Determination of earthquake source mechanisms using teleseismic 30-140 s waves: The January 17, 1994, Northridge earthquake

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Abstract. Complete seismic waveforms with periods between 30 and 140 s recorded at teleseismic distances (≥ 10°) are used to determine the source mechanism of the January 17, 1994, Northridge earthquake (Ms = 6.8). An algorithm called the Earth simplifying transformation (EST) is applied to the mainshock (primary event) data to calibrate and remove propagation effects of Earth’s lateral and radial heterogeneities that are not predicted by a reference Earth model. The procedure involves frequency domain division of each pair of observed and synthetic waveforms for an auxiliary event with a well-constrained source mechanism in proximity to the primary event, with the synthetics being computed for a reference Earth model. If the auxiliary event source mechanism is correct, the differences between the data and synthetics represent the residual effects of the Earth’s structure that are not predicted by the Earth model. The frequency dependent deconvolution filters are convolved with the corresponding observed signals for the primary event. The resulting waveforms are then used to determine the source mechanism of the primary event by forward modeling and by centroid-moment-tensor inversion, with excellent fits to the complete waveforms being achieved. The resulting solution for the Northridge event has a centroid depth of 12 ± 1 km and a mechanism with strike, φ = 115°, dip, δ = 40°, and rake, λ = 95°, with 5° uncertainty for each parameter. The EST procedure holds promise for reliable source inversion for small earthquakes in a region calibrated by large events with well-determined source parameters.

Introduction

The January 17, 1994, Northridge (Ms = 6.8) earthquake ruptured a south dipping fault beneath the San Fernando Valley in the Los Angeles region [e.g., Hauksson et al., 1995]. This was the most damaging large earthquake in the region, which has suffered an increased level of seismicity in the past decade. Recent progress in the worldwide deployment of broadband seismic stations allows the seismic radiation from earthquakes in southern California to be analyzed with unprecedented azimuthal coverage. For the assessment of regional seismic hazards, source mechanisms of large earthquakes in this region are routinely determined using various data sets, typically including P wave first motions, and body and surface waves recorded at local, regional, and teleseismic distances [e.g., Bent and Helmberger, 1989; Bent et al., 1989; Kanamori et al., 1990; Wald et al., 1990; Patton and Zandt, 1991; Romanowicz et al., 1992, 1993, 1994; Kanamori et al., 1992; Uhrhammer, 1992; Uhrhammer et al., 1994; Zhao and Helmberger, 1994; Dreger, 1994; Risema and Lay, 1995; Thio and Kanamori, 1995; Pasyanos et al., 1996].

Table 1 lists hypocentral parameters of the Northridge earthquake (hereinafter referred to as event I) along with focal mechanisms obtained previously using three methods: P wave first motions, regional waveform inversion, and routine centroid-moment-tensor (CMT) inversion, which are referred to as PFM, RWI and CMT mechanisms, respectively. The regional waveform solution obtained by Dreger [1994] is referred to as RWI1 and that by Song et al. [1995] to as RWI2. Figure 1 shows that there are moderate differences between these estimated source mechanisms for the Northridge earthquake. Precise determination of the rupture geometry and slip distribution requires high precision in the point-source model [e.g., Wald et al., 1996], so it is important to resolve which model is optimal and to reduce model errors which may bias the solutions.

Table 1 also lists mechanisms for two nearby events: an aftershock (Ms = 6.0) of the Northridge event, and the June 28, 1991, Sierra Madre earthquake (Ms = 5.2) (hereinafter referred to as events II and III, respectively). Events II and III are located 19 km N50°W and 50 km N84°E from event I, respectively. In the regional waveform inversions, the data for event III included Pp, Sd, and Ss waves, while for events I and II both body and surface waves (with periods 20-100 s) were used. In the CMT inversions, the data used for events II and III are long-period body waves, while the data for event I include surface waves with periods longer than 135 s and long-period body waves.

The source mechanisms for each event listed in Table 1 differ mainly in dip angle and strike direction. For the south dipping nodal planes, the dip angle ranges from 33° to 33°, 43° to 45°, and 41° to 43° for events I, II, and III, respectively; and the strike ranges from 105° to 130°, 100° to 116°, and...
Table 1. Source Parameters of Earthquakes Used in This Study

