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1 **On the influence of the asthenospheric flow on the tectonics and topography at a collision-**
2 **subduction transition zones: comparison with the eastern Tibetan margin**

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11

12 **Abstract**

13 **The tectonic and topographic evolution of southeast Asia is attributed to the**
14 **indentation of India into Eurasia, gravitational collapse of the uplifted terrains and the**
15 **dynamics of the Sunda and other western Pacific subduction zones, but their relative**
16 **contributions remain elusive. Here, we analyse 3D numerical geodynamic modeling results**
17 **involving a collision-subduction system and show that vigorous asthenospheric flow due to**
18 **differential along-strike slab kinematics may contribute to the surface strain and elevations**
19 **at collision-subduction transition zones. We argue that protracted northward migration of**
20 **the collisional front and Indian slab during south to south-westward rollback subduction**
21 **along the Sunda margin might have produced a similar asthenospheric flow. This flow**
22 **could have contributed to the southeast Asia extrusion tectonics and uplift of the terrains**
23 **around the eastern Himalayan syntaxis and protruding from southeast Tibet. Therefore, we**
24 **suggest that the tectonics and topographic growth east and southeast of Tibet are controlled**
25 **not only by crustal and lithospheric deformation but also by asthenospheric dynamics.**

26 **1. Introduction**

27 The growth of the Tibetan plateau and the magnificent mountains of Asia has been long
28 ascribed to lithospheric shortening and thickening along the India-Eurasia collisional margin
29 (Argand 1924), following the closure of the Tethyan ocean and on-going since ~50 Ma (Molnar &
30 Tapponnier 1975; Jolivet et al. 1990; Meyer et al. 1998; Cowgill et al. 2003). Unlike the abrupt
31 Himalayan mountain front, the gentler but still impressive topography along the eastern margin
32 of Tibet developed in a predominantly trans-tensional tectonic regime (Leloup et al. 1995; Wang
33 et al. 1998; Hall & Morley 2004). These fundamental observations and the geophysical evidence
34 suggesting the presence of a weak lower crust below Tibet (Nelson et al. 1996; Xu et al. 2007)
35 triggered a debate about the partitioning between clock-wise rigid rotation (Tapponnier et al.
36 1982; Armijo et al. 1986; Avouac & Tapponnier 1993; Leloup et al. 1995; Meade 2007) or
37 viscous eastward evacuation of the Eurasian crust and lithosphere, possibly involving crustal
38 channel flow (England & Houseman 1986; Royden et al. 1997; Clark & Royden 2000; Clark et
39 al. 2005; Schoenbohm et al. 2006; Copley & McKenzie 2007; Royden et al. 2008). These models
40 put different emphasis on strain localization, vertical gradients of strain due to depth-dependent
41 rheologies, the role of gravitational body forces and tractions at the base of the lithosphere, and
42 the influence of plate boundary dynamics. Unravelling the influence of these factors is
43 challenging also because analyses based on the distribution of gravitational potential energy
44 (*GPE*) (England & Molnar 1997) or simplified mechanical models such as the thin viscous sheet
45 (England & McKenzie 1983) might provide limited insights (Lechman et al. 2011; Schmalholz et
46 al. 2014).

47 Common to all proposed models are a focus on crustal dynamics and, regarding the
48 effects of plate boundary dynamics, the assumption that subduction of oceanic lithosphere

49 beneath eastern Eurasia and Indonesia chiefly created the accommodation space for
50 unconstrained continental extrusion. Some authors proposed that the suction exerted by these
51 subduction zones could have contributed to regional extension and extrusion (Burchfiel &
52 Royden 1985; Jolivet et al. 1990; Jolivet et al. 1994; Northrup et al. 1995; Fournier et al. 2004;
53 Flesch et al. 2005; Schellart & Lister 2005; Ghosh et al. 2006; Royden et al. 2008). Concerning
54 basal tractions, large-scale mantle convection might provide support to the topography of
55 southeast Asia through vertical stresses (Ricard et al. 1993; Lithgow-Bertelloni & Gurnis 1997)
56 and contribute to the overall India-Eurasia convergence (Ghosh et al. 2008; Ghosh et al. 2009;
57 Alvarez 2010; Becker & Faccenna 2011). Minor attention, however, has been given to the
58 potential forcing from the asthenospheric return flow owing to protracted northward migration of
59 the Indian slab and indentation front during rollback (late-Eocene to middle-Miocene) or stable
60 (middle-Miocene to present) subduction along the Sunda and western Pacific margins
61 (Tapponnier et al. 1986; van der Hilst & Seno 1993; Hall & Morley 2004; Sibuet et al. 2004;
62 Honza et al. 2004; Royden et al. 2008; Replumaz et al. 2013) (shortened to “differential along-
63 strike slab kinematics” or similar paraphrases in the following).

64 Our principal objective is thus to assess the potential role of the asthenospheric return
65 flow in response to differential along-strike kinematics in affecting the surface tectonics and
66 topography across coupled collision-subduction systems. To this aim, we present and discuss
67 results from numerical experiments involving joint continental collision and ocean-continent
68 subduction. The experiments suggest that vigorous mantle flow can arise from differential along-
69 strike slab kinematics and contribute to the surface strain and topography at the collision-
70 subduction transition zone by modulating the upper plate’s isostatic and dynamic balance. These
71 results are consistent with first-order structural, geophysical and geomorphic observations from

72 southeast Asia and build upon previously proposed models for the evolution of the eastern
73 Tibetan margin, which are first summarized hereafter.

74 **1.1 Tectonics and topography east and southeast of Tibet**

75 Intracontinental deformation of Eurasia involved a gradually increasing region after the
76 onset of continental collision with the Indian plate (Argand 1924; Tapponnier & Molnar 1977;
77 Tapponnier et al. 2001; Royden et al. 2008; Copley et al. 2010). While thrusting and topographic
78 growth in southern Tibet and Himalayas started as early as Eocene times (Aikman et al. 2008;
79 Gebelin et al. 2013), large fragments of the Eurasian lithosphere were extruded eastward out of
80 the collision zone toward southeast Asia (Tapponnier et al. 1982; Tapponnier et al. 1986;
81 Replumaz & Tapponnier 2003; Akciz et al. 2008) (Figs. 1a, b). Until activation of the Altyn-Tagh
82 fault zone during the Oligocene, the Red River shear zone marked the northern boundary of the
83 east-moving fragments (Tapponnier et al. 1986; Leloup et al. 1995; Wang et al. 1998; Replumaz
84 & Tapponnier 2003). High mantle temperatures inferred from alkali-rich magmatism between
85 ~30-50 Ma (Holbig and Grove 2008) may have enhanced continental extrusion, which occurred
86 during rapid trench retreat and slab rollback along the Sunda and western Pacific oceanic
87 subduction zones relative to the main collisional front and Indian slab (Tapponnier et al. 1986;
88 Jolivet et al. 1990; van der Hilst & Seno 1993; Jolivet et al. 1994; Fournier et al. 2004; Royden et
89 al. 2008; Replumaz et al. 2013). Differential along-strike kinematics of the Tethyan and western
90 Pacific slabs are implied by widespread early-Cenozoic extension within the upper plate
91 lithosphere of Indonesia (Hall and Morley 2004), Eocene-Oligocene extension in the South and
92 East China Seas (Sibuet, Hsu, and Debayle 2004), and Oligocene to middle-Miocene extension in
93 the Sea of Japan (Jolivet & Tamaki 1992; Tamaki 1992; Jolivet et al. 1994; Honza et al. 2004)
94 (Fig. 1a). Fast extrusion of the Eurasian lithosphere along the eastern Tibetan plateau margin has

95 slowed after ~15-20 Ma, when slab rollback along the Sunda and western Pacific margins also
96 diminished or ended (van der Hilst and Seno 1993). Similarly, backarc basin opening in the Sea
97 of Japan stopped at 10-12 Ma when the Philippine Sea Plate had moved northward and the
98 Eurasia-Pacific-Philippine Sea triple junction reached its present position (Jolivet et al. 1994).
99 The history of subduction along the southeast Eurasian and western Pacific margins as well as
100 lithospheric contrasts in the collision zone is also revealed by seismic tomography. Fast velocity
101 anomalies are observed beneath southern Tibet and northern India (Fig. 1c) together with dipping
102 positive perturbations associated with the Sunda and other western Pacific slabs, which penetrate
103 into the mantle transition zone in multiple locations (Kárason & van der Hilst 2000; Li et al.
104 2008; Li et al. 2008; Replumaz et al. 2013; Schaeffer & Lebedev 2013; Auer et al. 2014). Other
105 significant regional positive anomalies are seen in the Sichuan basin and eastern Siberia craton,
106 while beneath the rest of southeast Asia and between the Indian and Sunda slabs seismic
107 velocities are anomalously low.

