Dynamic Rupture Scenarios of the Cascadia Megathrust based on Interseismic Locking Models

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Abstract

The Cascadia subduction zone in the Pacific Northwest has well-documented geological records of megathrust earthquakes with the most recent Mw 9 rupture occurring in 1700 A.D. The paleoseismic observations suggest that Southern Cascadia is mature for future earthquakes since the last event. Consequently, it is crucial to investigate the potential rupture scenarios. Various interseismic locking models are developed along Cascadia, including offshore uncertainties and different material assumptions. Although they all share similar moment deficits, whether future earthquakes may rupture the entire margin or be segmented, as found in the paleoseismic records, remains unknown. Accordingly, we aim to investigate: (1) possible rupture segmentation patterns, (2) whether south Cascadia can host margin-wide ruptures, and (3) whether the existing locking models suggest similar future rupture scenarios. We estimate the stress distribution constrained by the locking models from static calculation and discover that they lead to different stress distributions, indicating distinct seismic potentials despite their similar moment deficits. Our dynamic rupture scenarios show that the south can generate both segmented ruptures (> Mw 7.3 - 8.4) and margin-wide ruptures (> Mw 8.6) depending on hypocenter locations. The extent of Schmalzle-based segmented scenarios matches the proposed historical segmented events, and the margin-wide scenarios are well consistent with the coastal subsidence records of 1700 A.D. Therefore, we propose that three high-slip trench-breaching patches are sufficient for reproducing historical subsidence records. Our reasonable dynamic simulations can be applied in future studies for assessing seismic and tsunami hazards, and also serve as a comparison for non-trench-breaching scenarios.

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Key Points:

- We conduct 3D dynamic rupture simulations for the future possible scenarios in Cascadia with constraints from interseismic locking models
- Application of different hypocenter locations reveals rupture segmentation and rupture directivity
- Our dynamic rupture scenarios have reasonably consistent segmentation extents and coastal subsidence patterns with paleoseismic observations

Abstract

The Cascadia subduction zone in the Pacific Northwest has well-documented geological records of megathrust earthquakes with the most recent Mw 9 rupture occurring in 1700 A.D. The paleoseismic observations suggest that Southern Cascadia is mature for future earthquakes since the last event. Consequently, it is crucial to investigate the potential rupture scenarios. Various interseismic locking models are developed along Cascadia, including offshore uncertainties and different material assumptions. Although they all share similar moment deficits, whether future earthquakes may rupture the entire margin or be segmented, as found in the paleoseismic records, remains unknown. Accordingly, we aim to investigate: (1) possible rupture segmentation patterns, (2) whether south Cascadia can host margin-wide ruptures, and (3) whether the existing locking models suggest similar future rupture scenarios. We estimate the stress distribution constrained by the locking models from static calculation and discover that they lead to different stress distributions, indicating distinct seismic potentials despite their similar moment deficits. Our dynamic rupture scenarios show that the south can generate both segmented ruptures (> Mw 7.3 - 8.4) and margin-wide ruptures (> Mw 8.6) depending on hypocenter locations. The extent of Schmalzle-based segmented scenarios matches the proposed historical segmented events, and the margin-wide scenarios are well consistent with the coastal subsidence records of 1700 A.D. Therefore, we propose that three high-slip trench-breaching patches are sufficient for reproducing historical subsidence records. Our reasonable dynamic simulations can be applied in future studies for assessing seismic and tsunami hazards, and also serve as a comparison for non-trench-breaching scenarios.

Plain Language Summary

Earthquakes occur when the shear stresses on a fault overcome the frictional resistance to cause a sudden slip. In subduction zones, the tectonic plates converge and the stresses accumulate at the contact between the plates. As more stresses accumulate on the interface, great earthquakes are possible. Although there are no significant earthquakes (> Mw 8) since 1700 A.D., the Cascadia subduction zone in the Pacific Northwest is known to have historical Mw 9 earthquakes based on geological studies. Interseismic locking models describe the relative motion of the fault. For instance, 1 means fully locked where the two sides do not move against each other, thus accumulating stress. We infer stress distributions from interseismic locking models and conduct 3D dynamic simulations based on the stresses to explore possible future earthquake extents. Our results demonstrate various scenarios, including single-segment (> Mw 7.3 – 8.2), multiple-segments (> Mw 8.2 – 8.4), and full-margin ruptures (> Mw 8.6). Most of these scenarios are consistent with geological records, suggesting our scenarios are reasonable future earthquake estimates.

1. Introduction

The Cascadia subduction zone is known to host great megathrust earthquakes as large as moment magnitude (Mw) 9 (Wang and Tréhu, 2016; Walton et al., 2021). Based on paleoseismic records (Long and Shennan 1998; Kelsey et al. 2005; Goldfinger et al. 2012; Engelhart et al. 2015), the average recurrence interval of these events is about 500 yrs but with large variations. It has been over 322 years since the latest great earthquake, an M ~9 margin-wide rupture in A.D. 1700 accompanied with a large, trans-Pacific tsunami (Atwater and Hemphill-Haley 1997; Goldfinger et al. 2012, 2017; Satake et al., 2003). Modern interseismic geodetic observations indicate accumulation of energy along almost the entire Cascadia margin towards a future earthquake (Flück et al., 1997; Wang et al., 2003; Burgette et al., 2009; McCaffrey et al., 2013; Schmalzle et al. 2014; Pollitz and Evans, 2017; Li et al. 2018; Michel et al. 2019; Lindsey et al. 2021).

One challenge in seismic hazard assessment at Cascadia is estimating the potential of rupture segmentation along the megathrust. There are questions regarding whether past events were predominantly full-margin ruptures or sequences of smaller ruptures that were too closely spaced in time to be resolved by paleoseismic records (Wang et al., 2013; Atwater et al. 2014; Frankel et al. 2015). Along-strike heterogeneities in megathrust and crustal structure are thought to have the potential to cause rupture segmentation in various parts of the margin (Tréhu et al., 2012; Wang and Tréhu, 2016; Watt and Brothers, 2021). Based on the interpretation of offshore turbidity records, megathrust earthquakes occurred more frequently in southern Cascadia, especially south of Cape Blanco (Goldfinger et al., 2017). The average recurrence interval is inferred to increase from around 200 years in the south to around 300 years in the central segment and 400-500 years in the north (Witter et al., 2012; Goldfinger et al. 2017). If the A.D.1700 event was a full-margin rupture as inferred by Satake et al. (2003), then at present the short-recurrence southern segment is statistically expected to be more ready for the next rupture. The first scientific question we address in this study is whether the next large earthquake is more likely a full-margin rupture or to be confined in the south.

Dynamic rupture scenarios based on interseismic locking models can contribute to estimating the magnitude, rupture extent, and potential segmentation of future earthquakes (Yang et al., 2019a; Li and Liu, 2021; Ramos et al., 2021; Yao and Yang, 2022). For instance, Yang et al. (2019a) derived dynamic scenarios for the Costa Rica subduction zone by using interseismic locking models to derive the initial stress of the megathrust prior to the rupture and were able to explain the rupture extent and magnitude of the 2012 Nicoya Mw 7.6 earthquake. Using a similar approach, Ramos et al. (2021) conducted dynamic rupture simulations for Cascadia with the initial stress based on the interseismic locking model of Schmalzle et al. (2014). By nucleating ruptures from a high-stress location either in the south or in the north, they obtained scenarios of margin-wide rupture. Li and Liu (2021) conducted quasi-dynamic numerical simulation of long-term fault behavior in Cascadia. They inferred fault rate-state friction stability from interseismic locking models (Schmalzle et al. 2014; Burgette et al., 2009). They found that whether the rupture was full-margin depended on what locking model was used.

Besides rupture extent and earthquake magnitude, the effect of rupture directivity on ground motion should be further investigated using dynamic rupture simulations. It is well understood that, with a heterogeneous initial stress distribution along the fault, different hypocenter locations can lead to different rupture directivities (Yang et al., 2019b; Yao and Yang, 2022). Even with a similar rupture extent, a different rupture directivity leads to a very different pattern of ground motion intensity (Yao and Yang, 2022). Therefore, the second scientific question we address in this study is how hypocenter location controls rupture directivity to impact ground motion.

To investigate the above questions, we carry out dynamic rupture simulation to obtain self-consistent rupture scenarios. We consider different Cascadia megathrust locking models (Figure 1), namely those by Schmalzle et al. (2014), Li et al. (2018), and Lindsey et al. (2021). Our research aims to derive rupture scenarios originating from South Cascadia. Assuming the same stress accumulation time, we investigate the role of stress distribution and hypocenter location in producing possible segmentation patterns and ground motion patterns. We further compare the rupture scenarios with the proposed segmented paleoearthquakes as well as coseismic subsidence amplitudes.

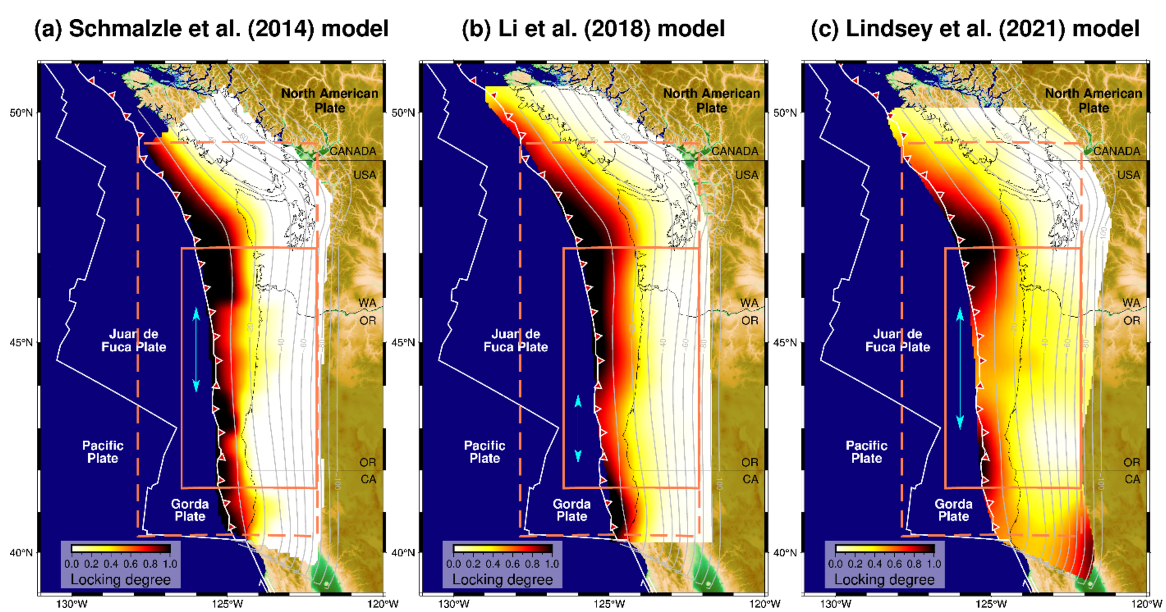

Figure 1. Interseismic locking models for CSZ. (a) Model from Schmalzle et al. (2014). (b) Model from Li et al. (2018). (c) Model from Lindsey et al. (2021). Coral dashed line: our static calculation domain. Coral solid line: our dynamic simulation domain. Cyan arrow: central creeping segments.

2. Interseismic locking models of the Cascadia megathrust

Since solutions for the inversion of geodetic measurements are nonunique, different assumptions are applied in deriving interseismic locking models, governing the smoothness of slip distribution and the degree of locking at the trench (McCaffrey et al., 2013; Schmalzle et al. 2014; Pollitz and Evans, 2017; Li et al. 2018; Michel et al. 2019; Lindsey et al. 2021). Here we summarize the three locking models adopted in this work, all derived by inverting land-based GNSS observations (Figure 1). Although Cascadia does not have a geomorphological trench because of the thick sediment cover, we refer to the deformation front as the "trench" in the following discussion for wording convenience.

Because land-based GNSS measurements cannot resolve the locking state of the shallowest portion of the megathrust which is far offshore, Schmalzle et al. (2014), following McCaffrey et al. (2013), proposed two models of opposite, prescribed near-trench locking states which fit the GNSS data equally well. One model assumes full locking at the trench with the locking degree monotonically decreasing downdip following the Gamma function designed by Wang et al. (2003) (Gamma model). The other model assumes a Gaussian-like locking distribution so that creeping occurs at the trench and full locking occurs farther downdip (Gaussian model). Creeping of the shallowest part of the fault may occur in a transient fashion such as during earthquake afterslip or slow slip events but is unlikely a sustained behavior over the interseismic period (Wang and Dixon, 2004; Wang, 2007). Thus, in this study we only use the Gamma model, referred to as the Schmalzle model hereafter (Figure 1a).

Following the explanation of Wang and Dixon (2004) and Wang (2007), Lindsey et al. (2021) included in their locking model the effect of stress shadowing in which a frictionally unlocked shallow segment of the fault may have little motion because of the neighboring frictionally locked patches immediately downdip. Although stress shadowing is explicitly invoked, the kinematic behavior of the megathrust in this model is similar to that described by the aforementioned Gamma model. The difference in inversion results is caused mainly by assumed inversion parameters that constrain slip deficit distribution. In this study, we use their best-fit locking model, referred to as the Lindsey model (Figure 1c).

The above two locking models assume an elastic Earth, but the real Earth is viscoelastic, and viscoelastic stress relaxation plays an important role not only in postseismic but also interseismic deformation (Wang et al., 2012). To address this effect, Pollitz and Evans (2017) and Li et al. (2018) inverted Cascadia interseismic geodetic data based on analytical solutions and finite element models, respectively. Li et al. (2018) constructed many locking models that fit the geodetic data equally well. Here we only use their "preferred" locking model, referred to as the Li model (Figure 1b).

Because of the lack of near-field, seafloor geodetic constraints, all these models suffer from a high degree of nonuniqueness and thus contain large errors. By using these models to design initial fault stress distribution, we do not intend to construct a "correct" dynamic rupture model. Instead, we use these models to explore how different initial stress distributions may affect the rupture process. As such, these models may be considered as ad-hoc to each other. Improved understanding of the dynamic rupture process will help the design of kinematic rupture models for the purpose of probabilistic seismic hazard analyses and the appraisal of model uncertainties. We think the three models shown in Figure 1 adequately represent the range of assumptions used in constructing Cascadia megathrust locking models by different research groups in terms of Earth rheology, near-trench locking state, and smoothness of slip deficit distribution. Since stress accumulation is mostly determined by the spatial gradient of the locking distribution and the major first-order features of active faulting could be governed by the spatial gradients of stress (Nur, 1978), it is important to ask whether the slip deficit heterogeneities in these locking models can lead to consistent rupture scenarios.

3. Method and model parameter

We use open-source finite-element code PyLith which is developed for dynamic and quasi-static simulations of crustal deformation (Aagaard et al., 2017a). Input parameters for our dynamic simulation include fault geometry, material properties, initial stresses (τ_0), and fault frictional law parameters (Harris et al., 2018).

3.1 Megathrust geometry and mesh

We adopt the Cascadia megathrust model of McCrory et al. (2004), which has a smoothly curved geometry starting from a 5 km depth below sea level and use an exponential curve to approximate the shape of the upper plate near the trench as shown in Figure 2a.

