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## Plume-induced break-up of a subducting plate: Microcontinent formation without cessation of the subduction process

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### ► To cite this version:

Alexander Koptev, Anouk Beniest, Taras Gerya, T.A. Ehlers, Laurent Jolivet, et al.. Plume-induced break-up of a subducting plate: Microcontinent formation without cessation of the subduction process. *Geophysical Research Letters*, 2019, 10.1029/2018GL081295 . hal-02325347v1

**HAL Id: hal-02325347**

**<https://hal.science/hal-02325347v1>**

Submitted on 25 Nov 2020 (v1), last revised 19 May 2021 (v2)

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1 **Plume-induced break-up of a subducting plate:**  
2 **Microcontinent formation without cessation of the subduction process**  
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11

12 **Key Points:**

- 13 • 3D thermo-mechanical modeling was performed to investigate the impact of the arrival of  
14 a mantle plume below a subducting lithospheric plate.
- 15 • The thermal and buoyancy effects of a mantle plume prove sufficient to induce the  
16 separation of a microcontinent from the subducting plate.
- 17 • The proposed scenario is consistent with a separation of the Apulian microcontinent from  
18 Africa during subduction of the Neo-Tethys.  
19

## 20 **Abstract**

21 Separation of microcontinents is explained by a ridge jump toward the passive margin as a  
22 possible consequence of plume-induced rheological weakening, ultimately leading to break-up  
23 followed by accretion of the oceanic crust along a new spreading center. In contrast to such a  
24 purely extensional case, the separation of continental micro-blocks from the main body of the  
25 African plate during its continuous northward motion and subduction under Eurasia is still  
26 poorly understood. Our numerical experiments show the thermal and buoyancy effects of mantle  
27 plume impingement on the bottom of the continental part of a subducting plate are sufficient to  
28 induce separation of an isolated micro-continental block from the main subducting continent,  
29 even during induced plate motion necessary for uninterrupted oceanic and continental  
30 subduction. Subsequent continental accretion occurs by decoupling upper-crustal nappes from  
31 the newly formed subducting microcontinent, which is in agreement with the Late Cretaceous –  
32 Eocene evolution of the eastern Mediterranean.

## 33 **Plain Language Summary**

34 Separation of micro-continental blocks from their parent continent is usually attributed to abrupt  
35 relocation of concentrated extension from the mid-oceanic ridge to the adjacent continental  
36 margin. In the context of extensional passive margin evolution, previous extensive numerical and  
37 analogue studies have revealed that hot upwelling mantle flow plays a key role in the mechanical  
38 weakening of the passive margin lithosphere needed to initiate a ridge jump. This, in turn, results  
39 in continental break-up and subsequent microcontinent isolation. However, the consequences of  
40 mantle plume impingement on the base of a moving lithospheric plate that is already involved  
41 into subduction are still unexplored quantitatively. Here we present the results of 3D thermo-  
42 mechanical models showing that even in the context of induced plate motion (contractional  
43 boundary conditions), which are necessary to sustain continuous convergence, thermal and  
44 buoyancy effects of the mantle plume emplaced at the bottom of the continental part of the  
45 subducting plate are sufficient to initiate continental break-up and the subsequent opening of a  
46 new oceanic basin that separates the micro-continental block from the main body of the  
47 continent. With these models, we show that it is physically possible to form microcontinents in a  
48 convergent setting without the cessation of subduction.

49

## 50 **1 Introduction**

51 Microcontinents are defined as independent blocks of continental lithosphere that are isolated  
52 from any other continental domains and are surrounded by ocean floor (Scrutton, 1976; Peron-  
53 Pinvidic & Manatschal, 2010). Jan Mayen in the Norwegian-Greenland Sea (Myhre et al., 1984;  
54 Gudlaugsson et al., 1988; Peron-Pinvidic et al., 2012; Blischke et al., 2016; Schiffer et al., 2018),  
55 the Seychelles in the Indian Ocean (Dickin et al., 1986; Plummer & Belle, 1995; Collier et al.,  
56 2009; Ganerød et al., 2011), and the East Tasman Plateau and the Gilbert Seamount Complex in  
57 the Tasman Sea (Exon et al., 1997; Gaina et al., 1998, 2003) are usually cited as “typical”  
58 examples of separated micro-continental blocks (Fig. 1a) as they are formed in a purely  
59 extensional tectonic setting. The best example of such microcontinents is the Jan Mayen platform  
60 separated from East Greenland in the Oligocene – Early Miocene due to a ridge jump toward the  
61 passive continental margin with the cessation of spreading at the existing mid-oceanic ridge  
62 (Lundin & Doré, 2002; Gaina et al., 2009). In this and certain other cases, rifting and

63 subsequent break-up along a passive margin is triggered by the thermal effect of an arriving  
64 mantle plume. The increase in temperature causes a decrease in the strength of the lithosphere  
65 that is large enough to re-localize the extension center from the mid-oceanic ridge toward the  
66 inner flank of the continental margin (Müller et al., 2001 and references therein). Apart from a  
67 severe weakening due to plume-induced reheating, intrinsic differences in lithospheric strength  
68 between different parts of the passive margin (Vink et al., 1984; Molnar et al., 2018) and plate  
69 reorganisation events (Whittaker et al., 2016) have also been considered to influence the  
70 segregation of continental fragments.

71 Another group of microcontinents refers to micro-blocks that are detached from the continental  
72 segment of lithosphere during continuous subduction of its oceanic counterpart under the active  
73 margin of the opposite continent (Fig. 1b). Natural examples have been found in the eastern  
74 Mediterranean, where an intricate assemblage of tectonic units implies that several continental  
75 fragments were successively separated from Africa and then accreted to the Eurasian margin  
76 (Sengör & Yilmaz, 1981; Dercourt et al., 1986; Barrier & Vrielynck, 2008; Menant et al., 2016;  
77 Barrier et al., 2018), while the Neo-Tethys was consumed by subduction (Takin, 1972; Dercourt  
78 et al., 1993; Hafkenscheid et al., 2006; Agard et al., 2011; Jolivet et al., 2016). In particular, the  
79 Apulian microcontinent detached from Africa in the Jurassic (Ricou, 1994; Frizon de Lamotte et  
80 al., 2011; Jolivet et al., 2016) and drifted away from Africa to subduct and collide with Eurasia  
81 in the Late Cretaceous – Eocene resulting in the building of the Taurides in the east and the  
82 Hellenides in the west (Dercourt et al., 1986; Jolivet et al., 2003; Van Hinsbergen et al., 2005).  
83 Later, in the Late Eocene – Early Oligocene, continental rifting started in the Red Sea and Gulf  
84 of Aden (Roger et al., 1989; Ghebreab, 1998; Watchorn et al., 1998; Leroy et al., 2012), which  
85 ultimately led to Africa-Arabia break-up in the Early Miocene (Joffe & Garfunkel, 1987;  
86 Bosworth et al., 2005; Leroy et al., 2010; Mohriak & Leroy, 2013). In contrast to the Apulian  
87 microcontinent, Arabia separated from Africa at the very last stage of the Neo-Tethys during  
88 Africa/Arabia-Eurasia collision (Jolivet & Faccenna, 2000; Bellahsen et al., 2003; Koptev et al.,  
89 2018c). The northern Appalachians (e.g. Van Staal et al. 2009, 2012) and southern Alaska (e.g.  
90 Plafker & Berg, 1994; Bruhn et al., 2004; Enkelmann et al., 2017) are other classical examples of  
91 complex mountain belts formed by accretion of tectonic terranes during the Early to Middle  
92 Paleozoic and in the Cretaceous – Cenozoic, respectively.

