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8 3-D Pn tomography reveals continental subduction at the  
9 boundaries of the Adriatic microplate in the absence of a  
10 precursor oceanic slab

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23 **Abstract** Slab pull generated by subducting oceanic lithosphere is generally  
24 considered as a major trigger for the onset of continental subduction. However, this  
25 may be in conflict with the occurrence of UHP terranes bearing no evidence of  
26 oceanic lithospheric rocks involved in the exhumation cycle. Here, we image the  
27 uppermost mantle P velocity structure beneath the Central Mediterranean, suggesting  
28 the possibility that the initiation of continental subduction may not require a precursor  
29 oceanic slab. We combine (i) a three-step inverted 3-D Pn tomography model of the  
30 Adriatic microplate with (ii) available geologic constraints and palinspastic  
31 reconstructions of the Africa-Eurasia plate-boundary zone. Our Pn tomography model  
32 reveals elongated regions with  $V_p < 7.6$  km/s around the Adriatic microplate, clearly

33 connected with the slab structure inferred from teleseismic P wave tomography and  
34 supportive of continental subduction along the Dinaric, Alpine and Apenninic  
35 subduction zones. Contrasting styles of subduction are observed on the opposite sides  
36 of the Adriatic microplate: a laterally variable SW-dipping subduction is documented  
37 beneath the Apennines, continental to the north and oceanic to the south, where  
38 rollback is faster; a laterally continuous NE-dipping continental subduction is  
39 documented under the Dinarides. The lack of a precursor oceanic slab under the  
40 Dinarides demonstrates that the onset of continental subduction, in complex plate-  
41 boundary zones, can be controlled by plate-tectonic processes far away from the  
42 subduction initiation site, and may take place without the contribution of the negative  
43 buoyancy of an old oceanic lithosphere.

#### 44 **1. Introduction**

45 Continental subduction has long been dismissed in the light of the intrinsic  
46 buoyancy of continental crust (McKenzie 1969) but it is now considered as a  
47 relatively common geodynamic process, either demonstrated by the occurrence of  
48 ultra-high pressure (UHP) metamorphic rocks of continental origin exposed at the  
49 Earth's surface (e.g., Guillot et al., 2009 and references therein) or by high-resolution  
50 geophysical imagery (e.g., Schneider et al., 2013; Zhao et al., 2015). The driving  
51 force for continental subduction is generally thought to be provided by the negative  
52 buoyancy of old oceanic lithosphere that enters the subduction zone before the  
53 adjoining continental margin (e.g., Davies, 1999). However, there are examples of  
54 UHP terranes that bear no evidence of oceanic lithosphere involved in the exhumation  
55 cycle, e.g., the Dabie Shan in China (Zhang et al., 2009) or the Western Gneiss  
56 Region in Norway (Hacker, 2007). This may either suggest that the negative  
57 buoyancy of eclogitized oceanic lithosphere is too high to permit its partial  
58 exhumation, or that the continental subduction does not necessarily need the trigger of  
59 a large oceanic slab. And in some specific cases the role of slab pull may be  
60 negligible to enhance the subduction (e.g., Guillaume et al., 2013). Such a challenging  
61 scenario requires validation in a well-constrained subduction setting where the

62 physiography of the subducting plate is independently assessed, and the occurrence of  
63 a precursor oceanic slab can be safely excluded.

64 The Adriatic microplate, located within the complex plate boundary zone  
65 between Africa and Eurasia (e.g., [Faccenna et al., 2014](#)) is an ideal setting to test this  
66 hypothesis (Fig. 1a). This microplate is bounded by three orogenic belts (Alps,  
67 Apennines and Dinarides) mainly formed during Cenozoic subductions (e.g., [Handy  
68 et al., 2010](#); [Malusà et al., 2015](#)), and the physiography inherited from the Mesozoic  
69 opening of the Alpine Tethys is exceptionally well preserved (e.g., [Winterer and  
70 Bosellini, 1981](#); [Fantoni and Franciosi, 2010](#)). However, along the boundaries of the  
71 Adriatic microplate, UHP rocks attesting Cenozoic continental subduction are only  
72 observed atop the European slab in the Western Alps ([Handy et al., 2010](#); [Zhao et al.,  
73 2015](#)). No Adria-derived UHP rock has been exhumed in the Apenninic and Dinaric  
74 belts ([Jolivet et al., 2003](#); [Ustaszewski et al., 2008](#); [Malusà et al., 2015](#)), which means  
75 that the occurrence of Adriatic continental crust possibly reaching mantle depths has  
76 to be confirmed by geophysical evidence.

77 In this article, we employ the inversion of Pn phases to map the seismic velocity  
78 signature of continental subduction along the boundaries of the Adriatic microplate.  
79 Pn phases, the first arrivals at regional distance, propagate through the crust, penetrate  
80 into the uppermost mantle and are finally refracted through the Moho (e.g., [Hearn,  
81 1999](#)) and return to the free surface. Geophysical images based on the analysis of Pn  
82 phases are already available for the Mediterranean region (e.g., [Pei et al., 2011](#); [Diaz  
83 et al., 2013](#); [Lü et al., 2017](#)), but these images have low resolution and are based on a  
84 2-D inversion scheme ([Hearn, 1999](#)) that strongly depends on crustal depth  
85 corrections. This implies that large errors may affect the inversion of these Pn  
86 datasets, especially in complicated tectonic settings as the Africa-Eurasia plate  
87 boundary zone. Here we apply a three-step Pn traveltimes inversion scheme to invert  
88 the 3-D wavespeed structure of the uppermost mantle beneath the Adriatic microplate  
89 using Pn arrivals recorded since the 1960s by permanent and temporary stations (Fig.  
90 1b). The resulting 3-D tomography model with resolution of  $0.75^\circ \times 0.75^\circ$ , when

91 interpreted within the framework of available geologic and geodynamic constraints,  
92 provides seismic evidence of continental subduction in the lack of a precursor oceanic  
93 slab, with implications for our understanding of subduction processes in general.

## 94 **2. Tectonic setting**

95 The Adriatic microplate is the result of breakup of the northern Gondwana  
96 margin and consequent opening of the Alpine Tethys in the Jurassic (e.g., [Handy et](#)  
97 [al., 2018](#); [von Raumer et al., 2013](#)). The architecture inherited from the Mesozoic  
98 Tethyan rifting is largely preserved within the major thrust sheets of the Southern  
99 Alps, the Apennines and the Dinarides (Fig. 1a), and it is also imaged by oil-industry  
100 surveys beneath the Cenozoic sedimentary successions of the Po Plain and the  
101 Adriatic Sea (e.g., [Fantoni and Franciosi, 2010](#)). Beneath the Po Plain, the Adriatic  
102 microplate preserves evidence of a series of Mesozoic platforms, ridges and plateaus  
103 (e.g., Lugano ridge and Trento plateau) separated by NNE-SSW trending basins  
104 floored by continental crust (e.g., Lombardian and Belluno basins) (Figs. 1a, 2a). At  
105 the boundary with the Western Alps, a NNE-SSW trending gravimetric anomaly,  
106 classically referred to as the Ivrea body (IV in Fig. 1a; [Nicolas et al., 1990](#)), likely  
107 marks the lithospheric necking zone of the southern Tethyan margin ([Malusà et al.,](#)  
108 [2015](#)). Long-lasting carbonate platforms (Apulian, Dalmatia and Latium-Abruzzi) are  
109 preserved to the SE, where they are separated by an elongated basin floored by  
110 continental crust (Adriatic basin in Fig. 1a). Mesozoic oceanic crust is preserved  
111 farther south in the Ionian basin (Fig. 1a).

112 The NNE-SSW structural grain inherited from Tethyan rifting shares the same  
113 orientation with several major faults mapped in the Alpine and Apenninic wedges  
114 (Fig. 1a). For example: the westernmost segment of the Insubric Fault (IF in Fig. 1a)  
115 is aligned to the Ivrea body in the Western Alps; the Giudicarie Fault (GF in Fig. 1a)  
116 is located at the boundary between the Mesozoic Lombardian basin and the Trento  
117 plateau; the NNE-SSW faults at the transition between the Northern and the Southern  
118 Apennines bound the Mesozoic Latium-Abruzzi platform (Fig. 1a).

119 The WNW-ESE trend of the Insubric Fault in the Eastern Alps may mirror the  
120 orientation of transform faults at the time of continental breakup, and a similar trend  
121 has been also proposed for the original boundary between the Adriatic continental  
122 crust and the Ionian oceanic crust farther south (Faccenna et al. 2014) (Fig. 2b, c).  
123 Within this scenario, the inferred Mesozoic spreading direction for the Adriatic  
124 margin would be nearly perpendicular to the direction of Adria-Europe convergence  
125 documented since the Late Cretaceous by paleomagnetic data (purple arrows in Fig.  
126 2b, c). In the Late Cretaceous, oceanic domains were thus expected to be present to  
127 the NW of the Adriatic microplate, in front of the incipient Alpine and Apenninic  
128 subduction zones, but not in correspondence with the future Dinarides.