108 Eastward to southeastward continental extrusion is going on at present as indicated from
109 the pattern of active faulting (e.g., Molnar and Tapponnier, 1975; Tapponnier and Molnar, 1977),
110 although eastward extrusion of Tibet is partially absorbed by shortening in the Nan Shan and
111 Qilian Shan (Meyer et al, 1998) and to a lesser degree in the Longmen Shan thrust (Burchfiel
112 2004; Rongjun et al. 2007). GPS data relative to stable Eurasia (Gan et al. 2007) (Fig. 1b) show
113 convergence along the India-Eurasia margin and a prominent clockwise rotation around the
114 Eastern Himalayan Syntax (EHS), consistent with trench retreat and slab rollback in the Indo-
115 Burman region (Hall and Morley 2004). Differently from within the plateau, where uplift related
116 to the early India-Eurasia collision raised terrains already above sea level (Murphy et al. 1997;
117 Kapp et al. 2007), rapid uplift and erosion of relict low-relief terrains in eastern Tibet is coeval

118 with the structural transition to the present-day kinematic pattern (Clark et al. 2006).

119 East-west extension in central and southern Tibet during the late Cenozoic (Armijo et al.
120 1986; Williams et al. 2001; Blisniuk et al. 2001; Zhang et al. 2004) is also ascribed to body forces
121 due to high elevations (England & Houseman 1986; England & Molnar 1997; Flesch et al. 2001;
122 Liu & Yang 2003; Flesch et al. 2005). It is not clear, however, whether a gravitationally driven
123 flow would coherently affect the deformation down to upper mantle levels or end within the
124 crust, in turn implying intra-lithospheric rheological decoupling. Surface-wave tomography and
125 receiver functions suggest the presence of a relatively weak lower crust confined along the
126 eastern and southern plateau margin by rigid craton-like lithosphere beneath the Sichuan basin
127 and the Indian plate, respectively (Nelson et al. 1996; Xu et al. 2007). Overall, the geometry of
128 the plateau margins seems consistent with the underlying crustal rheology: the gently sloping
129 north- and south-eastern plateau margin overlies a weak lower crust, while the steeper southern
130 front overlies a strong crust throughout (Royden 1996; Clark & Royden 2000; Jordan & Watts
131 2005). Rapid eastward flow in the deeper crust would explain relatively late uplift of the extruded
132 terrains in absence of crustal shortening.

133 **1.2. Proposed models**

134 Classical models of the India-Eurasia indentation involve plane horizontal strain of a
135 plastic medium by a rigid indenter (Tapponnier et al. 1982; Peltzer and Tapponnier 1988) or
136 analogue experiments allowing for crustal thickening (Peltzer et al. 1982; Cobbold & Davy
137 1988), which led to successful qualitative comparisons of the slip lines across the plastic medium
138 to the orientations of major Asian strike-slip fault zones. Plane strain experiments, however,
139 neglect the vertical deformation and therefore the topographic evolution and the forcing from

140 surface elevation variations to the overall geometry of the deformation.

141 The regional lithospheric strain has been compared to numerical and analytical solutions
142 of the deformation field within a thin viscous sheet (Bird & Piper 1980; England & McKenzie
143 1983; England et al. 1985; England & Houseman 1986; Jiménez-Munt & Platt 2006), a
144 formulation that circumvents difficulties arising from the lack of knowledge on the crustal rock
145 rheology by solving the 3D strain field for depth-averaged stresses or velocities. Jointly with the
146 assumption of isostatic compensation this approach provided important insights amongst which
147 are, for example, estimates of the average viscosity of the Tibetan lithosphere (England and
148 Molnar 1997) and quantifications of crust-mantle coupling and the contribution of the *GPE* or
149 sub-lithospheric dynamics to the overall stress field (Flesch, Haines & Holt 2001; Flesch et al.
150 2005; Ghosh et al. 2006; Copley et al. 2011). Yet, inferring a compensation depth where complex
151 collisional and subduction dynamics altered the usual crust-mantle-asthenosphere stratification
152 and imply vigorous sub-crustal motion is, at best, speculative. Moreover, neglecting depth-
153 dependent behaviour across southeast Asia may be inappropriate if the lower crust beneath Tibet
154 is weaker relative to the upper crust (Nelson et al. 1996; Xu et al. 2007). More recent models
155 accounting for lower crustal channel flow relative to the upper crust have shown that depth-
156 dependent rheological variations affect the surface strain and topography consistently with the
157 observed surface structures and kinematics (Willett et al. 1993; Royden 1996; Clark & Royden
158 2000; Beaumont et al. 2004; Lechman et al. 2011; Capitanio 2014; Capitanio et al. 2015).
159 However, incomplete constraints on rheological variations of crustal and mantle rocks associated
160 with changes of the chemical, thermal and pressure conditions at depth throughout the collisional
161 and subduction history of the southeast Asia remain a substantial limitation for depth-dependent
162 strain predictions. The relative contributions from each deformation mechanism to the total strain

163 evolution remain elusive. Model formulations that account simultaneously for multiple
164 deformation mechanisms are therefore desirable for the aim of quantitatively characterizing the
165 overall deformation pattern or recognize additional potential contributions to the total strain.

166 **2. Methods**

167 In this section we outline the modeling approach and the methods of analysis, while more
168 details about the numerical solutions can be found in Gerya & Yuen (2007) and Gerya (2010).
169 The setup and boundary conditions (Fig. 2) as well as the parametric study involving the
170 numerical experiments discussed in this manuscript are also described in detail in Sternai et al.
171 2014, where we focused on extrusion dynamics rather than the forcing on the topography. Here,
172 we report the aspects that are more relevant to this study, a summary of the material properties
173 (Table 1) and videos showing the evolution of the reference model run (Videos S1-S4).

174 **2.1 Numerical modeling approach and reference setup**

175 We simulate the geodynamic evolution of a convergent collision-subduction system
176 characterized by along-strike differential slab kinematics constrained by the wealth of
177 observations from southeast Asia through self-consistent high-resolution 3D numerical thermo-
178 mechanical modeling. We use the numerical model I3ELVIS to solve the 3D momentum,
179 continuity and energy equations with the finite differences method, accounting for depth-
180 dependent non-Newtonian visco-plastic crustal and mantle rheologies (Gerya 2010). The
181 numerical setup and boundary conditions of the reference model are specified in Fig. 2;
182 additional simulations, whose setup is similar to that of the reference model (differences are
183 mentioned explicitly), are described in section 3.2. We imposed the upper plate crust as 35 km
184 thick where z (i.e., the along-strike direction) ≤ 490 km and 45 km thick where $z > 510$ km (linear

185 interpolation in between). The initial thermal structure of all continental plates is laterally
186 uniform with 0 °C at the surface and 1300 °C at 90, 140 and 150 km depth for the upper plate,
187 indenter and backstop continent, respectively. The oceanic domain, characterised by a trench-
188 parallel weak fracture zone to initiate subduction and a trench-perpendicular weak fracture zone
189 to allow for slab tearing, separates the three continental plates. The thermal structure of the
190 oceanic lithosphere is that of a half-space cooling age of 120 Ma (e.g., Turcotte & Schubert,
191 2002). Uniform and constant in time x -parallel velocities equal to $\sim 2 \text{ cm a}^{-1}$ (convergence) are
192 imposed to the $x = 2000 \text{ km}$ boundary.

193 Material properties, initially distributed on ~ 130 million randomly distributed Lagrangian
194 markers, are advected through the marker-in-cell technique (e.g., Gerya, 2010) accordingly to the
195 computed velocity field and a fourth-order Runge-Kutta scheme. Advected properties are then
196 interpolated by weighted distance averaging on a regularly spaced Eulerian grid (501 by 165 by
197 197 nodes to discretize the 2000 by 328 by 1000 km model domain – i.e., resolution of $\sim 4, 2$ and
198 5 km – in the x, y and z dimension, respectively). This operation enables the integration of
199 stresses or other parameter estimates (as described in the following section) to remain unaffected
200 by numerical diffusion of sharp gradients.

201 Predictions from multi-layered numerical models are particularly affected by rheological
202 parameters that cannot be measured directly (the limitation, however, is common to all
203 formulations), but can be constrained by laboratory experiments (Ranalli 1995) and observables
204 (England et al. 1985; Baumann & Kaus 2015). The ability to account simultaneously for multiple
205 deformation mechanisms (i.e., thickening/thinning, buckling, viscous flow, etc.) and geodynamic
206 processes (i.e., continental collision and ocean-continent subduction) and quantify their relative

207 contributions to the overall strain and topographic evolution without being tied to depth-averaged
208 values (as, for example, with thin viscous sheet-like models) enables to identify potential sub-
209 crustal forcing. Complex mantle-crustal interactions in collision or subduction systems, with
210 several implications for the large-scale tectonics and topography, have been demonstrated
211 through 2D multi-layered numerical models (Willett et al. 1993; Beaumont et al. 2004; Faccenda
212 et al. 2009; Nikolaeva et al. 2010). Solving also for the lateral dimension provides us with the
213 possibility to account for along-strike rheological variations, which are inescapable across
214 coupled collision-subduction systems and established along, for instance, the India-Eurasia and
215 Sunda margins. The focus is put on the potential contribution from the resulting asthenospheric
216 return flow to the surface topography and tectonics at the collision-subduction transition zone.
217 We compare the inferred tectonics and topography to those observed across southeast Asia during
218 the Cenozoic and find a certain degree of similarity. However, a number of differences between
219 our model setups and the natural setting can be readily recognized (e.g., uniform convergence
220 velocity in space and time, somewhat smaller scale, simplification of the plate geometry, etc.)
221 and we do not claim that our numerical model is representative of the entire subduction and
222 collisional history of the India-Eurasia and Sunda margins. We further stress that assessments as
223 to the causes of differential along-strike slab kinematics in southeast Asia are beyond the purpose
224 of this study. As such, our work is primarily meant to test the hypothesis that asthenospheric
225 dynamics across coupled collision-subduction systems can affect the surface evolution rather than
226 produce a realistic representation of the Cenozoic history of the India-Eurasia and Sunda
227 margins.

