



**Contrasted continental rifting via plume-craton  
interaction: applications to Central East African rift**  
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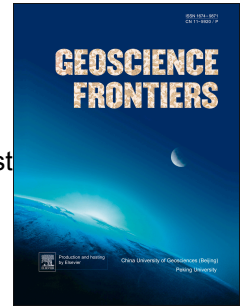


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# Accepted Manuscript

Contrasted continental rifting via plume-craton interaction: applications to Central East African rift

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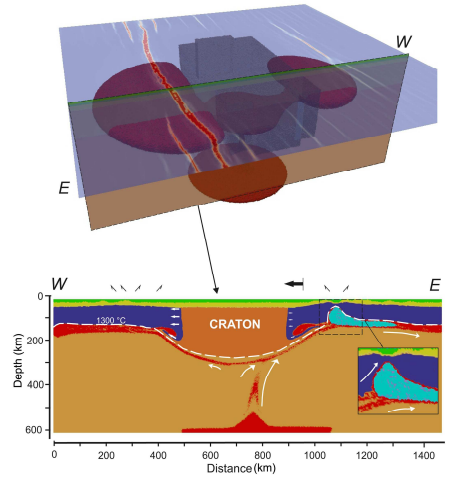
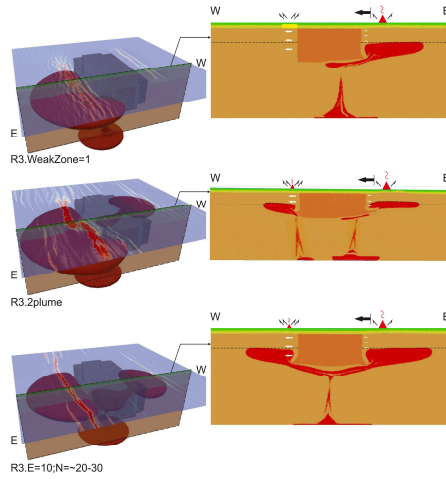
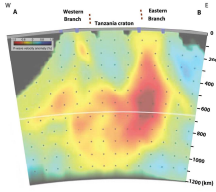
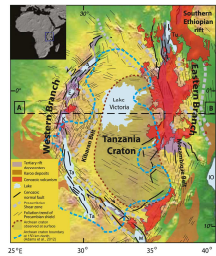
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ACCEPTED MANUSCRIPT

1           **Contrasted continental rifting via plume-craton interaction:**  
2                           **applications to Central East African rift**

3  
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17  
18           **Abstract**

19  
20           The East African Rift system (EARS) provides a unique system with the juxtaposition  
21   of two contrasting yet simultaneously formed rift branches, the eastern, magma-rich, and the  
22   western, magma-poor, on either sides of the old thick Tanzanian craton embedded in a  
23   younger lithosphere. Data on the pre-rift, syn-rift and post-rift far-field volcanic and tectonic  
24   activity show that the EARS formed in the context of the interaction between a deep mantle

25 plume and a horizontally and vertically heterogeneous lithosphere under far-field tectonic  
26 extension. We bring quantitative insights into this evolution by implementing high-resolution  
27 3D thermo-mechanical numerical deformation models of a lithosphere of realistic rheology.  
28 The models focus on the central part of the EARS. We explore scenarios of plume-lithosphere  
29 interaction with plumes of various size and initial position rising beneath a tectonically pre-  
30 stretched lithosphere. We test the impact of the inherited rheological discontinuities (suture  
31 zones) along the craton borders, of the rheological structure, of lithosphere plate thickness  
32 variations, and of physical and mechanical contrasts between the craton and the embedding  
33 lithosphere. Our experiments indicate that the ascending plume material is deflected by the  
34 cratonic keel and preferentially channeled along one of its sides, leading to the formation of a  
35 large rift zone along the eastern side of the craton, with significant magmatic activity and  
36 substantial melt amount derived from the mantle plume material. We show that the observed  
37 asymmetry of the central EARS, with coeval amagmatic (western) and magmatic (eastern)  
38 branches), can be explained by the splitting of warm material rising from a broad plume head  
39 whose initial position is slightly shifted to the eastern side of the craton. In that case, neither a  
40 mechanical weakness of the contact between the craton and the embedding lithosphere nor the  
41 presence of second plume are required to produce simulations that match observations. This  
42 result reconciles the passive and active rift models and demonstrates the possibility of  
43 development of both magmatic and amagmatic rifts in identical geotectonic environments.

44

45 *Keywords:* plume-lithosphere interaction, continental rifting, East African Rift System, 3D  
46 numerical modeling.

47

## 1. Introduction

48  
49  
50 Rifting of continental lithosphere is a fundamental geodynamic process that controls  
51 the growth and evolution of continents and the birth of ocean basins (e.g., Buck, 1991; Buck,  
52 2007). It involves the entire mantle-lithosphere system through heat transfer, active or passive  
53 mantle flow and magmatism, stretching and thinning of the crust/upper mantle due to far-field  
54 forces, and, possibly, viscous coupling between mantle flow and lithospheric deformation. In  
55 the active or plume scenario, rifting occurs as a result of dynamic stresses imparted by large  
56 mantle diapirs or sheet-like mantle upwelling, rising through the mantle, that advect sufficient  
57 heat to produce large amounts of surface volcanism (Sengör and Burke, 1978). In the passive  
58 or plate scenario, rifting occurs as a result of tensional intra-plate far-field forces transmitted  
59 within lithospheric plates, while mantle upwelling and melting is a consequence of  
60 lithospheric stretching (McKenzie, 1978). At oceanic spreading centres the major driving  
61 mechanism is mantle upwelling. However, ridge push forces associated with mantle  
62 upwelling and near ridge topographic gradients are either initially smaller or become  
63 progressively smaller than the far-field forces associated with slab pull of the subducting  
64 lithosphere (Olson et al., 2001). Hence, at large scales, even oceanic rifting would be  
65 impossible without far-field forces driving mature oceanic lithosphere away from the ridge,  
66 thus allowing for continuous creation of space for the accretion of new lithosphere. In  
67 continents, rifting and passive margin development concepts and models are based on the so-  
68 called “passive rifting” mechanism where mantle upwellings are not playing a significant role  
69 in rift dynamics (Buck, 1991; Burov and Poliakov, 2001; Whitmarsh et al., 2001; Huisman  
70 and Beaumont, 2003; Buck, 2007; Cloetingh et al., 2013; Burov et al., 2014). However, it is  
71 generally accepted that some continental rifts, such as the Afar and Central Africa rift regions

72 (Ritsema et al., 1999; Nyblade et al., 2000) or the Rio Grande system (Satsukawa et al., 2011)  
73 involve significant mantle contribution. The dynamic contribution of large mantle upwellings  
74 in other continental rifts such as Baikal (e.g., Burov et al., 1994; Petit et al., 1997), the Rhine  
75 graben, or the Pannonian basin (e.g., Cloetingh et al., 1999) is still debated due to the absence  
76 of the records of magmatic pre-rift activity and clear seismological signatures of deeply-  
77 rooted mantle upwellings. However, while the observational signature of mantle-induced  
78 rifting is often equivocal, a purely passive mechanism also meets a number of problems,  
79 specifically in zones of ultra-slow rifting. This issue was termed the “tectonic force paradox”,  
80 which states that far-field tectonic forces transmitted in the lithosphere are not sufficient to  
81 rupture normal continental lithosphere, unless it is previously weakened (e.g., Behn et al.,  
82 2006; Buck, 2006, 2007). Indeed, simple estimation of the forces needed to extend lithosphere  
83 (Buck, 2006) shows that for reasonable driving force levels ( $5 \times 10^{12} \text{ N m}^{-1}$ ), only extremely  
84 thin lithosphere ( $< 30 \text{ km}$ ) can be rifted tectonically in absence of magmatic dyke intrusion.  
85 Hence, it is often suggested that continental rifting and breakup either require meso-scale  
86 strain softening (Behn et al., 2006; Buck, 2007; Huisman and Beaumont, 2007; Precigout et  
87 al., 2007) or additional strain localization mechanisms associated, for example, with the  
88 interactions between mantle plume and the overlying lithosphere (Burov et al., 2007; Burov  
89 and Gerya, 2014; Koptev et al., 2015; Stamps et al., 2015).

90 The role of mantle upwellings in continental rift dynamics is a long debated topic,  
91 illustrated by the classic “passive versus active” rifting debate (Fitton, 1983; Foulger et al.,  
92 2000; Foulger and Hamilton, 2014). Seismic tomography reveals deep-seated low-velocity  
93 anomalies in the mantle underneath several rift zones (e.g., Ritsema et al., 1999; Nyblade et  
94 al., 2000; Nolet et al., 2006; Nolet et al., 2007) that cannot be interpreted as a consequence of  
95 passive lithospheric stretching. In particular, broad low seismic velocity zones observed  
96 throughout the upper mantle cannot be easily inferred from small-scale mantle convection



97 induced by passive stretching of the lithosphere, for example in the East African (Ritsema et  
98 al., 1999; Nyblade et al., 2000; Adams et al., 2012) or Rio Grande (Satsukawa et al., 2011) rift  
99 systems. Evidence for strain accommodation by magma intrusion in young continental rift  
100 basins (Calais et al., 2008) is also indicative of magma-assisted rifting (Kendall et al., 2005;  
101 Kendall et al., 2006). Finally, the EARS shows petrological evidence for pervasive elevated  
102 mantle temperature under the rift requires significant heating from below and/or fluid-assisted  
103 melting (Rooney et al., 2012; Fergusson et al., 2013; Armitage et al., 2014). These  
104 observations are indicative of a contribution of deep mantle processes in the evolution some  
105 continental rifts, in particular the EARS.

