Physical Mechanisms of Earthquake Nucleation and Foreshocks: Cascade Triggering, Aseismic Slip, or Fluid Flows?

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Abstract Earthquakes are caused by the rapid rupture of their seismogenic faults. Whether large or small, there is inevitably a certain nucleation process involved before the dynamic rupture. At the same time, although not very common, significant foreshock activity has been observed before some large earthquakes. Understanding the nucleation process and foreshocks of earthquakes, especially large damaging ones, is crucial for accurate earthquake prediction and seismic hazard mitigation. The physical mechanism of earthquake nucleation and foreshock generation is still in debate. While earthquake nucleation process is present in the laboratory experiments and numerical simulations, it is difficult to observe such process directly in the field. In addition, it is currently impossible to effectively distinguish foreshocks from ordinary earthquake sequences. In this article we first summarize foreshock observations in the last decades and attempt to classify them into different types based on their temporal behaviors. Next, we present different mechanisms for earthquake nucleation and foreshocks that have been proposed so far. These physical models can be largely grouped into the following three categories: elastic stress triggering, aseismic slip, and fluid flows. We also review several recent studies of foreshock sequences before moderate to large earthquakes, focusing on how different results/conclusions can be made by different datasets/methods. Finally, we offer some suggestions on how to move forward on the research topic of earthquake nucleation and foreshock mechanisms and their governing factors.

Introduction

Earthquakes are caused by the rapid dynamic rupture of their source faults. Recent studies have found a wide range of fault slip behaviors, ranging from creep, aseismic fault slip, pre-slip, afterslip, and regular earthquake ruptures (Peng and Gomberg, 2010; Avouac, 2015; Harris, 2017; Bürgmann, 2018; He et al., 2023; Ide and Beroza, 2023). However, the unstable fault slip must involve a development process (i.e., nucleation) from static to dynamic including a pre-slip phase (Dieterich, 1992; Ohnaka, 1992, 2013; Ampuero and Rubin, 2008). Here we define **pre-slip** as the aseismic slip that precedes the irreversible dynamic fault rupture. Laboratory experiments (Lei and Ma, 2014; McLaskey, 2019; Yamashita et al., 2021; Goebel et al., 2024) and numerical studies (Cattania and Segall, 2021; He et al., 2023) have shown that this process can greatly vary in time and space scales depending on factors such as the uniformity of the fault geometry and frictional property. In general, the more uniform the fault, the shorter duration time and smaller the nucleation size, making it impossible for us to observe any precursory phenomena (Yamashita et al., 2021). Other recent studies also suggest that geometric complexities, frictional properties and fault orientations with respect to the maximum ambient stress directions play important roles in determining the fault slip mode of individual earthquakes, their collective behaviors, and nucleation processes before large earthquakes (Ben-Zion, 2008; Gabriel et al., 2024; Lee et al., 2024).

Foreshocks are a group of small to moderate-size earthquakes occurring in a relatively short space-time window before a larger earthquake (i.e., the **mainshock**) (Mogi, 1963). According to this definition, foreshocks can only be defined after the occurrence of the **mainshock** (i.e., the largest event in a sequence). Indeed, about 1/3 to 1/2 of the large mainshocks have foreshocks (Reasenberg, 1999; Chen and Shearer, 2016). However, this number depends strongly on which earthquake catalogs are used, and how foreshocks are defined (Trugman and Ross, 2019; van den Ende and Ampuero, 2020; Moutote et al., 2021). On the other hand, only 5-10% of small-size earthquakes are followed within one week of a larger magnitude event (Reasenberg, 1999; Christophersen and Smith, 2008). For example, the successful short-term earthquake forecast for the 1975 M7.3 Haicheng earthquake in northeast China (Wang et al., 2006) was based on observed foreshock swarm initiated two days before the earthquake. However, there were many similar swarms observed in Chinese mainland, which were not followed by larger earthquakes (Chen et al., 1999; Lei et al., 2024). These numbers and rare individual cases highlight the challenges to discriminate foreshocks from other earthquake sequences (Ogata et al., 1996) and use them as a reliable precursor for upcoming large earthquake (Zaccagnino et al., 2024). In this paper, the term "**earthquake precursor**" refers to a specific concept: it describes phenomena occurring near the epicenter before a major earthquake that can be reasonably explained based on current understanding and are intrinsically linked to the nucleation process of the earthquake.

An alternative approach is to estimate the probability of future earthquakes in near real time based on the past and current seismic activity and statistical forecasting models, such as the Epidemic Type Aftershock-Sequences (ETAS) model (Ogata, 1988). The ETAS model uses the established relationship in statistical seismology (Utsu et al., 1995; Utsu, 2002) such as the Omori's law for aftershock decay (Omori, 1894) and the Gutenberg-Richter magnitude-frequency relationship (Gutenberg and Richter, 1944). This is also known as **operational earthquake forecasting (OEF)** (Jordan et al., 2011, 2014), which can be used for large aftershock forecasting (Ogata, 2017; Hardebeck et al., 2024; Mizrahi et al., 2024), or twin ruptures (i.e., two earthquakes of similar sizes) (Fukushima et al., 2023). The most recent example is the 'megaquake advisory' along the Nankai trough in southwest Japan issued by Japan Meteorological Agency (JMA) following a magnitude 7.1 earthquake off the coast of Kyushu Island in southwest Japan on August 8, 2024 (Toda et al., 2024). However, its ability to accurately forecast future large mainshocks is likely limited, given the fact that most triggered earthquakes are aftershocks that are smaller than the parent triggering event, and only a small percentage of them are larger than the triggering event (i.e., foreshocks).

In this article, we provide a brief review of precursory signals before large earthquakes, focusing primarily on seismic (i.e., foreshocks) and aseismic slip behaviors, and various physical mechanisms that have been proposed so far for earthquake nucleation. We highlight primary differences in the results from several well-studied foreshock sequences and offer possible explanations. We do not provide a comprehensive review of other types of precursory signals, but instead refer the readers to the following papers on aseismic deformation before large earthquakes (Roeloffs, 2006), radon gases (Riggio and Santulin, 2015), laboratory acoustic emissions (Lei and Ma, 2014), earthquake nucleation (Ohnaka, 2013), seismo-electromagnetic precursors (Freund, 2011; Chen et al., 2022), foreshocks and mainshock nucleation (Ellsworth, 2019; Kato and Ben-Zion, 2021; He et al., 2023; Martínez-Garzón and Poli, 2024), and insights gained from injection-induced earthquakes (Lei et al., 2020; Ge and Saar, 2022; Moein et al., 2023). Finally, and most importantly, we point out potential future directions focusing on physical mechanisms of earthquake nucleation and foreshocks and their governing factors.

Foreshocks and Other Precursory Signals

There is a long history of both failed attempts and a few limited successes in studying earthquake precursors and earthquake prediction (Bakun et al., 2005; Roeloffs, 2006; Wang et al., 2006; Ouzounov et al., 2018). For example, the aforementioned 1975 M7.3 Haicheng earthquake was preceded by a swarm of foreshocks that occurred on the conjugate faults in the extensional step-over zone in the middle of the main rupture fault and lasted for about two days (Jones et al., 1982; Chen et al., 1999; Lei et al., 2024). While there is still some argument on whether the Haicheng earthquake prediction was considered as a true success or not, the foreshock sequence was one of the most important precursors (Wang et al., 2006). In any case, this is the only instance so far where the government issued an earthquake prediction and evacuation mobilization, significantly reducing casualties from the subsequent mainshock. As the 50-year anniversary of the Haicheng earthquake prediction is approaching, it is important to revisit what progress has been made in the past half century on our understanding of foreshocks, earthquake nucleation and precursory signals.

Partially motivated by the apparent success of the Haicheng earthquake prediction, and the development of plate tectonics and rock dilatancy theory (Scholz et al., 1973) in the early 1970s, there was a heightened sense of optimism that reliable earthquake precursors existed, and earthquake prediction was just around the corner. Based on the apparent 22-year recurrence of similar M6-type earthquake along the Parkfield section of the San Andreas Fault in Central California and their similar characteristics (Bakun and McEvilly, 1979), the Parkfield earthquake prediction experiment was proposed (Bakun and Lindh, 1985), with the expected event between 1988 and 1993. The long-waited M6 Parkfield earthquake did occur in 2004, arguably the most well recorded event around the world at that time. However, it was not preceded by any observable precursors (i.e., foreshocks or pre-slip) (Bakun et al., 2005). This event apparently initiated at the other side of fault segment that was broken during the 1934 and the 1996 M6 Parkfield earthquake, which were preceded by M5 foreshocks (Bakun and McEvilly, 1979). This study, along with other studies during early 2000s (Roeloffs, 2006), highlighted the difficulty in detecting reliable precursors for short-term earthquake prediction.

