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THE ROYAL SOCIETY

Denudation and uplift at passive margins: the record on the Atlantic Margin of southern Africa

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The onshore region of a passive margin forms an integral part of the geological evolution from continental break-up to later sedimentation in the offshore basins. The dominant surface process in the onshore region is denudation, which acts to remove any direct evidence of surface uplift. However, denudation can be constrained on geological time-scales through low temperature thermochronological data, such as that obtained through apatite fission track analysis. Here, we present a suite of such data from the Atlantic Margin of southern Africa. The data have been modelled in terms of their temperature histories since the Jurassic. These temperature histories have been combined with heat-flow data to estimate the equivalent depth of denudation over these time-scales. Average denudation rates are of the order of a few tens of metres per million years, but show considerable variations both temporally and spatially. These results demonstrate that passive margins experience complex patterns of denudation. Three landscape-evolution models are considered. Our results imply the downwarping model is inappropriate. The other two models, scarp retreat and pinned drainage divide, predict trends similar to those observed but the complexities inherent in the data and the evolution of passive margin topography do not allow us to resolve one from the other. In practice, both models probably operate to some extent as a margin evolves. Estimates of palaeotopography have been made, assuming simple isostatic response models to denudational unloading. Flexural models with effective elastic thickness (EET) of 25 km predict elevations 2 km and more above the present-day values, while models with EET of 0 km predict elevations up to 750 m higher than the present day. These models ignore any post-break-up tectonic uplift and need independent constraints on surface elevation to assess their validity.

Keywords: denudation; fission track analysis; passive margin; southern Africa; Namibia

1. Background

Passive continental margins are formed within a regional extensional environment dominated by crustal thinning and subsidence. However, a complete explanation of passive margin evolution must account for the evolution of the subaerial topography of the margins in addition to the subsidence of the offshore marginal basins. Physical mechanisms proposed to explain elevated margins can be conveniently divided into transient and permanent mechanisms. The former rely on the thermally induced effects of rifting, such as dynamic secondary convection or the more passive isostatic response to lateral heat flow or depth-dependent extension. These mechanisms can only explain uplift on the time-scale of the thermal time constant of the lithosphere (*ca.* 60–70 Ma), as the thermal perturbation driving uplift will decay. Thus, these models are inappropriate for margins like the South Atlantic which formed more than 130 Ma ago. In contrast, permanent uplift mechanisms do not predict such time-dependent decay. These mechanisms rely on the isostatic response of the lithosphere to rifting, such as mechanical unloading and plastic necking incorporating flexure, or to magmatic underplating.

Models producing tectonic uplift have proven particularly difficult to constrain because, apart from the thermal decay, the models do not address how the topography evolves with time. More problematic is that, unlike the stratigraphy of the offshore basins, there is no obvious record of vertical motion. In principle, it is possible to make some estimate of surface uplift using markers such as shallow marine sediments, coals (generally assumed to be low elevation), laterite surfaces (often assumed to form at low elevation, but probably only require low relief) or palaeobotany. However, these will tell us little about rates of uplift unless we have features of different ages and most of the techniques provide information on denudation processes operating over relatively short time-scales (thousands of years). More problematically, denudation acts to remove these potential records of surface motion, and this problem becomes more acute for older margins.

In an attempt to address the long-term evolution of passive margin topography, considerable effort has been expended over the past five years or so by geophysicists developing numerical simulations (Gilchrist *et al*. 1994; Kooi & Beaumont 1994; Braun & Sambridge 1997; van der Beek & Braun 1998). These simulations are generally based around simple deterministic rules and can produce extremely realistic looking landscapes. However, it is not always obvious what controls the landscape evolution nor the physical meaning of various input parameters. Moreover, a variety of deterministic models can produce effectively identical final landscapes. While these models are an important advance in our understanding of surface process and the link to tectonics, it is desirable to constrain these models with independent observations.

Therefore, one starting point for considering uplift in the context of passive margin evolution is to address the question of constraining long-term denudation. If we can quantify the denudation chronology in a region, we can at least estimate the contribution of isostatic rebound to the present-day elevation, which may lead to the characteristic upwarp of many passive margin escarpments (Gilchrist & Summerfield 1990). However, the denudation chronology of a passive margin is important in itself in that denudation of the passive margin mountains provides much of the detritus ultimately deposited in the offshore basins.

2. Constraining long-term denudation

Summerfield (1991, ch. 15) gives an overview of the classical approaches for constraining landscape evolution and the nature of denudation. These include characterizing the nature of solid and solute loads in rivers and quantifying erosion into surfaces of known age (e.g. valley incision into a dated volcano). The first approach suffers from considerable uncertainties such as measuring both the suspended and bedloads. Moreover, both approaches, and indeed many related geomorphological techniques, may rely on extremely short-term samples, which include anthropogenic influences.

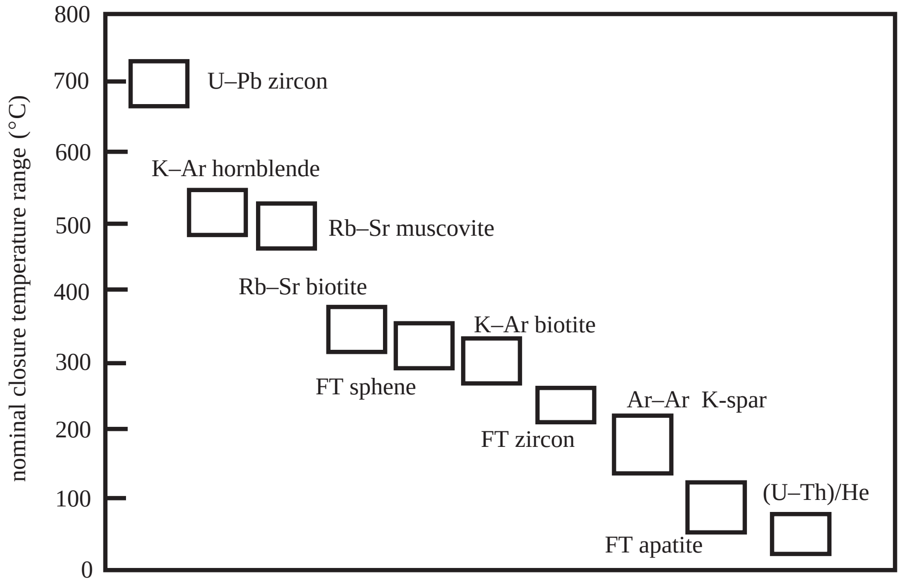

Figure 1. Nominal closure temperature ranges for a selection of mineral-isotope systems. The low-temperature systems such as apatite fission track analysis and (U–Th)/He are most suitable for constraining denudation and surface processes on geological time-scales.

For longer time-scale denudation rates, we need to consider the geological record. Rust & Summerfield (1990) considered the volumes of sediment preserved in the Orange Basin, offshore South Africa and used these to infer equivalent denudation rates. The major limitation with this approach remains reliable constraints on the area of the eroding source region over time.

Geochronological methods, particularly those sensitive to temperature, seem to have the most promise for constraining long-term denudation in the eroding source region. These methods are generally based somewhat loosely on the concept of a closure temperature (Dodson 1973). Thus, the measured age or date for a given mineral-isotopic decay system is interpreted in terms of the time at which the host rock passed through the closure temperature for that particular mineral system. As originally defined by Dodson (1973), the closure temperature represents the temperature reached during cooling where the daughter element is lost by some physical process more rapidly than it is produced by decay of the parent element. Figure 1 shows some representative closure temperature estimates for a variety of different mineral-isotope systems.

Although the effective closure temperature is cooling rate dependent and typically there is a range in temperature over which the loss of the daughter element occurs, figure 1 serves to illustrate the applicability of these techniques for constraining timing and rates of geological processes. For quantifying denudation on passive margins, where we expect a few kilometres of section to be removed, the methods sensitive to lower temperatures such as apatite fission track analysis or (U–Th)/He are the most appropriate. Apatite fission track analysis is a well-established technique and has been widely applied to passive margin studies (see, for example, Brown *et al*. 1999; Gallagher & Brown 1997). In contrast, (U–Th)/He thermochronometry (Zeitler *et* *al*. 1987; Wolf *et al*. 1997, 1998) has not been widely exploited to date, although it has great potential for yielding key information on cooling histories between 75 and 40 °C.

The basis of fission track analysis and the applications to geological problems have recently been reviewed by Gallagher & Brown (1997), Gallagher *et al*. (1998) and Gleadow & Brown (1999), and we will not go over the details presented therein. However, there are some basic points worth reiterating for the sake of continuity.

