



HAL
open science

Numerical models for continental break-up: Implications for the South Atlantic

A Beniést, A Koptev, Evgenii Burov

► To cite this version:

A Beniést, A Koptev, Evgenii Burov. Numerical models for continental break-up: Implications for the South Atlantic. *Earth and Planetary Science Letters*, 2017, 461, pp.176-189. 10.1016/j.epsl.2016.12.034 . hal-01467320

HAL Id: hal-01467320

<https://hal.sorbonne-universite.fr/hal-01467320>

Submitted on 14 Feb 2017

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 **Numerical models for continental break-up: implications for the South Atlantic**

2
3 Beniest, A.^{1,2}, Koptev, A.¹ Burov, E.^{1†}

4 Email corresponding author: a.beniest@gmail.com

5 1. Sorbonne Universités, UPMC University Paris 06, CNRS, Institut des Sciences de la Terre de
6 Paris (iSTeP), 4 Place Jussieu 75005 Paris, France

7 2. IFP Energies nouvelles, Geosciences, 1&4 Avenue du Bois-Préau 92852, Rueil-Malmaison,
8 France

9
10 † Deceased 9 October 2015

11
12 **Abstract**

13 We propose a mechanism that explains in one unified framework the presence of continental break-up
14 features such as failed rift arms and high-velocity and high-density bodies that occur along the South
15 Atlantic rifted continental margins. We used 2D and 3D numerical models to investigate the impact of
16 thermo-rheological structure of the continental lithosphere and initial plume position on continental rifting
17 and breakup processes. 2D experiments show that break-up can be 1) “centred”, mantle plume-induced and
18 directly located above the centre of the mantle anomaly, 2) “shifted”, mantle plume-induced and 50 to 250
19 km shifted from the initial plume location or 3) “distant”, self-induced due to convection and/or slab-
20 subduction/delamination and 300 to 800 km off-set from the original plume location. With a 3D, perfectly
21 symmetrical and laterally homogenous setup, the location of continental break-up can be shifted hundreds
22 of km’s from the initial position of the mantle anomaly. We demonstrate that in case of shifted or distant
23 continental break-up with respect to the original plume location, multiple features can be explained. Its
24 deep-seated source can remain below the continent at one or both sides of the newly-formed ocean. This
25 mantle material, glued underneath the margins at lower crustal levels, resembles the geometry and location
26 of high velocity/high density bodies observed along the South Atlantic conjugate margins. Impingement of

27 vertically up-welled plume material on the base of the lithosphere results in pre-break-up topography
28 variations that are located just above this initial anomaly impingement. This can be interpreted as aborted
29 rift features that are also observed along the rifted margins. When extension continues after continental
30 break-up, high strain rates can relocalize. This relocation has been so far attributed to rift jumps. Most
31 importantly, this study shows that there is not one, single rift mode for plume-induced crustal break-up.

32

33 **1. Introduction**

34 Over the last decades a large variety of rift features have been recognised and explained with different
35 methods and different concepts. These features include for example aborted rift structures, anomalous
36 topography or anomalously high velocity/high density bodies located in the lower crust. Explanations for
37 anomalous features often link one mechanism with one observed rift feature. For example, on plume
38 impingement, a stratified lithospheric rheology (e.g. D'Acremont et al., 2003; Burov et al., 2007;) would
39 result in topographic uplift, as has been modelled with thermo-mechanical modelling. Forward modelling
40 shows that magmatic underplating can cause topographic variations (Hirsch et al., 2010). Anomalously high
41 velocity/high density bodies have been observed on tomographic images below the continents, implying
42 that in some regions magmatic processes dominate rifting (Cornwell et al., 2006). The latter is also suggested
43 by gravity modelling that revealed the presence of anomalously high-density bodies in e.g. the South
44 Atlantic domain, implying that volcanic activity played a key role in margin development (Blaich et al.,
45 2011; Maystrenko et al., 2013).

46 Review papers combine all these studies on one specific topic. Examples are the role of the Moho in
47 extensional settings (Cloetingh et al., 2013), the effect of volcanism in rifting and continental break-up
48 (Franke, 2013) or the dynamic processes that control rifting (Ziegler and Cloetingh, 2004).

49 With this study we demonstrate how one break-up mechanism can explain a multitude of features. We use
50 the South Atlantic break-up as our case study for plume-induced continental break-up. Since the South
51 Atlantic developed diachronously and it is a very complex region requiring a 3D approach, we have not the
52 intention to reproduce the South Atlantic evolution as such, including along-axis northward break-up

53 propagation to close to the plume (Franke, 2013), but rather to address general observations on continental
54 break-up. Our fully coupled lithospheric-grade 2D and 3D models have an explicit elasto-visco-plastic
55 rheology that accounts for realistic deformation of the lithosphere and a slip free surface that can calculate
56 vertical motions. The 2D model has proven to be very useful to investigate plume-lithosphere interactions
57 (e.g. d’Acremont et al., 2003; Burov et al., 2007). We take it one step further by developing one scenario to
58 explain multiple anomalous features, such as high-velocity/high-density bodies and anomalous topographic
59 variations with one single model. The 3D model is used to include the lateral component in a very simple,
60 completely lateral homogeneous setting (Koptev et al., 2016).

61

62 **2. Geological and geophysical setting**

63 *2.1. South Atlantic opening*

64 Initial extension between Africa and South America was accommodated along a former fold-and-thrust belt
65 (present-day location see Fig. 1), known as the Gondwana Fold Belt (GFB) or the Cape Fold Belt (CFB).
66 This fold-and-thrust-belt was reactivated during the Early Mesozoic as a strike-slip system before the
67 opening of the South Atlantic (Cobbold et al., 1992). During this reactivation it weakened the South
68 American plate prior to the development of the Atlantic Mid-Oceanic Ridge, forming a first set of
69 extensional basins (Autin et al., 2013), their axes oriented obliquely to the present-day orientation of the
70 spreading centre (Fig. 1). Several extensional pulses caused the opening of the South Atlantic between 134
71 Ma and 113 Ma (e.g. Torsvik et al., 2009; Moulin et al., 2010). Voluminous volcanic activity, recognised
72 on seismic reflection profiles as ‘Seaward Dipping Reflectors’ (SDR’s) in the form of aerial flood basalts
73 (extrusive) and/or underplating (intrusive) accompanied an episode of extension that created the South
74 Segment (Fig. 1), starting between 134 Ma and 132 Ma (Moulin et al., 2010). Another pulse contributed to
75 the formation of the Central Segment, starting around 112 Ma (Moulin et al., 2010) and is marked by
76 massive salt deposits that have not been found along the margins of the South Segment (Fig. 1, Torsvik et
77 al., 2009). Only minor volcanic activity has been recorded in this segment as the typical SDR’s are mostly
78 absent, except just north of the Rio Grande Rise (Franke, 2013). The opening of the South Atlantic and

79 formation of the Mid-Atlantic Ridge is considered to be due to a combination of passive far-field forces
80 (Husson et al., 2012) and the presence of different hotspots (Torsvik et al., 2009). A major far-field stress
81 component that enhanced the growth of the South Atlantic domain during the Mesozoic is the subducting
82 and ‘pulling’ Nazca plate to the west of the South American continent, which also resulted in the faster
83 west-ward migration of the South American plate with respect to the almost stationary African plate (e.g.
84 Husson et al., 2012). The South-African super plume rises from the core-mantle boundary (CMB) to below
85 the mantle transition zone (Hassan et al., 2015) which is reflected in present-day topography by a
86 “superswell” at the margins of the south-west African continent (Nyblade and Robinson, 1994; Davaille et
87 al., 2005). As shown by a.o. Lithgow-Bertelloni and Silver (1998), this excess of topography elevation is
88 dynamically supported by upwelling flow of buoyant material through the mantle. From this large-scale,
89 lower mantle low-velocity anomaly, the hotspots and their tracks (e.g. the Bouvet (Meteor), the Trinidad
90 and St Helena Hotspots (Torsvik et al., 2006,2009)) and the only deep rooted Tristan plume (Fig. 1, Torsvik
91 et al. (2009) and references therein) might develop over a long period of time (~200 Myr).

92

93 *2.2. Lithosphere structure margins*

94 The selected profile for our 2D model connects the offshore southwest Africa Orange Basin and its
95 conjugate with the Colorado basin on the South American side (pink line, Fig. 1). The Tristan hotspot lies
96 actually in the middle of the two transects (Fig. 1). The present-day crustal and lithosphere structure of these
97 margins is constrained by combining published work on deep seismic refraction data, tomography, gravity
98 and magnetic studies (Fig. 2).

99 On the African side of the transect (Fig. 2a) the lithosphere thickness ranges from 120 km below the oceanic
100 crust to 200 km below the continent (Fishwick, 2010). With gravity modelling and seismic interpretation
101 the Moho-depth has been estimated to be less than 10 km below the oceanic crust of the Orange basin and
102 over 40 km below the continent (Maystrenko et al., 2013). Even though crustal movements have been
103 observed in central Africa, the southern African plate is considered relatively stable with a strong rheology
104 (Heine et al., 2013).

105 On the South American side the lithosphere-asthenosphere boundary (LAB) reaches depths of 160km below
106 the stable continent in Eastern Brazil and 120 km in below the continent in Central Argentina (Heit et al.,
107 2007). The Moho depth varies between 70 km below the plateau in the Andean orogeny to 25-35 km below
108 the flat continent (Van Der Meijde et al., 2013). For the Colorado basin specifically, deep refraction seismic
109 studies reveal a crustal thickness of the margins of 30 km (Franke et al., 2006).

110 We assume that before continental break-up, the lithosphere thickness of the South American plate was
111 similar to that of the African plate. However, the South American plate underwent an earlier deformation
112 phase prior to the formation of the South Atlantic domain (Autin et al., 2013). Extensional deformation does
113 result in lithospheric thinning and weakening (Ziegler and Cloetingh, 2004). We, therefore, adopt a weaker
114 strength compared to the African plate and a thickness of 180 km, which is the mean between the 200 km
115 of the African lithosphere and the present-day 160 km South American lithosphere, to account for this earlier
116 deformation phase.

117

118 *2.3. High velocity/high density bodies and aborted rift structures*

119 Along the South Atlantic conjugate margins high velocity/high density bodies have been described at lower
120 crustal depths below the continent and the margin (Fig. 1) using seismic data and gravity modelling.
121 Anomalous gravity and velocity bodies have been noted in the Central segment on the African side from
122 Gabon (Dupré et al., 2007) to the Lower Congo (Contrucci et al., 2004) and the Kwanza Basin (Blaich et
123 al., 2011) and on the South American side from the Sergipe-Alagoas Basin (Mohriak et al., 2000), to the
124 Camamu-Atmada basin (Blaich et al., 2011) and the Santos basin (Blaich et al., 2011). In the South segment
125 on the African side these bodies have been observed in the Walvis Basin (Blaich et al., 2011) and the Orange
126 Basin (Dressel et al., 2015) and on the South American side in the Colorado basin, along the Uruguayan
127 margin (Clerc et al., 2015) and in the deep Argentina Basin (Franke et al., 2006; Schnabel et al., 2008).
128 These bodies differ from the seaward-dipping reflectors (SDR) as they are situated at the base of the
129 lithosphere or at lower crustal levels and do not necessarily have a magmatic origin. They could be
130 serpentinized mantle or mafic and ultramafic crustal rocks (Fig. 2, Blaich et al., 2011).

131 Graben structures or aborted rift structures onshore along the whole South American margin of the South
132 Atlantic domain (Burke, 1976), are located near the anomalously high velocity/high density bodies. In the
133 South Segment, graben structures and failed rift structures are less-abundant along the African margin,
134 where they appear mainly along the south South-African margin and in the Central segment along the Gabon
135 and Benin margins. On the South American side of the South Segment, the basins oriented perpendicular to
136 the present-day ridge extend onshore as aborted rift features (Burke, 1976). Another failed rift feature is
137 observed in the southwestern part of the Santos Basin, where the now aborted Abimael ridge is located
138 parallel to the present-day Mid Oceanic Ridge (Heine et al., 2013).

139

140 **3. Model setup**

141 The 2D thermo-mechanical numerical code FLAMAR, based on the FLAC-Para(o)voz algorithm (Cundall,
142 1989; Poliakov et al., 1993) has been used to investigate the effect of plume location on continental break-
143 up using the South-Atlantic as an example of a fully developed rift-to-spreading system. We built our case
144 on the continuation of earlier parametric studies on the rheology of the lithosphere and plume-continental
145 lithosphere interactions (D’Acremont et al., 2003; Burov et al., 2007). Where needed, we adjust the
146 parameters according to the geological and geophysical evidence described above. A symmetric simulation
147 that is not area-specific are carried out with the 3D viscous-plastic numerical code 3DELVIS (Gerya and
148 Yuen, 2007). All mechanical and thermo-rheological parameters are listed in Table 1. We have performed
149 a series of 36 experiments. Controlling parameters and principal results are summarized in Table 2.

