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► **To cite this version:**

P. Pitard, A. Replumaz, C. Thieulot, M.-p. Doin. Modeling Deep Rooted Thrust Mechanism of Crustal Thickening in Eastern Tibet. *Geophysical Research Letters*, 2023, 50 (15), pp.e2023GL104134. 10.1029/2023GL104134 . hal-04260881

HAL Id: hal-04260881

<https://hal.science/hal-04260881>

Submitted on 26 Oct 2023

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1 **Modelling deep rooted thrust mechanism of crustal thickening**
2 **in Eastern Tibet**

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9
10 **Key Points:**

- 11 • 2-D numerical models of thrusts embedded in the high viscosity upper crust, to test
12 thermo-kinematic models based on thermochronology data.
- 13 • accommodation in the lower crust by ductile flow of the deformation induced by the
14 high angle thrust in the upper crust.
- 15 • predicting exhumation rates and subsidence patterns that are compatible with the
16 measured ones in Eastern Tibet.

17
18 **Abstract**

19 To test Eastern Tibet crustal thickening modes, we compare 2-D numerical models
20 of two emblematic end-member models, with either an obstacle in the low viscosity lower
21 crust or a thrust embedded in the high viscosity one. We show that the obstacle halts the
22 viscous lower crustal flow potentially initiated by the weight of the high Central Tibet,
23 generating a smooth exhumation gradient at the edge of the plateau, not observed in Eastern
24 Tibet. On the contrary, including a low viscosity discontinuity in the upper crust, mimicking
25 a shallow steep listric fault as inferred in the region, reproduces a sharper exhumation

26 profile, as constrained from thermo-kinematic inversions of thermochronological data, and
27 the lack of foreland basin, as observed in the field. Moreover, such fault drives deformation
28 throughout the entire crust, suggesting a deep crustal ductile shear zone limiting the more
29 ductile deformation in the lower crust although no discontinuity is imposed.

30

31 **Plain Language Summary**

32 The role of thrusting in crustal thickening during the formation of Tibet, the world's
33 largest and highest orogenic plateau, constitutes one of the main controversies in earth
34 sciences. In Eastern Tibet in particular, two end-members based on two contrasting
35 controversial hypotheses can be tested: the thickening is dominated either by the flow of the
36 lower Tibetan crust halted by the hard Sichuan craton, or by thrusting of the Tibetan upper
37 crust. Here, we present 2-D crustal numerical models of a shallow steep listric thrust (as
38 inferred in the region) embedded in the high viscosity upper crust, and we show that such
39 model reproduces the exhumation profile constrained from thermochronological data and
40 the lack of foreland basin observed in the field. Interestingly, we also show that such upper
41 crustal thrust drives upward the more ductile lower crust albeit no discontinuity is imposed.
42 On the contrary, by using a model driven by an overpressure in the lower crust, we show
43 that the obstacle halts the viscous lower crustal flow and generates a smooth exhumation
44 gradient at the edge of the plateau, not observed in Eastern Tibet.

45

46

47 **1. Introduction**

48 Despite decades of controversy, our understanding of the formation of the Tibetan
49 Plateau remains limited. The role of competing mechanisms, such as distributed crustal
50 thickening versus lateral propagation of thrust faulting at crustal or lithospheric scales, is
51 still poorly understood. Conceptual models explaining observations at the continental
52 scale are based on hypotheses that are hard to reconcile. On the one hand, buoyancy
53 forces associated with very thick crust and a low-viscosity channel in the Tibetan crust
54 may drive distributed outward flow of the partially molten middle crust towards the east
55 caused by the weight of the high Central Tibet, with entrenchment of the rivers within
56 relict penneplains contemporaneous of the uplift ("channel flow model", Royden et al.,
57 2001; Clark et al., 2005a). On the other hand, surface geological observations of fault
58 systems favor the role of thrust faulting in the upper crustal thickening, while the lower
59 crust subducts attached to the lithospheric mantle, with discrete outward propagation of
60 thrusts since the onset of collision at ~50 Ma, through strike-slip faults, such as the
61 XianShuiHe fault to the east ("stepwise model", Tapponnier et al., 2001; Jiao et al.,
62 2023). However, no numerical model considering the role of thrust faulting in the
63 interaction between the lower and upper crusts has yet been implemented in Eastern Tibet
64 in the stepwise model to be compared with the analytical dynamic pressure generated by
65 the lower crustal flow used as the driving load for the flexure equation to compute a
66 vertical deflection done for the channel flow model. Therefore, it remains difficult to test
67 the mechanical and rheological consistency, and the ability to explain observations, of
68 these end-member conceptual models.

