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1 **Mid-ocean ridge unroofing revealed by magmatic intrusions**

2
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16 **Mid-ocean ridges are quintessential sites of tectonic extension¹⁻⁴, where divergence between**
17 **lithospheric plates shapes abyssal hills that cover ~2/3 of Earth's surface^{5, 6}. Here we show**
18 **that tectonic extension at the ridge axis can be partially undone by tectonic shortening across**
19 **the ridge flanks. This process is evidenced by recent sequences of reverse-faulting**
20 **earthquakes ~15 km off axis at the Mid-Atlantic and Carlsberg Ridges. Using mechanical**
21 **models, we show that shallow compression of the ridge flanks up to the brittle failure point**
22 **is a natural consequence of lithosphere unbending away from the axial relief. Intrusion of**
23 **magma-filled fractures, which manifests as migrating swarms of extensional seismicity along**
24 **the ridge axis, can provide the small increment of compressive stress that triggers reverse-**
25 **faulting earthquakes. Through bathymetric analyses, we further find that reverse**
26 **reactivation of mid-ocean ridge normal faults is a widely occurring process that can reduce**
27 **the amplitude of abyssal hills by as much as ~50%, shortly after they form at the ridge axis.**
28 **This "unroofing" mechanism exerts a first-order influence on the fabric of the global ocean**
29 **floor, and provides a physical explanation for reverse-faulting earthquakes in an extensional**
30 **environment.**

31
32 Normal faults are major contributors to the seismic activity of mid-ocean ridges¹⁻⁴ (MORs).
33 Over 100s of kyrs, these faults offset and tilt newly accreted volcanic seafloor, shaping regularly
34 spaced abyssal hills⁵⁻⁷. Remarkably, normal faults with the largest offsets are typically found on
35 the edges of axial valleys, but almost never in the outer ridge flanks (Fig. 1). This observation,
36 made in the 1970s before multibeam bathymetric data became widely available, led several authors
37 to formulate the concept of abyssal hill unroofing, i.e., the notion that MOR normal faults
38 experience reverse slip that shortens their offsets as they migrate up the ridge flanks⁸⁻¹¹. This idea
39 lost favor because no clear driving mechanism had been identified and, most notably, because
40 reverse-faulting earthquakes compatible with ridge-normal compression near the axis had not been

41 reported¹². Reverse-faulting events are well documented hundreds of km from MOR axes¹³, in
42 oceanic lithosphere older than 10–20 Ma, where a compressive stress state is expected¹⁴. Rare
43 reverse-faulting events have also been observed in lithosphere as young as 3 Ma (~30 km from the
44 axis at slow spreading rates¹³), or even closer to the axis in the vicinity of transform faults¹⁵.
45 Surprisingly, these events indicate both ridge-parallel and ridge-normal compression (Extended
46 Data Fig. 1a), and have been attributed to either thermal stresses¹⁶, the complex stress state of
47 ridge-transform intersections¹⁷, or recent shifts in plate motion¹⁸. Clear manifestations of near-axis
48 (< 30 km) compressive seismicity have overall remained elusive, until late 2022.

49

50 **Swarms of reverse-faulting earthquakes**

51

52 On September 26th, 2022, a particularly active swarm of earthquakes with moment magnitudes
53 $M_w > 4$ began at 54°N on the Northern Mid-Atlantic Ridge^{19, 20} (Fig. 1a). This slow-spreading
54 segment (2.2 cm/yr of full plate separation rate) is located north of the Charlie-Gibbs Fracture
55 Zone, and bounded by two non-transform offsets. Its symmetric morphology is typical of a
56 magmatically-robust ridge section (in the sense of ref. 21), with a ~1.3 km deep axial valley
57 flanked by two shoulders, and abyssal hill bounding normal faults with a characteristic spacing of
58 2–4 km (Fig. 1c). During the first 3 days of the 2022 swarm, relocated seismicity (Methods,
59 Supplementary Table 1) showed southward migration along the axial valley by ~45 km (Extended
60 Data Fig. 2a), and entirely consisted of E-W extensional mechanisms. This normal-faulting activity
61 continued for 27 more days without clear signs of further migration. 80 hours into the swarm, a
62 magnitude-5.1 reverse-faulting earthquake indicative of ridge-normal compression occurred
63 below the summit of the ridge shoulder, 15 km east of the neovolcanic axis. Between September
64 29, 2022 and January 4, 2023, 11 more reverse-faulting events with M_w up to 5.9 occurred on N-
65 S striking, ~45–50°-dipping planes, outlining two narrow bands symmetrically located ~15 km
66 east and west of the ridge axis (Fig. 1a). Inspection of the teleseismic P waveforms from the largest
67 compressive events shows a very short delay between the direct P arrival and phases reflected off
68 the seafloor and sea surface. This implies remarkably shallow hypocenter depths, which we
69 estimate within 2–5 km below seafloor (Methods).

70

71 We identified a similar pattern in a November 2014 earthquake sequence on the slow-
72 spreading Carlsberg Ridge at 6°N in the Indian Ocean (full spreading rate: 2.4 cm/yr; Fig. 1b, d).
73 There, a less active swarm of $M_w \sim 5$ normal-faulting seismicity at the ridge axis (with possible,
74 though unclear, southeastward migration: Extended Data Fig. 2e–h) preceded 5 reverse-faulting
75 events ~15 km NE and SW of the ridge axis. Interestingly, both ridge flanks had produced similar
76 M_w 5–5.3 reverse-faulting events in 2005 and 2009 that were not preceded by a detectable swarm
77 of normal-faulting earthquakes.

78

79 **Compression of mid-ocean ridge flanks**

80

81 Sparse bathymetric data shows that the 6°N segment of the Carlsberg Ridge is morphologically
82 similar to the 54°N segment of the Mid-Atlantic Ridge, with even more pronounced axial relief
83 (Fig. 1c, d). By summing the offsets of major normal faults on both sides of the axis^{22–24} (Methods),
84 we estimate the tectonically-accommodated fraction of plate separation T at ~15% in both
85 segments (Fig. 2). The remaining ~85% of plate divergence occurs through magmatic
86 emplacement and could be achieved by, for example, the intrusion of a meter-wide dike through
87 the axial lithosphere every ~50 years^{25, 26}. Such partitioning of magmatic and tectonic strain likely
88 explains the formation of regularly-spaced abyssal hills bounded by faults that accumulate normal
89 offsets as large as ~500 m when they grow along the edges of the axial valley^{7, 27, 28}. The geometry
90 of the compressive events detected in both segments however indicates that the shallow portion of
91 these normal faults can be reactivated in a reverse sense once they migrate out of the axial valley
92 and reach the top of the ridge shoulders.

93

94 Standard numerical simulations of MORs show that the formation of an axial valley results
95 in compressional faulting on the ridge flanks, even when dike intrusion accounts for most of the
96 plate spreading (Fig. 3a). Here we model an idealized 2-D cross-axis section of a slow-spreading
97 center where a fraction $M (= 1 - T)$ of the far-field extension ($2U$) is accommodated by continuous
98 magmatic emplacement in a narrow axial zone^{27–29}. In this simulation, the value of M is a self-
99 evolving function of the axial relief, which accounts for mechanical feedbacks between relief
100 development and dike intrusions in the axial zone^{30, 31} (Methods). M is adjusted to yield a
101 characteristic abyssal-hill spacing of ~3 km following ref. 7. The development of shoulders is a
102 natural consequence of the vertical displacement that accompanies tectonic extension at the ridge
103 axis: as it moves off axis, uplifted lithosphere must unbend to transition to a state of rigid horizontal
104 motion³². This occurs over a characteristic flexural length scale α modulated by the integrated
105 strength of young oceanic lithosphere. In the simulation shown in Fig. 3a, the thermal structure of
106 the ridge is fixed, and chosen such that the visco-elasto-plastic lithosphere flexes into shoulders
107 similar to those of the 54°N segment of the Mid-Atlantic Ridge. This produces a zone of shallow
108 horizontal compression ~25 km wide and ~2 km deep, beginning 6 km from the magma injection
109 zone on both ridge flanks (Fig. 3a). The associated horizontal deviatoric stresses exceed tens of
110 MPas and reach the point of compressive Mohr-Coulomb failure, with stresses that peak ~15 km
111 off-axis. This stress field is consistent with the occurrence of shallow, ridge-perpendicular reverse-
112 faulting earthquakes near the summit of the ridge flanks (Fig. 1c, d).

113

114 The shallow compressive strain produced by unbending of the ridge flanks can be estimated
115 through an order-of-magnitude approach, as the ratio of the compressive zone thickness D (~2 km
116 in Fig. 3a) to the lithosphere's radius of curvature³³. A simple unbending model^{34, 20} relates the
117 radius of curvature to the amplitude h of the axial relief (~1.5 km) and the flexural length scale α
118 (~10 km). This yields a strain $\epsilon \sim Dh / \alpha^2$ on the order of 0.03, large enough to leave a detectable
119 signature in seafloor relief.

120

121 The bathymetric signature of unfauling

122

123 At both the Northern Mid-Atlantic Ridge and Carlsberg Ridge, we detect a sharp break in
124 the trend of cumulative fault offset vs. off-axis distance, at about 15 km, roughly co-located with
125 the bands of compressive earthquakes (Fig. 2). The break in slope corresponds to a reduction in
126 the apparent fraction T of tectonic extension by 0.06–0.1. We interpret this break in slope as a
127 reduction of the average offset on abyssal hill bounding faults caused by cumulative reverse slip
128 (Fig. 4a). To first order, T corresponds to the characteristic normal fault heave (the horizontal
129 component of fault offset) divided by the characteristic fault spacing. Therefore, a decrease from
130 $T = 0.13$ to $T' = 0.07$ (Fig. 2a) can represent a reduction in heave from ~400 m down to ~200 m
131 for a population of faults evenly spaced by ~3 km, a scenario consistent with the morphology of
132 the 54°N segment of the Mid-Atlantic Ridge (Fig. 1c).

133

134 The Carlsberg and Northern Mid-Atlantic Ridge are not unique in hosting near-axis
135 reverse-faulting earthquakes. Over a dozen more comparable events can be found throughout the
136 Global CMT Catalog^{35, 20}, along the ultraslow-spreading Southwest Indian Ridge, the slow-
137 spreading Mid-Atlantic Ridge and Carlsberg Ridge, as well as the intermediate-spreading
138 Southeast Indian Ridge (Extended Data Figs. 1, 3). Based on low-resolution bathymetric and
139 gravity data, ref. 20 attributed these events to unbending in the footwall of large-offset detachment
140 faults, where small magnitude reverse seismicity had previously been documented³⁶. This
141 interpretation is however at odds with available high-resolution bathymetry data: out of the 13
142 ridge sections that have hosted reverse-faulting earthquakes and have been mapped with shipboard
143 multibeam echosounders (Fig. 4b, Supplementary Tables 2, 3), ten display a magmatically robust
144 morphology (in the sense of ref. 21) with symmetric, staircase-like abyssal hills bounded by short-
145 offset faults (Extended Data Figs. 1, 3). While the other three do host scattered detachment faults,
146 a systematic co-location with the reverse quakes is difficult to establish given the large location
147 uncertainties of teleseismic events. Analyzing representative bathymetric profiles from all these
148 ridge sections systematically yields a reduction in the apparent T fraction by 0.02–0.18 at distances
149 ranging from ~10 to 30 km off-axis (Extended Data Figs. 1, 4, 5). In addition, revisiting a published
150 database of 2157 normal faults identified along the intermediate-spreading Chile Ridge²⁴—where
151 we found no near-axis compressive event in the CMT catalog—yields a similar pattern at the scale
152 of the entire ridge (Extended Data Figs. 6–8, Supplementary Tables 3, 4). There, $T-T' \sim 0.07$, but
153 the shift occurs slightly closer to the ridge axis, within ~7 to ~16 km from the neovolcanic zone.

154

155 It is important to note that processes other than unfauling could contribute to a reduction
156 in apparent T fraction across the flanks of magmatically-robust ridge sections. These include
157 sedimentation filling bathymetric lows and reducing the apparent amplitude of abyssal hills.
158 Median sediment thickness however only increases by ~5 m per Myr of seafloor age^{37–40}.
159 Sedimentary infilling is thus unlikely to account for more than ~30 m of apparent offset reduction
160 (e.g., $T-T' < 10^{-2}$) in areas of suspected unfauling. On the other hand, gravitational mass wasting

161 is known to alter the slopes of axis-facing fault scarps by maintaining them at or below the
162 frictional angle of repose ($\sim 35^\circ$) through repeated rocksliding^{41, 42}. The progressive nature of this
163 process however appears at odds with the abruptness of the reduction in apparent T fraction (Fig.
164 2), unless the reduction marks the point where normal slip on abyssal hill faults ceases and mass
165 wasting begins to overcome tectonic uplift. We do not favor this endmember interpretation because
166 fault scarps appear to migrate across the unfauling zone without any detectable change in their
167 average slope (Extended Data Fig. 6d). Furthermore, axis-facing scarps on the fast-spreading East-
168 Pacific Rise—which presumably experience gravitational mass wasting like other ridge sections—
169 —show an increase in apparent T fraction ~ 17 km off-axis (Extended Data Fig. 9). This increase is
170 readily explained by unbending away from an axial high³⁴ ($T-T' < 0$), mirroring the unbending away
171 from an axial valley ($T-T' > 0$; Fig. 4a). We conclude that changes in apparent T fraction primarily
172 reflect shallow flexural strains reworking the tectonic fabric of abyssal hills, as opposed to surface
173 processes.

