SEISMIC PHENOMENA CONNECTED WITH EARTHQUAKES AND EXPLOSIONS

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The spatial and frequency distribution of path-corrected Rayleigh waves from the San Fernando earthquake are systematically related to the faulting process as a multiple event. The surface wave source is taken to be a depth-distributed set of double couples. A least-squares inversion is used to find the set of source parameters which optimally fit the variance-weighted data. In this study the strength, the slip angle, and the rupture delay times are optimally determined for the spatially
20. Abstract Cont’d.

assumed distribution of double couples. A sophisticated error analysis is performed to estimate the uncertainties of the calculated model variables.

One of the interesting side results from this inversion technique is the \textit{a priori} determination of what azimuthal distribution of observations is most beneficial in determining the source parameters, the uncertainties in the calculated model parameters, and the relative coupling between various parameters. Of the 18 WWSSN stations used for this event, two stations contain a total of 30\% of the total information in the data set. In fact, the data set could have been reduced by one-half and only 25\% of the total information which constrains the model variables would have been lost.

Transversely polarized shear waves (SH) from well located earthquakes are used to determine an upper mantle shear velocity structure compatible with both travel times and waveforms of observed seismograms. Vertically polarized shear waves (SV) are shown to be inadequate in identifying triplications because of contamination by PL-coupled shear waves and other P-SV interactions. The observed SH travel times are considerably slower than the JB values out to 30 degrees.
## CONTENTS

<table>
<thead>
<tr>
<th>I.</th>
<th>Summary</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>II.</td>
<td>Abstracts of publications and reports during this contract period</td>
<td>II-5</td>
</tr>
<tr>
<td>III.</td>
<td>Least-squares inversion of surface wave spectra to obtain source parameters: The 1971 San Fernando earthquake</td>
<td>III-12</td>
</tr>
<tr>
<td>IV.</td>
<td>Upper-mantle shear structure</td>
<td>IV-113</td>
</tr>
<tr>
<td>V.</td>
<td>Previous technical reports</td>
<td>V-162</td>
</tr>
</tbody>
</table>

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D. W. TAYLOR
Technical Information Officer
I. Summary

This year’s research in Seismic Phenomena Connected with Earthquakes and Explosions has concentrated on four task areas.

1) Evaluation of seismic discriminants
2) Understanding and determination of seismic source characteristics
3) Determination of the earth’s seismic properties (i.e. velocity and density structure and $Q^{-1}$)
4) Inversion of observations using crust and upper mantle properties to obtain the spatial and temporal characteristics of earthquakes and explosions

In the semiannual technical report, we reported on two different approaches for investigating the characteristics of the seismic source, in particular the determination of the corner frequency for P and S waves. The first uses theoretical volume source models of tectonic release of the Archambeau type and obtains numerically their P- and S-wave spectra. The spectra for these particular models show a definite difference in the P-waves and S-wave corner frequencies. This approach yields theoretical $M_S$ vs $m_b$ curves which can be discussed in terms of scaling by the various source parameters. The second approach is to obtain the earthquake P-wave source history from comparison of theoretical and observed seismograms. This history is then used as the S-wave source history to obtain theoretical seismograms which are compared with S-wave observations of the same events.

Two possible criticisms of the first approach is that the Archambeau type tectonic release source does not model the tectonic region realistically and that the
presence of a zero frequency minimum on the far-field spectrum is due to a questionable termination of the tectonic release region at a finite distance from the source. Archambeau claims that the approximation using a finite termination is realistic and closer to actual field situations than using the same spatial release distribution to infinity. We are continuing research on tectonic release sources using more realistic models of tectonic stress distributions. In particular we are trying to determine for what stress and body force distributions the zero frequency minimum is a valid result.

The use of synthetic body and surface waves is becoming an extremely valuable tool in not only determining the character of the seismic source but also in defining the seismic propagation properties of the earth. In the semiannual report, we gave an example of a method using amplitudes and arrival times of body and Rayleigh waves from NTS events to determine an average crustal model from NTS to mid-Arizona. In this report, transversely polarized shear waves (SH) from well located earthquakes are used to determine an upper mantle shear velocity structure. It is further demonstrated that observations of travel times for vertically polarized shear waves (SV), the most common technique, can lead to erroneous shear structure models due to continuation of P-SV interactions.

In Section III, the spatial and frequency distribution of path-corrected Rayleigh waves from the San Fernando earthquake are systematically related to the faulting process as a multiple event. The surface wave source is taken to be a depth-distributed set of double couples. A least-squares inversion is used to find the set of source parameters which optimally fit
the variance-weighted data. In this study the strength, the slip angle, and the rupture delay times are optimally determined for the spatially assumed distribution of double couples. A sophisticated error analysis is performed to estimate the uncertainties of the calculated model variables.

One of the interesting side results from this inversion technique is the a priori determination of what azimuthal distribution of observations is most beneficial in determining the source parameters, the uncertainties in the calculated model parameters, and the relative coupling between various parameters. Of the 18 WWSSN stations used for this event, two stations contain a total of 30% of the total information in the data set. In fact, the data set could have been reduced by one-half and only 25% of the total information which constrains the model variables would have been lost.

One could visualize, using these inversion operators to design multiple explosion events which would make discrimination criteria unreliable except in azimuths where data was difficult to obtain. Of course, the inversion operators used here were for Rayleigh wave data only. For discrimination techniques based on Love and Rayleigh waves possibly combined with body waves the design would probably be extremely complicated and economically unfeasible.

Of equal importance, this section describes and uses techniques for correcting surface wave data for mixed path crustal structure between source and receiver, and for multipath propagation interference. The error bounds associated with these corrections are estimated, and their influence on the reliability of the model are determined.

Since the crust and upper-mantle shear structure dominate the propagation characteristics of Rayleigh and Love surface waves, Section IV deals
with an attempt at constructing a preliminary shear velocity model compatible with both travel times and waveforms of observed seismograms. Transversely polarized (SH) LRSM and WWSSN observations are used in the study. Vertically polarized shear waves (SV) are shown to be inadequate in identifying triplications because of contamination by PL-coupled shear waves and other P-SV interactions. Accurately located west coast earthquakes, modeled as shear dislocations, are used as sources. The observed SH travel times are considerably slower than the JB values out to 30 degrees. Select profiles of observed waveforms in conjunction with the travel-times provide the data for model determinations. This is accomplished by fitting the observed waveforms with synthetics.

Research is continuing on the relations between seismic source characteristics and magnitude. Recently Anderson and Kanamori have made a study on the relationships between energy, moment, magnitude and source dimensions. Using both deterministic and statistical source models, they have shown that the empirical Gutenberg-Richter magnitude-energy relationship has strong theoretical support and implies that rupture times for large earthquakes are much longer than rise times. In general, earthquake magnitudes (M) are proportional to $L^2$, where $L$ is a linear source dimension, but $M \sim L^3$ for small events and $M \sim L$ for very large events. Interplate and intraplate earthquakes form separate populations in moment-magnitude, or moment-dimension plots. In each population the inferred stress drop and effective stress is roughly constant. The large scatter in plots involving magnitude is considerably reduced when earthquakes are grouped according to type. The scatter virtually disappears when magnitude is removed as a variable, e.g., dimensions versus moment plots.
II. Abstracts of Publications and Reports

During This Contract Period

Displacements at the surface of a layered halfspace produced by a point source dislocation is investigated. Expressing the source in terms of P, SV and SH displacement potentials allows the solution to be expanded in generalized rays. The transient response for each ray is obtained by the Cagniard-deHoop method. First-motion and high-frequency asymptotic approximations of the exact solutions are discussed.

Numerical experiments on the far-field response produced by shallow dislocations are presented. Synthetic seismograms computed for a strike-slip dislocation are compared with some observations from the Borrego Mountain earthquake. Preliminary results indicate an equivalent point source depth of 9 km with the far-field time function approximated by a step function with an exponential decay. This time function fits both the P and S waveforms. The apparent shift in corner frequency between the P and S waves for shallow events as reported by some investigators is explained by surface reflections.


In order to study regional variations in the upper mantle structure in the central United States, the second step of a two step procedure formulated in Kurita (1972a) has been taken and experimentally shown to be very powerful method for elucidating the fine configuration of the low-velocity zone. The strong advantage of this method as compared with the other methods, is a mutual independency of the extent of the velocity decrease and the depth to the bottom of the low-velocity zone. The upper mantle structure down to about 220 km in this region has been inferred from complex transfer ratios of long period P and S waves from deep-focus earthquakes recorded at FLO, OXF, and SHA on the basis of the crustal models obtained in Kurita (1972b). From the Interior Plain to the Gulf of Mexico the low-velocity zone shifts to a shallower depth while increasing its thickness and decreasing its velocity. This zone is about 50 km thick layer ranging in depth from about 150 to 200 km under the Interior Plain, about 75 km thick layer from about 120 to 195 km under the Gulf Coastal Plain, and about 80 km thick layer from about 95 to 175 km under the continental shelf of the Gulf of Mexico, all nearly along 89°N longitude. The decrease in S wave velocity at the top of this zone is about 0.30, 0.45, and 0.70 km/sec under each of the areas above, although the last value may be somewhat an overestimate. Both boundaries of this zone are sharp, the transition occurring over at most about 10 km. In this region the existence of the high-velocity lid zone is possible.

When an earthquake point source is represented in the most general way (a stress drop source) six independent functions of time (the six stress drop elements) are required to describe the source. The common seismological practice of using a single time function for a point source model of actual earthquakes is not valid unless all six time functions are independently obtained from observed data and found to be the same. The difficulty first noted by Knopoff and Randall (1970) (that one double couple lacked sufficient degrees of freedom to represent an arbitrary point source but that there was not a unique decomposition of a source into a combination of a double couple and a compensated linear vector dipole (CLVD)) is shown to result from their choice of non-orthogonal basis elements. The unique equivalent force representation of an earthquake source is shown to be three perpendicular force doublets parallel to the principal axes of the stress drop tensor. The success of the double couple source model, a special case of the stress drop source, in matching most observations of shallow earthquakes implies, but does not require that most shallow earthquakes result from shear dislocations along essentially flat surfaces.


The Green's function solutions for a shear dislocation with symmetry in the direction of infinite length have been integrated analytically over the direction of infinite length. The displacement solution is reduced from a surface integral over the fault plane to a line integral of temporal convolutions over the width of the fault. If the time history function of fault displacement is any piecewise linear function the convolutions can be integrated analytically, reducing the solutions to line integrals. Numerical results and plots are presented for a simple example. The two-dimensional solutions give exact first motions at points "over" the fault. Also the solutions can be used to test two-dimensional finite difference and finite element computer programs.


This is an attempt at constructing a preliminary shear velocity model compatible with both travel times and waveforms of observed seismograms. Transversely polarized LRSM and WWSS observations are used in this study. SV signatures are demonstrated to be inadequate in identifying triplications because of contamination by PL-coupled shear waves and other P-SV interactions. Well located west coast earthquakes, modeled as shear dislocations, are used as sources. The observed SH travel times are considerably slower than the JB values out to 30 degrees. Select profiles
of observed waveforms in conjunction with the travel-times provide the data for model determinations. This is accomplished by fitting the observed waveforms with synthetics.

Synthetics are computed for a number of current models as preliminary attempts at fitting the data. Models containing large, sharp transitions predict strong second arrivals beyond 30°. Models containing low velocity zones between 400 and 700 km predict complicated waveshapes between 20 and 25°. None of the above features are apparent in observed SH waveforms. The final model is relatively smooth and exhibits the same characteristics as the HWB P-model, except that the percentage velocity jump is much less at 400 km and slightly larger at 500 km. The 600 km transition is subdued similar to the P-model.


A linear array of eight Caltech portable broad-band seismograph trailers was set out from NTS to near Phoenix, Arizona, for the pre-announced underground nuclear test, Osoyoos, on September 21, 1972. Travel time and amplitude information were used to find an average crustal model by calculating synthetic seismograms using the Cagniard-deHoop method. Rayleigh waves from other nuclear events at NTS, as recorded at the Tucson WWSS station, were examined as a control for determining the structure of the top half of the crust. Group velocity curves were found and synthetic Rayleigh waves calculated for Tucson and Kingman (LRSM). The formations and characteristics of Pn, a reflected Pn, and the Pg phases were examined. Pg is demonstrated to be composed of the primary P reflection from the mantle and contains multiple arrivals of P-SV conversions. Comparisons of synthetic and observed seismograms indicate a crustal thickness of 30 km with a Poisson's ratio of .23. The crust-mantle transition appears to be sharp jumping from 6.1 to 7.9 km/sec. The amplitude behavior of Pn shows little evidence of any lid structure.
Hanks, T.C., "Long-period Strong Ground Motion Following the San Fernando Earthquake."