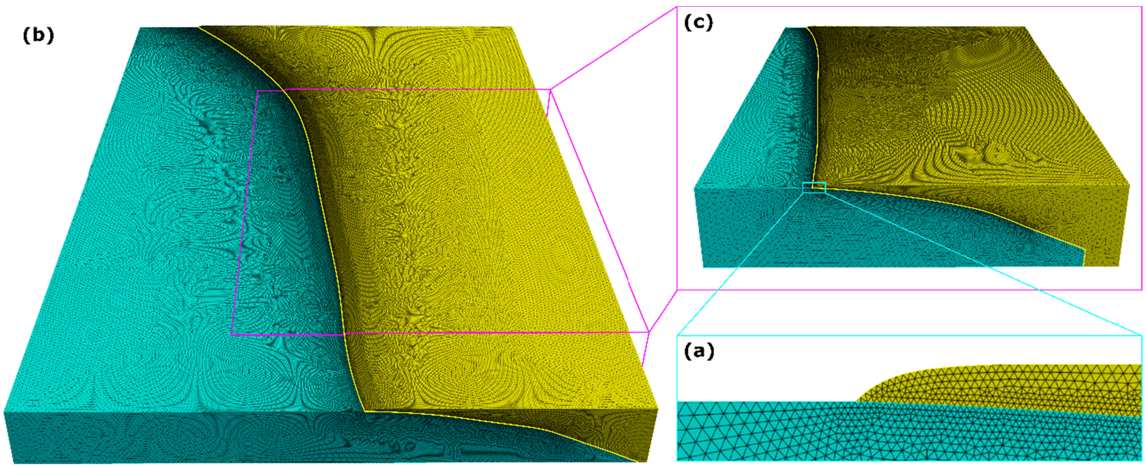

Figure 2. 3D model configuration. (a) Mesh geometry near the trench – continental block edge approximated with an exponential curve away from the trench. (b) Finite element mesh for static calculation with two model units: oceanic block (cyan) and continental block (yellow). (c) Finite element for the dynamic simulations.

We generate two 3D tetrahedral meshes for Cascadia using geometry and mesh generation software CUBIT (Blacker et al., 2016) to accommodate scientific purposes and computational cost. Both meshes each consists of two model units - the oceanic block and the continental block. We use the larger one of the two meshes, extending from 40.5°N to 49°N covering the whole megathrust (Figure 2) to calculate stress distribution from locking models. We apply a coordinate transformation to fix the origin at -129°E, 39°N. This larger mesh extends 970 km, 600 km, and 75 km in the strike, strike-normal, and depth dimensions, respectively (Figure 2b). The element size on the fault is 500 m above 35 km depth for the major locked zone and gradually increases downdip.

For computational efficiency, we use the smaller one of the two meshes, extending from 41.5°N to 47°N, to conduct dynamic rupture simulation in our area of focus. We are focused mainly on the scenarios of rupture initiation in the south and on the effect of the central segment. Geological evidence of ruptures limited to northern Cascadia is elusive (Petersen et al. 2014), suggesting that ruptures breaking the northern segment might eventually develop into margin-wide ruptures. This is consistent with the higher stress accumulation in the north provided by most locking models (Burgette et al. 2009; McCaffrey et al. 2013; Schmalzle et al. 2014; Pollitz and Evans, 2017; Li et al. 2018; Michel et al. 2019; Lindsey et al. 2021) as well as the dynamic simulation results from Ramos et al. (2021). The small mesh covers the entire southern and central Cascadia, extending 600 km, 420 km, and 95 km in the strike, strike-normal, and depth dimensions, respectively (Figure 2c). The element size is 500 m above 50 km depth and gradually increases further downdip.

To minimize potential artefacts due to mesh boundaries, we extend the small mesh for the dynamic simulation to 95 km depth and even deeper than the larger mesh for static calculation by 20 km. In comparison, interseismic locking occurs mostly shallower than 30 km depth (Figure 1) and, to be further explained in sections 3.3 and 4.1, the model-predicted rupture propagation does not extend far beyond this depth because of lack of inferred interseismic stress built-up farther downdip. As will be shown section 4.2, none of our simulations features rupture deeper than 50 km depth.

3.2 Material Properties

Similar to most other dynamic rupture models, we assume an elastic Earth and apply absorbing conditions to all boundaries except the free surface at the top. The material property structure is based on the 3D Community Velocity Model (CVM) of Cascadia (Stephenson et al. 2017) in which the body wave velocities of the oceanic block are approximately 30% higher than the continental block. The density is calculated from p-wave velocity based on the empirical relationship of Brocher (2005). We have tested two different material property structures in order to see how they affect the dynamic rupture process. One model is referred to as the 1-D velocity model, in which the material properties of the continental block are applied to the whole mesh. Another model is referred to as the bi-material velocity model, where material properties of both continental and oceanic blocks are considered (Figure S1). The two structures lead to very similar rupture scenarios. Between the two test models shown in Figure S2, the moment magnitude differs only by 0.01 (Figure S2). Hence, we use the 1-D velocity structure for the rest of our dynamic simulations (Figure 3a).

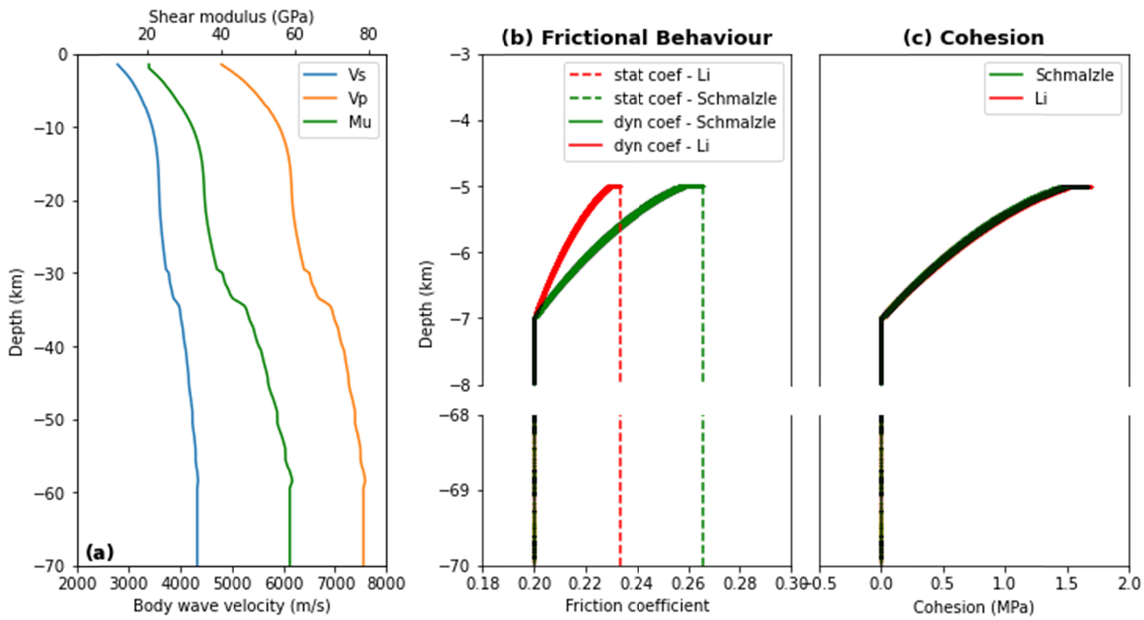

Figure 3. Depth-dependent parameters for dynamic rupture simulations. (a) 1-D velocity model calculated from Stephenson et al. (2017) and shear modulus. (b) Dynamic and static coefficients which remain constant below 7 km depth. (c) Cohesion, remaining constant below 7 km depth. Note that it appears dark because the two curves are overlapping.

3.3 Stress accumulation and initial stress on the megathrust

Following previous studies (Yang et al., 2019b; Ramos et al., 2021), we assume that the slip deficit has been continuously accumulated since the A.D.1700 earthquake. There are uncertainties associated with this assumption because there are no observational constraints on whether medium size earthquakes or significant creep occurred in the seismogenic depth range of the megathrust after 1700 but before the instrumental era. Upon interpreting GNSS velocity variations, Materna et al. (2019) proposed temporal variations in megathrust locking in southernmost Cascadia updip of the ETS zone associated with stress perturbations due to offshore M6+ earthquakes in the incoming oceanic plate. We do not include these complicated temporal variations in our calculation of slip deficit because neither the uniqueness of the GNSS data interpretation nor the physical mechanism of the proposed variations are well understood.

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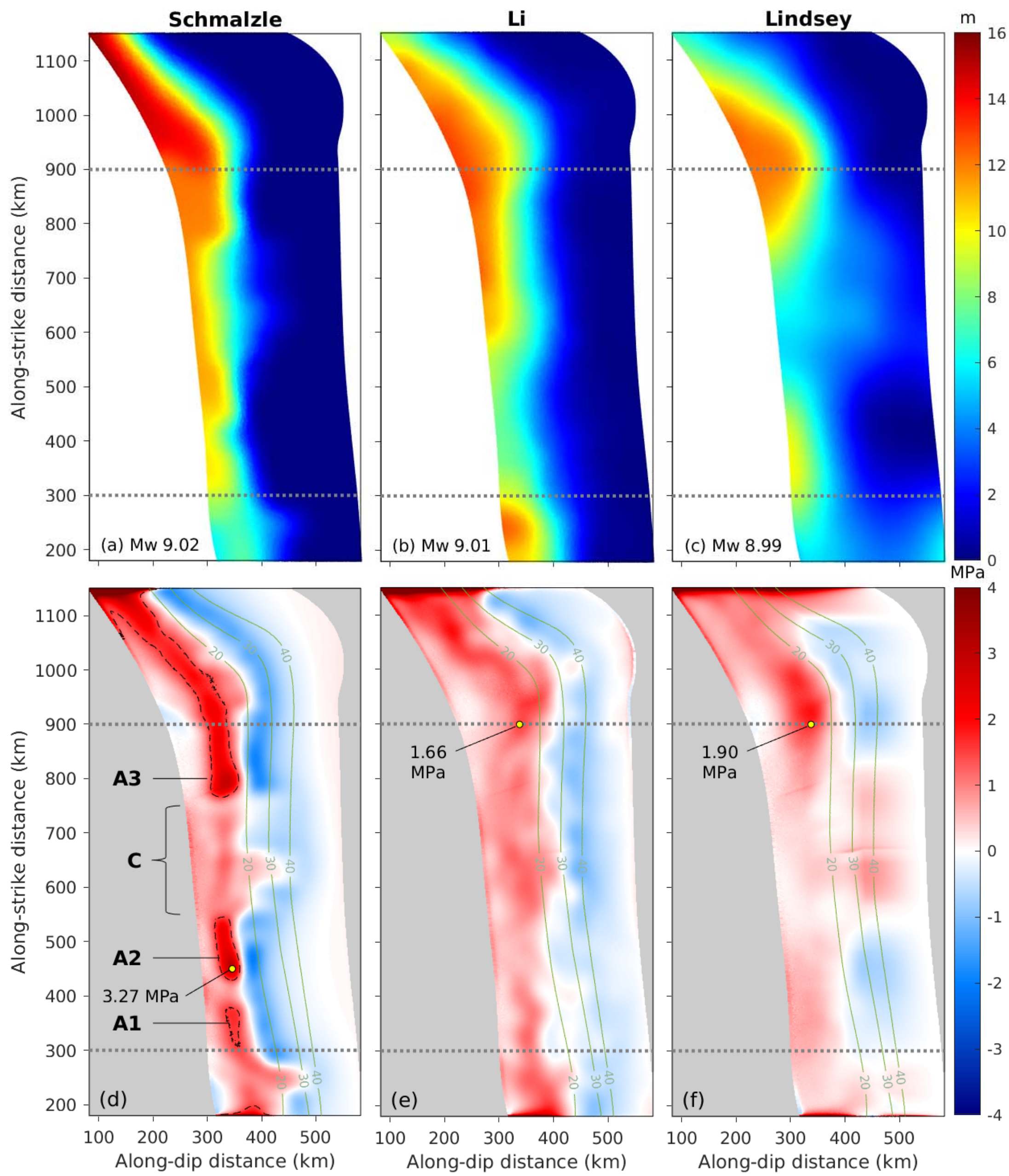

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269 Figure 4. Slip deficit and stress change. (a) – (c): Total slip deficit with a uniform stress
270 accumulation time of 320 years. Dotted lines: the boundaries of the dynamic simulation
271 domain. (d) – (f): Dip component of the stress build-up caused by the slip deficit in (a) – (c).
272 Dotted lines: the boundaries of the dynamic simulation domain. Yellow dots: the point of
273 highest stress change magnitude within the dynamic modelling domain. Dashed lines in (d):
274 1.5 MPa stress contours. A1, A2, and A3 refer to the stress asperities while C marks the
275 creeping segment extent.

With the interseismic locking distribution assumed to be time-independent, the slip deficit at present (Figures 5a – 5c) is simply the product of the subduction rate, slip deficit rate as a fraction of the subduction rate as given by the locking models (i.e., the locking degree in Figure 1), and the time since the A.D. 1700 great earthquake. In an elastic model, the incremental stress associated with the accumulation of this slip deficit can be readily determined from the slip deficit distribution (similar to the determination of static stress drop from coseismic slip distribution) (Figures 5d – 5f). Following Yang et al. (2019b) and Ramos et al. (2021), we assume that this incremental stress solely propels the next megathrust rupture (Figure 4), which implies that the "base level" of the fault stress plays no role, that is, whether the A.D.1700 event feature complete or partial stress drop is unimportant. It also means that the spatial heterogeneity of the fault stress distribution just after that earthquake is unimportant. This assumption is obviously a leap of faith, but it is theoretically consistent with the slip-weakening friction law invoked in our modelling which will be explained in section 3.4, and it makes it operationally possible to derive initial fault stress from interseismic locking models. Note that the incremental stress derived from one of locking models shown in Figures 1c and 5c occurs far deeper than the commonly assumed seismogenic depth limit of around 30 km in some areas along the margin (Figure 4f). To confine seismic rupture within a reasonable depth range, we use a cosine function to taper the fault stress in this model to zero from 35 km to 75 km depth (Figure 4f).

Effective normal stress is the normal stress minus pore fluid pressure. For simplicity, we assumed as a uniform effective normal stress of 50 MPa on the entire megathrust regardless of how the shear stress varies along the fault. This low effective normal stress is based on the notion of very high pore fluid pressurization at depths as inferred for global subduction zones (Saffer and Tobin, 2011). For example, given an average rock density 2500 kg/m^3, an effective normal stress 50 MPa at depth 20 km requires pore fluid pressure about 90% of the lithostatic pressure. There is no reason against other small values such as 30 or 60 MPa, and the lack of depth dependence is for the numerical convenience with little observational support. Nonetheless, the number 50 MPa we use is a typical value over the velocity-weakening region which is often used in earthquake simulation studies (Lapusta and Liu, 2009; Michel et al., 2017; Yang et al., 2019a).

Cascadia is well-known for Episodic Tremor and Slip (ETS) events (Rogers and Dragert, 2003). Gao and Wang (2017) suggest that although the effective normal stress in the ETS region is exceptionally low because of near-lithostatic fluid pressure, the ETS zone is rheological separated from the seismogenic zone and thus is not involved in dynamic rupture. As will be shown in section 4.2, in our models the rupture is arrested before reaching the ETS zone without additional constraints, which is consistent with the notion of Gao and Wang (2017).

