Effects of post-subduction processes on continental lithosphere

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Physical Sciences - Article

Keywords: subduction, post-subduction, plate tectonics

Posted Date: September 8th, 2021

DOI: <https://doi.org/10.21203/rs.3.rs-861968/v1>

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Effects of post-subduction processes on continental lithosphere

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Subduction is a key driver of plate tectonics on Earth¹. A range of observations indicate that the termination of subduction leads to diverse and unexplained tectonic and geological activity, including anomalous magmatism, exhumation and topographic subsidence, followed by rapid uplift. However, the mechanism driving this complex surface response remains enigmatic. A prime example of recent subduction termination can be found in northern Borneo (Malaysia), where subduction ceased in the late Miocene² and was followed by a puzzling tectonic response^{3,4,5,6,7,8} that cannot be reconciled with our current understanding of post-subduction tectonics. Here, we use new passive-seismic data to image, in unprecedented detail, a sub-vertical lithospheric drip that developed as a Rayleigh-Taylor gravitational instability⁹ from the root of a volcanic arc. We use thermo-mechanical simulations to reconcile these images with time-dependent dynamical processes within the crust and underlying mantle following subduction termination. Our model predictions illustrate how significant extension from a downwelling lithospheric drip can thin the crust in an adjacent orogenic belt, facilitating lower crustal melting and possible exhumation of subcontinental material. Our study provides a new paradigm for core-complex formation in other areas of recent subduction termination.

Most of Earth's sediments, oceanic crust, and mantle lithosphere returns to, and re-equilibrates with, Earth's mantle at subduction zones, where one tectonic plate dives beneath an adjacent plate and descends into Earth's interior, in the process driving plate motions at the surface¹⁰. Despite the diverse global implications of this subduction process, the dynamical mechanisms and structural changes that occur during and after subduction termination remain unresolved, leaving a fundamental gap in our understanding of the subduction cycle. Here we directly address this shortcoming and its implications for continental evolution through an inter-disciplinary study that specifically targets the post-subduction tectonic setting of northern Borneo (Sabah – eastern Malaysia, Fig. 1). This is achieved by combining novel geophysical and geological observations and inferences with state-of-the-art computational models.

The segment of continental lithosphere that is now northern Borneo bears the signature of two opposing subduction systems that ceased in the Miocene². Despite ongoing debate^{11,12,13}, oceanic lithosphere from the Proto-South China Sea (PSCS) is generally thought to have been consumed beneath north-western Borneo between ca 40 and 20 Ma^{8,11,14}. Subduction ended with continental collision of the Dangerous Grounds, which deformed and elevated the Crocker Formation above sea level (Fig. 1). To the northeast, seafloor spreading of the Sulu Sea began at ~21 Ma², driven by northwest subduction of the Celebes Sea beneath the Sulu arc, as indicated by calc-alkaline volcanism in the Dent and Semporna peninsulas¹⁵. Previous work suggests that trench-retreat, driven by slab rollback beneath the Celebes Sea, may have induced significant extension in northern Borneo⁸, leading to exhumation, accompanied by uplift, of subcontinental peridotite near Ranau and a rift-related magmatic episode (9.2-10.5 Ma) near Telupid⁷. The slab associated with Celebes Sea subduction is yet to be seismically imaged; however, based on the cessation of arc magmatism, subduction in this area is thought to have terminated at ~9 Ma².

Fig. 1: Geological map of Sabah with seismic stations of nBOSS and MetMalaysia. Surface geology information and geochemical dating from the study area². Inset in the upper left corner shows Malaysia in dark green and Sabah within the black rectangle. Inset in upper right corner summarises the tectonic evolution of Borneo and the Sulu Sea in the late Paleogene and mid-Miocene (modified from Hall, 2013).

Following subduction termination, a granite pluton was emplaced in a northwest-southeast extensional setting and crystallised between 7.8 and 7.2 Ma^{5,6}, with zircon inheritance patterns implying melting of the underthrust continental crust of the Dangerous Grounds. Thermochronological data from the pluton indicate remarkable peak exhumation rates of more than 7 mm/yr from the latest Miocene to the Early Pliocene⁶ (~6-4.5 Ma), resulting in Mt Kinabalu (4095 m elevation), which towers over the Crocker Range (average height 1500 m) and most peaks in southeast Asia. The timing and rate of subsidence and uplift in northern Borneo remain unresolved, although several lines of evidence suggest Neogene (23 – 2.5 Ma) uplift of 0.3 mm/year for western Sabah¹⁶, and subsidence in eastern Sabah (starting

at ~14 Ma with widespread sedimentation) followed by rapid uplift, making it fully emergent in the early Pliocene^{16,17} (~5 Ma). Plio-Pleistocene (~5-2 Ma) intraplate magmatism in the Semporna Peninsula (Fig. 1) indicates a transition from flux-driven volcanism associated with subduction to a system driven by decompression melting⁴. Observations from the mid-late Miocene (~10 Ma) to recent geological record are not easily reconciled with an active subducting system. The primary focus of this study is to reveal the mechanisms driving these more recent observations.