150

151 *3.1. 2D numerical model*

152 The FLAMAR code has been updated and modified over the last 20 years (Burov and Diament, 1995; Burov
153 and Poliakov, 2001; Le Pourhiet et al., 2004; Yamato et al., 2008). For the sake of coherency with previously
154 published papers we only describe the main features and essentials of the model used for this study. Detailed
155 descriptions of the code can be found in studies that have tested the code for many different geological cases
156 (D’Acremont et al., 2003; Le Pourhiet et al., 2004; Yamato et al., 2008). FLAMAR is a fully explicit, finite

157 element/finite difference code based on a Cartesian coordinate frame. It has a 2D strain formulation with a
 158 Lagrangian mesh that consists of quadrilateral elements consisting of two couples of triangular sub-elements
 159 containing tri-linear shape functions. It uses a large-strain, time-marching scheme. The code solves for full
 160 Newtonian equations of motions in a continuum mechanics approximation (3.1)

$$161 \quad \langle \rho \ddot{\mathbf{u}} \rangle - \nabla \sigma - \rho \mathbf{g} = 0 \quad (3.1)$$

162 where ρ , $\ddot{\mathbf{u}}$, σ and \mathbf{g} stands for density, acceleration of the object, stress and acceleration due to body forces
 163 or gravity, respectively.

164 It is coupled with constitutive laws (3.2) to quantify viscous, elastic and plastic characteristics by the heat
 165 transfer equation (3.3), where the heat advection term ($\dot{\mathbf{u}} \nabla T$) is included in the Lagrangian derivative
 166 (DT/Dt). Erosion and sedimentation is accounted for using a linear diffusion equation assuming
 167 conservation of mass (3.4).

$$168 \quad \frac{D\sigma}{Dt} = F(\sigma, \mathbf{u}, \dot{\mathbf{u}}, \nabla \dot{\mathbf{u}}, T) \quad (3.2)$$

$$169 \quad \rho C_p DT/Dt - k \nabla^2 T - \sum_i^n H_i = 0; \quad \rho = f(P, T) \quad (3.3)$$

$$170 \quad \frac{dh}{dt} = a \nabla^2 h \quad (3.4)$$

171 In this case, t stands for time, \mathbf{u} is the displacement vector, and T is temperature. The heat transfer equation
 172 relies on C_p for the specific heat, k for thermal conductivity respectively and H for the internal heat
 173 production, including radiogenic heat and frictional heat dissipation. P stands for pressure that become
 174 negative for compression. The linear diffusion equation uses a constant a and the height or thickness of the
 175 sediments h .

176 The code is capable of calculating realistic visco-elasto-plastic rheologies explicitly. Pressure-dependent
 177 deformation is maintained through the Mohr-Coulomb criteria for the plastic regime and the non-linear
 178 viscous flow law at depth. The free surface upper boundary condition calculates high-resolution topographic
 179 changes due to deformation of the lithosphere.

180

181 *3.1.1. Model geometries*

182 The model box is 2000 km wide and 400 km deep. The grid size is 400 x 80 elements, resulting in a
183 resolution of 5 km x 5 km per element. We have tested three different lithospheric setups with diverse
184 complexities (see 3.1.2, Fig. 3) and three different locations of a 1700°C thermal and compositional mantle
185 anomaly at 400 km depth (D’Acremont et al., 2003). The initial locations vary laterally at the base of the
186 model with the center of the anomaly positioned 1) at the centre of the model (i.e. plume location at 1000
187 km, see Table 2), 2) at 200 km to the right of the model box’s centre (i.e. plume location at 1200 km) and
188 3) at 200 km to the left of the model box’s centre (i.e. plume location at 800 km). Each location is tested in
189 a separate calculation. Following previous studies the base of the anomaly lies at 400 km depth as the deeper
190 mantle phase does not have a large impact on the crustal evolution (D’Acremont et al., 2003; Ribe and
191 Christensen, 1994). The anomaly has a simplified, symmetric, spherical shape since at depth viscous bodies
192 take a spherical shape and this follows the line of 2D and 3D numerical modelling experiments on plume-
193 lithosphere interaction (a.o. D’Acremont et al., 2003; Burov and Gerya, 2014; Koptev et al., 2016). In most
194 of the experiments, it has a diameter of 230 km. The effects of a mantle anomaly with a diameter of 100 km
195 were tested in a limited number of models. The composition of the mantle anomaly is olivine with a density
196 of 3250 kg/m³ (except for several models where it is 3310 kg/m³) which has been determined to be an
197 intermediate plume in previous studies (Turcotte and Schubert, 2002; D’Acremont et al., 2003). No
198 background density tests have been performed as the background density used for background calculations
199 is the same for the plume as well as the surrounding mantle. The thermal contrast between the plume and
200 the mantle varies as thermal exchanges happen between the plume and the mantle, decreasing the
201 temperature of the plume. The mantle also cools as the plume rises to shallower depths.

202

203 *3.1.2. Density and rheological structure*

204 The 2D model consists of four horizontal rheological layers. For Setup 1 (Fig. 3a and Fig. 3b), a laterally
205 homogeneous 40 km thick two-layered crust and a 160 km thick lithospheric mantle have been applied. We
206 test the model’s sensitivity for two different rheological properties of the crust. We use a “weaker”
207 rheological strength envelope (Setup 1a; Fig. 3a), composed of 1) a wet quartz upper crust with a density of

208 2500 kg/m³ and 2) a diabase lower crust with a density of 2750 kg/m³. Our second rheological strength
209 envelope has the characteristics of a “strong”, cratonic crust that consists of: 1) a dry quartz upper crust with
210 a density of 2600 kg/m³ and 2) a strong diabase lower crust with a density of 2850 kg/m³. The rheological
211 differences of the two strength envelopes represent a “weaker” crust that has been subject to an earlier
212 deformation phase, before the opening of the South Atlantic, which is the case for the South American side
213 (Autin et al., 2013), and a “stronger” crust of cratonic nature that represents the stable southern African
214 continent(after Burov and Diament, 1995). Dry olivine flow law has been assumed for both lithospheric and
215 sub-lithospheric mantle in all our experiments. The initial density of the mantle decreases from 3330 kg/m³
216 to 3310 kg/m³ at the lithosphere-asthenosphere boundary. Specific values of the rheological parameters used
217 are given in Table 1.

218 For Setups 2 and 3 we apply a laterally non-homogeneous crustal rheology: a “weak” crustal rheology for
219 the left half of the model and a “strong” crustal rheology for the right one (Fig. 3c-f). The crustal and
220 lithospheric thicknesses are laterally homogeneous in Setup 2a: 20 km for upper crust, 20 km for lower crust
221 and 160 km for lithospheric mantle (Fig. 3c). Setups 2b and 3 are characterized by laterally varying
222 lithospheric layers, based on the lithospheric scale structure described in section 2.2: the “weaker” left half
223 has a 15 km-thick upper crust, 15 km-thick lower crust and a 150 km-thick lithospheric mantle, whereas the
224 “stronger” right half has a 20 km-thick upper crust, a 20-km thick lower crust and a 160 km-thick
225 lithospheric mantle (Fig. 3c-f). Three different contacts between the rheological strengths are tested. Setup
226 2b has a straight vertical contact. For Setup 3 we have adopted a geometry resembling the old suture zone
227 that is reactivated during the first extensional phase. The suture is dipping either 40 degrees towards the
228 ‘strong’, African rheology (Setup 3a) or towards the ‘weak’ South American rheology (Setup 3b).

229 By the setups described above we have tested the following parameters: initial position of the plume, density
230 of the mantle plume (limited to Setup 3a) and different half-rate velocity boundary conditions (see Table 1).

231

232 *3.1.3. Mechanical and thermo-rheological boundary conditions*

233 We simulate tectonic forcing by applying a constant, time independent, extension rate along the vertical side
234 of the box of 25 mm/yr. An equal half-rate velocity is applied on both sides of the box (12.5 mm/yr) to one
235 set of models and 5 mm/yr on the right side and 20 mm/yr on the left side is applied to a second set of
236 models (Table 1). The half-rate velocities are adopted from (Müller et al., 2008). The bottom of the box is
237 defined by hydrostatic pressure with free slip in all directions. The upper side of the box is a free surface
238 boundary, implying a free stress and a free slip condition in all direction, allowing the lithosphere to develop
239 freely. A moderate erosion by diffusion is applied ($a = 500 \text{ m}^2/\text{yr}$).
240 The upper and bottom thermal boundary condition is a fixed temperature 10 °C and 1400 °C respectively to
241 represent a ‘cold’ geotherm. An old lithosphere of 500 Ma (Burov and Diament, 1995) has been assumed
242 for the tectonic age used to represent the super-continent Pangea before break-up. The geotherm used for
243 the models reaches 500 °C at Moho depth, 1330 °C at the lithosphere-asthenosphere boundary (LAB) after
244 which it becomes adiabatic until it reaches 1400 °C at the base of the model at 400 km (Ribe and Christensen,
245 1994).

246

247 *3.2. 3D numerical model*

248 A 3D model has been performed with the thermo-mechanical viscous-plastic 3DELVIS code (Gerya, 2010;
249 Gerya and Yuen, 2007) that combines the finite difference method with a marker-in-cell technique. The 3D
250 model box has the horizontal dimensions $1500 \times 1500 \times 635 \text{ km}$ and consists of $297 \times 297 \times 133$ nodes
251 offering spatial resolution of ca. $5 \times 5 \times 5 \text{ km}$ per grid cell. Not area-specific initial setup consists of a
252 stratified three-layer (upper/lower crust and lithospheric mantle) continental lithosphere underlain by an
253 asthenosphere. The total thickness of the two-layer crust is 36 km; the depth of lithosphere-asthenosphere
254 boundary is 150 km. The mantle plume has been seeded at the base of the modelled domain by a spherical
255 thermal anomaly of 370 °C with a radius of 200 km. The initial geotherm is piece-wise linear with fixed
256 temperatures at the surface (0°C), at the Moho (700°C), at the base of the lithosphere (1300°C), and at the
257 bottom of the model box (1630°C). Weak tectonic forcing has been simulated by applying a constant ultra-
258 slow divergent horizontal velocity of 3 mm/year along the sides of the model. More detailed information on

259 the 3D model setup and rheological and material properties used in our 3D experiments can be found in
260 Burov and Gerya (2014) and Koptev et al. (2015, 2016).

261

262 **4. Model results**

263 *4.1. 2D model results*

264 Three types of model scenarios result from our set of experiments. “Centred” break-up, when the mantle
265 anomaly moves vertically upwards and break-up happens directly above the original location of the centre
266 of the mantle anomaly. “Shifted” break-up, when the mantle anomaly first migrates vertically and, once it
267 reaches the base of the lithosphere, migrates laterally until break-up occurs with a 50 to 250 km offset with
268 respect to the initial anomaly position. “Distant” break-up, when a mantle anomaly rises to the base of the
269 lithosphere and remains there, while the location of break-up takes place more than 300 km away from the
270 initial site of the anomaly.

271 Experiment 6, characterized by a “strong” homogenous lithosphere, is an example of “centred” break-up
272 (Fig. 4a). The mantle anomaly reaches the base of the lithosphere rapidly within 2 Myr, after which it
273 penetrates into the lithosphere. The rising flow of plume material is strong enough to break apart the
274 overlying lithospheric mantle and crust between 7 and 8 Myr. The surface reacts by uplift, then subsidence
275 and alternating positive and negative vertical movements of the margins and the rift centre. Although the
276 initial position of break-up centre is situated directly above mantle plume, the continuous extensional
277 evolution, including strain relocation and changing temperature distribution, suggest a post-rift lateral shift
278 of the spreading axis. Note that after continental break-up mantle plume material reaches the surface where
279 it contributes to the formation of new oceanic lithosphere.

280 The “shifted” mode of continental break-up is illustrated by Experiment 12 where the mantle plume anomaly
281 has been seeded below a stronger lithosphere composing the right half of the model domain (Fig. 4b). As in
282 the case of Experiment 6, the onset of rifting starts with a rapid rise of the anomaly, impinging the
283 lithosphere not later than 2 Myr. Surface topography associated with localized crustal strain is formed
284 around 3-4 Myr with small offset (<50 km) from the point directly above mantle plume impingement.

285 Further upslope migration plume material leads to continental break-up between 7 and 10 Myr. A principal
286 difference from the “centred” experiment 6 is the lateral shift (50 to 250 km) of the newly formed spreading
287 axis with respect to the initial position of the mantle plume. Lateral migration of the plume head to this
288 break-up axis leads to an asymmetrical distribution of the plume material: some of the material reaches the
289 surface at the spreading centre, another part remains glued beneath the highly thinned lithosphere at depths
290 between 200 and 10 km. Similar to Experiment 6, the final stage of the “shifted” system development is the
291 strain relocation corresponding to 200 km-wide jump of the spreading axis.

292 Finally, experiment 16 illustrates the “distant” break-up mode that starts with a rapid uplift of the mantle
293 plume to the bottom of the lithosphere, an observation typical for all performed models. This expectedly
294 results in associated topography variations (Fig. 4c). In contrast with the two previously discussed break-up
295 modes (experiments 6 and 12), mantle plume material remains glued beneath the base of the lithosphere
296 without localized ascent towards the formed break-up centre. Lithosphere thinning that will result in break-
297 up occurs at large distance (more than 500 km) from the plume impingement. This appears to be related to
298 secondary mantle convection associated with plume-induced subduction of the lithospheric mantle that has
299 developed upon plume upwelling to the lithosphere-asthenosphere boundary on both sides of the plume
300 head. It is noteworthy that initial topographic changes created by the impingement of the plume remain
301 visible throughout the model evolution. Given the lack of near-surface plume material, this “distant” mode
302 cannot be considered as break-up directly induced by the impact of the mantle plume. Nevertheless, it might
303 reflect the implicit influence of the upwelled plume on “distant” break-up processes via plume-induced
304 subduction and mantle convection.

