

Tectonic loading rate controls magnitude–rupture-length scaling

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Tectonic loading rate controls magnitude–rupture-length scaling

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Abstract

Seismic hazard assessments rely on scaling laws that convert mapped fault length into the magnitude a fault can produce. Yet earthquakes of similar rupture length differ substantially in magnitude. We compiled a global catalog of magnitude 6+ strike-slip earthquakes and found that slowly loaded faults host systematically larger earthquakes than equally long ruptures on fast plate-boundary faults. Physics-based earthquake-cycle simulations explain this offset. Longer interseismic intervals allow greater healing, and colder lithosphere sustains a thicker seismogenic layer, increasing stress drop, slip, and downdip rupture area. For equivalent rupture lengths, slowly deforming Mongolia yields seismic moments up to six times larger than the San Andreas Fault. Loading rate is a measurable control on maximum earthquake size that belongs in hazard-assessment scaling laws.

Rupture length is a key scaling parameter for estimating the magnitude of the largest earthquake a tectonic fault is capable of producing, because it can be reconstructed for historical, prehistoric, and poorly instrumented events when direct magnitude estimates are unavailable. Empirical magnitude–rupture-length relations therefore underpin fault-based seismic hazard models, especially where instrumental records are short, serving two complementary roles. The rupture length of a mapped prehistoric

event is converted into an expected magnitude and, conversely, a specified magnitude is converted into a rupture geometry used to position finite sources relative to exposed sites. Yet the available data show that earthquakes with nearly identical rupture lengths can differ in moment magnitude (M_w) [1] by 0.5 units or more, corresponding to an approximately sixfold difference in seismic moment [2]. This scatter is commonly treated as irreducible variability. Its physical origin, however, determines whether standard scaling relations systematically underestimate the largest earthquakes that some faults can produce.

The great strike-slip earthquakes of the Mongolia–Altai region illustrate the problem. During the 20th century, the 1905 Tsetserleg ($M_w \sim 8.0$), 1905 Bulnay ($M_w \sim 8.4$), 1931 Fuyun ($M_w \sim 8.0$), and 1957 Gobi–Altai ($M_w \sim 8.1$) earthquakes ruptured discrete faults within the Eurasian plate interior (Fig. 1a) [3–5]. These faults accumulate strain far more slowly than major plate-boundary faults. Paleoseismic observations near Mogod record at least three surface-rupturing earthquakes during the past 50,000 years on a fault slipping at only $0.06 \pm 0.01 \text{ mm yr}^{-1}$ [4]. The Tsetserleg fault accommodates approximately 0.3 mm yr^{-1} [6], two orders of magnitude below the approximately 30 mm yr^{-1} rate of the San Andreas Fault (Fig. 1b)[7]. Geodetic and geological observations show similarly slow deformation across Mongolia and the broader Asian interior [8–15]. The occurrence of several $M_w \geq 8$ earthquakes on such slowly loaded faults has therefore posed a long-standing question. How can faults that accumulate strain at such low rates generate earthquakes that are unusually large for their rupture length [16, 17]?

Earthquake size is set by seismic moment [18],

$$M_0 = GDA, \quad (1)$$

where G is shear modulus, D is average coseismic slip, and A is rupture area. For a fixed rupture length and shear modulus, a larger moment therefore requires greater slip, greater down-dip rupture width, or both. Earlier work found that intraplate earthquakes tend to have larger stress drops and slip than interplate earthquakes of comparable length, with stress drops differing by about a factor of six on average [19]. That contrast was attributed to stronger intraplate faults, but the process that makes them stronger remained unresolved. Statistical studies later examined whether the corner moment of earthquake populations depends on plate rate and concluded that the dependence was weak [20, 21]. Those analyses, however, sampled rates only approximately ranging from 12 to 110 mm yr^{-1} [21], which spans only one order of magnitude. They also did not identify a mechanism connecting loading rate to event-scale stress drop, slip, and rupture area. In contrast, Mongolian tectonic slip rates are approximately two to three orders of magnitude lower than those of the San Andreas Fault, highlighting the potential difference between ultra-slow continental faults and fast plate-boundary faults.

Results

Observational evidence

To investigate the origin of this scatter, we compiled a global catalog of 50 historical and recent M_w 6+ strike-slip earthquakes from California, Turkey, Iran, China, Japan, New Zealand, Alaska, and Mongolia, extending the dataset of [22] with tectonic slip-rate estimates for each fault (table 1), and compared them with the widely used relation $M_w = 5.56 + 0.87 \log_{10}(L)$ [23], where L is rupture length; more recent self-consistent compilations obtain similar trends and comparable scatter [24]. We found that events on faults slipping slower than 5 mm yr^{-1} define a systematically higher magnitude–length trend than events on faster faults (Fig. 1c). The separation approaches 0.5 magnitude units for long ruptures; the four Mongolia–Altai earthquakes lie consistently above, and two San Andreas earthquakes below, the standard empirical curve. The catalog spans rupture lengths from approximately 20 to 460 km and tectonic slip rates from approximately 0.3 to 50 mm yr^{-1} ; the offset is therefore not restricted to one earthquake size or fault system but varies across a continuous range of loading conditions. This systematic offset is comparable in size to the intrinsic scatter of the length-based relations themselves, approximately 0.25–0.35 magnitude units [23], indicating that loading rate accounts for a substantial, physically explicable share of variability that fault-based hazard models otherwise treat as unstructured.

Mechanisms increasing seismic moment for a given rupture length

We propose that slow loading increases moment for a given rupture length through two complementary mechanisms affecting slip and rupture area in Eq. 1. First, slow loading lengthens the interseismic interval, allowing the fault state to evolve toward greater frictional strength before failure. The resulting earthquake then begins from a stronger state, produces a larger peak-to-residual stress drop, and accumulates more coseismic slip. Second, slowly deforming continental interiors are commonly underlain by cold, mechanically strong lithosphere, which allows brittle behavior to extend to greater depth, broadening the velocity-weakening seismogenic region and increasing the rupture area available for a given along-strike length. We tested this hypothesis with physics-based, dynamic simulations of the earthquake cycle, complemented by rate-and-state friction theory.

The two mechanisms are complementary. Healing sets the stress state from which rupture initiates, and hence the slip released over the available fault area, whereas thermal structure sets that area by fixing the depth over which unstable friction operates. Their combination is most effective for long strike-slip ruptures, for which additional width and slip raise moment without lengthening the mapped surface rupture – the quantity best preserved in the geological record.

Geotherm

Mongolia and the San Andreas Fault provide representative end-members for evaluating this coupled effect. The San Andreas system accommodates a large fraction of Pacific–North America motion and is characterized by a comparatively warm crustal geotherm. In contrast, the faults of Mongolia lie within slowly deforming continental lithosphere that is colder and mechanically stronger over the relevant depth range. This association between slow deformation rate and cold lithosphere is common in stable continental interiors [16, 17, 25, 26], although loading rate and temperature remain physically distinct controls. We constructed yield-strength envelopes using representative geotherms for Mongolia and the San Andreas system [27–31] and standard

116 flow laws (Methods) [32]. The Mongolian geotherm, approximately 18°C km^{-1} , is sub-
 117 stantially colder than the approximately 35°C km^{-1} San Andreas geotherm over the
 118 seismogenic depth range (Fig. 2a).

119 Because ductile strength decreases rapidly with temperature, the colder model
 120 retains appreciable strength to greater depth. The brittle–ductile transition occurs at
 121 approximately 24 km in the Mongolian end-member and 14 km in the San Andreas
 122 end-member (Fig. 2b). Brittle strength in the shallow crust remains broadly com-
 123 parable, but effective normal stress and frictional strength continue to increase with
 124 depth to the transition (Fig. 2c). We use this transition to prescribe the base of the
 125 velocity-weakening region in the earthquake-cycle models. Above it, the fault can host
 126 unstable slip; below it, velocity-strengthening friction promotes stable creep (Fig. 2d)
 127 [33, 34]. The colder end-member therefore contains a substantially wider rupture-
 128 capable fault surface. This geometric effect increases potential rupture area, while the
 129 greater depth-integrated strength also changes the stress available to drive coseismic
 130 slip.

131 **Loading rate, fault strength and coseismic slip**

132 We first isolated the response to loading rate in two-dimensional earthquake-cycle
 133 simulations governed by the aging form of rate-and-state friction [33, 35, 36]. In this
 134 formulation, the state variable increases during the interseismic period as fault contacts
 135 age and strengthen, consistent with laboratory slide–hold–slide experiments [36]. We
 136 imposed the depth-dependent stress and frictional structures of the cold- and hot-
 137 geotherm models and varied the tectonic loading rate V_p from 10^{-2} to 60 mm yr^{-1}
 138 (Methods) [32]. Comparing the Mongolia and San Andreas loading rates, 0.3 and
 139 30 mm yr^{-1} , respectively, shows that slower loading produces a much longer healing
 140 interval and a larger maximum state before instability (Fig. 3a). When rupture occurs,
 141 the slowly loaded model releases a larger stress drop and accumulates greater coseismic
 142 slip.

143 The time series show that slower loading does more than stretch the earthquake
 144 cycle in time. The state variable reaches a higher pre-event value in the slow model,
 145 and the coseismic step in cumulative slip is correspondingly larger (Fig. 3a). The
 146 loading interval therefore changes the mechanical initial condition of each earthquake,
 147 providing the constitutive link between geological loading rate and dynamic event size.

