Evidence for a serpentinized plate interface favouring continental subduction

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

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Lithosphere rheology is a key parameter to understand the dynamics of continental subduction following the closure of paleo-oceans during convergence between tectonic plates. Serpentinites are commonly inferred in many subduction zones\textsuperscript{1–3}. They are the principal water carrier toward the mantle, with effective viscosity several orders of magnitude lower than that of the dry upper mantle at geological strain rates\textsuperscript{4}. Due to these unique properties, serpentinites are expected to play a major role in controlling the bulk rheology of subduction zones\textsuperscript{5}.

The Western Alps is one of the best-preserved subduction wedges in the world\textsuperscript{6,7} (Fig. 1). They are the result of Cretaceous-to-Paleogene subduction of the Tethys Ocean and adjoining European continental margin beneath the Adriatic microplate\textsuperscript{8}. The Alpine subduction wedge includes abundant high-pressure serpentinites\textsuperscript{9–11} mainly exposed above (ultra) high-pressure (UHP) gneissic domes characterized by mineral assemblages attesting to continental subduction\textsuperscript{12,13} (Dora-Maira, Gran Paradiso, and Monte Rosa in Fig. 1b). The deep configuration of this subduction wedge is increasingly well documented by geophysical data\textsuperscript{14–22}, but a high-resolution image of seismic velocity structure for the deepest levels of the plate interface is still missing. As a consequence, potential rheology variations within the subduction channel remain largely unconstrained, which precludes a full understanding of the dynamics of crucial stages of continental subduction and exhumation of UHP rocks.

![Fig. 1 Geologic setting and depth slices of Vs.](https://example.com/fig1.png)

\textbf{Fig. 1 Geologic setting and depth slices of Vs.} \textbf{a} Tectonic sketch map of the Adria-Europe plate boundary zone. \textbf{b} Geological sketch map of the western Alpine region (modified from Malusà et al.\textsuperscript{45}). The magenta line indicates the location of the cross section of Fig. 2a. Numbers 1–6 indicate the sites of the depth-velocity profiles illustrated in Fig. 2b. The purple line indicates the gravity anomaly of the Ivrea body (0 mGal isoline). Acronyms: Br, Briançonnais; DM, Dora-Maira; FPF, Frontal Pennine Fault; GP, Gran Paradiso; IF, Insubric Fault; La, Lanzo; MR, Monte Rosa; Se, Sesia-Lanzo; SL, Schistes lustrés; Vi, Viso. \textbf{c, d} Vs structure at different depths (40 km and 60 km) after TransD inversion. The green star marks the deep-low-velocity body (yellow to wheat area).
Laboratory experiments show that serpentinization significantly reduces seismic velocity and density of the precursor peridotite\textsuperscript{23,24}, thus providing a viable approach for probing the distribution of serpentinites within a subduction zone. However, based on numerical models, the sustainable thickness of a serpentinized channel may be on the order of a few kilometers only\textsuperscript{25}, i.e., close to the limit of seismic detectability\textsuperscript{26}. This implies that seismic probing requires cutting-edge imaging techniques.

Here, we apply a Bayesian transdimensional (TransD) inversion technique\textsuperscript{27–29} to invert Rayleigh wave group velocity dispersion data (Supplementary Note 1 and Supplementary Figs. 1–7) for a shallow upper mantle shear-wave velocity model in the Western Alps. The inversion treats the model parameterization (e.g., number of velocity layers) as an unknown in the inversion to be determined by data. This avoids regularizing the inverse problem and yields more robust velocity amplitude estimates\textsuperscript{27}. The tomography is conducted from dispersion data\textsuperscript{19} measured in the area of the first densely spaced array of broadband seismic stations across the Western Alps (Fig. 1)\textsuperscript{15} (Methods and Supplementary Note 1). This approach provides absolute shear-wave velocity (Vs) information (Supplementary Figs. 8–14) and an unprecedented high-resolution image of the velocity structure at the plate interface of a continental subduction zone.

