FOCAL DEPTHS AND MECHANISMS OF MID-OCEAN RIDGE EARTHQUAKES
FROM BODY WAVEFORM INVERSION

by

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ABSTRACT

Accurate focal depths of mid-ocean ridge earthquakes are critical to tectonic interpretations of ridge-crest processes, notably the thickness of the brittle layer beneath the ridge axes and, implicitly, the depth of hydrothermal circulation. We determine in this thesis the focal depths and source characteristics of 29 ridge-axis earthquakes along the Mid-Atlantic Ridge, the Azores spreading center, the northern Red Sea, the Gulf of Aden, and the Carlsberg and Central Indian Ridges. The source parameters are obtained by a formal inversion of long-period P and SH waveforms [Nabelek, 1984]. Because of the special importance of focal depth, we devote particular attention to the resolution of centroid depth and to an assessment of the relationship between the formal error given by the inversion method and the actual uncertainty. We also stress by example the advantage of including SH waves as well as P waveforms in the inversion. We examine as well the effects of uncertainties in the source region velocity structure and of incomplete azimuthal coverage on the inversion results. The waveform inversion solutions are shown to be generally insensitive to details of source velocity structure; as long as azimuthal coverage spans more than 2 quadrants of the focal sphere for both P and SH waves, the solution obtained by waveform inversion is robust with respect to inclusion or deletion of selected subsets of the data.

The ridge-axis earthquakes of this study are all characterized by normal faulting on planes dipping at about 45° and generally striking parallel to the trend of local ridge segments. Seismic moments range from 2 to 10^24 dyn-cm, and source time functions are all of simple form. The P and S waveforms for all earthquakes can be well matched using
conventional values of $t^*$ (1 and 4 sec, respectively). The P waves from these earthquakes show strong water-column reverberations, suggesting that fault rupture generally extended to the seafloor. The predominant period of these reverberations constrains the water depth in the epicentral region. On the basis of estimated water depth and the epicentral location, all of these earthquakes can be shown to have occurred beneath the inner floor of the median valley. We cannot exclude the possibility that some of these events occurred near the inner wall of the rift valley (e.g., the August 15, 1966 and the May 31, 1971 earthquakes) or that they represent slip on major inward-dipping faults that contribute to the relief of the inner valley walls. It may be inferred with reasonable certainty, however, that none of these earthquakes occurred beneath the rift mountains.

The centroid depths of the earthquakes in this study provide important information on the vertical extent of brittle failure along slow spreading ridges. A few of the earthquakes included here, however, may not be representative. Two earthquakes in the Azores region may have occurred in response to the slow rifting of oceanic lithosphere rather than along a true oceanic spreading center. Two earthquakes in the northern Red Sea are likely to have occurred in thinned continental crust. Excluding these earthquakes, the centroid depths of the ridge-axis events with adequate azimuthal coverage are all very shallow, ranging from 1 to 4.2 km. If centroid depth marks the mean depth of fault slip, then earthquake faulting extended to 2-8 km beneath the sea floor during these earthquakes. The two Azores earthquakes have centroid depths of about 5 km. The two northern Red Sea earthquakes have centroid depths of about 6 km, consistent with the view that seismic faulting accompanying continental rifting extends to greater depths than does seismic activity along oceanic ridge axes. Two microearthquake experiments conducted in regions of the Mid-Atlantic Ridge near epicenters of one or more of the large earthquakes in this study show that microearthquakes occurred deeper than the probable slip zone of the larger earthquakes. However, the uncertainties both in centroid depths and in and microearthquake focal depths also permit a coincidence in depth of the two measures of activity.

The combined set of centroid depths for well-constrained source mechanisms of Atlantic and Indian Ocean ridge earthquakes show that the greatest depth of earthquake faulting tends to shoal with increasing spreading rate. This provides direct evidence for a general decrease with spreading rate in the maximum thickness of the zone of brittle behavior for slow-spreading ridges. The maximum depth of earthquake faulting probably outlines the depth of the brittle-ductile transition beneath the ridge axis. It is likely that the
lithosphere is colder and mechanically stronger beneath the slowest-spreading ridge segments because there is more time for the lithosphere to cool between major episodes of crustal magma injection and volcanism. Additional evidence for this interpretation is the fact that the largest earthquakes ($M_0 > 9 \times 10^{24}$ dyn·cm) occur on ridge segments with half-spreading rates less than 14 mm/yr.
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APPENDIX C. COMPARISON OF INVERSION SOLUTIONS WITH ALTERNATIVE SOURCE VELOCITY STRUCTURES 289
CHAPTER 1. INTRODUCTION

The discovery of a worldwide mid-ocean ridge system was probably the most important discovery in marine geophysical research. Mid-ocean ridges make up over half of the Earth's major plate boundaries; they are regarded as zones in which the upper mantle exerts an exceptionally strong influence on the Earth's crust and lithosphere. The ridge axis proper is the most active part of the ridge system, with tectonic processes evolving before our eyes. Despite the fact that ridges are largely inaccessible to direct observation and remote from seismographical stations, studies of mid-ocean ridges have contributed significantly to the theory of plate tectonics. In the 1950's, Bullard et al. [1954] discovered high heat flow in the rift zones of mid-ocean ridges. This and subsequent studies of ridge heat flux [Von Herzen and Maxwell, 1959; Von Herzen and Uyeda, 1963; Vacquier and Von Herzen, 1964] support the hypothesis that the genesis of mid-ocean ridges is related to the ascent of deep-seated mantle material [Hess, 1962]. Based on studies of magnetic lineations, Vine and Matthews [1963] suggested that the spreading of the crust is recorded by the imprinting of geomagnetic field reversals in the remanent magnetization of oceanic crust. Epicenter and focal mechanism determinations have revealed both the boundaries of plates [Barazangi and Dorman, 1969] and their relative motions [Sykes, 1967].

The mid-ocean ridge system displays significant variations in structure, tectonics, volcanism, and hydrothermal
circulation [Macdonald, 1982]. In the Indian and Atlantic Oceans the ridges are relatively narrow and rugged, with a prominent deep valley along the crest. In contrast, the fast-spreading East Pacific Rise is much wider and less rugged and has no prominent valley along its crest. Several models, including "hydraulic head loss" [Sleep, 1969; Sleep and Biehler, 1970; Sleep and Rosendahl, 1979], "material imbalance" [Deffeyes, 1970], and "steady state necking" [Tapponnier and Francheteau, 1978], have been proposed to explain the apparent correlation of rift valley characteristics with spreading rate. None of these models has been clearly shown to be superior to date, largely for lack of knowledge of the rheology of the deeper crust and of the mantle beneath the ridge axis.

In this thesis, we will determine the source characteristics of large ridge-axis earthquakes from an inversion of long-period P and SH waveforms. One of the main purposes of this study is to obtain reliable centroid depths. Since earthquakes are believed to be the result of brittle failure, the maximum depth at which earthquake faulting occurs can be interpreted as the depth of the brittle-ductile transition zone.

**VOlCANISM AND TECTONICS OF THE AXIAL ZONE**

The axial region of a mid-ocean ridge can be divided into a central zone of "crustal accretion" and a flanking "plate boundary" zone [Macdonald, 1982]. The crustal accretion zone is characterized by recent and ongoing magmatism and volcanism; the active volcanic portion of this region is sometimes called
the "neovolcanic zone." At slow-spreading ridges, the nonvolcanic zone can consist of isolated and discontinuous volcano chains [Needham and Francheteau, 1974; Macdonald et al., 1975]. In contrast, fast-spreading ridges display essentially continuous neovolcanic zones interrupted only by transform faults and offset overlapping spreading center segments [Searle et al., 1981; Macdonald and Fox, 1983]. The central neovolcanic zone is typically 1-2 km wide at all spreading rates [Normark, 1976; Bryan and Moore, 1977; Luyendyk and Macdonald, 1977; van Andel and Ballard, 1979; Spiess et al., 1980]. However, this is only an instantaneous view of the volcanic zone. From the width of the magnetic anomaly polarity transition, the crustal accretion zone has been estimated to be 1-3 km for fast-spreading ridges [Klitgord et al., 1975; Macdonald et al., 1980a] and 1-8 km for slow-spreading ridges [Macdonald, 1977]. This variation of crustal accretion zone width suggests that the structure of an individual median valley segment varies with time [Macdonald, 1977]. In support of this view, the median valley of slow-spreading ridges can range from a configuration with a wide inner floor and narrow flanking terraces (e.g., FAMOUS area) to one with a narrow inner floor and well-developed wall terraces (e.g., AMAR).

The plate boundary zone is characterized by active faulting and deformation. This zone can be subdivided into zones of fissuring and faulting. Fissures have been observed along all ridge axes. At both types of ridges, the most intense fissuring occurs in bands 1-2 km wide flanking the
central volcanic zone. These fissures are typically 1-3 m across, extending 10 m to 2 km along strike [Luyendyk and Macdonald, 1977; Ballard and van Andel, 1977]. They are parallel to the strike of the ridge, implying that they are formed by tension failure rather than cracking due to thermal contraction [Luyendyk and MacDonald, 1977]. It is likely that these fissures provide access for cold seawater to penetrate and cool the deeper levels of hot young crust [Lister, 1977].

At slow spreading ridges, fissures are also observed along the central spreading axis. At intermediate- to fast-spreading ridges, fissures probably also occur along the axis but are obscured by continuous central volcanoes [Macdonald, 1982].

Outside the fissure zone, significant vertical disruption of the crust indicates active normal faulting. At slow spreading ridges, individual faults have vertical throws of 200 m or greater, and a series of these faults creates scarps 600 m or higher, resulting in deep valleys [MacDonald and Luyendyk, 1977]. At fast spreading ridges, no rift valleys are evident, but normal faulting creates axially dipping fault scarps with throws of 50 m or less [Klitgord and Mudie, 1974].

Using precise bathymetric data, Macdonald and Atwater [1978] estimated that the active faulting continues up to 30 km from the axis in the rift mountains of the Mid-Atlantic Ridge. Using a similar analysis, Shih [1980] estimated that active faulting extends to 4 to 10 km from the spreading axis of intermediate- and fast-spreading ridges.
HYDROTHERMAL CIRCULATION

Both high and low heat flow values have been measured close to the axes of mid-ocean ridges, initially producing a confused picture of the heat flow pattern across ridges [Lee and Uyeda, 1965; Von Herzen and Langseth, 1965; Von Herzen and Lee, 1969; Langseth and Von Herzen, 1970; Lister, 1970]. In addition, all conductive models of the near-ridge thermal structure [McKenzie, 1967; McKenzie and Sclater, 1969; Sleep, 1969; Sclater et al., 1971] overestimate the heat flow in young oceanic lithosphere. These observations led to a realization of the importance of hydrothermal circulation as a mechanism of heat loss at mid-ocean ridges [Lister, 1972]. Hydrothermal circulation was first verified by measurement of helium isotopes, potential temperature and salinity in bottom water samples at the Galapagos spreading center [Weiss et al., 1977]. The hydrothermal field discovered in 1979 at the axis of the East Pacific Rise at 21°N [RISE Project Group, 1980] allowed the first direct estimate of axial hydrothermal heat flux [Macdonald et al., 1980b]. The total heat flux from the vents is overwhelmingly large, suggesting that these vents are not steady-state. Recently, hydrothermal venting phenomena, including black smokers, were discovered along the Mid-Atlantic Ridge at 26°N, providing direct evidence of high-temperature hydrothermal activity at slow-spreading oceanic ridges [Rona, 1985].

It is difficult to estimate the depth of penetration of hydrothermal circulation, but earthquake depths should provide
some bound. In the East Pacific Rise hydrothermal area at 21°N, microearthquakes are between 1 and 3 km deep [Macdonald et al., 1980c; Riedesel et al., 1982]. Focal depth studies of intraplate earthquakes [Chen and Molnar, 1983; Wiens and Stein, 1983] have shown that seismic failure is generally confined to regions at temperatures less than about 600°-800°C. At higher temperatures, the lithosphere cannot sustain significant differential stress because of ductile flow. Since earthquakes represent brittle failure, the maximum depth at which earthquakes occur can provide bounds on the depth of the brittle-ductile transition zone. The occurrence of earthquakes beneath a ridge axis indicates the presence of a significant thickness of crust at a temperature less than 600°C-800°C, presumably the result of on-axis cooling by hydrothermal circulation. Oxygen isotope studies in the Samail ophiolite provide additional evidence that hydrothermal circulation in oceanic crust may reach depths greater than 5 km [Gregory and Taylor, 1981].

RIDGE AXIS SEISMICITY

Earthquakes along mid-ocean ridges are a primary manifestation of the process of seafloor spreading. The fast-spreading East Pacific Rise appears aseismic at teleseismic distances; recorded earthquakes are confined to transform faults. Slow-spreading ridges, in contrast, are commonly marked by earthquakes with body-wave magnitudes as large as 6. Modern study of ridge-crest earthquakes dates from the
demonstration by Sykes [1967] that such events are characterized by normal faulting mechanisms having T axes that are approximately horizontal and oriented parallel to the direction of spreading. Aside from simple fault-plane solutions, few studies have since been made of the source characteristics of ridge-crest earthquakes. This is particularly true of focal depth, an extremely important quantity for understanding the depth extent of brittle behavior at ridge crests and its implications for local thermal structure and for such controlling processes as shallow magma injection and cooling by hydrothermal circulation [e.g., Lister, 1977].

While it is generally agreed that ridge-crest seismic activity is quite shallow [e.g., Isacks et al., 1968], well-constrained focal depths have been presented in the literature for only a small number of large ridge-crest earthquakes. An early effort to determine the focal depth of ridge-crest earthquakes was that of Tsai [1969], who fit Rayleigh-wave amplitude spectra from several events to synthetic spectra generated for various source depths and obtained focal depths of 30 to 65 km. Weidner and Aki [1973] demonstrated, however, that, when phase as well as amplitude spectra of Rayleigh waves are considered, the focal depths of ridge-crest earthquakes are required to be very shallow; they obtained depths of 3 ± 2 km beneath the seafloor for two normal-faulting events on the axis of the Mid-Atlantic Ridge. Duschenes and Solomon [1977] reported that the short-period P waveforms from these two
events contain apparent depth phases consistent with such shallow focal depths. Conventional fault-plane solutions derived from P-wave first motions for ridge-crest earthquakes often show non-orthogonal nodal planes [Sykes, 1967, 1970a; Thatcher and Brune, 1971; Solomon and Julian, 1974], in apparent disagreement with the expected double-couple source model. Hart [1978] suggested that this non-orthogonality is a consequence of the interference between direct and surface-reflected phases for extremely shallow sources. This suggestion was confirmed by Trehu et al. [1981], who inverted Rayleigh wave spectra for the source moment tensor and synthesized long-period P waveforms for two earthquakes on the Reykjanes Ridge. They showed that both the surface wave spectra and the P waveforms were compatible with double-couple source mechanisms and very shallow focal depths. Pearce [1981] modeled the short-period P waves from a small (mb = 4.7) earthquake on the axis of a spreading center segment in the Gulf of Aden and inferred a focal depth of 5 ± 2 km.

Beyond these few studies of large earthquakes, the remaining published information on the depth of seismogenic faulting along ridge axes comes from microearthquake surveys. While a number of microearthquake studies have been carried out along segments of the Mid-Atlantic Ridge and other ridge systems using either sonobuoys or ocean-bottom seismometers (ORS's), the majority of these experiments were conducted with an insufficient number of stations to resolve focal depths [Duschenes et al., 1983]. Two Mid-Atlantic Ridge microearth-
quake experiments with networks of more than 3 instruments are noteworthy exceptions. One was near the eastern intersection of St. Paul's Fracture Zone and the median valley of the ridge [Francis et al., 1978], and the second was within the median valley at a site (near 23°N) distant from major transform faults [Toomey et al., 1985]. Both regions displayed microearthquake activity within the median valley to depths of 7-8 km below the sea floor, and both are also the sites of large earthquakes well recorded at teleseismic distances. These data provide an opportunity to compare the centroid depths of the large earthquakes with the distribution of focal depths of microearthquakes in the same region.

THESIS OUTLINE

Accurate focal depths of mid-ocean ridge earthquakes are critical to tectonic interpretations, because limits on focal depths would place bounds the thickness of the brittle layer beneath the ridge axes and, implicitly, on the depth of hydrothermal circulation. The main objective of this study is to obtain reliable focal depths for a large number of ridge-crest earthquakes. Generally the focal depths of shallow earthquakes cannot be determined accurately from travel times, at least without arrival times at nearby stations, because the depth phases pP and sP are not clearly observed on teleseismic recordings. In contrast, very good depth determination is possible using waveform modeling for events well-recorded at teleseismic distances. Such efforts are worthwhile, because, although OBS's will give increasing numbers of precise depths
in the future, microearthquakes may not be similar to larger earthquakes.

We determine in this thesis the focal depths and source characteristics of 29 ridge-axis earthquakes in the North Atlantic and Indian Oceans. The source parameters are obtained by a formal inversion of long-period P and SH waveforms. We use the inversion procedure of Nabelek [1984], as applied to oceanic intraplate earthquakes by Bergman et al. [1984] and Bergman and Solomon [1984, 1985a]. Because of the special importance of focal depth, we devote particular attention to the resolution of centroid depth and to an assessment of the relationship between the formal error given by the inversion method and actual uncertainty. After a presentation of the results of individual inversions, we generalize to a discussion of the implications of the derived source parameters for median valley tectonics along slow-spreading ridge systems.

Chapter 2 describes the inversion technique of Nabelek [1984] and discusses the depth resolution of the inversion procedure. We also examine the effects of uncertainties in velocity structure and complete azimuthal coverage on the inversion results. We find that the waveform inversion solutions are generally insensitive to details of source velocity structure. Also, as long as the azimuthal coverage spans more than 2 quadrants of the focal sphere for both P and SH waves, the inversion result is stable.

In Chapter 3, we present source studies of 14 Mid-Atlantic Ridge earthquakes and 2 normal-faulting earthquakes on the
slowly spreading portion of the Azores-Gibraltar plate boundary. All of these earthquakes have mechanisms characterized principally by normal faulting on fault plates with dip angles near 45°. The water depths estimated from the predominant period of water-column reverberations in the P wave-trains indicate that the earthquakes all occurred beneath the inner floor of the median valley. The centroid depths of all Mid-Atlantic Ridge earthquakes are very shallow, ranging from 1 to 3 km. The two earthquakes in the Azores region have centroid depths of about 5 km.

In Chapter 4, we present source studies of 13 large ridge-axis earthquakes in the Indian Ocean region, including events in the Red Sea and the Gulf of Aden and on the Carlsberg and Central Indian Ridges. Again, a comparison of water depth above the epicenter determined from P waveforms with that from bathymetric maps indicates that these earthquakes occurred beneath the inner floor of the median valley. One of these earthquakes (October 7, 1981), however, is probably an off-ridge event, though the seismicity does not clearly define the plate boundary at this area. The two earthquakes in the Red Sea are likely to have occurred in thinned continental crust. Excluding these three events, the centroid depths range from 1 to 5 km. The deepest of these events (April 22, 1969) occurred near a ridge-transform intersection where the lithosphere is probably colder than normal. The two earthquakes in the Red Sea have greater centroid depths, providing additional evidence that seismogenic deformation within rifting
continental crust extends to a greater depth than along oceanic spreading centers.

In Chapter 5, we discuss some general characteristics of the ridge-axis earthquakes of this study. A question of particular interest is the confidence with which the depths of such earthquakes may be used to infer the extent of the brittle-ductile transition zone beneath the slowly spreading ridges. We also examine the issue of focal depth versus spreading rate.

In Chapter 6, the major conclusions of this thesis are summarized, and suggestions for future research are offered.
CHAPTER 2.
WAVEFORM INVERSION PROCEDURE AND RESOLUTION OF SOURCE PARAMETERS

INTRODUCTION

This chapter first introduces the concept of the centroid point source which is used to describe the general earthquake source that forms the basis for body waveform inversion. Next, we describe the inversion procedure, stressing by example the advantage of including SH waves as well as P waveforms. We then discuss at length the resolution of centroid depth obtainable with the inversion procedure. Finally we show the effects on the inversion results of uncertainties in the velocity structure and of less than optimal azimuthal coverage.

SOURCE PARAMETERIZATION

The formulation of source parameterization below is drawn primarily from Nabelek [1984]. The displacement field due to a distribution of moment density tensor can be written as

\[ u_k^+(x,t) = \int \sum_{i,j} m_{ij}(x',t) * g_{ki,j}(x',t; x,0) dV \] (1)

where \( u_k^+(x,t) \) is the displacement in the k direction at the point \( x \) and time \( t \); \( g_{ki,j} \) are Green's functions; \( x' = 3/axj \) taken at the source point \( x' \); and \( m_{ij} \) are components of the moment density tensor which are related to the body force density by \( F_i = -m_{ij,j} \). The symbol * denotes convolution, and the integral is over the volume \( V \) of nonzero \( m_{ij} \). Equation (1)
expresses the theoretical relationship between the data \( u_k \) and
the unknowns \( m_{ij}(\xi, t) \). An infinitely large data set encompassing
all time and space positions, however, is required to define
completely the moment density distribution \( m_{ij} \) [Ward, 1983].
Since we have only a finite data set, some sort of
parameterization of the source is necessary.

When the source dimension and duration are small compared
with the dominant wavelength and period of the motion,
respectively, we can expand the Green's function in a Taylor
series about a point \((\xi_0, \tau_0)\) in the source region:

\[
\begin{align*}
    u_k(x,t) &= g_{ki;j}M_{0ij} - g_{ki;j}M_{1ij} + g_{ki;j}M_{2ij} \\
             &\quad + \frac{1}{2} \ddot{g}_{ki;j}M_{3ij} - \dddot{g}_{ki;j}M_{4ij} + g_{ki;j}m_{0ij}m + \ldots 
\end{align*}
\]

(2)

where

\[
M_{nij}\ldots n = \int_0^\infty (\tau - \tau_0)^n M_{ij}\ldots n (\tau) d\tau 
\]

(3)

\[
M_{ij}\ldots n = \int_V (\xi_2 - \xi_2^0)\ldots(\xi_n - \xi_n^0)m_{ij}(\xi, t) dV 
\]

(4)

\[
\ddot{g}_{ki;j} = \frac{\sigma}{n} - \frac{\gamma_j^n}{Cn} g_{ki,n} = g_{ki,j} 
\]

(5)

where the dot denotes differentiation with respect to time, \( \gamma_j^n \)
are the direction cosines of the departing rays, \( n \) is an index for
upgoing and downgoing P and S waves, and \( Cn \) is the wave velocity
in the source region. The six zero-order terms, \( M_{0ij} \), are the
components of the moment tensor. We can choose \( \xi_0 \) and \( \tau_0 \) so that
\( M_{2ij} \) and \( M_{3ij} \) vanish; \( \xi_0 \) and \( \tau_0 \) are then the centroid location
and centroid time. Dziewonski et al. [1981] combine \( M_{0ij}, \xi_0 \) and
to form their centroid moment tensor (CMT) solution. The higher-order terms $M^2_{ij}, M^1_{ij}$, and $M^0_{ij}$ give estimates of the source duration, the rupture velocity, and the extent of the source about the centroid position. As pointed out by Ward [1983], second-order terms are smaller than the CMT terms by about a factor of 10, thus making second-order analysis difficult, so higher-order terms are usually dropped. However, Nabelek (personal communication, 1985) has argued that there is some trade-off between the source duration and centroid depth and therefore it is important to include the effect of the source duration in the inversion. Following Nabelek [1984], we grouped the higher-order terms together and called this group the source time function. It should be noted that because the absolute travel time is ignored, neither the absolute epicentral location nor the origin time of the centroid source can be determined directly. With the above parameterization, we can nonetheless study the average properties of the source. The unknowns in our problem are the spatial moment tensor of degree zero, its centroid depth, and the source time function.

**INVERSION PROCEDURE**

**General Formalism**

We set up the inverse problem in the form

$$d = m(P)$$  \hspace{1cm} (6)

where $d$ contains the observations and $m(P)$ is a synthetic seismogram for model parameters $P$. The least squares problem is based on the minimization of

$$S = [d-m(P)]^T C_d^{-1} [d-m(P)] + [P_o-P]^T C_P^{-1} [P_o-P]$$  \hspace{1cm} (7)
where $P_0$ are the parameters of an initial model, $C_d$ and $C_p$ are the covariance matrices for data and parameters, respectively, and $T$ denotes transpose. Since the problem is nonlinear, the least squares solution is found iteratively. Tarantola and Valette [1982a] show that an appropriate algorithm to compute $P$ is

$$P_{k+1} = P_k + (J^T C_d^{-1} J)^{-1} J^T C_d^{-1} \left[ d - m(P_k) \right] - C_p^{-1} (P_k - P_0)$$

where $J$ is the matrix of partial derivatives

$$J_{ij} = \frac{\partial m_i}{\partial P_j}$$

For the ridge earthquakes in our study we have almost no a priori knowledge of the model parameters, so we set the covariance matrix $C_p = \sigma^2 I$ in the limit where $\sigma^2 \rightarrow \infty$. In such a limit we are minimizing

$$S' = \left[ d - m(P) \right]^T C_d^{-1} \left[ d - m(P) \right]$$

and the corresponding algorithm can be expressed as

$$P_{k+1} = P_k + (J^T C_d^{-1} J)^{-1} J^T C_d^{-1} \left[ d - m(P_k) \right]$$

which is the solution of the classical nonlinear least squares problem.

The theoretical body wave Green's functions are calculated in the frequency domain using propagator matrices and the reciprocity theorem [Bouchon, 1976]. A set of these functions for each layer is stored for the computation of synthetic seismograms. We assume that ridge-axis earthquakes are caused by shear failure, so in our analysis the inversion solutions
are obtained using the constraint that the source be a double-couple. Trehu et al. [1981] and Nabelek [personal communication, 1985] have shown that any non-double-couple component is insignificant (< 10%) for two Mid-Atlantic Ridge earthquakes.

Procedural Details

Long-period P and S waves recorded by the World-Wide Standardized Seismograph Network (WWSSN) and the Global Digital Seismic Network (GDSN) at epicentral distances between 30° and 80° are hand digitized and then interpolated at intervals of 0.5 sec [Wiggins, 1976]. The SH waves are obtained by projecting the observed horizontal component seismograms onto a line perpendicular to the azimuth from the station to the epicenter. We try to obtain an optimum azimuthal coverage for each earthquake. When several stations with similar waveforms near each other are available, we select the one with the best signal-to-noise ratio. When the signal-to-noise ratios are about the same, we select the one with the largest magnification. The seismograms from different stations are equalized to a common instrument magnification of 3000, and through corrections for attenuation and geometric spreading, to a common epicentral distance of 40 degrees. Station weights are variable and are based on the background noise level. Additional weights may be applied to compensate for uneven station distributions. The time window containing the waveform to be inverted is selected a priori for each station; for P waves the window is usually extends about 20 sec from the first onset to the end of the first water wave. If the window
is too short, the source description may be incomplete and erroneous source parameters may result. Too long a time window may result in too much weight on the water waves, thus producing a poor fit to the earlier portion of the waveform.

