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1 3D numerical modeling of mantle flow, crustal dynamics and magma genesis associated with slab
2 roll-back and tearing: the eastern Mediterranean case

3

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21 **Abstract**

22 Interactions between subduction dynamics and magma genesis have been intensely
23 investigated, resulting in several conceptual models derived from geological, geochemical and
24 geophysical data. To provide physico-chemical constraints on these conceptual models, self-
25 consistent numerical simulations containing testable thermo-mechanical parameters are
26 required, especially considering the three-dimensional (3D) natural complexity of subduction
27 systems. Here, we use a 3D high-resolution petrological and thermo-mechanical numerical
28 model to quantify the relative contribution of oceanic and continental subduction/collision,
29 slab roll-back and tearing to magma genesis and transport processes. Our modeling results
30 suggest that the space and time distribution and composition of magmas in the overriding
31 plate is controlled by the 3D slab dynamics and related asthenospheric flow. Moreover, the
32 decrease of the bulk lithospheric strength induced by mantle- and crust-derived magmas
33 promotes the propagation of strike-slip and extensional fault zones through the overriding
34 crust as response to slab roll-back and continental collision. Reduction of the
35 lithosphere/asthenosphere rheological contrast by lithospheric weakening also favors the
36 transmission of velocities from the flowing mantle to the crust. Similarities between our
37 modeling results and the late Cenozoic tectonic and magmatic evolution across the eastern
38 Mediterranean region suggest an efficient control of mantle flow on the magmatic activity in
39 this region, which in turn promotes lithospheric deformation by mantle drag via melt-induced
40 weakening effects.

41

42 **Keywords**

43 3D numerical modeling, subduction-related magmatism, slab roll-back, slab tearing,
44 asthenospheric flow, lithospheric deformation

45 1. Introduction

46 Arc magmatism is one of the most typical surface expressions of subduction dynamics
47 tracking physico-chemical processes from the upper mantle to the crust [Arculus, 1994]. It is
48 broadly accepted that the genesis of these magmas is dominated by partial melting of a
49 hydrated mantle wedge source, as the result of the H₂O release from the subducting oceanic
50 lithosphere [Arculus, 1994; Grove *et al.*, 2006]. These primary mafic magmas then ascend
51 through the mantle, reaching the base of the overriding crust where they can be retained in the
52 melting assimilation storage and homogenization (MASH) zone [Hildreth and Moorbath,
53 1988]. During their storage and their subsequent ascent up to the upper crust, these magmas
54 undergo some degrees of fractionation and crustal assimilation, which provide the typical
55 calc-alkaline magmatic trend. To reach a coherent understanding of the interactions between
56 subduction dynamics and magma genesis, numerical experiments have been performed
57 involving mechanical, thermal and petrological constraints [Huw Davies and von
58 Blanckenburg, 1995; Nikolaeva *et al.*, 2008; Zhu *et al.*, 2009, 2013; Gerya and Meilick, 2011;
59 Bouilhol *et al.*, 2015]. However, in subduction environment involving slab roll-back and
60 tearing, the 3D effects of slab dynamics and related asthenospheric flow on the three-
61 dimensional (3D) space and time distribution and composition of magmas have been most
62 often assessed by conceptual models [e.g. Pe-Piper and Piper, 2006; Dilek and Altunkaynak,
63 2009; Ersoy and Palmer, 2013; Jolivet *et al.*, 2015].

64 In 3D oceanic-continental subduction system involving slab roll-back and tearing,
65 particular attention has been given to the control of subduction dynamics and related
66 asthenospheric flow on lithospheric deformation without considering the role of magmatic
67 processes [Jadamec and Billen, 2010; Li *et al.*, 2013; Capitanio, 2014; Duretz *et al.*, 2014;
68 Magni *et al.*, 2014; Sternai *et al.*, 2014]. Rheological weakening effect of magmas resulting
69 from the partial melting of the upper mantle and the crust has, however, major consequences

70 on lithospheric-scale tectonics [*Gerya and Meilick, 2011*], which need to be tested across
71 complex 3D subduction system.

72 Here we use a 3D high-resolution petrological and thermo-mechanical numerical
73 geodynamic model to assess (1) the role of subduction dynamics involving slab roll-back and
74 tearing processes on the magmatic activity within the overriding lithosphere and (2) the
75 weakening effect of magmas on the lithospheric deformation. The numerical setup accounts
76 for experimentally calibrated non-Newtonian visco-plastic rheologies and physico-chemical
77 processes, such as slab dehydration, fluid transport, partial melting, melt extraction and
78 crustal growth [e.g. *Nikolaeva et al., 2008; Zhu et al., 2013*]. Modeling results suggest that
79 slab dynamics and related asthenospheric flow affect the genesis, transport and emplacement
80 of magmatic bodies both in the arc and back-arc domain and that such melts weaken the
81 lithosphere, promoting lithospheric deformation and its coupling with asthenospheric flow.
82 Our results are then compared to observations from the eastern Mediterranean region, where
83 the late Cenozoic evolution of the subduction zone including slab roll-back and tearing
84 processes and associated magmatism is relatively well constrained (Fig. 1) [e.g. *Jolivet and*
85 *Faccenna, 2000; Pe-Piper and Piper, 2006; Dilek and Altunkaynak, 2009; Ersoy and Palmer,*
86 *2013; Jolivet et al., 2013*].

87

88 **2. Late Cenozoic evolution of the eastern Mediterranean subduction zone**

89 Since the Mesozoic, the geodynamic evolution of the Mediterranean realm is characterized
90 by the subduction and/or collision of successive oceanic and continental lithospheres,
91 governed by the Africa-Eurasia convergence [*Stampfli and Borel, 2002; Menant et al., 2016*].
92 Since the latest Eocene-early Oligocene (i.e. 35-30 Ma), increasing rate of slab retreat has
93 resulted in the opening of back-arc basins, such as in the eastern Mediterranean region where
94 the Hellenic southward slab roll-back led to the opening of the Aegean domain [*Jolivet and*

95 *Faccenna, 2000*]. In this extensional context, medium- to high-temperature rocks were
96 exhumed as several metamorphic core complexes (MCCs), such as in the Rhodope massif, the
97 Cyclades and the Menderes massif (Fig. 1) [*Lister et al., 1984; Bozkurt and Oberhänsli, 2001;*
98 *Vanderhaeghe, 2004; Brun and Sokoutis, 2007; Jolivet et al., 2013*]. At the same time,
99 subduction and collision of the Arabian plate with the southern margin of the Tauride belt
100 took place further east (Fig. 1c) [*Jolivet and Faccenna, 2000*]. The present-day kinematics of
101 this region is characterized by an overall counterclockwise rotation of the surface velocity
102 vectors as recorded by GPS measurements (Fig. 2a) [*Reilinger et al., 2006; Le Pichon and*
103 *Kreemer, 2010*] controlled by the westward extrusion of Anatolia, which is mainly
104 accommodated by the North Anatolian Fault (NAF) [*Şengör et al., 2005*].

105 At ~150 km-depth, tomographic models for the eastern Mediterranean region show trench-
106 parallel variations in seismic velocity, with low-velocity anomalies located in eastern and
107 western Anatolia and interpreted as major break-off or tears in the slab that displays a high-
108 velocity anomaly (see star symbols in Fig. 2b) [*de Boorder et al., 1998; Wortel and Spakman,*
109 *2000; Li et al., 2008*]. The complex geometry of the subducting lithosphere thus resulted in a
110 complex mantle convection pattern including both poloidal and toroidal flows [*Faccenna and*
111 *Becker, 2010*], the latter having been partly imaged by *SKS* splitting anisotropy (Fig. 2b)
112 [*Paul et al., 2014*].

113 Integrated in this complex tectonic frame, magmatic occurrences show a specific spatial
114 and temporal distribution associated with variable petrological and geochemical composition
115 [*Menant et al., 2016*]. Since the Oligocene (i.e. 34-23 Ma), high-potassic calc-alkaline to
116 shoshonitic magmatism was progressively emplaced southward from the Rhodope massif to
117 the Cyclades and the Menderes massif, associated with the exhumation of MCCs (Figs. 1a, c)
118 [*Pe-Piper and Piper, 2006; Dilek and Altunkaynak, 2009; Ersoy and Palmer, 2013*]. In
119 addition to this first-order migration, a secondary southwestward migration of the magmatic

120 activity is recorded during the Miocene (i.e. 23-5 Ma) from the Menderes massif to the
121 Cyclades (Fig. 1a) [Jolivet *et al.*, 2015]. Evolution of geochemical and isotopic composition
122 of this magmatism suggests an increase of mantle source component relative to the crustal
123 source component [Keskin, 2003; Dilek and Altunkaynak, 2009; Ersoy and Palmer, 2013;
124 Jolivet *et al.*, 2015], finally producing alkaline magmas with a major depleted asthenospheric
125 source component [Ersoy and Palmer, 2013]. In eastern Anatolia, a wide late Miocene-
126 Quaternary volcanic province was also emplaced with composition evolving from high-
127 potassic calc-alkaline to alkaline (Fig. 1c) [Keskin, 2003].

128

129 **3. Numerical modeling**

130 In this section we describe the modeling approach and the initial reference model setup and
131 then outline the main objectives of the parametric study.

132

133 **3.1. Modeling approach and setup**

134 We use the wealth of observations from the eastern Mediterranean region to design and
135 calibrate numerical experiments that could be qualitatively and quantitatively compared to the
136 first order evolution of this region during the late Cenozoic. We stress, however, that the main
137 objective of our work is to investigate the 3D effects of slab retreat and tearing on magmatism
138 and we do not claim that our models provide an exhaustive description of geodynamics across
139 the area. As such, results may also be representative of other regions similarly characterized
140 by fast evolving narrow subduction zones involving lateral variations of the subducting
141 material (e.g. central and western Mediterranean region, southeastern Asia).

142 Numerical modeling was performed using the code *I3ELVIS*, like in other recent studies
143 [Zhu *et al.*, 2009, 2013; Li *et al.*, 2013; Duretz *et al.*, 2014; Sternai *et al.*, 2014]. The code
144 solves the mass, momentum and heat conservation equations based on a conservative finite-

145 difference scheme combined with a marker-in-cell technique and includes non-Newtonian
146 visco-plastic rheologies (Table S1) [*Gerya and Yuen, 2007*]. In addition, this code accounts
147 for physico-chemical processes, such as slab dehydration, fluid transport, partial melting, melt
148 extraction and crustal growth at continental margin [*Zhu et al., 2013*]. Details on the
149 numerical approach and material properties used in our numerical experiments are available
150 as Supplementary Material.

151 In all our experiments, the model domain measures $1500 \times 680 \times 656$ km in the x , y and z
152 directions, respectively, resolved by $437 \times 277 \times 149$ grid nodes giving a resolution of $3.4 \times$
153 2.5×4.4 km (Fig. 3a). Around 130 million additional randomly distributed markers are used
154 for advecting the material properties and computing water release, transport and consumption
155 as well as partial melting and melt extraction. The initial setup is designed with an overriding
156 continental lithosphere and a subducting oceanic lithosphere including a small continent (i.e.
157 500×328 km in x and z directions, respectively), which represent the Eurasian plate and the
158 converging Mesogean and Arabian lithospheres, respectively (Figs. 1c and 3a). A 10 km thick
159 trench-parallel weak zone is prescribed at the interface between the overriding and subducting
160 lithospheres in order to initiate subduction. In our reference model, the overriding continental
161 lithosphere is composed by 30 km and 20 km thick upper and lower crust, respectively, while
162 the subducting continental crust is composed of 20 km and 10 km thick upper and lower crust,
163 respectively (Figs. 3a and 3b). The initial thermal structure of the continents is laterally
164 uniform and varies for each experiment (see *Section 3.2*). In the reference model, this is a
165 linear vertical gradient with 0 °C at the surface and 1300 °C at 89 km and 98 km depth for the
166 overriding and subducting continental lithospheres, respectively. The oceanic lithosphere
167 incorporates a 3 km thick hydrated basaltic crust and a 5 km thick gabbroic crust and its
168 thermal structure, also different for each experiment, is defined using a half-space cooling age
169 (of 60 Ma for the reference model) [*Turcotte and Schubert, 2002*]. These lithospheres overlay

170 an asthenospheric mantle characterized by an initial adiabatic temperature gradient of ~ 0.5 °C
171 km^{-1} .

172 Velocity boundary conditions are free slip at both back and front boundaries ($z = 0$ and $z =$
173 656 km). Material influx is imposed through the right and left boundaries ($x = 0$ and $x = 1500$
174 km). Velocities across these boundaries are null except for the first 250 km depth of the right
175 boundary where a ~ 1.9 cm yr^{-1} influx is applied (Fig. 3a), scaled on the African plate absolute
176 motion estimated between 1.2 and 2.5 cm yr^{-1} since the Oligocene [*Jolivet and Faccenna,*
177 2000]. The mass conservation is then ensured by material outflux through the upper and lower
178 permeable boundaries ($y = 0$ and $y = 680$ km). The top surface of the lithosphere is calculated
179 dynamically as an internal free surface by using an initial 12-15 km thick low-viscosity layer
180 simulating either air (for $y < 12$ km) or sea water. The large viscosity contrast caused by this
181 low-viscosity layer minimizes shear stresses at the top of the lithospheres that effectively
182 approximates a free surface [*Schmeling et al., 2008*].