Based on findings of high-rate friction experiments (e.g., Di Toro et al., 2011), we set a dynamic friction coefficient of 0.2 (i.e., dynamic stress level of 10 MPa) for the fault below 7 km and assume it to be constant. The southern Cascadia material for the frontal thrust is velocity-weakening while the northern Cascadia material is velocity-strengthening (Stanislowski et al., 2022). For simplicity, we assume that the frontal thrust is neutrally stable by increasing the dynamic coefficient linearly to the static coefficient levels of 0.2656 and 0.2332 for Schmalzle and Li models respectively (Figure 3b and Table 1) from 7 km updip to 5 km depth. The initial stress prior to the dynamic rupture is the sum of the dynamic stress and the interseismic stress accumulation inferred from the locking models.

Table 1 *Model parameters in dynamic rupture simulations*

Fault parameter	Schmalzle model	Li model
Static friction coefficient, f_s (yield strength/σ_n)	0.2656	0.2332
Dynamic friction coefficient, f_d (dynamic stress/σ_n)	0.2	0.2
Effective normal stress, σ_n (MPa)	50	50
Critical weakening distance (m)	0.6, 1	0.6, 1

3.4 Fault frictional law and resolution test

The fault is assumed to be governed by the linear slip-weakening law (Ida, 1972) in which fault shear stress τ_f is given by (Aagaard et al., 2017b),

$$\tau_f = \begin{cases} \tau_c - (\mu_s - (\mu_s - \mu_d)\frac{d}{d_c})\tau_n & d \leq d_c \; and \; \tau_n \leq 0 \\ \tau_c - \mu_d \tau_n & d > d_c \; and \; \tau_n \leq 0 \\ 0 & \tau_n > 0 \end{cases} \tag{1}$$

where μ_s is the static friction coefficient, μ_d is the dynamic friction coefficient, d_c is the slip-weakening distance, τ_n is the effective normal stress, τ_c is the cohesive stress, and d is the slip distance. The frictional resistance (τ_f) decreases linearly with increasing fault slip when $d < d_c$ but stays constant when $d > d_c$. It is should be emphasize that, according to the slip-weakening law, the rupture behavior is controlled by the difference between yield stress and initial stress instead of the absolute stress level.

We set the yield stress, which is the product of static friction coefficient and normal stress, to be uniformly 0.01 MPa above the maximum initial shear stress on the fault within the dynamic modelling domain (Figure 3b and Table 1). For example, the highest interseismic stress accumulation within the region of dynamic modelling in accordance with the Schmalzle slip deficit distribution is 3.27 MPa (Figure 4e) which, with the uniform dynamic stress 10 MPa, gives the highest initial stress 3.27 + 10 MPa = 13.27 MPa. The yield stress of the model based on the Schmalzle slip is thus l3.28 MPa, which translates to a static friction coefficient of 13.18/50 = 0.2656 (Table 1). The assumed homogeneity of the yield stress can be understood as an indication of relatively smooth megathrusts that are conducive to very large earthquakes, and its low amplitude reflects the low fault strength as inferred from heat flow data (Gao and Wang, 2014).

The highly compliant, frontal region of the accretionary prism could significantly impact the rupture scenarios as its inelastic deformation can act as an energy sink (Galvez et al., 2014). We tested the sensitivity of the assumed depth limit to the weak frontal prism. We found that strong free-surface reflections and amplified fault slip would be generated to facilitate trench-breaching rupture if a thinner frontal prism was used, but rupture would be halted if a thicker frontal prism was used (Figure S3). For simplicity, we adopt an average depth range (i.e. 5 – 7 km) for the frontal prism according to the velocity model of Stephenson et al. (2017) (Figure S4). Similar to Ramos et al. (2021), we add cohesion to the segment of the megathrust overlying the assumed frontal prism (Figure 3c) to suppress undesired rupture initiation near the trench.

Based on seismic observations, the critical weakening distance, d_c, has been suggested to be proportional to the local total slip, indicating spatially heterogeneous d_c on faults (Mikumo et al., 2003; Tinti et al. 2005; Fukuyama and Mikumo, 2007). However, because the slip distribution is now known a priori for future earthquakes, and the scaling relationship between slip and d_c has large uncertainties (Guatteri and Spudich, 2000; Chen and Yang, 2020), there is little information about d_c. Ensuring fair comparison among locking models is another challenge in deciding on d_c because the same d_c represents different fracture energy given the different initial stress and yield stress in each model. Dynamic models constructed by Weng and Yang (2018) and Yao and Yang (2020) show that a uniform d_c yields synthetic waveforms that compare with observations very well, and that a heterogeneous slip-scaled d_c does not lead to appreciable improvements. Therefore, we take the simpler approach of assuming a uniform d_c. We recognize the large uncertainties associated with the choice of d_c and test a range of uniform d_c values to see how the results are affected. In section 4.3, we will discuss the results using d_c of 1 m and 0.6 m.

Our models need to meet the resolution requirement. A cohesive zone refers to the fault plane portion behind the crack tip where shear stresses drop from static to dynamic value with a slip less than d_c (Ida 1972). The cohesive zone of in-plane (mode II) ruptures can be estimated by the following equation (Day et al., 2005)

$$\Lambda = \Lambda_0 A^{-1}(v), \qquad \Lambda_0 = \frac{9\pi}{32}\frac{\mu}{1-\nu}\frac{d_c}{\tau_s - \tau_d}, \tag{2}$$

where μ is shear modulus, ν is Poisson's ratio, and τ_s and τ_d are yield stress and dynamic shear stress, respectively. Considering d_c of 0.6 m - 1 m, the static cohesive zone sizes are around 7.5 - 25 km. Given a lower bound for shear wave speed Vs of 3.165 km/s and a rupture speed of 3.1 km/s, $A_{III}^{-1} = (1 - V_r^2/V_S^2)^{1/2} = 0.2$, the dynamic cohesive size can be as small as ~1.5 km. Aagaard et al. (2013) demonstrate that PyLith can resolve cohesive zones around 1.5 times the size of the tetrahedral elements. Therefore, our element size of 500 m on the fault can resolve cohesive zones in our models.

3.5 Rupture initiation

The nucleation zone refers to the area where the rupture begins. In the prescribed nucleation zone, the initial stress has to meet the yield stress to initiate the rupture. To initiate a rupture, we decrease the yield strength inside the designated nucleation zone by decreasing the static friction coefficient within the nucleation zone from the τ_s values shown in Table 1 to $\tau_s^i =$

0.2001. In order to initiate a spontaneous rupture, a nucleation zone needs to exceed a critical size (Yang et al., 2019a),

$$A_1 = \frac{(3\pi)^3}{2^{11}} \frac{\overline{\tau_0 - \tau_d}}{\tau_s - \tau_s^i} \frac{(\tau_s - \tau_d)^2}{(\overline{\tau_0 - \tau_d})^4} \mu^2 d_c^2 . \tag{3}$$

So $\overline{\tau_0 - \tau_d}$ and $(\tau_s - \tau_s^i)$ denote average static stress drop and strength decrease within the nucleation zone respectively. For instance, given an average static stress drop of 1.5 MPa and d_c of 0.6 m for the Schmalzle model, the critical radius of a circular nucleation zone is 13.6 km.

We tested nucleation zone radii of 10 km and 15 km, comparable with those adopted in dynamic modeling studies for the 2011 M9.1 Tohoku earthquake (Duan 2012; Ide and Aochi 2012; Galvez et al. 2014). With different nucleation sizes, the model-predicted rupture scenarios for the same hypocenter locations are similar (Figure S5). To ensure that our rupture scenarios could represent first-order features from the interseismic locking models instead of the interpolation methods, the nucleation zone size has to be comparable to the patch size or node spacing used during inversion (Figure 5), especially for the Li model. Thus, we adopted a larger radius of 15 km. We then conduct simulations on ten along-strike hypocenter locations in southern Cascadia spanning from 41.77° - 44.47° latitude. A range of depths is also tested depending on the positive stress change distribution of each locking model.

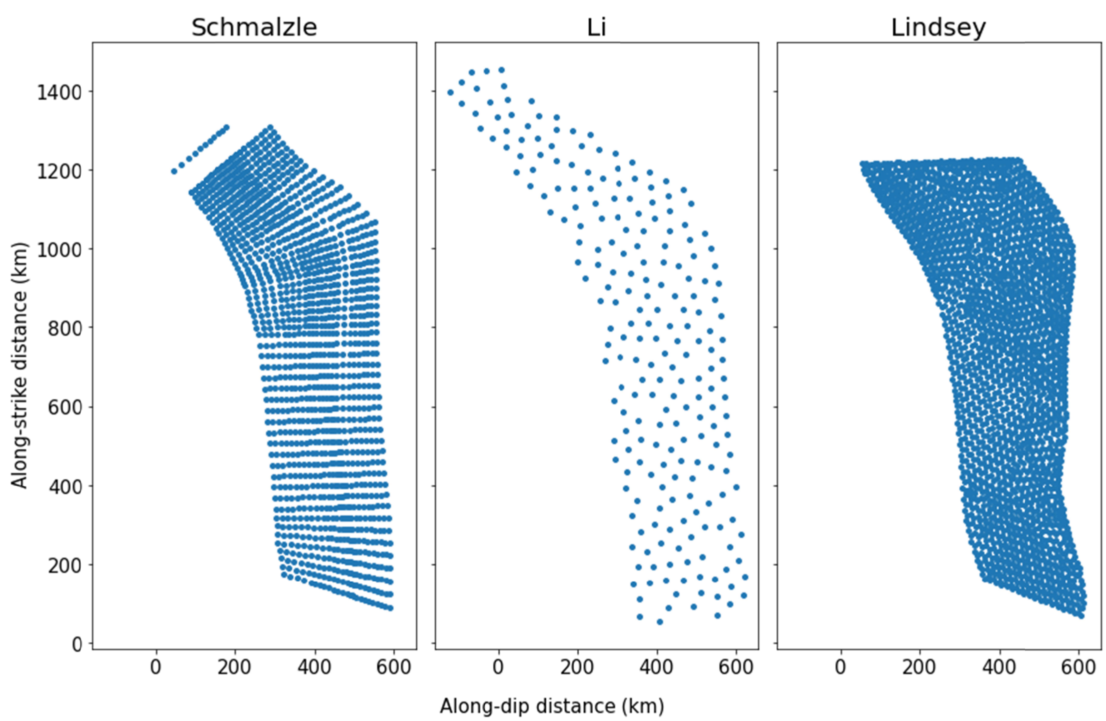

Figure 5. Spatial distribution of original data points for the three interseismic locking models.

4. Results

4.1 Stress build-up from locking distribution

From static calculations as described in section 3.3, we obtain distributions of stress accumulation from the total slip deficit (Figure 4). Since the stress change along strike is negligibly small, only the dip component is shown in Figure 4d – 5f. Nevertheless, the strike component is used in our dynamic simulations and the points of the highest stress change are determined by the magnitude of stress change vectors. The slip deficit distributions calculated from the locking models have similar patterns and their moment magnitude within the static model domain only differs by 0.03 (i.e. Mw 8.99-9.02) (Figure 4). All of the three locking models feature high slip deficit above 12 m in northern Cascadia (above 900 km along-strike distance in Figure 4). The largest contrast between the models is in the south and central segments. For example, the segment that exhibits more creep is located at 550-750 km, 300-550 km, and 500-700 km in the Schmalzle, Li, and Lindsey models, respectively, with different maximum slip deficits (Figures 5a – 5c). The derived stress accumulation distributions display a larger difference in along-strike variations among these models (Figures 5d – 5f).

The depth extent of positive stress build-up based on the Schmalzle model extends to ~20 km depth. We can locate three high-stress patches, labelled A1, A2, and A3 in Figure 4d in our dynamic model domain. A2 hosts the maximum stress build-up of 3.3 MPa. Between the A2 and A3, there is a creeping segment with obviously lower stress (labeled C in Figure 4d). Such stark along-strike variations are not that obvious in the slip deficit distribution (Figure 4a), because the stress accumulation is proportional to the second derivative of slip deficit. While A2 and A3 host sharp downdip decrease in slip deficit within a narrow locking zone, the C segment has a more gradual decrease with deeper locking depths. This illustrates that the stress distributions can reveal the seismic potentials that may not be identified as first-order features in slip deficit distributions.

The stress build-up based on the Li model shows a more uniform along-strike distribution, except in the northernmost region where the highest slip deficit takes place (Figure 4). The positive stress in this model extends deeper, to ~30 km depth. Although it has a longer zone of low slip deficit than in the Schmalzle model, there are no distinct high-stress patches but only a slightly low-stress patch at 400-500 km (Figure 4e). The maximum accumulated stress in the dynamic model domain is only 1.7 MPa.

The Lindsey model shows a somewhat similar along-strike variation of stress distribution to the Schmalzle model even though the amplitude is different (Figure 4). The slip deficit amplitude of the Lindsey model is significantly lower than the Schmalzle model, having maximum accumulated stress of 1.9 MPa in the north. Yet, their spatial gradient variations along strike are alike. The south and the north have steeper slip deficit gradients constrained in shallower depths ~30 km while the creeping segment has a noticeable gentle decrease in slip deficit until the slab bottom (Figure 4c). Considering that rupture segmentation is dominated by the spatial variation of stress instead of the amplitude, we expect results akin to the Schmalzle model given modifications of frictional parameters regarding the amplitude. Because of the poorly constrained down-dip locking depths, the Lindsey stress model is not further evaluated for dynamic simulations.

4.2 Predicted rupture scenarios

Using the initial stress which includes accumulated stress derived from locking models, we initiate the ruptures with a range of hypocenter locations. In some dynamic scenarios, the ruptures propagate outside with considerable rupture extent, classified as breakaway scenarios. The examples of breakaway scenarios using d_c of 0.6 m and 1 m are shown in Figure 6-7 and Figure S6 respectively. While in other cases, the rupture propagation stops immediately outside the nucleation zones due to the lack of elastic energy release to overcome the fracture energy required to weaken the fault, termed self-arresting events (Figure 8). The moment magnitude for scenarios is calculated according to the integral of the final slip (d) over the fault plane area (A) using an average shear modulus (μ) of 35 GPa ($M_0=\mu Ad$; $Mw=2/3*(\log_{10}(M_0) - 9.1)$). Our moment magnitude gives a lower limit for the scenarios that propagate out of the model domain (e.g. Figure 6e & 6g). The slip rate means the relative particle velocity across the fault while the rupture speed is the rate of rupture front movement (Rowe and Griffith, 2015), calculated every 10 seconds.

