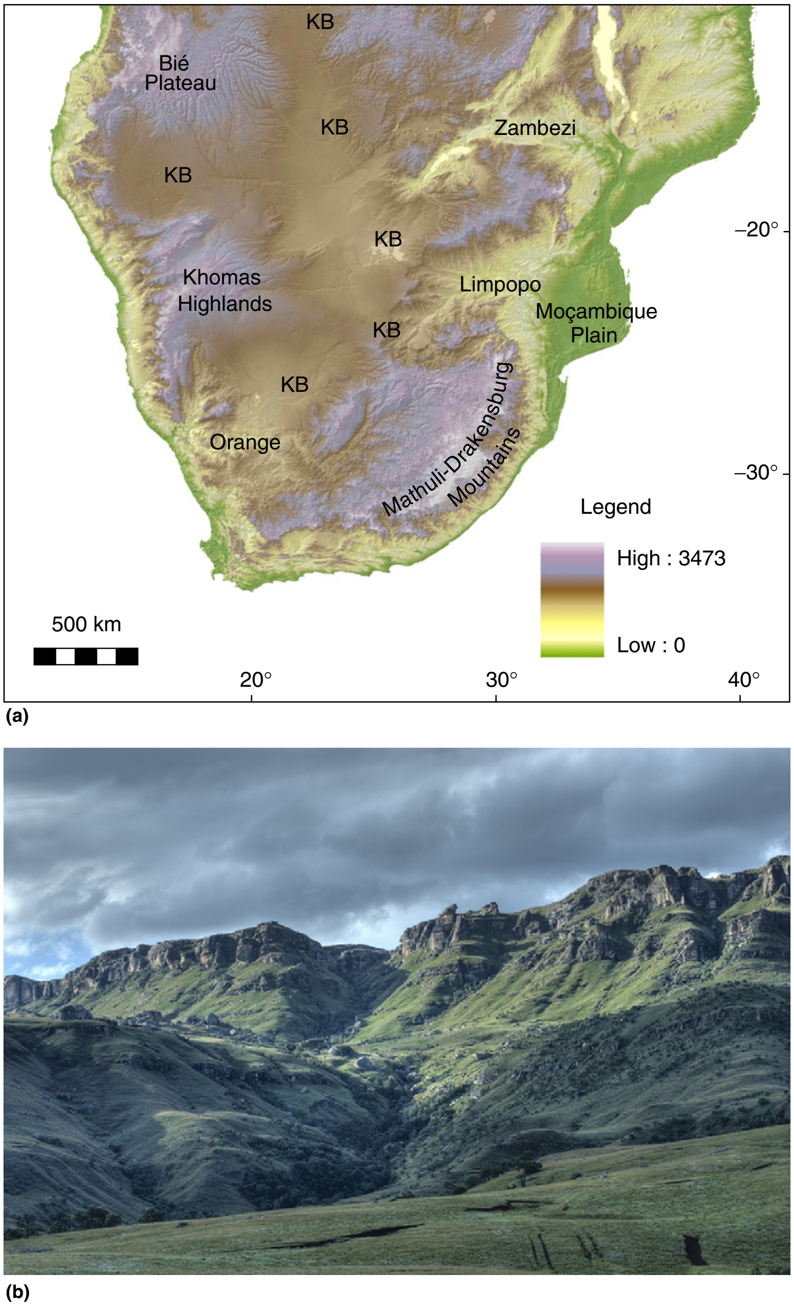

Figure 4 (a) SRTM digital image for southern Africa, illustrating how the low-relief coastal plain and continental interior are separated either by a dramatic escarpment cliff (see **Figure 4(b)**) or by a marked increase in slope. Prominent breaks in the escarpment (labeled) are associated with the Limpopo, Zambezi drainage basins. There is a less pronounced break in the western escarpment, where it is incised by the Orange River (with the major north bank Fish River tributary). (b) Prominent escarpment of the Drakensberg-Maluti mountains (also known as the Mathuli-Drakensberg mountains). Reproduced from African Imagery.

loading (e.g., U.S. Atlantic), and Type 3, an escarpment characterized by embayments and gorges caused by rivers that rise at a preexisting continental divide inboard of the escarpment (e.g., parts of southeast Australia; Nott et al., 1996). This genetic classification embodies some of the central controversies about escarpment evolution. Inland of these dramatic escarpments, the continental hinterland commonly defines a broad, low-relief basin (Ollier, 1984).

The tectonic geomorphology of passive margins and continental hinterlands is primarily concerned with vertical movements of the crust and associated landscape evolution, including epeirogenic (long wavelength) elevation changes. Some of the factors proposed to cause vertical movements are shown in Figure 6.

The aim of this study is to develop a tectonic and geomorphic framework to understand the processes responsible for sculpting the diverse landforms of passive continental margins and continental hinterlands. The study focuses on the origin of the highlands that form major sea-facing escarpments on some margins and how this has impacted landscape evolution. Many examples are taken from southern Africa, which has archetypal status in the study of passive margin and continental interior landscapes.

It is convenient to discuss passive margin and continental hinterland tectonic geomorphology in two separate time frames: in the first, the igneous and tectonic processes associated with rifting that precede continental breakup and generate the prerift topography are discussed. The second time frame deals with postrifting processes. The study of continental hinterlands is most naturally treated after the latter. Despite the above observation that passive margin and continental interior tectonics, dominated by vertical movements, can only be linked indirectly to plate tectonics, in which horizontal displacements are prevalent, it is argued that passive margin and continental interior evolution have to be viewed in the context of underlying plate tectonic controls.

5.5.2 Igneous and Tectonic Processes Associated with Rifting

Modern rifts, such as the East African Rift System (EARS) and Red Sea–Gulf of Suez, are bounded by uplifted marginal shoulders (Steckler and Omar, 1994), and by analogy, similar topography is anticipated to have formed during the early development of other rifted passive margins. A variety of mechanisms, not mutually exclusive, have been proposed for such elevated rift shoulders. Initial rifting may involve upwelling of hot, buoyant asthenosphere, leading to thermal expansion and crustal updoming mainly due to dynamic effects (e.g., White and Mckenzie, 1989). This asthenospheric-related uplift may be preceded or postdated by extension (active vs. passive rifting: e.g., Sengör and Burke, 1978; Turcotte and Emerman, 1983; Huismans et al., 2001). If volcanism accompanied these processes, penetrative magmatism associated with the intrusion of dikes and sills would provide further advective heating (Gilchrist and Summerfield, 1994). Plume acolytes and plume skeptics would join battle over the role of putative mantle plumes in contributing to lithospheric heating and uplift. Uplift of rift flanks can be predicted from models that allow nonuniform extension in the crust and mantle, whether the extension is continuous or discontinuous (Royden and Keen, 1980; Rowley and Sahagian, 1986). However, these effects would be transient, and are envisaged to decay over a timespan of ~60 Ma, so that they cannot account for the existence of escarpments on rifted margins of greater vintage.

Magmatic underplating of the lower crust by basic intrusive igneous rocks, less dense than the underlying ultrabasic mantle (McKenzie, 1984; White and Mckenzie, 1989), could in principle lead to a buoyancy of a rifted margin, thus permanently sustaining an initial thermal uplift (**Figure 6**). However, a number of considerations suggest that underplating may not play a major role in maintaining uplifted margins. Cox (1993) demonstrated that in a tholeitic LIP, the ratio of cumulates to extruded magma would be approximately 0.5, which would imply that a lava thickness of 2000 m (appropriate for the Drakensberg in South Africa) would be associated with ~1000 m of underplating, which would produce a very modest uplift. Far greater amounts of underplating would be implied in the case of the eruption of the estimated 7.5-km-thick succession of rhyolites interbedded

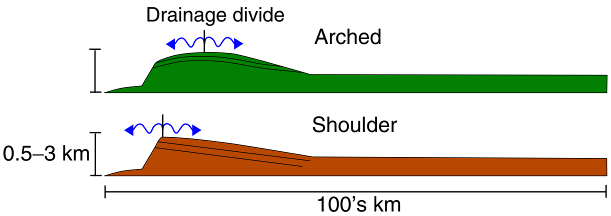

Figure 5 Arch vs. Shoulder Type margins. Vertical line with double arrows shows the drainage divide.

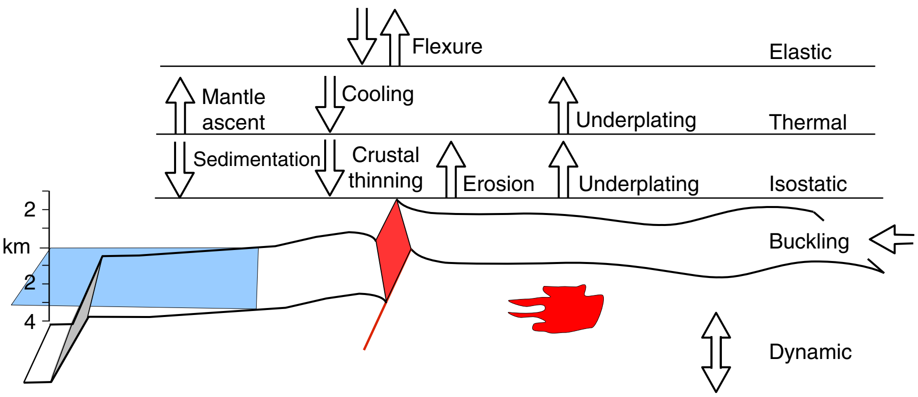

Figure 6 Factors that cause vertical movements at passive margins and continental interiors. The factors can be grouped into elastic, thermal, isostatic, buckling, and dynamic effects. Isostatic responses are obtained from sedimentation, crustal thinning, erosion, and underplating of intrusions. Opposite elastic responses occur on either side of normal faults. Thermal effects may include heating in the early stages of rifting, followed by cooling. Dynamic topography is created by mantle flow. The diagram includes different factors that operate at different times during passive margin evolution, so that it does not represent a single time in passive margin evolution.

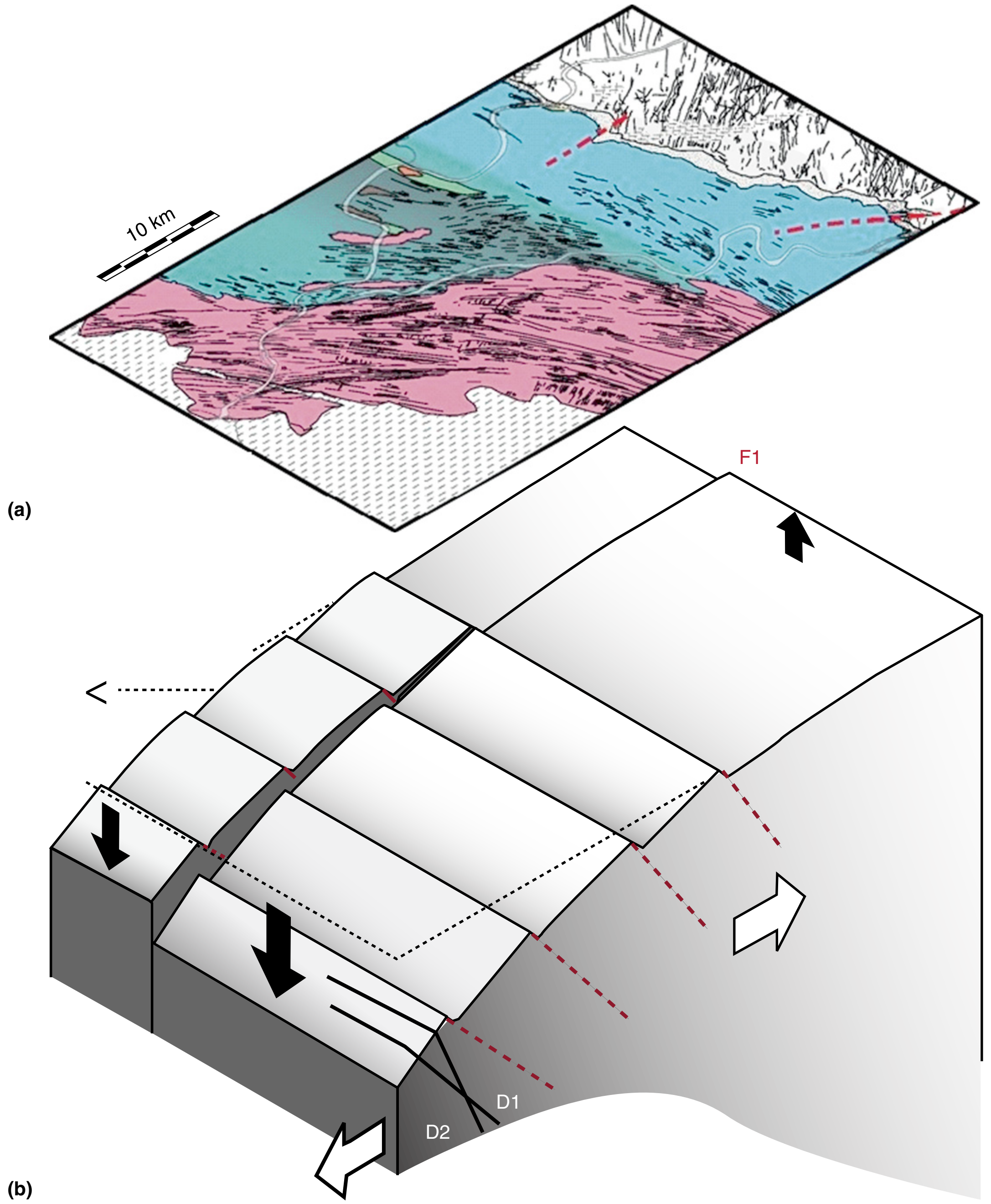

Figure 7 The Geology of the Lebombo monocline. (a) Geology: pink color indicates rhyolites, blue color indicates basalts, and black lines indicate bedding. (b) Model of the crustal structure, showing how extension occurs during intrusion of dykes in phases D1 and D2, and monoclinal downwarping is accommodated by domino block faulting with transverse fault zones (F1). Reproduced with permission from Klausen, M.B., 2009. The Lebombo monoclene and associated feeder dyke swarm: diagnostic of a successfully and highly volcanic margin. Tectonophysics 468, 42–62.

with basalts (Cox, 1993) of the low-lying (450–550 m high) Lebombo mountain range on the southern African east coast (**Figure 7**).