Fission track data can be considered in terms of the fission track age and track length distribution which, simplistically, provide information on timing and temperature respectively. Physically, the important process is the shortening or annealing of fission tracks with increasing temperature. On geological time-scales (1–100 Ma), significant annealing occurs over a temperature range of about 50–120 °C, and the length of an individual track is primarily a function of the maximum temperature to which it is exposed. As fission tracks are formed continuously over time, each track will experience a different proportion of the host rock's thermal history. Consequently, the combined fission track-length distribution and age data reflect the integrated thermal history, but it is the track-length data that are critical for extracting this information.

The other important factor is of course a predictive model for fission track annealing as a function of temperature and time. Currently, these models rely heavily on empirical calibrations (Laslett *et al*. 1987; Carlson 1990; Crowley *et al*. 1991), rather than a sound physical basis. However, the models provide a good description of the laboratory data and generally produce acceptable agreement with observations when extrapolated to geological time-scales. A variety of methods incorporating these empirical models has been proposed to model fission track data in terms of an optimal data-fitting thermal history. As the predictive models are generally nonlinear yet straightforward to evaluate, the data-fitting methods generally rely on some form of stochastic sampling/simulation technique (Corrigan 1991; Gallagher 1995; Willett 1998). These techniques are straightforward to implement, and one potential advantage of well-formulated sampling-based methods is that the sampling can be used to construct useful posterior probability distributions (Gilks *et al*. 1996).

3. Models of passive margin landscape evolution

The onshore regions of passive margins display a complex and variable range of morphologies with topographic profiles (both parallel to and perpendicular to the margin) varying significantly between and along margins (Brown *et al*. 1999). For example, some passive margins are characterized by a well-defined escarpment, separated the continental interior from the low elevation, typically low-relief coastal plain, which is adjacent to the offshore sedimentary basin. Classic examples of these high-elevation margins are western southern Africa (South Africa and Namibia), southeast Brazil, southeast Australia, western India and the Red Sea. There are also examples of margins with little topographic expression, such as Argentina and southern Australia, which have no obvious topographic escarpments but have well-developed offshore basins. The morphological differences onshore reflect a complex relationship between denudation, drainage, lithology and tectonics, implying that a single model of passive margin formation, however complex, is unlikely to provide a satisfactory explanation for all margins.

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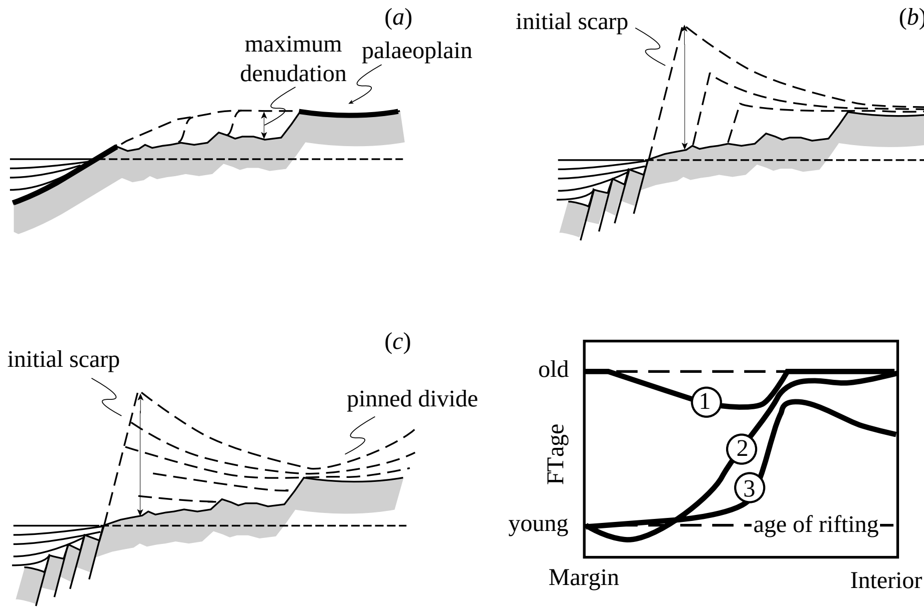

Figure 2. Schematic representation of passive margin landscape evolution and the spatial distribution of expected apatite fission track ages. See text for details.

To date, there are three broad classes of models for landscape evolution of high-elevation passive margins: the downwarp, scarp retreat and pinned-divide models (figure 2). The downwarp models (King 1962; Ollier & Pain 1997) propose that the margin is initially formed by long-wavelength downflexing of the lithosphere with limited faulting. This leads to a broad monocline. Denudation is expected to occur between remnants of the initial land surface preserved along the coast and the crest of the escarpment, inland of which only negligible amounts of denudation occur. Consequently, the distribution of apatite fission track age across the margin would be characterized by older ages (at least as old or older than the initial surface) occurring at the coast and inland of the escarpment with only moderately reduced ages in between.

These downwarp models neglect the isostatic response to the denudational unloading and this must be considered a serious limitation. In contrast, the scarp retreat models (Gilchrist & Summerfield 1990; Gilchrist *et al*. 1994; Tucker & Slingerland 1994) propose that an initial escarpment is formed at the new continental margin by differential vertical displacement across normal faults separating the subsiding rift basin from the subaerial part of the margin. The highest rates of denudation are predicted immediately seaward of the initial scarp as a consequence of the high relief. This relief-dependent erosion rate leads to scarp retreat as the scarp front always has the highest relief. The associated isostatic rebound can lead to the evolution of the coast-parallel upwarp geometry, characteristic of many passive margin ranges. Seaward of the retreating scarp, the coastal region is characterized by more moderate denudation rates and the interior region experiences relatively slow denudation. This would produce a strong gradient in apatite fission track age across the margin with the oldest ages occurring inland of the scarp and decreasing toward the

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coast. The youngest ages may be similar to, or younger than the time of rifting and continental break-up. An important prediction of these models is that the timing of maximum denudation decreases inland from the coast toward the final position of the escarpment.

The pinned-divide class of model (Kooi & Beaumont 1994; Gilchrist *et al*. 1994) also incorporates a steep escarpment formed during continental rifting. The key difference between these models and the scarp retreat described earlier is the incorporation of an inland drainage divide, defined by a gentle slope toward the rift scarp. The presence of this drainage divide leads to rapid incision of streams draining seaward of the divide and the formation of a new scarp close to the initial position of the drainage divide. This results in essentially uniform rates of denudation in the region seaward of the drainage divide, producing a downwearing rather than a scarp retreat pattern of denudation. Furthermore, if the base level of the interior drainage is lowered simultaneously with the formation of the initial scarp, then a significant amount of denudation may occur inland of the drainage divide. In this case the new scarp is still formed, and is 'pinned', at the position of the initial drainage divide (Kooi & Beaumont 1994). Thus, the important differences between the scarp retreat models and the pinned-divide models are the predicted timing of denudation across the margin and the amount of denudation predicted inland of the final position of the escarpment. As shown in figure 2, these differences are potentially resolvable from apatite fission track data.

4. The Atlantic Margin of southern Africa

(*a*) *Geological background and evidence for palaeo-elevation and denudation*

Brown *et al*. (1999) have discussed the morphological and geological characteristics of the South Atlantic Margins of both Africa and South America. Here we reiterate some points relevant to southern Africa, but refer the reader to the earlier paper for the detail.

The topography of the Atlantic Margin of southern Africa is variable. For example, the South West Cape region of South Africa is dominated by the high-relief mountains of the Cape Fold Belt, which locally reach elevations of over 2000 m and are underlain by the folded and resistant quartzite-dominated lithologies of the Siluro-Ordovician Table Mountain Group rocks. Further inland, a well-defined escarpment with a mean elevation of *ca.* 1500 m and relief of *ca.* 800m is underlain by the flat-lying sedimentary strata of the Permo-Triassic Karoo Sequence. Here the escarpment is coincidental with the southern limit of outcropping Mid-Jurassic dolerite sills. To the north the escarpment is less well defined and is replaced by a highland region underlain by the crystalline metamorphic rocks of the Namaqua metamorphic belt, which separates the high-relief coastal region from a low-relief interior plateau at *ca.* 1000 m elevation. North of the Orange River valley, the Margin topography is again dominated by a well-defined escarpment, defined by the western boundary of the Nama Group rocks, which form a near-horizontal capping to the highly deformed crystalline basement beneath. Between *ca.* 23° S and 28° S the region immediately inland of the escarpment has been dissected to varying degrees by the Fish River drainage system, which is a major tributary of the Orange River.

In Namibia, the escarpment extends northward to *ca.* 22° S, where it merges with the northeast-trending highland region of the Khomas Hochland, representing the

Damara metamorphic belt. Further north, the Margin topography is characterized by a gradual rise from the coast to the elevated interior. In southern Angola (north of 18° S), the Margin is once again characterized by a well-defined escarpment region, which reaches elevations of over 2500 m in places.

