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A Physical Scaling Relation Between the Size of an Earthquake and its Nucleation Zone Size

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Abstract—A specific model of the earthquake nucleation that proceeds on a non-uniform fault is put forward to explain seismological data on the nucleation in terms of the underlying physics. The model is compatible with Gutenberg-Richter's similarity law for earthquake frequency-magnitude relation. A theoretical approach in the framework of fracture mechanics, based on a laboratory-based slip-dependent constitutive law, leads to the conclusion that the earthquake moment M_o scales with the third power of the critical slip displacement D_c and the critical size $2L_c$ (L_c, half-length) of the nucleation zone. This scaling relation quantitatively explains seismological data published, and it predicts that $2L_c$ is of the order of 10 km for earthquakes with $M_o = 10^{21}$ Nm, 1 km for earthquakes with $M_o = 10^{18}$ Nm, and 100 m for earthquakes with $M_o = 10^{15}$ Nm, under the assumption that the breakdown stress drop $\Delta\tau_b = 10$ MPa. However, L_c depends on not only D_c but also $\Delta\tau_b$, so that the scaling relation between L_c and D_c may be violated by $\Delta\tau_b$, because $\Delta\tau_b$ potentially takes any value in a wide range from 1 to 10^2 MPa, depending on the seismogenic environment. The good agreement between the theoretical relation and observed results suggests that a large earthquake may result from the failure of a large patch of high rupture growth resistance, whereas a small earthquake may result from the breakdown of a small patch of high rupture growth resistance. The present result encourages one to pursue the prediction capability for large earthquakes.

Key words: Earthquake nucleation, inhomogeneous fault, high rupture growth resistance, a slip-dependent constitutive law.

Introduction

Physical nature of the shear rupture nucleation on an inhomogeneous fault has been revealed by high-resolution laboratory experiments (OHNAKA and KUWAHARA, 1990; OHNAKA and SHEN, 1999). In particular, OHNAKA and SHEN (1999) have conclusively demonstrated that the rupture nucleation consists of two phases: an initial, stable and quasi-static phase, and the subsequent, unstable and accelerating phase, and that the nucleation zone size and its duration are consistently scaled in terms of the slip-dependent constitutive law that governs the rupture process. One of the constitutive law parameters, which is referred to as the critical slip displacement, has been found to be a scaling parameter prescribed by the character-

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istic length representing geometric irregularity of the rupture surfaces. These provide a basis for modeling the earthquake nucleation in terms of the underlying physics, and allow one to discuss a scaling relation between the sizes of an earthquake and its nucleation zone specifically. The nucleation is defined here as the transition process from an initial phase of stable, quasi-static rupture up to the critical stage beyond which the rupture propagates at a high-speed close to sonic velocities.

As discussed in an earlier paper (OHNAKA and SHEN, 1999), there are two classes of physical quantities inherent in the shear rupture: scale-dependent quantities, and scale-independent quantities. The scale-dependent quantities include the nucleation zone size, its duration, the shear rupture energy, and the seismic moment. OHNAKA and SHEN (1999) have shown that scale dependence of scale-dependent physical quantities is commonly ascribed to the scale dependence of the critical slip displacement. The critical slip displacement D_c is defined as the slip displacement required for the local strength in the breakdown zone behind the rupture front to degrade to a residual friction stress level. The laboratory experiments have also demonstrated that D_c is greatly affected by the characteristic length representing geometric irregularity of the rupturing surfaces. On the other hand, theoretical analyses (RICE, 1980; OHNAKA and YAMASHITA, 1989) show that D_c is directly related to the breakdown zone size, which is relevant to the critical size of the nucleation zone. This suggests that the size of an earthquake should be related to the critical size of the nucleation zone, given that the larger the entire fault size, the larger characteristic length is in general included in the fault zone.

The objective of this paper is to show theoretically, on the basis of a specific model of the nucleation based on the laboratory experiments, how the size of a mainshock earthquake scales with the critical size of the nucleation zone. More specifically, a scaling relation that the earthquake moment is proportional to the third power of the critical size of the nucleation zone will be derived, based on a physical model of the nucleation. ELLSWORTH and BEROZA (1995) have empirically found, among other things, that the mainshock seismic moment scales with the seismic nucleation zone size, and hence the relation derived theoretically will be compared with seismological data on earthquake nucleation, particularly data analyzed by ELLSWORTH and BEROZA (1995). It will be shown that the theoretical relation agrees quantitatively with those data on earthquake nucleation.

A Constitutive Law for Earthquake Rupture

There are increasing amounts of evidence that the earthquake rupture that takes place in the brittle layer in the earth's crust is a mixed process between what is called frictional slip failure and fracture of intact rock mass, so that the constitutive law for the earthquake rupture should be formulated as a unifying law that governs both frictional slip failure and shear fracture of intact rock. In fact, shear fracture strength of intact material is the upper end member of the strength of frictional slip failure on the mating surfaces of the same material (OHNAKA, 1996; OHNAKA *et al.*, 1997). The slip-dependent constitutive law is a unifying law that governs both frictional slip failure and shear fracture of intact rock. We will derive a scaling relation that exists between the seismic moment of a mainshock earthquake and its nucleation zone size, by assuming a laboratory-based slip-dependent constitutive law as the governing law for the earthquake rupture in the framework of fracture mechanics.

The slip-dependent constitutive formulation presumes the slip displacement to be an independent and essential variable, and the rate- or time-dependence to be of secondary significance. That is, the shear traction is expressed as a function of the slip displacement in this formulation (Fig. 1), and the parameters prescribing the law, such as the peak shear strength τ_p, the breakdown stress drop $\Delta\tau_b$, and the critical slip displacement D_c, are assumed to be an implicit function of the slip velocity or time (OHNAKA *et al.*, 1997; OHNAKA, 1998). In Figure 1, τ_i denotes the initial strength on the verge of slip at the rupture front, τ_p the peak shear strength attained at the slip displacement D_a, τ_r the residual frictional stress, and D_c the critical slip displacement defined as the minimum amount of slip required for the shear strength to degrade to τ_r. The breakdown stress drop $\Delta\tau_b$ is defined as the shear stress difference between τ_p and τ_r.

The slip-dependent constitutive law is prescribed by a set of constitutive parameters (τ_i, τ_p, $\Delta\tau_b$, D_a, D_c), and the values of these parameters are affected by seismogenic environments. In particular, the rupturing surfaces are in mutual

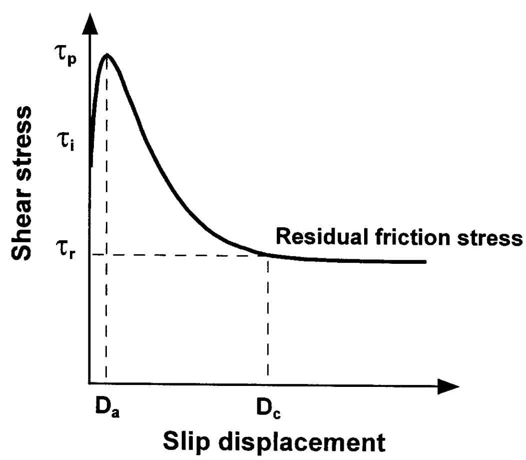

Figure 1
A slip-dependent constitutive relation for the shear rupture. τ_i denotes the initial shear stress on the verge of slip, τ_p denotes the peak shear strength, τ_r denotes the residual friction stress, D_a denotes the slip displacement at which the peak strength is attained, and D_c denotes the critical slip displacement.