The parameters of the best-fitting point source are those which minimize the mean square residual $r^2$, defined as

$$r^2 = \frac{\sum_{j=1}^{N} \sum_{i=1}^{M_j} w_j (s_{ij}-o_{ij})^2}{\sum_{j=1}^{N} M_j}$$

where $s_{ij}$ and $o_{ij}$ are the amplitudes of the synthetic and observed seismograms at station $j$ and time sample $i$, $w_j$ is the weight used for station $j$, $M_j$ is the number of samples in the time window used in the inversion for station $j$, and $N$ is the total number of stations. A station with both a P and an SH waveform used in the inversion is counted twice in (9). The station weights $w_j$ are in inverse proportion to the noise level prior to the start of the time window. The weighting is normalized to the average of all weighting factors by the formula

$$w_j = \frac{1/\sigma_j^2}{(1/N)\Sigma(1/\sigma_j^2)}$$

where $\sigma_j^2$ is the estimated noise variance at station $j$. If the errors in the data and model are normally distributed, minimization of equation (9) yields the maximum likelihood estimate.

In the inversion, the source is assumed to be represented by an average mechanism at its centroid location. The double couple mechanism is specified by the strike, slip and dip
Since we ignore source finiteness effects, the synthesized centroidal seismograms sometimes begin earlier or later than the observed first motion. During the inversion, we iteratively realign the observed and the synthesized waveforms if a better cross-correlation can be found. The minimum allowable shift is 0.5 sec, the same as our digitization angles, specified with the convention of Aki and Richards [1980] and given for all sources in the abbreviated form (strike, dip, slip), with all angles in degrees. The time history is allowed to be complicated. The source time function is represented by a series of overlapping triangle functions of equal duration and variable amplitude. The number and duration of the time function elements are chosen a priori, but the amplitudes are determined in the inversion. The length of each time function element should be set to the minimum resolvable time interval, which depends on the frequency content of the signal. If the time element length is chosen to be too short, the result is an instability in estimated amplitude. If, on the other hand, it is too long, the result is a poor description of the source and possibly biased estimates of other source parameters, most likely depth. The number of time function elements needed depends on the source duration. However, there is generally some trade-off between the centroid depth and the total duration of the source time function. In the inversion it is important not to let a significant portion of the source time function become negative; however, a small negative value of the last time element is allowed.
interval. Inclusion of this time shift is important; since horizontal mislocation is not included as a parameter in the inversion, we have to remove the lateral variations in travel time. The derived focal mechanism, source time function and centroid depth all depend on the proper alignment of synthetic and observed waveforms, so these time shifts are necessary to obtain an optimal solution.

For all mid-ocean ridge axis earthquakes the waveform inversions were performed with the same source velocity structure: a water layer with variable thickness, a single crustal layer of thickness 6 km and P and S-wave velocities $\alpha = 6.4 \text{ km/s}$ and $\beta = 3.7 \text{ km/s}$, respectively, and a mantle halfspace with $\alpha = 8.1 \text{ km/s}$ and $\beta = 4.6 \text{ km/s}$. The thickness of the water layer is first estimated from bathymetric maps, then further constrained by the predominant period of water column reverberations, seen as prominent phases immediately following the P wave arrival. A crustal thickness of about 6 km has been reported for several axial valley segments along the Mid-Atlantic Ridge [Fowler, 1976; Bunch and Kennett, 1980; Purdy and Detrick, 1985]. While uniform-velocity crust and mantle layers are oversimplifications to the actual structure, we demonstrate below that the waveform inversion results are generally insensitive to details of source structure.

Other aspects of data reduction and waveform synthesis follow the procedure of Bergman et al. [1984]. In particular, we employed the conventional values of 1.0 and 4.0 s for the attenuation parameter $t^*$ [Futterman, 1962] for long-period P and
SH waves, respectively. Because there is significant trade-off between $t^*$ and source time function, more precise values of $t^*$ cannot be independently determined from our inversion.

**Importance of Including SH waves**

By including SH waveforms as well as P waveforms in a formal inversion a better constrained focal mechanism is obtained than if P waves are used alone [Nabelek, 1984]. We illustrate this principle by example: the June 28, 1972 earthquake ($m_b = 5.5$, $M_S = 5.5$) in the northern Red Sea. Pearce [1977] obtained a strike-slip mechanism (335/75/2) from an analysis of the relative amplitudes of short-period P and pp waves. Since he did not give seismic moment and centroid depth, to obtain these parameters we first fix the double couple orientation as given by him and use only P waves in the inversion. Figure 2.1 shows that Pearce's solution does not fit the observed long-period waveforms well. Although most of the polarity in the P and SH waves is correct, some amplitudes are definitely wrong. Synthetic P waves at COL, MAT, KBL, and NDI have amplitudes too small and all synthetic SH waves are too large.

Figure 2.2 shows the solution after 5 iterations using only P waveforms and Pearce's [1977] solution as the starting model. The mechanism (21/60/345) is still strike-slip, although it is very different from Pearce's solution. Most of the P waves fit quite well, but SH waves at NIL, CHG, and MAL have the wrong first motions. When we then include the SH waveforms in the inversion, after 5 more iterations the solution converges to a nearly pure normal faulting mechanism (303/40/278). All waveforms are quite well fit (Figure 2.3). This focal mechanism
is very close to the final solution (288/40/260) obtained in Chapter 4 using different station weightings and a pure normal faulting mechanism as the starting model.

From the above example, it is clear that SH waves can provide essential information in focal mechanism studies. Both phase and amplitude information should be employed to avoid ambiguous solutions.

DEPTH RESOLUTION

Since the centroid depths found for ridge axis events are perhaps the most important new source parameters reported here, the depth resolution of our inversion procedure deserves particular attention. For deeper earthquakes, the reflected phases from the free surface are useful in determining the correct source depth. However, for shallow earthquakes, the source duration becomes much greater than the travel time between the direct and reflected waves, and the source depth is difficult to estimate. A major limitation in determining the centroid depth using long-period body waveforms is the trade-off between the duration of the source time function and the best-fitting depth [Stein and Kroeger, 1980; Forsyth, 1982]. A further drawback in our inversion [Nahelek, 1984] is that the number of time function elements and their durations are chosen a priori. This method, if not used with care, can act to predetermine the best fitting depth since the centroid depth and total source time function duration are closely related.

To address this problem, we first fix the centroid depth at each of a range of preassigned values and invert for the
remaining source parameters. At each depth, we try many different combinations (i.e., number and duration) of time function elements and choose the source time function that gives the smallest residual. At the same time, as noted above, during the iteration we constantly realign the synthetic and observed waveforms to improve the overall fit. Once we obtain the best solution at a fixed depth, we then relax all constraints in the inversion; further iteration usually yields the best overall solution in the neighborhood of the best fixed-depth solution. Following these steps we obtain a plot of residual versus depth; we estimate the uncertainty in the best fitting depth according to guidelines described in the next section. This overall procedure is analogous to the one commonly used (in the frequency domain) to estimate earthquake source depths from surface wave amplitude and phase spectra [e.g., Weidner and Aki, 1973; Aki and Patton, 1978; Romanowicz, 1981, 1982].

A number of other methods have been used to determine a best-fitting focal depth from body wave data, but they generally require prior knowledge of the focal mechanism. Forsyth [1982] determined the depth by deconvolving the windowed seismogram with a set of three delta functions representing the relative amplitudes of P, pP and sP. Delay times for the reflected phases were estimated from trial depths. He found that the complexity of the deconvolved signal varies with trial depth. Using the criterion of maximum compactness (or simplicity), the preferred focal depth is identified as that depth which gives the simplest source time
function. Wiens and Stein [1984] used a method that is very similar to our technique to determine focal depth. First the dip of one nodal plane is determined from P-wave first motions; then the strike and slip angles are estimated using Rayleigh wave data. After determining the focal mechanism, they modeled the P waveforms at several trial depths. The depth that gives the lowest error (defined as the sum of the squares of the observed minus the theoretical seismograms) was chosen as the depth for the event. The problem of trade-off between source time function and depth is handled by including models with a range of trial time function durations and depths [Stein and Kroeger, 1980]. Wiens and Stein [1984] and Engeln et al. [1985] used a variation of the deconvolution procedure of Kikuchi and Kanamori [1982] to determine the source time function that gives the smallest residual for a fixed depth. By repeating this process at a series of depths they obtained a plot of residual versus depth. As mentioned before, all these methods are based on prior knowledge of source mechanism. We consider that our procedure is less susceptible to bias because it does not require prior assumption of either the focal mechanism or the source time function.

Estimation of the Uncertainty in Centroid Depth

Numerical tests have shown that the inversion technique provides good estimates of basic centroidal parameters when azimuthal station coverage is adequate and noise contamination is not serious [Nabelek, 1984]. As described in later chapters, the formal error ($2\sigma$) is centroid depth is 0.1 to 0.6 km for the earthquakes in our study, but because of
bias introduced by factors not reflected in the estimate of a posteriori parameter variances, such as the trade-off between the duration of the source-time function and the centroid depth and the trade-off between source mechanism and centroid depth, the true uncertainty is greater. Strictly speaking, the actual error in a reported estimate, that is, the magnitude and sign of its deviation from the true solution, is usually unknowable. Limits to the error, however, can generally be inferred--with some risk of being incorrect--from the precision of the measurement process by which the reported value was obtained and from reasonable limits to the possible bias of the measurement process.

To estimate the uncertainty in the centroid depth we conducted a test of significance for the residuals between the best-fitting depth and those obtained when the inversion is performed with the depth fixed at nearby values. A test of significance is a rule for deciding whether to accept or reject the null hypothesis. We want to examine whether we would be justified in concluding that there is a difference between the fit of the best depth and other depths. A nonsignificant result does not prove that the null hypothesis is correct, merely that it is tenable, i.e., our data do not give adequate grounds for rejecting it.

Since we are dealing with time series, the data points we used in equation (9) are not independent. To get around this problem, we use the residuals at individual stations as our data instead of the residuals at discrete time samples. This
approach underestimates the number of degrees of freedom and will yield a very conservative result, but we are assured of independent data. The mean square residual for station $j$ is

$$r_j^2 = \frac{1}{M_j} \sum_{i=1}^{M_j} (s_{ij} - o_{ij})^2$$  \hspace{1cm} (11)

We investigated the distribution of individual station residuals $r_j^2$ with the $\chi^2$ test for goodness of fit and found that the residuals can generally be considered to follow a normal distribution. This is because the set of $r_j^2$ should follow a $\chi^2$ distribution, and as the number of degrees of freedom increases the $\chi^2$ distribution approaches a normal distribution [Huntsberger and Billingsley, 1981]. We also redefine the mean square residual for the model as

$$R^2 = \frac{1}{N} \sum_{j=1}^{N} r_j^2$$  \hspace{1cm} (12)

The quantities $r^2$ and $R^2$ in (9) and (12) would be equal only if all stations have the same inversion window length and weight. However, we cannot use $R^2$ as a measure of $\chi^2$ in an F-test of significance because the $r_j^2$ are not homogeneous [Steel and Torrie, 1982], that is, the amount of information differs from seismogram to seismogram. For example, depth information is mostly contained in the P waves while much of the constraint on focal mechanism is contained in the SH waves. Also, the information contained in seismograms of the same kind of wave differ because of the radiation pattern.
If heterogeneity of the $r_j^2$ samples is significant, then it may be wise to consider the individual station residuals rather than $R^2$ [Steel and Torrie, 1982]. Following this idea, we consider the difference in $r_j^2$ at each station for two different focal depths. That is, we consider that our experimental observations (residuals) are self-paired. If the members of the pair tend to be positively correlated, that is, if members of a pair tend to be large or small together, an increase in the ability of the experiment to detect a small difference is possible.

We now consider the matched station residuals at two depths $A$ and $B$ and the set of differences

$$d_j = r_{jA}^2 - r_{jB}^2 \quad j = 1, 2, \ldots, N \quad (13)$$

where $B$ is the depth at which the smallest average residual $R^2$ occurs. We denote the mean and standard deviation of the set of $d_j$ by $\mu_d$ and $\sigma_d$, respectively. The set of $d_j$ may be regarded as sampling a normally-distributed population of station residual differences with mean $\mu_d$ and standard deviation $\sigma_d$. The null hypothesis tested is that the mean of the population of differences is zero; the alternative is that the mean is larger than zero. The test statistic is

$$t = \frac{\mu_d}{\sigma_d/\sqrt{N}} \quad (14)$$

which follows the $t$-distribution with $N-1$ degrees of freedom. Since we are testing the hypothesis that the solution at one
depth is better than the other, we use the one-tailed (or one-sided) test.

We investigated the depth resolution of the inversion by considering the differences between residuals at the best-fitting depth and those at nearby depths, both shallower and deeper. Specifically, we determined the depths at which the t statistic exceeds the allowed value for a significance level of 0.1. That is, we have a probability of 0.1 to conclude that there is a difference where in fact there is none (Type I error, see Appendix A). The depth ranges determined in this manner for four ridge-axis earthquakes are shown in Figure 2.4, along with the mean square residuals $r^2$ and $R^2$ as functions of source depth. In general, the resolution of the centroid depth estimated by the t-test is consistent with a formal error of about 10 $\sigma_h$, where $\sigma_h$ is the standard error in the centroid depth $h$ determined in the waveform inversion.

The range in uncertainty about the best-fitting depth is not symmetric; it is usually smaller at the shallow end than at the deeper end. The uncertainty range tends to be smaller for larger earthquakes, probably because more stations are used for larger events, so that the focal mechanism is better constrained. It should be recalled that we have underestimated the number of degrees of freedom for all these tests in order to be assured of independent data.

Finally, it should be noted that statistical considerations are not the sole basis for drawing inferences; a physical appreciation of the problem and experience may also be brought
profitably into the judgment. To illustrate how details of waveform shapes which are poorly reflected in the mean square residual can be important clues in judging the success of a particular solution, we have plotted in Figure 2-5 the observed and synthetic P waveforms at 3 stations generated with the best-fitting source parameters at 5 depths for the Mid-Atlantic Ridge earthquake of June 6, 1972. For this earthquake, the plot of mean squared residual versus centroid depth (Figure 2.4) is very flat at depths greater than about 1.5 km, and the range of statistical uncertainty is large. On this basis alone, we would conclude that long-period waveform data place virtually no constraint on the depth of this event. On the other hand, the character of the first half-cycle of motion of the synthetic P waveforms in Figure 2.5 allows us to distinguish a centroid depth of 1.8 km as being superior to solutions even 1 km shallower or deeper. The solution at a depth of 1 km does not produce a large enough dilatation at CAR. The solution at 3 km produces too large a dilatation at AAE and AQU, and the width of the first pulse at CAR is too wide. The solutions at 4 and 5 km progressively aggravate these problems, although the overall residual error remains nearly constant. We conclude that the range in acceptable centroid depths determined in the statistical study is a good, although rather conservative, estimate of the true uncertainty in the depth estimated in the inversion.

The Effect of the Assumed Velocity Structure

The accuracy of source parameters depends on the velocity
structure of the source region. Since we do not have direct knowledge about the velocity structure at most earthquake sites, we use a generalized velocity model for all of our ridge-axis events, as noted earlier.

For most of the events in this study, we found that the mean square residual versus depth displayed two minima, one located in the crust and one just below the crust-mantle interface (Figure 2.4). The question of depth resolution therefore involves two separate issues. The first concerns the choice between the two minima in the relation for residual versus depth. The second issue is the estimation of uncertainty in the depth at which the preferred minimum occurs.

With the exception of one earthquake (October 7, 1981, on the Central Indian Ridge), the solution within the crust has a smaller residual than the sub-Moho solution and would therefore be the preferred solution. However, the other source parameters are similar for both depths and the difference in residuals is usually small. Because the deeper solution lies immediately below the crust-mantle interface in the source velocity structure for most earthquakes and occurs within about 2 km of the Moho for all events, we suspect that it is an artifact of the assumed sharp velocity contrast.

To explore the reasons for this apparent ambiguity in solutions, we show in Figure 2.6 several P and SH waveforms for the crustal and mantle solutions for the Reykjanes Ridge earthquake of September 20, 1969, which has sharp minima in the residual-versus-depth curve at 1.5 and 6.0 km (Figure 2.4).
Other than the centroid depth, all source parameters are similar for the two solutions. It may seem, at first glance, remarkable that the waveforms in Figure 2.6 are so similar for the solutions at the two centroid depths. That this is so is basically a result of the limited bandwidth of long-period teleseismic signals. Figure 2.7 shows elementary seismograms for teleseismic P waves predicted by the two solutions at Matsushiro (an elementary seismogram is a series of delta functions of relative amplitude and timing appropriate to the principal body wave phases, without the smoothing effect of instrument response, time function duration, and anelastic attenuation). The only notable difference between the solutions at 1.5 and 6.0 km depth is the delay time between the P and pP phases (about 0.5 and 2.0 sec for the two solutions, respectively). After the pP phase, the principal phases are the water reverberations, and for this section of the seismogram the two solutions are almost identical.

Distinguishing the solutions at the two alternative depths hinges entirely on our ability to identify the contribution to the signal from the free surface. When we include the effects of recording instrument, source time function, and anelastic attenuation, however, the broadening and coalescence of P and pP phases result in a significant deterioration of depth discrimination. For very shallow large earthquakes, we usually cannot identify the onset of the first motion to better than 1 sec on long-period records. A small difference between the arrival times of P and pP can therefore be taken up by
realignment of synthetic and observed waveforms. For the synthetic waveforms of the 1.5 km and 6.0 km solutions, including the effects of all transfer functions, the beginning of the inversion window which corresponds to the onset of the earthquake at MAT (Figure 2.7) differs by only 0.5 sec. For all seismograms used in the inversion for this earthquake, in fact, the difference between the beginning of the inversion windows for the solutions at the two centroid depths is less than or equal to 1 sec. Figure 2.8 shows comparative synthetic seismograms for the two solutions at five other stations. As can be seen, the waveforms for solutions at 1.5 and 6 km centroid depth are very similar, and if we cannot identify the onset of the first motion to better than 1 sec then choosing between the two solutions is difficult.

To investigate further the choice of solutions, we performed a series of inversions for the September 20, 1969 earthquake using a variety of crustal models. In our first series of tests we examined the effect of crustal thicknesses on the depths and relative magnitude of the two minima in $r^2$. For these tests, we kept the crustal and mantle velocities constant and varied the crustal thickness. For a halfspace model with crustal P and S velocities there is only one minimum, at 1.4 km below the sea floor. When a crustal layer greater than 1.5 km in thickness is assumed for the source structure, a stable minimum in $r^2$ develops at about 1.5 km depth and the second minimum appears immediately below the crust-mantle interface. With a progressively thicker crustal layer this second minimum
moves with the layer boundary. These results suggest that the sharp velocity contrast at the Moho acts to reflect energy downward in a manner similar to the water-crust boundary. For a source located immediately below the Moho, the upgoing P wave reflected downward at the Moho has the same arrival time as direct P but opposite polarity. This effect makes the amplitude of the first P arrival smaller, and the seismogram mimics that of a shallower crustal source. The effect is largest when this destructive interference is maximum, that is, when the focus is just below the Moho; as the earthquake focus moves deeper, the effect diminishes and the residual becomes larger.

To test this hypothesis, we repeated the inversions with a velocity model which has an additional layer in the lower crust to simulate a gradual increase in velocity across the crust-mantle transition. The sub-Moho solution becomes significantly worse than the best crustal solution as the velocity contrast at the Moho proper is reduced. For the 2-layer crustal model, an additional local minimum also appears just below the interface between the crustal layers.

In principle, short-period records can be used to distinguish the crustal and sub-Moho solutions indicated by the inversion of long-period waveforms. As an illustrative example, we consider the June 2, 1965 Mid-Atlantic Ridge earthquake. The long-period body waveform inversion shows two minima in the residual-versus depth curve (see Chapter 3 and Appendix B), similar to the double minima shown in Figure 2.4. The best-fitting solution is at 3 km depth; a secondary solution occurs
immediately below the Moho. The synthetic long-period P and SH waveforms are very similar for the two solutions, and our statistical test cannot distinguish them. Figure 2.9 compares P waveforms for this earthquake recorded on four short-period vertical WWSSN instruments with synthetic short-period waveforms generated for the solutions at 3 km and 6 km centroid depth. Because the absolute amplitudes of the short-period waveforms appear to be controlled by path effects, only the shapes of the waveforms are used for comparison. Figure 2.9 shows that the 3-km-deep solution is preferable because all phases before the first water-surface reflection pwP are in good agreement with the observed waveforms. For the 6-km-deep sub-Moho solution, the pwP phase in the synthetic waveform is late by about 1 sec with respect to the observed seismogram. This discrepancy is particularly clear at stations SJG, WIN, and PEL. The ability of the short-period waveforms to discriminate between the two solutions is due to greater bandwidth in the recorded signals; the beginning of first motion can be resolved to better than a few tenths of a second.

From the above analysis, we infer that the solution immediately below the Moho indicated by the inversion of long-period body waveforms from the ridge-axis earthquakes of this study is an artifact of the sharp velocity discontinuity at the crust-mantle transition in the assumed source velocity model. It is quite probable that our simple source structure overestimates the Moho velocity contrast beneath a slowly-
spreading ridge [e.g., Purdy and Detrick, 1985]. We conclude that the shallow crustal solution is stable for any reasonable crustal thickness and that it probably represents the true solution for these earthquakes. Tarantola and Valette [1982b] also noted the existence of secondary maximum-likelihood solutions at interfaces between layers of constant velocity. This is because the discontinuities in velocity cause the probability density functions $\sigma(z)$ to have discontinuities in slope, where $\sigma(z)$ is the probability that the true solution lies within a small depth interval $dz$ at depth $z$.

Other source parameters were consistent for all of the different structural models we have tested, including the halfspace models and the model with 2 crustal layers. The seismic moment varied by less than 20%, strike and dip varied by less than 10°, and slip varied by less than 20°. The results of the inversions with different velocity structures are shown in Table 2.1.

The second part of our test compares the inversion results using our generalized ridge structure with that using a known ridge velocity structure. This comparison can give us an idea about any bias introduced by using a simplified velocity model. A refraction experiment [Purdy and Detrick, 1985] was conducted along a section of Mid-Atlantic Ridge median valley near 23°N where several earthquakes of this study (June 3, 1962 and the June 28, 1977 swarm). Figure 2.7 shows the compressional velocity versus depth derived by Purdy and Detrick [1985] as well as the simplified models with a uniform crust and 3
homogeneous crustal layers. The shear velocities used in the
inversions are estimated by assuming a Poisson's ratio of
\( \nu = 0.25 \).

Figure 2.8 shows the residual versus depth curves for two
different velocity structures for the largest earthquake in the
1977 swarm. For the model with a uniform crust, there are two
clearly defined local minima, one within the crust at 1.6 km
beneath the sea floor and the other immediately beneath the
Moho. The uncertainty in the depth of the shallow solution
spans a depth range from 1 to 2 km. The local minimum below the
Moho lies outside the uncertainty range. For the model with the
3-layer crustal structure, the residual curve is generally
flatter. There are three local minima: the global minimum
occurs in the crust at 2 km below the sea floor, the second
minimum lies just beneath the interface between the second and
third crustal layers, and the third minimum lies just beneath
the Moho. Since the residual curve is flatter, the uncertainty
range for centroid depth is greater (1.2 to 3 km and 4.4 km to
5 km depth). This analysis indicates that the depth of the
global minimum probably is generally insensitive to the details
of the velocity structure. The uncertainty range for the best
fitting centroid depth, however, is probably underestimated when
we use a simple velocity model for the source region.

The Effect of Incomplete Azimuthal Coverage

Ideally in a body waveform inversion study, the waveforms
to be inverted sample fully the lower focal hemispheres for both
P and SH waves. All too often, however, the sampling by
available seismic stations is incomplete, particularly in azimuth. It is therefore important to estimate the effects of incomplete coverage on the estimation and resolution of source parameters. We chose the September 20, 1969 Reykjanes Ridge earthquake to investigate these effects. This earthquake is the largest ridge axis earthquake we have studied, and the signal-to-noise ratio for all observed seismograms is quite good. We vary the azimuthal coverage by selective removal of data, and we recompute the residual versus depth curve. The residual versus depth curves using all 24 P and SH waveforms appears in Figure 2.9 and is used as the standard result for comparison. The focal sphere coverage for the full data set is also shown in the figure.

4-quadrant coverage. In the first test, we omitted some stations that are close to other stations but we still retain full azimuthal coverage. The station coverage for both P and SH waves is shown in Figure 2.10; also shown in the same figure is the residual-versus-depth curve. The best-fit model parameters are almost identical to the results obtained using all 24 waveforms. The strike, dip and slip angles differ by less than 2°. The best centroid depth at 1.5 km is the same as in the full solution. The uncertainty ranges for centroid depths are also identical for both data sets.

3-quadrant coverage. In this test, we eliminated data from stations in the first quadrant for both P and SH waves. The station coverage and its residual-versus-depth curve are shown in Figure 2.11. All centroid parameters are reasonably well
determined. The depth is overestimated by 0.2 km, and double couple angles differ by less than 7° from the standard result. However, the uncertainty range for centroid depth (1-4 km and 6-9 km) is much larger than in the full solution. This is mostly because the strike angle is less well constrained, varying from 18° to 38° compared with the range 22° to 29° for the standard solution.

2-quadrant P wave coverage. In this test, we eliminated P-wave data from stations in the first and second quadrants, but we retain all SH wave data. The station coverage and its residual curve are shown in Figure 2.12. The estimated best-fit model parameters for this coverage are quite different from the standard result. The strike and slip angles are overestimated by about 15°, while the dip angle is about the same. The best centroid depth is at 6 km (just below the Moho) compared with the standard result of 1.5 km. The uncertainty range for centroid depth is 0.5 to 10 km. This large range is due to a significant trade-off between source mechanism and focal depth. The second minimum in the residual curve occurs at 1.5 km, the best-fitting centroid depth for the standard solution.

The above analysis underscores the importance of good azimuthal coverage. As long as the azimuthal coverage includes 3 or 4 quadrants, the best-fit model parameters are very stable. For lesser azimuthal coverage, however, the best-fitting solution can be significantly different from the actual source. As we might have expected, the uncertainty range for centroid depth increases as the azimuthal coverage decreases. Based on
our inference that the minimum beneath the Moho is an artifact of the velocity discontinuity, then the best-fitting solutions (crustal) obtained under different degrees of azimuthal coverages are all quite similar. This internal consistency gives us confidence in the results presented in the chapters to follow.

COMPARISON WITH SURFACE WAVE METHODS

Two Mid-Atlantic Ridge earthquakes (June 2, 1965 and November 16, 1965) included in our study were also examined in detail by Weidner and Aki [1973] using the phase and amplitude spectra of Rayleigh waves. The body wave inversion results for these two earthquakes are summarized in Chapter 3. These two earthquakes give us an opportunity to compare our estimated depth uncertainty with that obtained by a detailed surface wave analysis. For both of these earthquakes, our inversion indicates a possible solution immediately beneath the Moho. If we exclude the mantle solutions, for reasons described above, the centroid depth (3 km) and depth resolution (± 2 km) for the June 2, 1965 earthquake are identical using these two methods. For the November 16, 1965 event, our body waveform inversion result suggests a slightly shallower centroid depth (2.1 km) and a slightly better depth resolution (± 1 km) than that indicated by Weidner and Aki [1973] (3 ± 2 km).

