Paper 4:

Seismic methods for determining earthquake source parameters and lithospheric structure

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ABSTRACT

The seismologic methods most commonly used in studies of earthquakes and the structure of the continental lithosphere are reviewed in three main sections: earthquake source parameter determinations, the determination of earth structure using natural sources, and controlled-source seismology. The emphasis in each section is on a description of data, the principles behind the analysis techniques, and the assumptions and uncertainties in interpretation. Rather than focusing on future directions in seismology, the goal here is to summarize past and current practice as a companion to the review papers in this volume.

Reliable earthquake hypocenters and focal mechanisms require seismograph locations with a broad distribution in azimuth and distance from the earthquakes; a recording within one focal depth of the epicenter provides excellent hypocentral depth control. For earthquakes of magnitude greater than 4.5, waveform modeling methods may be used to determine source parameters. The seismic moment tensor provides the most complete and accurate measure of earthquake source parameters, and offers a dynamic picture of the faulting process.

Methods for determining the Earth's structure from natural sources exist for local, regional, and teleseismic sources. One-dimensional models of structure are obtained from body and surface waves using both forward and inverse modeling. Forward-modeling methods include consideration of seismic amplitudes and waveforms, but lack the formal resolution estimates obtained with inverse methods. Two- and three-dimensional lithospheric models are derived using various inverse methods, but at present most of these methods consider only travel-times of body waves.

Controlled-source studies of the Earth's structure are generally divided by method into seismic refraction/wide-angle reflection and seismic reflection studies. Seismic refraction profiles are usually interpreted in terms of two-dimensional structure by forward modeling of traveltimes and amplitudes. The refraction method gives excellent estimates of seismic velocities, but relatively low resolution of structure. Formal resolution estimates are not possible for models derived from forward modeling, but informal estimates can be obtained by perturbing the best-fitting model. Inversion

methods for seismic refraction data for one-dimensional models are well established, and two- and three-dimensional methods, including tomography, have recently been developed.

Seismic reflection data provide the highest resolution of crustal structure, and have provided many important geological insights in the past decade. The acquisition and processing of these data have been greatly advanced by the hydrocarbon exploration industry. However, reliable crustal velocity control is generally lacking, and the origin of deep crustal reflections remains unclear, resulting in non-unique interpretations. A new form of lithospheric seismology has recently emerged that combines the advantages of seismic refraction and seismic reflection profiles, and the distinction between the two methods is steadily diminishing.

Major challenges for future work will be the collection of data that are more densely sampled in space, and the development of interpretation methods that provide quantitative estimates of the uncertainties in the calculated models.

INTRODUCTION

A wide range of seismologic methods are used to study the structure of the lithosphere and the seismicity of the Earth. This chapter reviews seismic methods in three sections: the determination of earthquake source parameters; the determination of the Earth's structure using natural (earthquake) sources; and the determination of structure using man made (controlled) sources. After a brief historical introduction, data-analysis techniques and assumptions and uncertainties associated with each method are discussed. Seismologic theory and practice are far too extensive to be completely covered in the available space; existing texts serving that function include Richter (1958), Grant and West (1965), Dobrin (1976), Claerbout (1976, 1985), Telford and others (1976), Cerveny and others (1977), Coffeen (1978), Garland (1979), Aki and Richards (1980), Ben-Menahem and Singh (1981), Lee and Stewart (1981), Sheriff and Geldart (1982, 1983), Kennett (1983), Bullen and Bolt (1985), Tarantola (1987), Waters (1987), and Yilmaz (1987). The purpose here is to provide a critical review of the methods used in other chapters of this volume, and particularly to discuss their uncertainties and limitations, and to indicate where more detailed information may be obtained.

Historical background

Wherever documents of human history are uncovered, hints of interest in earthquakes can be found. Chinese records of earthquakes date from the Chou dynasty (ca. 1122 to 221 B.C.), the Bible recounts the collapse of the walls of Jericho (ca. 1100 B.C.), and Aristotle (b. 384 B.C.) devoted his attention to the study and classification of earthquakes. Until the end of the 19th century, the cause of earthquakes was rarely associated with faults.

Fault scarps observed after two California earthquakes (Fort Tejon, 1857, and Owens Valley, 1872) played an important role in establishing the earthquake-fault association (Richter, 1958). The 1906 San Francisco earthquake was the most thoroughly studied of its time (Lawson, 1908), and provided data for the elastic rebound concept of Reid (1910), which is still the generally accepted theory of earthquake rupture and the origin of seismic waves (Howell, 1986).

The transition from a purely descriptive study of earthquakes to a quantitative study came with the development, in 1892, of a reliable seismograph by John Milne (1850-1913) in Japan. In 1897, Robert Oldham was the first to merge seismologic observations with theory when he identified compressional, shear, and surface waves on an actual seismogram. At the time of the 1906 San Francisco earthquake, there were already dozens of seismographs of various designs operating worldwide, most prominently in Germany, Italy, and Japan. Some 50 years later it had become clear that a more reliable and uniform determination of world seismicity required a network of standardized seismic stations. Thus, the World-Wide Standardized Seismographic Network (WWSSN), consisting of 120 continuously recording stations, began operation in the 1960s (Oliver and Murphy, 1971). It was recognized as well that regional networks generally provide much-improved accuracy and sensitivity for determining local seismicity compared with worldwide networks. By the 1970s, about 100 microearthquake networks were operating at various locations throughout the world (Lee and Stewart, 1981).

The history of early controlled-source seismic investigations of Earth's structure is summarized by DeGolyer (1935), Bullen and Bolt (1985), James and Steinhart (1966), and Sheriff and Geldart (1982). An Irish scientist, Robert Mallet, apparently was the first, in 1848, to report on the use of an explosive source in seismology (Aki and Richards, 1980, p. 268). One hundred years passed before a systematic program of controlled-source seismic profiling of the deep crust was begun. In the meantime, earthquake traveltime data were used to define the crust/mantle boundary (Mohorovicic, 1909) and seismic velocity layering within the crust (Conrad, 1925).

Economic incentives spurred the development of seismic refraction exploration methods in the 1920s. In 1924 an oil field was discovered in Texas using seismic refraction fan-shooting techniques. Marine seismic refraction investigations were initiated some 10 years later when Ewing and others (1937) recorded explosions at four stations located south of Woods Hole, Massachusetts. Subsequently, deep-water marine measurements were begun in the 1950s by Ewing and his colleagues. In Germany, seismic refraction measurements of the deep continental crust began in 1947, and in the United States the active pursuit of the method was begun shortly thereafter by the Carnegie Institution of Washington (Tuve, 1951; Tatel and Tuve, 1955). As of the early 1960s, seismic refraction profiles were collected with regularity both in the United States and worldwide.

The history of seismic reflection profiling is much more closely tied to hydrocarbon exploration activities (Sheriff and Geldart, 1982). The modern seismic reflection system as we know it today dates from the 1930s when field parties recorded 10 to 12 channels of data on photographic paper, using some signal processing and enhancement capabilities provided by vacuum tube circuits. The common depth point (CDP) recording method (Mayne, 1962, 1967) was invented in 1950 as a way of providing further signal enhancement, and it came into routine use in the mid-1960s. The VIBROSEIS method was developed in 1953 as an alternative to explosive sources; by the mid-1970s it accounted for about 60 percent of land recording and today is the most common source for on-land seismic reflection studies.

Investigations of the use of the seismic reflection method for studies of the deep crust began about the same time as seismic refraction studies, but required many more years to become a routine method (Mintrop, 1949). Much of the earliest work was done in Europe, but results in the United States were also reported by Junger (1951) and Dix (1965). Later, experiments on three continents-in Australia by the Bureau of Mineral Resources, in Canada (Kanasewich and Cumming, 1965; Clowes and others, 1968), and in Germany (Meissner, 1967; Dohr, 1970) - gave conclusive evidence for intracrustal and Moho reflections at near-vertical incidence. The Consortium for Continental Reflection Profiling (COCORP) undertook a pilot study of deep reflection seismology in northern Texas in 1975 (Oliver and others, 1976), and shortly thereafter a continuously operating seismic crew was engaged. To date, more than 10,000 km of data have been collected in the United States over abroad geographic area by COCORP and other institutions (Brown and others, 1986; Phinney and Roy-Chowshury, this volume; Smithson and Johnson, this volume). The high resolution of structure obtained with the seismic reflection method accounts for its extensive application.

Reflection profiling is also highly effective in the marine environment due to the excellent coupling of the source (air-guns) with the water medium, the lack of surface waves, and the relatively uniform bottom conditions, which minimize static corrections. These advantages have been effectively exploited by scientists in the British Institutions for Reflection Profiling Syndicate (BIRPS; Matthews and Cheadle, 1986; Mat thews and others, 1987). High-quality marine data exist for both the Atlantic and Pacific margins of the United States (Hutchinson and others, 1988; Trehu and others, this volume; Couch and Riddihough, this volume; Phinney and Roy-Chowdhury, this volume), and recently, deep reflection data were collected on the Great Lakes in a joint U.S.-Canadian project (Behrendt and others, 1988).

DETERMINATION OF EARTHQUAKE SOURCE PARAMETERS

Earthquake data

There are three basic sources of data used in earthquake studies: permanent regional seismic networks, temporary dense networks of portable seismographs, and global seismic observatories (broad-band seismic installations). Permanent regional networks are used where the level of seismic activity or potential seismic hazard is high enough to pose a significant public safety risk. Typically, each seismograph consists of a single vertical- component 1-Hz geophone, and high-gain analogue telemetry electronics, which are of limited use for extracting waveform information due to signal clipping; however, some networks use three-component digital recording and telemetry. There are some 50 seismic networks in the United States, each with an average of 30 stations spaced some 30 to 100 km apart. The largest permanent network is operated in California by the U.S. Geological Survey (USGS) and the California Institute of Technology; it consists of about 600 stations, with the densest coverage within 100 km of the San Andreas fault.

After a major earthquake occurs, a temporary network of portable seismographs may be installed in the epicentral region. Until recently, these networks, typically 10 to 20 seismographs spaced 5 to 10 km apart, consisted of vertical-component smoked-drum or pen-and-ink seismographs. Currently, these networks generally include three-component seismographs (e.g., Archuleta and others, 1982; Borcherdt and others, 1985) and accelerometers that provide more useful information for the determination of locations, magnitudes, and seismic moments than do vertical-component seismographs.

For larger earthquakes (magnitude 4.5 or greater), the study of broad-band waveforms (teleseisms) recorded at global seismic observatories provides additional information on the hypocentral depth and nature of faulting. Teleseismic data are particularly useful because they include waves emanating at a wide range of angles from the earthquake source, and usually include seismic information over a wider frequency band (e.g., Langston and Helmberger, 1975; Dziewonski and others, 1981).

Analysis of earthquake data

A three-component seismograph will register P- and S-wave (among other) arrivals from an earthquake; the differences in their respective arrival times are sufficient to calculate the distance from the seismograph to the earthquake, provided that the crustal velocity structure is approximately known. If three local seismographs with abroad azimuthal distribution record the earthquake, the hypocentral location can be determined accurately, if again the crustal velocity structure is well known. In practice, most micro-earthquakes are recorded by regional vertical-component seismographs, with only limited three-component recordings, and the local velocity structure is subject to numerous uncertainties. The impact of these limitations can be reduced by recording the earthquakes with many proximal stations. Therefore, one philosophy behind micro-earthquake network design is to operate as many vertical-component seismographs as possible, distributed evenly over the study area. An alternative philosophy is to operate three-component, broadband, high-dynamic range seismographs distributed more broadly, and to use the complete waveforms to determine locations and source parameters. Richter (1958) and Lee and Stewart (1981) have discussed standard methods for locating earthquakes using network data.

