Empirical Green Function Deconvolution
of Broadband Surface Waves: Rupture Directivity
of the 1992 Landers, California ($M_w = 7.3$), Earthquake

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Abstract  We present a simple method to determine gross temporal and spatial characteristics of faulting that can uniquely define the fault plane for a relatively large earthquake. The technique involves deconvolution of teleseismic and regional broadband surface waves (and/or body waves) of a small event (an empirical Green function) from the corresponding signals at the same stations for a large mainshock. This deconvolution corrects each mainshock seismic record for propagation and instrument response, and if the focal mechanisms and centroid depths are identical, deconvolution provides relative source time functions for the mainshock. Azimuthal variations of source time function width or subevent interference reveal the mainshock rupture directivity. Changes in the source mechanism during rupture can be evaluated for some classes of events. Our empirical Green function procedure is novel in its application to surface waves and its use for large ($M \geq 7$) earthquakes. We perform synthetic tests to establish intrinsic limitations of the method. The procedure requires very little data processing and can be applied in near-real time with the current distribution of seismic stations that have dial-up data retrieval capability. It also provides a straightforward procedure for reducing bias in source duration estimates resulting from aspherical heterogeneity and for detecting anomalous long-period radiation. We apply the procedure to the 28 June 1992 Landers, California ($M_w = 7.3$), earthquake. The Landers earthquake source duration lasted about 25 sec, and was dominated by two subevents with predominantly north-northwestward rupture extending 60 to 70 km. The subevents show counterclockwise rotations of at least 12° in strike that correlate well with mapped surface rupture.

Introduction
Routine analysis of all moderate-to-large earthquakes using Centroid Moment Tensor (CMT) approaches (e.g., Dziewonski et al., 1981; Dziewonski and Woodhouse, 1983; Giardini et al., 1993; Ritsema and Lay, 1993) and body-wave modeling procedures (e.g., Sipkin, 1982; Sipkin and Needham, 1992) provides estimates of the basic faulting parameters for most significant earthquakes within a few months of their occurrence. These point-source moment tensor solutions provide the average faulting geometry, seismic moment, and centroid depth, along with a simply parameterized estimate of the source time function duration. However, the fault-plane ambiguity remains in these solutions, and with the exception of very broadband CMT inversions (Ekström, 1989), neither source function complexity nor slip distribution are quickly or routinely determined. Ongoing modeling efforts are concentrated on refining descriptions of earthquake sources by estimation of the spatial and temporal distribution of moment release and resolution of possible mechanism changes during rupture. These problems are not new; much work has already been done in these areas. Teleseismic body waves have been employed extensively to estimate the details of great subduction zone earthquakes (e.g., Ruff and Kanamori, 1983; Beck and Ruff, 1984; Schwartz and Ruff, 1985; Kikuchi and Fukao, 1986). Teleseismic body waves, strong-motion recordings, and regional body waves have also been analyzed to characterize the sources of
large- and moderate-size events (e.g., Hartzell and Heaton, 1983; Sipkin, 1982; Beroza, 1991; Kikuchi and Kanamori, 1991; Steidl et al., 1991; Wald et al., 1991; Dreger and Helmberger, 1992; and many others). In the following, we present a new method to estimate gross finiteness features of earthquake ruptures using teleseismic and regional surface waves. The technique is primarily useful for large earthquakes ($M_w > 6.5$) and attains much higher resolution than previously possible using surface waves.

Given the high quality and near-real time availability of modern broadband seismic data, it is important to establish techniques that not only further our basic understanding of earthquakes, but also provide useful information quickly after a large event. Reliable estimates of source parameters for major earthquakes are now commonly available within a few hours of the event (e.g., Ekström et al., 1986; Nakanishi et al., 1991; Thio and Kanamori, 1992; Pasayanos and Romanowicz, 1992; Lay et al., 1993; Ritsma and Lay, 1993). These methods incorporate a variety of procedures and data, but generally only determine point-source parameters. Yet, finite-fault information may be very important soon after an event. In particular, information on rupture propagation may indicate where to expect possible increased damage as a result of rupture directivity, thereby helping to coordinate disaster response teams. Rapid identification of the causal fault is also important for postevent hazard assessment, such as calculation of stress transfer to adjacent faults which may fail subsequently (e.g., Harris and Simpson, 1992; Stein et al., 1992; Jaumé and Sykes, 1992). For submarine events, tsunami excitation is influenced by the actual fault plane as well as by details of the source function. The simplicity of the method described below allows for an estimate of first-order rupture finiteness in the first hour or two after a large earthquake, limited by the time it takes to determine the gross faulting parameters and to identify a suitable empirical Green function.

In this article, we describe the deconvolution technique and assess its limitations using synthetic tests. We extend the analysis of the 1992 Landers, California ($M_w = 7.3$), earthquake in Ammon et al. (1993) by including additional surface-wave arrivals as well as body waves. This large strike-slip event has been extensively studied, which allows us to test the consistency of our results with field observations (surface rupture) and other seismic modeling procedures. We find that for the Landers event, our method proves stable and robust, and the rupture directivity is unambiguous. A fairly detailed model of the source is obtained, including a change in mechanism during rupture and nonuniform slip along the fault. With the procedure now being automated, results can be obtained within 1 to 2 hr after comparable size future events.

