

Title: Stopping phase reveals abrupt arrest of large strike-slip earthquakes

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Abstract: Earthquake magnitude is controlled by where and when rupture propagation stops. Yet rupture arrest has rarely been directly observed in near-field seismic records of natural earthquakes. Here we present systematic near-field observations of ground-motion stopping phases from large strike-slip earthquakes. Analysis of 12 global events shows that transient overshoot in fault-parallel ground surface displacement is a robust diagnostic signature of abrupt termination of rupture propagation. Dynamic rupture simulations reveal that near-field ground motions are strongly amplified by low wavespeed rocks at shallow depth, which enhance the amplitude of displacement overshoot recorded at the surface. The occurrence of stopping-phases at near-fault locations far from mapped rupture termini implies that large strike-slip earthquakes rupture in a segmented manner, with dynamic rupture propagation punctuated by abrupt arrest and reinitiation at internal fault-segment boundaries.

Main Text: Understanding how earthquake ruptures propagate and ultimately arrest is a major challenge in seismology. Paleoseismic and geological data show that strike-slip fault systems, such as the San Andreas and Alpine Faults, are composed of a series of discrete structural segments that may fail independently in moderate sized events, or link together dynamically during larger earthquake ruptures that propagate across hundreds of kilometers (1–4). Whether rupture remains confined to a single fault segment or grows into a multi-segment, large-magnitude event ultimately depends on whether rupture arrests before propagating across segment boundaries (5–8). Despite the fundamental role of rupture arrest in controlling earthquake size, its seismic signature—the stopping phase, a body wave carrying dynamic stresses of opposite polarity to the direction of fault slip—is rarely observed in large earthquakes, leaving the physical processes governing rupture termination largely hidden from direct observation.

Stopping phases were first identified in laboratory fracture experiments (9) and subsequently used to infer source properties of distant (10–12) and small earthquakes (13,14). With the advent of finite-fault inversion techniques (15–17), stopping phase observations have been used less frequently to constrain earthquake source properties. Nevertheless, stopping phases remain an inherent feature of dynamic rupture models and are thought to control the high-frequency decay of earthquake spectra (18–21). If generated during natural earthquakes, these phases should be most apparent in near-field ground motions, yet such observations are extremely rare (22–24), and the reasons for their scarcity remain unresolved.

Here, we report clear evidence of stopping phases in large strike-slip earthquakes. Using near-field observations from 12 surface-breaking events (table S1), we identified systematic stopping phase signatures in ground motions recorded near rupture terminations and reproduced these features using dynamic rupture simulations. These results provide direct seismic evidence that large earthquakes arrest abruptly rather than decelerate gradually. They further indicate that large strike-slip ruptures propagate in a segmented fashion, with cascades across multiple fault segments punctuated by transient arrest at internal segment boundaries.

Near-field observations of large strike-slip earthquakes

Near-field seismic recordings, within several kilometers of the source fault, provide unique constraints on slip dynamics during earthquake rupture. Such recordings are uncommon and often limited to a single instrument, yet they capture aspects of rupture behavior that cannot be resolved from far-field observations (25, 26). During the 2025 Mw 7.7 Mandalay (Myanmar) strike-slip earthquake, rupture propagated along a 500 km section of the Sagaing fault, and a single seismometer recorded near-field seismic waves radiated by the propagating rupture at a distance of 2.6 km from the fault (27). The recorded ground motion closely resembles other near-field records of large strike-slip earthquakes (Fig. 1; fig. S1); an initial phase of rapid slip-parallel motion ($> 1 \text{ m s}^{-1}$) was followed by a secondary phase of negative polarity (i.e., ground motion opposite to the fault slip direction), producing a ‘whiplash’ type of motion. Previous studies showed that the initial phase reflects dynamic slip on the nearby fault, commonly referred to as the “rupture phase” (Fig. 1) (28–30). The physical origin of the secondary phase, however, remains enigmatic. Owing to its reverse polarity, one possibility is that it reflects a transient episode of negative slip on the fault plane; however, to our knowledge, clear evidence of back-slip during dynamic rupture has not been reported for natural earthquakes. Another possibility is that it is generated by the wave effect, in which the static near-field contribution decays more rapidly with distance than the radiated far-field waves, although this mechanism has not been

demonstrated and appears unlikely given the close proximity of these observations to the fault. An alternative explanation, which we propose in this study, is that it represents a stopping phase produced during rupture arrest.

To test this hypothesis, we analyzed near-field seismic and geodetic data from 12 large strike-slip earthquakes (Fig. 1 and fig. S1). Five well-instrumented events, with multiple near-fault stations distributed along strike, form the primary basis for our analysis and span diverse tectonic settings, enabling us to distinguish ground motions generated within the central portions of the fault from those radiated during rupture arrest at the ends of the earthquake (Fig. 2). The limited number of well-instrumented events reflects the rarity of dense near-fault seismic networks rather than selective event choice.

The dataset from these events comprises 24 near-field records, including strong-motion accelerations and 1–10 Hz GNSS displacement waveforms, with a mean off-fault distance of 3 km (see Supplementary Materials). We computed fault-parallel displacements and quantified waveform shape using a metric of dynamic overshoot, η :

$$\eta = (u_d - u_s) / u_d$$

Where u_d denotes the peak dynamic fault-parallel displacement, and u_s is the final static displacement (e.g., Fig. 1). Independent geodetic measurements of coseismic deformation were used to validate our instrumentally measured displacement waveforms and to quantify the distribution of surface fault slip (Fig. 2 and figs. S2-S7).

Near-field displacement waveforms exhibit systematic differences between central portions of the rupture and regions near rupture termination. Waveforms near the epicenters and central parts of the fault rupture show step-like fault-parallel displacement envelopes with low values ($\eta \leq 0.15$) of dynamic overshoot (Fig. 2; stations IZT, H16, KEKS, EKZ1, and 143). These waveform shapes reflect the progressive accumulation of slip on the nearby fault plane and are consistent with previous kinematic (28, 31) and dynamic models (32, 33) of near-field particle displacements for strike-slip faulting. In contrast, waveforms recorded adjacent to the ends of the fault where the rupture propagation was terminated display large dynamic overshoot ($\eta = 0.2 - 0.5$), expressed as a prominent and transient peak in fault-parallel displacement (Fig. 2; stations ARC, 011, CMBL, 4612, and 141). Displacement overshoot shows no systematic relationship with off-fault distance, indicating that propagating seismic waves are not the primary cause of the negative displacement observed in the near field (fig. S8). The spatial localization of overshoot near rupture termini suggests a link between the observed waveform reversal and the arrest of rupture propagation.

The 2023 Mw 7.8 Pazarcık earthquake provides a particularly clear example of this behavior within a single rupture. The southwestern branch of the event was densely recorded by 11 near-field sensors distributed along a 100-km section of the Amanos segment of the East Anatolian fault (Fig. 2E). Rupture propagation along the Amanos segment was influenced by fault geometry, with two prominent bends in the fault trace (Fig. 2, E and J; green bars), acting as structural barriers that correlate with high-frequency radiators, delayed rupture propagation, reduced surface slip, and enhanced off-fault deformation (30, 34–38). Fault-parallel displacement waveforms recorded near these bends exhibit systematically larger dynamic overshoot (Fig. 2, E and J; stations 708, 145, 137, and 144) than those recorded near the centers of the intervening segments (Fig. 2, E and J; stations 712, 143, and 142).

Dynamic model of rupture propagation and arrest

The systematic pattern of near-field ground motions observed for large strike-slip earthquakes suggests that dynamic overshoot in fault-parallel displacement is associated with rupture arrest. To test this interpretation, we conducted dynamic rupture simulations of a vertical right-lateral strike-slip fault that is 45 km long and 15 km wide, generating synthetic earthquakes of Mw 6.9 (Fig. 3A, fig. S9, and table S3). Rupture nucleated at 7.5 km depth and propagated bilaterally before terminating against strength barriers at the fault ends, which represent abrupt rupture arrest (fig. S10).

Simulations with homogeneous elastic properties and homogeneous initial stress conditions failed to reproduce the observed near-field displacement overshoot (fig. S11; see Supplementary Materials). In these models, fault-parallel ground motions exhibit step-like displacement envelopes both near the fault center and near the rupture terminations, without significant overshoot (Fig. 3D and fig. S11). This behavior contrasts with observations from natural earthquakes.

Introducing depth-dependent seismic velocity structure markedly altered the modeled ground motions (table S4). Models with low near-surface wavespeeds that increase with depth produced pronounced along-strike variations in fault-parallel displacement (Fig. 3; see Supplementary Materials). Near rupture terminations, these simulations generated large dynamic overshoot (Fig. 3A; station S3), whereas locations farther from the fault ends exhibited step-like displacement histories (Fig. 3A; station S1). This spatial pattern closely matches the observations (Fig. 3C).

The modeled overshoot results from stopping-phase radiation generated during abrupt rupture arrest. When rupture encounters a sharp strength barrier, a stopping phase is produced that propagates back along the fault, rapidly decelerating slip (Fig. 3A and fig. S10). Because the rupture front is curved, deeper portions of the fault reach the barrier first, and rupture arrest progresses upward over a finite duration of approximately 2 s. During this interval, stopping phases are generated continuously as the locus of termination migrates vertically along the barrier edge. As this wave propagates upward toward Earth's free surface, through low-velocity material, its amplitude increases (fig. S12). Upon reaching the surface, the stopping phase produces transient fault-parallel ground velocities (0.5 m s^{-1}) of opposite polarity to fault slip, leading to displacement overshoot near rupture terminations (Fig. 3A; station S3). We verified that the spatial pattern and amplitude of overshoot in our models are not significantly affected by the assumption of a linear edge barrier by testing non-linear barrier geometries (fig. S13 and fig. S14).

Although stopping phases are generated in models with homogeneous material properties, their displacement amplitudes at the free surface do not result in overshoot (Fig. 3D and fig. S15). This result indicates that low wavespeed material at shallow depth is essential for amplifying stopping-phase ground motions to amplitudes detectable in near-field displacement records.

To distinguish abrupt rupture arrest from gradual termination, we also simulated scenarios in which rupture decelerated progressively toward the fault end (Fig. 3B and figs. S16 and S17; see Supplementary Materials). In these models, the sharp strength barrier was replaced by a smooth reduction in initial shear stress over a finite distance (Fig. 3B). Under these conditions, stopping-phase amplitudes were suppressed and fault-parallel displacement waveforms lacked significant overshoot (Fig. 3D), in contrast to near-field observations (Fig. 3C). An additional simulation incorporating segmented faults and a step-over geometry confirmed that transient rupture arrest

at internal segment boundaries produces stopping-phase signatures that are similar to those generated in the single-fault, abrupt arrest model (fig. S18).

Abrupt arrest of large strike-slip earthquakes.

Our results demonstrate that abrupt arrest of large strike-slip earthquakes produces a distinct overshoot in near-field fault-parallel ground displacement. This signature has likely gone undetected in many past earthquakes because near-fault seismic instrumentation has been sparse, limiting systematic observations of rupture arrest. The consistent appearance of this signature in several well-instrumented recent earthquakes underscores the increasing capability of modern seismic networks to resolve rupture dynamics with high fidelity (39). Near-fault records from the additional seven events included in our analysis may reflect the same underlying process (Fig. 1 and fig. S1); however, sparse station coverage in these cases prevents a definitive assessment.

Dynamic rupture simulations clarify the conditions under which displacement overshoot can occur. Our preferred models produced mildly crack-like rupture, where slip at a single point on the fault continued long after the rupture front had passed. However, we found that large overshoot can also arise in simulations dominated by pulse-like rupture behavior, characterized by a self-healing slip pulse of shorter duration (fig. S15). In these simulations, rapid fault re-strengthening behind the rupture front generated a negative phase of fault-parallel motion that trails the propagating rupture, producing pronounced overshoot along much of the fault length (fig. S19). This behavior differs from observations, as near-epicentral stations from the 1999 Mw 7.4 Izmit, 2016 Mw 7.0 Kumamoto, and 2023 Mw 7.7 Elbistan earthquakes consistently exhibited low η values (Fig. 2), more consistent with crack-like rupture behavior. Regardless of whether natural earthquakes preferentially rupture in crack-like or pulse-like modes, an arrest mechanism is required to explain the pronounced increase in η values observed near rupture terminations.

Our results provide near-field seismic evidence that large strike-slip earthquakes terminate abruptly, consistent with long-held theoretical models that predict stopping phase radiation upon rupture arrest (19). Abrupt termination of rupture propagation is an efficient mechanism to explain the observed high-frequency decay (ω^2) of global source spectra (19,40,41). On the other hand, earthquakes that terminate by gradual deceleration of rupture (42, 43) do not produce near-field stopping phases (Fig. 3B), and lack high-frequency energy (41, 44), suggesting gradual arrest is unlikely to be common for large earthquake ruptures.