106 Our understanding of rift formation and evolution has matured thanks to our ability to  
107 quantify surface kinematics from geodetic data (e.g., Stamps et al., 2008; Saria et al., 2014),  
108 seismically image crustal and lithospheric structures (Whitmarsh et al., 2001; Buck 2007;  
109 Nolet et al., 2007), and to model the mechanical behaviour of a rheologically layered  
110 lithosphere in physically consistent frameworks (Burov and Poliakov, 2001; Huisman and  
111 Beaumont, 2003; Burov et al., 2007; Huisman and Beaumont, 2007; Guillou-Frottier et al.,  
112 2012; Burov et al., 2014; Burov and Gerya, 2014; Koptev et al., 2015). However, most  
113 continental rift models explore 2D passive rifting scenarios, probably because most of the  
114 available observational data is derived from studies of passive margins, where the records of  
115 the initial stage of rifting are buried under thick post-rift sedimentary sequences, or to fossil  
116 rifts, where the tectonic, thermal, and magmatic signatures of rifting have long decayed away.  
117 The seismically and volcanically active EARS therefore provides a unique complementary  
118 setting of a young and on-going continental rift that develops in the presence of large-scale  
119 mantle upwelling (Lithgow-Bertelloni and Silver, 1998) and slow far-field plate motions  
120 (Stamps et al., 2008; Saria et al., 2014). Here we focus on the central part of the EARS, where  
121 seismic tomography shows warm mantle material rising under the old, thick, and cold

122 Tanzania craton (Mulibo and Nyblade, 2013a; Fig. 1). This context requires 3-dimensional  
123 thermo-mechanical models with sufficiently high resolution to accurately capture strain  
124 localization in the brittle crust, as shown in a previous, generic, study (Koptev et al., 2015).  
125 Here we follow-up on this study with a series of numerical experiments that explore various  
126 boundary conditions and initial geometrical and thermo-rheological configuration of the rift  
127 system. We test the upwelling of more than one upper mantle plume below the central EARS.  
128 Indeed, body-wave tomography shows strong indication of small-scale upper-mantle plumes  
129 rising from a single lower-mantle mega-plume ponded below the 660 km phase transition  
130 boundary (Mulibo and Nyblade, 2013a,b) . We also tested the impact of plume size,  
131 temperature, composition and initial position below the Tanzanian craton, as well as the  
132 impact of the lithosphere structure and of inherited structures such as rheologically weakened  
133 suture zones along the borders of the craton.

134

## 2. Geological settings and data

135

136

137 The EARS is a linear active volcano-tectonic structure that cuts across the 1300 km-  
138 wide, 1100 m-high Ethiopian and East African plateaux (Fig. 1), whose high elevation is  
139 dynamically supported by whole-mantle convective upwelling (Lithgow-Bertelloni and  
140 Silver, 1998; Nyblade et al, 2000) that initiated at 30–40 Ma (Burke, 1996; Ebinger and Sleep,  
141 1998). Passive mechanisms of EARS formation due to gravity-driven far-field forces caused  
142 by crustal thickness gradients have been also considered in earlier studies (e.g., Logatchev et  
143 al., 1972).

144 The lithospheric structure of the African continent is highly heterogeneous as many  
145 old suture zones of Proterozoic mobile belts were reactivated as rifts during the Paleozoic and  
146 Cretaceous (Burke, 1996). Small yet well-preserved thick cratons such as the Tanzania,  
147 Congo and Kaapvaal cratons are found throughout the EARS. These cratons, characterized by  
148 greenstone belts, tonalites, and various other high-grade metamorphic rocks, may play an  
149 important role in the localization and reactivation of deformation thanks to rheological  
150 contrasts with ancient suture zones running along their borders (e.g., McConnell, 1972; Mohr,  
151 1982; Morley, 1988; Versfelt and Rosendahl, 1989; Ring, 1994; Corti et al., 2007; Guillou-  
152 Frottier et al., 2012).

153 Two eastern and western rift branches in the central EARS are superimposed onto  
154 sutures and shear zones formed by Proterozoic mobile belts that embrace the rigid Archean  
155 Tanzanian craton. Intense magmatism and continental volcanism are largely present in the  
156 eastern rift branch, while other branches such as the western rift to the west of the Tanzanian  
157 craton and the Malawi rift to the south, show only small amounts of Cenozoic volcanics. The  
158 eastern rift is characterized by a southward progression of the onset of volcanism (Baker,  
159 1987; Ebinger, 1989; Forster et al., 1997; George et al., 1998), with widespread extension and

160 uplift of rift shoulders between 30 and 20 Ma (Ebinger et al., 1989; Morley et al., 1992;  
161 McDougall et al., 2009; Wichura et al., 2011) and the establishment of localized rift basins  
162 around 20 Ma (Ebinger, 1989; Wolfeden et al., 2004; Chorowicz, 2005; Stab et al., 2015).  
163 Using a combination of detrital zircon geochronology, tephro- and magnetostratigraphy,  
164 Robert et al. (2012) documented the synchronous initiation and development of volcanism  
165 and basin development in the western and eastern branches of the EARS, in contrast to  
166 previous geological models that inferred a considerably younger western rift that initiated  
167 around 12 Ma only (Ebinger et al., 1989; Cohen et al., 1993; Lezzar et al., 1996; Tiercelin and  
168 Lezzar, 2002).

169 Most of the seismicity of the central EARS is concentrated in the narrow, amagmatic  
170 Western rift, with hypocenters reaching depths of 30–40 km and large normal faults indicative  
171 of large historical events (Yang and Chen, 2010; Moucha and Forte, 2011). In contrast, the  
172 magma-rich eastern rift is characterized by earthquake hypocenters confined to the upper ~15  
173 km and heat flow anomalies reaching  $110 \text{ mW/m}^2$  (Nyblade, 1997). These two rift branches  
174 are separated by a relatively aseismic domain centred on the 2.5–3 Ga old Tanzanian craton  
175 where seismic (Ritsema et al., 1998; Nyblade et al., 2000; Nyblade and Brazier, 2002;  
176 Weeraratne et al., 2003; Venkataraman et al., 2004; Adams et al., 2012), xenolith (Chesley et  
177 al., 1999; Lee and Rudnick, 1999), and gravity (Petit and Ebinger, 2000) data showed a 170–  
178 250 km-thick keel and a largely resisted extensional Cenozoic tectonism lithosphere that is  
179 colder and stronger than the surrounding orogenic belts. Neogene kinematics of the Nubia-  
180 Somalia plate system refers to 2 mm/yr divergence between the onset of rifting (25–30 Ma)  
181 and 4 Ma, accelerating to 4 mm/yr after 4 Ma (Stamps et al., 2008; Saria et al., 2014).

182 The Tanzanian craton (Fig. 1) is underlain by a broad low seismic velocity anomaly  
183 extending across the 410 km discontinuity down to the transition zone (660 km) (Nyblade et  
184 al., 2000; Huerta et al., 2009; Nyblade, 2011; Mulibo and Nyblade, 2013a,b). This anomaly is

185 indicative of high temperature and melt presence and is consistent with the spreading of a  
186 mantle plume head beneath the craton (Weeraratne et al., 2003; Adams et al., 2012). Below  
187 the transition zone, this plume may connect with the African Superplume, a large-scale low  
188 shear-wave velocity anomaly extending from the core-mantle boundary into the mid-mantle  
189 under eastern Africa (Ritsema et al., 1999; Masters et al., 2000; Mégnin and Romanowicz,  
190 2000; Gu et al., 2001; Grand, 2002) – though seismic data is equivocal (Ritsema et al., 2011;  
191 Simmons et al., 2011). Despite the debate on one versus two mantle plumes below the EARS  
192 based on geochemical (Rogers et al., 2000; MacDonald et al., 2001; Pik et al., 2006; Nelson et  
193 al., 2008; Nelson et al., 2012) and geophysical (Chang and van der Lee, 2011; Hansen et al.,  
194 2012) data as well as on the results of numerical modelling (Ebinger and Sleep, 1998; Lin et  
195 al., 2005), new He, Ar, Nd, Sr and Ne isotopic data and major and trace element compositions  
196 from Neogene volcanics across the EARS suggest a common heterogeneous deep mantle  
197 source for the whole rift system (Furman et al., 2006; Furman, 2007; Chakrabarti et al., 2009;  
198 Hilton et al., 2011; Halldórsson et al., 2014), possibly indicating a source rooted in the  
199 African Superplume (Ershov and Nikishin, 2004; Bagley and Nyblade, 2013) with upward  
200 transport via localized thermal upwellings (Nyblade, 2011; Mulibo and Nyblade, 2013a,b).  
201 Here we take advantage of these recent improvements in our understanding of deep structures,  
202 geological evolution and recent kinematics together with new cutting edge numerical  
203 modelling techniques (Gerya and Yuen, 2007, see Supplementary Methods) to design a 3D  
204 ultra-high resolution viscous plastic thermo-mechanical numerical model that accounts for  
205 thermo-rheological structure of the lithosphere and hence captures the essential geophysical  
206 features of the central EARS.

207

### 3. Numerical model

#### 3.1. 3D Model Setup

We use the staggered grid/particle-in-cell 3D viscous-plastic 3DELVIS code (Gerya and Yuen, 2007), based on a combination of a finite difference method applied on a staggered Eulerian grid with a marker-in-cell technique (see Supplementary Methods for more details).

##### 3.1.1. Spatial dimensions and resolution

The spatial dimensions of the 3D model are  $1500 \text{ km} \times 1500 \text{ km} \times 635 \text{ km}$ . The regular rectangular Eulerian grid of the model domain consists of  $297 \times 297 \times 133$  nodes and offers spatial resolution of  $5 \text{ km} \times 5 \text{ km} \times 5 \text{ km}$  per grid element (Fig. 2). This implies very large mesh dimensions (more than ten million elements and hundred million randomly distributed Lagrangian markers) and hence requires unprecedented numerical efforts. The computations have taken 200 years of cumulated computing time in single CPU core equivalent (with average 4 years of single CPU core time per experiment) on SGI shared (NUMA) fat-node cluster with 2.8 Ghz Intel Xeon CPU cores.

##### 3.1.2. Internal model structure and rheological parameters

The initial model setup comprises a stratified three-layer continental lithosphere composed of an upper and lower crust and lithospheric mantle overlaying the upper mantle. The lithosphere mantle embeds a rectangular ( $800 \text{ km} \times 400 \text{ km}$ ) cratonic block characterized by greater thickness (250 km; Smith, 1994; Ritsema et al., 1998; Mulibo and Nyblade,

233 2013a,b) and smaller density due to its depleted mantle composition (Conolly, 2005). The  
234 total crustal thickness is 36 km. Depth to the bottom of the embedding “normal” lithosphere is  
235 150 km, except for the cases specified in the next section.

236 The mantle plume(s) was (were) initiated by seeding a temperature anomaly(ies) at the  
237 base of the upper mantle. Following Burov et al. (2007) and Koptev et al. (2015), its (their)  
238 starting geometry is modeled as a hemisphere with a radius of 200 km, except for the several  
239 test models with smaller and larger plumes (Table 1). The initial position of the mantle  
240 plume(s) with respect to the craton is one of the parameters tested in this study (see section  
241 3.2).