129 Adria-Europe convergence in the Late Cretaceous was initially accommodated by  
130 SE-dipping Alpine subduction (e.g., Handy et al., 2010) and by progressive  
131 consumption of the Alpine Tethys and adjoined European paleomargin beneath the  
132 Adriatic microplate. The Apenninic and Dinaric subduction zones started developing  
133 along the western and eastern boundaries of the Adriatic microplate during the  
134 Paleogene (Fig. 2b). The Apenninic slab progressively shifted northward beneath  
135 Corsica-Sardinia (Malusà et al., 2015), and started interacting with the European slab  
136 by the end of the Oligocene (Malusà et al., 2016; Zhao et al., 2016), when the onset of  
137 Apenninic slab retreat determined the Neogene opening of the Ligurian-Provençal and  
138 Tyrrhenian backarc basins (Faccenna et al., 2014) (Fig. 2c). Apenninic slab retreat is  
139 associated with a progressive migration of calc-alkaline magmatism in the Apenninic  
140 forearc (Peccerillo and Frezzotti, 2015). A number of buried thrust fronts (a-to-d in  
141 Fig. 1a) developed during ongoing Adria-Europe convergence in front of the Northern  
142 Apennines (Malusà and Balestrieri, 2012), where the Plio-Quaternary sedimentary  
143 successions of the Po Plain record differential subsidence along the orogen strike  
144 (Bigi et al., 1990). Cenozoic continental subduction, well documented by geophysical  
145 data in the Western Alps and in the Apennines (Chiarabba et al., 2014; Zhao et al.,  
146 2015), still remains conjectural and poorly constrained in the Eastern Alps and the  
147 Dinarides.

148 **3. Methods and datasets**

149 The Pn tomography model presented in this work is based on the three-step 3-D  
150 approach proposed by Sun and Kennett (2016a; 2016b). The main steps of this  
151 approach include: (i) building of a 3-D broad-scale initial model; (ii) relocating all the  
152 events by a nonlinear relocater in the 3-D initial model; and (iii) performing  
153 traveltimes residuals inversion in spherical coordinates.

154 *3.1 The 3-D initial model*

155 A good 3-D initial model can significantly speed up the convergence of  
156 tomographic inversion, thus reducing the computational burden and the variance of  
157 traveltimes residuals (see supplementary material). Our 3-D model is defined on a  
158 regular computational grid interval of  $0.25^\circ$  in longitude (from  $1^\circ\text{W}$  to  $22^\circ\text{E}$ ) and  
159 latitude (from  $38^\circ\text{N}$  to  $52^\circ\text{N}$ ) and 5 km in depth (from 0 to 280 km) in the spherical  
160 coordinate system, for a total of  $\sim 0.3$  million grids. The Pn phases propagate from the  
161 source in the crust to the uppermost mantle and are refracted back to the crust due to  
162 vertical velocity gradients.

163 A good crustal velocity model may thus enhance the reliability of inversion of  
164 uppermost mantle velocities. The EPcrust model (Molinari and Morelli, 2011), which  
165 integrates a range of previous data by previous authors, provides a reliable image of  
166 continental-scale crustal properties, but locally underestimates the Moho depth. For  
167 example beneath the Western Alps ( $<40$  km instead of  $\sim 40\text{-}70$  km as inferred from  
168 recent receiver function analysis by Zhao et al., 2015). The ESC Moho model of Grad  
169 et al. (2009) better agrees with the recent results of Zhao et al. (2015), and is also  
170 considered to define the crust components of the 3-D initial model (Fig. 3).

171 For the mantle components, we use a mantle P velocity model derived from the  
172 global shear wavespeed model SL2013sv (Schaeffer and Lebedev, 2013) using the  
173 empirical relations. The P velocity deviations are first converted from the S velocity  
174 deviations relative to the ak135 model (Kennett et al., 1995) and then the absolute P

175 velocities are obtained from summation between the P deviations and the ak135  
176 model. This procedure was successfully tested in recent studies dealing with the  
177 inversion of uppermost mantle structures beneath Australia (Sun and Kennett, 2016a)  
178 and eastern China (Sun and Kennett, 2016b).

### 179 *3.2 Pn arrival data*

180 Pn phases sample deeper levels of the uppermost mantle as the epicentral distance  
181 of events becomes larger. In our study, the region under consideration is sufficiently  
182 wide to allow the penetration of Pn with long epicentral distance into the uppermost  
183 mantle. Figure 4 compares the Pn raypaths in the 1-D ak135 model and in the 3-D  
184 initial and inverted velocity models. In the 1-D model, the Pn phases propagate  
185 downward to 50 km at most at an epicentral distance of 10°, whereas in both 3-D  
186 models the Pn phases dive to ~100 km depth. This illustrates that the actual Pn wave  
187 propagation is much more complicated than the simplified assumption of running  
188 along paths immediately below the Moho, as expected for a stratified medium.

189 In this study, we used the Pn arrivals for events with magnitude > 3.0 archived at  
190 the International Seismological Centre (ISC, <http://www.isc.ac.uk>) from January 1960  
191 to November 2013. The waveforms of events with magnitude > 3.5 recorded from  
192 December 2013 to November 2016 were fetched from the EIDA Data Archives at  
193 GEOFON (<http://www.webdc.eu/webdc3/>). We also collected events recorded during  
194 the CIFALPS experiment in the Western Alps from July 2012 to September 2013, in  
195 order to improve the ray path coverage in the complex Alps-Apennines transition  
196 zone (Malusà and Balestrieri, 2012; Malusà et al., 2017). The first arrivals were first  
197 automatically picked using the autoregressive technique applied to Z-component  
198 traces (Leonard and Kennett, 1999). Then, all the automatically picked arrivals were  
199 examined visually and calibrated manually, and only arrivals with clear onsets were  
200 selected.

201 We performed rigid selections of available Pn arrivals before inverting the  
202 uppermost mantle structures. The first arrivals with epicentral distance between 2.0°

203 and  $12^\circ$  were considered to be Pn arrivals, which screen out the possible Pg phases at  
204 shorter epicentral distances. We also excluded any raypaths with Pn traveltime  
205 residuals larger than 8 s relative to the ak135 model (Kennett et al., 1995) to avoid  
206 erroneous pickings and other inconsistencies, although the 3-D model can tolerate  
207 larger residuals with considerations of lateral velocity perturbations. We dropped  
208 events with location differences greater than 10 km when comparing catalogue and  
209 relocation in the 3-D initial model (see following section), and kept the earthquake  
210 clusters of Fig. 1b to enhance the reliability of inverted Pn wavespeed. This led to a  
211 final dataset of 395,918 selected Pn arrivals for 9,519 events and 1,080 stations.

### 212 *3.3 Event relocation*

213 Focal parameters influence the traveltime of regional phases. These parameters  
214 are determined using different models, such as the JB Earth model for events archived  
215 at ISC before 2006, and the ak135 model for events archived since January 1st 2006.  
216 In order to eliminate potential bias in Pn traveltime using different models, we have  
217 relocated all the events from 1960 to 2016 by a nonlinear event locator using the 3-D  
218 initial model.

219 The nonlinear event locator precomputed the seismic traveltime at all 3-D grids  
220 for all permanent and portable stations using the multi-stage fast marching method  
221 (FMM) (de Kool et al., 2006). All parameters are defined on the regular grids in the  
222 spherical coordinates leading to avoid the Earth flatten approximation at regional  
223 scale. The predicted traveltime at an arbitrary location in the 3-D model is obtained by  
224 trilinear interpolation, and the four focal parameters (i.e., origin time, two epicentral  
225 parameters and depth) are determined by the fully nonlinear Neighborhood Algorithm  
226 (NA, Sambridge and Kennett, 2001). According to previous studies (de Kool and  
227 Kennett, 2014; Sun and Kennett, 2016a), this procedure of relocating events in a 3-D  
228 model before regular Pn tomography can effectively decrease the variance of  
229 traveltime residuals. In the study area, earthquake relocation reduced the variance of  
230 traveltime residuals from 10.8 to 1.99 (see supplementary material).

### 231 3.4 Inversion of $P_n$ arrivals

232 We used the Fast Marching TOMOgraphy (FMTOMO) scheme (Rawlinson and  
233 Urvoy, 2006), a nonlinear tomography approach assuming local linearity, to invert the  
234 uppermost mantle P velocity structures from  $P_n$  traveltimes residuals. The FMTOMO  
235 package can simultaneously invert multiple classes of body wave datasets including  
236 refractions, reflections and teleseismic data from passive and active source datasets.  
237 The multi-stage FMM (de Kool et al., 2006) is designed to solve the forward problem  
238 of traveltimes prediction in the 3-D model. The algorithm is the same used for event  
239 relocation, which makes traveltimes predictions compatible in the two procedures.

