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► To cite this version:

Vincent Roche, Pietro Sternai, Laurent Guillou-Frottier, Armel Menant, Laurent Jolivet, et al.. Emplacement of metamorphic core complexes and associated geothermal systems controlled by slab dynamics. *Earth and Planetary Science Letters*, 2018, 498, pp.322-333. 10.1016/j.epsl.2018.06.043 . insu-01856826

HAL Id: insu-01856826

<https://insu.hal.science/insu-01856826>

Submitted on 14 Aug 2018

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4 *Emplacement of metamorphic core complexes and associated geothermal*
5 *systems controlled by slab dynamics*

6

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33 <https://doi.org/10.1016/j.epsl.2018.06.043>

34

35 **ABSTRACT**

36 Slab rollback results in the development of low-angle normal faults (detachments) and
37 metamorphic core complexes (MCCs) in back-arc domains. Although the mechanical
38 consequences of slab dynamics on lithospheric and crustal behaviors have already been studied,

39 thermal effects have not been investigated yet. This study shows that slab rollback produces
40 lithospheric-scale thermal perturbations intrinsically associated with emplacement of amagmatic
41 high-enthalpy geothermal systems. Using a multi-scale numerical modeling approach, with
42 lithospheric-scale 3-D thermo-mechanical models of subduction, and 2-D models of fluid flow
43 at the scale of detachments, we demonstrate that subduction-induced extensional tectonics
44 controls the genesis and distribution of crustal-scale thermal domes from the base of the crust,
45 and the location of high-energy geothermal systems. We find that when slab tearing occurs, Moho
46 temperatures can temporarily increase by up to 250 °C due to significant shear heating in the
47 flowing upper mantle. Associated thermal anomalies (with characteristic width and spacing of
48 tens and hundreds of km, for crustal and lithospheric scales, respectively) then migrate
49 systematically toward the retreating trench. These thermal domes weaken the crust, localize
50 deformation and enhance the development of crustal-scale detachments. These thermo-
51 mechanical instabilities mimic genesis of high-temperature MCCs with migmatitic cores in the
52 back-arc domain, such as those of the Menderes (western Anatolia, Turkey) and Larderello
53 (southern Tuscany) provinces in the Mediterranean realm, and those in the Basin and Range
54 (western United States), where detachments control the bulk of the heat transport. At the scale of
55 MCCs, the bulk fluid flow pattern is controlled by topography-driven flow while buoyancy-
56 driven flow dominates within the permeable detachments, focusing reservoir location of high-
57 energy geothermal systems at shallow depth beneath the detachments.

58

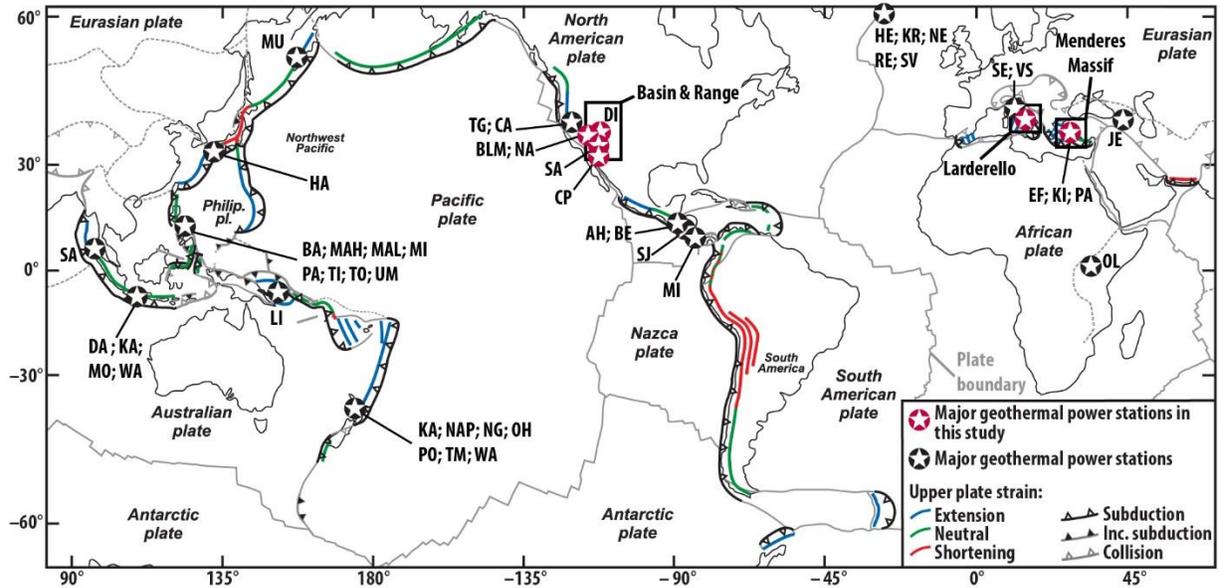
59 **1. INTRODUCTION**

60 The development of geothermal power plants has been increasing since the 1970s,
61 building upon more than 100 years of history in geothermal energy extraction. Currently, the

62 installed global capacity is estimated at 12.6 GWe (Gigawatt electrical), and forecasts for 2050
63 point to a worldwide capacity of 140 GWe, approximately 8.3% of total world electricity
64 production (Bertani, 2016). Many of the high enthalpy geothermal resources (HEGRs) that make
65 such an ambitious goal possible are located in the vicinity of subduction zones and volcanic arcs,
66 where both magmatic and tectonic processes operate (Fig. 1). Others are located in “amagmatic”
67 provinces such as the Menderes Massif in Western Turkey and the Basin and Range in the
68 Western United States. Particularly noteworthy for this study is that while geothermal systems
69 associated with magmatism have been studied in detail (e.g. Cumming, 2009) those located in
70 “amagmatic” provinces have received less attention (Moeck *et al.*, 2014; Roche *et al.* submitted).

71 Recent to present (i.e. Pliocene-Quaternary) magmatism in the upper crust across the
72 Menderes Massif and the Basin and Range is rare (Blackwell *et al.*, 2009; Faulds *et al.*, 2010)
73 compared to that of the Larderello geothermal field, located in the southern Tuscany (i.e. Italy)
74 (Santilano *et al.*, 2015) (Fig. 1). Open discussions on the possible role of hidden magmatic
75 intrusions on these geothermal systems remains debated. Even if present, however, magmatic
76 intrusions alone cannot explain the extent of these geothermal provinces (several thousand km²
77 each), and therefore cannot account for high concentration of HEGRs. We must thus identify
78 other sources of heat (deeper and larger-scale) possibly associated with deep, large-scale
79 geodynamic processes involved by the nearby subduction systems.

80



81

82 Figure 1: The major subduction zones on Earth (modified from Schellart *et al.* 2007). Black stars
 83 show the locations of major geothermal power stations that are larger than 50 MWe that are
 84 currently operational or under construction. Red stars are reserved for highlighting those within
 85 the study areas mentioned in this study. Geothermal stations: JE, Jermaghbyur (Armenia); Mi,
 86 Miravalles (Costa Rica); AH, Ahuachapán, BE, Berlín (El Salvador); HE, Hellisheidi, KR,
 87 Krafla, NE, Nesjavellir, RE, Reykjanes, SV, Svartsengi (Iceland); DA, Darajat, KA, Kamojang,
 88 MO, Mount Salak, SA, Sarulla, WA, Wayang Windu (Indonesia); LA, Larderello, SE,
 89 Serrazzano, VS, Valle Secolo (Italy); HA, Hatchobaru (Japan); OL, Olkaria (Kenya); CP, Cerro
 90 Prieto (Mexico); KA, Kawerau, NAP, Nga Awa Purua, NG, Ngatamariki, OH, Ohaaki, PO,
 91 Poihipi, TM, Te Mihi, WA, Wairakei (New Zealand); SJ, San Jacinto Tizate (Nicaragua); LI,
 92 Linhir (Papua New Guinea); BA, Bacman I, MAH, Mahanagdong, MAL, Malitbog, MI,
 93 Mindanao I-II, PA, Palinpinon I-II, TI, Tiwi A-B-C, TO, Tongonan 1, UM, Upper Mahiao
 94 (Philippines); MU, Mutnovskaya, Russia; EF, Efeler, KI, Kızıldere, PA, Pamukören (Turkey);
 95 BLM, CA, Calistoga, DV, Dixie Valley, NA, Navy, SA, Salton Sea, TG, The Geysers (United
 96 States).

97

98 Slab rollback induces lithospheric extension in the overriding plate where low-angle
 99 normal faults (detachments) control both the exhumation of metamorphic core complexes
 100 (MCCs) and the magma ascent and/or fluid circulation (e.g. Reynolds and Lister, 1987; Huet *et*

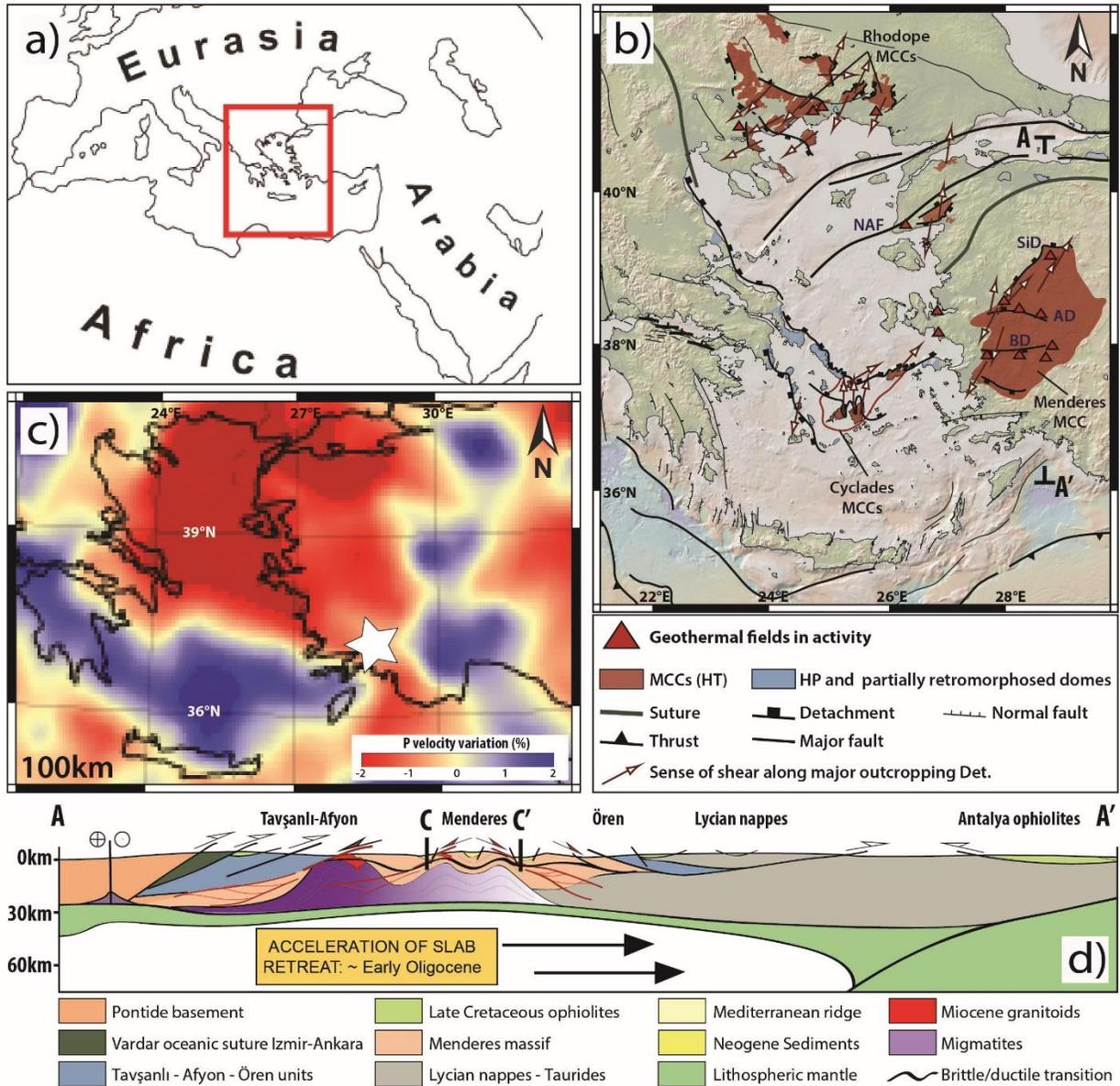
101 *al.*, 2011). Because detachments may correspond to permeable structures deeply rooted down to
102 the brittle-ductile transition (Famin *et al.*, 2004; Mezri *et al.*, 2015), active extensional domains
103 in back-arc regions could represent favorable settings for HEGRs. The possible sources of heat
104 responsible for regional high temperature-low pressure (HT-LP) metamorphic overprint recorded
105 in these MCCs include (i) heating associated with thermal diffusion of excess heat generated by
106 nappe stacking (increased radiogenic crustal heat, e.g. Bousquet *et al.*, 1997), (ii) shear-heating
107 in the mantle (e.g. Schubert and Yuen, 1978), or (iii) advection of hot asthenosphere to shallow
108 depths during slab retreat (Wannamaker *et al.*, 2008; Jolivet *et al.*, 2015). However, no unifying
109 mechanism responsible for both the generation of MCCs and emplacement of high-enthalpy
110 geothermal systems in the same regions has yet been recognized. This is however a crucial
111 question since HEGRs represent a major economic interest in terms of exploration for carbon-
112 free energy resources.