174
175 Available bathymetric data suggests that the unfauling process as summarized in Fig. 4a
176 primarily occurs at magmatically robust ($T \leq 0.2$) MOR sections with an axial valley flanked by
177 pronounced ridge shoulders. It does not rule out the occurrence of unbending quakes in the footwall
178 of oceanic detachments as proposed by ref. 20, but the impact of such events on ocean floor
179 physiography is not as clear. As a measure of normal fault shortening, $T-T'$ provides a proxy for
180 the horizontal strain associated with shoulder compression. Its typical value (~ 0.07) however
181 exceeds our initial order-of-magnitude estimate of $\varepsilon \sim 0.03$. One reason could be that the depth
182 extent of reverse slip is greater than predicted by our numerical models (Fig. 3a), e.g., $\varepsilon \sim 0.07$ if
183 D extends to ~ 5 km instead of ~ 2 km below seafloor, consistent with the depths of near-axis
184 reverse-faulting earthquakes²⁰ (Supplementary Table 1). This could be achieved if the effective
185 frictional strength of the shallow, compressed lithosphere was reduced, for example by the
186 pressurization of pore fluids trapped in abyssal hill bounding faults. Supra-hydrostatic pore
187 pressures could plausibly deepen the neutral bending plane in MOR shoulders down to ~ 5 km or
188 deeper (Extended Data Fig. 10), and would also have the effect of narrowing the cross-axis extent
189 of the unfauling zone, contributing to a more abrupt decrease in apparent T (see Supplementary
190 Information). Further, assuming $D = 5$ km and using the unfauling distance (the position x_c of the
191 sudden decrease in T) as a proxy for the flexural wavelength α to estimate the unfauling strain ε
192 yields a reasonable agreement with our measurements of $T-T'$ across all candidate unfauling sites
193 (Fig. 4c).

194

195 **Role of magmatic intrusions**

196

197 The widespread bathymetric signature of abyssal hill unfauling contrasts with its relative
198 scarcity in the seismic record (Fig. 4b). The moment rate associated with extensional faulting at
199 the axis straightforwardly scales as $\dot{M}_E \sim GUTH_0$ (per unit length along the ridge axis, with G the
200 shear modulus and H_0 the thickness of the axial lithosphere; 43). The seafloor shortening rate in

201 the ridge flanks can be estimated as the spreading half-rate times the reduction in fault heave per
202 unit distance away from the axis, i.e., $\varepsilon U = (T-T')U$. Along the neutral plane located at depth D
203 (Figs. 3a, 4a), there is no shortening. Consequently, the shortening rate averaged over the thickness
204 of the compressive domain is $(T-T')U/2$. It follows that moment rate associated with reverse slip
205 should scale as $\dot{M}_R \sim GU(T - T')D/2$, and $\dot{M}_R / \dot{M}_E \sim \frac{(T-T')D}{2TH_0}$. Assuming $D = H_0 \sim 5$ km, $T-T' \sim$
206 0.07 and $T \sim 0.15$ yields $\dot{M}_R / \dot{M}_E \sim 0.2$. Ref. 20 estimated the total moment rate of near-axis
207 reverse-faulting events in the CMT catalog as the equivalent of 1.5 Mw5.5 earthquakes per year
208 for the period 1985–2023, which is about 5% of the moment rate associated with normal-faulting
209 events ($\dot{M}_R / \dot{M}_E \sim 0.05$). One explanation for this discrepancy could be that reverse slip in the
210 shallow ridge flanks is more aseismic than normal slip on the edge of the axial valley. This would
211 be consistent with supra-hydrostatic pore fluid pressures in the unfaulting zone, which increase the
212 characteristic nucleation size of earthquakes and favor a regime where aseismic reverse slip
213 transients account for greater cumulative displacements than reverse-faulting earthquakes^{43–45}.

214

215 The idea that flexural stresses alone cause reverse-faulting earthquakes only sporadically
216 is consistent with the fact that the two clearest manifestations of the unfaulting process, the 2014
217 Carlsberg and 2022 Mid-Atlantic Ridge swarms, were plausibly triggered by a dike intrusion
218 event¹⁹. In both instances, the reverse-faulting earthquakes were preceded by on-axis swarms of
219 extensional seismicity, with a pattern of southward migration at ~ 0.2 m/s in the Atlantic case.
220 Migrating seismicity at rates ~ 0.1 – 1 m/s has unambiguously been linked to dike propagation in
221 Iceland and Ethiopia^{46,47}, and similar swarms have been observed at magmatically active segments
222 of the Juan de Fuca Ridge⁴⁸ and Gakkel Ridge⁴⁹. It is thus likely that the initial swarm of
223 extensional earthquakes was the manifestation of a magmatic fissure propagating along the ridge
224 axis. Standard elastic dislocation modeling (Fig. 3b, Methods) shows that opening of a vertical
225 dike between 4 and 10 km below the axial valley floor induces static Coulomb stress changes that
226 promote normal slip on axial valley faults while promoting shallow reverse slip ~ 15 km off-axis.
227 Magmatic fissures are typically ~ 1 meter wide at oceanic spreading centers²⁵. A dike intrusion
228 such as those inferred at the Carlsberg and Mid-Atlantic Ridge may thus represent a one-in-fifty
229 year event that precipitated the occurrence of otherwise elusive unfaulting earthquakes (Fig. 4a).
230 Although not necessary to drive reverse slip on MOR normal faults, magmatic intrusions may have
231 been instrumental in revealing this fundamental component of seafloor spreading, which over
232 geological time strongly reworks the fabric of the global ocean floor.

233

234

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405 **Main figure captions**

406

407 **Figure 1 | Swarms of normal and reverse seismicity at two slow-spreading mid-ocean ridges.**

408 **a.** Bathymetry of the 54°N segment of the Northern Mid-Atlantic Ridge⁵⁰, showing relocated
409 earthquakes from the 2022 swarm. Red and blue beachballs show normal and reverse mechanisms,
410 respectively. NTO: Non-transform offset. AV: Axial valley. **b.** Bathymetry of the Carlsberg Ridge
411 near 6°N, with relocated 1990–2020 seismicity (Global CMT Catalog³⁵). Shading reveals zones
412 with multibeam data (visible fault scarps) vs. interpolated bathymetry (textureless). **c.** Topographic
413 cross-section along dotted line in panel a, with fault scarps shown in red (west side) and yellow
414 (east side). Blue bars indicate projected location of reverse-faulting events. **d.** Topographic cross-
415 section along dotted line in panel b: closest available transect with shipboard bathymetry. V.E. =
416 vertical exaggeration.

417

418 **Figure 2 | Tectonic strain and the bathymetric signature of unfauling.** Plots of cumulative

419 fault heave vs. distance from the axis for **a.** the Mid-Atlantic Ridge at 54°N, and **b.** the Carlsberg
420 Ridge at 6°N. Blue bands mark location of reverse-faulting events. Black lines show best fitting
421 piece-wise linear function, with slope T (the tectonic fraction of plate separation) near the ridge
422 axis, and slope T' further off-axis.

423

424 **Figure 3 | The stress state of mid-ocean ridge shoulders. a.** Cross-section of the horizontal

425 deviatoric stress in a 2-D numerical model of seafloor spreading developing an axial valley and
426 symmetric shoulders^{30,31}. See text for meaning of symbols. **b.** Change in shear stress ($\Delta\tau$) resolved
427 on 45°-dipping faults (>0 and <0 indicates normal and reverse motion, respectively) induced by 1
428 meter of dike opening (purple line), shown only in areas where the Coulomb stress change (ΔCFS)
429 brings faults closer to failure. Black line shows relief of the 54° segment of the Mid-Atlantic Ridge.

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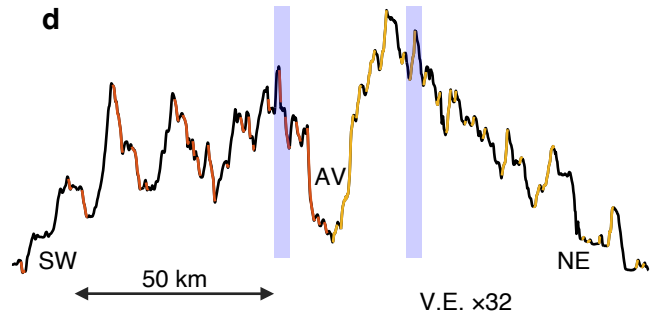
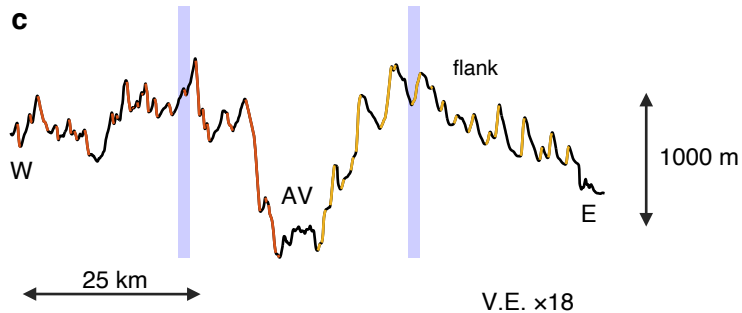
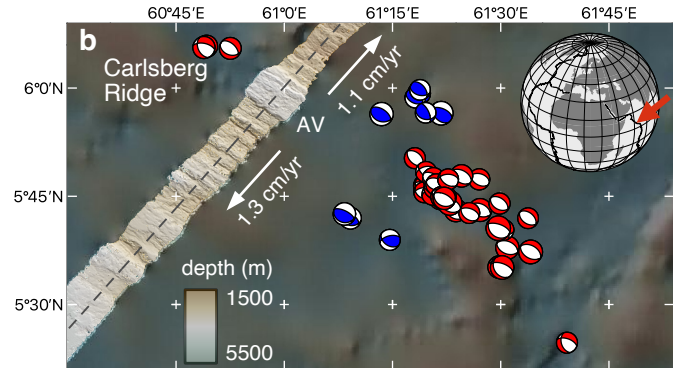
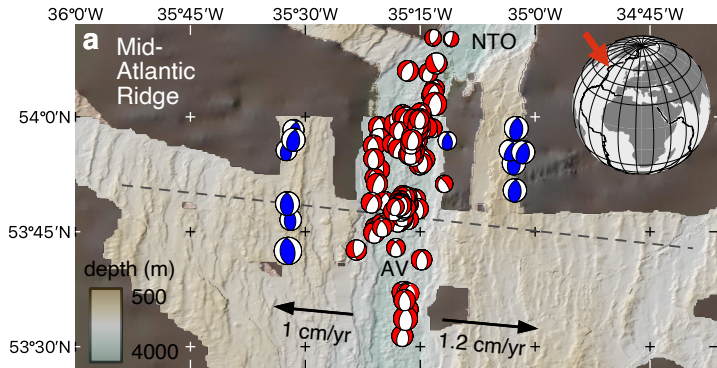
431 **Figure 4 | Mid-ocean ridge unfauling. a.** Cartoon showing the shallow flexural compression of

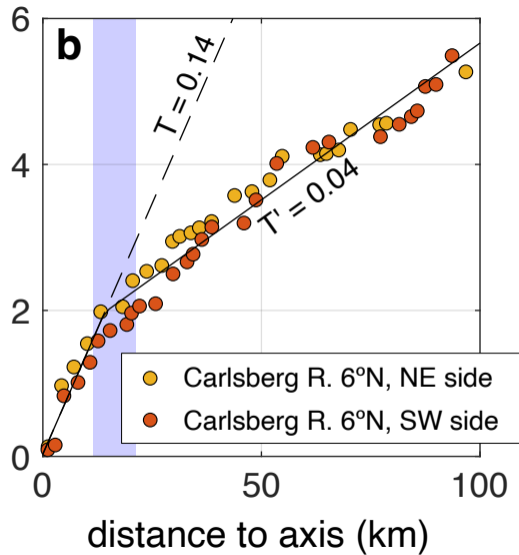
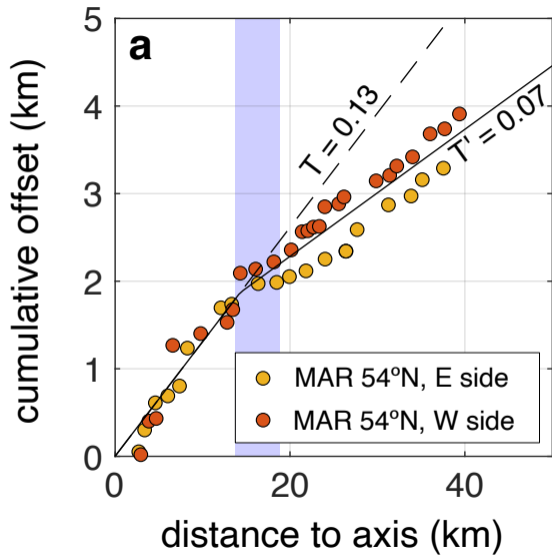
432 mid-ocean ridge flanks, and the occasional triggering of reverse slip by a magmatic intrusion. Red
433 and blue stars represent extensional and reverse-faulting earthquakes. Inset illustrates the depth-
434 dependent stress state of unbending lithosphere, with shallow compression (blue) and deep
435 extension (red). **b.** Spilhaus projection map of mid-ocean ridge strike-slip and normal-faulting
436 earthquakes from the Global CMT catalog. Blue dots show reverse-faulting earthquakes that
437 occurred in areas with a documented signature of bathymetric unfauling (see Supplementary
438 Information). **c.** Relationship between the change in apparent T fraction ($T-T'$) and unfauling strain
439 (ε), estimated as the product of axial relief (h) by the depth of the unfauling zone (D , here assumed
440 to equal 5 km everywhere) divided by the unfauling distance x_c (here used as a proxy for flexural
441 wavelength α) squared. Error bars are derived from the non-linear least squares fit to the
442 cumulative offset plots (e.g., Fig. 2, see Methods). Size of the symbols scales with x_c .