Source Time Function Extraction and Interpretation

Most efforts to quantify the spatial and temporal history of a rupture using seismic waves involve estimation of the source time function or moment-rate function at various geometries around the fault (e.g., Ruff, 1987). Source time function information contained in seismic signals is obscured by propagation and recording effects. To isolate the far-field source time function from propagation and instrument effects, we deconvolve surface waves (or body waves) of a moderate-size event from the surface waves (or body waves) of a large event. The small event must be near the mainshock and have a similar centroid depth and focal mechanism. At least a magnitude unit difference in size is desirable so that the source duration of the small event is much less than that of the larger event. An event satisfying these criteria is called an empirical Green function (EGF), as it constitutes an approximate point-source impulse response for a given moment tensor. Previous researchers have used empirical Green functions to estimate source parameters from body waves from small events (e.g., Mueller, 1985; Mori and Frankel, 1990; Hough et al., 1991), strong-motion recordings of large events (e.g., Hartzell, 1978), and surface waves from moderate-size events (Weidner and Aki, 1973; Patton, 1980; Burger and Langston, 1985). The result of the deconvolution is a time series that depends on both the mainshock and the EGF. We refer to this time series as a relative source time function (RSTF). The long-period spectrum of the RSTF is related to the ratio of moments for the two events. The azimuthal duration and amplitude variations of the RSTF's reveal the directivity and complexity of the mainshock rupture. For EGF's with very short durations, the RSTF duration corresponds closely to the actual rupture duration of the large event.

Two approaches are available for analyzing the resulting RSTF's. The simplest is a relative location analysis of discrete features of the time function, such as the onset or end of moment release, or relative extrema in the time functions (Fig. 1). In the absence of complications such as a change in focal mechanism during rupture or differences between the EGF and mainshock focal mechanisms, the RSTF is a moment-normalized slant-stack of the moment-rate density distribution over the fault plane (Ruff, 1984; Schwartz and Ruff, 1985; Ruff, 1987). Thus, a more formal analysis tool is the finite inverse Radon Transform. This is a more objective approach since no time picks are necessary and the entire RSTF is incorporated into imaging the spatial and temporal moment release. The choice of analysis technique depends on the suitability of the EGF. The relative location technique can be applied to almost any set of time functions, as long as care is used in identifying deconvolution effects that may be the result of changes in the
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...mechanism during rupture of the mainshock or differences between the mainshock and EGF mechanisms. Radon Transform methods are more sensitive to amplitude fluctuations resulting from mechanism changes or inadequacies of the EGF, making them difficult to routinely employ. However, they can provide realistic resolution estimates if such complexities are not present.

Selecting the Empirical Green Function

The applicability of this technique depends on the existence and suitability of the EGF. The suitability of an EGF-mainshock pair depends on the location, depth, size, and focal mechanism of both events. Ideally, the mainshock and the EGF should be identical in focal mechanism and co-located with similar depths. We assume that these attributes of a given event have been determined using point-source inversion procedures. The EGF must be significantly smaller than the mainshock, yet large enough to be recorded with high signal-to-noise by a reasonable number of stations. The difference in size implicitly means that the depth and co-location requirements can only be approximately satisfied. To attain a very precise slip distribution, one must have the EGF rupture overlap with the mainshock fault plane, but since we are primarily interested in robust first-order spatial-temporal features, we can use EGF's at greater distances. Many deconvolutions using theoretical, single-depth Green functions have extracted useful source function estimates for large events (e.g., Ruff and Kanamori, 1983), and we limit our analysis to a single EGF. We primarily use surface waves since they have the highest signal-to-noise ratios, but our method readily incorporates body waves as well. We have found that the two-dimensional radiation patterns of surface waves generally provide more stable deconvolutions, and the range of directivity parameter sampled is larger than for body waves. Thus, we present surface-wave applications with an emphasis toward rapid analysis capabilities.

Ammon et al. (1993) showed that surface-wave deconvolutions can be successfully applied to large earthquakes: the 1992 Landers ($M_w = 7.3$) and 1992 Cape Mendocino ($M_w = 7.1$), California, earthquakes, each of which had a preshock with $M_w = 6$ that could be used as an EGF. Velasco et al. (1993) further utilize this technique to investigate recent large earthquakes near the Mendocino Triple Junction and in the Gorda plate. Not all large earthquakes that have occurred in the last 3 yr have had an immediate preshock that could be used as an EGF, a notable case being the Loma Prieta earthquake of 1989. However, we can use any event that has occurred recently enough to be recorded at the same stations as a given mainshock, as long as it satisfies the criteria for an EGF. We searched the Harvard CMT catalog for suitable EGF's for 14 earthquakes that occurred within the last 3 yr with $M_w > 7.1$. We found that each event has a candidate EGF located within 70 km of the mainshock with both a similar mechanism and depth. Since subsets of digital data are available on-line from IRIS and POSEIDON data centers for most recent events larger than $M_w = 6$, we can usually access data for a suitable EGF very rapidly.

Data Windowing and Deconvolution

Once an EGF is selected for a given mainshock, the surface-wave seismograms (usually, R1 and G1) are windowed using group-velocity values of 5.3 to 2.9 km/sec for Love waves, and 4.5 to 2.3 km/sec for Rayleigh waves (Fig. 2). The windows are adjusted depending on the propagation characteristics between source and receiver to ensure the signal is completely captured. Body waves (typically, $P$ and $SH$) are windowed manually. Each window is referenced to the origin time and the epicenter of each event, so that the phase of the deconvolution is sensitive to these parameters. The mean of each seismogram is removed and a Hanning window is applied to taper the seismogram near the ends. The traces are padded with zeros to twice their original length and the deconvolution is performed in the frequency domain using the water-level method (Helmbberger and Wiggins, 1971; Clayton and Wiggins, 1976). Although not always the most effective deconvolution procedure (Sipkin and Lerner-Lam, 1992), water-level deconvolution is a reasonable choice when little information on noise is available. For first-order information, the water-level technique is adequate, although care must be used to select only high-quality data for the analysis. For higher-resolution investigations, time-domain approaches based on linear-inverse theory may be more appropriate since they provide detailed information on the uncertainties of the resulting time functions and allow easy incorporation of
a priori constraints (Hartzell and Heaton, 1985; Sipkin and Lerner-Lam, 1992; Ammon, 1992).