183

184 **3.2. Parametric study on lithospheric geothermal gradient and rheological layering**

185 To assess the role of subduction dynamics on magma genesis and transport, we explore a
186 wide range of rheological profiles of interacting continental and oceanic plates by changing
187 their thermal gradients (Table S2). In addition, since there is no consensus on the rheological
188 structure of the lithosphere, especially in the eastern Mediterranean region whose
189 geodynamical history is long and complex, different rheological layering have been tested for
190 the continental lithospheres by defining either one or two crustal layers characterized by
191 distinct non-Newtonian rheologies (Figs. 3a and b). We thereby control the overall
192 rheological contrast and therefore the degree of plate coupling between the interacting
193 lithospheres in our experiments [*Gerya and Meilick, 2011*] leading to reproduce different
194 regimes of subduction with either advancing or retreating trench. Changing the lithospheric

195 thermal gradients also modifies the degree of partial melting of the lithosphere, allowing us to
196 investigate the weakening effect of magmas on lithospheric deformation. In the following, we
197 use acronyms like *2c89-60* for model with two-layer continental crust, an overriding
198 lithosphere 1300 °C isotherm at 89 km depth and 60-Myr old oceanic lithosphere (Fig. 3a).

199

200 **4. Results**

201 In this section, we outline the main numerical results through a series of figures obtained
202 from selected experiments (Table S2) focusing on the overall evolution of these models, on
203 the resulting asthenospheric and lithospheric flow pattern and on the genesis and evolution of
204 partially molten rocks from the mantle to the crust. Figs. 4 and 5 illustrate the evolution of
205 models *2c89-60* and *1c113-70*, characterized by two- and one-layer continental crust,
206 respectively. Fig. 6 shows the temporal variations of the velocity of trench migration from
207 selected experiments and from kinematic reconstructions of the eastern Mediterranean region
208 [*Menant et al.*, 2016]. Fig. 7 highlights the mantle and crustal flow pattern and the strain rate
209 distribution in the overriding lithosphere from selected models. Finally, Figs. 8 and 9 illustrate
210 the temporal distribution of partially molten rocks from models *2c89-60* and *1c113-70*.
211 Additional results are also described in this section, although related figures are available as
212 Supplementary Material.

213

214 **4.1. Succession of events: oceanic and continental subduction followed by slab tearing**

215 In our reference model *2c89-60* (Fig. 4; Videos S1 and S2), subduction initiates through
216 progressive bending and roll-back of the oceanic slab, associated with a homogeneous
217 stretching of the overriding continental lithosphere (Fig. 4a). Subsequent continental
218 subduction then induces along-trench buoyancy variations resulting in progressive curvature
219 of the trench (Fig. 4b). When the continental slab reaches ~150 km depth, trench-parallel

220 tearing occurs at the transition between the more buoyant continental lithosphere and the
221 oceanic lithosphere (Fig. 4b). The slab tear then propagates along this compositional
222 boundary inducing trench-perpendicular tearing (Fig. 4c). Once decoupled from the oceanic
223 lithosphere, the torn continent bends up and collides into the overriding plate while the still
224 active oceanic subduction zone continues to retreat associated with slab flattening. This
225 succession of events is robust through all experiments with two-layer continental crust (model
226 *2c84-100* with higher lithospheric thermal gradient; Fig. S1), except when the lithospheric
227 thermal gradient is low, which leads to overall compression in the overriding lithosphere
228 (model *2c98-100*; Fig. S2a). This transient evolution is also consistent in models with one-
229 layer continental crust (model *1c113-70*; Fig. 5; Videos S3 and S4) with the exception of a
230 higher stretching of the subducting continental lithosphere also associated with crustal
231 decoupling, which thus delays slab tearing (Figs. 5b and c). It is noteworthy that the apparent
232 lack of lithospheric mantle below the overriding stretched continental crust (model *1c113-70*;
233 Fig. 5) does not imply that the crust lies directly on the asthenosphere. Depending on the
234 evolution of the thermo-mechanical properties of the peridotite, the lithosphere-asthenosphere
235 boundary is therefore more accurately defined in our experiments by considering the 1200 °C
236 isotherm (Fig. 3b).

237 In all numerical experiments, laterally varying subduction dynamics exerts a major control
238 on the surface kinematics. Coeval subduction of oceanic and continental lithospheres results
239 in an along-trench variation of the rate of trench retreat evolving in the reference model *2c89-*
240 *60* from $\sim 2 \text{ cm yr}^{-1}$ to $3\text{-}6 \text{ cm yr}^{-1}$ above the continental and oceanic slab, respectively (Fig.
241 6a). Once slab tearing has occurred, trench retreat accelerates above the continuing oceanic
242 subduction zone reaching up to $\sim 6 \text{ cm yr}^{-1}$ in this model. Models with higher lithospheric
243 thermal gradient show a sensibly faster trench retreat (model *2c84-100*; $\sim 9 \text{ cm yr}^{-1}$), while
244 those with a lower lithospheric thermal gradient are characterized by a slowly advancing

245 oceanic trench (model *2c98-100*; 0.5-2 cm yr⁻¹; Fig. 6b). Finally, models with one-layer
246 continental crust (model *1c113-70*) also show faster trench retreat than the reference model
247 despite a lower lithospheric thermal gradient (reaching up to ~7 cm yr⁻¹ above the fast
248 retreating oceanic slab; Fig. 6c). The magnitude of these trench velocities is broadly similar to
249 the rate of trench retreat for the Hellenic subduction system, estimated between 0.5 and 3
250 cm yr⁻¹ during the late Cenozoic (see solid line in Fig. 6d) [*Menant et al.*, 2016].

251

252 **4.2. Asthenospheric vs. lithospheric flow pattern**

253 Except for model *2c98-100* characterized by major compressional tectonics (Fig. S2), all
254 performed numerical experiments show an initial dominant poloidal asthenospheric flow,
255 initiated at the slab tip and upwelling into the mantle wedge (Fig. 7a). When slab tearing
256 occurs, the asthenospheric flow changes to a poloidal flow coming through the newly formed
257 slab windows then evolving in a toroidal flow, which finally reaches the mantle wedge (Fig.
258 7b). Meanwhile, extrusion of the overriding lithosphere towards the oceanic subduction
259 occurs (Figs. 4c and 5c). Extrusion tectonics in models with two-layer continental crust takes
260 place in a restricted zone above the trench-perpendicular slab tear, where the deformation is
261 localized along a strike-slip fault zone (Fig. 7c). Lithospheric and asthenospheric flows in the
262 extensional back-arc domain are then sub-parallel and controlled by the slab roll-back (Fig.
263 7d). However, the horizontal component of flow velocity at the lithosphere-asthenosphere
264 boundary is subject to an abrupt decrease (see velocity profile in inset of Fig. 7c). Conversely,
265 models with one-layer continental crust are characterized by more pronounced strike-slip
266 faulting during upper plate extrusion reaching a wide extensional region above the fast
267 retreating oceanic slab (Fig. 7e). In addition, the comparatively weaker continental lithosphere
268 in this extensional domain allows for more efficient transmission of the asthenospheric flow
269 to the overriding lithosphere, as shown by a progressive decrease of the horizontal component

270 of flow velocity from the asthenosphere to the upper part of the continental crust (see velocity
271 profile in inset of Fig. 7e).

272

273 **4.3. *Genesis and transport of partially molten rocks***

274 In our reference model *2c89-60* (Fig. 8), the first stage of oceanic subduction is associated
275 with (1) partial melting of the lower continental crust of the overriding lithosphere (Figs. 8a
276 and c) and (2) water release from the hydrated subducting oceanic crust at ~70 km depth
277 (Fig. 8d). Subsequent hydration of the mantle wedge contributes to its partial melting
278 resulting in a wave-like plume geometry of partially molten mantle (according to the
279 magmatic plume description of *Zhu et al.* [2009]) that starts to ascend at ~150 km depth (Figs.
280 4a and 8d). While the oceanic slab retreats, partially molten crust and mantle are stored at the
281 base of the overriding crust, being passively transferred from the arc to the back-arc domain
282 through time (see vertical cross-sections in Fig. 8). The up-bending of the subducting
283 continental lithosphere resulting from slab tearing is accompanied by fast rising and storage of
284 partially molten mantle and continental crust as well as of dry asthenosphere below the active
285 collision zone (Figs. 8e and h). A significant amount of partially molten sediments is then
286 added to the stacked material, which is laterally transferred through the collision zone into the
287 back-arc domain and exhumed through the stretching crust up to the surface (Figs. 8i and k).
288 Lateral migration of partially molten mantle at the base of the stretched crust above the
289 ongoing subduction zone also takes place (Fig. 8e) and is progressively replaced by dry
290 asthenosphere (Fig. 8i).

291 In model *2c98-100* displaying slow subduction associated with a compressional regime
292 (Figs. S2b and c), the magmatic activity is restricted to small amounts of partially molten
293 continental crust and sediments slowly ascending within the thickening overriding
294 lithosphere. In model *2c84-100* characterized by two-layer continental crust and fast slab roll-

295 back (Fig. S1), the hydrated and partially molten rocks evolution is faster, similar to the
296 reference model *2c89-60*, except that the wave-like plume of partially molten mantle in the
297 mantle wedge develops later (Fig. S3). Finally, the magmatic evolution of model *1c113-70*
298 with one-layer continental crust shows a number of differences. (1) A major partial melting
299 event occurs at the base of the overriding crust (Fig. 5). The partially molten material is then
300 progressively exhumed through the stretched crust forming a high temperature dome-like
301 structure (Figs. 9a, c and j). (2) The ascent of partially molten mantle in the mantle wedge
302 starts deeper, from ~300 km depth (Fig. 5a) and occurs as a wave-like plume (Fig. 5b), except
303 around the torn continental slab where it ascends as regularly-spaced finger-like plumes (Fig.
304 5c). (3) The long residence time of continental slab at depth before tearing results in
305 increasing partial melting of the crust and subsequently finger-like upwelling of melts
306 (Fig. 9g).

307

308 **5. Discussion**

309 We discuss below the relationships between magmatism, mantle flow and lithospheric
310 deformation assuming that partially molten rocks mimic magmatic sources from which
311 magmas with specific composition are potentially extracted. Finally, we compare our results
312 to observations from the eastern Mediterranean region.

313

314 **5.1. Ascent of partially molten mantle**

315 During the initial subduction of the oceanic lithosphere, slab dehydration and subsequent
316 hydrated partial melting result in the accumulation of mantle-derived melts within the mantle
317 wedge [Grove *et al.*, 2006]. According to our models, partially molten mantle with lower
318 viscosity and density rises up as plumes (Figs. 8b and 9b). This ascent is controlled by the
319 competition between the buoyancy of partially molten rocks and the subduction drag

320 (depending on viscosity and slab dip). The predominance of the buoyancy force results in the
321 development of Rayleigh-Taylor instabilities at the interface between hydrated/partially
322 molten mantle and dry mantle, initiating plume rising. *Zhu et al.* [2013] suggest that plate
323 convergence rate affects the depth of the plume growth with faster plate convergence (i.e.
324 $\sim 6.5 \text{ cm yr}^{-1}$) carrying more efficiently the hydrated oceanic crust down the subduction zone
325 than slower plate convergence (i.e. $\sim 2 \text{ cm yr}^{-1}$). We can add upon this evidence that the
326 asthenospheric flow related to slab roll-back and tearing is another crucial process affecting
327 the initiation and trajectories of partially molten rock upwelling through the mantle wedge.
328 Indeed, as soon as subduction and subsequent partial melting of the hydrated mantle start, a
329 sub-horizontal asthenospheric flow develops in the mantle wedge as a result of the dominant
330 poloidal flow component initiated at the slab tip ($\sim 5 \text{ cm yr}^{-1}$; model *2c89-60*; Figs. 8d and
331 10a). This sub-horizontal asthenospheric flow, particularly fast for high rate of slab roll-back,
332 is then deflected downward close to the slab-mantle interface ($\sim 10 \text{ cm yr}^{-1}$; model *1c113-70*,
333 Figs. 9d and 10b) resulting in an asthenospheric return flow that tends to delay the growth of
334 the plumes (initiated $\sim 300 \text{ km}$ depth in model *1c113-70*, Fig. 5a). Conversely, slab roll-back
335 and tearing promotes an asthenospheric upwelling which drags upward partially molten rocks
336 in the mantle wedge (Figs. 8k and 9k). In our experiments, this magmatic source mainly
337 ascends as wave-like plumes (Figs. 4b and 5b) but also as finger-like plumes, especially close
338 to the torn continental slab (model *1c113-70*, Figs. 5b and c). *Zhu et al.* [2009] argued that the
339 main controlling factor on the geometry of these plumes is the viscosity variation of the
340 partially molten material. In this study, the constant viscosity of 10^{19} Pa s for these partially
341 molten phases (constrained by imposing a lower cut-off value for viscosities in our
342 experiments, see Supplementary Material) suggests that other factors also control the plume
343 structure in the mantle wedge. We find here that finger-like plumes initiate above a steep and
344 slow sinking slab associated with slow asthenospheric return flow (Fig. 9h; $\sim 1.6 \text{ cm yr}^{-1}$),

345 when the continental slab buoyancy force counterbalances the slab pull force. The reduced
346 stress imposed by this slow asthenospheric flow then promotes the development of along-
347 trench Rayleigh-Taylor instabilities at the slab-mantle interface enabling the growth of finger-
348 like plumes. Conversely, in the case of fast retreating slab (Figs. 8j and 9j), related faster
349 asthenospheric flow (i.e. $\sim 6 \text{ cm yr}^{-1}$) avoids the initiation of these along-trench Rayleigh-
350 Taylor instabilities resulting in the ascent of partially molten rocks as a wave-like plume.

