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Small-scale convection induced by passive rifting: the cause for uplift of rift shoulders

W. Roger Buck *

Department of Earth Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139 (U.S.A.)

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Areas adjacent to rifts, or rift shoulders, are often observed to be uplifted as much as a kilometer or more. In some of these regions geologic data indicate a passive origin for the rifting itself (i.e. there was no anomalous heating of the regions before rifting). Purely conductive heat transport between the rift, where the lithosphere has been thinned, and the rift flanks cannot account for the magnitude of the uplift. Small-scale convection will be induced in the mantle beneath a rift due to the lateral temperature gradients there. Numerical experiments show that convection increases the amount of heat transported vertically into the rift and laterally out of it. In these calculations, the viscosity is taken to be dependent on temperature and pressure and, in some cases, stress. The mantle flow results in thinning of the adjacent lithosphere causing flanking uplift as well as slowing of the subsidence of the middle of the rift. The magnitude of the uplift is dependent on the geometry of the rift and the importance of stress-dependence in the rheology of the mantle. For viscosity parameters which are consistent with the pre-rift temperature structure small-scale convection can produce uplift at least twice as great as would be produced by lateral conduction alone.

1. Introduction

Rifting is the pulling apart of the crust and lithosphere. Sleep [1] showed that conductive cooling of the thinned lithosphere at a continental margin is consistent with the long term subsidence of the Atlantic margin. The possibility that the thinning of the lithosphere is caused by stresses transmitted horizontally, or passively rifted, has been suggested by Salveson [2] and McKenzie [3]. A thermal model based on passive rifting, called the uniform stretching or the extension model has been suggested by McKenzie [3] to explain the subsidence of rifts. Analysis of data from intracratonal basins [4,5] and deep well holes on the Atlantic margin [6–8] indicates that large corrections need to be made to the uniform extension model in many cases where the data on subsidence is complete to the earliest stages of sedimentation. Data on recent rifts most clearly show the need for modification of the uniform stretching model. For several rifts, it has been shown that uplift of the flanks does not predate rifting and that this uplift extends well beyond the area where rifting has thinned the crust [9–11]. Uplift of the areas adjacent to rifted crust is not predicted by the stretching model. Convection, which is induced by the large horizontal temperature gradients in the mantle where the lithosphere has been thinned by rifting, may explain some of these features, as noted by Buck [12]. In studying how tensional stresses could produce extension of the lithosphere, Keen [13] found that small-scale convection should accompany lithospheric thinning and suggested that it might contribute to uplift of areas flanking rifts.

The purpose of this study is to determine if convection beneath passive rifts can produce the observed uplift of rift flanks. To do this, I assume a rift temperature structure which is based on the stretching model and a mantle viscosity relation and then calculate the time evolution of the flow and the changes in the temperature distribution under the rift. The numerical method used is an extension of standard finite difference methods used to study mantle convection [14,15] which allows for curved flow boundaries, and is de-

* Present address: Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964, U.S.A.

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scribed in Buck [16]. A major difference between this problem and the studies of small-scale convection under a horizontally uniform lithosphere [17–20] is that convective flow will occur regardless of the viscosity parameters assumed, since the flow is driven by the temperature structure of the rift. The rate of flow, and its effect on the cooling of the rift and the uplift of the flanks, depends on the viscosity parameters assumed and on the geometry of the rift. Therefore, the model geometry and the viscosity parameters are varied to show their effect on the results.

I first discuss the justification for a mechanically simple model of rifting caused by tectonic stresses and outline the simple thermal model, called the extensional or stretching model [3], which is based on it. Next, I review the subsidence data for several areas and data on the uplift of rift shoulders which indicates limitations in the model. Then I describe numerical calculations on the effect of mantle flow induced by large horizontal temperature gradients in the mantle produced by passive rifting. Finally, I discuss how the resulting convective flow modifies the thickness of the lithosphere in a rifted region as a function of time and the model parameters.

2. Models of rifting

In these calculations I will assume that rifting is passive. Here passive refers to the role of the asthenosphere in the rifting. Passive rifting is driven by stress transmitted by the mechanically strong lithosphere. Active rifting is produced by upwelling of anomalously hot asthenosphere which thins and causes uplift of the lithosphere. Active rifting results in volcanism and doming preceding rifting while for passive rifting rifts form first and then doming may follow [31]. Areas which are clearly associated in space and time with stresses manifest in continental convergence such as the Rhinegraben [22] and the Baikal Rift [23,24] are thought to be passive examples [25]. The great length of continental margins argues for a passive origin for at least some of the length of these rifted areas, since it is likely that anomalous upwelling of asthenosphere is concentrated in individual hotspots. Fig. 1 shows topographic profiles across several of these rifts and one continental margin.

