

# Buoyancy and localizing properties of continental mantle lithosphere: Insights from thermomechanical models of the eastern Gulf of Aden

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- Buoyancy and localizing properties of continental
- mantle lithosphere: Insights from thermomechanical
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X - 2 WATREMEZ ET AL.: BUOYANCY AND PLASTICITY OF LITHOSPHERE Physical properties of the mantle lithosphere have a strong Abstract. 4 influence on the rifting processes and rifted structures. In particular, in con-5 text of rifting, two of these properties have been overlooked: (1) Mohr-Coulomb 6 plasticity (localizing pressure-dependent) may not be valid at mantle depths 7 as opposed to non-localizing pressure-independent plasticity (hereafter, per-8 ect plasticity), and (2) lithosphere buoyancy can vary, depending on the petro-9 logical composition of the mantle. Focussing on the Arabian plate, we show 10 that the lithosphere may be negatively buoyant. We use thermomechanical 11 modeling to investigate the importance of mantle rheology and composition 12 on the formation of a passive margin, ocean-continent transition (OCT) and 13 oceanic basin. We compare the results of this parametric study to observa-14 tions in the eastern Gulf of Aden (heat-flow, refraction seismics and topog-15 raphy) and show that (1) mantle lithosphere rheology controls the margin 16 geometry and timing of the rifting; (2) lithosphere buoyancy has a large im-17 pact on the seafloor depth and the timing of partial melting; and (3) a per-18 fectly plastic mantle lithosphere 20 kg m<sup>-3</sup> denser than the asthenosphere 19 best fits with observed elevation in the Gulf of Aden. Finally, thermomechan-20 ical models suggest that partial melting can occur in the mantle during the 21 Arabian crustal break-up. We postulate that the produced melt could then 22 infiltrate through the remnant continental mantle lithosphere, reach the sur-23 face and generate oceanic crust. This is in agreement with the observed nar-24 row OCT composed of exhumed continental mantle intruded by volcanic rocks 25 in the eastern Gulf of Aden. 26

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#### 1. Introduction

Whether or not a large amount of magmatic activity occurs during rifting has led to 27 two separate classes of rifted passive margins: volcanic or non-volcanic margins. Volcanic 28 margins are commonly associated with high thermal anomalies in the mantle, which are 29 often due to mantle plumes like the Afar Plume in the western Gulf of Aden [e.g. Courtillot 30 et al., 1999] or the Islandic Plume in Greenland and Norway [e.g. Eldholm and Grue, 31 1994]. These are characterized by basalt flows (seaward dipping reflectors), volcanoes, and 32 mafic underplating produced during the rifting process [e.g. Mutter et al., 1982; White 33 and McKenzie, 1989; Geoffroy, 2005]. These margins present a sharp ocean-continent 34 boundary or a very narrow ocean continent transition zone [OCT, e.g. Bauer et al., 2000; 35 M jelde et al., 2007]. 36

Non-volcanic rifted margins are usually compared to the Iberia margin and are char-37 acterized by (1) tectonized features (tilted blocks) and a well identified transitional zone 38 between the continental crust and the oceanic crust, the OCT [e.g. Whitmarsh et al., 1991; 39 Louden et al., 1997; Lavier and Manatschal, 2006, or (2) a wide zone of hyper-extended 40 crust [e.g. Contrucci et al., 2004; Unternehr et al., 2010]. Non-volcanic margins actually 41 contain a limited amount of magmatism and there is now a general agreement on de-42 scribing them as magma-poor margin. The nature of the OCT varies along magma-poor 43 margins [e.g. Leroy et al., 2010a; Gerlings et al., 2011]. It is usually either made of (1) a 44 Zone of Exhumed Continental Mantle [Z.E.C.M., e.g. Manatschal, 2004], (2) deeper man-45 the serpentinized by percolation of the seawater through the sediments and faulted crust [e.g. Boillot et al., 1987; Pérez-Gussinyé and Reston, 2001], or (3) highly tectonized oceanic 47

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crust formed by ultra-slow spreading [e.g. Srivastava and Roest, 1995; Sibuet et al., 2007].
Another type of magma-poor margin, like the Angolan margin, shows hyper-extended
continental crust and evidence of shallow sediments deposition ('sag' basins) during the
late phase of rifting [e.g. Moulin et al., 2005].

The factors that control the amount of melt produced at magma-poor margins, either by decompressional melting during crustal thinning [e.g. *Minshull*, 2002] or post-rift thermal anomaly [e.g. *Lucazeau et al.*, 2009], are yet to be understood. Especially, it is still not always clear how the properties of the mantle lithosphere affect the rift evolution and the production of melt, i.e. the nature – volcanic or not – of the margin.

To answer these types of questions, several studies have shown that comparing me-57 chanical models of rifted margin formation to data and field observations is an effec-58 tive approach [e.g. Brun and Beslier, 1996; Lavier and Manatschal, 2006; Huismans and 59 Beaumont, 2011]. In particular, Huismans and Beaumont [2011] show the importance 60 of a viscous lower crust on the necking of the mantle lithosphere and the geometry of 61 Angolan-type magma-poor margins. The eastern Gulf of Aden has been chosen to tackle 62 the problem of Iberia-type magma-poor margin. Indeed, (1) it is well documented, (2) 63 the conjugate passive margins are known as magma-poor and the width and nature of 64 the OCT vary along the margin [presence of magmatism where the OCT is narrow and 65 exhumed mantle where the OCT is wider, *Leroy et al.*, 2010a], and (3) the oceanic basin 66 is young (oceanic accretion began at least 17.6 Ma ago), so the conjugate margins are 67 easily correlable and numerical modeling of the whole basin is possible. 68

Thermomechanical modeling has been widely used to study the mechanisms of lithoro sphere extension and the parameters governing the geometry of rifted structures (e.g.

influence of the extension rate, Bassi, 1995; Van Wijk and Cloetingh, 2002; Huismans and 71 Beaumont, 2003; Burov, 2007; or of the thermal structure of the lithosphere, Chéry et al., 72 1989). However, beyond the thermal state and ductile strength of the lithosphere, the 73 effects of (1) plastic behaviour of the mantle lithosphere and (2) buoyancy of the litho-74 sphere with respect to the underlying asthenosphere, on the structure of passive margins, 75 are not yet well understood. It is indeed questionable whether the standard brittle failure 76 criterion [Byerlee, 1978] is applicable to mantle lithosphere [e.g. Watts and Burov, 2003; 77 Bürgmann and Dresen, 2008; Burov, 2011] or if weaker mechanisms must be included. 78 Similarly, it has become standard to include phase transition and realistic composition 79 in the crust [e.g. Yamato et al., 2007; Gerya et al., 2008] to study the dynamic of active 80 margins. However, the effect of petrological composition - and especially the extent of 81 depletion – on the buoyancy of mantle lithosphere during continental extension is rarely 82 taken into account, except to explain post rift subsidence anomalies [Kaus et al., 2005]. 83 This well might be an important parameter for rifts, which affect continental lithosphere, 84 such as the Rio Grande rift [Van Wijk et al., 2008], the Baikal rift [Petit and Déverchère, 85 2006], or the Gulf of Aden. 86

In this paper, we therefore explore the effect of two overlooked key parameters on rift evolution: (1) the failure criterion acting in mantle lithosphere and (2) the buoyancy of the lithosphere. Then, we compare the results of this parametric study to the heat-flow measurements and wide-angle seismic models at the northern margin, and elevation data across the eastern Gulf of Aden. This allows us to better understand how these two parameters influence the presence of magmatism at the OCT together with the nature and width of the OCT.

