

2 **A plate tectonics oddity: Caterpillar-walk exhumation of subducted continental**
3 **crust**

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15 **1. Numerical method, model setup and experiment description.**

16 **a) Numerical model**

17 We use thermo-mechanical code Flamar v12 to assess the response of multi-layered
18 visco-elasto-plastic lithosphere. The code is based on the FLAC algorithm (Cundall,
19 1989; Poliakov et al., 1993) and is described, following its evolution, in many previous
20 studies (Burov and Poliakov, 2001; Burov et al., 2001; Burov et al., 2003; Burov and
21 Guillou-Frottier, 2005; Burov and Yamato, 2008; Burov and Cloetingh, 2009). In

22 difference from its predecessors (the above-cited references before Burov and Yamato,
 23 2008) Flamar is thermodynamically coupled (accounts for mineralogical phase
 24 changes) and extensively relies on marker in-cell technique to trace material properties,
 25 P-T-t paths and to interpolate physical parameter fields during dynamic remeshings of
 26 the Lagrangian framework. Here we limit the description of the code to most essential
 27 features: the ability to handle (1) large strains and multiple visco-elastic-plastic
 28 rheologies including Mohr-Coulomb failure (faulting) and non-linear pressure-
 29 temperature and strain-rate dependent ductile creep; (2) stain localization and strain
 30 softening/hardening; (3) thermo-dynamic phase transitions; (4) internal heat sources;
 31 (5) free surface boundary conditions and surface processes (diffusive erosion and
 32 sedimentation).

33 **Basic equations.** Flamar has a mixed finite-difference/finite element numerical
 34 scheme, with a Cartesian coordinate frame and 2D plane strain formulation. The
 35 Lagrangian mesh is composed of quadrilateral elements subdivided into 2 couples of
 36 triangular sub-elements with tri-linear shape functions. Flamar uses a large strain fully
 37 explicit time-marching scheme. It locally solves full Newtonian equations of motion in
 38 a continuum mechanics approximation:

$$39 \quad \rho \frac{Dv_i}{Dt} - \frac{\partial \sigma_{ij}}{\partial x_j} = \rho g_i \quad (1)$$

40 coupled with constitutive equations:

$$41 \quad \frac{D\sigma}{Dt} = F(\sigma, \mathbf{u}, \mathbf{v}, \nabla \mathbf{u}, \nabla \mathbf{v} \dots T \dots) \quad (2),$$

42 and with equations of heat transfer (where heat advection terms $\mathbf{v}\nabla T$ are included in
 43 the Lagrangian derivative DT/Dt) and state equations:

44
$$\rho C_p DT/Dt - k\nabla^2 T - \sum_i^n H_i, \quad (3)$$

45
$$\rho = f(P, T) \quad (4).$$

46 Here $\mathbf{v}, \mathbf{u}, \boldsymbol{\sigma}, \mathbf{g}, k$ are the respective terms for velocity, displacement, stress,
 47 acceleration due to body forces and thermal conductivity. P is pressure (negative for
 48 compression). The terms t, ρ, C_p, T, H_i designate respectively time, density, specific
 49 heat, temperature, internal heat production, thermal expansion coefficient and
 50 isothermal compressibility. The symbol Σ means summation of various heats sources
 51 H_i . The state equation $\rho = f(P, T)$ refers to the formulation, in which phase changes are
 52 taken into account and density is computed by using free Gibbs energy minimization
 53 algorithm that evaluates the equilibrium density of constituent mineralogical phases for
 54 given P and T as well as latent heat contribution H_l to the term $\sum_i^n H_i$ ($\sum_i^n H_i = H_r + H_f$
 55 $+ H_l + \dots$). The term $\sum_i^n H_i$ also accounts for radiogenic heat H_r and frictional
 56 dissipation H_f . The terms $D\boldsymbol{\sigma}/Dt, F$ are the objective Jaumann stress time derivative
 57 and a functional, respectively. In the Lagrangian method, incremental displacements
 58 are added to the grid coordinates allowing the mesh to move and deform with the
 59 material. This allows for the solution of large-strain problems while using locally the
 60 small-strain formulation: on each time step the solution is obtained in local
 61 coordinates, which are then updated in a large-strain mode.

62 Solution of (1) provides velocities at mesh points used for computation of element
 63 strains and of heat advection $\mathbf{v}\nabla T$. These strains are used to calculate element stresses
 64 and equivalent forces used to compute velocities for the next time step. Due to the

65 explicit approach, there are no convergence issues in case of non-linear rheologies. The
 66 algorithm automatically checks and adopts the internal time step using Courant's
 67 criterion of stability. Once Courant's criterion is satisfied, physical instabilities are
 68 treated in correct way since in this case the algorithm cannot miss forks in physical
 69 processes.

