How much variability in process and parameters is required to explain rifted margins?

by

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There is no unified theory of continental rifting and passive margin formation. To highlight this fact, but also to illustrate that some progress has been made in defining problems of rifted margins, I discuss three topics. One overarching theme is that we need to understand what is the minimum variability in processes and parameters needed to explain the various types of observed continental rifting.

1) **Is rifting quantized? Or controls on the large-scale structure of rifts.**

One view is that there are two basic types of rifts- “passive rifts” which were stretched apart by far-field forces and “active rifts” where the mantle might have pushed up the surface causing the local rifting. Active rifts are characterized by voluminous volcanism before the rifting. Seismic observations show that many rifts, previously thought to have essentially amagmatic and “passive”, such as the North America-Africa conjugate margins, were sites of massive intrusion and extrusion of magma.

More recently, continental extension has been categorized in terms of the strain distribution in the rift including the **rift width**. Some rifts are narrow (of order a 100 km wide) and others are many times that width. The first explanation suggested for variable rift widths was that it depends on the rate of extension. Slow rifts could strengthen as strong mantle replaces weaker crust as long as thermal advective heating of the rift is largely countered by conductive cooling. This can only happen if the rate of extension is very small. In most treatments of this mechanism the initial lithospheric and crustal thickness is the same and only the strain rate varies. However, a number of regions of present day extension, like the Basin and Range Province, are places of anomalously thick crust and may have been regions of high heat flow even before extension began.

I argue that variations in the thickness of the crust and lithosphere at the start of rifting lead to fundamentally different modes of extension. During rifting the crust and lithosphere can be locally thinned. The interaction of strength differences, related to lithospheric thickness variations, with buoyancy forces, related to crustal thickness variations, lead to three distinct modes of extension. In metamorphic core complexes the upper crust extends in a narrow region, while the lower crust thins over a broad region. Core complexes form in extremely thin lithosphere and very high heat flow and in continents this is generally where the crust is anomalously thick. Wide rifts occur in regions of high heat flow and thick crust, but not in as extreme conditions as needed for core complexes. Crustal and mantle melting may occur during the development of core complexes and wide rifts since melting temperatures are reached at shallow depths in these regions.

Related to this topic is the "**tectonic force problem**". There may not be sufficient tectonic force to initiate a passive narrow rift in thick lithosphere. Mantle plume derived magmatism may be a necessary condition for the initiation of narrow rifting in such regions. Magmatic accommodation of extension can occur at much lower stress differences than needed for faulting through thick brittle lithosphere. Red Sea rift may be an area where early magmatism weakened the lithosphere enough for rifting to proceed with little or no continued magmatic input. One thing, which is particularly unclear at present, is whether dikes propagate laterally for as much as a thousand kilometers from a source region or whether the region of rapid mantle melting can be thousands of kilometers wide.

References

Buck, W. R. (1991). "Modes of continental lithospheric extension." J. of Geophysical Res., B, Solid Earth and Planets **96**(12): 20,161-20,178.

Kusznir, N. J. and R. G. Park (1987). "The extensional strength of the continental lithosphere: its dependence on geothermal gradient, and crustal composition and thickness." in Coward, M.P., J.F. Dewey and P.L. Hancock (eds) Continental Extensional Tectonics, Geol. Soc. London, Spec. Pub. 28: 35-52.

Sengor, A. M. C. and K. Burke (1978). "Relative timing of rifting and volcanism on Earth and its tectonic implications." Geophys. Res. Lett. **5**: 419-421.

2) **How are different normal fault patterns formed? Or controls on the fine-scale structure of rifts.**

Field mapping and seismic observations show that the pattern of high angle normal faults is different across different rifts. The widths of sub-basins differ, the throw on faults is not uniform and fault sets can locally dip in the same direction or be variable in dip. In continental core complexes, as well as at inside corners of mid-ocean ridge-transform faults in slow spreading ridges, normal faults are mapped that dip at very low, sometimes negative angles. We have very little insight into what controls the distribution and dips of normal faulting in extensional environments. One idea is that some faults have fundamentally different properties, such as a lower friction coefficient, perhaps due to pore fluid pressures, than do other faults. In contrast to this view is the idea that it is mainly the thickness of the lithosphere and flow properties of the crust that may control the pattern of crustal faulting. Recently developed numerical models, that allow the spontaneous formation of faults according to rules for strain weakening, may give some insight into the problem. They show that not only the amount of weakening on a fault, but the rate of weakening with strain is important tot the pattern of faulting developed. Very slow weakening with strain leads to no localization of slip on fault-like structures. Very fast strain weakening can lead to shattering of the brittle crust and in extreme cases to very small offset on faults before they are replaced by other faults.

Related to this topic is the "**weak fault paradox**". Some contend that low-angle normal faults imply that some special faults had almost no strength when they slipped. We find that for moderate levels of cohesion and rates of cohesion reduction with strain, low-angle like fault structures can be produced by extension of thin brittle layers. No active slip on low-angle faults is needed in these models where inactive portions of the fault rotate over to flat or even negative dips.

References

Buck, W. R. (1993). "Effect of Lithospheric thickness on the formation of high-and low-angle normal faults." Geology **21**: 933-936.

Forsyth, D. W. (1992). "Finite extension and low-angle normal faulting." Geology **20**: 27-30.

Lavier, L., W. R. Buck, et al. (1999). "Self-consistent rolling-hinge model for the evolution of large-offset low-angle normal faults." Geology **27**(12): 1127-1130.

3) What causes extra post rift subsidence? Or Dynamic Subsidence of Passive Margins.

Analysis of passive continental margins frequently indicates greater subsidence than can be explained by the estimated crustal and lithospheric thinning. Often extra subsidence begins at the start of seafloor spreading in adjacent areas.

These observations have been used to support the idea that margin-scale "detachment " structures are important. One side of the margin is viewed as the upper plate and the other is the lower plate of gently dipping normal fault or detachment. On the upper plate side, a small amount of stretching near the surface would mask the fact that the lower crust and mantle lithosphere below have been severely thinned. This thinning would lead to syn- and post-rift subsidence. However, as conjugate margins have been studied it seems that both margins of the rift want to subside the way this model predicts for the upper plate margin. This is sometimes described as the "upper plate paradox".

A possible explanation of this paradox may involve mantle flow. This pulse of subsidence may result from flow of anomalously low-density asthenosphere from under rifted lithosphere to the area of even thinner lithosphere under a nascent ocean basin. Plumes may have brought hot, low-density mantle to shallow depths, where it could pool in a region of locally thin lithosphere. Recent work suggests that plume injected asthenosphere may have a sufficiently high viscosity that it does not convectively cool on a short time scale. Thus, hot asthenosphere might reside under a rift for many millions of years before it can flow out. The thinning of the layer of low-density asthenosphere could cause as much as a kilometer of water loaded subsidence. The rate of subsidence would depend on the initial thickness and width of the pooled asthenosphere and on the rate of plate divergence. For wide rifts and slow rates of seafloor spreading the rate of added subsidence would be similar to that of thermal subsidence. This dynamic subsidence of passive margins may provide a constraint on mantle convection and melting.

References

Driscoll, N. W. and G. D. Karner (1998). "Lower crustal extension across the Northern Carnarvon basin, Australia: Evidence for an eastward dipping detachment." Journal of Geophysical Research **103**(4975-4991).

McKenzie, D. P. (1978). "Some remarks on the development of sedimentary basins." EPSL **40**: 25-32.

White, R. S. and D. McKenzie (1989). "Magmatism at rift zones: the generation of volcanic continental margins and flood basalts." J. Geophys. Res. **94**: 7685-7729.