#### 2. Geodynamic settings

The Arabian plate shows a complex history, with the accretion of terranes in the Neo-94 Proterozoic [e.g. Al Husseini, 2000], the occurrence of the Oman obduction and continental 95 subduction during late Cretaceous [e.g. Searle, 1983; Agard et al., 2010] the rifting episode 96 in the Mesozoic [Bosence, 1997]. The formation of the Arabian lithosphere is the result of 97 two major tectonic events [Al Husseini, 2000]: the Amar collision (640 - 620 Ma) and the 98 Najd rifting (530 - 570 Ma). The ages of the basement rocks of Socotra (southern margin 99 of the eastern Gulf of Aden) are 600 to 860 Ma [U-Pb dating, Denele et al., 2012]. In our 100 models, we therefore consider an approximate age of the Arabian lithosphere of 700 Ma, 101 defining it as a tecton [Neo-Proterozoic lithosphere, e.g. Griffin et al., 2003]. 102

The current Gulf of Aden is a young oceanic basin, which separates Arabia from Somalia, 103 accommodating the difference between their plate velocities (Fig. 1). In absolute motion, 104 both plates move to the north. However, the Arabian plate is faster than the Somalian 105 plate [Vigny et al., 2006]. The difference in velocity leads to a present day opening rate of 106 approximately  $2 \text{ cm yr}^{-1}$  in the eastern Gulf of Aden [e.g. Fournier et al., 2001]. The rifting 107 of the Arabian-Nubian tector began  $\sim 34$  Ma ago [e.g. Leroy et al., 2012]. Subsequently, 108 seafloor spreading occurred at the latest at 17.6 Ma in the east of the Shukra-el-Sheik 109 Fracture Zone [e.g. Leroy et al., 2004; d'Acremont et al., 2006; Leroy et al., 2012]. A 110 post-rift high thermal regime is observed in the whole Gulf of Aden and is interpreted as 111 being in relation to the Afar hotspot activity [Lucazeau et al., 2009; Basuyau et al., 2010: 112 Leroy et al., 2010b; Chang and Van der Lee, 2011]. 113

Studies of dynamic topography show that the doming due to the presence of the Afar hotspot does not influence the topography of the eastern Gulf of Aden [e.g. *Lithgow*- <sup>116</sup> Bertelloni and Silver, 1998]. Indeed, the eastern Gulf of Aden is roughly 200 km away <sup>117</sup> from the apex of the hotspot. Furthermore, the wavelengths of the structures of the <sup>118</sup> dynamic topography are greater than 4000 km, as they are due to deep mantle sources <sup>119</sup> [e.g. Hager and Richards, 1989]. Thus, we do not take the dynamic topography into <sup>120</sup> account in this study.

Tomographic studies show that the Arabian-Nubian lithosphere is approximately 250 121 km thick [Ritsema and van Heijst, 2000; Debayle et al., 2001; Pasyanos and Nyblade, 122 2007]. Receiver functions [Sandvol et al., 1998; Al Amri, 1999; Pasyanos and Walter, 123 2002; Al-Damegh et al., 2005; Tiberi et al., 2007; Al-Hashmi et al., 2011; Al-Lazki et al., 124 2012; Ahmed et al., 2013] and refraction seismic profiles [Mooney et al., 1985] constrain the 125 thickness of the Arabian crust, which varies from 35 km (partly thinned crust, close to the 126 Red Sea and the Gulf of Aden) to 49 km (thickened crust, Oman mountains). We adopt a 127 mean crustal thickness of 44 km for our models, corresponding to the average value of non-128 thinned and non-thickened Arabian crust. Geophysical studies on the north-eastern Gulf 129 of Aden magma-poor margin show that the OCT is narrow (15 km to 50 km) and presents 130 strong along-margin variations with serpentinization and little volcanism [Lucazeau et al., 131 2008, 2009, 2010; Autin et al., 2010; Leroy et al., 2010a; Watremez et al., 2011a]. 132

This information will be used both to constrain the initial and boundary conditions of the thermomechanical models presented in section 3 and to compare them with the present structures of the Gulf of Aden. In the parametric study presented here, the aim of the modeling is to better understand the mechanisms of margin formation, such as the north-eastern Gulf of Aden, as well as the nature of their OCT.

#### 3. Parametrization and modeling

#### 3.1. Mantle lithosphere plasticity

The Byerlee [1978] failure criterion is commonly used to represent the yield strength 138 behaviour of rocks. Byerlee's law is written in terms of principal stresses. Hence, it 139 may account for pore-fluid pressure or any other deviations from lithostatic pressure. 140 Extrapolating Byerlee's law at pressures corresponding to depths greater than 40 km 141 implies unrealistically high yield strength for the lithosphere (up to several GPa). Thus, 142 it may not be applicable to the mantle lithosphere [e.g. Watts and Burov, 2003; Précigout 143 et al., 2007; Burov, 2011]. At this confining pressure, other mechanisms may limit the 144 strength of rocks, including compaction bands, grain boundary sliding [GBS, e.g. Préciquet 145 et al., 2007] and Peierls creep [high stress creep for stresses  $> 10^3$  MPa, Kameyama 146 et al., 1999]. All these mechanisms limit the strength of the lithosphere at high deviatoric 147 stresses by reducing the pressure dependance of the strength, and decreasing the ability 148 of the lithosphere to localize strain at high strain-rate (except GBS). Furthermore, field 149 observations of exhumed mantle shear zones show strain localization at all scales and no 150 brittle deformation [see a review in *Préciqout et al.*, 2007; *Bürgmann and Dresen*, 2008]. 151 In contrast, geophysical imaging of active strike slip faults in the mantle suggests that 152 they are broad (>10-100 km wide) zones of distributed shearing [Molnar et al., 1999; 153 Little et al., 2002; Sol et al., 2007]. There is obviously a scale effect in this apparent 154 contradiction. 155

To model the brittle-ductile behaviour of rocks, one can use visco-elasto-plastic rheologies based on Maxwell summation for deviatoric strain-rate ( $\dot{\epsilon}^d$ ):

$$\dot{\epsilon}^d = \dot{\epsilon}^d_{vis} + \dot{\epsilon}^d_{elas} + \dot{\epsilon}^d_{plas} \tag{1}$$

The viscous part  $(\dot{\epsilon}_{vis}^d)$  corresponds to the ductile behaviour of rocks and accounts for deformation by creep. The elastic strain rate  $(\dot{\epsilon}_{elas}^d)$  is very small but is needed to model the path dependance of the rheology. Finally, the brittle part  $(\dot{\epsilon}_{plas}^d)$  uses a plastic flow rule to limit the maximal rock strength when a brittle failure criterion is reached. For Byerlee behaviour, the plastic flow rule follows the *Vermeer* [1990] Mohr-Coulomb plastic model, but is limited to an incompressible flow (zero dilatation).

