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From continental rifting to seafloor spreading: Insight from 3D thermo-mechanical modeling

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A B S T R A C T

Continental rifting to seafloor spreading is a continuous process, and rifting history influences the following spreading process. However, the complete process is scarcely simulated. Using 3D thermo-mechanical coupled visco-plastic numerical models, we investigate the complete extension process and the inheritance of continental rifting in oceanic spreading. Our modeling results show that the initial continental lithosphere rheological coupling/decoupling at the Moho affects oceanic spreading in two manners: (1) coupled model (a strong lower crust mechanically couples upper crust and upper mantle lithosphere) generates large lithospheric shear zones and fast rifting, which promotes symmetric oceanic accretion (i.e. oceanic crust growth) and leads to a relatively straight oceanic ridge, while (2) decoupled model (a weak ductile lower crust mechanically decouples upper crust and upper mantle lithosphere) generates separate crustal and mantle shear zones and favors asymmetric oceanic accretion involving development of active detachment faults with 3D features. Complex ridge geometries (e.g. overlapping ridge segments and curved ridges) are generated in the decoupled models. Two types of detachment faults termed continental and oceanic detachment faults are established in the coupled and decoupled models, respectively. Continental detachment faults are generated through rotation of high angle normal faults during rifting, and terminated by magmatism during continental breakup. Oceanic detachment faults form in oceanic crust in the late rifting–early spreading stage, and dominates asymmetric oceanic accretion. The life cycle of oceanic detachment faults has been revealed in this study.

1. Introduction

The continuous process from continental rifting to seafloor spreading is a key step in Wilson's cycle (Wilson, 1966), but the complete process is scarcely simulated. Understanding the inheritance of continental rifting in seafloor spreading is crucial to study the incipient oceanic ridge evolution and remains a big challenge (Lister et al., 1986; Taylor et al., 1999; Ebinger and Casey, 2001; Nielsen and Hopper, 2004; Taylor et al., 2009; Gerya, 2012). Compared to the continental rifting process, which has been widely studied by numerical and analogue modeling (Buck, 1991; Buck et al., 1999; Huismans and Beaumont, 2003; Corti, 2008, 2012), few studies have been done on the complete rifting–spreading process due to several difficulties. One difficulty is that the long extension process requires large strain to reach the final steady state of seafloor spreading from the initial intact continental lithosphere, and it includes many complex geodynamic processes, such as partial melting of the asthenospheric mantle, melt extraction and percolation towards surface, magmatic accretion of new oceanic crust and hydrothermal circulation at the axis of ridges resulting in excess cooling of oceanic crust (Gerya, 2010, 2013). Another difficulty is that oceanic spreading is by its nature a 3D problem, as large heterogeneities are present along oceanic ridges. Ultra-slow spreading ridges are mostly oblique to spreading directions and consist of alternating magmatic–amagmatic segments (Dick et al., 2003). Symmetric and asymmetric oceanic accretion (i.e. oceanic crust growth) are alternatively distributed along slow spreading ridges (Escartin et al., 2008). Overlapping spreading centers and transform faults are two common ways that oceanic segments interact ((Gerya, 2012), and references therein). These 3D features of oceanic ridges naturally require 3D models. Furthermore, the deficit of direct natural observations makes the rifting–spreading transition enigmatic. Unlike the widely distributed examples of continental rifts and oceanic ridges, very few natural examples record the rifting–spreading transition, and some good examples are the Woodlark Basin in the southwestern Pacific ocean (Taylor et al., 1999, 2009), the Laptev Sea margin in the Arctic ocean (Franke et al., 2001; Engen et al., 2003) and the Red Sea-Gulf of Aden system (d'Acremont et al., 2010; Leroy et al., 2010; Ligi et al., 2012; Brune and Autin, 2013).

As commonly believed (Dunbar and Sawyer, 1988, and references therein), continental rifts do not occur randomly, but tend to follow pre-existing weaknesses (such as fault zones, suture zones, failed rifts, and other tectonic boundaries) in the lithosphere, although explanations for the earliest weaknesses formed in the proto-lithosphere are debated. Many natural examples prove that the way rifts develop is to

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follow pre-existing lithospheric structures, for instance, the western branch of East African Rift (Nyblade and Brazier, 2002; Corti et al., 2007), the Main Ethiopian Rift (Corti, 2009), and the Baikal rift (Petit and Deverchere, 2006). The early-stage formed rift can be a template for the future rift development and continental breakup. However, numerous heterogeneities are generated during continental rifting, and rift propagation may divert from the initial rift trend (i.e. trend of pre-existing weakness), just as it has been proposed for the Main Ethiopian Rift (Keranen and Klemperer, 2008; Corti, 2009). To what extent rifting history influences continental breakup and seafloor spreading is hardly constrained. Moreover, the influence of rifting history on continental breakup and seafloor spreading is related to the formation of transform faults. One possible mechanism of transform fault formation which is widely proposed is related to the inheritance of pre-existing lithospheric weaknesses. Transform faults may initiate and develop along pre-existing zones of weakness that are nearly perpendicular to oceanic ridges, such as the long transform fault in the equatorial Atlantic ridge between South America and Africa (Wilson, 1965). The correspondence between the passive margins and the transform faults in the Gulf of Aden also suggests the possibility of inheritance of pre-existing weakness in some transform fault formation (d'Acremont et al., 2010).

