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A genetic link between transform and hyper-extended margins

Laetitia Le Pourhiet [a,*], Dave A. May [b], Lucas Huille [a], Louise Watremez [a], Sylvie Leroy [a]

[a] Sorbonne Universités, UPMC Univ Paris 06, CNRS, Institut des Sciences de la Terre de Paris (iSTeP), 4 place Jussieu, 75005 Paris, France
[b] Department of Earth Sciences, University of Oxford, Oxford, United Kingdom

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ABSTRACT

The similarity between the geometry of the West African and South American coastlines is among one of the strongest natural observations supporting the plate tectonic paradigm. However, using classical plate tectonic approaches to model these conjugate transform margins results in a high degree of variability in palaeogeographic reconstructions. Using state-of-the-art 3D coupled thermo-mechanical numerical models, we simulate for the first time, crustal deformation at the onset of oceanisation along large offset oblique margins. Our models show that obliquity causes oceanic rift propagation to stall, resulting in an apparent polyphased tectonic evolution, and in some circumstances leads to the formation of hyper-extended margins. As a result, conjugate margins located at the edge of future fracture zones are highly asymmetric from rifting to spreading, with their lengths differing by a factor of 5 to 10, before the final phase of break-up occurs. Accounting for this discrepancy should ameliorate future palaeogeographic reconstructions.

1. Introduction

The conceptual model of a transform margin (Fig. 1) is an adaptation from the kinematic model of a transform plate boundary of Francheteau and Le Pichon (1972). In this model, the interior of the plates are assumed to be rigid, except along diverging boundary segments (dark grey regions in Fig. 1). Even in deforming areas, the direction of transport is everywhere parallel to the relative plate motion. This model was subsequently adopted by marine geologists to understand the dynamics of extremely oblique passive margins (Mascle and Blarez, 1987).

Extending this idea, Basile et al. (1993) recognised that the edge of the divergent and transform segments display different morphologies and vertical displacement history. They introduced the terms of concave and convex transform divergent segments based on the shape of the of oceanic domain (Fig. 1A). Within the rigid transform margin models, these differences are explained by the interaction of the margin with the ridge during the early stage of drifting (Fig. 1B) which causes a secondary heat pulse and uplift (Rüpke et al., 2010). This process also affects the transform margin between the stages depicted in Fig. 1B and Fig. 1C, and according to the rigid kinematic model, without resulting in any horizontal deformation.

Using the transform margin model to reconstruct plates motion is known to fail at large offsets. Attempts to geometrically match conjugate margins typically lead to isolated pockets of oceanic crust along rifted segments (Bullard et al., 1965; Rabinowitz and LaBrecque, 1979; Unternehr et al., 1988), or overlapping regions of continental blocks (Moulin et al., 2009). The failure of the classical model is most frequently explained by pre break-up deformation (Torsvik et al., 2009), or, as suggested by Vink (1982), Turner et al. (2003), by a delay in the break-up along the oblique segment. Both explanations clearly indicate that tectonic plates deviate from the rigid kinematic block theory outlined in Fig. 1.

Paleogeographic reconstructions can account for distributed strain, but they require quantitative constraints on the rate of deformation, and the obliquity of the structures which accommodate the strain. Recently, high-resolution two-dimensional dynamic models of break-up have been utilised to better understand the syn-rift deformation and provide constraints for paleogeographic reconstructions (Brune et al., 2014, 2016). Nevertheless, the conclusions drawn from such models need to be demonstrated in oblique rifting context. Modelling oblique setting naturally mandates the usage of three-dimensional dynamic models.

To date, two classes of 3D rift models have been considered. The first class have been designed to focus on the localisation of the deformation within large oblique weak zone (Tron and Brun, 1991; Mart and Dauteuil, 1999; McClay, 1990; van Wijk, 2005; Brune, 2014; Heine and Brune, 2014), whilst the second class focus on the strain localisation within the transfer zone between two

* Corresponding author.
E-mail address: laetitia.le_pourhiet@upmc.fr (L. Le Pourhiet).

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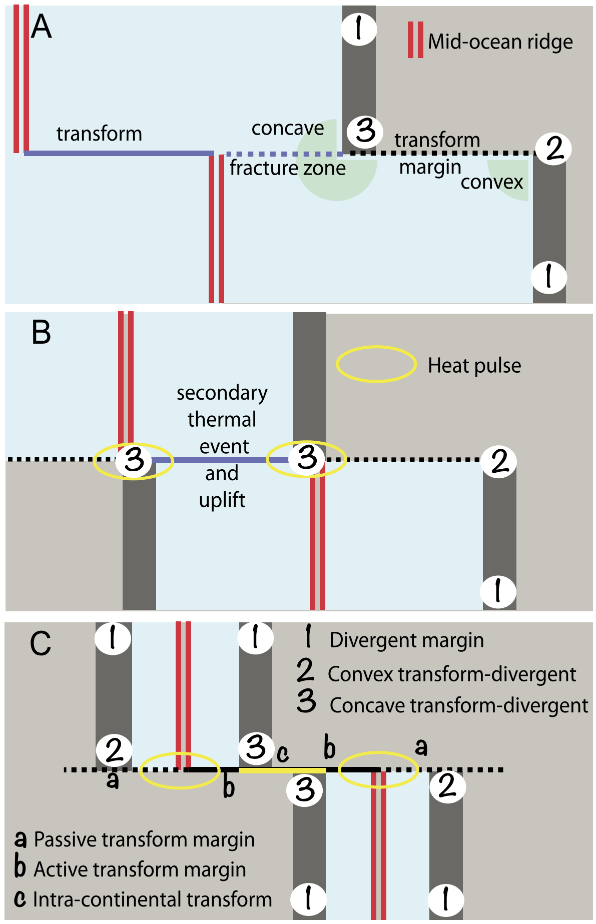

