Imaging Earthquake Source Complexity

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Abstract
Recent improvement in analysis methods, computational resources, and seismological and geological observations enable us to image earthquake source in detail. This paper reviews various studies for each component of source imaging: model parameterization, data, theoretical wave calculation, and optimization. Differences in analysis in each of these aspects can lead to significant differences among models for the same event. The comparison between two recent large earthquakes demonstrates that the quality and quantity of data set is critical in the source imaging problem. While the models of 1999 Izmit, Turkey, earthquake show wide variation, those of the 1999 Chi-Chi, Taiwan, earthquake have common features. This is attributable to the fact that the data for the Izmit was sparse, while the data for the Chi-Chi earthquake was the most complete for any earthquake in history. Improved knowledge of source complexity and diversity of rupture behavior is leading to important new areas of study. As examples, we review current research problems concerning the frequency dependence of the earthquake source, imaging of dynamic rupture properties, and characterization of source complexity and scaling.

1. INTRODUCTION
This paper reviews studies of earthquake source imaging using seismic waves. There are varying degrees of source characterizations and research frontiers are different for each degree. For example, source dimension and rupture directivity are easily resolved for large events with ample near-field data, but they can still be problematic without such data and new imaging methods are being developed to retrieve these parameters for large events recorded teleseismically [e.g., McGuire et al., 2000] or small, regionally recorded events [e.g., Imanishi and Takeo, 2002]. The assumption that earthquakes occur as shear slip on faults is not always valid for events around volcanoes [e.g., Kanamori et al., 1993; Dreger et al., 2000], and studies of earthquakes in such areas must account for this. Rapidly resolvable earthquake features are another topic that has recently received increased attention [Kanamori et al., 1997; Dreger and Kaverina, 2000] due to its usefulness in guiding emergency response after large earthquakes. In this paper we focus on methods that have been used to reveal complexity of ordinary earthquakes, which occur as shear slip across faults. For a more general review, see Beroza [1995] and Yoshida [1995].

The most detailed images of earthquakes have been obtained using data recorded near to the earthquake source, where Green's functions from different parts of the fault are strongly variable. At short distances, most seismic instrumentation is driven off-scale, and only instruments designed to stay on scale during strong ground motion record the motion of the ground with fidelity. Attempts to image the earthquake source with strong motion data date back to studies of the 1966 Parkfield earthquake [Aki, 1968; Haskell, 1969], which parameterized the finite-source with a handful of parameters and constrained them using a single seismogram. Subsequently images of the same earthquake were developed that allowed for spatially variable slip [Trifunac and Udwadia, 1974]. Over much of the last 35 years, strong ground motion studies have been data limited, and progress in the field has been impelled by the occurrence of earthquakes that are exceptionally well recorded relative to the earthquakes that preceded them. The 1979 Imperial Valley, California earthquake is a good example.

The 1979 Imperial Valley earthquake was recorded by dozens of near-source strong motion instruments in the United States and Mexico. It stimulated a great deal of activity in strong motion modeling. The slip distribution of this earthquake was imaged by a number of investigators [Olson and Apsel, 1982; Hartzell and Heaton, 1983;
Archuleta, 1984]. During the following two decades, related approaches have been applied to many earthquakes in California and Japan. While less widely used, a similar approach is also applicable to teleseismic data for large global events [e.g., Mendoza, 1993]. The method is currently useful for global earthquakes of $M \sim 7$ or larger, regional events of $M \sim 5$ or larger within dense strong-motion seismometer networks, and $M \sim 3$ or larger events within dense high-sensitivity seismometer networks. The framework of finite-source imaging is not much different from what it was in 1980’s, but the scale of computation has grown and improved techniques have been imported from statistics and optimization. The use of recently developed data such as Global Positioning System (GPS) and synthetic aperture radar (SAR) interferometry can help to decrease trade-offs between the temporal and spatial distribution of earthquake slip and thus increase the resolution and stability of imaging. In this paper we review methods and data in section 2.

The field of imaging the earthquake source tends to get driven by large, well-recorded events. A number of models have been presented for the 1989 Loma Prieta, California, earthquake [Beroza, 1991; Hartzell et al., 1991; Steidl et al., 1991; Wald et al., 1991; Zeng et al., 1993; Horton, 1996], the 1992 Landers, California, earthquake [Cohee and Beroza, 1994; Wald and Heaton, 1994; Cotton and Campillo, 1995; Hernandez et al., 1999], the 1994 Northridge, California, earthquake [Dreger, 1994; Hartnell et al., 1996; Wald et al., 1996; Zeng and Anderson, 1996], and the 1995 Kobe, Japan, earthquake [Horikawa et al., 1996; Ide and Takeo, 1997; Kakehi et al., 1996; Sekiguchi et al., 1996a, 2000; Wald, 1996; Nakahara et al., 1999; Cho and Nakanishi, 2000]. Although the models for each event have similarities to some extent, significant differences are often present. In section 3, we will compare and contrast models of two recent events, the 1999 İzmit, Turkey earthquake [Yagi and Kikuchi, 2000; Bouchon et al., 2002; Delouis et al., 2002; Sekiguchi and Iwata, 2002] and the 1999 Chi-Chi, Taiwan, earthquake [Chi et al., 2001; Ma et al., 2001; Wu et al., 2001; Zeng and Chen, 2001].

In section 4 we introduce some recent topics in this field: what is the difference in source images determined from low and high frequency waves? what aspect of dynamic behavior of the seismic source can be resolved from seismic source imaging? among numerous possible models, what are the common properties of earthquake sources and how do these complexities scale, or not, with earthquake size? These are some of the questions that are subjects of current research.

2. METHOD AND DATA

2.1 Fault Models and Parameters

Earthquake faults are usually expressed as a plane or a set of planes. The location and orientation of the planes are based on the hypocenter of the mainshock and aftershocks. If good hypocentral locations are not available, for example in the case of teleseismic events or offshore events, the orientation is based on the focal mechanism or moment tensor. When surface offsets are available they can constrain the fault geometry in detail. In the study of the 1992 Landers earthquake [Cohee and Beroza, 1994; Wald and Heaton, 1994; Cotton and Campillo, 1995], a set of plane faults with varying azimuth of about 30 degrees are assumed along the surface offsets of Camp Rock/Emerson, Homestead Valley, and Johnson Valley faults. The fault planes can be more complex. In the case of the 1997 Kagoshima, Japan, earthquake ($M_s$ 5.9), the fault models required two perpendicular planes [Miyake et al., 1999; Horikawa 2002] based on the L-shaped aftershock distribution. When strong-motion records are available in the source area, shear wave polarities are useful to determine more precise fault location than that suggested from the aftershock distribution [Sekiguchi et al., 1996b]. Although fault offsets provide clear evidence of the fault location at surface, the shape of their deep extension can be ambiguous even with precisely determined aftershock locations. Assuming three-dimensional boxes instead of planes, Cho and Nakanishi [2000] determined three-dimensional moment release distribution and showed the non-planar spatial pattern of moment release coincides with the non-planar spatial distribution of precise aftershock locations. Although such an approach introduces
a large number of unknowns, requires well-resolved structure, and a high-quality data set, it may be the only way to analyze events for which faulting is truly complex and impossible to characterize by simple planes.

