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Anthony Jourdon, Laetitia Le Pourhiet, Frédéric Mouthereau, Dave A. May. Modes of propagation of continental break-up and associated oblique rift structures. *Journal of Geophysical Research : Solid Earth*, 2020, 125 (9), <10.1029/2020JB019906>. <hal-02985979>

HAL Id: hal-02985979

<https://hal.sorbonne-universite.fr/hal-02985979v1>

Submitted on 2 Nov 2020

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Modes of propagation of continental break-up and associated oblique rift structures

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Abstract

V-shaped oceanic propagators are widespread around the world. Their geometry combined with magnetic anomalies associated to their opening shows at first order that ridge propagation in the third dimension occurs by pulses. In this study we use 3D thermomechanical numerical models to show how oblique kinematic boundary conditions control both the intracontinental rift development and the oceanic ridge propagation. To do so, we apply a shortening velocity boundary condition in the direction perpendicular to the extension for “strong” and “weak” crustal rheologies. Numerical models results highlight that three ridge propagation modes can occur. For low out of plane velocities (12% to 15% of the extension rate), the ridge propagation is fast (>1.5 cm/yr) and straight. Higher shortening velocities (15% to 17%) lead to a ridge propagation by pulses alternating between fast propagation (~1.5 cm/yr) and stalling phases. Finally, for higher velocities (17% to 20%) a ridge jump propagation mode occurs, localizing a new spreading centre between 100 and 200 km far from the initial ridge. We also show that ridge propagation phases are associated with dip-slip dominated deformation while stalling phases are dominated by strike-slip deformation. These deformation regimes are marked by structures reorientation while kinematic boundary conditions remain constant. We discuss these results in term of plate tectonics reconstructions and regional geological studies.

1 Introduction

Magnetic anomalies record the past direction and rate of seafloor spreading (Vine & Matthews, 1963). As such they are our best constrain for global plate tectonic reconstructions (Müller et al., 2008; 2016; Seton et al., 2012). In perfectly cylindrical condition the orientation of the first magnetic anomaly is parallel to the rifted margin of continents. However, in some areas the first magnetic anomalies intercept the rifted margins of continents at acute angle instead of being parallel to it. These remnants are precious as they allow us to quantify the rate at which continental break-up propagates along the strike of the mid-oceanic ridges (Courillot, 1982). Indeed, using only basic kinematic considerations, it is clear that if the propagation occurs at the same velocity as the spreading rate, the angle between the margin and the magnetic anomaly should be roughly 45°. Faster propagation leads to the parallelisation of the magnetic anomaly with the margin of the oceanic basin matching with the 2D cylindrical model. Slower propagation produces magnetic anomalies that normalise to the rifted margin, until the limit case of a transform margin (no ridge propagation at all).

In map, these phases of sluggish ridge propagation produce V-shaped margins. V-shaped propagators are found in many places on Earth (Figure 1) and seafloor age map compilation by Müller et al. (2008) is often sufficient to unravel the first order features of the timing of continental break-up at regional or more global scale. Examples displayed in Figure 1 show propagators separated by very long segments

41 where magnetic anomalies are parallel to the margin. This indicates that pulsation in the rate of
42 continental break-up propagation first introduced by Courtillot (1982) in the Gulf of Aden is a major
43 feature of divergent plate boundaries. Despite the well documented geological record of these changes
44 of rate, and their implications for plate tectonics and kinematic reconstructions or the thermal regime
45 of oblique margins, the dynamics of the propagation at long/geodynamic timescale remains poorly
46 studied or understood.

47 Kinematic interpretations of V-shaped propagators are based on models established in the late 1970s
48 and early 1980s, which are still in use today. The first model, called the rotational extension (Hey et
49 al., 1980; Molnar et al., 2017; Mondy et al., 2018; Zwaan et al., 2020), assumes that seafloor spreading
50 occurs near the Euler pole and is therefore described by very large variations in the opening rate in the
51 direction of the ridge. In this case, the first magnetic anomaly forms parallel to the rifted margin then
52 the linearly increasing rates of opening with distance from Euler pole produce a series concentric V's
53 over time Figure 2c. This model infers that very rapid changes in tectonic regime must occur at the tip
54 of the V-shaped ocean domain. The second kinematic model, published by Vink (1982), considers that
55 the divergence remains constant in the direction of rift axis but it is variably partitioned between the
56 oceanic accretion and the continental rifting (Figure 2b). This model applies only when magnetic
57 anomalies indicate that the Euler Pole is far enough so that the rate of divergence is constant at
58 regional scale. This is generally the case for a distance greater than 3000 km or 30° from the Euler
59 pole. The Vink model allows the formation of V-shaped propagators when the propagation rate is
60 similar to the oceanic accretion rate. For faster propagation, magnetic anomalies are parallel to the
61 margin (Figure 2a). The Vink model is the basis for Courtillot's (1982) proposal (Figure 2d) who
62 interpreted the change in angle between magnetic anomalies and the continental margins along the
63 direction in the Gulf of Aden as the geological record of pulsating continental break-up propagation. It
64 should be noted here that these two kinematic models are not contradictory but rather complement
65 each other as two end-member cases. However, Vink's and Courtillot's models predictions in terms of
66 magnetic anomalies and associated tectonic structures at the front of the propagator are different and
67 therefore permit the description of different natural examples of V-shaped propagators. The Woodlark
68 Basin only opens 500 km from the Euler Pole. It displays a strong segmentation, variations from
69 extensional to strike-slip structure (Holm et al., 2016; Mondy et al., 2018) and only the youngest ridge
70 ages at the tip which conform with characteristics of a rotational extension opening basin. While the
71 South China Sea that opened at about 4500 km from its pole of rotation (Brais et al., 1993; Mazur et
72 al., 2012; Le Pourhiet et al., 2018) with very little variations in the spreading rate along the strike
73 would better illustrate the Vink kinematic model.

74 Dynamically speaking, on the one hand, the rotational opening model has been reproduced by a
75 number of dynamic simulations based either on analogue materials (Molnar et al., 2017; Zwaan et al.,
76 2020) or numerical methods (Mondy et al., 2018) while, on the other hand, almost no properly scaled
77 dynamic simulations, including numerical (Allken et al., 2012; Beauissier et al., 2019; Liao & Gerya,
78 2015; Laetitia Le Pourhiet et al., 2017) or analogue models (Ding & Li, 2016; Zwaan et al., 2016),
79 reproduces the kinematics proposed by Vink (1982) at lithospheric scale. With these boundary
80 conditions and a realistic viscosity in the asthenospheric mantle, lithospheric necking should propagate
81 approximately ten time faster than plates diverge (Allken et al., 2012; Laetitia Le Pourhiet et al., 2018)
82 resulting invariably in the fast propagation kinematic displayed in Figure 2a. Therefore, to produce a
83 kinematic solution, which resembles Vink's (1982) model (Figure 2b) and a number of geological
84 observations (Figure 1), some additional ingredients must therefore be added to the dynamic
85 simulations to stop or slow down the propagation in order to form a V-shaped oceanic propagator.

86 The obvious candidate to slow down the propagation of continental break-up is the presence of a
87 strong rheological heterogeneity, so called strong ribbon (Figure 2d). This hypothesis was first
88 proposed by Morgan & Parmentier (1985) based on the theory of crack propagation. It was supported
89 by small strain numerical simulations of the propagation of continental rifting published by Van Wijk
90 & Blackman (2005). The results of this first study are consistent with the results presented by Allken
91 et al. (2012): the propagation rate of a neck (continental rift) in a frictional layer (upper crust) lying
92 above a viscous layer (lower crust) decreases with the viscosity of the lower layer. Many similar
93 analogue or numerical experiments (Allken et al., 2012; Brune et al., 2017; Corti, 2012) have proved
94 that increasing the viscosity of the lower crust resulted in a slower lengthening of grabens in the upper
95 crust. However, at lithosphere scale there is nothing like a high viscosity asthenosphere to delay or
96 slow the propagation of continental break-up. At that scale, the strength of the lower crust has been
97 shown to enhance fast lithospheric necking and tends to increase the rate of propagation of continental
98 break-up (Benes & Scott, 1996; Le Pourhiet et al., 2018).

99 Oblique continental rifting controlled by the presence of pre-existing viscous weak zones oblique to
100 the extension direction creates en-échelon rift basins, which are separated by the transfer zones that
101 accommodate strain differences between segments (Ammann et al., 2017; Brune, 2014; Duclaux et al.,
102 2020; Mart & Dauteuil, 2000; McClay & White, 1995; Molnar et al., 2017). These isolated basins
103 form small segments of oceanic accretion before complete lithosphere continental plate break-up is
104 achieved. As they grow in width they also grow in length, propagating from the basin center towards
105 both the extremity. Yet their propagation is limited by the compressional stress build-up that arises
106 from the growth of nearby segment. Le Calvez & Vendeville, (2002) proposed a model set-up to study
107 in details the dynamic of linkage between en-échelon segments. This set-up consists of two offset
108 weak zones which grow laterally in map view and link or not depending on spacing, frictional layer
109 thickness and viscosity of the underlying layer. Allken et al. (2012) presented the result for a similar
110 set of numerical experiments and showed the importance of the rate of softening in the frictional layer
111 to produce a transform fault. More recently, Le Pourhiet et al. (2017) have extended this model to
112 study continental break-up at lithosphere scale (Figure 2e). At that scale, the linkage between
113 segments can be delayed by 10's of millions of years. Meanwhile, two asymmetric V-shaped
114 propagators form, as shown in Figure 2e. Full linkage occurs only after oblique necking of the mantle
115 lithosphere is reached showing similarities with the initial conditions of Ammann et al. (2017) study in
116 which an oblique weak zone is imposed. These asymmetric propagators better fit the observations in
117 the Gulf of Aden or in the equatorial Atlantic (Le Pourhiet et al., 2017) than the original propagator
118 models do (Figure 2a, b), but they do not apply to all geological examples of V-shaped propagators.

119 To produce oblique extension, pure shear boundary conditions in map view similar to the Alboran
120 Sea set up in Le Pourhiet et al., (2014) (i.e. stretching velocity in one horizontal direction and a
121 shortening velocity in the other horizontal direction) are a good alternative to the introduction of
122 oblique heterogeneities in cylindrical boundary conditions. Indeed, strain localisation causes the
123 formation of structures that can rotate internally within the pure shear far field strain and which
124 favours the emergence of instable 4 plates system that stabilizes into stable 3 plates systems (Gerya &
125 Burov, 2018). It has been demonstrated that pure shear conditions produce V-shaped propagators both
126 during the reorganisation of oceanic plates (Gerya & Burov, 2018) and during continental break-up
127 (Le Pourhiet et al., 2018). This results together with the analogue experiments of Ding & Li (2016),
128 who introduced a gradient in gravitational potential energy facing the propagation direction, suggests
129 that applying a small amount of shortening or compression out of the plane of the main driving
130 extension slows down continental rifting propagation rate. One should note that in 3D the transition
131 from transtensional to transpressional regime in pure shear map setting is related to change in surface

132 of the model and that as long as the surface remain in dilatational regime no shortening related
 133 structures form (Le Pourhiet et al., 2014). GPS velocities vectors plotted relative to fix Arabian plate
 134 in figure 1 clearly show frontal displacement towards actively propagating oceanic ridge exist in active
 135 propagating continental rifts of Aden and Red Sea. In the first case, one might suspect the influence of
 136 the Afar plume and in the second case the vicinity of the Mediterranean subduction zone. In both cases
 137 the frontal shortening is not associated to shortening tectonic structures which could “fossilize” its
 138 existence.

139 Here, we therefore intend to focus our attention in this still largely unexplored direction. We adopt the
 140 same boundary conditions as Le Pourhiet et al. (2018) (Figures 2f and 3) in order to analyse the
 141 sensitivity of continental break-up propagation rate to both the rate of ridge parallel shortening and
 142 lower crustal rheology. We first extract from the simulations synthetic propagation rates and expected
 143 magnetic anomalies in opening basins that corresponds to data that are widely available, relate them to
 144 large scale plate deformations. We then focus on the tectonic structures that accommodates the frontal
 145 shortening in the models in order to discuss what are key tectonic observations that might help to
 146 decipher past kinematics and lithospheric dynamics from seismic lines and thermal dataset.

147 **2 Set Up of the thermo-mechanical simulations**

148 *2.1 Modelling approach*

149 In order to model the long term deformation of the lithosphere we use pTatin3D (May et al., 2014,
 150 2015), a highly scalable, massively parallel implementation of the finite element method which
 151 employs an Arbitrary Lagrangian-Eulerian (ALE) discretization together with the material point
 152 method to solve the conservation of momentum for an incompressible fluid:

$$153 \quad \nabla \cdot (2\eta \dot{\boldsymbol{\epsilon}}) - \nabla P = \rho \mathbf{g} \quad (1)$$

$$154 \quad \nabla \cdot \mathbf{v} = 0 \quad (2)$$

155 where η is the non-linear effective viscosity, $\dot{\boldsymbol{\epsilon}}$ the strain rate tensor, P the pressure, ρ the density, \mathbf{g} the
 156 gravity acceleration vector and \mathbf{v} the velocity vector of the fluid. This Stokes flow is coupled with the
 157 time dependent heat conservation equation:

$$158 \quad \frac{\partial T}{\partial t} + \mathbf{v} \cdot \nabla T = \nabla \cdot (\kappa \nabla T) + \frac{H}{\rho C_p} \quad (3)$$

159 where T is the temperature, \mathbf{v} the velocity vector of the fluid, κ the thermal diffusivity, H the heat
 160 source (here only radioactive heat sources are considered) and C_p is the heat capacity. The material
 161 density is considered to vary with temperature such as:

$$162 \quad \rho = \rho_0 (1 - \alpha(T - T_0)) \quad (4)$$

163 where ρ_0 is the initial material density and α the thermal expansion coefficient.