69 In order to generate new insights in deformation modes in Tibet, we model the
70 mechanical behavior of the crust under boundary conditions corresponding to both end-
71 member models. We focus on the very contrasted Tibet eastern edge, with to the

72 northeast the Longmenshan thrust belt (LMTB) bounding the rigid Sichuan craton with a
73 steep topographic gradient, and to the southeast the Yalong thrust belt (YTB) crossing the
74 Yalong margin bounding the South China craton with a gentler topographic gradient.
75 This key location has long been studied to propose different models of Tibetan Plateau
76 formation (Fig. 1). In this region, a collection of multi-scale data is available that we use
77 to constrain the models, including a dense thermochronologic dataset thanks to numerous
78 granite outcrops in Eastern Tibet compared to mostly sedimentary rocks in Central Tibet
79 (e.g., Clark et al., 2005; Pitard et al., 2021). At depth, receiver function profiles (Robert et
80 al. 2010) and regional shear waves tomography images (Yao et al., 2008) show a sharp
81 ~20 km step of the Moho depth between the plateau (~60-63 km) and the Sichuan craton
82 (~40 km) across the LMTB, and a ~15 km smoother transition in crustal thickness across
83 the YTB between the plateau (~60-63 km) and the South China craton (45-50 km) (Fig.
84 1). In the channel flow model, mid-crustal flow is halted in the Longmenshan by the rigid
85 Sichuan craton, inducing topographic growth at its border as in the high-standing LMTB
86 (Fig. 1C). Across the Yunnan margin, flow is continuous in the absence of obstacle in the
87 lower crust, generating a smooth topographic gradient with no spatial discontinuity and
88 no thickening of the upper crust (Clark et al., 2005a). By contrast, in the stepwise model
89 of Tapponnier et al. (2001), discrete thrust faults of the YTB merge on an intracrustal
90 decollement, allowing the thickening of the upper crust and forming topographic steps
91 (Fig. 1B).

92 In this paper, we present 2-D numerical models of thrust embedded in the high viscosity
93 upper crust (model 1 under “lateral constant velocity” representing the eastward motion
94 of Tibet shown by the GPS velocity field, reproducing the “stepwise model” boundary
95 conditions), in which upper crustal thrust geometry is based upon thermo-kinematic
96 modelling of thermochronology data. We compare with models with no fault but

97 showcasing an obstacle in the lower crust (model 2 under “Poiseuille velocity”
98 representing the velocity field generating by the weight of the high Central Tibet in a
99 lower crustal channel, reproducing the "channel flow model" boundary conditions),
100 which setup is based upon analytical dynamic pressure result. We test the short-term
101 mechanical consistency of the fault geometry for varying fault and lower crust viscosity
102 to shed light on end-member mechanisms of Tibetan plateau formation. Such modelling
103 allows to include the localisation of deformation on faults in the long-term history of
104 building the Tibetan Plateau, to quantify the mechanisms of crustal thickening up to the
105 modern value of ~65 km (Fig. 1).

106
107 **2. Geological and geophysical constraints on the crustal structure of Tibet eastern**
108 **edge**

109 The LMTB at the western margin of the Sichuan basin (Fig. 1) is active, as attested by
110 major earthquakes, such as the 2008 Mw7.9 Wenchuan earthquake. During the earthquake,
111 thrusts ruptured with steep shallow ramps (up to 70°), rooting between 15 and 25 km-depth
112 on a sub-horizontal décollement zone at the brittle-ductile transition (e.g., Hubbard and Shaw,
113 2009; deMichele et al., 2010; Fielding et al., 2013). On the longer term, high exhumation rate
114 above the ramp and low above the décollement have been measured using thermochronology
115 (e.g., Li et al., 2023). It has been proposed that brittle deformation occurs in the thick upper
116 crust above the décollement, a Cretaceous passive margin with a thick sedimentary wedge,
117 while the thin lower crust remain mostly passive (e.g. Airaghi et al., 2018). It has also been
118 proposed that the brittle deformation observed in the upper crust has been induced by the
119 lower crust flow, above the location where the flow is deviated upward, with no foreland
120 flexural basin in the footwall (e.g. Burchfiel et al., 2008). Last, considering that the LMTB
121 coincides with the limit of a low shear velocity layer in the middle and lower crusts which

122 limits the southeastward extent of the thickened Tibetan crust, it has been proposed that
123 thrusts should extend in the lower crust (Robert et al., 2010) (Fig. 1C).

124 Across the Yalong margin, the YTB is present-day inactive, with no reverse focal
125 mechanism observed. On the longer term, the Muli thrust of the YTB shows a
126 thermochronology age zonation parallel to the fault, with Miocene ages (<10 Ma) close to the
127 thrust, rapidly becoming older away from it (Clark et al., 2005b; Pitard et al., 2021). Such
128 spatial distribution of the ages is best reproduced by a listric thrust geometry, at a velocity
129 rate of 0.26 km/Ma until 11.5 Ma then at a faster rate of 0.52 km/Ma, yielding since 50 Ma to
130 a total of ~15 km of exhumation along the uppermost crustal high-angle ramp and ~8 km
131 above the sub-horizontal deep décollement (Pitard et al., 2021) (Fig. 1E). The depth of the
132 intra-crustal décollement is not well constrained at depth greater than 20 km, as the
133 thermochronometers are only sensible to upper crustal thermal evolution. On the contrary, the
134 age zonation could not be reproduced only by the entrenchment of the Yalong River as
135 proposed in the channel flow model (Clark et al., 2005), with too old ZHe and AFT ages
136 predicted (Fig. 1E).