These volcanic rocks were supplied by a dense local network of feeder dikes, which show that the Lebombo was an area of major crustal extension at the time of eruption (du Toit, 1930). The rhyolites have Sr isotopic signatures identical to the associated basalts, suggesting that they were either derived from the latter by fractional crystallization or by partial melting of source rocks similar to the basalts. Either mechanism requires that the rhyolites represent ~11–16% of the original basaltic source rocks, which thus would have to be 45–75 km thick (Cox, 1993). A lower, but still substantial amount of underplating would be implied on the basis of a thinner (3.5 km) sequence of rhyolites inferred by Klausen (2009). Underplating of this magnitude would be expected to produce substantial permanent uplift of the margin, whereas the Lebombo rises only slightly above the 'low veld' of the surrounding low-lying (<300 m) east coast plain of southern Africa.

The Lebombo volcanic belt has further important implications for understanding the processes associated with rifted margins. It has a monoclinal structure, dipping eastwards toward the coast, where it is covered beneath younger Cretaceous sedimentary rocks (Dingle and Scrutton, 1974). The lavas are cut by dikes with compositions that can be matched with those of the volcanic rocks, indicating that they represent the feeder system. The highest dyke density is associated with the monoclinal axis and the displacement resulting from these intrusions indicates a crustal extension of at least 15% (du Toit, 1930). The oldest dikes dip to the west (**Figure 7**), showing that they have been tilted by the monoclinal warping, whereas the youngest intrusive rocks are close to vertical. These direct field observations led du Toit to conclude that the maximum period of flexure must have been during the interval of volcanic eruption of the Lebombo belt – a conclusion recently endorsed by Klausen (2009). Cox (1993) noted that had the monocline been below sea level, it would certainly have been identified as a sequence of seaward-dipping reflectors by marine geophysicists.

The Lebombo belt shows a sharp change in strike at the Limpopo river from almost north-south to trending north northeast. This point of inflection marks the eastern end of a major dyke swarm that trends west northwest through southern Zimbabwe and across Botswana for a distance of approximately 1200 km. Reeves (1978) suggested that the dykes exploited the failed arm of a triple junction, with the trends of the Lebombo volcanic rocks marking the line of failure associated with opening of the Indian Ocean.

The Deccan Traps flood basalts in India are associated with the rifted passive margin that forms the west coast of the country. They occur on the northern end of the major seaward-facing escarpment known as the Western Ghats (**Figure 8**). In the extreme north of the outcrop, the lavas show a seaward-dipping monoclinal flexure, which dies out to the south (Ollier and Powar, 1985). The northern margin of the Traps is associated with a rift, which intersects the west coast at the Gulf of Cambay. The Tapti and Narmada Rivers exploit this rift to flow to the west coast, and thus in the opposite direction to the major drainages further to the south, which rise off the western Ghats and flow eastwards to the Bay of Bengal. Cox (1989) interpreted the rift, which was exploited by the two major west-flowing rivers, to represent the failed arm of a triple junction. This points to a common tectonic setting at a triple junction for the monoclinal flexures of the Lebombo and the northern extremity of the Western Ghats (**Figure 8**). Both are characterized by a low-relief interior that contrasts with the well-defined escarpment elsewhere on the southern African and West Indian passive margins.

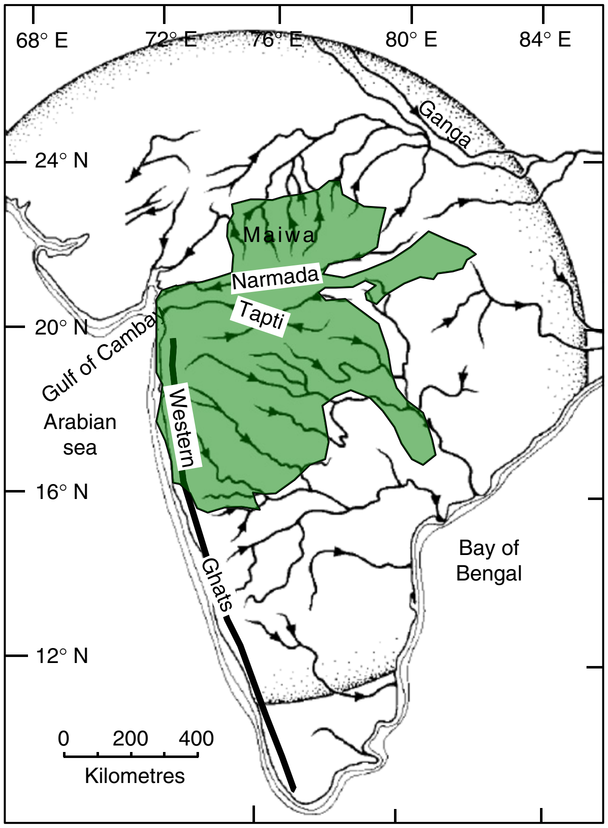

Figure 8 Drainage patterns in India and the Deccan Traps. Most of peninsular India is characterized by dome flank drainage, but the Narmada and Tapti drainages, flowing in the opposite direction, are examples of rift-related drainage. The extent of the inferred plume is shown by the stipple and the extent of the Deccan traps is shown in green. Western Ghats is the line of the escarpment. Diagram partly based on Cox (1989).

5.5.3 Prerifting Continental Topography and Elevation

The interior of southern Africa has an anomalously high elevation (generally >1000 m asl) compared with average elevations of 400–500 m asl that characterize cratonic areas in other continents (Lithgow-Bertelloni and Silver, 1998; Gurnis et al., 2000). A critical issue in accounting for such differences is the elevation and nature of the topography of the prerift land surface. At some rifted margins, there are very tight constraints, as in the opening of the Red Sea–Gulf of Suez, where normal faults related to rifting, dated at 21–19 Ma, cut across a sedimentary sequence capped by shallow water Eocene marine carbonates (Steckler and Omar, 1994). However, by 17 Ma coarse conglomerates deposited into the rifted marine basin provide evidence for the existence of high marginal relief, reflecting rift flank uplift, which at the present time attains altitudes in excess of 2000 m along the Red Sea–Gulf of the Suez margin (Steckler and Omar, 1994). However, Japson et al. (2006, 2009) used landscape analysis and interpretation of apatite fission track (AFT) data to show that following rifting at ~65 Ma, the Greenland west coast was low lying, and experienced Palaeogene and Miocene marine transgressions. The modern relatively high elevation (up to 2000 m asl) and rugged topography of this margin was inferred to have developed relatively recently following two phases of Neogene uplift (11–10 and 7–2 Ma). Japson et al. (2006, 2009) suggested that several other elevated plateaux in continental hinterlands may also postdate breakup, and may be relatively recent.

In southern Africa, the Lower Triassic Beaufort sediments are interpreted to have been deposited in a terrestrial low-relief environment of marshes crossed by intermittent streams (Truswell, 1970). Triassic uplift of the Cape Fold Belt resulted in highlands in the south of the region, but absolute elevations are not at all constrained. Little geological evidence exists for the configuration of the drainage system in the Jurassic, which was marked by the eruption of the Karoo flood basalts in the early part of this period (~182 Ma; Marsh et al., 1997). However, by the early Cretaceous, following opening of both the Indian and the Atlantic Oceans, the configuration of continental drainage of south-central Africa changed drastically from the Triassic river system that preceded the Karoo volcanism (Moore and Blenkinsop, 2002).

Cox (1989) ascribed the development of the post-Gondwana drainage network in southern Africa to updoming over the Karoo and Paraná plumes, centered on the east and west coasts, respectively. This concept was extended by Moore and Blenkinsop (2002), who noted that initial uplift, related to the Karoo volcanic event in the east, would have produced a westward-draining network, but that the subsequent Paraná-Etendeka volcanism, centered on the Atlantic margin of southern Africa, would have reversed the postrift paleo-slopes, initiating the major eastward-flowing rivers that dominate the modern drainage system. It should be stressed, however, that these drainage reconstructions are independent of whether Gondwana disruption was initiated by the putative Karoo and Paraná mantle plumes. Advective heating by penetrative volcanism related to the extensive dike swarms associated with the Karoo and Paraná-Etendeka volcanism, possibly coupled with rift-flank uplift, could also explain the doming invoked by Cox (1989).

The magnitude of the prerift topography in southern Africa remains a highly contentious issue. Partridge and Maud (1987) envisaged a relatively high (though not specified) altitude for southern Africa before rifting. This was supported by Gilchrist and Summerfield (1991), who calculated the prerifting continental elevation to be ~1200 m asl by reloading the mean continental denudation of ~1800 m estimated by Rust and Summerfield (1990) for the Atlantic-draining catchment and making an adjustment for Airy isostasy. In contrast, Burke and Gunnell (2008) argued that the continent was low lying, and characterized by low relief immediately before continental breakup. On the basis of topographic modeling, Doucouré and de Wit (2003) concluded that the modern bimodal topography of Africa had been established by the Cretaceous, with elevated areas centered on east and southern Africa, a conclusion that was also reached by interpretation of AFT data (Kounov et al., 2008, 2009; Tinker et al., 2008). However, Doucouré and de Wit (2003) envisaged that both areas were at considerably lower average elevations (200–400 m) than the modern topography (>1000 m average elevations asl). The difference between their estimate and that of Gilchrist and Summerfield (1991) serves to underline the uncertainties inherent in the assumptions underlying geophysical modeling.

The arguments for a low elevation of Africa immediately before, and for an extensive period following, continental breakup must also be seen against the firm geological evidence that in the Red Sea–Gulf of the Suez passive margin, rift flank uplift, erosion, and consequent isostatic rebound had produced a strong inland relief in less than 3 Ma following rifting, and elevations in excess of 2000 m asl at the present time, some 21 Ma after rifting (Steckler and Omar, 1994). The topography around the time of rifting of regions with anomalously high present-day elevations, as in the case of east and southern Africa, remains strongly debated, and may vary considerably in different geological/tectonic settings (e.g., Greenland vs. Red Sea–Gulf of Suez).

5.5.4 Postrifting Evolution of Marginal Escarpments

One of the most controversial issues in postrifting landscape evolution has been the evolution of the dramatic seaward-facing escarpments that occur on some passive margins. This debate must naturally be extended to explain why they are absent on other margins and why some pairs of margins have apparently complementary geometries, such as the well-developed elevated escarpment of Namibia in southern Africa and the extensive low-relief plain on the margin of the Argentinean seaboard.

5.5.4.1 King's Scarp Retreat Model – Backwearing

The elevated mountainland formed by the Drakensberg Karoo basalts (**Figure 4**), which represents the highest point of the almost continuous coast-parallel escarpment that bounds southern Africa, has long held central stage in the debate regarding the evolution of passive margins. King (1955, 1963) envisaged that the coastal plain seaward of the Drakensberg mountainland had been planed by a process of parallel scarp retreat (backwearing), analogous to that originally proposed by Penck (1924) (**Figures 9(b)** and **9(c)**). King considered that the original scarp had formed at the coast at the time of disruption of Gondwana and had advanced inland over a period of 120–140 Ma. The coastal plain would accordingly be regarded as a pediplain.

