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1 **Evidence for a serpentized plate interface favouring**
2 **continental subduction**

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23 **Abstract**

24 The dynamics of continental subduction is largely controlled by the rheological properties
25 of rocks involved along the subduction channel. Serpentinites have low viscosity at
26 geological strain rates. However, compelling geophysical evidence of a serpentinite
27 channel during continental subduction is still lacking. Here we show that anomalously low
28 shear-wave seismic velocities are found beneath the Western Alps, along the plate
29 interface between the European slab and the overlying Adriatic mantle. We propose that
30 these seismic velocities indicate the stacked remnants of a weak fossilised serpentinite
31 channel, which includes both slivers of abyssal serpentinite formed at the ocean floor and
32 mantle-wedge serpentinite formed by fluid release from the subducting slab. Our results
33 suggest that this serpentinitized plate interface may have favoured the subduction of
34 continental crust into the upper mantle and the formation/exhumation of ultra-high
35 pressure metamorphic rocks, providing new constraints to develop the conceptual and
36 quantitative understanding of continental-subduction dynamics.

37 **Introduction**

38 Lithosphere rheology is a key parameter to understand the dynamics of continental
39 subduction following the closure of paleo-oceans during convergence between
40 tectonic plates. Serpentinites are commonly inferred in many subduction zones¹⁻³.
41 They are the principal water carrier towards the mantle, with effective viscosity
42 several orders of magnitude lower than that of the dry upper mantle at geological
43 strain rates⁴. Due to these unique properties, serpentinites are expected to play a major

44 role in controlling the bulk rheology of subduction zones⁵.

45 The Western Alps is one of the best preserved subduction wedges in the world^{6,7}
46 (Fig. 1). They are the result of Cretaceous-to-Paleogene subduction of the Tethys
47 Ocean and adjoining European continental margin beneath the Adriatic microplate⁸.
48 The Alpine subduction wedge includes abundant high pressure serpentinites⁹⁻¹¹ mainly
49 exposed above (ultra)high pressure (UHP) gneissic domes characterized by mineral
50 assemblages attesting to continental subduction^{12,13} (Dora Maira, Gran Paradiso and
51 Monte Rosa in Fig. 1b). The deep configuration of this subduction wedge is
52 increasingly well documented by geophysical data¹⁴⁻²², but a high-resolution image of
53 seismic velocity structure for the deepest levels of the plate interface is still missing.
54 As a consequence, potential rheology variations within the subduction channel remain
55 largely unconstrained, which precludes a full understanding of the dynamics of crucial
56 stages of continental subduction and exhumation of UHP rocks.

57 Laboratory experiments show that serpentinization significantly reduces seismic
58 velocity and density of the precursor peridotite^{23,24}, thus providing a viable approach
59 for probing the distribution of serpentinites within a subduction zone. However, based
60 on numerical models, the sustainable thickness of a serpentinized channel may be on
61 the order of few kilometers only²⁵, i.e., close to the limit of seismic detectability²⁶.
62 This implies that seismic probing requires cutting-edge imaging techniques.

63 Here, we apply a Bayesian transdimensional (TransD) inversion technique²⁷⁻²⁹ to
64 invert Rayleigh wave group velocity dispersion data (Supplementary Information) for

65 a shallow upper mantle shear-wave velocity model in the Western Alps. The inversion
66 treats the model parameterization (e.g., number of velocity layers) as an unknown in
67 the inversion to be determined by data. This avoids regularizing the inverse problem
68 and yields more robust velocity amplitude estimates²⁷. The tomography is conducted
69 from dispersion data¹⁹ measured in the area of the first densely-spaced array of
70 broadband seismic stations across the Western Alps (Fig. 1)¹⁵ (see Methods and
71 Supplementary Note 1). This approach provides absolute shear wave velocity (V_s)
72 information (Supplementary Information) and an unprecedented high-resolution
73 image of the velocity structure at the plate interface of a continental subduction zone.

