

Fault volume digital twin to reproduce the full slip spectrum, scaling and statistical laws

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Abstract

Seismological and geodetic observations of fault zones reveal diverse slip dynamics, scaling, and statistical laws. Existing mechanisms explain some but not all of these behaviors. We show that incorporating an off-fault damage zone—characterized by distributed fractures surrounding a main fault—can reproduce many key features observed in seismic and geodetic data. We model a 2D shear fault zone in which off-fault cracks follow power-law size and density distributions, and are oriented either optimally or parallel to the main fault. All fractures follow rate-and-state friction with parameters enabling slip instabilities. We do not introduce spatial heterogeneities in frictional properties. Using quasi-dynamic boundary integral simulations accelerated by hierarchical matrices, we simulate slip dynamics and analyze events produced both on and off the main fault. Despite spatially uniform frictional properties, we observe a natural continuum from slow to fast ruptures, as seen in nature. Our simulations reproduce the Omori law, inverse Omori law, Gutenberg-Richter scaling, and moment-duration scaling. We observe seismicity localizing toward the main fault before nucleation of main-fault events. During slow slip events, off-fault seismicity migrates in patterns resembling fluid diffusion fronts, despite the absence of fluids. We show that tremors, Very Low Frequency Earthquakes, Low Frequency Earthquakes, Slow Slip Events, and earthquakes can all emerge naturally within this fault volume framework, making it an ideal digital twin for testing hypotheses, performing ground-truth inversions, and probing mechanical properties inaccessible with natural observations.

Plain Language Summary

Earthquake faults exhibit complex behavior ranging from slow creeping movements to fast destructive ruptures, but the physical mechanisms underlying this remain unclear. We investigate whether the geometric arrangement of smaller fractures around a main fault—called the “damage zone”—can explain these diverse slip behaviors. We develop simulations that incorporate both a rough main fault and surrounding

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networks of fractures, all governed by the same frictional properties. We find that geometric complexity alone reproduces the full spectrum of fault slip observed in nature—from barely detectable slow slip events to regular earthquakes—without requiring spatially varying friction. Our simulations naturally generate realistic aftershock sequences following Omori’s law, earthquake size-frequency distributions matching the Gutenberg-Richter relation, spatiotemporal clustering of seismicity, and complex source time functions resembling tectonic tremor. This fault volume model serves as a digital twin that can generate physically consistent synthetic earthquake catalogs for training machine learning algorithms, testing how faults respond to external perturbations like tidal stresses or fluid injection, and developing improved methods for interpreting seismic and geodetic observations. Our results demonstrate that the network of small fractures surrounding major faults provides a fundamental organizing principle for fault dynamics, offering a geometric foundation upon which frictional and hydraulic effects can be evaluated.

Key Points

- A fault volume model reproduces the full spectrum of slip behaviors—from slow slip events to fast earthquakes—using geometric complexity alone.
- Damage zone fractures surrounding a rough fault generate realistic aftershock sequences, foreshock activity, and spatiotemporal clustering without imposed frictional heterogeneity.
- The model naturally recovers empirical scaling laws including Gutenberg-Richter statistics, Omori decay, moment-duration scaling, and source-time-function complexity.
- Geometric complexity provides a foundational baseline for fault slip dynamics, enabling a digital twin framework for testing physical mechanisms and developing inverse methods.

1 Introduction

In the brittle upper crust, fault zones display a large variety of slip dynamics and moment release. Until the discovery of slow slip events, SSEs, (*Hirose et al. 1999; Dragert et al. 2001*) and tremor (*Obara 2002*), faults were thought to either remain locked during the interseismic period or continually creep—the former leading to stick-slip-like release of stored strain energy (*Brace & Byerlee 1966*), the latter to continuous strain release (*Steinbrugge et al. 1960*). Modern seismological and geodetic observations have significantly advanced our understanding of fault slip behavior, revealing a continuum of deformation modes spanning from regular fast earthquakes to a diverse family of slow earthquakes and/or steady creep. Slow earthquakes include “seismic” members such as low-frequency earthquakes (LFEs), tectonic tremor, and very low-frequency earthquakes (VLFs), as well as “geodetic” members—slow slip events (SSEs)—typically classified as short-term (days to weeks) or long-term (months to years) (*Nishikawa et al. 2023*). These phenomena are increasingly recognized as different manifestations of a shared underlying physical process (*Beroza & Ide 2011; Ide & Beroza 2023*). SSEs, characterized by the gradual release of tectonic stress, have been observed in numerous subduction zones, including Cascadia, Central Ecuador, Guerrero, Hikurangi, Northern Chile, and southwest Japan, as well as along continental plate boundaries such as the San Andreas fault system in California, North Anatolian fault system, Haiyuan fault (*Lowry et al. 2001; Dragert et al. 2001; Rogers & Dragert 2003; Douglas et al. 2005; Jolivet et al. 2013; Vallée et al. 2013; Ruiz et al. 2014; Rousset et al. 2016; Shelly 2017; Michel et al. 2019; Dal Zilio et al. 2020*). These events often co-occur with LFEs and tremor, forming Episodic Tremor and Slip (ETS) sequences, particularly well documented in Cascadia and Nankai (*Rogers & Dragert 2003; Michel et al. 2019*). Since their discovery, observations and models have emphasized the potential role of SSEs in the earthquake cycle. Occurring near or

within the seismogenic zone, SSEs are thought to influence the initiation or modulation of large earthquakes (Rogers & Dragert 2003; Segall & Bradley 2012; Obara & Kato 2016; Cruz-Atienza et al. 2021). Moreover, aseismic slip phenomena are not limited to traditionally aseismic regions. Increasingly, observations show that slow and fast slip events can coexist within the seismogenic zone itself (Schwartz & Rokosky 2007; Ito et al. 2013; Ruiz et al. 2014; Thomas et al. 2017). Aseismic slip has notably been observed prior to several major earthquakes, including the 2011 Tōhoku-Oki (Ito et al. 2013), 2014 Iquique (Ruiz et al. 2014), and 2017 Valparaíso earthquakes (Ruiz et al. 2017; Caballero et al. 2021). Postseismic aseismic slip in seismic patches has also been documented in several subduction/collision zones (Johnson et al. 2012; Thomas et al. 2017; Villafuerte et al. 2025), further highlighting the complex interaction between seismic and aseismic processes.

Although slip dynamics may appear highly complex across timescales, robust empirical laws continue to emerge from seismological observations of fast events. The Gutenberg-Richter law describes the magnitude-frequency distribution of earthquakes, following a power law with a b-value near 1 (Gutenberg & Richter 1942). The Omori law characterizes the decay rate of aftershocks following a mainshock (Utsu et al. 1995), while an inverse Omori law describes the acceleration of foreshock activity before major events, first noted by Papazachos (1973). For fast ruptures, seismic moment scales cubically with rupture duration (Ide et al. 2008; Ide & Beroza 2023). For slow ruptures, debate remains between a linear scaling (Ide et al. 2008) and a cubic one (Gomberg et al. 2016; Michel et al. 2019). More recently, Kato & Ben-Zion (2020) reported a phenomenon of localization followed by delocalization of deformation before and after major earthquakes, respectively—a pattern observed in southern California for several large events, including the 1992 M_w 7.3 Landers, 1999 M_w 7.1 Hector Mine, and 2019 M_w 7.1 Ridgecrest earthquakes (Ben-Zion & Zaliapin 2020). Furthermore, seismicity has been observed to migrate during slow-slip events, with spatiotemporal features resembling a diffusion front, suggesting that fluid diffusion may govern these dynamics (e.g. Danré et al. 2024).

The wide spectrum of observed slip behaviors has prompted investigation into various physical mechanisms, with fault friction heterogeneity emerging as the most commonly invoked (Brace & Byerlee 1966; Scholz 2019). Laboratory experiments have consistently shown that frictional properties are central to the transition from stable to unstable slip, particularly near the brittle-ductile transition zone (e.g. Leeman et al. 2016; Tinti et al. 2016; Scuderi et al. 2016; Scuderi et al. 2017; Leeman et al. 2018; Sirorattanakul 2024; Pignalberi et al. 2024; Yuan et al. 2024; Meyer et al. 2024; Salazar Vásquez et al. 2024). Numerical simulations incorporating variable constitutive friction parameters within rate-and-state frameworks have successfully reproduced a spectrum of slip modes, including slow ruptures (Yoshida & Kato 2003; Liu & Rice 2005; Barbot 2019; Nie & Barbot 2021). Modeling and experimental studies indicate that frictional heterogeneity not only modulates slip dynamics and triggering (Aochi & Ide 2009; Dublanche et al. 2013) but also governs complex precursory processes that may culminate in a mainshock (Kato et al. 1997; Ariyoshi et al. 2012; Dublanche 2017; Gounon et al. 2022; Wang et al. 2023b). Other mechanisms have also been proposed, including fluid-related effects such as dilatant strengthening (Segall & Rice 1995; Segall et al. 2010; Liu & Rubin 2010), spatiotemporal variations in permeability and pore fluid pressure (Skarbak & Rempel 2016; Cruz-Atienza et al. 2018; Zhu et al. 2020; Perez-Silva et al. 2023; Ozawa et al. 2024), elastic or poroelastic bimaterial effects (Heimisson et al. 2019; Abdelmeguid & Elbanna 2022), brittle patches embedded in ductile matrices (Ando et al. 2012), and transient stress perturbations due to nearby earthquakes (Liu & Rice 2007) or failed nucleation

processes (*Rubin 2008*). Notably, the earliest model of slow slip events was proposed by *Perfettini et al. (2001)*, who showed that a spring-block system obeying rate-and-state friction and subjected to normal traction perturbations could, under certain conditions, exhibit “aseismic stick-slip” behavior—prior to the observational discovery of SSEs in Cascadia and the Bungo Channel. All of the above mechanisms primarily focus on modeling the fault as a single frictional surface.

However, faults are far more complex than simple, narrow shear zones idealized as frictional surfaces. Geological observations show that, surrounding the fine-grained fault core—where most slip localizes—is a broader damage zone composed of pervasively fractured rock with an intricate three-dimensional geometry across multiple length scales (*Sibson 1977; Chester et al. 1993; Biegel & Sammis 2004*). This hierarchical geometric complexity ranges from tens of kilometers (*see, for e.g., Fletcher et al. 2014*) down to the millimetric scale (*Sowers et al. 1994; Mitchell & Faulkner 2009; Fagereng et al. 2010*). Figure 1 of *Okubo et al. (2019)* illustrates this hierarchical structure of fault systems over a wide range of length scales. Fault slip typically occurs on a localized plane—the fault core—which lies within a damage zone that generally extends several hundreds of meters in width (*Chester & Logan 1986; Sibson 1986; Power et al. 1987; Ben-Zion & Sammis 2003; Sibson 2003; Savage & Brodsky 2011; Ostermeijer et al. 2020; Rodriguez Padilla et al. 2022; Liu et al. 2025*). Various studies have shown that fault geometry—such as roughness (*Candela et al. 2011; Cattania & Segall 2021; Ozawa & Ando 2021*), bends (*King & Nábělek 1985; Ritz et al. 2015; Romanet et al. 2020; Ozawa et al. 2023*), branches (*Aochi et al. 2000; Kame & Yamashita 2003; Kame et al. 2003; Oglesby 2003; Bhat 2004; Bhat et al. 2007b; Marschall & Douilly 2024; Templeton et al. 2010*), and step-overs (*Oglesby 2005; Biasi & Wesnousky 2016; Rodriguez Padilla et al. 2024*), play an important role in the dynamics of earthquake ruptures. Recent work has also underlined the key role of active faults networks (*Im & Avouac 2023; Cheng et al. 2025*). Large earthquakes ($M_w > 7$) often exhibit complex ruptures involving multiple faults (*Stein & Bird 2024*). For example, the 1992 Landers earthquake involved the activation of multiple faults during a single seismic event (*Sowers et al. 1994; Fliss et al. 2005*), and the 2016 Kaikōura earthquake involved at least 15 faults (*Klinger et al. 2018*). More recently, fault system geometry was found to significantly influence the slip distribution of the M_w 7.1 Ridgecrest earthquake (*Nevitt et al. 2023*) and the 2023 M_w 7.8 Kahramanmaraş, Turkey, earthquake (*Chen et al. 2024; Yao & Yang 2025*). Furthermore, recent work by *Lee et al. (2024)* suggests that fault slip stability may be controlled by the orientation and complexity of surrounding fault networks, with complex fault systems associated with locked segments that promote stick-slip behavior, whereas simpler geometries favor stable creep.

Several studies have also focused on understanding the role of newly created fracture - or dynamic damage - in the propagation of a single dynamic rupture (*Xu et al. 2015; Bhat et al. 2012, among others*). This damage can significantly affect the high-frequency radiation generated during rupture propagation (*Thomas & Bhat 2018; Okubo et al. 2019; Marty et al. 2019; Okubo et al. 2020*). Recent findings highlight radiation from such multifracture structures in subduction zones (*Chalumeau et al. 2024*), and the anisotropy imposed by the preferred orientation of off-fault fractures (*Huang et al. 2025*). Moreover, as a dynamic rupture propagates, the energy dissipated by off-fault fracturing processes (creation of new fractures or reactivation of preexisting ones) becomes critical to consider (*Okubo et al. 2019*). The fracture energy dissipated within the off-fault volume can be substantial—often comparable to the fracture energy on the main fault itself (*Andrews 2005; Okubo et al. 2019; Okubo et al. 2020*). Measurements of off-fault inelastic deformation using radar and optical imagery, when compared with aftershock sequences, indicate

that fault systems exhibiting greater off-fault damage tend to produce relatively higher numbers of aftershocks (*Milliner et al. 2025*). Additionally, dynamic damage has been shown to play a crucial role in controlling the transition to supershear rupture (*Jara et al. 2021*).

Thus, while fault slip behavior is strongly influenced by the properties of the interface itself, the overall mechanical behavior of faults is equally shaped by the structural complexity of the entire fault zone. Its physical properties evolve over timescales ranging from seconds to millions of years, accommodating displacements from millimeters to tens of kilometers. In particular, dynamic ruptures can induce significant changes in both on-fault and off-fault mechanical properties. These changes, in turn, influence rupture nucleation, propagation, timing, seismic wave radiation, and postseismic deformation. This broader structure alters the rheology of both the fault core and the surrounding rock, introduces complex interactions, and thereby affects the style and dynamics of fault slip (*e.g. Andrews 2005; Collettini et al. 2009; Niemeijer et al. 2010; Thomas et al. 2014; Faulkner et al. 2006; Dor et al. 2006b; Dor et al. 2006a; Mitchell & Faulkner 2009; Bhat et al. 2010; Biegel et al. 2010; Bhat et al. 2012; Okubo et al. 2019; Okubo et al. 2020*). Understanding such processes inevitably calls for a comprehensive approach that considers the fault zone as a whole, rather than focusing solely on the fault core.

Keeping this in mind, recent studies increasingly focus on understanding the role of complex and irregular geometries throughout the full seismic cycle. *Romanet et al. (2018)* incorporated geometrical complexity in numerical models and found that stress interactions between parallel faults can naturally generate slow slip events. Similarly, *Cattania & Segall (2021)* showed that foreshock sequences can emerge spontaneously when rough fault surfaces are considered. In a related study, *Ozawa & Ando (2021)* modeled a fault network with subsidiary fractures surrounding a primary rough fault and demonstrated the spontaneous emergence of an Omori-like aftershock decay (*Utsu et al. 1995*). Incorporating the realistic geometry of northern Cascadia into numerical models enabled the generation of slow slip events that closely match GPS observations in the region (*Li & Liu 2016*). Although an increasing number of studies (*Yin et al. 2023; Im & Avouac 2024; Zhai et al. 2025*) explore the effects of geometry over multiple seismic cycles, no current model reproduces the full range of observed slip dynamics.

We aim to investigate the role of this realistic fault geometry and its associated off-fault damage—hereafter referred to as the fault volume—in governing slip event dynamics. We propose a simplified model of a 2D fault volume consisting of a main self-similar rough fault surrounded by a hierarchy of off-fault slip planes. All fractures are frictionally homogeneous (rate-weakening) and capable of dynamic slip. To enable simulation of this geometrically complex system, we develop a fast quasi-dynamic earthquake cycle model accelerated using hierarchical matrices. Hierarchical matrices compress the stress interaction matrix by approximating distant interactions with low-rank blocks, enabling simulations of large, complex systems within reasonable computational time (see Supporting Information for more details). Remarkably, this purely geometric complexity is sufficient to reproduce the full spectrum of observed slip behaviors. The model can generate a continuum of slip events and recover all major empirical scaling laws associated with the seismic cycle. This establishes the model framework as a “digital twin” of a fault zone where hypotheses about fault slip mechanisms can be tested systematically, inversion techniques can be validated against known ground-truth, and fundamental mechanical properties—such as stress evolution, energy partitioning, and damage zone dynamics—can be probed directly in ways that are impossible with natural observations.

In the following section, we introduce the fault volume model, stress loading, and initial conditions used in our study. We then outline the methodology for event detection, distinguishing between fast and slow ruptures, and describe our approach for constructing the synthetic earthquake catalog. In the results section, we begin with a case study illustrating the spatiotemporal evolution of seismic cycles and the emergence of complex slip dynamics. We analyze the moment rate functions of events across scales, revealing distinct characteristics between slow and fast ruptures. We then demonstrate the model’s ability to reproduce key empirical laws, including the Gutenberg-Richter relation, moment-rupture area scaling, moment-duration scaling, and both Omori and inverse Omori laws. Additionally, we highlight the model’s capacity to replicate the localization-delocalization transition of deformation and the apparent diffusion-like migration of seismicity during slow slip events. We conclude by discussing the model’s implications and limitations, identifying testable hypotheses, and evaluating its potential as a digital twin framework for fault systems.