240 The subsurface inversion scheme is applied to minimize the objective function.  
241 The parametrisation is defined in 3-D spherical coordinates to avoid the Earth flatten  
242 approximation, which makes the FMTOMO approach suitable to solve regional and  
243 global tomography problems. With the FMTOMO scheme, we inverted the 3-D P  
244 velocity model of the uppermost mantle with resolution as high as  $0.75^\circ \times 0.75^\circ$ .  
245 During the  $P_n$  tomography, we allowed the crustal velocity to be updated so that the  
246 effects of velocity variations of the crust on the uppermost mantle were taken into  
247 account. A series of inversions with different damping and smoothing regularizations  
248 were performed to determine the optimal regularization, which was reached with  
249 damping and smoothing factors equal to 20. We performed 17 iterations of inverting  
250  $P_n$  traveltimes residuals. As shown in Fig. S2, the inversion converged after the first  
251 four iterations.

## 252 4. Resolution tests

253 We used the checkerboard approach to examine the ability of available data and  
254 FMTOMO scheme to discern small velocity heterogeneities in the uppermost mantle.  
255 By considering the variations of Moho topography and the 3-D crustal and mantle  
256 velocities, we placed a layered checkerboard velocity anomaly overlaid on the 3-D  
257 initial model beneath the Moho (see Figs. 5b, c). This setting is much different from

258 the traditional one with alternating fixed layers of velocity anomaly, which is not a  
259 realistic representation of the heterogeneous Earth (e.g., [Fichtner et al., 2009](#)).

260 We followed a well-established procedure: 1) taking the checkerboard perturbed  
261 velocity as true model and synthesizing the traveltimes using the same source-station  
262 configuration of the real data; 2) applying the inversion from the 3-D initial model  
263 and synthetic residuals. In the resolution tests, the velocity anomaly ranges from -0.3  
264 km/s to 0.3 km/s, about 3.75% perturbation from the globally average velocity in the  
265 uppermost mantle. We reduced the smoothing and damping parameters to 5 to  
266 balance the tradeoff between the resolution of data and the proximity of the inverted  
267 model to the starting model. Figure 5a illustrates the recovered checkerboard with  
268 one-layered velocity anomaly centered at depth of 50 km, and the corresponding  
269 vertical sections along latitude and longitude are displayed in Figs. 5b, c. The  
270 checkerboard patterns are recovered quite well beneath the Adriatic microplate at  
271 resolution of  $0.75^\circ \times 0.75^\circ$ .

272 The resolution is poorer for the northern and eastern margins of the study region,  
273 and for the area to the SW of the Italian peninsula. For this reason, only the region  
274 between  $4^\circ\text{E}$ - $20^\circ\text{E}$  and  $38^\circ\text{N}$ - $48^\circ\text{N}$  with raypath density greater than 1000 is  
275 considered for further interpretation. These resolution tests suggest that the 3-D  
276 FMTOMO inversion scheme can readily identify not only lateral changes, but also  
277 vertical changes in P wavespeed within the Adriatic microplate, based on the fact the  
278 regional phases dive deeper at larger distances (see Fig. 4). However, the size of the  
279 well-resolved area gets progressively smaller with increasing depth (Figs. 6a, b),  
280 because of less seismic rays sampling the deeper uppermost mantle (Fig. 4c).

## 281 **5. Results and interpretation**

### 282 *5.1 Seismic signature of continental subduction around the Adriatic microplate*

283 The 3-D Pn wavespeed structure of the Adriatic microplate is illustrated in Fig. 7,  
284 by three horizontal slices of the inverted 3-D P velocity model, corresponding to

285 depths of 50, 60 and 70 km. Four representative cross-sections of the initial and  
286 inverted models are shown in Fig. 8.

287 In our 3-D Pn tomography model, the expected seismic velocity signature of  
288 continental subduction in the upper mantle is given by P velocities  $<7.8 \pm 0.2$  km/s.  
289 Vp values  $>8$  km/s should in fact characterize the dry peridotites of the Alpine  
290 lithospheric mantle (e.g., [Solarino et al., 2018](#)), whereas the heterogeneous Adriatic  
291 lower crust may yield Vp values ranging between  $\sim 6.7$  km/s (for granulites with felsic  
292 to intermediate composition) to  $\sim 7.2$  km/s (for granulite-facies metapelites),  
293 increasing up to  $\sim 7.6$  km/s because of metamorphic phase changes and progressive  
294 eclogitization during subduction at depth  $>40$  km (e.g., [Solarino et al., 2018](#) and  
295 references therein). Oceanic subduction likely remains undetected in our tomography  
296 model, because of the high Vp values characterizing the oceanic crust after  
297 eclogitization (Vp  $>8.0$  km/s for mafic eclogites), which are virtually  
298 undistinguishable from the Vp values characterizing the lithospheric mantle of the  
299 upper plate.

300 Elongated regions characterized by Vp  $<7.8$  km/s (and locally  $<7.6$  km/s) are  
301 observed, in Fig. 7, all along the southwestern, northwestern and northeastern borders  
302 of the Adriatic microplate. In the Alps, a low Vp belt is confined between the south-  
303 to east-dipping frontal Alpine thrusts and the north-dipping thrusts of the Adriatic  
304 retroforeland (Fig. 7a). To the south, two belts of low Vp match the frontal thrusts of  
305 the Northern Apennines and the Dinarides. We interpret these low Vp belts as the  
306 evidence of continental subduction at the boundaries of the Adriatic microplate along  
307 the Dinaric, Alpine and Apenninic subduction zones.

### 308 *5.2 Continental subduction in cross section*

309 Our Pn traveltime inversion, when analyzed in cross section and compared to the  
310 initial broad-scale velocity model (reported in the left panel of Fig. 8), introduces  
311 additional velocity heterogeneities that are shown in the right panel of Fig. 8. In cross-  
312 section A-A', Vp  $<7.6$  km/s can be traced down to depths  $>70$  km beneath the

313 Northern Apennines, and down to depths >65 km beneath the Dinarides (yellow-to  
314 red colors in Fig. 8). The wedge-shaped yellow-to-red regions in cross-section A-A'  
315 are asymmetric. They document Adriatic continental crust located more than 20 km  
316 deeper than the Moho (i.e., the 7.8 km/s velocity interface) imaged at the western and  
317 eastern boundaries of cross-section A-A'. These findings are not consistent with  
318 simple crustal shortening across the Dinarides and the Northern Apennines, but are  
319 instead supportive of opposite-dipping continental subduction zones at the  
320 northeastern and southwestern boundaries of the Adriatic microplate (DS and AS in  
321 Fig. 8).

322 Cross-section B-B' highlights a progressive northward deepening of the 7.8 km/s  
323 velocity interface along the orogen strike of the Apennines, from ~40 km in the  
324 Southern Apennines to >70 km in the Northern Apennines.

325 The heterogeneities in P velocity observed along cross-section C-C' are consistent  
326 with an eastward-dipping continental subduction beneath the Western Alps (WA in  
327 Fig. 8), in agreement with receiver function results (Zhao et al., 2015). On the upper  
328 plate side of the Western Alps subduction, the 7.8 km/s velocity interface gets  
329 shallower in correspondence with the Ivrea body (IV in Fig. 8), in agreement with  
330 recent local earthquake tomography results (Solarino et al., 2018), and reaches a depth  
331 of ~50 km beneath the eastern Po Plain.

332 Cross-section D-D' highlights the relationships between the Apenninic subduction  
333 (AS in Fig. 8) and the southward-dipping Central Alps subduction (CS in Fig. 8),  
334 providing further evidence supporting the complex lithospheric structure beneath the  
335 Po Plain recently described by Malusà et al. (2018).

### 336 *5.3 Along-strike changes in Pn velocity structure*

337 The most relevant along-strike changes in velocity structure observed in the Pn  
338 tomography model are indicated by bold numbers 1 to 6 in Fig. 7a. As observed in  
339 Fig. 7, the low velocity belt corresponding to Dinaric subduction displays a

340 remarkable continuity in the 50 km depth slice from Austria to Albania, providing  
341 geophysical evidence of a laterally continuous Dinaric subduction zone. The Dinaric  
342 low-velocity belt terminates at high-angle (1 in Fig. 7) against the low-velocity belt of  
343 the Eastern Alps, which shows an ENE-WSW trend consistent with major thrust  
344 faults formed during Alpine subduction. Notably, this ENE-WSW low-velocity belt is  
345 not observed, beneath the Eastern Alps, at depths greater than 50-60 km (Fig. 7b, c).