<table>
<thead>
<tr>
<th>Event</th>
<th>Origin Time, UTC</th>
<th>Location</th>
<th>Depth, km</th>
<th>( M_s )</th>
<th>Strike/dip/slip</th>
<th>First Motion Solution</th>
<th>Regional Wave Solution</th>
<th>CMT Solution</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>Jan 17, 1994, 1330:55.4</td>
<td>34:21(^{3})N, 118:53(^{6})W</td>
<td>18.2</td>
<td>6.8</td>
<td>105/35/105; 273/56/83</td>
<td>100/45/70; 307/48/109</td>
<td>242/50/82; 74/41/99</td>
<td>34.22(^{4})N, 118.70(^{6})W</td>
</tr>
<tr>
<td>II</td>
<td>Jan 17, 1994, 2333:30.7</td>
<td>34:32(^{6})N, 118:69(^{8})W</td>
<td>9.8</td>
<td>6.0</td>
<td>121/43/105; 280/49/76</td>
<td>116/43/92; 294/47/88</td>
<td>235/50/74; 79/43/108</td>
<td>34.22(^{4})N, 118.70(^{6})W</td>
</tr>
<tr>
<td>III</td>
<td>June 28, 1991, 1443:54.5</td>
<td>34:26(^{6})N, 118.00(^{6})W</td>
<td>12.0</td>
<td>5.2</td>
<td>128/33/106; 289/53/80</td>
<td>115/43/89; 296/45/91</td>
<td>223/38/38; 93/43/130</td>
<td>34.26(^{4})N, 118.00(^{6})W</td>
</tr>
</tbody>
</table>

The origin time, hypocentral location and depth, and \( M_s \) are from the Southern California Seismographic Network operated by California Institute of Technology and U.S. Geological Survey.

\(^a\) Hauksson [1994, also personal communication, 1994], Hauksson et al. [1995].
\(^b\) Dreger and Heimberger [1991]; Dreger [1994, also personal communication, 1994].
\(^c\) Song et al. [1994].
\(^d\) Dziewonski et al. [1992, 1994].

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Figure 1. Epicentral location (pluses) and source mechanisms of the 1994 Northridge (I. mainshock; II. aftershock) and 1991 Sierra Madre (III) earthquakes (Table 1). The mechanism for event II is the solution from regional wave inversions [Dreger, 1994], and for event III the solution is from P wave first motions. First motion data for the Northridge mainshock were obtained by Hauksson et al. [1995] and are shown below in an equal-area projection with the confidence level indicated by the size of each symbol. Three focal mechanism solutions listed in Table 1 are also shown (regional wave solution is from Dreger [1994]).
73° to 93° for events I, II, and III, respectively. Note the common tendency for the strike direction in the CMT mechanisms to be rotated clockwise by 15° to 25° relative to the PFM mechanisms. The estimated uncertainties of the dip and strike angles for PFM and RW11 mechanisms are in general less than 5° (E. Hauksson and D. Dreger, personal communication, 1994). This is small in comparison with the variations in estimated mechanisms.

For the Northridge mainshock, the differences between point-source mechanism estimates obtained from various methods (Figure 2) may be attributed to uncertainties or biases in these estimates, which must be identified and reduced when possible, or attributed to mechanism changes in the rupture process. Hauksson et al. [1995] pointed out that there is an increase in dip along strike in the distribution of aftershocks, and they attributed the difference between the mainshock mechanisms obtained from various methods to a curved rupture surface. However, comparable differences between the estimated source mechanisms are also found for relatively small events in this region, such as events II and III (Table 1), which suggests that the inconsistency may instead reflect model errors associated with the different methods. Systematic mechanism inconsistencies are often found in the Preliminary Determination of Epicenters Monthly Listings for earthquakes in southern California, in particular, for events with large magnitude ($M_e \geq 6.5$). Usually, the formal uncertainty of the mechanism for each method is much smaller than the differences between the mechanisms, reflecting the fact that the formal uncertainties do not take into account model uncertainties.

In addition to differences between source mechanism estimates for the Northridge mainshock, there are also moderate differences between the centroid depth estimates obtained in various studies. Thio and Kanamori [1996] obtained a source model from inversions of teleseismic body wave data, which consists of three subevents located at depths of 19, 17.5, and 14 km with more than 75% of the seismic moment being released by the two deeper subevents. This is consistent with the CMT depth of 16.8 km. Studies of strong motions by Wald and Heaton [1994] and regional waveforms by Dreger [1994] also favor slip occurring at large depths. However, analyses of geodetic data by Hudnut et al. [1996] and Shen et al. [1996] prefer a shallower centroid depth (about 10 km).
The combined model for displacement time histories obtained by Wald et al. [1996] from joint inversion of strong-motion, teleseismic, Global Positioning System (GPS), and leveling data shows that large displacements were concentrated at shallow depths updip from the hypocenter. Using this model, we estimated the centroid depth at between 11 and 12.5 km (Table 2 and Figure 2), which is slightly deeper than the 10 km centroid depth obtained from the geodetic data analysis. This difference may be attributed to the difficulty of recovering slip near the downdip edge of the rupture plane with the available GPS data [Hudnut et al., 1996]. However, it remains to be seen whether the resolving power of teleseismic data can be improved to allow us to obtain a shallow centroid depth consistent with that for combined seismological and geodetic data.