Across southwestern Gondwana, aeolian sandstones indicate terrestrial, semi-arid palaeoclimatic from the latest Triassic to earliest Cretaceous (Martin 1973; Dingle *et al*. 1983; Visser 1995). The Early Cretaceous Paraná-Etendeka flood basalts were erupted subaerially and directly onto these aeolian sandstones. In both Namibia and Brazil, intercalated sand dunes tens of metres thick between some flows show that the continental conditions prevailed during the eruptions (Milner *et al*. 1995*a*). The Paraná-Etendeka flood basalt province is attributed to the influence of a mantle plume (Hawkesworth *et al*. 1992; Gallagher & Hawkesworth 1994), which may produce some dynamic uplift, of the order of 500–1000 m. Because of the terrestrial nature of the sedimentation, there is no direct stratigraphic evidence for such uplift. Cox (1989) suggested that present-day drainage patterns in Namibia reflect long-lived uplift associated with plume-induced magmatic underplating in the Late Jurassic–Early Cretaceous. However, this argument is based primarily on the existence of drainage away from the present-day elevated Margin, which may be a consequence of erosional rebound and flexure of the Margin following break-up (Gilchrist & Summerfield 1990).

Early Cretaceous mafic dykes related to final break-up around 132–135 Ma are widespread along the African Margin (Reid *et al*. 1990, 1991). The depth of intrusion of these dykes is difficult to constrain, but depths less than 1–2 km seem unlikely, and would indicate a minimum level of post break-up erosion. Similarly, Early Cretaceous alkaline intrusive complexes (135–125 Ma; Watkins *et al*. 1994; Milner *et al*. 1995*b*) provide some indirect constraints on post-intrusive denudation. For example, the Early Cretaceous Brandberg complex in northeast Namibia has a peak elevation of *ca.* 2600 m, although the regional land surface is *ca.* 2000 m lower and Okenyenya (*ca.* 70 km further east, 1902 m) has *ca.* 1000 m relief indicating that at least this much denudation has occurred locally. The Cape Cross and Messum centres nearer the continental margin are the same age but are deeply eroded, which suggests 1–2 km as a minimum for post-intrusive denudation by analogy with the two previous locations.

The post-Etendeka geology in western southern Africa is dominated by the extensive but thin (generally less than 200 m) Kalahari Basin (Thomas & Shaw 1990). The age of the base of the Kalahari Basin sequence is thought to be Late Cretaceous to earliest Tertiary (Thomas & Shaw 1990; Partridge 1993). In the northern Etendeka province of Namibia, the lava sequence is preserved within a narrow, coast-parallel, fault-bounded half-graben (downthrown to the west). In the same area, a conglomerate deposit, consisting entirely of basaltic clasts derived from the west, was deposited within an active half-graben structure (Ward & Martin 1987). These structures clearly post-date the volcanism, and indicate significant tectonism and erosion of the lava sequence at some time after *ca.* 124 Ma.

Numerous intrusive kimberlite and associated alkaline pipes of Late Cretaceous age occur within South Africa and southern Namibia. At some of these sites, crater facies sedimentary sequences are still preserved indicating minimal denudation (less than 100 m) in these areas since the Late Cretaceous–Early Tertiary (Smith 1986; de Wit *et al*. 1992). In Namibia, terrestrial sedimentation during the Cenozoic was largely restricted to a 150 km wide zone along the southern sector of the Margin, which is the current region of the Namib Sand Sea (Ward 1987, 1988). The sequence is thin (generally less than 300 m) and the oldest unit, the Tsondab Sandstone Formation, which overlies the basement unconformity surface, is probably diachronous with an Early Palaeocene maximum age (Ward 1987; Ward & Corbett 1990). It is highly significant that remnants of this unit occur immediately seaward of the present escarpment region in the vicinity of the Kuiseb River in Namibia, indicating that the escarpment at this locality has not retreated appreciably (less than 20–30 km) since the earliest Tertiary (Ward 1987).

Offshore, the Margin of Namibia and South Africa is dominated by four main basins. From south to north these are the Orange, Luderitz, Walvis and Namibe Basins, the latter occurring to the north of the Walvis Ridge and having more in common with the Angolan Shelf to the north (Light *et al*. 1993). The main depocentres of these basins contain thick sequences of clastic sediments and volcanics, generally greater than 6 km and reaching over 12 km in the northern Walvis Basin (Maslanyj *et al*. 1992; Rust & Summerfield 1990; Gerrard & Smith 1982). Sedimentation within the developing marginal rift basins offshore of southwest Africa was dominated by oxidized, terrestrial, volcaniclastic sandstones, evaporites and subaerial volcanics. Within the central rift the transition to marine depositional environments occurred shortly after break-up and is marked everywhere by a well-developed ‘drift-onset’ unconformity (6At-1 Sequence Boundary; Brown *et al*. 1995) at the end of the Hauterivian (*ca.* 130 Ma ago). In addition to this major unconformity, locally variable unconformities occur in the Cenomanian–Turonian (97–88 Ma ago) and Upper Maastrichtian (*ca.* 68 Ma ago) sequences.

Although rifting began in the Middle Jurassic and break-up finally occurred during the Early Cretaceous, the major volume of sediment within the Orange and Walvis Basins was deposited during the Late Cretaceous–Early Tertiary (Brown *et al*. 1995; Rust & Summerfield 1990). Rust & Summerfield (1990) determined that this volume of *ca.* 2.8×10^6 km^3 (adjusted to equivalent rock volume) is equivalent to an average depth of denudation of 1.8 km over the whole of the Orange River catchment (an area of 1.55×10^6 km^2). However, the pattern of denudation onshore is likely to have been highly variable, being dependent on post-break-up tectonics, the pattern of drainage development, style of landscape evolution and long-term climatic and lithological variations.

(*b*) *Constraining the long-term denudation chronology of the Atlantic Margin of southern Africa*

A substantial suite of fission track data has been collected from the southwest African Margin (Haack 1983; Brown 1992) and the locations of the samples are shown in figure 3. The stratigraphic ages of the samples from South Africa range from Precambrian (Namaqua metamorphic belt) through to Late Triassic (Stormberg Group, Upper Karoo Sequence), while the samples from Namibia were collected from Precambrian to Cambrian metamorphic, igneous and sedimentary rocks. However, the apatite fission track ages of these samples lie between 449 ± 20 Ma and 59 ± 3 Ma. The fact that all of the apatite fission track ages are significantly younger than the stratigraphic age of the host rocks indicates that all the sampled rocks have been subjected to substantially higher temperatures in the past (mostly in

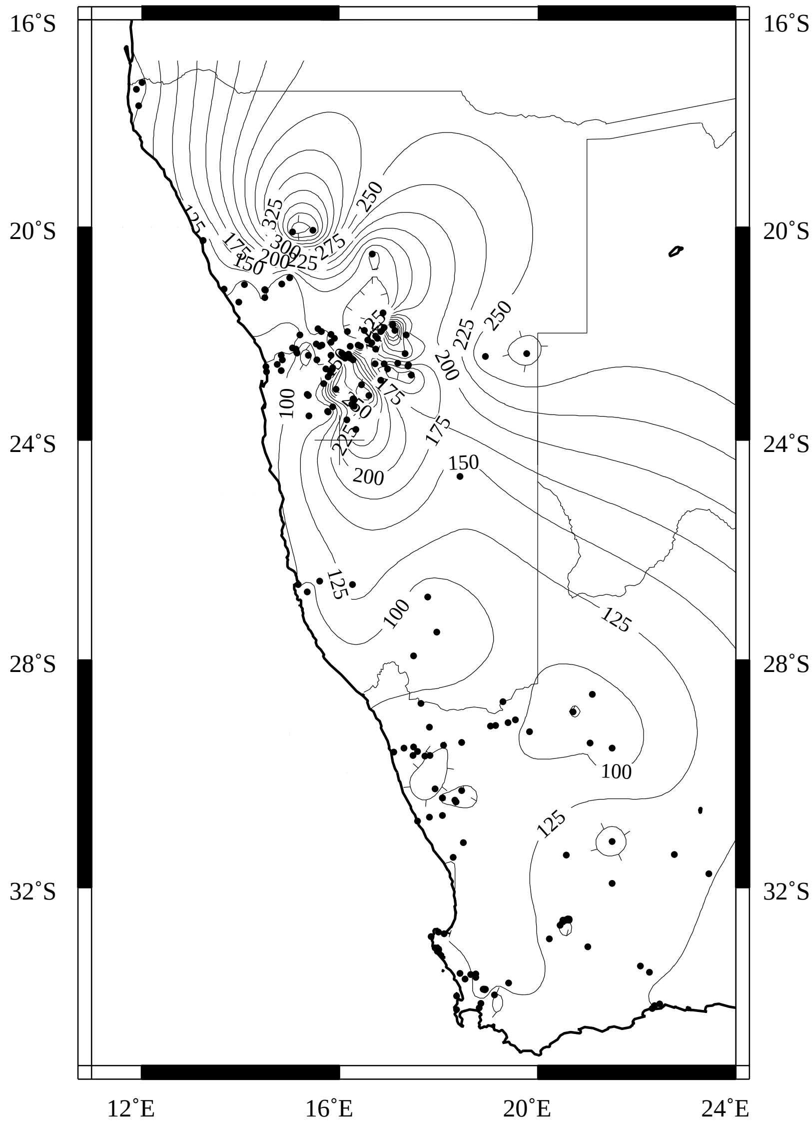

Figure 3. Location of apatite fission track samples on the Atlantic Margin of southern Africa and the contoured fission track age data. The ages contours include data from Brown (1992) and Haack (1983), whereas all modelling discussed later only uses the data of Brown (1992) as no track length data were available for the Haack (1983) data.