We consider that the depth resolution using body waveform inversion and that obtainable by surface wave analysis are comparable when the source radiation pattern is well sampled and signal-to-noise ratios are high. Both of these methods, it should be noted, are extremely time-consuming. The surface wave
method requires a great deal of digitizing and computer processing, and the body wave method requires testing of many different source time functions. In addition, the surface wave method usually requires precise information on phase velocity along the ray path, or in the region between nearby sources, whereas body waveform inversion does not need such additional information. For sufficiently large events, we prefer the body wave method for its relatively more compact data sets and robustness with respect to source velocity structure. However, since surface waves are the most energetic phases for shallow events, surface wave analysis sometimes is the only feasible method for smaller events when azimuthal coverage for body waves is inadequate [e.g., Kafka and Weidner, 1981].
Table 2.1
Results of Inversions for the Centroidal Parameters of the September 20, 1969 Reykjanes Ridge Earthquake

<table>
<thead>
<tr>
<th>Model no.</th>
<th>Crustal thickness, km</th>
<th>Centroid depth (crustal solution)</th>
<th>Centroid depth (mantle solution)</th>
<th>Seismic moment, $10^{25}$ dyn-cm</th>
<th>Strike/dip/slip</th>
<th>Residual, mm²</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$\infty$</td>
<td>1.40</td>
<td>----</td>
<td>13.5</td>
<td>20/42/258</td>
<td>10.9</td>
<td>Crustal halfspace</td>
</tr>
<tr>
<td>2</td>
<td>1</td>
<td>0.99</td>
<td>2.77</td>
<td>16.1</td>
<td>30/41/266</td>
<td>13.3</td>
<td>1.77 km below Moho</td>
</tr>
<tr>
<td>3</td>
<td>2</td>
<td>1.41</td>
<td>2.82</td>
<td>15.6</td>
<td>21/41/257</td>
<td>10.8</td>
<td>0.82 km below Moho</td>
</tr>
<tr>
<td>4</td>
<td>3</td>
<td>1.44</td>
<td>3.15</td>
<td>16.0</td>
<td>25/39/263</td>
<td>11.4</td>
<td>0.15 km below Moho</td>
</tr>
<tr>
<td>5</td>
<td>4</td>
<td>1.45</td>
<td>4.02</td>
<td>15.7</td>
<td>25/37/261</td>
<td>10.7</td>
<td>0.02 km below Moho</td>
</tr>
<tr>
<td>6</td>
<td>5</td>
<td>1.45</td>
<td>5.01</td>
<td>15.5</td>
<td>21/43/260</td>
<td>10.1</td>
<td>0.01 km below Moho</td>
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<tr>
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<td>1.50</td>
<td>6.00</td>
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<tr>
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<td>0</td>
<td>----</td>
<td>3.09</td>
<td>16.4</td>
<td>24/42/260</td>
<td>11.9</td>
<td>Mantle halfspace</td>
</tr>
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<td>4.07*</td>
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<td>25/44/263</td>
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<td>6.02</td>
<td>16.4</td>
<td>24/42/260</td>
<td>11.9</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Notes to Table 2.1

Model 1 is a crustal half-space; models 2-7 include a single layer crust over a mantle half-space; model 8 is a mantle half-space, and model 9 includes a two-layer crust over a mantle half space.

For model 1-8, the seismic velocities are

Crust: \( \alpha = 6.4 \text{ km/sec}, \beta = 3.7 \text{ km/sec} \)
Mantle: \( \alpha = 8.1 \text{ km/sec}, \beta = 4.6 \text{ km/sec} \).

For model 9,

Layer 1: \( \alpha = 6.4 \text{ km/sec}, \beta = 3.7 \text{ km/sec}, \text{ thickness} = 4 \text{ km} \)
Layer 2: \( \alpha = 7.2 \text{ km/sec}, \beta = 4.2 \text{ km/sec}, \text{ thickness} = 2 \text{ km} \)
Mantle: \( \alpha = 8.1 \text{ km/sec}, \beta = 4.6 \text{ km/sec} \).

* A second local minimum occurs in the crust, immediately below the interface between the two crustal layers.
FIGURE CAPTIONS

Figure 2.1. Comparison of observed (solid line) and synthetic (dashed line) long-period P and SH waves for the June 28, 1972, northern Red Sea earthquake, with the focal mechanism solution obtained from Pearce [1977] plotted on the lower focal hemisphere (equal-area projection). The seismic moment, centroid depth and source time function are obtained from the inversion of P waveforms with the double-couple orientation fixed at values given by Pearce [1977]. All amplitudes are normalized to an instrument magnification of 3000; the amplitude scales correspond to the waveforms that would be observed on an original seismogram for such an instrument. The two vertical lines in the P waves delimit the portion of each time series used in the inversion. Symbols for both types of waves are open circle, dilatation; solid circle, compression; cross, emergent arrival. For SH waves, compression corresponds to positive motion as defined by Aki and Richards [1980]. The source time function obtained from the inversion is also shown.

Figure 2.2. Comparison of the observed and synthetic P and SH waves for the June 28, 1972 event, with the focal mechanism solution obtained from the inversion of P waveforms without an imposed constraint on the source mechanism. See Figure 2.1 for further explanation.

Figure 2.3. Comparison of the observed and synthetic P and SH waves for the June 28, 1972 event, with the focal mechanism
solution obtained from the unconstrained inversion of both P and SH waveforms. See Figure 2.1 for further explanation.

Figure 2.4. Mean squared residual versus centroid depth for four earthquakes in this study. The solid and dashed lines indicate residuals $r^2$ and $R^2$, defined in equations (9) and (12), respectively. The horizontal bar indicates the 90% confidence interval about the best-fitting depth found in the body-waveform inversion, which is indicated by an arrow.

Figure 2.5. Comparison of observed (solid) and synthetic (dashed) P waveforms at 3 stations for the Mid-Atlantic Ridge earthquake of June 6, 1972. The synthetic seismograms are generated from the best-fitting solution obtained with the centroid depth fixed at the value indicated. Additional waveforms not shown (see Figure 3.16) are used in the inversions. The preferred centroid depth is 1.8 km. Vertical tic marks delimit the portions of the waveforms included in the inversions.

Figure 2.6. Comparison of the crustal (1.5 km centroid depth) and mantle (6.0 km centroid depth) solutions for the Reykjanes Ridge earthquake of September 20, 1969. The preferred centroid depth is 1.5 km. Vertical tic marks delimit the portions of the waveforms included in the inversions.

Figure 2.7. Comparison of synthetic and observed P waveforms at MAT for the best-fitting solutions at 1.5 km and 6 km centroid depth for the Reykjanes Ridge earthquake of September 20, 1969. At the top are the relative amplitudes of the different phases predicted by the radiation pattern
and respective time delays; subsequent traces show the effect of convolution with the instrument response of a WWSSN long-period seismometer, with the source time function (shown on the bottom trace), and with the attenuation operator ($\delta t^* = 1$ sec).

Figure 2.8. Comparison of synthetic P waveforms at 3 stations and SH waveforms at 2 stations for the two alternative solutions for the Reykjanes Ridge earthquake of September 20, 1969. Also shown for each station and solution are elementary seismograms depicting the relative amplitudes and arrival times of different phases predicted by the radiation pattern and source depth.

Figure 2.9. Comparison of observed (solid) and synthetic (dashed) short-period P waveforms at 4 stations for the Mid-Atlantic Ridge earthquake of June 2, 1965. The synthetic short-period seismograms are generated from the best-fitting solution (3.0 km centroid depth) and the secondary solution (6.0 km centroid depth) obtained from long-period waveform inversion. The preferred centroid depth is 3.0 km.

Figure 2.10. Models for compressional velocity versus depth in the source region of ridge-crest earthquakes. The solid line shows the velocity-depth function obtained from refraction data along the median valley of the Mid-Atlantic Ridge near 23°N [Purdy and Detrick, 1985]. The dashed line shows the model with a homogeneous crustal layer overlying a uniform mantle assumed in the calculation of most synthetic seismograms. A model with three crustal layers, used for
testing the bias introduced by the uniform crust model, is also shown. See text for a discussion of the effects of varying the assumed velocity model on centroid depth determination.

Figure 2.11. Comparison of the mean-squared residual $R^2$ versus depth for the June 28, 1977, earthquake using two different velocity structures (Figure 2.7). The horizontal bar indicates the 90% confidence interval about the best-fitting depth found in the body-waveform inversion, as indicated by an arrow. The vertical dashed lines indicate the assumed layer boundary.

Figure 2.12. The mean-squared residual $R^2$ versus centroid depth for the September 20, 1969 earthquake using all stations. The station coverage and focal mechanism for P and SH waves are also shown. The residual at each depth is normalized by the mean-squared waveform over the same time window. With this normalization, a perfect fit would have a residual of zero and a source with zero seismic moment would have a residual of unity. The horizontal bars indicate the 90% confidence interval about the best-fitting depth found in the body-waveform inversion, indicated by an arrow.

Figure 2.13. The mean-squared residual $R^2$ versus centroid depth for the September 20, 1969 earthquake when some stations are omitted but full azimuthal coverage is still retained. See Figure 2.9 for further explanation.

Figure 2.14. The mean-squared residual $R^2$ versus centroid depth for the September 20, 1969 earthquake when the data from
stations in the first quadrant for both P and SH waves are eliminated. See Figure 2.9 for further explanation.

Figure 2.15. The mean-squared residual $R^2$ versus centroid depth for the September 20, 1969 earthquake when the P-wave data from stations in the first and second quadrants are eliminated. See Figure 2.9 for further explanation.
Figure 2.1
Figure 2.3
Figure 2.4
Figure 2.5
## SEPTEMBER 20, 1969

<table>
<thead>
<tr>
<th>1.5 km</th>
<th>6 km</th>
<th>1.5 km</th>
<th>6 km</th>
</tr>
</thead>
<tbody>
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<td><img src="image1.png" alt="Waveform 1" /></td>
<td>MAT P</td>
<td><img src="image2.png" alt="Waveform 1" /></td>
<td>SDB P</td>
</tr>
<tr>
<td><img src="image3.png" alt="Waveform 2" /></td>
<td>SHL P</td>
<td><img src="image4.png" alt="Waveform 2" /></td>
<td>NAT P</td>
</tr>
<tr>
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<td>NIL P</td>
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<td>SJG P</td>
</tr>
<tr>
<td><img src="image7.png" alt="Waveform 4" /></td>
<td>TAB P</td>
<td><img src="image8.png" alt="Waveform 4" /></td>
<td>BEC P</td>
</tr>
<tr>
<td><img src="image9.png" alt="Waveform 5" /></td>
<td>TRI P</td>
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</tr>
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<td><img src="image11.png" alt="Waveform 6" /></td>
<td>NAI P</td>
<td><img src="image12.png" alt="Waveform 6" /></td>
<td>AAM P</td>
</tr>
</tbody>
</table>

**Figure 2.6a**
Figure 2.6b
Figure 2.7
Figure 2.8
Figure 2.9
Figure 2.10
Figure 2.11

Centroid Depth, km

Uniform Crust

Three-layer Crust

\[ R^2 \]

\[ 0 \rightarrow 0.4 \]

\[ 0 \rightarrow 0.2 \]

\[ 0 \rightarrow 10 \]
SEPTEMBER 20, 1969
All stations

Figure 2.12

Centroid Depth, km

Figure 2.12
SEPTEMBER 20, 1969
Selected 4 quadrant coverage

Figure 2.13
SEPTEMBER 20, 1969
3 quadrant coverage

Figure 2.14
SEPTEMBER 20, 1969
2 quadrant P-wave coverage

Figure 2.15
CHAPTER 3. RIDGE-AXIS EARTHQUAKES IN THE
NORTH ATLANTIC OCEAN

INTRODUCTION

In this chapter we determine the focal depths and source mechanisms of 16 normal-faulting earthquakes in the North Atlantic Ocean. Fourteen earthquakes occurred along the Mid-Atlantic Ridge, and 2 earthquakes occurred along the spreading portion of the Azores-Gibraltar plate boundary. The source mechanisms of these 16 ridge-axis earthquakes are determined by the body-waveform inversion technique described in Chapter 2. All events show fault planes dipping about 45° and striking parallel to the local trend of the ridge axis. The water depths estimated from the ringing part of the P wave-trains indicate that the ridge earthquakes discussed in this chapter all occurred beneath the inner floor of the median valley. The centroid depths of these earthquakes are all very shallow, and calculated depth uncertainty ranges for these events are generally small. The similarities in source mechanisms and centroid depths may be a reflection of the relatively uniform tectonic setting along this ocean ridge.

THE MID-OCEAN RIDGE SYSTEM IN THE NORTH ATLANTIC

The present-day structure of the seafloor in the north Atlantic is dominated by the Mid-Atlantic Ridge and the four major plates (Eurasian, North American, South American and African) that meet along this boundary. The area of study (Figure 3.1) includes the Mid-Atlantic Ridge from the equator
to the Mohns Ridge as well as the western part of the Azores-Gibraltar seismic zone marking the western boundary between the Eurasian and African plates.

The Mid-Atlantic Ridge nearly bisects the North Atlantic basin. The ridge is slowly spreading (10-18 mm/yr half rate), and it has rugged, block-faulted topography with numerous prominent fracture zones. By analyzing the relationship between bathymetry and gravity, McKenzie and Bowin [1976] have shown that isostatic compensation in the Atlantic Ocean begins at a wavelength of about 100 km and increases with increasing wavelength. They interpreted this relationship in terms of the flexural support of topographic relief produced near the ridge axis by an elastic plate about 10 km in thickness. A later analysis by Cochran [1979] yielded a similar effective thickness (7-13 km) for the elastic lithosphere near most segments of the Mid-Atlantic Ridge except for the Reykjanes Ridge at 60°-62°N, where the effective thickness must be in the range 2-6 km. The extent to which these estimates are measures of the thickness of the zone of elastic or brittle behavior at the rise axis proper, however, is uncertain. Macdonald [1977] estimated that the width of the crustal accretion zone at 37°N on the Mid-Atlantic Ridge is between 1 and 8 km, and he suggested that the crustal accretion zone at slow spreading ridges may vary with time and space. The magnetic anomalies from the Atlantic are generally not as clear and sharp as those from the faster spreading East Pacific Rise, perhaps an indication that crustal accretion at slow spreading rates is not continuous in time and space [Macdonald, 1982].
There have been numerous seismic refraction studies along the Mid-Atlantic Ridge. The structure beneath the median valley, however, has been well resolved only in three small areas at $37^\circ$, $45^\circ$, and $60^\circ$N [Purdy and Ewing, 1985]. The main results of these studies were that: (1) No evidence for an axial crustal magma chamber was found [Whitmarsh, 1975; Fowler, 1976]. However, it is possible that narrow magma chambers do exist but could not be detected because of the limited power of the conventional refraction experiment [Fowler, 1978]. (2) The formation of a normal, 6-to-7-km-thick crustal section, with a 7.2 km/sec basal layer and an 8.1 km/sec upper mantle, occurs within 10 km of the ridge axis [Fowler, 1976, 1978; Fowler and Keen, 1979].

Purdy and Ewing [1985] and Macdonald [1982] pointed out that most of these experiments were conducted on spreading segments that are too short to resolve deep crustal structure. In contrast, a recent experiment [Purdy and Detrick, 1985] was performed on a long ridge segment at $23^\circ$N, and it was found that the oceanic crust appears mature, with a well-defined Moho, upper mantle velocities of ~8 km/sec, and a total crustal thickness of 6-7 km. The ocean crust was interpreted to be in a cooled state between episodes of major magmatic injection. Two well-constrained refraction studies [Fowler, 1976; Bunch and Kennett, 1980] suggest the presence of low-velocity mantle material beneath the Mid-Atlantic Ridge median valley. However, no strong evidence exists to suggest that low-velocity upper mantle is a ubiquitous characteristic of axial valley segments.
of slow spreading ridges [Purdy and Ewing, 1985].

Two of the earthquakes studied in this chapter occurred along the Azores spreading center. The Azores region is the site of the triple junction between the North American, Eurasian, and African plates. East of the Azores, the Eurasia-Africa plate boundary is dominantly a transform fault called the Gloria fault [Laughton et al., 1972; Laughton and Whitmarsh, 1974]. At its western end, the Eurasia-Africa plate boundary is a spreading center with a very slow rate of opening [Krause and Watkins, 1970; McKenzie, 1972; Searle, 1980]. There are several hypotheses concerning the evolution of the Azores spreading center. (1) The triple junction began as a ridge-fault-fault (RFF) junction and changed to the present ridge-ridge-ridge (RRR) junction following a change of spreading direction from E-W to ESE-WNW [Krause and Watkins, 1970]. (2) The present triple junction (RRR) has been stable for an extended period [McKenzie, 1972]. (3) Short segments of the Mid-Atlantic Ridge are displaced to the east and are connected to the main ridge by several ESE-WNW leaky transforms [Machado et al., 1972]. (4) The relative motion of Africa with respect to Europe changed several times from ENE to WSW to SW then back to WSW during the last 72 M.y. [Searle, 1980]. According to this view, the Azores spreading center was probably initiated around 36 M.y. B.P. when the relative motion last changed direction. In this scenario, the triple junction was probably RFF during most of its history and then suddenly, rather than gradually, moved to the north.
The Azores plateau appears to have been formed by the Azores hot spot [Schilling, 1975; Schilling et al., 1977]. Using Rayleigh wave dispersion data, Searle [1976] found that the crust of the Azores Plateau is about 60% thicker than the surrounding oceanic crust and that uppermost mantle velocities are anomalously low. Most of the thickening seems to have taken place in layer 2, probably due to increased production of basaltic lava. The magnetic data in the Azores region show a N120°E trend, similar to that of the tectonic lineaments. Recent seismic and volcanic activity suggests that the present-day spreading center may run from the Mid-Atlantic Ridge at 38°40'N, almost due east to Terceira, and then down the Terceira Rift to the Gloria fault [Searle, 1980]. The Terceira Rift consists of a series of basins and ridges [Machado, 1959]. The basins have well-developed normal fault scarps trending N142°-154°E, and square cross sections unlike the V-shaped Mid-Atlantic Ridge median valley. The ridges also have normal faults but with smaller amplitudes. It is not clear whether the spreading center runs through the basins or ridges or both [Searle, 1980]. Lachenbruch [1973] suggested that a very slow-spreading ridge like the Azores should also have a median valley. On the other hand, Sleep and Rosendahl [1980] argued that a spreading center near a hot spot may not have a median valley; it may even have axial high. It is possible that the basins were formed first by rifting of oceanic lithosphere. When the spreading center evolved to hot spot ridge structure, axial volcanic ridges flanking the rifted basins may have developed [Searle, 1980].
EARTHQUAKE DATA SET

We began our study with a search for earthquakes on the Mid-Atlantic Ridge large enough to yield clear long-period P and S waves at teleseismic distances and having epicenters within the median valley of an active ridge segment. The search included the catalog of Rothe [1969] for 1962-1965, the bulletins of the International Seismological Centre (ISC) for the years 1964-1982, the monthly listings of the National Earthquake Information Service (NEIS) for the years 1982-1983, and published studies of fault plane solutions for Mid-Atlantic Ridge earthquakes [Sykes, 1967, 1970; Solomon and Julian, 1974; Udias et al., 1976; Einarsson, 1979; Savostin and Karasik, 1981; Dziewonski et al., 1983a,b]. Particular care was taken to exclude near-ridge intraplate events [Bergman and Solomon, 1984]. We restricted the search to the northern Mid-Atlantic Ridge, between latitudes 0° and 80°N, primarily to ensure a good azimuthal distribution of teleseismic stations.

The 16 selected earthquakes are listed in Table 3.1; their epicenters are shown in Figure 3.1. For all of the events in the table, we have found at least 10 clear P and SH waves recorded on long-period instruments at stations of the World Wide Standard Seismograph Network (WWSSN) or the Global Digital Seismic Network (GDSN) at epicentral distances between 30° and 80°. The events range between latitudes 0.4° and 72.2°N and have epicenters within median valley segments on bathymetric maps of the GEBCO (General Bathymetric Chart of the Oceans) series [Laughton and Monahan, 1978; Heezen and Tharp, 1978;
Johnson et al., 1979; Searle et al., 1982]. The local half-spreading rates vary from 0.8 to 1.8 mm/yr according to the plate kinematic model of Minster and Jordan [1978]. The earthquakes have body-wave magnitudes $m_b$ between 5.1 and 5.9, and, as demonstrated below, all have focal mechanisms characterized by normal faulting. We believe that this list is exhaustive, given our selection criteria, for the period 1962 through mid-1983.

**WAVEFORM INVERSION RESULTS**

In this section, we present the source parameters obtained from the inversion of P and SH waveforms for the 16 ridge-axis earthquakes. The best-fitting double-couple orientation, seismic moment, and centroid depth are given in Table 3.2. The convention for describing double-couple orientation is that of Aki and Richards [1980]. Centroid depths are given relative to the top of the crust. Also listed in Table 3.2 is the average water depth above each hypocenter, obtained by matching the synthetic and observed seismograms in the ringing portion of the P wave train. All waveform inversions were performed with the same source structure: a water layer of variable thickness, a single crustal layer of thickness 6 km and seismic velocities $V_p = 6.4$ km/s and $V_s = 3.7$ km/s, and a mantle half-space with $V_p = 8.1$ km/s and $V_s = 4.6$ km/s. For the July 4, 1966, Azores region earthquake, we also examined the influence of uncertainty in structure on inversion solutions; details are given below in the discussion of this earthquake.
June 3, 1962

The earthquake of June 3, 1962 occurred in the Mid-Atlantic Ridge median valley about 100 km south of the Kane Fracture Zone (Figure 3.2). Because this event predates the establishment of many WWSSN stations, station coverage is rather poor and resolution of the focal mechanism is less than ideal. The inversion solution (4/43/274) represents nearly pure dip-slip motion (Figure 3.3) with a seismic moment of $6.1 \times 10^{24}$ dyn-cm. The centroid depth of 1.2 km is quite well-constrained. The 3.9 km water depth estimated from the predominant period of water reverberations in the P waveforms is slightly greater than the depth of the median valley inner floor at the nominal latitude of the epicenter.

Because of the poor station coverage, there is significant trade-off between the centroid depth and the focal mechanism. A second minimum in the residual-depth curve occurs at 6 km (immediately below the Moho discontinuity). The focal mechanism (-4/48/258) for the mantle solution has a larger strike-slip component than that of the best solution at 1.2 km, but the seismic moment of $6.1 \times 10^{24}$ dyn-cm is the same as the best solution. The source time function for the mantle solution is about 2.4 sec in duration compared with 3 sec for the best solution. For both solutions, all observed waveforms are well-fit. Using the t-test, we estimate the range of acceptable centroid depths to be from 1-2 km to 6-7 km for a significance level $\alpha$ of 0.1 (Appendix B). To simulate a more graduated Moho transition and to test the stability of the mantle solution, we
repeated the inversion with a source model in which the lowest 2 km of crust has a P wave velocity of 7.1 km/s. We found that the mantle solution is degraded with such a source structure, but it is still well within the uncertainty range with $\alpha = 0.1$. Since the residuals for the two solutions are similar, the test power ($1-\beta$, see Appendix A) is small ($\sim 0.1$), implying that the statistical test in this case cannot distinguish these two solutions without more data. On the basis of the analysis in Chapter 2, however, we believe that the deeper minimum is an artifact of inadequately modeled crustal structure and poor sampling of the focal sphere.

June 2, 1965

The June 2, 1965 earthquake, which occurred in the median valley about 50 km north of the 15°20' Fracture Zone (Figure 3.4), has been the subject of several previous studies. On the basis of P-wave first motion polarities, Sykes [1970] found a normal faulting mechanism with non-orthogonal nodal planes. Tsai [1969] estimated the seismic moment to be $2.8 \times 10^{25}$ dyn-cm from the Rayleigh wave amplitude spectrum at one station, but this value was based on an adopted focal depth of 65 km. Using both the amplitude and phase spectra of Rayleigh waves from a number of stations, Weidner and Aki [1973] demonstrated that the focus is actually quite shallow ($3 \pm 2$ km). They also obtained a best-fitting double couple ($0/40/260$) and a seismic moment of $7.8 \times 10^{24}$ dyn-cm. From the identification of pP on short-period P waveforms, Duschenes and Solomon (1977) estimated a focal depth of $2 \pm 1$ km.
The body-wave inversion solution (335/38/244) is shown in Figure 3.5. The SH waves provide exceptional coverage, and all of the observed waveforms are well fit. The westward-dipping nodal plane has a strike slightly east of north, parallel to the local trend of the ridge axis. A small component of strike-slip motion is required in order to fit the SH waveforms to the northeast, particularly GDH and KTG, where the small dilatation of the direct SH phase is clearly seen. The best-fitting seismic moment is $8.3 \times 10^{24}$ dyn-cm, and the centroid depth is 3.0 km. Both values agree quite well with those of Weidner and Aki [1973]. The water depth inferred from the P waveforms is 3.4 km.

Although the azimuthal coverage for this equation is quite good, there is still some trade-off between source mechanism and centroid depth. The range in acceptable centroid depth at $\alpha = 0.1$ is 1-5 km and 6-8 km.

A secondary solution occurs at 6 km centroid depth, immediately below the Moho. This mantle solution (331/40/237) is very similar to the best-fitting solution; however, some seismograms give a slightly worse fit than for the solution at 3 km depth. For example, the first half cycle of the synthetic P wave at LPB and of the synthetic SH waveform at SHA are too wide. Also the fit to the water reverberations is worse for the mantle solution because the predicted amplitudes are somewhat smaller. On statistical grounds, both solutions are quite acceptable. When a more gradual Moho transition is assumed for the velocity structure at the source, the mantle solution deteriorates and can be rejected at $\alpha = 0.1$. The estimated
depth uncertainty for the preferred shallow solution in our study is the same as that of Weidner and Aki [1973] but greater than the ± 1 km determined by Duschenes and Solomon [1977]. We have to remember that our depth estimate is the centroid depth and not necessarily the point of rupture as seen using the short-period waveforms.

November 16, 1965

The earthquake of November 16, 1965 occurred in the median valley about 100 km north of the Atlantis Fracture Zone (Figure 3.6). From P-wave first motions, Sykes [1967] found a normal-faulting mechanism with non-orthogonal nodal planes. Tsai [1969] estimated the seismic moment to be $5.2 \times 10^{25}$ dyn-cm from the Rayleigh wave amplitude spectrum at one station, based on an assumed focal depth of 45 km. Weidner and Aki [1973] estimated a focal depth of $3 \pm 2$ km using both amplitude and phase spectra of Rayleigh waves. They also reported a best-fitting double couple of 0/59/234 and a seismic moment of $1.2 \times 10^{25}$ dyn-cm. From an identification of the short-period pP phase, Duschenes and Solomon [1977] obtained a focal depth of $4 \pm 2$ km.