346 On the northwestern edge of the Adriatic microplate, the abrupt change in  
347 orientation of the low- $V_p$  belts (2 in Fig. 7a) marks the boundary between the Alpine  
348 and Apenninic subduction zones. Based on our tomography model, underthrusting of  
349 continental crust by Apenninic subduction can be detected as far north as the Emilia  
350 thrust front (b in Fig. 1a). Farther west, the inferred  $V_p$  values (7.7-7.8 km/s at 50 km  
351 depth) are in line with a tectonic scenario including exhumation of hydrated mantle-  
352 wedge rocks at shallow depth (Liao et al., 2018), and with the P wave velocities  
353 documented by local earthquake tomography (Solarino et al., 2018).

354 The low  $V_p$  belt of the Northern Apennines shows a prominent break (3 in Fig.  
355 7a) and is displaced northeastward in correspondence with the boundary between the  
356 Emilia and Ferrara-Romagna thrust fronts (b and c, respectively, in Fig. 1a), where  
357 the thickness of the Pliocene-Quaternary foredeep successions of the Po Plain sharply  
358 increases from ~3 km to >8 km (Pieri and Groppi, 1981). Farther south, major along-  
359 strike velocity breaks are observed at the boundary between the Northern and the  
360 Southern Apennines (4 in Fig. 7a), in correspondence with major tectonic structures  
361 mapped at the surface (Fig. 1a). The low-velocity belt parallel to the Apenninic thrust  
362 fronts is no longer observed in the Southern Apennines (5 in Fig. 7a), which suggests  
363 that subduction in southern Italy was dominantly oceanic. Offshore Albania (6 in Fig.  
364 7a), the sharp increase in  $V_p$  (>8.0 km/s) at 60 km depth may mark the presence of  
365 oceanic crust, originally located to the south of the Adriatic basin and now squeezed  
366 within the Dinaric subduction zone.

367 *5.4 Comparison with previous images based on a 2-D inversion scheme*

368 Previous Pn tomography models of the area analyzed in this work (e.g., Diaz et  
369 al., 2013; Lü et al., 2017) are based on 2-D inversion schemes that are prone to  
370 average the uppermost mantle structure over depth. In the model assuming isotropic  
371 propagation that was presented by Diaz et al. (2013) (see their Fig. 8a), no major  
372 linear trend of Pn velocity anomalies is observed outside of the Alps. When an  
373 anisotropic term is included in their calculation (Diaz et al., 2013, their Fig. 8b),  
374 continuous low Pn velocity anomalies are instead imaged all along the Dinarides, the  
375 Apennines of central and southern Italy, and more discontinuously in the Alps.  
376 Unlike our 3-D Pn tomography model of Fig. 7a, no along-strike change from  
377 continental to oceanic subduction was detected by Diaz et al (2013) along the  
378 Apennines, and the relationships between Alpine and Apenninic subductions are not  
379 well resolved in their model. Similar velocity features are also observed in the model  
380 presented by Lü et al. (2017), which anyway highlights a southward increase in Pn  
381 velocities from the Northern (~7.7 km/s) to the Southern (~7.8 km/s) Apennines (Lü  
382 et al., 2017, their Fig. 10). The comparison of our results with previous Pn  
383 tomography models based on 2-D inversion suggests the importance of applying a 3-  
384 D inversion scheme, which can delineate vertical velocity variations, to get the Pn  
385 wavespeed structure of the upper mantle in tectonically complicated regions.

## 386 **6. Discussion**

### 387 *6.1 Comparison with available teleseismic P wave tomography models*

388 In Fig. 9, the main features of our 3-D Pn tomography model are summarized and  
389 compared with the slab structure highlighted by available teleseismic P wave  
390 tomography models (Zhao et al., 2016 and references therein). Outside of the well-  
391 resolved areas of the recent Zhao et al. (2016)'s model, slab traces in Fig. 9 are based  
392 on a previous, lower-resolution tomography model by Piromallo and Morelli (2003).  
393 In this figure, we observe a good match between the slab structure inferred from  
394 teleseismic P wave tomography (thick blue lines in Fig. 9) and the sites of European  
395 and Adriatic continental subductions documented by our 3-D Pn tomography (thick

396 brown lines in Fig. 9). Beneath the Dinarides, a NE-dipping slab was already detected  
397 by [Lippitsch et al. \(2003\)](#) and [Zhao et al. \(2016\)](#), but it was only resolved beneath the  
398 northernmost part of the Dinaric belt. Here, we provide evidence for a continuous belt  
399 of underthrust continental material down to >50 km depth from Austria to Albania,  
400 which is clearly connected with the NE-dipping slab previously imaged by teleseismic  
401 tomography beneath the Dinarides.

402 The relationships between the Alpine and Dinaric subductions become clearer  
403 when the pattern of continental subduction from Pn tomography is combined with the  
404 slab structure from 100 to 300 km depth as constrained by teleseismic P wave  
405 tomography (Fig. 9). At 50 km depth, the NW-SE low-velocity belt marking the NE-  
406 dipping subduction of Adriatic continental crust under the Dinarides terminates, to the  
407 north, against the ENE-WSW trending remnants of the eastern Alpine subduction  
408 zone. At 100 to 200 km depth, evidence of Alpine subduction is no longer observed  
409 beneath the Eastern Alps, but the NE-dipping Adriatic slab is instead detected farther  
410 north, beneath the remnants of the former Alpine subduction zone (see Fig. 9). This  
411 complex slab structure may possibly require a Paleogene slab-breakoff event in the  
412 Eastern Alps, an hypothesis that would deserve further geophysical investigations. To  
413 the west of the Giudicarie Fault, the trace of continental subduction inferred from Pn  
414 tomography is clearly connected with a SE-dipping European slab continuously  
415 imaged from the Central to the Western Alps, which is steeper in the vicinity of the  
416 Giudicarie Fault. In the Northern Apennines, the segmented low-Vp belt attesting  
417 SW-ward continental subduction of Adriatic crust matches with the trace of the  
418 Apenninic slab documented by teleseismic P wave tomography, where such a  
419 segmentation is not observed at 100 km or greater depths. In central Italy, at the  
420 boundary between the Northern and the Southern Apennines, the low-velocity belt  
421 marking SW-ward continental subduction is located on top of a gap in low-Vp  
422 anomaly centered at ~100 km depth, and interpreted as a possible slab window (e.g.,  
423 [Zhao et al., 2016](#)).

424 *6.2 Contrasting styles of subduction around the Adriatic microplate*

425 Our Pn tomography model supports the idea of contrasting styles of subduction of  
426 the opposite sides of the Adriatic microplate (Fig. 10): (i) to the NE, a laterally  
427 continuous NE-dipping continental subduction is observed under the Dinarides,  
428 possibly resuming to the north an older SE-dipping Alpine subduction; and (ii) to the  
429 SE, a laterally variable subduction is documented under the Apennines, continental to  
430 the north and oceanic to the south.

431 This scenario is in agreement with the presence of unsubducted Mesozoic oceanic  
432 crust in the Ionian Sea offshore Calabria (Fig. 9), which contrasts with the occurrence  
433 of Adriatic continental crust under the Adriatic sea floor. An oceanic slab beneath the  
434 Southern Apennines (5 in Fig. 10) may explain the faster retreat of the southern  
435 segment of the Apenninic trench compared to the northern one (Faccenna et al.,  
436 2014), where the interaction between the Alpine and Apenninic slabs may have  
437 controlled the location of the Corsica-Sardinia pole of rotation during the scissor-type  
438 opening of the Ligurian-Provençal and Tyrrhenian backarc basins (Malusà et al.,  
439 2016). It has been long recognized that Apenninic slab rollback was responsible for  
440 the progressive migration of Cenozoic orogenic magmatism from the southern French  
441 coast to the Apennines (yellow to purple marks in Fig. 9) (e.g., Carminati and  
442 Doglioni, 2012). Notably, the laterally variable composition of rocks subducted at the  
443 Apenninic trench, as determined by our Pn tomography model, is also mirrored by  
444 contrasting geochemical compositions of Quaternary magmas that are supportive,  
445 from the Campania province to the Aeolian Islands, of a major contamination of  
446 fluids released from the Ionian slab (purple lozenges in Fig. 9) (Peccerillo and  
447 Frezzotti, 2015). The slab window located at the boundary between the Northern and  
448 the Southern Apennines (4 in Fig. 10) may have formed right at the transition between  
449 oceanic lithosphere and more buoyant continental lithosphere subducted under the  
450 Northern Apennines.

451 Slab retreat is obviously hindered to the north by the interaction between the  
452 Apenninic and Alpine slab (2 in Fig. 10). Differential rates of slab retreat in the  
453 Northern Apennines are possibly accommodated by lithospheric-scale tectonic

454 discontinuities, as observed in the Pn tomography model thanks to the offset of  
455 velocity structure at the transition between the Emilia and Ferrara-Romagna thrust  
456 fronts (3 in Fig. 10). The picture provided by Pn tomography is confirmed not only by  
457 the differential subsidence in the foreland basin north of the Apennines, hosting a >8  
458 km thick Pliocene-Quaternary succession to the east compared to only ~3 km to the  
459 west (Pieri and Groppi, 1981), but also by the along-strike variations in erosional  
460 exhumation documented in the orogenic belt by low-temperature thermochronology  
461 (Malusà and Balestrieri, 2012).

