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INVERSE STUDIES OF THE EARTHQUAKE SOURCE MECHANISM FROM NEAR-FIELD STRONG MOTION RECORDS

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ABSTRACT

This paper presents a review of finite fault modeling, i.e. inversion of the earthquake source mechanism using near-source strong motion records as input. The first earthquakes for which there was adequate strong motion data for such studies were the 1966 Parkfield and 1971 San Fernando, California earthquakes. Since then, the finite fault modeling has become an integral part of the modern earthquake studies. Knowledge of the features of the earthquake source mechanism, as retrieved from near-field strong motion records, is valuable for understanding the physical processes related to brittle release of tectonic strain during earthquakes, and for prediction of strong ground motion from future earthquakes in seismic design of structures. The paper includes a review of (1) the published literature on this topic, (2) the methodology (theoretical source models and solution of the inverse problem) with emphasis on the difficulties and limitations, and (3) the strong motion data that has been used as input in these studies. The paper also comments on the source mechanism of submarine earthquakes, slides and slumps, and discusses the possibility of using maregrams for inversion of their source mechanism.

KEYWORDS: Earthquake Source Mechanism, Inverse Problem, Near-Field Strong Motion Data, Tsunami Sources

INTRODUCTION

There are many scientific reasons why detailed knowledge of the earthquake source is valuable, but the destructive effects of earthquakes are a sufficient driving force to learn as much as possible about their nature and effects. Because direct measurement of the source activity is not feasible, one can learn about the source only from distant (weak motion) and near (strong-motion) records. The inversion can be associated with different source parameters, depending on the mathematical model, but ultimately the goal is to learn as much as possible about the nature of the source. In earthquake engineering, understanding of the earthquake source is important for prediction of ground motion from future possible earthquakes, which is needed for seismic design of structures.

In the last several decades, much research has been done in strong motion seismology, which has led to better understanding of the nature of earthquake source. The early studies focused on determining the orientation of fault plane. An important factor in the further development of source mechanism research was the deployment of strong-motion accelerographs. Developed mainly for engineering purposes, to measure the seismic forces in structures, the accelerographs are now the main source of near-field data used to study the earthquake source (Trifunac and Todorovska, 2001), and have contributed significantly to the earthquake source investigations.

The problem of characterizing the earthquake source from recorded data is an inverse type of problem, i.e. one in which the input of a system has to be determined from the output. The system function here describes the modification of the motion radiated from the source by the propagation path. A major difficulty in finding a unique solution to this problem is detailed specification of the system function. Even though the associated wave propagation phenomena are well understood at this time, too detailed modeling of the system function would (1) make the inverse problem too complex to solve numerically, and (2) even be useless for practical applications, because the three-dimensional geologic structure along the propagation path is usually known only on a rough scale. One remedy for this problem is to use records (system output) from stations very close to the fault, which are least influenced by the propagation path. The likelihood of recording many such records for every damaging earthquake, however, is not very high because the exact time and place of occurrence of such earthquakes is not known, and it is still prohibitively expensive to saturate the earth with sufficiently dense strong motion arrays. Hence, the propagation path effects have to be included in the model. Because of these constraints, the system function implemented in the model is only an approximation of the real world, and cannot predict all the features contained in the recorded motions (system output). This problem is treated as "noise" in the data. Another type of noise results from approximations due to the finite arithmetics of the digital computers. All of this leads to the inverse problem being "ill-posed". The finite source modeling, hence, reduces mathematically to a solution of an inherently ill-posed inverse problem, which does not have a unique solution. Solving this problem means finding an acceptable solution, using some regularization techniques and imposing physical constraints.

This paper presents a review of the developments in seismic source inversion using near-source records, typically recorded by the strong motion accelerograph networks. This problem is also known as finite source modeling (the source has a finite size), in contrast to point source modeling in the studies that use distant records, typically recorded by the seismological networks. The paper first reviews published literature on this topic, which is followed by a review of the methodology, including theoretical source models and formulation and solution of the inverse problem (which, for a number of reasons mentioned in the preceding paragraphs, is ill-posed). Finally, as the progress in finite source inversion has been strongly conditioned on the quantity and quality of strong motion records available for such studies, this paper also reviews the developments in recording strong motion data that has been used in such studies. The consideration of other types of data, such as teleseismic body waves, surface waves, and static fault offset and geodetic data, and their use in source inversion is beyond the scope of this paper. This paper also comments on the possibility of source inversion for submarine earthquakes, slides and slumps using maregrams.

HISTORICAL REVIEW

The first useful recordings for inverse analyses, resulting from the systematic deployment of accelerographs in California, were those of the 1966 Parkfield earthquake (Housner and Trifunac, 1967), and of the 1971 San Fernando earthquake (Trifunac and Hudson, 1971). The proximity of the accelerographs to the source (< 50 km) and their ability to record high frequencies ($f < 25$ Hz) produced records at distances comparable to the wave lengths of the energy radiated by the source, and of the same order as the source dimensions, thus reducing the complications caused by the propagation path. The 1966 Parkfield earthquake was recorded by a linear strong motion array, perpendicular to and near the southeastern end of the surface rupture (Trifunac and Udwadia, 1974). The 1971 San Fernando earthquake was recorded by more than 90 accelerographs, in the free-field or in building basements. Five of these stations (four surrounding the fault, with fault-to-station distances less than 50 km, and one, the Pacoima Dam site, centered above the fault) could be used for inversion of the fault slip (Trifunac, 1974).

Figure 1 illustrates the accumulation of strong motion data for the period between 1933 and 1994. It shows the total number of "free field" strong motion accelerograms recorded during main events (solid points), and the cumulative number of strong motion accelerographs in U.S. and in Japan up to 1980. The times of selected California earthquakes contributing to the strong motion database for southern California are also shown on the same time scale.

The trial and error, forward simulations of recorded strong ground motion near a source first appeared following the Parkfield, California, earthquake of 1966 (Aki, 1968, 1979; Haskell, 1969; Boore et al., 1971; Tsai and Patton, 1973; Niazy, 1973; Boore and Zoback, 1974a; Anderson, 1974, 1976; Kawasaki, 1975; Levy and Mall, 1975; Hartzell et al, 1978; Bouchon, 1979, 1982; Murray, 1967; Luco and Anderson, 1985; Anderson, 1976; Liu and Helmberger, 1983), Borego Mountain earthquake of 1968 (Heaton and Helmberger, 1977) and San Fernando, California earthquake of 1971 (Mikumo, 1973; Boore and Zoback, 1974b; Niazy, 1975; Bouchon and Aki, 1977; Bouchon, 1978; Heaton and Helmberger, 1979). The number of forward studies of more recent well-recorded earthquakes is too large to enumerate here.

