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► **To cite this version:**

Laetitia Le Pourhiet, Nicolas Chamot-Rooke, Matthias Delescluse, Dave A. May, Louise Watremez, et al.. Continental break-up of the South China Sea stalled by far-field compression. *Nature Geoscience*, 2018, 11, pp.605-609. 10.1038/s41561-018-0178-5 . hal-01828278

**HAL Id: hal-01828278**

**<https://hal.science/hal-01828278>**

Submitted on 8 Jan 2024

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1 **Continental break-up of the South China Sea stalled by far field compression**

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15

16 **Outcome of decades of two-dimensional modeling of lithosphere deformation**  
17 **under extension is that mechanical coupling between continental crust and the**  
18 **underlying mantle controls how a continent breaks apart to form a new ocean.**  
19 **However, geological observations unequivocally show that continental break-up**  
20 **propagates in the third dimension at rates that do not scale with the rate of**  
21 **opening. Here, we perform three-dimensional numerical simulations and**  
22 **compare them with observations from the South China Sea to show that tectonic**  
23 **loading in the direction of propagation exerts a first order control on these**  
24 **propagation rates. The simulations show that in the absence of compression in**  
25 **that direction, continental break-up propagates fast, forming narrow**

26 **continental margins independently of the coupling. When compression is**  
27 **applied, propagation stagnates, forming V-shaped oceanic basins and wide**  
28 **margins. Changes in out-of-plane loading therefore explain the alternation of**  
29 **fast propagation and relative stagnation. These new dynamic constraints suggest**  
30 **that the West-to-East topographic gradient across the Indochinese Peninsula**  
31 **prevented continental break-up propagation through the 1000 km wide**  
32 **continental rift of central and west basin of the South China Sea, until the**  
33 **direction of stretching changed 23 million years ago resulting in bypassing and**  
34 **acceleration of continental break-up propagation.**

35  
36 The relative movement of rigid plates on the sphere is described by rotation around a pole.  
37 Assuming that continental break-up occurs after some constant amount of continental extension,  
38 the propagation rate of break-up should scale with the gradient of stretching rate along strike. Far  
39 from the pole of rotation, the velocity of opening is nearly constant over hundreds of kilometres  
40 and the model predicts that break-up should occur almost instantaneously<sup>1</sup> (Figure 1a). Within  
41 this theory, V-shaped oceanic basins can form only in the close vicinity of the pole of rotation,  
42 where variations in the stretching rate are important (Figure 1b). This scissors opening  
43 kinematic<sup>2</sup> model has been reproduced by dynamic experiments<sup>3</sup> and numerical simulations<sup>1</sup>. A  
44 consequence of this model is that active shortening and thickening of the crust must be observed  
45 in the close vicinity of the propagator (Figure 1b). Yet, several well-documented V-shaped  
46 propagators, like the East Basin of the South China Sea<sup>4</sup>, or the Coral Sea<sup>5,6</sup>, are forming far  
47 from the pole of rotation within zones of diffuse continental rifting<sup>7</sup> (Figure 1c).

48 It is widely accepted that the rheological layering of the lithosphere controls the width and  
49 structure of passive margins<sup>8,9,10</sup>. Dynamic models have pointed out that mechanical  
50 heterogeneities delay the localisation of deformation at the onset of continental rifting<sup>11,12</sup>.

51 Extrapolating these models to continental break-up stage, it is generally hypothesized that Vink  
52 kinematic model<sup>7</sup> (Figure 1c) arises from variation in lithosphere strength. Yet, to date, V-shaped  
53 oceanic basins have only emerged from dynamic simulations and experiments using scissor  
54 opening, and or, simulations with two offset oceanic basins propagating in opposite direction<sup>13,14</sup>.  
55 An alternative explanation is therefore needed to explain the stagnation of uni-directional break-  
56 up far from the rotation pole.

### 57 **3D tectonic loading impacts continental break-up**

58 Assuming that break-up propagation is similar to the opening of cracks in an elastic media,  
59 local variations in the rate of propagation are expected with variation in loading<sup>15,16</sup>. Here, we  
60 study how a small amount of compression, and/or extension, acting normal to the direction of  
61 propagation influences the rate of break-up propagation at lithospheric scale, discarding the local  
62 effect of magma supply<sup>17</sup> (see methods). In order to study continental break-up propagation, a  
63 lithospheric scale weak zone is place on one side of the models to trigger rapid continental  
64 break-up after 10 to 15 Myr of stretching at 20 mm/yr. Three 3D simulations are presented in  
65 Figure 2: a reference cylindrical experiment, and two experiments in which a small amount of  
66 shortening or stretching is applied in the third dimension.

