



# Plume-induced continental rifting and break-up in ultra-slow extension context: Insights from 3D numerical modeling

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## ABSTRACT

Breaking the lithosphere in extension without exceeding the driving far-field forces available on Earth is a tough quantitative modeling problem. One can tear it apart by propagation of an existing oceanic basin or weakness zone or one must assist rifting with magmatic processes, which drop the effective stress and weakens locally the lithosphere. While previous 3D models have demonstrated that non-cylindrical plumes produce almost cylindrical rift structures in a lithosphere under slight far-field loading, our contribution goes one step further by producing models of complete continental break-up. We investigate in details how the rheological stratification of the continental lithosphere interacting with active mantle plume influences the geometry and dynamics of rifting to continental break-up in 3D. We find that, irrespective of the rheological stratification, a plume-induced rifting process always occurs in two stages: an early crustal rifting stage and a late lithospheric necking (breakup) stage. In case of a rheologically decoupled lithosphere, initial brittle deformation is concentrated in the upper crust and strongly localized due to compensating ductile flow of lower-crustal material (core complex extension mode). On the contrary, rheological coupling between upper crust and lithospheric mantle results in highly distributed brittle deformation in the crust above mantle plume head (wide rift mode). Both core complex-like and wide rifting are followed by an abrupt transition to narrow rift stage when a localized ascent of mantle plume material focuses high strain along faults zones breaking through the entire lithosphere. The Main Ethiopian Rift, the Basin and Range province, and the East Shetland Basin may be natural examples of regions that have passed through these two stages of extension. Across-strike and along-strike asymmetry of break-up patterns arising spontaneously within initially symmetrical and laterally homogenous environment seems to be an intrinsic characteristic of plume-induced rifting.

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## 1. Introduction

Continental rifting is a fundamental geodynamic process that is described as stretching and thinning of the whole lithosphere, ultimately leading to a rupture (break-up) of the continent and the birth of new ocean basins (e.g., Buck, 1991, 2007). Depending upon the relative role of upwelling mantle flow and far-field forces in rift dynamics, rifting is conventionally considered to be either “active” (Sengör and Burke, 1978) or “passive” (McKenzie, 1978).

Despite significant progress in understanding of the rifting evolution from numerical (e.g., Burov and Poliakov, 2001; Huisman and Beaumont, 2003, 2008, 2011; Van Wijk and Blackman, 2005;

Watremez et al., 2013; Brune et al., 2014; Brune, 2016; Brune et al., 2017; Le Pourhiet et al., 2017) and analogue modeling (e.g., Benes and Davy, 1996; Brun, 1999, 2002; Chemenda et al., 2002; Sokoutis et al., 2007; Bonini et al., 2007; Corti, 2008, 2009; Corti et al., 2011, 2013), many unanswered questions remain. Some unresolved problems are the rapid strain transfer from rift-bounding faults to narrow magmatic segments (Ebinger and Casey, 2001; Corti, 2008), the enigmatic coexistence of magmatic and amagmatic overlapping rift branches (Ebinger et al., 2000; Nyblade and Brazier, 2002; Ring, 2014), the cause of significant along-axis rift width variations (Zeyen et al., 1997; Corti et al., 2013), or the origin of along-strike tectonic/magmatic segmentation of volcanic margins (Geoffroy, 2001, 2005). Most of these questions are directly or indirectly related to asymmetrical 3D features of the structural patterns and, as a result, require 3D models.

Recent technical and conceptual advances in numerical modeling have enlarged the possibility to investigate geodynamic processes

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with the help of 3D numerical experiments (e.g., Gerya and Yuen, 2007; Popov and Sobolev, 2008; Braun et al., 2008; Gerya, 2010; Thieulot, 2011; May et al., 2014, 2015). The results of 3D ultra-high resolution models that couple mantle/lithosphere system in a dynamically and rheologically consistent framework show that an upwelling mantle plume combined with far-field extension leads to the localization of continental extension along well-defined rift basins, thus reconciling the passive (external tectonic stresses) and active (plume-activated) rift concepts (Burov and Gerya, 2014; Koptev et al., 2015).

We follow-up on the study of Burov and Gerya (2014) with a series of numerical experiments of ascending mantle plume interacting with a rheologically realistic continental lithosphere subjected to ultra-slow far-field extension. Indeed, a number of geodetic studies now indicate that continental rifting initiates at extension rates of only a few millimeters per year (e.g., Calais et al., 1998; Saria et al., 2014; Birhanu et al., 2016). Our study focuses on the variations of plume-induced rifting style – mode of strain localization, deformation zone width, spatial location of break-up and spreading axes – as a function of variations in thermo-rheological properties of the continental lithosphere. We investigate a broad spectrum of thermo-rheological lithospheric profiles ranging from complete mechanical coupling to complete decoupling by varying either the lithology of the lower crust (wet quartzite and plagioclase flow laws, respectively) or the temperature at the Moho for a wet quartzite flow law. We also test the influence of varying the far-field extension rate and the mantle plume density.

In our previous works (Koptev et al., 2015, 2016), we have shown that the variations in the plume size and/or temperature have negligible effect on the principal output features of the models. Thus, here we intentionally focus on the analysis of thermo-rheological properties of the lithosphere and far-field tectonic forces.

We do not address the role of structural inheritance, which has been previously investigated by 2D simulations of continental rifting and breakup processes (e.g., Le Pourhiet et al., 2004; Huet et al., 2011; Chenin and Beaumont, 2013; Manatschal et al., 2015). Yet, recent 3D modeling studies of continental extension have demonstrated that complexities and multiphase tectonic evolution often attributed to inherited heterogeneities can naturally emerge from spontaneous 3D interactions (Brune et al., 2012; Le Pourhiet et al., 2012; Brune and Autin, 2013; Brune, 2014, 2016; Gerya, 2013; Liao and Gerya, 2014, 2015; Beniest et al., 2017; Le Pourhiet et al., 2017). Therefore, we believe that before studying the role of pre-existing heterogeneities in 3D, it is important to evaluate the complexity that can emerge spontaneously from the intrinsic dynamics of plume-induced continental rifting and breakup under slow far-field extension.

## 2. Numerical model description

### 2.1. 3D model design

#### 2.1.1. Spatial dimensions and resolution

We produced the simulations presented in this contribution using viscous-plastic I3ELVIS code (Gerya and Yuen, 2007; Gerya, 2010). This code is based on conservative finite differences and a marker-in-cell technique (see Supplementary methods for more details).

The regular rectangular model box used in this study consists of  $297 \times 297 \times 133$  nodes (corresponding to dimensions of  $1500 \times 1500 \times 635$  km) and offers spatial resolution of ca.  $5 \times 5 \times 5$  km per grid cell. These simulations require large computational efforts. At each timestep, the code solves a system of algebraic equations that comprises thirty million degree of freedom and advects a hundred million randomly distributed Lagrangian markers. The total model time of experiments varies between 40 and 100 Myr, representing 50,000 timesteps on average. Simulations were run on a SGI shared (NUMA) fat-node cluster with 2.8 GHz Intel Xeon CPU cores. Each simulation represents 5 years of CPU time in total.

#### 2.1.2. Internal model structure

We impose a 30 km-thick “sticky air” layer at the uppermost part of the model domain to approximate the upper surface of the crust as a free surface. The viscosity of the “sticky air” is  $10^{18}$  Pa s and its density is  $1 \text{ kg/m}^3$ , according to optimal parameters established by Cramer et al. (2012). We use a vertically stratified continental lithosphere with an upper crust, lower crust, and lithospheric mantle lying on the upper mantle.

The total crustal thickness is 36 km and comprises two layers of 18 km each. The ductile rheology of the upper crust is of minor importance since the corresponding depth interval of 0–18 km is mainly dominated by rock type independent brittle failure (Byerlee, 1978; Brace and Kohlstedt, 1980; Kohlstedt et al., 1995; Burov, 2011). Therefore, we employ a felsic upper crust described by wet quartzite rheology (Supplementary Table 1) in all experiments.

In contrast, the effective rheology of the lower crust is strongly controlled by temperature and lithology. Moreover, the rheological behavior of the lower crust, which controls the mechanical coupling within lithospheric layers and the integrated strength of the lithosphere (Burov and Diament, 1995), is known to influence the mode of continental rifting (e.g. Buck, 1991; Burov and Poliakov, 2001, 2003; Buck, 2007; Gueydan et al., 2008), even in 3D (Le Pourhiet et al., 2017). In order to explore different levels of (de)coupling, we varied the temperature at the Moho for a wet quartzite lithology (Table 1). However, because this lithology does not permit to capture cases with strong coupling, we also performed a strongly coupled experiment (model 2, Table 1) using a strong mafic lower crust described by plagioclase rheology (Supplementary Table 1).

The ductile rheology of the mantle lithosphere (dry olivine flow law, Supplementary Table 1) is controlled by dry olivine dislocation and Peierls creeping flow, while the hot asthenospheric mantle (dry olivine flow law as well) deforms predominantly by diffusion creep (Evans and Goetze, 1979; Caristan, 1982; Karato and Wu, 1993; Durham et al., 2009). The initial depth of the lithosphere–asthenosphere boundary is 150 km.

We initiate a mantle plume by seeding a temperature anomaly at the base of the model box (635 km). Following Burov and Gerya (2014) and Koptev et al. (2015, 2016), this initial anomaly is modeled as a hemisphere with a radius of 200 km. In order to explore the sensitivity of rifting style to variations in the buoyancy of the mantle plume, we decrease the mantle plume density calculated from standard thermodynamic petrological models (Perple\_X; Connolly, 2005) by  $30 \text{ kg/m}^3$  in models 13–14 (Table 1). We assume that the mantle plume contains a small amount of water so that its deformation follows a wet olivine flow law (see Supplementary Table 1).

As mentioned above, we intentionally avoid any pre-imposed structuration within crust and lithospheric mantle. We address the readers Koptev et al. (2015, 2016) for the detailed analysis of the role of the inherited rheological discontinuities (cratonic blocks, suture zones).

#### 2.1.3. Initial temperature distribution and thermal boundary conditions

The initial geotherm is piece-wise linear, with  $0 \text{ }^\circ\text{C}$  at the surface ( $\leq 30$  km, the air),  $500\text{--}800 \text{ }^\circ\text{C}$  at the Moho (see Table 1),  $1300 \text{ }^\circ\text{C}$  at the lithospheric base and  $1630 \text{ }^\circ\text{C}$  at the bottom of the model domain (635 km depth). The resulting adiabatic thermal gradient in the mantle is  $0.5\text{--}0.7 \text{ }^\circ\text{C/km}$ . The mantle plume has an initial temperature of  $2000 \text{ }^\circ\text{C}$ , which corresponds to a 300 to  $370 \text{ }^\circ\text{C}$  temperature excess between the plume and the surrounding mantle. This relatively large temperature excess is consistent with seismic data showing the 20–40 km depression of the 410 km discontinuity beneath the eastern branch of the East African rift system in Kenya and Tanzania (Huerta et al., 2009) where low-velocity anomalies indicate the upper-mantle thermal structure related to the African superplume (Nyblade, 2011).

We use a constant temperature condition at the upper surface ( $0 \text{ }^\circ\text{C}$ ) and at the bottom ( $1630 \text{ }^\circ\text{C}$ ) of the model, and a thermal insulating boundary condition (i.e. zero conductive heat flux) to all vertical boundaries.