74 **Results and discussions**

75 **Geophysical evidence of a serpentized subduction channel**

76 Figure 2a shows a cross section of the inverted velocity model across the southern
77 Western Alps down to 90 km depth, and the main tectonic features, indicated by black
78 lines, based on available geologic constraints^{30, 14, 17}. At shallow depth, the inverted
79 velocity model shows low V_s regions corresponding to the main sedimentary basins
80 formed on European and Adriatic crusts (Southeast basin and Po Plain, respectively).
81 The European lithospheric mantle (in blue in Fig. 2) is clearly underthrust beneath
82 the Adriatic mantle imaged on the eastern part of the profile. In general, the
83 subduction wedge beneath the Western Alps exhibits a normal velocity structure with
84 V_s progressively increasing with depth, from $\sim 3.3 \text{ km s}^{-1}$ in the uppermost levels, to
85 $\sim 3.6 \text{ km s}^{-1}$ (wheat colours) at $\sim 40 \text{ km}$ depth (profile 3 in Fig. 2a). On the Adriatic

86 side of the cross section (around +60 km distance, between profiles 4 and 5) the
87 culmination of the region with $V_s > 4.0 \text{ km s}^{-1}$ at ~10-15 km depth matches with the
88 Ivrea gravity anomaly (Fig. 1b; see also Supplementary Fig. 8), and with the
89 high-velocity regions beneath the Dora-Maira dome independently observed in
90 receiver function¹⁴ and local earthquake tomography models²¹ (see also
91 Supplementary Fig.15 and 17).

92 The most intriguing feature of the inverted model is the low-velocity region with
93 $V_s \leq 3.7 \text{ km s}^{-1}$ imaged between 50 km and 70 km depth and around +75 km distance
94 (profile 5, green star in Fig. 2a). This low-velocity region, squeezed between the
95 European lithospheric mantle and the overlying Adriatic mantle, may mark the fossil
96 subduction channel beneath the Western Alps. Along this channel, V_s initially
97 increases from $\sim 3.6 \text{ km s}^{-1}$ at ~45 km depth (region a in Fig. 2a) to $\sim 3.9 \text{ km s}^{-1}$ at ~55
98 km depth (region b in Fig. 2a), and then decreases again to $\sim 3.6 \text{ km s}^{-1}$ down to ~70
99 km depth (region c in Fig. 2a). The distinctive velocity structure of region c is even
100 more evident in the depth-velocity profiles reported in Fig. 2b, where the blue lines
101 and associated grey bands indicate the velocity and associated 1σ error after the
102 TransD inversion. All of these profiles show crustal velocities increasing with depth,
103 apart from profile 5 that shows a major velocity drop at ~45 km depth. This “inverted”
104 crustal velocity structure agrees well with previous receiver function results showing
105 a negative velocity gradient beneath the Dora Maira¹⁴ (see also Supplementary
106 Fig.17).

107 The lateral extent of the deep low-velocity region can be evaluated in the horizontal
108 slices of Fig. 1c,d. As shown in Fig. 1d, the region with $V_s < 3.7 \text{ km s}^{-1}$ is observed, at
109 60 km depth, all along the arc of the Western Alps, from the Dora Maira to the Monte
110 Rosa domes (DM and MR, respectively, in Fig. 1b). This attests to a clear linkage
111 between the Eocene eclogite belt, which is part of the Alpine subduction wedge, and
112 the deep low-velocity region highlighted by TransD images. Similar to Figure 2, two
113 additional cross-sections are presented in Supplementary Fig. 10 and 11.

114 The geologic interpretation of the peculiar velocity structure beneath the Alpine
115 wedge requires independent knowledge of the V_s ranges for the different rock types
116 likely involved in the Alpine subduction zone. These V_s ranges are provided by a
117 number of laboratory experiments^{23, 31-33} that are summarized in Fig. 3.