Active rifting is generally considered to be produced by anomalous heat transport from the asthenosphere to the lithosphere which causes the lithosphere to thin. Spohn and Schubert [26] have shown that if the heat flux out of the asthenosphere were to increase 5 to 10 times over an average value that the continental lithosphere could be thinned to crustal levels in a few tens of millions of years. This would cause isostatic uplift of the region of thinned lithosphere. The tensional stresses generated by this uplift may be of sufficient magnitude to cause rifting [27,28]. The heat in this model would be carried by convection, but the high rate of heat transport requires anomalously high asthenospheric temperature, perhaps due to a mantle plume. This process is, therefore,

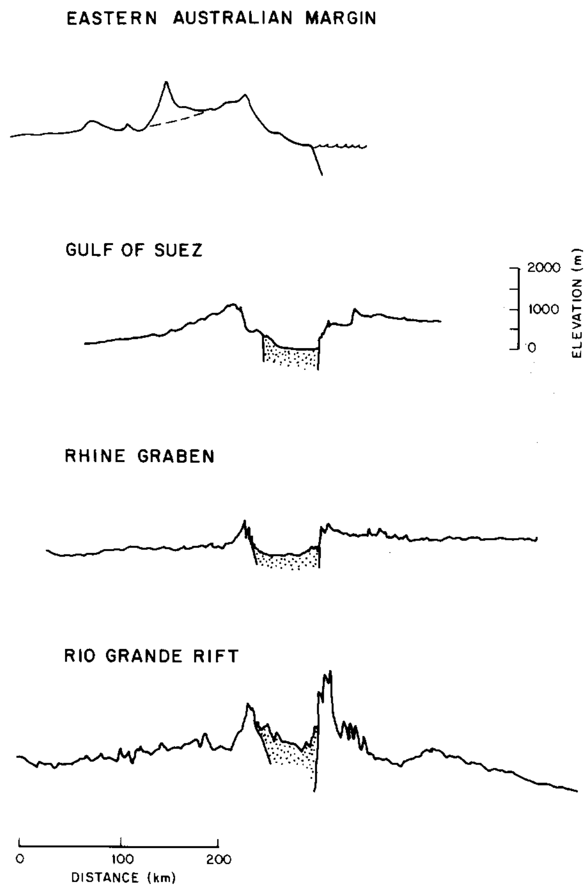

Fig. 1. Topographic profiles across selected rifts and one continental margin where the transition between unstretched continental crust and oceanic crust is narrow. The vertical exaggeration is 40 to 1. Gulf of Suez profile is from Steckler [11] and Australian margin profile is from Weissel and Karner [10]. The peaked topography on the Australian profile is Mt. Kosciusko which may have predated rifting of the margin [10].

fundamentally different than the process being considered here where convection is a response to the large lateral temperature gradients in a rift. Active rifting may be important in areas such as the East African Rift where broad scale doming seems to precede rifting [29] and it is often difficult to unequivocally determine that a given rift is either active or passive. The amount of uplift associated with active rifting depends completely on the assumed amount of anomalous mantle heat transport.

A mechanical model of passive rifting was put forward by Salveson [2]. The thermal consequences of this model have been discussed in terms of a simple thermal model by McKenzie [3]. He considered the instantaneous extension of each vertical column of the lithosphere and crust by equal amounts, and assumed that asthenosphere upwells passively to maintain isostatic equilibrium. When a vertical column of the lithosphere is stretched by a factor β, it then thins to $1/\beta$ times its original thickness. The subsidence has two components. There is an initial component due to the thinning of the crust and a long-term component due to the cooling of the lithosphere back to an assumed equilibrium thickness. The initial subsidence or uplift depends on the original crustal thickness and the amount of stretching. The modeling of subsidence and heat flow are done with constant temperature boundary conditions at 125 km analogous to the plate model for the oceanic lithosphere [30]. The initial thermal structure is derived from the simple movement of temperatures, along with material, vertically up in the lithosphere according to the amount of thinning. The geometry of this is shown in Fig. 2 for the case of equal thinning of the crust and lithosphere. For the simple model of extension, the thermal gradient is assumed to be linear with 0°C at the surface and 1300°C at the base of the lithosphere and the temperature of the asthenosphere, which moves up to replace the thinned lithosphere is 1300°C.

Several mechanisms, beside small-scale convection, may contribute to the elevation of regions flanking passive rifts. Vening Meinesz [31] considered rift basins to be an elastically strong wedge bounded by normal faults. The uplift of the rift shoulders, in his model, is a flexural response to the subsidence of the basin. Bott [27] has modified this model to consider the elastic layer to be the upper crust and that the lower crust deforms by ductile flow. This model allows for crustal thinning under rift basins as is inferred from seismic data [32]. Artemjev and Artyushkov [33] qualitatively considered rifting due to necking of ductile crust. Zuber and Parmentier [34] have shown how the thickness and rheology of a strong viscous or plastic crustal layer control the width of a rift. An uplifted region flanking the rift is a consequence of the necking in this model. The magnitude of uplift for all these models appears to be too small to explain the amount of uplift observed without a major contribution from other processes.