242 Mantle densities, thermal expansion, adiabatic compressibility, and heat capacity are  
243 computed as function of pressure and temperature in accordance with a thermodynamic  
244 petrology model *Perple\_X* (Conolly, 2005), which insures thermodynamically consistent  
245 variation of material properties, including phase changes. *Perple\_X* was used in all models  
246 except for one experiment specified below. Uncertainties in mineralogical composition may  
247 result in 15–30 kg/m<sup>3</sup> bias in thermodynamic estimates of mantle density (Watremez et al.,  
248 2013). This specifically refers to cratons, whose mantle composition may be subject to larger  
249 variations than normal lithosphere. Accordingly, we artificially decreased the craton density  
250 calculated from the standard petrology model by 15 kg/m<sup>3</sup> to ensure initial isostatic  
251 equilibrium of the system. For the crustal rocks we used a simple Boussinesq approximation  
252 (see Supplementary Table 1) since metamorphic changes in these rocks would be of minor  
253 importance in the context of our problem.

254 A series of numerical experiments also explores the impact of the rheological  
255 properties of the lower crust – wet granite (*WetQz*) or granulite (*An<sub>75</sub>*) – whereas the ductile  
256 part of the upper crust was represented by wet granite (*WetQz*) in all experiments. The latter  
257 assumption is valid since in bi-layer crusts the ductile rheology of the upper crust is of minor

258 importance since the corresponding depth interval of 0–15 km is mainly dominated by rock-  
259 type independent brittle failure (e.g. Burov, 2011). In the models the ductile rheology of the  
260 mantle lithosphere (dry olivine) is controlled by dry olivine dislocation and Peierls creeping  
261 flow, while the sub-lithospheric mantle (dry olivine as well) deforms by diffusion creep  
262 (Caristan, 1982; Karato and Wu, 1993; Durham et al., 2009). The mantle plume is supposed  
263 to be slightly “moist” and has the rheology of wet olivine. The complete list of the rheological  
264 parameters of the model materials is provided in Supplementary Table 1.

265 The effectively free surface topography is implemented by inserting a 30 km thick  
266 low-viscosity “sticky air” layer between the upper interface of the model box and the surface  
267 of the crust. The viscosity of the “sticky air” is  $10^{18}$  Pa s and its density is  $1 \text{ kg/m}^3$ , according  
268 to optimal parameters established in the previous studies (Duretz et al., 2011; Cramer et al.,  
269 2012; Burov and Gerya, 2015).

270

### 271 *3.1.3. Velocity boundary conditions*

272

273 Although some have proposed dominant deviatoric compression acting on the African  
274 plate in consideration of surrounding mid-ocean ridges (e.g., Zeyen et al., 1997), calculations  
275 of deviatoric stresses arising from lateral gradient of gravitational potential energy (GPE) due  
276 to elevation – most of eastern and southern Africa being at elevations  $> 1500 \text{ m}$  – and lateral  
277 density variations show that the EARS undergoes  $\sim 10 \text{ MPa}$  of E–W extensional deviatoric  
278 stresses (Coblentz and Sandiford, 1994; Stamps et al., 2010; Stamps et al., 2014). This is  
279 equivalent to a force per unit length of  $1 \text{ TN/m}$  for a 100 km-thick lithosphere, of the same  
280 order as slab pull forces. This extensional deviatoric stress regime is the source of the far-field  
281 extension, which, in our models, is applied as a kinematic boundary condition.



282 We simulate this weak tectonic forcing by applying a constant divergent velocity  
283 normal to the “eastern” and “western” sides of the model box. Following geological and  
284 geodetic estimates for the EARS extension rates, we varied this velocity between 1.5 and 6  
285 mm/yr (Table 1). The corresponding horizontal forces on the borders of the model are small  
286 (on the order of typical ridge push forces, i.e.  $(1\sim 2)\times 10^{11}$  N per unit length). Free slip  
287 boundary conditions are used on the “northern” and “southern” sides of the model, which are  
288 not subject to extension. Compensating vertical influx velocities through the upper and lower  
289 boundaries are introduced to ensure mass conservation in the model domain (Gerya, 2010).

290

#### 291 ***3.1.4. Initial temperature distribution and thermal boundary condition***

292

293 The initial geotherm is one of the variable parameters of our experiments. In the  
294 reference experiment, the initial geotherm is piece-wise linear, with 0 °C at the surface ( $\leq 30$   
295 km, the air), 400 °C at the upper/lower crustal interface, 700 °C at the Moho, 1300°C at the  
296 bottom of the lithosphere (i.e. deeper below the craton and shallower below the embedding  
297 lithosphere) and 1630 °C at the bottom of the model domain at 635 km depth. The resulting  
298 adiabatic thermal gradient in the mantle is 0.5–07 °C/km. The mantle plume(s) has (have) an  
299 initial temperature of 2000 °C (except for one model, see Section 3.2). We chose an initial  
300 mantle plume temperature of 2000 °C, 300 °C warmer than the surroundings, consistent with  
301 the 20–40 km depression of the 410 km discontinuity observed seismically beneath the  
302 Tanzanian craton (Huerta et al., 2009).

303 The thermal boundary conditions correspond to fixed temperature values at the upper  
304 surface and the bottom of the model (0 and 1630 °C, respectively; Koptev et al., 2015) and  
305 zero horizontal heat flux across the vertical boundaries.

306

### 307 **3.2. Experiments and key variable parameters**

308

309 We tested 34 different experimental settings by varying ten controlling parameters that  
310 characterize the properties of plume(s) and lithosphere and the velocity boundary conditions  
311 (Table 1):

312 1. Number of mantle plumes;

313 2. Initial position of the mantle plume(s) with respect to the craton;

314 3. Initial size of the mantle plume(s);

315 4. Density of the mantle plume(s);

316 5. Temperature of the mantle plume(s);

317 6. Presence, number and shape of the weak rheological interface(s) along the craton  
318 border(s);

319 7. Rheology of the lower crust;

320 8. Craton thickness;

321 9. Normal (non-cratonic) lithosphere thickness;

322 10. Horizontal extension velocity.

323 We started our experiments with a reference model (Fig. 3, model 1.R2) characterized  
324 by a single mantle plume (initial hemisphere radius of 200 km, initial temperature of 2000°C,  
325 and dynamic P-T dependent density structure defined by the thermo-dynamic model  
326 *Perple\_X*, Conolly, 2005). In this reference experiment the initial plume is seeded exactly  
327 below the central part of the craton and the lithosphere does not contain any weak predefined  
328 rheological zones; the rheology of its lower crust refers to wet quartzite (WetQz). The  
329 thicknesses of the “normal” and “cratonic” lithosphere correspond to commonly inferred  
330 values of 150 and 250 km, respectively (e.g., Burov et al., 2007; Guillou-Frottier et al., 2012).

331 The horizontal velocities applied along the “eastern” and “western” sides of the model are 3  
332 mm/yr.

333 The first parameter that has been varied in the experiments is the initial position of the  
334 plume with respect to the center of the craton. The models **2.R3.PosPL=North-East**,  
335 **3.R3.PosPl=North** and **4.R3.PosPl=East** correspond to a lateral shift of the mantle plume to  
336 the NE (225 km), to the north (200 km) and to the east (100 km), respectively. Note that all of  
337 the models listed below refer to either central plume position (prefix R2) or to that  
338 characterized by a north-east shift (prefix R3), except for the models with a different  
339 thicknesses of the “normal” lithosphere (models 28–31), where the initial plume position has  
340 been shifted westward (see below).

341 The next series of experiments (models 5–10) is characterized by a weak narrow  
342 vertical interface(s) between the craton and the embedding lithosphere that mimics suture  
343 zones. These zones have rheological parameters of the upper crust: for the “weak” zone(s)  
344 along long the side(s) of the craton we used granite (WetQz) rheology whereas the  
345 surrounding area of the lower crust consists of mafic ( $An_{75}$ ) rocks. Therefore the implemented  
346 weak zones are not over-softened and are weaker than the surroundings within the depths  
347 below the upper-lower crustal interface. Several models with one and two weak interfaces (in  
348 presence of centered and NE shifted plume) have been implemented (models  
349 5.R2.WeakZone=1, 6.R2.WeakZone=2, 7.R3.WeakZone=1 and 8.R3.WeakZone=2). Also we  
350 have tested the configurations with one weak zone welded into stronger crust within the  
351 western part of model and weak lower crust for the entire opposite half of the studied domain  
352 (model 9.R3.WeakZone=3). Finally, the curved shape of two weak interfaces embracing  
353 cratonic bloc has been tested in the model 10.R2.LongWeakZone.

354 The experiments with two plumes (models 11–17) containing a second mantle plume  
355 shifted to the southwest (225 km) from the center of the craton. The first two experiments of

356 this series (11.R3.2plume and 12.R3.2plume+LowCrust=An75) refer to different rheologies of  
 357 the lower crust (WetQz and An75, respectively) whereas the second plume has the same  
 358 parameters as the first one, except for a slightly smaller radius of 150 km. The next 3 models  
 359 were implemented with the goal to explore the sensitivity of the model 12 to the properties of  
 360 the second plume. In these experiments the second plume has been made, respectively, bigger  
 361 (initial radius of 175 km, model 13.R3.2plume+BigPlume), hotter (initial temperature of  
 362 2100°C, 14.R3.2plume+HotPlume), and lighter (Perple\_X-derived density was artificially  
 363 reduced by 30 kg/m<sup>3</sup>, 15.R3.2plume+LightPlume). Two plumes of the same size (200 km)  
 364 were tested in model 16.R3.2plume.EqualSize whereas model 17.R3.2plume+WeakZone=2  
 365 refers to the additional introduction of two weak interfaces along the craton borders into the  
 366 setup of the model 11.

367 The impact of much smaller initial NE shifts and of slightly bigger ( $r=250$  km) initial  
 368 plume are studied in experiments 18.R3.E=50; N=100 (shift is 112 km; eastward component  
 369 is 50 km, northward component is 100 km), 19.R3.E=5; N=10 (shift is 11 km; eastward  
 370 component is 5 km, northward component is 10 km) and 20.R3.E=10; N= $\sim$ 20–30 (shift is  $\sim$ 25  
 371 km; eastward component is 10 km, northward component is 20–30 km). Model 21.R3.E=10;  
 372 N= $\sim$ 20–30+R=200 refers to a shift of  $\sim$ 25 km for the reference plume size ( $r=200$  km).  
 373 Inserting two weak zones into the model 20 yields model 22.R3.E=10; N= $\sim$ 20–  
 374 30+WeakZone=2.