305 “Centred” break-up preferably takes place using initial Setup 1, where the crust and lithospheric mantle are
306 laterally homogeneous and no inherited structures are given (Fig. 4a, Table 1), but other setups can also
307 evolve according to this mode (Fig. 5a-c). Break-up occurs between 6 and 10 Myr, directly above the initial
308 location of the mantle anomaly. Mantle material reaches the surface at the point of impingement that evolves
309 into the break-up axis. Almost all plume material is involved in formation of new oceanic lithosphere. As a
310 result, after continuous (more than 10 Myr) calculations, only little material remains below the thinned

311 continental lithosphere. Note that, even though central located plumes are expected to develop to a
312 symmetric or “central” mode, a central located plume evolves the least likely into “central” type of break-
313 up (table 1). “Shifted” break-up is favoured by Setup 2a where the thickness of the lithospheric layers is
314 laterally homogeneous but crustal rheology differs (Fig. 5d-f). The mantle anomaly rises and break-up also
315 occurs between 6 and 10 Myr, but in this case it is shifted from the initial point of impingement. Most mantle
316 material remains below the lithosphere, but through migration along the bottom of the lithosphere some
317 material still reaches the surface. This mode of break-up only occurs when the lithosphere properties
318 (rheology and/or thickness) varies laterally, but it does not completely control “shifted” break-up, because
319 not all laterally varying rheology experiments result in “shifted” break-up. Plume location is also not a
320 controlling factor for the model to result in “shifted” break-up as all three plume locations can result in
321 “shifted” break-up. The “distant” break-up experiments have a preference for Setup 2 and 3, where both the
322 lithospheric layers’ thickness and the crustal rheology are laterally different (Fig. 5g-i). Crustal break-up
323 happens slightly later compared to the “centred” and “shifted” experiments: between 9 and 12 Myr. Mantle
324 anomaly material does not reach the surface, but remains completely glued to the bottom of the lithosphere.
325 Most of the continental break-up modelled with the “distant” experiments occurs in the lithospheric segment
326 that is characterized by a strong crust.

327 Almost half (12 out of 25) of the equal half-rate velocities boundary condition results in “centred” break-
328 up. More than half (6 out of 11) of the unequal half-rate velocity boundary conditions result in “distant”
329 break-up mode. The different velocity parameters do have a preference for a certain break-up style, but it is
330 not a controlling factor.

331 The models that resulted in “shifted” break-up have a mantle anomaly that rises to the base of the lithosphere
332 and upon arrival, migrates, in most cases, towards the weaker lithosphere to break through this less strong
333 segment. This is in contrast with the “distant” model results that develop crustal break-up in the stronger
334 lithosphere (11 out of 11 models) when the plume remains glued below the weaker lithosphere and does not
335 break through. In case of “distant” break-up mode, the rheology is very important and strongly controls this
336 mode of break-up.

337

338 *4.2. 3D model results*

339 Similarly to the 2D experiments, the 3D model shows a quick (<2 Myr) upwelling of the plume material up
340 to lithosphere-asthenosphere boundary (Fig. 6a). After this, the plume head starts to spread laterally within
341 lower part of lithospheric mantle (Fig. 6b-d). When the mantle plume impinges on the base of lithosphere,
342 almost all plume material is partially melted (Fig. 6a). Following spreading and cooling expectedly leads to
343 gradual solidification of the plume (Fig. 6b-c), which has been completed at 50 Myr (Fig. 6d).

344 The interplay between far-field forces and active mantle upwelling results in a “classical” single rift that
345 crosses the entire model domain in the direction perpendicular to external extension (Fig. 6a). Continuous
346 evolution shows the formation of a wide rift valley where localized brittle deformation is concentrated along
347 the boundary fault (Fig. 6b-c). This rift basin opens rapidly (Fig. 6b) reaching a width of 600 km in less than
348 35 Myr while passive extension applied at the boundaries would result in only 200 km width (Fig. 6c). This
349 highlights the important role of plume-related buoyancy forces in the context of studied “active-passive”
350 rifting.

351 The next stage of the system evolution (65 Myr) is a quick switch of deformation localization from rift-
352 bounding faults to narrow zones inside the rift valley (Fig. 6e). This change in rifting style is caused by
353 initiation of localized upwelling of plume material along stretched zone(s) highlighted at the surface by
354 localized high strain rates (Fig. 6e). Further localized plume ascent associated with decompression melting
355 of plume material increases the rate of lithospheric thinning leading to continental break-up along a
356 spreading axis that has shifted laterally outwards from the centre of the plume head (Fig. 6f). This
357 asymmetrical position of the spreading zone arises spontaneously within initially symmetrical and laterally
358 homogenous lithosphere and is likely controlled by melting and cooling processes into head of mantle
359 plume.

360 Thus, a lateral shift of plume-induced break-up centres with respect to initial plume impingement revealed
361 in certain 2D experiments (see for example Experiment 6, Fig. 4a) appears to be an intrinsic characteristic
362 of self-induced plume-related processes that do not necessary requires fast (>1 cm/y) external extension nor

363 any lateral lithospheric heterogeneity (see also Experiment 2 for a 2D example of “distant” break-up in the
364 context of laterally homogenous lithosphere).

365

366 **5. Discussions**

367 *5.1. General aspects*

368 The results of our models are important in the context of ongoing discussion on plume-induced continental
369 break-up. We show that continental break-up can be initiated by just one single mantle plume under different
370 initial and boundary conditions. In most of our 2D models (32 out of 36) continental break-up takes place
371 as a result of the direct (“centred” or “shifted” modes) or indirect (“distant” mode) impact of the mantle
372 plume (Fig. 8). Four remaining models do not result in break-up. On one hand, the models that develop
373 according to “centred” (Fig. 8b) or “shifted” (Fig. 8c) modes are directly induced by the plume anomaly
374 which results in penetration of plume material up to the surface. On the other hand, the “distant” mode is
375 characterized by secondary mantle convection associated with plume-induced *subduction* and/or
376 *convection*. In this case, the mantle plume is not involved directly in break-up processes and remains glued
377 at the base of adjacent unbroken lithosphere (Fig. 8d).

378 In a very early phase, strain rate localizes and topographic variations develop directly above the initial plume
379 impingement location (Fig. 8a). They remain visible only in the “shifted” and “distant” models and can be
380 interpreted as aborted rifts. It was commonly accepted and almost self-evident that in the case of plume-
381 induced continental break-up, its axis should be situated directly above the initial plume impingement
382 position (D’Acremont et al., 2003). However, observations such as failed rifts and deep-seated mantle
383 sources beneath a strong continent that are significantly remote from the mid-oceanic ridge, actually imply
384 that continental rifting and break-up are not a purely symmetric and “plume-centred” processes. Our
385 modelling demonstrates that symmetric development of mantle material ascent and subsequent continental
386 break-up are not a definite outcome. More than half of our models (19 out of 32) result in “shifted” and
387 “distant” break-up modes, suggesting that these modes should also be expected in a wide range of initial
388 and boundary conditions. Even so, our perfectly symmetric and lateral homogeneous 3D model shows that

389 in a purely symmetric setting, with no lithospheric rheological heterogeneities, continental break-up shifted
390 from the original centre of the mantle anomaly is possible. We argue here that “centred” symmetric
391 continental break-up developed directly above mantle plume is only one particular case of possible
392 evolutions of plume-induced break-up systems.

393

394 *5.2. The case of the South Atlantic*

395 The South Atlantic is considered to be a good example of plume-induced continental break-up (e.g. Torsvik
396 et al., 2009). Some of the observations such as failed rifts (Heine et al., 2013) and high velocity bodies (e.g.
397 Blaich et al., 2011) cannot be explained with conventional models, usually assuming a “centred”-like break-
398 up style (e.g. Burov et al., 2007; D’Acremont et al., 2003). Yet, our experiments showing “shifted” break-
399 up mode (Fig. 7; Fig. 8c) can be used to explain these features. In these models, initial crustal deformation
400 associated with mantle plume impingement (Fig.7a-b; Fig.8a) are formed within the first 5 Myr. Significant
401 topography variations developed during this initial stage of rift evolution (i.e. before break-up) can be
402 interpreted as very early failed rift features (e.g. the failed Abimael rift in the southwest of the Santos Basin).
403 The topographic plateaus that remain elevated long after break-up have also been observed with dynamic
404 topography studies (Nyblade and Robinson, 1994). Next, localized strain becomes concentrated close to the
405 boundary between strong and weak lithosphere that is laterally offset (~400 km) from the area of initial
406 plume impingement (Fig.7c). Transition from wide rift stage to lithospheric break-up is marked by
407 narrowing a broad rift region (over 1000 km width) down to narrow rift valley (Fig. 7c-d) with the width of
408 10’s of kilometers between the two rift-shoulders. Associated lithospheric thinning leads to a separation of
409 the two plates along a spreading centre corresponding to South Atlantic ridge (Fig.7e; Fig.8c). Part of strong
410 crust that remains attached to the weaker lithosphere segment, could correspond to the Brazilian craton that
411 was once bordering the African continent (Heine et al., 2013).

412 Simultaneously with thinning of the lithosphere below the future break-up centre, the plume material
413 migrates along the base of the lithosphere and rises towards the deformed crust where it breaks through.

414 This migration can go as far as 300 km from the plume impingement point and only ceases when the material

415 that is still at the base of the lithosphere (at depths between 200 km and 10 km) reaches thermal equilibrium
416 (in this case after 22 Myr). This confirms the hypothesis that one mantle anomaly (or plume) can flow
417 laterally over significant distances below a slow-moving continent, after being risen to the base of the
418 lithosphere (e.g. Sleep, 2006). When thermal equilibrium is reached, the mantle material glued to the base
419 of the lithosphere at shallower depths corresponds geometrically and location-wise to high-velocity/high-
420 density bodies observed on seismic data below the thinned continental lithosphere and the transition zone
421 of the South Atlantic domain (Clerc et al., 2015). During migration, products of partial melting of the mantle
422 material can move vertically to (shallow) lower crustal levels. They might resemble high density bodies
423 observed at lower crustal levels inside continental crust with similar geometries observed with gravity
424 modelling (Blaich et al., 2011). These processes cannot be reproduced by our 2D modelling, because no
425 melt production and extraction have been simulated.

426 Note that our 2D study has not the intention to capture such 3D features like along-axis northward
427 propagation of the break-up axis (Franke, 2013) up to the centre of the surface manifestation of Tristan
428 plume activity – the Paraná-Etendeka continental flood basalts province.

429 After continental break-up, the mantle plume anomaly continues to play an important role in the spreading
430 evolution of the system. Strain rate relocation takes place around 18 Myr, when the spreading axis shifts
431 another 200 km towards the left from the original position of the break-up centre (Fig. 7f). This phenomenon
432 could correspond to a rift-jump that have also been both observed and modelled (Brune et al., 2014) in the
433 South Atlantic domain.

434 The question we raised about the initial position of the mantle plume anomaly responsible for continental
435 break-up in the South Atlantic remains open. On figure 8e, we show a reconstructed configuration of the
436 slow velocity anomaly (corresponding to the South African Super Plume) at the CMB based on present-day
437 seismic tomography model (after Davaille et al., 2005). The orange dots refer to the three possible principal
438 locations of initial thermal anomaly at the upper/lower mantle boundary corresponding to the Tristan plume.
439 The central point refers to the most evident “centered” scenario (Fig. 8b) when the deep-seated thermal
440 anomaly in the upper mantle is supposed to be located directly below its surface manifestation and hints to

441 voluminous Paranà-Etendeka continental flood basalts province (Fig. 8e, e.g. Torsvik et al., 2009; Heine et
442 al., 2013). This scenario, however, is not consistent with the commonly considered concept that plumes
443 emerge from the edges of the large low-velocity anomalies at the CMB that has been confirmed by both
444 numerical modelling (Hassan et al., 2015) and by empirically established correlation between downward
445 projected plume-associated large igneous provinces and the margins of the large low shear velocity province
446 beneath Africa (Torsvik et al., 2006). Moreover, with this ‘centred’ scenario we cannot explain additional
447 features such as failed rift arms and anomalous bodies at lower crustal levels. The ‘distant’ break-up mode
448 (Fig. 8d), where the initial plume centre is located below the South American section and remains there after
449 “plume-independent” continental break-up, does not fit well with geological observations of the voluminous
450 Paranà-Etendeka continental flood basalts that are supposed to be related with direct influence of the Tristan
451 hot spot (Torsvik et al., 2009). Finally, initial plume position slightly moved to African side (right dot on
452 Fig. 8e) refers to “shifted” scenario that seems to be preferable (Fig.8c). The time length of the modelled
453 rift phase (10 Myr +/- 3 Myr) is much shorter than has been inferred from geological and geophysical
454 observations (160 Ma to 134 Ma, (Franke, 2013)) in the South Segment of the South Atlantic. Despite this,
455 with the eastward offset initial position of the mantle plume with respect to a heterogeneous thermo-
456 rheological lithospheric structure we are able to explain not only plume induced flood basalts but also a set
457 of anomalous features such as failed rift systems, and deep crustal bodies.