An assumed fault is usually divided into a number of subfaults. The lower part of Figure 1 illustrates typical subfault parameterizations. Some studies use only one point for each subfault while others arrange many points. In the first case, the calculated time function for each subfault is assumed to contain the effect of propagation within the subfault and spatial smoothing is applied to prevent aliasing. In the second case rupture front propagation within the subfault is accounted for by the timing of the multiple point sources and spatial smoothing is not required to prevent aliasing; however, this latter approach requires much more computation. The first arrangement corresponds to a delta function in space, while the latter case corresponds to a boxcar function in the limit of short spacing. Other spatial basis functions, e.g., linear or cubic splines, are possible as well.

When good data and well-known structure are available, we may determine slip direction by determining relative slip amplitude in two directions through linear inversion or by solving for rake angles using a nonlinear inversion. While difficult to resolve, time-dependent slip angle rotation is an important parameter, since it can be used to constrain the absolute stress level [Spudich, 1992; Guatteri and Spudich, 1998].

The temporal aspects of slip behavior—the shape of the slip function and the rupture propagation velocity—are difficult to resolve, but they provide important information about the dynamic faulting process since they are influenced by the local energy balance and fault constitutive relation, including the surface fracture energy. The way in which the temporal dependence of slip is parameterized is critical to source imaging because it determines the nature of both the solution and the inverse problem. A multiple time window method was developed by Hartzell and Heaton [1983] and this parameterization is frequently used [e.g., Hartzell and Heaton 1986; Wald et al., 1991]. In this approach the time function is expressed by successive overlapping time windows, which allows for the possibility that a subfault may rupture more than once [Olson and Apsel, 1982], and be considered to be a linear summation of temporal basis functions [Yoshida, 1992]. The start time of the first time window is usually determined by a propagation time from the hypocenter at a fixed velocity, which prohibits acausal rupture propagation. As long as these functions are fixed in the time domain, the problem of determining their amplitude is linear. In this approach the duration of the slip function (rise time) is adjusted discretely depending on the arrangement of basis functions. Local rupture propagation velocities are also changed only discretely and the assumed propagation velocity is presumed for the rupture within a subfault. It is possible to make a flexible model that can represent variable rise time and local rupture propagation by using a finer system of subfaults and time windows, but this approach has not been widely used because it requires a very large number of parameters.

Other parameterizations assume the functional form of the time dependence of slip using small number of parameters [Archuleta, 1984; Yoshida, 1986; Takeo, 1987; Beroza and Spudich, 1988]. A dynamically realistic function is easily expressed in this way, and the rise time can be parameterized explicitly. Since the timing of this function can also included as a parameter, changes of local rupture velocity are easily modeled. This approach leads to a nonlinear inverse problem in the time domain. An alternative approach, proposed by Olson and Anderson [1988], solves for variable rise time in the frequency domain as a linear inverse problem, though it is only applicable for fixed rupture velocity.

There are other approaches to determine subevents in source area rather than determining spatio-temporal functional form of the slip function. Kikuchi and Kanamori [1982, 1991] developed an iterative forward method to extract subevents, each of which is represented as a point source, from observed waveforms. The subevents may be located in a plane as a fault system [Kikuchi and Fukao, 1985]. However, it should be noted that the assumption of discrete subevents is not always valid and Ihmélé [1998] showed that teleseismic datasets can be easily misinterpreted using the subevent parameterization, especially for a fault with small aspect ratio. This method is extensively used for many earthquakes recorded at teleseismic distances [e.g., Kikuchi and Kanamori, 1994] and some recent results using this approach have been reported quite rapidly, within a few days, via the

Extracting information about the seismic source that goes beyond the centroid and moment-tensor when only teleseismic data is available is a difficult process. The seismic moment and centroid location represent the zeroth and first order terms in a Taylor series expansion of the moment-release distribution, respectively. The simplest description of an earthquake source that contains information about the spatial and temporal extent of moment release is the extension of this Taylor series to include the second order terms, i.e. the space-time variances of the moment-release distribution [Backus 1977a, b; Okal, 1982; Doornbos 1982a, b; Silver and Jordan, 1983; Gusev and Pavlov, 1988; Pavlov 1994; Das and Kostrov, 1997]. The “second moments” describe the length, width, duration, and propagation velocity of the moment-release distribution in a weighted average sense. McGuire et al. [2001] showed that by linearizing around an estimate of the zeroth and first moments (i.e. the CMT parameters) it is possible to solve for the 2nd degree moments using anomalies in the low-frequency amplitudes of surface and body waves. Since about 1995, the various global seismic networks have become dense enough that these parameters can be routinely estimated for events larger than $M_w \sim 7.0$.

2.2 Data

Seismograms used for imaging source complexity are categorized into near-field and far-field, and those at intermediate distances are rarely used. Near-field data is recorded within a few multiples of the source size and usually less than 150 km where $P_n$ often becomes the first arrival in continental crust and the interpretation of waveforms becomes quite difficult. Far-field waves are recorded at azimuthal distance between 30° and 100°, where direct arrival of body waves are separated from other arrivals and easily interpreted. The true definition of far-field depends on the source size. When $L$, $\lambda$, and $r$ are source dimension, wave length and hypocentral distance, the condition is written as $L^2 \ll \lambda r$ [Aki and Richards, 1980]. For an event with length and corner frequency of 1 km and 1 Hz (typical for M4), the hypocentral distance of 30 km is enough to analyze waves at frequencies higher an order of magnitude higher than the corner frequency. Hence, we use global records as far-field data for large earthquakes and local records as near-field data for large events and as far-field data for small and moderate earthquakes.