Two hundred thirty-four components of ground displacement are the basis of an investigation of long-period (approximately 2-10 seconds) strong ground motion following the San Fernando, California, earthquake at local (Δ ≤ 100 km) distances. Seismic moment, source dimension, radiation pattern, source propagation, the development of surface waves together with their subsequent dispersion, and the local structure of the Los Angeles Basin all appear to have first order significance in the shaping of observed ground displacements and their variation with azimuth and epicentral distance. Long-period strong ground motion occurred with considerably greater amplitude at southern azimuths than at northern azimuths; at comparable epicentral distances amplitude differences approached a factor of 10. Radiation pattern alone seems insufficient to explain the observed variations. Especially strong signals at azimuths due south of the epicenter and the development of large amplitude (3-5 cm), long duration surface wave trains at stations along the southeastern margin of the Los Angeles Basin suggest the significant effect of this geologic structure on the observed ground displacements. Data redundancy at closely spaced stations in Los Angeles and at Lake Isabella Dam suggest that errors in ground displacement at periods near 8 seconds are approximately 1 cm and at periods near 15 seconds are several cm, for a record sensitivity of 7.6 cm/g.

Newton, C.A., "Temporal Changes in P-Wave Velocity Near Riverside, California."

P travel times from NIS to Goldstone and Riverside, California have been measured for the last eight years. During this period there have been four or more cycles of compressional velocity changes in the crust NNE of Riverside, but not necessarily limited to that area. Periods of lower velocity are associated with decreased seismicity and periods of higher velocity are associated with increased seismicity. These data suggest that this area which includes portions of the San Jacinto and San Andreas fault zones have repeatedly strained to a dilatant state and then fractured. The largest earthquake related to these velocity changes was a magnitude 5.4 shock which followed a low velocity period of approximately 450 days. It appears that this area is now in a dilatant state which began on August 1972.

Whitcomb, J.H., "P'P' Times and Average Mantle Velocities Under Continents and Oceans."

P'P' times are an important data type because they provide one of the few means at hand for comparison of mantle structure under different lithospheric regions such as continents and oceans. However, scatter of the
data for this phase has long discouraged investigators from using the readings for anything more than an approximate check on P'P' times. A partial explanation for this scatter is asymmetric P'P' which arrives earlier than the symmetrically reflected phases, but at least some of the difficulty lies in assignment of the incorrect P'P' branch to a reading and in assumption of the incorrect velocity structure at the reflection point. These problems are avoided and the analysis suggests that one-way travel of P waves through oceanic mantle is delayed by 0.65 to 0.95 sec relative to United States mid-continental mantle.

American Geophysical Union, Fifth-Fifth Annual Meeting, April 8-12, 1974, Washington, D.C.


Long-period seismograms recorded at Pasadena are used to determine the focal process of the 1960 Chilean earthquake. Synthetic seismograms computed for various fault models are matched with the observed strain seismograms to determine the fault parameters. A low-angle thrust model with rupture length of 800 km and rupture velocity of 3.5 km/sec is consistent with the observed Rayleigh to Love wave ratio and the radiation asymmetry. A seismic moment of 1 to 3 x 10^{30} dyne cm, the largest ever reported, is obtained over a period range from 200 to 300 sec. This value together with the estimated fault area of about 2 x 10^{5} km^2 fits into the recently established moment vs. fault area systematics. The excitation efficiency seems to increase with the period. The strain seismogram clearly shown unusually long-period (300 to 600 sec) wave arriving at the P time of a large foreshock which occurred 15 minutes before the main shock, suggesting a large slow deformation in the epicentral area prior to the major failure; the entire focal process may be envisaged in terms of a large-scale deformation which started rather gradually and eventually triggered the foreshocks and the "main" shock. This mechanism may explain the large premonitory deformations documented, but not recorded instrumentally, for several Japanese earthquakes.


A linear inversion scheme is presented to fit in a least-squares sense discretized surface wave spectra. With this technique, an estimate of the uncertainty in the predicted source model is given when errors in the observed data are considered. An earthquake source is represented by a series of double couples which can be offset both spatially and temporally to approximate a moving rupture. A comparison of this representation of a moving source and that classically representing a finite source is given. The inversion scheme yields the source parameters (moment and slip angle)
for the individual double couples as well as rupture velocity. Rayleigh wave spectra from a suite of stations recording the 1971 San Fernando, California earthquake are examined and the resolvability of the inverted source parameters are discussed.
III. Least-Squares Inversion of Surface Above Spectra to Obtain Source Parameters: The 1971 San Fernando Earthquake

This section is Chapter 5 and Appendices 3 and 4 of Ralph Wilson Alewine's thesis (1974) titled "Application of Linear Inversion Theory Toward the Estimation of Seismic Source Parameters". References and details cited in this section can be found in the thesis. This Thesis has been submitted and successfully defended as partial fulfillment of the requirements of the Degree of Doctor of Philosophy at the California Institute of Technology, Pasadena, California.
Chapter 5

Least-Squares Inversion of Surface Wave Spectra to Obtain Source Parameters: The 1971 San Fernando Earthquake

5.1 Introduction.

In this chapter, we will examine the spatial and frequency distribution of Rayleigh waves from the 1971 San Fernando, California, earthquake and how this distribution is related to the faulting processes of that event. The use of surface waves to determine gross source parameters has been a common tool in recent seismological investigations, e.g. Wu (1968), Tsai and Aki (1970, 1971), Canitez and Toksöz (1971, 1972), and Mitchell (1973). Unfortunately, the problem of systematically finding values of the source variables which optimally fit the observed data has not yet received the attention due the problem. With the exception of a Monte Carlo technique by Tsai (1972) and an iterative technique by Turnbull et al. (1973), the "trial and error" method was usually employed in finding suitable variables which describe the data.

We hope to extend the usefulness of surface observations by showing how an inversion procedure can be applied to surface wave observations in order to see what information can and cannot be gained about the earthquake source.
Along with this inversion, we will use a method of expressing the finiteness of the fault plane by numerically summing point source solutions. This method will allow for the variation of parameters over the fault surface. Finally, we would like to compare the solution that we obtained for the San Fernando source mechanism from the static study in the previous chapter to that obtained by the dynamical surface waves.

5.2 Theoretical Model.

Spectra from a Point Source. Theoretical surface wave spectra will be computed for a source in a multilayered medium which approximates the San Fernando area. The elastic parameters for this medium are given in Appendix 5. Far-field amplitude spectra depend on the surface wave amplitude response of the layered medium and also the source type, depth and orientation. Expressions for the frequency dependent media response factor, which is dependent of the source, are given by Harkrider (1964). Computation of this factor involves use of the Thomson-Haskell matrices for the multilayered half-space (Haskell, 1953). The formulation for the component of the surface wave spectra which is due to different source types and the orientation of the source was developed by Ben-Menahem and Harkrider (1964). The far-field solution for Rayleigh wave
spectra for a point double-couple can be written as

\[ A(\omega) = \frac{S(\omega) k A_R X(e,h) e^{-i(\kappa r + \frac{3\pi}{4})}}{\sqrt{\kappa r}}, \quad (5.1) \]

where

\[ S(\omega) = \text{transformed spectral source function} \]
\[ k = \text{Rayleigh wave wave number} \]
\[ A_R = \text{medium response factor} \]
\[ r = \text{radial distance to receiver} \]
\[ X(e,h) = \text{complex radiation pattern function, where } e \]
\[ \text{is the azimuth of the station relative to the} \]
\[ \text{source and } h \text{ is the source depth.} \]

The radiation pattern function is given by Ben-Menahem and Harkrider (1964) and by Harkrider (1970). This function is dependent on the source parameters and the source depth. At a given depth, this azimuth dependent function is characterized by the slip, or rake, angle and the dip of the source-equivalent fault. The numerical techniques used in the computer programs which calculate these functions are discussed by Harkrider (1970).

The source function is assumed to be a step function in moment. This is given by

\[ S(t) = \begin{cases} 
0 & \text{for } t < 0 \\
M_0 & \text{for } t > 0 
\end{cases}, \quad (5.2) \]
where $t$ is the origin time of the event.

**Source Finiteness Effects.** In the above discussion, the source was represented by a point source. However, if the dislocation on the actual fault surface initiates at one point and propagates to another point, the fault finiteness and rupture velocity can become important. Ben-Menahem (1961) has shown how the dimensions of the source and the speed of rupture play an important role in the resulting spectral radiation patterns. This author shows that spectra from a source which radiates energy evenly as propagation occurs was modulated by the shift factor,

$$
\frac{\sin(X)}{X} e^{ix},
$$

(5.3)

where

$$
X = \frac{\omega \cdot b}{2C} \left[ \frac{C}{v} - \cos(\theta_0) \right].
$$

The following parameters are defined:

- $\omega$ = angular frequency
- $C$ = Rayleigh wave phase velocity
- $b$ = horizontal rupture length
- $v$ = horizontal rupture velocity
- $\theta_0$ = angle measured from the direction of rupture.
However, if the source radiates energy unevenly during propagation, then this simple modulation no longer applies. One such method of uneven radiation can occur by having the dislocation (i.e., moment) vary along the fault. The fact that surface wave excitation is depth dependent means that a fault which has a vertical component of propagation can also give rise to uneven radiation during propagation. The analytic modeling of such rupture would be very difficult for a layered medium, however, we give here a numerical approximation to this phenomena.

**Numerical Approximation of a Propagating Fault.** In this section, we will present a method of numerically approximating the surface wave spectra due to a propagating rupture. To do this, we will consider the fault to be modeled by a discrete number of double couple sources arranged spatially to reflect the length of a fault. Propagation along the fault occurs by "turning on" each of the sources sequentially at a time given by

$$T_{d_i} = \frac{\Delta s_i}{V_R}$$

where $\Delta s_i$ is the distance to the $i$th individual source from some origin and $V_R$ is some rupture, or propagation, velocity that we wish to approximate. The spectra from
each of the sources will have a different phase at some fixed observation point. The total spectra measured at this point will be the complex sum of the individual contributions. The phase delay of the individual propagation sources relative to the initial source taken to occur at the origin is given by

$$\tau_{p1} = \frac{\hat{R}_1 - R_0}{C(\omega)} + T_{d1}$$

where $R_0$ is the distance from the initial, or reference source, to the exterior observation point and $\hat{R}_1$ are the distances from the other sources to the observation point. $C(\omega)$ is the phase velocity. The geometry of this problem is illustrated in Figure 5.1. From this figure, we can write

$$\hat{R}_1 = [d_{s_1}^2 + R_0^2 - 2 d_{s_1} R_0 \cos(\lambda - \theta_{sp})]^{1/2},$$

where $d_{s_1}$ is the horizontal separation of the individual sources relative to the initial source taken in the direction of rupture, $\lambda$. $\theta_{sp}$ is the "azimuth" to the station relative to the initial source. Appendix 4 goes into the details of how the spectra are summed given this phase information. Both strike slip and dip slip faulting can be approximated with this geometry.
Figure 5.1. Geometry of the assumed multiple double couple source.
We caution here that the angle \( \theta_F \) as defined in Figure 5.1 is not the same angle as the angle \( \theta \) in the complex radiation pattern term of equation (5.1). A preliminary examination of Figure 1 in the Ben-Menahem and Harkrider (1964) paper would indicate that these two angles are equivalent, but their definition of \( \theta \) is in error. The angle \( \theta \) should be measured positively clockwise from the positive strike direction when viewed above the half-space.

As a check to this numerical approximation, we will model a vertical strike slip earthquake and check the resulting spectra with that calculated by the method of Ben-Menahem (1961). This model consisted of 16 sources, each at a depth of 20 km, with dip 90°, slip angle 0° (pure left lateral faulting), and identical moments. The sources were spaced at 2 km increments along the strike of the origin source (\( \lambda_s = 180° \)) so that the fault length and propagation length were 30 km. Spectra were computed for apparent propagation velocities of 1.5, 3.0 and 4.5 km/sec at a point in a direction 45° to the direction of propagation in order to avoid a node in the radiation pattern of strike slip events. These spectra were then compared to that calculated for a single point source with the shift term for this fault length and propagation velocities. The moment of the single source was taken as the simple sum of
the moments in the discretized model. The results are shown in Figure 5.2.