**Table 1**  
Controlling parameters of the numerical experiments.

Experiment title	Controlling parameters				
	Lithosphere properties		Temperature at the Moho (°C)	Horizontal extension velocities (mm/yr)	Plume density
	Type (rheology) of lower crust (Supplementary Table 1)	WetQz – wet quartzite flow law; An <sub>75</sub> – plagioclase flow law			
1. R1	1 (WetQz)	700	3	Perple_X	
2. R1.Lc = An <sub>75</sub>	2 (An <sub>75</sub> )	700	3	Perple_X	
3. R1.T <sub>Mh</sub> = 500 °C	1 (WetQz)	500	3	Perple_X	
4. R1.T <sub>Mh</sub> = 600 °C	1 (WetQz)	600	3	Perple_X	
5. R1.T <sub>Mh</sub> = 800 °C	1 (WetQz)	800	3	Perple_X	
6. R1.V <sub>ext</sub> = 1.5	1 (WetQz)	700	1.5	Perple_X	
7. R1.V <sub>ext</sub> = 2	1 (WetQz)	700	2	Perple_X	
8. R1.V <sub>ext</sub> = 6	1 (WetQz)	700	6	Perple_X	
9. R1.T <sub>Mh</sub> = 600 °C + V <sub>ext</sub> = 1.5	1 (WetQz)	600	1.5	Perple_X	
10. R1.T <sub>Mh</sub> = 800 °C + V <sub>ext</sub> = 1.5	1 (WetQz)	800	1.5	Perple_X	
11. R1.T <sub>Mh</sub> = 600 °C + V <sub>ext</sub> = 6	1 (WetQz)	600	6	Perple_X	
12. R1.T <sub>Mh</sub> = 800 °C + V <sub>ext</sub> = 6	1 (WetQz)	800	6	Perple_X	
13. R1.LightPlume	1 (WetQz)	700	3	Perple_X-30 kg/m <sup>3</sup>	
14. R1.LightPlume + V <sub>ext</sub> = 6	1 (WetQz)	700	6	Perple_X-30 kg/m <sup>3</sup>	

### 2.1.4. Velocity boundary conditions

We simulate weak tectonic forcing by applying a constant, time-independent, extension rate along the entire length of the right (“eastern”) and left (“western”) sides of the model domain. We use a half rate of extension of 3 mm/yr for the reference experiment, a typical value for pre-break-up continental rifts affected by mantle upwellings, such as the Rio Grande and East African rifts (Buck, 2007; Stamps et al., 2008; Saria et al., 2014). We test the influence of this parameter by varying it within a range of 1.5 to 6 mm/yr (see Table 1). We chose the upper limit in order to ensure that the resulting horizontal forces along the borders of the models are of the same order of magnitude as “ridge push” ( $2-3 \times 10^{12}$  N per unit length) and “slab pull” ( $4-6 \times 10^{12}$  N per unit length) forces (e.g. Parsons and Richter, 1980; Schellart, 2004). This ensures that we do not violate the far-field tectonic forces available within typical continental lithosphere in plate interiors (e.g., Forsyth and Uyeda, 1975; Buck, 2006, 2007). We use free slip boundary conditions at the “northern” and “southern” sides of the models, which we do not subject to extension. Compensating vertical velocities along the upper and lower model boundaries are calculated on the basis of mass conservation within the model domain.

### 2.2. Modeling procedure: Experiments and key variable parameters

We tested 14 different experimental settings (Table 1) by varying four controlling parameters that characterize the properties of the lithosphere (lower-crustal rheology and Moho temperature), boundary conditions (extension half rate), and mantle plume density.

In the reference experiment (model 1), an initial temperature of 700 °C at the Moho results in rheological decoupling between the upper crust and the lithospheric mantle. In contrast, model 2 shows strong rheological coupling due to the stronger (mafic) rheology imposed for the lower crust. Moho temperature from 500 °C to 800 °C (models 3–5) permits to investigate intermediate coupling cases. Models 6–8 illustrate the impact of the velocity boundary conditions: slower (1.5 and 2 mm/yr; models 6 and 7, respectively) and faster (6 mm/yr; model 8) far-field extension with respect to assumed in the reference model 1 (3 mm/yr). Combined variations of Moho temperature and extensional velocity are represented by models 8–12. In model 13, Perple\_X-derived density of the mantle plume is reduced by 30 kg/m<sup>3</sup> to investigate the effect of the plume chemical buoyancy. Model 14 combines a lighter mantle plume and faster far-field extension.

Before proceeding to the description of the results, we should note here that resolution tests were performed in order to assess the robustness of our results. The results presented in Supplementary Fig. 1 show that a notable decrease in the resolution (up to 10 km per grid cell) leads to distributed small-offset faulting (Supplementary Fig. 1a–c). On the contrary, the middle-resolution experiments (7.5 km per grid cell) produce similar localization pattern as the one produced with a higher resolution used in our study (5 km per grid cell, Supplementary Fig. 1d–f). We conclude that the linear localized sub-parallel faults are robust model feature as they systematically appear independently of the size of the grid cell. The resolution of 5 km per cell adopted in the series of experiments presented is therefore optimal and further increasing the resolution would not change significantly the interpretation of the models.

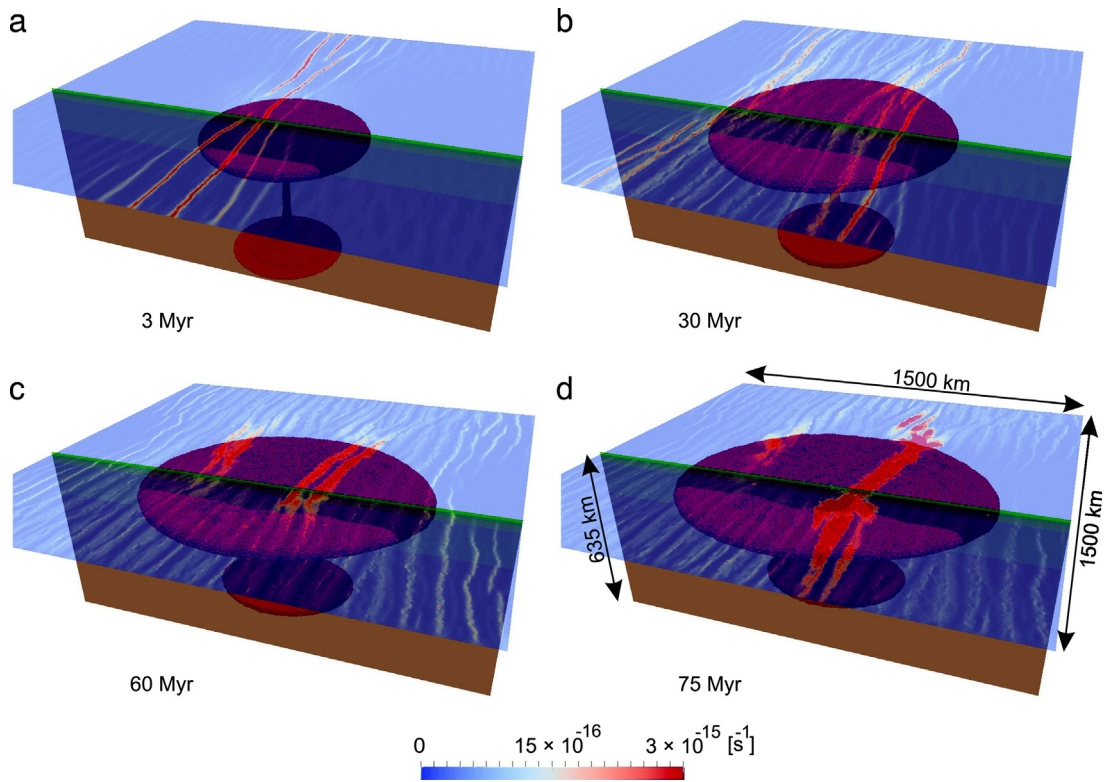
## 3. Experimental results

### 3.1. Reference model 1

For the chosen parameters, the reference model (Figs. 1–3) produces a rapid mantle ascent as the plume material reaches the base of the lithosphere in 0.5 Myr. Surface topography first reacts by domal uplift (Fig. 2a), soon after (<3 Myr) replaced by subsidence and coeval initiation of series of linear, localized, and sub-parallel normal faults associated with long and narrow rifted basins (Fig. 1a; Fig. 2b). These faults merge into a localized, linear rift, which crosses the entire model domain in the direction perpendicular to the far-field extension, with considerable narrowing in its central part (Fig. 1a). In cross-section, the model shows concentrated brittle strain in the upper crust underlain by a lower crust subjected to distributed ductile flow (Fig. 3a).

As shown by Burov and Gerya (2014), this model indicates that a hot mantle upwelling allows for this strain localization within a laterally homogenous lithosphere subjected to weak far-field forcing. However, this does not lead to immediate lithospheric scale break-up: localized deformation is restricted to the brittle upper crust whereas the high-strength lithospheric mantle remains almost undeformed slowly flowing under applied external extension (Fig. 3a).

Ductile flow in the lower crust leads to crustal boudinage within the central part of the model domain (see zoom inset of Fig. 3d). In more details, as the lower crust rises, it appears to be harder to localize brittle deformation, which causes lateral migration of brittle upper-crustal strain towards adjacent undeformed areas (Fig. 3b–d). This results in gradual in-sequence opening of the rift basin (Fig. 1b; Fig. 2c–f),

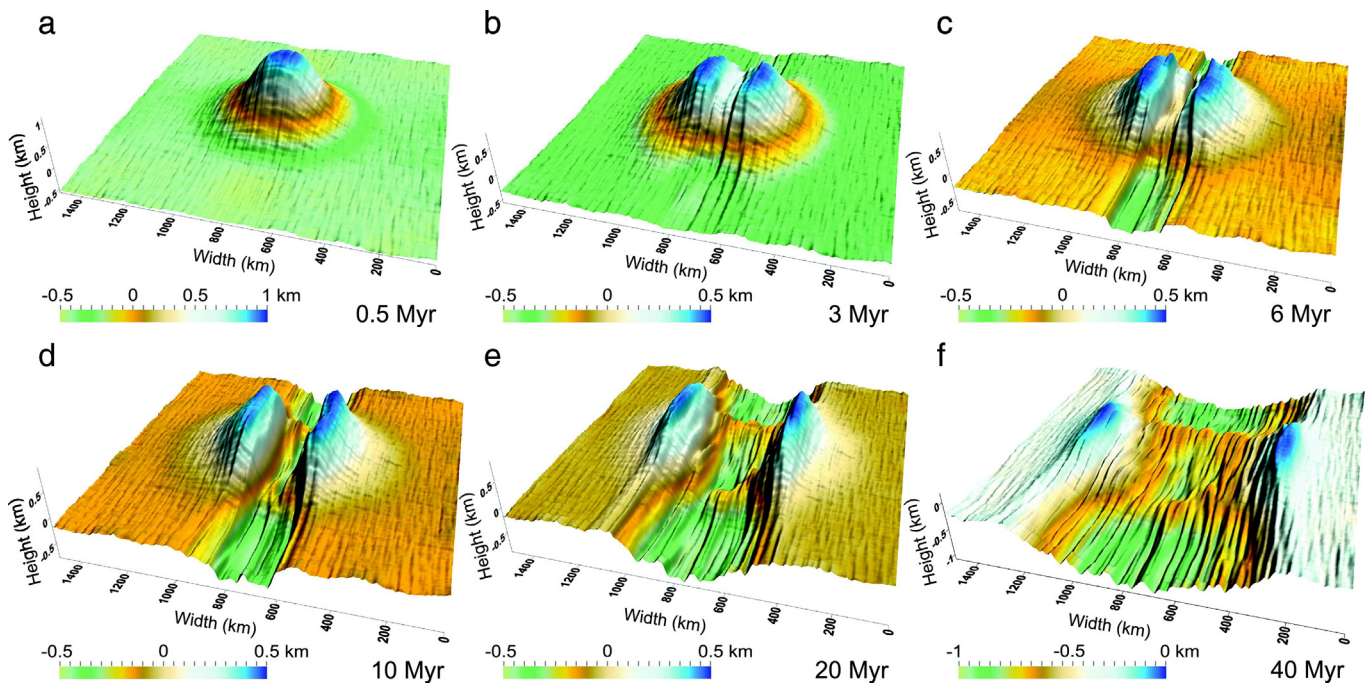


**Fig. 1.** Reference model 1. 3D view of experiment evolution: a) rapid plume uplift and formation of localized zone of brittle deformation in the upper crust; b) fast opening of 600-km wide rift basin due to combined effect of “active” and “passive” rifting; c) appearing of two narrow rifts associated with initiation of localized plume ascent; d) break-up of the continental lithosphere and development of single spreading zone considerably shifted from the mantle plume center. The plume material is shown in dark red. Blue to red colors indicate strain rate at the level of 10 km (i.e. in the upper crust).