The inversion solution (Figure 3.7) shows nearly pure normal faulting (25/46/265) on a fault plane whose strike is much closer to the trend of the ridge axis than the mechanism of Weidner and Aki [1973]. The distribution of P and SH waves provides a good sampling of the focal spheres. In particular, the nodal SH waveforms at MDS and LPB constrain the strike of the double couple. The seismic moment is $8.1 \times 10^{24}$ dyn-cm, the centroid depth is 2.1 km and well-constrained, and the depth
uncertainty is only \( \pm 1 \) km. The inferred water depth is 3.2 km. A secondary solution is located just below the Moho, but this depth is outside the range of acceptable values at \( a = 0.1 \). The narrowness of the depth uncertainty range can be readily seen on synthetic seismograms; for 3 km depth, the P-wave first motions at KON, MAL, and WIN are too large and most SH waves do not fit well. Our estimated depth uncertainty is comparable to the results of the surface-wave and short-period P-wave studies mentioned above, indicating agreement among these methods.

July 4, 1966

This earthquake occurred slightly to the east of Sao Miguel within a small depression at the eastern end of the Azores spreading center (Figure 3.8). Using P-wave first motions, Banghar and Sykes [1969] found a mechanism involving normal faulting with a large strike-slip component of motion (112/70/220). McKenzie [1972] restudied this earthquake with P-wave first motions and found a similar solution (103/70/230). Grimison and Chen [1985] used both P and SH waves to infer a pure normal-faulting mechanism (126/56/270) and suggested that oblique spreading is occurring at the Azores spreading center. They estimated the centroid depth to be about 12 \( \pm 5 \) km and the seismic moment to be \( 4.0 \times 10^{24} \) dyn-cm.

Because the depth for this and another nearby earthquake (April 20, 1968) obtained by Grimison and Chen [1985] are significantly deeper than the maximum centroid depth of about 3 km that we have obtained for Mid-Atlantic Ridge events, we decided to examine these two earthquakes in detail as part of
our study.

Our inversion solution (135/49/274) shown in Figure 3.9 is similar to that of Grimison and Chen [1985]. The strike of our solution is very close to the overall trend of the Azores spreading center. The centroid depth is 5.4 km, and the seismic moment is $4.0 \times 10^{24}$ dyn·cm. The difference in centroid depth between our study and that of Grimison and Chen [1985] is significant since both studies used the inversion routine of Nabelek [1984]. Several factors may have contributed to this discrepancy. First, we used many more stations (25) and have much better azimuthal coverage for both P and SH waves. Grimison and Chen [1985] only used 3 SH waveforms which gave almost no constraints on source mechanism. We used 10 SH waves and have very good constraints on the nodal planes. Second, our source-time function is 5 sec in duration, while theirs was 2 sec. Grimison and Chen [1985] argued that there is little trade-off between the focal depth and a source time function of such short duration. Third, different velocity structures were used in the inversion. We used a generalized oceanic crust-mantle structure while Grimison and Chen [1985] used a mantle halfspace.

To estimate the depth uncertainty, we followed the procedures of Chapter 2. Since the azimuthal coverage for this event is quite good, the source mechanism is stable for a centroid depth fixed in the range 1 to 20 km. The range of acceptable centroid depths is 4-5 km. A secondary local minimum in residual-versus-depth occurs at 10.8 km but may be rejected.
at $\alpha = 1$. For the 10.8-km deep solution, the P waveforms fit well, but the SH waves at NOR, SJG, BEC, and SCP are not well fit. This event occurred in an area that may be anomalous relative to conventional oceanic ridges, both because of the extremely low rate of extension and the proximity to the Azores hot spot. We therefore repeated our inversion with alternative source velocity structures. First, we used a mantle halfspace ($V_p = 8.0$ km/s, $V_s = 4.6$ km/s) identical to that used by Grimison and Chen [1985] and obtained a centroid depth of 6.8 km. A secondary local minimum in residual-versus-centroid depth occurred at 11.2 km, very close to the centroid depth inferred by Grimison and Chen [1985]. For a source structure approximated as a crustal halfspace ($V_p = 6.4$ km/s, $V_s = 3.7$ km/s) we obtained a centroid depth of 3.7 km. Finally, we used a model with a two-layer crust over a mantle halfspace ($V_p = 4.5, 6.9, 7.6$ km/s; $V_s = 2.6, 4.0, 4.4$ km/s; layer thicknesses = 3.8 and 4.2 km, respectively), following Searle [1976], and obtained a centroid depth of 6.3 km. This solution gives the smallest overall residual among the best-fitting solutions for all of the source structures we have examined.

A comparison of inversion results using different source structures is given in Table 4.3. Secondary local minima in the residual-versus-depth curves are also given in this table. As we can see, the best-fitting double-couple orientation is quite stable with respect to changes in the assumed source structure. However, the centroid depth varies from 3.7 to 6.8 km depending on the source structure used. We consider the 3 km difference
to be small in view of the shallowness of this earthquake and the very different source structures used in the tests. The length of the source time function ranges from 4 to 7.5 sec, slightly larger than for Mid-Atlantic Ridge earthquakes. All of the secondary local minima are deeper and have much shorter source time functions.

From this analysis, we believe that this event is deeper than typical Mid-Atlantic Ridge earthquakes and has a longer duration of rupture. Because the amplitudes of the water reverberations are much smaller than the P wave amplitudes, we believe that earthquake rupture was confined within the crust and did not extend to the seafloor. Comparisons of observed and synthetic seismograms for inversion results with different source structures, as well as the secondary solutions, are given in Appendix C.

April 20, 1968

The April 20, 1968 earthquake occurred about 50 km southeast of Terceira (Figure 3.8). Grimison and Chen [1985] studied this event using both P and SH waves and found a normal faulting mechanism (302/54/270). They estimated the depth to be 15 ± 5 km and the seismic moment to be 4.0 x 10^{24} dyn-cm. Our inversion solution (328/34/293), shown in Figure 3.10, has a slightly greater strike-slip component than that of Grimison and Chen [1985]. The centroid depth is 4.6 km, and the seismic moment is 2.0 x 10^{24} dyn-cm. Both the focal mechanism and centroid depth are very similar to those of the June 4, 1966 event. Because this earthquake is very small, the station coverage is not as good as that for the earlier earthquake. The
range in centroid depth estimated using the t-test is 3-16 km. A secondary solution occurs at 9 km, with a source-time function duration of 1 sec. The 9-km deep solution fits the waveforms quite well except for the P wave at TRN. The small size of this earthquake is probably the limiting size for which we can expect reliable results from long-period body waveform inversion. At the level $\alpha = 0.15$, the range in acceptable centroid depths is between 3 and 6 km. We think this earthquake is similar to the larger July 4, 1966, Azores event; the centroid depth lies between 3 and 6 km.

**September 20, 1969**

The September 20, 1969 earthquake occurred on the Reykjanes Ridge (Figure 3.11), about 100 km south of a transition in the character both of ridge axis seismicity and axial valley morphology [Francis, 1973; Vogt and Johnson, 1975]. To the north of about 59°N the ridge is anomalously shallow, lacks a prominent median valley, and is nearly aseismic, presumably a result of the influence of the Iceland hot spot; to the south of 59°N the seismicity and median valley morphology are more typical of other sections of the Mid-Atlantic Ridge. Using P-wave first-motion data, Solomon and Julian [1974] obtained a normal-faulting mechanism with nonorthogonal nodal planes. They showed that if the projection of first-motion data onto the focal sphere were corrected for propagation through a laterally heterogeneous model of the velocity structure beneath the ridge axis, a focal mechanism with orthogonal nodal planes could be found (9/41/263). Hart [1978] modeled the P waveforms from this earthquake as a forward problem. He argued that with a shallow
focal depth (2.1 ± 0.2 km) the waveforms, including those with apparently anomalous first motions, can be matched with a double-couple source (9/35/260). He obtained a moment of 2 x 10^{25} \text{dyn-cm}.

The observed P and SH waves show good signal-to-noise ratios and are well-distributed, so the waveform inversion problem is unusually well-constrained (Figure 3.12). The E-W component at NAT and both horizontal components at NIL were found to have reversed polarity and were corrected prior to inversion. The inversion solution (23/42/261) differs slightly in strike from that reported by Hart [1978], but the P waveforms alone allow only loose constraints to be placed on the fault strike; the P waveforms in Figure 11 are exceptionally well fit by the synthetic seismograms. We obtained a moment of 1.5 x 10^{25} \text{dyn-cm} and a centroid depth of 1.5 km, both slightly smaller than the values reported by Hart [1978]. The indicated water depth is 2.2 km, consistent with an epicenter in the central median valley of this shallow ridge system (Figure 3.11). Because of good azimuthal coverage, the source-mechanism-versus-depth trade-off is small. The 90% confidence interval for centroid depth as given by the t-test is 1-3 km. The behavior of a mantle solution has been extensively studied in Chapter 2, where we suggested that the mantle solution is an artifact of the sharp Moho assumed in the source structure used in the inversion.

May 31, 1971

The May 31, 1971 earthquake occurred north of Iceland on
the Mohns Ridge (Figure 3.13). Conant [1972] obtained a normal-faulting mechanism with nonorthogonal nodal planes from P-wave first motions. Savostin and Karasik [1981] used a larger data set to constrain the focal mechanism to a combination of normal and strike-slip faulting (141/59/220). Such a mechanism, however, does not fit the observed P and SH waveforms. From waveform inversion we obtained a normal-faulting mechanism (Figure 3.14) with only a small strike-slip component (51/51/284). We have excellent azimuthal coverage for both P and SH waves, and most observed waveforms are well fit. The seismic moment is $6.8 \times 10^{24}$ dyn-cm, and the focal depth is 2.3 km and well-constrained (Table 3.2).

Two slightly different water depths were indicated by the water-column reverberations in the P wavetrains. Stations to the east (SHL, KBL, MSH, IST, AOU and TOL) are consistent with a water depth of 3.1 km, while stations to the west (TRN, WES, AAM, DUG, and COL) suggest a somewhat shallower water depth of 2.8 km. We interpret this pattern as indicating an epicenter near the northwestern wall of the median valley inner floor, a location about 15 km north or northwest of the ISC epicenter (Figure 3.13). Due to good azimuthal coverage, the source mechanism is very stable. The range in acceptable centroid depths is 2-3 and 6-9 km. A secondary solution at 8.2 km depth has a focal mechanism (53/50/290) similar to that of the best-fitting solution at 2.3 km. The seismic moment for the mantle solution is $6.8 \times 10^{24}$ dyn-cm, and the source time function is 3 sec in duration. When we use a crustal structure
with a more gradual Moho transition the mantle solution deteriorates and can be rejected for $a = 0.1$.

April 3, 1972

The April 3, 1972 earthquake occurred on the southern Reykjanes Ridge, about 150 km north of its intersection with the Gibbs Fracture Zone (Figure 3.11). Einarsson [1979] obtained a normal-faulting mechanism with nonorthogonal nodal planes from P-wave first motions. From a moment-tensor inversion of the Rayleigh-wave radiation pattern and forward modeling of long-period P waveforms, Trehu et al. [1981] found a pure normal-faulting mechanism (4/46/270) and suggested that the apparent nonorthogonality is due to the shallowness of the source. They matched the P waveforms with a finite source of length 13 km, width 3 km, and moment $7.5 \times 10^{24}$ dyn-cm, with rupture initiating near the center of the fault and extending to the seafloor. The focal mechanism of Trehu et al. [1981] provides a good match to the P waveforms, but there are significant discrepancies with the SH waveforms at AAE, JER, and TUC. The inversion solution (8/48/254) has slightly more strike-slip motion (Figure 3.15) and matches all P and SH waveforms. The centroid depth of 1.9 km obtained from the inversion is somewhat deeper than the 1.1 km indicated by the finite fault model of Trehu et al. [1981]. The seismic moment of $4.4 \times 10^{24}$ dyn-cm given by the body-waveform inversion is smaller than that obtained by Trehu et al. [1981] from the forward modeling of P waveforms and from the moment-tensor inversion of Rayleigh waves. The inferred water depth of 2.8 km
is consistent with an epicenter in a locally deep portion of the median valley inner floor (Figure 3.11).

The depth uncertainty for this earthquake is small, with the range in acceptable values between 1.5 and 3 km, because the focal mechanism is very well constrained by SH waves. A secondary minimum in residual-versus-depth occurs just below the Moho at 6 km beneath the sea floor. Despite the fact that the mantle solution is very similar to the best solution, it may be rejected at the $\alpha = 0.1$ level. This is because the best-fitting solution is slightly better than the mantle solution at every station, thus making $\sigma_d$ small and the statistic $t$ large.

June 6, 1972

The June 6, 1972 earthquake occurred in the Mid-Atlantic Ridge median valley about 100 km south of the Hayes Fracture Zone (Figure 3.6). Station coverage for both P and SH waves is mostly confined to the northern hemisphere. The inversion solution (20/51/253) represents predominantly normal faulting (Figure 3.16) with a small strike-slip component. The strike of the fault planes is constrained by the SH waves at KTG, KEV, and CAR. The seismic moment is $4.1 \times 10^{24}$ dyn-cm, the centroid depth is 1.8 km, and the indicated water depth is 3.2 km.

The range in centroid depth at $\alpha = 0.1$ for this earthquake is 1 to 7 km. A secondary solution occurs just below the Moho at 6 km below the seafloor. The focal mechanism (25/50/260) and seismic moment ($3.5 \times 10^{24}$ dyn-cm) are very similar to those of the best-fitting solution at 1.8 km depth. Upon close examination, we noticed that the first half cycles of most
synthetic SH waves for the mantle solution are too wide when compared with the observed seismograms. At 6 km depth, the probability value \( P \) (see Appendix A) is 0.15; that is, we should make a type I error only 3 out of 20 times. Also, the mantle solution can be rejected at \( \alpha = 0.1 \) when a gradual Moho transition is used in the inversion. We therefore believe the mantle solution is not correct, and a more reasonable statement of the range in centroid depth is 1 to 6 km.

**June 28, 1977**

On June 28, 1977 an earthquake swarm occurred near the epicenter of the June 3, 1962 earthquake discussed above (Figure 3.2). The ISC lists 5 events over a 4-hour period. The first in the series (1538:37.8) had \( m_b = 5.3, M_S = 5.6 \); the fourth and largest in the series (1918:36) had \( m_b = 5.9, M_S = 6.0 \).

The inversion solution for the largest earthquake of the swarm (Figure 3.17) represents nearly pure normal faulting (1/44/255), a solution very similar to that of the June 3, 1962 earthquake. All of the observed waveforms are well-fit by the solution. The seismic moment is \( 1.1 \times 10^{25} \) dyn-cm, and the centroid depth is 1.6 km. P-wave coverage spans about 3 quadrants, and SH-wave coverage is well constrained about the nodal planes, so resolution of the focal mechanism is very good. A secondary solution occurs just beneath the Moho, but this solution can be rejected at \( \alpha = 0.1 \). Using the t-test we estimate that the range in centroid depth is 1 to 2 km at the 0.1 significance level. This estimate is probably accurate, because
the probability of failing to detect the difference is only 0.1 (i.e., $\beta = 0.1$) for depths outside this range. The inferred water depth is 3.5 km, indicating that the ISC epicenter is mislocated 6-7 km to the west (Figure 3.2).

The best overall solution for the first swarm event (Figure 3.18) is characterized by normal faulting with a large strike-slip component (355/50/241) and a centroid depth of 5.5 km. A secondary solution occurs at 2.5 km depth and has a focal mechanism (1/46/254) with a smaller strike-slip component that is nearly identical to that of the large event. Because all of the stations for this event are located in the northern hemisphere, the inversion solution is not well-constrained. If we constrained the slip angle to be larger than 250°, then the shallower solution has the smallest overall residual. We prefer the shallow solution because its mechanism is similar to those of other large earthquakes (with better station coverage) and microearthquakes in this area. Also, a preliminary analysis of short-period body-waveforms for this event indicates that the shallow solution is preferable [Bergman and Solomon, 1985b].

The shallower solution (1/46/254) is shown in Figure 2-17. The seismic moment is $3.0 \times 10^{24}$ dyn-cm, and the inferred water depth is 4.0 km, greater than that of the larger event but also consistent with an epicenter in the central median valley about 10 km east of the ISC location (Figure 3.2).

The P waveforms of another event in this swarm (1618:16, $m_b = 5.5$, $M_s = 5.7$) are obscured by the surface waves from the earthquake 40 minutes before. A normal faulting mechanism
(0/45/255), with a seismic moment of $5 \times 10^{24}$ dyn-cm and a centroid depth of 1.5 km, fits the observed SH waveforms for this earthquake quite well.

**January 28, 1979**

The January 28, 1979 earthquake occurred in the median valley near its intersection with a small fracture zone about 100 km north of the Vema Fracture Zone (Figure 3.4). The focal mechanism of this earthquake, as determined in the inversion, is pure normal faulting (20/46/270), with a seismic moment of $6.3 \times 10^{24}$ dyn-cm and a centroid depth of 2.2 km (Figure 3.19). The strike of the nodal planes differs by about 10° from the strike of the ridge axis. The SH waves are reasonably well-fit, but the station coverage for SH waves is inadequate to rule out a mechanism striking N10°E rather than N20°E. The inferred water depth of 3.7 km is consistent with the epicenter shown in Figure 3.4.

Because the fault planes are not well-constrained, there is some trade-off between the centroid depth and focal mechanism. Using the t-test, we estimate the range of acceptable depths to be from 1.5 to 9 km at a 0.1 significance level. However, we think the uncertainty in centroid depth is smaller than this range indicates, for several reasons. The local minima in the residual-versus-depth curve are sharp between 2 and 3 km and between 6 and 7 km. At 3 km centroid depth, the P value at which the solution differs significantly from the best-fitting solution is about 0.15; that is, we would make a type I error only 3 out of 20 times. At an α level of 0.1, β = 0.6 for the 3-km-deep solution; that is, we have about a 3-in-5 chance of
being unable to detect the difference between the best and the 3-km-deep solutions. The use of $\alpha = 0.15$ slightly improves our chance of detecting a difference ($\beta = 0.5$). Also, when we use a more gradual Moho in the inversion, the mantle solution deteriorates and lies outside the uncertainty range at $\alpha = 0.15$. From the above statistical considerations, our best estimate of the range in centroid depth is 2 to 3 km.

April 22, 1979

The earthquake on April 22, 1979 occurred very near the epicenter of the June 6, 1972 event (Figure 3.6). Station coverage for both P and SH waves is excellent, and the observed waveforms are very well fit by a normal-faulting mechanism (17/52/262), as shown in Figure 3.20. The centroid depth is 1.8 km and is very well-constrained. The indicated water depth is 3.3 km, and the seismic moment is $9.9 \times 10^{24}$ dyn-cm. The double-couple orientation and both the centroid and water depths are very similar to those of the June 6, 1972 earthquake.

The station coverage for this event is fairly good, particularly for the SH waves which gives very good constraints for the fault planes. There is little trade-off between centroid depth and focal mechanism for the depth range 1 to 10 km. Using the t-test, we estimate that the range in acceptable centroid depth is between 1.5-2.5 km and 6-7 km at a significance level of 0.1. The focal mechanism (13/54/255) and seismic moment ($8.3 \times 10^{24}$ dyn-cm) for the mantle solution are very similar to those of the best solution. This mantle solution deteriorates, however, and lies outside the uncertainty
range at $\alpha = 0.15$ when a gradual Moho structure is used in the inversion. Based on the above analysis, we believe that the correct centroid depth lies between 1.5 and 2.5 km.

June 28, 1979

The June 28, 1979 earthquake occurred in the median valley about 30 km south of its eastern intersection with the St. Paul's Fracture Zone (Figure 3.21). Station coverage for both P and SH waves is concentrated in the northern hemisphere. The focal mechanism of this event (Figure 3.22) is characterized by nearly pure normal faulting (346/40/280). The observed P and SH waves are well fit by this mechanism, with a centroid depth of 2.5 km and a seismic moment of $3.6 \times 10^{24}$ dyn-cm. The inferred water depth is 3.5 km, consistent with the epicentral position and bathymetry shown in Figure 3.21. A secondary minimum in residual-versus-depth occurs at 6 km, just beneath the Moho. The focal mechanism (340/44/263) and seismic moment ($3.5 \times 10^{24}$ dyn-cm) for this solution are similar to those of the best-fitting solution. The fits of all waveforms with the mantle solution are quite good. However, at a significance level of 0.1 the mantle solution can be rejected. The resolution of the centroid depth therefore lies in the range 1.5-4.5 km.

January 29, 1982

The January 29, 1982 earthquake occurred in the median valley about 200 km north of the Kane Fracture Zone (Figure 3.23). Using a semi-automated centroid moment tensor inversion and data from the GDSN, Dziewonski et al. [1983a] obtained a normal-faulting mechanism with a significant strike-slip
component (353/61/246). Station coverage, particularly for P waves, is rather poor for this event because many records have not yet been distributed. From the body-waveform inversion, we obtained a nearly pure normal-faulting mechanism (9/44/264), with fault planes striking subparallel to the local trend of the ridge axis (Figure 3.24). Most of the waveforms are well-fit by the synthetic seismograms, especially the nodal character of the SH wave at KBS. The seismic moment of $5.4 \times 10^{24}$ dyn-cm is somewhat smaller than the $8.1 \times 10^{24}$ dyn-cm obtained by Dziewonski et al. [1983a]. The centroid depth is 2.4 km, and the water depth 3.8 km.

The range in centroid depth for this earthquake is 1-4 km and 6-7 km at $\alpha = 0.1$. A secondary minimum in residual-versus-depth occurs just beneath the Moho at 6 km depth. The focal mechanism (13/43/271) and seismic moment ($6.3 \times 10^{24}$ dyn-cm) of the mantle solution are very similar to those of the best solution. However, the fits of most waveforms are slightly worse than those of the best-fitting solution. Using a gradual Moho, the mantle solution deteriorates and lies outside the uncertainty range. We therefore believe that the centroid depth is between 1 and 4 km.

May 12, 1983

The earthquake of May 12, 1983 occurred in the median valley about 250 km north of the $15^\circ 20'$ Fracture Zone (Figure 3.24). A centroid moment-tensor solution (354/50/259) by Dziewonski et al. [1983b] shows predominantly normal-faulting motion. Station coverage for this event is good, and the focal
mechanism is well-resolved. Our inversion solution (341/48/251) shown in Figure 3.25 is very similar to that of Dziewonski et al. One of the nodal planes is subparallel to the local trend of the ridge axis. The seismic moment of $7.1 \times 10^{24}$ dyn-cm is similar to the $8.1 \times 10^{24}$ dyn-cm obtained by Dziewonski et al. [1983b]. The centroid depth is 3.1 km, and the inferred water depth is 4.1 km.

There is some trade-off between the centroid depth and the source mechanism for fixed depths in the range 1-10 km. A secondary minimum in residual-versus-depth occurs at 6 km, immediately below the Moho, but this solution can be rejected at $\alpha = 0.1$. We estimate that the centroid depth of this earthquake lies in the range 2-6 km at a significance level of 0.1.

DISCUSSION

The 16 earthquakes under consideration in this chapter occurred along four different plate boundaries. Except for the two Azores region earthquakes, all 14 earthquakes occurred on spreading segments of the Mid-Atlantic Ridge. Since the Azores earthquakes occurred beneath basins probably formed by the rifting of oceanic lithosphere rather than in a mid-ocean ridge proper, we will discuss them separately.

The 14 Mid-Atlantic Ridge ridge-axis earthquakes of this study are remarkably similar in their source characteristics. All show normal-faulting mechanisms with at least one nodal plane dipping at about 45° and striking approximately parallel to the local trend of the ridge axis. The inferred T axes are all nearly horizontal and are approximately aligned with the
spreading direction. The seismic moments span a range of only a factor of 5, from \(3 \text{ to } 15 \times 10^{24} \text{ dyn-cm}\), and the source time functions are generally of similar duration. The centroid depths are all very shallow, between 1.2 and 3.1 km beneath the seafloor.

The water depth at the epicenter of each of these earthquakes is well-constrained by the predominant period of the large water-column reverberations in the P wavetrains. The water depth estimated from the P waveform is compared in Table 3.2 with the maximum local depth of the median valley as indicated on available bathymetric maps. Given the uncertainties in the epicentral coordinates and in the depths obtained from contour maps constructed from ship-track data of variable sampling density and accuracy, the agreement between the two depth estimates is surprisingly good, to within a few hundred meters for all earthquakes in Table 3.2. We conclude from this comparison that all of the earthquakes in this study occurred beneath the inner floor of the median valley. We cannot exclude the possibility that some of these events occurred near the inner walls of the rift valley (e.g., the May 31, 1971 earthquake) or that they represent slip on major inward-dipping faults that contribute to the relief of the inner valley walls [Macdonald and Luyendyk, 1977]. It may be inferred with reasonable certainty, however, that none of the Mid-Atlantic Ridge earthquakes considered in this chapter occurred beneath the rift mountains.

From the centroid depth, the source-time function, and the
seismic moment, we can obtain rough estimates of fault dimensions, average slip, and average stress drop for these earthquakes. For simplicity, we assume that the zone of slip is rectangular, with horizontal length $L$ and down-dip width $w$. We estimate $L$ from the length $t_s$ of the source time function. On the basis of the large water-column reverberations we assume that rupture extended to the surface, and we estimate $w$ from the centroid depth. For unilateral rupture, $L = V_r t_s$, where the rupture velocity $V_r$ is taken to be equal to 0.8 times the shear velocity, or 3.0 km/s. We estimate $w$ from $w = 2.8 h$, where $h$ is the centroid depth; this relation follows from a fault dip of about 45° and a centroid depth that marks the average depth of fault slip. If the fault width $w$ given by this relation is larger than the fault length, we set $w = L$ so that the vertical extent of faulting is no greater than the horizontal extent.

The average slip $D$ can be estimated from the relation $M_o = \mu L w D$ [Aki, 1966], where the rigidity $\mu$ of the crust is taken to be $4.0 \times 10^{11}$ dyn/cm$^2$ from our assumed crustal structure. The stress drop $\Delta \sigma$ can be estimated from the relation $\Delta \sigma = c M_o/(L w^2)$, where $c$ is a geometrical constant near unity [Knopoff, 1958; Aki, 1966].

