

1 **Title**

2 Do Coupled Megathrusts Rupture?

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This is a non-peer-reviewed preprint submitted to EarthArXiv.

16 Abstract

17 Megathrust earthquakes are among the most destructive and least predictable natural hazards.
18 Kinematic geodetic coupling models, which identify regions of the plate interface where
19 interseismic strain accumulates, are essential for seismic and tsunami hazard assessment¹⁻⁴. Yet
20 their reliability remains debated: geodetic records are short^{5,6}, offshore resolution remains poor⁷,
21 and earthquake ruptures may propagate dynamically across both creeping and locked slab
22 portions⁸⁻¹⁰. This uncertainty has limited our ability to assess whether large earthquakes release
23 strain where coupling is strongest or rupture independently of interseismic coupling. Here we
24 present the first systematic global analysis of slip-coupling relationships, compiling 61
25 earthquake slip models (Mw 6.7–9.1) across 12 subduction zones. We find that slip–coupling
26 correlations strengthen with earthquake magnitude across margins. Events above Mw 8.7 release
27 about half of their slip where coupling exceeds 0.8, compared with a tenth for events below Mw
28 7.5. Large earthquakes therefore release strain accumulated in strongly coupled regions, whereas
29 coupling provides weaker constraints on moderate-earthquake rupture, which may reflect limited
30 resolution, evolving locking or rupture arrest. Our results link coupling models to coseismic slip,
31 showing where large earthquakes release strain and why anticipating the hazard posed by the
32 more frequent moderate earthquakes require denser, longer geodetic records.

33

34 MAIN TEXT

35 Introduction

36 The largest earthquakes on Earth occur at subduction zones, where a dense plate sinks into the
37 mantle, sliding below an upper plate¹¹. The boundary between these plates, known as the
38 megathrust, contains asperities¹² that remain fixed during the interseismic period, accumulating
39 strain over decades to centuries before rupturing in an earthquake. As strain gradually builds up,
40 the overriding plate deforms, generating surface displacements¹³ that are routinely measured by
41 geodetic networks. Geodesists invert these displacements for coupling models, which identify
42 portions of the megathrust that creep at the plate convergence rate (kinematic coupling ~ 0) or
43 remain fixed and accumulate slip deficit (kinematic coupling ~ 1)¹⁴ (Fig. 1). In this framework,
44 strongly coupled regions are expected to rupture coseismically and weakly coupled regions to
45 creep aseismically, an assumption that underpins many contemporary assessments of seismic
46 potential in subduction zones¹⁻⁴.

47 Observations increasingly show that this relationship is more complex. The 2020 Mw 7.8
48 Shumagin earthquake ruptured a partially weakly coupled region¹⁵, the 2007 Sumatra
49 earthquakes ruptured only part of a strongly coupled patch¹⁶, and the 2025 Mw 8.8 Kamchatka
50 earthquake released more slip than had accumulated since the last large event¹⁷. Such
51 discrepancies may arise from limitations in how coupling models represent megathrust locking.

Coupling models are based on only a few decades of observations^{14,18}, despite earthquake cycles spanning hundreds to thousands of years^{5,6}, and coupling can vary substantially over just a few years^{19–22}. Measured presumably elastic interseismic deformation may also reflect viscoelastic deformation^{13,23} or contributions from slow-slip events that can last for decades²⁴.

In addition, spatial constraints such as limited offshore geodetic coverage reduce the resolution of coupling estimates^{25,26}. Deformation near the trench is particularly poorly resolved, leading to large uncertainties in shallow megathrust coupling, where coseismic slip may result in devastating tsunamis^{7,27}. Beyond these observational limitations, theoretical and numerical studies suggest that high coupling is not a prerequisite for earthquake rupture. Dynamically propagating slip²⁸ may penetrate creeping regions while strongly coupled areas may act as rupture barriers^{8–10,29}.

A Global Comparison of Coupling and Slip

To address the important question of whether geodetically inferred coupling correlates with, or even predicts, megathrust coseismic slip, we establish systematic coupling-slip relationships. We assemble a global dataset of kinematic coupling models spanning twelve active subduction margins (Materials and Methods), together with 61 coseismic slip models (Mw 6.7 to 9.1) describing individual megathrust earthquakes that ruptured these margins (Figs. 1 and S1; Tables S1-S2; Text S1). Coseismic slip is often represented by finite-fault models³⁰ that resolve the spatial distribution of slip, and in some cases its temporal evolution, across multiple subfaults spanning the rupture surface (Fig. 2A). Each subfault is assigned a slip or rupture rise time value, inferred from teleseismic, regional seismic, geodetic, satellite, or tsunami observations^{31,32}.

Point-Based Representation of Slip

To compare interseismic coupling and coseismic slip, we discretize finite-fault slip models into point clouds whose density reflects the slip distribution and sample the coupling field at those locations. For each finite-fault model, we generate samples by uniformly distributing points within each subfault, with sample size reflecting subfault slip, area, and a sampling density constant κ (Eq. 1). We aggregate the samples across all subfaults such that the resulting point cloud reproduces the finite-fault slip distribution. As an example, expressing the Mw 7.7 2007 Tocopilla earthquake (Fig. 2A) as sample points shows that the discrete slip-distribution resolution improves with increasing κ and sample count (Fig. 2B1–4). From this point cloud, we assign coupling values to each sample (Figs. 2C1–2), aggregating them into a slip-weighted cumulative distribution function (CDF) describing how slip is distributed across the coupling spectrum. We use CDFs because they avoid additional hyperparameter choices and converge quickly toward the underlying coupling-slip distribution, allowing us to use relatively small sample counts to reduce computational costs (Text S2, Figs. S2-S3). The Tocopilla earthquake illustrates this, where CDFs computed from widely different sample sizes align, revealing that

approximately 60% of the coseismic slip occurred in areas with coupling values below 0.4, and only about 15% in regions where coupling exceeds 0.8 (Fig. 2D).

Quantifying Coupling Model Uncertainty

Coupling models are often ill-constrained, particularly offshore, yet they are usually reported as a single best-fitting solution without explicit uncertainty estimates. We provide a first-order assessment of coupling uncertainty by estimating interseismic posterior covariance along all twelve megathrusts. We discretize all twelve subduction interfaces into triangular dislocations and compute the Green's functions for the geodetic observations that were used in original coupling studies to estimate the posterior covariance (Eq. 5). The mesh resolution and regularization parameters are chosen to limit imposed smoothing while providing conservative upper-bound estimates of coupling uncertainty. In Chile, for example, the interface is discretized into 151 triangular fault patches with Green's functions calculated for ~300 nearby GNSS stations, yielding interseismic slip uncertainty that ranges from ~50% of the plate convergence rate in poorly resolved regions to ~5% in well-constrained areas (Fig. 2E). We propagate this uncertainty into the slip-coupling CDFs (Methods).

Results

Do Coupled Megathrusts Rupture?

Whether megathrust ruptures align with strongly coupled regions provides a fundamental test of the predictive value of geodetically inferred coupling models^{33–35}. We systematically test this by discretizing 61 (Fig. S1) earthquakes into ~12.5 million slip points and assigning coupling values according to their position within the twelve corresponding coupling distributions. The resulting CDF, constructed from the mean coupling models, shows that 60% of slip occurs where coupling is above 0.7 and 15% is released in fully coupled areas (Fig. 3A). To account for uncertainty in the coupling field, we generate 500 realizations of the coupling distribution for each subduction zone and recompute the corresponding slip-weighted coupling CDF for each realization. This slip-coupling correlation, with its uncertainty, is stronger than the correlation between slow slip and coupling that we compute for the Hikurangi subduction zone (Figs. 3A and S4). It is, however, weaker than in our idealized seismic-cycle simulations (Methods), in which models with both strong and moderate dynamic weakening place more than half of the coseismic slip in fully coupled regions, far exceeding the ~15% observed in our global compilation (Fig. 3A). These idealized end-members set an upper reference for the slip-coupling correlation of a persistent asperity, but do not on their own demonstrate whether megathrust slip favors coupled regions.

To analyze this quantitatively, we construct a null model in which the spatial association between slip and coupling is completely random. For each megathrust earthquake in our database (each represented by a finite-fault model) we generate 500 synthetic realizations by translating and

rotating its slip distribution across the coupling field (Methods). For each of the 500 realizations we aggregate the slip samples of all earthquakes and compute the slip-weighted coupling cumulative distribution function (CDF). We then compare each of the 500 observed CDFs, which include coupling uncertainty, with each of the 500 null CDFs. Evaluating at each coupling level whether the observed fraction of slip above that level exceeds the null (Eq. 8) yields 250,000 pairwise comparisons. The observed distributions exceed the null in more than 90% of comparisons across the entire coupling spectrum, exceeding the 68% threshold we adopt for a statistically meaningful slip–coupling relationship (Fig. 3A). This supports the conclusion that slip is preferentially concentrated in coupled regions, particularly within the most strongly coupled portions of the interface.

Variation in slip–coupling correlation across earthquake magnitudes

Our compilation of earthquake slip models ranges from moderate events to ruptures hundreds of kilometers long, allowing us to analyze whether slip–coupling relationships vary with magnitude. We bin slip samples in ~ 0.4 -magnitude intervals and compute slip–coupling CDFs for each group. We observe slip–coupling correlations that strengthen with moment magnitude. For events larger than Mw 8.7, $\sim 55\%$ of slip occurs above coupling of 0.8, whereas for events smaller than Mw 7.5, only $\sim 10\%$ of slip occurs above 0.8 (Fig. 3B). Comparison of each magnitude-binned CDF with its corresponding null model sharpens this contrast. For Mw < 7.1 earthquakes, the observed CDF exceeds the null (Eq. 8) only for coupling values below ~ 0.2 , and the null dominates at higher values (Fig. 3C1). Mw 7.1–7.5 (Fig. 3C2) and Mw 7.5–7.9 (Fig. 3C3) events show similar behavior, with observed slip exceeding the null only up to coupling values of ~ 0.55 and ~ 0.75 , respectively. Smaller events therefore show limited positive deviations from the null and do not robustly sample strongly coupled regions. By contrast, earthquakes larger than Mw 7.9 generally exceed the null across much of the coupling spectrum, with slip predominantly concentrated in higher-coupling regions (Figs. 3C4–C6). We note that for Mw 8.3–8.7 events (Fig. 3C5), the observed distributions exceed the 68% threshold only above coupling values of ~ 0.6 . To analyze whether this magnitude dependence holds at the level of individual events, we compute CDFs for each of the 61 megathrust slip models (Fig. 4A1). These individual-event CDFs show event-to-event variability, as expected given coupling-model uncertainty and margin-specific coupling structure. For example, the Mw 6.9 Valparaíso earthquake released more of its slip in strongly coupled regions than events of comparable magnitude (Figs. 4A1 and S5), but ruptured a sparsely instrumented region where coupling is highly uncertain (Fig. 2E). By comparison, the Mw 6.8 Vallenar earthquake ruptured a region with denser geodetic coverage and lower coupling uncertainty (Fig. 2E), and shows a slip–coupling relationship more consistent with other moderate-size events (Figs. 4A1 and S5-6). Despite this variability, the individual-event CDFs show a slip–coupling correlation that strengthens with magnitude (Fig. 4A1). A principal component analysis (PCA) of all 61 event-level CDFs, which requires no magnitude binning, supports this trend. The first principal component captures 73.3% of the variance, with a positive relationship between PC1 scores and

moment magnitude whose slope exceeds that of the corresponding null ensemble in ~95% of paired realizations (Text S3, Fig. S7).

To quantify this event-to-event variability more systematically, we compare individual slip-coupling CDFs using the Cramér–von Mises distance (CvM; Eq. 3), which quantifies disparity between CDFs, providing a direct measure of heterogeneity in slip–coupling relationships. Using CvM we observe that earthquakes below Mw 7.5 show average CvM values about 3.5 times greater than those of larger earthquakes and standard deviations 2.6 times higher (Fig. 4B1). We obtain similar results using an alternative L2 norm distance metric (Fig. S8). We interpret this as evidence that lower-magnitude earthquakes have more heterogeneous slip–coupling relationships, whereas larger ruptures exhibit consistent correlation of slip and coupled regions. Part of this contrast may reflect the smaller rupture areas of moderate events, which sample fewer independent portions of the coupling field.

Variation in slip–coupling correlation across subduction zones

Differences in slip–coupling among subduction zones may reflect distinct rupture behavior or differences in baseline coupling shaped by margin properties such as sediment input, thermal structure, fluids, geometry and roughness^{36–39}. We therefore test whether inter-margin variability reflects rupture behavior or the underlying coupling fields. We compute slip–coupling CDFs for each of the twelve subduction zones (Fig. 4A2) and quantify pairwise disparities using the CvM distance. Chile⁴⁰ emerges as the most distinct margin, with an average CvM value of 0.13, more than twice the global mean of 0.06. This reflects both the strongest alignment of slip with highly coupled regions and the largest mean CvM distance from other margins (Fig. 4B2). However, because Chile also exhibits the highest baseline kinematic coupling of all margins (Fig. 1), its distinctive signature may primarily reflect elevated coupling rather than a fundamentally different slip–coupling behavior.

To account for such baseline differences, we subtract each margin’s null CDF from its observed CDF. This null solution, constructed from synthetic megathrust slip distributions that randomly sample each margin’s coupling field, already encodes baseline coupling, so subtracting it may isolate the rupture-specific slip–coupling correlation. Recomputing CvM distances on these residuals produces a tighter inter-margin cluster. For example, the global mean distance decreases by about half, and Chile drops more than fourfold, converging to the corrected global mean (Fig. 4B2). This indicates that much of the apparent inter-margin contrast may arise from differences in baseline coupling rather than intrinsically different slip–coupling behavior, a result we also obtain using the L2 norm (Fig. S8).

The Role of Dynamic Weakening in Slip–Coupling Correlations

Our observations raise the question of whether the weaker slip–coupling correlations, especially for smaller events, can be explained by dynamic rupture processes. Earthquake rupture is

inherently dynamic, involving non-linear interactions of seismic waves, rapid fault stress changes, and frictional weakening that unfold within seconds to minutes^{9,28,41,42}. In this framework, rupture nucleates in regions that are relatively highly stressed and coupled, but dynamic weakening at seismic slip rates reduces frictional resistance by processes such as flash heating or thermal pressurization^{28,42}, and allows slip to propagate into surrounding regions that are less critically stressed and weakly coupled⁹. Consistent with this, our idealized seismic-cycle simulations, with either conventional rate-and-state friction^{43,44} or strong rate-weakening at seismic slip rates⁴⁵ (Methods), show that stronger dynamic weakening shifts a larger fraction of coseismic slip toward lower-coupling regions relative to the hypocenter (Figs. 3A1,5).

To assess whether this signature is present in observations, we compute the probability density function (PDF) of slip as a function of rupture time and coupling relative to the hypocenter for the 25 megathrust earthquakes (Mw 6.7–8.4) with resolved rupture timing (Fig. S9). We find no systematic influence of dynamic weakening on slip partitioning. Ruptures (Fig S10) concentrate most slip within ± 0.2 of the hypocenter's initial coupling and are equally likely to propagate into lower-coupling regions, which are presumably less critically stressed and may require enhanced dynamic weakening to slip⁹, as into higher-coupling regions (Fig. 5, Fig. S11-S12). A similar pattern emerges for the corresponding spatiotemporal null solution, where randomized slip–coupling–time correlations result in slip evenly distributed above and below its initial coupling (Fig. 5). Slip relative to hypocentral coupling for all 61 earthquakes, without timing information, shows the same result with no preference for higher or lower coupling than at the hypocenter, at any magnitude (Fig. S13). This first-order analysis does not rule out dynamic weakening in individual earthquakes, but shows no consistent large-scale signature of it in the slip–coupling relationships of our dataset.

Discussion and Conclusions

We identify a systematic magnitude dependence in slip–coupling relationships: large megathrust earthquakes rupture strongly coupled regions, whereas moderate events show weaker and more variable slip–coupling relationships and release much of their slip in weakly to intermediately coupled regions (Figs. 3C1-3C6, 4B1, S7A). This is surprising because many earthquake characteristics exhibit scale-independent behavior⁴⁶ and it contrasts with the expectation that earthquakes of all sizes release strain accumulated during the interseismic period⁴⁷. Part of this signal may reflect the resolution of geodetic coupling models. Moderate ruptures approach the scales that existing observations resolve, so that coupling is averaged over areas larger than the rupture and the coupling state of the failed patch may be obscured. The Mw 6.8 Vallenar earthquake illustrates this limit. Its rupture area of $\sim 1,800$ km² is only slightly larger than the optimal dislocation size required to resolve coupling variations from the available geodetic observations in Chile (Fig. S14) and nearly 20 times smaller than the Mw 8.3 Illapel rupture (Fig. 2E). Denser InSAR coverage and offshore geodesy are needed to better resolve

236 moderate-earthquake asperities and to determine whether their apparent weak coupling is
237 physical or unresolved.

238 Resolution limits alone, however, are unlikely to explain the magnitude-dependent signal, as the
239 slip–coupling correlation continues to strengthen with moment magnitude even for ruptures far
240 larger than the resolvable dislocation size (Fig. 3B; Text S3). Moderate earthquakes may instead
241 preferentially rupture regions where asperities evolve through creep-front migration or erosion⁴⁸,
242 so that present-day coupling captures only part of the strain accumulated before rupture. This
243 interpretation is consistent with geodetic observations of transient changes in coupling and
244 decadal migration of locked–creeping boundaries^{19–22,49}. Inversions that allow asperities to
245 evolve, together with paleogeodetic and geomorphic archives that integrate deformation over
246 longer timescales than modern geodesy^{22,50,51}, may help distinguish persistent strain accumulation
247 from transient locking.

248 A complementary explanation is that moderate earthquakes nucleate on small asperities within
249 heterogeneous coupled regions and arrest before growing into larger ruptures. The energy release
250 rate of a small rupture may be insufficient to break into adjacent coupled regions that are not yet
251 critically loaded, where the available stress drop is small⁵². Laboratory and numerical studies
252 show that regions which creep interseismically, and are frictionally stable under slow loading,
253 may host coseismic slip only if dynamic weakening at high slip rates lowers their strength below
254 the ambient shear stress^{9,10,53}. However, the spatiotemporal distribution of slip shows no
255 systematic preference for low-coupling regions, where enhanced dynamic weakening might be
256 required to sustain rupture (Fig. 5). Moreover, if dynamic weakening were the dominant control,
257 larger ruptures should propagate more readily into low-coupling regions, producing the opposite
258 magnitude dependence to that observed (Fig. 3). Thus, dynamic weakening may contribute to
259 rupture propagation, but does not by itself explain the weaker slip–coupling correlations
260 observed for moderate events.

261 Across all earthquake sizes, the observed slip–coupling relationships are weaker than those of
262 our idealized seismic-cycle simulations, even with strong dynamic weakening (Fig. 3). Spatial
263 averaging in geodetic inversions likely amplifies this discrepancy, but regularization alone is
264 unlikely to explain it (Text S6). The discrepancy more likely reflects the complexity of natural
265 megathrusts, where variations in temperature, rheology, stress, pore-fluid pressure, geometry,
266 roughness, and frictional properties produce more heterogeneous coupling and ruptures than our
267 idealized models^{36–39,54}. The same factors likely contribute to differences in baseline coupling
268 among subduction zones and thus to the variability of slip–coupling relationships among
269 margins. However, once differences in baseline coupling are accounted for, a consistent
270 slip–coupling relationship emerges across subduction zones (Figs. 4B2, S8 and S7D), suggesting
271 that accumulated strain is released in a similar manner worldwide.

272 In summary, we present the first global analysis of slip–coupling correlations, providing new
273 constraints on the link between geodetically inferred kinematic coupling patterns and earthquake
274 slip distributions. Across 61 megathrust earthquakes in twelve subduction zones, the
275 slip–coupling relationship strengthens with earthquake magnitude. Larger ruptures increasingly
276 concentrate slip in highly coupled regions, with the strongest correlation observed for $M_w \geq 7.9$
277 events, whereas smaller events do not. Inter-margin comparisons reveal that once baseline
278 differences in coupling strength are removed, slip–coupling relationships converge to a
279 consistent global pattern. This suggests that despite regional variations in convergence rate, slab
280 geometry, sediment supply, and thermal structure, the mechanics connecting coupling and
281 rupture are fundamentally similar across subduction zones.

282 **Acknowledgements:** We thank (in alphabetical order) M. Chlieh, B. Cosenza-Murales, L. Dal
283 Zilio, C. DeMets, A. Ellis, L. Feng, J. Freymueller, R. Bürgmann, R. Hanifa, G. Hilley, J.
284 Loveless, B. Loeferer, L. Maubant, M. Métois, T. Nishimura, D. Panda, R. Plata-Martinez, M.
285 Radiguet, and L. Xue for sharing their coupling models with us. BO and AAG acknowledge the
286 IGPP Green Foundation, National Science Foundation (grant no. OAC-2311208) and
287 CRESCENT Award 2017Y0SD for their support. AAG acknowledges additional support by the
288 National Science Foundation (grant nos. OAC-2139536, OAC-2311208, EAR-2225286,
289 EAR-2121568, RISE-2531036), USGS (NEHRP award No. G26AP00141), Horizon Europe
290 (ChEESE-2P, grant number 101093038; Geo-INQUIRE, grant number 101058518) and through
291 a gift from Pacific Gas and Electric Company (PG&E).

292 **Data availability:** Finite fault models, slip samples and code to sample finite fault models can be
293 found at [10.5281/zenodo.17160782](https://doi.org/10.5281/zenodo.17160782). Coupling models can be found in the Coupling Cloud
294 database⁵⁵. Finite-slip models in our dataset were obtained from the [SRCMOD database](#)⁵⁶ and the
295 USGS.

296 **Competing interest declaration:** The authors declare no competing interests.

297 **Author contributions:** Conceptualization: B.O., A.A.G.; Data curation: B.O.; Formal analysis:
298 B.O.; Methodology: B.O.; Validation: B.O., A.A.G.; Supervision: A.A.G.; Visualization: B.O.;
299 Funding acquisition: B.O., A.A.G.; Writing – original draft: B.O., A.A.G.; Writing – review &
300 editing: B.O., A.A.G.

301 Materials and Methods

302 Data selection

303 To evaluate whether coseismic slip preferentially occurs where the plate interface is most
304 strongly coupled, we assemble a global dataset that pairs interseismic kinematic coupling maps
305 with finite-fault slip models. We highlight that kinematic coupling should not be confused with

frictional locking, which describes the fault’s mechanical response to slip, where “locked” implies unstable behavior, or with seismic coupling, which measures the fraction of total slip released seismically, with 1 indicating entirely seismic slip⁷

The coupling model portion of our dataset includes twelve subduction margins (Fig. 1), with coupling maps chosen to favor models that use the latest geodetic data and cover the largest area (Text S1). These include the Kamchatka⁵⁷, Chile⁴⁰, Peru⁵⁸, Colombia and Ecuador⁵⁹, Mexico⁶⁰, Costa-Rica⁶¹, Sumatra⁶², Alaska^{63,64}, Japan⁶⁵, Nankai⁶⁶, Sagami²⁶ and the Himalayas⁶⁷ subduction zones.