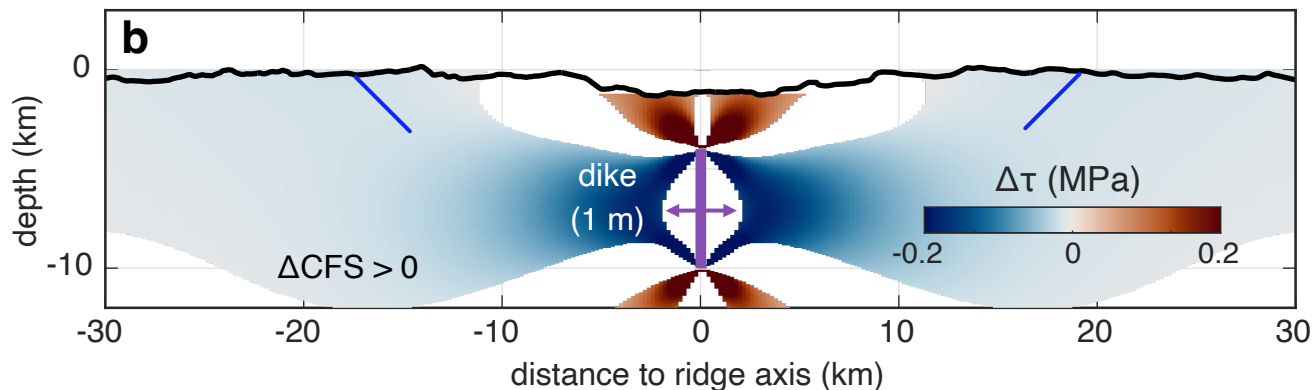
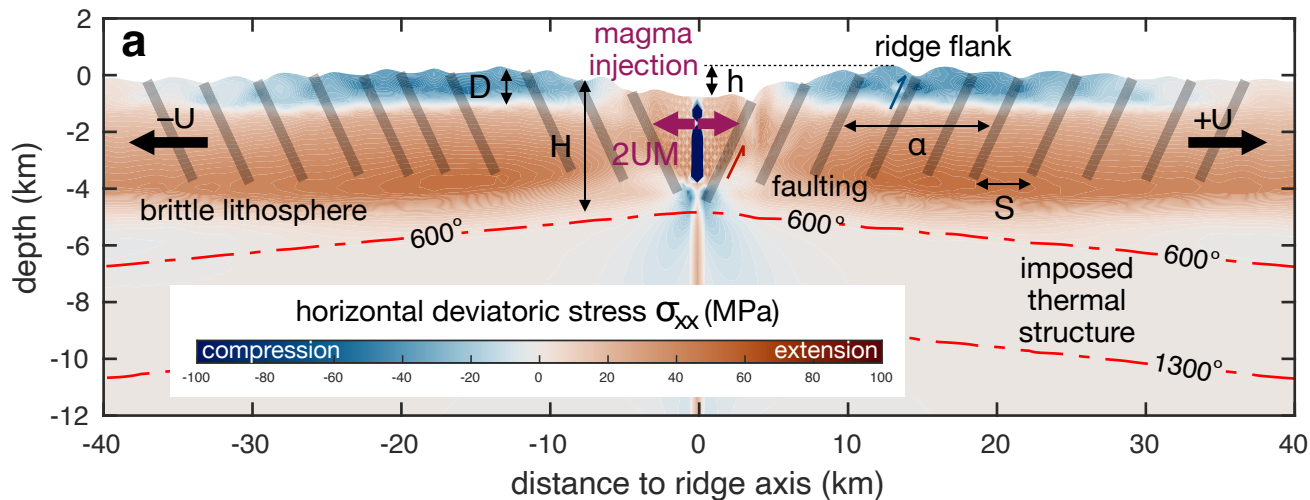
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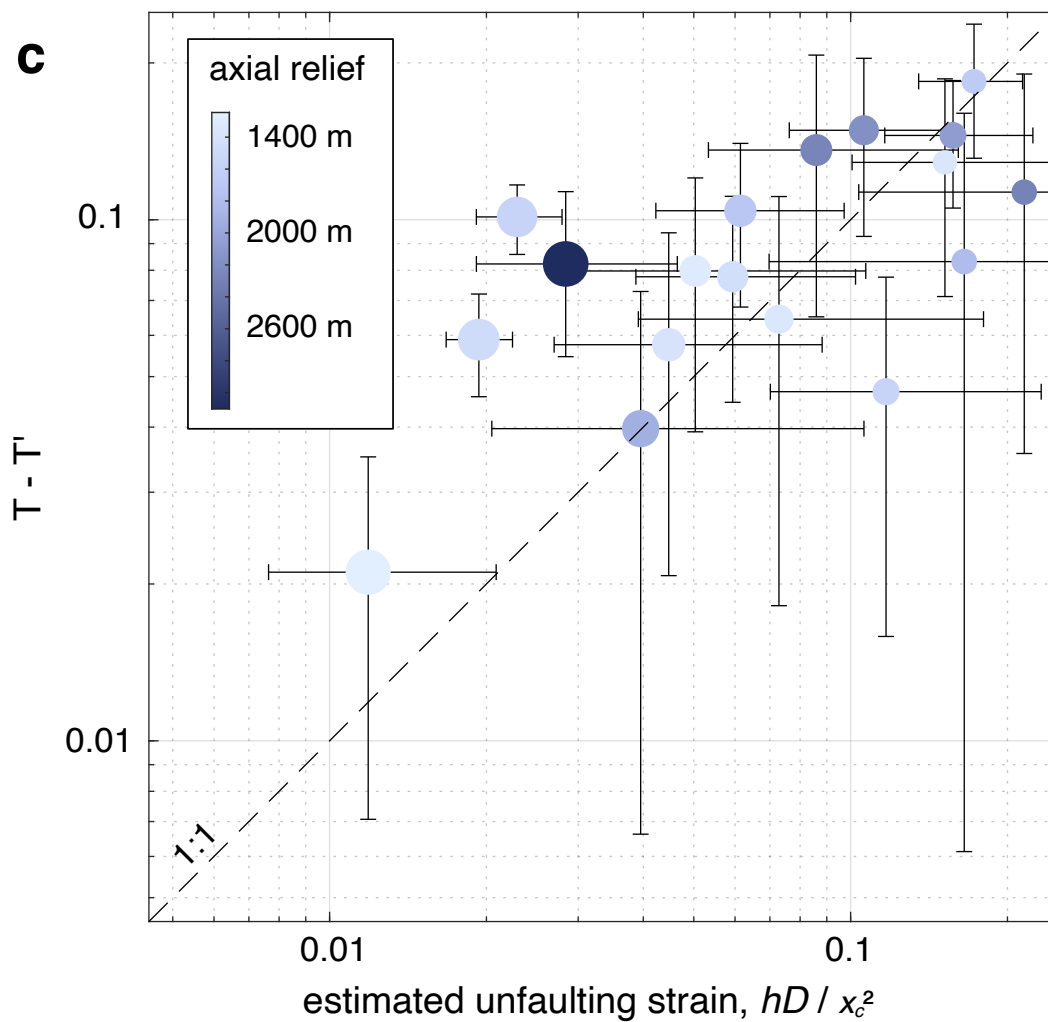
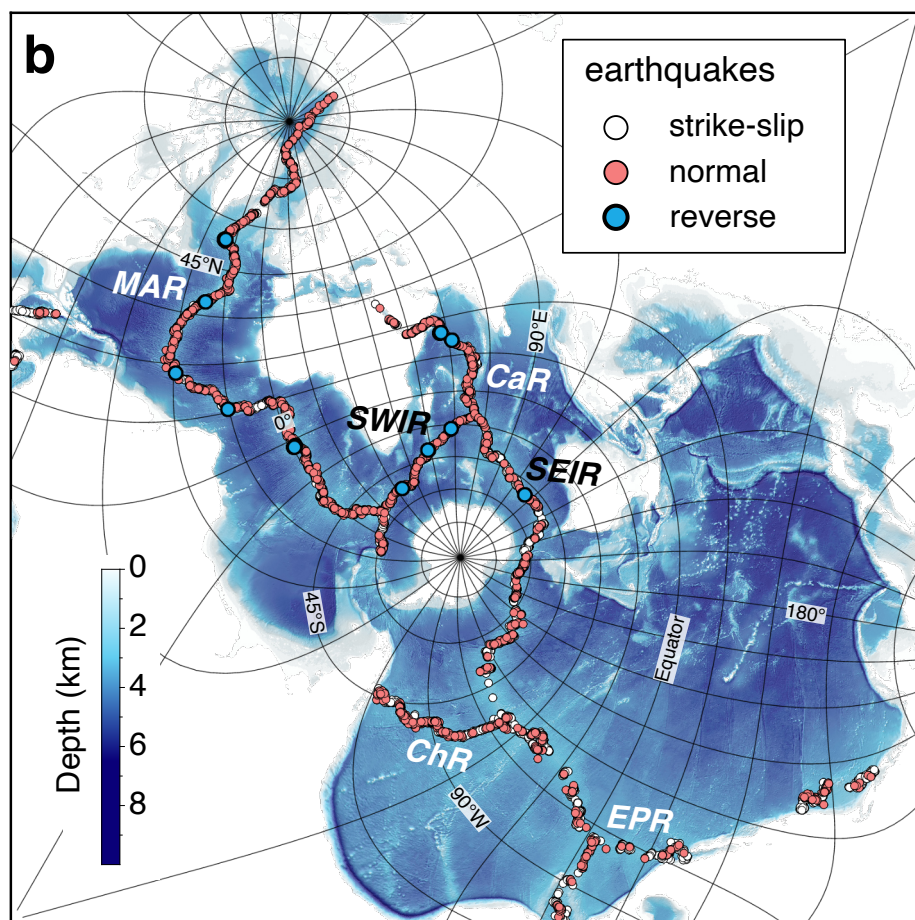
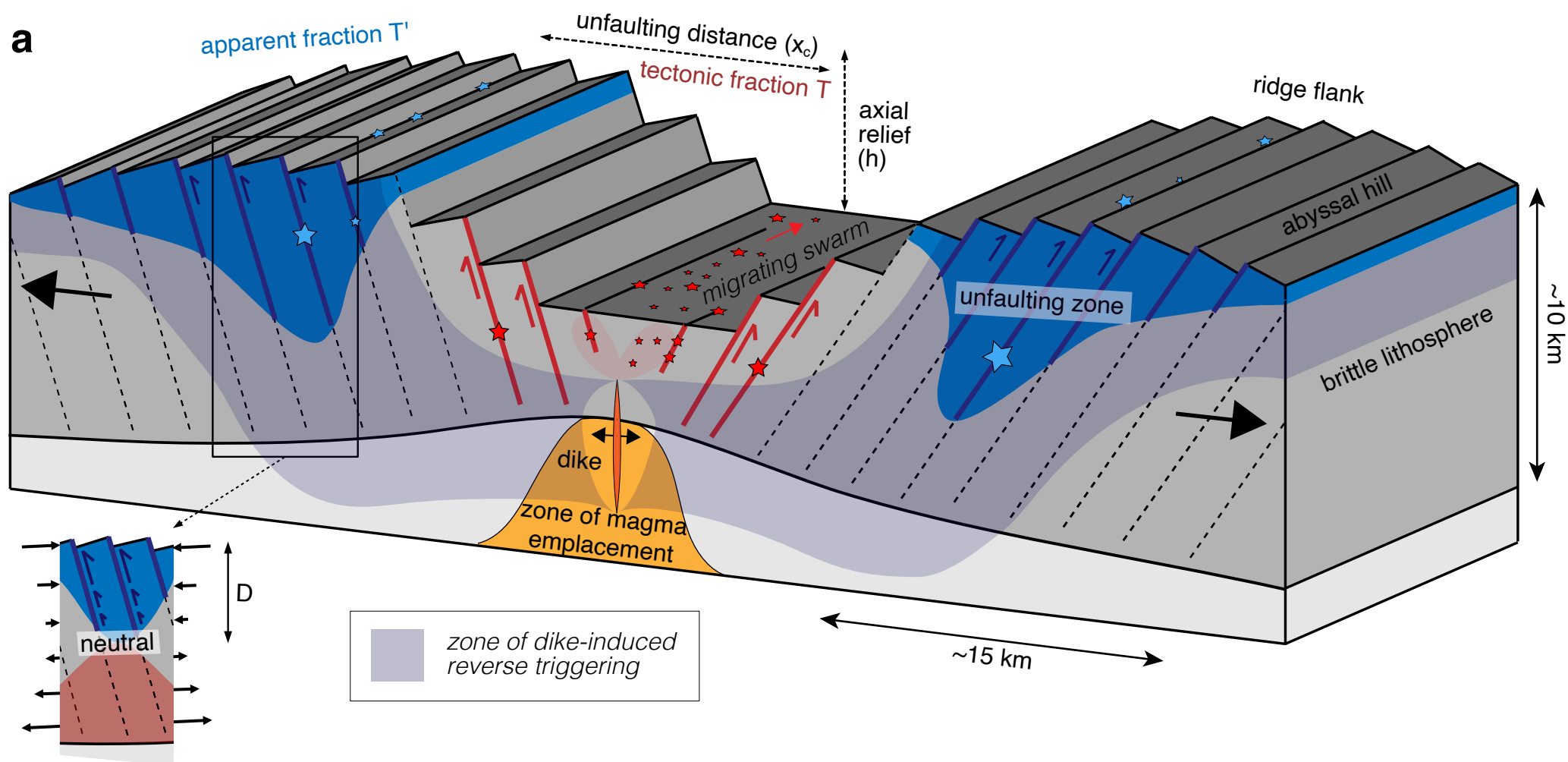
444 **Methods**

445









446 **Earthquake locations and mechanisms**

447 Earthquakes on the northern Mid-Atlantic Ridge are far from seismographic stations, and event
448 detection is therefore limited to moderate and larger earthquakes ($M > 4.0$). In this study, we rely
449 primarily on hypocenters reported by the National Earthquake Information Center (NEIC,
450 [https://www.usgs.gov/programs/earthquake-hazards/national-earthquake-information-center-
451 neic](https://www.usgs.gov/programs/earthquake-hazards/national-earthquake-information-center-
451 neic)). The NEIC routinely reports hypocenters for earthquakes $M \geq 4.5$ in this area, but only rarely
452 for events smaller than $M 4.2$. To complement the NEIC hypocenters, we make use of surface-
453 wave detections and locations calculated using intermediate-period (35-150 s) surface waves
454 recorded on the Global Seismographic Network (GSN) using a global grid search⁵¹. The threshold
455 for the surface-wave detector is ~ 4.7 , but occasionally an event of that size will be missed in the
456 NEIC catalog.