351

352 **5.2. *Magmatic evolution in the lower crust and mantle flow***

353 Besides controlling the ascending trajectories of magmatic sources through the mantle
354 wedge, the dominant toroidal asthenospheric flow related to slab tearing (Figs. 10c and d)
355 drives the trench-parallel migration of magmas and their emplacement at the base of the
356 overriding continental crust. Here, both partially molten mantle and crust are stored during the
357 first stage of subduction (Figs. 8 and 9) with a predominance of crustal-derived magmas,
358 especially in model *1c113-70* where the prescribed one-layer continental crust is less
359 refractory than the two-layer continental crust defined in model *2c89-60*. When slab tearing
360 occurs, large amount of both partially molten crust and mantle are stored at the base of the
361 overriding crust (Figs. 8e and 9e). Subsequent toroidal asthenospheric flow due to protracted
362 roll-back of the oceanic slab then drags these magmatic sources from the collision zone to the
363 back-arc domain, above the retreating subduction (Figs. 10c and d). In addition, we found a
364 lateral increase of mantle-derived relative to crust-derived melts in the back-arc domain (Fig.
365 5e) and a final upwelling of hot and dry asthenosphere potentially resulting in the adiabatic
366 melting of depleted mantle, which is not considered in the model (Figs. 8i and 9i).
367 Implementation of this magmatic process should be therefore a major improvement for future
368 works in order to discuss the control of asthenospheric upwelling on the genesis of magmas
369 with alkaline composition in the back-arc domain.

370 No significant contribution from partial melting of the oceanic crust was recognized in our
371 experiments, in turn suggesting that most magmas within the mantle wedge are generated
372 from hydrated mantle, even in warm subduction zones [Bouilhol *et al.*, 2015]. In the MASH
373 zone, mantle- and crustal-derived magmas may mix and therefore change composition, as if
374 they undergo fractionation and crustal assimilation processes [Hildreth and Moorbath, 1988].
375 Sedimentary material likely contributes to the maturation of these magmas in the MASH
376 zone, especially across continental subduction zone where the strong coupling of the
377 continental lithospheres enhances the basal erosion of the upper plate along the subduction
378 interface and therefore the burial of detrital continental crust (Figs. 8i and k).

379

380 **5.3. Magmatism and lithospheric deformation**

381 Low-viscosity partially molten rocks stored in the overriding lithosphere allow for
382 decreasing the bulk strength of the upper plate [Watts and Burov, 2003] thus favoring partial
383 decoupling between the continental crust and the underlying lithospheric mantle (Figs. 9b and
384 f). This melt-induced lithospheric weakening also enhances the decoupling at the subduction
385 interface, which in turn increases the rate of slab roll-back, thereby affecting lateral extrusion
386 and stretching of the upper plate (Fig. 7; comparison between slow evolving models *2c89-60*
387 and *2c98-100* and fast evolving models *1c113-70* and *2c84-100*).

388 In our experiments, low-viscosity partially molten rocks exhumed through the overriding
389 crust also provide a strong rheological contrast with the surrounding high-viscosity rocks
390 (Figs. 8b and 9b) affecting the distribution of crustal deformation. Strike-slip faulting and
391 back-arc-related intense lithospheric stretching thus localize in the weaker parts of the crust
392 accommodating the lateral extrusion of continental block and fast slab roll-back, respectively
393 (model *2c89-60*; Fig. 7c). Moreover, in accordance with previous studies [Gerya and Meilick,
394 2011; Capitano, 2014], our experiments show that a large amount of partial melting of the

395 overriding crust decreases the bulk lithospheric strength and favors the propagation of the
396 strike-slip fault zone and the opening of a wider back-arc basin (model *Ic113-70*; Fig. 7e).

397 The driving forces of plate tectonics across subduction-collision environment have been
398 largely discussed in term of crustal and mantle contributions [e.g. *Royden, 1996; Jolivet et al.,*
399 *2009; Faccenna and Becker, 2010*]. Recently, *Sternai et al.* [2014] proposed that extrusion
400 tectonics may be jointly driven by continental collision, slab roll-back and associated mantle
401 flow, the latter being particularly effective across hot and thin lithospheres. Indeed, the
402 viscosity contrast between such lithosphere and the underlying asthenosphere is low (i.e. ~ 10
403 Pa s; see Fig. 4 of *Sternai et al.* [2014]) promoting lithosphere/asthenosphere coupling. Our
404 results further suggest that the decrease of the bulk lithospheric strength by the magmatic
405 input results in a lower viscosity contrast along the lithosphere-asthenosphere boundary (i.e.
406 less than 10^2 Pa s; inset of Fig. 7e) therefore favoring the lithosphere/asthenosphere coupling.
407 This is a crucial mechanism for transmission of the velocities from the flowing mantle to the
408 crust (inset of Fig. 7e). In addition, sub-parallel crustal and asthenospheric velocity fields in
409 the back-arc domain where the upper lithosphere is thin, hot and therefore weak (Figs. 7d and
410 f), suggest that the strain is, at least to some extent, controlled by the underlying
411 asthenospheric flow in agreement with field observations [*Jolivet et al., 2009*].

412 These results emphasize the role of mantle and crustal magmas on upper plate deformation
413 and lithosphere/asthenosphere coupling in subduction zone by decreasing the bulk
414 lithospheric strength. Applied to the strong downgoing lithosphere, such a melt-induced
415 lithospheric weakening could be considered as an efficient and required mechanism for slab
416 bending. Absence of slab partial melting in our experiments does not confirm this hypothesis
417 but fluid-induced weakening [*Gerya and Meilick, 2011*] related to slab hydration could be an
418 alternative explanation to reduce the slab strength. Normal faulting occurring in the upper,

419 brittle part of the slab would then have promoted this weakening effect by enhancing deep
420 slab hydration [Faccenda et al., 2009].

421

422 **5.4. Implications for the eastern Mediterranean region**

423 The late Cenozoic geodynamics of the eastern Mediterranean region, in particular coeval
424 retreat of the Hellenic subduction zone and collision across the Arabia-Eurasia margin
425 (Fig. 1c) [Jolivet and Faccenna, 2000], are consistently reproduced by our numerical
426 experiments (Figs. 4, 5 and S1). At depth, slab tearing occurs in these models along the
427 ocean-continent transition, which is consistent with seismic tomographic models proposing
428 the existence of two slab tears beneath western and eastern Anatolia (Fig. 2b) [de Boorder et
429 al., 1998; Wortel and Spakman, 2000; Li et al., 2008]. In addition, the toroidal asthenospheric
430 flow associated with slab roll-back and tearing in our experiments (Fig. 7) is consistent with
431 instantaneous 3D velocity field calculations for the flowing mantle [Faccenna and Becker,
432 2010] and other self-consistent 3D numerical studies applied to the eastern Mediterranean
433 region [Li et al., 2013; Capitanio, 2014; Duretz et al., 2014; Sternai et al., 2014]. No clear
434 correlation, however, can be done between this toroidal flow and the SKS fast splitting
435 directions (Fig. 2b) [Paul et al., 2014] suggesting a more complex relationship between the
436 SKS orientations and the asthenospheric flow pattern in this region (see discussion of Le
437 Pichon and Kreemer [2010]).

438 According to our results, slab tearing is correlated with the extrusion of crustal block from
439 the collisional domain toward the retreating subduction zone resulting in an overall
440 counterclockwise crustal velocity pattern (Figs. 7c and e), similar to the present-day crustal
441 kinematics across the study region deduced from GPS measurements (Fig. 2a) [Reilinger et
442 al., 2006; Le Pichon and Kreemer, 2010]. The magnitude of these crustal velocities is also
443 consistent, with trench motion evolving from -2 cm yr^{-1} (advancing trench) to $3\text{-}7 \text{ cm yr}^{-1}$

444 (retreating trench) in models *2c89-60* and *1c113-70* and from 0.5 to 3 cm yr⁻¹ for the Hellenic
445 trench (Fig. 6) [Menant *et al.*, 2016]. Acceleration of trench retreat, combined with an
446 advancing collisional front in our experiments as well as in the natural case can be related to
447 slab tears, which would confirm the middle-late Miocene age for slab tearing in the eastern
448 Mediterranean region [Faccenna *et al.*, 2006; Dilek and Altunkaynak, 2009; Jolivet *et al.*,
449 2015]. Small differences in these velocities may be, to some extent, due to the imposed free
450 slip conditions at the front and back boundaries ($z = 0$ and $z = 656$ km) laterally constraining
451 the crustal and mantle flow. Additional numerical experiments accounting for open
452 boundaries as well as the neighboring plates would then allow for more realistic boundary
453 conditions and therefore more accurate crustal velocities [Yamato *et al.*, 2009].

454 Mimicking the NAF accommodating the lateral extrusion of Anatolia since ~12 Ma,
455 localized strike-slip fault zone is fairly well reproduced in models characterized by a weak
456 overriding lithosphere (model *1c113-70*; Fig. 7e). Conversely, the propagation of this
457 localized strike-slip deformation is reduced within a strong lithosphere (model *2c89-60*; Fig.
458 7c). A weak overriding lithosphere also allows for opening of a wide back-arc basin (model
459 *1c113-70*; Fig. 7e) characterized by a ~50 km thick lithosphere (inset of Fig. 7e), which is
460 consistent with the present-day 40-50 km thick Aegean lithosphere [Endrun *et al.*, 2008]. In
461 addition, the exhumation in the back-arc domain of partially molten rocks as dome-like
462 structures (model *1c113-70*, Fig. 8c) mimic high-temperature MCCs with migmatitic core,
463 recognized in the Rhodope and Menderes massifs as well as in the Cyclades archipelago
464 [Lister *et al.*, 1984; Bozkurt and Oberhänsli, 2001; Vanderhaeghe, 2004; Brun and Sokoutis,
465 2007; Jolivet *et al.*, 2013]. These similarities suggest that the late Cenozoic deformation
466 pattern in the eastern Mediterranean region is controlled by the melt-induced weakening of
467 the lithosphere, especially in the Aegean back-arc domain where the emplacement of

468 magmatic bodies with a major crustal source component confirms an intense partial melting
469 of the lower crust [*Pe-Piper and Piper, 2006; Ersoy and Palmer, 2013*].

470 Space and time distribution of these magmatic products is also characterized by a first-
471 order southward migration driven by slab roll-back (see orange arrow, Fig. 1a) [*Pe-Piper and*
472 *Piper, 2006; Dilek and Altunkaynak, 2009; Ersoy and Palmer, 2013*], which is fairly well
473 reproduced in our models (Figs. 8 and 9). During the Miocene, this magmatic trend is
474 modulated by a second-order southwestward migration from the Menderes massif to the
475 Cyclades (see yellow arrow, Fig. 1a), which is also faithfully reproduced in our model with an
476 along-trench migration of partially molten rocks at the base of the crust controlled by the
477 toroidal asthenospheric flow related to slab tearing (Figs. 10c and d). This along-trench
478 migration of magmatic activity in the Aegean domain is accompanied by an increase of the
479 mantle-sourced magmas [*Jolivet et al., 2015*], a feature that some experiments fail to
480 reproduce (model *1c113-70*; Fig. 9i), probably due to the insufficient numerical resolution
481 that buffered the mantle signature in the intensively partially molten continental crust.
482 Conversely, this trend is recognized in model *2c89-60* considering two-layer continental crust
483 (Fig. 8e), where increasing mantle-sourced magmatic component at the base of the overriding
484 crust mimics observations from the Miocene intrusions across the Cyclades [*Jolivet et al.,*
485 *2015*].

486 Eastern Anatolia is further characterized by a wide calc-alkaline to alkaline volcanic
487 province of late Miocene to Quaternary age (Fig. 1c) [*Keskin, 2003*]. In our experiments,
488 partially molten mantle and crust accumulate at the base of the collision zone due to the sub-
489 collisional slab detachment potentially resulting in such a wide magmatic activity (Figs. 8e
490 and 9e). Later alkaline volcanism emplaced in the eastern Mediterranean region resulting
491 from the adiabatic partial melting of depleted asthenosphere [*Keskin, 2003; Ersoy and*

492 *Palmer, 2013*] which can be related to the asthenospheric upwelling associated with slab
493 tearing and intense lithospheric stretching reproduced in our experiments (Figs. 8i and 9i).

494 Our results clearly emphasize the role of slab roll-back and tearing and related
495 asthenospheric flow on the late Cenozoic magmatic and tectonic evolution of the eastern
496 Mediterranean region. Of course, these 3D models cannot fully assess the natural complexity
497 of the eastern Mediterranean subduction zone, characterized by successive oceanic and
498 continental subduction events and, possibly, several slab tearing processes [*Stampfli and*
499 *Borel, 2002; Menant et al., 2016*]. However, consistency between our experiments and the
500 natural case in terms of crustal kinematics, deformation and magmatic evolution suggests that
501 we reproduce the first-order crustal and mantle dynamics of the eastern Mediterranean
502 subduction zone during the late Cenozoic.

