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DETERMINISTIC INVERSE APPROACHES FOR NEAR-SOURCE HIGH-FRECUENCY STRONG MOTION

Masahiro Iida

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DETERMINISTIC INVERSE APPROACHES
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CONTENTS

Preface			9
Lecture 1	Source Inversion		11
	Subject 1	Source complexity of the 1987 Whittier Narrows, California, earthquake from the inversion of strong motion records	11
	Subject 2	Source inversion of high-frequency strong motion records in the near-source region	27
Lecture 2	Array Layout for Source Inversion		35
	Subject 1	Relationship between strong-motion array parameters and the accuracy of source inversion and physical waves	35
	Subject 2	Optimum strong-motion array geometry for source inversion	54
	Subject 3	Preliminary analysis of resolving power of existing strong-motion arrays for source inversion	65
	Subject 4	Strong seismic ground motion array layout for source studies: three attempts	85
	Subject 5	Source effects on strong-motion records and resolving power of strong-motion arrays for source inversion	91
Lecture 3	Scatterer Inversion		109
	Subject 1	Seismic path and site effect estimation based upon scatterer inversion	109
	Subject 2	The seismic coda, site effects, and scattering in alluvial basins studied using aftershocks of the 1986 North Palm Springs, California, Earthquake as source arrays	126

DETERMINISTIC INVERSE APPROACHES FOR NEAR-SOURCE HIGH-FREQUENCY STRONG MOTION

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PREFACE

This is a collection of the materials which were used as texts in Dr. Masahiro Iida's three lectures at CENAPRED and the Engineering Institute of UNAM on October to November, 1993, on the topic, "Deterministic Inverse Approaches for Near-Source High-Frequency Strong Motion".

The topics of the three lectures are: (1) Source inversion, (2) Array layout for source inversion and (3) Scatterer inversion. The common purpose of the three lectures is to demonstrate the effectiveness of a deterministic inverse approach for near-source high-frequency strong motions. In the first lecture, he indicates that source effects are dominant in the near-source regions, and that the detailed source modeling is required, while, in the second lecture, effects of array geometries for source studies are estimated. In the third lecture, lateral heterogeneity in shallow underground structure is found to be considerable.

The author expresses his appreciation to Japan International Cooperation Agency (JICA) and National Center for Disaster Prevention (CENAPRED) for giving him chances to make those lectures and for recommending him to publish the lecture notes here. The author also thanks Mrs. Mie Tanaka de Aréchiga for typing texts.

Lecture 1 SOURCE INVERSION

Subject 1 Source Complexity of the 1987 Whittier Narrows, California, Earthquake from the Inversion of Strong Motion Records

1. INTRODUCTION

The October 1, 1987, Whittier Narrows earthquake (origin time 1442:19.48 UT) was a moderate sized event with a local magnitude of 5.9. It generated a great deal of interest because of its location within the major metropolitan area of Los Angeles, and because of its apparent association with a previously unidentified, north-dipping, blind thrust [Davis *et al.*, 1989; Hauksson and Jones, 1989]. Blind thrusts are so named because they do not reach to the Earth's surface and are typically overlain by a deformed section of anticlinal folds. The earthquake hazard in California and the rest of the world due to blind thrusts is just beginning to be appreciated. Other recent earthquakes which have been associated with blind thrusts include the 1983 (M_L = 6.5) Coalinga earthquake [Eaton *et al.*, 1983] and the 1985 (M_S = 6.6, 6.9) Nahanni earthquakes [Wetmiller *et al.*, 1988]. Stein and Yeats [1989] give an informative worldwide survey of blind thrust regions. To assess more accurately the seismic hazard that future blind thrust earthquakes pose for the Los Angeles basin, we attempt in this study to recover as much detail as possible about the source for the Whittier Narrows earthquake. We are fortunate in that a large number of strong motion records were recorded close to the epicenter of this earthquake. In fact, it is one of the best instrumented earthquakes to date. These records from the basis of our study, in which the strong motion waveforms are inverted to obtain the history of slip on a finite fault plane.

2. DATA

The strong motion data used in this study come from three main sources: the California Division of Mines and Geology network (California Strong Motion Instrumentation Program) [Shakal *et al.*, 1987], the U.S. Geological Survey [Etheredge and Porcella, 1987; Brady *et al.*, 1988], and the University of Southern California [Trifunac, 1988]. To minimize propagation path effects, which are often difficult to distinguish from source effects, only stations within 15 km of the epicenter are used. With this cutoff, the station ranges are comparable to or less than the source depths, which emphasize the direct body waves. The preponderance of direct body waves in the data is substantiated for the stations to the northwest of the epicenter by the simple waveforms recorded for the magnitude 5.3 aftershock of October 4, 1987 [Levine *et al.*, 1988]. These records do not exhibit any obvious complex propagation path effects, which would make unraveling the source history difficult. Comparing the aftershock records with the mainshock records from the same stations, shows the mainshock records to be much more complicated, reflecting a considerably more complex source. Table 1 lists the 17 stations used and their locations. The station distribution is shown in Figure 1, which also shows the map view or surface projection of the model fault plane. There is good 360° azimuthal coverage of the source. The station pattern is similar to an array configuration tested by Iida *et al.* [1988] and found to give good resolution of the source.