93 Over the last decade, several numerical studies have been performed to investigate the origin of  
94 microcontinents in extensional settings. Possible mechanisms for intra-oceanic ridge-jumps  
95 (Mittelstaedt et al., 2008, 2011; d'Acremont et al., 2010) and for mid-oceanic ridge jumps to the  
96 passive continental margins have been explored by both numerical (Beniest et al., 2017a, 2017b;  
97 Lavecchia et al., 2017) and analogue (Dubinin et al., 2018) modeling studies. However, it is still  
98 not apparent how microcontinents form in a convergent tectonic system like Africa during its  
99 motion toward Eurasia. Part of the enigmatic nature of these microcontinents is the overall  
100 contractional tectonic regime in which diverging forces act locally to separate slivers of  
101 continental lithosphere from the main body. Such extensional episodes resulting in the stripping  
102 of the microcontinents have been attributed to the push of mantle plumes periodically arriving at  
103 the base of the African lithosphere (Faccenna et al., 2013; Jolivet et al., 2016). These mantle  
104 plume events and associated large igneous provinces (LIPs) of the past 180-300 Myr (Torsvik et  
105 al., 2006; Burke et al., 2008; Jourdan et al., 2008; Ernst, 2014) are related to a long-lived thermal  
106 upwelling through the mantle beneath Africa (Forte & Mitrovica, 2001) that is often referred as  
107 “African Superplume” (Gurnis et al., 2000). It likely emanates from a large-scale, low seismic  
108 velocity province at the core-mantle boundary (Ritsema et al., 1999). The recent numerical

109 models show that the mantle superplumes might be a result of an upwelling return flow in  
110 response to deeply subducting slabs penetrating the lower mantle (Zhang et al., 2010; Heron et  
111 al., 2015; Dal Zilio, 2018; Dal Zilio et al., 2018).

112 However, the exact consequences of the arrival of the ascending mantle plume at the base of the  
113 continental part of the subducting lithospheric plate on microcontinents separation and their  
114 subsequent subduction and/or accretion at the opposite continental margin have never been  
115 estimated quantitatively and thus remain in many aspects unclear.

116 Here we present the results of a 3D high-resolution thermo-mechanical model that investigates  
117 the effect of mantle plume impingement on the base of an already moving subducting continental  
118 lithospheric plate that contains an oceanic domain continuously sinking below the opposite  
119 continent. With these experiments we test the hypothesis that the buoyancy and thermal  
120 influence of a mantle plume are sufficient to induce separation and isolation of micro-continental  
121 blocks even in the context of uninterrupted plates' convergence.

122

## 123 **2 Numerical model description**

124 We produced the simulations presented in this contribution using the viscous-plastic code  
125 "I3ELVIS" (see Supporting information, and Gerya & Yuen (2007) and Gerya (2010) for more  
126 details).

127 The model domain (Fig. 2) measures  $1000 \times 400 \times 200$  km in the x, y and z dimensions. This  
128 model box is resolved by  $501 \times 201 \times 101$  grid points thus offering a spatial resolution of ca. 2 km  
129 per grid cell in each direction.

130 The initial model setup consists of two parts: 1) an overriding continental lithosphere (400 km  
131 wide), and 2) a subducting slab that includes both oceanic and continental lithospheric segments  
132 (200 km and 400 km wide). The continental crust consists of 20 km of felsic (wet quartzite  
133 ductile flow law) upper crust and 15 km of mafic (plagioclase flow law) lower crust. The oceanic  
134 crust is represented by 8 km of basalt and gabbroic rocks (plagioclase flow law). Both the  
135 lithospheric and the sublithospheric mantle are composed of anhydrous peridotite (dry olivine  
136 flow law).

137 The velocity boundary conditions are free slip at both the front and back sides of the model box  
138 ( $y=0$  km and  $y=400$  km). The right boundary ( $x=1000$  km) uses a uniform and constant-in-time  
139 x-parallel velocity (varied from 0 to 4.5 cm/yr), which defines the material influx. The left  
140 boundary ( $x=0$  km) is fixed, meaning zero displacement in all directions. Mass conservation is  
141 ensured by the material outflux through the upper and lower permeable boundaries ( $z=0$  and  
142  $z=200$  km).

143 The laterally uniform thermal structure of the continents is piece-wise linear with  $0^\circ\text{C}$  at the  
144 surface (i.e. at the boundary between sticky air and upper crust) and  $1350^\circ\text{C}$  at 100 km ("normal"  
145 continental geotherm) or 90 km depth ("elevated" continental geotherm). The initial thermal  
146 distribution in the oceanic lithosphere is defined by a half-space cooling age at 60 Ma (Turcotte  
147 & Schubert, 2002). Both continental and oceanic lithospheres overlay a sublithospheric mantle  
148 with an initial adiabatic temperature gradient of  $\sim 0.5^\circ\text{C}/\text{km}$ .

149 We start our modeling procedure with the "no-plume" model **1** (see Table S1), including a  
150 "normal" continental geotherm ( $1350^\circ\text{C}$  isotherm at 100 km depth) in both the overriding and

151 subducting plates and a horizontal material influx at the right model side ( $V_x$ ) of 4.5 cm/yr. Next,  
152 we introduce in this setup a mantle plume anomaly below the subducting continent. This  
153 anomaly is characterized by different initial temperatures – “normal” (1900°C, model 2) and  
154 relatively “high” (2000°C, model 3). Finally, we test the impact of an “elevated” continental  
155 geotherm (1350°C isotherm at 90 km) with the experiments characterized by a “normal”  
156 (1900°C) plume temperature and different boundary velocities of material influx ( $V_x$ ) varying  
157 from 4.5 cm/yr (model 4) through 3.0 cm/yr (model 5), 1.5 cm/yr (model 6) and 0.3 cm/yr  
158 (model 7) to 0 cm/yr (model 8). Note that in these experiments (model 4-8), the “elevated”  
159 continental geotherm is only applied to the subducting plate while the continental lithosphere of  
160 the overriding plate keeps the “normal” thermal gradient.

161

### 162 **3 Results**

163 In the “no plume” experiment (model 1, Fig. 3a), subduction of the oceanic lithosphere is  
164 followed by sinking of the continental part of the subducting plate. Initiation of the continental  
165 subduction coincides in time (~5 Myr) with the onset of partial melting of the hydrated mantle  
166 material composing the initial “weak zone” seeded between the overriding and subducting plates.  
167 At a later stage of the system’s development (~10 Myr), the subducted continent reaches depths  
168 of 100 km where its upper crust is also involved in melting. Similar temporal transitions of  
169 subduction-related magmatism from mafic with mantle origin to felsic with crustal origin  
170 geochemical signature are reported in the Malayer-Boroujerd plutonic complex in western Iran  
171 (Deevsalar et al., 2018).

172 In the remaining experiments (models 2-8; Figs. 3b-d, 4-5), the mantle plume seeded by a  
173 temperature anomaly at the bottom of the model box ascends quickly (~0.1 My) to the base of the  
174 continental part of the overlying plate. The thermal impact of the upwelling hot mantle leads to  
175 partial melting of not only the plume itself but also of adjacent lithospheric and sublithospheric  
176 material. For the models with the “normal” temperature gradient (models 2-3), this mixed melt is  
177 attached to the bottom of the lithosphere and moves together with the plate into the subduction  
178 zone (Fig. 3b-c) without a strong influence on the system’s evolution compared to that of the “no  
179 plume” model 1 (Fig. 3a). Note, however, that model 3 with a relatively “hot” thermal anomaly  
180 (initial temperature of 2000°C) shows a deeper penetration of the plume into the lithospheric  
181 mantle and a more intense percolation of melts that are continuously generated from the melt-  
182 bearing material, thereby resulting in the underplating of mafic magmas at the crust-mantle  
183 boundary (Fig. 3c).