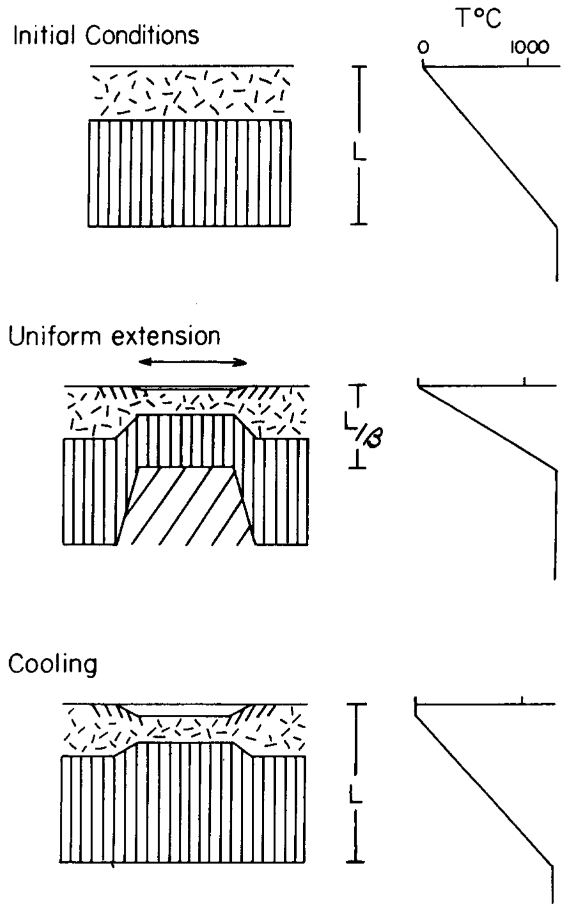

Fig. 2. Illustration of the thermal model of McKenzie [3]. The cross-section shows an area of the lithosphere which initially has a uniform thickness (L). The temperature profile is linear with depth to the base of the lithosphere. The crust (stippled) and lithosphere are instantaneously extended by a factor (β) and so are thinned by a factor ($1/\beta$). The area where the lithosphere was thinned is replaced by isothermal asthenosphere (diagonal lines). Cooling causes thermal subsidence as the temperature profile returns to the original profile.

3. Geologic data on rifts

The subsidence of passive margins estimated using data on sediments from deep wells, as described by Sclater and Christie [4], has been used to test the stretching model. Using gravity and seismic data the thickness of the continental crust beneath sediments is estimated at a well site. The ratio of the average crustal thickness on shore to the well site crustal thickness gives the stretching factor for the crust there. It is then assumed that the rest of the lithosphere was thinned by the same amount. Using this method, the subsidence of rifted continental margins and intracratonic basins has been shown to be consistent with the uniform extensional model [4,6,7,32,35–38]. In the areas where the agreement between the subsidence data and model predictions is good there are few data on the earliest subsidence of the basin.

Subsidence data for the early period of subsidence of rifted areas (i.e. the first 25 m.y.) have been shown to require some modification to the uniform extension model. Royden and Keen [6] claimed that the simple stretching model would not fit the data for wells on the margin of the Labrador Sea. They had to modify it to allow for greater thinning of the mantle lithosphere than for the crust. This conclusion may be questioned since the wells were drilled only on the higher points of tilted crustal blocks. Other workers considering intracratonal rifting [5,9] have found this same need to modify the model to include two layers of stretching, with the amount of subcrustal lithospheric thinning depending on the site.

A related set of data on the uplift of the flanks of rifts do not easily fit into the uniform extensional model. As noted by Morgan [39], a broad regional uplift is usually associated with rifting. The shoulders of the Rhinegraben rift have been uplifted 1000 m since the time of rifting [22]. The Rio Grande rift [40] and the Baikal Rift [24] also show uplifted flanks. The flanks of the Red Sea and the Gulf of Suez rifts show > 1 km uplift which post-dates rifting [11]. One can argue that all these cases are examples of passive rifting, but the uniform extension model cannot fit these data. Thus, small-scale convection, which is a necessary consequence of passive rifting, should be studied as a possible mechanism for producing flanking uplifts.