Each of the 17 stations recorded three components of ground acceleration. However, only the horizontal components are used in the source inversions for the following reason. We have some information on the general shape of the seismic velocity versus depth function in the Los Angeles basin for both P and S waves. However, considerably less is known about the ratio of the P to S wave velocities in this area. Because of strict timing requirements in the inversion, accurate knowledge of this ratio as a function of depth is needed to simultaneously model P and S waves. For this reason we have chosen to invert only the horizontal components of motion, which are dominated by S wave energy Because of the difficulty in modeling high frequencies, velocity records (rather

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* This lecture note is based on the paper of the same title written by Stephen Hartzell and Masahiro Iida, published on "Journal of Geophysical Research, Vol. 95, No. B8, pp. 12,475-12,485, August 10, 1990"

than accelerations) are used in the inversion. The instrument-corrected ground acceleration is first integrated to velocity. The velocity records are then band-pass filtered from 0 2 to 3.0 Hz using a zero-phase-shift Butterworth filter [Oppenheim and Schafer, 1975]. The filtering is done to remove frequencies outside the bandwidth for which Green's functions could be conveniently calculated. The records are then interpolated to a uniform time step of 0.05 sec (Nyquist frequency of 10 Hz)

TABLE 1 Strong Motion Station Location and Orientation

Station	Latitude, °N	Longitude, °W	Owner	Components
FRS Alhambra–Fremont School	34.070	118.150	CDMG	180,270
SWA San Marino–SWestern Acad.	34.115	118.130	CDMG	270,360
OBP Los Angeles–Obregon Park	34.037	118.178	CDMG	270,360
ECP Altadena–Eaton Canyon Park	34.177	118.096	CDMG	360,90
DOW Downey–County Maint. Bldg.	33.924	118.167	CDMG	180,270
GVR Garvey Reservoir	34.05	118.11	LAMWD	330,60
WND Whittier Narrows Dam–Upstream	34.03	118.05	ACOE	62,152
ALH Alhambra 900 S. Fremont	34.09	118.15	USGS	360,90
WTR Whittier 7215 Bright Ave.	33.977	118.036	USGS	90,180
LBM Los Angeles–Bulk Mail Center	33.99	118.16	USGS	280,10
VRN Vernon 4814 Loma Vista Ave.	34.00	118.20	USGS	277,7
NWK Norwalk 12400 Imperial Hiway	33.92	118.07	USGS/BECH	360,90
U19 San Gabriel 600 E. Grand Ave	34.091	118.093	USC	270,180
U66 El Monte 11338 Fairview Ave	34.093	118.018	USC	185,95
U71 West Covina 1307 S. Orange	34.064	117.952	USC	320,230
U73 Hacienda Heights 16750 Colma	33 990	117 942	USC	230.140
U93 Arcadia 180 Campus Dr.	34 130	118.036	USC	9,279

CDMG, California Division of Mines and Geology; LAMWD, Los Angeles Metropolitan Water District; ACOE, Army Corps of Engineers, BECH, Bechtel Power Corporation, USGS, U.S. Geological Survey; USC, University of Southern California.

3. FAULT MODEL AND INVERSION METHOD

The model fault for the Whittier Narrows earthquake is taken to be a square planar region 10 km on an edge. The hypocenter is located at the center of the fault plane at a depth of 14.6 km [Hauksson and Jones, 1989]. The epicenter is located at 34°2.96'N, 118°4.86'W [Hauksson and Jones, 1989]. Bent and Helmberger [1989] modeled teleseismic body waves and obtained a strike of 280° and a dip of 40°. Hauksson and Jones [1989] obtained a similar strike of 270° from local short-period first motion data. However, their preferred value for the dip is 25°. Lin and Stein [1989] modeled geodetic data and obtained a preferred dip of 30°. In general, we favor the use of fault plane parameters obtained from longer-period data, since these values should be more representative of the majority of the moment release. We fix the strike of our model fault plane at 280°. Two different values of dip were tried in the inversion of the strong motion data, 30° and 40°. A dip of 30° gave a marginally better fit to the data, regardless of how the fault was parametrized. We show only the results for a 30° dip. The surface projection of the model fault can be seen in Figure 1