503

504 **6. Conclusions**

505 Using a 3D high-resolution coupled petrological and thermo-mechanical model integrating
506 experimentally calibrated non-Newtonian crustal and mantle rheologies, we are able to
507 reproduce the first-order geodynamic evolution of a coeval oceanic and continental
508 subduction zone characterized by slab roll-back and tearing. Integrating slab dehydration,
509 fluid transport and partial melting processes, our experiments demonstrated (1) the control of
510 3D subduction dynamics and related asthenospheric flow on space and time evolution of
511 magmas from the upper mantle to the overriding crust and (2) the significant role of melt-
512 induced rheological weakening on lithospheric deformation. In detail, we reach the following
513 conclusions.

514 - Slab dynamics and associated mantle flow affect magma genesis and transport within the
515 hydrated mantle wedge. A typical asthenospheric return flow in the mantle wedge tends to
516 delay the ascent of partially molten rocks. Conversely, asthenospheric upwelling associated

517 with slab roll-back and tearing allows for a fast ascent of the magmatic sources as plumes
518 with variable morphologies. A slow sinking slab thus promotes the development of along-
519 trench Rayleigh-Taylor instabilities at the slab-mantle interface favoring finger-like plumes,
520 while fast evolving subduction zone and associated asthenospheric flow facilitate the ascent
521 of large amount of partially molten material as wave-like plumes.

522 - During their emplacement, both mantle- and crust-derived magmas contribute to the
523 reduction of the bulk strength of the overriding lithosphere thereby affecting the deformation
524 pattern. The ascent of large amounts of partially molten material within the stretched crust
525 localizes deformation along dome-like structures. The crustal deformation is then
526 characterized by strike-slip faulting and stretching that accommodate the lateral extrusion of
527 crustal block and slab roll-back, respectively. This melt-induced weakening of the crust also
528 promotes the lithosphere/asthenosphere coupling and the transmission of the velocities from
529 the flowing asthenosphere to the crust, supporting the hypothesis that the mantle flow may
530 drag the crust thereby affecting surface deformation.

531 - The kinematic, strain and magmatic evolution of the eastern Mediterranean region
532 deduced from geological and geophysical constraints are reasonably well reproduced by our
533 experiments suggesting a major control of subduction dynamics and related asthenospheric
534 flow on the magmatic evolution. As such, we propose that roll-back and tearing of the eastern
535 Mediterranean slab may have controlled, respectively, the southward and southwestward
536 migration of magmatic centers since the late Cenozoic.

537

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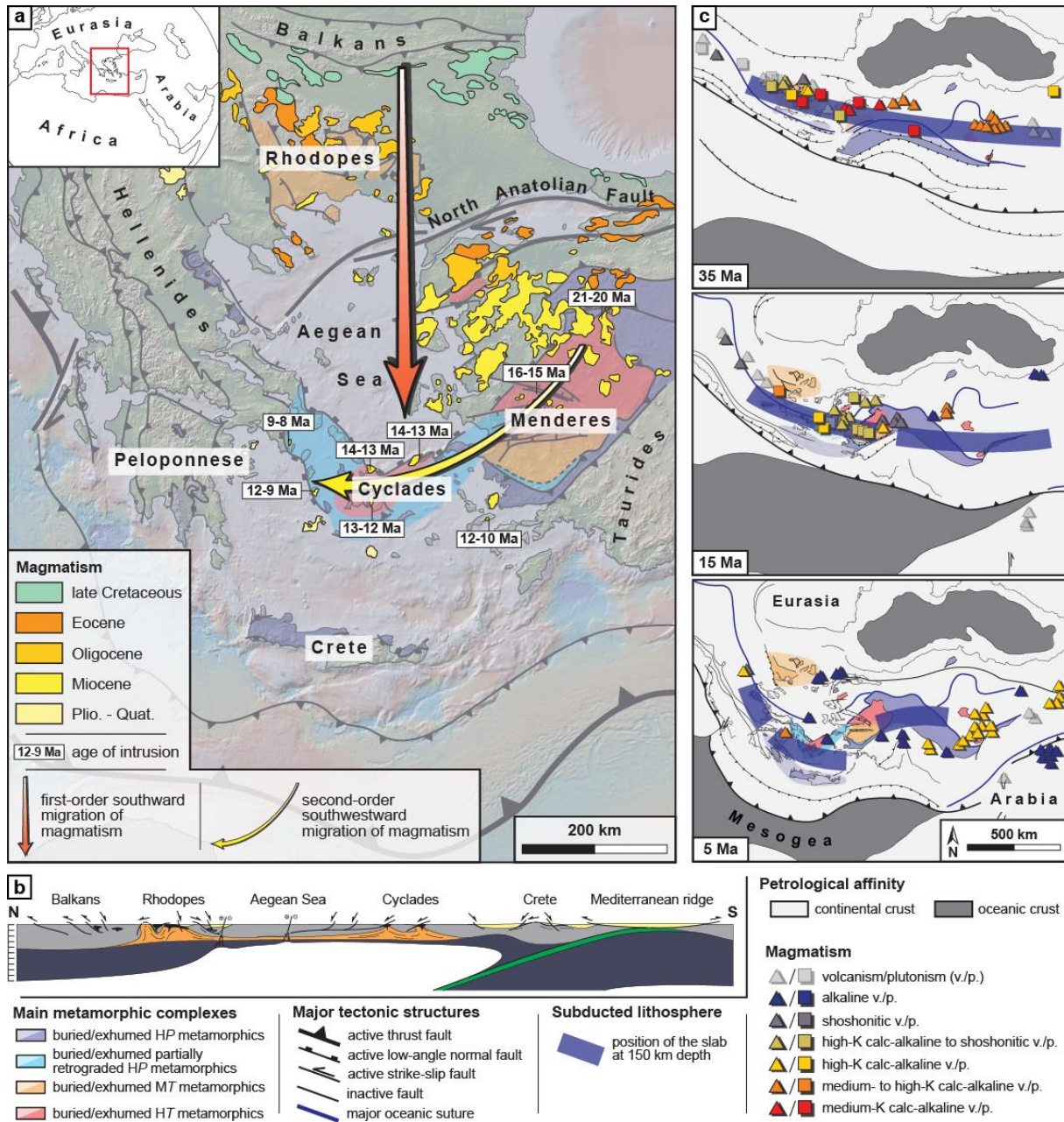
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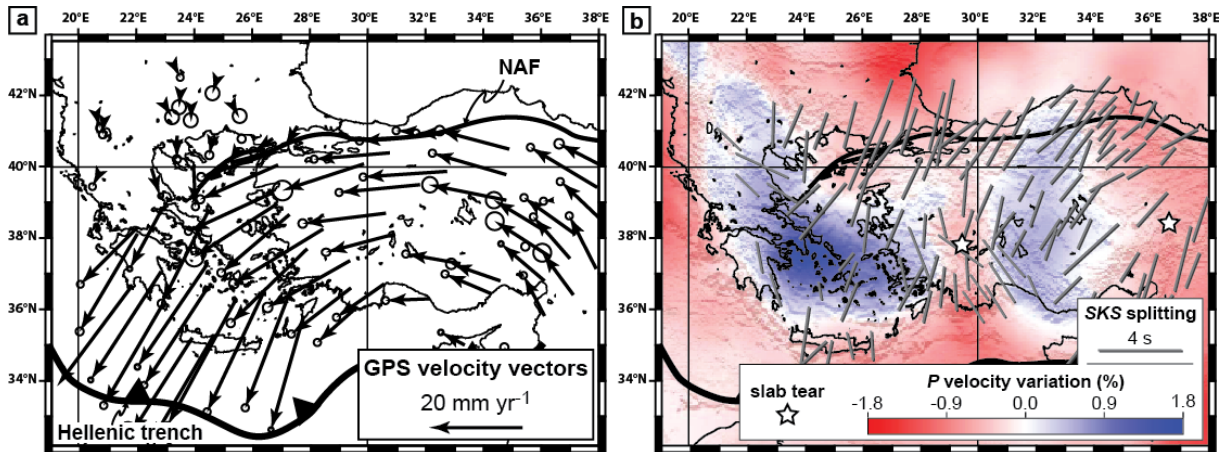
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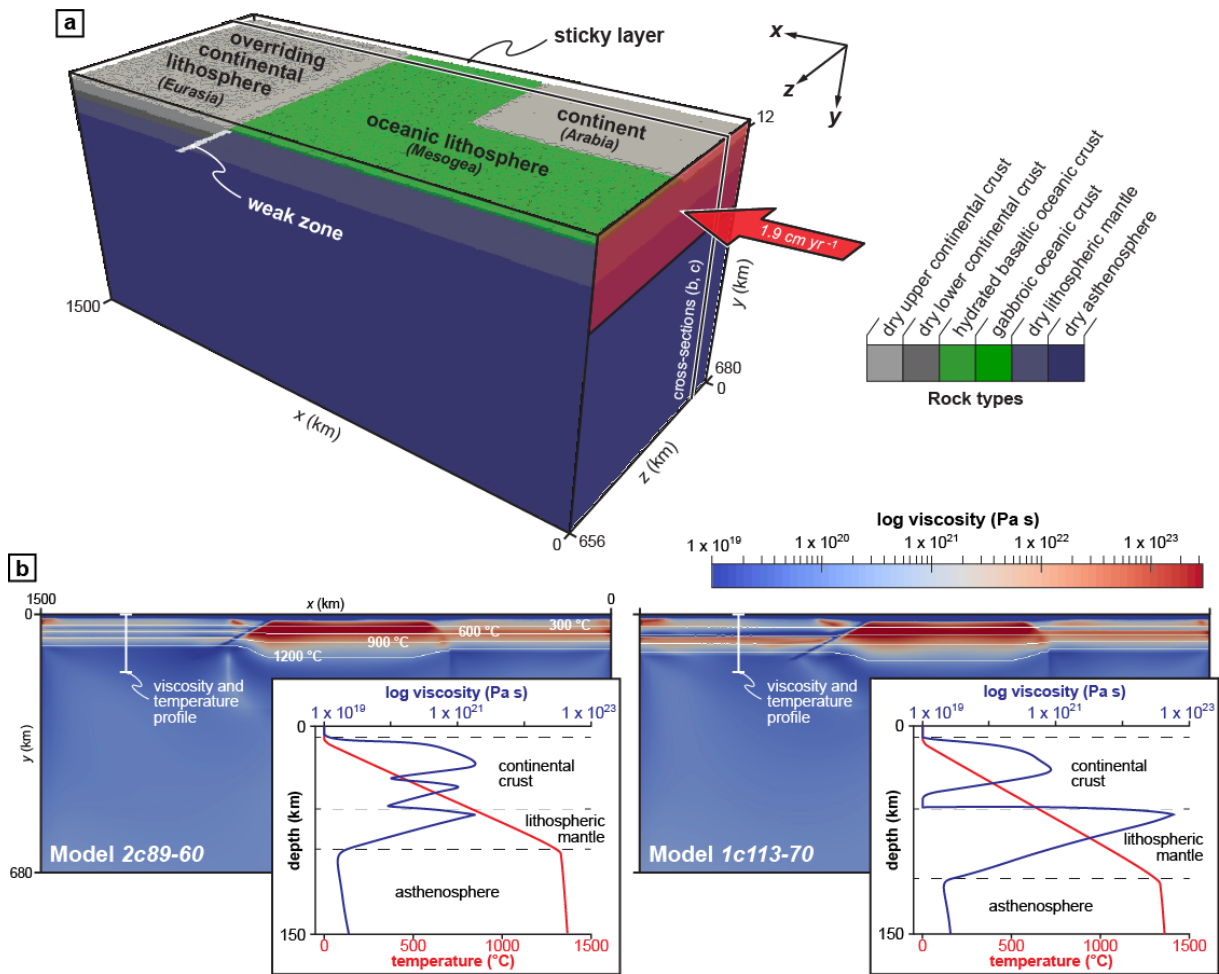


668

669 **Fig. 1:** Late Cenozoic geodynamic evolution of the eastern Mediterranean region. (a) Tectonic map of
 670 the eastern Mediterranean region showing the spatial distribution of the magmatism since the late
 671 Cretaceous. Red and yellow arrows show, respectively, the first-order southward migration of the
 672 magmatism since the late Cretaceous and the second-order southwestward migration of the
 673 magmatism during the Miocene. (b) Present-day N-S cross-section across the Rhodope-Aegean
 674 region. (c) Paleotectonic maps of the eastern Mediterranean region at 35, 15 and 5 Ma showing the
 675 spatial distribution of contemporaneous magmatic occurrences. Modified from *Menant et al.* [2016].