118 Based on the composition of the oceanic slivers now exposed in the Alpine belt,
119 various proportions of serpentinitised mantle, basalts and sedimentary rocks were
120 subducted at the Alpine trench since the Cretaceous, during the oceanic subduction
121 stage. Subsequent Paleogene continental subduction additionally involved granitic
122 and (meta)sedimentary rocks from the hyperextended European margin. All of these
123 rocks display a progressive increase in V_s with depth and metamorphic grade, which
124 is particularly evident either in pelitic, granitic or mafic rocks (Fig. 3). For a given
125 rock type, the effects of increasing pressure with depth, which would imply an
126 increase in V_s , are partly compensated by the effects of increasing temperature that
127 would imply a V_s decrease. As a result, major changes in V_s with depth are mainly

128 due to changes in metamorphic mineral assemblages.

129 In metapelites, V_s values provided by laboratory experiments progressively
130 increase from 3.3-3.8 km s⁻¹ in low-grade metapelites³⁴ to 3.8-4.1 km s⁻¹ in high-grade
131 metapelites³². In mafic rocks, V_s values progressively increase from 3.9-4.0 km s⁻¹ in
132 gabbros to 4.3-4.5 km s⁻¹ in blueschists to 4.6-5.0 km s⁻¹ in mafic eclogites³³. V_s
133 values of 3.5-3.7 km s⁻¹ measured in granites³⁵ are expected to increase up to 4.0-4.3
134 km s⁻¹ in quartzofeldspathic gneisses under eclogite facies conditions³⁶. In fully
135 hydrated ultramafic rocks, measured V_s values increase from 2.4-3.0 km s⁻¹ in
136 low-temperature lizardite serpentinite³⁷ to 3.4-3.8 km s⁻¹ in high-temperature
137 antigorite serpentinite^{33, 37}, but V_s as low as 3.2 km s⁻¹ could be observed in antigorite
138 mylonites due to the platy shape of antigorite crystals and the related strong fabric³⁸.
139 A sharp change in V_s , towards values of 4.4-5.1 km s⁻¹ that are typical of peridotite, is
140 however expected across the antigorite-out curve^{32, 33, 39}, with V_s linearly increasing
141 for decreasing volume fraction of antigorite^{37, 40} (Fig. 3).

142 The normal velocity structure of the Alpine subduction wedge down to 50 km is
143 fully consistent with the V_s values, progressively increasing with depth, measured in
144 metapelites and gneisses that are exposed in UHP gneissic domes at the Earth surface.
145 By contrast, the progressive V_s decrease from region b to region c within the
146 subduction channel (Fig. 2a) cannot be explained by metamorphic phase changes in
147 metapelites and gneisses. The only rock characterized by V_s values of ~3.6 km s⁻¹ in
148 the 55-70 km depth range is serpentinite. We propose that Fig. 2a may provide

149 evidence of a weak serpentinite channel preserved in the 55-70 km depth range along
150 the plate interface (Methods). Alternatively, the low V_s values in region c may be due
151 to high pore-fluid pressures, which would in turn promote ongoing serpentinitization.
152 Serpentinitization favours aseismic creep⁴, in line with the aseismic character of the
153 deep-low velocity body (Fig. 2a). At greater depth, the sharp increase in V_s along the
154 subduction channel attests to the destabilization of antigorite and transformation of
155 serpentinite into peridotite.

156 **Progressive development of the serpentinitized plate interface**

157 Serpentinite preserved along the continental subduction channel may have formed
158 by three different processes in different settings (Fig. 4), including hydration of
159 oceanic peridotite by seafloor hydrothermal activity/seawater alteration (abyssal
160 serpentinite), fluids released from the subducted slab to the overlying mantle wedge
161 (mantle wedge serpentinite) and fluids circulating and affecting ultramafic rocks
162 within the subduction wedge (subducted serpentinite)⁴¹.