137

138 **3. Method: 2-D mechanical models of a viscous crust**

139 A 2-D code has been developed in which the Stokes equations are solved for an
140 incompressible Newtonian fluid on a Lagrangian grid (Pitard et al., 2023). We define a
141 rectangular box of 200 km × 50 km with four different zones for which we can assign
142 different values of viscosities and densities (Fig. 2). Two horizontal layers correspond to the
143 upper and lower crust. Viscosity is constant within the lower crust in model 1, while in model
144 2, the lower crust is separated in two blocks, with the high viscosity one on the right
145 mimicking a rigid obstacle. Viscosity within the upper crust is computed with a plastic yield
146 stress increasing with depth to simulate a more realistic crustal rheological profile

147 corresponding to a traditional model of a quartzo-feldspathic crust (Burov and Watt, 2006;
148 Chen and Molnar, 1983). It defines a high strength upper crust and a ductile lower crust with
149 analogy to the jam layer of a jelly sandwich (Burgmann and Dresen, 2008). In model 1, a
150 fault is mimicked in the upper crust by a thinly meshed constant-width zone of lower
151 viscosity (Fig. 2). Such low viscosity in the fault, three order of magnitude lower than for the
152 upper crust, simulates a weak fault with low shear stresses, often reproduced in model by
153 strain-softening (e.g., Huismans and Beaumont, 2003; Buiter et al., 2016).

154 The triangle mesh is deforming at each time step. The time stepping is adaptive,
155 following the CFL condition (van Zelst et al., 2022). The boundary conditions of the model
156 are as follows: convergence is imposed on the left side of the box, with a constant horizontal
157 velocity (model 1, “stepwise model” boundary conditions) or a Poiseuille horizontal velocity
158 in the lower crust and zero velocity in the upper crust (model 2, “lower channel flow model”
159 boundary conditions). The right side is a closed boundary. Vertical velocity is allowed on
160 both sides with a zero shear stress boundary condition. The top boundary of the box is a free
161 surface, allowing the growth of topography, while the bottom boundary condition is no slip
162 with no vertical velocity.

163

164 **4. Results for endmember models: “lateral constant velocity” versus “Poiseuille** 165 **velocity”**

166 In the model 1 (stepwise model), to mechanically test the accommodation at depth by
167 ductile flow of the shallow kinematic velocity field induced by the high angle of the Muli
168 thrust, constant horizontal velocity condition of 0.05 cm/yr are imposed on the western side
169 of the model, representing the eastward motion of Tibet shown by the GPS velocity field
170 (e.g. Wang et al., 2001). To mimick a shallow high angle listric fault (70° dip), a
171 discontinuity with a lower viscosity of 10^{20} Pa.s is introduced in the upper crust with a

172 listric shape and a steep shallow dip. Under this constant lateral kinematical boundary
173 condition, the upper crust moves horizontally close to the moving boundary, and
174 progressively moves upward toward the fault, reaching a maximum vertical velocity of
175 ~ 0.06 cm/yr, i.e., ~ 0.6 km/Ma (Fig. 3B). It corresponds to an increase of the uplift of the
176 surface toward the fault, less than 10 meters 100 km away from the fault and more than 60
177 meters near the fault after 110 000 years (Fig. 3A). Interestingly, the lower crust moves
178 horizontally close to the moving boundary, but also progressively moves upward toward the
179 fault, even if the fault does not extend at depth. The velocity magnitude is higher in the
180 lower crust than in the upper one until the flow reaches the position of the fault, with a rapid
181 decrease of the horizontal velocity. A complete parametric study has been carried out (see
182 supplementary material), to systematically test the influence of each parameter (viscosities
183 within the fault and the lower crust, lower crustal thickness, fault width and geometry,
184 lateral velocity, background strain rate, cohesive strength, friction angle in the upper crust),
185 before choosing the parameter values of model 1, so that the velocity gradient toward the
186 fault is similar to the exhumation gradient obtained by thermo-kinematic modelling based
187 on thermochronology data (Fig. 1E). By changing the geometry of the fault, we show that
188 with a straight fault the surface velocity is much more symmetric, showing a strong
189 subsidence in the footwall (Fig. 3A, S5). It is worth noting that adding a rigid lower crust
190 obstacle with a listric upper crustal fault does not change the velocity profile (Fig. S9).
191 The fault is absent in model 2 (lower channel flow model) but a rigid block is present in the
192 lower crust, acting as a barrier to the flow, with a Poiseuille velocity boundary condition in
193 the lower crust (Fig. 3D). The maximum horizontal velocity imposed on the left boundary is
194 8 cm/yr (Clark and Royden, 2000), the optimum published value to reproduce the
195 topography of the LMTB (Fig. 1C). It has been obtained with analytical model of dynamic
196 overpressure within a lower crustal channel of 15 km width. Under these Poiseuille velocity

197 boundary conditions, but considering the upper and lower crust as incompressible
198 Newtonian fluid with different values of viscosities and densities (Fig. 2), the horizontal
199 lower crustal motion is transmitted to vertical motion in the upper crust (Fig. 3D). The
200 maximum vertical velocity is of 0.94 cm/yr on the left boundary of the box, leading after
201 ~100 000 years to ~1 km of positive deflection, decreasing to ~300m above the obstacle
202 edge, and no deflection above the rigid obstacle (Fig. 3D). A complete parametric study has
203 also been carried out (see supplementary material). In particular, we show that without the
204 obstacle in the lower crust, the velocity profile is much smoother, almost horizontal (Fig.
205 3C). Interestingly, with a fault in the upper crust, only a small local perturbation is observed
206 above the fault (Fig. 3C), corresponding to a minor flow in the upper crust driven by the
207 fault (S15).