We previously performed a CMT-type analysis of the source mechanism of the Northridge earthquake, in which we used the laterally homogeneous preliminary reference Earth model (PREM) [Dziewonski and Anderson, 1981], and found strong data dependence of the solution [Zhang et al., 1994]. For many stations, the observed data differ significantly from the model predictions. In order to reduce the biases due to unmodeled effects of Earth’s structure, we incorporated recently developed three-dimensional models of the Earth in our long-period surface wave CMT inversions, basically replicating the procedure currently used at Harvard. The aspherical Earth model corrections did not reduce the discrepancies to a significant extent. We infer that even for long period (> 135 s) surface waves the existing aspherical models cannot eliminate all actual propagation effects, and this certainly holds for the shorter period signals.

In this study we use 30 to 140 s surface waves recorded at teleseismic distances to further constrain the source mechanism of the 1994 Northridge earthquake. Our data set consists of Global Seismographic Network seismic waves recorded at epicentral distances larger than 10° for the 1994 Northridge mainshock and two events near the mainshock epicenter (Figure 1). The data are dominated by high signal-to-noise ratio surface waves, which must be accurately corrected for propagation effects. At present this can only be done empirically. Our objectives are to exploit the neglected surface wave information to refine the Northridge source mechanism and to develop a general strategy for including this energy in systematic inversions for smaller events in a calibrated region.

The procedure followed consists of the following steps. First, we compute synthetics for a reference Earth model using an a priori, well-constrained focal mechanism for a nearby auxiliary event. By deconvolving the synthetic signals from data for the auxiliary event, we obtain complex transfer functions or deconvolution filters that represent the propagation effects which are not predicted by the Earth model. This is equivalent to complete waveform calibration of wave propagation effects for each path to optimize the performance of source inversion methods. By removing the empirically determined propagation effects relative to a known model from the observed waveforms for the Northridge mainshock (primary event), we obtain simplified waveforms, which are intrinsically well modeled by the reference Earth model. These waveforms are then compared with the synthetics for the reference Earth model using several source mechanism solutions obtained previously for the Northridge mainshock, along with waveform inversion of the complete signals. The excellent fit to the waveforms that is obtained indicates the general utility of this method for high-precision focal mechanism determination. This approach constitutes a complete waveform generalization of empirical path correction methods such as those of Weidner and Aki [1973] and Patton [1980].

### Method

Following Gilbert and Dziewonski [1975], the spectrum of a component of ground motion excited by a point source at angular frequency \( \omega \) for a reference Earth model may be given by

\[
u_k(x, \omega) = \sum_{i=1}^{6} \psi_{ki}(x, x_i, \omega) f_i(\omega)
\]

where \( u_k \) is the \( k \)th record in a set of seisograms, with the receiver at position \( x \) and the source at \( x_i \); \( \psi_{ki} \) are excitation kernels, and the \( f_i \) represents six independent components of the moment-rate tensor. For another event located at \( x_i' \) with moment-rate tensor \( f_i' \), the spectrum may be given by

\[
u_k'(x, \omega) = \sum_{i=1}^{6} \psi_{ki}(x, x_i', \omega) f_i'(\omega).
\]

Taking into account our imperfect knowledge of Earth structure and background noise, the spectra of observed ground motion for these events may be expressed as

\[
U_k(x, \omega) = u_k(x, \omega) \alpha_k(x, \omega) + \epsilon_k(x, \omega)
\]

\[
U'_k(x, \omega) = u'_k(x, \omega) \alpha'_k(x, \omega) + \epsilon'(x, \omega)
\]

where \( \alpha \) and \( \alpha' \) represent effects of the deviation of the Earth's structure from the Earth model, assuming that source parameters are perfectly known. In the following analysis the noise terms \( \epsilon \) and \( \epsilon' \) are ignored and the event at \( x_i \) is referred to as the primary event, while the event at \( x_i' \) is an auxiliary event. In conventional algorithms, the moment-rate tensor of the primary event, \( f_i \), is determined from data for the event, \( U_i \), by solving the following equation

### Table 2. Comparison of Source Models for the Northridge Mainshock

<table>
<thead>
<tr>
<th>Method</th>
<th>Centroid Depth, km</th>
<th>Strike</th>
<th>Dip</th>
<th>rake</th>
<th>Moment, 10^3 N m</th>
</tr>
</thead>
<tbody>
<tr>
<td>COM</td>
<td>11-12.5</td>
<td>122°</td>
<td>40°</td>
<td>101°</td>
<td>1.4</td>
</tr>
<tr>
<td>GPS</td>
<td>10.1</td>
<td>110°</td>
<td>41°</td>
<td>91°</td>
<td>1.1</td>
</tr>
<tr>
<td>EST</td>
<td>12</td>
<td>115°</td>
<td>40°</td>
<td>95°</td>
<td>1.2</td>
</tr>
</tbody>
</table>

COM, Wald et al. [1996]; GPS, Hudnut et al. [1996]; EST, this study.