113 Here, we first document the self-consistent formation of crustal domes in the overriding
114 plate as a result of thermo-mechanical instabilities in 3-D numerical simulations of ocean-
115 continent subduction dynamics. These models provide crucial thermal constraints for the mantle
116 and crust, and show the importance of shear heating and fast mantle flow on the overall heat
117 budget. Subduction-induced thermal signature in the overriding crust obtained from these
118 experiments is then used as basal thermal boundary condition in 2-D numerical models dedicated
119 to the understanding of fluid flow in the upper crust in presence of detachments that accommodate
120 the formation of these domes. Results show that deep upper crustal hot fluids are drained upward
121 by the permeable detachment. These results will be first compared with geological observations
122 from the Menderes Massif, and then in the discussion, with other cases in the Mediterranean
123 realm (Anatolia and Tuscany) and in the western United States.

124

125 **2. GEODYNAMIC AND GEOTHERMAL SETTINGS OF THE MENDERES MASSIF**

126 During the Cenozoic, the eastern Mediterranean region (Figs. 2a and 2b) has undergone
127 a two-step tectono-metamorphic evolution. In the late Cretaceous-Eocene, the convergence of
128 Africa and Eurasia first led to the closure of the Izmir-Ankara Ocean. At this time, the accretion
129 of subducting continental and oceanic tectonic units (e.g. [Jolivet and Brun, 2010](#)) led to formation
130 of a south-verging crustal-scale orogenic wedge. Since the Oligo-Miocene, collapse of the
131 Hellenides-Taurides belt in this region is mainly controlled by the southward retreat of the
132 African slab, further accelerated in the middle Miocene by a major slab tearing event evidenced
133 by tomographic models below western Turkey (Fig. 2c) (e.g. [Piromallo and Morelli, 2003](#)).
134 Extension in the overriding plate has thus led to exhumation of different MCCs such as the
135 Cyclades in the Aegean Sea and the Menderes Massif in western Anatolia, accommodated by
136 crustal-scale low-angle normal faults such as the Simav, Alaşehir and Büyük Menderes
137 detachments ([Hetzl et al., 1995](#); [Bozkurt et al., 2011](#)) (Figs. 2b and 2c). Currently, plate
138 kinematics in this region are characterized by more localized extension, mainly controlled by the
139 westward motion of Anatolia ([Reilinger et al., 2006](#)) and by N-S extension accommodated by
140 steep normal faults in the Gediz and Büyük Menderes Grabens, both consequences of slab
141 dynamics (e.g. [Jolivet et al., 2013](#); [Gessner et al., 2013](#); [Sternai et al., 2014](#)).



142

143 Figure 2: (a) Global map with red box showing the location on the study area. (b) Simplified
 144 tectonic maps showing the main metamorphic core complexes and associated detachments faults
 145 in the Aegean region (modified from Jolivet *et al.*, (2015)). SiD: Simav Detachment, AD:
 146 Alaşehir Detachment, BD: Büyük Detachment, NAF: North Anatolian Fault. (c) Tomographic
 147 models from Piromallo and Morelli (2003) showing the V_p anomalies at the 100 km depth. White
 148 star shows the location of a slab tear below the Menderes Massif. (d) Simplified cross-sections
 149 highlighting slab retreat and formation of crustal detachments. CC' corresponds to the 2-D
 150 numerical model cross-section.

151

152 The Menderes Massif is also recognized as an active geothermal area (see Fig. S1a in
153 Appendix 1). This region is characterized by long wavelength east-west variations of surface heat
154 flow density with values locally exceeding 90 mW m^{-2} (Erkan, 2015). At depth, it seems that a
155 similar east-west long wavelength thermal anomaly is also present, as suggested by the Curie-
156 point isotherm map of Aydin *et al.* (2005). According to these authors, the depth of the $580 \text{ }^{\circ}\text{C}$
157 isotherm would be lower than 10 km in western Anatolia (see Fig. S1b in Appendix 1). It turns
158 out that most of the geothermal systems in Turkey are located in the Menderes Massif (Fig. 2b)
159 where recent magmatism is almost absent. Here, geothermal fields are characterized by medium-
160 to high-enthalpy, with reservoir temperature values ranging from 120 to $287 \text{ }^{\circ}\text{C}$ (e.g. Roche *et al.*
161 *in revision*). Most are used for district heating systems (e.g. Balçova and Salihli), whereas a few
162 are exploited for electricity production (e.g. Germencik, Salavatlı; Kızıldere; Alaşehir). Some
163 authors have suggested that most of the geothermal activity can be considered as amagmatic in
164 origin (Faulds *et al.*, 2010; Gessner *et al.*, 2017). However, Ozdemir *et al.* (2017) considered a
165 probable magmatic intrusion in the crust to explain the heat source. Since there is presently no
166 evidence for the existence of such plutons, some authors suggest a deeper and larger heat source
167 triggered by large-scale tectonic processes such as subduction (Kaya, 2015; Gessner *et al.*, 2017;
168 Roche *et al.*, submitted). In that sense, Kaya (2015) invoked a possible lithospheric thinning as
169 the origin of the anomalously high heat flow. This example show that large-scale processes
170 related to lithosphere dynamics could be involved in the genesis of geothermal systems. We thus
171 need to employ a multi-scale approach (lithospheric- and crustal-scales), to improve the
172 characterization of heat transfer possesses from the mantle to the geothermal system.

173

174 3. 3-D THERMAL EVOLUTION OF SUBDUCTION

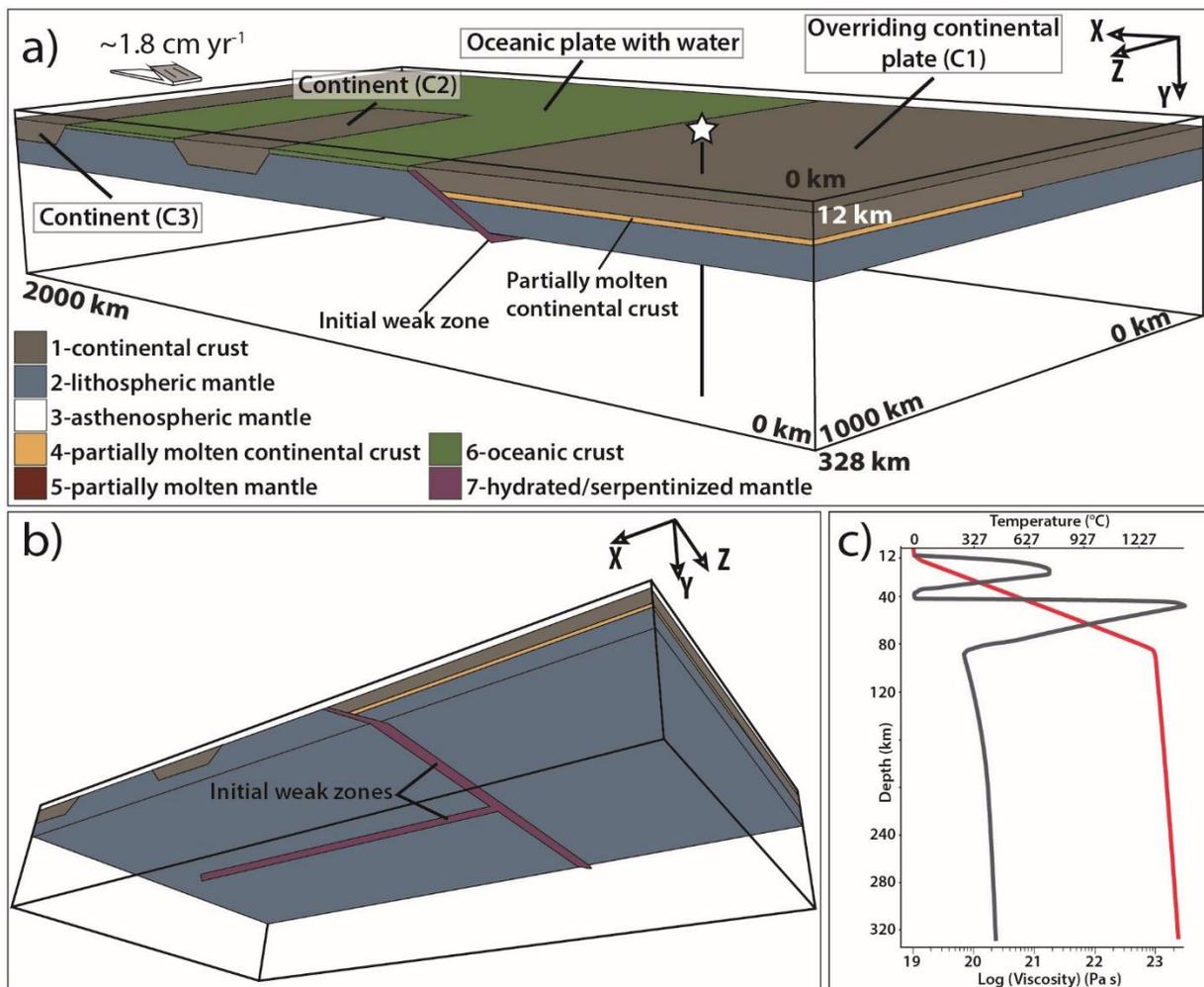
175 In a previous study dedicated to the analysis of surface deformation in the Aegean-
176 Anatolian domain, Sternai *et al.* (2014) found that mantle flow due to slab rollback and tearing
177 can drag the upper plate from below, and explain upper plate extension and the formation of the
178 North Anatolian Fault. Their results emphasize the active role of mantle flow in driving the
179 surface strain and horizontal crustal velocities. However, the associated thermal processes were
180 not described by these authors. Because heat transport is inherently coupled with deformation,
181 mantle shearing may provide a strong heat source. Thus, we must consider it, especially if we
182 want to describe in detail the thermal features associated with 3-D subduction dynamics when
183 slab rollback and tearing occur. Indeed, while radioactive heat production is considered as the
184 main heat source in the continental crust, shear heating has been suggested to be important in the
185 mantle (Shubert and Yuen, 1978) and thus probably also in arc and back-arc settings where fast
186 slab dynamics may be associated with intense asthenospheric flow. Here, we perform similar
187 numerical simulations than Sternai *et al.* (2014), but improve the analysis by investigating the
188 effects of shear heating. We describe the numerical results from a thermal point of view, by
189 focusing on the localization of crustal thermal anomalies in the overriding plate.

190

191 3.1. Model setup

192 The numerical simulations were performed with the code “I3ELVIS” (Gerya, 2010) that
193 solves the momentum, continuity and energy equations based on a staggered finite difference
194 scheme combined with a marker-in-cell technique. Realistic, non-Newtonian visco-plastic
195 rheologies are used in the model, and includes temperature- and pressure-dependent viscosities
196 (see Appendix 2). The model setup is shown in Figure 3. The model domain measures

197 2000 x 328 x 1000 km in the x, y and z dimensions, respectively and is resolved by
 198 501 x 165 x 197 grid points resulting in a resolution of 4, 2 and 5 km in the x, y and z dimensions,
 199 respectively. It includes three continental plates (C1, C2 and C3), allowing reproduction of
 200 similar tectonic events that affected the Tethys convergent domains (subduction, collision, slab
 201 tearing, trench retreat; [Sternai et al. 2014](#)). These three continental plates are defined by a crust
 202 of 35 km, 50 km and 45 km thick (Fig. 3). In addition to the trench-parallel weak fracture zone
 203 allowing for subduction initiation, the model includes a trench perpendicular weak fracture zone
 204 within the oceanic domain, which allows for slab tearing (Fig. 3).



205

206 Figure 3: Numerical setup. (a) 3D initial model setup where the top layer ("sticky air", $y < \sim 12$
 207 km") and the water are cut off for clarity. Colours showing different rock types. (b) Location of

208 the "weak zones" (i.e. hydrated/serpentinized mantle) into the lithosphere to trigger subduction
209 (z-parallel) and to allow slab tearing (x-parallel) in the initial model. (c) Initial vertical profiles
210 of viscosity (grey) and temperature (red), taken at the location shown by the white star.

211

212 The initial thermal structure of C1, C2 and C3 is laterally uniform with 0 °C at the surface
213 and 1300 °C at 90, 140 and 150 km depth, respectively. The initial thermal structure of the oceanic
214 lithosphere corresponds to the cooling age of 120 Ma (e.g. [Turcotte and Schubert 2002](#)). The
215 initial temperature gradient in the asthenospheric mantle is $\sim 0.5 \text{ }^\circ\text{C km}^{-1}$ (adiabatic). Boundary
216 conditions involve free slip on the $z = 0$ and $z = 1000$ km sides of the model domain, while
217 uniform and constant x-parallel velocities equal to $\sim 1.8 \text{ cm yr}^{-1}$ (convergence) are imposed at
218 the $x = 2000$ km boundary, and the $x = 0$ km is fixed. Global mass conservation is ensured by
219 material outflow through the permeable lower boundary ($y = 328$ km). Surface processes are also
220 implemented using a highly simplified gross-scale erosion-sedimentation law. The parametric
221 study on various boundary conditions and parameters is presented in [Sternai *et al.* \(2014\)](#).

