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Mechanisms of extensional basin formation and vertical motions at rift flanks: Constraints from tectonic modelling and fission-track thermochronology

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Abstract

We present reconstructions of vertical motions at the Saudi Arabian Red Sea margin and the Transantarctic Mountains in order to test the validity of models of rift flank uplift. The eroded rift flanks were flexurally backstacked using apatite fission-track thermochronology to determine the amount of erosion. Basin sediments were simultaneously flexurally backstripped, constrained by seismic reflection profiles. Flexural rigidities were estimated from published coherence studies of gravity and topography. When adopting these lithospheric strength values it appears that significant tectonic uplift (of the order of 3–5 km) has taken place in the present-day coastal plain areas. Forward thermomechanical modelling of rift flank uplift indicates that thermal mechanisms fail to explain the observed tectonic uplift/subsidence pattern and crustal structure; flexural uplift as a result of lithospheric necking appears to be a much more plausible mechanism. Best-fit models consistently predict mid-crustal kinematic necking levels, with depths between 10 and 30 km. Whereas uplift of the Saudi Arabian Red Sea margin can be modelled adopting a pure-shear necking model, the large uplift of the Transantarctic Mountains requires a simple shear mechanism with lithospheric stretching concentrated under the uplifted mountain range. A comparison of these model results with rheological strength profiles and dynamic models for extension of continental lithosphere suggests that the kinematics of extension are determined to a large extent by the rheological coupling of strong upper/middle crust and weak lower crust.

1. Introduction

Rift flank uplifts are commonly observed features that border many extensional basins and rifted margins [1] and they are an important element to be incorporated in tectonic models of rift basin formation. Several classes of models have been proposed to explain their origin.

Early models proposed that rift flank uplifts were supported by thermal processes and involved mechanisms by which extra heat can be put into the flank regions, for instance by two-layer depth-dependent stretching of the lithosphere [2,3]. An alternative thermal model suggests that small-scale convection in the sublithospheric mantle is induced by extension, producing excess lithospheric thinning by thermal erosion [4]. These models generally assume that local isostasy prevails during rifting, since this is neces-

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sary to produce sufficient uplift of rift flanks employing relatively small thermal forces [5].

More recently, however, a number of observations have raised doubts on the concept of thermal support of rift flank uplifts. For instance, strongly negative Bouguer gravity anomalies under rift flanks, expected if large amounts of hot, low-density material are present, are not generally observed. Also, the existence of 'old' (Mesozoic) rift margin uplifts suggests that they can be supported for periods longer than the thermal time constant of the lithosphere (~ 60 m.y.). This has led some authors to propose that the lithosphere retains finite strength during rifting and that rift flanks are mechanically supported by upward flexure acting on a basin during rifting [6–8].

A third class of models explains uplift adjacent to subsiding basins by large-scale simple shear, either along a shear zone cutting across the entire lithosphere [9] or along intracrustal detachments accommodating shear strain between the upper crust and mantle, with an offset between the loci of crustal and subcrustal extension [10,11]. In these models both thermal and mechanical processes may interact to produce uplifted rift flanks.

Finally, a model which was proposed to explain permanent uplift of regions bordering volcanically active rifted margin involves crustal thickening by underplating due to partial melting of the underlying mantle [12]. Large-radius domes may also develop during the evolution of a rift, for which dynamic compensation by hot and partially molten upwelling asthenosphere has been proposed [13]. These mechanisms should, however, have clear seismic and gravity expressions. Whereas crustal underplating can be inferred from seismic data for volcanic rifts and the oceanward side of volcanic rifted margins [e.g., 12], no clear evidence exists for underplating of the uplifted shoulders adjacent to rifted margins. Furthermore, these mechanisms will tend to produce more regional (~ 1000 km wavelength) plateau uplift associated with magmatic activity as opposed to rift flank uplift with a wavelength of around 200 km.

In this paper, we explore the roles of these different mechanisms in creating rifted margin

uplifts and derive a method to compare predicted uplift values with observations. While the vertical motions within rift basins have been routinely studied for more than a decade using backstripping techniques [e.g., 14] the absence of a stratigraphic record on rift flanks has long hampered comparison of model results with data. Recently, Brown [15] proposed a method to quantify tectonic uplift using apatite fission-track data. Here we extend this technique to the two-dimensional case, including a flexural response of the lithosphere. The results are compared to the predictions of a thermomechanical forward model [8] with the purpose of determining the mechanisms causing rift flank uplift and the kinematics of extension. This approach is applied to two areas (the Saudi Arabian Red Sea margin and the Transantarctic Mountains) for which input data for both two-dimensional backstripping (i.e., seismic reflection profiles) and backstacking (fission-track data) exist. This work extends previous models for the uplift of these rift flanks [16,17] by explicitly taking into account the effects of erosion on the derived tectonic uplift and by using a forward model to explain quantitatively the observed uplift patterns.