- a Inferred from aftershock distributions.
- b Moment-tensor solution has a nondouble couple component less than 1%.
which is correct to zeroth order in terms of $\alpha$, since implicit in the algorithms is the assumption

$$\alpha_k(x, \omega) = 1.$$  

(5)

The question that we shall address in this study is as follows. Given a set of seismograms, $U_k$ and $U_k'$, at common stations from two events, which are located in proximity (in comparison with the event station distance), along with synthetics for the auxiliary event, $U_j'$, computed for a given (correct) focal mechanism and Earth model, is it possible to determine the moment-rate tensor of the primary event, $f_j$? For this purpose we define the following transformation

$$u_k(x, \omega) = u_k'(x, \omega) U_k(x, \omega) / U_k'(x, \omega),$$

(6)

which is referred to as the Earth simplifying transformation (EST) in the following analysis. If there is good correlation between the data for the primary and auxiliary events despite their differences in polarity and instantaneous phase for various wave packets, we may assume

$$\alpha_k(x, \omega) = \alpha_k'(x, \omega).$$

(7)

In this study, our data sets for various events listed in Table 1 show good correlations, and the excellent fit to the simplified signals that is obtained indicates the reliability of this assumption. The accuracy of this assumption is influenced by the proximity of the events and the frequency content of the data. Given multiple auxiliary events in a region, one could average deconvolution filters or spatially interpolate them to strengthen this approximation, and to reduce error due to incorrect auxiliary event source parameters. Assuming (7), the right-hand side of (6) becomes $u_k(x, \omega)$, thus the spectrum of the EST waveform may be expressed as

$$u_k^2(x, \omega) = \sum_{i=1}^{s} \psi_i(x, x_i, \omega) f_i(\omega)$$

(8)

which is correct to first order in terms of $\alpha$ and will be used to determine the moment-rate tensor ($f_j$). For simplicity of analysis, we adopt a frequently used assumption: the $f_j$ are considered to be independent of frequency except for a common correction for an assumed duration of the source and are regarded as the moment tensor.

In this study, various experiments were made to explore the sensitivity of primary event source parameters obtained from inversions of EST waveforms to the prescribed source parameters of the auxiliary event. Clearly, errors in auxiliary event centroid estimates (time, location, and depth) will project into the primary event solution, influencing not only centroid but also source mechanism estimates, so good independent constraints on the auxiliary event centroid parameters are critical. However, an interesting and important question is how errors in the auxiliary event source mechanism project into the primary event solution.

To address this question, we use an arbitrarily given moment tensor $g'_j$ in the calculation of synthetics for the auxiliary event. If $f_j$ is the true moment tensor of the auxiliary event, the errors in the auxiliary event source mechanism are $g'_j - f_j$. The errors can be also described by the linear relation:

$$g'_j = f_j d_i, \quad (i = 1, ..., 6).$$

Then the EST waveforms corresponding to the incorrect auxiliary event moment tensor $g'_j$ may be written as

$$u_k^2(x) = \sum_{i=1}^{s} \psi_i(x, x_i) g_i,$$

(9)

with $g_i = f_i d_i, \quad (i = 1, ..., 6)$, where $f_i$ is the true moment tensor of the primary event. Clearly, one will recover a biased source mechanism for the primary event, with linear dependence on the auxiliary event solution.

However, since both the EST waveforms in (9) and (8) correspond to the same set of excitation kernels, inversions of biased EST waveforms (9) will yield the same centroid estimates as for correct EST waveforms (8). Therefore, errors in the auxiliary event source mechanism do not project into errors in the primary event centroid estimates. It follows that using well-determined centroid parameters of the auxiliary event, reliable centroid parameters of the primary event can be obtained even by inverting EST waveforms that are computed for an arbitrarily given auxiliary event moment tensor. Our experiments on the Northridge earthquake data set confirm that variation of the prescribed auxiliary event moment tensor does not affect the primary event centroid estimate.

The linear dependence of the derived primary event moment tensor on the prescribed auxiliary event moment tensor allows us to estimate any possible biases in the primary event moment-tensor solution, if the biases in the auxiliary event moment tensor are determined. This is achieved by first estimating bias factors $d_i$ for the prescribed auxiliary event moment tensor and then calculating the resulting errors in the primary event moment-tensor solution: $g_i - g_i / d_i, \quad (i = 1, ..., 6)$.