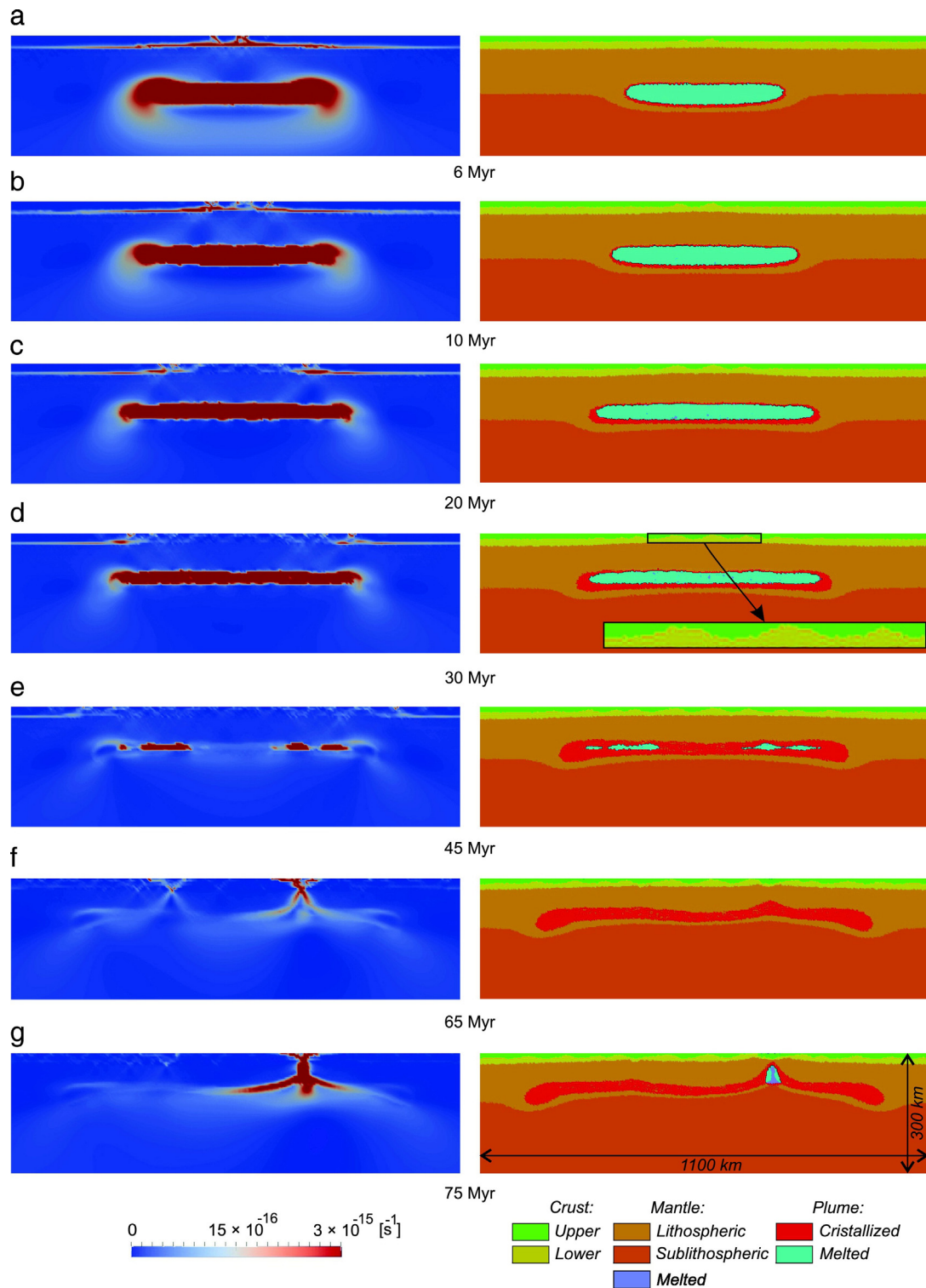
which is bounded by active normal faults that are always located at the edge of the thinned and previously deformed crust (Fig. 3b–d).

The generated rift basin opens ~3 times faster than the imposed far-field extension rate: a 600-km wide rift forms in <35 Myr while

the amount of passive extension applied at the boundaries is only 200 km (Fig. 1b, Fig. 3d). This confirms that, for plume-induced “active-passive” rifting, the active mantle upwelling is the main driver of rapid rifting.



**Fig. 2.** Reference model 1. Surface topography evolution: domal uplift (a) is quickly replaced by subsidence and formation of rifted basin (b) that opens (c–e) significantly faster compared to the imposed external extension (half rate is 3 mm/yr): at 40 Myr the amount of passive extension applied at the boundaries is 240 km while the width of formed depression is >700 km (f).



**Fig. 3.** Reference model 1. Strain rate (left column) and compositional (right column) cross-sections: a) localized brittle deformation in the upper crust and distributed ductile strain into weak lower crust; b–d) crustal boudinage due to ductile flow in the lower crust; in-sequence migration of the brittle deformation leads to the opening of the rift basin; e) almost complete recrystallization of the plume material (note spatial position of the last melted segments); wide rift basin (~800 km) with less expressed fault boundaries; f) initiation of localized plume uplift and development of two faults zones cutting the entire lithosphere; g) continental break-up along right narrow rift (note that left rift is abandoned); decompressional melting of plume and lithospheric material due to its localized uplift underneath the spreading axis. Spatial position of vertical cross-sections is shown on Fig. 1. It is shifted to 250 km from the central part of the model domain and, consequently, does not pass through the plume tail. Given the quick lateral penetration of the plume material into the lower part of the lithospheric mantle, in the corresponding cross-sections the plume head is underlain by a thin layer of the remnant lithospheric mantle.

Partial melting of almost whole plume head material at the early stage of the model evolution (Fig. 3a–d) is consistent with geochemical data that suggest a deep mantle source for the pre-rift Oligocene flood

basalts from the NW-Ethiopian plateau (e.g. Pik et al., 1999, 2006; Nelson et al., 2008, 2012; Halldórsson et al., 2014) and for mafic to ultra-mafic dykes and sills South China and Gairdner Dyke Swarm in Australia

(~825 Ma) preceding Rodinia rifting and break-up (e.g. Li et al., 1999; Li et al., 2005; Wang et al., 2007).

The next stage of the system evolution starts at ~60–65 Myr with the development of lithospheric scale faults associated with incipient and localized ascent of plume material through the lithosphere (Fig. 3f). In contrast to the former stage, the entire lithosphere is now subjected to localized deformation, with necking and boudinage (Fig. 3f) at two narrow and parallel areas located inside the rift valley that formed at the previous stage (Fig. 1c). Their spatial position seems to be correlated with the location of the last melted portions of the plume head material (blue blobs on Fig. 3e) before complete solidification. As the model evolves, one of these narrow rifts further develops into a spreading center while the other decays (Fig. 1d; Fig. 3g).

Thus, in the context of a laterally homogenous lithosphere, slow external extension and active mantle plume are not able to localize brittle deformation in the mantle lithosphere instantly. The high-strength lithospheric mantle remains almost undeformed over the first 60 Myr while highly localized brittle deformation concentrates in the upper crust, mechanically decoupled from the mantle by the ductile lower crust. This localized brittle deformation in the upper crust underlain by lower crust subjected to ductile flow corresponds to core complex mode of extension (see Buck, 1991). Upper-crustal deformation is localized along normal faults bounding a broad rift system with wide-spread in-sequence boudinage of the upper crust. Each localization event corresponds to the formation of a single core-complex before deformation migrates outward. Further model evolution shows a quick transfer to narrow rift mode (Buck, 1991) when the localized ascent of mantle plume material triggers strain localization along normal faults cutting through the whole lithosphere (~60–65 Myr). Pre-break-up narrow rifting is followed by post-break-up spreading that initiates at ~75 Myr. In this model, the spreading axis is shifted laterally from the center of the plume head.

### 3.2. Model with plagioclase rheology of the lower crust (Model 2)

In contrast to the reference model 1, model 2 (Figs. 4, 5, and 6) is characterized by rheological coupling between the upper crust and

the mantle lithosphere as a result of the stronger plagioclase rheology used for the lower crust (Table 1). This precludes ductile flow of lower-crustal material and leads to diffuse brittle deformation in the crust and the lithospheric mantle overlaying the plume head (Fig. 6a). Further model evolution shows a gradual broadening of the area undergoing distributed deformation (Fig. 6b–e). As a result, until about 30–40 Myr a large number of small-offset, parallel, and linear faults cover a ~1000 km-wide dome of slightly elevated (~400 m) topography (Fig. 4a; Fig. 5a).

Burov and Gerya (2014) showed that ultra-slow tectonic stretching of the continental lithosphere in the absence of active mantle upwelling results in broadly distributed small-offset parallel faults, which are not localized within any particular zone (see also Supplementary Fig. 2), and that in the same far-field tectonic context, the simultaneous impingement of active mantle plume leads to progressive focusing and amplification of localized non-axisymmetric deformation and to the development of major large-offset highly localized faults. Our results show that this rapid localization of the deformation at the surface only occurs when the lower crust can flow (see Section 3.1). When the lower crust is coupled to the lithospheric mantle the initial extension phase is not as focused as in Burov and Gerya (2014). Instead a first phase of distributed normal faulting (Fig. 4a) reassembling the deformation of “no-plume” experiment by Burov and Gerya (2014) precedes the plume-induced break-up.