184 An increase of the continental geotherm leads to large changes in the system’s behavior (models  
185 4-8 with “elevated” thermal gradient; Figs. 3d, 4-5, S1-2). Mantle plume impingement on the  
186 bottom of the lithosphere (Fig. S1a) is followed by fast (~25 cm/yr) penetration of hot plume  
187 material through the mantle (Fig. S1b). This leads to thinning of the continental lithosphere and  
188 decompressional melting of both lithospheric and asthenospheric mantle in the area of plume  
189 emplacement and along the spontaneously localizing trench-parallel rifting zone that crosses the  
190 entire model domain (Fig. 4a; Fig. S2b). This spontaneous rifting characterized by a quick  
191 transition in time from wide to narrow mode (Fig. S2a-b) likely results from the presence of  
192 intraplate extensional stresses produced by both the plume-induced divergence and the growing  
193 pulling force of the continuously subducting oceanic slab. These extensional stresses result in a  
194 rapid and focused continental break-up in less than 0.5 Myr after the model’s onset (Fig. 4b).

195 Note that the break-up first occurs in the central part of the model domain (i.e. above the plume  
196 apex) while in the peripheral segments the transition from the pre-break-up rifting to post-break  
197 spreading is delayed by ~0.2-0.3 Myr (Figs. S1c, S2c). This break-up leads to the separation of  
198 the newly formed microcontinent from the main continental body of the subducting plate (Fig.  
199 4c).

200 The combined effect of the plume push and pulling force of a negatively buoyant slab results in a  
201 lateral displacement of the separated micro-block that is considerably faster (~10-15 cm/yr) than  
202 the imposed boundary condition velocity (~4.5 cm/yr) at the right side of the subducting plate  
203 (Fig. S3). Therefore, despite this continuous external push, a trench-parallel oceanic basin opens  
204 within a few Myr (Figs. S2d) reaching a width of ~100 kilometers wide above the plume  
205 impingement point (Fig. S1d).

206 The sensitivity analysis of the previous results with different boundary conditions (models **5-8**)  
207 shows that fast plume-induced continental break-up occurs within 0.5 Myr in all experiments,  
208 regardless the magnitude of the applied external push (Fig. 5). In contrast, the maximum width of  
209 the newly formed ocean basin varies as a function of the boundary velocities: it increases from  
210 100 km to 200 km as the external push decreases from 3.0 cm/yr to 0 cm/yr (Fig. 5).

211 As soon as the initial ocean is completely subducted, the separated micro-continental block starts  
212 to sink below the overriding continental plate (Fig. 4d). After having reached its maximum depth  
213 (~100 km), buoyant upper crust of subducting microcontinent decouples from the lower crust  
214 and mantle and migrates upwards to the surface in a return flow along the subduction channel  
215 (Fig. 4e-f). The fluid-saturated oceanic crust of the newly formed basin, which is also involved in  
216 the subduction, may play a lubricating role in this exhumation process. Note that more advanced  
217 exhumation occurs in the central segment of the model domain (Fig. 4f) possibly due to the  
218 thermal impact of the mantle plume. A comparison of the model sensitivity to different boundary  
219 velocities of material influx (models **4-7**) shows that faster subduction leads to an earlier onset of  
220 exhumation of the upper micro-continental crust. The time of exhumation initiation varies from  
221 7-9 Myr through 12 Myr to 45 Myr with a decrease of applied push ( $V_x$ ) from 4.5-3.0 cm/yr  
222 through 1.5 cm/yr to 0.3 cm/yr (see Figs. 3d, 4e, 5a-c). Note also that in the case of relatively  
223 “fast” subduction (models **4-5**;  $V_x=4.5-3.0$  cm/yr), exhumation of the micro-continental upper  
224 crust occurs simultaneously with the subduction of the main continent. As a result, exhumed  
225 micro-continental crust overthrusts the crust of the subducted main continent while newly  
226 formed oceanic crust is deeply (>100 km) subducted (Figs. 3d, 4f, 5a). This is different for the  
227 “intermediate” case (model **6**,  $V_x=1.5$  cm/yr) where the exhumed crust of the microcontinent is  
228 underlain by oceanic material of the new basin because the exhumation process was already  
229 ongoing before subduction of the main continental body (Fig. 5b). In the “slow” subduction case  
230 of model **7** ( $V_x=0.3$  cm/yr), micro-continental upper crust accretes at the boundary between the  
231 subducting and upper plates without deep subduction below the overriding continent (Fig. 5c).

232 Even in the absence of external material influx (model **8**;  $V_x=0$  cm/yr) an oceanic subduction is  
233 quickly initiated because of the gravitational slab-pull force. However, the active subduction  
234 continues only until the microcontinent reaches the overriding plate (3 Myr after onset of the  
235 experiment). After 10 Myr, the subduction process is stopped and the system can be considered  
236 as steady-state (Fig. 5d). This implies that an external push and contractional boundary  
237 conditions are necessary to sustain ongoing subduction of the oceanic lithosphere and isolated  
238 micro-block(s) as well as their moving towards the opposite continent, as is the case for the  
239 Africa-Eurasia convergence.

240

241 **4 Discussion**

242 The formation of isolated micro-continental blocks and their subsequent accretion are common  
243 in the geologic record. The separation of microcontinents in a purely extensional setting of the  
244 passive rifted margins is attributed to a ridge jump toward the continent due to rheological  
245 weakening of the passive margin lithosphere resulting from the thermal impact of the mantle  
246 plume (Fig. 1a). Repeated jumps of mid-oceanic ridges toward nearby hot spots within oceanic  
247 lithosphere have been reproduced numerically by Mittelstaedt et al. (2008, 2011) and observed  
248 by d'Acremont et al. (2010). In the context of the continental lithosphere, the 2D thermo-  
249 mechanical modeling by Lavecchia et al. (2017) showed an abandonment of the rift initially  
250 localized in a pre-existing zone of lithospheric weakness and associated rift jump toward the  
251 thermally activated mantle plume. Mantle upwelling, thus, was confirmed to be one of the key  
252 factors in the contrasting continental rifting characterized by both “passive/amagmatic” and  
253 “active/magmatic” branches (see also 3D experiments by Koptev et al., 2015, 2016, 2018a,  
254 2018b) that may further evolve into break-up centers embracing isolated continental block  
255 (Beniest et al., 2017b). In analogue models, a complete separation of the micro-continental body  
256 from a parent continent is also strongly connected to the thermal weakening due to the presence  
257 of a hot spot (Dubinin et al., 2018).

258 Continental micro-blocks separated from the main body of the African plate during its northward  
259 motion toward Eurasia accommodated by uninterrupted oceanic and continental subduction  
260 define a second type of microcontinent behavior (Fig. 2b). The northern margin of Africa is  
261 known to have experienced fragmentation since the Paleozoic when a series of small continental  
262 plates have been rifted away from Gondwana (the Avalonia and Armorica terranes in the  
263 Ordovician, the European Hunic terranes in the Silurian, and the Cimmerian blocks in the  
264 Permian) and subsequently accreted along the active southern margin of Laurussia (Matte, 2001;  
265 Stampfli & Borel, 2002; Von Raumer et al., 2003). The European Hunic terranes have been  
266 detached from the active northern Gondwana margin by oceanward migration of the arc-trench  
267 systems due to slab rollback of the Rheic ocean (Stampfli & Borel, 2002; Stampfli et al., 2002)  
268 and thus were bounded by different – active and passive – plate margins. Evolution and possible  
269 mechanisms for formation of such back-arc basins separating the volcanic arcs from the  
270 mainland have been thoroughly investigated by several numerical studies over the last decade  
271 (Clark et al., 2008; Sizova et al., 2010; Gerya & Meilick, 2011; Dal Zilio et al., 2018). In  
272 contrast, other continental domains separated from northern Gondwana in the Paleozoic (the  
273 Avalonia and Armorica terranes, and the Cimmerian blocks) had initially two passive margins  
274 (Matte, 2001; Stampfli & Borel, 2002) similar to the Mesozoic Apulian microcontinent  
275 segregated from the passive African margin (Jolivet et al., 2016). Formation of these  
276 microcontinents with two passive margins but formed on the subducting plate (Fig. 1b) is not  
277 apparent since they represent the results of extensional processes while developed in an overall  
278 context of convergence.