70 **Explicit elastic-viscous-plastic rheology.** We use common serial Maxwell-type solid,
 71 in which the total strain increment in each numeric element is defined by a sum of
 72 elastic, viscous and brittle strain increments. In contrast to fluid dynamic approaches,
 73 where non-viscous rheological terms are simulated using pseudo-plastic and pseudo-
 74 elastic viscous terms (Solomatov and Moresi, 2000; Bercovici and Ricard, 2001),
 75 Flamar explicitly treats all rheological three terms. The parameters of elastic-ductile-
 76 plastic rheology laws for crust and mantle are derived from rock mechanics data
 77 (Supplementary Table1).

78 **Plastic (brittle) behavior.** The brittle behaviour of rocks is described by Byerlee's law
 79 (Byerlee, 1978), which corresponds to a Mohr-Coulomb material with friction angle ϕ
 80 = 30° and cohesion $|C_0| < 20$ MPa:

$$81 \quad |\tau| = C_0 - \sigma_n \tan \phi, \quad (5)$$

82 where σ_n is normal stress $\sigma_n = \frac{1}{3}\sigma_I + \sigma_{II}^{\text{dev}} \sin \phi$, $\frac{1}{3}\sigma_I = P$ is the effective pressure
 83 (negative for compression), σ_{II}^{dev} is the second invariant of deviatoric stress, or
 84 effective shear stress. The condition of the transition to brittle deformation (function
 85 rupture f) reads as: $f = \sigma_{II}^{\text{dev}} + P \sin \phi - C_0 \cos \phi = 0$ and $\partial f / \partial t = 0$. In terms of principal
 86 stresses, the equivalent of the yield criterion (6) reads as:

$$87 \quad \sigma_1 - \sigma_3 = -\sin \phi (\sigma_1 + \sigma_3 - 2C_0 / \tan \phi) \quad (6)$$

88 **Elastic behaviour.** Elastic behaviour is described by the linear Hooke's law:

$$89 \quad \sigma_{ij} = \lambda \varepsilon_{kk} \delta_{ij} + 2G \varepsilon_{ij}, \quad (7)$$

90 where λ and G are Lamé's constants. Repeating indexes mean summation and δ is the
91 Kronecker's operator.

92 **Viscous (ductile) behaviour.** Within deep lithosphere and underlying mantle regions,
93 creeping flow is highly dependent on temperature and is non-linear non-Newtonian
94 since the effective viscosity also varies as function of differential stress (Kirby and
95 Kronenberg, 1987; Turcotte and Schubert, 2002):

$$96 \quad \dot{\varepsilon}^d = A (\sigma_1 - \sigma_3)^n \exp(-Q/RT), \quad (8)$$

97 where $\dot{\varepsilon}^d$ is effective shear strain rate, A is a material constant, n is the power-law
98 exponent, $Q = E_a + PV$ is the activation enthalpy, E_a is activation energy, V is
99 activation volume, P is pressure, R is the universal gas constant, T is temperature in K,
100 σ_1 and σ_3 are the principal stresses. The effective viscosity μ_{eff} for this law is defined
101 as:

$$102 \quad \mu_{\text{eff}} = \dot{\varepsilon}^{(1-n)/n} A^{-1/n} \exp(Q/nRT). \quad (9)$$

103 For non-uniaxial deformation, the law (9) is converted to a triaxial form, using the
104 invariant of strain rate and geometrical proportionality factors:

$$105 \quad \mu_{\text{eff}} = \dot{\varepsilon}_{\text{II}}^d (1-n)/n (A^*)^{-1/n} \exp(Q/nRT), \quad (10)$$

106 where $\dot{\varepsilon}_{\text{II}}^d = (\text{Inv}_{\text{II}}(\dot{\varepsilon}_{ij}))^{1/2}$ and $A^* = 1/2 A \cdot 3^{(n+1)/2}$. The parameters A , n , Q are the
107 experimentally determined material constants (Supplementary Table 1). Using olivine
108 flow parameters, we verify that the predicted effective viscosity at lithosphere
109 asthenosphere boundary is $10^{19} - 5 \times 10^{19}$ Pa s matching post-glacial rebound data

110 (Turcotte and Schubert, 2002). Due to temperature dependence of the effective
111 viscosity, the viscosity decreases from 10^{25} - 10^{27} Pa s to asthenospheric values of 10^{19}
112 Pa s in the depth interval 0-152 km. Within the adiabatic temperature interval in the
113 convective upper mantle (152 km –700km), the dislocation flow law (9) is replaced by
114 a nearly Newtonian diffusion creep. We also use Peierl’s stress limiter that limits
115 deviatoric stresses in mantle (7) assuming Peierl’s stress of 5GPa.