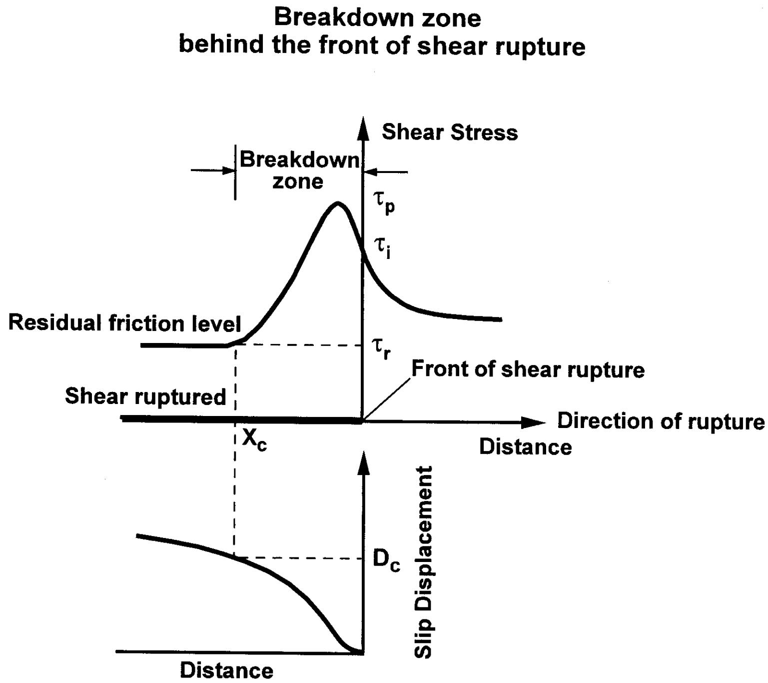

Figure 2
A physical model of the breakdown zone behind the rupture front, derived from a constitutive relation as shown in Figure 1. τ_i denotes the initial shear stress on the verge of slip at the rupture front, τ_p denotes the peak shear strength, τ_r denotes the residual friction stress, D_c denotes the critical slip displacement, and X_c denotes the breakdown zone size.

contact and are interacting throughout the breakdown process of the shear rupture, and hence D_a and D_c are severely influenced by geometric irregularity of the rupturing surfaces. As noted in the previous section, D_c is scale-dependent, and the constitutive law for shear rupture includes this scaling parameter. In this sense, the constitutive law is also scale-dependent. This scale-dependence is a very important property that plays an essential role in scaling scale-dependent physical quantities inherent in the shear rupture (OHNAKA and SHEN, 1999).

The breakdown zone is defined as the zone behind the rupture front over which the shear strength degrades transitionally to a residual friction stress level along the fault. Figure 2 shows a model of the breakdown zone behind a rupture front, which can be specified once a slip-dependent constitutive relation as shown in Figure 1 is given. The breakdown zone size X_c in the phase of dynamic rupture propagating at a high speed V_c is closely related to the critical slip displacement D_c by the relation (OHNAKA and YAMASHITA, 1989):

$$\frac{D_c}{X_c} = k\frac{\Delta\tau_b}{\mu} \tag{1}$$

where μ is the rigidity, and k is a well-defined, dimensionless quantity. The dimensionless quantity k has been calculated as

$$k = \frac{\Gamma}{\pi^2 \xi C(V_c)} \tag{2}$$

for a dynamic slip-weakening model (OHNAKA and YAMASHITA, 1989). Here, ξ in equation (2) represents a numerical parameter, $C(V_c)$ represents a known function of the rupture velocity V_c and Γ is a dimensionless parameter defined by (OHNAKA and YAMASHITA, 1989)

$$\Gamma = \int_0^1 \frac{\sigma(Y)}{\sqrt{Y}} dY \tag{3}$$

where $\sigma(Y)$ is the non-dimensional shear strength at a non-dimensional distance Y measured from the rupture front in the breakdown zone. $C(V_c)$ has a different functional form according to either in-plane shear (mode II) or anti-plane shear (mode III) (see OHNAKA and YAMASHITA, 1989). Relation (1) will be used later for deriving a scaling relation between the seismic moment of a mainshock earthquake and its nucleation zone size.

An Earthquake Nucleation Model

There is commanding evidence that earthquake faults in the seismogenic layer are inherently inhomogeneous. For instance, an "asperity" (KANAMORI and STEWART, 1978; KANAMORI, 1981) or "barrier" (AKI, 1979, 1984) on earthquake faults is a local patch of high rupture growth resistance on the fault. Such a local patch of high rupture growth resistance may be a mainfestation of geometric structure and/or irregularity of the fault (zone). An earthquake fault in general exhibits a geometrically irregular structure of various scales, and high resistance of the rupture growth can be attained at portions of fault bend or stepover, and at interlocking asperities on the fault surfaces with geometric irregularity.

Recent high-resolution laboratory experiments (OHNAKA and SHEN, 1999) have demonstrated that fault inhomogeneity plays an important role in scaling the rupture nucleation process; in other words, the characteristic length representing geometric irregularity of the rupturing surfaces is a key to scaling scale-dependent physical quantities inherent in the rupture, including the nucleation. It is therefore unrealistic to assume a uniform fault, and hence we assume an inhomogeneous fault model. For simplicity, however, the fault model assumed here comprises local, strong patches of high rupture growth resistance, and the remaining weak portion. There is a physical constraint to be imposed on the strength of such local patches; that is, the upper limit of the patch strength equals the shear strength of intact rock at lithospheric conditions. Such a local patch of high rupture growth resistance on the fault may be a physical manifestation of what has been called "barrier" (AKI, 1979, 1984) or "asperity" (KANAMORI and STEWART, 1978; KANAMORI, 1981). The size of a local patch of high rupture growth resistance may represent a characteristic distance on the fault.

We presume that the earthquake source at shallow depths is a shear rupture instability that takes place on such an inhomogeneous fault in the seismogenic layer. The laboratory experiments (OHNAKA and KUWAHARA, 1990; OHNAKA and SHEN, 1999) reveal that the rupture begins to nucleate at a position of the lowest resistance of rupture growth (or weakest toughness), and that the nucleation necessarily proceeds toward the remaining unbroken area of higher rupture growth resistance on the fault. One may argue that the rupture should nucleate at a position of high stress concentration. However, this is not the case for the rupture that nucleates on an inhomogeneous fault where the strength distribution is non-uniform, because the stress cannot concentrate beyond the strength. It will thus be reasonable to assume that the earthquake nucleation begins to occur, as a consequence of the tectonic loading, somewhere at a weakest portion on the fault, and that the nucleation is necessarily required to proceed toward the remaining unbroken area of higher rupture growth resistance. At this stage, the nucleation proceeds stably and quasi-statically, because no elastic strain energy stored in the medium surrounding the fault has been released. For the rupture to develop spontaneously, the elastic strain energy stored in the surrounding medium needs to be released.

As noted in the previous section, it has been demonstrated conclusively (OHNAKA and SHEN, 1999) that the nucleation process consists of two phases: an initial, stable and quasi-static phase (phase I), and the subsequent, unstable and accelerating phase (phase II). Phase I is a steady rupture growth controlled by the rate of an applied load, such as the tectonic loading. On the other hand, phase II is a spontaneous rupture extension driven by the release of the elastic strain energy stored in the surrounding medium (OHNAKA and SHEN, 1999). This physically means that the rupture cannot begin to propagate abruptly at a high speed close to sonic velocities immediately after the stored elastic strain energy is released. When the constitutive property of the fault is inhomogeneous, the rupture is required to grow at accelerating speeds from the quasi-static phase to the phase of high-speed rupture propagation. It can be inferred for major earthquakes that the time required for the rupture to grow from a slow phase of the rupture velocity being of the order of a few mm/s to the phase of high-speed rupture close to the shear wave velocity, is of the order of a few to a few tens of hours (OHNAKA and SHEN, 1999).