### 462 *6.3 Continental subduction beneath the Dinarides without a precursor oceanic slab*

463 Our Pn tomography model is supportive of a laterally continuous continental  
464 subduction beneath the Dinarides. Notably, the Dinaric subduction imaged by  
465 geophysical data is a relatively young feature (Paleogene) of the Adria-Europe plate  
466 boundary zone. At the time of Dinaric subduction initiation, the Tethyan oceanic crust  
467 was already largely consumed, mainly by SE-dipping Alpine subduction (Fig. 2b).  
468 Mesozoic Tethyan crust was only preserved south of Adria, and was later subducted  
469 beneath the European continental margin in front of Sardinia (Fig. 2b, c). Remnants of  
470 Tethyan oceanic crust that underwent subduction to be exhumed as (meta)ophiolites  
471 are well documented in the suture zones overlying the main slabs (dashed green lines  
472 in Fig. 9).

473 Based on available reconstructions of Adria physiography inherited from  
474 Mesozoic rifting (see Figs. 1, 2), we can infer that rocks subducted beneath the  
475 Dinarides, in the Paleogene Bundva-Pindos trough (e.g., Kovács et al., 2007), likely  
476 had a continental affinity. Notably, ophiolitic rocks are exposed in the Sava and  
477 Vardar units farther east, which may suggest that continental subduction was preceded  
478 by oceanic subduction (see discussion in Kovács et al., 2007). However, this latter  
479 hypothesis can be safely ruled out by considering the spatial relationships between the  
480 Dinaric slab imaged by Pn tomography, the Sava and Vardar suture zones, and the  
481 subduction-related magmatism (Fig. 9). The (meta)ophiolites mapped all along the

482 boundaries of the Adriatic microplate systematically show a close spatial relationship  
483 with the underlying slabs, which is not observed in the case of the Dinarides. These  
484 relationships are preserved even for slabs that experienced major retreat, such as in  
485 the Apennines. Moreover, unlike the Apennines, the age of magmatism atop the  
486 Dinaric slab is not supportive of Dinaric slab retreat. We can thus conclude that the  
487 Vardar and Sava zones are relics of Mesozoic subduction zones, and that continental  
488 subduction in the Dinarides was not triggered by a precursor oceanic slab entering the  
489 subduction zone before the adjoining Adriatic paleomargin.

490       The joint geophysical-geologic analysis of the Adriatic microplate and  
491 surrounding areas presented in this work thus demonstrates that different styles of  
492 subduction can be expected in complex plate boundary zones such as the  
493 Mediterranean, and that the role of slab pull during the onset of continental  
494 subduction can be negligible. Although the presence of an oceanic slab may favour  
495 slab rollback, thus determining systematic relationships between trench retreat,  
496 backarc extension and magmatism as observed in the Apennines, continental  
497 subduction may also be triggered by changes in plate motion and in the force balance  
498 away from the subduction initiation site (e.g., [Stern, 2004](#)). This was likely the case of  
499 the Dinarides, where subduction was the result of convergent motion between Africa  
500 and Eurasia during coeval opening of the Atlantic Ocean, but there is no evidence of  
501 subducted oceanic lithosphere involved in the subduction-exhumation cycle. A similar  
502 mechanism may also be invoked for continental subduction in the Hindu Kush, where  
503 there has been no oceanic crust involved since at least the Late Cretaceous (e.g.,  
504 [Searle et al., 2001](#)).

## 505 **7. Conclusions**

506       Our three-step 3-D Pn tomographic velocity model of the Adriatic microplate and  
507 surrounding areas, analyzed within the framework of available geologic and  
508 geodynamic constraints, leads to the following main conclusions:

- 509 - Elongated regions with  $V_p < 7.8$  km/s down to depths  $> 50$  km around the  
510 Adriatic microplate, clearly connected with the slab structure inferred from  
511 teleseismic P wave tomography, are supportive of continental subduction  
512 along the Dinaric, Alpine and northern Apenninic subduction zones.
- 513 - The NW-SE low- $V_p$  belt marking Dinaric subduction displays a remarkable  
514 continuity at 50 km depth slice from Austria to Albania; it terminates to the  
515 north against the NNE-SSW low- $V_p$  belt of the Eastern Alps at deeper  
516 depths, which marks the remnants of the eastern Alpine subduction zone.
- 517 - A laterally variable SE-dipping subduction is documented beneath the  
518 Apennines, continental to the north and oceanic to the south; an oceanic slab  
519 beneath the Southern Apennines may explain the faster retreat of the southern  
520 segment of the Apenninic trench.
- 521 - The lack of a precursor oceanic slab under the Dinarides demonstrates that the  
522 onset of continental subduction, in complex plate-boundary zones, do not  
523 necessarily need a major contribution of the negative buoyancy of old oceanic  
524 lithosphere, but can be triggered by plate-tectonic processes far away from the  
525 subduction initiation site.

526 Our results may find application to other subduction zones, where the  
527 mechanisms of continental subduction initiation are still poorly understood.