116

117 **b) Initial and boundary conditions**

118 The model geometry (Supplementary Fig. 1) comprises a rectangular box (2000 x 700
119 km) with variable spatial resolution: 85 x 115 quadrilateral bilinear elements (4 km x 4
120 km) for the first 340 km depth and 36 x 115 quadrilateral bilinear elements (10 km x 4
121 km) below 300 km depth. Each element of the numerical grid is assigned its specific
122 material phase defined as a subset of physical parameters of the corresponding
123 material: density, thermal and rheology parameters (Supplementary Table 1). The
124 initial geometry assumes an already initiated oceanic subduction, with the subducting
125 slab reaching down to 272-km depth. Two blocks of continental crust are embedded in
126 the subducting plate, fated to be accreted to the overriding plate in the course of the
127 ongoing subduction. These continental blocks are 400 km wide and with 44-km thick
128 crust (representing characteristic dimensions of micro-continental terrains and
129 reconstructed volumes of accreted metamorphic rocks (Brun and Faccenna, 2008;
130 Jolivet and Brun, 2010)). The two continental blocks are at 150 km away from each
131 other (following those structural reconstructions), with oceanic lithosphere in between.
132 The continental crust is assumed to be homogeneous and of quartzite-type composition
133 (Ranalli and Murphy, 1987), with a density of 2800 kg m^{-3} ; sediments (Kronenberg

134 and Tullis, 1984) have a density of 2400 kg m^{-3} . Oceanic crust is represented by
135 gabbro basalt and crust-mantle interface – by serpentinite (Hilairt, 2007), both with a
136 density of 2900 kg m^{-3} and the subducting mantle lithosphere is represented by olivine
137 (Brace and Kohlstedt, 1980) with density of 3350 kg m^{-3} . The mechanical boundary
138 conditions assigned on the four sides of the box are: at the left and right side are fixed
139 horizontally; at the bottom: hydrostatic pressure with free slip in all directions; the
140 upper surface is explicitly free (free stress and free slip/displacement condition in all
141 directions), with moderate diffusion surface erosion ($k_e = 500 \text{ m}^2/\text{yr}$, see e.g. Burov and
142 Yamato, 2008). The bottom is pliable Winkler basement (= hydrostatic condition). A
143 bottom layer is introduced with a viscosity and a density higher than for the subducting
144 slab, to simulate pressure-induced viscosity growth toward the transition at 660 km
145 depth and to avoid extreme deformation of the bottom of the box when the slab reaches
146 it. Plastic and viscous materials soften with accumulated strain through a decrease of
147 the angle of friction and cohesion and decrease in viscosity, respectively
148 (Supplementary Table 1).

149

150

151 **c) Thermal structure of the lithosphere**

152 The thermal structure is one of the key parameters defining the mechanical strength
153 and buoyancy of the lithosphere. In this model, the thermal base of the lithosphere
154 (1330°C) is placed at 200 km depth corresponding to old oceanic lithosphere
155 approaching subduction zone (Schubert et al., 2001). The surface temperature is fixed
156 at 0°C and zero heat flow (no heat exchanges with the surrounding region) is set as the
157 lateral thermal boundary condition. The initial background geotherm for the mantle

158 lithosphere system is obtained by joining the lithospheric and deep mantle adiabatic
159 geotherms (approx. $0.3^{\circ}\text{C}/\text{km}$, Sleep, 2003). Fixed temperature of 1600°C is used as
160 bottom boundary condition at 700 km depth. The initial geotherms in the lithosphere
161 are age-dependent (Parsons and Sclater, 1977; Burov and Diament, 1992; 1995).
162 Moreover, the initial geotherm within the subducting slab is represented by an oceanic
163 lithosphere with a 0°C at surface and 1330°C at the base following the slab
164 approximation described in Burov and Diament (1995) for an old oceanic geotherm.

165

166 **2. Deformation of the blocks in slab rollback process**

167 Supplementary Fig. 2a represent step-by-step evolution of the central part of the
168 model. Additional passive markers (color circles) allow for tracing of key parts of the
169 model (see also Supplementary Fig 3). Supplementary Fig. 2b shows the step-by-step
170 evolution of the strain rate field, and instantaneous deformation at each time step. The
171 first phase of evolution is characterized by the accretion of the block 1 underthrust
172 below the accretionary wedge and oceanic suture zone. The consequence of this
173 accretion is the burial of a part of continental block, that remains cold, and the increase
174 of the slab dip due to the negative buoyancy of the buried continental crust. The
175 ongoing thrusting is responsible for splitting of the continental block on two parts, its
176 left part dragged under its right part along the new major thrust. When this part is
177 buried it pushes up the previously buried part of the block, forcing it to the surface.
178 After that the first major thrust is changed in a normal detachment to accommodate the
179 exhumation of the right part of the first block. The oceanic subduction that continues
180 after the accretion of block 1, while this latter is still partly attached to the subducted
181 slab, provokes a rotation of the entire block still at the favor of the slip of the normal

182 detachment. As result of this rotation, exhumed crustal structure is inversed, with
183 lower crust exposed to the surface. When the oceanic subduction advances enough to
184 decouple the block 1 from its subducting mantle part, the dip of the slab decreases and
185 the block is no longer undergoing rotation. The fully decoupled block 1 now belongs to
186 the overriding plate. The change of the dip of the slab combined with the acceleration
187 of the slab retreat result in the ascent of the asthenosphere below the right tip of the
188 block. When the block 2 arrives at the trench, the dip of the slab increases, leading to
189 the rising of the asthenosphere below the entire first block. The warming of the block 1
190 can be seen on the P-T-t paths in Supplementary Fig 3. The association between the
191 extension and the vigorous warming of the first continental block results in a
192 metamorphic core complex mode of exhumation. This deformation mode, based on a
193 crustal flow, enables a considerable amount of stretching while the Moho stays
194 practically flat. The extension is mostly localized in this part because the ductile
195 potential is high. The accretion of the block 2 occurs in the same way than the first
196 one. The only difference is that the left tip of the block stays somehow connected to
197 the trench, and is not completely decoupled from the slab as it is for the block 1.