**Application to Teleseismic Surface Waves**

In comparison with conventional algorithms, in which the moment tensor of an event is directly determined using data for the event itself, the EST algorithm introduced here requires additional data from the auxiliary event, which must be located at a close distance to the primary event (in comparison with the event station distance) and must have a known moment tensor. Note that the relaxation of the strong assumption (5) involved in conventional algorithms allows us to make use of all seismic waves, in particular surface waves with periods less than about 140 s; these waves are typically so strongly affected by the complexity of the Earth's structure, including both lateral and radial heterogeneities, that direct inversion of the short-period surface waves is not possible.

In this section we first show several EST waveforms obtained for the Northridge mainshock (the primary event) using events II and III as auxiliary events. Note that the focal mechanism of event III (the 1991 Sierra Madre earthquake) is characterized by nodal planes dipping in the NW-SE direction, in clear contrast with the NE-SW dipping directions of nodal planes for the Northridge mainshock (Figure 1). Using two auxiliary events with different locations, depths, and source mechanisms allows us to examine the validity of the assumptions in the proposed algorithm and to explore any biases in the results due to our choice of auxiliary event.

Figure 3 shows displacement waveforms, band-pass filtered between 7 and 30 mHz, recorded at station WMQ from events I and III for vertical and radial components and from events I and II for the transverse component (the transverse component at this station from event III is very noisy and is not shown here, but transverse components for many other stations were used in our analysis). The station is located at teleseismic distance from the source region, and the waveforms are dominat-
ed by large-amplitude fundamental mode surface waves. Comparison of the waveforms indicates that for each component there is good coherence between the observed waveforms for the primary and auxiliary events. This indicates that for this station the signals are primarily shaped by the common propagation effects, implying that (7) is a good approximation for paths to this station.

Figure 3 also shows the synthetics computed for Earth model PREM for the Northridge mainshock (primary event) using the PFM mechanism (dotted lines). It is clear that there is considerable discrepancy between the arrival time and dispersion characteristics of the observed waveforms \( Z_I, R_I, \) and \( T_I \) and synthetics for the primary event. Synthetics calculated using other focal mechanisms listed in Table 1 for the Northridge mainshock have waveforms similar to those for the PFM mechanism, but for several stations there are significant differences in amplitudes. The amplitude effects will be discussed later, but it is clear that direct inversion of these signals using the PREM model will fail to resolve the source. For this reason, standard CMT inversions completely filter out all of the surface wave signal seen in Figure 3.

Figure 3 further shows the EST waveforms, with \( Z_{II}^M, R_{II}^M, \) and \( T_{II}^M \) (solid lines) corresponding to vertical, radial, and transverse components, respectively, where the superscript indicates the auxiliary event used. In the calculation of the EST waveforms, the synthetics for the auxiliary event are computed for a point source with a step-time function located at the hypocenter with the PFM mechanism used for event III and the RWII mechanism used for event II. Note that for \( Z_{II}^M \) and \( R_{II}^M \), although the mechanism of the auxiliary event (event III) is very different from that of the primary event (Figure 1 and Table 1), the fit of the PREM synthetics to the EST waveforms is excellent, demonstrating the validity of assumption (7).

In the calculation of each EST waveform, the convolution and deconvolution of data and synthetics are performed in the frequency domain, and truncation is used when peaks and holes of the spectra for the two events mismatch, resulting in exceptionally large peaks in the EST spectra. Other common techniques, such as smoothing or filtering, may also be used to reduce the instabilities associated with deconvolution.

Now we compare EST waveforms with the synthetics predicted for various source mechanisms. Our objective is to find a source mechanism best matching the EST waveforms. This source mechanism may be found using a forward modeling approach or an inversion approach by solving (8). The question that we will address in the following is whether the EST algorithm enables us to use the shorter period surface wave energy to reliably determine the source mechanism solution.
We found that for most of the stations the data from event II have higher signal-to-noise ratio than for event III. This may be related to weather conditions in the summer for many stations for event III or may reflect the difference in magnitude. The event II data set is also much larger, since it includes many recently deployed stations. In the following analysis we use the EST waveforms derived with event II as the auxiliary event, although the EST waveforms obtained with event III as the auxiliary event yield essentially the same results, indicating the reliability of source parameters of these auxiliary events. In the calculation of the EST waveforms, the synthetics for event II are computed for a point source with a step-time function located at the hypocenter and the corresponding RW11 mechanism. Given several auxiliary events, one could stack the EST waveforms to reduce sensitivity to auxiliary event mechanisms, but we have not attempted that here.

We compare the EST waveforms with synthetics computed for the primary event for a point source located at the epicenter with a half duration of 5 s at a depth of 14 km. A 14-km centroid depth was obtained from the regional waveform inversions for the Northridge mainshock [Dreger, 1994].