228 **2.2 Analytical investigation of the numerical results**

229 Results are analysed in terms of *GPE* and lithospheric strain and stress variations quantified

230 through depth-integration (depth-integrated values are equal to depth-averaged values multiplied
 231 by the corresponding thickness) across the model domain. The force balance equation,
 232 $\partial\sigma_{ij}/\partial x_j = -\rho g_i$, where σ_{ij} is the total stress tensor, x_j is the j^{th} coordinate axis, ρ is the density
 233 and g_i is the i^{th} component of gravity, can be used to describe the deformation of a continuous
 234 lithosphere. In the above equation, i denotes x , y and z coordinate axes and the repeated index j
 235 represents the summation over x , y and z (we use a right-handed coordinate system where x , y and
 236 z point south, down and east, respectively). If the horizontal length scale is large in comparison
 237 with the thickness of the lithosphere and local density and elevation contrasts are isostatically
 238 compensated, the force balance equation implies that

$$239 \quad \sigma_{yy}(y) = -g \int_{s.t.}^y \rho(y') dy' \quad (1)$$

240 where **s. t.** stands for surface topography, stating that the weight per unit area of any column of
 241 rock is supported by the vertical normal stress on its base, $\sigma_{yy}(y)$. Depth-integration of eq. (1)
 242 from the surface topography to the reference level (or compensation depth, given the assumption
 243 of isostatic equilibrium), L , leads to the definition of **GPE** as the negative depth-integrated
 244 vertical normal stress (Fleitout & Froidevaux 1983)

$$245 \quad \bar{\sigma}_{yy} = \int_{s.t.}^L (L - y) \rho(y) g dy = \text{GPE} \quad (2)$$

246 the over bar representing depth-integration. As previously described by e.g., England &
 247 McKenzie (1983), England & Molnar (1997), Flesch et al. (2001) or Ghosh et al. (2009),
 248 substituting into the horizontal (i.e., x - z plane) force balance equation the relationship
 249 $\tau_{ij} = \sigma_{ij} - \delta_{ij} \frac{1}{3} \sigma_{kk}$ (where τ_{ij} is the deviatoric stress tensor, δ_{ij} is the Kronecker delta and $\frac{1}{3} \sigma_{kk}$
 250 is the negative of pressure) and integrating at depth from the surface topography to the reference

251 level L leads to

$$252 \quad \partial \bar{\tau}_{xx} / \partial x - \partial \bar{\tau}_{yy} / \partial x + \partial \bar{\tau}_{xz} / \partial z = -\partial \bar{\sigma}_{yy} / \partial x + \tau_{xy}(L)$$

$$253 \quad \partial \bar{\tau}_{xx} / \partial x + \partial \bar{\tau}_{zz} / \partial z - \partial \bar{\tau}_{yy} / \partial z = -\partial \bar{\sigma}_{yy} / \partial z + \tau_{zy}(L) \quad (3)$$

254 Horizontal gradients of the depth-integrated deviatoric stress required to deform the lithosphere
 255 (i.e., the left-hand side of eq. 3) are balanced by horizontal gradients of **GPE** and the shear
 256 traction applied at the reference level by the underlying dynamics (i.e., the right-hand side of eq.
 257 3).

258 We infer the contributions from crustal and sub-crustal dynamics to the surface evolution
 259 across the model domain by comparing the shear stress applied by the mantle flow associated
 260 with differential along-strike slab kinematics to the base of the crust and horizontal gradients of
 261 **GPE** determined by depth-integration throughout the crustal thickness (i.e., the reference level, L ,
 262 corresponds to the base of the crust). This comparison also provides insights on the discrepancy
 263 between collisional strain pattern predictions assuming depth-independent behaviour and
 264 neglecting basal shear tractions (England & Houseman 1986; England & Molnar 1997) and those
 265 from a multi-layered rheology that accounts for the crustal and mantle dynamics arising from
 266 joint collision and subduction forcing. The choice of the base of the crust as reference level
 267 implies isostatic compensation at this depth. However, because vertical deviatoric traction arises
 268 from the mantle dynamics, a certain departure from the isostatic equilibrium is expected. The
 269 constitutive law

$$270 \quad \tau_{ij} = H \dot{\epsilon}_{ij}^{1/n-1} \dot{\epsilon}_{ij} \quad (4)$$

271 where $\dot{\epsilon} = \frac{1}{2} \left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right)$ (v_{ij} being the velocity tensor) is the strain rate tensor, $\epsilon'_{II} = \sqrt{1/2 \epsilon'_{ij} \epsilon'_{ij}}$ is
 272 the second invariant of the strain rate tensor and H is a rheological coefficient, expresses the
 273 rheology of a “power law” fluid whose effective viscosity, $\eta = 1/2H \epsilon'_{II}^{1/n-1}$, is determined
 274 through experimental parameterization of common crustal and mantle rocks coefficients (Ranalli
 275 1995). The vertical deviatoric stress at depth, $\tau_{yy}(y)$, determined via eq. 4, further elucidates the
 276 contribution from the sub-surface dynamics to the lithospheric strain and surface topography in
 277 terms of dynamic upward or downward deviatoric stress. Horizontal variations of $\tau_{yy}(y)$ provide
 278 information on the degree of isostatic compensation and therefore on the suitability of the choice
 279 of base of the crust, or any other depth, as compensation depth.

280 As a post-processing operation, we apply the above analytical relationships to the values
 281 interpolated on the numerical Eulerian grid (i.e., after application of the fourth-order Runge-Kutta
 282 advection scheme). The GPE , in particular, is computed from such a discretized set of values as
 283 the depth-averaged vertical normal stress at nodes between (including those at) the surface
 284 topography and base of the crust multiplied by the corresponding crustal thickness. This provides
 285 a discrete bi-dimensional grid map of GPE values from which gradients in the x and z (i.e.,
 286 horizontal) directions are derived. Horizontal GPE gradients are then compared to the shear stress
 287 as advected at the lowermost crustal Eulerian nodes.

288 **2.3 Testing the analysis**

289 As a test of our analysis, we consider a three-layer model configuration including
 290 lithospheric (crust and mantle lithosphere) and asthenospheric materials (Fig. 3a), based on
 291 experiments presented by Schmalholz et al. 2014. The horizontal boundary between the mantle
 292 and asthenosphere is at 120 km depth below the lowlands, and the asthenospheric layer extends

293 to a depth of 200 km. While the top surface of the lithosphere is calculated as an internal free
294 surface through a 15 km thick layer of “sticky air” and an “infinity-like” condition is applied to
295 the bottom boundary (Gerya 2010), free-slip conditions are imposed on all lateral boundaries,
296 which do not move (i.e., no imposed shortening or stretching).

297 Because this model is let to evolve under the only action of gravity, the flow and
298 topography (Fig. 3b) within the model domain results only from lateral variations of *GPE* as
299 dictated by the initial variations of the crustal thickness. Thus, the estimated *GPE* variations
300 obtained by applying equation 2 to the set of discrete values on the numerical Eulerian grid (that
301 we compare to the analytical solution of equation 2 in Fig. 3c) would explain the topography
302 once the isostatic equilibrium is reached. This test also validates that, when shear tractions at the
303 base of the crust (also shown in Fig. 3c) are negligible, the thin viscous sheet approximation
304 reproduces reasonably well the dynamics of multi-layered numerical models, as previously
305 pointed out by, for instance, Schmalholz et al. 2014.