The first inverse studies were performed by Trifunac (1974) and by Trifunac and Udwadia (1974). In this pioneering work, the fault surface was divided into subfault segments, and the slip in each segment was chosen to give the best fit, in the least square sense, to the observed displacements integrated from the recorded accelerograms. The fault geometry, the dislocation velocity, the dislocation rise time and the sense of the dislocation propagation were chosen using other independent information, or by a trial and error procedure. Ten years later, following the Imperial Valley California earthquake of 1979 (Hartzell and Helmberger, 1982; Hartzell and Heaton, 1983), the inverse method for selecting the dislocation amplitudes was revived and further improved (Olson and Apsel, 1982; Jordanovski et al., 1986; Jordanovski and Trifunac, 1990a, 1990b). The reason for this delay was that up to that time, there were very few good quality near-field records that could be used for source inversion, and that the theoretical model of the source required too many *a priori* prescribed parameters, which led to non-uniqueness of the solution.

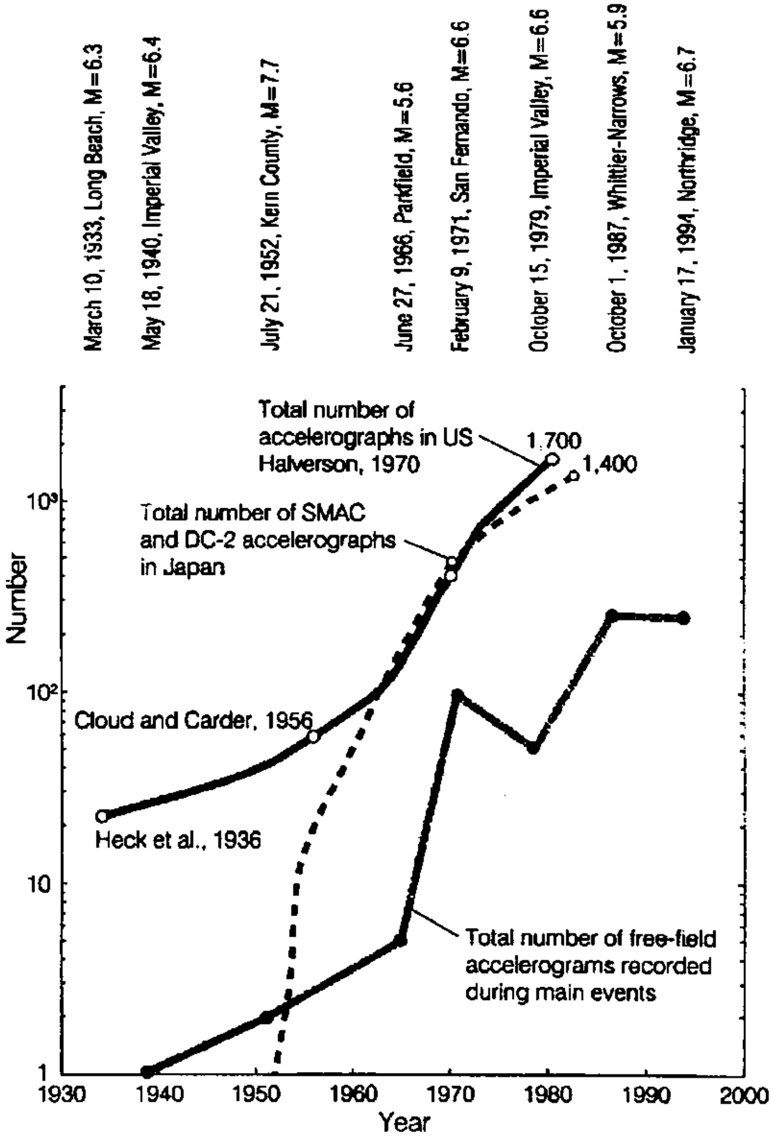

Fig. 1 Total number of "free field" strong motion records recorded during main events (solid points), and the cumulative number of strong motion accelerographs in U.S. and in Japan up to 1980 (selected California earthquakes contributing to the strong motion database for southern California are also shown on the same time scale)

The next well-recorded earthquake which provided data for source inversion was the October 1, 1987, Whittier-Narrows earthquake. This was also the first significant event to be recorded by the Los Angeles strong motion array (Trifunac, 1988; Trifunac and Todorovska, 2001). Together with the stations operated by the United States Geological Survey (USGS) and California Division of Mines and Geology

(SDMG), with many strong motion stations surrounding the source, it provided unprecedented new possibilities for advanced source inversion studies (Hartzell, 1990). Since 1987 and through 1990s, it has became common to have hundreds of accelerograms from major earthquakes (Figure 1), and since the Loma Prieta, 1989, California earthquake, the inversion of the fault slip based on strong motion records has became one of the central topics in the study of all well-recorded major earthquakes.

Other earthquakes, for which inverse studies of the fault slip could be carried out, are: Parkfield, 1966, California (Trifunac and Udwadia, 1974); San Fernando, 1971, California (Trifunac, 1974); Tabas, 1978, Iran (Hartzell and Mendoza, 1991); Imperial Valley, 1989, California (Jordanovski and Trifunac, 1990a,1990b; Hartzell and Heaton, 1983; Olson and Apsel, 1982); Izu-Hanto-Toho-Oki, 1980, Japan (Takeo, 1988); Naganoken-Seibu, 1984, Japan (Takeo, 1987; Takeo and Mikami, 1987); Morgan Hill, 1984, California (Beroza and Spudich, 1988; Hartzell and Heaton, 1986; Mikumo and Miyatake, 1995); Central Chile, 1985, Chile (Mendoza et al., 1994); Michoacan, 1985, Mexico (Mendoza and Hartzell, 1989); North Palm Springs, 1986, California (Hartzell, 1989); Superstition Hills, 1987, California (Wald et al., 1990); Whittier-Narrows, 1987, California (Hartzell, 1990; Hartzell and Iida, 1990; Iida and Hartzell, 1991; Zeng et al., 1993); Loma Prieta, 1989, California (Beroza, 1991; Steidl et al., 1991; Wald et al., 1991); Sierra Madre, 1991, California (Wald, 1992); Landers, 1992, California (Wald and Heaton, 1994a; Hartzell and Liu, 1995); Northridge, 1994, California (Wald and Heaton, 1994b; Wald et al., 1996; Hartzell et al., 1996); and Chi-Chi, 1999, Taiwan (Chi et al., 2001; Ma et al., 2001; Zeng and Chen, 2001; Oglesby and Day, 2001; Wu et al., 2001; Wang et al., 2001).