Now we determine which mechanism listed in Table 1 best matches the EST signals for the mainshock. In comparison with the RW11 model the RW12 model differs more from the fault geometry that is inferred from the aftershock distribution (Figure 2); for simplicity the RW12 model is not discussed here. Figure 4 shows the transverse components of the EST waveforms, along with synthetics for several stations in the azimuth range 3° to 12°. These signals are dominated by Love waves. For each station the synthetics for different source mechanisms have similar waveforms, indicating that variations of the instantaneous phase for surface waves are not sensitive to the modest difference between these source mechanisms. However, the synthetic amplitudes vary substantially. The amplitudes of the synthetics for the RW11 mechanism (mechanism B in Figure 4) have the best agreement with the EST amplitudes. For other stations, differences between amplitudes of the synthetics for these different source mechanisms are smaller than for the stations shown in Figure 4.

In order to examine the sensitivity of the amplitudes of the synthetics to the dip, strike, and rake of the south dipping nodal plane, we computed synthetic seismograms for many more source mechanisms. We found that for the transverse component seismograms at the stations shown in Figure 4, the amplitudes of the synthetics are mainly sensitive to the strike of the nodal plane, reflecting the fact that these stations are located near the nodal direction of the Love wave radiation pattern for the Northridge earthquake. This indicates that the amplitudes of Love waves are useful for resolving the strike direction for thrust fault earthquakes when the EST algorithm is used. For the transverse components of EST waveforms our forward modeling analysis indicates that the strike of the south dipping nodal plane is between 110° and 120°, compatible with the aftershock distributions (Figure 2).

The EST waveforms shown in Figure 4 are computed using the RW11 mechanism for the auxiliary event. We also computed EST waveforms using the PFM mechanism for the auxiliary event, which differs in the strike of the south dipping nodal plane by 16° from the RW11 mechanism. We found that the amplitudes of the EST waveforms are not very dependent on this choice, demonstrating the stability of EST waveforms relative to small differences in the source mechanism for the auxiliary event. This indicates that using EST waveforms to constrain the source mechanism of the primary event is a robust approach, as long as uncertainties in the auxiliary event mechanism are not too large.

Because of the nonlinear relation between data and fault parameters strike, dip, and rake, it is very complicated to quantify error estimates for the primary event fault parameters associated with small perturbations in the auxiliary event fault parameters; but this is offset by the advantage of being able to quantify error estimates for moment tensor elements, as discussed in the previous section, and by being able to very closely model the EST waveforms.

We also model vertical components of the EST waveforms. The vertical and radial components are dominated by Rayleigh waves. For the data set used in this study, the radial components have poor signal to noise ratios and are not used for the forward modeling. Figure 5 shows the vertical components of the EST waveforms along with synthetics computed for different source mechanisms for stations at various azimuths.

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**Figure 4.** Transverse components of EST waveforms ($T_{II}^i$) for the Northridge mainshock at various stations, which are obtained with the synthetics for the auxiliary event (II) being computed for the focal mechanism from regional wave inversions. Amplitudes are given in parentheses in units of microns (amplitudes of $T_{II}$ with the synthetics computed for the focal mechanism from $P$ wave first motions are shown in brackets for comparison). Symbols A, B, and C indicate synthetics computed for the Northridge mainshock for a seismic moment of $1.2 \times 10^{19}$ N m and various focal mechanisms listed in Table 1: first motion solution, A; regional wave solution, B; and CMT solution, C.
For comparison with the forward modeling analysis described above, we inverted the EST waveforms for the moment tensor using the CMT inversion method [Dziewonski et al., 1981]. Our data set included vertical, transverse, and radial components for a total of 31 stations and 64 channels. Many stations clustered along the azimuth range between 210° and 240° (mostly in China) were eliminated to obtain uniform azimuthal coverage. Our moment tensor solution (hereinafter referred to as EST solution) is essentially a double couple with strike, φ = 115°, dip, δ = 40°, and rake, λ = 95° for the south dipping nodal plane (the fault plane) and φ = 289°, δ = 50°, and λ = 86° for the north dipping nodal plane (Figure 2 and Table 2). The EST solution has a nondouble couple component less than 1%, which is smaller than for the routine CMT solution (about 5%).

This mechanism is in close agreement with the combined source model obtained by Wald et al. [1996] from joint inversion of strong-motion, teleseismic, and GPS data, and the source model determined independently by Hudnut et al. [1996] from GPS data, and aftershock distribution. It differs somewhat from solutions for P wave first motions and routine CMT inversion (Figure 2 and Table 2). Using the aftershock distribution, we estimated the fault dip at between 37° and 42° and fault strike at between 105° and 120° (Figure 2). In the combined source model of Wald et al. [1996], the rake is independently determined, while the strike and dip are inferred from the aftershock distribution. By comparing the EST solution with the combined source model, the fault model from the geodetic data analysis, and aftershock distributions, we estimate the uncertainties for the strike, dip, and rake in our EST solution at less than 5° (Figure 2).