If the fault is tough enough, a large amount of the elastic strain energy can be stored in the medium surrounding the fault. However, if the fault is very weak everywhere on the entire fault, an adequate amount of the strain energy to bring about a large earthquake cannot be stored. We assume that patches of high rupture growth resistance on a fault are tough enough for an adequate amount of the elastic

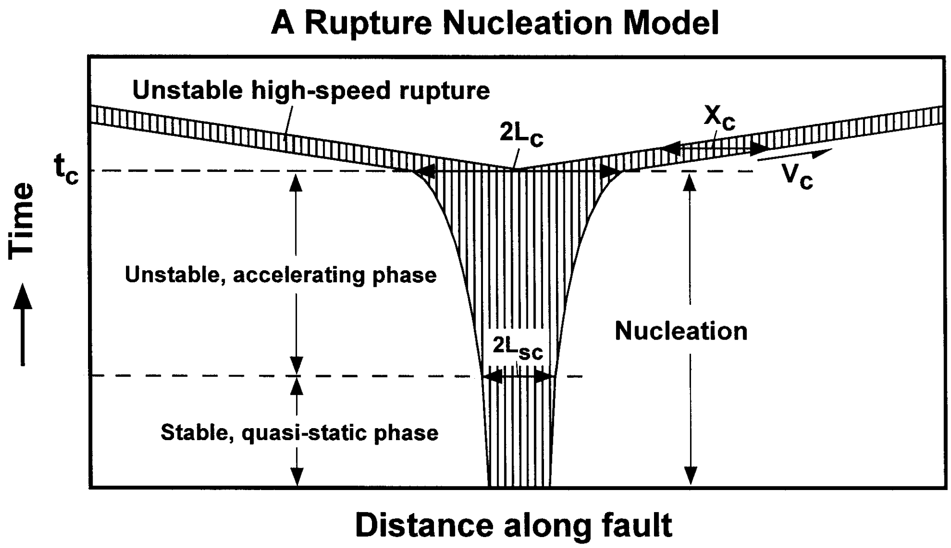

Figure 3
Schematic diagram of a rupture nucleation model. In this model, the rupture initially grows stably and quasi-statically at a steady speed to a critical length $2L_{sc}$, from which it extends spontaneously at accelerating speeds reaching another critical length $2L_c$ at a time $t = t_c$. Beyond the critical length $2L_c$, the rupture propagates bi-directionally at a constant, high-speed V_c close to the shear-wave velocity. The hatched portion represents the zone in which the breakdown (or slip-weakening) is proceeding with time. X_c denotes the breakdown zone size, and $2L_c$ denotes the critical size of the nucleation zone.

strain energy to be stored, but that the remaining portion of the fault is too weak to sustain an adequate amount of the elastic strain energy. In this case, the rupture may not be able to propagate spontaneously until the critical condition at which one of the patches of high rupture growth resistance is broken down by slow growth of rupture nucleation is met. If the patch size is geometrically large, the total amount of the elastic strain energy stored in the surrounding medium is large, so that the resulting earthquake will be large. On the other hand, a large amount of D_c is by definition required for the breakdown of a geometrically large patch, and the large amount of D_c necessarily leads to a large size of the nucleation zone. This qualitatively predicts that the size of an earthquake scales with the nucleation zone size.

With all the above mentioned in mind, we assume a specific rupture nucleation model shown in Figure 3. The hatched portion in Figure 3 shows the zone in which the breakdown (or slip-weakening) is proceeding with time. In this model, the nucleation initially proceeds stably and quasi-statically at a steady speed V_{st} (phase I) up to a critical length L_{sc}, beyond which the nucleation extends spontaneously at accelerating speeds (phase II) to another critical length L_c (at a time $t = t_c$ in Fig. 3). It has been found that the rupture growth rate at phase II increases with an increase in the rupture growth length according to a power law (OHNAKA and SHEN, 1999). After the rupture growth length has reached the critical length L_c at $t = t_c$, the rupture propagates at a constant, high-speed V_c close to the shear-wave velocity. This model is based on the fact revealed with recent high-resolution laboratory experiments (OHNAKA and SHEN, 1999). The behavior of rupture growth changes at the critical length $2L_{sc}$ (L_{sc}, half-length) from a quasi-static phase controlled by the rate of the tectonic loading to a self-driven, dynamic phase controlled by the inertia. Note, however, that the critical size of the nucleation zone is not defined here as $2L_{sc}$, but defined as $2L_c$, so as to make it possible to compare the present model with seismological data on the nucleation.

Although the model shown in Figure 3 is bi-directional, we can also assume a similar model in which the rupture extends uni-directionally. Since whether the rupture extends bi-directionally or uni-directionally is extraneous to the subject to be discussed here, we simply employ a bi-directional rupture model discussed previously (OHNAKA, 1996; OHNAKA and SHEN, 1999). In the present bi-directional rupture model, the critical size $2L_c$ (L_c, half-length) of the nucleation zone is related to the breakdown zone size X_c in the phase of dynamic, high-speed rupture as follows:

$$\left.\begin{array}{lll} L < X_c & \text{for} & t < t_c \\ L_c = X_c & \text{at} & t = t_c \end{array}\right\}. \tag{4}$$

We thus have from (1) and (4)

$$L_c = \frac{1}{k}\frac{\mu}{\Delta\tau_b} D_c \tag{5}$$

which shows that the critical size L_c of the nucleation zone can be expressed explicitly in terms of the slip-dependent constitutive law parameters D_c and $\Delta\tau_b$ (OHNAKA and SHEN, 1999). It may be worthwhile noting here that $\Delta\tau_b$ and D_c represent the breakdown stress drop and the critical slip displacement, respectively, around the critical stage $t = t_c$ beyond which the rupture propagates at a high-speed V_c close to sonic velocities.

Rupture Growth Resistance

The rupture growth resistance is a distinct physical quantity, in the framework of fracture mechanics, defined as the shear rupture energy required for the rupture front to further grow. The shear rupture energy G_c is defined by (PALMER and RICE, 1973)

$$G_c = \int_0^{D_c} [\tau(D) - \tau_r] \, dD \tag{6}$$

where $\tau(D)$ represents a slip-dependent constitutive relation, which governs the relation between the shear traction τ and the slip displacement D on the fault in the breakdown zone behind the rupture front. Once a slip-weakening constitutive relation $\tau(D)$ is specified, G_c can be calculated from (6), and expressed in terms of the constitutive law parameters $\Delta\tau_b$ and D_c as (OHNAKA and YAMASHITA, 1989)

$$G_c = \frac{1}{2} \Gamma \Delta\tau_b D_c \tag{7}$$

where Γ has been defined in equation (3).

Γ can be regarded as virtually constant. If a simplified, linear slip-weakening relation is assumed, it is derived from (6) that Γ is exactly unity. However, laboratory-based slip-dependent constitutive relations are found to be nonlinear. OHNAKA and YAMASHITA (1989) have estimated Γ to be about 1/2 from (3) by numerical calculation, using nonlinear constitutive relations observed during dynamic frictional slip failure on a simulated fault in the laboratory. On the other hand, $\Gamma = 1$ has been evaluated from experimental data on nonlinear constitutive relations for the shear fracture of intact rock sample tested at lithospheric conditions (OHNAKA *et al.*, 1997). It is thus confirmed that the assumption that Γ is of the order of unity is valid even for nonlinear constitutive relations, and hence, $\Gamma = 1$ will be used later in the present analysis.

