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The Barrier Model and Strong Ground Motion

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Abstract—An overview of the most important developments in Engineering (or Strong Motion) Seismology is presented alongside Professor Keiiti Aki's contributions, who is one of the founders of this field. The mechanics of earthquake rupture are discussed with due emphasis on the various physical phenomena. The presentation is made in a tutorial manner, borrowing freely from Keiiti Aki's papers, and endeavoring to emulate his unique style of clarity, simplicity and synthetic ability.

Key words: Barrier, specific barrier model, earthquake source, strong ground motion.

Introduction

The work of Professor Keiiti Aki in the discipline of Seismology is unprecedented in its breadth, depth and originality. His contributions span virtually the entire frequency range (i.e., normal modes, surface waves, body waves, strong motion, seismic coda, harmonic tremor). In the present article, we aim to survey his contributions to the field of *Engineering* (or *Strong Motion*) *Seismology*. The presentation is made in a tutorial manner, borrowing freely from Keiiti Aki's papers, and trying to emulate his unique style of clarity, simplicity and synthetic ability.

Earthquake Seismology deals with the study of the generation, propagation, and recording of elastic waves in the earth, and of the physical processes occurring at the source of an earthquake. By the term *Engineering* (or *Strong-Motion*) *Seismology* we mean that part of seismology dealing with earthquakes close enough to the causative source where ground motion is strong enough to pose a threat to engineering structures. The principle problem of engineering seismology is the estimation of strength, frequency content, duration and spatial variability of the most destructive (in terms of its effects on a particular structure) ground-shaking that is likely to occur at a site. This estimation should be based on the physics of the generation and propagation of seismic waves. According to AKI (1980b), the ultimate objective of current research efforts is to "*compute seismic motion expected at a specific site of an*

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engineering structure when the fault mapped by geologists breaks.'' In the early days of earthquake studies, before the development of sensitive seismographs, all seismology was of necessity ''strong motion seismology,'' as this is evident in the work of ROBERT MALLET (1810–1881), who established the basis of observational field seismology in his detailed study of the destructive Neapolitan earthquake of 16 December 1857 in Italy (MALLET, 1862).

Tectonic Processes and the Mechanics of Earthquake Rupture

On the basis of overwhelming evidence it is now widely accepted that earthquakes are caused by the dynamic spreading of shear rupture on a fault plane (AKI, 1972a). This model of earthquake source is the ''*fault model*,'' initially proposed by REID (1910) in his ''*elastic rebound theory*.'' On the basis of deformations observed on the surface or measured by geodetic methods and seismic data obtained at local and distant stations, Reid proposed that the San Francisco earthquake of 1906 was the release of strain energy stored in the vicinity of the San Andreas fault by a slip along the fault. This theory stirred considerable controversy. AKI (1979b, 1988) gives an historical account of the controversies which the fault model has survived from its early days until it was firmly established in the mid-60s, when a quantitative test of the model became possible with the use of the global network of calibrated stations, the advent of large-scale digital computers and the development of an appropriate mathematical framework, the so-called ''*dislocation theory*,'' which relates the observed seismogram with the slip motion across a fault plane. Furthermore, the success of the theory of plate tectonics provides the strongest support for the fault model. The ''*theory of plate tectonics*,'' which describes the kinematics of the upper layer of the earth, was implicit in Reid's elastic rebound theory. It is based upon the assumption that the upper part of the crust, called the lithosphere, is decidedly more rigid than the underlying asthenosphere. The lithosphere is composed of a number of plates which move relative to the mantle and to each other. Indeed, the consistency of plate motions with the direction and amount of slip during earthquakes everywhere is remarkable.

Kinematics of Fault Rupture

We have already pointed out above that earthquake ground motion results from unstable slip accompanying a sudden drop in shear stress on a geologic fault. Therefore, an earthquake is primarily a mechanical process. During the short span of this process, the earth, except in the earthquake source, behaves as an elastic body. Consequently, seismic waves are linear elastic waves propagating in a very complex, nonhomogeneous, dissipative, prestressed medium (the earth is in a prestressed state due to internal deformation and its own gravitational field). Therefore, the basic analytical tool for studying earthquakes is *classical elastodynamic theory* (e.g., GURTIN, 1972; ACHENBACH, 1973; ERINGEN and SUHUBI, 1975; MIKLOWITZ, 1978) supplemented with *fracture mechanics* (e.g., FREUND, 1990; KOSTROV and DAS, 1988; BROBERG, 1999).

In order to express mathematically the ground motion induced by an earthquake, we need a formula for the displacement—at a general point in space and time—in terms of the physical parameters that originated the motion. This formula is provided by the *elastodynamic representation theorem*. As noted by AKI and RICHARDS (1980), the representation theorem is a bookkeeping device by which the displacement from realistic source models is synthesized from the displacement produced by the simplest of sources—namely, the unidirectional unit impulse, which is localized precisely both in space and time. The displacement response due to such a singular source is referred to as *Green's function*.

A mathematical statement of the representation theorem for the *faulting source* (BURRIDGE and KNOPOFF, 1964)

$$u_n(\mathbf{x}, t) = \int_{-\infty}^{+\infty} d\tau \iint_{\Sigma} \Delta u_i(\boldsymbol{\xi}, \tau) c_{ijpq} \nu_j \frac{\partial}{\partial \xi_q} G_{np}(\mathbf{x}, t - \tau; \boldsymbol{\xi}, 0)\, d\Sigma(\boldsymbol{\xi}) \tag{1}$$

where c_{ijpq} are the components of the elasticity tensor, $G_{np}(\mathbf{x}, t-\tau; \boldsymbol{\xi}, 0)$ is the Green's tensor which represents the n-th component of displacement at $(\mathbf{x}, t-\tau)$ due to an impulsive concentrated unit force acting in the p-th direction at $(\boldsymbol{\xi}, 0)$, $\Delta u_i(\boldsymbol{\xi}, \tau) = u_i(\boldsymbol{\xi}, \tau)\mid_{\Sigma^+} - u_i(\boldsymbol{\xi}, \tau)\mid_{\Sigma^-}$ is the displacement discontinuity across Σ (i.e., the slip on the fault plane) and $\boldsymbol{\nu}$ is a unit vector normal on surface Σ and pointing from Σ^- to Σ^+.

Starting with eq. (1) and making use of the properties of the Dirac delta function it may be demonstrated (e.g., AKI and RICHARDS, 1980) that a point shear dislocation is equivalent to a *double-couple*, i.e., the double-couple is the body force which would have to be applied in the absence of the fault to produce the same radiation as a given point dislocation. The first person to obtain the double-couple equivalence for an effective point source of slip was VVEDENSKAYA (1956).

In the far-field and for periods with wavelengths much larger than the source size, the fault appears as a point dislocation. The scalar value of the moment of one of the couples in the double-couple representation of the point dislocation is the *seismic moment* M_0. Assuming an *average slip* $\overline{\Delta u}$ over the fault plane then

$$M_0 \equiv \mu \overline{\Delta u} A = \mu \times \text{average slip} \times \text{fault area} \tag{2}$$

where μ is the rigidity (i.e., shear modulus) of the lithosphere and A is the area of the fault plane which slipped.

The first precise determination of the seismic moment was accomplished by AKI (1966) for the 1964 Niigata earthquake using long-period Love waves observed by

WWSSN. It is the most important static parameter of the source; the value of which at the end of the rupture process measures the permanent inelastic strain produced by the event, and thus it is the simplest way to measure the strength or size of an earthquake. It can be reliably inferred from seismic observations such as displacement spectra of long-period surface waves, free oscillations of the earth or directly from field observations, as suggested by eq. (2) and thus serves as a direct link between seismological, geological and geodetic observations.

Kinematic Models

One of the fundamental problems of seismology is the inference of the earthquake faulting process from the analysis of seismic waves radiated from the source. In order to render such an *inverse problem* tractable, physically reasonable assumptions must be made about the rupture process and its evolution in time. Thus, as a first approximation to the solution of the above problem, dislocation models were introduced as simple kinematic descriptions of the evolution of faulting with time [for a discussion of the meaning of the terms "*kinematic*" and "*dynamic*" see AKI and RICHARDS (1980), BOX 5.3]. In general, in kinematic models the faulting process is represented in terms of the *slip* (or *source*) *function*, the form of which usually is chosen intuitively, without rigorous analysis of the time-dependent stresses acting on the area. In particular, dislocation models represent simple geometrical idealizations of actual faulting in the earth, and as such they are extremely simplified, averaged out versions of the rupture process. Analyses and inversion studies using dislocation models have provided significant insights into the effects of *fault finiteness* and *fault geometry* on the radiation of elastic wave energy.

