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Evidence of oceanic plate delamination in the Northern Atlantic

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The Earth's surface is constantly being recycled by plate tectonics. Subduction of oceanic lithosphere and delamination of continental lithosphere constitute the two most important mechanisms by which the Earth's lithosphere is recycled into the mantle¹⁻⁵. Delamination or detachment in continental regions typically occurs below mountain belts due to a weight excess of overthickened lithospheric mantle, which detaches from overlying lighter crust, aided by the existence of weak layers within the continental lithosphere⁵. Oceanic lithosphere is classically pictured as a rigid plate with a strong core that does not allow for delamination to occur⁶. Here, we propose that active delamination of oceanic lithosphere occurs offshore Southwest Iberia. The process is assisted by the existence of a lithospheric serpentized layer that allows the lower part of the lithosphere to decouple from the overlying crust. Tomography images reveal a sub-lithospheric high-velocity anomaly below this region, which we interpret as a delaminating block of old oceanic lithosphere. We present numerical models showing that for a geological setting mimicking offshore Southwest Iberia delamination of oceanic lithosphere is possible and may herald subduction initiation, which is a long-unsolved problem in the theory of plate tectonics⁷. We further propose that such oceanic delamination is responsible for the highest-magnitude earthquakes in Europe, including the M8.5-8.7 Great Lisbon Earthquake of 1755 and the M7.9 San Vicente earthquake of 1969⁸. In particular, our numerical models, in combination with calculations on seismic potential, provide a solution for the instrumentally recorded 1969 event below the flat Horseshoe abyssal plain, away from mapped tectonics faults⁹. Delamination of old oceanic lithosphere near passive margins constitutes a new class of subduction initiation mechanisms, with fundamental implications for the dynamics of the Wilson cycle¹⁰⁻¹¹.

The Southwest Iberian Margin lies along the complex Azores-Gibraltar plate boundary zone¹² (Fig. 1). The region is characterized by WNW-ESE oblique convergence of ~ 4 mm/yr between Africa and Eurasia¹³. The resulting deformation is accommodated on the seafloor by a set of visible WNW-ESE strike-slip faults and NE-SW thrusts, which control the uplift of seamounts that separate abyssal plains¹⁴. The Southwest Iberian Margin has unleashed the highest-magnitude earthquakes in Europe⁹. These include the M8.5-8.7 Great Lisbon Earthquake of 1755, whose ground-motion, tsunami and fires caused widespread destruction in Portugal and killed tens of thousands^{15,16}, the M7.5-8.5 earthquakes of 1356 and 1761¹⁷, and the more recent M7.9 earthquake of 1969, located ~ 200 km SW off Cape St. Vincent⁹ (Fig. 1).

Figure 1 – Eastern segment of the Azores-Gibraltar plate boundary that separates the Eurasian and the African plates. Here it is possible to recognize two main fault zones: the Gloria Fault in the north, which is generally considered the main trace of the plate boundary, and a seemingly less active fracture to the south, here named Gloria South Fracture. Both these structures have experienced high-magnitude earthquakes. To the east, their linear traces progressively give place to major thrust faults that elevate the seafloor forming two major bathymetric highs, the Coral Patch Seamount and the Gorringer Bank, that delimit the Horseshoe Abyssal Plain. This region has unleashed some of the major historical and instrumental earthquakes in the Atlantic region^{8,9,12,15,17}.

The 1356, 1755 and 1761 earthquakes are poorly understood given their offshore location and occurrence within the historical period⁸. However, the more recent M7.9 1969 earthquake has been instrumentally recorded. Remarkably, it occurred below the flat Horseshoe Abyssal Plain, away from mapped tectonic faults (Fig. 1). Although of limited quality, aftershock locations were used to argue that the event occurred on a north-dipping thrust structure^{9,18}. The same authors proposed that the 1969 earthquake was generated by a large trans-lithospheric thrust fault beneath the Horseshoe Abyssal Plain, responsible for the consumption of oceanic lithosphere¹⁸. However, the expression of such north dipping lithospheric thrust was never found. In fact, currently mapped faults at crustal levels have an overall south-dipping vergence, such as the NE-SW Gorringe Northern Thrust that uplifts a massive 5-km-high seamount (Fig. 1). Besides, what mechanism could explain the existence of such trans-lithospheric fault, capable of generating seismic events of very high magnitude, typical of subduction zones, below a flat abyssal plain overlain by almost undisturbed sediments?

Over the past 50 years, intensive surveying searched for the elusive sources of the 1755 and 1969 earthquakes. Numerous models were proposed^{16,19,20}, but no consensus has emerged. Here, we integrate recent seismic and geological observations to propose a novel geodynamic interpretation for the region that is supported by geodynamic models, that can explain the generation of high-magnitude earthquakes, and can be reconciled with deep mantle structure.

High-resolution teleseismic *P*-wave travel-time tomography covering the SW Iberian margin and encompassing ocean bottom seismometer (OBS) data has recently confirmed the existence of a previously imaged high velocity anomaly below the Horseshoe Abyssal Plain (Fig. 2, S3, Methods)^{21,22,23}. This anomaly extends down to a depth of 250 km and has been proposed to correspond to a small slab that would be the expression of subduction initiation²². However, this model would imply >200 km of convergence and crustal displacement

accommodated in the SW Iberia region, along a NW-SE direction (see Fig. S2). Such amount of convergence contradicts palinspastic restorations that show a maximum tectonic shortening of 20 to 50 km at crustal levels^{24,25,26}. A possible explanation for this mismatch would be the existence of decoupling levels within the lithosphere. This would result in different amounts of shortening accommodated between the deep lithosphere and the shallow crust.