208

209 **5. Discussion: comparison of model predictions with morphotectonic observations**

210 We compare the modelled deflection profiles with the exhumation profile constrained by
211 thermochronologic ages for the Muli thrust of the YTP crossing the Yalong margin (Fig.
212 1E). Such exhumation profile is more pertinent for a quantitative comparison with the
213 modelled deflection profile, as erosion that shaped the topography during and after the
214 tectonic uplift is not included in our models. Previously, only the topographic profiles have
215 been used, showing smoother topographic steps across the YTB than across the LMTB (fig.
216 1).

217 **5.1. Lower crust channel flow model**

218 In the channel flow model, no obstacle has been inferred for the Yalong margin and a
219 smooth topographic step has been calculated (Clark et al., 2005a). Our numerical results
220 showcase a very different deflection profile which is almost flat (Fig. 3C). Such profile is
221 not comparable with the sharp exhumation profile of the Muli thrust, with a sharp positive

222 deflection above the thrust (Fig. 1). Furthermore, in the Yalong margin region, the channel
223 flow scenario has been related to a regional uplift of low-relief surfaces where exhumation
224 is mainly driven by deeply entrenched river incision. By testing such scenario on the
225 complete thermochronologic dataset available for the Muli thrust (Pitard et al., 2021), we
226 show that incision alone is not reproducing the thermochronologic ages (Fig. 1E).

227 Nevertheless, assuming that the South China craton is comparable with the Sichuan
228 craton, as seen in tomography models (Yao et al., 2008), with a positive velocity anomaly in
229 the foreland of the Yalong margin (Fig. 4), we also compare the Yalong margin with the
230 model including an obstacle in the lower crust. In the channel flow model, such obstacle is
231 mimicking the Sichuan craton stopping the eastward weak lower crust flow below the
232 LongmenShan range (Clark and Royden, 2000), the dynamic pressure at the edge of the
233 obstacle generating an overpressure localizing the thrusts in the upper crust (Burchfiel et al.,
234 2008). Using the published strikingly high velocity calculated for the flow to fit the
235 topographic step observed in the LongmenShan range above the LMTB (8 cm/yr), a value
236 that is unrealistic in nature, a symmetric negative vertical flexural deflection has been
237 obtained on the other side of the obstacle (Clark and Royden, 2000). First, such negative
238 deflection in the footwall of the LMTB is not observed in the field, and such a lack of
239 foreland flexural basin has been pointed out as a peculiar characteristic for such high
240 mountain range (Burchfiel et al., 2008), so that regional topography deduced from flexure
241 equation (Clark and Royden, 2000) is not pertinent. Second, by testing such boundary
242 conditions in our model 2, we show that the rapid flow in the lower crust leads to the
243 thickening of the crust with unrealistic rapid growth of the topography (1 km in 110.000
244 years, Fig. 3D), but also with a very different deflection profile, smooth with a gradual
245 decrease of the velocity being almost null at the edge of the obstacle (Fig. 3C). Such smooth
246 deflection profile is not comparable with the sharp exhumation profile of the YTB (Fig. 1E)

247 at the edge of the South China craton (Fig. 4), similar to the one of the LMTB at the edge of
248 the Sichuan craton (e.g., Li et al., 2023). Indeed, by adding a thrust in the upper crust at the
249 edge of the obstacle (Figure S15), as the lower crust flow is not reaching the edge of the
250 obstacle, such thrust is not playing a significant role in limiting the deformation, unlike
251 what it is observed in SE Tibet (Figure 1).

252

253 **5.2. Low viscosity thrust embedded in the upper crust**

254 With a straight fault geometry, the deflection profile is symmetric, with a negative
255 deflection in the footwall of the thrust (Fig. 3A), which does not agree with the lack of
256 foreland flexural basin observed in the footwall of both the YTB and LMTB (Burchfiel et
257 al., 2008). On the contrary, the deflection profile obtained in our model 1 with a listric fault
258 (Fig. 3B), is comparable to the exhumation profile measured for the Yalong margin, with a
259 sharp positive deflection above the Muli thrust and no symmetric negative deflection above
260 the foreland (Fig. 1C). For the hanging wall, the thermo-kinematic inversion of the
261 thermochronological data obtains an exhumation rate between 0.26 and 0.52 km/Ma (Pitard
262 et al., 2021), of the same order of magnitude than the rate at 0.6 km/Ma obtained in our
263 model (Fig. 3A). The maximum total exhumation is of ~15 km close to the Muli fault trace
264 (Fig. 1E), corresponding in nature, where erosion at the edge of the plateau is active, to a
265 smooth topographic step of ~2 km (Fig. 1B). Such quantitative and qualitative difference
266 makes the comparison of the model deflection with only the topographic profile not
267 pertinent.