2 Fault Volume Model

2.1 Overview of the Fault Volume Model

This study focuses primarily on the impact of the geometry and architecture of a fault zone on slip dynamics. Hence, this fault volume model is a simplified representation of a fault and its damage zone. We begin by defining the elastic properties of the medium, its loading and initial conditions. All length scales are non-dimensionalized using frictional length scales (either the nucleation length or the cohesive zone size) computed using the initial stress state of the medium. We remark that these frictional length scales (see Supporting Information for more details) are merely approximations, as they are based on assumptions of a single planar fault undergoing no change in normal and shear traction over time and always at steady state. We then establish the frictional properties of the main fault and set its length to four times the nucleation length. Next, we define a damage zone with a width, W , five times the cohesive zone size of the main fault. Within this damage zone, we assume a hierarchical distribution of off-fault fractures. The density of these fractures decreases according to a power law with a fixed exponent and maximum value. The size distribution also follows a power law with a fixed exponent, where the smallest length scale is approximately 1m and the largest is on the order of the main fault length. We select the orientation distribution of these fractures from one of four possible orientations with respect to the main fault and principal stress directions (Section 2.2). We assume each off-fault fracture is twice its nucleation length to determine the critical slip distance, d_c , for each fracture. This ensures d_c scales automatically with fault length as described above. Finally, we statistically sample five times from the distributions (length, orientation, and density) to generate five samples of the fault volume model for a given orientation of the off-fault fractures and its rate and state friction properties. We thus generate a total of 40 different fault volume models for this study each of them including around 700-1000 faults. In the following sections we describe each of the steps in more detail.

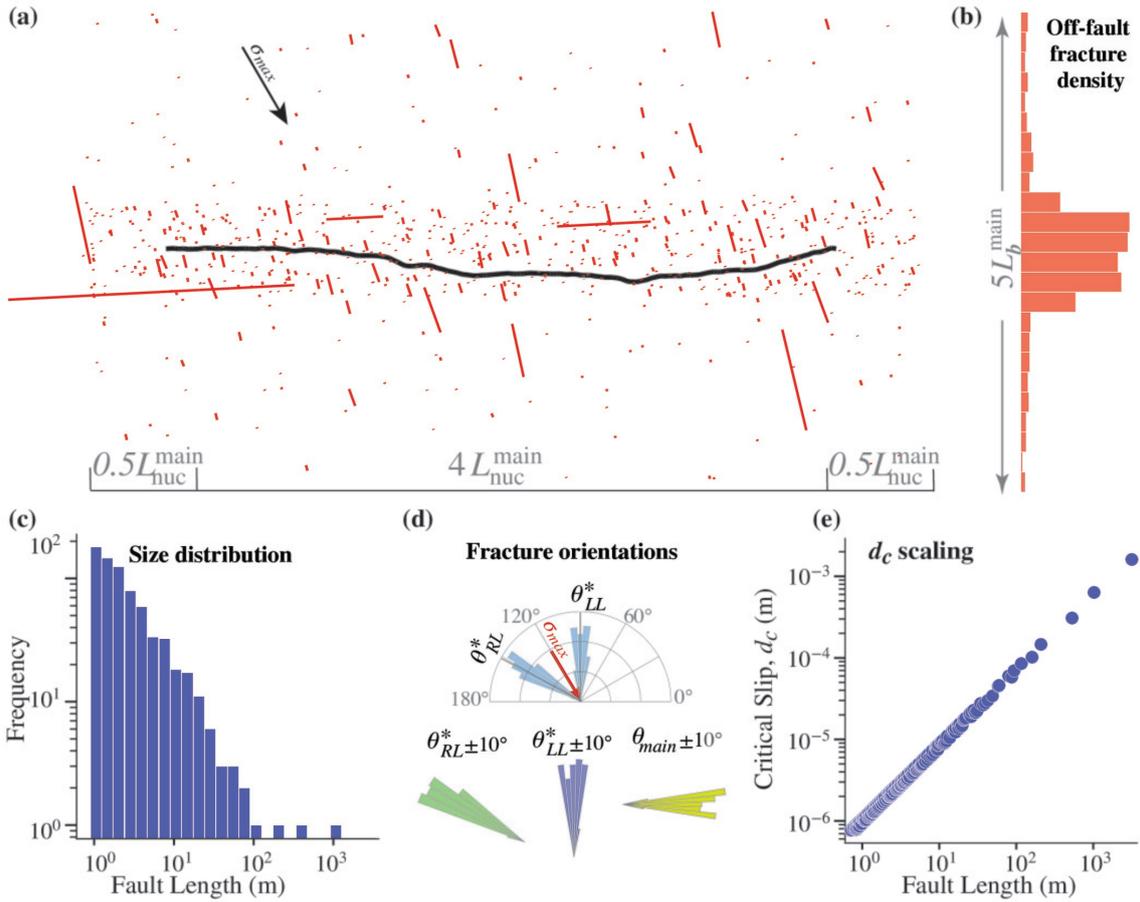


Figure 1: Fault volume geometry of a case study: a) sketch of fault volume geometry (not to scale): the main fault in black and the off-fault fractures in red b) off-fault fracture density perpendicular to the main fault (linear scale) c) Frequency of length distribution of the faults, including the main fault d) rose diagram of off-fault fracture orientations and principal stress direction e) critical slip distance of all fractures with respect to their length. RL and LL stands for right and left lateral slip respectively.

2.2 Geometry of the Main Fault and its Damage Zone

The fault volume in the model comprises a main fault exhibiting right-lateral movement in an isotropic linear elastic medium ($\mu = 30$ GPa, $\nu = 0.25$), with fixed frictional properties, accompanied by off-fault fractures within a damage zone surrounding the main fault. Since geometric roughness is a characteristic observed at multiple scales in natural faults (*Power et al. 1987; Schmittbuhl et al. 1993; Lee & Bruhn 1996; Renard et al. 2006; Candela et al. 2009; Candela et al. 2012*), in this fault volume model, the main fault is characterized by self-similar roughness with $\alpha = 0.002$ (*Dunham et al. 2011*). Its extent is four times its nucleation length $L_{\text{main}} = 4L_{\text{nuc}}^{\text{main}}$ (Figure 1a). Note that the nucleation length is just an estimation, calculated for a single planar infinite fault, following *Rubin & Ampuero (2005) and Viesca (2016)* (See Supporting Information for more details). We set the damage zone width as $W = 5L_b^{\text{main}}$ making it proportional to the main fault's cohesive zone size (L_b^{main}). This choice reflects the fact that the cohesive zone size governs the decay rate of stress perturbations from the main fault, which in turn controls the characteristic length scale over which main and off-fault fractures interact (*Okubo et al. 2019*). We also add a fracture damage zone at each end of the main fault whose fracture density is assumed to be uniformly distributed.

From natural observations, the length distribution of off-fault fractures follows a power law (*Bon-*

net et al. 2001; Ben-Zion & Sammis 2003). This modeling choice is grounded in the understanding that such distributions effectively capture the natural variability of fault lengths. Accordingly, we consider a power law distribution of fractures length with an exponent of 1 (Figure 1c). The smallest length scale is set to be around 1 m and the largest being of the order of the length of the main fault. Each off-fault fracture is set to be twice its nucleation length, ensuring it is large enough to rupture independently (*Rubin & Ampuero 2005; Viesca 2016*).

We next set the orientation statistics of the off-fault fractures. We experiment with four different orientations of fractures. In the first three cases, the fractures are deliberately well oriented to maximize reactivation and thus interaction with the main fault. We examine favorable orientations for right-lateral failure, left-lateral failure, and a combination of both (see example in Figure 1d). Optimal planes make an angle of $\theta = \pm (\pi/4 - 1/2 \tan^{-1} f_0)$ with the direction of maximum principal stress (*Bhat et al. 2007a*), where f_0 is the reference friction coefficient. Here $f_0 = 0.6$, so $\theta \approx \pm 30^\circ$. The initial maximum principal stress makes an angle of 120° with the horizontal axis. We thus consider right-lateral fractures (oriented at $\theta = 150^\circ \pm 10^\circ$), left-lateral fractures (oriented at $\theta = 90^\circ \pm 10^\circ$), both right and left-lateral fractures (conjugate planes at $\theta = [90^\circ; 150^\circ] \pm 10^\circ$). For the final case, the fractures run parallel to the main fault, as it has been observed that the growth process of faults reveals that off-fault cracks tend to be oriented at angles of $[0^\circ - 20^\circ]$ to the trace of major faults (*Perrin et al. 2015*). For this case, we consider the fracture orientations in the range $[0^\circ \pm 10^\circ]$ (Figure 1d).

Observations from fault zones consistently show a significant decrease in the density of off-fault cracks with increasing distance from the main fault. This observation has been well documented in fault zones (*Chester & Logan 1986; Faulkner et al. 2006; Powers & Jordan 2010; Rodriguez Padilla et al. 2022*). Based on these observations of damage zones surrounding major faults, we model a power-law decay of fracture density with distance normal to the main fault, using an exponent of 1 (Figure 1b shows the sampled distribution). This exponent is higher than values reported in some studies (*Ostermeijer et al. 2020; Savage & Brodsky 2011*) but lower than the values suggested by *Powers & Jordan (2010)*. We fix the peak density of off-fault fractures to be around 4 fractures per square meter, which is consistent with observations (*Chester & Logan 1986; Mitchell & Faulkner 2009; Faulkner et al. 2011; Rodriguez Padilla et al. 2022*). We further assume that all of the off-fault fractures are pure mode II fractures. The main numerical bottleneck in simulating tensile off-fault fractures in an earthquake cycle model is the lack of a good cohesive law that allows for the healing of the cohesive strength after rupture. We therefore restrict ourselves to shear fractures that follow rate-and-state friction law.

2.3 Frictional Parameters

Friction is assumed to follow regularized rate-and-state friction, spatially uniform and rate-weakening on all faults with $f_0 = 0.6$ (See Supporting Information for more details). Specifically, for the main fault, the ratio a/b is fixed at 0.75 and d_c is set at 2 mm. Here, a is the direct effect parameter that governs the instantaneous change in friction with a change in slip rate, b is the evolution effect parameter which controls how friction evolves over time via changes in the state variable and d_c is the critical slip distance. Since a/b is a crucial parameter in controlling the frictional length scales (see Supporting Information for more details) we parametrize the problem by varying this parameter. All off-fault fractures within the model share the same a/b ratio, with tested values

of 0.4 and 0.5. The characteristic slip distance (d_c) is scaled with the length of off-fault fractures (Figure 1e) such that each off-fault fracture is twice the nucleation length.

Earthquake sources are conventionally modeled as shear ruptures occurring on preexisting faults within the seismogenic zone. Nevertheless, these faults are characterized by complex geometric irregularities and mechanical heterogeneities that manifest across a wide range of spatial scales throughout the fault zone. Be it in the lab scale or the field scale, the net weakening distance is a homogenized manifestation of weakening processes occurring at smaller scales (*Ohnaka & Shen 1999; Gabriel et al. 2024*). This approach is also inspired by the observation that the so called fracture energy (G') scales with slip (*Ohnaka 2003; Abercrombie & Rice 2005*), and slip is itself related to fracture length, through elasticity. Furthermore, *Rubin & Ampuero (2005)* showed that fracture energy is approximately proportional to d_c within the rate-and-state friction framework. These observations together justify scaling d_c with fracture length in our model. This also has the additional advantage of keeping the computational cost reasonable. To keep the values of d_c realistic, we ensure that the smallest fracture is around 1m, the length scale of laboratory experiments and use laboratory inferred values of d_c (*Marone 1998; Scuderi et al. 2016; Leeman et al. 2018*) of the order of several tens of microns.

2.4 Loading and Initial Conditions

The medium has an initial prestress state with $\sigma_{11}^0 = -6.46$ MPa, $\sigma_{12}^0 = 3.88$ MPa, $\sigma_{22}^0 = -10.94$ MPa. It is also loaded with uniform far-field stressing rate, $\dot{\sigma}_{11}^\infty = -0.0064$ Pa s⁻¹, $\dot{\sigma}_{12}^\infty = 0.0038$ Pa s⁻¹, $\dot{\sigma}_{22}^\infty = -0.0108$ Pa s⁻¹. This corresponds to strain rates $\dot{\epsilon}_{11}^\infty = -4.93 \times 10^{-14}$ s⁻¹, $\dot{\epsilon}_{22}^\infty = -1.23 \times 10^{-13}$ s⁻¹ and $\dot{\epsilon}_{12}^\infty = 6.33 \times 10^{-14}$ s⁻¹. The direction of principal stress for far-field stress rate and initial stress are the same, and is assumed here to make an angle of 120° with the horizontal axis. At each fault, $\tau_0 = \sigma_{ij}^0 n_j s_i$ and $\sigma_0 = \sigma_{ij}^0 n_j n_i$, where τ_0 and σ_0 are initial shear and normal stress on the fault, \mathbf{n} and \mathbf{s} are the unit normal and unit tangent vectors respectively. All faults are assumed to be initially at steady state, and initial slip rates are considered constant: $V_{\text{init}} = V_0 = 10^{-9}$ m/s where V_0 is the reference slip rate for rate-and-state friction. The system initiates through the activation of a high slip rate patch located on the main fault. We neglect the first artificial earthquake on the main fault forced by loading, and only study the subsequent cycles after the system loses the memory of the initial conditions. To avoid an unusual buildup of normal traction on the fault, we simply cap the normal traction at 10 MPa in an elastic-rigidly plastic sense. This needs to be examined in more detail in future work as it is beyond the scope of this paper. The slip rate is capped at a minimum value of 10^{-20} m/s at the end of each time step to avoid unnecessary numerical artifacts as suggested in the SEAS benchmark (*Jiang et al. 2022*).

2.5 Numerical Implementation

We implement the fault volume model using a quasi-dynamic boundary integral method with spatial convolution accelerated using hierarchical matrices. To assess the sensitivity of our fault volume model to various parameters, we conducted multiple simulations exploring different fracture orientations and frictional parameters of off-fault fractures. To ensure robustness of our results, each off-fault orientation was subjected to five different statistical samplings, guaranteeing a com-

prehensive understanding of the system’s behavior across various geometries. Furthermore, we repeated the entire analysis, initially with an a/b ratio of 0.5 for the off-fault fractures, and subsequently with a new a/b ratio of 0.4, while maintaining homogeneous, rate-weakening friction conditions. In total, we tested 40 different configurations with various combinations of off-fault fracture orientations and frictional parameters.

We account for the full elastic interaction between all faults in the system, including the main fault and all off-fault fractures. To ensure numerical accuracy, we discretize each of the faults such that at least 5 grid points are within the cohesive zone of each fault. This results in about 30000–50000 unknowns in the system per simulation. Each of the simulations is run until we produce between 8 and 10 major earthquakes on the main fault, which corresponds to between 2 and 4 million adaptive time steps. More details on the numerical implementation can be found in Supporting Information.

2.6 Event Detection and Catalog Generation

The detection and classification of events in the system involve several steps. Events are identified when the maximum slip rate exceeds predefined thresholds, with categorization into “fast” or “slow” depending on the specific threshold exceeded: 10^{-3} m/s for fast and 10^{-6} m/s for slow. Additional values for the slow rupture threshold (10^{-7} , 10^{-8} , and 10^{-9} m/s) are also tested. The effect of this threshold on the catalog and the scaling laws will be discussed in the following sections. Subsequently, events that fall within this threshold range are delineated spatiotemporally through the application of the connected-component labeling (CCL) algorithm, commonly utilized in image processing. The detailed event detection algorithm is presented in Supporting Information. This process yields information on rupture length L_{rup} and duration T , which allows us to estimate an average rupture velocity $v_r = L_{\text{rup}}/T$. We also estimate for each event the final slip δ and subsequently the seismic moment as follows: $M_0 = \mu S \delta$, where μ is the shear modulus and S is the rupture area. Since the model is 2D, we use an equivalent rupture surface S defined as $\pi L_{\text{rup}}^2/4$, assuming a circular rupture. We then estimate the moment magnitude $M_w = 0.67 \log_{10} M_0 - 6.06$ (*Hanks & Kanamori 1979*). Finally, we apply a filtering process to the detected events, eliminating those with fewer than 5 time steps and spanning less than 5 grid points so that the catalog properties can be accurately computed. An example of the generated catalog is shown in Figure 2e.

3 Influence of the Fault Volume on the Slip Dynamics of the System

3.1 Seismic Cycles and Slip Dynamics

In this section, we provide a concise summary of results based on a case study using the fault geometry shown in Figure 1.