The next step of the procedure is bandpass filtering the time functions to eliminate high-frequency noise as well as any frequencies inappropriate for the available EGF-mainshock pair. The high-frequency corner of the filter depends primarily on the EGF corner frequency and dispersion effects caused by differences in location between the EGF and mainshock. For the earthquakes we investigate, the bandwidth of the data is large (periods from less than ten to several hundred seconds) and unstable deconvolutions occur primarily at azimuths near radiation pattern nodes of the mainshock or EGF. The reduced number of nodes in surface-wave radiation patterns accounts in part for their enhanced stability in the deconvolutions. The stability of the deconvolution is judged by inspection, and often, those deconvolutions first judged unstable are found to actually contain information on differences in the focal mechanisms of the EGF and mainshock. Amplitude variations of discrete subevents are sensitive to both directivity and focal mechanism changes and in certain circumstances can be used to constrain the focal mechanism of both events (Ammon et al., 1993).

Discrete Source Function Complexity Analysis

To identify the direction of rupture of an earthquake, we use unilateral directivity calculations, such as have been extensively done for deconvolutions of teleseismic P-wave source time functions (e.g., Hirasawa, 1965; Schwartz and Ruff, 1985). The analysis is based on the linear relationship

\[ \delta t = t_0 - \Gamma \Delta, \]

where \( \delta t \) is the measured time difference between the

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**Figure 2.** Example of source time function extraction for the deconvolution of surface waves recorded at ANMO, NNA and BKS. Each surface-wave arrival is group velocity windowed (vertical lines in seismogram window). The Joshua Tree (\( M_w = 6.2 \)) earthquake recordings are deconvolved from the Landers (\( M_w = 7.3 \)) earthquake recordings. Each deconvolution gives a relative source time function at each station for the Landers earthquake (right).
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chosen features (e.g., onset to peak; onset to end; onset to reference EGF start time; and peak to peak for two subevents), $t_i$ is the time between the features, and $\Delta$ is the distance between the two features (Fig. 1). The directivity parameter, $\Gamma$, is given by

$$\Gamma = \frac{\cos \theta}{c},$$  \hspace{1cm} (2)

where $\theta$ is the azimuth of the station relative to the azimuth of a line connecting the spatial offset between the features, and $c$ is the wave phase velocity, discussed in more detail below. To estimate $\Delta$ and $t_i$, the model (1) is investigated systematically for each azimuth. The preferred azimuth corresponds to the direction producing the best linear correlation coefficient for (1).

Radon Transform Analysis

Ruff (1984, 1987) has presented a thorough description of the relationship between observed source time functions and the moment-rate density distribution on the fault. For our application to a vertical strike-slip fault, the Ribbon Fault model is appropriate, and the relationship between a far-field time function, $F(t, \Gamma)$, and the moment-rate density on the fault, $m(t, x)$, is given by a Radon Transform

$$F(t, \Gamma) = \int_0^t m(t - \Gamma x, x)dx.$$ \hspace{1cm} (3)

Thus, by performing an inverse Radon Transform we can recover the moment-rate distribution for a given earthquake.

$$m(t, x) = \int_{-\infty}^{\infty} I(t) * F(t - \Gamma x, \Gamma)d\Gamma.$$ \hspace{1cm} (4)

The term $I(t)$ is the inverse Radon Kernel (Ruff, 1987) and has a Fourier Transform equal to $\omega_0$. The operator $*$ represents convolution. Expression (4) is general and can be in theory reconstruct complicated faulting histories using this relationship. In practice, obtaining a numerical solution of (4) requires very stable estimates of the far-field time functions and a good sampling in both azimuth and phase velocity. Additionally, any change of focal mechanism and any intrarupture dispersion, which affects the source time function, will affect the utility of (4).

Surface-Wave Dispersion

A distinguishing characteristic of our approach to isolating the source time function is our use of both the amplitude and phase of surface waves. Early directivity studies using surface waves (Press et al., 1961; Ben-Menahem, 1961) employed only the amplitude spectrum of surface waves. Some later methods have included model calculations including finite-fault parameterizations (e.g., Kanamori and Given, 1981; Zhang and Kanamori, 1988), but these are subject to inaccuracies in the models used for propagation corrections, which have generally prevented analysis of periods less than 100 sec. In our procedure, we use time-domain signals obtained by deconvolution. Surface-wave dispersion will have an increasing, deleterious effect on estimating the time function as the distance between the EGF and the earthquake increases to a significant fraction of the wavelength. If the EGF and mainshock were co-located point sources with different time functions, the deconvolution would directly extract the relative time function, no matter how dispersed the original waveforms, because the dispersion is common to both signals. More commonly, dispersion complicates the interpretation of the time functions in two ways. First, dispersion affects the waveform shape; second, we must choose a single phase velocity to calculate relative locations using the discrete event directivity analysis. We first address the choice of a single phase velocity for directivity calculations, and then illustrate the effects of dispersion on the wave shapes using calculations with synthetic seismograms.

We must choose one velocity [$c$ in (2)] for the simplified directivity analysis to calculate the relative location of discrete features of the RSTF's. The straight-line fit determines the distance between subevents (the slope), the time between subevents (the intercept), and the rupture direction (the optimal linear correlation coefficient). Our choice is to use the average phase velocity associated with the Airy phase of the Rayleigh or Love waves. The phase velocity is slowly varying near the Airy phase, and thus this phase velocity corresponds to the largest packet of energy arriving at a given station. The difference in phase of the dominant energy in the Airy phase between the two source locations yields the primary source function characteristics.

