ARCHITECTURE OF CONTINENTAL RIFTS

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Architecture of continental rifts

1. Introduction

This paper describes the mechanism of continental rifting which lead to the formation of modern continental rift zones like the East African Rift, the Rio Grande Rift of the western United States, the Rhine Graben of Europe and the Baikal Rift of central Asia. At first, the paper gives an overview about the main features of the continental rift systems and then it explains the process of rifting.

2. Characteristics of modern continental rifts

Rifts are divergent intraplate structures of the continental lithosphere and are formed in response to horizontal extensional stresses. Regions of rifting can extend to about 1000 km along strike with a rift width of 50 to 100 km (or more) and an evaluation of 1 to 2 km.

An idealized cross-section shows a central graben flanked by steep boundaries made up of a series of faults. The fault structures may be either symmetric or asymmetric across the rift, depending on the geometry of the fault system. During rifting the graben shoulders are uplifted, sometimes higher than the surrounding plateau (shoulder uplift). Seismic studies suggest that rifts are characterized by a series of half-grabens and that some steep faults near the surface become listric near-horizontal detachments at depth. Along these detachments a transition from brittle to ductile rock deformation takes place. Extensional faults which appear to be planar and inclined at 30° to 60° generate large magnitude earthquakes. Continental rifting is nearly always accompanied by a series of parallel dykes, volcanism along fissures and central volcanoes which forms if a region along fissures begins to dominate the outpouring of magma. In greater depths and on the crust-mantle boundary impinge mafic layered intrusions known as sills.

Many contemporary rifts are characterized by negative Bouguer gravity anomalies, high heat flow and more or less volcanic activity. High heat flow means 90-115 mWm$^{-2}$, which is in contrast to provinces older than 250 Ma, where heat flow amounts to ~50 mWm$^{-2}$ and in younger provinces to ~77 mWm$^{-2}$ (Allen & Allen 1992). The negative Bouguer gravity anomalies are associated with upwelled low-density asthenosphere. The uprisen asthenosphere causing an moving upward of the lithosphere-asthenosphere boundary which is an isotherm of 1300°C. Therefore, the heat flow below the axial region of the rift valley is higher than normal. In contrast, heat flow on the flanks of the rift domes may be normal. These points suggest a shallow Moho position below rift zones.

3. Rifting process

3.1 Active versus passive rifting

Continental rifting can be classified into two categories, depending on the mechanism of rifting. There is active and passive rifting. The two styles are caused by two recognizable sources of lithospheric tension and have different surface expressions.
In case of the *active rifting* (Fig.1.A), the rifting process is a result of local tension associated with an underlying upper mantle plume and characterized by initial regional doming of the crust, extensive volcanism and late stage graben formation.

*Passive rifting* (Fig.1.B) is a passive response to a regional stress field resulting from plate boundary forces on opposite sides of a large predominantly continental plate (BOTT 1992). Initial graben formation, local uplift of the graben shoulders, and little or no volcanism characterize this type of rifting. Rifting takes place first and doming may follow.

Real world cases may exhibit aspects of both sources of tension.

### 3.2 Active rifting

The impinging of a mantle plume on the base of the lithosphere initiates the process of active continental rifting. Thermal expansion and magmatic underplating generate dynamic uplift above the plume. This dynamic uplift leads to regional crustal and mantle lithosphere doming.

#### Mantle plume to rift evolution

A plume from a convective instability near the core-mantle boundary (~2900km) rises as the result of the thermal buoyancy (Fig.2.1). The plume consists of two parts, a large mushroom head followed by a relatively narrow tail (or feeder conduit) which connects the head to the source region. During ascent, the head grows by entrainment and reaches a diameter of approximately 400-600 km (Fig.2.2). If the plume passes the discontinuity at 670 km (lower-upper mantle boundary), he can separate from the tail and maybe a new plume is forming on the core-mantle boundary. The diameter of the rising plume enlarges up to 800-1200 km due to entrained lower mantle material. If the plume head approaches the base of the lithosphere the head flattens out into a disc with a diameter of 1500-2500 km and a thickness of 100-200 km (CAMPBELL & GRIFFITHS 1990) (Fig.2.3). The plume head is a mixture of material both from the hot source (20-40%) and from the lower mantle. Upper mantle material will be displaced by the head, but not wrapped into.
If the rising plume approaches the lithosphere, the head begins to replace the thin boundary layer of surrounding mantle material and is heated by conduction until its temperature and density becomes comparable to that of the plume. The heat transfer mechanism changes from predominantly convective to mainly conductive. Its average temperature amounts 1350-1400 °C and differs +70-120°C from the surrounding mantle temperatures (BOILLOT & COULON 1996). The amount of uplift above a plume is thus sensitively related to the initial thermal structures of the overlying material.