Similarly to reference model 1, the next stage of the system development (~50–55 Myr) shows localized deformation along two parallel zones (Fig. 4b–c) marked by narrow topographic depressions (Fig. 5b–c) underlain by localized plume ascent (Fig. 6f). Only one of these narrow rifts evolves into a spreading center (Fig. 4d; Fig. 5d–f; Fig. 6g) whereas the other aborts, despite an apparent higher level of initial maturity (Fig. 6f). It is noteworthy that the hot plume material does not produce uplift along the future break-up axis uniformly (see compositional distribution at 70 Myr on Fig. 7). A vertical cross-section along-strike the future line of break-up shows that the solidified mantle plume remains attached to the lithospheric base in its central part, whereas segments on either side show uplift and decompression

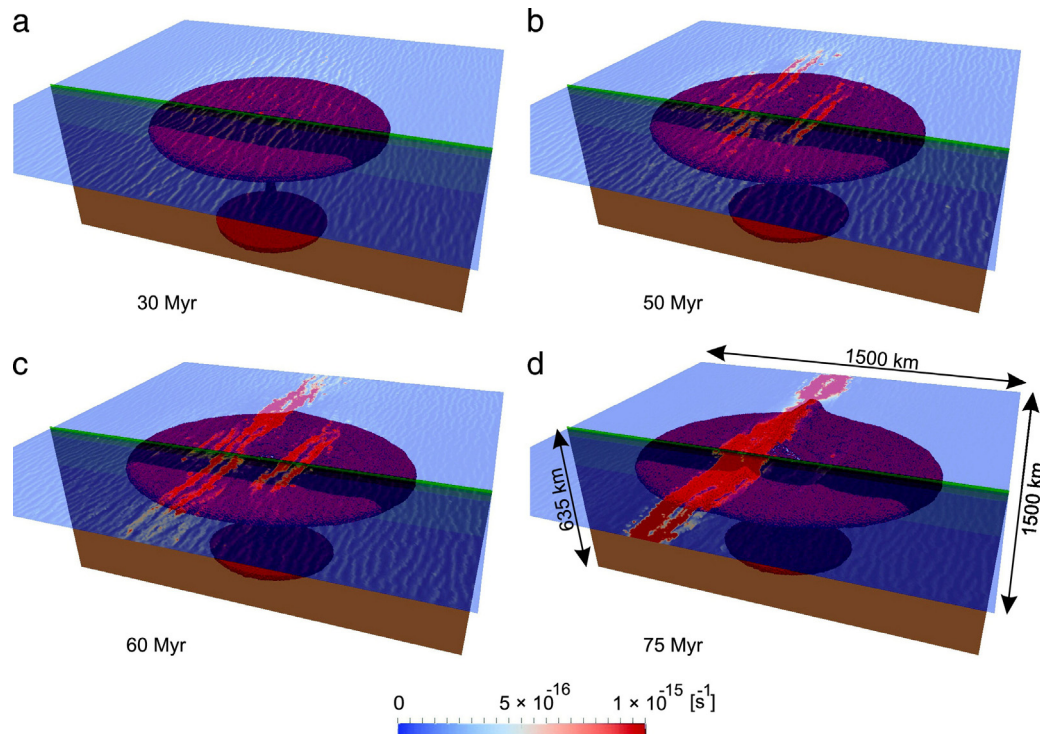
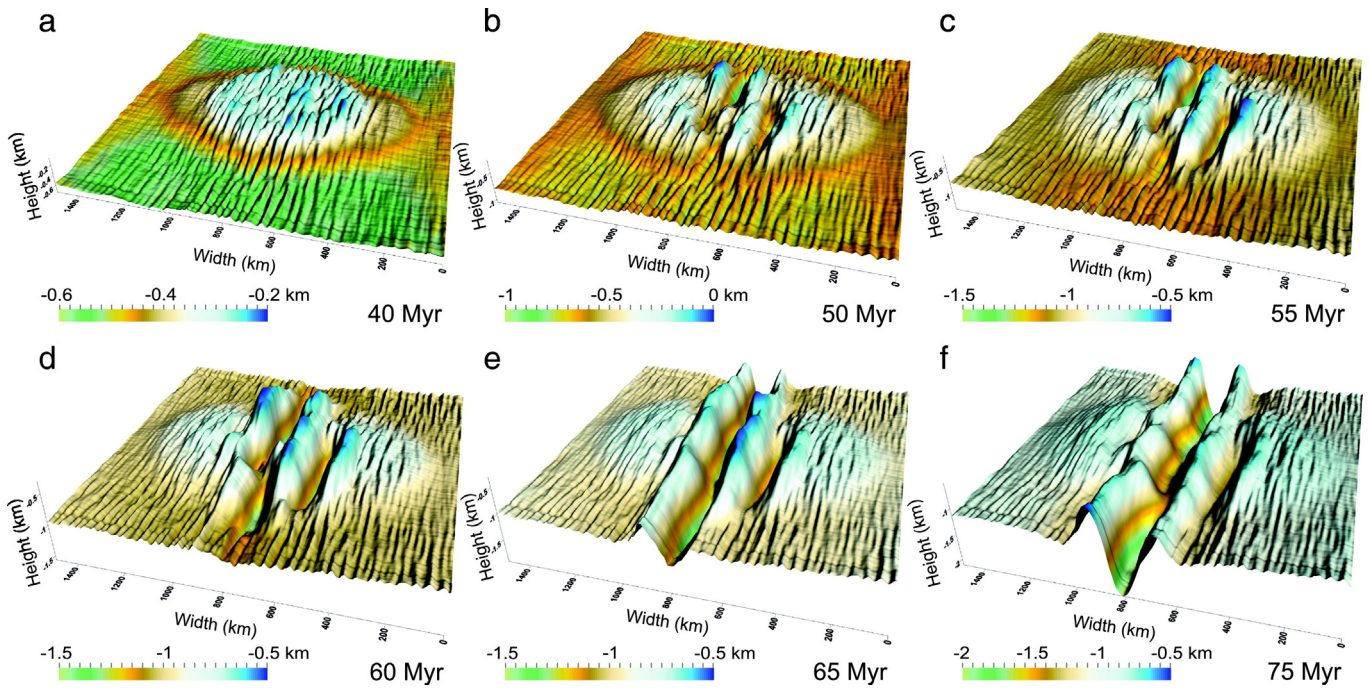


Fig. 4. Model with a mafic lower crust (Model 2). 3D view of experiment evolution: a) highly distributed linear faults above laterally spread plume head; b–c) narrow rifting along two stretched zones; d) break-up of the continental lithosphere along one spreading axis situated above mantle plume center.



**Fig. 5.** Model with a mafic lower crust (Model 2). Surface topography evolution: a) distributed small-offset faults over plume-induced central domal uplift; b–d) formation of two parallel depressions bounding by rift shoulders; e–f) one of these rift structures continues to develop into spreading axis while another one is decaying.

melting (Fig. 7d). These localized melting zones rise through the lithosphere up to Moho depth and lead to development of separated crustal magma reservoirs feeding volcanic centers along future “passive” margin.

The simultaneous distributed deformation in the crust and the mantle lithosphere during the initial stage of the model evolution contrasts greatly with the decoupled model. Brittle strain is spread over a broad region subject to the thermal influence of the mantle plume. In contrast to reference model 1 where the opening of the rift basin occurs by lateral migration of localized upper-crustal strain (Fig. 3a–e), here the broadening of the rift zone occurs (Fig. 6a–e) as a result of the lateral flow of the hot plume material within the lower part of the mantle lithosphere. As a result, strain accumulates predominately in the central part of the box, which favors a quasi-centered position of the narrow rift during subsequent model stages. This model therefore predicts that, in a rheologically coupled lithosphere, initial wide rifting is followed by narrow rifting localized at the location of maximum cumulative deformation, close to the plume source center. On the contrary, the in-sequence core complex extension mode in reference model 1 does not favor a particular location for narrow rift localization, which appears to be controlled by complex internal processes within the deep mantle plume. In this case, the final position of the narrow rift/spreading axis can be considerably shifted with respect to the area of initial plume impingement.

### 3.3. Dependence of initial rift width and final break-up position on Moho temperature

In order to study different degrees of the rheological (de)coupling of the lithosphere, we performed a series of models with different temperatures at Moho interface (models 1, 3–5).

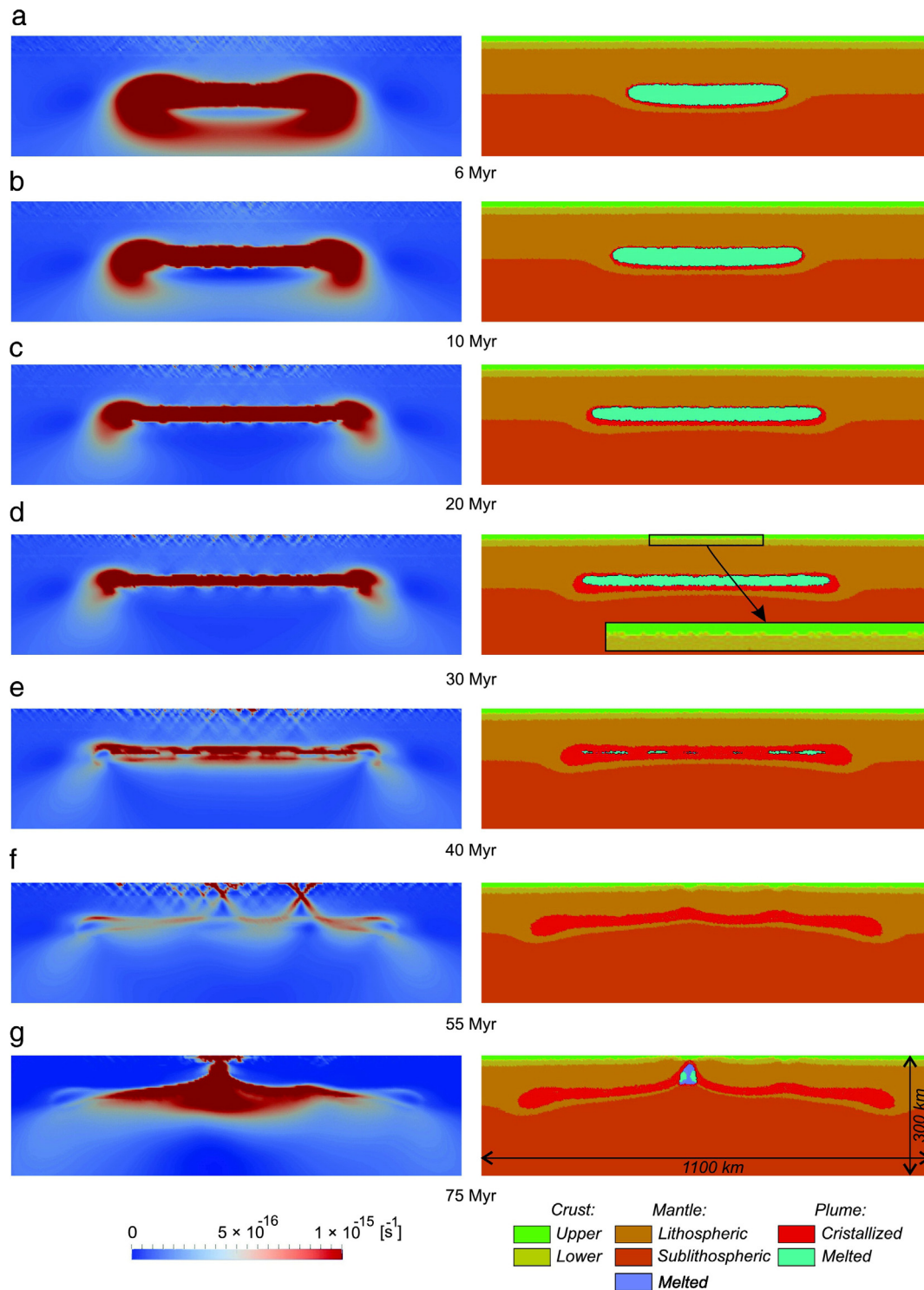
Similarly to reference model 1 (temperature at Moho depth of 700 °C), models 4 and 5 (Moho temperature of 600 °C and 800 °C, respectively) show localized brittle deformation in the upper crust soon after (~3 Myr) mantle plume impingement at the base of the lithosphere (Fig. 8b–d). On the contrary, a colder crustal geotherm (500 °C at Moho interface, model 3) results in a wide (~500 km) zone of highly distributed brittle strain (Fig. 8a). Decreasing Moho temperature leads

to increasing lower-crustal strength. As a result, the area of brittle upper-crustal deformation widens from 100 km to 200–300 km then 500 km with decreasing of Moho temperature from 800 °C to 500 °C (models 5, 1, 4 and 3, respectively, see Fig. 8).