Figure 2 – Tomography data off Iberia showing a high-velocity anomaly interpreted as sinking lithospheric block. a, Slice at a sub-lithospheric depth of 150 km. b, Cross-section A-B cutting across the high-velocity anomaly off SW Iberia. Strikingly, this anomaly is located precisely below the flat Horseshoe Abyssal Plain, the region where the 1969 earthquake occurred (“beach-ball”), which was the strongest instrumental event ever registered in Europe and Africa. In this region, there are also a clusters of relatively deep hypocenters (up to 60 km; red circles) that were recorded during the NEAREST OBS experiment. Note that the Gorringer Bank is offset to the NE from this feature. Blue dots are the background seismicity recorded by the IPMA seismic network between 2000 and 2015. A resolution test is provided as Supplementary Information (SI-I)

More recently, a cluster of earthquakes located below the Horseshoe Abyssal Plain was characterized in detail using the NEAREST OBS experiment^{22,27}. These earthquakes occur at depths of 20 to 60 km, within the lithospheric mantle (Figs 2, S3). The small number of earthquakes above 20 km suggests mechanical decoupling between the unserpentinized lithospheric mantle (below 20 km) and a roofing serpentized mantle layer and crust²⁷. A seismic refraction study²⁸ further revealed that the basaltic oceanic crust is missing in the Horseshoe Abyssal Plain and that the sedimentary cover lies directly on top of exhumed serpentized lithospheric mantle, while to the south of the plate boundary, the African domain is made of typical oceanic crust.

A mechanism that would allow explaining these observations is a delamination of the lithosphere and/or the existence of a drip-like structure, similar to what is sometimes observed under continental orogens. We propose that the weak serpentized lithospheric mantle acts as a layer that decouples the positively buoyant crust from the underlying negatively buoyant lithospheric mantle. Once partially delaminated, the slab's negative buoyancy drives progressive sinking. This would explain why shortening at crustal levels is significantly less than would be implied by the existence of a typical subduction-related lithospheric slab penetrating the asthenosphere to depths of more than 200 km. Besides, if the delaminating block is still attached to the surface, it may still be pulling the crust down, explaining the sag of the Horseshoe Abyssal Plain and the formation of the thick sedimentary infill.

To test this hypothesis, we carried out 2D numerical models using the code Underworld²⁹. The models include two oceanic lithospheric plates simulating Africa and Eurasia (see Methods, Fig. E3). A far-field convergence velocity of 8 mm/yr was imposed to the African lithosphere; this value is higher than the present-day 4 mm/yr to account for the faster velocities during the Cenozoic³⁰. The Eurasian plate lacks a well-developed basaltic crust, which is instead replaced by a weak serpentized layer, mimicking the structure

obtained by refraction data²⁸. We tested the existence of either a single boundary separating the two plates (a weak fracture/transform fault) or a plate boundary zone limited by two main vertical fractures zones, which more closely resembles the structure of the less compressionaly-deformed plate boundary zone that can be observed further to the west (mimicking the Gloria Fault and the Gloria South Fracture; Fig. 1).

In the model with one plate boundary, the vertical weak zone is converted into an incipient subduction zone, with the thinner (Eurasia) plate subducting under the thicker one (Africa) (See Supplementaty Information SI-II; Fig. S2, Movie E1). In this case, the weaker serpentinized mantle on the Iberian side is converted into a subduction channel with the overlying crust accommodating about 200 km of displacement, which is not observed in the natural example. In the model with two vertical weak zones, something different and unexpected occurs. The block limited by the two weak zones sinks to the North under the thinner Eurasian plate (Fig. 3, Movie E2). The weak serpentinized layer thickens and accommodates most of the shortening, significantly decoupling the bulk deep deformation from the crust. A new main north-dipping fault forms at the contact between a delaminating lithospheric mantle block and the overlying serpentinized mantle (and crust). Right above this main fault, the sedimentary crust remains relatively flat. Surface deformation is accommodated mostly at the limits of the delaminating block. Interestingly, a south-dipping low angle thrust forms in the sedimentary cover (in a Gorringe-like location; see Fig. 3c and 4). Deformation (shortening) at crustal levels is less pervasive than that at deeper levels because most of the shallower deformation is accommodated within the weak serpentinized mantle. The overall geometry of the body resembles a case of flake tectonics, with a block of lithospheric mantle delaminating into the asthenosphere, driven by the plate's convergence and its own negative buoyancy.

Figure 3. Results of the numerical model with two plate boundaries showing the formation of a lithospheric delaminating block. Note the formation of a flake-like structure with a Gorringe-like south-dipping thrust at crustal levels and a deeper lithospheric north-dipping fault below a flat region comparable with the Horseshoe Abyssal Plain, where a 1969-like event may have occurred (see schematic illustration in Fig. 4). This model shows that under these circumstances the oceanic lithosphere may delaminate leading to unexpected lithospheric configurations similar to the ones observed offshore Southwest Iberia. The black arrows represent velocity vectors, and the red lines isotherms in degrees Celsius. Strain rate for values higher than $0.1 \times 10^{-13} \text{ s}^{-1}$ is also shown.

At first order, the numerical model with the two weak fractures explains most of the observations (Fig. 4): 1) The existence of a fast velocity anomaly that could reach depths of ~200 km below the flat Horseshoe Abyssal Plain limited by two deforming regions (that could mimic the Gorringe Bank, to the north, and the Coral Patch Ridge, to the south, see Fig. 1); 2) The decoupling of the less deformed crust, with the formation of a major low-angle fault dipping to the south (mimicking the Gorringe northern thrust), from the underlying unserpentinized lithosphere; 3) The existence of a hidden large north dipping fault separating the delaminating block from the overriding block; 4) This fault would be responsible for the observed low to moderate seismicity at depths of 20-60 km and could have been the elusive source of the 1969 earthquake.