In this figure, we have normalized the amplitudes to the values calculated for a non-propagating point source at the depth of the source at the point of initial rupture. By utilizing this normalization, we cancel the media source depth effects on the spectra. The ratio shows only the effects of the fault finiteness. In Figure 5.2a, we see that the directivity calculated by this numerical technique matches very closely that given by Ben-Menahem (1961). For propagation velocities 3.0 km/sec and greater, there is less than 1% difference in the two calculations. We see that the directivity is a very pronounced function in the period range of interest for slow propagation velocities.

Now that we have shown that the directivity function can be accurately calculated numerically, we now examine the effect of dip slip propagation on surface wave amplitude spectra. A model of a 45° pure dip slip fault was constructed. The initial motion occurs at a depth of 20 km and propagates at a constant velocity up the fault to a depth of 2 km. A total of 10 double couple sources were spaced at a 2 km depth and lateral increments to approximate this fault. The propagation distance along the fault is 28.3 km, and the pure horizontal propagation distance is
Figure 5.2. Comparison of the frequency-dependent directivity function given by the Ben-Menahem formulation and the numerical approximation. Upper—results for a 30 km long left lateral strike slip fault. Lower—results for a 28.3 km wide 45° dip slip fault.
18 km. The spectra were calculated at an exterior point which is in the direction of propagation. Spectra were also calculated using the Ben-Menahem formulism for the same parameters. All the amplitudes were normalized to a non-propagating point source at a depth of 20 km. Figure 5.2b shows a comparison of the resulting directivities for this dip slip case.

We see in this figure that there is a vast difference in the spectra computed by the two methods. The most significant difference is the accentuation of the amplitudes in the 25-60 sec period range. As a check as to whether the large amplitudes calculated in the numerical case were due to a Doppler effect or due to the non-uniform depth excitation along the propagation length, we normalized these amplitudes to the spectrum which was calculated using the numerical approach when an infinite propagation velocity was used. This normalization cancels any non-uniform depth excitation since the depth distribution was identical in both cases. The resulting directivity was almost identical to that calculated by the Ben-Menahem method and shown in the lower part of the figure.

We thus conclude that the effect of a source radiating energy as it propagates from one depth to another during faulting can be much larger than the Doppler effect and
cannot be ignored when spectra due to this type of faulting are being modeled. The depth effect on the spectra will vary from medium to medium and from source type to source type. No attempt was made here to catalogue this effect for various media and propagation parameters.

**Approximation of San Fernando Faulting.** In the previous chapter, we discussed the geophysical observations which constrained the fault model geometry for the 1971 San Fernando, California, earthquake. In this chapter, we would like to adopt an approximation to this same geometry to try to explain the radiation of surface wave energy from this fault. This approximation is accomplished by using four double-couple sources. The primary constraint for limiting the approximation to only four sources was economic, since the forward problem had to be recalculated after each iteration for the partial derivative matrix. However, we consider that the approximation chosen is adequate in light of the periods of the surface waves used. The parameters which describe this fault geometry and remain fixed in the inversion are listed below.

<table>
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<th>Source No.</th>
<th>depth(km)</th>
<th>dip</th>
<th>d_s(km)</th>
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<tr>
<td>1</td>
<td>14.5</td>
<td>52°</td>
<td>0.0</td>
</tr>
<tr>
<td>2</td>
<td>10.5</td>
<td>52°</td>
<td>2.49</td>
</tr>
<tr>
<td>3</td>
<td>6.25</td>
<td>35°</td>
<td>7.51</td>
</tr>
<tr>
<td>4</td>
<td>2.00</td>
<td>35°</td>
<td>14.36</td>
</tr>
</tbody>
</table>
These loci of the four point sources do not lie on a straight depth-distance line, but rather define the center line of a fault plane whose dip increases with depth. In calculating the propagation delay time for each of the elements, the distance, \( \Delta_1 \), is used. \( \Delta_1 \) is given by

\[
\Delta_1^2 = (h_{s_1} - h_{s_1})^2 + d_{s_1}^2.
\]

These distances are the straight line distances from the initial source to the three other double-couples. The delay time error caused by using this distance rather than the distance measured along the curved "fault plane" is less than 1 sec in the extreme case.

5.3 Forward Problem Formulation.

In this section, we wish to pose a relationship between the source model parameters and the observed spectra in a linear form as in equation (2.3). As we did for the static problem, we will propose some fault geometry and use an inversion scheme to solve the fault model parameters for this geometry. In order to describe the data, we will use a set of double-couple sources that are arranged spatially to approximate the fault model system which was presented in the previous chapter. Some of the parameters of the individual sources, specifically those which
describe the geometry of the fault system, will be set initially and not changed in the inversion. These parameters are the depth and dip associated with individual source elements and the spatial relationship between elements. The variable parameters for each element are taken to be the moment and slip angle. In addition, we will include the propagation velocity and propagation direction as variables for the entire system.

The model response matrix, \( A \), for this problem is given to be composed of partial derivatives of the form

\[
\frac{\partial S_i}{\partial X_k}, \tag{5.4}
\]

where \( S_i \) is the spectral amplitude value at some frequency \( f_i \) (\( i=1,L \)) measured at some "azimuth" \( \theta_{Fj} \) (\( j=1,J \)) with respect to the fault system. \( X_k \) (\( k=1,K \)) is the source parameter which is allowed to be varied. The matrix \( A \) has \( K \) columns and \( L \times J \) rows, where \( K \) is the number of model variables (specified below), \( L \) is the number of discrete spectral points at a single station and \( J \) is the number of stations used. For this case, \( L=10, J=18, K=10 \).

We see from Appendix 4 that the addition of multiple sources is a non-linear operation. This problem can be overcome by mapping only small increments of the model into
changes in the data. The criterion for this linearization is the same as was discussed in Chapter 2. The partial derivatives which compose the matrix $A$ were computed numerically for a given model. This was done by the simple differencing,

$$\frac{\partial S_i}{\partial x_k} \approx \frac{S_i(m + \Delta m_k) - S_i(m)}{\Delta m_k}.$$  

The incremental values of the source parameters used were:

- $M_0$ - moment = $0.1 \times 10^{26}$ dyne-cm
- $\Delta \lambda$ - slip angle = $10^\circ$
- $\Delta \lambda_R$ - propagation direction = $10^\circ$
- $\Delta V_R$ - propagation velocity = 0.25 km/sec.

5.4 Inversion Procedure.

An inversion procedure will be adopted to find the model parameters which best fit the observed spectral data. Since the number of data points far exceed the number of variables, we will fit the data in a least-squares sense. Appendix 3 gives the derivation of the inversion operators for the general least-squares case.

Since we have mixed units in the source model parameters and thus mixed units in $A$ and $\delta m$, a problem of
scaling arises. One way of correcting for the dimensionality of the parameters is through the use of the weighting, or model correlation, coefficients. The weights of the source parameters can be adjusted so that the weighted coefficients of the rows of $A$ are about equal. For a non-linear problem, this scaling would be model dependent and thus would probably vary in the inversion. This method, although it can give satisfactory results in obtaining a fit to the data, is not esthetically pleasing. A satisfactory method was found to solve this scaling problem. This was accomplished by non-dimensionalizing the problem,

$$ A \delta m = \delta d , $$

by dividing through by the data and model parameters. The dimensional matrices and vectors are replaced by their non-dimensional counterparts in the inversion. The components of the quantities are defined as:

$$ \hat{A}_{ij} = \frac{m_j}{d_{0i}} A_{ij} \quad \delta \hat{a}_i = \frac{1}{d_{0i}} \delta d_i $$

$$ \hat{c}_{nn_{1i}} = \frac{1}{d_{0i}^2} c_{nn_{1i}} \quad \delta \hat{m}_j = \frac{1}{m_j} \delta m_j . $$

We note that this scaling is also non-linear.
Throughout the inversion calculations carried out in this chapter, we will limit the absolute value of a model perturbation in one iteration to that used in the differencing method of calculating the partial derivatives in the forward problem. This was done to avoid projecting any model perturbation outside the range of assumed linearity.

As a test to the inversion scheme, we will calculate the theoretical spectra at 18 far-field points (corresponding to the location of the 18 stations which are described in the next section) for a given set of source parameters. The non-variable source parameters are the same as those used to describe the San Fernando faulting. These theoretical spectra were then input into the inversion scheme as "data" to see how accurately the corresponding variable source parameters could be recovered. The results of this test are shown in Figure 5.3. The initial values for the source parameters are that shown on the left at Iteration = 0. The source values used in calculating the "data" are shown by the arrows on the right hand side of the figure. We see that at least 6 iterations are required before the solution starts to converge to the true values, and after 11 iterations, the estimated values are very close to the true values. The restriction that the maximum size of the perturbations can be no larger than the differencing
Figure 5.3. Test to check the amplitude spectra inversion technique. Initial values of the fault parameters are shown on the left; the theoretical values are indicated by arrows on the right.
values slows the convergence somewhat. Not shown in this figure is the estimate of the propagation velocity and propagation direction for each iteration. These two quantities varied more smoothly and converged more quickly to the theoretical values than did the moments and slip angles. The propagation velocity and propagation direction converged to 4.2 km/sec and 120° from an initial value of 3.2 km/sec and 100°, respectively.

In this test, the variance of each "datum" was taken to be 5% of the calculated value. The test was repeated with the data variances set at 25% of the calculated values. Although the convergence to the true model values was not as good as in the previous example, the solution remained stable through the iterations, and a fair estimate of the model parameters was obtained. The results from these experiments were encouraging enough to attempt to apply the technique to actual observed surface wave spectra.

5.5 Data.

The source of the surface wave data used in this study was from the long period vertical component instruments of the World-Wide Standard Seismograph Network (WWSSN). Only the vertical Rayleigh wave component at each station was considered in this study. Data from the Canadian Standard Stations and the WWSSN stations located in the continental
United States could not be used since either the long period vertical component was inoperative or off-scale during the passage of the Rayleigh wave train. The original intent in collecting this data was to limit the stations used to those whose epicentral distances were less than 60°-70°. However, to provide a more uniform and extensive azimuthal coverage of stations, this condition was relaxed somewhat. The closest station in a given azimuthal increment from the epicenter which gave a readable record was chosen. A total of 18 records was chosen to be examined, and these stations provided a fairly good azimuthal coverage for this event. Table 5.1 gives a list of all these stations used, the epicentral distances, and the geodetic azimuth of the station with respect to the epicenter.

The Rayleigh wave signal was identified on each of the records and digitized at an irregular interval taking adequate care to sufficiently define the trace. These data were then linearly interpolated to an equal increment rate of one sample every two seconds. The digitization was restricted to a velocity window of between about 4.3 km/sec to about 1.5 km/sec. There was a slight variance to this window depending on the distance and source-receiver path. The choice of the velocity window was based on a visual judgment as to when the signal arrived and essentially
### Table 5.1

<table>
<thead>
<tr>
<th>Station</th>
<th>Geodetic Azimuth</th>
<th>Distance (Deg)</th>
<th>% Land</th>
<th>% Ocean</th>
<th>Geom. Fact.</th>
<th>Geographic Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>NOR</td>
<td>10</td>
<td>58</td>
<td>90</td>
<td>10</td>
<td>1.66</td>
<td>Nord, Greenland</td>
</tr>
<tr>
<td>KTG</td>
<td>23</td>
<td>60</td>
<td>77</td>
<td>23</td>
<td>1.68</td>
<td>Kap Tobin, Greenland</td>
</tr>
<tr>
<td>AKU</td>
<td>27</td>
<td>63</td>
<td>75</td>
<td>25</td>
<td>1.70</td>
<td>Akureyri, Iceland</td>
</tr>
<tr>
<td>VAL</td>
<td>38</td>
<td>74</td>
<td>55</td>
<td>45</td>
<td>1.76</td>
<td>Valentia, Ireland</td>
</tr>
<tr>
<td>MAL</td>
<td>47</td>
<td>86</td>
<td>61</td>
<td>39</td>
<td>1.80</td>
<td>Malaga, Spain</td>
</tr>
<tr>
<td>BEC</td>
<td>77</td>
<td>45</td>
<td>78</td>
<td>22</td>
<td>1.51</td>
<td>Bermuda</td>
</tr>
<tr>
<td>SJG</td>
<td>96</td>
<td>49</td>
<td>69</td>
<td>31</td>
<td>1.56</td>
<td>San Juan, Puerto Rico</td>
</tr>
<tr>
<td>CAR</td>
<td>104</td>
<td>53</td>
<td>43</td>
<td>57</td>
<td>1.61</td>
<td>Caracas, Venezuela</td>
</tr>
<tr>
<td>BHP</td>
<td>116</td>
<td>44</td>
<td>70</td>
<td>30</td>
<td>1.50</td>
<td>Balboa Heights, Canal Zone</td>
</tr>
<tr>
<td>CJE</td>
<td>137</td>
<td>44</td>
<td>70</td>
<td>30</td>
<td>1.50</td>
<td>Galapagos Islands</td>
</tr>
<tr>
<td>PEL</td>
<td>141</td>
<td>81</td>
<td>16</td>
<td>84</td>
<td>1.79</td>
<td>Peldehue, Chile</td>
</tr>
<tr>
<td>SBA</td>
<td>194</td>
<td>120</td>
<td>0</td>
<td>100</td>
<td>1.67</td>
<td>Scott Base, Antarctica</td>
</tr>
<tr>
<td>WEL</td>
<td>224</td>
<td>97</td>
<td>0</td>
<td>100</td>
<td>1.79</td>
<td>Wellington, New Zealand</td>
</tr>
<tr>
<td>API</td>
<td>236</td>
<td>70</td>
<td>0</td>
<td>100</td>
<td>1.74</td>
<td>Alifamalu, Samoa</td>
</tr>
<tr>
<td>KIP</td>
<td>260</td>
<td>38</td>
<td>0</td>
<td>100</td>
<td>1.41</td>
<td>Kipapa, Hawaii</td>
</tr>
<tr>
<td>GUA</td>
<td>285</td>
<td>88</td>
<td>0</td>
<td>100</td>
<td>1.80</td>
<td>Guam Island</td>
</tr>
<tr>
<td>MAT</td>
<td>307</td>
<td>80</td>
<td>1</td>
<td>99</td>
<td>1.79</td>
<td>Matsusicho, Japan</td>
</tr>
<tr>
<td>COL</td>
<td>339</td>
<td>35</td>
<td>69</td>
<td>31</td>
<td>1.36</td>
<td>College, Alaska</td>
</tr>
</tbody>
</table>