4. Formulation of rifting calculations

To study the effects of convection on the cooling of a rift, I must first define the initial temperatures in the rift after passive rifting, but before convection and conduction has altered the temperatures. As in McKenzie [3], I consider the instantaneous thinning of an assumed pre-rift, horizontally uniform temperature structure for the lithosphere and constant temperatures in the asthenosphere. Two types of initial temperature structures are considered. First, the temperature resulting from half space cooling for a set length of time was used. This allows comparison of the effect of varying the geometry and the viscosity parameters, but is probably not a very realistic geotherm, since it allows for no mantle heat flux. Next, a set of calculations was done using a linear initial temperature profile, for simplicity, and because this allows for easy comparison with the results of McKenzie [3]. This is a more reasonable way to set up the calculations since the linear temperature gradient implies that the lithosphere is in equilibrium with the heat flux being transported by convection below the lithosphere, as discussed below. The initial rift temperature structure is derived from the profile by stretching by an amount which depends on position within the rift.

The subsidence or uplift of a point in the rift is calculated assuming that pressure variations at the base of the lithosphere and temperature variations within the lithosphere cause an Airy isostatic response. Table 1 gives the physical parameters used to calculate the deformations. Here the vertical displacement of a point is defined as the change in the elevation at a given time from the elevation due to the initial temperature structure. Thus, the

TABLE 1

Parameters used for calculating the model results

Symbol	Name	Value	Units
κ	diffusivity	10^{-6}	m^2/s
α	thermal expansion coefficient	4.0×10^{-5}	1/°C
g	acceleration of gravity	9.8	m/s^2
ρ_m	mantle density	3500	kg/m^3
ρ_w	water density	1000	kg/m^3
K	conductivity	3.2	J/m s °C
c_p	specific heat	900	J/kg °C

initial subsidence due to crustal thinning is not included.

My main interest is in the effect of convection on subsidence and uplift of the surface in the area of a rift. I also compare the vertical subsidence or uplift at each point with an estimate of the surface deformation in the absence of convection driven by the rift temperature structure. This is done by using the same initial temperature structure as for the convective calculation, but considering only conductive heat transport.

The effects of finite amplitude convection are studied by numerically solving the Navier-Stokes equations of mass, momentum and energy conservation in two dimensions [14]. The finite difference method I use is similar to the method discussed in Parmentier [15] but involves curved flow boundaries. This is necessary because the thickness of the lithosphere (the top of the convecting region) in this problem varies by up to a factor of 5 for the initial temperature and viscosity structure. In Buck [16] an approximate method for solving for the flow adjacent to a curved boundary is described.

I consider viscosity to be a function of temperature, pressure and, in some cases, strain rate. For viscosity (μ) that depends only on temperature (T) and pressure (P). I use a relation appropriate for creep in olivine [41] to define the viscosity at a point:

$$\mu(T, P) = A \exp[(E + PV)/RT] \tag{1}$$

where A is a constant which determines the average level of viscosity, E is the activation energy which determines the temperature dependence of viscosity and V is the activation volume which sets the pressure dependence and R is Boltzman's constant. In several cases, I take the viscosity to depend on deviatoric stress, or strain rate ($\dot{e}$), as well as temperature and pressure. I then define viscosity as:

$$\mu(T, P, \dot{e}) = \frac{\mu(T, P)}{1 + A_e[\mu(T, P) \cdot \dot{e}]^{2/3}} \tag{2}$$

where $\mu(T, P)$ is defined by equation (1) and the power 2/3 comes from assuming a power-law rheology [42] with $n = 3$. A_e determines the strength of the strain rate dependence.

5. Models considered

Several parameters of these models are varied in an effort to understand what effects the cooling of a rift. The viscosity is changed in two ways. First, the average viscosity is changed through parameter A in equation (1) and described by a reference viscosity which is the viscosity at 1300°C and at a pressure corresponding to 150 km depth. The temperature dependence of viscosity (E) is also varied. When non-Newtonian viscosity is considered the value of A_e controls the strength of the strain rate dependence. The total width of the initial rift temperature structure (W_r) (see Fig. 3) is also varied between 50 and 100 km and the width of the region where the initial rift temperature structure varies in thickness (W_t) is also varied. Table 2 lists the model cases considered.

As noted before the viscosity parameters should not be treated as being independent of the pre-rift temperature structure. If the temperature structure before rifting is not in steady-state then there will be changes in the model lithospheric thickness