148 This behavior persists across the full loading-rate range and in both thermal struc-
 149 tures (Fig. 3b). As V_p decreases, stress drop rises from approximately 0.5–1 to 4–5
 150 MPa, while slip per event increases from approximately 0.5–1 to 6–8 m. The maxi-
 151 mum state variable also increases systematically as recurrence intervals lengthen. At
 152 any fixed loading rate, the cold-geotherm model produces larger stress drops and slip
 153 than the hot-geotherm model because its velocity-weakening region extends deeper
 154 and samples greater effective normal stress. The simulations therefore separate two
 155 effects within the same framework. Loading rate governs the duration of interseismic
 156 healing, whereas thermal structure governs the depth and strength of the seismogenic
 157 region. Their common direction explains why cold, slowly loaded continental faults can
 158 release substantially more moment than warm, rapidly loaded plate-boundary faults
 159 of the same along-strike length.

The Mongolia–San Andreas comparison is an end-member test. Natural faults occupy a continuum between these settings, and the predicted magnitude shift should vary continuously with loading rate, seismogenic width, and effective stress. Consistent with this expectation, intermediate-rate faults in Fig. 1c generally fall between the two limiting trends, allowing the framework to be applied beyond a binary intraplate–interplate classification.

Rupture area

Rupture area cannot be represented in two dimensions; we therefore next performed three-dimensional earthquake-cycle simulations with PyQuake3D [32, 37] (Fig. 4a). We modeled the same hot/fast and cold/slow end-members while varying fault length and retaining the prescribed frictional and stress structure within each end-member. In both cases, ruptures nucleate inside the velocity-weakening region and propagate laterally along strike (Fig. 4b). In the cold model, rupture also extends farther down dip because the unstable region reaches the deeper brittle–ductile transition (Fig. 4c). Following the earthquake, velocity-strengthening regions surrounding the ruptured patch creep and reload the fault. The simulations thus resolve the complete geometric consequence of a thicker seismogenic layer rather than representing it through an assumed rupture width.

For a representative fault length, the cold/slow rupture accumulates approximately 16 m of coseismic slip, compared with approximately 13 m in the hot/fast model, and ruptures an area about 1.7 times larger (Fig. 5a,b). Through Eq. 1, these differences yield approximately twice the seismic moment for that event. Across the complete suite of modeled fault lengths, the separation between the two end-members increases with rupture length and approaches 0.5 magnitude units for the longest ruptures (Fig. 5c). The simulation results reproduce the relative positions of the slow- and fast-fault observations, while the empirical relation of [23] lies between the two modeled trends. This behavior is consistent with the unusually large slip observed on the Mongolia–Altai ruptures. The 1905 Bulnay earthquake, for example, produced approximately 6 m of average slip along a roughly 375-km rupture (table 1), comparable to or greater than the slip of substantially longer plate-boundary ruptures.

Analytical approximation

To connect loading rate, fault strength, and moment analytically, we derived a reduced rate-and-state model for the interseismic evolution (Methods). In the long-recurrence limit, the maximum state variable scales approximately as $\theta_s \sim V_p^{-1} \ln(V_p^{-1})$. Frictional strength depends logarithmically on state; the corresponding healing contribution to peak strength and stress drop therefore grows approximately with $\ln(V_p^{-1})$. A two-order-of-magnitude decrease in loading rate can therefore produce a several-megapascal increase in stress drop for representative rate-and-state parameters. This distinction is central. Loading rate strongly controls the time available for state evolution, while the constitutive law converts that increase into logarithmic frictional strengthening. The effect becomes most pronounced on ultra-slow faults because their recurrence intervals extend far beyond those sampled at active plate boundaries.

The healing-derived stress drop enters a surface-rupture moment relation of the form $M_0 \propto \frac{L^3}{\Lambda^{1.79}} \Delta\tau_{p-r}$ (Supplementary Text), where $\Lambda = L/W$ is the rupture length-to-width ratio, W is seismogenic width, and $\Delta\tau_{p-r}$ is the peak-to-residual stress drop.

The exponents reflect the geometry of elongated strike-slip ruptures, for which width saturates before length, rather than a circular crack. Using the loading rates and seismogenic widths of the two end-members, the theoretical curves reproduce the separation between the three-dimensional simulations and the observational trends (Fig. 5c). At short rupture lengths, the catalog and simulations approach circular-crack scaling [38, 39]; as rupture length grows relative to width, they transition to elongated-rupture behavior. This transition explains why the magnitude difference between the end-members becomes larger for long surface ruptures.

A sensitivity analysis evaluates the relative roles of healing and seismogenic width (Supplementary Text, fig. 3). For widths typical of continental strike-slip faults ($W \gtrsim 3$ km), the loading-rate contribution to stress drop exceeds the incremental width contribution by about an order of magnitude or more, reaching a factor of approximately 50 for the parameters adopted for Mongolia. The width effect diminishes once the seismogenic zone is substantially wider than the critical nucleation length, whereas frictional healing continues to increase as loading rate decreases. In the present models, enhanced healing is therefore the dominant source of the magnitude offset, and the deeper seismogenic layer amplifies it by increasing rupture area. This interpretation is consistent with global evidence for systematically larger stress drops in intraplate than interplate earthquakes [40–42], with the contrast identified by [19], and with recent evidence that long hold times can strengthen faults sufficiently to affect induced-earthquake behavior [43].

Discussion

The observational catalog, two-dimensional simulations, three-dimensional simulations, and analytical model address different parts of the same problem. The catalog identifies a systematic offset in magnitude at a given rupture length (Fig. 1c). The two-dimensional calculations show how loading rate changes healing, stress drop, and slip (Fig. 3). The three-dimensional calculations demonstrate how that frictional effect combines with seismogenic thickness to change both slip and rupture area (Fig. 4). The analytical model then links these quantities to the observed magnitude–length scaling (Fig. 5c). Their agreement supports a common physical interpretation. Long loading intervals allow slowly slipping faults to restrengthen between earthquakes, and the cold lithosphere typical of many continental interiors provides a larger fault area over which that stored stress can be released.

This agreement is most informative at the level of intermediate observables. The models predict the full sequence from longer recurrence to greater state evolution, larger stress drop and slip, and deeper instability with larger rupture area; each step can be tested against distinct observations, including paleoseismic recurrence, stress-drop compilations, geological slip, and seismogenic thickness. The magnitude–length offset is thus the integrated expression of a chain of physically testable processes.

The models necessarily simplify natural faults, using a single rate-and-state friction law constrained mainly by laboratory experiments [33, 36], one-dimensional geotherms rather than three-dimensional thermal structures, and idealized fault geometries. The

catalog is likewise limited to 50 earthquakes, with rupture-length and slip-rate estimates of variable quality (table 1). These limitations affect the precise size of the inferred offset, but not the direction of the modeled response. Longer interseismic intervals increase healing and stress drop, whereas a deeper seismogenic layer increases rupture area. The results therefore describe a mechanism for a systematic component of magnitude-length variability rather than a complete account of every source of scatter.

The broader implication is that earthquake magnitude reflects a fault’s loading history, not only the length that ruptures. Slowly deforming continental interiors can accumulate strain over intervals far longer than the instrumental and geodetic records used to characterize them. Their low deformation rates imply long recurrence intervals, but not necessarily a low maximum credible magnitude. A rupture-length relation calibrated mainly on active plate boundaries can therefore underestimate the upper range of earthquake size on ultra-slow faults. The Mongolia–Altai earthquakes represent an extreme expression of this behavior, but the mechanism should apply more broadly to cold intraplate strike-slip faults with long recurrence times [16, 17].

These results bear directly on how magnitude–rupture-length scaling is used in seismic hazard assessment. Wells and Coppersmith [23] developed their relations as a statistical bridge between geological fault length and earthquake magnitude precisely because the largest event a fault can generate rarely occurs within the historical record; the same logic underlies most fault-based hazard models today. In probabilistic hazard analysis, a magnitude–rupture-length or magnitude–area relation converts each magnitude in a fault’s magnitude–frequency distribution into a rupture of finite size, which is then floated along the fault trace to compute source-to-site distance. Because ground motion depends strongly on the shortest distance to the rupture plane rather than to the epicenter or fault center, the assumed rupture length directly controls which sites are treated as near-fault for a given event. A scaling relation calibrated on fast-loading plate-boundary earthquakes systematically underestimates the moment, slip, and rupture area associated with a given length on slowly loaded, cold intraplate faults, which in turn underestimates both the maximum magnitude assigned to such a source and the ground motion at sites close to the true rupture. The same bias acts in the opposite direction for magnitude-to-length conversions used in scenario and paleoseismic interpretation, since a mapped rupture length on an ultra-slow fault corresponds to a larger characteristic magnitude than the global median relation implies. Because rare, large earthquakes dominate the low-probability tail of a hazard curve, this bias is largest precisely where it matters most for critical infrastructure and long-return-period design. Our results provide a physical basis, rather than an arbitrary adjustment, for weighting branches of a scaling-relation logic tree by tectonic setting. Loading rate identifies, in advance, which faults are likely to fall above the global median curve, and by how much.

A single empirical magnitude–length curve assigns the same expected earthquake size to faults with fundamentally different loading histories. Our results show why that approximation fails at the slowest end of the tectonic spectrum. Tectonic loading rate governs the time available for fault healing, while thermal structure sets the seismogenic width over which stored elastic energy can be released. Incorporating these

292 boundary conditions into earthquake-cycle models constrains a range of physically
293 admissible magnitudes for a mapped rupture length rather than one universal value.
294 For the fastest and slowest faults considered here, that range differs by approximately
295 0.5 magnitude units, and it points in a specific direction, toward larger magnitude,
296 larger rupture dimensions, and closer near-fault distances than a plate-boundary-
297 calibrated relation would assign. Loading rate is already estimated for most major
298 faults from geodesy or paleoseismology; treating it as an explicit input to magnitude-
299 length scaling therefore offers an immediate, physically grounded refinement to seismic
300 hazard assessment.