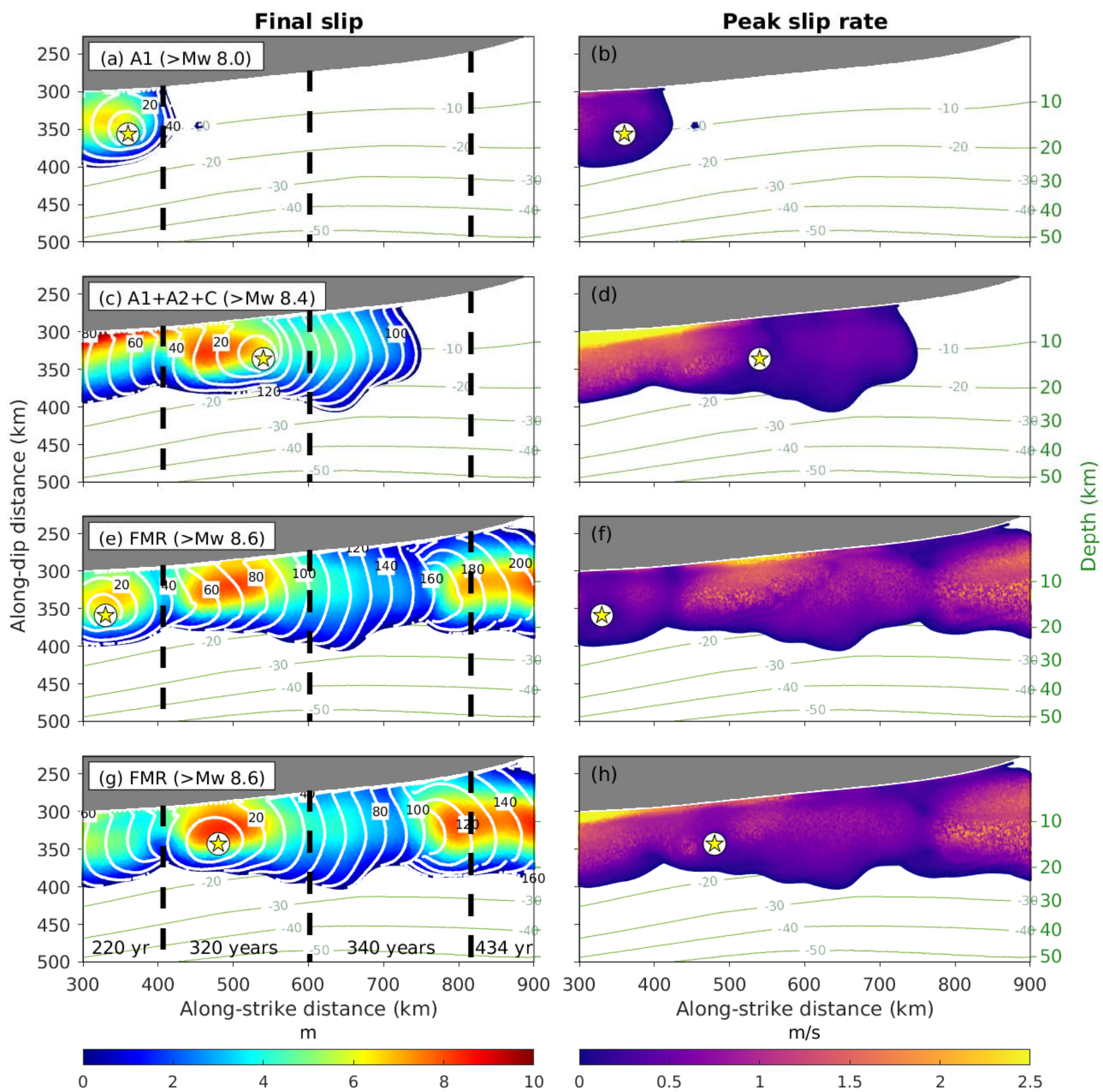

Figure 6. Dynamic rupture scenarios derived from the stress distribution of the Schmalzle model using a dc of 0.6 m. (a), (c), (e), (g): Final slip distribution. Stars: hypocenter locations. Olive-green contours: slab depth contours. Rupture fronts (white contours) are displayed every 10 seconds and numbered every 20 seconds. Black dashed lines: recurrence time intervals of 220, 320, 340, and 434 years (Goldfinger et al., 2017) as written in (g). The labeled Mw is calculated by slip within the model domain, thus scenarios with slip extending outside the domain should have larger magnitudes. (b), (d), (f), (h): Peak slip rate throughout the rupture. Stars: hypocenter locations. Olive-green contours: slab depth contours. (a) – (b): Scenario rupturing the A1 asperity. (c) – (d): Scenario rupturing A1 and A2 asperities and part of the creeping segment C. (e) – (f): Full-margin rupture (FMR) initiated from A1. (g) – (h): Full-margin rupture initiated from A2.

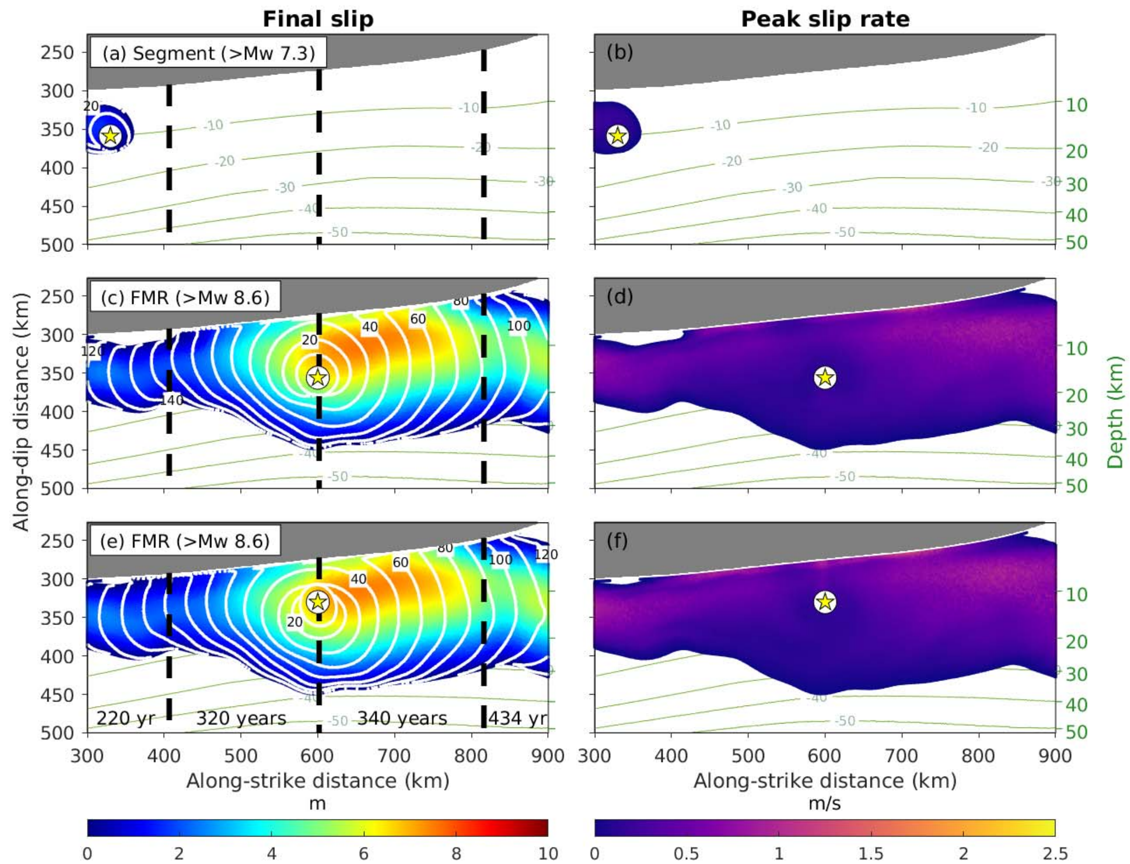

Figure 7. Dynamic rupture scenarios derived from the stress distribution of the Li model using a dc of 0.6 m. Same as Figure 6, except for the Li model. (a) – (b): Segmented rupture scenario. (c) – (f): FMR for different hypocenter locations.

We further classify the self-arresting and breakaway events explicitly. According to the empirical relationships between the rupture area and magnitude, the rupture within the nucleation zone is around Mw 6.5 (Wells and Coppersmith, 1994). Earthquakes generally have rupture velocities higher than 1 km/s (Rowe and Griffith, 2015) and demonstrate a ratio between rupture velocity and v_s starting from around 0.4 (Weng and Ampuero, 2020). Since our v_s at trench (5 km depth) is 3.17 km/s, we expect breakaway ruptures to reach rupture velocities higher than 1.27 km/s (0.4 v_s). Consequently, we define the scenarios with Mw < 6.5 and rupture speed less than 1.27 km/s as self-arresting ruptures, and those above as breakaway ruptures. Our analysis will only focus on the breakaway ruptures, considering self-arresting ruptures are merely the results of artificial nucleation.

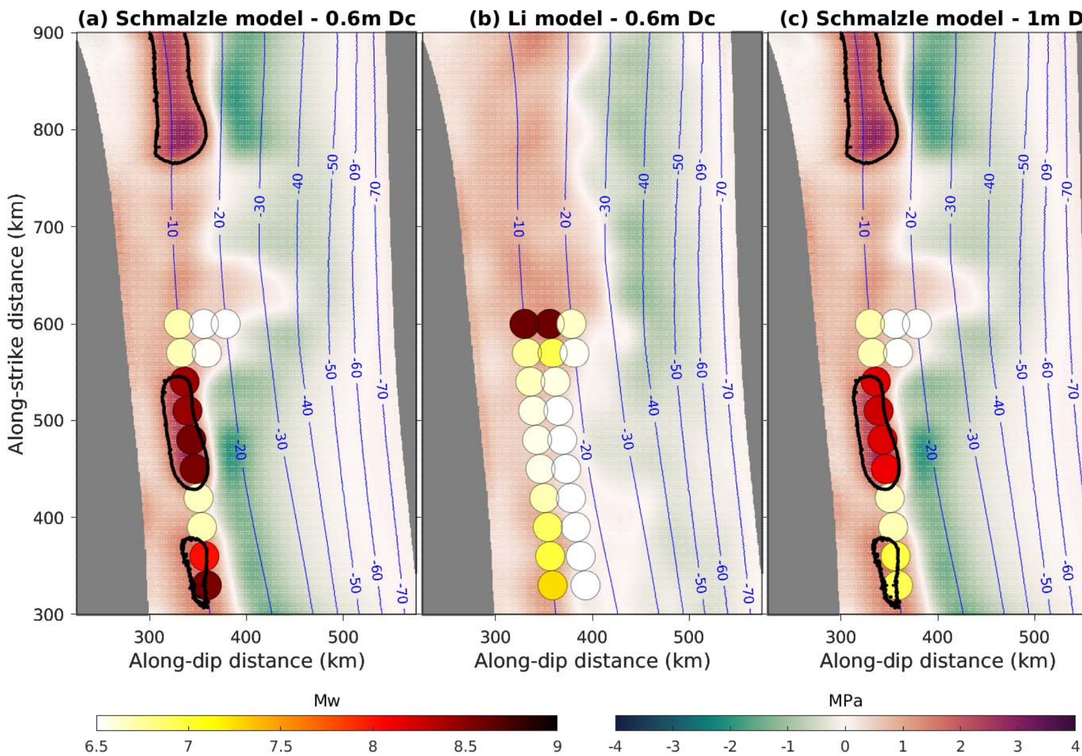

Figure 8. Moment magnitude dependence on hypocenter locations. Map view of the moment magnitudes of rupture scenarios nucleated at each location (circles) with the stress build-up in the background, and slab depth contours (blue lines). (a) Scenarios derived from the Schmalzle model using a dc of 0.6 m. Black lines: 1.5 MPa stress contour, same as in Figure 4. (b) Scenarios acquired from the Li model using a dc of 0.6 m. (c) Same as (a) except for a dc of 1 m (Figure S6).

We further divide the breakaway ruptures into segmented ruptures and full-margin ruptures. “Full-margin ruptures” represent rupture scenarios that propagate out of the entire model domain. Because the northern Cascadia holds the highest accumulated stress and our model domain includes a part of the northern segment, it is reasonable to assume that the ruptures propagating out of the domain’s northern boundary would eventually rupture the entire northern Cascadia. Similarly, the southern segment inside the domain has consistent stress levels with the southernmost Cascadia outside of the domain, hence we assume the “full-margin ruptures” can break the southern Cascadia as well. For this reason, we name the ruptures propagating out of the south and north of the domain as “full-margin ruptures” in the following context. In contrast, the scenarios where their along-strike rupture extents within the model domain are regarded as segmented ruptures.

Full-margin ruptures are shown in both the Schmalzle-based (Figure 6e-h) and the Li-based scenarios (Figure 7c-f) with maximum final slips of 8.5 m and 7.6 m respectively. Those of the Schmalzle model are larger than Mw 8.6, reaching a rupture speed of 3.1 km/s and a peak slip rate of 4.5 m/s. The source durations last for more than 150-200 seconds depending on the hypocentre location (Figure 9a). On the other hand, the moment magnitudes of full margin ruptures from the Li model are also higher than Mw 8.6. They have a slightly lower rupture velocity of 2.7 km/s and a peak slip rate of 1.4 m/s. The source duration is less than 140 seconds (Figure 9a). The full-margin ruptures of the Schmalzle model and the Li model halted at 30 km and 40 km depths respectively. All are initially predominated by crack-like ruptures, evolving into pulse-like ruptures (Movies S1-S4).

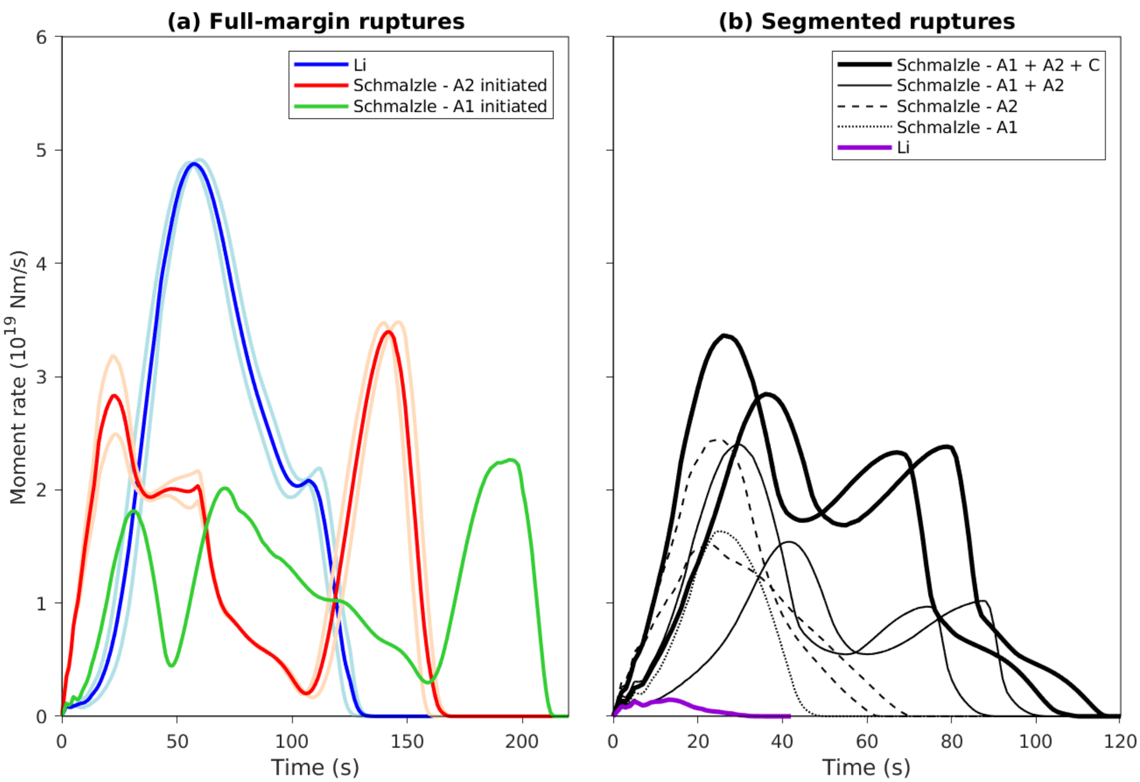

Figure 9. Moment rate functions from dynamic rupture simulations. (a) Moment rate of all full-margin rupture scenarios. Moment rate functions of individual neighboring rupture scenarios are indicated by lighter colors (light blue for the Li model and pink for those initiated from A2 in the Schmalzle model) and the average is marked by solid colors (blue for the Li

model and red for the A2 initiation in the Schmalzle model). Note that there is only one event initiated from A2 in the Schmalzle model (green line). (b) Moment rate of all segmented ruptures for the Schmalzle model (black lines) and Li model (purple line). A1 and A1+A2+C ruptures were derived using a dc of 0.6 m while A2 and A1+A2 scenarios were simulated with a dc of 1 m (Figure S6). The only segmented event from the Li model utilizes a dc of 0.6 m.