306 **3. Results**

307 **3.1 Reference model**

308 **3.1.1 Description of the numerical results**

309 Like in southeast Asia, the geodynamic evolution of our model is characterized by pre-
310 collisional oceanic subduction along the entire overriding plate margin, followed by simultaneous
311 continental indentation and protracted subduction at the side of the collisional domain (videoS1-
312 videoS4). Increased buoyancy across a thicker portion of the overriding plate crust rise the
313 surface topography to ~3 km elevation prior to the onset of collision, which is in agreement with

314 geological observations (Murphy et al. 1997; Kapp et al. 2007). The imposed crustal thickness
315 variations result in a basal layer of partially molten, less viscous material below the collisional
316 margin, which is also consistent with observations (Nelson et al. 1996; Xu et al. 2007). Following
317 the onset of continental indentation, a plateau with a similar morphology to that of the current
318 eastern half of Tibet is generated through thrusting along the collisional domain (Fig. 1b and 4a).
319 Syn-collisional continental extrusion toward the subduction zone takes place through trans-
320 tensional deformation (Fig. 4b), with a resulting distribution of elevations across the collision-
321 subduction transition zone to a first order comparable to that currently present east and southeast
322 of Tibet (Fig. 1b and 4a).

323 At depth, an ascending/toroidal asthenospheric return flow develops below the upper plate
324 and through the opening slab window triggered by the fast descent of the pre-collisional slab and
325 enhanced by the increasing topographic weight during continental collision. The sub-collisional
326 slab then hangs almost vertically during continental extrusion and rollback of the neighbouring
327 slab (Fig. 4c). The rheological stratification of the upper continental plate is essentially
328 unaffected by this early subduction event and the associated asthenospheric flow, thus tractions
329 due to the mantle flow are imposed at the base of the lithosphere. The post-collisional ocean-
330 continent subduction event is conditioned by the precedent flow of hot asthenosphere. The major
331 effect is that the upwelling asthenosphere warmed the incipiently subducting lithosphere close to
332 the slab tear, enhancing slab rollback and trench retreat and contributing to a further vigorous
333 lateral/upwelling asthenospheric flow of up to ~ 12 and ~ 9 cm a^{-1} in the horizontal and vertical
334 direction, respectively, due to differential along-strike slab kinematics at the collision-subduction
335 transition zone (Fig. 4 and 6). Such an asthenospheric flow is able to thermally erode the upper
336 plate mantle lithosphere below the back-arc extensional domain. During advanced stages of the

337 model run, the rheological stratification of the upper plate in the back-arc extensional domain is
338 thus characterised by the absence of a rigid lithospheric mantle, which translates into shear
339 stresses of up to ~100 MPa applied directly to the base of the crust by the asthenospheric flow
340 (Fig. 7a,b).

341 Amongst the many limitations that our reference model is subject to, probably the most
342 important is that free-slip back and front boundaries imply a symmetric collision-subduction
343 setting, while the India-Eurasia-Sunda-Western Pacific setting is inherently asymmetric, with
344 major subduction zones and predominant extrusion towards the east. However, at the stage of the
345 model run shown by, for instance, Fig. 4, a rotational trajectory of the surface particles around the
346 collision-subduction transition zone is the only possible given that the continental indenter
347 prevents the material from moving parallel to the x-axis in the collisional domain, forcing the
348 flow towards the extending back-arc domain. An imposed influx or outflux across the $z = 0$ km or
349 $z = 1000$ km boundaries would modify the geometry of such rotational motion thereby better
350 reproducing, at least to some extent, the natural complexity. However, in this context, we favour
351 a simplifying approach in which this additional complexity is neglected.

352 **3.1.2 Results from the analytical investigation**

353 As should be expected if the regional instantaneous deformation follows eqs. 1-4, *GPE*
354 from our numerical model reaches the highest magnitude in the plateau and gradually decrease
355 across the collision-subduction transition zone to reach minimum values in the back-arc
356 extensional domain (Fig. 4a). There is, however, a significant discrepancy between the
357 topographic and *GPE* gradients across the extruded terrains arising because of the deviatoric
358 shear tractions applied by the mantle flow to the base of the crust and the simplifying assumption

359 of isostatic compensation at this depth. We show in Fig. 5 and 7a,b that horizontal *GPE* gradients
 360 and deviatoric shear stresses at the base of the crust are similar in magnitude across both the
 361 plateau proper, the plateau margins and the extruded terrains. They therefore exert a similar
 362 control on the overall surface strain and elevations. Also noteworthy is that higher basal shear
 363 stresses (and horizontal *GPE* gradients) are found toward the plateau margins and in the back arc
 364 domain, where the most topographic changes in space and time are expected, while they assume
 365 relatively low values beneath the elevated terrains (Fig. 5). High basal shear stresses at the
 366 plateau margins imply a mechanical coupling between the crust and mantle and an active
 367 contribution from the mantle flow to the crustal deformation. Similar mantle and crustal
 368 (especially lower crustal) flow patterns below the plateau proper and across the extruded terrains
 369 (Fig. 6a) associated with relatively low basal shear stresses (Fig. 5) suggest that these layers flow
 370 jointly in response to a common driver: differential along strike slab kinematics.

371 The fit between predictions of topography from our model and those assuming isostatic
 372 equilibrium and approximating the lithosphere to a uniform viscous sheet with an average density
 373 of $\sim 3000 \text{ kg/m}^3$ is minimized for compensation depths of $\sim 150 \text{ km}$ or higher, but still shows
 374 considerable differences (Fig. 7c,d, see the blue and red profiles). A similar outcome arises from
 375 the comparison between the topography predicted by our multi-layered model (i.e., accounting
 376 for the static and dynamic contributions to the surface elevations) and the isostatically-balanced
 377 topography resulting from the inferred lithospheric and asthenospheric
 378 structures, $Y_{iso} = \frac{(\rho_a - \rho_c) l_c}{\rho_a} + \frac{(\rho_a - \rho_m) l_m}{\rho_a}$ where ρ_c , ρ_m and ρ_a are the average crustal, mantle
 379 lithosphere and asthenospheric densities, respectively, and l_c and l_m are the crustal and mantle
 380 thickness (Fig. 7c,d see the blue and green profiles). In both cases, an overall agreement between
 381 elevation predictions in the plateau and a considerable discrepancy between the topographic

382 estimates across the extruded terrains and in the proximity of the subduction zone can be
383 observed, suggesting that dynamic contributions to the surface elevations outside of the
384 collisional domain cannot be neglected.

385 In our numerical experiment, continental extrusion is enhanced by the suction exerted by
386 self-consistent slab rollback in the subduction domain and, to a minor extent, by the imposed
387 convergence. The balance between these driving factors and the resulting forcing from the
388 asthenospheric flow to the horizontal crustal strain (that we do not investigate here, but have been
389 addressed in Sternai et al. 2014) determines whether continental extrusion is accompanied by
390 surface extension or shortening. In both cases the horizontal crustal velocity field is characterized
391 by rotational trajectories (rigid dislocation or viscous flow of the upper and lower crust,
392 respectively) with low angle to the lateral asthenospheric return flow generated by differential
393 along-strike slab kinematics at the collision-subduction transition zone (Figs. 4c and 6a). The
394 suction exerted by subduction of the oceanic mantle lithosphere, however, implies an upwelling
395 component of motion in the asthenospheric return flow and a down-welling component of motion
396 in the overriding plate lower crust across the extruded terrains, resulting in a modification of the
397 initial rheological stratification due to thermal erosion and delamination of the upper plate mantle
398 lithosphere (Fig. 6). In Figure 6b we also show that, unlike in the collisional domain where
399 deviatoric upward stresses responsible for the plateau uplift are ascribed to upper crustal levels,
400 the upwelling asthenosphere provides most of the upward deviatoric traction supporting the
401 topography of the extruded terrains.

402 **3.2 The role of convergence velocity, crustal thickness and geothermal gradient**

403 We performed an additional numerical experiment with x-parallel velocities imposed to

404 the $x=2000$ km boundary of $\sim 6 \text{ cm a}^{-1}$ (i.e., three time faster convergence rates than those of the
405 reference model). While the overall geodynamic evolution (i.e., pre-collisional oceanic
406 subduction along the entire overriding plate margin, subsequent continental indentation and
407 protracted subduction) is similar to that of the reference model, the principal effect of increased
408 convergence velocities is that of reducing slab rollback and trench retreat (Fig. 8), as also
409 recognized by previous studies (e.g., Schellart, 2005). The amount of upper plate extension and
410 the asthenospheric return flow are thus reduced compared to those in the reference model at
411 similar modelled times. As expected, because the asthenospheric forcing on the surface strain is
412 reduced, *GPE* and topographic trends across the upper plate are similar.

413 Nikolaeva and co-workers presented an extensive investigation of the parameters that
414 codetermine the initiation and evolution of an ocean-continent subduction on a 2D version of the
415 numerical model used in this study (Nikolaeva et al. 2010). They found that increased thickness
416 of the upper plate crust enhances subduction rates and leads to a faster geodynamic evolution. On
417 the contrary, an increased upper plate lithospheric thickness leads to slower subduction rates and
418 overall evolution. In the light of this previous work, we ran an additional simulation in which the
419 initial $1300 \text{ }^\circ\text{C}$ isotherm is at 70 km (i.e., 20 km higher than in the reference model) depth and the
420 crust measures 30 km where $z \leq 490 \text{ km}$ and 40 km where $z > 510 \text{ km}$, with linear interpolation in
421 between (i.e., 5 km thinner than in the reference model). Again, the overall geodynamic evolution
422 is similar to that of the reference model, but some differences in the timing and amount of surface
423 strain can be recognized (Fig. 9). A warmer lithosphere implies particularly fast slab rollback and
424 asthenospheric return flow (Nikolaeva et al. 2010) while a thinner crust implies an overall lower
425 topography across the upper plate. As a result, the amount of upper plate extrusion and trench
426 retreat is larger than in the reference model at similar modelled times. A particularly fast

427 asthenospheric return flow (Fig. 9c) produces similar effects on the upper plate strain and
428 topography to those observed in the reference model (Fig. 4). Also noteworthy is that the amount
429 of strike-slip deformation dominates on normal faulting in a warmer environment characterized
430 by a thinner crust.