Following Hartzell and Heaton [1983] the fault plane is divided up into small rectangular regions of equal area which we will call subfaults (Figure 2). Each subfault is 1 km^2. The ground motions at the strong motion stations are calculated for both a dip-slip mechanism (thrust) and a strike-slip mechanism (right-lateral) on each of the individual subfaults using the discrete wave number/finite element method of Olson *et al.* [1984]. The Green's functions include all theoretical arrivals for the specified structure and time interval and are valid in the frequency band from 0 to 3.0 Hz. The synthetic ground motions for each subfault are band-pass filtered in the same manner as the observed data with a 0.2- to 3.0-Hz Butterworth filter. The velocity structure is shown in Figure 3 and is based on the work of Wald *et al.* [1988], which in turn is based on the results of Hauksson [1987] and Apsel *et al.* [1981]. The most important feature of the model is its high velocity gradient in the top 6 km, which simulates low-velocity sediments Most of the strong motion stations in this study are located on thick sediments, and the velocity structure is constructed in accordance with this fact.

The inversion method is discussed in detail by Hartzell [1989] and is reviewed here. If we wish to solve for the slip amplitudes for a prescribed rupture velocity, the problem is linear. The observed records and the subfault synthetic records then form an overdetermined system of linear equations,

$$\mathbf{A}x \cong b \qquad (1)$$

where **A** is the matrix of synthetics, b is the data vector, and x is the solution vector of the subfault dislocation weights. Each column of **A** is composed of the synthetics, strung end to end, for a particular subfault and a particular mechanism (either strike-slip or dip-slip) for all the stations in the inversion. Similarly, b is formed by stringing all the observation records end to end. Thus each time point on each record is explicitly included in the inversion. The number of columns of **A** depends on the number of elements in x. The elements of x are the amounts of strike-slip and dip-slip dislocations to be applied to each subfault to fit the observations. Equation (1) can be solved by linear least squares, but the solution is unstable. The instability arises because **A** is an ill-conditioned matrix, meaning that a small change in the data results in a large change in the solution. The problem is stabilized by appending linear constraints giving

$$\begin{pmatrix} \mathbf{C}_d^{-1}\mathbf{A} \\ \lambda_1\mathbf{S} \\ \lambda_2\mathbf{M} \end{pmatrix} x \cong \begin{pmatrix} \mathbf{C}_d^{-1}b \\ 0 \\ 0 \end{pmatrix} \qquad (2)$$

S is a matrix of smoothing constraints where the difference between the slip on adjacent subfaults is set equal to zero. **M** is a matrix of minimization constraints where the slip on each subfault is set equal to zero. λ_1, and λ_2 are linear weights, whose magnitudes control the trade-off between satisfying the constraints and fitting the data. $\mathbf{C}_d$ is an a priori data covariance matrix, which is used as a data scaling matrix. The data covariance matrix is diagonal and normalizes each data record to have a peak amplitude of 1.0. Thus each record has nearly an equal weight in the inversion. The solution vector **x** is solved for using a Householder reduction method that invokes a positivity constraint on the solution [Lawson and Hanson, 1974], that is, each of the values in the vector **x** are greater than or equal to zero.

If we wish to solve simultaneously for the magnitude of the slip and the rupture initiation time for each subfault, the problem is nonlinear and is solved in an iterative manner. Let x_k be an initial guess at the solution vector and $g(x_k)$ be the calculated ground motions for the initial guess based on our fault model. Define the residual vector

$$R = b - g(x_k)$$

to be the difference between the data and the prediction of the model. Next, define the Jacobian matrix **G** of partial derivatives of the model predictions with respect to the model parameters, where

$$G_{ij} = \partial g_i / \partial x_j$$

If a perturbation to the model parameters, Δx_k, satisfies the data,

$$g(x_k + \Delta x_k) = g(x_k) + G\Delta x_k = b$$

or

$$\mathbf{G}\Delta x_k \cong R \tag{3}$$

We now have an overdetermined problem similar to (1) which is solved using a least squares criterion for the model parameter perturbation vector Δx_k. The $k + 1$ solution is obtained from $x_{k+1} = x_k + \Delta x_k$, a new residual vector is calculated, and (3) is solved again. This process is continued until the solution converges.