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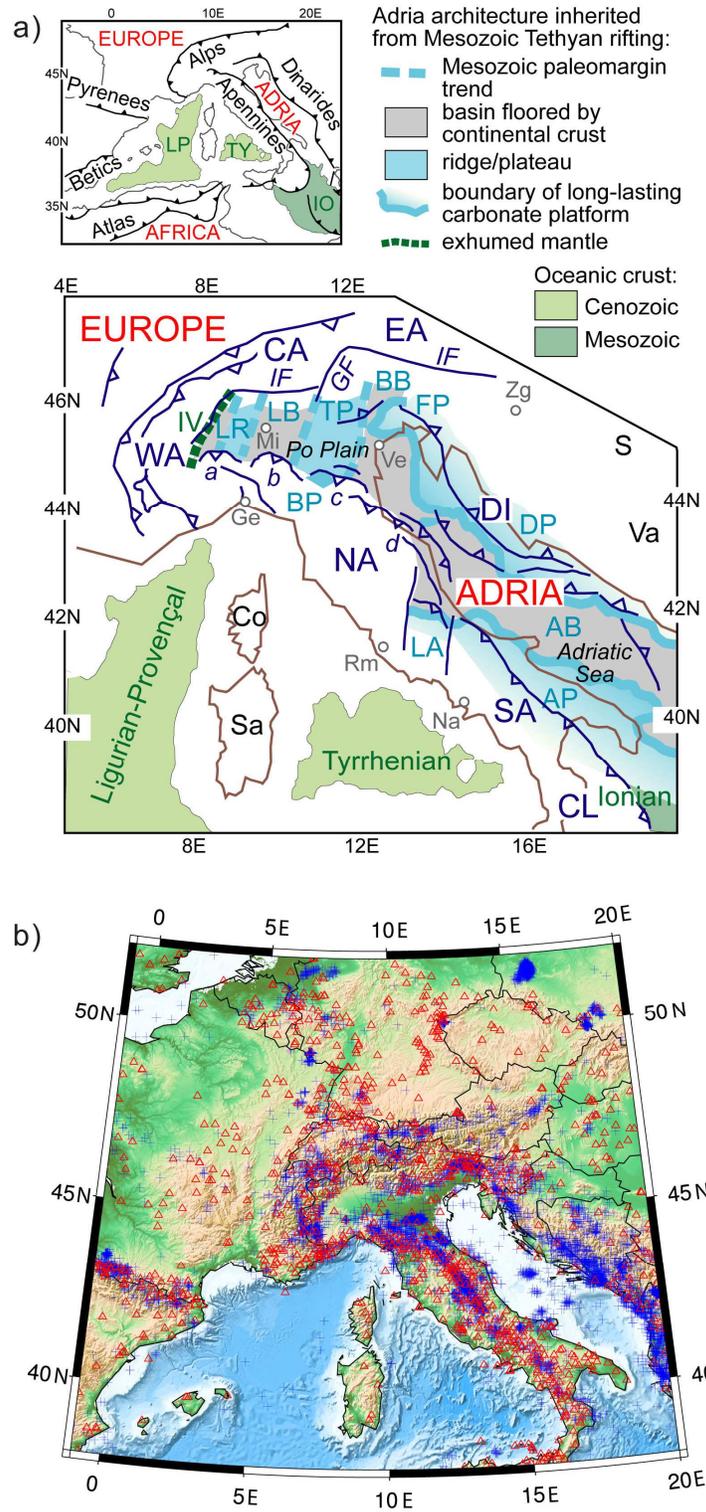
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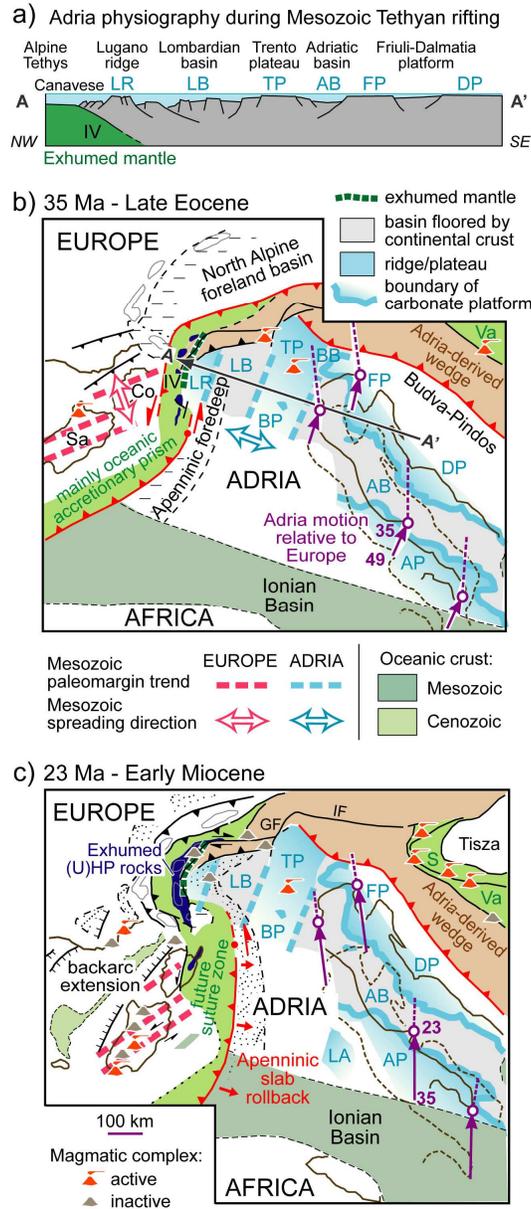
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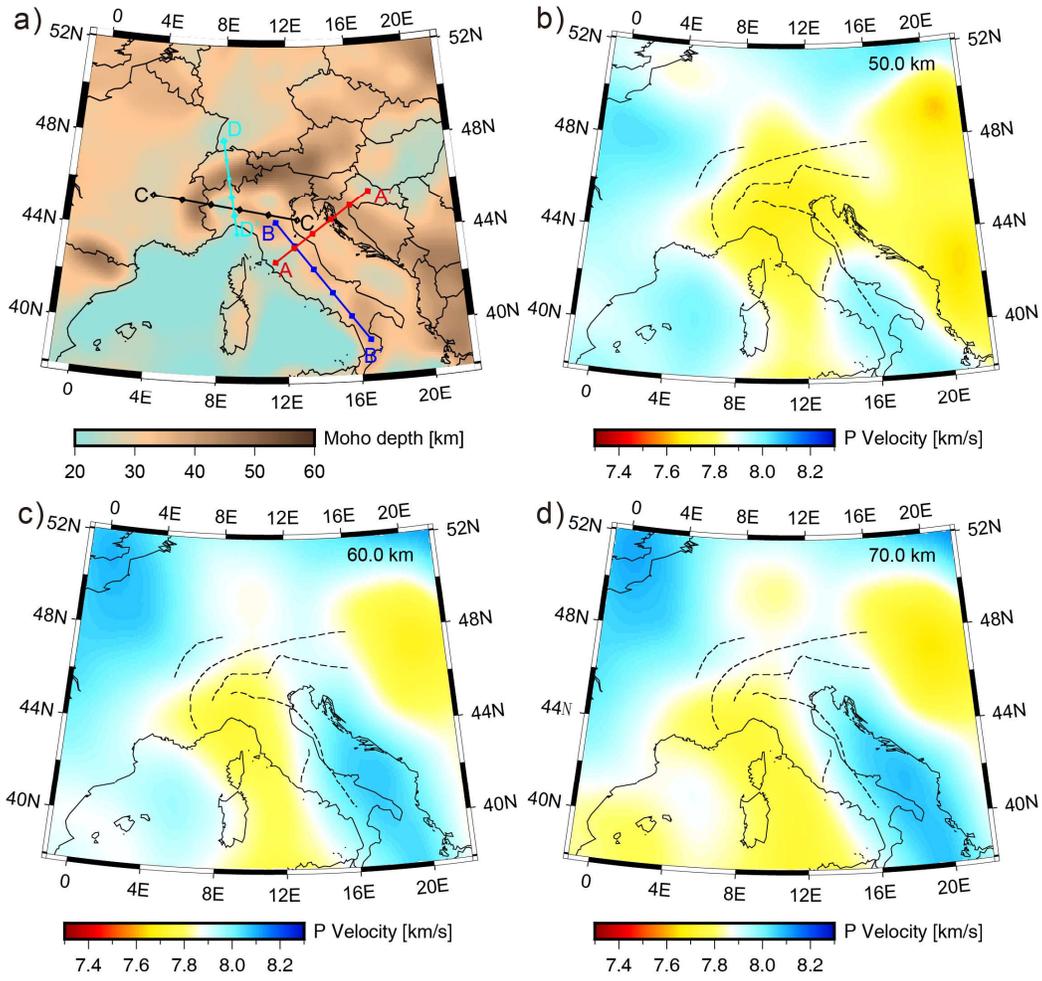
700 **Figure 1.** (a) Tectonic sketch map of the Central Mediterranean (top-left) and main  
 701 tectonic structures of the Adriatic microplate. Tectonic domains: CA, Central Alps;  
 702 DI, Dinarides; EA, Eastern Alps; IO, Ionian basin; LP, Ligurian-Provençal basin; NA,

703 Northern Apennines; SA, Southern Apennines; TY, Tyrrhenian basin; WA, Western  
704 Alps. Other keys: AB, Adriatic basin; AP, Apulian platform; BB, Belluno basin; BP,  
705 Bagnolo platform; Co, Corsica; DP, Dalmatia platform; FP, Friuli platform; GF,  
706 Giudicarie Fault; IF, Insubric Fault; IV, Ivrea body; LA, Latium-Abruzzi platform;  
707 LB, Lombardian basin; LR, Lugano ridge; S, Sava zone; Sa, Sardinia; TP, Trento  
708 plateau; Va, Vardar. Northern Apennine Thrust Fronts (a-to-d): a, Monferrato; b,  
709 Emilia; c, Ferrara-Romagna; d, Ancona. Main towns (in grey): Ge, Genoa; Mi, Milan;  
710 Na, Naples; Rm, Rome; Ve, Venice; Zg, Zagreb (based on [Bigi et al., 1990](#); [Jolivet et](#)  
711 [al., 2003](#); [Kovács et al., 2007](#); [Fantoni and Franciosi, 2010](#)). **(b)** Location of seismic  
712 stations (red triangles) and seismic events (blue crosses) considered in this study (grey  
713 lines = country borders).



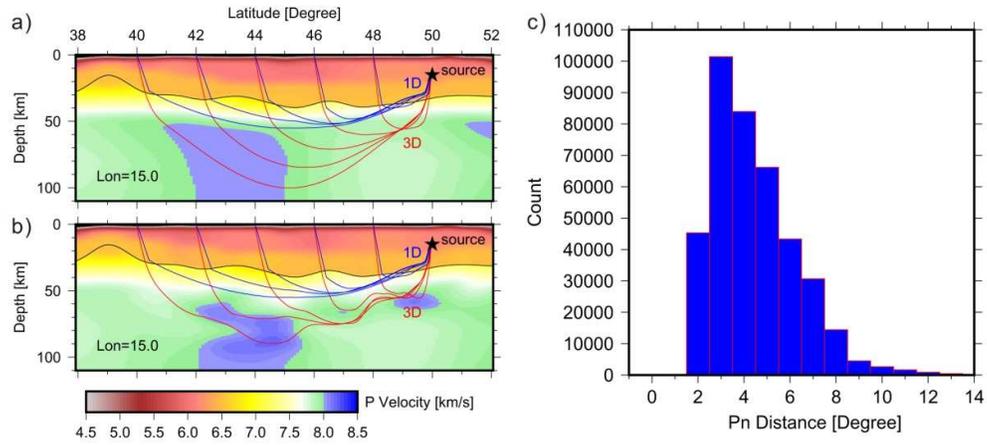
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715 **Figure 2.** (a) Extensional architecture of the Adriatic microplate during Tethyan  
 716 rifting, 2× vertical exaggeration, see location of cross-section in frame (b) (based on  
 717 Winterer and Bosellini, 1981; Bertotti et al., 1993). (b-c) Palinspastic reconstruction  
 718 of the Adria-Europe plate boundary zone during Eocene (U)HP rock exhumation in  
 719 the Western Alps (b) and the onset of Apenninic slab rollback (c). Purple arrows  
 720 indicate relative Adria-Europe relative plate motion (numbers = age in Ma).  
 721 Acronyms as in Fig. 1. Based on Malusà et al. (2015) and Fig. 1, Ionian-Africa  
 722 relationships after Faccenna et al. (2014).