The centroid depth from the inversion is about 12 km, which is 2 km deeper than for the auxiliary event (event II). This depth is in excellent agreement with the combined source model obtained by Wald et al. [1996] but is slightly deeper than the 10 km centroid depth obtained by Hudnut et al. [1996] from GPS data and shallower than the 14 km centroid depth obtained by Dreger [1994] from regional waveform inversions (Figure 2). We estimate the uncertainty of the depth in the EST solution at less than 2 km, although this is weakly dependent on our choice of PREM as a reference model. As discussed previously, the uncertainty of the estimate of the primary event centroid depth reflects uncertainties of the auxiliary event centroid estimates and is independent of errors in the prescribed source mechanism of the auxiliary event.

The centroid location obtained from the CMT inversion of EST waveforms is about 5 km to the north of the epicenter, which is also in close agreement with the direction of main rupture directivity [Wald et al., 1996; Zeng and Anderson, 1996]. The epicentral (or centroidal) estimate obtained here is subject to errors in the prescribed source depth and mechanism of the auxiliary event; however, it is consistent with the analysis of combined strong-motion, teleseismic, and GPS data. While the EST results for the Northridge event are not surprising, they do represent a validation of the EST methodology and a refinement of the teleseismic point-source solution which brings it to better agreement with independent constraints. The methodology exploits the mechanism and depth sensitivity of 30-140 s surface waves, which are usually ignored. The methodology offers promise in application to remote areas where large events can be used as auxiliary events to process solutions for small events recorded at only a few stations.

Figure 5. Similar to Figure 5 for vertical components of EST waveforms for various stations.
Discussion

The significance of using the EST procedure in earthquake source studies is the reduction of propagation effects of Earth’s lateral and radial heterogeneities that are not predicted by a reference Earth model. The path calibration is made for complete waveforms and for each path to optimize the performance of source inversion methods. The calibration allows us to make use of 30-140 s waves, which may be so strongly affected by the complexity of the Earth’s structure that direct inversion of the signals is not possible.

The EST algorithm can be applied to analysis of source parameters of a large earthquake using a small earthquake with well-constrained source mechanism as the auxiliary event, as demonstrated in this study. However, for optimal performance of the algorithm it is preferable to use an auxiliary event which is larger than the primary event. Since errors in auxiliary event data will project into EST waveforms, using a large auxiliary event allows us to enhance the signal-to-noise ratio in the EST waveforms. Moreover, source mechanisms of large events are usually more reliable, because they can be better constrained by larger number of recordings. In this study we used two earthquakes with different centroid and source mechanism estimates as the auxiliary event, the purpose being to test the reliability of the EST algorithm. However, the algorithm requires only one auxiliary event.

The EST procedure holds promise for reliable source inversion for small earthquakes in a calibrated region. Since small events of $m_o$ of 4 or less are recorded only at local or regional distances and only have good signal-to-noise ratio in passbands with periods less than about 40 s, path calibration is essential. In practice, it is necessary to use an accurate model of the crustal structure in the calculation of synthetics for the auxiliary event in order to exploit the short-period energy. The crustal structure can be estimated using data from large, auxiliary events in the region. The source mechanisms of these events must be independently determined, perhaps by teleseismic signals, to avoid circularity in modeling the regional waveforms. This is possible because large events produce adequate teleseismic signals. Depths of the events can be constrained robustly using teleseismic body waves and regional S-P times. If one has a suite of large events with well constrained source parameters, the EST method can be used to develop spatially varying deconvolution filters to allow small events to be inverted throughout the region.

This study was partly motivated by the fact that source inversions of teleseismic surface waves with aspherical model corrections do not always yield a solution compatible with independently determined solutions, and partly by a desire to use an unexploited portion of the waveforms. In our previous CMT-type analysis of the source mechanism of the Northridge earthquake (Zhang et al., 1994), we found that for body waves low-pass filtered with a 30-s cutoff period and surface waves low pass filtered with a 135-s cutoff period, separate inversions gave a body wave mechanism that is significantly rotated relative to the surface wave mechanism, while a simultaneous inversion yielded intermediate parameters, consistent with the independent solution obtained from regional waveform inversions. One interpretation for the discrepancy between body and surface wave inversions involves changes in the source mechanism during rupture, sensed to a different degree by different wave types. Alternatively, the inconsistency may reflect different model errors in the inversions.