458

459 **6. Conclusion**

460 Different lithospheric strengths comparable to the South American and African continental crust, inherited
461 structures, boundary velocity conditions corresponding to average spreading rates, and initial location of a
462 thermal mantle anomaly (i.e. plume) have been tested to investigate the dynamics of plume induced
463 continental break-up. A set of 36 models shows that with only one anomaly three very different scenarios
464 for continental break-up can be realized depending on the rheological structure, anomaly location and
465 inherited structures. Continental break-up does not necessarily occur above the centre of the initial location

466 of a mantle anomaly. As mentioned above, our models show three types of break-up 1) “centred” break-up,
467 2) “shifted” break-up and 3) “distant” break-up.

468 “Centred” and “shifted” break-up types of models refer to plume-induced type of break-up. For the first
469 mode, mantle material rises vertically towards the bottom of the lithosphere after which it breaks through
470 the crust and reaches the surface directly above the initial plume position. The “shifted” type of break-up
471 shows continental break-up that is 50 to 200 km shifted from the initial location of the mantle anomaly. In
472 this case, the mantle plume rises and impinges the lithosphere, after that it migrates laterally and cuts through
473 the lithosphere reaching the surface at a break-up point considerably shifted from the area of initial, pre-
474 break-up impingement. Some material remains glued underneath the lithosphere at depths between 200 and
475 10 km. These deep-seated bodies, at depths of 200 km, are not situated directly below the break-up centre,
476 but are spread over a large area below the continental margins. The shallower bodies geometrically resemble
477 high density/high velocity bodies detected by seismic profiling and gravity modelling along the margins of
478 the South Atlantic domain and at lower crustal levels.

479 The “distant” break-up mode refers to continental rupture that is indirectly induced by the presence of the
480 mantle plume ponding at the bottom of the weaker continental lithosphere, when “plume-independent”
481 break-up of adjacent stronger lithosphere appears to be considerably (from 250 and 800 km) displaced from
482 the location of plume-lithosphere interaction. In this case, laterally widely spread plume material remains
483 glued below unbroken segments of the lithosphere.

484 Topographic changes that occur very early during initial rifting stage remain visible for a long period and
485 can possibly be interpreted as failed rift systems (in the cases of “shifted” and “distance” modes). Strain
486 relocation after continuous post-break-up extension could be interpreted as rift jumps. A simple 3D model
487 has been built to illustrate that even in a fully symmetric setup, rift-to-break-up processes are not by default
488 symmetric and can very well evolve asymmetrically.

489 There is no controlling parameter for one of the three types of rifting, with a combination of parameters
490 determining the outcome, but the location of the mantle anomaly with respect to the rheology is the most

491 essential. The most important result of this study is that there is not one single rift mode for plume-induced
492 crustal break-up.

493

494 **7. Acknowledgment**

495 This study is co-funded by the Advanced ERC Grant 290864 RHEOLITH to A. Koptev and E.Burov. 3D
496 numerical simulations were performed on the ERC-funded SGI Ulysse cluster of ISTEP. We thank William
497 Sassi (IFPE), Xavier Guichet (IFPE) and Sylvie Leroy (UPMC) for fruitful discussions in the early phases
498 of the project. We would also like to thank Sierd Cloetingh (Utrecht University) for constructive and
499 thorough reading of the manuscript. The two anonymous reviewers are also thanked for their helpful
500 remarks and critical points of view.

501

502 **8. References**

- 503 Autin, J., Scheck-Wenderoth, M., Loegering, M.J., Anka, Z., Vallejo, E., Rodriguez, J.F., Dominguez, F.,
504 Marchal, D., Reichert, C., di Primio, R., Götze, H.-J., 2013. Colorado Basin 3D structure and
505 evolution, Argentine passive margin. *Tectonophysics* 604, 264–279. doi:10.1016/j.tecto.2013.05.019
- 506 Blaich, O.A., Faleide, J.I., Tsikalas, F., 2011. Crustal breakup and continent-ocean transition at South
507 Atlantic conjugate margins. *J. Geophys. Res.* 116, 1–38. doi:10.1029/2010JB007686
- 508 Brune, S., Heine, C., Pérez-Gussinyé, M., Sobolev, S. V., 2014. Rift migration explains continental margin
509 asymmetry and crustal hyper-extension. *Nat. Commun.* 5, 1–9. doi:10.1038/ncomms5014
- 510 Burke, K., 1976. Development of graben associated with the initial ruptures of the Atlantic Ocean.
511 *Tectonophysics* 36, 93–112.
- 512 Burov, E., Gerya, T., 2014. Asymmetric three-dimensional topography over mantle plumes. *Nature* 513,
513 85–89. doi:10.1038/nature13703
- 514 Burov, E., Guillou-Frottier, L., D’Acromont, E., Le Pourhiet, L., Cloetingh, S., 2007. Plume head-
515 lithosphere interactions near intra-continental plate boundaries. *Tectonophysics* 434, 15–38.

516 doi:10.1016/j.tecto.2007.01.002

517 Burov, E., Poliakov, A., 2001. Erosion and rheology controls on synrift and postrift evolution: Verifying
518 old and new ideas using a fully coupled numerical model. *J. Geophys. Res.* 106, 16461–16481.
519 doi:10.1029/2001JB000433

520 Burov, E.B., Diament, M., 1995. The effective elastic thickness (T_e) of continental lithosphere: What does
521 it really mean? *J. Geophys. Res.* 100, 3905–3927. doi:10.1029/94JB02770

522 Clerc, C., Jolivet, L., Ringenbach, J.C., 2015. Ductile extensional shear zones in the lower crust of a
523 passive margin. *Earth Planet. Sci. Lett.* 431, 1–7. doi:10.1016/j.epsl.2015.08.038

524 Cloetingh, S., Burov, E., Matenco, L., Beekman, F., Roure, F., Ziegler, P.A., 2013. The Moho in
525 extensional tectonic settings: Insights from thermo-mechanical models. *Tectonophysics* 609, 558–
526 604. doi:10.1016/j.tecto.2013.06.010

527 Cobbold, P.R., Gapais, D., Rossello, E.A., Milani, E.J., Szatmari, P., 1992. Permo-Triassic
528 intracontinental deformation in SW Gondwana. *Invers. Tectonics Cape Fold Belt, Karoo Cretac.*
529 *Basins South. Africa* 23–26.

530 Contrucci, I., Matias, L., Moulin, M., Géli, L., Klingelhofer, F., Nouzé, H., Aslanian, D., Olivet, J.L.,
531 Réhault, J.P., Sibuet, J.C., 2004. Deep structure of the West African continental margin (Congo,
532 Zaïre, Angola), between 5°S and 8°S, from reflection/refraction seismics and gravity data. *Geophys.*
533 *J. Int.* 158, 529–553. doi:10.1111/j.1365-246X.2004.02303.x

534 Cornwell, D.G., Mackenzie, G.D., England, R.W., Maguire, P.K.H., Asfaw, L.M., Oluma, B., 2006.
535 Northern Main Ethiopian Rift crustal structure from new high-precision gravity data. *Geol. Soc.*
536 *London, Spec. Publ.* 259, 307–321. doi:10.1144/GSL.SP.2006.259.01.23

537 Cundall, P.A., 1989. Numerical experiments on localization in frictional materials. *Ingenieur-Archiv* 59,
538 148–159.

539 D’Acremont, E., Leroy, S., Burov, E.B., 2003. Numerical modelling of a mantle plume: the plume head–
540 lithosphere interaction in the formation of an oceanic large igneous province. *Earth Planet. Sci. Lett.*
541 206, 379–396. doi:10.1016/S0012-821X(02)01058-0

542 Davaille, A., Stutzmann, E., Silveira, G., Besse, J., Courtillot, V., 2005. Convective patterns under the
543 Indo-Atlantic « box ». *Earth Planet. Sci. Lett.* 239, 233–252. doi:10.1016/j.epsl.2005.07.024

544 Dressel, I., Scheck-Wenderoth, M., Cacace, M., Lewerenz, B., Götze, H.-J., Reichert, C., 2015.
545 Reconstruction of the southwestern African continental margin by backward modelling. *Mar. Pet.*
546 *Geol.* 67, 544–555. doi:10.1016/j.marpetgeo.2015.06.006

547 Dupré, S., Bertotti, G., Cloetingh, S., 2007. Tectonic history along the South Gabon Basin: anomalous
548 early post-rift subsidence. *Mar. Pet. Geol.* 24, 151–172. doi:10.1016/j.marpetgeo.2006.11.003

549 Fishwick, S., 2010. Surface wave tomography: Imaging of the lithosphere – asthenosphere boundary
550 beneath central and southern Africa? *Lithos* 120, 63–73. doi:10.1016/j.lithos.2010.05.011

551 Franke, D., 2013. Rifting, lithosphere breakup and volcanism : Comparison of magma-poor and volcanic
552 rifted margins. *Mar. Pet. Geol.* 43, 63–87. doi:10.1016/j.marpetgeo.2012.11.003

553 Franke, D., Neben, S., Schreckenberger, B., Schulze, A., Stiller, M., Krawczyk, C.M., 2006. Crustal
554 structure across the Colorado Basin, offshore Argentina. *Geophys. J. Int.* 165, 850–864.
555 doi:10.1111/j.1365-246X.2006.02907.x

556 Gerya, T., 2010. Dynamical instability produces transform faults at mid-ocean ridges. *Science* (80-.). 329,
557 1047–1050.

558 Gerya, T. V., Yuen, D.A., 2007. Robust characteristics method for modelling multiphase visco-elasto-
559 plastic thermo-mechanical problems. *Phys. Earth Planet. Inter.* 163, 83–105.
560 doi:10.1016/j.pepi.2007.04.015

561 Hassan, R., Flament, N., Gurnis, M., Bower, D.J., Müller, M., 2015. Geochemistry, Geophysics,
562 Geosystems. *Geochemistry Geophys. Geosystems* 18, 1541–1576. doi:10.1002/2014GC005684.Key

563 Heine, C., Zoethout, J., Müller, R.D., 2013. Kinematics of the South Atlantic rift. *Solid Earth* 4, 215–253.
564 doi:10.5194/se-4-215-2013

565 Heit, B., Sodoudi, F., Yuan, X., Bianchi, M., Kind, R., 2007. An S receiver function analysis of the
566 lithospheric structure in South America. *Geophys. Res. Lett.* 34, 1–5. doi:10.1029/2007GL030317

567 Hirsch, K.K., Scheck-Wenderoth, M., Wees, J.-D. Van, Kuhlmann, G., Paton, D.A., 2010. Tectonic

568 subsidence history and thermal evolution of the Orange Basin. *Mar. Pet. Geol.* 27, 565–584.
569 doi:10.1016/j.marpetgeo.2009.06.009

570 Husson, L., Conrad, C.P., Faccenna, C., 2012. Plate motions, Andean orogeny, and volcanism above the
571 South Atlantic convection cell. *Earth Planet. Sci. Lett.* 317–318, 126–135.

572 Koptev, A., Burov, E., Calais, E., Leroy, S., Gerya, T., Guillou-Frottier, L., Cloetingh, S., 2016.
573 Contrasted continental rifting via plume-craton interaction: Applications to Central East African Rift.
574 *Geosci. Front.* 7, 221–236. doi:10.1016/j.gsf.2015.11.002

575 Koptev, A., Calais, E., Burov, E., Leroy, S., Gerya, T., 2015. Dual continental rift systems generated by
576 plume – lithosphere interaction. *Nat. Geosci.* 8, 388–392. doi:10.1038/NGEO2401

577 Le Pourhiet, L., Burov, E., Moretti, I., 2004. Rifting through a stack of inhomogeneous thrusts (the
578 dipping pie concept). *Tectonics* 23, 1–14. doi:10.1029/2003TC001584

579 Lithgow-Bertelloni, C., Silver, P., 1998. Dynamic topography, plate driving forces and the African
580 superswell. *Nature* 395, 345–348. doi:10.1038/26212

581 Maystrenko, Y.P., Scheck-Wenderoth, M., Hartwig, A., Anka, Z., Watts, A.B., Hirsch, K.K., Fishwick, S.,
582 2013. Structural features of the Southwest African continental margin according to results of
583 lithosphere-scale 3D gravity and thermal modelling. *Tectonophysics* 604, 104–121.
584 doi:10.1016/j.tecto.2013.04.014

585 Mohriak, W.U., Mello, M.R., Vieira, I.S., Bassetto, M., Koutsoukos, E.A.M., 2000. Crustal architecture,
586 sedimentation, and petroleum systems in the Sergipe-Alagoas Basin, Northeastern Brazil. *AAPG*
587 *Spec. Vol.* 273–300.

588 Moulin, M., Aslanian, D., Unternehr, P., 2010. A new starting point for the South and Equatorial Atlantic
589 Ocean. *Earth-Science Rev.* 98, 1–37. doi:10.1016/j.earscirev.2009.08.001

590 Müller, R.D., Sdrolias, M., Gaina, C., Roest, W.R., 2008. Age, spreading rates, and spreading asymmetry
591 of the world's ocean crust. *Geochemistry, Geophys. Geosystems* 9, 1–19.
592 doi:10.1029/2007GC001743

593 Nyblade, A.A., Robinson, S.W., 1994. The African Superswell. *Geophys. Res. Lett.* 21, 765–768.

594 Poliakov, A.N.B., Podladchikov, Y., Talbot, C., 1993. Initiation of salt diapirs with frictional overburdens:
595 numerical experiments. *Tectonophysics* 228, 199–210. doi:10.1016/0040-1951(93)90341-G

596 Ranalli, G., 1995. *Rheology of the Earth*, 2nd edition. Chapman and Hall, 413 pp.

597 Ribe, N.M., Christensen, U.R., 1994. Three-dimensional modeling of plume-lithosphere interaction. *J.*
598 *Geophys. Res.* 99, 669–682.