163 Abyssal serpentinite formed when the Tethys ocean was still open, whereas mantle
164 wedge serpentinites formed by fluid release during oceanic subduction (Fig. 4a). In
165 oceanic domains characterized by slow spreading rates as in the Western Alps,
166 serpentinitization may have occurred over a thickness of 2-4 km reaching extents of
167 70-80%⁴². During subsequent continental subduction, these abyssal serpentinites were
168 either stacked in the subduction wedge, or dragged towards the subduction channel
169 together with basalts and sediments of the Tethys seafloor and rocks of the European

170 paleomargin (Fig. 4b). An example of such stacking is provided by the 10 km-thick
171 Viso unit (Vi in Fig. 1), which includes different slivers of serpentinites with
172 eclogites⁶. During subduction, meta-ophiolites experienced breakdown of amphibole
173 and zoisite, and partial dehydration of antigorite that was likely completed at ~180 km
174 depth along the cold alpine geothermal gradient ($< 7^{\circ}\text{C km}^{-1}$)^{6,43,44}. Continental rocks
175 experienced progressive dehydration breakdown of zoisite, chlorite, biotite, talc and
176 lawsonite, but the low geothermal gradient of the Alpine subduction zone prevented
177 the breakdown of phengite in UHP continental rocks. Dehydration of metasediments
178 and mafic rocks was likely completed at 200-250 km⁴⁴. Substantial amounts of
179 aqueous fluids were thus released at sub-arc depths (80 to 180 km), promoting further
180 serpentinization both in the subduction wedge and along the plate interface above the
181 subduction channel. Minor amounts of fluids are possibly released from the slab even
182 today, despite subduction is no longer active, thus promoting further serpentinization
183 and lowering seismic velocities⁴⁴.

184 We propose that the resulting serpentinized plate interface made of subducted
185 serpentinites and mantle wedge serpentinites may have controlled the bulk rheology
186 of the Alpine subduction zone during the Cenozoic, facilitating the subduction of
187 European continental lithosphere into the Adriatic upper mantle, and enhanced the
188 formation of UHP metamorphic rocks that are now exposed in the Dora-Maira dome.
189 During fast exhumation of the Dora-Maira rocks in the Late Eocene, which was
190 possibly triggered by a component of divergent motion of the upper plate^{7, 45, 46},

191 subducted abyssal serpentinites and the adjoining mantle wedge serpentinites raised
192 along the subduction channel (Fig. 4c). Numerical models ⁴⁶ suggest that the
193 subduction channel may become much thicker after upper-plate divergent motion
194 compared to the previous stages of plate convergence, and may include slivers of
195 serpentinite formed in different settings. Moreover, the subduction channel is
196 squeezed between the two plates during the transition from continental subduction to
197 continental collision favoring its thickening. Numerical models also predict that the
198 transient uplift of isothermal surfaces during fast exhumation is rapidly followed by
199 thermal relaxation. In the Western Alps, this likely allowed the preservation of a thick
200 composite volume of partly serpentinitised mantle rocks down to depths of ~70 km that
201 can be now detected by cutting edge tomographic methods. However, the precise
202 values of thickness and amount of serpentinitisation of this composite rock volume
203 remain uncertain due to the tradeoff between them (Methods). Our observations may
204 provide an ideal example for developing conceptual frameworks of continental
205 subduction, and then have the potential of rendering more accurate dynamic modeling
206 and comparisons with multidisciplinary evidence.

207

208 **Methods**

209 In this work, we applied a Bayesian transdimensional (TransD) tomography
210 technique^{27, 28, 47} to invert a high-quality Rayleigh wave dispersion dataset¹⁹ for crustal
211 and upper mantle shear-wave velocity structure in the Western Alps. The TransD

212 inversion treats the number of velocity layers as an unknown parameter which is
213 determined in the inversion⁴⁸. This avoids subjective choice in the inversion such as
214 manual tuning of damping and regularization parameters. The technique is capable of
215 retrieving high resolution velocity models of crust and uppermost mantle^{29, 49, 50}.