One of the first and most widely used dislocation models is *Haskell's model* (HASKELL, 1964, 1966, 1969) which represents faulting on a rectangular plane of length L and width W. According to this model, rupture initiates at one end of the fault with the appearance of a dislocation line segment spanning the width of the fault. This rupture front propagates along the length of the fault with velocity V. At each point of the fault plane, slippage is initiated when the rupture front reaches the point. The time that it takes for slip at a point to reach the final value Δu_0 is called the *rise time* τ. MADARIAGA (1978) calculated exactly the elastic waves radiated by Haskell's model in the far-and near-field.

When we take a closer look at the nucleation of the rupture process, we realize that the unidirectional propagation of rupture in Haskell's rectangular fault model is an oversimplification of physical reality. For this reason, other dislocation models were proposed that allow rupture to initiate at a point (rather than simultaneously everywhere along a line segment) and then spread out radially (rather than propagate in a single direction) at a uniform velocity until it covers an arbitrary two-dimensional surface on the fault plane (SAVAGE 1966, MOLNAR *et al.*, 1973).

Focusing our attention to the near-field, it is fair to say that modern quantitative analysis of strong ground motion observations started with the now famous Station No. 2 record (HOUSNER and TRIFUNAC, 1967) obtained from the 1966 Parkfield earthquake at a distance of only 80 m from the fault break. AKI (1968) and HASKELL (1969) demonstrated that the observed transverse component of displacement of the above ground motion record, which exhibited a simple impulsive form, was precisely what is expected for a right-lateral strike-slip rupture propagating from northwest to southeast. They used the five parameter $(L, W, \Delta u_0, V, \tau)$ *Haskell model* described above. Within this model, the most important parameters affecting the level of strong ground motion are τ and D. Of these two parameters, the rise time τ is the one that is most difficult to determine. It is relevant to point out that the *slip velocity* $\Delta\dot{u}$, an estimate of which may be obtained by the ratio $(\Delta u_0/\tau)$, was investigated by AKI (1983) and was found to exhibit a large variation. AKI (1983) attributed this significant variation of the value of the average slip velocity to the inability to resolve (i.e., infer accurate estimates) of the short rise times.

In the beginning, most simulation methods, including those presented in the pioneering studies of AKI (1968) and HASKELL (1969), used Green's function for an unbounded homogeneous medium. Subsequently, in order to make the propagation medium more realistic, the free surface effect and the effect of sedimentary layers were included in the simulations (e.g., BOUCHON, 1979; for a review of modeling studies extending to the beginning of the 1980s, see AKI, 1982, 1983).

It gradually became evident, though, that the assumption of uniform slip over the fault plane (as required by Haskell's model) was not adequate to simulate simultaneously the motions of an earthquake event recorded at more than one recording station. Therefore, the original source model had to be modified in the following two ways:

The first important modification was to allow the slip function to vary from place to place on the fault plane i.e., the fault had to be subdivided into subfaults, each with a different slip vector. The first attempt to allow non-uniform slip was made by TRIFUNAC and UDWADIA (1974) for the 1966 Parkfield earthquake. It was the study of the 1979 Imperial Valley earthquake, however, that marked the beginning of systematic waveform inversions (HARTZELL and HELMBERGER, 1982; OLSEN and APSEL, 1982; HARTZELL and HEATON, 1983; ARCHULETA, 1984). Similar inversion studies followed for other California earthquakes, and since then inversions of this type have become routine for well recorded earthquake events.

The above *deterministic kinematic modeling* approach has been limited to a rather smooth picture obtainable from the frequency range lower than about 1 Hz (i.e., spatial resolution $\sim$5 km). Despite this limitation, the above inversion studies demonstrated that the spatial and temporal behavior of earthquake faults is rich in *complexity* and *heterogeneity*. The slip distributions inferred by all the above inversion studies exhibit two important features: (i) the existence of relatively localized areas of large static slip, and (ii) short dislocation rise time τ. The shortness of the dislocation rise time also has been confirmed recently by eyewitness observations of the coseismic fault movement during the 1990 Luzon, Phillipines, earthquake (M_s 7.8) as reported by YOMOGIDA and NAKATA (1994) (see also WALLACE, 1984, for a pertinent eyewitness account regarding the 1983 Borah Peak, Idaho, earthquake).

The short rise time was recognized earlier by AKI (1968, 1979) for the 1966 Parkfield earthquake; he introduced the *barrier model* to explain the observed shortness of rise time. According to the barrier model, the shortness of rise time is a consequence of the strong segmentation of the fault plane, so that rupture involves sequences of crack-like propagation over a small patch, arrest at its borders, renucleation on a neighboring patch, etc. The mathematical description of such a process is provided by the ''*specific barrier model*'' proposed by PAPAGEORGIOU and AKI (1982) and PAPAGEORGIOU and AKI (1983a) to be described below. An alternative explanation for the shortness of rise time has been proposed by HEATON (1990) who attributes it to a ''*self-healing*'' process due to a particular friction law with strong velocity dependence at high slip rate. However the ''self-healing'' hypothesis was found by BEROZA and MIKUMO (1996), IDE and TAKEO (1997) and DAY *et al.* (1998) unnecessary to explain earthquake kinematics. [The interested reader may find informative discussions related to the ''self-healing'' hypothesis in BIZZARI *et al.* (2001) and GUATTERI and SPUDICH (2000). In a related study ANDREWS and BEN-ZION (1997) investigated the conditions for and properties of dynamic ruptures consisting of narrow ''*propagating slip pulses*'' associated with variations of normal stress].

The other important modification in the source model originally used by AKI (1968) and HASKELL (1969) to simulate strong ground motion was the introduction of stochastic elements, thus reducing the number of parameters needed to describe the details of the slip function. We will elaborate on this in our discussion of the ''specific barrier model.''

Rupture Dynamics on a Heterogeneous Fault

In the previous section we pointed out the usefulness of dislocation models in earthquake seismology. However, such models are associated with very strong singularities which are physically unacceptable. Specifically, strong stress singularities of the type r^{-1} are found on the fault surface around the edges of the fault. These singularities, which are a consequence of the assumption of constant slip over the entire fault plane, are so strong that an infinite strain energy change is predicted independently of any source parameters. Since faulting is a failure along a fault plane, we expect that the fault plane, once ruptured, cannot sustain stress beyond the frictional stress.

Furthermore, from a purely continuum mechanical point of view, the constant slip is inadmissible because near the borders of the dislocation model there is interpenetration of matter. Therefore, the elastic solutions are not valid inside a core region around the edges of the dislocation model. The withdrawal from these physical inconsistencies is to eliminate these singularities by smoothing the slip near the rupture front, and thus rendering a slip function that is not only kinematically satisfactory for shear faults, but also compatible with a physically plausible stress distribution on the fault plane. This requirement forms the basis of crack models. The physics of dynamic crack growth is the domain of study of *dynamic fracture mechanics* (MADARIAGA, 1978; MADARIAGA, 1983a; AKI and RICHARDS, 1980; FREUND, 1990; DMOWSKA and RICE, 1986).

We consider a 2-D crack model in a homogeneous isotropic elastic half space as shown in Figure 1. A plane crack representing the fault lies on the plane (x, y) with its rupture front parallel to the y axis. The position of the rupture front as a function of time is described by the function $x = l(t)$. The material is assumed to be elastic everywhere (even at the crack tip) and the applied external stress (*tectonic stress*) $\sigma^0_{zj}(x)$ (where $j = x$ or y) is assumed to be uniform. Inside the crack, after the passage of the rupture the stress drops to the *dynamic friction* $\sigma^f_{zj}(x)$ and the difference $\Delta\sigma(x) = \sigma^0_{zj}(x) - \sigma^f_{zj}(x)$ is the *stress drop*.