457 Uncertainties for earthquake locations based on teleseismic P-wave arrivals are typically
458 10–20 km⁵². They can be even larger for surface-wave detections⁵¹, which gives rise to significant
459 scatter in earthquake epicenters. We relocate them using cross-correlations of their Rayleigh and
460 Love waveforms, without correcting for specific surface wave radiation patterns⁵³. This yields
461 clearly defined bands of seismicity with relative location uncertainty smaller than 2 km, based on
462 the misfit of the very large numbers of differential travel times (Supplementary Table 1). Empirical
463 location estimates, based on inversions of subsets of the available data, are smaller than ~ 5 km for
464 all but five of the earthquakes. Because the absolute location of the earthquakes is not well
465 constrained, we shift their latitudes by 0.054° (~ 6 km) to the West in a manner that centers the
466 central band of seismicity over the neovolcanic axis as identified from bathymetric maps of the
467 54°N segment (Fig. 1a).

468 We calculate focal mechanisms for these events using the centroid-moment-tensor
469 algorithm employed in the Global CMT (GCMT) Project^{35, 54}. Long-period (>40 s) body- and
470 surface-wave seismograms recorded on the GSN are matched to estimate the six elements of the
471 moment tensor, and an earthquake centroid location in time and space. We follow the procedure
472 used in the standard GCMT processing³⁵, and attempt analysis of all events with at least one
473 reported magnitude of 4.7 or larger. The focal mechanisms for all earthquakes shown here are
474 available in the GCMT catalog.

475 Between September 26, 2022 and January 30, 2023, 125 events yield stable and robust
476 moment-tensor results, using the same quality criteria used in the standard GCMT processing. Fig.
477 1a shows the focal mechanisms of the 125 earthquakes plotted at their reported epicenters. A
478 majority of the events ($n = 113$) have normal-faulting geometries, with the remaining events ($n =$
479 12) having reverse mechanisms. The fault strikes of both normal and reverse mechanisms are
480 generally North–South, with some variability. Fault dips also show variability. It is likely that
481 much of the variation can be attributed to uncertainties in the moment-tensor estimates as many
482 events are small ($M < 5$) and the dip-slip components of the moment tensor are difficult to constrain
483 for shallow sources using long-period seismograms⁵⁵.

484 The 47 earthquakes shown in Fig. 1b occurred between August 10, 1990 and June 14, 2019
485 on the Carlsberg Ridge. These have been analyzed previously as part of the GCMT project. Here

486 we relocated them using the surface wave cross-correlation method described above, again
487 yielding a clear central band of extensional events (Supplementary Table 1). All earthquake
488 locations were shifted by 0.0581° (~ 6 km) to the East and by 0.0436° (~ 5 km) to the North to align
489 the central band with the ridge axis.

490

491 **Hypocenter depths for the 2022 Mid-Atlantic Ridge swarm**

492 Hypocenter depths are not well constrained in standard location algorithms that mainly rely on
493 first-arriving P waves. Uncertainties are typically on the order of 10 km. The long-period body and
494 surface waves used in the CMT analysis are similarly insensitive to the earthquake depth. To obtain
495 better depth estimates, we examine the P wave waveforms recorded at teleseismic distances. At
496 these distances, the first-arriving P wave is followed by phases reflected off the surface of the
497 Earth, or off internal boundaries above the source. The delay times of the most prominent
498 reflections pP and sP can sometimes be used to infer a depth when the duration of the source is
499 shorter than the difference in the travel time of the two phases. For very shallow earthquakes, the
500 reflected phases overlap the direct phase, creating a complicated waveform.

501 We examined P-wave waveforms for the M5.7 normal-faulting earthquake on September
502 26 and the three largest reverse-faulting earthquakes on September 29 (M5.8), October 1 (M5.9)
503 and November 29, 2022 (M5.5). The waveforms show P-wave polarities consistent with the
504 GCMT focal mechanisms. When the pulses are filtered to reflect ground displacement it becomes
505 clear that at many stations the initial P-wave onset is followed nearly immediately (within 1–3
506 seconds) by an arrival of opposite polarity. We interpret this as interference by ocean-bottom and
507 ocean-surface reflected phases. A delay of ~ 1 second between the P-wave onset and the change
508 of displacement polarity observed at some stations is consistent with very shallow earthquake
509 depth (< 5 km). To obtain a more robust depth estimate it is necessary to model the interference of
510 the direct and reflected phases. We model the broadband teleseismic P wave using the method of
511 ref. 56. In this technique, the broadband P-wave waveforms are synthesized in a layered elastic
512 structure including a water layer. The depth, focal mechanism, and moment-rate function are
513 varied to obtain, in a least-squares sense, the best fit between observed and synthetic waveforms.
514 The CMT solution is used as a soft constraint on the focal mechanisms to stabilize the inversion.
515 We used a simple elastic model with an eight-kilometer-thick oceanic crust overlain by an ocean
516 layer. We estimated the water depth at the locations of each of the four earthquakes from the
517 bathymetry. For the normal-faulting earthquake, we used a water depth of 2000 meters (bottom of
518 the axial valley), and for the reverse-faulting earthquakes we used depths between 1200 and 1500
519 meters (top of the ridge flanks).

520 The results from the broadband P-wave inversions are consistent with the qualitative
521 interpretation of the polarity change as resulting from early reflected phases. The best-fitting point-
522 source focal depths below the ocean surface range between 2.7 and 4.2 km, corresponding to
523 depths in the top 3 km in the crust. In the inversion, these depths are primarily constrained by the
524 early part of the waveform, and the observed early polarity change. Since the earthquake durations
525 are longer than the delay between the direct and reflected phases, the depth of the full rupture may

526 be larger than the estimated point-source depth. The inversion results constrain the initial part of
527 the rupture to have occurred at very shallow depth in the crust (<3 km). The later part of the rupture
528 may have extended deeper. A summary of our depth estimates is included in Supplementary Table
529 1.

530

531 **Fault strain from bathymetric analyses**

532 To estimate the fraction of plate separation T taken up by slip on major faults, we use a standard
533 method to calculate the cumulative vertical displacement (throw) of faults^{22–24}. This method
534 consists of identifying fault scarps in ridge-normal transects of high-resolution shipboard
535 bathymetry and summing their vertical offsets as a function of along-profile distance. We
536 specifically select steep axis-facing slopes bounded by sharp breaks (Fig. 1c, d), to minimize
537 confusion with volcanic features⁵⁷. Because of mass wasting and other seafloor reworking
538 processes^{41, 42}, the vertical offset on a scarp is generally considered a more reliable estimator of
539 vertical fault offset than its horizontal extent⁵⁸. We therefore use scarp throw as a proxy for fault
540 throw, and divide it by the tangent of the assumed fault dip to assess fault heave. Numerical models
541 suggest that MOR faults rapidly rotate from their steep initial angles ($>50^\circ$) to shallower dips closer
542 to 45° (59), consistent with the focal mechanisms in our catalog. We therefore assume a dip of 45°
543 for all faults in this study. We then construct plots of the cumulative horizontal offset on faults as
544 a function of distance to the axis (the position of the mid-point of each fault scarp). These typically
545 show a linear trend with slope T close to the axis, and a sharp decrease in slope ($T' < T$) ~ 15 km
546 off-axis. It should be noted that assuming a dip of 45° only affects the absolute value of T , which
547 can be straightforwardly adjusted to account for a different fault dip. We use a non-linear least-
548 squares method to fit a continuous, piece-wise linear function with slopes T and T' before and after
549 a cutoff-distance x_c , with y-intercept y_0 , to all our plots of cumulative offset vs. distance. The only
550 constraint enforced on T , T' , and y_0 is that they are positive, while x_c must be greater than the
551 distance of the first fault to the axis. Bathymetric data for the 54°N segment of the Mid-Atlantic
552 Ridge is from ref. 50.

553

554 **Candidate unfauling events across the global MOR system**

555 We searched the Global CMT catalog³⁵ for reverse-faulting events indicative of ridge-normal
556 compression within ~ 50 km of a MOR axis²⁰. We retained those located in areas with sufficient
557 shipboard bathymetric coverage —as compiled in the NCEI (NOAA) and GMRT (LDEO/NSF)
558 repositories, and by additional sources listed in the corresponding figure captions— to perform the
559 fault strain analysis described in the above section. These earthquakes are plotted in Fig. 4b and
560 listed in Supplementary Table 2. They occurred in 13 sections of the Southwest Indian Ridge
561 (SWIR), Carlsberg Ridge, Mid-Atlantic Ridge (MAR), and Southeast Indian Ridge (SEIR),
562 spanning ultraslow, slow and intermediate spreading rates. Corresponding maps are shown in Fig.
563 1 and Extended Data Figs. 1 and 3. Ten of these 13 sections feature regularly spaced abyssal hills
564 bounded by steep axis-facing normal faults, and can be unambiguously classified as magmatically-
565 robust following the criteria of ref. 21. The other three (Extended Data Fig. 3d, e, h) feature

566 corrugated surfaces typical of large-offset detachment faults, although the reverse-faulting
567 earthquakes typically occurred beneath a terrain dissected by steep, closely-spaced faults (within
568 a typical location uncertainty of 10–20 km). This is evident from the bathymetric profiles we
569 gathered in each location, which are plotted in Extended Data Figs. 4 and 5. Results of our strain
570 analyses are compiled in Supplementary Table 3. All sections showed evidence for unfauling,
571 with $T-T'$ averaging 0.09 ± 0.04 (Fig. 4c).

572

573 **Lack of unfauling near an axial high: the East Pacific Rise at 9°30'N**

574 We performed one additional strain analysis in a bathymetric profile across the fast-spreading East
575 Pacific Rise (EPR) at 9°30'N (Extended Data Fig. 9). Of all the ridge sections we analyzed, this is
576 the only one where the non-linear least-squares fit yielded a value of T' (0.04) greater than T (0.01).
577 This is consistent with the model of ref. 34, in which lithospheric unbending away from an axial
578 high results in shallow extension. This configuration can be viewed as reciprocal to the shallow
579 compression induced by unbending away from an axial valley, which drives unfauling (Fig. 3a).
580 It further suggests that the decrease in apparent T fraction observed in all profiles that straddle
581 ridge valley shoulders cannot be solely due to seafloor reworking processes such as mass wasting
582 or sedimentation. This is because these processes are likely active at the EPR, where the apparent
583 T fraction increases away from the ridge axis.

584

585 **Numerical modeling of mid-ocean ridge relief and stress**

586 Using the approach of ref. 30 we conducted 2-D numerical simulations of magmatic injection and
587 fault growth at an idealized mid-ocean ridge with a fixed thermal structure. This approach builds
588 on a series of spreading center models^{7, 27, 29} that used FLAC (Fast Lagrangian Analysis of
589 Continua), an explicit hybrid finite-element and finite-difference technique, to solve the equations
590 of mass, momentum and energy conservation in a visco-elastic-plastic continuum. This method is
591 well suited to simulating localized deformation, approximating faulting, and is described in detail
592 elsewhere^{60–62}.

593 We consider deformation in a 2-D model domain 150 km wide and 20 km deep (Fig.
594 3a). The top boundary is stress free and the bottom boundary is a Winkler foundation, simulating
595 flotation on an inviscid substrate with a mantle density. The sides are pulled at a constant velocity
596 (full rate: $2U = 2.5$ cm/yr) and crustal material is added via diking and lower crustal intrusion at
597 the model center as described below. Regridding of the distorted Lagrangian mesh occurs
598 regularly, and returns the base and sides of the model domain to their original positions. Mantle
599 material is added or subtracted at the base during re-gridding.

600 Both the crustal and mantle material are assigned a visco-elastic–plastic rheology with a
601 viscosity that reflects a power law relating strain rate to the differential stress to the power $n = 3$,
602 and an Arrhenius dependence on temperature. The parameter values are those of dry diabase⁶³,
603 which ensures that the region cooler than $\sim 600^\circ\text{C}$ (lithosphere) behaves essentially elasto-
604 plastically. Brittle-plastic deformation is described by a Mohr-Coulomb failure criterion with a
605 friction coefficient of 0.63 and initial cohesion of 25 MPa. When this yield criterion is met, the

606 material weakens with strain beyond yield, leading to the localization of brittle deformation in
 607 model fault zones. Following previous work that simulated both bending and stretching related
 608 faulting²⁷, we consider 2 phases of strain weakening. 10 MPa of cohesion loss occurs over 1% of
 609 plastic strain, followed by 13 MPa of additional cohesion loss over 30% of strain. Our simulations
 610 do not account for the possibility of tensile failure. The model elements are 250 m wide and the
 611 fault zones in the model are typically ~ 4 elements wide⁶². The amount of fault slip necessary for
 612 total fault weakening (30% strain) is thus ~ 300 m. Both the crust and mantle are assigned a
 613 Young's modulus of 30 GPa and a Poisson's ratio of 0.25.

614 For simplicity, the thermal structure was fixed for the model shown in Fig. 3a, although
 615 similar models allowing self-consistent evolution of the thermal structure were used by ref. 31 to
 616 compare simulations with observed axial relief and faulting patterns as functions of spreading rate
 617 and crustal thickness. For the case shown in Fig. 3a, the seafloor was set to 20°C and the depth of
 618 the 600°C isotherm is set to 5 km at the ridge axis, and increases to 7 km, 40 km away from the
 619 axis. Temperature increases linearly with depth from the seafloor to the 600 °C isotherm. It also
 620 increases linearly between the 600°C and 1300°C isotherms, but with five times the vertical
 621 temperature gradient. Temperature is capped at 1300°C throughout the deeper parts of the model.