268 However, the velocity field near the Muli thrust is constrained only in the first 20 km
269 of the crust by thermochronological data (Pitard et al., 2021). Our 2-D numerical modelling
270 overcome such limitations raised by thermo-kinematic modelling, in particular for the
271 implied viscous flow in the lower crust. Indeed, our model 1 shows that the listric thrust in

272 the upper crust influences deformation throughout the entire crust, limiting the propagation
273 of the flow in the lower crust below the fault, albeit no discontinuity is imposed (Fig. 3B).
274 The regional tomographic images show that the crustal thickening stops at the Muli thrust
275 (Yao et al., 2008). It suggests that the limit of the more ductile deformation in the lower
276 crust is driven by the upper crustal Muli thrust, extending at depth by a ductile shear zone
277 that could reach the base of the crust (Robert et al., 2010), acting as a ramp to exhume
278 crustal material to the surface (Fig. 4). On the contrary, in North Tibet, the lower crust
279 appears rigid and subducts attached to the lithospheric mantle, while thrusts are branching at
280 depth on an intracrustal decollement, thickening only the upper crust (e.g., Meyer et al.,
281 1998; Lin et al., 2022).

282 Adding a rigid obstacle in the lower crust mimicking the South China craton is not
283 modifying the deflection profile (Fig. 3A), but in our setup we impose the position of the
284 upper crust thrust (Fig. 2), while in nature the thrust belts are located at the edge of the
285 cratons (Fig. 4). Indeed, the LMTB is located at the edge of the Sichuan craton, extending at
286 depth and marking a strong crustal contrast (Robert et al., 2010) (Fig. 1C). The YTB is also
287 at the edge of the South China craton, and we conclude to a structure at depth similar to the
288 one of the LMTB, with the Muli thrust being the dominant structure (Fig. 4). Nevertheless,
289 with respect to the LMTB, the YTB is spreading more over the craton, making a smoother
290 topographic transition between the high plateau and the foreland (Fig. 1B). Such spreading
291 is most probably due to the propagation of crustal thickening along the strike-slip fault of
292 XianShuiHe, as shown in the north of the plateau along the Altyn Tagh fault (Tapponnier et
293 al., 2001; Jiao et al., 2023).

294

295

296

297 **6. Conclusion: implication for crustal thickening processes of Eastern Tibet**

298 We conclude that structural, topographic, tomography, thermochronology, and
299 thermo-kinematic modelling data from the Yalong thrust belt region contradict the
300 hypothesis of a fast flow in a lower crust channel, but favor the model where shallow high
301 angle listric thrusts (70° dip) act as a ramp to exhume crustal material to the surface. Our
302 modelling yields the thickening of the entire crust, with a thick, warmer and therefore more
303 ductile lower crust (Fig. 4) as seen on tomographic images. In this region (LongmenShan
304 and Yalong margins), the former passive margin structure of the SongPan Ganze block with
305 a thick sedimentary wedge and a thinned lower crust (e.g., Roger et al., 2010; Airaghi et al.,
306 2018), bounding the rigid Sichuan and South China craton (Yao et al., 2008), probably leads
307 to this peculiar deformation regime.