216 **Dispersion dataset and transdimensional inversion**

217 The dispersion dataset¹⁹ takes advantage of the large number of seismic stations in
218 Europe, in particular in the great Alpine region with the high-density AlpArray and
219 CIFALPS temporary experiments. This dataset provides the highest possible data
220 coverage (0.15° grid spacing) in our study area. At each geographical location, we
221 carried out a 1D inversion of the local dispersion curve, and inverted for Vs
222 perturbations around a reference model. The reference model was set to a
223 homogeneous half-space with no discontinuities, and the a priori distribution for
224 velocity variations was set to a uniform distribution (3.8 km s⁻¹) with relatively wide
225 range (±50%). The latter is the major difference compared with the Bayesian
226 inversion¹⁹ which imposed a monotonous velocity increase with depth, and therefore
227 could not detect the low velocity zones featured in the present study. Finally, the set of
228 1D models computed at each location are combined to produce a laterally varying 3D
229 velocity model (more details in Supplementary Note 1).

230 We formulated the inverse problem in a Bayesian framework where all information
231 is represented in probabilistic terms. The solution of the problem is the posterior
232 probability distribution which represents the probability density of the model

233 parameters V_s . We used the reversible-jump Monte Carlo algorithm⁵¹ to generate
234 samples from the posterior distribution. At each iteration of the algorithm, a new 1D
235 velocity model is proposed and its estimated dispersion curve is compared to
236 observations. The proposed model is then either accepted or rejected in the ensemble
237 solution. After a large number of iterations, the ensemble of accepted models is used
238 to approximate the posterior distribution. The algorithm prevents over-fitting the data
239 by penalizing over-complex solution models²⁷. Uncertainties in the solution and
240 tradeoffs between model parameters are evaluated using the posterior model variance
241 and covariance. For a detailed description of the model parameterization, the
242 algorithm and tomographic inversion, as well as model robustness and effects of the
243 anisotropic serpentinite layer, we refer the reader to methodology papers^{27, 29} and to
244 the Supplementary Information.

245 **Model robustness**

246 To illustrate the robustness of the inverted shear-wave velocity model, we first
247 compare our inversion results with the V_s model¹⁹ along the three cross-sections
248 (Supplementary Figs. 12 to 14). Both studies use the same dispersion maps for the
249 final 3D velocity model, but our TransD inversion allows negative velocity gradient
250 with respect to depth (i.e., low velocity zones). We compare misfit between the data
251 and predicted dispersions of the two models (Supplementary Figs 7 to 9, panel b and
252 c), and a form of data variance reduction (panel d),

253
$$\left(1 - \frac{\text{variance}(\text{data} - \text{synthetics})}{\text{variance}(\text{data})} \right) \times 100.$$

254 Here synthetics are the predicted group velocities for the two models. Our new
255 model shows in general a better fit over the previous Vs model¹⁹ across the low
256 velocity zone described in the text.

257 We further compare the velocity features with other available seismic models along
258 the CIFALPS and ECORS-CROP⁵⁵ lines (Supplementary Fig. 15 to 17). Our model
259 shows many consistent crustal features, such as the high-velocity Ivrea body and the
260 low-velocity basins, that are spatially comparable with higher resolution models
261 (Supplementary Fig.15 and 16).

262 In Supplementary Fig.17, we compare the receiver function
263 common-conversion-point (CCP) stacked model¹⁴ along the CIFALPS line with the
264 depth gradient of the inverted velocity model. The depth gradient was computed with
265 0.25 km vertical spacing. This comparison shows that both the CCP section and the
266 Vs model detect the Moho of the subducted European lithosphere, and most
267 interestingly, the top boundary of the low velocity zone. Although the amplitude of
268 P-to-S converted phases at velocity boundaries mapped with the CCP section cannot
269 be simply explained by vertical velocity gradients, the correspondence between
270 Supplementary Fig.17a and 17b in the distance range 50-80 km is striking. In
271 Supplementary Fig.17a, the deep Ps phase at 70-75 km depth, 60-80 km distance that
272 was interpreted as the deepest trace of the European Moho ever imaged in the Western
273 Alps¹⁴ well coincides with the strong positive Vs gradient at the same location in

274 Supplementary Fig.17b. A similar coincidence is also striking on top of the LVZ at
275 45-50 km depth (same 60-80 km distance range) with an "inverted" Moho of negative
276 Ps amplitude in Supplementary Fig. 17a and a negative velocity gradient in
277 Supplementary Fig.17b. However, the strong negative Ps phase at 20-35 km depth and
278 25-70 km distance in Supplementary Fig.17a is not that well explained by a strong
279 negative velocity gradient in Supplementary Fig.17b, showing the limits of this
280 comparison. Nevertheless, we think that the discovery of the low-velocity deep body
281 at the subduction interface from the TransD inversion of group velocity dispersion
282 data, which compares well with the independent CCP image may be attributed to the
283 high resolution of the dispersion dataset associated with the high station density along
284 the CIFALPS line.