Nevertheless, there is a renewed interest in studying earthquake precursory phenomenon in the last few years (Pritchard et al., 2020), largely driven by improved observations from both near-field (including off-shore) seismic and geodetic recordings (Ben-Zion et al., 2022), availability of big data in seismology (Arrowsmith et al., 2022), and promises in artificial intelligence (AI)/Machine Learning (ML) methods in seismology (Beroza et al., 2021; Mousavi and Beroza, 2023) and other domains (Wang et al., 2023). Using spatio-temporal changes in background seismicity and continuous GNSS recordings, recent studies have found clear anomalies up to several years for earthquakes in Japan (Ogata, 2005, 2007). A recent analysis of high-rate GNSS time series also found a 2-hour-long exponential acceleration of aseismic slip before 90 Mw>7 earthquakes around the globe (Bletery and Nocquet, 2023, 2024), although the results were challenged by subsequent re-analysis (Bradley and Hubbard, 2023, 2024). Similarly, migrating foreshocks, repeating earthquakes (Kato et al., 2012) and offshore geodetic observations (Ito et al., 2013) all suggested the existence of slow slip before the 2011 M9 Tohoku-Oki earthquake. However, stacked tiltmeter records did not show accelerating-like deformation starting two hours before the mainshock (Hirose et al., 2024), suggesting that such aseismic slip, if exists, would be less than a moment magnitude Mw 6.4 earthquake. In addition, it cannot be determined whether it links to the pre-slip of the mainshock. On the other hand, migratory aseismic slip and abundant foreshocks are frequently observed for daily ice stream earthquakes at the Whillans Ice Plain, West Antarctica (Barcheck et al., 2021). The month-long swarm-like foreshocks before the 2014 M8.1 Iquique earthquake sequence in Northern Chile was also likely driven by geodetically observable slow-slip events, likely reflecting the pre-slip of the mainshock (Ruiz et al., 2014).

Aseismic slip or fluid migration have been invoked to as the primary physical mechanisms driving the foreshock sequence along oceanic transform faults (McGuire et al., 2005; Liu et al., 2020) and continental settings (Zhou et al., 2022; Wang et al., 2024a; Lei et al., 2024). More recently, both aseismic slip and elevated fluids are found to work in concert to drive a prolonged earthquake swarm that started beneath the Noto Peninsula in Central Japan since November 2020 (Amezawa et al., 2023; Kato, 2024; Nishimura et al., 2023; Yoshida et al., 2023a, b), which culminated in the 2024 M7.5 Noto mainshock (Toda and Stein, 2024; Ishikawa and Bai, 2024). The Noto mainshock was preceded by three immediate foreshocks in the magnitude ranges of M5-6 in the last 4, 2 minutes and 14 seconds before, likely reflecting a change from the long-term swarm-like behavior to a burst-like foreshock sequence (Peng et al., 2024). Geochemical analysis in the Noto epicentral region also indicated that the fluid sources are likely from upper mantle (Umeda et al., 2024). Similarly, several studies also found possible evidence of prolonged preparation processes before the 2023 M7.8 Pazarcik, Türkiye earthquake (Picozzi et al., 2023; Kwiatek et al., 2023), although no immediate foreshocks were identified.

In summary, while in some cases aseismic process or fluid flows have been observed before large earthquakes (e.g., the most recent M7.5 Noto sequence), these processes are not clearly present in other well studied cases (i.e., the 2004 M6 Parkfield earthquake), highlighting again the challenge of identifying a reliable precursory signal before moderate to large earthquakes. Here, understanding the dominant factors that cause such differences is a crucial task.

Foreshock Types

Mogi (1985) attempted to define foreshocks of two types. In the first type (type C in his definition), seismic activity gradually increases toward the mainshock. In the second type (type D in his definition), the increasing foreshock activity dies down before the occurrence of the mainshock. However, foreshock observations in the past few decades suggest a much more complex patterns than those two types. Based on their temporal behaviors, here we attempt to classify foreshock behaviors into the following types: no immediate foreshocks/quiescence, accelerating, burst, swarm and two hybrid types (Fig. 1). Sometimes large earthquakes occur without obvious foreshocks and show an apparent quiescence in the last few days/weeks preceding the mainshock (Fig. 1a). For example, while there were some increases of seismic activity and slow slip derived from repeating earthquakes in a broader region starting a few years to a few months before, there were no immediate foreshocks (i.e., in the last few days) before the 2008 M7.9 Wenchuan, China earthquake (Li et al., 2011; Yuan et al., 2017) and the 2023 M 7.8 Pazarcik, Türkiye, earthquake, the first mainshock of the Kahramanmaraş sequence (Kwiatek et al., 2023; Picozzi et al., 2023). The *accelerating-type* foreshocks (Fig. 1b) are mostly found in laboratory settings as acoustic emissions (Lei & Ma, 2014; McLaskey, 2019; Yamashita et al., 2021; Bolton et al., 2023; Goebel et al., 2024; Lei, 2024). In natural setting, except in a few cases (e.g., Bouchon et al., 2011) or for stacked sequences (Bouchon et al., 2013; Felzer et al., 2015; Shearer et al., 2023), most foreshocks do not show clear accelerating patterns before the mainshock. In the *burst-type* (Fig. 1c), a moderate-size foreshock would be followed by its own aftershocks that decays with time, and then the larger mainshock happens. Examples include the 2010 M6.7 Yushu earthquake (Ni et al., 2010; Chuang et al., 2023; Huang et al., 2023), the 2014

M7.3 Yutian earthquake (Li et al., 2024), the 2016 M7.0 Kumamoto earthquake (Kato et al., 2016) and the 2019 M7.1 Ridgecrest earthquake (Ross et al., 2019b). In the *swarm-type* (Fig. 1d), the mainshock is preceded by an earthquake swarm without a clear dominant event or there are some events with progressively increase of magnitude before the largest foreshock (i.e., opposite of the third burst-type). Examples of this type include the 1975 M7.3 Haicheng earthquake (Jones et al., 1982; Chen et al., 1999; Lei et al., 2024) and the 2010 M7.2 El-Mayor Cuccapah earthquake (Hauksson et al., 2011; Yao et al., 2020). In the *first hybrid-type* (Fig. 1e), the foreshock activity typically starts with a swarm-type sequence that can last weeks or months. Right before the mainshock, there is clear change of the behaviors from swarm-type to burst-type, which would include one or a few moderate-size foreshocks with their own aftershocks, and the mainshock itself. Examples of this type include the 2009 M6.1 L'Aquila earthquake (Cabrera et al., 2022), the 2011 M9.1 Tohoku-Oki earthquake (Kato et al., 2012), the 2014 M8.1 Iquique earthquake (Ruiz et al., 2014; Kato and Nakagawa, 2014), and the 2021 M6.1 Yangbi earthquake (Lei et al., 2021; Zhou et al., 2022; Zhu et al., 2022; Liu et al., 2022; Wang et al., 2024a). In the *second hybrid-type* (Fig. 1f), a mainshock was preceded by a swarm-like sequence a few months or years before the eventual mainshock, followed by an apparent quiescence, and then accelerating-type foreshocks occurred right before the mainshock. A well-known example is the 1923 M7.9 Great Kanto earthquake sequence in Japan (Ohnaka et al., 1984). Depending on how the last 4 minutes of the foreshock activity is classified, the foreshock sequence of the 2024 Noto Peninsula earthquake can be viewed as either of the two hybrid types (Shelly, 2024; Peng et al., 2024).