As with the linear problem, the matrix is generally ill-conditioned, and the solution of equation (3) must be stabilized by appending stabilization criterion:

$$\begin{pmatrix} \mathbf{C}_d^{-1}(\mathbf{G}_s|\lambda_1\mathbf{G}_t \\ \lambda_2\mathbf{S}_s \\ \lambda_3\mathbf{M}_s \\ \lambda_4\mathbf{S}_t \\ \lambda_5\mathbf{M}_t \end{pmatrix} \Delta x_k \cong \begin{pmatrix} \mathbf{C}_d^{-1}R \\ 0 \\ 0 \\ 0 \\ 0 \end{pmatrix} \tag{4}$$

$\mathbf{C}_d$ has the same meaning as in equation (2). The Jacobian matrix written here is partitioned into the partials with respect to slip amplitude, $\mathbf{G}_s$, and the partials with respect to rupture time, $\mathbf{G}_t \cdot \lambda_1$ is a scalar weight to account for the dimensionality difference between the partials with respect to slip amplitude and rupture time. After each iteration, the rupture time perturbations are obtained by $\Delta x_k = \lambda_1 \Delta x_k$. The matrices $\mathbf{S}_s$ and $\mathbf{S}_t$ contain spatial smoothing constraint equations for the perturbations to slip amplitude and rupture initiation time, respectively. The matrices $\mathbf{M}_s$ and $\mathbf{M}_t$ contain minimizing constraints for the perturbations to slip amplitude and rupture time. λ_2, λ_3, λ_4 and λ_5 are the corresponding scalar weights on these constraints. All the weights $\lambda_1, \ldots, \lambda_5$ are obtained by a trial and error procedure. If the weights are set improperly, the solution will diverge or converge at a slower rate. The elements of the solution vector x_k at each iteration are required to be nonnegative by setting any model parameter equal to zero that becomes negative. This constraint prohibits backward slip on the fault and rupture initiation times earlier than the origin time. The model parameter which is set to zero is free to take on a positive value at any subsequent iteration. Because the problem is nonlinear and the model space can be a complexly corrugated surface, it would be ill-advised to constrain a model parameter, that initially goes negative or approaches zero, to be zero up through the final iteration.

The covariance of the model parameters due to a variance in the data of σ_d^2 is

$$[\text{cov } \mathbf{x}] = \sigma_d^2 \mathbf{V}\Lambda^{-2}\mathbf{V}^T \tag{5}$$

where $\mathbf{U}\Lambda\mathbf{V}^T$ is the singular value decomposition of the matrix on the left-hand side of (4). Following Olson and Apsel [1982] and Mendez and Luco [1990], the variation in the model parameters for a given data tolerance σ_d can be written as

$$\delta\mathbf{x} = \pm\sigma_d[\text{diag } (\mathbf{V}\Lambda^{-2}\mathbf{V}^T)]^{1/2} \tag{6}$$

Since the solution to the linear least squares problem is unique, equation (6) is an expression for the variation in model parameters for the global minimum. However, the calculation of slip amplitude and rupture initiation time is a nonlinear problem, and (6) is then a local estimate of the variation in model parameters.

A few words about the application of equation (6) are called for. We need to do a singular value decomposition with a positivity constraint on the model parameters. This calculation is done by performing SVD on the modified matrix constructed from the left-hand side of equation (4) by removing the columns associated with the predetermined zero model parameters. This procedure is justified because an identical solution is obtained using SVD on this diminished matrix as using Householder transformations on the full matrix of synthetics and constraint equations with a positivity constraint in place. Then, given a measure of the misfit to the data that one is willing to accept, the data tolerance σ_d, equation (6) gives the corresponding perturbation to the model parameters. A convenient measure for σ_d is a percentage of the Euclidean norm of the misfit between the data and synthetics. With a positivity constraint, certain of the model parameters are identically zero, and this analysis returns no model perturbations for these parameters. The SVD analysis is performed on the synthetics and constraints matrix including the premultiplication by the diagonal matrix of data normalization weights, $\mathbf{C}_d$. Applying equation (6) without $\mathbf{C}_d$ would result in model perturbations for a different formulation of the problem than the one used in this paper. However, a linear relationship does exist between the size of the weights in $\mathbf{C}_d$, the size of the singular values of the synthetics matrix, and the Euclidean norm, which is offset in equation (6) by the inverse relationship between the size of the singular values and the model perturbations.

4. RESULTS

Three different types of inversions of the strong motion waveforms were performed. The first and simplest approach assumes a constant rupture velocity with each subfault rupturing once. The second formulation also uses a fixed rupture velocity, but each subfault is allowed to rupture twice, to allow for a more complex source-time function. The third type of inversion allows each subfault to rupture once, but the rupture velocity and the rupture time of each subfault may vary. In every case the source-time function associated with the rupturing of a single subfault is a triangle with a 0.2-s duration. A duration of 0.3 s was also tried, but this value gave an inferior fit to the data.