285 **Effects of slow-velocity layer and seismic anisotropy**

286 We address here the question of the effective thickness of the slow-velocity
287 serpentinite layer recoverable by our dispersion dataset. We also test the influence of
288 the anisotropy of the serpentinite layer in our isotropic two-step tomographic
289 inversion.

290 **Recovery of Layer thickness.** We constructed simple 1D layered models that include
291 a slow velocity layer whose top is at 40km depth with 10, 15, 20, 25 and 30 km layer
292 thickness, respectively. Synthetic group velocities were computed and randomly
293 distributed Gaussian noise with $\sim 100 \text{ m s}^{-1}$ standard deviation was added. We then
294 carried out the transdimensional inversion with the same uniform prior model used for

295 the real data inversions. The modeling results are presented in Supplementary Fig.18.
296 The inversion tests show that the inversions can recognize a slow velocity input layer,
297 but the amplitude recovery is not satisfactory for layer thicknesses smaller than 15km.
298 The exact depth of the velocity discontinuities is less well recovered because surface
299 wave dispersion data are not very sensitive to velocity gradient. It has been shown
300 that by incorporating a smooth 1D background model with discontinuity depth
301 inferred from receiver functions, the transdimensional inversion may better recover
302 the velocity jump at a discontinuity²⁹. The synthetic tests shown here, however, did
303 not use the this approach²⁹ to be consistent with the inversion of real data used in this
304 manuscript.

305 **Effects of an anisotropic layer on the isotropic inversion.** Serpentinites are highly
306 anisotropic rocks with a slow velocity axis of hexagonal symmetry⁵⁶. Our two-step
307 tomographic inversion, i.e. the group velocity map inversion¹⁹ and the
308 transdimensional inversion conducted here, is isotropic only. It is well known that the
309 presence of seismic anisotropic may cause artifacts in tomographic inversions⁵⁷.
310 However, the evaluation of the full 3D effects of such an anisotropic layer in noise
311 tomographic inversion may require numerical simulations of noise correlations, which
312 is beyond the scope of this study.

313 Here we take a simple approach by inverting back-azimuthally dependent synthetic
314 dispersion data for isotropic 1D velocity profile to assess whether imaging artifacts
315 may be raised by our isotropic tomographic approach. We assume that: anisotropy in

316 the serpentinite layer may be simulated by a slow velocity with hexagonal symmetry;
317 the effect of anisotropy in the dipping layer may be approximated by that of a
318 horizontal layer with dipping symmetry axis; and the anisotropic signal used for the
319 transdimensional inversion is effectively smoothed by the group velocity inversion,
320 especially in the western Alps where the azimuthal distribution of interstation paths is
321 optimum thanks to the array geometry.

322 We used the anisotropic reflectivity package ANIPROP⁵⁸ to compute the group
323 velocities at the same periods as in the real data. To maximize the anisotropy effects,
324 we took the 30 km low-velocity layer model of the previous thickness tests. The
325 strength of anisotropy is fixed at -8%^{56,58}. We computed the dispersion curves from 0
326 to 350° back-azimuths with 10° increments; then we averaged to obtain the effective
327 group velocity dispersion curve and inverted it to obtain the 1D velocity profile. This
328 procedure assumes that the isotropic ambient noise group velocity inversion has
329 averaged the anisotropic signal over the range of back-azimuths. The inverted velocity
330 model is then compared with the input model for the anisotropic effects.