excess of *ca.* 110 °C). Ages pre-dating break-up at *ca.* 134 Ma were only obtained from samples from the interior regions of the continent. Ages post-dating break-up (*ca.* 134 Ma) are typical of the coastal region but they also occur within the interior. This regional pattern is in marked contrast to other passive margins where there is a clear relationship between apatite age and distance from the continental margin (Gallagher & Brown 1997; Gallagher *et al.* 1998).

More direct information on the chronology of cooling and the range of palaeotem-

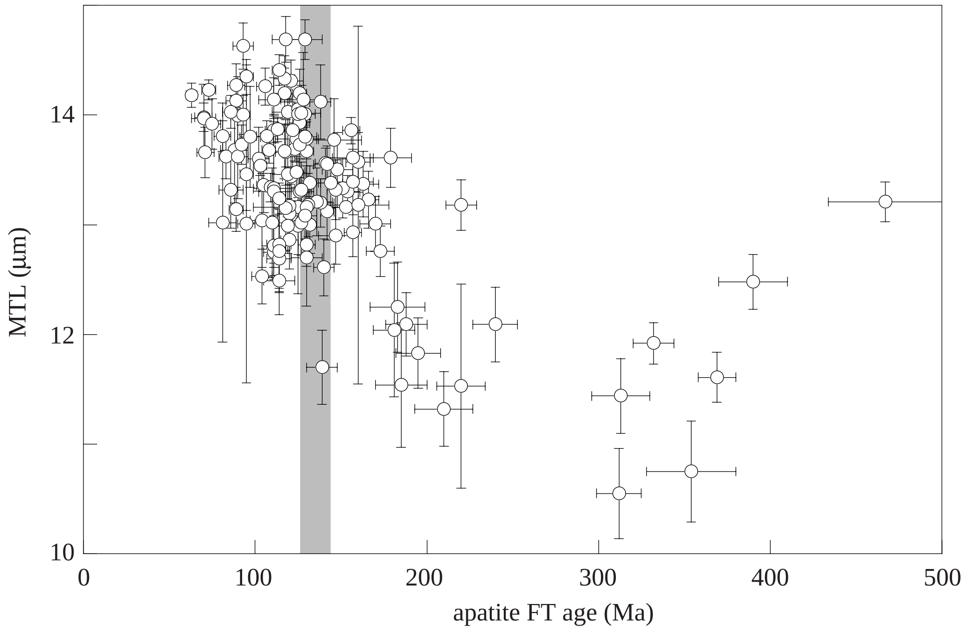

Figure 4. Relationship between apatite fission track age and mean track length (MTL) for southern Africa data (Brown 1992). The grey band marks the time of continental break-up.

peratures can be derived by examining the variation of mean track length with apatite age (figure 4). The youngest ages (60–75 Ma) are associated with narrow, unimodal track length distributions (standard deviations of 0.9–1.6 μm) with long mean lengths (14.6 ± 0.1 to 13.6 ± 0.12 μm). Older ages are associated with broader more complex and sometimes distinctly bimodal track length distributions with short mean lengths (*ca.* 11–12.5 μm). Long mean track lengths suggest discrete cooling episodes occurred at approximately the timing indicated by the associated fission track age (Brown *et al*. 1994; Gallagher & Brown 1997). The overall pattern indicates that those samples with mean track lengths greater than *ca.* 14 μm cooled from palaeotemperatures in excess of *ca.* 110 °C to near surface temperatures during the Early Cretaceous (140–120 Ma) while others have cooled from similar temperatures during the Mid-Cretaceous (100–80 Ma). The significantly shortened mean track lengths for most of the remaining samples indicates that they spent a protracted period at temperatures between 110 and 60 °C before cooling to surface temperatures during the Late Cretaceous.

The most significant difference between the data from Namibia and South Africa is the maximum age and the proportion of older apatite ages. All the samples yielding apatite ages older than *ca.* 150 Ma occur in northern Namibia, and the age distribution defines a distinct NE-trending corridor of ages younger than *ca.* 100 Ma separating regions characterized by substantially older ages (greater than *ca.* 200 Ma). The orientation of this corridor is similar to the regional structural trend of the intracontinental branch of the Pan-African Damara metamorphic belt, and appears to be coincident with the northern central zone as defined by the structural and metamorphic architecture of the belt (Miller 1983).

Although there are insufficient data from southern Namibia to clearly define a regional pattern, it is nevertheless significant that apatite ages of *ca.* 70 Ma were obtained from samples within the vicinity of the Karas Mountains (*ca.* 28° S). The NW orientation of the major shear zone structures bounding these Mountains is similar to other major transcurrent structures adjacent to the western margin of the Kaapvaal Craton, and samples from this region in South Africa yielded latest Cretaceous ages with long mean track lengths.

Overall, the data from Namibia indicate substantial cooling during the Cretaceous with a discrete phase of accelerated cooling identified in the latest Cretaceous (*ca.* 80–60 Ma). The geographical extent and magnitude of this latter episode was clearly highly variable, and the pattern of cooling within the interior appears to be closely related to the geometry of pre-existing crustal structures. Consequently, the pattern of denudation implied by these data cannot be simply related to the formation of the passive margin during the Late Jurassic–Early Cretaceous. An appropriate regional model must account for an Early Cretaceous phase of erosion, probably linked to the early rift stage of Margin development, as well as a later more variable phase of accelerated denudation associated with the tectonic reactivation of regional structures. The evolution of the Orange Drainage Basin (Dingle & Hendey 1983; Rust & Summerfield 1990; de Wit 1993), regional lithological differences (e.g. the distribution, both vertically and horizontally, of basalts, Karoo sediments and basement) and a change from a temperate, seasonal and wet climate to an arid climate during the latest Cretaceous–Early Tertiary (Ward 1987; Ward & Corbett 1990; Partridge 1993) no doubt also played an important role in controlling denudation rates along the Margin.

(*c*) *Quantifying long-term regional denudation*

As mentioned earlier, empirical annealing models can be used to model thermal histories from fission track data. For regional data-sets collected from outcrop samples, the thermal history information is best viewed in terms of contours of palaeotemperature at a given time. These individual time-slices can be easily combined into a computer animation to view the cooling history as overburden was removed and the rock samples approached the surface.†

In figure 5, we show a selection of these palaeotemperature maps, constructed using the modelling approach described by Gallagher (1995). The general trend is one of progressive cooling since continental break-up in the Early Cretaceous, but there are also temporal and spatial variations as anticipated from the earlier discussion of this data.

The temperature histories can be converted to equivalent depth, using the following relationship for one-dimensional conductive heat transfer,

$$z = (T - T_{\mathrm{s}})\frac{k}{Q}, \tag{4.1}$$

where z is depth, T is the modelled temperature, T_{s} is the surface temperature, k is the thermal conductivity, and Q is the heat flow. The typical magnitude of denudation rates in passive margin settings (less than $100\ \mathrm{m\ Ma^{-1}}$) in this setting means that advective heat transfer is not significant, and we have also neglected the influence of topography on subsurface temperature, which is a reasonable assumption

† See http://www.diamond.ge.ic.ac.uk/kgalla/WWW/Namibia.html for examples.

