SOURCE CHARACTERIZATION OF THE
OCTOBER 30, 1983 NARMAN-HORASAN EARTHQUAKE

by

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ABSTRACT

Source parameters are determined for the October 30, 1983 Narmar-Horasan earthquake in northeastern Turkey, $M_e = 6.9$, using far-field body wave and near-field synthetic seismograms. A double-couple point source solution is obtained from the inversion of teleseismic P and SH long-period WWSSN and GDSN data. The preferred source mechanism is left-lateral strike-slip with a small thrust component (strike $= 215^\circ$, dip $= 64^\circ$, slip $= 7^\circ$). The centroid depth is estimated at 10 km and seismic moment at $9.0 \times 10^{23}$ dyn-cm. The duration of the source time function is approximately 5 s.

Dimensions of an assumed rectangular fault are determined from the point source solution and aftershock distribution. Fault width is estimated at 13 km and length at 15 km. Based on these values the average displacement on the fault is 1.2 m. It is speculated that source region structure is responsible for the complicated nature and long duration of the body wave records.

Forward modeling of the single three-component strong motion record from a station 25 km south of the epicenter is carried out using a discrete wavenumber calculation. The study reveals that rupture to the southwest, propagating at sub-shear velocity, explains the outstanding characteristics of the processed displacement records. A 1/2 to 1 km thick low velocity surface layer representing sedimentary cover is an essential part of the model.

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Professor of Geophysics
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SECTION I  INTRODUCTION

On the morning of October 30, 1983 at 4:12:27 UTC, a devastating earthquake occurred in the province of Erzurum in northeastern Turkey. According to the USGS, fifty villages experienced extensive damage or were completely destroyed, leaving 25,000 homeless. There were at least 1,342 deaths. The reported magnitudes are $m_b = 6.1$ and $M_s = 6.9$. The epicenter is located between Narman and Horasan at 40.3°N, 42.2°E. The maximum recorded acceleration from the single operating instrument in the area, at an epicentral distance of 25 km, is 0.17 g along a transverse direction.

Field investigations reported intermittent surface faulting over a distance of 12 km, oriented N-E/S-W, and showing left-lateral strike-slip displacement. A Modified Mercalli intensity XIII isoseismal is drawn approximately 5 km by 25 km, running southwest from the epicenter. The maximum intensity zone coincides with the aftershock distribution determined from local data.

The purpose of this study is to investigate the rupture processes of the mainshock using teleseismic and strong motion seismograms. A body wave inversion scheme is used to analyze the abundance of teleseismic data. The seismic source is modeled as a point source, representing a centroid, i.e., "average", parameterization. The strong motion records are forward modeled with a discrete wavenumber calculation of the near-field displacements. The objective is to understand the space-time history of the rupture process.

Body wave inversion theory and application are presented in Section II. The discrete wavenumber technique and strong motion modeling are presented in Section III. Conclusions are given in Section IV. Tables and figures follow the text. Mathematical details are given in the appendices.
SECTION II  BODY WAVE INVERSION

2.1 Introduction

The modeling of teleseismic body wave data for determination of source parameters is motivated by several factors. Geometric ray theory views this seismic radiation as wave packets traveling along paths specified by Snell's law. In the epicentral distance range of $30^\circ$ to $90^\circ$, the rays travel steeply through the crust and upper mantle. These phases are mostly exposed to vertical variation in the near-surface region and are unaffected by upper-mantle triplications. Lateral distance is mainly traversed in the relatively homogeneous mantle without interference from the core-mantle boundary. Ray theory helps surmount the complication of crustal stratification by providing a basis for the calculation of the effects of conversion and reflection at boundaries (Langston and Helmberger, 1975; Bouchon, 1976). Body wave analysis in this distance range offers the further convenience that the wave packets representing P and S phases arrive well separated in time.

Seismic source characterization through the quantitative comparison of observed body wave seismograms with synthetic seismograms calculated for a postulated source model is a well-established technique (Langston and Helmberger, 1975; Kanamori and Stewart, 1976; Bouchon, 1976). With the current availability of high quality digital data and the anticipated improvement of coverage, the utility of automatic inversion schemes is expected to increase. They make possible routine estimation of source parameters, detailed source studies, and regional studies involving many events. A brief well-referenced review of the development of body wave inversion is given by Nábělek (1984).

The inversion algorithm used in this study (Nábělek, 1984) is a versatile implementation of the technique suitable for earthquakes of magnitude 5.0 or greater. Source parameterization is in terms of point double-couple mechanism,
seismic moment, time function, and centroid depth. Estimation of source finiteness is also possible. The inclusion of centroid depth directly in the formal inversion and a source time function parameterization which more closely matches observed spectra are unique to this formulation. The inversion scheme is an interactive iterative process minimizing the mismatch between synthetic and observed seismograms in a least-squares sense.
2.2 THEORY

2.2.1 Forward Problem

*Point Source and Earth Model*

The physical model of the seismic source employed in this study is a displacement discontinuity parallel to a planar surface, presumably caused by brittle fracture, in a linear isotropic homogeneous (LHH) elastic medium. At the frequencies of interest, the source is viewed as a point shear dislocation. The mathematical description of this model is found within a general moment tensor formulation. The solution is an average over the fault surface, representing centroid properties (Backus and Mulchany, 1976; Nábělek, 1984). Since the source is torqueless and without an isotropic pressure component, the moment tensor is a symmetric traceless matrix corresponding to a pure force double couple. In an arbitrary frame, this representation is often decomposed into the sum of three elementary moment tensors, according to the desirability of physical description (Aki and Richards, 1980; Ben Menahem and Singh, 1981) or for computational convenience (Ward, 1980; Nábělek, 1984).

The P- and S-wave displacement fields in an LHH elastic medium due to a double-couple source are directly related to those excited by a point body force \( f(r_o,t) \) represented by

\[
\mathbf{u}^X(r,t) = [\mathbf{G}^X(r | r_o,t) \cdot \mathbf{f}(r_o)] \ast f(t), \tag{1}
\]

where \( \mathbf{G} \) is the elastodynamic Green's dyadic, \( X \) signifies the wave type, P or S, and \( (\ast) \) denotes time convolution. The cartesian components of the displacement field are

\[
\mathbf{u}^X(r,t) = C^X(r,t) \ast f_j(t), \tag{2}
\]

with the source located at the origin. Developments of the Green's dyadic are given by Aki and Richards (1980) and Ben Menahem and Singh (1981).
The displacement field is determined by the Navier equation,

$$\alpha^2 \text{grad div } \mathbf{u} - \beta^2 \text{curl curl } \mathbf{u} + \mathbf{f}/\rho = \frac{\partial^2 \mathbf{u}}{\partial t^2},$$  \hspace{1cm} (3)$$

where $\alpha$ is compressional velocity, $\beta$ is shear velocity, and $\rho$ is density. In expressing the displacement in terms of potentials, the problem reduces to a pair of decoupled wave equations. The solutions are given in Aki and Richards (1980). Following an integration over volume and separation of the medium response from the source term, the Green’s functions are

$$G_{ij}^p = \frac{1}{4\pi \alpha^2 r} \gamma_i \gamma_j \delta(t - \frac{r}{\alpha})$$ \hspace{1cm} (1)$$

and

$$G_{ij}^s = \frac{1}{4\pi \beta^2 r} (\delta_{ij} - \gamma_i \gamma_j) \delta(t - \frac{r}{\beta}),$$

where $\gamma_i$ is the direction cosine $\mathbf{z}_i/r$, $\delta_{ij}$ is the Kronecker delta, and $\delta(t)$ is the Dirac delta function.

The moment tensor component $M_{jk}$ is the limiting case of a body force acting in the direction $\mathbf{e}_j$, separated from the origin by an infinitesimal distance $\varepsilon$ along $\mathbf{e}_k$, such that as $\varepsilon$ approaches 0, the product $\varepsilon f(t)$ approaches $M_{jk}(t)$. The spatial derivative along $\mathbf{e}_k$ of the Green’s function is therefore required. The $i^{th}$ component of displacement due to $M_{jk}$ is then

$$u_i^X(r,t) = G_{ij,k}(r,t) \cdot M_{jk}(t),$$ \hspace{1cm} (5)$$

where

$$G_{ij,k} = \frac{\partial G_{ij}}{\partial x_k}.$$ 

In the far-field, terms which attenuate faster than $1/r$ are neglected, so that spatial differentiation may be replaced with time differentiation,

$$G_{ij,k}^X = \frac{\gamma_k}{c^{X}} G_{ij}^X,$$ \hspace{1cm} (6)$$
where \( c^X \) is wave velocity. The far-field medium response to the source moment tensor component \( M_{jk} \) is

\[
G_{ij,k}^P = \frac{1}{4\pi\rho a^3 r} \gamma_i \gamma_j \gamma_k \delta(t - \frac{r}{a})
\]

and

\[
G_{ij,k}^S = \frac{1}{4\pi\rho \beta^3 r} (\delta_{ij} - \gamma_i \gamma_j) \gamma_k \delta(t - \frac{r}{\beta})
\]

This fundamental result shows the dependence on material constants, radiation pattern, and \( 1/r \) decay. The specification of a double-couple source mechanism, requiring three parameters, determines the moment tensor components. The convention for source specification in terms of strike, dip, and slip is given in Figure 1.