331 We tested two models including a slow velocity layer with horizontal and 45° tilted
332 symmetry axis; for comparison we also included the fast velocity symmetry axis cases.
333 As shown in Supplementary Fig.19, individual Rayleigh wave dispersion curves
334 respond to different incident back-azimuths in both the fast and slow velocity
335 horizontal symmetry axis cases. Their average is very close to the dispersion
336 computed from the isotropic velocity model with the same 30-km slow-velocity

337 thickness. The inverted velocity models well recover the slow velocity of the input
338 model. The horizontal fast symmetry axis model slightly pushes down the bottom of
339 the layer velocity layer. The effect of anisotropy is maximum in the tilted symmetry
340 axis examples. In the case of a tilted slow velocity symmetry axis, the recovered
341 isotropic slow velocity is overestimated ($\sim 0.1 \text{ km s}^{-1}$) while the recovered layer
342 thickness is larger ($\sim 5 \text{ km}$) than in the average case. For the tilted fast velocity
343 symmetry axis case, the recovered amplitude of the layer is underestimated, and the
344 recovered layer is slightly thinner than in the averaged case.

345 In conclusion, in the presence of a thick tilted serpentinite layer, anisotropy in the
346 dispersion signal may cause the isotropic inversion to over-estimate both the
347 slow-velocity amplitude and the layer thickness. Considering the results from
348 thickness tests we suggest that our isotropic inversions may be capable of revealing a
349 20- to 25-km thick slow-velocity (~ 3.7 to 3.8 km s^{-1}) layer, however, the amplitude of
350 the slow-velocity layer remains uncertain.

351 **Velocity of the serpentinite layer.** To further illustrate the robustness of the
352 velocity estimate of the serpentinite layer in the preferred model, we operated 2 more
353 tests using: a model with larger variations in the lower velocity channel, and a
354 homogenous model of 4.0 km s^{-1} from 10km downwards (Supplementary Fig.20).
355 These are compared with the final model at Location 5 along the CIFALPS line. As
356 shown in Supplementary Figure 20a, both test models lie pretty much within the 1σ
357 model uncertain area, and the predicted dispersion curves are close to that of the final

358 model (Supplementary Fig. 20b). However, the misfit distribution of these two test
359 models is not centered, and the 4 km s^{-1} homogenous model even shows a bi-modal
360 distribution. This implies that these two test models provide a poorer fit to the data
361 than the final model.

362 In addition, the error at a given depth is not independent from the error at another
363 depth. For example, an increase of V_s at 40 km needs a decrease of V_s at 65 km, and
364 vice versa. Our previous receiver function results¹⁴ indicate a strong upward velocity
365 increase attested by a strong negative energy of P-to-S conversion (Supplementary
366 Fig.17). Therefore, we can preclude an increase of V_s at 65 km, otherwise a decrease
367 of V_s at 40 km would give a smooth discontinuity. The other possibility, to the
368 extreme, is having $V_s=5 \text{ km s}^{-1}$ at 40 km to get 3.3 km s^{-1} at 65 km to fit the data, but
369 this is very unlikely because even 100% serpentinite cannot satisfy this low velocity
370 value of 3.3 km s^{-1} .

371 **Data availability.** The authors declare that the data supporting the findings of this
372 study are available within the paper and its supplementary information file
373 (Supplementary Data 1).

374 **Code Sharing.** Software package of Bayesian Transdimensional tomography
375 technique^{27, 28, 46} is available, and may be accessible through co-author.

