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Topographic asymmetry of the South Atlantic from global models of mantle flow and lithospheric stretching

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ABSTRACT

The relief of the South Atlantic is characterized by elevated passive continental margins along southern Africa and eastern Brazil, and by the bathymetric asymmetry of the southern oceanic basin where the western flank is much deeper than the eastern flank. We investigate the origin of these topographic features in the present and over time since the Jurassic with a model of global mantle flow and lithospheric deformation. The model progressively assimilates plate kinematics, plate boundaries and lithospheric age derived from global tectonic reconstructions with deforming plates, and predicts the evolution of mantle temperature, continental crustal thickness, long-wavelength dynamic topography, and isostatic topography. Mantle viscosity and the kinematics of the opening of the South Atlantic are adjustable parameters in thirteen model cases. Model predictions are compared to observables both for the present-day and in the past. Present-day predictions are compared to topography, mantle tomography, and an estimate of residual topography. Predictions for the past are compared to tectonic subsidence from backstripped borehole data along the South American passive margin, and to dynamic uplift as constrained by thermochronology in southern Africa. Comparison between model predictions and observations suggests that the first-order features of the topography of the South Atlantic are due to long-wavelength dynamic topography, rather than to asthenospheric processes. The uplift of southern Africa is best reproduced with a lower mantle that is at least 40 times more viscous than the upper mantle.

1. Introduction

The South Atlantic Ocean exhibits asymmetric bathymetry and topography. The bathymetric asymmetry (Hayes, 1988; Marty and Cazenave, 1989) is particularly pronounced in the southern South Atlantic (Morgan and Smith, 1992), with the deep Argentine Basin on the western flank (*e.g.* Shephard et al., 2012) and a shallow eastern flank, continuous with the anomalously elevated African Plateau (Nyblade and Robinson, 1994). Continental crust is ∼30 km thick in the hinterland of both conjugate margins, yet it is covered by ∼1.5 km of post-rift sediments on the South American side while being eroded on the African side (Blaich et al., 2009). The topographic asymmetry of the South Atlantic is characterized by the southern African and Brazilian plateaus in the hinterland of two segments of elevated passive continental margin (*e.g.* Japsen et al., 2011).

Several models have been proposed to explain the observed bathymetric asymmetry of the South Atlantic Ocean that constitutes a deviation from the relationship between depth and age of ocean floor (*e.g.* McKenzie, 1967). Morgan and Smith (1992) proposed that asthenospheric flow could explain the flattening of old ocean floor, and attributed the asymmetry to differences in asthenospheric flow due to the asymmetric migration rates of the South American and African plates with respect to the deep mantle. In contrast, Doglioni et al. (2003) proposed that a general westward motion of the lithosphere with respect to the mantle results in an eastward flow of the asthenosphere, such that the eastern rift flank could be underlain by depleted, lighter asthenosphere, resulting in shallower bathymetry. Although quite different, these models both propose a shallow mantle origin for the topographic asymmetry of the South Atlantic; they are also exclusively inspired by and compared to present-day observations. Neither of these models address the elevated passive continental margins in South Africa and in Brazil nor the evolution of the region.

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Here, we investigate the effect of mantle convection on the evolution of the topography of the South Atlantic. We introduce Earth deformation models that consider both mantle flow and lithospheric deformation. These models are forward in time and progressively assimilate space- and time-dependent boundary conditions derived from global tectonic reconstructions with continuously closing and deforming plates. They predict the evolution of mantle temperature, thickness of the continental crust, dynamic topography and isostatic topography. This global approach makes it possible to investigate the consequences of a given tectonic reconstruction for the dynamic evolution of a passive margin at continental scale and over tens of millions of year, including the post-rift phase. We implement two alternative kinematic models for the opening of the South Atlantic Ocean (Torsvik et al., 2009; Heine et al., 2013) and compare the predicted crustal thickness for South America, where results are most contrasted, to continent-scale Moho maps (Lloyd et al., 2010; Assumpção et al., in press; Chulick et al., 2013). The models are also compared to present-day topography and bathymetry, and to an estimate of residual topography (Winterbourne et al., 2009) for the Atlantic. Furthermore, the predicted subsidence history is quantitatively compared to borehole tectonic subsidence, while the uplift history of southern Africa is qualitatively compared to the unroofing history constrained by thermochronology (Flowers and Schoene, 2010; Zhang et al., 2012).

2. A numerical model of mantle flow and lithospheric deformation

We introduce a new generation of forward models of mantle flow and lithospheric deformation with progressive data assimilation. We first present the equations solved for mantle convection and shearing of the lithosphere, and then describe the methods we use to obtain compositionally-distinct continents in global mantle flow models, and to implement deforming areas in global plate reconstructions.

2.1. Governing equations

We solve the finite-element problem of thermochemical convection within the Earth's mantle under the Boussinesq approximation (*e.g.* Spiegel and Veronis, 1960) using *CitcomS* (*e.g.* Zhong et al., 2008) modified to allow assimilation of data derived from global plate reconstructions.