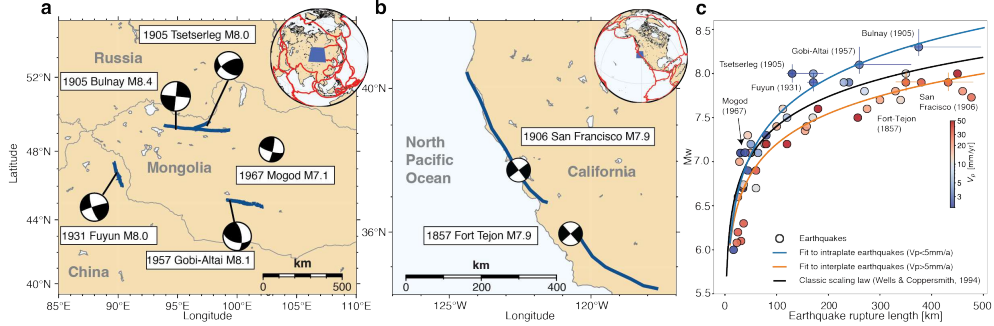


Fig. 1: Great strike-slip intraplate earthquakes in Mongolia and interplate earthquakes in San Andreas fault zone with their rupture-length–magnitude scaling relative to global strike-slip earthquakes. (a) Map of Mongolia and major earthquakes that occurred in the 20th century, with rupture lengths (blue) and focal mechanisms. **(b)** Map of the San Andreas fault zone and two major earthquakes, with rupture lengths (red) and focal mechanisms. Insets in panels A and B show the regional locations of Mongolia and the San Andreas fault zone relative to major plate boundaries (red). **(c)** Moment magnitude versus rupture length for the global strike-slip earthquake catalog (Table 1), colored by tectonic slip rate (V_p). Blue and orange curves are least-squares fits to events on faults with V_p below and above 5 mm yr^{-1} , respectively; the black curve is the empirical relation of [23]. The earthquakes labeled in panels a and b are marked by their reported minimum and maximum rupture lengths and magnitudes (error bars).

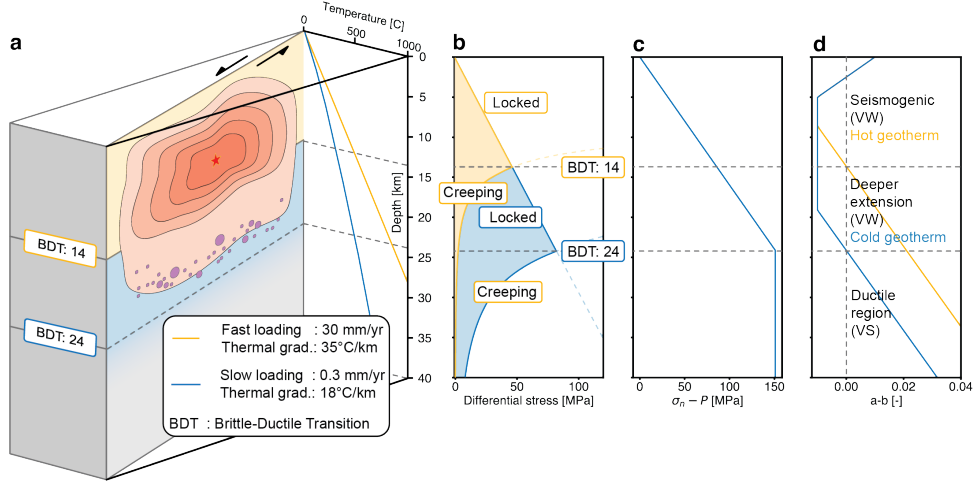


Fig. 2: Thermal and rheological framework linking tectonic loading rate to seismogenic thickness. (a) Schematic strike-slip fault block (left) showing the coseismic slip distribution of a representative rupture (contours) and hypocenter (red star) within the seismogenic zone, alongside temperature-depth profiles for the cold, slowly loaded Mongolian setting (blue; 18 °C/km) and the warmer, rapidly loaded San Andreas Fault setting (orange; 35 °C/km) [27, 31]. (b) Brittle (shallow) and ductile (deep) strength envelopes for both settings; the brittle-ductile transition (BDT), where the two intersect, is markedly deeper for the cold geotherm (24 km) than for the hot geotherm (14 km). (c) Corresponding effective normal stress with depth. (d) Depth-dependence of the rate-and-state friction parameter ($a - b$) inferred from the BDT, distinguishing the velocity-weakening (VW) seismogenic zone from velocity-strengthening (VS) creep at greater depth; the deeper BDT of the cold geotherm yields a thicker seismogenic zone. Panels (a-d) share a common depth axis (Methods).

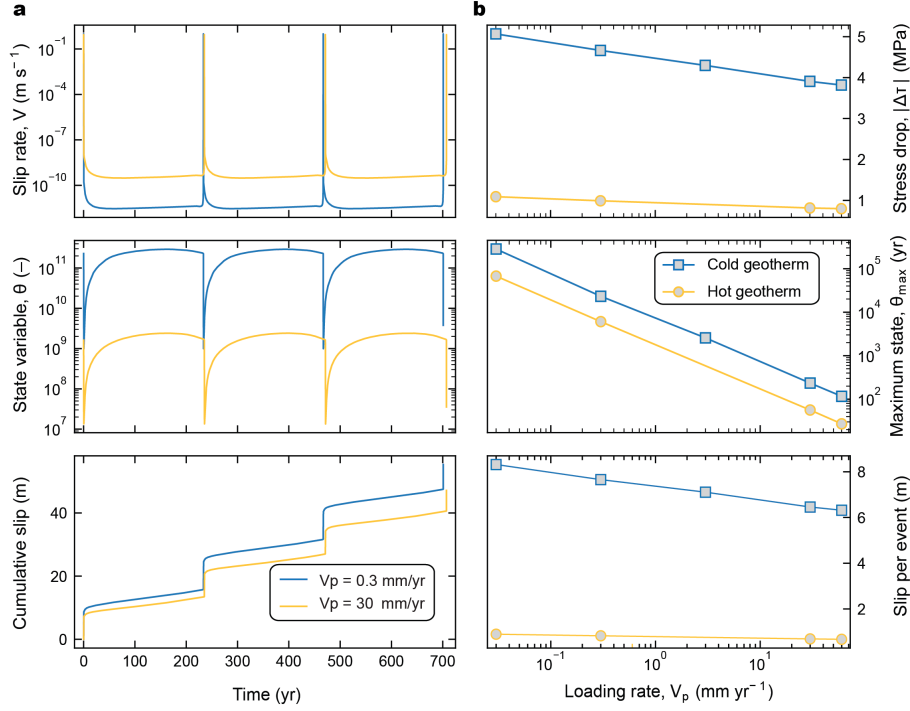


Fig. 3: Two-dimensional earthquake-cycle simulations show that slower loading enhances interseismic healing, stress drop, and coseismic slip. (a) Time series of slip rate, state variable (a proxy for interseismic healing), and cumulative slip for the cold-geotherm fault model loaded at $V_p = 0.3 \text{ mm/yr}$ (blue, Mongolia end-member) and $V_p = 30 \text{ mm/yr}$ (orange, San Andreas end-member). The time axis for the slowly loaded fault is compressed by the ratio of the two loading rates (10^2) to align earthquake cycles for comparison. **(b)** Stress drop, maximum state variable, and slip per event as a function of loading rate V_p , for the cold-geotherm (blue squares) and hot-geotherm (orange circles) fault models. All three quantities increase systematically as V_p decreases, approximately following a power law (Methods).

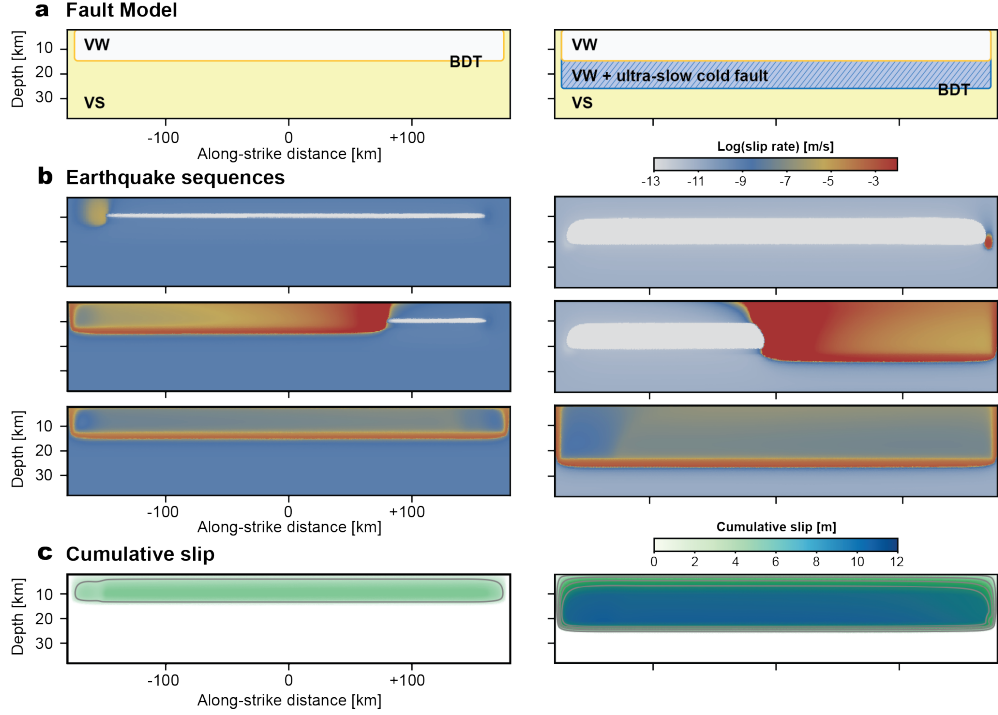


Fig. 4: Three-dimensional simulations show that a thicker seismogenic zone and enhanced healing combine to produce larger, deeper ruptures under slow tectonic loading. (a) Fault models for the hot-geotherm, fast-loading setting (left) and the cold-geotherm, ultra-slow-loading setting (right), showing the shallow velocity-weakening seismogenic zone (VW, white), the additional deep velocity-weakening zone associated with ultra-slow loading (blue hatching, right), the velocity-strengthening creeping zone (VS, yellow), and the BDT. (b) Snapshots of slip rate on the fault plane during an earthquake sequence – nucleation (top), coseismic rupture propagation (middle), and post-seismic afterslip (bottom) – for the hot/fast (left) and cold/ultra-slow (right) settings. The cold/ultra-slow rupture propagates over a larger, deeper fault area. (c) Cumulative coseismic slip for the same event in each setting, showing substantially larger slip in the cold/ultra-slow case. Color scales for slip rate and cumulative slip are shared across columns (Methods).