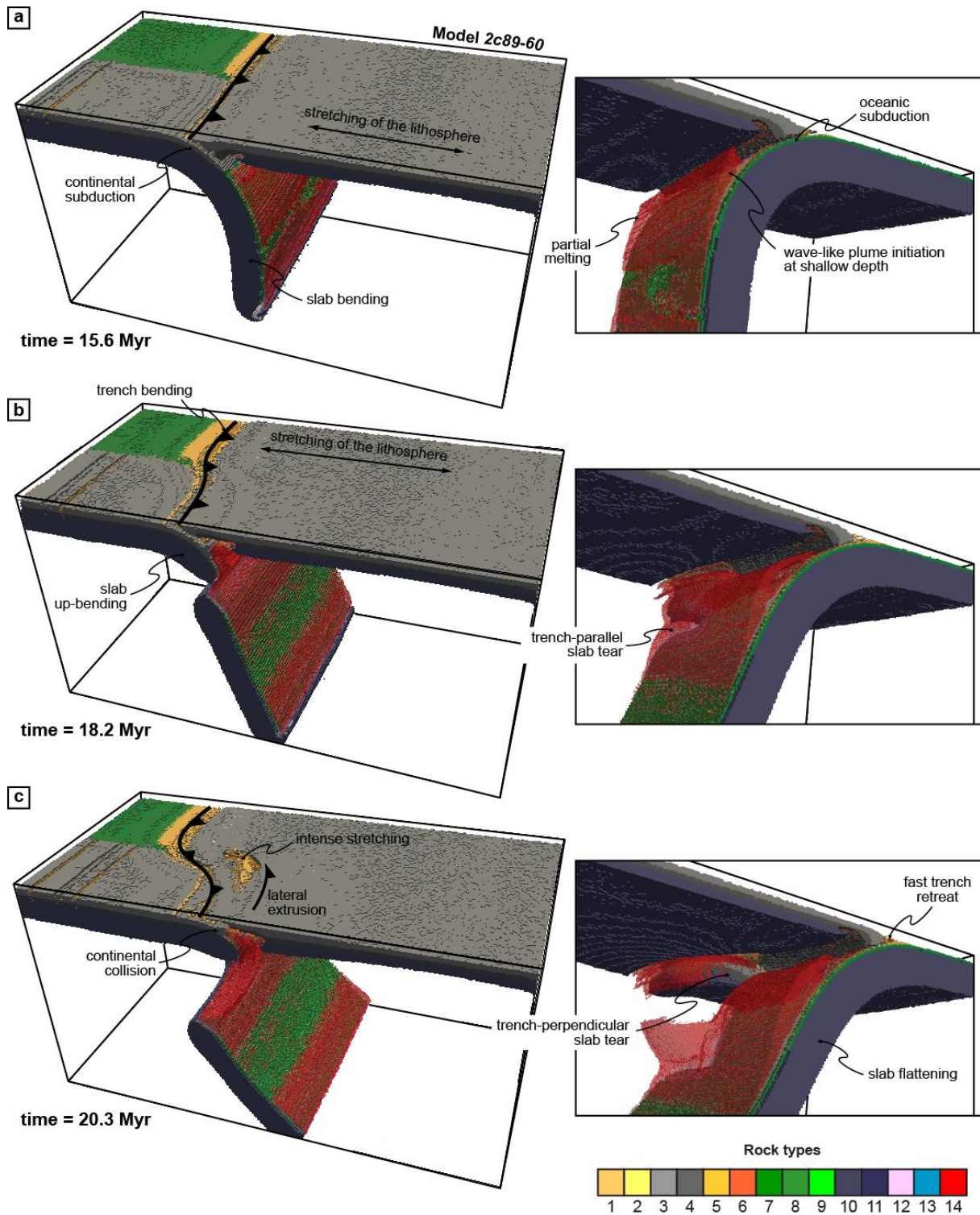


676
 677 **Fig. 2:** Geophysical maps of the eastern Mediterranean region. (a) GPS velocities relative to stable
 678 Eurasia with 1σ confidence ellipses [Reilinger *et al.*, 2006]. (b) Seismic tomographic model showing
 679 the V_p anomalies at ~ 150 km depth [Li *et al.*, 2008]. Slab tears suspected in the region are highlighted
 680 by white stars. Station-averaged SKS splitting data are also highlighted [Paul *et al.*, 2014].
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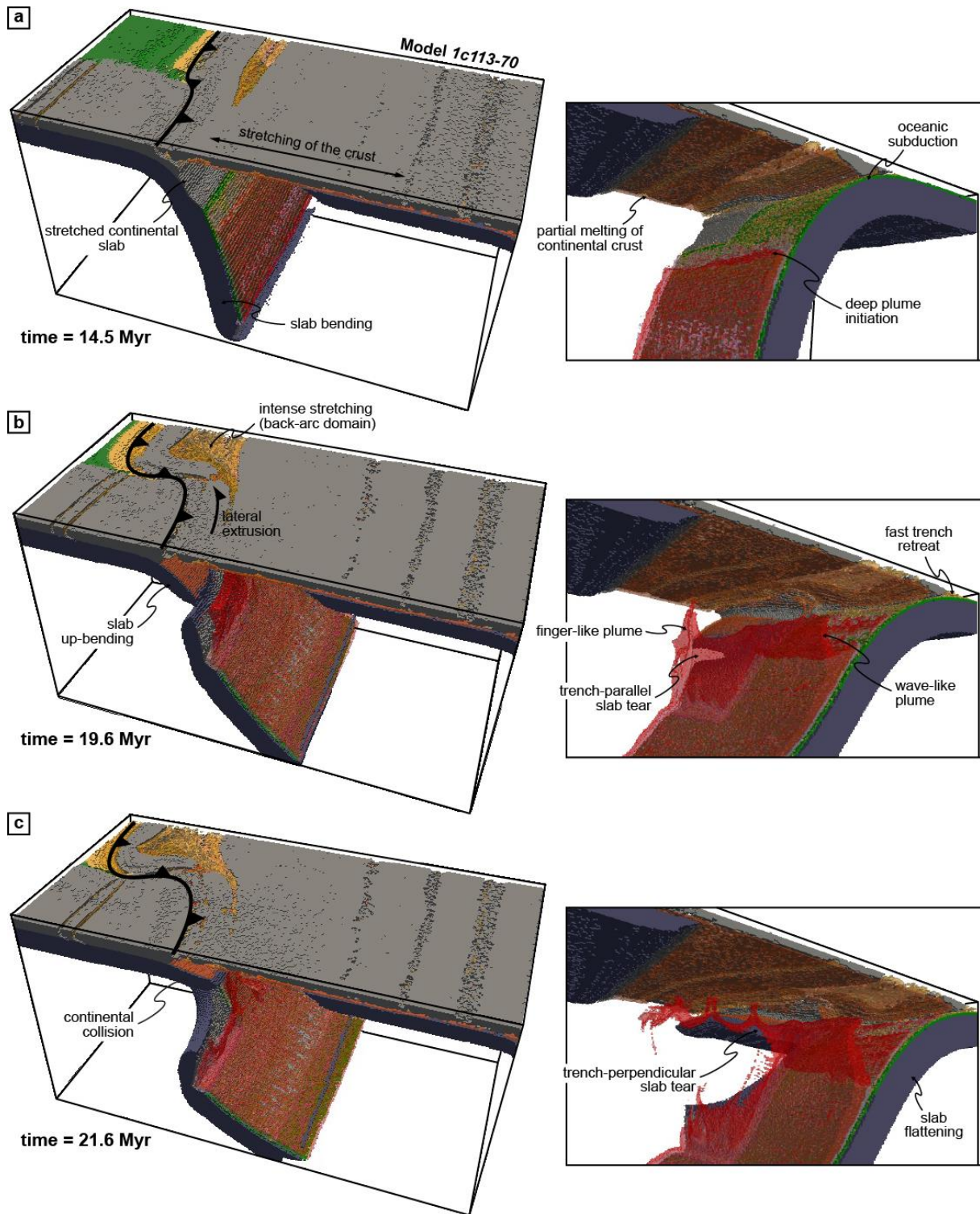
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Fig. 3: Initial model setup. (a) 3D model domain with colors representing the different rock types. Air and sea water phases (composing the sticky layer) are cut off for clarity. (b) x - y viscosity cross-sections of the initial model domain with a two-layer (left panel) and one-layer (right panel) continental crust. White lines represent 300, 600, 900 and 1200 °C isotherms. Vertical viscosity (blue) and temperature (red) profiles are shown as inset for each model.



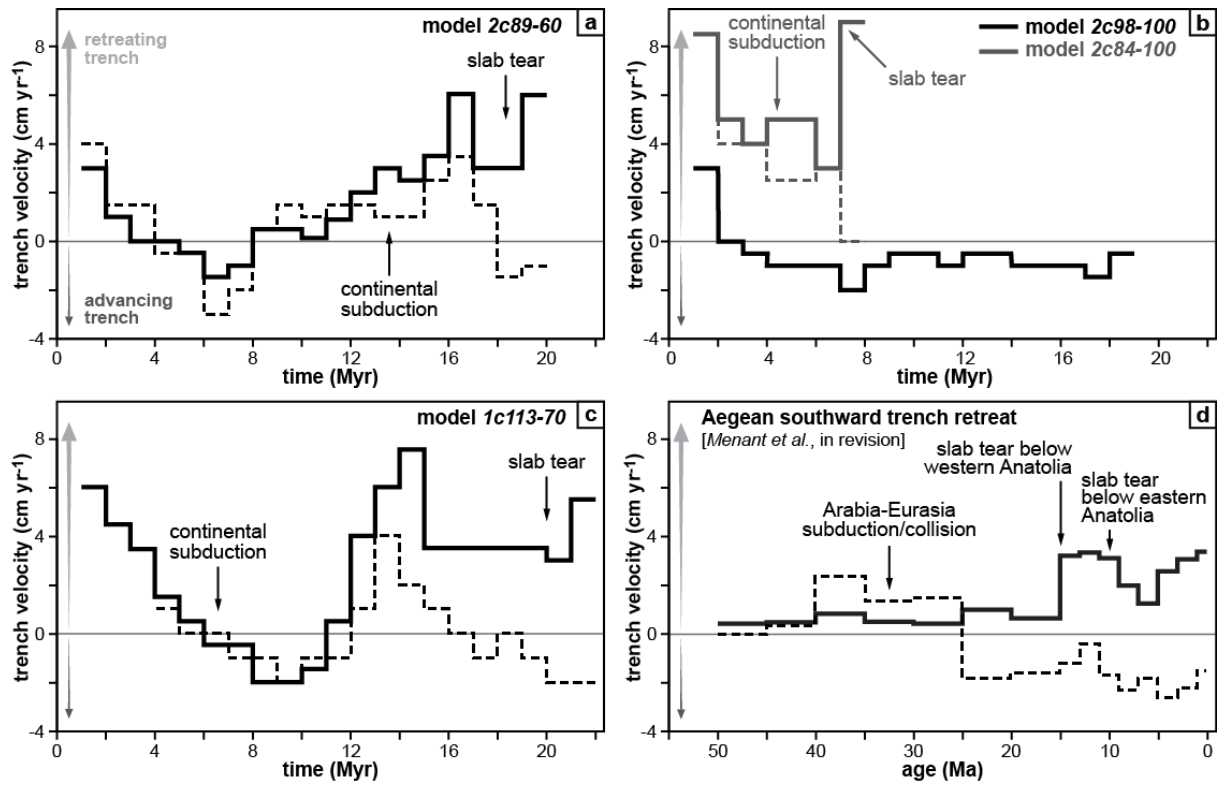
688

689 **Fig. 4:** Evolution of the reference model 2c93-60 (with a two-layer continental crust). (a, b, c) 3D
 690 lateral view of the subduction zone (left panels) and zoom on the mantle wedge (right panels) for
 691 selected time steps. Colors represent the different rock types: 1 – sediments; 2 – partially molten
 692 sediments; 3/4 – dry upper/lower continental crust; 5/6 – hydrated/partially molten continental crust; 7
 693 – dry oceanic crust; 8/9 – hydrated/partially molten oceanic crust; 10 – dry lithospheric mantle; 11 –
 694 dry asthenosphere; 12/13 – hydrated/serpentinized mantle; 14 – partially molten mantle.
 695 Asthenosphere and other phases are cut off for clarity.