Perhaps the most important results relevant to our discussion here have been obtained by KOSTROV and NITIKIN (1970) and FREUND (1972, 1979) who demonstrated that the solutions of these crack problems have a number of universal features which are independent of the details of the rupture front motion $l(t)$ and the

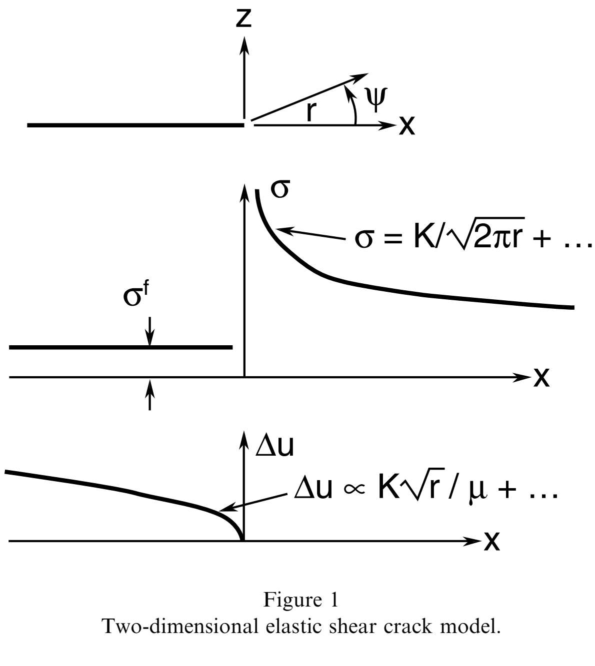

Figure 1
Two-dimensional elastic shear crack model.

stress drop distribution $\Delta\sigma(x)$. Specifically, the stress and velocity fields present characteristic inverse square-root singularities in the vicinity of the crack tip. The singularity of the stress field at the crack tip is a result of the assumption that the material remains elastic even in the immediate vicinity of the surface front. The inverse square-root singularity appears because this is the only way the elastic field can ensure a finite energy flow into the rupture front. This energy, referred to as *Griffith specific fracture energy G*, is used to create a new fault surface and is spent in the nonlinear processes taking place in the breakdown zone that exists in the vicinity of the crack tip. Thus, *linear elastic fracture mechanics* is only a large-scale approximation of the rupture process, according to which all the inelastic processes at the breakdown zone are characterized by one parameter, the *dynamic stress intensity factor K*.

To describe the fracture at an earthquake source using the crack model, two important pieces of information are necessary: (1) the initial distribution of stress on the fault surface before the earthquake, and (2) the constitutive law governing the fracture propagation, or as this law is referred to in linear elastic fracture mechanics, the *fracture criterion*.

Following ANDREWS (1978), the stress applied on the fault zone can be separated into two terms: (i) the *self-stress* that arises from irregular slip and whose sources are therefore local, and (ii) the *ambient tectonic stress* that has distant sources such as slip on distant faults, fault creep at depth and viscous drag at the base of the lithosphere. The tectonic stress arising from distant sources varies smoothly over the fault plane and has significant components at wavelengths of the order of the depth of the seismogenic region. On the contrary, the self-stress must vary strongly across the fault surface in order to explain the stationary occurrence of numerous small earthquakes.

In fracture mechanics, there are two different types of fracture criteria: (i) *Griffith's criterion* and (ii) *Irwin's criterion*, both of which have been used to describe crack growth in earthquake seismology (e.g., Box 15.2 in AKI and RICHARDS, 1980). Under both of the above criteria, we have the stress singularity in front of the tip of the advancing crack. In reality, however, no real material can sustain infinite stresses because as KNOPOFF (1981) points out, paraphrasing Spinoza, "*nature abhors an infinity*." Thus, a zone of cohesive forces at the crack tip has been proposed to remove the stress singularity (BARENBLAT, 1959). This zone is used to model the breakdown process, small-scale yielding, microcrack formation, etc.; that takes place over a zone of finite area at the circumference of the crack (for a thorough review of the subject see e.g., RICE, 1980).

The most realistic cohesive zone model for a number of geophysical applications is the so-called *slip weakening model*. This model was used for the first time in a seismological context by IDA (1972, 1973) and its consequences on fault rupture evolution were investigated by IDA and AKI (1972) (see also ANDREWS, 1976a, 1976b, 1985, 1994). According to this model, in the simplest case slippage is modeled as rate insensitive. The strength of the fault zone reaches a peak value σ_u (also referred to as "*yield-stress*") which corresponds to the onset of slipping for fresh fractures or it is preceded by slip at lower stresses for pre-existing faults. The stress to maintain slippage reduces as the amount of slip increases up to a critical amount D (referred to as the "*characteristic weakening slip*"), above which the stress to maintain slippage remains constant, equal to the *dynamic friction* σ_f. Such a constitutive law of the fault gauge is depicted in Figure 2. The crosshatched area shown in Figure 2 represents the energy per unit area of crack absorbed at the crack tip by the breakdown process (as noted above $G =$ *Griffith's specific fracture energy*). The region of the crack near the tip where the applied stress is greater than the frictional stress is the *cohesive* (or *break-down*) *zone d*. The *average value* of the (*cohesive*) *stress*

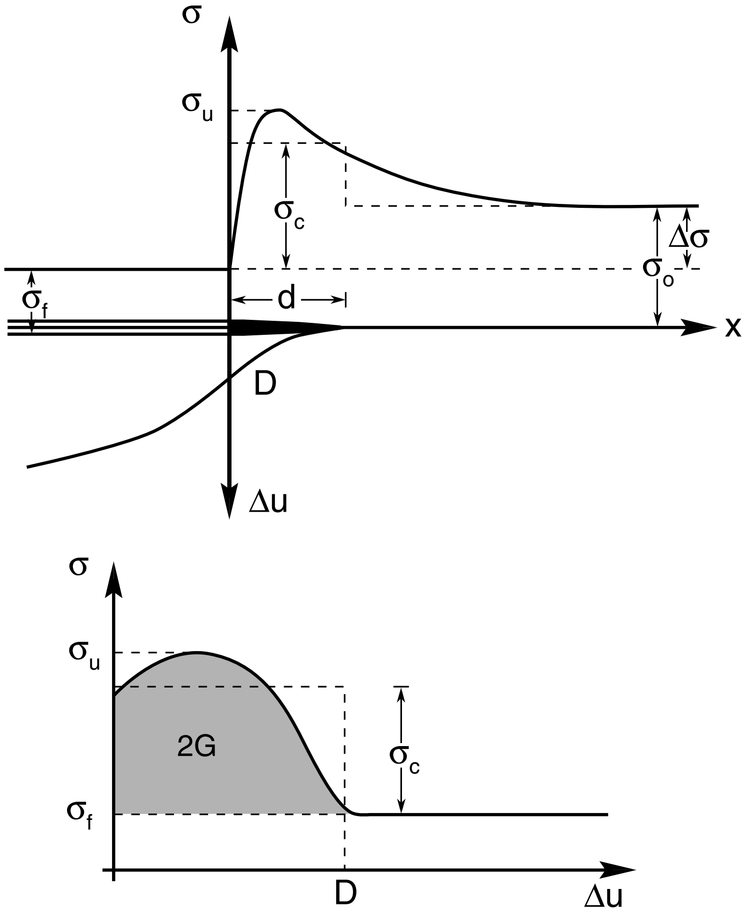

Figure 2
Constitutive law of the "slip-weakening" model.

distributed over the break-down zone is denoted by σ_c. The differences $(\sigma_u - \sigma_0)$ and $(\sigma_0 - \sigma_f)$ are referred to as the "*strength excess*" and "*stress drop*", respectively, and the ratio $S = (\sigma_u - \sigma_0)/(\sigma_0 - \sigma_f)$ is referred to as the "*strength parameter*." Due to the finite strength of the material which is depicted by the constitutive law, the stress singularity and the slip distribution shown in Figure 1 have been replaced by a continuous stress distribution and a smooth slip distribution respectively, shown in Figure 2. [A competing constitutive law is the "*rate- and state-dependent friction law*" (DIETERICH, 1979, 1992; RUINA, 1983; PERRIN *et al.*, 1995). In contrast to the "slip-weakening" constitutive law which assumes that friction (or total traction) is a function of the fault slip only, the "rate- and state-dependent" constitutive law implies that the friction is a function of slip velocity and state variables (BIZZARI *et al.*, 2001)].

If the geometry and material properties of the fault zone are homogeneous and the tectonic stress uniform, then the crack, once it starts moving dynamically, will never stop. Stating this differently, without strength or stress heterogeneity, earthquake fault rupture would never stop. The only way that shear fracture may remain limited in space is that there be strong patches/ligaments on the fault surface to stop the rupture (e.g., HUSSEINI *et al.*, 1975; MADARIAGA, 1979), or that the rupture would break into previously relaxed areas of the fault (i.e., the crack would "*run out of gas*" AKI, 1988; AKI and RICHARDS, 1980). Thus heterogeneity is a fundamental part of the earthquake process. The observed complexity of earthquake phenomena, which was extensively documented in the past two decades (for reviews see for example AKI *et al.*, 1977; AKI, 1979a; AKI, 1980a; PAPAGEORGIOU and AKI, 1982), is a direct consequence of the heterogeneity of the physical properties of the fault zone. Once the earthquake starts, the growth and arrest of fracture is controlled in a very complex way by the distribution of stress and strength.