Figure 2a depicts the time series of the moment-release rates (\dot{M}_0), with the black curve illustrating the contribution originating from the main rough fault. In contrast, the red curve represents the sum of the moment-release rates of off-fault fractures. In Figure 2a, in order to measure the impact of the fault volume on slip dynamics, we normalize time by the recurrence time of earthquakes when considering only a single rough fault, without a damage zone. Notably, for the fault volume

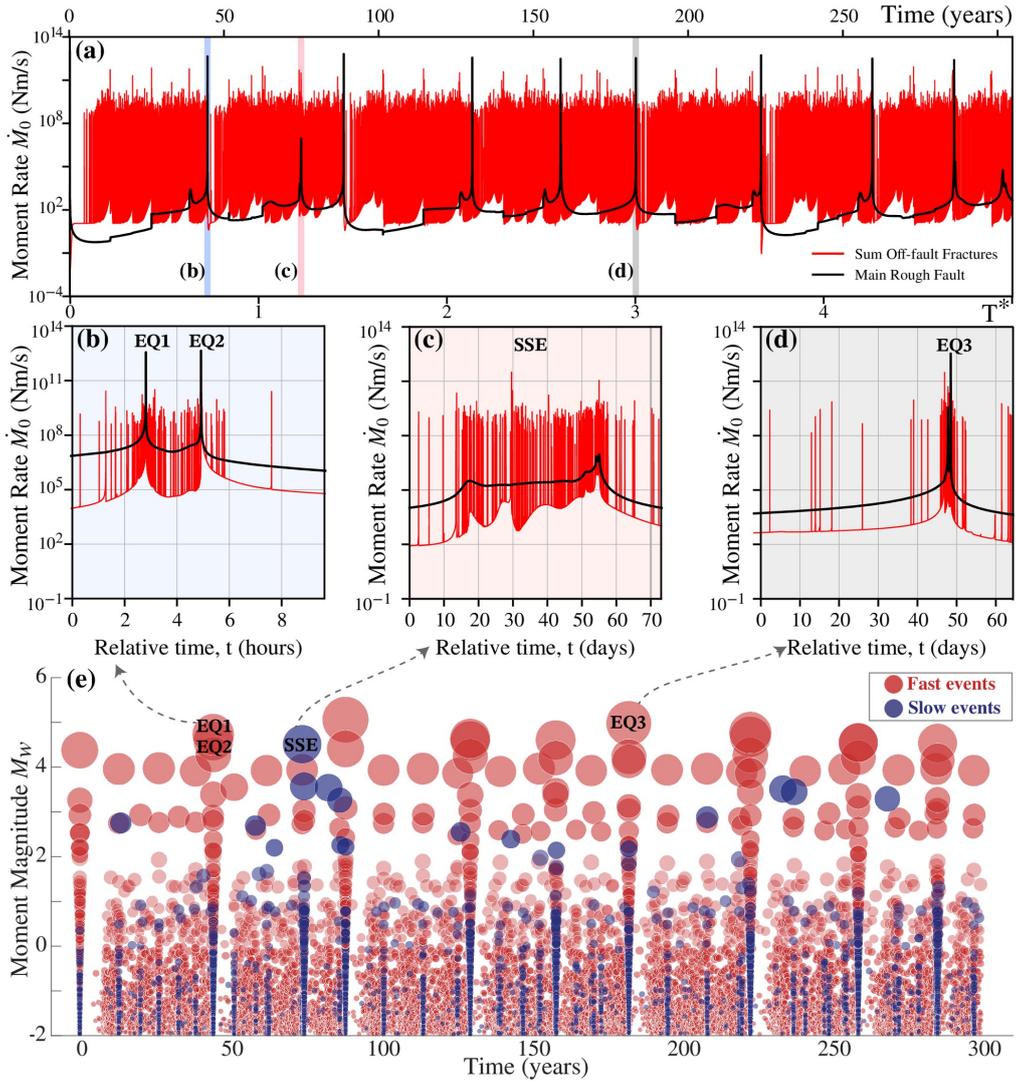


Figure 2: Time series of the moment rate \dot{M}_0 . The black curve represents the contribution coming from the main rough fault, and the red curve represents the summation of the moment rate released by all the fractures. Panel a) The entire seismic cycle. Top x-axis is absolute time in years, and bottom x-axis is time normalized by recurrence time of earthquakes when considering only the main rough fault, without a damage zone (~ 62 years). Panel b), c) and d) show time snapshots of the seismic cycle to highlight certain events. Here x-axis is relative time for the corresponding time snapshot. The spatiotemporal description of these events are shown in Figure 3. Panel e) shows the catalog built from the above continuous moment-rate results using slip velocity thresholds as described in section 2.6.

scenario presented here, we observe eight ruptures, twice as many as in the single-fault case, over the same time period (Figure 2a). This substantial difference emphasizes the significant perturbation introduced by the fault volume on the seismic behavior of the system, resulting in a discernibly shorter recurrence interval. Importantly, the recurrence time is not constant but rather exhibits variability, giving rise to intricate seismic and aseismic events. While the main fault experiences seismic cycles, individual off-fault fractures also undergo their own seismic cycles, but their collective moment release appears relatively steady throughout the main fault’s seismic cycle. We will examine this in more detail in later sections.

We observe both slow and fast earthquakes on the main fault, in comparison to the dynamics of a single main fault where only fast ruptures are observed. Figures 2b–d shed light on the complex seismic behavior driven by the presence of a fault volume. Examples include instances of multiple ruptures occurring within a brief timeframe, with intervals as short as a few hours between them

(Figure 2b). Additionally, we observe an increase in seismic activity recorded on the fractures before the rupture and a subsequent decrease afterward. This will be further discussed in the following sections. Furthermore, Figure 2c reveals the occurrence of prolonged slow slip events spanning a couple of months. These findings collectively underscore the complex nature of seismic activity influenced by the presence of a fault volume.

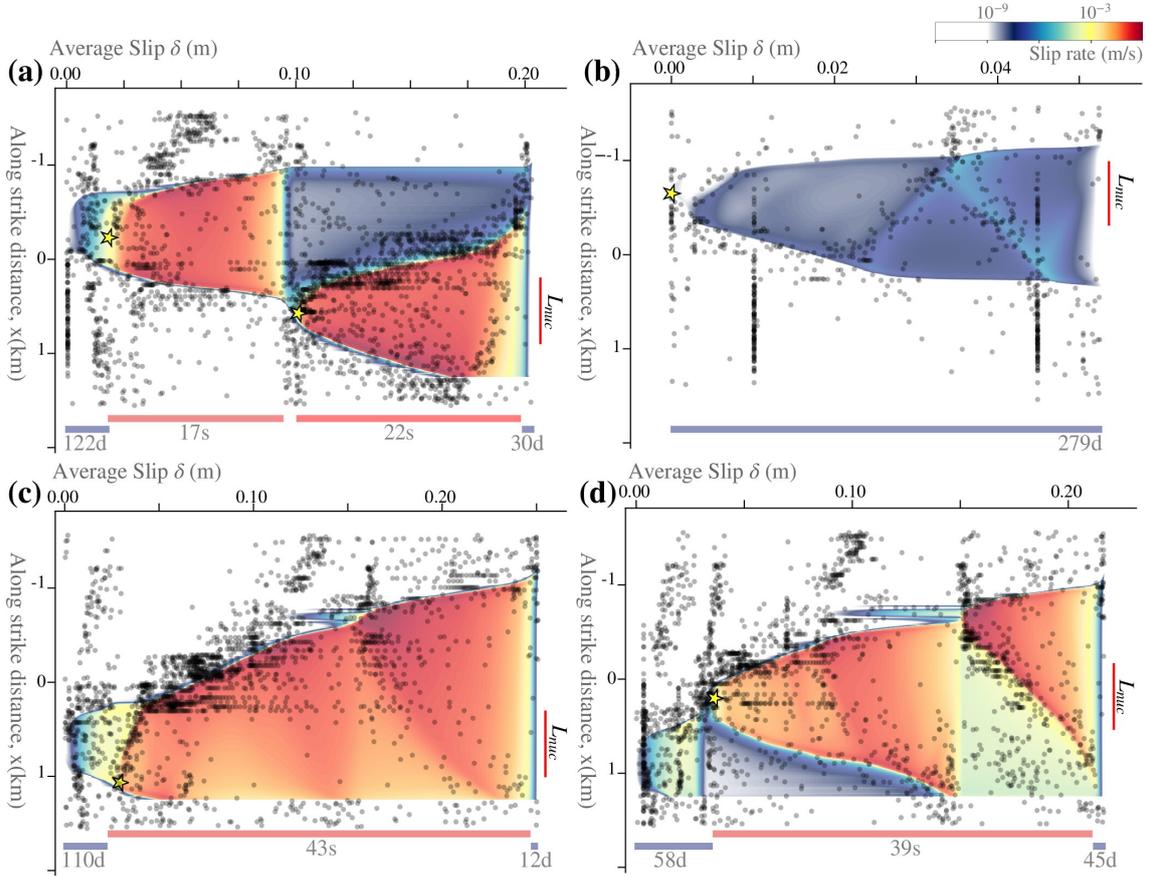


Figure 3: Sequences of slip rate profiles on the main fault. Top x-axis is average slip accumulated during the specific time sequence. Left y-axis represents the along-strike distance on the main fault. Horizontal colored lines show the specific durations for different phases of the rupture sequence, with blue lines indicating slow phases and red lines indicating fast phases. Nucleation and afterslip phases are indicated by two blue lines at the beginning and end of fast events. The colormap represents slip rate. Black circles represent events detected on the fractures, and projected onto the main fault strike. Yellow stars represent the epicenters of the different events. Panel a) shows complex partial ruptures on the main fault. Panel b) shows one slow slip event. Panel c) shows a full rupture on the main fault. Panel d) shows an event, nucleated from one end of the fault and accelerated in a cascading process.

Having explored temporal dynamics, we now examine spatial complexity of the seismic events recorded on the main fault (Figure 3). We delve into the spatial complexity of seismic events recorded on the main fault by looking at the slip rate profiles along strike as a function of spatially averaged cumulative slip during the event. In addition, we project all off-fault activity during this event onto the main fault and denote them by simple black circles. For our analysis, we carefully selected several representative events from the case study, encompassing a diverse range of slip occurrences. This includes partial ruptures, full ruptures, and slow slip events, each characterized by distinct magnitudes and durations. For the fast ruptures, we present both the nucleation phase, where the slip rate starts to accelerate from an arbitrarily small threshold, i.e., 10^{-9} m/s, prior to reaching some geodetically detectable threshold, i.e., 10^{-6} m/s, and the afterslip phase, where the slip rate decelerates to the same limits. Figure 3a features two partial ruptures, each encompassing around half of the main fault. Notably, the second rupture (moment magnitude $M_w = 4.70$)

nucleates at the edge of the first one ($M_w = 4.56$), demonstrating a cascading effect. Despite the rapid rupture of coseismic phases, which last 17 and 22 seconds, respectively, the nucleation phase lasting 122 days and the afterslip phase taking place over 30 days are significantly slower.

In Figure 3b, one slow slip event is showcased, rupturing only part of the main fault. This slow slip event spans 279 days with a moment magnitude $M_w = 4.48$. It is essential to acknowledge the subjectivity in detecting slow events due to the slip rate threshold used (here defined by slip rates between 10^{-9} and 10^{-3} m/s). This threshold effect will be discussed later in the text. We observe continuous seismic activity on the off-fault fractures accompanying the slow event on the main fault, as previously seen in Figure 2c. Figure 3c unfolds a full rupture on the main fault with a moment magnitude of $M_w = 5.03$, revealing nuanced variations in slip rate acceleration and deceleration within the rupture and along the fault's strike. The nucleation and afterslip phases of this event last 110 and 12 days, respectively. Figure 3d features another full rupture on the main fault ($M_w = 4.95$), preceded by a nucleation phase that ruptures around one-third of the main fault over 58 days. The rupture accelerates from the left corner of the nucleation patch. Then, the full rupture on the main fault lasts for 39 seconds; after a 45-day afterslip phase, the slip rate falls below 10^{-9} m/s. The figure illustrates a distinct cluster of off-fault events following the rupture front on the main fault.

Beyond demonstrating that slow and fast events emerge naturally within the same fault system under identical constitutive parameters, our model is also able to produce spatially localized regions where slow slip is persistent. This emerges naturally from the local geometric configuration of the main fault and surrounding off-fault fractures, without requiring imposed frictional or hydraulic heterogeneity. The coexistence of slow and fast slip events on the same fault or within the same region (see Figure S2), as produced by our simulations, is also supported by a growing body of observational, geological, and laboratory evidence. Seismological observations document the coexistence or close spatial association of slow slip, tremor, very-low-frequency earthquakes, and regular earthquakes in several tectonic settings (*e.g. Kato et al. 2012; Johnson et al. 2012; Ito et al. 2013; Ruiz et al. 2014; Tsang et al. 2015; Lin et al. 2020; Villafuerte et al. 2025; Woods et al. 2024*). Geological studies indicate that deformation mechanisms associated with slow and fast slip can operate within the same fault zones over geological timescales (*e.g. Fagereng et al. 2019*), and laboratory experiments demonstrate the coexistence and interaction of slow and fast slip under controlled conditions (*e.g. Aben & Brantut 2023*).

3.2 Slow to fast events across scales

We present a general overview of the source time functions (STFs) generated within our fault system model, focusing on their variability in moment magnitude, duration, and spectral characteristics. The analysis includes moment-rate functions (MRFs) from both slow and fast events, considering a representative fault volume geometry in which off-fault fractures are oriented parallel to the main fault (unlike the case discussed earlier where the off-faults are optimally oriented right-lateral faults).

While we get a continuum of slow and fast events for all off-fault geometries, we selected this particular geometry because it exhibits a remarkably broad and diverse range of temporal behaviors, capturing both small-magnitude slow events and large slow slip events occurring on both off-faults and the main fault. This suggests that adjusting frictional parameters could potentially

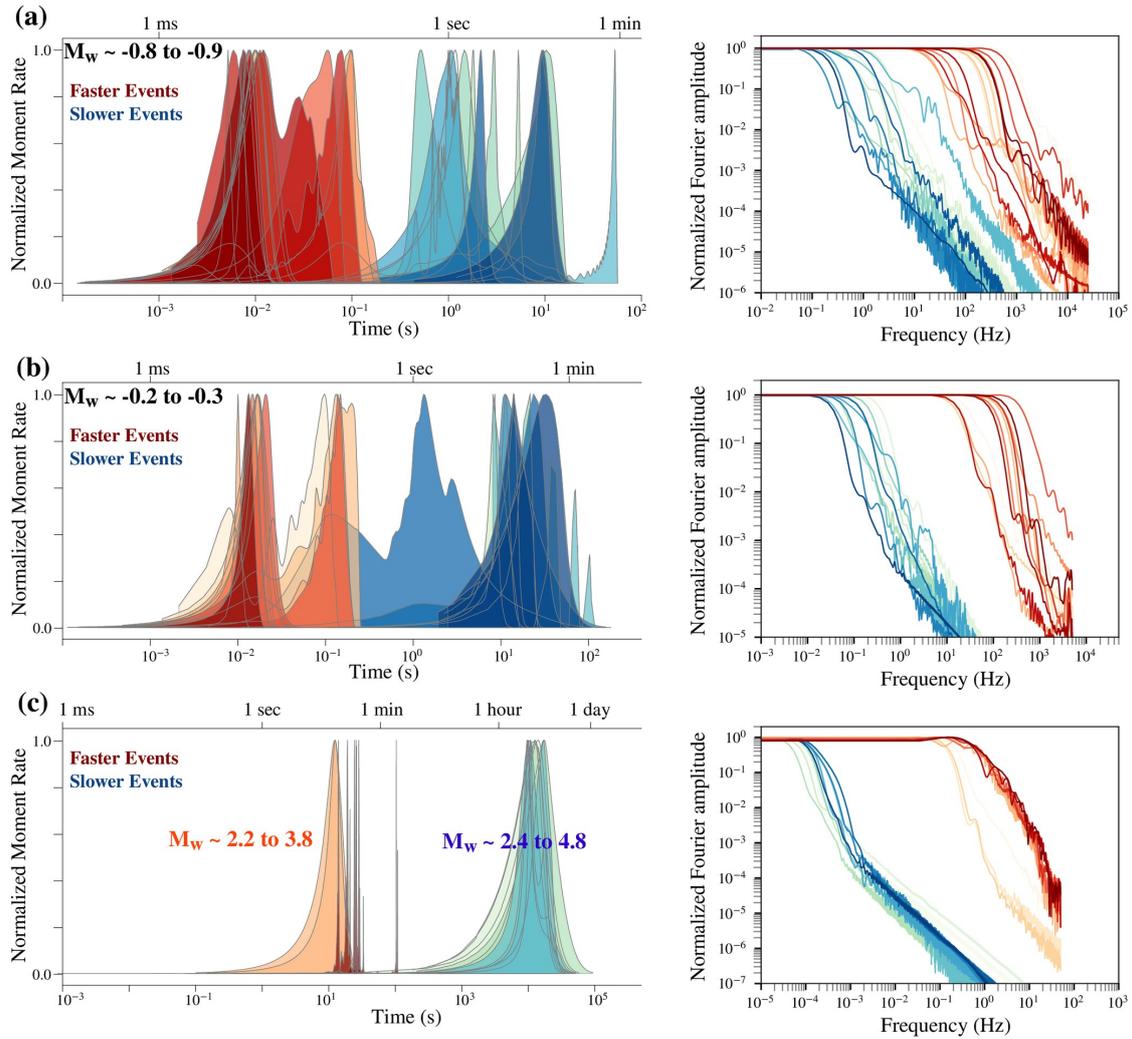


Figure 4: Moment-rate functions (MRFs) (left panels), and their spectra (right panels), across different M_w ranges for both off-fault and main fault events. (a) Selected MRFs from slow and fast events in the off-fault region within a narrow range of $-0.9 \leq M_w \leq -0.8$. (b) Same as (a), but for $-0.3 \leq M_w \leq -0.2$. (c) MRFs from main fault events, and their spectra, with fast ruptures in the range $2.2 \leq M_w \leq 3.8$ and slow slip events in the range $2.4 \leq M_w \leq 4.8$.

produce similar behavior across different geometries, though this requires further investigation.