The conservation of mass is expressed as

$$\nabla \cdot \mathbf{u} = 0 \tag{1}$$

where $\mathbf{u}$ is the velocity. The conservation of momentum is expressed as

$$-\nabla p + \nabla \cdot \left[\eta\left(\nabla \mathbf{u} + \nabla^{\top} \mathbf{u}\right)\right] = (-\alpha T + \delta\rho_{ch} C + \delta\rho_{ph} \Gamma) Ra\, \bar{g}\, \hat{\mathbf{r}} \tag{2}$$

where p is the dynamic pressure, η the mantle viscosity, $\top$ the transpose, α the coefficient of thermal expansivity, T the temperature, $\delta\rho_{ch}$ the chemical density, C the composition, $\delta\rho_{ph}$ the density difference across a phase change, $\bar{g}$ a depth-dependent, non-dimensional acceleration of the gravity field, $\hat{\mathbf{r}}$ the radial unit vector and Ra is the Rayleigh number defined below. $\delta\rho_{ch}$ is non-dimensionalized and described in terms of the buoyancy ratio

$$B = \delta\rho_{ch} / \alpha_0 \rho_0 \Delta T \tag{3}$$

with ρ_0 the density and ΔT the superadiabatic temperature change across the mantle. Parameters are listed in Table 1. The phase function (*e.g.* Richter, 1973) is

Table 1
Generic model parameters. Subscript "0" indicates reference values.

Parameter	Symbol	Value	Units
Rayleigh number	Ra	7.8×10^8	–
Thermal expansion coefficient	α_0	3×10^{-5}	K^{-1}
Density	ρ_0	4000	$kg\,m^{-3}$
Gravity acceleration	g_0	9.81	$m\,s^{-2}$
Temperature change	ΔT	1912 or 2412	K
Mantle thickness	h_M	2867	km
Thermal diffusivity	κ_0	1×10^{-6}	$m^2\,s^{-1}$
Viscosity	η_0	1×10^{21}	Pa s
Chemical density (Archean lithosphere)	$\delta\rho_{ch\,arch}$	−1.74	%
Chemical density (Proterozoic lithosphere)	$\delta\rho_{ch\,prot}$	−1.52	%
Chemical density (Phanerozoic lithosphere)	$\delta\rho_{ch\,phan}$	−1.37	%
Earth radius	R_0	6371	km
Activation energy (upper mantle)	$E_{\eta UM}$	100	$kJ\,mol^{-1}$
Activation energy (lower mantle)	$E_{\eta LM}$	33	$kJ\,mol^{-1}$
Activation temperature	T_η	452	K
Compositional viscosity pre-factor	η_C	100	–
Background mantle temperature	T_b	1685	K

$$\Gamma = 1/2\left[1 + \tanh\left(\frac{\rho\bar{g}(1 - r - d_{ph}) - \gamma_{ph}(T - T_{ph})}{\rho g w_{ph}}\right)\right] \tag{4}$$

where r is the depth, d_{ph} and T_{ph} are the ambient depth and temperature of a phase change, γ_{ph} is the Clapeyron slope of a phase change, and w_{ph} is the width of the phase transition. In order to investigate the effect on dynamic topography of a phase change at $d_{ph} = 670$ km (case TC6, Table 2), we use $\delta\rho_{ph} = 7\%$ and $\gamma_{ph} = -2$ MPa K^{-1} (Billen, 2008), $w_{ph} = 40$ km and $T_{ph} = 1667$ K.

The Rayleigh number Ra is defined by

$$Ra = \frac{\alpha_0 \rho_0 g_0 \Delta T h_M^3}{\kappa_0 \eta_0} \tag{5}$$

where h_M is the thickness of the mantle, η_0 the reference viscosity, and κ_0 the thermal diffusivity.

The conservation of energy is

$$c_p \frac{\partial T}{\partial t} = -c_p\, \mathbf{u} \cdot \nabla T + \nabla \cdot (c_p \kappa \nabla T) \tag{6}$$

where c_p is the heat capacity and t is time. The equation for advection of the composition field is

$$\frac{\partial C}{\partial t} + (\mathbf{u} \cdot \nabla) C = 0. \tag{7}$$

The composition field, C, is represented by tracers that are advected using a predictor-corrector scheme (McNamara and Zhong, 2004). The composition field is determined using the ratio method (Tackley and King, 2003), modified to give $C = 0$ if an element contains no tracers. This modification allows us to considerably limit the total number of tracers required to track compositionally distinct material in the uppermost and lowermost mantle.

Eqs. (1)–(7) are routinely used to solve the problem of thermochemical mantle convection in a viscous shell. They can also be used to solve for heat advection and diffusion upon stretching of the lithosphere (*e.g.* McKenzie, 1978; Jarvis and McKenzie, 1980), allowing us to solve both problems simultaneously given sufficient numerical resolution.

2.2. Data assimilation

Resolving the fully-dynamic, time-dependent problem of thermochemical mantle convection requires a resolution of the order