We note that the above classification is not strict or mutually exclusive. Besides the two hybrid types, transitional types may also occur. For example, the entire foreshock sequence of the 2021 M6.1 Yangbi earthquake can be either viewed as the swarm-type. However, it includes three episodes, and the first two started with moderate-size earthquakes followed by their own aftershocks (Lei et al., 2021; Zhu et al., 2022; Zhou et al., 2022). Additionally, depending on the inspected time windows, we would end up with different types. For example, while the overall foreshock behavior of the foreshock for the 2010 M7.2 El-Mayor Cuccapah mainshock Hauksson et al., 2011; Yao et al., 2020) was swarm-like, if we zoom in in the last day and especially in the last hours, there were several M>3 earthquakes in the last 6 minutes of the 2010 M7.2 El-Mayor Cuccapah mainshock (Yao et al., 2020), which clearly did not follow the overall swarm-type behaviors in the last two days. In this case, it would likely be classified as one of the hybrid-type, rather than the swarm-type. Again, it is an important issue to understand the governing factors behind each foreshock type.

Physical Mechanisms of Foreshocks

One motivation for such classification for foreshocks is that it may reflect differences in their driving mechanisms and structures within the source region. In general, each burst in the burst-type foreshocks (Fig. 1c) behaves like a mainshock-aftershock sequence, and the total foreshocks could be simply modeled by statistical processes such as the Omori's law for aftershock decay (Utsu et al., 1995; Utsu, 2002; Kanamori and Brodsky, 2004) or the ETAS model, which encompasses both the Omori law and the Gutenberg-Richter law of magnitude-frequency distribution (Ogata, 1988, 2017). Their similar behaviors to aftershock sequence also suggest a similar physical mechanism for both foreshocks and aftershocks. On the other hand, swarm-like behaviors can also be modeled using ETAS but requires considering external driving factors that may vary over time (Petrillo et al., 2024), such as episodic fluid flow or slow-slip events, within the background seismic activity (Vidale and Shearer, 2006). Similar approaches have been successively applied to injection-induced seismic activities to account the role of injection operations (e.g., Lei et al., 2019; Lei et al., 2020; Jia et al., 2023; Moein et al., 2023).

Fig. 2 summarize five physical models of earthquake nucleation and foreshocks that have been proposed so far. The first three models are presented in McLaskey (2019), while the last two are from two recent studies (Lei et al., 2024; Wang et al., 2024a). The *pre-slip* model (Fig. 2a) asserts that precursory aseismic slip occurs during earthquake nucleation, which can drive both foreshocks as well as the initial rupture of the mainshock (Dieterich, 1992; Ohnaka, 1992, 2013; Dodge et al., 1996; Bouchon et al., 2011; He et al., 2023). The *cascade* or *cascade triggering* model (Fig. 2b), on the other hand, describes that both foreshocks and the mainshock initiation as a random process that are linked by elastic stress triggering (Ellsworth and Bulut, 2018). In other words, a mainshock is considered as an event triggered by the preceding earthquake which happens to be larger (Helmstetter et al., 2003; Felzer et al., 2004). Martínez-Garzón and Poli (2024) argues that both *pre-slip* and *cascade* models are likely end-member models that may oversimplify the real processes. Mignan (2014) found that studies with smaller events listed in the earthquake catalogs favored more on the *pre-slip* model, suggesting a potential bias due to catalog completeness. In the *rate-dependent cascade-up* model (Fig. 2c), the nucleation process includes *pre-slip* that can drive foreshocks, but the occurrence of foreshocks can trigger a rapid dynamic rupture of the mainshock (i.e., *cascade triggering*) (Noda et al., 2013; McLaskey and Lockner, 2014; McLaskey, 2019).

In the *fluid-driven* model (Fig. 2d), episodic fluid flow (most likely from larger depth) is the primary driver for the foreshock activity as well as the initiation of the mainshock (Jansen et al., 2019; Lei et al., 2024). In this case, foreshock seismicity fronts expand from the initial injection point (if the fluid source can be approximated as a point) with sqrt(t) (Shapiro et al., 1997), and likely cover a much wider area than the critical nucleation length scale Lc of the mainshock. In the *migratory slow-slip* model (Fig. 2e), aseismic slip, rather than fluid flow, is hypothesized to be the main driver of the foreshock and the mainshock initiation (Barcheck et al., 2021; Wang et al., 2024a; Wang et al., 2024c). A major difference between this (Fig. 2e) and the *pre-slip* models (Fig. 2a) is that the aseismic slip area here is much larger than the critical nucleation length scale Lc for dynamic slip of the subsequent mainshock. A recent study by Li et al. (2024) suggests that the wavelength of normal stress variation λ with respect to the local critical nucleation length scale Lc^* controls the selection of nucleation models, and *migratory slow-slip* model could emerge when λ lies between Lc^* for high normal stress patch and that for low normal stress patch. Moreover, Moutote et al. (2023) argue that the same aseismic process not only drive the foreshock and mainshock nucleation, but also aftershock migration.

We note that these five models are the most representative ones, but other competing models have also been introduced. For example, Kato and Ben-Zion (2021) introduced a *progressive localization* model that involves an evolution of deformation from distributed damage in a rock volume to more localized shear slip along future rupture zones. During such localization process, elevated seismicity would occur in multiple clusters, and one of them would evolve into a foreshock sequence and finally initiate the mainshock. Possible examples include the 2014 M8.1 Iquique earthquake sequence in Northern Chile (Socquet et al., 2017), the 2016 Central Italy seismic sequence (Sugan et al., 2023), and likely the 2024 M7.6 Noto earthquake and preceding swarms (Kato, 2024; Peng et al., 2024). Such localization of deformation may occur in either geometrically simple or complex fault structures (Kato and Ben-Zion, 2021).

Based on numerical modeling of the nucleation process on homogenous fault that is governed by the rate-and-state friction law, He et al. (2023) found that a stress-releasing zone (termed weakening-zone core) expands first and then shrinks in the later stage right before the mainshock. They argued that the evolution of foreshock activity including its time-space migration before the 2014 M8.1 Iquque earthquake was consistent with such an expansion and shrinking pattern (Yagi et al., 2014). However, to reasonably explain both *pre-slip* and foreshocks, the "homogeneous" fault must contain some distributed asperities or sub faults. Just like what has been observed in the laboratory, before a large-scale rupture occurs, the asperities within the nucleation zone of the fault need to break sequentially, forcing *pre-slip* to reach a critical scale (Lei, 2003). Finally, numerical modeling results suggest that *afterslip* (or more generally triggered aseismic slip) from previous earthquake can drive foreshocks and the next larger mainshock (Ito and Kaneko, 2023). Like the *cascade* model, their critical nucleation scale Lc can be much smaller than the region that hosts aseismic slip.

Recently, Stein and Bird (2024) proposed an alternative *cascade* model where large continental earthquakes (such as the 2023 M7.8 Pazarcik, Türkiye, earthquake) likely nucleated on a branch or splay fault before jumping onto the major strike-slip fault (Fig. 3a). The initial rupture on the branch fault could be considered as a foreshock, although it would be viewed as part of the mainshock initial rupture at teleseismic distances. For example, the 2024 M7.5 Noto mainshock was preceded by a M5.9 foreshock 14 seconds earlier (Peng et al., 2024). But this event was only listed in the JMA catalog, not by the USGS or other global catalogs. Ma et al. (2024) showed that there was likely more than one event in these 14 seconds, and hence they argued for a continuous initial slow rupture, rather than one single foreshock (Fig. 3b). Ozacar and Beck (2004) also showed that the initial ruptures of the 2001 M7.9 Kunlun fault and the 2002 M7.9 Denali fault earthquakes all started on fault structures with different faulting styles than the main strike-slip faults (Fig. 3a). Regardless of the name of the branched fault rupture and its faulting style, it likely reflects an alternative view to consider how large ruptures on continental faults are initiated.

Martínez-Garzón and Poli (2024) recently reviewed several of these models and proposed an *integrated* model with the initial phase being the localization model, and the last stage being the *rate-dependent cascade-up* model. Similarly, based on precursory acoustic emission during nucleation of laboratory stick-slip experiment, Marty et al. (2023) argued that the nucleation process is almost entirely aseismic at the beginning, and is followed by increased proportion of elastic stress triggering right before the onset of mainshock rupture. This is similar to the *rate-dependent cascade-up* model (McLaskey, 2019).