For the Mid-Atlantic Ridge earthquakes, the estimated fault area $L w$ ranges from 30 to 100 km$^2$ (Table 3.5) and shows no resolvable correlation with moment for the limited set of events studied here. The average fault slip varies from 10 to 80 cm. The estimated stress drops lie between 7 and 70 bars, within the range shown by other tectonic earthquakes [Kanamori and Anderson, 1975].
The two earthquakes in the Azores region have slightly larger centroid depths than the Mid-Atlantic Ridge earthquakes. Using different source structures in our inversion, the best-fitting centroid depth varies from 3.7 km to 6.8 km for the July 4, 1966, earthquake. We believe that these two earthquakes occurred in a different environment from those along the Mid-Atlantic Ridge. It is possible that fault rupture within these two earthquakes did not extend to the seafloor. The estimates of fault width and fault area for these two earthquakes in the Azores region (Table 3.5) are probably too large, and the average slip and stress drop are probably too small.
Table 3.1 Epicentral Data for North Atlantic Ridge-Axis Earthquakes

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Epicentral and magnitude data are taken from the International Seismological Summary for 1962 and the ISC for 1964-83.

a Half spreading rate calculated from model RM2 of Minster and Jordan [1978]

b Magnitude M from Rothe [1969]
<table>
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<tr>
<th>Date</th>
<th>$M_o$</th>
<th>$t_S$</th>
<th>Strike</th>
<th>Dip</th>
<th>Slip</th>
<th>Centroid Depth</th>
<th>Water Depth</th>
<th>Depth of Median Valley</th>
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<td>June 3, 1962</td>
<td>6.1±0.5</td>
<td>3.0±1.0</td>
<td>4.4±0.5</td>
<td>42.8±1.7</td>
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<td>1.2±0.11</td>
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<td>3.4-4.2</td>
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<td>June 2, 1965</td>
<td>8.3±0.7</td>
<td>3.0±1.0</td>
<td>335.0±1.9</td>
<td>37.5±0.8</td>
<td>243.5±2.1</td>
<td>3.0±0.06</td>
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<td>3.5-4.0</td>
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<td>Nov. 16, 1965</td>
<td>8.1±0.8</td>
<td>3.0±1.0</td>
<td>24.6±2.5</td>
<td>45.9±1.0</td>
<td>265.3±2.8</td>
<td>2.1±0.14</td>
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<td>3.0-3.5</td>
</tr>
<tr>
<td>July 4, 1966</td>
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<td>5.0±1.0</td>
<td>135.4±1.0</td>
<td>49.0±0.4</td>
<td>274.3±1.1</td>
<td>5.4±0.10</td>
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<td>4.2±1.4</td>
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<td>50.9±1.2</td>
<td>52.1±0.5</td>
<td>283.8±1.2</td>
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<td>3.2±0.8</td>
<td>8.2±3.0</td>
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<td>2.0-2.5</td>
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<td>3.0-3.5</td>
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<td>5.0±1.0</td>
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<td>3.4-4.2</td>
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<td>2.0±1.0</td>
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<td>16.6±1.3</td>
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<td>261.7±1.5</td>
<td>1.8±0.06</td>
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<td>3.0-3.5</td>
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<td>3.6±0.9</td>
<td>345.5±1.5</td>
<td>40.0±0.9</td>
<td>279.7±2.1</td>
<td>2.5±0.09</td>
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<td>3.5-4.0</td>
</tr>
<tr>
<td>Jan. 29, 1982</td>
<td>5.4±0.8</td>
<td>3.0±1.0</td>
<td>9.0±3.3</td>
<td>43.8±0.6</td>
<td>264.1±3.3</td>
<td>2.4±0.08</td>
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<td>3.5-4.0</td>
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<td>341.0±2.0</td>
<td>48.2±0.7</td>
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</tbody>
</table>
Notes to Table 3.2

a) The range indicated for source parameters ($M_0$, strike, dip, slip, and centroid depth) is one standard deviation (formal error).

b) $x 10^{24}$ dyne-cm ($10^{17}$ N-m)

c) Total duration of the source time function, sec

d) Relative to the seafloor, km

e) Inferred from modeling the water-column reverberations in the later portions of the P wavetrains.

f) Local depth of the median valley inner floor indicated by bathymetric maps [Laughton and Monahan, 1978; Heezen and Tharp, 1978; Johnson et al., 1979; Searle et al., 1982; Toomey et al., 1985].
### Table 3.3 Inversion results for the July 4, 1966 Azores spreading center earthquake

<table>
<thead>
<tr>
<th>Source structure</th>
<th>Seismic moment, (10^{24}) dyn-cm</th>
<th>Double couple (strike/dip/slip)</th>
<th>Centroid depth, km</th>
<th>Duration of source time function, sec</th>
<th>RMS residual, mm</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>A. Overall best solution</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1. Crustal halfspace</td>
<td>4.0</td>
<td>131/49/270</td>
<td>3.7</td>
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<td>1.90</td>
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<tr>
<td>2. Uniform crust over mantle halfspace</td>
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<td>135/49/274</td>
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<td>5</td>
<td>1.89</td>
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<tr>
<td>3. Two-layer crust over mantle halfspace</td>
<td>4.7</td>
<td>133/50/271</td>
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<td>6</td>
<td>1.80</td>
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<td>4. Mantle halfspace</td>
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<td>130/50/267</td>
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<td><strong>B. Secondary local minimum</strong></td>
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<td>1. Crustal halfspace</td>
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<td>4. Mantle halfspace</td>
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<td>146/47/285</td>
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Table 3.4a  Summary of Source Parameter Uncertainty Ranges for Atlantic Ocean Ridge-Axis Earthquakes

<table>
<thead>
<tr>
<th>Date</th>
<th>Range in centroid depth, in km, given by t-test, $\alpha = 0.1$</th>
<th>Secondary solution with sharp Moho</th>
<th>Secondary solution with gradual Moho</th>
<th>Adopted uncertainty ranges</th>
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<td></td>
<td>Range in centroid depth, in km, given by t-test, $\alpha = 0.1$</td>
<td>Depth, km within uncertainty range for $\alpha = 0.1$</td>
<td>Depth, km within uncertainty range for $\alpha = 0.1$</td>
<td>Depth, km $\alpha$</td>
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<td>1-2 and 6-7</td>
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<td>1-2</td>
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<td>1-5 and 6-8</td>
<td>6</td>
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<td>1-5</td>
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<tr>
<td>Nov. 16, 1965</td>
<td>1-3</td>
<td>6</td>
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<td>1-3</td>
</tr>
<tr>
<td>July 4, 1966</td>
<td>5-6</td>
<td>10.8</td>
<td>no</td>
<td>5-6</td>
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<td>1-3 and 6-7</td>
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Table 3.4b: Summary of Source Parameter Uncertainty Ranges for Atlantic Ocean Ridge-Axis Earthquakes

<table>
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<th>Dip, deg</th>
<th>Slip, deg</th>
<th>Seismic Moment, $10^{24}$ dyn-cm</th>
<th>Source time function duration, sec</th>
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<td>33-46</td>
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<td>45-46</td>
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<td>285-296</td>
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<tr>
<td>Date</td>
<td>Fault length,\textsuperscript{a} km</td>
<td>Fault width,\textsuperscript{b} km</td>
<td>Fault area, km\textsuperscript{2}</td>
<td>Average slip, cm</td>
<td>Stress drop, bars</td>
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<td>June 28, 1977</td>
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<td>70</td>
<td>40</td>
<td>40</td>
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<tr>
<td>Jan. 28, 1979</td>
<td>6</td>
<td>6</td>
<td>40</td>
<td>40</td>
<td>30</td>
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<tr>
<td>April 22, 1979</td>
<td>12</td>
<td>5</td>
<td>60</td>
<td>40</td>
<td>30</td>
</tr>
<tr>
<td>June 28, 1979</td>
<td>11</td>
<td>7</td>
<td>70</td>
<td>10</td>
<td>7</td>
</tr>
<tr>
<td>Jan. 29, 1982</td>
<td>9</td>
<td>7</td>
<td>60</td>
<td>20</td>
<td>10</td>
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<tr>
<td>May 12, 1983</td>
<td>7</td>
<td>7</td>
<td>50</td>
<td>40</td>
<td>20</td>
</tr>
</tbody>
</table>

\textsuperscript{a} Estimated from duration of source time function. Rupture velocity is assumed to be 0.8 times shear wave velocity.

\textsuperscript{b} Calculated from the assumptions: \( w < \ell \) and \( w < 2.8h \).
FIGURE CAPTIONS

Figure 3.1. Location of Mid-Atlantic Ridge and Azores region earthquakes studied in this chapter, along with the overall distribution of seismicity [Tarr, 1974]. Filled circles show earthquakes with \( m_b > 4.5 \) during 1963-1972; open circles denote events in this century with \( M_s > 8 \).

Figure 3.2. Detailed bathymetry of the Mid-Atlantic Ridge median valley near the epicenters of the earthquakes of June 3, 1962, and June 28, 1977. Bathymetric contours, in m, are from Toomey et al. [1985]. The location of this region relative to neighboring elements of the Mid-Atlantic Ridge system is shown in Figure 3.23. Fault plane solutions are equal area projections of the lower focal hemisphere; compressional quadrants are shaded. Mechanisms labeled A and B for February 1982 are composite fault plane solutions determined by Toomey et al. [1985] for two clusters of microearthquakes recorded by a network of ocean-bottom seismometers and ocean-bottom hydrophones located at the positions shown by numbered circles and triangles, respectively.

Figure 3.3. Comparison of observed (solid line) and synthetic (dashed line) long-period P and SH waves for the June 3, 1962, earthquake, with the focal mechanism solution obtained from the inversion plotted on the lower focal hemisphere (equal-area projection). All amplitudes are normalized to an instrument magnification of 3000; the amplitude scales correspond to the waveforms that would be
observed on an original seismogram from such an instrument. The two vertical lines delimit the portion of each time series, digitized at 0.5 s intervals, used in the inversion. Symbols for both types of waves are open circle, dilatation; solid circle, compression; cross, emergent arrival. For SH waves, compression corresponds to positive motion as defined by Aki and Richards [1980]. Some first motions are shown for stations not used in the waveform inversion. The source time function obtained from the inversion is also shown.

Figure 3.4. Bathymetry and seismicity of the Mid-Atlantic Ridge near the epicenters of the earthquakes of June 2, 1965; January 28, 1979; and May 12, 1983 (Mercator projection). Bathymetric contours, in km, are from Searle et al. [1982]; regions shallower than 3 km are shaded. Earthquake epicenters are from the ISC for 1964-1979; larger symbols denote events with $m_b > 5.4$. Fault plane solutions for events of this study are equal area projections of the lower focal hemisphere; compressional quadrants are shaded.

Figure 3.5. Comparison of the observed and synthetic P and SH waves for the June 2, 1965, event. See Figure 3.3 for further explanation.

Figure 3.6. Bathymetry and seismicity of the Mid-Atlantic Ridge near the epicenters of the earthquakes of November 16, 1965; June 6, 1972; and April 22, 1979. Bathymetry is from Searle et al. [1982]; see Figure 3.4 for further explanation.
Figure 3.7. Comparison of the observed and synthetic P and SH waves for the November 16, 1965, event. See Figure 3.3 for further explanation.

Figure 3.8. Bathymetry and seismicity of the spreading centers near the Azores triple junction, including the epicenters and mechanisms of the earthquakes of July 4, 1966, and April 20, 1968. Bathymetric contours, in km, are from Searle et al. [1982]; regions shallower than 2 km are shaded. See Figure 3.4 for further explanation.

Figure 3.9. Comparison of the observed and synthetic P and SH waves for the July 4, 1966, event. See Figure 3.3 for further explanation.

Figure 3.10. Comparison of the observed and synthetic P and SH waves for the April 20, 1968, event. See Figure 3.3 for further explanation.

Figure 3.11. Bathymetry and seismicity of the Reykjanes and Mid-Atlantic Ridges near the epicenters of the earthquakes of September 20, 1969, and April 3, 1972. Bathymetric contours, in km, are from Laughton and Monahan [1978]; regions shallower than 2 km are shaded. See Figure 3.4 for further explanation.

Figure 3.12. Comparison of the observed and synthetic P and SH waves for the September 20, 1969, event. See Figure 3.3 for further explanation.

Figure 3.13. Bathymetry and seismicity of the Mohns Ridge near the epicenter of the earthquake of May 31, 1971 (polar stereographic projection). Bathymetric contours, in km,
are from Johnson et al. [1979]; regions shallower than 2 km are shaded. See Figure 3.4 for further explanation.

Figure 3.14. Comparison of the observed and synthetic P and SH waves for the May 31, 1971, event. See Figure 3.3 for further explanation.

Figure 3.15. Comparison of the observed and synthetic P and SH waves for the April 3, 1972, event. See Figure 3.3 for further explanation.

Figure 3.16. Comparison of the observed and synthetic P and SH waves for the June 6, 1972, event. See Figure 3.3 for further explanation.

Figure 3.17. Comparison of the observed and synthetic P and SH waves for the mb = 5.9 event of June 28, 1977. See Figure 3.3 for further explanation.

Figure 3.18. Comparison of the observed and synthetic P and SH waves for the mb = 5.3 event of June 28, 1977. This earthquake was the first in a series of 5 events within 4 hours, including the earthquake of Figure 3.17. See Figure 3.3 for further explanation.

Figure 3.19. Comparison of the observed and synthetic P and SH waves for the January 28, 1979, event. See Figure 3.3 for further explanation.

Figure 3.20. Comparison of the observed and synthetic P and SH waves for the April 22, 1979, event. See Figure 3.3 for further explanation.

Figure 3.21. Bathymetry and seismicity of the Mid-Atlantic Ridge near the epicenter of the earthquake of June 28, 1977.
1979. Bathymetric contours, in km, are from Heezen and Tharp [1978]; regions shallower than 3 km are shaded. See Figure 3.4 for further explanation.

Figure 3.22. Comparison of the observed and synthetic P and SH waves for the June 28, 1979, event. See Figure 3.3 for further explanation.

Figure 3.23. Bathymetry and seismicity of the Mid-Atlantic Ridge near the epicenter of the earthquake of January 29, 1982. Bathymetric contours, in km, are from Searle et al. [1982]; regions shallower than 3 km are shaded. The small box centered on the median valley near 23°N outlines the area shown in greater detail in Figure 3.2. See Figure 6 for further explanation.

Figure 3.24. Comparison of the observed and synthetic P and SH waves for the January 29, 1982, event. See Figure 3.3 for further explanation.

Figure 3.25. Comparison of the observed and synthetic P and SH waves for the May 12, 1983, event. See Figure 3.3 for further explanation.
Figure 3.1
Figure 3.5
Figure 3.6
Figure 3.7
Figure 3.8

APRIL 20, 1968

JULY 4, 1966

EAST AZORES F. Z.
Figure 3.10
Figure 3.11
Figure 3.12
Figure 3.15
Figure 3.16
Figure 3.17
Figure 3.18
Figure 3.20
Figure 3.21
Figure 3.24
CHAPTER 4. RIDGE-AXIS EARTHQUAKES IN THE
NORTHWEST AND CENTRAL INDIAN OCEAN

INTRODUCTION

In this chapter, we determine the focal depths and source mechanisms of 11 ridge-axis earthquakes in the Indian Ocean, 2 rift earthquakes in the northern Red Sea, and one near-ridge earthquake. Source mechanism studies of mid-ocean ridge earthquakes in the North Atlantic [Trehu et al., 1981; Chapter 3] and Arctic [Jemsek et al., 1984] Oceans have shown that ridge-axis earthquakes have several consistent features: they are extremely shallow, located within the median valley, and characterized by nearly pure normal faulting, with nodal planes dipping at about 45° and striking parallel to the overall trend of the ridge axis. However, both the Mid-Arctic and northern Mid-Atlantic Ridges have relatively simple tectonic histories. In contrast, while mid-ocean ridges in the Indian Ocean have generally comparable spreading rates (half-spreading rates between 6-22 mm/yr), they are the products of a more complex tectonic evolution. Thus, a synthesis of the source mechanisms of ridge events in the Indian Ocean will provide us a chance to compare the focal depths and source mechanisms of earthquakes over a range of both spreading rates and different tectonic settings. As in the previous chapter, we pay special attention to focal depth, as it can be an extremely important quantity for understanding the depth extent of brittle behavior beneath the ridge axis.
SPREADING CENTERS OF THE NORTHWEST INDIAN OCEAN REGION

The active ridge system in the northwestern Indian Ocean includes the E-W trending Sheba Ridge in the Gulf of Aden, the NW-SE-trending Carlsberg Ridge, and the N-S-trending Central Indian Ridge. To the west, the Sheba Ridge joins the southern end of the Red Sea and the East African Rift system. The account of regional tectonics presented below is drawn primarily from the syntheses of Schlich [1982] and Shazly [1982].

The Red Sea

The Red Sea is a NNW-SSE-trending depression separating the Arabian Peninsula and the African Shield. It extends some 1800 km from the tip of the Sinai Peninsula to the Straits of Bab El Mandab [Shazly, 1982]. The Red Sea consists of continental shelves, a main trough from 15°N to the Sinai Peninsula, and a deeper and narrower axial trough between 15° and 24°N [Cochran, 1983]. The axial trough is associated with large-amplitude magnetic anomalies [Drake and Girdler, 1964]. Seismic refraction data [Drake and Girdler, 1964] have shown that the basement velocities beneath the main trough are similar to those of the continental rocks. A later seismic refraction study [Tramontini and Davies, 1969] found higher velocity for the inner half of the main trough between 22° and 23°N, suggesting that the main trough is underlain by oceanic crust.

There are several schools of thought about the nature of the crust beneath the Red Sea. One hypothesis, based on seismic and geologic studies, is that the oceanic crust is limited to the axial trough and that most of the main trough is underlain
by continental crust [Huchinson and Engel, 1972; Lowell and Genik, 1972; Ross and Schlee, 1973; Owell et al., 1975]. A competing hypothesis, based on plate kinematic and magnetic anomaly studies, is that the Red Sea is almost entirely underlain by oceanic crust [McKenzie et al., 1970; Girdler and Styles, 1974, 1976; Roeser, 1975; Styles and Hall, 1980].

Cochran [1983] has made a convincing case for an alternative model in which (1) sea floor spreading is occurring in the southern Red Sea, (2) continental extension is ongoing but no mid-ocean ridge is present in the northern Red Sea, and (3) a transition area separates the two tectonic regimes.

The Sheba Ridge

The Sheba Ridge is a part of the Mid-Indian Ridge system. To the east, the ridge is offset by about 300 km from the Carlsberg Ridge by the NE-SW-trending Owen Fracture Zone. To the west, the ridge joins with the East African Rift to the south and the Red Sea axial spreading center to the north. The western portion of the Sheba Ridge consists of a single deep valley, while it breaks into a succession of parallel ridges to the east. The depth of the valleys increases progressively from west to east [Schlich, 1982]. The trend of the ridge axis varies from E-W at about 56°E to SE-NW at the Owen Fracture Zone. The Sheba Ridge is offset by several NE-SW-trending fracture zones. Earthquake epicenters clearly delineate the accretionary plate boundary between the African and Arabian [Sykes and Landisman, 1964; Sykes, 1970b] or Indo-Arabian [Wiens et al., 1985] plates. Seismic refraction data [Laughton and
Tramontini, 1969] and magnetic lineations [Girdler, 1963; Drake and Girdler, 1964; Laughton, 1966] have shown that the crust in the Gulf of Aden is oceanic. Laughton et al. [1970] calculated the half-spreading rate to be about 10 mm/yr, perhaps slightly greater to the east (11 mm/yr) than to the west (9 mm/yr).

The Carlsberg Ridge

The Carlsberg Ridge trends NW-SE between the equator and the Owen Fracture Zone and separates the Arabian Sea and the East Somali Basin. The ridge has been well-studied [Matthews et al., 1965; Cann and Vine, 1966; Le Pichon and Heirtzler, 1968; Fisher et al., 1968; McKenzie and Sclater, 1971]. The ridge axis is clearly defined by earthquake epicenters [Barazangi and Dorman, 1969] and axial magnetic anomalies. The present half-spreading rate, based on axial magnetic anomalies, is about 12 to 13 mm/yr [Le Pichon and Heirtzler, 1968; McKenzie and Sclater, 1971].

The Central Indian Ridge

The N-S-trending Central Indian Ridge extends from the equator at the southern end of the Carlsberg Ridge to the African-Indian-Antarctic triple junction near 25°S. It is bounded on the east by the Chagos-Laccadive Ridge and on the west by the Mascarene Plateau. Langseth and Taylor [1967] and Fisher et al. [1971] have shown that the Central Indian Ridge consists of a series of short ridge segments cut by many northeast-southwest-trending fracture zones. The major fracture zones are the Vema Trench, the Argo Fracture Zone, and the Marie Celeste Fracture Zone. Earthquake epicenters and axial magnetic
anomalies clearly delineate the plate boundary. The present half-spreading rate varies between 18 and 24 mm/yr from north to south [Fisher et al., 1971]; the spreading direction is about N40°E. Fisher et al. [1971] suggest that the bathymetric complexity was caused by a change in spreading direction from N-S in the late Cretaceous and early Tertiary to NW-SE in Miocene time. The Chagos Fracture Zone was then broken up into a series of ridge segments joined by transform faults.

EARTHQUAKE DATA SET

We first assembled a list of mid-ocean ridge earthquakes that occurred on the ridge systems of the Indian Ocean between latitudes 25°S and 30°N and over the interval 1964-1984. The search was based primarily on the Bulletins of the International Seismological Center (ISC) for the years 1964-1982, the monthly listings of the National Earthquake Information Service (NEIS) for the years 1982-1984, and published studies of fault plane solutions for the study area [Banghar and Sykes, 1969; McKenzie et al., 1969; Ben-Menahem and Aboodi, 1971; Dziewonski et al., 1983; Dziewonski and Woodhouse, 1983a; Dziewonski et al., 1984]. We chose only the earthquakes large enough to yield clear long-period P and S waves at teleseismic distances. We excluded known near-ridge intraplate earthquakes [Bergman and Solomon, 1984], strike-slip events on transform faults, and a complex normal-faulting earthquake (October 18, 1964) for which we were unable to obtain a satisfactory solution.

Epicentral data for the 14 earthquakes for which we have obtained reliable solutions are listed in Table 4.1; their
epicenters are shown in Figure 4.1. For all of the events in the table, we have used P and SH waves recorded on long-period instruments at stations of the WWSSN or GDSN at epicentral distances between 30° and 80°. All these events have epicenters within median valley segments on bathymetric maps of the GEBCO (General Bathymetric Chart of the Ocean) series [Laughton, 1975; Fisher et al., 1982]. The local half-spreading rates vary from 6 to 22 mm/yr, according to plate kinematic model RM2 of Minster and Jordan [1978]. The earthquakes have body wave magnitudes between 5.4 and 6.1, and all but the one near-ridge event have normal faulting mechanisms.

WAVEFORM INVERSION RESULTS

We present here the details of the source mechanisms obtained from the inversion of P and SH waveforms for the 11 Indian Ocean ridge-axis, 2 northern Red Sea, and 1 near-ridge earthquakes. The best-fitting double-couple orientations, seismic moments, and centroid depths obtained from the formal inversions are given in Table 4.2. The convention for describing a double-couple orientation is that of Aki and Richards [1980]. Centroid depths are relative to the seafloor. The average water depth above each hypocenter is inferred by matching the synthetic and observed seismograms in the ringing portion of the P wavetrain. A discussion of the uncertainty in centroid depth is given for each earthquake. The uncertainty ranges are calculated at a significance level of 0.1. In some cases, we also give P and B values (see Appendix A for a discussion of these quantities).
Except for two earthquakes in the northern Red Sea, all waveform inversions were performed with the same source structure: a water layer of variable thickness, a single crustal layer of thickness 6 km and seismic velocities $V_p = 6.4$ km/s and $V_s = 3.7$ km/s, and a mantle half-space with $V_p = 8.1$ km/s and $V_s = 4.6$ km/s. For the northern Red Sea earthquakes, which probably occurred in thinned continental crust [McKenzie et al., 1970; Cochran, 1983], we used a crustal half-space with $V_p = 6.0$ km/s and $V_s = 3.46$ km/s. For the March 31, 1969 earthquake, we have also examined the influence of structure models on inversion solutions. Details of the study are given later in the discussion of this earthquake.

March 19, 1964

This earthquake occurred on the East Sheba Ridge where the ridge axis changes direction (Figure 4.2). To the east, the trend of the ridge axis is SE-NW. To the west, the trend of the ridge axis is ESE-WNW and the ridge becomes shallower. Station coverage for this event is poor, particularly for the SH waves. The inversion solution for this event (Figure 4.3) is characterized by normal faulting with a small strike-slip component (282/38/241). The observed P and SH waveforms are well fit, and the dip of one nodal plane is constrained by the P wave at BUL which shows a clear upward first motion. The strike of one nodal plane agrees well with the trend of the ridge axis to the west; the strike of the second (317°) is parallel to the trend of the ridge axis to the east. The seismic moment of this earthquake is $5.0 \times 10^{24}$ dyn-cm, the
centroid depth is 1.8 km, and the average water depth at the
epicenter is about 3.3 km, which is consistent with the
bathymetry.

Because of the poor station coverage, there is a
significant trade-off between the centroid depth and the focal
mechanism. For a fixed centroid depth varying from 0 to 15 km,
the strike of the best-fitting mechanism varies from 280° to
340°, the dip varies from 25° to 41°, and the slip varies from
230° to 315°. Using the t-test we estimate the range of
acceptable depths to be from 0.5 to 13 km at an 0.1 significance
level. However, we think that the depth uncertainty is smaller
than this range would indicate because the P waveform at ATU is
not very well-modeled when the centroid depth is greater than
4 km. Also, at depths between 4 and 6 km and greater than 8 km,
the synthetic SH waves at SHL, CHG, BAG, and NHA show a small
upward motion which is not observed. At 4 km centroid depth,
the P value is 0.15; that is, we would make a type I error only
3 out of 20 times. At an α level of 0.1, β = 0.6 for the 4 km
solution. That is, we have 3 chances in 5 of being unable to
detect the difference in solutions. The use of α = 0.15
(β = 0.5) improves our chance of detecting the difference
slightly. From the above analysis, our best estimate of the
range in centroid depth is 1 to 4 km.

December 3, 1964

This event occurred on the Central Indian Ridge between the
Argo and Marie Celeste Fracture Zones (Figure 4.4). The ridge
in the epicentral region is marked by two parallel valleys, and
the earthquake epicenters are scattered over both valleys and the dividing ridge. This earthquake occurred near a bend in the eastern valley. Using first motions of the phases P and PKP, as well as the polarization of S waves, Banghar and Sykes [1969] proposed a normal-faulting mechanism with a large component of strike-slip motion for this event (314/60/216). Tsai [1969] estimated the seismic moment to be $3.4 \times 10^{25}$ dyn-cm, using the vertical Rayleigh-wave amplitude spectra at two stations, but this estimate was based on an assumed focal depth of 65 km. The inversion solution (Figure 4.5) is similar to the solution of Banghar and Sykes [1969], but with a smaller strike-slip component (302/46/249). The centroid depth is 1.6 km. The seismic moment is $6.8 \times 10^{24}$ dyn-cm, and the water depth is 3.4 km. All observed waveforms are very well fit by the synthetics. The nodal plane striking at 330°, subparallel to the trend of the ridge axis, is probably the fault plane.

The station coverage for this event is quite good: both P and SH waves have almost three-quadrant coverage. The focal mechanism for this earthquake is very stable: all 3 angles vary less than 10° for centroid depths held fixed between 1 and 10 km. Using the t-test, we estimate that the range in acceptable centroid depths is 1 to 6 km at $\alpha = 0.1$. We believe that this estimate is accurate because the probability of failing to detect the difference ($\beta$) is smaller than 0.4 for depths outside this range.