For each margin, we compile published finite-fault models locally coinciding with the corresponding coupling region, yielding 61 distinct earthquakes (Mw 6.7–9.1) that occurred between 1923 and 2024 (Figs. 1 and S1; Table S1). We rely primarily on slip solutions derived by the USGS based on body- and surface-wave inversions, which reduces methodological variability and extends coverage to many otherwise undocumented megathrust earthquakes^{30,68–93}. We supplement our dataset with slip models from other published studies that document slip for historic earthquakes and were made available via the SRCMOD database^{56,94–109}.

Data sampling

We compare slip distributions with kinematic coupling maps, $C(x, y)$, by sampling each finite-fault model and expressing the slip distribution as a discrete point cloud (Figs. 2B1, D1-4). For each finite-fault model, all subfaults are sampled such that for a given subfault i we draw N_i points uniformly within its bounds, with the number of points proportional to its planar subsurface area A_i , and slip, S_i :

$$N_i = \kappa \cdot A_i S_i, (1)$$

Where κ is a density proportionality constant set to 3500 km^{-3} for our primary analysis and 750 km^{-3} for the null and spatial uncertainty analyses. These choices ensure that all 61 CDFs converge, with sampling uncertainties smaller than 1% and 5%, respectively (Text S2; Fig S2).

Next, we use the surface positions x_j, y_j of each sampling point, j , to evaluate the coupling, $C_j = C(x_j, y_j)$ (Fig. 2B2). The resulting sets of coupling values, $\{C_j\}_{j=0}^{N_f}$, represent a slip-weighted sampling of the coupling field, modulated by how coseismic slip is distributed across the range of coupling values. We note that for the case where we sample the rupture time (Fig. 5) we assign a uniform rise time for all samples drawn from subfault i .

Lastly, we assume that coupling and finite-fault models share a common geometry, which may not always hold. We compare multiple finite-fault models for the Tōhoku and Maule earthquakes based on different geometries and data, and further assess finite-fault uncertainty across the full

earthquake dataset (Text S4). These yield only second-order variations in slip–coupling relationships (Figs. S15-S18), indicating that geometry differences have a limited impact, likely because coupling varies over longer spatial scales than finite-fault slip models.

Characterizing slip–coupling relationships

We compile all N_t coupling assigned samples, $\{C_q\}_{q=0}^{N_t}$, from the finite-slip models, considering either the full dataset spanning 61 megathrusts (Fig. 3A) or any subset of interest, such as specific earthquakes (Fig. 4A1), subduction-related groups (Fig. 4A2) or selected magnitude ranges (Fig. 3C). We then compute the CDFs using:

$$\hat{P}(C) = \frac{1}{N_t} \sum_{q=1}^{N_t} \mathbf{1}_{\{C_q \leq C\}}, \quad (2)$$

where $\mathbf{1}_{\{C_q \leq C\}}$ is the indicator function returning 1 if $C_q \leq C$ and 0 otherwise.

We compute the distance between two cumulative distribution functions $\hat{P}_1(C)$ and $\hat{P}_2(C)$ using the Cramér-von Mises (CvM) distance:

$$\omega = \int_0^1 (\hat{P}_1(C) - \hat{P}_2(C))^2 dC. \quad (3)$$

We evaluate the joint spatiotemporal distribution of slip and coupling by computing a probability density function (PDF) of rupture rise time, T , normalized to 0–1, and coupling relative to the hypocenter, $\Delta C = C - C_{\text{hyp}}$ and ΔC which spans -1 to 1 . We use $\sim 700,000$ samples from 25 megathrust earthquakes which resolve spatiotemporal slip and employ a kernel density estimator to compute the PDF:

$$\hat{f}(T, \Delta C) = \frac{1}{N_t |H|^{1/2}} \sum_{q=1}^{N_t} \frac{1}{2\pi} \exp \left(-\frac{1}{2} \left[H^{-1/2} \begin{bmatrix} T - T_q \\ \Delta C - \Delta C_q \end{bmatrix} \right]^\top \left[H^{-1/2} \begin{bmatrix} T - T_q \\ \Delta C - \Delta C_q \end{bmatrix} \right] \right), \quad (4)$$

where H is a bandwidth matrix defined by $H = N_t^{-1/6} \Sigma^{1/2}$ and Σ is the covariance matrix computed from the full set of samples, N_t .

Quantifying spatial uncertainty in coupling fields

We quantify and propagate uncertainty in interseismic slip rate models using linear inverse theory, where the posterior model covariance, C_{post} , is expressed as¹¹⁰:

$$\mathbf{C}_{post} = (\mathbf{G}_{hf}^T \mathbf{C}_d^{-1} \mathbf{G}_{hf} + \alpha^2 \mathbf{L}^T \mathbf{L})^{-1}, \quad (5)$$

where G_{hf} is the Green's function that maps the unit slip on a patch h in the dip-slip and strike-slip directions to the corresponding displacement at station f in the east, north and vertical directions, C_d is the data covariance, L is the Laplacian smoothing operator and α is the regularization weight.

We estimate C_{post} for each margin by computing the Green's functions, G_{hf} , for all geodetic observations positions used in the original studies. We assume an elastic half-space with a Poisson ratio of 0.25 and discretize the twelve subduction margins into twelve uniform meshes of triangular dislocations¹¹¹ based on the original geometry used in each publication (Table S2). The number of patches is selected such that the available three-component geodetic observations³⁵ are roughly three times the number of slip parameters in the dip and strike directions. We construct C_d , assuming uncorrelated data uncertainties with standard deviations of 4 mm/yr and 2 mm/yr in the vertical and horizontal directions, respectively. For the models derived from InSAR^{60,61} we assign uncertainties twice those of GNSS, distribute observations uniformly to match the reported spatial density, and assume spatially correlated errors that decay exponentially with a characteristic length scale of 10 km¹¹². Lastly, we apply a small regularization weight of $\alpha=0.1$, which limits imposed smoothing and makes the resulting uncertainty estimates conservative upper bounds (Text S5; Figs. S14 and S19).

Subsequently, for each subduction we draw five hundred correlated interseismic slip realizations from the multivariate normal distribution $\mathcal{N}(\mathbf{S}, \mathbf{C}_{post})$, bounded¹¹³ between zero and the plate convergence rate S_{cov} ¹¹⁴ where the mean interseismic slip rate, S , is linked to coupling field, $C(x,y)$, and S_{cov} :

$$S(x, y) = S_{cov} (1 - C(x, y)). \quad (7)$$

For each slip realization, we compute the corresponding coupling field using Equation (7). We then sample each of the 61 finite fault models against the 500 coupling realizations to obtain a distribution of slip-coupling correlations, which we use to compute the deviation from our mean slip-coupling correlation. We acknowledge that our discretization choices and regularization parameters may differ from those in the original coupling models. Our goal is to provide first-order upper-bound estimates that are consistent across all subduction zones rather than to strictly reproduce the original inversions (Text S5). Other sources of nonuniqueness, including uncertainties in coseismic slip models (Text S4; Figs. S15–S18; Tables S3–S4), smoothing constraints (Text S6; Figs. S20–S22), and the choice of coupling model (Text S7; Fig. S23–S27), produce variations in slip-coupling relationships that remain second order, in particular relative to the null hypothesis, and do not affect the primary patterns.

398 Null model

399 We assess whether the observed correlation between slip and coupling exceeds that of a null
400 model in which the spatial slip-coupling association is purely random. For each of the 61
401 megathrust earthquakes, we generate 500 synthetic realizations, resulting in 30,500 null
402 earthquake slip models across ~ 1.2 billion samples. For each realization, we translate the original
403 finite fault model to a location drawn uniformly within the subduction margin where the original
404 megathrust earthquake model ruptured (e.g., Fig. S28). We then rotate the model about a pivot
405 point chosen at random within the projected realization, with an angle drawn uniformly from 0 to
406 360 degrees. We only retain synthetic earthquake models for which at least 95% of slip samples
407 fall within regions where coupling is defined. We preserve the original finite-fault down-dip
408 orientation and ignore geometric discrepancies between the coupling model and accepted
409 realizations. The null model is limited to the spatial extent of the published coupling models,
410 some of which do not cover the full seismogenic zone or extend beyond it.

411 We consider that slip favors coupling up to level c (a coupling threshold between 0 and 1) when
412 the fraction of slip released above c is larger for the observed CDFs, CDF^{obs} , than for the null
413 realization, CDF^{null} , in more than 68% of the 500×500 pairwise comparisons:

$$414 \quad H(c) = \frac{1}{500 \times 500} \sum_{i=1}^{500} \sum_{j=1}^{500} \mathbf{1} [CDF_i^{\text{obs}}(c) < CDF_j^{\text{null}}(c)] . \quad (8)$$

415 The exceedance fraction is the probability that a randomly drawn observed realization exceeds a
416 randomly drawn null realization and serves as a probability-of-superiority measure. It equals
417 50% when the observed and null ensembles are indistinguishable. We adopt 68% as the threshold
418 for a systematic deviation.

419 Slip-coupling correlation for idealized quasi-dynamic 3D megathrust models

420 We simulate sequences of earthquakes and aseismic slip¹¹⁵ on an idealized 3D megathrust using
421 the open-source boundary integral code HBI¹¹⁶. The fault is loaded by a backslip boundary
422 condition assuming the deep portion of the megathrust which we do not model slips steadily at
423 the plate rate V_{pl} . The system evolves in time by solving for the slip velocity, V , and the state
424 variable, θ , along the megathrust, which obey the regularized formulation of rate-and-state
425 friction¹¹⁷:

$$426 \quad F(V, \theta) = a \sinh^{-1} \left(\frac{V e^{\theta/a}}{2V_0} \right) , \quad (9)$$

427 where V_0 is the reference slip velocity and a is the friction parameter describing the direct
428 velocity effect. The state variable evolves according to⁴⁵:

429

$$\frac{d\theta}{dt} = \frac{-V}{D_{RS}} [F(V, \theta) - F_{ss}(V)], \quad (10)$$

430 where D_{RS} is the characteristic state evolution distance and $F_{ss}(V)$ is⁴⁵:

431

$$F_{ss}(V) = f_w + \frac{f_0 - (b - a) \ln \left(\frac{V}{V_0} \right) - f_w}{\left[1 + \left(\frac{V}{V_w} \right)^8 \right]^{1/8}}, \quad (11)$$

432 where f_0 is the reference friction coefficient at $V = V_0$, $f_w = 0.2$ is the fully weakened friction
 433 coefficient⁴⁵ approached for $V \gg V_w$, V_w is the weakening velocity that controls the transition
 434 to strong dynamic weakening, and b is the friction parameter describing the evolution effect.

435 This formulation allows two end-member scenarios: (1) $V_w = 0.1 \text{ m/s}$ ⁴⁵, which produces strong
 436 dynamic weakening at coseismic slip rates, and (2) $V_w = 10^6 \text{ m/s}$, which behaves like classical
 437 rate-and-state friction adhering to the slip law⁴³ with only moderate dynamic weakening.

438 For both models we use our simulated interseismic period and estimate kinematic coupling along
 439 fault nodes, $C_t(x)$, using:

440

$$C_t(x) = 1 - \frac{s(x, t_{eq}) - s(x, t_{30})}{V_{pl}(t_{eq} - t_{30})}, \quad (12)$$

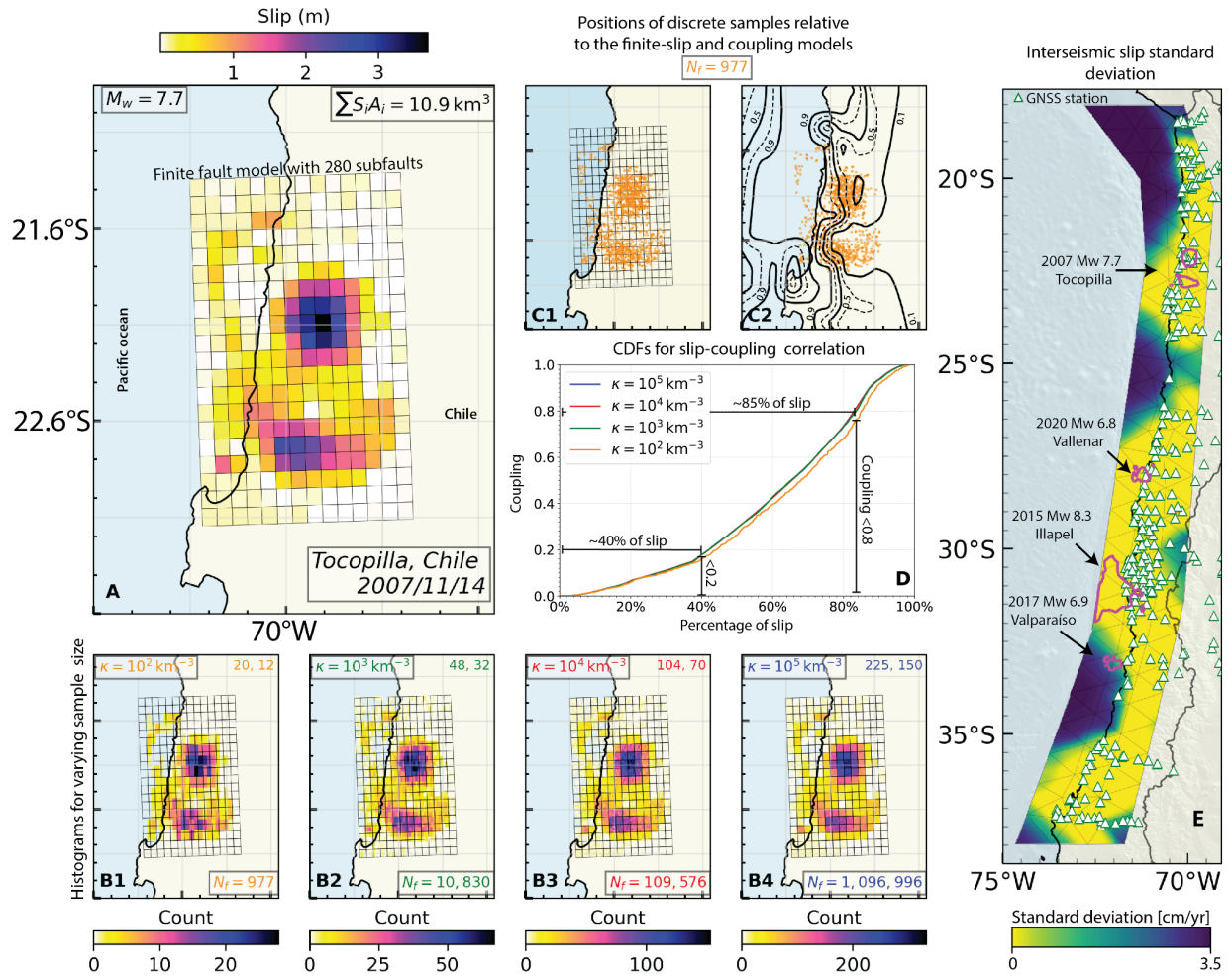
441 where $s(x, t_{eq})$ and $s(x, t_{30})$ are the cumulative slip at location x at the earthquake onset and 30
 442 years prior (Fig. S29). These simulations generate many megathrust earthquakes. However, after
 443 the initial spin-up phase, the events become quasi-periodic, and we analyze a single
 444 representative event. Table S5 lists all model parameters. A 2D simulation (Fig. S30) with
 445 different frictional properties and run with the volumetric code Tandem^{118,119}, yields a similar
 446 slip-coupling correlation to the 3D HBI¹¹⁶ simulation (Fig. S31).



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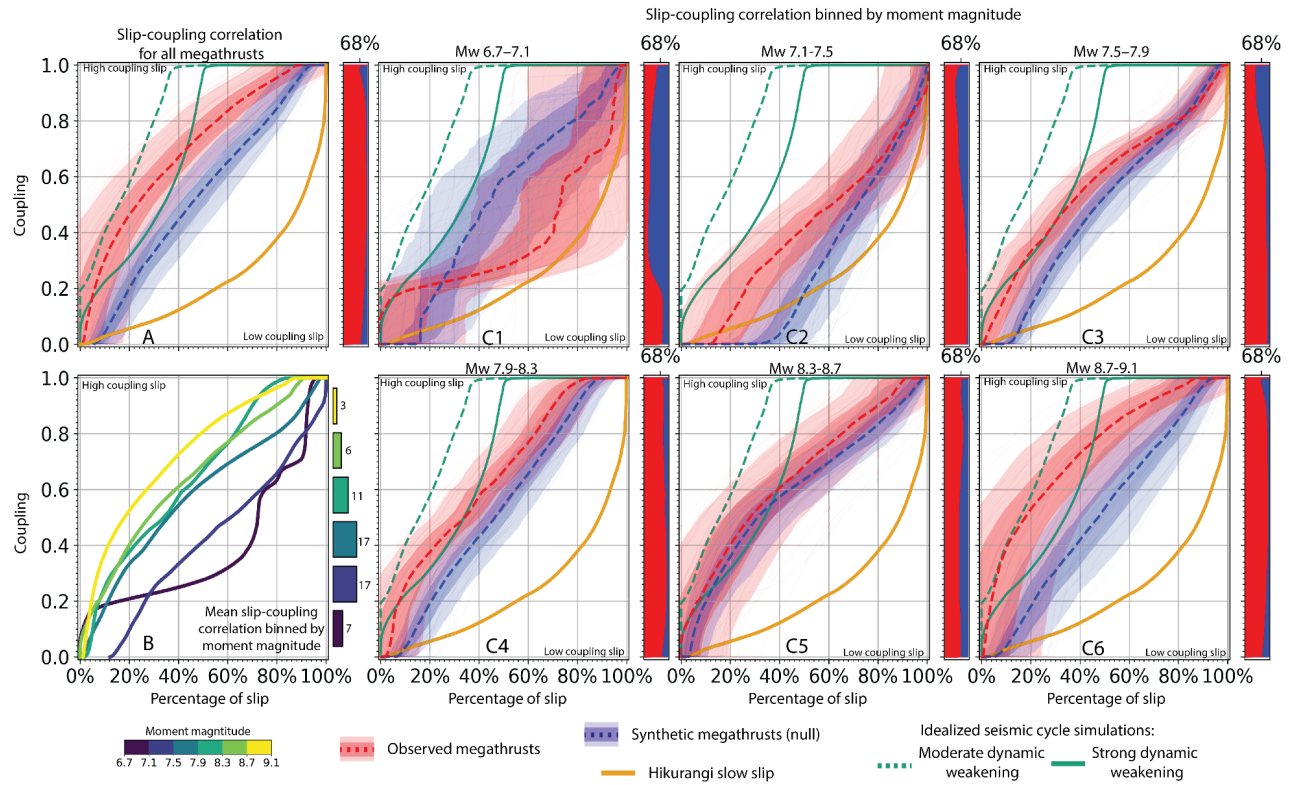


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Figure 2. Coseismic slip distribution and point-based discretization of the 2007 Mw 7.7 Tocopilla megathrust earthquake³⁰. A – Coseismic slip resolved on 280 rectangular subfaults. B1-B4 - Number of samples within bins along the fault surface. The number of bins (shown in the upper right of each panel) is determined using the Freedman–Diaconis rule¹²⁰ and reflects the effective resolution of the discretized slip distribution. C1-C2 – Point-based representation of slip. Dots indicate 977 samples drawn. In C2, contours show the coupling distribution⁴⁰ used to assign coupling values to each sample. D – Cumulative distribution functions (CDFs) of slip-coupling correlation for varying κ and sample count. E- Standard deviation of interseismic slip along the Chilean megathrust. Green triangles mark GNSS stations, black lines 151 fault patches, and magenta contours the rupture areas of the Tocopilla, Vallenar, Illapel and Valparaíso megathrust earthquakes.

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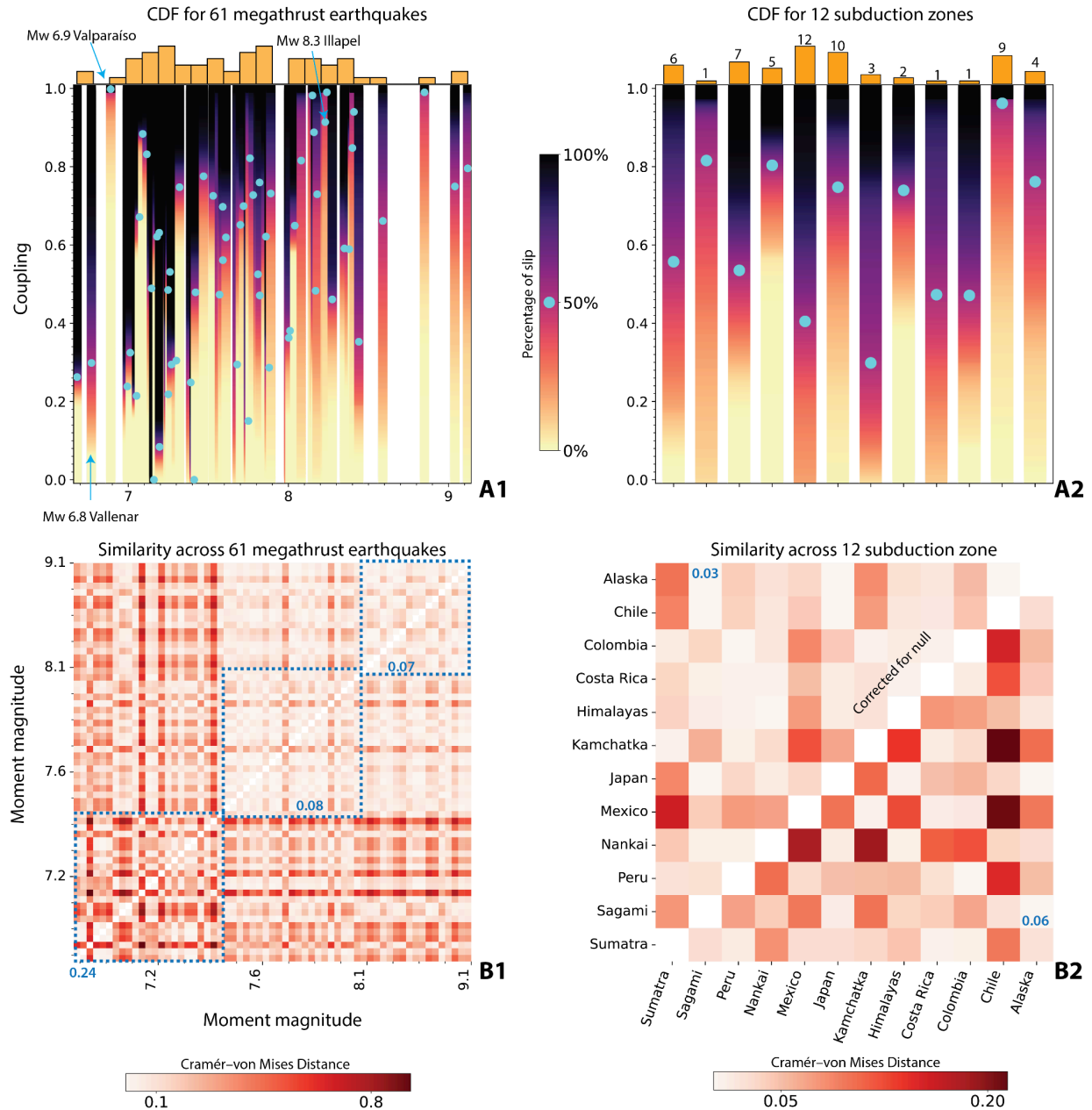
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Figure 3. Slip–coupling correlations for observed (red) and synthetic null (blue) megathrust earthquakes. Dashed red and blue lines show the mean, with shaded bands indicating where 50% and 95% of the 500 coupling-uncertainty and null realizations lie. Marginal bars indicate the percentage of 500×500 pairwise comparisons in which observed slip exceeds (red) or falls below (blue) the null, with the 68% threshold for a statistically meaningful deviation from the null (Eq. 8). Slip–coupling correlations from seismic cycle simulations and from slow slip at the Hikurangi subduction zone are shown in green and orange, respectively. A – All events combined. B – Mean CDFs for six magnitude bins, with event counts per bin shown by bars. C1–C6 – Full distributions for each magnitude bin.