622 Dike intrusions accommodate the separation of plates at the spreading axis with a uniform
 623 average opening rate described by the M parameter (the fraction of plate spreading accommodated
 624 by diking as defined by ref. 27), such that the dike opening rate in the numerical model is defined
 625 as $2UM$. Instead of using a constant value for M , we follow the analysis of ref. 30 and relate it to
 626 crustal thickness (H_C) and on-axis lithosphere thickness (H_L) as follows:

$$627 \quad M = \frac{H_C}{H_L + 0.5H_G}. \quad (1)$$

628 In the above equation, H_G is the thickness of the magma column intruded into the asthenosphere
 629 due to the excess pressure associated with the development of the axial relief³⁰, and can be
 630 expressed as:

$$631 \quad H_G = 0.25H_L + 0.5 \sqrt{\frac{9}{4}H_L^2 + \frac{6P_{DL}H_L}{(\rho_c - \rho_f)g}}, \quad (2)$$

632 where ρ_c is the density of the crust, ρ_f is the density of the magma, and P_{DL} is the driving pressure
 633 needed to open a meter-wide dike. In the simulation shown in Fig. 3a, H_C is 7 km, ρ_c and ρ_f are
 634 set to 3000 kg/m³ and 2700 kg/m³, and P_{DL} is 10 MPa. We further prescribe minor oscillations in
 635 the amount of magma delivered to dikes, implemented as:

$$636 \quad H_C(t) = H_C + \Delta M H_C \sin(2\pi \omega t). \quad (3)$$

637 In equation (3), ΔM and $1/\omega$ are the amplitude and period of the magma supply fluctuations, set to
 638 0.15 and 0.2 Myr respectively. $H_C(t)$ is then used in equation (1) to compute temporal fluctuations
 639 in M . These small fluctuations are set to enable the regular growth of normal faults that are evenly
 640 spaced by ~ 3 km.

641

642 **Elastic stress changes caused by a dike intrusion**

643 We calculate the stress changes caused by 1 meter of uniform opening on a vertical, 6 km tall by
 644 50 km long rectangular dislocation embedded in an elastic half-space. The calculation is carried

645 out with the MATLAB routines of ref. 64, which allow us to mesh the dike with two triangular
646 dislocations. The top of the dike is set at 4 km below the surface. The Young's modulus and
647 Poisson's ratio of the material are set to 10 GPa and 0.25, respectively⁶⁵. We compute the change
648 in Coulomb Failure Stress⁶⁶ on 45°-dipping receiver faults striking parallel to the dike, assuming
649 a friction coefficient of 0.6. Wherever this stress change is positive, faults are brought closer to
650 failure. We assess their favored sense of slip by calculating their resolved shear stress, with the
651 convention of negative shear stress indicating reverse motion. Results are shown in Fig. 3b.
652 Because of the linearity of the problem, all stresses can be thought of as normalized by meter of
653 opening on the axial dike. We note that the compressive stress changes imparted by diking across
654 the ridge shoulders are ~100 times smaller than the absolute compressive stresses imparted by
655 unbending (Fig. 3). Dike-induced compression is thus not sufficient to reverse the sense of slip on
656 abyssal hill bounding faults, but could plausibly precipitate failure on faults already close to
657 compressive yielding. Finally, it is important to note that the assumption of a homogeneous elastic
658 half-space is warranted for rapid deformation events, but likely overestimates stress changes
659 particularly if low-viscosity lower crust / asthenosphere can rapidly relax some of these stresses.

660

661 **Data Availability**

662 All bathymetric data used in this study are from the published literature as referenced^{50, 67, 68}, or
663 openly available in the GMRT (<https://www.gmrt.org/>) / NOAA-NCEI
664 (<https://www.ncei.noaa.gov/maps/bathymetry/>) repositories. The fault scarp dataset of ref. 24 is
665 provided in Supplementary Table 4. Earthquake data are from the CGMT Catalog
666 (<https://www.globalcmt.org/CMTsearch.html>), except for earthquake relocations, which are
667 provided in Supplementary Table 1.

668

669 **Code Availability**

670 The simulation shown in Fig. 3a was run with the version of the FLAC code⁶⁰ developed by refs.
671 30 and 31. This code and the corresponding visualization scripts are available from the
672 corresponding author upon reasonable request. The stress calculations shown in Fig. 3a were done
673 with the code openly distributed with ref. 64.

674

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745

746 **Author contributions**

747 J.-A.O. designed the study, conducted the bathymetric analyses with J.E. and M.B., carried out the
748 elastic stress modeling, and wrote the initial manuscript. G.E. compiled and analyzed the
749 earthquake data. W.R.B. and Z.L. designed the models of ridge flank flexure. All authors discussed
750 and analyzed the results, and provided feedback on the manuscript.
751

752 **Additional information**

753 Supplementary Information is available for this paper. Correspondence and requests for materials
754 should be addressed to J.-A.O. at olive@geologie.ens.fr. Reprints and permissions information is
755 available at www.nature.com/reprints. The authors declare no competing interests.
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760 **Extended data figure captions**

761

762 **Extended Data Figure 1 | Examples of near-axis compression at mid-ocean ridges. (a)**
763 Bathymetric map of the Southeast Indian Ridge (SEIR) near 115°E, with focal mechanisms from
764 the CMT catalog. **(b)** Bathymetric map of the Mid-Atlantic Ridge (MAR) south of the Marathon
765 transform fault (TF), with focal mechanisms from the CMT catalog. **(c)** Bathymetric profile across

766 the SEIR (dashed line in panel a), with colored segments indicating fault scarps. **(d)** Cumulative
767 fault heave vs. distance from the axis for the southern (red) and northern (yellow) sides of the
768 SEIR, with best-fitting piece-wise linear functions shown as dashed lines ($T = 0.13$, $T' = 0.05$ on
769 the South side; $T = 0.09$, $T' = 0.07$ on the North side). **(e)** Bathymetric profile across the western
770 flank of the MAR (dashed line in panel b), with colored segments indicating fault scarps. **(f)**
771 Cumulative fault heave vs. distance from the axis for the western side of the MAR, with best-
772 fitting piece-wise linear function shown as dashed lines ($T = 0.21$, $T' = 0.08$).

773

774 **Extended Data Figure 2 | The 2022 Northern Mid-Atlantic Ridge and 2014 Carlsberg Ridge**
775 **seismic sequences.** Latitude and moment magnitude of earthquakes, color-coded by mechanism
776 (blue: reverse faulting, red: normal faulting) and location, throughout the **(a–d)** 2022 MAR
777 sequence and **(e–h)** 2014 Carlsberg Ridge sequence.

778

779 **Extended Data Figure 3 | Examples of near-axis reverse-faulting earthquakes at mid-ocean**
780 **ridges.** Bathymetric maps, reverse focal mechanism and profile used for strain analyses at **(a)**
781 MAR 0°50'N near the St. Paul transform fault^{67, 68}; **(b)** MAR 7°50'S; **(c)** MAR 34°50'N; **(d)**
782 Carlsberg Ridge 9°45'N; **(e)** MAR 1°20'S; **(f)** SWIR 18°30'E; **(g)** MAR 17°S; **(h)** SWIR 43°E near
783 the Discovery II transform fault; and **(i)** SWIR 57°E.

784

785 **Extended Data Figure 4 | Bathymetric signatures of unfauling in the Atlantic Ocean.** Left:
786 Bathymetric cross sections with fault scarps highlighted in color. Right: Plots of cumulative fault
787 heave vs. distance at selected sections of the Mid-Atlantic Ridge. Corresponding values of T , T' ,
788 and x_c are listed in Supplementary Table 3.

789

790 **Extended Data Figure 5 | Bathymetric signatures of unfauling in the Indian Ocean.** Left:
791 Bathymetric cross sections with fault scarps highlighted in color. Right: Plots of cumulative fault
792 heave vs. distance at selected sections of the Southwest Indian Ridge (SWIR) and Carlsberg Ridge.
793 Corresponding values of T , T' , and x_c are listed in Supplementary Table 3.

794

795 **Extended Data Figure 6 | Unfauling along the intermediate-spreading Chile Ridge. (a)**
796 Bathymetric map of the Chile Ridge axis outlining its major segments, adapted from ref. 24. Insets
797 detail bathymetry of segments N1 and N9N–N9S. **(b)** Position and **(c)** amplitude of the change in
798 apparent T at the Chile Ridge: histograms of x_c and $(T-T')$ for all transects across the Chile Ridge,
799 excluding poor fits highlighted in gray in Supplementary Table 3. **(d)** Average slope of axis-facing
800 scarps vs. distance to the ridge axis. Data are from the fault scarp compilation of ref. 24.

801

802 **Extended Data Figure 7 | Tectonically-accommodated strain along the Chile Ridge.**
803 Cumulative fault heave vs. distance from the axis along individual transects from each segment of
804 the Chile Ridge, based on the fault scarp compilation of ref. 24. Red and yellow dots correspond
805 to the western and eastern side of the axis, respectively, with best-fitting piece-wise linear

806 functions shown as black lines. This excludes poor fits highlighted in gray in Supplementary Table
807 3.

808

809 **Extended Data Figure 8 | Amplitude and position of the change in apparent T along the Chile**

810 **Ridge. (a–i)** Histograms of the amplitude of the change ($T-T'$) in apparent T in each bathymetric
811 transect, grouped by segment. **(j–r)** Histograms of the distance x_c where the change in apparent T
812 occurs in each bathymetric transect. These plots exclude poor fits highlighted in gray in
813 Supplementary Table 3.

814

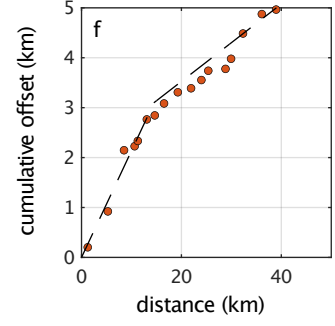
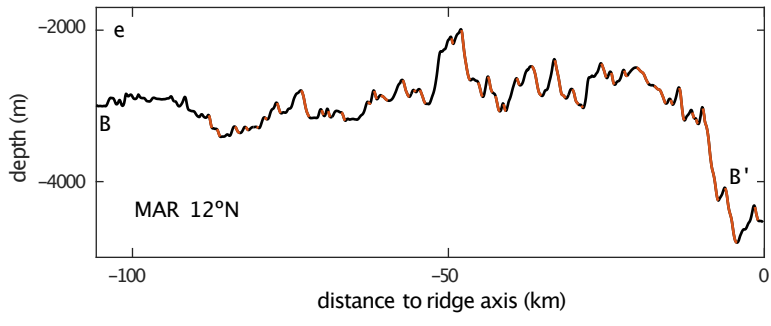
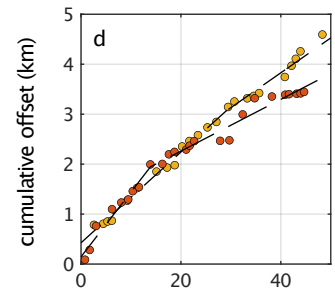
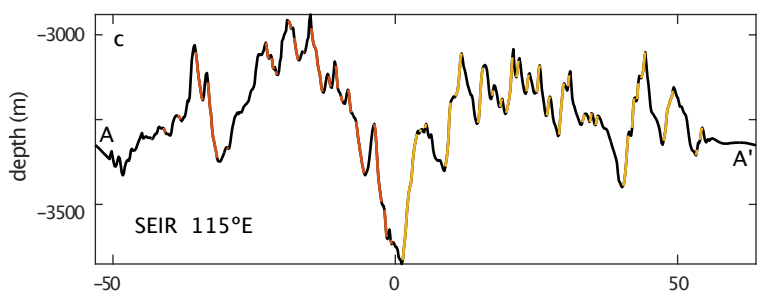
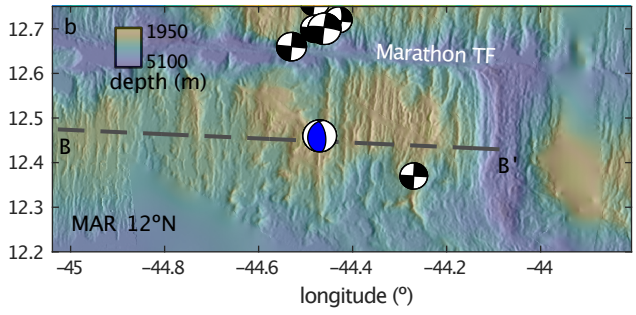
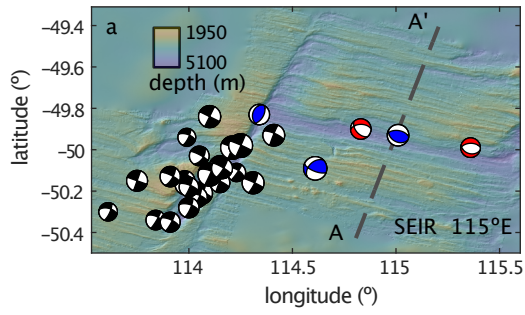
815 **Extended Data Figure 9 | Lack of unfaulting near an axial high. (a)** Bathymetric map of the

