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Dynamic Topography in Rift Zones: Implications for Lithospheric Heating [Extended Abstract]

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Dynamic topography in rift zones: implications for lithospheric heating [extended abstract]

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Extension of the lithosphere in rift zones results in heating of the crust and mantle beneath the rift. Based on simple stretching models (McKenzie 1978), the subsidence history of rifts has been attributed to crustal thinning during extension and subsequent conductive cooling of the lithosphere beneath the rift. The amount of subsidence caused by cooling should provide an estimate of the amount of heat added to a column of lithosphere as it extends. Models of this type, which assume local isostatic compensation and uniform extension over the whole thickness of the lithosphere, underpredict the amount of post-rift thermal subsidence (Royden & Keen 1980) and do not explain the flanking uplifts associated with rifts. Predictions of central subsidence from models that do not explain flanking uplifts should be suspect since the same mechanisms producing the uplift may also affect the central subsidence.

The origin of flanking uplifts may have important implications for the amount and horizontal distribution of lithospheric heating beneath a rift. In a study of the Gulf of Suez rift, Steckler (1985) proposes that flanking uplifts are isostatically supported by thermal expansion due to heating. Because horizontal heat conduction occurs too slowly to explain the amplitude and width of flanking uplifts, convection due to stretching-induced horizontal temperature variations beneath the rift (Buck 1986) has been suggested as a heat transfer mechanism.

However, the strength of the lithosphere may also contribute in several ways to flanking uplifts. Vening-Meinesz (1950) first suggested that flanking uplifts were due to regional compensation of a mass deficit of crust/mantle columns within the rift by elastic flexure of the lithosphere. The Vening-Meinesz theory also predicts the width of a rift zone based on the elastic thickness or flexural rigidity of the lithosphere. The predicted width of flanking uplifts of the Suez rift, based on a flexural rigidity consistent with the rift width, is very similar to that observed. However, the amplitude of the uplift due to the mass deficit associated with sediments filling the rift alone is only about half that observed. This suggests that additional vertical forces are required to produce the uplift.

Vertical forces generated by the horizontal extension of a lithosphere of varying thickness may contribute to the observed flanking uplift and central subsidence of rift zones. This is illustrated by the simple model in figure 1, which considers a strong layer that has thinned due to extension beneath the rift zone. The lithosphere or strong layer is treated as a viscous or plastic continuum overlying a much weaker fluid mantle substrate. Horizontal equilibrium of the small darkly shaded material element at the bottom of the layer requires that an applied horizontal extensional stress $\bar{\sigma}_{xx}$ induce a shear stress σ_{xz} within the layer. Flow within the layer produced by this induced shear stress results in surface subsidence at the centre of the thinned region and flanking uplift. In the absence of flow caused by the induced shear stresses, surface topography would decay in response to gravitational forces. The amplitude of the

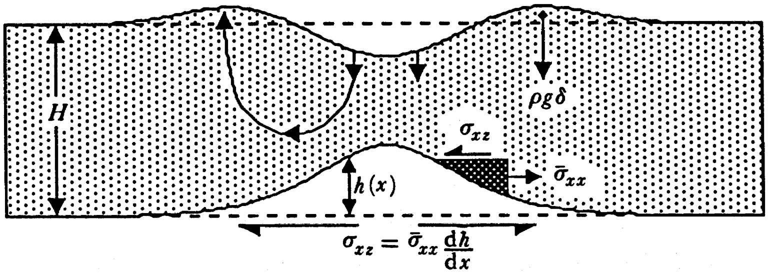

Figure 1. In a layer with thickness variation $h(x)$ extending due to a horizontal stress $\bar{\sigma}_{xx}$, flow due to the induced shear stress σ_{xz} causes surface topography. The equilibrium surface topography represents a balance between vertical flow at the surface due to gravity and the induced shear stress.

equilibrium surface topography is therefore determined by a balance of vertical motion at the surface due to gravitational forces on the topography and the induced shear stresses. It is important to emphasize that, in this simple example, the density of the layer is the same as its substrate so that surface topography is supported entirely by stresses within the layer and is not due to vertical isostatic forces. The present analytical formulation of the model assumes that the layer thickness variation is small, so that flow within the layer can be calculated by the induced shear stresses acting on the bottom of a layer of constant thickness. Figure 2 shows the surface topography resulting from the layer thickness variation shown. The amplitude of the surface relief is proportional to magnitude of the layer thickness variation and depends on a single non-dimensional parameter S defined in figure 2.

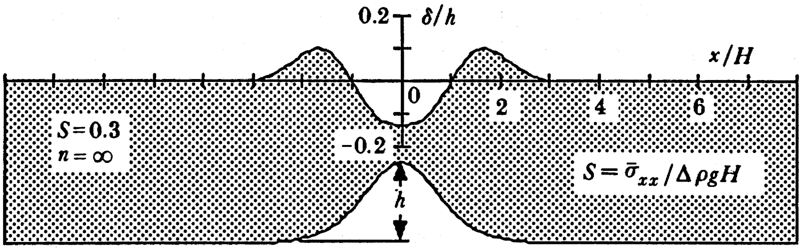

Figure 2. Stress-supported surface topography δ caused by a layer thickness variation of the form shown with amplitude h. The strong layer is taken to be a perfectly plastic material with stress exponent $n=\infty$, a continuum representation of deformation due to distributed faulting.

For application to rifts like the Suez rift, the strong layer is assumed to represent the brittle upper crust, and a perfectly plastic layer is adopted as a continuum idealization of deformation due to faulting. The formation of rift zones by the growth of a small amplitude strong layer thickness disturbance, as considered by Zuber & Parmentier (1986), predicts a flank-to-flank rift width of about four layer thicknesses. A brittle layer thickness of about 15 km would produce a rift zone comparable in width to the Suez rift. If the brittle crust thins by 30% at the centre of the rift zone and the applied horizontal stress is controlled by frictional sliding on faults, the surface relief predicted by results like that in figure 2 will be on the order of 1 km. Dynamic, stress-supported topography may therefore contribute significantly to uplifting the flanks and deepening the central depression of rifts. This may substantially reduce the amount of heating of the crust and mantle required to explain uplift and subsidence in rift zones.

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