A systematic analysis of the models with different combinations of initial crustal geotherm and applied far-field extension rate (models 1, 4–6; 8–12; Fig. 9; Supplementary Table 2) shows that this trend persists regardless of the far-field velocities applied: increasing the Moho temperature from 600 °C to 800 °C leads to decreasing the rift width from 425 km to 150 km for the fastest half-velocity (6 mm/yr) and from 175 km to 75 km for the slowest one (1.5 mm/yr).

Further evolution of the model with a relatively hot geotherm (800 °C at Moho, model 5, Fig. 10d) shows opening of the rift basin due to lateral migration of localized upper-crustal deformation similar to the core complex extension mode of reference model 1 (700 °C at Moho; Figs. 1–3; Fig. 10c). On the contrary, the model with the coldest Moho temperature (500 °C, model 3, Fig. 10a) shows a gradual broadening of the area experiencing distributed brittle strain, up to a width of ~750 km at 20 Myr. As in the model with a mafic lower crust (model 2, Fig. 4–6), this deformation mode can be classified as wide rifting. The lithosphere remains slightly decoupled even with the coldest geotherm, thus, restricting brittle deformation to the upper crust and keeping lithospheric mantle undeformed (Fig. 8a; Fig. 10a). Model 4 (600 °C at Moho depth) displays an intermediate extension mode (Fig. 10b): the highest brittle strain rate is concentrated along normal faults bounding wide rift valley but distributed internal deformation continues within the basin.

The rate of rift basin opening seems to be controlled by the lateral spreading of mantle plume material. In the case of the hottest geotherm (800 °C at Moho; model 5), the plume material penetrates profoundly into the lithospheric mantle, which limits its lateral flow. This is reflected in the relatively narrow (200 km-wide at 20 Myr) rift valley formed above (Fig. 10d). On the contrary, a colder lithospheric geotherm (600–700 °C at Moho; models 1, 4) prevents the vertical penetration of the plume head but favors its lateral spreading. This results in considerably broader (350–500 km-wide at 20 Myr) rift basins (Fig. 10b–c). Decreasing Moho temperature to 500 °C (model 3) leads to further increasing the rift basin width, up to ~750 km (Fig. 10a). This



**Fig. 6.** Model with a mafic lower crust (Model 2). Strain rate (left column) and compositional (right column) cross-sections: a–d) gradual broadening of diffuse brittle deformation in the crust and the lithospheric mantle over spreading plume head (note the lack of lower-crustal ductile flow and the absence of crustal boudinage); e) almost complete solidification of the plume material (note quasi-uniform lateral distribution of melted remains); f) localized deformation along two narrow rifts and associated with them localized ascent of the plume material; g) continental break-up along quasi-centered axis (note that another narrow rift is failed).

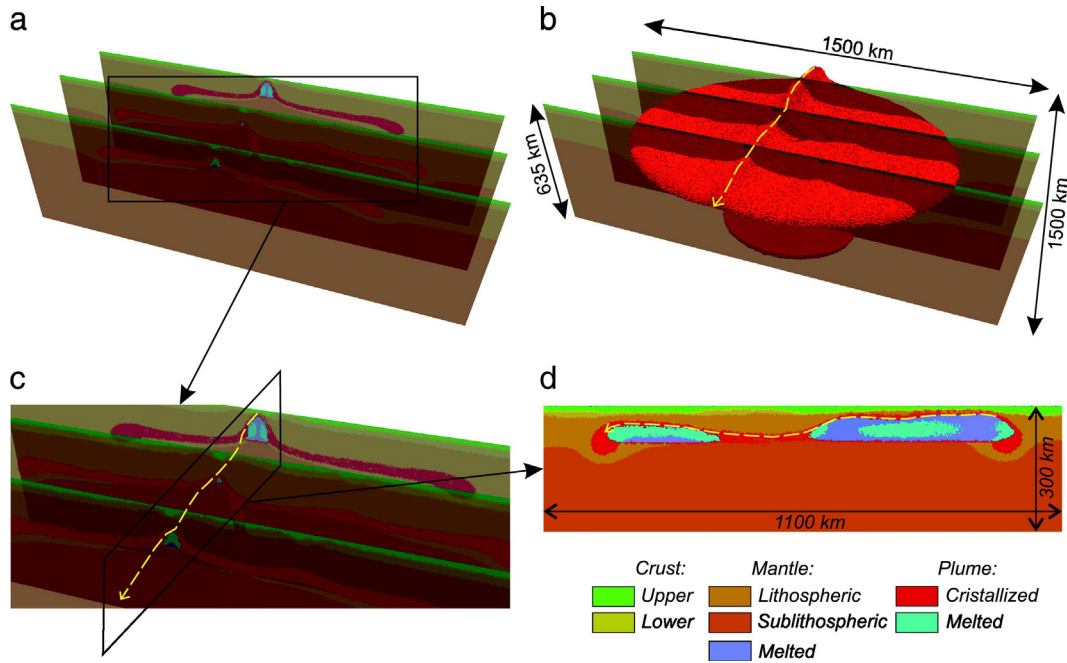
results from wider lateral spreading of the plume material, together with a change of extension mode to wide rifting.

Thus, the width of initially localized deformation appears to be governed by the degree of rheological (de)coupling between the brittle upper crust and the lithospheric mantle, whereas the subsequent opening of the rift basin is controlled by lateral flow of the mantle plume, which, in turn, is conditioned by the lithospheric geotherm (Supplementary Fig. 3).

As mentioned above (see Section 3.1–3.2), both the core complex and wide rift modes of extension are condemned to be succeeded by narrow rifting and spreading.

In case of reference model 1 (700 °C at Moho; Fig. 11c), the spatially decentered narrow rifts are linked to the last melted fragments remaining inside the mantle plume material before its complete solidification (see also Fig. 3e–f). Meanwhile, due to its hotter lithospheric geotherm (800 °C at Moho), model 5 shows a transition to narrow rift mode before complete





**Fig. 7.** Model with a mafic lower crust (Model 2). Compositional cross-sections (at 70 Myr) showing that localized plume ascent varies not only across-strike but also along-strike. Yellow line highlights the position of the plume apex; the arrow indicates the direction of its dipping. Bulk of mantle plume is shown in red. Note that the vertical variations in the position of the plume apex reach 70–80 km that in rough accordance with conceptual model of volcanic margin segmentation based on magnetic fabric studies (Callot et al., 2001, 2002; Geoffroy, 2005; Gac and Geoffroy, 2009). See Section 4.2 for more discussion.

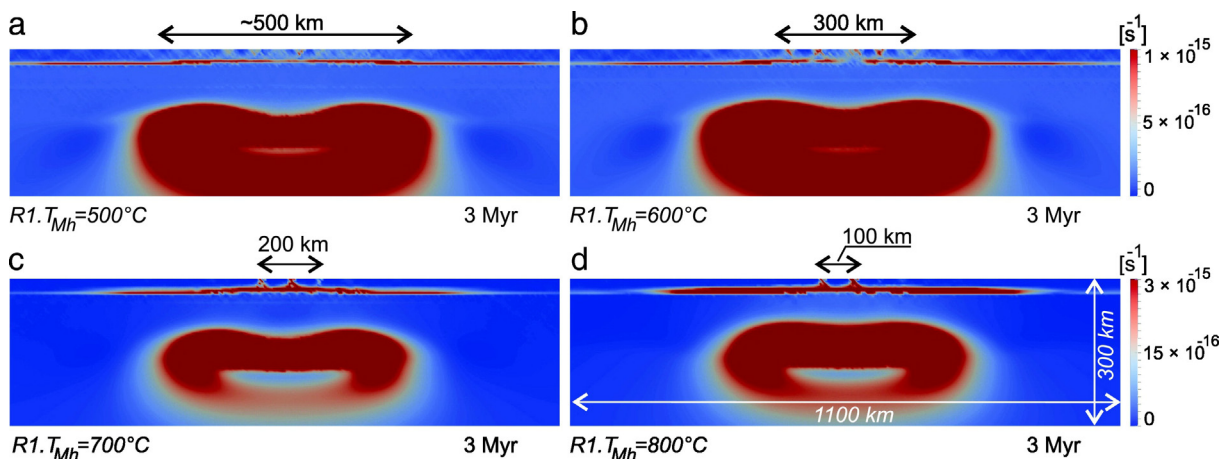
plume solidification: the remaining fragments of the melted plume material are involved in its localized ascent along two axes shifted symmetrically with respect to the center of the plume head (Fig. 11d). They both later evolve into spreading axes (Fig. 13c) in contrast to reference model 1 where only one of the narrow rifts survives (Fig. 1d; Fig. 3g; Fig. 13a).

Experiments with a colder Moho temperature (500–600 °C; models 3–4) result in a (quasi)central position of the localized narrow rifts (Fig. 11a–b). This appears to be typical for models characterized by a (quasi)“wide rift” mode of initial crustal deformation (see also model 2 with mafic lower crust; Fig. 6). The gradual broadening of diffuse brittle deformation leads to strain accumulation within the central part of the model domain resulting in a (quasi)centered position of the newly formed narrow rifts.