198

199 **3. Pressure-temperature-time paths**

200

201 Supplementary Fig. 3 shows the final structure with several colored markers located
202 both in blocks 1 and 2. Their PTt paths are shown in Supplementary Figure 3b. The
203 PTt paths of two blocks show different patterns that we can qualify as “cold” or “hot”.
204 The “hot” paths of the block 1 are the blue and the pink one, yellow for the block 2.
205 They are characterized by an effective burial to ~2 to 3GPa followed by a rapid cold

206 exhumation and then a strong increase of temperature related to the detachment of the
207 blocks from the slab and the consequent asthenospheric warming. The “cold” paths (all
208 the other ones) are characterized by an effective burial to ~2 to 3GPa followed by a
209 rapid cold exhumation.

210

211 **4. Reference experiment with laterally homogeneous lithosphere (no continental**
212 **blocks)**

213

214 Supplementary Fig. 4 shows several steps of an experiment without the continental
215 blocks in the subducting slab. As for the experiments with heterogeneous lithosphere,
216 slab retreat starts immediately as well as the extensional deformation in the overriding
217 crust. A part of the continental lithosphere that forms the overriding plate is extremely
218 stretched to accommodate the extension due to the slab rollback (from 0 to 42.5 Myr).
219 Then high slab break-off occurs, stopping the subduction process as well as the
220 extensional deformation in the overriding plate (70.3 Myr). This experiments shows
221 that in the absence of structural heterogeneities, the subduction, roll-back and
222 exhumation are only one-staged progressive, and the roll-back has a much smaller
223 amplitude.

224

224 **Supplementary references:**

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279 **Supplementary figures, table and movie captions:**

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281 **Supplementary Fig. FT1:** a) Initial and boundary conditions of the experiment, b) initial
282 thermal conditions

283 **Supplementary Fig. FT2:** a) Zoom to the uppermost part of the plate contact zone in the
284 model, with selected passive markers (coloured circles), b) strain rate distributions in
285 the of the upper part of the experiment (showing localization of deformation, in
286 particular crustal detachments). This figure is spread on to three sheets. Figure shows
287 step-by-step evolution of the central part of the model where two embedded
288 continental crustal blocks first subduct and then get exhumed to the surface, with the
289 material of the second block sliding below and behind the exhumed preceding block.
290 Delamination of the crustal blocks has a major impact on the evolution of the
291 subduction zone and, in the particular, on slab roll-back and back-arc extension. On
292 their way to the surface blocs undergo reversing rotation, just like a caterpillar bent,
293 resulting in inversion of the crustal structure at surface. At surface, both crustal blocks
294 stretch laterally forming large scale metamorphic terrain structurally resembling
295 structural reconstructions of the Aegean realm (Fig. 1)

296 **Supplementary Fig. FT3:** a) Final structure of the experiment with the final position of
297 the colored markers shown in Fig. S2, b) pressure-temperature paths through time
298 (PTt) of the markers of block 1, c) pressure-temperature paths through time (PTt) of
299 the markers of block 2.

300 **Supplementary Fig. FT4:** Several snapshots of the experiment with laterally
301 homogeneous lithosphere ,i.e. without continental blocks in the subducting slab.

302 **Supplementary Table FT1:** Parameters of the experiments.

303 **Supplementary Movie FT1:** This movie shows the evolution of the structural pattern of
304 the experiment.