Long before the plume head reaches the lithosphere the elevation of the Earth's surface is increasing, reaches a maximum and then declines slowly. The simple uplift of the Earth's surface induces horizontally-directed deviatoric stresses.

The uplift above the plume can be considered in three separate parts. At first, the significant initial uplift which occurs above the centre of the plume while it is still rising. This uplift ceases by the following onset of voluminous basaltic volcanism 10 to 20 Ma later, the continental flood basalt event.

At first, the volcanism starts suddenly and extends over a large area of 2000-2500 km and then contracts rapidly to a narrow linear volcanic chain, 100-300 km across and marks the later track of the hot spot. In an initial short period of 1-4 Ma the bulk of the basalts erupt, although the volcanism can extend over a period of at least 7-17 Ma as long as the thermal anomaly continues to rise (CAMPBELL & GRIFFITHS 1990).

The later phase of uplift is characterized by broader-scale uplift, which results as the plume head collapses and spreads laterally within the upper mantle. As the plume head flattens within the uppermost mantle, uplift will be followed by subsidence above the plume axis (to last for 20-60 Ma) (HILL 1991). Near the margins of the plume the uplift can continue.

The crustal extension depends on the tensional effects imposed by the rising plume and may lead to gravitationally-induced rifting without splitting the continent and producing new ocean crust. The rifting process is accompanied by thinning of the lithosphere beneath the axis and under the dome. Crustal thinning is accompanied by deep crust faults that produce steps in the crustal thickness. The lithosphere can extend or stretch until crustal thickness.

Heat flow in central region increases and significant volumes of basalt from the asthenosphere intrude into the crust and solidify. This significant decompression melting within the plume head originates from ~70 km below the surface, because at a depth of about 70 km there is the "dry" basalt solidus (HILL 1991). To form flood basalts, there has to be a partial melting of 10-15%.

**Continental flood basalts**

Active continental rifting is not necessary connected with the occurrence of continental flood basalts, but in many cases continental rifting followed soon after a period of flood basalt volcanism. Four cases are described in detail:

- rifting of the Seychelles Platform from India (Deccan Traps)
- opening of the North Atlantic (basalts of the Tertiary North Atlantic Province)
- opening of the South Atlantic (Paraña-Etendeka)
- beginning of extension in the Red Sea (Ethiopian Traps)

Flood basalts appear as the first volcanic expression of a mantle plume and are brief geologically events, typically occurring over 2 to 3 Ma. The flood basalts extend over a large equidimensional area of 2000 to 2500 km and are associated with picritic basalts which are
found close to hotspot tracks. Continental flood basalts are formed by fractional crystallisation of a picritic parent magma (Fig. 3) within the plume head, whereas the associated picrites originate from the hot and relatively uncontaminated source mantle of the plume axis or tail.

The plume head is cooler than the axis, so it undergoes a lower degree of partial melting, giving basalts a lower MgO content than the picrites. Picrites are enriched in MgO (15-25%) and in highly incompatible elements (Rb, Ba, K, Nb, La...), but the degree of enrichment is highly variable. Those with MgO contents between 12-18% consist mainly of glass and quenched olivine. Picrites with about 18% MgO contain olivine phenocryst (SAUNDERS 1992).

### 3.3 Passive rifting
Passive rifting occurs above ordinary mantle and is initiated by tension in the plate interior. Such intraplate tensions are generated by plate boundary forces like subduction pull on opposite sides of a large predominantly continental plate (BOTT 1992) (Fig. 4). The continental plate may be unstable when subjected to regional tension. The resulting stress may cause stretching and thinning of the lithosphere and may lead to the failure of the continental lithosphere. The disturbed isostatic equilibrium leads to a compensating rise of the asthenosphere. Such anomalous subsurface loading and the associated isostatic surface loading (crustal thinning) produce opposing stress fields. The resulting stress depends on the relative thickness of crust and lithosphere (BOTT 1992). Consequently, local variations in the pre-rift strength of the continental lithosphere can be reactivated like relatively young orogenic belts. Examples include the opening of the North Atlantic, which closely followed Paleozoic mobile belts, and the rifting of Antarctica from Africa, which was largely confined to Late Proterozoic mobile belts.

Several models for the stretching and thinning of the continental lithosphere were developed to explain the formation of extensional basins. On the one hand, there are models dominated
by pure shear and on the other hand by simple shear. The latter results in asymmetrical rift valley.

3.3.1 Pure Shear – Mechanical Stretching
If pure shear deformation takes place, a symmetrical rift valley forms above upwelling asthenosphere.

The following uniform and non-uniform stretching models describe how the lithosphere accommodates extensional strain and how the main basin features - central graben and uplifted rift flanks - are formed.

**Uniform, instantaneous stretching**
(McKenzie model)
This simple geometrical stretching model is depth independent and explains the common occurrence of relatively unfaulited sedimentary basins overlying rifts.