Whereas dynamic models [e.g., 6,18] can provide an answer to the question of what is physically possible, kinematic studies are more appropriate to address what is geologically reasonable. Since it is our objective to model actual geological examples we choose to employ a kinematic rather than a dynamic model. We subsequently interpret the results of our kinematic approach by a comparison with rheological models and predictions of dynamic models of extension.

2. Vertical motions at rift flanks

The various models for rift flank uplift should have different effects on the evolution of the rift zone, most noticeably on the spatial and temporal distribution of subsidence and uplift, present-day topography, gravity and heat flow [e.g., 19]. In this section we explore the consequences for vertical motions at rift flanks and crustal structure, using a thermomechanical model developed by

Kooi [20, see also 8]. By varying the stretching distributions and the mechanism of isostatic compensation (incorporating either local isostasy or a

lithosphere which retains finite strength during extension, with the equivalent elastic thickness T_e defined by the depth to a specific isotherm) we

Fig. 1. Predictions of forward models of rift flank uplift for vertical motions at rifted margins and crustal structure. (A) Thermal uplift of rift flanks adopting local isostasy and different isotherm-defined flexural rigidities. (B) Pure-shear mechanical (flexural) uplift as a result of lithospheric necking for different necking depths. (C) Simple-shear detachment model for local isostasy (thermal uplift only) and a lithosphere that maintains finite strength (thermal and mechanical uplift) for different detachment depths. The tectonic uplift/subsidence and crustal structure at the end of rifting are shown. β factors are similar for all models.

quantify the various conceptual models for rift flank uplift. This should be considered as a parameter study to obtain an insight into the effects of different assumptions. The dynamic implications of these assumptions will be discussed later.

Fig. 1 shows predicted tectonic uplift/subsidence patterns and crustal structure for a model of thermal uplift, mechanical uplift (induced by necking) and a detachment model. Crustal extension factors (β) are the same in all model runs. Thermal uplift is modelled by two-layer extension with lithospheric stretching (δ) 5 times as large as β and extending twice as far inland as it, in order to maximize the resulting uplift. Dynamic models of sublithospheric small-scale convection suggest that it can thin the lithosphere by a factor 2 to 3 more than the crust and widens the zone of lithospheric extension with respect to that of crustal extension [21].

The kinematics of extension in the mechanical uplift model are controlled by a kinematic reference level or 'depth of necking' z_c ; upon extension with a factor β a kinematic surface depression S is produced given by [7]:

$$S = \left(1 - \frac{1}{\beta}\right) z_c \quad (1)$$

Subsequently, all loads (i.e., both thermal and isostatic) acting on the extended lithosphere are integrated and regionally compensated [cf. 5]. When z_c is sufficiently deep a surface depression which is deeper than locally compensated basin depth is created and a regionally upward state of flexure develops, creating rift flank uplift.

Finally, a detachment model is defined by an offset between δ and β , with δ concentrated below the upper plate margin [cf. 22]. In this model z_c is interpreted as the depth to the de-

Fig. 2. Dependence of model results on adopted parameter values. Upper panels show maximum tectonic uplift for narrow margins (50 km transition from unextended continental to oceanic lithosphere) and lower panels show the same for wide margins (200 km transition zone). 'Young' lithosphere denotes a model with 30 km crustal thickness and 125 km lithospheric thickness, 'old' lithosphere for 40 km crustal thickness and 200 km lithospheric thickness. 'Short' rifting time is 5 m.y. from onset of extension to crustal separation, 'long' rifting time is 50 m.y. Complete uplift/subsidence patterns and crustal structure for the model with old lithosphere and short rifting time are shown in Fig. 1.

tachment [7,11]. In the case of local isostasy, uplift will be thermally supported [e.g., 19,22]. If the lithosphere retains strength the uplift will be supported flexurally as well as thermally with the two mechanisms reinforcing each other at the upper plate margin and counteracting at the lower plate margin.

Fig. 2 shows the dependence of model results on (1) width of the rifted margin (a narrow margin where the transition from unextended continental crust to oceanic crust takes place within 50 km, and a wide one where this transition zone is 200 km wide), (2) thermotectonic ‘age’ of the lithosphere (a ‘young’ continental setting with crustal thickness of 30 km and a lithospheric thickness of 125 km vs. an ‘old’ or ‘cratonic’ setting with 40 km crust and 200 km lithosphere) and the (3) rifting velocity (5 m.y. rifting time vs. 50 m.y. rifting time).