305

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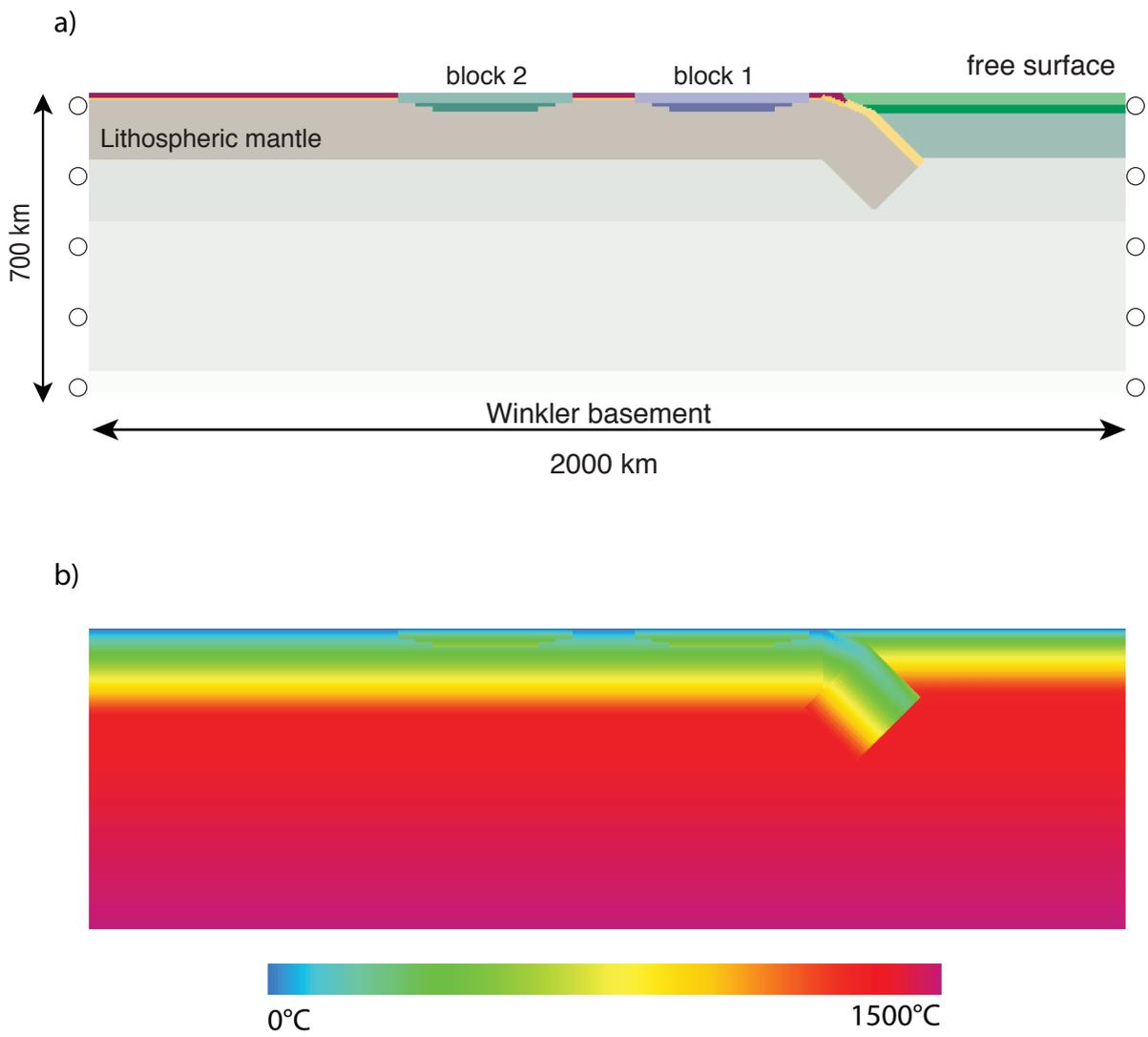
Supplementary Table 8 F 1: Numerical parameters

		Continental crust	Sediments	Oceanic crust	Lithospheric mantle	Mantle	660km boundary
		Quartzite dry ¹	Quartzite wet ²	Serpentinite ³	Olivine ⁴	Olivine ⁴	Olivine ⁴
Viscosity parameters	n	2.4	2.7	3.8	3	3	3
	A, MPa ⁻ⁿ .s ⁻¹	6.3 x 10 ⁻⁶	2.18 x 10 ⁻⁶	4.5 x 10 ⁻¹⁵	7x 10 ⁴	7x 10 ⁴	7x 10 ⁴
	E, kJ.mol ⁻¹	156	120	89	520	520	520
Minimum viscosity					10 ²¹		10 ²⁵
Density kg.m ⁻³		2700 (3350 [*])	2400	2900 (3350 ^{**})	3350	3320-3330	3360
Elastic parameters	λ, Pa	3 x 10 ¹⁰	3 x 10 ¹⁰	1 x 10 ¹⁰	3 x 10 ¹⁰	1 x 10 ¹⁰	4 x 10 ¹⁰
	μ, Pa	3 x 10 ¹⁰	3 x 10 ¹⁰	1 x 10 ¹⁰	3 x 10 ¹⁰	1 x 10 ¹⁰	4 x 10 ¹⁰
Cohesion, Pa		20 to 0	20 to 0	20 to 0	20	20	300
Frictional angle, °		10 to 2	10 to 2	10 to 2	30	30	2
Thermal conductivity, W.m ⁻¹ .K ⁻¹		2.5	2.5	3.5	3.5	3.5	3.5
Specific heat Cp, J.kg ⁻¹ .K ⁻¹		1 x 10 ³	1 x 10 ³				
Internal heat production at surface, hs, W.kg ⁻¹		1 x 10 ⁻⁹	1 x 10 ⁻⁹				
Production decay depth, hr, km		10	10				

¹ from Ranalli and Murphy 1987, ² from Kronenberg and Tullis, 1984, ³ from Hilairet et al, 2007, ⁴ from Brace and Kohlstedt, 1980

* change of density when pressure > 5 x 10⁹ and temperature > 550°C

** deep oceanic crust at the onset of experiments



Supplementary Figure 1: a) Initial and boundary conditions of the experiment, b) initial thermal conditions