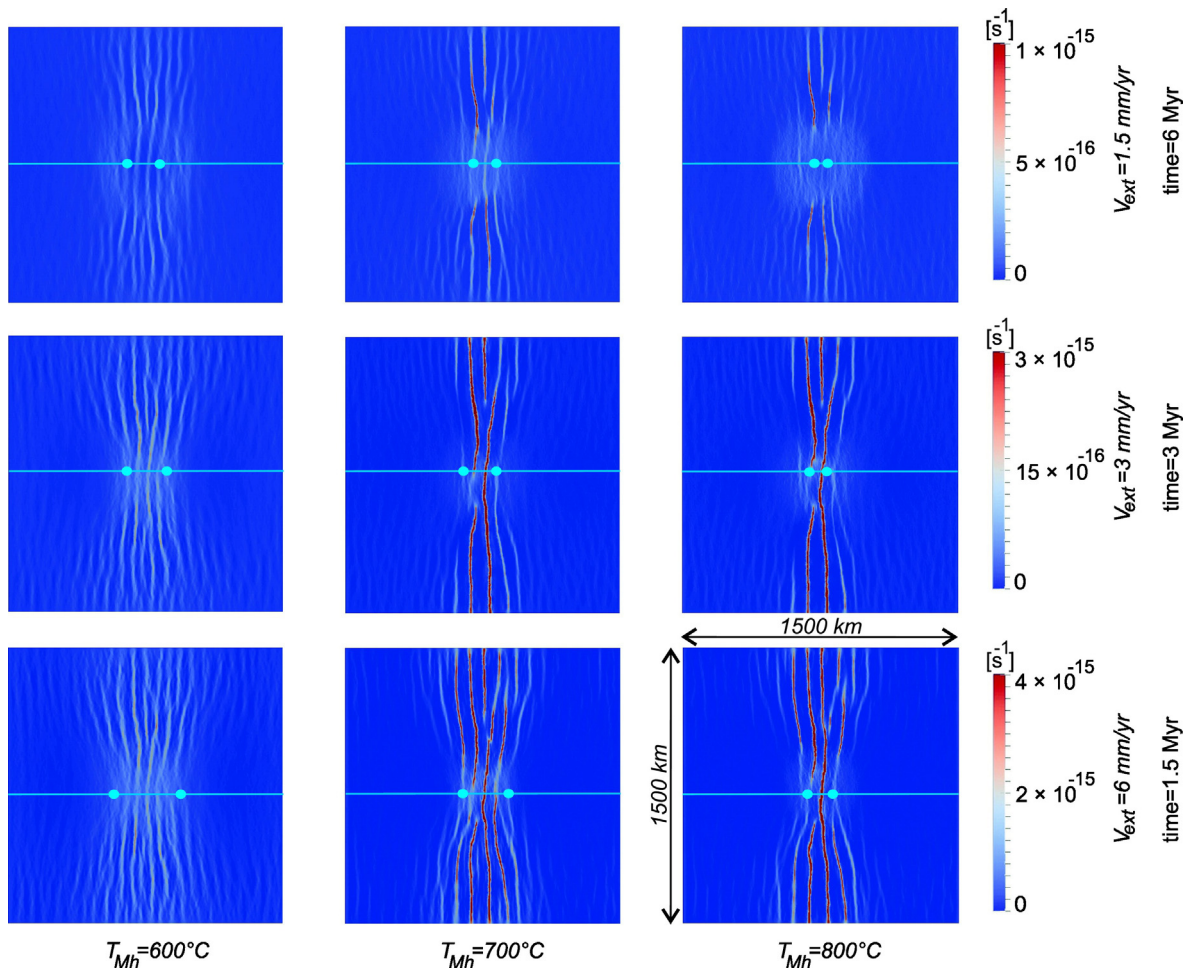
Thus, the initial mode of crustal deformation is crucial for the spatial location of future narrow rifts and spreading centers. In the case of core complex-like extension, the break-up location is likely to be shifted with

respect to the area of initial plume impingement. On the contrary, a wide rift mode results in narrow rifting and break-up above the plume head center (Supplementary Fig. 3).

The onset of the narrow rifting stage is triggered by the isolated uplift of hot plume material along narrow stretched zones (Fig. 3f; Fig. 6f). Fig. 12 summarizes the timing for initiation of this localized plume ascent. It increases linearly from 50 Myr through 65–80 Myr to 95 Myr with decreasing Moho temperature from 800 °C to 500 °C (models 5, 1, 4 and 3, respectively; see white triangles and squares on Fig. 12). However, the model with a mafic lower crust (700 °C; model 2; see gray square on Fig. 12) shows a rapid transition to narrow rifting (55 Myr), considerably faster than in model 3 (500 °C), which is similar in terms of initial deformation mode (wide rift). Thus, at constant stretching rate, the timing of lithospheric break-up is controlled by the lithospheric geotherm rather than by the mode of initial crustal deformation. Yet, increasing extension rate (models 8; black triangle on



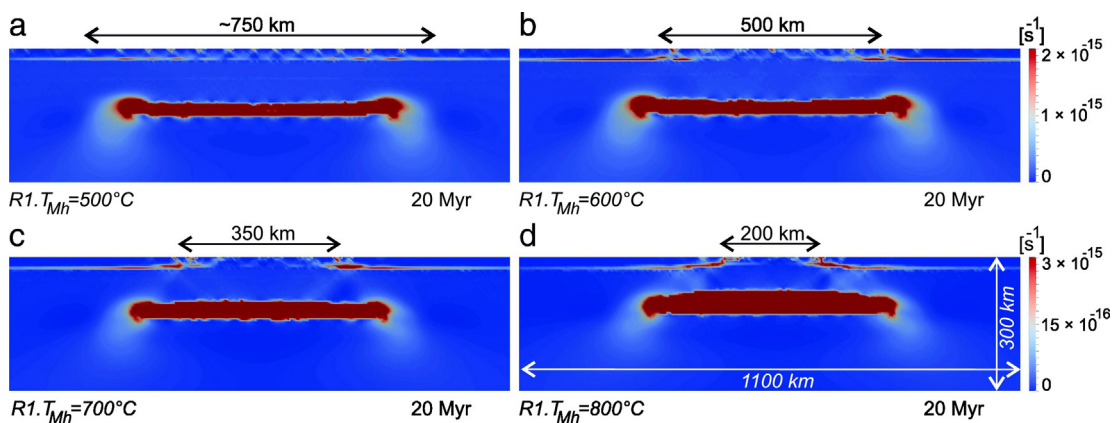
**Fig. 8.** Strain rate cross-sections through a) model 3 (Moho temperature of 500 °C); b) model 4 (Moho temperature of 600 °C); c) reference model 1 (Moho temperature of 700 °C) and d) model 5 (Moho temperature of 800 °C) at 3 Myr. Note that the area of brittle upper-crustal deformation narrows systematically with increasing Moho temperature.



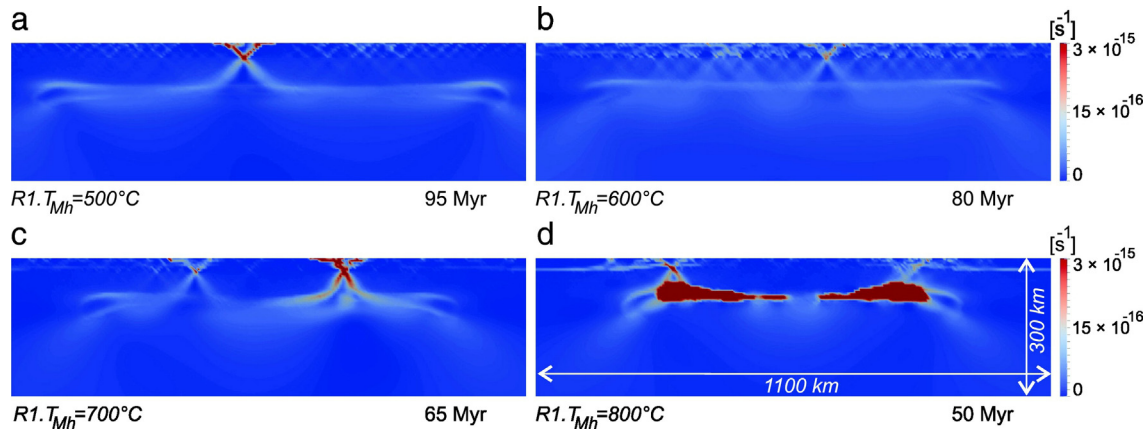
**Fig. 9.** Strain rate distributions at 10 km depth (the upper crust) for models **1**, **4–6** and **8–12** characterized by different Moho temperatures ( $T_{Mh}$ ): 600 °C (left column), 700 °C (middle column) and 800 °C (right column) and far-field extension rates ( $V_{ext}$ ): 1.5 mm/yr (upper row), 3 mm/yr (middle row) and 6 mm/yr (lower row). Elapsed time intervals are chosen to ensure identical amount of extension for all models shown. Blue circles indicate the borders of deformation zones in the central (narrowest) model sections (blue lines). Note the systematic increase of the width of the deformation zone with decreasing geotherm and increasing far-field extensional rate.

Fig. 12) significantly decreases the time between plume impact and break-up. Doubling the extension rate from 3 mm/yr (model **1**) to 6 mm/yr (model **8**) indeed reduces the time to onset of narrow rifting time (65 Myr and 20 Myr, respectively), while a reduction of the half rate from 3 to 2 mm/yr (model **7**) results in considerable prolongation

of the initial stage of deformation (up to 90 Myr). An important consequence is that, for “active-passive” rifting, far-field extension rates cannot be directly derived from geological observations of rift basin opening because of the important role played by active mantle upwelling in the force balance.



**Fig. 10.** Strain rate cross-sections through a) model **3** (Moho temperature of 500 °C); b) model **4** (Moho temperature of 600 °C); c) reference model **1** (Moho temperature of 700 °C) and d) model **5** (Moho temperature of 800 °C) at 20 Myr. Note the core complex mode of extension for model **1** (700 °C) and **5** (800 °C), wide rift mode for model **3** (500 °C), and intermediate mode for model **4** (600 °C). The width of the rift basin is governed by the lateral spreading of mantle plume material.



**Fig. 11.** Strain rate cross-sections through a) model 3 (Moho temperature of 500 °C); b) model 4 (Moho temperature of 600 °C); c) reference model 1 (Moho temperature of 700 °C) and d) model 5 (Moho temperature of 800 °C) at the time of narrow rift localization. Note the (quasi)centered position of narrow rift in case of (quasi)wide rift mode of initial crustal deformation (500 °C and 600 °C at Moho depth; models 3 and 4, respectively), and considerably shifted narrow rifts for core complex-like models (700 °C and 800 °C at Moho depth; models 1 and 5, respectively).

### 3.4. Spatial position of narrow rifts/spreading axis

As mentioned above, narrow rifting tends to localize directly above the plume head center in models with relatively high rheological coupling between the upper crust and the mantle lithosphere. On the contrary, a rheologically decoupled lithosphere favors considerable (~100 km) shift of the break-up center(s) with respect to the area of initial plume impingement.

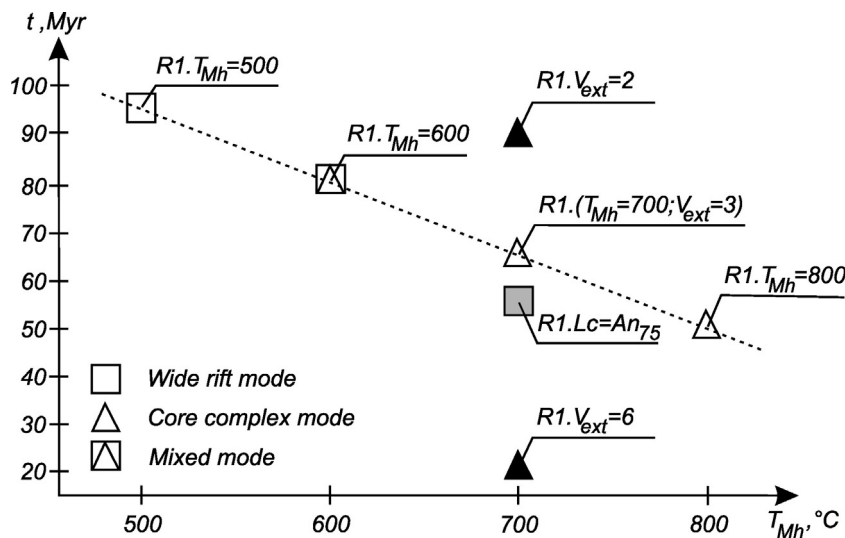
Additional experiments that combine a rheologically decoupled continental lithosphere with faster far-field extension (models 8, 14) and decreased plume density (models 13, 14), result in more complex break-up configurations (Fig. 13d–f). Models with faster extension rate (models 8, 14) shows two en echelon overlapping spreading zones interconnected by a > 200 km long orthogonal transfer zone (Fig. 13d,f) that could be a precursor structure for oceanic transform faults observed in nature. Model 13, with a lighter mantle plume together with the reference far-field extension value (Fig. 13e) evolves into an interesting pattern with an “eastern” and overlapping “western” elongated break-up zone that resemble the two branches observed in the East African rift around the Tanzanian craton (Fig. 14a).