In order to describe heterogeneities on the fault plane and gain conceptual understanding of the complexity of rupture process, the terms "*asperities*" and "*barriers*" have been used in the published literature. "*Barrier*" is a strong patch on the fault plane which remains unbroken after the passage of the rupture front. The presence of barriers on the fault surface explains aftershocks as release of stress concentration through static fatigue. "*Asperity*" on the other hand, is a strong patch surrounded by a region where stress has been released by preslips and aftershocks (e.g., AKI, 1979a; AKI, 1984; PAPAGEORGIOU and AKI, 1982).

In the course of dynamic faulting, seismic radiation (and in particular high-frequency radiation) is controlled by the slip velocity field. In view of the fact that the most significant feature of slip velocity is the singularity at the rupture front, it follows naturally that the dominating part of seismic radiation is emitted by the rupture front. As the rupture front moves smoothly, it radiates continuously, generating the low-frequency part of the field. High frequency waves are produced by jumps in the rupture velocity and/or abrupt changes in the stress intensity factor. Accelerograms are dominated by these impulsive waves. Therefore, the radiation of high frequency waves is controlled by the motion of the rupture front. Because the rupture front is only a geometrical definition, it does not have "*inertia*" and hence its speed can change abruptly when the rupture reaches differing stress or frictional regimes which are controlled by the fault heterogeneities. As MADARIAGA (1983a) demonstrated there are two ways to produce jumps in the particle velocity radiation, and consequently strong acceleration pulses: (1) the rupture front stumbles on a barrier where the strength or rupture resistance increases suddenly, the rupture velocity changes abruptly and a strong wave (step change in particle velocity) is generated; (2) the rupture front encounters an asperity due to a previously unbroken ligament on the fault. Regardless of whether the rupture velocity changes or not, this generates a step of particle velocity. Therefore, barrier and asperities are the source of high frequency waves. The wave front discontinuities created in this fashion are evaluated by *asymptotic methods* and may be propagated away from the source by *ray theoretical methods* (MADARIAGA, 1977; MADARIAGA, 1983a; MADARIAGA, 1983b; MADARIAGA and BERNARD, 1985; ACHENBACH and HARRIS, 1978; ACHENBACH and HARRIS, 1987; BERNARD and MADARIAGA, 1984a; BERNARD and MADARIAGA, 1984b; SPUDICH and FRAZIER, 1984; ACHENBACH *et al.*, 1982).

The above analytical results were confirmed observationally by SPUDICH and CRANSWICK (1984) who analyzed motions of the 1979 Imperial Valley earthquake recorded at the 5-element El Centro differential array, and by ZENG *et al.* (1993b), who by inversion mapped on the fault plane the sources of high frequency radiation of the 1989 Loma Prieta earthquake. Since both the state-of-stress on a fault and the strength of the fault material may be quite heterogeneous over a real fault surface, it is reasonable to expect that the advancement of the rupture front may be uneven, jumping around unyielding barriers—as was clearly demonstrated by the numerical studies of DAS and AKI (1977a, 1977b), MIKUMO and MIYATAKE (1978) and DAY (1982) and was verified by inversion studies such as that of BEROZA and SPUDICH (1988) and OLSEN *et al.* (1997)—and resulting in a pattern of broken and unbroken regions such as that of the 1966 Parkfield earthquake suggested by AKI (1979a) and shown in Figure 3. In particular, Figure 3 was obtained as follows: The hypocenters of the Parkfield aftershocks were projected on the fault plane. According to the barrier model few aftershocks are expected over a section of the fault that slipped smoothly. On the contrary, areas that act as barriers to the rupture experience little slip and are stress concentrators. This induced stress increase combined with static fatigue causes a sequence of aftershocks. With this reasoning, AKI (1979a) drew boundaries between regions with no aftershocks (slipped sections, indicated as white in Figure 3) and regions with aftershocks (unbroken barriers with little slip, indicated as gray in Figure 3). This *complementarity relation* between fault slip and aftershocks was further verified by MENDOZA and HARTZELL (1988) and TAKEO (1988).

However, the distribution of stress and strength on real faults is unknown, and thus it is impossible to describe deterministically the details of the rupture process which, as we argued above, are responsible for the generation of high frequency

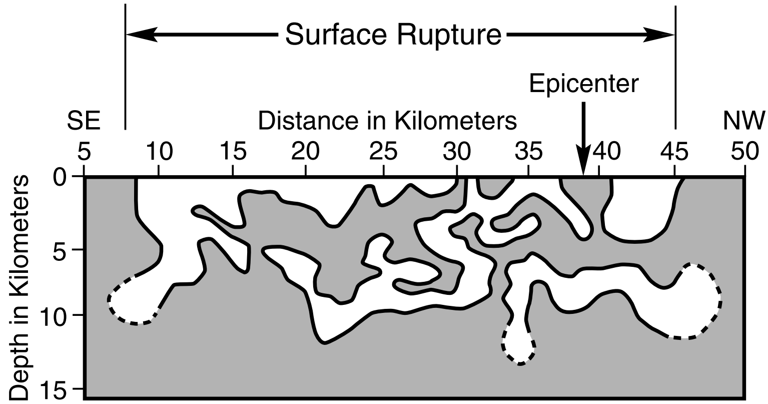

Figure 3
Heterogeneous rupture during the 1966 Parkfield, California, earthquake (modified from AKI, 1979a).

waves that dominate accelerograms. Hence, beginning with the works of HASKELL (1966) and AKI (1967), investigators have tried to introduce stochastic elements in the description of the source, and several attempts have been made to introduce a *hybrid* of deterministic and stochastic models, in which gross features of rupture propagation are specified deterministically while the details of the rupture process are described by a stochastic model specified by a small number of parameters (BOORE and JOYNER, 1978; HANKS, 1979; ANDREWS, 1981; IZUTANI, 1981; PAPAGEORGIOU and AKI, 1983a; BOATWRIGHT, 1982; BOATWRIGHT, 1988; KOYAMA, 1985).

In order to conceptualize rupture on a heterogeneous fault plane and provide the framework for its mathematical description, let us consider an idealized geometry consisting of a rectangular area containing small circles. As AKI (1982, 1983, 1984) points out, there are two opposing views of how slip can take place over this fault plane. In one of them, the circles represent strong ligaments resisting fracture, while the regions between circles have already slipped aseismically. Once the rupture starts, the ligaments will break in a more or less independent manner and will generate the high frequency waves that are observed in accelerograms. After the rupture, the entire area of the fault is broken, and the residual stress will be uniform over the fault plane, equal to the static friction. This viewpoint, that is referred to as the "*asperity model*," was adopted by KANAMORI and STEWART (1978) in interpreting the teleseismic *P* waveforms of the Guatemala earthquake of 1976, and was described by RUDNICKI and KANAMORI (1981) [For numerical studies of the rupture of an asperity see DAS and KOSTROV (1983, 1986) and FUKUYAMA and MADARIAGA (1998)].

In the other view, the circle represents a crack where a slip occurs during the fault rupture, but the region between cracks remains unbroken after the rupture. The possibility of such segmented ruptures was demonstrated by DAS and AKI (1977a) using numerical experiments. A rupture front may be stopped by a barrier, but elastic waves generated by the slip can break the fault plane ahead of the barrier in the case of shear crack. Thereafter, the rupture can propagate over the entire fault plane leaving unbroken barriers behind. The resultant irregular slip can explain observed accelerograms. This model is called the "*barrier model*" by AKI *et al.* (1977) and is supported by numerous examples of fault segmentation mapped by geologists (AKI, 1980a). In contrast to the asperity model, the residual stress over the fault plane is not uniform after the rupture. Excess stress will be induced at the unbroken barriers and may become the cause of aftershocks.

A real fault plane may contain a mixture of strong ligaments that during earthquake rupture may behave as asperities or barriers. In fact, in the numerical experiments of DAS and AKI (1977a), the following three situations were found when a crack tip passes a barrier, depending on the initial stress: (1) The barrier is broken immediately; (2) the crack-tip proceeds beyond the barrier, leaving behind an unbroken barrier; and (3) the barrier is not broken at the initial passage of the crack tip, but eventually breaks due to a subsequent dynamic increase in stress, effectively behaving as an asperity.