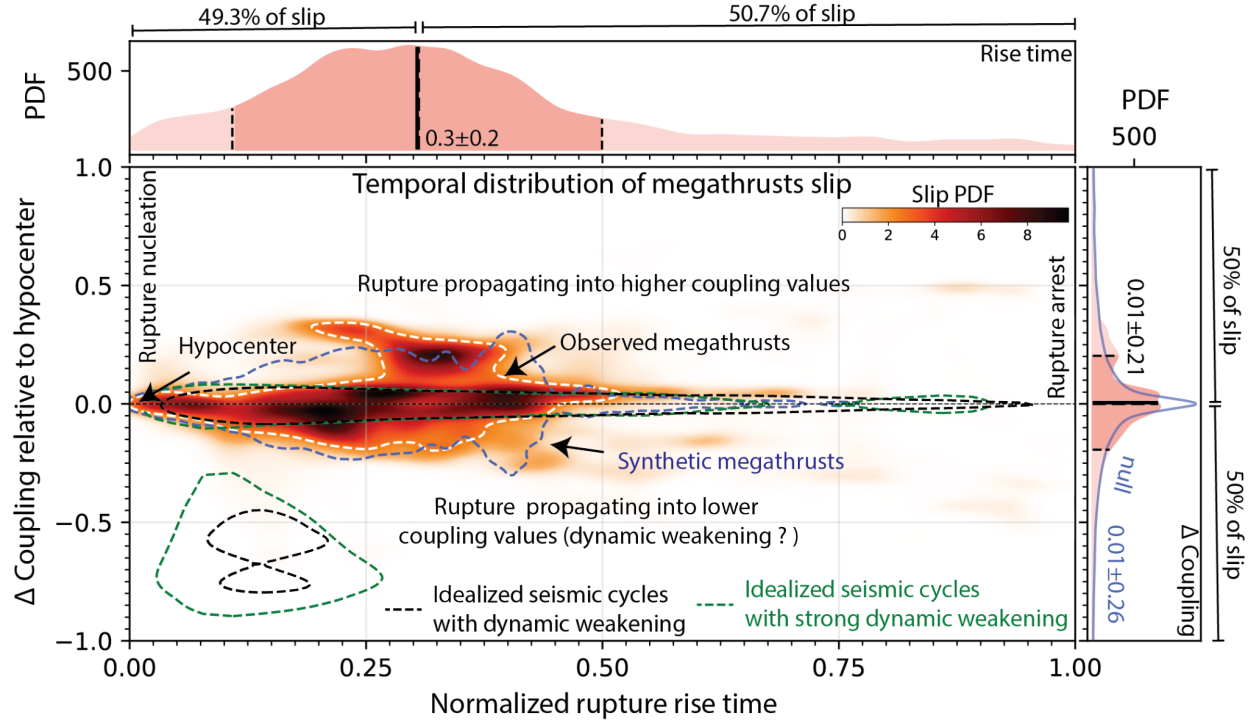
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483 **Figure 4. Slip-coupling similarities across megathrust earthquakes and subduction**
 484 **margins.** A1- CDFs for 61 observed megathrust earthquakes, with bars showing event counts
 485 per 0.1 magnitude bin A2 - CDFs aggregated across 12 subduction zones with bars indicating
 486 event counts per margin. B1 - Pairwise Cramér-von-Mises (CvM) distances among the 61
 487 megathrust earthquakes; blue numbers denote mean within-box values. B2 - Pairwise CvM
 488 distances among the 12 margins; blue numbers give mean distances for both null-corrected and
 489 uncorrected cases.

490



491

492 **Figure 5. Spatiotemporal patterns of megathrust slip.** Probability density of slip as a function
 493 of normalized rupture rise time and coupling relative to the hypocenter for 25 megathrust
 494 earthquakes. White, blue, black, and green contours enclose 68% of slip for observed, synthetic,
 495 and idealized models with moderate and strong dynamic weakening, respectively. Marginal
 496 panels show the corresponding PDFs of coupling and rise time for observed and synthetic
 497 earthquakes, with means indicated by solid lines and standard deviations by dashed lines.

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Supplementary Materials for

Do Coupled Megathrusts Rupture?

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This PDF file includes:

Supplementary Text S1-S8

Figs. S1- S31

Table S5

Other Supplementary Materials (see table captions below):

Table S1

Table S2

Table S3

Table S4

Text S1 - Choice of coupling models

Kamchatka

To the best of our knowledge, the only geodetic coupling model available for the section of the Kamchatka subduction zone along the Pacific–Eurasian plate interface—and one of the first kinematic coupling models ever established—is that of Bürgmann et al.¹. Bürgmann et al.¹ used 12 permanent and 18 campaign GPS stations, operating over 4 and 3 years respectively, to estimate slip deficit rates across a range of smoothing values. Notably, Bürgmann et al.¹ do not report coupling values directly, but rather slip deficit rates along the megathrust. We here estimate coupling from their inference by normalizing the reported slip deficit rates by the maximum plate convergence rate of $8 \text{ cm}\cdot\text{yr}^{-1}$ and adopting a smoothing value of $\beta=0.05$, as in their preferred model.

Alaska

The subduction of the Pacific Plate beneath the North American Plate in southern Alaska has long been recognized as a source of large megathrust earthquakes, rendering the estimation of slip deficit and seismic potential a key research focus ^{2–4}. For our analysis, we rely on the coupling model compiled by Basset et al.⁵, which synthesizes results from several studies, primarily based on GNSS observations, to quantify slip deficit along the Alaska margin. These sources include Elliott & Freymueller³, who incorporated data from 84 continuous GNSS stations and 327 campaign sites, with most observations collected between 2002 and 2016. Additionally, Basset et al.⁵ incorporated kinematic coupling estimates from Li & Freymueller⁴, based on 78 campaign and continuous GPS stations spanning the period from 1992 to 2017.

Central America

Northern Mexico

Kinematic coupling along the northern Cocos plate beneath Mexico has been estimated by several studies over the past few decades ^{6–8}. Our here preferred model is that of Maubant et al.⁷, who used recent geodetic observations, including GNSS and Sentinel-1 InSAR velocities collected over a 3.5-year period, to estimate coupling across the 3D megathrust geometry as constrained by Slab2. Out of the models we examined, this study reports the most spatially extensive coupling, spanning from western to eastern Mexico (106°W to 95°W), thereby enabling the inclusion of the largest number of earthquakes in our analysis. We note that their model occasionally reports negative coupling values, due to slow slip events in the region. In our analysis, we set those negative values to zero.

Costa Rica

A few estimates of interseismic coupling along the Cocos–Caribbean Plate interface in Costa Rica, primarily focused on the Nicoya Peninsula, have been published to date^{9,10}. For our analysis, we rely on the model of Xue et al.⁹, which incorporates the most up-to-date geodetic observations available at the time, including three years of L-band InSAR interferograms and GNSS velocity data from 19 continuous and 29 campaign GPS stations.

Sumatra

Surprisingly, despite Sumatra producing some of the largest megathrust earthquakes, we found only one study that provides coupling estimates spanning much of the island¹¹. Chlieh et al.¹¹ combines vertical displacement rates inferred from coral growth rings at 44 sites, spanning the period from 1962 to 2000, with horizontal GPS velocities from 33 sites surveyed during campaign-style observations between 1991 and 2001. The model also incorporates data from six continuous GPS stations deployed during the two years leading up to the 2004 Great Sumatra earthquake. For their inversion, Chlieh et al.¹¹ simplified the megathrust interface to a planar dislocations dipping uniformly at 13° , beginning at the trench.

Colombia, Ecuador , and Peru

Colombia & Ecuador

The segment of the Nazca Plate that subducts beneath the South American Plate along the Ecuador–Colombia trench extends for approximately 900 km along-strike and has produced several large megathrust earthquakes. To the best of our knowledge, two estimates of interseismic coupling have been published for this region^{12,13}. Our preferred model¹³ incorporates the most up-to-date geodetic observations, relying on data from 113 GNSS sites distributed across Colombia, Panama, Ecuador, and Venezuela to compute coupling along the megathrust, constrained using the Slab2 geometry¹⁴.

Peru

To our knowledge, two estimates of slip deficit rates have been published for the Central Andes subduction zone, where the Nazca Plate subducts beneath the South American Plate^{15,16}. Here our preferred model, from Lovers et al.¹⁵, relies on more recent geodetic observations, incorporating data from 47 permanent and 26 campaign GNSS sites collected between 1993 and 2022 to

estimate interseismic coupling along the megathrust from Lima to Arica. The model uses a 3D fault geometry derived from Slab2.0 and explores a range of smoothing parameters and reference frames. For our analysis, we adopt their preferred solution, computed in the Peruvian Sliver reference frame, using a 30 km smoothing constraint.

Chile

The Chilean margin is a well-studied region where numerous estimates of interseismic coupling along the subducting Nazca Plate have been published^{17–22}. For our analysis, we use the model of Métois et al.²⁰, which spans nearly the entire Chilean margin, extending over 3,000 km from 18°S to 38°S, and encompasses four earthquake slip models for events larger than Mw 8.0 (Fig. 1). The model incorporates 396 horizontal GNSS velocities along with 70 vertical slip rate estimates across the region, with some sites including data collected as early as the 1990s. To estimate slip deficit, the authors discretized the slab interface into a grid of 93 along-strike nodes (spaced every 0.25°) and 11 along-dip nodes, based on the Slab1 geometry, and solved for coupling at each node as well as for the Euler pole.

Coupling in Japan

Japan hosts three distinct subduction zones. In the southwest, the Nankai Trough marks the subduction of the Philippine Sea Plate beneath the Eurasian Plate. To the east, the Sagami Subduction Zone represents a relatively short (~350 km) segment where the Philippine Sea Plate subducts almost perpendicularly to the Nankai Trough. Further north, the Japan Trench marks the subduction of the Pacific Plate beneath the Eurasian Plate, extending through Tohoku and into Hokkaido.

Interseismic coupling along these three subduction zones has been estimated by several groups, typically focusing on individual segments^{23–29}. For our analysis, we adopt the model of Loveless & Meade (2015), for the Japan trench. Loveless & Meade simultaneously estimated slip deficit across all three subduction zones using a 3D elastic block-model consisting of 20 tectonic blocks and relying on approximately 19 years of GNSS observations from ~800 sites. Their model extends all the way to Hokkaido allowing us to include a number of megathrust earthquakes that would not align with other coupling maps (Fig. 1). We note that Loveless & Meade reported two coupling models, one in which the plate interface near the trench is locked, and another where it is freely slipping. We use the model in which the trench is locked, consistent with interpretations that high apparent coupling near the trench results from stress shadow effects^{24,30}. For Sagami we rely on coupling inversions that employed 840 landbase GNSS sites as well as 23 campaign

style acoustic offshore geodetic measurements²⁷. For the Nankai section, we rely on a study that used the same dataset as Sagami but applied a Markov Chain algorithm to probe the posterior distribution, together with Green's functions based on realistic Earth properties and structures, to obtain coupling fields with minimal assumptions²⁸.

Himalayas

Numerous studies have sought to characterize interseismic coupling along the Himalayan plate boundary, where the Indian Plate converges with the Eurasian Plate along the Main Himalayan Thrust^{31–37}. Our preferred model³⁶ incorporates the most up-to-date geodetic observations and estimates coupling across the entire margin, from 74°E to 92°E. Panda & Lindsey (2024) analyzed data from 839 GNSS sites spanning the Indian Plate, the Tibetan Plateau, and the Indo-Burmese Arc. They tested over 200 block model configurations in an iterative framework, systematically accounting for block motions and internal strain to isolate the coupling signal along the Main Himalayan Thrust. Their three best-fitting models consistently indicate that the megathrust is largely locked. For our analysis, we adopt their preferred solution, based on the block geometry that provides the best fit to the observed data.

Kinematic coupling models not used

We would like to note a few additional coupling studies that we obtained but did not use, as we did not find records of finite slip models that ruptured these segments. Both coupling models discussed below are shown in Fig. 1 for completeness, but are not included in the primary analysis.

Southern Mexico, Guatemala, El Salvador, and Nicaragua

Along the Cocos–Caribbean Plate interface, coupling has not been studied in detail, and we identified only one study³⁸ that provides a quantitative estimate for its northern section. Ellis et al.³⁸ used data from over 200 GNSS stations to model the elastic and kinematic behavior of southern Mexico, Guatemala, El Salvador, and Nicaragua, resolving motion across eight tectonic plates and blocks using the TDEFNODE software. For our analysis, we use the coupling they estimated for the Cocos–forearc sliver, spanning from southern Mexico to Nicaragua.

Java

Two estimates of slip rate have been proposed for the western and eastern segments of the Sunda subduction zone in Java. The first is based on three years of data from 13 GNSS stations

collected between 2008 and late 2010 and was used to estimate coupling in western Java³⁹. The second incorporates data from 37 GNSS stations distributed across central and eastern Java, spanning the period from 2008 to 2014⁴⁰. Notably, neither study reports coupling values directly; instead, they provide slip rate estimates, which we normalize by the maximum slip rate to approximate coupling.

Text S2 - Sensitivity of slip–coupling correlation to choice of κ

To test the sensitivity of our results to the choice of κ , we compute the CDFs for all megathrust events using 13 different κ values ranging from 21 to 10^5 km^{-3} . For $\kappa=3500 \text{ km}^{-3}$ and $\kappa=750 \text{ km}^{-3}$, the maximum differences are less than 1% and 5%, respectively (Fig. S2). We therefore adopt $\kappa = 3500 \text{ km}^{-3}$ for our primary analysis and 750 km^{-3} for the null model and spatial uncertainty tests.

We further examine the convergence of the probability density function (PDF) and CDF (Fig. S3) using the Tocopilla megathrust earthquake ⁴¹ (Fig. 2). The CDF stabilizes with just over 10,000 samples, whereas the PDF fails to converge even with one million samples, due in part to the clustering of samples near the edges of the distribution.

Text S3 - Principal component analysis for the moment-magnitude dependence of slip-coupling correlation

To evaluate whether the magnitude dependence inferred from the binned analysis (Fig. 3) is also expressed at the event-level, we use Principal component analysis (PCA)^{42,43}. We construct a data matrix, X_{ij} in which each row, i , is one megathrust earthquake and each column, j , is the value of its slip-coupling CDF at a given coupling value. PCA identifies the dominant orthogonal modes of variability in $X_{ij} - \bar{X}_j$, where $\bar{X}_j$ is the mean CDF value across earthquakes at coupling threshold j . In our case, the leading principal component explains 73.3% of the event-level mean slip-coupling CDF variance, indicating that most differences among earthquake slip-coupling CDFs can be described by a single dominant mode. We then project each centered earthquake CDF, $X_i - \bar{X}$, onto this leading CDF-variability axis, commonly referred to as the PC1. To test whether this dominant mode is related to earthquake size, we project each coupling-uncertainty realization onto the same leading CDF-variability axis and compute the corresponding PC1 scores. We also apply the same projection to the event-level null realizations to obtain the corresponding null distribution. For the observed and null ensembles separately, we then fit $PC1 = aMw + b$, to each of the 500 realizations, where the slope a quantifies the magnitude dependence along the observed PC1 axis.

We find that the magnitude dependence remains robustly positive under coupling uncertainty, with mean slope $a = 3.76$ and a 95% range of 1.45 to 5.52 (Fig. S7A) In contrast, the event-level null ensemble has a weaker and broader slope distribution, with mean $a = 1.73$, and a 95% range of -0.48 to 3.93 (Fig. S7B) Importantly, the observed slope exceeds the corresponding null slope in 95% of comparisons between observed and null realizations, demonstrating that the observed magnitude dependence is stronger than expected under the null hypothesis and is not explained by our bin choices (Fig. 3).

We note that even the null exhibits a weak positive trend, with only ~7% of realizations having negative slopes. This arises because larger synthetic ruptures sample more of the megathrust and therefore better reflect the background coupling distribution. As the largest earthquakes occur on highly coupled margins such as Japan and Chile, the null acquires a weak positive slope despite random rupture placement. However, as shown above, the observed slope remains larger than the null, indicating that this geometric effect alone cannot explain the observed magnitude dependence.

Lastly, to test whether the observed magnitude dependence is controlled by a small number of earthquakes or subduction zones, we repeated the PCA analysis after removing one earthquake at a time and, separately, removing all earthquakes associated with each subduction zone. For each leave-one-out experiment, we recomputed the observed and event-level null mean slopes. The

magnitude dependence remains positive when any single earthquake, or any entire subduction zone, is removed from the analysis (Fig. S7C-D) demonstrating that the observed trend is a robust feature of the global dataset rather than being controlled by any individual earthquake or subduction zone. We note that excluding the Nankai subduction zone produced the largest reduction in the observed–null slope difference; however, the observed mean slope remained larger than the corresponding null mean slope.

Text S4 – Robustness of the slip–coupling correlation to discrepancies between finite-fault and coupling models

One limitation of our analysis is that finite-fault and interseismic coupling models are derived from different datasets, inversion strategies, and assumptions, which can introduce systematic differences between the two fields. In particular, differences in the assumed elastic structure, fault geometry, and observational constraints may affect the inferred slip amplitudes and their spatial distribution. To evaluate the impact of these factors, we analyze multiple independent finite-fault models for two well-studied earthquakes for which numerous alternative models are available: twelve models for the 2011 Tōhoku earthquake and five models for the 2010 Maule earthquake (Tables S3 and S4). These models span seismic, geodetic, tsunami, and joint inversions, and employ different fault geometries and parameterizations (Figs. S15 and S16).

Across both earthquakes, the inferred slip–coupling relationships remain consistent despite the substantial differences among the finite-fault models (Fig. S17). Although individual models differ in the detailed distribution of slip, these variations produce only small changes in the overall slip–coupling relationship, with the exception of one Maule model in which the dominant slip patch is shifted. This result is consistent with previous intercomparisons of finite-fault models⁴⁴, which show that large-scale rupture characteristics are generally robust across inversion approaches, whereas smaller-scale slip variations are less well resolved. Together, these results indicate that our conclusions are controlled primarily by the long-wavelength structure of the rupture rather than by the particular finite-fault inversion adopted.

To further demonstrate that uncertainties associated with the finite-fault models have only a limited impact on our results, we generated 500 spatial realizations of each of the 61 megathrust earthquakes by randomly shifting the finite-fault model within a 20 km radius, corresponding to the average spread in hypocenter locations among the alternative Tōhoku and Maule finite-fault models. We then repeated the complete PCA for all 500 realizations, propagating earthquake-location uncertainty through the analysis. The resulting distribution of regression slopes remained strongly positive and was centered on an even steeper slope than obtained from the coupling-uncertainty analysis (4.34 versus 3.76; Figs. S7A,S18), demonstrating that the inferred magnitude dependence is robust to uncertainties in earthquake location. This weaker sensitivity likely reflects the different spatial scales resolved by the two datasets. Coupling models represent smooth, long-wavelength deformation, whereas finite-fault models resolve relatively localized slip. Thus uncertainty in slip produces only minor changes in the sampled coupling and the resulting slip–coupling relationship.

Text S5. Sensitivity of uncertainty estimates to discretization and regularization

To address the sensitivity of model uncertainty to discretization and regularization, we performed an additional resolution-based analysis. The goal of this analysis is not to replace the uniform first-order uncertainty estimate used in the main text, but to test whether our conclusions depend strongly on the assumed mesh density or choice of regularization weight.

Because the original dislocation meshes and regularization weights are generally not available, we construct a comparable discretization for each margin using the resolving power of the geodetic data. We start from an initially dense uniform triangular mesh and adjust the number of dislocations until the mean diagonal of the model resolution matrix:

$$\mathbf{R} = (\mathbf{G}_{hf}^T \mathbf{C}_d^{-1} \mathbf{G}_{hf} + \alpha^2 \mathbf{L}^T \mathbf{L})^{-1} \mathbf{G}_{hf}^T \mathbf{C}_d^{-1} \mathbf{G}_{hf}, \quad (1)$$

reaches 0.97 under weak regularization (Fig S14). The diagonal elements of $\mathbf{R}$ quantify how well each model parameter is resolved, such that values close to one indicate high resolvability. This step ensures that the number of model parameters is consistent with the resolving power of the observations, rather than imposed a priori.

For each resulting mesh, we evaluate the posterior covariance and resolution matrix over a range of regularization weights. We retain the weakest regularization for which the mean model resolution reaches 0.85, and terminate the search when the maximum posterior standard deviation falls below one half of the plate convergence rate, consistent with the largest uncertainties reported for the Lesser Antilles⁴⁵, Cascadia⁴⁶, and Hikurangi⁴⁷. This algorithm ensures that uncertainty is estimated in a manner consistent with the resolving power of the data and with uncertainties observed across subduction zones. This approach yields lower uncertainty estimates (Fig. S19) than those obtained in the main-text analysis (Fig. 3), and therefore we retain the main-text results as a more conservative upper bound.

Text S6 - Sensitivity of slip–coupling correlation to coupling smoothing constraints

Kinematic coupling models are typically estimated through linear inversion, in which the number of slip parameters along the megathrust exceeds the number of available geodetic observations. This leads to an underdetermined problem in which many slip solutions can fit the data equally well. To stabilize the inversion, spatial smoothing constraints are commonly imposed to penalize abrupt slip variations and favor more spatially coherent coupling patterns. While this regularization reduces non-uniqueness, it can also alter the resulting coupling distribution.

