Review

Modeling fast and slow earthquakes at various scales

By Satoshi IDE*1,†

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Abstract: Earthquake sources represent dynamic rupture within rocky materials at depth and often can be modeled as propagating shear slip controlled by friction laws. These laws provide boundary conditions on fault planes embedded in elastic media. Recent developments in observation networks, laboratory experiments, and methods of data analysis have expanded our knowledge of the physics of earthquakes. Newly discovered slow earthquakes are qualitatively different phenomena from ordinary fast earthquakes and provide independent information on slow deformation at depth. Many numerical simulations have been carried out to model both fast and slow earthquakes, but problems remain, especially with scaling laws. Some mechanisms are required to explain the power-law nature of earthquake rupture and the lack of characteristic length. Conceptual models that include a hierarchical structure over a wide range of scales would be helpful for characterizing diverse behavior in different seismic regions and for improving probabilistic forecasts of earthquakes.

Keywords: earthquake, dynamic rupture, friction law, slow earthquake, hierarchical structure

Introduction

An earthquake is a unified physical process originating from the release of energy accumulated by long-term plate motion, followed by the propagation of seismic waves in underground elastic materials, and surface shaking that may cause significant structural damage. Seismic energy is released during seismic rupture, which is a mixture of shear fracture and frictional slip along near-planar surfaces (fault planes) in rocks at depth. This definition of an earthquake was established in the early 1960s, coinciding with the emergence of the theory of plate tectonics. The mathematical formulation of seismic wave radiation, based on simple models of seismic rupture, has been systematically described by Aki and Richards.1) The propagation of seismic waves radiating away from a source is described primarily by the linear elastodynamic equation and can be accurately predicted given knowledge of the physical properties of the subsurface materials. Thus, the theory of earthquakes was thought to have been established in the 1980s. However, the complexity and diversity of seismic sources, which represent the real problem in earthquake physics, remain poorly understood.

Dynamic rupture at a seismic source is controlled mainly by the elastodynamic equation, fracture criteria, and friction laws. Although fracture theory based on energetics had been mathematically established by the 1960s,2,3) the mechanisms of frictional sliding are still being studied. Since the 1970s, laboratory experiments have revealed important aspects of the friction laws for rocky materials, such as the temporal increase in static friction4) and the dependence of dynamic friction on slip rate.5) These characteristics were integrated into an empirical friction law, the rate- and state-dependent friction (RSF) law,6,7) which is mathematically simple and quite useful for numerical simulations, as discussed in the next section.

In addition to understanding the basic physical laws, characterizing earthquakes is complicated further by the large difference in scale between
laboratory experiments and natural fault systems. For example, the RSF law requires knowledge of several parameters whose scaling properties have not been established. Fault planes can be macroscopically approximated as flat, but microscopically they are fractal surfaces over broad scale ranges, including branches, steps, and jogs, which are difficult to constrain at depth. Dynamic rupture produces many small secondary faults, or microfractures, in surrounding rocks, which are macroscopically described as inelastic deformation. Seismological observation can resolve only macroscopic images of seismic rupture. Therefore, the microscopic realizations of friction and fracture on complex faults in inelastic media must consistently be related back to the macroscopic image of fault slip on a near-planar fault in elastic media. The scaling properties of earthquakes help to quantify these relationships and have been studied since the 1970s using seismic data analysis.

The analysis of seismic data has been enhanced by the rapid development of regional and global seismometer networks and the digitalization of data analysis. The amount of data has been increasing exponentially since the 1970s, similar to Moore’s law for computers. This increase can be demonstrated using the number of earthquakes detected and located by the Japan Meteorological Agency (JMA) (Fig. 1). This increase does not represent an increase in the number of earthquakes in Japan, but reflects their increased detectability owing to the development of seismic observation networks and to improvements in digital data handling. Similar improvements have occurred at the global scale as well. Since the 1990s, seismic observation and data analysis, together with the knowledge of fracture and friction laws of dynamic rupture, have led to an improved physical understanding of earthquake rupture.

The number of earthquakes in the JMA catalog shows a limited increase after 2000, suggesting that some technical limit has been reached in Japan, unlike Moore’s law still surviving owing to various innovations. However, at the same time, there have been qualitative improvements in the JMA catalog. Around 2000, the JMA discovered low-frequency earthquakes (LFEs) and included them in their catalog (Fig. 1). These are a kind of slow earthquake, as discussed further below in this review. Figure 2 compares seismic waves radiated from an LFE with those from two ordinary earthquakes of magnitudes M4.3 and M0.4. The amplitude of seismic waves for the LFE is smaller than for the M0.4 earthquake,
but for frequency content, the LFE bears greater similarity to the M4.3 earthquake. The small signals from LFEs had been recorded at many seismic stations prior to 2000, but several simultaneous observations were required to recognize them as natural events. The discovery of LFEs and slow earthquakes provides a breakthrough into the study of earthquake sources. We now recognize that slow deformation, likely controlled by viscous or diffusive deformation, is as important as fast slip, controlled by fracture and friction, for developing a comprehensive understanding of the physics of earthquakes.