Lithospheric foundering in the form of a Rayleigh-Taylor instability⁹, or ‘drip’, has been invoked to explain a number of perplexing geological observations across a wide variety of tectonic settings^{18,19,20,21}. These drips are rarely detected with seismic imaging due to their relatively small size, but numerical simulations show the association between downwelling of dense material and topographic subsidence, followed by the removal of lithospheric mantle and topographic uplift^{20,22}. Lithospheric thinning through sinking of gravitational instabilities is also thought to facilitate melting under continents^{23,24}: typically, the void created by the missing lower lithosphere is replaced by upwelling asthenospheric mantle, which decompresses, triggering magmatism. Here, we present new seismic observations and novel geodynamic simulations that, when combined with existing geological and geochemical constraints, reveal the central role of such a drip in controlling the post-subduction evolution of northern Borneo.

We use data from 46 seismic stations that comprise the temporary northern Borneo Orogeny Seismic Survey (nBOSS) network (Fig. 1, Extended Data Fig. 1), deployed between March 2018 and January 2020. Our dataset is augmented by 28 stations from the national seismic network operated by MetMalaysia. We use analyses of P-to-S receiver functions to determine the thickness of the crust (Methods, Extended Data Fig. 2), while relative arrival-time residuals from teleseismic P waves are mapped as 3-D perturbations in the crust and upper

mantle (Methods, Extended Data Figs. 4-7). Anisotropy in the lithospheric mantle is constrained using core-refracted shear phases (Methods, Extended Data Fig. 3), and a shear-wave velocity model is used to generate a lithospheric thickness map (Methods).

Receiver-function results document large variations in crustal thickness: the regional pattern suggests an alternation of thin and thick crust striking to the NE (Fig. 2a). Surprisingly, the western region of the Crocker Range, including Mt Kinabalu, is underlain by relatively thin crust. Recent work indicates that rapid extension at ~10 Ma led to tectonic removal of crustal material and exhumation of the Ranau peridotites⁷, which include subcontinental lithospheric mantle rocks. If we assume that the region of crustal thickening (50-55 km in Fig. 2a) was produced by continental collision between the Dangerous Grounds and western Sabah, this indicates that the crust beneath the Kinabalu and Ranau area (~35 km) has thinned by about 30%. Another area of even thinner crust (<30 km) is located around Telupid, where recent U-Pb analysis⁷ found evidence for extension-related magmatism at ~9 Ma (Fig. 1).

The spatial variation of azimuthal anisotropy revealed in our study (Fig. 2c) exhibits two orthogonal orientations that imply two dominant deformational scenarios in the lower lithosphere. The fast orientations of anisotropy in the Crocker Range correlates strongly with the strike of the orogenic belt, suggesting that lattice preferred orientation of uppermost-mantle minerals may have developed by ductile deformation due to belt-orthogonal compressive stress. Anisotropy orientations near Mt Kinabalu appear to be slightly perturbed from the dominant orogen-parallel pattern in the Crocker Range, and assume a northwest-southeast trend in eastern Sabah. Despite approximate alignment with the direction of absolute plate motion, it is likely that these orientations are parallel to the back-arc spreading direction created during slab roll-back of the Celebes Sea. Our map of lithospheric thickness, averaging 100 km in the study area, demonstrates that the lower lithosphere is absent beneath the Semporna Peninsula (~40-50 km depth). This is also an area where the splitting analysis returns null

observations, which can be explained by vertically oriented flow (Methods, Extended Data Fig. 3). Additional direct evidence for thin lithosphere beneath Semporna comes from the estimated melting depth of basalts from sample SBK13⁴ (see Fig. 2d for location), which contains no evidence of contamination from crustal or subduction related processes. This yielded unusually shallow values of 55 ± 8 km²⁵, implying that intraplate magmatism in Semporna may have been produced by an upper mantle source that likely experienced decompression melting as it upwelled beneath recently thinned lithosphere.