To assess the impact of smoothing on our results, we compare slip–coupling correlation derived from three smoothing levels in Peru¹⁵ and four in Kamchatka¹, where coupling maps are reported for each smoothing level (Fig. S20). We treat each coupling map obtained with a distinct smoothing value as an independent distribution and compute its mean slip–coupling correlation together with the corresponding null solution. We observe that although variations in smoothing can produce differences in the inferred coupling patterns, the first-order relationship between observed and null slip–coupling correlations remains consistent. In both Peru and Kamchatka the same general trends relative to the null hypothesis persist across smoothing levels (Figs S21 and S22). Distributions that lie above the null remain above and those below remain below, indicating that smoothing alters the magnitude of the slip–coupling correlation but not its sign relative to the null hypothesis, nor our interpretation of the results.

One exception occurs for the smallest earthquake magnitudes in Kamchatka, where the strongest smoothing levels, which deviate most from the minimum data misfit, produce some divergence relative to lower smoothing cases (Fig. S22). Nevertheless, the lowest smoothing levels, which are the most reliable, still indicate that coupling weakly predicts smaller earthquakes, reinforcing our original interpretation. Overall, these tests suggest that our results are robust to first order with respect to smoothing choices, although a fully consistent comparison would benefit from deriving all coupling models using a uniform inversion and smoothing framework.

Text S7 - Sensitivity of slip–coupling correlation to choice of coupling model

Coupling models can differ fundamentally across studies due to differences in the duration and type of geodetic observations, forward model assumptions, and inversion strategies⁴⁸. To evaluate how these choices affect the robustness of our slip–coupling results, we repeat our analysis using multiple published coupling models for five subduction zones where alternatives are available beyond our preferred model, including Costa Rica^{9,10}, the Himalayas^{32,36,37}, Japan^{24,49}, Nankai^{28,46,49,50}, and Mexico^{6,7}.

To ensure a consistent comparison across models, we restrict the analysis to the region of the megathrust where all coupling models overlap and include only earthquakes whose rupture areas span this domain. We then compute the mean observed and null slip–coupling correlations within this shared region. Because this requires discarding earthquakes and interface areas that fall outside the overlap, the resulting CDFs may differ from those in the main text. The purpose here is not to reproduce the primary global analysis but to isolate the sensitivity of slip–coupling correlations to the choice of coupling model. Across margins, the alternative coupling models yield broadly similar slip–coupling patterns (Figs. S23-S27). In particular, the first-order behavior is consistent across models, with the same large-scale features remaining either above or below the null expectation, although the offset between the observed and null curves varies among coupling models. Overall, this variability suggests that our qualitative conclusions about the slip–coupling relationship are robust, with most differences among coupling models confined to second-order features.

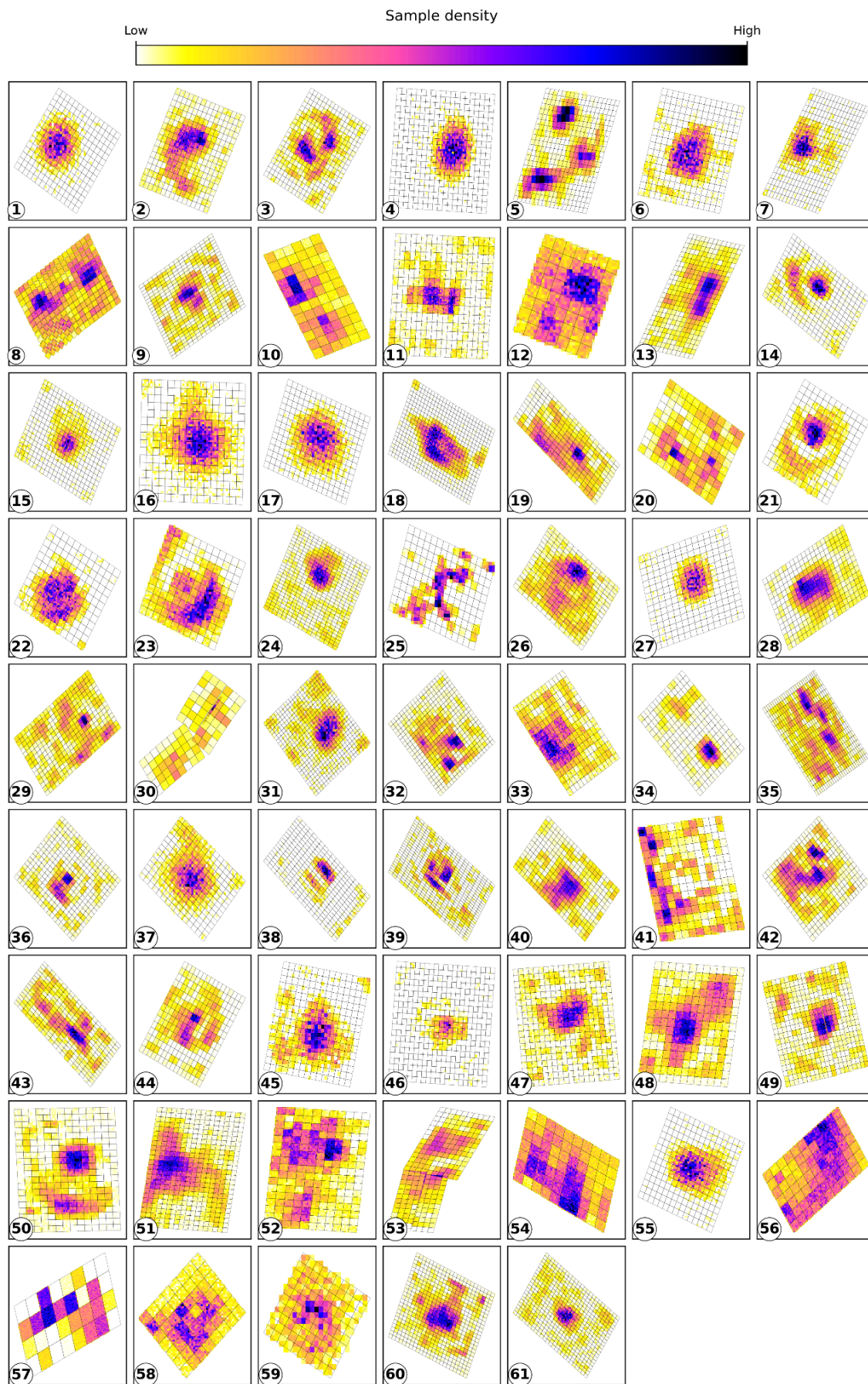


Fig. S1 – Sixty one finite slip models and their discretization. Black lines indicate the positions of subfaults comprising each model. The colormap shows the density of samples across the fault surface, with total sample counts for each model provided in Table S1. Event properties are listed in Table S1.

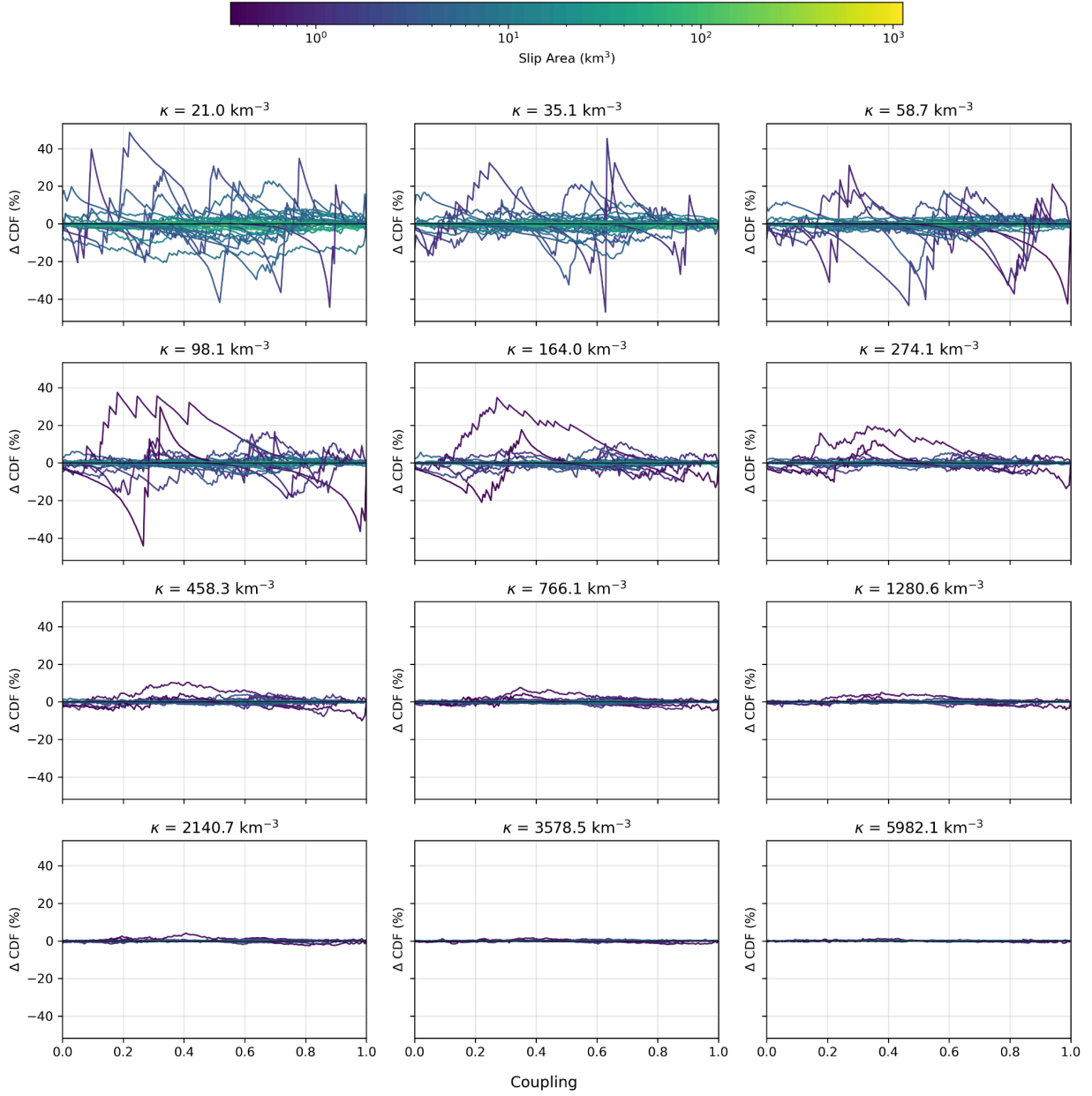


Fig. S2 – Slip–coupling difference CDFs for varying κ . Each panel shows the difference between CDFs for 61 megathrust events at a given κ , computed relative to the CDF estimated using $\kappa = 10^5 \text{ km}^{-3}$. Curves are color-coded by the total slip area of each megathrust.

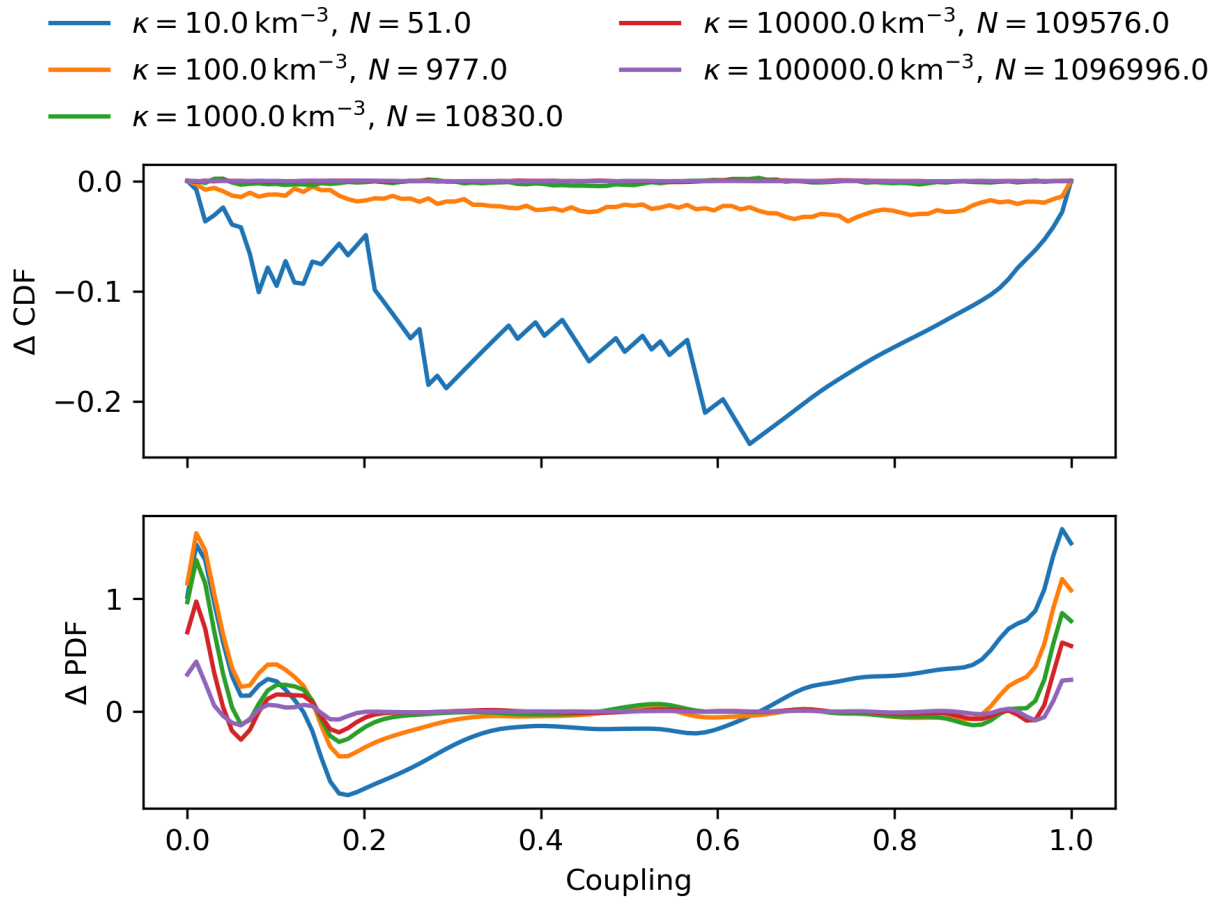


Fig. S3 – Slip-coupling difference CDFs and PDFs for varying κ . Panels show the difference between CDFs and PDFs for the 2017 Mw 7.7 Tocopilla megathrust earthquake ⁴¹ at a given κ , computed relative to those estimated using $\kappa = 5 \cdot 10^6 \text{ km}^{-3}$.

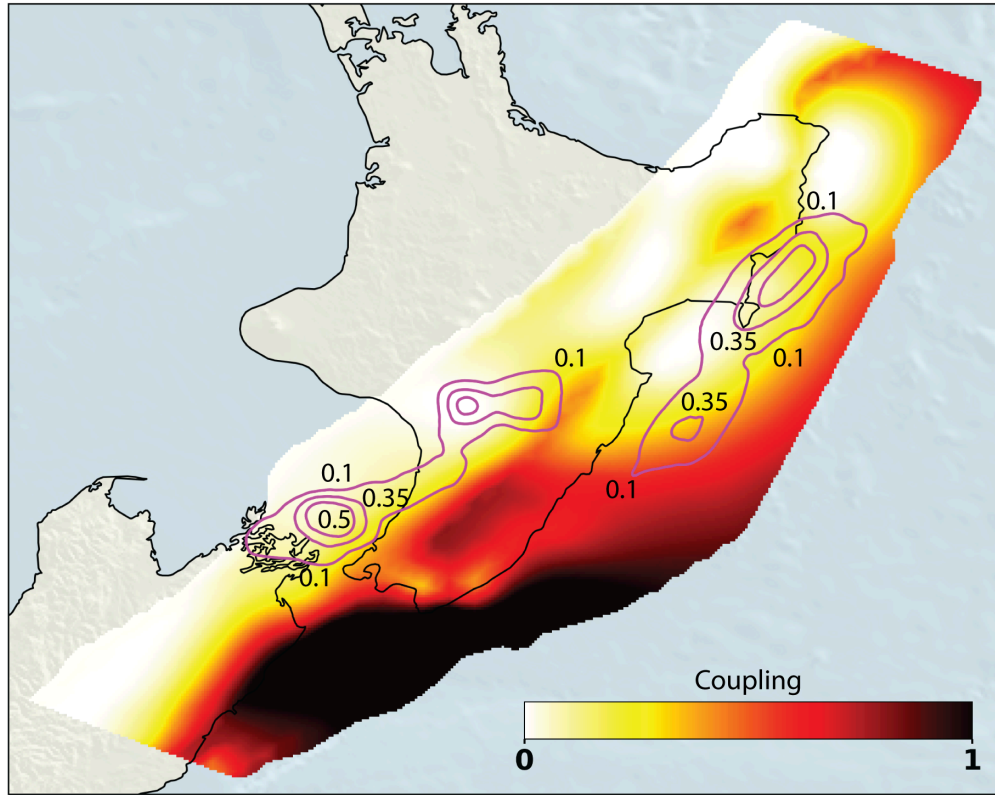


Fig. S4 – Accumulated slow slip and coupling for the Hikurangi margin. Magenta contours show three intervals of accumulated slow slip (0.1,0.35 and 0.5 meters) between 2002-2014 in New Zealand's Northern Island⁵¹. Coupling is determined from GNSS data recorded between 2006-2016⁵². We use these to compute the slow-slip-coupling correlation.

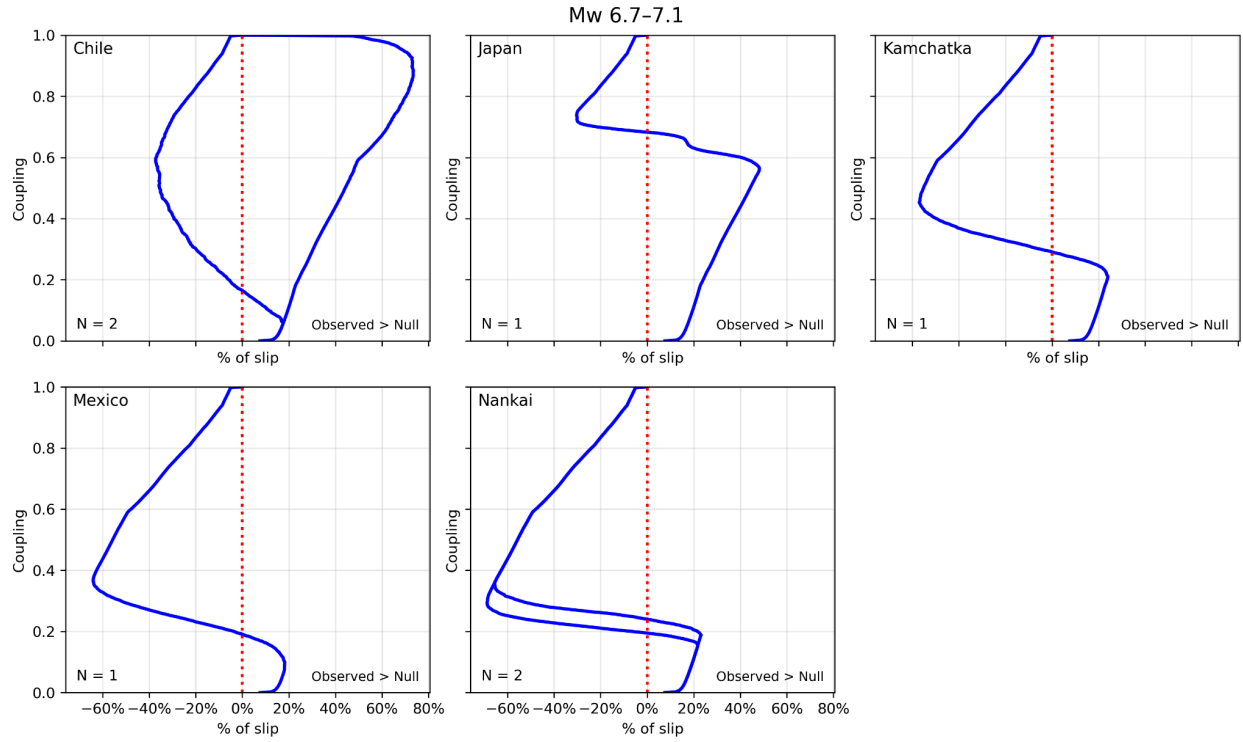


Fig. S5. Slip–coupling distributions relative to the null model for Mw 6.7–7.1 earthquakes, grouped by subduction zone. In each panel, blue curves show the difference between the observed slip–coupling cumulative distribution function (CDF) for each earthquake and the mean null CDF for the same magnitude bin. Positive values indicate excess slip relative to the null model, whereas negative values indicate less slip than expected under the null hypothesis. The vertical red dashed line marks the null solution. Numbers in each panel indicate the number of earthquakes included from that region.

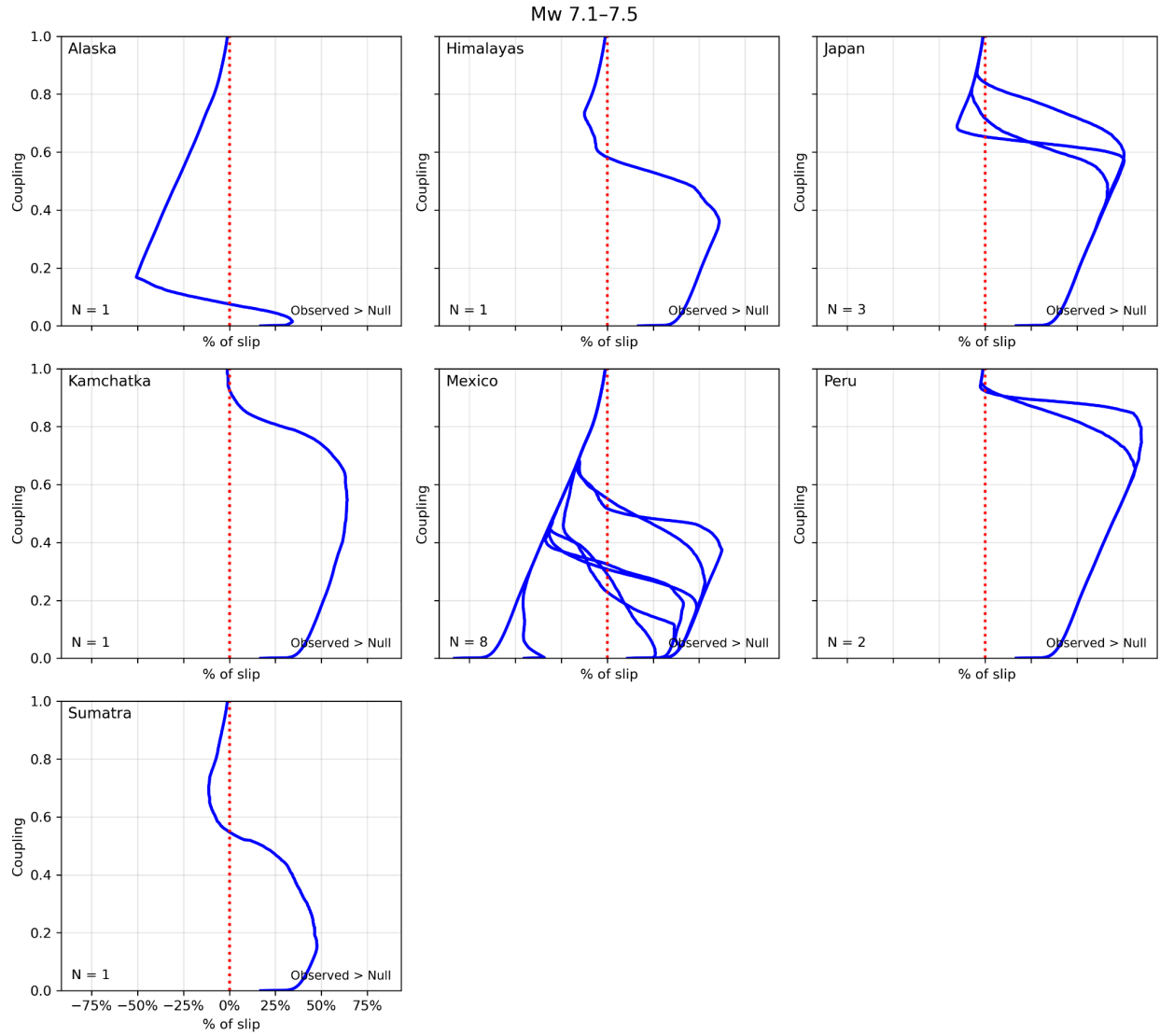


Fig. S6. Slip–coupling distributions relative to the null model for Mw 7.1–7.5 earthquakes, grouped by subduction zone. See Fig. S5 for complete figure caption.