Stopping-phase signatures are also observed at near-fault locations distant from the mapped rupture termini of large strike-slip earthquakes (Fig. 1). The 2023 Mw 7.8 Pazarcık earthquake provides a clear example, as multiple stations along the 100 km Amanos segment recorded stopping phases near local fault bends and step-overs. Our results suggest these observations record transient rupture arrest at structural barriers, consistent with dynamic rupture modeling of delayed rupture propagation through bends on the Amanos segment (30), and reductions in local slip amplitude (37, 45).

The prevalence of stopping phases adjacent to segment boundaries along strike-slip faults globally indicates that internal barriers to rupture propagation are common (Fig. 1 and fig. S1). This implies that ruptures propagate along strike-slip faults in a segmented manner within a single event, with multiple episodes of rupture arrest occurring at internal segment boundaries (Fig. 4 and fig. S18). Although such discontinuities in rupture propagation have been inferred from surface deformation patterns (46–48), and explored extensively in numerical modeling (8, 20, 49), the results presented here provide direct in-situ seismic evidence for this process.

These findings also have important implications for near-fault ground motion hazard. Engineering ground-motion models typically represent permanent fault-parallel displacement, using a one-sided pulse (referred to as 'fling step') (50, 51). Our results show that this displacement is frequently accompanied by a stopping phase that produces transient reverse fault-parallel motion (Fig. 1). Such large-amplitude displacement pulses are particularly damaging for base-isolated and other long-period structures in urban settings (52, 53). Stopping-phase signatures represent a generic earthquake source effect that should be explicitly considered in assessments of near-field ground motion hazard near rupture endpoints and internal fault-segment boundaries of large strike-slip earthquakes.

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Competing interests: Authors declare that they have no competing interests.

Data, code, and materials availability: Strong motion data for the 2023 Mw 7.8 Pazarcık (<https://tdvms.afad.gov.tr/list-station/543428/37.043/37.288>) and Elbistan earthquakes (<https://tdvms.afad.gov.tr/list-station/543593/37.239/38.089>) are available from the Turkish Disaster and Emergency Management Authority (AFAD). 10-Hz GNSS displacement data for the 2016 Mw 7.8 Kaikōura earthquake station CMBL was sourced from (66), and 1-Hz GNSS displacement data for 2016 Mw 7.0 Kumamoto earthquake station 701 are available at (67). Strong-motion data at Kasho Dam for the 2000 Mw 6.6 Tottori earthquake were used with permission from the local government of Tottori Prefecture. All other strong-motion seismic data, satellite-based coseismic displacement data, and LiDAR data used in this study are from previously published sources cited in the manuscript. The SPECSEM3D software is available at (78), and numerical data generated from dynamic rupture simulations are available at (68). All other data needed to evaluate the conclusions in this paper can be found in the paper or in the supplemental materials. No physical samples or materials were collected or generated for this study.

Supplementary Materials

Materials and Methods

Figs. S1 to S19

Tables S1 to S4

References (54-78)

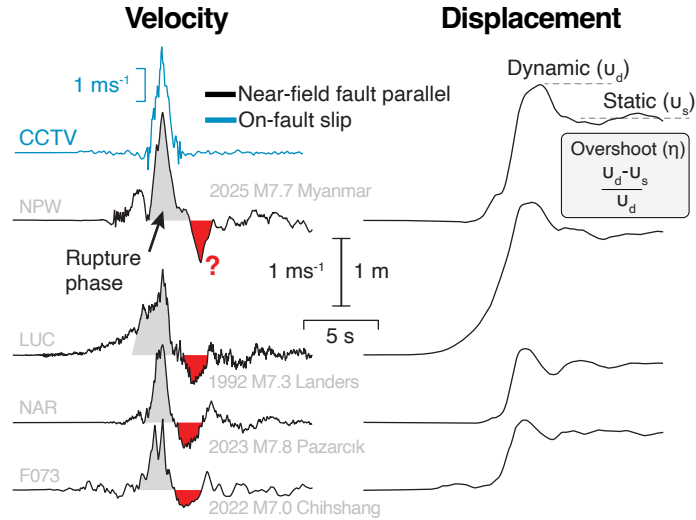


Fig. 1. Dynamic overshoot in near-field ground motions of large surface-breaking earthquakes. Fault-parallel velocity and displacement waveforms recorded by near-field (0.3-3 km) strong-motion instruments in four large surface-breaking strike-slip earthquakes (black): NPW (27), LUC (54), NAR (55), F073 (56). On-fault slip rate derived from Closed Circuit Television (CCTV) footage is shown for the 2025 Mw 7.7 Mandalay earthquake (27) (blue). All near-field records show a clear rupture phase (gray fill) where the ground moves in the direction of fault slip, followed by a negative phase where the ground moves in the opposing direction (red fill), which could be a stopping phase. The amount of dynamic overshoot (η) is defined by the difference in dynamic (u_d) and static (u_s) fault-parallel displacement normalized by u_d .

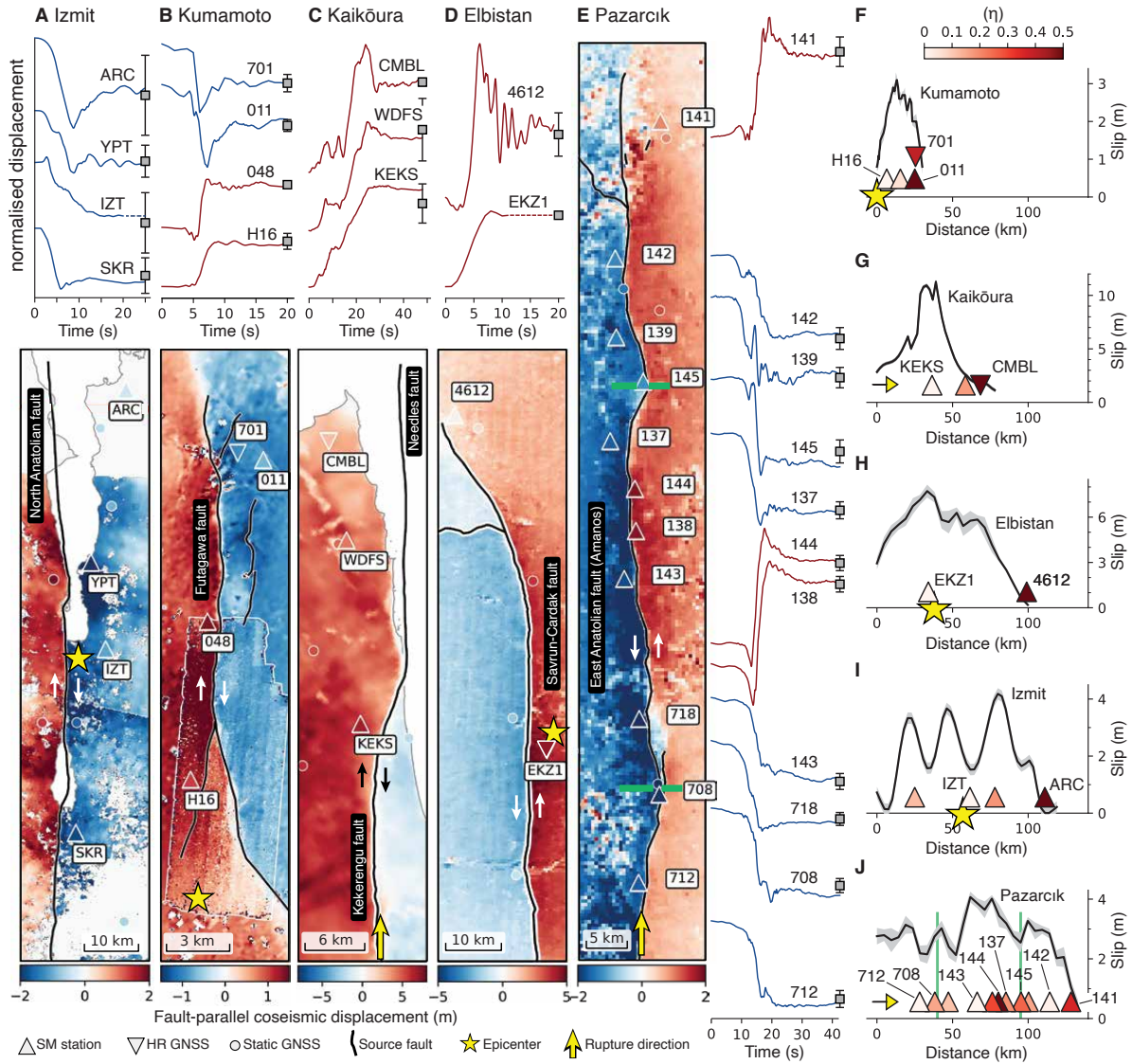


Fig. 2. Static and dynamic displacement for five large strike-slip earthquakes. Fault-parallel displacement waveforms are normalized by their final displacements and shifted along the time axis. Geodetic measurements of fault-parallel offset (gray squares) and their root-mean-square error are shown for comparison. Map colors show the magnitude of coseismic strike-slip displacement, and instruments are colored by their measured fault-parallel displacement (table S2). (A) 1999 Mw 7.4 Izmit earthquake (Türkiye) (57,58). (B) 2016 Mw 7.0 Kumamoto earthquake (Japan) (59–61). (C) 2016 Mw 7.8 Kaikōura earthquake (New Zealand) (62–64). (D) 2023 Mw 7.7 Elbistan earthquake (Kahramanmaraş doublet, Türkiye) (57,65). (E) 2023 Mw 7.8 Pazarcık earthquake (Kahramanmaraş doublet, Türkiye) (57,65), fault bends are shown (green bars). Waveforms near epicenters and fault centers show step-like envelopes, whereas large overshoot is recorded near rupture terminations. (F–J) Instrumental records are colored by overshoot (η). Distribution of fault-parallel slip (black curves) and uncertainties (gray shading) are from (57–65). Fault bends are shown (green bars). Rupture direction is towards the right.

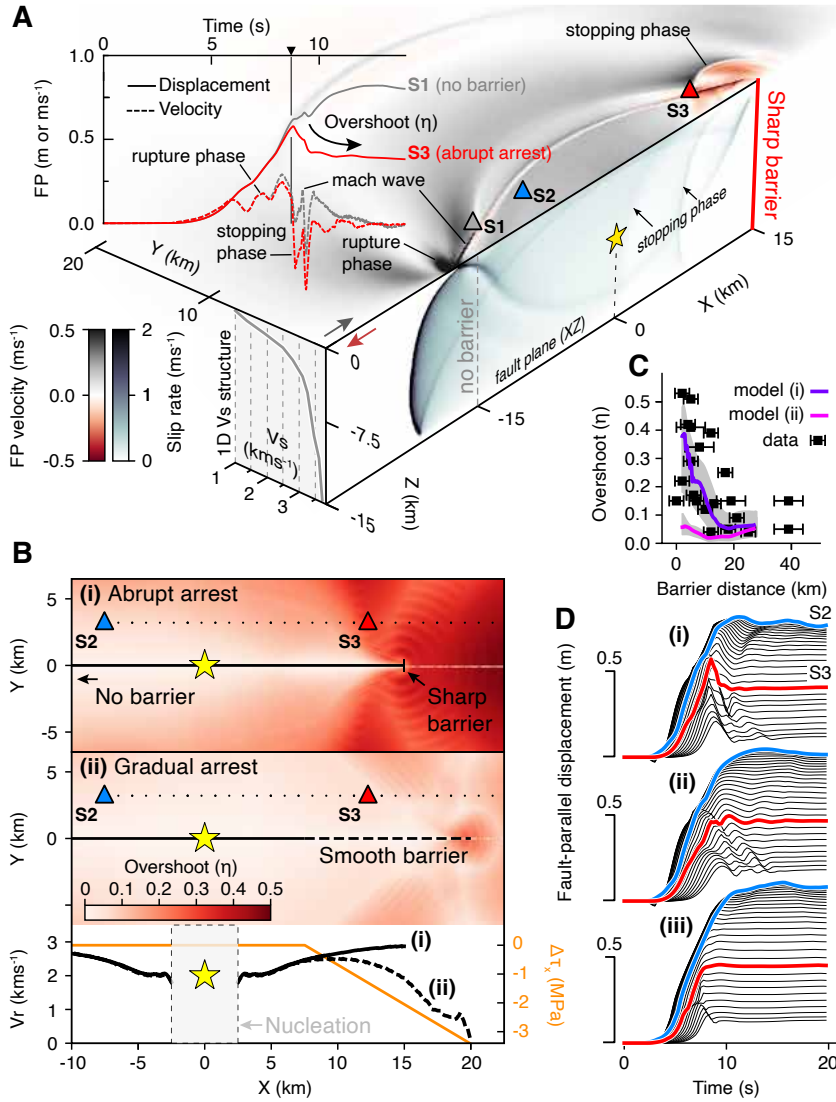


Fig. 3. Dynamic rupture model of stopping phase ground motions. (A) Snapshots of slip rate and fault-parallel (FP) ground velocity 8.5 s after nucleation of a M6.9 strike-slip earthquake at 7.5 km depth. Waveforms at synthetic stations S1 and S3 are shown. S3 shows stopping phase due to rupture arrest at sharp barrier, while S1 shows step-like displacements (low η) due to unimpeded rupture. (B) Distribution on Earth's free surface of overshoot (η) in abrupt (i) and gradual (ii) arrest models. Rupture velocity (V_r) and initial shear stress reduction ($\Delta\tau_x$) within the smooth barrier is shown. (C) Overshoot (η) comparison with observations for abrupt (i) (purple) and gradual arrest models (ii) (magenta), as a function of along-strike distance to rupture termination (or nearest barrier). Model curves show overshoot (η) at off-fault distance of 3 km. Error bars represent uncertainty in the location of rupture arrest (Materials and Methods), and gray curves show distribution (1σ) of model overshoot (η) between 0 and 6.5 km from the fault. (D) Waveforms for abrupt arrest (i), gradual arrest (ii), and homogeneous velocity and stress with abrupt arrest (iii). Traces are spaced every 1 km (black), with those from stations S2 and S3 highlighted in blue and red, respectively.