The results indicate that flexural uplift of rift flanks is much more effective than thermal uplift. Maximum thermal uplift reaches 2.5 km for local isostasy, a thick lithosphere and short rifting times, but also in this extreme case it is < 1.5 km when a strong lithosphere is employed. If mantle thinning is spread out over a much wider area than crustal thinning (as in the ‘wide margin’ case) the difference between the local and regional isostasy models becomes much smaller. Such a widespread zone of excessive lithospheric thinning cannot be explained by convection models and would require a regional thermal ‘dome’ and/or widespread underplating.

Mechanical uplift can reach more than 6 km for relatively deep levels of z_c at narrow margins and even more at upper plate margins where it is aided by thermal uplift. The strongest control on the amount of mechanical uplift is exerted by the depth of z_c . Both lithospheric ‘age’ (thickness) and rifting duration have a more significant effect on thermal uplift patterns (lithospheric thickness has a positive effect, the duration of rifting a negative one) than on mechanical uplift. Finally, the results suggest that the width of the rifted margin exerts a major control on the predicted amount of flank uplift, with maximum uplift taking place at narrow margins. For wide margins the difference in uplift patterns and crustal struc-

Fig. 3. Variation of fission-track age and mean track length with temperature, calculated using a quantitative annealing model [29], for a 100 m.y. stable geotherm. A depth scale is added assuming a $30^{\circ}\text{C}/\text{km}$ geotherm. Both fission-track ages and mean lengths decrease rapidly between 80 and 120°C in the partial annealing zone (PAZ).

ture between the various models is much smaller than for narrow margins.

3. Quantitative estimates of rift flank uplift from fission-track data

The annealing of apatite fission-tracks at temperatures below 120°C over geologically relevant timescales (10^6 – 10^8 y) makes fission-track thermochronology suitable for studying cooling and exhumation in the upper few kilometres of the continental crust. Under tectonically stable conditions, apatite fission-track ages and mean track lengths will decrease systematically with temperature (and hence depth), the main control on the pattern being the geotherm (Fig. 3). Secondary controls will be exerted by the duration of tectonically stable conditions and the chemical composition of the apatites [23]. At temperatures between 80 and 120°C , in the partial annealing zone [24], the amount of annealing increases rapidly towards total. Samples exhumed from different

depths (paleotemperatures) will, therefore, have different fission-track ages and mean lengths. At rifted continental margins characterised by an elevated escarpment, fission-track ages generally increase from the coastal plain (where they are often lower than the age of rifting) to the uplands, inland of the escarpment, reflecting differential erosion [25,26] (Fig. 4a). Consideration of the length distribution of confined tracks [27] allows reconstruction of the cooling history. A plot of fission-track age against mean track length of an area that has undergone erosion and cooling will show a characteristic boomerang shape, from which the amount and timing of exhumation can be inferred [28]; the low-age peak in the plot dates the exhumation of samples from paleotemperatures $\geq 120^\circ\text{C}$ (Fig. 4b).

It should be noted that apatite fission-track data record cooling of a rock sample which can be translated into exhumation (or erosion) by adopting a reasonable geotherm model. Recently,

a number of authors have drawn attention to the various components which constitute exhumation [e.g., 15,30,31]. Brown [15] in particular suggested a method for resolving between tectonic uplift and isostatic rebound as a response to erosion. By analogy with the backstripping technique in sedimentary basins [14], where:

$$u_T = w + s \frac{(\rho_s - \rho_m)}{(\rho_w - \rho_m)} \quad (2)$$

with u_T = tectonic subsidence, w = present-day water depth, s = sediment thickness and ρ_w , ρ_s and ρ_m = the densities of seawater, sediments and sublithospheric mantle respectively, the back-stacked tectonic uplift of an eroded area is:

$$u_T = (H_0 + \Delta E) - \frac{\Delta E \rho_c}{\rho_m} - H_i \quad (3)$$

with H_0 and H_i = present-day and initial elevation respectively, ΔE = the amount of erosion/exhumation as derived, for instance, from apatite

Fig. 4. (a) Pattern of fission-track ages (■) and mean track lengths (○) expected at uplifted rifted margins where coastal plain samples were exhumed from greatest depths (paleotemperatures). Inset shows exhumation history adopted (linear exhumation of 100 Ma old stable block from 30 Ma onward). Fission-track parameters were calculated using the model of Green et al. [29]. (b) Resulting age-length plot; shading = mean length $\pm$ standard deviation for samples from different paleotemperatures (scale on top). Insets show modelled fission-track length distributions for samples originating from above (70°C), within (100°C) and below (130°C) the fossil PAZ.

fission-track data, and ρ_c and ρ_m = the densities of eroded crustal material and sublithospheric mantle respectively.