The P-wave tomographic model has several key implications for reconciling the geological and seismic observations with underlying mantle structure and evolution. We image a high-velocity perturbation subparallel to the western coastline of Sabah (Fig. 2b). This anomaly is clearly visible in map view at 300 km depth with a width of ~ 70 km, consistent with the thickness of oceanic lithosphere. Most of the PSCS slab is thought to be in the lower mantle²⁶ but we consider a plausible interpretation of this anomaly to be a final upper-mantle remnant of this subducted slab. The vertical section in Fig. 2d indicates that the high-velocity anomaly is a steeply dipping feature; while this suggests that Paleogene subduction could have been at a high angle, it is also plausible that the steep angle was acquired when trench motion became quasi-stationary due to continent-continent collision. This latter process may have led to slab-detachment at the continent-ocean transition, which can occur even 10 Ma after onset of continental collision^{27,28}, whilst subsequent plate motion can explain why the PSCS slab is not found further to the east. This result provides important spatio-temporal constraints for unravelling the tectonic history of the PSCS and further refining regional plate motion reconstructions.

Figure 2: Crustal thickness, lithospheric thickness, anisotropy and P-wave tomographic model. **a)** Depth-to-Moho estimates determined by receiver function analysis. **b)** Horizontal slice at 300 km depth taken from the final tomographic solution model. Black line and red dots indicate the location of the vertical profile presented in **d)**. **c)** Fast-polarization orientations from core-refracted shear phases superimposed onto a lithospheric thickness map. Arrow indicates the absolute plate motion direction. White diamond in **d)** denotes the approximate location of sample SBK13. SD: Semporna Drip; PSCS: Proto-South China Sea; mk: Mt Kinabalu; cr: Crocker Range; sp: Semporna Peninsula; tlp: Telupid.

In addition to the PSCS slab, a vertical profile running from the Semporna Peninsula to Mt Kinabalu (Fig. 2d) in the P-wave tomographic model reveals an elongated high-velocity anomaly extending from ~125-350 km depth. This anomaly does not show lateral continuity (as expected from a slab remnant of the Celebes Sea) and has a dip orientation (tilted from the vertical towards the southeast) broadly consistent with the current direction of absolute plate

motion. Significantly, when accounting for an absolute ESE plate motion of ~ 2.7 cm/yr, its restored surface location coincides with the region where we inferred vertical flow and lack of lower lithosphere in the Semporna Peninsula (Fig. 2c), pointing towards its initiation at the root of the former volcanic arc, as has been postulated for Rayleigh-Taylor instabilities elsewhere on Earth^{18,20}.

Based upon the joint analysis of our new seismic results with the surface geological evidence, we interpret the northwest dipping high-velocity anomaly as a Rayleigh-Taylor gravitational instability that developed from the Dent-Semporna volcanic arc root, which we name the Semporna Drip (SD in Fig. 2d). We hypothesise that such a lithospheric drip can induce widespread extension in areas of thick crust, at distances that extend 100s of kilometres from the locus of the instability, as the lower lithosphere is removed and descends into the asthenospheric mantle. In the context of Sabah, we propose that the SD has thinned the crust of the Crocker Range, facilitating melting of the lower crust and emplacement of Mt Kinabalu. We also postulate that subsidence and uplift in eastern Sabah, in addition to observed intra-plate volcanism in the Semporna Peninsula⁴, and possibly Telupid, can be reconciled with the spatial and temporal evolution of the SD.

The integration of modern seismic data with structural geology and geochronology allows us to constrain a new thermo-mechanical simulation that aims to approximate, to first order, the recent tectonic evolution of northern Borneo and validate our hypothesis (Fig. 3). For simplicity, we perform the simulation in 2-D (model setup in Extended Data Fig. 8), which is sufficient to capture the fundamental behaviour of this 3-D system. Our simulation provides a dynamical mechanism that explains why a mountain as high as Mt Kinabalu is situated in the Crocker Range, underpinned by relatively thin crust. It demonstrates that a 1% positive density anomaly of possible mafic/ultramafic cumulates (e.g., pyroxenite, eclogite) from the sub-arc lithosphere is sufficient to induce dynamic downwelling (Fig. 3 and Supplementary video 1).