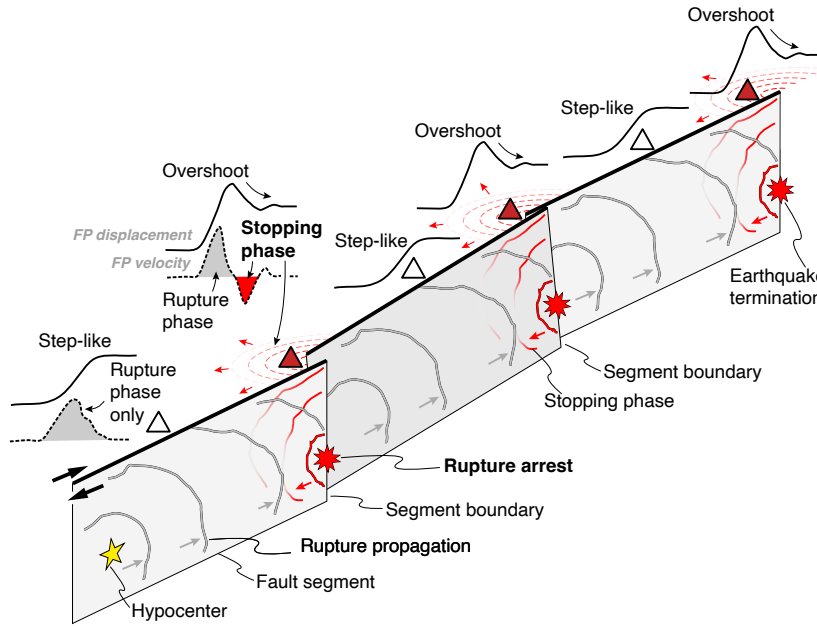


Fig. 4. Dynamics of segmented strike-slip ruptures. Conceptual model of dynamic rupture propagation through a segmented strike-slip fault, illustrating the fault-parallel (FP) ground motions expected near the segmented boundaries and at the centers of fault segments. Rupture initiates at one end of the fault (yellow star) and propagates unilaterally across the three fault segments. Earthquake propagation is punctuated by transient rupture arrest at segment boundaries (red stars), which radiates stopping phases and produces dynamic overshoot in near-fault ground motions. Rupture propagation is unimpeded at the centers of fault segments, and near-fault ground motions there are characterized by step-like displacements due to a lack of stopping phase radiation.



Supplementary Materials for
Stopping phase reveals abrupt arrest of large strike-slip earthquakes

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Materials and Methods

Strong-motion and HR-GNSS data

We analyze near-field strong-motion and HR-GNSS observations from several large strike-slip earthquakes using consistent baseline correction procedures. We use published strong-motion waveforms for the 2016 Mw 7.8 Kaikōura earthquake (66) and 2016 Mw 7.0 Kumamoto earthquake (69), which were previously corrected for baseline drift. For the 1999 Mw 7.4 Izmit earthquake records, we follow the method of (70), and obtain results consistent with previous analysis of the strong-motion data (26). For station IZT, and stations from the 2023 Kahramanmaraş earthquakes, we use a manual correction procedure because the automated method of (70) produced residual displacements that are inconsistent with independent geodetic offsets. Although the geodetic data may contain postseismic displacement accumulated in the days following the coseismic rupture, this is considered to be negligible.

For each station, we first remove the instrument response from the raw waveforms and demean each trace using a pre-event 10 second window. We then integrate the acceleration time series to obtain velocity waveforms. After inspecting the velocity data, we apply a linear correction to remove instrument tilt before integrating the velocity to obtain displacement time series. An example of the linear correction is shown in Figure S2. Residual displacements remain flat after strong shaking subsides, and we compare these results to static offsets measured by GNSS-calibrated satellite data. Comparison of the instrumentally-derived final displacements and the geodetically-derived static offsets along the fault-parallel direction are shown in Figure 2 and Table S2. Along-strike distances between the stations and the rupture terminations are measured relative to the end of mapped surface breaks along the main fault ruptures, with uncertainties of 5 km for onshore terminations, and 10 km assigned to offshore rupture termination (error bars Fig. 3C and fig. S16).

Geodetic validation of strong-motion displacements

Satellite-based images (e.g., optical images or synthetic aperture radar) taken before and after coseismic deformation are routinely used to measure the coseismic slip at the ground surface across faults (61), and to estimate slip at depth. These methods resolve relative displacements across the fault, but they are limited in their ability to determine ground displacement relative to a terrestrial

or inertial frame of reference. To validate displacements derived from double integration of strong-motion acceleration waveforms, we first align the image-derived deformation field to a terrestrial reference frame and correct for any long-wavelength distortions in the satellite images using GNSS observations.

Coseismic displacements measured by GNSS networks surrounding each earthquake source are used to perform this correction. For each GNSS station located within or adjacent to the footprint of the satellite data, we use the mean displacement of pixels within 100 m, and compute the residual difference between the image displacement and the GNSS displacement:

$$\Delta u_i = u_i^{\text{image}} - u_i^{\text{GNSS}},$$

where Δu_i is the residual at location i , u_i^{image} is the image-derived displacement, and u_i^{GNSS} is the GNSS-measured displacement.

We then model the residuals as a planar function of the image coordinates (x_i, y_i) :

$$\Delta u_i = ax_i + by_i + d,$$

where a , b , and d are coefficients representing a best-fitting plane. These coefficients are solved using least squares, and the fitted plane is subtracted from the image displacement field $u^{\text{image}}(x, y)$ to obtain a corrected displacement field:

$$u^{\text{corr}}(x, y) = u^{\text{image}}(x, y) - (ax + by + d).$$

To quantify the uncertainty of the corrected displacement field, we compare the corrected image displacements u_i^{corr} with the corresponding GNSS displacements u_i^{GNSS} . We use the root-mean-square error (RMSE) as a measure of the misfit:

$$\text{RMSE} = \sqrt{\frac{1}{N} \sum_{i=1}^N (u_i^{\text{corr}} - u_i^{\text{GNSS}})^2},$$

This procedure is applied separately to the east and north components of the coseismic displacement field. The data sources and results are shown in Figures S3 to S7. For some earthquakes analyzed in this study, the pre- and post-event satellite images span time periods during which additional surface deformation occurred. This includes deformation from an M6.5 foreshock preceding the 2016 Mw 7.0 Kumamoto earthquake (59) and overlapping deformation fields associated with

the 2023 Kahramanmaraş earthquake doublet (65). In these cases, the summed deformation from both events is used as the coseismic GNSS displacement to correct the satellite-based geodetic data.

In addition to the satellite-based images, the 2016 Mw 7.0 Kumamoto earthquake offers high-resolution pre- and post-earthquake airborne LiDAR topographic data, which is used to derive coseismic deformation across the source faults (60). We use these LiDAR-based displacements instead of satellite-based data to validate strong-motion stations that fall inside the footprint of the LiDAR surveys (stations H16 and 048; Fig. 2B). The geodetically-derived static offsets are displayed alongside the instrumentally measured displacement timeseries (gray squares in Figure 2) and are listed in Table S2.

Dynamic model setup

We simulate spontaneous dynamic rupture using a model inspired by exercise TPV32 from the Statewide California Earthquake Center and United States Geological Survey Spontaneous Rupture Code Verification Project (71). While full details of the benchmark problem are available online (<https://strike.scec.org/cvws/>), we summarize here the key aspects of the model configuration relevant to this study. Dynamic rupture is simulated on a vertical, right-lateral strike-slip fault that is 45 km long and 15 km wide, embedded in an elastic half-space (fig. S9a).

Fault friction is governed by a linear slip-weakening law (72, 73). When the fault is sliding, the shear strength τ at a given point on the fault is defined as:

$$\tau = C_0 + \mu \max(0, \sigma_n).$$

where C_0 is the frictional cohesion, μ is the coefficient of friction, and σ_n is the effective normal stress. The coefficient of friction evolves with slip according to

$$\mu = \mu_s + (\mu_d - \mu_s) \min\left(\frac{D}{d_c}, 1\right).$$

where μ_s and μ_d are the static and dynamic friction coefficients, D is the accumulated slip, and d_c is the critical slip distance. Backward slip is not permitted in the model, so left-lateral motion across the fault plane does not occur.

Following exercise TPV32, we assume frictional cohesion on the fault at shallow depth:

$$C_0 = \begin{cases} (0.000425 \text{ MPa/m})(2400 \text{ m} - \text{depth}), & \text{if depth} \leq 2400 \text{ m}, \\ 0 \text{ MPa}, & \text{if depth} \geq 2400 \text{ m}. \end{cases}$$

Under this formulation, C_0 is 1.02 MPa at the Earth's surface and decreases linearly to zero at a depth of 2400 m. At greater depths, frictional cohesion is zero.

The fault is initially at rest, and rupture is nucleated by applying an additional shear stress within a circular patch of radius 2.0 km centered at the hypocenter, such that the local stress exceeds the fault yield strength (Fig. S9a). A defining feature of the TPV32 model is a one-dimensional, depth-dependent structure in seismic velocity, fault strength, and initial stress that increases continuously with depth (Fig. S9, Table S4). This velocity and stress structure is intended to approximate generic, but realistic crustal conditions.

Both the effective normal stress σ_n and the initial shear stress τ_x scale with the ratio of the local shear modulus to a reference shear modulus of 32.0 GPa, resulting in depth-dependent initial stress conditions (Fig. S9e). The lateral and basal edges of the fault are surrounded by regions of elevated μ_s , which prevent rupture from propagating beyond the modeled fault area, while the top of the fault coincides with a free surface. Dynamic rupture simulations are performed using a spectral element method (74).

While the fault geometry in exercise TPV32 is 30 km long and 15 km wide, we extend the fault length to 45 km to isolate the dynamics of unilateral rupture termination. Specifically, the fault is extended by 15 km in the negative X direction, from $X = -15$ km to $X = -30$ km (Fig. S9a), so that waves radiated during rupture arrest at one end of the fault do not interfere with near-field ground motions at the opposite end. This asymmetric geometry also enables direct comparison between rupture propagation toward a strong edge barrier and unimpeded rupture propagation in the opposite direction. When the bilateral rupture fronts reach $X = \pm 15$ km, rupture propagating toward $X = 15$ km encounters a strong barrier, while rupture propagating toward $X = -15$ km continues without obstruction (Fig. 3A).

Model results are identical to those of exercise TPV32 until rupture reaches $X = \pm 15$ km. Beyond this point, ground motions begin to deviate slightly due to wave arrivals generated by rupture propagating along the extended portion of the fault between $X = -15$ km and $X = -30$

km. Snapshots of rupture evolution and near-fault ground velocity at the free surface are shown in Figure S10. Because this model closely follows the TPV32 configuration, we refer to it as the TPV32 model.

Smooth arrest model

To investigate rupture termination by gradual deceleration rather than abrupt arrest in the TPV32 model, we implement a smooth arrest model by extending the fault length from $X = 15$ km to $X = 22.5$ km and reducing the initial shear stress near the end of the fault. This configuration prevents rupture from encountering a sharp edge barrier and instead promotes progressive deceleration of rupture propagation. The initial shear stress τ_x is reduced according to

$$\tau_x(X) = \tau_{x0} (1 - 0.01 (X - 7.5)), \quad 7.5 < X < 22.5 \text{ km.}$$

Figure 3B illustrates the resulting reduction in τ_x at a depth of -7.5 km. As rupture propagates into this region of reduced shear stress, rupture velocity decreases gradually before spontaneous arrest occurs at $X = 20$ km. Snapshots of rupture propagation, near-fault ground velocity, and the resulting spontaneous arrest are shown in Figure S17.

We also simulate gradual rupture arrest over shorter distances, by increasing the degree of stress reduction (Fig. S16). In these models, rupture velocity deceleration occurs over distances between approximately 2-5 km. Such models do not produce overshoot consistent with near-field observations of rupture arrest (Fig. S16). This suggests that rupture propagation in our models needs to decelerate and arrest over distances shorter < 2 km, in order to produce stopping phase ground motions consistent with near-field observations.