This paper first reviews the current physical understanding of dynamic earthquake rupture, followed by a discussion on the scaling of rupture. Then, recently discovered slow earthquakes are introduced and their implications for the physics of earthquakes are discussed. Finally, a conceptual model is proposed that explains various earthquake phenomena over a broad range of scales.

**Physical models of earthquake rupture**

Because earthquake rupture includes frictional slip, it is often modeled using a slider (block) connected to a loading point using a spring (Fig. 3a). The slider has static and dynamic frictions, and shows stick-slip motion. When the force of the spring reaches the static friction, the slider moves and then stops after releasing elastic energy stored in the spring. This is an analogy for shear slip on a fault due to long-term plate loading, as illustrated for a subduction zone in Fig. 3b. A set of many similar sliders is also used to study statistical features of seismic activity over the long term.9,10 These models are useful for simulating some aspects of earthquake phenomena, but may also lead to misunderstandings. It is important to appreciate the limitations of these models.

A single slider like that shown in Fig. 3a slips repeatedly in the same manner. In some regions, earthquakes of similar sizes have occurred at almost the same location several times. A well-known example is the repeating earthquake sequence on the plate interface beneath Kamaishi, eastern Japan,11 where ten earthquakes of about M4.9 ± 0.1 occurred from 1957 to 2008. A sequence of earthquakes of about M6 at Parkfield, California, had been considered as repeating slip at almost the same area on the San Andreas Fault.12 Even larger earthquakes of about M8 occurred at almost the same location off the southeast coast of Hokkaido, Japan, as the 1952 and 2003 Tokachi-Oki earthquakes.13 These
earthquakes may be represented as slip of a slider-like underground structure. Although there are many examples of such repeating events, especially for small earthquakes, these are yet rare examples among various occurrence patterns of earthquakes, and most earthquakes cannot be modeled using a simple spring and slider model.

An essential difference between a spring-slider model and earthquake rupture is that the latter represents internal motion in an elastic material. For a spring-slider model, the magnitude of the pulling force from the spring can be different from that of the frictional force on the slider. On an internal surface in the elastic medium, stress is always equal to the frictional stress, which might be governed by some friction laws. Given appropriate initial conditions and outer boundary conditions, the displacement at any point in an elastic material is determined by the stress on the internal surface, namely, frictional stress. However, if displacement on the internal surface is assumed, frictional stress is automatically determined from the elastic properties without assuming friction laws. These conditions are dictated by the uniqueness theorem.1)

The spatial and temporal distribution of slip on the fault plane during an earthquake has been estimated using seismic waveforms since the 1980s14–16 by slip inversion.17) Figure 4 shows an example of the result of slip inversion for the 1995 Kobe earthquake.18) Based on the uniqueness
theorem, we can determine stress on the fault using the slip distribution (Fig. 4d). Comparing slip and stress at any points on the fault plane, we can recover the relationship between slip and frictional stress (Fig. 4c). Such an analysis is impossible if we characterize an earthquake using a spring and slider model. The relationships are not simple, but some of them resemble a slip-weakening friction law, in which frictional stress decreases as slip increases. A similar relationship between slip and stress has been obtained in laboratory experiments. An important parameter in the slip-weakening law is the slip-weakening distance, $D_c$, which was estimated as 0.5–1 m for the Kobe earthquake. In real data analysis, the available frequency range of data is limited and we cannot constrain $D_c$ accurately. The fracture energy, $G_c$, which is the work done on the fault plane before reaching the dynamic friction, is better constrained. This fracture energy is different from the surface fracture energy required to separate a material. Rather, it includes all inelastic energy consumption around the fault plane, such as secondary faults and inelastic deformation, and should be referred to as breakdown work, or more generally, as seismological fracture energy.

Another problem in the spring and slider model is related to the strength of a fault. With a simple slip-weakening law for the slider (Fig. 4b), the initiation of the slider movement is determined by the maximum stress. Therefore, the maximum stress level is often regarded as the strength of the system. This is not true for slip on the internal surface of an elastic medium. When a point on the plane starts slipping after reaching the maximum stress level, the slip changes the stress on the surrounding region. It is not clear whether the surrounding region also slips, and the initial slip may be stopped if the surrounding region does not slip. The initial rupture point may not be accelerated sufficiently to radiate seismic waves. Any points in an elastic medium cannot behave independently from the surrounding region. The initiation of slip that is fast enough to radiate seismic energy is determined not by a local friction law but also by the energy balance in a wider region. The rupture propagates if energy released by slip is sufficiently larger than energy consumed in the surrounding material. Slip across a wider fault area releases elastic energy more efficiently, whereas energy consumption may have some upper limit determined by material properties. Therefore, there is a characteristic size of a slipped area required to start dynamic rupture. This is the critical size of the rupture nucleus, which is controlled by the seismological fracture energy. Once the slip area exceeds this limit, the energy balance cannot be sustained quasi-statistically and dynamic rupture with a high slip rate starts propagating. With homogeneous stress and friction conditions on a planar fault, the rupture speed soon reaches a terminal velocity (P or S waves), and the rupture does not stop, because elastic energy release always exceeds the fracture energy. The arrest of rupture requires some special mechanisms such as heterogeneous stress levels, inelastic deformation, and irregular fault geometry.