These results illustrate that the geometry of the plume-induced rifting/break-up pattern can be strongly variable and asymmetric even in the absence of inherited crustal/lithospheric heterogeneities. The weakening effect (i.e. plastic strain softening (e.g. Huismans and Beaumont, 2003), see Supplementary methods) amplify initial small perturbations that are introduced by the randomly distributed markers, thus, resulting in the asymmetry that seems to be an intrinsic characteristic of plume-induced rifting/spreading process that develops spontaneously from initially symmetrical plume-lithosphere geometry. The final position of the continental break-up zone(s) is likely to be offset with respect to the plume impingement and, consequently, does not necessarily coincide with the area of pre-rift uplift and magmatism.

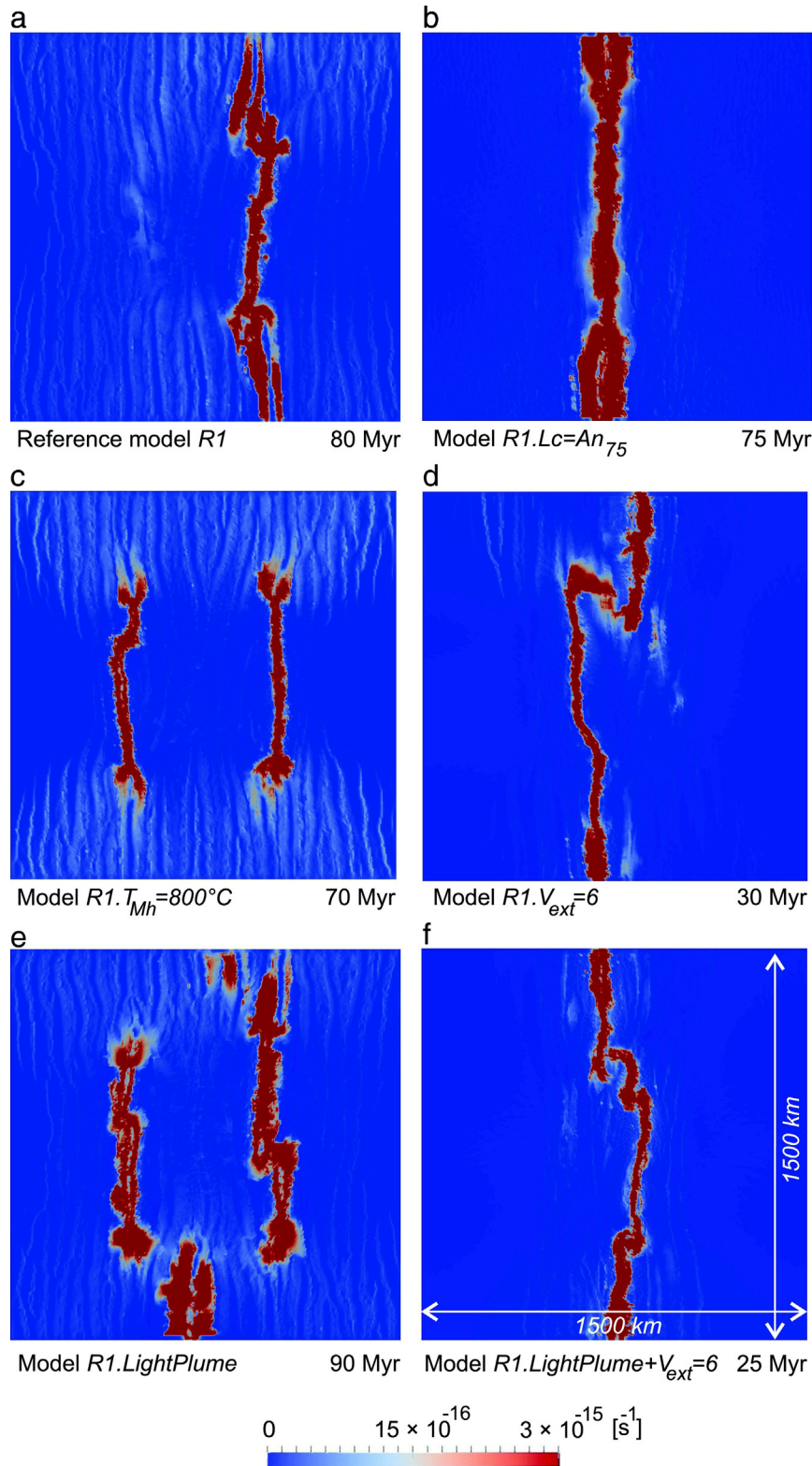
## 4. Discussion

### 4.1. Two-stage continental rifting

Rapid transfer from extensional deformation on normal faults bounding the rift valley to localized strain accommodated within a



**Fig. 12.** Timing of the transition from wide to narrow rift as a function of initial Moho temperature. Triangle and squares represent core complex and wide rift modes of initial crustal extension, respectively, whereas triangle inside square refers to mixed mode. White triangles and squares show models with weak felsic lower crust and the reference value (3 mm/yr) of far-field extension (models 1, 3–5). Gray square represents model 2 characterized by a stronger mafic lower-crustal rheology; black triangles show the models with different external spreading half-rate (2 and 6 mm/yr; models 7 and 8, respectively). Note the linear dependence between time and Moho isotherm, strongly non-linear effect of the variations in far-field velocities and close time for different modes of extension (compare the reference model 1 (white triangle at 700 °C) and model 2 (gray square)).

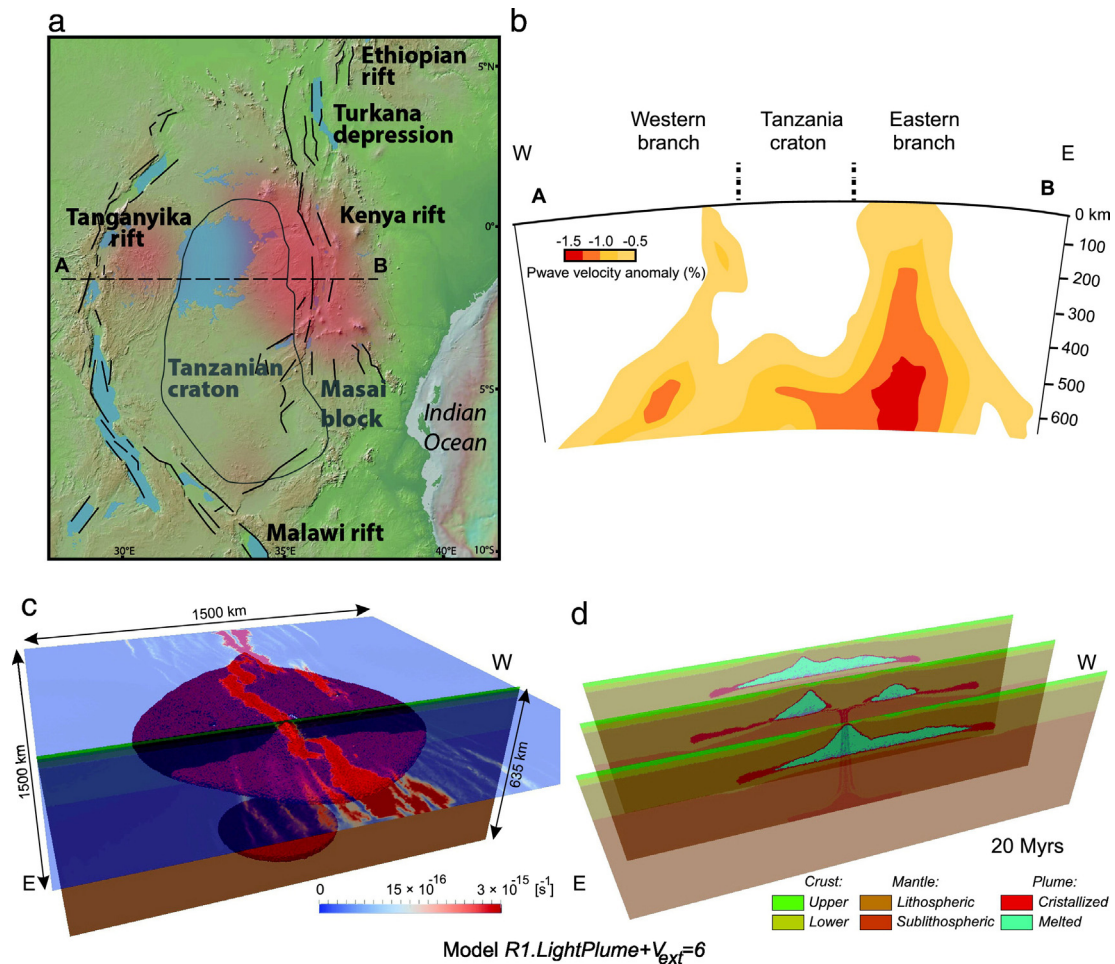


**Fig. 13.** Variability of final continental break-up geometries in models 1–2, 5, 8, 13–14. Models 1 (a) and 2 (b) are characterized by single break-up zone; model 5 (c) demonstrates two symmetrical spreading centers; in case of models 8 (d) and 14 (f), two overlapping break-up zones are interconnected by strike-slip faults; model 13 (e) shows the most complex pattern resembling the configuration of East African rifts around Tanzanian craton (see Fig. 14a).

narrow zone associated with volcanic activity in the rift valley has been argued for in the Main Ethiopian rift (Ebinger and Casey, 2001). Present-day strain accommodation along narrow magmatic Wonji segments may be related to magma-assisted rifting by repeated magma intrusion (Kendall et al., 2005, 2006). Alternatively, this strain transfer from

border faults to Wonji segments may be controlled either by rift kinematics variations (Boccaletti et al., 1999) or by obliquity of far-field extension to the trend of the rift (Corti, 2008, 2009).

Rheologically decoupled model 1, however, shows that the rapid transition from wide rift to strain localization along narrow zones can simply



**Fig. 14.** a) Tectonic setting of the central part of the East African Rift System: Tanzanian craton embraced by active magmatic (eastern) and amagmatic (western) rift branches. Black lines indicate major faults (Corti et al., 2013). The gray line shows the edges of the Tanzanian craton. Red color indicates low velocity zones imaged at the depth of 200 km by S-wave tomography (O'Donnell et al., 2013); b) east-west seismic P-wave velocity mantle tomography profile (Mulibo and Nyblade, 2013) showing two zones of hot material around Tanzanian craton: more intensive negative anomaly underneath magmatic Eastern branch and less intensive one below amagmatic Western branch; c) 3D view of the main features for the model **14**; d) vertical cross-sections of the material phase distribution for the model **14**. Note that very trivial in terms of initial setting model **14** evolves into complex strain and material phase patterns providing good fit with observed contrasted East African rift branches.

result from plume uplift triggering a transition from in-sequence crustal scale core complex to narrow rift mode of deformation (Fig. 1c; Fig. 3f). Models show that this transition takes place considerably earlier (~15 Myr) than the penetration of melted plume material in the crust (Fig. 1d; Fig. 3g) and does not require pre-defined oblique heterogeneities or variations in rift kinematics. Consequently, the polyphase evolution of the Main Ethiopian rift may be explained in the framework of self-induced processes resulting from the interaction between a mantle plume and a homogeneous continental lithosphere under constant far-field extension.