Case (2) above was the basis of the "specific barrier model" proposed and developed by PAPAGEORGIOU and AKI (1983a, 1983b) to model and interpret strong motion acceleration spectra of major California earthquakes.

The "*specific barrier model*" consists of circular cracks of equal diameter $2\rho_0$, filling up a rectangular fault of length L and width W, as shown in Figure 4. As the rupture front sweeps the fault plane with the "*sweeping velocity*" V, a stress drop $\Delta\sigma$ (referred to as the "*local stress drop*") takes place in each crack starting from its center and spreading with a "*spreading velocity*" v.

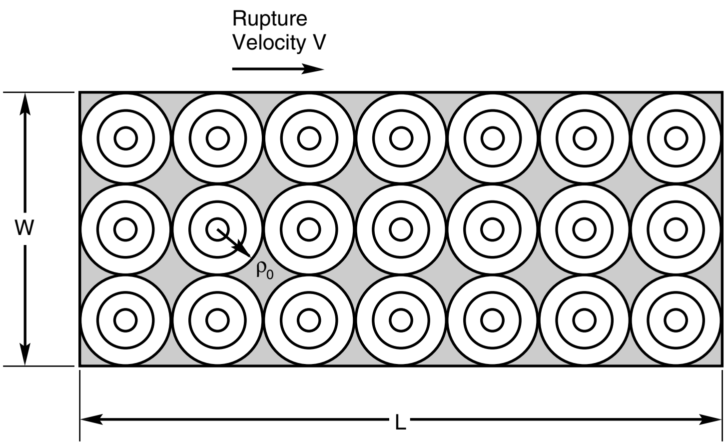

Figure 4
Specific Barrier Model (PAPAGEORGIOU and AKI, 1983a).

The slip stops abruptly when the crack radius reaches ρ_0. SATO and HIRASAWA (1973) proposed a kinematic (dislocation) model simulating the rupture of such a circular crack. The compact closed form expression of the far-field displacement waveform that they obtained has all the essential features of the waveform obtained from the more realistic numerical models studied by MADARIAGA (1976). Furthermore, PAPAGEORGIOU and AKI (1983a) demonstrated that the *stopping phases* (i.e., the phases radiated when the rupture front is arrested abruptly by the barrier that exists at the periphery of the crack) of the SATO and HIRASAWA (1973) model simulate to within a multiplicative factor the exact results obtained by MADARIAGA (1977). BERNARD and MADARIAGA (1984a) showed that these multiplicative factors may be calculated by approximate consideration of the healing waves on the fault. Finally, SATO (1994) investigated the effect that the finite deceleration time, at the final stage of rupture of the circular crack model, has on the stopping phases. Focusing again our attention on Figure 4, we point out that the region between circular cracks represents barriers left unbroken after the passage of the rupture front. The ruptures of individual cracks are statistically assumed to take place *independently*. Thus, the "specific barrier model" is a hybrid of a deterministic and stochastic one and is described by five parameters, namely L, W, $V\,(=v)$, $2\rho_0$ and $\Delta\sigma$.

PAPAGEORGIOU and AKI (1983b) chose major California earthquakes for which L, W, V and the maximum slip $\Delta u_{\max}$ were already known from observations other than strong ground acceleration data. Then, they estimated the barrier interval $2\rho_0$ and the local stress drop $\Delta\sigma$ by fitting the acceleration power spectra predicted by the model to the observed ones. In this process, they had to introduce a sixth parameter to define a cut-off frequency (called $f_{\max}$ by HANKS 1982) beyond which the acceleration spectrum decays sharply with increasing frequency.

PAPAGEORGIOU and AKI (1983a, 1983b) attributed $f_{\max}$ to the smoothing effect in the break-down zone at the crack tip and, following IDA (1973) and AKI (1979a), related it to the size d of the break-down zone using the following relation

$$f_{\max} = \frac{v}{d}. \tag{3}$$

The above smoothing effect of the presence of the break-down zone was confirmed numerically by GABRIEL and CAMPILLO (1989) and FUJIWARA and IRIKURA (1991), and analytically by ACHENBACH and HARRIS (1978) and SATO (1994).

Furthermore, the parameters G, σ_c, d and D that were introduced in connection with Figure 2 and which characterize the fracture strength of a fault zone, may be determined from the observed acceleration power spectrum and eq. (3) as elaborated by PAPAGEORGIOU and AKI (1983a). The values of the above parameters thus inferred have the following physical interpretation: G represents the fracture energy of the barrier which is necessary to arrest the propagation of rupture of a subevent; d represents the length of the inelastic zone over which rupture is arrested; σ_c is the average cohesive force, distributed on the inelastic zone; D represents the slip that occurs in the break-down zone, which is required to break the bond completely. The above parameters determine the coseismic and long-term behavior of faults (e.g., CAO and AKI, 1984). For instance, it has been demonstrated theoretically that the critical slip displacement D plays a key role in determining the rupture nucleation dimension (DIETERICH, 1992), precursory deformation (YAMASHITA and OHNAKA, 1992), and high-frequency radiation of acceleration (IDA, 1973; PAPAGEORGIOU and AKI, 1983a; OHNAKA and YAMASHITA, 1989).

In the process of inferring the above parameters, PAPAGEORGIOU and AKI (1983b) corrected the observed acceleration spectra for propagation path effects, however they did not consider any correction for the local site effect of recording stations, inasmuch as they did not find obvious differences between rock and soil sites within their data set (PAPAGEORGIOU, 1988).

Later, however, a study of the site effect made by PHILLIPS and AKI (1986) at most of the stations of the USGS Central California network using the *coda method* (AKI, 1969; AKI and CHOUET, 1975) revealed a strong frequency-dependent site amplification for the average of all the network stations relative to the average of stations located on granitic rocks. Assuming that the latter site may be approximated by a homogeneous half space, the average of amplification factors for all stations was adopted in correcting the acceleration power spectra for the recording site effect (AKI and PAPAGEORGIOU, 1989). The revised source parameters (along with the corresponding ones of the Loma Prieta, 1989, earthquake which were inferred by CHIN and AKI (1991), using the "specific barrier model") are shown in Figure 5 as a function of Magnitude (M_s). It is evident that these source parameters show a remarkably systematic dependence on magnitude over the range 6.1 to 7.5. In particular, we point out the constancy (within a factor of 2) with magnitude of $\Delta\sigma$, and the linear variation of the sub-event size $2\rho_0$ with earthquake size. It is relevant also to point out that recently BERESNEV and ATKINSON (1999), in their simulations of strong ground motion using a model virtually identical to the "specific barrier model" (except that they use BRUNE's (1970) model to describe the radiation of the subevents), were compelled to use a subevent size that increases linearly with earthquake size in order to achieve best fits to the observed spectral amplitudes of the events that they analyzed. Finally, as pointed out originally by AKI *et al.* (1977) (see also PAPAGEORGIOU and AKI, 1982), the "barrier interval" $2\rho_0$ and the "rise time" τ are related as follows:

$$\text{(Barrier interval)} \sim \text{(Rise Time)} \cdot \text{(Rupture Velocity)}.$$

This relation suggests that knowledge of one of the two parameters i.e., the barrier interval or the rise time, permits the estimation of the other, given that the average rupture velocity ("sweeping velocity") V is a fairly stable parameter ($V = 0.6$ to $0.9\ \beta$, where $\beta =$ shear wave velocity).