The superposition of modified forms of (7) account for the effects of the earth model, consisting of a homogeneous spherical medium with near-surface radial stratification. The modifications are understood in the context of ray theory. The appropriate rays leaving the source are traced through the source crustal region, multiplying with transmission and reflection at boundaries. At the bottom of the crust, the multiples form the wave packet destined for a particular azimuth and epicentral distance. Appendix A outlines the framework for handling a vertically stratified source region. The homogeneous mantle is modeled with geometrical spreading (Bullen, 1963), anelastic attenuation (Futterman, 1962), and traveltimes. The effect of different material parameters between source and receiver regions and the free surface effect (Bullen, 1963) are taken into account. The displacement is convolved with the instrument response if the synthetics are to be directly compared with the observed seismograms. The computational form of the synthetic seismograms, representing the forward problem, is given in Appendix B. The time domain expression is a weighted and time delayed sum of elementary seismograms.
In the case of a homogeneous earth model, the wave packet consists of the phases P, pP, and sP in the P-SV system, and S and sS in the SH system. The simple expressions representing this case are given in the work of Kanamori and Stewart (1976). Since the individual phases are responsible for the characteristic features of the synthetics, familiarity with their contributions is helpful in modeling. Source region structure, with the exception of extreme crustal models, is a secondary effect.

The inversion scheme requires partial derivatives of the expression for the synthetics. The derivatives with respect to the source parameters are simple functions of the elementary seismograms (Nábělek, 1984). The time derivatives are obtained by finite differences.

**Source Time Function and Extended Source**

The parameterization of the source time function, representing the time derivative of the seismic moment, is in terms of overlapping equal-width isosceles triangles. The expression is

\[ \Omega(t) = \sum_k w_k \, T_{\Delta\tau}(t - \tau_k) , \] (5)

with

\[ \tau_k = (k - 1)\Delta\tau , \]

where \( T_{\Delta\tau}(t - \tau_k) \) denotes a triangular function of width \( 2\Delta\tau \), starting at time \( \tau_k \) (Nábělek, 1984). The relative amplitudes \( w_k \) of the time function elements are parameters in the formal inversion while the number and duration are specified.

The temporal alignment of the synthetics and data records is made outside of the formal inversion. Initial inversions are performed using theoretical arrival times. Large misalignments are corrected early. Later the alignment is adjusted based on the theoretical arrival times and short-period first motions. The final solution is usually preceded by a two step procedure. Convergence of a series of
iterations is followed by realignment of the synthetics with the data records for better cross-correlation. This is repeated until a stable solution is found.

The inversion scheme offers three ways of estimating source finiteness and rupture propagation direction (Nábělek, 1984). The first is to assume that short-period first-motion pulses originate from a common nucleation point. A sinusoidal time shift with respect to azimuth is interpreted as a separation of centroid and nucleation points. The time shift is of the form

\[ \Delta t_0 = \eta \Delta h + p \Delta r \cos(\Delta \varphi), \]  

(9)

where \( p \) is the ray parameter, \( \eta \) is vertical slowness, and \( \Delta h, \Delta r, \Delta \varphi \) are separation in depth, radial position, and azimuth. The separation parameters may be entered into the formal inversion.

The second possibility is the introduction of a simple kinematic source model consisting of a uniformly propagating horizontal rupture. This deforms the original point source radiation patterns and produces a Doppler shift in frequency (Ben Menahem and Singh, 1981). The effect is incorporated with a variation in the duration of trapezoidal time function elements according to the factor

\[ v_\tau p \cos(\varphi - \delta), \]  

(10)

where \( v_\tau \) is rupture velocity along strike, \( \varphi \) is azimuth, and \( \delta \) is strike (Nábělek, 1984).

If appropriate, the event under study may be modeled as a multiple event. Subevents with different source mechanisms are displaced in space and time. The scheme is a more general form of that used for the separation of the nucleation and centroid points. Time delay and spatial separation become inversion parameters.
2.2.2 Inverse Problem

The present purpose is to describe the method by which the optimum source model is determined. The measure of model suitability is provided by quantitative comparison of the synthetic seismograms to the data records. In addition to noise, uncertainties and approximations of the seismic source and earth structure are responsible for the mismatch between the synthetics and data. In describing this discrepancy and using it to adjust model parameters automatically, one formulates an inverse problem.

\textit{Levenberg-Marquardt Algorithm}

A least-squares criterion is used as a measure of mismatch between model response and data. This is motivated by considering data and model parameters as independent random variables with a normal distribution and maximizing their joint probability function, or by viewing the error as a vector in data space and minimizing length. The residuals are

\[ e(m) = f(m) - d, \quad (11) \]

where the response to the model \( m \) is

\[ f_i = f(x_i, m), \quad (12) \]

and the \( n \) observed amplitudes are

\[ d_i = d(x_i), \quad (13) \]

at some point \( x_i \) in space and time.

The optimum model is found with the minimization of

\[ S(m) = \sum_i e_i^2. \quad (14) \]

In the case that \( f(x_i, m) \) is linear in \( m \), \( S(m) \) determines an upward-concave paraboloid in model space, the minimum of which is easily found (Strang, 1984). For nonlinear \( f \), the contours of constant \( S \) are no longer ellipsoids and once situated
on the surface, an iterative method is required to arrive at the minimum. The commonly severe distortion of the surface $S$ (Marquardt, 1963) is responsible for the need to closely check convergence.

Approaches to the minimization of (14) include the methods of Gauss-Newton and of steepest decent. Gauss-Newton approximates the surface $S$ around the current estimate with a linear function so that the correction to model parameters is the normal equation solution of that linear system. With the sensitivity matrix defined as

$$J(m) = [\nabla_m f^T(m)]^T,$$

where $\nabla_m$ is the gradient with respect to model parameters, the model correction is

$$\Delta m = -(J^T J)^{-1} J^T e.$$  \hspace{1cm} (15)

The steepest decent correction takes a heedless plunge in the direction of the negative gradient of the surface $S$,

$$\Delta m = -k J^T e,$$

where $k$ is a constant. The steepest decent correction is more effective high-up on the surface $S$ while the Gauss-Newton correction is better near the minimum. Both methods require careful attention to the control of step size and convergence is often problematic (Marquardt, 1963). The Levenberg-Marquardt (LM) algorithm is an interpolation between these two methods and allows for an estimation of step size.

The simultaneous minimization of the sum of squares of both residuals and parameter increments leads to the LM correction to nonlinear parameters (Levenberg, 1944). The correction

$$\Delta m = -(J^T J + \lambda D)^{-1} J^T e$$

is the least squares solution to the linear system
\[
\begin{bmatrix}
J \\
(\lambda D)^N
\end{bmatrix}
\Delta m = -
\begin{bmatrix}
e \\
0
\end{bmatrix}.
\] (19)

The LM parameter \(\lambda\) is a non-negative number and \(D\) is a diagonal matrix containing the relative weights of the parameter increments. The damping to the model parameter correction is aimed at staying within a neighborhood where the linear approximation to \(S\) is valid. Initially, when \(\lambda\) is large, the steepest descent correction controls the correction. As the minimum is approached, \(\lambda\) is reduced and the Gauss-Newton term takes over. Still, the LM parameter, which represents the relative importance of \(|\Delta m|\) to \(|\Delta e|\) in the minimization, must be specified.

The choice of the LM parameter in the body wave inversion algorithm depends on a step bound determined at each position (Moré, 1978). This calculation is based on a measure of the agreement between the linear approximation and the nonlinear surface, the ratio of the actual reduction to the predicted reduction,

\[
\frac{S(m)^2 - S(m + \Delta m)^2}{S(m)^2 - (S(m) + S(m) \Delta m)^2}.
\] (20)

The LM parameter is determined to keep \(\Delta m\) within the step bound.