PC1 score uncertainty projected onto the fixed observed PC1 axis

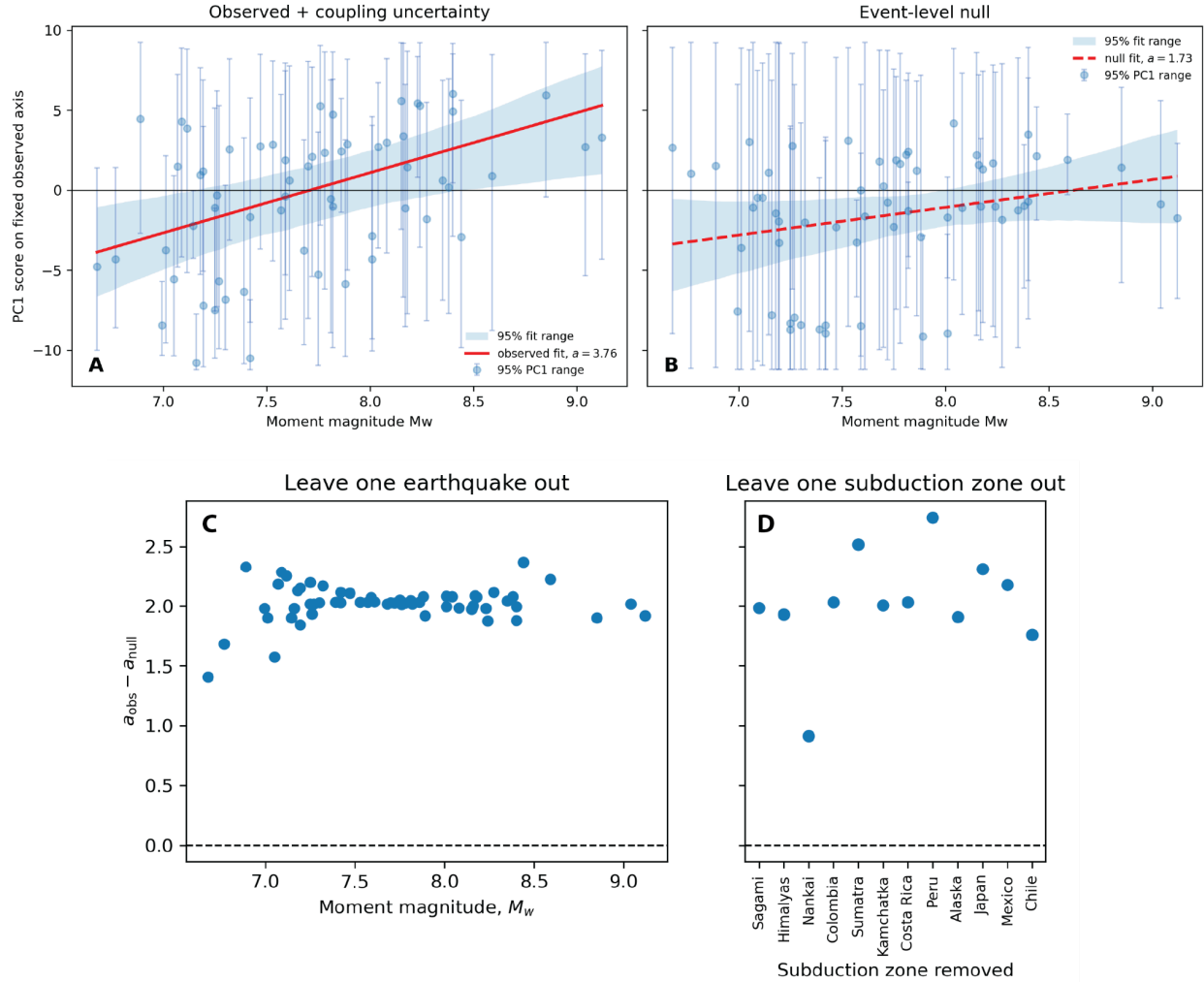


Figure S7. PCA slope analysis and leave-one-out robustness analysis. Linear relationship between PC1 and moment magnitude for the observed coupling-uncertainty ensemble (A) and the event-level null ensemble (B). Points show the mean PC1 score for each earthquake across 500 realizations, with vertical bars indicating the 95% range. The red line shows the linear fit to the mean PC1 scores, and the blue shaded region indicates the empirical 95% range of fitted relations obtained from the 500 realizations. Difference between the observed and corresponding null mean slopes after removing one earthquake at a time (C) and all earthquakes associated with one subduction zone (D).

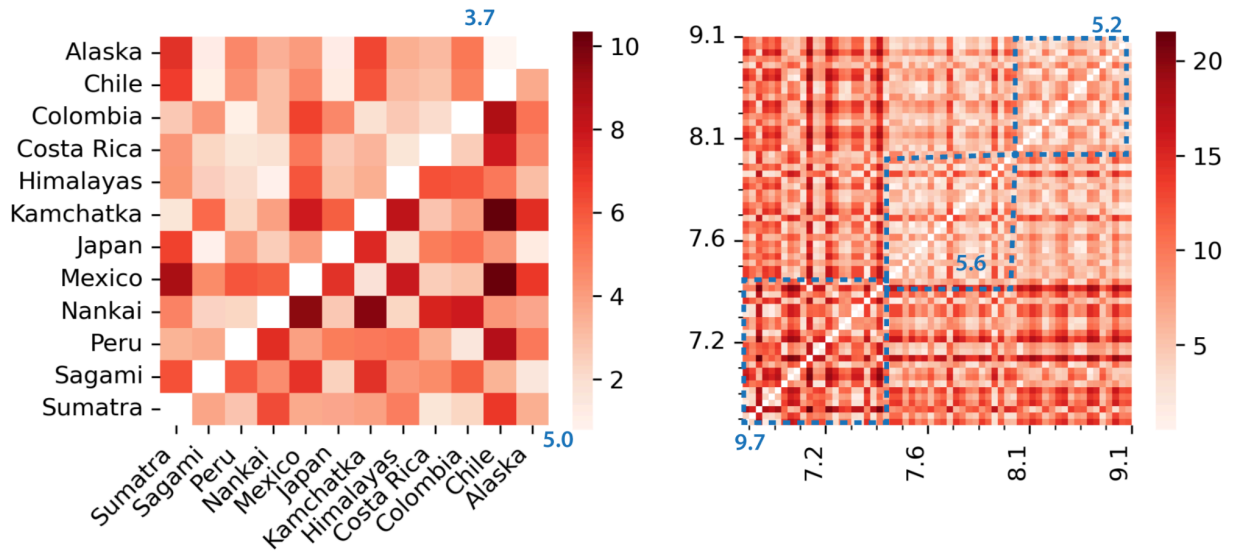


Fig S8 – Slip-coupling similarities across individual earthquakes and subduction margins. Left - Pairwise L2 distances among the 12 margins; blue numbers give mean distances for both null-corrected and uncorrected cases. Right - Pairwise L2 distances among the 61 megathrust earthquakes; blue numbers denote mean within-box values.

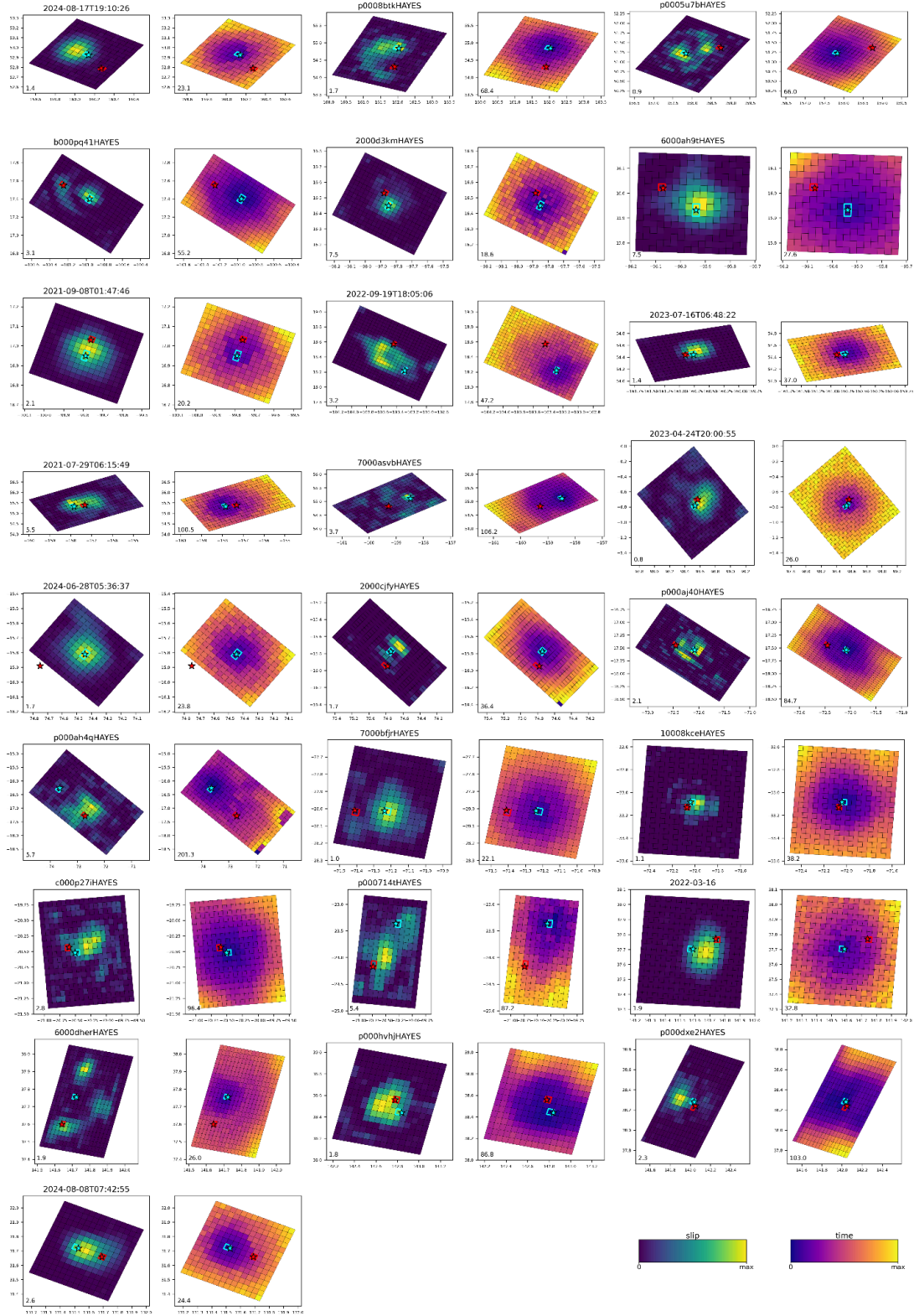


Fig S9 – Spatiotemporal patterns of slip for 25 megathrust earthquakes . Colormaps show slip and rupture rise time for each event. The upper left of each panel pair lists the event

identifier (Table S1), and the lower left gives the maximum slip and rupture rise time for that event. Red and cyan stars mark the GCMT centroid⁵³ and hypocenter, respectively.

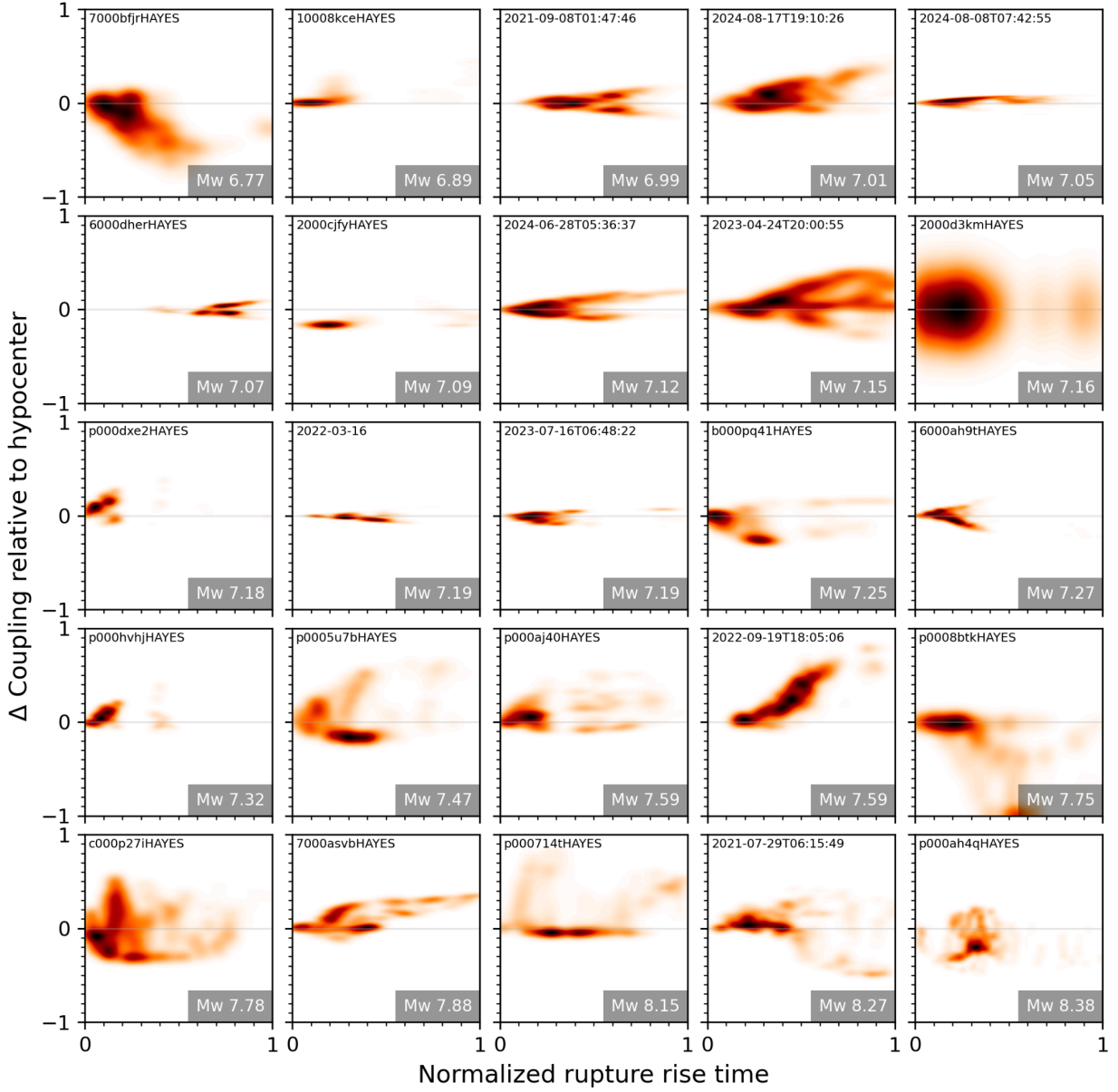


Fig S10 – Spatiotemporal patterns of slip and coupling for 25 megathrust. Colormap shows Probability density of normalized rupture rise time and coupling relative to the hypocenter for each of the 25 megathrust earthquakes. The upper left corner shows the unique identifier of each event (Table S1).

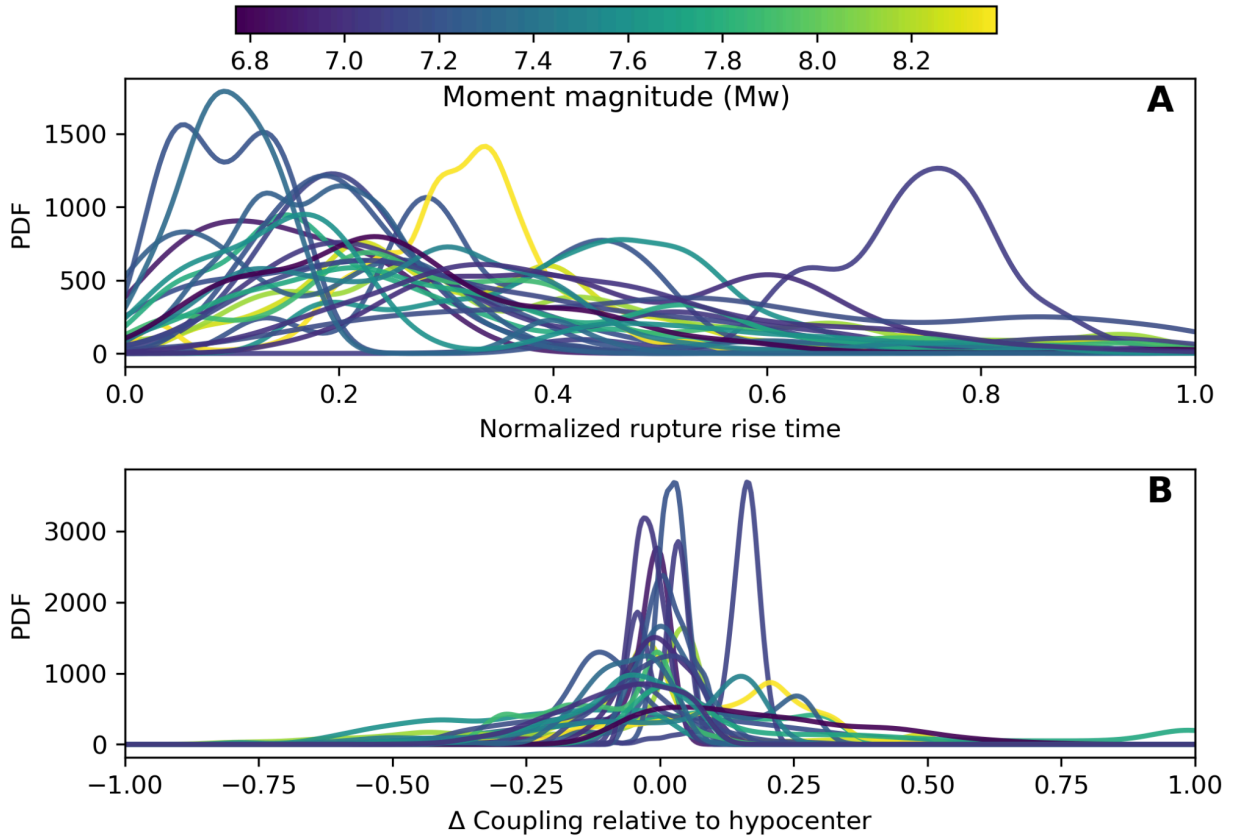


Figure S11 – Spatiotemporal patterns of slip for 25 megathrust earthquakes. Panels show PDFs of coupling relative to the hypocenter and normalized rise time for each event, color-coded by moment magnitude.

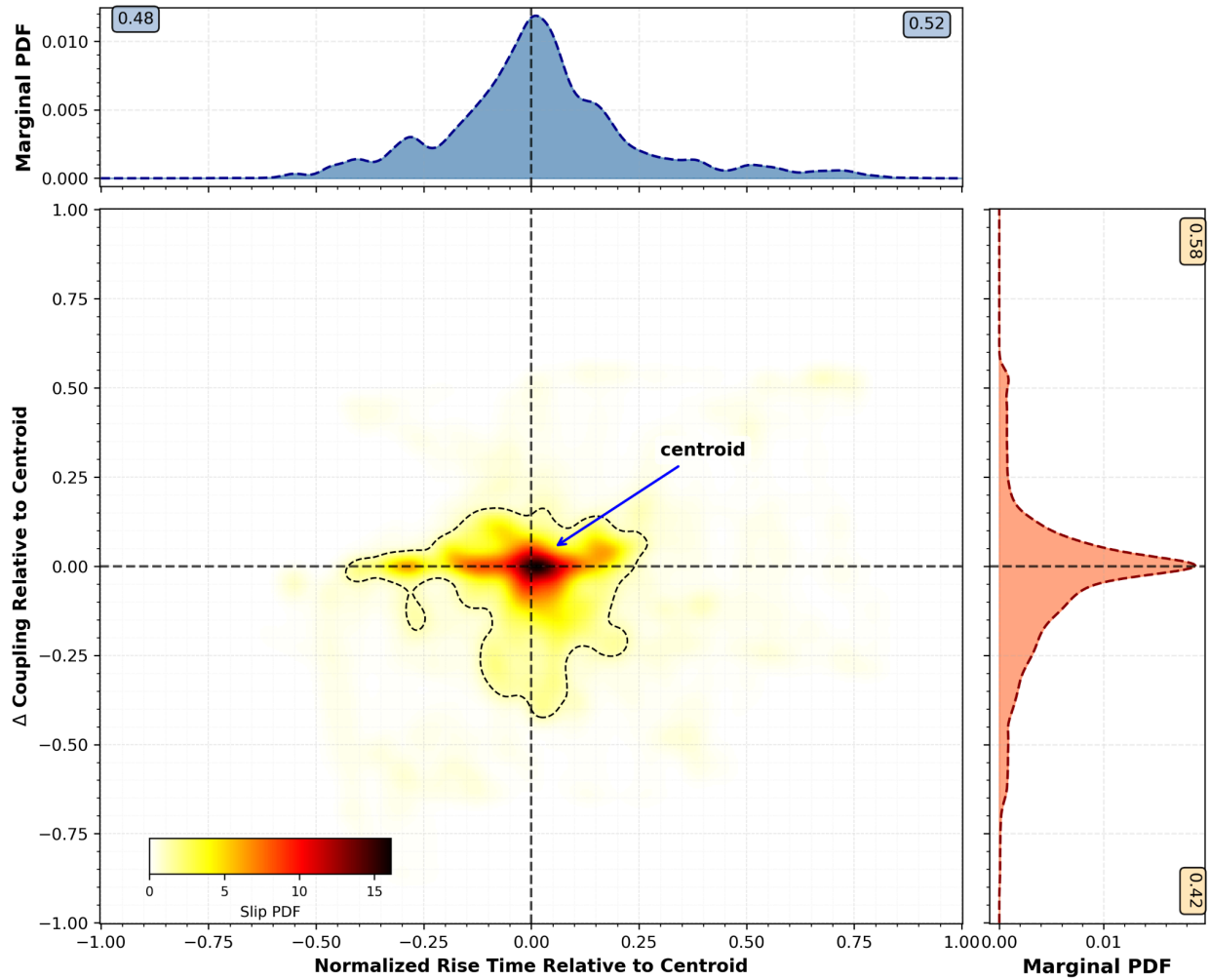


Fig S12 Spatiotemporal patterns of megathrust slip. Probability density of slip for the 24 megathrust earthquakes with a GCMT centroid as a function of normalized rupture rise time and coupling relative to the position of the centroid located at (0,0). The top and right panels show the marginal PDFs of rise time and coupling, respectively. Numbers in the marginal panels give the fraction of total weight on either side of zero.