Table 5.1. Stations used in this surface wave study.
ended at a particular station and on the dispersion curves expected for the traveled path. Typical of the dispersion curves used were those of Brune and Dorman (1963), McEvilly (1964), Santo (1963), Tryggvason (1962), and Brune (1969).

The Fourier spectra of these records were then taken with a numerical transform routine. The spectra were calculated for the frequency range from 60 sec period to 8 sec period, with 512 discrete frequency points in this range. The spectra were smoothed slightly with a two-point Hanning filter such that for the $i$th frequency component of a particular record

$$f_i(\omega) = \frac{2}{5} \left\{ \frac{1}{4}[f_{i-2}(\omega)+f_{i+2}(\omega)]+\frac{1}{2}[f_{i-1}(\omega)+f_{i+1}(\omega)]+f_i(\omega) \right\}.$$ 

**Instrument Correction.** The instrument frequency and amplitude response for the World-Wide long period seismograph system were calculated from the published instrument and coupling constants for that instrument. These calculations are based on the relations given by Hagiwara (1958) and Mitchell and Landisman (1969). Both an amplitude and frequency correction were supplied. Since some of the stations that were used in this study operate at different instrument gains during different parts of the year, a check was made on each seismogram to verify the listed
instrument gain. This was done by comparing the height of the calibration pulse to the applied current impulse. The magnification computed in this fashion was within about 10% of the listed magnification except for station SBA which appeared to be off by a factor of 2.

**Geometrical Spreading Correction.** All of the amplitude spectra $A(\omega)$ are corrected back to a common distance, $A_c$, by multiplying by the well-known relation for spreading on the surface of a sphere

$$C_{GS} = \sqrt{\frac{\sin \Delta}{\sin \Delta_c}} ,$$

(5.6)

where $\Delta$ and $\Delta_c$ are given in radians. For the data in this case, we chose our reference distance, $\Delta_c$, to be 2000 km.

**Attenuation Correction.** Following the instrumental and geometrical spreading corrections, the spectral amplitudes were corrected for attenuation. This correction takes the form

$$A_c(\omega) = A(\omega) e^{\gamma(\omega) x} ,$$

(5.7)

where $\gamma(\omega)$ is the amplitude attenuation coefficient appropriate for the particular source to receiver path, and $x$ is the epicentral distance. Since the paths from the San Fernando epicenter to most of the WWSSN stations used in
Figure 5.4. Amplitude attenuation coefficients assumed for the two structures used in this study.
this study involve mixed crustal type paths, the appropriate attenuation coefficient would necessarily be a path-averaged value. We attempted to take this into consideration in the following manner.

We assume that each of the transmission paths can be divided into two types: continental and oceanic. From the literature, we compiled estimates of the fundamental model Rayleigh wave attenuation coefficients for these two media. These estimates are summarized in Figure 5.4. The continental attenuation coefficients are reduced from data given by Gutenberg (1945), Gutenberg and Richter (1936), Tryggvason (1965), Nuttli (1973) and Mitchell (1973). The oceanic attenuation coefficient curve was reduced from data in papers by Gutenberg (1945), Ben-Menahem (1965), and Tsai and Aki (1969). The partitioning of continental and oceanic travel paths was estimated for each station, and the spectral amplitudes measured at that station were corrected by multiplying by the following relation,

\[ C_A = e^{\left[ x_L \gamma_L(\omega) + x_O \gamma_O(\omega) \right]} \]

(5.8)

where

\[ \gamma_L(\omega) = \text{the continental, or land, amplitude dissipation coefficient} \]
\[ \gamma_0(\omega) = \text{the oceanic amplitude dissipation coefficient} \]
\[ x_L = \text{the distance traveled over a continental type path} \]
\[ x_0 = \text{the distance traveled over an oceanic type path}. \]

Table 5.1 lists the percentage of the total path length that is spent in each of the assumed attenuation provinces. Also listed in this table is the geometrical spreading factor given by the equation in the above paragraph.

**Local Crustal Correction.** This correction attempts to compensate for the effects of surface waves traveling laterally from one structure to another. In the case under consideration, there are marked variations in the structure of the crust and upper mantle between the area in which the surface waves were generated and the area around the stations at which they are measured. Fortunately, the corrections applied along the path between these two points cancel out so that only a local crustal correction at the receiver needs to be applied to correct the spectral amplitudes back to the medium in which they were generated. An approximation which has given satisfactory results in predicting the amplitude changes of Rayleigh waves traveling across a lateral boundary has been given by McGarr (1969) and McGarr and Alsop (1967). The validity of this
approximation has been verified by wave propagation studies in laterally heterogenous media using the finite-difference numerical technique (D. Boore, personal communication, 1974).

If we assume that the total energy in the fundamental mode Rayleigh wave remains constant in the process of transmissions across the boundary margin, then we can write

\[
\frac{A_R}{A_S} = \sqrt{\frac{\omega_S(\omega)}{\omega_R(\omega)}},
\]

where

\[ A_S(\omega) = \text{the amplitude in the source medium} \]
\[ A_R(\omega) = \text{the amplitude in the receiver medium} \]
\[ \omega_S(\omega) = \text{the normalized energy flux in the source medium} \]
\[ \omega_R(\omega) = \text{the normalized energy flux in the receiver medium}. \]

The normalized energy flux is given by

\[ \omega(\omega) = 2 \, E(\omega) \, U(\omega), \]

where \( E(\omega) \) is the total potential or kinetic energy at a particular frequency normalized to the surface displacement. The calculation of this energy excited in a given layered earth model is given by Harkrider and Anderson.
Figure 5.5. Energy flux as a function of frequency calculated for various crustal and upper-mantle models. The velocity-density structures for the various models are given in Appendix 5.
### Table 5.2

<table>
<thead>
<tr>
<th>Crust &amp; Upper Mantle Type</th>
<th>Station</th>
</tr>
</thead>
<tbody>
<tr>
<td>San Fernando Crust over</td>
<td>Source Medium</td>
</tr>
<tr>
<td>Gutenberg Continent</td>
<td></td>
</tr>
<tr>
<td>Greenlandic Shield with Ice Cap</td>
<td>NOR, KTG, SBA</td>
</tr>
<tr>
<td>Iceland</td>
<td>AKU</td>
</tr>
<tr>
<td>Irish Continental Shelf</td>
<td>VAL</td>
</tr>
<tr>
<td>Iberian Shield</td>
<td>MAL</td>
</tr>
<tr>
<td>Mid-Atlantic Ocean</td>
<td>BEC, SJG</td>
</tr>
<tr>
<td>Western America Tectonic</td>
<td>CAR, BHP, COL</td>
</tr>
<tr>
<td>Pacific Ocean East</td>
<td>GIE</td>
</tr>
<tr>
<td>Pacific Ocean West</td>
<td>AFI, KIP, GUA, WEL</td>
</tr>
<tr>
<td>Andean</td>
<td>PEL</td>
</tr>
<tr>
<td>Central Japan</td>
<td>MAT</td>
</tr>
</tbody>
</table>

Table 5.2. Crust and upper mantle structure used for the local crustal structure spectral amplitude correction. Velocity-density models for the different structures are given in Appendix 5.
(1966, equation 1). \( U(\omega) \) is the group velocity for the particular medium.

At each of the 18 WWSSN stations used in this study, the crustal structure at that station was approximated and the velocity dispersion and the spectral energy density calculated. The crustal models used in this calculation and the references for these models are given in Appendix 5. The energy flux, \( W(\omega) \), given by these calculations is summarized in Figure 5.5. The stations for which each crustal model correction was used are listed in Table 5.2.

In order to correct the spectral amplitudes back to a common medium, we multiply by a correction which normalizes the amplitudes to the source medium. This correction takes the form

\[
C_{LC} = \sqrt{\frac{W_R(\omega)}{W_S(\omega)}}.
\]

Multipath Propagation Removal. The complex cepstrum technique which was established by Schafer (1969) has proven to be an effective means of removing spectral modulations caused by the interference of simultaneously arriving signals. Two or more signals can arrive simultaneously at a given station due to either multipathing of a single signal or can result from the addition of several seismic
signals, multisourcing, or some combination of these two effects. In this study, we want to eliminate as best we can from our observed spectra that modulation which is due to any multipathing effects and retain any modulations which may arise from the addition of multiple sources. The spectra which have been de-modulated from multipathing effects can then be examined by this method to try to determine the existence of whether the source is a single or multiple event.

Cohen (1970) and Flinn et al. (1973) have used the technique to detect the separation of the body phases P and pP for very shallow events, while Linville (1971) and Tsai (1972) have applied the method to surface waves. In the cepstrum technique, a homomorphic deconvolution is applied to separate the components of a convolution of a seismic signal and a multipath operator. Basically, the method acts like a filter applied to the amplitude spectra where the interference effect appears as a scalloping of the spectra over a wide frequency range. When a long pass filter is applied, the spectrum modulating effects are removed. This filtering is done by applying the appropriate bandpass to the cepstrum. Tsai (1972) and Linville (1971) give a summary of the technique involved and the appropriate filter bands to use in correcting spectra in the
Figure 5.6. Example of cepstral filtering at two stations. Original spectra are shown by dashed line; the refined, or filtered, spectra are shown by the solid line.
bandwidth which is involved in this study. Such a long pass filter was applied to the observed spectra. This spectral filtering had little effect except to smooth the spectra. A typical example of the removal of the spectral modulations is shown in Figure 5.6. The spectra are presented as a function of frequency to show how the modulations are frequency distributed. In this figure, the original spectra are shown by the dashed line, and the spectra after the application of the cepstrum technique are shown as a solid line.

We saw in the last chapter from the static dislocation models of the San Fernando earthquake that there appeared to be two areas of large dislocation -- one in the hypocentral region and one near the surface. If the dislocations from these two areas acted like two separate sources, separated spatially and temporally, then the interference of these two sources should appear in the complex cepstrum. The cepstra were calculated at several of the stations chosen in the opposite direction of the apparent rupture. This was done to maximize the effect of the final separate-ness of the sources. At station COL, there appeared to be some signal interference at a delay time of about 8 sec and at 11.5 sec. This might be interpreted as the interference of the signal excited at the hypocenter by another
signal which was excited when the fault became near or broke the surface. This effect was not as apparent at station NOR, so we feel that no firm conclusions should be drawn from this particular analysis.