Despite having the same accretionary wedge setting as in Figure S4b, all the scenarios shown here, except case 6c, do not demonstrate the large near-trench slip as tested above because of the different hypocenter locations and stress distribution. For the Schmalzle model, as the rupture initiates in the south, the combined effects from rupture directivity and free surface reflection in the south are smaller as compared to initiation from the north. As for the case of 6c, its hypocenter is located further north, thus allowing a stronger directivity. However, such high slip trench features are also absent in the north even with hypocenters in the south. This is because while the rupture propagates through the central creeping segment, the energy depletes and it is insufficient to cause a large slip until it reaches the high-stress asperity at the north. For the Li model, the high-slip trench is also absent because there are no particular high-stress asperities that could trigger larger slip near the trench.

Rupture segmentation is observed in both models. From the Schmalzle model dynamic scenarios, we observed one scenario breaking A1 (Figure 6a) and two scenarios rupturing A1, A2, and partly C (Figure 6c). The A1 segmented rupture (Figure 6a) is initiated by a hypocenter location at A1 asperity and the source duration continues for 40 seconds (Figure 9b), with rupture stopped above 20 km depth. Both A1+A2+C scenarios (Figure 6c) are triggered by nucleation at A2 asperity, and the source durations last for 110-120 seconds (Figure 9b), having slip above 30 km depth. For the Li model, only one dynamic segmented scenario is found rupturing the southernmost segment. Since the rupture initiation is close to the domain boundary, the rupture propagates out of the south quickly after nucleation while being arrested in the north and above 20 km depth (Figure 7a), resulting in a duration time as short as 30 seconds (Figure 9b). Except for the A1+A2+C dynamic models which have similar rupture evolution behaviors to the full-margin ruptures (Movie S5), the short segment ruptures (Figure 6a and 7a) are primarily crack-like ruptures as the rupture duration is insufficient for them to grow into pulses (Movies S6-7).

4.3 Hypocentral effects on the potential moment magnitude and ground surface response

In view of the different resulting scenarios, we investigate the effect of different hypocenters in both models with a d_c of 0.6 m. For the heterogeneous Schmalzle model, there is a strong along-strike variation in moment magnitude with respect to the stress distribution (Figure 8a). The nucleation zones within the highest stress patch A2 result in scenarios with Mw > 8.4-8.6 and the events within A1 have Mw >8.0-8.6. All the nucleation centers lying outside of the stress asperity results in self-arresting ruptures. This demonstrates the hypocentral dependency of magnitudes in the Schmalzle model.

Meanwhile, the Li geodetic locking model gives a smoother and more homogeneous stress distribution within the model domain that does not flavor rupture segmentation except in the southernmost region where the initial stress is slightly higher (Figure 8b). Although full-margin ruptures take place with hypocenters in a particular region, it by no means suggests that the ruptures are larger on that site. It shows that the initiation of full-margin ruptures is sensitive to slight stress perturbations on the fault. Recalling our assumptions of linear stress accumulation and uniform background stress level, small deviations on these assumptions (e.g., spatial variations in stress accumulation time and material properties) could cause comparable stress distribution perturbations in the Li model while the perturbations would be relatively insignificant in the Schmalzle model. Thus, the Schmalzle model shows clearer seismic potential while the Li model is more ambiguous considering the uncertainties in stress accumulation evolution and background stress field.

Apart from the moment magnitude, the hypocentral effects on ground surface response are also noticeable. We compare the velocity magnitude of synthetic stations near major cities derived from the margin-wide scenarios in both models (Figure 10). Although the rupture extent from the scenarios of different hypocentres in each model is highly similar (Figure 10a), the amplitude of peak ground velocity can differ twice. For instance, in the Li model, the two hypocenters have the same along-strike distance but different downdip depths - 10 km and 15 km. The deeper nucleation event (15 km) clearly demonstrates larger peak ground velocities than the shallower one (10 km) at stations CAVE, DBO, BUCK, and LOKI. Such a difference is primarily due to the rupture propagation. For the 15 km event, the rupture propagates updip since initiation, setting off a strong wavefront (Figure 7c). However, the 10 km event starts by propagating downdip and is followed by updip fault slip along the sides of the nucleation zone, creating two wavefronts shortly after the nucleation (Figure 7e). The interference of these seismic waves and those from downdip fault slip leads to a more ambiguous waveform slightly lagging behind the 15 km event even though the 10 km one is in closer proximity to the surface.

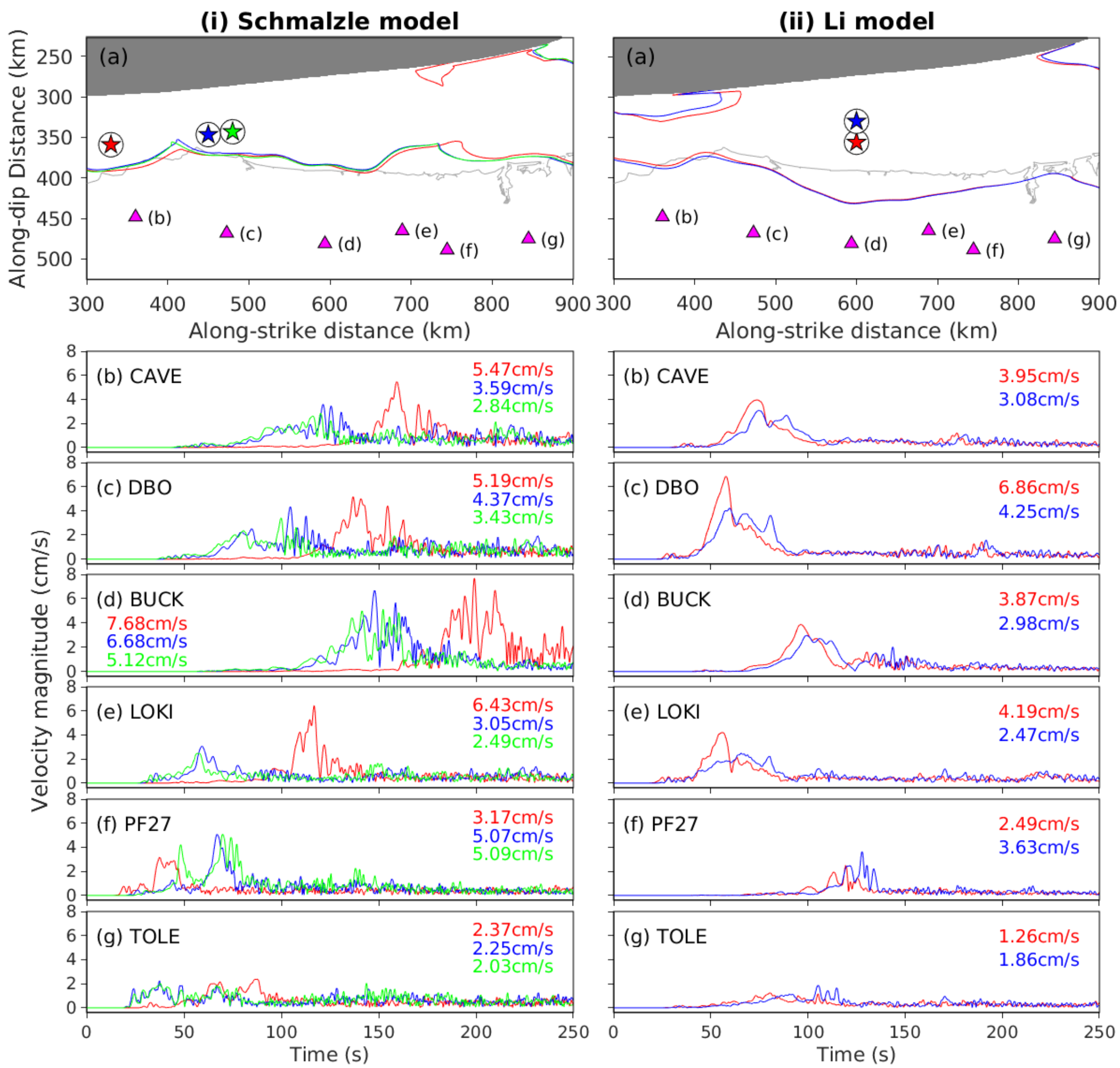

Figure 10. Synthetic velocity magnitude at stations for FMR. (a) The 1 m final slip contour of rupture scenarios with the coastline (light grey). The colors of the slip contours match with the star (hypocenter location) colors. Magenta triangles: station locations. Labels beside stations: the plot number. The stations near major cities along the strike are selected from the Pacific Northwest Seismic Network. (b) – (g) Comparison of velocity magnitudes (three-component combined) among the rupture scenarios in (a) with matching colors. The corresponding peak velocity magnitudes and station names are marked on each trace. (i) Schmalzle model. (ii) Li model.

The hypocentre at a different location along-strike also results in different waveforms. In the Schmalzle model, there are two hypocentres in A2 and one hypocentre in A1 contributing to margin-wide ruptures. Since the A1 hypocentre event propagates from the southernmost region to the north of the domain, the strong directivity causes a distinct pulse as compared to the A2 hypocentre events. For example, the LOKI station has peak velocity magnitudes of 2.5-3 cm/s for the A2 hypocentres but 6.4 cm/s for the A1 hypocentre (Figure 10ei).

4.4 Seafloor deformation and coastal subsidence in margin-wide scenarios

We also evaluate the surface deformation patterns for our margin-wide rupture scenarios. The peak vertical ground displacements for the Schmalzle-based and Li-based scenarios are similar in magnitude, ranging from -1.1 m to +1.0 m and from -1.2 m to +1.1 m respectively. On the other hand, the maximum peak ground velocity of the Schmalzle-based scenarios (i.e. 2.3 m/s) is remarkably higher than that of the Li-based scenarios (i.e. 1.2 m/s) by almost double. Both models show the highest peak ground velocity towards the tip of the continental crust and the northernmost region of the domain.

Coseismic hingeline refers to the point where there is zero seafloor vertical displacement. Compared to the Schmalzle model (Figure 11b), the coseismic hingeline for the Li model (Figure 11d) is further inland, especially for the central and northern segments because the down-dip rupture extent of these regions in the Li's model is deeper (Ramos et al., 2021). However, it is noted that the down-dip locking depth of the seismogenic zone is poorly constrained by geodetic data (Wang and Tréhu, 2016), thus we only focus on the along-strike variations of coastal subsidence instead of the absolute amplitudes.

The average coastal subsidence is then extracted from the peak vertical ground displacement of the data points closest to the coastline in all the margin-wide scenarios in both models (Figure 11a). We then compare the synthetics with the subsidence records of the A.D. 1700 M9 earthquake. In our scenarios, Li's coastal subsidence gives a more distinct pattern compared to the observations, having the largest amount of deformation at the high slip patch in the north and decreasing further away. For scenarios from Schmalzle model, the along-strike coseismic subsidence appears to fluctuate with slightly larger deformation in high-slip segments region.

The subsidence records of the A.D.1700 M9 rupture also exhibit heterogeneous along-strike pattern, which can be matched by models with several high slip patches (Wang et al., 2013). In our case, the scenarios from Schmalzle model can reproduce a similar along-strike variation with the observations, mainly due to the higher slip heterogeneity with three high-slip patches (Figure 6) compared to those from Li model (Figure 7).

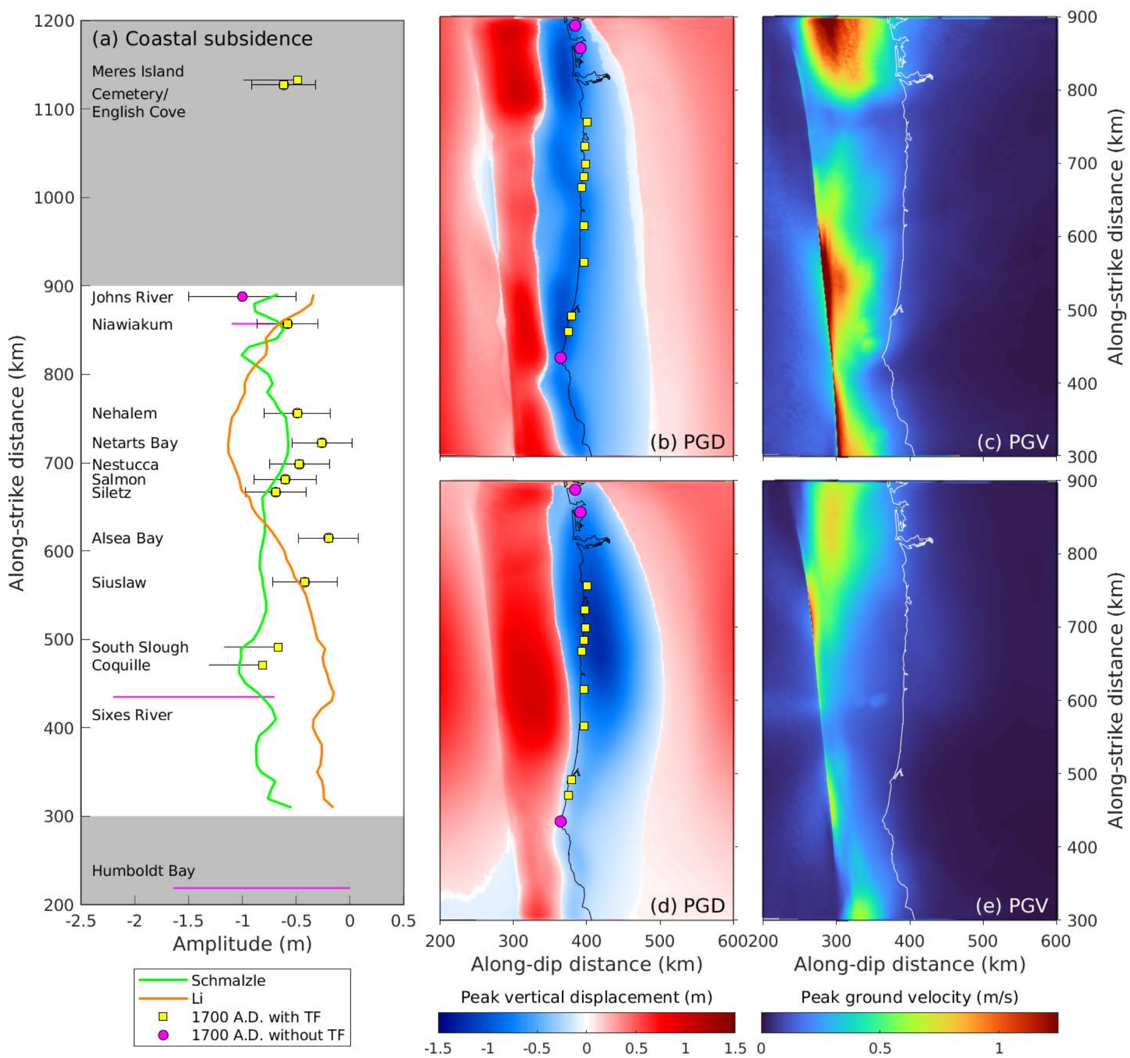

Figure 11. Ground motion intensities of full-margin ruptures. (a) Average peak vertical ground displacement along the coastline for the Schmalzle model (green line) and the Li model (orange line). Yellow squares: observations sites with transfer function analysis (TF). Pink circles: sites without TF. Error bars: one standard deviation. Black lines with yellow squares at one end: yellow squares as the minimum estimates. Pink line: uniform distribution. Grey patches: regions outside of the model domain. (b) Peak vertical ground displacement of full-margin ruptures in Schmalzle model with the coastline (black line). Observation sites with

(yellow square) and without (pink circles) TF analysis (Wang et al., 2013). (c) Average peak ground velocity PGV of FMR derived in the Schmalzle model with the coastline (white line). (d) – (e): Same as (b) and (c) respectively except for the Li model.