The magnitude of an earthquake may be defined in various ways, depending on the portion and type of seismogram used in the measurement. Local earthquakes are generally assigned a magnitude in one of two ways. Local magnitude (M_L) is based on the maximum amplitude (normalized to a distance of 100 km) recorded on a standard Wood-Anderson seismograph (Richter, 1958). Coda magnitude (M_c), designed to carry M_L to values of less than 4.0, is derived from the duration of the recorded seismogram (the coda) after correction for the event-seismograph distance and regional attenuation variations (Lee and Stewart, 1981). Teleseismic earthquakes are often assigned a body-wave magnitude (m_b)), which is determined from the amplitudes of short-period body waves, and/or a surface-wave magnitude (M_s), which is determined from the amplitude of surface waves with a period of 20 sec, a measurement requiring data from a long-period seismograph (Richter, 1958). However, deep earthquakes do not necessarily generate large surface waves, which makes M_s and M_L estimates unreliable. Moment magnitude (M_w) is a magnitude scale based directly on the radiated energy (Hanks and Kanamori, 1979) and is related to the seismic moment (M_o, measured in dyne-centimeters) by:

$$M_w = (2/3) \log M_o - 10.67 \qquad (1)$$

and

$$M_o = \mu SA \tag{2}$$

where μ is the shear modulus in $dyne/cm^2$; S is the average fault slip in cm; A is the area of the fault surface in cm^2.

The earliest instrumental recordings of earthquakes showed a systematic variation in first ground motion with azimuth, and it was soon recognized that this first-motion variation could be used to infer the sense and nature of faulting at the source. A double-couple source (a pair of perpendicular coupled forces) generates either compressions or dilatations, depending on the azimuth with respect to the fault plane (Fig. 1A). Null records, or nodal phases, are observed where the motion changes from one sense to the other. The pressure axis (P-axis) is defined as the center of the dilatational quadrant and the tensional axis (T-axis) is defined as the center of the compressional quadrant. These relations follow from basic principles of rock mechanics. As can be inferred from Figure 1A, the most important criterion for a reliable determination of an earthquake's focal mechanism is that it be recorded at many azimuths and distances, thereby giving uniform coverage of the focal sphere. Figure 1B illustrates three common fault types, and the sense of P-wave first motions (compressional or dilatational) in lower hemisphere projections, as is most commonly used. Focal mechanisms provide a powerful means of determining fault planes and slip directions, and consequently, crustal and lithospheric plate motions. Focal mechanisms for a variety of fault geometries and tectonic settings

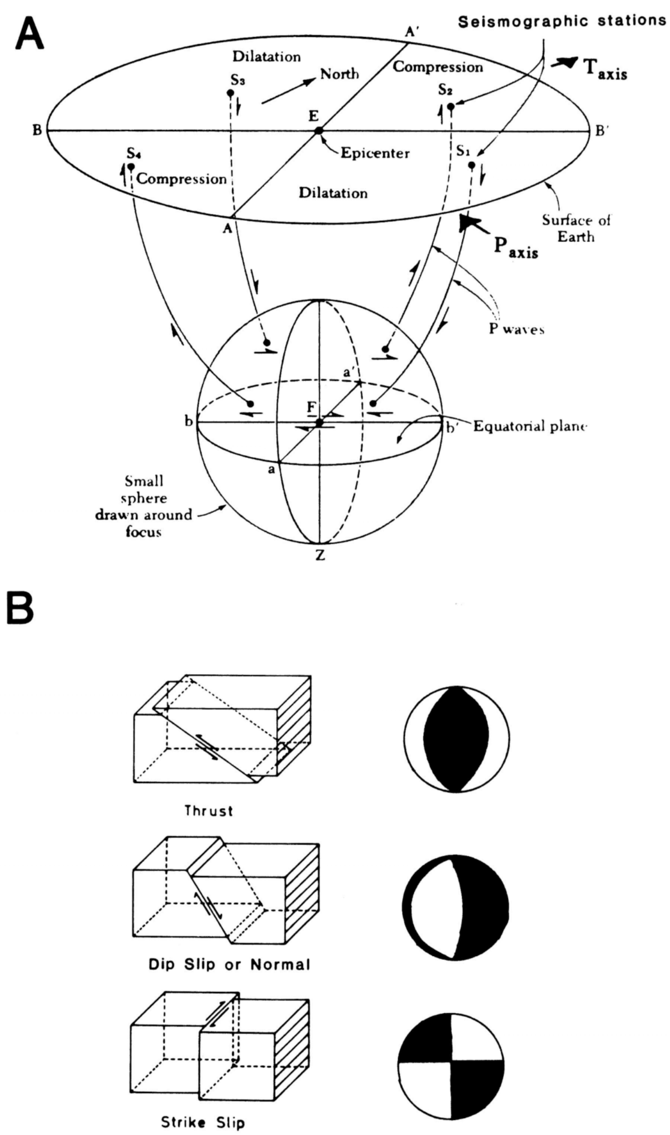


Figure 1. A, Upper-hemisphere compressional and dilatational quadrants in a three-dimensional view around a pure strike-slip earthquake focus. The focal mechanism of an earthquake can be determined if first-motion data are available covering all four quadrants. From Bolt (1982). B, Block models of three simple fault motions (thrust, normal, and strike slip), and the corresponding lower-hemisphere projection of first motions (black for compression, white for dilation).

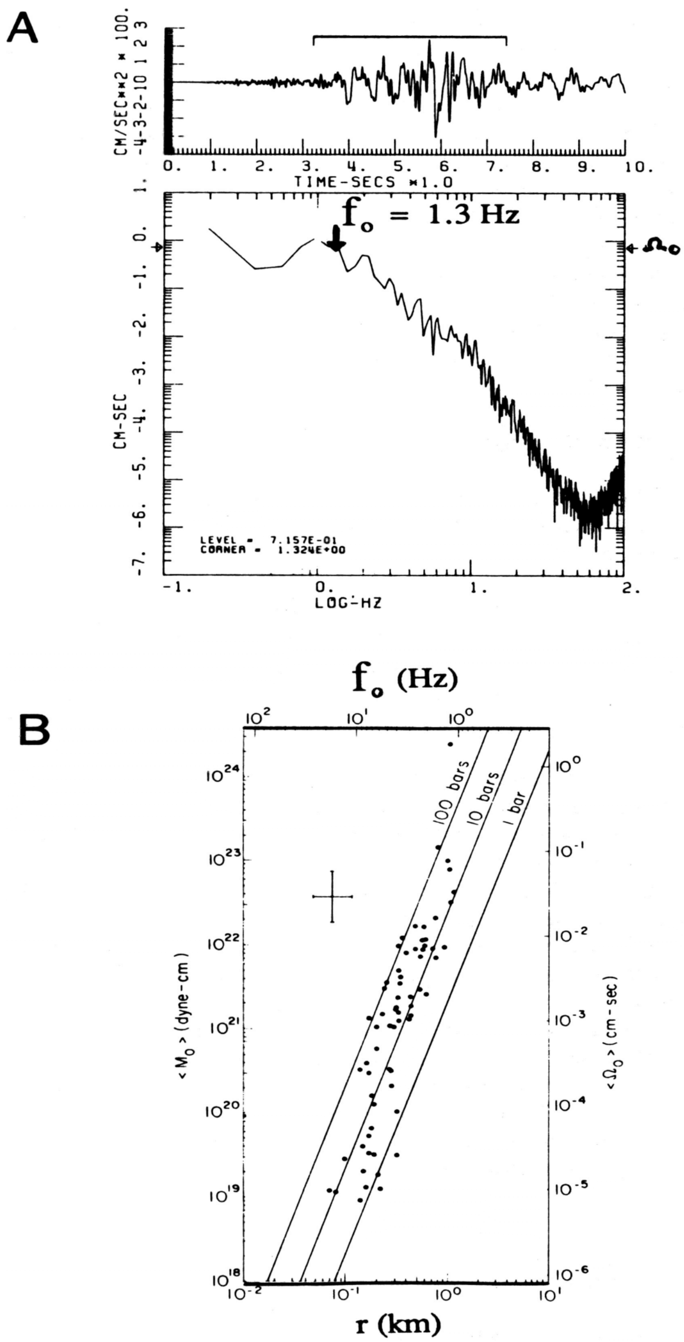


Figure 2. A, Seismogram and corresponding shear-wave frequency spectrum for a microearthquake. Source radius and stress drop are calculated from spectral values [equations (3) and (4)]. B, Seismic moment (M_0) versus source radius (r), low-frequency spectral level (Ω_0) and corner frequency (f_0) for microearthquakes in Mammoth Lakes, California. Stress drops of 1, 10, and 100 bars are indicated by sloping lines. Note error bar indicating uncertainty. From Archuleta and others (1982).

are presented by Dewey and others (this volume). For larger earthquakes it is now common to proceed in the analysis beyond first motions and find a model of fault slip that matches the observed regional and teleseismic waveforms and amplitudes (Helmberger, 1974; Langston and Helmberger, 1975; Nabelek, 1985). The determination of focal mechanisms from the analysis of long-period body waves has been discussed by Dziewonski and others (1981).

An important physical parameter associated with movement on a fault is the stress drop. Aki (1967) first suggested a striking similarity among all earthquakes, namely, a constant stress drop. Hanks (1977) concluded that, for 12 orders of magnitude in seismic moment, the stress drop is constant, with an uncertainty of ± 1 order of magnitude. The stress drop ($\Delta\ \sigma$, measured in bars) associated with a locally recorded earthquake is generally calculated from the frequency spectrum of the shear wave. According to the model of Brune (1970):

$$r=2.34V_s/(2f_o) \qquad (3)$$

$$\Delta\ \sigma=7M_o/16r^3) \qquad (4)$$

where r is the source radius in km, V_s is the shear-wave velocity in km/s, f_o is the corner frequency in the displacement spectrum (Fig. 2), and M_o is the seismic moment in dyne-cm.

Stress drop is most reliably determined from these

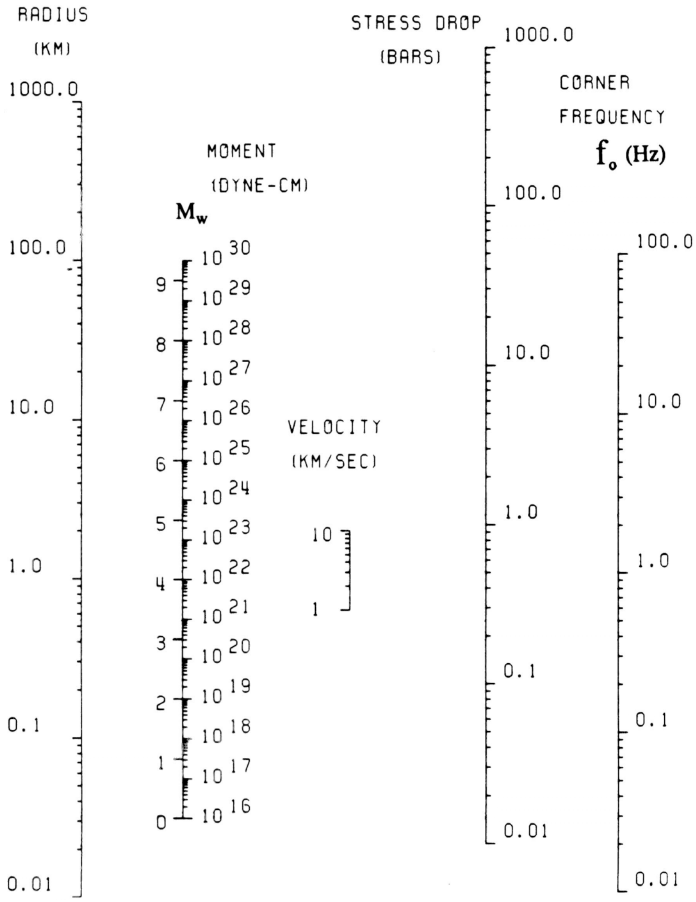


Figure 3. Composite nomograph of source parameters derived from equations (1), (3), and (4) in text. For the example in Figure 2A, the corner frequency (f_0) in the shear-wave spectrum is 1.3 Hz, and we may assume a crustal shear-wave velocity (V_s) of 3.0 km/sec. A line connecting these two values yields a source radius (r) of 0.8 km, a seismic moment (M_0) of 7.0 x 10^{21} dyne-cm, a moment magnitude (M_w) of 3.8, and a stress drop ($\Delta\ \sigma$) of 0.6 bars. From Mahdyiar (1987).

relationships using horizontal-component records that are used to calculate the S-wave spectrum (Fig. 2).