Figure 4. Conceptual model combining the knowledge obtained from both the natural prototype and the results of the numerical models. A delaminating block could reproduce a drip-like structure as the one observed in the tomography model. The serpentinized layer decouples the crustal deformation from the one occurring at deeper lithospheric levels, giving rise to a flake-like structure, forming the south-dipping Gorringe thrust and the north-dipping 1969 fault.

We have also estimated a potential rupture area of $\sim 20,000 \text{ km}^2$ for this north dipping fault based on the along-strike length of the tomographic anomaly ($\sim 160 \text{ km}$) and the dip angle obtained from 1969 focal mechanism⁹ (Supplementary Information SI-III). According to our calculations, this fault would be capable of generating an M8.6 event in case of a full rupture and a M7.9 earthquake for a partial rupture of 80 km, which is the rupture area proposed for the 1969 event. This would solve the long-lasting problem of the apparent inexistence of a fault with the characteristics necessary to explain the occurrence of very high magnitude events in the region. Further work will be needed to image the deep lithosphere in this location with a higher resolution, such as the collection of more refraction and teleseismic data.

Our identification of the first example of delamination of oceanic lithosphere has important geodynamic implications. In this case, delamination is forced by the convergence between Africa and Eurasia and aided by a serpentinization front that mechanically decouples the crust from the underlying lithosphere, in between two fracture zones. The lack of other examples worldwide could be due to the process occurring only under very specific conditions: it requires very old, thick and serpentinized oceanic lithosphere, in the context of a diffuse plate boundary, subject to a minimum amount of far-field convergence. There is also an observational challenge – the process is not visible at the surface. While this phenomenon may be rare in the present-day Earth, it might have been more common in periods over which wider areas of older and unstable oceanic lithosphere covered the Earth's seafloor, and potentially earlier in the Earth's history such as in the Archean. Delamination of old oceanic lithosphere may play an important role in the initial stages of subduction initiation, facilitating the foundering of oceanic plates. Such a process would explain the very high magnitude seismicity off SW Iberia, providing a robust driving mechanism for the St. Vincent earthquake of 1969 and, possibly, for the Great Lisbon Earthquake of 1755.

- 207 1. Göğüş, O. H., Pysklywec, R. N., Şengör, A. M. C., Gün. E. Drip Tectonics and the enigmatic
208 uplift of the Central Anatolian Plateau. *Nature Communications* **8**, 1538,
209 <https://doi.org/10.1038/s41467-017-01611-3> (2017).
- 210 2. Houseman, G. A., McKenzie, D. P. & Molnar, P. Convective instability of a thickened boundary
211 layer and its relevance for the thermal evolution of continental convergent belts. *Journal of*
212 *Geophysical Research* **86**, 6115–6132, <https://doi.org/10.1029/JB086iB07p06115> (1981).
- 213 3. Stern, R.J., 2002. Subduction Zones. *Review of Geophysics* **40**, 3-1-3-38, (2002).
- 214 4. Magni, V. Plate tectonics: Crustal recycling evolution. *Nature Geoscience* **10**, 623-624,
215 <https://doi.org/10.1038/ngeo3015> (2017).
- 216 5. Göğüş, O. H., Ueda, K. Peeling back the lithosphere: Controlling parameters, Surface
217 expressions and the future directions in delamination modeling. *Journal of Geodynamics* **117**,
218 21-40, <https://doi.org/10.1016/j.jog.2018.03.003> (2018).
- 219 6. Burov, E. B. Rheology and strength of the lithosphere. *Marine and Petroleum Geology* **28**,
220 1402–1443, <https://doi.org/10.1016/j.marpetgeo.2011.05.008> (2011).
- 221 7. Cramer, F., Magni, V., Domeier, M. et al. A transdisciplinary and community-driven database
222 to unravel subduction zone initiation. *Nature Communications* **11**, 3750,
223 <https://doi.org/10.1038/s41467-020-17522-9> (2020).
- 224 8. Custódio, S., Lima, V., Vales, D., Cesca, S., Carrilho, F. Imaging active faults in a region of
225 distributed deformation from the joint clustering of focal mechanisms and hypocenters:
226 Application to the Azores-Western Mediterranean region. *Tectonophysics* **676**, 70-89,
227 [Doi:10.1016/j.tecto.2016.03.013](https://doi.org/10.1016/j.tecto.2016.03.013) (2016).
- 228 9. Fukao, Y. Thrust faulting at a lithospheric plate boundary: The Portugal earthquake of 1969.
229 *Earth and Planetary Science Letters* **18**, 205–216, [https://doi.org/10.1016/0012-](https://doi.org/10.1016/0012-821X(73)90058-7)
230 [821X\(73\)90058-7](https://doi.org/10.1016/0012-821X(73)90058-7) (1973).
- 231 10. Dewey, J.F. Continental margins: a model for conversion from Atlantic-type to Andean-type.
232 *Earth Planetary Science Letters* **6**, 189-197, [https://doi.org/10.1016/0012-821X\(69\)90089-2](https://doi.org/10.1016/0012-821X(69)90089-2)
233 (1969).
- 234 11. Wilson, J. T. Did the Atlantic close and then reopen? *Nature* **211**, 676 – 681,
235 [Doi:10.1038/211676a0](https://doi.org/10.1038/211676a0) (1966).
- 236 12. Grimison, N.L., Chen, W.P. The Azores–Gibraltar plate boundary: focal mechanisms, depths
237 of earthquakes, and their tectonic implications. *Journal of Geophysical Research Solid Earth*
238 **91**, 2029–2047, <http://dx.doi.org/10.1029/JB091iB02p02029> (1986).
- 239 13. Nocquet, J.M. Present-day kinematics of the Mediterranean: a comprehensive overview of GPS
240 results. *Tectonophysics* **579**, 220–242, <http://dx.doi.org/10.1016/j.tecto.2012.03.037> (2012).
- 241 14. Duarte, J.a.C., Rosas, F.M., Terrinha, P., Schellart, W.P., Boutelier, D., Gutscher, M.A., Ribeiro,
242 A. Are subduction zones invading the Atlantic? Evidence from the southwest Iberia margin.
243 *Geology* **41**, 839–842. <http://dx.doi.org/10.1130/G34100.1> (2013).
- 244 15. Martinez Solares J., Lopez Arroyo A. The great historical 1755 earthquake, effects and damage
245 in Spain. *Journal of Seismology* **8**, 275-294,
246 <https://doi.org/10.1023/B:JOSE.0000021365.94606.03> (2004).
- 247 16. Gutscher, M.-A. What caused the Great Lisbon Earthquake? *Science* **305**, 1247–1248.
248 [DOI:10.1126/science.1101351](https://doi.org/10.1126/science.1101351) (2004).
- 249 17. Stucchi, M., Rovida, A., Gomez Capera, A., Alexandre, P., Camelbeeck, T., Demircioglu, M.,
250 Gasperini, P., Kouskouna, V., Musson, R., Radulian, M., Sesetyan, K., Vilanova, S., Baumont,
251 D., Bungum, H., Fäh, D., Lenhardt, W., Makropoulos, K., Martinez Solares, J., Scotti, O., Živčić,
252 M., Albini, P., Batllo, J., Papaioannou, C., Tatevossian, R., Locati, M., Meletti, C., Viganò, D.,
253 Giardini, D. The SHARE European Earthquake Catalogue (SHEEC) 1000–1899. *Journal of*
254 *Seismology* **17**, 523–544, <http://dx.doi.org/10.1007/s10950-012-9335-2> (2013).