Here we attempt to further classify these foreshock physical models into three categories: elastic stress changes, aseismic slip, and fluid flows (Fig. 4). Some physical models presented before (Fig. 3) can be combined into one category. For example, the *pre-slip* model (e.g., Dieterich, 1992; Ohnaka, 1992; Bouchon et al., 2011), the recently proposed *migratory slow-slip* model (Barcheck et al., 2021; Wang et al., 2024a) and the *afterslip* model (Ito and Kaneko, 2023) can be all combined into the aseismic slip category. Similarly, the elastic stress change category would include both static and dynamic stress changes from nearby earthquake ruptures (i.e., the original and alternative cascade models), and possibly dynamic stress changes from nearby earthquake ruptures (Antonioli et al., 2006; Ding et al., 2023; Dong et al., 2024) and large distant earthquakes (Walter et al., 2015; Martínez-Garzón and Poli, 2024). It is also obvious that the physical process in each category is not mutually exclusive and can 'trigger' each other. For example, an earthquake can trigger aseismic slip surrounding the mainshock rupture plane, which is termed **afterslip**. Afterslip is known to drive aftershocks (Peng and Zhao, 2009; Ross et al., 2017; Perfettini et al., 2018), and if the triggered aftershocks happen to be larger than the previous earthquake, we now have a case where afterslip (or more generally aseismic slip) from previous earthquake can drive foreshocks and the next larger mainshock (Ito and Kaneko, 2023). Similarly, based on the physical mechanisms of dilatancy hardening and fault slip affecting its hydraulic diffusivity, there are also interactions between fluid action and slow pre-slip that can either promote or inhibit each other (Liu et al., 2020; Lei, 2024). Thus, while fluid flows can drive seismicity, the occurrence of moderate to large earthquakes are also known to break seals within the fault zones, resulting in rapid changes of fluid flows and subsequent seismicity through a fault-valve mechanism (Sibson, 2007; Kato, 2024). Finally, as mentioned before, aseismic slip and fluid flow can work in concert to drive foreshocks and earthquake swarms (Sirorattanakul et al., 2022; Peng et al., 2024).

We note that these models/categories have long been proposed to explain aftershocks or remotely triggered seismicity (Freed, 2005; Hill and Prejean, 2015; Hardebeck et al., 2024). Based on these statistical seismological characteristics, one could potentially argue that we cannot tell foreshocks and aftershocks apart, until the largest mainshock occurs (Helmstetter et al., 2003; Felzer et al., 2004; Zaccagnino et al., 2024). When we observe seismic activity with characteristics like typical foreshocks known so far, determining whether it is a precursor to a major earthquake requires a comprehensive consideration of many factors, including the scale of local faults and their stress criticality and potential fluids. Despite decades of research progress, this still presents a challenging seismological issue.

Differences in Foreshock Observations and Interpretations

In the past twenty years, several foreshock sequences have been well studied, although different conclusions were made on the physical mechanisms of mainshock nucleation. Examples include the 1999 M7.6 Izmit (Bouchon et al., 2011; Ellsworth and Bulut, 2018), the 2019 M7.1 Ridgecrest earthquake (Ross et al., 2019b; Huang et al., 2020; Yue et al., 2021), and the 2021 M6.1 Yangbi earthquake (e.g., Zhu et al., 2022; Liu et al., 2022; Zhou et al., 2022; Wang et al., 2024a).

Below we attempt to highlight a few areas where majority of the differences reside. Fig. 5a shows the Coulomb stress changes from an M4.6 foreshock that occurred more than 2 hours before the 2010 M6.7 Yushu earthquake in the Tibetan Plateau in Western China (Chuang et al., 2023). The M4.6 foreshock occurred on a conjugate fault plane and casted a negative Coulomb stress on the hypocenter of the M6.7 mainshock. However, Huang et al. (2023) showed that while the M4.6 foreshock occurred on the conjugate fault plane, both events were close to the same E-W trending Ganzi-Yushu fault. Such an alternative location would result in a small but positive Coulomb stress changes on the mainshock hypocenter, which is consistent with the *cascade* model. Similarly, Peng et al. (2024) found that an M5.5 foreshock occurred 4 minutes before the 2024 M7.5 Noto Peninsula mainshock also casted an apparent stress shadow (i.e., negative Coulomb stress change) at the mainshock hypocenter. Zhou et al. (2022) and Wang et al. (2024a) also found that by moving the hypocenter of the 2021 M6.1 Yangbi mainshock from 4-5 km to 7-8 km depth, the Coulomb stress changes from preceding foreshocks on the Yangbi mainshock hypocenter became much smaller as compared to the previous results (Zhu et al., 2022). Clearly, a change of relative location of a few kilometers between the foreshock and the mainshock can result in a significant change in the amplitude of the static Coulomb stress changes, sometimes even the sign. These comparisons highlight the need to compare different methods used to determine both the absolute and relative locations of the foreshock and the mainshock initial hypocenter. Besides event location, dynamic stress change could play a significant role, e.g., the mainshock hypocenter could be first loaded by a positive dynamic Coulomb stress change and then a positive or negative static Coulomb stress change from a nearby foreshock (Antonioli et al., 2006; Ding et al., 2023; Dong et al., 2024). Slow slip or fluid flow may also generate stress changes at the mainshock hypocenter, but it is not generally considered when computing stress changes.

Repeating earthquakes (or repeaters) refer to a group of earthquakes occurring at essentially the same region, or more strictly speaking, correspond to the failure of the same asperity (Fig. 5b) and generating nearly identical waveforms (Vidale et al., 1994; Nadeau and McEvilly, 1999; Uchida, 2019). They are generally considered to be frictionally locked asperities surrounded by rate-strengthening region that either creep constantly, or slip during aseismic transient (Beeler et al., 2001; Uchida and Burgmann, 2019; Nakajima and Hasegawa, 2023). Hence, their recurrence intervals and co-seismic slip can be used together to infer the amount of ambient aseismic slip or creep at depth (Nadeau and McEvilly, 1999; Materna et al., 2018; Deng et al., 2020; Nakajima and Hasegawa, 2023), pre-slip or aseismic slip before large earthquakes (Li et al., 2011; Kato et al., 2012; Chen and Li, 2018; He et al., 2023) or afterslip/postseismic deformation (Peng et al., 2005; Zhao and Peng, 2009; Chen et al., 2010; Yao et al., 2017). However, so far there is no universal criteria for defining repeating earthquakes. Depending on the threshold of the cross-correlation coefficient (CCC) and other parameters (e.g., time window length, frequency band), different groups may reach totally different conclusions on whether repeating earthquakes exist and their detailed clustering behaviors. Gao et al. (2021) recently pointed out that using CCC alone is not sufficient to distinguish between true tight repeaters (i.e., overlapping source regions) and loose repeaters (i.e., partially or close but non-overlapping source regions) (Fig. 4b). A combination of CCC and interevent overlapping region (based on relocations or differential S-P arrival times) is needed to identify those tight repeating clusters (Peng et al., 2005; Zhao and Peng, 2009; Gao et al., 2021; Sugan et al., 2022). Additionally, the mechanism of repeating earthquakes does not necessarily require aseismic slip (Ellsworth and Bulut, 2018). Laboratory studies have shown that continuous/cycled injection of fluid may also trigger repeated ruptures due to a decrease in the effective normal stress, as expected from the Mohr-Coulomb failure law (Zhu et al., 2021). In summary, inaccurate estimation of repeating earthquakes may lead to biased understanding of the physical process (aseismic slip or fluid flows) leading up to the mainshock.