The seismic moment was defined above as the product of the fault area, average fault slip, and rock shear strength; the total seismic moment integrated over the source volume is the seismic moment tensor (Madariaga, 1983). The moment tensor measures the inelastic deformation at the source during the earthquake, and its value at the end of the rupture process measures the permanent inelastic strain produced by the event. For large earthquakes, the seismic moment and seismic moment tensor provide more accurate quantitative measurements of fault rupture than seismic magnitudes or stress drops (Aki and Richards, 1980). Mahdyiar (1987) has presented a nomograph that allows convenient visualization of the relationships between seismic source radius (r), moment magnitude (Mw), seismic moment (M_o), stress drop Δ σ), and the corner frequency (f_o) (Fig. 3).

Uncertainties in earthquake source parameter determinations

Errors in source parameters depend on the distribution and density of station coverage. For an earthquake within a regional network, epicentra1 determinations are more accurate than hypocentral determinations. As a general rule, epicentral locations are accurate to one-tenth the average station spacing, and hypocentral locations are two to three times less accurate. The most accurate focal depths are obtained when the nearest seismograph is within one focal depth of the epicenter. Earthquake locations outside the networks will have substantially larger errors. Likewise, the uncertainty in a focal mechanism depends on the azimuthal and distance coverage and on the data quality. For earthquakes with magnitude greater than 4.5 that have been recorded since the mid-

1960s, a reliable focal mechanism can be obtained by waveform (synthetic seismogram) modeling (e.g., Langston and Helmberger, 1975; Nabelek, 1985). In many cases, the source properties of an earthquake of magnitude greater than 6 that occurred more than 30 years ago can be determined from the analysis of the surviving seismograms (e.g., Dozer, 1986).

Uncertainties in the determination of seismic moments and stress drops are, like location uncertainties, dependent on station coverage. The amplitude measurements on which magnitude calculations are based are affected by two major factors. The first, seismic attenuation (Q) along the propagation path, acts to reduce the measured amplitude. In general, attenuation effects are much larger for the western United States than the eastern. The second factor, the local station site response, may, in extreme cases, play the dominant role in determining seismic amplitudes. Uncertainties in seismic moments are estimated to be 50 percent or less; stress drop uncertainties are larger: half an order of magnitude or more (Archuleta and others, 1982; Fig. 2).

DETERMINING EARTH STRUCTURE FROM NATURAL SOURCES

Data

The use of earthquake sources to determine the Earth's structure has certain advantages over manmade sources: worldwide distribution (particularly the distribution with depth), strong seismic energy, and particularly high-amplitude shear and surface waves. By comparison, manmade sources with high yields (very large chemical explosions or nuclear sources) are limited in number and distribution, and are relatively weak in shear-wave energy. The disadvantages of natural sources arise in part from their unpredictability in time and space, and the uncertainties in hypocentral location and origin time.

The data used to determine earth structure from natural sources come from the same networks used to determine earthquake source parameters: permanent regional networks, temporary aftershock networks, and intermediate-period and broad-base seismic observatories such as WWSSN stations or other global networks. In addition, data are obtained from special temporary deployments of intermediate-period seismographs (as much as 20 sec) at broad spacings (25 to 100 km) that record teleseismic body- and surface-wave arrivals. Typically, these temporary deployments last 2 to 4 months, in order to record a sufficient number of earthquakes.

Methods of analysis

More than a dozen methods exist for the determination of Earth's structure using earthquake sources. Some of these methods can also be applied to man made sources, particularly nuclear explosions.

One-dimensional methods: Body waves

The observed arrival times (or seismograms) from earthquake sources are commonly displayed in a traveltime plot (or record section), with distance from the source plotted against time. In such a display, the travel time curves of refracted and reflected arrivals may be identified in the data. These curves can then be inverted for velocity-depth structure subject to several assumptions. The most important assumption is that the average structure between the source (or sources) and the recording array can be reasonably approximated by a one-dimensional velocity-depth function (i.e., assuming no lateral velocity variations). The questionable validity of this assumption is probably the largest source of error. A second source of error is the common assumption that no low- velocity zones occur within the one-dimensional velocity-depth function.

Once a P- or S-wave traveltime curve has been measured, the Herglotz-Wiechert integral (Bullen and Bolt, 1985) can be used to determine the one-dimensional velocity-depth function that matches an observed traveltime curve. This integral relates the velocity-depth function to the slowness (reciprocal apparent velocity) versus distance curve:

$$Z(p_1)= {}^1/\pi \int_0^{x1} \cosh^{-1} (p_0/p_1)dx \qquad (5)$$

where $dT/dX = p$, p_0 is the apparent slowness at the surface, and p_1 is the apparent slowness at a distance X_1.

Once the traveltime curve has been inverted using this integral, the resultant velocity-depth model can be evaluated and modified by iterative forward modeling, wherein a series of small adjustments are made to the initial model to improve the overall traveltime fit. This is essentially the procedure used by Jeffreys and Bullen (1935, 1940) to develop the first models for the velocity structure of the Earth.

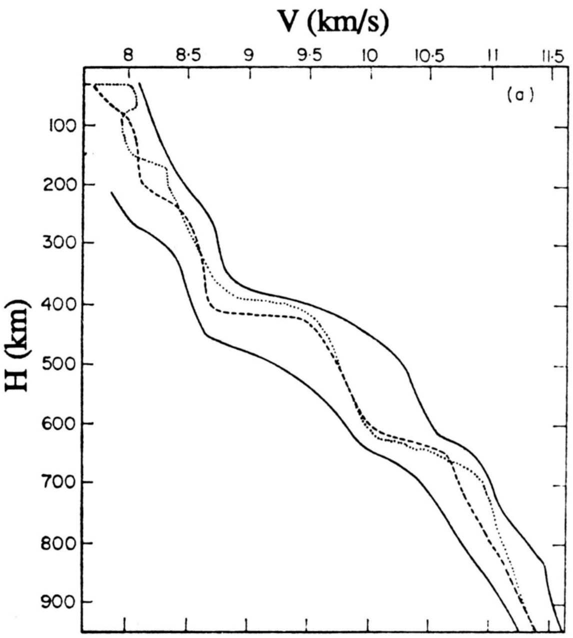


Figure 4. Application of "tau-*p*" inversion of earthquake traveltime for the determination of upper mantle structure. Shown are velocity-depth extremal bounds for 98 percent confidence level (solid lines), "best fitting" velocity model (dashed line), and comparison with previous model CIT 208 (dotted line). From Bessonova and others (1976).

All traveltime observations are in some way incomplete or contain scatter due to near-surface effects, lateral velocity variations, and incomplete sampling of the wavefield. For this reason, it is often desirable to know the permissible bounds on the velocity-depth structure that a given traveltime curve provides rather than simply deriving one particular velocity-depth model. This question has been approached from various points of view, the most common one being "tau-*p* inversion" (Bessonova and others, 1974, 1976). In this approach, the observed time-distance curve is transformed into a delay time (tau)-apparent slowness (*p*) curve that is a single-valued function (i.e., triplications in the travel-time curve do not occur). This simple transformation allows the inversion process to conveniently include the extremal bounds (i.e., limits) on the velocity-depth structure. Bessonova and others (1976) illustrated the application of the method to earthquake travel-time data (Fig. 4). Walck and Clayton (1984) applied a variation of the tau-*p* method wherein an entire earthquake record section is directly transformed by means of slant stacking (e.g., McMechan and others, 1982; Waters, 1987) to yield the tau-*p* curve needed for inversion.

An alternative approach to tau-*p* inversion is to search the parameter space for many models that fit the data. Because there is no direct means of finding all models that fit a given data set, an indirect means is used: a random search of the valid model space. In this so-called "hedgehog" method, many thousands of models (or even millions if sufficient computer time is available) are obtained by perturbing an originally satisfactory model and evaluating the new model for its fit to the observed data. Muller and Mueller (1979) and Knopoff (1972) effectively applied this method to seismic refraction and surface-wave data analysis, respectively.

Seismic amplitudes provide an important additional constraint on velocity structure because amplitudes are particularly sensitive to the details of velocity gradients and discontinuities. In fact, two velocity-depth models that are nearly indistinguishable in terms of their travel-times will generally have important differences in amplitude-distance behavior. These differences can be determined by the calculation of synthetic seismograms (the numerical calculation of the response of an idealized medium to seismic wave propagation). Ideally, these seismograms should be calculated with the same frequency band as the observed seismograms and should include converted and multiple phases. Synthetic seismogram calculations are far more complicated than travel-time calculations and therefore require considerable computer resources. Synthetic seismogram modeling usually enters the process after a best-fitting velocity-depth model has been obtained from travel-times.

Trial-and-error amplitude fitting is the most common method used to refine a velocity-depth model with synthetic seismograms. A series of adjustments are made to the initial model until a satisfactory amplitude and traveltime fit is obtained. A variety of methods exist for the calculation of synthetic seismograms for earthquake and explosive sources, including Cagniard-de Hoop (Helmberger and Wiggins, 1971), reflectivity (Fuchs and Muller, 1971; Kind, 1978), finite differences (Boore, 1972; Kelly, 1976), finite elements (Smith, 1975), the "WKBJ" method (Chapman, 1978), recursive methods (Kennett, 1974, 1983), and discrete wave-number methods (Bouchon, 1982; Luco and Apsel, 1983). Trial-and-error modeling using synthetic seismograms has been applied to the determination of upper mantle structure by Helmberger and Wiggins (1971), Grand and Helmberger (1984), and Burdick (1981). Aki and Richards (1980) and Spudich and Archuleta (1987) have given a comprehensive treatment of the theoretical basis for several of these methods.

Synthetic seismogram modeling can also be applied to determine the velocity structure beneath a single station that records nearly vertically incident teleseismic body waves. In this analysis, the complexity in the waveform is used to identify crustal and upper-mantle converted and multiply reflected phases (Langston, 1977; Owens and others, 1984). Based on the amplitudes and times of these phases, the depths and velocity contrasts of the major crustal and upper-mantle discontinuities can be calculated. Owens and others (1984) and Owens and Zandt (1985) have applied this method to crustal studies and obtained excellent agreement with neighboring seismic refraction data (Fig. 5).

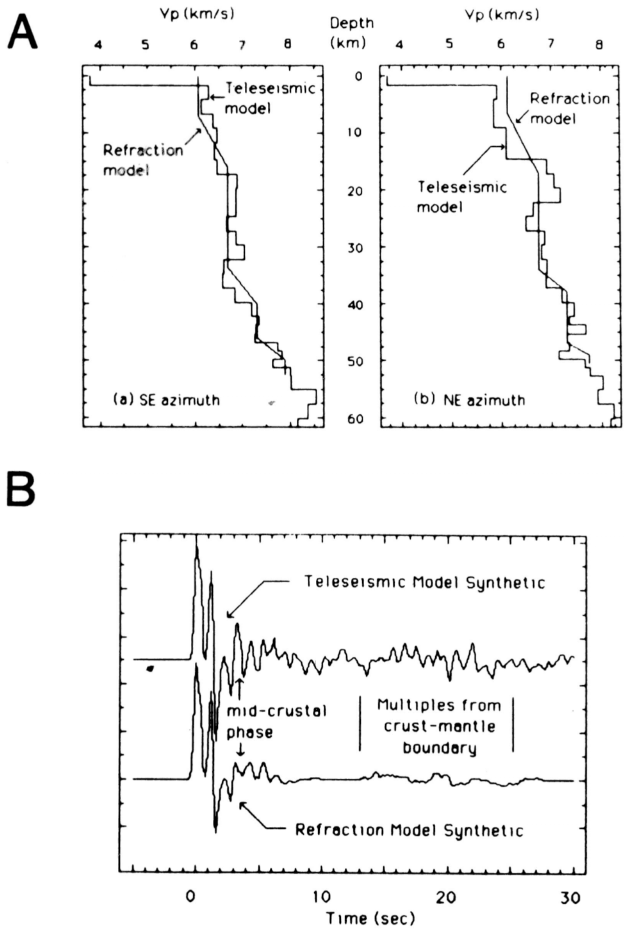


Figure 5. Teleseismic waveform modeling (also called receiver transfer function modeling) applied to data from the Cumberland Plateau, Tennessee (Zandt and Owens, 1986). A, Comparison of models derived from waveform modeling with nearby seismic refraction model. The two models agree well in average velocity structure. B, Example of two synthetic seismograms for teleseismic waveform model and refraction model showing more complex (and more realistic) synthetic for the teleseismic waveform model due to greater complexity in the velocity structure.