599 Sandwell, D.T., Smith, W.H.F., 2009. Global marine gravity from retracked Geosat and ERS-1 altimetry:
600 Ridge segmentation versus spreading rate. *J. Geophys. Res. Solid Earth* 114, 1–18.
601 doi:10.1029/2008JB006008

602 Schnabel, M., Franke, D., Engels, M., Hinz, K., Neben, S., Damm, V., Grassmann, S., Pelliza, H., Dos
603 Santos, P.R., 2008. The structure of the lower crust at the Argentine continental margin, South
604 Atlantic at 44°S. *Tectonophysics* 454, 14–22. doi:10.1016/j.tecto.2008.01.019

605 Sleep, N.H., 2006. Mantle plumes from top to bottom. *Earth-Science Rev.* 77, 231–271.
606 doi:10.1016/j.earscirev.2006.03.007

607 Torsvik, T.H., Rouse, S., Labails, C., Smethurst, M.A., 2009. A new scheme for the opening of the South
608 Atlantic Ocean and the dissection of an Aptian salt basin. *Geophys. J. Int.* 177, 1315–1333.
609 doi:10.1111/j.1365-246X.2009.04137.x

610 Torsvik, T.H., Smethurst, M.A., Burke, K., Steinberger, B., 2006. Large igneous provinces generated from
611 the margins of the large low-velocity provinces in the deep mantle. *Geophys. J. Int.* 167, 1447–1460.
612 doi:10.1111/j.1365-246X.2006.03158.x

613 Tsenn, M.C., Carter, N.L., 1987. Upper limits of power law creep of rocks. *Tectonophysics* 136, 1–26.
614 doi:10.1016/0040-1951(87)90332-5

615 Turcotte, D.L., Schubert, G., 2002. *Geodynamics*. Cambridge Univ. Press, Cambridge, UK.

616 Van Der Meijde, M., Julià, J., Assumpção, M., 2013. Gravity derived Moho for South America.
617 *Tectonophysics* 609, 456–467. doi:10.1016/j.tecto.2013.03.023

618 Yamato, P., Burov, E., Agard, P., Le Pourhiet, L., Jolivet, L., 2008. HP-UHP exhumation during slow
619 continental subduction: Self-consistent thermodynamically and thermomechanically coupled model

620 with application to the Western Alps. *Earth Planet. Sci. Lett.* 271, 63–74.
621 doi:10.1016/j.epsl.2008.03.049
622 Ziegler, P.A., Cloetingh, S., 2004. Dynamic processes controlling evolution of rifted basins. *Earth-Science*
623 *Rev.* 64, 1–50. doi:10.1016/S0012-8252(03)00041-2
624

625 **Figure captions**

626

627 Figure 1. Map of the South Atlantic domain with the location of large fracture zones, high velocity bodies
628 (red ellipsoids), onshore graben structures (black dashed lines), the outline of the African Super Plume
629 (dashed green line, after Davaille et al. (2015)) and the African superswell (dashed red line, after Nyblade
630 and Robinson, 1994). Also shown are the extent of the Seaward Dipping Reflectors (SDR's, blue after
631 Moulin et al., 2010 and green after Torsvik et al., 2009) and the Aptian salt (yellow, after Torsvik et al.,
632 2009) deposits. The orange line gives the location of the Gondwana Fold-and-Thrust-Belt. Pink solid lines
633 mark locations of the lithosphere-scale cross-sections (South America: A-A'; South Africa: B-B', Fig. 2).
634 Hotspots (red stars): Tr = Trinidad hotspot; StH = Saint Helena hotspot; Bv = Bouvet (Meteor) hotspot;
635 Deep-rooted mantle plume (yellow star): TdC = Tristan deep-rooted hotspot.

636

637 Figure 2. Lithosphere scale cross-sections of present-day South Atlantic margins. The Moho depth varies
638 from 10 to 30 km on the South American side from ocean to continent (Schnabel et al., 2008). On the
639 African side the depth varies from less than 10 km to over 40 km, ocean to continent (Maystrenko et al.,
640 2013). The lithosphere-asthenosphere boundary (LAB) varies from 50 km to 120 km, ocean to continent for
641 Colorado basin on the Argentinean margin (Heit et al., 2007). The Orange Basin on the South African
642 margin has a LAB depth ranging between 80 km and 200 km from ocean to continent (Fishwick, 2010). The
643 location of anomalous bodies is depicted (in green) for the Colorado Basin (Schnabel et al., 2008) and the
644 Orange Basin (Maystrenko et al., 2013). The gravity profile has been extracted from the global marine
645 gravity map of Sandwell and Smith v18.1 (Sandwell and Smith, 2009).

646

647 Figure 3. Six tested numerical setups. a) Setup 1a: 4-layered weak rheology, crust 40 km thick, lithosphere
648 200 km thick. b) Setup 1b: strong 4-layered rheology, crust 40 km thick, lithosphere 200 km thick. c) Setup
649 2a: combined rheological profiles (weak on the left side, strong on the right side), crust 40 km thick, and

650 lithosphere 200 km (equal for both rheologies). d) Setup 2b: combined rheological strength envelopes,
651 (weak on the left side, strong on the right side), crust 30 km thick on the right side and 40 km thick on the
652 left side, lithosphere 180 km thick on the left side and 200 km thick on the right side, no complex contact
653 geometries. e) Setup 3a: combined rheological strength envelopes, (weak on the left side, strong on the right
654 side), crust 30 km thick on the right side and 40 km thick on the left side; lithosphere 180 km thick on the
655 left side and 200 km thick on the right side. The contact between the two different crustal thicknesses is a
656 low-angle geometry, dipping towards the right. f). Setup 3b: combined rheological strength envelopes,
657 (weak on the left side, strong on the right side), crust 30 km on the right side and 40 km thick on the left
658 side, lithosphere 180 km thick on the left side and 200 km thick on the right side. The contact between the
659 two different crustal thicknesses is a low-angle geometry dipping towards the left.

660
661 Figure 4. Models with different rheology and plume location showing the most representative examples of
662 the three modes of continental break-up. a) “Centred”: Experiment 6, Setup 1, with a strong rheology and
663 the anomaly located at 1200 km (200 km offset from the centre towards the right). At 2.1 Myr the first
664 topographic response occurs. The break-up axis develops directly above the initial mantle plume position
665 and mantle material reaches the surface. b) “Shifted”: Experiment 12, Setup 2, with a laterally varying
666 rheology and the anomaly positioned in the centre at 1000 km. At 2.1 Myr the first topographic variation
667 shows with a larger extend than the “centred” break-up model. The break-up axis develops offset from the
668 original mantle plume location and mantle material migrates towards the spreading centre, reaching the
669 surface. c) “Distant”: Experiment 23, Setup 3, has a laterally varying rheology and the anomaly is positioned
670 in the centre at 1000 km. At 2.1 Myr minor topographic variation. The break-up axis develops far offset
671 from the original mantle plume location and the mantle plume remains glued to the base of the lithosphere.
672 The initial topographic variations remain visible after break-up.

673
674 Figure 5. Examples of models with different setup, plume location rheological structure showing the
675 different modes of break-up; a-c) “centred” examples, d-f) “shifted” examples, g-i) “distant” examples. In

676 red the initial location of the mantle anomaly is drawn. The graphs below show the normalised, statistical
677 likelihood of a mode (“centred”, “shifted” or “distant”) for a given setups.

678
679 Figure 6. Evolution of 3D model a) rapid plume uplift leading to formation of linear extension-perpendicular
680 rift at the crustal level; b-c) development of wide rift basin with localized crustal high strain along bounding
681 normal faults; gradual cooling and solidification of plume head material; d) widely distributed rift above
682 completely crystallized plume head ponding lithosphere-asthenosphere boundary; e) rapid transition from
683 deformation localized in normal faults bounding wide rift valley to localized strain within narrow zones
684 associated with localized plume ascent; f) breakup of the continental lithosphere along spreading zone
685 considerably shifted from centre of the mantle plume. Bulk of plume material is shown in pale orange. Green
686 to red colours indicate strain rate at the level of 10 km (i.e. within upper crust). Component distribution is
687 shown for vertical cross-sections trough central part of the model domain.

688
689 Figure 7. Detailed display of experiment 12. Setup 2 is the base for this experiment with a laterally varying
690 rheology. The anomaly is positioned in the centre of the model at 1000 km. At 2.1 Myr the first topographic
691 variation shows with a larger extend than the “centred” break-up model. Strain rate localizes slightly shifted
692 from the plume impingement point at 3.1 Myr. Strain localizes, topography variations grow and crustal
693 break-up occurs at 8.1 Myr. Mantle material slowly migrates towards the spreading centre reaching the
694 surface until the material that remains below the lithosphere reaches thermal equilibrium around 22 Myr.

695
696 Figure 8. Schematic representation of the three modes of break-up. a) The very early phase of rifting (0 –
697 5 Myr) is very similar for all three modes after which they develop into b) “Centred”; c) “shifted” and d)
698 “distant”. An example of a simplified interpretation of the Uruguayan margin (after Clerc et al., 2015) is
699 used to demonstrate the resemblance of the “shifted” mode of break-up, like experiment 12, with the
700 South Atlantic domain. e) Map showing the outline of the plate configuration at the moment of break-up

701 between Africa and South America on the lower mantle low velocity zone (South African Super Plume)
702 (from Davaille et al., 2005). The Parana-Etendeka flood basalts are depicted in green (after Torsvik et al.,
703 2009). The orange dots refer to the three possible principal locations of initial thermal anomaly at the
704 upper/lower mantle boundary corresponding to the Tristan plume

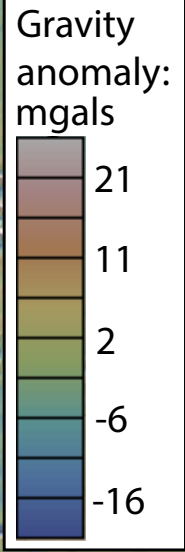
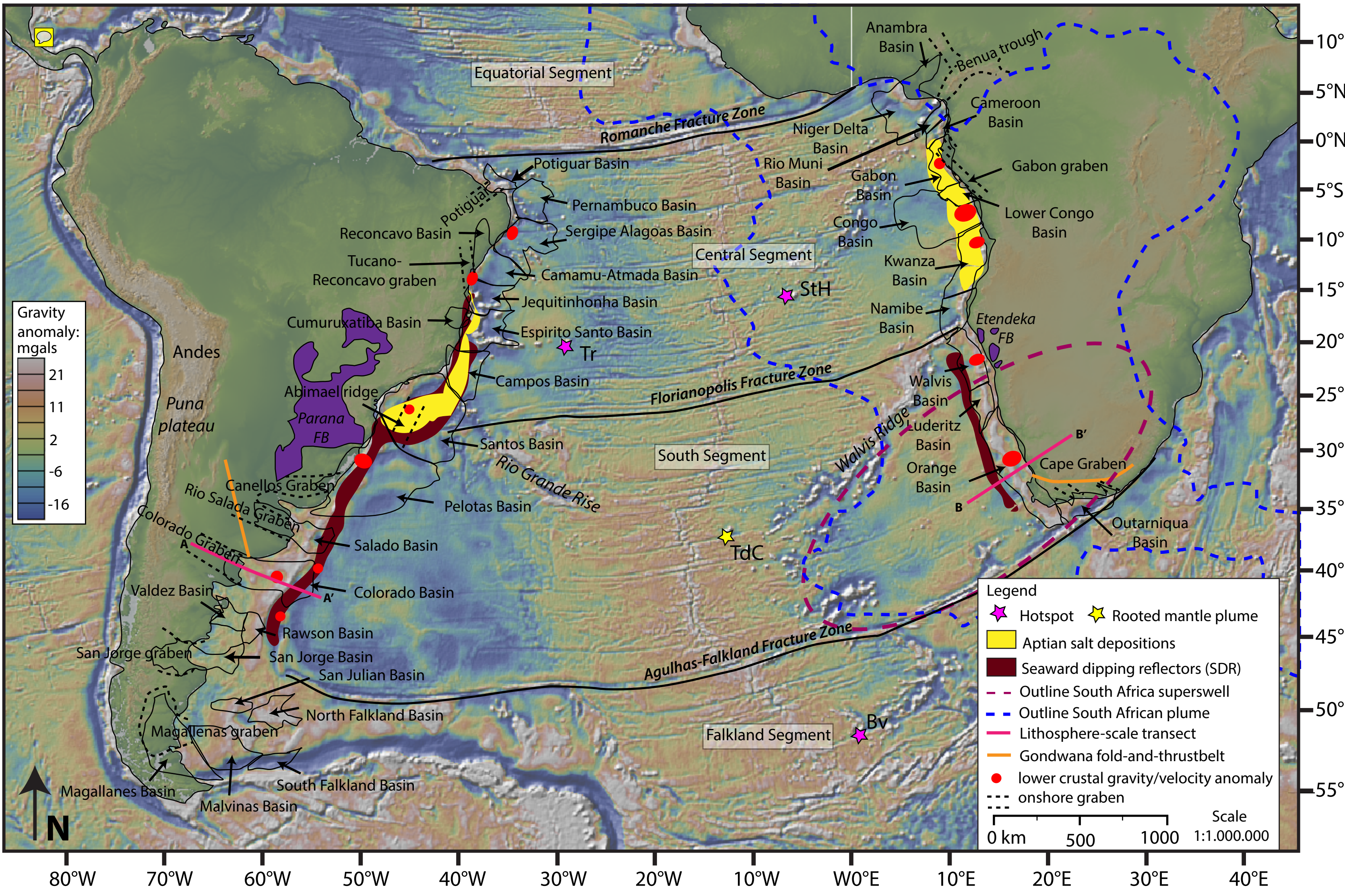
705

706 **Table caption**

707 Table 1. Summary of the thermal and mechanical parameters used for this study. 1) Turcotte and Schubert
708 (2002), 2) Ranalli (1995); 3) d’Acremont et al. (2003) and references therein; 4) Tsenn and Carter (1987);
709 5) Burov and Poliakov (2001).