Model with a homogeneous elastic half-space

To isolate the role of depth-dependent strength and stresses, and velocity structure in controlling rupture arrest and near-fault ground motions, we simulate a model with homogeneous elastic properties and homogeneous initial stress conditions. Material properties are chosen to match those at a depth of $Z = -5$ km in the reference TPV32 model. Specifically, we adopt a P-wave speed of $V_p = 5500$ m/s, an S-wave speed of $V_s = 3250$ m/s, an initial shear stress of $\tau_x = 26.2$ MPa, and

an effective normal stress of $\sigma_n = 52.4$ MPa. Under these conditions, the resulting dynamic stress drop is 6.81 MPa.

Snapshots of the dynamic rupture evolution are shown in Figure S11. Compared with the reference TPV32 model, these simulations exhibit a pronounced reduction in negative fault-parallel ground velocities in the negative X direction near the rupture termination at $X = 15$ km. As a result, the large displacement overshoot (η) observed in the reference model is strongly suppressed in the homogeneous case (e.g., Fig. S15c versus Fig. S15a).

To further examine the influence of shallow velocity structure, we perform additional simulations with varying degrees of near-surface velocity reduction (Fig. S12a). We find that fault-parallel displacement overshoot at station S3 increases systematically with decreasing shallow shear-wave velocity, while the fully homogeneous model produces the smallest overshoot values, with $\eta < 0.05$. In the default model that includes shallow low-velocity structure, overshoot magnitude depends strongly on the depth-dependent V_s structure beneath S3 (Fig. S12c). This behavior indicates that as stopping-phase radiation propagates upward toward the free surface, its contribution to fault-parallel displacement is increasingly amplified as it encounters materials with lower seismic wavespeeds.

Pulse model

We simulate pulse-like rupture propagation by modifying the slip-weakening friction law used in the TPV32 model (Fig. S9d). The dynamic friction coefficient is reduced from $\mu_d = 0.45$ to $\mu_d = 0.40$, and rapid fault re-strengthening is introduced behind the rupture front. In this formulation, the coefficient of friction increases from $\mu_d = 0.40$ to $\mu_s = 0.58$ over a slip distance of 0.3 m. Fault re-strengthening initiates once the fault has accumulated 0.72 m of slip, corresponding to four times the critical slip distance ($d_c = 0.18$ m) (Fig. S9d).

These frictional parameters produce a strongly pulse-like rupture, as illustrated by snapshots of rupture evolution in Figure S19. At near-fault locations, a pronounced negative phase of fault-parallel ground velocity follows the rupture phase, resulting in large fault-parallel displacement overshoot η along much of the simulated rupture length (Fig. S15b). In contrast, the stopping-phase signatures associated with rupture arrest at the fault edge are similar to those produced by the crack-like rupture model, indicating that the characteristics of stopping phases are largely insensitive to whether rupture propagates in a pulse-like or crack-like mode (Fig. S19).

The spatial pattern of overshoot produced by the pulse model differs from that observed in natural earthquakes. In particular, the pulse-like rupture generates large near-fault overshoot in the epicentral region (Fig. S15b), whereas the large strike-slip earthquakes analyzed in this study consistently exhibit low η values near their epicenters. This discrepancy indicates that the observed near-field ground motions are more consistent with crack-like rupture behavior than with strongly pulse-like rupture propagation (Fig. S15).

Non-linear barrier model

We explore the influence of the idealized sharp barrier on the overshoot near the rupture termination. The sharp linear barrier presented in Figure 3 is replaced with a non-linear barrier geometry defined by a sinusoidal curve as a function of depth (z) on the fault plane:

$$x(z) = A \sin\left(\frac{2\pi}{\lambda}z\right), \quad (\text{S1})$$

where $A = 1000$ m and $\lambda = 5000$ m. The geometry of this non-linear edge barrier is shown in fig. S13a. Snapshots of rupture encountering this non-linear barrier are shown in figure S14. Both the fault-parallel displacement waveforms and the spatial distribution of near-fault overshoot (η) are similar for the linear and non-linear barrier geometries (fig. S13), demonstrating that modeled overshoot does not strongly depend on the assumption of an idealized linear barrier geometry.

Segmented fault model

To examine rupture propagation and arrest at internal fault-segment boundaries, we simulate a segmented two-fault system separated by a right-stepping extensional stepover. This dynamic rupture model is similar to exercise TPV22 from the Statewide California Earthquake Center and United States Geological Survey Spontaneous Rupture Code Verification Project (71). Full details of the benchmark configuration are available online (<https://strike.scec.org/cvws/>), and we summarize here the aspects relevant to this study. We modify this model to replicate the material and stress conditions of TPV32, including the continuous 1D structure in seismic velocity, fault strength, and initial stress. To increase the initial shear stresses and promote rupture nucleation on the secondary fault, we increase the component of the stress tensor parallel to the fault (σ_{13}) from 30 MPa to 31 MPa.

In this model, rupture initiates on fault 1 (Fig. S18) and propagates along strike until it becomes transiently arrested at the segment boundary at the edge of fault 1. Following this temporary arrest, a secondary rupture is dynamically triggered on fault 2 near the segment boundary and subsequently propagates along fault 2. Figure S18 shows a snapshot of the simulated rupture evolution at 10 s after the origin time. The simulation captures both the stopping phase radiated during transient rupture arrest at the segment boundary and the emergence of the rupture phase associated with the dynamically triggered rupture on fault 2.

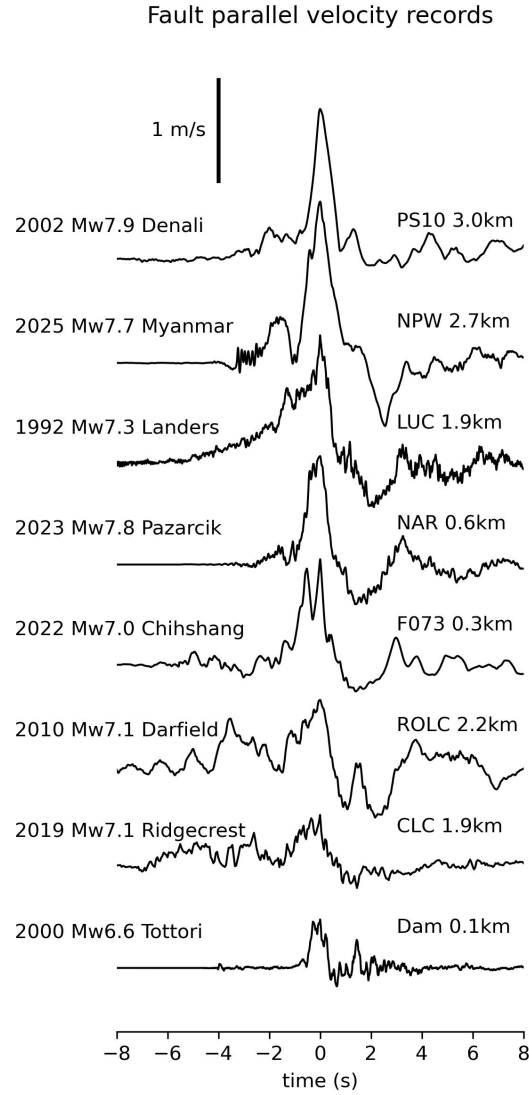


Figure S1: Unfiltered near-field fault-parallel velocity seismograms from strike-slip earthquake ruptures not included in the detailed analysis in Fig. 2 in the main text. Traces are aligned by their maximum velocity, and distance to surface fault rupture is shown next to the station code. References: 2002 Mw 7.9 Denali (75), 2025 Mw 7.7 Mandalay (27), 1992 Mw 7.3 Landers (54), 2023 Mw 7.8 Pazarcık (55), 2022 Mw 7.0 Chihshang (56), 2010 Mw 7.1 Darfield (76), 2019 7.1 Ridgecrest (70), 2000 Mw 6.6 Tottori (77). Note that the 2022 Mw 7.0 Chihshang earthquake was dominantly strike-slip but contained a significant reverse component of slip.

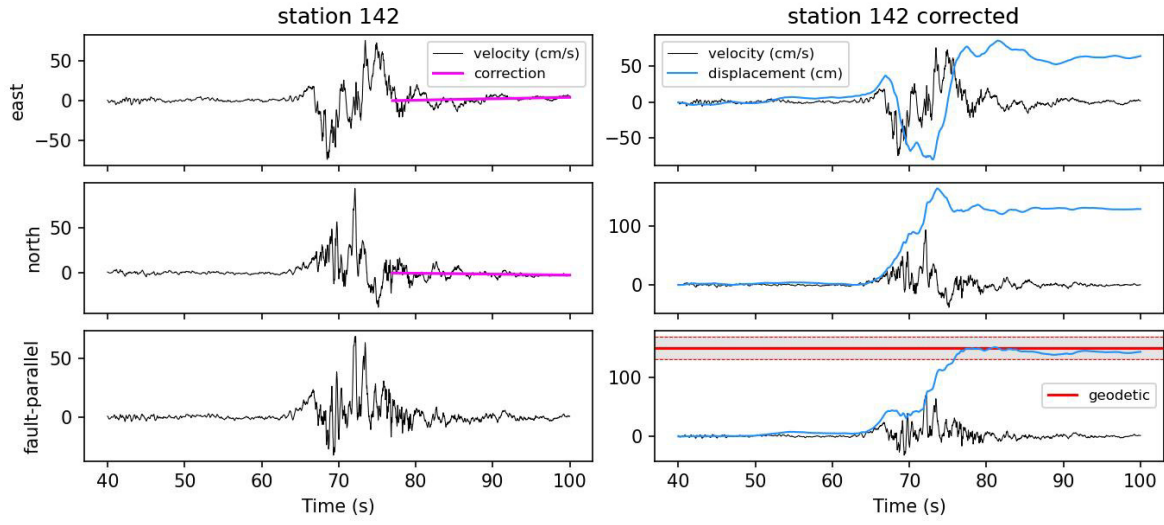


Figure S2: Velocity and displacement seismograms for station 142 from the 2023 Mw 7.8 Pazarcık earthquake. The left column shows the ground velocity before baseline correction, and also shows the linear baseline correction function (magenta). The right column shows the ground velocity (black) and displacement after the correction is made (blue). Horizontal red line shows the fault-parallel displacement from the independent geodetic dataset, the width of which corresponds to the RMSE from the correction of the geodetic image using coseismic GNSS data.

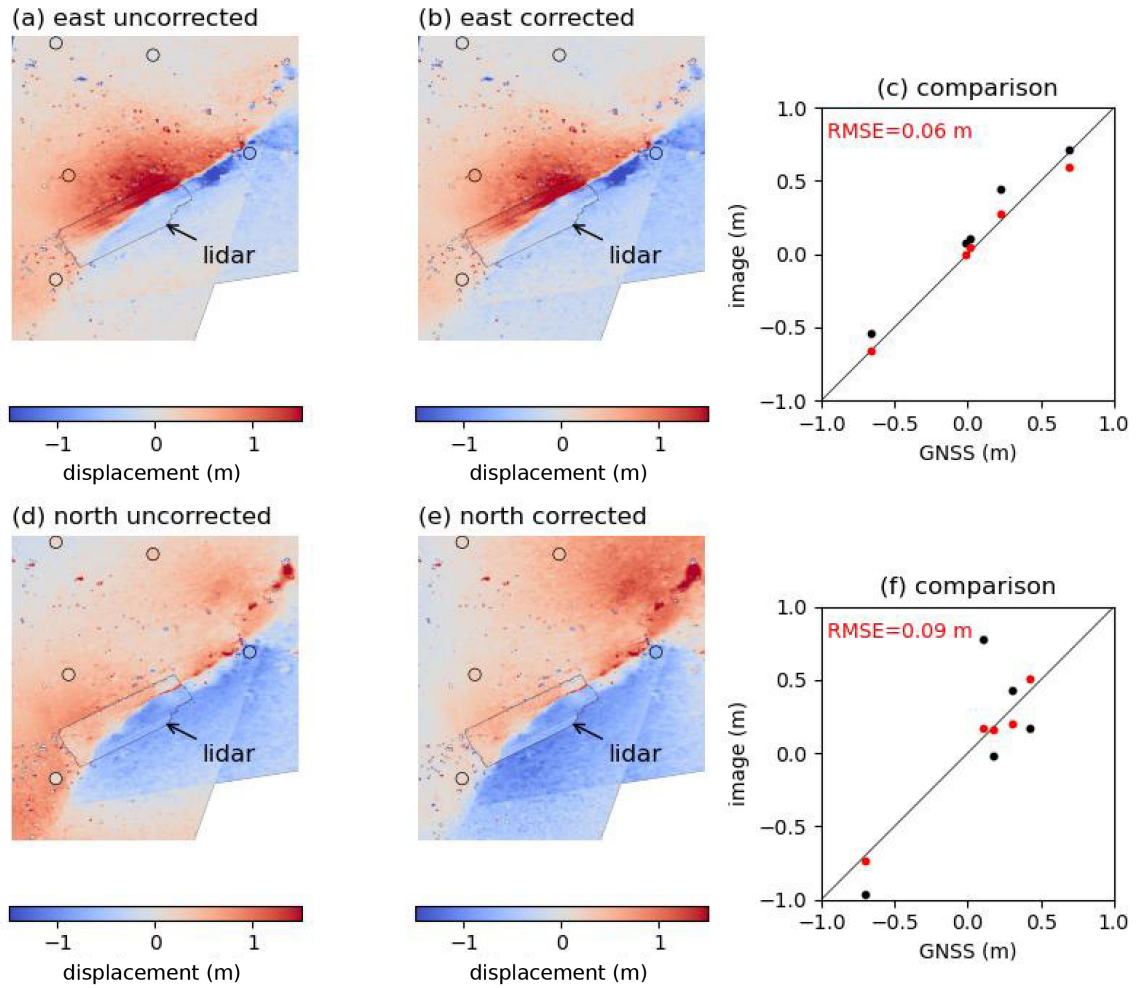


Figure S3: Horizontal coseismic displacements for the 2016 Mw 7.0 Kumamoto earthquake. Uncorrected east and north (a and d) and corrected (b and e) components of the displacement field, respectively. Red-blue shading represents displacements derived from satellite images (61), while the colored circles show the coseismic GNSS displacements plotted to the same scale (59). Displacements extracted from pre- and post-earthquake LiDAR topography are also displayed (60). Comparison of GNSS and satellite-based measurements of coseismic displacement before (black) and after correction is applied (red) are shown for the east (c) and north (f) components.