164 The Stokes problem is solved using a Q_2 - P_1 discretization while the heat equation is
 165 discretized on Q_1 elements. pTatin3D uses a free surface boundary condition which can dynamically
 166 evolve due to deformation and surface processes that we model with a 2D diffusion equation
 167 discretized on Q_1 elements:

$$168 \quad \frac{\partial h}{\partial t} = -\nabla \cdot (\kappa \nabla h) \quad (5)$$

169 where h is the altitude and k the diffusion coefficient. The equation is solved using a first order
170 explicit time integration.

171 2.2 *Rheological model*

172 To simulate the long time behaviour of the lithosphere we use temperature and pressure dependent
173 non-linear rheologies. Brittle parts of the lithosphere follow the Drucker-Prager pseudo-plastic yield
174 criterion adapted to continuum mechanics:

$$175 \quad \eta_p = \frac{C \cos(\phi) + P \sin(\phi)}{\dot{\epsilon}^{II}} \quad (6)$$

176 where C is the cohesion, ϕ the friction coefficient, P the pressure and $\dot{\epsilon}^{II}$ the second invariant of the
177 strain rate tensor. To simulate the friction drop in deforming faults, we use a simple linear decrease of
178 the friction angle from 30° to 5° with accumulated plastic strain from 0 to 1. However, laboratory
179 experiments show that under high confining pressures (> 1 GPa) brittle behaviour changes to plastic
180 behaviour (e.g. Kameyama et al., 1999; Precigout et al., 2007). We account for that change by
181 limiting the Drucker-Prager yield stress to a maximum deviatoric stress of 400 MPa according to the
182 calibration performed by Watremez et al. (2013).

183 Viscous parts of the lithosphere are simulated using an Arrhenius flow law for dislocation
184 creep:

$$185 \quad \eta_v = A^{-\frac{1}{n}} (\dot{\epsilon}^{II})^{\frac{1}{n-1}} \exp\left(\frac{Q+PV}{nRT}\right) \quad (7)$$

186 where A , n and Q are material dependant parameters (see Table 1), R is the gas constant and V is the
187 activation volume.

188 2.3 *Initial conditions*

189 Our experiments represent a 1200 (in the x direction) x 600 (in the z direction) x 250 (in the y
190 direction) km^3 domain discretized with 512x256x128 elements respectively (Figure 3a). In the y
191 direction, two third of the 128 elements are concentrated in the first third of the 250 km giving
192 approximately 8 km^3 ($2 \times 2 \times 2 \text{ km}^3$) elements in the first 80 km. Four rheological layers subdivide the
193 models domain to represent the upper and lower crust (each of 20 km thick), the lithosphere mantle
194 (80 km thick) and the asthenosphere (170 km thick). The Moho is located at 40 km depth and the
195 lithosphere-asthenosphere boundary is at 120 km depth (Figure 3a). Lithosphere mantle and
196 asthenosphere are both simulated using a dry olivine flow law (Hirth & Kohlstedt, 2003) (Table 1).

197 Depending on the model, the crust rheology may vary. The rheology of the lower crust has been
198 proven to be of first order importance for the strain localization and propagation in 3D (Allken et al.,
199 2012; Le Pourhiet et al. 2014; Brune et al., 2017; Corti, 2012; Le Pourhiet et al., 2017; 2018).
200 Therefore, in this study we use two different crustal lithologies denoted as weak and strong (Figure
201 3c). The weak crust corresponds to an upper and a lower crust modelled with a quartz rheology
202 (Ranalli & Murphy, 1987) while the strong crust corresponds to an upper crust modelled with a quartz
203 rheology and the lower crust with an anorthite rheology (Rybacki & Dresen, 2000). We modelled
204 those two rheological layering with the same boundary and initial conditions in order to assess the role
205 of the lower crust on the evolution of the rifting and continental break-up.

206

207

	Units	Quartz	Anorthite	Olivine
Reference		Ranalli and Murphy 1987	Rybacki and Dresen 2000	Hirth and Kohlstedt 2003
A	Mpa ⁻ⁿ .s ⁻¹	6.3x10 ⁻⁶	13.4637	1.1x10 ⁵
n		2.4	3	3.5
Q	KJ/mol	156	345	530
V	m ³ .mol ⁻¹	0	3.8x10 ⁻⁵	18x10 ⁻⁶

208

209 The initial temperature field is computed with a steady-state analytical solution (Turcotte &
210 Schubert, 2002)

211

$$212 \quad T_{\text{init}} = T_{y=0} + \frac{-yq_m}{k} + H \frac{y_p^2}{k} \left(1 - \exp\left(\frac{-y}{y_p}\right) \right) \quad (8)$$

213 for a $T_{y=0} = 0^\circ\text{C}$, an incoming mantle heat flux $q_m = 20 \text{ mW.m}^{-2}$, a radiogenic heat production $H =$
214 $1.2 \times 10^{-6} \text{ W.m}^{-3}$, a characteristic radiogenic layer of $y_p = 40 \text{ km}$ and a conductivity of $3.3 \text{ W.m}^{-1}.\text{K}^{-1}$.
215 This analytical solution gives 610°C at 40 km (depth of the Moho, Figure 3c) and 1300°C at 120 km
216 (depth of the LAB, Figure 3c). Then from 120 km to 250 km we compute a linear increase of
217 temperature for an adiabatic gradient of 0.5°C/km (Figure 3c). Although this second part of the
218 geotherm is not steady state, the cooling by diffusivity is very slow (less than 2°C/Myr for the
219 maximum cooling rate) and it allows to keep reasonable conductivity values in the asthenosphere (3.3
220 $\text{W.m}^{-1}.\text{K}^{-1}$).

221 The initial radiogenic heat production is set as an exponential decay of heat production with depth
222 according to Turcotte & Schubert (2002) as follow:

$$223 \quad H = H_0 \exp\left(\frac{-y}{y_p}\right) \quad (9)$$

224 for a surface production H_0 of $1.2 \times 10^{-6} \text{ W.m}^{-3}$.

225 In order to initiate localisation on one side of the modelling domain, we define a weak zone in
226 which a random amount of initial plastic strain ranging from 0 to 1 is applied to the particles located in
227 a cube of coordinates $x=550$ to $x=650$, $y=0$ to $y=100$ and $z=0$ to $z=100$ instead of 0.0 to 0.03
228 background value. Since we use plastic strain softening, the initial amount of plastic strain will
229 decrease the initial friction angle in the concerned area as:

$$\phi = \phi_0 - \frac{\epsilon_p - \epsilon_{\min}}{\epsilon_{\max} - \epsilon_{\min}} (\phi_0 - \phi_\infty)$$

230 Where ϕ is the friction angle, ϕ_0 the initial maximum friction angle (30°), ϕ_∞ the minimal friction
231 angle reached after softening (5°), ϵ_{\max} and ϵ_{\min} are the bounds of maximum and minimum plastic
232 strain amount between which softening is applied and ϵ_p the actual plastic strain computed from
233 equation (6).

234 2.4 *Boundary conditions*

235 The boundary conditions to solve the heat conservation equation are $T=0^{\circ}\text{C}$ at the top boundary ($y=0$
236 km), $T=1440^{\circ}\text{C}$ at the bottom boundary ($y=250\text{km}$) and null heat fluxes on all vertical boundaries of
237 the domain. To solve the conservation of momentum, we impose a free surface on the top boundary, a
238 free slip boundary conditions on the face of normal z at coordinates $x, y, z=0$, and Dirichlet boundary
239 conditions on the normal component of velocity on all other vertical sides. An extension velocity $v_x=5$
240 mm/yr is imposed on both faces of normal x and a shortening velocity v_z is imposed on the face of
241 normal z at coordinates $x, y, z=600$ (Figure 3a, b). This shortening velocity may vary between models
242 (from 1.2 mm/yr to 2.0 mm/yr). It introduces a part of constriction in the model domain that
243 intrinsically modifies the ratio between normal stretching and thinning as compared to a cylindrical
244 model. In this range of shortening rates, propagation rates are indeed of the same order of magnitude
245 (less than 1.5 cm/yr) of the plates velocity imposed in the model (1 cm/yr) and it is for this range of
246 velocities that we expect the highest dependence of deformation pattern to the rheology of the
247 lithosphere (Le Pourhiet et al., 2018).

248 2.5 *Post processing*

249 In order to compare the numerical models results with natural examples and known geological and
250 geophysical data, we computed an equivalent of seafloor magnetic anomalies. Magnetic anomalies
251 measured on the seafloor are used to model seafloor spreading rates and oceanic crust ages in oceans (
252 Müller et al., 2008). In our models, the formation of oceanic crust by partial melting of the mantle and
253 magma crystallization is not implemented. Therefore, we compute an age of mantle exhumation below
254 the 800°C isotherm which can be roughly approximated as the Curie Temperature of ferromagnetic
255 material (Rushbrooke & Wood, 1958). This age corresponds to the time in the model at which the
256 mantle cooled beneath 800°C . In the models we use 0 Myr as the beginning and time goes forward
257 from this initial value while geological time is counted from present (0 Ma) to past. Therefore, the age
258 of the mantle exhumed during the latest stages of the models is greater than the age of the mantle
259 exhumed in the earlier stages. Those ages values need to be interpreted as the greater is also the
260 younger.

261

262 Figures 8, 10 and 12 and supplementary figures 11 and 12 display the deformation regime computed
263 from the stress tensor orientation also known as the Regime Stress Ratio (Brune, 2014; Brune &
264 Autin, 2013 and reference therein). The colour intensity is related to the strain rate second invariant
265 value.

266 Figures 4, 5, 7, 8 and 9 show the plastic and/or viscous strain deformation. Those values represent
267 respectively the cumulated finite strain computed with the equations (6) and (7). Therefore, the plastic
268 strain displays the deformations related to frictional (or brittle) parts of the lithosphere while the
269 viscous strain shows the deformations related to the dislocation creep flow laws in the “ductile” parts
270 of the lithosphere.

271 Supplementary Figure 1 shows the β factor of the crust computed as

$$272 \quad \beta = \frac{h_c^{t=0}}{h_c^{t=n}} \quad (10)$$

273 where $h_c^{t=0}$ is the initial thickness of the crust at the beginning of the simulation ($h_c = 40$ km, $t = 0$
274 Myr) and $h_c^{t=n}$ is the thickness of the crust at a given time n . Thus, the β factor value represents the
275 thinning ratio of the crust at a given time.

276 Figure 6a displays the position of the tip and the ocean propagation with respect to time. The position
 277 of the tip is measured as the higher z coordinates of the mantle-continent transition along the strike of
 278 the oceanic ridge. Therefore, it represents a particular area at the frontier between a plate boundary
 279 (the oceanic ridge) and an intra-continental deformation zone (the rift). As discuss in the numerical
 280 results section, this area may evolve with time in a triple junction or remain as a plate boundary to
 281 intra-continental deformation transition. The ridge propagation (Figure 6b) is the time derivative of
 282 this position computed as:

$$283 \quad Vrp = \frac{\Delta z}{\Delta t} = \frac{z_{t=n} - z_{t=n-1}}{t_n - t_{n-1}} \quad (11)$$

284 where Vrp is the ridge propagation velocity, $z_{t=n}$ and $z_{t=n-1}$ are the z coordinates at time t_n and t_{n-1}
 285 respectively. Figure 6c shows the tectonic forces evolution through time. F_x and F_z represent the
 286 forces normal to the faces of normal x and z respectively. To compute forces we first compute the
 287 deviatoric stress tensor as:

$$288 \quad \boldsymbol{\sigma}' = 2\eta\dot{\boldsymbol{\epsilon}} \quad (12)$$

289 where η is the viscosity and $\dot{\boldsymbol{\epsilon}}$ the strain rate tensor. We then compute the traction along faces of
 290 normal x and z respectively:

$$291 \quad \vec{T}_x = \iint \boldsymbol{\sigma}' \cdot \vec{x} \, dzdy \quad (13a)$$

292 and:

$$293 \quad \vec{T}_z = \iint \boldsymbol{\sigma}' \cdot \vec{z} \, dx dy \quad (13b)$$

294 where \vec{x} and \vec{z} are the unit vectors in x and z directions respectively. We then compute the tectonic
 295 forces per linear meter as:

$$296 \quad F_x = \frac{\|\vec{T}_x\|}{L_z} \quad (14a)$$

297 and:

$$298 \quad F_z = \frac{\|\vec{T}_z\|}{L_x} \quad (14b)$$

299 where L_x and L_z are the length of the domain along x and z axis respectively.