Although a slip-weakening friction law on the fault plane in elastic media can describe the dynamic rupture process, it is not appropriate for modeling slow deformation before the initiation and after the termination of fast rupture. Without fast fault slip, materials around the fault plane heal, that is, become strong, due mainly to adhesion, as demonstrated by the increase in static stress with time of stationary contact in rock friction. The RSF law, mentioned in the introduction, is an empirical friction law based on friction experiments and is applicable for slow deformation. Friction coefficients change depending on the state of the frictional surface, which is related to the real contact area between the asperities on the two surfaces. By defining the strength as the stress at a reference slip rate, the frictional slip rate is expressed by an exponential function of applied stress. The temporal change in state (or strength) can be described by an evolution law, which describes physical and chemical processes related to the abrasion and adhesion of materials around a fault. Although several candidates for the evolution law have been suggested by several researchers, there is no single law that can explain all aspects of rock experiments consistently. RSF laws have been applied to dynamic problems also, though their applicability has yet to be confirmed experimentally. The dynamic behavior of frictional slip governed by the RSF law is similar to that of the slip-weakening law, with an apparent slip-weakening distance controlled by an intrinsic length scale of the RSF law ($L$).

Using the RSF law as a boundary condition on a fault plane embedded in an elastic medium, we can simulate the behavior of frictional slip, not only during a high-speed rupture but also for the whole period including long quiescence periods. These so-called earthquake cycle simulations were first applied by Tse and Rice to a strike-slip system mimicking the San Andreas Fault. Other similar simulations have been studied by many groups using various fault
geometries and configurations. In such RSF simulations, we can observe slow slip accelerating up to the initiation of dynamic rupture (preslip) and decelerating after the termination of dynamic rupture (afterslip). The preslip area corresponds to the aforementioned rupture nucleus. As weak evidence exists for preslip and strong evidence exists for afterslip, RSF simulations enable comparisons to be made between observations and model calculations. However, it is not clear whether the RSF law derived from laboratory experiments can be directly applied to natural fault systems. Major problems include determining the behavior at high slip rates, and the scaling problem discussed in the next section.

The physical models of earthquakes explained here are over-simplified. The intrinsic length scale of friction laws ($Dc$ for the slip-weakening law and $L$ for the RSF law) controls the scale of phenomena related to these laws. Therefore, typical RSF simulations produce repeating rupture events of the same size, which resemble the repeating earthquake sequences observed in Kamaishi and elsewhere. However, these repeating events are fairly rare in nature, and in most seismic regions earthquake statistics lack a characteristic size, as discussed in the next section.

### Scaling earthquake rupture

Various sizes of earthquakes are observed, from very large M9-class earthquakes to small events with negative magnitude. Even very small fractures in rocks produced by laboratory experiments are mechanically similar to earthquake rupture. Earthquakes are a well-known example of phenomena that obey power laws. The most well known power law for earthquakes is the Gutenberg-Richter frequency-magnitude relation (the GR law). Although it was originally introduced as the relationship between the local magnitude and the number of earthquakes, a contemporary definition should describe it as a power law relation between the number of earthquakes and the seismic moment, or the seismically radiated energy (Fig. 5). The negative slope between the log cumulative number of events and the magnitude Mw is the $b$-value, which is close to 1 and considered to be related to the ambient stress level. Small $b$-values are suggestive of high stress and vice versa. As the largest size of earthquake is restricted by the earth structure, the scaling relation for very large earthquakes is different. Figure 5 shows a slight deviation from a power law for magnitudes above M8, suggesting that a distribution with an upper limit, such as a Gamma distribution, better describes the size-frequency statistics. However, the lower limit of the GR law, or even the smallest earthquake size, has not been clearly identified.