431 **4. Discussion and implications**

432 A considerable discrepancy between horizontal *GPE* gradients and topography across
433 southeast Asia (England & Molnar 1997), to a first order comparable to our model predictions
434 (Fig. 4a), questions isostatic equilibrium outside Tibet. Dynamic contributions to the regional
435 elevations in eastern Tibet and Yunnan have been ascribed to an eastward flow in a mid-crustal
436 channel below Tibet diverted to the north and south by the stronger Sichuan craton (Clark &
437 Royden 2000; Clark et al. 2005) and can, to some extent, explain this discrepancy. This
438 interpretation, however, neglects relatively short wavelengths (i.e., a few hundred kilometres)
439 gravity perturbations (Jin et al. 1994; Balmino et al. 2011) suggesting the presence of mass
440 anomalies, in addition to the dynamic flow, involved in the support of the topography. Because
441 the crust along portions of south-eastern Tibet has been stretching since at least ~4Ma (possibly
442 for more than 10 Ma locally) (Williams et al. 2001; Blisniuk et al. 2001; Zhang et al. 2004;
443 Meade 2007; Royden et al. 2008) and several detachment-like extensional faults and back-arc
444 basins, unlikely related solely to gravitational collapse, can be observed in the Indochina,
445 Yunnan, Sunda and east China provinces, one may assume that the Sunda and western Pacific
446 subduction zones provided an active contribution to southeast Asian tectonics from the Eocene to
447 the present (Jolivet et al. 1990; van der Hilst & Seno 1993; Leloup et al. 1995; Northrup et al.
448 1995; Jolivet et al. 1999; Jolivet et al. 2001; Fournier et al. 2004; Hall & Morley 2004; Schellart
449 and Lister 2005). An alternative interpretation may, therefore, involve the lateral/upwelling

450 asthenospheric return flow in response to northward motion of the Indian slab during south- to
451 south-westward rollback or stable subduction along the Sunda region throughout the Cenozoic
452 (Fig. 10).

453 The asthenospheric flow generated by differential along-strike slab kinematics, consistent
454 with relatively short wavelengths gravity anomalies across the Eurasian plate (Jin et al. 1994;
455 Balmino et al. 2011), may provide support to the topography of the upper plate by modulating the
456 isostatic balance and applying dynamic upward deviatoric stresses (Figs. 4-7). Consistently with
457 this view, tomographic models show negative anomalies in seismic waves velocity across
458 southeast Asia (Káráson & van der Hilst 2000; Li et al. 2008; Replumaz et al. 2013; Schaeffer &
459 Lebedev 2013; Auer et al. 2014) suggesting the presence of hot asthenospheric material at
460 relatively low depths (Fig. 1c). In addition, the absence of foredeep basins of Cenozoic age along
461 the eastern margin of the plateau (Royden et al. 1997; Royden et al. 2008), very low crustal
462 seismic velocities (Meltzer et al. 1998; Yang et al. 2010), rapid exhumation and high-temperature
463 metamorphism of Pleistocene age (Burg et al. 1998; Burg et al. 1997; Liu & Zhong 1997) suggest
464 a shallow brittle-ductile transition and rapid advection through a primary flow path into the
465 massifs from depth rather than a shallow detachment related to lower crustal flow. Extreme
466 exhumation rates, in particular, have been related to either exceptionally high rates of rock uplift
467 owing to crustal-scale buckling (Burg et al. 1998) or a positive feedback among erosion, heat
468 advection, rock strength and deformation (Zeitler et al. 2001). Recent chronological constraints
469 and reconstructions of the former Yarlung-Tsangpo valley bottom (Wang et al. 2014), however,
470 relate high erosion rates to rapid tectonic rock uplift and disprove any effective erosional control
471 to the tectonic deformation. Based on our numerical results (Figs. 4-7), we propose that the
472 asthenospheric return flow in response to along-strike differential slab kinematics across the

473 collision-subduction transition contributes to such a tectonic upward push and sustains here the
474 high strain rates and topography. Noteworthy is that accounting for surface processes in our
475 experiment would lead to a coupling between erosion and crustal deformation (Avouac & Burov
476 1996; Zeitler et al. 2001), in turn enhancing localized rapid uplift along the edge of the plateau as
477 well as in the syntax region.

478 Testing further our modeling results by comparing predictions with seismic anisotropy
479 (Fig. 1c), currently the only technique at hand offering insights on the mantle strain, is beyond
480 our possibilities for it would require accounting for the elastic tensor during deformation in order
481 to convert the computed strain field into seismic anisotropy. However, an overall consistency
482 between the crustal strain and seismic anisotropy in south-eastern Tibet has been interpreted to
483 indicate coupling between mantle and crustal flow (Sol et al. 2007) (Fig. 6a). This observation
484 does not necessarily rule out the presence of a weaker lower crust relative to the upper crust, but
485 suggests that the lower crust is strong enough to transmit to upper crustal levels the stress arising
486 from the mantle dynamics. In particular, previous investigations based on the thin viscous sheet
487 formulation (Ghosh et al. 2006; Ghosh et al. 2009) demonstrate that deviatoric stresses at the base
488 of the lithosphere from large-scale mantle convection contribute as much as those related to *GPE*
489 gradients to the style and magnitude of the deformation worldwide, especially if the viscosity
490 contrast between the asthenosphere and the lithosphere is between 10^2 - 10^4 (Ghosh et al. 2008).
491 Accounting for a multi-layered rheology, our model shows that non-negligible deviatoric stresses
492 also arise from more local mantle dynamics (Figs. 4-7). The effectiveness of these stresses is
493 locally improved in our model by removal of the upper plate mantle lithosphere, in which case a
494 viscosity contrast between the asthenosphere and the lower crust of 10-100 at ~40-60 km depth
495 (Fig. 6b) is sufficient to produce deviatoric stresses comparable to those associated with

496 horizontal *GPE* gradients (Figs. 5 and 7b,c). Long-wavelength traction fields to the base of the
497 Indian plate generated by large-scale mantle convection due to subducted lithosphere are thought
498 to foster the northward motion of India into Eurasia (Ghosh et al. 2009; Alvarez 2010; Becker &
499 Faccenna 2011) but the opposite view, i.e., that basal traction might resist plate motion, has been
500 advocated as well (Copley et al. 2010). In any case, more local asthenospheric dynamics may
501 also influence the surface tectonics and topography across southeast Asia (Fig. 10), where long-
502 lived collision-subduction dynamics likely altered the natural lithospheric rheological
503 stratification.

504 Consistently with our numerical results, most Precambrian collisional orogens formed
505 atop of warmer mantle relative to the Cenozoic and involved high topographies owing to
506 protracted deformation over long periods (varying broadly between 50 and 200 Ma) with
507 homogeneous thickening by mass redistribution in the upper and lower crust (Taylor &
508 McLennan 1995; Windley 1995; Nironen 1997; Cagnard et al. 2006; Gerya 2014). We thus
509 speculate that the geodynamic significance of crustal deformation and topographic growth driven
510 by the asthenospheric flow is not peculiar to recent times, but also finds expression in
511 Precambrian orogenesis.

512 **5. Conclusions**

513 In conclusion, several aspects related to the deformation and topographic evolution east
514 and southeast of Tibet appear consistent with a forcing from the asthenospheric return flow
515 owing to differential slab kinematics across the collision-subduction transition zone, a
516 contribution that was neglected by previous models. While our experiment might exaggerate the
517 influence of the asthenospheric flow on the lithospheric deformation, especially if compared to

518 the present-day, the comparison between the observed and modelled surface kinematics is
519 striking (Figs. 1, 4, and 6). There is, in addition, geological evidence that the early phase of
520 extrusion of Indochina was affected by differential along-strike slab kinematics and the
521 associated mantle return flow. Such a mantle forcing is a non-exclusive alternative to previously
522 proposed causes of deformation and is untie to unrealistic assumptions such as those of depth-
523 independent behaviour, total isostatic compensation at crustal depths or a uniform and unaltered
524 lithospheric rheological stratification despite long-evolving coupled collision-subduction
525 dynamics. More comprehensive models of southeast Asia geodynamics, and continental
526 dynamics in general, should therefore account for sub-crustal dynamics.