The most important data for detailed imaging of large earthquakes are near-field strong motion seismograms. Recently, dense regional networks of strong-motion seismographs have been installed in Japan [K-NET: Kinoshita, 1998; KiK-NET: Aoi et al., 2000] and Taiwan [TSIMP: Liu et al., 1999]. The development of recording instruments with high dynamic range makes it possible to use high-sensitivity records for detailed source analysis. Data from the Trinet array, downhole strong motion instruments within the proposed Plate Boundary Observatory (PBO), and the proposed 7000 strong motion instruments of the Advanced National Seismic System (ANSS) should ensure rich strong motion data sets from future significant earthquakes in the United States as well. The focus of the ANSS initiative on urbanized regions and the limited extent of borehole observations within the PBO, however, means that earthquakes occurring in rural areas of the United States, such as the 1999 Hector Mine earthquake, will continue to be relatively poorly recorded.

The data used in the source imaging inversion problem are usually displacement or velocity while most original records are either velocity or acceleration. Therefore the original data usually must be bandpass filtered and numerically integrated. The choice of displacement or velocity and pass band of filter determines what aspects of the waveforms are emphasized and affects the resolution of the inversion [Mendoza and Hartzell, 1988; Ide, 1999]. Even when the pass band of filter is clarified, it is not always apparent what frequencies of the original data are explained by the model in time domain, although Cohee and Beroza [1994] explored that issue using large aftershocks of the 1992 Landers earthquake. There are some studies that analyze seismograms in the frequency
domain [Olson and Anderson, 1988; Cotton and Campillo, 1995] and frequency domain analysis can provide important independent information. In waveform inversion, the large amplitude waves of low frequency tend to be well explained and while high-frequency waves of small amplitude are often neglected. Ji et al. (2002a) proposed a method to make a model that can explain high-frequency waves as well as low-frequency waves, using a wavelet transform. If we had a complete description of the prediction error, the choice of whether to model the velocity or the displacement fields would not matter. Clearly, a better accounting of errors in weighting the inverse problem and a description of the uncertainties in the resulting rupture model is desirable.

Except for accelerometers that have flat response down to zero frequency, all seismometers have a lower frequency limit. In realistic conditions, noise limits the lowermost frequency even for broadband instruments. Knowledge of static displacement could be quite useful for constraining the total seismic moment. Towards the same end, various geodetic data have been used: triangulation [e.g., Yoshida and Koketsu, 1990], leveling [e.g., Horikawa et al., 1996; Yoshida et al., 1996], Global Positioning System (GPS), [e.g., Wald and Heaton, 1994; Horikawa et al., 1996; Yoshida et al., 1996], and synthetic aperture radar (SAR) interferometry [Hernandez et al., 1999; Delouis et al., 2002] in source imaging. The analysis of the 1999 Hector Mine earthquake provides a good example of the importance of including geodetic data in finite-fault inversions [Ji et al., 2002a, 2002b; Kaverina et al., 2002]. Numerous models of static slip distributions based on geodetic data exist, though we do not refer them in this paper. When combining different data sets, such as seismic and geodetic data, it is important to account for the relative uncertainty in each. As long as data weighting schemes used in modeling strong motion data are largely ad hoc, it will be difficult to obtain optimal contributions from multiple data sets when developing an image of the earthquake source.

2.3 Green's Function

There are various methods and computer algorithms for Green's functions calculation in 1D layered structure [Kennet and Kerry, 1979; Bouchon, 1981; Yao and Harkrider, 1983; Koketsu, 1985; Takeo, 1985; Saikia, 1994]. The combination of ray theory and isochrone integration [Bernard and Madariaga, 1984; Spudich and Fraser, 1984] is also effective in the high-frequency, near-source approximation. In practice, however, waveform calculations for layered Earth structure only provides reasonable Green’s functions only at frequencies lower than ~1 Hz. Even at lower frequencies, time alignment is often used to reduce the error in the Green's function calculation [Zhu and Helmberger, 1996].

If three-dimensional structure is available, Green's functions can be computed for it numerically. Graves and Wald [2001] carried out inversion simulation using both 1D and 3D structures. As expected, accurate 3D Green's function can increase the resolution of model; however, if the assumed 3D structure is different from the true structure, the results can be misleading. Thus, although 3D Green's function will ultimately improve seismic source imaging, incorporating it into source inversions will require a careful examination of the validity of the assumed structure and the predictive value of Green’s functions that are determined for it.

For analysis of higher frequency waveforms, it may be essentially impossible to predict waveforms and much more practical to use the empirical Green's function (eGf) method [Hartzell, 1978], in which small earthquakes are used as Green’s functions for the analysis of larger earthquakes. Of course they are not true Green’s functions because they are the response of the Earth to a dislocation, rather than a directed point force. The approximation will also only be valid at frequencies below the corner frequency of the small event. Nevertheless, the eGf approach is a powerful one and variations of it have been applied to earthquake source imaging for large earthquakes such as the 1983 Japan-Sea earthquake (Mw 7.7), [Fukuyama and Irikura, 1986] and small events of M 3-4 [Mori and Hartzell, 1990; Courboulex et al., 1996; Fletcher and Spudich, 1998; Hellweg and Boatwright, 1999; Ide, 2001]. The eGf method has also been applied to high frequency recordings of small events [e.g., Frankel et al., 1986]. In the eGf method, we assume that the source duration of small event is negligible and that
all the effects of path and site are equal for both small and large events; however, a single eGf event will likely not be a good approximation to the actual Green's function over the entire source area of a large event. In this case we need a set of eGf events each of which represents a Green's function on a part of the fault plane of the large event [Fukuyama and Irikura, 1986]. Since finding eGf events can be a difficult task, the application of eGf analysis to large events has been limited.

2.4 Solving the Problem

When the data is connected to a model by linear equations, we can solve this problem using the many techniques of linear inverse theory [e.g., Menke, 1984] and obtain a unique solution with an associated estimate of its uncertainty. The validity of the uncertainty estimate, of course, may only be as good as the uncertainty estimates for the data or the Green’s functions used to fit the data. When a problem is nonlinear, however, uniqueness is no longer guaranteed and the estimation of error becomes yet more difficult. The most popular and traditional way to solve nonlinear problems is through linearization, which was introduced to source imaging in the late 1980s [e.g., Yoshida, 1986; Takeo, 1987; Beroza and Spudich, 1988; Hartzell, 1989]. In the case of a multi-modal problem like seismic waveform inversion, however, it is often hard to find the global minimum by a simple linearized scheme as Gauss-Newton methods or the Levenberg-Marquart method. This might seem to argue against nonlinear parameterizations of the inverse problem; however, there is no guarantee that a linear parameterization will yield a more reliable answer. If the assumptions that lead to a nonlinear parameterization are better than those that lead to a linear parameterization, then the solution to the nonlinear problem may be a closer approximation of reality.