Using a single phase velocity is clearly an approximation. Using a different constant phase velocity in the directivity calculations changes only the slope of the line and hence only influences our estimate of the relative location of features in the RSTF's. The change has no effect on the estimated direction of rupture since that depends on the correlation coefficient. Ammon et al. (1993) found that changing the phase-velocity values used for Rayleigh waves from 3.85 km/sec to a significantly slower value of 3.45 km/sec changed the slope estimates less than 10%, well within the uncertainty in locations presented in that study. While no single phase velocity can hold for the full bandwidth of the deconvolved pulse, the fact that deconvolution eliminates the primary dispersion and restores the pulse-like signal greatly reduces sensitivity to the choice of a reference velocity. Li and Toksöz (1993) avoid this issue by relying on the ampli-
tudes of the surface-wave deconvolutions rather than timing measurements, but the amplitudes can be rather unstable and we prefer to emphasize timing of features in the RSTF’s.

To investigate the effects on the time function shape, we process synthetic seismograms using the deconvolution and directivity procedures described above. The synthetic seismograms were calculated for an isotropic, plane-layered velocity model approximating PREM (Dziewonski and Anderson, 1981) containing a 20-km-thick crust. The calculations were performed using the method and programs of Herrmann (1987) and are appropriate for plane-layered earth models. Since the important dispersion effects are the intrarupture dispersion and the dispersion over the path-length differences between the EGF and the mainshock to each station, the flat-earth approximation is appropriate for small to moderate-size events (less than a few hundred kilometers between events and subevents). Finite-faulting effects and rupture directivity for the mainshock were simulated by summing equally spaced point sources along the length of the mainshock. We present the results for two synthetic tests; the first is for a small earthquake, 25 km in length, and illustrates the lower limit of resolution of the technique. The second is designed to approximate the geometry of a larger earthquake with a 100-km rupture. The basic result of these tests is that the appropriateness of the EGF depends on the distance from the mainshock. The larger the separation between the two events, the stronger the dispersion effects, and hence large EGF-mainshock separations or large mainshocks require analysis at longer periods.

Figure 3 shows the geometry and RSTF’s for the 25-km-long unilateral rupture and two different EGF locations. The first EGF location is at the onset of the fault and the second is located 25 km from the end of the fault. The mainshock synthetic was simulated by summing equally spaced point sources along the length of the mainshock. We present the results for two synthetic tests; the first is for a small earthquake, 25 km in length, and illustrates the lower limit of resolution of the technique. The second is designed to approximate the geometry of a larger earthquake with a 100-km rupture. The basic result of these tests is that the appropriateness of the EGF depends on the distance from the mainshock. The larger the separation between the two events, the stronger the dispersion effects, and hence large EGF-mainshock separations or large mainshocks require analysis at longer periods.

For the second test, with a 100-km-long rupture, the synthetic seismograms for the mainshock were again}

![Figure 3. Synthetic tests of the effects of intrarupture dispersion on deconvolutions for a 25-km-long fault. The time functions correspond to recordings at opposite azimuths along the rupture direction. The time functions have been low-pass filtered using a two-pole, zero-phase Butterworth filter with a short-period corner at 10 sec. The upper plot demonstrates the results for an EGF that is located at the onset of rupture; the bottom is for an EGF 25 km from the end of the fault.](image-url)
created by summing point sources spaced 3 km apart and using a rupture velocity of 3 km/sec (Fig. 4). The EGF was located at one end of the rupture. The results are presented in Figure 4 for two different bandwidths; the first has a 10-sec-period low-pass cutoff and the second has a 20-sec-period low-pass cutoff. As expected, the dispersive effects increase with the length of the fault. However, even the 10-sec bandwidth has only minor dispersive effects. The directivity can be clearly estimated by the variation in pulse widths. Below 20 sec, the effects of intrarupture dispersion are all but invisible. The source functions in the direction of rupture resemble the low-pass filter response, and the side lobes on the source functions in directions away from rupture are from the filter. We can locate the onset of the mainshock moment release and the centroid of the mainshock rupture by azimuthal variations in the timing of the onsets of the primary RSTF pulses and centroid of the RSTF’s. For lower frequencies, the estimates of these locations are degraded in proportion with the rupture length of the event, but still only modest errors are incurred.

For small events, the EGF should be located within one or two fault lengths of the mainshock, since higher frequencies are necessary to identify directivity. For larger events, the dispersive effects at high frequencies are significant, but corresponding larger directivity effects permit a robust analysis using lower frequencies. The dispersive effect could be minimized by applying an effective propagation correction to “move” the EGF to the estimated centroid of the mainshock, but this requires a priori knowledge of the fault orientation, which is usually not available. Lacking that, one might equalize the effective epicentral locations. Such a correction requires an estimate of the local structure, but a reasonable model will improve the deconvolved time functions. For the Landers earthquake, this correction is unnecessary since the EGF centroid is within 20 to 30 km of the epicenter of the Landers mainshock. We verified this directly by computing appropriate dispersion corrections.

Source Depth Considerations

In addition to spatial offsets in the relative epicenters of the EGF and mainshock, the depth of the EGF relative to the depth of moment release of the mainshock can affect the resolution of the RSTF’s. This well-known effect is stronger for body waves than for surface waves, and only important for short-period surface waves (for crustal earthquakes). The suitability of the EGF depends on the difference in depth relative to the wavelength of interest. Velasco et al. (1993) explore this effect with synthetic tests and find that Love wave RSTF’s are stable for depth differences as great as 20 km, while body waves and Rayleigh waves are stable for depth differences no greater than 10 to 15 km. For the Landers strike-slip event presented in this article, the depth effect is small since the depth extent of major moment release for the EGF and the mainshock most likely overlap, given their respective aftershock distributions (Hauksson, 1992, personal comm.).