At high confining pressure, Byerlee's law [1978] shows that maximal brittle strength is 164 proportional to 0.6 P, where P is the total pressure. Within our formulation, we model 165 this experimental law using the cohesion  $C_0 = 20$  MPa and the internal friction angle 166  $\phi = 30^{\circ}$ . If  $\phi$  is not equal to zero, this results in strain localization (shear banding) due to 167 the apparent strain weakening behaviour [Vermeer, 1990; Le Pourhiet, 2013]. Mechanisms 168 such as Peierls [for lithospheric extension, e.g. Popov and Sobolev, 2008] and GBS [e.g. 169 *Précigout et al.*, 2007] creeps are shown to limit rock strength to 400 to 700 MPa [Burov, 170 2011] and to be weakly pressure dependent. Using the creep parameters of *Hirth and* 171 Kohlstedt [2003] and considering the mean strain rate in the rift area of  $10^{-14}$  s<sup>-1</sup>, a 172 mantle temperature of  $650^{\circ}$ C, corresponding to depths of approximately 100 km and an 173 average grain size of 100  $\mu$ m, the value of the GBS yield stress is approximately 400-500 174 MPa. Thus, we roughly approximated these mechanisms with  $C_0 = 450$  MPa and  $\phi = 0^{\circ}$ . 175 As a result, the lithosphere behaves as a perfectly plastic material and no significant 176 localization, without an additional softening law, is predicted to occur. This mechanism 177 is also known as "stress limiter" [van Hunen et al., 2002]. 178

Thus, while Byerlee's parameters predict a large strength peak at  $\sim 90$  km depth (more than 1 GPa), the maximal strength in the perfectly plastic mantle lithosphere is constant and limited to a much smaller value (in this study, 450 MPa) for the whole depth interval in the mantle lithosphere, up to 100 km depth (Fig. 2).

#### 3.2. Lithosphere buoyancy

In order to get direct information on the buoyancy of the Arabian mantle lithosphere, 183 density profiles have been calculated for the chemical compositions of Arabian plate mantle 184 xenoliths available in the literature. These calculations have been carried out using the 185 free energy minimization thermodynamic code PERPLEX07 [Connolly, 2009], together 186 with a recent thermodynamic database and activity models developed for mantle pressure 187 and temperature conditions and compositions [Xu et al., 2008]. Four average compositions 188 of tecton subcontinental mantle lithosphere [Griffin et al., 2009] were also considered as 189 standard compositions for lithospheres having similar age as the Arabian plate. Since the 190 composition of the asthenosphere below the Arabian plate is unknown, we used the fertile 191 Hawaiian pyrolite composition [Wallace and Green, 1991]. This standard is used as a 192 proxy for the density of the asthenosphere influenced by the Afar hot-spot. The buoyancy 193 is calculated as the density difference between asthenosphere and mantle lithosphere: 194  $\rho_{asth} - \rho_{lith}$ . The dataset of chemical compositions, the description of the method and the 195 density curves are presented in the Supplementary material. 196

The chemical composition of the xenoliths and the standards are plotted in a Mg# vs. %Al<sub>2</sub>O<sub>3</sub> diagram (Fig. 3A). The xenoliths from the Arabian plate are divided into two chemical groups. The depleted group (lherzolites and harzburgites) plot in the depleted part of the tecton xenoliths (dotted red zone on Fig. 3A). The corresponding compositions <sup>201</sup> are also more depleted than the tecton and asthenosphere standards. The fertile group <sup>202</sup> (clinopyroxenites and websterites) shows greater scatter.

The clinopyroxenites and websterites are partly [Henjes-Kunst et al., 1990] or entirely 203 [Stein et al., 1993] interpreted as cumulates formed in Pan-African times. These authors 204 do not explicitly specify the geodynamic context for the formation of these rocks. Com-205 parison with modern magmatic arcs [e.g. Saleeby et al., 2003] suggests that they formed 206 as cumulates below arcs, which is consistent with the accretion of the Arabian shield 207 during the end of the Pan-African orogeny [Frisch and Al-Shanti, 1977; Hargrove et al., 208 2006. These rocks could also correspond to recycled subducted oceanic crust that has 209 been incorporated into the lithosphere during cooling of the mantle [Hofmann and White, 210 1982]. Amounts of recycled oceanic crust in the asthenosphere range mainly between 10%211 and 20% and reach up to 30% [Sobolev et al., 2007]. 212

The calculated depth profiles of buoyancy are presented in Fig. 3B. Chemical variations are directly reflected by the buoyancy. The more depleted is the composition (low % of Al<sub>2</sub>O<sub>3</sub> and high Mg#), the greater is the buoyancy. Lherzolites and harzburgites are buoyant or have a slightly negative buoyancy, whereas all clinopyroxenites and the websterites have a very negative buoyancy. Mean buoyancies averaged over the lithosphere are approximately 20 kg m<sup>-3</sup> for the lherzolite-harburgite group, -100 kg m<sup>-3</sup> for the clinopyroxenite-websterite group and 5 kg m<sup>-3</sup> for the tecton standards.

The Arabian plate mantle lithosphere is a mixing of depleted and fertile rocks, both groups having different buoyancies. The buoyancy of the whole lithosphere can therefore be estimated by a weighted average of the buoyancy of these two end-members. In order to take into account the complex history of the Arabian plate, we propose two models

(Fig. 3C). In the first model, the depletion of lherzolites and harzburgites is inherited from 224 partial melting coeval with the formation of the Arabian shield [Stein et al., 1993]. The 225 depleted end-member corresponds to the mean of the lherzolite-harzburgite group. The 226 fertile end-member corresponds to the mean of the clinopyroxenite-websterite group (blue 227 line on 3C). A mantle lithosphere composed of 35% clinopyroxenite-websterite and 65%228 lherzolite-harzburgite has a negative buoyancy of  $\sim 20 \text{ kg m}^{-3}$ . In the second model, the 229 depletion is due to Miocene partial melting that postdates the onset of rifting [McGuire, 230 1988]. The depleted end-member is the mean of the tecton standards and the fertile 231 end-member corresponds to the mean of the clinopyroxenite-websterite group (red line 232 on 3C). 25% clinopyroxenite-websterite is sufficient to explain a negative buoyancy of 233  $\sim 20$  kg m<sup>-3</sup>. A few 10s kg m<sup>-3</sup> negative buoyancy can therefore be explained by a 234 significant proportion ( $\sim 30\%$ ) of heavy clinopyroxenite and websterite. Using a different 235 as the nosphere composition indicates a difference of  $\sim 10 \text{ kg m}^{-3}$ . Hence, a reduction of the 236 amount of clinopyroxenite and websterite by 15-25% would induce an average buoyancy 237 of -20 kg m<sup>-3</sup>. In the parametric study, we apply a  $\pm 30$  kg m<sup>-3</sup> buoyancy in order to test 238 extreme parameters. 239

#### 3.3. Parametrization and numerical modeling

We use Flamar 12 [an outgrowth of Paravoz, *Burov and Yamato*, 2008] that is inspired by the FLAC v3 algorithm [*Cundall*, 1989; *Poliakov et al.*, 1993]. The numerical method has been described in full detail in previous studies [e.g. *Burov and Poliakov*, 2001; *Burov and Cloetingh*, 2009; *François et al.*, in press]. This method can handle a free upper surface boundary condition and almost any visco-elasto-plastic rheology. The algorithm explicitly takes into account elastic-brittle-ductile properties of lithosphere and asthenosphere. Due to the explicit nature of this code for elastic compressibility, temperature, gravity and pressure-dependent body forces are computed without the necessity to use assumptions such as potential temperature. The description is limited here to the details that are the most essential for this study.