**August 15, 1966**

This earthquake occurred in the Carlsberg Ridge median
valley (Figure 4.6). Using P-wave first motions, Banghar and Sykes [1969] found a normal-faulting mechanism (272/28/270) for this event. One of the nodal planes in their solution is not well-determined. Tsai [1969] estimated a seismic moment of $1.26 \times 10^{25}$ dyn-cm for this event, using the vertical Rayleigh-wave amplitude spectrum at one station and an assumed focal depth of 45 km. The best inversion solution (Figure 4.7) is similar to that of Banghar and Sykes [1969] but has a small strike-slip component (294/51/260), a centroid depth of 3.4 km, and a seismic moment of $4.4 \times 10^{24}$ dyn-cm. The emergent character of the SH wave at WIN is revealed by the clear arrival of SV energy on the radial component. The shapes of P and SH waveforms are very well-fit by our inversion mechanism. However, using a higher value of $t^*$ for S waves would produce a better overall fit. The inferred water depth of 3.8 km is consistent with the ISC epicenter and the local bathymetry. The maximum water depth in the median valley is between 4.0 and 4.5 km. This earthquake may have occurred near one of the the inner walls of the rift mountains.

The station coverage for this event is fairly good, particularly for the P waves. The focal mechanism for this event is quite stable for centroid depths held fixed between 1 and 10 km. Using the t-test, we estimate the centroid depth to lie between 2 and 6 km at the 0.1 significance level. This estimate is probably accurate; $8 < 0.3$ for depths outside this range.
March 31, 1969

This earthquake occurred at the mouth of the Gulf of Suez at the northern end of the Red Sea (Figure 4.8). McKenzie et al. [1969] used P wave first motion polarities to infer a pure normal-faulting mechanism (310/34/270), with the T axis roughly perpendicular to the ridge axis. Ben Menaham and Aboodi [1971] found that the surface-wave radiation pattern fits a normal-faulting solution with small strike-slip component (327/44/310). They estimated the centroid depth to be about 15±5 km and the seismic moment to be 1.6 x 10^26 dyn-cm.

The best-fitting double-couple mechanism (294/37/271) indicated by body waveform inversion (Figure 4.9) is close to that of McKenzie et al. [1969]. The centroid depth of 6.2 km is shallower than that of Ben Menaham and Aboodi [1971], and the seismic moment (1.1 x 10^26 dyn-cm) is somewhat smaller than their value. The source structure used for this inversion is a halfspace with velocities appropriate to continental crust (V_p = 6.0 km/sec, V_s = 3.46 km/sec).

The station coverage for both P and SH waves is excellent, and the observed waveforms are all very well fit. The fault strike is controlled mostly by the SH waves. Small downward motions of the P waves at SDB and BUL constrain the dip of one of the nodal planes. Because the water wave for this event is small, we could not estimate the water depth using the P wavetrain. The water depth used in the inversion (0.2 km) is estimated from the bathymetric map.

There is some trade-off between the centroid depth and the
strike (which varies between 285° to 310°), but the dip (35°-40°) and slip (260°-280°) are quite stable for centroid depths fixed in the range 1-20 km. Using the t-test, we estimate that the centroid depth is between 5.5 and 8 km at $\alpha = 0.1$. For this earthquake, the waveform fits appear to be good between centroid depths of 5 and 12 km, with the source time function varying from 5 to 10 sec in duration.

Since the velocity structure in the epicentral area is not well-known, we have repeated the waveform inversion for this event with 2 other velocity structures. First, we used our standard oceanic structure and obtained a centroid depth of 12 km and a more complex time function. Second, we used a model consisting of a 15-km-thick continental crust ($V_p = 6.0 \text{ km/s}$, $V_s = 3.46 \text{ km/s}$) overlying a mantle halfspace ($V_p = 8.1 \text{ km/s}$, $V_s = 4.6 \text{ km/s}$), and we obtained a centroid depth of 9.2 km. A weak motivation for such a velocity structure is given by the proposal of McKenzie et al. [1970] that the crust in the Gulf of Suez might have been thinned to half its original thickness. Table 4.3 summarizes the inversion results of this earthquake with three different velocity structures; see also Appendix C. Since we are not certain about the velocity structure in this area, we cannot rule out a centroid depth as great as 9 or 12 km.

April 22, 1969

This earthquake occurred near the intersection between the southern end of the East Sheba Ridge and the Owen Fracture Zone (Figure 4.2). A very good fit to the observed P and SH waveforms from this earthquake (Figure 4.10) is obtained with a
normal faulting mechanism with some strike-slip component (350/63/305). The station distribution of P waves for this event is poor, but the P waveforms clearly require some strike-slip component. The clear upward first motions at stations SHL and CHG provide an excellent constraint for one of the nodal planes. The seismic moment is $2.9 \times 10^{24} \text{ dyn-cm}$, and the centroid depth is 5.1 km. Although the station coverage for P waves is poor, the source mechanism is very stable. For solutions in which we fixed the focal depth and inverted for other source parameters, the seismic moment varied by less than 20%, and strike, dip, and slip varied by less than $10^\circ$. Since this mechanism is well-constrained, the focal depth is probably quite accurate. We determined that the centroid depth of this earthquake is in the range 3-8 km at a significance level of 0.1. A secondary minimum in the residual-versus-depth curve occurs at 7 km, 1 km below the Moho. The focal mechanism (344/63/302) and seismic moment ($2.5 \times 10^{24} \text{ dyn-cm}$) for the mantle solution are very similar to those of the best solution. The t-statistic calculated for this depth is only slightly smaller than the threshold $t_a$. This solution becomes degraded and lies outside the uncertainty range, however, when a gradual Moho structure is used for the inversion. Based on the above analysis, we believe the centroid depth is between 3 and 6 km.

The water depth determined from the period of the water-column reverberations in the P wavetrains is about 4.9 km. The water depth in the median valley indicated on the bathymetric map is between 4.5 and 5.0 km. Unlike pure
normal-faulting ridge-axis events, the amplitudes of water waves for this earthquake are small compared with the first cycle of the P waves, probably due to its significant strike-slip component. If we take the nodal plane striking at 293° as the fault plane, then the strike agrees with the trend of the ridge axis. We think this earthquake probably occurred in a ridge-transform fault transition region where the crust and upper mantle are colder than in the normal ridge environment.

December 14, 1969

The December 14, 1969 earthquake is located about 300 km south of the Owen Fracture Zone on the Carlsberg Ridge (Figure 4.2). The seismicity pattern suggests that the easternmost of the 3 parallel valleys in the vicinity of the epicenter probably marks the ridge axis. The ISC epicenter for this earthquake lies within the next valley to the west, and thus it is probably an off-ridge event. South of the epicentral region, the Carlsberg Ridge bends and the median valley shifts to the southwest.

The focal mechanism (317/63/61, Figure 4.11) is characterized by thrust-faulting with some strike-slip component. The seismic moment is $10.3 \times 10^{24}$ dyn·cm, the centroid depth is 3.1 km, and the inferred water depth is 3.0 km. The station coverage for this event is very good and all observed waveforms are well fit. One nodal plane is very well-constrained by P waves at NIL, SHL, CHG, and SNG. The inferred P axis is nearly horizontal and is approximately aligned with the spreading direction. One nodal plane is
subparallel to the trend of the ridge, and the other fault plane striking N-S is parallel to a linear depression north of the epicentral region. The epicenter of this event is about 30 km west of the presumed median valley and thus within the African plate. We estimate that the age of the lithosphere in the epicentral region is about 3 M.y. from the half-spreading rate predicted by model RM2 [Minster and Jordan, 1978]. This event is thus one of the closest known thrust-faulting earthquake to an actively spreading ridge.

There is very little trade-off between focal mechanism and centroid depth for this event. Using the t-test, we estimate that the centroid depth is between 1 and 7 km at \( \alpha = 0.1 \). For this earthquake, the fits are quite good for fixed depths between 2 and 6 km. At depths greater than 7 km, the fit of the P waveforms deteriorates. We believe that the estimate of 1-7 km for the depth uncertainty range is probably accurate.

Another thrust-faulting earthquake (March 29, 1969, \( m_b = 5.6 \), at 10.38°N and 56.83°E) occurred in young lithosphere of the African plate near the intersection of the Owen Fracture Zone and the Carlsberg Ridge, but we were unable to match the long-period P and S waveforms in detail.

**June 28, 1972**

This earthquake occurred in the northern Red Sea about 15 km northwest of the epicenter of the March 31, 1969 earthquake (Figure 4.8). The inversion solution (Figure 4.12) is characterized by nearly pure normal faulting (288/40/260). The station distribution for Indian this event is fairly poor,
however, especially for the SH waves. The centroid depth is 6.1 km, and the seismic moment is $3.7 \times 10^{24}$ dyn-cm. The source structure used for the inversion is a continental crustal halfspace. Both the focal mechanism and centroid depth for this earthquake are very similar to those of the March 31, 1969 event. Due to the poor station coverage, there is a significant trade-off between focal mechanism and centroid depth; the strike varies from $260^\circ$ to $300^\circ$, the dip varies from $30^\circ$ to $50^\circ$, and the slip varies from $230^\circ$ to $300^\circ$ for centroid depths fixed between 1 and 15 km. The centroid depth range estimated by the t-test is 5-13 km. A closer examination of the seismograms shows that at depths greater than 11 km the first dilatational half cycles at NDI, CHG, SNG, MAL, PTO, VAL, and COP are too wide and too large. At a depth of 11 km, $\beta = 0.6$. The P value for a depth of 11 km is 0.13. We think that the true centroid depth is probably between 5 and 11 km.

September 7, 1972

This event occurred at the southern end of the Carlsberg Ridge just north of the Mabahiss Fracture Zone (Figure 4.13). The inversion solution (Figure 4.14) is a nearly pure normal-faulting mechanism (333/55/276). The nodal planes are subparallel to the ridge axis. All observed waveforms are well fit. The seismic moment is $3.5 \times 10^{24}$ dyn-cm, the centroid depth is 1.6 km, and the inferred water depth is 3.4 km. A secondary minimum in the residual-versus-depth curve occurs at 6 km, just beneath the Moho. The focal mechanism at this depth (319/46/257) has slightly more strike-slip component than the best solution. This deeper solution is worsened when a more
gradual Moho is used in the inversion; however, it is still within the uncertainty range at $\alpha = 0.1$. At $\alpha = 0.1$, the probability of making a type II error, $\beta$, is equal to 0.6 for the 6 km solution. That is, in 3 out of 5 times the t-test will not detect a difference between the two solutions. If we use $\alpha = 0.15$, however, then the difference between the mantle solution and the best-fitting solution is significant. For this reason, we suggest that centroid depth lies in the range 1-2 km at a significance level of 0.15.

January 3, 1980

This event occurred just north of the equator on the Carlsberg Ridge (Figure 4.13). The station coverage for this earthquake is excellent for both P and SH waves. From the waveform inversion we obtained a nearly pure normal-faulting mechanism (322/42/254). All P and SH waveforms are very well fit by the synthetics (Figure 4.15). The seismic moment is $8.8 \times 10^{24}$ dyn-cm, the centroid depth is 1.1 km, and the inferred water depth is 3.5 km. This earthquake is the shallowest in this study. The water depth suggests that the epicenter of this earthquake is a little to the south of the ISC location.

The shallowness of this earthquake is reflected in the lack of dilatational first motions in the P waves at SHL, HKC, GRM, PRE, and WIN. The focal mechanism is quite stable for centroid depths fixed in the range 1-10 km; strike, dip, and slip all varied less than 15°. The range in acceptable centroid depth at $\alpha = 0.1$ is 0.5-4 km and 6-9 km. A secondary solution occurs at a depth of 7 km; however, this solution becomes degraded and can
be rejected when a gradual Moho transition is used in the inversion. We therefore believe that the depth range 0.5 to 4 km reflects the true uncertainty in centroid depth.

October 7, 1981

The October 7, 1981 earthquake occurred on the Central Indian Ridge about 100 km south of the Vema Fracture Zone (Figure 4.4). Whether the epicenter lies along a short median-valley segment or is within the Indian plate is ambiguous on the basis of seismicity and bathymetry data alone (Figure 4.4). Using a semi-automated centroid tensor inversion and data from the GDSN, Dziewonski and Woodhouse [1983] obtained a mechanism combining normal and strike-slip faulting (356/44/318). Station coverage for P waves is rather poor for this event. We obtained a normal-faulting mechanism (311/50/257) with the T axis roughly perpendicular to the ridge axis. Most of the waveforms are well fit by the synthetic seismograms (Figure 4.16). The seismic moment of $4.5 \times 10^{24}$ dyn-cm estimated in our inversion is in excellent agreement with the moment of $4.7 \times 10^{24}$ dyn-cm obtained by Dziewonski and Woodhouse [1983]. The centroid depth is 8.2 km, and the inferred water depth is 3.4 km.

This earthquake is the deepest we have studied. There is a significant trade-off, however, between focal mechanism and centroid depth because of the poor P-wave coverage. A secondary minimum in the residual-versus-depth curve occurs at 1.9 km with a normal-faulting focal mechanism (336/56/282) that differs from the best solution and is closer to the solution of Dziewonski and Woodhouse [1983]; the seismic moment for this second
solution is $4.8 \times 10^{24}$ dyn-cm. Most of the waveforms are also quite well fit by the synthetic seismograms for this solution (Figure 4.17). With a gradual Moho transition included in the source region structure, both solutions become degraded but the mantle solution is still the best solution. Because of the poor station coverage, however, we do not have confidence in our centroid depth. The centroid depth is regarded as lying between 1 and 9 km.

October 25, 1982

This earthquake occurred on the Carlsberg Ridge just south of a fracture zone (Figure 4.6). Directly south of this area, the Carlsberg Ridge changes its direction from NW-SE to almost N-S. A centroid moment-tensor solution (293/73/260) by Dziewonski et al. [1983a] shows normal faulting with one steeply-dipping nodal plane. Our inversion solution (323/60/292) shown in Figure 4.18 has a greater strike-slip component than that of Dziewonski et al. Station coverage for this event is good, and all waveforms are well fit by the synthetic seismograms, especially the nodal character of the P wave at SHL. The seismic moment of $4.1 \times 10^{24}$ dyn-cm is somewhat smaller than the estimate of $7.4 \times 10^{24}$ dyn-cm obtained by Dziewonski et al. [1983a]. The centroid depth is 2.9 km, and the inferred water depth is 3.2 km. Neither the earthquake epicenters nor the bathymetric map clearly define the median valley in this area. However, the water depth suggests that this earthquake may have occurred a little to the north of the ISC location. We also investigated the nearby earthquake (October 18, 1964) inferred by Banghar and Sykes [1969] to
involve primarily normal faulting, but because it is a complex
double event we were unable to obtain a reliable centroid depth
using the procedures we have followed here for other ridge axis
earthquakes.

The range in acceptable centroid depth for the October 1982
earthquake is 1 to 7 km. A secondary minimum in the residual-
versus-depth curve occurs just below the Moho, at 6 km beneath
the sea floor. The residuals for the solutions at 2.9 km and
6 km depth are approximately equal. Also, the focal mechanism
(324/57/297) and seismic moment (3.9 x 10^{24} dyn-cm) for the
6-km-deep solution are almost identical to those of the solution
at 2.9 km. However, the mantle solution becomes degraded and can
be rejected at $\alpha = 0.1$ when a gradual Moho transition is assumed
for the inversion. We believe that this earthquake lies in the
crust and that the centroid depth is between 1 and 6 km.

December 8, 1982

This earthquake occurred on the West Sheba Ridge in the
western Gulf of Aden (Figure 4.19). Dziewonski et al. [1983a]
obtained a normal faulting mechanism (288/51/272), using the
method of centroid moment-tensor inversion. Our inversion
solution (290/56/254) shown in Figure 4.20 is very similar to
that of Dziewonski et al. [1983a] but has a small strike-slip
component. The seismic moment of $3.4 \times 10^{24}$ dyn-cm is very
similar to the $3.2 \times 10^{24}$ dyn-cm obtained by Dziewonski et al.
[1983a]. The centroid depth is 4.2 km, and the inferred water
depth is 2.2 km. Pearce [1981] found a focal depth of $5 \pm 2$ km
for a small ($m_b = 4.7$) ridge-axis earthquake about 300 km to
the east.
Although the number of stations used for the inversion is small, the stations are well-located to give adequate coverage for both P and SH waves. The range in acceptable centroid depths is 3 to 6 km at $\alpha = 0.1$. The plot of mean-square-residual versus centroid depth is very flat, but the t-statistic has a clear minimum in the depth range 3 to 6 km. This is because the standard deviation of the difference, $\sigma_d$, is very small.

April 11, 1983

This earthquake occurred in the median valley of the Carlsberg Ridge (Figure 4.6). Station coverage for this event is poor because many records have not yet been distributed. From the body waveform inversion, we obtained a normal faulting mechanism (Figure 4.21) with a significant component of strike-slip motion (304/51/239). The nodal plane striking at 304°, subparallel to the ridge axis, is probably the fault plane. The seismic moment is $3.6 \times 10^{24}$ dyn-cm, the centroid depth is 2.2 km, and the inferred water depth is 3.5 km.

There is significant trade-off between focal mechanism and centroid depth. For the centroid depth fixed in the range 0.5 km to 15 km, the strike varies from 290°-310°, the dip varies from 35°-55°, and the slip varies from 230° to 280°. The range in acceptable centroid depths is 0.5-10 km. A secondary minimum in the residual-versus-depth curve occurs at 8 km below the sea floor. The focal mechanism (303/50/243) and seismic moment ($3.4 \times 10^{24}$ dyn-cm) of the mantle solution are very similar to those of the best-fitting solution. All waveforms
are also quite well fit by the synthetics for the 8-km-deep solution. Using a gradual Moho transition, the residual increases slightly for the mantle solution, but the change is quite small. From the present data, the centroid depth must be regarded as lying between 0.5 and 10 km. More data may reduce the uncertainty range.

December 8, 1983

This event occurred in the median valley of the Carlsberg Ridge about 110 km southeast of the epicenter of the April 11, 1983, event (Figure 4.6). Dziewonski et al. [1984] obtained a normal faulting mechanism with a small strike-slip component (287/62/249). The available data give reasonably good P coverage, but SH coverage is poor. The number of stations used in the inversion is small because many records have not yet been distributed. The observed waveforms are very well fit (Figure 4.22) by a normal faulting mechanism (307/45/268) at a depth of 1.5 km. The seismic moment of $2.9 \times 10^{24}$ dyn-cm is smaller than the $3.9 \times 10^{24}$ figure dyn-cm obtained by Dziewonski et al. [1984]. The inferred water depth is 3.5 km.

The residual-versus-depth plot shows another local minimum at 7 km beneath the sea floor. Using the t-test, the estimated range in acceptable centroid depths is between 1 and 2 km and between 6 and 9 km. However, the solutions between 6 and 9 km show too much dilatational first motion at BUL. Also, using a gradual Moho transition the inversion solutions at these depths become worse than the crustal solution, and they can be rejected on the basis of the t-test.
DISCUSSION

The portion of the Indian Ocean under consideration in this chapter is quite complicated, and the tectonic history is not well-understood. At least three different plates are separated by the ridge system in this area. Spreading rates vary from 6 to 22 mm/yr [Minster and Jordan, 1978]. The source mechanisms of the 11 Indian Ocean ridge earthquakes show great similarity. The seismic moments span only a factor of 3, from 3 to $9 \times 10^{24}$ dyn-cm. As with ridge-axis earthquakes in the north Atlantic, the inferred T axes are all nearly horizontal and are approximately aligned with the spreading direction. However, the range of centroid depths is from 1.1 to 8.2 km, which is three times the range for Mid-Atlantic Ridge earthquakes (1.5 to 3.0 km).

For most of these earthquakes, there are two local minima in the residual-versus-centroid depth curve, one within the crust and one near the top of the mantle. The mantle solution usually degrades and falls outside the range of acceptable solutions when a gradual rather than a sharp Moho transition is adopted in the inversion. We believe that these mantle solutions are artifacts of the simple source velocity model used in our inversion. A summary of the depth uncertainty for the earthquakes examined in this chapter is given in Table 4.4.

For only one earthquake (October 7, 1981) is the best solution within the mantle, a result that holds even for a gradual Moho transition. Because the station coverage is quite poor for this earthquake, however, we do not have much
confidence in the centroid depth. If this earthquake actually occurred in the mantle and the rupture extended to the seafloor, the stress drop would be only 0.7 bar, an order of magnitude smaller than that of other ridge-axis events. We suspect either that this event is shallow or that rupture did not extend to the seafloor.

In this thesis, we have noticed that there is considerable trade-off between mechanism and depth if station coverage is not adequate. Generally, the earthquakes with good azimuthal coverage have smaller intervals of permissible centroid depth. We therefore believe that station coverage is probably the most important factor in determining earthquake centroid depth using long-period body waves. Excluding the earthquakes with inadequate coverage, the centroid depths for ridge-axis events in the Indian Ocean range from 1.1 to 4.2 km. This range is similar to that for events along the northern Mid-Atlantic Ridge.

Applying the same statistical test as for the Atlantic events, we have determined the uncertainty ranges of all of our Indian Ocean ridge earthquakes. From the statistical analysis, we conclude that the range of uncertainty in centroid depth for earthquakes on Indian Ocean ridges is about ± 4 km. This value is somewhat larger than that for the Mid-Atlantic Ridge earthquakes and is mainly the result of poorer station coverage. It should be recalled that we have underestimated the number of degrees of freedom in our analysis in order to be assured of independent data.
The water depths estimated from the P waveforms are compared in Table 4.2 with the maximum local depth of the median valley as indicated on bathymetric maps. The agreement between the two depth estimates is fairly good. Given the uncertainties in the epicentral coordinates and in the depths obtained from the contour map, most of these earthquakes seem to have occurred beneath the inner floor of the median valley. Again, we may not rule out the possibility that some of these events occurred near the inner walls of the rift valley (e.g., August 15, 1966). However, we are sure that none of these ridge earthquakes occurred beneath the upper rift mountains.

We also calculated the fault length, fault width, fault area, average slip, and stress drop for the earthquakes treated in this chapter (Table 4.5). Excluding the two Red Sea earthquakes and earthquakes with inadequate azimuthal coverage, the estimated fault area Lw ranges from 30 to 170 km$^2$. The average fault slip varies from 5 to 60 cm. The estimated stress drop lies between 20 and 80 bars. All these values are similar to those of the Mid-Atlantic Ridge earthquakes in Chapter 3.

The March 31, 1969 Red Sea earthquake is the largest earthquake in this study. At present, the interpretation of the nature of the seafloor beneath the Red Sea is still disputed. We have performed a body wave inversion with 3 different velocity structures and found very similar focal mechanisms (Table 4.3). The inversion solution with oceanic crust has the best overall fit and the deepest centroid depth. The signal-to-noise ratios are very high for all P waves. The other
Red Sea earthquake (June 28, 1972) occurred slightly to the north of the larger 1969 event. Although station coverage for the 1972 event is poor, its focal mechanism and centroid depth are similar to those of the 1969 earthquake. These Red Sea events are probably the deepest events we have studied. (The October 7, 1981 event is nominally deeper, but it has a very large depth uncertainty.) The half-spreading rate for the northern Red Sea, calculated from Minster and Jordan's kinematic model [1978], is about 5.9 mm/yr, except for the Azores the slowest spreading rate we have encountered in this study.

