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Influence of a pre-existing basement weakness on normal fault growth during oblique extension: Insights from discrete element modeling

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ABSTRACT

We use discrete element modeling to investigate three-dimensional fault geometry and the three-dimensional evolution of a fault network that develops above a 60° dipping planar pre-existing weakness striking 60° relative to the extension direction. The evolution of the fault network comprises three stages: (i) reactivation of pre-existing structure and nucleation of new faults (0–10% extension); (ii) radial propagation and interaction between reactivated structure and new faults (15%–20% extension); and (iii) linkage between reactivated structure and adjacent faults (20%–25% extension). During the first stage, the pre-existing structure mostly reactivates, forming a long and under-displaced fault. New faults are mainly extension-perpendicular and dip at 60°. During the second stage, 'saw-tooth' fringes grow upwards from the upper tip of the reactivated structure (which becomes the major fault) and influence the density and orientation of surrounding faults. During the third stage, the reactivated structure links laterally and vertically with adjacent faults, creating non-planar fault geometries. Following linkage, the reactivated structure enhances the displacement of linked faults along branch lines. Our study demonstrates that pre-existing weak faults can be reactivated, propagating upwards in an irregular ('saw-tooth') pattern, and affecting fault density, orientation, dip and displacement, and providing the nucleation site of new faults.

1. Introduction

Normal faults developing during a single rift phase ideally strike perpendicular to the extension direction and show approximately collinear configuration (e.g. Anderson, 1951; Gawthorpe and Leeder, 2000; Cowie et al., 2000, 2005). The general evolution of a rift-related normal fault population in homogeneous crust is commonly considered in terms of a three-stage model: (i) fault initiation, characterized by the nucleation of numerous short, small-displacement fault segments; (ii) interaction and linkage between adjacent fault segments, and; (iii) continued activity on a few large, through-going fault systems that bound half graben depocenters (e.g. Cowie et al., 2000, 2005; Gawthorpe and Leeder, 2000; McLeod et al., 2000; Meyer et al., 2002; Gawthorpe et al., 2003).

Multiphase rift basins and rifts that are built on a previously faulted or folded basement are prone to develop arrays of non-collinear faults, with interaction between reactivated and secondary faults. Examples of non-collinear fault arrays that are interpreted to result from multiphase rifting include the NW Shelf of Australia (e.g. Frankowicz and McClay, 2010), Gulf of Thailand (e.g. Morley et al., 2004, 2007), Gulf of Aden (e.g. Lepvrier et al., 2002; Bellahsen et al., 2006), the northern North Sea (e.g. Badley et al., 1988; Færseth, 1996; Færseth et al., 1997; Odinsen et al., 2000; Whipp et al., 2014; Duffy et al., 2015), and Milne Point, Alaska (Nixon et al., 2014). Nixon et al. (2014) found that second-phase faults abut against reactivated first-phase faults, and showed that two abutting faults can link kinematically by reactivating a segment of the first-phase fault. Duffy et al. (2015) found similar evidence that second-phase faults abut against or were retarded by a reactivated first-phase fault in the northern North Sea. Such observations indicate that fault evolution in a multiphase rift basin is more complicated than that predicted by the aforementioned three-stage normal fault evolution model.

Physical models greatly help us understand how non-collinear faults and fault interactions evolve during two-phase extension (e.g. McClay and White, 1995; Keep and McClay, 1997; Henza et al., 2010, 2011). Henza et al. (2011) suggested that reactivated first-phase faults can interact and link with second-phase faults to form non-collinear fault geometries with a moderately developed first-phase fault population. However, these models have difficulty in visualizing the model interior and the three-dimensional fault geometry during extension. Although

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Nixon et al. (2014) and Duffy et al. (2015) illustrated three-dimensional fault geometry and classified fault interaction styles based on final fault displacement analysis, the complete fault array is still not well understood because the 'root' of first-phase faults is deeply buried. Questions remain about the evolution of normal fault networks where reactivation of pre-existing structures influences their development. Specific questions include: i) How are pre-existing faults reactivated and how do they propagate during subsequent rifting? ii) How does a reactivated normal fault influence and interact with newly formed faults as rifting progresses? iii) How different is the normal fault geometry near a reactivated fault compared to the geometry of more distant faults? In order to answer these questions we employ a three-dimensional discrete element model to simulate crustal extension with a pre-existing planar weakness striking oblique to extension direction. The model enables us to observe three-dimensional fault geometry during extension, making it possible to analyze fault growth history and the effect of the pre-existing weakness on fault network evolution in space and time.

2. Methodology

2.1. Discrete element model

The discrete element model used in the paper simulates the crust as an assembly of spherical elements (e.g. Cundall and Strack, 1979; Mora and Place, 1993, 1994), and has been successfully used to investigate crustal deformation, such as the growth of faults (e.g., Imber et al., 2004; Hardy and Finch, 2006, 2007; Schöpfer et al., 2006, 2007a, b; Egholm et al., 2007; Hardy, 2013; Finch and Gawthorpe, 2017), folding (Finch et al., 2003, 2004; Hardy and Finch, 2005), boudinage (Komoróczi et al., 2013) and contractional wedges in mechanical stratigraphy (Wenk and Huhn, 2013). In this discrete element model, the crust consists of an upper part that deforms in a brittle manner and a lower part that behaves in a firmo-viscous way (Ranalli, 1995) (Fig. 1a–b).