\textit{Inversion of Body Wave Data}

Many factors are involved in performing an inversion. Besides the influence exerted by the researcher during the initial data preparation, the final solution is affected by decisions made throughout the inversion process. These include the choice of free parameters, time element number and duration, adjustment of data weighting, temporal alignment of seismograms, and the inclusion of source finiteness effects. Physical plausibility and other sources of information, such as aftershock distribution, observed surface faulting, and intensity isoseismals, help determine the directions pursued. A set of such decisions, together with the least-squares minimization, may be viewed as one iteration toward the final solution. The process is very much an interactive one.
The inversion routine of Nábělek (1984) is a tool for detailed study of seismic sources. The full potential is realized only with both knowledge of seismic source modeling in general and experience with the iterative process. Much of the effort in arriving at a final solution is similar to that of a forward modeling study. Initially, the inversion routine provides clues of possible approaches and later, fine adjustments to the solution.

It should be noted that the program allows for the incorporation of a priori information in the inversion. This is accomplished through a more general stochastic inversion scheme, of which the present scheme is a restricted form. The operational difference is the possibility of adding soft constraints to model parameters. The scheme and philosophical background are discussed by Jackson (1979), Tarantola and Valette (1982), and Nábělek (1984).

2.2.3 Limitations

Contributions to the mismatch between synthetics and data include assumptions and approximations in the mathematical description of the seismic source and earth structure, data quality and completeness, and bias introduced by the researcher during data preparation and inversion. Starting with the simplest model consisting of a single point source of short duration in a half-space, the solution hopefully may be refined with more detailed space-time history and crustal structure. However, for a large complicated event, non-uniqueness may overcome the mathematical method and the researcher's experience to leave a portion of the records unaccounted for.

A close examination of the resolution of source parameters is made by Nábělek (1984). The discussion of the resolvability of moment tensor components is in terms of the medium response functions $T^k$ defined in Appendix A. Of concern in the present case is the determination of the strike-slip component. As seen in the next section, the long-period P-wave data alone has very good azimuthal
coverage. With the addition sampling of the focal sphere by the SH data, no problems are foreseen in this respect.

Nábělek (1984) presents several examples illustrating the effect on the apparent time history of an event due to the point source approximation of a finite fault surface. Also of interest is the trade-off between source depth and source time function.

Possible sources of error in the data include instrument calibration, incomplete knowledge of receiver region crustal structure, and the digitization of analog records. A record of anomalous amplitude may be downweighted in the inversion or simply discarded if data coverage allows. Manual digitization errors are kept to a minimum by a final comparison with the originals. In the case of rotated SH records, comparisons are made between two separate digitizations.

The inversion routine supplies standard errors calculated from the partial derivatives as part of the solution. Though these give some idea of the uncertainty of a particular solution, confidence limits are generally wider. This is due to nonlinearity and adjustments to the calculation made outside of the inversion scheme, such as seismogram alignment. The true uncertainty is based on the several intermediate solutions along the way to the final solution, the standard errors, and the results of tests conducted on synthetic data by Nábělek (1984).
2.3 Application

2.3.1 Earthquake

Information apart from the seismic records has a direct bearing on the following analyses. Aftershock distribution, observed surface faulting, and isoseismal mapping influence the directions pursued during, and the conclusions to be drawn from, both the body wave inversion and the near-field modeling. Also, previous crustal structure studies of the epicentral region are important.

Eastern Turkey is less seismic than the highly active western region. The fault zones and fault plane solutions of previous events of eastern Turkey are shown in Figure 2. The epicentral region is northeast of the intersection of the North Anatolian and East Anatolian fault zones. The area has experienced large shocks in 1924, $m_b = 6.8$, and in 1952, $m_b = 5.8$. The geology is a complex melange of volcanic material overlain by sediment and is little studied.

The isoseismals for the Narman-Horasan earthquake reported by Toksöz et al. (1984) are shown in Figure 3. The outstanding feature is the elongated Modified Mercalli intensity XIII zone running southwest from the USGS epicenter. The zone is between the two largest towns of the area, Narman and Horasan. These experienced the greatest damage.

The location of the most prominent surface breaks (Özgül, 1983) are also shown in Figure 3. Strike ranges from N15°E to N40°E and length from 100 m to 1 km. All show a left-lateral strike-slip component of tens of centimeters to 1 m. A large dip-slip component, the eastern block displaced downward, was observed for some segments. It is conjectured that the complex near-surface geology is responsible for the segmented surface manifestation of dislocation at depth (Toksöz et al., 1984). The deformed melange overlain by sediment could have damped rupture propagation to the surface. The breaks are generally aligned parallel and along the eastern boundary of the maximum intensity zone.
The distribution of aftershock epicenters determined from local data for the 3-week period after the mainshock are shown in Figure 4. The distribution is coincident with the maximum intensity zone. The distribution of aftershock depth is given by Toksöz et al. (1984). The estimated depths extend to 20 km.

Crustal structure studies of the area include those of Chen et al. (1980), Canitez and Toksöz (1980), and Kenar and Toksöz (1980). The first study uses travel-time residuals to estimate the uppermost mantle P-wave velocity at $7.73 \pm 0.08$ km/s. The second study confirms this result for eastern Turkey. Also, an average crustal thickness of 40 to 45 km is obtained from surface wave and Bouguer anomaly modeling. The final study models the crustal S-wave velocity structure by inverting surface wave data. The crustal thickness is estimated at 40 km and the shear velocity of the underlying mantle is close to the Poisson solid counterpart of the P-wave result from the other studies.

2.3.2 Data and Processing

Data

The data used in the body wave inversion includes analog records from the Worldwide Standardized Seismograph Network (WWSSN) and digital seismograms from the Global Digital Seismograph Network (GDSN). The WWSSN records are from long- and short-period instruments with peak responses at 15 and 1 s. The GDSN data are distributed in long-, intermediate-, and short-period bands, representing responses centered at 28, 10, and 1 s. The subnetwork to which a specific station belongs, Seismic Research Observatories (SRO), Abbreviated SRO (ASRO), Regional Seismic Test Network (RSTN), or Digital WWSSN (DWWSSN), determines which bands are available.

The long-period analog and digital data, representing the most useful frequency bands for determining centroid parameters of large events, is abundant in
the epicentral distance range of interest. The signal-to-noise ratio is almost universally high. Samples of Z-component GDSN seismograms are given in Figure 5.

Processing

The WWSSN analog records are put into the same digital format as the GDSN data by hand-digitization and equal-space interpolation. The P-wave data set consists of Z-component seismograms. For the SH data set, transverse components are retrieved by rotation of the N-S and E-W records. The seismograms are normalized with respect to magnification and epicentral distance in the inversion routine.

The GDSN short-period data is deconvolved to remove the instrument response and applied digital filtering. The deconvolution broadens the bandwidth to include periods greater than 1 s. The longer periods add stability to the inversion. The low-frequency content is limited by a high-pass filter in order to exclude noise amplified by deconvolution.

Records are chosen from stations which fall within the allowable epicentral distance ranges. Distances between $30^\circ$ and $90^\circ$ are permitted for P-wave records. The arrival of the ScS phase reduces the furthest distance of the SH range to $75^\circ$. The maximum number of seismograms per inversion is 25.

The time-sampling rate of the long-period digital data is 1 point per second. The analog records are digitized at a rate of 2 points per second. Although the frequency content of the data is adequate at the lower sampling rate, the precision of temporal alignment of the seismograms is limited by the sampling interval. The short- and intermediate-period digital data are decimated to 4 or 5 points per second.

The assignment of relative weights to seismograms allows for the adjustment of the influence of various subsets of the data. If a subset, whether determined by
phase, instrument type, of frequency band, has adequate azimuthal coverage, then the power of that subset reflects its importance. Weighting factors are based on the power and number of stations in the subsets. A cluster of n seismograms in azimuth and distance is down-weighted by a factor of \(1/\sqrt{n}\). Individual records may also be down-weighted according to quality.

**Short-Period Records**

Short-period arrival times are important for proper seismogram alignment. The finite fault effects which may be included in the inversion scheme are directly dependent on the accuracy with which the short-period records are read. For the Narman-Horasan earthquake, the short-period P-wave arrivals, both analog and digital, are emergent in character. Many arrivals are picked with estimated errors of ±1 s, while others are much less certain.

2.3.3 Inversions

The analysis of the long-period data from the Narman-Horasan earthquake includes separate inversions of the analog and digital seismograms. The purpose of inverting data from single frequency bands is to search for evidence of source multiplicity. Coherent energy might be more prevalent in one frequency band than another. The search is motivated by the presence of significant energy beyond the first few cycles after the P and S arrivals. The separate inversions are followed by the inversion of a combined data set. The mixing of data from different instruments adds stability to the solution.