Figures 4a and 4b show selected MRFs from off-fault events within narrow magnitude ranges: $-0.9 \leq M_w \leq -0.8$ and $-0.3 \leq M_w \leq -0.2$, respectively. These MRFs exhibit a continuous and broad distribution of slip durations, with significant variability in their shape, ranging from abrupt moment release to smoother and more gradual evolutions. While the fast events are classical earthquakes, the slower events can easily be called LFE's and VLFES. To confirm this, we need to radiate these events dynamically and look at their far-field waveforms, which we plan to do in future work. Although not explored here, our model framework could also explain tectonic tremors as successive ruptures of shear cracks as recently proposed by *Yabe & Ujiie (2025)*. This diversity highlights the influence of structural complexity and distributed deformation in controlling rupture behavior, emphasizing that seismic complexity arises not only from the main fault, as shown previously, but also from off-fault activity.

Figure 4c shows MRFs of events on the main fault, which exhibit a bimodal behavior. Due to the greater extent and roughness, the main fault hosts the largest magnitude events, characterized by

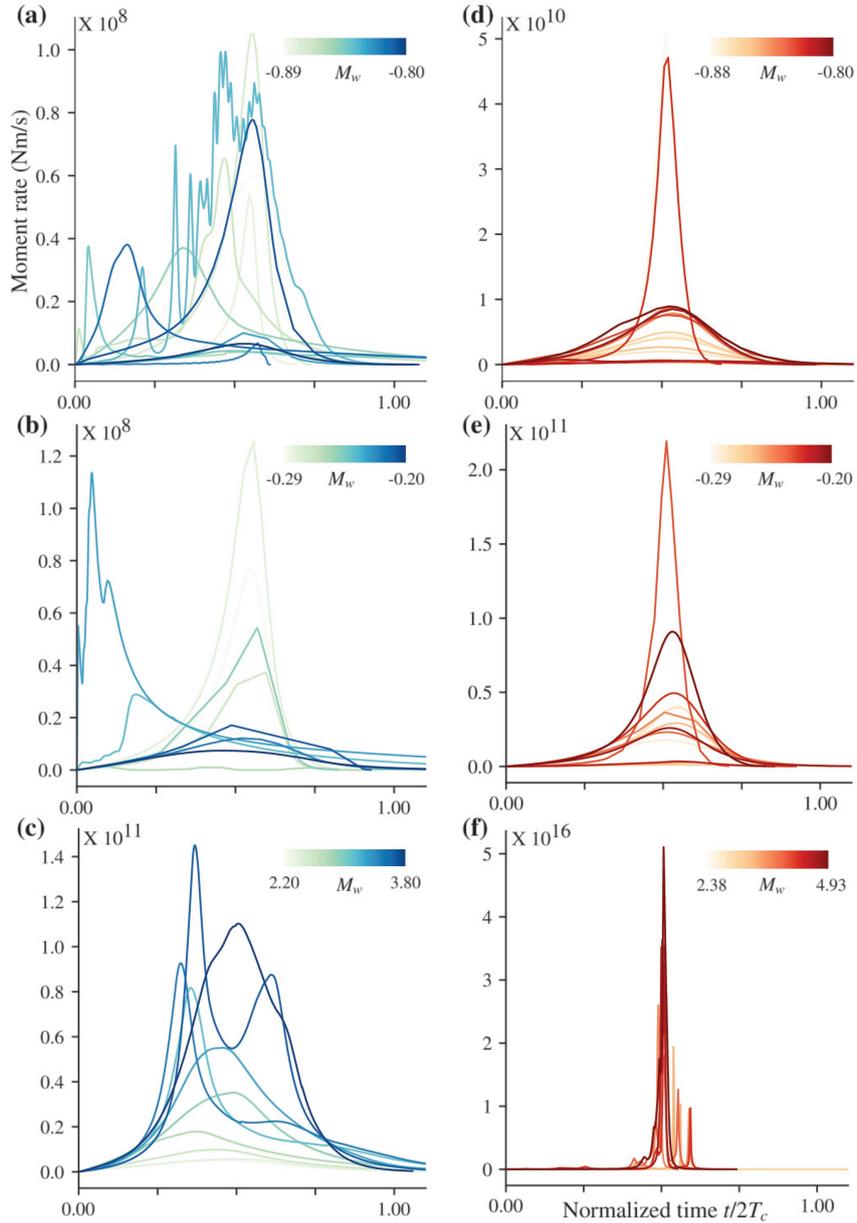


Figure 5: MRF shapes for slow and fast events after normalization by centroid time with colors indicating moment magnitude. Panels (a) and (b) correspond to slow events in the off-fault region for the same magnitude ranges as in Figures 4a and 4b. (c) shows slow events on the main fault. Panels (d) and (e) correspond to fast off-fault events with the same M_w ranges as in (a) and (b). (f) Fast events on the main fault.

more complex STFs, including multiple acceleration and deceleration phases during the ruptures. The magnitude range considered here includes both large-magnitude, fast ruptures, partial or full, ($2.2 \leq M_w \leq 3.8$) and long-duration slow slip events ($2.4 \leq M_w \leq 4.8$).

We now examine the spectral characteristics of these events. The corresponding source spectra for the same events show, as expected, that the longer durations of the slow events lead to significantly lower corner frequencies than fast events across all three magnitude ranges, on both the main and off-faults. Recently, *Wang et al. (2023a)* analyzed the S-wave displacement spectral signature of low-frequency earthquakes in the Nankai Trough, after correcting for empirically derived attenuation, and concluded that the spectra are consistent with the classical earthquake model with vastly different rupture velocities and stress drops. Our results align very well with

the conclusions drawn by *Wang et al. (2023a)*. Beyond the corner frequency, the spectral decay of slow events appears similar to that of fast events. However, the high-frequency content is less reliably estimated, as it would be predominantly influenced by the fully dynamic response of both event types, an aspect that our quasi-dynamic model does not account for, but which we aim to address in future investigations. To further examine MRF characteristics across these events, we normalized the time axis of each MRF with respect to its centroid time, T_c (Figure 5). The centroid time is the moment rate weighted average time (*Dziewonski et al. 1981; Duputel et al. 2013*) and is defined as $T_c = \int_{t_1}^{t_2} t \dot{M}_0(t) dt / \int_{t_1}^{t_2} \dot{M}_0(t)$. Here $\dot{M}_0(t)$ is the moment rate, and t_1 and t_2 correspond to the onset and termination times of the MRF, defined by when the slip rate crosses a detection threshold.

One key distinction between slow and fast STFs lies in their symmetry. MRFs of slow events (Figures 5a–c) tend to be more asymmetric, typically exhibiting a sharp rise followed by a longer, gradual decay—indicating a positively skewed moment release. In contrast, fast earthquakes generally show more symmetric MRFs with a single, well-defined peak (Figures 5d–f). However, for the largest slow slip events (Figure 5c), the STFs of both fast and slow events become more irregular, and the asymmetry for some events is less distinguishable. Consistent with the findings of *Meier et al. (2017)*, our results suggest that fast earthquakes tend to have simpler MRFs, with fewer subevents and a well-defined peak and gradual decay (Figures 5d–f). While this section focuses on a specific fault geometry, similar results were obtained for all the 40 fault volume configurations tested in this study.

4 Statistical and Scaling Laws of the Events in the Catalog

In this section, we demonstrate the ability of our numerical model to replicate observed empirical statistical relationships and scaling laws governing seismic activity. These empirical relationships include the Omori law for the decay in the rate of aftershocks, Gutenberg-Richter magnitude-frequency distribution, the inverse-Omori law for foreshock escalation, moment-area scaling, and moment-duration scaling. We posit that any reasonable ‘digital twin’ of a fault system should be able to reproduce these empirical laws in addition to the range of slip dynamics presented earlier. Through this analysis, we plan to show that our fault volume model not only produces a diverse range of slip dynamics but also satisfies key empirical laws observed in natural seismicity. We show that for each geometrical configuration and friction parameters, the above listed empirical laws are satisfied (see Figures 2e, S3, S4). For the sake of brevity here we show the robustness of these results across all the different cases of off-fault orientations and frictional parameters.

4.1 Omori Law

In seismological studies, the Omori law characterizes the temporal decay of aftershock activity following a mainshock event. Specifically, it states that the number of aftershocks decreases inversely with time after the mainshock, following a power-law decay as follows: $n(t) = k/(c+t)^p$ where k is aftershock productivity, c is a time offset, and p is a constant that describes the decay rate and typically falls in the range 0.8–1.2 in most cases (*Omori 1894; Utsu et al. 1995*). To evaluate the Omori law in our simulations, for each orientation distribution of the off-fault fractures, we stack the aftershock sequences following each major rupture on the main fault. We then fit the stacked sequences to the Omori law to estimate the p exponent. Figure 6a illustrates the

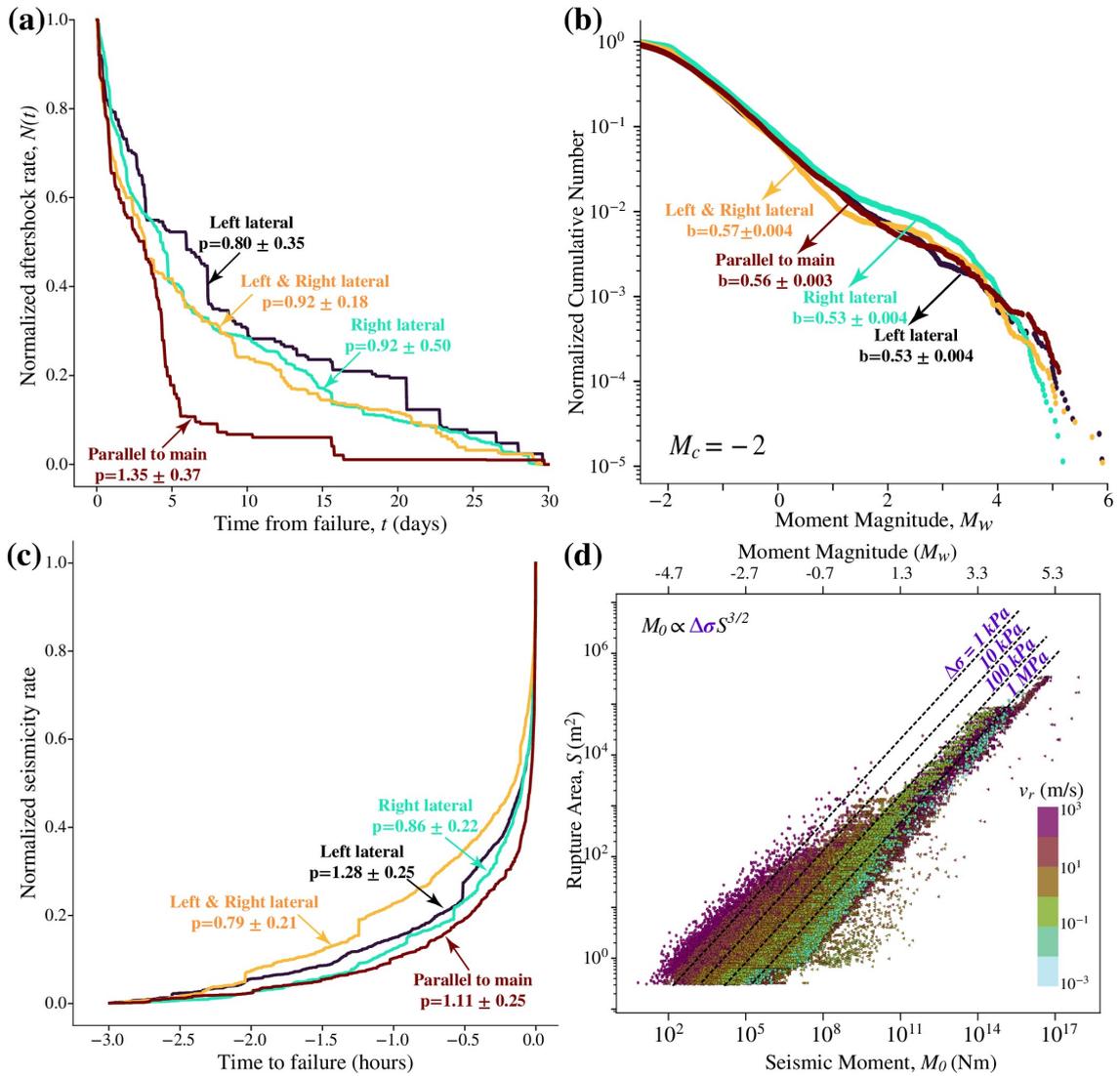


Figure 6: (a) Omori Law: Effect of off-fault fracture orientation. The average Omori decay curve over a period of 30 days is calculated from the stacked curves of different fracture orientation, and the average p coefficient is reported (b) Magnitude Distribution: Effect of off-fault fracture orientation on the b -value found from stacked catalog of fast ruptures with same off-fault fracture orientation (c) Inverse Omori: Effect of off-fault fracture orientation (d) Moment-area scaling: The a/b ratio of the off-fault fractures is 0.5.

Omori law coefficient p across different fracture orientations when $a/b = 0.5$. The coefficient remains consistently around 1 across all orientations, indicating a uniform behavior regardless of these parameters. While off-fault orientation does influence the duration of the aftershock period (see the case when off-faults are aligned with the main fault), we defer this investigation to future studies as it falls outside the scope of our current research.

4.2 Gutenberg-Richter Magnitude Frequency Distribution

Figure 6b illustrates the magnitude-frequency distribution of ruptures across different fracture orientations. The x-axis denotes the magnitude range, while the y-axis represents the frequency of occurrences. The distribution follows the Gutenberg-Richter law, exhibiting a logarithmic relationship between magnitude and frequency. This distribution provides crucial insights into the relative occurrence rates of seismic events of varying magnitudes, offering valuable information

for seismic hazard assessment and earthquake forecasting. To estimate the b -value, we utilize the maximum likelihood method (Aki 1965), compensating for the binning error (Marzocchi & Sandri 2009). We note that $M_w = -2$ approximately serves as the completeness magnitude of our catalog. All plotted distributions conform to the Gutenberg-Richter law, displaying a similar b -value across orientations. Interestingly, there is no apparent influence of off-fault fracture orientation on the magnitude-frequency distribution, suggesting that this aspect does not significantly impact slip dynamics.

As shown by Aki (1981), a b -value between 0.5 and 1 can be envisaged when the fractal dimension is between 1 and 2 where the fault lines are distributed along a plane. This needs to be examined in much more detail. Although there is still room to investigate the convexity of the magnitude–frequency distribution using truncated G–R distributions (Huang & Beroza 2015), we leave this investigation for future studies. It is important to note that our primary objective is not to study the origin of the b -value derived from our fault volume model. Rather, our focus is to demonstrate that considering the dynamics of a fault volume with fractures surrounding the main fault results in a logarithmic relationship between magnitude and frequency, consistent with the observed Gutenberg-Richter law in nature. While we suspect that the b -value may be influenced by factors such as the power-law distribution of fracture length and density distribution of fractures along fault-normal distance, we defer this exploration to future studies, as it falls beyond the scope of our current investigation.

4.3 Inverse Omori Law

In addition to the Omori-law-like reduction in the earthquake rate after a given event (section 4.1), the catalog of events in this paper’s simulation follows the empirical inverse Omori law that characterizes the increase in foreshock activity preceding a major rupture event. Unlike the aftershock decay described by the traditional Omori law, the inverse Omori law predicts a gradual rise in foreshock activity leading up to a significant seismic event, where foreshock rate increases as an inverse of the time to the mainshock as follows: $n(t) = 1/(c + \Delta t)^p$ (Jones & Molnar 1979; Shearer et al. 2023), where p is a constant that describes the increase rate of foreshocks and is typically around 1. In Figure 6c, we show the increasing rate of seismicity in a 3-hour window prior to the main shock, grouped by fracture orientation. The increasing rate of seismic events off-fault exhibits an inverse Omori law, with p -values close to 1. The alignment with the Omori law underscores its applicability in describing the temporal evolution of aftershock activity within our simulation. Similarly, the consistent trend of foreshock rate increase observed across our simulated seismic events aligns well with the predictive power of the inverse Omori law, demonstrating its ability to capture precursory behavior preceding major seismic events and laboratory earthquakes as shown in Marty et al. (2023).