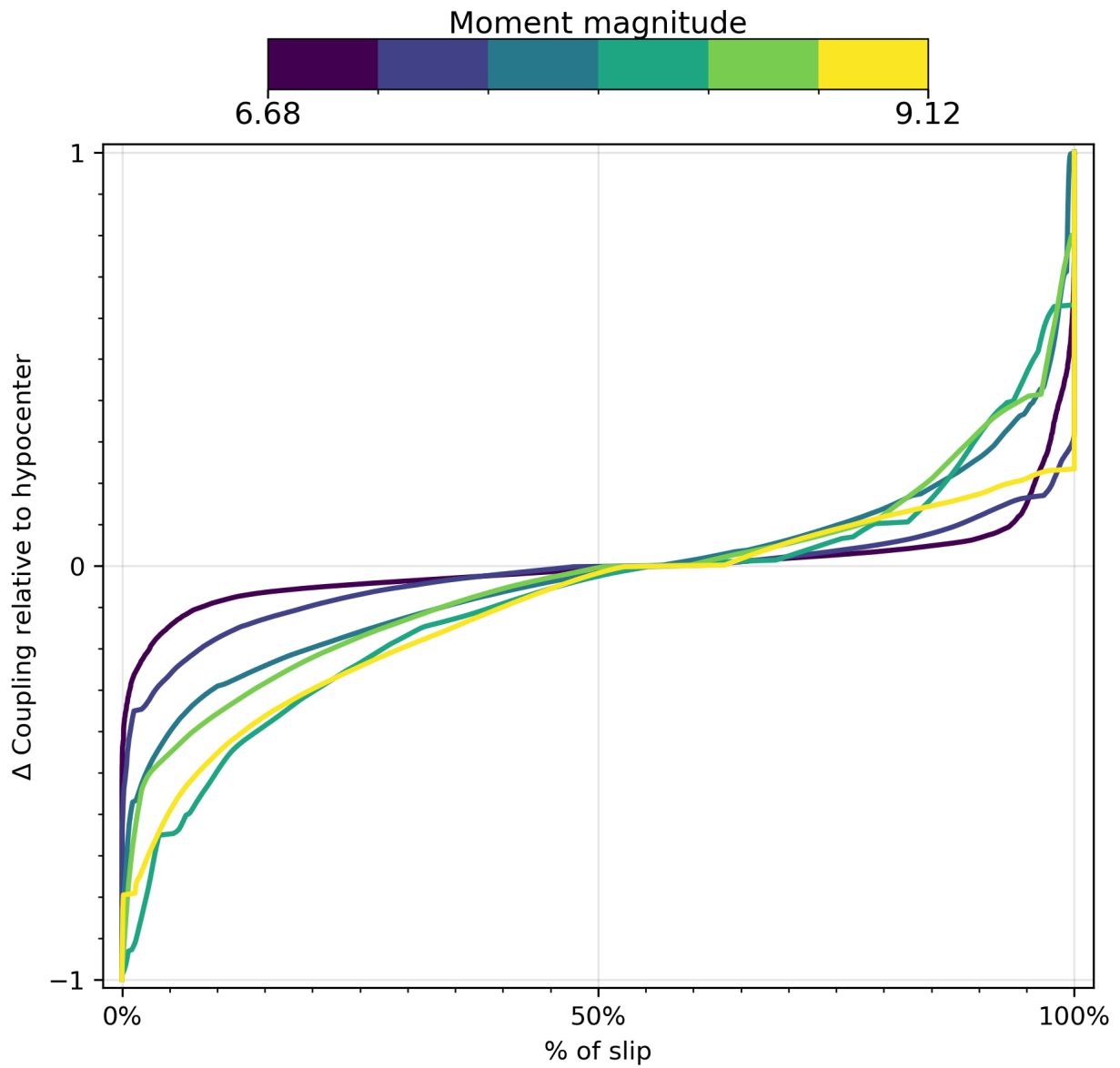


Figure S13 – CDFs of coupling relative to the hypocenter for 61 megathrust earthquakes binned into 6 groups based on moment magnitude. Slip is concentrated almost equally above coupling values above and below the initial hypocenter and does not show a strong preference across magnitudes.

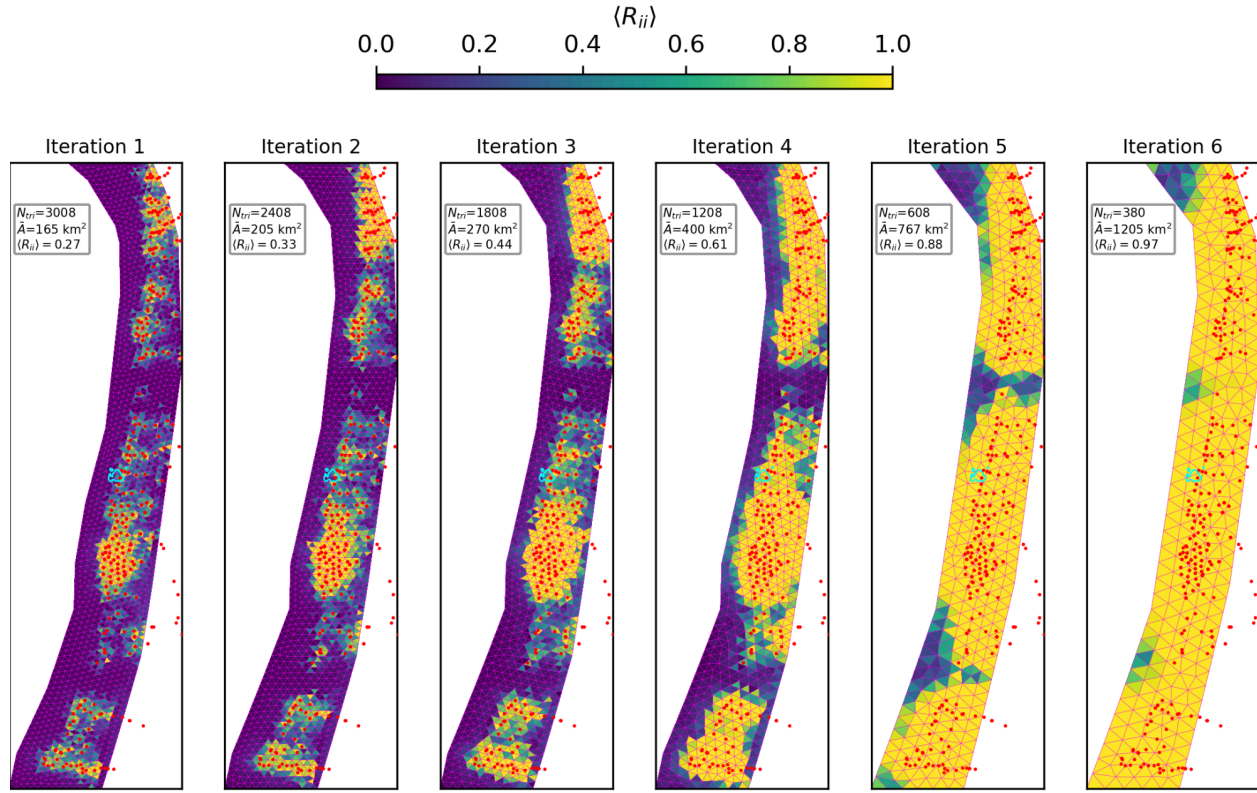


Fig. S14 – Resolution-based adaptive mesh refinement for the Chilean margin. Panels show the iterative adjustment of the triangular discretization until the mean diagonal resolution, $\langle R_{ii} \rangle$, reaches approximately 0.97 (eq. S1). Triangle colors indicate R_{ii} , the diagonal elements of the model resolution matrix, and red dots show GNSS station locations. The cyan contour outlines the rupture area of the 2020 Mw 6.8 Vallenar earthquake, approximately 1800 km². Each panel reports the number of triangles, mean triangle area, and $\langle R_{ii} \rangle$. See Text S5 for details.

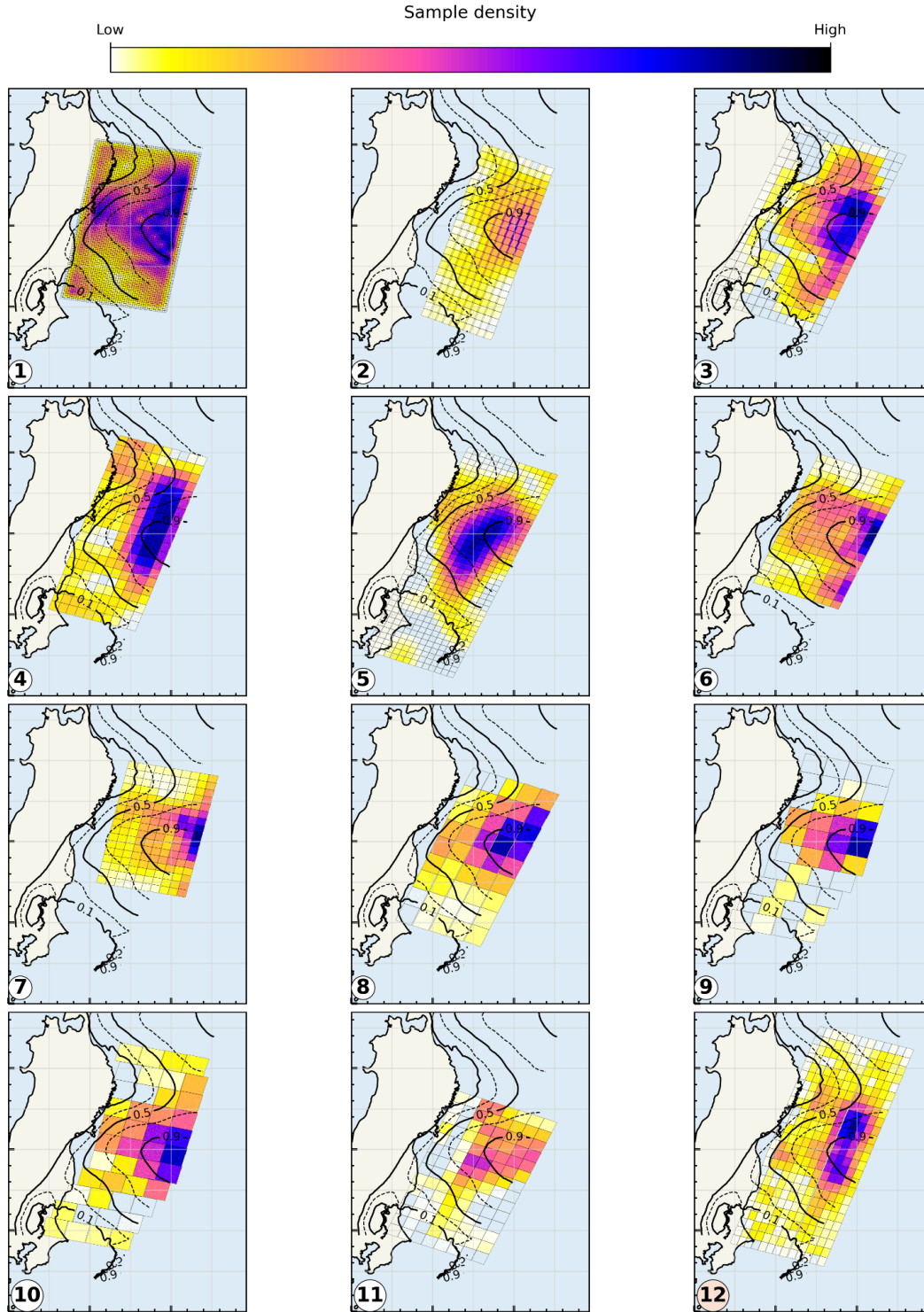


Fig. S15 – Twelve Tōhoku finite slip models^{54–64}. Black lines indicate the positions of subfaults comprising each model. The colormap shows the density of samples across the fault surface, with total sample counts for each model provided in Table S2. Black contours mark coupling values⁴⁹. Event properties are also listed in Table S2; event 12 corresponds to the model used in the main text⁴¹.

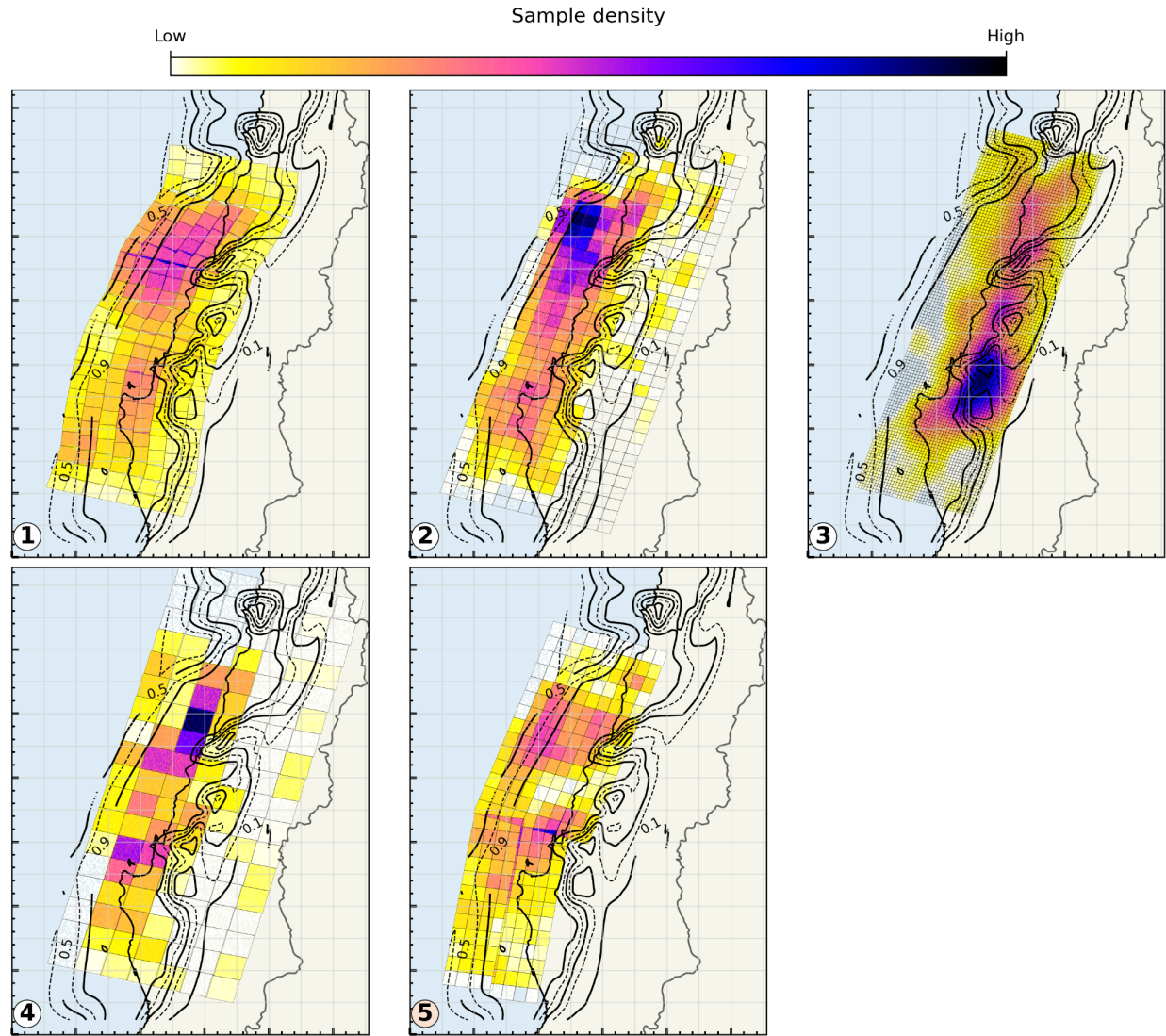


Fig. S16 – Five Maule finite slip models^{65–68} and their discretization. Black lines indicate the positions of subfaults comprising each model. The colormap shows the density of samples across the fault surface, with total sample counts for each model provided in Table S2. Black contours mark coupling values²⁰. Event properties are also listed in Table S2; event 5 corresponds to the model used in the main text⁴¹.

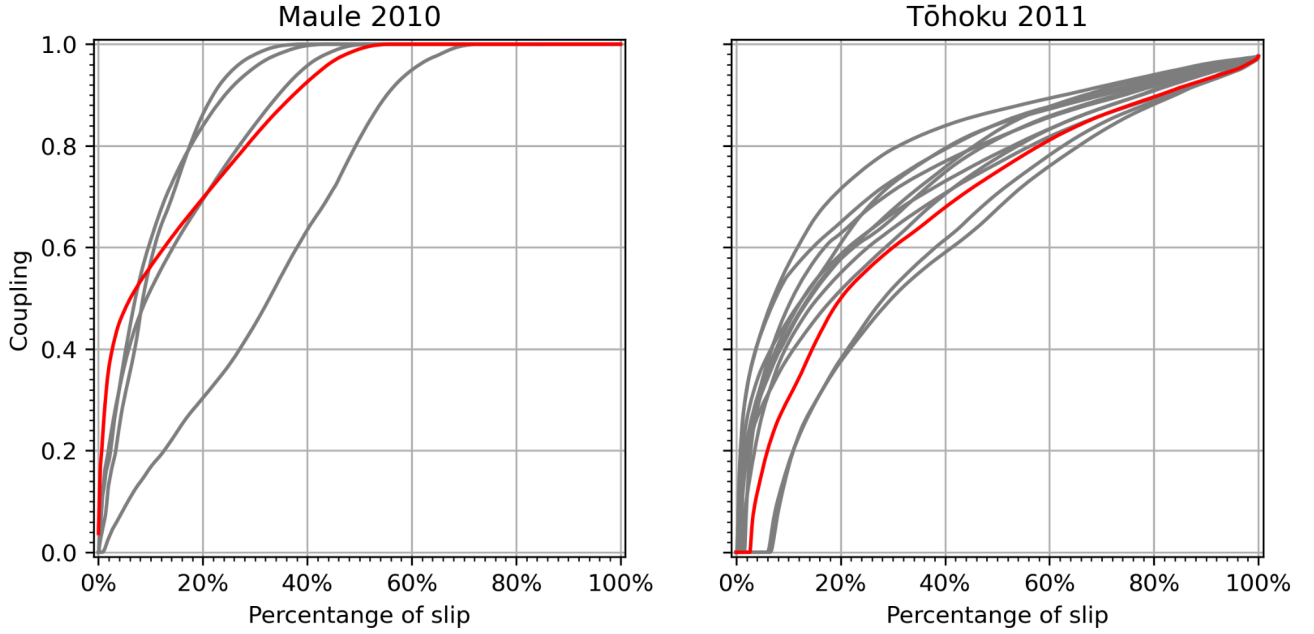


Fig. S17 – Slip-Coupling correlations for different finite slip models for the Maule and Tohoku megathrust earthquakes. Red and grey lines indicate finite slip models we used in the main text⁴¹ and additional models^{54–68} we examine, respectively.

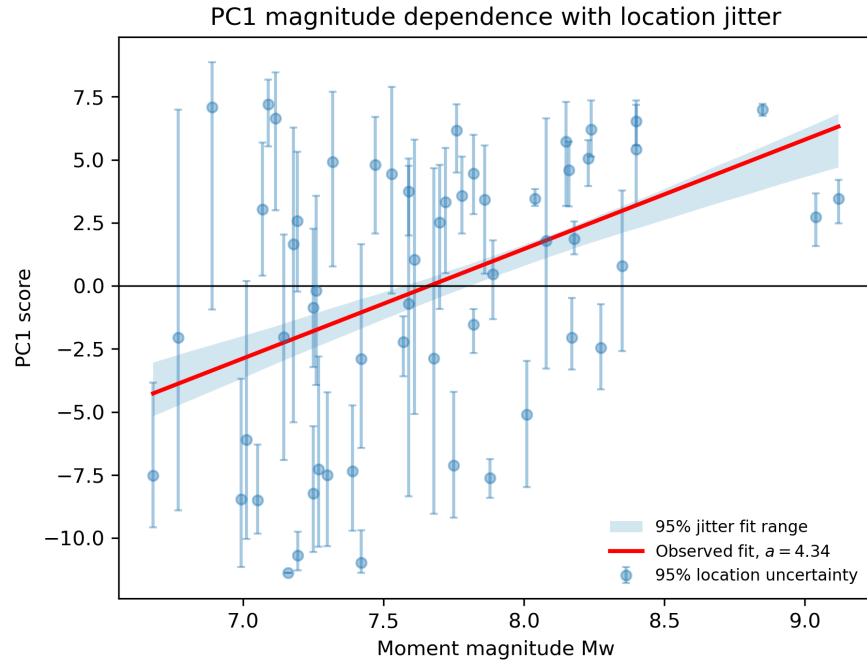


Figure S18. PCA slope analysis incorporating spatial uncertainty in finite-fault location. Linear relationship between PC1 and moment magnitude obtained from 500 spatial perturbations of the finite-fault models. Points show the mean PC1 score for each earthquake, with vertical bars indicating the 95% range across the perturbation ensemble. The red line shows the linear fit to the mean PC1 scores, and the blue shaded region denotes the empirical 95% range of regression lines obtained from the 500 realizations.