222 Heat production by viscous or plastic shearing (H_s), consisting of the dissipation of the
223 mechanical energy during irreversible non-elastic deformation, is controlled by the deviatoric
224 stresses σ'_{ij} and strain rates $\dot{\epsilon}'_{ij}$, as follow ([Gerya, 2010](#)):

$$225 \quad H_s = \sigma'_{ij} \dot{\epsilon}'_{ij} \quad (1)$$

226 The possibility to investigate H_s in our numerical simulations is a significant advance with respect
227 to the previous models presented in [Sternai *et al.* \(2014\)](#). In order to test the possibility that slab
228 tearing would enhance mantle shear heating, two series of numerical models, with and without
229 slab tearing, have been performed, all other settings being equal (i.e. same boundary conditions,
230 material properties, see more information in Appendix 2).

231

232 **3.2. Results**

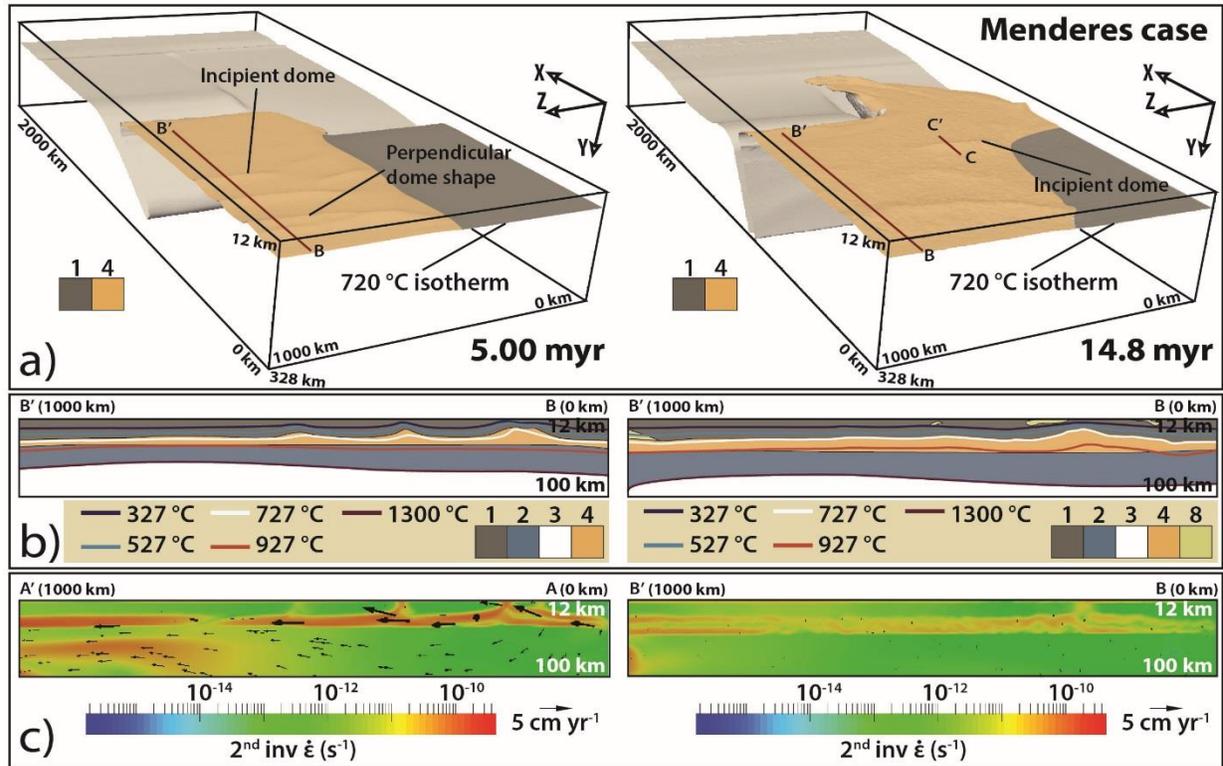
233 Before focusing on thermal effects associated with slab dynamics, we briefly present the
234 model evolution. The overall sequence of events is robust and includes first an early subduction
235 of the oceanic domain between C1 and C2 with localized extensional deformation in the
236 overriding plate. In our models, it includes the magmatic arc and the back-arc domain (see video
237 S1 in Appendix 1). During this first subduction event, a toroidal mantle flow occurs below C1,
238 as induced by the down-going slab. This event also controls crustal flow and surface
239 deformation. Afterward, continental collision between C1 and C2 occurs, locally increasing slab
240 dip and thus favoring slab tearing. Later subduction of the oceanic domain between C1 and C3
241 induces widespread extensional deformation in the overriding plate. This last event is highly
242 conditioned by the previous horizontal to sub-horizontal return flow of hot asthenosphere, which
243 warms up the incipiently subducting lithosphere close to the slab tear, enhancing here a non-
244 cylindrical (i.e. not parallel to the initial upper plate margin) slab rollback. Rheological
245 stratification of the upper plate is characterized at this stage by a thin rigid lithospheric mantle
246 (~ 10 km). Deformation is localized in the overriding plate with strike-slip deformation
247 propagating toward the subducting plate (Sternai *et al.*, 2014).

248

249 **3.2.1. Thermo-mechanical instabilities**

250 The early subduction event is characterized by the development of several positive
251 thermal anomalies (i.e. anomalously high temperatures) (i) at the base of the crust with
252 wavelengths around 200 km, and (ii) in the overriding lower crust, within a partially molten low-
253 viscosity layer, with wavelengths around 30 km (Figs. 4a and 4b). The thermal instabilities in the
254 low-viscosity crust generate dome-like structures cored by partially molten low-viscosity

255 material (orange colour in Figs. 4a and 4b). These domes are elongated parallel to the subduction
256 trench and separated by 200 – 300 km (Fig. 4a). With time, new thermal domes appear toward
257 the trench, controlled by the flow of low viscosity material in response to both slab rollback and
258 tearing (Fig. 4b; Videos S2-S3 in Appendix 1). The strain rate distribution shows that mature
259 crustal domes preferentially localize deformation, rather than incipient crustal domes closer to
260 the trench (Fig. 4c). In addition, deformation localizes around the edges of these domes, showing
261 the importance of lateral crustal strength contrasts on strain localization (Video S4 in Appendix
262 1). At 14.8 Myr, two domes have disappeared but the thickness of the partially molten layer at
263 the base of the crust has tripled (from ~ 4 km to 15 km thick, Fig. 4b; Video S3 in Appendix 1).
264 Meanwhile, the horizontal component of flow velocity in the partially molten layer has decreased
265 (from ~ 5 to ~ 1 cm yr⁻¹, Fig. 4c). On the opposite, above the slab tear, the deformation is localized
266 (see Fig. 4b in Sternai et al. (2014)) and horizontal component of crustal and mantle velocities
267 have increased up to ~ 6 and ~ 9 cm yr⁻¹, respectively. The shear heating and the advective mantle
268 heat flow components near the fast velocity zones are enhanced accordingly, as detailed below.
269



270

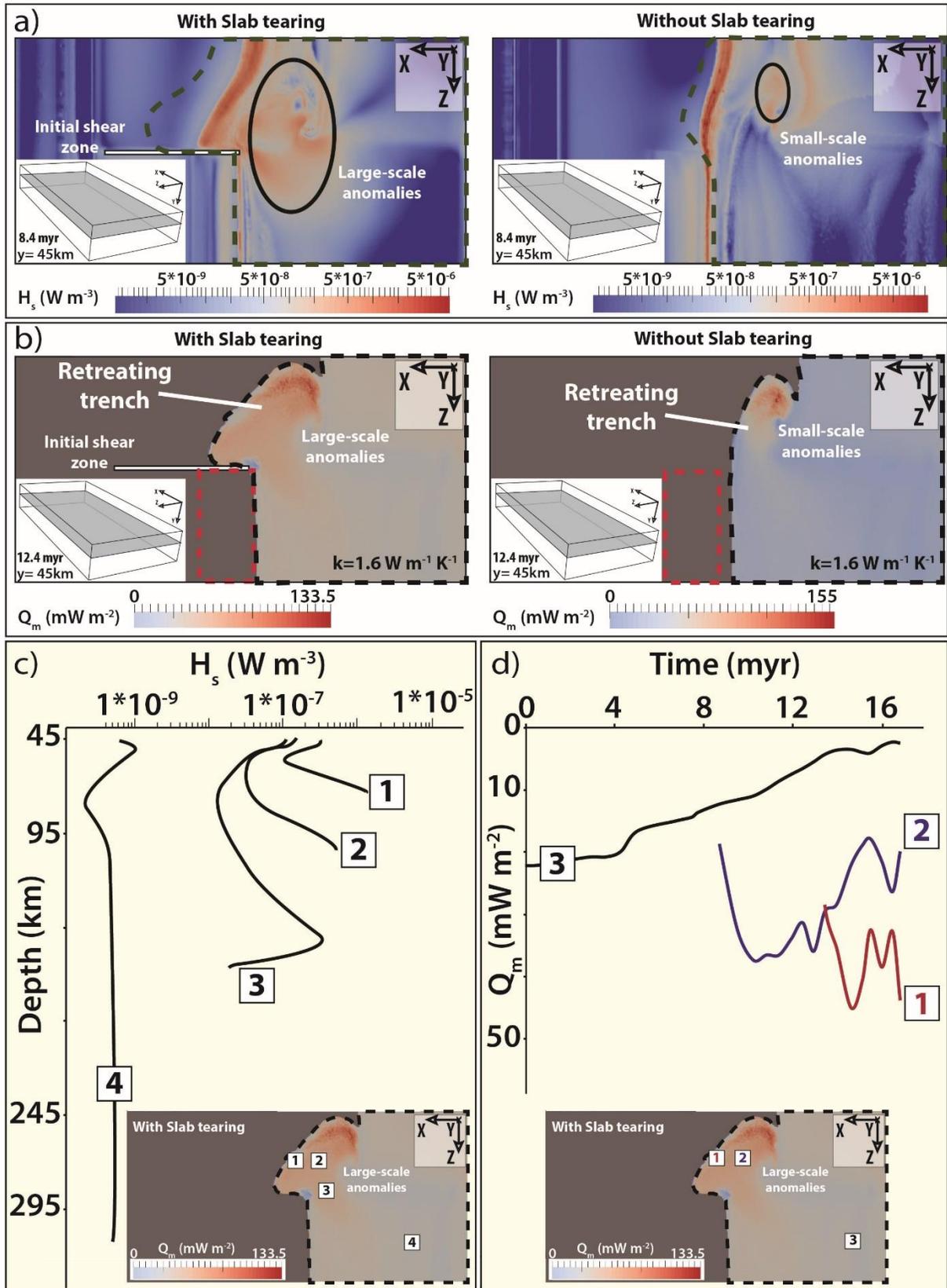
271 Figure 4: Thermo-mechanical evolution of the model. (a) The 720 °C isotherm at 5 myr and at
 272 14.8 myr, respectively. Colours showing different rock types: 1, 4 indicate upper and lower crust,
 273 respectively, see also Figure 3. Note the position of CC' that may correspond to the 2-D numerical
 274 model cross-section. (b) Vertical cross-sections showing different rock types and isotherms.
 275 Colours showing different rock types: 1-2-3-4 as defined in Figure 3; 8-hydrated crust. (c)
 276 Vertical cross-sections showing the second invariant of the strain rate tensor and velocity vectors.
 277

278 3.2.2. Slab dynamics and shear heating

279 The contribution of shear heating to heat production in our reference model (with slab
 280 tearing between C2 and the oceanic domain) is compared to similar simulations without slab
 281 tearing (Fig. 5; see Fig. S3 in Appendix 1). The frictional heat produced in the overriding plate
 282 locally reaches values of $\sim 5.0 \mu\text{W m}^{-3}$ regardless of slab tearing (Fig. 5). However, when slab
 283 tearing occurs, a larger-scale thermal perturbation appears and develops below and at the base of
 284 the crust ($\sim 400 \times 200$ km at 12.4 Myr; see Fig. 5b). Anomalies associated with shear heating are

285 thus distributed over a larger domain in the case of tearing because of a more widespread mantle
286 flow around the tear, and appear more localized and parallel to the subduction trench without
287 tearing (Fig. 5a, see right panel). In addition, our results show that the amount of heat produced
288 by shearing leads to an increase of mantle heat flow (Q_m) at the base of the crust, and particularly
289 above the tear. In particular, for average shear heating of $\sim 2.0 \mu\text{W m}^{-3}$ and mantle thickness of
290 30 km (Fig. 5c, profile 1), the contribution to the heat flow nearby the subduction front is of ~ 60
291 mW m^{-2} at 12.4 Myr. Conversely, far from the subduction front, shear heating is insignificant ($<$
292 1.0 nW m^{-3} , see Fig. 5c, profile 4), implying no contribution of this mechanism to heat production
293 by this mechanism (Fig. 5d, profile 1).

294 Shear heating in the mantle during slab retreat and tearing plays a key role in increasing
295 temperatures at the base of the overriding crust, where mantle heat flow values are around 100
296 mW m^{-2} and reach locally up to 130 mW m^{-2} above the slab tear at 12.4 Myr (Fig. 5b, see left
297 panels). We found that 46 % of the heat flow at the base of the crust is due to mantle shear heating,
298 the remaining part being due to the advection of hot material. In the absence of slab tearing, the
299 average of mantle heat flow at similar times is around 80 mW m^{-2} , reaching locally up to 155
300 mW m^{-2} but over a much smaller area, (Fig. 5b, right). We therefore suggest that advection of hot
301 asthenosphere to shallow depths during slab retreat and tearing plays an important role in
302 increasing temperatures at the base of the crust (Fig. 6). Furthermore, mantle heat flow values
303 may reach steady-state after ~ 4 Myr (values around $40 - 50 \text{ mW m}^{-2}$, Fig. 5d).