There are several scaling relations between macroscopic parameters of earthquake rupture. Linear length parameters such as fault length, fault width, and average slip are almost proportional to each other. Therefore, the changes in stress and strain during an earthquake can be considered to be scale independent. These length scales are also almost proportional to rupture duration, because the rupture propagation velocity is also scale independent. Although the scale independence of the velocity of rupture propagation is due to the limitation of elastic wave (S-wave) speed, it is not clear what controls the scale independence of stress and strain changes. Since the discovery that large (M > 6) events occur with a near-constant stress drop between 1 and 10 MPa, stress drop estimates for many earthquakes worldwide have fallen within this range. Another candidate for a scale invariant parameter is
the ratio between seismic energy and seismic moment. These are two key quantities that characterize earthquake size: the former is a dynamic measure related to seismic waves and the latter is a static measure related to fault size. These two values span over 20 orders of magnitude for earthquakes, but their ratio, which is a non-dimensional quantity often referred to as scaled energy,\textsuperscript{40) varies over just two orders of magnitude around $10^{-5}$, and is regarded as a constant.\textsuperscript{41)} However, the scale dependence of scaled energy within this range remains a matter of debate. Some studies have found a slight increase in scaled energy as earthquake size increases,\textsuperscript{42),43) whereas other studies found an irrelevance of scale.\textsuperscript{44),45) During earthquake rupture, the fault slip develops spatially in a complex manner, which can be resolved using observed seismic waves (Fig. 4d). Occasionally, the slip region extends in only one direction, jumps to a distant area, or returns to a previously slipped region. Such complex behavior is clearly observed for very large earthquakes down to tiny earthquakes with magnitudes of approximately M1.\textsuperscript{46) It is often asked whether rupture at the initial stage of a large earthquake is similar to the whole rupture for a small earthquake. In fact, the very beginning of the rupture in a large earthquake is also complex, and the growth process from a tiny to a large earthquake appears self-similar unless affected by the finiteness of the seismogenic layer, as demonstrated for the 2004 Chuetsu\textsuperscript{47) and 2004 Parkfield earthquakes.\textsuperscript{48) Therefore, it is often difficult to discern the difference between the early seismic waves from small and large earthquakes. Figure 6 shows a set of seismograms from six closely located earthquakes that occurred in the Tohoku-Oki region, one of which is the 2011 Tohoku-Oki earthquake. In the first 0.5 s, all of the events between M4.8 and M9.0 radiated similar seismic waves, and even at 2.0 s the difference between events from M6.3 to M9.0 is not significant. The moment magnitudes of the Tohoku-Oki earthquake in the first 3 s and 10 s are 6 and 7.2 respectively, and no anomalies are observed\textsuperscript{49) within the frequency range for seismological observations. This may not be the case for lower frequencies because there is some evidence that slow deformation occurred before the Tohoku-Oki earthquake.\textsuperscript{50),51) Further study is required to clarify the relationship between very slow deformation and the initial stage of rupture. Various observations suggest that earthquake rupture has near scale-invariant properties. Until the termination of propagation, rupture does not seem to have a characteristic length. This contradicts the characteristics of friction laws, explained in the previous section, because most friction laws have some intrinsic length scale, which defines the size of rupture that can be stopped. Therefore RSF simulations with a characteristic length produce repeated earthquakes that cannot stop spontaneously and are artificially stopped by the model boundary. What kind of numerical simulations can resolve this contradiction, between various sizes in observation and single size in RSF simulation? It is known that RSF simulations can produce earthquakes of various
sizes if the spatial discretization is insufficient to express slow deformation in the rupture nucleus. The elastic continuum is not properly modeled with such conditions and slip events are generated that are similar to those from a discrete model with many springs and sliders. This is an artifact, but it may reveal a real property of earthquakes. Even if a completely correct friction law is available in a mathematical form, we cannot apply it on a flat fault plane to simulate real earthquakes. Rather, the geometrical heterogeneity of real fault systems may end up discretizing the continuum with a perfect friction law into a discrete system with an approximated friction law.

Microscopically observable geometrical irregularities of a fault system may be approximated by a spatial distribution of macroscopic parameters on a fault plane. Fracture energy, \( G_c \), is one candidate for this approximation because geometrical irregularities must consume energy through inelastic deformation during rupture propagation. As a slip-weakening distance, \( D_c \), in a slip-weakening law depends on the characteristic wavelength of the geometry of the frictional surface, \( \tilde{D} \) we may assign \( D_c \) or \( G_c \) spatially, corresponding to the distribution of irregularities. This idea has been realized using fractal circular patches, each of which has \( D_c \) proportional to the radius of the patch. This model explains the scale-independent characteristics of earthquake rupture, including statistically self-similar rupture growth, near-constant but locally variable rupture propagation velocity, the independence of the initial seismic signal from the final size, and the spontaneous termination of rupture without special mechanisms. The model has also been applied to explain long-term seismicity and the rupture process of the 2011 Tohoku-Oki earthquake, as summarized in a recent review paper. Similar hierarchical models with RSF have also been studied. Different kinds of numerical simulations may also provide a better understanding of the effects of hierarchical fault structure, which is quite important but often neglected in continuum simulations with RSF. A relationship must exist between the power-law characteristics of dynamic earthquake rupture and the fractal nature of fault structure, but earthquake science has yet to reveal it.