We note here the energy required for fault rupture. The shear rupture energy $\overline{G_c}$ averaged over the entire fault area S is defined by:

$$\overline{G_c} = \frac{1}{S} \int_S G_c \, dS. \tag{8}$$

Let S_{A1} be the area of the geometrically largest patch of high rupture growth resistance (which may be called asperity) on the fault. Equation (8) may be rewritten as follows:

$$\overline{G_c} = \frac{1}{S} \left[\int_{S_{A1}} G_c \, dS + \int_{S - S_{A1}} G_c \, dS \right] \tag{9}$$

where the first term of the right-hand side of equation (9) denotes the integral over the area S_{A1} of the geometrically largest asperity on the fault, and the second term denotes the integral over the rest $S - S_{A1}$ of the fault area. The first term of the right-hand side of equation (9) is a fraction a (< 1) of $S\overline{G_c}$, so that we have

$$\overline{G_c} = \frac{1}{aS} \int_{S_{A1}} G_c \, dS = \left(\frac{S_{A1}}{aS} \right) \overline{G_c^{A1}} \tag{10}$$

where

$$\overline{G_c^{A1}} = \frac{1}{2}\Gamma\Delta\tau_b^{A1}D_c^{A1}. \tag{11}$$

Here, $\Delta\tau_b^{A1}$ and D_c^{A1} represent the breakdown stress drop and the critical slip displacement, respectively, required for the geometrically largest asperity on the fault to break down.

Equation (7) or (11) shows that the shear rupture energy is directly related to the critical slip displacement. This indicates that G_c is necessarily scale-dependent, because D_c is scale-dependent. The scale-dependence of D_c has been more fully discussed by OHNAKA (1998), and OHNAKA and SHEN (1999). The scale-dependence of G_c may be understood from the following consideration. Real rupture surfaces cannot be plane, but they have geometric irregularity (or roughness). The rougher the ruptured surfaces, the larger the real surface area becomes. However, the irregularity of real rupture surfaces is not taken into consideration to evaluate G_c, and hence G_c may be called the apparent rupture energy. It thus follows that the apparent rupture energy G_c becomes larger as the rupture surfaces become rougher (OHNAKA *et al.*, 1997).

More specifically, the rupture surfaces in general have a band-limited fractal nature (see Discussion section). In this case, there is a characteristic length scale representing geometric irregularity of the rupture surfaces, and this characteristic length becomes longer on the rougher rupture surfaces. Since D_c scales with the characteristic length, G_c also scales with the characteristic length. This will be more fully discussed elsewhere, since in-depth discussion about this is beyond the scope of the present paper. A large-scale fault tends to include a large characteristic length scale. Thus, G_c defined by (6) or (7) is not only dependent on the fault material property, but also scale-dependent.

A Scaling Relation between Seismic Moment and Critical Size of Nucleation Zone

We first consider how the seismic moment M_o of a mainshock earthquake scales with the critical slip displacement D_c. Let $\bar{D}$ be the slip amount averaged over the entire fault area S, and $\overline{\Delta\tau}$ be the stress drop $\Delta\tau$ averaged over S. The seismic moment M_o is defined by (AKI, 1966)

$$M_o = \mu\bar{D}S. \tag{12}$$

The earthquake rupture finally arrests when the driving force has become lower than the rupture growth resistance. The condition of the rupture arrest may thus be written as $L(\Delta\tau)^2 \leq \kappa\mu G_c$ (κ being a constant), if the breakdown zone size X_c behind the rupture front is sufficiently small compared with the final fault length L. The sign of equality in this equation represents the condition at which the rupture grows quasi-statically. However, this may be regarded as the 'critical' condition below which the dynamic rupture must arrest. In the situation that the driving force (or $\Delta\tau$) gradually diminishes with distance toward the fault end, or that the resistance G_c gradually increases with outward distance from the fault end, we thus expect that the following relation:

$$(\overline{\Delta\tau})^2 = \kappa\mu \frac{\overline{G_c}}{L} \tag{13}$$

holds between $\overline{G_c}$, $\overline{\Delta\tau}$, and L.

If we further assume the following scaling relations:

$$S = c_1 L^2 \tag{14}$$

and

$$\bar{D} = c_2 L \tag{15}$$

where c_1 and c_2 are numerical constants, we have from (10)–(15)

$$M_o = c_1 c_2 \left(\frac{\kappa\Gamma}{2}\right)^3 \left(\frac{S_{A1}}{aS}\right)^3 \left(\frac{\mu}{\overline{\Delta\tau}}\right)^3 \left(\frac{\Delta\tau_b^{A1}}{\overline{\Delta\tau}}\right)^3 \mu (D_3^{A1})^3. \tag{16}$$

On the right-hand side of equation (16), the asperity area S_{A1}, the entire fault area S, and the critical slip displacement D_c^{A1} are scale-dependent, while the rest of the parameters are scale-independent. If it is assumed that the ratio $S_{A1}/(aS) = \overline{G_c}/\overline{G_c^{A1}}$ is scale-independent, we find that the critical slip displacement D_c^{A1} is the only scale-dependent parameter on the right-hand side of equation (16). If we assume that the stress drop $\overline{\Delta\tau}$ averaged over the fault area and the breakdown stress drop $\Delta\tau_b$ in the breakdown zone behind the rupture front are virtually constant (in a statistical sense), the equation (16) predicts theoretically that the mainshock seismic moment is proportional to the third power of the critical slip displacement at a geometrically largest patch of high rupture growth resistance on the fault.

The scaling relation $M_o \propto D_c^3$ at an asperity of the geometrically largest size on the fault, is well grounded, because D_c is by definition the critical slip displacement required for the breakdown of a local patch of the high rupture growth resistance, and because a large amount of the critical slip displacement is required for the breakdown of a patch of geometrically large size.

Similarly, from (5) and (16), we have the relation between the seismic moment M_o and the critical size $2L_c$ of the nucleation zone:

$$M_o = c_1 c_2 \left(\frac{k\kappa\Gamma}{4}\right)^3 \left(\frac{S_{A1}}{aS}\right)^3 \left(\frac{\Delta\tau_b^{A1}}{\overline{\Delta\tau}}\right)^6 \mu (2L_c)^3. \tag{17}$$

Equations (16) and (17) lead to the conclusion that the seismic moment is primarily prescribed by the property of the geometrically largest asperity on the fault, and that the seismic moment be proportional to the third power of the critical slip displacement and the critical size of the nucleation zone, if the assumptions made are physically reasonable.

The relations (16) and (17) have been derived theoretically by assuming an asperity model of earthquake fault, and scaling laws, in the framework of fracture mechanics based on a slip-dependent constitutive law. To what extent these theoretical relations are justified can be checked by comparing them with seismological data on the nucleation. This will be done in next section.