In the 1990s several techniques for obtaining global solutions to nonlinear problems were applied to earthquake source inversion. The simulated annealing (SA) method is a sophisticated version of the Monte Carlo method that includes a mechanism to escape local minima based on thermodynamics. Hartzell and Liu [1995] and Hartzell et al. [1996] used this and the simplex method for source imaging of the 1992 Landers earthquake and the 1994 Northridge earthquake, respectively. Courboulex et al. [1996] used SA to analyze the source process of a small (M3) earthquake. Genetic Algorithms (GA) is another popular global search method that simulates the process of the natural population's evolution according to the natural selection principle of survival of the fittest. It has been successfully used to generate a composite source model for the 1994 Northridge earthquake [Zeng and Anderson, 1996].

By minimizing the $L_2$ norm when solving an inverse problem by least squares, we implicitly assume the errors follow a Gaussian distribution. Real data, however, often have outliers that are not described by Gaussian statistics, in which case least squares solutions may give too much weight to such data. The use of $L_1$ norm instead of $L_2$ norm (least square) can provide a more robust solution and is more resistant to the effect of outliers [e.g. Menke, 1984], but it makes the problem nonlinear. There have been several applications of $L_1$ norm minimization in source imaging [e.g., Das and Kostrov, 1990; Hartzell et al., 1991; Ihmle, 1998]. The comparison between the results using $L_1$ and $L_2$ norm minimization can be informative to illustrate what part of the solution is dependent on this assumption [Hartzell et al., 1991].

The solution of the linear source inverse problem with a large number of model parameters requires some kind of constraint. A positivity constraint that prohibits slip in reverse direction is physically reasonable under high pressure and high dynamic friction. This constraint is frequently assumed by using non-negative least-squares (NNLS) algorithm [Lawson and Hanson, 1977]. Application of a positivity constraint can dramatically increase the resolution provided the constraint is valid (so-called super resolution). Minimum norm and minimum roughness constraints have also been used to regularize the inverse problem. In source imaging constraints to fix the total seismic moment of slip model to some independently estimated value [e.g., Das and Kostrov, 1990; Ji et al., 2002b] or to minimize the seismic moment [Hartzell, 1989] have also been used. Introducing such a constraint
introduces another sort of uncertainty; we need to determine the appropriate weight of a constraint using hyperparameters. The same problem arises when we use different types of data, such as waveform and GPS; relative weight should be determined by some objective method. Often an empirical approach is used to yield a solution that is, by some subjective measure deemed “reasonable”. An alternative approach is to determine the “optimal” hyperparameters using a Bayesian model and Akaike’s Bayesian Information Criterion (ABIC), [Akaike, 1980], which treats constraints as prior information. ABIC has been successfully introduced into earthquake source inversion [e.g., Yoshida, 1989; Ide et al., 1996; Sekiguchi et al., 2000]. It should be noted that previous applications have an error in the case of multiple constraints [Fukahata et al., 2002], although it is not significant difference if there are many more data than constraints. We also emphasize that the objectively estimated hyperparameter is the optimal one only for $L^2$ norm minimization and depends upon the validity of the prior assumption, which may not be easily confirmed.

3 COMPARISON OF MODELS OF RECENT EARTHQUAKES

3.1 1999 İzmit, Turkey, Earthquake

The İzmit (Kocaeli) earthquake ($M_c$ 7.4) occurred on August 17, 1999, on the North Anatolian fault system in northwestern Turkey. The fault motion was almost pure right-lateral strike slip and surface offsets of as much as 5 m was observed along fault segments whose total length was about 100 km [e.g., Barka et al., 2002]. Accelerograms were recorded at 21 stations of Boğaziçi University and Earthquake Research Department of the General Directorate of Disaster Affairs, and rapidly made available to the public via the Internet. These include records at five stations within 20 km from the surface trace. A station at about 4 km from the surface trace (SKR) recorded curious waveforms with had an S-P time substantially shorter than the S-P time predicted for the hypocenter. This has been interpreted as a result of super shear rupture propagation [Ellsworth and Çelebi, 1999]. In addition to strong motion data, data for this event also included surface offsets, GPS, and InSAR observations, which are useful for constraining the fault geometry, the slip distribution and the total seismic moment.

Four research groups have developed fault slip models as summarized in Table 1 [B: Bouchon et al., 2002; D: Delouis et al., 2002; S: Sekiguchi and Iwata, 2002; Y: Yagi et al., 2000]. Delouis et al. [2002] also presented an independently estimated slip model for each data type and the result in each case is quite different. We show only the result of their joint inversion as model D. The B, D, and S models assumed fault planes based on the surface fault trace. The B and D models used simulated annealing (SA) to solve the nonlinear inverse problem in which the timing of source time functions are model parameters. The S and Y models applied non-negative least squares (NNLS) with ABIC to estimate appropriate smoothing constraints. The B and S models included propagation within each subfault while that information is not available for other models. It should be noted that B model is the average of 10 models that explain the data well in the SA solution.

Figure 2 shows the slip distributions for these models. Since the assumed hypocentral locations are slightly different, we aligned them at the latitude of station IZT. Some differences arise from model configuration. Since the subfaults of model D are rectangular of 7.5 km x 4.5 km, the slip distribution tends to be elongated in the strike direction. The model area of model D is larger in space and time than the others. There is slip100 km east of the hypocenter and 30-50 s after the origin time, which is also reported in a study of far-field waveforms [Gülen et al., 2002; Li et al., 2002]. The surface offsets also suggest the existence of slip in this region. The difference in the model area partly explains the difference in seismic moment between the models. The usage of near-field velocity instead of displacement in S model may lead to a less smooth slip distribution and more uncertain seismic moment value although they use the ABIC, which should lead to an objective degree of smoothness.
Even after taking such differences into account, we still find substantial unexplained differences between the models. For example, slip of 4-6 m at 60 km east from the hypocenter in model B is not present in other models. Slip concentration near the hypocenter is characteristic only of the D model. Large slip to the west of the hypocenter is common to all the models, but the depth is not well constrained. In the D and Y models, the slip area from the hypocenter to the station SKR is almost continuous while there is little slip in this area in the B and S models. To explain the short S-P time at SKR, the B model and S-model include super shear rupture propagation and slip triggered by the P wave, respectively. The D and Y model have no such feature and cannot explain the short S-P time. It is worth noting that there is some concern about the record at SKR in that it the horizontal component parallel to the fault did not record the earthquake. Whether this affected the horizontal component perpendicular to the fault, on which the inference of super shear rupture is based, is unknown.