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- 812

813 **Figure captions**

814 **Figure 1** (a) Tectonic map of southeast Asia and the Sunda and western Pacific subduction
 815 zones. 1: Indian (IND) continental plate; 2: Eurasian (EU) continental plate; 3: Actively opening

816 intra-continental rifts; 4: Pacific (PAC) oceanic plate. AFT: Altyn Tagh Fault; BB: Bohai Basin;
817 EHS: Eastern Himalayan Syntax; PBT: Pengguan and Beichuan Thrusts; PHS: Philippine Sea;
818 QLS: Quin Ling Fault; RRF: Red River Fault; SAS: Sumatra-Andaman (Sunda) subduction;
819 SCS: South China Sea; TF: Tanlu Fault. (b) Topography, GPS data (black arrows) from Wang et
820 al. (2001), Zhang et al. (2004) and Gan et al. (2007) and major tectonic structures. Arrows along
821 the strike-slip fault zones show the dominant sense of shear. Red and grey arrows along the Red
822 River Fault differentiate between the sense of shear during the middle- and late-Cenozoic,
823 respectively. (c) Average seismic wave velocity anomalies between 50 and 350km depth from
824 Schaeffer & Lebedev 2013 and seismic anisotropies (green markers) from Wüstefeld et al.
825 (2009). The low velocities beneath Tibet are due to its deep crustal root (which reaches to 80km
826 depth), while surface wave tomography suggests that a mantle lid with rather high seismic
827 velocities underlies most of Tibet (Priestley et al. 2008).

828

829 **Figure 2** 3D reference model setup with colours showing the initial rock type distribution: 1 –
830 continental crust; 2 – oceanic crust; 3 – lithospheric mantle; 4 – hydrated/serpentinized mantle
831 (initially imposed “weak fracture zone”). b) Initial model domain and location of the “weak
832 fracture zones” into the lithosphere to initialize subduction (z -parallel) and allow for differential
833 along-strike slab kinematics as described in Sternai et al. 2014. Other phases in (a) and (b) are cut
834 off for clarity. c) x - y viscosity profile (location shown by the dashed line in (a)) of the initial
835 model domain. Also shown in white are the initial 1300 °C, 900 °C and 500 °C isotherms. The
836 velocity boundary conditions are free slip at the top ($y = 0$ km) and at both the front and back
837 boundaries ($z = 0$ km and $z = 1000$ km). The left and right boundaries ($x = 0$ km and $x = 2000$ km)
838 use constant x -parallel velocities, which define the material influx. Global mass conservation is
839 ensured by material outflux through the lower permeable boundary ($y = 328$ km). The top surface

840 of the lithosphere is calculated dynamically as an internal free surface through a 12 km thick
 841 layer of “sticky air” (Gerya 2010).

842
 843 **Figure 3** a) x - y representation of the model setup of the test simulation following experiments
 844 described in Schmalholz et al., 2014 (parameters are the same along the z -axis). ρ_i is the material
 845 density. b) Plan view of a selected time step (modelled time ~ 4 Ma) of the test experiment
 846 showing the modelled topography (colours) and GPE (white isolines, units: N/m). (c) Profiles
 847 (the location is shown by the dashed line in (b)) of GPE calculated analytically from equation 2
 848 and from the numerical results as well as τ_{xy} at the base of the crust.

849
 850 **Figure 4** Plan view of a selected time step (modelled time ~ 15 Ma) of the numerical experiment
 851 showing (a) the modelled topography (colours) and GPE (black isolines, units: N/m) distribution,
 852 (b) the second invariant of the rate-of-strain tensor and major tectonic structures and (c) the slab
 853 geometry (visualized through an iso-viscosity surface equal to 10^{23} Pa s, color-coded by depth)
 854 and the velocity field in the crust (red arrows) and asthenosphere (blue arrows). The major
 855 subduction/thrust fronts are also shown in each panel. Dashed red and black lines represent the
 856 continental indenter and upper plate boundaries, respectively. The white dotted lines in (a) show
 857 the location of the profiles in Fig. 7.

858
 859 **Figure 5** Plan view of a selected time step (same as Fig. 4) showing (a) $\partial \bar{\sigma}_{yy} / \partial x$, (b) $\partial \bar{\sigma}_{yy} / \partial z$, (c)
 860 $\tau_{xy}(L)$, (d) $\tau_{zy}(L)$. Note that these quantities vary between the same order of magnitude, suggesting
 861 that horizontal gradients of GPE (i.e., $\partial \bar{\sigma}_{yy} / \partial x$ and $\partial \bar{\sigma}_{yy} / \partial z$) and deviatoric shear stresses at
 862 the base of the crust ($\tau_{xy}(L)$ and $\tau_{zy}(L)$) jointly contribute to the surface strain and elevation

863 variations. Major subduction/thrust fronts are also shown in each panel. Dashed red and black
 864 lines represent the continental indenter and upper plate boundaries, respectively. Also shown in
 865 black is the 1000 m topographic elevation isoline.

866
 867 **Figure 6** (a) Plan view of a selected time step (same as Fig. 4 and 5) showing the velocity field
 868 within the upper crust (left panel, red arrows), lower crust (central panel, green arrows) and
 869 asthenosphere (right panel, blue arrows). The thermal anomaly produced by the asthenospheric
 870 return flow in response to differential along-strike slab kinematics and responsible for thermal
 871 erosion of the overriding plate mantle lithosphere beneath the extruded terrains is shown on the
 872 right panel by the 1500 °C, 1300 °C and 900 °C isotherms at 100 km depth (black solid lines). (b)
 873 Vertical velocity (red), vertical deviatoric stress (blue) and viscosity (black) profiles at depth.
 874 Black (left panel) and yellow (right panel) crosses in (a) show the profile location. Horizontal
 875 dotted lines show the rheological stratification. Note the absence of mantle lithosphere to
 876 decouple the lower crust and asthenosphere and higher deviatoric upward stresses to an
 877 asthenospheric level in the profile within the extruded terrains (right panel).

878
 879 **Figure 7** a, b) Topographic profiles (location is shown in Fig. 4a) from our model (blue line),
 880 assuming isostatic equilibrium and approximating the lithosphere to a uniform thin viscous sheet
 881 (shortened to t.v.s. in the caption) with an imposed average density (red line), and assuming
 882 isostatic equilibrium but accounting for the density and lithospheric structure from our model
 883 (i.e., Y_{iso} , green line). The inset shows the L^2 -norm distance (i.e., fit integrated along the profile)
 884 between blue and red profiles for different imposed density and compensation depth values. The
 885 star shows the pair value used for the general plot. c, d) Profiles (same location as (a, b)) of the
 886 horizontal *GPE* gradients (i.e., $\partial\bar{\sigma}_{YY}/\partial X$ and $\partial\bar{\sigma}_{YY}/\partial Z$, blue lines) and deviatoric shear stress at

887 the base of the crust (i.e., $\tau_{XF}(L)$ and $\tau_{ZF}(L)$, red lines). Note that these quantities vary between
888 the same order of magnitude, suggesting that they jointly contribute to the surface strain and
889 elevation variations.

890
891 **Figure 8** Plan view of a selected time step (modelled time ~15 Ma) of a numerical experiment
892 similar to the reference model (Fig. 4) but with faster convergence rates ($\sim 6 \text{ cm a}^{-1}$). The figure is
893 showing (a) the modelled topography (colours) and GPE (black isolines, units: N/m) distribution,
894 (b) the second invariant of the rate-of-strain tensor and major tectonic structures and (c) the slab
895 geometry (visualized through an iso-viscosity surface equal to 10^{23} Pa s , color-coded by depth)
896 and the velocity field in the crust (red arrows) and asthenosphere (blue arrows). The major
897 subduction/thrust fronts are also shown in each panel. Dashed red and black lines represent the
898 continental indenter and upper plate boundaries, respectively.

899
900 **Figure 9** Plan view of a selected time step (modelled time ~15 Ma) of a numerical experiment
901 similar to the reference model (Fig. 4) but with different crustal and lithospheric thickness (see
902 text for details). The figure is showing (a) the modelled topography (colours) and GPE (black
903 isolines, units: N/m) distribution, (b) the second invariant of the rate-of-strain tensor and major
904 tectonic structures and (c) the slab geometry (visualized through an iso-viscosity surface equal to
905 10^{23} Pa s , color-coded by depth) and the velocity field in the crust (red arrows) and asthenosphere
906 (blue arrows). The major subduction/thrust fronts are also shown in each panel. Dashed red and
907 black lines represent the continental indenter and upper plate boundaries, respectively.