4. Flexural strength of the lithosphere during extension

Backstripping and backstacking analyses are extremely sensitive to assumptions about the isostatic compensation of sedimentation or erosion. The estimated amount of isostatic readjustment is inversely proportional to the adopted lithospheric strength and, therefore, the amount of inferred tectonic uplift/subsidence is proportional to it. This is because, in a flexural backstripping or backstacking scheme, the amount of isostatic compensation is 'filtered' by a flexural response function $\Psi_{(k)}$, which depends on the wavelength (wavenumber k) of the load and the flexural rigidity (D) of the lithosphere [32,33]:

$$\Psi_{(k)} = \left[1 + \frac{Dk^4}{\Delta\rho g} \right]^{-1} \quad (4)$$

with $\Delta\rho$ = the density contrast between eroded/deposited material and the buoyant asthenosphere.

Because of the relatively small density contrast between crustal and asthenospheric material, isostatic rebound after erosion will be substantial (approximately $\frac{5}{6}$ the amount of erosion) for local isostatic equilibrium or low flexural rigidities. This has led a number of authors [e.g., 15,34,35] to conclude that nonuniform erosion and isostatic rebound may be the dominant factor in producing the present-day elevation patterns at rifted continental margins, with tectonic uplift distributed much more homogeneously.

To derive a consistent tectonic uplift pattern it is therefore essential to have a reliable estimate of lithospheric strength. Previous models of flexural uplift of the rift flanks studied here found best fits for equivalent elastic thicknesses (T_e) greater than 40 km (corresponding flexural rigidity $D = 5.7 \times 10^{23}$ Nm) for the Saudi Arabian Red Sea margin [16] and up to 110 km ($D = 1.2 \times 10^{25}$ Nm) for the Transantarctic Mountains [17]. The

latter study, however, seems biased towards high values because the observed topographic deflection was modelled as being solely the result of a line load acting at the coastline. Most studies infer that flexural rigidities decrease towards the rift basin; T_e within the Ross Sea basins is only 20–24 km ($D = 7.1\text{--}12.0 \times 10^{22}$ Nm) [17,36].

Other studies [e.g., 37–39] arrived at much lower T_e 's of < 5 km ($D < 2.5 \times 10^{21}$ Nm) for extensional basins and margins. However, these studies may seriously underestimate the strength of the lithosphere because of their kinematic assumptions. The traditional sediment loading model [32] employed in the earlier analyses [37,38] implicitly assumes near-zero strength during the synrift phase [20] whereas the flexural cantilever model [39] directly translates upper crustal deformation (block faulting) into lithospheric strength, while these are probably decoupled [cf. 16].

A better estimate of lithospheric flexural rigidity can be made from coherence analyses of Bouguer gravity anomalies and topography. Several of these studies have been performed in similar tectonic settings (i.e., cratonic areas adjacent to rifts or rifted margins) and they come up with relatively large values: $D \approx 1.0 \times 10^{23}$ Nm ($T_e = 22$ km) for Southwest Australia [33], $D \approx 2.8 \times 10^{23}$ Nm ($T_e = 32$ km) for the Baikal Rift area [40], and D varying from 1.7×10^{23} Nm to 4.4×10^{24} Nm ($T_e = 27\text{--}79$ km) for East Africa [41]. These values are consistent with the results of forward models of flexure in these areas [42,43]. We conclude that significant flexural rigidity is maintained at rifted continental margins, with T_e ranging from around 20 km ($D = 7.1 \times 10^{22}$ Nm) within the basin to $> 40\text{--}50$ km ($D = 5.7\text{--}11.1 \times 10^{23}$ Nm) inland. This is compatible with models in which T_e is tracked by the 400–450°C isotherm. These values are substantially higher than the T_e values of ≤ 15 km ($D < 3.0 \times 10^{22}$ Nm) inferred by models where the topography is solely the result of differential erosion and isostatic rebound [34,35].

Furthermore, in a rifted margin setting, there will be an interplay of upward isostatic compensation as a result of erosion of the rift flank and downward isostatic compensation of the sedimentary infill of the basin, which will also influence

the derived tectonic uplift and subsidence patterns. In order to arrive at a consistent tectonic uplift/subsidence section across a rifted basin and its margins using a flexural model it is thus necessary to simultaneously backstrip the basin sediments and backstack the eroded rift flank. We have done this for the Saudi Arabian Red Sea margin and the Transantarctic Mountains–Ross Sea area using a two-dimensional finite difference flexure algorithm. While, in principle, the temporal evolution of tectonic subsidence and uplift can be calculated if well data and fission-track length distributions are available, we will restrict ourselves here to the total subsidence and uplift since rifting. At this stage, we are not aiming to predict basin stratigraphy or the morphotectonic evolution of the margin.