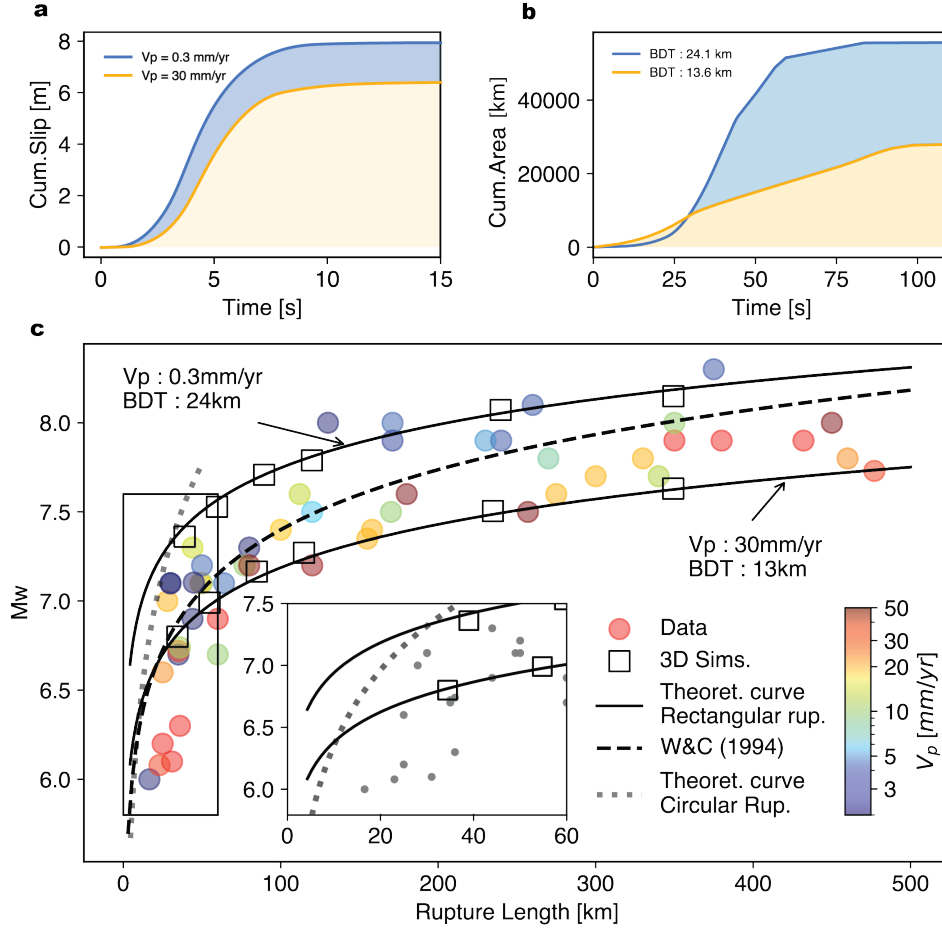


Fig. 5: Theoretical and numerical rupture length-moment magnitude scaling reproduces the observed intraplate-interplate offset. (a) Cumulative coseismic slip and (b) cumulative rupture area versus time during a single earthquake, from the 3D simulations of Fig. 4, for the cold-geotherm/ultra-slow (blue; BDT = 24.1 km) and hot-geotherm/fast (orange; BDT = 13.6 km) settings. (c) Moment magnitude versus rupture length. Filled circles are observational data from Fig. 1C, colored by tectonic slip rate; open squares are 3D simulation results. Solid black curves show the theoretical scaling relation (Methods) for the two end-member settings; the dashed curve is the empirical relation of [23]; the dotted curve is the circular-rupture scaling of [39], appropriate for small, near-circular ruptures before they transition to elongated-crack behavior [38], and it better describes the smallest events (inset).

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415 Methods

416 Strike-Slip Earthquake Catalog

417 We adopted the earthquake catalog of [22] as our primary dataset and extended
 418 it with 50 historical and recent strike-slip earthquakes, including the 2023 Türkiye
 419 earthquake doublet and the 2025 Myanmar earthquake, together with tectonic slip-rate
 420 estimates compiled from the literature for each event (table 1). Rupture length, mean
 421 coseismic slip, seismogenic depth, shear modulus, moment magnitude, and tectonic
 422 slip rate were compiled from the sources listed for each entry.

423 Yield-Strength Envelopes and Thermal Structure

424 To investigate how lithospheric thermal structure influences earthquake behavior,
 425 we constructed one-dimensional yield-strength envelopes (YSEs) for representative
 426 intraplate and plate-boundary environments. The yield-strength envelope combines
 427 brittle failure at shallow depths with ductile creep at greater depths, allowing estima-
 428 tion of the brittle-ductile transition (BDT), which is subsequently used to define the
 429 distribution of rate-and-state friction parameters in the numerical simulations.

430 Temperature was prescribed as a function of depth using representative geother-
 431 mal gradients for the two tectonic settings considered in this study. The cold-craton
 432 geotherm was based on estimates for the Mongolian lithosphere [27–29], whereas the
 433 plate-boundary geotherm was chosen to approximate conditions inferred for the San
 434 Andreas Fault system [30, 31]. A constant rock density of $\rho = 2800 \text{ kg m}^{-3}$ was
 435 assumed throughout the crust and uppermost mantle.

436 The brittle portion of the yield-strength envelope was computed using Anderson’s
 437 faulting theory for a vertical strike-slip fault [51]. The brittle differential stress for the
 438 strike-slip regime is given by

$$\Delta\sigma_b = \frac{2\mu_s(1-\lambda_f)\rho gz}{\sqrt{1+\mu_s^2}+\mu_s}, \quad (2)$$

439 where $g = 9.81 \text{ ms}^{-2}$, z is depth, λ_f is the pore-fluid pressure ratio, and μ_s is the
 440 static friction coefficient. This expression gives the depth-dependent brittle strength
 441 under the assumption of hydrostatic to moderately elevated pore-fluid pressure.

442 Ductile deformation was modeled using a power-law creep rheology,

$$\Delta\sigma_d = \left(\frac{\dot{\epsilon}}{A_p}\right)^{1/n} \exp\left(\frac{Q}{nRT}\right), \quad (3)$$

443 where $\Delta\sigma_d$ is the differential stress, $\dot{\epsilon} = V_p/H_d$ is the tectonic strain rate (the ratio
 444 of the tectonic slip rate V_p to the ductile shear-zone thickness H_d), A_p is a material
 445 constant, n is the stress exponent, Q is the activation energy, R is the universal gas
 446 constant, and T is temperature in kelvin. We adopted experimentally determined flow-
 447 law parameters for wet quartzite [52] and wet plagioclase [53] (table 1), because quartz-
 448 and feldspar-rich lithologies dominate deformation in much of the continental crust.

449 The BDT was defined as the depth at which the brittle and ductile strengths are
 450 equal, $\Delta\sigma_b(z) = \Delta\sigma_d(z)$. Above the BDT, deformation is dominated by frictional

Table 1: Rheological parameters used to construct the yield-strength envelopes. Flow-law parameters correspond to dislocation creep of wet quartz and wet plagioclase. The tectonic strain rate was computed as $\dot{\epsilon} = V_p/h_{\text{shear}}$, where h_{shear} is the thickness of the lower-crustal ductile shear zone.

Parameter	Symbol	Value	Units	Material	Reference
Pre-exponential factor	A	1.1×10^{-28}	$\text{Pa}^{-4} \text{s}^{-1}$	Wet quartz	[52]
Stress exponent	n	4.0	–	Wet quartz	[52]
Activation energy	Q	223×10^3	J mol^{-1}	Wet quartz	[52]
Pre-exponential factor	A	3.16×10^{-22}	$\text{Pa}^{-3} \text{s}^{-1}$	Wet plagioclase	[53]
Stress exponent	n	3.0	–	Wet plagioclase	[53]
Activation energy	Q	345×10^3	J mol^{-1}	Wet plagioclase	[53]
Rock density	ρ	2800	kg m^{-3}	All models	This study
Gas constant	R	8.314	$\text{J mol}^{-1} \text{K}^{-1}$	Constant	Standard
Strain rate	$\dot{\epsilon}$	V_p/h_{shear}	s^{-1}	All models	This study
Ductile shear-zone thickness	h_{shear}	5	km	Lower crust	[54]
Fluid pressure ratio	λ_{pf}	0.94	–	All Models	This Study

failure and earthquake generation; below the BDT, ductile creep dominates and long-term strain is accommodated aseismically. The colder cratonic geotherm therefore produces a deeper BDT and a thicker seismogenic layer than the hotter plate-boundary geotherm (Fig. 2).