Comparison of Theoretical Relations with Seismological Data

Slip-dependent constitutive law parameters have been estimated for actual earthquakes by PAPAGEORGIOU and AKI (1983), ELLSWORTH and BEROZA (1995), and IDE and TAKEO (1997). PAPAGEORGIOU and AKI (1983) estimated the breakdown stress drop $\Delta\tau_b$, the critical slip displacement D_c, the breakdown zone size X_c, and the shear rupture energy G_c for earthquakes with moderate to large earthquakes, based on a specific barrier model for earthquake faulting. ELLSWORTH and BEROZA (1995) analyzed near-source recordings of the slow initial P wave for earthquakes with a wide magnitude range 2.6 to 8.1, and they evaluated the breakdown stress drop $\Delta\tau_b$ and the nucleation zone size L_c. IDE and TAKEO (1997) estimated constitutive relations for the 1995 Kobe earthquake from near-field seismic waves by waveform inversion and the solution of elastodynamic equations using a finite difference method. The basic parameters of the present concern, $\Delta\tau_b$, D_c, X_c, and L_c, together with the seismic moment M_o, are compiled for these earthquakes in Table 1.

Since D_c for earthquakes was not evaluated by ELLSWORTH and BEROZA (1995), we have evaluated this parameter for earthquakes listed in Table 1 in their paper. D_c can be evaluated from equation (5). In order to evaluate $\Delta\tau_b$ and L_c, Ellsworth and Beroza assumed that the longitudinal wave velocity $V_P = 6$ km/s, the shear wave velocity $V_S = V_P/1.73$, the rupture velocity $V = 0.8V_S$, and $\mu = 30{,}000$ MPa. Under these assumptions, and assuming that $\Gamma/\xi = 3.3$ (OHNAKA and YAMASHITA, 1989), we have $k = 2.9$ for an in-plane shear crack (mode II), and $k = 3.5$ for an anti-plane shear crack (mode III). With this in mind, we have assumed that $k = 3$ for a circular crack model employed by Ellsworth and Beroza, and evaluated D_c for those earthquakes from (5). The evaluated values for D_c are listed in Table 1.

For the estimate of L_c for the Kobe earthquake, we have presumed that at its hypocenter, whose depth was determined to be 16 km, the nucleation reached the critical stage beyond which the rupture propagated at a high speed V_c. From (5), we have estimated $L_c = 1700$ m using the values $\Delta\tau_b = 3$ MPa and $D_c = 0.5$ m evaluated by IDE and TAKEO (1997). They suggest, however, that 0.5 m is the upper limit of D_c at a deeper part of the fault where the nucleation must have reached the critical stage. In this case, 1700 m will be the upper limit of L_c for the Kobe earthquake. If $\Delta\tau_b = 3$ MPa and, for instance, $D_c = 0.3$ m are assumed, we have $L_c = 1000$ m. These suggest that L_c for the Kobe earthquake was of the order of 1×10^3 m.

Table 1

Constitutive parameters for earthquakes

Event	M_o (Nm)	$\Delta\tau_b$ (MPa)	D_c (m)	X_c (m)	L_c (m)	References
Fort Tejon 1857	$(5.3–9.0) \times 10^{20}$	50–70	3–4	1000–2000		Papageorgiou and Aki (1983)
Kern County 1952	2.0×10^{20}	68	3	1000		Papageorgiou and Aki (1983)
San Fernando 1971	1.2×10^{19}	48	1	500		Papageorgiou and Aki (1983)
Borrego Mountain 1968	6.3×10^{18}	30–40	0.4	600		Papageorgiou and Aki (1983)
Long Beach 1933	2.8×10^{18}	20	0.4	600		Papageorgiou and Aki (1983)
Parkfield 1966	1.4×10^{18}	20	0.4	500		Papageorgiou and Aki (1983)
19 Sep 1985	1.4×10^{21}	5.0	3.2		6300	Ellsworth and Beroza (1995)
28 Jun 1992	9×10^{19}	4.1	1.4		3400	Ellsworth and Beroza (1995)
25 Apr 1989	2.4×10^{19}	3.4	2.9×10^{-1}		850	Ellsworth and Beroza (1995)
18 Oct 1989	2.8×10^{19}	1.9	4.2×10^{-1}		2200	Ellsworth and Beroza (1995)
17 Jan 1994	1×10^{19}	40	2.4		600	Ellsworth and Beroza (1995)
15 Oct 1979	6×10^{18}	14	1.4		1000	Ellsworth and Beroza (1995)
9 Jun 1980	4.8×10^{18}	6.0	2.5×10^{-1}		410	Ellsworth and Beroza (1995)
24 Oct 1993	5.8×10^{18}	5.5	2.9×10^{-1}		520	Ellsworth and Beroza (1995)
28 Jun 1992	2×10^{18}	8.1	4.5×10^{-1}		560	Ellsworth and Beroza (1995)
23 Apr 1992	1.4×10^{18}	17	2.6×10^{-1}		150	Ellsworth and Beroza (1995)
31 May 1990	7.5×10^{17}	64	8.3×10^{-1}		130	Ellsworth and Beroza (1995)
29 Jun 1992	4.8×10^{17}	2.9	1.2×10^{-1}		420	Ellsworth and Beroza (1995)
28 Jun 1991	2.8×10^{17}	18	8.6×10^{-1}		480	Ellsworth and Beroza (1995)
20 Mar 1994	8.9×10^{16}	42	9.7×10^{-1}		230	Ellsworth and Beroza (1995)
3 Dec 1988	2.4×10^{16}	15	3.3×10^{-1}		220	Ellsworth and Beroza (1995)
16 Jan 1993	2.4×10^{16}	8.8	2.6×10^{-1}		300	Ellsworth and Beroza (1995)
14 Nov 1993	2.0×10^{16}	1.1	4.5×10^{-2}		410	Ellsworth and Beroza (1995)
11 Aug 1993	1.3×10^{16}	8.3	2.2×10^{-1}		270	Ellsworth and Beroza (1995)
3 Feb 1994	2×10^{15}	9.6	7.0×10^{-2}		73	Ellsworth and Beroza (1995)
6 Feb 1994	1.4×10^{15}	8.0	3.7×10^{-2}		46	Ellsworth and Beroza (1995)
2 Feb 1994	5×10^{14}	2.3	1.5×10^{-2}		66	Ellsworth and Beroza (1995)
27 Oct 1991	3.5×10^{14}	6.5	3.4×10^{-2}		52	Ellsworth and Beroza (1995)
6 Feb 1994	3.5×10^{14}	8.3	3.0×10^{-2}		36	Ellsworth and Beroza (1995)
26 Oct 1992	2.5×10^{14}	100	2.8×10^{-2}		28	Ellsworth and Beroza (1995)
1 Feb 1994	2.5×10^{14}	3.4	1.1×10^{-2}		32	Ellsworth and Beroza (1995)
2 Jan 1990	1.8×10^{14}	8.2	3.2×10^{-2}		39	Ellsworth and Beroza (1995)
4 Feb 1994	1.8×10^{14}	8.4	2.9×10^{-2}		35	Ellsworth and Beroza (1995)
8 Mar 1994	1.3×10^{14}	0.4	4×10^{-3}		100	Ellsworth and Beroza (1995)
30 Jan 1988	8.1×10^{13}	60	1.1×10^{-1}		18	Ellsworth and Beroza (1995)
8 Nov 1992	7.9×10^{12}	5.4	5.9×10^{-3}		11	Ellsworth and Beroza (1995)
Kobe 17 Jan 1995	1.9×10^{19}	1.5	1		1000	Ide and Takeo (1997)
		3	< 0.5		1700	