We also found discrepancies between inversions of only Rayleigh waves and joint inversions of Rayleigh and Love waves. While joint inversions yield solutions with a shallow-dipping north plunging nodal plane, Rayleigh wave inversions prefer solutions with a shallow-dipping south plunging nodal plane. This discrepancy is most likely due to an inaccuracy of the Earth model involved in the source determination. Love waves usually have smaller signal-to-noise ratio than Rayleigh waves, thus including Love waves in inversions may increase uncertainty relative to inversions that use only Rayleigh waves. Love waves also commonly show larger deviations from predictions for standard Earth models than do Rayleigh waves. The greater variance in Love wave observations may cause different biases in resolving the true source radiation pattern, resulting in an apparent Rayleigh-Love discrepancy. Therefore including Love waves in inversions does not necessarily provide improved source mechanism resolution, unless propagation errors are reduced to a comparable level. Aspherical Earth model corrections did reduce the discrepancies between inversions of different surface wave data sets, but not to a significant extent. The limited accuracy of the aspherical corrections at long period (150 s and longer) and the low signal-to-noise ratio for the long-period Love waves still preclude a consistent inversion. Similar problems were analyzed by Kage et al. (1996) for the July 12, 1993, Hokkaido-Nansei-Oki, Japan, earthquake.

The objective of the procedure introduced in this study is to reduce the biases due to the effects of Earth’s lateral and radial heterogeneity in the source determination and to extend the passband beyond that possible with existing low-resolution aspherical models. The key to success is independent knowledge of the auxiliary event’s source mechanism and depth. The advantage of the method is that it enables use of much more complete waveform information than possible with theoretical propagation models.

In comparison with the discrepancy discussed above between results obtained from observed Rayleigh and Love waves, more consistent results are obtained from FST waveforms for only Rayleigh waves and from EST waveforms for both Rayleigh and Love waves. The EST approach appears to work better for stations at short distances, presumably due to both the higher signal-to-noise ratio and reduced levels of highly path dependent multipathing. The method allows broadband surface wave information to contribute to the source inversion and can complement empirical Green’s function methods used for estimating the source time function (e.g., Murphy, 1977; Harzfell, 1978; Frankel et al., 1986; Ammon et al., 1993; Li and Toksoz, 1993).

The limitations of the EST algorithm involve the fact that the source mechanism of the auxiliary event must be known a priori and that event must be large enough so that 30-140 s surface waves arc well recorded at stations that also record the primary event. For remote areas, using large earthquakes reported in the Harvard CMT catalog is often the only choice for the auxiliary event. In addition, the spectral division in the calculation of the EST waveform often limits the resolution of the method. The choice of method used for reducing error in the deconvolution is subject to how well the detailed characteristics of the signal and noise are known. In order to overcome these limitations, the EST algorithm may be extended to include multiple auxiliary events to optimize the deconvolution filter for the primary event; and it is also desirable to develop highly accurate deconvolution techniques.
Although routine CMT analysis or P wave first motion analysis can provide acceptable source parameter determinations for study of global and regional seismicity, we feel that it is warranted to identify the error bounds associated with various determinations and to resolve differences between these routine determinations. This is particularly valuable when there are other types of seismological and geophysical data available for important events such as the Northridge earthquake. Using this event as a calibration of the EST procedure, we find a solution consistent with strong motions, teleseismic and regional waveforms, geodetic measurements, and aftershock distributions. This gives us confidence for application of the methodology to other events and contributes to the overall characterization of the Northridge earthquake.

Conclusions

The Earth simplifying transformation (EST) technique allows complete teleseismic waveform information in the period range 30 to 140 s to be used to constrain earthquake source parameters. Normally, the propagation effects of Earth’s lateral and radial heterogeneities are so severe that teleseismic intermediate- and short-period surface waves are not used to study the earthquake source despite their superb signal to noise ratio and sensitivity to source depth. The method involves empirical path calibration and requires independent constraints on the source parameters of at least one auxiliary event. The method can be applied to events either larger or smaller than the calibration event, and regional or teleseismic phases can be used. Using two well-constrained, moderate-size earthquakes in southern California, we apply the EST method to successfully resolve the source parameters for the January 17, 1994, Northridge, California earthquake. Our solution, obtained using wave field information that is usually filtered out of standard inversions, is in close agreement with the results of a joint inversion of strong-motion, teleseismic, and GPS data. It differs somewhat from solutions for P wave first motions and routine CMT inversion and appears to eliminate biases affecting these solutions. The EST procedure enables much more complete teleseismic waveform information to be used in constraining the source mechanism of moderate size earthquakes ($M_s > 4$), as long as recordings for a nearby event with a well-determined mechanism are available. The auxiliary event may be either larger or smaller than the primary event, and multiple auxiliary events can be used to spatially interpolate the deconvolution filters. The EST method holds promise for enabling complete waveform inversion for small- and moderate-size events in a region calibrated by a few well-studied auxiliary events.

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