Finally, different migration patterns of foreshocks have been used to infer different physical processes driving the sequence. Fig. 5c summarizes three types of seismicity migration curves and the corresponding driving forces. Steady slow-slip events (SSEs) tend to be followed by deep tectonic tremor at larger depth or microseismicity at shallower seismogenic depth (Kato, 2019; Wickham-Piotrowski et al., 2024), and the seismicity typically track the slip front of the SSEs. If the SSE front is rather slow and steady, then we would expect to observe a slow and steady seismicity front. On the other hand, if fluid flow is the primary driver, the seismicity typically would expand rapidly first from the initial injection point following the approximate formula of sqrt(t) (Shapiro et al., 1997), where t is time since the initial injection. Examples include the 1975 M7.3 Haicheng foreshock sequence in Northeast China (Lei et al., 2024), the 2008 M4.9 Mogul foreshock sequence in Nevada, USA (Jansen et al., 2019), and the initial foreshocks of the 2021 M6.1 Yangbi sequence (Lei et al., 2021). If a moderate-size foreshock triggers afterslip, which in turn drive its own aftershock activity, then we would expect to observe a very rapid expansion at the beginning following the formula of log10(t) (Peng and Zhao, 2009; Ross et al., 2017; Perfettini et al., 2018). Examples of such logarithmic-type foreshock-aftershock expansion include the 2011 M9.0 Tohoku-Oki earthquake and its M7.3 foreshock (Kato et al., 2012) and the foreshock sequence of the 2013 M8.1 Iquique earthquake (Kato and Nakagawa, 2014). These different migration/expansion patterns are relatively easy to distinguish if the time scale of observation relatively large. On the other hand, if the time scale is relatively short, such as the half-hour difference between the largest M5.1 foreshock and the M6.1 Yangbi mainshock (Wang et al., 2024a), or if there are multiple fluid source points, it is relatively difficult to distinguish between different patterns. We also note that except in a few cases (Schoenball and Ellsworth, 2017; Kato and Obara, 2014; Wu et al., 2017), there is no quantitative way to examine whether a migration pattern would fit either a sqrt(t) or log10(t) with different expansion speed or diffusivity. Here, the chain of evidence from different aspects becomes crucial.

Additional Complexities

In addition to the issues mentioned above, in this section we summarize additional complexities that may not be well appreciated in previous studies. First, slow slip is not in steady state, e.g., it could be perturbed by spatial variation of normal stress, or fluid pressure. Based on decades of geodetic and seismological observations, Jolivet and Frank (2020) found that large-scale SSEs consist of many intermittent episodes of short-term bursts. Recent laboratory experiments (Wang et al., 2024c) also showed that the mainshock nucleation process could be intermittently mixed with short acceleration and deceleration episodes. In particular, a later decreasing migration speed of slow slip and foreshocks may result from the shadowing effect of a nearby locked asperity, without requiring afterslip or rate-strengthening rheology. Finally, Im and Avouac (2023) argued that elastic stress interaction in a discrete fault network governed by the rate-and-state-friction can also produce accelerating foreshock sequence and migrating earthquake swarm, without requiring aseismic slip. In other words, the interpretation for observed kinematic features of seismicity migration is not unique. Additionally, migration front of seismicity does not exactly follow the migration front of slow slip or fluid. In many cases, the former could lag the latter (if the latter is considered as the driving force), so only the lower bound of the latter can be estimated from the migration of the former (Yamashita et al., 2022). Possible mechanisms for the time delay include: (1) nucleation of seismicity takes time (Dieterich, 1994), (2) seismicity occurs on asperities, which fail later than the surrounding creep due to their relatively higher normal stress (Ohnaka, 2013). These complexities further highlight the difficulty of identifying the physical mechanism that drive mainshock nucleation based on foreshock seismicity alone, which will be summarized in the next section.

Which Model(s)?

As summarized before, many progresses have been made in the past decades on observing foreshocks and the mainshock nucleation processes based on field observations (Kato and Ben-Zion, 2021; Martínez-Garzón and Poli, 2024), laboratory experiments (McLaskey, 2019; Bolton et al., 2023; Geobel et al. 2024) and numerical modeling (Cattania and Segall, 2021; He et al., 2023). However, it is still difficult to identify a single physical model that can explain most if not all the observations. The challenges can be summarized as follows:

1. Inconsistent datasets or methods used to identify certain features, which are then used to distinguish between different physical models (Fig. 5). Open sharing of the several recent

datasets, such as the 2019 Ridgecrest sequence, and the 2024 Noto sequence, as well as organized efforts such as the community stress drop validation study (Baltay et al., 2024), would allow different groups to compare results and methods with each other to identify potential issues.

2. Additional observations/measurements are needed to distinguish between these models. While seismologists have been trying to use different characteristics of foreshock patterns to distinguish between these models, most observations are indirect and are subject to uncertainties (due to measurement errors, inconsistent methods and inherent insensitivities of seismic waves to these physical processes). For example, while migrating foreshocks (or low-frequency earthquakes/tremor) and repeating earthquakes have been used to infer SSEs (Shelly et al., 2007, 2011; Kato et al., 2012; Uchida and Burgmann, 2019), the most direct way to observe aseismic slip before large earthquakes is through geodetic measurements (Roeloffs, 2006; Avouac, 2015), which are also subject to high noises and uncertainties (Bletery and Nocquet, 2023, 2024; Bradley and Hubbard, 2023, 2024). Similarly, while seismic tomographic methods can be used to infer the presence of high fluid pressures (Nakajima, 2022), the interpretation is not unique. A better way to map fluids in the subsurface is through magnetotelluric (MT) methods (Unsworth and Rondenay, 2013), especially if the spatio-temporal evolution of subsurface fluid flows can be mapped in detail (Shelly, 2024). A combination of different geophysical methods, along with geochemical observations (Umeda et al., 2024), hydromechanical modeling (Wang et al., 2024b), may provide us a much more complete picture of how fluid flow, aseismic slip and elastic stress triggering drive foreshocks and mainshock nucleation at seismogenic depth.
3. Some models are end-member types, while others are not mutually exclusive. The *pre-slip* and *cascade* models (Fig. 2) are clearly end-member models (Martínez-Garzón and Poli, 2024), while the "*cascade-up*" model is a hybrid of these two (McLaskey, 2019). In addition, the three major categories (Fig. 4) of physical processes that drive seismicity (including foreshocks) can co-exist and influence each other. As mentioned before, one of the best examples is the recent 2024 M7.5 Noto peninsula sequence, where the preceding swarm since 2021 was likely driven by a fluid injection at larger depth, which triggered aseismic processes that lasted for a few years (Nishimura et al., 2023; Kato, 2024). While it is still not clear how the long-lasting swarm changed its mode into a short-term foreshock sequence in the last four minutes before the Noto mainshock, tectonic loading, aseismic slip driven by fluid flow, and elastic stress triggering likely worked in concert (Peng et al., 2024) to trigger the eventual M7.5 mainshock with a very slow initial rupture (Ma et al., 2024). From this perspective, the primary mechanisms of different cases can vary over different time scales or spatial ranges (Martínez-Garzón and Poli, 2024). Finally, multiple physical processes can operate simultaneously, as revealed by recent observations of foreshocks (McLaskey, 2019; Yao et al., 2020; Cabrera et al., 2022; Liu et al., 2022; Zhou et al., 2022; Wang et al., 2024a; Martínez-Garzón and Poli, 2024) and aftershocks or triggered seismicity (Meng and Peng, 2014; Ross et al., 2017; Hardebeck et al., 2024). It might be 'useless' to argue which model(s) are correct, because as we all know, "*All models are wrong, but some are useful*" (Box, 1980; Field, 2014). Perhaps we can modify the sentence slightly as "*All models are incomplete, and some work together*".

What's Next?

While the jury is still out on which physical model(s) play(s) the most important role in driving foreshocks and earthquake nucleation, we have made significant processes over the past few decades on this topic. Looking forward to the upcoming decade, we expect potential breakthrough in the following directions:

1. **Ultra-dense and near field/deep borehole geophysical observations**. Recent development in ultra-dense sensors such as high-frequency nodal instruments, distributed acoustic sensing (DAS) and other fiber-sensing technologies (Zhan, 2020; Lindsey and Martin, 2021) allow us to observe earthquake sequences in a much higher density than before and simultaneously obtain more detailed source and surrounding structural information through geophysical inversion, which can be used to better image subtle and sub-events before, during and after the mainshock, as well as their relationship with the fine-scale fault structure involved in earthquake nucleation (Shearer et al., 2023; Zhai et al., 2024; Ma et al., 2024). Deep borehole geophysical observations at the North Anatolian Fault in the Eastern Sea of Marmara (GONAF) in Western Türkiye (Kılıç et al., 2020), the BedrettoLab, Swiss Alps (Volpe et al., 2023) and elsewhere around the world may provide new insights on the nucleation process of microearthquakes at ultra-fine scale.
2. **Extracting hidden patterns from high-dimensional, high-resolution earthquake catalogs**. In recent years, advanced methods such as template matching (Peng and Zhao, 2009; Zhang and Wen, 2015; Beaucé et al., 2018; Chamberlain et al., 2018) and machine learning (Kong et al., 2019; Mousavi and Beroza, 2023) have been applied to years of continuous waveforms, resulting in **high-resolution earthquake catalogs** that are several times to tens of times more than listed in standard earthquake catalog (Ross et al., 2019a; Zhai et al., 2021; Tan et al. 2021; Neves et al., 2023). These high-resolution catalogs may contain additional information that are not revealed by typical statistical seismology methods such as ETAS. Deep neural point process models (Zhu et al., 2022; Stockman et al., 2023; Dascher-Cousineau et al., 2023; Zlydenko et al. 2023) have shown promising results that warrant continuing development in this direction. In addition, these high-resolution catalogs (together with focal mechanisms and other derived high-dimensional parameters) can be used to examine how microearthquakes respond to external stress perturbations such as tidal stress and dynamic stress from large distant earthquakes (Li et al., 2023; Wang et al., 2022), which have been shown to evolve before large earthquakes (Beaucé et al., 2023; Liu et al., 2024) and volcanic eruptions (Bell et al., 2021).
3. **Revisiting historical earthquakes**. Major earthquakes are rare and sudden events. On average, around 300 earthquakes of magnitude 6 or greater, 30 of magnitude 7 or greater, and fewer than 2 of magnitude 8 or greater occur worldwide each year. Revisiting historical earthquakes is of special significance. For many past major earthquakes, the understanding of fault structures may have been quite rough due to the limitations of observational technology at that time. For earthquakes with ongoing aftershock activity (e.g., the 1975 Haicheng earthquake), refined data such as precise hypocenters of recent evets recorded by improved networks can enhance our understanding of the foreshock mechanisms (Lei et al., 2024). It is also necessary to obtain more detailed physical fields (e.g., high-resolution catalogs) in regions where major earthquakes have occurred in the past with dense array observations and advanced methods as described above, and to

compare and analyze the dominant mechanisms of whether and what types of foreshocks occurred (e.g., Manganiello et al. 2022; Wetzler et al. 2023).

4. **Integration of statistical models and physical models**. Earthquakes are fundamentally processes where fault rupture release the strain energy accumulated in the surrounding rock volume. The main factors controlling foreshock activity include the fault structure itself, the host rock volume, regional stress field, and potential deep fluids and other transient events that could influence all three of the above factors. These controlling factors evolve independently but with interaction among each other over time. Spatially, they exhibit strong heterogeneity, hierarchical features, and self-similarity, while our understanding of these factors remains uncertain. This results in both physical and statistical models having their respective strengths and limitations. Statistical models, based on empirical relationships of seismic activity, often account for complexity and interactions to some extent. Physical models are generally simplifications of the real-world scenarios and hence may not capture the full complexity. As a result, statistical models are generally more effective than current physical models in reproducing past events and predicting future ones (Mancini et al., 2019; Hardebeck, 2021; Hardebeck et al., 2024). Hence, integrating both approaches is undoubtedly the future direction. In constructing physical models, considerations should extend beyond hydraulics and rock mechanics to include geochemical factors, such as the dissolution and crystallization of fault materials under hydrothermal conditions. Finally, recent numerical modelling results (Xu et al., 2015) and field observations (Xu et al., 2024) suggest a possible phase diagram between self-arresting and runaway ruptures. This likely indicates a fundamental difference between earthquakes that follow the Gutenberg-Richter frequency-magnitude distribution, and the largest earthquake, i.e., the **dragon king event** that clearly deviate from the expected Gutenberg-Richter distribution (Ben-Zion, 2008; Sornette and Ouillon, 2012; Sornette et al., 2024). Identifying the physical processes that lead to these dragon king events with statistical tools (Beaucé et al., 2023; Liu et al., 2024) can offer new insight into earthquake nucleation, and answer fundamental questions on whether small and large earthquakes start the same or not (Ide, 2019).
5. **Imaging fault zone heterogeneity and fluids**. The heterogeneity of seismogenic faults, including their geometries and frictional properties of fault planes, appears to be one of the necessary conditions for foreshocks and other precursory phenomena (Yamashita et al., 2021; Bolton et al., 2023; Cheng & Wang, 2016; Goebel et al., 2024). Fluid activity, which is episodic in time and localized in space (Lei et al., 2014; Ross et al., 2020), can enhance these heterogeneities, thereby generating observable precursor phenomena (Nishimura et al., 2023). Additionally, this effect causes foreshock activity to be more sensitive to tidal forces, resulting in modulation of their occurrence times (Lei et al., 2023; Yu et al., 2022). Therefore, in-depth studies of the heterogeneity of seismogenic structures and fluid effects, along with the related statistical characteristics of seismic activity, may uncover crucial evidence chains for foreshock identification. This should be one of the key directions for future research, can be achieved by high-resolution field studies of exhumed faults (Ostermeijer et al. 2020) or recent surface ruptures (Barnhart et al., 2020), a combination of high-resolution, high-dimensional earthquake catalogs (Shelly et al., 2023), as well as high-resolution fault zone imaging with advanced recording/methods (Atterhol et al., 2024; Li and Ben-Zion, 2024), possibly in 4D (Wang et al., 2024b).

6. **Integration of research across different scales.** At this point, we have primarily focused on the nucleation and foreshocks of natural earthquakes. Over the past two decades, there has been a continuous increase in seismic activity induced by industrial fluid injection, with numerous moderate to strong earthquakes observed (e.g. Barbour et al., 2017; Ellsworth et al., 2019; Meng et al., 2021; Lei et al., 2022; Moein et al., 2023; Zhao et al., 2023), comparable in magnitude to typical foreshocks. Compared to natural earthquakes, injection-induced earthquakes offer more detailed known conditions, such as fluid pressure, injection volume, and detailed geological structures. This provides an opportunity for in-depth research on the role of fluids in fault activation. Indeed, this research requires close collaboration between academia and industry. The results not only support the safety and efficiency of related industrial operations but also significantly advance the study of precursory processes of fluid-driven or fluid-involved natural earthquakes. The field scale of injection-induced earthquakes lies between laboratory experiments and natural earthquakes, making the integration of research across different scales an important direction.

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Figure Captions

Figure 1. Schematic diagrams showing temporal behaviors of different types of foreshock sequences: (a) No obvious foreshocks right before the mainshock; (b) Accelerating-type foreshocks where the seismicity rate increases exponentially before the mainshock; (c) Burst-type foreshocks where each moderate-size foreshocks are followed by their own aftershocks; (d) Swarm-type foreshocks where the foreshock sequence can be classified as a swarm; (e) Swarm-burst-type foreshocks where the sequence include a swarm and a subsequent moderate-size event (and its aftershocks) right before the mainshock; (f) Swarm-accelerating-type foreshocks where the foreshock sequence starts as an swarm, a subsequent quiescence, and finally an accelerating sequence right before the mainshock.

Figure 2. Schematic diagrams showing five representative foreshock models that have been proposed so far. (a) Pre-slip; (b) Cascade; (c) Rate-dependent cascade-up; (d) Fluid-driven; and (d) Migratory slow-slip model. In each panel, Lc represents the critical nucleation length needed for the mainshock rupture to go unstable. Modified from McLaskey (2019) and Lei et al. (2024).

Figure 3. (a) A schematic diagram showing a fault model where a large earthquake rupture on the main fault which is strong and is preceded by a foreshock. The initial mainshock rupture occurs on a subsidiary fault that has either a different fault orientation or different faulting style (Stein and Bird, 2024). The mainshock ruptures unilaterally to the right and is followed by numerous aftershocks. (b) A normalized moment rate function showing the initial and primary mainshock ruptures, which is preceded by a foreshock and followed by many aftershocks.

Figure 4. An updated summary of three primary physical models that drive seismicity (foreshocks, aftershocks, swarms) and potential interactions between them.

Figure 5. (a) Static Coulomb stress changes induced by the M4.6 foreshock on the 2010 M6.7 Yushu mainshock epicenter and surrounding region in the Tibetan Plateau, Western China (modified from Chuang et al., 2023). The dark green star marks an alternative location of the Yushu mainshock (Huang et al., 2023) where the Coulomb stress change is small but positive. (b) A schematic diagram showing a tight repeating cluster where their source ruptures large overlap, and a loose repeating cluster where their ruptures partially or do not overlap. (c) Expected migration patterns of seismicity driven by a steady-state slow-slip event (SSE), fluid flow, and afterslip. Modified from Zhu et al. (2022).