Both the crust and subcrustal lithosphere stretch rapidly by the same amount of vertical thinning regardless of the different rheology of the lithospheric layers (Fig.5.). Lithospheric thinning allows passive upwelling of asthenosphere below the stretched lithosphere to maintain isostatic equilibrium. The isostatic compensation results in fault controlled subsidence of the top surface of the lithosphere and a rise of the Moho. This initial subsidence (syn-stretching) determines the depth of the newly formed basin. Due to this geologically instantaneous stretching event there exists a thermal anomaly below the newly formed basin as a result of the uprisen asthenosphere. In contrast to the active rifting model, there is no regional doming of the rift region. With time the thermal anomaly decays due to vertical conduction only and there is no heat loss to the rift margins. Moreover, the relatively high-density mantle lithosphere replaces relatively low-density asthenosphere which loads the overlying basin from below. This time-dependent period called thermal subsidence (post-stretching) and is not accompanied with lithospheric stretching. (ROBERTS & YIELDING 1994)

**Fig.5. Illustration of the lithospheric stretching model of McKenzie (1978) (after ROBERTS & YIELDING 1994)**

**Non-uniform stretching**
In many rift basins the initial and thermal components are not in the proportions predicted by the uniform stretching model. The thermal subsidence is too great, so that a greater anomaly than predicted in the McKenzie model has developed below the basin. To explain the
The depth-dependent stretching model bases on the view that the lithosphere layered into different rheologies.

Following that crust and mantle lithosphere extend by different amounts and that the amount of vertical thinning varies with depth. Consequently, the subcrustal region stretches over a larger region than the crust which results in the creation of a greater thermal anomaly as predicted in the uniform model.

The distribution of extension with depth may either be discontinuous or continuous.

Discontinuous stretching requires a physical decoupling of the crust and lithospheric mantle and implies that mass is not preserved during stretching.

In the continuous stretching model the physical decoupling within the lithosphere is not a prerequisite. In this model exists a gradation of stretching between crust and mantle lithosphere and mass is preserved during extension.

The initial fault controlled subsidence and the thermal subsidence are given by a solution of the same general form as that provided by McKenzie. In contrast the initial subsidence depends much more on crustal thinning than whole lithosphere thinning (ALLEN & ALLEN 1992). In contrast to the uniform model, where only vertical conduction takes place, there is also lateral conduction of heat into the cooler, unstretched lithosphere of the basin margins.

The basin itself undergoes rapid initial subsidence, while the rift flanks isostatical rise and expose to erosion (Fig. 6). The basin margins and the basin subside during the following thermal subsidence. If erosion occurred, the basin margins will subside below their initial, pre-stretch elevation. In contrast the rift flanks will return to their initial position, if no erosion takes place. (ROBERTS & YIELDING 1990)

### 3.3.2 Simple Shear

The simple shear model describes the asymmetrical stretching of lithosphere in which extension is controlled by a major shear zone or detachment.

There is a combination of upper crustal fault controlled extension and a much smaller region of concentrated lower crustal/mantle lithosphere extension. The lower crustal/subcrustal thinning produces thermal domes and erosional unconformities.

The asymmetrical stretching patterns of uplift and subsidence vary across the basin. A typical basin has an outer zone where only the upper crust stretches in contrast to an
inner zone where the upper crust stretches by a smaller amount as the lower crust and mantle lithosphere (Fig. 7). The outer zone is characterized by initial subsidence due to thinning of upper crust and there is no thermal subsidence. In contrast the inner zone shows both initial tectonic and later thermal subsidence. The additional stretching in the lower crust and lithospheric mantle cause early uplift (ALLEN & ALLEN 1992).

4. Conclusions

Continental rifting is a thermomechanical process which can lead to the break-up of continents. There are two mechanisms of rifting which play an important role: active and passive rifting. In active rifting an underlying mantle plume originating from the core-mantle boundary induces regional tension into the continental lithosphere by impinging on the base of the lithosphere. The continental lithosphere elevates and causes regional crustal and mantle lithosphere doming. This tension can lead to rifting of the continental lithosphere but not to continental break-up. Passive rifting is a passive response to a regional stress field resulting from plate boundary forces on opposite side of a large predominantly continental plate. The resulting stress cause stretching and thinning of the lithosphere and lead to the failure of the continental lithosphere. Local variations in the pre-rift strength of the continental lithosphere can be reactivated.

Both sources of tension result in stretching and thinning of the lithosphere. It is not clear that major continental break-up could occur readily in response to just one of these two sources of tension. The most favourable mechanism for continental break-up would be that a large continental mass is made unstable by plate boundary forces and the development of a hot spot in the underlying upper mantle triggers break-up.


**Literature**