300 **3. Dynamic of continental break up propagation**

301 Numerical models are aimed at investigating the role of both the crust rheology and the intensity of rift
 302 parallel shortening applied in our simulations. We show here that if we impose a sufficient velocity v_z
 303 (perpendicular to the extension direction) the propagation rate of the continental breakup is not
 304 instantaneous and does not exceed 1.5 cm/yr. Moreover, this is independent of the rheology of the
 305 crust. For the same amount of shortening velocity applied in the direction perpendicular to the
 306 extension, the main difference between strong and weak lower crust models resides in the propagation
 307 mode. Furthermore, for large shortening velocities, weak and strong lower crust models tend to show
 308 the same behaviour forming oblique rifts (Figures 4 and 5). In order to highlight the principal results
 309 of this study we first focus on the ridge propagation mode as a function of rift parallel inflow
 310 (obliquity).

3.1. Rheology of the crust and sensitivity to the shortening velocity

Our numerical modelling results show that the rift-parallel shortening velocity and the crust rheology strongly affect the propagation of the rift and its structure (Figures 4, 5 and 6). With a strong lower crust, the minimum shortening velocity necessary to form a strike slip conjugate and isolate a new micro-plate is 17% of the extension velocity ($v_z=1.7$ mm/yr, Figures 4e and 5e, model 3). With a weak lower crust, the rift propagation starts to be sensitive to the shortening velocity when the latter reaches 12% of the total extension velocity (Figure 4b and 5b, model 1). The influence of this third boundary condition becomes very important for 15% of the extension velocity (Figure 4d and 5d, model 2). For both lower crust rheologies (strong and weak), the shortening velocity represents a threshold, across which pure extension gives place to a transtensional and strike-slip deformation regime. A strong lower crust favours the formation of three tectonic plates joining at a Ridge-Ridge-Fault triple junction because it allows for a more localised strain than weaker crust rheologies (Figures 4e, g, h and 5e, g, h). Margins are narrow and few shear zones accommodate the crustal thinning (as shown by the beta factor varying from 1 to 5 in less than 50 km, Supplementary Figure 1). In contrast, a weak lower crust rheology favours a more distributed deformation (Figure 4b, d, f and 5b, d, f). Wide rifts are formed and many shear zones develop to accommodate the crustal thinning (beta factor of ~ 3 along 300 km from the exhumed mantle, Supplementary Figure 1). This distribution prevents the formation of three distinct plates and the effects of the out-of-plane boundary condition are visible in the rift evolution and propagation. Finally, for both weak and strong lower crusts a too large amount of shortening (20% for the strong case, Figures 4h and 5h and 17% for the weak case, Figures 4f and 5f) results in the formation of an oblique continental rift followed by oblique spreading.

3.2. Ridge propagation

The numerical models highlight three types of rifts propagation modes (Figure 4 and Figure 6a). The first mode is a very fast one (from 3 cm/yr to 1.5 cm/yr, Figure 6b) that results in an oceanic opening almost simultaneous along the rift axis. Models with a strong lower crust and a shortening velocity less than 17% of the total extension velocity (1.7 mm/yr) display this evolution (Figure 4 a, c and Figure 6).

The second propagation mode is characterized by a succession of fast propagation and stalling phases (Figure 6). This mode is principally visible in the weak lower crust models for shortening velocities less than 1.7 mm/yr (Figure 4b, d). Due to the weak lower crust, a long phase of continental rifting occurs (~ 40 Myr) before ridge propagation takes place (Figure 6d). The propagation velocity starts at about 1.5 cm/yr before slowing down in ~ 10 Myr to less than 5 mm/yr (Figure 6b). This propagation velocity drop can be related to crustal thickness variations and competition between extension velocity and shortening velocity. Indeed, the initial continental rift phase efficiently thins the crust along 200 km to 300 km (in z direction) where the velocity field is principally perpendicular to the rift. Therefore, once crustal breakup occurs, the propagation along this already thinned crust is fast. However, once the rift tip reaches a less thinned crust the propagation stops or slows down to allow the continental crust domain in front of the tip to thin enough before resuming the propagation (Supplementary Figures 4, 7 and 9). During the whole simulations time (~ 120 Myr) models were able to produce two or three stalling phases, each one lasting less than the previous one. This feature can be explained by the decreasing crustal thickness with time in front of the rift propagator.

Finally, a third propagation mode reveals a “jump” of the spreading centre from the tip of the stalling rift to a new basin. Models with strong and weak lower crust both present this propagation mode for shortening velocities larger than 1.7 mm/yr (Figure 4e, g, h and Figure 6a). As in the two previous propagation modes, the rift starts with a fast propagation (~ 1.5 cm/yr, Figure 6b) before stalling. But instead of propagating again from where it has stopped, a new spreading centre forms 200

357 km to 300 km farther (Figure 4e and Figure 6a). The main characteristic of this propagation mode
358 resides in the deformation regime between the onset of the stalling phase and the formation of a new
359 spreading centre. As soon as the along strike oceanic propagation stops, the intra-continental
360 deformation at the front of the ridge switches from a dip-slip dominated regime to a strike-slip
361 dominated. During this deformation regime, the rifting continues but localizes on an offset branch and
362 leads to the formation of two offset oceanic segment and an ultra oblique margin (Figure 4e, g, h).

363 *3.3. Tectonic forces evolution*

364 Figure 6c shows the tectonic forces F_x (equation 14a) and F_z (equation 14b) evolution along
365 boundaries of normal x and z respectively. Along the faces of normal x the F_x force corresponds to
366 the extensional forces. Strong lower crust models display initial extensional forces (from 15 TN/m to
367 17 TN/m) greater than weak lower crust ones (from 10 TN/m to 12 TN/m) at the beginning of the
368 experiments and through time. The F_x force display a small dependency to the rift parallel shortening
369 magnitude at the beginning of the simulations. However, since the rift parallel shortening influences
370 the strain localization, the force evolution differs between models. At first order, the F_x force
371 decreases with the lithosphere thinning through time for all models. The force decrease is directly
372 related to the thinning and propagation rates of the models: the faster the lithosphere thins, the faster
373 the force decreases. Then, once the continental break-up occurred along the whole length of the
374 modelled domain the force reaches or tend to approach a plateau representing the strength of the
375 lithosphere.

376 Along the face of normal z a first order observation is that the F_z force is not dependent to the
377 rift parallel shortening velocity but to the strength of the lithosphere. Indeed, the models with a strong
378 lower crust display an initial state between 4 TN/m and 5 TN/m while the models with a weak lower
379 crust show an initial state between at ~ 3 TN/m. During the deformation of the lithosphere, the F_z force
380 shows few variations related to strain localization events in the models except for models with a strong
381 lower crust and a rift parallel shortening of 1.2 mm/a to 1.5 mm/a. These two models display a fast
382 force increase associated with the fast ridge propagation and a plateau between 8 TN/m and 9 TN/m
383 when the ridge had propagated all along the domain. In absolute values, the F_z force is 2 to 5 times
384 lower than the extensional force F_x .

385 **4. Characteristic structures associated to propagation of break-up**

386 In order to decipher the geological record of the different propagation modes and at the same time to
387 gain insights on the transition from straight mode of propagation to partitioned mode we have chosen
388 to details the evolution of models with rates of shortening 1.2 mm/a (model 1), 1.5 mm/a (model 2)
389 and 1.7 mm/a (model 3).

390 *4.1. Weak Lower crust, 12% of shortening velocity*

391 The first model (model 1) is characterized by a weak lower crust and a shortening velocity
392 $v_z=1.2$ mm/yr (12% of the total extension velocity). Continental rifting lasts 40 Myr before continental
393 breakup occurs at the southern boundary of the model (Figure 7a, and Figure 8a). During this initial
394 phase, normal faulting is perpendicular to the extension direction. The weak lower crust results in
395 strain distribution over a wide continental rift (Figure 7a, b, cross sections 2 and 3 and Figure 8a, b).
396 At 50 Myr, the continental breakup reaches $z=200$ km and a localized strain band replaces the wide
397 strain distribution in the exhuming mantle between $z=0$ and $z=200$ km (Figures 7b, cross section 3 and
398 Figure 8b), while in the rest of the model continental rifting is still active (Sections 1 and 2 of Figure
399 7b). At 60 Myr, the domain of exhumed mantle reaches $z=300$ km and the oceanic rift enters a second
400 phase characterized by short episodes of propagation and stalling take place. Propagation first slows

401 down for 10 Myr while the stretching of the continental crust increases in a narrow domain and finally
402 localized as crustal breakup occurs, allowing oceanic rift propagation to continue (Figure 7c, d).
403 During this phase, a minor strike-slip crustal deformation initiates at the edge of the propagator tip
404 (coordinates: $300 < z < 400$; $700 < x < 800$, Figure 7d, cross section 2 and Figure 8d) that accommodates
405 the differential movement between the stalling ridge and ongoing crustal extension away from the
406 ridge. Then, from 80 Myr to 100 Myr the ridge propagation resumes and as the mantle is exhumed, the
407 active deformation in the crust progressively vanishes (Figure 7e-g, Figure 8e-g). The age map of
408 mantle exhumation displays a typical pattern where the oldest ages are decreasing in the direction of
409 propagation (Figure 7g). At the end of the model, the angle between the oceanic rift and the
410 continental margins varies along strike. This angle variation developed during the slowing down of the
411 propagation phase. The direction of normal shear zones changes in time and along the rift axis. The
412 first faults formed during the continental rift phase have been rotated during the first stalling phase
413 (between 50 Myr and 60 Myr, Figure 7c, d). Then, as the rift propagates the shear zones formed in the
414 crust continue to rotate and align with the rifted margins.

415 *4.2. Weak Lower crust, 15% of shortening velocity*

416 The second model (model 2) also presents a weak lower crust but with a velocity $v_z=1.5$
417 mm/yr (15% of the total extension velocity). Here again continental rifting takes place during
418 approximately 40 Myr before reaching the crustal breakup and mantle exhumation (Figure 8a and
419 Supplementary Figures 5a and 6a). Like the previous model a wide rift develops but deformation
420 appears to be more localized. At 50 Myr, the ridge is propagating northward (up to $z=200$) and normal
421 faulting is reduced (Figure 9b and Figure 10b). At this stage, the propagation starts to slow down and
422 the ridge propagation enters a stalling phase until 80 Myr. During these 30 Myr, the exhumed mantle
423 domain widens, forming stripes of exhumation ages parallel to the direction of extension but
424 perpendicular to the continental margins located at the edge of the propagator (Figure 9b, c, d). This
425 stalling phase results in drastic modification of the rift architecture. The rifted margins located around
426 the tip and normal faults rotate until they form an angle of 90° with the active ridge (Figure 9c, d, e
427 and Figure 10c, d, e). As a result, the rifted margins located at the propagator tip are striking parallel to
428 the extension direction. This rotation is accompanied with the development of a left-lateral strike-slip
429 fault zone in the crust (Figures 9c, d, e and Figure 10c, d, e). Transcurrent deformation first develops
430 at around 50 Myrs as ridge propagation stops (Figure 10b) and evolves as a highly localized crustal-
431 scale strike-slip shear zone (Figure 10e). The rotated normal faults of the continental rift reach an
432 obliquity of $\sim 45^\circ$ with respect to the stretching direction and ultimately form releasing bends aligned
433 in the strike-slip fault direction. During this phase, the continental crust is thinned under a
434 transtensional strain regime, exhuming the lower crust in elongated dome structures parallel to
435 stretching direction as described on the field (e.g. Jolivet et al., 2004) and reproduced previous
436 modelling studies (e.g. Le Pourhiet et al., 2012, 2014). Around 80 Myr, the propagation resumes and
437 the exhumed mantle domain opens obliquely to the stretching direction along the former strike-slip
438 zone (Figures 9e and Figure 10e). At 90 Myr, the ridge propagation slows down once again and a
439 second left-lateral strike-slip structure develops (Figure 9f and Figure 10f). The most remarkable result
440 is that the rift orientation change associated with the development of strike-slip deformation occurs
441 spontaneously without involving changes in plate kinematics (Figure 9 and 10).

442 *4.3. Strong Lower crust, 17% of shortening velocity*

443 The third model (model 3) involves a strong lower crust and a velocity $v_z=1.7$ mm/yr (17% of the
444 total extension velocity). In this model, the continental rifting phase stands for less than 30 Myr before
445 crustal breakup occurs (Figure 6d). Crustal deformation is also more localized than in the previous
446 models (Figure 11 and 12). At 30 Myr, the ridge reaches $z=200$ km (Figures 11a and Figure 12a) and

447 its propagation slows down. This propagation rate change coincides with the formation of two new
448 deformation zones (Figure 12a): a right-lateral strike-slip structure to the left and a left-lateral
449 transtensional shear zone to the right. At 40 Myr, the ridge propagation enters in a stalling phase and
450 stops (Figure 11b). The two branches localize now all the intra-continental deformation at the front of
451 the ridge and a ridge-rift-fault triple junction which rapidly evolves in ridge-ridge-fault forms at the tip
452 of the propagator (Figure 12b). The strike-slip motion in both branches intensifies and becomes
453 predominant (Figure 12b, c) until a new exhumed mantle domain appears at 60 Myr ($800 < x < 900$;
454 $400 < z < 500$, Figures 11d, cross-section 1). Between the initial propagator and this new basin, a thin
455 band of continental crust remains before the two exhumed mantle domains join at 70 Myr (Figures
456 11e, cross section 2). As soon as the new exhumed mantle domain forms, the deformation regime on
457 the right branch changes once again from a strike-slip regime to transtensional extension (Figure 12e).
458 As in the second model (weak lower crust, $v_z=1.5$ mm/yr), the strike-slip deformation in the right
459 branch accommodates the rotation of the continental margin previously located at the edge of the
460 initial propagator. The rifted margin not only rotates but also thins during the transtensional
461 deformation phase, before being torn apart when both exhumed mantle domains join (Figure 11e,
462 cross-section 2). The finite rotation results in a rifted margin oriented parallel to the extension
463 direction, while it was formed perpendicularly to it. This process may be responsible for the initiation
464 of a transform margin. Here again, the lower crust exhumes preferentially in areas where strike-slip
465 deformation is dominant. In the left branch, the right-lateral strike-slip deformation is active from 30
466 Myr to the end of the simulation (Figure 11 and Figure 12). This localized vertical shear zone
467 progressively offsets the former rifted margin structures (Figure 11, cross section 3) and forms a
468 vertical margin where the transition from continent to exhumed mantle is sharp (Figures 11e, f, cross
469 section 4). Compared with the previous models, the lithosphere-scale strain partitioning is a main
470 characteristic of model 3 and results in the formation of a triple junction and three distinct tectonic
471 plates. As in the model 2, the change in deformation regime and rift orientation is provoked by the
472 stalling phase while kinematic boundary conditions are kept constant.