Figure DR1

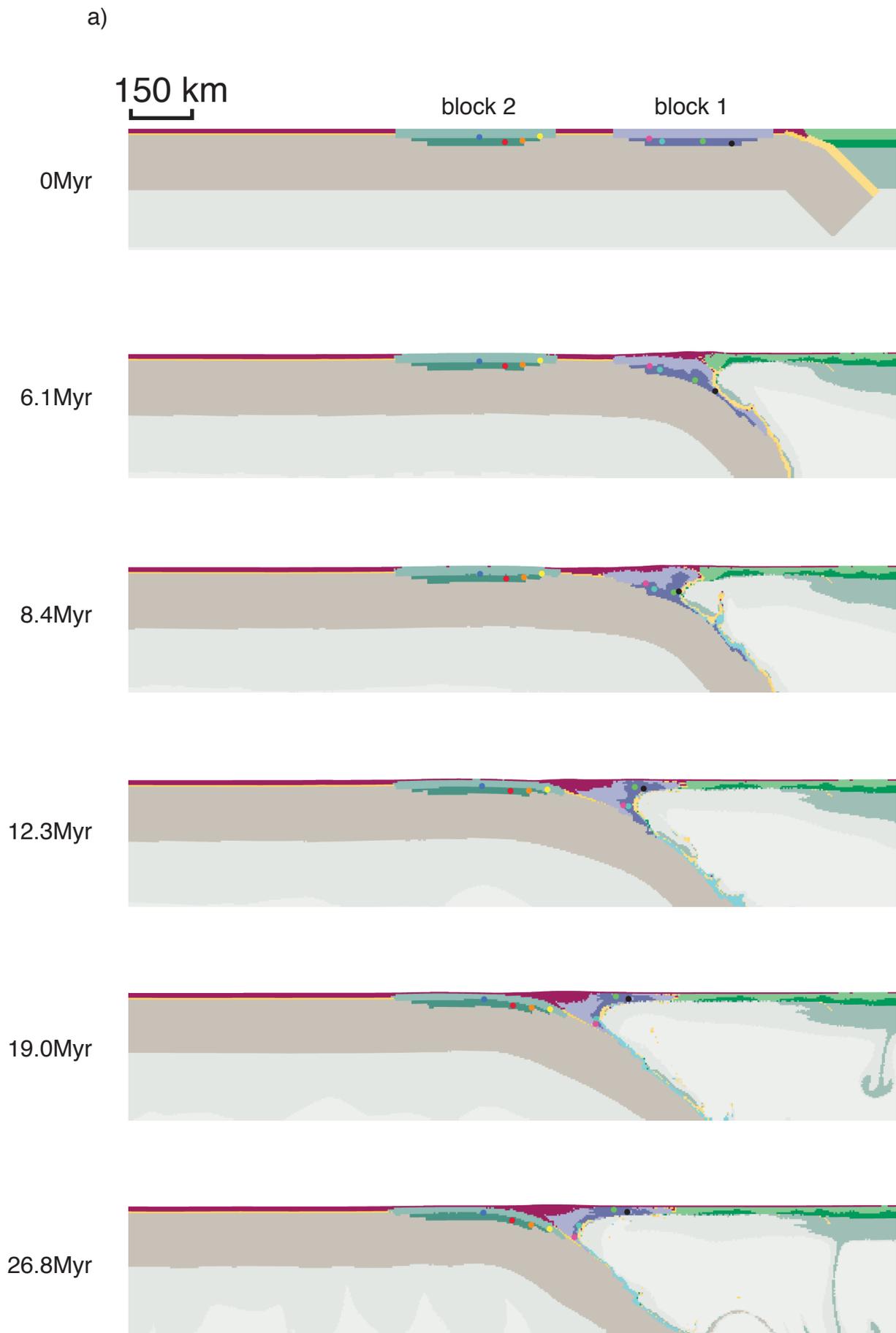


Figure AÜ2

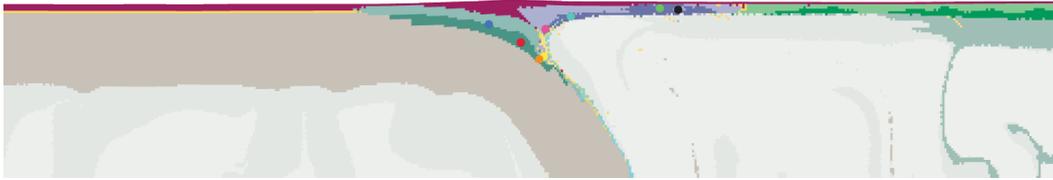
a)