Characteristics of slow earthquakes

Since 2000, anomalous signals have been detected in subduction zones around the world. These signals appear as weak and repetitive seismic signals and anomalies in long-term geodetic deformation. The former is radiated by tremors, which are considered as successive occurrences of tiny shear slip on the plate interface. They are often referred to as non-volcanic tremors, or tectonic tremors, to distinguish them from tremors associated with volcanoes. Short isolated signals, as shown in Fig. 2, are observed and characterized as a low-frequency earthquake. These signals are observed at frequencies greater than 1 Hz. The latter geodetic signals are caused by slow slip events (SSEs) on the plate interface that last for days, weeks, or even more than a year. Tremors and SSEs often occur simultaneously in approximately the same location and are referred to as episodic tremor and slip (ETS). At frequencies between tremors and SSEs, signals with periods of around 20 s are also observed. The sources of these signals are also shear slip on the plate interfaces and are often referred to as very low frequency (VLF) earthquakes, although they may be a part of broadband phenomena that bridge the gap between tremors and SSEs. These earthquake-like phenomena from tremors and SSEs are classified within the family of slow earthquakes. Several reviews of these phenomena are available.

Slow earthquakes have attracted the attention of researchers because they occur adjacent to the source areas of very large earthquakes. Figure 7 shows the locations of LFEs, VLF events, and SSEs in the Nankai subduction zone. Slow earthquakes delineate the bottom edge of the source areas of previous megathrust earthquakes. There are also shallow VLF events observed near the trough axis with accompanying tremors. Various combinations of LFEs, tremors, VLF events, and SSEs are observed in many subduction zones, such as Alaska, Cascadia, Mexico, Costa Rica, Ecuador, Chile, and New Zealand, as summarized in the aforementioned review papers. Similar tremors are also observed on strike-slip faults along the San Andreas Fault in California. Despite the prevalence of slow earthquakes, they have not been observed in the Tohoku-Oki region, where the density and quality of seismic observation networks are comparable to those in the Nankai region. The only similar phenomenon in the Tohoku-Oki region has been slow deformation observed before the Tohoku-Oki earthquake. This deformation was inferred from the slow migration of seismicity and from records of ocean bottom pressure gauges.

The characteristics of slow earthquakes are quite different from those of ordinary earthquakes. The
seismic moment is estimated for various slow earthquakes and compared with the duration of events (Fig. 8). Although the seismic moment of ordinary earthquakes is proportional to the cube of duration, the seismic moment of slow earthquakes is proportional to the duration. In other words, the seismic moment rate of slow earthquakes is almost constant, between $10^{12}$ and $10^{13}$ Nm/s. The difference between slow and fast (ordinary) earthquakes increases with the seismic moment. The difference is also observed in the stress drop associated with these events. Although the stress drop for fast earthquakes is in the range of $1-10$ MPa, the stress drop for SSEs larger than Mw 6 is estimated to lie in the range of 0.01–0.1 MPa. Similarly, the scaled energies are about $10^{-5}$ and $10^{-10}$ for fast and slow earthquakes, respectively.62),73)

The big differences in scaling laws and scaled energies suggest that fast and slow earthquakes are qualitatively different phenomena. It is unlikely that decelerated rupture propagation in ordinary earthquakes is similar to that in slow earthquakes. This difference is manifested through another kind of slow seismic event, tsunami earthquakes.74) These are rare, and fewer than ten tsunami earthquakes have been observed by digital observation networks. Tsunami earthquakes occur on the plate interface at subduction zones near the trench axis and generate tsunamis that are much larger than would be expected from the accompanying shaking. Major seafloor deformation occurs slowly for a long duration without radiating strong seismic energy. An example of a tsunami earthquake is the 1992 Nicaragua earthquake of Mw 7.6. Although the rupture process of a typical Mw 7.6 earthquake continues for about 30 s, the rupture propagation for the Nicaragua earthquake was slow and continued for about 100 s.75) This is long, but not as long as the duration of slow earthquakes. From the scaling law of slow earthquakes,71) the duration of a Mw 7.6 SSE is more than three years. Almost no events are observed between tsunami earthquakes and SSEs. For example, no earthquake-like phenomena of Mw 7.6 have lasted for one hour. Clearly, this is not a problem of
Detectability, as shown in Fig. 8. The same gap is also visible in the scaled-energy diagram (Fig. 8b). Although the scaled energy of tsunami earthquakes is about $10^{-6}$, which is smaller than that of ordinary earthquakes by about one order of magnitude, the scaled energies for VLF events and SSEs are far less than for tsunami earthquakes. Tsunami earthquakes likely represent ordinary earthquakes that occur in relatively soft materials near the trench axis, with low effective stress due to high pore pressure. They are qualitatively different from slow earthquakes accompanied by tremors.