375 The series of experiments with different thicknesses of “normal” lithosphere (150 km  
 376 within eastern half of model domain, 200 km within western one) starts with a central initial  
 377 position of a plume of reference size ( $r=200$  km) (model 23.R2.H\_lit=150–200). Then we  
 378 sequentially shift the mantle plume to 25, 50 and 100 km westward (models  
 379 24.R3.H\_lit=150–200+W=25, 25.R3.H\_lit=150–200+W=50 and 26.R3.H\_lit=150–  
 380 200+W=100, respectively). We combine a model with a hemispherical plume shifted by 50–

381 km has with a hemi-ellipsoidal plume of bigger size (horizontal radius of 400 km; vertical  
382 radius of 200 km) to design model 27.R3.H\_lit=150–200+E=50+BigPlume.

383 Models 28–31 refer to different sizes ( $r=150, 200$  and  $300$  km) of the north-east  
384 shifted (225 km) plume in presence of different lower crustal rheologies (WetQz or An<sub>75</sub>).

385 Finally, the last 3 models (models 32–34) illustrate the impact of craton thickness  
386 (model 32.R3.H\_crater=200) and velocity boundary conditions (the models with slower  
387 (33.R3.Vext=1.5) and faster (34.R3.LC=An<sub>75</sub>+Vext=6) external extension).

388 Every model run took about 4 years of CPU time (2 month of physical run time on a  
389 shared memory SGI parallel supercomputer).

390

## 4. Experimental results

### 4.1. Reference model (Model 1)

The reference model 1.R2 shows a rapid plume ascent after the experiment onset: the mantle plume reaches the bottom of the cratonic lithosphere in 0.5 Myrs, which is similar to previous models with cratons (Burov et al., 2007; Koptev et al., 2015). The cratonic block causes the plume head to split into two initially nearly symmetrical parts, each of which flows towards the base (LAB) of the “normal” lithosphere near the craton borders (Fig. 3a). As shown in Koptev et al. (2015), brittle strain localization in the crust, initially caused by far-field stresses, is amplified by heat transport and serves to channelize the plume material, without requiring regions of pre-existing thinning or rheological weakness. This channeling helps localizing strain in two symmetric narrow north-south rifts above the zones of plume heads emplacement (Fig. 3b). This positive feedback between lithospheric thinning and channelized flow of the plume material is a key mechanism for strain localization in the models.

The next stage of the system development corresponds to localized ascent of the plume material (at 55 Myr) along the narrow and stretched zones (Fig. 3c) that further leads to fast (<1 Myrs) destruction of the continental crust (at 75 Myr, Fig. 3d) by hot mantle material and transition from pre-breakup rifting to post-breakup spreading (>75 Myr).

Strain distribution within the crust shows two symmetric N–S stretched (i.e. perpendicular to far-field extension) rifting zones appearing simultaneously on either side of craton just above mantle hot material concentrated below the lithosphere-asthenosphere boundary (Fig. 3).

416 **4.2. Initial position of the mantle plume with respect to the craton**417 *(Models 2–4)*

418

419 The initial position of the mantle plume with respect to the craton is one of the most  
420 important parameters tested in this study (Table 1). Figure 4 shows that different initial plume  
421 positions (models 2–4) result in very different evolutions of the system compared to the  
422 reference model (model 1). A common feature of these models is the deflection of plume  
423 material by the craton and the formation of a local uplift centered above a secondary plume  
424 head. The direction of this deviation of the plume material by the cratonic block is controlled  
425 by the initial position of the plume, whereas strain localization within the upper crust  
426 (horizontal slices on the Fig. 4) is conditioned by the spatial distribution of hot material  
427 ponding under “normal” lithosphere. The model with a plume head deflected to the north  
428 (3.R3.PosPl=North, Fig. 4b) shows strain localization within the central part of the model  
429 domain, while models with a east and north-east shifted plume (4.R3.PosPl=East (Fig. 4a) and  
430 2.R3.PosPl=North-East (Fig. 4c)) both show an eastward displaced rift.

431 The central cratonic block, less deformable than the surrounding lithosphere, moves  
432 eastward and rotates slowly anticlockwise (model 2.R3.PosPl=North-East, Fig. 4c). This  
433 rotation is consistent with observed geodetic displacements (Stamps et al., 2008; Saria et al.,  
434 2014). It results from the torque due to asymmetrically distributed forces exerted by the plume  
435 material on the craton keel. The deflection of the plume material towards the eastern rift  
436 basins, together with the lateral motion of the cratonic block driven by the plume, preserves  
437 the craton from thermo-mechanical erosion until the system reaches steady state at 20 Myr  
438 (Sleep et al., 2002). This provides new insights for understanding of the survival of small  
439 cratonic terrains.

440 As shown above, models **2** and **4** (Fig. 4a,c) are in good agreement with the  
441 observations in the EARS, as they reproduce the magmatic eastern branch and of the  
442 anticlockwise rotation of the craton. However, these models do not reproduce the observed  
443 strain localization along the western margin of the cratonic bloc in the western rift branch.

444

#### 445 **4.3. Rheological properties along the craton borders (Models 5–10)**

446

447 The most obvious way to produce strain localization is to locally impose weaker  
448 rheological properties (e.g. McConnell, 1972; Mohr, 1982; Ring, 1994; Corti et al., 2007;  
449 Guillou-Frotier et al., 2012). Thus, to allow for rift formation along the western side of the  
450 craton, we perform model **7.R3.WeakZone=1** (Supplementary Fig. 1c), a variation of model **2**  
451 (Fig. 4c) where we insert of zone of weaker (WetQz) rheology between the western side of  
452 the craton and the embedding lithosphere. To test more symmetrical and geologically  
453 consistent cases we also conducted experiment **8.R3.WeakZone=2** (Supplementary Fig. 1d)  
454 with two weak interfaces along both sides of the craton.

455 The results of these models expectedly show additional zones of deformation along the  
456 western boundary of craton (Supplementary Fig. 1c,d). The only difference between these two  
457 models is the more restricted fault distribution within the “magmatic” rift branch in model **8**  
458 (Supplementary Fig. 1d), due to the predefined eastern weak zone.

459 To make the style of deformation within the eastern branch more similar to that of  
460 model **2**, we designed model **9.R3.WeakZone=3** (Supplementary Fig. 1e) with only one  
461 western weak interface (as in model **7**, Supplementary Fig. 1c) and a weaker rheology of the  
462 lower crust for the entire eastern part of the model domain (note that the lower crustal  
463 rheology of the model **2** (Fig. 4c) is that of wet quartzite (weak) everywhere). This model  
464 shows the same timing and style of deformation on the eastern “hot” side of the craton as in



465 model 2, but also shows the formation on an additional zone of deformation along the  
466 opposite craton boundary, a feature in better agreement with geological observations within  
467 the EARS (Supplementary Fig. 1e).

468 Supplementary Fig. 1f shows the results of model 10.R2.LongWeakZones  
469 characterized by a central plume position and a more complex geometry of weak zones. It  
470 aims at reproducing more accurately the embracing shape of the EARS rift branches around  
471 the Tanzanian craton.

472 Models with a central plume position in the presence of weak zones (models  
473 5.R2.WeakZone=1 and 6.R2.WeakZone=2) are shown in Supplementary Fig. 1a,b.

474 Note that model 5 that contains only one weak interface (along the western edge of the  
475 craton, Supplementary Fig. 1a) demonstrates the asymmetry not only at the crustal level as  
476 deformation expectedly develops only within this predefined weak zone, but also in the deep  
477 mantle plume where the plume ascent takes place only within its westward deflected half.  
478 This indicates that the rheological properties of the continental crust not only impact the  
479 surface morphology and crustal strain patterns, but also influence the distribution of plume  
480 head material at depth, which, in turns, bears consequences for magmatic processes and  
481 mantle lithosphere stability.

#### 482

#### 483 **4.4. Two-plumes (Models 11–17)**

#### 484

485 Recent seismic tomography data indicates the presence of a second, possibly smaller,  
486 mantle plume under the western branch of the EARS (Mulibo and Nyblade, 2013a,b). Such  
487 secondary upper mantle plumes could be stemming from the ponding of superplume material  
488 beneath the 660 km discontinuity (Yuen et al., 2007).

489 We tested this hypothesis with a series of experiments containing two mantle plumes.  
490 The first one is shifted to the NE as in most of previous models, the second one is smaller  
491 ( $r=150$  km versus  $r=200$  km for the “ordinary” first plume) and shifted to the SW.

492 The first model in this series (model **11.R3.2plume**, Fig. 5) shows that the upwelling  
493 of a bigger plume starts much faster (Fig. 5a,b), which causes the initiation of the eastern  
494 branch in the absence of visible deformation at the opposite side of the craton. After 15 Myr  
495 the head of the second plume, deflected by the craton, reaches the bottom of “normal”  
496 lithosphere (Fig. 5c) to the west where it causes strain localization less pronounced than in the  
497 east (Fig. 5d).

498 These results are weakly sensitive to variations of lower crust rheology  
499 (Supplementary Fig. 2a) and properties of the second mantle plume (size (Supplementary Fig.  
500 2b), initial temperature (Supplementary Fig. 2c) and density (Supplementary Fig. 2d)).  
501 Models **12–15** show mostly differences in timing and initiation of the western branch  
502 (Supplementary Fig. 2).

503 Model **16.R3.2plume.EqualSize** (Supplementary Fig. 3) with mantle plumes of equal  
504 size ( $r=200$  km) show simultaneous and symmetric upwelling and deflection causing pure  
505 anti-clockwise rotation of craton block in contrast of all previous “asymmetric” models where  
506 anti-clockwise rotation of the craton is combined with its westward motion. Note that the  
507 position of the craton in reference model **1**, on the contrary, remains stationary over the 80  
508 Myrs of the model (Fig. 3).

509 Model **17.R3.2plume+WeakZone=2** that combines an additional small plume shifted  
510 to the SE and weak interfaces along the craton borders, does not show much difference with  
511 the other models in this series, except for more localized strain distribution in the rifts as  
512 expected given the narrower predefined weak zones (Supplementary Fig. 4).