5. Backstacking rifted margin uplifts

5.1. Saudi Arabian Red Sea margin

Apatite fission-track data for the Saudi Arabian Red Sea (SARS) margin are shown in Fig. 5a. Fission-track ages (FTA) on the SARS margin [28] increase from < 20 Ma on the coastal plain up to > 400 Ma in the continental interior, indicating that several kilometres of erosion affected the coastal plain area after rifting (at 30–25 Ma, [44]). We have converted FTA to pre-erosion paleotemperatures assuming linear exhumation of a 400 Ma old crustal block from 30 Ma onward. Samples that have experienced temperatures of $> 90^\circ\text{C}$ (FTA < 152 Ma) are confined to the western side of the escarpment, and samples

Fig. 5. Apatite fission-track data from the Saudi Arabian Red Sea margin and the Transantarctic Mountains. (a) Pattern of fission-track ages (●, bars represent standard errors) and mean track lengths (△) against distance from the coastline for the Saudi Arabian Red Sea margin (data from [26]). (b) Samples plotted at their true elevations, labelled according to inferred paleo-(pre-exhumation) temperature adopting idealized exhumation model (linear exhumation of 400 Ma old crust since 30 Ma). ● = samples exhumed from temperatures $> 120^\circ\text{C}$; □ = 90 – 120°C ; △ = 60 – 90°C ; ▲ = $< 60^\circ\text{C}$; shading = indicates present-day topography; dashed lines labelled 90 and 120 are paleoisotherms. (c) and (d) Same as (a) and (b) but for the Transantarctic Mountains (data from [46,47]). Here exhumation model is linear exhumation of 150 Ma old crust since 60 Ma.

that were at temperatures of $> 120^{\circ}\text{C}$ (FTA < 22 Ma) are found only in the coastal plain area, at elevations of < 1 km. The pattern of paleoisotherms indicates that the geothermal gradient prior to rifting was around $30^{\circ}\text{C}/\text{km}$ and that maximum erosion of around 4 km (assuming a surface temperature of 30°C) has occurred in the coastal plain area. Paleoisotherms dip inland, indicating that maximum erosion took place at the present-day shoreline. The amount of erosion inland of the escarpment is more difficult to estimate; extrapolation of the paleoisotherms would suggest significant erosion inland of the escarpment but the fact that FTA do not seem to correlate with elevation inland of the escarpment argues against this.

The amount of sediment deposited in the Red Sea basin was determined from seismic reflection profiles [45]. We have backstacked/stripped the amounts of eroded and deposited material using different models for the strength of the litho-

sphere (Fig. 6a). Results are shown for a local isostatic model and constant flexural rigidity models with T_c 's of 15 and 60 km ($D = 3.0 \times 10^{22} - 1.9 \times 10^{24}$ Nm) encompassing the minimum and maximum strength values we expect at rifted margins. A local isostatic model suggests that tectonic subsidence decreases from the present-day elevation maximum towards the coastline, as a result of the large isostatic rebound component. Such a pattern is, however, inconsistent with geomorphological studies which suggest that the present-day position of the escarpment is mainly controlled by erosional processes, i.e. escarpment retreat [1].

In contrast to local isostatic backstacking, all models incorporating flexural strength of the lithosphere predict significant tectonic uplift of the coastal plain, as a result of filtering of the isostatic response and interaction with downward compensation of the sediments in the basin. The maximum tectonic uplift of the SARS margin is

Fig. 6. Results of backstripping/backstacking for the Saudi Arabian Red Sea margin and the Transantarctic Mountains–Victoria Land Basin for different models of lithospheric strength. Lower parts show sediment fill of the basin determined from seismic data (from [45] for the Red Sea and [48] for the Ross Sea) and estimated amount of erosion on the flanks. Mesozoic sediments in the Ross Sea basin were not backstripped, in contrast to the continental ice sheet. Upper parts show reconstructed tectonic uplift/subsidence patterns for different isostatic compensation models. For the Saudi Arabian Red Sea margin, a model with erosion up to 200 km inland is also shown (thin line).

calculated to be 3.5 km for $T_c = 60$ km and 2.7 km for $T_c = 15$ km. The difference between the minimum and maximum reasonable strength estimates is mainly reflected in the flexural wavelength of the tectonic uplift pattern. Assuming that there has been erosion up to 200 km inland of the escarpment also affects the wavelength of the tectonic uplift pattern but only slightly suppresses the maximum tectonic uplift (to 2.6–3.3 km), which is located just inland of the coastline.

5.2. Transantarctic Mountains

Contrary to the SARS margin, the youngest FTA [46,47] in the Transantarctic Mountains

(TAM) are not encountered at the coastline but somewhat inland (Fig. 5c). Samples which have experienced paleotemperatures of $> 120^\circ\text{C}$ (FTA < 55 m.y., adopting an onset of exhumation at 60 Ma) have only been encountered 20 km inland, at elevations of < 900 m. Paleoisotherms reach an elevation maximum here and dip both towards the coast (as a result of normal faulting) and towards the hinterland (Fig. 5d). Maximum erosion is estimated to be around 4.5 km, again using a geothermal gradient of 30°C but a surface temperature of 0°C . No fission-track data are available for areas more than 50 km inland. We have assumed that no erosion has taken place landward of the present-day elevation maximum of around 3 km, some 80 km inland.