376

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578 **Author contributions**

579 L.Z. conceived this experiment; H.Y., L.Z., A.P., Y.L., L.S. and T.B. analysed the
580 dataset; S.S. and E.E. provided relocated foci distribution; H.Y., L.Z., A.P. and Y.L.
581 plotted the tomography images; M.M., L.Z., S.G. and H.Y. interpreted the results;
582 L.Z., M.M., H.Y., S.G., A.P. and G.L. wrote the manuscript, with the contribution of
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584

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587 **Additional information**

588 Supplementary Information is available for this paper at xxxx

589

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597 **Figure Captions**

598 **Fig. 1** Geologic setting and depth slices of Vs. **a**, Tectonic sketch map of the
599 Adria-Europe plate boundary zone. **b**, Geological sketch map of the western Alpine
600 region (modified from Malusà et al.⁴⁴). The magenta line indicates the location of the
601 cross-section of Fig. 2a. Numbers 1 to 6 indicate the sites of the depth-velocity
602 profiles illustrated in Fig. 2b. The purple line indicates the gravity anomaly of the
603 Ivrea body (0 mGal isoline). Acronyms: Br, Briançonnais; DM, Dora-Maira; FPF,
604 Frontal Pennine Fault; GP, Gran Paradiso; IF, Insubric Fault; La, Lanzo; MR, Monte
605 Rosa; Se, Sesia-Lanzo; SL, Schistes lustrés; Vi, Viso. **c**, **d**, Vs structure at different
606 depths (40 km and 60 km) after TransD inversion. The green star marks the deep
607 low-velocity body (yellow to wheat area).

608

609 **Fig. 2** Shear wave velocity model across the Western Alps. **a**, Cross section showing
610 the absolute Vs in the 0-90 km depth range (see location in Fig. 1b). The white lines
611 indicate the 3.7 and 3.8 km s⁻¹ contours. The green star marks the deep low-velocity
612 body. Earthquakes foci (depth >30 km) in a 40-km wide swath parallel to the profile
613 are plotted as purple circles. Letters a-to-c indicate regions of the model discussed in
614 the text. Main crustal features (black lines) after Zhao et al.¹⁴; Malusà et al.¹⁷
615 (acronyms as in Fig. 1b). **b**, Depth-velocity profiles ($\pm 1\sigma$) beneath 6 typical sites
616 along the transect (see location in Fig 1b and panel a). The orange vertical line is the
617 3.8 km s⁻¹ isovelocity.

618

619 **Fig. 3** Measured V_s (km s^{-1}) for different rocks (green = ultramafic; blue = mafic;
620 purple = granitic; brown = pelitic) at ambient conditions. The size of the circles is
621 proportional to V_s , while circle location depends on the pressure-temperature range at
622 equilibrium for each rock type^{23, 31-39, 52}. The V_s range of 4.0-4.3 km s^{-1} in
623 quartzofeldspathic gneisses is based on calculations for modeled rocks at 700°C and 2
624 GPa³⁶. The lower bound for HT serpentinite (3.4 km s^{-1}) is calculated from
625 single-crystal elastic data and referred to Reuss approximation and 2 GPa²³. Lizardite
626 (Lz) and antigorite (Atg) stability fields^{53, 54}. The dashed gray lines indicate the
627 metamorphic facies boundaries (Am = Amphibolite; Bs = Blueschist; Ec = Eclogite;
628 Gr = Granulite; Gs = Greenschist). Note the progressive increase in V_s with depth and
629 metamorphic grade in pelitic, granitic and mafic rocks, and the sharp change in V_s
630 across the Atg-out curve in ultramafic rocks. The green star marks the rock types
631 consistent with V_s and depth of the analyzed low-velocity body.

632

633 **Fig. 4** Environments of serpentinization during subduction and geologic interpretation
634 of the deep low-velocity body. **a**, Oceanic subduction. **b**, Continental subduction. **c**,
635 Upper plate divergent motion and exhumation of UHP rocks (crustal features color
636 coded as in Fig. 1b). Lithospheric structure after Zhao et al.^{14, 15}, Lyu et al.¹⁶ and
637 Solarino et al.²¹. Environments of serpentinization and fluid circulation after
638 Deschamps et al.⁴¹. Evolution of lizardite (Lz) and antigorite (Atg) destabilization

639 isotherms after Liao et al.⁴⁶. Percentage serpentinite in the exhumed mantle wedge
640 after Solarino et al.²¹. The deep low-velocity body imaged in the subduction channel
641 consists of abyssal and mantle wedge serpentinites that lubricate the plate interface
642 and facilitate continental subduction.