**Corrected Data.** All of the above corrections were applied to the measured amplitude spectra at the WWSSN stations. These spectra are shown in Figure 5.7. The upper part of this figure shows the location and azimuthal distribution of the stations used. The lower part of the figure gives the corrected amplitude spectra at each of the stations. The spectra are arranged by columns with geodetic azimuth increasing from top to bottom. The spectral values indicated by a dashed line in the figure are the values for which the total correction exceeded a factor of 10. These data will have to be considered less reliable than the longer period data because of the uncertainties in the corrections. Samples of the spectral values were chosen at ten discrete periods: 60, 50, 40, 34, 30, 26, 22, 20, 18, and 16 sec. It was felt that this sampling interval was adequate to reflect the spectra shape at the individual stations. Periods shorter than 16 sec were not included into the inversion data set because of the larger uncertainty in the corrections at these periods, particularly in the attenuation factor.
Figure 5.7. Observed spectral amplitudes after all corrections have been applied. Dotted line indicates when the corrections exceed a factor of 10.
Estimated Errors in the Data. Errors to the spectra which we have measured for this event arise from two sources: 1) errors in measuring the spectra at the different stations and 2) errors in applying the corrections to the spectra.

Errors from actually measuring the spectra could arise from digitizing errors and from taking too short a signal to get all the spectral information at certain frequencies. One station, KIP, was redigitized to include a much larger signal and the spectral amplitudes recomputed. The recalculated amplitudes were found to be at most 3-4% different on the average than that originally computed. A value of 3% error was taken as the estimate for this error.

By far, the largest error involved in the application of the spectral corrections arises from the attenuation correction. If we assume that we know the attenuation values to only 25% of the values given in the figure, then the estimated error for this correction would be

\[ E_c(\omega) = e^{\frac{\gamma}{4} \Delta} \]

We next can assume that we know the crustal corrections to only 25%. This turns out to be a small error, except at very short periods, because the original correction is small.
Distribution of Data Information. In Appendix 3, we have derived an operator, $J$, which tells how the information in the data set is distributed. The use of this operator has been demonstrated by Minster et al. (1974), and these authors give a lucid description of its properties. The importance of a particular datum, which is a spectral amplitude at a particular frequency measured at a particular spatial location, depends on both the location and the accuracy of the datum. If we consider each WWSSN station to be a subset of the entire data set, then the relative importance of that station is simply the sum of the importances of the data in that subset. Figure 5.8 shows how the information in the chosen data set is distributed. In this figure, we have summed the relative importances of the frequencies at the individual stations and plotted this importance as a function of the geodetic azimuth of those stations. The results of this plot are somewhat surprising. We see that two stations, COL and KTG, contain a total of 30% of the total information in the data set. In fact, the data set could be reduced by half and only 25% of the total information which constrains the model variables would be lost.

From Appendix 3, we see that the importance of a datum does not depend on the actual value of the datum
Figure 5.8. Distribution of data information with respect to summed information density at a single station. Information density is normalized such that the total from all stations is 100%.
itself but rather the estimated error in the datum and the model response operator, A. The importance values are dependent on the source model only in the sense that A is model dependent. Although not done in this case, this type of calculation could be carried out before the data set is gathered to determine an optimum distribution of stations to use.

5.6 Inversion Results.

Fault Model. The ten model parameters which describe the source system were found by the least-squares fit to the observed spectral data measured at the 18 WWSSN stations. The best fit model parameters as determined from the inversion procedure are as follows:

<table>
<thead>
<tr>
<th>Source #</th>
<th>Depth (km)</th>
<th>Moment (10^{26})dyne-cm</th>
<th>Slip Angle (\lambda_1)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>14.5</td>
<td>0.62</td>
<td>66°</td>
</tr>
<tr>
<td>2</td>
<td>10.5</td>
<td>0.20</td>
<td>78°</td>
</tr>
<tr>
<td>3</td>
<td>6.25</td>
<td>0.51</td>
<td>82°</td>
</tr>
<tr>
<td>4</td>
<td>2.0</td>
<td>0.39</td>
<td>70°</td>
</tr>
</tbody>
</table>

propagation direction \(\lambda_R\) = 99°
propagation velocity = 2.95 km/sec.

Figure 5.9 illustrates the degree of fit to the data that this model exhibits. The dashed straight line in the center of each radiation pattern indicates the orientation
Figure 5.9. Calculated and observed radiation patterns. Orientation of the fault plane (N70°W) is shown by the dashed line in the center of each pattern. The amplitudes of the patterns are in units of cm-sec.
of the strike of the fault system, taken to be in the direction of N70°W. Rupture initiates at a depth of 14.5 km and propagates in the direction of S11°W. The fit to the long period data (40-60 sec) shows that the radiation pattern is a symmetric dipole pattern. These two patterns show very little, if any, signs of fault finiteness. The finiteness of the fault starts to become apparent in the middle two figures. For these periods (26-30 sec), the patterns are symmetric in a direction perpendicular to the direction of propagation but not in the direction of propagation. For the shorter periods (16-20 sec), the effects of the rupture length and non-uniform excitation become very important in shaping the radiation patterns.

**Uncertainty in Model.** We would now like to estimate the uncertainties in these best fit model parameters. As done in Chapter 2 for the stochastic inverse case, we will calculate these uncertainties at a particular confidence level by mapping the estimated errors in the data into errors in the model. The least-squares solution matrix for unweighted model parameters which performs this mapping is the matrix given by Mathews and Walker (1964, p. 366-367)

\[
Q = A^T C_{nn}^{-1} A
\]  

(5.10)

We note here that this solution does not take into account
the possible weighting of the different model parameters. This is no problem in this case, or in any other case, since the problem can be non-dimensionalized as we have done here. The non-dimensional form of this operator has the same form as that given above.

The bilinear product of the matrix, Q, and some model perturbation, q, yields a confidence region of resolvability. That is,

\[ q^* Q q = k^2(c) \]  \hspace{1cm} (5.11)

This equation defines the hyperellipsoid of uncertainty for a particular model perturbation at a confidence interval, k(c).

By fixing k(c) to a particular value, taken to be 1.96 in this problem, we calculate possible model perturbations for which the equation of the ellipsoid is solved. Any model perturbation vector which lies inside this ellipsoid is unresolvable at the confidence interval, k(c), while any model perturbation vector which protrudes this ellipsoid is resolvable at that confidence level. In the example considered here, we will limit the analysis to two dimensions. With this limitation, we can define an ellipse of uncertainty for combinations of model parameter pairs. As we will see in the examples below, this analysis also allows
the consideration of the coupling between the various model parameters.

The ellipse of uncertainty is shown for various fault model parameter pairs in Figure 5.10. In Figure 5.10, the moments are given in units of $10^{26}$ dyne-cm, and the angles are expressed in degrees. In Figure 5.10a, we have considered the area of uncertainty when the moments of the two deepest source double couples are allowed to vary. We see that $M_{o1}$ has an uncertainty of only $0.05 \times 10^{26}$ dyne-cm when it is allowed to vary alone. Perturbations larger than this value would be resolvable by the data unless the magnitude of the moment on the adjacent double couple ($M_{o2}$) was also allowed to vary. The maximum uncertainty that $M_{o1}$ can take in this coupled system is shown by the projection of the maximum excursion of the ellipse onto the $M_{o1}$ axis. The maximum uncertainty of $M_{o2}$ is similarly calculated.

The ratio of the uncertainty measured at the point where the ellipse crosses the axis of a model variable and the maximum uncertainty of that variable defines a coefficient which reflects the coupling of the two variables. We will call this ratio the model co-variance coefficient (MCVC). For a co-variance coefficient equal to 1.0, there is no coupling between parameters. That is, the maximum
Figure 5.10 (a–e). Uncertainty ellipses for various fault model parameter pairs. Combinations of perturbations which fall inside the ellipses are not detectable by the data at the 95% confidence level. Moments are given in units of $10^{26}$ dyne-cm; angles are expressed in degrees.
Figure 5.10 (f-i) Uncertainty ellipses for various fault model parameter pairs. Combinations of perturbations which fall inside the ellipses are not detectable by the data at the 95\% confidence level. Moments are given in units of 10^{26} dyne-cm; angles are expressed in degrees.
uncertainty of one variable is not dependent on the uncertainty of the other variable. For coefficient values less than 1.0, the coupling increases with decreasing values. Figures 5.10a-c show how the coupling between the individual sources decreases with increasing distance between the sources. We see that there is strong coupling between $M_{o1}$ and $M_{o2}$ (MCVC=.30) but very little coupling between $M_{o1}$ and $M_{o4}$ (MCVC=.85). In part a of the figure, we see that the coupling is in a positive sense between $M_{o1}$ and $M_{o2}$, i.e., an increase in one moment is traded-off with an increase in the other. Between $M_{o1}$ and $M_{o3}$, the coupling is in the opposite sense, and between $M_{o1}$ and $M_{o4}$ the coupling that exists is in a positive direction. Figures 5.10d-e show that adjacent slip angles are slightly coupled, but much less so than their respective moment values. The MCVC for the $\lambda_1-\lambda_2$ pair is .75. As the distance increases between sources, the relative coupling between the slip angles of the sources decreases rapidly. Figures 5.10f-i show how the moment and slip angle for the individual sources are related. These results are somewhat surprising in that the slip angle and moment for individual sources are almost completely decoupled. The MCVCs for these parameter pairs range from .90 to .95.

In these figures, we observe that the maximum coupled
uncertainty a particular variable can acquire is constant from comparison to comparison. This uncertainty is the standard error for that variable. Mathews and Walker (1964) show that this standard error is given by

\[ \varepsilon_1 = \sqrt{Q^{-1}_{11}}. \]

Generalizing this expression, we can say that the values of the model parameters are known at a confidence level \( c \) when the errors are expressed as

\[ \varepsilon_1 = \sqrt{Q^{-1}_{11}} k(c). \]  \hspace{1cm} (5.12)

This is the uncertainty which we will assume for the model variables. These uncertainties at the 95% confidence level are given below:

\begin{align*}
\Delta M_{o1} &= 0.15 \times 10^{26} \text{ dyne-cm} \\
\Delta M_{o2} &= 0.21 \times 10^{26} \text{ dyne-cm} \\
\Delta M_{o3} &= 0.13 \times 10^{26} \text{ dyne-cm} \\
\Delta M_{o4} &= 0.11 \times 10^{26} \text{ dyne-cm} \\
\Delta \lambda_1 &= 13^\circ \\
\Delta \lambda_2 &= 18^\circ
\end{align*}
$\Delta \lambda_3 = 10^\circ$

$\Delta \lambda_4 = 9^\circ$

$\Delta \lambda_R = 5^\circ$

$\Delta V_R = 0.2 \text{ km/sec.}$

**Comparison with Static Solution.** We would now like to compare the solution for the fault model of the San Fernando earthquake derived with surface data to that obtained with the static displacements. To do this, we will sum the moments of the individual static fault elements for the elements that fall within a depth range around the location of the double couple sources. We will assume that the actual static moment distribution will be bracketed by the acceptable dislocation models which gave the maximum and minimum moments. The moments for the two models are shown in Figure 5.11. We see here that for the upper 10 km of the fault system there is very good agreement in the static and dynamical solutions. The only significant difference between the two solutions is the moment in the hypocentral region. The surface wave study indicates a moment which is a factor of 2-3 times greater than that predicted by the static data. We recall from the previous chapter that the uncertainties in this area were larger than for the more shallow areas. In the surface wave study, this was not the case. The uncertainty in the hypocentral region was not
Figure 5.11. Comparison of the moment-depth distribution function found from the static study and the surface wave study. Static moments are averaged over the indicated depth ranges. The static moment distribution from the dislocation model shown in Figure 4.5 and 4.10 is shown as a solid and dashed line respectively.
substantially different from the other areas. If we assume that the fault geometry used in the last chapter is correct, then the surface wave study indicates 5.5 m of displacement over an area of 36 km² in the hypocentral region. This compares favorably with that derived by Hanks (1974) of 7.5 m over about the same area. This last calculation was based on still another data set of this event, namely near-field strong motion records.

5.7 Conclusions.

In this chapter, we have seen how theoretical spectral values of surface waves can be calculated numerically for an approximation to a finite source. We have shown that when non-uniform excitation of energy occurs during faulting over some depth-distributed fault surface, the resulting spectral amplitudes can be far different from that of a non-depth-distributed source at the same depth of initiation.

Rayleigh wave data were calculated from a set of 18 WWSSN stations and normalized by correcting for the various source to receiver paths. An inversion scheme was used to find the set of fault model parameters which fit the observed spectral data in a least-squares sense. The best fit solution gave a distributed moment of $1.7 \times 10^{26}$ dyne-cm, and it was found that both the moment and slip angle for this event varied along the down-dip width of the fault.
The slip angle varied in such a way that the displacements became more predominantly dip slip as the dislocation propagated upward from the point of initial rupture at about 3.0 km/sec. Also indicated by the surface wave data is that the propagation proceeded in a direction 5°-10° away from the normal of the strike of the fault. The fault model solution obtained by the surface wave inversion compares very favorably with that given by the inversion of the static displacement data with the exception of the initial dislocation in the hypocentral region where the static data field gives poor resolution.

