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2021

Mallick, R., Meltzner, A. J., Tsang, L. L. H., Lindsey, E. O., Feng, L. & Hill, E. M. (2021). Long-lived shallow slow-slip events on the Sunda megathrust. Nature Geoscience, 14(5), 327-333. https://dx.doi.org/10.1038/s41561-021-00727-y

https://hdl.handle.net/10356/152990

https://doi.org/10.1038/s41561-021-00727-y

Long-lived shallow slow-slip events on the Sunda megathrust

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Abstract: During most of the time between large earthquakes at tectonic plate boundaries, surface displacement timeseries are generally observed to be linear. This linear trend is interpreted to result from steady stress accumulation at frictionally locked asperities on the fault interface. However, due to the short geodetic record it is still unknown whether all interseismic periods show similar rates, and whether frictionally locked asperities remain stationary. Here we show that two consecutive interseismic periods at Simeulue Island, Indonesia experienced significantly different displacement rates, which cannot be explained by a sudden reorganization of locked and unlocked regions. Rather, these observations necessitate the occurrence of a 32-year slow-slip event on a shallow, frictionally stable area of the megathrust. We develop a self-consistent numerical model of such events driven by pore-fluid migration during the earthquake cycle. The resulting slow-slip events appear as abrupt velocity changes in geodetic timeseries. Due to their long-lived nature, we may be missing or mismodelling these transient phenomena in a number of settings globally; we highlight one such ongoing example at Enggano Island, Indonesia. We provide a method for detecting these slow-slip events which will enable a substantial revision to the earthquake and tsunami hazard and risk for populations living close to these faults.

Tectonic slow-slip phenomena span a wide range of slip rates, from *mm* yr^{-1} to *mm* s^{-1}, and represent a mode of tectonic fault slip behavior that can be detected by geodetic instruments but do not release the seismic radiation typically associated with earthquakes. While post-earthquake accelerated fault creep (afterslip), earthquakes deficient in high frequency radiation (slow earthquakes), and spontaneous slow-slip events (SSEs) are all under the purview of slow-slip phenomena, in this study we are concerned only with the kinematics and dynamics of SSEs. It is unclear whether SSEs are simply earthquakes in slow motion[1] or a separate group of phenomena caused by significantly different underlying mechanisms[2,3]. The most common hypothesis regarding the mechanics of SSEs on faults governed by frictional processes involves large earthquake (instability) nucleation sizes on velocity-weakening fault zones[4–8] in the presence of low effective confining stresses, which means that instabilities can nucleate but not run away to seismic slip rates. Such low confining stresses generally result from elevated pore pressures[3,9–11]. Additional SSE mechanisms include a switch from velocity-weakening behavior at low velocities to velocity-strengthening behavior at higher velocities[12,13], and the fluid-driven interplay of frictional and ductile processes at depths where rocks become hot enough that flow-like behavior is activated[14–17].

Here we show that a decadal scale SSE nucleated on a shallow velocity-strengthening section of the Sunda megathrust. Velocity-strengthening faults are generally thought to be stable, and therefore unable to nucleate frictional instabilities like earthquakes and SSEs[18,19]. The observations and inferences we present challenge the typical model of SSE nucleation and therefore necessitate a different mechanism to explain its kinematics.

Transient deformation in the Simeulue coral record

In this study, we focus on the deformation history of Simeulue Island in the time period 1738-1861[20,21] (Figure 1). This is thought to be an inter-seismic period for this section of the Sunda megathrust leading up to a plate boundary rupture of M_w~8.5 in 1861[20,22]. The magnitude and spatial pattern of the coseismic deformation in the 1861 event bears remarkable similarity to the 2005 M_w 8.6 Nias-Simeulue megathrust earthquake[20,22] indicating that fault zone properties have remained largely static for at least two earthquake cycles. This implies that it is reasonable to use inferences of frictional properties from the better documented earthquake cycle, the 2005 earthquake[23,24] (Figure 1a), to study fault zone mechanics from the 18^{th}-19^{th} centuries.