5.5.4.2 Pinned Drainage Divide – Downwearing

Kooi and Beaumont (1994, 1996) used numerical landscape modeling to show that if an escarpment formed at the coast at the time of continental breakup, it would very rapidly degrade if there was an antecedent drainage system with an inland drainage divide (**Figure 9(d)**). They demonstrated that a new escarpment could then rapidly be initiated at the location of the original drainage divide, followed by very slow rates of inland retreat. This model for location of the escarpment at a pinned drainage divide is strongly dependent on a number of poorly known variables, and is inapplicable to the Eastern Australian coastal escarpment and other arched-type escarpments, where the modern divide is located well inland of the scarp. Nevertheless, it provides a potential explanation for the broad coincidence of the escarpment and modern coastal drainage divide in southern Africa and the Western Ghats of India.

Fleming et al. (1999) first proposed that the pinned drainage divide concept was more appropriate to the evolution of

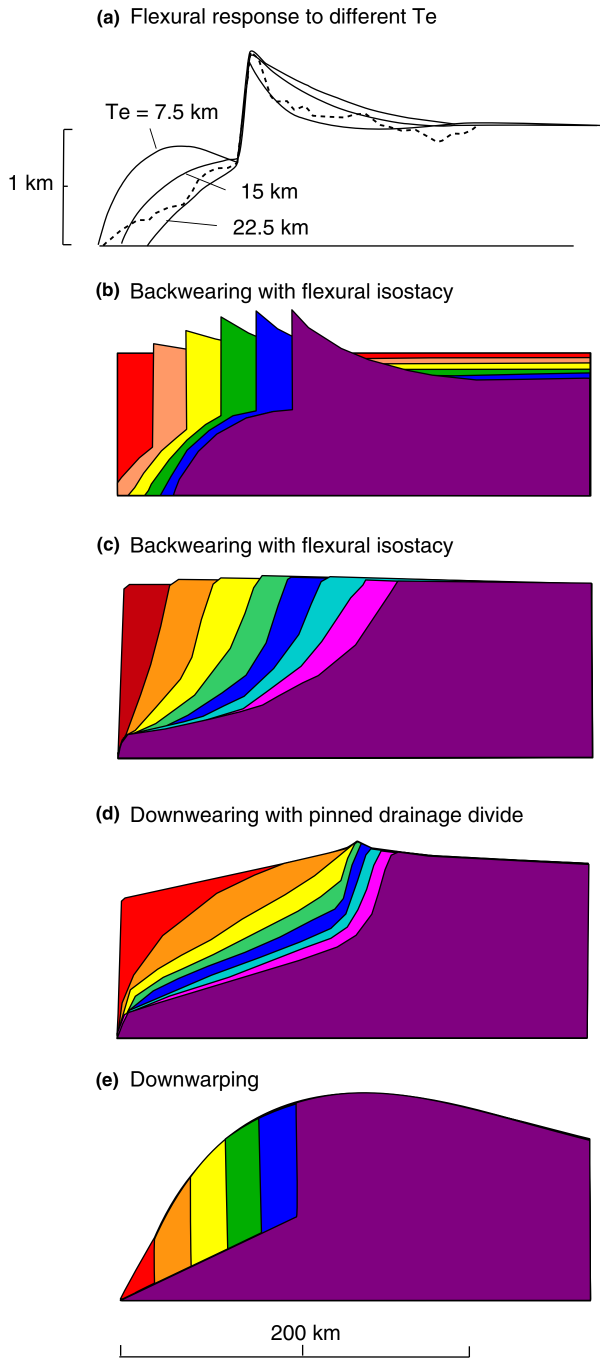

Figure 9 Escarpment evolution at passive margins. (a) Modeled rift-flank uplift associated with flexural isostasy for various values of elastic thickness T_e. The typical profile across the western margin of southern Africa is shown as a dashed line (Gilchrist and Summerfield, 1990). (b) Models of scarp retreat and rift-flank uplift associated with flexural isostasy (Gilchrist and Summerfield, 1990). Colors in all diagrams represent successive stages. (c) Models of scarp retreat with flexural isostasy (Van der Beek et al., 2002). (d) Plateau degradation model, pinned drainage divide (Van der Beek et al., 2002). (e) Model of downwarping of the Continental margin (Ollier and Pain, 1997). All diagrams redrawn from original sources and rescaled to approximately the same scale.

the Drakensberg escarpment than King's scarp retreat model on the basis of cosmogenic isotope studies, also carried out by Kounov et al. (2007) with similar implications. These indicated very slow modern retreat rates of this major scarp, incompatible with inland migration from an original coastal position at a constant rate since continental breakup. This view was subsequently supported by Brown et al. (2002) and Van der Beek et al. (2002) on the basis of AFT studies and numerical landscape modeling, which also assumed a constant rate of scarp retreat subsequent to breakup.

The models for downwearing of the Drakensberg escarpment to a pinned drainage divide were criticized by Moore and Blenkinsop (2006). They noted that a major shortcoming was the dependence of the models on variables that were difficult to constrain and the fact that the models failed to account for the observation that the Drakensberg is bounded by major scarps on three sides. Further, the modern scarp is capped by very thick, resistant basalt flows, and the present-day basalt outcrop is surrounded by a dense boxwork of dykes, 40–60 km wide, that was the source of the lavas, suggesting a major local focus of volcanic activity. Moore and Blenkinsop (2006) argued that massive lava flows would be more likely in close proximity to the boxwork of feeder dykes and that erosion rates would be dramatically slowed once the lavas were exposed. Erosion rates would further be influenced by climatic factors and the late Neogene evolution of the C4 grasses. Modern erosion rates, determined on the basis of cosmogenic nuclides, may have no bearing on historic rates, and cannot therefore be used as a basis to reject King's scarp retreat model.

The importance of lithology in controlling the inland migration of an initial escarpment formed in response to rifting is further illustrated in the classic investigation of the Red Sea–Gulf of Suez passive margin of Steckler and Omar (1994). In this area, rifting has incised a sequence of massive, resistant Eocene limestones that cap the friable and readily eroded 'Nubian sandstone.' The entire sequence thins from the north to the south, and as a result, erosion has resulted in the initial removal of the resistant Eocene limestone capping to expose the 'Nubian Sandstone' in the south. Erosion of the latter formation has caused rapid undercutting of the overlying limestone cap and the greatest inland advance of the scarp in the south.

5.5.4.3 Downwarping of the Continental Margin

Ollier drew attention to the presence of a major continental divide inland of the escarpment that overlooks the coastal plain of much of eastern Australia (Ollier, 1982; Ollier and Pain, 1997; Ollier and Stevens, 1989). These authors suggested that marginal downwarping of eastern Australia at the time of continental breakup had occurred along an axis corresponding to the inland divide. Thereafter, headward river erosion into the downwarped shoulder of the continent resulted in the development of the coastal escarpment (**Figure 9(e)**). Although Ollier (1982) envisaged that scarp retreat continues at the present time, a Pliocene basalt that has flowed down one of the escarpment valleys shows relatively little evidence of erosion. The implied slow recent rates of erosion of the eastern Australian escarpment might reflect the drastic vegetation change linked to the late Neogene evolution of widespread C4 grasses (Moore and Blenkinsop, 2006). This is broadly consistent with AFT data that suggest rapid erosion linked to Cretaceous breakup (e.g., O'Sullivan et al., 1995), followed by slow erosion through the Cenozoic.

Ollier's model has been criticized by Nott and Horton (2000), who presented field evidence that along the northern sector of the eastern Australian escarpment, the inland drainage divide was already extant in the early Jurassic and thus predates the Late-Jurassic to early Cretaceous formation of the Queensland and Townsville Basins near the modern coastline (Struckmeyer and Symonds, 1997). Nevertheless, the Lebombo monocline in the east of southern Africa formed synchronously with the associated lavas, erupted at ~182 Ma (Du Toit, 1930; Klausen, 2009), and thus significantly before 155 Ma, when opening of the Indian Ocean is generally inferred to have commenced (Reeves and de Wit, 2000; McMillan, 2003). If this sequence of events is correct, the existence of a drainage divide in eastern Australia before continental breakup in this area would not be incompatible with a marginal downwarp that has remained stationary and existed for longer than generally assumed (cf. Young, 1989). It should be noted, however, that Klausen (2009) has proposed that continental rifting in the east of southern Africa to initiate the Indian Ocean commenced immediately after eruption of the Lebombo volcanics. A monoclinal geometry is also suggested by Ghebreab and Talbot (2000) for the Red Sea margin in Eritrea. Here, extension began at 40 Ma, followed by a second event at 23 Ma, which formed the monoclinal flexure, uplifting 30 Ma flood basalts in its hanging wall, at about the time that the Arabian and Nubian shields separated. Flood basalts, monoclinal flexure, and ocean opening therefore appear to have occurred simultaneously here, similar to the scenario proposed for southern Africa by Klausen (2009).

5.5.4.4 Isostasy and Flexure

King (1955) proposed that erosional isostatic rebound of the coastal plain would occur once a critical threshold of ~480 km of inland erosion had occurred. Thereafter, uplift would take place, initiating a new cycle of scarp retreat and pediplanation, giving rise to the development of cyclic land surfaces. King's (1955) envisaged multicycle erosion surfaces have been strongly criticized (e.g., Summerfield, 1988) because geophysical models show that the isostatic response to erosional unloading would be immediate and continuous, rather than delayed and episodic. Furthermore, King (1955) did not account for the flexural response of the lithosphere. Nevertheless, as elaborated in a later section, objections to the mechanism proposed by King (1955) cannot be used to counter the field evidence for polycyclic erosion in southern Africa. The concept of flexural isostacy was applied to escarpments by Gilchrist and Summerfield (1990, 1991, 1994), who pointed out that one of the consequences of rifting would be higher local relief and a marked lowering of the erosional base level on the coastal margin of a rift flank, leading to higher denudation rates than in the relatively low-relief interior. Numerical modeling showed that if the lithosphere was treated as an elastic plate, this differential denudation would result in landward erosion, coupled with progressive elevation of the escarpment separating the coastal plain from the interior plateau (**Figures 9(a)** and **(b)**).

One of the main uncertainties in applying the flexural isostacy model to passive margins is the elasticity of the lithosphere, encapsulated by the flexural rigidity *Te*. Estimates of *Te* for continental margins are quite variable, even along the same margin. For example, Tassara et al. (2007) showed estimates of <10 to 30 km for the passive margins of South America. Pazzaglia and Gardner (1994) suggested an average value of 40 km for the U.S. east coast, within the range of 20–60 km for previous estimates, and much larger than the 15 km value for the SW African margin (Gilchrist and Summerfield, 1991). Although *Te* estimates are subject to some uncertainty, there appears to be a real variation between different passive margins and variation along the length of the same margin.

The flexural isostacy model was applied to the escarpment along the west coast of South Africa (Gilchrist and Summerfield, 1991), where up to 5 km of postrift denudation has occurred on the coastal plain (Brown et al., 2002; Kounov et al., 2009). Gilchrist and Summerfield (1991) concluded that the best correspondence between modeled topography and that observed for the coastal plain, marginal escarpment, and inland plateau was obtained when *Te* = 15.0 km (**Figure 9(a)**). A better fit was obtained for the escarpment and interior plateau for a lower elastic thickness (7.5 km), but this also predicted a major crustal warp on the coastal plain, which contrasts with the very subdued upwards convex profile of the west coast of South Africa. In contrast, it is noted that scarp retreat initiated by the Red Sea–Gulf of Suez rift has been associated with deep erosion (2–5 km) and major uplift on the coastal plain (Steckler and Omar, 1994), similar to that predicted by the Gilchrist and Summerfield (1991) for their model assuming a low elastic thickness for the continental margin. The implication is that the observed differences in *Te* for different continental margins could to lead to differences in the isostatic flexural response to differential denudation associated with the development of a marginal escarpment.

An important prediction of the model put forward by Gilchrist and Summerfield (1991) is that inland sediment dips would be expected in the vicinity of the escarpment. This is indeed the case for the Karoo sequence on the coastal plain below the Drakensberg escarpment in the east of South Africa, where a minimum of 4.5 km of denudation (Brown et al., 2002) is estimated to have occurred in the ~150 Ma since initiating of rifting (**Figure 10**; de Wit, 2003). Differential denudation and flexural isostasy thus appear to have played a major role in the evolution of this portion of the southern African horseshoe escarpment.