**Results and discussion**

**Geophysical evidence of a serpentinized subduction channel.**

Figure 2a shows a cross-section of the inverted velocity model across the southern Western Alps down to 90-km depth, and the main tectonic features, indicated by black lines, based on available geologic constraints\textsuperscript{14,17,30}. At shallow depth, the inverted velocity model shows low Vs regions corresponding to the main sedimentary basins formed on European and Adriatic crusts (Southeast basin and Po Plain, respectively). The European lithospheric mantle (in blue in Fig. 2) is clearly underthrust beneath the Adriatic mantle imaged on the eastern part of the profile. In general, the subduction wedge beneath the Western Alps exhibits a normal velocity structure with Vs progressively increasing with depth, from ~3.3 km s\textsuperscript{−1} in the uppermost levels, to ~3.6 km s\textsuperscript{−1} (wheat colors) at ~40-km depth (profile 3 in Fig. 2a). On the Adriatic side of the cross section (around +60-km distance, between profiles 4 and 5), the culmination of the region with Vs >4.0 km s\textsuperscript{−1} at ~10–15-km depth matches with the Ivrea gravity anomaly (Fig. 1b; see also Supplementary Fig. 8), and with the high-velocity regions beneath the Dora-Maira dome independently observed in receiver function\textsuperscript{15} and local earthquake tomography models\textsuperscript{21} (see also Supplementary Figs. 15–17).

The most intriguing feature of the inverted model is the low-velocity region with Vs \(\leq 3.7\) km s\textsuperscript{−1} imaged between 50- and 70-km depth and around +75-km distance (profile 5, green star in Fig. 2a). This low-velocity region, squeezed between the European lithospheric mantle and the overlying Adriatic mantle, may mark the fossil subduction channel beneath the Western Alps. Along this channel, Vs initially increases from ~3.6 km s\textsuperscript{−1} at ~45-km depth (region a in Fig. 2a) to ~3.9 km s\textsuperscript{−1} at ~55-km depth (region b in Fig. 2a) to ~3.9 km s\textsuperscript{−1} at ~55-km depth (region c in Fig. 2a).
depth (region b in Fig. 2a), and then decreases again to ~3.6 km s\(^{-1}\) down to ~70-km depth (region c in Fig. 2a). The distinctive velocity structure of region c is even more evident in the depth-velocity profiles reported in Fig. 2b, where the blue lines and associated gray bands indicate the velocity and associated 1σ error after the TransD inversion. All of these profiles show crustal velocities increasing with depth, apart from profile 5 that shows a major velocity drop at ~45-km depth. This “inverted” crustal velocity structure agrees well with previous receiver function results showing a negative velocity gradient beneath the Dora-Maira\(^{15}\) (see also Supplementary Fig. 17).

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

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

Based on the composition of the oceanic slivers now exposed in the Alpine belt, various proportions of serpentinized mantle, basalts, and sedimentary rocks were subducted at the Alpine trench since the Cretaceous, during the oceanic subduction stage. Subsequent Paleogene continental subduction additionally involved granitic and (meta)sedimentary rocks from the hyper-extended European margin. All of these rocks display a progressive increase in Vs with depth and metamorphic grade, which is particularly evident in pelitic, granitic, or mafic rocks (Fig. 3). For a given rock type, the effects of increasing pressure with depth, which would imply an increase in Vs, are partly compensated by the effects of increasing temperature that would imply a Vs decrease. As a result, major changes in Vs with depth are mainly due to changes in metamorphic mineral assemblages.

In metametamorphic, Vs values provided by laboratory experiments progressively increase from 3.3 to 3.8 km s\(^{-1}\) in low-grade metabedules\(^{24}\) to 3.8–4.1 km s\(^{-1}\) in high-grade metabedules\(^{35}\). In mafic rocks, Vs values progressively increase from 3.9 to 4.0 km s\(^{-1}\) in gabbros to 4.3–4.5 km s\(^{-1}\) in blueschists to 4.6–5.0 km s\(^{-1}\) in mafic eclogites\(^{33}\). Vs values of 3.5–3.7 km s\(^{-1}\) measured in granites\(^{36}\) are expected to increase up to 4.0–4.3 km s\(^{-1}\) in quartzofeldspathic gneisses under eclogite facies conditions\(^{36}\). In fully hydrated ultrama
crocks, measured Vs values increase from 2.4 to 3.0 km s\(^{-1}\) in low-temperature lizardite serpentinite\(^{37}\) to 3.4–3.8 km s\(^{-1}\) in high-temperature antigorite serpentinite\(^{33,37}\), but Vs as low as 3.2 km s\(^{-1}\) could be observed in antigorite mylonites due to the platy shape of antigorite crystals and the related strong fabric\(^{38}\). A sharp change in Vs, toward values of 4.4–5.1 km s\(^{-1}\) that are typical of peridotite, is however expected across the antigorite-out curve\(^{32,33,39}\), with Vs linearly increasing for decreasing volume fraction of antigorite\(^{37,40}\) (Fig. 3).