The shallow marine corals on Simeulue record the combined effects of long-term sea-level and land-height change. These corals recorded a linear subsidence at a rate of 1-2 mm/yr from 1738 to 1829 (Figure 1b, 2a). Around 1829 (±3 years), an abrupt velocity change occurred at each of the coral sites (Extended Data Figure 1), causing significantly greater subsidence rates of up to 10 mm/yr (Figure 1b,2a), which persisted until the great earthquake in 1861. Previous studies have shown that these velocity changes are tectonic in origin, and interpreted them as spontaneous variations in the spatial extent of frictionally locked asperities and stably creeping regions of the Sunda megathrust[20,21]. Specifically, the locked section of the megathrust was thought to have expanded down-dip to a depth of ~55 km[21]. While variations in dimensions of locked asperities have been interpreted to occur on megathrusts in relation to stress transfer from nearby earthquakes[25,26], there is no evidence to suggest moderate/great earthquakes occurred within ~500 km of Simeulue around the time of the abrupt velocity change in the coral records[20,22]. Additionally, frictional locking is not expected to persist beyond the brittle-plastic transition, which is thermally controlled in subduction zones, and is estimated to be around 30 km depth for this location[27–29]. In light of these inconsistencies, we reevaluate the observations and inferences from the paleogeodetic record for Simeulue using a physics-based inverse method to relate fault slip rates to subsidence rates, and test the hypothesis: can steady frictional locking/sliding on a megathrust explain the observed vertical deformation history?

Steady and unsteady inter-seismic behavior

We develop an inverse method in which we solve for the locked ($v = 0$) and unlocked ($v > 0$) distribution of fault segments during each time period separately, for steady frictional behavior on a megathrust driven by the stresses arising from subducting an elastic slab into the mantle (in plane strain - see Methods). In our formulation, fault slip rates in the unlocked segments are governed by the stressing rates afforded by the nearby locked segments[30,31] i.e. we seek a simply connected locked domain and the surrounding region is treated as a crack driven at constant stress (Extended Data Figure 2a); this is inherently different from typical over-parameterized inter-seismic coupling inversions which solve directly for slip rates and are forced to use slip rate regularization to stabilize the inverse problem. We pose the inverse problem in terms of the misfit between the observed coral subsidence rates and the predicted rates as a function of the geometric parameters that describe the locked and unlocked domains. This is a non-linear problem for which we employ a Markov Chain Monte Carlo sampling scheme (see Methods for details).

From the inferred fault segment locking/unlocking distributions we show that under the assumption of piece-wise steady-state inter-seismic processes, a down-dip migration of the inter-seismically locked zone is necessary to fit the 1829-1861 coral data (Extended Data Figure 2c,3); this feature has been noted by previous studies[20,21]. Additionally, the previously locked shallow megathrust needs to unlock and remain that way, in order to improve the fit to the 1829-1861 subsidence rates (Extended Data Figure 2b). However, this configuration is physically implausible for three reasons:

(1) The transition from down-dip frictional locking to creep at the full plate subduction velocity would need to occur over infinitesimally narrow widths on the megathrust (Extended Data Figure 2d) in order to fit the 1829-1861 vertical velocities. This would create nearly infinite stressing rates which cannot be supported by the elastic medium.

(2) If the down-dip extent of the megathrust were to relock and remain in that configuration from 1829 to 1861, it needed to have completed within 6 years (uncertainties in the timing of the velocity change) (Figure 1b). This implies sustained creep deceleration of at least ~10mm yr^{-2} on the megathrust at depths of ~30-50 km, followed by ~3 decades of elastic strain accumulation within a domain which is at temperatures greater than 500ºC (Extended Data Figure 2C)[27,29]. We are currently unaware of any mechanism that can cause such a process spontaneously on a plate boundary fault or shear zone under these conditions; rock deformation at these temperatures is dominated by either viscous or plastic creep, not stick-slip.

(3) Earthquake cycle simulations show that unlocking of previously locked fault segments, as we infer for the shallow megathrust from our piece-wise steady-state inverse models (Extended Data Figure 2b), does not occur abruptly. Gradual unlocking of frictional asperities is seen in numerical models when creep from the fault segments surrounding the asperity erode its boundaries[32]. However, this process leads to nonlinear displacement timeseries[32–34] which is again inconsistent with our observations.