279 In the 3D thermo-mechanical numerical results presented here, we combine a pre-imposed  
280 convergence rate with the self-driven dynamics of a mantle plume. The push of the mantle plume  
281 impinged at the base of the African plate triggering local extension is shown to be sufficient for  
282 micro-block separation. Our results reveal the key role of the initial thermal structure of the  
283 overlying continent in this process because continental break-up and microcontinent separation



284 occurs only under condition of a relatively “elevated” geotherm (see models **4-8**; Figs. 3d, 4-5).  
285 In contrast, variations in plume temperature do not appear to be crucial (models **2-3**; Fig. 3b-c).  
286 The presence of a continuous external push (possibly resulting from a large-scale mantle flow  
287 due to the large plume postulated by Jolivet et al. (2016)) is shown to be indispensable to sustain  
288 continental subduction of the separated micro-block (compare Fig. 5a-c and Fig. 5d). In the final  
289 stages of the experiments, the subduction of the separated micro-block is followed by tectonic  
290 exhumation of the upper crust (Fig. 5a-c). This is consistent with continental accretion in the  
291 eastern Mediterranean zone during the Cenozoic where it has occurred by decoupling crustal  
292 nappes from the subducting continental lithosphere in contrast to the eastern Anatolia  
293 characterized by “true” collisional events (Menant et al., 2018).

294 The thermo-mechanical models presented here consider the mantle plume as a necessary  
295 ingredient for the generation of locally-extensional tectonics in a regionally-compressive  
296 environment that ultimately results in microcontinent formation. Note though, that we do not rule  
297 out other factors possibly contributing to the separation of a micro-block from the subducting  
298 plate such as (1) regional-scale plate reorganization associated to sharp convergence acceleration  
299 (Agard et al., 2006, 2007) or (2) lateral transition from subduction to collision along the  
300 convergent boundary (Decourt et al., 1993; Bellahsen et al., 2003; Koptev et al., 2018c) in the  
301 presence of (3) pre-existing zones of weakness (Dixon et al., 1987; Ligi et al., 2012).

302

## 303 **5 Conclusions**

304 The modeling results presented here reveal that even in a convergent tectonic setting, an  
305 upwelling flow of hot mantle material can trigger localized extension, rifting and breakup. This  
306 ultimately leads to the opening of a new oceanic basin that separates a micro-continental sliver  
307 from the main continent belonging to a subducting lithospheric plate. The scenario evaluated  
308 involves plume-induced separation of a microcontinent away from the main continent during  
309 uninterrupted subduction. This can be compared to the Paleozoic – Cenozoic evolution of Africa,  
310 which is characterized by subsequent separation of several micro-continental pieces while  
311 subduction of the oceanic and continental lithosphere below the Eurasian margin was ongoing.

312

## 313 **Acknowledgments**

314 We thank two reviewers for their helpful and constructive comments. This study is co-funded by  
315 the Advanced ERC Grant 290864 RHEOLITH (L. Jolivet – A. Koptev) and ERC Consolidator  
316 Grant 615703 EXTREME (T. Ehlers – A. Koptev). The numerical simulations were performed  
317 on the ERC-funded SGI Ulysse cluster of ISTeP. The computer code I3ELVIS used to generate  
318 our 3D thermo-mechanical numerical model is provided in Gerya (2010). Open source software  
319 ParaView (<http://www.paraview.org>) was used for 3D visualization. The figures in the  
320 supporting information contain the numerical simulation data.

321

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620

621 **Figure captions**

622 **Figure 1.** Plate reconstructions and schematic scenarios illustrating two principal types of the  
 623 microcontinents: a) microcontinent formed in a purely extensional tectonic setting: the North  
 624 Atlantic region, separation of the Jan Mayen microcontinent from East Greenland (reconstruction  
 625 after Torsvik et al., 2015); b) microcontinent formed in the context of continuous convergence of  
 626 the plates: Africa and the Neo-Tethys ocean, the Apulian microcontinent detached from Africa  
 627 and then accreted to the Eurasia during ongoing subduction of the Neo-Tethys lithosphere. The  
 628 layer colors on the schematic profiles correspond to the rock types shown on Fig. 2.

629 **Figure 2.** Initial model setup that consists of an overriding continental lithosphere and  
 630 subducting lithospheric plate including both oceanic and continental lithosphere. A mantle plume  
 631 (40-km radius thermal anomaly of 1900-2000°C) is seeded underneath the continental part of  
 632 subducting plate. A weak zone characterized by low plastic strength and wet olivine rheology  
 633 has been implemented in order to provide a rheological decoupling between the upper and  
 634 subducting plate, enabling subduction of the incoming oceanic lithosphere beneath the overriding  
 635 continent (e.g. Burg & Gerya, 2005; Vogt et al., 2017, 2018; Willingshofer et al., 2013). The top  
 636 surface of the lithosphere is defined as an internal free surface through a 12 km-thick layer of  
 637 “sticky air” (Gerya, 2010; Duretz et al., 2011; Cramer et al., 2012). The code “I3ELVIS”  
 638 accounts for mineralogical phase changes by thermodynamic solution for density, obtained from  
 639 the optimization of Gibbs free energy for a typical mineralogical composition of the  
 640 lithospheric/sublithospheric mantle and plume material (Connolly, 2005). Partial melting is  
 641 implemented using the most common parameterization (Katz et al., 2003; Gerya, 2013) of  
 642 hydrous mantle melting processes. The color legend for different types of the rocks includes the  
 643 following: 0 – sticky air, 1 – sediments, 2 – upper continental crust, 3 – lower continental crust, 4  
 644 – oceanic crust/underplated mafic material, 5 – lithospheric mantle, 6 – sublithospheric mantle, 7  
 645 – hydrated continental crust, 8 – hydrated mantle/mantle plume, 9 – melt-bearing continental  
 646 crust, 10 – melt-bearing gabbro, 11 – melt-bearing lithospheric mantle, 12 – melt-bearing  
 647 sublithospheric mantle, 13 – melt-bearing hydrated mantle/mantle plume, 14 – quenched mantle.  
 648 Hydrated, partially molten and quenched rocks not shown on Fig. 2 will appear over the  
 649 evolution of the experiments (see Figs. 3-5).

650 **Figure 3.** Central vertical cross-sections showing the evolution over time for the following  
 651 experiments: a) model 1, no plume; b) model 2, mantle plume of “normal” temperature  
 652 (1900°C); c) model 3, “hot” mantle plume (2000°C); d) model 4, “elevated” continental  
 653 geotherm (1350°C isotherm at 90 km) above “normal” (1900°C) mantle plume. All these  
 654 experiments are characterized by the same velocity of horizontal influx:  $V_x=4.5$  cm/yr. Colors of  
 655 rock types are as on Fig. 2.

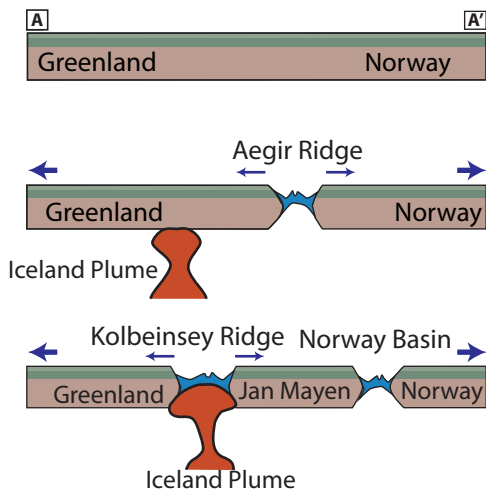
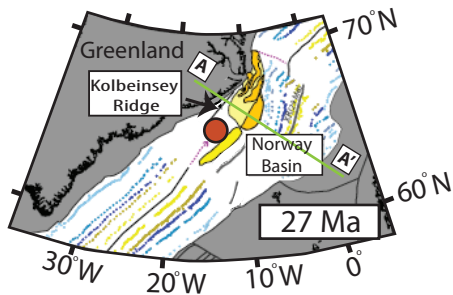
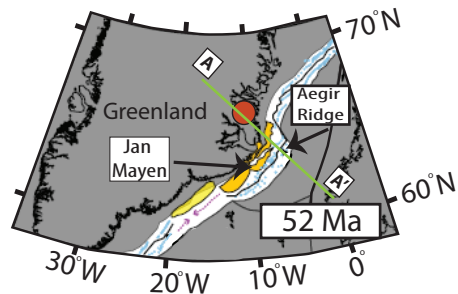
656 **Figure 4.** 3D view of the temporal evolution for the model 4 (“elevated” continental geotherm  
 657 above “normal” mantle plume,  $V_x=4.5$  cm/yr; see Table S1). Colors of rock types are as on Fig.  
 658 2.