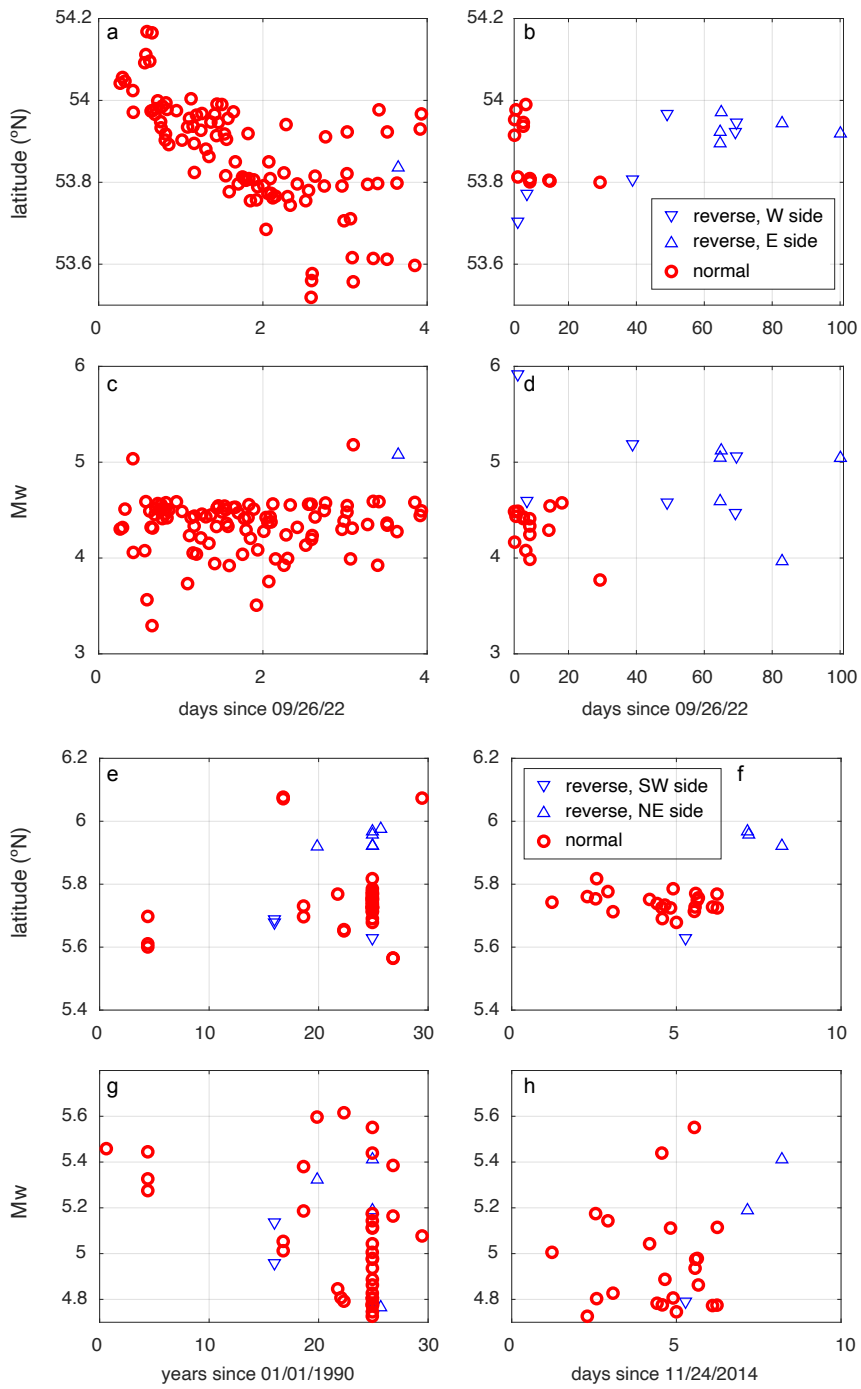
816 fast-spreading East Pacific Rise at 9°30'N, which unlike every other ridge section studied here
817 features an axial high instead of an axial valley. White line indicates location of bathymetric
818 transect. **(b)** Bathymetric cross sections with fault scarps highlighted in color. **(c)** Cumulative fault
819 heave vs. distance, with best-fitting piece-wise linear function shown as black lines. In this case T'
820 $> T$.

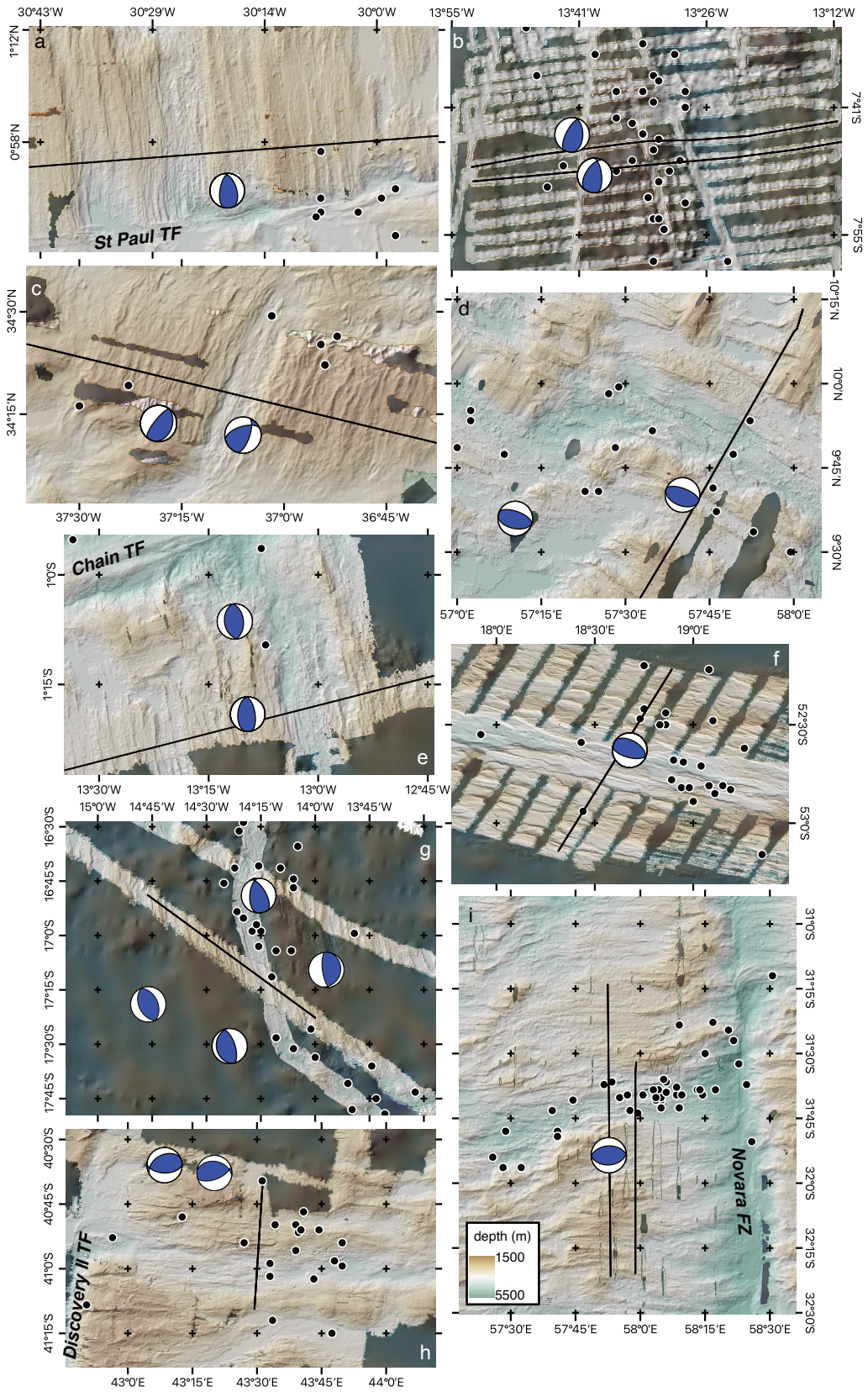
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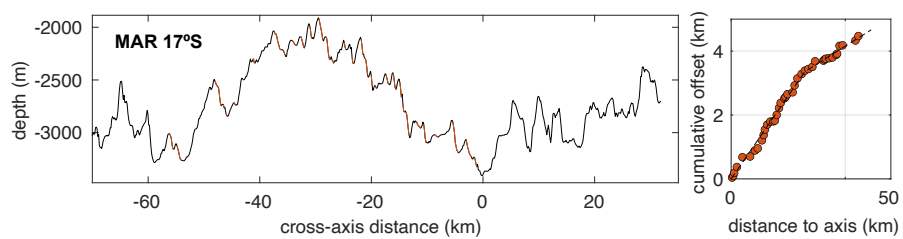
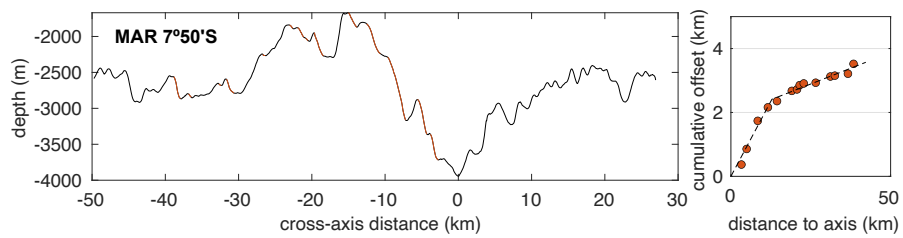
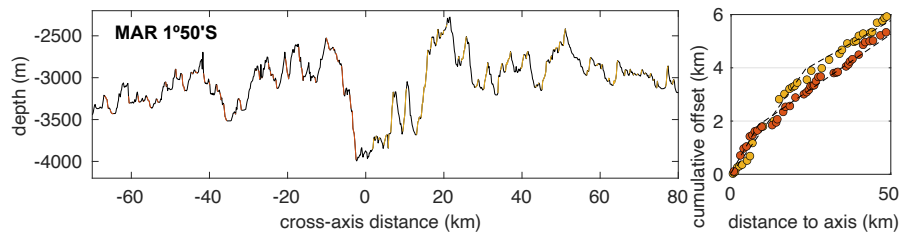
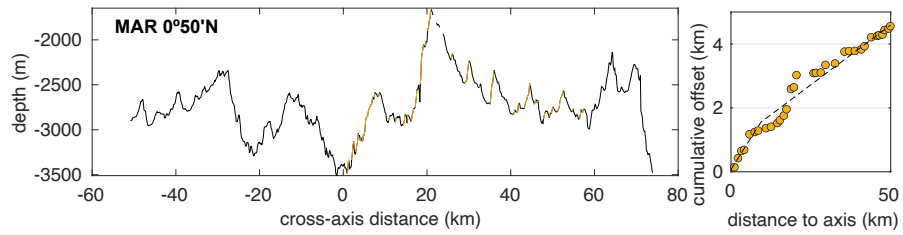
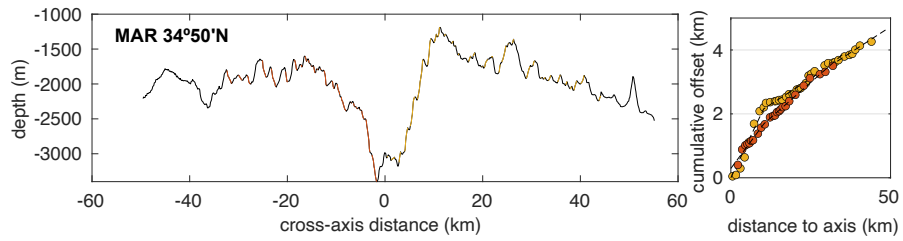
822 **Extended Data Figure 10 | Mechanics of ridge shoulder unbending.** The lithosphere's yield

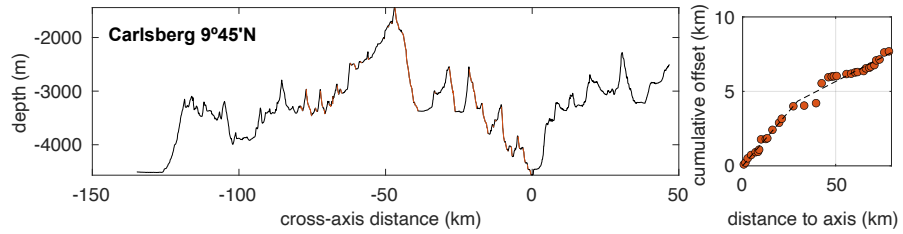
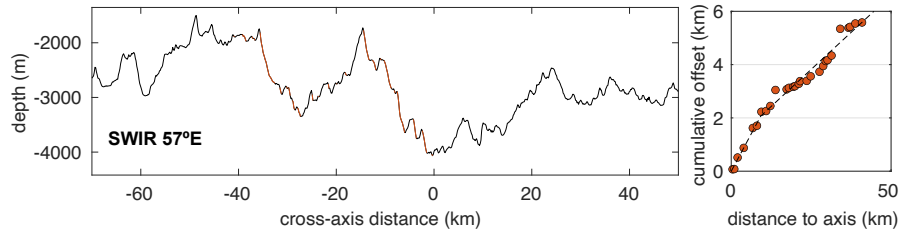
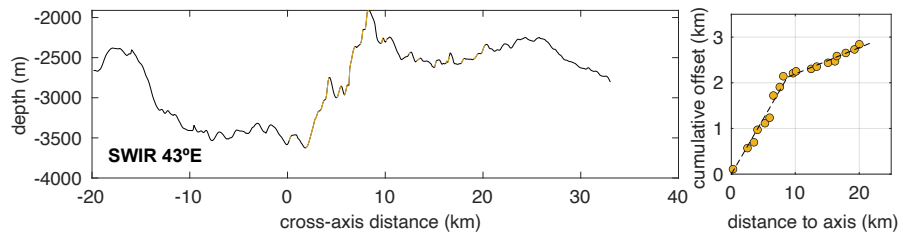
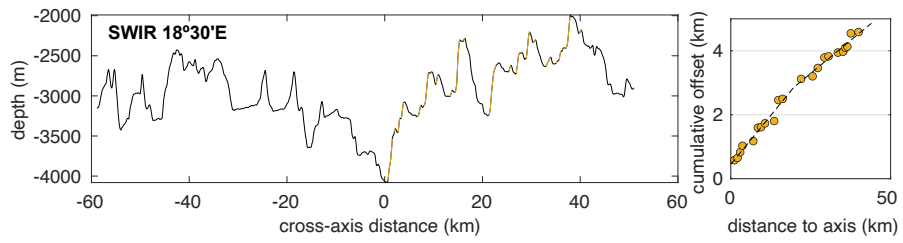
823 stress σ_y is defined as the difference between the vertical and horizontal stresses needed to produce
824 fault slip. **(a)** shows an idealized plate that was accreted with curvature and no bending stresses.
825 σ_0 is an assumed background stress difference. The simple case shown has no cohesion and
826 parameters A and B depend strongly on the friction coefficient f and assumed pore pressure P_P on
827 the faults, as defined in the Supplementary Information text. **(b)** shows a case with reverse faults
828 that are three times stronger than normal faults, while **(c)** illustrates deepening of the neutral depth,
829 D , when the reverse faults are assumed to be weaker than the normal faults. **(d)** Analytical estimate
830 of the neutral depth D , which marks the base of the compressive zone in an unbending ridge
831 shoulder. The ratio of D to the layer thickness, H , is plotted versus the ratio of pore pressures on
832 reverse versus normal faults assuming $f = 0.75$ and that the pore pressure on normal faults is 1/3
833 the lithostatic pressure (Blue curve). Assuming a rock density of 3000 kg/m³ and water density of
834 1000 kg/m³, the left limit is for hydrostatic pore pressure on the reverse faults while the right limit
835 is for lithostatic pore pressure on the reverse faults. Red curve shows the effect on the neutral depth
836 of a regional horizontal extensional stress difference equal to 20% of the extensional yield stress
837 at the base of the layer. See Supplementary Information text for details.