To explain the abrupt change in vertical velocities in the observational record, our preferred model is one where the locking configuration from 1738 to 1829 persisted on the fault, while a multi-decadal SSE spontaneously nucleated and was sustained on the shallow megathrust from 1829 to 1861 (Figure 2, Extended Data Figure 3-4). We model this SSE as uniform velocity fault creep over a simply connected domain (Methods). We infer the plausible range of transient fault slip rates to be 0.7-2.5 times the long-term trench-normal convergence rate and the SSE released 0.8-3.0 m of slip (Figure 2d). If we assume that this deformation occurred uniformly over the width of our transect (100 km), we predict a moment release equivalent to an event of M_w 7.2-7.9 (Figure 2d). This event, along with the Banyak Islands SSE[35], SSEs in the Mexican subduction zone[36], Alaskan SSEs[37] and the pre-Tohoku SSE[3,33,34], represent some of the largest recorded SSEs. As a result of the low slip rates, such events can be difficult to detect in geodetic time series, since they may not be fast enough to cause a reversal in motion or show the obvious nonlinear patterns in the time series typically expected for SSEs. However, if we apply mechanical constraints on kinematic models we can test if the observations are consistent with steady plate motion or if they are more compatible with a long-lived transient slip event. These methods can also be applied to estimate a more realistic seismic and tsunami hazard for subduction zones where near-trench model resolution is limited[31]. A similar method was used to detect slow, sustained creep migration along the

Cascadia megathrust[38]. Such an approach has the potential to lower the threshold at which we can detect transient slip behavior on a megathrust.

SSEs on velocity-strengthening faults

Comparing the region inferred to slip in the 1829-1861 SSE with coseismic and postseismic slip from the 2005 M_w 8.6 Nias earthquake (Figure 2c), it appears that the 1829-1861 SSE occurred in a velocity-strengthening region[23]. Rate-and-state friction modeling of the postseismic slip for the 2005 earthquake suggests that the shallow megathrust is likely under low effective confining pressures and exhibits weakly velocity-strengthening behavior[24].

We constructed a rate-and-state frictional model (Figure 3) which shows that the observations of ~100 years of quiescence of shallow fault motion (1738-1829) can be explained by the response of a velocity-strengthening fault to a sudden fluid expulsion event early in the earthquake cycle (Figure 3a, Extended Data Figure 5). As faults can be regulatory agents for the transfer of fluids (fault-valve behavior)[39,40], the release of fluids due to the breakage of permeability seals, possibly caused by the propagation of a past earthquake into this section of the megathrust[41–43] (at some time before 1738) or by strong shaking caused by passing earthquake waves[44], abruptly increases the confining pressures on the fault segment and causes it to become clamped (Figure 3a). We model this as an instantaneous drop in pore fluid pressure (ΔP_F = -2 MPa) with a resulting normal stress increase. This is followed by a period of strain accumulation on the partially drained fault for up to centuries (Figure 3b, 4). As the stresses and the stressing rates on the clamped fault increase over the following ~100 years, the fault is perturbed away from steady-state (due to elastic interaction from the creep front migrating inwards) and nucleates a transient slip event with slip rates comparable to the plate velocity[40] (Figure 3). Assuming normal stress recovery at a kinematically imposed linear rate as a proxy for slip rate modulated gouge compaction, which raises pore-pressure[41,45,46] (see Methods and Supplementary Section S1), the fault which had just initiated accelerated creep is further destabilized as it returns to its undrained state (Figure 4). The resulting SSE continues for decades with near-constant slip rates comparable to those observed between 1829-1861 at Simeulue (Figure 4).