67 All these models include a mechanically weak lower crust, which is prone to slow  
68 localisation of deformation and favour the development of wide rift structure<sup>8,10,18</sup>. Further  
69 details are provided in the supplementary method section. Time evolution of the models in 3D  
70 and cross-section are provided in supplementary movies 1, 2 and 3.

71 After 23 Myr of evolution, the break-up has propagated in the pristine part of the three models.  
72 Finite plastic strain outlines the orientation of major faults bounding the passive margin. The  
73 passive margins are narrow and symmetric and very little differences in width are observed along  
74 strike in the models with zero out-of-plane loading (Figure 2b), or with 2.5 mm/yr out-of-plane

75 extension (Figure 2c). In contrast, the model with 2.5 mm/yr out-of-plane compression (Figure  
76 2a) shows the formation of a wide continental rift at the front of a V-shaped propagator. The  
77 width of the rift increases with distance to the propagator and is bounded by minor strike slip  
78 structures that make a 45° angle with the direction of extension.

79 Surface heat flow at the seafloor serves as a proxy for oceanic domain age<sup>19</sup> and shows that  
80 with fast break-up propagation, the angle between the seafloor age and the margin is only a few  
81 degrees. On the contrary, owing to the slow propagation of break-up, the model with out-of-  
82 plane compression displays a typical V-shaped oceanic propagator with angles of more than 30°.

83 The location of the tip of the oceanic domain with time has been tracked for the three  
84 models (Figure 2d). The models with no tectonic forcing or with out-of-plane extension display a  
85 clear linear trend with propagation rates of approximately 150 mm/yr. The rate of propagation in  
86 the experiment in compression is not constant and the location of the tip of the oceanic domain  
87 with time is best approximated by a square root function, meaning that the propagation scales  
88 with  $t^{-1/2}$ , and rate decreases with  $t^{-3/2}$  and tends towards zero. The initial break-up is also found  
89 to occur slightly later in the model in compression than in the other two models.

90 The width of passive margin and continental rift are often interpreted as a proxy for the  
91 viscosity of the lower crust<sup>8</sup>. All the models presented here have a weak lower crust and  
92 therefore should result in the formation of wide continental rifts, according to the current  
93 paradigm in the field<sup>9,10,18,20</sup>. Yet, the width of continental rift produced varies from 360 km to  
94 1000 km (see cross-sections in Figure 2).

95 In order to evaluate the relative effects of lower crustal rheology and boundary conditions  
96 on propagation, we provide the results of two supplementary simulations with a ten times larger  
97 lower crustal viscosity. Comparing the amount of crustal thinning (Figure 3) indicates that the  
98 rheology of the lower crust indeed influences both the rate of propagation of continental break-

99 up and the width of passive margin when compression is applied. Yet, in extension, rheological  
100 effects are negligible as propagation occurs 10 times faster (150 mm/yr) than the far field plate  
101 velocity (15 mm/yr).

102 Propagation of continental break-up modifies the dynamics of continental rifting because  
103 the horizontal strain rate increases at the tip of the propagator together with the viscous stress in  
104 the lower crust. This favours localised brittle strain at depth and rapid necking. For the fast  
105 propagation cases, this leads to the formation of narrow continental margins in a lithosphere with  
106 weak lower crust. For the slow propagation case, the width of the continental rift depends on the  
107 strength of the lithosphere.

#### 108 **The South China Sea propagator stalled due to Indochina**

109 The South China Sea (SCS) is the largest area of immersed continental crust in the world.  
110 The SCS has all the characteristics of a wide continental rift, but crust is thinner and extension  
111 has led to continental break-up<sup>21</sup>. In map view, the width of the passive margin increases from  
112 400 km up to 900 km towards the west<sup>22</sup> as the age of sea floor decreases, an observation that is  
113 clearly incompatible with typical scissor opening kinematics (Figure 1b). In the following,  
114 geological ages are given in million annum (Ma) to contrast with time that remains in million  
115 years (Myr).