513

514 **4.5. Small NE shifts of the initial position of the mantle plume (Models 18–22)**

515

516 This series of experiments shows that the asymmetrical distribution of hot mantle  
517 material on the both sides of the craton causing EARS-like rifting with two coeval  
518 asymmetric branches can be reproduced not only under the assumption of two mantle plumes  
519 of different size shifted in opposite directions, but also by adjusting the initial position of a  
520 slightly scaled-up ( $r=250$  km) single plume.

521 Model **18.R3.E=50; N=100** (Fig. 6a) is characterized by an initial NE shift of the  
522 plume that is twice smaller than in previous models (112 km instead of 225 km). However, it  
523 still shows full deflection of the plume material toward one side of the craton as it is observed  
524 in above-mentioned models with initially “shifted” plumes. On the contrary, a small (only 11  
525 km) shift of the plume position to the NE (model **19.R3.E=5; N=10**) results in an almost  
526 symmetrical plume head splitting in both directions and quasi-symmetrical crustal strain  
527 distribution similar to reference model **1** (Fig. 6b). However, increasing the initial shift to ~25  
528 km (model **20.R3.E=10; N=~20–30**) leads to plume head separation into two non-equal parts,  
529 which results in a distribution of mantle plume material and crustal deformation roughly  
530 similar to models with two plumes (Fig. 6c).

531 Model **20** shows large amounts of melt produced on the rifted eastern side of the  
532 craton whereas the western border remains less deformed and relatively magma-poor. Melt is  
533 produced as a result of both adiabatic decompression as the plume rises, and of the extra heat  
534 advected by the plume itself, leading to generation of both, plume-derived and mantle-  
535 lithosphere-derived melts (Fig. 7). The mixing of plume-derived and lithospheric mantle-  
536 derived melts is consistent with geochemical data from Kenyan rift volcanics (Spath et al.,  
537 2001). This melting, in turn, increases the rate of lithospheric thinning under the eastern rift  
538 branch.

539 Model **21.R3.E=10**;  $N \sim 20-30+R=200$  illustrates the important role of the plume size  
540 in asymmetric plume head separation: a “standard-size” plume ( $r=200$  km) does not provide  
541 enough material for splitting in both directions, which leads to single-side deflection of the  
542 plume head as in the “long-shifted” models (Supplementary Fig. 5).

543 The addition of two weak zones along the craton borders to the most “successful”  
544 model of this series (model **20**) does not significantly modify the results (model **22.R3.E=10**;  
545  $N \sim 20-30+WeakZone=2$ ; Supplementary Fig. 6).

546

#### 547 **4.6. Thicknesses of the embedding lithosphere (Models 23–27)**

548

549 There is geological and geophysical evidence that the thickness of the lithosphere  
550 embedding the Tanzanian craton is larger to the west than to the east (Artemieva and Mooney,  
551 2001; Artemieva, 2006). We conducted experiments (models **23–27**) with an embedding  
552 lithosphere considerably thicker in the western half of the model domain (200 km) than in the  
553 eastern one (150 km) while keeping a 250 km-thick craton.

554 The most interesting feature of this model series is that a central initial plume position  
555 (**23.R2.H\_lit=150–200**, Fig. 8) and even a slightly (25 km and 50 km (Supplementary Fig.  
556 7a,e)) west-shifted plume (**24.R3.H\_lit=150–200+E $\sim$ 25–50**) lead to the complete deflection  
557 of the plume head to the east as in model **4**). Models with a larger initial plume show similar  
558 results (**27.R3.H\_lit=150–200+W=50+BigPlume**, Supplementary Fig. 7b,f). Only models  
559 with a westward shift of the plume position by 100 km (model **26.R3.H\_lit=150–200+E=100**)  
560 provides the deviation of the bulk of the plume material to the western side of the craton  
561 (Supplementary Fig. 7c,g). An initial shift of 75 km (the model **25.R3.H\_lit=150–200+W=75**;  
562 Supplementary Fig. 7d,h) leads to plume head separation into two non-equal parts as in model  
563 **20.R3.E=10**;  $N \sim 20-30$  (Fig. 6c).

564 The magmatic eastern branch of the EARS is associated with thinner lithosphere (150  
565 km; Artemieva and Mooney, 2001; Artemieva, 2006) while the magma-poor western rift  
566 branch develops in a much thicker one (200 km). In case of equal thickness of “normal”  
567 lithosphere this contrasted distribution of the magmatic activity can be explained by a  
568 significant eastward shift of the uprising plume with respect to the craton, which results in the  
569 eastward deflection of the plume head by the cratonic keel (e.g., models **2** and **4**).

570 However, models with different (in western and eastern segments) thicknesses of the  
571 embedding “normal” lithosphere show that only considerable westward shift of the initial  
572 plume position (about 75–100 km, i.e.  $\frac{1}{4}$  of craton width) (models **25–26**) result in large-scale  
573 magmatism to the west of the craton whereas central (models **23–24, 27**) and, obviously,  
574 eastward-shifted initial plume position result in deviation of the uprising hot material to the  
575 east.

576

#### 577 **4.7. Plume size (Models 28-31)**

578

579 We performed models with significant (225 km) NE shifts in the initial plume position  
580 to explore the impact of the plume size ( $r=150, 200$  and  $300$  km, models **28–31**). In general,  
581 these models show a similar evolution of strain within the upper crust (Supplementary Figure  
582 8). Only the combination of a small plume ( $r=150$  km) with weak (WetQz) lower crust  
583 rheology (model **28.R3.R=150+LC=WetQz**) leads to considerable differences in timing of the  
584 rifting processes (Supplementary Fig. 8a).

585 These results, however, do not permit to conclude that the plume-head size has no  
586 effective impact on system evolution since the above mentioned experiments with smaller  
587 initial shift (models **18–22**) demonstrate different modes of system development (from

588 asymmetric splitting (model **20**) to full deflection of plume head (model **21**)) resulting from to  
589 plume-head size variation.

590

#### 591 **4.8. Additional experiments (Models 32–34)**

592

593 Model **32.R3.H\_crat=200** with a thinner (200 km) craton show different strain  
594 distribution within the upper crust compared to all previous models (Supplementary Fig. 9c).  
595 Varying the boundary velocities (models **33.R3.Vext=1.5** and **34.R3.Vext=6**) only affects the  
596 timing of the main events (onset of rifting, beginning of the plume ascent, continental crust  
597 break-up, and transition to spreading) without significant impact on the other model output  
598 features (Supplementary Fig. 9a,b).

599

## 5. Discussion and conclusions

600

601

602 Our experiments show that a complex double rift system can develop from relatively  
603 simple initial conditions. In our preferred scenario the system, submitted to weak far-field  
604 tensional stress, evolves as a consequence of the deflection of a rising mantle plume by a  
605 craton keel. This preferred model produces features that bear strong similarities with first-  
606 order geological and geophysical observations in the EARS. Overall, our results reconcile the  
607 active (plume-activated) and passive (far-field tectonic stresses) rift concepts demonstrating  
608 that both magmatic and a-magmatic rifts may develop in identical geotectonic environments.

609 A feature common to all experiments is the rapid ascent of a mantle plume toward the  
610 bottom of the craton, followed by the deflection and/or splitting of the plume head, depending  
611 on the initial position of the plume. This results in the ponding and lateral spreading of the  
612 plume material at the base of the thinner lithosphere that embeds the craton, as also observed  
613 in previous 2D experiments (Burov and Guillou-Frottier, 2005; Burov et al. 2007; Burov and  
614 Cloetingh, 2010; Guillou-Frottier et al., 2012). The initial position of the strain localization  
615 zones – the future rift basins – within the upper crust is controlled by the presence of weak  
616 zones in the crust and by the distribution of plume material ponding below the lithosphere that  
617 surrounds the craton.

618 A small asymmetry in the initial position of the plume can lead to a strongly  
619 asymmetric system evolution. A rift zone forms along the eastern side of the craton with  
620 significant melt production from mantle plume material (Baker et al., 1987; Ebinger et al.,  
621 1989), analogous to the eastern magmatic branch of the EARS. To explain the formation of an  
622 asymmetric system with the coeval initiation of the amagmatic western branch and magmatic

623 eastern branch as observed in the central EARS, we experimentally explored several scenarios  
624 of which three can be retained as specifically pertaining to the EARS (Fig. 9):

625 (1) The most trivial scenario assumes mechanically weak vertical interfaces simulating  
626 the suture zone observed in the geology along the western border of the craton only (model 7).  
627 In this case the initial position and the size of the plume are relatively unimportant.

628 (2) A second scenario involves a second smaller plume initially shifted to the SW  
629 (model 11). In this case, rift basins develop on both sides of the craton with no need for  
630 weakening the interface between the craton and the embedding lithosphere.

631 (3) Finally, a broad mantle plume whose initial position is slightly shifted to the  
632 eastern side of the craton also results in contrasted double-rifting with an asymmetric  
633 distribution of mantle material on either side of the craton (model 20). This model does not  
634 require weakening the interface between the craton and the embedding lithosphere.

635 It is not possible at this point to choose a preferred scenario because adequate data in  
636 the central EARS are still quite sparse. However, it is noteworthy that only the third scenario  
637 is compatible with two important features of the geological evolution of the EARS, (1) the  
638 quasi-simultaneous initiation of both rift branches (Robert et al., 2012) and (2) their feeding  
639 from a single mantle source according to geochemistry data (Chakrabarti et al., 2009; Hilton  
640 et al., 2011; Halldórsson et al., 2014). Under this scenario, models with a thicker lithosphere  
641 to the west of the craton, as indicated by geophysical observations (Artemieva and Mooney,  
642 2001; Artemieva, 2006) that provided the best fit to observations by further increasing rift  
643 asymmetry and favoring intense magmatism along the eastern border of the Tanzania craton.

644

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950

ACCEPTED MANUSCRIPT

951 **FIGURE CAPTIONS**

952

953 **Figure 1. Geological and geophysical context.** Top: Geological map of the EARS showing  
954 the surface extent of the Tanzanian craton, surrounded on both sides by active rift branches –  
955 the magma-poor western rift characterized by low-volume volcanic activity, large ( $M > 6.5$ )  
956 magnitude earthquakes, and hypocenters at depths up to 30–40 km, while the magma-rich  
957 eastern rift is characterized by a broad zone of shallow (5–15 km) and smaller magnitude  
958 seismicity, but voluminous Cenozoic volcanism. Note the geometry (dashed line) of the  
959 craton boundary at 150 km depth (Adams et al., 2012). Bottom: East-west cross-section  
960 showing P-wave velocity mantle tomography observations (Mulibo and Nyblade, 2013a) that  
961 illustrate the thick Tanzanian craton underlain by hot mantle material deflected towards the  
962 eastern rift branch.