659 **Figure 5.** Central vertical cross-sections showing the evolution for the experiments characterized  
 660 by normal temperature of the mantle plume (1900°C), “elevated” continental geotherm (1350°C  
 661 isotherm at 90 km) in the overriding plate and varied velocities of horizontal influx at the right  
 662 model boundary: a) model 5,  $V_x=3.0$  cm/yr; b) model 6,  $V_x=1.5$  cm/yr; c) model 7,  $V_x=0.3$   
 663 cm/yr, d) model 8,  $V_x=0.0$  cm/yr. Note a gradual increase of the maximum width of newly

664 formed oceanic basin from 100 km through 125-175 km to 200 km with decreasing  $V_x$  from 3.0  
665 cm/yr through 1.5-0.3 cm/yr to 0 cm/yr. Colors of rock types are as on Fig. 2.

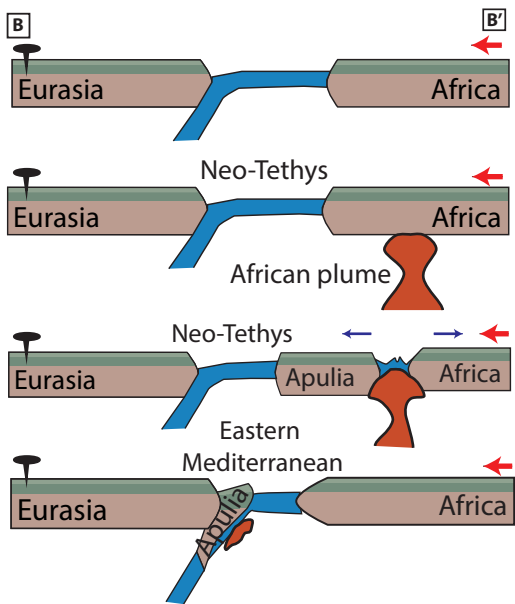
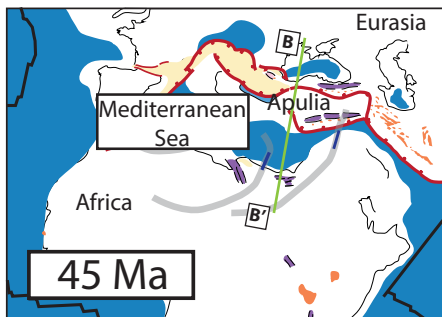
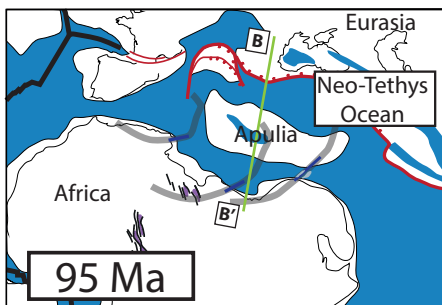
Figure 1.

## a. Divergent tectonic setting



- major continental entity □ oceanic crust
- main tectonic blocks and southward of extension of Jan Mayen microcontinent
- magnetic anomalies ● Iceland plume

## b. Convergent tectonic setting



- major continental entity □ oceanic crust
- active compression
- active rifting
- volcanism
- mid-oceanic ridges
- subduction zone
- motion paths of Africa
- normal faults

Figure 2.

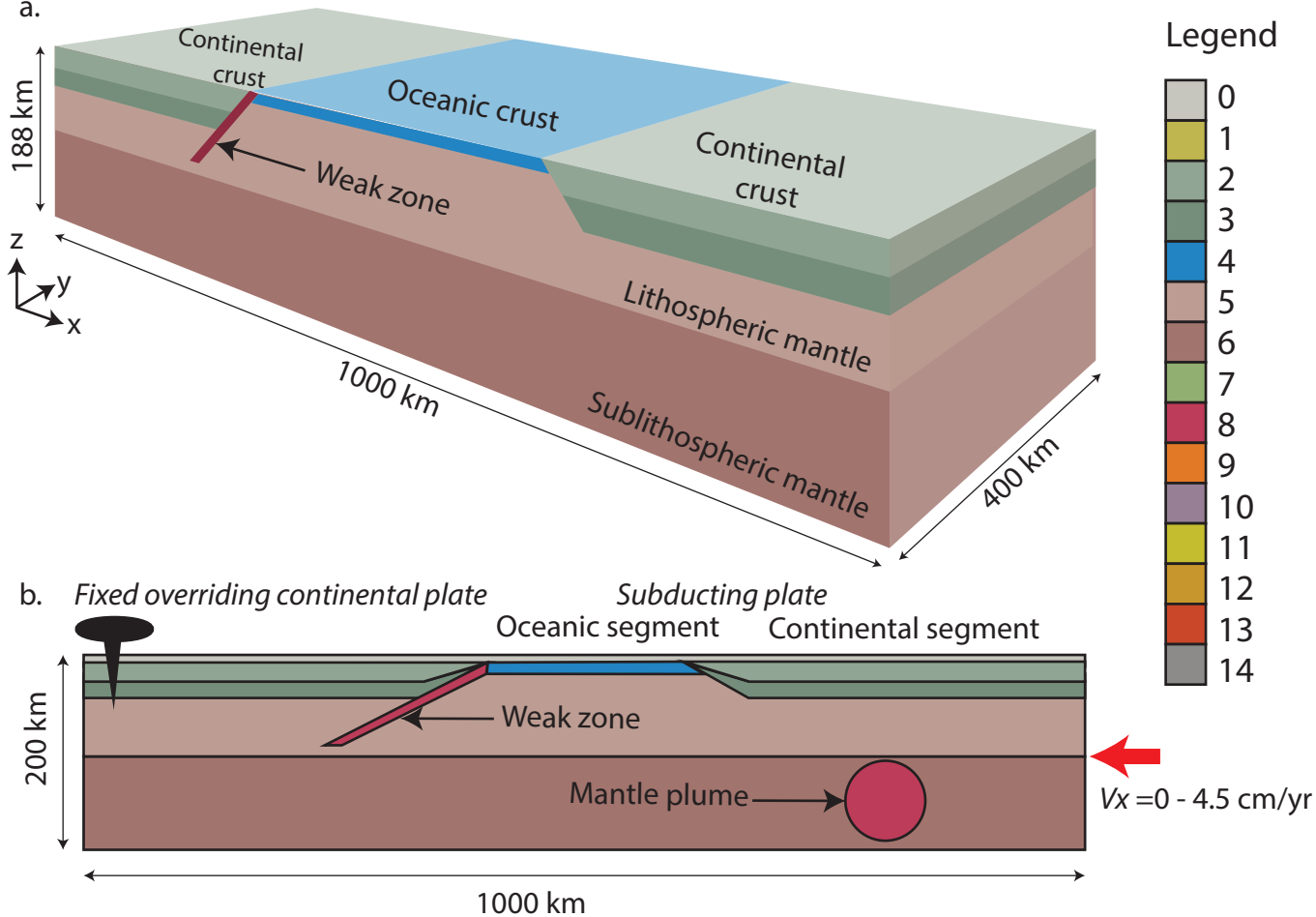
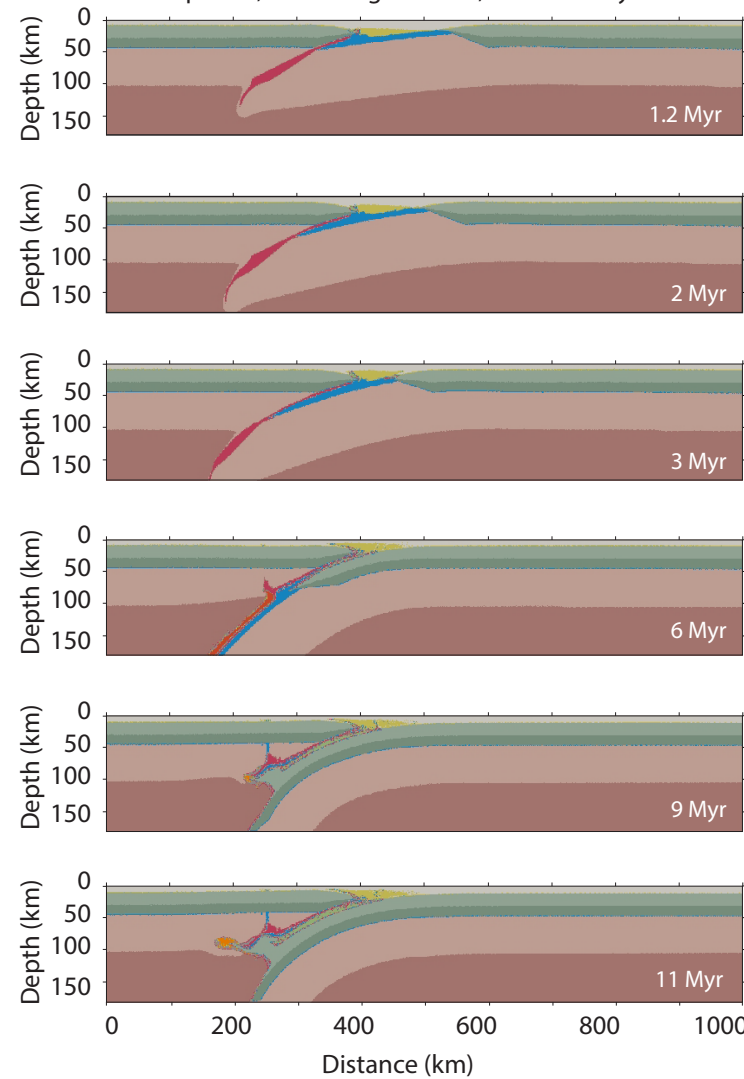