116 From 32 Ma (magnetic anomaly 11) to 23 Ma (7-6b) (Stage 1, Figure 4b), the East SCS sub-  
117 basin is the only one that accommodates regional extension by spreading. The North-South  
118 opening direction at the time is best described by a pole of rotation located E65/N09, far enough  
119 from the SCS to dismiss again the scissor-type model. From anomaly 6a the direction of opening  
120 changes by 15°<sup>23</sup> as outlined by the change of orientation in magnetic anomaly in the yellow  
121 oceanic domain (Figure 4a). The break-up propagates rapidly (Stage 2, Figure 4b) through the  
122 central sub-basin, before stalling for 2 Myr until 20 Ma (6/5e) when the last increment of

123 propagation (Stage 3, Figure 4) took place. The last recorded magnetic anomaly (5c) marks the  
124 end of spreading in the SCS<sup>24,25</sup> at 16 Ma.

125         Discussing the controversial origin, and end, of opening is beyond the scope of this study,  
126 since we focus here on the mechanism of propagation. Our interpretations mainly rely on the  
127 relative timing of continental break-up of the three distinct oceanic sub-basins, each associated to  
128 a V-shaped propagator, and on the change in the direction of spreading and propagation at a time  
129 close to magnetic anomalies 7 or 6b (Figure 4). Both observations are robust<sup>23,25,26</sup>.

130         Our study aims at understanding the dynamics of the wide continental rift which preceded the  
131 opening of the central and west sub-basin, outlined in brown on Figure 4b and c. This continental  
132 rift formed when the East basin was spreading and propagating slowly (Figure 4a and b, blue  
133 spreading phase). Our modelling shows that from 32 to 23 Ma, the simultaneity of active  
134 continental rifting over the 1000 km width in the west SCS basin, with spreading in the East  
135 SCS, cannot be explained solely by a rheological contrast. The slow propagation of continental  
136 break-up during this 9 Myr period (Figure 4a and b, blue) is only compatible in rate with models  
137 with out-of-plane compression. On the contrary, the fast propagation periods observed in the  
138 central and west basin (Figure 4a and b, yellow and pink) are compatible in rate with the  
139 experiments with extensional or no tectonic loading at the front.

140         From the onset of rifting to 23 Ma, rifting occurs on structures that are perpendicular to the  
141 Indochinese Peninsula, as recorded by the magnetic anomaly in the East sub-basin and the  
142 abandoned rift branch of the Qiongdongnan basin (Figure 4a). As large topographic gradient can  
143 propagate compressive stress over hundreds of kilometers<sup>27,28</sup> and dynamic sandbox experiments  
144 have witnessed non-instantaneous continental break-up propagation in presence of topographic  
145 gradient<sup>29</sup>, we propose that the slow propagation results from the topographic load of Indochina.  
146         After the change in regional kinematics, the natural direction of propagation points obliquely to

147 the peninsula. From this moment, continental break-up starts to propagate through central and  
148 west basin at rates, which can reach 200 km per Myr, as predicted by models with no out-of  
149 plane compression.

150 The structure of the passive margin imaged along a refraction profile that cuts across the  
151 West and central SCS displays a series of basins overlying an almost flat Moho<sup>30</sup> (Figure 4c).  
152 These basins lay on top of the thinned lower crust region, like in the model displayed in Figure  
153 2a. These structures do not form with a stronger lower crust (Figure 3). Therefore, if the  
154 differences in width of passive margin from East to West also result from the well-documented  
155 change in the nature of the lithosphere<sup>31</sup>, our simulations suggest the lower crust should be  
156 weaker rather than stronger<sup>11,22</sup> in the West and Central Basin. Yet, the mechanical heterogeneity  
157 of the crust in this region<sup>26,31</sup> could well be responsible for the 6/5e stagnation period that our  
158 simplistic models do not explain.

### 159 **Oceanic propagators affect the width of passive margins**

160 The width and architecture of magma-poor passive margins are often used as a proxy to  
161 measure the initial effective strength of lithosphere, and particularly, the viscosity of the  
162 continental lower crust. Here, we have shown that with cylindrical or tri-axial extension  
163 boundary conditions, continental break-up propagates ten times faster than tectonic plate velocity  
164 erasing the sensitivity of continental rifting structures on crustal rheology.