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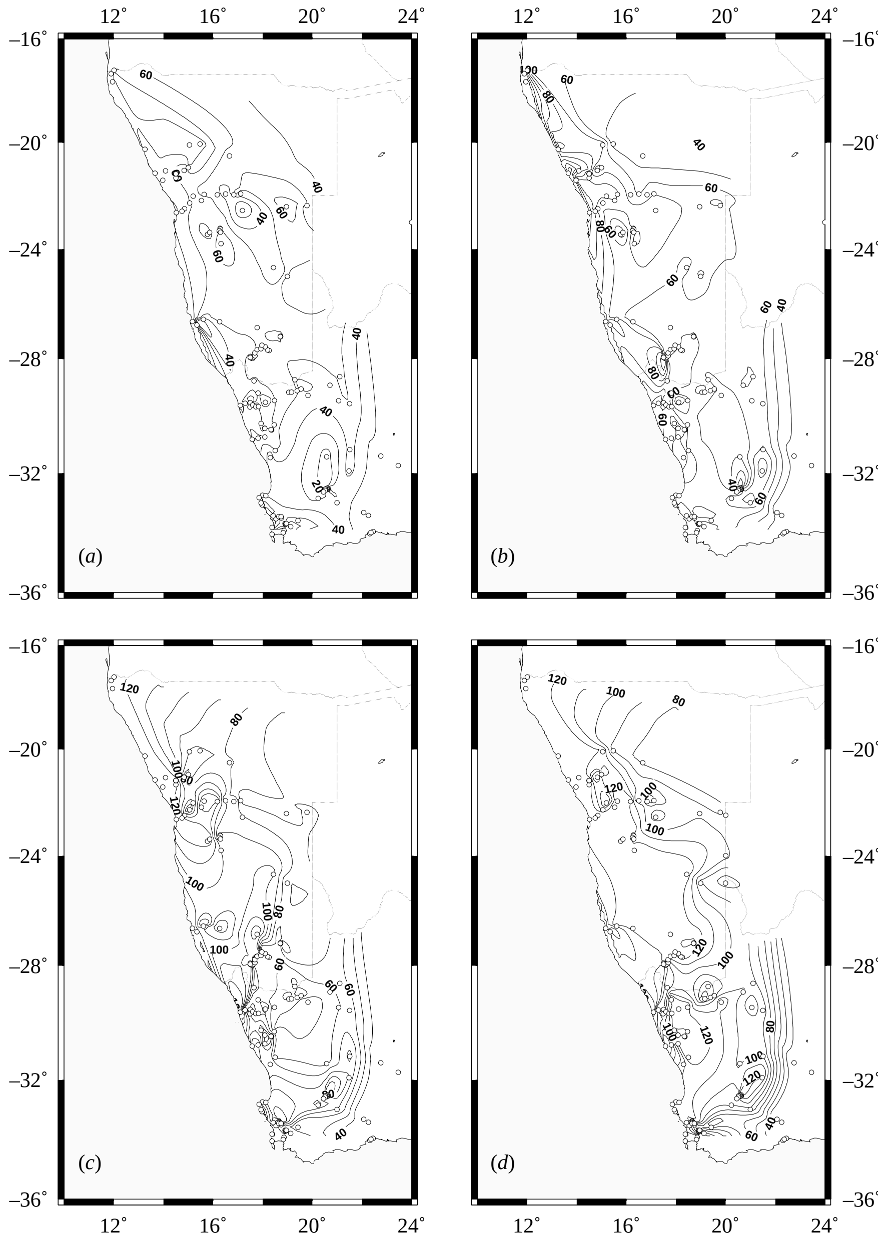

Figure 5. Palaeotemperature maps for (*a*) 40, (*b*) 70, (*c*) 100, (*d*) 140 Ma, based on the temperature history models derived from the apatite fission track data. The contours are in °C.

given the temperature range of 50–120 °C over which significant annealing occurs (Brown 1991). We have used the available present-day heat-flow data from southern Africa (Pollack *et al*. 1993). Gallagher *et al*. (1994) have shown that although heat

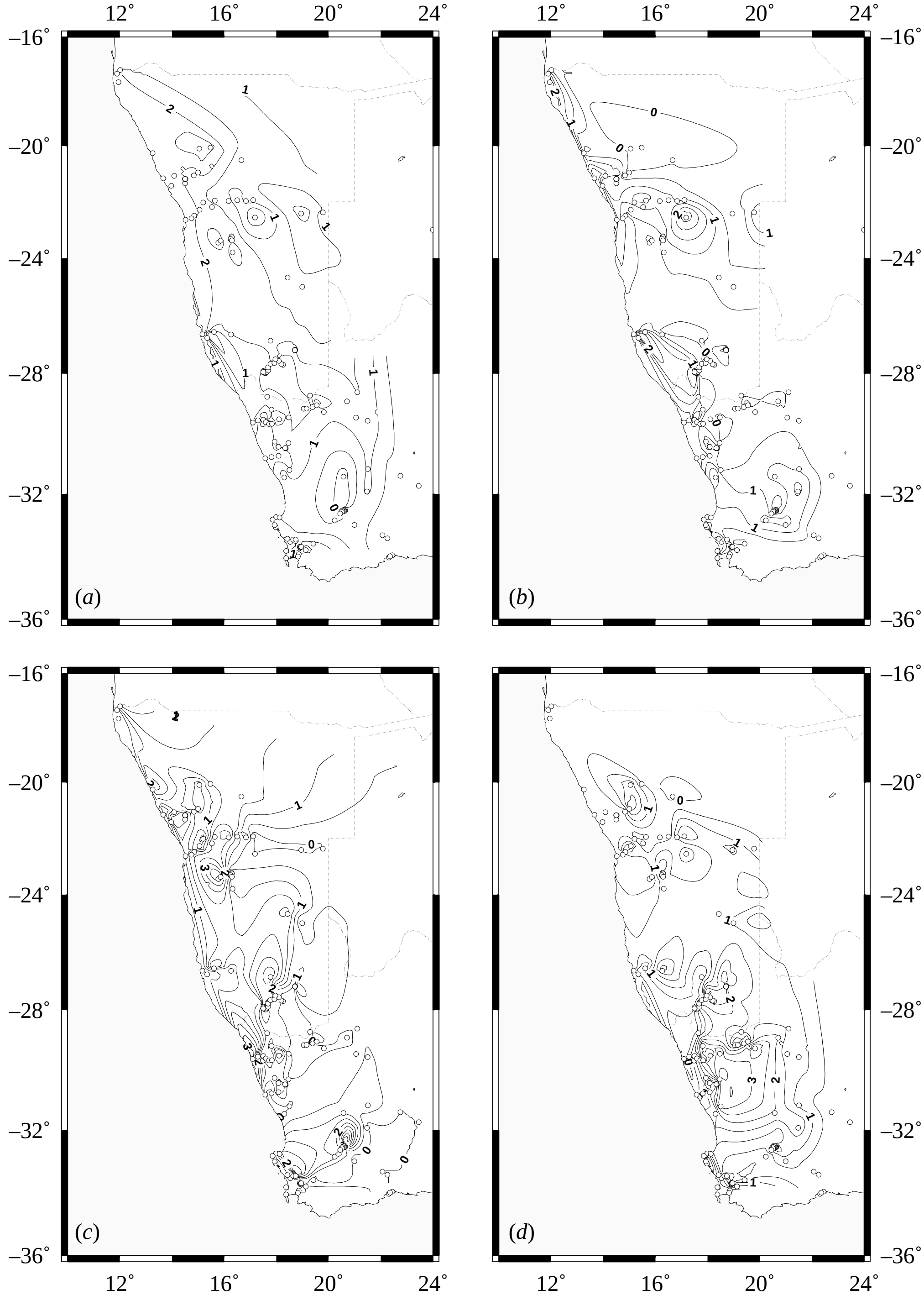

Figure 6. Denudation maps for four time-intervals (*a*) 0–40, (*b*) 40–70, (*c*) 70–100, (*d*) 100–140 Ma derived from the palaeotemperature maps, and using the present-day heat-flow data from southern Africa. The contours are in kilometres.

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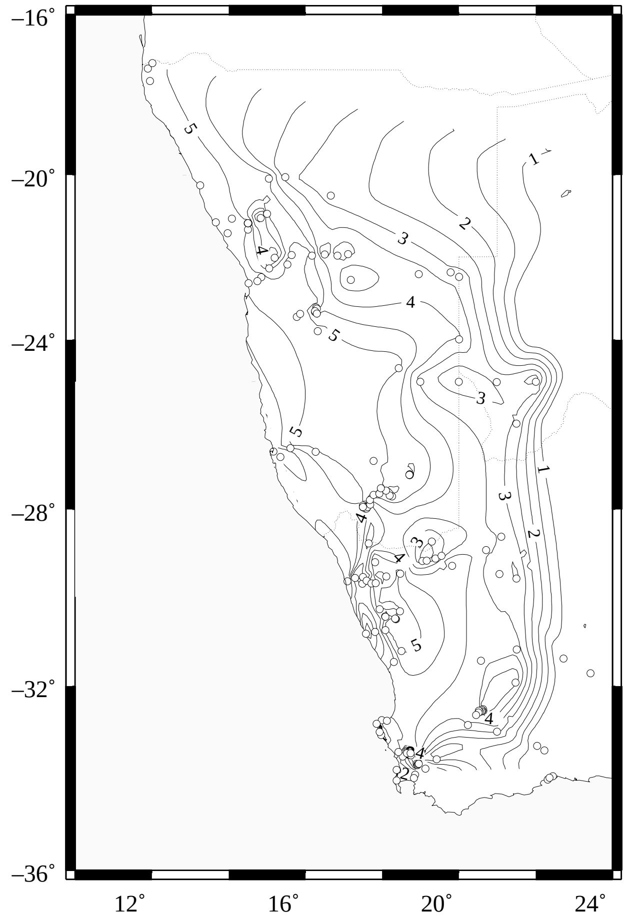

Figure 7. The total amount of denudation since 140 Ma, inferred from the apatite fission track data. The contours are in kilometres.

flow is elevated during extension, the increase in heat flow experienced in the rift flanks is relatively subdued. Given that we have no constraints on how heat flow might have varied over time, we choose the least complex model. Therefore, we do not vary heat flow over time, although in principle this is straightforward to do. We also assume a constant thermal conductivity of 2.5 W m^{-1} K^{-1} for the eroded section and a constant surface temperature of 20 °C.