963

964 **Figure 2. Model setup.** Grey arrows show the velocity boundary conditions, applied in a  
965 direction perpendicular to the model domain. The initial radius of the plume is 200 km.

966

967 **Figure 3. Reference model (Model 1, Table 1).** Craton is the dark gray quasi-rectangular  
968 volume. The plume material is shown in dark red. Blue to red colors at the model surface  
969 indicate cumulative strain due to faulting.

970

971 **Figure 4. Models with different initial position of mantle plume with respect to craton**  
972 **(Models 2–4, Table 1).**

973

974 **Figure 5. Model with 2 mantle plumes (Model 11, Table 1).**

975

976 **Figure 6. Models with small NE shifts of initial position of bigger ( $R=250$  km) mantle**  
977 **plume (Models 18–20, Table 1).**

978

979 **Figure 7. Distribution of plume material and melt in the model 20 (Table 1).** EW model  
980 cross-section at 45 Ma that best-fits observations in the EARS, shows plume head separation  
981 onto two non-equal parts. The  $1300^{\circ}\text{C}$  isotherm delineates the base of the lithosphere. The  
982 plume splitting and deflection preserves the craton keel while the deflected material thermally  
983 erodes the mantle lithosphere to the east of the craton and pushes the craton to the west. The  
984 produced melt percolates within the partially molten region and accumulates below the rift  
985 axis. It combines plume-derived and mantle-lithosphere components and has a strong effect  
986 on the upwelling velocity within asthenospheric wedge below the axis of the “a stern” rift  
987 (right). Black arrow indicates initial position of craton border.

988

989 **Figure 8. Model with different thicknesses (150 km within eastern half of the model**  
990 **domain and 200 km within western one) of the embedding “normal” lithosphere (Model**  
991 **23, Table 1).** Brown surface on Fig. 8e,f,g,h corresponds to the lithospheric bottom.

992

993 **Figure 9. Three possible scenarios explaining main EARS features.** (a) the assumption of  
994 rheologically weak interface along the western border of the craton (Model 7); (b) the  
995 presence of second smaller plume initially shifted in SW direction (Model 11). (c) the unequal  
996 splitting of relatively big plume which initial position is slightly shifted to the eastern side of  
997 the craton (Model 20).



998 Table 1. Controlling parameters of the experiments.

Experiment title	Controlling parameters										
	Mantle plume(s) properties					Lithosphere properties				Boundary conditions	
	Number	Initial position	Initial size (R, km)	Density	Temperature (K)	Weak rheological interface(s) craton border(s)	Rheology of lower crust	Thickness (km)		Horizontal extension velocity (mm/year)	
Craton								“Normal” lithosphere			
1.R2	1	Centre	200	Perple_X	2000	-	WetQz	250	150	3	
2.R3.PosPL=North-East	1	NE (225 km) shift	200	Perple_X	2000	-	WetQz	250	150	3	
3.R3.PosPI=North	1	N (200 km) shift	200	Perple_X	2000	-	WetQz	250	150	3	
4.R3.PosPI=East	1	E (100 km) shift	200	Perple_X	2000	-	WetQz	250	150	3	
5.R2.WeakZone=1	1	Centre	200	Perple_X	2000	One interface along west craton border	An75	250	150	3	
6.R2.WeakZone=2	1	Centre	200	Perple_X	2000	Two interfaces along west and east craton borders	An75	250	150	3	
7.R3.WeakZone=1	1	NE (225 km) shift	200	Perple_X	2000	One interface along west craton border	An75	250	150	3	
8.R3.WeakZone=2	1	NE (225 km) shift	200	Perple_X	2000	Two interfaces along west and east craton borders	An75	250	150	3	
9.R3.WeakZone=3	1	NE (225 km) shift	200	Perple_X	2000	One interface along west craton border	west	An75	250	150	3
							east	WetQz			
10.R2.LongWeakZones	1	Centre	200	Perple_X	2000	Two curved interfaces along west and east craton borders	An75	250	150	3	

11.R3.2plume	2	NE (225 km) shift	200	Perple_X	2000	-	WetQz	250	150	3	
		SW (225 km) shift	150								
12.R3.2plume+ LowCrust=An75	2	NE (225 km) shift	200	Perple_X	2000	-	An75	250	150	3	
		SW (225 km) shift	150								
13.R3.2plume+ BigPlume	2	NE (225 km) shift	200	Perple_X	2000	-	An75	250	150	3	
		SW (225 km) shift	175								
14.R3.2plume+ HotPlume	2	NE (225 km) shift	200	Perple_X	2000	-	An75	250	150	3	
		SW (225 km) shift	150		2100						
15.R3.2plume+ LightPlume	2	NE (225 km) shift	200	Perple_X	2000	-	An75	250	150	3	
		SW (225 km) shift	150	Perple_X - 30 kg/m <sup>3</sup>							
16.R3.2plume.EqualSize	2	NE (225 km) shift	200	Perple_X	2000	-	WetQz	250	150	3	
		SW (225 km) shift									
17.R3.2plume+ WeakZone=2	2	NE (225 km) shift	200	Perple_X	2000	Two interfaces along west and east craton borders	An75	250	150	3	
		SW (225 km) shift	150								
18.R3.E=50;N=100	1	NE (112 km) shift	250	Perple_X	2000	-	WetQz	250	150	3	
19.R3.E=5;N=10	1	NE (11 km) shift	250	Perple_X	2000	-	WetQz	250	150	3	
20.R3.E=10;N=~20-30	1	NE (~25 km) shift	250	Perple_X	2000	-	WetQz	250	150	3	
21.R3.E=10;N=~20-30+ R=200	1	NE (~25 km) shift	200	Perple_X	2000	-	WetQz	250	150	3	
22.R3.E=10;N=~20-30+ WeakZone=2	1	NE (~25 km) shift	250	Perple_X	2000	Two interfaces along west and east craton borders	WetQz	250	150	3	
23.R2.H_lit=150-200	1	Centre	200	Perple_X	2000	-	WetQz	250	west	200	3
									east	150	

<b>24.</b> R3.H_lit=150–200+ W=~25-50	1	W (~25-50 km) shift	200	Perple_X	2000	-	WetQz	250	west	200	3
									east	150	
<b>25.</b> R3.H_lit=150–200+ W=75	1	W (75 km) shift	200	Perple_X	2000	-	WetQz	250	east	150	3
									west	200	
<b>26.</b> R3.H_lit=150–200+ W=100	1	W (100 km) shift	200	Perple_X	2000	-	WetQz	250	east	150	3
									west	200	
<b>27.</b> R3.H_lit=150–200+ W=50+BigPlume	1	W (50 km) shift	400- 200	Perple_X	2000	-	WetQz	250	east	150	3
									west	200	
<b>28.</b> R3.R=150+LC=WetQz	1	NE (225 km) shift	150	Perple_X	2000	-	WetQz	250	150	3	
<b>29.</b> R3.R=150+LC=An <sub>75</sub>	1	NE (225 km) shift	150	Perple_X	2000	-	An <sub>75</sub>	250	150	3	
<b>30.</b> R3.R=200+LC=An <sub>75</sub>	1	NE (225 km) shift	200	Perple_X	2000	-	An <sub>75</sub>	250	150	3	
<b>31.</b> R3.R=300+LC=An <sub>75</sub>	1	NE (225 km) shift	300	Perple_X	2000	-	An <sub>75</sub>	250	150	3	
<b>32.</b> R3.H_crat=200	1	NE (225 km) shift	200	Perple_X	2000	-	WetQz	200	150	3	
<b>33.</b> R3.Vext=1.5	1	NE (225 km) shift	200	Perple_X	2000	-	WetQz	250	150	1.5	
<b>34.</b> R3.LC=An <sub>75</sub> +Vext=6	1	NE (225 km) shift	200	Perple_X	2000	-	An <sub>75</sub>	250	150	6	



**Alexander Koptev** is a post-doctoral researcher at University of Pierre and Marie Curie (Paris VI). After his M.Sc in Geology (2008) from the Geological Department of Lomonosov Moscow State University, Russia, he started to work on geophysical processes in the Earth interior and completed his Ph.D from the same University in 2011. At present time, he is a member of ISTEP (the Institute of Earth Sciences of Paris) research group and his research is primarily focused on plume-lithosphere interactions and rift formation and, more broadly, numerical modeling of geodynamical processes.

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**Evgueni Burov** is a professor at the Institute of Earth Sciences of Paris (ISTEP) of the University of Pierre and Marie Curie. His work has a special focus on concepts in long-term lithosphere rheology, the mechanisms of intraplate deformation, such as plate bending, mountain building and basin formation. He established quantitative links between observations of flexural deformation of the continental lithosphere and its non-linear rheology and multi-layer structure. He has also studied the interplays of mantle flow and plumes and lithosphere deformation, and the links between short-term and long-term deformation in the lithosphere. His work is also devoted to implementation of new conceptual and methodological approaches for frontier thermo-mechanical numerical models capable of handling strong non-linear rheologies, surface processes, and large strain tectonic deformation. He was awarded Stephan Mueller Medal of European Geosciences Union in 2015.

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**Eric Calais** is a professor and Head of the Geosciences Department at the Ecole Normale Supérieure in Paris, France. His research focuses on the kinematics and dynamics of active tectonic processes which he studies by combining observations from space geodesy and models of lithospheric deformation. He initiated and led field experiments in the Caribbean, central Asia, and east Africa to study active deformation processes at spatial and temporal scales ranging from individual earthquakes or volcanic events to the deformation of plate margins. He is also involved in research on large earthquakes in intraplate regions such as the Central Eastern U.S. and western Europe. He was science advisor for the United Nations in 2010–2012 in the aftermath of the Haiti earthquake.

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**Laurent Guillou-Frottier** is a research engineer at the Bureau de Recherches Géologiques et Minières (BRGM, the French Geological Survey) working mainly in the fields of mineral resources and geothermics. After undergraduate diplomas in Earth Sciences and Applied Physics at University Pierre et Marie Curie and University Paris Diderot, he received his PhD and his Habilitation in Geophysics at the Institut de Physique du Globe de Paris (IPGP). His recent research interests include the role of subduction-related processes in the emplacement of porphyry-copper deposits, the role of mantle plumes in metallogenic crises, the occurrence of hydrothermal processes during weathering of ultramafic rocks, and the establishment of thermal anomalies in sedimentary basins.