304

305

Figure 5: Mantle heat source with and without tearing at the mantle-crust transition. (a) Plan view

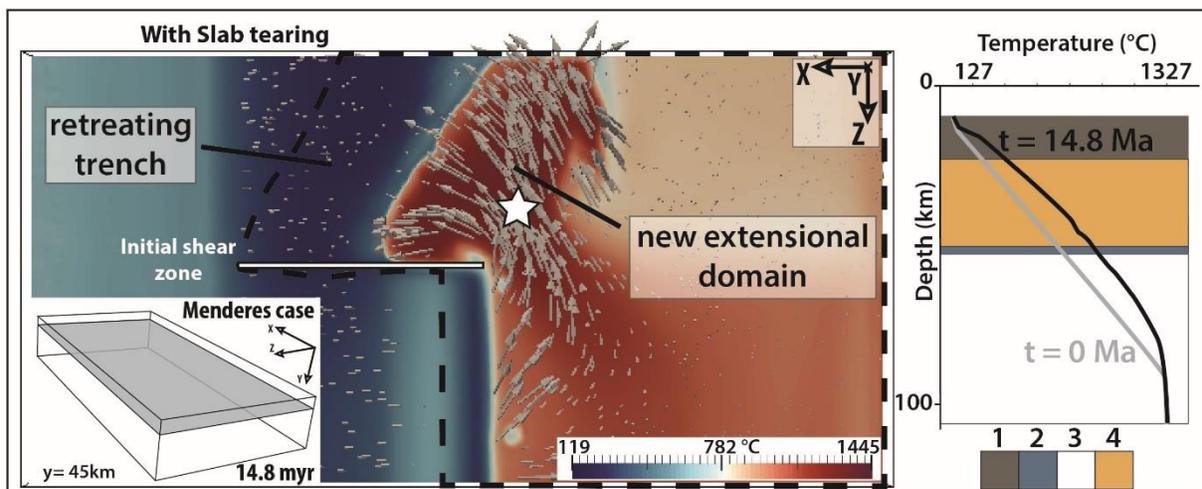
306 of shear heating (H_s) at the mantle-crust transition at 8.4 Myr. Dashed line on both panels shows
 307 the plan view of the overriding plate C1. (b) Plan view of mantle heat flow (Q_m) at the mantle-
 308 crust transition at 12.4 Myr. Black and red dashed lines on both panels show the plan view of the
 309 overriding mantle, and of the overriding C2, respectively. (c) Vertical depth-shear heating (H_s)
 310 profiles in the mantle below the stretched overriding plate at $t_0 + 14.8$ myr, taken at the location
 311 shown by the white squares. (d) Evolution of mantle heat flow (Q_m) at the mantle-crust transition.
 312 The values are taken at the location shown by the white squares.

313

314 To summarize, slab tearing facilitates the rollback of the subducting plate and extension
 315 within the upper plate. Consequently, slab tearing and its associated mantle flow contributes to
 316 the increase in heat flux at the base of the crust by (i) enhancing mantle shear heating and (ii)
 317 controlling the return flow of hot asthenosphere (e.g. toroidal and poloidal mantle flow
 318 components). Both mechanisms together appear to be able to increase the Moho temperature by
 319 up to 250 °C (Fig. 6). Therefore, slab tearing may (i) affect the heat source responsible for
 320 regional HT-LP metamorphic rocks in MCCs and (ii) explain regional high-temperature
 321 anomalies across amagmatic geothermal systems in post-subduction settings.

322

323



324 Figure 6: Plan view of temperature anomaly at the mantle-crust transition (~ 45 km depth) in the
325 case of slab tearing. Grey arrows represent the velocity vectors in the asthenosphere. Black
326 dashed line on the left panel shows the plan view of the overriding plate C1. Right panel: vertical
327 temperature profiles at t_0 and $t_0 + 14.8$ myr, taken at the location shown by the white star. Colours
328 showing different rock types: 1-2-3-4, as defined in Figure 3.

329

330 **4. FROM SUBCRUSTAL THERMAL ANOMALIES TO GEOTHERMAL SYSTEMS:**

331 **2-D MODEL OF FLUID CIRCULATION**

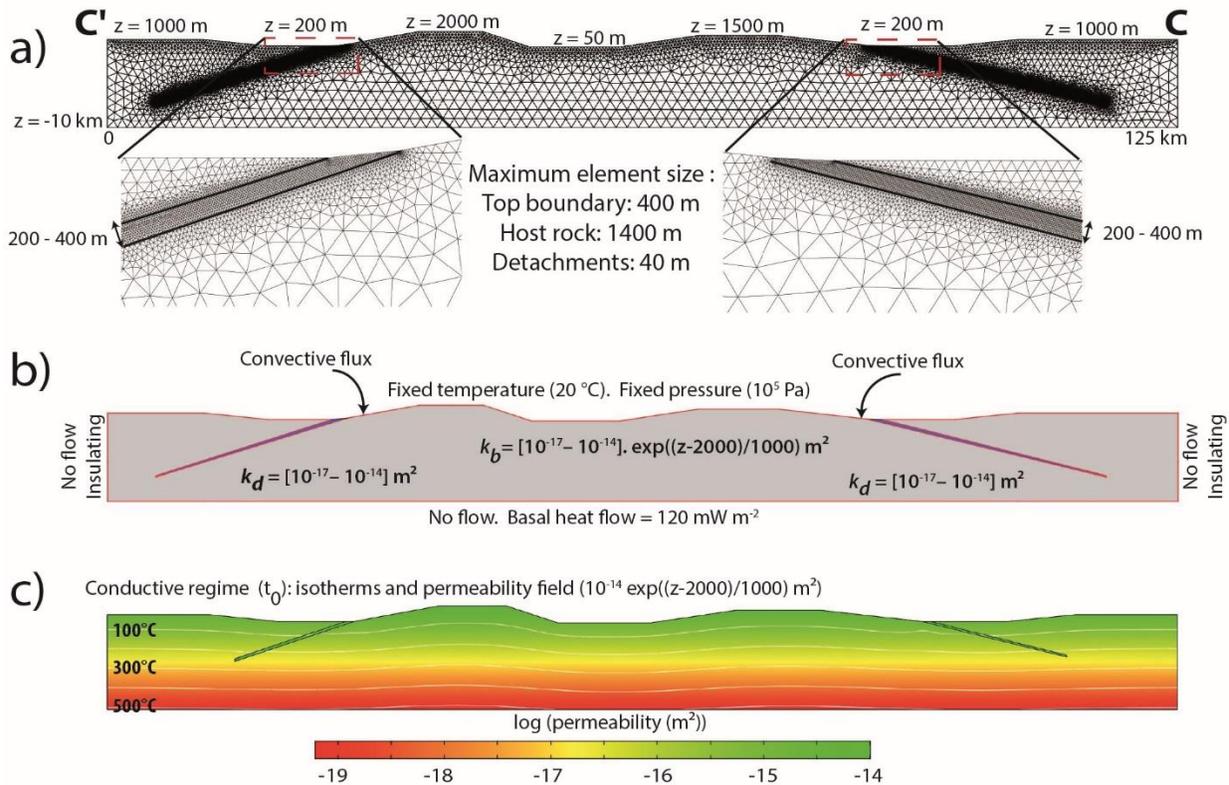
332 Results of the 3-D model show that at 12.4 Myr, mantle heat flow can reach anomalously
333 high values exceeding 130 mW m^{-2} (Fig. 5b). At 10 km depth above the tearing, temperature
334 values vary between 450 and 500 °C, and are thus similar to estimated temperatures from
335 aeromagnetic data at the same depth in the Menderes Massif (~ 580 °C) (Aydin *et al.*, 2005).
336 Using these results, we defined a 2-D fluid flow model in a porous media to analyse the thermal
337 consequences of anomalously high basal heat flow on fluid circulation in the upper crust. The
338 advantage of this model is a good spatial resolution (see Appendix 2), representative
339 permeability, and realistic rock and fluid properties.

340 The model measures 120 x 10 km and is designed to represent a SSW-NNE cross section
341 of the Menderes massif. We choose the basal heat flow in order to (1) account for previous 3-D
342 numerical results where mantle heat flow could reach values up to 130 mW m^{-2} (Fig. 5b), and (2)
343 reproduce temperatures above 500 °C at 10 km depth, as suggested by Curie-point depths in
344 western Anatolia (Aydin *et al.*, 2005). The objective of this model is to understand the effects of
345 high basal heat flux on the fluid flow circulation pattern, in particular around permeable
346 detachments. In the following, we describe the main characteristics of the model setup and
347 present the main results.

348

349 4.1. Model setup

350 Coupling between Darcy's law, heat transfer equation and mass conservation is
351 performed using the Comsol Multiphysics™ software (finite element method). Details about the
352 numerical procedure (benchmark experiments, model setup, fluid and rock properties) are given
353 in Appendix 2. The model is shown in Figure 7, where two detachment faults (with a dip angle
354 varying from 10 to 20°) represent the Büyük Menderes and the Alaşehir Menderes detachments
355 (Fig. 2b). Figure 7a illustrates the mesh refinement in permeable zones (close to the surface and
356 within the detachments). A convective flux condition was imposed at the top of the detachments,
357 in order to allow the emergence of hot fluids at the surface. In addition, we use temperature-
358 dependent density and viscosity laws for fluids (see Appendix 2). No magmatic source is present
359 in the model, and the chosen basal heat flow value at a depth of 10 km corresponds to 120 mW
360 m⁻² (Fig. 7b). This value is therefore consistent with both criteria (i.e. mantle heat flow from our
361 3-D model and Curie-point depths estimations). A constant pressure of 10⁵ Pa is imposed at the
362 surface allowing fluid influx while no flow is allowed across the others boundaries of the box
363 (Fig. 7b). At time $t = 0$, thermal regime is purely conductive (see white isotherms in Fig. 7c).
364 Then, permeability values are switched on (color code of Figure 7c; Appendix 2) and the transient
365 evolution of temperature and velocity fields are recorded during 5 Myr. Permeability in the
366 basement (k_b) decreases exponentially with depth (Manning and Ingebritsen, 1999) (see Fig. 7c
367 and Appendix 2) while detachment permeability is assumed constant (k_d). Despite a number of
368 arguments in such extensional setting suggest high permeability at the surface (highly fractured
369 and karstified rock, see Özen *et al.* (2012)) numerical values remain unconstrained. We therefore
370 decided to test a wide range of realistic permeabilities for the basement (i.e. from 10⁻¹⁷ to 10⁻¹⁴
371 m²) and the detachment fault with same range of values (Fig. 7b).



373

374 Figure 7: (a) Geometry and mesh used for 2-D models of fluid circulation in the upper crust of
 375 the Menderes area, where MCCs and detachments are reproduced. This cross section would
 376 correspond, from left to right, to a N-S cross section, from Salihli to Salavatli geothermal areas.
 377 Mesh is refined at the top surface and within detachments, where permeability is the highest. (b)
 378 Boundary conditions and range of values for permeability (note the depth-dependence of the host
 379 rock permeability). (c) Steady-state conductive regime, where isotherms in white are separated
 380 by 100°C . Colors refer to permeability values affected at time $t > 0$ (see text) for the case detailed
 381 in Figure 9.