Wyss (1971) obtains an average moment from the long period surface waves at two European stations (which were not included in this data set) of 1.9 x 10^{26} dyne-cm. Canitez and Toksöz (1972) fit the surface wave spectra at 6 stations in the northeast and southeast geodetic azimuths with a single point source at a depth of 14 km. These authors obtain a moment of 0.75 x 10^{26} dyne-cm and a rake angle of 45°. Obvious reasons for obtaining different values of moment and slip angle by these authors and the present study are 1) the non-depth-distributed source gives rise to a different radiation pattern than a depth-distributed source, and 2) the azimuthal coverage of stations was more limited in the Canitez-Toksöz data set than
in the data set used here. Comparing their data set to that used here, we find that, if the spectra at the stations used in those authors' study were discretized in the same manner as in this chapter, then their data set would have only approximately 40% of the total information which constrains the model variables as was used in this study.

Finally, we feel that the technique of the hybridization of the numerical approximation of a finite fault and the least-squares inversion formalism allows the maximum derivation of information about the source from a set of corrected surface wave observations. Use of the operators associated with the inversion methods provides a powerful tool to determine a priori which data would be most beneficial in determining the source parameters and to estimate the uncertainties in the calculated model parameters and to determine the relative coupling between various parameters. It is hoped that these techniques can be extended to the inversion of other seismic data. The simplest extension would be the inversion of Rayleigh/Love spectral ratios. This type of data would yield far better results than when each is analyzed separately because of the cancellation of error inducing path effects.


Santo, T. A., Division of the Pacific area into seven regions in each of which Rayleigh waves have the same group velocities, Bull. Earthquake Res. Inst., Tokyo Univ., 41, 719-741, 1953.


APPENDICES
Appendix 3

Derivation of Least-Squares Inversion Operators

In this appendix, we will derive the least-squares inverse and the associated operators for three cases. For convenience with respect to manuscript preparation, matrices will be given here in regular typed form, not boldface as in the rest of this thesis. This practice should cause no confusion in this case.

**Case I**: Least-Squares Inverse with no weighting of the model components and no inclusion of the data variances.

We define the forward problem to be

\[ \delta m = \delta d_o \]  \hspace{1cm} (A3.1)

where \( \delta d_o \) is the observed data vector, \( \delta m \) is the true model, and \( A \) is the operator which maps a function from the model space to the data space. We define the vector \( b = \bar{\delta m} \) to be the best estimate of the model, \( \delta m \), that we can obtain.

In this case, we wish to minimize the fit to the data such that

\[ \varepsilon_1^2(b) = \| \delta d_o - \bar{\delta d} \|^2 \]  \hspace{1cm} (A3.2)
is minimized. Here we have $||Z||^2 = Z^*Z$, and $\delta d$ is the calculated fit to the data given by

$$A b = \delta d.$$  \hspace{1cm} (A3.3)

Substituting (A3.3) into (A3.2) and expanding, we get

$$\varepsilon_1^2(b) = \delta d_o \delta d_o - \delta d_o A^* A b - b^* A^* \delta d_o + b^* A^* A b.$$  \hspace{1cm} (A3.4)

Now performing a first order perturbation in $\delta b$ of (A3.4) and setting to zero implies

$$b = \delta m = (A^* A)^+ A^* \delta d_o,$$  \hspace{1cm} (A3.5)

which is the result we are seeking.

**Case II:** Least-Squares Inverse with no weighting of model components and with the inclusion of the data variance.

Here we have the same forward problem as before, however, each datum, $d_{01}$, has associated with it some variance, $\sigma_1^2$. If we assume that the data are unbiased and are statistically independent, then we can write the variances in diagonal form.
We want to minimize the fit to the data in a least-squares sense such that

\[ \epsilon_1^2(b) = \| \delta d_o - \bar{d} \|_{C_{nn}}^2 \tag{A3.6} \]

is minimized. This weighted norm, \( \| \cdot \|_{C_{nn}} \), is defined by

\[ \| z \|_{C_{nn}}^2 = z^* C_{nn}^{-1} z. \]

Substituting (A3.3) into (A3.6) and expanding, we get

\[ \epsilon_1^2(b) = \delta d_o C_{nn}^{-1} \delta d_o - \delta d_o C_{nn}^{-1} A b - (A b)^* C_{nn}^{-1} \delta d_o + (A b)^* C_{nn}^{-1} (A b) \]

Performing the first order perturbation of \( \delta b \) on this last equation and setting the result to zero, we obtain

\[ b = \delta m = (A^* C_{nn}^{-1} A)^+ A^* C_{nn}^{-1} \delta d_o. \tag{A3.7} \]
Case III: Least-Squares Inverse with weighting of the model components and with the inclusion of the data variance.

For some problems, it might be useful to consider a set of non-zero positive weights for the model components. This weighting, or model correlation coefficients, can be defined in the following manner,

\[
C_{ss} = \begin{bmatrix}
W_1 & \cdots & 0 \\
. & W_2 & . \\
. & . & . \\
. & . & . \\
0 & . & \cdots & W_M
\end{bmatrix}
\]

The error induced through the use of this correlation coefficients is given as the following,

\[
\epsilon_2^2(b) = b^T C_{ss}^{-1} b .
\]  \hspace{1cm} (A3.8)

The error of the fit to the data is still given by equation (A3.6). For this case, we want to minimize some quadratic sum of these two errors.

\[
\epsilon^2(b) = \epsilon_1^2(b) \cos \theta + \epsilon_2^2(b) \sin \theta .
\]
This minimization of $\varepsilon^2(b)$ proceeds as before giving

$$b = \delta m = Q^+(\theta) A^* C_{nn}^{-1} \delta d_o \quad (A3.9)$$

where $Q^+(\theta) = (A^* C_{nn}^{-1} A + \tan \theta C_{ss}^{-1})^+$. 

Substitution of (A3.9) into (A3.3) enables us to write

$$\delta d = J \delta d_o.$$ 

$J$ is called the data importance matrix and is given by

$$J = A Q^+(\theta) A^* C_{nn}^{-1}.$$ 

The data importance matrix tells how the information in the data set is distributed (Wiggins, 1972). This operator serves the same role in the least-squares inverse as does the model response operator (equation 2.29) in the stochastic inverse. The data importance matrix gives us information as to how the model components as a whole "see" how large an effect individual data values have on the estimation of a model. Minster et al. (1974) show that the trace of this operator is invariant and given to be the number of independent linear combinations of the
data. For the inversion problem to be determinant, this sum must equal the number of independent model parameters.

In both the inversion procedures discussed in this thesis, we have neglected an obvious operator. We will call this operator the perturbation relaxation operator. These operators are defined below.

\[
\delta d = P_S \delta d_o \quad \text{(Stochastic inverse)}
\]

\[
\delta m = P_{L.S} \delta m \quad \text{(Least-squares inverse)}
\]

where

\[
P_S = A W A^+(W A^+ C_{nn} \tan \theta)^+
\]

\[
P_{L.S} = (A C_{nn}^{-1} A + C_{ss}^{-1} \tan \theta)^+(A C_{nn}^{-1} A)
\]

These operators show simply that in the stochastic inverse case, when \(C_{nn}\) has large values for certain data, these data will be essentially neglected in the fit for the sake of fitting other better determined data. In the least-squares inverse case, when \(C_{ss}\) has a very small value for certain model components, the model will try to fit the data by ignoring the contributions from these particular model components. Also, we see that for \(C_{nn}\) and \(C_{ss}\) fixed,
the point along the trade-off curve influences the perturbation sizes in the iterative inversion scheme.
Appendix 4
Spectral Summation of Two Sources in the Frequency Domain

Consider the addition of two seismic sources by spectral summation. The first source is represented by an amplitude and a phase spectrum of the form $A_1(\omega) e^{i\phi_1(\omega)}$. The phase spectra $\phi_1(\omega)$ contains information about both the source and the propagation path. To this spectra we wish to add an additional source represented by $A_2(\omega) e^{i\phi_2(\omega)}$, where $A_2(\omega)$ is the amplitude spectra of the secondary source, $\phi_2(\omega)$ is the phase spectra of the secondary source, and $\Delta\phi_2(\omega)$ is the additional phase of the secondary source caused by a spatial or temporal offset of the second source from the first source. Propagation effects for traveling a slightly different path from origin to receiver are taken into account with this term. If we let $z$ be the total of the two sources, then we can write,

$$z = A_1(\omega) e^{i\phi_1(\omega)} + A_2(\omega) e^{i[\phi_2(\omega) + \Delta\phi_2(\omega)]}.$$  

(A4.1)
Thus,

\[ Z(\omega) = \left\{ A_1^2(\omega) + A_2^2(\omega) + 2A_1(\omega)A_2(\omega)\cos[\phi_1(\omega) - \phi_2(\omega) - \Delta\phi_2(\omega)] \right\}^{1/2} \]

(A4.2)

and

\[ \psi(\omega) = \tan^{-1} \frac{A_1(\omega)\sin[\phi_1(\omega)] + A_2(\omega)\sin[\phi_2(\omega) + \Delta\phi_2(\omega)]}{A_1(\omega)\cos[\phi_1(\omega)] + A_2(\omega)\cos[\phi_2(\omega) + \Delta\phi_2(\omega)]} \].

(A4.3)

The phase delay of the single original source is given by

\[ \tau_{p_1} = \frac{\phi_1(\omega)}{\omega} \].

(A4.4)

and the new phase delay of the summed signal is

\[ \tau_{p_z} = \frac{\psi(\omega)}{\omega} \].

(A4.5)

The group delay of the summed signal is given by

\[ \tau_{g_z} = \frac{\partial\psi(\omega)}{\partial\omega} \].

(A4.6)

In practice in order to avoid as much as possible the difficulties encountered by the multivaluedness of the
trigonometric functions, we wish to compute only the additional group delay the additional source has over the group delay of the single original source. If we define this additional group delay by \( \delta \tau_g \), then

\[
\delta \tau_g = \tau g_z - \tau g_1,
\]  

(A4.7)

where \( \tau g_1 \) is the group delay of the original signal. By defining \( \frac{\partial \phi_2(\omega)}{\partial \omega} = \tau g_2 \) and \( \frac{\partial \Delta \phi_2(\omega)}{\partial \omega} = \tau g_{2s} \), we have by equation (A4.6)

\[
\tau g_z = \frac{A_1^2(\omega) \tau g_1 + A_2^2(\omega) \frac{\Gamma_+}{2} + \Gamma_+ A_1(\omega) A_2(\omega) \cos(\alpha)}{A_1^2(\omega) + A_2^2(\omega) + 2 A_1(\omega) A_2(\omega) \cos(\alpha)}
\]  

(A4.8)

where

\[
\alpha = \phi_2 + \Delta \phi_2 - \phi_1
\]

\[
\Gamma_+ = \tau g_1 + \tau g_2 + \tau g_{2s}
\]

\[
\Gamma_- = \tau g_2 + \tau g_{2s} - \tau g_1
\]

Substituting this expression into equation (A4.7) we get,

\[
\delta \tau_g = \frac{A_2^2(\omega) \Gamma_- + A_1(\omega) A_2(\omega) \cos(\alpha) \Gamma_-}{A_1^2(\omega) + A_2^2(\omega) + 2 A_1(\omega) A_2(\omega) \cos(\alpha)}
\]  

(A4.9)
The additional phase delay caused by the spatial and temporal offset of the secondary source is given by

\[
\tau_{p2s} = \frac{\Delta \phi_2(\omega)}{\omega} = \frac{\hat{R} - R}{C(\omega)} + T_{d2}.
\] (A4.10)

Here \( R \) is the distance from the hypocenter of the first signal to the observer and \( \hat{R} \) is the distance from the hypocenter of the second signal to the observer. \( C(\omega) \) is the phase velocity of the medium. \( T_{d2} \) is the delay in time after the initiation of the first source before the origin of the second source. This time delay can be expressed as an apparent rupture velocity given by \( T_{d2} = \frac{R_s}{V_R} \), where \( R_s \) is the separation distance for the two events.

Differentiation of \( (\tau_{p2s} \cdot \omega) \) with respect to \( \omega \) yields the secondary group delay, \( \tau_{g2s} \) given by

\[
\tau_{g2s} = \frac{\hat{R} - R}{U(\omega)} + T_{d2},
\] (A4.11)

where \( U(\omega) \) is the group velocity of the medium.