473 **5. Discussion**

474 *5.1. Geodynamic interpretation of rift parallel inflow*

475 Lithospheric modelling aims at understanding the dynamics that lead to the emergence of new
476 localised plate boundaries at regional scale. These plate boundaries are of 3 types (Fault, Ridge,
477 Trench), which can be studied separately. But plates tectonics usually deal with triple junctions and
478 the velocity field surrounding triple junction is not cylindrical in map, it is rather a combination of
479 pure and simple shear components. The geodynamic pertinence of non cylindrical boundary
480 conditions makes therefore no doubts if one wants to study the emergence of triple junction.

481 In term of modelling, Le Pourhiet et al. (2014) have shown that transtensional tectonic
482 structures (A type domes) emerging from pure shear boundary conditions are similar to those
483 produced by imposing simple shear boundary conditions. Yet, pure shear boundary conditions have
484 the advantage over simple shear boundary conditions because they allow the formation of large strike
485 slip structures without having to impose a step function in the boundary conditions of the models, that
486 is, without prescribing the location of the strike slip faults nor their occurrence at all. They are
487 therefore useful to study the emergence or not of localised strike slip plate boundary rather than
488 oblique rift.

489 Finally, as for all models run with kinematic boundary conditions, one need to discuss the
490 forces needed to deform the model. Our results and especially the force calculation show some
491 unexpected results which need to be discuss.

492 For 3D deformation patterns to emerge within cylindrical boundary conditions one must
493 include a form of inheritance in the initial conditions. Models having large area with random noise
494 (Duclaux et al., 2020; Naliboff et al., 2020) demonstrates that the offset of oblique structures in
495 cylindrical conditions is not very large without strong kinematic forcing (Ammann et al., 2017; Brune
496 et al., 2012; Allken et al., 2012; Gerya, 2013; Liao & Gerya, 2015; Le Pourhiet et al., 2017)
497 assuming that regional deformation is cylindrical. But also through boundary conditions such as
498 oblique extension (Brune, 2014; Brune et al., 2012; Brune & Autin, 2013), pure shear (e.g. Gerya &
499 Burov, 2018; Le Pourhiet et al., 2014) or combination between pure and simple shear (Le Pourhiet et
500 al., 2018). Kinematic boundary conditions involving pure shear showed the ability to reproduce
501 formation of more than two tectonic plates and triple junctions (Gerya & Burov, 2018).

502 In this study we used Le Pourhiet et al. (2018) kinematic boundary conditions to impose
503 obliquity. As for pure shear boundary conditions we showed that accounting for sufficient rift parallel
504 shortening leads to the formation of a triple junction between three tectonic plates in models with
505 strong lower crust. The coupled rheology of the crust seems to be a first order constraint to produce
506 large scale strike-slip fault in extensional contexts as also shown in analogue models (Zwaan et al.,
507 2019).

508 In terms of tectonic forces, the models show that the forces are more dependants to the
509 rheology of the lithosphere than to the rift parallel shortening magnitude. In a weak lower crust
510 lithosphere, the force associated to the rift parallel shortening (F_z) is in the same order of magnitude
511 than the force generated by gravitational potential energy (Bird et al., 2006). For a strong lower crust
512 lithosphere, the gravitational potential energy may not be sufficient to explain such rift parallel
513 shortening and an active tectonic process like a subduction or a collision may be required in the
514 vicinity of the rift to allow rift parallel shortening. However, the force generated by this rift parallel
515 shortening is ten times lower than the extensional force and likely represents the resistance of the
516 strike-slip fault developing in strong lower crust models which is several hundreds of kilometres long.
517 Such a force may be lowered by a more efficient strain softening in the brittle crust.

518 *5.2. Ridge migration and comparison with previous modelling work*

519 Models presented in this study show that a small amount (12% to 20%) of rift parallel shortening
520 leads to alternation of ridge propagation phases and ridge stalling phases. These alternating phases are
521 each associated with strain regime changes. During ridge propagation, the deformation in the
522 continental crust located at the front of the propagator is dip-slip dominated, while during stalling
523 phases the deformation in the continental crust is strike-slip dominated.

524 According to our models results, the ridge propagation occurs under three different mechanisms
525 (Figure 6 and Figure 14). For low rift-parallel shortening velocities (12% to 15% of the extension rate,
526 Figure 4a, b, c) the ridge propagation is fast (between 1.5 cm/yr and 3.5 cm/yr, Figure 6b) and straight.
527 Therefore, in this case the ridge propagation velocity exceeds the total extension rate (1 cm/yr). This
528 propagation mode has been documented in crustal scale models involving a low viscosity layer
529 beneath a plastic layer (Allken et al., 2012) where the lower viscosities lead to faster propagation. The
530 same behaviour is observed in lithospheric scale models with cylindrical boundary conditions (Le
531 Pourhiet et al., 2018) where free slip boundary conditions prevent deformation in the rift parallel

532 direction (which is equivalent to a boundary opposing no resistance to propagation) but with the
533 mantle acting as the viscous layer controlling the propagation rate.

534 Higher rift-parallel shortening velocities (15% to 17%, Figure 4d) lead to ridge propagation by
535 pulses alternating between fast propagation (between 0.5 cm/yr and 1.5 cm/yr, Figure 6b) and stalling
536 phases. This propagation mode obviously requires that the ridge propagation stalls at some point. Until
537 now, this behaviour has been modelled only during soft linkage of en-échelon propagators (Ammann
538 et al., 2017; Liao & Gerya, 2015; Le Pourhiet et al., 2017) and a rift parallel inflow boundary
539 condition model (Le Pourhiet et al., 2018, and this study). The common feature between these models
540 is that during the stalling phase the velocity field at the tip of propagator changes and leads to a strike-
541 slip dominated deformation. This mode of propagation was also proposed by Courtillot (1982) in the
542 Gulf of Aden to explain the obliquity between seafloor magnetic anomalies and the passive margin. In
543 Courtillot's (1982) model, a hard node in the lithosphere structure would explain why the ridge
544 propagation stopped. All these thermo-mechanical and conceptual models show that to enter in a
545 stalling ridge phase, a non-cylindrical parameter is required. It can be rheological (Courtillot, 1982) or
546 kinematic (Ammann et al., 2017; Liao & Gerya, 2015; Le Pourhiet et al., 2017, 2018).

547 Finally, for higher velocities (17% to 20%, Figure 4e, f, g) a ridge jump propagation mode occurs.
548 This ridge propagation mode also requires a ridge stalling phase, but rather than resuming from the
549 ridge tip, a new spreading centre localizes between 100 and 200 km away from the ridge (Figure 6a
550 and Figure 11d). This propagation mode is particularly well observed in models evolving towards
551 three distinct tectonic plates (Figure 4e, g, h), therefore for highly oblique kinematic conditions. A
552 propagation by jump was also obtained in analogue models with rotational extension coupled with an
553 oblique weak seed (Molnar et al., 2017) where only highly oblique weak zones lead to jumps of
554 spreading centres. Effectively, a weak zone oblique to the ridge propagation direction has the same
555 effect as a strong zone or an inflow in the rift parallel direction. In the first case it will guide the
556 deformation, while in the second case it will force the deformation to propagate elsewhere.

557 These different propagation modes are highly dependent of the lithosphere ability to deform in the
558 rift parallel direction, whether it is due to rheological heterogeneities (e.g. Courtillot, 1982; Molnar et
559 al., 2017) or velocity field variations (e.g. Ammann et al., 2017; Le Pourhiet et al., 2017, 2018).

560 *5.3. Comparison with natural cases*

561 Although the four oceanic basins presented in Figure 1 are not at same time and space scales, they
562 all share similar features between each other and with our models. All of them are well documented
563 examples of ridge propagation characterized by pulses of alternating fast propagation along straight
564 margins and stalling phases which often corresponds to the location of oblique and transform margins.
565 Our numerical simulations show that during propagation phases the deformation occurs orthogonal to
566 the rift axis whereas oblique extension and strike-slip deformation characterises stalling phases.
567 Initiation of strike-slip deformation results in the reorientation of pre-existing rift normal faults. As
568 stretching continues an oblique continental rift forms which may result in the formation of a transform
569 margin or oblique continental break-up propagation (Figure 14). A first order observation is that as in
570 the natural systems described in this study, none of the models involving rift parallel shortening
571 produce shortening structures (as thrusts or folds). However, the non-cylindrical component of the
572 velocity field is well expressed in the GPS velocities of active oblique rift systems (Gulf of Aden and
573 Northern Red Sea basins).

574 Among the four oceanic basins in Figure 1, the South China Sea is probably the most compelling
575 example of secondary oblique continental break-up propagation following a stalling phase. Figure 1b

576 outlines three V-shaped propagators which indicate that continental break-up propagated from east to
577 west by pulses of rapid propagation, separated by two periods of stalling. The longest period occurred
578 from 32 to 23 Ma where magnetic anomalies in the propagator region form at a 60° angle with the
579 continental margin indicating that the propagation rate is much lower than the spreading rate. Le
580 Pourhiet et al. (2018) proposed that rift parallel inflow could explain this first phase of stalling using a
581 model similar to the one with a weak crust and 1.2 mm/yr rift parallel inflow. They then posit that this
582 long stalling period ends when both the spreading and propagating direction of the mid-oceanic ridge
583 changed by 15° (Sibuet et al., 2016) in the South China Sea (Figure 1b). The origin of this rift parallel
584 inflow has been proposed as the expression of gravitational potential energy gradient (e.g. Ding & Li,
585 2016; Le Pourhiet et al., 2018) between the developing basin and the higher Indochina to the west.
586 Bird et al. (2006) showed that a topographic load can propagate vertical forces of 1 TN/m to 2 TN/m
587 over hundreds of kilometres. Here, with a more thorough exploration of the effect of shortening on the
588 propagation of continental break up, we show that increasing the shortening rate to 1.5 mm/yr
589 reproduces the 15° rotation of the spreading direction reported by Sibuet et al. (2016) and the
590 propagation of a younger oceanic propagator self consistently. Moreover, associated to the strike-slip
591 deformation, lower crust domes exhume in the continental margin as in the South China Sea. This
592 finding questions whether magnetic anomalies do record a change in far field conditions or simply the
593 localization of the deformation on an oblique continental rift branch.

594 This result may have significant implications on plates reconstruction and geodynamic
595 interpretations. Indeed, changes of structural directions are common during continental rifting. These
596 phases coincide with a distribution of the deformation in the crust in front of the oceanic propagator as
597 in the Atlantic realm (Enachescu, 2006; Mohn et al., 2015; Nirrengarten et al., 2018; Sibuet et al.,
598 2007), in the Gulf of Aden (Fournier et al., 2010; Leroy et al., 2012) or in the northern Red Sea
599 (Almalki et al., 2016; Bartov et al., 1980; Bosworth, 2015; 2020) where major transform fault zones
600 are segmenting the oceanic basins at locations where stalling phases occurred.

601 Previous studies proposed that these stalling phases can be related to several processes acting as
602 an opposing force to the rift propagation in the undeformed continental lithosphere like the presence of
603 high strength continental domains or inherited structures in the lithosphere (Courtilot, 1982; Morgan
604 & Parmentier, 1985; Nirrengarten et al., 2018). As a result they are generally interpreted as indicating
605 varying kinematic conditions or inheritance. This hypothesis was already questioned by numerical
606 model of rift linkage at oblique continental margin which showed that with increasing strain
607 localisation the tectonic regime changes from initially dip-slip dominated to strike-slip dominated (Le
608 Pourhiet et al., 2017). Here, we show that strain partitioning and change in deformation regime can
609 occur at the tip of oceanic propagators under constant boundary kinematic conditions (Figures 9 and
610 11). This implies that a relationship seems to exist between major transform margins and the rift
611 stalling phase during ocean opening. In the North Atlantic Ocean we observe that the Newfoundland-
612 Azores-Gibraltar fault zone and the Charlie-Gibbs fault are the two major transform faults separate
613 segment of different spreading ages (Nirrengarten et al., 2018 and references therein). In the same
614 way, the Gulf of Aden (Figure 1a) opened almost synchronously between the Alula-Fartak fault zone
615 and the Shukra el Sheik fault zone but was stalling along the Shukra el Sheik fault zone for 9 Ma to 12
616 Ma before propagation restarted (Fournier et al., 2010; Leroy et al., 2012). However, in our models we
617 were not able to produce oceanic transform faults segmenting the ridge. Thermo-mechanical models
618 tend to show that magmatism and serpentinization are essential to produce oceanic transform faults
619 (e.g. Ammann et al., 2017; Gerya, 2012).