Figures 4, 5 and 6 show the contours of slip in centimeters of the solutions for the three inversion problems. Model L15 (Figure 4) assumes a fixed rupture velocity of 2.5 km/s with each subfault rupturing once. Model L18 (Figure 5) has a fixed rupture velocity of 2.5 km/s with each subfault allowed to rupture twice with a time separation of 0.2 s. Model NL22 (Figure 6) is the variable rupture velocity model with each subfault rupturing once. Figure 7 shows the contours of the initial and final rupture front positions for model NL22 at intervals of 0.5 s. The initial rupture front timing is for a constant rupture velocity of 2.5 km/s. In each of these figures the hypocenter is indicated by a solid dot. The slip is resolved into the strike-slip component (right-lateral), the dip-slip component (thrust), and the vector sum of the two. In the case of model L18, in which each subfault is allowed to rupture twice, Figure 5 displays the total slip that has occurred during both rupture intervals.

A rupture velocity of 2.5 km/s corresponds to about 0.8 V_s in the source region in our model. Other fixed rupture velocities were tried from 2 3 to 3.0 km/s. Rupture velocities of 2.7 and 3.0 km/s did not fit quite as well as 2.5 km/s. A rupture velocity of 2.3 km/s gave a smaller misfit to the data than 2.5 km/s, measured in terms of the Euclidean norm $\| \mathbf{A}x - b \|$. However, this improvement was accomplished largely by the addition of slip near the edges of the fault plane in an attempt to fit smaller amplitude, later arriving energy in the waveforms, and with the accompanying degrading of the fit to the beginning of the waveforms. Our confidence in properly identifying an arrival in the records as being due to a source effect and not a propagation effect, decreases the greater the arrival time after the origin time. For this reason fitting the later, lower amplitude arrivals in the records is not given a high priority. Therefore, the best fitting constant rupture velocity is estimated to be 2.5 km/s.

Table 2 lists the moment, Euclidean norm, and variance for models L15, L18 and NL22. The variance has the usual definition of the Euclidean norm squared divided by the number of degrees of freedom. The number of degrees of freedom is $N - 1$, where N is the number of data values in the inversion minus the number of nonzero model parameters in the solution. From Table 2 we see that model L18 fits the data the best. However, the difference in the fits is not dramatic. The closeness of the models is also seen in the similarity of the slip distributions in Figures 4, 5 and 6. A greater difference is seen for model NL22, which is understandable given that its rupture fronts are considerably different from models L15 and L18. In general, model NL22 has lower peak values of displacement. By varying the rupture velocity, NL22 is able to obtain as good a fit to the data with lower values of slip.

TABLE 2. Inversion Model Results

Model	Moment, dyn cm	Euclidean Norm	Variance
L15	7.9×10^{24}	19 05	0.054
L18	1.0×10^{25}	17.92	0.048
NL22	7.4×10^{24}	18.31	0.051

Perturbations in the model parameters for a given data tolerance σ_d are calculated by using the diagonal elements of the unit model covariance matrix and equation (6). The choice of the data tolerance is subjective, and we have arbitrarily selected a value of 1.0. For comparison, 10% of the Euclidean norm gives a value of 1.8. The contours of δx for model NL22 are shown in Figure 8. The perturbation to slip amplitude is about ± 6 cm. The perturbation to rupture time is about ± 0.1 s. For 10% of the Euclidean norm, these values would be approximately ± 12 cm and ± 0.2 s. However, these estimates do not address the errors in the model due to uncertainties in the velocity structure and the focal mechanism. There are also differences between the solutions L15, L18 and NL22 which are larger than these estimates, that reflect the constraints of the different model parametrizations. For these reasons the slip distributions in Figures 4, 5 and 6 have been contoured with large steps of 15 cm to emphasis the most important features. Similarly, a contour interval of 0.5 s is used in Figure 7 for the rupture front.

The data records are compared with the synthetics for models L18 and NL22 in Figure 9. In general, the waveforms are fit well in both shape and amplitude. The 280° component at station LBM shows the greatest disparity in amplitude, suggesting a localized propagation phenomenon.

The inferred distribution of slip can be used in a forward calculation to predict the ground motion in the epicentral region. This calculation is done in Figure 10 for model L18, where peak velocities are contoured. The values are for band-pass filtered synthetics from 0.2 to 3 0 Hz and thus are smaller than the actual peak velocities. The north-south component is contoured, which is the largest. Peak whole record amplitudes are used, so that different phases may be responsible for the peak motions at different distances and azimuths. However, the largest amplitudes are from direct S phases. The strong motion stations used in the inversion are indicated by triangles. The area of highest expected velocities is near the town of Whittier. The second largest amplitudes are to the west and northwest of the epicenter. Whittier experienced the greatest damage during the earthquake [Hauksson *et al.*, 1988; Leyendecker *et al.*, 1988], with some houses coming off of their foundations. Alhambra and Monterey Park had less damage, but more than the surrounding areas, with numerous broken chimneys. These damage records are well predicted by model L18. Kawase and Aki [1990] have explained the heavy damage in Whittier by critically incident SV waves in conjunction with a topographic effect. The results of this study indicate that although these processes may be partly responsible for the damage in the Whittier area, the ground motion can also be explained by source effects.