The depth-dependent distribution of rate-and-state friction properties was prescribed using the computed yield-strength envelopes. Regions above the BDT were assigned velocity-weakening friction ($a - b < 0$), representing the locked seismogenic zone capable of generating earthquakes. Regions below the BDT were assigned velocity-strengthening friction ($a - b > 0$), representing stable creeping behavior. To avoid abrupt rheological transitions, $(a - b)$ was varied smoothly across the BDT over a finite depth interval. This procedure produces a self-consistent link between lithospheric thermal structure, fault rheology, and earthquake-cycle dynamics: colder lithosphere yields a deeper BDT, a thicker velocity-weakening zone, and larger potential rupture dimensions than hotter plate-boundary settings. Parameters for the cold and hot geotherms are listed in table 1.

Numerical Earthquake-Cycle Simulations

We modeled a planar strike-slip fault as an infinitesimally thin interface governed by aging-law rate-and-state friction [33], embedded in a purely elastic medium, and solved the quasi-dynamic governing equations with boundary-element methods in both two and three dimensions using BEM2D [56] and PyQuake3D [37], respectively. Depth-dependent friction parameters (a , b , $a - b$) were assigned from the yield-strength envelopes described above, with velocity-weakening friction confined above the BDT and velocity-strengthening friction below it. The 2D simulations impose the along-dip friction and stress profile for the cold- and hot-geotherm models and vary only the tectonic loading rate V_p , spanning from 10^{-2} to 60 mm/yr, to isolate the effect of loading rate on interseismic healing, stress drop, and slip (Fig. 3). The 3D simulations use the same two end-member geotherms (cold: BDT = 24.1 km; hot: BDT = 13.6 km) and vary fault length while holding the loading rate at each end-member's

representative value ($V_p = 0.3$ and 30 mm/yr, respectively) and all other parameters fixed, to additionally capture the effect of rupture area on seismic moment (Figs. 4 and 5). To prevent numerical artifacts, we used at least seven grid points per critical nucleation size ($L_b/dx \sim 7$), where (dx) is the grid spacing and $L_b = \frac{Gd_c}{b\sigma}$. To keep the constitutive parameters consistent, we defined the critical slip distance d_c from the seismogenic width of the fast-loading fault (San Andreas),

$$d_c = \frac{\pi}{W_c} \frac{b\sigma}{G} \left(\frac{a}{b} - 1 \right)^2,$$

where the ratio of seismogenic width to critical nucleation length was set to $W_c = W_{seis}/L_c = 10$, $\sigma = 100$ MPa, and $L_c \sim \frac{1}{\pi} \frac{b}{b-a} L_b$ [58]. A 0D spring-slider system was used to derive and verify the analytical relationships presented in the Supplementary Text (Figs. 1 and 2).

Rate-and-State Friction

To investigate how tectonic loading rate influences earthquake magnitude, we require a constitutive law that explicitly accounts for fault healing during the interseismic period. We therefore employ the rate-and-state friction (RSF) framework [33, 35], which is widely used to describe laboratory observations of frictional sliding and earthquake-cycle dynamics.

In the classical RSF formulation, the friction coefficient depends on both slip velocity (V) and a state variable (θ),

$$\mu(V, \theta) = \mu_0 + a \ln \left(\frac{V}{V_0} \right) + b \ln \left(\frac{\theta}{\theta_0} \right), \quad (4)$$

where (a) and (b) are empirical parameters describing the direct velocity effect and the state evolution effect, respectively. The parameter (μ_0) is the steady-state friction coefficient at the reference velocity (V_0), and ($\theta_0 = d_c/V_0$) is the corresponding reference state. The quantity (d_c) denotes the critical slip distance.

The state variable (θ) represents the evolving physical condition of the fault interface. Although its microscopic interpretation depends on the fault material and deformation mechanism, it is commonly associated with the age and maturity of asperity contacts and related microstructural evolution within the fault gouge. Larger values of (θ) correspond to a more strongly healed fault and therefore greater frictional resistance.

Several state-evolution laws have been proposed. In this study we adopt the aging law [33],

$$\frac{d\theta}{dt} = 1 - \Omega, \quad (5)$$

where $\Omega = \frac{V\theta}{d_c}$ measures the distance from steady state. When ($\Omega = 1$), the fault is at steady state and the state variable ceases to evolve ($\dot{\theta} = 0$). Under this special condition, Eq. (4) reduces to the familiar steady-state friction relation,

$$\mu_{ss}(V) = \mu_0 + (a - b) \ln \left(\frac{V}{V_0} \right). \quad (6)$$

The choice of the aging law is central to this study. Unlike alternative formulations such as the slip law, the aging law permits state evolution even when the fault is stationary ($V = 0$). Consequently, it reproduces the progressive increase in static friction observed in laboratory slide-hold-slide experiments [36], in which friction grows approximately logarithmically with hold time. Because our hypothesis is that slowly loaded faults experience substantially longer interseismic healing periods, a constitutive law capable of representing time-dependent frictional strengthening is essential.

The analytical derivations presented below show that the maximum attainable state variable increases approximately as the inverse of tectonic loading rate. Through Eq. (4), this enhanced healing increases fault strength prior to failure, providing the physical link between slow tectonic loading, larger stress drops, and ultimately larger earthquake magnitudes.

Theoretical Analysis of the Magnitude-Loading-Rate Dependence

The numerical simulations presented in the main text indicate that earthquake magnitude increases systematically as tectonic loading rate decreases. To understand the origin of this dependence, we derive an analytical approximation based on the aging form of rate-and-state friction. Our objective is to determine how the tectonic slip rate V_p influences the maximum state variable attained during the interseismic period and, consequently, the coseismic stress drop and seismic moment.

Interseismic healing regime ($\Omega \ll 1$)

During the interseismic phase, the state variable remains far below steady state ($\Omega = V\theta/d_c \ll 1$). Under this condition, the state evolution of the aging law simplifies to

$$\dot{\theta} \approx 1, \quad (7)$$

such that the state variable increases approximately linearly with time,

$$\theta(t) \approx \theta_i + t, \quad (8)$$

where the subscript i denotes the initial value hereafter and θ_i is the initial state variable at an instance of interseismic time ($\Omega \ll 1$).

The rate-and-state Eq. (4) can be rewritten as:

$$\mu = \mu_i + a \ln \left(\frac{V}{V_i} \right) + b \ln \left(\frac{t}{\theta_i} + 1 \right), \quad (9)$$

Taking the time derivative of this equation gives

$$\dot{\mu} = a \frac{\dot{V}}{V} + \frac{b}{t + \theta_i}. \quad (10)$$

Under quasi-static conditions and $V \ll V_p$, changes in shear stress through tectonic loading ($\tau_l \sim KV_p t$) are balanced by changes in frictional resistance, where K denotes elastic stiffness. Assuming constant effective normal stress and taking the time derivative,

$$\dot{\tau}_l \approx \sigma \dot{\mu} \quad (11)$$

Combining this equation with Eq. (10), and introducing the characteristic nucleation time [61]

$$t_r = a\sigma/\dot{\tau}_l, \quad (12)$$

we get

$$\frac{1}{t_r} \approx \frac{\dot{V}}{V} + \frac{b/a}{t + \theta_i}. \quad (13)$$

Using the definition of Ω , we get

$$\frac{\dot{\Omega}(t)}{\Omega(t)} = \frac{\dot{\theta}}{\theta} + \frac{\dot{V}}{V}. \quad (14)$$

We define the velocity-strengthening parameter

$$\epsilon = 1 - b/a \quad (15)$$

which measures the distance from velocity-neutral friction ($a = b$). Combining Eq. (14) with Eqs. (7), (8) and (13) gives

$$\frac{\dot{\Omega}(t)}{\Omega(t)} = \frac{1}{t_r} + \frac{\epsilon}{t + \theta_i}. \quad (16)$$

By integrating the equation over the interval $t \in [0, t_s]$, where t_s denotes the time instant at which the fault crosses the steady state, and imposing the steady-state condition $\Omega(t_s) = 1$ together with the initial condition $\Omega(0) = \Omega_i$, yields

$$\int_{\Omega_i}^1 \frac{d\Omega}{\Omega} = \int_0^{t_s} \left(\frac{1}{t_r} + \frac{\epsilon}{t + \theta_i} \right) dt. \quad (17)$$

Its solution is

$$-\ln(\Omega_i) = \frac{t_s}{t_r} + \epsilon \ln \left(\frac{\theta_i + t_s}{\theta_i} \right). \quad (18)$$

Inserting $\theta_s = \theta_i + t_s$ and rearranging for the steady-state crossing state variable θ_s ,

$$\frac{\theta_s}{t_r \epsilon} e^{\frac{\theta_s}{t_r \epsilon}} = \Omega_i^{-\frac{1}{\epsilon}} \frac{\theta_i}{t_r \epsilon} e^{\frac{\theta_i}{t_r \epsilon}} \quad (19)$$

Eq. (19) is transcendental and can be recast into a form involving the Lambert-W function. Introducing $x = \frac{\theta_s}{t_r \epsilon}$, and $y = \frac{\theta_i}{t_r \epsilon}$

$$xe^x = \frac{ye^y}{\Omega_i^{1/\epsilon}} \quad (20)$$

And from the definition,

$$x = W \left(\frac{ye^y}{\Omega_i^{1/\epsilon}} \right) \quad (21)$$

542 where W is the Lambert-W function. For any real argument (z) , the Lambert-W
 543 function has two real branches. The principal branch satisfies $(-1 < W_0(z) < 0)$,
 544 which can correspond to the dynamic slip velocity when Ω is constant, just before the
 545 relaxation phase starts [58]. The lower branch ($W_{-1}(z) \leq -1$) requires $(\theta_s > \theta_i)$ and
 546 is appropriate for the large-healing solution.

$$\frac{\theta_s}{t_r} = \epsilon W_{-1} \left(\frac{\theta_i \Omega_i^{-1/\epsilon}}{t_r \epsilon} e^{\frac{\theta_i}{t_r \epsilon}} \right). \quad (22)$$

547 The characteristic relaxation time (t_r) is a large number that depends on the
 548 tectonic slip rate (V_p) . The exponential term inside the Lambert-W function can be
 549 expanded as $e^y = 1 + y + O(y^2)$, since $t_r \rightarrow \infty$, we have $y \rightarrow 0^-$. Substituting this
 550 approximation into Eq. (22),

$$z = \frac{\theta_i \Omega_i^{-1/\epsilon}}{\epsilon t_r} + O(t_r^{-2}) \quad (23)$$

551 since $y = O(t_r^{-1})$.