Gunnell and Fleitout (2000) demonstrated that a lithospheric isostatic flexural response to the differential rates of erosion of the inland plateau and coastal plain could in principle account for the major escarpment formed by the Western Ghats (**Figure 8**). The best fit between the actual and the modeled topography was obtained when the lithosphere was treated as a broken plate, with a high flexural rigidity (*Te* = 70 km) inland of the rifted margin and a low rigidity (*Te* = 1.3 km) on the seaward side of the break. However, Gunnell and Fleitout (2000) stressed that this was a

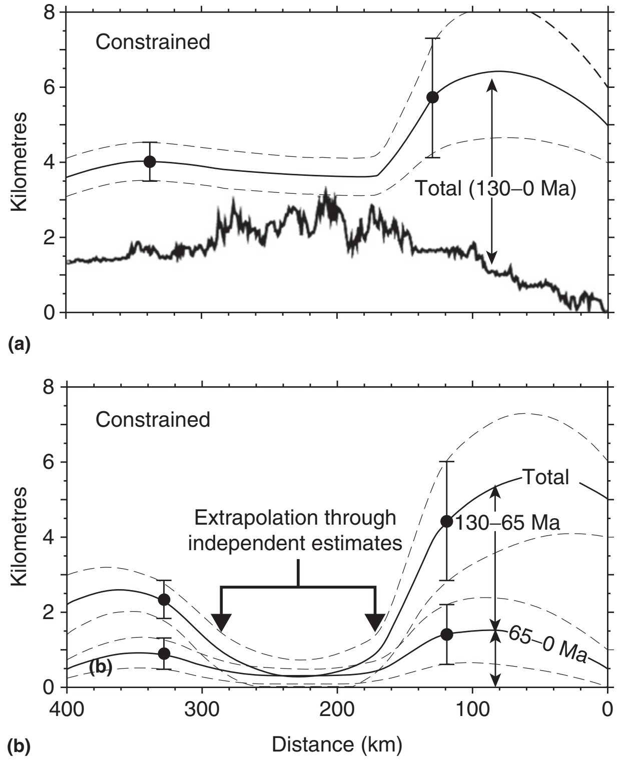

Figure 10 Interpolation between the denudation estimates at the two borehole sites constrained across the Lesotho Highlands by independent estimates derived from zeolite distribution and *in situ*-produced cosmogenic 36Cl concentrations. (a) Total denudation, shown relative to present-day surface. (b) Denudation estimates for Late Cretaceous and Tertiary periods. Reproduced with permission from Brown, R.W., Summerfield, M.A., Gleadow, A.J.W., 2002. Denudational history along a transect across the Drakensberg Escarpment of southern Africa derived from apatite fission track thermochronology. Journal of Geophysical Research 107, 2350, doi:10.1029/2001JB000745.

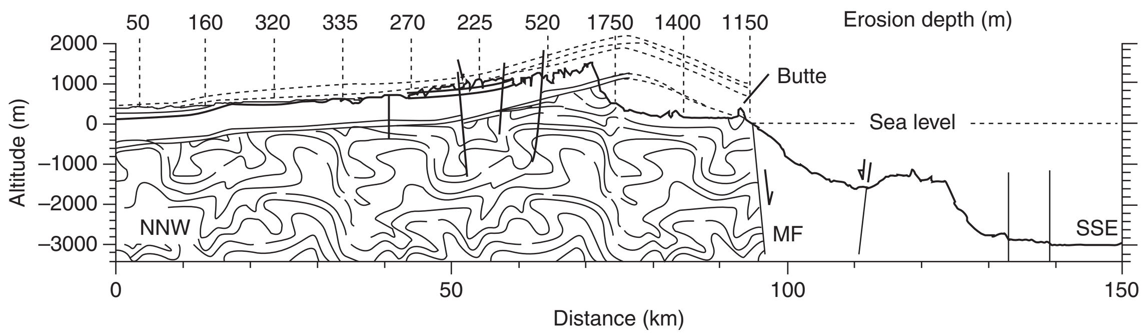

Figure 11 Strike-perpendicular cross section of Jabal Samhan, Arabian peninsula. MF – master fault. Marginal downwarping is attributed to sediment loading. Reproduced with permission from Gunnell, Y., Carter, A. Petit, C., Fournier, M., 2007. Post-rift seaward downwarping at passive margins: new insights from southern Oman using stratigraphy to constrain apatite fission-track and (U-Th)/He dating. Geology 35, 647–650.

nonunique solution in view of the poorly known variables involved in their models.

A study by Gunnell et al. (2007) has potentially important implications for the roles of both the flexural isostasy and the marginal downwarping of the continent in controlling escarpment evolution. These authors noted that the sediments forming the escarpment on the Oman coast (Arabian Peninsula) dip inland, as predicted by the modeled flexural response to differential erosion of the interior plateau and coastal plain. However, the unconformity at the base of the escarpment sedimentary sequence is preserved at a lower level closer to the coast, and the elevation differential cannot be explained by faulting. It was therefore inferred that marginal downwarping of the continent had occurred along an axis of flexure on the coastal side of the escarpment (Figure 11) and that 1.75 km of denudation had occurred on the coastal plain. On the basis of numerical modeling and AFT constraints, the seaward flexure was inferred to be the result of sediment loading on the continental shelf. This contrasts with the model for early continental flexure linked directly to continental rifting proposed by Ollier (2004), which is supported by the synchronicity of the Lebombo Karoo age volcanics and coastwards monoclinal flexure of the eastern southern African coastal plain. An implication of the modeling carried out by Gunnell et al. (2007) is that coastal flexures may have more than one origin, and by extension, different temporal links to continental rifting.

The passive margins discussed above have all experienced deep (several km) erosion extending well inland of the present escarpment (Brown et al., 1999). A common feature of these margins and their hinterland was an originally thick cover of readily eroded sediments. In contrast, cover sediments were either thin or absent over the eastern third of Madagascar (Emmel et al., 2004) before rifting in the late Cretaceous (De Wit, 2003). AFT studies across the eastern Madagascar (Emmel et al., 2004; Seaward et al., 2004) indicate that there has been minimal erosion inland of the escarpment since the time of rifting. This suggests that the limited depth of post-rifting erosion along this margin reflects the absence of a thick readily eroded sedimentary cover overlying the more resistant granitic basement.

The studies discussed above indicate that flexural isostasy has played a major role in the formation and growth of escarpments on passive margins. The possibility of variable flexural properties of passive margins along strike suggests the potential for along-strike variations in modes of landscape development and rates of erosion. Incision rates may vary by an order of magnitude, and early landscapes may be very different from one point to another, as documented for the southeastern Australian margin by Van der Beek et al. (2001).

5.5.4.5 Perspectives from Integrating Low-Temperature Geochronology and Numerical Modeling

Ollier and Pain (1997) and Gallagher et al. (1998) suggested that AFT analyses offer a useful way to distinguish downwarping from flexural isostasy models for passive margin evolution. Because the base of escarpments in a downwarped margin will be exhuming deeper stratigraphic levels, AFT ages should decrease from the coast toward the escarpment base. However, the rapid and deep exhumation of scarps near the coast in the flexural models should lead to AFT ages that increase inland toward the escarpment (Gallagher et al., 1998). A compilation of AFT data from Southeast Australia, Southeast Brazil, Southwest Africa, Yemen, and West India shows a general increase in age inland, and none of the data sets show the predicted pattern for downwarping. However, this is not the pattern observed in Scandinavia, where the youngest AFT ages occur at the heads of the fiords (Rohrman et al., 1995). Ollier and Pain (1997) interpret these data to confirm the downwarping model.

These conclusions need to be reevaluated in the light of recent numerical modeling. Braun and Van der Beek (2004) stated that "A retreating escarpment would progressively reset the thermochronological ages as it migrates from the coastline to its present-day position and therefore lead to a clear decrease in thermochronological ages from the coastline toward the escarpment." The modeling of Braun and van der Beek (2004) revealed the reason for this contradiction with previous concepts. The spatial pattern of low-temperature geochronological results is sensitive to a range of variables, including the geotherm and erosion rates, in addition to the mode of passive margin evolution, which requires evaluation by numerical modeling. The modeling suggests that the low-temperature geochronology of transects from the coast to the escarpment cannot generally discriminate between backwearing and downwearing (**Figure 12**): this has been confirmed for (U-Th)/He data from the Southeast Australian escarpment (Persano et al., 2002). To constrain the mode of escarpment development, transects are needed parallel as well as perpendicular to the escarpment (Braun and Van der Beek, 2004).

5.5.4.6 Sinuosity of Escarpments

The sinuosity of escarpments appears to evolve with age and to depend on the type of margin. Matmon et al. (2002) showed that escarpments at passive margins generally had higher sinuosity than those of rifts, implying an evolution toward more complex margins through time. Arch-Type margins have higher sinuosity than Shoulder Type. Both these aspects of escarpment geomorphology are consistent with higher measured rates of retreat at knick points compared with interfluves, which would lead to greater sinuosity over time. The greater frequency and size of drainages that cut Arch Type escarpments are also compatible with this model for escarpment evolution.

5.5.4.7 Low-Relief Passive Margins without a Marginal Escarpment

The controversy surrounding the origin of escarpments inland of many passive margins is mirrored by the question of why they are absent along others. This issue is well illustrated by the margin of Namibia, with its well-developed escarpment and high interior, and the conjugate low-lying Argentinean coast line, where the Saldo rift basin intersects the coast at right angles along the low-relief Argentinean portion of the South American margin. The Saldo basin was associated with strike–slip faulting that accommodated crustal reorganization during the early phase of separation of Africa and South America (Nürnberg and Müller, 1991) and represents a failed rift system. The major Paraná-Plate river, with headwaters inland of the escarpment located to the north, enters the Atlantic via the Saldo Rift.

In southern Africa, there is a prominent break in the escarpment associated with the Lebombo triple junction (**Figure 4(a)**), which is exploited by the Limpopo river, with headwaters well inland of the coastal escarpment. In India, the major Tapti and Narmada Rivers empty into the Arabian Sea at the major inflection in the coastline forming the Gulf of Cambay, at the northern end of the escarpment formed by the western Ghats (**Figure 8**). Cox (1989) inferred that this was a triple junction and that these two rivers have exploited a failed rift arm.

There are thus several examples of low-relief passive margins lacking an inland escarpment that are associated with failed rift arms linked to the disruption of Gondwana. These are invariably exploited by major rivers, with headwaters well inland. The geomorphic expression of some passive margins is

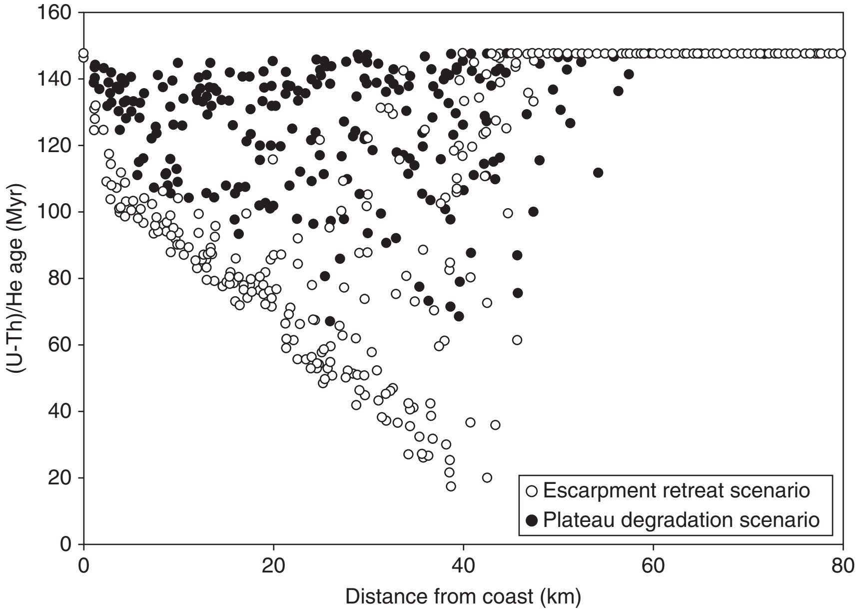

Figure 12 Distribution of apatite (U–Th)/He ages as a function of distance from the coastline as computed from Pecube temperature predictions under the two end-member scenarios. Reproduced with permission from Braun J, van der Beek P., 2004. Evolution of passive margin escarpments: what can we learn from low-temperature thermochronology? Journal of Geophysical Research 109, F04009. doi: 10.1029/2004JF000147

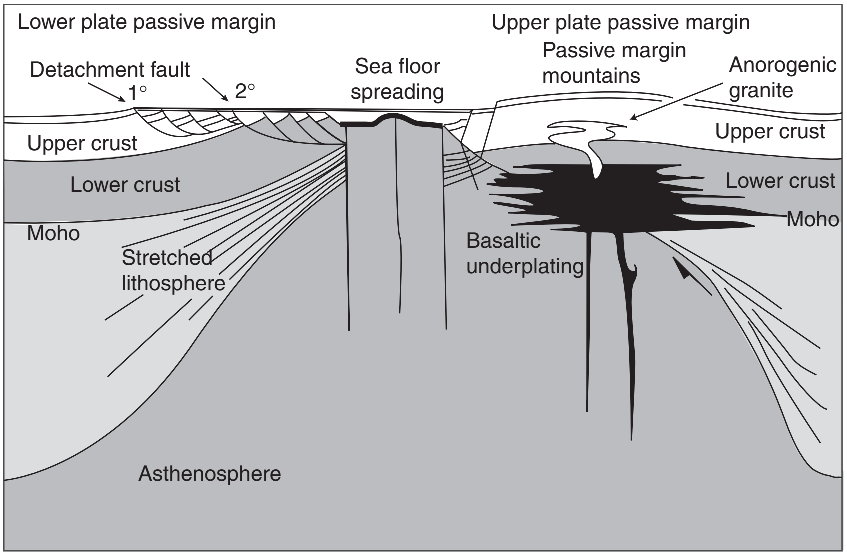

Figure 13 Conjugate margins differentiated as the Upper plate (hanging wall of detachment) with high elevations and Lower plate (footwall) with low elevations. Redrawn from Lister, G.S., Etheridge M.A,. Symonds P.A., 1991. Detachment models for the formation of passive continental margins. Tectonics 10, 1038–1064.

closely linked to structural controls, in particular the location of rifts associated with triple junctions.