620 The South China Sea and the Gulf of Aden are both good examples to illustrate the rifting of a
621 weak continental lithosphere (Figure 4b, d and Figures 7 and 9) due to the geodynamic context in

622 which those two oceanic basins have formed. The Gulf of Aden is located near an important mantle
623 thermal anomaly (the Afar hot spot, *e.g.* Fournier et al., 2010; Leroy et al., 2012) while the South
624 China sea opening followed a long-term subduction associated with magmatism (Pubellier et al., 2003;
625 Wang et al., 2013). The Gulf of Aden presents the advantage of being an active plate boundary
626 monitored with GPS stations. The GPS velocities with respect to Arabia fixed (ArRajehi et al., 2010)
627 display the Eastward motion (*i.e.* parallel to the ridge) of the continent at the tip of the oceanic
628 propagator and an oblique North-eastward motion for Africa fixed. The oceanic basin and continental
629 margins display stalling phases and propagation phases and the rifting phase is characterized as a wide
630 oblique rift. Figure 13c shows the relative velocity field in the Model 1 (1.2 mm/yr rift parallel inflow)
631 which display the oblique velocity at the tip of the propagator. In the Gulf of Aden, the force that
632 triggers the rift parallel inflow as displayed by GPS velocities could come from the differential
633 gravitational potential energy between the high topography above the Afar hot spot and the opening
634 Gulf of Aden.

635 However, the strain localization and the evolution of the deformation in a stronger lithosphere
636 (Figure 4e, g, h and Figure 11) results in the development of a localized strike-slip fault and the
637 formation of a micro-plate. These models display similarities with the Northern Red Sea and its two
638 branches, the Gulf of Aqaba to the East and the Gulf of Suez to the West (Almalki et al., 2016; Bartov
639 et al., 1980; Bosworth, 2015; 2020). In the Northern Red Sea, the Sinai block behaves as a third
640 tectonic micro-plate separated from Arabia by the Levant transform Fault and from Africa by an active
641 oblique rift (with respect to the Red Sea). Although a continental rift is not a complete plate boundary
642 it is the initiation. The GPS velocities with respect to Arabia display the southward motion of the Sinai
643 block along the Levant Fault. Therefore, regarding the absence of shortening structures in the northern
644 Red Sea, the GPS velocities showing the rift parallel displacement and the presence of the transform
645 continental fault of the Levant, we propose that a rotational extension model (Molnar et al., 2017;
646 Smith, 1993; Zwaan et al., 2020) may not be the best candidate to explain the dynamics of the
647 Northern Red Sea. However, the formation of the triple junction between a continental transform fault,
648 an active oblique rift and a ridge is well reproduced in our strong lower crust models when the rift
649 parallel inflow is greater than 16% of extension velocity (Figure 14b, model 3). The geodynamic
650 origin of this regional rift parallel inflow arise from the contrast between the fast motion of the
651 Arabian plate driven by subduction in Zagros and Makran and the slow northward motion of Africa in
652 collision with the European plate.

653 Moreover, the lithosphere-scale strain partitioning between a highly localized strike-slip fault and
654 a very oblique rift leads to the development of two types of transform margins with extremely
655 different kinematic, dynamics and thermal history.

656 The classical model for transform margins formation involves the development of transform faults
657 during the intra-continental rifting phase (Basile, 2015; Francheteau & Le Pichon, 1972; Mascle &
658 Blarez, 1987). The evolution from rifting to oceanic spreading should preserve the continental
659 structures and develop oceanic transform faults along the previous continental ones. This model has
660 major implications on development of the margin structure and temperature. It implies that oceanic
661 ridge migrates along the transform margin as a triple junction at half the spreading rate (for large offset
662 transform margin it may take several million years) heating the continental margin and its basins
663 (Ruepke et al. 2010). It also implies that the strike-slip faults responsible for the steep structure of the
664 transform margins (Mercier de Lépinay et al., 2016) are active at the beginning of the rifting phase and
665 that the transform margin was born with characteristics observable today on the fossil transform
666 margins. However, models of this study show different characteristic and evolution for transform
667 margins.

668 On the one hand, models with high obliquity display the typical kinematic context of a transform
669 margin (Basile, 2015; Francheteau & Le Pichon, 1972). In this condition one of the oblique
670 continental rift branch (Figure 11, cross sections 1 and 2, Figure 12) reach continental break-up and
671 the system evolves into two en-échellon ridges separated by a zone of soft linkage that evolves into an
672 oblique rift with strike-slip deformation similarly to previous studies (Ammann et al., 2017; Liao &
673 Gerya, 2015; Laetitia Le Pourhiet et al., 2017). However, structural properties of these modelled
674 margins correspond to hyper-thinned continental crust but not steep margins along a localized
675 transform fault. On the other hand, simulations with high obliquity and strong lower crust produce
676 steep margins along a pure strike-slip fault (e.g. Figure 11f, corss section 4) that display the structural
677 characteristics of a transform margin as a slope break, a marginal ridge close to the slope break and a
678 strike-slip fault separating the continental crust from the exhumed mantle (or oceanic crust)(Mercier
679 de Lépinay et al., 2016). However, these strike-slip margins form in atypical kinematic context for a
680 transform margin. From an initially hyper thinned margin developed during the initial extension phase
681 (until 30Myr), a steep margin develops (Figure 11e, f, cross section 4) as the deformation increases
682 along the strike-slip fault. This strike-slip shear zone puts in contact a 50 km thick continental
683 lithosphere with the asthenosphere. As a result, the formation these steep margins starts 10 Myr after
684 the continental break-up meaning that the margin geometry is acquired after the intra-continental
685 rifting phase. These results highlight that oblique and transform margins can acquire their final
686 structure several million years after continental break-up due to the strike-slip deformation.

687 **6. Conclusion**

688 3D thermo-mechanical numerical models integrating a rift parallel inflow show that:

- 689 - The 3D boundary conditions can be interpreted as an absolute velocity field that can be
690 converted to a relative velocity field comparable with natural cases of triple junctions
691 involving continental transform faults.
- 692 - At least three propagation modes can exist. A fast propagation resulting in a straight oceanic
693 basin (2 plates), an alternation of propagation and stalling phases (extension opposing velocity
694 of 12% to 15% for weak lower crust) and propagation with a “jump” of the spreading centre
695 (opposing extension velocity over 17% for strong and weak lower crust).
- 696 - .
- 697 - Continental break-up propagation slows down when rift parallel inflow is applied (Le Pourhiet
698 et al., 2018) and we show that during these stalling phases the continental margins and
699 tectonic structures rotate and the deformation regime changes from dip-slip dominated to
700 strike-slip dominated while boundary kinematic conditions are kept constant.
- 701 - There might be a relationship between stalling phases, transform margin formation and the
702 subsequent location of large oceanic transform fault zones due to margins rotation
- 703 - Two modes of transform margin can develop, steep margin associated with strike slip faults
704 and hyper-extended ones associated with offset spreading segment similar to Le Pourhiet
705 (2017)
- 706 - All oblique margins might not emerge from a single geodynamic process and that it could be
707 worthwhile in the future to revisit the available data using the soft linkage oblique margin and
708 the strike-slip oblique margin models as a classification

Acknowledgment

709
710 This study is part of the Orogen program, a tripartite research collaboration between CNRS, BRGM,
711 and TOTAL SA. We thank TOTAL SA for the computing time on Pangea supercomputer.
712 pTatin3D is an open source code, which is available publicly at
713 <https://bitbucket.org/ptatin/ptatin3d/src/master/>. The version of the code used in this study, an example
714 options file to reproduce the results and the models results are available publicly at
715 <http://doi.org/10.5281/zenodo.3967184>
716 Maps from figure 1 were made with GeomapApp and post-processing with Paraview.
717

Figure captions

718
719 Figure 1: Topographic and bathymetric maps (GMRT, (Ryan et al., 2009) with oceanic crust ages of
720 a) the Gulf of Aden (Fournier et al., 2010; Leroy et al., 2012), c) the South China Sea (Li et al., 2014;
721 Le Pourhiet et al., 2018), e) the Red Sea (Almalki et al., 2016; Bosworth et al., 2020), g) the North
722 Atlantic Ocean (Klitgord & Schouten, 1986). Main transform fault zones are reported with thick black
723 lines. GPS velocities are from (ArRajehi et al., 2010; Reilinger & McClusky, 2011). Graphs with red
724 curves represent the ridge propagation in b) the Gulf of Aden, d) the South China Sea, f) the Red Sea,
725 h) the North Atlantic Ocean interpreted from the oceanic seafloor ages distribution along strike.
726 Topographic and bathymetric maps made with GeoMapApp (www.geomapapp.org) / CC BY.

727 Figure 2: Kinematic models of ridge propagation and continental rift geometry. a) Fast ridge
728 propagation (Vink, 1982), b) slow ridge propagation (Vink, 1982), c) Scissor opening (Hey et al.,
729 1980). Dynamic models of the formation of oceanic V-shaped propagators and continental rifts d)
730 strong ribbon in the continental crust slowing ridge propagation (Courtillot, 1982), e) offset
731 propagators (Le Pourhiet et al., 2017), f) shortening perpendicular to extension direction (Le Pourhiet
732 et al., 2018).

733 Figure 3: a) Numerical models setup. Black arrows display the initial velocity field in the models
734 domain. b) Velocity boundary conditions to solve the conservation of momentum. c) Yield stress
735 envelopes of the models domain computed for a constant strain rate of 10^{-14} s^{-1} and an initial geotherm
736 represented by the red curve.

737 Figure 4: Map view of nine models with a strong and weak lower crust and varying v_z velocities
738 displaying the plastic strain and the mantle exhumation age. The plastic strain represents all the
739 deformations that occurred under the regime described by the equation 6. The mantle exhumation age
740 is set when the mantle cools beneath the isotherm 800°C .

741 Figure 5: Serial cross-sections of nine models with a strong and weak lower crust and varying v_z
742 velocities representing the plastic strain and the viscous strain. For every models, the cross-sections
743 are regularly spaced every 50 km from $z=150$ km to $z=500$ km. The plastic strain represents all the
744 deformations that occurred under the regime described by the equation 6. The viscous strain represents
745 all the deformations that occurred under the regime described by the equation 7.

746 Figure 6: a) Curves for every models representing the position of the tip along the z axis (in km) with
747 respect to time (in Myr). b) Curves for every models representing the velocity of the rift propagation
748 (in mm/yr) with respect to time (in Myr). The velocity of the rift propagation is computed as the time
749 derivative of the position of the tip. c) Evolution of the tectonic forces through time. F_x (left panel)
750 corresponds to the forces applied in the direction normal to the face of normal x (equation 14a) and F_z

751 corresponds to the forces applied in the direction normal to the face of normal z (equation 14b) d)
752 Diagram showing the rift evolution of every model in time for both strong and weak lower crust.

753 Figure 7: Numerical model with a weak lower crust and a v_z shortening velocity of 1.2 mm/yr (model
754 1). Left panels show a map view evolution of the plastic strain (deformation related to equation 6) in
755 the crust and exhumation ages of the mantle below the 800°C isotherm. Red dashed lines shows the
756 locations of cross-sections displayed in the right panels. On the cross-sections, background colours
757 correspond to the simulated lithologies presented in Figure 3. The deformation drawing is interpreted
758 from Supplementary Figures 2 and 3.

759 Figure 8: Numerical model with a weak lower crust and a v_z shortening velocity of 1.2 mm/yr (model
760 1). The background colours represent the lithologies detailed in Figure 3. The superimposed colours
761 represent the stress inferred deformation regime (Brune, 2014; Brune & Autin, 2013). The intensity of
762 the colourscale is related to the intensity of the strain rate tensor second invariant.

763 Figure 9: Numerical model with a weak lower crust and a v_z shortening velocity of 1.5 mm/yr (model
764 2). Left panels show a map view evolution of the plastic strain (deformation related to equation 6) in
765 the crust and exhumation ages of the mantle below the 800°C isotherm. Red dashed lines shows the
766 locations of cross-sections displayed in the right panels. On the cross-sections, background colours
767 correspond to the simulated lithologies presented in Figure 3. The deformation drawing is interpreted
768 from Supplementary Figures 5 and 6.

769 Figure 10: Numerical model with a weak lower crust and a v_z shortening velocity of 1.5 mm/yr (model
770 2). The background colours represent the lithologies detailed in Figure 3. The superimposed colours
771 represent the stress inferred deformation regime (Brune, 2014; Brune & Autin, 2013). The intensity of
772 the colourscale is related to the intensity of the strain rate tensor second invariant.

773 Figure 11: Numerical model with a strong lower crust and a v_z shortening velocity of 1.7 mm/yr
774 (model 3). Left panels show a map view evolution of the plastic strain (deformation related to equation
775 6) in the crust and exhumation ages of the mantle below the 800°C isotherm. Red dashed lines shows
776 the locations of cross-sections displayed in the right panels. On the cross-sections, background colours
777 correspond to the simulated lithologies presented in Figure 3. The deformation drawing is interpreted
778 from Supplementary Figures 8 and 9.