308

309 **Acknowledgments**

310 This work was supported by the Agence Nationale de Recherche (ANR-20-CE49-0008
311 “Tibetan Orchestra”).

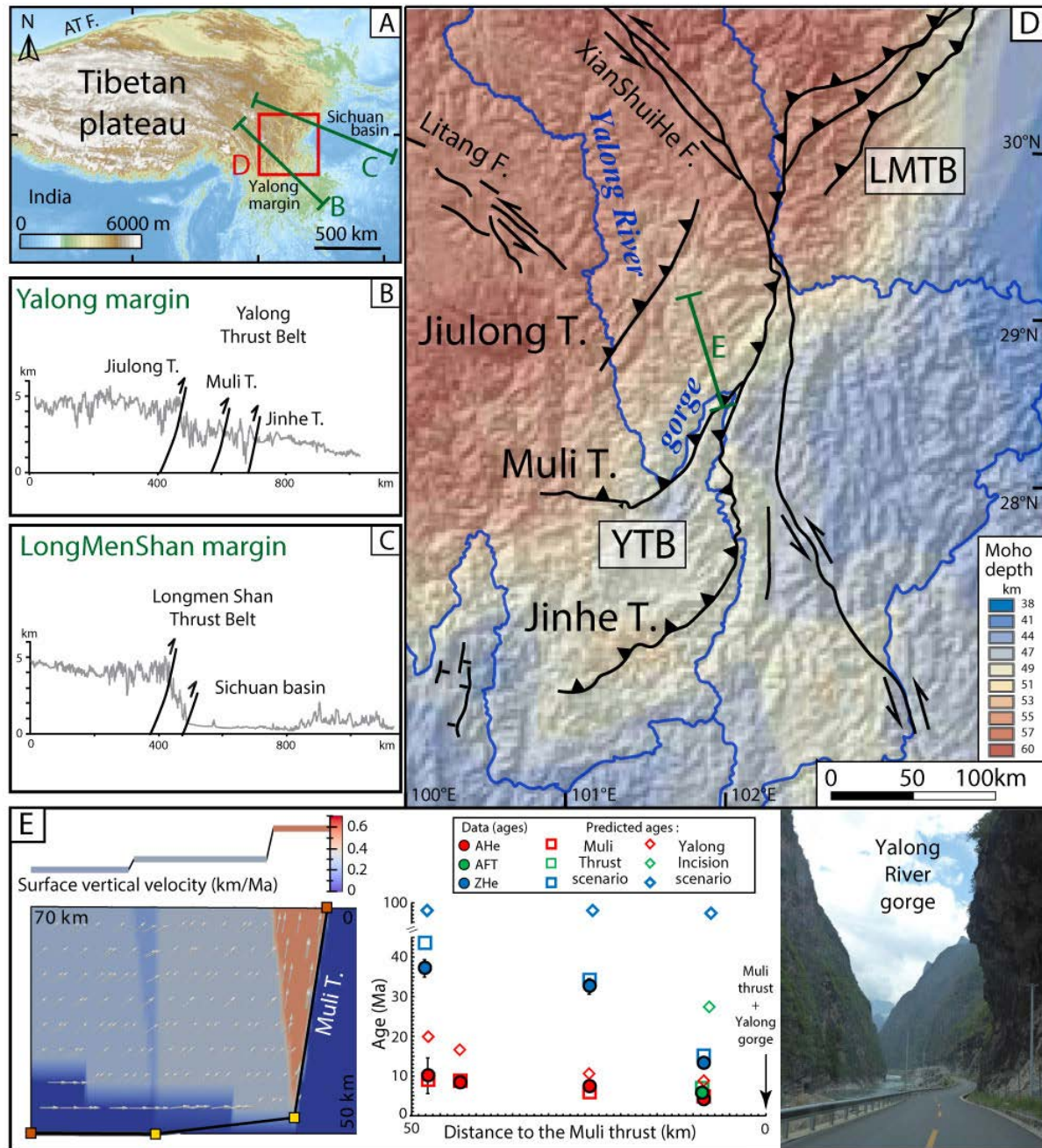
312 **Open Research**

313 The code used for this paper can be found at <https://doi.org/10.5281/zenodo.7558920> (Pitard
314 et al., 2023).

315 **Supplemental Material**

316 8 models with varying basic parameters from the two models presented in the paper
317 (viscosities within the fault and the lower crust, lower crustal thickness, fault width and
318 geometry, lateral velocity), and parametric study for optimization of physical parameters of
319 models 1&2 are joined to this paper as Supporting Information.

320



321

322 **Figure 1.** A/ Location of the Yalong margin, South-East Tibet. AT F.: Altyn Tagh fault. B/

323 & C/ Topographic profiles across the Yalong and the LongMenShan margins (see A/ for

324 location). D/ Moho depth map (Yao et al., 2008), with main faults, LMTB : Longmenshan

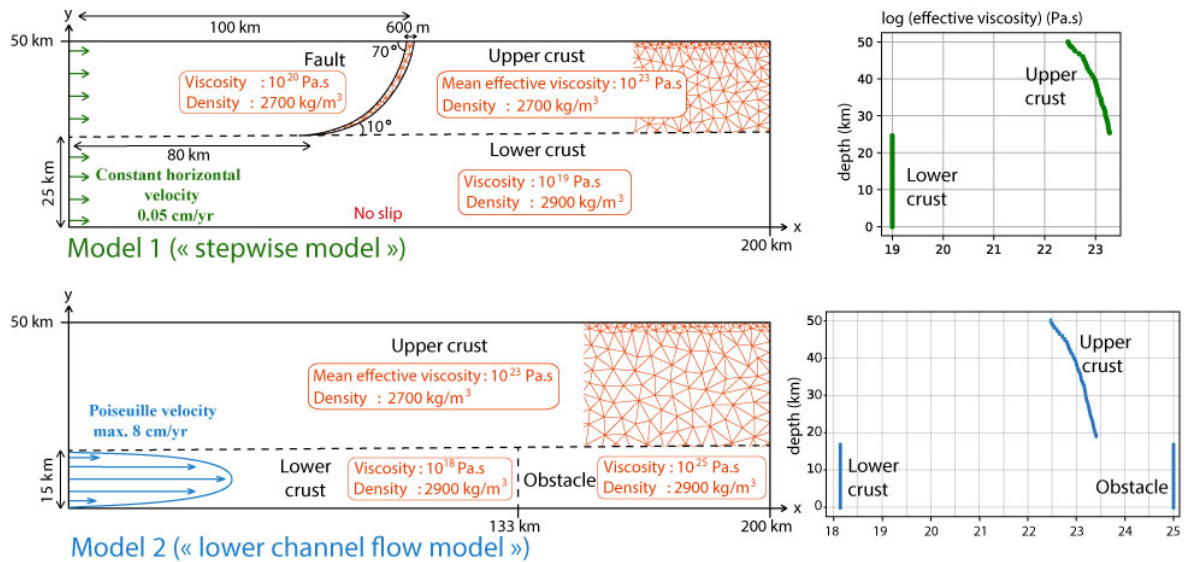
325 thrust belt, YTB : Yalong thrust belt including the Jiulong, Muli and Jinhe thrusts (T.),