552 The standard asymptotic expansion for $z \rightarrow 0^-$ follows the form

$$W_{-1}(z) = L_1 - L_2 + o(1), \quad (24)$$

553 where $L_1 = \ln(-z)$ and $L_2 = \ln(-\ln(-z))$. The leading orders are

$$L_1 = \ln \left(-\frac{\theta_i}{\epsilon} \right) - \ln(t_r) - \frac{\ln(\Omega_i)}{\epsilon}, \quad (25)$$

554 and

$$L_2 = \ln(t_r \ln(t_r)) + o(1), \quad (26)$$

555 since t_r is a large number. Hence,

$$W_{-1}(z) \approx -\ln(t_r) - \ln(\ln(t_r)) + \ln \left(-\frac{\theta_i}{\epsilon} \right) - \frac{\ln(\Omega_i)}{\epsilon}. \quad (27)$$

556 Replacing the physical variables, the steady-state crossing yields

$$\theta_s \approx -t_r \epsilon \left[\ln(t_r) + \ln(\ln(t_r)) + \frac{\ln(\Omega_i)}{\epsilon} - \ln \left(-\frac{\theta_i}{\epsilon} \right) \right]. \quad (28)$$

557 The above asymptotic approximation shows

$$\theta_s \sim -\epsilon t_r \ln(t_r) = \frac{(b-a)\sigma}{KV_p} \ln \left(\frac{a\sigma}{KV_p} \right). \quad (29)$$

558 The dominant dependence is approximately inverse in tectonic loading rate, with a
 559 weak logarithmic correction arising from the Lambert-W asymptotics,

$$\theta_s \sim V_p^{-1} \ln(V_p^{-1}). \quad (30)$$

560 Eq. (28) therefore predicts that slowly loaded faults can accumulate substantially
 561 larger state variables before reaching the steady-state crossing condition. Physically,
 562 this implies greater frictional healing, higher fault strength prior to failure, and
 563 consequently larger stress drops during subsequent earthquakes. The slip rate at
 564 steady-state crossing is

$$V_s = \frac{d_c}{\theta_s},$$

565 which serves as the natural velocity scale for the subsequent acceleration phase.

566 To test the analytical prediction, we compare Eq. (28) against numerical sim-
 567 ulations of a spring-slider system (Fig. 1). Two simulations differ only in tectonic
 568 loading rate and represent idealized plate-boundary ($V_p = 10^{-9} \text{ m s}^{-1}$) and intraplate
 569 ($V_p = 10^{-11} \text{ m s}^{-1}$) conditions, respectively. Both use identical frictional ($b-a = 0.005$)
 570 and rheological properties. The slowly loaded system develops substantially larger
 571 values of (θ_s), consistent with the inverse dependence predicted by Eq. (29).

572 To evaluate the influence of frictional properties, we additionally performed a simu-
 573 lation with twice as strong velocity-weakening friction ($b-a = 0.01$), within the range
 574 of typical laboratory estimates, while keeping the plate rate fixed at ($V_p = 10^{-9} \text{ m s}^{-1}$).
 575 Increasing ($b-a$) modifies the slope of Ω versus θ and changes the trajectory. How-
 576 ever, its effect on the maximum attainable healing remains considerably smaller than
 577 that produced by a two-order-of-magnitude reduction in tectonic loading rate. These
 578 results indicate that tectonic loading rate is the primary control on the maximum
 579 potential healing time, whereas variations in frictional parameters provide a secondary
 580 modulation.

581 **Acceleration regime ($\Omega \gg 1$)**

582 Later in the earthquake cycle, the state evolves with slip once $\Omega \gg 1$ (Eq. 5). A
 583 suitable approximation was developed by [62]. We begin applying this approximation
 584 when the fault is somewhat above steady state, with a starting value $\Omega^+ \gtrsim 1$ chosen
 585 empirically. Because the steady-state crossing corresponds to $\dot{\theta} = 0$, the value of θ
 586 remains close to its steady-state crossing value θ_s as Ω grows above 1. We thus set the
 587 starting value of state as $\theta^+ = \theta_s$, and the corresponding starting value of slip velocity
 588 as $V^+ = \Omega^+ d_c / \theta_s$.

589 We set the time t when $\Omega = \Omega^+$, and normalize it with the characteristic nucleation
 590 time $T = t/t_r$ [62]. We define a non-dimensional parameter $\alpha = \frac{V_p \theta_s}{d_c} \frac{L_B}{W - L_B}$, combining
 591 the ratio of the potential slip deficit at the steady-state crossing to the critical slip
 592 distance ($V_p \theta_s / d_c$), and the ratio of the critical nucleation length to the difference
 593 between fault width and critical nucleation length ($L_B / (W - L_B)$), where $L_B =$
 594 $G d_c / (b \sigma)$.

595 Following [62], slip velocity, state, and the distance from steady state evolve as

$$\frac{V(T)}{d_c / \theta_s} = \frac{\Omega^+ e^T}{\frac{\Omega^+}{\alpha} (1 - e^T) + 1}, \quad (31)$$

$$\frac{\theta(T)}{\theta_s} = \left(\frac{\Omega^+}{\alpha} (1 - e^T) + 1 \right)^{\frac{W}{W-L_B} \frac{a}{b}}, \quad (32)$$

596 and

$$\Omega(T) = \Omega^+ e^T \left(\frac{\Omega^+}{\alpha} (1 - e^T) + 1 \right)^{\frac{W}{W-L_B} \frac{a}{b} - 1}. \quad (33)$$

597 **Prediction of Peak-to-Residual Stress Drop**

598 We next estimate the peak-to-residual stress drop analytically. Given the initial
599 conditions – the healing time since the last earthquake, θ_i , and the distance from
600 steady state, $\Omega_i \sim 10^{-4} - 10^{-8}$ – the peak-to-residual stress drop is

$$\Delta\tau_{p-r} = \Delta\tau_h + \Delta\tau_n, \quad (34)$$

601 where $\Delta\tau_h$ and $\Delta\tau_n$ denote the shear-stress contributions accumulated during the heal-
602 ing phase (interseismic, $\Omega < 1$) and during earthquake nucleation ($\Omega > 1$), respectively.
603 These two terms can be rewritten as

$$\Delta\tau_h = \sigma_n (a - b) \ln \left(\frac{W_0(z)}{W_{-1}(z)} \right) \quad (35)$$

604 and

$$\Delta\tau_n = a \sigma_n \ln \left(\frac{W_{-1}(z)}{W_0(z)} \right) + b \sigma_n \ln \left(\frac{\theta_{\tau_{peak}}}{\theta_s} \right), \quad (36)$$

605 where the subscript τ_{peak} denotes the peak stress prior to the earthquake, approxi-
606 mated at the point where the fault slip rate reaches the dynamic range. Previous work
607 identified two regimes of dynamic slip rate using a radiation-damping term: no healing
608 ($\Omega \gg 1$) and constant weakening (Ω constant) [58]. The principal branch of the Lam-
609 bert function W_0 corresponds to the latter (i.e., $\Omega = 1$), while the lower branch W_{-1}
610 is associated with the steady-state crossing in Eq. (21), i.e., the maximum potential
611 healing. The argument $z = \theta_i \Omega_i^{-1/\epsilon} / (\epsilon t_r)$ was introduced in Eq. (22). Equation (34)
612 then simplifies to

$$\Delta\tau_{p-r} = b \sigma \ln \left(\frac{\theta_{\tau_{peak}}}{W_0(z) \epsilon t_r} \right), \quad (37)$$

613 where $\theta_{\tau_{peak}}$ denotes the state variable at peak stress, corresponding to the onset of
614 dynamic rupture. Evaluating Eq. (31) at the dynamic slip rate from the principal
615 branch $W_0(z)$ of the Lambert-W function yields the non-dimensional time to reach
616 peak stress:

$$e^{T_{\tau_{peak}}} = \frac{1 + \frac{\Omega^+}{\alpha}}{\frac{d_c}{\theta_s V_{\tau_{peak}}} + \frac{\Omega^+}{\alpha}} \quad (38)$$

617 where $V_{\tau_{peak}}$ is the slip rate at peak stress, approximated following the asymptotic
618 solution of [58] as $V_{\tau_{peak}} \approx 2(a\sigma/G)c_s \sim 1 \text{ cm s}^{-1}$, with c_s the shear-wave speed.
619 Substituting Eq. (38) into Eq. (32) gives

$$\theta_{\tau_{peak}} = \theta_s \left(1 + \frac{1 - e^{T_{\tau_{peak}}}}{\alpha} \right)^{\frac{a}{b} \frac{W}{W-L_B}}. \quad (39)$$

620 Finally, the peak-to-residual stress drop can be written as

$$\Delta\tau_{p-r} = b\sigma \ln(\theta_s) + \frac{a\sigma W}{W - L_B} \ln\left(1 + \frac{1 - e^{T_{peak}}}{\alpha}\right) - b\sigma \ln(W_0(z)\epsilon t_r). \quad (40)$$