779 Figure 12: Numerical model with a strong lower crust and a v_z shortening velocity of 1.7 mm/yr
780 (model 3). The background colours represent the lithologies detailed in Figure 3. The superimposed
781 colours represent the stress inferred deformation regime (Brune, 2014; Brune & Autin, 2013). The
782 intensity of the colourscale is related to the intensity of the strain rate tensor second invariant.

783 Figure 13: Map view of the horizontal velocity field extracted from model 3 ($v_z = 1.7$ mm/a, strong
784 lower crust) at 50 Myr. a) Absolute horizontal velocity field. b) Relative horizontal velocity field
785 (plate A fixed). c) Relative horizontal velocity field (plate C fixed). d) Relative horizontal velocity
786 field (plate B fixed)

787 Figure 14: Synthetic sketch representing the three propagation modes highlighted by models of this
788 study. Propagation modes are related to the degree of obliquity and strain partitioning. Relative
789 velocities vector directions are extracted from the model with a strong lower crust and a shortening
790 velocity $v_z=1.7$ mm/yr.

791

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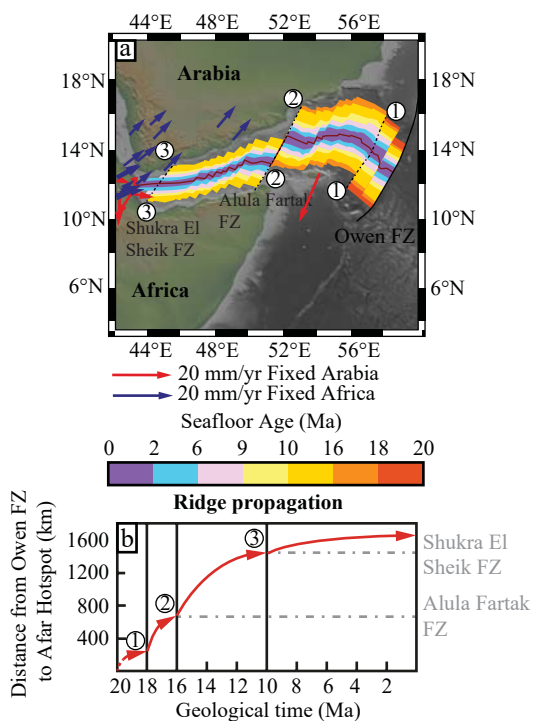
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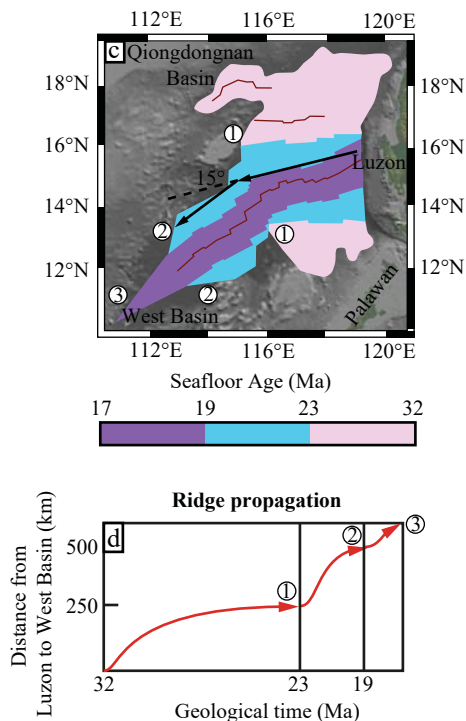
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Figure 1.

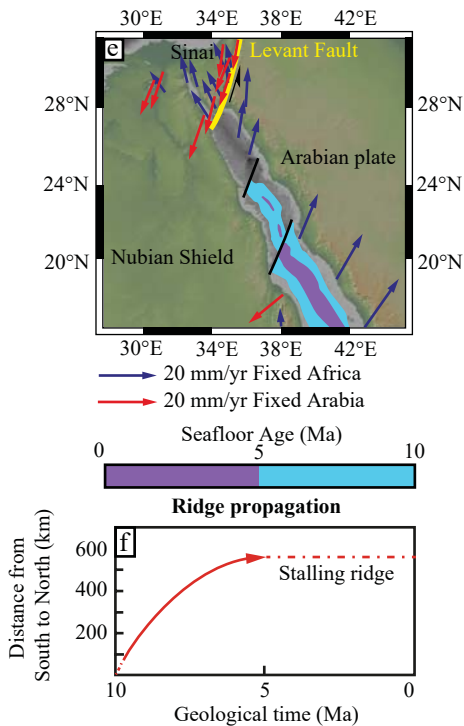
Gulf of Aden



South China Sea



Red Sea



North Atlantic Ocean

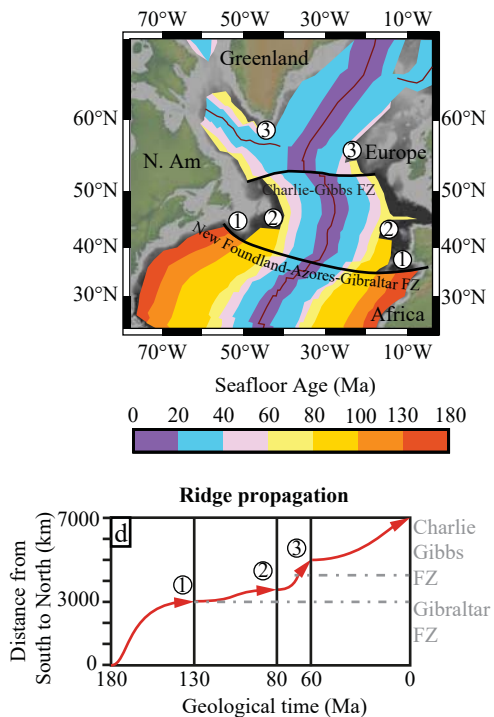


Figure 2.

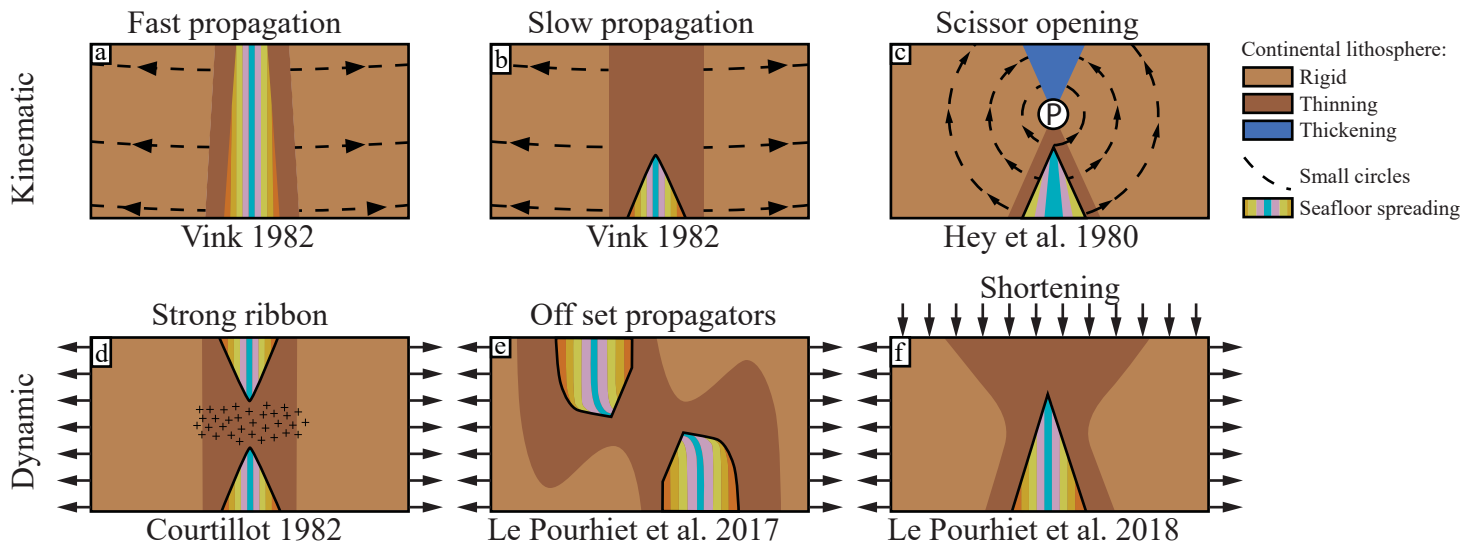


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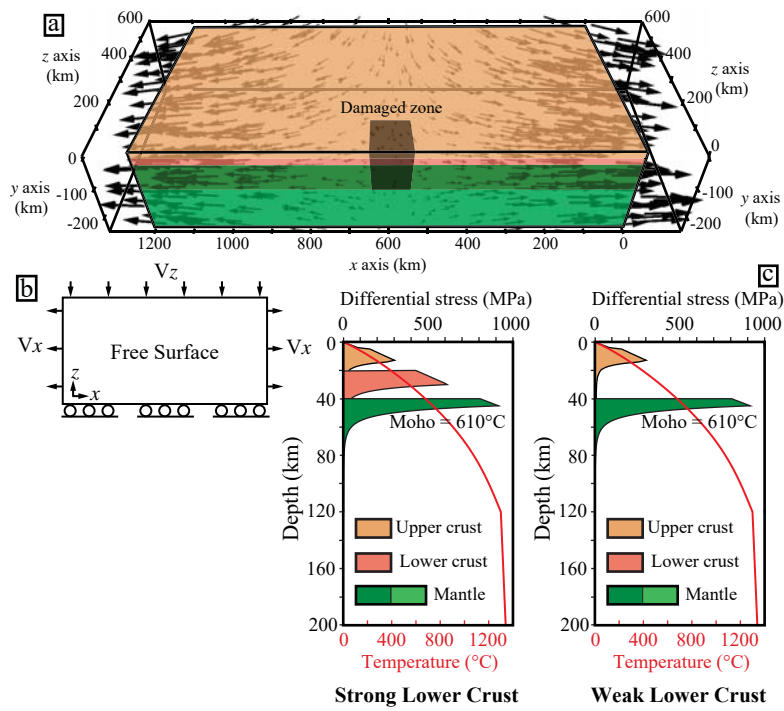


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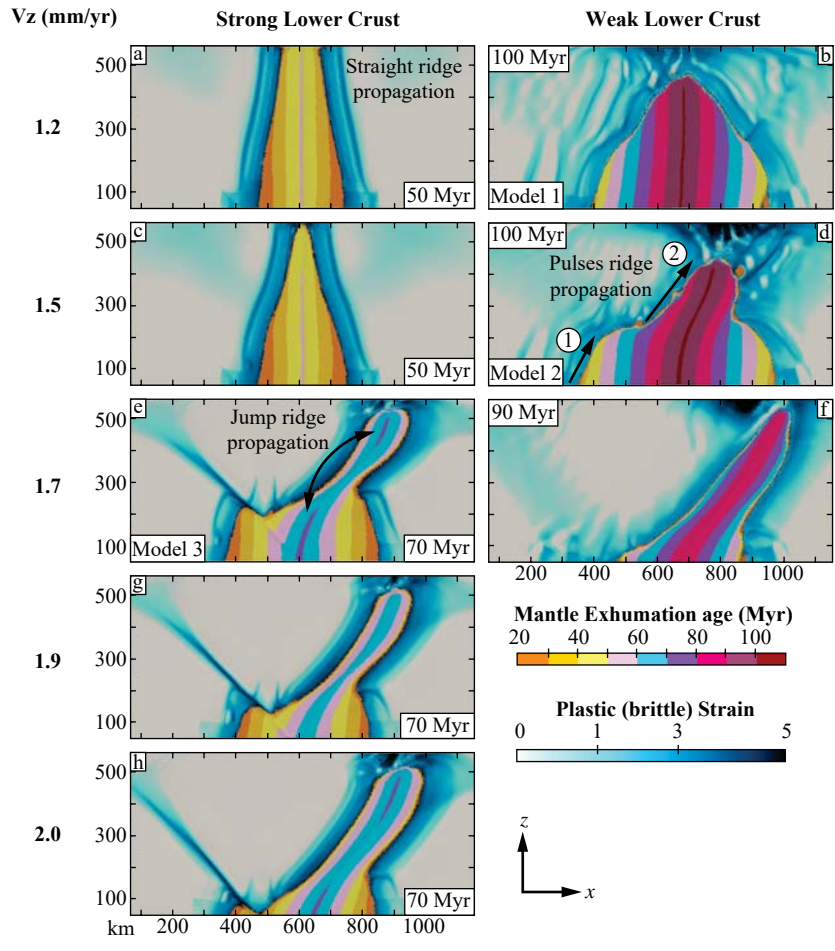