326 XianShuiHe and Litang strike-slip faults (F.). E/ 2-D crustal velocity field obtained by

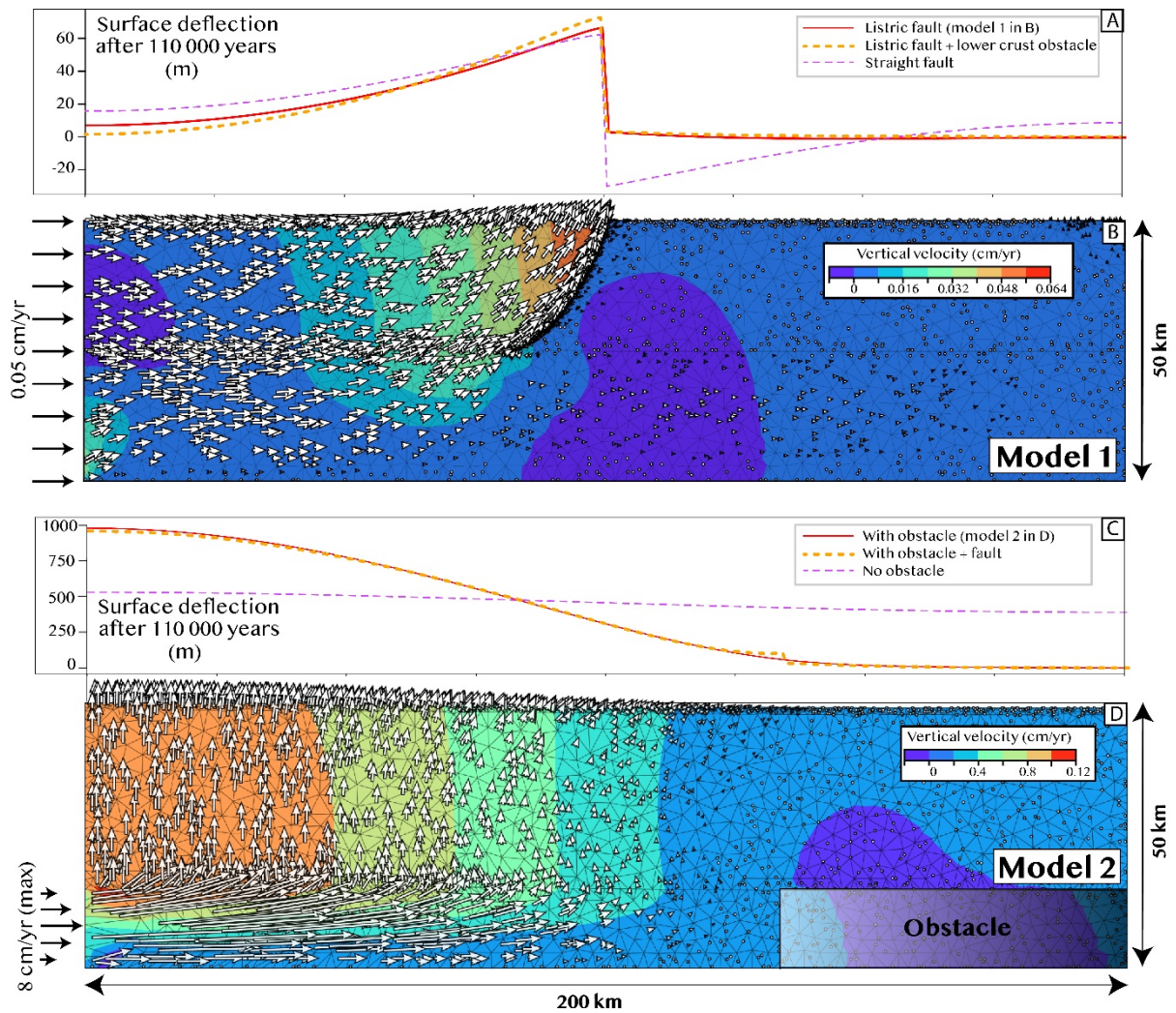
327 inversion of thermochronology ages with corresponding measured and predicted ages (AHe

328 and ZHe: Apatite or Zircon (U-Th)/He, AFT: Apatite fission track) for the tectonics

329 scenario of exhumation above the Muli thrust (square)(Pitard et al., 2021), to be compared
330 with an incision scenario (diamond) of exhumation due to the incision of the Yalong River,
331 illustrated by a picture of the deep gorge where the river crosses the thrust.
332