Our model setup is inspired by previous rifting models achieved with the same numerical 250 thermomechanical code [Burov, 2007, Fig 2]. The modeling box is 400 km thick and 1200 251 km wide; grid resolution is 4 km in both directions. The properties of each element of 252 the numerical grid (cell) are defined by the density and the thermal and elasto-visco-253 plastic parameters of its material. We use a density and rheology structure from the 254 lithosphere and asthenosphere derived from Ranalli [1995]. All models include a 44 km 255 thick crust with four horizontal rheological layers. The parameters of the reference model 256 are presented in Table 1. Materials in the specific numerical formulation used for the 257 study are elasto-visco-plastic. To avoid activation of Mohr-Coulomb plastic deformation 258 in the asthenosphere [which is unlikely, e.g. Watts and Burov, 2003], we assigned high 259 cohesion values to the asthenosphere. 260

The boundary conditions are (1) horizontal extension velocities  $(V_{ext})$  applied at each 261 vertical side of the model, (2) a Winckler pliable basement that simulates the response to 262 lithostatic pressure variations and (3) a free upper surface boundary condition (free stress 263 and free slip conditions in all directions) combined with a moderate diffusion erosion 264 and sedimentation  $(k_{eros} = 50 \text{ m}^2 \text{ yr}^{-1})$  and water-column pressure dependent on the 265 surface topography (Fig. 2). The elevation at each point of the free surface is calculated 266 following the model of *Culling* [1960], which takes into account the effect of erosion and 267 sedimentation. 268

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The assumed initial thermal structure is constrained by the radiogenic heat production in the crust, the age the lithosphere and the temperature at the base of the lithosphere. It is noteworthy that, after 600 Ma, the thermal structure of thermally undisturbed lithosphere reaches a permanent state. Thus, there is no practical difference in the thermal structure of a 600 Ma or, for example, a 2000 Ma old lithosphere [e.g. *Burov and Diament*, 1995]. The thermal age in the models is 700 Ma, to be consistent with the real age of the Arabian shield.

The main parameters tested in this study are the plastic failure pressure dependence and the buoyancy of the mantle lithosphere.

#### 3.4. Areas of possible partial melting

Partial melting is not supposed to have a strong impact on the overall rift evolution 278 of the Eastern Gulf of Aden, as its rifted margins are magma-poor. The lack of a small 279 production of melt at break-up is therefore a non-negligible constraint on the validity of 280 the models. Some codes include partial melting in the modeling, in order to simulate 281 mantle convection [e.g. ONeill et al., 2006]. However, for the sake of simplicity, it is 282 not necessary to directly include partial melting calculations. We follow the simplified 283 approach developed by *McKenzie and Bickle* [1988], which consists in calculating the 284 areas where partial melting can occur. We compute the location of these areas at the 285 post-processing stage. The method uses the model-predicted pressure and temperature 286 conditions in the mantle as input, and output a melt fraction. We consider the areas 287 where the melt fraction [defined in *McKenzie and Bickle*, 1988] is between 0 and 1. This 288 post-processing allows us to assess whether melt can be produced during the rifting and 289 to characterize the nature of the OCT in the model. 290

#### 4. Parametric study

#### 4.1. Influence of mantle lithosphere rheology

Figure 4 compares two models with different mantle rheology and highlights the influence of this parameter on rifting evolution and rifted margins formation.

As expected from the rheology law, deformation is more diffuse in the perfectly plastic 293 mantle than in a Byerlee's mantle. Meanwhile, strain localization is also enhanced by the 294 higher strength of a Byerlee's lithosphere and the imposed velocity boundary condition. 295 This lack of localization results in delaying the break-up of continental crust by 1.5 Myr 296 for the model with the perfectly plastic mantle. As the strain localization is more in-297 tense in the Mohr-Coulomb mantle lithosphere, deformation propagates faster through 298 the continental crust. Moreover, the strain rate patterns 0.5 Ma before the break-up of 299 the continental crust show an asymmetric deformation in the upper mantle beneath the 300 continental rift for the Mohr-Coulomb model, while the perfectly plastic model remains 301 symmetric (Fig. 4A). 302

Crustal flexure (shape of the continental crust aside the rift) – and hence flexural stresses 303 at the time of crustal break-up is more intense for the Mohr-Coulomb model than for 304 the perfectly plastic model, contributing to an earlier break-up for the former (Fig. 4A). 305 As can be seen the local extrema of topography coincide in either case, producing close 306 values of flexural wavelength ( $\sim 250$  km) and, hence, of elastic thickness (Fig. 4B). The 307 corresponding equivalent flexural thickness estimated using analytical formulae for the 308 flexural parameter [Turcotte and Schubert, 2002] is on the order of 40 km. This wavelength 309 is largely controlled by the strongest part of the mantle lithosphere [Burov and Diament, 310 1995]. The difference in amplitude at rift flanks refers to smaller levels of flexural stress in 311

the crust in case of a less localizing, perfectly-plastic rheology. We observe that no partial melting can occur in the mantle lithosphere for either model, suggesting that continental mantle can be exhumed without production of oceanic crust in either case.

In order to compare the models with the present day Gulf of Aden, we compare their 315 geometries and bathymetries 18 Ma after break-up. The geometry of the structures at 316 this moment also shows a slightly stronger flexure for the Mohr-Coulomb model than 317 for the perfectly plastic model, while conjugate margins are still asymmetric (Fig. 4A). 318 The slope of the Moho discontinuity is also steeper for the Mohr-Coulomb model than 319 for the perfectly plastic model, implying a wider distal margin with the perfectly plastic 320 configuration. Moreover, the ridge location shows a strong asymmetry in the Mohr-321 Coulomb model, contrary to the perfectly plastic model. This can be attributed to the 322 fact that a Mohr-Coulomb mantle localizes the deformation better, allowing for further 323 strain to occur on the fault formed along the margin at the break-up of the continental 324 crust. Comparison of the topography 18 Ma after continental crust break-up provides a 325 similar amplitude of the topography for both models, but the same asymmetry is recovered 326 as the oceanic ridge is highlighted by a deeper zone along the model (Fig. 4B). We also 327 observe the development of this asymmetry on the evolution of the topography (distance 328 versus time, Fig. 4C). Indeed, after continental break-up, the ridge tends to be localized 329 close to the margin during approximately 10-12 Myr for the Mohr-Coulomb model. The 330 ridge remains stable until the fault zone reaches a hardening sufficient to cause a relocation 331 of the ridge. On the contrary, the ridge in the perfectly plastic model is centered from 332 the onset of the formation of the oceanic basin. This difference is due to the fact that 333 the perfectly plastic mantle has a more diffuse deformation mechanism than the Mohr-334

<sup>335</sup> Coulomb mantle. The model with the perfectly plastic mantle also shows an oceanic <sup>336</sup> ridge that is centered in the same way as observed in the Gulf of Aden, giving additional <sup>337</sup> preference to the perfectly plastic model over the Mohr-Coulomb model.

It can be concluded that the choice of the failure criterion in the mantle lithosphere has 338 a significant impact on the extensional deformation in the upper mantle, and indirectly 339 in the continental crust. Compared to the Mohr-Coulomb mantle, the assumption of a 340 perfectly plastic mantle allows for a more diffuse deformation, with the following conse-341 quences: (1) more extension is needed to break the crust apart, (2) the flexure is more 342 gentle, and (3) the horizontal position of the oceanic ridge, as observed in the Gulf of 343 Aden, is more consistent with observation. However, oceanic basins predicted by both 344 models are deeper than the observed topography profiles across the Gulf of Aden. Hence, 345 additional factors such as mantle density should be explored to explain the shallowness 346 of the oceanic basin of the Gulf of Aden. 347

#### 4.2. Influence of lithosphere buoyancy

The three model-setups are identical to the Mohr-Coulomb setup previously presented (Table 1), except for the density of the asthenosphere, in order to test the influence of the buoyancy of the lithosphere on the rifting and the formation of rifted structures (Fig. 5). Thus, we compare three series of experiments with different lithosphere buoyancy (see paragraph 3.2 for definition). The three values of buoyancy tested here are -30, 0 and +30 kg m<sup>-3</sup>. The model with a positive buoyancy (30 kg m<sup>-3</sup>) is the same as the Mohr-Coulomb configuration described in paragraph 4.1. Our first observation is that the variation of the buoyancy of the lithosphere does not affect the timing of the break-up of the continental crust. Indeed, this break-up occurs 7 Ma after the beginning of the extension for all three models.