5. DISCUSSION AND CONCLUSIONS

The result that model L18 gives a lower variance than model L15 suggests that the true source-time function for each subfault of our model for the Whittier earthquake is more complicated than a simple triangle. The result that model NL22 gives a lower variance than model L15 suggests that the true rupture velocity is not constant. Both of these conclusions should not be surprising. The fact that models L18 and NL22 vary the source-time function and the rupture time, respectively, but not both, is a computational expedient. Both of these functions most likely varied during the Whittier earthquake. In particular, Figure 7 indicates some important features of the advancement of the rupture front for model NL22. Although the rupture has an average velocity of 2.5 km/s, there is an obvious asymmetry in the advancement of the rupture front, with some directions being faster than others. However, there appears to be no simple correlation of these fast and slow areas with features in the slip distribution.

The complexity of the slip for the Whittier earthquake is the most obvious result of this study. There are at least four distinguishable sources. The most apparent sources are two associated with the hypocenter, and three or four other source areas surrounding these two. An approximately circular region of low slip, with a radius of 2-3 km, lies around the hypocenter. As shown in Figure 11, the aftershocks located by Hauksson and Jones [1989] fall within this low slip zone. In this figure, the aftershocks occurring within the first five days and located within ± 2 km of the closest point on the model fault plane, are perpendicularly projected onto the plane. The comparison in Figure 11 is with model L18, but any one of the three solutions in this paper would yield a similar result. Wald *et al.* [1988], using a forward simulation technique, concluded that the acceleration records could be adequately modeled by putting most of the slip where the aftershocks did not occur. The Whittier earthquake appears to be another example of the occurrence of aftershocks where major slip did not occur during the mainshock [Mendoza and Hartzell, 1988].

The question may be raised why additional aftershocks did not occur outside of the sources which surround the aftershock pattern. Let us conjecture that these sources should lie within the ring of aftershocks in Figure 11. To test this hypothesis, an inversion was performed in which the fault plane was restricted to the smaller area within the aftershock pattern, with a fault length and width of 6 km. Each subfault was allowed to rupture twice with a separation in time of 1.0 s. This model forces the mapping of the outlying sources into the area of the fault plane encompassed by the aftershocks. This inversion gave the worst fit to the data in terms of Eucidean norm of any model tried and also required slips in excess of 2 m, leading us to conclude that these sources occurred outside of the aftershock pattern. We did not try a larger fault plane than 10 km^2. So the possibility exists that the source region may be even larger. However, given the good fits to the strong motion records by models L18 and NL22. we conclude that most of the source is resolved by these models. Furthermore, if a larger fault were used, sources would be added to the model to explain later arriving energy, which we have low confidence in being due to source effects.

Another interpretation of the slip distributions in Figures 4, 5 and 6 would put the outlying sources on nearby fault planes suggested by the complex pattern of aftershock focal mechanisms. Studies by Hauksson and Jones [1989] and Magistrale and Kanamori [1989] suggest additional lineations of both thrust and strike-slip mechanisms besides the mainshock fault plane. In particular, the largest aftershock of October 4 ($M_L = 5.3$) has a right-lateral strike-slip mechanism, and its hypocenter is plotted in Figure 11. The teleseismic body wave modeling results of Bent and Helmberger [1989] resolved two sources in the mainshock, with the second approximately 5 times larger than the first and delayed by 1.0 s. Both of these sources were found to have thrust mechanisms. The small teleseismic amplitude of the body wave phase SH compared to the amplitude of SV rules out any significant amount of strike-slip motion during the mainshock [Bent and Helmberger, 1989]. Our slip models are in good agreement with these teleseismic results. If the outlying sources in Figures 4, 5 and 6 did occur on subsidiary fault planes, they would have to be predominantly thrust faults closely associated with the mainshock fault.

Lin and Stein [1989] modeled geodetic data for the Whittier Narrows earthquake and obtained estimates of the dimensions of the fault plane. The minimum error solution they found by restricting the fault plane to the aftershock region has a fault length of 4.5 km, a down dip width of 6 km, a dip of 30°, a slip of 110 cm, and a moment of 9.6×10^{24} dyn · cm. However, if they do not limit the fault plane to the area encompassed by the aftershocks, they find another minimum solution with a smaller error. This solution has a fault length of 12 km, a down dip width of 4 km, a dip angle of 34°, a slip of 71 cm, and a moment of 1.09×10^{25}. This model is in good agreement with the strong motion inversion results presented in this paper.