The dense volcanic root sinks vertically through the mantle, reaching a depth of approximately 400 km in 10 Ma. Its tilt arises naturally due to relative motion between the surface plate (to the right in our model) and underlying mantle; indeed, lateral motion of the plate leads to a horizontal offset of 170 km between the near-surface departure point and the base of the instability at 400 km, consistent with the seismological images and our estimate of regional lithospheric thickness. The progressive development and foundering of the lithospheric mantle results in a clear pattern of subsidence and uplift (Fig. 3); the vertical load of the density seed initially pulls down the crust, generating subsidence above the drip, but as the drip descends into the mantle and its impact at the surface wanes, we observe a systematic and gradual topographic rebound.

The most intriguing feature of our model predictions is the extensive thinning of the region of thickest crust, promoted by the lithospheric instability as it sinks into the underlying mantle. Analysis of the stress state in our simulations indicates that this region, which represents the Crocker Range, is under an extensional stress regime, as highlighted by the principal directions of tension (Extended Data Fig. 9). The distance between two points in the lower crust, measured via Lagrangian particles (Extended Data Fig. 10), is strongly time-dependent, with rapid stretching of ~ 4 mm/yr over the first Myr of the simulation, which progressively decreases to 1.2 mm/yr from 4-5 Myr and, subsequently, close to 0.5 mm/yr. This demonstrates a direct link between extension and the evolution of the downwelling drip. Indeed, 10 Myr after onset of the simulation, the Moho is $\sim 20\%$ shallower in the region of initially thickened crust. As soon as deformation localises in such a region, the feedback between the flow field and viscosity ensures that areas of weakness are maintained as the simulation evolves (Extended Data Fig. 11 and Supplementary video 2). As expected, considerable stretching of the crust leads to elevated temperatures at shallower depths (Fig. 3 and Extended Data Fig. 10), which would facilitate melting of the lower crust and emplacement

208 of Mt Kinabalu. In our geodynamical model, the instability does not fully detach from the
 209 lithosphere (Fig. 3 and Extended Data Fig. 12); it is important to note that such narrow
 210 connections lie beyond the resolution limit ($\sim$ station spacing) of our seismic imaging.
 211 Nonetheless, relatively low P-wave perturbations in the upper mantle beneath Semporna
 212 Peninsula (approximately -0.4 km/s dVp) do point towards possible asthenospheric upwelling,
 213 as do shallow upwelling velocities predicted in our simulations (Extended Data Fig. 12 and
 214 Supplementary video 1), a direct consequence of lower lithospheric removal.

215

216 **Figure 3: Temporal snapshots illustrating the evolution of a lithospheric drip from our**
 217 **simulation:** time (t) in Myr from start of the simulation. White lines are isotherms in Kelvin. A
 218 thick, higher density seed (3340 kg/m^3) is centred at $x=1000$ km, approximating mafic
 219 cumulates beneath a volcanic arc. Moho (interface between crustal and mantle material
 220 volume fraction fields) and surface elevation evolution are illustrated above each panel. The
 221 differential elevation is calculated by subtracting the surface elevation at time 0 from that at
 222 a specific timestep, accounting for the prescribed plate motion at 4 cm/yr . Note how the area

of thick crust (cr) undergoes significant thinning and the Moho becomes shallower. Similarly, subsidence accompanies initial sinking of the drip, which is then followed by vertical rebound as the lower lithosphere is removed. cr: Crocker Range; SD: Semporna drip.

The relatively small and transient nature of lithospheric drips mean that they are rarely imaged seismically; as such, their important role in the evolution of arc-complexes and subduction zones, following subduction termination, may have been overlooked. Our study delivers insights into post-subduction dynamics and processes, providing a conceptual template for predicting the widespread effects of lithospheric foundering following subduction termination. Our simulations imply that post-subduction lithospheric drips may be an important transitional process in the evolution of any trench-arc complex, causing extensive deformation even in areas far from the location of initial downwelling. The mechanism we propose here may play a major role in shaping regional tectonics, requiring reinterpretation of areas recently affected by subduction termination, including the Carpathians-Pannonian region, western United States, the Betic-Rif orogen and the eastern Mediterranean region.

Main References

1. Stern, R.J. Subduction zones. *Rev. Geophys.*, **40**, 3-1 (2002).
2. Lai, C.K., et al. Cenozoic Evolution of the Sulu Sea Arc-Basin System: An Overview. *Tectonics*, **40**, (2021).
3. Burton-Johnson, A., et al. A Triassic to Jurassic arc in north Borneo: Geochronology, geochemistry, and genesis of the Segama Valley Felsic Intrusions and the Sabah ophiolite. *Gondwana Res.*, **84**, 229-244 (2020).
4. Macpherson, C.G., et al. Plio-Pleistocene intra-plate magmatism from the southern Sulu Arc, Semporna peninsula, Sabah, Borneo: Implications for high-Nb basalt in subduction zones. *J. Volcanol. Geotherm. Res.*, **190**, 25-38 (2010).