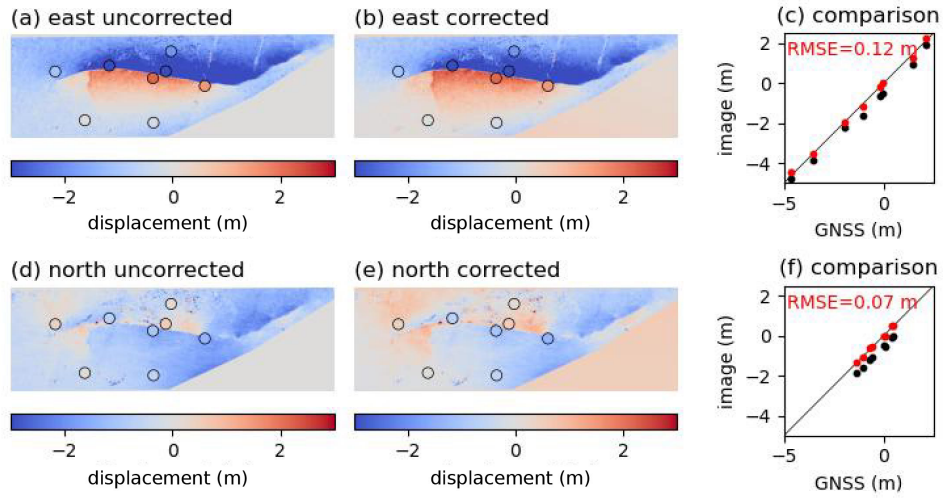


Figure S4: Horizontal coseismic displacements for the 2023 Mw 7.7 Elbistan earthquake. Uncorrected east and north (a and d) and corrected (b and e) components of the displacement field, respectively. Red-blue shading represents displacements derived from satellite images (57), while the colored circles show the coseismic GNSS displacements plotted to the same scale (65). Comparison of GNSS and satellite-based measurements of coseismic displacement before (black) and after correction is applied (red) are shown for the east (c) and north (f) components.

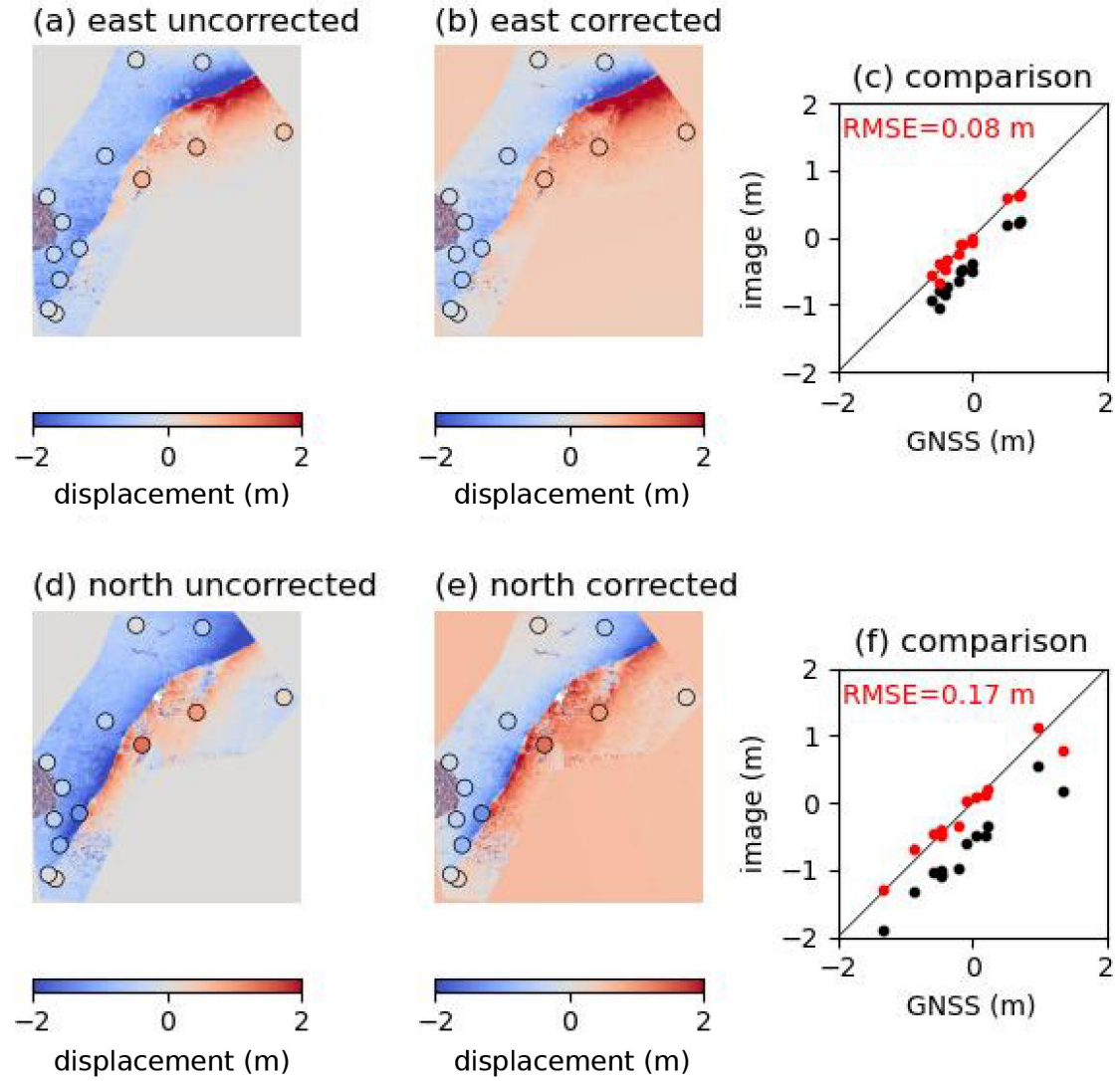


Figure S5: Horizontal coseismic displacements for the 2023 Mw 7.8 Pazarcık earthquake. Uncorrected east and north (a and d) and corrected (b and e) components of the displacement field, respectively. Red-blue shading represents displacements derived from satellite images (57), while the colored circles show the coseismic GNSS displacements plotted to the same scale (65). Comparison of GNSS and satellite-based measurements of coseismic displacement before (black) and after correction is applied (red) are shown for the east (c) and north (f) components.

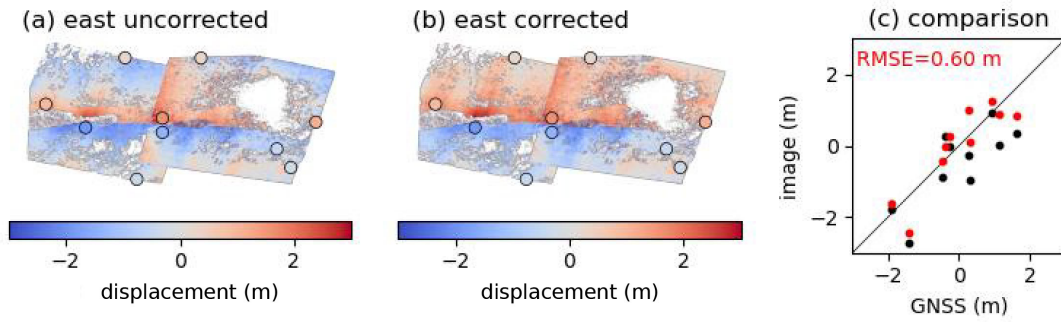


Figure S6: Horizontal coseismic displacements for the 1999 Mw 7.4 Izmit earthquake. Uncorrected (a) and corrected (b) east component of the displacement field. As the average fault strike is aligned east-west, the north component of displacement is not needed to extract the fault-parallel displacement. Red-blue shading represents displacements derived from satellite images (57), while the colored circles show the coseismic GNSS displacements plotted to the same scale (58). Comparison of GNSS and satellite-based measurements of east coseismic displacement before (black) and after correction is applied (red).

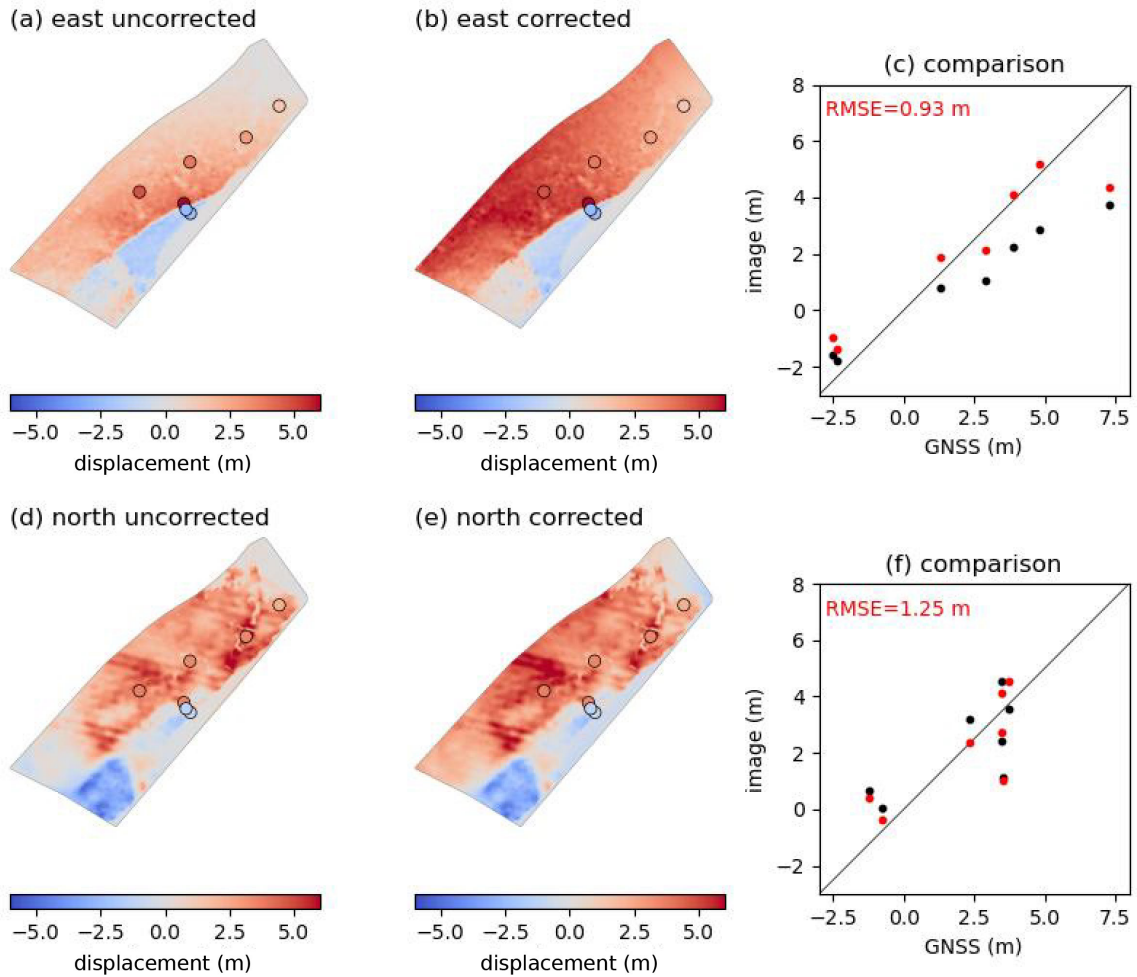


Figure S7: Horizontal coseismic displacements for the 2016 Mw 7.8 Kaikōura earthquake. Uncorrected east and north (a and d) and corrected (b and e) components of the displacement field, respectively. Red-blue shading represents displacements derived from satellite images (62), while the colored circles show the coseismic GNSS displacements plotted to the same scale (62, 64). Comparison of GNSS and satellite-based measurements of coseismic displacement before (black) and after correction is applied (red) are shown for the east (c) and north (f) components.