An entirely different perspective on low-relief passive margins is offered by the asymmetric rifting model of Lister et al. (1986, 1991; **Figure 13**). Based on the premise that asymmetry created by detachment faults is an inherent feature of crustal extension, Lister et al. (1986) proposed that passive margins can be considered in terms of an upper plate (the hanging wall to the detachment) and a lower plate (the footwall). The lower plate margin is envisaged to consist of rotational normal faults and tilt blocks, compared with the upper plate with higher angle normal faults. The upper plate is subject to uplift due to exposure to rising asthenosphere in the footwall of the detachment fault and thus becomes a High-Elevation passive margin, whereas in the lower plate, substitution of mantle for lower crustal material results in subsidence to form a Low-Elevation passive margin. An interesting result from numerical modeling of uplift in these models is that the only mechanism to ensure long-lived elevation of passive margins is underplating (Lister et al., 1991).

Typically, then, passive margins should consist of opposed, conjugate High- and Low-Elevation margins. Examples cited by Lister et al. (1991) include the Antarctic and South Australia, the Blake Plateau and the Guinea basin on the North Atlantic, and Southwest Africa and the Argentinean margin in the South Atlantic. Along-strike changes in polarity of the asymmetric extension may occur at transform faults, in a manner that is very similar to the alternating half-graben geometry of continental rifts (e.g., Rosendahl, 1987). Although the asymmetric rifting model is capable of explaining the diversity of passive margin expressions, one problem is the difficulty of finding strong observational evidence for its diagnostic feature: large displacement detachment zones.

5.5.5 Evolution of Continental Hinterlands

Southern Africa will be used as the backdrop to explore the nature of the controversies about landscape evolution inland of the escarpment following continental breakup. These center around two broad, but interrelated issues. The first is the concept of cyclical erosion episodes, which was championed for southern Africa by King (1949, 1955, 1963), and the second is the anomalous average elevation (>1000 m asl) of the interior plateau of the region. The latter problem raises the issue of dynamic topography, which is currently an area of great interest in tectonic geomorphology (e.g., Moucha et al., 2008).

5.5.5.1 Cyclic Erosion

King (1955, 1963) inferred after careful observation that landscape evolution involved successive phases of planation by scarp retreat, to produce a low-relief pediplain. He postulated that continental uplift was responsible for initiating a new cycle of erosion, which would incise the surface formed in the preceding cycle, and that six surfaces of differing ages could be identified in southern Africa. The oldest of these, termed the Gondwana Surface, was considered to have been planed before the disruption of Gondwana. Remnants of this surface were recognized in the highlands of the Drakensberg, and subsequently also the eastern highlands of Zimbabwe (Lister, 1987). The most widespread surface, termed the African Surface, formed following continental breakup, and was located over wide areas of the continent. Much of the early criticism of King's (1955) model centered on the problem of identifying planation surfaces and particularly their correlation over extensive distances. As discussed earlier, a more recent criticism has been leveled at the mechanism of punctuated isostasy that King (1955) proposed to account for episodic continental uplift.

In a major review, and arguably one of the most extensively cited references to landscape evolution in Africa, Partridge and Maud (1987) expressed strong support for cyclical erosion in southern Africa, but proposed a simplified model, involving three major cycles. They argued that Gondwana age surfaces are nowhere preserved in African, except where they were exhumed from beneath younger cover rocks. This view was subsequently endorsed by Fleming et al. (1999) on the basis

Figure 14 The African erosion surface (arrow) at Grahamstown, Eastern Cape, South Africa, with silcrete capping kaolinite.

Figure 15 Granite inselbergs characteristic of the Post-African I planation surface, Matopos Hills from Rhodes' Grave, Zimbabwe.

of erosion rates in the Drakensberg highlands, determined using cosmogenic nuclides, which implied substantial surface lowering since continental breakup.

Partridge and Maud (1987) proposed that there was a lengthy period of erosion following the disruption of Gondwana, with the early stages during the Cretaceous involving deep weathering under tropical conditions. This resulted in a senile surface with a deeply kaolinized mantle, capped by silcrete and ferricrete duricrusts, which was equated to King's African Surface (**Figure 14**). It was regarded as multicyclic in origin, but Partridge and Maud (1987) considered that there was insufficient evidence to confidently identify subcycles. The characteristic African Surface weathering profile was taken as a reference point for identifying younger surfaces. Partridge and Maud (1987) inferred that the African cycle of erosion was terminated by Miocene uplift, initiating a relatively short-lived, and therefore less extensive, Post-African I planation surface. Development of this surface primarily involved stripping of the weathered African Surface carapace, and could therefore be considered an etch surface (Moore et al., 2009). Granite inselbergs (**Figure 15**), which are commonly characteristic of this surface, have been interpreted to represent more resistant areas associated with lower joint densities, resulting in a shallower base to the African weathering profile (Twidale, 2002). The Post-African I planation surface was in turn envisaged to have been interrupted by major Plio-Pleistocene uplift of up to 900 m, focused on a line termed the Ciskei-Swaziland Axis, located on the eastern seaboard of southern Africa.

The broad synthesis proposed by Partridge and Maud (1987) has resulted in an interesting dichotomy of geological viewpoints. It is widely accepted in southern Africa that the diagnostic African Surface weathering mantle provides a reliable datum for long-range correlations and thus for recognizing younger cycles of erosion. However, as with criticisms leveled at King's (1955) original proposals for cyclic erosion surfaces, a major objection continues to be the absence of a mechanism to account for episodic uplift of the continent. Thus, Gilchrist and Summerfield (1991) suggested that uplift of a marginal scarp should be continuous rather than episodic. However, it should be emphasized once again that the criticism of the mechanism proposed by King (1955) does not negate the clear field evidence for multiple erosion cycles. A potential tectonic trigger to account for their initiation is presented in a later section.

King's (1955) model has been widely applied. For example in the Northern Territory, Australia, four distinct erosion surfaces have been identified and associated with cyclic erosion (e.g., Hays, 1968). However, this work has been criticized on the basis that the surfaces are structurally controlled, cannot necessarily be extrapolated over large distances, and have inappropriate ages, and therefore, the surfaces do not constitute evidence for cyclic erosion (Nott, 1994, 1995).

5.5.5.2 Dynamic Topography

Dynamic topography is caused by flow in the mantle that generates vertical stresses to support deformation of the Earth's surface (Hager, 1985; Lithgow-Bertelloni and Silver, 1998). There is currently debate about the possible magnitude of dynamic topographic effects, which have been predicted to provide over a kilometer of elevation (e.g., Lithgow-Bertelloni and Gurnis, 1997; Daradich et al., 2003; Conrad et al., 2004) or to have effects on the order of hundreds of meters (e.g., Wheeler and White, 2000). In either case, dynamic topography could therefore play an important role in passive margin and continental hinterland evolution.

Evidence for the role of relatively subdued dynamic topography has been presented for the continental interior of Australia by Sandiford (2007) and Sandiford and Quigley (2009). These studies point out that Australia is a particularly favorable location to examine dynamic topographic effects because of the exceptionally fast movement of the Australian plate over mantle anomalies, and because the aridity and low relief of the continent preserve some epeirogenic features very well.

The Australian continent is moving north toward an area of low dynamic topography at the convergent boundary between the Australian–Indian plate and the Eurasian plate. An area of low dynamic topography also occurs to the south of the continent (the Australia–Antarctic Discordance). The combination of motion toward low dynamic topography to the north and away from low dynamic topography to the south may have tilted the continent at a rate of 20 m Ma^{-1} over the last 15 Ma, and is responsible for the low elevation of the northern edge of the Australian plate, reflected in its broad continental shelf and the dearth of Cenozoic marine sediments in the northern part of the continent (**Figure 16**; Sandiford and Quigley, 2009). By contrast, Cenozoic sediments are found at elevations of 300 m up to 400 km inland from the southern coast.

The tilting of the Australian continent appears to be due mostly to dynamic topographic effects, but some 10% may also be attributed to movement of the continent across gradients in the geoid (Sandiford, 2007). There is a major geoid high over the convergent plate boundaries of the Indonesia–New Guinea region, which causes geoid height variations of 90 m over the continent. The anomaly is attributed to the deep mantle, caused by the subduction history of the Australian–Indian plate (Sandiford, 2007).

A variety of proposals to explain interior elevation of southern Africa by more dramatic dynamic topography related to plumes have been put forward. Burke (1996) and Burke and Gunnell (2008) argued that lithospheric heating and thermal uplift will be at a maximum when plate movement is slow relative to an underlying mantle plume. They suggested that the northeast extreme of the African continent came to rest over the Afar Plume at approximately 30 Ma. It was envisaged that the resultant lithospheric heating triggered the Afar volcanism, coupled with uplift that was responsible for the elevated topography of south and eastern Africa.

This model poses more problems than those it attempts to address. Firstly, plume skeptics would question whether the uplift associated with the Afar volcanism is a reflection of a deep mantle plume or rather the result of advective heating associated with penetrative magmatic processes, including the intrusion of dykes and sills linked to the Afar volcanism. It is further difficult to envisage the putative Afar plume being responsible for a broad region of high ground in southern Africa, at the southern extreme of the continent, when it is surrounded by much more proximal lower elevations immediately to the east and west in north Africa, although the 'channeled plume' explanation was offered for this phenomenon by Ebinger and Sleep (1998). Moreover, there is no evidence for a major volcanic episode in southern Africa at 30 Ma, and in this region, post-Gondwana alkaline volcanic pipe clusters show a systematic age pattern, younging from the hinterland toward the coastal margins (Moore et al., 2008). The youngest volcanic events (Early Palaeocene and Eocene–Oligocene) would not therefore relate to the high interior plateau.