710

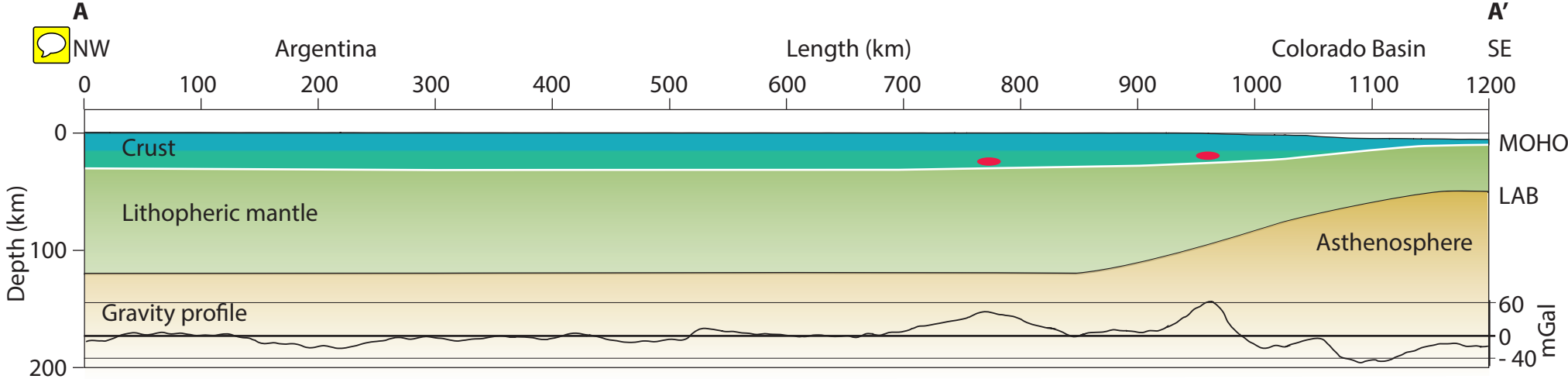
711 Table 2. Controlling parameters and principal results of the experiments.



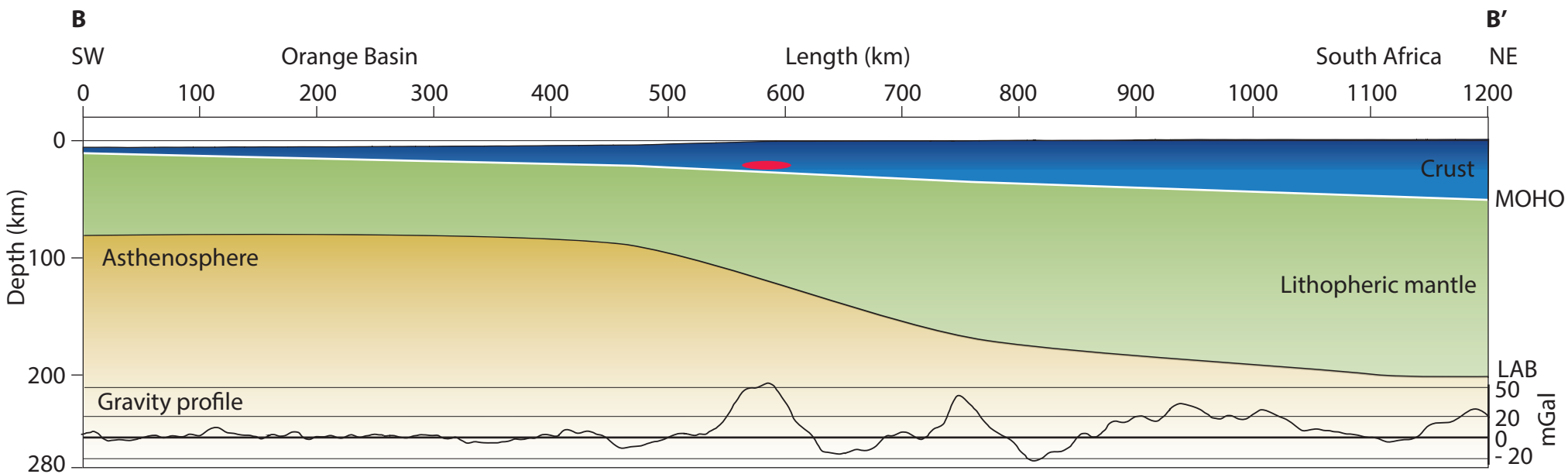
Legend

- ☆ Hotspot
- ☆ Rooted mantle plume
- Aptian salt depositions
- Seaward dipping reflectors (SDR)
- - - Outline South Africa superswell
- - - Outline South African plume
- Lithosphere-scale transect
- Gondwana fold-and-thrustbelt
- lower crustal gravity/velocity anomaly
- - - onshore graben

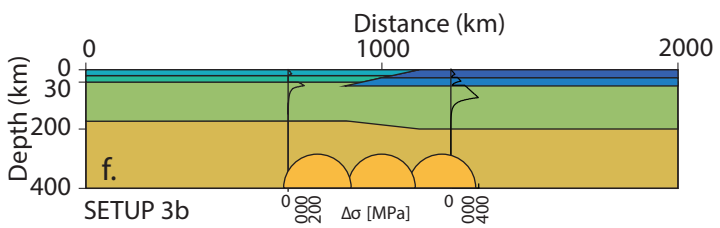
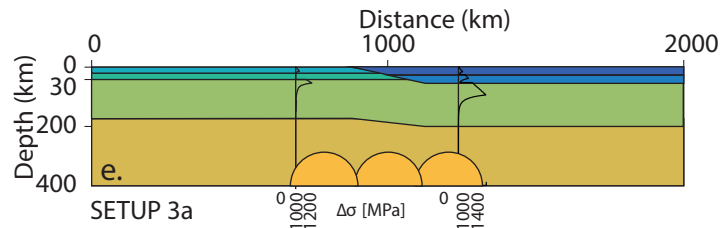
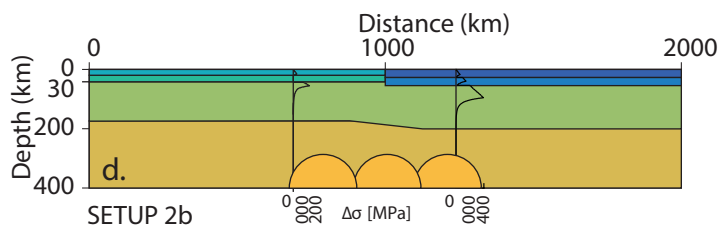
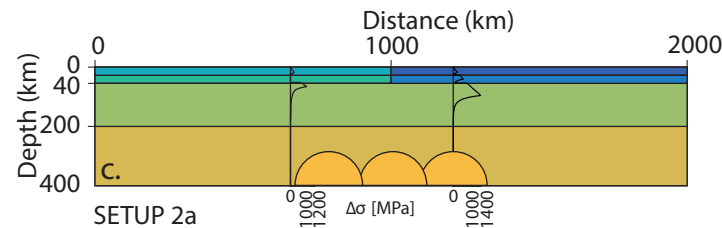
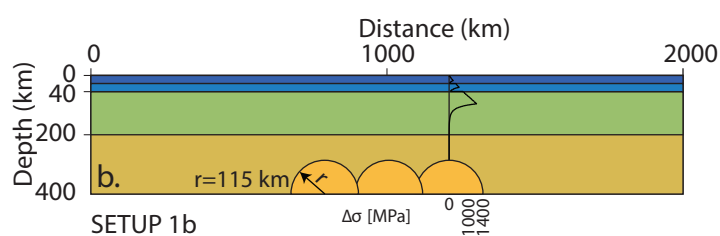
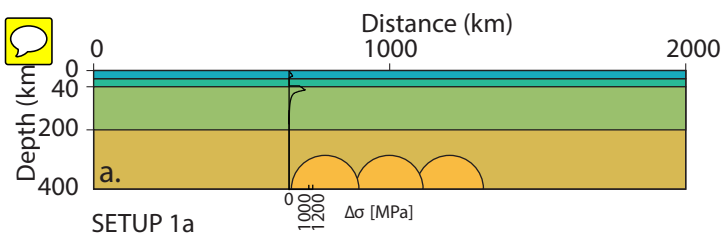
Scale
0 km 500 1000 1:1,000,000