The normal velocity structure of the Alpine subduction wedge down to 50 km is fully consistent with the Vs values, progressively increasing with depth, measured in metapelites and gneisses that are exposed in UHP gneissic domes at the Earth surface. By contrast, the progressive Vs decrease from region b to region c within the subduction channel (Fig. 2a) cannot be explained by metamorphic phase changes in metapelites and gneisses. The only rock characterized by Vs values of ~3.6 km s\(^{-1}\) in the 55–70-km depth range is serpentinite. We propose that Fig. 2a may provide evidence of a weak serpentinite channel preserved in the 55–70-km depth range along the plate interface (Methods). Alternatively, the low Vs values in region c may be due to high pore-fluid pressures, which would in turn promote ongoing serpentinization. Serpentinization favors aseismic creep\(^{4}\), in line with the aseismic character of the deep–low-velocity body (Fig. 2a). At greater depth, the sharp increase in Vs along the subduction channel attests to the destabilization of antigorite and transformation of serpentinite into peridotite.

Progressive development of the serpentinized plate interface. Serpentinite preserved along the continental subduction channel may have formed by three different processes in different settings (Fig. 4), including hydration of oceanic peridotite by seafloor hydrothermal activity/seawater alteration (abyssal serpentinite), fluids released from the subducted slab to the overlying mantle wedge (mantle-wedge serpentinite), and fluids circulating and affecting ultrama
crocks within the subduction wedge (subduced serpentinite)\(^{41}\).

Abyssal serpentinite formed when the Tethys ocean was still open, whereas mantle-wedge serpentinites formed by fluid release during oceanic subduction (Fig. 4a). In oceanic domains characterized by slow spreading rates as in the Western Alps, serpentinization may have occurred over a thickness of 2–4 km reaching extents of 70–80%\(^{42}\). During subsequent continental subduction, these abyssal serpentinites were either stacked in the subduction wedge, or dragged toward the subduction channel.
Figure 4 Environments of serpentinization during subduction and geologic interpretation of the deep-low-velocity body. a Oceanic subduction. b Continental subduction. c Upper-plate divergent motion and exhumation of UHP rocks (crustal features color coded as in Fig. 1b). Lithospheric structure after Zhao et al., Lyu et al., and Solarino et al. Environments of serpentinization and fluid circulation after Deschamps et al. Evolution of lizardite (Lz) and antigorite (Atg) destabilization isotherms after Liao et al. Percentage of serpentinite in the exhumed mantle wedge after Solarino et al. The deep-low-velocity body imaged in the subduction channel consists of abyssal and mantle-wedge serpentinites that lubricate the plate interface and facilitate continental subduction.

together with basalts and sediments of the Tethys seafloor and rocks of the European paleomargin (Fig. 4b). An example of such stacking is provided by the 10-km-thick Viso unit (V in Fig. 1), which includes different slivers of serpentinites with eclogites. During subduction, meta-ophiolites experienced breakdown of amphibole and zoisite, and partial dehydration of antigorite that was likely completed at ~180-km depth along the cold Alpine geothermal gradient (~7 °C km⁻¹). Continental rocks experienced progressive dehydration breakdown of zoisite, chlorite, biotite, talc, and lawsonite, but the low geothermal gradient of the Alpine subduction zone prevented the breakdown of phengite in UHP continental rocks. Dehydration of metasediments and mafic rocks was likely completed at 200–250 km. Substantial amounts of aqueous fluids were thus released at subarc depths (80–180 km), promoting further serpentinization both in the subduction wedge and along the plate interface above the subduction channel. Minor amounts of fluids are possibly released from the slab even today, despite the fact that subduction is no longer active, thus promoting further serpentinization and lowering seismic velocities.

We propose that the resulting serpentinized plate interface made of subducted serpentinites and mantle wedge serpentinites may have controlled the bulk rheology of the Alpine subduction zone during the Cenozoic, facilitating the subduction of European continental lithosphere into the Adriatic upper mantle, and enhanced the formation of UHP metamorphic rocks that are now exposed in the Dora-Maira dome. During fast exhumation of the Dora-Maira rocks in the Late Eocene, which was possibly triggered by a component of divergent motion of the upper plate, subducted abyssal serpentinites and the adjoining mantle-wedge serpentinites raised along the subduction channel (Fig. 4c). Numerical models suggest that the subduction channel may become much thicker after upper-plate divergent motion compared with the previous stages of plate convergence, and may include slivers of serpentinite formed in different settings. Moreover, the subduction channel is squeezed between the two plates during the transition from continental subduction to collision favoring its thickening. Numerical models also predict that the transient uplift of isothermal surfaces during fast exhumation is rapidly followed by thermal relaxation. In the Western Alps, this likely allowed the preservation of a thick composite volume of partly serpentinized mantle rocks down to depths of ~70 km that can be now detected by cutting-edge tomographic methods. However, the precise values of thickness and the amount of serpentinization of this composite rock volume remain uncertain due to the trade-off between them (Methods). Our observations may provide an ideal example for developing conceptual frameworks of continental subduction, and then have the potential of rendering more accurate dynamic modeling and comparisons with multidisciplinary evidence.