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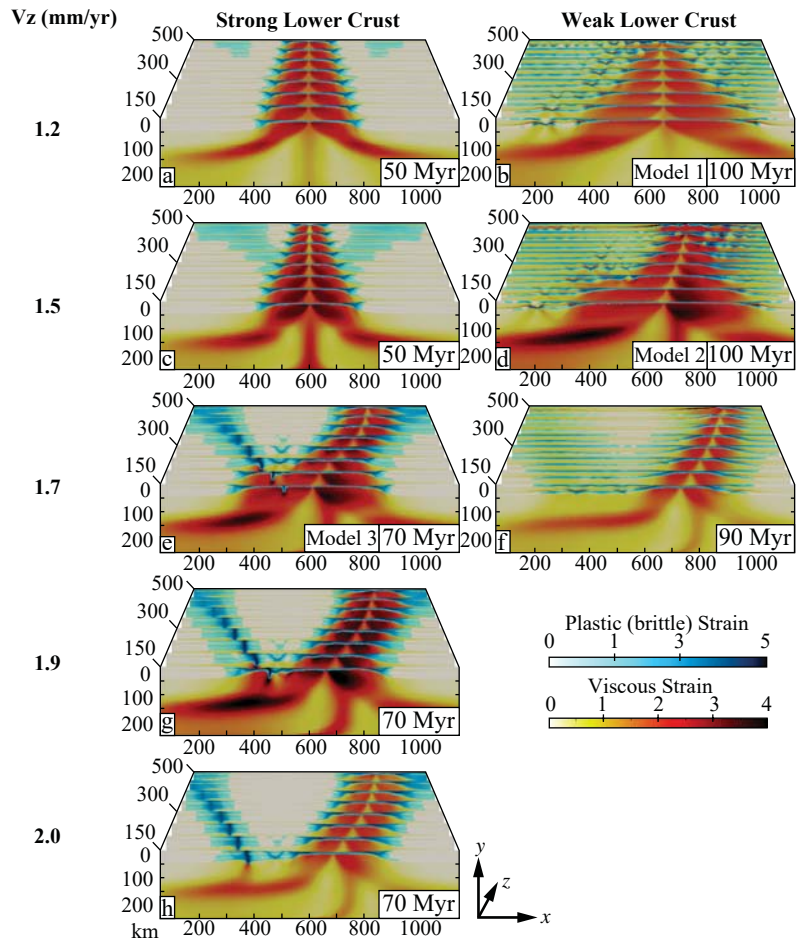


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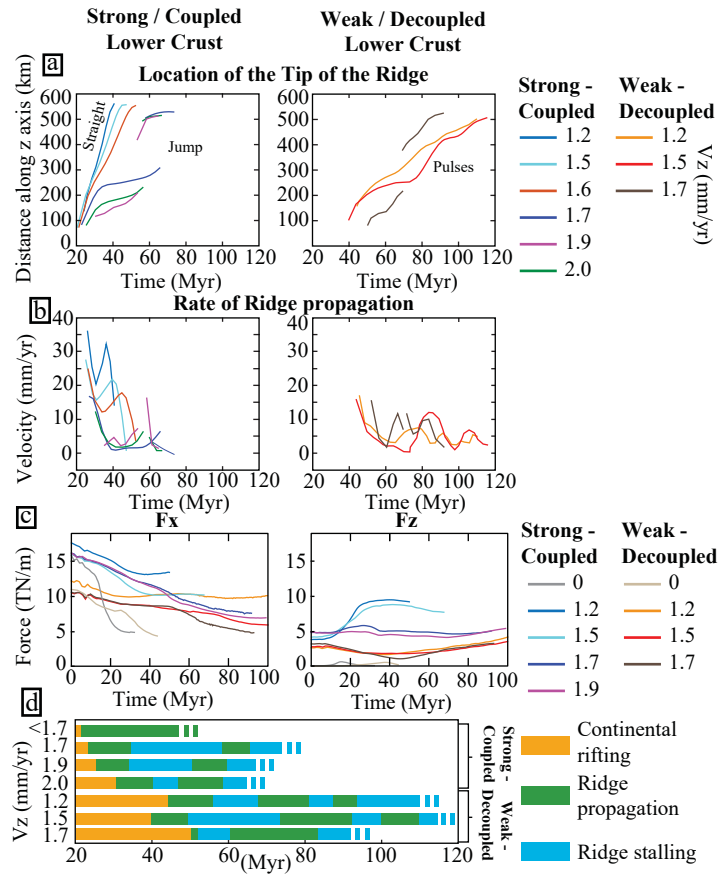


Figure 7.

Weak Lower Crust
 $V_z = 1.2 \text{ mm/yr}$ (model 1)

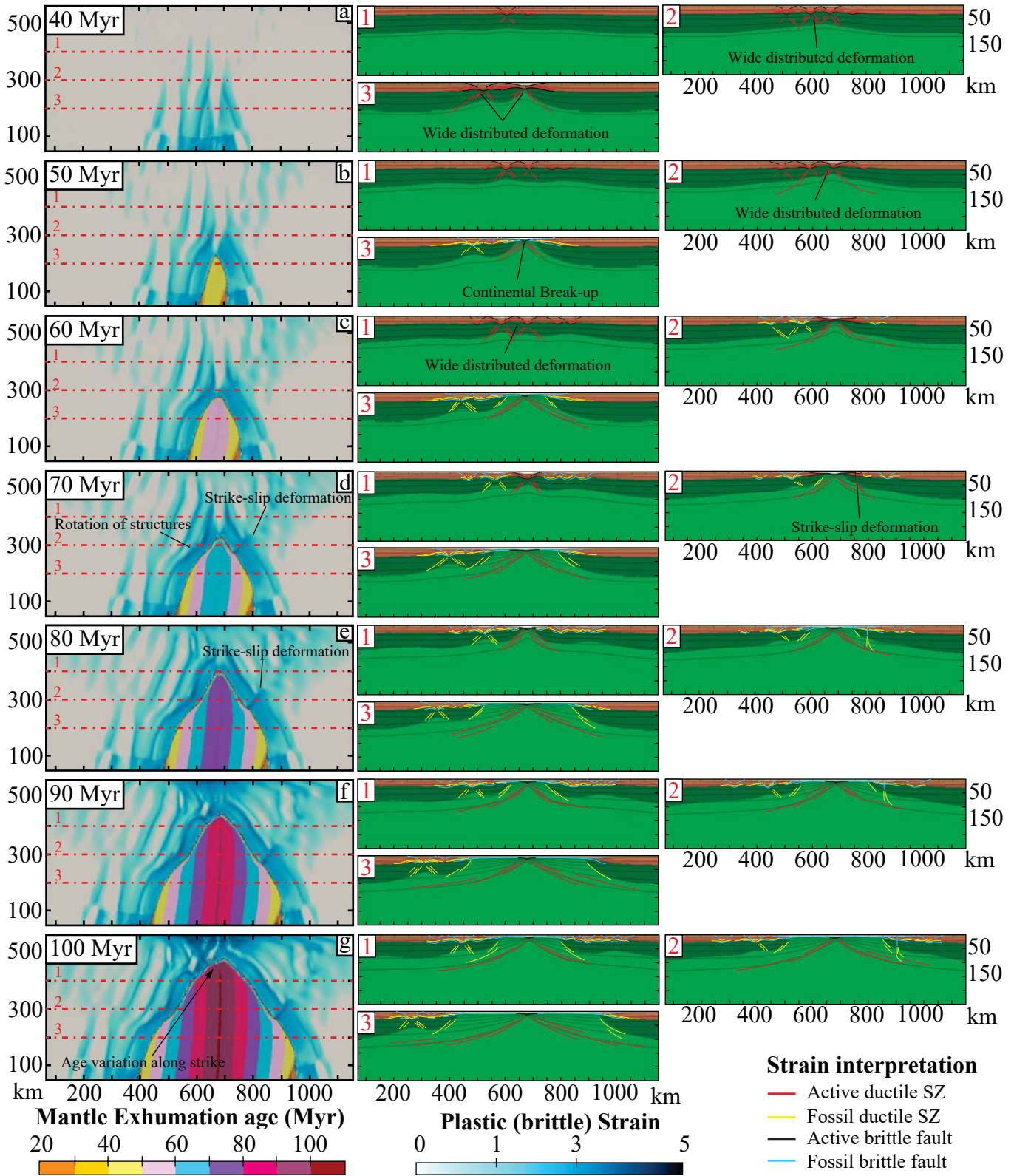


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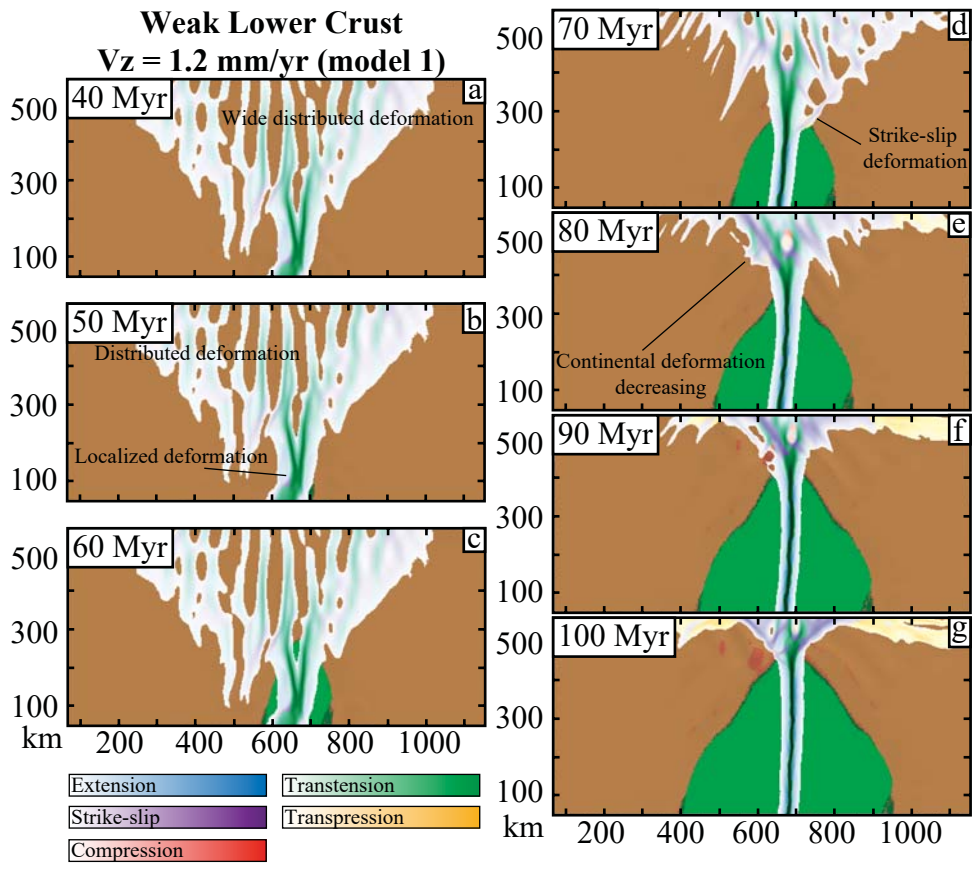


Figure 9.

Weak Lower Crust
 $V_z = 1.5 \text{ mm/yr}$ (model 2)

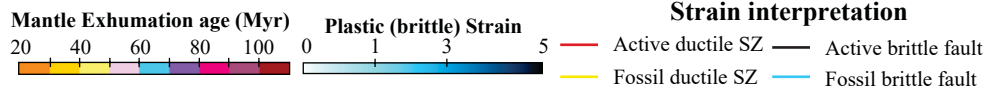
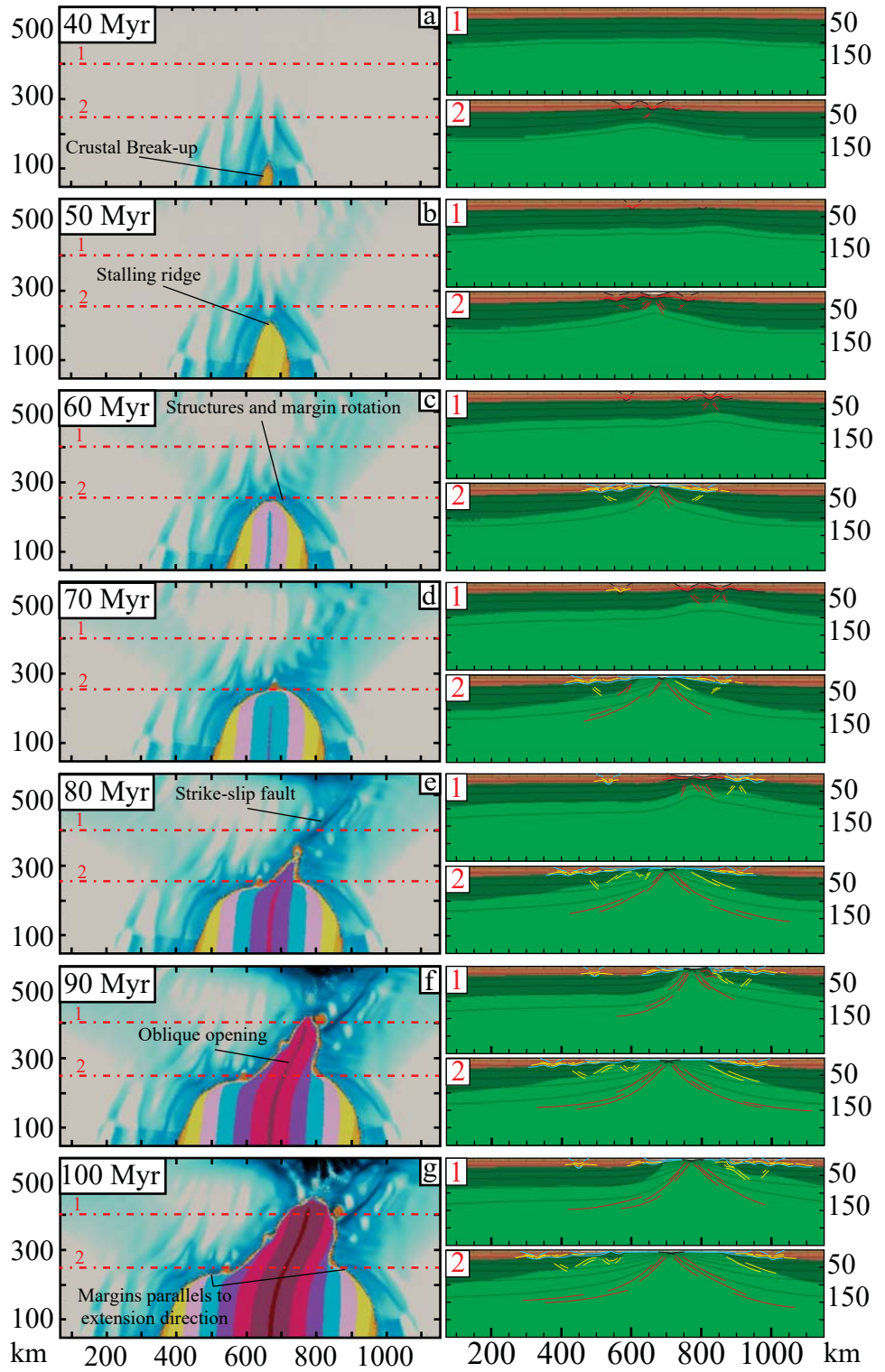


Figure 10.