The results of this study and Mendoza and Hartzell [1988] indicate that aftershock patterns do not define where slip occurred during the mainshock, but rather define where slip did not occur, or where it stopped. This conclusion is now supported by observation but was earlier postulated by Rybicki [1973] and Aki [1979], based on the theory that aftershock activity occurs in regions of concentrated stress following primary faulting during the mainshock. Stein and Lisowski [1983] calculated the postearthquake stress field caused by the coseismic slip for the 1979 Homestead Valley, California, earthquake. They were able to explain the occurrence of aftershocks by the regions which experienced an enhancement in stress. Schwartz *et al.* [1989] studied large subduction zone events in the Kurile Islands Arc. They found that nearly all of the smaller magnitude earthquakes located outside of the regions associated with major moment release of the great earthquakes. Similar results have been found for the 1986 Andreanof Islands, Alaska, earthquake [Engdahl *et al.*, 1989]. Oppenheimer *et al.* [1989] studied microearthquakes in detail along the Calaveras fault, California. They identify stationary aseismic zones or asperities which are the sites of magnitude 5+ earthquakes. In their study, not only are aftershocks not associated with mainshock slip, but neither are foreshocks or other background seismic activity. The results of these studies indicate that fault zones in both strike-slip and subduction zone regimes are characterized by regions of distinctly different properties. Some areas slip more or less continuously with strain released in many smaller earthquakes and as fault creep. Other areas behave as asperities, which are locked sections of the fault, and release strain energy in significant earthquakes, neither experiencing foreshocks or aftershocks. An important component of seismic hazard evaluation in the Los Angeles area and elsewhere will be the identification of these asperity regions.

6. REFERENCES

Aki, K. (1979). Characterization of barriers on an earthquake fault, *J. Geophys. Res.*, **84**, 6140-6148.

Apsel, R.J., G.A. Frazier, A. Jurkevics and J.C. Fried (1981). Ground motion prediction for the Los Angeles basin from a major San Andreas earthquake, *U.S. Geol. Surv. Open File Rep.*, **81-276**, 217.

Bent, A.L. and D.V. Helmberger (1989). Source complexity of the October 1, 1987, Whittier Narrows earthquake, *J. Geophys. Res.*, **94**, 9548-9556.

Brady, A.G., P.N. Mork and L.C. Seekins (1988). Processed strong-motion records Whittier Narrows, California earthquake October 1, 1987, *U.S. Geol. Surv. Open File Rep.*, **88-354**.

Davis, T.L., J. Namson and R.F. Yerkes (1989). A cross section of the Los Angeles area: Seismically active fold and thrust belt, the 1987 Whittier Narrows earthquake, and earthquake hazard, *J. Geophys. Res.*, **94**, 9644-9664.

Eaton, J., R. Cockerham and F. Lester (1983). Study of the May 2, 1983 Coalinga earthquake and its aftershocks, based on the USGS seismic network in northern California, The 1983 Coalinga, California Earthquakes, edited by J.H. Bennett and R W. Sherburne, *Spec. Publ.*, **66**, 261-273, Calif. Div. of Mines and Geol., Sacramento.

Engdahl, E.R., S. Billington and C. Kisslinger (1989). Teleseismically recorded seismicity before and after the May 7, 1986, Andreanof Islands, Alaska, earthquake, *J. Geophys. Res.*, **94**, 15,481-15,498.

Etheredge, E. and R. Porcella (1987). Strong-motion data from the October 1, 1987 Whittier Narrows earthquake, *U.S. Geol. Surv. Open File Rep.*, **87-616**.

Hartzell, S. (1989). Comparison of seismic waveform inversion results for the rupture history of a finite fault: Application to the 1986 North Palm Springs, California, earthquake, *J. Geophys. Res.*, **94**, 7515-7534.

Hartzell, S. and T. Heaton (1983). Inversion of strong ground motion and teleseismic waveform data for the fault rupture history of the 1979 Imperial Valley, California, earthquake, *Bull. Seismol. Soc. Am.*, **73**, 1553-1583.

Hauksson, E. (1987). Seismotectonics of the Newport-Inglewood fault zone in the Los Angeles basin, southern California, *Bull. Seismol. Soc. Am.*, **77**, 539-561.

Hauksson, E. and L.M. Jones (1989). The 1987 Whittier Narrows earthquake sequence in Los Angeles, southern California: Seismological and tectonic analysis, *J Geophys. Res.*, **94**, 9569-9589.