18. Purdy, G.M. The eastern end of the Azores–Gibraltar plate boundary. *Geophysical Journal of the Royal Astronomical Society* **43**, 973–1000, Doi:10.1111/j.1365-246X.1975.tb06206.x (1975).
19. Zitellini, N., Mendes-Victor, L., Córdoba, D., Dañobeitia, J., Nicolich, R., Pellis, G., Ribeiro, A., Sartori, R., Torelli, L., Bartolomé, R., Bortoluzzi, G., Calafato, A., Carrilho, F., Casoni, L., Chierici, F., Corela, C., Correggiari, A., Della Vedova, B., Gràcia, E., Jornet, P., Landuzzi, M., Ligi, M., Magagnoli, A., Marozzi, G., Matias, L.M., Penitenti, D., Rodriguez, P., Rovere, M., Terrinha, P., Vigliotti, L., Zahinos Ruiz, A. Source of the 1755 Lisbon earthquake and tsunامي investigated. *EOS Transactions American Geophysical Union* **82**, 285 (2001).
20. Gràcia, E., Dañobeitia, J.J., Vergés, J., PARSIFAL team. Mapping active faults offshore Portugal (36°N–38°N): implications for seismic hazard assessment along the southwest Iberian margin. *Geology* **31**, 83–86 (2003).
21. Gutscher, M.-A., Malod, J., Rehault, J.-P., Contrucci, I., Klingelhoefer, F., Spakman, W. and Mendes Victor, L. Evidence for active subduction beneath Gibraltar. *Geology* **30**, 1071–4, [https://doi.org/10.1130/0091-7613\(2002\)030<1071:EFASBG>2.0.CO;2](https://doi.org/10.1130/0091-7613(2002)030<1071:EFASBG>2.0.CO;2) (2002).
22. Monna, S., Cimini, G. B., Montuori, C., Matias, L., Geissler, W. H., Favali, P. New insights from seismic tomography on the complex geodynamic evolution of two adjacent domains: Gulf of Cadiz and Alboran Sea. *Journal of Geophysical Research - Solid Earth* **118**, 1587–1601, Doi:10.1029/2012JB009607 (2012).
23. Civiero, C., Strak, V., Custódio, S., Silveira, G., Rawlinson, N., Arroucau, P., & Corela, C. A common deep source for upper-mantle upwellings below the Ibero-western Maghreb region from teleseismic P -wave travel-time tomography. *Earth and Planetary Science Letters* **499**, 157–172, <https://doi.org/10.1016/j.epsl.2018.07.024> (2018).
24. Hayward, N., Watts, A.B., Westbrook, G.K., Collier, J.S. A seismic reflection and GLORIA study of compressional deformation in the Gorringe Bank region, eastern North Atlantic. *Geophysical Journal International* **138**, 831–850. (1999).
25. Galindo-Zaldívar, J.; Maldonado, A.; Schreider, A. A. Gorringe Ridge gravity and magnetic anomalies are compatible with thrusting at a crustal scale. *Geophysical Journal International* **153**, 586–594, doi:10.1046/j.1365-246X.2003.01922.x (2003).
26. Jiménez-Munt I., Fernández, M. Vergés, J., Afonso, J.C., Garcia-Castellanos, D., Fullea, J. The lithospheric structure of the Gorringe Bank: insights into its origin and tectonic evolution. *Tectonics* **29**, TC5019, doi:10.1029/2009TC002458 (2010).
27. Silva, S., Terrinha, P., Matias, L., Duarte, J.C., Roque, C., Ranero, C.R., Geissler, W.H., Zitellini, N. Micro-seismicity in the Gulf of Cadiz: Is there a link between micro-seismicity, high magnitude earthquakes and active faults? *Tectonophysics* **717**, 226–241, <https://doi.org/10.1016/j.tecto.2017.07.026> (2017).
28. Martínez-Loriente, S., Sallarès, V., Gràcia, E., Bartolome, R., Dañobeitia, J. J., Zitellini, N. Seismic and gravity constraints on the nature of the basement in the Africa-Eurasia plate boundary: New insights for the geodynamic evolution of the SW Iberian Margin. *Journal of Geophysical Research* **119**, 127–149, doi:10.1002/2013JB010476 (2014).
29. Moresi, L.-N., Betts, P. G., Miller, M. S., & Cayley, R. A. Dynamics of continental accretion. *Nature* **508**, 245 – 248, <https://doi.org/10.1038/nature13033> (2014).
30. G. Rosenbaum, G.S. Lister, C. Duboz. Relative motions of Africa, Iberia and Europe during Alpine orogeny. *Tectonophysics* **359**, 117–129, (2002).

