

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

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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**

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

The relative movement of rigid plates on the sphere is described by rotation around a pole. Assuming that continental break-up occurs after some constant amount of continental extension, the propagation rate of break-up should scale with the gradient of stretching rate along strike. Far from the pole of rotation, the velocity of opening is nearly constant over hundreds of kilometres and the model predicts that break-up should occur almost instantaneously¹ (Figure 1a). Within this theory, V-shaped oceanic basins can form only in the close vicinity of the pole of rotation, where variations in the stretching rate are important (Figure 1b). This scissors opening kinematic² model has been reproduced by dynamic experiments³ and numerical simulations¹. A consequence of this model is that active shortening and thickening of the crust must be observed in the close vicinity of the propagator (Figure 1b). Yet, several well-documented V-shaped propagators, like the East Basin of the South China Sea⁴, or the Coral Sea^{5,6}, are forming far from the pole of rotation within zones of diffuse continental rifting⁷ (Figure 1c).

It is widely accepted that the rheological layering of the lithosphere controls the width and structure of passive margins^{8,9,10}. Dynamic models have pointed out that mechanical heterogeneities delay the localisation of deformation at the onset of continental rifting^{11,12}.

51 Extrapolating these models to continental break-up stage, it is generally hypothesized that Vink
52 kinematic model⁷ (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^{13,14}.
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^{15,16}. 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 supply¹⁷ (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^{8,10,18}. 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 age¹⁹ 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 crust⁸. 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^{9,10,18,20}. 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-

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

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

The South China Sea propagator stalled due to Indochina

The South China Sea (SCS) is the largest area of immersed continental crust in the world. The SCS has all the characteristics of a wide continental rift, but crust is thinner and extension has led to continental break-up²¹. In map view, the width of the passive margin increases from 400 km up to 900 km towards the west²² as the age of sea floor decreases, an observation that is clearly incompatible with typical scissor opening kinematics (Figure 1b). In the following, geological ages are given in million annum (Ma) to contrast with time that remains in million years (Myr).

From 32 Ma (magnetic anomaly 11) to 23 Ma (7-6b) (Stage 1, Figure 4b), the East SCS sub-basin is the only one that accommodates regional extension by spreading. The North-South opening direction at the time is best described by a pole of rotation located E65/N09, far enough from the SCS to dismiss again the scissor-type model. From anomaly 6a the direction of opening changes by 15°²³ as outlined by the change of orientation in magnetic anomaly in the yellow oceanic domain (Figure 4a). The break-up propagates rapidly (Stage 2, Figure 4b) through the 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^{24,25} 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^{23,25,26}.

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^{27,28} and dynamic sandbox experiments
144 have witnessed non-instantaneous continental break-up propagation in presence of topographic
145 gradient²⁹, 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 Moho³⁰ (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 lithosphere³¹, our simulations suggest the lower crust should be
156 weaker rather than stronger^{11,22} in the West and Central Basin. Yet, the mechanical heterogeneity
157 of the crust in this region^{26,31} 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 gradient¹.

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^{32,33},
177 or in the Central Atlantic³⁴.

178 The propagation of continental break-up is frequently neglected in the analysis of passive margin
179 formation. Recent studies which include oceanic propagators^{13,14,35,36}, 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.

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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 pole²⁵). Colour of sea floor
320 corresponds to the calendar presented in b and is based on the magnetic anomalies²⁵ 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 profile²⁹ 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^{37,38} 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 (ρ) are small enough to
349 approximate conservation of mass by an incompressible flow

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

Where \mathbf{e} denotes the deviatoric strain rate tensor, \mathbf{v} the velocity, ρ_0 the specific density at T_0 . κ and α are respectively the thermal diffusivity and thermal expansion. Non-linear algebraic solvers are based on PETSc³⁹ and customized with specific solve options^{13,38}.

Rheology:

The rheology of the models has been kept to a minimum of complexity, i.e temperature dependence and brittle-plastic transition. In order to capture the brittle plastic transition at high confining pressure, Drucker Prager yield stress is limited by von Misses perfect plasticity at high deviatoric stress following $\tau_y = \min[P \sin \phi + C \cos \phi, \tau_m]$ ⁴⁰. Effective viscosity for brittle-plastic behaviour depends on local second invariant of strain rate e^II following $\eta_Y = \tau_y / e^II$ ⁴¹. In order to simulate plastic wear, the friction (ϕ) decreases linearly from initial value of 30° to a final value of 6° after 30 % of plastic strain⁴². τ_m is set to 200 MPa in the crust and 300 MPa in the mantle. The temperature dependence of the viscous strength of rock is approximated by the Frank-Kamenetskii flow rule $\eta_T = \eta_0 \exp(-T/\theta)$. The parameters η_0 and θ are set respectively to 10³⁰ Pa.s and 55.6°C in the mantle, 10²⁶ Pa.s and 50°C in the lower crust and 10²⁷ Pa.s and 50°C in the upper crust. Strong lower crust model have a lower crust viscosity 10 times larger than weak crust models. The effective viscosity η is the smallest of η_T and η_Y . We did not run model with dislocation creep law. This mode of deformation is however only expected to occur in the lower crust because at warmer temperatures, the mantle is expected to flow by diffusion creep. As we show in Figure 3 that models are only sensitive to lower crustal rheology in compression, we therefore believe that this approximation will not drastically affect our conclusions.