165 Furthermore, we also demonstrated that a slight component of shortening applied in the  
166 direction of propagation reduces propagation rate of continental break-up to the order of  
167 magnitude of plate tectonic velocity. This deceleration of rift propagation is recorded  
168 geologically by a V-Shape oceanic basin, which cuts across a continental rift zone. While, we  
169 point out the Indochinese Peninsula in the case of the SCS, any large gradient in topography can



170 cause the formation of V-shaped propagator. In the case of scissor opening, the stretching  
171 gradient alone is responsible for the topographic gradient<sup>1</sup>.

172 As continental rifting continues at the front and on the side of the propagator during the  
173 phase of stalling, the structural complexity increases at the front of the propagator. When/if  
174 propagation resumes due to a change in plate motion or simply continues in slow motion,  
175 spreading might well occur on an offset segment as evidenced in models with stronger lower  
176 crust. This might well be the case in actively propagating d'Entrecasteaux Island propagator<sup>32,33</sup>,  
177 or in the Central Atlantic<sup>34</sup>.

178 The propagation of continental break-up is frequently neglected in the analysis of passive margin  
179 formation. Recent studies which include oceanic propagators<sup>13,14,35,36</sup>, together with this study,  
180 all contradict the conventional idea that the width and the duration of continental rifting reflects  
181 the initial mechanical layering of the lithosphere. In comparison with the initial rheological  
182 coupling of the lithosphere, three-dimensional kinematic constraints imposed by oceanic  
183 propagators have equal or even greater influence on the width of passive margins.

184

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277

## 278 **Acknowledgements**

279 Maps were realised with GMT, post processing with Paraview. We thank Weiwei Ding and M.

280 Pérez-Gussinyé for their reviews. LLP acknowledge financial support of ERC Advanced

281 Research Grant RHEOLITH (grant agreement No.290864). DAM acknowledge financial support  
282 from (FP7/2007–2013)/ERC (Grant Agreement N° 279925) and the Alfred P. Sloan Foundation  
283 through the Deep Carbon Observatory (DCO) "Modeling and Visualisation" project.

284

285 **Author contributions**

286 LLP designed the experiments, analysed the results of the models and wrote the first draft of the  
287 paper.

288 N.C-R drafted the map, discussed the timing of the anomaly and opening of SCS, MD helped for  
289 the comparison with the refraction profile and MP discussed the geodynamic setting and the  
290 geological arguments.

291 DAM set up the solver options to run the numerical code efficiently.

292 LW discussed the results and participated writing the paper.

293

294 **The authors declare no competing financial interests**

295 **FIGURE CAPTIONS**

296 Figure 1: Continental break-up propagation models and observations. a: Almost instantaneous  
297 opening far from the rotation pole. b: Scissor opening close to the rotation pole with associated  
298 active thickening at the front of the propagator. c: Sketch of propagator observed to form far  
299 from the rotation pole with the same colour code, also known as the Vink kinematic model.  
300 Contrary to a and b, this kinematic has never been demonstrated with dynamic experiments or  
301 simulations.

302

303 Figure 2: Modelling set-up and results in term of margin width and propagation of continental  
304 break-up. a, b and c present the results of the simulations 23 Myr after onset of rifting with  
305 different boundary conditions in the direction of propagation illustrated on the bottom left  
306 corners. The cross-sections bb' represent mantle temperature and crustal geometry overlaid with  
307 brittle strain (>50%). Synthetic magnetic anomalies (blue-grey) highlight spreading in the  
308 oceanic domain. d: Location of the oceanic propagator tip with time for the three models.  
309 Interpretative sketches of structures associated with the two modes of propagation.

310

311 Figure 3: Effect of rheology versus tectonic loading. Crustal b-factor reflects the local thinning  
312 of the crust on passive margin. On the four maps, the white area corresponds to the oceanic  
313 domain. The label “Weak” refers to weak lower crust. These “weak” models are the same as  
314 those presented in Figure 2. Results labeled “Strong” use a 10 times larger viscosity in the lower  
315 crust. Labels “Compression” and “Extension” refers to the out-of-plane tectonic loading  
316 described in Figure 2.