5. Discussions

5.1 Potential rupture patterns in correlation with recorded segmentation and recurrence intervals

The rupture extents of segmented scenarios in the Schmalzle model are consistent with the recorded segmentation of paleoearthquakes (Goldfinger, 2012, 2017). The A1 scenario arrested around the 220-320 years recurrence interval boundary (Figure 6a), and all cases for A2 and A1+A2 ruptures stopped around the 320-340 years boundary as it enters the creeping segment (Figure S6). For A1+A2+C cases, the ruptures extend to part of the 340-year recurrence interval segment but not the whole (Figure 6c). This is because the A3 asperity is located slightly off the recurrence interval boundary. Therefore, the ruptures could either arrest before A3 or propagates to the rest of the high-stress northern region, causing full-margin ruptures. There are also ambiguities in determining the paleoseismic rupture limits due to limitations in core data. Hence, the A1+A2+C scenario is supported by the estimated minimum rupture limit in the segmented rupture model (Goldfinger et al., 2017) where rupture stops before the 340-year segment. On the other hand, the Li model does not share particular similarities with the recurrence interval segments within the model domain.

We also find that the margin-wide ruptures can be derived for all models given certain frictional parameters. Given the 320 years of silence and the recurrence intervals of 220-340 years in the south, all segmented scenarios are possible in the current stage. Although there are few constraints on the frictional parameters of the Cascadia megathrust, our combination of parameters allows variations in rupture scenarios, including segmented and margin-wide ruptures comparable with the geological records. This may suggest that the ratio between the frictional parameters and initial stresses is reasonable, if not the absolute amplitudes. The margin-wide rupture initiates at A1 and A2 in the Schmalzle model and the boundary between 320 and 340 recurrence intervals for the Li model. This reflects that at the current state, the possibility of ruptures initiating from the south or central Cascadia growing into a margin-wide rupture cannot be eliminated.

The diverse segmentation in our scenarios results from heterogeneous locking and various hypocentre locations. Ramos et al. (2021) initiated dynamic rupture simulations at locations of highest stress drop in the south, resulting in full-margin ruptures for scenarios with uniform stress accumulation time, and both full-margin and segmented rupture scenarios using heterogeneous stress accumulation time along strike which is determined empirically. Indeed, the uncertainties in the stress accumulation history could be introduced by a heterogeneous time interval. However, with strong along-strike differences in accumulation time, the stress distribution becomes largely affected by the empirical time interval instead of the locking distribution. Our study shows that segmented ruptures are possible even using a uniform stress accumulation time when different nucleation zones are used. The application of hypocentre locations discovers the possibilities of rupture initiation from a range of stress drops, thus more segmentation patterns are found apart from the largest possible margin-wide ruptures (Figure 12).

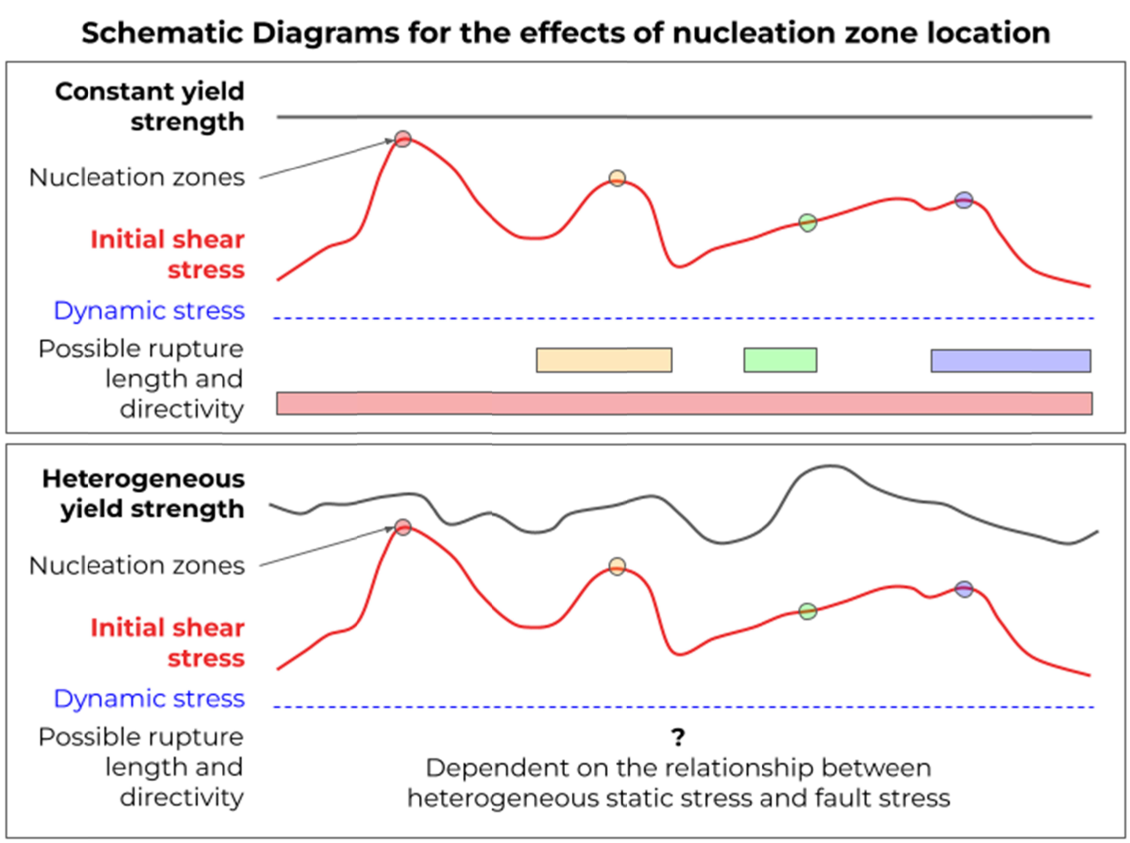

Figure 12. A schematic diagram demonstrating the potential rupture segmentation by applying different hypocenter locations.

5.2 Simulated coseismic subsidence compared with A.D. 1700 records

Although the along-strike variation of the heterogeneous Schmalzle-based coastal subsidence has a reasonable consistency with the paleoseismic records, our synthetic subsidence is generally slightly larger than the observations, exceeding one standard deviation in two sites (i.e., Alsea Bay and Siuslaw) at the central segment. Here we will provide possible reasons for such discrepancy.

Similar to the earthquake sequence simulations (Li and Liu, 2021), our subsidence is larger than the observational data at Alsea Bay. Our research focuses on estimating future earthquakes thus we assume homogenous background stress levels immediately after the A.D. 1700 margin-wide rupture. However, the background for A.D. 1700 could in fact be heterogeneous due to the spatial and temporal uncertainties in the geodetic locking and the slip history before the A.D. 1700 rupture. It is possible to reconstruct best-fit subsidence results by adjusting accumulation time empirically as in Ramos et al. (2021) but this is beyond the scope of our study.

Another important factor controlling subsidence is the inelastic accretionary prism deformation. One outstanding example is the 2011 Tohoku-Oki earthquake where the region of the largest slip does not cause the largest tsunami height possibly due to the inelastic deformation of the accretionary prism (Fujiwara et al. 2017; Wilson and Ma, 2021). Han et al. (2017) observed an along-strike variation for the consolidation state of the accreted sediments in Cascadia and propose that this could contribute to the megathrust slip behavior. For instance, offshore Washington has over-consolidated sediments incorporated into the mechanically strong outer wedge, and very little sediment is being subducted, flavoring potential near-trench rupture. On the other hand, a thick sequence of under-consolidated fluid-rich sediment is subducting offshore Central Oregon, possibly facilitating elevated pore pressure, thus promoting possible aseismic slip in this area. These factors may account for the slight deviation of our model subsidence from the data.

5.3 Comparison between dynamic simulations and static methods

Our dynamic simulation showcases a lower moment release in all rupture scenarios than estimations from variant static methods. Static methods commonly provide the upper bound of possible slip by assuming complete release of slip deficit in future earthquakes (Figure 13). The maximum slip deficits within the model domain for both models only differ slightly – 12.9 m for the Schmalzle model and 13.0 m for the Li model. Consequently, the maximum slip in the Schmalzle-based dynamic rupture model (8.5 m) contributes about 66% of the maximum slip deficit, and that of the Li model (7.6 m) is about 59%. This difference with the static locking models is observed in a number of studies, including the potential rupture segmentations for the Anninghe fault in west China (Yao and Yang, 2022), the central American subduction zone where the 2012 Nicoya Mw 7.6 earthquake occurred (Yang et al., 2019a), Himalaya front where the 2015 Nepal Mw 7.8 earthquake took place (Li et al., 2016), as well as the south American subduction zone where the 2010 Maule Mw 8.8 earthquake occurred (Moreno et al., 2010).

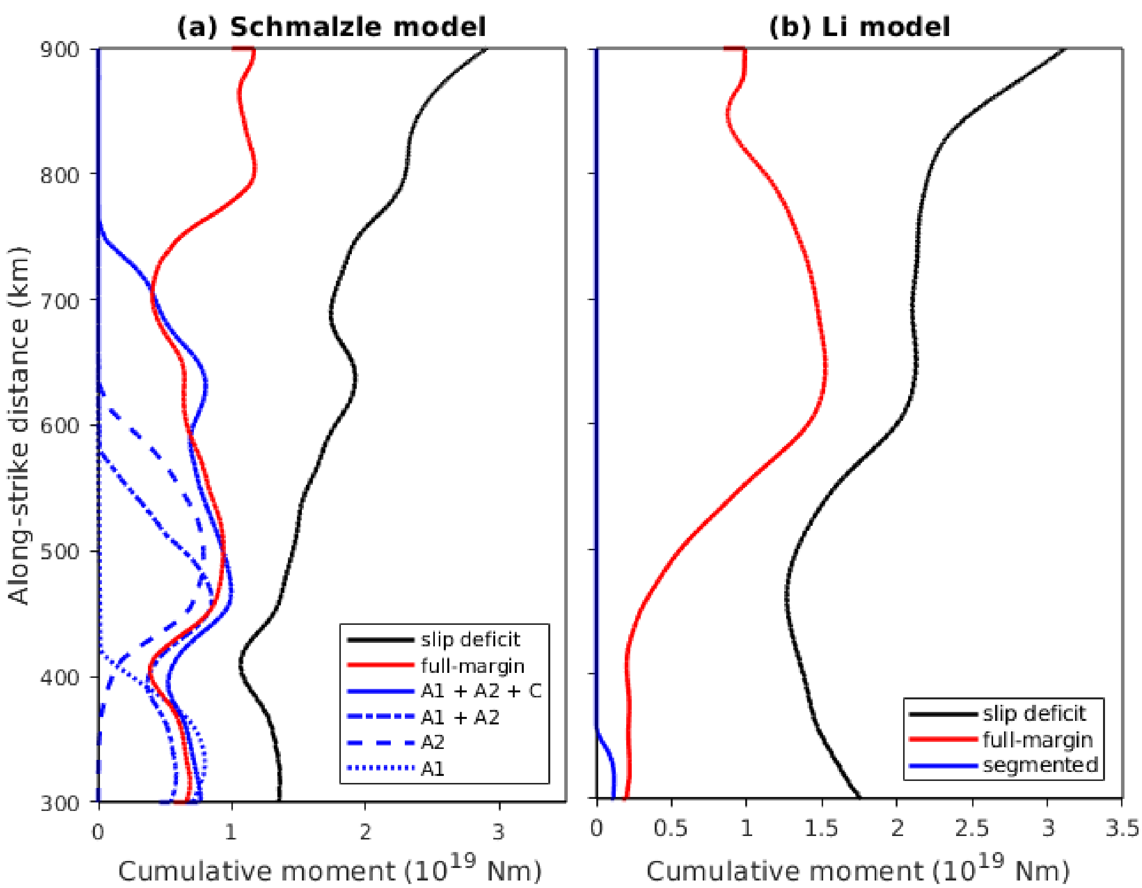

Figure 13. Cumulative moment versus the along-strike distance. The cumulative moment is the product of rigidity, slip, and area, integrated over every 500m width along strike. Blue lines: average cumulative moment for the segmented scenarios. Red lines: average cumulative moment for full-margin ruptures. Black lines: cumulative moment assuming all slip deficit in Figure 4 are released, also known as the moment deficit (Maurer et al. 2017). (a) Dominant dynamic rupture scenario types and slip deficit for the Schmalzle model. (b) Same as (a) except for the Li model.

These suggest that given our current frictional parameters, a considerable fraction of the slip deficit in regions of low to moderate stress drop is not released during dynamic simulations in our models (Figure 14 and 15). The stresses on these areas can be relieved later possibly in form of coseismic events and slow slip events. For instance, Cascadia is well-known for its episodic tremor and slow slip events. In addition, considering the poorly constrained downdip limit of the seismogenic zone using geodetic observations, the unreleased slip deficit may in fact represent uncertainties, including the portion for interseismic stress relaxation and the temporal variation in locking width (Wang and Tréhu, 2016). Therefore, the discrepancy highlights the necessity of conducting dynamic simulations on top of static calculations.