150 km

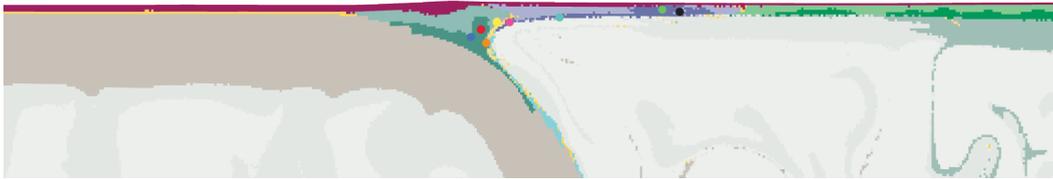
30.8Myr



36.6Myr



42.5Myr



46.4Myr



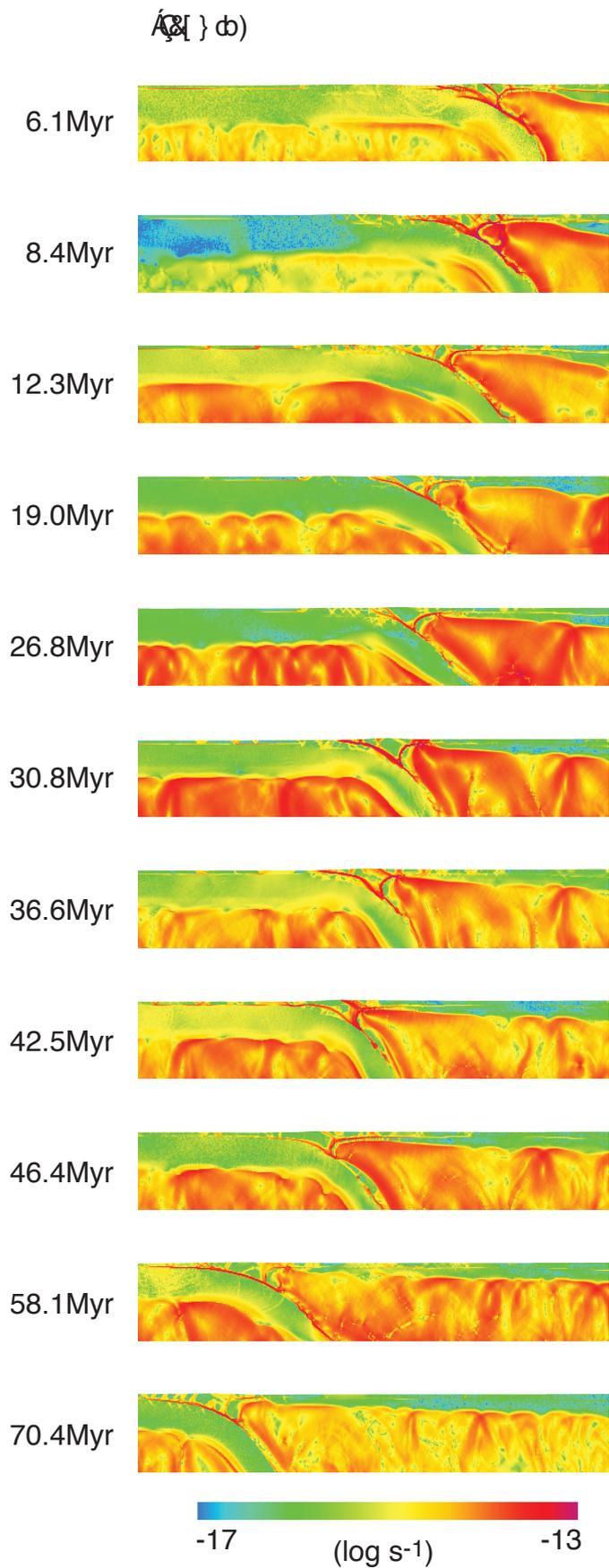
58.1Myr



70.4Myr

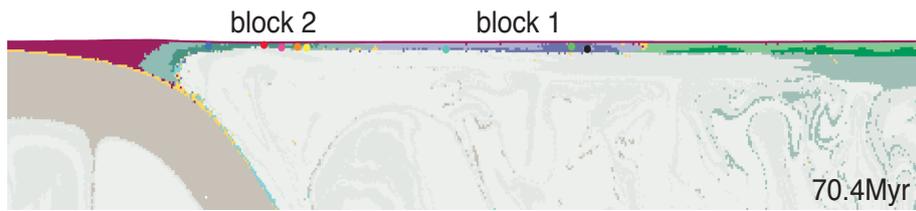


Figure 2.3.1

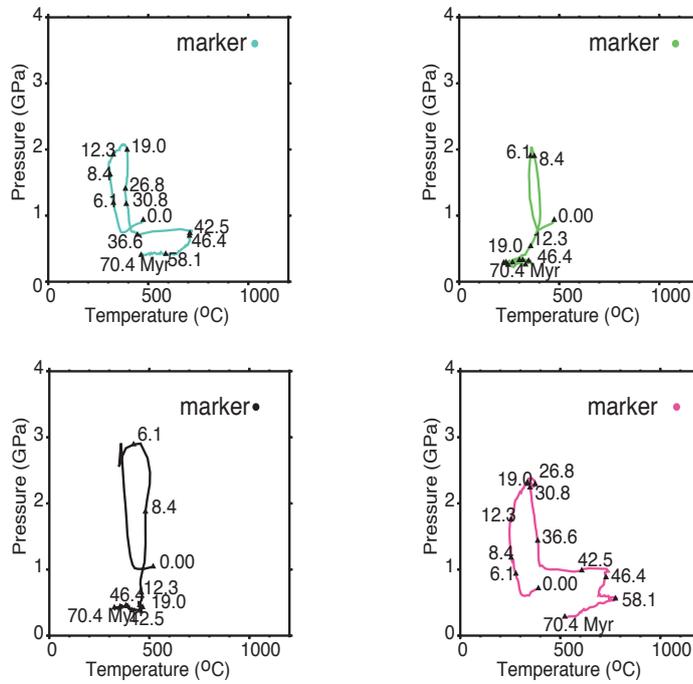


Supplementary Figure 2: a) structure of a upper part of the experiment with some markers, b) strain rate of the upper part of the experiment. The figure is spread on three sheets.