Differences and Changes in Mechanisms

Deconvolutions are sensitive to even slight differences of focal mechanism between the EGF and mainshock. Small differences in focal mechanism can largely be accounted for if discrete subevents can be identified, as demonstrated for the Landers earthquake by Ammon et al. (1993). Assuming EGF and mainshock vertical strike-slip focal mechanisms, they inverted the subevent pulse area fluctuations of the RSTF’s for the jth subevent/EGF seismic moment ratio, $A_j$, and slight changes in subevent strike using:

$$F_j = \frac{A_j \sin(2\theta_{ij})}{\sin(2\theta_{ij})},$$

where $F_j$ is the area of the subevent at the $i$th station for the $j$th subevent, $\theta_{ij}$ is the station azimuth relative to the subevent fault strike (with an adjustment for Love waves, $\theta_{ij} = \theta_{ij} + \pi/4$), and the subscript $o$ refers to the EGF. Figure 5 shows results of the subevent modeling for Landers and the observed RSTF’s at SNZO, which demon-

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**Figure 4.** Synthetic tests of the effects of intrarupture dispersion on deconvolutions for a 100-km-long fault, with bandwidths up to (top) 10 sec and (bottom) 20 sec. The EGF is located at the onset of rupture. The dispersive effects are stronger for short periods.
strates the amplitude difference of subevents caused by a change in strike.

Large differences in the focal mechanisms and depths of two events can be accounted for when no EGF's with similar mechanisms are available (Weidner and Aki, 1973), although the procedure loses its simplicity. When the mechanisms are known, azimuths expected to show strong sensitivity to the mechanism differences are easily identified. This sensitivity is the basis of the source modeling procedure employed by Weidner and Aki. They specifically looked at the spectral ratios of two events with different mechanisms to revise the mechanisms and depths of both events. An unfortunate consequence of changing or differing mechanisms is the invalidation of the assumptions necessary to perform an inverse Radon transform (Ruff, 1984) using the RSTF's to obtain a spatial and temporal moment-release image along the fault. For the Landers event, this particular problem rules out use of the Radon Transform. However, analysis of discrete source time function characteristics can still produce valuable information, as illustrated below.

**Application: the 28 June 1992 Landers, California, Earthquake**

The Landers earthquake (Fig. 6) occurred in the southern Mojave Block approximately 200 km east of Los Angeles (Mori et al., 1992). The event ruptured a series of faults totaling approximately 60 to 70 km in length (Sieh et al., 1993), and caused extensive damage in the towns of Yucca Valley and Landers. This event was the largest earthquake in southern California in 40 yr, and its close proximity to the San Andreas fault has raised much concern about changes in stress on that fault (e.g., Stein et al., 1992; Harris and Simpson, 1992; Jaumé and Sykes, 1992). Much is now known about the rupture process of this event from our initial surface-wave deconvolutions (Ammon et al., 1993) and from more detailed modeling of strong ground motions (Kanamori et al., 1992; Campillo and Archuleta, 1993; Cohee and Berroza, 1994; Dreger, 1994). Two strong pulses of energy dominate the far-field source time functions (Kanamori et al., 1992; Ammon et al., 1993) and the rupture propagated in a north-northwesterly direction along at least three different fault segments (Sieh et al., 1993; Massonnet et al., 1993; Murray et al., 1993). The second pulse is stronger than the first, and the strike of the rupture becomes more westerly toward the northern end of the rupture (Kanamori et al., 1992; Ammon et al., 1993). Tomographic imaging of the crustal velocities around the source region indicates that the mainshock and after-
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Table 1

<table>
<thead>
<tr>
<th>Event</th>
<th>Date (m/d/yr)</th>
<th>Origin Time</th>
<th>( M_s )</th>
<th>( M_w^* )</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Strike</th>
<th>Dip</th>
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<td>7.3</td>
<td>10.6</td>
<td>34.2°N</td>
<td>116.5°W</td>
<td>341°</td>
<td>70°</td>
<td>-172°</td>
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<tr>
<td>EGF (JT)</td>
<td>4/23/92</td>
<td>04:50:22</td>
<td>6.2</td>
<td>0.21</td>
<td>33.9°N</td>
<td>116.3°W</td>
<td>171°</td>
<td>89°</td>
<td>-177°</td>
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*In \( 10^{26} \) dynes-cm.

The Landers and Joshua Tree (EGF) Source Parameters from the NEIC and Dziewanski et al. (1993)

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shocks are mainly located in relatively high-velocity rock with surrounding low-velocity, aseismic zones (Zhao and Kanamori, 1993).

The Landers earthquake was preceded by the 23 April 1992 near-vertical strike-slip Joshua Tree earthquake \( (M_s = 6.2) \) located approximately 20 to 30 km to the south (Fig. 6; Table 1). Directivity estimates for this event using the regional TERRAscope stations indicate a northward rupture propagation extending to within about 20 km of the Landers epicenter (Kanamori, personal comm., 1992), but the duration was only a few seconds. Using the Joshua Tree waveforms as EGF's, Ammon et al. (1993) obtained 13 RSTF's using teleseismic and regional long-period surface waves recorded at stations from which data were retrieved by the IRIS Gopher automatic dial-up system. We expand the azimuthal coverage by adding surface waves from GSN stations NNA, AAK, KIV, and Japanese POSEIDON stations HKY, TKO, and YCU, as well as analyzing body waves at all stations. This doubles the number of observations used in the directivity analysis relative to Ammon et al. (1993).

The deconvolved time histories demonstrate a strong variation with azimuth and are plotted as a function of the directivity parameter \( \Gamma = \cos(\theta)/c \) in Figure 7. Two pulses are seen at stations with azimuths nearly perpendicular \( (\Gamma \approx 0) \) to the reference rupture direction of 335°. Stations with positive \( \Gamma \approx 0.3 \) demonstrate merging of the two subevents into one pulse as a result of the northward rupture propagation toward these stations. Stations with negative \( \Gamma \) (e.g., NNA) show an increasing separation between the subevents caused by the rupture propagating away from the southerly stations. These source time functions are very compatible with the two-subfault models of Kanamori et al. (1992), Campillo and Archuleta (1993), and Dreger (1994).