In the upper crust, neighboring elements (element i and neighbor j) interact in pairs through linear elastic repulsive-attractive force (Mora and Place, 1993) (Fig. 1b), which is represented by a breakable elastic bond that follows

$$F_{ijU} = \begin{cases} K(r - r_0), & r < r_b, \ \textit{intact bond} \\ K(r - r_0), & r < r_0, \ \textit{broken bond} \\ 0, & r \geq r_0, \ \textit{broken bond} \end{cases} \tag{1}$$

In Equation (1), K is the bond stiffness (elastic constant), r is the current separation between the element pair, r_0 is the equilibrium separation and r_b is the breaking separation. r_b is normally less than $1.1r_0$ (Mora and Place, 1994). Element i experiences an attractive force through the bond with the neighbor j (i.e. $r < r_b$), but no further attractive force when the bond is broken (i.e. $r > r_b$). A broken bond is not allowed to heal, and elements experience a repulsive force when they return to a compressive contact (i.e. $r < r_0$).

In the lower crust, elements interact through linear firmo-viscous (Newtonian fluid) forces including an elastic and a linearly viscous force in parallel, representing a firmo-viscous body (Fig. 1b). The component of the elastic force is

$$F_{ijL}^{elastic} = \begin{cases} K_c(r - r_0), & r < r_0 \\ 0, & r > r_0 \end{cases} \tag{2}$$

K_c is the bond stiffness in compression, consistent with K in the upper crustal elements. And the bond between elements in the lower crust has elastic properties in compression only (i.e. $r < r_0$). The component of viscous force is

$$F_{ijL}^{viscous} = -\eta \Delta \dot{x}_{ij} \tag{3}$$

Here η is the Kelvin viscosity and determined by empirical experiments, $\Delta \dot{x}_{ij}$ is relative velocity between elements i and j.

Since the lower crust behaves like a viscous fluid, elements within it can flow out of the model at the boundaries. To prevent this, the model is constrained by boundary walls in the x- and y-component directions. These walls exert a repulsive force on any element that crosses the boundary. In that way, the model is treated as a part of a larger system of elements that have the same mechanical properties. The force, F_{iB}, due to the boundary walls is given by

$$F_{iB} = -K_B r_B. \tag{4}$$

K_B is the elastic stiffness of the boundary wall and r_B is the distance by which the element exceeds the boundary (Wenk and Huhn, 2013).

As a whole, the crust is considered an elastic-brittle-plastic plate hydrostatically floating on a fluid mantle at a specific depth, in order to reach a hydrostatic equilibrium (King et al., 1988). This depth depends on a defined ratio of the mantle and crust densities. Under this circumstance, the force due to gravity and flotation, F_{iG}, is exerted on all elements in the vertical, z-component direction and follows

$$F_{iG} = g\left[(\rho_m - \rho_c)V_B - \rho_c V_A\right]. \tag{5}$$

Here g is the gravitational acceleration, ρ_m and ρ_c are mantle and crust densities respectively, and V_A and V_B are the volumes of an element above and below the hydrostatic equilibrium. The volume of an element above the hydrostatic equilibrium experiences a downward force, whereas the volume below the hydrostatic equilibrium experiences a resultant upward force. Additionally, a damping force that allows energy to be dissipated is applied to avoid kinetic energy building up in the closed system. This artificial viscous force is used to attenuate dynamic phenomena such as reflected waves from the boundary of models, in order to keep the system less dynamic and more quasi-static (Donzé et al., 1994; Mora and Place, 1994, 1998). The damping force, F_{iD}, is

$$F_{iD} = -\nu \Delta \dot{x}_{ij} \tag{6}$$

Here ν is the dynamic viscosity, and $\Delta \dot{x}_{ij}$ is the relative velocity between elements.

In order to reduce the cost of running a model, shear force caused by relative slip between elements is not considered within this technique as if the rock mass is frictionless (e.g., Donzé et al., 1994; Mora and Place, 1994, 1998; Hardy and Finch, 2007). Mora and Place (1994) successfully simulated the frictional stick-slip instability in a rock assemblage without shear force. Also, Finch et al. (2003, 2004) simulated normal faulting in mechanical stratigraphy above a basement structure. Previous studies suggested that realistic crustal deformation can be successfully simulated in frictionless rock mass. Therefore, the total force that an element in the upper crust experiences is

$$F_{iU}^{Total} = \sum_{j=1,n} F_{ijU} + F_{iB} + F_{iG} + F_{iD} \tag{7}$$

And the total force that an element in the lower crust experiences is

$$F_{iL}^{Total} = \sum_{j=1,n} \left(F_{ijL}^{elastic} + F_{ijL}^{viscous}\right) + F_{iB} + F_{iG} + F_{iD} \tag{8}$$

where n is the number of neighbors.

The boundary condition is implemented by imposing an external extension on all elements in the y-component direction to simulate movement of a rigid boundary wall while the opposite wall is static (Fig. 1a). The total run time is subdivided into numerous time steps, with each time step corresponding to a small increment. At each time step, elements are moving to new locations in the extension direction determined by equations of motion following Newtonian physics (Hardy and Finch, 2006). The new location of elements is:

$$Y_i(t+1) = Y_i(t) + \Delta Y\left(\frac{Y_i(t)}{Y_{\max}(t)}\right) \tag{9}$$

$Y_i\ (t+1)$ is the element location at time step $t+1$, $Y_i\ (t)$ is the element location at time step t, ΔY is the extension increment per time step and