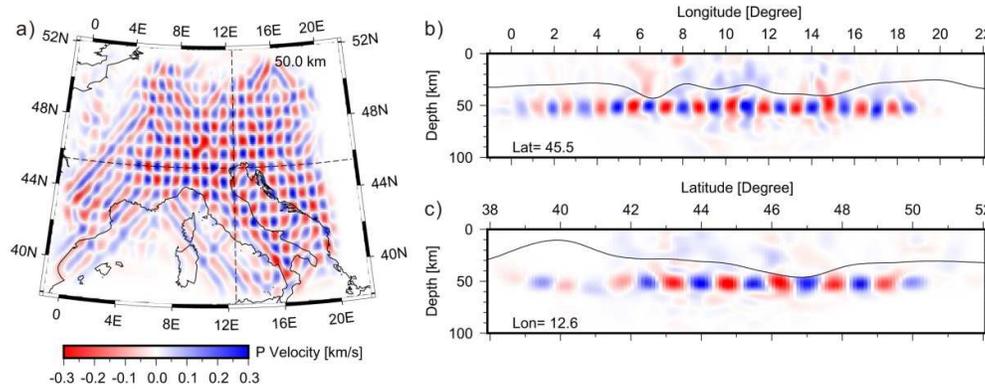
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724 **Figure 3.** (a) Moho depth from Grad et al. (2009), and constructed P velocity model  
 725 sliced at different depths: (b) 50 km, (c) 60 km and (d) 70 km. The initial model well  
 726 reflects the broad scale of the geologic settings. Lines in color in Figure 3a show the  
 727 locations of velocity cross-sections shown in Figure 8.



728

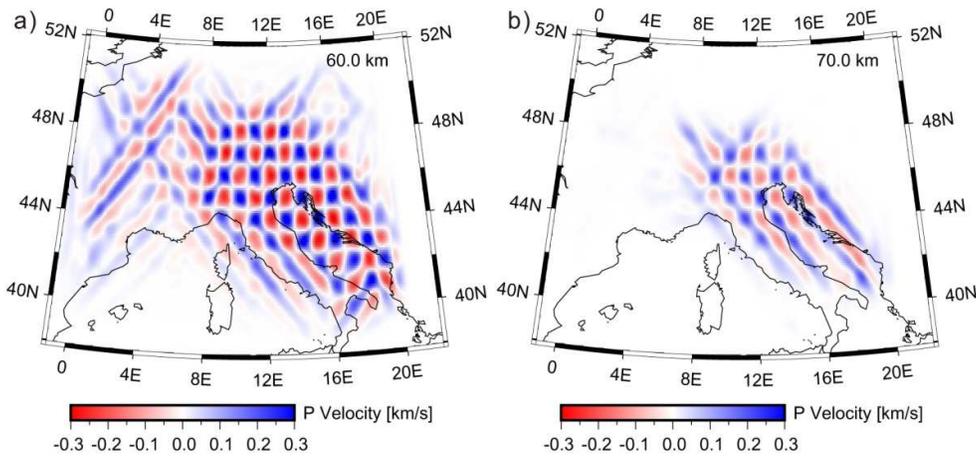
729 **Figure 4.** (a, b) Comparison of Pn raypaths between the 1-D ak135 model (blue lines)  
 730 and the 3-D initial (a) and inverted (b) models (red lines), to which the sliced velocity  
 731 profiles are referred to. (c) Histograms of Pn distances: most of them concentrates in  
 732 ranges of 2—8°, which indicates that the zone shallower than 80 km is well sampled.



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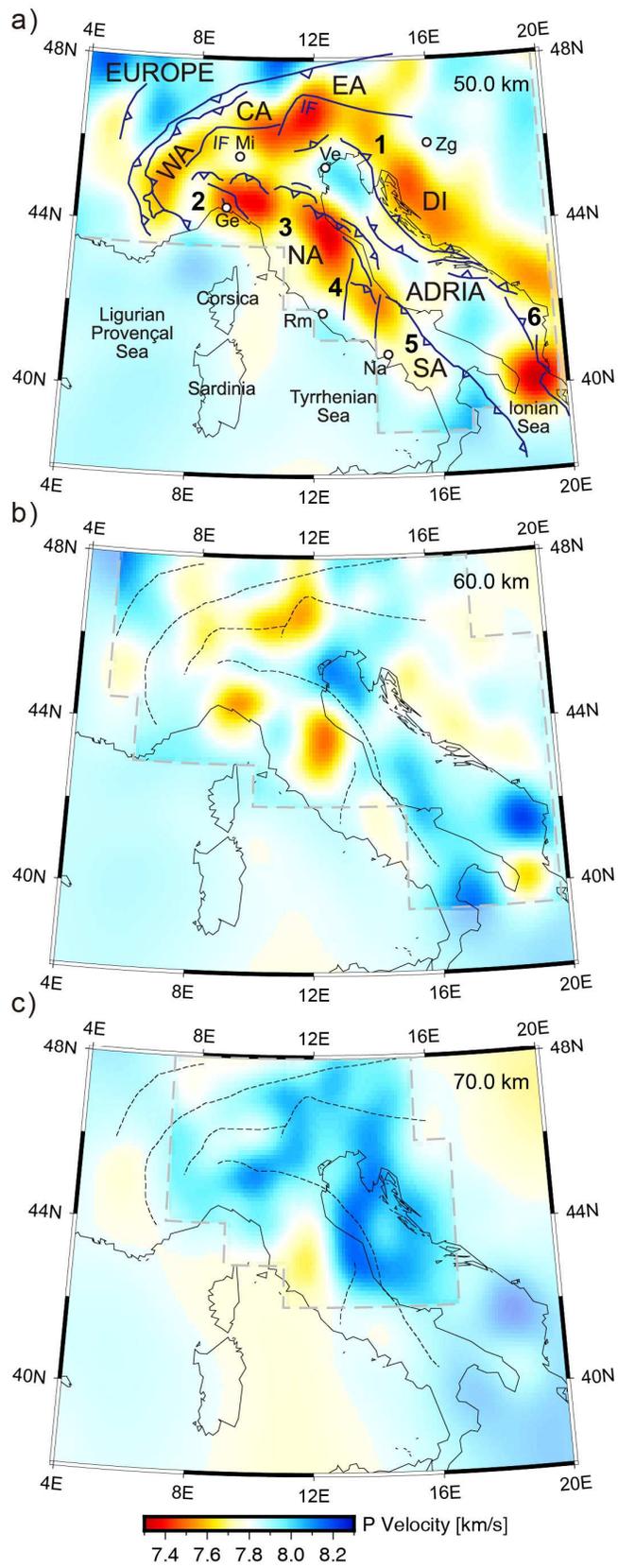
734 **Figure 5. (a)** Horizontal slice of the recovered checkerboard at depth of 50 km, and  
 735 vertical slices along latitude of  $45.5^\circ$  **(b)** and longitude of  $12.6^\circ$  **(c)**. The resolution is  
 736  $0.75^\circ \times 0.75^\circ$ . A layered velocity anomaly with 0.3 km/s is centred at depth of 50 km.  
 737 The dashed lines in Figure 5a show the locations of vertical slices in Figure 5b and 5c.