We performed directivity analyses using time picks for the onset, the two peaks, and the end of the RSTF's (Table 2). Sidelobe effects make timing of the onset and end difficult, as indicated by the lower correlation coefficients associated with these measures. The most robust measurement is between the two subevent peaks, which occurred 9.5 ± 1.1 sec apart, separated by 34 ± 7 km along an azimuth of 330 ± 20°. This indicates a rupture velocity of 3.6 ± 1.2 km/sec. Rupture velocity estimates from the onset and end measurements are systematically lower, from 1 to 2 km/sec, but are likely biased by the sidelobe contamination. An overall rupture velocity of about 2.5 km/sec appears most reasonable. The directivity results for the second subevent relative to the onset of motion of the Landers earthquake indicates that the best correlation occurs for an azimuth of 335° (Fig. 8, inset). The least-squares fit straight line corresponding to this azimuth is shown in Figure 8, and individual observations are presented as squares. There is a strong trend, and the scatter is largely attributable to the change in subevent mechanism described below.

The Landers event ruptured from just north of the Joshua Tree event, toward the north-northwest. The total duration of rupture lasted approximately 25 sec (Table 2). The formal directivity measurements suggest that the centroid of the first pulse is located a few kilometers to the east of the actual epicenter (Fig. 6). However, this is not reliably resolved, since there is complexity in the higher-frequency component of the first pulse. The more
Table 2

<table>
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<th>Event Pair</th>
<th>Time Correlation Distance Azimuth</th>
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<td>(From--To) Separation (sec) Coefficient (kan) (~)</td>
<td></td>
</tr>
<tr>
<td>Onset to EGF</td>
<td>2.3 ± 0.6 0.70 12.5 ± 3.9 170 ± 20</td>
</tr>
<tr>
<td>Onset to S1</td>
<td>7.7 ± 0.4 0.66 7.0 ± 2.4 80 ± 35</td>
</tr>
<tr>
<td>Onset to S2</td>
<td>16.9 ± 1.1 0.77 30.7 ± 7.6 335 ± 15</td>
</tr>
<tr>
<td>Onset to End</td>
<td>26.8 ± 1.5 0.60 25.3 ± 10.2 350 ± 20</td>
</tr>
<tr>
<td>S1 to S2</td>
<td>9.5 ± 1.1 0.83 34.2 ± 7.0 330 ± 20</td>
</tr>
</tbody>
</table>

Figure 8. Directivity analysis of the second subevent relative to the onset of motion of the Landers earthquake. The inset shows the variation of the corresponding linear correlation coefficient as a function of azimuth. The maximum value occurs at a strike of 335°, which is the case shown. Squares correspond to individual observations, the best fitting line is indicated by the solid line.

A change in focal mechanism strike from 353° for the Joshua Tree earthquake, to 344° for subevent 1, to 332° for subevent 2 should also be evident in body-wave deconvolutions. Thus, we test the surface-wave results by examining body-wave deconvolutions for consistency. Synthetic P and SH waves were computed for the Joshua Tree location with a vertical strike-slip mechanism and a strike of 353°. We model the Landers mainshock as two pulses, each with an 8-sec duration with vertical strike-slip mechanisms changing from a strike of 344° for subevent 1 to 332° for subevent 2. We lag the pulses to account for the directivity according to the peak-to-peak directivity results obtained from the surface-wave deconvolutions.

Deconvolving the synthetic Joshua Tree body waves from the Landers mainshock synthetics results in a close match to the observed source time functions for both P (Fig. 9) and SH waves (Fig. 10). Note that the P wave at station CCM (at an azimuth 72°) has a small positive pulse for subevent 1 and a strong negative pulse for subevent 2. The rotation of the focal mechanism from 353° to 332° has caused this station to cross over the P-wave node for the Joshua Tree earthquake, reversing the polarity of the second subevent. Other stations show similar polarity changes, confirming the change in mechanism found from the surface-wave modeling. This consistency confirms the model estimated from the surface waves and demonstrates that careful analysis of the RSTF’s can lead to detailed information about the mainshock faulting history. The prediction of the body waves is quite good, and we have not attempted any refinement of the model of Ammon et al. (1993).
We have pointed out that the influence of intrarupture dispersion and changing focal mechanism violates the assumptions of the Radon Transform relationship, and so it is not generally applicable to the Landers RSTF's. Still, an approximate description of the spatial distribution of slip would be useful. Since the EGF is located along an extension of the mainshock strike, the effects of dispersion for propagation perpendicular to the fault strike are small since the change in source-receiver distance for the entire mainshock and the EGF is very small. Additionally, directivity effects are minimal for wave propagation perpendicular to the rupture propagation. Using simple assumptions about the rupture, we can map the slip along the fault length using time functions observed nearly perpendicular to the fault. We assume that the rupture velocity, \( v_r \), was constant, the dislocation time function is relatively uniform across the fault, and that the fault dislocation function duration was less than the shortest period in the RSTF's (10 sec). With these assumptions, we can show that for stations nearly perpendicular to the fault, the slip distribution is simply a scaled and coordinate-stretched version of the time function (see the Appendix)

\[
S(v,t) = \frac{f_p(t)}{v_r \mu W},
\]

where \( S(v,t) \) is the slip, \( f_p(t) \) is the time function perpendicular to the fault, \( \mu \) is the shear modulus of the material, and \( W \) is the vertical extent of the faulting. To estimate \( f_p(t) \), we scale the RSTF's by the moment of the EGF. We assume \( \mu = 3 \times 10^{10} \text{ dyne/cm}^2 \), and \( W \) is estimated to be 15 km using aftershock distributions (Hauksson, 1992, personal comm.). Using this relationship, we estimate the slip distribution along the Landers fault and present the result in Figure 11. We used the average of the Love wave RSTF's observed at CCM, ANMO, and KIP (\( |\Gamma| \leq 0.06 \)) to generate the slip distri-

**Figure 9.** Landers P-wave synthetic and observed source time functions plotted as a function of azimuth. The strike change is constrained to that determined from the RSTF amplitude modeling of Ammon et al. (1993). Note the close match between synthetics and data, particularly the polarity reversals for CCM and COL P waves. Also plotted is the impulse response of the filter applied to the deconvolved time functions, which indicates the lower limit on time resolution.