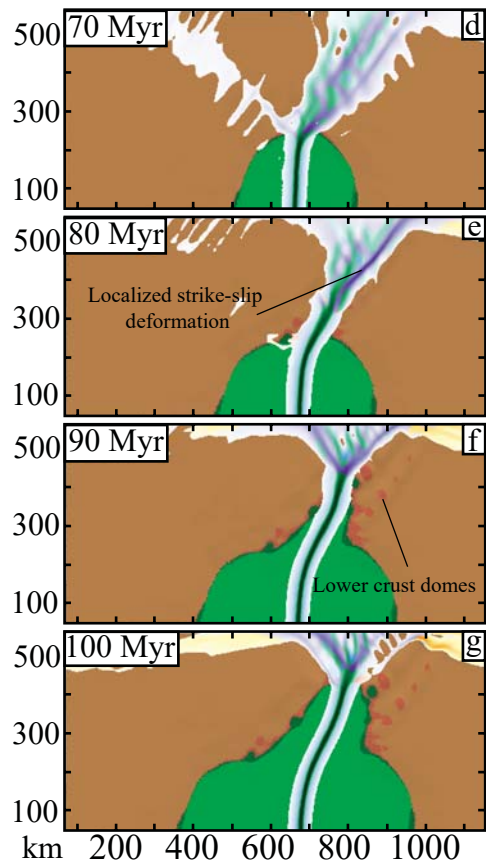
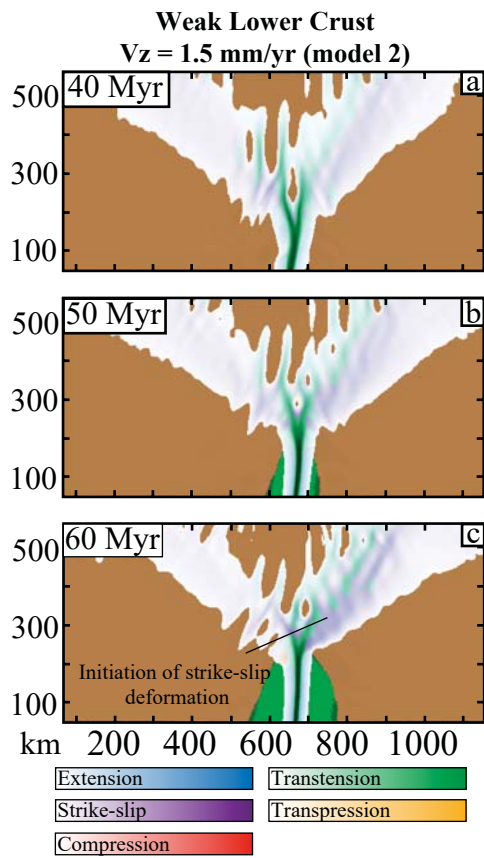


Figure 11.

Strong Lower Crust
 $V_z = 1.7$ mm/yr (model 3)

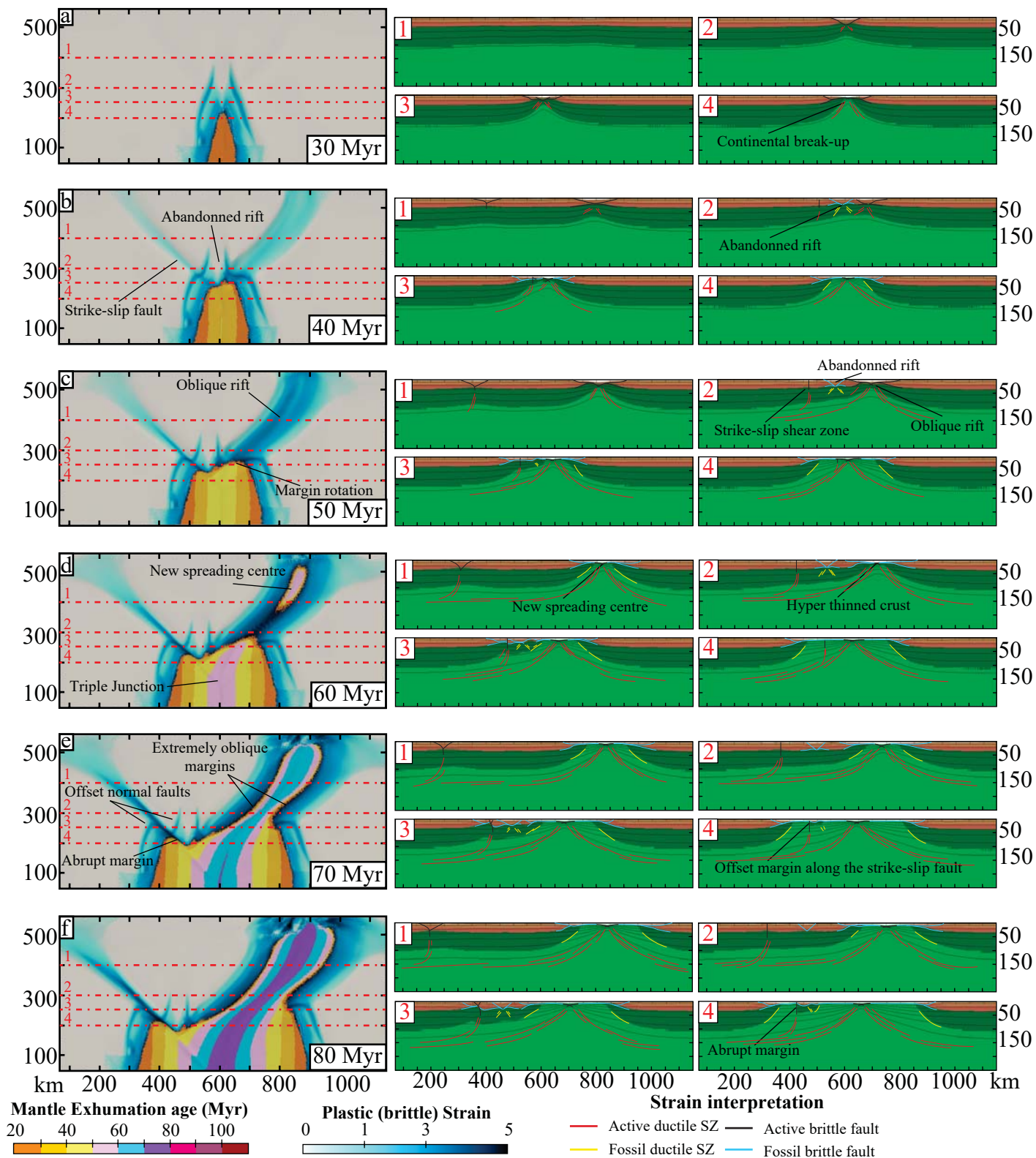


Figure 12.

Strong Lower Crust
 $V_z = 1.7$ mm/yr (model 3)

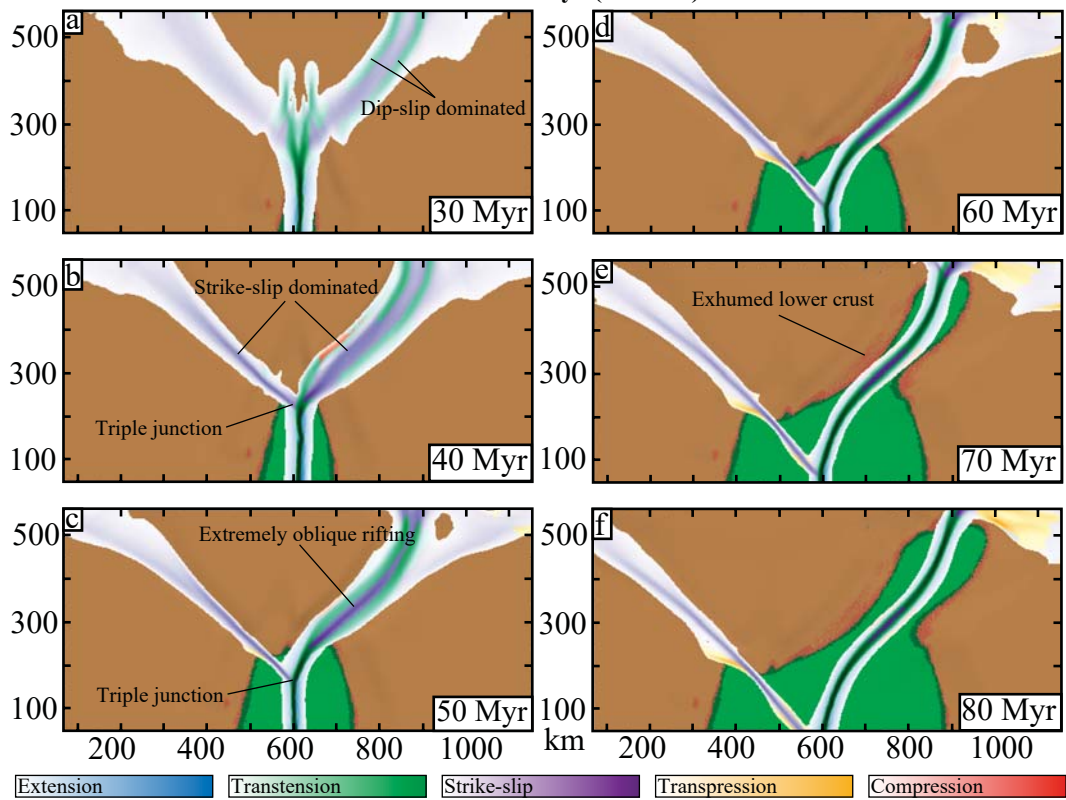


Figure 13.

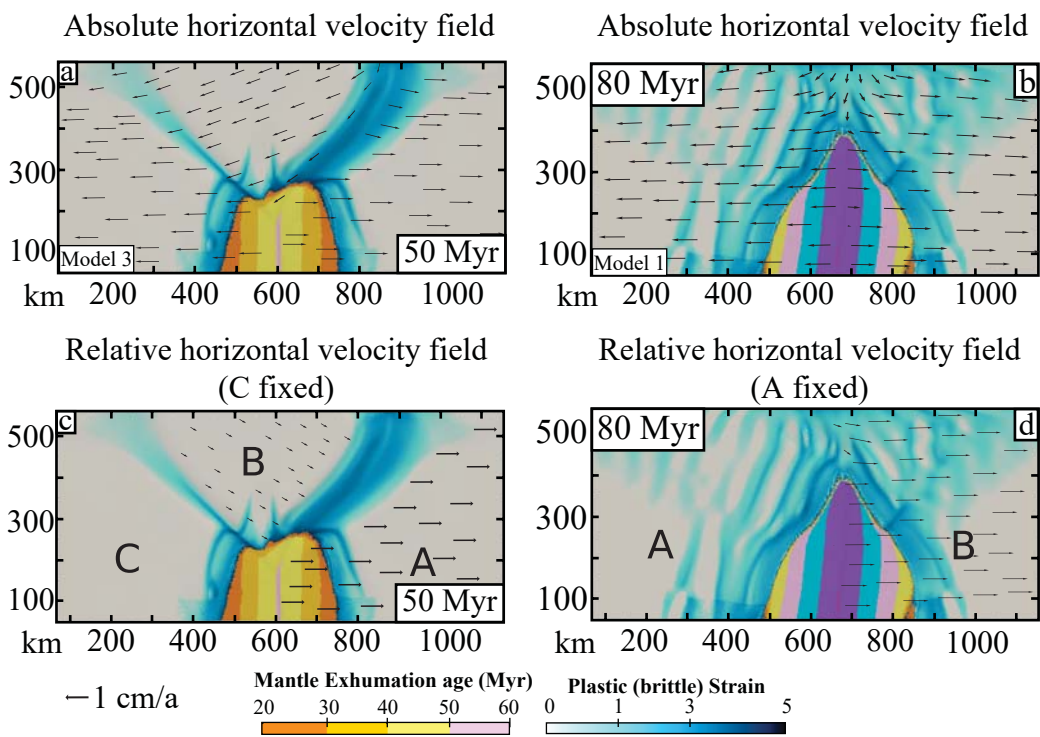


Figure 14.