382

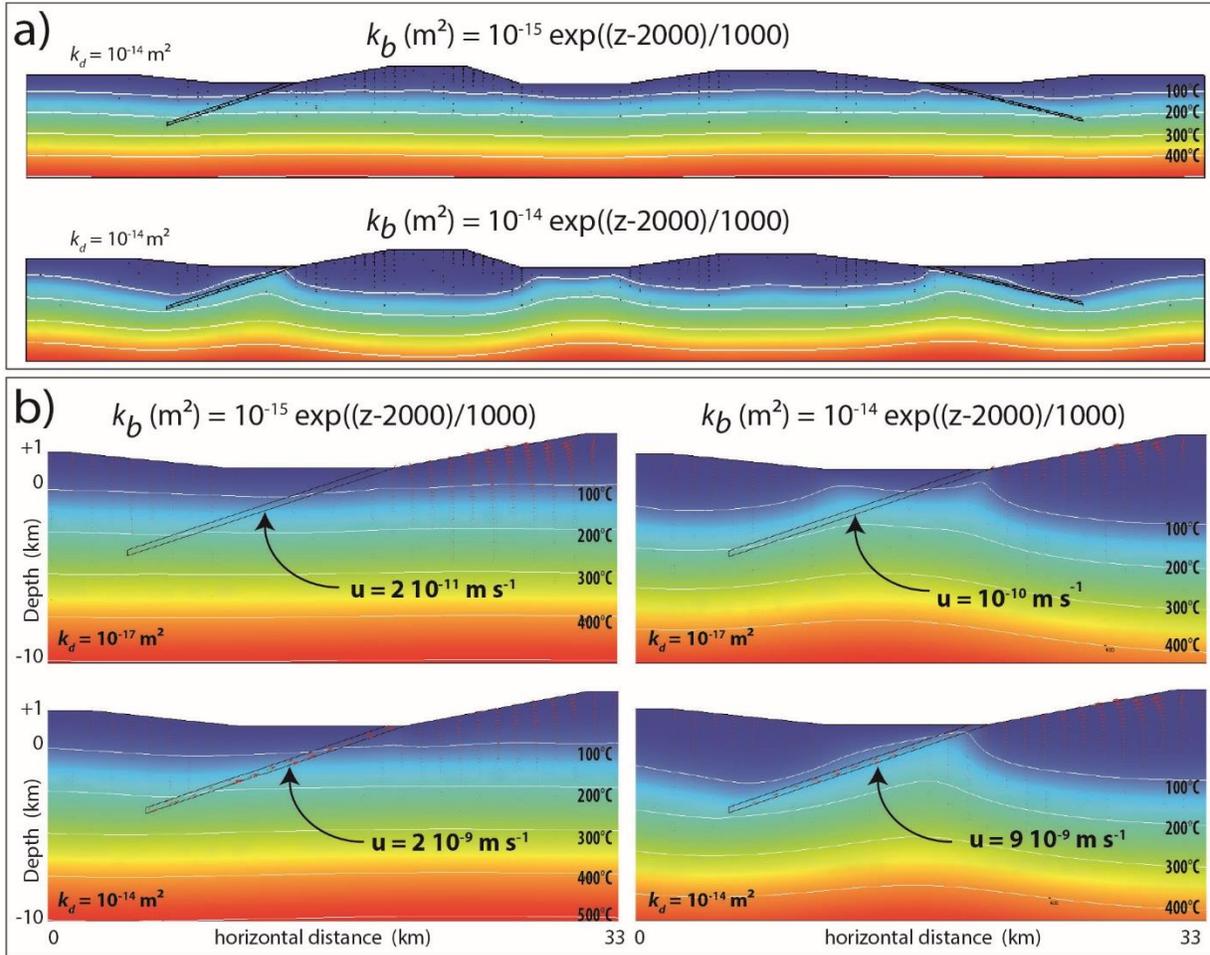
383 4.2. Results

384 4.2.1. Role of basement and detachment permeabilities

385 Figure 8 illustrates steady-state temperature fields for different combinations of both
 386 basement and detachment permeability. In Figure 8a, results show the major role of k_b to distort

387 significantly the isotherms, even if k_d is high (10^{-14} m²). In the upper case of Figure 8a, isotherms
388 remain flat and no thermal reservoir forms beneath the detachments. Alternatively, the lower case
389 with realistic higher permeability values shows a general fluid circulation pattern characterized
390 by: (i) surface meteoric fluids that flow from topographic highs to the top of the detachment fault
391 and other topographic lows, inducing a negative thermal anomaly in the hanging walls of
392 detachments; (ii) at depth, hot fluids are drained upward by the permeable detachments allowing
393 for isotherms to rise along the fault zones (see uplifted isotherms in bottom case of Fig. 8a). A
394 positive thermal anomaly is due to hot fluids rising up toward the top of the detachment while a
395 negative thermal anomaly induced by topography-driven flow grows in the hanging walls of
396 detachments, leading to the formation of hot fluid reservoirs beneath the detachments (near the
397 surface). To synthesize, at the scale of the whole massif, thermal undulations mimicking dome-
398 like structures are localized beneath topographic lows in the case of a high basement permeability
399 (Fig. 8a, bottom case).

400



401

402 Figure 8: Results for thermal perturbations due to fluid circulation in permeable zones for varying
 403 basement permeability (k_b) and for varying detachment permeability (k_d). (a) role of the basement
 404 permeability for a fixed value of k_d ($10^{-14} m^2$). (b) Zoom on the left detachment, with varying
 405 permeabilities (small k_d value in the top row; high k_d value in the bottom row). Left (right) column
 406 corresponds to a small (high) basement permeability.

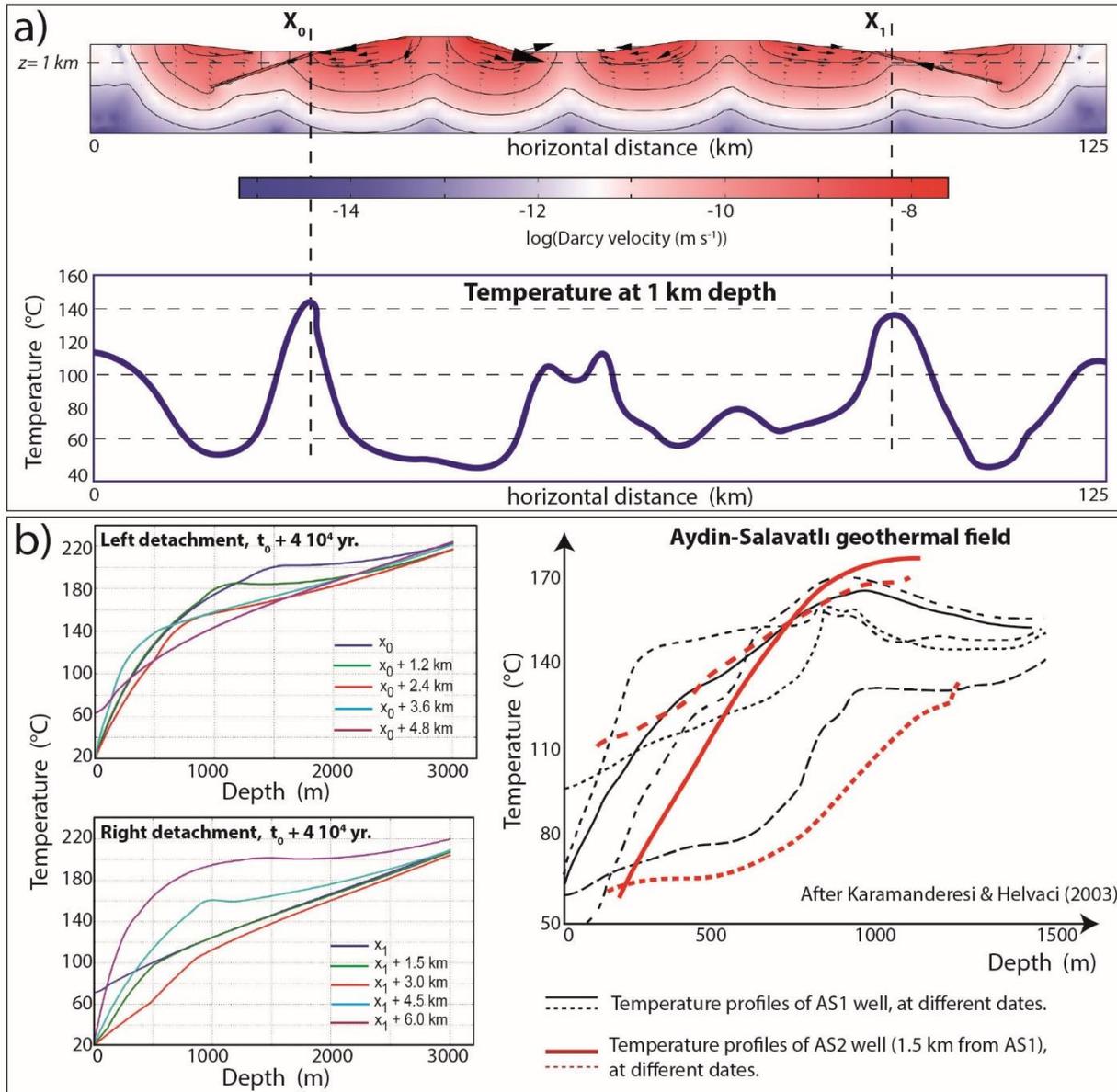
407

408 When k_d is varied (Fig. 8b), it turns out that thermal reservoirs establish at shallow depth
 409 when both k_d and maximum k_b values are important ($10^{-14} m^2$, bottom right case in Fig. 8b). In
 410 the case where k_d is low ($10^{-17} m^2$, top case in Fig. 8b) two thermal reservoirs appear only with a
 411 high maximum k_b value, suggesting that topography-driven flow is dominant. However,

412 temperatures at shallow depth and fluid velocity values within the detachment are much lower
413 than in the high k_d case (bottom right case in Fig. 8b). Crustal-scale permeable faults and their
414 induced high topography thus control the spatial distribution of upper-crust thermal anomalies at
415 depth, mimicking dome-like structures beneath topographic lows (Fig. 8a, bottom case, and Fig.
416 8b, right column).

417 ***4.2.2. Fluid flow field and temperature profiles***

418 Figure 9a illustrates the fluid flow velocity pattern in the case of Figure 8a (bottom case).
419 Fluid flow is favored by high-topography and permeable zones such as detachments. Located
420 within detachments, high velocity areas ($> 10^{-10}$ m / s – or 0.32 cm / yr) are recorded down to 5
421 km deep (red areas). The bottom panel of Figure 9a shows a horizontal temperature profile at 1
422 km depth. It is important to note that temperature anomalies of ~ 140 °C are focused close to the
423 top of the detachments. Figure 9b (left column) shows computed vertical temperature profiles at
424 different distances from the two detachments, at a given time (i.e. 4×10^4 yr). Figure 9b (right
425 column) illustrates temperature measurements at the Aydin-Salavatlı geothermal field, Büyük
426 Menderes Graben, within 2 boreholes ~ 1.5 km apart. Although the comparison of daily-measured
427 thermal profiles with these two sets of vertical temperature profiles is limited, one can observe
428 some similarities with (i) a high-temperature gradient at the surface, decreasing rapidly with
429 depth, and (ii) some negative values of the temperature gradient. In addition, modeled
430 temperatures are comparable, with emergence temperatures around 60 °C and reservoir
431 temperatures at depth around 180 – 200 °C. Because our density law is simplified (see Appendix
432 2), and because the vapour phase is not accounted for, the computed profiles show rather smooth
433 curves when compared to measurements. However, the typical convective signatures in
434 temperature profiles are similar.



436

437 Figure 9: Results for the experiment shown in the lower case of Figure 8a (and right case of
 438 Figure 8b). (a) Fluid flow velocity field (colors, isocontours and arrows) and horizontal
 439 temperature profile at 1 km depth (dashed line in the fluid flow pattern). Note that the largest
 440 temperature perturbations correspond to the top of the detachment locations (x_0 and x_1). (b)
 441 Computed vertical temperature profiles at time $t_0 + 4 \cdot 10^4 \text{ yr.}$ x_0 and x_1 are indicated in Figure 9a.
 442 On the right, measured temperature profiles at the Aydin-Salavatlı geothermal field; black curves

443 (plain and dashed) correspond to different measures in the same AS1 borehole (different dates of
444 measurements) and red lines are measured in the AS2 borehole, 1.5 km from the AS1 borehole
445 (after [Karamanderesi and Helvaci, 2003](#)).

446

447 **5. DISCUSSION**

448 **5.1. Large-scale thermo-mechanical boudinage, and emplacement of MCC in Western**

449 **Anatolia**

450 The modeled thermo-mechanical instabilities (Fig. 4) are not only analogous to crustal
451 and lithospheric boudinage ([Ricard and Froidevaux, 1986](#)), but they add a thermal component to
452 what was generally assumed to be purely mechanical. These instabilities developed in the
453 overriding plate are triggered by both shear heating and hot mantle flows. These mechanisms
454 indeed induce a thermal weakening of the lithosphere, thus controlling the development of
455 boudinage at different scales, and the localization of MCCs in the extensional crust. It is important
456 to note that these domes are also promoted by the low-viscosity crust, which is related to the
457 chosen initial thermal regime in our models. When the slab tear is accounted for, the crust is weak
458 over a wider domain above the tear, increasing its ability to flow (e.g. [Block and Royden, 1990](#);
459 [Rey et al., 2009](#)) and to form HT domes. In other words, shear heating and mantle flow control
460 the position of positive thermal anomalies in the overriding crust, which, in turn, control the
461 deformation pattern in the lower crust during extension. Such a lithospheric thermal weakening
462 is further enhanced by the accumulation of magmas in the back-arc domain, which is widely
463 promoted by slab rollback and tearing processes ([Menant et al., 2016b](#)).

464 In the Cyclades (Greece), the dominant wavelength of MCCs (several tens of kilometres,
465 similar to the spacing between the different islands) is compatible with that expected for crustal-
466 scale boudinage (Fig. 4). At the scale of the whole Aegean domain, long-wavelength variations

467 of Moho depth (Tirel *et al.*, 2004; Karabulut *et al.*, 2013) are more reminiscent of lithospheric-
468 scale boudinage (several hundreds of km; Fig. 4). This boudinage of the crust induces migmatite-
469 cored MCCs which were first exhumed in the northern Menderes Massif in the Miocene (Cenki-
470 Tok *et al.*, 2016), and then in the central and eastern Cyclades since the lower Miocene (Jolivet
471 *et al.*, 2015), coevally with a further acceleration of trench retreat related to the Hellenic slab
472 tearing. This is consistent with our model, where dome-like structures, cored by partially molten
473 low-viscosity material, develop first in the overriding plate, and then migrate toward the trench,
474 in response to both slab rollback and tearing (Figs. 4a and 4b). In addition, we suggest also that
475 excess heat due to these subduction dynamics controlled the retrogression in the greenschist-
476 facies of high-pressure metamorphic rocks in Cyclades (post-orogenic extension), consistently
477 with acceleration of slab retreat around 35 Ma (Jolivet *et al.*, 2015). Furthermore, our numerical
478 experiments highlight an increase of Moho temperature during a few Myrs related to
479 asthenosphere upwelling below the overriding plate (Figs. 5d and 6). We therefore suggest a
480 similar heating for the mantle-crust boundary in the Anatolian-Aegean domain, which is
481 consistent with the presence of HT metamorphic domes and the formation of syn-extension
482 granites and migmatites throughout the entire western Anatolian-Aegean domain between 20 Ma
483 and 8 Ma (Jolivet *et al.*, 2015; Menant *et al.*, 2016a).