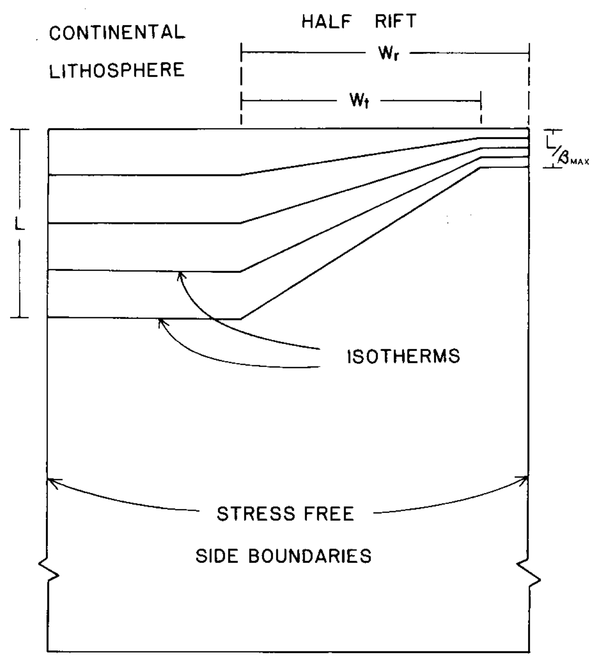

Fig. 3. The geometry of the initial rift temperature structure for a flow calculation is shown. The half-width of the rifted region (W_r) is varied in the models considered as is the width of the transition between zero stretching and maximum stretching (W_t) and the maximum stretching factor (β_{max}).

TABLE 2

The parameters which define the cases considered in this paper are shown. The half width of the rift is W_r and W_t is the width of the region over which the rift temperature structure varies laterally. The viscosity is defined by μ_{ref}, which is the initial viscosity at 150 km in the model, the activation energy (E) and the value of A_e defines the non-Newtonian rheology described by equation (2). For all cases, the activation volume (V) was 10^{-5} m^3/mole. The initial temperature structure for cases 1–4 resulted from 100 m.y. cooling of a 1300°C half space. For cases 5–7, the initial temperature profile was linear to a depth of 125 km and constant at 1300°C deeper

Case No.	Geometry			Viscosity		
	W_r (km)	W_t (km)	max β	μ_{ref} (10^{18} Pa s)	E (kcal/mole)	A_e
1	100	100	5	1.0	100	0
2	100	100	5	4.0	100	1.00×10^{-3}
3	100	100	5	4.0	70	0
4	50	50	5	1.0	100	0
5	100	100	5	1.0	100	0
6	50	16	2	1.0	100	0
7	50	16	2	5.0	70	0

after rifting which are not an effect of the passive thinning assumed here. The heat transported by convection depends on the temperature of the asthenosphere since the viscosity of that region depends on temperature. This heat transport must match the heat transported through the lithosphere of a given thickness. Therefore, in the self-consistent calculations 5–7, the viscosity parameters of E, V and μ_{ref} are combined so that the heat transported, when the asthenospheric temperature is 1300°C, equals the heat conducted by the linear gradient through a lithosphere of the given initial thickness.

In calculations of the effects of small-scale convection below a lithosphere which did not have an imposed rift temperature structure [20] it was shown that for convective wavelengths up to about 400 km the stress dependence of viscosity had little effect on the heat transport across the asthenosphere. This is a result of the small deviatoric stresses and strain rates for such systems. As noted by Fleitout and Yuen [19] when the deviatoric stresses in a system are below about 10 bars then the effects of non-Newtonian rheology should be negligible. However, when a rift temperature structure is introduced, larger deviatoric stresses and strain rates result from the large horizontal temperature gradients of this situation. The strain rate dependence of viscosity may play an important role in affecting the flow under the flanks of rifts. The effective viscosity defined by equation (2) is lowered in the regions of high stress and, therefore, will increase flow rates in those regions. In a sense, the strain rate dependence is the only free viscosity parameter, since the others are tied to the pre-rift temperature structure.

The calculations presented here differ in several ways from those of Keen [13]. In that work, the viscosity is taken to depend only approximately on temperature and to be a simple function of depth, in order to simplify the calculation of the flow field. Here, I consider explicit temperature, pressure and strain rate dependent rheologies. In the previous work [13] it has been suggested that when asthenospheric viscosities are sufficiently low small-scale convection may cause uplift of rift shoulders, but no quantitative calculation of the amount of that uplift are given. Finally, in no previous work has consistency in the pre-rift heat transport by the asthenosphere and that conducted by the lithosphere been required, as it is in some of the present models.

6. Results

Several of the model parameters which were varied in this study had a large effect on the geologically relevant results. I considered the effect of variations in the average viscosity, the activation energy, inclusion of stress-dependent viscosity, different initial temperature profiles and different widths of the rift zone (see Table 2).