317

318 Figure 4: Rifting and spreading history of the South China Sea a: Bathymetry (ETOPO1)  
319 projected with pole of opening of stage 1 located E65/N09 (mean pole<sup>25</sup>). Colour of sea floor  
320 corresponds to the calendar presented in b and is based on the magnetic anomalies<sup>25</sup> shown as  
321 blue lines. The red line represents today's location of the bb' cross section of the compression  
322 model (Figure 2a). b: Continental break-up calendar (red arrows) is represented with the same  
323 axes as Figure 2d to highlight periods of acceleration and stalling. The red circle indicates the bb'  
324 cross-section. c: Interpreted seismic refraction profile<sup>29</sup> with the Moho in white and the base of  
325 the upper crust in black. See Methods for more details on the projection and basic reconstruction  
326 possibilities.



327

328 **Methods**

329 **Map:**

330 Indochina has not move relative to the north margin of the SCS during the time period  
331 considered so rather than proposing an ultimate version of these reconstructions that can be  
332 found in a number of papers, we use an oblique Mercator projection with the pole of projection  
333 being the mean pole of opening for the time period considered here.

334 The motion are then straight lines (small circles), just like in the model, and tracking the tip of  
335 the propagator or the margins at any time is straightforward: northward margin will not move,  
336 while southern margin will move straight. This holds until accretion switches to more SW-NE,  
337 i.e. the last propagating phase, which we are not comparing to the model.

338 **Modelling Method:**

339 This study uses pTatin3D<sup>37,38</sup> a highly scalable, massively parallel implementation of the  
340 finite element method, which employs an Arbitrary Lagrangian Eulerian discretisation, together  
341 with the material point method to solve for stokes flow

342 (1)  $\nabla \cdot (2\eta\mathbf{e}) - \nabla P = \rho\mathbf{g}$ .

343 It is coupled with time ( $t$ ) dependant heat conservation

344 (2)  $\nabla \cdot (\kappa\nabla T) = DT/Dt$

345 through non linear temperature ( $T$ ) and pressure ( $P$ ) dependant rheologies, detailed thereafter, as  
346 well as temperature dependence of buoyancy

347 (3)  $\rho = \rho_0(1 - \alpha(T - T_0))$

348 using Boussinesq approximation, which considers changes in density ( $\rho$ ) are small enough to  
349 approximate conservation of mass by an incompressible flow

350 (4)  $\nabla \cdot \mathbf{v} = 0$ ,

351 Where  $\mathbf{e}$  denotes the deviatoric strain rate tensor,  $\mathbf{v}$  the velocity,  $\rho_0$  the specific density at  $T_0$ .  $\kappa$   
352 and  $\alpha$  are respectively the thermal diffusivity and thermal expansion. Non-linear algebraic  
353 solvers are based on PETSc<sup>39</sup> and customized with specific solve options<sup>13,38</sup>.

#### 354 **Rheology:**

355 The rheology of the models has been kept to a minimum of complexity, i.e temperature  
356 dependence and brittle-plastic transition. In order to capture the brittle plastic transition at high  
357 confining pressure, Drucker Prager yield stress is limited by von Misses perfect plasticity at high  
358 deviatoric stress following  $\tau_y = \min[P \sin \phi + C \cos \phi, \tau_m]$ <sup>40</sup>. Effective viscosity for brittle-plastic  
359 behaviour depends on local second invariant of strain rate  $e^II$  following  $\eta_Y = \tau_y / e^II$ <sup>41</sup>. In order to  
360 simulate plastic wear, the friction ( $\phi$ ) decreases linearly from initial value of 30° to a final value  
361 of 6° after 30 % of plastic strain<sup>42</sup>.  $\tau_m$  is set to 200 MPa in the crust and 300 MPa in the mantle.  
362 The temperature dependence of the viscous strength of rock is approximated by the Frank-  
363 Kamenetskii flow rule  $\eta_T = \eta_0 \exp(-T/\theta)$ . The parameters  $\eta_0$  and  $\theta$  are set respectively to 10<sup>30</sup>  
364 Pa.s and 55.6°C in the mantle, 10<sup>26</sup> Pa.s and 50°C in the lower crust and 10<sup>27</sup> Pa.s and 50°C in  
365 the upper crust. Strong lower crust model have a lower crust viscosity 10 times larger than weak  
366 crust models. The effective viscosity  $\eta$  is the smallest of  $\eta_T$  and  $\eta_Y$ . We did not run model with  
367 dislocation creep law. This mode of deformation is however only expected to occur in the lower  
368 crust because at warmer temperatures, the mantle is expected to flow by diffusion creep. As we  
369 show in Figure 3 that models are only sensitive to lower crustal rheology in compression, we  
370 therefore believe that this approximation will not drastically affect our conclusions.