Symmetric and asymmetric accretion are two oceanic ridge spreading modes. Symmetric oceanic accretion is featured by the roughly symmetrically distributed abyssal hills on both flanks of an oceanic ridge, while asymmetric oceanic accretion is characterized by the development of active detachment faults along one flank, and most of the detachment faults have 3D curved (convex towards ridge) geometries on a map view (Buck et al., 2005; Smith et al., 2006; Escartin et al., 2008). On a gross scale, oceanic accretion pattern has a strong relation with spreading rates. Asymmetric accretion is favored by slow spreading, while symmetric accretion is promoted by fast spreading (Buck et al., 2005; Puthe and Gerya, 2014, and references therein). On a small scale, however, alternating segments of symmetric and asymmetric oceanic accretion are observed along slow spreading ridges where the spreading rate varies slightly, such as the northern Mid-Atlantic Ridge (Escartin et al., 2008). In this case, the alternating symmetric/asymmetric oceanic accretion pattern is difficult to explain by variations in the spreading rate alone.

Allken et al. (2011) and Allken and Huismans (2012) investigated continental rift propagation and interaction in the upper crust through 3D numerical modeling using relatively simple temperature-independent (visco)-plastic rheologies. Extension strain in these studies was insufficient to reach continental breakup. Oblique extension of continental rifting is investigated based on 3D numerical modeling (Brune et al., 2012; Brune and Autin, 2013; Brune, 2014). Although continental breakup was reached in these models, seafloor spreading (e.g. oceanic crust accretion) was not simulated. Gerya (2010), Gerya (2013) and Puthe and Gerya (2014) numerically modeled the transform fault initiation (and development) and ridge segment interaction in 3D based on an incipient oceanic ridge or an idealized thinned continental lithosphere. Besides, several other 3D numerical modelings have been conducted with focuses on different stages of the rifting–spreading process (Van Wijk, 2005; Van Wijk and Blackman, 2005; Choi et al., 2008; Gregg et al., 2009), but the complete process is scarcely simulated.

In this study, we aim to investigate inheritance of continental rifting on incipient seafloor spreading by modeling the complete rifting–spreading process, with particular attention paid on the two oceanic accretion modes (symmetric and asymmetric accretion). The initial rheological structure of continental lithosphere and the geometry of a pre-existing weak zone are the two key parameters that we study. Two rheological structures named decoupled (DCP) and coupled (CP) are distinguished by the presence/absence of a strong lower crustal layer. Two types of model setup in terms of the weak zone geometry are investigated, one is 2D-like setup (i.e. a long planar extension-orthogonal weak zone going through the entire model box) and the other is 3D-like setup (i.e. a short weak zone going through part of the model box). The influence of the weak zone location (i.e. in upper crust or in uppermost lithospheric mantle) has also been investigated. Our preliminary modeling results show that the asymmetric oceanic accretion involved with active detachments is favored by DCP models, while CP models typically generate symmetric oceanic accretion. Initiation, development and termination of detachment faults, curved geometry of oceanic detachment faults, inheritance of continental rifting on seafloor spreading, and magmatism along passive margins are discussed.

2. Numerical implementation and model setup

2.1. Governing equations

The 3D thermo-mechanical coupled numerical code (Gerya, 2013) based on conservative finite-differences and marker-in-cell techniques is used to solve the mass, momentum and energy conservation equations for incompressible media:

$$\frac{\partial v_i}{\partial x_i} = 0 \tag{1}$$

$$\frac{\partial \sigma_{ij}}{\partial x_j} - \frac{\partial P_i}{\partial x_i} = -\rho g_i \tag{2}$$

$$\rho C_{p,eff} \frac{DT}{Dt} = \frac{\partial}{\partial x_i}\left(k \frac{\partial T}{\partial x_i}\right) + H_r + H_s + H_a \tag{3}$$

where v is velocity, σ the deviatoric stress tensor, P the total pressure (mean normal stress), ρ the density, g the gravitational acceleration, $C_{p,eff}$ the effective heat capacity (explained in below), T the temperature, k the thermal conductivity, H_r the radioactive heating, $H_s = \sigma\,\dot{\epsilon}$ the shear heating (product of deviatoric stress and strain rate), and $H_a = T\alpha \frac{DP}{Dt}$ the adiabatic heating. The Einstein notation is used for the indexes i and j, which denote spatial directions i = (x, y, z) and j = (x, y, z) in 3D. Lagrangian temperature equation is solved on the Eulerian nodes, and temperature increments are interpolated from nodes to markers by using the subgrid diffusion operation (Gerya and Yuen, 2003, 2007; Gerya, 2010), which can ensure physical consistence between nodal and marker thermal fields. Advection of temperature is implemented through marker advection. The Multigrid method is used to speed up the convergence of the Gauss-Seidel iterations for coupled solving of mass and momentum conservation equations. OpenMp-based parallelization is used in computation.

2.2. Rock rheology implementation

Visco-plastic rheology is implemented in our numerical models and Drucker-Prager yield criterion is used to determine whether viscous deformation or plastic deformation occurs. Viscous creep dominates model deformation when second invariant of deviatoric stress $\left(\sigma_{II} = \left(\frac{1}{2}\sigma_{ij}\sigma_{ij}\right)^{\frac{1}{2}}\right)$ is less than the plastic yielding criterion (σ_y). Effective creep viscosity which represents the competition between diffusion and dislocation creeps (Ranalli, 1995) is expressed as: $\eta = 1/(1/\eta_{diff} + 1/\eta_{disl})$, where η_{diff} and η_{disl} are computed as:

$$\eta_{diff} = \frac{1}{2} A_d {\sigma_{crit}}^{1-n} \exp\left(\frac{PV_a + E_a}{RT}\right) \tag{4}$$

$$\eta_{disl} = \frac{1}{2} A_d^{1/n} \dot{\epsilon}_{II}^{(1-n)/n} \exp\left(\frac{PV_a + E_a}{nRT}\right) \tag{5}$$