The characteristic rupture dimensions derived from the 2nd moments of the İzmit rupture models are given in Table 2. In all cases, the average velocity of the instantaneous centroid is along-strike to the east at about 2-4 km/s. This term has the largest effect on the wavefield of any of the higher order terms and appears to be somewhat well resolved in terms of direction and to a lesser extent magnitude [see also Clevede et al., 2002]. The variation in the magnitude of characteristic rupture length, $L_c$, 44-86 km, indicates a lack of resolution of the spatial extent of the slip distribution, and probably results from the variations in datasets and smoothing constraints between the studies. The D model study has a larger characteristic duration and rupture length, as well as a more unilateral $v_0/v_c$ ratio owing to the large late subevent to the east.

In conclusion, the differences in the model suggest that the data for this earthquake may not to be adequate to produce a robust image of the earthquake source. The assumption of the fault plane, velocity structure, and, especially, the choice of time alignment between data and synthetic waves all strongly affect the result. If super shear rupture occurred in this earthquake, then the usual assumption that the first arriving $S$ wave is from the hypocenter, will not be valid for all stations and the alignment of waveforms becomes complicated even with precise absolute time. Evidently, there is not enough data for this earthquake to overcome the differences introduced by different modeling assumptions made in the four studies of this event. In that sense, the source image of this earthquake is not robust. We can contrast this situation with that for a very well-recorded earthquake of similar size, the 1999 Chi-Chi event.

### 3.2 1999 Chi-Chi, Taiwan, Earthquake

The Chi-Chi earthquake (Mw 7.6) occurred on September 20, 1999 (UT), one month after the İzmit event, in central Taiwan. This event is located in the accretionary zone between the Eurasia plate and the Philippine Sea plate, and the Harvard CMT solution shows shallow (27˚) dipping thrust movement dipping to the east. Surface offsets appeared along an 85 km segment of the Chelungpu fault striking primarily in a north-south direction. The offset has a maximum of about 8 m at near the northern end of the fault, where the strike of the surface rupture turns to the east. The seismic waveforms were recorded by strong-motion seismometers at 441 stations of the Taiwan Strong-Motion Instrumentation Program, 60 of which are within 20 km of fault rupture [Lee et al., 2002]. As a result the Chi-Chi earthquake can be considered to be by far the best-recorded large earthquake to date. These mainshock data were distributed by Lee et al. [1999] by December of 1999 and stimulated the study of various aspects of this important event. In addition to the strong motion data, there are also dense GPS observations [Yu et al., 2001] in this area, which strongly constrain the static displacement field.

Since the northern edge of this event is quite complex and some data are in this area, we need to account for this complexity in the fault parameterization, particularly for strong motion data recorded in this area. For example, station TCU068 is located on the hanging wall within 1 km from the surface trace of the east-west striking part of the surface rupture and it would be impossible to model this data with a north-south striking fault. For this reason, studies adopted more than one fault arrangement for this earthquake, although overall characteristics of the
different models are similar. In this review, we compare only the simplest model in each study. The models are summarized in Table 3 (C: Chi et al., 2001; M: Ma et al., 2001; W: Wu et al., 2001; Z: Zeng and Chen, 2001). Ma et al. (2001) have a two-fault model and Wu et al. (2001) have two and three fault models in addition to the models shown here.

The Z model is constructed using a Genetic Algorithm to solve the non-linear problem in which timing and rise time of the source time function and rake angle for individual subfaults cause nonlinearity. The validity of this model is measured by cross correlation between data and synthetic waveforms and the $L_1$ norm of the static displacement, which is also a source of nonlinearity. The other three models are solved using a linear inversion with positivity constraints (NNLS), though this is not explicitly stated in the paper for models C and M. For these three models, smoothing constraints are also introduced and the weight is determined by the ABIC (W model) and to match the total seismic moment (C model).

Figure 3 compares the final slip distribution and the rupture propagation pattern for these models. Although we cannot compare the smoothness because the measure of smoothness is different in each case, there are many common features in this figure. The total slip distributions consistently show the largest slip of about 20 m at the shallowest region 40 km north of the hypocenter. Wide areas of high slip extend below this region. Slip near the hypocenter is about 10 m. In the temporal change figure, we see 1) shallower slip area than the hypocenter during first 10 s, 2) two patches of slip in 20-25 s, and 3) termination by 30 s. The Z model has these features, too.

In contrast to the Izmit earthquake, the 2nd moments of the various Chi-Chi rupture models are very similar in terms of rupture length, width, duration, and propagation velocity (Table 4). In all models the average velocity of the instantaneous centroid points roughly along strike to the north and is around 2 km/s in magnitude. The overall spatial and temporal dimensions, $L_c$, $W_c$, and $\tau_c$, are very similar indicating the high resolution of the dataset. The C model has a longer duration of slip at the large asperities causing a longer overall $\tau_c$.

The fact that clock of the seismometer is not accurate and that structure in this area is not simple suggests that there ought to be significant modeling error and thus we might expect significant differences in the models; however, there are substantial similarities in these models, which were independently derived by independent research groups with different assumed velocity structure and different time alignments of the data. This implies that the quality and dense distribution of the data overcome the uncertainties that the differences in the modeling introduced, so in that sense the image of this earthquake is robust. These models are examples of well-constrained source images.

4 CURRENT TOPICS

4.1 Frequency Dependence of Source Image

Most source imaging studies have resolved fault slip behavior at relatively low frequencies ($f < 1$ Hz); however, the source image is dependent on the frequency range used in the analysis. Using high-frequency waveform data has the potential to reveal more details of source process [e.g., Mendoza and Hartzell, 1988; Ide, 1999] and in addition, the mechanism by which damaging high frequency waves are generated. While low-frequency waveforms are coherent and linear summation of wave amplitude is possible, the phase of high-frequency waveform is usually incoherent and linear summation (under the assumption of random phase) can be assumed for the squared waveform. Using this assumption, Zeng et al. [1993] determined the spatial distribution of high-frequency wave generation for the Loma Prieta earthquake. They took mean squared displacement waveforms higher than 5 Hz for 0.6 s time windows and inverted this data into "energy radiation intensity," which is the product of the fault element area and the square of fault slip. Green's functions were calculated using the
The isochrone method of Spudich and Frazer [1984]. They found large energy radiation intensity near the hypocenter and around the high slip areas determined in the independent inversion of low frequency (0.3-2.0 Hz) waveforms.