The two northern Red Sea earthquakes are important in determining the motion in the Gulf of Suez. Girdler [1966] suggested that the total opening of the Red Sea is about 190 km. On the other hand, Freund [1965] has shown from field evidence that only 110 km of slip has occurred along the Arava-Dead Sea fault. The question is how this missing motion is being accommodated. Picard [1966] observed that the Gulf of Suez is dominated by vertical tectonic movements and suggested that the Gulf is the expression of graben tectonics. Combining Picard's [1966] notion, McKenzie et al. [1970] hypothesized that the Gulf of Suez is opening in the same sense as the Red Sea, and that the missing slip is taken up in the Gulf by normal faulting. However, Freund [1965] suggested that folds in the Suez area imply some left-lateral motion along the Gulf. Abdel-Gawad [1969] suggested that the missing motion can be taken up entirely by left-lateral motion in the Gulf. Ben-Menahem et al. [1976] suggested that earthquake focal mechanisms are
consistent with the interpretation of folds in the Suez area by Freund [1965]. Later focal mechanism studies [Pearce, 1977, 1980] also supported the hypothesis of left-lateral motion in the Gulf of Suez. However, among the 4 earthquakes used to support the left-lateral motion, one earthquake occurred outside the Gulf near Cairo, about 300 km away from the other earthquakes. The other 3 earthquakes all occurred at the mouth of the Gulf of Suez very close to each other. We have studied 2 of these (March 31, 1969 and June 28, 1972) but found that they have nearly pure normal-faulting mechanisms. The remaining earthquake (January 12, 1972) is too small for long-period body-wave analysis. Our results are in qualitative agreement with the suggestion of opening of the Gulf of Suez by McKenzie et al. [1970]. However, we cannot rule out some strike-slip motion also occurring in the Gulf.
<table>
<thead>
<tr>
<th>Date</th>
<th>Origin Time</th>
<th>Latitude, °N</th>
<th>Longitude, °E</th>
<th>Half spreading rate, mm/yr</th>
<th>mb</th>
<th>Ms</th>
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<td>March 19, 1964</td>
<td>0942:35.8</td>
<td>14.42</td>
<td>56.36</td>
<td>10.8</td>
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<td>5.7b</td>
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<td>0350:01.7</td>
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<td>22.1</td>
<td>5.7</td>
<td>6.3b</td>
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<td>Aug. 15, 1966</td>
<td>1020:47.0</td>
<td>3.73</td>
<td>64.00</td>
<td>13.6</td>
<td>5.2</td>
<td></td>
</tr>
<tr>
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<td>0715:54.4</td>
<td>27.61</td>
<td>33.91</td>
<td>5.9</td>
<td>6.1</td>
<td>6.8</td>
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<tr>
<td>April 22, 1969</td>
<td>2234:40.0</td>
<td>12.80</td>
<td>58.25</td>
<td>11.2</td>
<td>5.6</td>
<td>5.2</td>
</tr>
<tr>
<td>Dec. 14, 1969c</td>
<td>1837:09.0</td>
<td>8.20</td>
<td>58.49</td>
<td>9.4</td>
<td>5.9</td>
<td>5.6</td>
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<td>June 28, 1972</td>
<td>0949:35.0</td>
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<td>33.80</td>
<td>5.9</td>
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<td>0255:00.0</td>
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<td>17.4</td>
<td>5.6</td>
<td>5.7</td>
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<td>67.17</td>
<td>16.3</td>
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<td>5.6</td>
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<td>66.30</td>
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<td>5.6</td>
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<td>Oct. 25, 1982</td>
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<td>65.93</td>
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<td>46.08</td>
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<td>5.1</td>
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Epicentral and magnitude data are taken from the ISC for 1964-April 1983 and the NEIS for December 1983.

a Half spreading rate calculated from model RM2 of Minster and Jordan [1978]
b Magnitude M from Rothe [1969]
c Near-ridge event.
Table 4.2 Source Mechanisms Obtained from Body Waveform Inversions

<table>
<thead>
<tr>
<th>Date</th>
<th>$M_0^a$</th>
<th>$t_s^c$</th>
<th>Strike</th>
<th>Dip</th>
<th>Slip</th>
<th>Centroid Depth$^d$</th>
<th>Water Depth$^e$</th>
<th>Depth of Median Valley$^f$</th>
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<td>March 19, 1964</td>
<td>4.6±0.5</td>
<td>4.0±1.0</td>
<td>281.9±3.0</td>
<td>38.1±4.7</td>
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<td>3.0-3.5</td>
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<td>2.4±0.8</td>
<td>301.6±3.7</td>
<td>46.4±2.1</td>
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<td>1.6±0.12</td>
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<td>3.5-4.0</td>
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<td>Aug. 15, 1966</td>
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<td>3.0±1.0</td>
<td>293.6±1.8</td>
<td>50.7±0.7</td>
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<td>3.4±0.12</td>
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<td>3.5-4.0</td>
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<td>March 31, 1969</td>
<td>106.3±3.4</td>
<td>9.6±1.6</td>
<td>294.0±1.7</td>
<td>36.9±0.4</td>
<td>271.3±1.0</td>
<td>6.2±0.08</td>
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<tr>
<td>April 22, 1969</td>
<td>2.9±0.3</td>
<td>4.0±1.0</td>
<td>349.9±1.5</td>
<td>63.4±0.9</td>
<td>304.7±2.1</td>
<td>5.1±0.07</td>
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<td>Dec. 14, 1969</td>
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<td>317.1±1.0</td>
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<tr>
<td>June 28, 1972</td>
<td>3.8±1.1</td>
<td>4.5±1.5</td>
<td>288.4±1.2</td>
<td>40.0±0.4</td>
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<td>6.1±0.04</td>
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<td>4.0±1.0</td>
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<td>42.1±2.6</td>
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<td>254.1±3.6</td>
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<td>305.2±4.0</td>
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<td>2.8±0.17</td>
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<td>1.5±0.20</td>
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</tr>
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</table>
Notes to Table 4.2

a) The range indicated for source parameters ($M_0$, strike, dip, slip, and centroid depth) is one standard deviation (formal error).

b) $x \times 10^{24}$ dyne-cm ($10^{17}$ N-m)

c) Total duration of the source time function, sec

d) Relative to the seafloor, km

e) Inferred from modeling the water-column reverberations in the later portions of the P wavetrains.

f) Local depth of the median valley inner floor indicated by bathymetric maps [Fisher et al., 1982; Laughton, 1975]

* Taken from bathymetric map.
Table 4.3 Inversion results for the March 31, 1969
northern Red Sea earthquake

<table>
<thead>
<tr>
<th>Source structure</th>
<th>Seismic moment, 10^24 dyn-cm</th>
<th>Double couple orientation (strike/dip/slip)</th>
<th>Centroid depth, km</th>
<th>Duration of source time function, sec</th>
<th>RMS residual, mm</th>
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<tr>
<td>1. Continental crustal halfspace</td>
<td>106.3</td>
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<td>5.36</td>
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<td>2. Single oceanic crustal layer over mantle halfspace</td>
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<td>297/38/272</td>
<td>12.0</td>
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<td>5.28</td>
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<td>3. 15 km continental crustal layer over mantle halfspace</td>
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<td>301/39/274</td>
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<td>Date</td>
<td>Range in centroid depth, in km, given by t-test, $\alpha = 0.1$</td>
<td>Secondary solution with sharp Moho</td>
<td>Secondary solution with gradual Moho</td>
<td>Adopted uncertainty ranges</td>
<td></td>
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<td>---------------------------------------------------------------</td>
<td>-----------------------------------</td>
<td>-------------------------------------</td>
<td>---------------------------</td>
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</tr>
<tr>
<td></td>
<td>Depth, km within uncertainty range for $\alpha = 0.1$?</td>
<td>Depth, km within uncertainty range for $\alpha = 0.1$?</td>
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<td>--</td>
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<td>--</td>
<td>--</td>
<td>5-11 0.15</td>
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<td>--</td>
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Table 4.4b Summary of Source Parameter Uncertainty Ranges for Indian Ocean Ridge Earthquakes

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<tr>
<th>Date</th>
<th>Strike, deg</th>
<th>Dip, deg</th>
<th>Slip, deg</th>
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<th>Fault length, a km</th>
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<th>Fault area, km²</th>
<th>Average slip, cm</th>
<th>Stress drop, bars</th>
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<td>5</td>
<td>2</td>
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<td>Dec. 8, 1983</td>
<td>12</td>
<td>4</td>
<td>50</td>
<td>10</td>
<td>10</td>
</tr>
</tbody>
</table>

a Estimated from duration of source time function. Rupture velocity is assumed to be 0.8 times shear wave velocity.

b Calculated from the assumptions: w < l and w < 2.8h.

c Calculated with rigidity and shear velocity corresponding to continental crustal material.

d Calculated with rigidity and shear velocity corresponding to mantle material.
FIGURE CAPTIONS

Figure 4.1. Location of ridge-axis earthquakes in the northwest Indian Ocean and Red Sea studied in this chapter, along with the overall distribution of seismicity (ISC epicenters, 1964-1982). Earthquakes in the regions of Iran and Pakistan are not shown because they are too numerous.

Figure 4.2. Detailed bathymetry of the East Sheba Ridge near the epicenters of the earthquakes of March 19, 1964, and April 22, 1969 (Mercator projection). Bathymetric contours, in km, are from Laughton [1975]; regions shallower than 3 km are shaded. Earthquake epicenters are from the ISC for 1964-1982; large symbols denote events with mb > 5.4. Fault plane solutions for events of this study are equal-area projections of the lower hemisphere; compressional quadrants are shaded.

Figure 4.3. Comparison of observed (solid line) and synthetic (dashed line) long-period P and SH waves for the March 19, 1964 earthquake, with the focal mechanism solution obtained from the inversion plotted on the lower focal hemisphere (equal-area projection). All amplitudes are normalized to an instrument magnification of 3000; the amplitude scales correspond to the waveforms that would be observed on an original seismogram for such an instrument. The two vertical lines delimit the portion of each time series, digitized at 0.5 second intervals, used in the inversion. Symbols for both types of waves are open circle, dilatation; solid circle, compression; cross, emergent
arrival. For SH waves, compression corresponds to positive motion as defined by Aki and Richards [1980]. Some first motions are shown for stations not used in the waveform inversion. The source time function obtained from the inversion is also shown.

Figure 4.4. Bathymetry and seismicity of the Central Indian Ridge near the epicenters of the earthquakes of December 3, 1964, and October 7, 1981. Bathymetric contours, in km, are from Fisher et al. [1982]; regions shallower than 3 km are shaded. See Figure 4.2 for further explanation.

Figure 4.5. Comparison of the observed and synthetic P and SH waves for the December 3, 1964, event. See Figure 4.3 for further explanation.

Figure 4.6. Bathymetry and seismicity of the Carlsberg Ridge median valley near the epicenters of the earthquakes of August 15, 1966; October 25, 1982; April 11, 1983; and December 8, 1983. Bathymetry is from Laughton [1975]; regions shallower than 3 km are shaded. See Figure 4.2 for further explanation.

Figure 4.7. Comparison of the observed and synthetic P and SH waves for the August 15, 1966, event. See Figure 4.3 for further explanation.

Figure 4.8. Bathymetry and seismicity of the central and northern Red Sea. The March 31, 1969 and June 28, 1972 earthquakes occurred in the Red Sea at the mouth of the Gulf of Suez. Bathymetric contours, in km, are from Laughton [1975]; regions deeper than 1 km are shaded. See Figure 4.2 for further explanation.
Figure 4.9. Comparison of the observed and synthetic P and SH waves for the March 31, 1969, event. See Figure 4.3 for further explanation.
Figure 4.10. Comparison of the observed and synthetic P and SH waves for the April 22, 1969, event. See Figure 4.3 for further explanation.
Figure 4.11. Comparison of the observed and synthetic P and SH waves for the December 14, 1969, event. See Figure 4.3 for further explanation.
Figure 4.12. Comparison of the observed and synthetic P and SH waves for the June 28, 1972, event. See Figure 4.3 for further explanation.
Figure 4.13. Bathymetry and seismicity of the Carlsberg Ridge near the epicenters of the earthquakes of September 7, 1972, and January 3, 1980. Bathymetry is from Laughton [1975] and Fisher et al. [1982]; regions shallower than 3 km are shaded. See Figure 4.2 for further explanation.
Figure 4.14. Comparison of the observed and synthetic P and SH waves for the September 7, 1972, event. See Figure 4.3 for further explanation.
Figure 4.15. Comparison of the observed and synthetic P and SH waves for the January 3, 1980, event. See Figure 4.3 for further explanation.
Figure 4.16. Comparison of observed and synthetic P and SH waves for the October 7, 1981, event. The synthetics are for a centroid depth of 8.2 km. Station NWAO is in the GDSN; see separate amplitude scale. See Figure 4.3 for further explanation.
Figure 4.17. Comparison of observed and synthetic P and SH waves for the October 7, 1981, event. The synthetics are for a centroid depth of 1.9 km. See Figure 4.3 for further explanation.

Figure 4.18. Comparison of observed and synthetic P and SH waves for the October 25, 1982, event. See Figure 4.3 for further explanation.

Figure 4.19. Bathymetry and seismicity of the West Sheba Ridge near the epicenter of the earthquake of December 8, 1982. Bathymetric contours, in km, are from Laughton [1975]; regions shallower than 3 km are shaded. See Figure 4.2 for further explanation.

Figure 4.20. Comparison of the observed and synthetic P and SH waves for the December 8, 1982, event. See Figure 4.3 for further explanation.

Figure 4.21. Comparison of the observed and synthetic P and SH waves for the April 11, 1983, event. See Figure 4.3 for further explanation.

Figure 4.22. Comparison of the observed and synthetic P and SH waves for the December 8, 1983, event. See Figure 4.3 for further explanation.
Figure 4.1
Figure 4.2
Figure 4.3
Figure 4.4
Figure 4.7
Figure 4.9
Figure 4.10
Figure 4.12
Figure 4.14
Figure 4.16
Figure 4.18
Figure 4.20
Figure 4.21
Figure 4.22
CHAPTER 5. SOME GENERALIZATIONS ON THE CHARACTERISTICS OF RIDGE-AXIS EARTHQUAKES

INTRODUCTION

In previous chapters, we presented the source mechanisms of 29 large ridge-axis earthquakes that occurred in the North Atlantic and Indian Oceans. In this chapter we consider the implication for ridge tectonics of the earthquake source mechanisms and focal depths determined in previous chapters, and we interpret our results in terms of what is known about rifting at slow-spreading ridges.

We begin with a retrospective view of the uncertainty in centroid depth from long-period body waveform inversion as a function of azimuthal coverage and seismic moment. We then apply the most reliable centroid depths for ridge-axis earthquakes to a discussion of the depth extent of brittle behavior beneath slow-spreading mid-ocean ridges. In particular, we consider the distribution of centroid depth and other source parameters of ridge earthquakes versus spreading rate.

UNCERTAINTY IN CENTROID DEPTH: ANOTHER LOOK

For each of the 29 ridge-axis normal-faulting events studied with body-wave inversion, we have estimated the range in uncertainty in centroid depth by the statistical methods discussed in Chapter 2. These uncertainty ranges are listed in Tables 3.4 and 4.4. As indicated by the numerical tests in
Chapter 2, we expect that the uncertainty in centroid depth should be generally greatest for events with the poorest azimuthal coverage of P and SH waveforms. That this expectation is borne out is displayed in Figure 5.1, in which the range in acceptable centroid depths (Tables 3.4a and 4.4a) is plotted versus the greatest azimuthal gap between neighboring stations in the coverage of P waveforms. For all events with a P-wave gap of less than 120°, the range in acceptable centroid depths is less than 4 km (i.e., the uncertainty is about ± 2 km). For all events with a P-wave gap of less than 180°, the range in acceptable depths is less than 5 km. Because the uncertainty estimated from our statistical method is generally asymmetrical about the best-fitting centroid depth, the 5 km range translates to uncertainty limits of about -2 and +4 km. If the range of acceptable centroid depths is plotted against the average gap for P- and SH-wave coverage, a similar relationship is shown (Figure 5.2). A corresponding but weaker relationship exists for SH-wave coverage alone. The above analysis underscores the importance of adequate azimuthal coverage, especially for P-wave coverage, in order to estimate reliably the centroid depth of shallow earthquakes.

In Figure 5.3, the range in uncertainty in centroid depth is plotted against seismic moment. As expected, the largest ranges in uncertainty are all associated with the smaller earthquakes of this study. All earthquakes with a seismic moment larger than 5 \times 10^{24} \text{ dyn-cm} have ranges in acceptable
centroid depths of less than 4 km (or uncertainties less than ± 2 km). The uncertainty range tends to be smaller for larger earthquakes, probably because more stations are used so that the focal mechanisms are better constrained.

COMPARISON OF SOURCE MECHANISM WITH HARVARD CMT SOLUTIONS

For the most recent earthquakes of this study, centroid moment tensor solutions have also been obtained by Dziewonski and Woodhouse [1983] and Dziewonski et al. [1983a,b, 1984] and may be compared with the results of our inversion. These comparisons are shown in Table 5.1. For most of these events, the double-couple orientations and seismic moments obtained from the two different methods agree quite well, particular considering the very long periods (> 45 sec) of the waves used in the CMT inversion [Dziewonski et al., 1981]. One event for which the results of the two methods disagree significantly is the October 7, 1981 Indian Ocean earthquake. However, a solution corresponding to a secondary minimum in the residual-versus-depth curve has a double-couple orientation more similar to that of the Harvard solution (see Chapter 4). From the above comparisons, we believe that the Harvard CMT solutions are generally quite good and are an excellent starting point for more detailed study with body waveform modeling.

MODELING WATER COLUMN REVERBERATIONS

The water depth above an earthquake epicenter can be constrained by the predominant period of water column reverberations, seen as prominent phases immediately following
the P-wave arrivals [e.g., Ward, 1979]. The water depth estimated from the P waveform is in general agreement with the maximum local depth of the median valley as indicated on available bathymetric maps. However, the amplitudes of observed water waves are usually much larger than the amplitudes of synthetics. Given the very shallow centroid depths, it is quite possible that fault rupture broke the surface and thus generated larger water waves than predicted by the synthetics. It is also possible that the median valley can act to focus water reverberations, producing larger amplitudes than predicted after the first cycle of such motion.

Another problem is that the first observed water reverberation can arrive later than in the synthetic waveform (e.g., the June 2, 1965 Mid-Atlantic Ridge earthquake). If the earthquake occurs beneath the center of the median valley, then the first water reverberation samples the deeper part of the median valley, while later water wave reverberations may sample a shallower "average" median valley depth. In this scenario, the first water wave arrives later than the synthetic generated with a flat layered model adjusted to provide a best overall fit to the full P waveform.

Numerical modeling of P waves from synthetic sources located in structures with non-planar seafloor topography are needed to test these suggestions.

DEPTH RANGE OF BRITTLE BEHAVIOR

The best-determined centroid depths constitute an important constraint on the depth extent of brittle behavior in
the axial regions of slow-spreading ridges. To discuss the implications of our results for typical ridge segments, it is necessary to exclude the 2 earthquakes in the northern Red Sea and the two events in the Azores region from our data set. The Azores events are likely to have occurred in basins formed by the rifting of Atlantic lithosphere [Searle, 1980]. The Red Sea events probably occurred in thinned continental crust [McKenzie et al., 1970; Cochran, 1983]. We also exclude all earthquakes with less reliable centroid depths. On the basis of the tests described in Chapter 2, we include only earthquakes with more than two-quadrant azimuthal coverage for both P and SH waves. The final data set with which we consider the depth extent of faulting at mid-ocean ridges includes 18 earthquakes, 12 in the Atlantic and 6 in the Indian Ocean.

All 18 earthquakes have centroid depths less than or equal to 4.2 km beneath the seafloor. The two deepest earthquakes (August 15, 1966 and December 8, 1982) both occurred in the Indian Ocean. On the basis of the numerical experiments and statistical tests described earlier, these centroid depths are estimated to have uncertainties of ± 3 km for the smaller events and somewhat lesser uncertainties for the larger events.

In the interpretation of these centroid depths, it is important to note that the centroid depth does not mark the maximum depth of fault slip. If the fault is rectangular or circular in shape and the slip is approximately uniform, then the centroid should lie near the geometric center of the fault. Since prominent water-column reverberations are evident in the
P waves from these events, it is reasonable to infer that fault rupture extends to the seafloor. Under these conditions, the maximum depth of fault slip is twice the centroid depth, or 2 to 8 km below the sea floor for the earthquakes of this study. Given the uncertainty in centroid depth and the possibility of irregular fault shape or nonuniform slip, however, the maximum depth of fault slip may have exceeded 8 km for some of these events.

Two microearthquake experiments conducted with a sufficient number of ocean-bottom sensors to resolve focal depths [Francis et al., 1978; Toomey et al., 1985] were conducted in regions of the Mid-Atlantic Ridge near epicenters of one or more of the large earthquakes of this study. It is of interest to compare the distribution of microearthquake focal depths with the centroid depths obtained for the large earthquakes.

Francis et al. [1978] reported the locations of a number of microearthquakes recorded during December 1974 by a network of 4 ocean-bottom seismometers near the eastern intersection of the St. Paul's Fracture Zone and the Mid-Atlantic Ridge median valley. Focal depths concentrated in the ranges 0-1 km and 6-8 km. Francis et al. [1978] commented particularly on the absence of median valley events with focal depths between 1 and 5 km, which they attributed to the presence of an intracrustal layer highly cracked and weakened by pervasive hydrothermal circulation. The earthquake of June 28, 1979 (Figure 3.21) occurred beneath the median valley at the southern end of the zone of microearthquakes described by Francis et al. [1978].
The centroid depth of 2.5 km and the large water-column reverberations shown in the P waveforms (Figure 3.22), if taken at face value, would indicate that seismic slip was concentrated between the seafloor and 5 km depth. If correct, this result suggests that the microearthquakes in 1974 may have outlined the top and bottom of the slip zone of the large earthquake that followed 4.5 years later. Of course, the uncertainties in both the centroid depth and the microearthquake focal depths also permit a coincidence in depth of the two measures of activity.

The second microearthquake experiment was conducted early in 1982 in the Mid-Atlantic Ridge median valley near 23°N, approximately 90 km south of the Kane Fracture Zone [Toomey et al., 1985]. Because of the large network (10 stations) and knowledge of the local crustal velocity structure [Purdy and Detrick, 1985], the focal depths of many of the microearthquakes were determined to within ±1 km formal error at 95% confidence. Most of the well-resolved hypocenters were located between 5 and 8 km beneath the seafloor. Three large earthquakes in our study occurred in the same region in 1962 and 1977 (Figure 3.2). Their fault plane solutions are quite similar to composite fault plane solutions obtained for two clusters of microearthquakes [Toomey et al., 1985], indicating that both sets of events are probably responding to the same tectonic process. The centroid depths of the large earthquakes are 1.2 to 2.5 km, values most readily interpreted as indicating that slip extended between the seafloor and about 5 km depth. Most of the 1982 microearthquakes thus occurred deeper than this nominal slip zone of the
large earthquakes 5 and 20 years earlier, but the uncertainties in both sets of quantities also permit the large earthquake slip zones to have extended well into the depth range of microearthquake activity. Unfortunately, no microearthquake experiment has been conducted in the Indian Ocean that would allow us to compare the focal depths of large earthquakes and microearthquakes.

If the difference in depth between the slip zones of the large earthquakes and the regions of most intense microearthquake activity along these two segments of the Mid-Atlantic Ridge is real, then two possible explanations present themselves. One possibility is that microearthquakes in the median valley outline a strong portion of the crust [Kanamori, 1981] between 1-2 and 5-6 km depth, and that fault rupture during large earthquakes initiates near the base of the crust and propagates upward to the seafloor. If a large proportion of fault area or slip were concentrated in the upper crust for these large events, the depth of the centroid would be shallower than half the maximum depth of fault slip. According to this scenario, the microearthquake focal depths provide the most accurate measure of the maximum depth of brittle behavior. A second possibility is based on the premise that the microearthquakes at 5-8 km depth mark the strongest portion of the median valley faults. In this scenario, the strong zone acts as a barrier [Aki, 1977] to the downward propagation of fault slip during the large earthquakes of this study. Continued microearthquake activity is the result of the large stress
differences remaining near the base of the slip zone of the larger events. This scenario admits the possibility of a rare, very large ridge-axis earthquake that breaks through the barrier and has a significantly greater centroid depth and moment than any of the events of this study. While we favor the first scenario, we are unable to exclude the second one. Possible avenues of research to distinguish between them include searching for directivity effects in short-period waveforms, monitoring aftershocks shortly following a large event, and discovering an unusually large ($M_o \gtrsim 10^{26}$ dyne-cm) normal-faulting earthquake along ridge segments.

**CHOICE OF FAULT PLANES**

Generally, the preferred fault plane may be determined from a fault-plane solution if independent data such as the trends of aftershocks, ground breakage, or geological structures are available. For ridge-axis earthquakes, the choice of fault plane can often be made on the criterion that the fault strike be similar to the local trend of the ridge. For ridge earthquakes with nearly pure normal faulting mechanisms, of course, this method fails because the strike directions for the two nodal planes are nearly identical. In this section, we try another approach based on the predicted plate motions from model RM2. This approach gives us a chance both to test the relation between RM2 spreading directions and the geometry of ridge-axis faulting and to quantify the selection of fault plane and associated parameters (dip angle, slip direction).
For each of the normal-faulting earthquakes we have studied, we first determined the horizontal direction normal to the local small circle about the RM2 spreading pole. We then chose as the fault plane the nodal plane with the smallest difference between this direction and the observed strike direction. Tables 5.2 and 5.3 show these calculations for the Atlantic and Indian Ocean events, respectively. For most of these events the choice of nodal plane is very clear. Figures 5.4 and 5.5 show a histogram of observed-versus-predicted strike angles for the Atlantic and Indian Ocean earthquakes, respectively. The July 4, 1966, Azores event has a large discrepancy between observed and predicted strike directions, supporting the suggestion [Searle, 1980] of oblique spreading along the Azores spreading center. The average differences between the observed and theoretical strike angles are $-1.1^\circ$ and $+4.1^\circ$ for the Atlantic and Indian Ocean events, respectively; the rms differences are, respectively, 11.8° and 8.9°. These small differences suggest that RM2 provides fairly reliable estimates of the strike directions of major normal faults at divergent plate boundaries.

The fault planes chosen on the basis of agreement with RM2 are the same as the fault planes that would be chosen on the basis of agreement with local ridge trend except for the August 15, 1966, and January 3, 1980 Indian Ocean earthquakes. In the epicentral regions of these two earthquakes, according to available bathymetric maps, the local ridge axis appears to intersect the nearest transform fault at an oblique angle. Histograms of the dip angles of the preferred fault planes
are also shown in Figures 5.4 and 5.5. The average dip is 45.3 \( \pm 5.5^\circ \) for Mid-Atlantic Ridge earthquakes and 45.4\(^\circ \) \( \pm 7.3^\circ \) for ridge events in the Indian Ocean.

**STRESS DROP**

Stress drops were estimated in Chapters 3 and 4 for all earthquakes. These stress drops are very sensitive to assumed fault widths. For very shallow events, this problem is magnified. For example, when we have a centroid depth of 3 \( \pm 2 \) km and assume that faulting extends to the surface, the calculated stress drop for a 1-km-wide fault width is 25 times that for a 5-km-wide fault width. Figure 5.6 shows the calculated stress drop plotted versus centroid depth; the error bars for the stress drop are estimated from the errors in the centroid depth and in the duration of the source time function. It is clear that the large error bars prevent any conclusions from being drawn from this data set. The biggest factors in the large uncertainties are the assumptions that the centroid depth is half the maximum depth of fault slip and that the earthquake ruptured to the seafloor.

In an effort to minimize the effect of the assumed fault width on the calculated stress drop, we tested two other assumptions concerning fault widths. First, we assumed that all faults have a constant width of 2.8 km. Figure 5.7 shows inferred stress drop versus centroid depth under this assumption. No clear trend can be drawn from this plot. We next assumed that all faults have the same aspect ratio and that the fault width is half the fault length inferred from the
duration of source time function. Again, a plot of stress drop versus centroid depth does not yield any clear trend under this assumption (Figure 5.8). From these analyses, we believe either that stress drop does not vary systematically with centroid depth for ridge-crest earthquakes or that any such variation cannot be resolved.

SEISMIC MOMENT VS. CENTROID DEPTH

We also tested for a relationship between centroid depth and seismic moment. We again use only the earthquakes with adequate azimuthal coverage. For the Mid-Atlantic Ridge earthquakes, these quantities show no significant correlation (Figure 5.9). The Indian Ocean events, however, appear to show an inverse correlation between seismic moment and centroid depth (Figure 5.10). Combining the Atlantic and Indian Ocean data, the inverse correlation appears somewhat stronger (Figure 5.11). All larger ridge-axis earthquakes \(M_o \sim 10^{25} \text{ dyn-cm}\) have centroid depths of 1-2 km, while smaller events occur throughout the centroid depth range 1-5 km. One possible explanation is that slip during the larger earthquakes occurs preferentially in the uppermost crust, so that the centroid depths of the larger earthquakes are biased shallow relative to the mid-depth of fault rupture.

DEPENDENCE OF SOURCE PARAMETERS ON SPREADING RATE

We next consider the variation of source parameters with local spreading rate. In Figure 5.12, the centroid depth of Mid-Atlantic Ridge events is plotted against the half-spreading rate as predicted by the RM2 velocities of Minster and Jordan
[1978]. The 90% confidence ranges for centroid depths of events with adequate azimuthal coverage are indicated by vertical bars. A similar plot of centroid depth versus half-spreading rate for the Indian Ocean events is given in Figure 5.13. If we exclude events with less than adequate azimuthal coverage, the combined Atlantic and Indian data show that the earthquake faulting tends to shoal with increasing spreading rate (Figure 5.14).

This provides direct evidence for a general decrease with spreading rate in the maximum thickness of the zone of brittle behavior for slow-spreading ridges. The maximum depth of earthquake faulting probably outlines the depth of the brittle-ductile transition beneath the ridge axis. It is likely that the lithosphere is colder and mechanically stronger beneath the slowest-spreading ridge segments because there is more time for the lithosphere to cool down between major episodes of crustal magma injection and volcanism.

We next consider the variation of seismic moment with local spreading rate. Since seismic moment is more reliably determined than the centroid depth, we include all 29 ridge-axis earthquakes of this study. Figure 5.15 shows a plot of seismic moment versus half-spreading rate as given by RM2. The combined Atlantic and Indian Ocean data sets illustrate that the larger events \((M_0 > 9 \times 10^{24} \text{ dyn-cm})\) all occurred on ridges with half-spreading rates less than 14 mm/yr, while smaller events occurred for all spreading rates in the range 1-25 mm/yr. The largest earthquake (March 31, 1969) occurred in the Red Sea where the spreading rate is about 6 mm/yr. This plot supports
the hypothesis that the lithosphere is colder and mechanically stronger beneath slow-spreading ridges. However, more data are needed to confirm this hypothesis.