Arrival times of local earthquakes provide sufficient information for the calculation of an average one-dimensional model of the velocity structure beneath a seismic network. Although such models are not optimal for interpreting tectonic structure, a one-dimensional velocity model with appropriate station corrections provides the easiest and most efficient means of obtaining accurate earthquake locations. To obtain this model, traveltime residuals are computed for network stations using an assumed initial velocity model. A linear set of equations is generated relating the traveltime residuals to a series of small perturbations in origin times, hypocentral coordinates, and the velocity of the model. This set of equations is solved using damped least squares, and the initial model and hypocentral parameters are adjusted accordingly. The final model is obtained after several iterations, and includes individual station terms which to some degree correct for lateral velocity heterogeneities and site-dependent variations (Crosson, 1976; Eberhart-Phillips and Oppenheimer, 1984).

One-dimensional methods: Surface waves

The use of seismic surface waves (Rayleigh and Love waves) for investigations of crustal and upper-mantle structure began in the 1950s with the advent of high-quality seismic observatories and digital computers. Surface-wave analysis is an effective complement to seismic-refraction and earthquake body-wave studies for several reasons. Surface-wave propagation is dependent mainly on the shear-wave velocity structure (Bullen and Bolt, 1985), whereas many seismic-refraction and body-wave studies are restricted only to the compressional-wave structure. Together the shear- and compressional-wave velocities can be used to determine Poisson's ratio, and infer

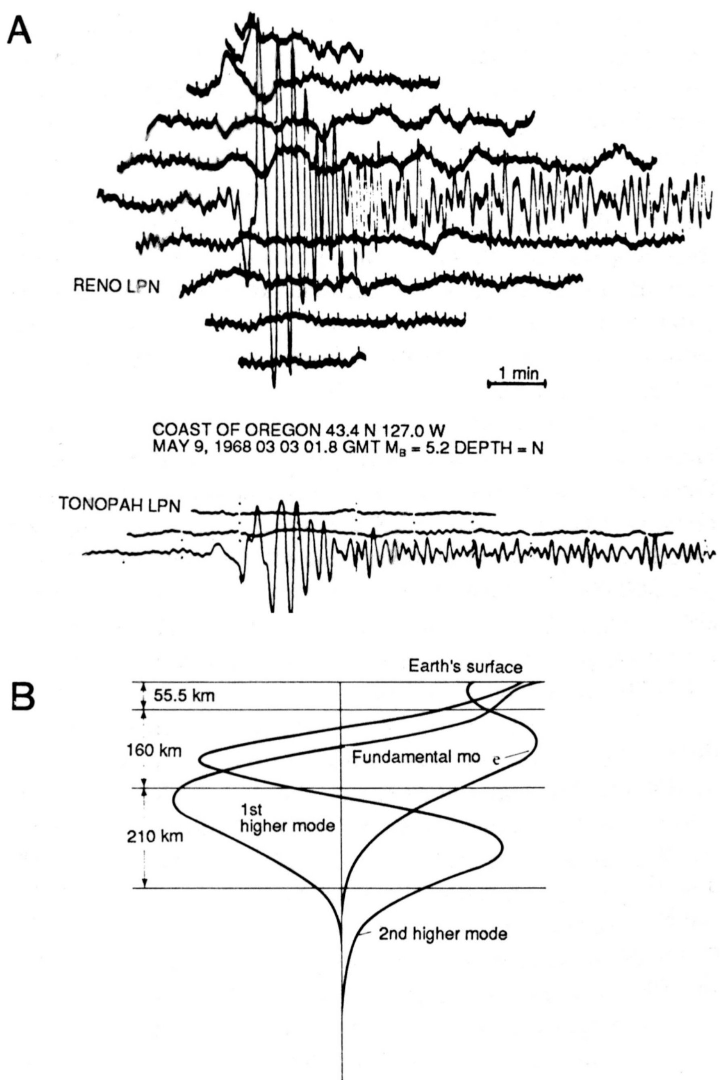


Figure 6. A, Typical surface-wave seismograms: Love waves recorded on long-period seismographs located at Reno and Tonopah, Nevada, for a magnitude 5.2 earthquake off the coast of Oregon. Well-dispersed wave-forms are evident. From Priestley and Brune (1978). B, The patterns of displacement with depth for three modes of Love waves for a period of 30 sec. It is apparent that each mode will be sensitive to somewhat different portions of the Earth's structure. From Bolt (1982).

composition and other physical proper- ties. Also, surface-wave investigations can easily provide information regarding deep structure: periods between 10 and 100 sec are routinely analyzed, corresponding to depths ranging from the upper crust to about 200 km (Bullen and Bolt, 1985).

The most useful surface-wave data for crustal and upper- mantle studies are recorded by seismographs with a long-period response (100 sec or longer; Fig. 6). Stable temperature, barometric pressure, and tilt are required for these seismographs, which are generally operated at isolated observatories, such as the WWSSN locations. Although the number of stations is limited, the worldwide distribution of earthquakes provides an abundant source of surface-wave observations. Recently, digital recording of long-period seismographs has made possible computer processing of surface-wave recordings, thereby permitting the interpretation of seismic attenuation and anisotropy (Anderson and Hart, 1976; Liu and others, 1976; Mitchell and others, 1976; Mitchell and Herrmann, 1979; Tanimoto and Anderson, 1985; Yomogida and Aki, 1985).

Long-period surface-wave propagation depends on the entire crustal and upper-mantle velocity structure, but as a general rule, the shear-wave velocity structure at a depth approximately 0.4 times the Rayleigh wave wavelength (0.25 times the wave-length, and the near-surface velocity, for Love waves) has the greatest influence on the phase velocity for a wave with a given period (Fig. 6B). Surface waves are observed as dispersed seismograms, with the longer period waves arriving ahead of the shorter period waves because they sample greater depths (which generally have higher velocities; Fig. 6A). If a broad range of wave-lengths is recorded, the lithospheric velocity structure can be estimated by working from the lowest phase velocity (shortest period and shallowest depth) to the highest phase velocity (longest period and greatest depth). Methods of measuring phase velocities of surface waves and inverting them for shear-wave structure have been discussed by Press (1956), Brune and others (1960), Brune (1969), and Kovach (1978).

Despite the differences in the equations of motion-governing surface waves and body waves, many of the analysis techniques previously discussed for body waves can be applied to surface-wave analysis after appropriate modifications. Trial-and-error fitting or least-squares inversion of the observed dispersion curves is used to find models that fit the data. This process can usually be constrained by converting an existing compressional-wave velocity structure to a shear-wave and density model and using it as a starting model. Generalized inverse theory (Backus and Gilbert, 1967; Wiggins, 1972; Aki and Richards, 1980; Menke, 1984; Tarantola, 1987) provides a method of evaluating uniqueness and estimating errors in the derived models. Monte Carlo random model-selection techniques (Press, 1968) and the "hedgehog" procedure of varying the physical parameters (Knopoff, 1972) provide a means of evaluating the range of models that will satisfy the observed data.

One of the most important results of surface-wave analysis is the clear distinction in shear-wave structure between different tectonic provinces (Brune, 1969; Knopoff, 1972). Surface waves have also been highly effective in defining the base of the lithosphere, generally taken as the top of an upper-mantle shear-wave low-velocity zone. Measurements of seismic attenuation from surface waves indicate that attenuation is significantly higher in the upper mantle than in the crust, and a maximum in attenuation appears to correlate with the base of the lithosphere (Anderson and Hart, 1976; Mitchell and others, 1976), supporting theories of a more plastic asthenosphere.

Assumptions and uncertainties in one-dimensional analysis

The assumption that a one-dimensional velocity-depth model is a reasonable representation of Earth structure is the largest source of uncertainty in the methods discussed above. However, two points are worth noting. First, the one-dimensional assumption substantially reduces the degrees of freedom in the solution, thereby allowing for a well-constrained estimate of the average velocity structure. Second, a general stratification of the crust and upper mantle on the horizontal scale of 100 km and greater is evidenced by the excellent fit of one-dimensional synthetic seismograms to seismic-refraction and surface-wave data, and the gross results of seismic reflection profiling.

The presence of low-velocity zones within the crust will cause errors in the interpretation of seismic data when some one-dimensional inverse methods are used. Neither Herglotz- Wiechert integration nor tau-*p* inversion methods, as generally applied, allow for low-velocity zones. If a low-velocity zone is present, layers below the zone will appear shallower than their true depths. Even when a low-velocity zone is correctly identified, the velocity within the zone is subject to a large uncertainty.

Amplitude modeling using synthetic seismograms significantly reduces uncertainties in the one-dimensional interpretation of seismic data unless the one-dimensional assumption is seriously in error. Seismic amplitudes are particularly sensitive to velocity gradients and velocity discontinuities; thus a one-dimensional model derived from synthetic seismogram analysis will be better constrained and include more detail than a model derived only from traveltime analysis.

Zandt and Owens (1986) presented a comparison of teleseismic waveform modeling for crustal structure with a seismic refraction model for the same area. Their comparison shows that the two methods were in excellent agreement in one study area, with a mean difference of about 0.2 km/sec at any depth (Fig. 5A). Given this agreement, further investigations of the methodology, assumptions, and uncertainties in teleseismic waveform modeling are warranted.

For surface-wave modeling, consideration of the range of models that fit a given surface-wave data set indicates that while particular details within a model may not be required by the data, the gross shear-wave velocity structure will be well estimated (Kovach, 1978). Since the compressional- and shear-wave structure is closely correlated, the uncertainties of the model will be significantly reduced if seismic refraction data are available to constrain the crustal thickness and compressional-wave velocity structure.

Two- and three-dimensional methods of determining Earth's structure

It is often of greater interest to determine two- or three- dimensional velocity variations in the Earth than to characterize the average one-dimensional structure. Examples include studies of continental margins, the roots of mountain belts, detecting subducted slabs, and studies of calderas and geothermal areas. In these studies, a two-dimensional model is derived from a linear seismic array, and a three-dimensional model is derived from an areal array.