How slow is the slip for slow earthquakes? For ordinary earthquakes, the slip rate at a point on the fault plane is about 1 m/s and the rupture propagation velocity is close to the S-wave speed. It is not easy to estimate the local slip rate for slow earthquakes, but the slip ("rupture" is not an appropriate word for a slow earthquake) propagation velocity can be estimated using tremor hypocenters as proxies. The migration of tremors over 10 km occurs within 10 min, which is in the order of 10 m/s. Migration over 100 km occurs within 10 days, which is in the order of 0.1 m/s. The propagation velocity lies between these values when the tremor front moves backwards. These values are significantly slower than the S-wave speed. The wide range in propagation speed and its apparent scale dependence suggests that slow earthquakes encompass some diffusion processes.

The difference between slow and fast earthquakes also appears in their frequency-size statistics. The magnitude of LFEs can be estimated with a traditional method using the amplitude of seismic waves. If we estimate the $b$-value of the GR law, it is more than 3, though the distribution cannot be explained using a power law; rather, it has a clear cutoff around M0.5. This contrasts with the frequency-size statistics for volcanic LFEs, which follow a power-law distribution with a $b$-value close to 2.83 It is not easy to define a magnitude for successive tremors because the definition of a tremor event is not clear. Instead, the cumulative distribution for the amplitude of the velocity envelope can be approximated using an exponential distribution, rather than a power law. As an exponential law has a characteristic size, the amplitude of the velocity envelope has a weak upper limit, which also limits the seismic moment rate and the energy rate of slow earthquakes. Such a limit can be observed only at a very short distance from the fault.
The stress change associated with slow earthquakes is very small. Therefore, it is not surprising that slow earthquakes are sensitive to small changes in stress, such as tidal stress\(^85\),\(^86\) and dynamic stress change due to surface waves radiated from distant large earthquakes\(^87\),\(^88\). Even stress variations of a few kPa can affect tremor occurrence. There is evidence that SSEs are also controlled by tidal stress.\(^89\) The relationship between tidal stress and tremor activity can be used to estimate fault rheology at deep plate interfaces.\(^90\) There is a long history of studies on the relationship between earthquakes and tidal stress. Although statistically significant correlation between seismicity and tides has been observed in some cases, the correlation is generally weak.\(^91\),\(^92\) However, significant correlation has been identified for earthquakes in subduction zones that occur prior to large earthquakes,\(^93\) including the 2011 Tohoku-Oki earthquake.\(^94\) In the Tohoku-Oki case, the correlation might be related to the observed slow slip before the earthquake.\(^95\),\(^96\) The relationships between tidal stress, slow earthquakes, and large earthquakes are important and can potentially help to improve probability forecasts of earthquakes.

**Modeling fast and slow earthquakes**

Slow earthquakes can be modeled in a manner similar to ordinary fast earthquakes, by assuming friction laws on the fault plane embedded in an elastic medium. Early RSF simulations for two-dimensional strike-slip faults produced slow episodic events for simulations where the slip-weakening distance was large.\(^97\) When the size of the seismic nucleus is comparable to the size of the seismogenic layer, the preslip cannot be accelerated into dynamic rupture. More realistic simulations in three-dimensional elastic media at subduction zones can generate slow slip events,\(^98\) but the size and recurrence interval of these events limit their applicability for explaining a wide range of slow earthquake behaviors. Introducing a cut-off slip rate into RSF laws enables slip rate weakening at low speeds and slip rate strengthening at high speeds, and better explains various slow slip events.\(^99\) Slip is accelerated near a rupture nucleus, but is soon decelerated at high speeds. Although the validity of using a cut-off slip rate in RSF laws has not been confirmed in laboratory experiments for typical rocks in seismogenic zones, the cut-off can be explained in terms of macroscopic behavior, as is explained further below. If the spatial distribution of the parameters in RSF laws is carefully controlled, both SSEs and large, fast earthquakes repeatedly occur in one model region.\(^100\) Some simulations have revealed temporal changes in SSE behavior during the earthquake cycle.\(^99\) The magnitude and recurrence interval of SSEs may change, reflecting stress accumulation on the source area for fast earthquakes. Although such behavior has not been confirmed in nature, due mainly to the limited period of observation, this kind of information may be utilized to improve probabilistic forecasts.

Although episodic SSEs are modeled in fairly narrow scale ranges in space and time, broadband slow earthquakes from tremors to SSEs are difficult to explain using ordinary RSF simulations. Even the seismological frequency range, from 0.01 Hz to 10 Hz, is too wide to be simulated numerically for a model with spatial variation. If we focus only on the time sequence, stochastic modeling is possible. The statistical characteristics of tremor waveforms can be modeled using a variation of the Langevin equation, the Ornstein-Uhlenbeck process,\(^101\) which is similar to Brownian motion with a damping term to prevent the divergence of the moment rate function. With such a damping term, the moment rate function produces characteristic values and the cumulative density function of its amplitude is close to an exponential function. The Langevin equation and Brownian motion are often used to explain diffusional processes, which may physically govern slow earthquakes.