The source mechanism and near-field studies on land have benefited from the steady accumulation of recorded strong motion data (see Figure 1), and from the relative simplicity of the physical problem (linear wave propagation through a solid, from the source to the receiver). In contrast, the source mechanism studies of submarine earthquakes, slides and slumps, and analyses of near-field wave motion in an ocean (tsunami) caused by the movement of the ocean bottom are, at present, only in the early stages of development. First, the source mechanism studies using tsunami wave recordings are complicated, because the wave propagation through both the solid and fluid must be considered, and second, there are no near-field, open ocean recordings surrounding a tsunami source, so far. Traditional source studies, which used tsunami recordings, at teleseismic distances, were based on a point-source representation of the earthquakes (Ben-Menahem and Rosenman, 1972; Kanamori, 1972; Comer, 1982; Ward, 1980, 1981, 1982; Okal, 1988). More recent studies of localized tsunami run-up (e.g., Heinrich et al., 2000) and of unusually large tsunami run-ups (e.g., Pelayo and Wiens, 1992), suggest that slow earthquakes and submarine slides and slumps can be powerful, but spatially complicated sources of tsunami in the near-field (Todorovska and Trifunac, 2001; Todorovska et al., 2002; Trifunac et al., 2001a, 2001b, 2002a, 2002b; Trifunac and Todorovska, 2002). For "fast" earthquake dislocations, which spread with velocities between 1.5 and 3 km/s, wave propagation in the fluid layer often can be neglected, and the tsunami amplitudes above the source can be assumed to be the same as on the ocean bottom (Geist, 1999; Satake, 1987, 1989). Under those conditions, the presence of the fluid layer may be neglected. For "slow" source processes, when the uplift of the ocean bottom spreads with velocities comparable to the long period tsunami velocity $c_T = \sqrt{gh}$ (g is acceleration due to gravity, and h is the depth of the ocean), the near-field waves in the fluid cannot be ignored and the process of searching for the earthquake source mechanism or for the time-dependent evolution of slides and slumps becomes difficult (Trifunac et al., 2001a, 2001b, 2002a, 2002b, 2002c). Assuming that the records of water wave motion versus time in the near-field become available, it will be possible, in principle, to invert for the time-dependent vertical displacements of the ocean bottom (Trifunac et al., 2002c). After time and space-dependent movement of the ocean bottom has been deciphered, it will then be possible, in the second stage, to perform inverse studies of the causative fault slip, or of slide or slump motions.

REVIEW OF METHODOLOGY

1. Seismic Source Models

Solving an inverse or identification problem consists of matching theoretical results with observations. Hence, the first step is to define a theoretical model of the source process (de Hoop, 1958). This model has to represent the physical process with satisfactory accuracy, or at least to represent its overall characteristics (Luco, 1987).

During the 1950s, for the first time, the problem of seismic source radiation was examined from the viewpoint of dislocation theory, which assumes the source to be a displacement discontinuity across the fault surface. Veedenskaya (1956) defined a system of forces equivalent to a rupture, and a few years later, Knopoff and Gilbert (1959, 1960) presented a detailed analysis of the dislocation problems for modeling a seismic source. They showed that an earthquake could be modeled as a combination of several fundamental problems.

In the 1960s, Maruyama (1963) presented a rigorous formulation of the force-equivalent representation for a general dynamic dislocation, and a year later, Burridge and Knopoff (1964) reformulated the body force equivalents using the theory of distributions. These two approaches showed that the double couple representation is the appropriate way to model the source radiation. The third important paper was that by Haskell (1969) on elastic displacements in the near-field of a propagating fault, using body force equivalents for a full elastic space, a rectangular fault, and a dislocation with a ramp time function, moving with constant velocity along the fault length. Haskell's model has been used with slight modification in almost all source inversion analyses, and will be reviewed in more detail later in this paper.

After 1964, when the foundations of the mathematical models of the source were well established, many analyses have been carried out that use a source model in practical applications. Explicit solutions have been derived for a homogeneous and a layered half-space (Sato, 1975; Apsel, 1979). The solutions then have been used to synthesize near-field and far-field ground motions, and to explain the radiation properties of earthquakes by solving the forward (Aki, 1968), or inverse (Trifunac, 1974) problem. Two basic types of source models were developed, kinematic and dynamic models, as discussed in the following.

1.1 Kinematic Models

The kinematic models assume that the entire dislocation time history is known at each point on the fault. It enables one to calculate the magnitude of the double couple in the body force equivalent representation of the source, and then the response of the surrounding medium, by using an integral representation or the equilibrium equations. Historically, this was the first source model. Due to the relatively simple numerical computations, it has been applied extensively to source inversion problems.

The weakness of the kinematic model is its requirement for *a priori* knowledge of the dislocation time history. The dislocation time history is determined by the interaction of forces in the source region, and physically meaningful functions are selected from the closed form solution for the propagation of a simple shear crack (Kostrov, 1964), or from numericai solutions for dynamic source models (Madariaga, 1976; Das and Aki, 1977; Day, 1982; Archuleta and Frazier, 1978). The basic characteristics of the dislocation time function are: (1) the square root singularity of its derivative with respect to time at the tip of the dislocation (this implies a significant change in slope at the initial time of its growth), and (2) that it reaches the final offset value with no additional jumps or sudden changes in slope.

Aki (1968) used a Heaviside step function as a dislocation time function, and compared synthetic seismograms from a moving dislocation model with those observed (i.e., computed from recorded accelerograms) during the Parkfield earthquake of June 28. In spite of the simplicity of the dislocation time function, his model matched successfully the spike shape of the displacement component perpendicular to the fault trace.

The most common dislocation time function used in kinematic models is the ramp function, first considered by Haskell (1964, 1969), and referred to as "Haskell's source model". It is described by

$$a(\tau) = \begin{cases} \dfrac{D_0 \tau}{T_R} & 0 < \tau < T_R \\ D_0 & T_R < \tau \end{cases} \tag{1}$$

where D_0 is final dislocation amplitude (offset) and T_R is the rise time. Its main characteristics are: rectangular fault shape, one-dimensional rupture propagation along the fault length with constant velocity, instantaneous rupture propagation along the fault width, ramp time history, and a constant dislocation final offset. This model has been used by many researchers, e.g., by Trifunac (1974) and Trifunac and Udwadia (1974), to estimate the final dislocation offset for the 1971 San Fernando and the 1966 Parkfield earthquakes, and by a number of Japanese researchers to compute theoretical seismograms (Sato, 1975; Kawasaki et al., 1975; Sudo, 1972).