908
909 **Figure 10** Schematic representation of the proposed sub-crustal forcing to the surface tectonics
910 and topography of the southeast Asia (note that the aim here is to facilitate comprehension by

911 providing the readers with a visual representation of the proposed forcing and no physical
912 meaning is implied by this figure). The present-day topography is joined to our modeling results
913 (a different model time step with respect to that shown in Fig. 4 is displayed here) showing a
914 possible representation of the overall geodynamics during the Cenozoic. The slab is visualized
915 through an iso-viscosity surface equal to 10^{23} Pa s, while red and blue arrows show the crustal
916 and asthenospheric velocity field, respectively. The present-day location of the main strike-slip
917 fault zones and collisional/subduction front is also shown in black and reported on the model
918 results.
919

Material properties										
	k	ρ_0	C_p	E_a	V_a	n	η_0	H_r	Material	$\sin(\phi_{eff})$
	($W m^{-1} K^{-1}$)	($kg m^{-3}$)	($J kg^{-1} K^{-1}$)	($kJ mol^{-1}$)	($cm^3 mol^{-1}$)		($Pa^a s$)	($mW m^{-3}$)		
Sticky-air	20	1	100	0	0	1	1×10^{19}	0	Air	0
Water	20	1000	3330	0	0	1	1×10^{19}	0	Water	0
Sediment	$0.64+807$ $/(T+77)$	2600	1000	154	8	2.3	1.97×10^{17}	2	Wet Qz.	0.15
Crust (upper plate, backstop)	$0.64+807$ $/(T+77)$	2750	1000	154	8	2.3	1.17×10^{17}	2	Wet Qz.	0.15
Crust (indenter)	$1.18+807$ $/(T+77)$	2950	1000	238	8	3.2	4.8×10^{22}	2	Wet Qz.	0.15
Oc. Crust	$1.18+474$ $/(T+77)$	3000	1000	238	8	3.2	4.8×10^{22}	0.25	Wet Qz.	0.15
Mantle	$0.73+129$ $3/(T+77)$	3300	1000	532	8	3.5	3.98×10^{20}	0.02	Dry Ol.	0.6
Weak Zone	$0.73+129$ $3/(T+77)$	3300	1000	471	8	4	5×10^{20}	0.05	Wet Ol.	0

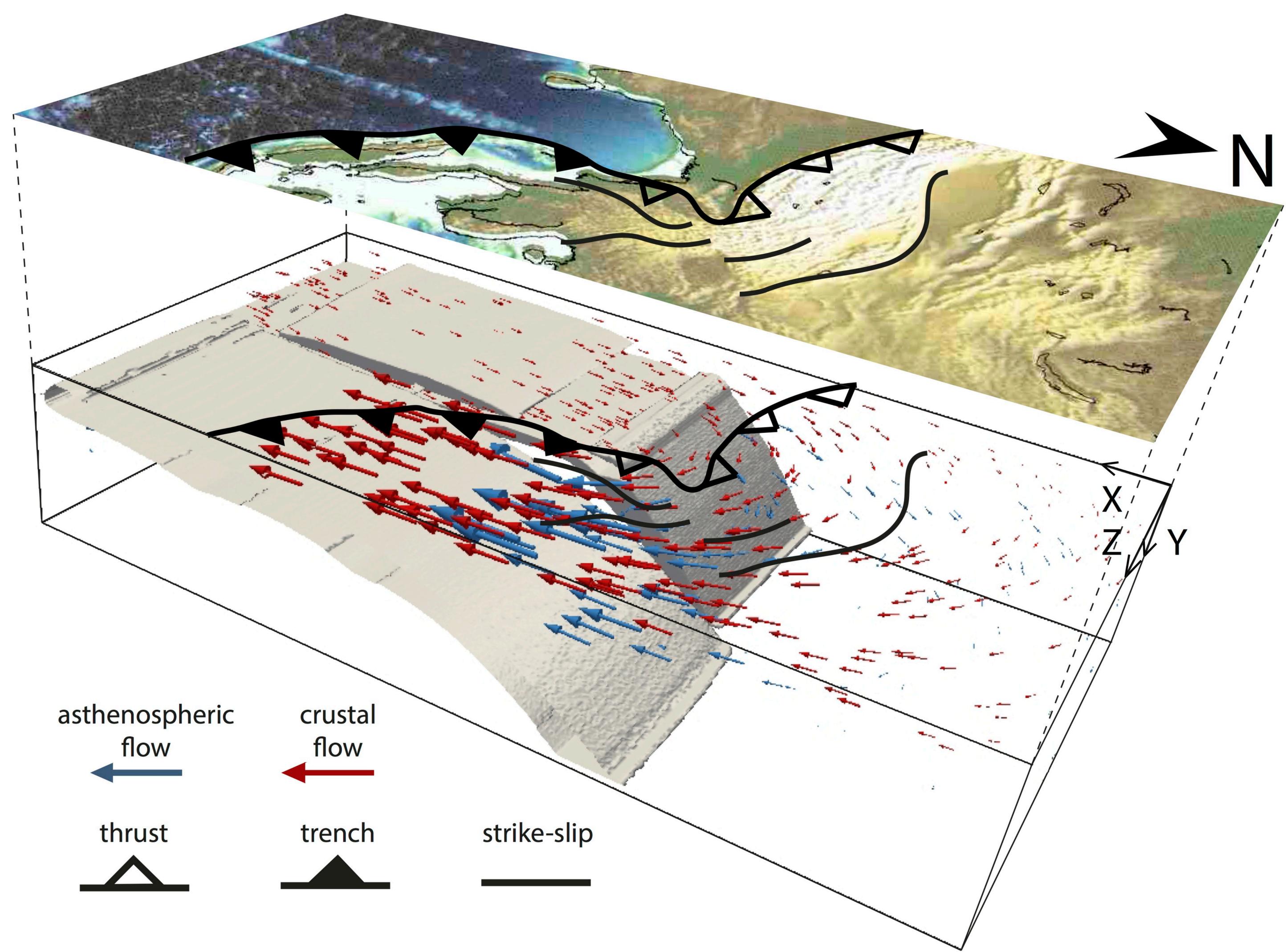
920 **Table 1:** Qz. and Ol. correspond to the abbreviations of Quartzite and Olivine. k denotes the
921 thermal conductivity, ρ_0 is the density, C_p is the specific heat capacity, E_a is the activation energy,
922 V_a is the activation volume, n is the stress exponent, η_0 is the reference viscosity, H_r is the