484 In the Menderes MCC (located above the slab tear, see Fig. 2), strain was first localized
485 in the north, along the Simav Detachment (early Miocene, stage 1 after Gessner *et al.* 2013)
486 accommodating the exhumation of an asymmetric migmatite-cored MCC (Rey *et al.*, 2009;
487 Cenki-Tok *et al.*, 2016). Then, strain migrated southward, toward the trench and localized within
488 the central part of the Menderes Massif (Fig. 2c). The Alaşehir and Büyük Menderes detachments
489 assisted exhumation of the lower crust during slab tearing (middle-late Miocene, stage 2 after

490 [Gessner et al. 2013](#)). This implies that the central part of the Menderes Massif had become weaker
491 than the previously activated Simav area, probably due to the extra source of heat associated with
492 slab tearing. According to [Labrousse et al. \(2016\)](#), the evolution of the style of extension from an
493 asymmetric MCC (localized deformation, case of the Simav detachment) to a double-dome
494 symmetric MCC (distributed deformation, case of the Menderes detachments) implies that heat
495 input at the base of the crust is higher than heat dissipation due to stretching. These observations
496 are consistent with the thermal features observed in our model, which emphasize the crucial role
497 of shear heating, as well as poloidal and toroidal components of mantle flow, to heat the lower
498 crust (Figs. 5 and 6). Both mechanisms are thus responsible for the presence of large-scale
499 thermal anomalies at depth (i.e. 580 °C at 10 km depth) and at the surface (i.e. heat flow values
500 around 100 mW m⁻², [Erkan, 2015](#)). As a consequence, the thermal state of the crust induced by
501 elevated mantle heat flow could control the style of extension (asymmetric versus symmetric
502 MCC).

503 Furthermore, hot fluids circulating within crustal-scale permeable faults help increasing
504 basal temperatures (around the ductile/brittle transitions) by several tens of °C (Fig. 8a), resulting
505 in a positive feedback for the formation of new shear zones and MCCs at depth. In this case,
506 magmatic bodies such as granitoids and migmatites may be developed and localized below the
507 detachment, as it seems to be the case in the Menderes with the Salihli diorite and Simav granites.

508 The progressive formation of HT domes topped with extensional detachments can thus
509 be summarized as follows: (1) mechanical (boudinage) and thermal instabilities within the
510 stretched lithosphere generate a series of domes, periodically spaced, producing rheological
511 heterogeneities (i.e. low-viscosity layers) that are likely to localize extensional shearing
512 deformation on their edges ([Le Pourhiet et al., 2003](#); [Huet et al., 2011](#)) and (2) upward flow of

513 deep hot fluids and downward flow of cold surface fluids further increase thermal contrasts across
514 the shear zones, enhancing the localization of deformation.

515

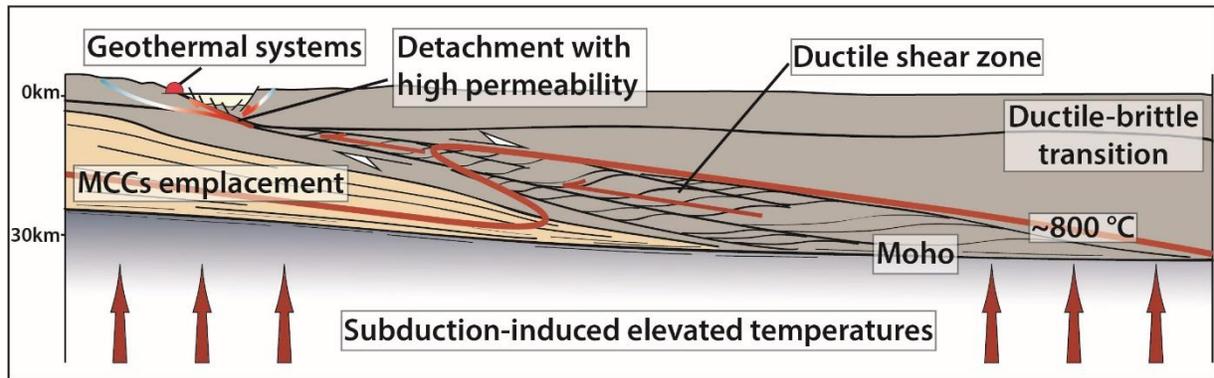
516 **5.2. Detachments and localisation of HEGRs in Western Anatolia**

517 The relationships between the localization of the deformation and thermal domes (Fig. 4)
518 are similar to that documented in MCCs, and may be associated with crustal-scale permeable
519 faults, which can provide pathways for meteoric waters to flow into the actively deforming
520 middle crust (e.g. [Famin et al., 2005](#); [Mulch et al., 2006](#)). As shown in Figure 5, the excess of
521 mantle heat flow is transferred by conduction into the crust, affecting the fluids present in the
522 crustal porosity. These fluids may then undergo deformation-driven flow in ductile shear zones
523 (e.g. [Oliver, 1996](#)), but also thermally driven (i.e. buoyancy-driven) flow through the crust, where
524 high permeability detachments may easily collect and bring up deep hot fluids. Indeed, our 2-D
525 numerical models suggest that an elevated basal temperature (580 °C) at 10 km depth would
526 induce temperatures of 300 – 350 °C at a depth of 6 km and thus a sufficiently high fluid density
527 contrast to permit upward flow along the low-angle fault in the shallower crust. In that case,
528 buoyancy-driven flow is superimposed to topography-driven flow (Fig. 8). In other words, the
529 topography-driven flow, which prevails in the regional fluid flow pattern, is dominated by
530 buoyancy-driven flow within highly permeable detachments.

531 Although the spatiotemporal evolution of permeability within detachments is poorly
532 known, field observations on these geological structures ([Famin et al., 2005](#)) suggest that they
533 could represent highly permeable zones of several hundreds of meters thick where fluid flow is
534 facilitated (Fig. 10). Moreover, if these channelized fluid flow systems are tectonically active,
535 earthquake faulting may be associated with seismic pumping that enhances fluid migration

536 toward dilation zones (Sibson *et al.*, 1975; Famin *et al.*, 2005). As a consequence, detachments
537 may be considered as the crustal-scale structural control for the circulation of deep geothermal
538 fluids related to HEGRs (e.g. Bellani *et al.*, 2004; Roche *et al.* submitted), driving the long-term
539 crustal-scale history of the geothermal systems (Fig. 10). However, it is noteworthy that most of
540 the geothermal systems in the Menderes Massif reside in fault intersection zones, where N-S
541 transfer faults and low-angle normal faults mostly formed dilational jogs that seem to promote
542 the emergence of hot waters and gas vents (Roche *et al.*, submitted). Fault interactions would
543 thus allow for fluid rising in the hanging wall from the reservoirs located below the detachment
544 (Fig. 8b, right column), and control the location of thermal springs. Further numerical simulations
545 would therefore be required to confirm this hypothesis based on field observations.

546



547

548 Figure 10: Schematic cross section showing fluid flow associated with detachment zone in back-
549 arc area. Red arrows indicate fluid circulation in ductile crust and circulation of meteoric fluids
550 in the upper crust are underlined by colored pathways (blue for cold and red for warm).

551

552 **5.3. Subduction dynamics and localisation of HERSs: the Larderello (Tuscany) and the**
553 **Basin and Range (Western United States) areas**

554 Our 3-D and 2-D model results are relevant to other regions, and thus may explain
555 relationships between subduction dynamics and the location of geothermal provinces; for
556 instance, (i) in the Larderello area post-orogenic extension is active in back-arc domain; or (ii) in
557 the Basin and Range Province (western United States). In both cases, MCCs are exhumed along
558 low-angle normal faults, and therefore represent a favorable setting for HEGRs. While some of
559 them are considered as “amagmatic” in origin (e.g. [Bellani *et al.*, 2004](#); [Faulds *et al.*, 2010](#)), some
560 authors require a magmatic heat source (e.g. [Koenig and McNitt, 1983](#); [Santilano *et al.* 2015](#)).
561 Through these two main additional examples, we show that the “magmatic” vs. “amagmatic”
562 classification may be ambiguous according to the long-term behavior of these geothermal
563 systems.

564 In the Northern Tyrrhenian Sea and Tuscany regions, subduction dynamics has also been
565 controlled by slab retreat since the Oligo-Miocene. This region is characterized by a shallow
566 Moho discontinuity (20 – 25 km depth), and a reduced lithosphere thickness due to uprising
567 asthenosphere and the delamination of crustal lithosphere ([Gianelli, 2008](#)). At crustal-scale a
568 series of low-angle normal faults crosscut the old thrusts (e.g. [Faccenna *et al.*, 1997](#); [Jolivet *et*](#)
569 [al., 1998](#)), and thus cannot be explained by a reactivation of pre-existing inherited discontinuities.
570 Instead, our numerical models suggest that thermo-mechanical boudinage induced by slab
571 rollback triggers the formation of localized shear zones and allows the exhumation of MCCs (e.g.
572 Elba Islands, Fig. S4 in Appendix 1). These shear zones are still active ([Bellani *et al.*, 2004](#)) and
573 may control the ascent of magmatic intrusions, such as beneath the Larderello geothermal field.

574 We therefore suggest that magmatic intrusions in the upper crust result directly from the deep
575 thermal anomaly associated with slab dynamics during extension.

576 Although subduction dynamics are quite different in the Basin and Range province (i.e.
577 first characterized by slab retreat since the Eocene at the northern edge of the Farallon slab
578 (Schellart *et al.*, 2010), and then by an asthenospheric upwelling above the Californian slab
579 window at 30 Ma (Atwater, 1970)), similar remarks also hold for this region. Indeed, our results
580 are consistent with geological observations where the presence of a conductor at 15 – 35 km
581 depth range (Wannamaker *et al.*, 2008) suggests that the central part of the Basin and Range
582 domain has become weaker, probably due to an extra source of heat associated with slab window
583 (i.e. magma underplating due to mantle flow and shear heating). In light of our numerical results,
584 we propose that fast-flowing asthenosphere also strongly heats the lithospheric mantle and the
585 lower crust via shear heating and mantle upwelling processes. Lithospheric weakening thus
586 promotes the propagation of major fault zones throughout the crust, leading to the connection
587 between the weaker conductive lower crust and the brittle upper crust (Fig. 10). Hot thermal
588 fluids at depth can then ascend within the damage fault zones, which thus control the location of
589 HEGRs at the surface (Faulds *et al.*, 2010). It is noteworthy that the presence of recent intrusions
590 (e.g. Steam Boat, Casa Diablo in the Walker Lane geothermal belt, Koenig and McNitt, 1983)
591 locally enhances thermally-driven flow, by increasing the temperatures of the geothermal systems
592 but this seems marginal as little active volcanism is recognized in this region (Faulds *et al.*,
593 2010).

594

595 **6. CONCLUSIONS**

596 3-D numerical experiments show that the mantle flow and shear heating resulting from
597 slab retreat and tearing have a significant effect on Moho temperatures during 5 – 10 Myrs. Such
598 a mechanism induces lithospheric and crustal thermal boudinage with different wavelengths,
599 which in turn creates thermally-driven rheological contrasts that localize deformation and lead to
600 the formation of crustal-scale permeable detachments. These structures allow for the exhumation
601 of MCCs and for the emplacement of HEGRs comparable to those observed in the Mediterranean
602 realm and in the Western United States. Although there is no consensus on a worldwide
603 classification of geothermal systems, our results highlight the importance of the spatial and
604 temporal scales of subduction dynamics in the control of genesis of geothermal systems.

605

606 **ACKNOWLEDGMENTS**

607 This work has been financially supported by the LABEX VOLTAIRE (ANR-10-LABX-
608 100-01) of the University of Orléans and by the French Geological Survey (BRGM). This paper
609 is a contribution of the ERC RHEOLITH Project (ERC advanced grant no. 290864). Pietro
610 Sternai is grateful to the Swiss NSF Ambizione grant PZ00P2_168113/1. Numerical simulations
611 were performed on the ETH-Zürich clusters BRUTUS and EULER. The paper benefited from
612 relevant revisions by David Whipp and one anonymous reviewer.