The other kinematic models are more or less smoother versions of the ramp function, with continuous time derivatives, converging continuously to zero at time t = 0, as opposed to the ramp function which has a discontinuous (rectangular shaped) time derivative at t = 0. For example, Ohnaka (1973) used the combination of a linear and exponential time functions, while Hartzell (1978) used a quadratic dislocation during the rise time.

A qualitatively different dislocation function used by the kinematic models is the one obtained from Kostrov's self-similar solution for shear-crack propagation. However, Anderson and Richards (1975) have shown that the differences in the near-field displacements between the ramp dislocation and the Kostrov's function are small.

1.2 Dynamic Models

The dynamic models describe the fault behavior by prescribing the force and stress distribution over the fault surface. Archambeau (1968) was the first one who formulated the source process as a stress relaxation problem. Under the influence of a pre-stress system, failure occurs within a given region where the material properties suddenly change (plastic flow), and strain energy is released. A part of that energy generates motion of the fault walls, and another part is radiated into the surrounding medium. Obviously, the dynamic approach is a more general and natural one, and the displacement history on the fault is obtained by solving the dynamic problem. Unfortunately, the physical processes that take place in the source region during an earthquake are still mathematically too difficult to model, making it impossible to solve the problem without a great simplification.

The first solution of the stress relaxation problem was presented by Kostrov (1964). He considered motion produced by a uniformly expanding circular shear crack in an infinite homogeneous medium pre-stressed by a constant stress. This problem is self-similar and, although it represents a very simplified earthquake model, it is of great importance for understanding the dislocation behavior. Kostrov presented a closed form solution for the dislocation $a(\xi,t)$

$$a(\xi,t)=\begin{cases} D_0\sqrt{v^2t^2-|\xi|^2} & |\xi|<vt \\ 0 & |\xi|<vt \end{cases} \tag{2}$$

where t is time, ξ is the coordinate of a point on the fault, v is the rupture velocity, and D_0 is a constant depending on the material properties and stress drop. In space, the dislocation is an ellipse, growing hyperbolically with time. Burridge and Willis (1969) extended Kostrov's solution by considering an elliptical fault expanding with different velocities in the prestressed and in the perpendicular direction. Their solution is analogous to Kostrov's self-similar solution.

Richards (1973, 1976) examined the properties of an elliptical crack in detail. His results showed that, in order to have a finite value of the stress in the front of the rupture tip, in the prestressed direction the rupture velocity has to be equal to the Rayleigh velocity, and in the perpendicular direction it must approach the shear wave velocity. However, for other directions, the stresses are still singular. Richards' analysis also showed that the breaking phase causes sudden changes in the particle accelerations, but these changes are almost impossible to notice in the displacement time histories.

Brune (1970) proposed a very simple dynamic model, in which the stress drop occurs instantaneously over the entire fault surface. By solving the equation of motion, he obtained a linear rise of the dislocation amplitudes with time. To account for the effects of the finite dimensions of the fault, Brune introduced an exponential decay factor such that

$$\frac{d}{dt}a(\xi,t)=\frac{\Delta\tau\beta}{\mu}e^{-t/T} \tag{3}$$

and

$$a(\xi,t)=\frac{\Delta\tau\beta T}{\mu}\left(1-e^{-t/T}\right) \tag{4}$$

where, $\Delta\tau$ is the shear stress drop, β is the shear wave velocity of the medium, and μ is Lame constant (shear modulus). Although his model is intuitive, it has proven to be very useful to obtain the global source parameters like seismic moment, stress drop or corner frequency (Trifunac, 1972a, 1972b; Hanks, 1972).

The models reviewed so far did not consider quantitatively the effects due to the finite fault dimensions. By introducing a stopping phase and finiteness of the fault, the dynamic model becomes very complicated to analyze analytically. However, several authors have investigated these effects by using numerical analysis. Madariaga (1976) examined the sudden stopping of a circular rupture, and showed that Kostrov's solution is good up to the time when the effects of the boundaries reach the observation point.

Day (1982) considered a rectangular fault, and his results are similar to those for a circular finite fault except for the point near the edge where he observed a very strong stopping phase in the stress time history. Das and Aki (1977) investigated two-dimensional spontaneous rupture propagation for an in-plane and an anti-plane shear crack.

In general, the dynamic models, so far, have been used mostly to investigate the nature of the dislocation, and have not yet found wide application in source inversion, with the exception of Brune's model. However, they have been very important for the kinematic models by setting realistic criteria for selection of their dislocation time functions.

2. Formulation of the Inverse Problem

Let Σ and V be three-dimensional vector spaces representing the fault surface and the surrounding medium. Then, a mathematical model of the earthquake source radiation can be used either to solve for the displacement at a point $\mathbf{x} \in \mathrm{V}$, given a dislocation $a(\xi,t)$, $\xi \in \Sigma$ (the forward problem), or to solve for the dislocation $a(\xi,t)$, $\xi \in \Sigma$, given the displacements at points $\mathbf{x} \in \mathrm{V}$ (the inverse problem).

Based on the body force equivalent representation, the displacement at any point $\mathbf{x} \in \mathrm{V}$ due to a displacement discontinuity $a(\xi,t)$ across the fault plane Σ, for $t \in (-\infty, T)$, where T is the duration of the source process, is given by

$$u_n(\mathbf{x},t) = -\int_{-\infty}^{t}\int_{\Sigma} a_i(\xi,\tau) K_{in}(\mathbf{x},\xi,t-\tau)\, d\Sigma\, d\tau \tag{5}$$

where

$$K_{in}(\mathbf{x},\xi,t-\tau) = C_{ijpq}\, n_j(\xi) \frac{\partial}{\partial \xi_q} G^P_{n,q}(x,\xi,t-\tau) \tag{6}$$

is the kernel of the integral Equation (5), which depends on the tensor C_{ijnq} of the material properties of the geological medium; $n_j(\xi)$ is the normal to the fault surface; and $G^P_n(\mathbf{x},\xi,\, t-\tau)$ is the fundamental solution of a wave created by double couple forces. If one could find the inverse operator $K^{-1}_{in}(\mathbf{x},\xi,\, t-\tau)$, the dislocation $a(\xi,t)$ would be obtained easily. However, this is not an easy task. The usual approach consists of the following steps:

1 Choosing a kernel (6) consistent with the boundary conditions for integral Equation (5): The kernel $K_{in}(\mathbf{x},\xi,\, t-\tau)$ given by Equation (6) is the fundamental solution of a wave created by double couple forces, propagating through inhomogeneous media, and can be very complicated and difficult to use in practical applications. Consequently, many researchers have used simpler models of the propagation path, assuming a homogeneous, isotropic, and elastic full-space, and accounting for the free surface effect by doubling the value of the displacement (Trifunac, 1974; Trifunac and Udwadia, 1974; Jordanovski et al., 1986). Some authors have considered layered viscous half-space (Olson and Apsel, 1982; Hartzell and Heaton, 1983), but their results show that in some cases, simple full or half-space solutions are justified, considering a trade-off between the complexity of the model and the accuracy of the final results. On the other side, the propagation path effects can be reduced if the recording stations are close to the fault.

The use of empirical Green's functions has attracted much attention. This method consists of representing the ground motion from a large earthquake, by using motions from small and medium earthquakes as kernels, and by convolution of these motions with the dislocation time history. In principle, because the empirical Green's functions intrinsically include propagation path effects and site characteristics, this method has an advantage that strong ground motion from a large earthquake can be predicted with satisfactory accuracy in a wider frequency range. The weakness of this method is the lack of records of small earthquakes in the fault region of a large earthquake, and the lack of small earthquakes with similar source characteristics (such as radiation pattern and dislocation time history, for example) to the ones for the main event.

2. *Defining a dislocation function $a(\xi,t)$ in which the parameters to be assessed are unknown:* In general, there are several basic characteristics of the dislocation which must be specified *a priori*: its temporal behavior, the rupture velocity or more generally the time when it reaches a given point on the fault, the rise time (time it takes to reach the final dislocation value), and the rake (direction) of the dislocation vector $a(\xi,t)$, $\xi \in \Sigma$. Another set of parameters that have to be specified *a priori* are those that define the geometry and position of the fault. In almost all source studies, the fault is considered as a plane (Olson and Apsel, 1982; Hartzell and Heaton, 1983; Jordanovski et al., 1986), or as a combination of planes to fit a more complex fault geometry (Trifunac, 1974). The only unknown parameters remaining to be estimated by inversion are the final amplitudes of the dislocation. Obviously, many of the parameters of the dislocation and of the fault must be defined *a priori*, or otherwise the inverse problem would be highly non-linear. Consequently, the current inversions of the earthquake source mechanism rely heavily on independent information about the earthquake that can help narrow down the set of admissible solutions.

According to the available theoretical and numerical solutions of dynamic earthquake source models, the dislocation function has the following properties: (a) its time derivative, $\dot{a}(\xi,t)$, has a finite jump at time $t = 0$; (b) its time derivative, $\dot{a}(\xi,t)$, does not change sign, so that the dislocation is monotonic function of time (nondecreasing or nonincreasing), and (c) the final offset value is reached during a specified time, called rise time, T_R, and $\dot{a}(\xi,t)$ is allowed to have a finite discontinuity at time $t = T_R$. Physically, this means that a sudden stop in the further dislocation growth is allowed. Condition (b) also assumes the possibility that the dislocation velocity is zero at some instant $t < T_R$, which implies that the dislocation can stop and later on continue to grow. The dislocation behavior at time $t < T_R$ is not uniquely defined by condition (b). Hence, there remains a certain amount of ambiguity. Anderson and Richards (1975) showed that quite different fault motions can produce very similar near-field displacements.

3. *Solving the forward problem by solving eqns (5):* The entire fault is divided into subfaults and, for each subfault, the theoretical displacement at a given point is calculated by convolving the kernel with a dislocation time history of unit amplitude. Many authors have used Haskell or Haskell-like dislocation for this purpose. The main feature of Haskell's model is that the ramp time function for the dislocation allows closed form solution. Further, he assumed that the dislocation occurs instantaneously along the entire fault width W, and propagates with constant velocity v along the fault length. Considering the entire fault as one rectangular plane is obviously a very crude approximation. Consequently, one has to divide the fault into a number of smaller rectangular subfault elements. On each such subfault, the dislocation will be a ramp function in time but with different amplitudes of the final offset. The size of the elements will depend on the desired resolution.

4. *Scaling the theoretical displacements to fit the observed displacements:* The final theoretical displacement is a superposition of the displacements due to the individual subfaults, and the final dislocation amplitudes of the subfaults are the unknowns. The solution is required to be such that the agreement between the observed and theoretical displacements is "best" in some sense. The method of trial and error was often used to solve the problem, until 1974 when Trifunac (1974) showed that direct inversion is possible. In almost all inversion works, the "best" solution is the least squares (LSQ) solution, i.e. the one that minimizes the sum of the squares of the difference between the recorded and theoretical

data. The LSQ approach can be thought of as a minimization of energy type criterion. If only the final dislocation amplitudes are unknown, the LSQ model becomes linear.