Methods

Tomography

The tomography data²³ resulted from the integration of broadband teleseismic P-wave travel-time data, recorded between 2007 and 2013, from 387 seismic land stations deployed throughout the Ibero-Maghrebian region (Fig. E1). To obtain good ray coverage offshore SW Iberia 24 temporary broadband OBSs deployed during the NEAREST experiment between 2007 and 2008 were included³¹, which were crucial to robustly image the investigated region. Data from 5 additional broadband OBSs from the TOPOMED project, which recorded between 2009 and 2010, were also added³².

A total of 25644 arrival-time residuals from 451 teleseismic events, with a moment magnitude of at least 5.5, in the teleseismic epicentral distance range $30^\circ < \Delta < 95^\circ$ were examined. All the recorded traces were aligned using predictions from the global reference model ak135l³³ and low-pass filtered at 5 Hz. The adaptive stacking procedure³⁴ was then used to achieve final alignment, thus obtaining a set of arrival time residual estimates for each source.

The structure is represented by a regular grid of nodes spaced 0.4° in latitude (from 26°N to 46°N) and longitude (from 19°W to 5°E) and $\sim 35\text{-km}$ in depth, extending from the Moho down to 800-km depth.

Relative arrival-time residuals were inverted using the FMTOMO package³⁵ to recover the upper mantle seismic structure. FMTOMO solves the forward problem of travel-time prediction using a grid-eikonal solver, known as the Fast Marching Method^{36,37,38}. Model parameters are adjusted with a subspace inversion technique³⁹. The forward and inverse steps are applied iteratively in order to address the weakly non-linear nature of the inverse problem. Because Moho topography and crustal velocity structure can have a significant influence on teleseismic travel times⁴⁰, an a priori 3D crust and Moho model (PRISM3D) was applied to the

starting model²³, which has the effect of correcting for unresolved crustal contributions to the measured arrival-time residuals^{40,41}.

Numerical Modelling

Mathematical formulation

We use Underworld to solve the equations of momentum, continuity and energy, and to track material properties⁴². The continuity and the momentum equations are given by

$$\nabla \cdot \mathbf{v} = 0 \quad (1)$$

and

$$\nabla \cdot \boldsymbol{\sigma} = \rho \mathbf{g} \quad (2)$$

in which $\mathbf{v}$ is the velocity vector, $\boldsymbol{\sigma}$ the stress tensor, ρ is the density and $\mathbf{g}$ is the gravity acceleration vector. The stress tensor is given by

$$\boldsymbol{\sigma} = \boldsymbol{\sigma}' - P\mathbf{I} \quad (3)$$

in which $\boldsymbol{\sigma}'$ is the deviatoric stress tensor, P the pressure and $\mathbf{I}$ the identity tensor.

The deviatoric stress tensor ($\boldsymbol{\sigma}'$) is defined as a function of both strain rate and effective viscosity:

$$\boldsymbol{\sigma}' = 2\eta_{\text{eff}}\dot{\boldsymbol{\epsilon}} \quad (4)$$

with η_{eff} being the effective viscosity and $\dot{\boldsymbol{\epsilon}}$ the strain rate tensor defined as:

354

355

356
$$\dot{\epsilon} = \frac{1}{2} [\nabla \mathbf{v} + (\nabla \mathbf{v})^T] \quad (5)$$

357

358 The heat equation is given by:

359

360
$$\rho C_p \left(\frac{\partial T}{\partial t} + \mathbf{v} \cdot \nabla T \right) = \kappa \nabla^2 T + H_s + H_A \quad (6)$$

361

362 in which C_p is the specific heat, T is the temperature, t the time, κ the thermal diffusivity, H_s

363 the shear heating and H_A the adiabatic heating. The shear heating term is given by:

364

365
$$H_s = \boldsymbol{\sigma}' : \dot{\epsilon} \quad (7)$$

366

367 The adiabatic heating term is given by:

368

369
$$H_A = \frac{-T\alpha \mathbf{v}_y g}{c_p} \quad (8)$$

370

371 with α corresponding to the thermal expansivity and $\mathbf{v}_y$ representing the vertical velocity.

372

373 The effective viscosity is non-linear and calculated according to the following model:

374

375
$$\eta_{\text{eff}} = A \frac{1}{n} \dot{\epsilon}^{\frac{1}{n}-n} \exp \left(\frac{E_a + V_a P}{nRT} \right) \quad (9)$$

376

A being the pre-exponential factor, n the stress exponent, E_a the activation energy, V_a the activation volume, and R the universal gas constant.

Compressibility for lower crustal and mantle rheologies is achieved by relaxing the incompressible constraint of equation (1) such as:

$$\nabla \cdot \mathbf{v} = -\frac{P}{\lambda} \quad (10)$$

In which λ is a parameter equivalent to a viscosity.