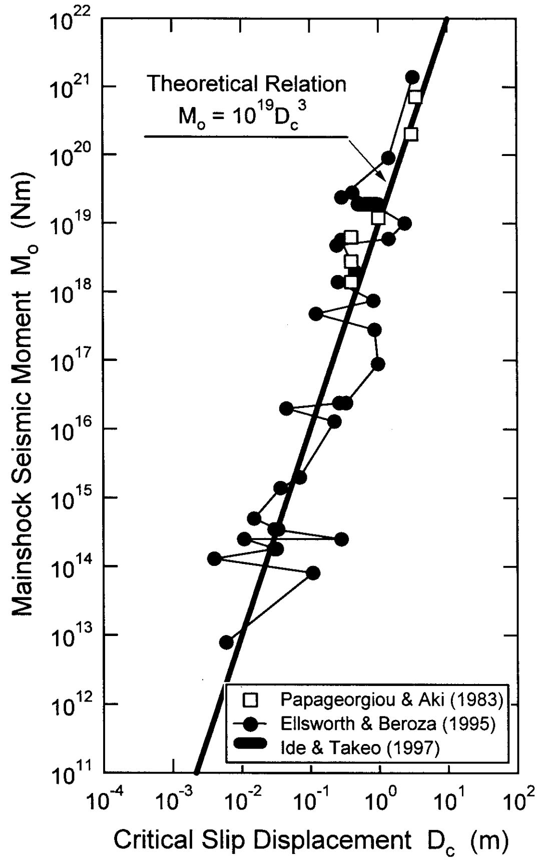

Figure 4
A plot of the seismic moment M_o against the critical slip displacement D_c for earthquakes listed in Table 1. The theoretical scaling relation, denoted by a thick line in the figure, is compared with seismological data.

SHIBAZAKI and MATSU'URA (1998) computed the far-field velocity waveform from a circular fault model with the slip-time function obtained by the numerical simulation of a slip failure event governed by a slip-dependent constitutive law, and they demonstrated that the seismic nucleation phase defined by ELLSWORTH and BEROZA (1995) on a seismogram corresponds to the critical nucleation phase at $t = t_c$ shown in Figure 3, at the earthquake source. This demonstrates that the critical size $2L_c$ of the nucleation zone defined here should equal twice the nucleation zone radius estimated by ELLSWORTH and BEROZA (1995), and concurrently, justifies the presumption that the critical size of the nucleation zone defined in this paper is compared directly with the seismic nucleation zone size estimated by Ellsworth and Beroza.

Figure 4 shows a plot of the logarithm of the mainshock seismic moment M_o against the logarithm of the critical slip displacement D_c for earthquake data compiled in Table 1. In Figure 4, white squares are data points taken from PAPAGEORGIOU and AKI (1983), black circles from ELLSWORTH and BEROZA (1995), and a black rectangle from IDE and TAKEO (1997). A thick straight line in the figure indicates the theoretical scaling relation:

$$M_o = 1 \times 10^{19}(D_c)^3 \tag{18}$$

where M_o and D_c are measured in Nm and m, respectively. It will be shown later that relation (18) is derived from the theoretical equation (16). It is found from Figure 4 that there is a trend for earthquake data that the seismic moment increases with an increase in the critical slip displacement, although there is a considerable fluctuation. This empirical trend is compared with the theoretical scaling relation (18), and we find that the trend agrees well with the theoretical scaling relation.

Figure 5 presents a plot of the logarithm of the seismic moment M_o against the logarithm of the breakdown stress drop $\Delta\tau_b$ for earthquake data compiled in Table 1. In contrast with Figure 4, it is found from Figure 5 that there is no such trend for the relation between M_o and $\Delta\tau_b$, indicating that the breakdown stress drop is independent of the seismic moment. This justifies the presumption that the breakdown stress drop is scale-independent. It is seen from Figure 5 that the average value for $\Delta\tau_b$ is roughly 10 MPa, though $\Delta\tau_b$ fluctuates in a range from 1 to 100 MPa. This average value will be used later to derive the scaling relation (18) from the theoretical relation (16).

In Figure 6, the logarithm of the mainshock seismic moment M_o is plotted against the logarithm of the critical size $2L_c$ of the nucleation zone for earthquakes compiled in Table 1. Figure 6 shows that the mainshock seismic moment scales with the critical size of the nucleation zone. This scaling relation has been found empirically by ELLSWORTH and BEROZA (1995). The nucleation zone size L_c has not been estimated for earthquakes analyzed by PAPAGEORGIOU and AKI (1983). Note, however, that the nucleation zone size L_c (half-length) equals the breakdown zone size X_c in the context of the present model. Contemplating that L_c is of the order of X_c, M_o for earthquakes analyzed by Papageorgiou and Aki has been overplotted in Figure 6 against $2L_c$ calculated from relation (4). White squares in Figure 6 are data points from PAPAGEORGIOU and AKI (1983), black circles from ELLSWORTH and BEROZA (1995), and a black rectangle from IDE and TAKEO (1997). A thick straight line in the figure represents the theoretical scaling relation:

$$M_o = 1 \times 10^{9}(2L_c)^3 \tag{19}$$

where M_o and L_c are measured in Nm and m, respectively. This relation is derived from equation (17), which will be shown below. Figure 6 shows good agreement between the theoretical prediction and seismological data.

We discuss here how relations (18) and (19) can be derived from the theoretical relations (16) and (17), respectively. To derive relations (18) and (19), we must assume appropriate values for the parameters: c_1, c_2, $S_{A1}/(aS)$, μ, $\overline{\Delta\tau}$, and $\Delta\tau_b^{A1}$, under the constraint that $\overline{\Delta\tau} < \Delta\tau_b^{A1}$. We assume here $c_1 = 0.5$ and $c_2 = 5 \times 10^{-5}$ in view of the values given by ABE (1975). We further assume $\mu = 30{,}000$ MPa, $\overline{\Delta\tau} = 3$ MPa, and $\Delta\tau_b^{A1} = 10$ MPa. Although we at present have no specific information as to what values both S_{A1}/S and a should take, it will not be unreasonable to assume $S_{A1}/S = 0.4$ and $a = 0.6$ (and hence $S_{A1}/(aS) = 2/3$) for the order-of-estimates. Under these assumptions, and given that $k = 3$, $\Gamma = 1$, and $\kappa = 2$, equation (16) is reduced to the relation: $M_o = 0.8 \times 10^{19} D_c^3 \approx 1 \times 10^{19} D_c^3$, and equation (17) is reduced to the relation: $M_o = 1.0 \times 10^9 (2L_c)^3$. It has thus been demonstrated that equations (18) and (19) are derived from (16) and (17), respectively. The agreement

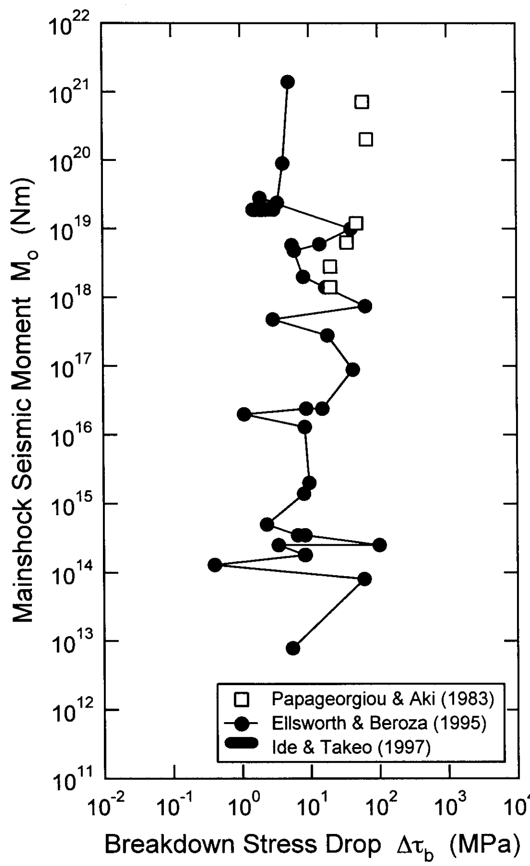

Figure 5
A plot of the seismic moment M_o against the breakdown stress drop $\Delta\tau_b$ for earthquakes listed in Table 1.