Fig. 1. Illustration of the kinematic model of transform margin formation. Modified from Francheteau and Le Pichon (1972) and Basile et al. (1993). Light grey represents normal thickness continental crust, dark grey regions represent the thinned continental crust and light blue regions indicate oceanic crust. A) Post rift stage illustrating the location of concave and convex transform divergent margins; B) Latest stage of transform margin activity showing that the mid-ocean ridge interacts thermally with the concave transform margin, long after break-up; C) Formation of the transform margin. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

en-echelon grabens by using two offset weak zones (Vendeville and Le Calvez, 1995; Gerya, 2010; Allken et al., 2011, 2012, 2013; Liao and Gerya, 2015).

Performing dynamic numerical models of large offset (100's km), magma poor, oblique passive margins such as Rio Muni in West Africa (Turner et al., 2003), Alula Fartak in the Gulf of Aden (Leroy et al., 2012), or the Ghana transform margin (Basile et al., 1993), requires three-dimensional experiments which include: full thermo-mechanical coupling, large deformation spanning at least 30 Myr of evolution and sufficiently high spatial resolution to capture accurately the deformation across a 300 km × 200 km transfer zone.

Due to technical limitations, none of the previous 3D numerical studies could simultaneously satisfy all of these requirements. Time scales of evolution rarely exceeded 5 Myr (Choi et al., 2008; Gerya, 2010, 2013; Allken et al., 2011, 2012, 2013), and the offset across the diverging segments reached, at maximum 50 km, within a model domain (in map view) of 100 km × 100 km (Choi et al., 2008; Gerya, 2010, 2013; Allken et al., 2012; Liao and Gerya, 2015). Heine and Brune (2014) modelled oblique rifting at the scale of the equatorial Atlantic, however their models did not have sufficient spatial resolution to resolve fault patterns in the crust. In this work, we utilise a newly developed, state-of-the-art massively parallel computational framework for studying 3D lithospheric deformation to surmount all of the aforementioned technical challenges required to simulate models of oblique passive margins.

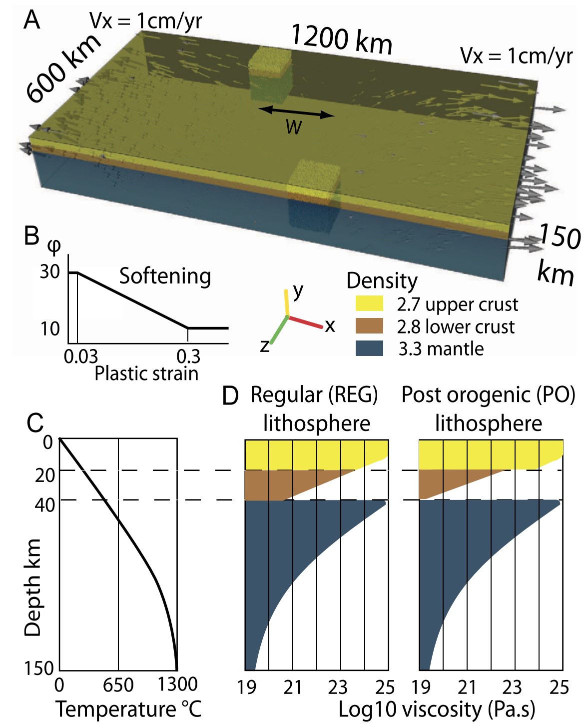

Fig. 2. Model definition. A) 3D domain with extensional boundary conditions and the outline of pre-existing weak zones offset by a distance W; B) Rheological softening law in the crust; C) Initial geotherm; D) Viscosity profile for the coupled (REG) and decoupled (PO) lithosphere obtained using the initial geotherm.

2. Method

2.1. Model set-up

The original kinematic model (Mascle and Blarez, 1987) supports that transform margins are expected form at bridging rift segments, therefore we adopted the model design which includes two offset weak zones. We consider a lithosphere scale model domain (Fig. 2A) representing 600 × 150 × 1200 km^3. Apart from the dimensions of the physical domain, the model configuration is similar to other models designed to simulate strain localisation of oblique transfer zones (Vendeville and Le Calvez, 1995; Gerya, 2010; Allken et al., 2011, 2012, 2013; Liao and Gerya, 2015) and incorporates a minimum amount of complexity.

All models consist of an initially horizontally layered hypothetical lithosphere which includes mantle, lower crust and upper crust (Fig. 2A). The Moho is originally located at 40 km depth and the boundary between lower and upper crust is located at 20 km depth. To simulate extensional systems, we impose a normal velocity with magnitude 1 cm/yr, along each face whose normal points in the x direction. This magnitude of velocity is quite standard for modelling continental extension and was previously used by Watremez et al. (2013) to model the opening of the Gulf of Aden, a young example of oblique continental extension. Material leaving the domain across these faces is compensated by a constant infilling normal velocity (2.5 mm/yr) at the base of the models. Both the inflow and outflow boundaries prescribed that the shear stress is zero. Zero normal velocity and shear stress (free slip) are prescribed on the domain faces with normal pointing in the