The plastic flow is ensured by employing a Drucker-Prager yield criterion, where the material yields on the stress surpasses the yield value:

$$\sqrt{\sigma'_{II}} = p \sin(\phi) + C \cos(\phi) \quad (11)$$

With σ'_{II} representing the second invariant of the deviatoric stress tensor, ϕ the internal friction angle and C the cohesion. Strain weakening is modeled by linearly decreasing the internal frictional angle and cohesion as accumulated brittle strain increases up to 50% (See Table E1).

Model description

The models consist of a 2800×660 km 2-D box with a resolution of 1400×200 finite elements (Fig. E2). The grid spacing is non-uniform along the vertical axis with an element size of 1.25 km in the lithosphere and 5.0 km in the upper mantle. The modelled region comprises a 15 km thick air layer, and two oceanic plates embedded in a 660 km-deep upper mantle. The plate on the right represents an older Africa-like oceanic plate covered by a 10

km-thick basaltic crust overlaid by a 5 km-thick sedimentary layer. The plate on the left simulates a Eurasia-like oceanic plate in which the basaltic crust is replaced by a 10 km-thick serpentinized upper mantle layer (and eventually hydrated, but almost inexistent, basaltic crust) that tapers to the north. The rheological parameters used in this study are presented in Table E1. The initial temperature profiles of the oceanic plates are prescribed according to the half-space cooling model⁴³. The thickness of the boundary layer is chosen to be 100 km and the half-space cooling age is 120 and 170 Myr for the northern and southern segments, respectively.

The models were run with two configurations. The first one included a vertical 90 km-deep, and 10 km-wide weak zone (I) in the contact between the two plates to simulate a weakened plate boundary. In a second configuration, another weak zone (II) was added 200 km to the north, more closely resembling the natural example (see Fig. 1) with two initial parallel fracture/transform faults defining the Africa-Eurasia plate boundary.

The boundary conditions were set to be no slip on the bottom, free slip on the side and free surface on the top. To simulate the northward motion of the African plate, a velocity of 8 mm/yr was imposed to the southern oceanic plate while the northern segment was kept fixed. This velocity is representative of the Cenozoic Africa-Eurasia convergence velocity. Notwithstanding, we performed kinematic sensitivity tests by running the models at 2, 4, 6, and 8 mm/yr, and the results were the same, only changing the total time that the experiment took to reach the same end result.

References Methods

31. Carrara, G., NEAREST Team. NEAREST 2008 Cruise Preliminary Report R/V Urania, 1st Aug 2008–04th Sept 2008, Technical Report ISMAR, (2008).
32. Grevenmeyer, I. e. a. FS Poseidon POS389 & POS393 & FS Maria S. Merian MSM15/5: TOPOMED - Topographic, structural and seismotectonic consequences of plate reorganization in the Gulf of Cadiz and Alboran Sea, POS389: Valletta, Malta – Malaga, Spain, 06.–17.08.2009; POS393: Malaga, Spain – Faro, Portugal, 14.–24.01.2010, MSM15/5: Valletta, Malta - Rostock, Germany, 17.–29.07.2010. IFM-GEOMAR Report, 45. IFM-GEOMAR, Kiel, page 41 pp (2011).

33. Kennett, B.L.N., Engdahl, E.R., Buland, R. Constraints on seismic velocities in the Earth from travel times. *Geophysical Journal International* 122, 108–124, (1995).
34. Rawlinson, N., Kennett, B.L.N.. Rapid estimation of relative and absolute delay times across a network by adaptive stacking. *Geophysical Journal International* 157, 332–340, (2004).
35. Rawlinson, N., Reading, A.M., Kennett, B.L.N.. Lithospheric structure of Tasmania from a novel form of teleseismic tomography. *Journal of Geophysical Research, Solid Earth* 111, 1–21, (2006).
36. de Kool, M., Rawlinson, N., Sambridge, M.. A practical grid-based method for tracking multiple refraction and reflection phases in three-dimensional heterogeneous media. *Geophysical Journal International* 167, 253–270, (2006).
37. Rawlinson, N. and Sambridge, M.. Wave front evolution in strongly heterogeneous layered media using the fast marching method. *Geophysical Journal International* 156, 631–647 (2004).
38. Sethian, J.A.. A fast marching level set method for monotonically advancing fronts. *Proc. Natl. Acad. Sci. USA* 93, 1591–1595, (1996).
39. Kennett, B.L.N., Sambridge, M., Williamson, P.R.. Subspace methods for large inverse problems with multiple parameter classes. *Geophysical Journal International* 94, 237–247 (1988).
40. Rawlinson, N., Pozgay, S., Fishwick, S. Seismic tomography: a window into deep. *Physics of the Earth and Planetary Interiors*. 178, 101–135 (2010).
41. Arroucau, P., Custódio, S., Civiero, C., Silveira, G., Dias, N., Díaz, J., Villaseñor, A., & Bodin, T.. PRISM3D – A three-dimensional reference seismic model for Iberia and adjacent areas. *Geophysical Journal International* 225, 789–810, (2021).
42. Moresi, L.-N., Quenette, S., Lemiale, V., Meriaux, C., Appelbe, B., Mühlhaus, H.-B.. Computational approaches to studying non-linear dynamics of the crust and mantle. *Earth Planetary Science Letters*, 305, pp. 149–168, (2007).
43. Turcotte, D.L., Schubert, G.. *Geodynamycs*. Cambridge University Press, 456 pp. (2002).

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Author Contribution

JCD conceived and coordinated the work and wrote the first draft of the manuscript. JCD, NR, WPS, FR, PT and AR developed the conceptual model. JD, NR and JA designed the numerical models. NR run and processed the numerical model's results and produced the respective figures. CC and SC developed the tomographic model. CC and SS produced the tomography figures. CC carried out the tomography resolution tests. SS compiled and processed the NEAREST OBS data and produced the respective figures. JD, PT, FR, SC and AR developed the tectonic interpretation of the study region. SC and SS reviewed the data related to historical and instrumental earthquakes. FR calculated the seismogenic potential of the studied structure. All authors contributed towards analyzing and interpreting the integrated data and results, and towards editing the paper.