This work is important, but there are some potential difficulties with it: viz., 1) neglecting the effect of complex structure and coda waves, 2) use of displacement data, which due to the rapid roll-off of high frequency displacement with increasing frequency, results in a very narrowband measure of the strength of the radiation, and 3) ambiguity of the physical meaning of the model parameter. The first two were solved by Kakehi and Irikura [1996], who use the empirical synthetic waveform calculation method of Irikura [1986] and root mean square of acceleration waveforms between 2 and 10 Hz. They applied this method to the 1993 Kushiro-Oki, Japan, earthquake (Mw 7.6) and found that the areas of high energy radiation intensity correspond to the edges of slip determined in low frequencies by Takeo et al. [1993]. This method was also applied to the 1993 Hokkaido-Nansei-Oki, Japan, earthquake (Mw 7.5), [Kakehi and Irikura, 1997] and the 1995 Kobe earthquake [Kakehi et al., 1996]. They found, with some exceptions, that the areas of high energy radiation intensity are not co-located with the high slip areas of low-frequency inversion. However, using a similar method, Hartzell et al. [1996] found the opposite result for the 1994 Northridge earthquake in which slip areas are common both for high and low frequency analysis.

Instead of using the eGf approach, Nakahara et al. [1998] calculated theoretical Green's functions based on the radiative transfer theory of Sato et al. [1997]. They inverted root mean square of velocity wave into the radiated energy on an assumed fault plane. To investigate the wide frequency range in velocity waveforms, they divided the 1-16 Hz frequency range into four bands and analyzed each band separately. Applying this method to the 1994 Off-Sanriku earthquake (Mw 7.7), they found an area of high energy radiation corresponding to the location where the largely unilateral rupture visible imaged from the low-frequency data [Nakayama and Takeo, 1997] terminated. The same method was also applied to the 1995 Kobe earthquake [Nakahara et al., 1999], in which high energy radiation was found to correspond to the initiation, surface breakage, and termination of rupture. In summary, most, though not all, studies of high frequency wave generation find that the distribution where high frequency waves are generated do not match the slip area revealed by low-frequency waves. However, this is an area that deserves more study in the future.

4.2 Dynamic Imaging

Fault slip inversion studies explained so far depend on the elastodynamic representation theorem [e.g., Aki and Richards, 1980], which allows calculation of waveforms from any functional form of spatio-temporal slip. However, a real earthquake will rupture in a manner that satisfies some constitutive relation for frictional slip or fracture. We can investigate what kind of constitutive relations are satisfied using the fault slip model by waveform inversion [Ide and Takeo, 1997; Bouchon, 1997], however, it is also possible to limit the source models to be consistent to known friction laws and fracture criteria when the source image is determined. This is a problem of dynamic source imaging.

Forward modeling of dynamic fault rupture has been possible since the 1970’s [Andrews, 1976; Das and Aki, 1977; Day, 1982]. The first attempt to connect dynamic modeling and source inversion was for the 1979 Imperial Valley earthquake [Quin, 1990]. He determined a distribution of stress drop and strength that explained the slip history in the kinematic model of Archuleta [1984]. The assumption of the friction law in this study was a simple static and dynamic friction. Similar analysis was applied to the inland earthquakes of Japan [Miyatake, 1992] to explain kinematic models of Takeo and Mikami [1990]. In these studies, waveforms were not compared. Fukuyama and Mikumo [1993] constructed a dynamic model to match synthetic waveforms and observed data for the 1990 Izu-Oshima earthquake (Mw 6.4) by iteratively calculating dynamic rupture process and solving waveform inversion. A similar attempt was made by Ide and Takeo [1996] for the 1993 Kushiro-Oki earthquake.
All the above-mentioned dynamic models assumed a simple friction law that has only static and dynamic friction coefficients. As a more physically reasonable law, Olsen et al. [1997] and Peyrat et al. [2001] used the slip-weakening friction law [e.g. Ida, 1972; Andrews, 1976] to simulate the dynamic rupture process in the 1992 Landers earthquake assuming a planar fault. They showed that a critical slip weakening distance, $D_c$, of 0.8 m can simulate the event and that a small change on initial stress state can result in quite different slip behavior. The geometry of fault system is another important factor. Aochi and Fukuyama [2002] considered a nonplanar fault system to simulate the Landers earthquake, successfully explaining the slip pattern, the unilateral rupture propagation, and seismic waveforms without assuming any laterally heterogeneous distribution of frictional parameters.

What can these dynamic models tell us about earthquake physics? A surprising aspect of kinematic rupture models has been that the rise time inferred from strong motion modeling is short compared with the time required for seismic waves to propagate healing information across the fault surface. [Heaton, 1990] proposed a self-healing slip pulse as a general feature of earthquake rupture. There are at least two possibilities for producing such behavior: the dynamics of fault friction law and inhomogeneity of stress or strength in the earth. The possibility of the latter was supported by a model for the rupture process of the 1984 Morgan Hill earthquake [Beroza and Mikumo, 1996]. The slip-weakening law used by Peyrat et al. [2001] also successfully produced pulse like rupture propagation for the 1992 Landers earthquake.

For the slip-weakening law, the slip-weakening distance $D_c$ is an important parameter. Seismological estimates of $D_c$ have been much larger than values found in the laboratory and previously thought to apply to rupture in the Earth. Ide and Takeo [1997] estimated $D_c$ as 0.5-1.0 m for the Kobe earthquake and a similar value of 0.8 m has been found for the Landers earthquake [Olsen et al., 1997]. As Guatteri and Spudich [2000] point out, however, the estimate of $D_c$ is difficult without good observations of high-frequency waves, and fracture surface energy $G_c$ rather than $D_c$ is more easily measured through source imaging. Guatteri et al. [2001] estimated $G_c$ for the Kobe earthquake to be $1.5 \times 10^6$ J/m$^2$.

### 4.3 Generalization and Scaling of Complexity

A remarkable property of the earthquake source is that some seismic source parameters, such as the stress drop and the apparent stress, do not vary systematically with magnitude over essentially the entire observable range of earthquake size [Abercrombie, 1995; Ide and Beroza, 2001]. The similarity between large and small earthquakes extends to the heterogeneity of the slip on the fault as well. Images of the slip distribution in earthquakes indicate that slip varies strongly over the fault from the smallest events ($M \sim 3$) for which such behavior is resolvable [Frankel et al., 1986] to the very largest earthquakes ($M \sim 9$) that have been recorded [Ruff and Kanamori, 1983]. Because slip in earthquakes is strong and variable, earthquake scaling behavior is usually cast in terms of average properties, such as the average slip [Wells and Coppersmith, 1994]. This can lead to difficulties for properties that are not so easily averaged, such as length or width; however recognition that slip is heterogeneous can lead to straightforward generalizations of effective width and length [Mai and Beroza, 2000].