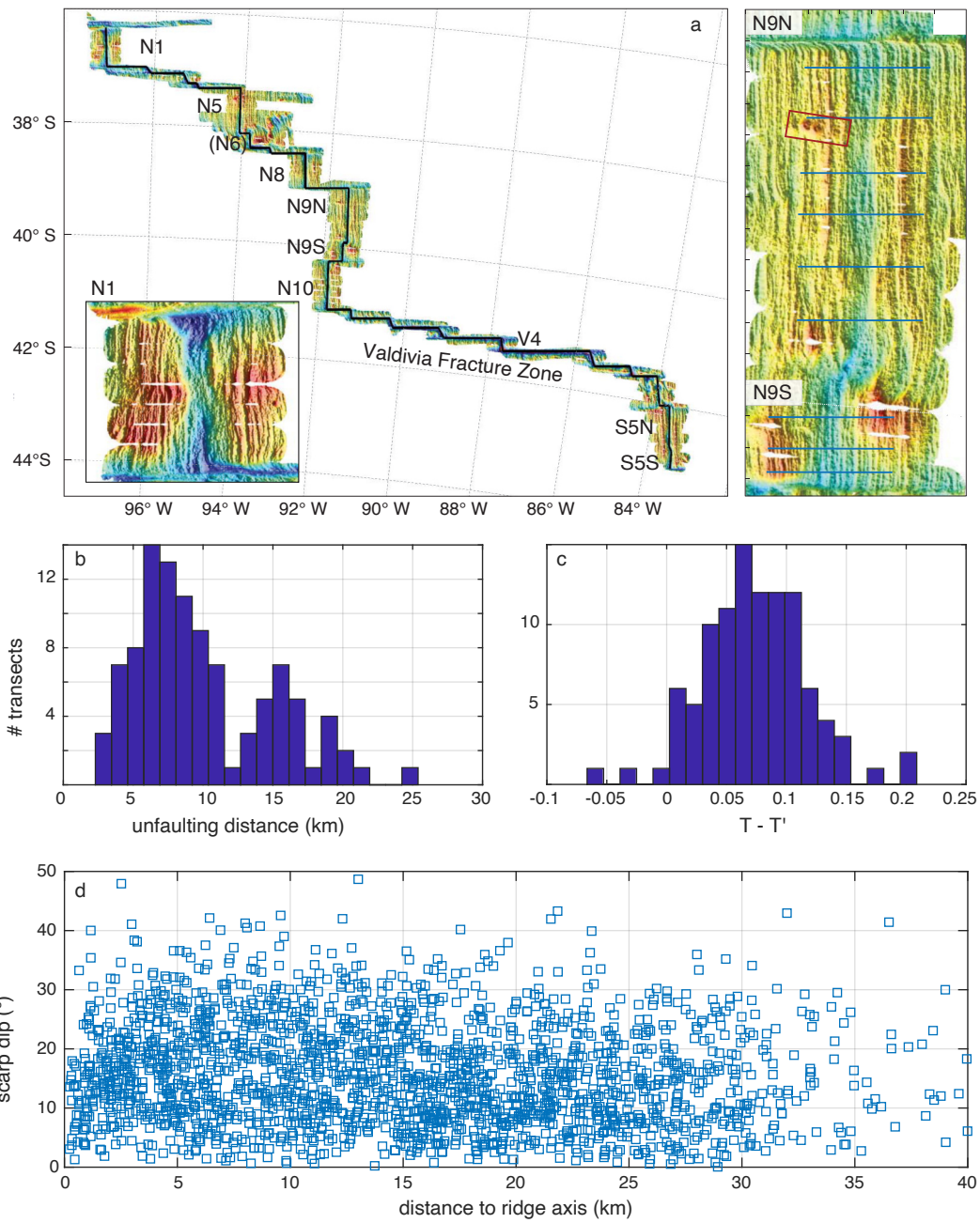


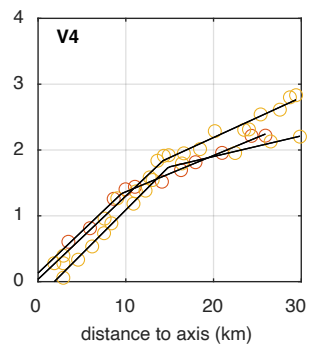
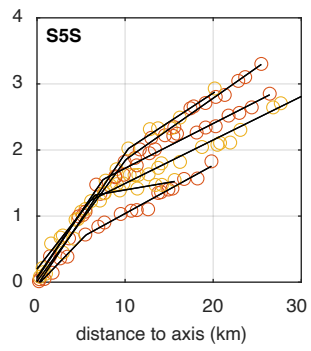
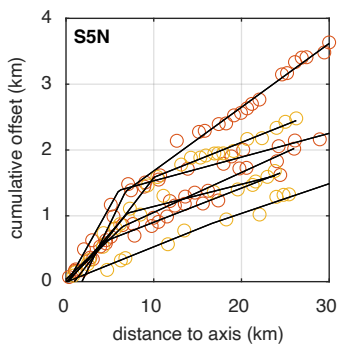
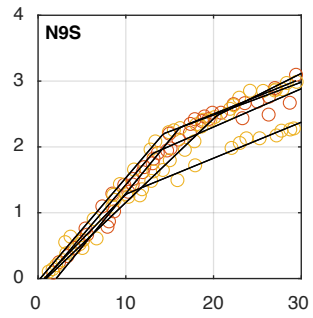
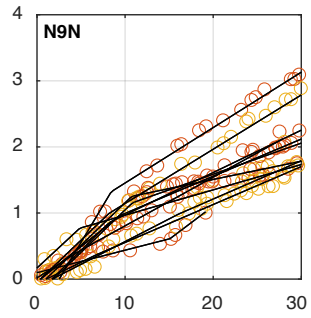
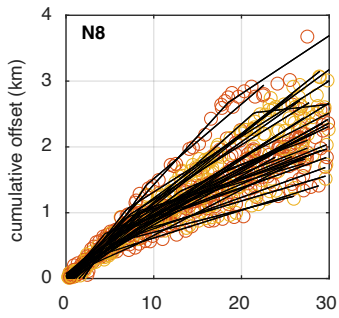
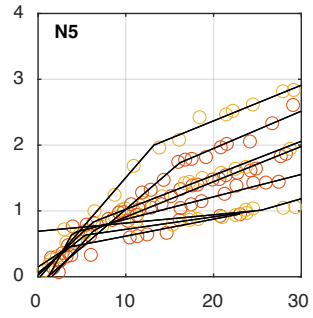
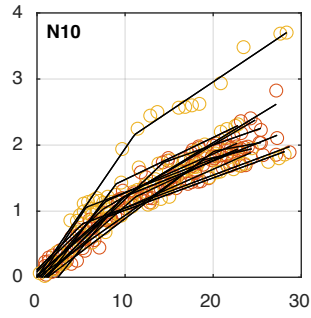
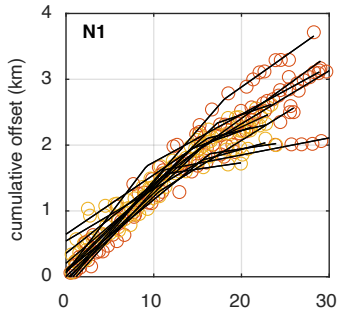


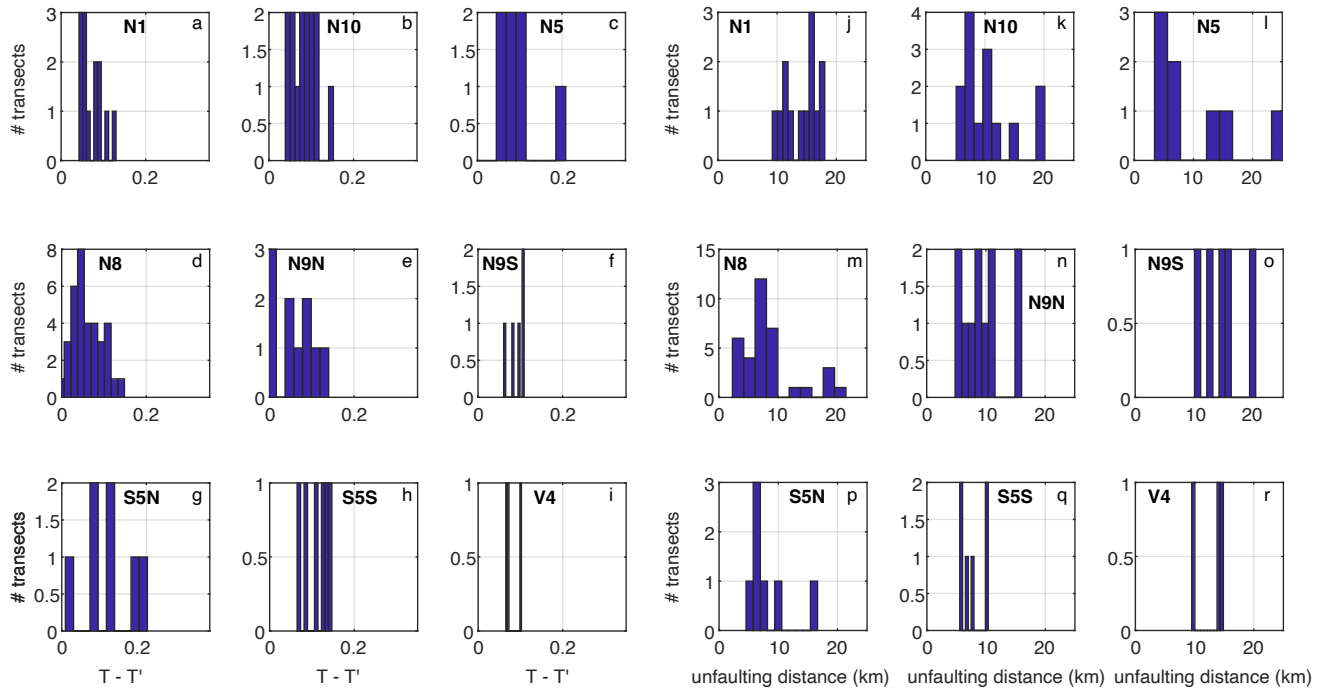


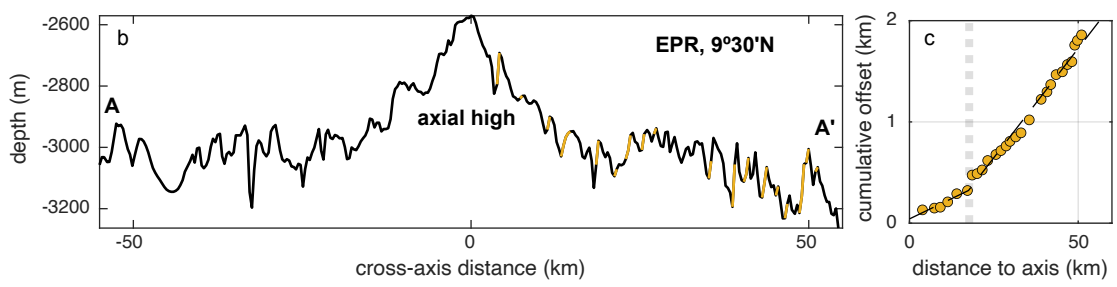
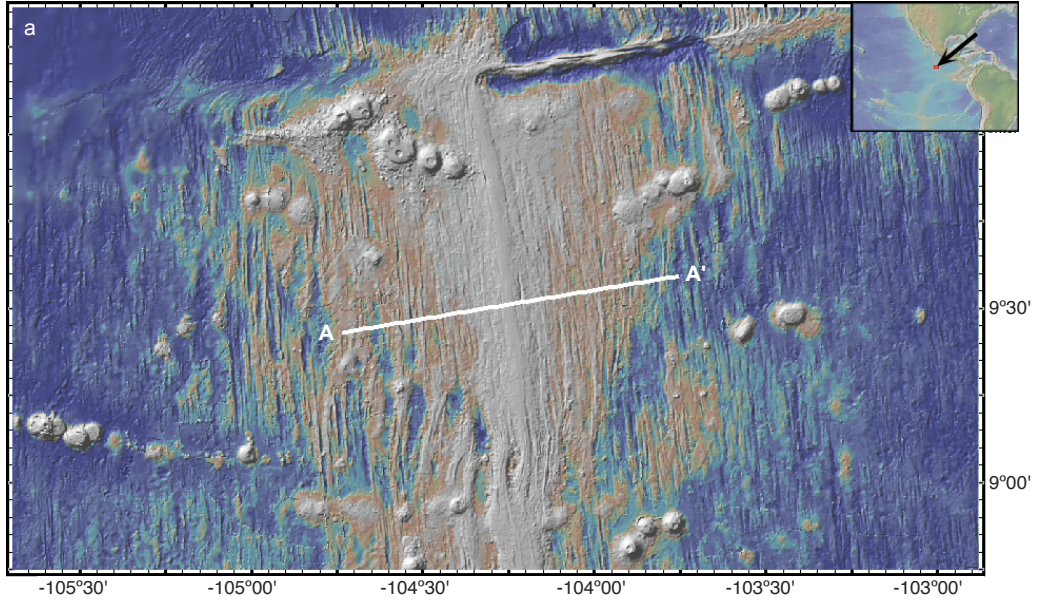




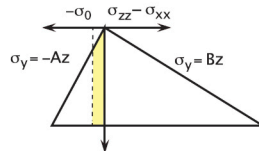




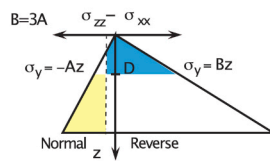
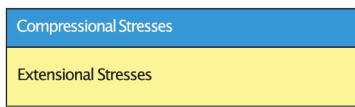




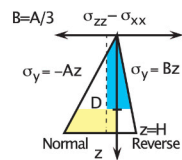
a. Layer accreted with zero bending stresses



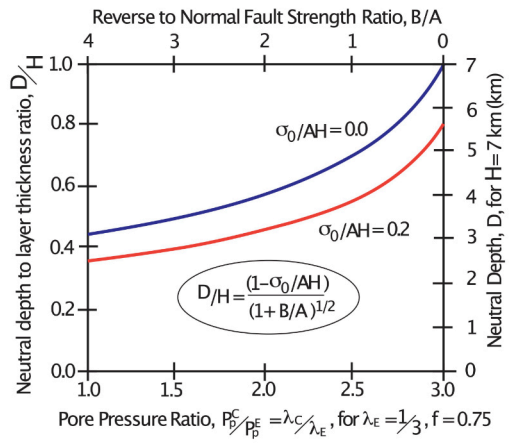
b. Flattened layer with strong reverse faults



c. Flattened layer with weak reverse faults



d. Depth extent of ridge shoulder compression



Supplementary Information for "Mid-ocean ridge unfaulting revealed by magmatic intrusions"

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Content

This supplementary file contains:

- A description of Supplementary Tables 1–4.
- An extended Methods section describing the fault scarp dataset from the Chile Ridge, and detailing a simple analytical model for the depth of the compressive zone taking into account the effect of pore fluid pressure.