It is seen that calculation of the group delay does not involve any inverse trigonometric functions so that the multivaluedness of the functions is not important. However, this is not the case for the calculation of \( \psi(\omega) \). In practice, the values of \( \psi(\omega) \) are numerically differenced.
to obtain an approximate expression for the group delay. If this approximate group delay is more than one period different from the exact calculated group delay, then a period is added to the phase delay, \( \tau_{pz} \), and the approximate group delay is recalculated. By starting with the phase and group delays of the long period end of the spectrum, the true phase can be unwound from that calculated numerically.

For the addition of many different sources, each offset in time and space with respect to one another, the above calculations are repeated for each source, with the original amplitude and phase spectrum replaced by the partial sum spectra after each individual summation.
Appendix 5

Velocity-Density Models for Local Crustal Corrections

This appendix lists the multi-layered velocity-density models of the crust and upper mantle used in the computation of the local crustal corrections for the WWSSN stations used in the surface wave study in Chapter 5.

The following symbols are employed in the tables:

- **D** = layer thickness in km
- **Alpha** = compressional velocity in km/sec
- **Beta** = shear velocity in km/sec
- **Rho** = density in gram/cm³.

References for the origin of the data used in the models are given below each table.
San Fernando Crust over Gutenberg Continent

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Source: Jungels (1973), Gutenberg (1944), Wood and Richter (1933), Harkrider et al. (1963).
Greenlandic Shield with Ice Cap

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Irish Continental Shelf

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Source: Ewing (1969), Talwani et al. (1965), Ewing et al. (1966), Katz and Ewing (1956), Officer et al. (1952).
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IV. UPPER-MANTLE SHEAR STRUCTURE

Donald V. Heimberger and Gladys R. Engen

Seismological Laboratory
California Institute of Technology
Pasadena, California 91109

Abstract

This is an attempt at constructing a preliminary shear velocity model compatible with both travel times and waveforms of observed seismograms. Transversely polarized LRSM and WWSS observations are used in this study. SV signatures are demonstrated to be inadequate in identifying triplications because of contamination by PI-coupled shear waves and other P-SV interactions. Well located west coast earthquakes, modeled as shear dislocations, are used as sources. The observed SH travel times are considerably slower than the \( L_0 \) values out to 30 degrees. Select profiles of observed waveforms in conjunction with the travel-times provide the data for model determinations. This is accomplished by fitting the observed waveforms with synthetics.

Synthetics are computed for a number of current models as preliminary attempts at fitting the data. Models containing large, sharp transitions predict strong second arrivals beyond 30\(^\circ\). Models containing low velocity zones between 400 and 700 km predict complicated waveshapes between 20 and 25\(^\circ\). None of the above features are apparent in observed SH waveforms. The final model is relatively smooth and exhibits the same characteristics as the HWB P-model, except that the percentage velocity jump is much less at 400 km and slightly larger at 500 km. The 600 km transition is subdued similar to the P-model.

\(^*\) Contribution No. 2460, Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91109.
INTRODUCTION

Considerable effort in recent years has gone into interpreting upper mantle structure in terms of mineralogic and chemical compositions. A fundamental contribution to this subject was given by Anderson [1967] who suggested that the smooth transition over the 413 to 1000 km depth interval described by the Jeffreys model was incorrect and, that instead, the velocity increased rapidly over two regions near depths of 400 and 600 km. He interpreted these transitions in terms of an olivine-to-spinel phase change at 400 km and a spinel-to-post spinel near 600 km.

More recently, McGetchin and Silver [1970] and others demonstrated the likelihood of other coexisting minerals such as garnet and pyroxene. The rather complicated problem of predicting the changes of seismic properties across phase boundaries involving a multi-mineral assemblage has been attempted by Ahrens [1973]. By varying the temperature and the iron content, assuming a uniform distribution with depth, he was able to match a number of detailed P-models; however, the corresponding predicted S-models bear little resemblance to shear models determined by seismic observations (see Figure 1). Of course, this shear profile can be easily altered in a seemingly infinite number of ways, changing the Fe content as a function of depth being an obvious example. It appears that much of this ambiguity can be eliminated by inverting detailed observed profiles of P and S together. For this reason, we think a serious review of existing shear models is in order.

There are essentially two types of seismic observations used in
shear model determinations, namely, eigenperiods from free oscillations and travel-times of body phases. Earth structure from free oscillations has been studied extensively (see for example Jordan and Anderson, 1973, and Dziemowski and Gilbert, 1972). These methods are quite successful in determining the gross structure of the earth but probably lack the resolving power necessary for detailed upper mantle modeling. Shear models based on body phases are relatively rare, some major contributors are Kovach and Robinson [1969], Robinson and Kovach [1971], Ibrahim and Nuttli [1967] and Nuttli [1972].

In the Kovach and Robinson [1969] study shear waveforms are not used except to measure the apparent velocity of the starting pulse across the TFO LRSM array. We repeated many of their measurements and found good agreement, but it should be realized that dt/dΔ values of only the first arrivals does not lead to unique models [for example, see STAN2 of Kovach and Robinson [1969], or Wiggins [1970]. Attempts at measuring the dT/dΔ values for second arrivals were not particularly successful as pointed out by the above authors.

On the other hand, Ibrahim and Nuttli [1967] and Nuttli [1972] used particle motion diagrams to determine the onset of the multi-arrivals produced by the various triplications. The method is successful in identifying signals, however, it appears, as demonstrated later, that many such arrivals are not true upper mantle shear arrivals but are PL-coupled shear waves and other P-SV crustal arrivals. We will avoid these problems by working exclusively with the SH component although this creates considerable observational difficulties.
SYNTHEZISING BODY WAVES

In order to interpret the rather complicated seismograms we must separate the various contributions due to upper mantle structure from other effects such as source complexity. To facilitate this task we calculate synthetic seismograms assuming various types of sources. The problem can be somewhat simplified by computing the SH component and performing the synthesizing in two stages. First, we compute a step function response of a layered model neglecting the radiation field and surface effects [see Helmberger [1973a] and Gilbert and Helmberger [197]]. Secondly, we generate an effective source function which includes the surface interaction for an event at a given depth and vertical radiation field by assuming a particular ray parameter as discussed by Helmberger [1974]. This procedure allows us to store the relatively expensive step function response and test various source configurations by convolution. Note that this procedure is not particularly accurate for cases where the vertical radiation pattern is varying rapidly at ray parameters appropriate for upper mantle triplications. We will attempt to avoid such sources in this study by relying mostly on strike-slip events. The effective source for the strike-slip and dip-slip cases as a function of epicentral depth is given in Figure 2. The synthetics for depth = 6 km, SH6, are appropriate for the Borrego Mountain Event [see Helmberger, 1974]. Synthetic responses for models STAN3, IN, and US26 are given in Figures 3, 4, and 5 using SH6 as the effective source.

The general features of these profiles are controlled by the relative jumps in velocity at the two transition zones. In Figure 4 the response is almost completely dominated by the two generalized rays
describing this interaction. The reflections are indicated by the peaks and the refractions by the preceding ramps. These two responses cross near 24°. In STAN3, the transition zones are much more subdued and, consequently, the reflected pulses are less obvious. The responses cross near 20° and produce a rather strong synthetic, a very important feature as we will see shortly. The US26 model is a reinterpretation of the travel-time data of Ibrahim and Nuttli [1967] to produce a shear model with transition zones compatible with Johnson's [1967] P velocity structure. The low velocity zone starting at 500 km is required to fit the triplication data. The synthetics from this model are similar to those from model IN. Models STAN3 and IN although derived from essentially the same data bank, differ in travel times, triplication points and waveshape. In fact, their predicted waveshapes are remarkably discordant.

OBSERVATIONS

The observations used in this study are the long-period LRSM and WWSS responses to earthquakes and nuclear events. Figures 6 and 7 show the locations of the various stations and events. Most of the data is relevant to Western United States although we included the Arctic event to sample some of the expected lateral variation. Depth coverage is from about 150 to 1500 km. The region sampled is predominantly that used in the P-wave inversion discussed by Helmberger and Wiggins [1971], and Wiggins and Helmberger [1973].

Samples of the WWSS observations used are given in Figures 8, 9 and 10. Some of the data in Figure 9 has been rotated into SH whereas in Figures 8 and 10 the observations are predominantly SH by choice of
station component. Complete seismograms of some sample LRSM observations, showing the SH pulse emerging from the Love wave are displayed in Figures 11 and 12. Before giving a detailed account of the data we will first discuss the arguments against using SV type motions.

Due to a fortunate set of circumstances the observations from the Borrego Mountain earthquake as recorded in southeastern Canada are naturally polarized into SH(NS) and SV(EW) (Figure 13). As expected from knowledge of the radiation pattern of this strike slip event, we find the first downgoing peak of the SV component to be roughly half the amplitude of SH at SCB and slightly less at STJ. There is good agreement between the SH pulses and the synthetics in Figure 2, SH6. The glitch occurring in the 1st peak can be interpreted as the reflection, sS, from the surface [Helmberger, 1974]. Unlike the SH pulses which do not appear to change shape with distances, the SV waveform is elongated and shows a complex behavior as a function of distance, becoming somewhat simpler at $\Delta = 49^\circ$. We interpret this interesting behavior as primarily due to shear-coupled PL waves and not the result of an upper mantle triplication as previous workers have suggested.

Extensive literature exists on the theory of PL waves [see for example Phinney, 1961; Su and Dorman, 1965]. Most PL waves are produced by the coupling between P and SV waves in the crustal waveguide. The type discussed have followed the SV wave, displaying prograde particle motion, and are sometimes called shear-coupled PL waves after Chander et al. [1968]. Particle motion diagrams along with the product trace ($Z^*EW$) used by Ibrahim and Muttli [1967] and Robinson and Kovach [1971] for arrival identifications are displayed in Figure 13. The onset of the PL phase can be determined by the start of the prograde motion which is especially easy to identify on the SCB record. Arrows mark the beginning of PL on the
various traces. Note the change of period as a function of distance. This type of dispersion has been studied by Chander et al. [1968] and can be used to establish the range at which the PL-coupling occurs. Using their figure on phase velocity versus period indicates that for the SCB observation the coupling occurred near 20°. The PL onset at this range is about 20 sec behind the first arrival and falls on the travel time branch associated with the bottom transition zone as defined by Ibrahim and Nuttli [1967], thus the confusion. In fact, a number of workers have pointed out that for some reason the arrival from the bottom transition is most easily identified on the vertical or radial component, for example see Figure 8 of Kovach and Robinson [1969].

There are other difficulties with working with SV waveshapes such as converted P waves arriving before the direct SV [see Kanasewicz et al., 1974]. This feature can be seen in Figure 13 where the S onset is unclear. There is also the problem of waveform distortion caused by receiver function complexity near 20°. That is, when the reciprocal of the SV ray parameter approaches the surface compressional velocity one obtains a complex interaction. The result is a waveshape distortion that is a function of receiver geology and, obviously, difficult to assess. For the above reasons we will rely entirely on SH although this made the data search much more demanding.

Table 1 lists the travel times and events used in this study. The times have been corrected for source depth and instrumental delay. That is, we found that the WWSS instrument has about a two second lag in its transient response compared to three seconds for the LRSM instrument. These values were determined from Figure 2 so they also contain the correction for an assumed Q and source description. The data are plotted in Figure 14. We included the SH data from NTS events as given by Nuttli [1969] and found good agreement.

The data from the Arctic event plot near the Borrego observations at large ranges but plot progressively early moving towards smaller ranges.
Figure 14). This feature suggests shallow lateral variations as might be caused by the absence of a low velocity zone. This conclusion is substantiated by the presence of a very strong Sn or high frequency S for ranges less than 20° for shield observations. The waveforms beyond 20° are dominated by deep structure and appear similar to the other observations.

SHEAR MODEL

Our search technique for finding acceptable models was essentially trial and error with the goals of fitting the travel times of Figure 14 and observed high quality SH waveforms. We were guided in this search by similarities in previously generated synthetic seismograms and observed waveforms. For instance, the synthetic at Δ = 14° for model STAN3 (Figure 3) looks similar to the observations at Δ = 15.6° displayed in Figure 12. From this comparison, one can roughly determine the relative position of the triplications by examining the step function response of S. Proceeding in this fashion we obtained the models given in Figure 15. The corresponding travel time triplications for the preferred model SHR14 are given in Figure 14 for comparison with the first arrivals. The synthetics are displayed in Figures 16 and 17.

Some of the general features of our final model SHR14 are quite similar to those of STAN3. The two major triplications associated with the 400 and 650 km transition zones still exist although considerably subdued. Reflections from these two zones cross at 20°. The small triplication produced by the velocity jump near 500 km in conjunction with the constructive interference caused by the above cross-over produces extremely strong shear waves near 20°. Note the amplitudes near 20° are nearly an order of magnitude greater than at 30°.
Beyond 20°, the synthetics are dominated by the first arrival with some interference near 28° and with little sign of a second arrival beyond 30°.