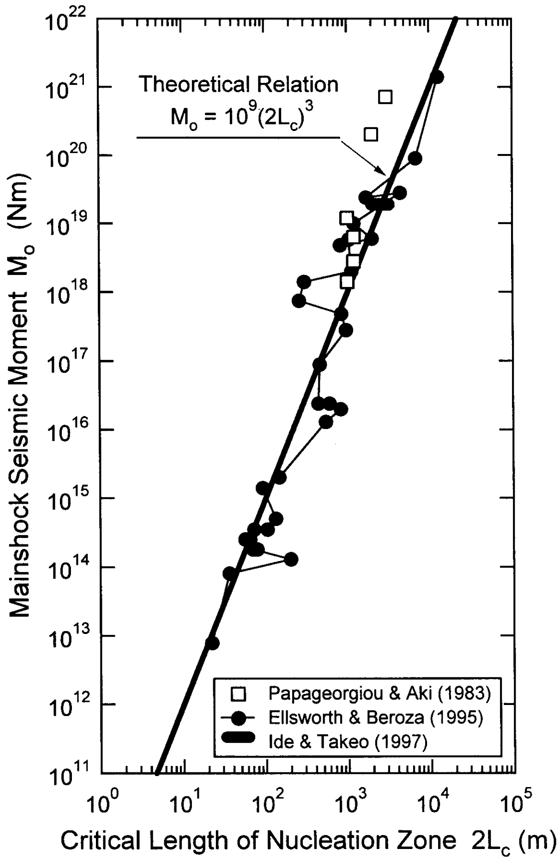

Figure 6
A plot of the seismic moment M_o against the critical size $2L_c$ of the nucleation zone for earthquakes listed in Table 1. The theoretical scaling relation denoted by a thick line in the figure is compared with seismological data.

of both equations (18) and (19) with seismological data suggests that the assumption of $S_{A1}/(aS) = 2/3$ is not unreasonable for earthquakes compiled in Table 1.

From (19) we have, for instance, $2L_c = 10$ km for earthquakes with $M_o = 10^{21}$ Nm, $2L_c = 1$ km for earthquakes with $M_o = 10^{18}$ Nm, and $2L_c = 100$ m for earthquakes with $M_o = 10^{15}$ Nm. These are estimates for the nucleation zone size for 'normal' earthquakes for which $\Delta\tau_b$ has been assumed to be 10 MPa. In reality, however, individual earthquakes have different values for $\Delta\tau_b$ (see Fig. 5), and equation (5) indeed shows that L_c depends on both D_c and $\Delta\tau_b$. Although $\Delta\tau_b$ is scale-independent, this does not mean that $\Delta\tau_b$ has a constant value for any earthquake. If a strong patch of high rupture growth resistance whose strength equals shear strength of intact rock mass is broken down, $\Delta\tau_b$ necessarily has a high value of the order of 10 to 100 MPa, depending on the individual, seismogenic environment (OHNAKA *et al.*, 1997). By contrast, if frictional slip failure occurs along a pre-existing fault of very low rupture growth resistance, $\Delta\tau_b$ may have a low value of the order of 1 MPa or less (OHNAKA and SHEN, 1999). It thus follows that $\Delta\tau_b$ can take any value in a wide range over 10^2 MPa, depending on the seismogenic environment. Hence, the proportional relationship between L_c and D_c may be severely violated by a fluctuation of $\Delta\tau_b$. Figure 7 illustrates how the proportional relation between L_c and D_c is violated by the fluctuation of $\Delta\tau_b$ for actual earthquakes. In Figure 7, white squares denote earthquake data from PAPAGEORGIOU and AKI (1983), black circles from ELLSWORTH and BEROZA (1995), and a black rectangle from IDE and TAKEO (1997). Four straight lines in Figure 7 represent theoretical scaling relations between L_c and D_c under the assumption that the parameter $\Delta\tau_b$ has constant values: 0.1, 1, 10, and 100 MPa, respectively.

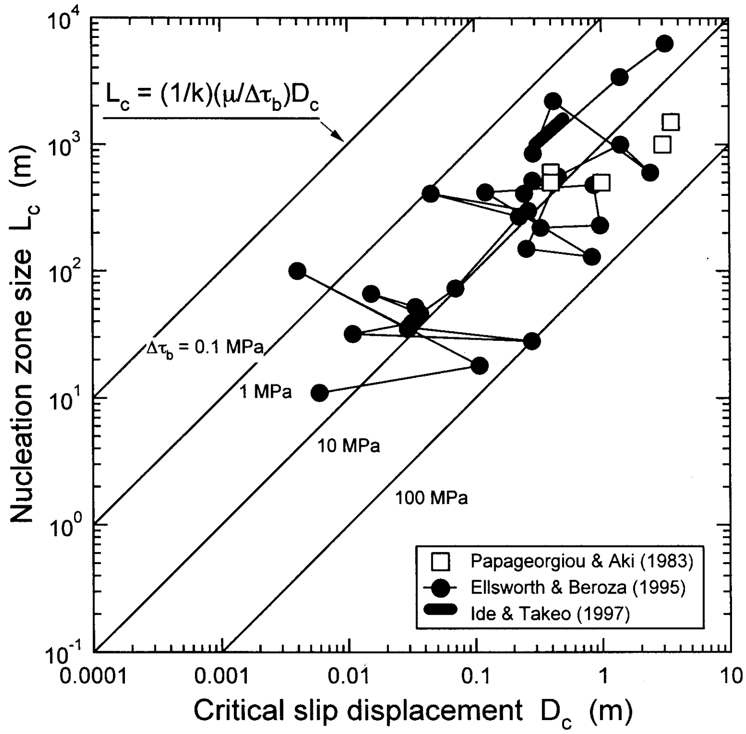

Figure 7

Scaling relation between L_c and D_c. Four straight lines in the figure represent theoretical relations between L_c and D_c with $\Delta\tau_b$ having constant values: 0.1, 1, 10, and 100 MPa, respectively. White squares, black circles, and a black rectangle denote earthquake data listed in Table 1. The scaling relation between L_c and D_c for actual earthquakes is violated by a fluctuation of $\Delta\tau_b$.

Discussion

We have proposed a fault model that the nucleation zone size and the final size of the resulting earthquake are both prescribed by a common patch (or asperity) of high rupture growth resistance on the fault. In this model, the nucleation includes the process during which the elastic strain energy sustained by a patch of high rupture growth resistance is released by failure of the patch. If the patch size is geometrically large, the resulting earthquake is large, because a geometrically larger patch of high rupture growth resistance can sustain a larger amount of the elastic strain energy, and because the failure of the geometrically larger patch results in the release of the larger amount of the elastic strain energy stored in the surrounding medium. On the other hand, a large amount of the critical slip displacement is required for the large patch of high rupture growth resistance to break down, resulting in the large size of the nucleation zone. Although the model may be overly simplified, it explains published data on the earthquake nucleation quantitatively, and allows one to provide a consistent comprehension for the earthquake nucleation in terms of the underlying physics. This suggests that an essential element of physical nature of the earthquake generation that occurs on a fault characterized by inhomogeneity has been incorporated into the model.