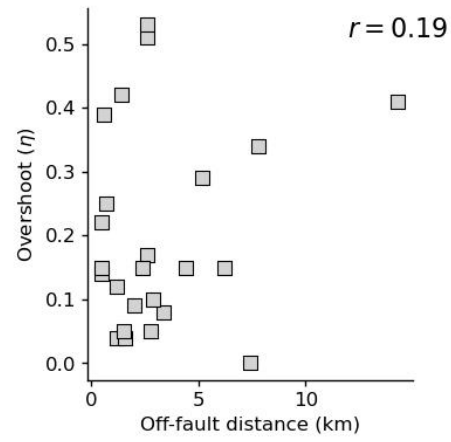


Figure S8: Dynamic overshoot (η) as a function of off-fault distance for near-fault seismo-geodetic data. "r" is the Pearson coefficient.

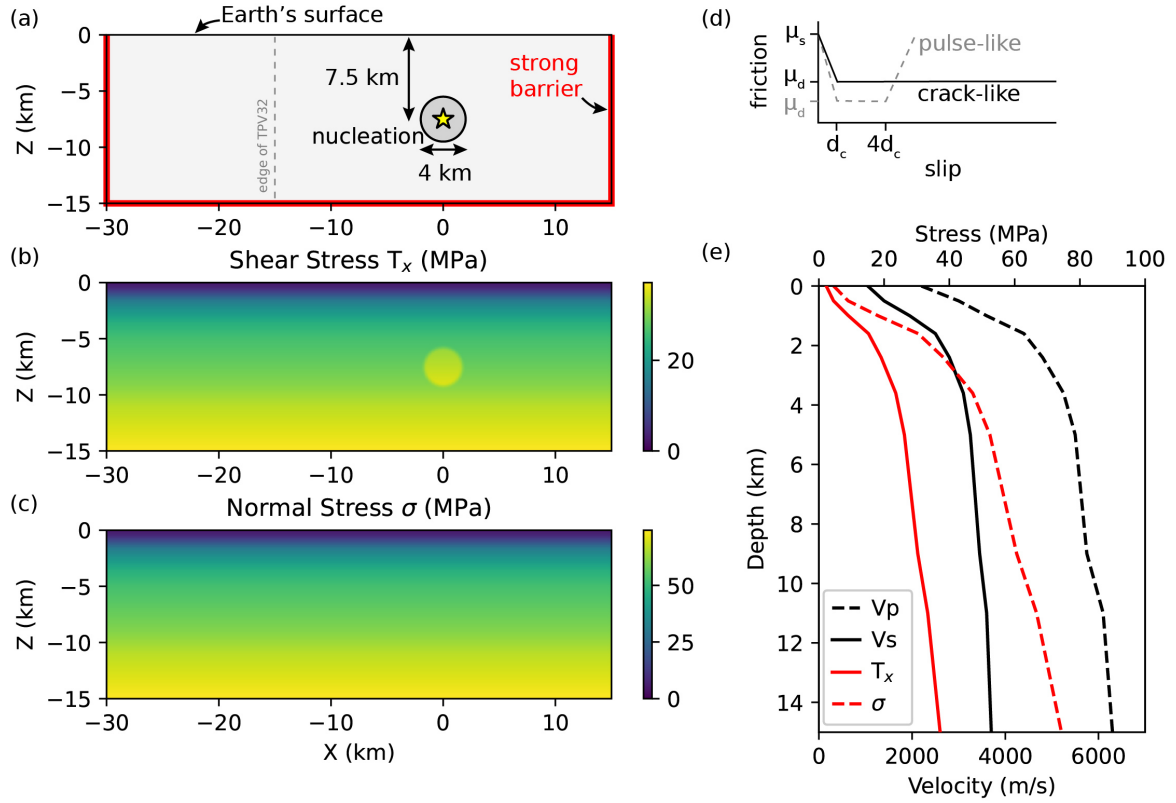


Figure S9: Dynamic rupture model initial conditions, illustrating the fault geometry and nucleation location (a), initial right-lateral shear stress distribution (along-dip initial stresses are 0 MPa everywhere) (b), initial normal stress distribution (c), slip-weakening friction for crack-like and pulse-like rupture modes (d), and 1D distribution of velocity and on-fault stresses (e).

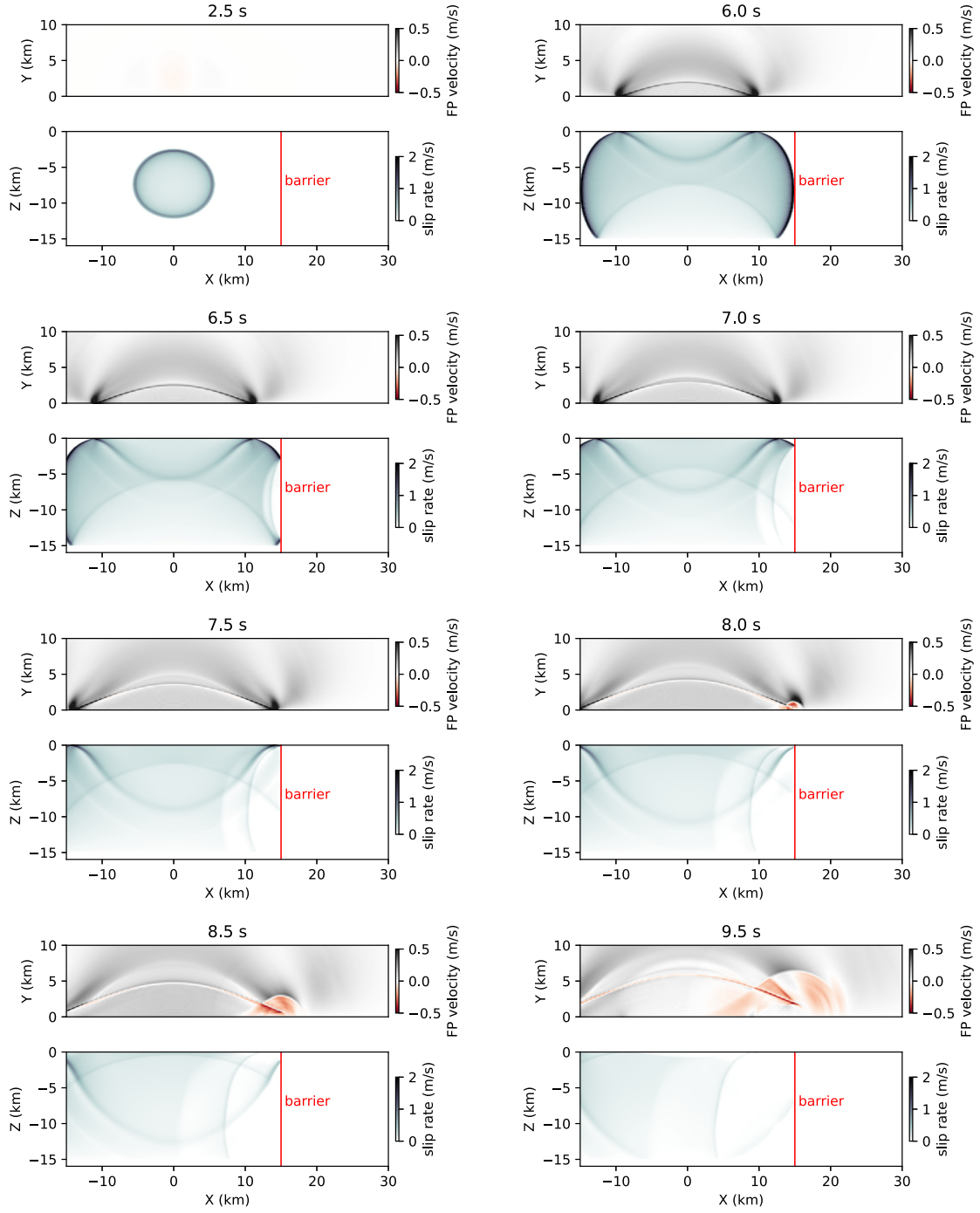


Figure S10: Snapshots rupture and surface deformation. Each pair of images represents a different timestep of the dynamic rupture simulation. Top panel shows a map view of the fault-parallel ground velocity at the free surface adjacent to the fault (fault is located at $Y=0$ km). Bottom panel shows the on-fault slip rate. Barrier to rupture is located at $X=15$ km.

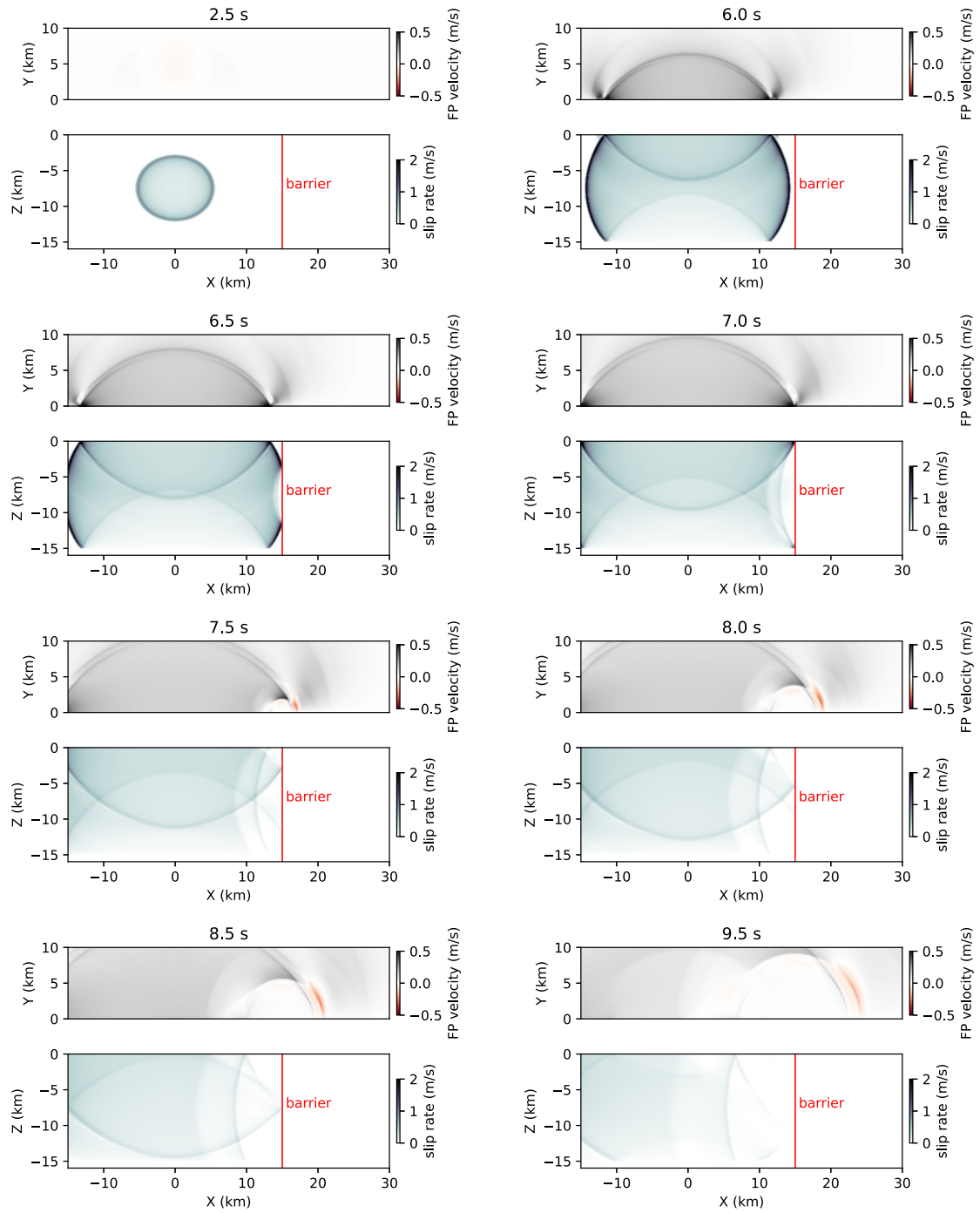


Figure S11: Same as Figure S10 but for a model with homogeneous material properties and homogeneous stress distribution.