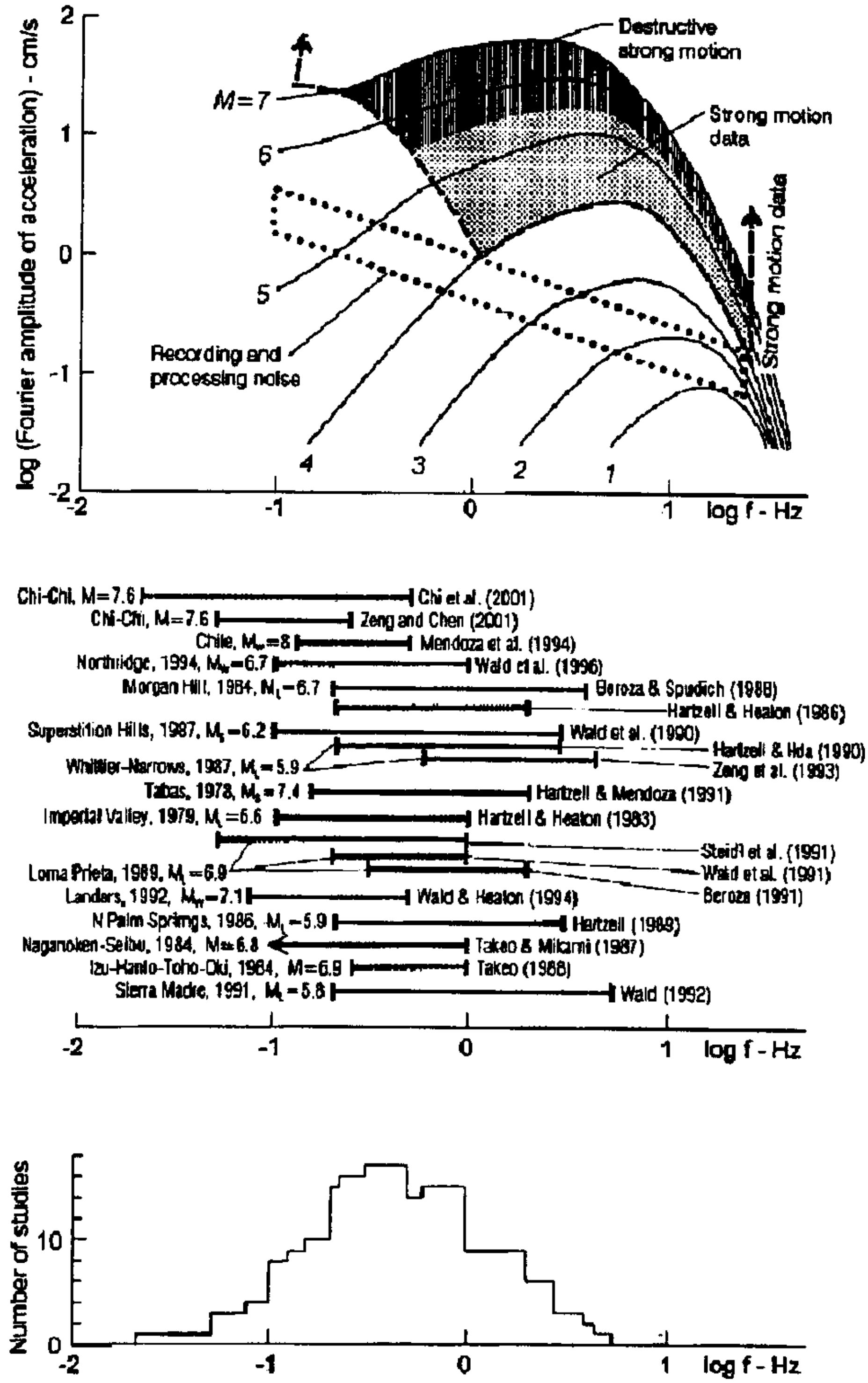

Fig. 2 Top: amplitude-frequency domain of recorded strong motion accelerograph data (also shown is the typical digitization noise for analog records (parallelogram shaped zone with dotted borders)). center: the frequency bands of strong motion data selected for nineteen inverse source mechanism studies (shown by horizontal gray bars); bottom: histogram of the number of studies cited above that use particular frequency bands of strong motion data

Most inverse source mechanism studies are able to resolve only the frequencies of strong motion smaller than 1 to 2 Hz. Furthermore, at small frequencies, the signal-to-noise ratio of most strong motion accelerographs becomes small (Lee et al., 1982; Trifunac and Lee, 1973), so that the useful frequency band that can be used in inverse studies is limited to the range from about 0.1 to 2 Hz. This is illustrated in Figure 2. The top part illustrates the spectral content and amplitudes for the bulk of the strong motion data that has been used so far for source inversion studies (except for the recent 1999 Hector Mine, California, and the 2001 Chi-Chi, Taiwan, earthquakes, most of the strong motion records used for source inversion studies have been recorded by analog accelerographs). The amplitude levels for each magnitude correspond to the average levels predicted by regression models. The dotted parallelogram illustrates the amplitudes of typical digitization and processing noise for analog accelerograms. The central part shows the frequency bands (shown by horizontal gray bars) of strong motion data used in nineteen inverse source mechanism studies. The bottom part shows a histogram of the number of studies cited in the central part that use particular frequency bands of strong motion data. It is seen that approximately 90 percent of all studies used at least the interval between 0.2 and 1 Hz, while only ~ 50 percent of the studies used at least the interval from 0.1 to 2 Hz.

3. Least Square Solution of the Inverse Problem

The displacement vector on the fault surface, $\mathbf{a}$, is in general a function of two spatial variables (ξ_1, ξ_2) and of time, and can be represented as

$$\mathbf{a}(\xi_1, \xi_2, t) = D(\xi_1, \xi_2) d(t) \tag{7}$$

where vector $\mathbf{d} = [D_1, D_2, ... D_J]$ represents the dislocation offset amplitudes on different subfaults, and function $d(t)$ describes the time dependence that is common for all the subfaults. The unknown to be determined is vector $\mathbf{D}$. In most studies, satisfactory spatial resolution has been achieved by dividing the fault into rectangular subfaults. For each subfault, the displacement at an observation point, due to unit dislocations in the ξ_1 and ξ_2 directions, are calculated using the model. The total theoretical response is represented as a linear combination of the contributions from all subfaults. The number of unknowns is equal to twice the number of subfaults, and depends on: a) the desired resolution of the fault elements, b) the stability of the numerical model, and c) the available computer storage.

The error function to be minimized is given by

$$E(\mathbf{D}) = \sum_{i=1}^{I} \int_0^T [f_i(t) - \sum_{j=1}^{J} \phi_{ij} D_j]^2 dt \tag{8}$$

where $f_i(t)$, $i = 1, 2, ...I$ are the observed displacements, $\phi_{ij}(t)$, $i = 1, 2, ...I$; $j = 1, 2, ...J$ is the theoretical prediction of the displacement at the i-th observation channel due to the j-th unit dislocation, I is the total number of observation channels to be matched (based on three components per site), J is the total number of displacements on the subfaults (based on two displacements per subfault), and $\sum_{j=1}^{J} D_j \phi_{ij}(t)$, $i = 1, 2, ...I$ is the total theoretical displacement at the i-th observation channel.