4.4 Moment-Area Scaling Law

Next, in Figure 6d, we explore the relationship between seismic moment (M_0) and rupture area (S) for both fast and slow ruptures. Notably, while determining rupture area (S) in nature can often be challenging, our model offers a straightforward approach as we have precise knowledge of fault behavior. Given our 2D model, we assume a circular rupture shape, simplifying the calculation

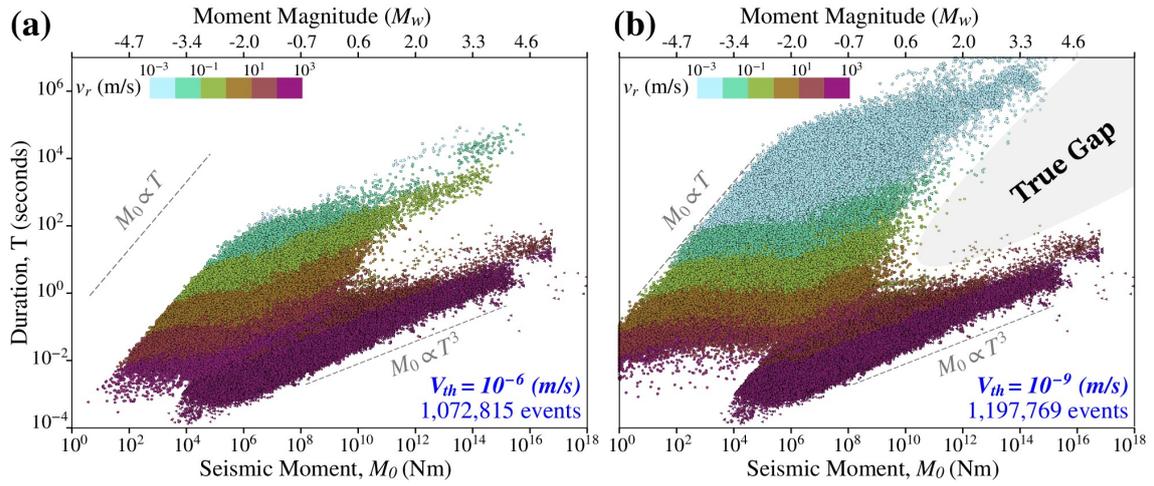


Figure 7: Moment-duration scaling for the comprehensive catalog compiled from all simulations : a) with a threshold of detection of slow ruptures at 10^{-6} m/s, showing a $M_0 \propto T^3$ relation for fast and slow ruptures; b) with a threshold of detection of slow ruptures at 10^{-9} m/s. Events are color-coded based on rupture velocity v_r .

of rupture area as equal to the square of rupture length. The figure reveals a scaling relation of $M_0 \propto S^{3/2}$, mirroring patterns observed in natural seismicity (*Kanamori & Brodsky 2001*). It is essential to note that this scaling law is consistent across both fast and slow ruptures. This scaling law underscores the fault volume model's fidelity in capturing the spatial distribution of seismic energy release across various rupture sizes. Importantly, the observation that both fast and slow ruptures exhibit almost the same scaling between moment and rupture area is noteworthy. Given that seismic moment relates to stress drop and rupture area as $M_0 = C \Delta \sigma S^{3/2}$ (*Eshelby 1957; Kanamori & Brodsky 2001; Madariaga 2009*), where C is a geometric constant ($C = 16/7\pi^{3/2}$ for a circular crack), the variability in stress drop must be smaller than that of rupture velocity. In fact the above relationship can be rearranged to show that $\bar{V} \mu / \bar{v}_r \Delta \sigma$ is a constant. Here \bar{V} is the average slip rate of the event and \bar{v}_r is the average rupture velocity of the event. In our simulations, average slip rates vary over 8 to 9 orders of magnitude between slow and fast events, while rupture velocities vary over 5 to 6 orders of magnitude. This implies that stress drops should vary over 2 to 3 orders of magnitude across the entire catalog of events. It is also clear that average slip rate, stress drop and average rupture velocity are three fundamental quantities of an event that cannot vary independently and are linked through the above relationship. *Supino et al. (2020)*, for instance, found that low-frequency earthquakes propagate more slowly and have lower stress drops than regular earthquakes.

4.5 Moment-Duration Scaling Law

In Figure 7, we delve into the moment-duration scaling, a crucial aspect of slow and fast ruptures. Observations in natural seismicity have highlighted a cubic relation between moment and duration for fast ruptures: $M_0 \propto T^3$ (*Kanamori & Anderson 1975; Kanamori & Brodsky 2004; Houston 2001*). However, for slow ruptures, the situation appears more complicated, with some observations suggesting a linear relation: $M_0 \propto T$ (*Ide et al. 2007; Ide & Beroza 2023*), while others indicating a cubic one (*Gomberg et al. 2016; Michel et al. 2019*). To investigate this further, we assess whether the detection threshold for slow ruptures influences the observed scaling relationship. Specifically, we examine Figure 7b, where the detection threshold is set to 10^{-9} m/s, and

compare it to Figure 7a, which uses a threshold of 10^{-6} m/s. This comparison allows us to evaluate how the choice of detection threshold affects the apparent moment–duration scaling. Notably, the latter threshold is still three orders of magnitude smaller than the threshold for fast ruptures, set at 10^{-3} m/s. The color scale represents the rupture velocity. The bottom x-axis displays seismic moment (M_0), while the top x-axis shows moment magnitude (M_w) for clarity and comparison with observational data. For the fast ruptures, we observe a clear cubic scaling ($M_0 \propto T^3$) between moment and duration, consistent with previous findings.

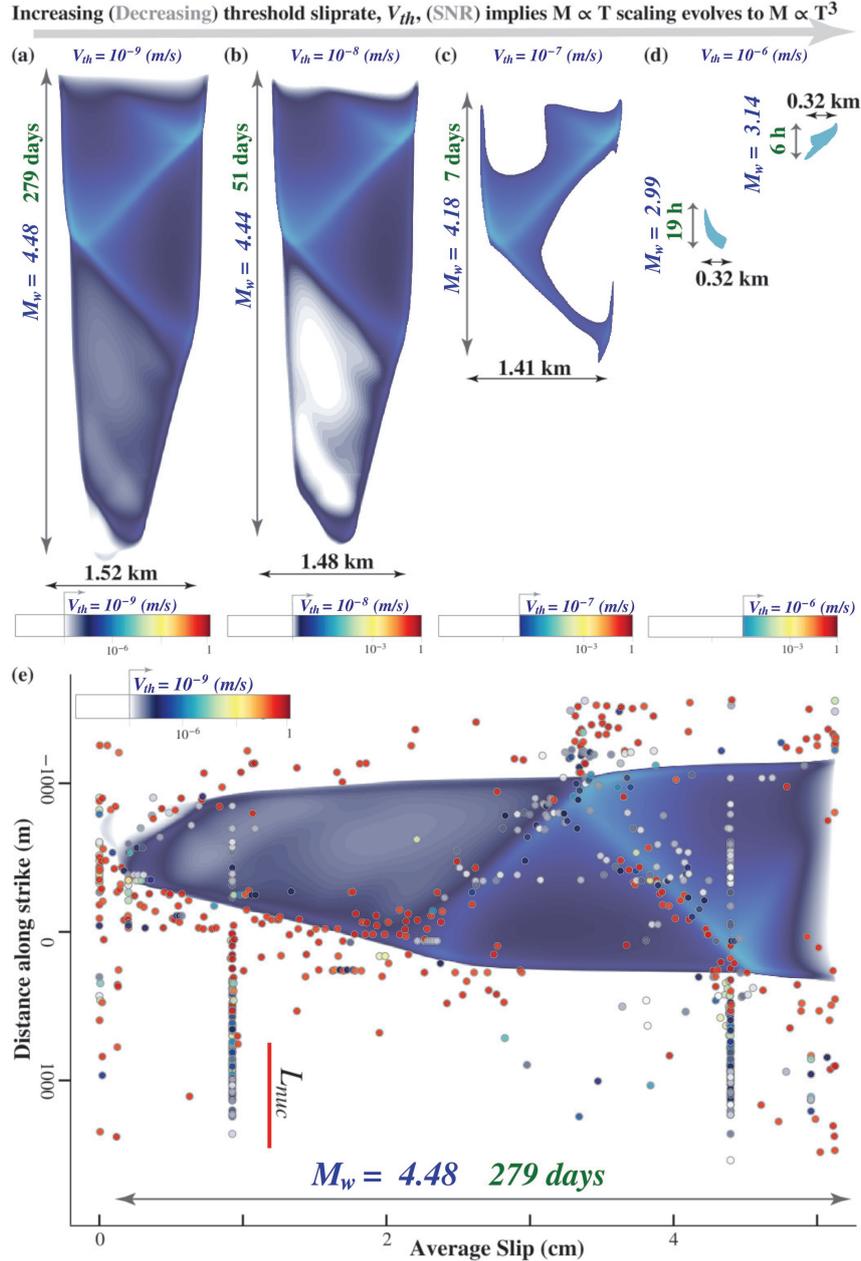


Figure 8: Effect of slip rate threshold (V_{th}), or signal to noise ratio (SNR), on the inference of the magnitude and duration of a slow slip event. The slip rate is plotted with four different minimum detection levels: (a) 10^{-9} m/s, (b) 10^{-8} m/s, (c) 10^{-7} m/s, and (d) 10^{-6} m/s. (e) Same as (a) with off-fault seismicity. Circles represent ruptures on the off-fault fractures, and are color-coded with respect to their maximum slip rate.

However, the behavior of slow ruptures varies between the two detection thresholds. With a lower threshold (Figure 7b), the durations of slow events are larger and we also obtain more events compared to the higher threshold case (Figure 7a). The overall trend of the slow events thus

changes as a function of the detection threshold. This discrepancy suggests that the scaling between moment and duration for slow ruptures is highly dependent on the detection threshold in our numerical model, akin to the sensitivity or threshold settings of field recording instruments or detection strategies employed on field data. In a recent work *Costantino et al. (2026)*, similar threshold-dependent effects were observed, in natural settings, when applying deep learning denoising techniques to reveal a continuum of slow slip events in the Cascadia subduction zone, demonstrating that detection threshold fundamentally shapes our understanding of slip behavior. The effect of the detection threshold, ranging from 10^{-9} m/s to 10^{-6} m/s, on the detectable moment and magnitude of an example slow-slip rupture is shown in Figure 8a-d, highlighting how inferences drawn from the duration and moment of a slow-slip rupture could be changed if the lowest slip rate at which the rupture is initiated cannot be detected.

While the slip rate on the main fault may be very low and remain undetectable, the pattern of off-fault seismicity migration provides valuable information about the rupture process on the main fault, Figure 8e. The dot symbols on the plot represent off-fault events projected onto the main fault, marking the extents of the slow slip rupture. Further analysis is warranted to assess how reliably the pattern of off-fault seismic activity can be used to infer the magnitude and duration of a slow event, and we defer this investigation to future studies.

Since each of the events is color-coded based on its average rupture velocity in Figure 7, it is apparent that slow or fast events are identical in nature except for their rupture velocity. In fact, it is evident that $M_0 \propto T^3$ and $M_0 \propto T$ are limiting bounds in the moment-duration scaling as proposed by *Ide & Beroza (2023)*. This suggests that slow ruptures are simply fast ruptures with different slip rates, rupture velocities, and stress drops. This also implies that the slow ruptures do not emerge due to a special frictional constitutive law since it is the same mechanism that generates fast ruptures as well. What thus produces the difference in rupture velocity is the spatiotemporally complex traction on the main fault due to the presence of the fault volume. We also observe small events tend to have a continuum of rupture velocities; however, as events become larger, a distinct gap emerges between fast and slow ruptures. This gap mirrors observations reported by *Ide & Beroza (2023)*, suggesting events that are challenging to detect or exceedingly rare. Our numerical model's ability to detect these events with confidence reinforces the presence of this gap, raising the possibility of a mechanical constraint where large events predominantly exhibit either fast or slow rupture velocities, rather than a continuum. This strongly suggests that both fast and slow ruptures do originate from the same mechanical model, despite their differing rupture velocities.

4.6 Localization and Migration of Seismicity

In this section, we address the question of localization and delocalization of deformation throughout the seismic cycle, aiming to understand how deformation occurs in a fault volume over a seismic cycle. Over decadal timescales during the interseismic period, elevated background seismic activity may reflect progressive damage accumulation across a broad region that will eventually host a major earthquake. In several large events including the 1992 M_w 7.3 Landers, the 1999 M_w 7.1 Hector Mine, the 2019 M_w 7.1 Ridgecrest, and the 2023 M_w 7.8 Kahramanmaraş earthquakes, seismicity has been observed to migrate and concentrate toward the eventual rupture plane over a few years to several months preceding the mainshock. This localization appears to facilitate

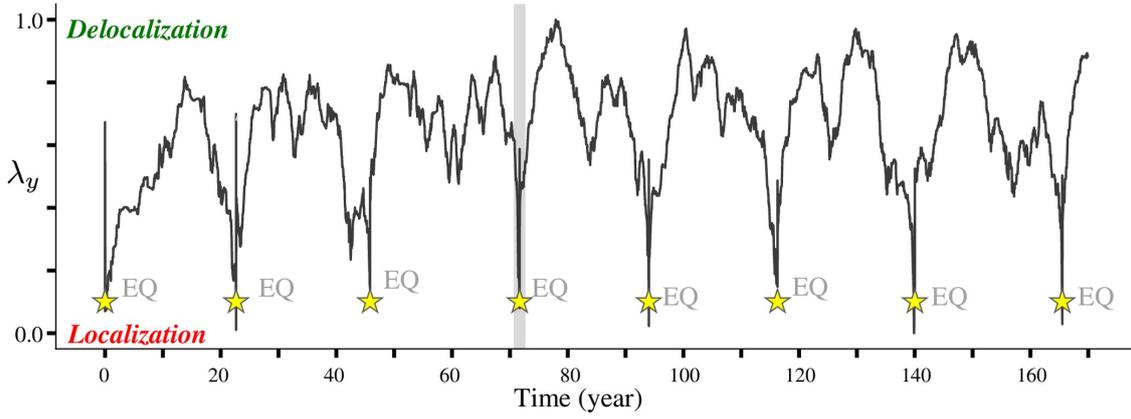


Figure 9: Localization and delocalization of off-fault seismic activity, measured by the standard deviation of hypocentral distances from the main fault in sequential 500-event batches, λ_y . The timing of earthquakes is indicated by yellow stars. The standard deviation values are normalized by their maximum value over all batches. The gray region highlights the time window shown in Figure 10.

the interaction and coalescence of fault segments and fractures that delineate the future principal slip zone (*Ben-Zion & Zaliapin 2019; Ben-Zion & Zaliapin 2020; Pritchard et al. 2020; Kato & Ben-Zion 2020; Kwiatek et al. 2023; Núñez-Jara et al. 2025*).

In our simulations, we quantify this localization by looking at the location of off-fault events over time. To this end, we calculate the standard deviation of hypocentral distances from the main fault (λ_y) in sequential 500-event batches and normalize the values between 0 and 1. The events tend to localize toward the main fault plane when $\lambda_y \rightarrow 0$ and delocalize into the damage zone when $\lambda_y \rightarrow 1$. Figure 9 shows the time series of λ_y . A reduction in this time series before the main fault earthquake, followed by an increase after it, is clearly distinguishable, highlighting the localization of seismicity onto the main fault as the time of the mainshock approaches, and its delocalization after the occurrence of the earthquake.

A closer look at the location of off-fault events reveals a migration toward the epicenter of the upcoming mainshock. We observe this feature somewhat systematically in our simulations but show only the visually clearest example below. Figure 10 shows the location of off-fault events during the time interval highlighted by the gray hatch in Figure 9. Specifically, Figure 10a shows a one-year window of foreshocks, while Figure 10b shows a one-year window of aftershocks. Events are shown by circles, color-coded based on time to failure in panel (a) and time from failure in panel (b). This figure clearly illustrates the migration of off-fault events toward the epicenter, emphasizing the importance of refined event catalogs for detecting such patterns and their potential forecasting value as a mainshock approaches. Following the rupture on the main fault, λ_y increases, and off-fault seismicity no longer exhibits a migration pattern.

4.7 Apparent Diffusion

It has been observed that the migration of seismicity in some fault zones follows a specific space-time evolution, similar to the one obtained from fluid diffusion (*e.g., Danré et al. 2024*). In other words, the distance r from the epicenter versus the time from the main shock t is observed to be $r \propto D\sqrt{t}$. In our simulations, we observe the same evolution despite the absence of any underlying fluid diffusion process. This observation is illustrated in Figure 11, where two slow-

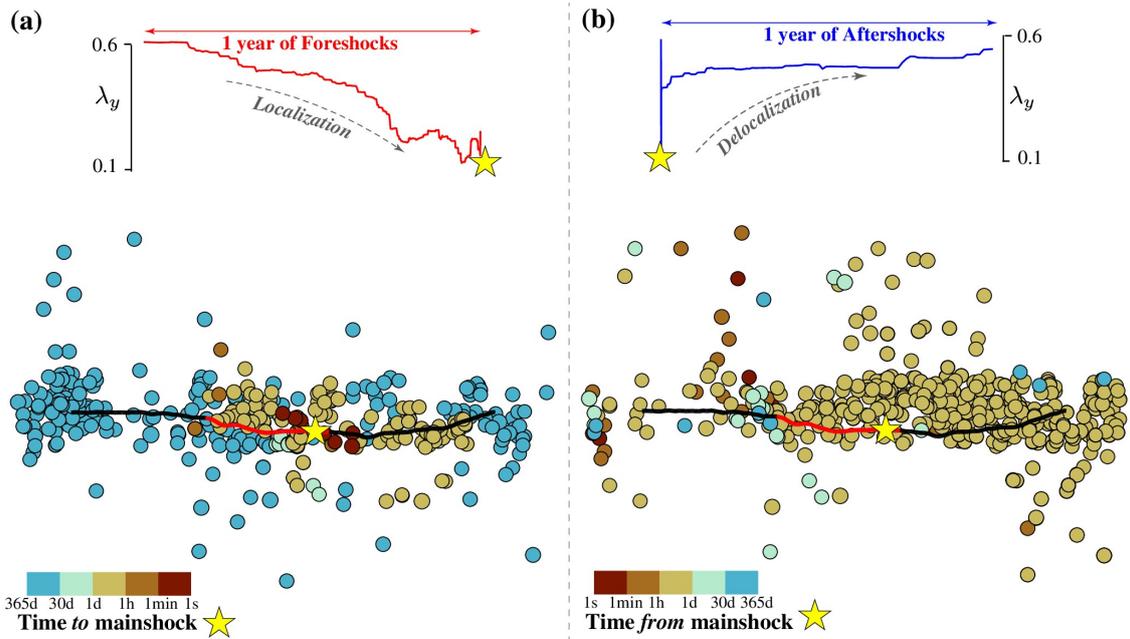


Figure 10: Zoomed-in view of the gray-hatched window shown in Figure 9, illustrating the standard deviation of the fault-normal component of earthquake locations over a one-year window before (a) and after the earthquake (b): Migration of seismicity toward the future epicenter during the foreshock period, followed by a return to background seismicity after the earthquake. Events are shown as circles, color-coded by time before and after the rupture.

slip ruptures are shown (panels a,c) along with seismic events occurring off-fault overlaid on the slip rate panels. The relative distance is calculated as the distance between the off-fault event and the detected nucleation point on the main fault, using a slip rate threshold of 10^{-9} m/s. The relative time is determined by the time difference between the occurrence of the off-fault event and that of the main fault. This off-fault seismicity migration is shown in panels (b,d). The values of the inferred diffusivity parameter D (0.02–0.06 m^2/s) fall within the observed range for the fluid diffusivity parameter (Amezawa *et al.* 2021).