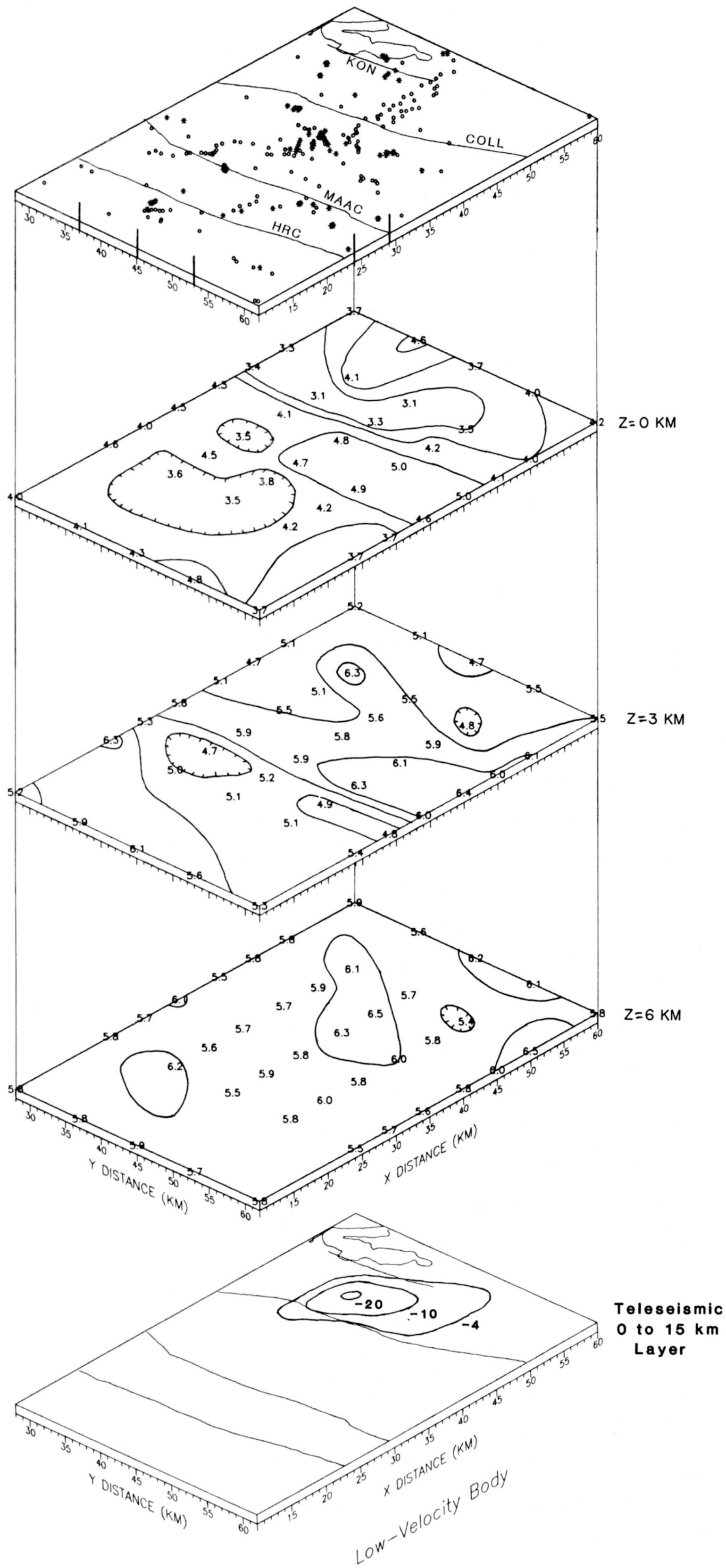
The time-term method (Willmore and Bancroft, 1960) is one of the oldest methods for determining two- and three-dimensional structure. The name is derived from the fact that the method separates the total travel time of a particular source- receiver pair into three parts:

$$T_{ij} = A_i + A_j + X_{ij}/_{ref} \tag{6}$$

where T_{ij} is the observed traveltime from source i to station j; A_i, and A_j are the source and station time-terms, respectively; $x_{i,j}$ is the distance between source i and station j, and V_{ref} is the refractor velocity.

The advantage of expressing the traveltimes in this way is that it assigns a common time-term to each station and source, and fixes the refractor velocity for each arrival branch. Having done so, well-developed methods for the least-squares solution of a set of simultaneous equations may be used, if there are more equations than unknowns. The solution requires that a sparse N by N matrix be inverted, where N is the number of stations or sources, whichever is larger. The method is well suited to a variety of situations, such as where a large number of earthquakes have been recorded by a seismic network (Oppenheimer and Eaton, 1984; Heam, 1984). Seismic anisotropy may also be included in the method simply by including an azimuthal dependence in the refractor velocity term (Bamford, 1977; Kohler and others, 1982; Walck and Minster, 1982).

The time-term method, as generally applied, assumes that the ray paths from source to receiver are along a series of nearly horizontal refractors with constant velocity. In areas of complex structure, delay-time methods are more effective than time-terms



because delay times more accurately account for actual propagation paths (O'Brien, 1968).

In alternative methods, the Earth is modeled by a set of discrete blocks, each with constant velocity, or by a non-uniform three-dimensional grid, with velocity values specified at the intersections (grid points). Either method allows for an arbitrarily complex three-dimensional structure as the size of the blocks or the grid-point spacing is decreased (assuming that sufficiently

Figure 7. Local earthquake traveltime inversion: perspective plot of planar cross-sections of the upper crustal velocity structure beneath northern California (Eberhart-Phillips, 1986). Epicentral locations are shown in top (geographic) section. Isovelocity contours at three depths (0, 3, and 6 km) are shown; velocities are in kilometers per second. Bottom plane shows the upper-crustal low- velocity body determined from teleseismic residuals by Oppenheimer and Herkenoff (1981); contours indicate percentage velocity decrease. Teleseismic velocity decrease of as much as 20 percent is not seen in any of the three depths in the local earthquake inversion, which in fact shows high seismic velocities in this same region at a depth of 6 km.

dense data are available to determine the velocity structure). In the case of the three-dimensional grid model, the velocity (and its spatial partial derivatives) at a particular point along a seismic ray path may be computed by linear interpolation between the surrounding eight grid-points. This permits rapid calculation of approximate travel-times to be used in an iterative simultaneous inversion for the three-dimensional velocity structure and hypocentral parameters using the travel-time residuals (Thurber, 1983). Parameter separation (Pavlis and Booker, 1980) may be used on the matrix of the hypocentral and velocity partial derivatives to separate the location problem from the velocity calculation, thereby reducing the size of the problem and increasing the amount of data that can be used. Eberhart-Phillips (1986) illustrates the effective use of this method in an inversion for crustal velocity structure in northern California using 200 earthquakes and 200 velocity grid-points (Fig. 7).

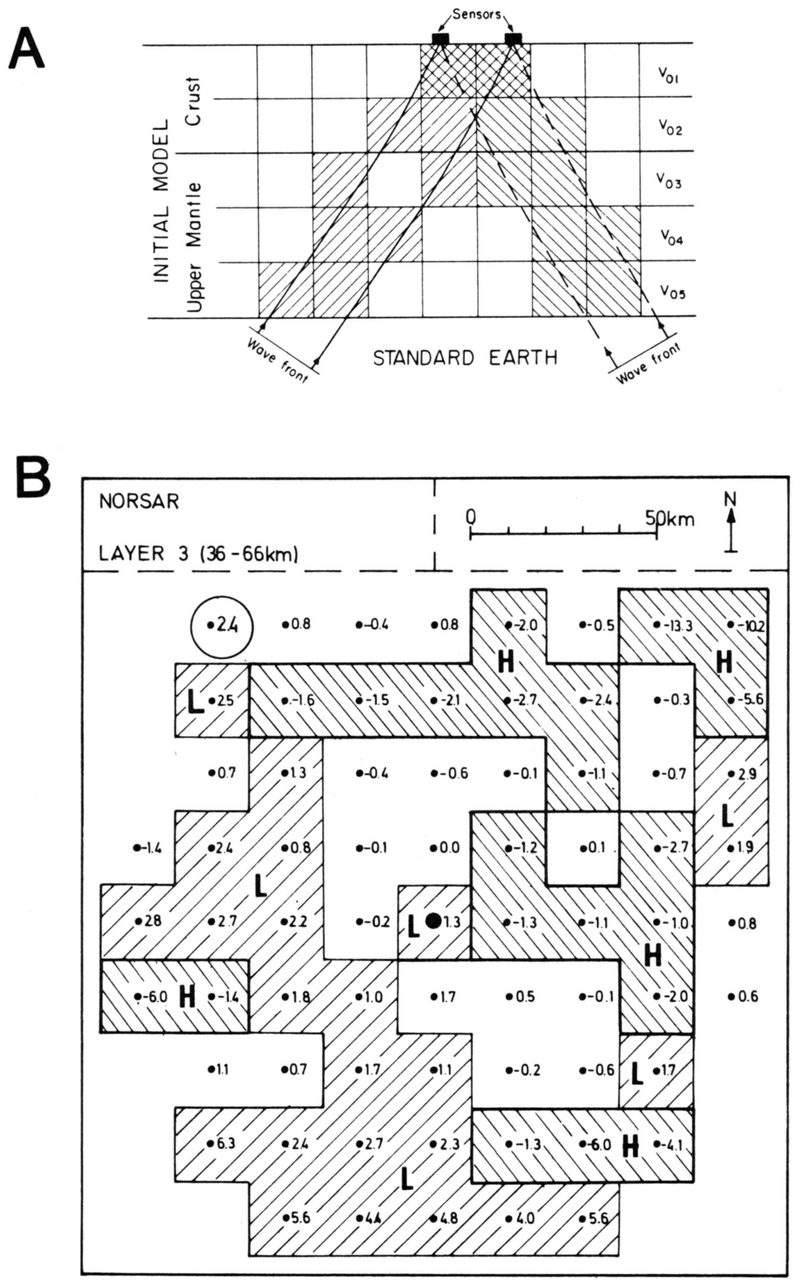


Figure 8. Teleseismic inversion method. A, Two wavefronts from opposite azimuths approaching surface sensors, traversing different blocks in the model, except for the two upper crustal blocks. B, Example from Norwegian "NORSAR" seismic array of a result from teleseismic inversion: high- and low-velocity regions (numbers in percentage of deviation from initial model) within blocks at a depth of 36 to 66 km. From Aki and others (1977).

In some applications of three-dimensional inversion, very large data sets are available, thereby requiring methods other than those using matrix inversion. For example, Hearn and Clayton (1986) considered about 45,000 earthquakes, each recorded by a subset of the southern California network (160 stations), yielding over 300,000 Pg arrival times. Such a large data set is not amenable to methods using matrix inversions. Instead, a form of tomography (Worthington, 1984; Humphreys and Clayton, 1988), a method widely used in medicine for imaging of the interior of the body, is used. Similar to time-term analysis, each travel-time is associated with terms corresponding to the source, receiver, and the connecting ray path, which is divided into N discrete calls. An initial reference model is derived from a least-squares fit to all data, and seismic ray paths from all sources to all stations are traced. The travel time residuals calculated for all source-station pairs are then distributed into the ray-path cells and the station and source terms, thereby yielding a revised model. The final model is obtained when the travel-time residuals converge after several iterations. An advantage of this method is that the calculation is done on a ray-by-ray basis, so there is no limit on the amount of data that can be used.

A powerful method for determining two- and three- dimensional lithospheric models is based on the inversion of teleseismic body waves. The inversion procedure is similar to those previously described, except that there is no need to determine simultaneously the hypocentral location and origin time; the modeling is based on relative arrival times throughout the array. The earth structure in the volume to be inverted generally consists of a series of layers that are divided into a number of blocks (Fig. 8). Within each block, deviation from the initially assigned velocity is determined from the travel-time data. The minimum block size that makes physical sense for this method is that equal to one wavelength (e.g., 6 to 8 km for 1 Hz) of the seismic energy recorded. However, in many cases, the block size is larger than a wavelength due to the limited amount of data available for the inversion.

Aki and others (1977) presented the mathematical formation of the teleseismic inversion problem, including the consideration of the effects of errors in the standard earth model, the initial block model, and the source parameters for the events used. The block inversion of teleseismic data has been widely applied in the past 10 years. Aki (1982) presented a summary of these results, and Iyer (1984a, b) discussed the application of the method to the study of calderas and magma chambers beneath volcanoes.

Assumptions and uncertainties in two- and three-dimensional methods

The time-term method is subject to several restrictive assumptions. The method will not work well if there is high relief on the refractor because the ray path from the refractor to a source or receiver will have an azimuthal dependence. Furthermore, the method does not account for lateral velocity variations or velocity gradients within the refractor. The result is that, if the refractor velocity is underestimated, the time-terms will be systematically underestimated as well (and vice versa).

Tomographic methods can be applied to laterally varying media, but often have blurred solutions because of the parameterization of the problem and because the solution is obtained iteratively rather than by a matrix inversion. The solution may be sharpened by using various processing techniques, although such processes may introduce artifacts.

Of the presently used three-dimensional methods, the matrix inversion method described by Thurber (1983) appears to result in the fewest artifacts, and a resolution matrix provides a measure of how well-constrained the velocity is at each grid-point. Teleseismic inversion methods (Aki and others, 1977) also provide resolution matrices that indicate those portions of the block model with an insufficient number of transmitted rays to reliably constrain the velocity anomaly. It is important to note that error analysis in inverse modeling is parameter-specific, and that if the model parameters are not correct, the actual "answer" will be much different from the inverse "solution," even though the error analysis will not reflect this.