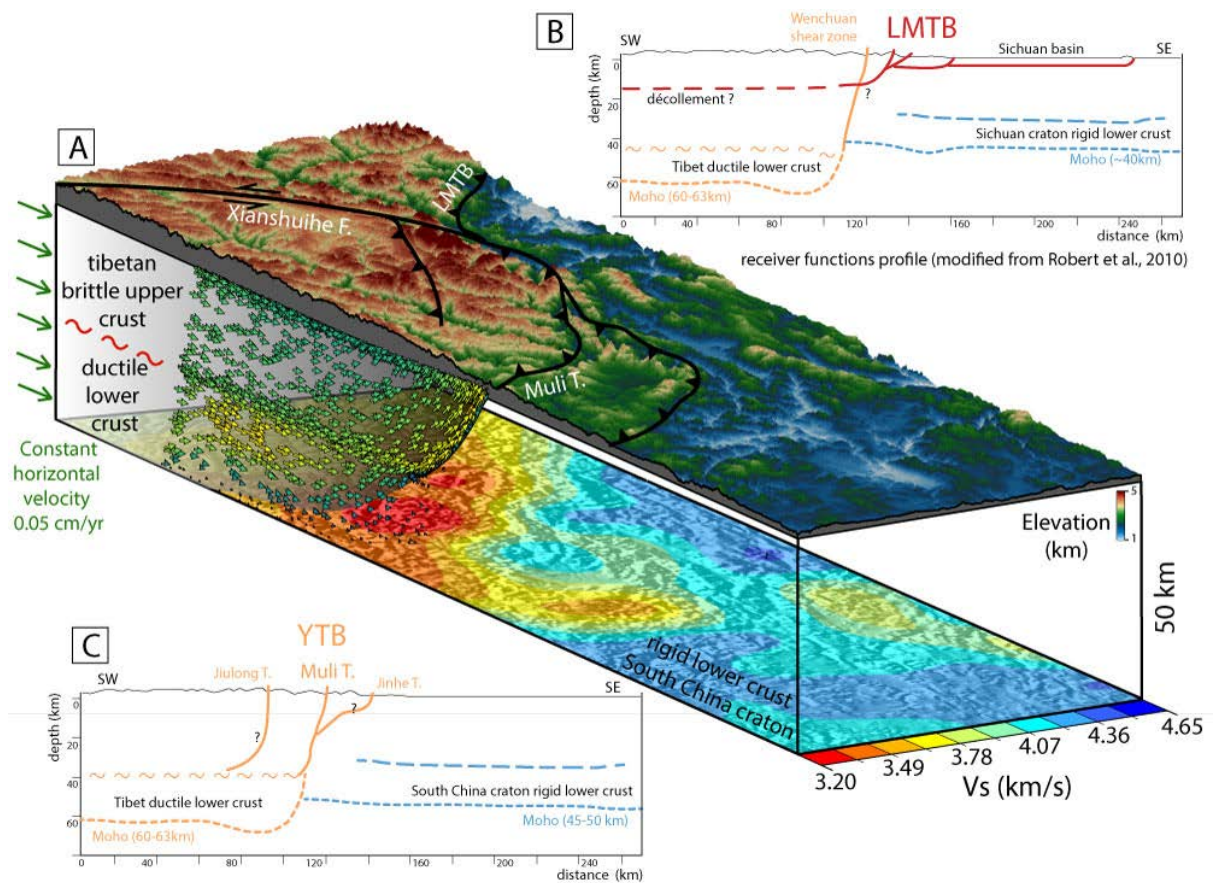
333
334
335 **Figure 2.** 2D crustal mechanical models setup with effective viscosity of a pseudo-plastic
336 behavior for the upper crust (mean value 10^{23} Pa.s), constant viscosity for the lower crust
337 (10^{19} Pa.s in model 1, 10^{18} Pa.s in model 2), the fault in model 1 (10^{20} Pa.s) and the obstacle
338 in model 2 (10^{25} Pa.s). Deformation under a constant horizontal velocity in model 1, or
339 under a Poiseuille flow in model 2, no-slip basal condition.



340

341 **Figure 3.** A/ Surface deflection and B/ velocity field of model 1, deforming under constant
342 horizontal velocity boundary condition, and a fault embedded in the upper crust. C/ Surface
343 deflection and D/ Velocity field of model 2, deforming under Poiseuille flow condition in
344 the lower crust.

345



346

347 **Figure 4.** A/ 3D block diagram of the Yalong margin with surface topography (2X
348 vertical exaggeration, Pitard et al., 2021), tomography section at 50 km-depth (Yao et al.,
349 2008), and the 2D crustal velocity field of our model 1, with constant lateral push, where
350 shallow the high angle listric Muli thrust acts as a ramp to exhume crustal material. B/ 2D
351 vertical section across the LongMenShan Thrust Belt (LMTB) drawn from the receiver
352 function profile of Robert et al. (2010). C/ 2D vertical section across the Yalong margin,
353 modified from B, showing a similar crustal geometry inferred for the Yalong Thrust Belt
354 (YTB).

355

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