Minimization of the error function $E(\mathbf{D})$ then leads to the following normal equations

$$\mathbf{A}^T \mathbf{A} \mathbf{d} = \mathbf{A}^T \mathbf{f} \tag{9}$$

where matrix A and vector f are given by

$$[\mathbf{A}^T \mathbf{A}]_{jk} = \int_0^T \phi_{ij}(t) \phi_{ik}(t) dt \tag{10}$$

and

$$[\mathbf{f}]_j = \int_0^T \phi_{ij}(t) f_i(t) dt \tag{11}$$

The normal Equation (9) can be derived also from the condition that, at each discrete time, the theoretical displacement must be equal to the recorded time function. This leads to the equation

$$\mathbf{Ad} = \mathbf{f} \tag{12}$$

For a large number of unknowns, the system of the normal equations can be singular. To deal with this problem, some kind of regularization has to be utilized. Olson and Apsel (1982) used dumped least squares to overcome this difficulty, and Jordanovski et al. (1986) applied Tihonov regularization which reduced the problem to minimization of the functional

$$M_\alpha(\mathbf{d},\mathbf{f}) = \|\mathbf{Ad} - \mathbf{f}\|^2 + \alpha^2 \|\mathbf{Ld}\|^2 \tag{13}$$

where α is the regularization parameter or dumping factor, and **L** is a matrix satisfying some physical requirements. Minimization of $M_\alpha(\mathbf{d},\mathbf{f})$ leads to the system of the equations

$$\begin{bmatrix} \mathbf{A} \\ \alpha\mathbf{L} \end{bmatrix} \mathbf{d} = \begin{Bmatrix} \mathbf{f} \\ 0 \end{Bmatrix} \tag{14}$$

In practical applications, for example, if **L** is a discrete version of the $p-$th differentiation operator, then the regularization given by (14) attempts to minimize the $p-$th derivative of **d**, and this has a smoothing overall effect on the solution **d**. A further improvement in obtaining a physically meaningful solution was achieved by imposing appropriate constraints on the dislocation amplitudes (Olson and Apsel, 1982; Jordanovski et al., 1986). Constraints such as the requirement that the dislocations at neighboring subfaults point in the same general direction, has led to satisfactory and physically acceptable solutions.

Jordanovski et al. (1986), and Jordanovski and Trifunac (1990a) also analyzed what they called "time shifting error". This error is due to the fact that the majority of strong motion records used for source inversion so far do not have absolute time, and that the strong-motion instruments do not record the very first strong-ground motion because of the relatively high trigger levels. On the other hand, for the theoretical displacements, the exact time of all arrivals at a station are known exactly. As a consequence, if the theoretical first motions are used as reference time in the LSQ fit, an error is introduced, which is difficult to estimate. The numerical method proposed by Jordanovski and Trifunac (1990a) to reduce this error improved noticeably the solution. It consists of expanding the theoretical displacements into Fourier series, and shifting the displacements in time, which is equivalent to multiplication of the coefficient of the Fourier expansion by a shifting rotation matrix. The optimization then consists of calculating these rotation matrices. This proposed approach and numerical method are effective if the timing error is not too large. It should be noted here that, with the use of digital accelerographs, that are very sensitive and have absolute time and pre-event memory, the "time shift error" can be reduced but not eliminated, because of the errors in the model predictions of the arrival times, based on a simplified model of the geology and of the wave phenomena along the propagation path.

The spatial resolution of the inverted dislocation amplitudes along the fault depends on the number of subfaults used in the inversion. However, while increasing the number of subfaults increases the resolution, it also increases the number of unknowns, and the LSQ system of equations eventually becomes ill-conditioned. Jordanovski and Trifunac (1990a) noted that, for a typical case, the sampling rate of the time histories is not a significant factor affecting the choice of the number of subfaults. The decision on the number of subfaults to be used is mostly influenced by the disk storage and by the numerical stability of the inversion. In typical applications, the dimension of the subfaults is of the order of several kilometers. This does not yield good spatial resolution, which limits the reliability of the estimates of the high frequencies by the model. Jordanovski and Trifunac (1990b) approached this problem by expanding the final offset over the fault in a series of appropriate spatial functions in the ξ_1 and ξ_2 directions, which reduced the LSQ problem to determining the unknown coefficients of the expansion. This approach is known as Galerkin method and it offers several benefits. Firstly, with appropriate selection of the expansion functions, the boundary condition can be satisfied exactly. For example, the dislocation is zero at the boundary of the ruptured fault, and if the boundary reaches the free surface, then the zero-dislocation condition can be relaxed. Secondly, when the coefficients of the expansion are evaluated, the final offset is known at each point on the fault. Thirdly, the high frequencies generated as artifacts of the discretization (the dislocation offset across the boundaries between adjacent subfault is discontinuous) are avoided. Finally, the number of unknowns is greatly reduced leading to a more stable LSQ system.

4. Source Mechanism of Submarine Earthquakes, Slides and Slumps

For inverse studies of the source mechanism of shallow submarine earthquakes, the procedures outlined above remain the same, provided (1) the speed of the spreading dislocation is an order of magnitude greater than the long period speed of tsunami, $\sqrt{gh}$ (Geist, 1999; Satake, 1987, 1989), and (2) the inversion is feasible in terms of only the vertical components of motion of the ocean bottom.

When the speed of spreading of the uplift of the ocean bottom is of the same order as the long wavelength speed of tsunami, $\sqrt{gh}$, the above procedure, again only in terms of the vertical components of motion, can be broken down into two consecutive inversion schemes. In the first stage, maregrams in the near-field surrounding the source can be inverted to determine the time-space evolution of the history of the ocean bottom uplifts. Unfortunately, near-field, open-sea maregrams of tsunami do not exist at present, and in this paper, we only comment that such inversions would be possible. It can be formulated in terms of a Monte Carlo simulation, where the movement of the ocean bottom is represented as a superposition of elementary "uplifts" spreading in one or two directions. Examples of such elementary tsunami sources can be found in Todorovska and Trifunac (2001), Trifunac et al. (2001a, 2001b), and Trifunac et al. (2002a, 2002b, 2002c). Once the vertical motion of the ocean floor is determined in time and space, the second stage of inversion can be initiated, again using only the vertical displacement components, to determine the temporal and spatial evolution of the slip on the causative fault.