DETERMINING EARTH STRUCTURE FROM CONTROLLED SOURCES

Seismic refraction profiling

Data. The work of Tuve (1951) is the earliest reference to a crustal-scale seismic refraction profile in the compilation by Braile and others (this volume). In the intervening 35 years (1951 to 1986), some 150 additional profiles have been reported, at the rate of four to five a year. Compared with earlier profiles, the most significant advances in data acquisiton have come in the number and spacing of shot points and recorders that constitute modern seismic refraction profiling. Many pre-1970 profiles were of a reconnaissance nature and used shot-point spacings of 200 km or larger and recorder spacings of 10 to 20 km. These experiments succeeded in outlining the average compressional-wave velocity structure of the crust and upper mantle on regional scales. In some cases, large explosions were recorded to ranges well in excess of 1,000 km, yielding measurements of the seismic velocity of the upper mantle that remain unsurpassed in terms of quality (Iyer and Hitchcock, this volume). The refraction experiments of the 1960s provide sufficient information to allow for the first contour plots of crustal thickness in the United States (Pakiser and Steinhart, 1964; Warren and Healy, 1973). James and Steinhart (1966) provided an excellent critical review of both crustal reflection and refraction seismology up to the mid-1960s.

Since the mid-1970s, the need for more geologically realistic crustal models has led to considerably more interest in the two-dimensional details of the velocity structure of the crust. This need has necessitated profiles with closer spaced seismographs and shot points. The improvement in the density with which data are collected is about an order of magnitude: 15 to 25 km shot-point intervals and 0.3 to 2.0 km seismograph spacings are now common (Fig: 9). Airgun sources fired in lakes and over continental margins provide continental refraction data with 100-m shot-point intervals. Recently, these airgun profiles have been recorded by land arrays using 100 to 300 m recorder intervals, yielding data of unprecedented density. Prodehl (1984), Meissner (1986), and Mooney and Brocher (1987) have provided modern summaries of results from seismic refraction profiling.

Methods of analysis. For most seismic refraction profiles, the most prominent arrivals are refractions within "layers" in the crust, and wide-angle reflections from velocity discontinuities. In general, middle- and lower-crustal wide-angle reflections have higher amplitudes than the refractions, and the interpretation process emphasizes matching the arrival times and amplitudes of these reflections. In recognition of the importance of these wide-angle reflections, some authors prefer the term

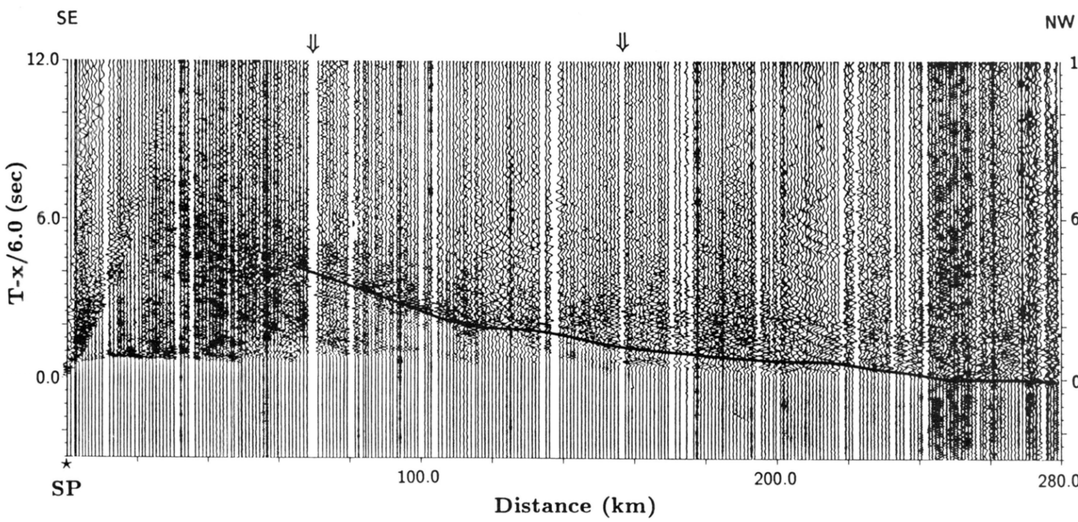


Figure 9. Modern seismic refraction/wide-angle reflection data recorded in central Alaska by the U.S. Geological Survey with an instrument spacing of I km. Crustal and upper-mantle phases are clearly recorded. Data are plotted with a reduction velocity of 6.0 km/sec. Moho reflection is indicated by solid line. Data of this quality and density are typical of those recorded over the past 10 years. More recently, data have been collected with recorded spacings of as little as 50 m, yielding data that can be processed similar to seismic reflection data (see text).

"refraction/wide-angle reflection" profiling to "refraction profiling"; here I use the briefer term for economy.

There have been steady improvements in methods for analyzing seismic refraction data since the first profiles were collected. Early investigators were often well aware of the potential for detailed travel-time and amplitude interpretation of two- and three-dimensional structure, but lacked both the requisite data density and the theoretical and computational capabilities necessary to make such interpretations. Prior to about 1970, most seismic refraction interpretations were based on analysis of refraction and wide-angle reflection arrival times, with the amplitudes of the most important phases generally interpreted in rather general terms (e.g., Healy, 1963; Hill and Pakiser, 1966). A variety of methods were applied in individual studies, in some cases with statistical analysis of errors in velocities and depths (e.g., Steinhart and Meyer, 1961). Two-dimensional models of crustal structure were derived from the interpretation of reversed profiles in terms of dipping layers, or by joining one-dimensional solutions for individual shot points.

The time-term method was often applied to determine depth variations on a given refractor (e.g., James and Steinhart, 1966). Smith and others (1966) gave a detailed mathematical treatment of the method as it was developed to that time. The Herglotz-Wiechert integral (equation 5) was also employed to derive one-dimensional models.

An alternative interpretational method was widely applied to North American refraction data by Prodehl (1970, 1979) and Prodehl and Pakiser (1980). This method is essentially a modification of the Herglotz- Wiechert integral to allow for low-velocity zones; it is well suited to the computation of one-dimensional velocity structures with either positive or negative velocity gradients and with transitional velocity discontinuities rather than sharp interfaces. The method and its application to real data are described by Prodehl (1979). Typically, the results from several shot points are presented in the form of isovelocity cross sections (c.f., Prodehl and Lipman, this volume), as opposed to layered cross sections.

Since the mid-1970s, major improvements in the analysis of crustal seismic refraction data have been made possible by two new capabilities: the processing of digital seismic traces and numerical modeling algorithms for two- and three-dimensional structures. The most important advances are:

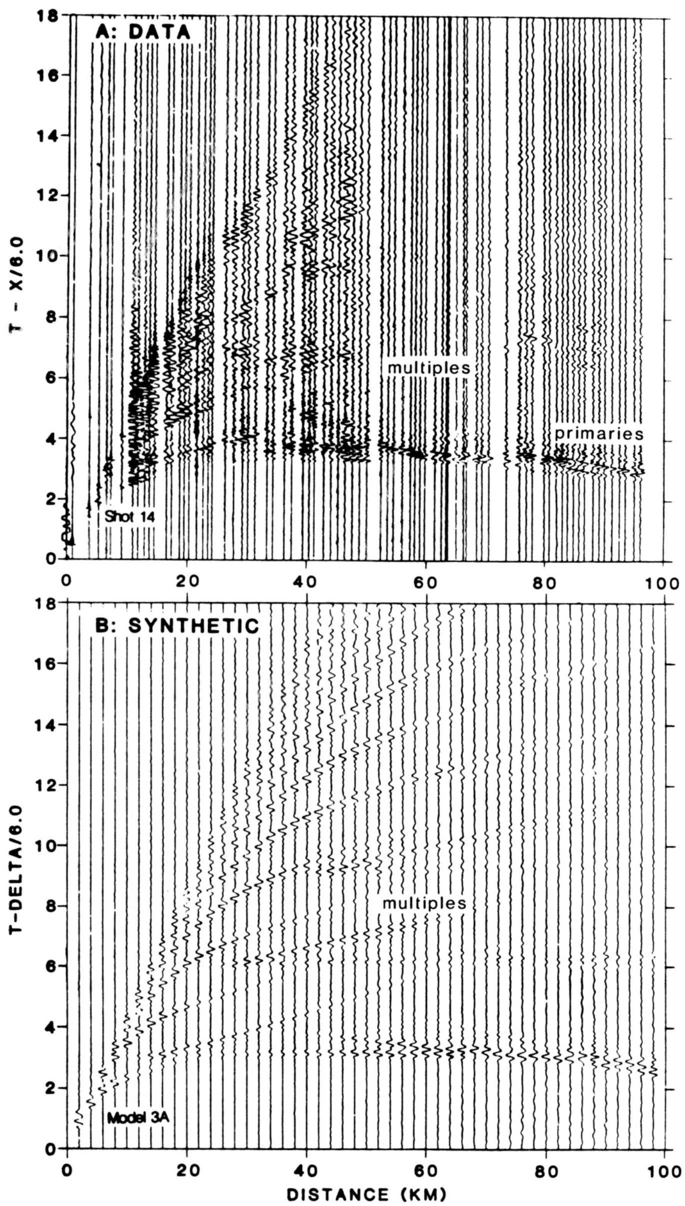


Figure 10. Synthetic seismogram modeling of seismic refraction data. A, Observed data from the Great Valley of California with prominent free-surface multiples. B, Synthetic seismograms calculated with the reflectivity method, including primary and multiple phases, and seismic attenuation. From Hwang and Mooney (1986).

1. Digital data processing allows for the routine application of methods that formerly were more cumbersome to apply to analogue data. These processes include frequency and velocity filtering, slant stacking of traces, and various modifications to the display of the traces (e.g., automatic gain control and variable-area shading).

2. Lateral variations in the velocity structure are accounted for with traveltime and amplitude calculation methods, as are the presence of seismic low-velocity zones within the crust and upper mantle (Mooney and Prodehl, 1984).

3. Ray theoretical and full-wave (complete) synthetic seismograms are calculated for the hypothesized crustal velocity model and are quantitatively compared with the observed data (Braile and Smith, 1974; Cerveny and others, 1977; McMechan and Mooney, 1980; Fig. 10).

4. The effects of apparent Q ("quality factor": inverse seismic attenuation) due to intrinsic attenuation and scattering may be accounted for.

5. Some quantitative estimates of model accuracy can be made when various inversion procedures are applied to the data. In addition to these advances, seismic interpretation methods, which were previously applied

only to seismic reflection data, are now applied to densely recorded refraction data. For example, McMechan and others (1982), Milkereit and others (1985) and McMechan and Fuis (1987) have applied automatic inversion methods based on slant stacking to refraction data, and Klemperer and Luetgert (1987) have applied normal-movement (NMO) velocity analysis and common midpoint (CMP) stacking methods. As denser seismic refraction/wide-angle reflection data are collected, the somewhat arbitrary line between reflection and refraction seismology will disappear.

Assumptions and uncertainties. There are important and restrictive assumptions made in many (but not all) seismic refraction interpretations published in the past 30 years: (1) that a planar, layered structure is a valid approximation; (2) that lateral velocity variations are smaller than vertical variations and occur on the same distance scale as the shot-point interval; and (3) that the principal crustal phases are correctly identified and have not been confused with multiples, phase conversions, or noise. This identification process is referred to as phase correlation.

The first assumption will not be valid in many situations, such as in highly folded and faulted areas or portions of the crust that have been heavily intruded. In such cases, known geologic constraints should be incorporated into the model, or the results should be qualified by a statement that they represent only the average properties of the crustal structure. The second assumption is derived from the observation that there is a general horizontal stratification of seismic velocity due to increases in pressure, temperature, density, and metamorphic grade with depth. The third assumption, concerning seismic phase correlation, is continuously reevaluated during the interpretation process on the basis of reciprocal traveltimes, amplitude and frequency behavior of the phases, and subjective input based on experience.