The major features of the data that dictate this model are most easily discussed with respect to Figure 14, where the major triplication points and branch crossings are labelled. By fitting the observed waveforms we verify the relative positions of these critical points and associated models. We will devote the remainder of this section to such a detailed comparison.

Comparing the synthetics generated in Figure 16 with observations requires a great deal of patch work due to the lack of a complete profile of data. This situation requires using short profiles of observations with particular source functions; one such comparison is given in Figure 18. By fitting these observations we hope to determine the position of branch DB with respect to branch AB. This determination is especially important since the apparent velocities as measured by first arrivals between BE are considerably higher than the branch DB indicates. First, we note that the two LRSM observations in Figure 11 are very much alike, and that they also agree with the synthetics of Figure 16. We assume that our SH6 source function is appropriate and group these two events together at shorter ranges as given in Figure 18. The dashed line in this figure is the same as branch AB in the travel time plot. The scatter in time picks as given in that plot can be easily seen by examining the observations in Figure 18. There are also considerable differences in waveshapes although some of this effect is probably instrumental as in the case of HRAZ. These distortions or waveshape instabilities could be
caused by lateral structure or perhaps vertical radiation at the source. We will be primarily concerned with the first 50 secs of waveform since the latter portion of the recordings are influenced by shallow structure (Love waves, see Figure 12). It is rather difficult to establish the goodness of fit of the synthetics to the observations without a formal comparison but we will characterize the fit by the relative separation between the major positive peaks as a function of distance. A traced overlay is useful for this purpose. The fit is good although the synthetic first arrival at 16° appears small. Part of this disagreement is caused by inadequate ray modeling in that we neglected the vertical radiation pattern which would enhance the first arrival about 10%.

The waveshapes of the WWSS observations are considerably more interesting than the LRSN examples of Figure 18 although they are rather sparse (see Figures 9 and 10). Most of these events are strike-slip as determined in the literature and can be compared directly with the synthetics of Figure 16. However, their source time histories are expected to be somewhat different due to variations in source depth and dislocation history. The match of the synthetic to observed shapes of Figure 10 is particularly convincing. The interference in the first peak of the synthetic near 17° is observed as well as the apparent pulse broadening. It should also be noted that the SH amplitudes increase rapidly approaching 20° as can be seen by comparing the SH pulse directly with the Love wave onset. It is not uncommon to observe the SH wave stronger than the surface waves near 20°. This strong focusing of energy allows relatively small earthquakes, magnitudes less than 5, to be examined at this range,
Strong amplitudes near 20° can also be observed in Nuttil [1969] data (see his Figure 6). The synthetics from model SHR14 display these features very well. This is not the case for the models IN and US26. STAN 3 predicts strong amplitudes near 20° but the shapes are somewhat discordant, especially at 17°.

Another comparison of LRSM observations with synthetics is given in Figure 19. We assumed a dip-slip source DS14 (see Figure 2) since this effective source looked the most like the observed record at 21° and is compatible with the P wave radiation. We explain the elongated signal at 26° as a combination of branches BC and CD.

Beyond 20° it becomes much more difficult to obtain quality data due to the shear coupled PL wave as discussed earlier. We have to rely on source-receiver geometry to resolve SH and SV since rotating does not seem too effective. The latter portion of the BLC waveshape (Figure 9) is dotted because of possible contamination by strong SV motion (PL problem). The front portion of the recording indicating interference appears accurate since this shape is reproduced by many events such as the observed shape at AAM from the Borrego event. The interference at this range is the only indication of the presence of a triplication. This feature is used to fix the branch HI. Faced with the lack of quality waveshapes at these ranges we attempted to explore the domain of possible models that satisfy the travel time data, the 28° interference, and waveshapes beyond 30° (lack of second arrivals). The model ED42 given in Figure 15 is an attempt at including a low velocity zone between 400 and 600 km. The interference at 28° is relatively easy to produce; however, to fit
the travel times requires that the 600 km transition be very large which produces strong second arrivals beyond 30°. One can suppress this feature by making the first arrival stronger (see Figure 17). The increase in velocity occurring at 850 km produces this effect from 28 to 32° but the problem recurs at 35° (see Figure 20). Synthetics for some other models are included for comparison. We don't feel that the data supports much of a second arrival at these ranges (see Figure 8). This model also predicts complicated waveforms from 20 to 24° as does the models US26 and IN. The first arrival is essentially diffracted energy and is thus small compared to the strong arrival returning from the 600 km transition.

On the other hand, most observations in the range 20 to 24° appear quite simple, for example, see MNN in Figure 9. This observation is especially clear since it occurs on a P-SV radiation node. There is little evidence of any motion on all three components other than SH and Love. However, a small first arrival could still be lost in the background noise. To resolve the question we examined relatively strong signals as displayed in Figure 21. The OXF observation of the Borrego event which is well located has a back azimuth of 274°. The trace goes off scale as expected but there does not appear to be a significant precursor. The same is true for the 5 July event as recorded at MNN. This event is also on a P-SV radiation node. A detailed search for this small arrival proved fruitless. For this reason, we conclude that the strong arrival occurring within the first few seconds of onset at these ranges is coming from the 600 km transition and that, therefore, the velocity must increase below 500 km. Any decrease would move this strong arrival back in time as demonstrated in Figure 17.
DISCUSSION

At this point it seems useful to compare the travel times predicted by the proposed models and discuss the various disagreements with respect to inadequate inversion techniques versus problems of lateral variation. The reduced travel times relative to the Jeffreys-Bullen [1958] tables are displayed in Figure 22 assuming a surface focus. The difference between IN and SHR14 in travel times is due to interpretation of seismograms as mentioned earlier. There does not appear to be any significant difference in Sn travel times as given by Nuttli [1969] even though we used different events. We think the reason for this is that he used explosions which are well located and we used mainly well located and timed earthquakes such as Borrego Mountain, San Fernando, and Truckee. The travel times predicted from the STAN3 model are substantially faster. However, the change of slope was the actual measurement and the rapid increases at 20 and 24° are also apparent in SHR14. The difference between 17° and 19° is due to the disagreement in branch crossings at B (see Figure 14). The above differences are mostly disagreements in data handling and probably not caused by lateral variation. On the other hand, the travel time difference between SHR14 and the Jeffreys-Bullen [1958] tables could easily be caused by regional structure. That is, the travel times appropriate for midwestern United States may be anomalous compared to Europe and other continental shield regions as suggested by the results from the Arctic event (Figures 8 and 14). There is also the problem of distinguishing between Sn and S when processing shield observations for ranges less than 20°. This may be the reason why the major transition zone at 400 km was not discovered earlier.
RESULTS

We have attempted to determine a shear model that fits both travel times and waveshapes, essentially by trial and error. Fitting fine structure by this method is extremely difficult and somewhat inefficient. For this reason we have tried to produce a starting model for formal inversion now being developed by exploring the possible domains of solutions available. We think SHR14 is a promising candidate. A slightly smoothed version along with a P-model, essentially L14 proposed by Helmberger and Higgins [1971] that has transition zones in roughly the same locations, is given in Figure 23. We also included a mod1 proposed by Jordan and Anderson [1967] for comparison. The model parameters are given in Table 2.

In summary, we will review the major structural features of our model that predict realistic seismograms fitting the data. However, due to the data distribution we have determined some parts of the model considerably better than others. We will attempt to convey the features which we think are well constrained by this study. First, the structure near the surface is taken from Stewart and Pakiser [1962] who have observed both P_n (8.2) and S_n (4.45) on the same profile of the Gnome explosion. These velocities are in agreement with measurements made in southern California using quarry blasts presently being investigated. The structure of the low velocity zone, LVZ, is taken from Helmberger [1973]. But since the structure of the LVZ is highly variable in western United States [York and Helmberger, 1973], this model can only be considered as representative. Fortunately, the details of the LVZ
do not greatly influence our upper mantle structure except by slight vertical shifting. The velocity below 200 km is fixed by the (dT/dΔ) values, essentially the slope of the branch AB in Figure 14. This branch appears to start at about 13°, in that the first arrival contains higher frequency than the corresponding Love wave with these high phase velocities (see Figure 9). At larger ranges, the AB branch bends downward indicating a positive velocity gradient. This feature suggests relatively strong first arrivals which can be seen in Figure 18. The overall fit of the synthetics to observed waveforms over the range 14 to 19° is quite good which leads us to conclude that the model from 200 to 450 km is reasonably accurate. (We will postpone questions about how accurate to a later date.)

The transition at 500 km is based mostly on amplitude and waveshape data. Since the velocity must increase substantially below 450 km to obey the first arrival times (see Figure 14) we simply added this structure to produce the proper waveshapes and produce the large amplitudes near 20°. At larger ranges the apparent velocity increases rather abruptly (see Figure 14) at about 23°. Similar values were obtained by Kovach and Robinson [1969] as is evident from the slope of the reduced travel time curve displayed in Figure 22. The magnitude of the 600 km transition zone can be determined by the jump in velocity whereas its smoothness can be ascertained by the lack of prominent second arrivals as indicated in Figure 8. At greater depths the velocity approaches B1.

In conclusion, a preliminary model of both compressional and shear velocities is presented. The importance of the seismic parameter,
determined from these velocities, in assessing phase changes versus compositional changes is obvious. We expect the model to be altered somewhat with the addition of wide-band array measurements and better inversion techniques. Nevertheless, a mineralogic interpretation of these combined profiles should prove interesting.

**Acknowledgments**

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REFERENCES


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FIGURE CAPTIONS

Figure 1. Comparison of models proposed from shear wave observations with a predicted shear profile based on theoretical mineralogic considerations.

Figure 2. Effective source function containing surface interaction as a function of epicentral depth; contains time history, instrument and absorption.

Figure 3. Synthetic profile based on STAN3 assuming SH6 as the effective source function.

Figure 4. Synthetic profile based on IN assuming SH6 as the effective source function.

Figure 5. Synthetic profile based on US26 assuming SH6 as the effective source function.

Figure 6. Map locating stations and events. Events are identified in Table 1 according to date (see Table 1).

Figure 7. Map locating an event in the Greenland Sea relative to stations using an Arctic projection.

Figure 8. Profile of observations from events 23 Nov 67 (Arctic) and 9 Apr 68 (Borrego).

Figure 9. Sample recording of WWSS observations of simple events. Note that FBC and BLC are of the Arctic event.

Figure 10. Examples of large amplitude arrivals near 20°.

Figure 11. LRSM observations of two simple events off Oregon as recorded by the same station.
Figure 12. Example of interference caused by upper mantle triplication.
Observations are of one of the events displayed in Figure 11.

Figure 13. Observations of the Borrego event as recorded on three components.
Particle motions as well as a component product Z*EW is included for studying the PL problem.

Figure 14. Reduced travel time plot based on SH data.

Figure 15. Proposed upper mantle models.

Figure 16. Synthetic profile based on SHR14 assuming SH6 as the effective source function.

Figure 17. Synthetic profile based on ED42 assuming SH6 as the effective source function.

Figure 18. Comparison of synthetics with observations assuming SH6 as the effective source function.

Figure 19. Comparison of synthetics with 4 June 64 observations assuming a dip-slip source (DS14).

Figure 20. Comparison of proposed models at 35° displaying the magnitude of the second arrival.

Figure 21. Example seismograms showing the strong onset of arrivals. Arrows mark the arrivals as determined by the authors.

Figure 22. Comparison of residuals of travel times as referenced to the Jeffrey-Bullen tables.

Figure 23. Comparison of SHR14 with a model determined from free-oscillations by Jordan and Anderson.
Fig. 1
Fig. 2
Fig. 8
Fig. 9

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Oct. 1, 1964

RTNM $\Delta = 18.92^\circ$ LOVE

Sep. 24, 1964

1 min.
Oregon Event, Sep. 24, 1964
HRAZ $\Delta = 15.69^\circ$

\[ \text{S I} \quad \text{LOVE} \]

W O A Z $\Delta = 15.64^\circ$

\[ \text{S I} \quad \text{LOVE} \]

1 min.
Fig. 13
Fig. 14
Shear velocity, km/sec

Depth, km

ED42

SHR14

Fig. 15
Fig. 16
Fig. 15
Figure 20
Fig. 21
Shear velocity, km/sec

Compressional velocity, km/sec

Depth, km

Fig. 23
Previous Technical Reports
1969 - Present