The strain rate patterns 0.5 Ma before the break-up of continental crust show an asymmetry beneath the center of the rift zone for the two models with non-zero buoyancy (Fig. 5A). This is probably due to the fact that the zero lithosphere buoyancy reduces the mechanical contrast between the mantle lithosphere and the asthenosphere, avoiding asymmetric processes to occur at this stage.

The geometry of the models at the moment of the continental crust break-up shows 363 a shallower boundary between mantle lithosphere and asthenosphere, as asthenosphere 364 density is lower than for the other tested models (Fig. 5A). This is simply due to the 365 fact that the lighter the material is, the more rapidly it uplifts. We observe that a zone 366 of possible partial melting occurs at the center of the rift for the model with a negative 367 buoyancy (-30 kg m<sup>-3</sup>). Indeed, in this model the 1300°C isotherm (approximately the 368 base of the lithosphere) is shallow enough at the time of break-up of the continental 369 crust to allow for partial melting. This implies that oceanic crust can be produced at 370 this time, while magmatism may occur earlier, producing underplated and/or intruded 371 magma, as the continental crust is not yet broken apart. This process is characteristic of 372 the formation of volcanic passive margins. 373

Model geometries 18 Ma after the break-up of the continental crust show asymmetric margins in all cases. We observe that the lateral position of the oceanic ridge is more centered when buoyancy is negative (Fig. 5A-B, see also topographic profiles). Indeed, predicted evolution of the topography shows that the oceanic ridge relocates after 10<sup>376</sup> 12 Ma for the model with positive buoyancy and after 6-7 Ma for the model with zero <sup>379</sup> buoyancy. The model with negative buoyancy shows an early relocation at the beginning <sup>380</sup> of the formation of the oceanic basin (Fig. 5C). We also observe that the depth of the <sup>381</sup> oceanic basin increases when asthenosphere is denser, which is consistent with the isostasy <sup>382</sup> effect, and that the model with a negative lithosphere buoyancy provides a better fit to <sup>383</sup> the present depth of the Gulf of Aden oceanic basin.

Thus, the buoyancy of the lithosphere controls (1) the depth of the oceanic basin, (2) the position and the relocation of the oceanic ridge in the basin, and (3) the volcanic or magma-poor nature of the rifted margins. A negative buoyancy of the lithosphere allows for a depth of the oceanic basin comparable with observations in the eastern Gulf of Aden.

#### 5. Discussion

#### 5.1. Comparison to the eastern Gulf of Aden

The choice of parameters for this model is governed by the results of the parametric 388 study (failure criterion in the mantle lithosphere and buoyancy of the lithosphere) and the 389 geodynamic context (kinematics). Thus, we use a perfectly plastic mantle and a negative 390 lithosphere buoyancy of  $-20 \text{ kg m}^{-3}$ , which is petrologically more realistic than the value 391 of  $-30 \text{ kg m}^{-3}$  used in the parametric study. Then, we modify the boundary conditions 392 to better fit the actual kinematics (see Section 2): a velocity of 0.5 cm  $yr^{-1}$  is applied on 393 the left side (South), while a velocity of  $1.5 \text{ cm yr}^{-1}$  is applied on the right side (North), 394 to keep a total opening velocity of 2 cm  $yr^{-1}$ . The small initial thermal anomaly used 395 to localize the rift is set closer to the left side of the box, to ensure that the rift is at 396 the center of the box at the final stage (18 Ma after crustal break-up), comparable to the 397 present-day Gulf of Aden. 398

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We compare the model at 18 Ma after the break-up of the continental crust to the data 399 available at the same scale in the eastern Gulf of Aden (Fig. 6). First, heat-flow at the 400 surface of the model is similar to heat-flow measurements for large wave-lengths on the 401 northern margin [Fig. 6A, Lucazeau et al., 2008]. Next, geometry of the crustal thinning 402 is comparable with refraction and wide-angle reflection seismic velocity model along the 403 same line [Fig. 6B, Leroy et al., 2010a]. However, the crude resolution of the model does 404 not allow for a more detailed match with the observed structures in the sediments and 405 crust. Finally, the topography of the model shows the same amplitude as profiles across 406 the eastern Gulf of Aden, even though the morphology of the rift shoulders is steeper than 407 the observation (Fig. 6C). The oceanic basin is also slightly deeper in the model than in 408 the data. In the eastern Gulf of Aden, higher than expected mantle heat-flow and lower 409 than expected seismic velocity and density values, inferred from wide-angle seismic and 410 gravity modeling, might be explained by a persistent post-rift thermal anomaly [Lucazeau 411 et al., 2008; Watremez et al., 2011a]. This topographic discrepancy might be explained by 412 the high thermal regime affecting the entire Gulf [e.g. Lucazeau et al., 2008, 2009, 2010; 413 d'Acremont et al., 2010; Leroy et al., 2010b]. 414

We also observe that pressure and temperature conditions in the asthenosphere at the center of the rift allow for partial melting at the exact moment of the break-up of the continental crust (Fig. 6B). The melt produced can infiltrate toward the surface and either (1) become trapped beneath the margin (underplating, magmatism) to form a volcanic margin and/or (2) generate oceanic crust, leaving very little time to serpentinize or exhume continental mantle, forming a magma-poor margin. This is consistent with the north-eastern Gulf of Aden margin that exhibits a very narrow OCT and even some magmatism [Autin et al., 2010; Leroy et al., 2010a; Watremez et al., 2011a; Leroy et al.,
2012], which may have triggered the continental break-up, as shown by Bronner et al.
[2011] for the Iberia margin.

#### 5.2. Perfectly plastic failure criterion

The parametric study shows that a perfectly plastic mantle lithosphere allows for stronger ridge relocation than a Mohr-Coulomb lithosphere, resulting in a ridge topography more comparable to the present Gulf of Aden morphology.

The choice of the failure criterion in the mantle governs the deformation in the mantle lithosphere and then the deformation in the continental crust. The perfectly plastic criterion makes deformation more diffuse in the lithosphere, while deformation is more localized with the Mohr-Coulomb mantle (narrower shear zones), using the same extension velocity (or strain rate). This explains why the break-up of the continental crust occurs earlier in the case of the Mohr-Coulomb mantle model.

These results are compatible with recent rheological and modeling studies [e.g. 434 Kameyama et al., 1999; Watts and Burov, 2003; Popov and Sobolev, 2008, which propose 435 that a Mohr-Coulomb rheology (in the form of Byerlee's frictional plastic law) is less suit-436 able for localized deformation of rocks in a high stress regime, especially in extension, than 437 Peierls or GBS rheologies. An alternative model advances the potential role of diking that 438 is supposed to weaken the lithosphere prior to rifting [e.g. Buck, 2004]. However, pre-rift 439 diking would lead to extensive pre-rift magmatism and formation of volcanic margins, 440 excluding the possibility to form a magma-poor rifted margin such as the eastern Gulf of 441 Aden. 442

#### 5.3. Negative buoyancy of the continental lithosphere

The parametric study shows that the buoyancy of the mantle lithosphere has a strong influence on the volcanic or magma-poor nature of the modeled rifted margin. Indeed, a highly negatively buoyant mantle lithosphere (-30 kg m<sup>-3</sup>) leads to the formation of melt in the asthenosphere prior to crustal breakup, as observed along volcanic margins. On the contrary, a greater buoyancy leads to the formation of typical magma-poor margins, with a zone of exhumed continental mantle.