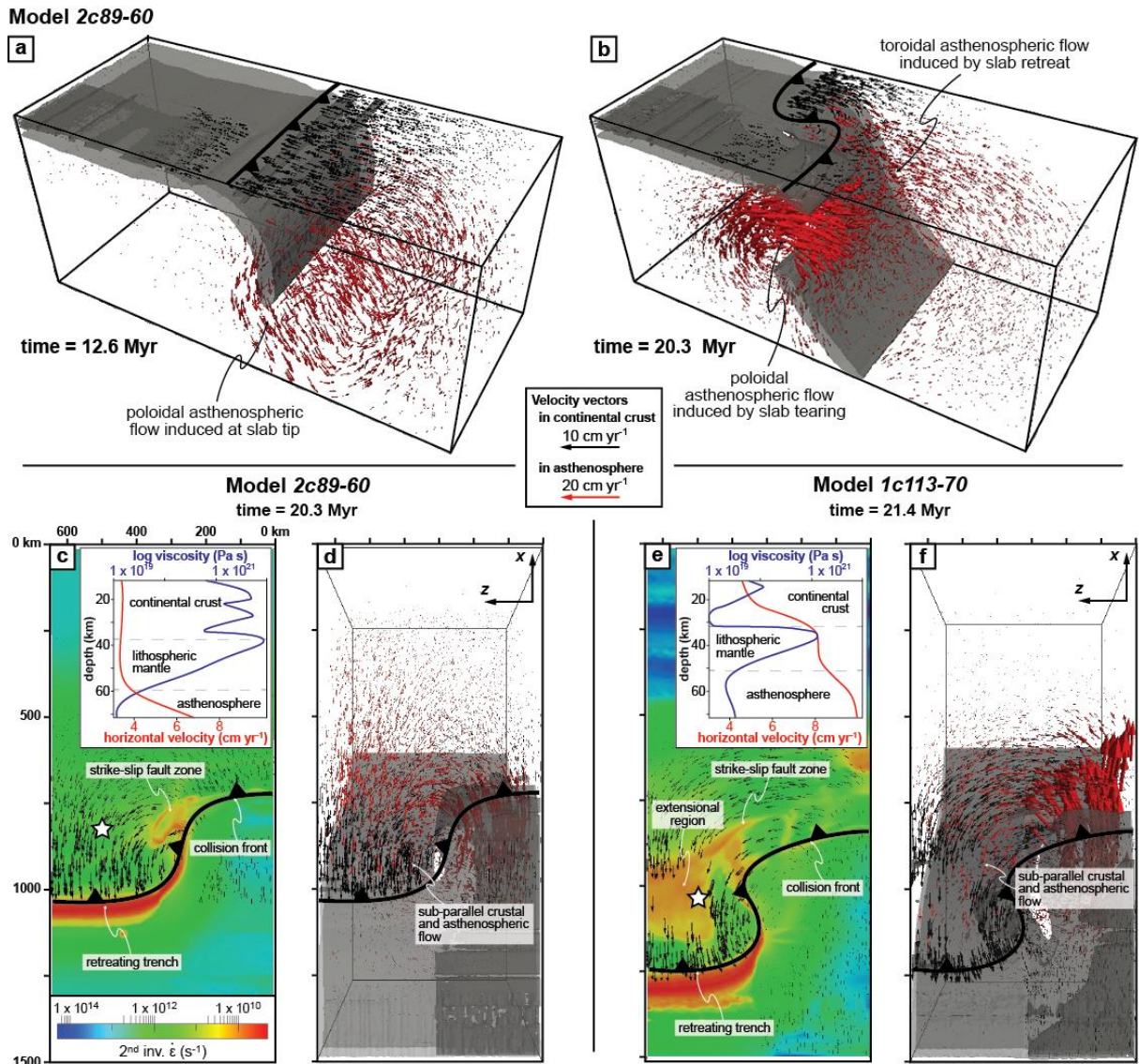
696

697 **Fig. 5:** Evolution of the model 1c113-70 (with a one-layer continental crust). (a, b, c) 3D lateral view
 698 of the subduction zone (left panels) and zoom on the mantle wedge (right panels) for selected time
 699 steps. Colors represent the different rock types (see Fig. 4 for detailed legend). Asthenosphere and
 700 other phases are cut off for clarity.

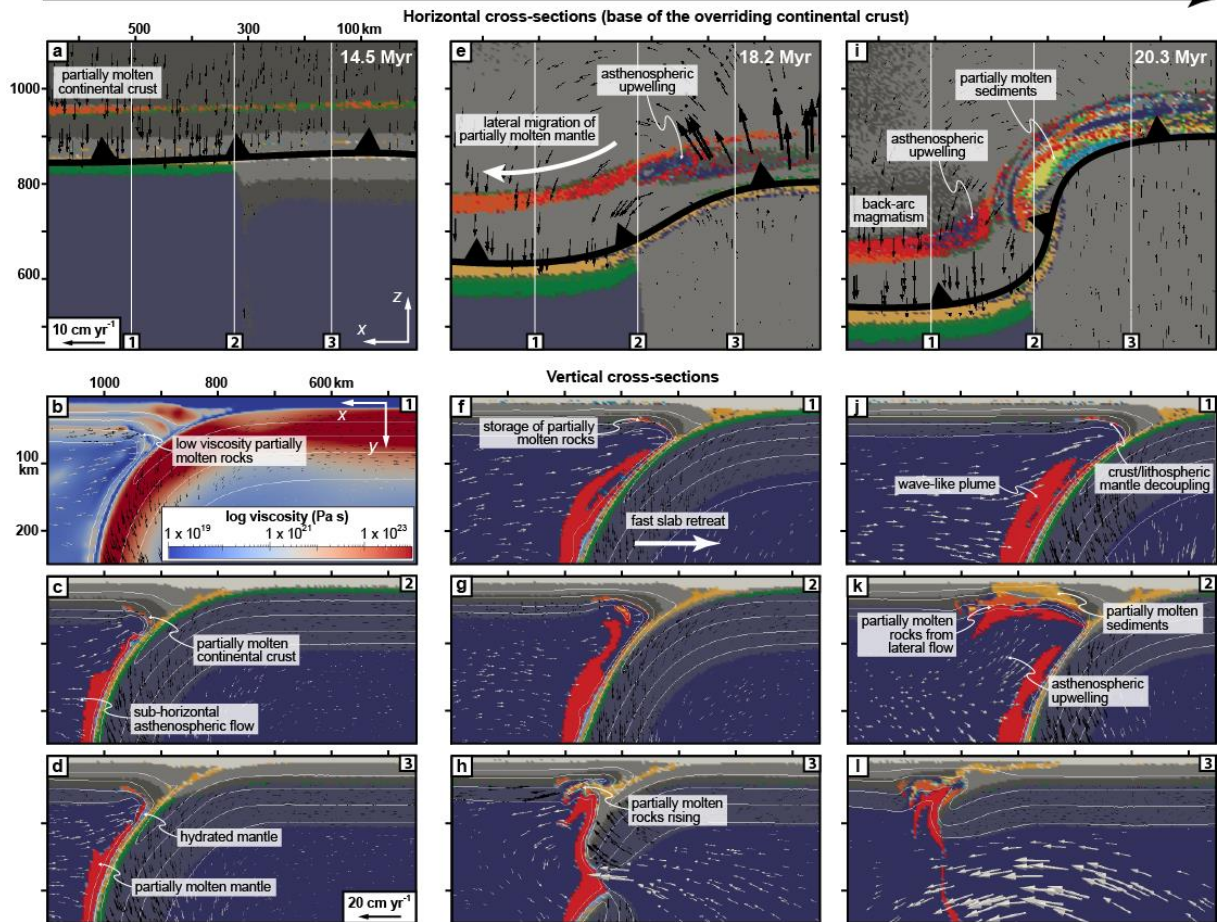


701

702 **Fig. 6:** Time-trench velocity diagram showing the trench migration of both oceanic and oceanic-
 703 continental subduction zones obtained from models 2c89-60 (a), 2c98-100, 2c84-100 (b) and 1c113-70
 704 (c) and from kinematic reconstructions of the eastern Mediterranean region (d) [Menant *et al.*, 2016].
 705 Solid and dashed lines correspond to only oceanic (or Hellenic) and oceanic-continent (or eastern
 706 Anatolia) subduction zones, respectively. Positive and negative distances correspond to retreating and
 707 advancing trench, respectively. The initiation of continental subduction and slab tearing processes are
 708 highlighted for each diagram.

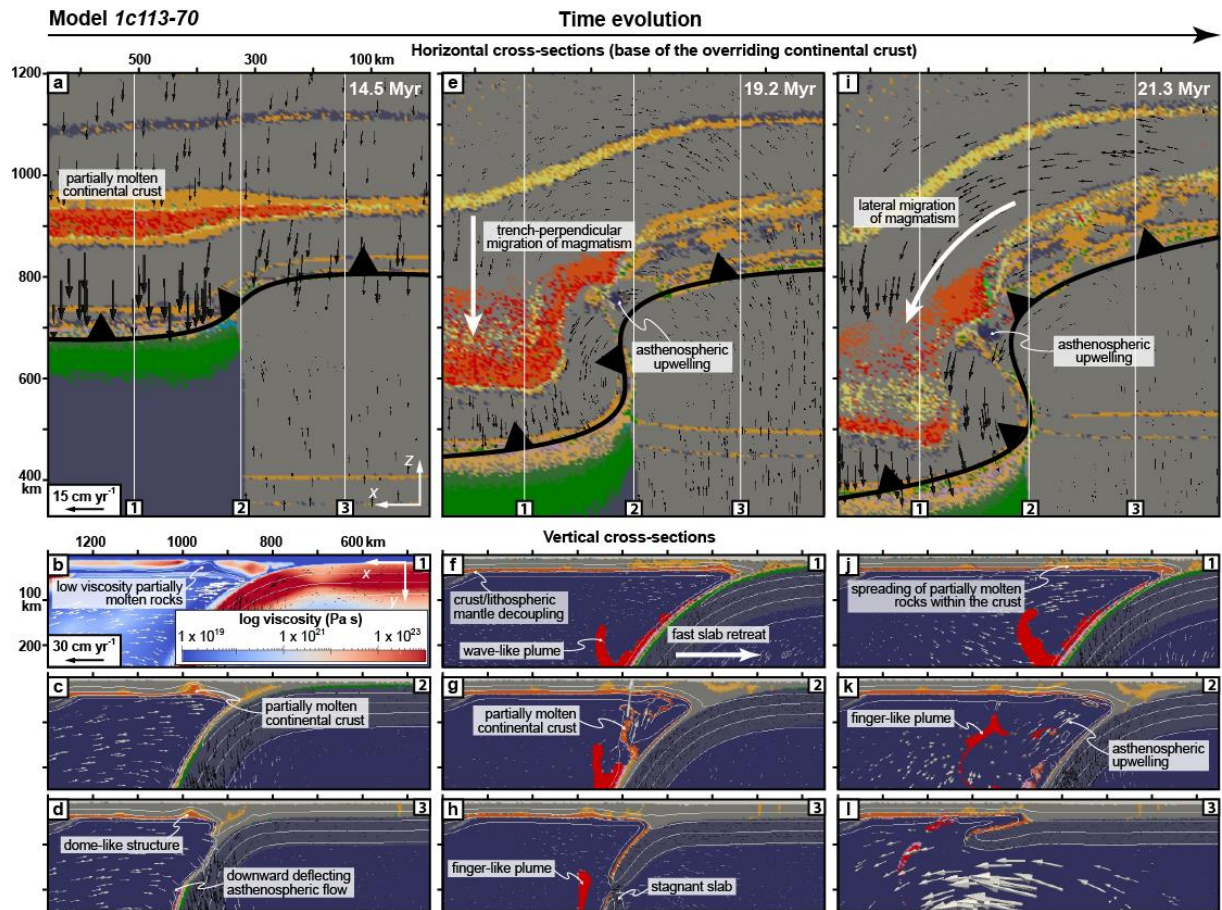


709
 710 **Fig. 7:** Asthenospheric and crustal flow evolution and crustal strain rate in performed numerical
 711 experiments. (a, b) 3D view of the subduction zone of model 2c93-60 at two selected time steps (i.e.
 712 before and after slab tearing) showing the slab geometry (iso-viscosity contour equal to 10^{23} Pa.s) and
 713 velocity vectors in the continental crust (black arrows) and asthenosphere (red arrows). (c, e) x - z plan
 714 view of selected time step showing the second invariant of the strain rate tensor in interacting
 715 lithospheres of models 2c93-60 and 1c113-70. Vertical viscosity (blue) and velocity (red) profiles are
 716 shown as inset for each extensional domain (see the white star for the location of the vertical profile).
 717 (d, f) x - z plan view of selected time step showing the velocity vectors in the continental crust and
 718 asthenosphere. Scale of velocity vectors is twice larger in continental crust than in asthenosphere.