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Data availability

The datasets generated and analysed during the current study are available from the corresponding author on reasonable request. The data necessary to evaluate this work are provided in the manuscript.

Code availability

The Underworld code is an open-source numerical particle-in-cell finite element code and is available at <http://www.underworldcode.org>. Input files are available from the corresponding author on reasonable request.

Supplementary Information

SI-I: Tomography Resolution Tests

To test the resolving ability of our dataset Southwest off Iberia, we have performed a synthetic spike resolution test, where two input velocity anomalies with a width of ~ 100 km and an amplitude of 0.30 km/s were superimposed on our starting model at a depth of ~ 90 km (Fig. S1). The comparison between the input and output models suggests a good resolution for high-velocity features in the uppermost mantle although some vertical smearing occurs. Nonetheless, the separate input anomalies can be clearly identified. This further supports that the high-velocity anomaly imaged in the tomography is a real feature and not an artefact of the inversion.

Figure S1. Resolution test with spike anomalies. **a)** Slice at 90 km depth showing the locations of the input spikes. **b)** Slice at 90 km depth showing the recovered spike anomalies. **c)** Cross-section of the two spikes through the input model. **d)** Cross-section through the output model. Orientation of the cross-section is shown on the map of panel a) in black. The test shows a good resolving ability of the inversion for the studied area SW off Iberia.

509

510 **Figure S2. Result of the numerical models with one discrete plate boundary showing the**
 511 **transformation of a transform plate boundary into a subduction zone.** The serpentinized layer
 512 transforms in the effective plate interface. Note that the initiation of a proper subduction zone, with a slab
 513 reaching a depth of 250 km leads to an extreme crustal displacement of more than 200 km, which contradicts
 514 the natural observations. Note that these models should be regarded as a proof of concept, and not as a replica
 515 of nature. Strain rate for values higher than $0.1 \times 10^{-13} \text{ s}^{-1}$ is also shown.

517

518 **Figure S3. Simplified crustal tectonic map of Southwest Iberia overlaying the tomography data²³.**

519 The main thrust faults are shown in black, the vertical/strike-slip faults in white and the Gibraltar
 520 accretionary wedge in grey¹⁴. The Horseshoe Abyssal Plain seismicity cluster imaged during the
 521 NEAREST OBS experiment is also shown with coloured circles²⁷. There are two families of events: a
 522 shallower one (blue colours) and a deeper north-dipping one (orange-yellow colours; see also Fig. 2b).
 523 The “beach-ball” marks the epicenter of the 1969 St. Vincent earthquake⁹, far from any of the major
 524 mapped faults. The blue dots are the background seismicity recorded by the IPMA seismic network
 525 between 2000 and 2015²⁷.

526

527 *SI-III: Estimated rupture area and implied seismogenic potential*

528 We have calculated the seismogenic potential of the north dipping blind-thrust
 529 potentially hidden below the Horseshoe Abyssal Plain that was inferred based on the
 530 computation of the 1969 focal mechanisms and on the distribution of its aftershocks^{9,18}. The
 531 existence of this blind-thrust is consistent with our numerical models and with the distribution
 532 of the NEAREST OBS experiment (see Fig. 2b). The dip of this fault was obtained from the

1969 focal mechanism and considered to be 52°NE⁹, while its length was approximated to the main length of the anomaly observed in the tomography (see Fig. S4).

Figure S4. Illustration of the rationale behind determining the maximum fault rupture lengths (L) ascribed to the postulated seismogenic structure of the 1969 (M7.9) earthquake. The ~82 km long segment marked in blue corresponds to the fault length implied by the seismogenic domain of the 1969 (M7.9) earthquake as proposed by Fukao⁹ based on the distribution of the aftershocks; The ~267 km long segment marked in yellow corresponds to the possible maximum fault length implied by the NW-dipping blind fault discussed in this work and deduced from the projected contour of the depicted seismic tomography.

The relationship between the length of a fault and its magnitude has been empirically established based on differently assumed rheological and seismological parameters for different tectonic settings in different parts of the world⁴⁴. Building on a previous approach⁴⁵ a new model has been recently proposed for the study of high magnitude earthquakes in SW Iberia⁴⁶. This model acknowledges different assumptions to infer an average fault slip (d) from the length (L) of the considered seismogenic active fault (Fig. S4). This fault-slip is then used

together with the fault width (W) and an assumed shear modulus (μ), to obtain the seismic moment (M_o , eq. 1) and the moment magnitude (M_w , eq. 2):

$$M_o = \mu \cdot A \cdot d \quad (\text{eq. 1})$$

$$M_w = \frac{2}{3} \cdot \log_{10}(M_o) - 6.07 \quad (\text{eq. 2})$$

in which A is the rupture area ($A = W \cdot L$), d is the average slip and μ the shear modulus.

Figure S5. Fault Parameters. Geometric fault parameters considered in the discussed seismic scaling laws of Manighetti⁴⁵ et al.