Our simulations reveal an apparent diffusive process driven by slow slip rather than fluids, demonstrating that seismicity migration following $r \propto \sqrt{t}$ does not necessarily imply underlying fluid-driven processes. This behavior can also emerge from the physics of interacting faults governed by rate-and-state friction along with a slow slip event occurring on the main fault.

5 Discussion

These findings, when taken in total, highlight the fault volume framework’s capacity to capture global, fault-related scaling laws and statistics, both on the main fault and within the fractured medium. Moreover, the fault volume model accurately reproduces diverse slip dynamics, slow to fast earthquakes across scales, all within a unified framework. Our simulations also replicate the observed localization of deformation leading up to a rupture and subsequent delocalization of deformation into the fractured medium afterward, as recently reported by Kato & Ben-Zion (2020) and Ben-Zion & Zaliapin (2020). This comprehensive representation of seismic activity underscores the importance of considering fault volume effects in seismic simulations. While several existing models produce one, or many, of the above-discussed features, the fault volume model stands out as the first to reproduce the entirety of observations and statistics simultaneously, demonstrating its efficacy in capturing the multifaceted nature of seismicity within fault systems.

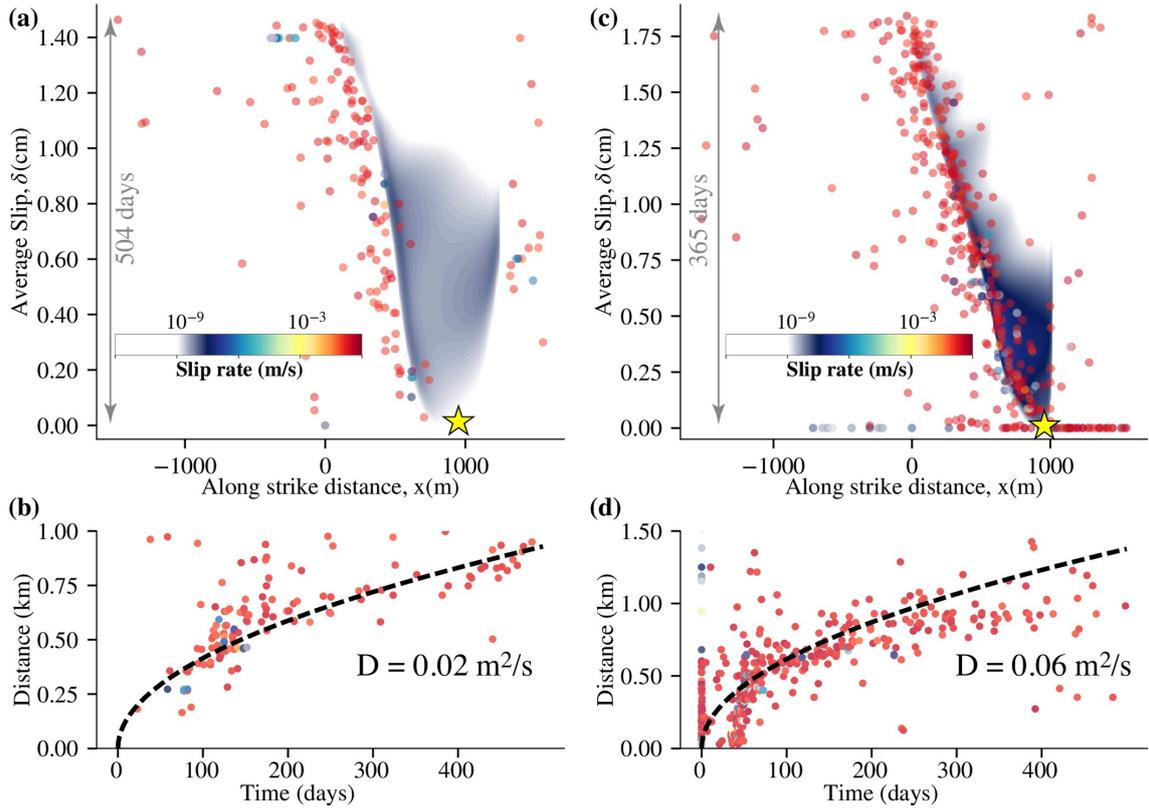


Figure 11: Off-fault Seismicity during two examples of slow-slip ruptures. a-c) Slip rate profiles of two slow slip events. b-d) The temporal migration of off-fault seismicity follows a distance from epicenter (r)–time (t) evolution (black dashed curve) similar to that of a diffusion front ($r \propto D\sqrt{t}$). The relative distance is calculated as the distance between the off-fault event and the detected nucleation point on the main fault, using a slip rate threshold of 10^{-9} m/s. The corresponding diffusion coefficient (D) is noted at the top of each curve.

We note here this is not the first attempt in our community to explain the diversity of slip dynamics. In the late 1970’s and early 1980’s there was a flurry of papers by *Blandford (1975)*, *Nur (1978)*, *Aki (1979)*, *Andrews (1981)*, *Hanks & McGuire (1981)*, and *Gusev (1983)* (among many others) that have proposed heterogeneities in tractions, friction and material properties play a major role in fault mechanics. Within this historical context, our work provides a natural framework to introduce traction heterogeneities through geometric complexity and reproduces a broader spectrum of slip dynamics than just earthquakes.

This fault volume model achieves all of the above without the explicit incorporation of frictional heterogeneity or fluid presence within the medium, indicating its efficacy in explaining complex slip behaviors solely through geometric complexity, which is an independently measurable quantity. A notable distinction arises when comparing the behavior of a single fault with frictional heterogeneity to that of a fault volume. In the former case, frictional properties remain constant throughout seismic cycles, while with a fault volume model, traction heterogeneities (asperities or barriers) naturally develop and disappear over various spatiotemporal scales. What may appear as a barrier, region of high-coupling for example, during one seismic cycle (few decades) may become an asperity during the next cycle, and vice-versa. This raises the question of what we mean by “barriers” or “asperities” in the context of geodetic observations, which typically span only a few decades. Such observations are often interpreted as evidence of frictional asperities, yet in reality they simply show that slip occurs when fault traction exceeds frictional resistance. Slip dynamics - whether creep, slow slip events (SSEs), broadband earthquakes or earthquakes -

are additionally governed by the temporal evolution of tractions: they also control the nucleation length and thus determines whether the fault can accelerate to seismic slip velocities. Moreover, no matter how good observations become, it will be impossible to determine the frictional heterogeneity of a fault unless very strict model assumptions are made. Therefore, we argue that modeling the full spectrum of slip dynamics through geometric complexity offers a simpler, integrated, and comprehensive approach opening up a new avenue to develop a digital twin framework for fault systems.

Recent advances in computational earthquake physics have enabled the development of digital twin frameworks for fault systems. *Henneking et al. (2025)* developed a Bayesian inversion-based digital twin that couples physics-based models of earthquake rupture and tsunami generation with real-time seafloor observations, demonstrating feasibility of real-time forecasting for subduction zone events. Their approach combines 3D coupled acoustic-gravity wave equations with seafloor pressure sensor data to infer earthquake-induced seafloor motion and forecast tsunami propagation with quantified uncertainties. Complementing this operational focus, *Abdelmeguid & Elbanna (2022)* and *Mia et al. (2023)* have developed computational platforms for simulating sequences of earthquakes and aseismic slip (SEAS) with explicit representation of fault zone complexity, including off-fault plasticity and secondary fractures. Their hybrid finite element-spectral boundary integral method aims toward “realizing Digital Twins for crustal volumes.” These efforts demonstrate complementary approaches to digital twins in earthquake science: real-time data assimilation for operational forecasting and high-fidelity physics-based simulation of fault zone processes across multiple spatiotemporal scales.

We adopt the latter definition of a digital twin for our physics-based fault volume modeling framework following the broader definition used in computational mechanics as “a high accuracy physics-based model” validated against observations (*Wright & Davidson 2020*). Our framework provides a simple yet high-fidelity virtual representation reproducing the full spectrum of observed fault slip behaviors from first principles, its statistical and scaling laws, enabling hypothesis testing and scenario exploration impossible on natural fault systems. This digital twin generates synthetic data with perfectly known sources and complete observational coverage, allowing us to develop improved metrics and inverse methods for interpreting fault slip dynamics that we can then apply to real observations. Future incorporation of data assimilation will enhance predictive capabilities, but the current physics-based approach already establishes this framework as a powerful digital twin for fault slip dynamics.

While our framework demonstrates the digital twin concept for fault slip dynamics, several avenues merit further investigation. An inexhaustive list is compiled below. First, exploring a broader range of friction parameters, and laws, for both off-fault fractures and the main fault would help establish the generality of our findings across different geological settings. The power-law exponent of the off-fault fracture size distribution represents another key parameter whose systematic variation could reveal how fault network geometry controls the emergent spectrum of seismic and aseismic phenomena along with the statistical properties. While we observed that off-fault orientation influences aftershock duration—particularly when fractures align with the main fault—a comprehensive study of orientation effects remains for future work. Additionally, our current framework’s exclusive focus on shear fractures can lead to unrealistic normal traction accumulation; implementing a strategy to properly manage normal traction loading and unloading would better capture the full traction evolution during slip events. To definitively confirm broad-

band earthquakes, source time functions should be propagated through elastic waves to synthetic seismometers. Moving toward fully dynamic simulations would eliminate quasi-dynamic approximations and allow investigation of inertial effects, though at significantly higher computational cost. Finally, the physical origin of the observed convexity in the truncated Gutenberg-Richter distribution warrants theoretical investigation to understand whether it reflects fundamental constraints on maximum earthquake size or emerges from specific properties of our fault network geometry.

The model makes several testable predictions that can be evaluated using independent observations not directly targeted in this study, including (i) systematic localization and migration of microseismicity toward the eventual rupture plane prior to large events, (ii) threshold-dependent scaling of slow-slip duration inferred from high-resolution geodetic or seismic catalogs, (iii) characteristic spatiotemporal clustering of off-fault seismicity during slow slip, and (iv) statistical properties of source time functions across scales. High-resolution seismicity catalogs, dense nodal arrays, and improved geodetic inversions provide promising avenues to quantitatively test these predictions and potentially falsify the model. Such evaluations will help refine the fault volume framework and its applicability to real fault systems.

6 Conclusion

In this paper, we investigate how fault zone architecture can reproduce the observed spectrum of slip dynamics and their statistical properties. We perform quasi-dynamic simulations on randomly generated 2D fault zone models, where the fault zone consists of a main fault with self-similar roughness and a power-law size distribution of off-fault fractures spanning lengths from about 1 meter to kilometers. The off-fault fracture density decays with distance from the main fault, over a few cohesive zone length scales, following a power-law distribution. All faults are assigned spatially uniform, slip rate-weakening frictional properties, and the critical slip distance d_c is scaled with fault length such that each fracture, if simulated independently, can undergo seismic cycles. We show that these ingredients are sufficient to reproduce the full spectrum of slip dynamics, from slow slip events to fast earthquakes, and their statistical properties.

Our simulations reproduce all of the observed scaling and statistical properties of slip dynamics, including the Gutenberg-Richter magnitude-frequency distribution, Omori and inverse-Omori behavior of seismicity rates before and after mainshocks, moment–duration and moment–area scaling relationships, and the observed localization-delocalization of seismicity around an event. We also observe that off-fault seismicity migrates in a manner similar to fluid diffusion in a homogeneous porous medium and raise a caveat emptor that square-root-time-like migration can emerge from the physics of interacting faults governed by rate-and-state friction, even in the absence of fluid diffusion.

Through an extensive analysis of the moment rate functions (MRFs), we find that off-fault MRF durations are continuous, whereas MRFs on the main fault show a bimodal behavior—i.e., either slow slip events or large fast earthquakes emerge. Additionally, the MRFs of small-magnitude slow-slip events tend to be asymmetric, while those of fast-slip events display a more symmetric pattern. As magnitude increases, the MRFs exhibit more pronounced acceleration and deceleration phases, deviating from simple functional forms and revealing complexities such as multiple phases of moment release.

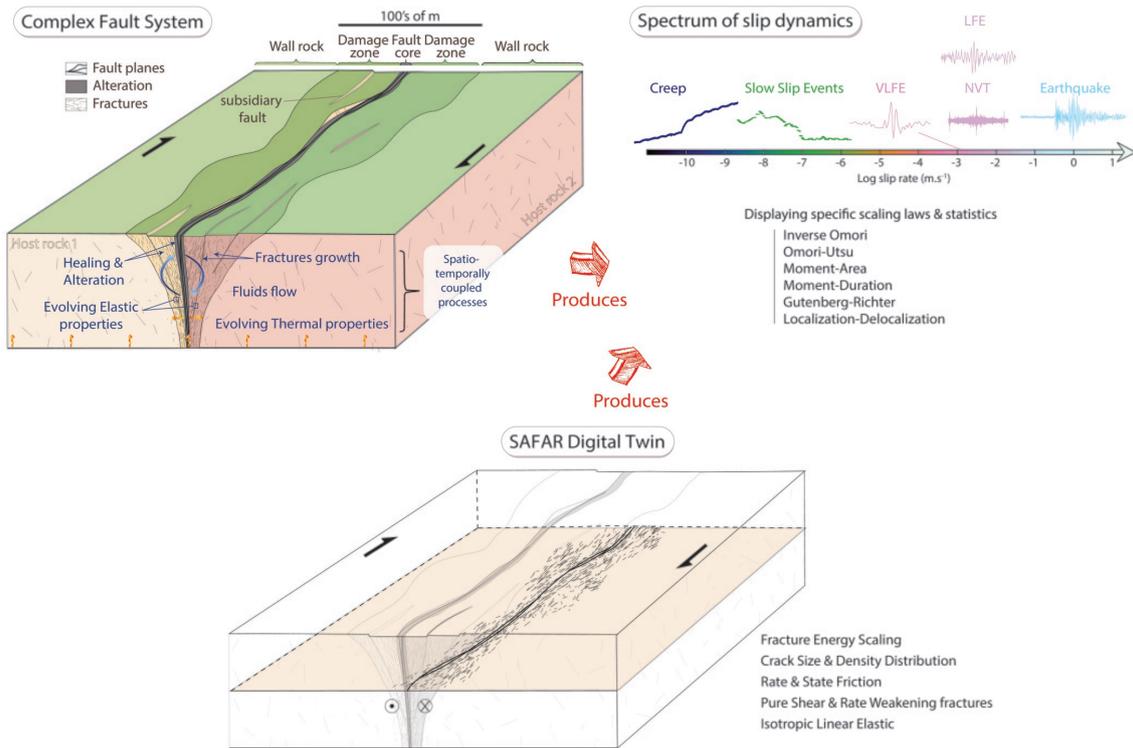


Figure 12: Summary of the digital twin model, SAFAR, that reproduces broadband slip dynamics, statistics and scaling laws of a fault volume.

We thus propose a geometry-driven modeling framework in which fault geometry provides a primary organizational structure for slip dynamics. Fault-zone geometry is an independently measurable and unavoidable property of natural systems across scales; in our framework, it naturally generates spatiotemporally evolving traction heterogeneities that govern nucleation, rupture velocity, and event interactions across the full spectrum of slip behaviors. This perspective helps reduce reliance on poorly constrained parameters and provides a physically grounded reference framework.

We do not intend to replace frictional or hydraulic heterogeneity models, but rather propose that geometric complexity constitutes a necessary baseline upon which these processes operate. Frictional and hydraulic heterogeneities may further modulate slip dynamics, but they act within a fault system whose stress organization is primarily shaped by geometry. This geometry-driven baseline provides a foundation within which additional complexities can be meaningfully evaluated and their relative contributions assessed. Such a framework enables the construction of digital twins of real fault systems—computational analogs that capture essential geometric and mechanical properties while generating realistic slip dynamics.