Fig. 7. Results of forward modelling of vertical motions and crustal structure for the Saudi Arabian Red Sea margin, for a thermal uplift model assuming local isostasy and different mechanical (flexural) uplift models (see legend in (B) and (C)). (A) Vertical motions. Shading = inferred range of tectonic uplift/subsidence from backstripping/stacking. (B) Crustal structure without erosion, and (C) including erosion of the rift flank. Shaded crustal structure is as observed from seismic and gravity data [49]. (D) Calculated free air gravity anomaly patterns for the different models, including erosion of the coastal plain. Anomalies resulting from both crustal (basement and Moho depth) and thermal structure are calculated [cf. 55]; the compensation depth lies at the lithosphere/asthenosphere boundary. Adopted densities (at 0°C) are 2400 kg m^{-3} for sediments, 2800 kg m^{-3} for continental crust and 3300 kg m^{-3} for mantle material. Sediment compaction is not corrected for, which may underestimate the gravity anomalies in the basins. Observed gravity anomalies are from the *Bureau Gravimétrique International* database.

The Tertiary Victoria Land Basin in the Ross Sea reaches a depth of 7 km and is underlain by another 8.5 km of presumably Mesozoic sediments [48]. Seeing as we have fission-track control on the Cenozoic exhumation of the Transantarctic Mountains only, we will not consider older vertical motions. Another load which should be taken into account is the ice sheet which covers the Antarctic continental interior and reaches a thickness of up to 3 km. This load acts to suppress the inland topography to below sea level.

Backstripping reveals that maximum tectonic uplift in the TAM reaches 5.8 km for $T_c = 60$ km, or 3.7 km for $T_c = 15$ km, and occurs at the same location as maximum erosion, i.e. 20 km inland (Fig. 6b). The upward directed effect of backstripping the basin sediments is so large that, for large T_c values, the backstacked rift flank uplift overlaps the restored topography.

6. Forward modelling of rift flank uplift

Now that we have established the patterns of tectonic uplift and subsidence for the two basin/flank systems, we can compare these to forward modelling predictions. We employ the same model as in section 2; model parameters include a 40 km thick crust and 200 km lithosphere prior to extension. These parameters are chosen to model a cratonic prerift tectonic setting. The strength of the lithosphere (T_c) is determined by the depth to the 400°C isotherm, corresponding to $T_c = 60$ km ($D = 1.9 \times 10^{24}$ Nm) in unextended lithosphere. We adopt this T_c as representing a good estimate from both coherence analyses and dynamic models of lithospheric extension. Stretching distributions and the kinematics of extension (i.e., depth of the kinematic reference level z_c) are taken as free parameters. β values are chosen so as to provide a fit to the tectonic subsidence of the basin for the chosen kinematic model; the model predicts tectonic uplift of the rift flank and crustal structure.

6.1. Saudi Arabian Red Sea margin

For the SARS margin we adopt a rifting period from 30 to 25 Ma [44]. From 25 Ma onwards, the locus of extension is concentrated at the axis of the Red Sea and proceeds towards ocean spreading (modelled by taking $\beta = 8$ and $\delta = 50$). The occurrence of Oligocene shallow-marine sediments [26] indicates that the area was at sea level prior to the onset of rifting. The present-day elevation of the Arabian continental interior is around 1 km; the rift flank topography is superimposed on this regional elevation. The regional uplift of the Arabian peninsula therefore seems to be the result of a much larger scale dynamic process. This is implemented in the forward model by a general uplift from 30 Ma onward.

A model of thermal uplift of the rift flanks is capable of explaining the present-day topography of the SARS flanks when local isostasy is adopted (Fig. 7). However, in this case the large isostatic rebound component related to erosion of the coastal plain area predicts a degree of crustal attenuation below the coastal plain not observed in the data; the SARS margin is characterised by a steep Moho step under the edge of the basin [49]. This is consistent with gravity modelling; the local isostatic model produces free air anomalies over the coastal plain that are far too high.

The crust below the coastal plain could have been thickened by magmatic underplating [e.g., 44] although there seems to be no evidence for this from the seismic refraction data [49]. In any case, the presence or absence of a crustal underplate does not affect the requirement of crustal attenuation from the present-day escarpment towards the coast in order to explain the topography and erosion patterns, if local isostasy is assumed.

The pattern of residual isostatic anomalies over the Red Sea margin, which is strongly negative (i.e., local isostatic anomalies are much higher than those observed) seaward, suggests that this margin is in an upward state of flexure [cf. 8] and that mechanical uplift of the rift flank plays a major role. A mechanical pure-shear (necking) model for rift flank uplift is able to reproduce both the tectonic uplift/subsidence patterns and

the crustal structure for the SARS margin (Fig. 7). Best-fit z_c levels lie in the middle crust (10–15 km).