Figure 6 shows contour maps of denudation between different time-intervals, and in figure 7 we show the total denudation since 140 Ma. It is clear that the region closest to the Margin has generally experienced more erosion that the inland regions, but there are local variations in denudation, and significant amounts of denudation can occur well inland from the Margin. These model results show that the downwarp models discussed earlier are not compatible with the observed data, neither in terms of the spatial distribution of denudation nor its evolution over time. Overall, the patterns of denudation are complex, reflecting a combination of different processes, of which the scarp-retreat/pinned-divide models might be regarded as end members.

Perhaps a more accessible representation of these results for the Margin as a whole

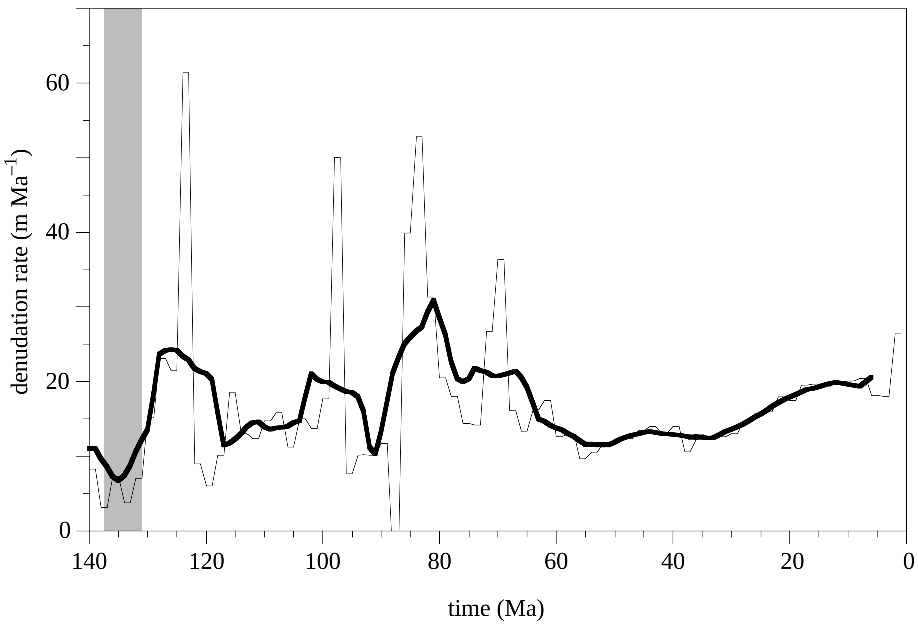

Figure 8. Spatially integrated denudation rate chronology for the Atlantic Margin of southern Africa. The lighter curve is the raw estimate from the denudation maps, and the heavy curve is a smoothed version. The shaded band indicates the inferred time of initiation of sea-floor spreading and the Paraná-Etendeka volcanism.

is in terms of a spatially integrated denudation chronology or denudation rate history (figure 8). This implies a period of accelerated denudation occurring around the time of break-up which may represent a major change on base-level related to the creation of rift margin topography. Rifting may have been activated as early as the Triassic but was certainly active by the Late Jurassic at the latest and led to ocean spreading by the earliest Cretaceous (135–130 Ma; Light *et al*. 1993; Clemson *et al*. 1997). The major volcanic episode associated with break-up produced the majority of the Paraná-Etendeka flood basalt province between 138 Ma and 128 Ma (Turner *et al*. 1994; Stewart *et al*. 1996). As can be seen in figure 8, there is an apparent lag between the inferred timing of peak denudation related to rifting and the timing of break-up by 5–10 Ma. However, it is not clear that this relatively short lag is real, and the resolution of the timing of cooling events inferred from apatite fission track analysis depends on the temperature from which cooling is initiated, the rate and amount of cooling (Gallagher 1995). In order to characterize features in these denudation chronologies that can be considered well resolved both in magnitude and timing, we require the addition of robust confidence intervals allowing for the propagation of uncertainties in the thermal histories through to the integrated denudation chronology. This is an area of ongoing work.

In addition to the enhanced denudation around the timing of break-up, there are also episodes of accelerated denudation occurring well after break-up. The most notable is around 70–90 Ma, and this reflects the Late Cretaceous episode discussed earlier, some 40–60 Ma after break-up and is not considered to reflect a long-term lag between any rift-related uplift and subsequent rapid denudation. In terms of

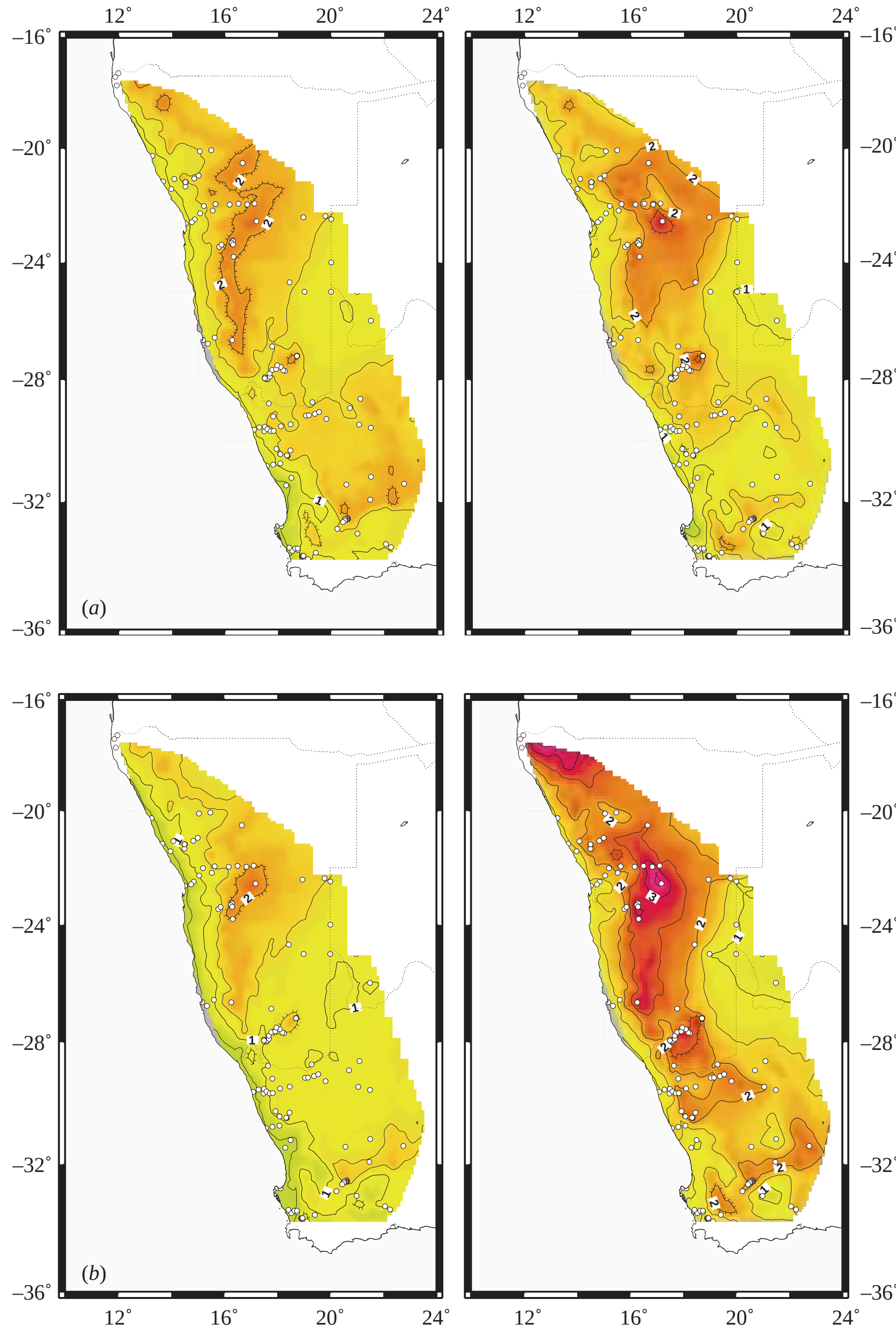

Figure 9. Reconstructed palaeotopography for the Atlantic Margin of southern Africa, calculated by loading the present-day topography with the estimated denudation, and adjusting for isostasy assuming EET of 0 and 25 km (left and right columns, respectively). (*a*) 40, (*b*) 70, (*c*) 100, (*d*) 140 Ma. The contours are in kilometres.