The crustal structure at the source and receivers is modeled as a half-space. The material constants assumed for both regions are \(\alpha = 6.14 \text{ km/s}, \beta = 3.54 \text{ km/s}, \) and \(\rho = 2.75 \text{ g/cm}^3\). The anelastic attenuation factors assumed for the mantle are \(t^* = 1 \text{ s for P-waves and } t^* = 4 \text{ s for S-waves.} \)
In the inversions of the separate WWSSN and GDSN long-period data sets, a short duration time function was used to obtain a preliminary solution. A long duration source time function was then employed with only the total moment and relative strengths of the individual time elements allowed to vary. For both the WWSSN and GDSN data sets, no significant contribution from later time elements appeared upon inversion. A positive result would indicate coherent energy farther back in the waveforms, justifying the use of an extended time function or a multiple event representation. The time window was reduced and the effects of seismogram alignment and source time function element number and duration were investigated. The final WWSSN and GDSN solutions were very similar. A final search for coherent energy later in the waveforms was unsuccessful in both cases.

The inversion of the combined WWSSN and GDSN long-period data set provides the final point source solution. Weighting factors compensated for differences in instrument and wave type. Information about the data set is found in Table 1 and station coverage is shown in Figure 6. Since the higher frequency analog records provide a more precise estimation of proper alignment between synthetics and data records, these guided the manual adjustment of the alignment. The minimum resolvable duration for the source time function elements was 2 s. Intermediate solutions did not show any notable variation in the parameters. The final solution is close to the solutions obtained from the separate WWSSN and GDSN data sets. A final search for evidence of source multiplicity was unsuccessful. The source parameters, standard errors, and estimated uncertainties are given in Table 2. The synthetic and observed seismograms are shown in Figures 7 and 8.

The effect of a propagating line source was examined using the short- and intermediate-period GDSN data. The short-period records were deconvolved and high-pass filtered at 10 s. A best point source alignment was determined using the long-period solution. The effect of a unilateral propagating line source with a
rupture velocity of 3 km/s, for both directions along strike, was then examined. The moment was the only free parameter in the subsequent inversions. Start times were first held at the best point source alignment and later allowed to vary. The results were inconclusive as to which of the two directions of rupture propagation is more suitable. The estimated moments of the two models were within a few percent of each other and the difference in total rms error was small.
2.4 Discussion

The solution obtained from the long-period body wave inversion provides a good fit to the first cycle of the data records. The standard errors and estimated uncertainties are small, reflecting the high stability of the solution throughout the several inversions. The emergent character of the short-period records do not allow an estimation of nucleation and centroid separation. Also, the short duration of the source time function prevents the introduction of a propagating line source from having much influence. Prominent finite fault effects though are not expected for only a moderately large event.

The point source solution makes possible an estimation of source dimensions. Since the 6 s duration of the source time function includes 2 s of slow initial rise, the duration of rupture contributing significantly to the total moment is taken to be 5 s. Assuming a rupture velocity close to shear velocity, this corresponds to a rectangular fault length of approximately 15 km. This is consistent with the observed surface faulting and the maximum intensity isoseismal. The width is estimated at roughly 13 km. This is based on the centroid depth and on the assumptions that the fault plane does not break the surface and that it is within the aftershock depth distribution given by Toksöz et al. (1984). The seismic moment obtained from inversion corresponds to an average displacement of 1.2 m over the 13 km × 15 km fault plane. The stress drop is estimated at 27 bar.

The energy present farther back in the data records was found incoherent with respect to the attempted introduction of source multiplicity. Seismograms from both the mainshock and the large aftershocks, Figure 9, exhibit the same characteristic late arrivals. This suggests that the complicated nature of the waveforms is due to source region structure. The effect of simple crustal models on the synthetics was investigated. The source parameters were held constant while the region structure was varied. A discontinuity at 40 km depth, representing the crust-mantle boundary, produced oscillations similar to those seen in the
data records immediately after the P and S arrivals. However, the introduction of much deeper structure in the source region would be required to match the energy appearing 30 to 60 s after the P and S arrivals on the data records.
SECTION III STRONG MOTION ANALYSIS

3.1 Introduction

Modeling of strong ground motion records reveals more detailed information about the faulting process. Usually centroid parameters obtained from teleseismic body or surface wave studies aid in the near-field analysis. Detailed studies of rupture history in space and time such as those of Hartzell and Helmberger (1982), Olson (1982), and Bouchon (1982), are of events for which ample strong motion data is available.

At the time of the Narman-Horasan earthquake, only one strong motion station was operating in the vicinity, 25 km to the south of the epicenter. The main objective of this near-field study is to determine the space-time history of the rupture process from the 3-component accelerogram. This is especially desired since the teleseismic data was found insensitive to finite fault effects.

Waveforms representing ground displacement obtained by integrating the acceleration records are forward modeled following the discrete wavenumber approach of Bouchon (1981). Though an asymptotic method based on the use of a high-frequency Green's function was considered (Bernard and Madariaga, 1984; Spudich and Frazer, 1984), the apparent surface wave contribution to the integrated records precluded this approach. The discrete wavenumber routine of Bouchon (1981) calculates the displacement field for a double-couple point source in a layered medium. A brief description of the underlying theory precedes the application.
3.2 Discrete Wavenumber Technique

Theory

The discrete wavenumber technique (Bouchon, 1981) facilitates calculation of the frequency domain impulse response for a double-couple point source in an unbounded LiK elastic medium. The essential feature is the replacement of the single source with a periodic array of sources. This leads to an exact discretization of the integration over wavenumber in the expression for the displacement potential. With a restriction in space and time, the array potential is equivalent to the single-source potential.

The one-dimensional wave equation for the compressional displacement potential field excited by a sinusoidal point source at the origin is

$$\frac{\partial^2 \varphi}{\partial t^2} - \alpha^2 \nabla^2 \varphi = 4\pi \alpha^2 \delta(x) e^{i\omega t}.\quad (21)$$

Potential theory offers the convenience of considering only this equation and its homogeneous counterpart to obtain the solution with which the general displacement field due to a point force may be constructed. In a cylindrical coordinate system, the inhomogeneous solution is

$$\varphi_0(r,z;\omega) = -ie^{i\omega t} \int_0^{\infty} \frac{k}{\nu} j_0(kr) e^{-i\nu z} dk,\quad (22)$$

where

$$\nu = \sqrt{k_a^2 - k^2}, \quad h_n(\nu) < 0, \quad k_a = \omega/\alpha.$$ 

The integral over wavenumber is the Sommerfeld integral. The homogeneous solution, of similar form though azimuth dependent, is obtained for a source located away from the origin. This is integrated over azimuth to represent the radiated displacement potential $\varphi_c(r,z;L;\omega)$ for a ring source of diameter L. The potential for a radially periodic arrangement of such sources with separation interval L is added to the point source potential at the origin.
\[ \varphi_a = \varphi_0 + \sum_{m=1}^{\infty} \varphi_c (r, z; mL; \omega), \tag{23} \]

to arrive at the potential for the source array with which the original point source is replaced.

The source substitution introduces a periodicity in the wave field which allows replacement of the wavenumber integral with an infinite sum. This exact discretization is accomplished using a delta function representation of the integrand. Applying the sifting property of the delta function, the array potential becomes

\[ \varphi_a (r, z; \omega) = -i \pi \sum_{n=0}^{\infty} \epsilon_n \frac{k_n}{\nu_n} J_0 (k_n r) e^{-i \nu_n |z|}, \tag{24} \]

where

\[ k_n = 2n \pi / L, \quad \nu_n = \sqrt{k_n^2 - k_n^2}, \quad \text{Im} (\nu_n) < 0, \]

and \( \epsilon_n \) is Neumann's factor.

Considering distance \( r \) and time \( t \) such that arrivals from the ring sources are excluded from the time window,

\[ \tau < \frac{L}{2} \quad \text{and} \quad \sqrt{(L - r)^2 + z^2} > at, \tag{25} \]

it is seen that the potential for the source array must be the same as that for the single source. So restricted, the discretized cylindrical representation for the potential (24) is equivalent to the solution of (21) in spherical coordinates,

\[ \varphi (R; \omega) = \frac{e^{i \omega (t - R/a)}}{R}. \tag{26} \]

With this identification, the displacement field due to a single force is found with substitution into the result of Lamb (1904). The general moment tensor solution follows by taking spatial derivatives. The use of propagator matrices extends the solution to layered media. Fourier transformation into the time domain produces the synthetic seismograms.
Model Specification

In the discrete wavenumber routine of Bouchon (1981), model specification includes source-receiver geometry, double-couple source mechanism, source function rise time, layered crustal model, time window and sampling rate, and source array spacing. The source time function, representing the slip time history, is

\[ s(t) = \frac{1}{2} (1 + \tanh \frac{2t}{t_0}) , \]  

(27)

where \( t_0 \) is the rise time. Each of the crustal structure layers is described with \( \alpha \), \( \beta \), \( \rho \), and P- and S-wave Q factors, \( q_p \) and \( q_s \). The attenuation model is that given by Aki and Richards (1960).