Hauksson, E., et al. (1988). The Whittier Narrows earthquake in the Los Angeles metropolitan area, California, *Science*, **239**, 1409-1412.

Iida, M., T. Miyatake and K. Shimazaki (1988). Optimum strong-motion array geometry for source inversion, *Earthquake Eng. Struct. Dyn.*, **16**, 1213-1225.

Kawase, H. and K. Aki (1990). Topography effect at the critical SV wave incidence: possible explanation of damage pattern by the Whittier Narrows, California earthquake of October 1, 1987, *Bull. Seismol. Soc. Am.*, in press.

Lawson, C. and R. Hanson (1974). Solving least squares problems, Prentice-Hall, Englewood Cliffs, N.J.

Levine, M.B., J.L. Beck, W.D. Iwan, P.C. Jennings and R. Relles (1988). Accelerograms recorded at Caltech during the Whittier Narrows earthquakes of October 1 and 4, 1987: A preliminary report, *Rep. EERL 88-01*, Earthquake Engineering Res. Lab., Calif. Inst. of Technol., Pasadena.

Leyendecker, E.V., L.M. Highland, M. Hopper, E.P. Arnold, P. Thenhaus and P. Powers (1988). The Whittier Narrows, California earthquake of October 1, 1987 - Early results of isoseismal studies and damage surveys, *Earthquake Spectra*, **4**, 1-10.

Lin, J. and R. Stein (1989). Coseismic folding, earthquake recurrence, and the source mechanism at Whittier Narrows, Los Angeles basin, California, *J. Geophys. Res.*, **94**, 9614-9632.

Magistrale, H. and H. Kanamori (1989). A 3-D southern California velocity model to relocate Los Angeles basin earthquakes, *Seismol. Res Lett.*, **60**, 24.

Mendez, A.J. and J.E. Luco (1990). Steady state, near-source models of the Parkfield, Imperial Valley, and Mexicali Valley earthquakes, *J. Geophys. Res.*, **95**, 327-340.

Mendoza, C. and S.H. Hartzell (1988). Aftershock patterns and main shock faulting, *Bull. Seismol. Soc. Am.*, **78**, 1438-1449.

Olson, A. and R. Apsel (1982). Finite faults and inverse theory with applications to the 1979 Imperial Valley earthquake, *Bull. Seismol. Soc. Am.*, **72**, 1969-2001.

Olson, A.H., J. Orcutt and G. Frazier (1984). The discrete wavenumber/finite element method for synthetic seismograms, *Geophys. J.R. Astron. Soc.*, **77**, 421-460.

Oppenheim, A. and R. Schafer (1975). Digital signal processing, Prentice-Hall, Englewood Cliffs, N.J.

Oppenheimer, D.H , W.H. Bakun and A.G. Lindh (1990). Slip partitioning of the Calaveras Fault, California, and prospects for future earthquakes, California, *J. Geophys. Res.*, **95**, 8483-8498.

Rybicki, K. (1973). Analysis of aftershocks on the basis of dislocation theory, *Phys Earth Planet. Inter.*, **7**, 409-422.

Schwartz, S.Y., J.W. Dewey and T. Lay (1989). Influence of fault plane heterogeneity on the seismic behavior in the southern Kurile Islands Arc, *J. Geophys. Res.*, **94**, 5637-5649.

Shakal, A.F., M.J. Huang, C.E. Ventura, D.L. Parke, T.Q. Cao, R.W. Sherburne and R. Blazquez (1987). CSMIP strong-motion records from the Whittier, California earthquake of 1 October 1987, *Rep. OSMS 87-05*, Calif. Div. of Mines and Geol., Sacramento.

Stein, R.S. and M. Lisowski (1983). The 1979 Homestead Valley earthquake sequence, California: Control of aftershocks and postseismic deformation, *J. Geophys. Res.*, **88**, 6477-6490.

Stein, R.S. and R.S. Yeats (1989). Hidden earthquakes, *Sci. Am.*, **260**, 48-57.

Trifunac, M.D. (1988). The Whittier Narrows, California earthquake of October 1, 1987 - Note on peak accelerations during the 1 and 4 October earthquakes, *Earthquake Spectra*, **4**, 101-113.

Wald, D., P.G. Somerville and L.J. Burdick (1988). The Whittier Narrows, California earthquake of October 1, 1987 - simulation of recorded accelerations, *Earthquake Spectra*, **4**, 139-156.

Wetmiller, R.J., R.B. Horner, H.S. Hasegawa, R.G. North, M. Lamontagne, D.H. Weichert and S.G. Evans (1988). An analysis of the 1985 Nahanni earthquakes, *Bull. Seismol. Soc. Am.*, **78**, 590-616.