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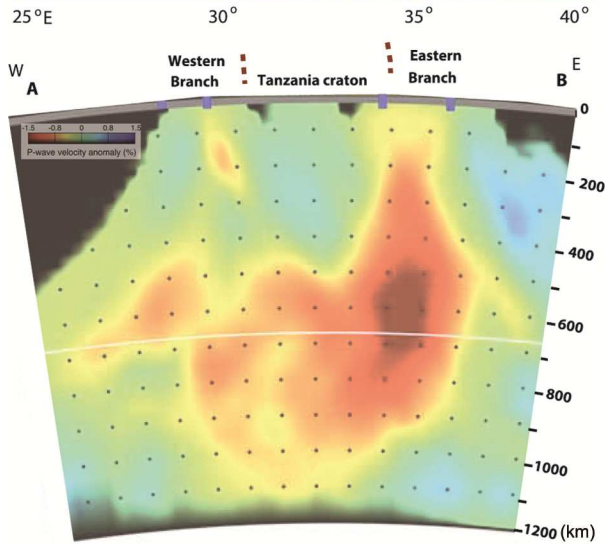
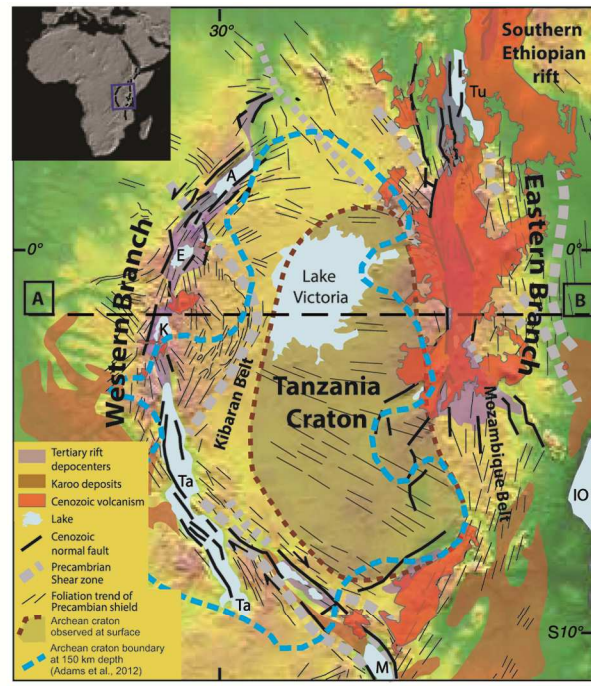
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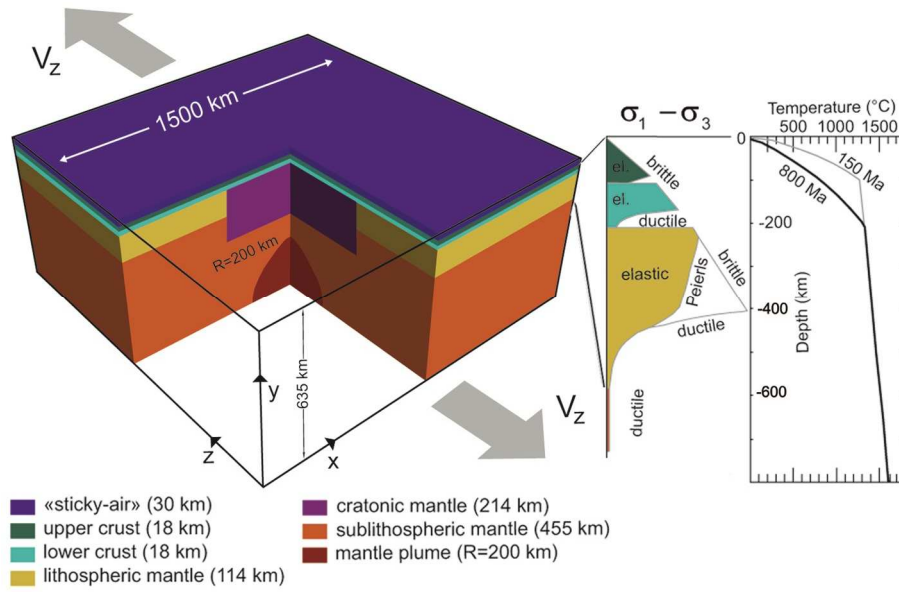
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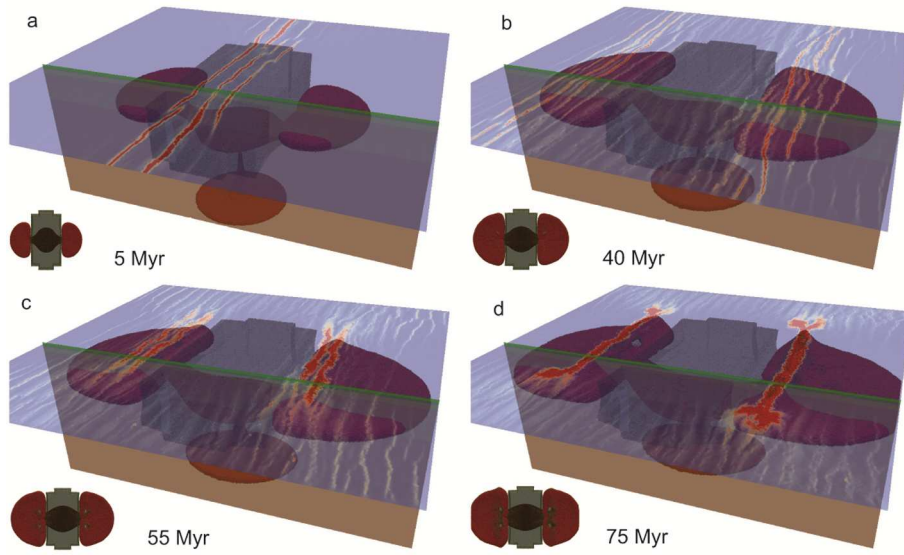


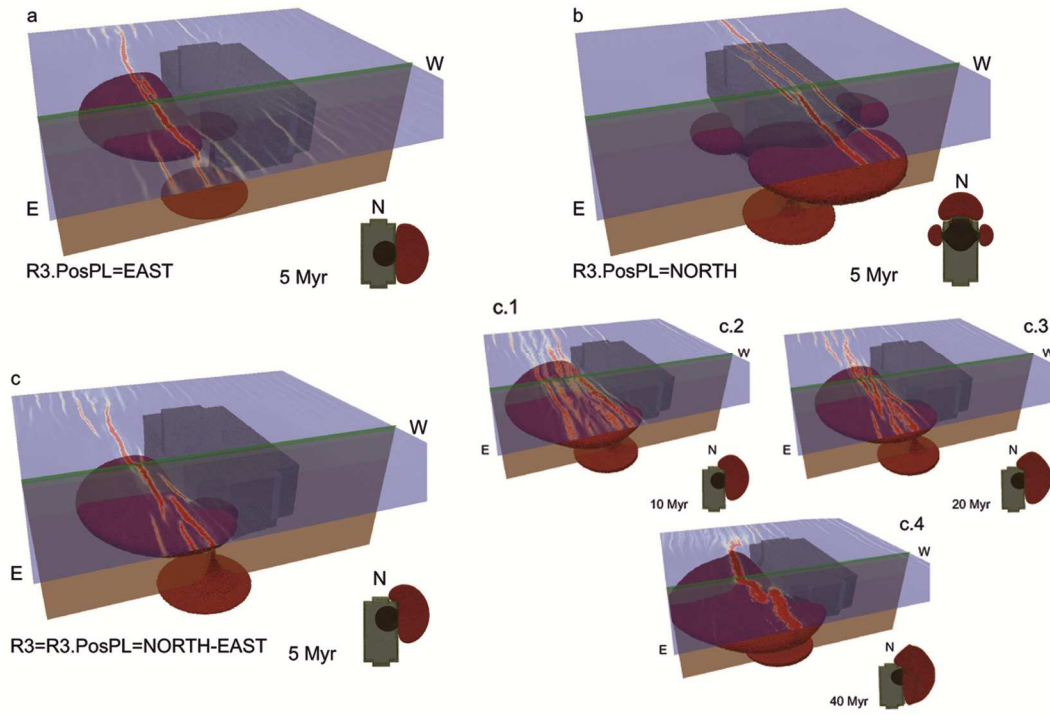
**Sierd Cloetingh** is Royal Netherlands Academy Professor of Earth Sciences at Utrecht University. He published more than 300 papers in international peer-reviewed journals and has been promotor of more than 70 PhD students of 18 different nationalities. He served the Earth Science community in various functions, including Presidency of the European Geophysical Society. He is currently the President of the International Lithosphere Programme, Editor-in-Chief of the International Journal “Global and Planetary Change” and Chairman of the Scientific Committee of the ESF Large Scale Collaborative Research Programme (EUROCORES) TOPO-EUROPE. He received honorary doctorates from five European universities and numerous honours and awards, including the Stephan Mueller Medal, Arthur Holmes Medal and honorary membership of the European Geosciences Union, Fellow and Honorary Fellow of the American Geophysical Union and the Geological Society of America, the Leopold von Buch Medal of the German Geological Society and the Alexander von Humboldt Research Award. He is member of the Royal Netherlands Academy of Arts and Sciences and Foreign member of the Royal Norwegian Academy of Sciences, the Royal Danish Academy of Sciences, the Heidelberg Academy, the Bavarian Academy and the German Academy for Technical Sciences, Acatech. He was distinguished in 2006 as Chevalier de Legion d’Honneur and in 2014 as Knight of the Royal Order of the Netherlands Lion for his contributions to science and European scientific cooperation in research and education. He was elected member of Academia Europaea in 1994 and served Academia Europaea as Chair of the Earth and Marine and Earth and Cosmic Sciences Sections and as Vice-President. In 2014 he was elected as President of Academia Europaea. He is a member of the Scientific Council of the ERC since 2009. In 2015 he was appointed as Vice-President of the ERC and coordinator of the ERC domain Physical Sciences and Engineering (PE).