The time interval between calculated points of the synthetics is determined by the highest frequency that is to be represented. The time window used throughout this study, \( t_w = 24 \) s, is first sampled with 256 points and later with 128 points. The effect of aliasing due to the discrete Fourier transform is reduced by the use of complex frequency (Bouchon, 1979). The imaginary component, \( \omega_i = \pi / t_w \) s\(^{-1}\), exponentially attenuates contributions outside of the specified time window. The sum over wavenumber is truncated when the last calculated contributions to all three displacement components are less than .001 times their current sums.
3.3 Application

3.3.1 Data

The single accelerogram for this event is from the strong motion station in Horasan. The instrument is a Kinematics SMA-1 accelerograph, with $\omega_0 = 25$ Hz and damping at 60% critical. According to Yilmaz (1984), it is installed in a small one story structure showing small cracks along the walls due to the Narman-Horasan earthquake.

Corrected records have been available from two sources. The first set (Toksoz et al., 1984), Figure 10, contains corrected acceleration, velocity, and displacement for the three components, N-S, E-W, and Z. A description of the processing is not included. The method of transducer and linear trend correction, filtering, and integration is assumed to be similar to those given by Hudson (1979). The second set (Yilmaz, 1984) is from a study based on spectral analysis aimed at engineering application. In this case the processing method is described. Noteworthy is the band-pass filtering between periods of .02 s and 20 s. A comparison of the displacement records is made in Figure 11. Since the processing of accelerograms can be problematic (Hudson, 1979), it is promising that the most characteristic traits of the two sets of corrected records are essentially the same. In general, the records are complex and of long duration.

The absolute time is unknown but it is assumed that the instrument triggered on the first P arrival. The high amplitude signal seen on both transverse acceleration records at $\sim 4$ s is identified as the S arrival. Using average crustal velocities, the travelt ime difference between P and S corresponds to a nucleation point distance of $\sim 30$ km. This is consistent with the USGS epicenter location.
3.3.2 Modeling

**Point Source**

Successful near-field modeling should explain certain outstanding characteristics of the records. These include the oscillation of 3 s period in the E-W component, the prolonged oscillation in both transverse components, and the relatively low amplitude of the vertical component. Also, the absolute amplitudes of the records should be matched.

The source mechanism used throughout this study is that obtained from the body wave inversion. The epicenter-receiver geometry and coordinate system used are shown in Figure 12a. The diagram of Figure 12b and Table 3 define the various crustal models. The layer thickness is given later for the particular calculation.

The point source depth is held constant at 5 km. Since the centroid depth was estimated at 10 km, these point source synthetics represent energy from the shallow part of a finite fault contributing to the total moment. It is this region that will more effectively excite surface waves. The prominent low-frequency oscillation seen on all records motivates the basic assumption that they dominate the near-field record.

Waveforms are synthesized for different source positions along strike. The radial, transverse, and vertical components relative to the epicenter-receiver separation are calculated. Later in the study, the horizontal components are rotated into N-S and E-W pairs so that the synthetics may be compared directly with the data records. Beside the low-pass filtering imposed by the time sampling rate, the synthetics are not filtered. Any DC level that they may exhibit will not be present in the band-pass filtered data records. The P-wave radiation pattern at the receiver is at a minimum when the source is located near position $z = 0$, rising to a maximum near position $z = 16.5$. The S-wave radiation pattern at the
receiver is at a maximum when the source is located near position $x = 0$, going through a node near position $x = 16.5$.

Note that the start time of all synthetics in this study is at $t = 2$ s. The amplitude scale is in terms of centimeters of displacement per source moment of $1 \times 10^{25}$ dyn-cm. This amplitude scale is used in all the following figures except where the scale is specified. The tails seen on some waveforms result from the exponential amplification of numerical noise. The amplification occurs during the correction in the time domain for the attenuation of energy outside of the time window in the frequency domain. Though the problem recurs later in the study, the results are not effected.

The first set of waveforms is generated for point sources at angles corresponding to three positions along strike, covering the maximum intensity zone, with three different crustal structures. The crustal models include a half-space, a 15 km layer over a higher velocity half-space, and a 500 m low velocity layer over a half-space. All models are without attenuation. The source time function is essentially a step, with $t_0 = 0.5$ s. The epicenter-receiver separation is 28 km. The synthetics are shown in Figure 13.

The half-space calculation clearly shows the P, SH, and Rayleigh arrivals. The noncausal tail of the P wave is visible before the calculated arrival time at 5.2 s. The second downward arrival on the radial component is the intermediate P arrival as defined by equation 4.32 of Aki and Richards (1980). The SV arrival appears as a glitch followed by the large amplitude Rayleigh phase in the vertical component.

The model with the layer boundary at 15 km depth is not very informative. Besides the appearance of a few reflected phases, the synthetics are essentially unchanged. The boundary is too deep to effect the excitation of surface waves with the close source-receiver configuration.
The low velocity layer over a half-space produces interesting results. Prolonged high amplitude reverberation follows both the SH and Rayleigh arrivals. The frequency of the oscillation is low enough so that the synthetics resemble the data records. The Rayleigh wave traveltime is slightly greater due to the lower average medium velocity. It is promising at this early stage that using the moment obtained from the body wave inversion, the amplitudes are close to those of the data records. For $M_o=8 \times 10^{25}$ dyn-cm, the amplitudes of the transverse components, excluding the sharp SH arrivals, are in the range of 10 to 20 cm.

To further investigate the effect of a thin low velocity surface layer, waveforms are generated for models with varying layer thickness and material constants. Also, two models are recalculated with attenuation. The synthetics, Figure 14, clearly show the dependence of the reverberation period on layer thickness and velocity, i.e., the traveltime between boundaries. The low velocity layer acts as a waveguide for both SH and Rayleigh radiation. The material attenuation smooths the waveforms and has greater effect on higher frequency oscillation.

The effect of layer thickness is examined in more detail. The average sedimentary layer properties chosen are $\alpha = 2.0$ km/s, $\beta = 1.15$ km/s, and $\rho = 1.7$ cm$^2$/gm$^3$. The synthetics from three source positions along strike for a range of layer thickness from 200 m to 1200 m are shown in Figures 15-18. Attenuation is not included in this calculation. The smoothing effect of a long rise time, $t_r = 1.0$ s, and the rotated N-S and E-W components are shown for source position $x = 0$. The crustal models with overlying layers of 600 m and 800 m produce synthetics with reverberation of 2 to 3 s period on both horizontal components. Also, the vertical component is of relatively low amplitude. These results encourage a finite fault calculation.
Extended Source

In order to model rupture over a finite fault surface, waveforms are generated at intervals along the fault. These are delayed and summed to represent rupture velocity and direction. Since the depth is constant, the model is that of a propagating line source. The source time function duration is chosen to be approximately the same as the rupture propagation time between point sources. This smooths the contributions to the discrete sum so that the result is more representative of a continuous rupture.

In the present case, a point source separation of 2.5 km is used. The synthetics will not vary drastically from one position to the next. The corresponding rise time is \(~1.0\) s since the model rupture velocities will be close to the shear velocity of the half-space. Waveforms are generated for source positions between $z = 2.5$ to $z = 20$ for two surface layer thicknesses, 800 m and 800 m, without attenuation. The synthetics from the model with layer thickness of 800 m, Figures 19 and 20, exhibit more pronounced reverberation than those for the model with layer thickness of 600 m.

Simple inspection of the synthetics provides evidence for a southward propagating rupture. The delay due to rupture propagation time will add to the decreasing source-receiver traveltime for positions further south. The oscillatory character of the single-source waveforms will be preserved in the stacking of their contributions. Summations representing northward propagation will smear the synthetics, leaving the large amplitude SH arrivals as the only distinctive features. These observations are born out in the summations shown in Figure 21.

The waveforms generated from the model with a 800 m thick surface layer stack more constructively than those from the model with layer thickness 600 m at a given rupture velocity and duration. The longer period of the former set is less sensitive to mismatch in the arrival time of significant energy from individual sources. The oscillatory character of the summations is enhanced for higher
rupture velocities since the source-receiver traveltime difference between neighboring sources is less than the rupture propagation time difference. Since super-shear rupture velocities are a rare exception in modeling studies, and with consideration of the 5 s duration of the source time function obtained from the body wave inversion, the model with sedimentary layer thickness of 800 m, with point source contributions from positions $x = 20$ to $7.5$ and $v_r = 3.1 \text{ km/s} = .98 \beta$, is preferred. This model represents a rupture duration of 4.8 s and a length of 15 km. The reverberation amplitude of the transverse components for $M_0 = 8.0 \times 10^{25}$ dyn-cm are approximately half of those of the data records. The better match in amplitude of the synthetics from individual point sources is lost with the imperfect stacking of their contributions.