The fault volume model presented in this paper demonstrates the viability of this approach. Despite its simplicity and reliance on broadly accepted assumptions, it successfully reproduces many of the observed key physical and statistical features of fault zones (see Figure 12). We name this digital twin framework *SAFAR* (Système Analogue des FAilles Réelles). Our synthetic catalogs can enhance machine learning algorithms for seismic hazard forecasting by providing physically consistent training data that spans the full spectrum of seismic behaviors. The framework also provides data to analytically model various physical mechanisms that underlie the observed scaling laws, statistics, and slip dynamics. In addition, it can be used to characterize the response of a fault

system to various types of perturbations, such as fluid injection or earth tides, and to understand the underlying physics of earthquake triggering. By demonstrating that geometric complexity alone can generate rich fault dynamics without requiring unmeasurable frictional heterogeneities, this work provides a novel framework for more predictive and physically grounded seismic hazard assessment.

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Open Research Section

The datasets generated and analyzed during this study, and codes used to analyze, are available on the Zenodo repository: <https://doi.org/10.5281/zenodo.18599125>.

Supplementary Information for

“Fault volume digital twin to reproduce the full slip spectrum, scaling and statistical laws”

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Introduction

This Supplementary Information provides detailed technical documentation for the numerical methods, algorithms, and additional results supporting the main manuscript. The document is organized as follows:

Section S1 describes the quasi-dynamic earthquake cycle model

Section S2 presents the catalog building algorithm used to identify and characterize individual slip events from spatiotemporal slip rate matrices.

Supplementary Figures:

Figure S1 illustrates the catalog generation workflow.

Figure S2 demonstrates spatial segregation of fast and slow events in a fault system with off-fault fractures parallel to the main fault, using the same configuration as the main manuscript.

Figure S3 presents statistical properties and scaling laws

Figure S4 shows moment-duration scaling relationships with different detection thresholds.

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S1 Model of earthquake cycle

S1.1 Governing equation

A classic model of earthquake cycles contains: (1) an elastic medium that stores strain energy, (2) a physical mechanism that loads the system (plate tectonics), and finally (3) a frictional resistance on the fault, that allows the accumulation of strain on the fault. To build the model of earthquake cycles in this paper, we will use the linear momentum balance equation that states that the stress loading on the fault, and the elastic stress response due to a slip distribution on the fault, must be equal to the strength of this fault at each curvilinear point s along the fault. Mathematically, it is written as:

$$\tau^f(s) = \tau_t^{\text{el}}(s) + \tau^{\text{rad}}(s) + \tau^{\text{load}}(s) \quad (\text{S1.1})$$

Where $\tau^f(s)$ is the strength (model by a friction law) of the fault, $\tau_t^{\text{el}}(s)$ is the elastostatic tangential traction response due to the slip distribution on the fault, $\tau^{\text{rad}}(s)$ is the instantaneous response of the system that allows some inertia control (*Rice 1993*), these two latter modeling the reaction of the elastic medium due to the slip. Finally, $\tau^{\text{load}}(s)$ is the loading shear traction on the fault. Each of these terms will be described with more detailed in the following sections.

S1.2 Strength of the fault

We will model the strength of the fault by the regularized rate-and-state friction law (*Lapusta et al. 2000*):

$$\tau^f(s, t) = -f(V, \theta)\sigma_n(s, t) = -a\sigma_n(s, t) \sinh^{-1} \left[\frac{V(s, t)}{2V_0} \exp \left\{ \frac{f_0 + b \ln[V_0\theta(s, t)/d_c]}{a} \right\} \right] \quad (\text{S1.2})$$

with aging state evolution (*Dieterich 1979; Ruina 1983*):

$$\frac{d\theta(s, t)}{dt} = 1 - \frac{\theta(s, t)V(s, t)}{d_c} \quad (\text{S1.3})$$

a is the direct effect parameter that governs the instantaneous change in friction with a change in slip rate, b is the evolution effect parameter which controls how friction evolves over time via changes in the state variable, d_c is the characteristic slip for state evolution, V_0 is reference slip rate and f_0 is the reference friction at V_0 . V and θ represent respectively the slip velocity and the state variable. σ_n is the normal traction on the fault and d_c is the critical slip distance. The term $f_0 + b \log(V_0\theta/d_c)$ is sometimes denoted by Ψ .

S1.3 Elastostatic response of the fault due to a slip distribution

The normal traction and tangential traction on the fault, given a slip distribution $\Delta\mathbf{u}(s)$, and the fault local curvature $\kappa^t(s)$, in an infinite and homogeneous medium can be calculated using boundary element method (*Tada & Yamashita 1997; Romanet et al. 2020; Romanet et al. 2024*). For

convenience we will note the tangential slip on the fault $\Delta u_t(s) = \mathbf{t}(s) \cdot \Delta \mathbf{u}(s)$:

$$\tau_t^{\text{el}}(s) = \mathbf{t}(s) \cdot \bar{\bar{\sigma}}(\mathbf{y}(s)) \cdot \mathbf{n}(s) = \int_{\Gamma} K_{\text{grad}}^t(s, \xi) \frac{\partial}{\partial \xi} \Delta u_t(\xi) d\xi + \int_{\Gamma} K_{\text{curv}}^t(s, \xi) \kappa^t(\xi) \Delta u_t(\xi) d\xi \quad (\text{S1.4})$$

$$\sigma_n^{\text{el}}(s) = \mathbf{n}(s) \cdot \bar{\bar{\sigma}}(\mathbf{y}(s)) \cdot \mathbf{n}(s) = \int_{\Gamma} K_{\text{grad}}^n(s, \xi) \frac{\partial}{\partial \xi} \Delta u_t(\xi) d\xi + \int_{\Gamma} K_{\text{curv}}^n(s, \xi) \kappa^t(\xi) \Delta u_t(\xi) d\xi \quad (\text{S1.5})$$

In order to set the model of earthquake cycle, we chose the compression negative sign convention. s is the curvilinear location along the fault, $\Gamma \equiv \mathbf{y}(s)$, at which the stress is evaluated. ξ is a curvilinear location, at which a source of stress is located. $\mathbf{n}(\xi)$ and $\mathbf{t}(\xi)$ are respectively the normal and tangential vector to the fault at point ξ . Their component in the global coordinate system are written with the corresponding subscript. Finally, K_{grad}^t is the kernel for the tangential traction, and K_{grad}^n and K_{curv}^n are the kernel for the normal traction. K_{grad}^t and K_{curv}^t are the kernels for the tangential traction associated respectively to the effect of the gradient of tangential slip and local curvature that multiplies the tangential slip. They both derive from the fact that a derivative of the tangential slip vector along the fault is:

$$\begin{aligned} \frac{\partial}{\partial \xi} [\Delta u_t(\xi) \mathbf{t}(\xi)] &= \frac{\partial}{\partial \xi} [\Delta u_t(\xi)] \mathbf{t}(\xi) + \Delta u_t(\xi) \frac{\partial}{\partial \xi} \mathbf{t}(\xi) \\ &= \frac{\partial}{\partial \xi} [\Delta u_t(\xi)] \mathbf{t}(\xi) + \Delta u_t(\xi) \kappa^t(\xi) \mathbf{n}(\xi) \end{aligned} \quad (\text{S1.6})$$

where we have used the relationship $\frac{\partial}{\partial \xi} \mathbf{t}(\xi) = \kappa^t(\xi) \mathbf{n}$.

Similarly, K_{grad}^n and K_{curv}^n are the kernels for the normal traction. More details about the derivation of these equations can be found in [Romanet et al. \(2024\)](#). It should be noted here that these kernels take into account the full elastic interaction between all faults in the system, including the main fault and all off-fault fractures.

S1.4 Radiation damping term

The radiation damping term was first introduced by [Rice \(1993\)](#) in the context of quasidynamic modeling. In fact [Andrews \(1980\)](#) showed that this is the impedance of the fault (ratio of shear stress to slip velocity) in the long wavelength limit. It was later shown that this term is exactly accounting for the instantaneous shear stress drop of the fault due to sliding ([Cochard & Madariaga 1994](#)):

$$\tau^{\text{rad}}(s) = -\frac{\mu}{2c_s} V(s) \quad (\text{S1.7})$$

where c_s is the shear wave speed. We use this term together with the static kernel to account for some dynamics in the system. Without this term, the slip on the fault during one event would be unbounded ([Rice 1993](#)).

S1.5 Set of ordinary differential equations for Quasi-Dynamic earthquake cycle models

Balance of forces requires the strength of the fault to be equal to the elastic shear traction (due to slip) plus the far-field loading traction plus radiation damping term (eq. S1.1).

By differentiating eq. S1.1 with time t , it can be recast into a set of coupled ODEs. The slip acceleration becomes:

$$\frac{dV(s, t)}{dt} = \frac{\dot{\tau}^{\text{load}}(s, t) + \dot{\tau}_t^{\text{el}}(s, t) + \frac{\partial f(V, \theta)}{\partial \theta} \frac{d\theta}{dt} \sigma_n(s, t) + f(V, \theta) \dot{\sigma}_n(s, t)}{\frac{\mu}{2c_s} - \frac{\partial f(V, \theta)}{\partial V} \sigma_n(s, t)} \quad (\text{S1.8})$$

From this equation, it is easy to see that the denominator would go to zero and hence the acceleration would go to infinity if there was no radiation damping term (*Rice 1993*).

The elastic shear traction rate is given by:

$$\dot{\tau}_t^{\text{el}}(s, t) = \int_{\Gamma} K_{\text{grad}}^t(s, \xi) \frac{\partial}{\partial \xi} V(\xi, t) d\xi + \int_{\Gamma} K_{\text{curv}}^t(s, \xi) \kappa^t(\xi) V(\xi, t) d\xi \quad (\text{S1.9})$$

The change of normal traction has two sources: one due from the loading rate and another one due to the reaction to slip of the elastic medium:

$$\frac{d\sigma_n(s, t)}{dt} = \int_{\text{faults}} K_{\text{grad}}^n(s, \xi) \frac{\partial}{\partial \xi} V(\xi, t) d\xi + \int_{\text{faults}} K_{\text{curv}}^n(s, \xi) \kappa^t(\xi) V(\xi, t) d\xi + \dot{\sigma}_n^{\text{load}}(s, t) \quad (\text{S1.10})$$

And we recall the evolution of the state variable eq. S1.3. This set of ODEs is then solved at each centre of element, using the Runge-Kutta 45 adaptive time step ODE solver algorithm (*Cash & Karp 1990*).

S1.6 Numerical discretisation of the Boundary Integral Equation

In order to evaluate the previous singular integrals in the sense of Cauchy principal values, we will assume piece-wise constant slip over fixed length Δs , centred on $\mathbf{y}(s_i)$. The slip is discretised as follows (*Rice 1993; Cochard & Madariaga 1994*):

$$\Delta u(s) = \sum_{j=1}^N \Delta u(s_j) [\mathcal{H}(s - s_j + \Delta s/2) - \mathcal{H}(s - s_j - \Delta s/2)] \quad (\text{S1.11})$$

where \mathcal{H} is the Heaviside function and N is the number of elements used to discretise the fault. Because the kernels imply the evaluation of the tangential vector, it is also needed to discretised

it along the fault:

$$\mathbf{t}(s) = \sum_{j=1}^N \mathbf{t}(s_j) [\mathcal{H}(s - s_j + \Delta s/2) - \mathcal{H}(s - s_j - \Delta s/2)] \quad (\text{S1.12})$$

Then the boundary integral equations S1.4 and S1.5 become a summation:

$$\begin{aligned} \tau(s_i) &= \sum_{j=1}^N [K_t(s_i, \xi_j) - K_t(s_i, \xi_{j+1})] \Delta u_j \\ \sigma(s_i) &= \sum_{j=1}^N [K_n(s_i, \xi_j) - K_n(s_i, \xi_{j+1})] \Delta u_j \end{aligned} \quad (\text{S1.13})$$

The implementation of this formulation requires to calculate for each $\tau(s_i)$ (N terms) a sum over N terms (for each j), which leads to a computational complexity of $\mathcal{O}(N^2)$. It means that the computational time will grow with the square of the problem size. This makes it difficult to handle large problems with a straightforward implementation. However, for the majority of the cases, the kernel K is smooth when the source point $\mathbf{y}(\xi_j)$ is far enough from the evaluated point $\mathbf{y}(s_i)$, or in other words: $|\mathbf{y}(s_i) - \mathbf{y}(\xi_j)| \gg 1$. Several methods can be used to accelerate the evaluation of eqns. S1.4 and S1.5. The Fast Fourier Transform was the first method used to accelerate the calculation (Andrews 1985), however it requires equispaced points on the fault. The only configuration where this can be achieved is on a simple planar fault. Fast Fourier Transform has been widely used in numerical models of seismic cycles (Lapusta et al. 2000; Lapusta & Liu 2009; Chen & Lapusta 2009; Michel et al. 2017). The Fast Multipole Method allows for considering a complex geometry, and was used by some numerical models (Hirahara et al. 2009; Romanet et al. 2018), however it requires an analytical development of the kernel. In this paper, we employ Hierarchical matrices which allows for an algebraic development of the kernel and thus can be applied to a large variety of problems. This approach is particularly optimal for static problems. This will be discussed in the next section.

S1.7 Hierarchical Matrices

Hierarchical matrices have already been used in the context of quasi dynamic modeling of faults (Bradley 2014; Romanet 2017; Cheng et al. 2025). Hierarchical matrices provide data-sparse approximations of non-sparse (dense) matrices (Hackbusch 1999; Börm et al. 2003). Hierarchical matrices provide an approximation requiring only $\mathcal{O}(N \log N)$ units of storage (instead of $\mathcal{O}(N^2)$). The two advantages of using H-matrices are that it reduces the memory for saving the full matrix of interaction, and it is also reducing the number of operations to perform the matrix-vector products. The use of H matrices is extremely pertinent in quasidynamic models. The kernel matrix of stress interaction S1.13 is not changing over time, making it a suitable matrix for compression, since the most expensive part, the construction of the data-sparse representation, is performed only once, while subsequent matrix-vector products are extremely fast. The accuracy of this method is completely controlled by the tolerance set for the compression, particularly if the required minimum accuracy approaches machine precision. The mathematical formulation of hierarchical matrices is beyond the scope of this paper but is very well explained in Börm et al.

(2003), *Desiderio (2017)*, *Romanet (2017)*, and *Cheng et al. (2025)*.

S1.8 Frictional length scales

Although the general answer is unavailable in the general framework of rate and state, it is possible to infer a length scale in some limiting cases for simple straight continuum faults with homogeneous frictional parameters. The first nucleation length-scale derived for continuous fault came from the spring slider modeling. If it is assumed that the stiffness of a fault k_{fault} is inversely proportional to its length, we can derive a nucleation length scale L_{nuc} (*Rice 1992*). At the critical length L_{nuc} , the stiffness of the fault equals the critical stiffness derived from a spring slider (*Rice 1992*).

$$k_c = \sigma_n \frac{b-a}{d_c} = \frac{\mu}{L_{nuc}} = k_{fault} \quad (S1.14)$$

Thus,

$$L_{nuc} = \frac{\mu d_c}{\sigma_n (b-a)} \quad (S1.15)$$

By studying the nucleation on fault with rate and state resistance, Dieterich derived another nucleation length scale L_b inversely proportional to the friction parameter b (*Dieterich 1992*).

$$L_b = \frac{\mu d_c}{\sigma_n b} \quad (S1.16)$$

More recent work shows that the nucleation actually depends on the ratio a/b (*Rubin & Ampuero 2005; Ampuero & Rubin 2008; Viesca 2016*). *Rubin & Ampuero (2005)* first derived analytical solution in the limiting case where $V\theta/D_c \gg 1$ for the aging state evolution law. They showed that this assumption would remain valid only if the ratio $a/b < 0.3781$. For high a/b however, they pointed out that the coefficient $V\theta/D_c$ was nearly constant at the interior of the nucleation patch. Using that as an assumption, with energetic consideration, they were able to derive another expression for the nucleation length scale when a/b approaches 1. We can summarize their result by:

$$\begin{cases} L_{nuc} = 2 \times 1.3774 L_b & 0 \leq a/b < 0.3781 \\ L_{nuc} = 2 \times \frac{L_b}{\pi(1-a/b)^2} & a/b \rightarrow 1 \end{cases} \quad (S1.17)$$

The nucleation length scale found by *Rubin & Ampuero (2005)*, was later shown to hold true by recasting the system of equations to look for instabilities and doing a linear perturbation analysis of these instabilities (*Viesca 2016*).