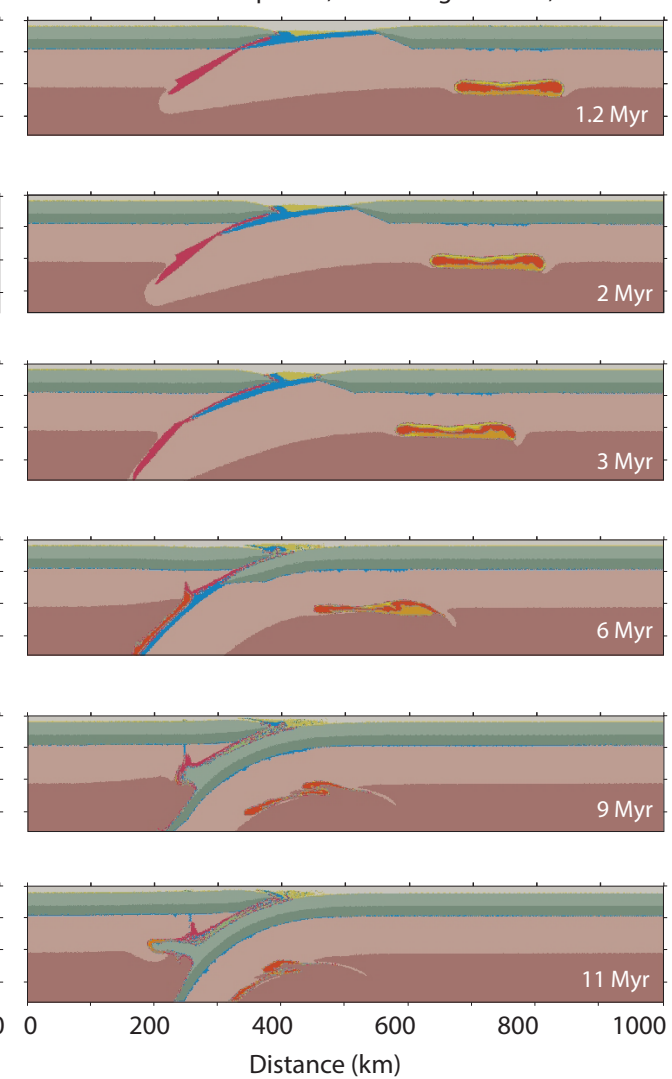
Figure 3.



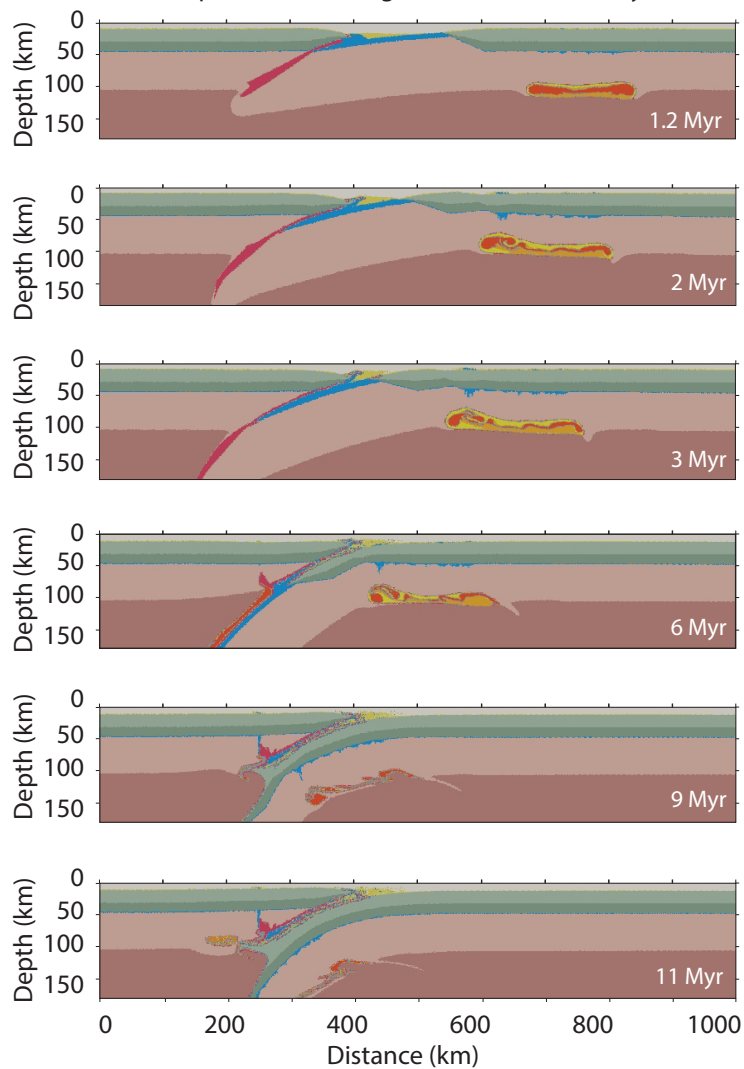
a. **Model 1:** No plume, "normal" geotherm,  $V_x = 4.5$  cm/yr