Several studies have tried to explain the proximity of tremor location to the lower edge of the potential sources of large earthquakes. A simple friction law with static and dynamic friction levels can produce tremor-like events below the seismogenic layer of strike-slip faulting, if the stress drop is close to zero.\(^102\) This is consistent with tremors being triggered by tiny stress changes. Similar models have been presented for subduction zones, with a neutral velocity zone below the seismogenic zone.\(^103\) RSF simulations have successfully modeled the repeated occurrence of earthquakes and VLF events by introducing small-scale heterogeneity to the spatial distribution of RSF parameters near the bottom of the seismogenic zone.\(^104\) A velocity-strengthening friction law with small brittle patches can simulate tremors within an SSE zone.\(^105\),\(^106\) Such brittle patch models are versatile enough to explain a wide variety of tremor behavior, from LFEs with short pulses to long-lasting tremor sequences, by changing the density of the patches and the degree of velocity strengthening, which may be referred to as viscosity.\(^106\)
As reviewed, most slow earthquake models contain mechanisms to decelerate slip rate. Without these mechanisms, dynamic rupture easily reaches seismic slip velocity of about 1 m/s, which is roughly constrained by seismic wave speed (~3 km/s) and by strain change (10^{-4}) during rupture. Before reaching this elastic limit, the feedback mechanisms decelerate the slip rate below a certain level. The feedback mechanism may be inherent in friction laws, as suggested by SSE simulations, {98,99} or may be spatially introduced. In the model with brittle patches on a viscous background, stress decreases when slip starts, but only in the brittle region. However, once the slip area expands into the viscous background, the stress increases because of the viscosity and suppresses slip in the brittle region. If we consider the average frictional properties for a slipped region at the beginning of rupture in the brittle patch, slip rate weakening dominates initially, but changes to slip rate strengthening because of the viscous nature of the surrounding area. Macroscopically, this might corresponds to RSF with a cut-off. Such a feedback mechanism corresponds to the damping term in the stochastic model. In this sense, slow earthquakes represent a process governed by the diffusion equation, whereas ordinary fast earthquakes are governed by the elastic wave equation. This was the reason that the term “diffusional earthquake” has been suggested to reflect the scaling law for slow earthquakes. {71}

A conceptual model for seismogenic zones and the predictability of earthquakes

Earthquakes can be modeled by assuming friction laws, fault geometry, and the properties of surrounding materials. Slow earthquakes are modeled in a similar manner. Thus, by making assumptions about the spatial variation of parameters of these models, we can try to explain the different characteristics of seismic activity in different seismic regions. For subduction zones, in the 1980s, Lay and Kanamori {108} proposed the asperity model. In the original model, an asperity was defined as a conceptual property with high strength that persists on the plate interface for a long time. This usage of asperity was similar to the real contact area in friction theory, and a binary spatial distribution of locked and slipping regions was used to explain the diversity of earthquake occurrences. In the studies that followed, the asperity concept was more widely used but with a looser definition, which yielded some confusion in seismology.

In addition to the original definition of an area of high strength, an asperity may be defined as (1) a large slip area during an earthquake, (2) a locked area during an interseismic period, and (3) the slip rate weakening area in the RSF framework. As explained above, frictional strength should not be used as a criterion for earthquake occurrence in an elastic continuum. Therefore, the old definition should be replaced by any of these new definitions, or something else. However, in general, these three definitions do not refer to the same thing and can be quite different. There is one case in which these definitions are consistent. Given a small region of slip rate weakening surrounded by a rate-strengthening region, where steady slip is occurring, the small region is locked and accumulates stress, which is released repeatedly by fast slip events. This is a characteristic earthquake model, explained by early RSF simulations. Quasi-periodic repeating events, like the Kamaishi sequence, are approximated well by this model. Using this model, probabilistic forecasts can be achieved by measuring recurrence intervals. Highly accurate predictions may even be possible with this model if preslip is observed, which is a natural product of friction laws with a single length scale. Unfortunately, such sequences are rare, particularly for large earthquakes.

The various scaling laws of earthquakes and the lack of characteristic length cannot be explained by simply assigning binary properties on the fault plane like in asperity models with any definition. Rather, an important, but generally overlooked, feature should be included in numerical simulations, namely the irregularity of fault planes. Mathematically, it is very complicated to introduce real fractal topography or fault networks over a wide range of scales. Therefore, they must be approximated in other ways. Figure 9 proposes one examples of approximation, which is similar to the fractal circular patch model presented by Ide and Aochi {52}. As mentioned above, the purpose of this model is to assign fracture energy that is proportional to the size of the patch. Earthquake sources have certain properties that obey power laws with upper and lower limits. The differences in these limits should characterize the behavior of earthquakes in different seismic regions.