Because slip in earthquakes is complex, however, it suggests the need for a general description of heterogeneity in terms of the statistics of spatial slip variation. Seismologists have long represented some aspects of the slip behavior in earthquakes in such a way. Haskell [1964] introduced a coherence length, shorter than the overall fault length, to explain the contribution of high frequency waves to the seismic spectrum. He subsequently formalized this [Haskell, 1966] using the autocorrelation of the slip acceleration on the fault. Aki [1967] used a similar approach, but used the autocorrelation of the slip velocity and related it to the observed scaling behavior of large earthquake spectra and magnitude.
Subsequent studies have concentrated more on spatial aspects of slip heterogeneity. Aki [1979] developed a barrier model to describe slip heterogeneity. Andrews [1980] subsequently developed a description of slip heterogeneity as a self-similar process and went on to include the temporal behavior of slip velocity as well [Andrews, 1981]. His model is notably different from earlier models in that it does not include a characteristic scale length. Frankel [1991] explained both slip heterogeneity and the frequency-magnitude statistics of earthquakes using a model of subfaults with scale invariant strength. Somerville et al. [1999] investigated the scaling properties of different parameters in finite-fault slip models. They developed relations between the seismic moment and: the radius of the largest asperity (high-slip region), the average number of asperities, and the slip duration. They used these observations to develop a self-similar slip model modified by low corner wavenumbers that scale with earthquake size. Mai and Beroza [2002] looked at a similar set of rupture models and characterized them as spatial random fields. They found that Gaussian models produced slip distributions that were inconsistent with slip distributions determined from strong motion modeling. They found that either a fractal model, or a von Karman distribution were most consistent with observations of past earthquakes. It is worth noting that the slip models on which these analyses are based may be biased to smoothness (more rapid decay) at high spatial wavenumbers since such constraints are typically used to regularize the inverse problem when finite-fault rupture models are estimated from strong motion and geodetic data. It is also possible that variations in rupture time are mapped into variations in the slip distribution, which might have the opposite effect.

One way to quantify the general properties of kinematic rupture models is by comparing their higher moments. The variances of the moment release distribution can be interpreted in terms of characteristic dimensions that are defined as twice the square-root of the variance tensor projected along a particular direction. For instance, the characteristic rupture length, $L_c$, is two times the square-root of the variance of the moment-release distribution along its primary spatial dimension. The characteristic duration, $\tau_c$, and characteristic width, $W_c$, have analogous definitions. For heterogeneous moment-release distributions, the characteristic dimensions are always smaller than the total dimensions, but provide an estimate of the region which contributed significantly to the moment-release [Backus 1977a,b]. The information about rupture propagation is derived from the mixed moment between space and time. This quantity specifies the average propagation velocity of the instantaneous centroid, $v_0$, which ranges from zero for a perfectly symmetric bilateral rupture to the full rupture velocity for a uniform slip unilateral rupture. The directivity of a rupture can be evaluated using the ratio of $v_0 \tau_c$ to $L_c$. This ratio ranges from 0 for a symmetric bilateral rupture to 1 for a uniform slip unilateral rupture. Figure 4 compares these two quantities for a population of 29 strong motion inversions performed on 17 earthquakes. Approximately 75% of these earthquakes have values of this ratio greater than 0.5 indicating a dominantly unilateral rupture. This observation has been confirmed for a catalog of large global events that were only recorded teleseismically [McGuire et al., 2002].

5 CONCLUSIONS

There has been considerable recent progress in seismic source imaging, much of it driven by new data sets from large earthquakes; nevertheless, there remain many research opportunities that present and future large-scale data gathering initiatives, such as Earthscope, will help to propel. The comparison between the Izmıt and the Chi-Chi earthquakes clearly demonstrates the value of improved data coverage. Dense and redundant data can help overcome the many shortcomings in different analyses of the earthquake source. The downhole strong-motion instrumentation to be provided by PBO will be the type of dataset that will drive new progress in source imaging. Even with ample data, however, it will be important for seismologists to use the data more carefully. This includes the use of empirical Green's functions and Green's functions calculated for three-dimensionally varying Earth structure to help to improve source images. Currently leading studies of the rupture dynamics of individual earthquakes, such as Landers [e.g., Peyrat et al., 2001], utilize only one-dimensional earth models. In the future, dynamic rupture modeling will be performed in more realistic three-dimensional structures. Accurately accounting for three dimensional propagation effects when interpreting strong-motion recordings and geodetic
data can greatly improve the resolution of seismic source imaging [Graves and Wald, 2001; Wald and Graves, 2001]. The need for this level of accuracy is even more critical when developing dynamical descriptions of an individual event’s rupture both because of the influence of the initial portion of the rupture on all later parts and because of the need for using data in as high a frequency band as possible [Guatteri and Spudich, 1999] to resolve short scale-length aspects of the earthquake source. The flexible array component of USArray, and associated passive and active seismic experiments, provide an opportunity to greatly increase our knowledge of the 3-D structure in regions where important strong-motion datasets already exist. The advanced active source techniques for imaging earth structure described in earlier chapters of this volume could be utilized to increase the power of existing strong-motion datasets to constrain rupture dynamics.

Improved utilization of the available datasets also means that uncertainties in the prediction of the data have to be taken into account quantitatively, that accompanying error analysis of the resulting model should be undertaken, and that better methods for regularizing the inverse problem must be used and/or developed. Doing so will allow diverse data sets, such as GPS and InSAR observations, to be combined with seismic observations in an optimal way. The types of imaging undertaken will continue to change with time as research goals evolve. Frequency-dependent source imaging, direct imaging of dynamic source parameters, and direct estimation of stochastic source parameters are all current research areas that have the potential to result in rapidly improving our understanding of the earthquake source.

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Figure Captions

Figure 1. Schematic illustrations of fault model parameterization. Subfault system (bottom) and source time function representation (top).

Figure 2. Comparison of source models for 1999 İzmit, Turkey, earthquake. Map (top) and slip distribution of the four source models.

Figure 3. Comparison of source models for the 1999 Chi-Chi, Taiwan, earthquake. Final slip distribution and slip during 5 s periods up to 35 s are shown for three source models.

Figure 4. Comparison of the characteristic rupture length, $L_c$, vs. the characteristic propagation distance of the centroid, $v_0^*\tau_c$, for a suite of strong-motion inversions [see McGuire et al., 2002; for references, definitions, and values]. Symmetric bilateral ruptures plot on a line with zero slope while uniform slip unilateral ruptures plot on a line with slope equal to 1.0. Stars and circles represent multiple strong motion models of İzmit and Chi-Chi earthquakes, respectively.
Tables

Table 1. Models of the 1999 İzmit, Turkey, earthquake.