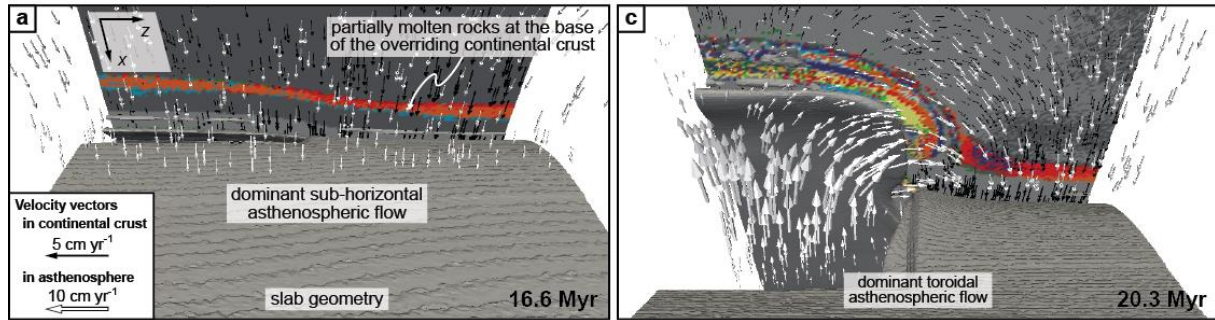


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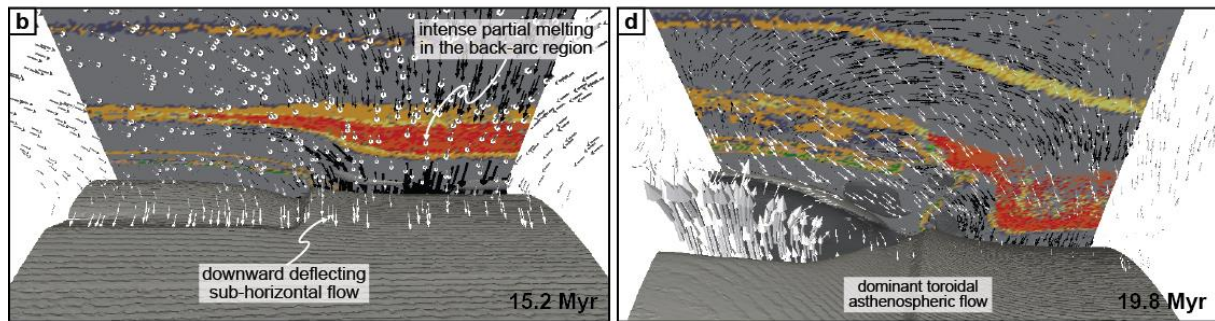
720 **Fig. 8:** Evolution of magmatism in reference model 2c93-60. x - z plan views (a, e, i) and x - y cross-
 721 sections (b, c, d, f, g, h, j, k, l) of the subduction zone for selected time steps. Colors represent the
 722 different rock types (see Fig. 4 for detailed legend), except for cross-section (b) where the viscosity is
 723 mapped. Black and white arrows represent velocity vectors in the continental crust (top panels) or
 724 lithospheric mantle (bottom panels) and the asthenosphere, respectively. Scale of velocity vectors is
 725 twice larger in top panels than in bottom panels. White lines (bottom panels) represent 300, 600, 900
 726 and 1200 °C isotherms.



Model 2c89-60



Model 1c113-70



735
 736 **Fig. 10:** View from below of the mantle wedge of selected models showing the relation between
 737 magmatic evolution at the base of the overriding crust and subduction-related asthenospheric flow for
 738 two selected time steps. (a, c) *x-z* plan view and slab geometry evolution of reference model 2c93-60.
 739 (b, d) *x-z* plan view and slab geometry evolution of reference model 1c113-70. On *x-z* plan view,
 740 colors represent the different rock types (see Fig. 4 for detailed legend). The slab geometry is
 741 visualized through an iso-viscosity contour equal to 10²³ Pa.s. Black and white arrows represent
 742 velocity vectors in the continental crust and the asthenosphere, respectively. Scale of velocity vectors
 743 is twice larger in continental crust than in asthenosphere.
 744

745 **Supplementary material**

746

747 **Numerical approach and governing equations**

748 *Conservation equations*

749 In our experiments, we model rocks as an incompressible fluid through the 3D mass
750 conservation equation:

$$751 \quad \nabla \cdot \vec{u} = \frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} + \frac{\partial v_z}{\partial z} = 0 \quad (1)$$

752 where v_x , v_y and v_z are the components of the velocity vector \vec{u} . The momentum of this
753 incompressible fluid thus occurs as a 3D creeping (or laminar) flow meaning that the viscous
754 forces are significantly higher than the inertial ones (i.e. Reynolds number $\ll 1$). This
755 momentum is then solved using Stokes equation:

$$756 \quad \frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xy}}{\partial y} + \frac{\partial \sigma_{xz}}{\partial z} = \frac{\partial p}{\partial x}$$
$$757 \quad \frac{\partial \sigma_{yx}}{\partial x} + \frac{\partial \sigma_{yy}}{\partial y} + \frac{\partial \sigma_{yz}}{\partial z} = \frac{\partial p}{\partial y} - g \rho \quad (2)$$
$$758 \quad \frac{\partial \sigma_{zx}}{\partial x} + \frac{\partial \sigma_{zy}}{\partial y} + \frac{\partial \sigma_{zz}}{\partial z} = \frac{\partial p}{\partial z}$$

759 where σ_{ij} are the components of the deviatoric stress tensor, p is the pressure, g is the
760 gravitational acceleration and ρ is the density (Table S1). Finally, heat diffusion in our
761 experiments is solved using the Lagrangian heat conservation equation:

$$762 \quad \rho C_P \frac{DT}{Dt} = - \frac{\partial q_x}{\partial x} - \frac{\partial q_y}{\partial y} - \frac{\partial q_z}{\partial z} + H_r + H_a + H_s \quad (3)$$

763 where C_p is the isobaric heat capacity, T is the temperature, t is the time, H_r is the
 764 radiogenic heat production, H_a is the adiabatic heat production, H_s is the shear heating and q_x ,
 765 q_y and q_z are the components of the 3D heat flow solved as:

$$\begin{aligned}
 766 \quad q_x &= -k \frac{\partial T}{\partial x} \\
 767 \quad q_y &= -k \frac{\partial T}{\partial y} \\
 768 \quad q_z &= -k \frac{\partial T}{\partial z}
 \end{aligned} \tag{4}$$

769 where k is the thermal conductivity depending on the temperature and rock composition
 770 (Table S1). Heat advection is then solved based on marker technics [Gerya and Yuen, 2007].
 771 Surface processes are implemented using a highly simplified gross-scale erosion-
 772 sedimentation law (see details in Gerya and Yuen [2007]). Considering these physical laws,
 773 the *I3ELVIS* code is able to model multiphase thermo-mechanical flow approximating ductile-
 774 brittle deformation of rocks for long time scales.

775

776 ***Fluid/melt dynamics: dehydration, fluid transport, partial melting and melt extraction***
 777 ***computing***

778 In our experiments, each defined lithology can be dry or hydrated, with pressure- and
 779 temperature-dependent stable mineralogical water content. Additional porous water content
 780 X_{H_2O} (wt.%) is prescribed in the sediments and hydrated oceanic crust and decreases linearly
 781 from the surface to 75 km depth mimicking the effects of low-temperature reactions:

$$782 \quad X_{H_2O} = X_{H_2O(0)} \left(1 - \frac{\Delta y}{75} \right) \tag{5}$$

783 where Δy is the depth below the surface and $X_{H_2O(0)}$ is the porous water content at the
 784 surface [Johnson and Pruis, 2003]. During subduction, the local water content of hydrated

785 sediments and oceanic crust can exceed their maximum water content generating
 786 independently moving water markers in the mantle wedge. The 3D velocity of these markers
 787 is then solved according to pressure gradient [*Faccenda et al.*, 2009]:

$$\begin{aligned}
 788 \quad V_{x(water)} &= v_x - A \frac{\partial p}{\partial x} \\
 789 \quad V_{y(water)} &= v_y - A \left(\frac{\partial p}{\partial y} - g \rho_{fluid} \right) \quad (6)
 \end{aligned}$$

$$790 \quad V_{z(water)} = v_z - A \frac{\partial p}{\partial z}$$

791 where ρ_{fluid} is the fluid density and A is the water percolation constant calculated as:

$$792 \quad A = \frac{v_{percolation}}{g(\rho_{mantle} - \rho_{fluid})} \quad (7)$$

793 where $v_{percolation}$ is the presumed standard water percolation velocity in the mantle
 794 [*Nikolaeva et al.*, 2008] and ρ_{mantle} is the mantle density. Mobile water is then absorbed by the
 795 surrounding mantle or crustal rocks by hydration process depending on pressure, temperature
 796 and rock composition (see details in [*Zhu et al.*, 2013]).

797 Partial melting process is also included in our experiments and the volumetric melt fraction
 798 is calculated for all rocks, excepted hydrated mantle, as:

$$\begin{aligned}
 799 \quad M_0 &= 0 \quad \text{at } T < T_{solidus} \\
 800 \quad M_0 &= \frac{T - T_{solidus}}{T_{liquidus} - T_{solidus}} \quad \text{at } T_{solidus} < T < T_{liquidus}
 \end{aligned}$$

801 (8)

$$802 \quad M_0 = 1 \quad \text{at } T > T_{liquidus}$$

803 where M_0 is the volumetric fraction of melt with no previous melt extraction, $T_{solidus}$ and
 804 $T_{liquidus}$ are, respectively, the solidus and liquidus temperatures depending on the pressure and

805 rock composition (Table S1). For the hydrated mantle, the degree of partial melting also
806 depends on the water content according to the parameterization of *Katz et al.* [2003] that is
807 constrained by thermodynamic experimental modeling. Resulting partially molten rocks can
808 then undergo a succession of melt extraction episodes depending on their melt extraction
809 threshold M_{max} and non-extractable melt fraction M_{min} , both equal to 1 %. Only the exceeding
810 volumetric melt fraction (i.e. > 1 %) is then extracted from partially molten rocks and tracked
811 by partially molten rock markers. For each extraction episode, the melt fraction M_{ext} recorded
812 in the partially molten rock markers is calculated as:

$$813 \quad M_{ext} = M - M_{min} \quad (9)$$

814 where M is the volumetric melt fraction in partially molten rock which accounts for
815 previous extraction episodes and is calculated as:

$$816 \quad M = M_0 - \sum M_{ext} \quad (10)$$

817 where $\sum M_{ext}$ is the total melt fraction from previous extraction episodes. M varies then
818 dynamically (Eqs. 8 and 10) until remaining solid rock is considered as refractory (i.e. unable
819 to undergo additional partial melting) when $\sum M_{ext} > M_0$. Partially molten rock markers
820 resulting from these extraction episodes are then instantaneously transmitted to the surface as
821 free melt is assumed to propagate upward much faster than rocks deform [*Hawkesworth,*
822 1997]. At the surface, these markers create volcanic rocks characterized by their volume and
823 composition (i.e. partially molten source).

824

825 ***Rock properties and rheology***

826 The physical properties of rocks used in our experiments are given in Table S1. Among
827 them, the density of solid rocks ρ_{solid} is calculated as:

828 $\rho_{solid} = \rho_{0solid} (1 - \alpha (T - 298)) (1 + \beta (p - 0.1))$ (11)

829 where ρ_{0solid} is the standard density of solid rocks and α and β are respectively the thermal
 830 expansion and compressibility of rock. For partially molten rocks, their effective density ρ_{eff} is
 831 calculated as:

832 $\rho_{eff} = \rho_{solid} \left(1 - M + M \frac{\rho_{0molten}}{\rho_{0solid}}\right)$ (12)

833 where $\rho_{0molten}$ is the standard density of molten rock. Non-Newtonian visco-plastic
 834 rheologies used in our experiments are implemented by both ductile and brittle/plastic
 835 experimentally constrained laws. Thus, the effective viscosity for dislocation creep depending
 836 on the strain rate, pressure and temperature is calculated as:

837 $\eta_{creep} = \dot{\epsilon}_{II}^{\frac{1-n}{n}} A_D^{\frac{-1}{n}} \exp\left(\frac{E+P V}{n R T}\right)$ (13)

838 where $\dot{\epsilon}_{II}$ is the second invariant of the strain rate tensor, n (creep exponent), A_D (pre-
 839 exponential factor), E (activation energy) and V (activation volume) are experimentally
 840 determined parameters specific to applied viscous flow law (i.e. wet quartzite, plagioclase
 841 (An₇₅), dry olivine and wet olivine flow laws; Table S1) [Ranalli, 1995]. R is the gas constant.
 842 This ductile behavior is limited with a brittle/plastic one, implemented by using the Drucker-
 843 Prager criterion σ_{yield} (i.e. plastic strength) [Ranalli, 1995] as:

844 $\eta_{creep} \leq \frac{\sigma_{yield}}{2 \dot{\epsilon}_{II}}$ (14)

845 where σ_{yield} is calculated as:

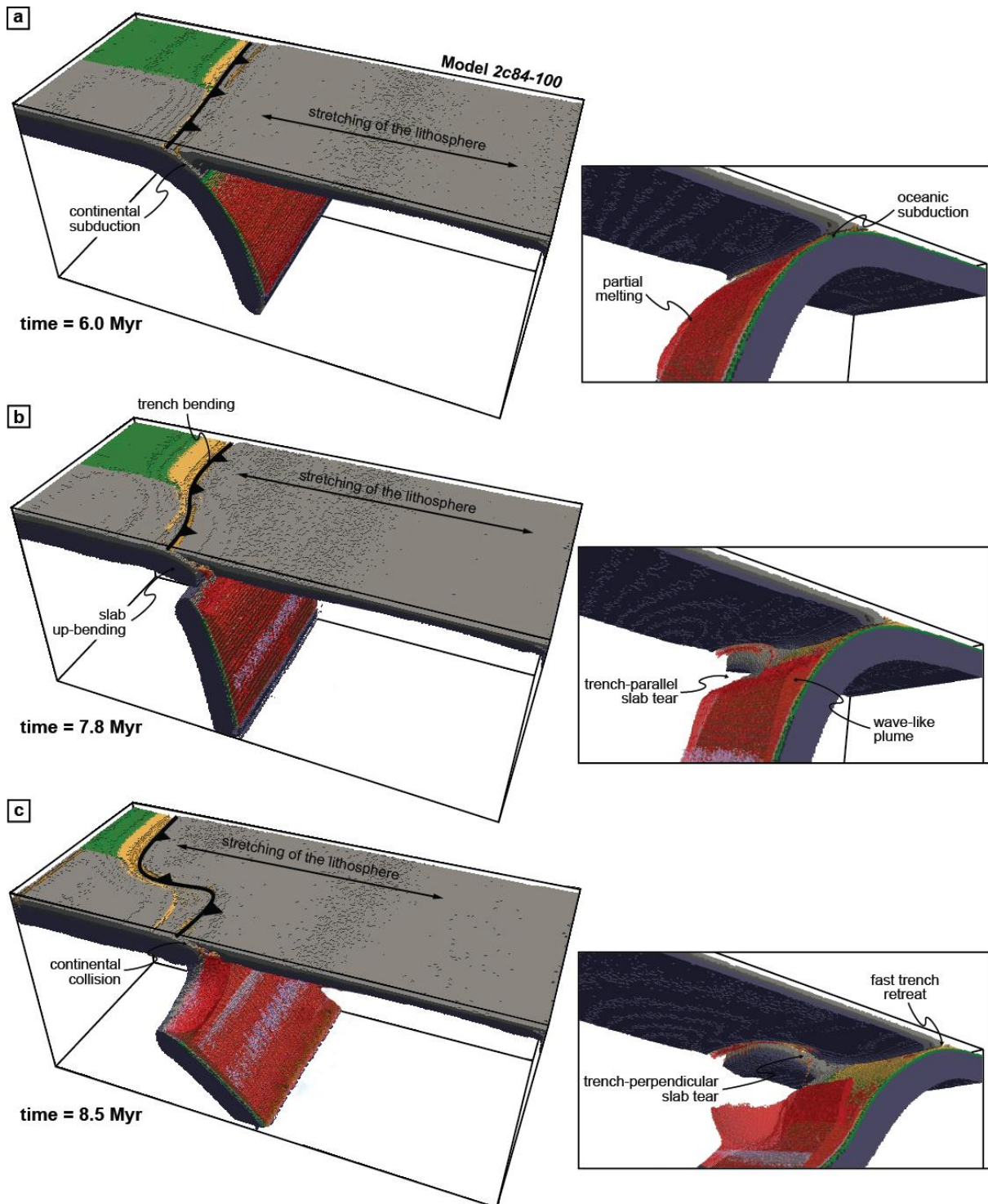
846 $\sigma_{yield} = c + p \sin(\varphi)$ (15)

847 where c is the cohesion of the rock and φ is the effective internal friction angle depending
848 on the fluid and melt contents (see details in *Gerya and Meilick* [2011]). The lower and upper
849 cut-off values for the viscosities of all rock compositions in our numerical experiments are
850 10^{19} and 10^{24} Pa s, respectively.

851

852 **Supplementary figures**

853



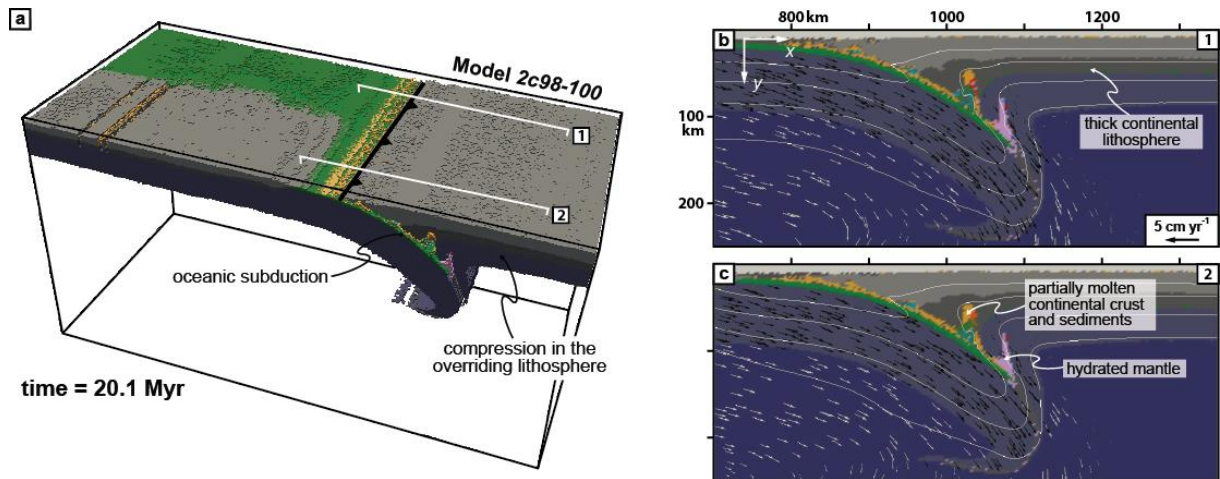
854

855 **Fig. S1:** Evolution of the model 2c84-100. a, b, c) 3D lateral view of the subduction zone (left panels)

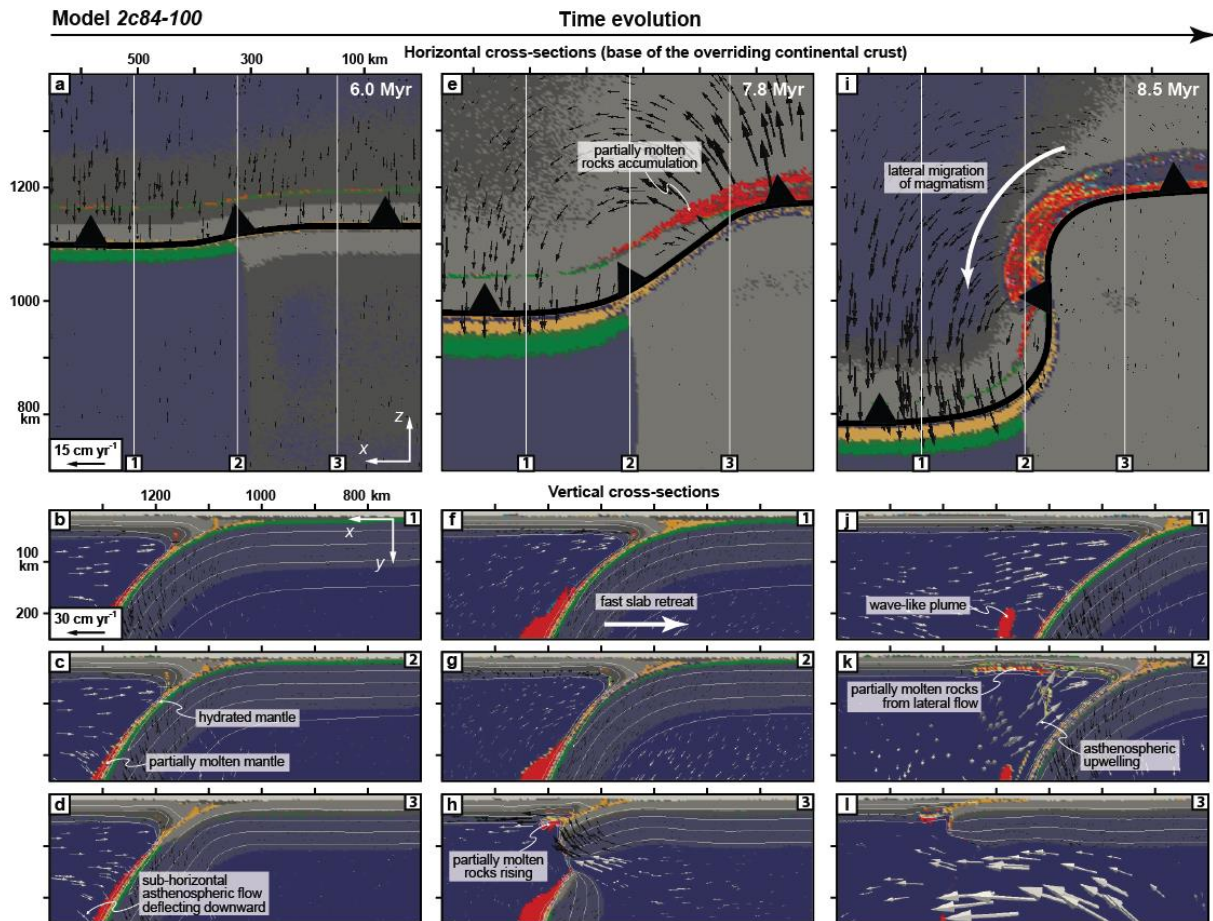
856 and zoom on the mantle wedge (right panels) for selected time steps. Colors represent the different

857 rock types (see Fig. 4 for detailed legend). Asthenosphere and other phases are cut off for clarity.

858



859
 860 **Fig. S2:** Subduction geometry of the model 2c98-100 displaying a relatively strong coupling between
 861 both interacting lithospheres. a) 3D lateral view of the subduction zone highlighting the compressional
 862 tectonics in the overriding continental lithosphere. b, c) *x-y* cross-sections of the subduction zone
 863 highlighting the restrained magmatic activity in the mantle wedge and the low velocity field in both
 864 the subducting lithosphere and the underlying asthenosphere (see Fig. 4 for detailed legend).
 865



866

867 **Fig. S3:** Evolution of magmatism in model 2c84-100. *x-z* plan views (a, e, i) and *x-y* cross-sections (b,

868 c, d, f, g, h, j, k, l) of the subduction zone for selected time steps. Colors represent the different rock

869 types (see Fig. 4 for detailed legend). Black and white arrows represent velocity vectors in the

870 continental crust (top panels) or lithospheric mantle (bottom panels) and the asthenosphere,

871 respectively. Scale of velocity vectors is twice larger in top panels than in bottom panels. White lines

872 (bottom panels) represent 300, 600, 900 and 1200 °C isotherms.

873

874 **Supplementary tables**

875

Material	Density ρ_0 (kg m ⁻³)	Isobaric heat capacity C_p (J kg ⁻¹ K ⁻¹)	Thermal conductivity k (W m ⁻¹ K ⁻¹)	Radiogenic heat production H_r (mW m ⁻³)	$T_{solidus}$ (K)	$T_{liquidus}$ (K)	Thermal expansion α (K ⁻¹)	Compressibility β (MPa ⁻¹)	Viscous flow law	Creep exponent n	Activation energy E (kJ mol ⁻¹)	Activation volume V (cm ³ mol ⁻¹)
Air	1	100	20	0			0	0	Air	1	0	0
Sea water	1000	3330	20	0			0	0	Water	1	0	0
Sediments	2600 (solid)	1000	0.64+807/($T-77$)	2	889+179/($P+54$)+20200/($P+54$) ² at $P < 1200$ MPa; 831+0.06 P at $P > 1200$ MPa	1262+0.09 P	3×10 ⁻⁵	3×10 ⁻³	wet quartzite	2.3	154	0
	2400 (molten)	1500								1		
Upper continental crust	2750 (solid)	1000	0.64+807/($T-77$)	1	889+179/($P+54$)+20200/($P+54$) ² at $P < 1200$ MPa; 831+0.06 P at $P > 1200$ MPa	1262+0.09 P	3×10 ⁻⁵	3×10 ⁻³	wet quartzite	2.3	154	0
	2700 (hydrated)									1		
	2400 (molten)									1500		
Lower continental crust	2950 (solid)	1000	0.64+807/($T-77$)	0.5	1327.15+0.0906 P	1423+0.105 P	3×10 ⁻⁵	3×10 ⁻³	wet quartzite	3.2	238	0
	2900 (hydrated)									1		
	2400 (molten)									1500		
Oceanic crust	3000 (solid)	1000	1.18+474/($T-77$)	0.25	1327.15+0.0906 P	1423+0.105 P	3×10 ⁻⁵	3×10 ⁻³	wet quartzite	3.2	238	0
	3000 (hydrated)									2.3		
	2900 (molten)									1500		
Mantle	3300 (solid)	1000	0.73+1293/($T-77$)	0.022	n/a	n/a	3×10 ⁻⁵	3×10 ⁻³	dry olivine	3.5	532	0.8
	3300 (hydrated/serpentinized)									4		
	2900 (molten)									1500		
Weak zone	3300	1000	0.73+1293/($T-77$)	0.05	n/a	n/a	3×10 ⁻⁵	3×10 ⁻³	wet olivine	4	470	0.8

876

877 **Table S1:** Material properties used in the numerical experiments. See details in the text.

Models with two-layer continental crust		thermal age of the oceanic lithosphere			
		40 Ma	60 Ma	80 Ma	100 Ma
depth of the 1300 °C isotherm in the overriding lithosphere	84 km	2c84-40	2c84-60	2c84-80	2c84-100
	89 km	2c89-40	2c89-60	2c89-80	2c89-100
	93 km	2c93-40	2c93-60	2c93-80	2c93-100
	98 km	2c98-40	2c98-60	2c98-80	2c98-100

878

Models with one-layer continental crust		thermal age of the oceanic lithosphere		
		60 Ma	70 Ma	80 Ma
depth of the 1300 °C isotherm in the overriding lithos.	111 km	1c111-60	1c111-70	1c111-80
	113 km	1c113-60	1c113-70	1c113-80
	115 km	1c115-60	1c115-70	1c115-80

879

880

881 **Table S2:** Range of performed models with variations in the continental crust structure and
882 geothermal gradient in the continental and oceanic lithospheres leading to explore the role of
883 lithosphere rheology on subduction dynamics and magmatic evolution. The overriding lithosphere in
884 dark grey, light grey and white models undergoes, respectively, a compressional, minor extensional
885 and major extensional tectonics. Models in bold are described in the text.

886