613

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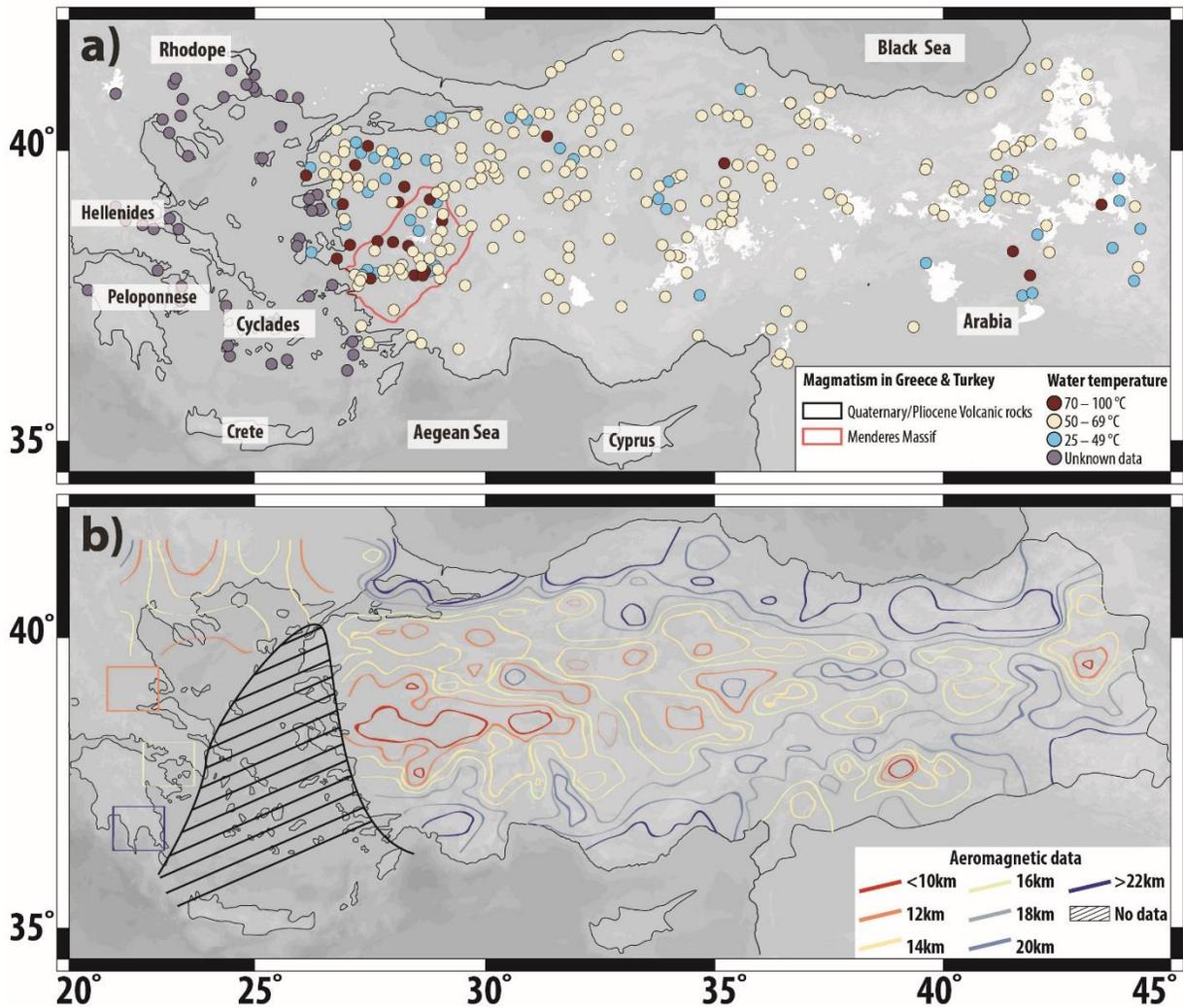
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758

760 1. Supplementary Figures and Videos

761



762

763 Figure S1: Maps of the Aegean-Anatolian region. (a) Geothermal map of eastern Mediterranean

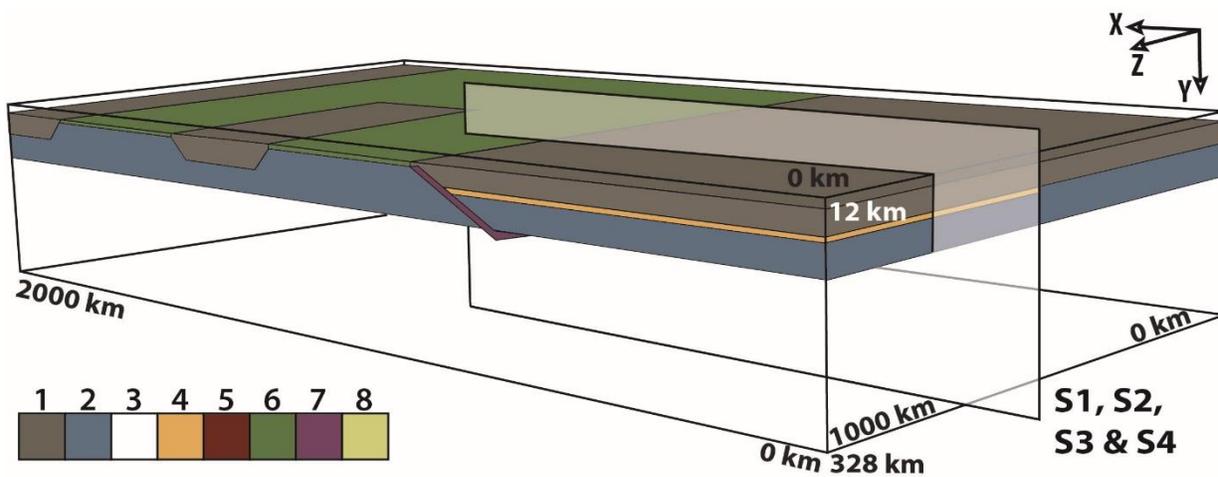
764 region highlighting major thermal occurrences based on a compilation of several data sources

765 (from the MTA (2005); Mendrinou *et al.* (2010) and Andritsos *et al.* (2015)), and spatial

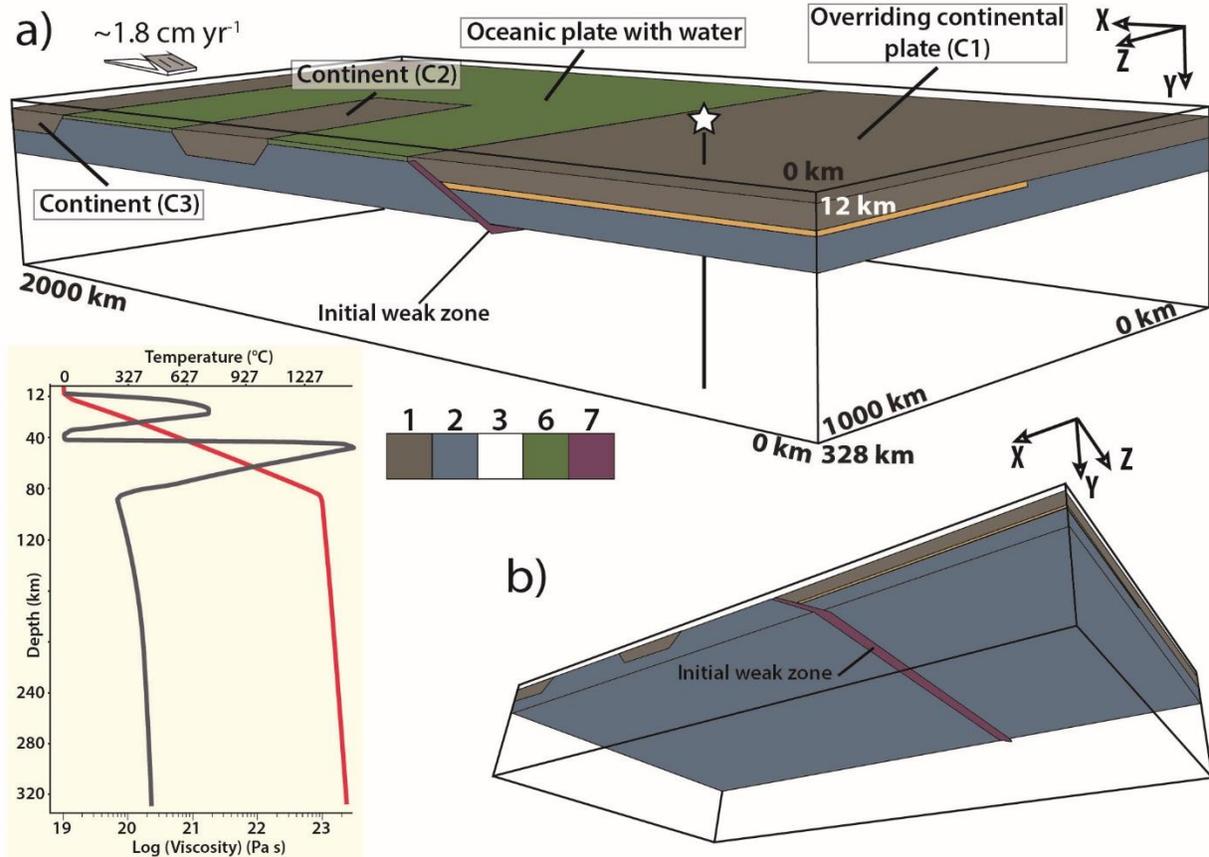
766 distribution of Pliocene-Quaternary volcanic rocks (from the geological map of the MTA). (b)

767 Curie-point depth map from Tselentis (1991); Stampolidis and Tsokas (2002) and Aydin *et al.*
768 (2005).

769



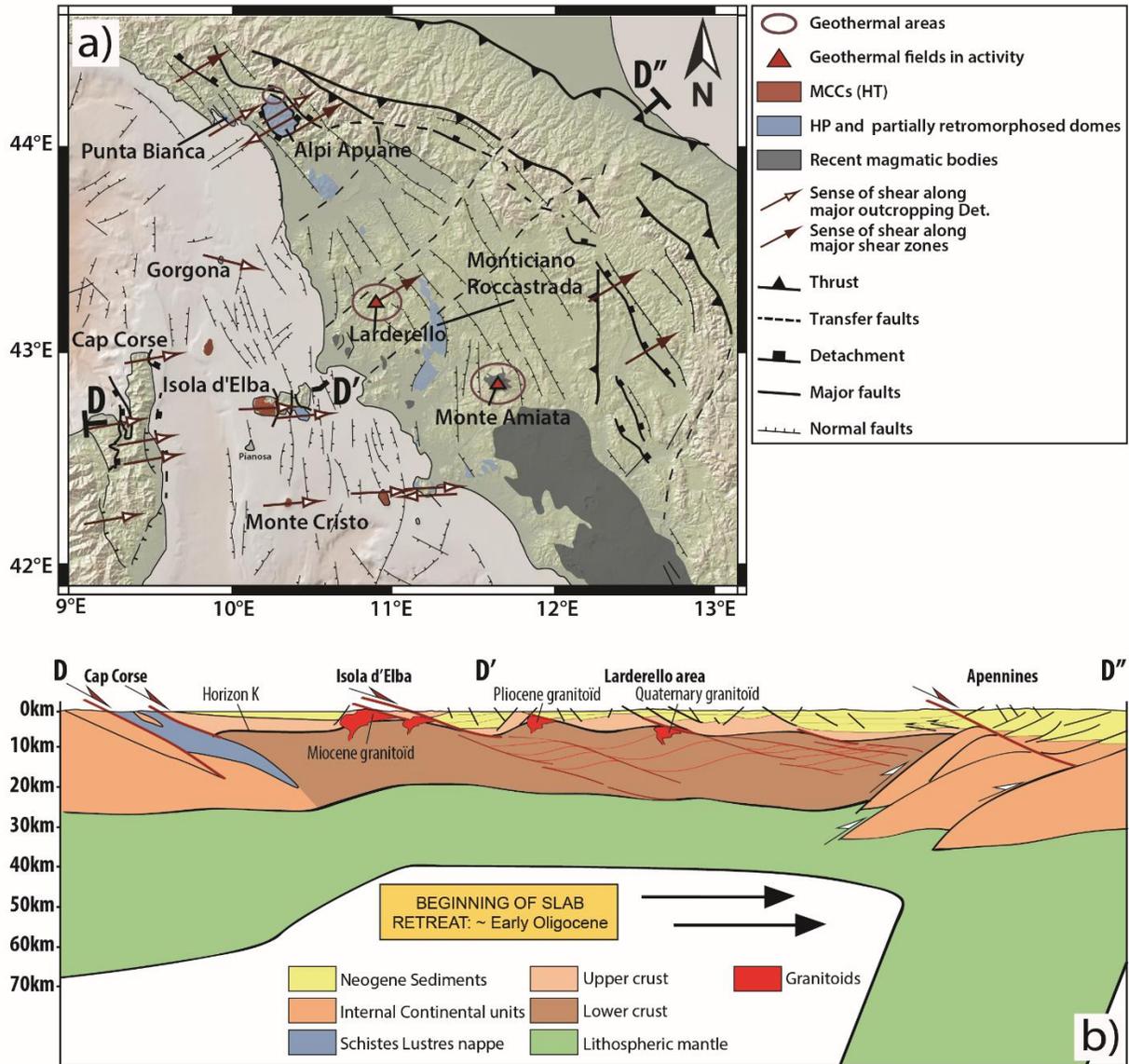
771 Figure S2: Location of videos shown by the vertical cross-section from our reference numerical
772 model. Colours showing different rock types: 1-2-3-4-5-6-7-8 as defined in Figure 3.



773

774 Figure S3: Experimental setup without slab tearing. (a) 3D initial model setup where the top layer
 775 (“sticky air”, $y < \sim 12 \text{ km}$) and the water are cut off for clarity. Colours showing different rock
 776 types as defined in Figure 3. Vertical viscosity (grey) and temperature (red) profiles at t_0 , taken
 777 at the location shown by the white star. (b) Location of the “weak zone” (i.e.
 778 hydrated/serpentinized mantle) into the lithosphere to initialise subduction (z-parallel).