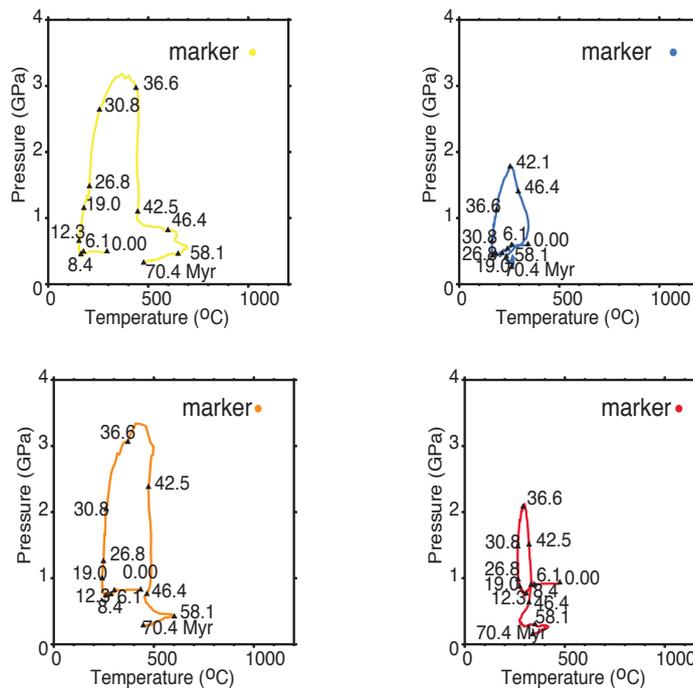
a) structure and colored marker at 70.4 Myr



b) Block 1



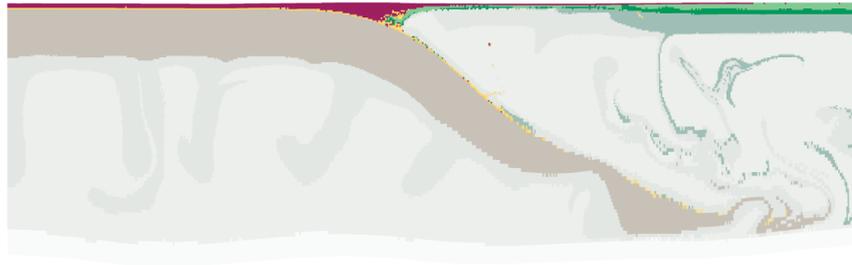
c) Block 2



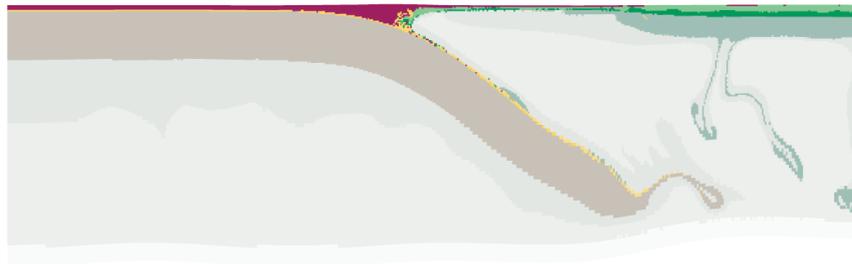
Supplementary Figure ÖÜ3: a) Final structure of the experiment with the position of the colored markers also presented in Supplementary Figure 2, b) pressure-temperature paths through time of the markers of block 1, c) pressure-temperature paths through time of the markers of block 2.



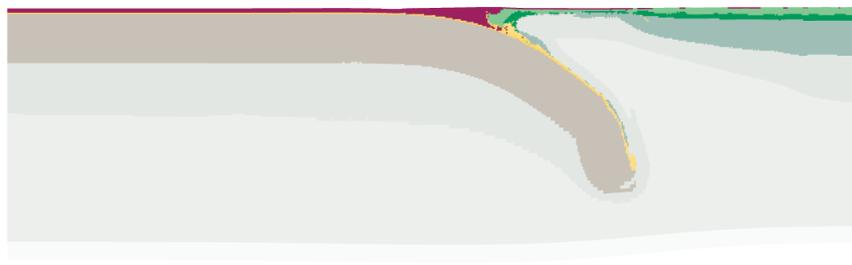
70.3Myr



42.5Myr



26.8Myr



7.15Myr

Supplementary Figure DR4: Snapshots of an experiment without any microcontinent.

Figure DR4