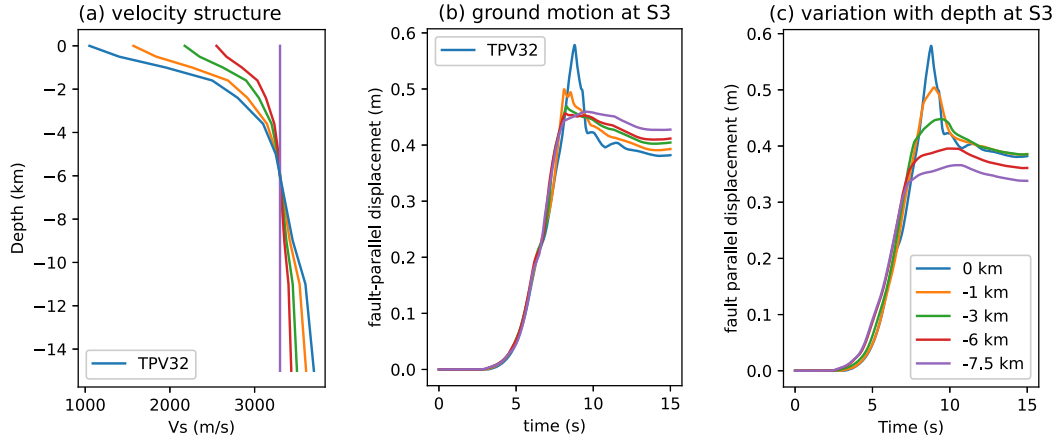


Figure S12: Effects of 1D V_s structure (a) on stopping phase ground motions at station S3 ($X=12$ km, $Y=3$ km) (b). Overshoot η is suppressed by more homogeneous velocity structures. We also illustrate for 1D velocity TPV32 (blue curve in part Fig S12a) the difference in overshoot with depth beneath station S3 (c), showing that large η is only observed at shallow depth where V_s is approximately 1000 m/s.

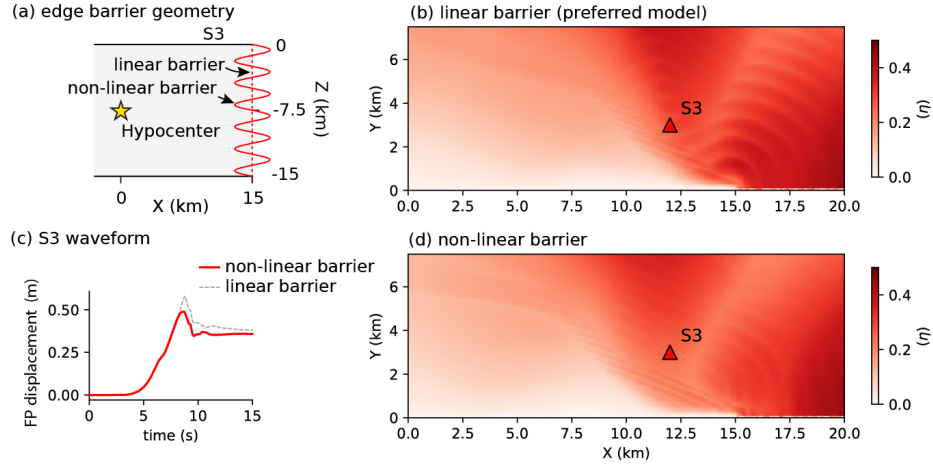


Figure S13: Rupture arrest against a non-linear barrier. (a) Geometry of the non-linear edge barrier model. (b) Spatial distribution of overshoot (η) surrounding the source fault for the preferred model with a sharp linear edge barrier geometry. (c) Waveforms for station S3 for models with a non-linear barrier (red) and linear barrier (dashed gray). (d) Spatial distribution of overshoot (η) surrounding the source fault for the preferred model with a sharp non-linear barrier geometry.

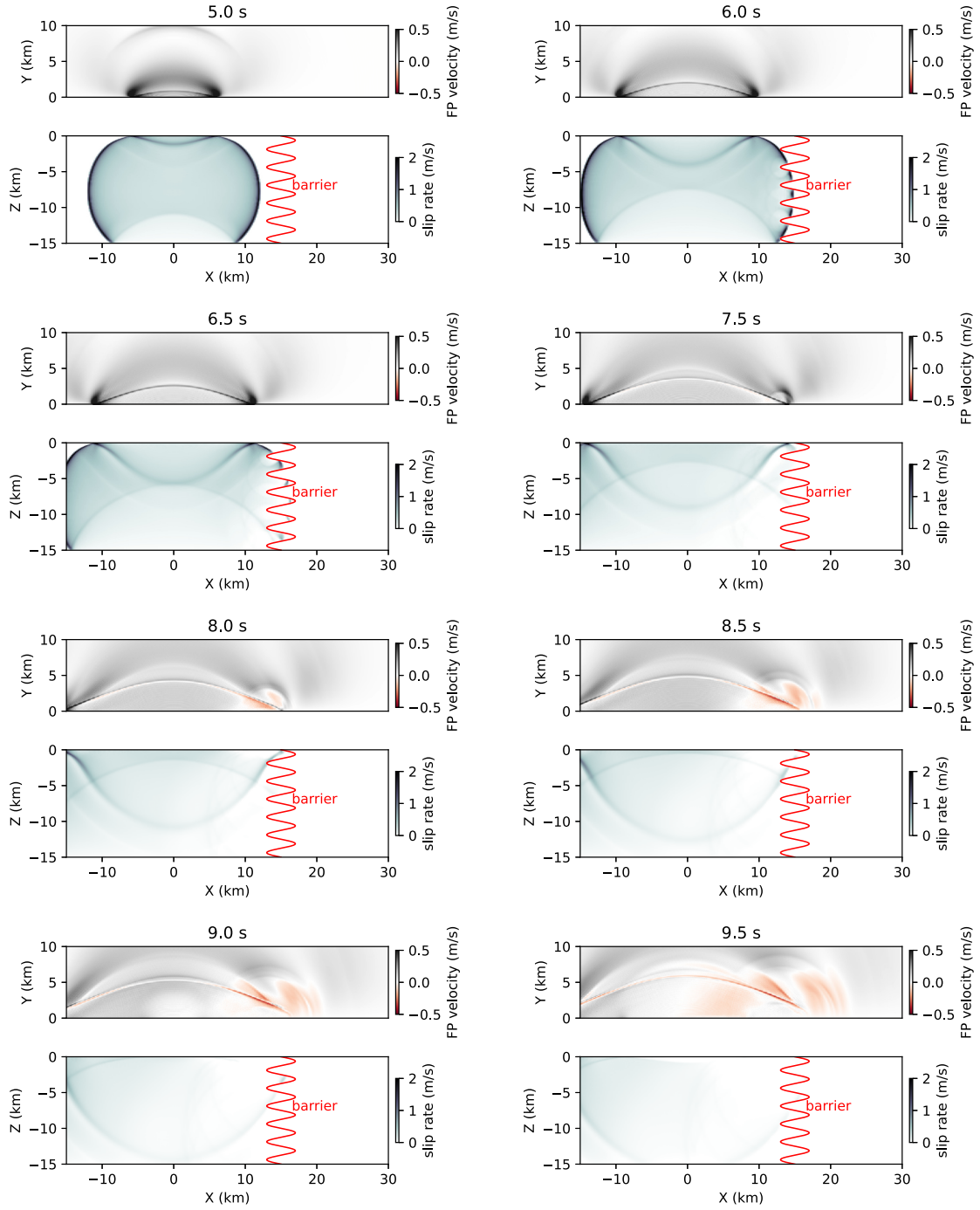


Figure S14: Same as Figure S10 but for a model with a curved edge barrier geometry, see fig. S13

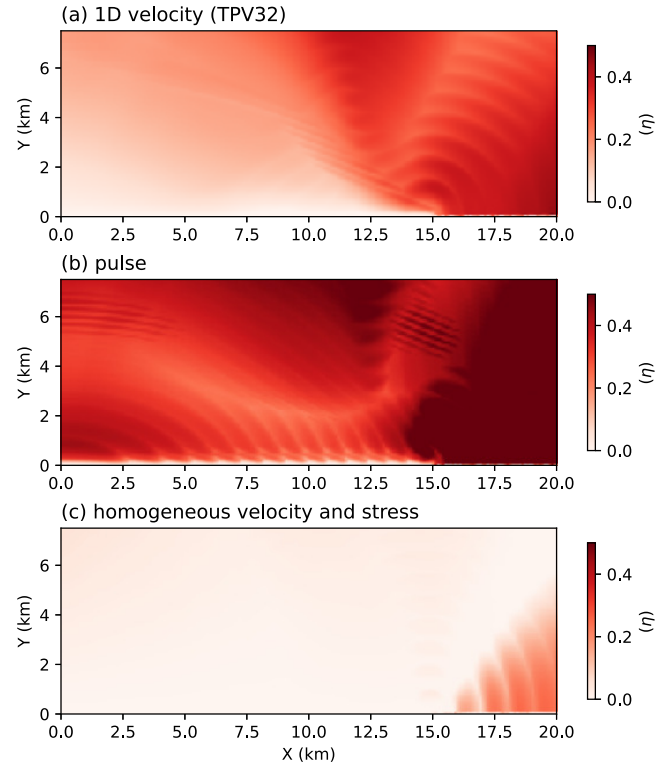


Figure S15: Near-fault distribution of η for different model parameters. (a) Default model shown in Fig. 3A (TPV32), (b) Model with strongly pulse-like rupture propagation. This model shows large η along the length of the fault. see also Fig. S19. (c) Model with homogeneous velocity structure and homogeneous on-fault stress.

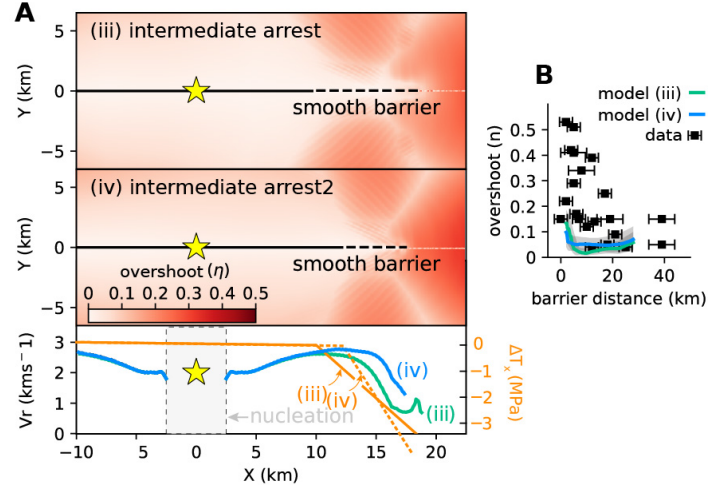


Figure S16: Gradual arrest models. (A) In model (iii) rupture velocity decelerates over approximately 5 km distance, whereas in model (iv) rupture velocity decelerates over approximately 2 km. In both cases, the amount of overshoot ($\eta < 0.2$), and not consistent with near-fault observations. (B) Model curves show overshoot (η) at off-fault distance of 3 km (green and blue curves), same as in Figure 3C. Error bars represent uncertainty in the location of rupture arrest (Materials and Methods), and gray curves show distribution (1σ) of model overshoot (η) between 0 and 6.5 km from the fault.