### Methods

In this work, we apply a Bayesian transdimensional (TransD) tomography technique to invert a high-quality Rayleigh wave dispersion dataset for crustal and upper mantle shear-wave velocity structure in the Western Alps. The TransD inversion treats the number of velocity layers as an unknown parameter that is determined in the inversion. This avoids subjective choice in the inversion, such as manual tuning of damping and regularization parameters. The technique is capable of retrieving high-resolution velocity models of the crust and uppermost mantle.

#### Dispersion dataset and transdimensional inversion

The dispersion dataset takes advantage of the large number of seismic stations in Europe, in particular in the great Alpine region with the high-density AlpArray and CIFALPS temporary experiments. This dataset provides the highest possible data coverage (0.15° grid spacing) in our study area. At each geographical location, we carry out a 1D inversion of the local dispersion curve, and invert for Vs perturbations around a reference model. The reference model is set to a homogeneous half-space with no discontinuities, and the a priori distribution for velocity variations is set to a uniform distribution (3.8 km s⁻¹) with a relatively wide range (±50%). The latter is the major difference compared with the Bayesian inversion that is determined in the inversion. This avoids subjective choice in the inversion, such as manual tuning of damping and regularization parameters. The technique is capable of retrieving high-resolution velocity models of the crust and uppermost mantle.

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inversion, as well as model robustness and effects of the anisotropic serpentinite layer, we refer the reader to methodology papers and to Supplementary Figs. 18–20.

Model robustness. To illustrate the robustness of the inverted shear-wave velocity model, we first compare our inversion results with the Vs model along the three cross sections (Supplementary Figs. 12–14). Both studies use the same dispersion maps for the final 3D velocity model, but our TransD inversion allows negative velocity values to reach the Moho (i.e., low-velocity zones). We compare misfit between the data and predicted dispersions of the two models (Supplementary Figs. 7 to 9, panels b and c), and a form of data variance reduction (panel d):

\[
\frac{1 - \text{variance(data – synthetics)}}{\text{variance(data)}} \times 100.
\]

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

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

In Supplementary Fig. 17, we compare the receiver function common-conversion-point (CCP) stacked model along the CIFALPS line with the depth gradient of the inverted velocity model. The depth gradient was computed with 0.25-km vertical spacing. This comparison shows that both the CCP section and the Vs model detect the Moho of the subducted European lithosphere, and most interestingly, the top boundary of the low-velocity zone. Although the amplitude of P- to S-converted phases at velocity boundaries mapped with the CCP section cannot be simply explained by vertical velocity gradients, the correspondence between Supplementary Figs. 17a and b in the distance range 50–80 km is striking. In Supplementary Fig. 17a, the deep Ps phase at 70–75-km depth, 60–80-km distance that was interpreted as the deepest trace of the European Moho ever imaged in the Western Alps, well coincides with the strong positive Vs gradient at the same location in Supplementary Fig. 17b. A similar coincidence is also striking on top of the ZAZ 45–50-km depth (the same 60–80-km distance range) with an “inverted” Moho of negative Ps amplitude in Supplementary Fig. 17a and a negative velocity gradient in Supplementary Fig. 17b. However, the strong negative Ps phase at 20–35-km depth and 25–70-km distance in Supplementary Fig. 17a is not well explained by a strong negative velocity gradient in Supplementary Fig. 17b, showing the limits of this comparison. Nevertheless, we think that the discovery of the low-velocity deep body at the subduction interface from the TransD inversion of group velocity dispersion data, which compares well with the independent CCP image may be attributed to the high resolution of the dispersion dataset associated with the high station density along the CIFALPS line.