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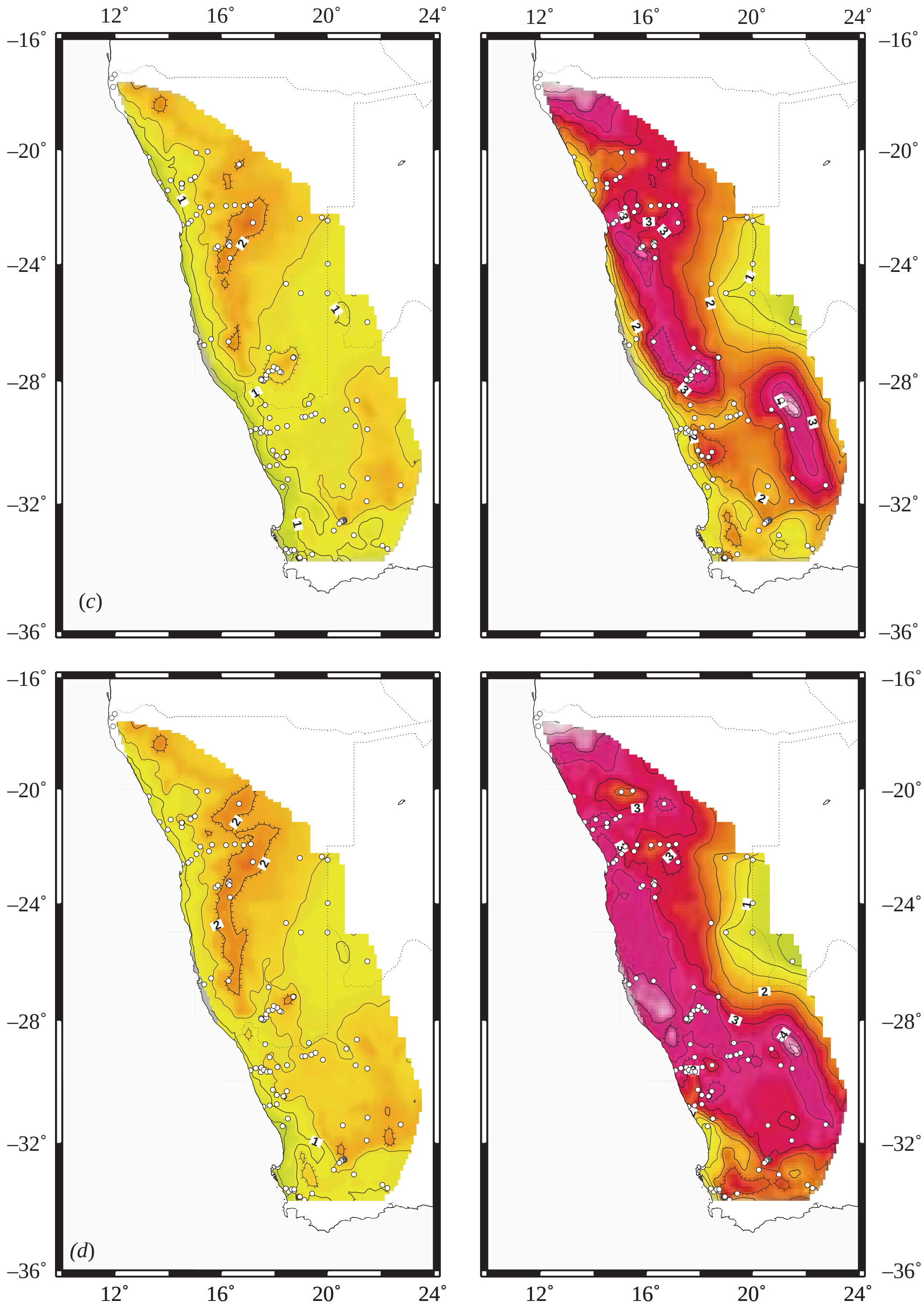

Figure 9. (*Cont.*)

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timing, this correlates with the inference of a global sea-level high in the latest Cretaceous (Haq *et al*. 1987). However, a higher sea-level would probably reduce river gradients and regional denudation rates would also tend to be reduced. This period of enhanced denudation appears to reflect in part regional reactivation of large structures in response to changes in plate motion in the Late Cretaceous (Brown *et al*. 1999).

Gallagher & Brown (1999) have linked a similar denudation chronology for Namibia alone to the offshore depositional record, using isopachs constructed from seismic reflection data. The major problem for correlating the denudation and depositional chronologies in this region is the relatively poor time resolution in the offshore stratigraphy. However, the flux of sediment shows a broad correlation to the onshore chronology, particularly in showing an increase in the Late Cretaceous. This occurs some 50 Ma after the initial break-up, by which time the tectonic subsidence is well into the thermal sag phase. This apparently rapid sedimentation therefore reflects accelerated delivery of clastic detritus from the onshore region, rather than a tectonic contribution in the offshore basin.

(*d*) *Implications for uplift and palaeotopography*

As mentioned earlier, it is extremely difficult to quantify surface elevation changes, and fission track data per se do not provide any direct constraints on uplift. However, we do expect vertical motion in response to denudation unloading, in much the same way as glacial rebound occurs in response to ice-sheet unloading. Indeed, the scarp-retreat model of Gilchrist & Summerfield (1990) provides an explanation for the characteristic upwarp morphology of many passive margin uplifts in terms of a regional isostatic response to denudation. Here, we have modelled the change in surface elevation as a consequence of the inferred denudation, using the following relationship

$$H = H_0 - D + R, \tag{4.2}$$

where H is the palaeo-elevation, H_0 is the present-day elevation, D is the denudation up to the present day, and R is the isostatic rebound associated with the denudational unloading. The simplest class of models for regional isostatic rebound relies on the thin elastic sheet approximation, and requires the specification of an effective elastic thickness (EET) or alternatively the flexural rigidity. The Airy or local isostatic model can be considered a limiting case for an EET of zero. The upper limit for EET in continents is somewhat more controversial, and estimates range up to 130 km. McKenzie & Fairhead (1997) have suggested that these estimates should be considered really as upper bounds as a consequence of low power in the short wavelength topography. They suggest 25 km as a more reliable upper limit.

In figure 9, we show the reconstructed palaeotopography using EETs of 0 and 25 km, calculated using a crustal density of 2750 kg m^{-3} and a mantle density of 3300 kg m^{-3}. Figure 10 shows the present-day elevation of the same region for comparison. An EET of zero is equivalent to Airy isostasy, and consequently, the elevation increases linearly as the denuded section is replaced and there is no lateral coupling between different regions. The elevation increases by a maximum of *ca.* 750 m, relative to the present day. For the case with an EET of 25 km, the maximum elevation is as much as 3 km higher than the present day. Also, the peak elevation in the

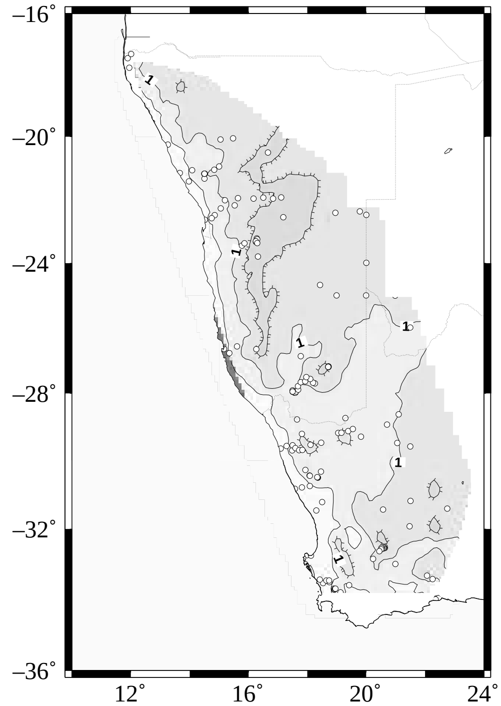

Figure 10. Present-day elevation for the region shown in figure 9. The contours are in kilometres.

model results moves inland with time.† Intriguingly, this implies a regional retreating scarp. However, this inference is model dependent and should be treated with due circumspection.