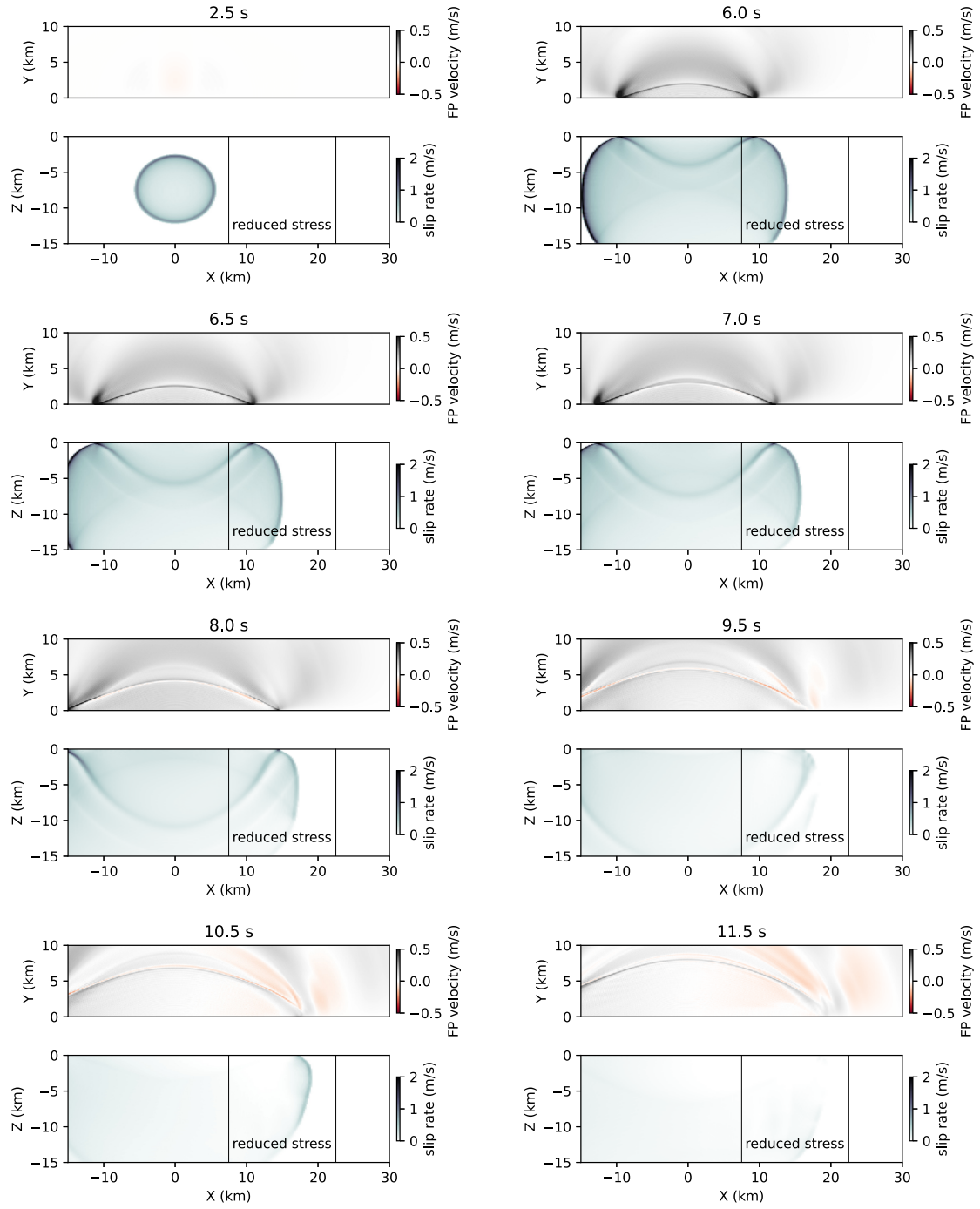


Figure S17: Same as Figure S10 but for a model without a sharp barrier. Instead, the initial shear stress driving the rupture is linearly decreasing between $7.5 < X < 22.5$ to produce a smooth arrest of the rupture.

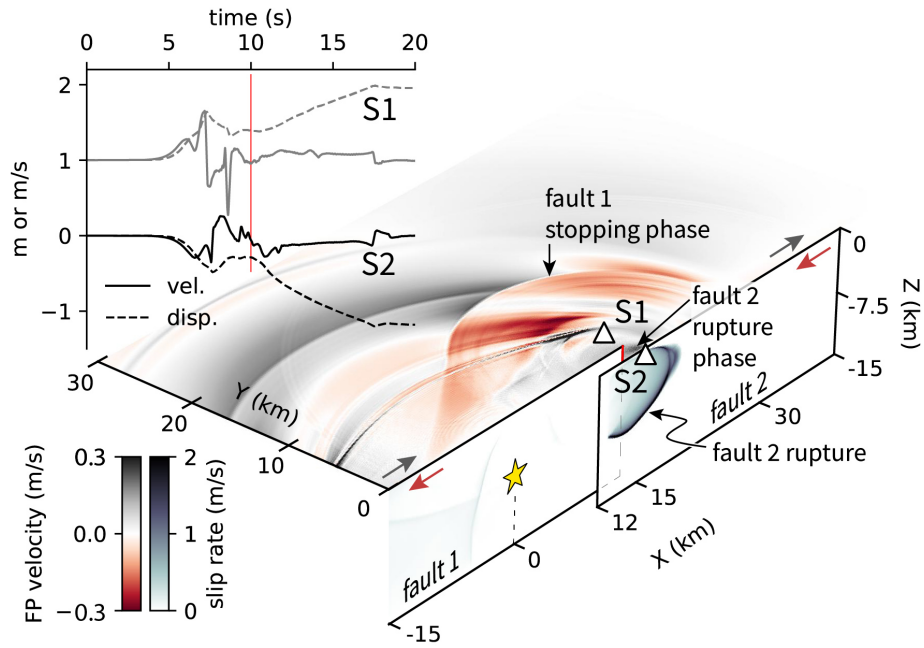


Figure S18: Dynamic rupture model of a segmented fault. Snapshot of dynamic rupture model based on SCEC benchmark exercise TPV22 (71), showing slip rate and free-surface fault-parallel ground velocity. At 10 s, rupture propagation has become arrested on fault 1 against the sharp segment boundary, producing a prominent stopping phase seen at the free surface, and a dynamically triggered secondary rupture has begun to propagate on fault 2 near the segment boundary. The emergence of this secondary rupture has generated a rupture phase that is following in the wake of the stopping phase that was produced during the previous episode of rupture arrest of fault 1.

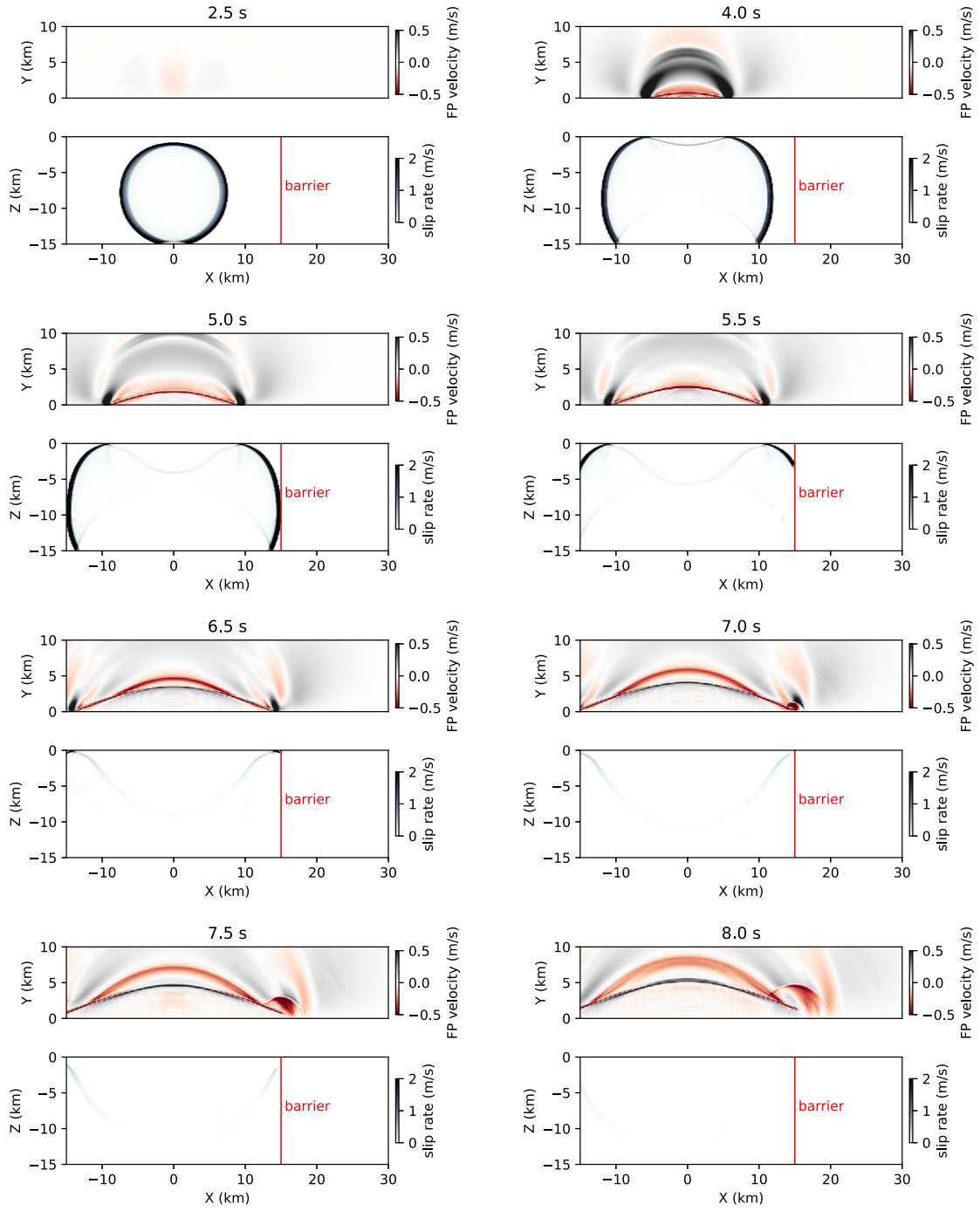


Figure S19: Same as Figure S10 but for a model where rupture is pulse-like. Notice the negative fault-parallel velocity at the free surface that trails the rupture front. This is responsible for the large η values along the entire length of the fault.

Table S1: 12 large strike-slip earthquakes with near-field instrumental records. Earthquakes are included if near-field records contain static offsets above 0.1 m.

Earthquake	location	Mw	stations < 15 km
2025 Myanmar	Myanmar	7.7	1
2023 Pazarcık	Türkiye	7.8	13
2023 Elbistan	Türkiye	7.7	2
2022 Chihshang	Taiwan	7.0	5
2019 Ridgecrest	United States	7.1	5
2016 Kaikōura	New Zealand	7.8	3
2016 Kumamoto	Japan	7.0	4
2010 Darfield	New Zealand	7.1	2
2002 Denali	United States	7.9	1
2000 Tottori	Japan	6.6	1
1999 Izmit	Türkiye	7.4	4
1992 Landers	United States	7.3	1

Table S2: Fault-parallel displacement data for Fig. 2. Distance refers to the shortest distance between the station to the fault rupture. Geodetic data are from (57–65)

Station	Lon	Lat	Distance (km)	Static U_s (m)	Dynamic U_d (m)	Geodetic (m)
701	130.9962	32.8707	1.4	-0.95	-1.65	-0.95 ± 0.20
011	131.0014	32.8582	2.6	-0.70	-1.43	-0.72 ± 0.10
048	130.9031	32.8349	0.5	1.40	1.63	1.45 ± 0.13
H16	130.8199	32.7967	2.0	1.04	1.15	1.14 ± 0.20
ARC	29.3610	40.8240	14.3	0.40	0.68	0.43 ± 0.30
YPT	29.7599	40.7641	4.4	2.06	2.35	1.90 ± 0.60
IZT	29.9600	40.7900	7.4	1.06	1.10	1.20 ± 0.60
SKR	30.3840	40.7370	1.8	2.90	2.05	1.60 ± 0.60
CMBL	174.2138	-41.7490	7.8	2.40	3.66	2.40 ± 0.10
WDFS	174.1384	-41.8274	6.2	4.20	4.96	4.60 ± 1.40
KEKS	173.9800	-41.9600	2.8	7.10	7.48	6.20 ± 1.40
4612	36.4819	38.0240	2.6	-0.33	-0.71	-0.29 ± 0.10
EKZ1	37.1873	38.0582	3.4	-4.44	-4.71	-4.44 ± 0.10
141	36.2197	36.3726	5.2	-1.05	-1.38	-1.08 ± 0.19
142	36.3661	36.4980	1.5	1.42	1.50	1.50 ± 0.19
139	36.4144	36.5838	2.4	1.52	1.75	1.60 ± 0.19
145	36.4064	36.6454	0.5	1.35	1.65	1.25 ± 0.19
137	36.4885	36.6929	2.6	1.82	2.12	1.78 ± 0.19
144	36.4857	36.7569	0.6	-2.04	-2.83	-1.95 ± 0.19
138	36.5112	36.8026	0.7	-2.00	-2.51	-1.92 ± 0.19
143	36.5571	36.8489	1.6	1.95	2.03	2.03 ± 0.19
718	36.6266	37.0078	1.2	2.40	2.68	2.32 ± 0.19
708	36.6484	37.0993	0.5	2.20	2.52	2.08 ± 0.19
712	36.7328	37.1840	1.2	1.83	1.90	1.72 ± 0.19

Table S3: Friction parameters for TPV32 (The Problem, Version 32) (71)

Parameter	Value
Static friction, μ_s	0.580
Dynamic friction, μ_d	0.450
Dynamic friction, $\mu_{d-pulse}$	0.400
Weakening distance, d_c	0.18 m
Cohesion at free surface, C_0	1.02 MPa

Table S4: Continuous 1D velocity structure of TPV32 (The Problem, Version 32) (71). In-between the listed depths, all properties are linearly interpolated to create a piecewise continuous structure.

Depth (m)	V_P (m/s)	V_S (m/s)	Density ρ (kg/m ³)
0	2200	1050	2200
500	3000	1400	2450
1000	3600	1950	2550
1600	4400	2500	2600
2400	4800	2800	2600
3600	5250	3100	2620
5000	5500	3250	2650
9000	5750	3450	2720
11000	6100	3600	2750
15000	6300	3700	2900
> 15000	6300	3700	2900