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

Recovery of layer thickness: We construct simple 1D layered models that include a slow-velocity layer whose top is at 40-km depth with 10°, 15°, 20°, and 30-km layer thickness, respectively. Synthetic group velocities are computed, and randomly distributed Gaussian noise with ~100 m s\(^{-1}\) standard deviation is added. We then carry out the transdimensional inversion with the same uniform prior model used for the real data inversions. The modeling results are presented in Supplementary Fig. 18. The inversion tests show that the inversions can recognize a slow-velocity input layer, but the amplitude recovery is not satisfactory for layer thicknesses smaller than 15 km. The exact depth of the velocity discontinuities is less well recovered because surface wave dispersion data are not very sensitive to velocity gradient. It has been shown that by incorporating a smooth 1D velocity gradient with respect to depth (i.e., low-velocity zones), it is possible to imaged in the Western Alps, well coincides with the strong positive Vs gradient at the same location in Supplementary Fig. 17b. A similar coincidence is also striking on top of the ZAZ 45–50-km depth (the same 60–80-km distance range) with an “inverted” Moho of negative Ps amplitude in Supplementary Fig. 17a and a negative velocity gradient in Supplementary Fig. 17b. However, the strong negative Ps phase at 20–35-km depth and 25–70-km distance in Supplementary Fig. 17a is not well explained by a strong negative velocity gradient in Supplementary Fig. 17b, showing the limits of this comparison. Nevertheless, we think that the discovery of the low-velocity deep body at the subduction interface from the TransD inversion of group velocity dispersion data, which compares well with the independent CCP image may be attributed to the high resolution of the dispersion dataset associated with the high station density along the CIFALPS line.

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Here we take a simple approach by inverting back-azimuthal data sets with the assumption of hexagonal anisotropy. With this assumption, we test the effect of anisotropy in the dipping layer: a) the anisotropic effect is approximated by that of a horizontal layer with dipping symmetry axis; the effect of anisotropy in the dipping layer may be approximated by that of a horizontal layer with dipping symmetry axis examples. In the case of a tilted slow-velocity symmetry axis, the inverted velocity model is then compared with the input model for the anisotropic case. We test two models, including a slow-velocity layer with horizontal and 45° tilted symmetry axis; for comparison, we also include the fast-velocity symmetry axis cases. As shown in Supplementary Fig. 19, individual Rayleigh wave dispersion curves respond to different incident back-azimuths in both the fast- and slow-velocity symmetry axis cases. Their average is very close to the dispersion computed from the isotropic velocity model with the same 30-km slow-velocity thickness. The inverted velocity models well recover the slow velocity of the input model. The horizontal fast-velocity symmetry axis model slightly pushes down the bottom of the velocity layer. The effect of anisotropy is maximum in the tilted symmetry axis examples. In the case of a tilted slow-velocity symmetry axis, the recovered isotropic slow velocity is underestimated (~0.1 km s\(^{-1}\)), while the recovered layer thickness is larger (~5 km) than in the average case. For the tilted fast-velocity symmetry axis case, the recovered amplitude of the layer is underestimated, and the recovered layer is slightly thinner than in the averaged case. In conclusion, in the presence of a thick tilted serpentinite layer, anisotropy in the dispersion signal may cause the isotropic inversion to overestimate both the slow-velocity amplitude and the layer thickness. Considering the results from thickness tests, we suggest that our isotropic inversions may be capable of revealing a 10-km thick slow-velocity (~3.7–3.8 km s\(^{-1}\)) layer, however, the amplitude of the slow-velocity layer remains uncertain.

Velocity of the serpentinite layer: To further illustrate the robustness of the velocity estimate of the serpentinite layer in the preferred model, we operate two more tests using a model with larger variations in the lower-velocity channel, and a homogenous model of 4.0 km s\(^{-1}\) from 10 km downward (Supplementary Fig. 20). These are compared with the final model at Location 5 along the CIFALPS line. As shown in Supplementary Fig. 20a, both test models lie pretty much within the 1σ uncertainty area, and the predicted dispersion curves are close to those of the final model (Supplementary Fig. 20b). However, the misfit distribution of these two test models is not centered, and the 4 km s\(^{-1}\) homogenous model even shows a bimodal distribution. This implies that these two test models provide a poorer fit to the data than the final model.

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

Data availability

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

Code sharing

Software package of Bayesian Transdimensional tomography technique is available, and may be accessible through coauthor (T.B., email: Thomas.bodin@ens-lyon.fr or H.Y., email: huaiyu.yuan@mq.edu.au).

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Author contributions

L.Z. conceived this experiment; H.Y., L.Z., A.P., Y.L., L.S., and T.B. analyzed the dataset; S.S. and E.E. provided relocated foci distribution; H.Y., L.Z., A.P., and Y.L. plotted the tomography images; M.M., L.Z., S.G., and H.Y. interpreted the results; L.Z., M.M., H.Y., S.G., A.P., and G.L. wrote the paper, with the contribution of other authors.

Competing interests

The authors declare no competing interests.

Additional information

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