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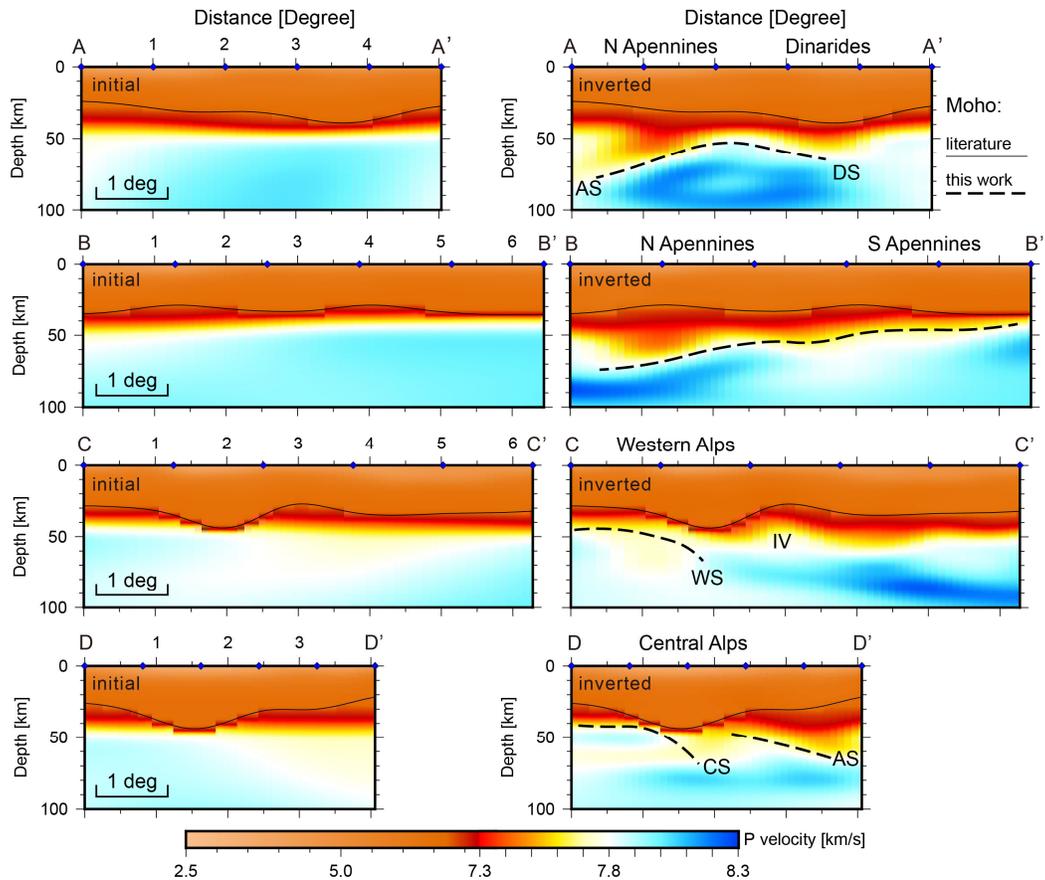


739

740 **Figure 6.** Recovered checkerboard of resolution tests horizontal sliced at depth of **(a)**  
 741 60 km and **(b)** 70 km. A layered velocity anomaly with 0.3 km/s is centred at depth of  
 742 60 km in Figure 6a while the layer is centred at 70 km in Figure 6b. The resolution is  
 743  $1^\circ \times 1^\circ$ .

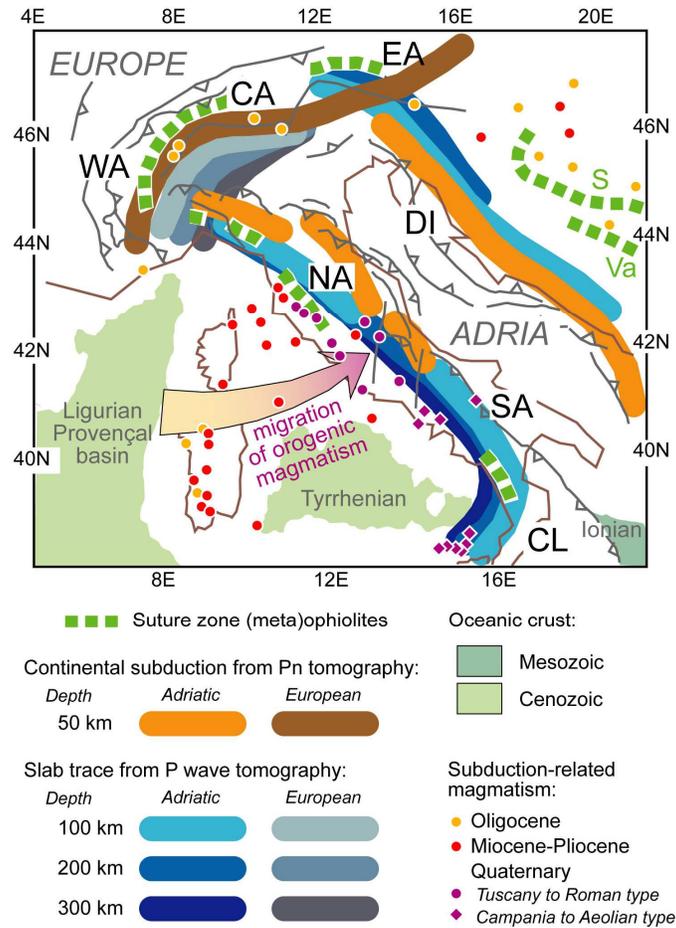


745 **Figure 7.** Horizontal slices of the inverted 3-D P wavespeed at different depths: **(a)**  
746 50 km, **(b)** 60 km and **(c)** 70 km. Gray dashed lines delimit regions with good  
747 resolution (other regions are masked). Major tectonic lines and acronyms as in Figure  
748 1a. Numbers 1 to 6 indicate the main features discussed in the text.



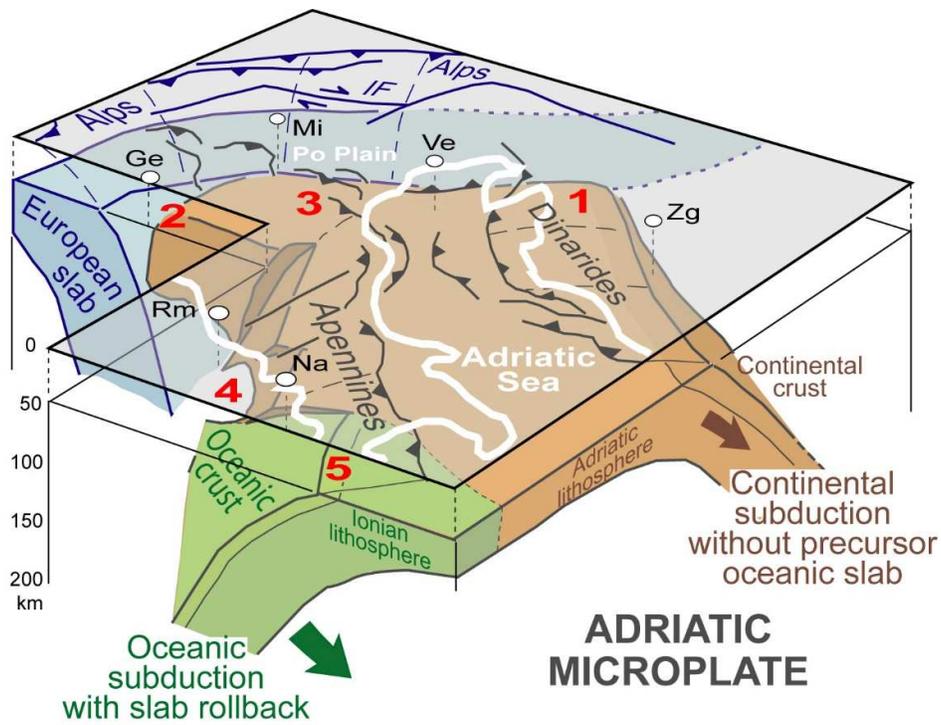
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750 Figure 8. Vertical slices comparing the initial (left) and inverted (right) 3-D P  
 751 wavespeed at different locations ( $\sim 2\times$  vertical exaggeration, see locations in Fig. 3a).  
 752 The thin black line is the Moho as compiled from the literature. Note the opposite  
 753 dipping subductions beneath the Apennines and the Dinarides (A-A'), the along-strike  
 754 northward deepening of low P velocity along the Apennines (B-B'), the E-dipping  
 755 subduction beneath the Western Alps (C-C'), and the S-dipping subduction beneath  
 756 the Central Alps (D-D'). Acronyms: AS, Apeninmic subduction; CS, Central Alps  
 757 subduction; DS, Dinaric subduction; IV, Ivrea body; WS, Western Alps subduction.



758

759 **Figure 9.** Relations between slab structure, distribution of orogenic magmatism (after  
 760 Kovács et al., 2007; Carminati and Doglioni, 2012) and accreted (meta)ophiolites  
 761 (after Bigi et al., 1990; Ustaszewski et al., 2008) around the Adriatic microplate. The  
 762 present-day slab traces at 100, 200 and 300 km depth are based on P wave  
 763 tomography models of Zhao et al. (2016) and Piromallo and Morelli (2003) (which  
 764 was only considered outside of the well-resolved areas of the Zhao et al. (2016)'s  
 765 model). Large arrow = migration of orogenic magmatism during Apenninic slab  
 766 rollback. Acronyms: CA, Central Alps; CL, Calabria; DI, Dinarides; EA, Eastern  
 767 Alps; NA, Northern Apennines; S, Sava zone; SA, Southern Apennines; Va, Vardar  
 768 ophiolites; WA, Western Alps.



769

770 Figure 10. Cartoon summarizing the main features of the 3-D Pn tomography model  
 771 of the Adriatic microplate. Continental subduction follows oceanic subduction in the  
 772 Northern Apennines, whereas it takes place without a precursor oceanic slab beneath  
 773 the Dinarides. In the lack of slab pull, the driving force to allow continental  
 774 subduction beneath the Dinarides is likely provided by the northward motion of  
 775 Africa. Acronyms as in Fig. 1a; numbers in red indicate the main features discussed in  
 776 the text.