**Figure 10.** Landers SH-wave synthetic and observed source time functions plotted as a function of azimuth. (See details in Fig. 9.)

**Figure 11.** Slip distribution (thick line) for the Landers earthquake obtained by averaging Love wave RSTF's observed at CCM, ANMO, KIP (\( |\Gamma| \leq 0.06 \)) assuming a rupture velocity of 3 km/sec. The maximum slip estimates are 5.5 \( \pm \) 1 m at approximately 10 to 12 km from the epicenter, and 8 \( \pm \) 1 m centered at approximately 40 km from the onset of rupture. The thin line indicates the observed surface offset from Sieh et al. (1993).
bution. For this azimuth, the Love waves are near a maximum in the radiation pattern and are thus least sensitive to changes in the fault strike. However, calculating the radiation pattern difference expected at these stations and averaging the possible bias in amplitudes, we expect the relative amplitude of the second subevent to be slightly underestimated by about 15%. We assumed a rupture velocity of 3 km/sec for the projection, although this value may vary. The DC offset of the slip distribution is a result of bandlimited data. The uncertainty in the estimate is about 1 m (measured from the noise level for distances greater than 90 km). The magnitude of the slip is inversely proportional to the rupture velocity, and so large rupture velocities spread out the rupture and decrease the estimated slip at any point. Our maximum slip estimates are $5.5 \pm 1$ m at approximately 10 to 12 km from the epicenter, and $8 \pm 1$ m centered at approximately 40 km from the onset of rupture. Our estimates of the slip are about 25% larger than those measured at the surface (Sieh et al., 1993), as shown in Figure 11, or those estimated using an analogous procedure for local signals (Kanamori et al., 1992). The slight difference in focal mechanisms may influence these amplitudes, or we may be underestimating the rupture velocity, or seeing slip beneath the seismogenic zone (underestimating W). The $SH$-wave deconvolutions in Figure 10 suggest that our model over predicts the first subevent amplitude by about 30%, which would reduce the disagreement. A small change in dip between the subevents, which was not allowed for, could account for this as well. Given the fact that other standard surface-wave inversion methods would fail to determine anything more than point-source parameters for this moderate event, the overall quality of our finite-source, changing mechanism solution is very good, especially considering that it was available within a few hours of the event.

Discussion

There are two basic applications of our deconvolution technique. The first is a rapid determination of fault directivity, and the second is to obtain long-period information on the character of large ruptures. The procedure is very successful in establishing the first-order finiteness characteristics of a large earthquake. Although the fault plane is not always uniquely determined from observable seismic-wave directivity, the only teleseismic method for specifying the fault plane is to establish finiteness effects in the source functions. This is an elementary assertion, as is the fact that no technique works in every case. Surface waves have an intrinsic advantage of increasing the range of the directivity parameter, $\Gamma$, to the maximum attainable for elastic waves, making the directivity more resolvable than from teleseismic body waves alone, given their limitation to large phase velocities. The broadband deconvolution of surface waves retains the bandwidth needed to exploit the increased range of directivity parameter. It is not correct to think that because our waves include long-periods we intrinsically have low spatial resolution. The temporal resolution of the broadband surface-wave deconvolutions is in fact enhanced by our bandwidth.

The success of this method depends on the availability of a suitable EGF. We have found that in most cases, suitable small events are available, whether they are foreshocks, aftershocks, or simply historical events. In selecting an EGF, we limit ourselves to events with a similar mechanism to the mainshock, to keep the method simple. The location and size of the EGF must be appropriate, and signal quality is critical since the EGF spectrum is the denominator of the deconvolution. The size of the EGF is very important, since the signal-to-noise ratio depends on the amplitude and distance traveled by these arrivals.

For near-real time applications, we need reliable estimates of the focal mechanisms of the mainshock to be determined quickly. For this purpose, we generally rely on the early Harvard CMT solutions (Ekström et al., 1986; Ekström, 1993) which are broadcast on computer mail. The reliability of the early CMT solutions has been addressed by other researchers, and the conclusion is that they are generally reliable. Several other rapid mechanism determination capabilities utilizing regional data are yielding quite stable source mechanisms within an hour or two of the event (e.g., Lay et al., 1993; Ritsema and Lay, 1993). While these various methods are still undergoing further development, the basic technology exists for quickly determining the point-source mechanism needed for selecting an EGF.

The on-line, digital databases of earthquake recordings are growing rapidly, and are critical for rapid determinations of fault directivity. Searching the Harvard CMT catalog to identify a suitable EGF is straightforward. Recent data can be acquired in minutes, since they are available on-line at the IRIS Data Management Center, Caltech TERRAscope, UC Berkeley Digital Seismic Network, and Earthquake Research Institute (University of Tokyo) POSEIDON data management centers. The Landers event provides a good example. Ammon et al. (1993) selected the EGF and were able to obtain data from the on-line sources, and they extracted reliable source function characteristics within a few hours of the event. In the future, identifying an EGF and transferring those data through computer connections could take less time than waiting for the surface waves from the larger shock to travel to all of the seismic stations with real-time access. This requires automated regional wave determinations of the focal mechanism, which are currently under development in many seismogenic regions.