6.2. Transantarctic Mountains

Uplift of the TAM commenced at around 60 Ma, coeval with renewed rifting in the Ross Sea [50]. Several arguments (e.g., the geomorphology of the mountain range, stratigraphic patterns within the sediments of the Ross Sea basins and the age of volcanics within the rift) suggest that rifting activity has been ongoing since then and is still active today [cf. 47,48,50]. We have therefore employed continuous extension since 60 Ma onwards in our forward model. The Victoria Land Basin in the Ross Sea is underlain by a basin of presumably Mesozoic age reaching a depth of 15.5 km [48], indicating that a previous phase of extension affected the area. We included this in the model by first removing the Cenozoic sedi-

ments from the basin and modelling the Mesozoic extension. The resulting crustal and thermal structure at 60 Ma is then the starting point for modelling of the Cenozoic extension. The load of the continental ice sheet covering Antarctica suppresses the present-day elevation to below sea level. When this load is backstripped the hinterland of the TAM has an elevation of around 500 m, similar to the pre-uplift elevation of the mountain range previously inferred [46,47]. We take this as the elevation of the area prior to rifting.

As was the case for the SARS margin it is, in principle, possible to explain the uplift of the TAM by local isostatic compensation in response to thermal processes (Fig. 8). However, the excessive mantle thinning ($\delta > 7$) that has to be employed to model uplift of the TAM leads to strongly negative free air anomalies, which are not observed. The model is also incapable of explaining the existence of the very deep (7 km) Victoria Land Basin; extreme stretching values

Fig. 8. Same as Fig. 7 for the Transantarctic Mountains–Ross Sea. Detachment model denotes model with z_c at 30 km and asymmetric stretching distribution, with δ concentrated beneath the Transantarctic Mountains uplift and β beneath the Ross Sea basins. Crustal structure is from [48,50], gravity data are compiled from [17,50].

($\beta > 10$) lead to crustal thinning greatly in excess of that observed but still too little basin subsidence.

Again, strongly negative residual isostatic anomalies over the Ross Sea basins suggest that the whole system is in an upward state of flexure and that mechanical uplift is important. Here, however, a pure-shear necking model is not very successful. Very deep necking levels (> 40 km) are required to explain the large uplift of the TAM, but these are incompatible with the crustal structure and gravity pattern. Deep levels of necking will produce a very smooth Moho topography and large negative gravity anomalies in the basin, which are not observed. A best-fit model (Fig. 8) adopts a lower crustal z_c level (30 km) with asymmetric extension (detachment model), where lithospheric stretching δ is concentrated under the TAM and crustal stretching β in the Ross Sea basins. This reproduces the observed tectonic uplift/subsidence patterns while maintaining a crustal structure that is compatible with observations. A simple-shear scenario for uplift of the TAM has been proposed previously and is consistent with petrological data that suggest the existence of anomalously hot upper mantle under the TAM [47,51]. The quantitative modelling performed here indicates that the TAM uplift is indeed partly thermally supported; the bulk of the uplift appears to be, however, a result of upward flexure.

7. Discussion and conclusions

Our backstacking results indicate that erosion of rifted margin uplifts is an important process in modifying topography as well as the geometry of the landward part of rifted basins. Therefore, backstacking rift flank erosion, constrained by apatite fission-track data, is essential in order to reconstruct the vertical motions that have affected the rift flank. For reasonable assumptions of lithospheric strength, the present-day topography combined with the amount of erosion that has taken place seaward of the present-day escarpment indicates significant rift-related tectonic uplift of the coastal plain. Maximum tec-

tonic uplift values vary from ~ 3 km for the Saudi Arabian Red Sea margin to > 5 km for the Transantarctic Mountains.

Thermal mechanisms for uplift of rift flanks have recently been questioned using arguments such as the long-lasting nature of rift flank uplifts and the observed gravity anomaly pattern over rifted margins [cf. 7,8]. Thermal mechanisms are only effective when combined with zero flexural rigidity (local isostasy) models for lithospheric rheology. Coherence studies of gravity and topography, in contrast, indicate that the lithosphere retains significant flexural strength adjacent to continental rifts and rifted margins. We have shown here that employing a local isostatic model in backstacking rift flank erosion leads to tectonic uplift patterns which are not consistent with current views on the morphotectonic evolution of uplifted rifted margins. Forward models employing thermal uplift and local isostasy result in crustal structures and gravity anomaly patterns that deviate strongly from those observed when erosion of the rift flank is taken into account.