The comparison between observed and modeled Gulf of Aden topography suggests that the mantle lithosphere of the Arabian shield has a negative buoyancy of approximately -20 kg m<sup>-3</sup>. Such values have also been proposed for the Baikal rift [*Gao et al.*, 1994; *Petit and Déverchère*, 2006]. Moreover, thermodynamical modeling of the density for a large mantle rock database has shown that mantle lithosphere of tectons can have slightly negative buoyancy for conductive geotherms [*Griffin et al.*, 2009].

The assumption of a subcontinental lithosphere composition close to lherzolitic leads to a positive buoyancy in the case of the Arabian shield (Fig. 3B), which is not in agreement with the result of the models (Section 4.2). However, the heterogeneity and the complex history of the mantle lithosphere are attested by many studies [*Kuo and Essene*, 1986; *McGuire*, 1988; *Stein et al.*, 1993; *Nasir and Safarjalani*, 2000; *Kaliwoda et al.*, 2007; *Shaw et al.*, 2007].

<sup>461</sup> One can doubt the long-term stability of a continental lithosphere with negative buoy-<sup>462</sup> ancy with respect to the asthenosphere. In this case, we expect Rayleigh-Taylor instabili-<sup>463</sup> ties after 300-400 Ma, which lead to the collapse of the continental lithosphere [e.g. *Burov* <sup>464</sup> and Watts, 2006]. However, *Lenardic and Moresi* [1999] show that a positive buoyancy of the lithosphere is not sufficient to create conditions for its long term stability. This
stability also depends on the strength of the lithosphere, an effective coupling between
the very buoyant crust and the mantle, as well as the occurrence of compressive stresses
[*François et al.*, in press]. The Arabian plate actually experienced such significant compressive stresses due to successive subductions [*Al Husseini*, 2000], potentially explaining
why it did not collapse.

Similarly to our work, former observations and numerical models show that a negative lithosphere buoyancy should lead to the development of Rayleigh-Taylor instabilities at the mantle-asthenosphere boundary during the rifting, resulting in removal of an essential amount of lithosphere mantle [e.g. *Burov*, 2007; *Van Wijk et al.*, 2008]. This has a strong influence on the rifted structures.

#### 6. Conclusions

The results of our modeling study show that the structure of the rifted continental margins is highly sensitive to the failure criterion of the mantle lithosphere while the buoyancy of the mantle lithosphere controls the depth of the oceanic basin and the relative timing of the partial melting. The model using a perfectly plastic rheology and negatively buoyant mantle lithosphere (-20 kg m<sup>-3</sup>) shows similar features as the eastern Gulf of Aden (heat-flow, topography, thinning of the continental crust, narrow OCT and position of the ridge).

The perfectly plastic behavior represents a proxi for low temperature plasticity [e.g. *Kameyama et al.*, 1999] whereas Mohr-Coulomb rheology produces higher stress and stress drop, which best represents brittle behaviour at geodynamic time-scale. In the case of the Gulf of Aden, structures are best fitted with a maximum yield strength of 450 MPa, which

is consistent with laboratory derived low temperature plasticity mecanism such as GBS 487 and Peierls creep [Kameyama et al., 1999; Précigout et al., 2007]. Popov and Sobolev [2008] 488 study, based on 3D model of continental rifting, argued that Mohr-Coulomb required 489 unrealistic stress build-up during continental rifting. Here, we reach similar conclusions 490 comparing our continental break-up model to the structures of the Gulf of Aden. Not 491 only, stress build-up is too strong, causing the formation of far too high rift shoulders 492 during continental rifting, but Mohr-Coulomb rheology also induced non realistic post-493 rift topography and extremely long lasting initial phase of asymmetric spreading (10-12) 494 Myr), which is neither observed in Aden nor anywhere else in world. 495

The Gulf of Aden represents the tipping point between the formation of (1) volcanic 496 rifted margins, where partial melting possibility occurs before the crustal break-up and 497 oceanic crust is created as soon as the continental crust is broken apart by dikes, and 498 (2) magma-poor rifted margins, where partial melting occurs after crustal break-up and a 499 zone of exhumed continental mantle forms between the last continental crust and the first 500 oceanic crust [e.g. White and McKenzie, 1989; Whitmarsh et al., 1991; Bauer et al., 2000]. 501 Our parametric study shows that negatively buoyant lithosphere favors early occurrence 502 of melting, whereas neutrally to positively buoyant mantle lithosphere causes the melting 503 to be delayed by up to 1.5 Myr [Watremez et al., 2011b]. Therefore, we posit that mantle 504 buoyancy variation could control the timing of melting and oceanization observed over a 505 wide range of passive margins. In the case of the Gulf of Aden, the first appearance of 506 partial melting in the rift is synchronous with the crustal break-up, which is shown to 507 correspond to a lithosphere that is negatively buoyant by a contrast of  $-20 \text{ kg m}^{-3}$ . This 508 density contrast is confirmed by the petrological study of xenolith of the Arabian mantle 509

<sup>510</sup> lithosphere and has also been shown to be a decisive factor for the occurrence of oceanic
<sup>511</sup> obduction in North Oman using different boundary condition and numerical approach
<sup>512</sup> [Duretz et al., 2013].

This panel of evidences implies that the negatively buoyant mantle lithosphere of the 513 Arabian plate has been stable for several hundreds of million years. It rises the question 514 of the dynamic forces, which could be responsible for the long lasting stability and rapid 515 destabilization of old continental lithosphere at the onset of rifting. Further thermome-516 chanical modeling of rifting, including petrological data to estimate the buoyancy of the 517 lithosphere and comparison to other geological region such as the Rio Grande Rift [Song 518 and Helmberger, 2007; Van Wijk et al., 2008], are needed to better constrain the links 519 between partial melting and the rheology and buoyancy of continental mantle lithosphere 520 highlighted by our study. 521

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#### Figure 1. Geodynamic settings of the Gulf of Aden.

Arrows show GPS vectors considering the Eurasian plate fixed [from Vigny et al., 2006]; numbers are the velocities in mm/yr. Black dots are epicentres of earthquakes (USGS database from 1973/01 to 2011/04,  $M_W \geq 3.5$  - http://earthquake.usgs.gov/earthquakes/ eqarchives/epic/). Hatched areas correspond to the Arabo-Nubian shield. Abbreviations: AFTZ, Alula-Fartak Transform Zone; AR, Aden Ridge; CR, Carlsberg Ridge; EAR: East African Rift; OFZ, Owen Fracture Zone; SHTZ, Socotra-Hadbeen Transform Zone; SR, Sheba Ridge; SSFZ, Shukra-el-Sheik Fracture Zone. The direction of extension of the Gulf of Aden is highlighted by the orientation of the transform zones. Relief is compiled from SRTM topography data [Farr et al., 2007] and gravity-predicted bathymetry [Sandwell and Smith, 1997].

#### Figure 2. Model setup.

An extension velocity of 1 cm/yr is applied on each side of the model. The springs represent the lithostatic pressure (Winkler basement) that is applied at the base of the model. The strength envelopes (blue and red curves) compare the different rheological behaviour in depth for the two different mantle failure criteria in the upper 150 km of the model. The black curve presents the evolution of the temperature with depth in the upper 150 km of the model. A 50°C thermal anomaly is applied at the base of the crust to localize the deformation at the center of the model.