Identification of the actual fault plane and general slip distribution is possible with our method. For rapid analysis, this constitutes the next logical step in the evo-
Empirical Green Function Deconvolution of Broadband Surface Waves

lution of source analysis procedures (Lay et al., 1993). This will be especially useful for offshore earthquakes or large earthquakes lacking ground rupture. Velasco et al. (1993) applied the deconvolution method to identify the fault plane for the 25 April Cape Mendocino, California ($M_w = 7.1$), earthquake. This thrust event occurred near the Mendocino Triple Junction (Pacific, Gorda, and North America plates), and did not rupture the surface. Furthermore, there were two large strike-slip aftershocks, making the interpretation of the fault plane very difficult. Velasco et al. (1993) identify the fault plane of the Cape Mendocino mainshock as the shallow thrust plane based on directivity, while finding that the two strike-slip aftershocks most likely ruptured conjugate fault planes. They also identified the rupture planes for two recent large ($M_w = 6.8, 7.1$) strike-slip events that occurred within the Gorda plate. This information is useful for understanding this complicated tectonic region. The only other way to obtain this information is through the modeling of strong motions and the location of aftershocks, but the latter are still quite ambiguous. It is unlikely that teleseismic body-wave inversions will reliably reveal the fault plane unambiguously, unless an event is very large and has strong directivity.

To extend the utility and resolution of the deconvolution method, a general procedure could be developed to identify and model focal mechanism changes between an EGF and a mainshock, or changes in focal mechanism during rupture. We have only done this for rotations of strike for vertical strike-slip faulting. As seen for the Landers earthquake, focal mechanism variations rule out simple imaging of the moment release without assumptions about the orientation of rupture. We circumvent this limitation by using RSTF's nearly perpendicular to the fault, and accounting for amplitude differences owing to mechanism changes. This will not be the case for all other earthquakes. Combined analysis of body- and surface-wave deconvolutions is viable, but complex faulting is just as much of a challenge for this approach as it is for procedures using model Green functions. Further applications are needed to fully explore the range of contributions that deconvolution methods can make to source analysis efforts.

Conclusion

We have adapted the empirical Green function strategy for removing propagation effects to broadband body- and surface-wave signals from large events. The method provides a powerful tool for isolating source time function duration and complexity. For events with moderate finiteness (20 to 50 km), the actual fault plane can be uniquely resolved for unilateral faults. Technological developments allow for application of the method in near-real time as well, which can contribute to postearthquake response (e.g., Mori et al., 1992; Lay et al., 1993). The synthetic tests establish guidelines regarding the frequencies that can be used stably in the method. These calculations show that it is important to use an empirical Green function that is close in location and centroid depth to the mainshock. The results obtained for the Landers earthquake demonstrate a strong unilateral rupture, with slight rotation of strike during rupture. The results are consistent with field evidence, as well as other investigations.

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References


Appendix

The Time Function for a Simple, Smooth Rupture

The Radon Transform, equation (3), expresses the relationship between the rupture of a ribbon fault of length \( L \) for a general moment-rate density function (recall that the moment-rate density of a ribbon fault varies with time and along the strike direction, \( x \)). If we make two simple assumptions about the form of the moment release along the fault, we can simplify this relationship. First, we assume that the entire rupture propagates with constant rupture velocity, \( V_r \). Next, we assume that the slip function, \( D(t) \), along the fault is independent of position. For the present, the shape of the function is completely general, but we require that it has unit area. The moment-rate density, \( m(x, t) \), is then

\[
m(x, t) = m(x)\delta(t - xv_r^{-1}). \tag{A1}
\]

The spatial function \( m(x) \) is equal to \( \mu WS(x) \), where \( \mu \) is the rigidity, \( W \) is the fault width, and \( S(x) \), the slip distribution along the fault, is zero outside the interval \( 0 \leq x \leq L \). Substituting (A1) into (3), we find

\[
f(t, \Gamma) = \int_{-\infty}^{\infty} m(x)\delta(t - xv_r^{-1} - \Gamma x)dx. \tag{A2}
\]

Since the shape of \( D(t) \) is independent of \( x \), we can remove it from the integration over \( x \)

\[
f(t, \Gamma) = D(t) * \int_{-\infty}^{\infty} m(x)\delta(t - xv_r^{-1} - \Gamma x)dx. \tag{A3}
\]

The operator * represents convolution. Next we use the relationship (valid for \( a \neq 0 \))

\[
\delta(ax + b) = \frac{1}{|a|} \delta(x + \frac{b}{a}) \tag{A4}
\]

such that

\[
a = -(v_r^{-1} + \Gamma)
b = t.
\]

The case \( \Gamma = -v_r^{-1} \), for which \( a \) vanishes, can be handled quite easily as a special case. Our primary interest is in the region where \( \Gamma \neq -v_r^{-1} \). We find

\[
f(t, \Gamma) = D(t) * \int_{-\infty}^{\infty} m(x)\delta(t - xv_r^{-1} - \Gamma x)dx. \tag{A5}
\]

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\[
f(t, \Gamma) = D(t) * \int_{-\infty}^{\infty} m(x)\delta(t - xv_r^{-1} - \Gamma x)dx. \tag{A5}
\]

We see that the observed time function is the convolution of the derivative of the dislocation time history with the moment-rate density function. If the dislocation function is a step, and the moment-rate function is constant over the fault length, we obtain a trapezoidal time function from the convolution of two boxcar functions. For a station perpendicular to the fault, \( \Gamma = 0 \), and we have

\[
f_{\perp}(t) = D(t) * v_r m(tv_r). \tag{A7}
\]

If we assume that the dislocation is relatively localized in time near the rupture front, such that for the periods of interest the dislocation is approximately a step function (for our purpose the duration of slip at any point on the fault is less than 10 to 20 sec), we can approximate \( D(t) \approx \delta(t) \) such that

\[
f_{\perp}(t) = v_r m(tv_r). \tag{A8}
\]

Finally, the relationship between the time function perpendicular to rupture and the slip on the fault is
With these simplifications, we see that the slip on the fault is a scaled version of the observed time function "stretched" using the coordinate transformation $x = v, t$. 

\[ S(v,t) = \frac{f_s(t)}{v, \mu W}. \]