621 The first term, representing the logarithmic healing time, strongly controls the peak-
 622 to-residual stress drop. The second term contributes to the reduction in shear stress
 623 during earthquake nucleation and the dynamic phase, while the third term has only
 624 a minor effect. The argument inside the third term corresponds to the state variable
 625 at the end of the dynamic rupture, which can be approximated as d_c/V_{dyn} , where
 626 $V_{dyn} \sim 1$ cm/s denotes the onset of relaxation. Eq. (40) then simplifies, for practical
 627 use, to

$$\Delta\tau_{p-r} = b\sigma \ln\left(\frac{\theta_s V_{dyn}}{d_c}\right) + \frac{a\sigma W}{W - L_B} \ln\left(1 + \frac{1 - e^{T_{peak}}}{\alpha}\right). \quad (41)$$

628 What controls stress drop: healing or seismogenic width?

629 Equation (41) shows that the peak-to-residual stress drop decomposes into two
 630 physically distinct contributions,

$$\Delta\tau_{p-r} = \Delta\tau_\theta + \Delta\tau_W \quad (42)$$

631 where $\Delta\tau_\theta = b\sigma \ln\left(\frac{\theta_s V_{dyn}}{d_c}\right)$ is the healing contribution, and $\Delta\tau_W = a\sigma\beta\frac{W}{W-L_B}$
 632 is the geometric contribution associated with the seismogenic width W , where $\beta =$
 633 $\ln\left(1 + \frac{1 - e^{T_{peak}}}{\alpha}\right)$. To determine which mechanism exerts the stronger control on
 634 stress drop, we evaluate the sensitivities of Eq. (41) to variations in healing time and
 635 seismogenic width. The logarithmic sensitivity to healing is

$$S_\theta = \frac{\partial\Delta\tau_{p-r}}{\partial\ln(\theta_s)} = b\sigma \quad (43)$$

636 whereas the sensitivity to seismogenic width is

$$S_W = \frac{\partial\Delta\tau_{p-r}}{\partial W} = -a\sigma\beta\frac{L_B}{(W - L_B)^2}. \quad (44)$$

637 The healing sensitivity is therefore constant and depends only on the state-
 638 evolution parameter b and the effective normal stress σ . In contrast, the width
 639 sensitivity decays as $(W - L_B)^{-2}$. Consequently, once the seismogenic width substan-
 640 tially exceeds the critical nucleation length ($W \gg L_B$), the influence of additional
 641 fault width rapidly diminishes.

642 Figure 3 presents the ratio S_θ/S_W for representative parameter values. For realistic
 643 values of β in the range -10 to -30 , the healing term becomes progressively more
 644 dominant with increasing seismogenic width. The transition typically occurs for $W \gtrsim$
 645 3 km, beyond which the contribution of fault healing surpasses that of seismogenic
 646 width by several factors.

For parameter values characteristic of Mongolia, with $W \sim 20\text{--}25$ km and $L_B \sim 0.5$ km, the ratio attains values of $S_\theta/S_W \sim 50$, indicating that interseismic healing exerts substantially stronger control on stress drop than variations in seismogenic width. This finding implies that the anomalously large earthquakes observed in slowly deforming continental interiors are primarily attributable to prolonged fault healing associated with low tectonic loading rates, whereas the deeper brittle-ductile transition and the resulting greater seismogenic thickness play a secondary, though non-negligible, role.

Interpolation from 0D to 3D: the Rectangular Crack

For simple geometries, a non-dimensional shape factor can be derived analytically; for rectangular cracks, however, a numerical solution is required. Numerical solutions for the shape factor relating stress drop to average fault offset were provided by [63]. We used their results for surface-breaking cracks ($d/W \rightarrow 0$), where d and W denote burial depth from the surface and the width of the fault, respectively. We fitted a power law for a given crack length-to-width ratio $\Lambda = L/W$,

$$C = \beta_1 \Lambda^{\beta_2}, \quad (45)$$

and estimated $\beta_1 = 2.20$ and $\beta_2 = -0.21$. The relationship between mean fault offset (slip) and stress drop is then given by

$$D = \frac{\Delta\tau_{p-r}}{C} W, \quad (46)$$

and the seismic moment is obtained as

$$M_0 = \frac{L^3}{\Lambda^{1.79}} \frac{\Delta\tau_{p-r}}{2.2}. \quad (47)$$

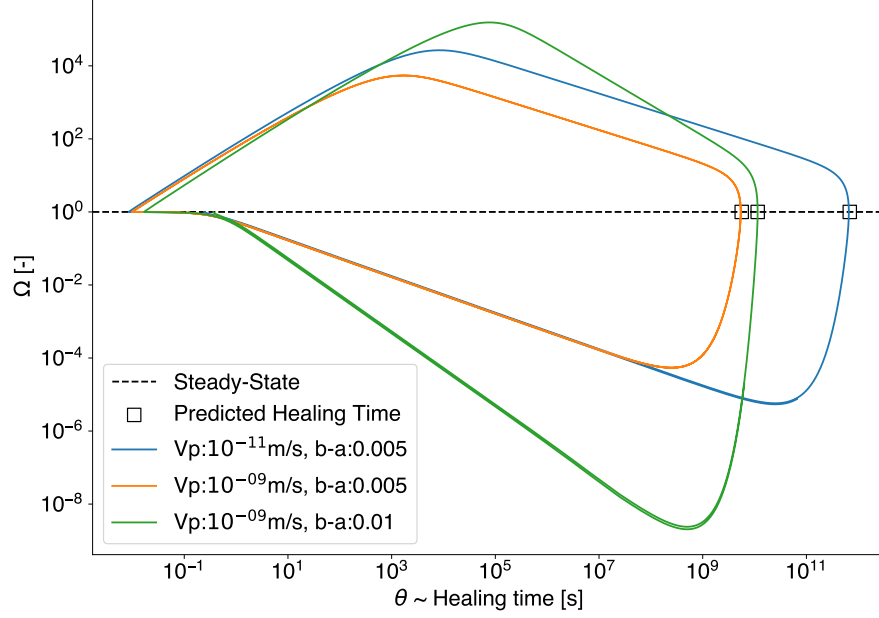
Here, the peak-to-residual stress drop is computed from Eq. (41).

Data Availability

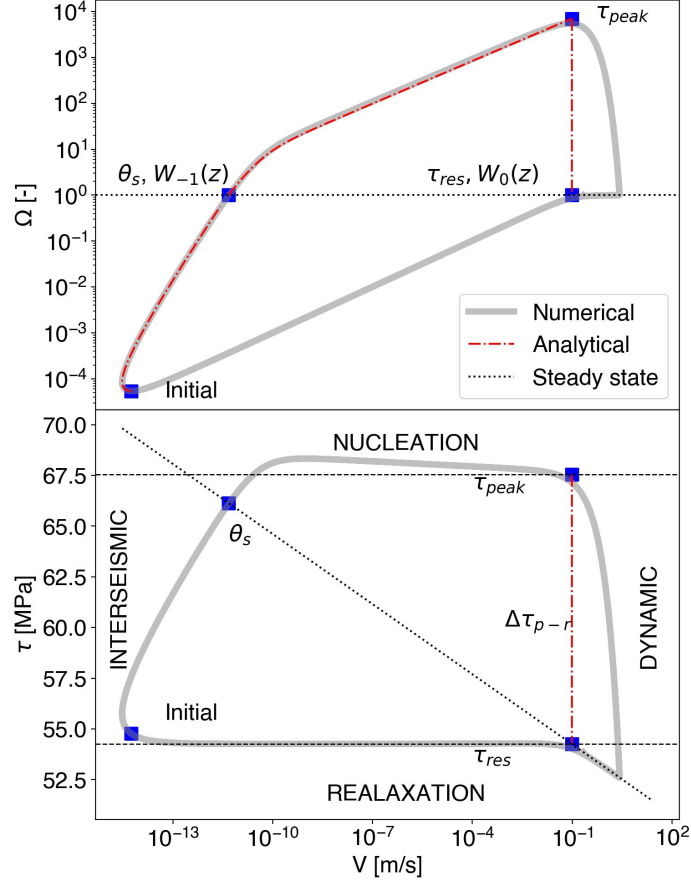
The data supporting the findings of this study are provided in Supplementary Table 1. The numerical simulation results are publicly available in the repository cited in Ref. [44].

Code availability

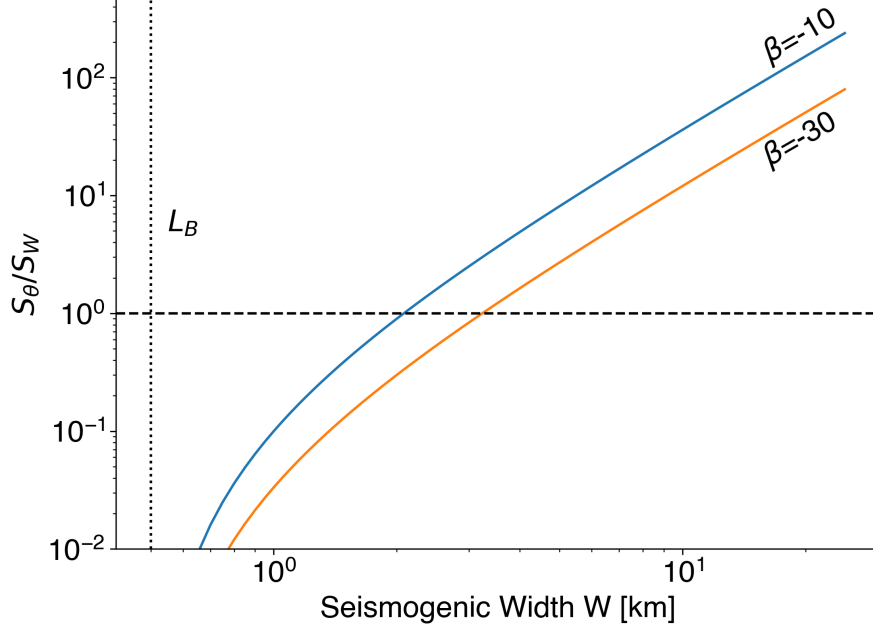
Earthquake-cycle simulations were performed using the open-source boundary-element code PyQuake3D [37]. The source code is publicly available at <https://github.com/Computational-Geophysics/PyQuake3D>. Comprehensive documentation, including installation instructions, user guides, benchmark examples, and tutorials, is available at <https://pyquake3d.readthedocs.io/en/latest/>. Maps were generated using the Generic Mapping Tools (GMT).