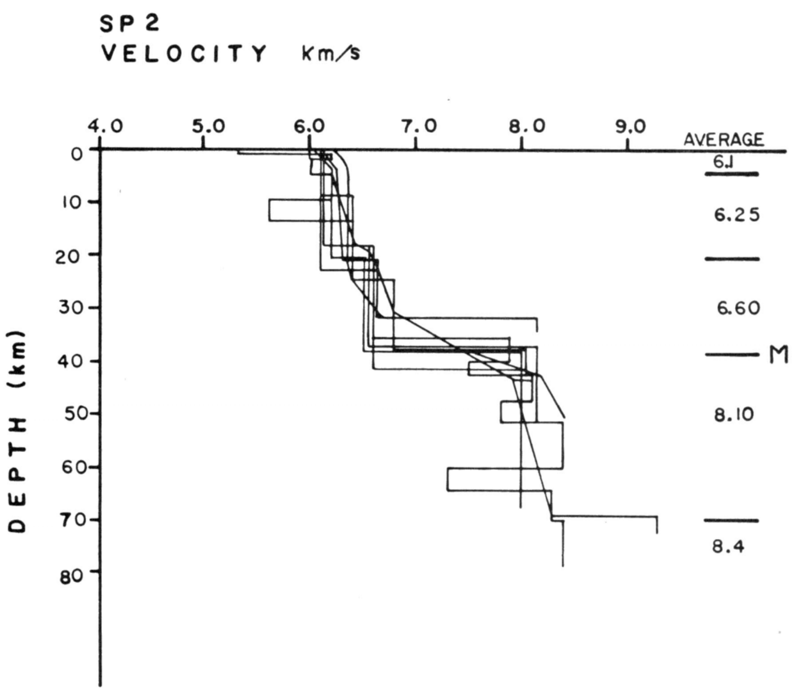


Figure 11. Comparison of seven separate traveltime interpretations of a seismic refraction profile recorded across the Saudi Arabian Shield in 1978. The average model is indicated in the column on the right-hand side of the figure. This comparison illustrates the uncertainty in seismic refraction data interpretation, particularly that arising from somewhat different phase correlations by different interpreters. Other comparisons (Walter and Mooney, 1987) show closer agreement when each interpretation includes amplitude modeling using synthetic seismograms. From Mooney and Prodehl (1984).

Detailed comparisons of interpretations of identical data sets have established that a final velocity model is more a function of the particular phase correlation used than of the interpretation method (Ansorge and others, 1982; Mooney and Prodehl, 1984; Finlayson and Ansorge, 1984; Walter and Mooney, 1987; Fig. 11).

Many of the existing refraction profiles in North America were collected in the 1960s, and Pakiser and Steinhart (1964) and James and Steinhart (1966) have summarized the uncertainties in the interpretation of refraction data as of that time. The determination of the seismic velocity of the upper crust has the lowest uncertainty, with an estimated error of 3 percent. This error estimate implies that a layer with a seismic velocity of 6.0 km/sec could actually have a velocity of 5.8 to 6.2 km/sec. Determinations of deeper crustal velocities and Moho depths have 10 percent estimated errors, but the upper-mantle velocity is uncertain to only 3 to 5 percent for the more detailed, reversed profiles due to the isolation of the Pn travel time curve over large offsets. When considering uncertainties in crustal thickness and Pn velocity, the ratio of the horizontal-to-vertical scale should be kept in mind. For example, the measurement of a 40-km-thick crust with a 400 km long refraction profile has a horizontal-to-vertical scale ratio of 10, which yields a very stable estimate in the vertical direction.

Since the 1970s, improved data quality and analysis methods have allowed more details of crustal velocity structure to be resolved, and uncertainties in layer depths, thicknesses, and velocities have been reduced approximately in half, particularly for the upper half of the crust. The assumption of planar layers is no longer necessary when two-dimensional ray-trace modeling is applied. Iterative forward modeling has been widely used to interpret this improved data. While forward modeling allows the inclusion in the velocity model of geologic and other constraints from geophysical data, it does not provide a quantitative estimate of errors (as does inverse modeling). Therefore, it is incumbent on the modeler to perturb the best-fitting model to assess the sensitivity of key model parameters to the fit of the data. In this way, some quantitative estimates of model resolution can be made. The statement by James and Steinhart (1966, p. 320) that "tradition has been strong in governing the types of velocity- depth functions fitted to the crust" remains true, and more formal means of evaluating errors and uncertainties in forward modeling remain to be developed.

Since seismic amplitudes are very sensitive to velocity gradients and discontinuities, these features can be refined through detailed synthetic seismogram modeling. The result is that velocity gradients (km/sec/km) are often cited along with average layer velocities. These velocity gradients are probably accurate ±50 percent or better when they are determined from high-quality data. The nature of seismic discontinuities has received much recent attention, and it is not uncommon that velocity transition zones and laminated zones are modeled in the crust instead of first-order discontinuities (Meissner, 1973; Deichmann and Ansorge, 1983; Braile and Chiang, 1986; Sandmeier and Wenzel, 1986). Estimates of transition-zone thickness are dependent on the frequency and bandwidth of the seismic energy used to probe the discontinuity, but as a general rule, an error of as much as ± 50 percent must be admitted.

Seismic reflection profiling

Data. Seismic reflection data provide the highest resolution information on the structural geometry of the crust. The application of reflection profiling to deep-crustal problems has led to great advances in the understanding of crustal structure, composition, and evolution. However, the resolution and data quality of reflection profiles is quite sensitive to the selection of data acquisition parameters. The selection process is often a difficult one; it involves a cost/benefit evaluation wherein factors such as the number of recording channels and shot points are balanced with the final cost per kilometer of profile. The earliest COCORP profiles were collected with a 48-channel recording system and 24-fold common depth point (CDP) coverage, 100-m geophone group intervals, and vibration points at every group. Four or five mechanical vibrator trucks were used as the seismic source array (Schilt and others, 1979; Oliver and others, 1983; Brown and others, 1986). Recent COCORP profiles were obtained with a 96- to 120-channel recording system, and the USGS has used an 800- to 1,000-channel sign-bit recording system, with 25-m group interval and 120-fold coverage (Stewart and others, 1986; Zoback and Wentworth, 1986). Field parameters vary for data collected by the University of Wyoming, Virginia Poly technical Institute, the California Consortium for Crustal Studies (CALCRUST), and other groups; they have been discussed by Barazangi and Brown (1986a, b) and Matthews and Smith (1987).

For studies of the deep crust, it is often possible to reprocess exploration industry VIBROSEIS data to recover longer two-way times by recorrelating with the original sweep and allowing the sweep to extend off the end of the trace, thereby obtaining approximate correlations to greater times (Okaya, 1986; Trehu and Wheeler, 1987). A limitation to the application of this method is that the original recording parameters of industry data are normally optimized for shallow structure.

Methods of analysis. The acquisition and processing of seismic reflection data has been well advanced by the hydrocarbon exploration industry. These techniques are described in numerous textbooks that range from introductory to advanced (Dobrin, 1976; Telford and others, 1976; Claerbout, 1976, 1985; Coffeen, 1978; Sheriff and Geldart, 1983; Jenyon and Fitch, 1985; Waters, 1987; Yilmaz, 1987). However, it should be kept in mind that the application of reflection seismology to the deep crystalline crust presents special problems not encountered in the exploration of the sedimentary rocks in the upper crust. Here, I emphasize those aspects of seismic processing that are of particular importance to deep crustal reflection seismology. Additional perspectives can be found in Taner and others (1970), Mair and Lyons (1976), Smithson and others (1980, 1986), Johnson and Smithson (1986), Mayer and Brown (1986), Zhu and Brown (1986), and Matthews and Smith (1987). Bally (1983) presented numerous examples of deep crustal seismic reflection data with an emphasis on the reflection signature of various tectonic regimes.

The processing of land seismic reflection profiling begins with a consideration of near-surface effects. In general, there are variations in both elevation and near-surface velocities along a seismic profile. Static corrections (time shifts) are derived for these variations. Usually a reference datum is selected, and both shot points and geophones are corrected to that datum. The proper derivation and application of static corrections is one of the most difficult problems in reflection seismology because of the severity of lateral velocity variations in the near-surface (Sheriff and Geldart, 1983).

Virtually all continental seismic reflection data are recorded using the common depth point (CDP) method, which results in substantial signal enhancement except in the most geologically complex situations. The summing of the field traces with a common depth point requires the application of a normal-moveout correction (NMO) for the hyperbolic curvature of reflection traveltimes. Some knowledge of the crustal velocity structure is needed for the proper application of the NMO correction, and an optimal velocity-depth function is commonly chosen from a series of stacking velocity, NMO, and semblance plots (Jenyon and Fitch, 1985). Standard seismic refraction analysis may also be applied to first arrivals, and the resulting shallow-velocity determinations may be used to supplement the analysis of the velocity panels. For the deeper crust, superior velocity estimates are obtained with longer array lengths (on the order of 10 to 20 km) because larger moveouts occur for larger offsets (Klemperer and Oliver, 1983). In theory, the root-mean-square velocity can be determined to depths of two or three times the maximum source-receiver offset (e.g., Bartelson and others, 1982; Hajnal, 1986). However, lateral velocity inhomogeneities and low signal-to-noise ratios degrade these velocity estimates, and in practice, few deep velocity estimates based on reflection data have been published. Coincident seismic refraction/wide-angle reflection measurements provide the most reliable deep seismic velocity estimates (Mooney and Brocher, 1987; Smithson and Johnson, this volume).

Many unwanted seismic signals that detract from the coherency of deep reflections are recorded during a seismic reflection survey. The most obvious unwanted signals are cultural noise; in many cases, noise can be removed by filtering or editing of the individual traces prior to stacking. Ground roll, shear waves, and other undesirable seismic energy can be largely eliminated from the unstacked data by a judicious combination of editing, muting, and velocity-filtering procedures (Sheriff and Geldart, 1983). Deconvolution is a process in which distortions of the wavefield due to multiples and filtering effects of the Earth are removed from the data (Sheriff and Geldart, 1983; Waters, 1987).

The CDP stacked section is a standard mode of presentation of seismic reflection data. Although further processing is often both desirable and possible, the CDP stacked section has the advantage of representing a standard level of processing, and is therefore the section with the fewest processing artifacts. The CDP section is often referred to as a "structural section" in recognition of the approximate image of the geologic structure it exhibits.

Lateral discontinuities and isolated velocity anomalies will produce diffractions in a CDP stacked section and may obscure or mimic deep reflections. True deep reflections may be well recorded, but may not appear in their correct subsurface positions because of the effects of dips and lateral velocity variations. Migration is a defocusing process that attempts to collapse diffraction hyperbolas and position reflectors at their actual locations in depth. Migration after CDP stacking is the most commonly applied method, and it is generally effective in structures with dips of less than about 30° and modest lateral velocity variations (Claerbout, 1976; Fig. 12). Other data artifacts can be reduced by (pre-stack) two-dimensional filters (fan, moveout, or pie-slice filters) that eliminate arrivals with particular apparent velocities as measured by the geophone array (Sheriff and Geldart, 1983).