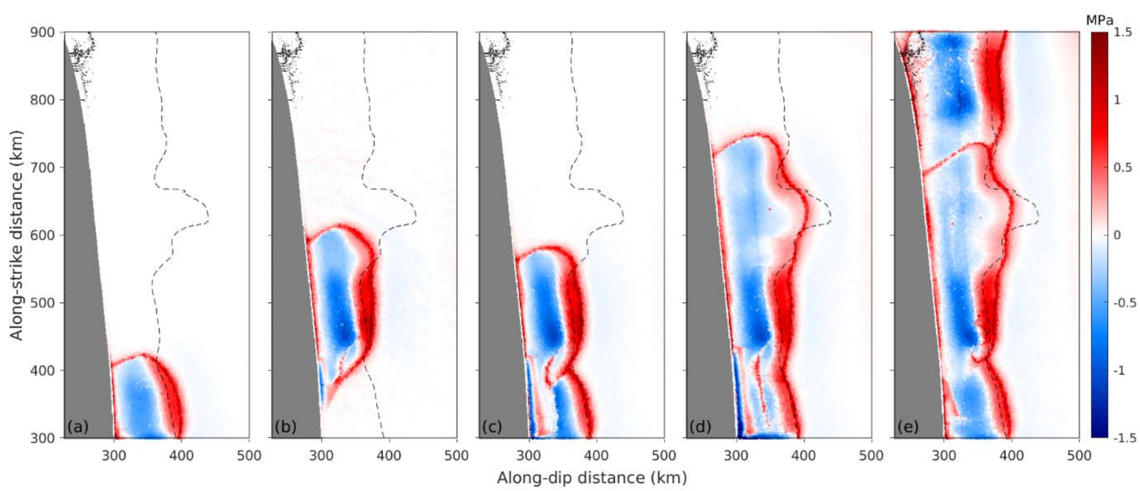

Figure 14. Stress change distributions for the Schmalzle -based rupture scenarios. Dashed lines: contours of zero stress change derived from locking models (Figure 4). The up-dip portion of the contour contains positive stress build-up. (a) Example of A1 rupture (Figure 6a). (b) Example of A2 rupture (Figure S6c). (c) Example of A1+A2 rupture (Figure S6e). (d) Example of A1+A2+C rupture (Figure 6c). (e) Example of a full-margin rupture (Figure 6g).

We further compared our results with the heterogeneous ruptures inferred from coastal subsidence estimates. Wang et al. (2013) proposed a range of heterogeneous slip models for the A.D. 1700 event using a 3D elastic dislocation model with reference to the subsidence estimates. Assuming the fault slip patches follow the bell-shaped function, they adjusted the slip patches parameters (e.g., size, location, and peak slip) to match the model-predicted surface deformation to the paleoseismic subsidence estimates using a trial-and-error approach. In particular, they preferred a model consisting of four high-slip patches for simplicity and having a reasonable fit with the observations. However, the models are limited by the large subsidence data gaps in northern and southern Cascadia (Figure 11).

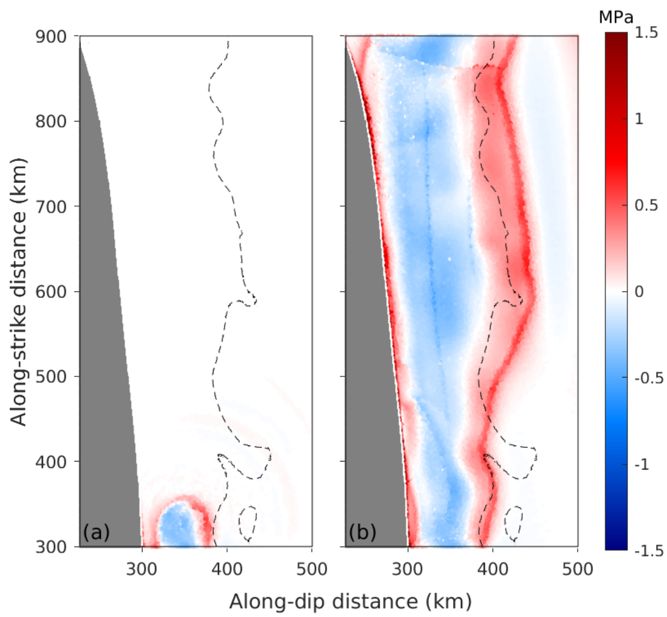

Figure 15. Stress change distributions for the Li-based rupture scenarios. (a) Example of segmented rupture (Figure 7a). (b) Example of a full-margin rupture (Figure 7e).

In comparison, our scenarios incorporate the locking models utilizing Global Navigation Satellite System GNSS data, which are more densely spaced along Cascadia, as the physical constraints on rupture depth and heterogeneities. We demonstrate that three high-slip patches in a dynamic rupture model could be sufficient to generate subsidence amplitudes similar to the observation of the A.D. 1700 megathrust earthquake. Similarly, the Schmalzle locking-constrained earthquake sequence simulation (Li and Liu, 2021) also suggests a three high-slip patches scenario, and its synthetic subsidence is in good agreement with the observational data. Therefore, the three high-slip patches scenarios could be close to the A.D.1700 event.

5.4 Limitations in deriving future coseismic slip

Although our dynamic models produce reasonable ground motions that match with multiple observational studies, there are limitations in constraining the up-dip frictional properties and rupture behaviors. The first concern comes from the frictional behaviors of the frontal prism. In our model, the strength drop (difference between static and dynamic stress) at the frontal prism decreases towards the trench, and the initial stress equals the addition of dynamic stress and stress drop from static simulation. Although cohesion could suppress fault failure at the beginning of the simulations, the stress perturbations from ruptures could induce higher slip rates at shallow depths as it easily overcomes the small strength drop, especially with the dynamic effects of free-surface reflection. However, in reality, velocity-strengthening materials are known to slip at low rates. Our models do not consider the plastic deformation of the frontal prism either. Indeed, cohesion could partly describe the energy absorption close to the free surface caused by the presence of unconsolidated gouge and clays (Galvez et al., 2014). However, the amplitude of cohesion in our case is not constrained by laboratory experiments, including local mineralogy, lithology, and fluid pressure. Moreover, the frictional behaviors in our model are prescribed for the fault interface and off-fault plasticity is neglected. Ulrich et al. (2022) and Wilson and Ma (2021) highlight the inelastic deformation of sediments as one of the dominant factors controlling seafloor deformation, hence tsunamic genesis. Incorporating off-fault plasticity and careful descriptions of frictional behaviors with respect to laboratory experiments and offshore geological studies would help establish realistic dynamic rupture scenarios.

Another major concern in estimating tsunami hazards comes from the uncertainty in future shallow rupture behavior. In our model, we assumed a simplified fault geometry where the fault extends to the top of the model domain, introducing trench-breaching ruptures in our dynamic models. Nevertheless, other rupture modes such as buried rupture, splay-faulting, and activation of thrusts and back-thrusts are possible (Wang and Tréhu, 2016). Gao et al. (2018) constructed hypothetical splay-fault geometries in addition to Priest et al. (2009) and a continuous along-strike frontal thrust model based on seismic profiles. Therefore, a more detailed 3D mapping of the complex fault geometry could help evaluate the possibility of different rupture mechanisms using dynamic rupture simulations.

6. Conclusion

In this study, we conducted 3D dynamic rupture simulations for Cascadia using different interseismic locking models with a range of hypocenter locations in the South. While the locking models have similar static moments and locking distributions, their heterogeneous stress distribution leads to distinct rupture scenarios. Both Schmalzle and Li models demonstrate that the south is capable of generating Mw >8 segmented ruptures and full-margin ruptures depending on the frictional parameters and hypocenter locations. For instance, both segmented and full-margin ruptures can occur with the same hypocenter location given different frictional parameters.

We found that the heterogeneity of interseismic locking models plays a key role in determining the rupture process. The more heterogeneous Schmalzle locking model yields a stress distribution with more asperities, thus facilitating segmented ruptures on the high-stress asperities. These segmented ruptures appear to have a reasonable correlation with the along-strike extent of the inferred recurrence intervals. On the other hand, the more homogeneous Li locking model gives a smoother stress distribution, hence the scenarios are either full-margin ruptures or self- arrested ruptures. The selection of hypocenter location is also a crucial parameter in controlling the potential segmentation patterns. For the more heterogeneous model, the scenarios that initiated from the higher stress asperities demonstrate a significantly larger moment magnitude.

Accordingly, surface deformation is also largely controlled by these factors. While the homogeneous locking model results in a simpler coastal subsidence pattern with the largest subsidence in the region of highest slip and decreasing further away, the heterogeneous model gives a more complex pattern depending on the stress asperities. This also suggests that the A.D.1700 earthquake may represent a possibly more heterogeneous slip model provided its fluctuating coastal subsidence pattern. In particular, our results show that a three high-slip patches scenario can reproduce a reasonably similar seafloor deformation with the A.D. 1700 earthquake. Apart from coastal subsidence, the synthetic ground shaking also demonstrates that rupture directivity is strongly controlled by prescribed hypocenter locations, leading to nearly double the peak ground velocity for scenarios initiated at different hypocenter locations even though the resulting slip distributions are almost the same.

This project can be further developed from multiple perspectives in the future, including the off-fault plasticity, the along-strike changes in accretionary prism geometry, and the addition of splay faults. Our simulation results can also be applied to tsunami modeling to evaluate the tsunami risks for each segmented rupture type. Furthermore, our models may help evaluate the present probabilistic seismic hazard analysis (PSHA) by providing possible slip distributions of the paleoearthquakes for source characterization as well as the synthetic ground motions for comparison with that generated by the empirical ground motion prediction equations.

On top of specific investigations on Cascadia, our findings could help understand the general relationship between interseismic locking models and the possible earthquake slip patterns, thus the moment magnitudes. Our study together with the dynamic simulations for the other fault zones, such as the Nicoya Peninsula subduction megathrust (Yang et al., 2019a) and the Anninghe fault (Yao and Yang, 2022), raises the possibility to provide new insights into more efficient slip estimations of seismic potentials for the fault zones worldwide in the future.

Data Availability Statement

All the data used in this work have been previously published, and references are provided in the paper. Dynamic rupture simulations were generated using the open-source software package PyLith, freely available at https://github.com/geodynamics/pylith. Locking model data may be found in the cited papers. All the important scripts and outputs are accessible in the temporal link https://drive.google.com/drive/folders/1sGSr1tuyvgm_kHL9mslxsHzwlLw3L2XX?usp=sharing during peer review. Upon acceptance of the manuscript, the data would be available in the CUHK Research Data Repository.

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Journal of Geophysical Research: Solid Earth

Supporting Information for

Dynamic Rupture Scenarios of the Cascadia Megathrust based on Interseismic Locking Models

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Contents of this file

Introduction

This supporting document explains the parameters used in the study and extends the results of the dynamic simulations. Specifically, Figure S1 shows the velocity models of the continental and oceanic blocks of Cascadia. Figure S2 is the comparison between dynamic simulations using bi-material (continental and oceanic) and 1D depth-dependent (only continental) material properties, illustrating the little difference between these two material assumptions. Figure S3 shows the dynamic simulation results assuming different thicknesses for the frontal prism. Figure S4 demonstrates the location of the sediment deposition at different depths, justifying our choice of the frontal prism depth range from 5 – 7 km. Figure S5 shows the moment magnitude of dynamic simulations initiated from different hypocenter locations with 10 km and 15 km radius nucleation zones, supporting our choice of 15 km radius nucleation zones. Figure S6 extends the Schmalzle-based dynamic simulations to a *dc* of 1 m.

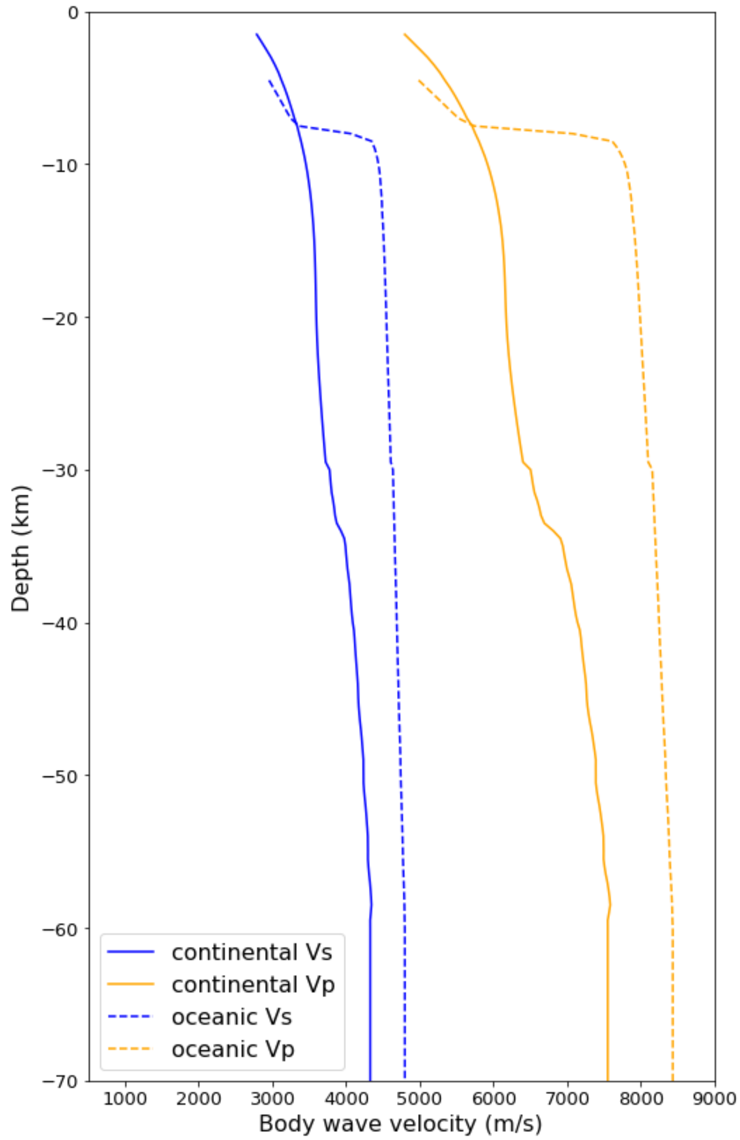

Figure S1. Velocity models of continental and oceanic blocks for Cascadia (Stephenson et al., 2017).