Extended Data Fig. 1: Dependence of the maximum healing time θ_s on tectonic loading rate V_p and frictional parameter $b - a$. Numerical solutions of the spring-slider system showing the evolution of Ω as a function of θ . Blue and orange curves correspond to identical frictional properties ($b - a = 0.005$) but different tectonic loading rates representative of intraplate ($V_p = 10^{-11} \text{ m s}^{-1}$) and plate-boundary ($V_p = 10^{-9} \text{ m s}^{-1}$) settings, respectively. The initial conditions are $\theta_i = d_c/V_{dyn}$ and $\Omega_i = 1$. The dynamic slip rate is $V_{dyn} = 2(a\sigma/G)c_s$, with $G = 32 \text{ GPa}$, $c_s = 3464 \text{ m s}^{-1}$, $a = 0.01$, and effective normal stress $\sigma = 100 \text{ MPa}$. The fault size is set to $W = 16 \text{ km}$, and the critical slip distance is computed using the theoretical nucleation size, with $W/L_{nuc} = 10$, $L_{nuc} = \frac{1}{\pi} \left(\frac{b}{b-a} \right)^2 L_b$, and $L_b = b\sigma/d_c$. Equation 28 is used to predict the maximum potential healing time θ_s from the initial conditions at the minimum value of Ω observed in the numerical simulations.



Extended Data Fig. 2: Theoretical estimates of the peak-to-residual stress drop using a simple spring-slider system. The distance from steady state, Ω , and shear stress, τ [MPa], are plotted against slip rate, V [m/s], from top to bottom. The numerical and analytical curves are shown as thick gray solid lines and thinner red dash-dotted lines, respectively, while steady states are indicated by dotted lines in both subpanels. The turning points corresponding to maximum healing (θ_s), peak stress before the fault slip enters the dynamic range (τ_{peak}), residual stress (τ_{res}), and the initial values for the analytical curves during the interseismic phase are marked with blue square markers. The analytical peak-to-residual stress drop using Eq. 40 ($\Delta\tau_{p-r}$) is shown in the lower panel.



Extended Data Fig. 3: Sensitivity of stress drop to fault healing and seismogenic width. Ratio between the healing sensitivity S_θ (Eq. 43) and the width sensitivity S_W (Eq. 44) as a function of seismogenic width W . Calculations use representative rate-and-state friction parameters $a = 0.005$, $b = 0.01$, effective normal stress $\sigma = 100$ MPa, and critical nucleation length $L_B = 5$ km. Colored curves correspond to values of β ranging from -10 to -30 . The sensitivity ratio increases rapidly with fault width, indicating that the influence of seismogenic width diminishes once $W \gg L_B$. For lithospheric conditions representative of Mongolia (BDT ~ 20 – 25 km), fault healing is approximately two orders of magnitude more influential than seismogenic width in controlling the peak-to-residual stress drop.

Extended Data Table 1: Historical strike-slip earthquakes. Length and Depth denote rupture length and seismogenic depth, respectively, and are given in km. Slip denotes the mean coseismic slip in m. The shear modulus (G) is expressed in units of $10 \times \text{GPa}$, and the tectonic slip rate V_p is given in mm/yr.

Date	Location	Length	Slip	Depth	G	M_w	V_p	Reference
10/28/1891	Neo-Dani (Japan)	80	3.1	15	3	7.3	2	[22, 64]
11/02/1930	Kita-Izu (Japan)	35	1.1	12	3.3	6.7	2	[22, 65]
12/25/1939	Erzincan (Turkey)	300	4.2	13	3.2	7.7	20	[22, 66]
05/19/1940	Imperial (CA)	60	1.6	13	2.5	6.9	34	[7, 22]
12/20/1942	Erbaa-Niksar (Turkey)	28	1.66	13	3.2	7.0	20	[7, 22, 66]
11/26/1943	Tosya (Turkey)	275	2.5	13	3.2	7.6	20	[22, 66]
02/01/1944	Gerede-Bolu (Turkey)	155	2.1	13	3.2	7.35	20	[22, 66]
07/22/1967	Mudurnu (Turkey)	60	0.9	12	2.4	6.7	10	[22, 66]
04/08/1968	Borrego Mtn (CA)	31	0.13	12	3.3	6.1	34	[7, 22]
10/15/1979	Imperial Valley (CA)	36	0.28	13	2.5	6.3	34	[7, 22]
07/29/1981	Sirch (Iran)	64	1.2	15	3.3	7.1	4.2	[22, 68]
11/23/1987	Super. Hills (CA)	25	0.3	12	3.2	6.2	34	[7, 22]
07/16/1990	Luzon (Philippines)	112	3.5	20	3.5	7.6	13	[22, 69]
06/28/1992	Landers (CA)	77	2.3	15	3	7.2	10	[22, 70]
10/16/1999	Hector Mine (CA)	44	1.56	12	3.0	6.9	1	[22, 71]
11/14/2001	Kunlun (China)	350	3.3	20	3	8.0	10	[22, 72]
01/07/1937	Kunlun (Tibet)	170	3.7	15	3	7.5	10	[7, 73]
01/15/2000	Yao'an (China)	16.6	0.63	11	3	6.0	1.8	[74]
01/09/1857	Fort Tejon (CA)	380	7	12	3	7.9	34	[7, 73]
04/18/1906	San Francisco (CA)	432	4.32	12	3	7.9	34	[7, 73, 75]
11/03/2002	Denali, Alaska (USA)	340	3.6	14	3	7.7	12	[73]
09/04/2010	Darfield-Greendale (NZ)	30	5.3	9	3	7.1	2	[76]
04/04/2010	El Mayor (Mexico)	120	1	15	3	7.2	46	[76]
04/16/2010	Kumamoto (Japan)	30	1.3	15	3	7.1	1	[77]
07/04/2019	Ridgecrest (CA)	50	1.5	15	3.3	7.1	10	[78]
02/06/2023	Pazarcik-Maras (Turkey)	270	5.6	13	3	7.8	8	[79]
02/06/2023	Cardak (Turkey)	120	3.5	18	3	7.5	6	[79]
12/26/1939	Erzincan (Türkiye)	330	3.6	12	3	7.8	20	[7, 73]
08/17/1999	Izmit (Turkey)	158	3	15	3	7.4	20	[80]
07/10/1894	Cinarcik (Turkey)	44	4	15	3	7.3	14	[80]
08/22/1949	Queen Charlotte (AK)	450	4.4	15	3	8.0	50	[73, 81]
07/30/1972	Sitka (AK)	180	4.9	15	3	7.6	50	[73, 81]
02/04/1976	Guatemala	257	3.1	13	3	7.5	49	[73]
05/22/1766	Central Marmara (Turkey)	49	3.25	15	3	7.1	18	[80]
08/09/1912	Ganos (Turkey)	100	2.5	15	3	7.4	19	[80, 82]
11/08/1980	Eureka (CA)	80	1.5	15	3	7.2	48	[73, 81]
05/25/1981	New Zealand (South Island)	25	0.8	12	3	6.6	23	[73, 83]
12/04/1957	Gobi (Mongolia)	260	5.5	15	3	8.1	3	[12]
05/01/1967	Bogod (Mongolia)	45	1.5	10	3	7.1	0.06	[4]
07/23/1905	Bulnay (Mongolia)	375	6	15	3	8.3	3	[3, 5, 87]
07/09/1905	Tsetserleg (Mongolia)	130	4	15	3	8.0	0.3	[6]
08/10/1931	Fuyun (China)	171	4	15	3	7.9	2.5	[90, 91]
11/22/1995	Gulf of Aqaba (Nuweiba)	50	1.4	15	3	7.2	4	[92]
10/17/1986	Loma Prieta	35.0	1.2	12.0	3.0	6.72	30	[93]
09/28/2004	Parkfield	23.0	0.27	8.8	3.0	6.08	30	[94]
01/24/2020	Elazığ (Turkey)	36	1	15	3	6.74	11	[95]
05/12/2008	Longmenshan (China)	240	4	23	3	7.9	3	[96–98]
08/10/1931	Koktokay-Ertai	171	6.3	15	3	8.0	4	[91]
04/24/1920	Haiyuan Fault, China	230	6	15	3	7.9	5	[99–102]
04/03/2025	Mandalay (Myanmar)	460	3.4	15	3	7.8	23	[103]

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844 **Author contributions**

845 E.S., Y.K. and L.D.Z. conceptualized the study and contributed to writing, review-
846 ing and editing the manuscript. E.S. and L.D.Z. performed data curation, formal
847 analysis and visualization, and wrote the original draft. L.D.Z. administered and super-
848 vised the project and provided resources. All authors reviewed and approved the final
849 manuscript.

850 **Competing interests**

851 There are no competing interests to declare.