Lithgow-Bertelloni and Silver (1998) and Gurnis et al. (2000) have suggested that the anomalous elevation of southern Africa reflects dynamic uplift related to a putative superplume. However, seismic tomography, utilizing an array of receiver stations, extending from the southern tip of the subcontinent ~2000 km to the northwest, failed to indicate a major thermal anomaly that could be ascribed to such a plume (Fouch et al., 2004). Further, Moore et al. (2009) point out that the geophysical plume models proposed imply domal uplift of southern Africa, which would be anticipated to result in a radial drainage pattern. In contrast, the most elevated ground is associated with the marginal escarpment and the continental interior is the relatively lower-lying site of the Kalahari basin (**Figure 4a**). Further, in contrast to the radial

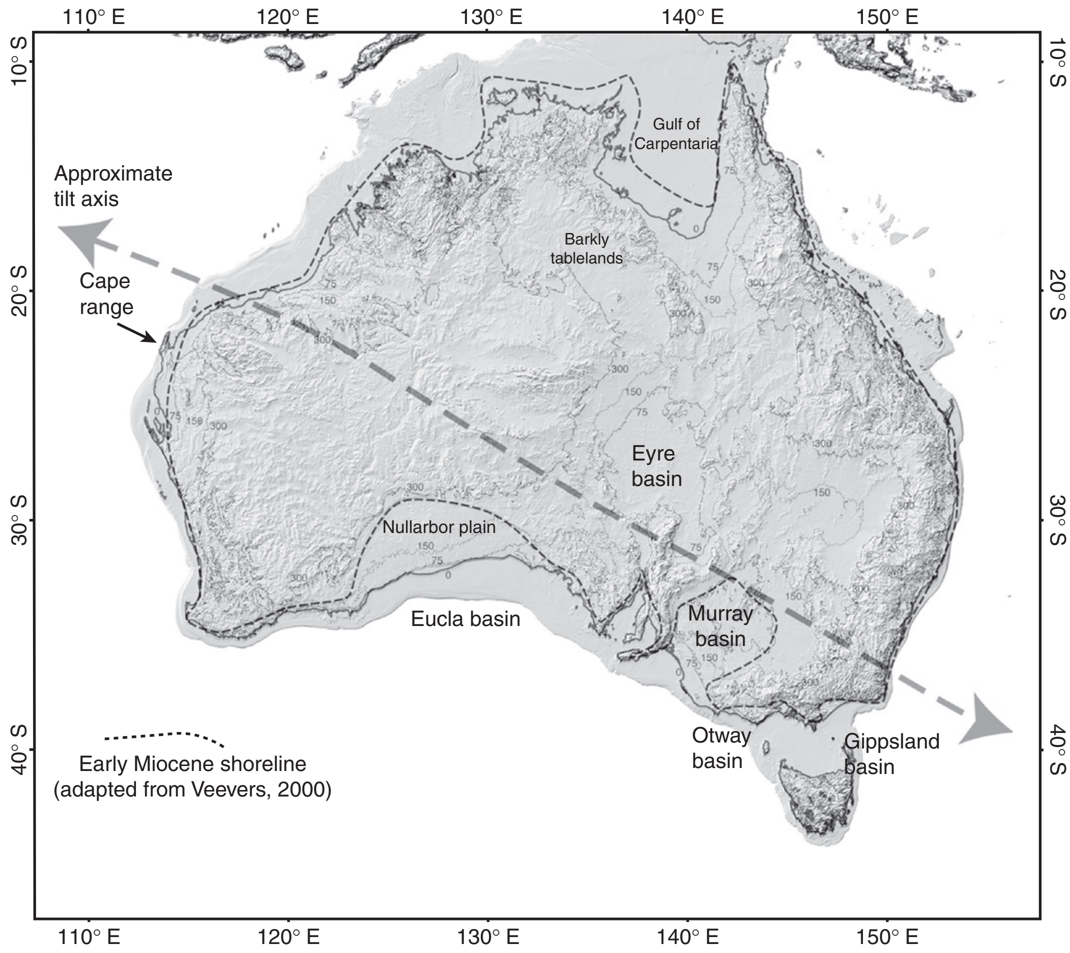

Figure 16 The tilting continent. Geological evidence, such as the position of the early Miocene shoreline and the asymmetry of the continental shelf, suggests that Australia is tilting northwards due to dynamic topographic effects. Reproduced with permission from Sandiford, M., 2007. The tilting continent: a new constraint on the dynamic topographic field from Australia. Earth and Planetary Science Letters. doi:10.1016/j.epsl.2007.06.023.

drainage predicted by the dynamic plume model, southern Africa is characterized by a remarkable pattern of three broadly concentric drainage divides, roughly parallel to the coastline, and also the mid-ocean ridges surrounding Africa (**Figure 17**). These three divides have been interpreted as flexure axes that reflect three episodes of epeirogenic uplift of southern Africa (du Toit, 1933; Moore, 1999), younging from the coast to the interior.

5.5.5.3 Plate Boundary Stresses and Lithospheric Buckling

Buckling of the whole lithosphere due to plate boundary stresses was first suggested for oceanic lithosphere (e.g., McAdoo and Sandwell, 1985), and has also been proposed for continental lithosphere on 100 km wavelengths that may be determined by the mechanical strength of the mantle (Tikoff and Maxson, 2001).

The ages inferred for the three flexure axes described above in southern Africa are broadly coeval with major changes in plate motion at the spreading ridges surrounding southern Africa. Thus, the marginal Escarpment Axis formed at the time of initial Gondwana breakup ($\sim$125 Ma) and would have migrated inland as an escarpment flexure, as envisaged by Gilchrist and Summerfield (1990). The age of the central Etosha-Griqualand-Transvaal (EGT) Axis correlates with a major shift in the pole of rotation of the Atlantic at $\sim$84 Ma (Nürnberg and Müller, 1991). The interior Ovambo-Kalahari-Zimbabwe (OKZ) Axis formed during the late Palaeogene, broadly coeval with both a reorganization of spreading regime in the Indian Ocean (Reeves and de Wit, 2000) and a marked increase in spreading rate in the mid-Atlantic (Nürnberg and Müller, 1991). The formation of each one of the flexure axes is marked in the offshore sedimentary record of Southern Africa by erosion and unconformities (McMillan, 2003), which also correspond with clusters of volcanic ages (Moore et al., 2008, 2009).

These correlations suggest that the flexure axes, and thus topography in southern Africa, could be primarily linked to deformation events associated with plate reorganizations. Although the exact mechanisms are speculative and require investigation, major episodes of erosion in the British Isles have also been correlated with plate boundary tectonic episodes (Hillis et al., 2008; Holford et al., 2009), which can also be recognized in distinct stratigaphic sequences separated by

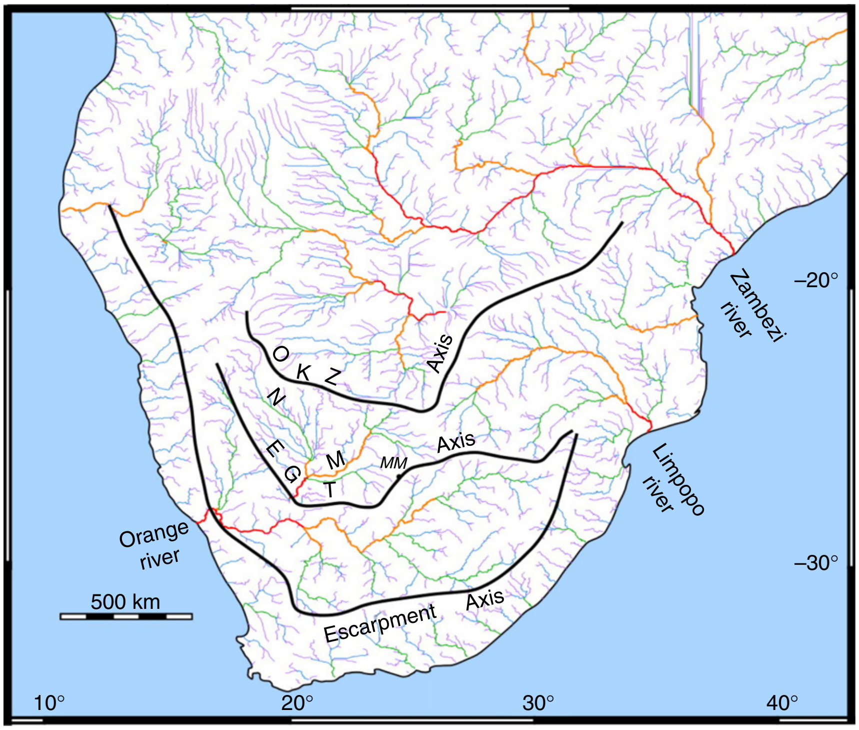

Figure 17 Concentric pattern of epeirogenic axes in Southern Africa. Colors denote stream order: purple - 1, blue - 2, green - 3, orange - 4, red - 5. M = Molopo River, N = Nossob River, MM = Mahura Muhtla. The major river divides are interpreted to reflect epeirogenic uplift Axes. EGT Axis = Etosha-Griqualand-Transvaal Axis; OKZ Axis = Ovambo-Kalahari-Zimbabwe Axis. Data from USGS EROS, http://eros.usgs.gov/products/elevation/gtopo30/hydro/af_streams.html, and Moore, A.E., Blenkinsop, T.G., Cotterill, F.P.D., 2009. Southern Topography and erosion history: plumes or plate tectonics. Terra Nova 21, 310–315.

unconformities (Stoker et al., 2010). The projection of compressional stresses over distances > 1000 km from oceanic ridges into the interiors of continental plates "can account for a broad spectrum of shortening-related intraplate deformation styles which vary in scale from upper crustal folding … to whole lithosphere buckling" (Holford et al., 2009).

On the scale of 100–1000 km, Australia preserves a record of Cenozoic uplift and sedimentary basin formation (e.g., Flinders ranges, Torrens and Frome basins) that shows topographic undulations of several hundred meters. These have been attributed to lithospheric buckling by Celerier et al. (2005) because of their correlation to Bouguer gravity anomalies. The state of stress in the Australian continent is largely compressional (The World Stress Map web site), and the orientation of the Flinders ranges (north–south) can be matched with the measured and modeled compressional (east–west) stresses prevailing from collisional boundary forces in the Himalaya, the Indonesian margin, and the New Zealand margin (Reynolds et al., 2002). In this example, there also appears to be a clear link between continental hinterland deformation and plate boundary stresses.

5.5.5.4 Implications for Continent-Wide Erosion Cycles and the Origin of Uplifts

A major criticism of the concept of cyclic erosion surfaces has always been the question of the mechanisms responsible for their initiation. A solution to this problem is hinted at by successive uplift along axes described above for southern Africa located progressively inland from the coast, initiated by intraplate stresses associated with spreading reorganizations at the ocean ridges. Each episode of epeirogenic tectonism would have rejuvenated the drainage network of southern Africa, thus providing a series of triggers to initiate new cycles of erosion in the continental hinterland (King, 1963; Lister, 1987; Partridge and Maud, 1987). The close correlation of the ages of these flexure axes with the major unconformities in the Congo Basin lends support to the development of contemporaneous erosion surfaces over wide areas of Africa, as postulated by King (1963). A reevaluation of erosion surfaces in southern African in relation to the three flexure axes offers a potential framework for refining understanding of their chronology and interrelationships.

The ages of the three flexure axes also correspond to continent-wide episodes of alkaline volcanism recognized by Bailey (1993), underlining the link between epeirogenesis and associated lithospheric stresses, and alkaline volcanism (Bailey, 1993; Moore et al., 2008). The broad upwarps represented by the flexure axes would be associated with relative tensional stresses in the upper surface of the plate. In contrast, the lower plate surface would experience relative tension beneath the basins surrounding the axes. This link between the uplift axes and the distribution of lithospheric stresses could explain the coastward younging of alkaline volcanism in southern Africa, which contrasts with the inland age progression of the three axes.

In summary, the observed topography of Southern Africa does not correlate well with dynamic topography predicted by plume models. Rather, the major features of the topography are determined by concentric flexural uplift axes, each coeval with an episode of plate boundary reorganization. This suggests that the dominant influence on the modern topography in southern Africa reflects stresses associated with plate kinetics, rather than mantle plumes.

5.5.6 Concluding Remarks

Passive margins are subject to a variety of sources of stress, mediated through several different processes, as summarized in **Figure 6**. The diversity of passive margins may reflect a corresponding variety in the relative importance of the processes shown in the figure. The clear temporal links between passive margin/continental hinterland evolution and plate tectonic reorganization demonstrated for southern Africa, and the directional links between basins and ranges and *in situ* stress in Australia, are suggestive evidence that the processes responsible for vertical movements are directly linked to first-order plate tectonic considerations. This link can also be extended, although with less confidence, to the earlier geological record of passive margins.

Responses to rifting vary along the length of passive margins and from one margin to another. Thus, marginal escarpments are typically absent in the vicinity of rifts associated with triple junctions. The local lithospheric flexural thickness will also influence the isostatic response and geomorphic evolution of passive margins. This is exemplified in the contrast between the marked flexural uplift on the Red Sea coastal plain and the more subdued response along the margin of southern Africa. Where rifted margins have a thick sedimentary cover, rapid deep erosion will follow, but this is not necessarily the case where basement rocks are exposed or where there is only a thin sedimentary capping.

Escarpments are perhaps the most spectacular and puzzling landforms associated with passive margins. There are three main current models for escarpment formation and evolution: flexural isostasy with backwearing, a pinned drainage divide with downwearing, and downwarping. Despite the application of newer techniques such as low-temperature thermochronology and numerical modeling, it is not yet clear which of these models is generally applicable, or whether more than one model, or combinations of them, may be appropriate for different passive margins. This remains an area for interesting future research.

This article has focused on Mesozoic–Holocene passive margins because their record is so much more visible, but passive margins can be traced in the geological record back to at least the late Archean, although their abundance and their duration seem to have changed through time (Bradley, 2008). The abundance of passive margins appears to be closely related to the assembly and breakup of Pangea, but not clearly to earlier hypothesized supercontinent cycles. The greater duration of passive margins in the Precambrian was interpreted by Bradley (2008) to indicate that, paradoxically, plate motion was slower in the past.

This article suggests that several fundamental issues still remain to be resolved about passive margin and continental hinterland tectonic geomorphology, which to some extent invoke larger tectonic questions. These issues can be summarized in the form of several questions:

1. Do mantle plumes play an important role, or indeed any role, in rifting and continental breakup?
2. How do escarpments form and evolve?
3. Is the high elevation of continental hinterland sustained over the long term, and if so, how?
4. Do detachments exist under passive margins?