The present model implies that regions of a few kilometers in dimension on a seismogenic fault would slip, initially at a very slow and steady rate, and subsequently at accelerating rates, over distances ranging one meter during the nucleation of a large earthquake. Note that such slips at an initially slow and steady rate, and subsequently at accelerating rates over distances of the order of a meter can occur with or without producing micro-earthquakes, depending on the fault zone structure and ambient conditions (such as temperature). For simplicity, we take for instance a non-uniform fault which has a sizeable asperity region on which irregularities (or micro-asperities) of short wavelength components are superimposed. In this case, a slow slip failure in the asperity region necessarily brings about fracture of micro-asperities in the region, and hence it carries micro-earthquakes (immediate foreshocks). A model for this has been proposed in an earlier paper (OHNAKA, 1992), and immediate foreshock activities induced during the nucleation process have been discussed for certain earthquakes (OHNAKA, 1993; DODGE *et al.*, 1995, 1996; SHIBAZAKI and MATSU'URA, 1995). It has been demonstrated in the laboratory that micro-seismic activities are indeed induced during the slip failure nucleation.

There is a pervasive idea that natural faults have a fractal nature at all scales, and that earthquake phenomena are explained by the model of self-organized criticality. If this is the case, earthquakes are unpredictable catastrophes. However, there is commanding counter-evidence against the above pervasive idea. Firstly, it is true that natural fault surfaces exhibit self-similarity over finite bandwidths (SCHOLZ and AVILES, 1986; AVILES *et al.*, 1987; OKUBO and AKI, 1987), but they cannot be self-similar at all scales. The self-similarity is limited by the depth of seismogenic layer and fault segment size. When self-similarity of the fault surfaces is band-limited, a different fractal dimension can be calculated for each band bounded by upper- and lower-corner wavelengths (AVILES *et al.*, 1987; OKUBO and AKI, 1987), and the corner wavelength that separates the neighboring two bands is a characteristic length representing geometric irregularity of the fault surfaces. In particular, pre-existing, matured faults, such as the San Andreas Fault, have a range of characteristic length scales departed from the self-similarity (AVILES *et al.*, 1987; OKUBO and AKI, 1987; AKI, 1992, 1996), and not only the depth of seismogenic layer and fault segment size, but also barrier or asperity size on the fault, and the thickness of fault zone will be representative characteristic length scales (AKI, 1992, 1996). The earthquake generation process and its eventual size are necessarily prescribed and characterized by these characteristic length scales (SCHOLZ, 1982, 1994; ROMANOWICZ, 1992; AKI, 1992, 1996; MATSU'URA and SATO, 1997). Secondly, a large earthquake can take place on such a pre-existing, large-scale fault, only after a large amount of the elastic strain energy has been stored in the medium surrounding the fault. The elastic strain energy is accumulated with the tectonic stress buildup after the earthquake occurrence. However, a prolonged time period is needed for the strain energy to be again stored up to a critical level which has the potential to produce an ensuing large earthquake on the same fault. These indicate that the model of self-organized criticality is not applicable to large earthquakes (see also KNOPOFF, 1996).

The well-known Gutenberg-Richter power-law relation for earthquakes is a scale-independent relation, and this scale-independence implies that the underlying physics is to be found in scale-independent processes (KNOPOFF, 1996). Thus, the Gutenberg-Richter relation has been explained by the model of self-organized criticality (e.g., BAK and TANG, 1989). I wish to show that the Gutenberg-Richter relation:

$$N(M) \propto 10^{-bM} \tag{20}$$

where $N(M)$ is the cumulative number of earthquakes with magnitudes greater than M, and b is a numerical constant close to unity, can be derived from the present model, if a power-law relation holds for the asperity (or patch) size distribution (AKI, 1981):

$$N(R) \propto R^{-d} \tag{21}$$

where $N(R)$ is the cumulative number of asperities whose characteristic size is greater than R, and d is the fractal dimension. The power-law relation with d close to 2 has commonly been observed for crater size distribution, and fragment size distribution resulting from the crushing of a heterogeneous body. Since D_c is by definition the critical amount of the displacement required for fracture of an asperity with the characteristic size R in the present context, a proportional relationship holds between D_c and R; that is,

$$D_c \propto R \tag{22}$$

On the other hand, the seismic moment M_o is related to the magnitude M by (KANAMORI and ANDERSON, 1975)

$$M_o \propto 10^{1.5M}. \tag{23}$$

We thus have from (18), (21), (22), and (23)

$$N(M) \propto 10^{-(d/2)M} \tag{24}$$

and from (20) and (24)

$$b \equiv \frac{d}{2} \tag{25}$$

which agrees with the relation derived by AKI (1981). When $b \approx 1$ we have $d \approx 2$ for the asperity size distribution from (25). The present model thus leads to the conclusion that the b-value is a manifestation of the fractal dimension for the asperity size distribution arising from heterogeneities inherent in the fault zone and the seismogenic layer.

The present model indicates that "an earthquake knows from the very beginning how large it is going to be," and it has been shown that the model is indeed applicable to a class of actual earthquakes complied in Table 1. Theoretically, this implies that an earthquake is basically predictable. Practically, however, an earthquake may still not be predictable, unless it is known beforehand how non-uniformly the rupture growth resistance is distributed on the real fault (or fault network) in the seismogenic layer, and unless the ongoing nucleation can successfully be identified and monitored by any observational means. Nevertheless, the present result encourages us to pursue the prediction capability for large earthquakes.

Conclusions

Assuming a specific model of the earthquake nucleation on a non-uniform fault, a scaling relation between the seismic moment of a mainshock earthquake and the critical slip displacement, or the critical size of the nucleation zone has been derived theoretically in the framework of fracture mechanics based on a laboratory-based slip-dependent constitutive law. The present approach leads to the conclusion that the earthquake moment M_o should be related to the critical slip displacement D_c by equation (16), and the critical size $2L_c$ (L_c, half-length) of the nucleation zone by equation (17). Equations (16) and (17) are reduced to equations (18) and (19), respectively, for 'normal' earthquakes for which local stress drop $\Delta\tau_b$ in the breakdown zone behind the rupture front and the stress drop $\overline{\Delta\tau}$ averaged over the entire fault area S have been assumed to be 10 MPa and 3 MPa, respectively. The scaling relations derived well explain seismological data published. In particular, equation (19) predicts that $2L_c$ is of the order of 10 km for earthquakes with $M_o = 10^{21}$ Nm, 1 km for earthquakes with $M_o = 10^{18}$ Nm, and 100 m for earthquakes with $M_o = 10^{15}$ Nm. Since L_c depends on not only D_c but also $\Delta\tau_b$, the scaling relation between L_c and D_c may be violated by $\Delta\tau_b$. This results because $\Delta\tau_b$ potentially takes any value in a wide range from 1 to 10^2 MPa, depending on the seismogenic environment. The good agreement between the theoretical relations and seismological data suggests that a large earthquake may result from the failure of a large patch of high rupture growth resistance which is capable of sustaining an adequate amount of the elastic strain energy stored in the surrounding medium, wheareas a small earthquake may result from the breakdown of a small patch of high rupture growth resistance. The Gutenberg-Richter frequency-magnitude relation can be derived from the present model. The present result encourages one to pursue the prediction capability for large earthquakes.

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