#### Figure 3. The Arabian plate mantle lithosphere buoyancy.

A. Composition of xenoliths from the Arabian plate mantle lithosphere (see Supplementary material for the references), of subcontinental mantle lithosphere standards from four tectons [*Griffin et al.*, 2009] and of the fertile Hawaiian pyrolite standard [*Wallace and Green*, 1991] plotted in a Mg# vs. %Al<sub>2</sub>O<sub>3</sub> diagram (Mg#=100 Mg/(Mg+Fe)). Also shown is the range of the tecton xenolith suites [*Griffin et al.*, 2009]. Two groups of xenoliths can be separated according to their chemistry. Lherzolites (lhz) and harzburgites (haz) have low Al<sub>2</sub>O<sub>3</sub> content and high Mg#. In spite of the scattering, clinopyroxenites (cpx) and websterites (web) have high Al<sub>2</sub>O<sub>3</sub> content and low Mg#.

B. Buoyancy calculated for the compositions of the xenoliths and the tecton standards along the initial thermal profile of the model (the buoyancy is calculated relative to the asthenosphere). The rapid buoyancy decrease between 50 km and 80 km shown by several curves of the cpx-web group corresponds to the plagioclase-spinel-garnet transitions.

C. Calculation of the Arabian plate mantle lithosphere buoyancy as a mixing of depleted and fertile rocks. The depleted fractions correspond to either the mean lhz-haz group (mean buoyancy ~20 kg m<sup>-3</sup>) or the mean of the tecton standards (mean buoyancy ~5 kg m<sup>-3</sup>). The fertile fraction corresponds to the cpx-web group (mean buoyancy ~-100 kg m<sup>-3</sup>). The buoyancy decreases with increasing fraction of cpx-web. The density of the "best fit model" (-20 kg m<sup>-3</sup>, see section 5.1) can be achieved for a mantle lithosphere containing approximately 30% clinopyroxenite and websterite.

#### Figure 4. Influence of the mantle lithosphere rheology.

The two columns correspond to the results of the two models comparing mantle lithosphere rheology.

A. The first line shows the strain rate in the models 0.5 Ma before the break-up of the continental crust, respecting the color scale shown below. The second and third lines show the geometry of the model at the time of the break-up of the continental crust and 18 Ma later, respectively. The color code for the materials is the same as in Figure 2. The black arrows show the position of the oceanic spreading ridge.

B. Topography of the models at the moment of crustal break-up and 18 Ma afterwards compared with the present-day topography of the Gulf of Aden.

C. Evolution of the model topography through time: at each time line correspond a 2D topographic line. Left panel: Mohr-Coulomb model . Right panel: Perfectly-plastic model (following the columns of panel A). The higher elevations (dark red) show the positions of the rift shoulders through time while the narrow green-to-blue zone shows the position of the oceanic ridge. The black circles highlight the places where and times when the oceanic ridge relocates (ridge jumps). The black horizontal lines correspond to the moment of crustal break-up.

#### Figure 5. Influence of the mantle lithosphere buoyancy.

The three columns correspond to the results of the three models comparing the buoyancy of the lithosphere. Organization of this figure is similar to Figure 4.

#### Figure 6. Comparison with the eastern Gulf of Aden.

We compare available information with the model 18 Ma after crustal break-up.

A. Comparison of the heat-flow of the model with the heat-flow profile across the northern margin [*Lucazeau et al.*, 2008]. The blue dots are heat-flow values corrected for sedimentation, topography and refraction and the green line is the modeled heat-flow.

B. Comparison of the geometry of the model with a refraction and wide-angle reflection seismic profile coincident with the heat-flow measurements in part A of this figure [Leroy et al., 2010a]. The grey box presents the geometry of the model and the possibility of partial melting in the mantle at the moment of crustal break-up. Z.E.C.M. means zone of exhumed continental mantle. C. Comparison the model topography with data along 4 profiles [Sandwell and Smith, 1997; Farr et al., 2007]. The positions of the profiles are shown on the map at the right of the figure. The red line corresponds to the position of the heat-flow/refraction line presented in A and B.

### Table 1. Reference model parameters.

Parameters of the reference model having a Mohr-Coulomb failure criterion in the mantle lithosphere and equal densities in the mantle lithosphere and asthenosphere. Upper and lower crustal

Parameters	Values						
Half extension velocity, $V_{ext}$	$1 \times 10^{-2} {\rm m}$	yr <sup>-1</sup>					
Temperature at the base of the lithosphere	1300 °C						
Radioactive heat production, $H_s$	$4.5 \text{x} 10^{-10} \text{ W kg}^{-1}$						
Radiogenic production decay length, $h_r$	$20 \mathrm{km}$						
Crust thermal conductivity, $k_c$	$2.5 \text{ W K}^{-1}$	$\mathrm{m}^{-1}$					
Mantle thermal conductivity, $k_m$	$3.3 \text{ W K}^{-1} \text{ m}^{-1}$						
Specific heat, $C_p$	$1000 \text{ J K}^{-1} \text{ kg}^{-1}$						
Thermal age 700 Ma							
	Upper	Lower	Mantle	Asthonosphoro			
	$\operatorname{crust}$	$\operatorname{crust}$	lithosphere	Asthenosphere			
Thickness (km)	24	20	206				
Density, $\rho$ (kg m <sup>-3</sup> )	2750	2900	3330	3360			
Initial scaling factor, A ( $Pa^{-n} s^{-1}$ )	$2.0 \times 10^{-4}$	$8.0 \times 10^{-3}$	$1.0 \times 10^4$	$1.0 \times 10^4$			
Power-law exponant, n	1.9	3.1	3.0	3.0			
Activation energy, $Q (J \text{ mol}^{-1})$	$137 \times 10^{3}$	$243 \times 10^{3}$	$520 \times 10^{3}$	$520 \times 10^{3}$			
Cohesion, $C_0$ (MPa)	20	20	20	300			
Internal angle of friction, $\phi$ (°)	30	30	30	2			
Elastic shear modulus, G (Pa)	$4.4 \times 10^{10}$	$4.4 \times 10^{10}$	$6.7 \times 10^{10}$	$7.0 \times 10^{10}$			

rheologies are dry granite and felsic granulite, respectively [Ranalli, 1995].









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# Supplementary material

# 1 Density of the Arabian mantle lithosphere and asthenosphere

Bulk rock chemical analyzes of xenoliths from the mantle lithosphere below the Arabian shield have been gathered from literature [Kuo and Essene, 1986; Nasir and Safarjalani, 2000; Stein et al., 1993; Ghent et al., 1980; Nasir, 1992; Nasir and Al-Fuqha, 1988]. These xenoliths are found in neogene alkali basalts of Yemen, Saudi Arabia, Jordan, Israel and Syria [McGuire, 1988]. The dataset contains analyzes of lherzolites (N=16), harzburgites (N=7), clinopyroxenites (N=13) and websterites (N=5). The analyzes of the xenoliths as well as those of the fertile Hawaiian pyrolite and of four standard compositions of mantle lithosphere below tectons (lithospheres younger than 1 Ga) are presented in table 1. The densities corresponding to these chemical analyzes have been computed with the free energy minimization program PERPLEX07 [Connolly, 2009]. We used a thermodynamic database that is dedicated to mantle rocks thermodynamics [Xu et al., 2008]. The activity model set takes into account solid solution in plagioclase, spinel, garnet, olivine, wadsleyite, clinopyroxene, orthopyroxene and high-pressure clinopyroxene (C2/C phase) in the six oxides Na<sub>2</sub>O-CaO-FeO-MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> system (NCF-MAS). Even if the method is fairly similar, our results are not directly comparable to those of Griffin et al. [2009] who have not considered Na<sub>2</sub>O in their computations. This assumption leads to higher density contrasts at depths below 400 km than with the six oxides system [Nakagawa et al., 2010]. Chemical analyzes are generally given in the Na<sub>2</sub>O-K<sub>2</sub>O-CaO-FeO-MnO-MgO-Fe<sub>2</sub>O<sub>3</sub>-Al<sub>2</sub>O<sub>3</sub>-Cr<sub>2</sub>O<sub>3</sub>-TiO<sub>2</sub>-SiO<sub>2</sub> eleven oxides system. Reduction from this whole system to the six oxides system has been carried out with the following assumptions. (1) The K molar amount, which is always minor, has been added to the Na molar amount. (2) The Mn molar amount, which is always minor, has been added to the Fe molar amount. (3) All iron is considered as  $Fe^{2+}$ . (4) Cr and Ti are assumed to be only present as oxides (chromite  $Cr_2O_3$  and rutile  $TiO_2$ ) and are therefore not considered. Amongst all tested assumptions, this set allows us avoiding large and probably unrealistic amount of excess  $Al_2O_3$  and  $SiO_2$  in the clinopyroxenites and the websterites. The density with depth for all the considered chemical compositions has been computed along the initial thermal profile of the models. These curves are presented on figure 1. Following Nakagawa et al. [2010], only buoyancy curves rather than absolute densities are presented in the text. We indeed consider that computing density differences reduces the uncertainties unavoidably arising from the chemical analyzes, the thermodynamic database and the activity models.