NEAR-RIDGE EARTHQUAKES

In Chapter 4, we noted that a thrust faulting earthquake (December 14, 1969) occurred at 3 km depth in very young oceanic lithosphere near the northern Carlsberg Ridge, just south of a small fracture zone. One of the nodal planes, striking at about 318°, is nearly parallel to the local trend of the ridge. As noted in Chapter 4, another event with a similar setting and mechanism occurred on March 29, 1969; a further thrust event (mb = 6.2) occurred near the Central Indian Ridge (18.88°S, 67.35°E) on November 6, 1984 [Dziewonski et al., 1985x]. Bergman and Solomon [1984] and Wiens and Stein [1984] have noted several near-ridge earthquakes which are characterized by thrust-faulting, at least two (November 25, 1965 and May 9, 1971) of which have P axes oriented nearly perpendicular to the nearby ridge axis. Both of these events occurred in the Pacific Ocean, in lithosphere older than 9 m.y. From bathymetry and earthquake epicenters, the December 14, 1969, earthquake occurred about 30 km from the median valley in lithosphere about 3 M.y. old.

Bergman and Solomon [1984] showed that a P axis oriented along the spreading direction is not a common feature for near-ridge earthquakes and suggested that there is little evidence of a "ridge-push" compressive stress associated with the cooling and subsidence of oceanic lithosphere [Lister, 1975;
Dahlen, 1981] to ages of at least 35 M.y. Further, the magnitude of the stress field associated with "ridge push" is negligible at lithosphere ages of a few million years [Dahlen, 1981]. The most likely cause for thrust events with P axes parallel to spreading direction in very young oceanic lithosphere is thermoelastic stress. Bratt et al. [1985] have shown that, when the approximate effect of cooling due to hydrothermal circulation is taken into account, thermal stress models predict a near-ridge region where $\sigma_1$ is nearly horizontal and large in magnitude in the uppermost 5 to 10 km of the lithosphere. Within the region of thrust faulting, the P axes are predicted to be in the direction of spreading.
Table 5.1 Comparison of the Harvard CMT solutions and body waveform inversion results for recent ridge-crest earthquakes

<table>
<thead>
<tr>
<th>Date</th>
<th>Harvard CMT solution</th>
<th>This study</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Double couple</td>
<td>Seismic momenta</td>
</tr>
<tr>
<td>7 October 1981</td>
<td>356/44/318</td>
<td>4.7</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>29 January 1982</td>
<td>353/61/246</td>
<td>8.1</td>
</tr>
<tr>
<td>25 October 1982</td>
<td>293/73/260</td>
<td>7.4</td>
</tr>
<tr>
<td>8 December 1982</td>
<td>288/51/272</td>
<td>3.2</td>
</tr>
<tr>
<td>12 May 1983</td>
<td>354/50/259</td>
<td>8.1</td>
</tr>
<tr>
<td>8 December 1983</td>
<td>287/62/249</td>
<td>3.9</td>
</tr>
</tbody>
</table>

\( a \times 10^{24} \) dyn cm

\( b \) Second-best solution (see Chapter 4)
Table 5.2. Comparison of observed strike angle with that predicted from plate velocity for Mid-Atlantic Ridge earthquakes

<table>
<thead>
<tr>
<th>Date</th>
<th>Nodal planes Strike/Dip</th>
<th>Observed-predicted strike angle</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 June 3, 1962</td>
<td>4/43 179/48</td>
<td>-6 -11</td>
</tr>
<tr>
<td>2 June 2, 1965</td>
<td>187/57* 335/38</td>
<td>-3 -35</td>
</tr>
<tr>
<td>3 Nov. 16, 1965</td>
<td>25/46 211/44</td>
<td>14 21</td>
</tr>
<tr>
<td>4 July 4, 1966</td>
<td>135/49 309/41</td>
<td>-30 -36</td>
</tr>
<tr>
<td>5 April 20, 1968</td>
<td>328/34 121/59</td>
<td>-11 -37</td>
</tr>
<tr>
<td>7 May 31, 1971</td>
<td>209/40 51/52</td>
<td>-1 22</td>
</tr>
<tr>
<td>8 April 3, 1972</td>
<td>8/48* 212/44</td>
<td>2 26</td>
</tr>
<tr>
<td>9 June 6, 1972</td>
<td>20/52* 227/41</td>
<td>9 36</td>
</tr>
<tr>
<td>10 June 28, 1977</td>
<td>1/46 204/46</td>
<td>-9 14</td>
</tr>
<tr>
<td>11 June 28, 1977</td>
<td>1/44 201/48</td>
<td>-10 11</td>
</tr>
<tr>
<td>12 Jan. 28, 1979</td>
<td>199/44 20/46</td>
<td>16 17</td>
</tr>
<tr>
<td>13 April 22, 1979</td>
<td>17/52 210/39</td>
<td>5 19</td>
</tr>
<tr>
<td>14 June 28, 1979</td>
<td>346/40* 153/51</td>
<td>-9 -22</td>
</tr>
<tr>
<td>15 Jan. 29, 1982</td>
<td>9/44 197/47</td>
<td>-1 7</td>
</tr>
<tr>
<td>16 May 12, 1983</td>
<td>188/45 341/48</td>
<td>-2 -29</td>
</tr>
</tbody>
</table>

For each earthquake, the fault plane listed first is the one preferred on the basis of best agreement with plate kinematic model RM2 [Minster and Jordan, 1978].

* indicates fault planes chosen on the basis of agreement with local ridge trend.
Table 5.3. Comparison of observed strike angle with that predicted from plate velocity for Indian Ocean ridge-axis earthquakes

<table>
<thead>
<tr>
<th>Date</th>
<th>Nodal planes Strike/Dip</th>
<th>Observed-predicted strike angle</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 March 19, 1964</td>
<td>282/38* 138/58</td>
<td>17 19</td>
</tr>
<tr>
<td>2 Dec. 3, 1964</td>
<td>151/48* 304/46</td>
<td>6 -25</td>
</tr>
<tr>
<td>3 Aug. 15, 1966</td>
<td>129/40 294-51*</td>
<td>0 -15</td>
</tr>
<tr>
<td>4 March 31, 1969</td>
<td>112/53 294/37</td>
<td>8 10</td>
</tr>
<tr>
<td>5 April 22, 1969</td>
<td>113/43* 350/63</td>
<td>-7 50</td>
</tr>
<tr>
<td>6 June 28, 1972</td>
<td>288/40 121/51</td>
<td>4 17</td>
</tr>
<tr>
<td>7 Sept. 7, 1972</td>
<td>143/36 333/55</td>
<td>10 20</td>
</tr>
<tr>
<td>8 Jan. 3, 1980</td>
<td>322/42 163/50*</td>
<td>12 32</td>
</tr>
<tr>
<td>9 Oct. 7, 1981</td>
<td>151/42 311/50</td>
<td>7 -14</td>
</tr>
<tr>
<td>10 Oct. 25, 1982</td>
<td>323/60 104/36</td>
<td>16 -24</td>
</tr>
<tr>
<td>11 Dec. 8, 1982</td>
<td>290/56* 137/37</td>
<td>-13 14</td>
</tr>
<tr>
<td>12 April 11, 1983</td>
<td>305/48* 161/48</td>
<td>-4 32</td>
</tr>
<tr>
<td>13 Dec. 8, 1983</td>
<td>130/45 307/45</td>
<td>0 -3</td>
</tr>
</tbody>
</table>

For each earthquake, the fault plane listed first is the one preferred on the basis of best agreement with plate kinematic model RM2 [Minster and Jordan, 1978].

* indicates fault planes chosen on the basis of agreement with local ridge trend.
FIGURE CAPTIONS

Figure 5.1. The greatest azimuthal gap in P waveform coverage versus the centroid depth uncertainty for all ridge-axis earthquakes studied. Depth uncertainties are listed in Tables 3.4a and 4.4a.

Figure 5.2. Average azimuthal gap (average of the gaps in P waveform and SH waveform coverage) versus the centroid depth uncertainty for all ridge-axis earthquakes studied.

Figure 5.3. Seismic moment versus centroid depth uncertainty for all ridge-axis earthquakes studied.

Figure 5.4. (Top) Histogram showing the difference between observed strike angle and the perpendicular to the spreading direction predicted by model RM2 [Minster and Jordan, 1978] for Mid-Atlantic Ridge earthquakes. (Bottom) Histogram of dip angles of the fault planes selected on the basis of minimizing the difference between predicted and observed strike angles.

Figure 5.5. (Top) Histogram showing the difference between observed strike angle and the perpendicular to the spreading direction predicted by model RM2 for the Indian Ocean ridge-axis earthquakes. (Bottom) Histogram of dip angles of the fault planes selected on the basis of minimizing the difference between predicted and observed strike angles.

Figure 5.6. Stress drop versus centroid depth for ridge-axis earthquakes. Error bars for stress drop are calculated
from the errors in source time function duration (Tables 3.2 and 4.2) and centroid depth (Tables 3.4a and 3.4b). Only events with a greatest azimuthal gap less than 180° for both P and SH waves are used.

Figure 5.7. Stress drop versus centroid depth under the assumption that the fault width is constant at 2.8 km. Such a width corresponds to a vertical extent of faulting of 2 km.

Figure 5.8. Stress drop versus centroid depth under the assumption that the fault width is half the fault length. The fault length is calculated from the duration of the source time function (Tables 3.2 and 4.2).

Figure 5.9. Seismic moment versus centroid depth for Mid-Atlantic Ridge earthquakes. Only events with a greatest azimuthal gap less than 180° for both P and SH waves are plotted.

Figure 5.10. Seismic moment versus centroid depth for Indian Ocean ridge-axis earthquakes. Only events with a greatest azimuthal gap less than 180° for both P and SH waves are used.

Figure 5.11. Seismic moment versus centroid depth for the combined data sets from Atlantic and Indian Ocean ridges.

Figure 5.12. Centroid depth versus half spreading rate for Mid-Atlantic ridge earthquakes. The solid circles represent earthquakes with a greatest azimuthal gap less than 180° for both P and SH waves. The vertical bar represents the uncertainty in the centroid depth at 90% confidence.
Figure 5.13. Centroid depth versus half spreading rate for Indian Ocean ridge-axis earthquakes. See Figure 5.12 for further explanation.

Figure 5.14. Centroid depth versus spreading rate for the combined data set from Atlantic and Indian Ocean ridges. Only events with a greatest azimuthal gap less than $180^\circ$ for both P and SH waves are included.

Figure 5.15. Seismic moment versus half spreading rate. All 29 ridge-axis earthquakes of this study are included.
Figure 5.1
Figure 5.2
Figure 5.3
ATLANTIC OCEAN

Figure 5.4
Figure 5.5
STRESS DROP, bars

FAULT WIDTH = 2.8km

figure 5.7
Figure 5.10
Figure 5.11

SEISMIC MOMENT, $x10^{24}$ dyne-cm

CENTROID DEPTH, km

0 5 10 15 20

0 1 2 3 4 5 6
ATLANTIC OCEAN

HALF SPREADING RATE, mm/yr

Figure 5.12
Figure 5.13
HALF SPREADING RATE, mm/yr

Figure 5.14
Figure 5.15

HALF SPREADING RATE, mm/yr

SEISMIC MOMENT × 10²⁴ dyne-cm
CHAPTER 6. THESIS SUMMARY AND SUGGESTIONS FOR FURTHER WORK

In this chapter we summarize the major results of this thesis. We have conducted detailed source studies for 29 large earthquakes along mid-ocean ridge spreading centers in the Atlantic and Indian Oceans. A major goal of our study has been to determine reliable centroid depths for ridge-axis earthquakes. We found that

(1) For long-period body waveform inversion, good azimuthal coverage, particularly for P waves, is very important in determining reliable centroid depths for shallow earthquakes. For best results the largest azimuthal gap should be less than 180° for both P and SH waves.

(2) In general, larger earthquakes have smaller uncertainties in source parameters, particularly centroid depth, probably because more stations are used and the mechanism is better constrained than for smaller events.

(3) All mid-ocean ridge earthquakes have mechanisms characterized predominantly by normal faulting on fault planes dipping at angles of about 45°. The T axes are all parallel to the direction of spreading.

(4) The water depths estimated from the predominant period of water-column reverberations in the P wavetrains indicate that the ridge earthquakes in this study all occurred beneath the inner floor of the median valley.

(5) The centroid depths of earthquakes with good azimuthal coverage are all very shallow, ranging from 1 to 4.2 km. If the
centroid depth marks the mean depth of fault slip, then
earthquake faulting extended to 2-8 km beneath the seafloor for
these events. A limited amount of slip at greater depth cannot
be excluded.

(6) The greatest centroid depths shoal with increasing
spreading rate. This provides direct evidence for a general
decrease with spreading rate in the maximum thickness of the
zone of brittle behavior for slow-spreading ridges.

(7) An example was found of thrust faulting in very young
(2 m.y.) lithosphere in the northern Indian Ocean; the axis of
greatest compression is parallel to the spreading direction.
It is likely that this event was caused by thermoelastic stress.

(8) The centroid depths for two earthquakes in the northern
Red Sea support the view that seismic faulting accompanying
continental rifting extends to greater depths than does seismic
faulting along the axes of oceanic spreading centers.

The findings of this thesis lead to several suggestions
for further work:

Numerical simulations. For each of the very shallow
earthquakes we have studied in this thesis, we have spent
considerable effort investigating the resolution of centroid
depth. We mainly carried out hypothesis testing, however, and
the depth uncertainties we have obtained are measures of
precision. To infer the accuracy obtainable from long-period
body waveform inversion, we should carry out numerical
simulations. Nabelek [1984] has carried out some such numerical
tests. A more detailed follow-on study, perhaps using a Monte
Carlo method, is necessary to document the true capability of the method. There are formalized procedures for learning from the details of a Monte Carlo experiment, details that may provide better estimates, suggest new measures of misfit, or both.

**Extension to other regions.** This study raises several issues concerning the tectonic interpretation of ridge crest earthquakes, including the relationship of microearthquakes and large earthquakes along a given ridge-axis segment, the relationship of earthquake activity within the inner median valley to fault structures in the adjacent rift mountains, and variations of some source properties with spreading rate. To address these issues will require extensions of this study to additional events, other regions, and an expanded frequency band. Spreading centers suitable for studies paralleling those of this thesis include the southern Mid-Atlantic Ridge, the Southwest Indian Ocean Ridge, the American-Antarctic Ridge, the Gorda Ridge, and the Galapagos Spreading Center.
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APPENDIX A.
TESTING HYPOTHESES

INTRODUCTION

A hypothesis is a statement or claim about the state of nature. Scientists often have hypotheses about their particular studies, hypotheses that need substantiation or verification—or rejection—for one purpose or another. To this end, they gather data and let the data support or cast doubt on their hypothesis. The process is termed the "testing of hypotheses." To accept a hypothesis does not mean the same thing to all investigators or in all situations. In a statistical experiment, accepting a hypothesis does not mean that the hypothesis is "proved" in any rigorous sense, because the data in a sample give only incomplete information about a population and can easily be misleading. To "accept" may mean to "believe," in the weak sense of being willing to act, or to proceed in some way, as though the hypothesis were true.

The hypothesis to be tested is called the null hypothesis, and the set of other states of nature or models admitted as possible for a given experiment is called the alternative hypothesis. A null hypothesis is often that of "no difference" in the effects of two treatments, an application that gave rise to the adjective "null." The null hypothesis will usually be denoted by $H_0$ and the alternative either by $H_0$ or $H_1$. 
TYPE I AND TYPE II ERRORS

The application of a statistical test may lead to the wrong conclusion. These wrong inferences are called errors of Types I and II [Folks, 1981]:

Type I error: Rejecting $H_0$ when $H_0$ is true.
Type II error: Accepting $H_0$ when $H_0$ is false.

Naturally, we would like to eliminate both errors, but this is impossible. Instead, we are forced to settle for controlling the probability of such errors. Introducing a little terminology from hypothesis testing, let

\[ \alpha = \text{probability of Type I error} = \text{significance level} \]
\[ \beta = \text{probability of Type II error} \]
\[ 1 - \beta = \text{power of the hypothesis test} \]

We would like both $\alpha$ and $\beta$ to be small, but as the $\alpha$ risk is reduced, the $\beta$ risk becomes larger; when we make $\beta$ smaller, $\alpha$ tends to become larger [Folks, 1981]. In application, it is a good idea to determine not merely whether the given hypothesis is accepted or rejected at the specified significance level, but also to determine the smallest significance level ($P$ value) at which the hypothesis can be rejected for the given evidence [Choi, 1978]. This number enables others to make a decision based on the significance level of their choice.

PROCEDURE

The components of a hypothesis test are as follows:

1) **Null and alternative hypotheses** for a parameter of an assumed probability distribution (i.e., $\mu_d = 0$ and $\mu_d > 0$, respectively, in our case).
(2) $\alpha$ and $\beta$, or $\alpha$ and $n$ (number of samples).

(3) A test statistic.

(4) A decision rule.

(5) A random sample from the population.

(6) Calculation of the test statistic.

(7) A decision either to accept or reject the null hypothesis obtained from the calculated value of the test statistic and the decision rule.

TESTS CONCERNING THE MEAN OF A NORMAL DISTRIBUTION WITH UNKNOWN VARIANCE

Let $X$ denote a random sample from a normally-distributed population with an unknown variance $\sigma^2$, whose mean $\mu$ is known to be either $\mu_0$ or greater than $\mu_0$. The problem is to test against

$H_0$: $\mu = \mu_0$.

$H_1$: $\mu > \mu_0$.

Let $\bar{x}$ and $S^2$ be the mean and variance of $X$; then the test statistic $t = \frac{\bar{x} - \mu}{S/\sqrt{N}}$ follows a t-distribution with $n-1$ degrees of freedom. For an intuitively reasonable test, we choose $H_1$ if $t$ is larger than some constant; otherwise we choose $H_0$. The critical region of such a test would be $t > t_\alpha$. An assessment of the errors for this test is as follows:

$\alpha = P$ (rejecting $H_0$ | $H_0$ is true)

$= P$ ($t > t_\alpha$ | $\mu=\mu_0$)

$= P \left( \frac{\bar{x} - \mu_0}{S/\sqrt{N}} > t_\alpha \right)$

In order to calculate $\beta$, let us assume that $\mu = \mu_1$ where
\( \mu_1 > \mu_0 \) when \( H_1 \) is true.

\[
\beta = P(\text{rejecting } H_1 | H_1 \text{ is true})
\]
\[
= P(t < t_\alpha | \mu = \mu_1)
\]
\[
= P\left( \frac{x - \mu_0}{S/\sqrt{N}} < t_\alpha \bigg| \mu = \mu_1 \right)
\]
\[
= P\left( \frac{x - \mu_1}{S/\sqrt{N}} + \frac{\mu_1 - \mu_0}{S/\sqrt{N}} < t_\alpha \right)
\]
\[
= P\left( t < \frac{\mu_0 - \mu_1}{S/\sqrt{N}} + t_\alpha \right)
\]

Note that \( \left( \frac{\mu_1 - \mu_0}{S/\sqrt{N}} \right) \) is a constant, making it possible to compute \( \beta \) from the t-table.

Let us now elaborate on the trade-off between \( \alpha \) and \( \beta \).

Figure A.1 shows the relationship between Type I and Type II errors. In this figure, the upper distribution represents the null hypothesis, \( H_0 \), with a true value for \( \mu \) of \( \mu_0 \). The Type I error, \( \alpha \), is shown on the right side of this distribution. The lower distribution represents the alternate hypothesis, \( H_1 \), with a true value for \( \mu \) of \( \mu_1 \), which is greater than \( \mu_0 \). The vertical dashed line represents the point at which \( H_0 \) can be rejected with \( \alpha \) risk and accepted with \( \beta \) risk. When the dashed line moves to the right, \( \alpha \) decreases but \( \beta \) increases; moving the line to the left increases \( \alpha \) and decreases \( \beta \).
FIGURE CAPTION

Figure A.1. Relation between the probability of type I error, \( \alpha \), and that of type II error, \( \beta \) [Folks, 1981]. Usually \( \alpha \) is chosen a priori and \( \beta \) is calculated after the test. \( \beta \) is a function of the difference \( \mu_1 - \mu_0 \) and is important in assessing the power of the test (see text).
Distribution under the null hypothesis; $$H_0 : \mu = \mu_0$$

Distribution under the alternative hypothesis; $$H_1 : \mu > \mu_0$$

Acceptance region

Rejection region

for null hypothesis

Figure A.1
APPENDIX B.
RESOLUTION OF CENTROID DEPTHS

In this Appendix we compile plots of mean-squared residual $R^2$ versus centroid depth and the test statistic $t$ versus centroid depth for all earthquakes of this study. Epicentral information for the earthquakes is given in Tables 3.1 and 4.1. The quantities $R^2$ and $t$ are defined in equations (12) and (14), respectively. The residual at each depth is normalized by the mean-squared waveform over the same time window. The horizontal bar in the plot of $R^2$ versus centroid depth indicates the 90% confidence interval about the best-fitting depth found in the body-waveform inversion. The horizontal line in the plot of $t$ versus centroid depth is the 90% confidence value of the statistic; when two horizontal lines are shown the lower line indicates the 85% confidence value.
ATLANTIC OCEAN
RIDGE EVENTS
JUNE 2, 1965

Centroid Depth, km

- Chart 1: Graph of t vs. Centroid Depth
- Chart 2: Graph of $R^2$ vs. Centroid Depth
Centroid Depth, km

\[ R^2 \]

\[ t \]
APRIL 20, 1968

Graph showing the relationship between Centroid Depth (km) and time (t) with a marked increase and decrease followed by a steady period. The graph also includes a plot of $R^2$ against Centroid Depth (km), showing a gradual increase.
JUNE 28, 1977

Centroid Depth, km

$R^2$

Centroid Depth, km
JUNE 28, 1977

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Graph 1:
- X-axis: Centroid Depth, km
- Y-axis: Some unspecified units
- Data points showing a trend

Graph 2:
- X-axis: Centroid Depth, km
- Y-axis: $R^2$
- Data points showing a trend with a horizontal line indicating a range
JANUARY 28, 1979

Graph showing the relationship between t and Centroid Depth, km.

Graph showing the relationship between $R^2$ and Centroid Depth, km.
APRIL 22, 1979

Centroid Depth, km

\( t \)

\( R^2 \)

Centroid Depth, km
JANUARY 29, 1982

Centroid Depth, km

Centroid Depth, km

R^2

0 5 10

2 4 6

0 .4 .6

Jan 29, 1982

Centroid Depth, km

R^2
MAY 12, 1983

Centroid Depth, km

\[ t \]

\[ .5 \]

\[ R^2 \]

\[ .3 \]
INDIAN OCEAN RIDGE EVENTS
DECEMBER 3, 1964

![Graph showing t vs. Centroid Depth, km and R² vs. Centroid Depth, km.](image)
MARCH 31, 1969

Centroid Depth, km

$R^2$

Centroid Depth, km
APRIL 22, 1969

Centroid Depth, km

\[ R^2 \]
JUNE 28, 1972

Centroid Depth, km

\[ R^2 \]

Centroid Depth, km
SEPTEMBER 7, 1972

Centroid Depth, km

$R^2$
APRIL 11, 1983

Centroid Depth, km

$t$

$R^2$

Centroid Depth, km
DECEMBER 3, 1983

![Graph showing relationship between t and R² over Centroid Depth, km.](image)
APPENDIX C.

COMPARISON OF INVERSION SOLUTIONS WITH
ALTERNATIVE SOURCE VELOCITY STRUCTURES

In this Appendix we compare the observed and synthetic seismograms for the July 4, 1966 Azores earthquake and the March 31, 1969 Red Sea earthquake using alternative source structures, as described in the text of Chapters 3 and 4, respectively.
FIGURE CAPTIONS

Figure C.1. Comparison of observed (solid line) long-period P and SH waves for the July 4, 1966, earthquake with synthetic waveforms (dashed line) from the secondary solution (10.8 km centroid depth) for a source structure consisting of a uniform crust over a mantle halfspace. The focal mechanism solution obtained from the inversion is plotted on the lower focal hemisphere (equal-area projection). All amplitudes are normalized to an instrument magnification of 3000; the amplitude scales correspond to the waveforms that would be observed on an original seismogram from such an instrument. The two vertical lines delimit the portion of each time series used in the inversion. Symbols for both types of waves are open circle, dilatation; solid circle, compression; cross, emergent arrival. For SH waves, compression corresponds to positive motion as defined by Aki and Richards [1980]. The source time function obtained from the inversion is also shown. The best-fitting solution for this source structure is shown in Figure 3.9.

Figure C.2. Comparison of observed P and SH waves for the July 4, 1966, earthquake with synthetic waveforms from the best-fitting solution (6.3 km centroid depth) when the source structure consists of a 2-layer crust over a mantle halfspace [Searle, 1976]. See Figure C.1 for further explanation.
Figure C.3. Comparison of observed P and SH waves for the July 4, 1966, earthquake with synthetic waveforms from the secondary solution (10.8 km centroid depth) for a source structure consisting of a 2-layer crust over a mantle halfspace. See Figure C.1 for further explanation.

Figure C.4. Comparison of observed P and SH waves for the July 4, 1966, earthquake with synthetic waveforms from the best-fitting solution (3.7 km centroid depth) for a crustal halfspace structure model. See Figure C.1 for further explanation.

Figure C.5. Comparison of observed P and SH waves for the July 4, 1966, earthquake with synthetic waveforms from the secondary solution (7.9 km centroid depth) for a crustal halfspace structure model. See Figure C.1 for further explanation.

Figure C.6. Comparison of observed P and SH waves for the July 4, 1966, earthquake with synthetic waveforms from the best-fitting solution (6.8 km centroid depth) for a mantle halfspace structure model. See Figure C.1 for further explanation.

Figure C.7. Comparison of observed P and SH waves for the July 4, 1966, earthquake with synthetic waveforms from the secondary solution (11.2 km centroid depth) for a mantle halfspace structure model. See Figure C.1 for further explanation.

Figure C.8. Comparison of observed long-period P and SH waves for the March 31, 1969, earthquake with synthetic
waveforms from the best-fitting solution for a source structure consisting of a 15-km-thick crust over a mantle halfspace. See Figure C.1 for further explanation.

Figure C.9. Comparison of the observed P and SH waves for the March 31, 1969 earthquake with synthetic waveforms from the best-fitting solution for an oceanic structure model. See Figure C.1 for further explanation.
JULY 4, 1966 (1-layer crust, secondary minimum)

Source Time Function

Figure C.1
Figure C.2
Figure C.3
JULY 4, 1966 (Crustal half-space, best solution)

Figure C.4
Figure C.5
Figure C.6

JULY 4, 1966 (Mantle half-space, best solution)
Figure C.7

JULY 4, 1966 (Mantle half-space, secondary minimum)

P WAVES

SH WAVES

Source Time Function (sec)

mm

sec

60

mm

sec

60

Figure C.7
Figure C.8
Figure C.9