A more significant qualification is that the reconstructed palaeotopography reflects only the passive response to denudation unloading. There is no allowance for any post-break-up tectonic uplift, nor indeed any changes in elevation associated with the dynamic processes of rifting or local fault reactivation. One refinement to this modelling approach would be to use palaeogeographical indicators (e.g. marine sediments of different ages, but unfortunately not applicable in southern Africa), and to make the palaeotopography honour such data. Similarly, there is some rationale for incorporating evidence for palaeodrainage directions, where available, to refine these models, or at least to improve their geological consistency.

5. Concluding remarks

Models of passive margin evolution require some consideration of the onshore region as this is the major source of detrital material for the offshore basins. The dominant surface process operating onshore is denudation, which means that only young margins are considered likely to preserve any direct record of surface uplift. To date,

† See http://diamond.ge.ic.ac.uk/kgalla/WWW/3Dtopo.qt.

such evidence has proved elusive and certainly low-temperature thermochronological methods, such as apatite fission track analysis, cannot provide direct constraints on surface uplift. However, it is possible to reconstruct the denudation chronology on older margins using appropriate thermochronological data. The inferred patterns of denudation for various margins are complex and vary both spatially and temporally, implying that no one model of landscape evolution will explain all passive margins. Denudation rates are of the order of a few tens of metres per million years. Periods of accelerated denudation can occur well after continental break-up and these episodes can be reflected in the depositional history recorded in the offshore basins. In terms of the three general models of landscape evolution considered, neither the data nor the modelling results support the downwarp model of Ollier & Pain (1997). However, it is likely that the other two models (scarp retreat and pinned divide/downwearing) both operate to some extent on geological time-scales and are manifested on a variety of length-scales in different margins. Southeast Australia is one example where the drainage divide lies inboard of the current escarpment and no single model can provide an adequate explanation for the evolution of this margin (P. Bishop, personal communication). Considering the problem of constraining uplift and palaeotopography, it is possible to reload the lithosphere with the estimated missing section and correct for isostasy. However, this approach assumes no tectonic uplift occurs in the post-break-up phase of the margin's evolution, and consequently provides limited insight into the surface uplift history. Additional palaeogeographical data (e.g. palaeoshorelines, palaeodrainage directions) need to be incorporated to assess the validity of the reconstructions.

Many of the maps presented here were constructed using GMT (Wessel & Smith 1991) and also interpolation code provided by Malcolm Sambridge. We also thank Mike Summerfield, Paul Bishop, Joe Cartwright, Roy Miller and Mike de Wit for many useful discussions. An anonymous reviewer provided a careful review of the original manuscript. This work was supported in part by De Beers, NAMCOR, the Namibian Geological Survey and NERC (grant GR9/1573).

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Discussion

M. Osmaston (*Woking, UK*). Can Dr Gallagher say a bit more about the spikes on the graph in figure 8 that he had ignored and smoothed out?

K. Gallagher. Well, that comes down to the way we modelled the data. I said we used the smoothest model and I was slightly disingenuous there. It is not quite the smoothest model. What we do is to try to stop unconstrained oscillations in the thermal history, but also to try to let periods of rapid cooling occur if those are what are required to fit the data. So what we end up with are thermal histories that are quite flat but that can contain steps, almost like seismic velocity–depth profiles. Consequently, you can get a period of quite rapid cooling over one or two million years because that is the type of information we are trying to extract from the data. However, I am not convinced that these features are all real because they include the statistical uncertainty in the solutions. For example, if you plot a confidence region, say around a particular time/temperature point, the time is quite poorly constrained although the temperature may be quite tightly constrained. Thus, you end up with

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confidence regions around a time/temperature point which are typically elongated along the time axis, although not necessarily elliptical in shape.

My feeling is that even though we model some quite discrete cooling periods they are present in our models because we are encouraging that to happen. I am not convinced they are always real, so the cooling could happen over 5 Ma, which would reduce the cooling rate by a factor of five relative to 1 Ma. So my justification for smoothing the denudation chronologies is that I think that is a way, at least qualitatively, of allowing for some of the uncertainty in the timing which we do not include explicitly in the denudation chronologies.

M. Osmaston. I feel that an incomplete understanding of epeirogenic processes could be leading you to throw out important information.

K. Gallagher. I think frankly that smoothing is the sensible thing to do at this stage. Of course, another parameter is required when deciding how to smooth things. Some of the discrete features evident in the denudation chronologies may well be real. We always show the unsmoothed version as well as the smoothed one, so you can pick what you want. I would leave that to you.

C. Ebinger (*University of Leeds, UK*). What independent sources of information are there for the denudation of the Kaapvaal craton in Cretaceous times?

K. Gallagher. Our model results indicate that 80–90 Ma ago there was a rapid erosion event over a quite large area of a supposedly stable craton. In terms of independent data, it is well known that around 85 ± 5 Ma ago widespread kimberlite intrusion occurred in the Kaapvaal craton. At the same time there was an increase in sedimentation rate in basins offshore the east coast (Mozambique) by a factor of about 30, and this was relatively short-lived (about 5–10 Ma). Also at this time, sediments deposited offshore Namibia reflect rapidly shallowing conditions and an influx of clastic material. Well, you may ask whether there is a process-oriented link between the generation of the kimberlites, the rapid offshore sedimentation and our inferred rapid denudation onshore, or are we seeing just a series of temporal coincidences? We believe the former may be correct. My colleague, Rod Brown, could give you a much more detailed exposition of the cause; he has been working with Bill Griffin from Macquarie University in Australia, who has done some work looking at the P–T conditions of xenoliths from the Kaapvaal craton. Griffin has divided these into two groups (pre- and post-90 Ma) based on the age of the host rock. Due to the differing mineralogical and chemical nature of these two groups, he has inferred that the geotherm (as inferred from the xenolith data) changed around 90 Ma but it is not particularly clear exactly when. It looks as though the lower lithosphere became hotter soon after 90 Ma ago. So the question is what caused this? One idea is that the bottom of the lithosphere drops off, and we all know the consequences of that from the modelling Greg Houseman did 15 years ago. This dropping, delaminating, 'deblobbing', or whatever you want to call it, can initially be a very slow process but when it actually gets underway it seems it can proceed quite catastrophically. Consequently, you can get rapid regional uplift leading to a major change in base level at the margins. 'Uplift and erosion' is a very dangerous phrase which many people use almost as one word; there is no particular reason why an uplifted region will begin to erode rapidly. However, if the continental interior's drainage is connected to the margins, you can imagine that this will look like quite rapid erosion with consequences for sediment delivery. In addition, as the cool lower lithosphere is replaced by hot asthenosphere, this can lead to melting of the more fusible parts of the remaining lithosphere, thereby producing the small quantities of melt we see emplaced around this time. Thus, the timing of kimberlite intrusion, the inference of a change in the lithosphere geotherm around 90 Ma ago and the denudation and sedimentation chronologies can be interpreted as reflecting a causal link related to a change in the lithospheric thickness.

P. Bishop (*University of Glasgow, UK*). I want to ask a general question related to these leads and lags. One of the things that we geomorphologists are often very concerned about at times is the very strong, but overly simplistic, temporal link that is made by some between uplift, denudation and sediment supply. From Dr Gallagher's data, does he have any sense of the sorts of lags between pulses of denudation onshore, as inferred from fission track analysis, and the corresponding offshore pulse of sedimentation? In other words, could you give us some sense of the lags between uplift, denudation and corresponding pulses of sedimentation offshore?

K. Gallagher. That is a difficult question to answer with our current database. We have a lot of isopach and seismic data from offshore Namibia and Minoru Aizawa; a PhD student at Imperial College, is working on putting all this together from Cape Town up to Angola. The two real problems he has had are, first, a lot of the syn-rift and early sedimentation sequences are in discrete fault-bounded regions, and consequently it is very hard to trace reflectors across the sub-basins. So even if we know the age of a reflector it is hard to correlate it through these faulted regions. The second problem in Namibia is that there are actually not many wells, and in addition the Tertiary sequence is mostly continental, so the sequence is difficult to date. We just do not know the age of a lot of these sediments well enough to be able to relate them to the discrete short-lived erosional spikes. However, we can say grossly for western southern Africa that there may have been a rift-related erosional event, and then a couple of 'bumps' in the Late Cretaceous. Broadly, these events correlate with the volume of sediment which we see offshore Namibia because we see more sediment in the Late Cretaceous than we do in the Early Cretaceous. So at that level the independently derived erosion and sedimentation chronologies do correlate.

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