STRONG MOTION DATA FOR SOURCE INVERSION STUDIES

The first multi-station recordings of strong earthquake motion that could be used for inverse studies occurred as a fortunate by-product of the ongoing overall deployment of strong motion accelerographs in Southern California, rather than from a carefully planned observation program (Housner and Trifunac, 1967; Trifunac and Todorovska, 2001). The first "arrays" of strong motion accelerographs were installed along lines perpendicular to faults, to measure the attenuation of peak accelerations perpendicular to the fault. Although intuition, as well as analysis (Miyatake et al., 1986; Iida et al., 1988), and practical experience in solving inverse problems suggest that such a configuration is one of the least convenient for strong motion studies, many accelerographs in United States and abroad have been arranged in such linear arrays (Hudson, 1983). If there were prior knowledge available about the mechanism of an earthquake to be recorded near a particular fault, it would be possible to search systematically for an optimum array configuration (Miyatake et al., 1986; Iida et al., 1988). However, at present, limitations associated with: (1) the lack of sufficiently detailed knowledge of the source mechanism and of the foci of future earthquakes, (2) the highly irregular geology between the source and the station, and (3) the lack of adequate funding for the required number and capability of the recording instruments have limited the number and quality of the strong motion records that have become available periodically for inversion studies. The early studies of the Parkfield, California, earthquake of 1966 discuss and illustrate these limitations. Such findings could have been used more systematically to improve the quality of the ongoing strong motion observation programs. However, because of the multiple objectives in deployment of strong motion arrays, the large number of organizations involved in strong motion observation, and the high costs involved, the overall quality of the observation programs has not improved dramatically following the 1966 Parkfield and the 1971 San Fernando earthquakes (Trifunac and Todorovska, 2001). The area covered and overall number of instruments has increased, and this has resulted in more of the earthquakes, that have occurred, to be recorded by multiple stations in the near-field, but the inter-station distance is still of the order of 10 km to several kilometers. This has contributed to increased understanding of the source mechanism, so far, more through the number of studies of different earthquakes than through an in-depth investigation of each recorded event.

Other earthquakes for which inverse studies of the fault slip could be carried out are: Parkfield, 1966, California; San Fernando, 1971, California; Tabas, 1978, Iran; Imperial Valley, 1989, California; Izu-Hanto-Toho-Oki, 1980, Japan; Naganoken-Seibu, 1984, Japan; Morgan Hill, 1984, California; Central Chile, 1985, Chile; Michoacan, 1985, Mexico; North Palm Springs, 1986, California; Superstition Hills, 1987, California; Whittier-Narrows, 1987, California; Loma Prieta, 1989, California; Sierra Madre, 1991, California; Landers, 1992, California; Northridge, 1994, California; and Chi-Chi, 1999, Taiwan. The numerous studies of these earthquakes are cited earlier in this paper.

A recent example of well-recorded strong motion is that of the $M_w = 7.6$, Chi-Chi, Taiwan, earthquake of 20 September 1999. The main shock triggered over 700 strong motion stations across the island, and yielded 65 near-field records (< 20 km from the epicenter). Many detailed inverse source mechanism studies of this event have already been published (Chi et al., 2001; Ma et al., 2001; Zeng and Chen, 2001; Oglesby and Day, 2001; Wu et al., 2001; Wang et al., 2001), and can be used to illustrate the relative differences in the final results. These studies use 15 to 47 strong motion stations surrounding the source, quote seismic moment estimates $M_o = (2 \text{ to } 4) \times 10^{27}$ dyne-cm, and consider source duration of 28 to 40 s. The estimate(s) of the maximum dislocation amplitudes are from 15 to 20 m, of rise times are 6 to 8 s, of average slip velocity is 2.5 m/s, of peak slip rate is 4.5 m/s, and of the average dislocation spreading velocity are 2.5 to 2.6 km/s. The calculated distributions of slip vectors across the fault show some similarity, but those are very different when compared in detail. Clearly, the resolving power of the state-of-the-art inversion methods is still restricted to long periods of strong motion, longer than 2 s in these examples.

The first strong motion array, with absolute timing and designed specifically for the inverse analyses of source mechanism, was installed by Trifunac in Bear Valley, California, in 1972 (Trifunac and Todorovska, 2001). Few weeks later, it recorded excellent data, which served to compare near, intermediate and far-field inverse analyses of the source mechanism (Dielman et al., 1975). To enable recordings and analyses of the propagating dislocations during moderate and large earthquakes, Trifunac and Hanks installed the 59 station San Jacinto array in 1973-74, along the active portion of the San Jacinto and San Andreas faults in Southern California (Trifunac and Todorovska, 2001). In 1979, this array contributed excellent data for inverse studies of the Imperial Valley, California earthquake. During subsequent years, this array recorded numerous earthquakes in this highly active area (e.g., Superstition Hills, 1987; Whittier-Narrows, 1987; Landers, 1992; Big Bear, 1992; Northridge, 1994; Hector Mine, 1999). In 1979-80, Trifunac and co-workers installed the two-dimensional strong motion array with absolute time in the Greater Los Angeles Metropolitan area (Trifunac and Todorovska, 2001) to study the source mechanism of the earthquakes in this area and the three-dimensional wave propagation through the sediments in the Los Angeles basin. The data this array recorded during the 1987 Whittier-Narrows and 1994 Northridge earthquakes is unprecedented.

Many of the deficiencies of the current strong motion programs in the U.S. (e.g., absolute time, high sensitivity and dynamic range, real-time telemetry and coverage of larger areas) will be addressed by the Advanced National Seismic System (ANSS) program, launched by the U.S. Geological Survey and partner seismological institutions. The deployment of the southern California element of ANSS (TriNet) has almost been completed, and will probably be one of the elements with the highest density of stations. TriNet is a high-tech and somewhat denser version of the San Jacinto Array and the Los Angeles and Vicinity Strong Motion networks combined, but with inter-station distance still of the same order of magnitude as for the Los Angeles and Vicinity Strong Motion network. So far, and while still in the process of implementation, TriNet has recorded only one significant event, the 1999 Hector Mine earthquake, for which finite source modeling studies have not yet been published.

CONCLUSIONS

The review of the above-discussed source mechanism studies suggests the following general conclusions:

1. At present, only the gross features of the distribution of dislocations over the fault area can be determined from strong motion data for shallow faults (source depth less than about 20 km) and for periods longer than about 1 s.
2. To improve the resolution of the motion on the fault, either many more strong motion records must be available around the causative fault, or a very detailed description of the three-dimensional geology between the source and the recording stations must be used in the inverse analyses.
3. All the strong motion records used in the inversion must have absolute time to determine the wave arrival times with accuracy better than 0.01 s (Jordanovski and Trifunac, 1990a).
4. More general dislocation models are being developed, which allow variable speed of spreading dislocations, repeated slips, smoother representation of the dislocation amplitudes, and more

complicated fault geometries. Many of these improvements can now be implemented because of the larger memory and speed of the digital computers.

5. Analogous inverse studies of the distribution of slip on submarine faults are possible, but will require recording of tsunami in open-sea and surrounding the causative fault (or slump or a slide).

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