There is considerable uplift produced by a combination of the lateral conduction of heat and the increase in the general advective heat flux in all these calculations. In Figs. 4 and 5 contours of the temperature field are shown which illustrate how the flow changes the temperatures under a rift. Fig. 6 shows that dynamical effects on the surface deformation are much smaller than the changes in the temperature structure. It is important to estimate the amount of uplift which would have occurred without convective heat transfer. The surface deformation with only conduction acting on the initial temperatures of case 4 is shown in Fig. 7, compared to the results from that convective calculation. The uplift is restricted to a narrower region and is of smaller amplitude than the

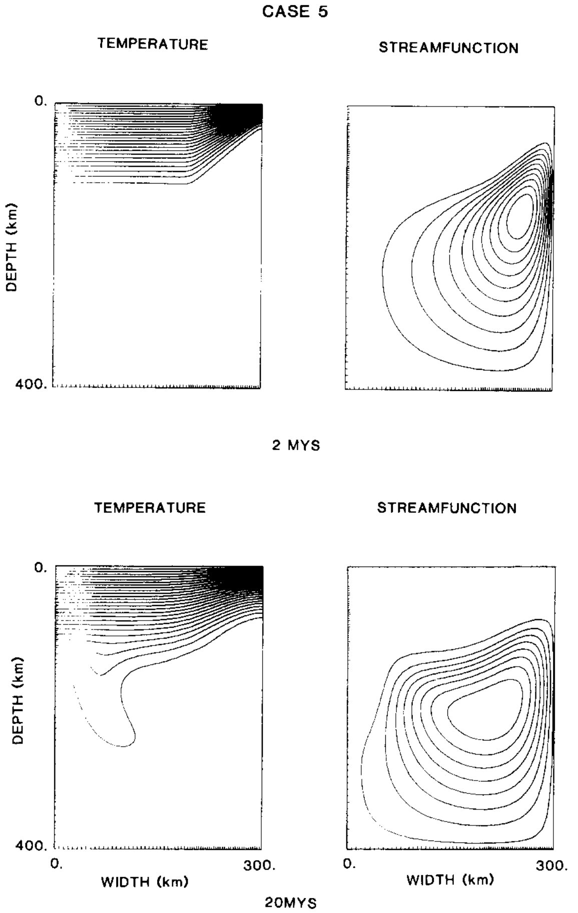

Fig. 4. Contours of constant values of temperature and stream function are shown for the indicated times for case 105 as defined in Table 2. The temperature contours are for every 50°C between 0 and 1300°C and the stream function contours are evenly spaced between zero and the maximum value in the box. The grid point positions are indicated by tick marks on the left and bottom sides of the box.

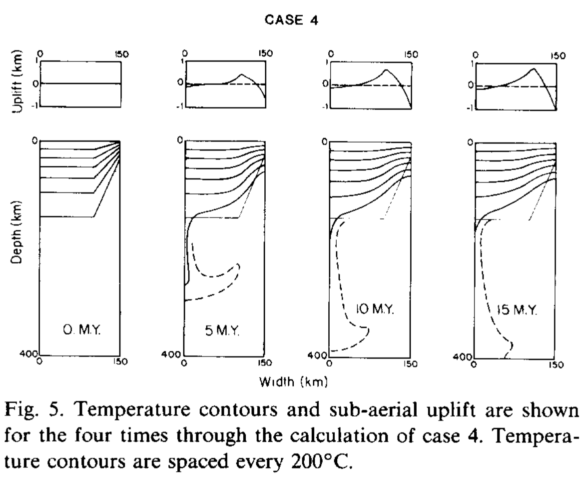

Fig. 5. Temperature contours and sub-aerial uplift are shown for the four times through the calculation of case 4. Temperature contours are spaced every 200°C.

uplift for the corresponding times for model. 4. The total uplift integrated over the area is about 2.5 times greater for this particular convective case compared to the conductive case.

In case 1, much of the uplift relative to the initial surface elevation is not centered over the unthinned flanks of the model rift but is closer to its center. This is due to the large width of the rift. The effect of the concentrated downwelling due to this pattern is illustrated in Fig. 6 which shows the components of the surface deformation for a time 25 m.y. into case 1. The part of the deformation which is due to the flow induced stresses at the base of the lithosphere is negative over the downwelling, but, as in all the cases, this component is small compared to the effect of temperature and thickness variations in the lithosphere. Case 3 was defined with the same parameters as case 1, but the average viscosity was higher by a factor of 4. This resulted in a similar pattern of surface deformation while the amplitude of the deformation was reduced by about 40% compared to case 1.