SSEs are typically modelled in the rate-and-state frictional framework as spontaneously nucleating instabilities on velocity-weakening faults[4–6,8]. The instability in these models is limited to an aseismic slip rate by a host of processes which ensure that with increasing slip, the elastic energy release exceeds frictional strength loss[4–6,8]. Other models proposed to explain SSE, not dependent on velocity-weakening friction, are typically applicable at depths hot enough that the rheology of the rocks are brittle-ductile[15–17] and hence are not relevant for our observations. Alternate models of SSE in the brittle domain include nucleation on a fault with velocity-weakening friction at low velocities and a switch to velocity-strengthening friction at increased velocities[12,13]. While this model is plausible, such behavior is not ubiquitous – it has mainly been seen in halite and certain types of serpentinite gouges[47]. The shallow velocity-strengthening section of the Sunda megathrust may have localized within layer silicates (within the subducting sedimentary package)[48]. The frictional properties of these rocks have not been tested at the slip rates relevant to this study; however, they have been noted to be purely velocity-strengthening over the ranges (0.1-10 μm/s) testable with experimental setups[47,49]. In this study, we provide a new mechanism to explain SSEs, by destabilizing a velocity-strengthening fault with normal stress variations likely related to pore fluid migration from the subducting sedimentary package into fault gouge (Figure 4). Recent theoretical work on this topic shows that the full incorporation of poroelastic effects into frictional behavior on faults indeed leads to mildly unstable behavior on velocity-strengthening faults, which may explain the mechanics of SSEs[50].

The results of our study force us to rethink our understanding of the nucleation and propagation of SSEs in a frictional framework. Given that the habitat of long-term SSEs at the down-dip edge of subduction zones is a region of high fluid pressure originating from dehydration reactions releasing fluids from sediments which are sealed by silica precipitation near the mantle wedge corner[9,11,41,51,52], it seems plausible that earthquake-cycle driven drainage and restoration of pore fluids along the fault may control the occurrence of megathrust SSEs (Figure 4) without any need to invoke complicated competing mechanisms to nucleate and then stabilize frictional instabilities on velocity-weakening faults.

A test for the applicability of this mechanism is to see if regions hosting long-term SSEs overlap with afterslip following large earthquakes, since afterslip is a typical sign of velocity-strengthening behavior. Sections of the megathrust in Nankai[53,54] as well as in Ecuador[55,56] appear to demonstrate this overlap of afterslip and SSE, thereby adding observational credence to our hypothesis. A different way to look for observations of this mechanism at play in the modern instrumental record is to compare locations and timing records of seismicity within the forearc and/or the subducting oceanic crust with megathrust creep rates. Compared to the SSE period, we expect a cessation of plate boundary creep, due to expulsion of pore-fluids, to preclude an increase in off-fault microseismicity (caused by released fluids travelling through the host rock). In other words, we expect an anti-correlation between off-fault seismicity and megathrust creep rates.

Implications for megathrust earthquakes

SSEs occurring on shallow megathrusts load neighboring seismogenic sections of the fault, bring them closer to failure, and pose a major earthquake and tsunami hazard[10]. There is precedent, in the case of the Sunda megathrust, for a close temporal relationship between multi-year SSEs and great plate boundary earthquakes. The 1829-1861 SSE discussed in this study was followed by a M_w ~8.4 earthquake[20], and the Andaman segment (further north) experienced a SSE from 2001-2004 along the shallow megathrust which was followed by the 2004 M_w 9.2 Sumatra-Andaman earthquake[57]. Creeping faults by themselves may not present a significant hazard, but their interactions with neighboring segments make them unexpectedly dangerous[58].

Looking forward, a potential area of concern is the southernmost section of the Sunda megathrust, near Enggano Island. There, we interpret the GPS data as indicative of a multi-year SSE along the shallow megathrust while the seismogenic domain is locked and accumulating strain (Figure 5). This section of the megathrust was previously assumed not to accumulate any strain due to the observation of negligible trench-normal motion in the horizontal velocities derived from a continuous GPS site[59]. However, once we account for the vertical velocities, which shows rapid subsidence (Figure 5B), we find that such a signal is incompatible with steady aseismic creep on the megathrust. Instead, we show that frictional locking of the seismogenic zone of the megathrust in concert with a long-lived shallow SSE may be canceling out landward motion while enhancing the subsidence of the island (Figure 5C). With an improved understanding of the interplay between the locked seismogenic zones, shallow SSEs and the associated time series, we conclude that the hazard to the communities living near this section of the megathrust is potentially higher than what was previously thought, and models of risk and mitigation strategies need updating.