- 248 5. Cottam, M., Hall, R., Sperber, C. & Armstrong, R. Pulsed emplacement of the Mount
249 Kinabalu granite, northern Borneo. *J. Geol. Soc.*, **167**, 49-60 (2010).
- 250 6. Cottam, M.A., Hall, R., Sperber, C., Kohn, B.P., Forster, M.A. and Batt, G.E. Neogene
251 rock uplift and erosion in northern Borneo: evidence from the Kinabalu granite, Mount
252 Kinabalu. *J. Geol. Soc.*, **170**, 805-816 (2013).
- 253 7. Tsikouras, B., Lai, C. K., Ifandi, E., Teo, C. H., & Xia, X. P. New zircon radiometric U-
254 Pb ages and Lu-Hf isotopic data from the ultramafic-mafic sequences of Ranau and
255 Telupid (Sabah, eastern Malaysia): Time to reconsider the geological evolution of
256 Southeast Asia?. *Geology*, (2021).
- 257 8. Hall, R. Contraction and extension in northern Borneo driven by subduction rollback. *J.*
258 *Asian Earth Sci.*, **76**, 399-411 (2013).
- 259 9. Houseman, G.A., McKenzie, D.P. and Molnar, P. Convective instability of a thickened
260 boundary layer and its relevance for the thermal evolution of continental convergent
261 belts. *J. Geophys. Res.*, **86**, 6115-6132 (1981).
- 262 10. Forsyth, D. and Uyeda, S. On the relative importance of the driving forces of plate
263 motion. *Geophys. J. Int.*, **43**, 163-200 (1975).
- 264 11. Hall, R. Reconstructing Cenozoic SE Asia. *Geol. Soc. Spec. Publ.* **106**, 153-184 (1996).
- 265 12. Cullen, A. B. Transverse segmentation of the Baram-Balabac Basin, NW Borneo:
266 Refining the model of Borneo's tectonic evolution. *Petroleum Geoscience*, **16**, 3–29
267 (2010).
- 268 13. Wu, J., & Suppe, J. Proto-South China Sea plate tectonics using subducted slab
269 constraints from tomography. *J. Earth Sci.*, **29**, 1304–1318 (2018).
- 270 14. Hutchison, C.S., Bergman, S.C., Swauger, D.A. & Graves, J.E. A Miocene collisional
271 belt in north Borneo: uplift mechanism and isostatic adjustment quantified by
272 thermochronology. *J. Geol. Soc.*, **157**, 783-793 (2000).

- 273 15. Bergman, S., Hutchison, C., Swauger, D., & Graves, J. K-Ar ages and geochemistry of
274 the Sabah Cenozoic volcanic rocks. *Bull. Geol. Soc. Malays.*, **44**, 165– 171 (2000).
- 275 16. Morley, C.K. & Back, S. Estimating hinterland exhumation from late orogenic basin
276 volume, NW Borneo. *J. Geol. Soc.*, **165**, 353-366 (2008).
- 277 17. Balaguru, A. & Nichols, G. Tertiary stratigraphy and basin evolution, southern Sabah
278 (Malaysian Borneo). *J. Asian Earth Sci.*, **23**, 537-554 (2004).
- 279 18. Zandt, G., et al. Active foundering of a continental arc root beneath the southern Sierra
280 Nevada in California. *Nature*, **431**, 41-46 (2004).
- 281 19. Levander, A., et al. Continuing Colorado plateau uplift by delamination-style convective
282 lithospheric downwelling. *Nature*, **472**, 461-465 (2011).
- 283 20. Göğüş, O.H., Pysklywec, R.N., Şengör, A.M.C. and Gün, E. Drip tectonics and the
284 enigmatic uplift of the Central Anatolian Plateau. *Nat. Comm.*, **8**, 1-9 (2017).
- 285 21. West, J.D., Fouch, M.J., Roth, J.B. and Elkins-Tanton, L.T. Vertical mantle flow
286 associated with a lithospheric drip beneath the Great Basin. *Nat. Geosci.*, **2**, 439-444
287 (2009).
- 288 22. Houseman, G.A. and Gemmer, L. Intra-orogenic extension driven by gravitational
289 instability: Carpathian-Pannonian orogeny. *Geology*, **35**, 1135-1138 (2007).
- 290 23. Elkins-Tanton, L.T. Continental magmatism, volatile recycling, and a heterogeneous
291 mantle caused by lithospheric gravitational instabilities. *J. Geophys. Res. Solid*
292 *Earth*, **112**, B03405, (2007).
- 293 24. Wang, H. and Currie, C.A. Magmatic expressions of continental lithosphere removal. *J.*
294 *Geophys. Res. Solid Earth*, **120**, 7239-7260 (2015).
- 295 25. Roberts, G.G., White, N., Hoggard, M.J., Ball, P.W. and Meenan, C. A Neogene history
296 of mantle convective support beneath Borneo. *Earth Planet. Sci. Lett.*, **496**, 42-158
297 (2018).