#### 371 **Domain, boundary conditions and initial conditions:**

372 The model domain is 600 km long in the direction of propagation ( $z$ ) and 1200 km long in  
373 the stretching direction ( $x$ ) and extends to 150 km depth. It is discretised using  $128 \times 32 \times 256$   
374 high order stable elements (Q2-P1), leading to spatial resolution of 5 km. Each simulation  
375 integrates deformation over a time span of 25 to 35 Myr, which corresponds to  $\square$  2000 time

376 steps on average. The initial geometry includes three horizontal layers: upper crust (20 km),  
377 lower crust (20 km) and mantle (110 km) with reference densities  $\rho_0$  set respectively to 2700,  
378 2800 and 3300 kg/m<sup>3</sup> at  $T_0 = 0^\circ\text{C}$ . The coefficient of thermal expansion,  $\alpha$  in equation 3, is  
379 constant and equal to  $2 \times 10^{-5} / ^\circ$  for all phases.

380         The initial geotherm corresponds to half-space cooling model for a thermal age of 300  
381 Myr and a constant thermal diffusivity of  $10^{-6} \text{ m}^2/\text{s}$ . This places the lithosphere-asthenosphere  
382 boundary as defined by the  $1300^\circ\text{C}$  isotherm at 120 km depth initially. Boundary conditions  
383 include symmetrical stretching at  $v_{x1/2} = 10 \text{ mm/yr}$  on the left and right sides, free slip on front  
384 boundary enforces the model to be cylindrical close to that side, the back side is either assigned  
385 free slip or a shortening/stretching velocity of  $v_z = \pm 2.5 \text{ mm/yr}$ . Infill of mantle material through  
386 the bottom boundary compensates for material leaving and entering through the side wall of the  
387 models.

388         In order to study propagation of break-up, it is necessary to ensure that break-up arise more  
389 rapidly on one side of the model. This initial localisation is obtained by ascribing larger  
390 amplitude (30% instead of a background value of 3%) of random brittle strain to a volume of  $100$   
391  $\times 100 \times 120 \text{ km}$  located at front of the domain. The deformation in this section of domain is not  
392 analyzed in the paper.

### 393         **Neglecting Magma supply:**

394         It has been proposed that in pristine normally thick lithosphere when the supply of magma  
395 is large, i.e. above large igneous provinces, individual dikes might propagate over  $1000 \text{ km}^{17}$ ,  
396 weakening the lithosphere over long distance. Our model neglects these magma supply effects.

397         Yet, modelling study and data show that if dike propagation is directional at short time  
398 scale (days), the polarity of propagation alternates at time scales of ten years or less. Upscaling  
399 to the timescale of one time step of our models (20 thousand years), dykes might act as a  
400 weakening factor but they would not introduce directionality. At long timescale, propagation of

401 diking is therefore slaved to the magma supply, which occurs beneath the zone of active necking  
402 of the mantle. Our model captures the propagation of these lithospheric weakness zones, which  
403 are driven by tectonic stress. The lithospheric column located under the ridge is indeed extremely  
404 weak and buoyant due to temperature (see Figure 2). When these highly localised zones of warm  
405 mantle material reach the surface, they approximate the mid-oceanic ridges at large scale and  
406 influences tectonic stress.

407

#### 408 **Data availability**

409 The data used for the refraction profile that support the findings of this study are available  
410 from the corresponding author upon request. The data used to produce the map are taken from  
411 ETOPO1 public database<sup>44</sup>, and magnetic anomalies have been digitalized from Briais et al.<sup>25</sup>.

#### 412 **Code availability**

413 The code used to generate the numerical simulation<sup>37,38</sup> can be access at  
414 <https://bitbucket.org/ptatin/ptatin3d>. Simulations can be reproduced using model RIFT3D\_T with  
415 example option file SCS.opts.