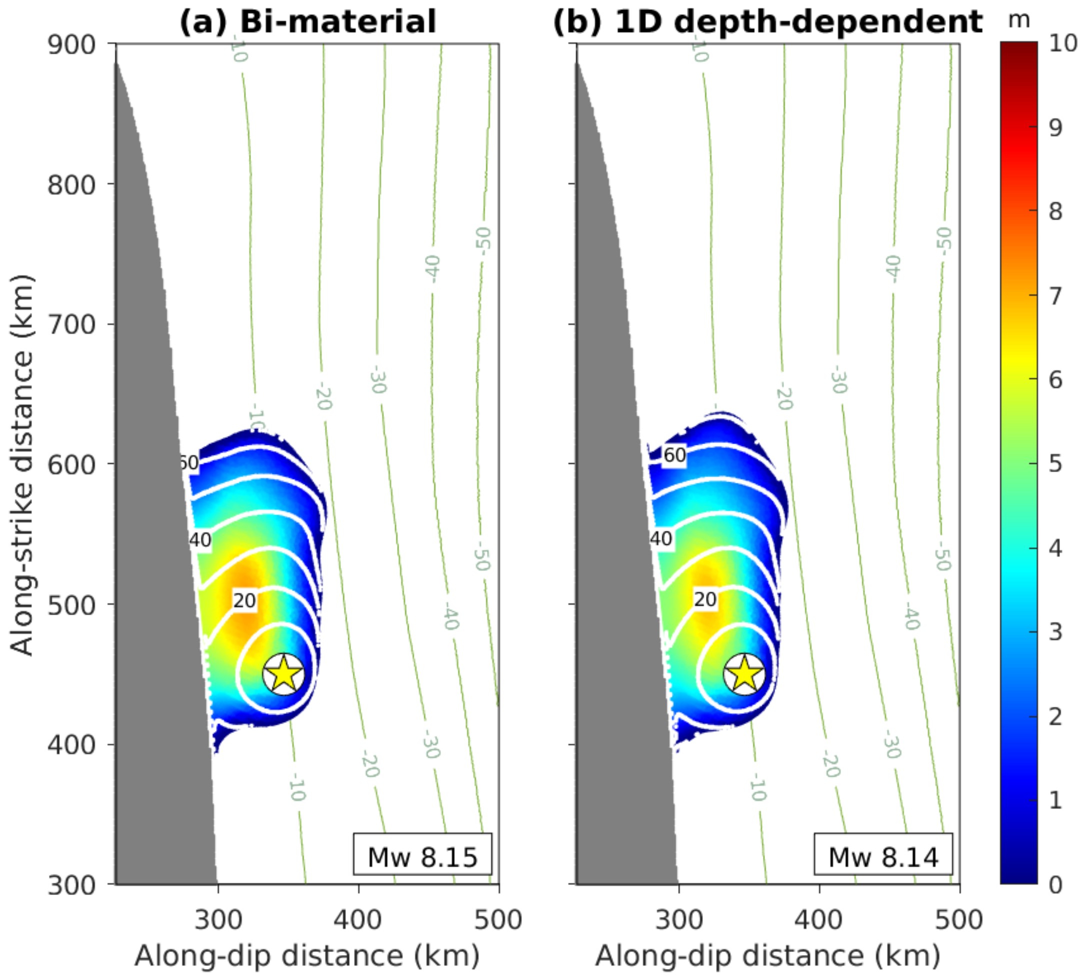

Figure S2. Final slip distribution of the same stress field utilizing (a) bi-material materials properties where the velocity contrast across two blocks is around 30%, and (a) 1D depth-dependent material properties (Figure 3a).

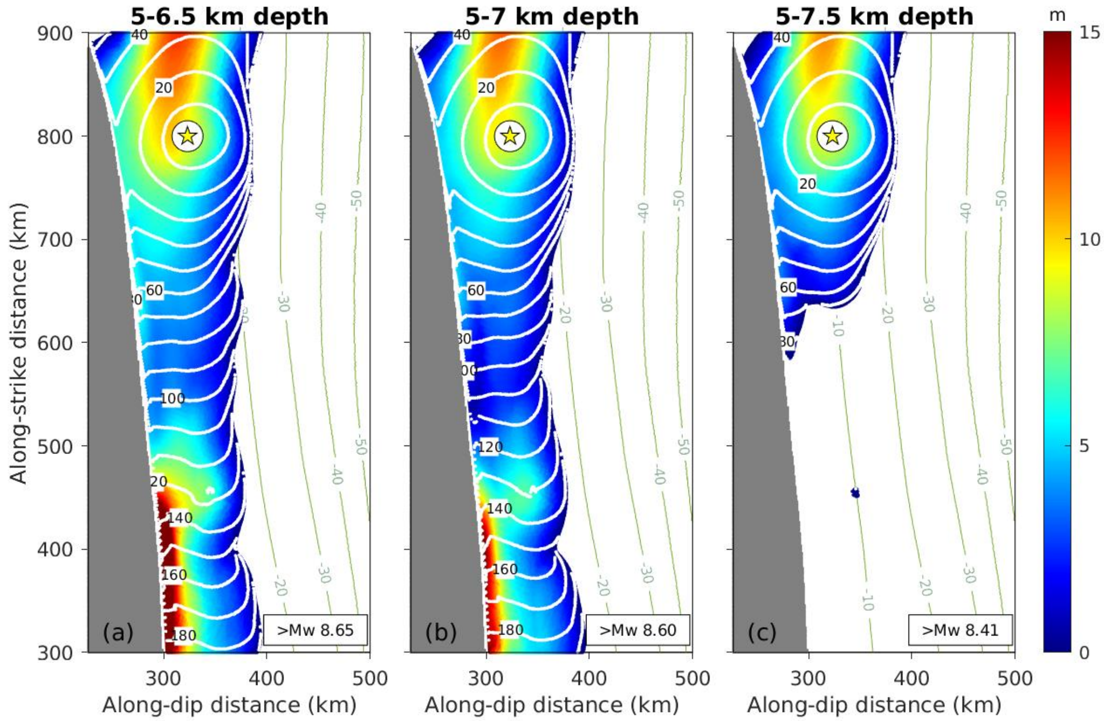

Figure S3. Final slip distribution derived from the Schmalzle model using a *dc* of 1 m assuming different thicknesses of the frontal prism. Stars: hypocenter locations. Olive-green contours: slab depth contours. Rupture fronts (white contours) are displayed every 10 seconds and numbered every 20 seconds. (a) Accretionary prism from 6.5 km depth to trench (5 km depth). Strong free-surface reflections are generated near the trench, causing an unphysical final slip that exceeds the slip deficit (b) From 5-7 km depth - our model assumption. (c) From 5 – 7.5 km depth.

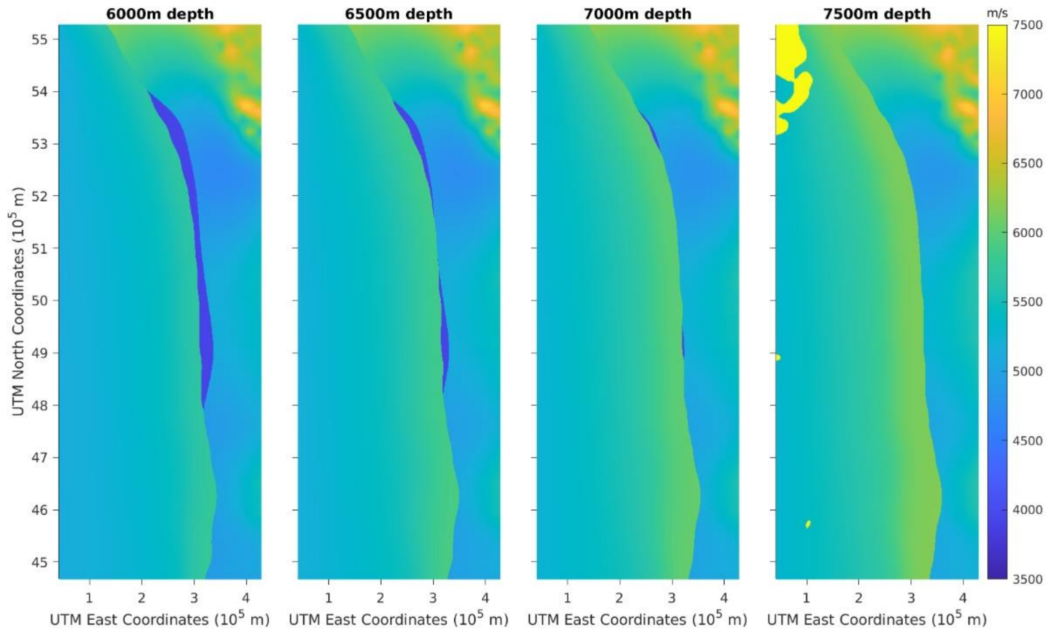

Figure S4. P wave velocity cross-sections of CSZ at different depths (Stephenson et al., 2017). The sediment deposition / accretionary prism is shown as the central gap with low velocity.

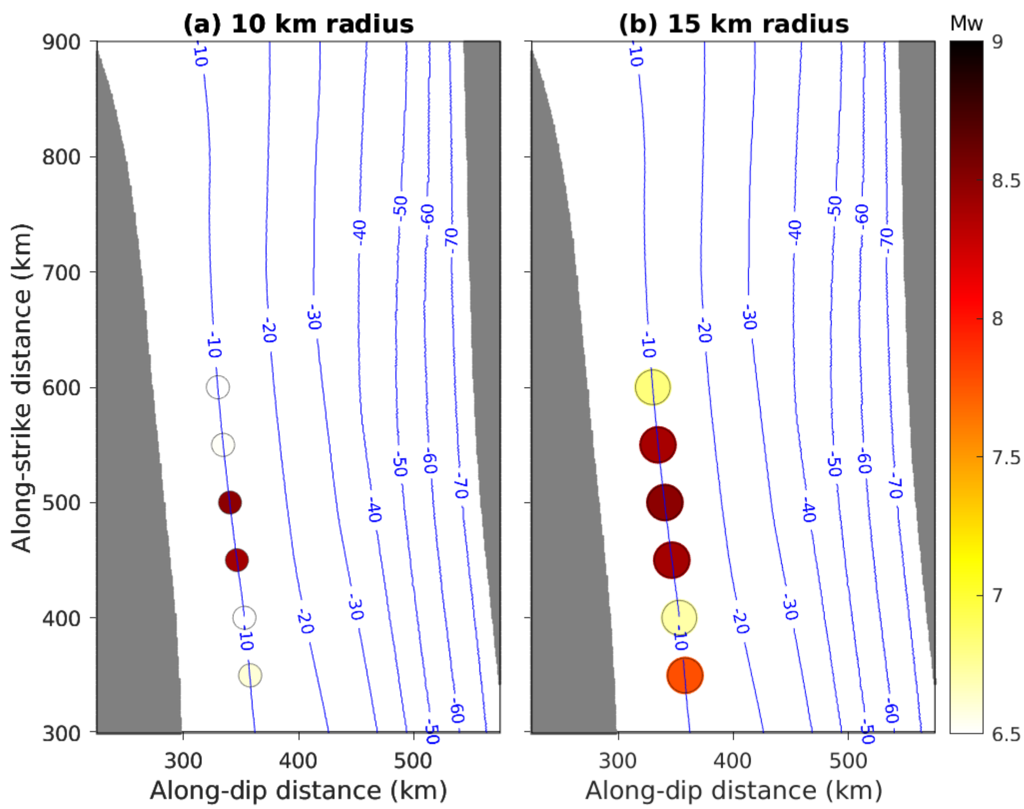

Figure S5. Map view of the moment magnitudes of rupture scenarios nucleated at each location (circles) using (a) 10 km radius and (b) 15 km radius nucleation zones.

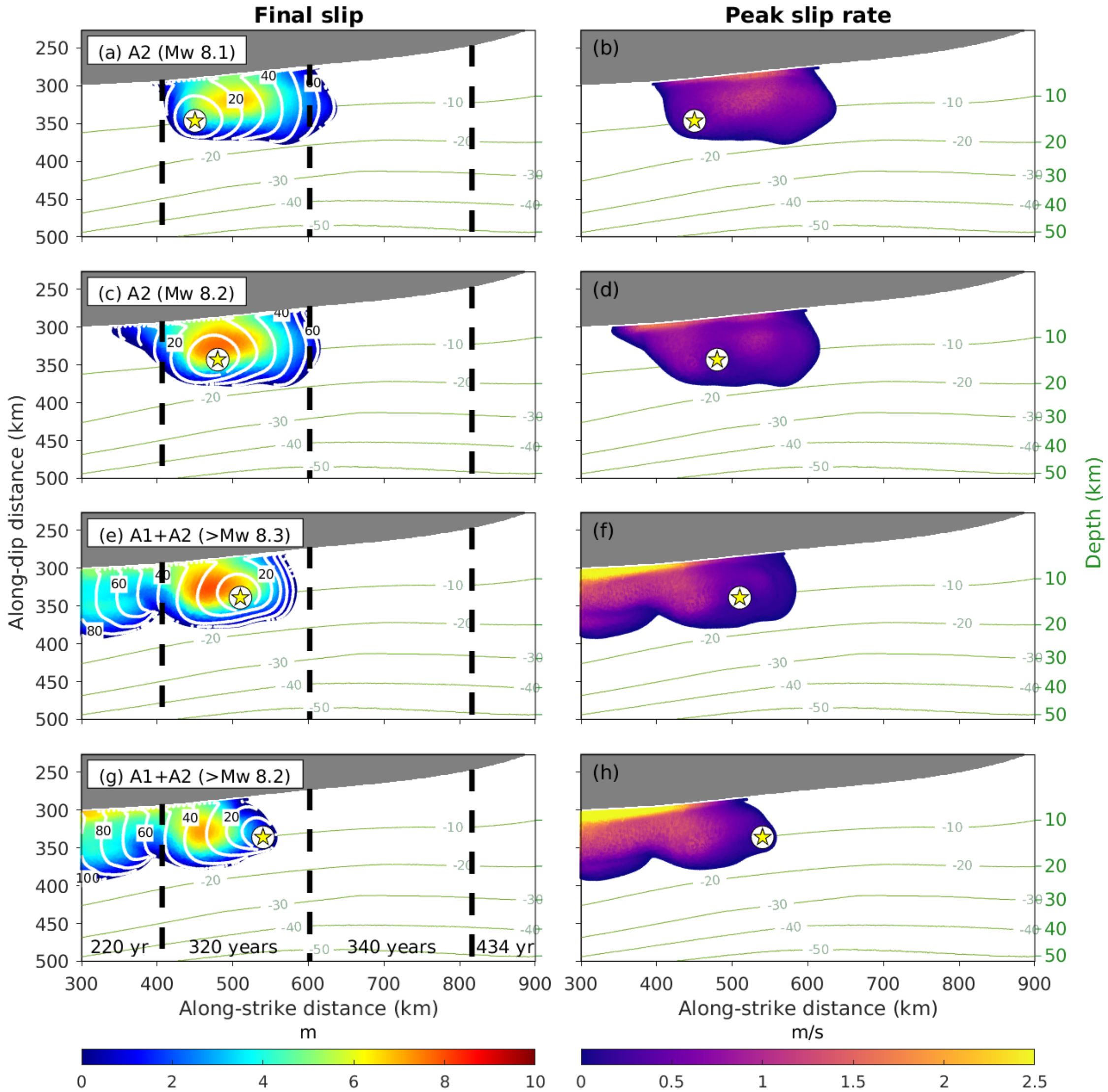

Figure S6. Same as Figure 6, except using a *dc* of 1 m. (a) & (b) Scenario rupturing the A1 asperity. (c) & (d) Scenario rupturing A1 and A2 asperities.

Supplementary Movies: Slip rate evolution in dynamic rupture scenarios
S1: Schmalzle model–FMR initiated fromA1(Figure 6e)
S2: Schmalzle model–FMR initiated from A2 (Figure 6g)
S3: Li model–FMR initiated at 15 km depth (Figure 7c)
S4: Li model–FMR initiated at 10 km depth (Figure 7e)
S5: Schmalzle model–A1+A2+C event (Figure 6c)
S6: Schmalzle model–A1 event (Figure 6a)
S7: Li model–segmented rupture (Figure 7a)