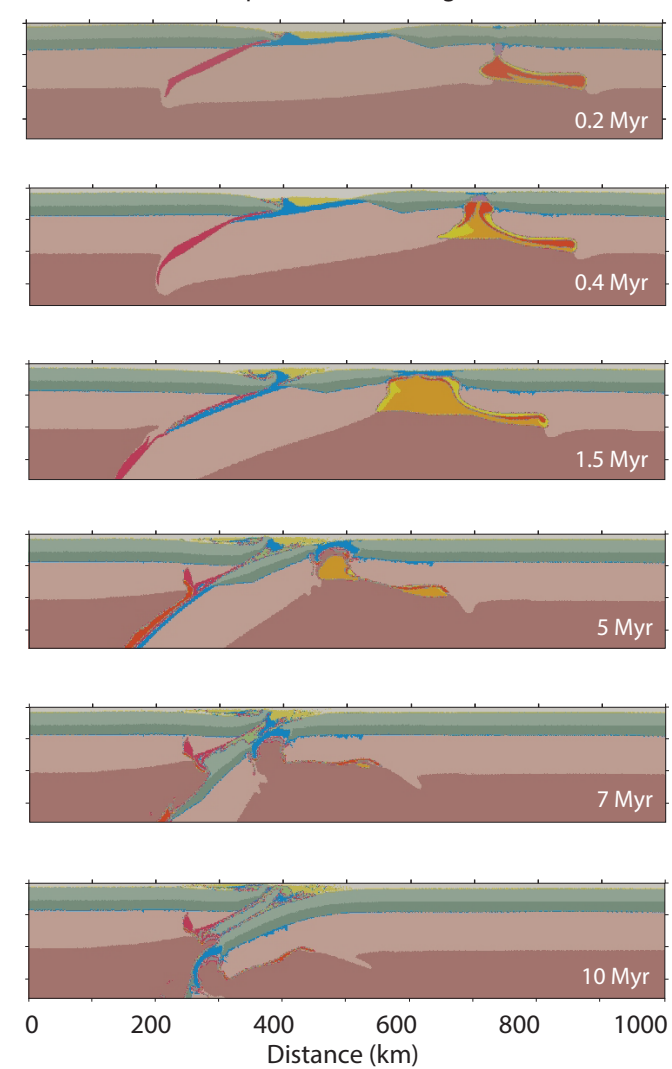
b. **Model 2:** "Normal" plume, "normal" geotherm,  $V_x = 4.5$  cm/yr



c. **Model 3:** "Hot" plume, "normal" geotherm,  $V_x = 4.5$  cm/yr

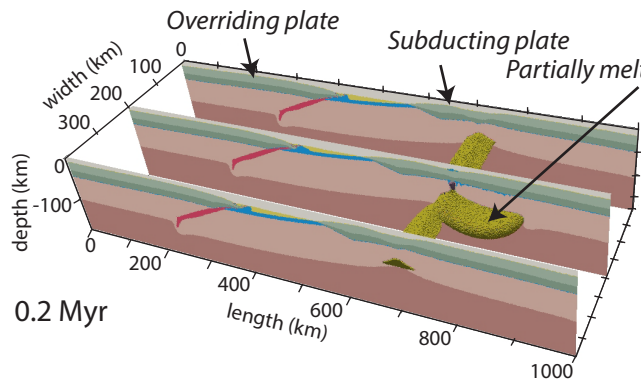


d. **Model 4:** "Normal" plume, "elevated" geotherm,  $V_x = 4.5$  cm/yr

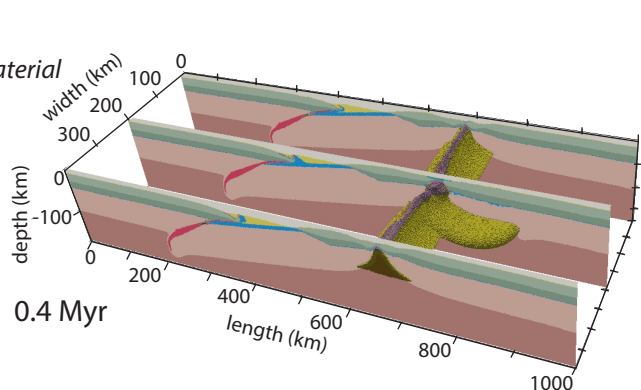


**Figure 4.**

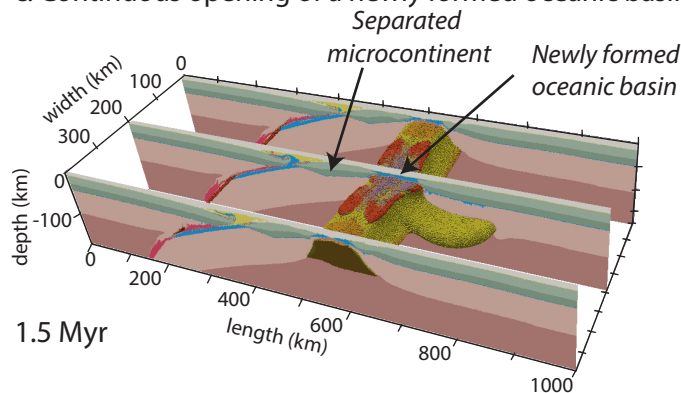
a. Mantle plume uplift leading to decompressional melting of lithospheric and sublithospheric mantle



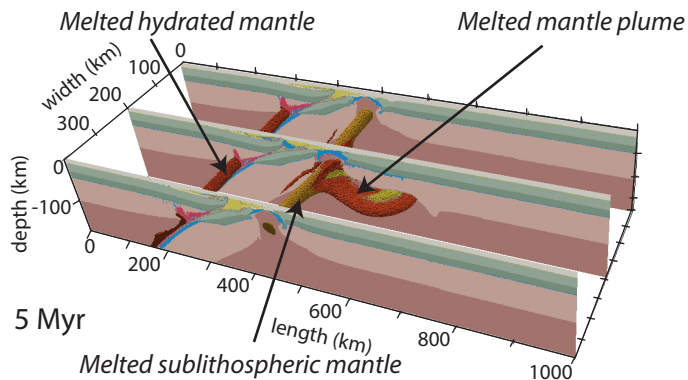
b. Continental break-up and separation of a microcontinent from the main subducting continent



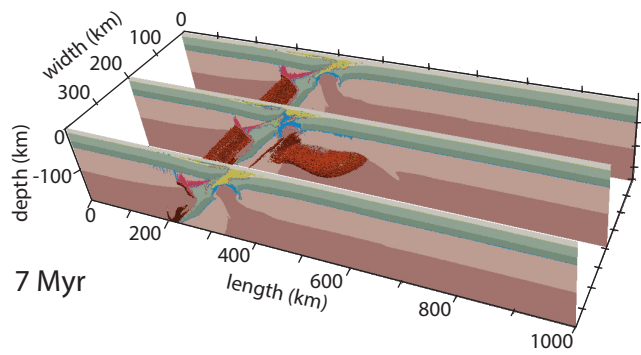
c. Continuous opening of a newly formed oceanic basin



d. Subduction of separated microcontinent



e. Subduction of newly formed basin; start of exhumation of the microcontinental upper crust



f. Exhumation and overthrusting of microcontinental crust onto the subducted crust of the main continent