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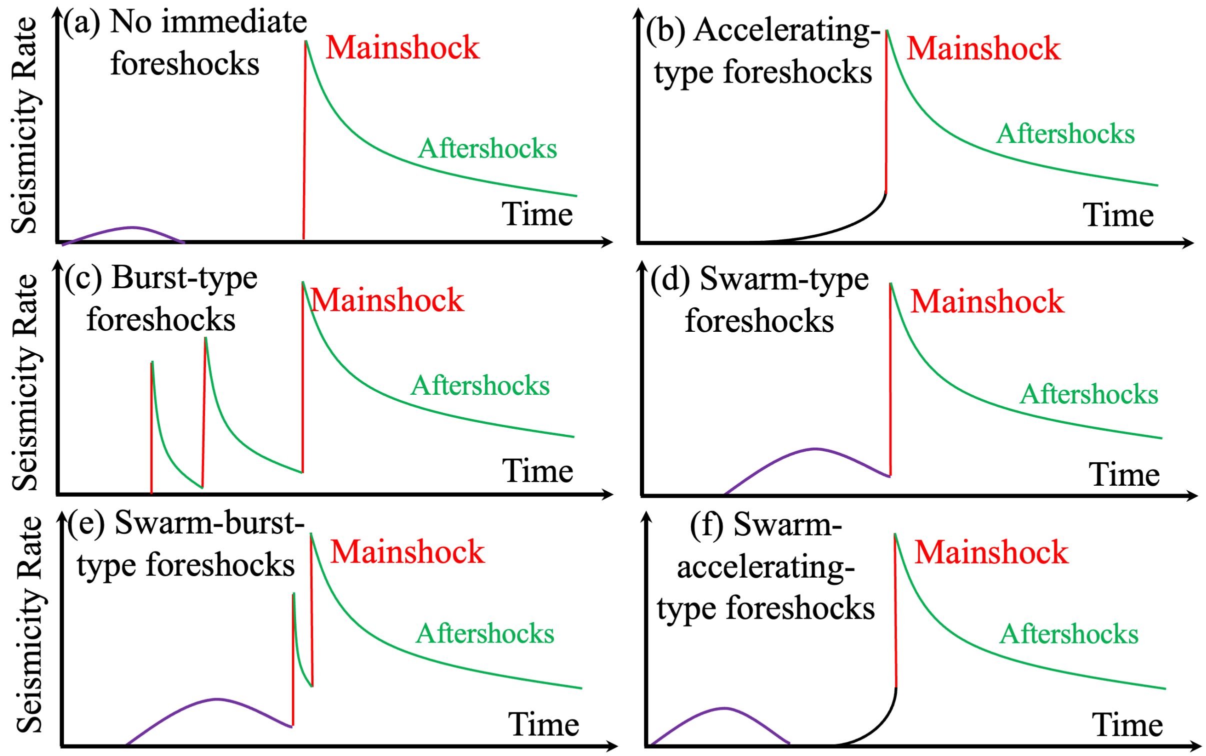

Figure 1. Schematic diagrams showing temporal behaviors of different types of foreshock sequences: (a) No obvious foreshocks right before the mainshock; (b) Accelerating-type foreshocks where the seismicity rate increases exponentially before the mainshock; (c) Burst-type foreshocks where each moderate-size foreshocks are followed by their own aftershocks; (d) Swarm-type foreshocks where the foreshock sequence can be classified as a swarm; (e) Swarm-burst-type foreshocks where the sequence include a swarm and a subsequent moderate-size event (and its aftershocks) right before the mainshock; (f) Swarm-accelerating-type foreshocks where the foreshock sequence starts as an swarm, a subsequent quiescence, and finally an accelerating sequence right before the mainshock.

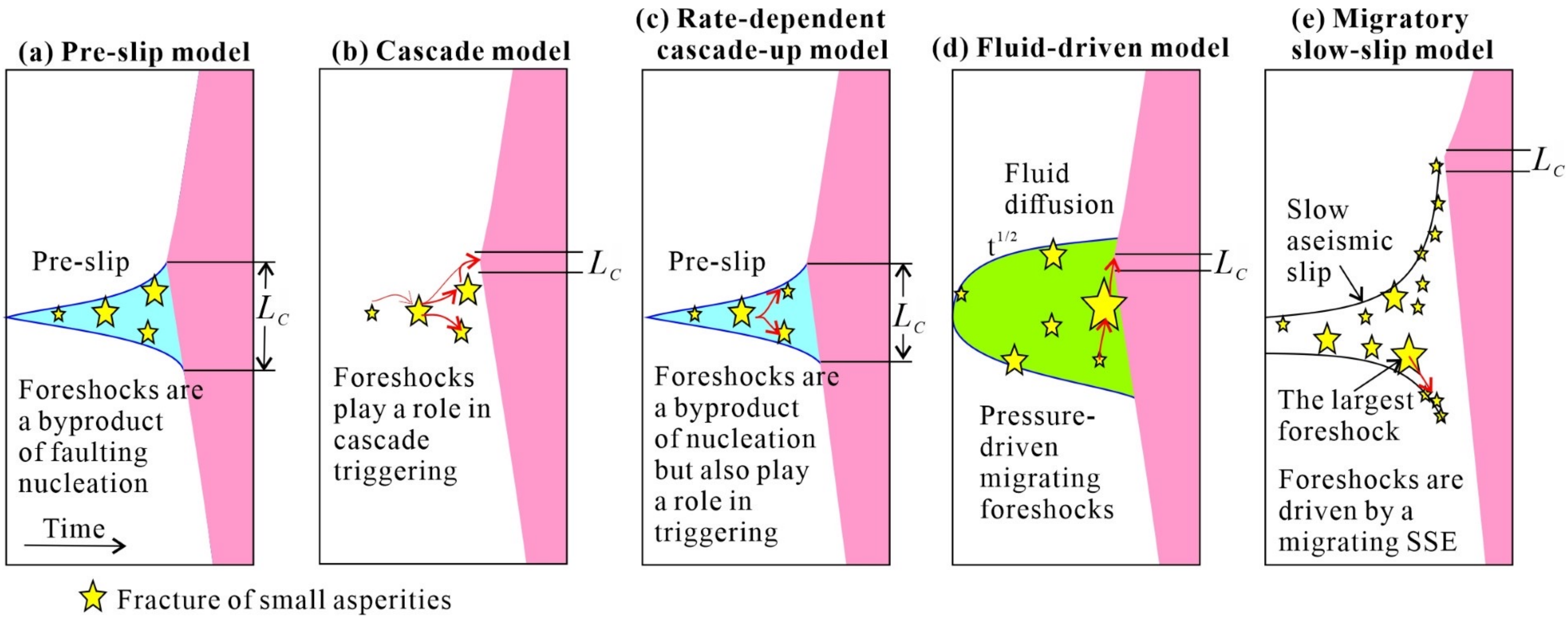

Figure 2. Schematic diagrams showing five representative foreshock models that have been proposed so far. (a) Pre-slip; (b) Cascade; (c) Rate-dependent cascade-up; (d) Fluid-driven; and (d) Migratory slow-slip model. In each panel, Lc represents the critical nucleation length needed for the mainshock rupture to go unstable. Modified from McLaskey (2019) and Lei et al. (2024).