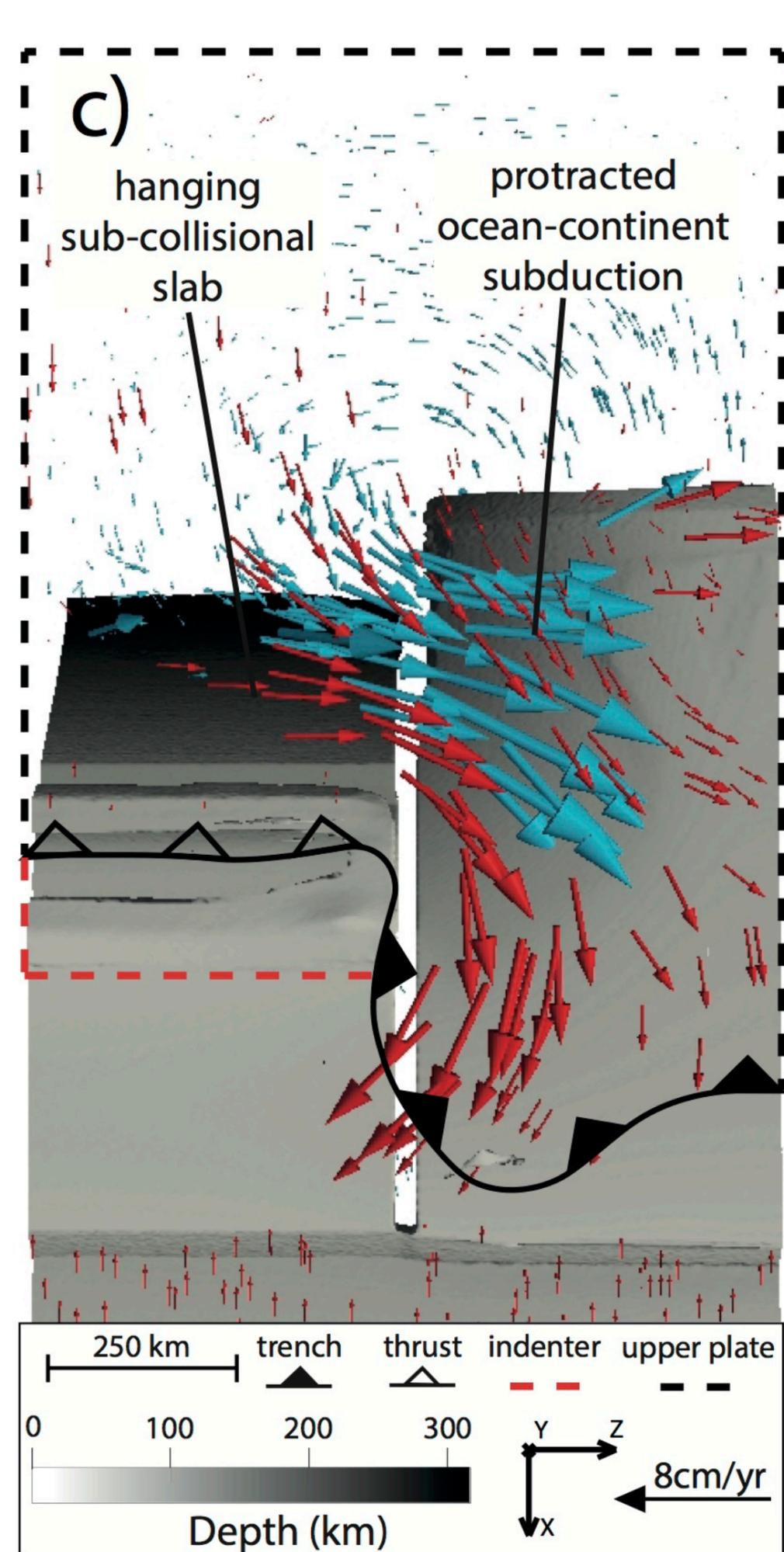
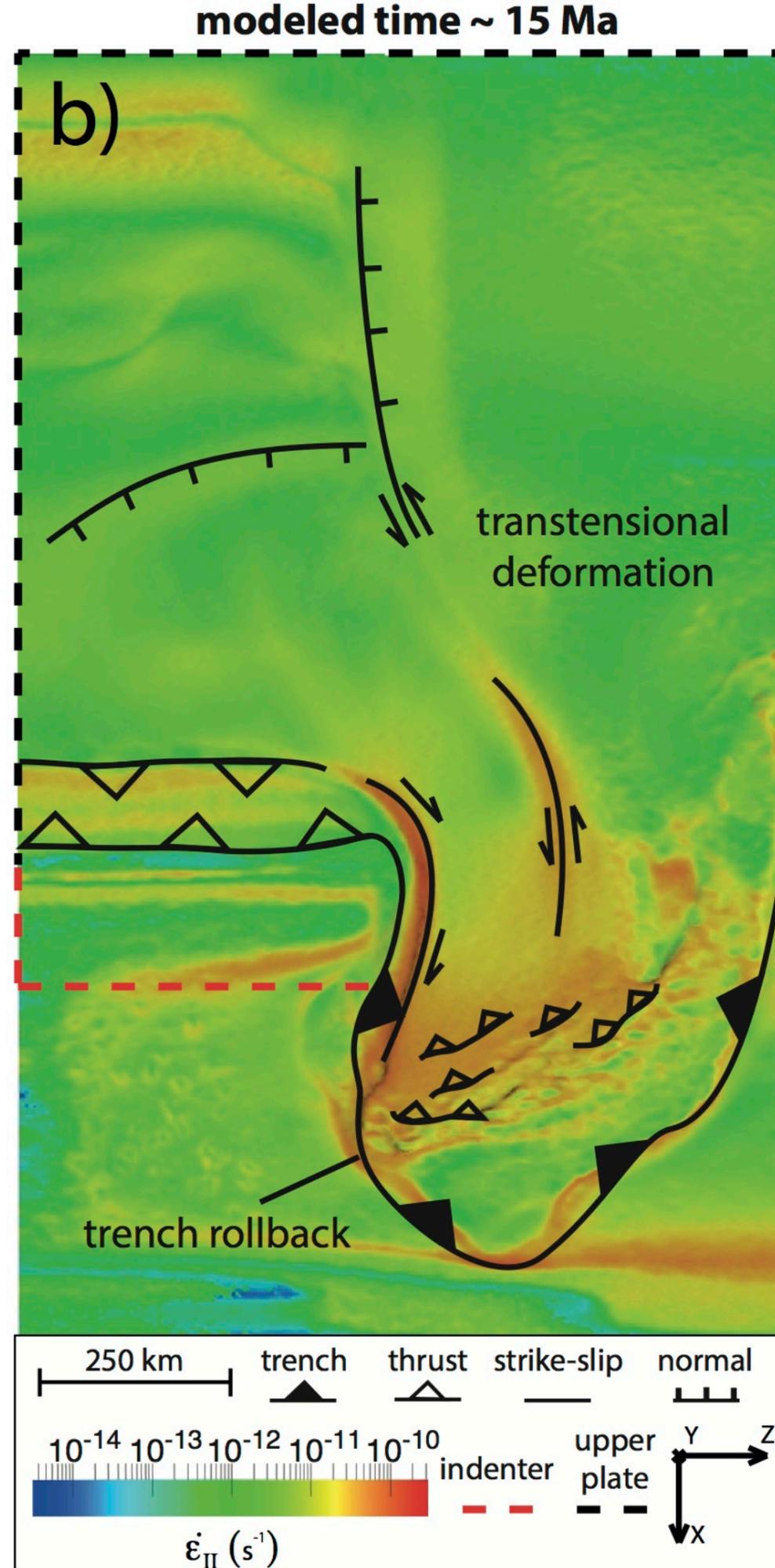
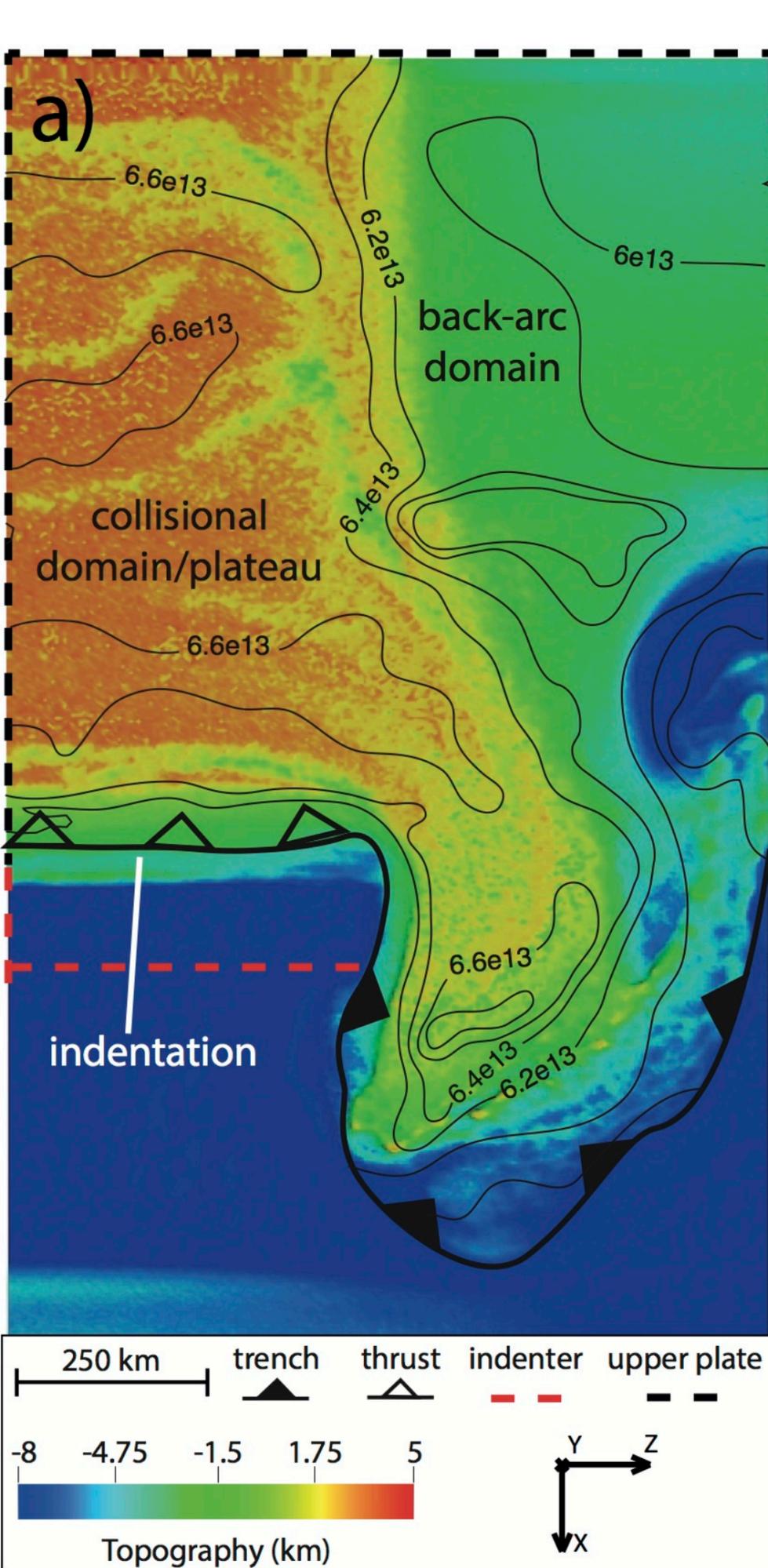
923 radiogenic heat production, ϕ_{eff} is the effective internal friction angle. Cohesion is 1 MPa for
924 each phase.

925

926

Accepted Manuscript





modeled time ~ 15 Ma