S2 Catalog building algorithm

To identify events that are spatiotemporally contiguous we use a method called Connected Component Labeling with Bounding Box Extraction. Several widely used software packages support connected component labeling (CCL) and region property extraction. In MATLAB, the `bwlabel` and `regionprops` functions provide robust tools for labeling and extracting geometric features such as bounding boxes and centroids (*The MathWorks, Inc. 2024*). In Python, the `scikit-image` library offers similar functionality (*Walt et al. 2014*). OpenCV, a popular C++ and Python computer vision library, includes highly optimized functions for fast labeling and region statistics (*Bradski*

Algorithm 1 Identify events

```

1: procedure CONNECTEDCOMPONENTSWITHPROPERTIES(image, rows, cols)
2:   // image : binary matrix where 1 indicates active pixel, 0 indicates background
3:   // labels : integer matrix where each nonzero entry is the region label
4:   // properties : array of structures with min/max row/column indices per region
5:   Initialize labels[rows, cols]  $\leftarrow$  0
6:   current_label  $\leftarrow$  0
7:   Define offsets:  $dr \leftarrow [-1, -1, -1, 0, 0, 1, 1, 1]$ ,  $dc \leftarrow [-1, 0, 1, -1, 1, -1, 0, 1]$ 
8:   for col  $\leftarrow$  1 to cols do
9:     for row  $\leftarrow$  1 to rows do
10:      if image[row, col] = 1 and labels[row, col] = 0 then
11:        current_label  $\leftarrow$  current_label + 1
12:        properties[current_label]: min/max row/col  $\leftarrow$  row, col
13:        LABELREGIONS(row, col, current_label)
14:      end if
15:    end for
16:  end for
17:  return labels, properties, current_label
18: end procedure
19: procedure LABELREGIONS(r, c, label)
20:   if r or c out of bounds or not 1 or already labeled then return
21:   end if
22:   labels[r, c]  $\leftarrow$  label
23:   Update properties[label] with min/max of r, c
24:   for i  $\leftarrow$  1 to 8 do
25:     nr  $\leftarrow$  r + dr[i], nc  $\leftarrow$  c + dc[i]
26:     LABELREGIONS(nr, nc, label)
27:   end for
28: end procedure

```

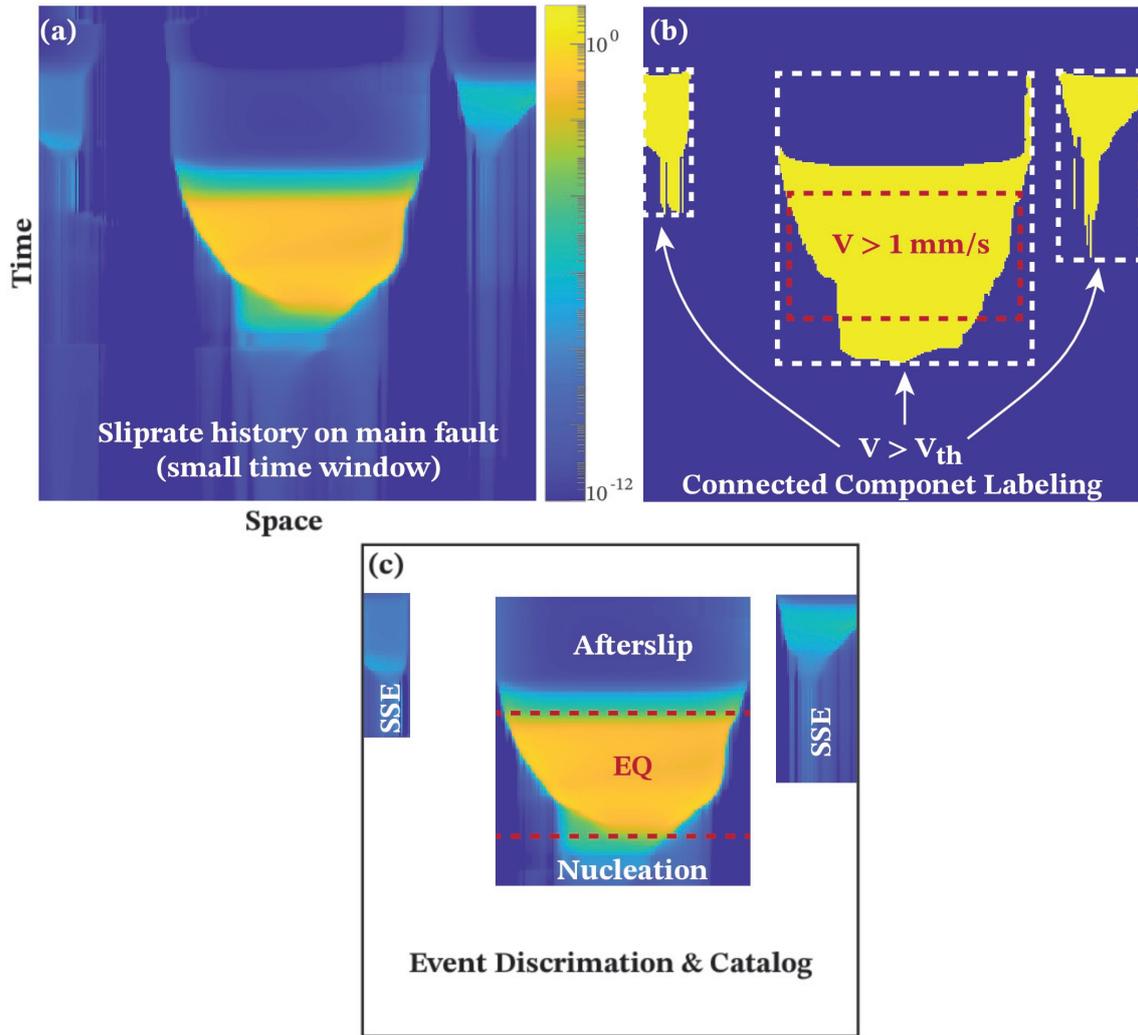


Figure S1: Steps involved in catalog generation from slip rate matrix. (a) Slip rate matrix (time, space). (b) Binary image after thresholding slip rates with Labeled connected components. (c) Bounding boxes around each connected component, representing individual events.

2000). In this work we decided to implement this ourselves as the algorithm is trivial and it's easier to adapt the results of our code to this algorithm (see Algorithm 1).

We first take a slip rate matrix (time, space) and convert into a binary image by setting to 0 all values that are below a velocity threshold. The algorithm then performs connected component analysis on this binary image using a recursive depth-first search (DFS) strategy to identify and label contiguous regions of foreground pixels (i.e., pixels with value 1). Each connected region is assigned a unique integer label, and its spatial extent is recorded as a bounding box defined by the minimum and maximum row and column indices. This type of region labeling is a fundamental technique in image analysis, commonly used for shape recognition, segmentation, and morphological operations. The DFS uses an 8-connectivity scheme, meaning that each pixel is connected to its horizontal, vertical, and diagonal neighbors. This is particularly useful in natural or irregular structures where connectivity extends beyond 4-neighbor (Manhattan) adjacency. The use of recursive DFS provides an intuitive implementation but is best suited for small to moderately sized images due to potential stack overflow risks. For larger datasets, stack-based or union-find methods may be preferred (Rosenfeld & Pfaltz 1966; Shapiro & Stockman 2001).

The procedure begins by scanning the image in a raster order. When an unlabeled foreground pixel

is encountered, a new label is assigned, and a recursive DFS is initiated from that pixel to visit all 8-connected neighboring pixels belonging to the same region. During traversal, the algorithm updates a property array that maintains the axis-aligned bounding box for each region in the form (min_row, max_row, min_col, max_col). Once the beginning and the end of the event (min_row, max_row) and its spatial extent (min_col, max_col) are obtained its trivial to compute the Moment, average slip, duration and other catalog based quantities.

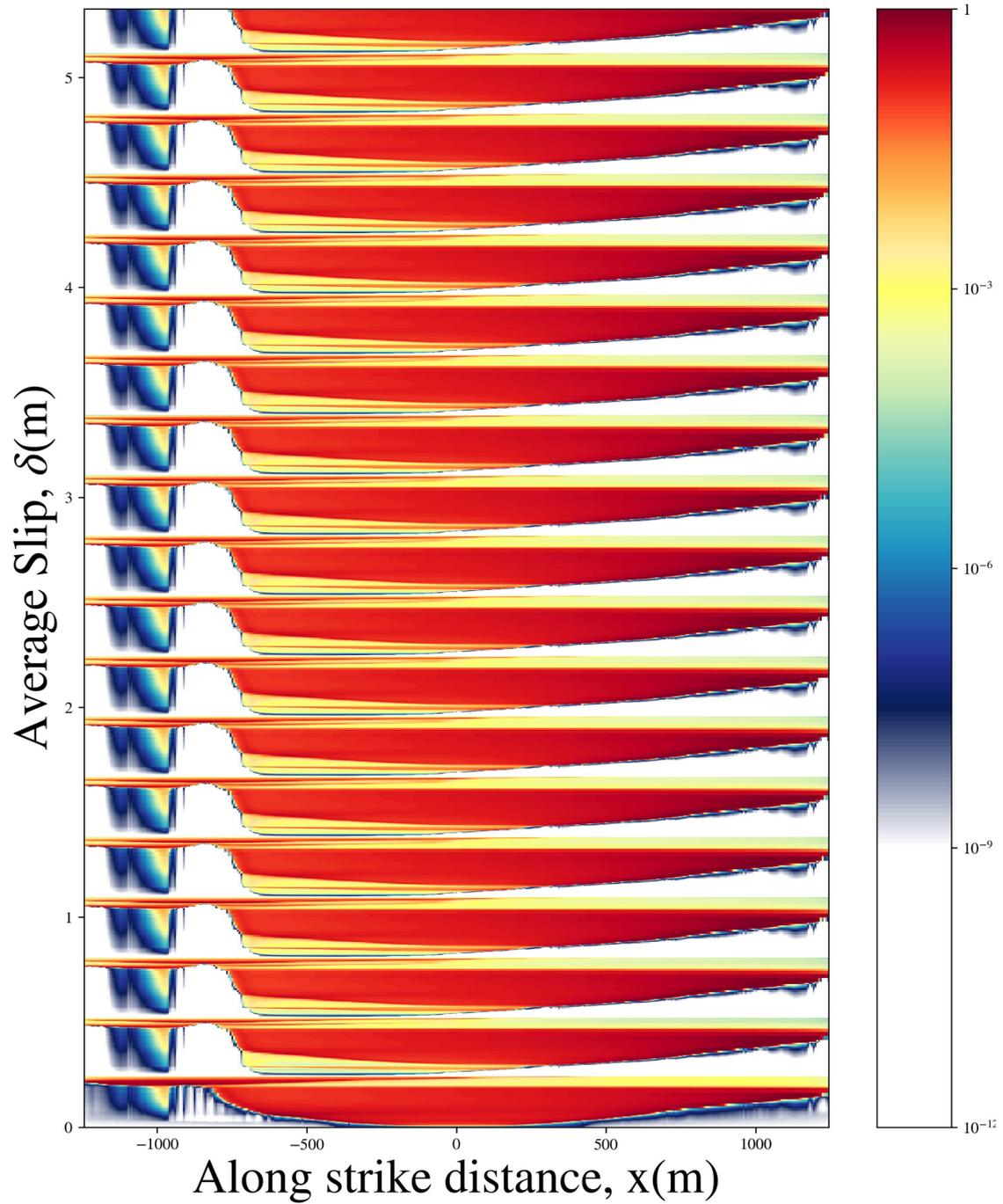


Figure S2: Segregation of slow and fast events at two separated regions of the fault. The geometrical configuration is the same as in Figure 1a except that the off-fault fractures are parallel to the main fault.

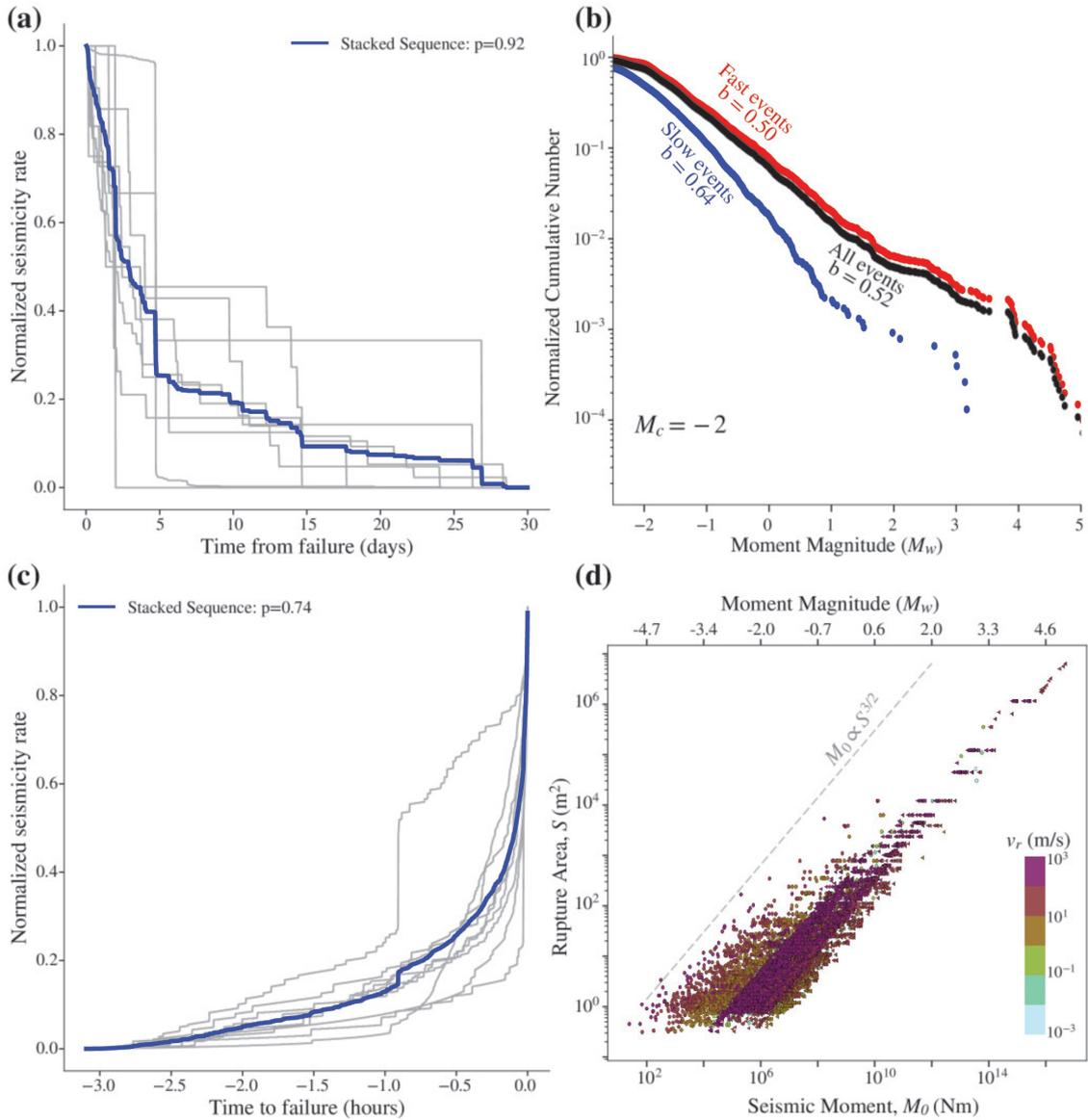


Figure S3: (a) Omori Law: Seismicity rate decreases over time following the mainshock. Each gray curve represents one earthquake cycle with one mainshock on the main fault, the blue curve represents the stacked sequences of all the seismic cycles in this simulation case (b) Magnitude-frequency distribution of the cataloged fast ruptures follows the Gutenberg-Richter distribution. Red and blue colors represent fast and slow ruptures respectively, while black color represents the total catalog (c) Inverse Omori: Seismicity rate increases inversely with time as the main rupture approaches. Each gray curve represents one earthquake cycle with one mainshock on the main fault, the blue curve represents the stacked sequences of all the seismic cycles in this simulation case (d) Scaling laws of inferred seismicity: Moment area scaling. Events are color-coded based on rupture velocity v_r .

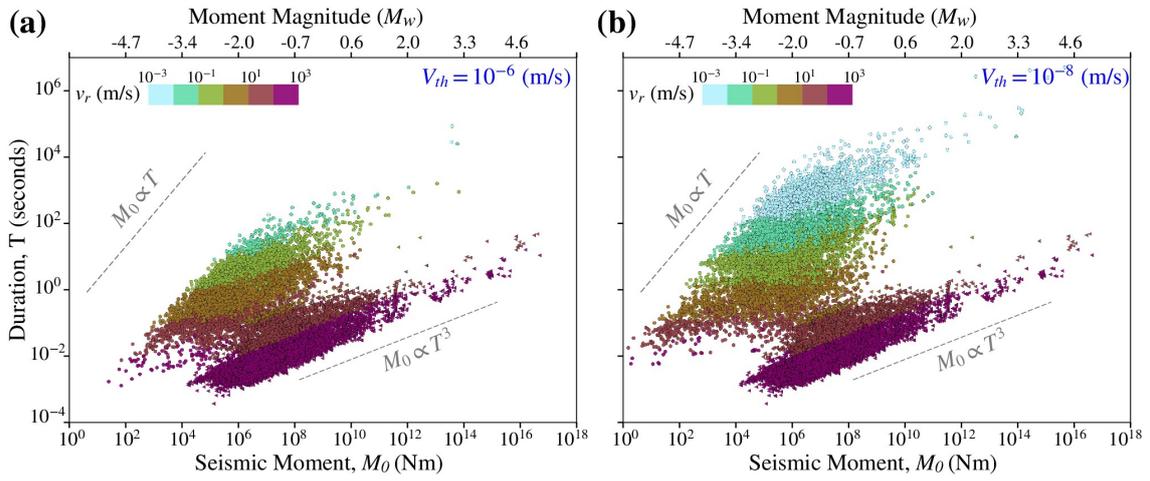


Figure S4: Scaling laws of inferred seismicity: Moment-duration scaling with a threshold of detection of slow ruptures at (a) 10^{-6} m/s and (b) 10^{-8} m/s. Events are color-coded based on rupture velocity v_r .

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