<table>
<thead>
<tr>
<th>Model</th>
<th># of faults</th>
<th>total size [km²]</th>
<th># of subfaults [°]</th>
<th>STF parameters</th>
<th>method</th>
<th>data [10¹⁰ Nm] [s]</th>
<th>Mw</th>
<th>Duration</th>
<th>Super S?</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td>N/A</td>
<td>155 x 18</td>
<td>31 x 4</td>
<td>N/A</td>
<td>timing, rise time</td>
<td>SA</td>
<td>Nd</td>
<td>2.5</td>
<td>24</td>
</tr>
<tr>
<td>D</td>
<td>4</td>
<td>165 x 23</td>
<td>23 x 5</td>
<td>180 [10] 20</td>
<td>timing, time windows</td>
<td>SA, F, G, S</td>
<td>2.4</td>
<td>50 (20)³</td>
<td>No</td>
</tr>
<tr>
<td>S</td>
<td>4</td>
<td>141 x 23</td>
<td>47 x 8</td>
<td>180 [45]</td>
<td>time windows</td>
<td>NNLS</td>
<td>Nv</td>
<td>1.5</td>
<td>20</td>
</tr>
<tr>
<td>Y</td>
<td>1</td>
<td>93.6 x 21.6</td>
<td>26 x 6</td>
<td>180</td>
<td>time windows</td>
<td>NNLS</td>
<td>Nd, F</td>
<td>1.7</td>
<td>18</td>
</tr>
</tbody>
</table>

⁴Nd: near-field displacement, Nv: near-field velocity, F: far-field displacement, G: GPS, S: InSAR.

⁵Within 50 km from the hypocenter.

Table 2. Second moment parameters of the İzmit earthquake

<table>
<thead>
<tr>
<th>Model</th>
<th>Lc (km)</th>
<th>Wc (km)</th>
<th>τc (s)</th>
<th>vc (km/s)</th>
<th>v₀ (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td>73</td>
<td>11</td>
<td>11</td>
<td>6.6</td>
<td>2.8</td>
</tr>
<tr>
<td>S</td>
<td>68</td>
<td>12</td>
<td>9</td>
<td>7.4</td>
<td>2.0</td>
</tr>
<tr>
<td>D</td>
<td>86</td>
<td>10</td>
<td>23</td>
<td>3.8</td>
<td>3.2</td>
</tr>
<tr>
<td>Y</td>
<td>44</td>
<td>10</td>
<td>8</td>
<td>5.5</td>
<td>4.0</td>
</tr>
</tbody>
</table>

Lc is the characteristic rupture length, Wc, the characteristic rupture width, and τc is the characteristic rupture duration. vc = Lc/τc and v₀ is the magnitude of the average propagation velocity of the instantaneous centroid (see McGuire et al., (2001) for definitions).

Table 3. Models of the 1999 Chi-Chi, Taiwan, earthquake.

<table>
<thead>
<tr>
<th>Model</th>
<th>total size [km²]</th>
<th># of subfaults [°]</th>
<th>rake</th>
<th>STF parameters</th>
<th>method</th>
<th>data [10¹⁰ Nm] [s]</th>
<th>Mw</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>105 x 45.5</td>
<td>32 x 13</td>
<td>45</td>
<td>time windows (16.5 s)</td>
<td>N/A</td>
<td>Nv</td>
<td>4.1</td>
</tr>
<tr>
<td>M</td>
<td>105 x 40</td>
<td>21 x 8</td>
<td>45</td>
<td>time windows (12 s)</td>
<td>N/A</td>
<td>Nd, F, G</td>
<td>4.7</td>
</tr>
<tr>
<td>W</td>
<td>84 x 44</td>
<td>20 x 11</td>
<td>80</td>
<td>time windows (24 s)</td>
<td>NNLS</td>
<td>Nv, G</td>
<td>2.7</td>
</tr>
<tr>
<td>Z</td>
<td>96 x 40</td>
<td>24 x 10</td>
<td>70</td>
<td>timing, rise time, shape</td>
<td>2.5⁴</td>
<td>GA</td>
<td>2.9</td>
</tr>
</tbody>
</table>

⁴Nd: near-field displacement, Nv: near-field velocity, F: far-field displacement, G: GPS.

Number of node point. *Vr, not maximum, changeable at each subfault.

Table 4. Second moment parameters of the Chi-Chi earthquake

<table>
<thead>
<tr>
<th>Model</th>
<th>Lc (km)</th>
<th>Wc (km)</th>
<th>τc (s)</th>
<th>vc (km/s)</th>
<th>v₀ (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>56</td>
<td>18</td>
<td>18</td>
<td>3.1</td>
<td>2.3</td>
</tr>
<tr>
<td>M</td>
<td>45</td>
<td>23</td>
<td>15</td>
<td>3.1</td>
<td>2.1</td>
</tr>
<tr>
<td>W</td>
<td>38</td>
<td>21</td>
<td>13</td>
<td>3.0</td>
<td>1.2</td>
</tr>
<tr>
<td>Z</td>
<td>40</td>
<td>20</td>
<td>11</td>
<td>3.7</td>
<td>2.7</td>
</tr>
<tr>
<td>I</td>
<td>46</td>
<td>22</td>
<td>12</td>
<td>3.9</td>
<td>2.6</td>
</tr>
</tbody>
</table>

Lc is the characteristic rupture length, Wc, the characteristic rupture width, and τc is the characteristic rupture duration. vc = Lc/τc and v₀ is the magnitude of the average propagation velocity of the instantaneous centroid (see McGuire et al., (2001) for definitions). The Z model was calculated using single fault-plane approximation (courtesy of M. Mai) to the reported model. ⁴ model of Iwata et al. [2000].
Fig. 1
Fig. 2

- **Bouchon et al. (2002)**: $M_s = 2.5 \times 10^{26} \text{Nm}$
- **Delouis et al. (2002)**: $M_s = 2.4 \times 10^{26} \text{Nm}$
- **Sekiguchi et al. (2002)**: $M_s = 1.5 \times 10^{26} \text{Nm}$
- **Yagi and Kikuchi (2000)**: $M_s = 1.7 \times 10^{27} \text{Nm}$

Slip (m) scale: 0, 2, 4, 6, 8

Scale: 0, 20, 40, 60, 80 km
Final slip

$Chi \ et \ al. \ \ 4.1 \times 10^{20} \ Nm$

$Ma \ et \ al. \ \ 4.7 \times 10^{20} \ Nm$

$Wu \ et \ al. \ \ 2.7 \times 10^{20} \ Nm$

Temporal change

Fig. 3
Fig. 4