26. Hall, R. & Spakman, W. Mantle structure and tectonic history of SE Asia. *Tectonophysics*, **658**, 14-45 (2015).
27. Duretz, T., Gerya, T.V. and May, D.A. Numerical modelling of spontaneous slab breakoff and subsequent topographic response. *Tectonophysics*, **502**, 244-256 (2011).
28. Magni, V., Faccenna, C., van Hunen, J. and Funiciello, F. Delamination vs. break-off: the fate of continental collision. *Geophys. Res. Lett.*, **40**, 285-289 (2013).

METHODS

Crustal thickness estimate

Teleseismic earthquake signals impinging upon the crust below a seismic station produce P-to-S (Pms) conversions at the Moho. Receiver function (RF) methods aim at isolating the Pms conversion from reverberations to constrain crustal velocity and thickness beneath each station. We use a total of 3862 high-quality radial P receiver functions from an initial set of 14447. Receiver functions were calculated using a time-domain iterative deconvolution method²⁹. Those with a poor fit exhibit oscillatory behaviour or are inconsistent with other RFs from the same backazimuth and distance range, and are removed from further analysis. The receiver functions are stacked at each seismometer, and jointly inverted with Rayleigh-wave group-velocity dispersion curves from the GDM52 global compilation³⁰, using *joint96*³¹, an iterative non-linear least squares inversion method, to produce a 1-D shear velocity model at each seismometer. The Moho is then picked from each velocity model at the depth that corresponds to the base of the steepest positive velocity gradient where shear velocity exceeds 4 km/s.

Shear-wave splitting analysis

Shear-wave splitting is a key and near-unambiguous indicator of seismic anisotropy. In total, 367 new splitting measurements were made in this study. The splitting analysis was

performed using the method of Silver and Chan³² as implemented in the SplitRacer toolkit³³. The waveform data were bandpass filtered between 3 and 25 s, with analysis limited to phase arrivals with a signal-to-noise ratio greater than 1.5. The splitting measurements were manually reviewed and classified as ‘good’, ‘null’, and ‘poor’. Final station estimates are determined by averaging ‘good’ and ‘null’ observations through the single-layer error surface stacking method of Wolfe and Silver³⁴.

Lithospheric thickness estimate

The shear-wave velocity (V_s) model is calculated by initially inverting for phase velocities at periods between 25 and 200 s using a total of 225 teleseismic sources. We implement a two-plane-wave approach³⁵ and analytical sensitivity kernels³⁶. Pseudo-dispersion curves are extracted on a 0.2° grid at locations where the formal error from the least-squares inversion is less than 0.6 km/s. These are inverted for V_s using a reversible jump, Markov-chain Monte-Carlo method³⁷. Based on synthetic recovery tests, we estimate that regions with more than 10 dispersion observations contain robust solutions. We convert our V_s model to temperature using the equation^{38,25} $V_s = (1 + b_v(z - 50))(m(T - 273) + c + A \exp(-E - PV_a)/RT)$, where z is the depth (in km), T is the temperature (in K) and b_v , c , A , E , m and V_a are empirically fitted constants with the values of $3.84 \times 10^{-4} \text{ km}^{-1}$, 4.72 km s^{-1} , $-1.8 \times 10^{16} \text{ m s}^{-1}$, 409 KJ mol^{-1} , $-2.8 \times 10^{-1} \text{ m s}^{-1}$ and $10 \times 10^{-6} \text{ m}^3 \text{ mol}^{-1}$, respectively. The pressure is assumed to be lithostatic and is calculated as $P = \rho g z$, where ρ is the density (3330 kg m^{-3}) and g is the acceleration due to gravity (9.81 ms^{-1}). The LAB is assumed to be a constant temperature of 1333°C . Thus, we assume the depth to the LAB is the depth to the 1333°C contour.