Since the USGS epicenter is located just east of position $x = 20$, that location is appropriate for the site of rupture initiation. Still, further investigation may constrain fault position. To this end, a new set of synthetics for the 800 m thick upper layer model is generated which includes positions further along strike in both directions. Attenuation is included in this calculation, with $q_p = q_s = 20$ for the upper layer and $q_p = q_s = 100$ for the half-space. The synthetics are shown in Figure 22.

Summations representing southward rupture, $v_r = 3.1 \text{ km/s} = .98 \beta$, starting with point source contributions from positions $x = 30$ through $x = 10$, are shown in Figure 23. The duration is 4.8 s and the length is 15 km. The general character of the seismograms does not vary much with fault segment position. The onset of the prominent reverberation on the E-W component reflects the change in source-receiver separation. The decreasing Z-component amplitude for positions toward the south is due to the node of the Rayleigh radiation pattern. The synthetics resulting from the summation of point source contributions from positions $x = 20$ to $x = 7.5$, which are the same as the preferred set of figure 21 though now with slightly reduced amplitude due to attenuation, are band-pass filtered
between .05 s and 20 s and displayed with the data in Figure 24.

In connection with the estimation of source strength, the assumption that a shallower source will more effectively excite reverberation is examined. Synthetics for point sources at position $x = 10$ and varying depth are shown in Figure 25. Reverberation is drastically reduced with increasing source depth. This is reassuring since energy radiated from distances in the down-dip direction could interfere destructively with that from the shallower region. The low amplitude level of the synthetics seen in Figure 24 is reached with the entire seismic source strength obtained from the body wave inversion lumped at 5 km depth. The overestimation of seismic moment in the present modeling scheme is therefore actually greater.

A bilateral rupture is represented by adding the synthetics for fault segments rupturing to the north and to the south from a common initiation point, Figure 26. Since northward rupture does not preserve the oscillatory character of the point source synthetics, that segment will not contribute to the reverberation amplitude of the synthetics representing bilateral rupture. However, the reverberation amplitude for a given total moment is decreased by the ratio of the strength assigned to the segments of northward and southward rupture. The effect is to increase the mismatch in the amplitudes between the synthetics and data records.
3.4 Discussion

The development of the near-field study depends on the initial result that a point source under a low velocity layer with reasonable layer thickness and material constants will generate waveforms that exhibit the reverberation which characterizes the data records. This crustal model is the simplest which reflects the known aspect of the epicentral region structure relevant to the study, the sedimentary surface layer. Since the specific crustal model chosen produces the desired effect due to its waveguide characteristics, a trade-off exists between layer velocity and thickness. Lower layer velocities though will result in an increase in displacement amplitudes. The study therefore provides only a rough estimate of sedimentary surface layer thickness. The estimated range is from $\frac{1}{2}$ to 1 km. The introduction of rather high attenuation produced only a slight decrease in amplitude.

It is surprising that although the crustal structure dominates the character of the data records, information about the source is still recoverable. A southward propagating rupture model is necessary to preserve the oscillatory character of the individual point source synthetics. Within this one-dimensional approximation of a fault surface, the match of the synthetic and data amplitudes improves with increasing rupture velocity beyond the near velocity of the half-space. The synthetics generated by the preferred model resemble the data records well in character and reproduce the relative amplitudes of the components if not the absolute amplitudes.

The estimate of source duration obtained from the body wave inversion and the corresponding fault length is maintained in the near-field study. Also, the position of the fault suggested by the epicenter location is used in the final model. With consideration of the observed surface faulting and the maximum intensity isoseismal, a unilateral southward propagating rupture presents a consistent picture. Although the mismatch between synthetic and observed amplitudes is
increased, the near-field modeling scheme is basically transparent to the effects of a bilateral rupture.

An important aspect in modeling the data records is the temporal alignment of the synthetic seismograms. The less than 2 s lag of the instrument trigger time with respect to the start time of the synthetics in Figure 24 would seem inconsistent with the epicenter-receiver separation. However, with the introduction of a planar rupture surface, Figure 27, this inconsistency is removed. With the assumption that nucleation occurs at depth, delay due to rupture propagation to the shallower region shows the alignment of the synthetics and data records in Figure 24 to be reasonable. Also, the curved rupture fronts would decrease the time delay between the point source contributions for a given rupture velocity. As seen by the synthetics of Figure 21 representing super-shear rupture velocity in the line source model, the effect would be to dramatically increase the amplitudes. Therefore, taking into account the second dimension of the fault surface may explain both the temporal alignment and the amplitude mismatch of the synthetics and data records.
SECTION IV CONCLUSIONS

Source characterization of the October 30, 1983 Narman-Horasan earthquake has been carried out using far-field body wave and near-field synthetic seismograms. The inversion of teleseismic P and SH long-period data provides the best point source solution: strike = 215°, dip = 64°, slip = 7°, depth = 10 km, scalar moment = 8.0 x10^{28} dyn-cm, and duration = 5 s. These parameters are very well constrained. The point source solution and aftershock distribution make possible an estimation of rectangular fault dimensions and accompanying parameters: width = 13 km, length = 15 km, displacement = 1.2 m, and stress drop = 27 bar. These values are much less certain.

The body wave inversion results were used in the forward modeling of the strong motion record. The study reveals that a unilateral rupture propagating to the southwest from the epicenter explains the outstanding characteristics of the processed displacement records. The model requires a low velocity surface layer, representing sedimentary cover. The estimated layer thickness is between ½ and 1 km. These near-field results do not call for adjustment of the body wave analysis. The thin surface layer is negligible. The effect of a propagating line source was previously found to be inconsequential in modeling intermediate-period body wave data. The final results of the combined study with the estimated uncertainties are given in Table 4.

Further progress in modeling the teleseismic records could be in the form of a source region study, including an analysis of body wave records from the large aftershocks. Another large event in the region would certainly provide an impetus for such an effort. The strong motion record might yield further information using a two dimensional fault model, as already suggested. Perhaps the constraints on rupture velocity, sedimentary layer thickness, and source dimensions could be strengthened.
Table 1  Parameters of Stations Used in Long Period Body Wave Inversion.

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Table 2  Long Period Body Wave Inversion Results - errors represent 1 standard deviation, estimated uncertainties in parentheses.

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<td>215.3 ± 0.7 ( ± 2 )</td>
</tr>
<tr>
<td>dip [deg]</td>
<td>63.7 ± 1.0 ( ± 2 )</td>
</tr>
<tr>
<td>slip [deg]</td>
<td>7.4 ± 0.6 ( ± 2 )</td>
</tr>
<tr>
<td>centroid depth [km]</td>
<td>9.9 ± 0.2 ( ± 1 )</td>
</tr>
<tr>
<td>scalar moment [dyn-cm]</td>
<td>8.0 ± 0.4x10^{25} ( ± 0.8 )</td>
</tr>
<tr>
<td>time function:</td>
<td></td>
</tr>
<tr>
<td>duration [s]</td>
<td>6 ( 5 ± 1 )</td>
</tr>
<tr>
<td>element 1(^1)</td>
<td>.22 ± .03</td>
</tr>
<tr>
<td>element 2</td>
<td>.78 ± .03</td>
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</tbody>
</table>

\(^1\) Relative amplitudes of 2 s duration elements.
Table 3  Crustal Structure Models - compressional and shear velocities, $\alpha$ and $\beta$ [km/s]; density $\rho$ [gm/cm$^3$]; P- and S-wave Q factors, $q_p$ and $q_s$.