Figures captions

Figure 1. Tectonic setting of the study (inset map). (**a**) The 2005 M_w 8.6 Nias-Simeulue Earthquake ruptured a portion of the Sunda megathrust; coseismic slip - green contours (2 m intervals), post-earthquake creep – yellow contours (0.5 m contours) [23]. The three coral sites we use in this study are labeled as LBJ, SMB and UTG. We show computed velocities from 1738-1829 (grey) and 1829-1861 (red). The site PBK recorded no velocity change before or after 1829. The black line represents the cross-section we use in the 2-d approximation in Figure 2. (**b**) Timeseries of the coral growth at three sites (1738-1861), and the estimated timing of the velocity change (1829±3). Only the filled circles (the highest level of growth prior to each diedown)[20] are considered in the velocity estimation (model predictions – thick line; 95% confidence interval – thin envelope).

Figure 2. Observations and modeling results for the 18th-19th century coral record in Simeulue projected along the cross-section shown in Figure 1A. (**a**) Subsidence rates for 1738-1829 (gray) and 1829-1861 (red). We assume the velocities from each epoch collectively show the average response of southern Simeulue Island to tectonic changes (filled circles). The individual site vertical velocities are plotted with error bars, while the model predictions are shown as polygons (67% confidence level, with darker colors showing envelopes closer to the median). (**b**) Estimated slip rate for 1738-1829 is shown as a gray polygon. We model the slip rate for 1829-1861 as an SSE (red) superimposed on the existing locking from 1738-1829. (**c**) We compare the probability distribution of the locked region (gray) and SSE location (red) with qualitative estimates of velocity-weakening (green – coseismic) and velocity-strengthening (yellow – afterslip) regions inferred from the 2005 Mw 8.6 Nias earthquake[23], along the transect shown in Figure 1a. (**d**) Probability distributions of slip rate, slip of the shallow SSE (black) and the equivalent moment release (orange) assuming a shear modulus of 30 GPa and along-strike width of 100 km.

Figure 3. Numerical model of an SSE on a velocity-strengthening fault. (**a**) Temporal evolution of fault velocity leading up to and during a shallow SSE (colors show normalized slip velocity; the vertical axis shows normalized distance along the fault and the horizontal axis is time). Model parameters used in the simulation (values of frictional parameters a-b in blue and effective confining stress (MPa) in black) are shown on the left. Fluid expulsion driven clamping is prescribed at t = 50 yrs. This is followed by ~150 yr long strain accumulation. A SSE nucleates spontaneously and continues to slip for ~25 years which corresponds to the period over which fluid pressures are restored

(inset figure). (**b**) Representative slip rate profiles (horizontal axis - normalized slip velocity) drawn for the pre-SSE at t = 150 yrs (gray line) and SSE at t = 210 yrs (red line).

Figure 4. Illustration of the pore-fluid driven SSE model presented in this paper. (**a**) The fractured fault gouge (governed by velocity-strengthening friction) is sealed and compacted by slow shear[41,45,46] during an SSE, and leads to an increase in pore fluid pressure (P_F) on the megathrust (decrease in effective normal stress; $\overline{\sigma} = \sigma - P_F$). Slip and $\overline{\sigma}$ are shown in (**b**) and (**c**) respectively. Following the SSE period, we show two possibilities: (1) an earthquake nucleates elsewhere and breaks permeability seals as it propagates into the shallow megathrust; this process drains the fault, increasing $\overline{\sigma}$ (bold line), (2) the SSE terminates and the unlocked section of the fault resumes its steady-state creep, with negligible change in $\overline{\sigma}$ (dashed line).