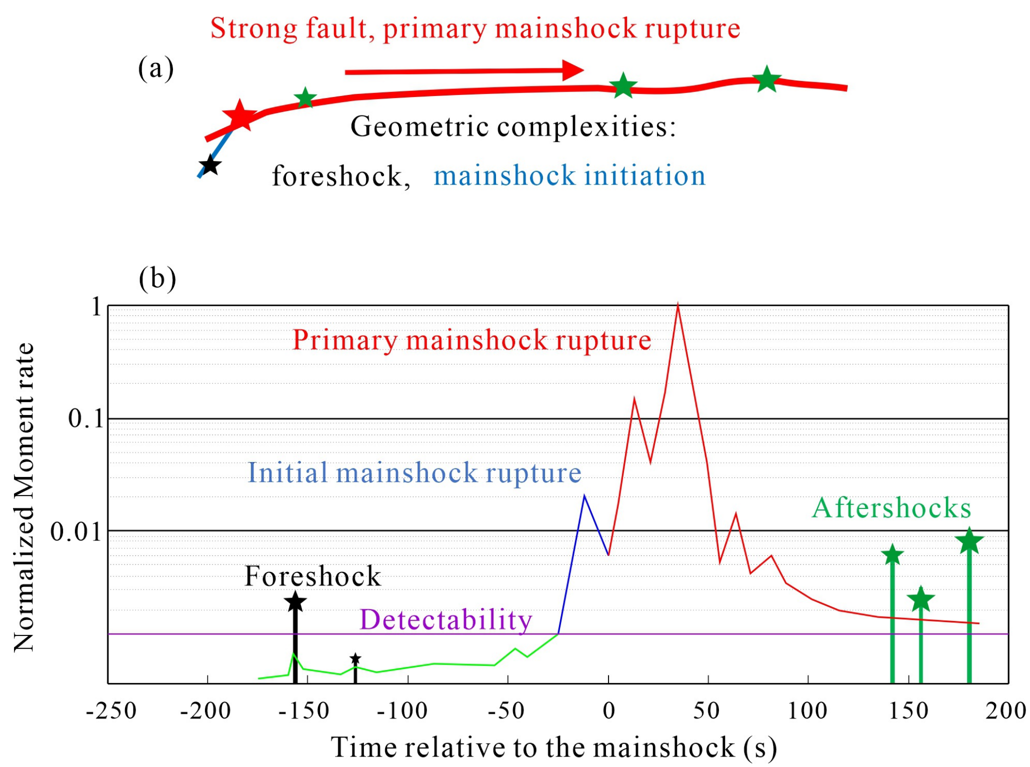

Figure 3. (a) A schematic diagram showing a fault model where a large earthquake rupture on the main fault which is strong and and is preceded by a foreshock. The initial mainshock rupture occurs on a subsidiary fault that has either a different fault orientation or different faulting style (Stein and Bird, 2024). The mainshock ruptures unilaterally to the right and is followed by numerous aftershocks. (b) A normalized moment rate function showing the initial and primary mainshock ruptures, which is preceded by a foreshock and followed by many aftershocks.

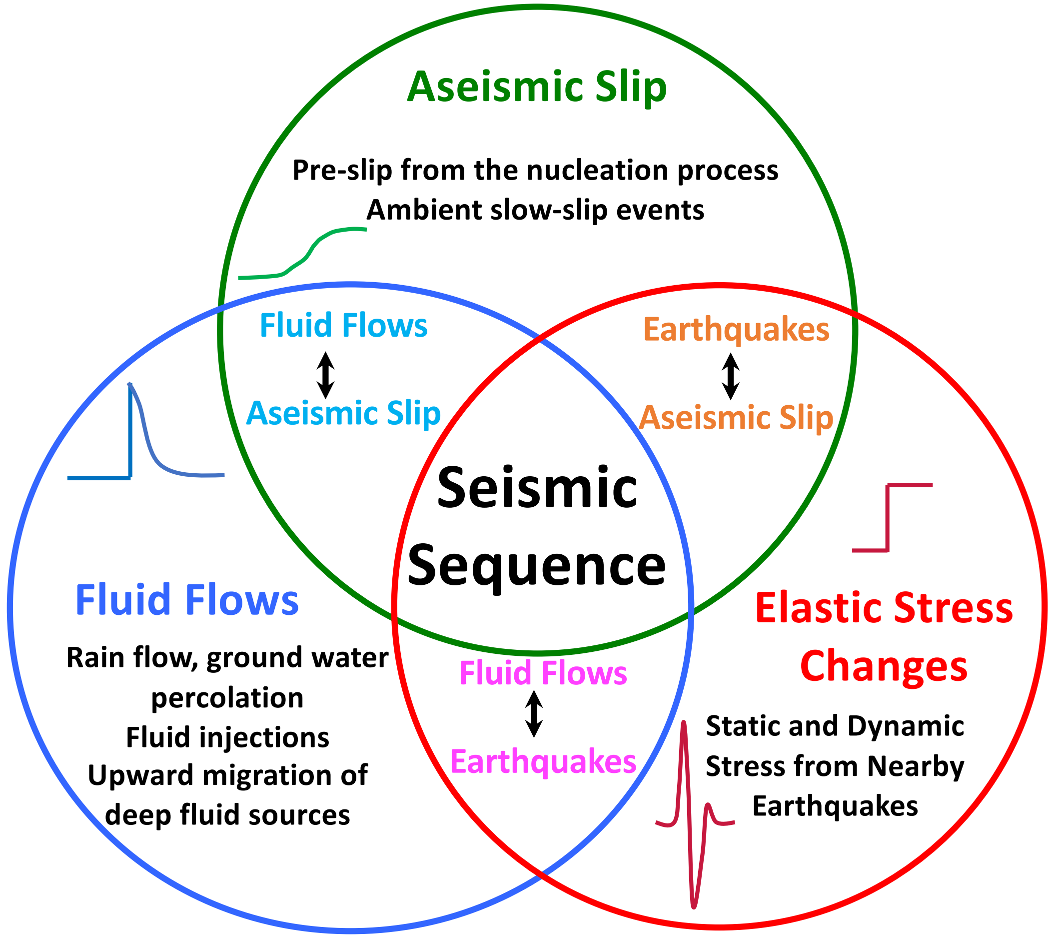

Figure 4. An updated summary of different physical models that drive seismicity (foreshocks, aftershocks, swarms) and potential interactions between them.

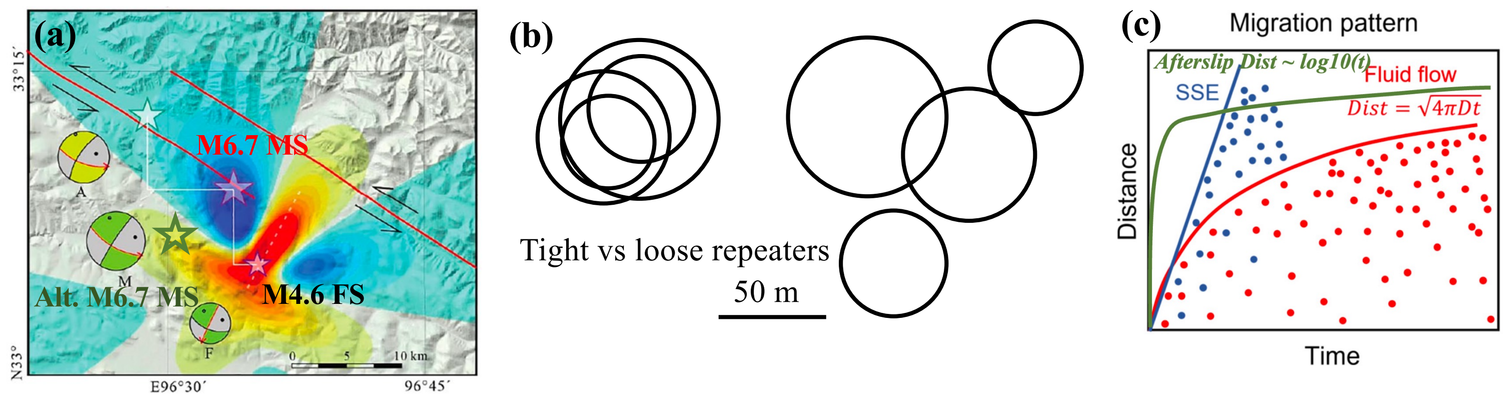

Figure 5. (a) Static Coulomb stress changes induced by the M4.6 foreshock on the 2010 M6.7 Yushu mainshock epicenter and surrounding region in the Tibetan Plateau, Western China (modified from Chuang et al., 2023). The dark green star marks an alternative location of the Yushu mainshock (Huang et al., 2023) where the Coulomb stress change is small but positive. (b) A schematic diagram showing a tight repeating cluster where their source ruptures large overlap, and a loose repeating cluster where their ruptures partially or do not overlap. (c) Expected migration patterns of seismicity driven by a steady-state slow-slip event (SSE), fluid flow, and afterslip. Modified from Zhu et al. (2022).