a. South America



b. Southwest Africa



LEGEND

Wet quartzite upper crust

Dry quartzite upper crust

Dry peridotite lithospheric mantle

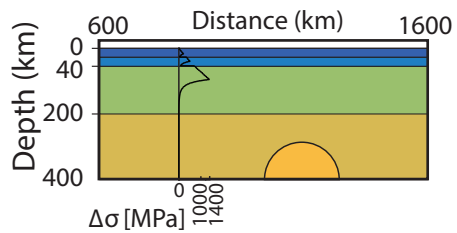
Olivine mantle anomaly

Wet diabase lower crust

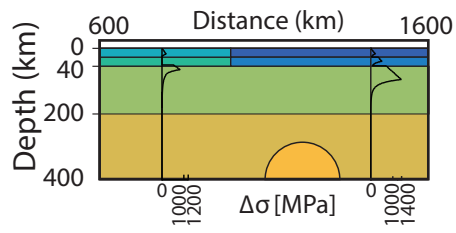
Dry diabase lower crust

Olivine asthenospheric mantle

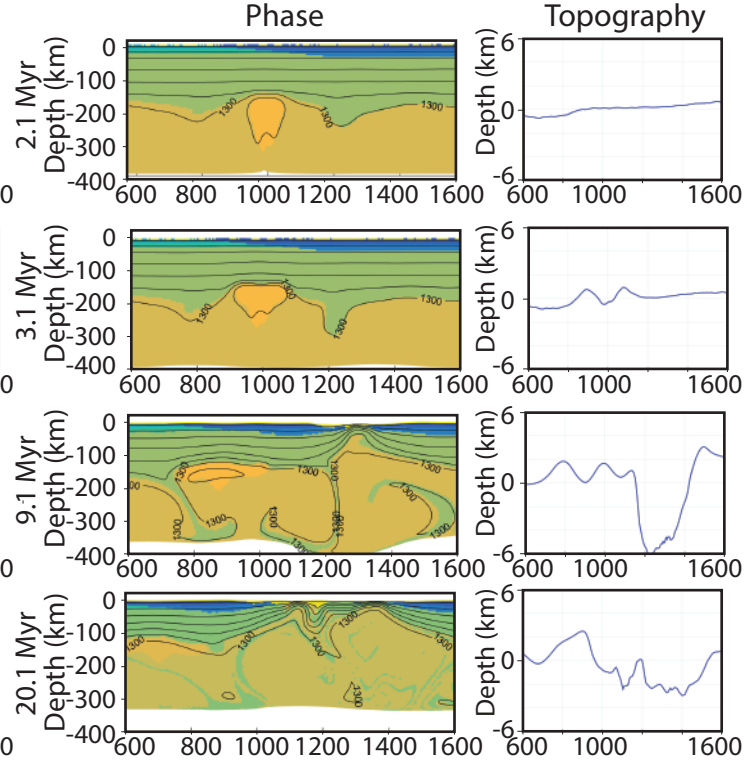
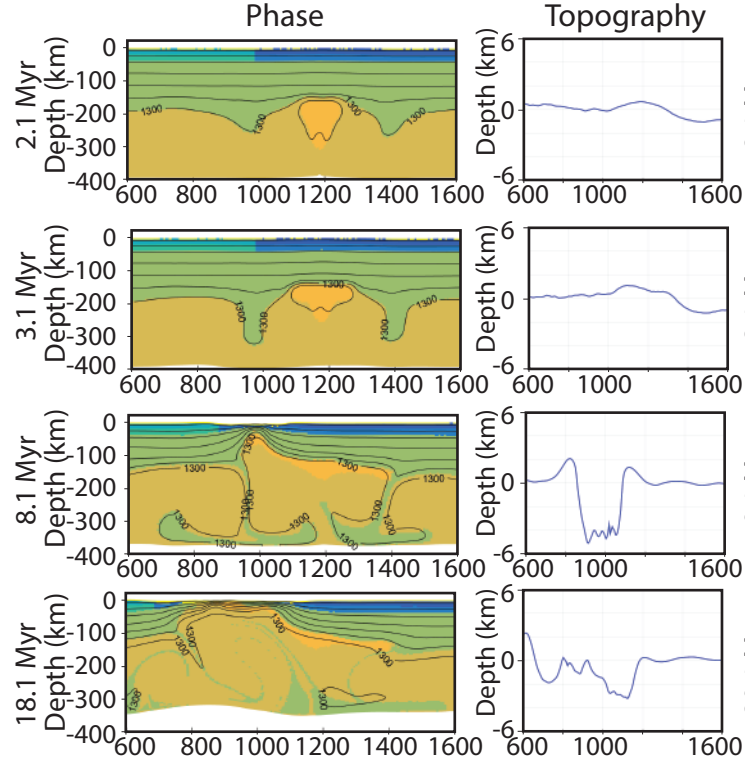
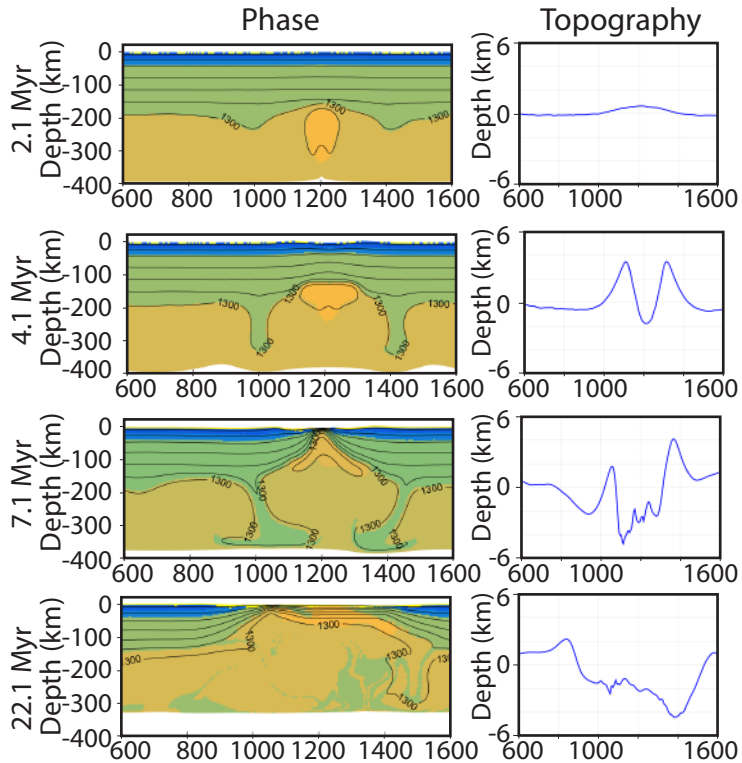
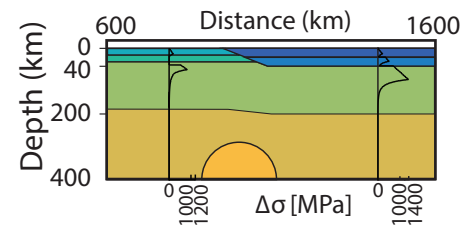
“Central” break-up



“Shifted” break-up



“Distant” break-up



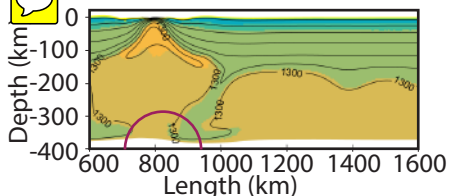
a. Experiment 6

b. Experiment 12

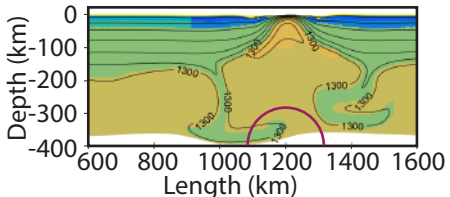
c. Experiment 23



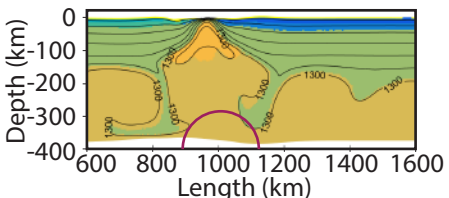
"Central" break-up



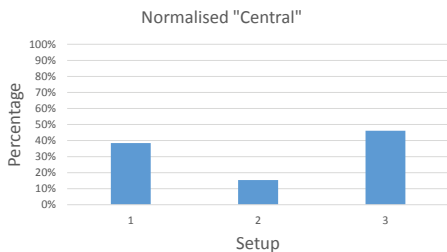
a. Experiment 1 @ 7 Myr



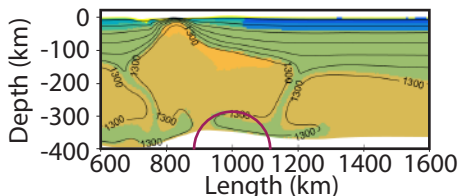
b. Experiment 9 @ 8 Myr



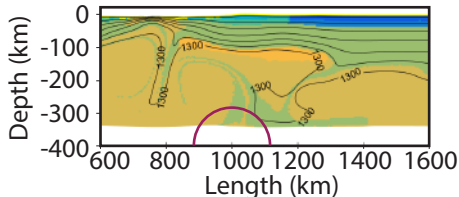
c. Experiment 17 @ 7 Myr



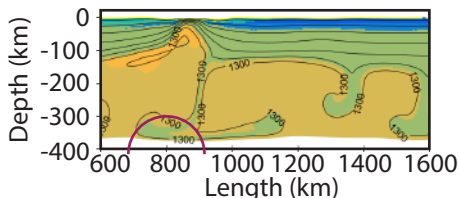
"Shifted" break-up



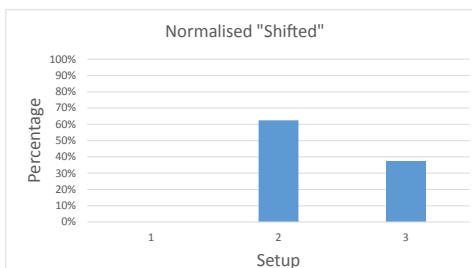
d. Experiment 11 @ 9 Myr



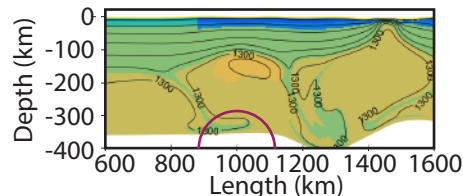
e. Experiment 8 @ 17 Myr



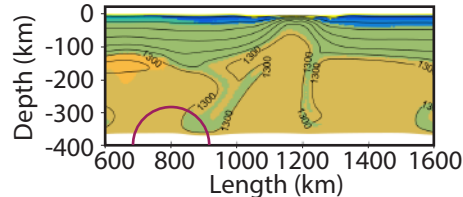
f. Experiment 34 @ 9 Myr



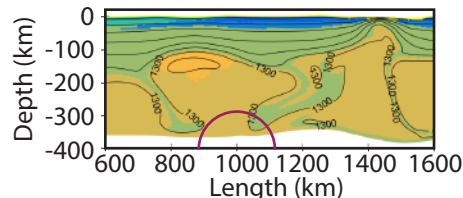
"Distant" break-up



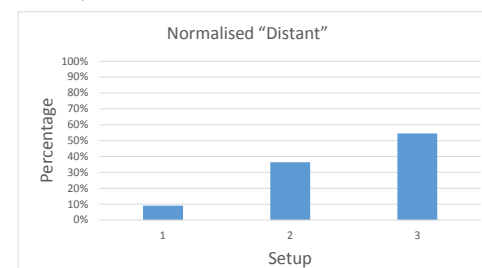
g. Experiment 14 @ 11 Myr

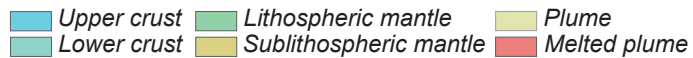
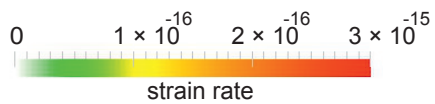
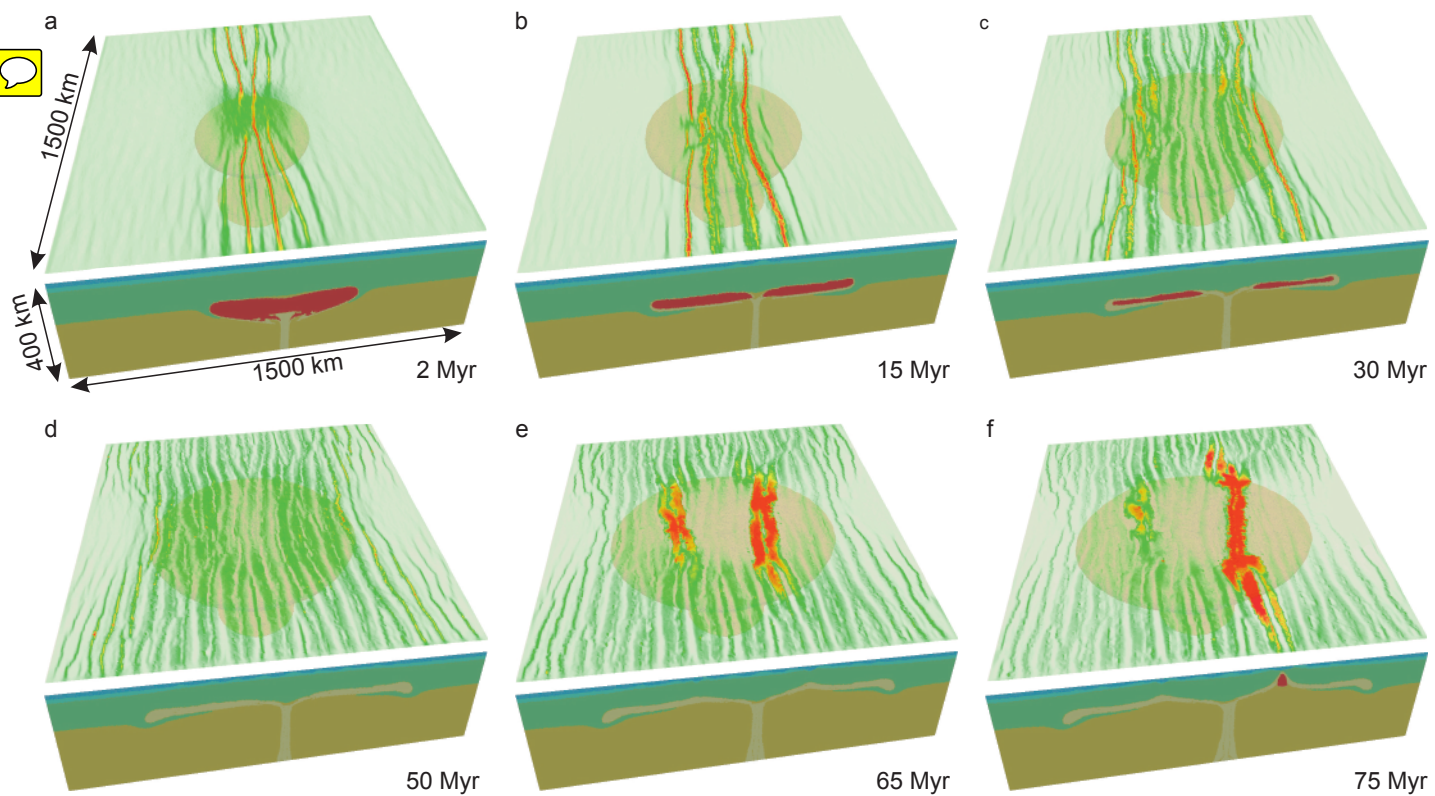