Figure 5. A possible ongoing long-lived SSE near Enggano Island. (**a**) The horizontal velocities from a campaign site (engg 1992-2001) and a continuous site (MLKN 2005-2015) show a marked change in amplitude and direction, which we infer as the joint effect of frictional locking and an SSE. In the inset we show a schematic cross-section of a locked seismogenic zone (cyan) and a shallow SSE (red) and their respective velocity predictions at Enggano. (**b**) Displacement timeseries for MLKN with coseismic steps in 2007 and 2012 removed (blue – trench-normal motion; black – vertical motion). (**c**) Schematic to show how the surface displacement timeseries from the locked and SSE region interact together to produce the observations at MLKN.

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Acknowledgements

This research was supported by the National Research Foundation Singapore under a Singapore NRF Investigatorship to EMH (Proposal ID NRF-NRFI05-2019-0009), Singapore NRF Fellowship to AJM (NRF-NRFF11-2019-0008), the Earth Observatory of Singapore (EOS), the National Research Foundation of Singapore, and the Singapore Ministry of Education under the Research Centers of Excellence initiative. Data for the Enggano GPS station MLKN were taken from the SuGAr network maintained by EOS and the Indonesian Institute of Sciences (LIPI). This is EOS contribution number 335.

Author contributions

RM, AJM, LLHT and EMH designed the study. RM and EOL developed the inverse method. RM conducted the data analysis and developed the numerical models for the study. LF processed the GPS data and provided the time series for MLKN. All authors jointly wrote the paper.

Competing interests

The authors declare no competing interests.

Data availability statement

The coral data used in this paper are from Meltzner et al. (2015) (*https://doi.org/10.1016/j.quascirev.2015.06.003*), also available at *https://doi.org/10.21979/N9/5QCLZX*. The daily RINEX files for the GNSS station MLKN are available for public download at *ftp://eos.ntu.edu.sg/SugarData*. The processed timeseries is provided at *https://doi.org/10.21979/N9/LMK36Z*. Topography and bathymetry plotted in Figure 1 and 5 are from the ETOPO1 dataset available at *https://doi.org/10.7289/V5C8276M*. The figures in this paper were made using MATLAB and Generic Mapping Tools[60].

Code availability statement

All computations in this study were carried out using MATLAB; code is available at *https://researchdata.ntu.edu.sg/dataverse/longlivedsse/*.

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Methods

Coral timeseries analysis

The coral timeseries for sites LBJ, UTG, SLR and SMB are taken from published data[20]. The measurement uncertainty at each point used in the velocity extraction is 2 cm. This uncertainty estimate comes from an observation that the largest measurement bias between two corals from the same site was 8 cm[20]; we interpret that as the 99% confidence level extrema (3σ).

It is obvious that a simple linear least-squares fit cannot explain the timeseries, and so we explore 3 different hypotheses: (1) a quadratic least-squares fit, invoking a continuous acceleration, (2) piece-wise linear fits with the timing of the change in velocities (T_{change}) being common to all the sites, (3) piece-wise linear fits treating each site independently.

We then assess the error improvement using the F-test. For the piece-wise linear fits, we add an equality constraint such that the fitted function is continuous across T_{change}. We estimate the mean and covariance of the fitted parameters using simple matrix inversion since the addition of equality constraints to the linear inverse problem can be thought of as a typical Least Squares problem with additional Lagrangian multipliers, which means the posterior distribution is still Gaussian. For a given dataset d, design matrix G (a function of T_{change}) and weighting function W, we estimate the distribution of the model parameters $m \sim N(\hat{m}, C_m)$ subject to an equality constraint $Am = b$ as,

$$\hat{m} = (G^TWG)^{-1}[G^Td - A^T(A(G^TWG)^{-1}A^T)^{-1}(A(G^TWG)^{-1}G^Td - b)] \tag{1}$$

$$C_m = \sigma^2((G^TWG)^{-1} - (G^TWG)^{-1}A^T(A(G^TWG)^{-1}A^T)^{-1}A(G^TWG)^{-1}) \tag{2}$$

The uncertainties in m associated with T_{change} are non-Gaussian but are an order of magnitude lower than C_m, which is why we report only C_m. σ^2 is the least-squares misfit when we use $\hat{m}$ to fit the data.