P-wave traveltimes inversion

Relative arrival-time residuals of various global P-wave phases are determined from teleseismic sources with an angular distance larger than 27° . Arrival-time residuals are

computed by using an adaptive stacking approach³⁹. Out of 787 teleseismic sources, we retain only those with an average uncertainty estimate of the residuals less than 0.12 seconds. This equates to a reduction in the effective number of sources to 570, hence a total of 30671 relative arrival-time residuals. These residuals are inverted for 3-D velocity perturbations using the multi-stage Fast Marching Tomography Method (FMTOMO) package⁴⁰. Crustal thickness variations inferred from receiver function are directly included in FMTOMO during the inversion. Similarly, shear-wave velocities derived from receiver function analysis are converted into P-wave velocities according to the formulation of Brocher⁴¹ and included in the inversion to correct for crustal-velocity variations. While the Moho is kept fixed during the inversion, crustal velocities are inverted for.

Thermo-mechanical simulation

We simulate the dynamics, evolution and tectonic and geological consequences of a lithospheric drip using Fluidity, a finite element, control-volume, adaptive mesh computational modelling framework^{42,43}. Fluidity has been extensively tested, verified and validated for geodynamical simulations of this nature. The numerical approach uses anisotropic, dynamically adaptive mesh optimization (Extended Data Fig. 12 and Supplementary video 2), to provide high-resolution in areas of dynamic importance. Our initial model setup is illustrated in Extended Data Figure 8, with key model parameters listed in Extended Data Table 1. The model consists of a 2-D domain of dimensions 2000 x 660 km. We solve the incompressible Stokes, continuity and energy equations, under the Boussinesq approximation. We also track volume fractions representing four different materials (upper crust, lower crust, seed and mantle), using a control-volume discretization with a minimally diffusive HyperC face-value scheme, that ensures conservation and boundedness by coupling the fluxes between the volume fractions⁴⁴. Mechanical boundary conditions include a free-surface top, incorporating mesh movement with a prescribed 4cm/yr horizontal velocity, a zero-slip base, and open hydrostatic

sides. Thermal boundary conditions are prescribed to 273 K at the surface and 1573 K at the base of the model. The Moho temperature is set to 973 K⁴⁵. We include internal heating, at different rates for upper crust, lower crust and underlying mantle. Initial thermal conditions are different above a density seed that is 120 km (approximate width of Sulu arc) in width and 180 km in height, centred at x=1000 km, where the Moho temperature is relatively higher (1173 K) and the thermal LAB is at 50 km depth. While the thickness of the density seed is larger than that used in other studies, its width is considerably narrower^{20,22,24}, but is consistent with the width inferred from seismic images. Our initial conditions include a region of thick crust, approximating the Crocker Range, believed to have formed by continental collision between the Dangerous Grounds and western Sabah^{8,11,14}. Adjacent to a region of thick crust we have a thinner crust (30 km thickness), believed to have been extended through rollback associated with trench retreat^{7,8}. We implement a composite viscosity (dependent on temperature, pressure and second invariant of strain-rate) based on four deformation mechanisms (diffusion creep, dislocation creep, Peierls creep and yielding)⁴⁶, with the effective viscosity determined via a harmonic mean⁴⁶.

Methods References

29. Ligorria, J. P., & Ammon, C. J. Iterative deconvolution and receiver-function estimation: *BSSA* **89**, 1395-1400 (1999).
30. Ekström, G. A global model of Love and Rayleigh surface wave dispersion and anisotropy, 25-250 s: *Geophys. J. Int.* **187**, 1668-1686 (2011)
31. Herrmann, R. B. Computer programs in seismology: An evolving tool for instruction and research. *Seismol. Res. Lett.* **84**, 1081-1088 (2013).
32. Silver, P. G., & Chan, W. W. Implications for continental structure and evolution from seismic anisotropy. *Nature*. (1988).