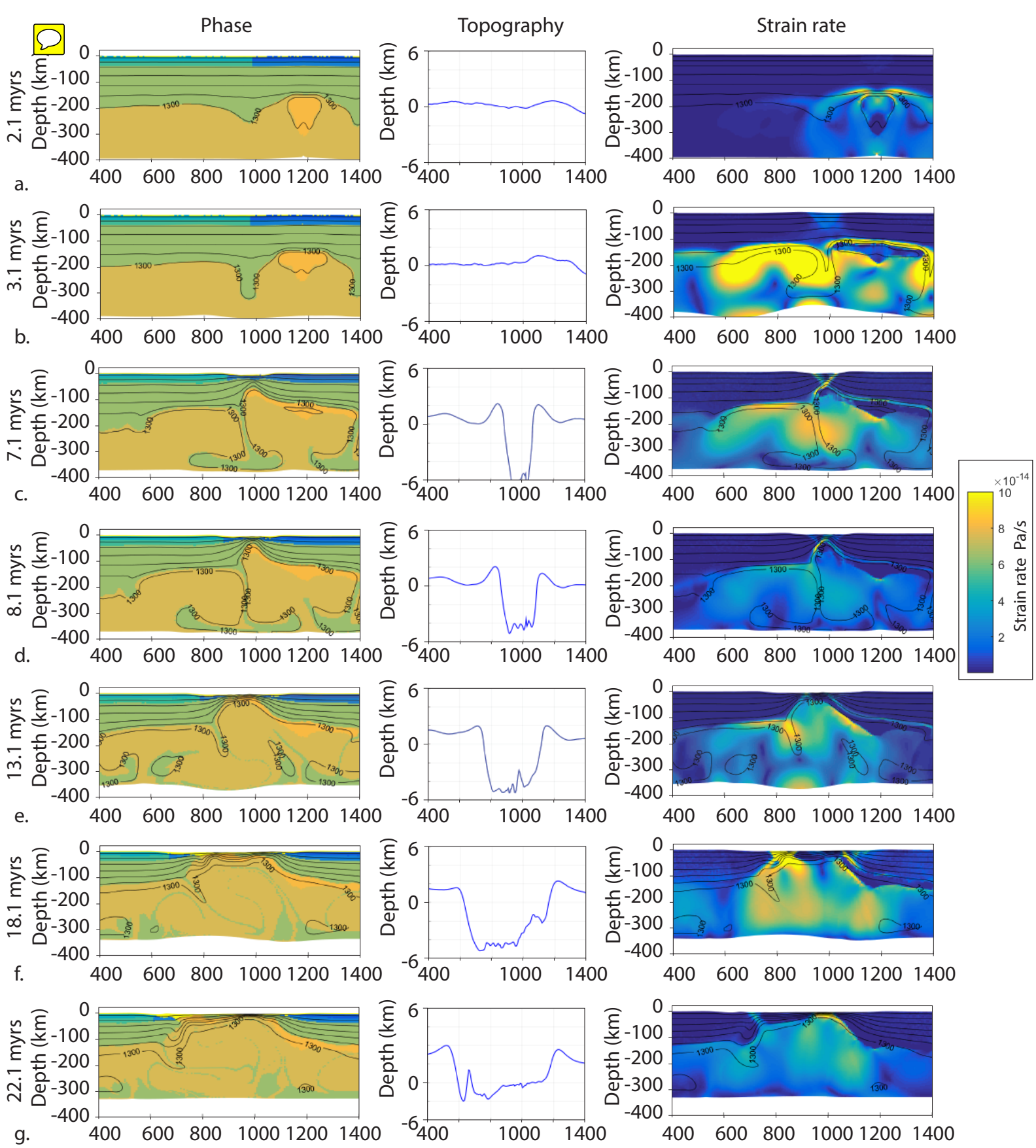
h. Experiment 22 @ 10 Myr

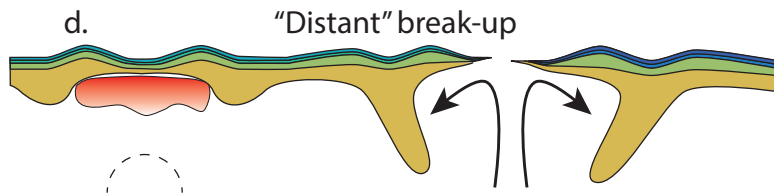
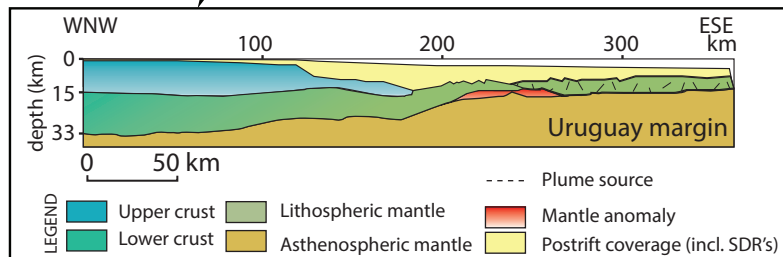
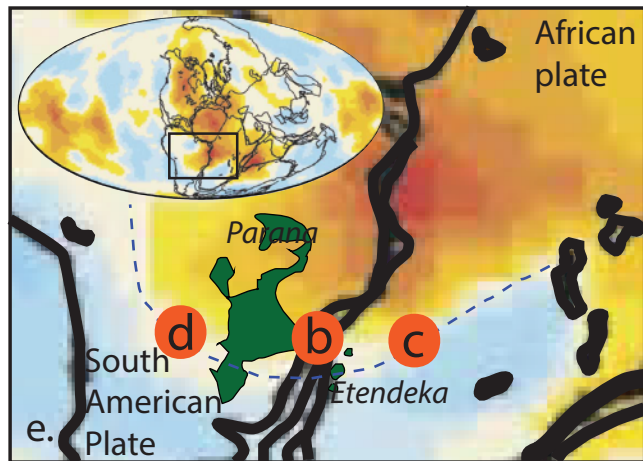
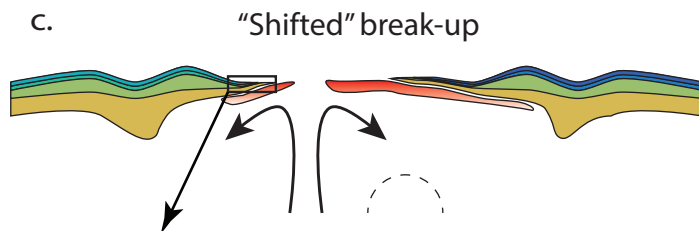
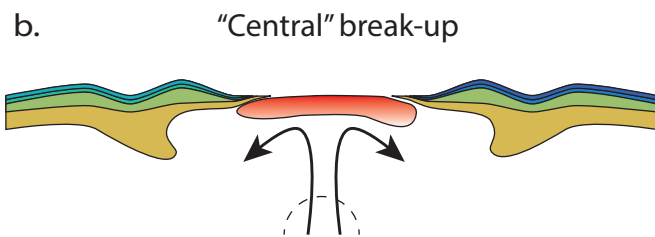
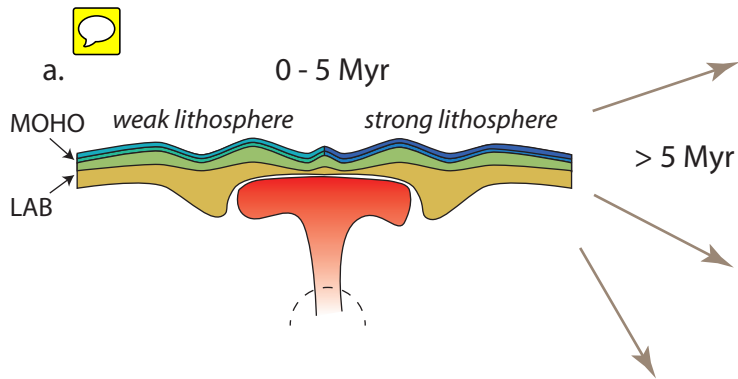


i. Experiment 35 @ 10 Myr









712 **Tables**713 **Table 1**

Thermal parameters	Thermal property	Value	Unit	Ref.
	Surface temperature	10	°C	1
	Temperature at the base of the thermal lithosphere	1330	°C	
	Temperature at the base of the upper mantle	1400	°C	
	Temperature mantle anomaly	1700	°C	
	Thermal conductivity crust	2.5	W/m °C	
	Thermal conductivity mantle	3.5	W/m °C	
	Radiogenic heat production at the surface	1.5E-9	W/kg	
	Radius radiogenic heat	10	km	
	Thermo-tectonic age of the lithosphere	500	myrs	
	Surface heat flow	40	mW/m ²	
	Mantle heat flow	15	mW/m ²	
Mechanical parameters	Mechanical property			
Strong upper crust <i>Dry quartz</i>	density	2600	kg/m ³	2
	Viscosity parameter N	3		
	Viscosity parameter A	6.8e-6	MPa ^{-n/s⁻¹}	
	Viscosity parameter E	1.56e5	J/mol	
Strong lower crust <i>Strong diabase</i>	density	2850	kg/m ³	3
	Viscosity parameter N	3.05		
	Viscosity parameter A	6.3e-2	Mpa ^{-n/s⁻¹}	
	Viscosity parameter E	2.76e5	J/mol	
Weak upper crust <i>Wet quartz</i>	density	2500	kg/m ³	2
	Viscosity parameter N	2.3		
	Viscosity parameter A	3.2e-4	Mpa ^{-n/s⁻¹}	
	Viscosity parameter E	1.54e5	J/mol	
Weak lower crust <i>Weak diabase</i>	density	2750	kg/m ³	4
	Viscosity parameter N	4.7		
	Viscosity parameter A	1.9e2	Mpa ^{-n/s⁻¹}	
	Viscosity parameter E	4.85e5	J/mol	
Lithospheric mantle <i>Peridotite</i>	density	3330	kg/m ³	2
	Viscosity parameter N	3.5		
	Viscosity parameter A	2.5e4	Mpa ^{-n/s⁻¹}	
	Viscosity parameter E	5.32e5	J/mol	
Asthenosphere <i>Olivine</i>	density	3310	kg/m ³	3
	Viscosity parameter N	3.2		
	Viscosity parameter A	7.0e3	Mpa ^{-n/s⁻¹}	
	Viscosity parameter E	5.1e5	J/mol	
Plume <i>Olivine</i>	density	3250	kg/m ³	3
	Viscosity parameter N	3.5		
	Viscosity parameter A	5.e14	Mpa ^{-n/s⁻¹}	
	Viscosity parameter E	5.2e5	J/mol	
	Friction angle	30	°	5
	Lamé elastic constant $\lambda=G$	25	GPa	
	Cohesion	20	MPa	
	Erosion coefficient (a)	500	m ² /yr	

714

715 **Table 2**

Controlling parameters Boundary conditions			Mantle plume properties			Setup	Results				
Extension rate (left)	Extension rate (right)		Plume location	Density (kg/m ³)	Diameter (km)	Initial geometry	Break-up point	Above center anomaly?	Break-up mechanism	Figure	
1	12.5 mm/yr	12.5 mm/yr	800 km	3250	230	Setup 1a	800 km	yes	Central	5a	
2	12.5 mm/yr	12.5 mm/yr	1000 km	3250	230	Setup 1a	1500 km	no	Distant		
3	12.5 mm/yr	12.5 mm/yr	1200 km	3250	230	Setup 1a	1200 km	yes	Central		
4	12.5 mm/yr	12.5 mm/yr	800 km	3250	230	Setup 1b	800 km	yes	Central		
5	12.5 mm/yr	12.5 mm/yr	1000 km	3250	230	Setup 1b	1000 km	yes	Central		
6	12.5 mm/yr	12.5 mm/yr	1200 km	3250	230	Setup 1b	1200 km	yes	Central		
7	12.5 mm/yr	12.5 mm/yr	800 km	3250	230	Setup 2a	850 km	no	Shifted	5e	
8	12.5 mm/yr	12.5 mm/yr	1000 km	3250	230	Setup 2a	800 km	no	Shifted		
9	12.5 mm/yr	12.5 mm/yr	1200 km	3250	230	Setup 2a	1200 km	yes	Central		
10	20 mm/yr	5 mm/yr	800 km	3250	230	Setup 2a	750 km	no	Shifted		
11	20 mm/yr	5 mm/yr	1000 km	3250	230	Setup 2a	800 km	no	Shifted		
12	20 mm/yr	5 mm/yr	1200 km	3250	230	Setup 2a	1000 km	no	Shifted		
13	12.5 mm/yr	12.5 mm/yr	800 km	3250	230	Setup 2b	1600 km	no	Distant		
14	12.5 mm/yr	12.5 mm/yr	1000 km	3250	230	Setup 2b	1450 km	no	Distant		
15	12.5 mm/yr	12.5 mm/yr	1200 km	3250	230	Setup 2b	xxx	xxx	no break-up		
16	20 mm/yr	5 mm/yr	800 km	3250	230	Setup 2b	1200 km	no	Distant		
17	20 mm/yr	5 mm/yr	1000 km	3250	230	Setup 2b	1000 km	yes	Central		
18	20 mm/yr	5 mm/yr	1200 km	3250	230	Setup 2b	1700 km	no	Distant		
19	12.5 mm/yr	12.5 mm/yr	800 km	3250	230	Setup 3a	1200 km	no	Distant		5h
20	12.5 mm/yr	12.5 mm/yr	1000 km	3250	230	Setup 3a	1000 km	yes	Central		
21	12.5 mm/yr	12.5 mm/yr	1200 km	3250	230	Setup 3a	1200 km	yes	Central		
22	20 mm/yr	5 mm/yr	800 km	3250	230	Setup 3a	1200 km	no	Distant		
23	20 mm/yr	5 mm/yr	1000 km	3250	230	Setup 3a	1300 km	no	Distant		
24	20 mm/yr	5 mm/yr	1200 km	3250	230	Setup 3a	1250 km	yes	Shifted		
31	12.5 mm/yr	12.5 mm/yr	800 km	3250	230	Setup 3b	800 km	yes	Central		
32	12.5 mm/yr	12.5 mm/yr	1000 km	3250	230	Setup 3b	1000 km	yes	Central		
33	12.5 mm/yr	12.5 mm/yr	1200 km	3250	230	Setup 3b	1200 km	yes	Central		
34	20 mm/yr	5 mm/yr	800 km	3250	230	Setup 3b	850 km	no	Shifted		
35	20 mm/yr	5 mm/yr	1000 km	3250	230	Setup 3b	1400 km	no	Distant		
36	20 mm/yr	5 mm/yr	1200 km	3250	230	Setup 3b	1700 km	no	Distant		