Domain, boundary conditions and initial conditions:

The model domain is 600 km long in the direction of propagation (z) and 1200 km long in the stretching direction (x) and extends to 150 km depth. It is discretised using $128 \times 32 \times 256$ high order stable elements (Q2-P1), leading to spatial resolution of 5 km. Each simulation 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 ρ_0 set respectively to 2700,
378 2800 and 3300 kg/m³ at $T_0 = 0^\circ\text{C}$. The coefficient of thermal expansion, α 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°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 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

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

Data availability

The data used for the refraction profile that support the findings of this study are available from the corresponding author upon request. The data used to produce the map are taken from ETOPO1 public database⁴⁴, and magnetic anomalies have been digitalized from Briais et al.²⁵.

Code availability

The code used to generate the numerical simulation^{37,38} can be access at <https://bitbucket.org/ptatin/ptatin3d>. Simulations can be reproduced using model RIFT3D_T with example option file SCS.opts.

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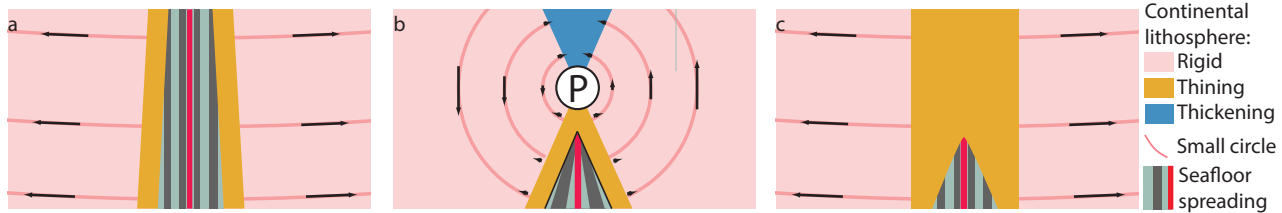
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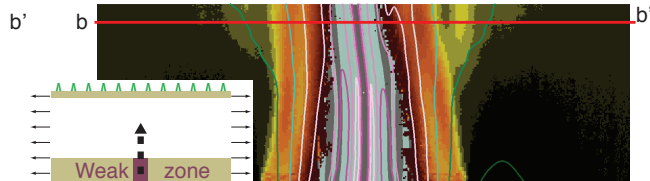
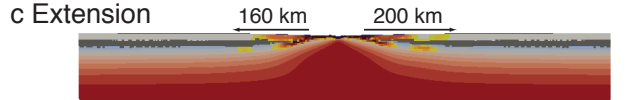
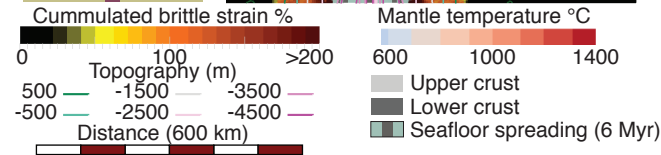
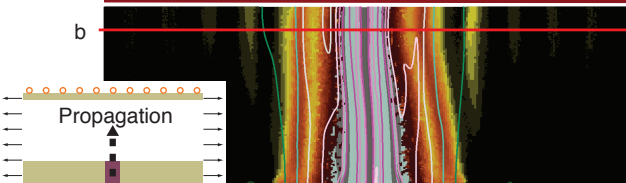
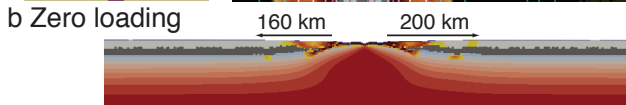
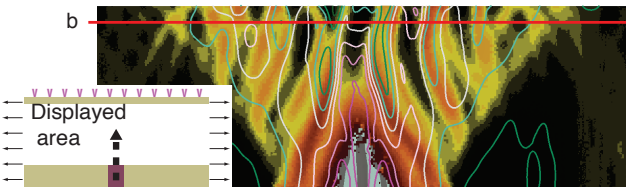
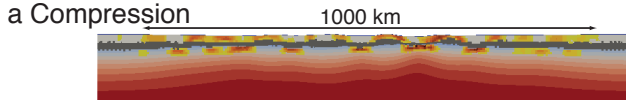
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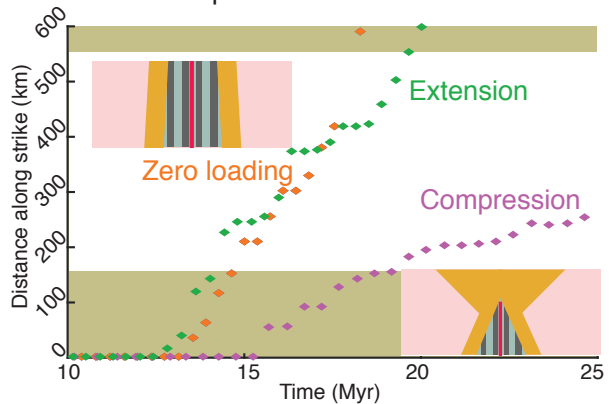
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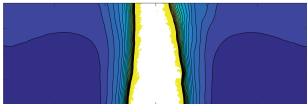
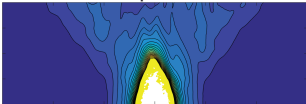
d Location of the tip with time



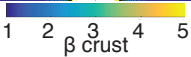
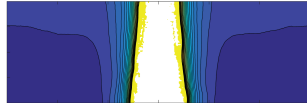
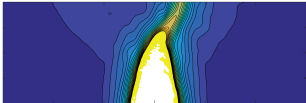
Compression

Extension

Weak



Strong



Distance (600 km)