The reduced-χ^2 from the different hypotheses is shown in Extended Data Figure 1b. From the F-test, we conclude that piece-wise linear fits for a common $T_{change} = 1829$ across all sites is the most likely hypothesis. We do not use the site SLR in subsequent analysis since the uncertainties in the velocities for the site are too large for the velocities to be meaningful.

SSE on a velocity-strengthening fault with normal stress perturbations

We can describe the variation of the friction coefficient μ as a function of velocity, v and a state parameter, θ using rate-and-state dependent friction[61,62].

$$\mu = \mu_0 + a\log\frac{v}{v_0} + b\log\frac{v_0\theta}{L} \tag{3}$$

μ_0, v_0 are reference values of the friction coefficient and velocity respectively, while L is the critical slip distance over which a population of frictional contacts evolve. We use the ageing law to describe state evolution[62]

$$\frac{d\theta}{dt} = 1 - \frac{v\theta}{L} \tag{4}$$

For a 2-d plane strain setup, at any given time t consider the force balance on any fault segment of length $d\zeta$, in the domain Ω, at a location ζ in the down-dip direction. There are 2 directions along which we resolve the action of tectonic driving forces: (1) in the plane of shear, $\hat{d}$, and (2) in the direction of the normal vector to the fault surface, $\hat{n}$. First, we consider the force balance in the fault slip direction:

$$\dot{\tau}^{\infty}t + \int_{\Omega} K_{\tau}\,(\zeta,\Omega)\delta(\zeta,t)d\zeta = \mu(v,\theta)\bar{\sigma} + \frac{G}{2v_s}v \tag{5}$$

Here $\dot{\tau}^{\infty}$ is the far field loading rate on the fault from the subducting plate model[63] with the trench-normal convergence rate, 40 mm/yr, taken from a regional plate motion model[64]. The integral over the fault domain Ω defines the elastic interaction kernel from the Green's function tensor K_{τ} and slip on the fault δ. The right hand side contains the frictional stress (with uniform $\bar{\sigma}$ = 60 MPa) and the radiation damping approximation (G, v_s are the shear modulus and shear wave speed of the elastic medium respectively) at seismic slip rates[5,65,66].
In the boundary element formulation, we can approximate the elastic interaction integral by a simple matrix multiplication for finite dislocations[67]. The time derivative of Equation 5 gives us the momentum balance in 2 directions:

$$\begin{aligned} &(\hat{d}) \qquad \dot{\tau}^{\infty} + K_{\tau}v(\zeta,t) = \dot{\mu}\bar{\sigma} + \mu\dot{\bar{\sigma}} + \frac{G}{2v_s}\dot{v} \\ &(\hat{n}) \qquad \dot{\bar{\sigma}} \rightarrow \text{ kinematically imposed} \end{aligned} \tag{6}$$

We use the far-field stressing rate in the dip direction, and kinematically impose the stressing rates in the $\hat{n}$ direction to model pore fluid migration. The overhead dots represent the time derivatives of the appropriate quantities.

We can numerically integrate the coupled ODEs in Equation 6 using a Runge-Kutta 4th order method, subject to an initial condition. The initial value is provided by means of a velocity step derived from Equation 5 for a sudden change in $\bar{\sigma}$ from t^- to t^+.

$$
\begin{aligned}
v^{+} &= (v^{-})^{\frac{\bar{\sigma}^{-}}{\bar{\sigma}^{+}}}(v_0)^{1-\frac{\bar{\sigma}^{-}}{\bar{\sigma}^{+}}}\exp\left(\frac{\left(\mu_0 + b\log\frac{v_0\theta^{-}}{L}\right)\left(\bar{\sigma}^{-} - \bar{\sigma}^{+}\right)}{a\bar{\sigma}^{+}}\right) \\
\theta^{+} &= \theta^{-}
\end{aligned}
\tag{7}
$$

We choose a value of $\bar{\sigma}^{-} - \bar{\sigma}^{+} = -2\mathrm{MPa}$ motivated by the magnitude of the stress drop necessary to generate a detectable transient[40] as well as by past studies that have proposed the MPa-scale rapid changes (drop) in pore-pressure that is expected during or soon after great earthquakes[43].