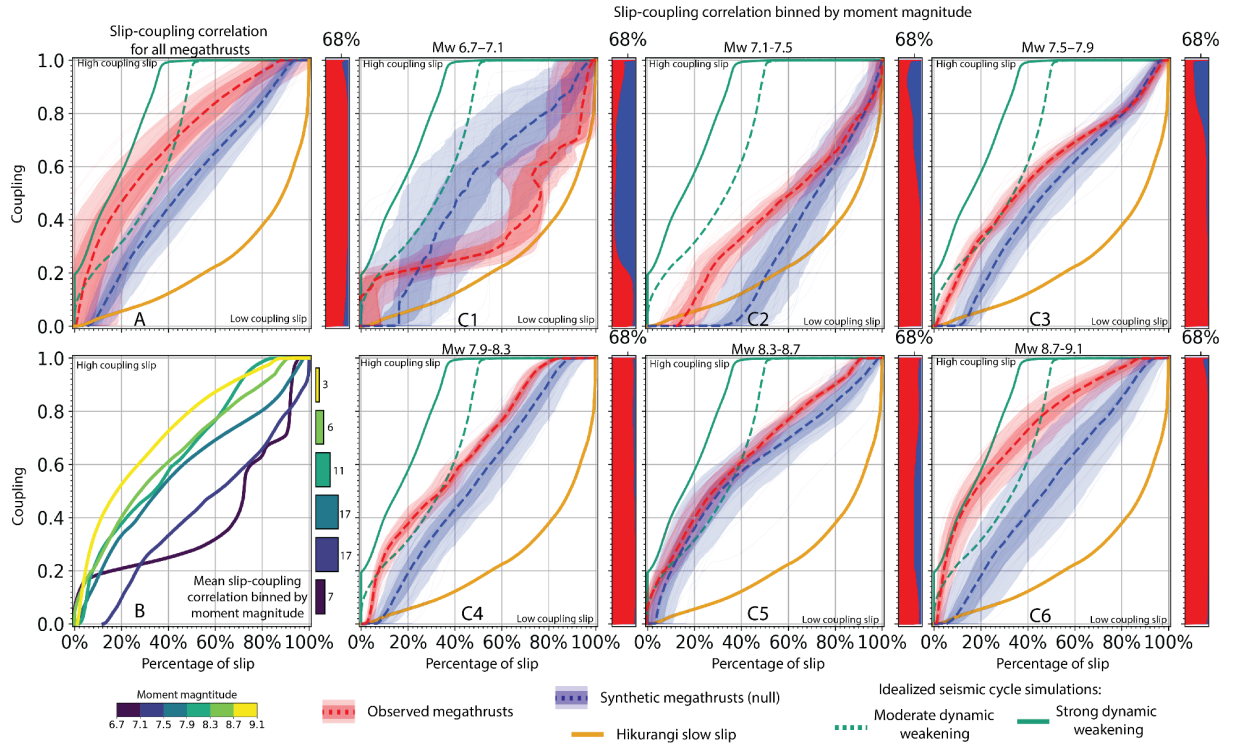


Figure S19. Slip–coupling correlations for observed (red) and synthetic null (blue) megathrust earthquakes for the uncertainty derived from the resolving power of the data (Text S5). See Fig. 3 for complete figure caption.

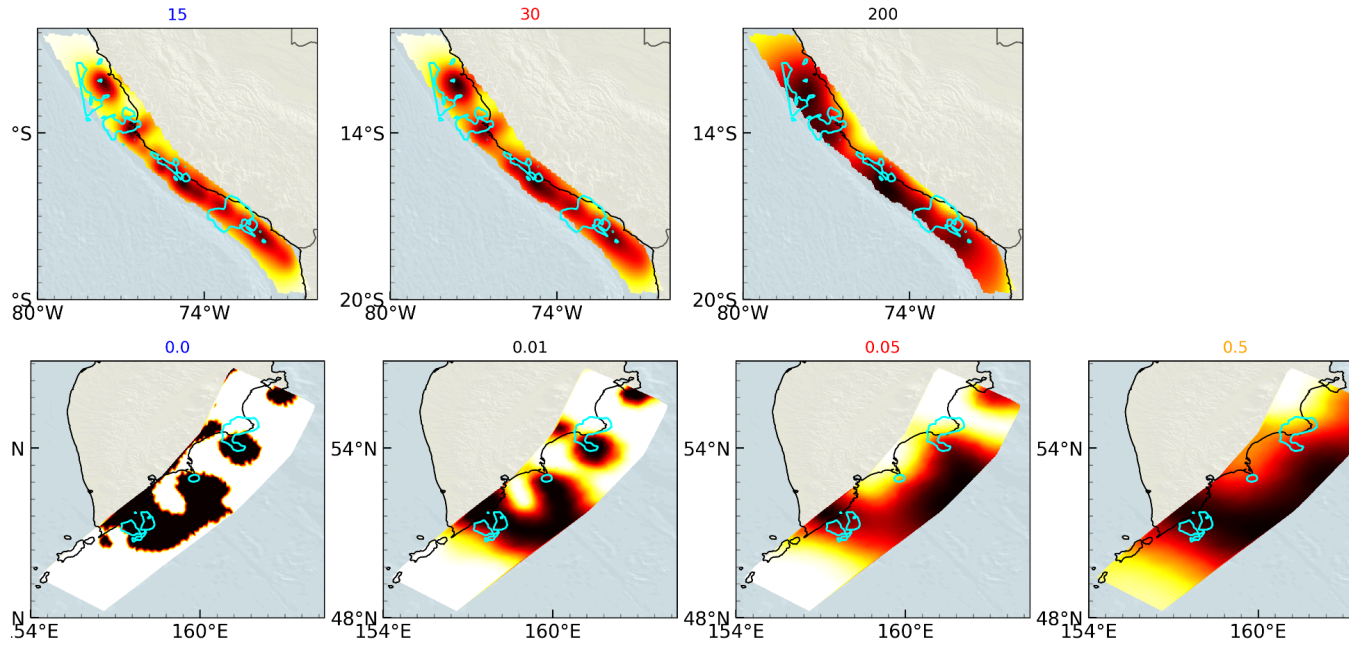


Fig. S20 – Peru¹⁵ and Kamchatka¹ kinematic coupling distributions for varying smoothing constraints. Cyan contours represent slip contours for regions where at least 30% of the maximum slip has been accumulated. Red Smoothing factors denote the preferred model used in the main text, corresponding to the authors’ favored solution.

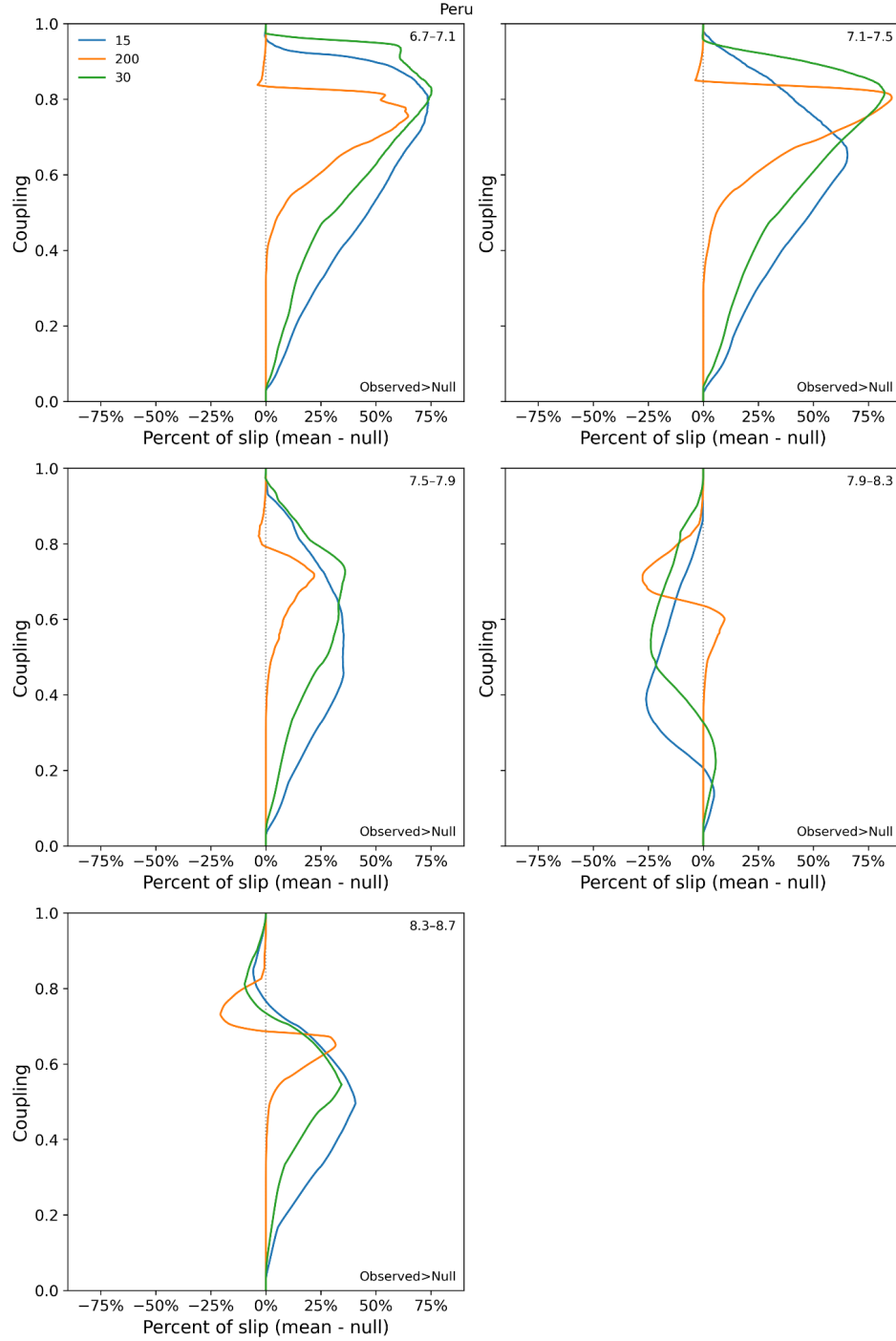


Fig. S21 – Difference between observed and mean null (computed from 500 synthetic solutions) slip-coupling correlation for varying smoothing factors in Peru¹⁵. Vertical dashed line denotes the null solution. Smoothing Value of 30 was used in the main text and is the preferred value by the authors¹⁵.

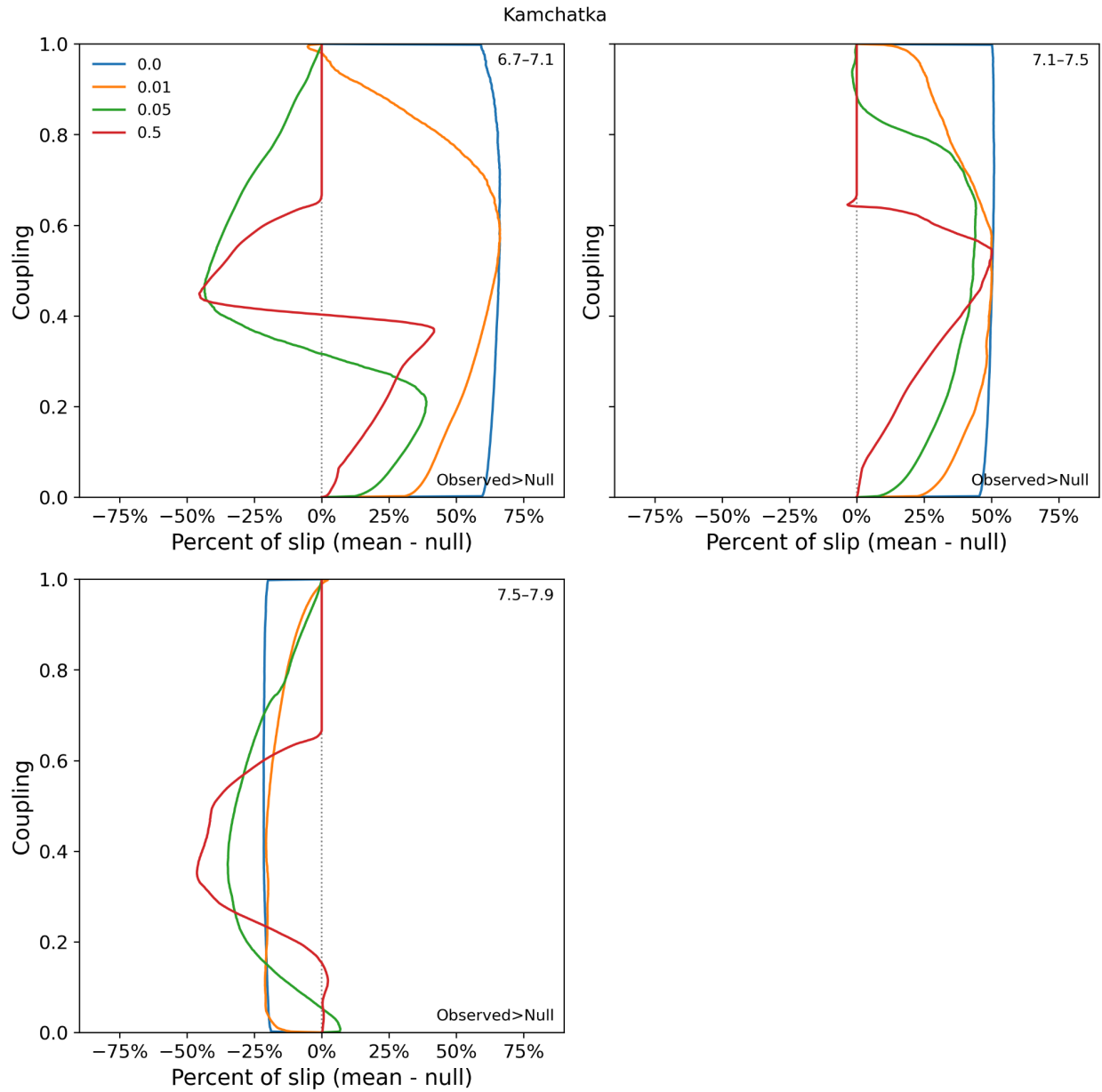


Fig. S22 – Difference between observed and mean null (computed from 500 synthetic solutions) slip-coupling correlation for varying smoothing factors in Kamchatka¹. Vertical dashed line denotes the null solution. Smoothing Value of 0.05 was used in the main text and is the preferred value by the authors.

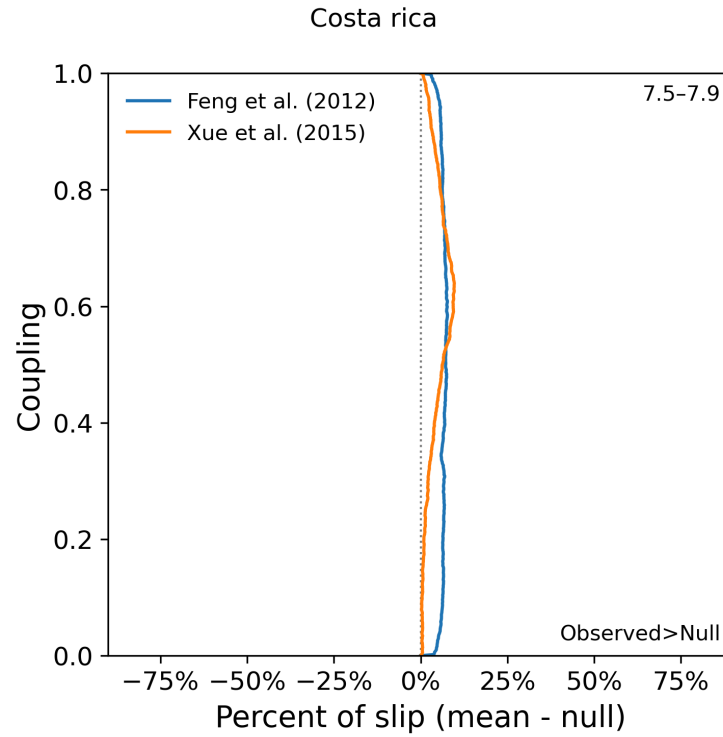


Fig. S23 – Difference between observed and mean null (computed from 500 synthetic solutions) slip-coupling correlation for different coupling distributions in Costa Rica^{9,10}. Vertical dashed line denotes the null solution.

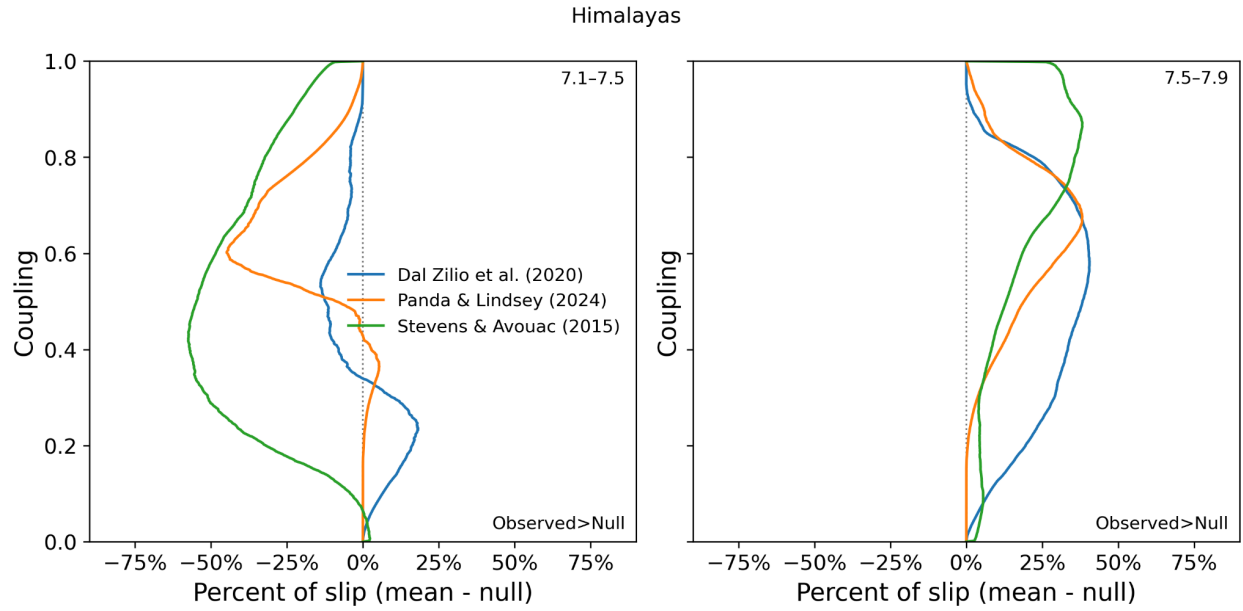


Fig. S24 – Difference between observed and mean null (computed from 500 synthetic solutions) slip-coupling correlation for different coupling distributions in Himalayas^{32,36,37}. Vertical dashed line denotes the null solution.

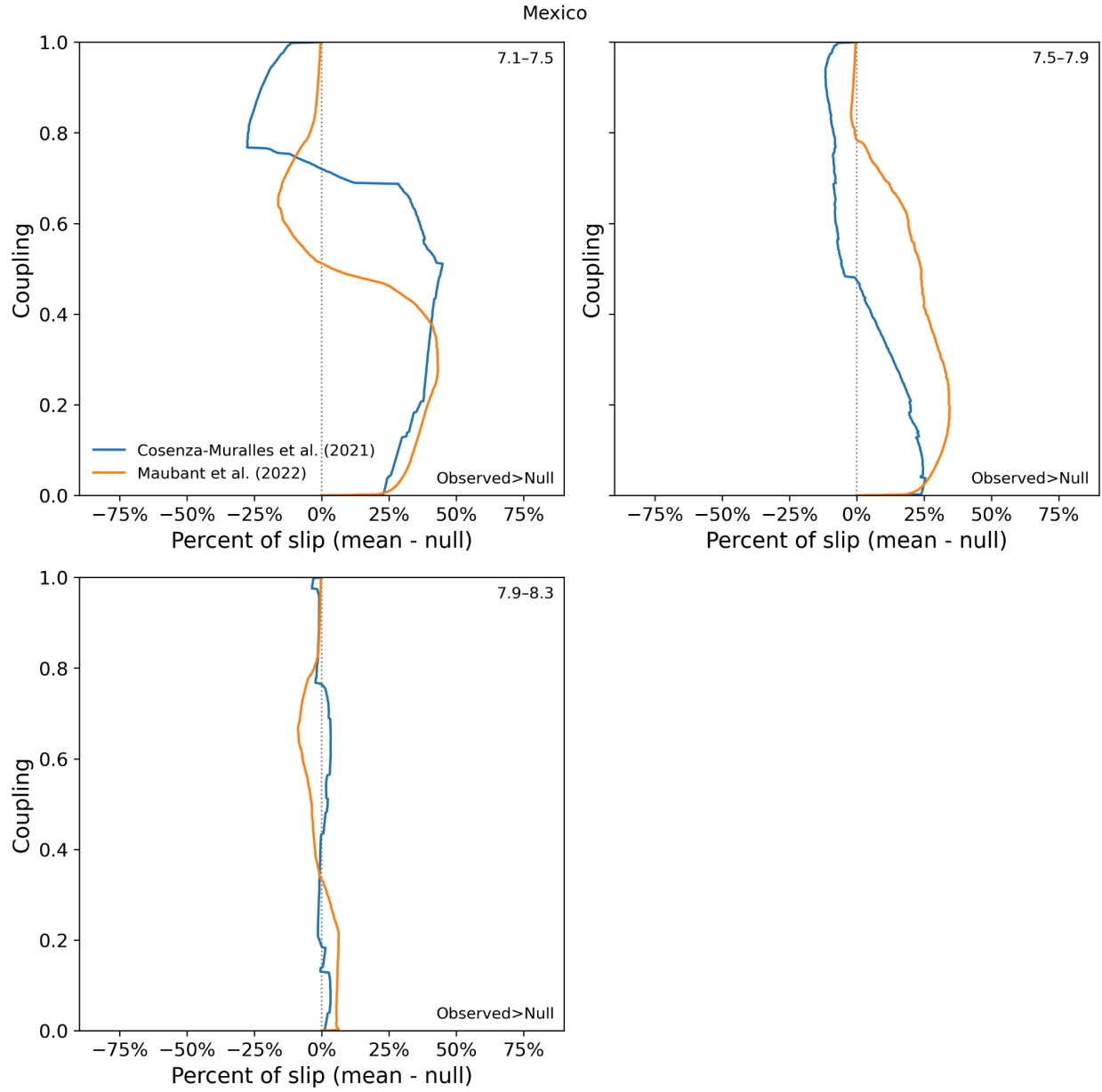


Fig. S25 – Difference between observed and mean null (computed from 500 synthetic solutions) slip-coupling correlation for different coupling distributions in Mexico^{6,7}. Vertical dashed line denotes the null solution.