The answers to these elusive questions are clearly critical in understanding the links between plate tectonics processes and intraplate responses.

Acknowledgments

We are grateful to Robert Dutton for assistance with the bibliography and to Jon Nott for valuable comments on an early draft. Excellent reviews by Paul Green and David Nash brought a necessary degree of circumspection to the first submission, and Lewis Owen did an excellent editorial job.

References

Bagguley, J., Prosser, S., 1999. The interpretation of passive margin depositional processes using seismic stratigraphy; examples from offshore Namibia. Geological Society Special Publications 153, 321–344.

Bailey, D.K., 1993. Petrogenic implications of the timing of alkaline, carbonatite and kimberlite igneous activity in Africa. South African Journal of Geology 96, 67–74.

Beaumont, C., Keen, C.E., Boutilier, R., 1982. On the evolution of rifted continental margins: comparison of models and observations for the Nova Scotia Margin. Royal Astronomical Society Geophysical Journal 70, 667–715.

Bishop, P., 2007. Long-term landscape evolution: linking tectonics and surface processes. Earth Surface Processes and Landforms 32, 329–365.

Bradley, D.W., 2008. Passive margins through Earth history. Earth-Science Reviews 91, 1–26.

Braun, J., van der Beek, P., 2004. Evolution of passive margin escarpments: what can we learn from low-temperature thermochronology? Journal of Geophysical Research 109, F04009. http://dx.doi.org/10.1029/2004JF000147.

Brown, R.W., Gallagher, K., Gleadow, A.J.W., Summerfield, M.A., 1999. Morphotectonic evolution of the south Atlantic margins of Africa and South America. In: Summerfield, M.A. (Ed.), Geomorphology and Global Tectonics. Wiley, Chichester, pp. 255–283.

Brown, R.W., Summerfield, M.A., Gleadow, A.J.W., 2002. Denudational history along a transect across the Drakensberg Escarpment of southern Africa derived from apatite fission track thermochronology. Journal of Geophysical Research 107, 2350. http://dx.doi.org/10.1029/2001JB000745.

Brown, R.W., Müller, R.D., Gaina, C., Struckmeyer, H.I.M., Stagg, H.M.J., Symonds, P.A., 2003. Formation and evolution of Australian passive margins: implications for locating the boundary between the continental and oceanic crust. Special Publication of the Geological Society of Australia 22, 223–243.

Burke, K., 1996. The African plate. South African Journal of Geology 99, 339–409.

Burke, K., Gunnell, Y., 2008. The African Erosion Surface: a continental-scale synthesis of geomorphology, tectonics, and environmental change over the past 180 million years. Geological Society of America Memoir 201, 1–66.

Celerier, J., Sandiford, M., Hansen, D.L., Quigley, M., 2005. Modes of active intraplate deformation, Flinders Ranges, Australia. Tectonics 24, http://dx.doi.org/10.1029/2004&C001679.

Conrad, C.P., Lithgow-Bertelloni, C., Louden, K.E., 2004. Iceland, the Farallon slab, and dynamic topography of the North Atlantic. Geology 32, 177–180.

Cox, K.G., 1989. The role of mantle plumes in the development of continental drainage patterns. Nature 342, 873–877.

Cox, K.G., 1993. Continental magmatic underplating. Philosophical Transactions of the Royal Society of London 342, 155–166.
Daradich, A., Mitrovica, J.X., Pysklywec, R.N., Willet, S.D., Forte, A.M., 2003. Mantle flow, dynamic topography, and rift-flank uplift of Arabia. Geology 31, 901–904.
De Wit, M.J., 2003. Madagascar: heads it's a continent, tails it's an island. Annual Reviews of Earth Sciences 31, 213–248.
Dingle, R.V., Scrutton, R.A., 1974. Continental break-up and the development of post-Paleozoic sedimentary basins around Africa. Bulletin of the Geological Society of America 85, 1167–1175.
Doucouré, C.M., de Wit, M.J., 2003. Old inherited origin for the present near bimodal topography of Africa. Journal of African Earth Sciences 36, 371–388.
Du Toit, A.L., 1930. The volcanic belt of the Lebombo: a region of tension. Transactions of the Royal Society of South Africa 18, 189–217.
Du Toit, A.L., 1933. Crustal movements as a factor in the geographical evolution of South Africa. South African Geographical Journal 16, 1–33.
Ebinger, C.J., Sleep, N.H., 1998. Cenozoic magmatism throughout east Africa resulting from the impact of a single plume. Nature 395, 788–791.
Emmel, B., Jakobs, J., Razakamanana, T., 2004. Titanite and apatite fission track analyses on basement rocks of central-southern Madagascar: constraints on exhumation and denudation rates along the eastern rift shoulder of the Morondava basin. Journal of African Earth Sciences 38, 343–361.
Fleming, A., Summerfield, M.A., Stone, J.O., Fifield, L.K., Cresswell, R.G., 1999. Denudation rates for the southern Drakensberg escarpment, SE Africa, derived from in-situ-produced cosmogenic ^{36}Cl: initial results. Journal of the Geological Society of London 156, 209–212.
Fouch, M.J., James, D.E., Vandecar, J.C., Van Der Lee, S.The Kaapvaal Seismic Group, 2004. Mantle seismic structure beneath the Kaapvaal and Zimbabwe Cratons. South African Journal of Geology 107, 33–44.
Fullard, H., Darby, H.C., 1974. The University Atlas (Sixteenth ed.). George Philip & Son. Ltd., London, 176 pp.
Gallagher, K., Brown, R., 1999. Denudation and uplift at passive margins: the record on the Atlantic margin of southern Africa. Philosophical Transactions of the Royal Society London A357, 835–859.
Gallagher, K., Brown, R.W., Johnson, C., 1998. Fission track analysis and its application to geological problems. Annual Review of Earth and Planetary Science 26, 519–572.
Geoffroy, L., 2005. Volcanic passive margins. Comptes Rendus Geoscience 337, 1395–1408.
Ghebreab, W., Talbot, C.J., 2000. Red Sea extension influenced by Pan-African tectonic grain in eastern Eritrea. Journal of Structural Geology 22, 931–946.
Gilchrist, A.R., Summerfield, M.A., 1990. Differential denudation and flexural isostacy in formation of rift-margin upwarps. Nature 346, 739–742.
Gilchrist, A.R., Summerfield, M.A., 1991. Denudation, isostacy and landscape evolution. Earth Surfaces Processes and Landforms 16, 555–562.
Gilchrist, A.R., Summerfield, M.A., 1994. Tectonic models of passive margin evolution and their implication for theories of long-term landscape development. In: Kirby, M.J. (Ed.), Process Models and Theoretical Geomorphology. Wiley, Chichester, pp. 55–84.
Gunnell, Y., Carter, A., Petit, C., Fournier, M., 2007. Post-rift seaward down-warping at passive margins: new insights from southern Oman using stratigraphy to constrain apatite fission-track and (U-Th)/He dating. Geology 35, 647–650.
Gunnell, Y., Fleitout, I., 2000. Morphotectonic evolution of the Western Ghats, India. In: Summerfield, M.A. (Ed.), Geomorphology and Global Tectonics. Wiley, Chichester, pp. 321–338.
Gurnis, M., Mitrovica, J.X., Ritsema, J., van Heijst, H.-J., 2000. Constraining mantle density structure using geological evidence of surface uplift rates: the case of the African superplume. Geochemistry, Geophysics, Geosystems 1, 1–44.
Hays, J., 1968. Notes on geomorphology. In: Walpole, B.P., Dunn, P.R., Randal, M.A. (Eds.), Geology of the Katherine-Darwin region, Northern Territory. Bureau of Mineral Resources, Geology and Geophysics Bulletin 82, Canberra, pp. 161–170.
Hillis, R.R., Holford, S.P., Green, P.F., et al., 2008. Cenozoic exhumation of the southern British Isles. Geology 36, 371–374.
Holford, S.P., Green, P.F., Duddy, I.R., Turner, J.P., Hillis, R.J., Stoker, M.S., 2009. Regional intraplate exhumation episodes related to plate boundary deformations. Geological Society of America Bulletin 121, 1611–1628.
Huismans, R.S., Podladchikov, Y.Y., Cloetingh, S.A.P.L., 2001. Transition from passive to active rifting; relative importance of asthenospheric doming and passive extension of the lithosphere. Journal of Geophysical Research 106(B6), 11,271–11,291.
Japson, P., Bonow, J.M., Green, P.F., Chalmers, J.A., Lidmar-Bergström, K., 2006. Elevated passive continental margins: long term highs or Neogene uplifts? New evidence from Greenland. Earth and Planetary Science Letters 248, 315–324.
Japson, P., Bonow, J.M., Green, P.F., Chalmers, J.A., Lidmar-Bergström, K., 2009. Formation, uplift and dissection of planation surfaces at passive continental margins – a new approach. Earth Surface Processes and Landforms 34, 683–699.
King, L.C., 1949. On the ages of African land-surfaces. Quarterly Journal of the Geological Society 104, 439–459.
King, L.C., 1955. Pedeplanation and isostacy: an example from South Africa. Quarterly Journal of the Geological Society 111, 353–359.
King, L.C., 1963. South African Scenery. Oliver and Boyd, Edinburgh, 308 pp.
King, L.C., 1976. Planation remnants upon high lands. Zeitschrift für Geomorphologie 20, 133–148.
Klausen, M.B., 2009. The Lebombo monocline and associated feeder dyke swarm: diagnostic of a successful and highly volcanic rifted margin. Tectonophysics 468, 42–62.
Kooi, H., Beaumont, C., 1996. Large-scale geomorphology: classical concepts reconciled and integrated with contemporary ideas via a surface processes model. Journal of Geophysical Research 101, 3361–3386.
Kooi, H., Beaumont, G., 1994. Escarpment evolution on high-elevation rifted margins: insights derived from a surface process model that combines diffusion, advection and reaction. Journal of Geophysical Research 99, 12191–12209.
Kounov, A., Niedermann, S., Viola, G., de Wit, M., Andreoli, G., 2007. Present denudation rates at selected sections of the South African escarpment and the elevated continental interior based on cosmogenic ^{3}He and ^{21}Ne. South African Journal of Geology 110, 235–248.
Kounov, A., Viola, G., de Wit, M., Andreoli, G., 2008. A Mid Cretaceous paleo-Karoo River valley across the Knersvlakte plain (northwestern coast of South Africa): Evidence from apatite fission-track analysis. South African Journal of Geology 111, 409–420.
Kounov, A., Viola, G., de Wit, M., Andreoli, G., 2009. Denudation along the Atlantic passive margin: new insights from apatite fission-track analysis on the western coast of South Africa. Geological Society of London Special Publication 324, 287–306.
Lister, G.S., Etheridge, M.A., Symonds, P.A., 1986. Detachment faulting and the evolution of passive margins. Geology 14, 246–250.
Lister, G.S., Etheridge, M.A., Symonds, P.A., 1991. Detachment models for the formation of passive continental margins. Tectonics 10, 1038–1064.
Lister, L.A., 1987. The erosion surfaces of Zimbabwe. Bulletin of the Zimbabwe Geological Survey, Harare 90, 163 pp.
Lithgow-Bertelloni, C., Gurnis, M., 1997. Cenozoic subsidence and uplift of continents from time-varying dynamic topography. Geology 25, 735–738.
Lithgow-Bertelloni, C., Silver, P., 1998. Dynamic topography, plate driving forces and the African superswell. Nature 395, 269–272.
Marsh, J.S., Hooper, P.R., Rehacek, J., Duncan, R.A., Duncan, A.R., 1997. Stratigraphy and Age of Karoo Basalts of Lesotho and Implications for correlations within the Karoo Igneous Province. American Geophysical Union Geophysical Monograph 100, 247–272.
Matmon, A., Bierman, P., Enzel, Y., 2002. Pattern and tempo of great escarpment erosion. Geology 30, 1135–1138.
McAdoo, D., Sandwell, D.T., 1985. Folding of oceanic lithosphere. Journal of Geophysical Research 90(B10), 8563–8569.
McKenzie, D.P., 1984. A possible mechanism for epeirogenic uplift. Nature 307, 616–618.
McMillan, I.K., 2003. Foraminiferally defined biostratigraphic episodes and sedimentary patterns of the Cretaceous drift succession (Early Barremian to Late Maastrichtian in seven basins on the South African and southern Namibian continental margin. South African Journal of Science 99, 537–576.
Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J., 2002. Characteristics of volcanic rifted margins. In: Menzies, M.A., Klemperer, S.L., Ebinger, C.J., Baker, J. (Eds.), Volcanic Rifted Margins. Geological Society of America (GSA), Boulder, CO, Special Paper 362, pp. 1–14.
Moore, A.E., 1999. A reappraisal of epeirogenic flexure axes in southern Africa. South African Journal of Geology 102, 363–376.
Moore, A.E., Blenkinsop, T.G., 2006. Scarp retreat versus pinned drainage divide in the formation of the Drakensberg escarpment, southern Africa. South African Journal of Geology 109, 455–456.