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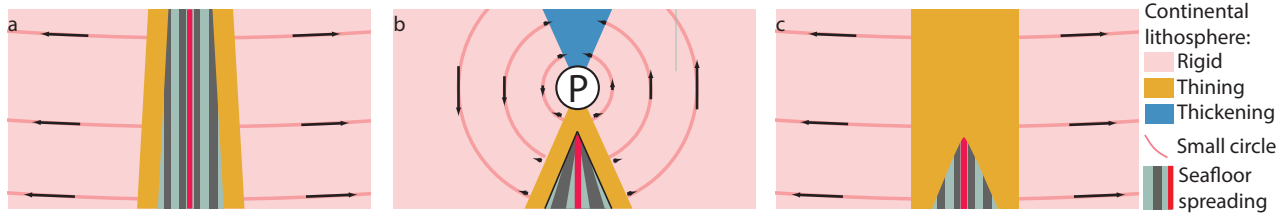
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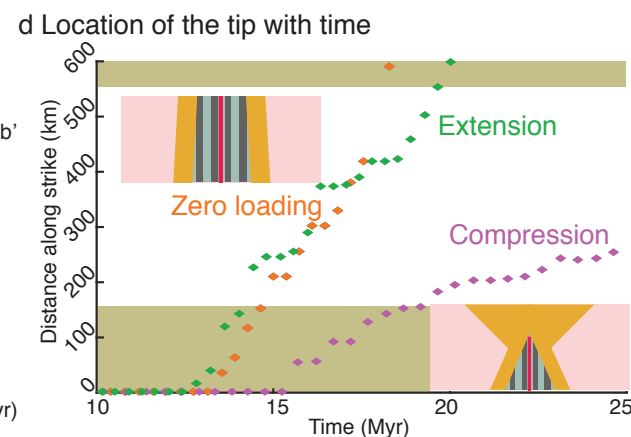
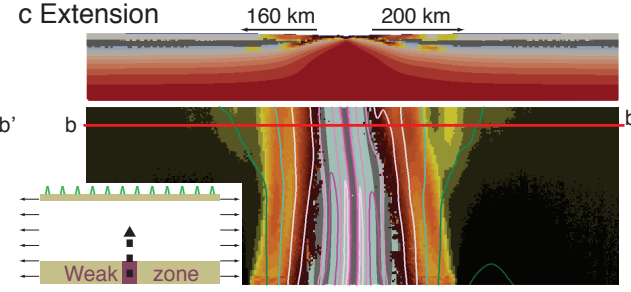
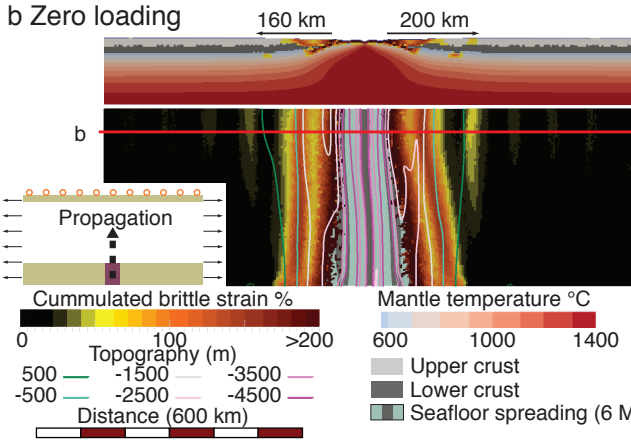
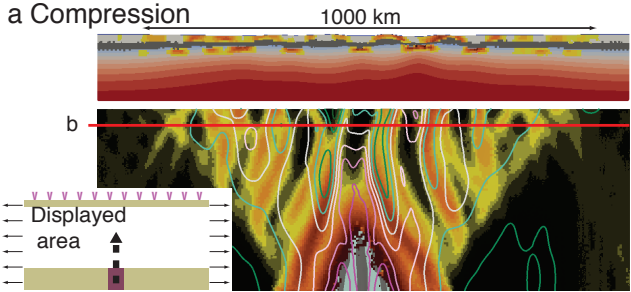
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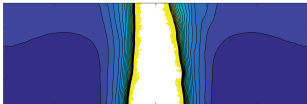
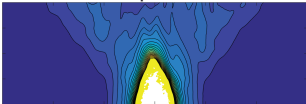




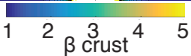
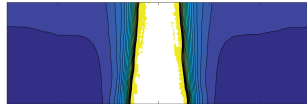
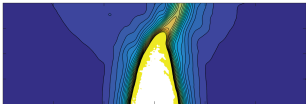
Compression

Extension

Weak



Strong



Distance (600 km)