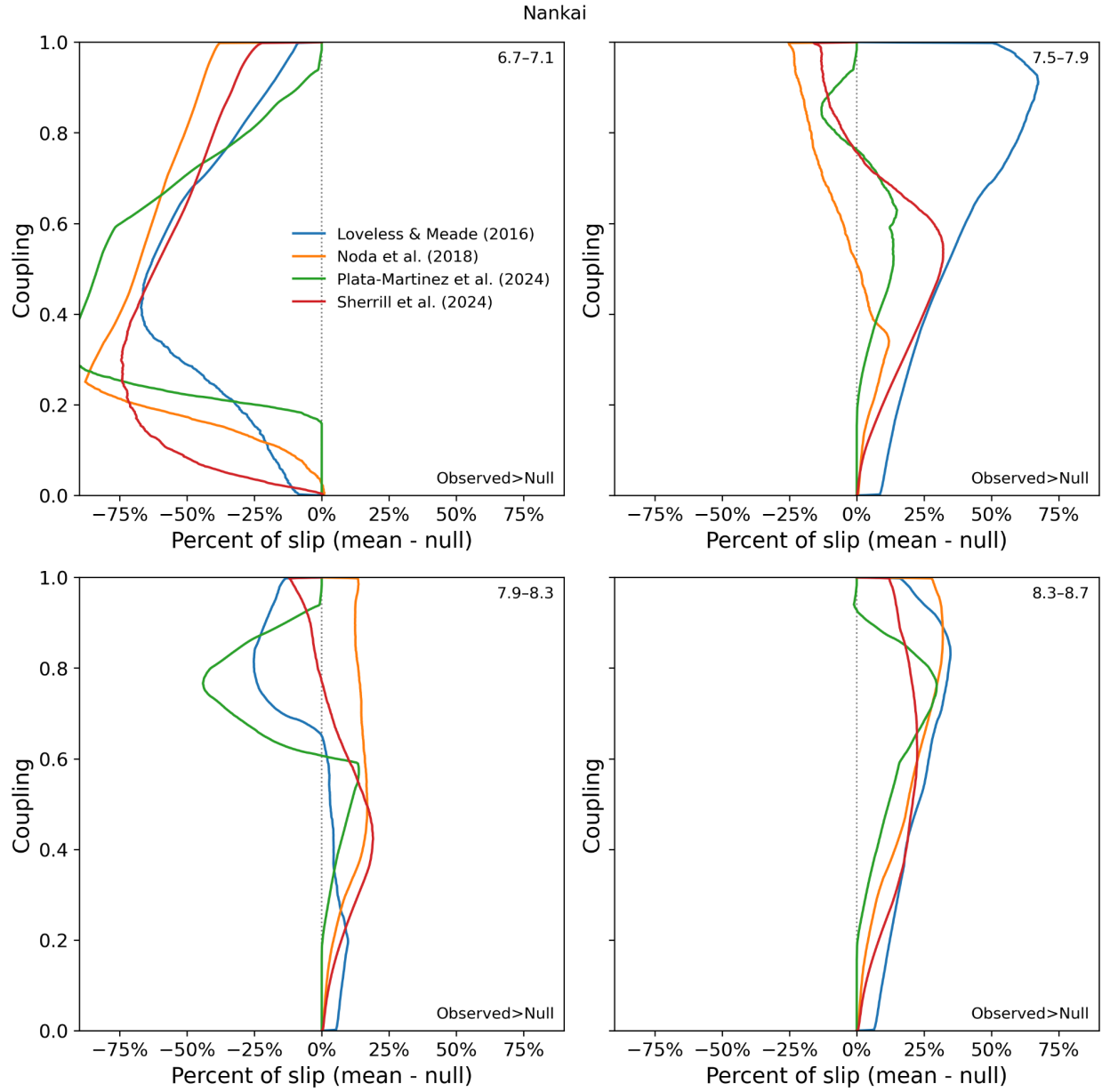


Fig. S26 – Difference between observed and mean null (computed from 500 synthetic solutions) slip-coupling correlation for different coupling distributions in Nankai^{28,49,50}. Vertical dashed line denotes the null solution.

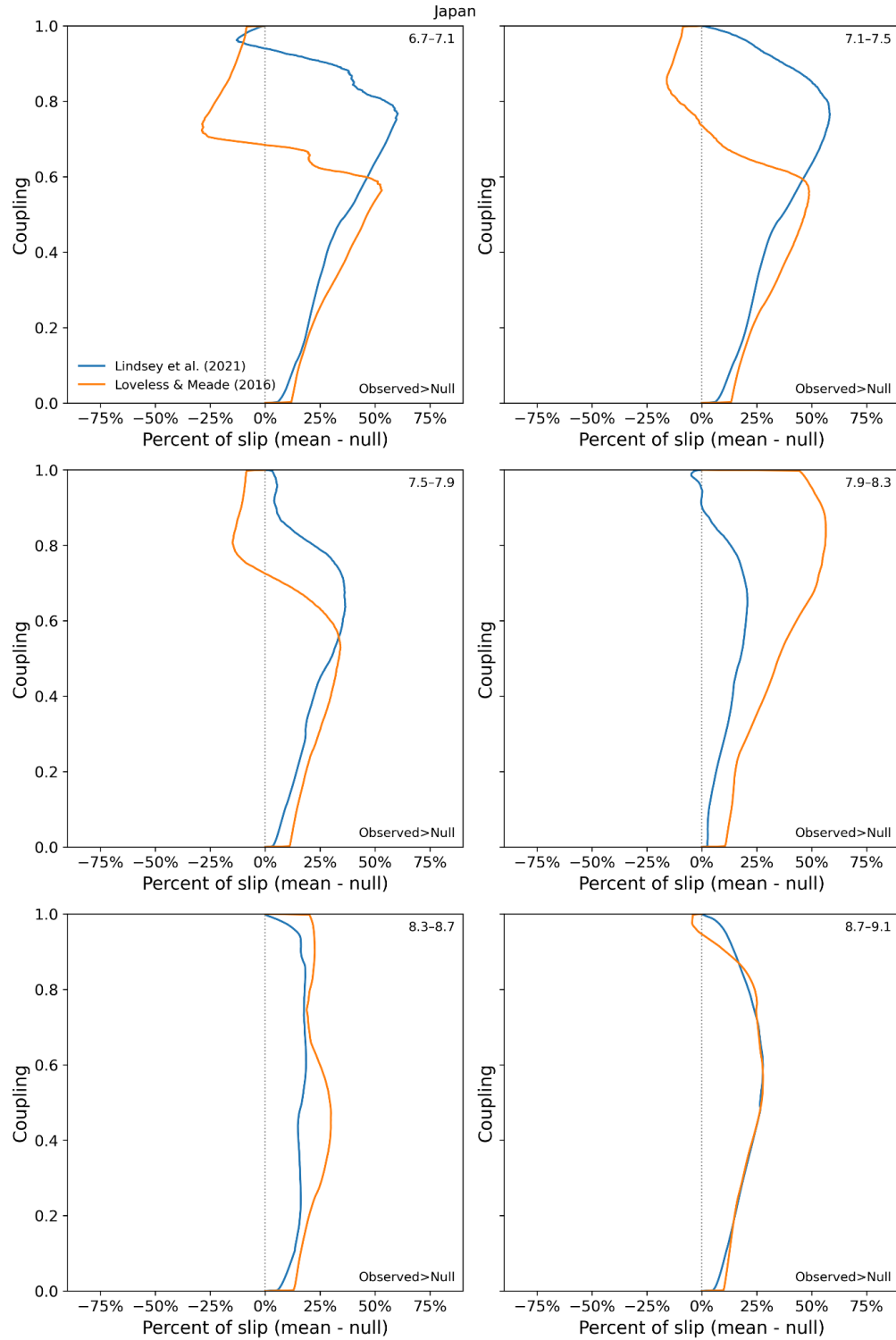


Fig. S27 – Difference between observed and mean null (computed from 500 synthetic solutions) slip-coupling correlation for different coupling distributions in Japan^{46,49}. Vertical dashed line denotes the null solution.

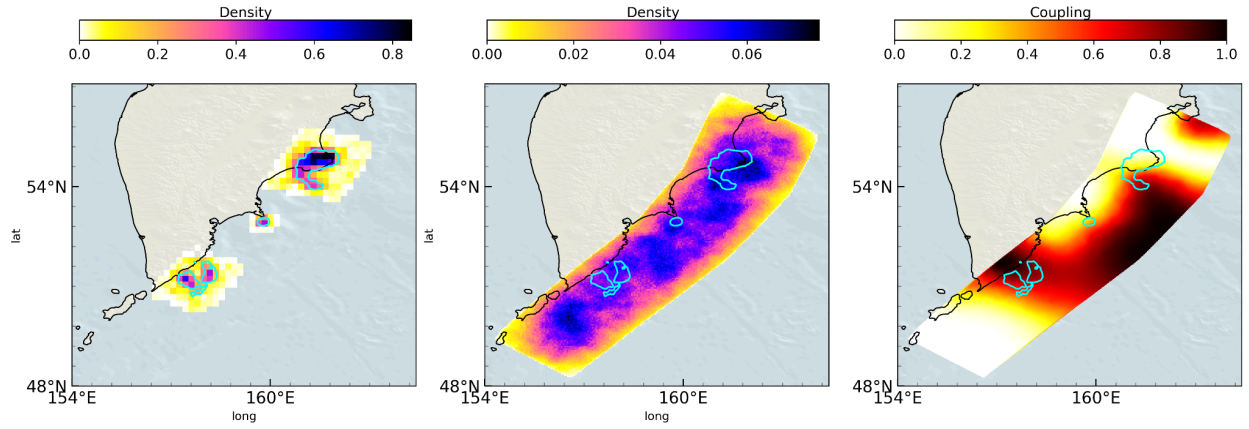


Fig S28 - Comparison between observed and synthetic (null) megathrust earthquake models in Kamchatka. A - Density of slip samples for three observed megathrusts. B - Density of slip samples from 1,500 synthetic realizations of the same megathrusts. C - Our preferred coupling model for Kamchatka¹. Cyan contours delineate regions where slip exceeded 30% of the maximum megathrust slip.

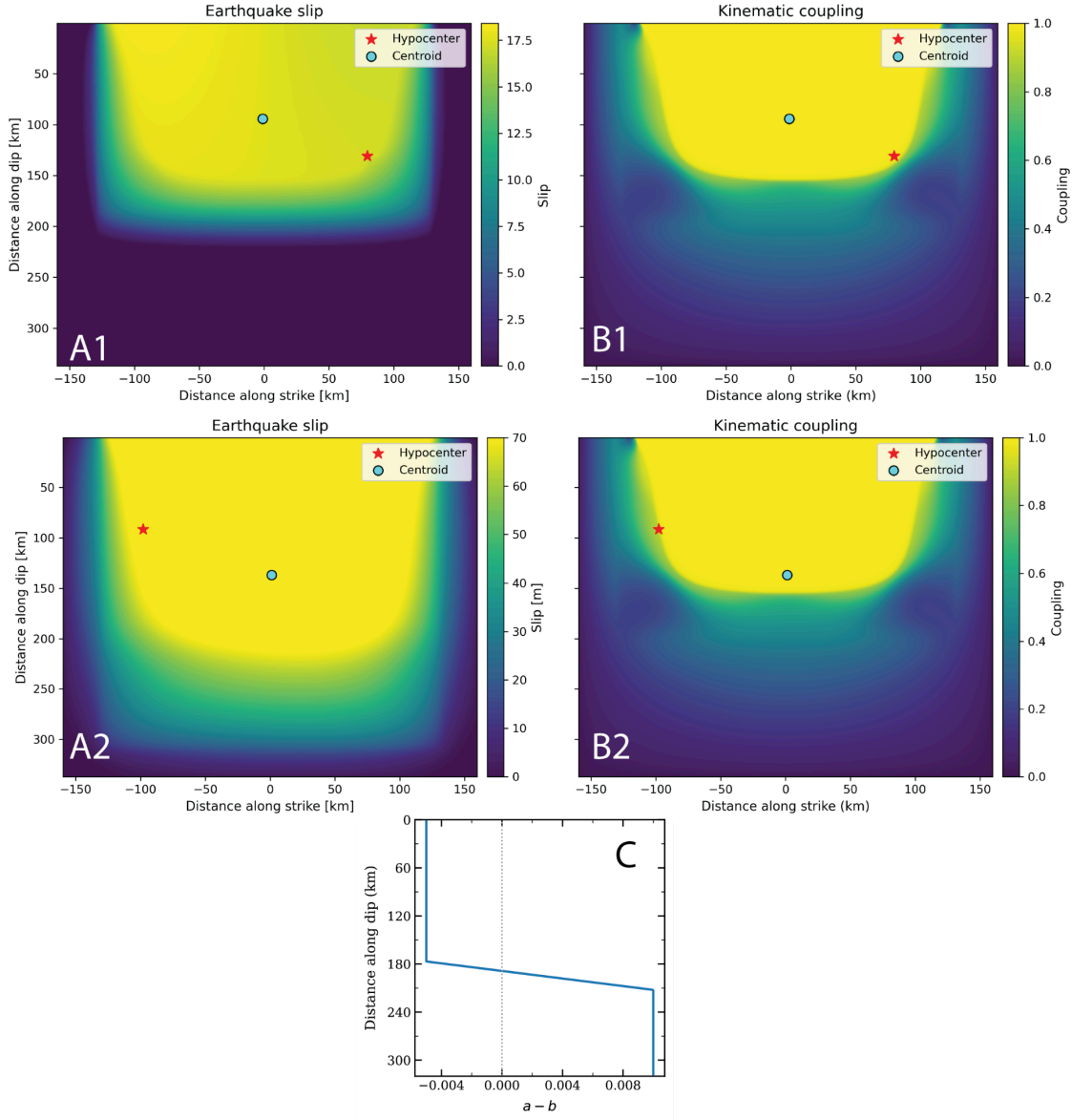


Fig. S29 - Megathrust earthquake slip and kinematic coupling in idealized 3D earthquake-cycle simulations. Results are shown for a 15° dipping planar megathrust with a seismogenic zone extending to 50 km depth and frictional properties chosen to be consistent with previous simulations. A - coseismic slip distributions for an earthquake following the spin-up phase. B - interseismic kinematic coupling fields (eq. 12). Case 1 uses moderate dynamic weakening, with $V_w = 10^6$ m/s, whereas Case 2 uses strong dynamic weakening, with $V_w = 0.1$ m/s. C - downdip frictional properties along the megathrust, shown as $a - b$. The 30 km lateral and bottom side regions are velocity strengthening, with $a - b = 0.01$ and $V_w = 10^6$ m/s.

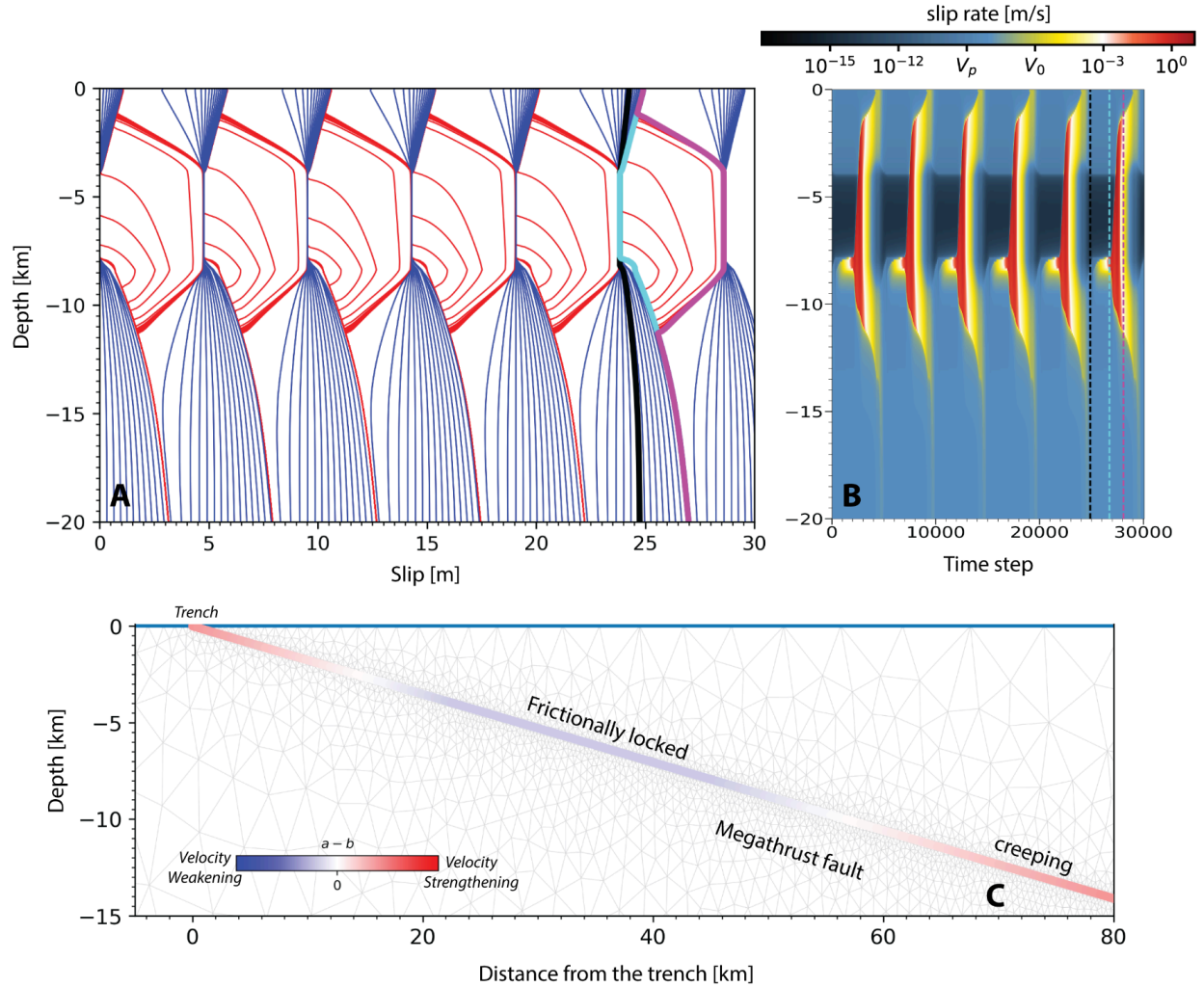


Fig. S30 - Numerical simulations of seismic and aseismic slip along an idealized 2D subduction zone over 1000 years (after the spin-up phase) using the volumetric code *tandem*⁶⁹. A - Slip contours on the megathrust, plotted every 10 seconds during coseismic rupture (red) and every 5 years during the interseismic period (blue). B - Slip rate along the megathrust. Black and cyan curves denote the 30-year interseismic period used to compute coupling (Eq. 10). Cyan and magenta mark the coseismic interval. C - Finite-element mesh used for the seismic cycle simulations. Grey lines show triangular elements, and the megathrust fault is color-coded by frictional properties guided by geodetic observations and prior simulations of Cascadia^{24,70,71}. See table S4 for complete descriptions of model parameters.

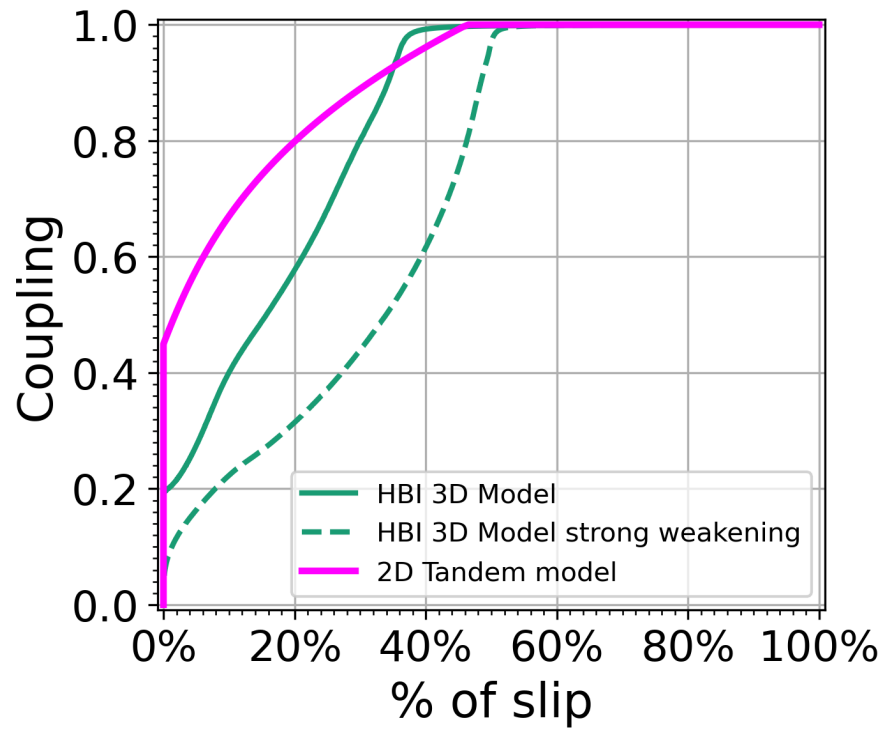


Fig. S31 - Comparison of slip-coupling correction for simulated seismic cycles, including a 2D volumetric simulation^{69,72} and a 3D boundary element model⁷³.

Table S1 - Location, date, and type and number of observations used to constrain all finite fault models used in our main text analysis. BODY refers to body-wave arrivals; GNSS to static or campaign Global Navigation Satellite System; SGM refers to strong ground motion records from accelerometers; SURF represents surface-wave observations; TELE corresponds to teleseismic broadband waveforms; TRIL indicates horizontal trilateration surveys based on distance measurements; Level refers to leveling surveys measuring vertical displacements; TIDE refers to tide-gauge records from tsunami waveforms; and Other includes miscellaneous sources not classified elsewhere. A value of 999 indicates that the data type was used in the inversion, but the number of observations is not reported. Total slip area is noted in km³.

Table S2 - Metadata for coupling used in main text including type and number of observations, dislocation size, type of geometry and forward model , smoothing type and weight.

Table S3 - Type of observations used to constrain eleven finite fault models for the Tohoku March 11, 2011, megathrust event in Japan. See Table S1 for a full description.

Table S4 - Types of observations used to constrain four finite fault models of the February 27, 2010 Mw 8.8 Maule megathrust earthquake in Chile. See Table S1 for a full description.

Parameter	Symbol	Unit	Value
Critical distance	D_{RS}	m	0.08
Reference slip rate	V_0	m/s	1e-6
Reference friction coefficient	f_0	–	0.6
Shear modulus	μ	GPa	32.04
Lamé parameter	λ	GPa	32.04
Half the shear-wave impedance for radiation damping	η	MPa·s/m	4.62
Initial normal stress	σ_n	MPa	50
Rate-and-state evolution effect parameter	b	–	0.015
Convergence rate	V_{pl}	cm/yr	3.1
Fully weakened friction coefficient	f_w	-	0.2
Weakening velocity	v_w	m/s	0.1 or 10^6

Table S5 - Model parameters used in the numerical simulations. f_w and v_w are uniform across the megathrust, except for the 30 km lateral and bottom side regions which are velocity strengthening, with $a - b = 0.01$ and $V_w = 10^6$ m/s. The depth profile of a is shown in Fig. S28C.

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