The presentation of seismic reflection data at page size presents a special challenge since such data are normally interpreted at large scale, and processed sections reproduce poorly when reduced. Line drawings are commonly made of the data to display the principal reflections visible in the original section. However, because the construction

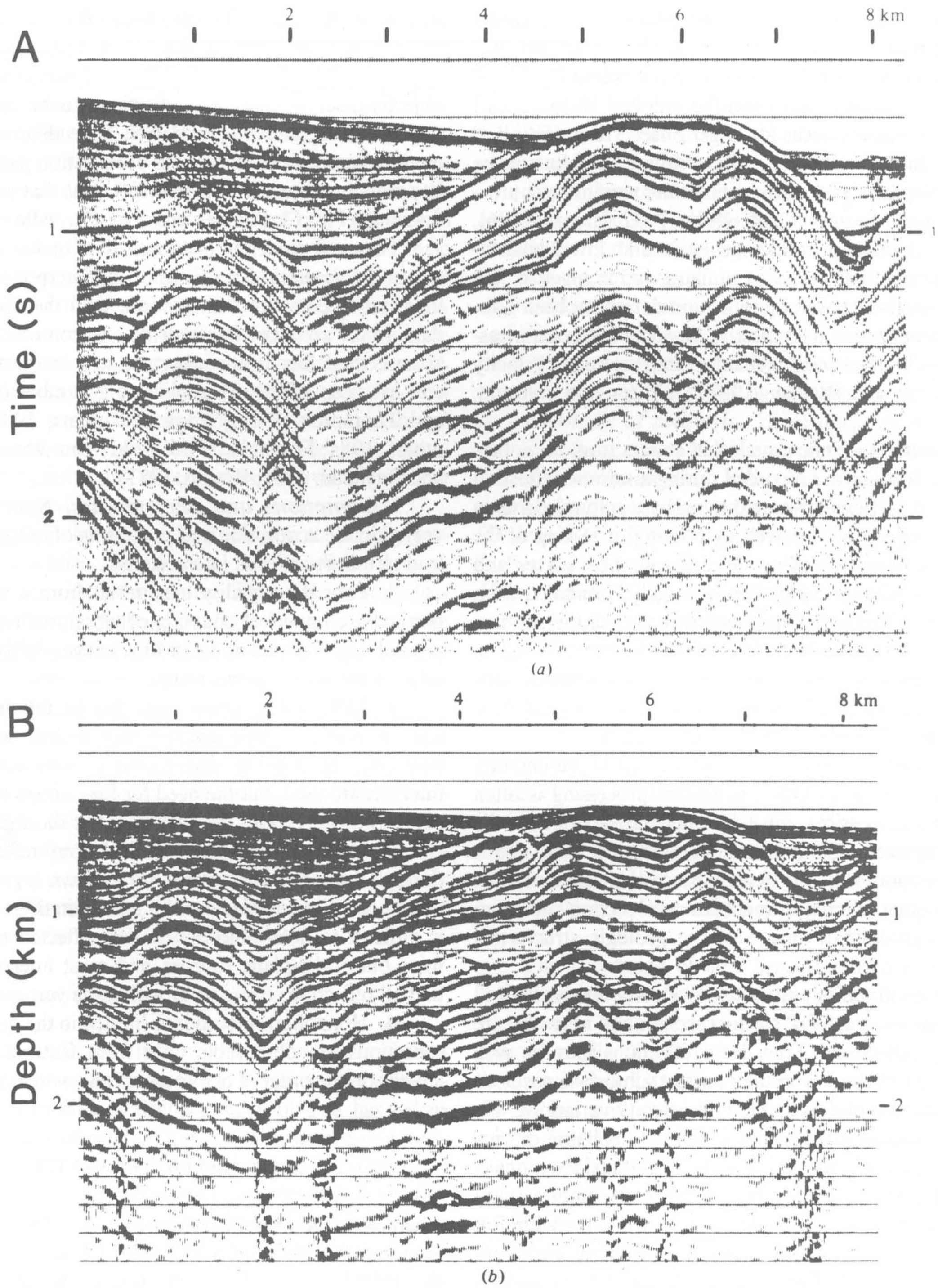


Figure 12. Example of depth migration of shallow seismic reflection data. A, Unmigrated time section. B, Finite-difference time migration with depth conversion. Note collapsed diffractions. From Hatton and others (1981) in Sheriff and Geldart (1983).

of a line drawing is highly subjective, recently there have been efforts to produce page-size plots of seismic reflection data that share the visual advantages of line drawings, but which use an automated, and therefore more objective, selection of the reflections to be highlighted in the display. Kong and others (1985) have described a coherency method based on slant stacking of reflection panels that succeeds in producing page-size reflection sections that are as clear as line drawings (Fig. 13; Phinney and Roy-Chowdhury, this volume).

Forward modeling of seismic data makes is possible to calculate the theoretical response of the interpreted geologic structure and to compare this response with the observed reflection data. In the absence of deep drilling to constrain crustal interpretations, modeling is an important step in crustal interpretation and has been effectively applied to a number of data sets (Hale and Thompson, 1982; Wong and others, 1982; Fountain and others, 1984; Fountain, 1986; Peddy and others, 1986; Blundell and Raynaud, 1986; Reston, 1987).

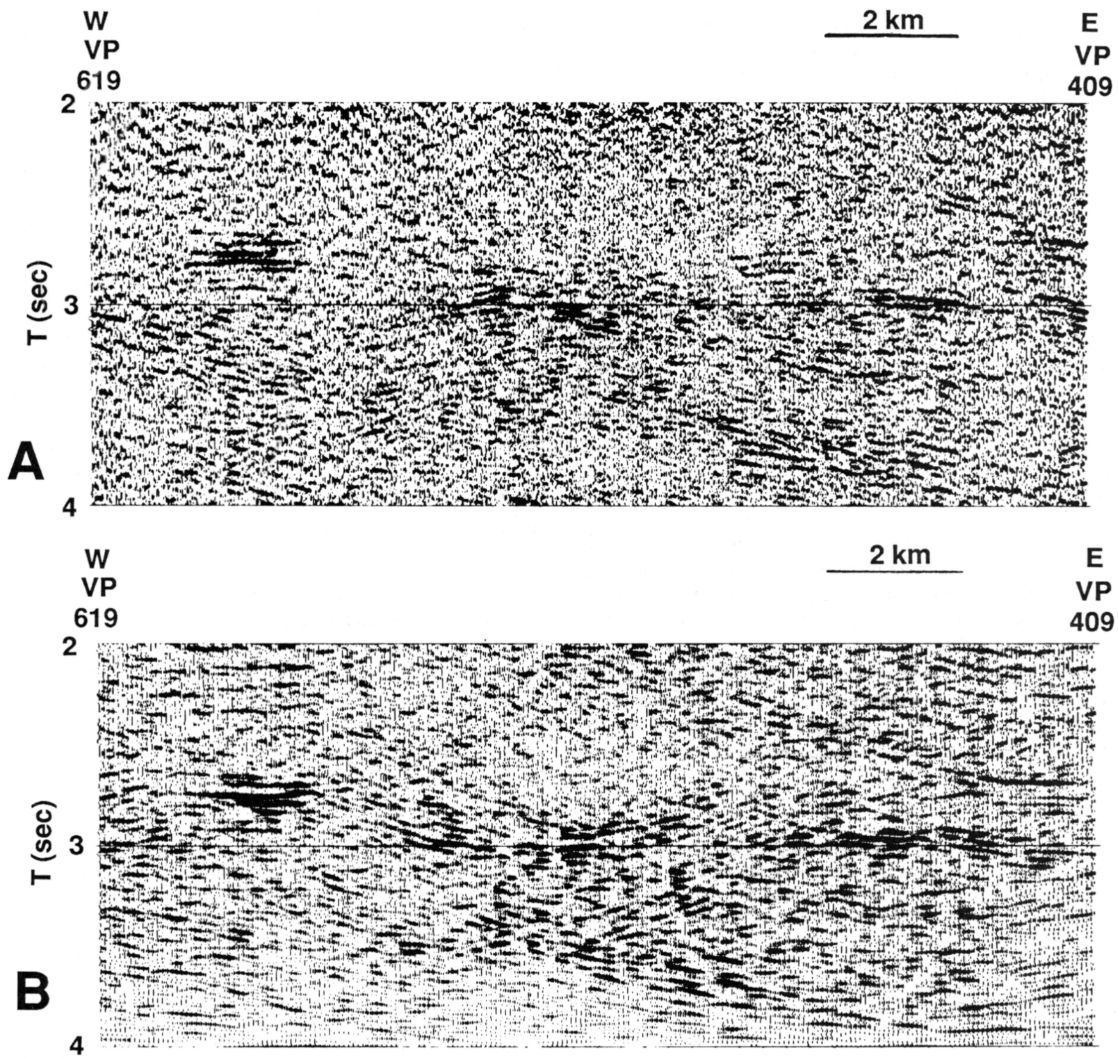


Figure 13. Seismic reflection data with coherency filtering applied to enhance reflections. Reproduction at small scale is significantly improved. A, Portion of the COCORP southern Appalachian line 4A stacked section from 2 to 4 sec, sampled at 4 msec, with 600 traces spaced at 33.5 m. B, Constant velocity (5.5 km/sec) time migration of the stacked section (A). From Kong and others (1985).

Assumptions and uncertainties. Numerous factors involved in the acquisition and processing of seismic reflection data lead to uncertainties in interpretation.

1. Poor data quality may result from a variety of causes, particularly imperfect coupling of vibrators or geophones to the ground, high cultural noise, or low, uneven-fold data arising from skips or crooked-line recording.

2. Large static corrections due to inhomogeneous near-surface conditions may result in poor stacked data. Static corrections can be reliably determined if sufficiently small group intervals are used, but the need for long arrays (for deep velocity control) conflicts with the advantages of short group intervals.

3. Three-dimensional control on reflector geometry, through the use of off-line recording arrays, is generally not avail- able because of logistical and budget limitations. Consequently, it is difficult to determine how much reflected energy is coming from out-of-the-plane sources, and most interpretations assume that all reflections come from within the vertical plane.

4. Horizontal resolution is limited in the deep crust. Seismic reflections are observable only from features with horizontal lengths on the order of one Fresnel zone, which may be expressed as:

$$R = (0.5\ LH) = 0.5\ V\ (2\ T/F) \tag{7}$$

where R is the Fresnel zone radius, L is the wavelength at the dominant frequency, H is the depth, V is the average velocity, T is the arrival time, and F is the frequency (Sheriff and Geldart, 1982). This relationship yields a Fresnel zone radius of 3 km for a 20-Hz reflection from 30 km depth in a crust with an average velocity of 6 km/sec.

5. Migration of deep seismic data often does not result in significant improvements in the quality of the image of the reflectors. This is due to generally low signal levels, the incomplete recording of the wavefield of deep reflections with conventional recording arrays, and distortions to the wavefield by near-surface features (Warner, 1987). If dips greater than 30° are present, pre-stack migration (Claerbout, 1985) is desirable; however, this process requires substantial computer resources.

A vertical seismic profile (VSP) (Gal'perin, 1974; Balch and Lee, 1984) is recorded in a borehole using surface sources. Since the stratigraphy of the hole may be inferred from borehole geophysical logs, the VSP provides a direct measure of the reflection response of the layers, and may be used to calibrate other seismic surveys. To date, VSPs have had only limited application to crustal reflection profiling (Smithson and Johnson, this volume).

In addition to uncertainties arising from data acquisition and processing, there are other more general issues of seismic wave propagation in the Earth that raise questions regarding seismic reflection interpretation. Blundell and Raynaud (1986) and Reston (1987) have shown that layered reflections, as commonly observed in the lower crust, may be produced by out-of-plane reflections in a model that includes only a single corrugated layer, or by spatial interference effects. Thus, the origin of deep crustal reflections remains a critical question in seismology. Future 3-D surveys will provide important new insights.

CLOSING REMARKS

This chapter has emphasized the data, analysis techniques, assumptions, and uncertainties of seismic methods as they are used to obtain the results summarized in the other chapters of the volume. Rather than emphasizing future directions in seismology, I have focused on current and past practice. From this perspective, it is appropriate to comment briefly on future needs in seismologic research.

The common reliance on forward modeling for the interpretation of seismic data presents a problem for those who seek to quantify resolution and uncertainties in geophysical models. While inverse modeling of traveltimes includes formal resolution estimates, forward modeling shows that waveforms and amplitudes are critical in confirming and refining model details. Waveforms and amplitudes have yet to be fully considered in published inverse solutions for two- or three-dimensional structure. Similarly, the determination of seismic source parameters within complex Earth structures is a promising field. Three-dimensional waveform modeling methods, which are feasible only on the most powerful computers (except for very long-wavelength data), are now being developed as seismologists strive to develop realistic Earth models.

As important as these issues of data modeling are, the highest priority in seismology is high-quality data acquisition; seismology, like all of the earth sciences, is a data-driven science. This is evident from the timing of the periods of the most rapid growth and progress, which have coincided with new data initiatives, such as the installation of the WWSSN network, the Early Rise / Lake Superior seismic refraction investigation (both early 1960s), the installation of the large seismograph networks (beginning in the late 1960s), and the COCORP seismic reflection program (1976 to present). The greatest impact on the future of seismology will come from our willingness and resourcefulness in collecting ample high-quality data, and using acquisition techniques that capture as much of the wavefield as possible.

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