Description of supplementary tables

Supplementary Table 1 | Relocated earthquakes at the Mid-Atlantic and Carlsberg Ridge.

Event code follows convention of GCMT Catalog. Table also includes depth estimates for 4 events from the 2022 MAR 53°N earthquake swarm.

Supplementary Table 2 | Unfaulting earthquakes.

List of reverse earthquakes from the GCMT catalog that occurred in areas mapped with high-resolution shipboard echo-sounders. Events labeled JM_X are part of the compilation from Jackson & McKenzie (2023).

Supplementary Table 3 | Compilation of unfaulting signatures in high-resolution seafloor bathymetry.

Best-fitting slope T of cumulative fault offset vs. distance to the ridge axis near the axis (T best, with maximum and minimum estimates), and away from the axis (T'), past critical distance x_c . Each row is a bathymetric transect on a specified side of the ridge axis. Shaded rows are transect that did not yield a satisfactory fit with a piece-wise linear function.

Supplementary Table 4 | Normal faults along the Chile Ridge. Original dataset from Howell et al. (2016). Each fault is identified along a ridge-normal transect by the position of the bottom and top of its seafloor scarp (locations 1 and 2). Positive and negative distance to the ridge axis correspond to fault located west and east of the axis, respectively. Segments are numbered as follows (#1–#9): N1, N10, N5, N8, N9N, N9S, S5N, S5S, and V4.

Supplementary Methods

Bathymetric evidence of unfauling without seismic manifestation: the Chile Ridge

A dataset of fault scarps previously identified along the entire Chile Ridge by Howell et al. (2016) (Supplementary Table 4) was also used to construct Extended Data Figs. 6–8. In Fig. 2, faults from both sides of the ridge are fitted together, while in Extended Data Figs. 1, 4, 5, 7, faults from each side of the ridge are fitted separately. We find that this choice has little influence on the results, which are summarized in Supplementary Table 3. At 12 transects out of 114 at the Chile Ridge, the non-linear least-squares method yielded unrealistically wide bounds on T and x_c . We interpret these as indicative of a non-robust fit (highlighted in gray in Supplementary Table 3), and exclude the corresponding transects from the plots shown in Extended Data Figs. 6–8. Overall, the mean unfauling strain across the Chile Ridge is 0.074 (± 0.044 of standard deviation, Extended Data Fig. 6), and this value is representative of all segments (Extended Data Fig. 8a–i). The unfauling distance displays a bimodal distribution, with peaks at ~ 7 and ~ 16 km (Extended Data Fig. 6b). This distribution reflects segments with a short unfauling distance (~ 5 – 10 km), such as N8, N9N, S5S and S5N, and segments with x_c between 10 and 20 km, such as N1, N9S, and V4. Segments N10 and N5 show a wider range in x_c (Extended Data Fig. 8j–r).

Extended Data Fig. 6d further shows no distinct change in the slopes of axis-facing (abyssal hill bounding) fault scarps as they migrate off axis past the unfauling distance. This suggests that the change in apparent T fraction identified in our fault strain plots is unlikely due to slope-reworking mass wasting processes, and instead have a tectonic origin that affects fault offset rather than scarp slope.

An analytical model for ridge flank compression modulated by pore fluid pressures

Analytical and numerical models suggest that the flattening of axial bathymetric relief occurs via lithospheric un-bending (e.g. Buck, 2001; Liu and Buck, 2018). The prefix “un-” is used because analysis suggests that lithosphere at spreading centers accretes in a “bent” state (i.e., with curvature “built in”). For ridges with an axial high (Extended Data Fig. 9), as at most fast-spreading centers, the initial curvature is concave up. For ridges with an axial valley, as at most slow-spreading centers, this curvature is concave down. The un-bending flanking an axial valley puts the top of the plate in compression and the base of the plate in extension.

Many tectonic models make a simplifying assumption that normal and reverse faults are subject to hydrostatic pore pressures (e.g., Brace and Kohlstedt, 1980). Models of axial valley formation that make this standard assumption predict a very shallow depth extent of reverse faulting during

unbending (e.g., ~ 2 km in Fig. 3). This may be a problem, in terms of explaining the observations discussed in this paper, for two reasons. First, it is hard to reconcile the large magnitude of the 2022 reverse faulting earthquakes on the MAR, up to magnitude 5.9, with a small depth extent of reverse fault slip. The moment magnitude should scale with fault area and fault slip. The fault area depends on the depth extent and axis-parallel extent of fault slip. Though bending faults appear to be longer than faults that cut a brittle layer (e.g., Supak et al., 2006) it is likely that such large magnitude earthquakes must happen on faults that cut more deeply than ~ 2 km. This is especially true since slip on bending faults should linearly decrease with distance from the top of the strong lithospheric plate and vanish at the “neutral depth” of bending (inset in Fig. 4a). The second problem with the assumption of hydrostatic pore pressures on reverse faults is that it results in a wide zone of un-bending strain with distance from the spreading axis. In the model of Fig. 3, the zone of unbending is ~ 30 km wide. This implies a wide zone of slip on reverse faults, and that would result in a similarly wide zone of reduction in abyssal hill relief. This is counter to the observed < 10 km wide zone of reduction in abyssal hill relief as shown in Fig. 2, as well as most T -fraction vs. distance plots shown in this study.

One possible explanation for these observations is that reverse faults flanking axial valleys are subject to the same high levels of pore pressures apparently affecting reverse faults in convergent settings. The idea of such high pore pressures on reverse faults comes from consideration of seismic reflection observations in some mountain ranges showing thrusting of several hundred-kilometer-wide packages of crustal rocks on almost flat reverse faults, or décollements (e.g., Bally et al., 1966). Hubbert and Rubey (1959) showed that these décollements had to be far weaker than expected to allow such large-scale thrusting. Those authors proposed that high pore water pressure, approaching lithostatic levels, was the likely mechanism allowing reverse slip at very low shear stress. Near-lithostatic pore pressure is also the accepted explanation for the shapes of many accretionary prisms above subducting plates (e.g., Davis et al., 1983). The fact that normal faults are not known to be associated with higher-than-hydrostatic pore pressures (e.g., Axen and Selverstone, 1994) makes mechanical sense because pore pressures greater than the minimum compressive stress should result in fracturing that relieves those high pore pressures.

High pore pressures on reverse faults should make those faults effectively weaker in that they would slip at lower levels of shear stress. Weakening reverse faults relative to normal faults should allow them to slip down to greater depths, as bending should not change the average horizontal stress in a layer. This is because flexure should not affect the horizontal force balance of a floating layer. Un-bending of an initially concave down layer, as shown in Extended Data Fig. 10, puts the top of the layer in compression and the bottom of the layer in extension, but the average stress change should be zero. More precisely, the integral of bending-induced horizontal stresses with respect to depth has to equal zero (e.g., Bodine et al., 1981). Compressional fault slip should occur down to the depth where the bending stress is zero, which is termed the neutral depth, D . Below, we estimate the effect of reverse fault weakening on the neutral depth of bending, using the thin plate flexure approximations (e.g., Watts, 2002).

Thin plate theory relates bending of a layer that is thin compared to the horizontal scale of bending. If h is the elevation of the plate surface relative to some datum then the plate has a curvature of dh^2/dx^2 where x is the horizontal distance from the axis in the direction of spreading. An approximate scaling gives the initial curvature of the plate at $x = 0$ as $2h_0/w^2$ where h_0 is the depth of the axial valley relative to the flanks and w is the half-width of the valley (Buck, 2001). If the plate is flattened out then the change in curvature is about $2h_0/w^2$. Thin plate theory holds that the horizontal strain, ε_{xx} , in a layer experiencing a change in curvature, $\Delta dh^2/dx^2$, varies with vertical distance, z' , from the depth in the plate where the bending strain is zero (the neutral depth D) as:

$$\varepsilon_{xx}(x, z) \cong z' \Delta dh^2/dx^2 \cong (D - z) 2e_0/w^2, \quad (1)$$

where z is depth below the surface.

We first consider the case of a plate of thickness h with uniform elastic properties and show that the bending stresses should be so large that parts of the plate would yield by faulting. The horizontal stress produced by bending is:

$$\sigma_{xx}(x, z) = \frac{E}{(1-\nu^2)} \varepsilon_{xx}(z) \quad (2)$$

where E is Young's Modulus and ν is Poisson's Ratio for the lithosphere. For unbending of axial valley lithosphere of thickness H the maximum compressional stress is at the top of the lithosphere where for uniform properties $z' = H/2$. Taking lithospheric thickness $H = 5 \text{ km}$, valley depth $e_0 = 1000 \text{ m}$, $w = 15 \text{ km}$ and assuming reasonable values for rock elastic constants of $E = 5 \times 10^{10} \text{ Pa}$ and $\nu = 0.25$ gives a maximum compressive stress of $\sim 10^{10} \text{ Pa}$. This is more than 2 orders of magnitude greater than the yield stress for faulting at several kilometers depth, as we will show below. For such elastic bending stress the top of the plate (i.e., shallower than D) is at the yield stress for reverse faulting and below z_N the plate is at extensional yield. In previous work on lithospheric bending the inclusion of brittle yielding decreases the effective horizontal length scale of yielding (e.g., Bodine et al., 1981; Buck, 1988). This effect may help explain the narrow zone of unbending that we infer from the observed patterns of cumulative fault offset with distance from a spreading center.

We follow many workers in defining the brittle yield stress in terms of the stress difference needed to slip on an ideally oriented fault in an Andersonian stress field (e.g., Brace and Kohlstedt, 1980). An Andersonian stress field is one in which the principal stresses are either parallel to or normal to the surface of the lithosphere (Anderson, 1951). Here we use the engineering convention that tension is positive. For compressional faulting the maximum principal stress is vertical (i.e., σ_{zz}) while for extensional faulting the maximum principal stress is horizontal (i.e., σ_{xx}). The ideal fault orientation is taken to give the minimum stress difference for slip on a fault with a friction coefficient f . A detailed derivation of the stress needed for slip (or yielding) on ideally-oriented

faults is given in Turcotte and Schubert (2002). It is common to represent the pore pressure as $P_p = \lambda \rho g z$, where λ is the ratio of pore pressure to vertical stress ($\sigma_{zz} = \rho g z$), ρ is the density of the lithosphere and g is the acceleration of gravity. Then, the brittle yield stress (a stress difference) for normal (extensional) faulting, writes:

$$\sigma_Y^E(z) = \sigma_{zz} - \sigma_{xx} = \frac{-2f(1-\lambda_E)\rho g z}{(1+f^2)^{\frac{1}{2}}+f} = -Az$$

and for reverse (compressional) faulting, assuming $P_p = \lambda_C \rho g z$, is: (3)

$$\sigma_Y^C(z) = \sigma_{zz} - \sigma_{xx} = \frac{2f(1-\lambda_C)\rho g z}{(1+f^2)^{\frac{1}{2}}-f} = Bz$$

This implies that at any depth the ratio of the magnitude of the yield stresses is:

$$\left| \frac{\sigma_Y^C(z)}{\sigma_Y^E(z)} \right| = \frac{B}{A} = \frac{(1-\lambda_C) \left[(1+f^2)^{\frac{1}{2}}+f \right]}{(1-\lambda_E) \left[(1+f^2)^{\frac{1}{2}}-f \right]} \quad (4)$$

The neutral depth is the depth that makes the integral of the bending stresses zero. This can be seen graphically in Extended Data Fig. 10 as requiring the area in blue to match the area in yellow on the yield stress plots. Since there is normal faulting within an axial valley, we expect that the average horizontal stress is non-zero and is extensional. Assuming that the average stress difference is $-\sigma_0$ the neutral depth for strong un-bending resulting in zero net bending stress is:

$$\frac{D}{H} = \frac{(1-\sigma_0/Ah)}{\sqrt{1+B/A}} \quad (5)$$

Extended Data Fig. 10 shows that the neutral depth is strongly dependent on the ratio of compressional to extensional frictional yield stress (i.e., B/A) and on the background stress difference σ_0 . It also shows how D varies with the pore pressure on reverse faults (panel d). For example, if the friction coefficient $f = 0.75$ and the same pore pressure acts on normal and reverse faults then $B/A=4$. Assuming a rock density of 3000 kg/m^3 , hydrostatic pore pressure on a normal fault implies that $\lambda_E = \rho_w/\rho \approx 1000/3000$ so that for $f = 0.75$, $A \approx (1/2)\rho g z$. To have $B/A=1$ then would require that $\lambda_C = 0.83$. This appears reasonable as larger pore pressure ratios have been estimated for convergent setting (e.g., Davis et al., 1983).

Other factors are likely to affect the depth range of unbending-related reverse faulting on the flanks of an axial valley. For example, inclusion of viscoelastic effects may result in a deepening of the maximum depth of thrusting. We expect this because on the long-time interval between bending earthquakes the lower lithosphere can flow viscously to relax stress differences. On the short time interval of an earthquake the lower lithosphere should maintain large stress differences. However, such effects are not simple to quantify since they require a numerical solution with very high spatial and temporal resolution.

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