If there is no brittle patch, slip occurs steadily or episodically in the form of SSEs, without radiating seismic waves. In nature, some SSEs occur without generating detectable seismic signals. If the upper limit is small and only small patches are used, these patches are fractured by the background process and
radiate small seismic signals. Small patches have low fracture energy and are easily influenced by tiny stress changes. Small seismic signals from sparsely distributed patches and clustered patches may be observed as LFEs and tremors, respectively. This agrees with the model of tremors and SSEs presented by Ando et al.,\(^{105}\) where most of the system is controlled by background diffusional processes. The repeating earthquakes and characteristic earthquakes occur where the upper and lower limits are close, or where the system is not fractal. These events are similar to LFEs, but their fracture energy is too large to be triggered by tidal stress. Therefore, the overall seismicity is controlled by the characteristics of the seismic region, not by the background. The model of a single large patch may correspond to subduction zones such as Nankai, Cascadia, and southern Chile, where very large earthquakes occur over the whole subduction zone but current seismicity is low relative to expectations based on the velocity of relative plate motion.\(^{108}\) As a large patch requires a longer nucleation process, precursors should be easier to detect in these regions. In contrast, in seismic regions with broad hierarchical scale structures, a large earthquake may occur as a cascading process from a small patch. Such regions render predictions based on precursors meaningless.

Thus, Fig. 9 represents an extension of the original asperity model,\(^{107}\) with a slightly more clear physical definition for rupture units based on fracture energy. Although conceptual, this type of model may provide a framework for statistically quantifying the diverse characteristics of seismicity for fast and slow earthquakes. Such a model can be used for probabilistically forecasting seismicity based on the structural characteristics of the seismic regions.

**Conclusions**

Characterizing the physical properties and processes of earthquakes has progressed significantly over the last three decades as a result of high-quality seismic and geodetic observations, precise laboratory...
Earthquakes are complex rupture processes on fault systems within elastic rocky media. Power-law scaling relationships govern some of the different behaviors observed from small and large earthquakes. Mathematically, earthquake rupture is modeled using friction laws on fault planes that are embedded in elastic media. Many examples have been presented using popular RSF simulations. However, our incomplete knowledge of friction laws and limited computer resources prevent us from explaining the scale-independent properties of earthquake rupture over a broad scale range.

The most significant advances in the last decade have been the discovery and understanding of slow earthquakes, which are important components of the comprehensive modeling of earthquakes from large-scale plate motion. Fast and slow earthquakes are qualitatively different phenomena, clearly separated by a gap in their observable frequencies and size range. Friction laws are critical for understanding ordinary earthquakes, whereas qualitatively different mechanisms such as plastic or viscous deformation are essential for understanding slow deformation. Studies of rheologies associated with slow deformation at depth are progressing by using LFEs and tremors as creep meters on the fault planes. Comprehensive modeling of fast and slow earthquakes using realistic physical laws may be possible in the near future, although the predictive power of such numerical models may be limited. These models are deterministic but their nonlinear nature suggests that they must be very sensitive to slight differences in initial conditions. Therefore, probabilistic approaches will likely be required for prediction.

In 2011, the International Association of Seismology and Physics of the Earth’s Interior endorsed the recommendations of the ICEF report, which suggest that probabilistic forecasts will be the principal means to help communities prepare for potentially destructive earthquakes, rather than deterministic predictions based on precursors. Although this paper does not review practical methods for generating probabilistic forecasts, some models, such as the epidemic-type aftershock sequence (ETAS) model, have shown statistically significant probability gains relative to random forecasts. ETAS models are constructed based on our knowledge of two well-known earthquake power laws, the GR law and the Omori law, which were discovered about a century ago. However, recently acquired knowledge on the physics of earthquakes, such as friction laws, hierarchical structure, slow earthquakes, and repeating earthquakes, has not been incorporated appropriately. Our future studies aim to improve practical forecasting methods and to expand our knowledge of earthquake physics.

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Profile

Satoshi Ide was born in Tokyo in 1969, and grown up in Chiba prefecture. After graduating from the School of Science, the University of Tokyo in 1997, he started his research career as a research associate at Earthquake Research Institute, the University of Tokyo. He stayed for one year as a visiting researcher in Stanford University from 2000 to 2001. He then returned to the School of Science as a lecturer in 2002, and was promoted to associate professor in 2008, and to professor in 2013. He is studying physical processes behind various earthquakes, such as the Kobe, and the Tohoku-Oki earthquakes, many medium and small earthquakes, and a family of slow earthquakes. He was awarded the Young Scientist Award of Seismological Society of Japan in 2004, the Young Scientists' Prize of MEXT in 2006, the JSPS Prize and the Japan Academy Medal in 2014.