The average fault-slip ($d = D_{max}/2$) is calculated as follows:

$$D_{max} \rightarrow \begin{cases} L \leq 2W_{sat} & D_{max} = a \frac{L}{2} \\ L > 2W_{sat} & D_{max} = a \frac{1}{\frac{1}{L} + \frac{1}{2W_{sat}}} \end{cases} \quad (\text{eq. 3})$$

in which a and W_{sat} are the slip/length ratio and the saturation fault width, respectively. These parameters are empirically deduced by the same authors for the study area using the few instrumental high-magnitude events that have occurred in this region. A slightly higher shear modulus of 65 GPa is assumed in this case. Table S1 summarizes the assumptions and input

parameters considered to determine the seismogenic potential of the northwards dipping blind-thrust explored in this work (e.g., its rupture area and implied M_0 and M_w). Two alternative main fault lengths (L) are considered (Fig. S4): one assuming seismic rupture only to the North of the main SWIM Fault; the other considering that the whole structure ruptures during a single event. The first scenario is clearly capable of justifying an event with a magnitude similar to the one of the 1969 ($M_{7.9}$) earthquake. Likewise, the second whole-rupture scenario is also compatible with the estimated ($M_{8.5-8.7}$) high magnitude generally ascribed to a 1755-like earthquake.

Table S2. Scaling of Law of Manighetti⁴⁵ et al. modified by Matias⁴⁶ et al. used to quantify the seismogenic potential of the postulated blind-fault in the study area.

	Model parameters	Scenario #1: rupture to the N of the SWIM fault	Scenario #2: whole rupture
Inputs	Fault length L (km)	82	267
	Saturation fault width – W_{sat} (km)	76	
	Slip-Length ratio (a - parameter)	$1.8e-4^{(46)}$	
	Seismogenic thickness Z (km)	60	
	Fault dip δ (°)	52	
	Shear modulus μ (GPa)	65	
Outputs	Average fault slip d (m)	3.7	8.7
	Rupture area $A = W.L$ (Km ²)	5603	20330
	Seismic moment M_0 (N.m)	$1.34e+21$	$1.15e+22$
	Moment magnitude M_w	8.0	8.6

References Supplementary Information

44. Johnston, A. C.: Seismic moment assessment of earthquakes in stable continental regions - III, New Madrid 1811–1812, Charleston 1886, and Lisbon 1755, *Geophys. J. Int.*, 126, 314–344, 1996.
45. Manighetti, I., M. Campillo, S. Bouleya, and F. Cottona (2007), Earthquake scaling, fault segmentation, and structural maturity, *Earth Planet. Sci. Lett.*, 253, 429–438.
46. Matias, L. M., T. Cunha, A. Annunziato, M. A. Baptista, and F. Carrilho (2013), Tsunamigenic earthquakes in the Gulf of Cadiz: fault model and recurrence, *Nat. Hazards Earth Syst. Sci.*, 13, 1–13.

Extended Data (Figures)

Figure E1. Location of the seismic stations (green symbols in main map) and events (red dots in inset global map) used for the tomographic model. The stations are coded according to their network and cover an area extending from the Pyrenees in the north to Morocco and the Canary Islands in the south. The six labelled stations are those for which residuals are shown in Civiero et al²³. Seismic experiment and station information can be found in their Supplemental Information Tables ST1 and ST2.

Figure E2. Numerical models initial set-up. Note that for the model shown in Fig. 3 both weak zones I and II has been implemented, while for the model shown in Fig. S2 only weak zone I has been implemented.

606 **E3. Table:** Physical parameters used in the numerical models.

Quantity	Symbol	Unit	Value
Gravity	g	m.s^{-2}	9.81
Thermal expansivity coefficient	α	K^{-1}	3.00×10^{-5}
Thermal diffusivity	κ	$\text{m}^2.\text{s}^{-1}$	1.00×10^{-6}
Sediments density at 273K	ρ_{uc}	kg.m^{-3}	2500
Dry mantle density at 273K	ρ_{dm}	kg.m^{-3}	3300
Hydrated mantle density at 273K	ρ_{hm}	kg.m^{-3}	2900
Top temperature	T_t	K	273
Bottom temperature	T_b	K	1748
Minimum viscosity	$\eta_{\min}$	Pa.s	1.00×10^{19}
Maximum viscosity	$\eta_{\max}$	Pa.s	1.00×10^{24}
Gas constant	R	$\text{J.K}^{-1}.\text{mol}^{-1}$	8.3145
Rheological parameters			
<i>Sediments (wet quartzite)</i>			
Stress exponent	n		2.3
Pre-exponential factor	A	$\text{MPa}^{-n}.\text{s}^{-1}$	3.1623×10^{-7}
Activation energy	E	kJ.mol^{-1}	1.54×10^5
Activation volume	V	$\text{m}^3.\text{mol}^{-1}$	8.0×10^{-6}
<i>Upper mantle (dry peridotite)</i>			
Stress exponent	n		3.5
Pre-exponential factor	A	$\text{MPa}^{-n}.\text{s}^{-1}$	2.5×10^4
Activation energy	E	kJ.mol^{-1}	5.32×10^5
Activation volume	V	$\text{m}^3.\text{mol}^{-1}$	12.0×10^{-6}
<i>Oceanic basaltic crust (plagioclase)</i>			
Stress exponent	n		3.2
Pre-exponential factor	A	$\text{MPa}^{-n}.\text{s}^{-1}$	3.1623×10^{-6}
Activation energy	E	kJ.mol^{-1}	2.38×10^5
Activation volume	V	$\text{m}^3.\text{mol}^{-1}$	0
<i>Hydrated/serpentinized mantle and weak zones (isoviscous)</i>			
Viscosity	η	Pa.s	1.0×10^{19}
Plastic parameters			
Cohesion	C	MPa	30.0
Cohesion after softening	C_s	MPa	0.1
Internal friction angle	α	$^\circ$	25.0
Internal friction angle after softening	α_s	$^\circ$	0.1
Max plastic strength (Peierls)			
Cohesion	C	MPa	300.0

607

608

609 **Extended Data (Movies)**

610

611 **Movie E1. Movie of the model with a single plate boundary**

612

613 **Movie E2. Movie of the model with two plate boundaries**

Supplementary Files

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