<table>
<thead>
<tr>
<th>MODEL</th>
<th>$\alpha$</th>
<th>$\beta$</th>
<th>$\rho$</th>
<th>$q_p$</th>
<th>$q_s$</th>
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<tr>
<td></td>
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<td>3.2</td>
<td>2.5</td>
<td>$10^5$</td>
<td>$10^5$</td>
</tr>
<tr>
<td>2</td>
<td>5.5</td>
<td>3.2</td>
<td>2.6</td>
<td>$10^5$</td>
<td>$10^5$</td>
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<tr>
<td></td>
<td>6.5</td>
<td>3.7</td>
<td>2.8</td>
<td>$10^5$</td>
<td>$10^5$</td>
</tr>
<tr>
<td>3</td>
<td>2.5</td>
<td>1.4</td>
<td>1.8</td>
<td>$10^5$</td>
<td>$10^5$</td>
</tr>
<tr>
<td></td>
<td>5.5</td>
<td>3.2</td>
<td>2.6</td>
<td>$10^5$</td>
<td>$10^5$</td>
</tr>
<tr>
<td>4</td>
<td>1.5</td>
<td>0.6</td>
<td>1.6</td>
<td>$10^5$</td>
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<tr>
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<td>5.5</td>
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<td>2.6</td>
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<td>0.6</td>
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<td>2.0</td>
<td>1.15</td>
<td>1.7</td>
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<tr>
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</tr>
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<td>origin time</td>
<td>4:12:27 UTC</td>
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<td>$m_b$</td>
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<td></td>
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<tr>
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</tr>
<tr>
<td>dip</td>
<td>$64 \pm 2^0$</td>
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<td>slip</td>
<td>$7 \pm 2^0$</td>
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</tr>
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<td>$10 \pm 1$ km</td>
<td></td>
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</tr>
<tr>
<td>moment</td>
<td>$8.0 \pm 0.8 \times 10^{25}$ dyn-cm</td>
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</tr>
<tr>
<td>duration</td>
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<tr>
<td>fault width</td>
<td>$&lt; 20$ km (assume 13 km)</td>
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<tr>
<td>fault length</td>
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<td>$\Delta u$</td>
<td>1.2 m</td>
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</tr>
<tr>
<td>$\Delta \sigma$</td>
<td>27 bar</td>
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<tr>
<td>rupture</td>
<td>unilateral to the southwest</td>
<td></td>
<td></td>
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</table>

$^1$ Above parameters reported by USGS.
$^2$ N-S component at 25 km epicentral distance.
FIGURE CAPTIONS

1 Convention of double-couple source specification given by Aki and Richards (1980). Strike $\phi$ is measured clockwise from north. Dip $\delta$ is measured down from horizontal to the right of strike direction. Motion of right-hand block is in the direction of slip $\lambda$.

2 Fault systems and previous events with fault plane solutions in northeastern Turkey.
Large events of the epicentral region:
   Event 1 - 1952, $m_b = 5.8$
   Event 2 - 1924, $m_b = 6.8$
   Event 3 - 1983, $m_b = 6.1$


4 Distribution of aftershock epicenters for 4/11/83 to 26/11/83.

5 Samples of long-period Z-component GDSN seismograms. Start of traces is 30 s before the P arrival.

6 Station coverage of combined long-period WWSSN and GDSN data set.

7 Synthetic and observed seismograms from final long-period inversion.

8 Synthetic and observed seismograms from final long-period inversion, continued.

9 Comparison of aftershock records (top traces) to those from the mainshock (bottom traces). Seismograms begin approximately 30 s before the P arrival.

10 Processed strong motion acceleration, velocity, and displacement records from Toksöz et al. (1984).


12 (a) Surface geometry and coordinate system. (b) Diagram for crustal structure models.

Note: All near-field synthetic seismograms start at $t = 2$ s.
The amplitude scale is centimeters per source moment of $1 \times 10^{25}$ dyn-cm unless otherwise specified.
R - radial
T - transverse
Z - depth, positive into earth
N-S - north-south
E-W - east-west

13 Synthetics for various crustal structure models, varying viewing angle (see Figure 12a), epicenter-source separation = 28 km, \( t_0 = 0.5 \) s, 256 points / trace.

14 Synthetics for various crustal structure models, viewing angle = 63\(^\circ\), epicenter-receiver separation = 28 km, \( t_0 = 0.5 \) s, 256 points / trace.

15 Synthetics for crustal structure model 6, varying layer thickness, source at position \( x = 20 \), \( t_0 = 0.5 \) s, 256 points / trace.

16 Synthetics for crustal structure model 6, varying layer thickness, source at position \( x = 10 \), \( t_0 = 0.5 \) s and 1.0 s, 256 points / trace.

17 Rotated synthetics of Figure 16.

18 Synthetics for crustal structure model 6, varying layer thickness, source at position \( x = 0 \), \( t_0 = 0.5 \) s, 256 points / trace.

19 Synthetics for crustal structure model 6, layer thickness = 800 m, varying source position, \( t_0 = 1.0 \) s, 128 points / trace.

20 Rotated synthetics of Figure 19.

21 Summations of synthetics displayed in Figure 20, representing southward and northward rupture, varying rupture velocity and duration. Positions referred to are of the contributing point sources.

22 Rotated synthetics for crustal structure model 7, layer thickness = 800 m, varying source position, \( t_0 = 1.0 \) s, 128 points / trace.

23 Summations of synthetics displayed in Figure 22, representing southward rupture initiating at varying position, \( \nu_r = 3.1 \) km/s = .98 \( \beta \), duration = 4.8 s.

24 Preferred summation of Figure 23, representing southward rupture, point source contributions from positions \( x = 20 \) through 7.5, \( \nu_r = 3.1 \) km/s = .98 \( \beta \), corresponding duration = 4.8 s and length = 15 km, for \( M_o = 8 \times 10^{28} \text{ dyn-cm} \) (bottom traces). Data records (top traces) from Toksöz et al. (1983) are displayed with same amplitude and time scales.

25 Synthetics for crustal structure model 6, layer thickness = 800 m, varying
depth, source at position $x = 10$, $t_0 = 1.0$ s, 128 points / trace.

26 Summations of synthetics displayed in Figure 22, $v_r = 3.1$ km/s = .98 $\beta$.

   Trace 1 - unilateral northward rupture, point source contributions from positions $x = 22.5$ through 30.
   Trace 2 - unilateral southward rupture, point source contributions from positions $x = 20$ through 12.5.
   Trace 3 - sum of traces 1 and 2.
   Trace 4 - unilateral northward rupture, point source contributions from positions $x = 20$ through 7.5.
   Trace 5 - sum of traces 1 and 4.

27 Diagram of rectangular fault model. The shaded region represents the area which is responsible for surface wave excitation, approximated by the point sources. The time-labeled rupture fronts correspond to $v_r = \beta$. Note the time delay from nucleation to the start of surface wave generation.
Figure 1
Figure 7
30/10/83  \( M_b = 5.3 \)
GAC  \( az. = 321^\circ \)  \( del. = 78^\circ \)

19/9/84  \( M_b = 5.3 \)
TATO  \( az. = 77^\circ \)  \( del. = 66^\circ \)

Figure 9
Figure 10
Figure 11
Figure 12
Figure 13
Figure 17
Figure 21

southern rupture from $X = 20.0$ to $12.5$

$southward$ rupture from $X = 20.0$ to $7.5$

northward rupture from $X = 12.5$ to $20.0$

northward rupture from $X = 7.5$ to $20.0$

duration (s)
rupture velocity (km/s)
Figure 22
Figure 24
BIBLIOGRAPHY


APPENDIX A Seismic Source in a Vertically Stratified Medium

The P- and S-wave displacement fields due to the moment tensor component $M_{jk}$ are

$$u_i^X(r,t) = G_{ij,k}^X(r,t) \ast M_{jk}(t). \quad (B1)$$

The far-field medium response functions are

$$G_{ij,k}^P = \frac{1}{4\pi \rho c^3 r} \gamma_i \gamma_j \gamma_k \delta(t - \frac{r}{c}) \quad (B2)$$

and

$$G_{ij,k}^S = \frac{1}{4\pi \rho \beta^3 r} (\delta_{ij} - \gamma_i \gamma_j) \gamma_k \delta(t - \frac{r}{\beta}).$$

Following the formulation of Nábělek (1984), the moment tensor components are assumed to have a common time dependence,

$$M_{ij}(t) = m_{ij} S(t). \quad (B3)$$

Since displacement in the vertical plane (P-SV radiation) is only excited by forces in that plane, and horizontal displacement perpendicular to propagation (SH radiation) only by forces in that direction, the moment tensor components may be collected into groups. In a cylindrical coordinate system, with $\chi$ denoting propagation direction $\hat{\kappa}$ for P-radiation and $\hat{\kappa} \times \varnothing$ for SV-radiation, the displacement is given by

$$u_\kappa(\varphi, \Delta, t) = (I^{PSV2} M_{PSV2} + I^{PSV1} M_{PSV1} + I^{PSV0} M_{PSV0}) \ast S(t) \quad (B4)$$

and

$$u_\varnothing(\varphi, \Delta, t) = (I^{SH2} M_{SH2} + I^{SH1} M_{SH1}) \ast S(t).$$

The $I^X$ are the azimuth independent medium response functions to couples exciting the two modes of radiation. In an infinite medium, the $I^X$ are simply the response functions (B2) with appropriate choice of indices to represent the various couples.
\[ f^{PSV2}(\Delta, t) = G_{X,r} \] horizontal dipole
\[ f^{PSV1}(\Delta, t) = G_{X,z} + G_{X,r} \] vertical couple
\[ f^{PSV0}(\Delta, t) = G_{X,z} \] vertical dipole
\[ f^{SH2}(\Delta, t) = G_{\varphi,r} \] horizontal couple
\[ f^{SH1}(\Delta, t) = G_{\varphi,z} \] vertical couple.