We assume that $\dot{\bar{\sigma}}$ is 0 during the strain accumulation phase and is a linearly decreasing function of time once the SSE spontaneously nucleates (t_{SSE}), such that the integrated normal stress change over the SSE duration ΔT_{SSE} equals the initial normal stress perturbation $\left(\bar{\sigma}^{-} - \bar{\sigma}^{+}\right) = \Delta p_F$ (Figure 3a). We add the linear taper on $\dot{\bar{\sigma}}$ to create a simple model of slip rate dependent gouge compaction which causes pore fluid pressures to rise, thereby lowering the effective confining stress ($\bar{\sigma}$) to its initial value before being perturbed $(\bar{\sigma}^{-})$ [41,46].

$$
\begin{aligned}
\dot{\bar{\sigma}} &= 0 \qquad (t \in [t^{+}, t_{SSE}]) \\
\dot{\bar{\sigma}} &= \frac{2\Delta p_F}{\Delta T_{SSE}}\left(1 - \frac{t - t_{SSE}}{\Delta T_{SSE}}\right) \qquad (t \in [t_{SSE}, t_{SSE} + \Delta T_{SSE}])
\end{aligned}
\tag{8}
$$

Inverse method to estimate SSE slip rates

We use physical constraints from Equation 6 to develop our inverse method as follows. In the steady inter-seismic period, locked parts of the fault accumulate strain while creeping parts of the fault slip at time invariant slip rates determined by the net shear stress rate at that location (Equation 6), with $\dot{\mu} = \dot{\bar{\sigma}} = 0$. We can write this as:

$$
\begin{aligned}
\dot{\tau}^{\infty} + K_{\tau}v(\zeta, t) &= 0 \qquad \text{steady creep at available } \dot{\tau} \\
\dot{\tau}^{\infty} + K_{\tau}v(\zeta, t) &> 0 \qquad \text{locked}
\end{aligned}
\tag{9}
$$

We segment the fault into 4 regions, using 3 free parameters $\zeta_{up}, \zeta_{down}, \zeta_{free}$ (Extended Data Figure 2a). Region 1 is a creeping zone which extends from the free surface to ζ_{up}; region 2 is a locked zone between ζ_{up} and ζ_{down}; region 3 is a creeping zone between ζ_{down} and ζ_{free}; region 4 creeps at the plate rate below ζ_{free}.

The data for the inverse problem is the vertical velocities we computed using Equations 1-2. Since we use a plane strain approximation and the stations do not exactly fall on a 2-d transect, we calculate weighted average velocities and uncertainties from the 3 sites to represent the vertical motion of Simeulue Island in a representative cross-section (Figure 2).

The slip rate to surface velocity relations are taken from analytical solutions for edge dislocations in plane strain for elastic media[67,68] and a geometric setup applicable to the Sunda megathrust[63,64,69]. The misfit between the data and model-predicted velocities is assumed to follow an unbiased multivariate Gaussian distribution with unknown variance. We sample the posterior distribution of the free parameters using a Markov-Chain-Monte-Carlo method (slice sampling)[70,71]. We use this method to estimate the steady-state slip rates on the Sunda megathrust for 1738-1829 as well as 1829-1861 (Figure 2, Extended Data Figure 3).

For the case where we assume a transient slip event occurs as a quasi-steady process, we parameterize a continuously connected section of fault to have a constant slip rate (within its domain, as well as over the time period 1829-1861). This velocity is superimposed on the jointly estimated 1738-1829 slip rate distribution from the steady-state inversion. We use slice sampling to sample the up-dip and down-dip extent of the transient as well as its slip rate $\left(\zeta_{up}^{trans}, \zeta_{down}^{trans}, V_{trans}\right)$ in addition to the steady-state locking/creeping parameters $\left(\zeta_{up}^{lock}, \zeta_{down}^{lock}, \zeta_{free}^{lock}\right)$ for the 1738-1829 period (Extended Data Figure 4).

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