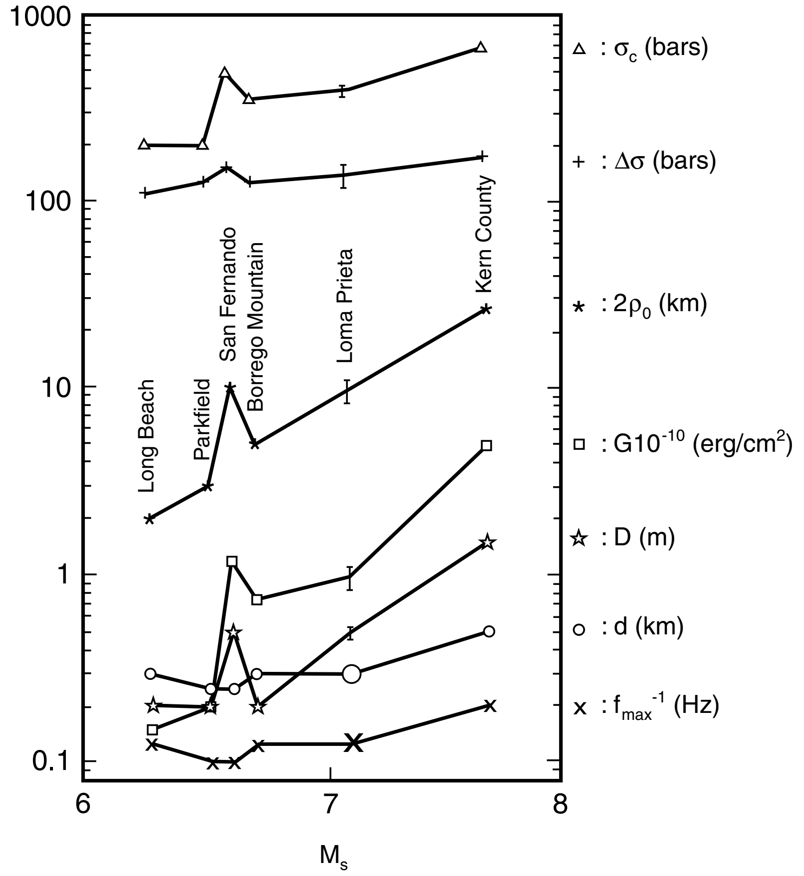

Figure 5
Source parameters of major California earthquakes.

The issue of f_{max}, whether it is due to source effects or recording site effects, has been controversial. HANKS (1982), ANDERSON and HOUGH (1984) and others found that f_{max} depends on the geologic condition of the recording site. On the other hand, AKI and PAPAGEORGIOU (1988) found that the f_{max} effect remained even after eliminating the site effect from the acceleration spectra. From an earthquake engineering point of view, f_{max} is an important parameter because it controls peak acceleration which is an important parameter for the seismic resistant design of structures. AKI and IRIKURA (1991) point out the work of KINOSHITA (1992) who found that f_{max} observed at the bottom of deep boreholes (about 3 km) in bedrock in Central Japan showed strong variation depending on the plate-tectonic setting of the seismic source. Furthermore, a weak but significant, in terms of its implications, increase of f_{max} with decreasing magnitude was observed by PAPAGEORGIOU and AKI

(1983b) for California earthquakes and by IRIKURA and YOKOI (1984) (see also AKI, 1988 and PAPAGEORGIOU, 1988) and UMEDA *et al.* (1984) for Japanese earthquakes, rendering another support for the source effect on f_{max}. The weak increase of the site of the break-down zone d with earthquake magnitude, which is evident in Figure 5, is a direct consequence of the above decrease of f_{max} with earthquake magnitude. As it was originally proposed by AKI (1988), (see also PAPAGEORGIOU, 1988, and references therein), the size of the break-down zone d is a measure of the width of the fault zone. With regard to this last statement, it is relevant to refer to the work of YAMASHITA and FUKUYAMA (1996). These investigators modeled that fault zone as a zone of densely distributed pre-existing cracks, consistent with seismological observations (LEARY *et al.*, 1987; LI *et al.*, 1994) and studied its behavior numerically. They found that the apparent critical slip displacement D is larger when the distribution density of the pre-existing cracks is larger and/or the fault zone width is greater. This is consistent with the observed variation of parameters D and d with earthquake magnitude shown in Figure 5. At any rate, we now have a more reliable estimator of the breakdown zone d, namely the low-velocity, low-Q zone measured by trapped modes in the fault zone (AKI, 2000, and references therein). It is relevant to point out that the estimates of d for the 1992 Landers earthquake zone (180 m; LI *et al.*, 2000) and the 1966 Parkfield earthquake zone (160 m; LI, *et al.*, 1997), based on the analysis of trapped modes, is in complete harmony with the estimates of d for the California earthquakes that PAPAGEORGIOU and AKI (1983b) analyzed (Fig. 5). Furthermore, for the 1992 Landers earthquake, a similar estimate of the fault zone width was made by an entirely different method at the same sites where the trapped modes were observed. From a detailed study of tension cracks on the surface, JOHNSON *et al.* (1994) concluded that the Landers fault rupture is *not* a distinct slip across a fault plane but rather a belt of localized shearing spread over a width of 50–100 m. AKI (2000) identifies this shear zone with the low-velocity, low-Q zone found from the trapped modes because their width is virtually the same at the same location on the fault. Since the trapped modes were observed from aftershocks with focal depths greater than 10 km, we conclude that the shear zone found by JOHNSON *et al.* (1994) extends to the same depth.

Regarding the issue of the origin of f_{max} of large earthquakes, we mention the work of AKI (1987) who proposed the hypothesis that f_{max} of large earthquakes is due to source effects and is causally related to the corner frequency of small earthquakes when the latter becomes constant below magnitude about 3 [for the tendency of the corner frequency to become constant for values of the seismic moment smaller than about 10^{21} dyn-cm (magnitude about 3 and corresponding source dimension of the order of 100 m) see CHOUET *et al.* (1978), ARCHULETA *et al.* (1982), among others]. AKI (1987) supported this hypothesis by analyzing borehole data (borehole located in the middle of the Newport-Inglewood fault) and demonstrating that there is a kink in the magnitude-frequency (of occurrence) relation at a magnitude around 3, reflecting a departure from self-similarity due to the effect of the fault width. ABERCROMBIE and LEARY (1993) and ABERCROMBIE (1995) investigated further Aki's above-mentioned hypothesis by analyzing also borehole data (borehole located at Cajon Pass, southern California) of small earthquakes. Based on the results of their analysis, the above authors concluded that there is no evidence of minimum source dimension at $\sim$100 m and that natural earthquakes are self-similar over a magnitude range $M \sim -2$ to $M \sim 8$. In correcting the analyzed data for attenuation, these authors assumed a constant Q. However, in a more recent study, ADAMS and ABERCROMBIE (1998), using data recorded in the same borehole and a robust method of analysis, found that Q is frequency-dependent (exhibiting strong frequency dependence for $f < 10$ Hz and weaker frequency dependence for $f > 10$ Hz). They conceded that the results (and therefore conclusions) of the previous study (that was based on the assumption of constant Q) may have been compromised. In fact, interpreting the variation of the observed frequency dependence of Q, ADAMS and ABERCROMBIE (1998) conclude that the earth's crust appears to be self-similar for length scales smaller than $\sim$100 m, and "smooth," possibly Gaussian, for longer length scales with correlation distances of a few kilometers. As a possible explanation of the above described apparent change, with length scale, of the crustal structure, these authors propose the presence of large crustal faults characterized by low velocity zones, about 100 m wide, in agreement with the findings of various other investigators (e.g., LI *et al.*, 1994).

Comparing several major California earthquake events (including those analyzed by PAPAGEORGIOU and AKI 1983b) with other major events from different tectonic environments AKI (1992b) observed that the source parameters of the California events deviate systematically from *self-similarity*. [According to the assumption of self-similarity, all earthquake events may be specified by a single parameter, say seismic moment M_0, and that small events are similar to large ones. Self-similarity implies *geometric similarity* i.e., length L, width W and slip Δu_0 all scale as $\sim M_0^{1/3}$, and *physical similarity* i.e., all nondimensional products of source parameters are the same, while the rupture velocity is constant and all parameters with the dimension of time scale as $\sim M_0^{1/3}$ (AKI, 1967; KANAMORI and ANDERSON, 1975)]. In particular, AKI (1992a) noticed that while fault length and width were not much different among these major California events, the decrease in moment was primarily due to decreasing slip (i.e., the amount of slip varies almost as $\sim M_0$ rather than $\sim M_0^{1/3}$, as one would expect if self-similarity were valid).

AKI (1992b) interpreted this departure from self-similarity in terms of both the "specific barrier-model" and the asperity model and he found the asperity model to be inconsistent with the above peculiarity of major California earthquakes. Furthermore, he found the asperity model to be inconsistent with the following observations: (1) The "asperity model" *cannot* explain the observed sharp impulsive displacement perpendicular to the fault plane (AKI, 1968; HASKELL, 1969; BOUCHON, 1979; MENDEZ and LUCO, 1990; YOMOGIDA, 1988; CAMPILLO *et al.*, 1989); (2) the observed *complementarity relation* between coseismic fault slip and aftershocks (AKI, 1979a; MENDOZA and HARTZELL, 1988; TAKEO, 1988; ZENG, 1991); (3) existence of barriers in the creeping segment of the San Andreas fault.