The \( M^X \) are the sums of the images of the moment tensor components in the directions specified by (B5).

\[ M^{PSV2}(\varphi) = m_{yy}\sin^2\varphi + m_{zz}\cos^2\varphi + 2m_{zy}\cos\varphi\sin\varphi \] (B6)
\[ M^{PSV1}(\varphi) = m_{yz}\sin\varphi + m_{zz}\cos\varphi \]
\[ M^{PSV0}(\varphi) = m_{zz} \]
\[ M^{SH2}(\varphi) = (m_{yy} - m_{zz})\cos\varphi\sin\varphi + m_{zy}(\cos^2 - \sin^2) \]
\[ M^{SH1}(\varphi) = m_{yz}\cos\varphi - m_{zz}\sin\varphi. \]

The labeling 2, 1, 0 is carried over from spherical harmonic analysis of Ward (1980). These indicate the symmetry of the couples in the \( \varphi \times \varphi \) direction. In terms of strike \( \theta \), dip \( \delta \), and slip \( \lambda \), the \( M^X \) are

\[ M^{PSV2} = \left[ \sin \delta \cos \lambda \sin 2(\varphi - \theta) - \sin 2 \delta \sin \lambda \sin^2(\varphi - \theta) \right] / \sqrt{2} \] (B7)
\[ M^{PSV1} = \left[ -\cos \delta \cos \lambda \cos (\varphi - \theta) + \cos 2 \delta \sin \lambda \sin (\varphi - \theta) \right] / \sqrt{2} \]
\[ M^{PSV0} = \left[ \sin 2 \delta \sin \lambda \right] / \sqrt{2} \]
\[ M^{SH2} = \left[ \cos \delta \cos \lambda \sin (\varphi - \theta) + \cos 2 \delta \sin \lambda \cos (\varphi - \theta) \right] / \sqrt{2} \]
\[ M^{SH1} = \left[ \sin \delta \cos \lambda \cos 2(\varphi - \theta) - \frac{1}{2} \sin 2 \delta \sin \lambda \sin 2(\varphi - \theta) \right] / \sqrt{2}. \]

In layered media, the calculation is complicated by the effects of conversion and reflection at the layer boundaries. Since the vertical stratification leaves the cylindrical symmetry intact, the radiation patterns of (B2) are still valid. However, terms dependent on depth \( h \) and ray parameter \( p \) appear. Since these will be the most time consuming part of the Green's function to calculate, the final
expression for the displacement field is written in terms of them. The wave packet destined for a particular $\Delta$ and $\phi$ includes contributions from both up- and down-going phases leaving the source. Each has a different elementary response function $E^X(h,p)$, labeled $X^+$ and $X^-$ respectively. These are generated using a propagator matrix method (Bouchon, 1976; Nábělek, 1984). As long as the source remains within the same layer, a simple time shift accounts for a change in source depth. With attention to writing equation (B5) in vector notation and factoring out the radiation patterns of (B2), the medium response functions are

$$I^{PSV_2}(h,\Delta,t) = \alpha^2 p^2 E^{P^+} + \alpha^2 p^2 E^{P^-} + \beta^2 p \eta_\beta E^{SV^+_\beta} + \beta^2 p \eta_\beta E^{SV^-_\beta},$$

$$I^{PSV_1}(h,\Delta,t) = -2\alpha^2 p \eta_\alpha E^{P^+_\alpha} + 2\alpha^2 p \eta_\alpha E^{P^-_\alpha} + \beta^2(\eta_\beta^2 - p^2) E^{SV^+_\beta} - \beta^2(\eta_\beta^2 - p^2) E^{SV^-_\beta},$$

$$I^{PSV_0}(h,\Delta,t) = \alpha^2 \eta_\alpha^2 E^{P^+} + \alpha^2 \eta_\alpha^2 E^{P^-} - \beta^2 p \eta_\beta E^{SV^+_\beta} + \beta^2 p \eta_\beta E^{SV^-_\beta},$$

$$I^{SH^2}(h,\Delta,t) = -\beta \eta_\beta E^{SH^+} + \beta \eta_\beta E^{SH^-},$$

$$I^{SH^1}(h,\Delta,t) = \beta p E^{SH^+} + \beta p E^{SH^-},$$

where the direction cosines have been written in terms of $\alpha$, $\beta$, $p$, and $P$- and $S$-wave vertical slowness, $\eta_\alpha$ and $\eta_\beta$.

The uncoupled P-SV and SH displacement fields in terms of the elementary response functions are

$$u^{PSV}(\phi,\Delta,t) = M_n (E^{P^+} R^{P^+} + E^{P^-} R^{P^-} + E^{SV^+_\alpha} R^{SV^+_\alpha} + E^{SV^-_\alpha} R^{SV^-_\alpha}) \cdot \dot{S}(t),$$

$$u^{SH}(\phi,\Delta,t) = M_n (E^{SH^+} R^{SH^+} + E^{SH^-} R^{SH^-}) \cdot \dot{S}(t),$$

with

$$M_n = \sqrt{2} M_0,$$

where $M_0$ is the scalar seismic moment. $\dot{S}(t)$ is the source time function. The elementary radiation patterns are,

$$R^{P^+} = \alpha^2 [M^{PSV_2} p^2 + M^{PSV_0} \eta_\alpha^2 - 2 M^{PSV_1} p \eta_\alpha]$$

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\[ R^{P-} = \alpha^2 [M^{PSV_2} p^2 + M^{PSV_0} \eta_\alpha^2 + 2 M^{PSV_1} p \eta_a] \]

\[ R^{SV+} = \beta^2 [M^{PSV_1} (p^2 - \eta_\beta^2) - (M^{PSV_0} - M^{PSV_2}) p \eta_\beta] \]

\[ R^{SV-} = \beta^2 [M^{PSV_1} (p^2 - \eta_\beta^2) + (M^{PSV_0} - M^{PSV_2}) p \eta_\beta] \]

\[ R^{SH+} = \beta [M^{SH_1} p - M^{SH_2} \eta_\beta] \]

\[ R^{SH-} = \beta [M^{SH_1} p + M^{SH_2} \eta_\beta] . \]
APPENDIX B  Computation of Body Wave Synthetics

The expression given by Nábelek (1984) representing a synthetic seismogram is

\[ s(t) = M_n \sum_{k=1}^{n} \sum_{i=1}^{m} w_k H_i(t - \tau_k - (\eta_i - \bar{\eta}_i)d + \bar{\eta}_i \xi + \rho \cos(\varphi - \psi) r_i , \]  

(C1)

where

\[ H_i(t) = h_i(t) * T_{\deltaT}(t) * M(t) * C(t) * R(t), \]  

(C2)

and

\[ M_n = \text{moment tensor norm} \]  

(C3)

\[ w_k = \text{weight of source time function element} \]  

\[ H_i(t) = \text{elementary seismograms} \]  

\[ t = \text{time} \]  

\[ \tau_k = \text{time shift of time function element} \]  

\[ \eta_i = \text{vertical slowness of the parent rays} \]  

\[ \bar{\eta}_i = \text{average vertical slowness over vertical distance of ray contributing to first motion} \]  

\[ d = \text{centroid depth with respect to the layer interface above} \]  

\[ \xi = \text{centroid depth with respect to the nucleation point} \]  

\[ p = \text{ray parameter} \]  

\[ \rho = \text{radial distance from epicenter of nucleation point to epicenter of centroid} \]  

\[ \psi = \text{azimuth from epicenter of nucleation point to epicenter of the centroid} \]  

\[ \varphi = \text{station azimuth} \]  

\[ r_i = \text{normalized source radiation pattern} \]  

\[ n = \text{number of time element function elements} \]  

\[ m = 4 \text{ for P-SV system, 2 for SH system} \]  

\[ h_i(t) = \text{response of the source crust to the parent rays from a point P, SV, and SH source} \]
\[ T_{\Delta r}(t) = \text{normalized time function element} \]

\[ M(t) = \text{mantle response function} \]

\[ C^R(t) = \text{receiver crustal response} \]

\[ R(t) = \text{instrument response} \]