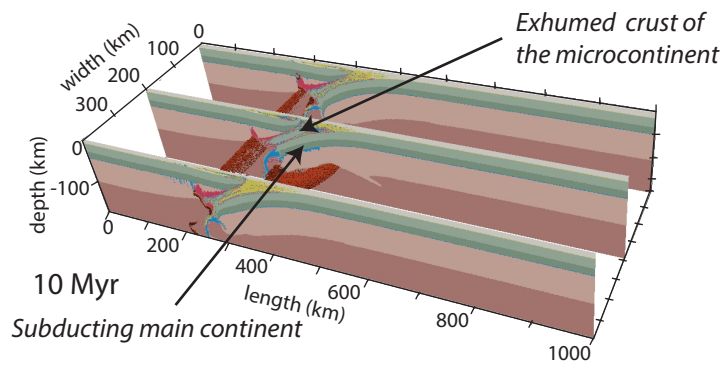
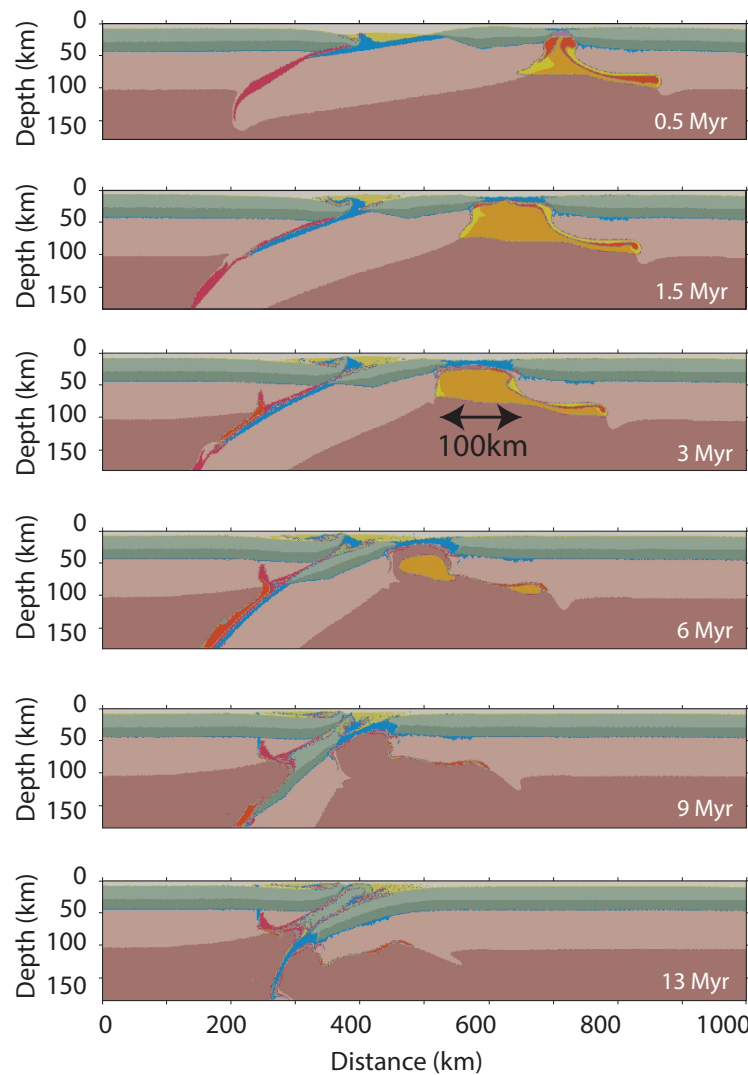
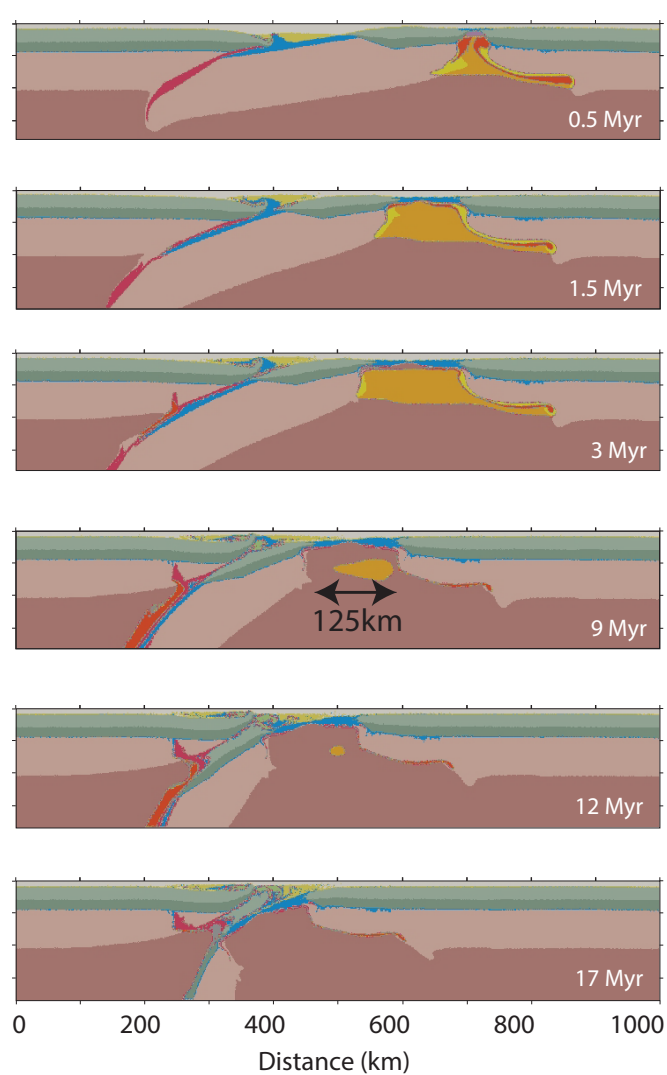
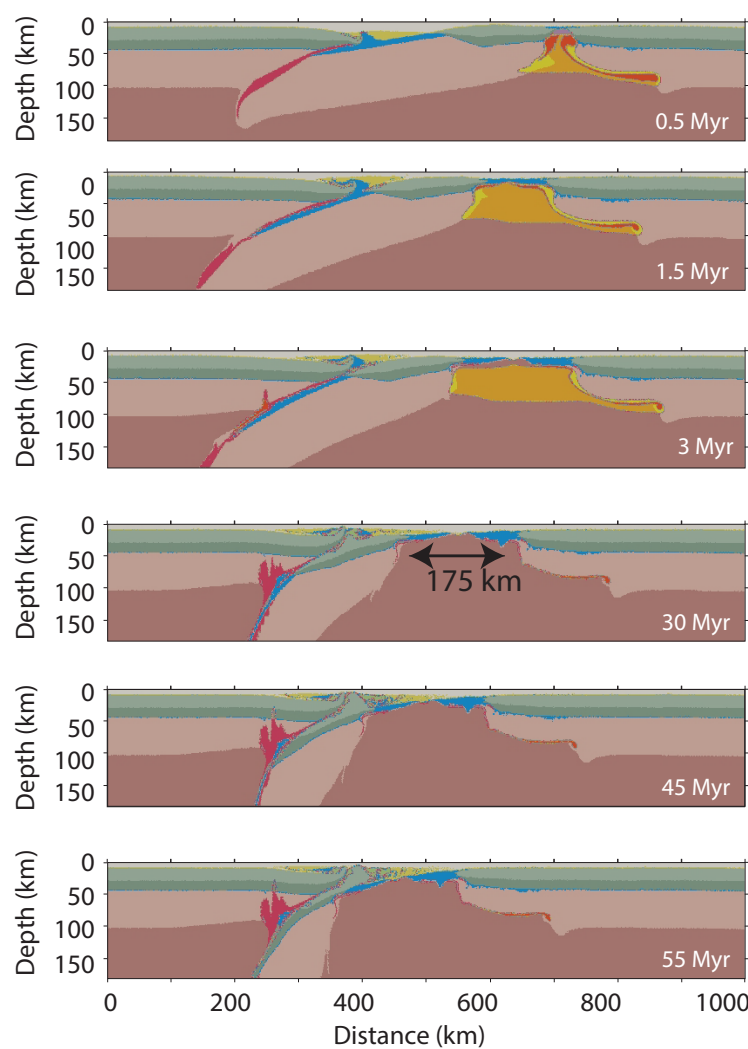


Figure 5.

**a. Model 5** ("Fast" subduction):"Normal" plume, "elevated" geotherm,  $V_x = 3.0$  cm/yr**b. Model 6** ("Intermediate" subduction):"Normal" plume, "elevated" geotherm,  $V_x = 1.5$  cm/yr**c. Model 7** ("Slow" subduction):"Normal" plume, "elevated" geotherm,  $V_x = 0.3$  cm/yr**d. Model 8** ("No push"):"Normal" plume, "elevated" geotherm,  $V_x = 0.0$  cm/yr