Therefore, a mechanical (flexural) support of rift flanks seems much more plausible. An upward state of flexure at rifted margins, which is required to support the flanks, is most logically explained as the consequence of stretching around an intracrustal kinematic level (z_c) which would stay horizontal in the absence of isostatic compensation. Such kinematics are predicted by dynamic models of lithospheric stretching but the exact physical significance of this level is not clear at the moment. Pure-shear models of lithospheric extension [e.g., 6,18,52] predict that necking will occur around the strongest layer within the lithosphere, which, depending on the tectonic setting, will be either the upper crust or the upper mantle. Models that include offset pre-existing weakness zones in the upper crust and upper mantle predict the development of a subhorizontal detachment in the weak lower crust and in this case the level of necking varies over the model, from a deep level under the hangingwall (upper plate) of extension to a more shallow one under the footwall or lower plate [53]. By comparing the best-fit z_c level derived from kinematic modelling studies of well-constrained tectonic settings such as the

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Fig. 9. Comparison of depth range of the level of necking inferred from kinematic models with depth-dependent rheological models of the lithosphere [54]. Rheological models employ a quartzitic upper crust, dioritic lower crust and an olivine mantle rheology. (a) 'Cratonic' or 'old' lithosphere (upper crustal thickness 20 km, lower crustal thickness 20 km, 200 km lithosphere), and (b) 'young' lithosphere (upper crustal thickness 15 km, lower crustal thickness 15 km, 125 km lithosphere). Vertical bars denote best-fit levels of necking for (1) Saudi Arabian Red Sea margin, (2) Transantarctic Mountains [this study], (3) East African rifts [43], (4) Northeast Brazilian margin [57], (5) Gulf of Lions [8], (6) Valencia Trough [55], (7) Barents Sea [56], (8) Tyrrhenian Sea [Spadini and Cloetingh, in prep.] and (9) Pannonian basin [Van Balen et al., in prep.]. (c) Cartoon showing possible relationship between mid-crustal necking and lower crustal detachment during evolution of rifted continental margin.

Transantarctic Mountains and the Saudi Arabian Red Sea margin to rheological models of the lithosphere it should be possible to evaluate the dynamic significance of this level (Fig. 9).

We have shown here that mid- to lower crustal z_c levels (10–15 km for the Saudi Arabian Red Sea margin, 30 km for the Transantarctic Mountains) are required to produce the observed amount of tectonic uplift and arrive at a realistic crustal structure. Deep (subcrustal) levels of necking that were proposed as the cause for large uplift of rift flanks (e.g., the Transantarctic Mountains [52]) are incompatible with gravity data and Moho topography. Modelling studies of rifted basins developed in ‘younger’ tectonic settings (margins influenced by Phanerozoic prerift thermotectonic events), using the same thermomechanical model, also consistently arrive at mid-crustal z_c levels of the order of 17–25 km [e.g., 9,55,56,unpublished data] (Fig. 9). A kinematically similar model [7] employed for the East African rifts [43] and the Northeast Brazilian margin [57] adopts a level at the base of the crust but the sensitivity of model predictions to this depth was not reported. Apparently, a mid- to lower crustal necking level is controlling the kinematics of extension in most of these different tectonic settings. This is also the depth to which most of the major normal faults bounding rifts can be traced on deep seismic reflection records [e.g., 58,59]. The tectonic significance of this level may, therefore, be that it represents the depth at which concentrated (brittle) deformation in the upper crust gives way to more distributed ductile deformation in the lower crust. Apparently, the continental lithosphere preferentially necks around the strongest crustal level near the brittle–ductile transition; the system is, therefore, probably detached from the strong upper mantle. This interpretation is in accord with dynamic models of continental extension which include a mechanism that weakens the subcrustal mantle [18,53]. In this case, intracrustal necking seems to be the preferred mechanism. Models which suggest subcrustal necking [e.g., 52] may overestimate lithospheric strength, as also suggested by the general absence of subcrustal seismicity in areas where rheological models predict signifi-

cant lithospheric strength [60]. An interesting discrepancy, however, exists between the vertical motions derived from fission-track data and from predictions of kinematic models on the one hand, and the motions inferred by dynamic models on the other. The latter predict maximum tectonic uplift of the flanks of the order of 2 km [18,52], which is a factor of 2 lower than the values derived here.

The need to adopt a lateral offset between crustal extension β and mantle extension δ in a number of kinematic modelling studies may suggest that intracrustal detachments were operating during extension [e.g., 57]. We have shown that for the Transantarctic Mountains–Ross Sea area mantle thinning should be concentrated under the rift flank in order to explain the uplift/subsidence patterns while maintaining a reasonable crustal structure. Mid-crustal necking levels may thus be rheologically coupled to lower crustal detachments [16,57] (Fig. 9c), suggesting that lower crustal weak zones exercise strong rheological control over the kinematics of extension.

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