779



780

781 Figure S4: Simplified tectonic map showing the main metamorphic core complexes and
 782 associated detachments faults. (a) Map of the Northern Tyrrhenian region and Northern
 783 Apennines, modified from Jolivet *et al.* (1998). (b) Simplified cross-sections highlighting slab
 784 retreat and formation of crustal detachments.

785

786 Videos S1 – S4 showing the ascent of partially molten material (the yellowish color), the thermal
 787 domes, the composition and the second invariant of the strain rate tensor evolution, respectively.

788 Note that the location of the modeled arc in our numerical model is the position of mantle
789 upwellings (i.e. molten material) projected at the surface. Even if the distance between arc and
790 trench changes in time and space, note that the distance should be around 100 – 150 km from the
791 trench toward the upper plate.

792

793 2. Supplementary materials

794

795 2.1. 3-D numerical approach

796

797 *Continuity and momentum equations*

798 The code I3ELVIS uses a staggered finite difference scheme combined with the marker-in-
799 cell technique (Gerya, 2010). Lagrangian advecting markers are thus combined with an immobile
800 Eulerian grid. The incompressible continuity equation describes the conservation of mass during
801 the deformation of a continuous medium (rocks) as:

$$802 \operatorname{div}(\vec{v}) = 0,$$

803 (2)

804 where \vec{v} is the local velocity and $\operatorname{div}()$ is the divergence operator.

805 The conservation of momentum is obtained through the solution of the Stokes equation:

$$806 \frac{\partial \sigma'_{ij}}{\partial x_j} - \frac{\partial P}{\partial x_i} + \rho g_i = 0,$$

807 (3)

808 where i and j denote the coordinate indices (i.e. x, y, z), σ' is the deviatoric stress tensor, x_i and
809 x_j are the spatial coordinates and g_i is the i^{th} component of the gravity vector \vec{g} .

810

811 *Heat conservation equation*

812 In order to predict changes in temperature due to heat transport, the Lagrangian heat
813 conservation equation is solved, which consider conductive heat transport, as well as internal heat
814 generation:

$$815 \rho C_p \frac{DT}{Dt} = -\frac{\partial q_i}{\partial x_i} + H,$$

816 (4)

817 where the heat flux vector \vec{q} is related to the temperature gradient in space as:

$$818 q_i = -k \frac{\partial T}{\partial x_i}.$$

819 (5)

820 T is temperature and k is the thermal conductivity of the material, which depends on pressure (P),
821 T , and composition of the material. C_p is the isobaric heat capacity, $\frac{DT}{Dt}$ is the substantive time
822 derivative of temperature and H is the volumetric heat productions. The latter term includes
823 several types of heat generation/consumption processes:

$$824 H = H_r + H_s + H_a ,$$

825 (6)

826 where H_r is the radiogenic heat production, H_s is the shear heating and H_a is the adiabatic heat
827 production. H_s is related to dissipation of the mechanical energy during irreversible non-elastic
828 (e.g. viscous) deformation. It is calculated as:

$$829 H_s = \sigma'_{ij} \dot{\epsilon}'_{ij} ,$$

830 (7)

831 where $\dot{\epsilon}'$ is the strain rate tensors. H_a corresponds to heat production related to pressure changes.

832 It is calculated as:

$$833 \quad H_a = T\alpha \frac{DP}{Dt},$$

834 (8)

835 where $\frac{DP}{Dt}$ is the substantive time derivative of pressure. Advection of temperature is performed

836 by using Lagrangian markers (Gerya, 2010).

837

838 *Viscous rheology of rocks*

839 The physical properties of rocks used in our experiments are given in Table 1. Among

840 them, the density of solid rocks depends on P , T , and composition. In our experiments, ρ_{solid} is

841 calculated as follow:

$$842 \quad \rho_{solid} = \rho_{0solid} [1 + \beta(P - P_r)] * [1 - \alpha(T - T_r)],$$

843 (9)

844 where ρ_{0solid} is the standard density of solid rocks at reference pressure P_r (10^5 Pa = 1 bar) and

845 temperature T_r (298.15 K = 25 °C); α and β are respectively the thermal expansion and

846 compressibility of rock. As density is also modified by phase changes (such as partial melting

847 which is implemented in this code), effective rock density ρ_{eff} is calculated as follow:

$$848 \quad \rho_{eff} = \rho_{solid} \left(1 - M + M \frac{\rho_{0molten}}{\rho_{0solid}}\right), \quad (10)$$

849 where $\rho_{0molten}$ is the standard density of molten rock, and M is the volumetric melt fraction in

850 partially molten rock (see below for more information). In addition, non-Newtonian visco-plastic

851 rheologies used in our experiments are implemented by both ductile and brittle/plastic

852 experimentally constrained laws. Effective viscosity for dislocation creep thus depends on the
853 strain rate, pressure and temperature. It is calculated as follow:

$$854 \quad \eta_{creep} = \dot{\epsilon}_{II}^{\frac{1-n}{n}} A_D^{\frac{-1}{n}} \exp\left(\frac{E+P V}{n R T}\right),$$

855 (11)

856 where $\dot{\epsilon}_{II}$ is the square root of the second invariant of the strain rate tensor, R is the gas constant,
857 n is the creep exponent, A_D is the pre-exponential factor, E is the activation energy) and V is the
858 activation volume. These parameters depend on the applied viscous flow law (i.e. wet quartzite,
859 plagioclase (An₇₅), dry olivine and wet olivine flow laws; Table 1) (Ranalli, 1995). This ductile
860 behavior is limited with a brittle/plastic one, implemented by using the Drucker-Prager criterion
861 σ_{yield} (i.e. plastic strength) (Ranalli, 1995) as follow:

$$862 \quad \eta_{creep} \leq \frac{\sigma_{yield}}{2\dot{\epsilon}_{II}},$$

863 (12)

864 where σ_{yield} is calculated as:

$$865 \quad \sigma_{yield} = \sigma_c + P \sin(\varphi), \tag{13}$$

866 where σ_c is the compressive strength of the rock, P is the dynamic pressure and φ is the effective
867 internal friction angle depending on the fluid and melt presence (see details in Gerya and Meilick
868 (2011)).

869

870 *Fluid/melt dynamics: example of partial melting*

871 In our experiments, partial melting process is also included. For all crustal rocks, the
872 volumetric melt fraction (M_0) is calculated as follow:

$$873 \quad M_0 = 0 \text{ at } T < T_{solidus}$$

874 $M_0 = \frac{T - T_{solidus}}{T_{liquidus} - T_{solidus}}$ at $T_{solidus} < T < T_{liquidus}$

875 (14)

876 $M_0 = 1$ at $T > T_{liquidus}$

877 where M_0 is the volumetric fraction of melt with no previous melt extraction, $T_{solidus}$ and $T_{liquidus}$
 878 are, respectively, the solidus and liquidus temperatures depending on the pressure and rock
 879 composition (Table 1). For the mantle, the degree of partial melting depends on the water
 880 content according to the parameterization of Katz *et al.* (2003). Resulting partially molten
 881 rocks can then undergo a succession of melt extraction episodes depending on their melt
 882 extraction threshold M_{max} and non-extractable melt fraction M_{min} , both equal to 1%. Only the
 883 exceeding volumetric melt fraction (i.e. > 1 %) is then extracted from partially molten rocks,
 884 thus generating magma markers. For each extraction episode, the melt fraction M_{ext} recorded in
 885 the partially molten rock markers is calculated as:

886 $M_{ext} = M - M_{min},$

887 (15)

888 where M is the volumetric melt fraction in partially molten rock, accounting for previous melt
 889 extraction events. It is calculated as:

890 $M = M_0 - \Sigma M_{ext},$

891 (16)

892 where ΣM_{ext} is the total melt fraction from previous extraction episodes. Then, M varies
 893 dynamically until remaining solid rock is no longer able to undergo additional partial melting
 894 (i.e. $\Sigma M_{ext} > M_0$). Magma markers resulting from these extraction episodes are then
 895 instantaneously transmitted to the surface as free melt is assumed to propagate upward much
 896 faster than rocks deform (Hawkesworth, 1997).

Material	k	H _r	ρ ₀	C _p	E _a	V _a	α	β	n	η ₀	Viscous flow law	sin (Φ _{eff})
	(W m ⁻¹ K ⁻¹)	(mW m ⁻³)	(kg m ⁻³)	(J kg ⁻¹ K ⁻¹)	(kg mol ⁻¹)	(m ³ mol ⁻¹)	(K ⁻¹)	(kbar ⁻¹)		(Pa ⁿ s)		
Sticky-air	20	0	1	100	0	0	0	0	1	1x10 ¹⁹	Air	0
Water	20	0	1000	3330	0	0	0	0	1	1x10 ¹⁹	Water	0
Sediment	0.64+807/(T+77)	2	2600	1000	154	8	3x10 ⁻⁵	1x10 ⁻³	2.3	1.97x10 ¹⁷	Wet Qz.	0.15
Cont. Crsut (C1 & C3)	0.64+807/(T+77)	2	2750	1000	154	8	3x10 ⁻⁵	1x10 ⁻³	2.3	1.17x10 ¹⁷	Wet Qz.	0.15
Cont. Crsut (2)	1.18+807/(T+77)	2	2950	1000	238	8	3x10 ⁻⁵	1x10 ⁻³	3.2	4.8x10 ²²	Wet Qz.	0.15
Oc. Crust	1.18+474/(T+77)	0.25	3000	1000	238	8	3x10 ⁻⁵	1x10 ⁻³	3.2	4.8x10 ²²	Wet Qz.	0.15
Mantle	0.73+1293/(T+77)	0.02	3300	1000	532	8	3x10 ⁻⁵	1x10 ⁻³	3.5	3.98x10 ¹⁶	Dry Ol.	0.6
Weak zone	0.73+1293/(T+77)	0.05	3300	1000	47	8	3x10 ⁻⁵	1x10 ⁻³	4	5x10 ²⁰	Wet Ol.	0

897

898 Table 1: Material properties used in the numerical experiments. k denotes the thermal
899 conductivity, H_r is the radiogenic heat production, ρ₀ is the density, C_p is the specific heat
900 capacity, E_a is the activation energy, V_a is the activation volume, α is the thermal expansion, β is
901 the compressibility coefficient, n is the stress exponent, η₀ is the reference viscosity and Φ_{eff} is
902 the effective internal friction angle. Qz. and Ol. correspond to the abbreviations of Quartzite and
903 Olivine.

904

905 2.2. 2-D numerical approach

906

907 The numerical model presented in Figures 7 and 8 has been built on the same basis as those
908 published by Guillou-Frotier *et al.* (2013). Darcy's law is coupled with the heat equation in a
909 porous medium and with mass conservation. We have used the Comsol Multiphysics™ software,
910 where temperature-dependent properties can be easily implemented. Benchmark experiments
911 have already been performed and published (Garibaldi *et al.*, 2010).

912 The numerical mesh consisted in more than 32,000 triangles, with a maximum size of 40 m
913 within the detachment. Petrophysical properties are given in Table 2. Fluid density law was
914 chosen to fit the experimental data for pure water between 20 and 350 °C:

$$915 \rho_L(T) = 1002.4 - 0.1905 * T - 0.0025 * T^2$$

916 (17)

917 where T is temperature in °C and ρ_L the fluid density in kg m^{-3} . For temperature greater than
 918 350°C, fluid density is assigned a constant value of 630 kg m^{-3} . This choice has no fundamental
 919 consequences on the results since at these high temperatures, fluid is located at depth greater than
 920 5 km, where permeability does not allow fluid circulation. For the fluid dynamic viscosity, the
 921 law used has been written as:

$$922 \quad \mu(T) = 2.414 \cdot 10^{-5} \times \exp\left(\frac{570}{T+133}\right)$$

923 (18)

924 where T is in °C and dynamic viscosity μ is in Pa s.

925 The initial pressure field increases linearly with depth. The initial temperature field is
 926 computed in a conductive regime, such that the surface temperature is 20 °C and the bottom of
 927 the box (10 km depth) is close to 500 °C, as it is suggested by Aydin *et al.* (2005). To reach a
 928 temperature around ~ 500 °C at a depth of 10 km, a basal heat flow of 120 mW m^{-2} is imposed
 929 during the transient behaviour. At $t > 0$, permeability values (see Table 2) are affected to the two
 930 units, and transient evolution of temperature and velocity fields are recorded. The chosen
 931 permeability law accounts for depth-dependence reaching a low value ($\sim 10^{-18} \text{ m}^2$) at the brittle-
 932 ductile transition, as suggested by Violay *et al.* (2017).

	Basement	Detachment
Eq. thermal conductivity(*)	2.8	1.6
Porosity (%)	5	30
Permeability (**) (m^2)	$[10^{-17} - 10^{-14}] \exp((z-2000)/1000)$	$[10^{-17} - 10^{-14}]$

Heat production ($\mu\text{W}\cdot\text{m}^{-3}$)	1.0	0
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933 Table 2: Material properties used in the numerical experiments. (*) the equivalent thermal
934 conductivity accounts for solid matrix and fluid-filled pores. (**) permeability decreases
935 exponentially with depth, except in the detachment.

936

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972