Inclusion of non-Newtonian rheology increases

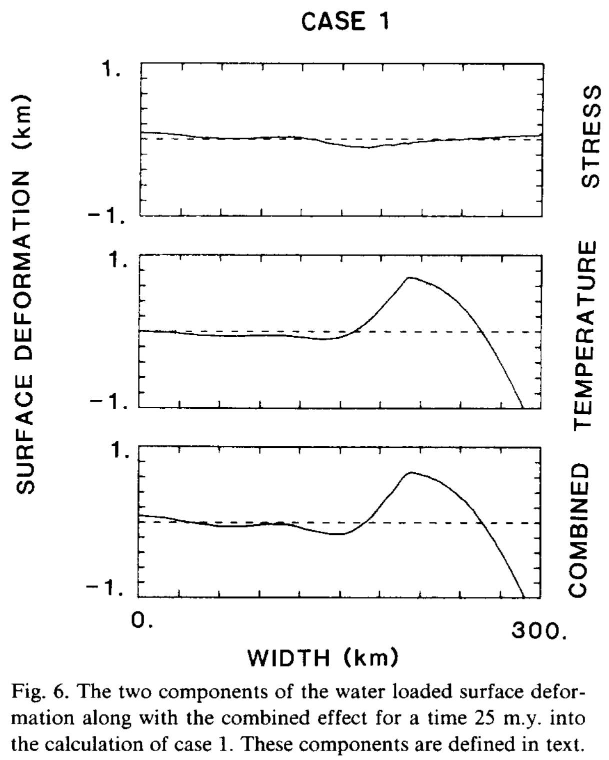

Fig. 6. The two components of the water loaded surface deformation along with the combined effect for a time 25 m.y. into the calculation of case 1. These components are defined in text. As for other cases, the component due to the convective stresses is small compared to that due to the temperature variations.

the uplift produced by convection. Case 2 and case 1 had the same amount of uplift on the rift flanks even though case 2 had a higher reference Newtonian viscosity than case 1 by a factor of 4. Simply increasing the reference viscosity acts to decrease the uplift. This is shown by case 3 which had the same reference viscosity as case 2 but had no stress dependence included in calculating the viscosity.

The effect of considering a linear initial temperature profile for the unrifted lithosphere is considered in the self-consistent cases 5–7. The linear profile is probably a better approximation to most rifts. In case 5 the geometry of thinning and the viscosity parameters are the same as those used in case 1. In case 6 the viscosity parameters are the same as those used in case 4 and the total amount of stretching represented in the rift temperature structure is the same, though the distribution is different. The most important difference between them is the initial lithospheric temperature profile which is linear for cases 5 and 6 and is described by an error function (i.e. curved) for cases 1 and 4. The maximum uplift in case 5 is 30% lower than that for case 1. For case 6, the difference between the convective calculation and a calculation using the same rift temperature structure where only conductive transfer of heat was allowed is shown in Fig. 8. The difference between the conductive and convective estimates of the shoulder uplift is not as large as for the same comparison for case 4. The linear profile minimizes the amount of material at the base of the lithosphere which is at a high temperature and so has a low viscosity. It is only the low-viscosity material which can be incorporated in the flow. The more material that is swept

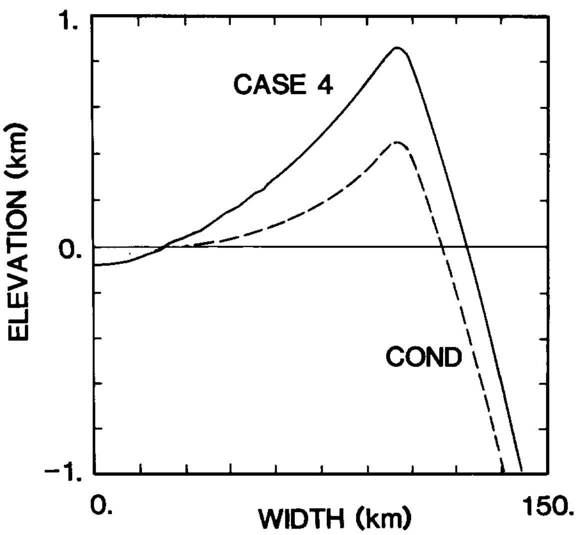

Fig. 7. The sub-aerial elevation for a conductive case with the same initial temperature structure as for case 4 is compared to the results of case 4 at a model time of 20 m.y. in each case. The conductive uplift averaged over area is about 40% of that for case 4 and the subsidence of the center of the rift is greater than for the convective case.

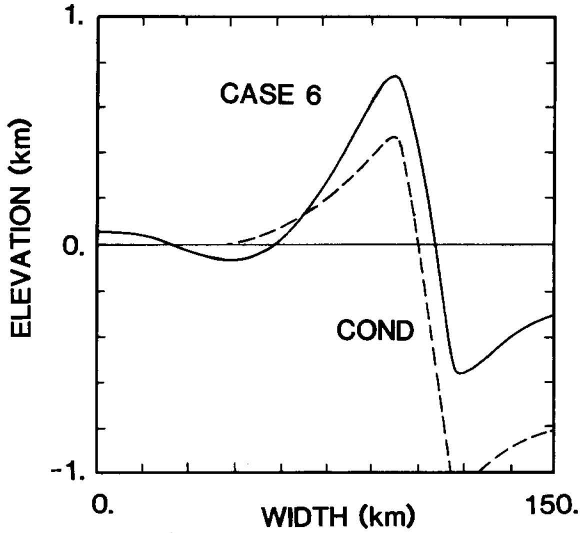

Fig. 8. The sub-aerial elevation for case 6, which is one of the calculations with a linear initial temperature profile and viscosity parameters which were consistent with the initial temperature structure. This is compared to the elevation for a conductive calculation with the same rift geometry. Both represent a model time of 20 m.y.