33. Reiss, M. C., & Rumpker, G. SplitRacer: MATLAB code and GUI for semiautomated analysis and interpretation of teleseismic shear-wave splitting. *Seismol. Res. Lett.*, **88**, 392-409 (2017).
34. Wolfe, C. J., & Silver, P. G. Seismic anisotropy of oceanic upper mantle: Shear wave splitting methodologies and observations. *J. Geophys. Res. Solid Earth*, **103**, 749-771 (1998).
35. Forsyth, D.W., Li, A., Levander, A. and Nolet, G. Array analysis of two-dimensional variations in surface wave phase velocity and azimuthal anisotropy in the presence of multipathing interference. *Geophys. Monogr.*, **157**, 81 (2005).
36. Yang, Y. and Forsyth, D.W. Regional tomographic inversion of the amplitude and phase of Rayleigh waves with 2-D sensitivity kernels. *Geophys. J. Int.*, **166**, 1148-1160 (2006).
37. Dreiling, J. & Tilmann, F. BayHunter - MCMC transdimensional Bayesian inversion of receiver functions and surface wave dispersion. 5 Files (GFZ Data Services, 2019).
38. Priestley, K. & McKenzie, D. The thermal structure of the lithosphere from shear wave velocities. *Earth Planet. Sci. Lett.*, **244**, 285–301 (2006).
39. Rawlinson, N., Kennett, B. L. N. Rapid estimation of relative and absolute delay times across a network by adaptive stacking. *Geophys. J. Int.*, **157**, 332-340 (2004).
40. Rawlinson, N., de Kool, M., Sambridge, M. Seismic wavefront tracking in 3-D heterogeneous media: applications with multiple data classes. *Explor. Geophys.*, **37**, 322-330 (2006).
41. Brocher, T.M. Empirical relations between elastic wavespeeds and density in the Earth's crust. *Bull. Seismol. Soc. Am.*, **95**, 2081-2092 (2005).
42. Davies, D. R., Wilson, C. R. and Kramer S. C. Fluidity: A fully unstructured anisotropic adaptive mesh computational modeling framework for geodynamics. *Geochem. Geophys. Geosyst.*, **12**, Q06001 (2011).

43. Kramer, S. C., Wilson, C. R. and Davies, D. R. An implicit free surface algorithm for geodynamical simulations. *Phys. Earth Planet. Inter.*, **194**, 25–37 (2012).
44. Wilson, C. R. (2009), Modelling multiple-material flows on adaptive unstructured meshes, PhD thesis, Imp. Coll. London, London.
45. Currie, C. A., Hyndman R. The thermal structure of subduction zone back arcs. *J. Geophys. Res. Solid Earth*, **111**, B08404 (2006).
46. Garel, F. et al. Interaction of subducted slabs with the mantle transition-zone: A regime diagram from 2-D thermo-mechanical models with a mobile trench and an overriding plate. *Geochem., Geophys., Geosy.*, **15**, 739-1765 (2014).
47. Kennett, B. L. N., Engdahl, E. R., and Buland, R. Constraints on seismic velocities in the Earth from traveltimes. *Geophys. J. Int.* **122**, 108-124 (1995).

Contributions

S.P. conceived this study, created the P-wave tomographic model, integrated all observational constraints, and made all figures. S.P. performed the numerical simulation with guidance from D.R.D, building on developments by D.R.D., S.C.K. and C. R. W. S.P., D.R.D and R.H. made the main interpretation. C.B. measured azimuthal anisotropy. A.G. created the crustal thickness map. T.G. created the lithospheric thickness map. S.P., C.B., A.G., T.G., F.T., D.C. and N.R. participated in the seismic field acquisition. N.R. acquired funding for this study. N.R. and F.T. conceived the nBOSS experiment. S.P. wrote the paper, following discussion with, and contributions from, all authors.

Competing interests

The authors declare no competing financial interests

Code availability

The Fluidity computational modelling framework, including source code and documentation, is available from <https://fluidityproject.github.io/>; the latest release, which was used for the simulations presented herein, has been archived at Zenodo <https://doi.org/10.5281/zenodo.5221157>. The source code and manual for FMTOMO are available at <http://www.earth.org.au/codes/FMTOMO/>.

Data availability

Seismic data of the nBOSS network will be accessible through the IRIS Data Management (<http://www.iris.edu/mda>) from February 2023 (see https://doi.org/10.7914/SN/YC_2018). Details on the status of this database may be obtained from N.R.

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Acknowledgments S.P. acknowledges support from the Natural Environmental Research Council (NERC) Grant NE/R013500/1 and from the European Union's Horizon 2020 Research and Innovation Program under Marie Skłodowska-Curie Grant Agreement 790203. We thank the TanDEM-X Science Communication Team (German Aerospace Center (DLR) e.V.) for providing TanDEM topographic data. We thank the NERC Geophysical Equipment Facility for loan 1038. Numerical simulations were undertaken on the NCI National Facility in Canberra, Australia, which is supported by the Australian Commonwealth Government.

Supplementary Files

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