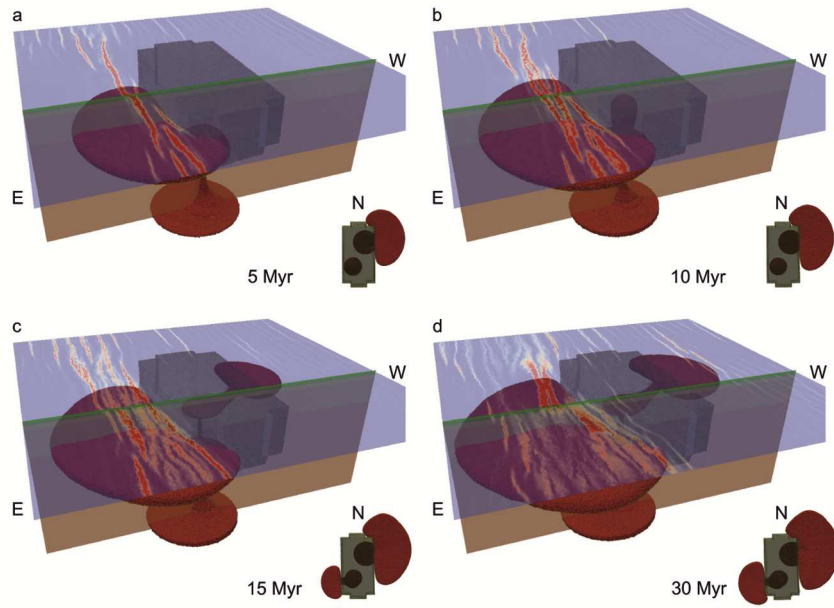


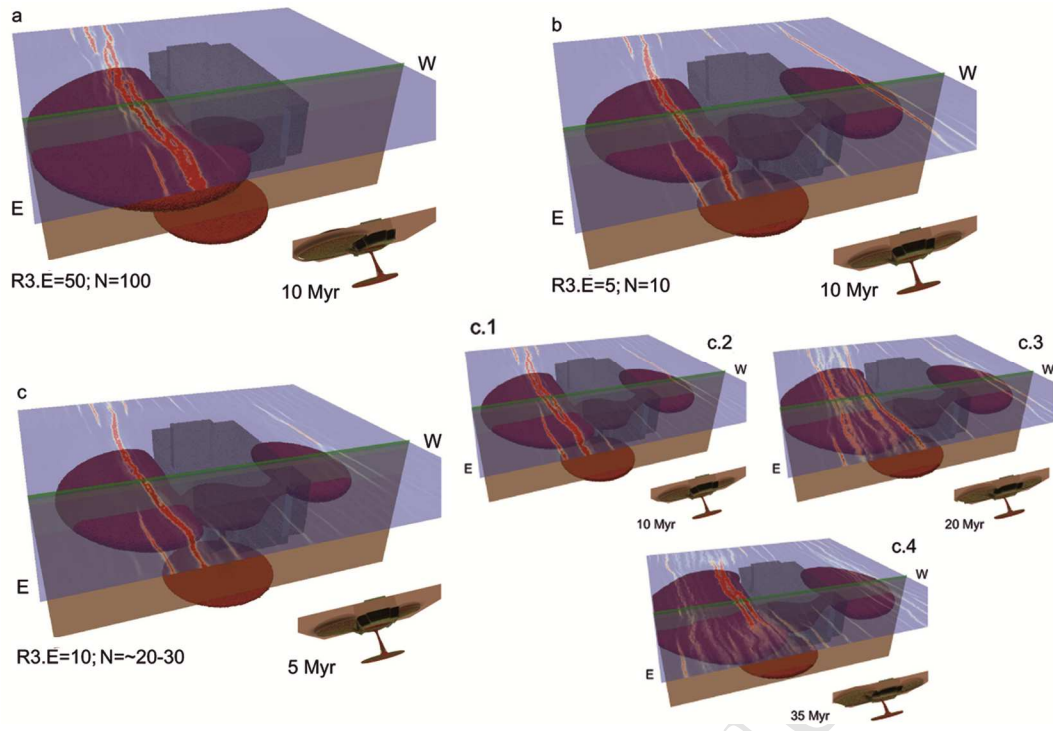


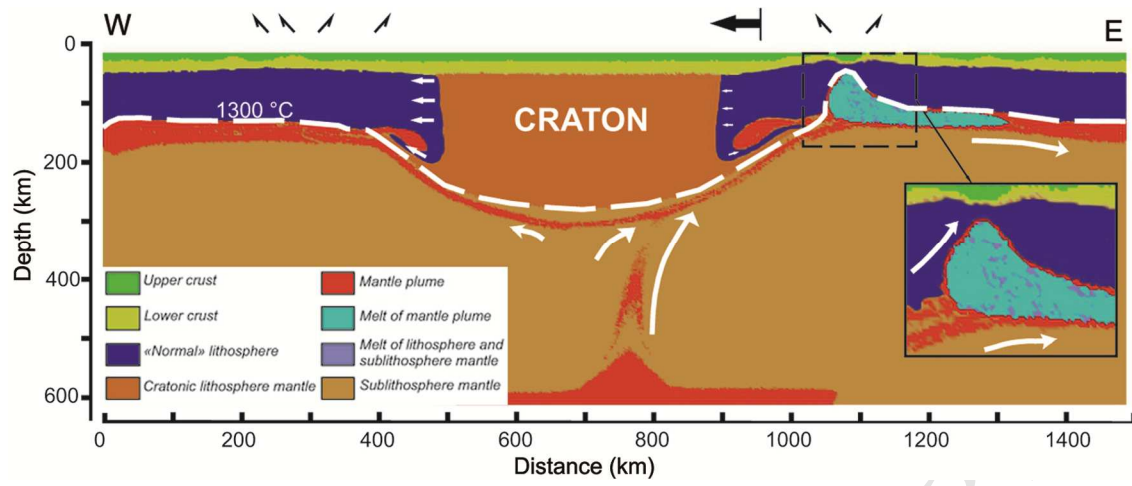


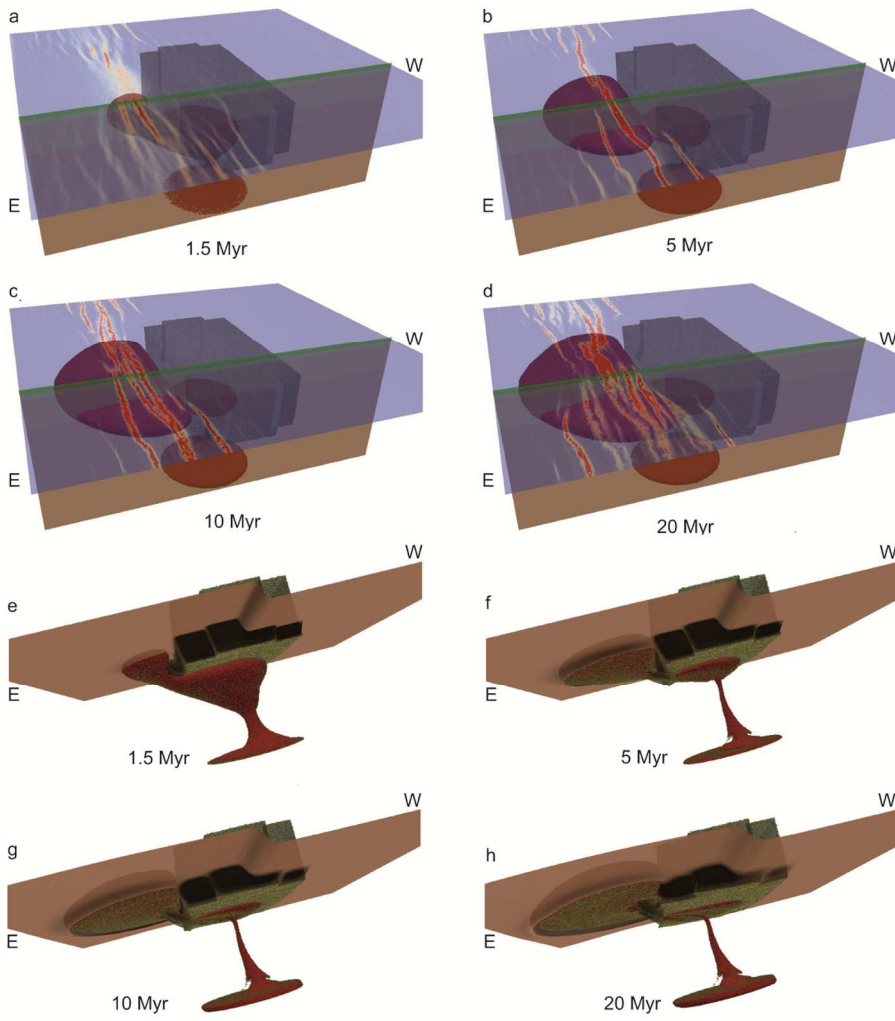


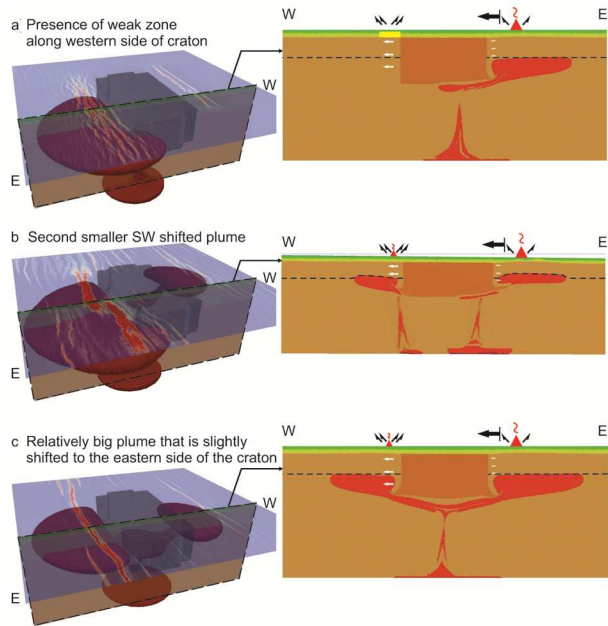














1 *Research highlights:*

- 2 1. Mantle plume is deflected by the cratonic keel and preferentially channeled along one of its
- 3 sides.
- 4 2. Simultaneous contrasted continental rifts (magma-rich and magma-poor) form due to
- 5 mantle plume interaction with a micro-craton.
- 6 3. Model reconciles the passive and active rift concept and demonstrates the possibility of the
- 7 development of both magmatic and amagmatic rifts in identical geotectonic environments.