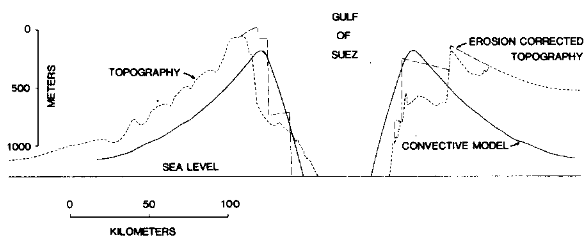

Fig. 9. A comparison of model results and topographic data across the Gulf of Suez shown at a vertical exaggeration of 75 : 1. The solid line is the elevation for case 4 at a model time of 20 m.y., which was shown in Fig. 7. The zero line for the model results was placed at a level of 200 m. The line of short dashes is the topographic data across the Gulf of Suez at 28°50′N. The line with longer dashes is the estimate of topography if eroded material were added back to the rift shoulders done by Steckler [11].

into the flow, the greater the uplift of the flanks of the rift. Case 7 was set up with the same geometry as in case 6, but the temperature dependence of viscosity was lowered so that the change in viscosity for a given change in temperature was nearly half as large. The reference viscosity was increased so that the pre-rift asthenospheric heat flux was the same as for case 6. The uplift for this model was essentially the same as was found for case 6.

7. Discussion and conclusion

The numerical difficulties in dealing with very narrow zones where temperature gradients and physical parameters vary rapidly and in treating non-Newtonian rheology has forced me to consider cases where the effects of convection may not be as large as they are in real rifts. Since I do consider two different model widths (50 and 100 km widths of the half-rifts), I have learned something about the importance of the width in changing the effect of convection beneath the rift. Just as with models where only conductive heat transport is allowed, the convective models show that the narrower the rift and the transition from rifted to unrifted regions the greater the predicted shoulder uplift. I was only able to include non-Newtonian effects in one calculation which had a rather high average viscosity and a wide rift transition. Even in this case, the effect of the lowering of the effective viscosity (equation 2) in the region of high strain rates significantly increased the shoulder uplift. Thus, when calculations are done with narrow rift zones and non-Newtonian rheology, I expect the predicted uplift to be larger than for any of the cases considered here. Also increasing the initial thickness of the lithosphere or greater total stretching should increase the predicted uplift.

To show that the results of these calculations are roughly consistent with uplift data for continental rift zones, I consider topographic data from several major rift zones. Fig. 1 shows profiles across the Gulf of Suez, the Rhine Graben and the Rio Grande Rift and the southeast Australian continental margin. Each shows some degree of elevation of the areas flanking the subsided center of the rift. Shown in Fig. 9 is the model uplift for case 4 to the same scale as the topographic profiles across the Gulf of Suez. Case 4 was constructed with an initial rift temperature structure representing the same average lithospheric extension and width for the rift zone as has been estimated for the Gulf of Suez [11]. This exercise shows that simple convective effects can produce uplifts which are of the same magnitude and have the same spatial distribution as those observed. Conductive transport of heat alone cannot do this.

Two processes which may bear on this problem have not been included in the models. One is flexure. The assumption of Airy isostasy may be good during rifting, but the elastic strength of the lithosphere should be important later. Flexure would tend to support the rift flank topography (i.e. freeze it in) even after the high temperatures there have been reduced by cooling. Without flexure, the topography would be lost on cooling. The other effect is partial melting of the base of the lithosphere under the flanks due to the heat brought there by convection and conduction. This may give rise to alkaline volcanism which is observed in many rift zones [39]. The advective transport of heat through the lithosphere by these magmas would add to the thermal uplift of the flanks.

The results of this preliminary study show that the effects of convection induced by a passive rift temperature structure can be a major cause of the uplift of the flanks of rifts. We have shown that the predicted uplift is greater for narrower rifts and for lower average viscosities. However, if we consider the rifting to be passive in origin, we are not free to choose any average viscosity. These results indicate the stress dependence of viscosity can add to the effects of uplift since this tends to reduce the viscosity in the high strain rate areas at the edge of the rift. The small-scale convection beneath a passive rift can also account for the apparent need for two levels of thinning of the lithosphere in conductive models of the subsidence of rifts since the subsidence all across a rift is affected by the convective transport of heat.

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