Model **2** (strongly coupled lithosphere) where wide rifting is followed by the formation of narrow rift can be compared with the southern Basin and Range, where the Gulf of California (Henry, 1989; Lizarralde et al., 2007; Ferrari et al., 2013) and Rio Grande rifts (Morgan et al., 1986; Baldrige et al., 2006) cut through an older network of distributed normal faults (Buck, 1991). Similar migration of fault activity through time can be also observed across the East Shetland Basin in the northern North Sea (Cowie et al., 2005).

#### 4.2. Along-strike and across-strike segmentation

Along-strike and across-strike asymmetry of break-up patterns appears to be intrinsic characteristic of most of our 3D experiments.

Model **2** shows non-uniform ascent of plume material along the future break-up center, which results in the development of separated

magma chambers (Fig. 7). Localized crustal magma reservoirs and their sources in the mantle have been inferred from magnetic fabric studies of magma flow in the East Greenland dyke swarm (Callot et al., 2001; Callot and Geoffroy, 2004). The along-strike tectono-magmatic segmentation of volcanic “passive” margins was interpreted to result from small-scale convection (Geoffroy, 2001, 2005; Callot et al., 2002; Gac and Geoffroy, 2009) resembling localized buoyancy-driven mantle flow, as proposed at the slow-spreading Mid-Atlantic Ridge (Lin et al., 1990). Model **2** shows that localized crustal zones of magma storage may develop as a result of self-induced spatial irregularities along-strike of a plume ascent channelized by localized crustal thinning, in a process that does not require pre-defined (sub)lithospheric density and/or thermal heterogeneities.

This non-uniform localized plume uplift can also reconcile the geophysical and geochemical evidence of a mantle plume and of the time progression of magmatism onset, when it is not consistent with plate motion. For example, the High Lava Plains (western part of Yellowstone hot spot track) progresses to the northwest over time in a direction that is not consistent with North American plate motion (Jordan et al., 2004; Meigs et al., 2009; Wagner et al., 2010). Model **2**, with along-strike sinking of the plume apex from 35 km to 70 km (Fig. 7d), provides a scenario that explains such time-progressive tracks of plume-related volcanism solely in terms of plume-lithosphere interaction without requiring the apparent plume track to align with plate motion.

Faster localized uplift of the plume material at the distal point from the plume center is consistent with the timing of magmatic and tectonic events in the South Atlantic where seafloor spreading propagated northward up to the Tristan da Cunha hot-spot (Franke, 2013).

Across-strike segmentation is represented in our models by complex asymmetric systems characterized by various velocities of plume material uplift along different localized axes leading to contrasting rifting. The East African rift system can be considered as a natural example of asymmetrical continental rifts that develop in the context of plume-lithosphere interaction (Fig. 14a–b). The simultaneous development of amagmatic and magmatic rift branches observed at opposite sides of the Tanzanian craton has been interpreted to result from the interaction of a plume with a craton (Koptev et al., 2016). Yet, the more simple and initially symmetrical model 14 (Fig. 13f; Fig. 14c–d) also reproduces a setting that bears similarities with the central East African rift without requiring a laterally heterogeneous lithosphere or an ad-hoc asymmetrical plume position with respect to a craton (Koptev et al., 2015). Another simple setup (model 13, Fig. 13e) results in a final break-up pattern that matches well that of the central East African rift system: a straightened eastern branch and curved western branch propagating further to south into the Malawi rift (Fig. 14a).

It is generally accepted that the geometry of the rifting/break-up pattern should be largely controlled either by pre-existing crustal and/or lithospheric heterogeneities (e.g., Ring, 1994; Corti et al., 2007; Bonini et al., 2007; Guillou-Frottier et al., 2012) or by the spatial position of hot mantle material ponding at the lithosphere-asthenosphere boundary with respect to lithospheric heterogeneities (Burov and Gerya, 2014; Koptev et al., 2016). Our models, however, generate spatial complexity similar to that observed at natural rifts without requiring that preexisting crustal or lithospheric structures control rift initiation and development.

#### 4.3. General discussions

Buck (1991) classifies the extension of continental lithosphere into three modes: narrow rift, wide rift, and core complex. A high-strength lithospheric mantle causes localized brittle deformation and the formation of a narrow rift. On the contrary, a ductile sub-Moho mantle leads to wide rifting or core complex modes of extension (Buck, 1991; Brun, 1999, 2002; Gueydan et al., 2008). Here we show that extension of a continental lithosphere with a high-strength uppermost mantle can initially result in both wide rift and core complex crustal deformation modes. Ultra-slow far-field extension applied to a homogenous lithosphere does not appear to be sufficient for propagation of large faults into the lithospheric mantle. As a result, during initial stage of extension, deformation concentrates in the crust with diffuse brittle strain in both upper and lower crust (wide rift mode) or localized brittle deformation in the upper crust underlain by lower crust subjected to ductile flow (core complex mode). Further extension and/or localized ascent of hot plume material results in concentrated deformation that eventually crosscuts the whole lithosphere and marks the transition to narrow rifting. Previous models assumed considerably larger far-field forcing (compared to the ultra-slow extension applied here, consistent with geodetic observations of present-day rifting) and/or pre-imposed lithospheric weaknesses. As a result, they lead to fast strain localization in the brittle sub-Moho mantle so that the initial stage of crustal extension observed here is absent. In contrast, our experiments predict a long (up to ~100 Myr) stage of extensional deformation in the crust, either in wide rift or core complex mode, before strain localization into a narrow rift. In our models, the thermal impact of the upwelling plume material under ultra-slow extension determines crustal strain localization within a laterally homogenous lithosphere, as also proposed by Burov and Gerya (2014). However, plume-induced perturbations are not sufficient for faulting to propagate through the high-strength lithospheric mantle. As a result, the narrow rift-like continental break-up at the final stage of model development is preceded by a long-lasting stage of wide rifting,

possibly with the development of in-sequence core complex structures, depending on the rheology of the crust. The Basin and Range and the Main Ethiopian Rift may be natural examples of regions that have went through those two modes of extension, with narrow rift propagation following either distributed crustal extension (wide rift mode) or localized faulting at the boundaries of wide rift depression (core complex mode). The evolution of “passive” margin magmatism where pre-break-up volcanism over a large area (up to  $10^6$  km in case of the North-Atlantic/Labrador-Baffin system) predates syn-breakup magmatic activity concentrated at the continent-ocean boundary (Geoffroy, 2001; Geoffroy et al., 2015), is also consistent with this two-stage development.

Numerous examples have proven that the degree of plate coupling have major implications on the deformational processes in extensional and compressional systems (e.g. Beaumont and Quinlan, 1994; Brun, 2002; Willingshofer et al., 2013; Jammes and Huisman, 2012; Vogt et al., 2017; Le Pourhiet et al., 2017). In our models, the style of initial crustal deformation (in-sequence core complex or synchronous distributed faulting) is controlled by the rheological coupling between the upper crust and the lithospheric mantle. When a weak lower crust acts as a decoupling layer, brittle strain in the upper crust is strongly localized due to compensating ductile flow of lower-crustal material (core complex mode). On the contrary, a strong lower crust that efficiently couples the upper crust to the mantle lithosphere leads to distributed brittle deformation in both the upper and lower crust (wide rift mode). A variety of intermediate cases is possible. For instance, a lowering of the temperature at the bottom of a weak felsic lower crust expectedly leads to a gradual increase of the width of core complex-like deformation area, which results in wide rift mode for the coldest geotherm. These results are consistent with inferences derived from analogue modeling (Brun, 2002).

We observe in our models that the location of the final narrow rift, i.e. the breakup location, is governed by the mode of initial crustal deformation. A strong lower crust favors continental break-up coincident with the region of maximum cumulative deformation above the plume head center. On the contrary, a weak lower crust usually results in break-up that is laterally shifted with respect to the initial plume position. In this case, strain localization in the lithospheric mantle is controlled by partial melting, crystallization fractionation, and other processes within the mantle plume ponding lithosphere-asthenosphere boundary. This results in a significant offset between the area of initial surface response to the plume impingement and the final axis of continental break-up.

Most of our 3D experiments, as well as other 3D experiments published in the last 10 years (Brune et al., 2012; Le Pourhiet et al., 2012; Brune and Autin, 2013; Gerya, 2013; Liao and Gerya, 2014, 2015; Brune et al., 2014, 2017; Le Pourhiet et al., 2017), produce self-induced temporal and spatial (across-strike and along-strike) break-up irregularities that were only produced in 2D using very strong lateral mechanical heterogeneities. Without dismissing that lithological heterogeneities are present in the crust and the lithosphere, our work provides new elements to evaluate their actual importance in the dynamics of continental rifting.

#### 5. Conclusions

Despite controversy concerning causes for continental break-up that persists between far-field lithospheric extension and mantle (super) plume(s) (Anderson, 1994; White and McKenzie, 1995; Storey, 1995; Courtillot et al., 1999; Anderson, 2000; Dalziel et al., 2000; Foulger and Hamilton, 2014), our previous studies (Burov and Gerya, 2014; Koptev et al., 2015, 2016) have permitted to reconcile the active and passive models.

The present study identifies the importance of the vertical stratification of lithospheric rheology on the pattern and evolution of continental rifting under the presence of a plume and weak far-field extension.

1. Continental rifting in our models always occurs in two stages, irrespective of the initial rheological stratification of the lithosphere: an early crustal rifting stage and a later lithospheric necking (breakup) stage. The crustal rifting stage influences the lithospheric breakup stage because it pre-structures the crust differently.
2. A rheologically decoupled lithosphere (i.e. with a weak lower crust) favors core complex extension mode with localized upper-crustal brittle strain while the lower crust is subject to compensating ductile flow. On the opposite, rheological coupling between the upper crust and the mantle lithosphere leads to diffuse brittle deformation (wide rift mode). Thus, plume-induced initial strain localization is not only restricted to the brittle upper crust but also requires the lithosphere which is rheologically decoupled by a ductile lower crust.
3. Plume-induced thermal perturbations are not sufficient for rapid faulting through high-strength lithospheric mantle. We observe that concentrated deformation along narrow faults zones cutting through the whole lithosphere is preceded by long-term (up to ~100 Myr long) stage of wide rift or core complex-like crustal extension. In the models, the transition to the subsequent narrow rift stage is triggered by localized ascent of plume material. Narrow rift localization following core-complex mode of extension seems to be governed by complex processes within the deep mantle plume, thus resulting in break-up axes considerably shifted with respect to the area of initial plume impingement. On the opposite, wide rifting favors more a predictable final position of the break-up axis close to the area of maximum cumulative deformation above the plume source center.
4. Most of the models with a rheologically decoupled lithosphere show a highly asymmetrical narrow rifting/spreading system that develops spontaneously from symmetrical and laterally homogenous experimental setups. This self-induced asymmetry seems to be an intrinsic characteristic of the rifting and break-up processes associated with the plume-lithosphere interaction under weak and constant far-field extension.

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## Appendix A. Supplementary Data

Supplementary data to this article can be found online at <http://dx.doi.org/10.1016/j.tecto.2017.03.025>.

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