Table 1: Chemical compositions of the mantle lithosphere xenoliths from the Arabian shield, tectons mantle lithosphere [*Griffin et al.*, 2009] and asthenosphere (fertile Hawaiian pyrolite) [*Wallace and Green*, 1991]. The compositions are given in the reduced NCFMAS system, which has been considered for density computation with the free energy minimization program PERPLEX07 [*Connolly*, 2009]. Also given is the Mg# = 100 Mg/(Mg+Fe) relative to the compositions.References: 1: *Ghent et al.* [1980], 2: *Griffin et al.* [2009], 3: *Kuo and Essene* [1986], 4: *Nasir* [1992], 5: *Nasir and Al-Fuqha* [1988], 6: *Nasir and Safarjalani* [2000], 7: *Stein et al.* [1993], 8: *Wallace and Green* [1991].

Lherzo	lites							
Sample	M-1	K-3	T-5	A-4	R-9	JAR1	JAR3	JAR9
$\mathrm{SiO}_2$	45.92	46.06	47.16	47.02	44.66	43.44	43.72	43.03
$Al_2O_3$	1.26	1.82	3.51	1.51	1.82	1.95	2.01	1.91
FeO	8.62	8.44	8.58	7.35	9.75	7.88	8.13	8.62
MgO	41.53	40.96	36.77	41.79	40.77	44.85	44.22	44.54
CaO	2.32	2.53	2.95	1.96	2.12	1.83	1.81	1.84
$Na_2O$	0.35	0.20	1.02	0.37	0.87	0.05	0.11	0.06
Mg#	0.90	0.90	0.88	0.91	0.88	0.91	0.91	0.90
Ref.	4	4	4	4	4	5	5	5
Sample	JAR15	JAR20	T6	B4	I1	TA-806	TA-842	MHZ-230
$\mathrm{SiO}_2$	43.62	43.39	44.70	44.69	43.01	45.16	44.69	46.81
$Al_2O_3$	1.72	1.78	3.42	3.30	3.18	2.11	2.60	3.23
FeO	7.92	8.44	8.84	6.89	8.82	8.08	8.01	8.97
MgO	44.83	44.38	40.32	42.23	41.55	42.35	41.99	37.32
CaO	1.86	1.92	2.36	2.57	3.06	2.01	2.40	3.44
$Na_2O$	0.06	0.09	0.35	0.32	0.38	0.29	0.31	0.23
Mg#	0.91	0.90	0.89	0.92	0.89	0.90	0.90	0.88
Ref.	5	5	6	6	6	7	7	7
Harzbu	rgites							
Sample	JT26A	$\rm JT26B$	HAK-1	T3	B2	I3	H2	
$\mathrm{SiO}_2$	44.70	45.05	44.25	42.85	44.32	43.54	42.24	
$Al_2O_3$	0.80	2.01	1.51	2.81	2.65	2.10	1.05	
FeO	8.19	8.19	7.98	9.02	7.90	9.57	9.15	
MgO	45.41	43.54	45.16	43.59	43.57	42.92	46.33	
CaO	0.80	1.01	0.92	1.46	1.40	1.70	1.17	
$Na_2O$	0.09	0.20	0.18	0.27	0.16	0.17	0.06	
Mg#	0.91	0.90	0.91	0.90	0.91	0.89	0.90	
Ref.	1	1	3	6	6	6	6	
Pyroxe	nites							
Sample	M-5	T-8	A-7	R-2	A-3	R-5	T1	T5
$\mathrm{SiO}_2$	54.51	53.87	53.63	52.18	51.93	51.00	44.71	46.04
$Al_2O_3$	8.51	9.60	6.87	7.86	11.44	10.38	12.67	11.40
FeO	5.45	4.31	5.81	7.26	5.92	8.27	10.72	10.46
MgO	15.86	16.98	24.34	24.53	13.77	15.78	16.13	17.38
CaO	14.69	14.76	8.48	7.35	15.79	13.34	14.42	13.37
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$Na_2O$	0.98	0.48	0.87	0.82	1.15	1.23	1.36	1.34
Mg#	0.84	0.88	0.88	0.86	0.81	0.77	0.73	0.75
Ref.	4	4	4	4	4	4	6	6
Sample	B7	Т8	I14	BS-701	KM-1402			
$SiO_2$	48.45	47.63	48.80	48.84	44.07			
$Al_2O_3$	8.69	9.64	10.53	12.01	12.65			
FeO	8.34	7.68	9.15	4.95	11.60			
MgO	17.65	18.86	19.57	17.02	14.55			
CaO	15.24	15.30	11.16	16.07	15.92			
$Na_2O$	1.62	0.89	0.78	1.10	1.21			
Mg#	0.79	0.81	0.79	0.86	0.69			
Ref.	6	6	6	7	7			
Webste	rites							
Sample	58376A	58376B	58377	T15	M11			
$SiO_2$	51.44	51.98	52.06	46.59	48.92			
$Al_2O_3$	7.39	6.29	5.48	11.03	9.52			
FeO	5.90	8.94	8.61	9.68	7.53			
MgO	17.72	19.49	18.37	23.87	23.89			
CaO	16.61	12.28	14.72	8.10	8.35			
$Na_2O$	0.93	1.02	0.77	0.73	1.78			
Mg#	0.84	0.80	0.79	0.81	0.85			
Ref.	1	1	1	6	6			
Asthen	osphere							
Sample	Fert-Pyr							
$\rm SiO_2$	45.80							
$Al_2O_3$	3.59							
FeO	8.74							
MgO	38.07							
CaO	3.13							
$Na_2O$	0.67							
Mg#	0.89							
Ref.	8							
Tecton	mantle lit	thospher	e					
Sample	Tc-1	Tc-2	Tc-3	Tc-4				
$SiO_2$	44.83	45.34	44.80	44.59				
$Al_2O_3$	3.53	3.93	2.62	2.33				
FeO	8.19	8.23	8.41	8.66				
MgO	40.09	38.99	41.47	41.95				
CaO	3.12	3.22	2.52	2.23				
$Na_2O$	0.24	0.28	0.18	0.24				
_ Mg#	0.90	0.89	0.90	0.90				
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Figure 1: Density evolution with depth for mantle lithosphere xenoliths from the Arabian shield, tectons mantle lithosphere and asthenosphere. The computation of the density has been carried out with the free energy minimization program PERPLEX07 [Connolly, 2009].

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