Moore, A.E., Blenkinsop, T.G., 2002. The role of mantle plumes in the development of continental-scale drainage patters: the southern African example revisited. South African Journal of Geology 105, 353–360.

Moore, A.E., Blenkinsop, T.G., Cotterill, F.P.D., 2008. Controls of post-Gondwana alkaline volcanism in southern Africa. Earth and Planetary Science Letters 268, 151–164.

Moore, A.E., Blenkinsop, T.G., Cotterill, F.P.D., 2009. Southern African topography and erosion history: plumes or plate tectonics. Terra Nova 21, 310–315.

Moore, A.E., Cotterill, F.P.D., Broderick, T.G., Plowes, D., 2009. Landscape evolution in Zimbabwe from the Permian to present, with implications for kimberlite prospecting. South African Journal of Geology 112, 65–86.

Moucha, R., Forte, A.M., Mitrovica, J.X., Rowley, D.B., Quere, S., Simmons, N.A., Grand, S.P., 2008. Dynamic topography and long-term sea-level variations; there is no such thing as a stable continental platform. Earth and Planetary Science Letters 271, 101–108.

Nott, J., 1994. Long-term landscape evolution in the Darwin region and its implications for the origin of landsurfaces in the north of the Northern Territory. Australian Journal of Earth Sciences 41, 407–415.

Nott, J., Horton, S., 2000. 180Ma continental drainage divide in northeastern Australia: role of passive margin tectonics. Geology 28, 763–766.

Nott, J.F., 1995. The antiquity of landscapes on the North Australian Craton and the implications for theories of long-term landscape evolution. Journal of Geology 103, 19–32.

Nott, J.F., Young, R.W., McDougall, I., 1996. Back-wearing versus down-wearing versus gorge extension in the long-term denudation of a highland mass: quantitative evidence from the Shoalhaven catchment, southeastern Australia. Journal of Geology 104, 224–233.

Nürnberg, D., Müller, R.D., 1991. The tectonic evolution of the south Atlantic from Late Jurassic to present. Tectonophysics 191, 27–53.

O'Grady, D.B., Syvitski, J.P.M., Pratson, L.F., Sarg, J.F., 2000. Categorizing the morphologic variability of siliciclastic passive continental margins. Geology 28, 207–210.

O'Sullivan, P.B., Kohn, B.P., Foster, D.A., Gleadow, A.J.W., 1995. Fission track data from the Bathurst Batholith: evidence for rapid mid-Cretaceous uplift and erosion within the eastern highlands of Australia. Australian Journal of Earth Sciences 42(6), 597–607.

Ollier, C.D., 1982. The Great Escarpment of eastern Australia: tectonic and geomorphic significance. Journal of the Geological Society of Australia 29, 13–23.

Ollier, C.D., 1984. Morphotectonics of continental margins with great escarpments. In: Morisawa, M., Hack, J.T. (Eds.), Tectonic Geomorphology. Allen & Unwin, Boston, pp. 3–24.

Ollier, C.D., 2004. The evolution of mountains on passive continental margins. In: Owens, P.N., Slaymaker, O. (Eds.), Mountain geomorphology. Edward Arnold, London, UK, pp. 59–88.

Ollier, C.D., Marker, M.E., 1985. The Great Escarpment of southern Africa. Zeitung für Geomorphologie 54, 37–56.

Ollier, C.D., Pain, C.F., 1997. Equating the basal unconformity with the palaeoplain: a model for passive margins. Geomorphology 19, 1–15.

Ollier, C.D., Powar, K.B., 1985. The Western Ghats and the morphotectonics of peninsular India. Zeitschrift für Geomorphologie 54, 57–69.

Ollier, C.D., Stevens, N.C., 1989. The Great Escarpment of Queensland. In: Maitre, Le (Ed.), Pathways in Geology – Essays in Honour of Edwin Sherbon Hills. Blackwells Scientific, Melbourne, pp. 140–152.

Pain, C.F., 1985. Morphotectonics of the continental margin of Australia. Zeitung für Geomorphologie 54, 23–35.

Partridge, T.C., Maud, R.R., 1987. Geomorphic evolution of southern Africa since the Mesozoic. South African Journal of Geology 90, 179–208.

Pazzaglia, F.J., Gardner, T.W., 1994. Late Cenozoic flexural deformation of the middle U.S. Atlantic passive margin. Journal of Geophysical Research 99(B6), 12,143–12,157.

Pazzaglia, F.J., Gardner, T.W., 2000. Late Cenozoic landscape evolution of the US Atlantic passive margin: insights into a North American Great Escarpment. In: Summerfield, M.A. (Ed.), Geomorphology and Global Tectonics. Wiley, Chichester, pp. 283–302.

Penck, W., 1924. Morphological Analysis of Landforms, English translation by Czeck, H. and Boswell, K.C., London 1953, 423 pp.

Persano, C., Stuart, F.M., Bishop, P., Barford, D.N., 2002. Apatite (U-Th)/He age constraints on the development of the Great Escarpment on the south-eastern Australian passive margin. Earth Planetary Science Letters 200, 1–14.

Reeves, C., de Wit, M.J., 2000. Making ends meet in Gondwana: retracing the transforms of the Indian Ocean and reconnecting the continental shear zones. Terra Nova 12, 272–280.

Reeves, C.V., 1978. A failed Gondwana spreading axis in southern Africa. Nature 273, 222–223.

Reynolds, S.D., Coblentz, D.D., Hillis, R.R., 2002. Tectonic forces controlling the regional intraplate stress field in continental Australia: results from new finite-element modelling. Journal of Geophysical Research 107, http://dx.doi.org/10.1029/2001JB000408.

Rohrman, M., Van der Beek, P., Andriessen, P., Cloetingh, S., 1995. Meso-Cenozoic morphotectonic evolution of southern Norway: neogene domal uplift inferred from apatite fission track thermochronology. Tectonics 14, 704–718.

Rosendahl, B.R., 1987. Architecture of continental rifts with special reference to East Africa. Annual Review of Earth and Planetary Sciences 15, 445–503.

Rowley, D.B., Sahagian, D., 1986. Depth-dependent stretching: a different approach. Geology 14, 32–35.

Royden, L., Keen, C.E., 1980. Rifting processes and the thermal evolution of the continental margin of eastern Canada determined from subsidence curves. Earth and Planetary Science Letters 51, 343–361.

Rust, D.J., Summerfield, M.A., 1990. Isopach and borehole data as indicators of rifted margin evolution in southwestern Africa. Marine and Petroleum Geology 7, 277–287.

Sandiford, M., 2007. The tilting continent: a new constraint on the dynamic topographic field from Australia. Earth and Planetary Science Letters 261, 152–163. http://dx.doi.org/10.1016/j.epsl.2007.06.023.

Sandiford, M., Quigley, M., 2009. TOPO-OZ: insights into the various modes of intraplate deformation in the Australian continent. Tectonophysics 474, 405–416.

Seaward, D., Grujie, D., Schreurs, G., 2004. An insight into the breakup of Gondwana: identifying events through low-temperature thermochronology from the basement rocks of Madagascar. Tectonophysics 23, TC3007. http://dx.doi.org/10.1029/2003TC001556,2004.

Sengör, A.M.C., Burke, K., 1978. Relative timing of rifting and volcanism on Earth and its tectonic implications. Geophysical Research Letters 5, 419–421.

Steckler, M.S., Omar, G.I., 1994. Controls on erosional retreat of the uplifted flanks at the Gulf of Suez and northern Red Sea. Journal of Geophysical Research 99(B6), 12159–12173.

Stoker, M.S., Holfrod, S.P., Hillis, R.R., Green, P.F., Duddy, I.R., 2010. Cenozoic post-rift sedimentation off northwest Britain: recording the detritus of episodic uplift on a passive continental margin. Geology 38, 595–598.

Struckmeyer, H.I.M., Symonds, P.A., 1997. Tectonostratigraphic evolution of the Townsville Basin, Townsville Trough, offshore northeastern Australia. Australian Journal of Earth Sciences 44, 799–817.

Summerfield, M.A., 1988. Global tectonics and landscape development. Progress in Physical Geography 12, 389–404.

Summerfield, M.A., 1991. Global Geomorphology: An Introduction to the Study of Landforms. Longman Scientific and Technical, London, 537 pp.

Summerfield, M.A. (Ed.), 2000. Geomorphology and Global Tectonics. Wiley, Chichester.

Tassara, A., Swain, C., Hackney, R., Kirby, J., 2007. Elastic thickness structure of South America estimated using wavelets and satellite-derived gravity data. Earth and Planetary Science Letters 253, 17–36.

Tikoff, B., Maxson, J., 2001. Lithospheric buckling of the Laramide foreland during Late Cretaceous and Paleogene, western United States Rocky Mountain. Geology 36, 13–15.

Tinker, J., de Wit, M., Brown, R., 2008. Mesozoic exhumation of the southern Cape, South Africa, quantified using apatite fission track thermochronology. Tectonophysics 455, 77–93.

Truswell, J.F., 1970. An Introduction to the Historical Geology of South Africa. Purnell, Cape Town, 167 pp.

Tucker, G.E., Slingerland, R.L., 1994. Erosional dynamics, flexural isostacy, and long-lived escarpments: a numerical study. Journal of geophysical Research 99, 12229–12243.

Turcotte, D.L., Emerman, S.H., 1983. Mechanisms of active and passive rifting. Tectonophysics 94, 39–50.

Twidale, C.R., 2002. The two-stage concept of landform and landscape development involving etching: origin, development and implications of an idea. Earth Science Reviews 57, 37–74.

Van der Beek, P., Pulford, A., Braun, J., 2001. Cenozoic development in the Blue Mountains (southeast Australia): lithological and tectonic control on rifted margin morphology. The Journal of Geology 109, 35–56.

Van der Beek, P., Summerfield, M.A., Braun, J., Brown, R.W., Fleming, A., 2002. Modelling post breakup landscape development and denudational history across the southeastern African (Drakensberg Escarpment) margin. Journal of Geophysical Research 107, B12. http://dx.doi.org/10.1029/2001JB000744.

Wheeler, P., White, N., 2000. Quest for dynamic topography: observations from Southeast Asia. Geology 28, 963–966.
White, R.S., Mckenzie, D.P., 1989. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. Journal of Geophysical Research 94, 7685–7730.
Young, R.W., 1989. Crustal constraints on the evolution of the continental divide of eastern Australia. Geology 17, 528–530.

Relevant Website

http://dc-app3-14.gfz-potsdam.de/pub/introduction/introduction_frame.html
World Stress map.

Biographical Sketch

Tom Blenkinsop is Professor and Head of Discipline in the School of Earth and Environmental Sciences, James Cook University, Australia, where he has been working since 2002 after teaching and researching at the University of Zimbabwe in Harare for 13 years. His main interests are in structural geology and the areas of overlap between this subdiscipline, tectonics, and landscape evolution. His interest and experience in passive margins and continental interiors derive from studies in southern Africa and Australia, which make several fascinating comparisons. As a structural geologist, he is particularly interested in faulting and fluid flow in the upper crust, and the role of structures in the formation of mineral deposits. These themes are linked by considerations of the past and present states of stress in the Earth's crust.

Andy Moore is Vice President (Exploration, Diamonds) of African Queen Mines – a Canadian-listed mineral exploration company, and also an honorary Research Associate of the Geology Department of Rhodes University, South Africa. He has been involved in diamond exploration in Botswana, Zimbabwe, Namibia, Madagascar, and South Africa since 1980. He initiated and managed private diamond exploration companies in Botswana – an undertaking that proved to be a far more interesting and character-building way to part with most of one's capital than the gaming tables. Then, concerned about possible character deficiencies, he commenced diamond prospecting in Zimbabwe. Initial research interests were focused on kimberlite mineralogy and geochemistry and their application to the economic evaluation of kimberlites. Kimberlite prospecting led to an interest in drainage evolution in southern Africa and its application to interpreting the dispersion of kimberlite pathfinder minerals away from the primary source rocks. An unexpected spin-off was exploring the linkage between landscape evolution and the evolution and dispersion of plant fish and mammal species. A more recent curiosity has been the temporal links between continental tectonics that modified drainage evolution, and both episodes of alkaline volcanism, and the spreading history of the ocean ridges surrounding southern Africa. In essence, an archetypal kimberlite prospector – jack of all trades and master of none.