Also, in support of the "barrier model" is the work of ZENG (1991), ZENG *et al.* (1993a, 1993b), who by inversion found that the high-frequency energy sources of the 1989 Loma Prieta and the 1987 Whittier Narrows earthquakes are located along or near the boundaries of localized large slip zones, which is consistent with the theoretical consideration that high frequencies are primarily generated from the rupture stopping areas or places with large slip variation. These conclusions have been confirmed by other investigators based on the analysis of other earthquake events (KAKEHI and IRIKURA 1996, 1997; NAKAHARA *et al.*, 1998). There are also however notable exceptions. Specifically, from an analysis of the 1994 Northridge, California, earthquake, HARTZELL *et al.* (1996) found that of the two major sources of high-frequency radiation that they identified, the one located at the hypocenter (an area associated with a large final slip) is associated with the initiation of rupture and, apparently, the breaking of a high-stress-drop "asperity," while the second is associated with abrupt stopping of the rupture in a westerly direction, apparently by a "barrier."

Concluding, based on the above we may state that California earthquakes are of the "*barrier type*" family (AKI, 1984, 1988). On the other hand, asperities may be more important for great earthquakes along plate boundaries such as subduction zones (AKI, 1984, 1988; YOMOGIDA, 1988; CAMPILLO *et al.*, 1989).

The Barrier Model vis-a-vis Kinematic and Dynamic Source Models Inferred by Inversion of Strong Ground Motion

In the last two decades since the publication of the "specific barrier model" a sufficient number of earthquake slip models have been inferred by inversion of strong motion data so that certain systematic features emerged regarding the slip variation over the fault plane (MENDOZA and HARTZELL, 1988; SOMERVILLE *et al.*, 1999). Furthermore, the inferred kinematics of the earthquake source have been used to compute the dynamic features of the rupture process (i.e., "stress drop" $\Delta\sigma$, "strength excess" $(\sigma_u - \sigma_0)$ etc.) [e.g., BOUCHON (1997); GUATTERI and SPUDICH (2000) and numerous references therein]. In addition, by postulating a constitutive law (such as the slip-weakening model) investigators inferred parameters such as the "characteristic weakening slip" D and the "Griffith's specific fracture energy" G [for a critical and thorough review of the inference of constitutive law parameters such as D and G, see GUATTERI and SPUDICH (2000)]. Specifically, for the 1979 Imperial Valley, California, earthquake GUATTERI and SPUDICH (2000) estimate $G = 2 - 6 \cdot 10^9$ erg/cm^2 (consistent with an earlier estimate of $2 \cdot 10^9$ erg/cm^2 by BEROZA and SPUDICH, 1988) and they point out that there is a trade-off between D

and "strength excess" $(\sigma_u - \sigma_0)$ ($D = 1$ m with low "strength excess" or $D = 0.3$ m with high "stress excess," both estimates producing indistinguishable ground motion waveforms in the 0–1.6 Hz frequency band). For the 1992 Landers, California, earthquake OLSEN *et al.* (1997) estimate $D \sim 0.8$ m while for the same earthquake PULIDO and IRIKURA (2000) estimate $D \sim 1$ m for the Johnson Valley (southern) and Camp Rock/Emerson (northern) fault segments and $D \sim 3.5$ m for the Homestead Valley (central) fault segment. Finally, for the 1995 Hyogo-ken Nanbu (Kobe), Japan, earthquake IDE and TAKEO (1997) estimate $D \sim 0.5$ m for the deeper part of the fault and $D \sim 1$ m for the upper part of the fault. Comparing the above estimates of D with the corresponding values shown in Figure 5, we notice that they are remarkably close to—and in any case bound from above as expected (GUATTERI and SPUDICH 2000) — the estimates obtained using the parameters of the "specific barrier model" and high frequency waves (i.e., $f_{\max}$). [Regarding the above comparison, we should keep in mind that the resolution of the kinematic inversion models is limited because they are based on the analysis of rather long period waves ($f < 1$ Hz) and due to the effects of spatial and temporal-smoothing constraints applied in such inverse-problem formulations].

Regarding the "local stress drop" $\Delta\sigma_L$ and "cohesive stress" σ_c parameters, we compare our estimates (Fig. 5) with those of BOUCHON (1997) [the latter properly averaged over the regions of high stress drop (which in most cases coincide with regions of high slip)] and of other investigators (e.g., GUATTERI and SPUDICH, 2000) and we find that they are in reasonable agreement.

Next, what caveats should one be aware of regarding the "specific barrier model"? Clearly, the model is an end member of a spectrum of models that represent a main earthquake event as a collection of subevents. For instance, the subevent is modeled as a crack, the rupture of which is arrested at the perimeter by a barrier. We have already seen above that heterogeneities on the fault plane may act also as "asperities," in which case a model such as that of KOSTROV and DAS (1988) of a *circular fault with a central asperity* (see also FUKUYAMA and MADARIAGA, 1998) may have to be considered in representing a subset of the subevents that compose the main event. For this we would need a closed-form mathematical expression of the far-field radiation of such a subevent model, analogous to the expression of SATO and HIRASAWA (1973) for the circular crack.

Another concern regarding the "specific barrier model" is that all subevents are assumed to be of the same size, contrary to the more complex picture that has emerged from the kinematic models that have been obtained by inversion. We performed preliminary calculations allowing for a distribution of crack size around a representative size (similar to the model proposed by BOATWRIGHT, 1982) and we have found that estimates of the "local stress drop" $\Delta\sigma_L$ are not affected significantly (less than 30% change). This should have been anticipated because, for an earthquake event of a given magnitude, there is an average/typical subevent size that contributes most of the radiation. [This is tacitly recognized by SOMERVILLE *et al.* (1999) when they plot "area of largest asperity" vs. "seismic moment".]

Then, in view of the above concerns, how does use of the "specific barrier model" provide estimates of various source parameters that are in general agreement with estimates of the same parameters obtained by other means? The answer to the above question, at least partially, lies in the following facts: In estimating source parameters using the "specific barrier model," we start with the "local stress drop" $\Delta\sigma_L$ which we estimate from the *power* spectrum of the "stationary" segment of the accelerogram. The geometric parameter that controls the power spectrum of the radiation emitted by the source is an "*effective width*" (see eq. (47) of PAPAGEORGIOU and AKI, 1983a) which, at least for the strike slip California events that were analyzed using the model, is well approximated by the "*nominal*" width of the fault. Next, securing a reliable estimate of the "local stress drop" $\Delta\sigma_L$ we proceed to estimate the barrier interval from the ratio $(\Delta\bar{u}/2\rho_0)$ which represents the "local strain drop" which in turn is proportional to the "local stress drop." The uncertainty here lies with estimates of $\Delta\bar{u}$ in view of the fact that it appears that only a fraction (say 50%) of the nominal fault plane slips significantly (and radiates seismic energy) while in the "specific barrier model" we assume a uniform distribution of cracks covering the entire nominal rupture area of the fault plane. This would involve an uncertainty of a factor of not more than 2. However, usually this is the uncertainty associated with estimates of the seismic moment M_0. [It is evident that as more data accumulate regarding the percentage of the nominal rupture area that slips significantly radiating seismic energy, the information could readily be incorporated in the procedure of estimating parameters of earthquake sources using the "specific barrier model".]

Summarizing, the "specific barrier model" provides the most complete, yet *parsimonious*, self-consistent description of the faulting processes that are responsible for the generation of high-frequency waves. The model, in spite of its simplicity, is robust enough to provide reasonable estimates of various important source parameters and provides an effective tool to simulate/model strong ground motion for engineering applications. Until the kinematic inversion studies of the earthquake source are based on considerably higher frequency waves than the present ones, the "specific barrier model" contributes to earthquake source studies.

Numerical Simulation of Strong Ground Motion – Forward Modeling

Considerable progress has been made in recent years in understanding strong ground motion in terms of source, propagation path and recording site effect. As a demonstration of the level of the achieved understanding one may refer to the plethora of successful numerical simulations of observed strong ground motions using various mathematical models of the earthquake source and earth medium.

The simplest and least expensive method for simulating strong ground motion is based on the assumption that accelerograms are realizations of a stochastic process with time varying intensity and, possibly, frequency content. HOUSNER (1947, 1955) interpreted the erratic appearance of strong-motion accelerograms by reasoning that seismic waves are initiated by irregular slippage along faults followed by numerous random reflections, refractions and attenuations along the propagation path. Following Housner's *paradigm* many investigators (e.g., HUDSON, 1956; BYCROFT, 1960; HOUSNER and JENNINGS, 1964; JENNINGS *et al.*, 1968; JOYNER and BOORE, 1988; SHINOZUKA, 1988, see last one for a recent review) developed stochastic models for the analysis of recorded accelerograms or the computation of synthetic ones.

At the time Housner formed the above hypothesis, earthquake source theory and methods for evaluating Green's functions for realistic earth media were not well developed. Thus, Housner proceeded by considering simple, yet effective for earthquake engineering purposes, functional forms for radiated waves. As it was elaborated in the previous section, now we know that high frequency waves emanate from the rupture front as it interacts with heterogeneities (i.e., barriers and asperities) of the fault plane, and that ground motion may be computed by convolving the slip function with the Green's tensor of the earth (see eq. (1)). By now it should be apparent to the reader that Housner's original idea of modeling high-frequency seismic radiation has obtained a more concrete expression with the "specific barrier model" of PAPAGEORGIOU and AKI (1983a) that was presented in the previous section.

The developments related to the stochastic modeling of accelerograms came about thanks primarily to the efforts of the engineering community. Not having a physical model to describe the frequency content of the elastic waves radiated by the earthquake source, earthquake engineers adopted simple and/or empirical spectral models (e.g., white noise spectrum, Kanai-Tajimi spectrum; see for example CLOUGH and PENZIEN, 1975).

Recently seismologists, recognizing the stochastic character of high-frequency waves, adopted the engineering approach in simulating strong motion accelerograms, based on the assumption that they are realizations of "*band limited gaussian white noise*" with time varying intensity (HANKS, 1982; HANKS and MCGUIRE, 1981; BOORE, 1983). The contribution that the seismologists made to this development consists of the fact that they used a physical model (instead of an empirical one) to describe the spectral content of the simulated motions. In particular, they adopted BRUNE's (1970) "ω^2-*model*" to describe the *source spectrum* (i.e., the spectrum of the elastic waves radiated by the source before these has been modified by the propagation path and site effects) and assumed self-similarity to establish the *scaling law* of the source spectrum (i.e., how the source spectrum scales/varies with earthquake size; AKI, 1967, 1972b).

The stochastic modeling of accelerograms, as was originally used by earthquake engineers or even in its most refined form proposed by seismologists (e.g., BOORE, 1983), has the following limitations: (1) The model is based on a point source; (2) the model is not adequate for simulating the long-period *near-field effect* expected in the *near-source region* from the slip on the fault; and (3) the model provides a description of the *temporal* variation of ground motion of a single point of the ground but cannot provide a description of the *spatial* variability of ground motion which is necessary for the analysis of extended structures (e.g., pipelines, tunnels, bridges, dams).

The above three limitations apply irrespective of the spectral model that one may adopt. In addition, the "ω^2-model" fails to explain the observed $M_s - M_0$ relation for large earthquakes (BOORE, 1986) and there appears to be a consensus that a single corner frequency model (such as the "ω^2-model") cannot explain observations for the entire frequency range of large events (e.g., GUSEV, 1983; PAPAGEORGIOU and AKI, 1985; PAPAGEORGIOU, 1988; ATKINSON, 1993; BOATWRIGHT, 1994; HADDON, 1995, 1996a). Furthermore, a fundamental problem with the "ω^2-model" is the ambiguous nature of the key parameter called "*stress parameter*" (BOORE and ATKINSON, 1987). [Parenthetically we point out that the stress parameter that appears in the "specific barrier model" and is referred to as the "local stress drop," has a clear physical meaning—being the stress drop of the subevents that compose the earthquake event—has been found to be a very stable parameter for a given tectonic region (see Fig. 5) and may be estimated even by geological exploration methods (*paleoseismology*) thus rendering the model potentially useful in predicting strong ground motion even for tectonic areas for which there are no recordings (AKI, 1984).]

The limitations of the engineering approach for simulating strong motion may be eliminated by using the deterministic kinematic modeling approach based on the representation theorem (eq. (1)). It was pointed out earlier that in order to apply eq. (1) one needs to know: (1) the slip history of the fault rupture (i.e., when, how much, for how long and in what direction each point of the fault slipped, or will slip, during an event), and (2) the Green's function of the earth with enough accuracy for the frequency range of interest.

Regarding the first requirement above, if one is interested in simulating ground motion to be generated by a fault which may potentially rupture, it is evident that the slip history involves many parameters that must be specified. One possible way to proceed is to adopt the slip history, inferred by inversion, of an event of similar size, with the same *source mechanism* (i.e., strike-slip, dip-slip, etc.) and preferably from the same tectonic region (e.g., SAIKIA, 1993; HEATON *et al.*, 1995). Alternatively, one may adopt the "specific barrier model" to parameterize the slip history and obtain reliable estimates of the size of the subevents.

Regarding the evaluation of the Green's function, the chief factor limiting its accuracy is ignorance of the earth structure at the source-site region. For example, in order to simulate deterministically ground motion reaching a maximum frequency of 5 Hz, it is necessary to know the 3-D structure of the earth on a scale of a few hundred meters. In view of the fact that both engineers and seismologists recognize the stochastic character of high frequency strong motion, the following approach for earthquake motion simulation may be proposed: Use the *Empirical Green's Function Method* (e.g., HARTZELL, 1978, 1989; WU, 1978; HUTCHINGS and WU, 1990; HADDON, 1996b) or stochastic modeling (e.g., ZENG *et al*., 1993a, 1995) to simulate ground motions in the high frequency range (say above 1 Hz) and combine these results with those obtained using deterministic kinematic modeling in the low frequency range (say below 1 Hz) (e.g., ZENG *et al*., 1993a; HEATON *et al*., 1995). The above recommended procedure tacitly recognizes the fact that ground motions at periods longer than 3 sec of past events have not been reliably recorded by the analog strong motion instruments.

Finally, it is well recognized that ground shaking and associated damage to engineered structures are strongly influenced by the geology in their vicinities. The *Coda Method*, exploiting the well established separability of source, path and site effects on coda waves of local seismic events originally proposed in the seminal work of AKI (1969), offers a cost-effective way to empirically determine the *site amplification factor* for regional microzonation (PHILLIPS and AKI, 1986; SU and AKI, 1995; AKI, 1993, see last one for review). Complications that are caused by the nonlinear behavior that unconsolidated sediments exhibit when subjected to large strains, were first detected seismologically by CHIN and AKI (1991) for the 1989 Loma Prieta earthquake. In order to correct the coda (i.e., weak motion) site amplification factor to be applicable to strong motions (i.e., motions exceeding a threshold peak acceleration), AKI and CHIN (1994) have proposed a very simple method that they tested in connection with 1992 Landers earthquake data. The method appears promising and requires further testing with strong motion data that have been recorded by accelerographs collocated with high frequency (i.e., weak motion) instruments or broad-band instruments (KATO *et al*., 1995). Since the publication of the work of CHIN and AKI (1991), various investigators have observed seismologically the effect of soil nonlinearities for other earthquakes (e.g., SU *et al*., 1998).

Ultimately, it should be evident that the intent of the engineering approach to strong motion simulation is to capture the essential characteristics of high-frequency motion at an average site from an average earthquake of specified size. Phrasing this differently, the accelerograms artificially generated by engineers do not duplicate any specific earthquake but rather embody certain average properties of past earthquakes of a given magnitude (SHINOZUKA, 1988). Contrastingly, the kinematic modeling approach adopted by seismologists involves the prediction of motions from a fault that was identified by geologists and which has specific dimensions and orientation in a specified geologic setting. This latter approach is useful for *site-specific* simulations.

Conclusions

The last three decades have witnessed remarkable advances in the field of Engineering Seismology. It is fair to say that earth scientists have developed the capabilities to synthesize ground motions generated by realistic sources embedded in realistic propagation media to such a degree, that they are currently capable of assessing the range of plausible ground motions at the site of an engineering structure. Every aspect of Strong Motion Seismology has been influenced by the seminal works of Keiiti Aki. His contributions include the first modeling of near-source ground motion, introduction in seismology of the concept of earthquake source spectrum and its scaling with magnitude, study of fault rupture using models of Fracture Mechanics, documentation and characterization of fault heterogeneity responsible for the generation of the short-period and high-frequency waves, introduction of the ''barrier model'' (and a mathematical expression of it referred to as the ''specific barrier model''), inversion study to identify the sources of high-frequency radiation on the fault plane, study and modeling of coda waves, analysis and numerical modeling of site effects including a cost-effective method for regional microzonation using the coda method. All of Keiiti Aki's contributions plowed new ground and opened new vistas of research. Like Galileo, he focused on fundamental seismological phenomena and quantified them. He left an indelible mark both as a superb scientist and as a great teacher. Let all of us follow in his footsteps.

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