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Excerpt
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1 Introduction

Every day there are about 50 earthquakes worldwide that are strong enough to be felt locally, and every few days an earthquake occurs that is capable of damaging structures. Each event radiates seismic waves that travel throughout Earth, and several earthquakes per day produce distant ground motions that, although too weak to be felt, are readily detected with modern instruments anywhere on the globe. Seismology is the science that studies these waves and what they tell us about the structure of Earth and the physics of earthquakes. It is the primary means by which scientists learn about Earth's deep interior, where direct observations are impossible, and has provided many of the most important discoveries regarding the nature of our planet. It is also directly concerned with understanding the physical processes that cause earthquakes and seeking ways to reduce their destructive impacts on humanity.

Seismology occupies an interesting position within the more general fields of geophysics and Earth sciences. It presents fascinating theoretical problems involving analysis of elastic wave propagation in complex media, but it can also be applied simply as a tool to examine different areas of interest. Applications range from studies of Earth's core, thousands of kilometers below the surface, to detailed mapping of shallow crustal structure to help locate petroleum deposits. Much of the underlying physics is no more advanced than Newton's second law ($F = ma$), but the complications introduced by realistic sources and structures have motivated sophisticated mathematical treatments and extensive use of powerful computers. Seismology is driven by observations, and improvements in instrumentation and data availability have often led to breakthroughs both in seismology theory and our understanding of Earth structure.

The information that seismology provides has widely varying degrees of uncertainty. Some parameters, such as the average compressional wave travel time through the mantle, are known to a fraction of a percent, while others, such as the degree of damping of seismic energy within the inner core, are known only very approximately. The average radial seismic velocity structure of Earth has been known fairly well for over 50 years, and the locations and seismic radiation patterns of earthquakes are now routinely mapped, but many important aspects of the physics of earthquakes themselves remain a mystery.

1.1 A Brief History of Seismology

Seismology is a comparatively young science that has only been studied quantitatively since about 1900. Reviews of the history of seismology include Dewey and Byerly (1969) and Agnew (2002). Early thinking about earthquakes was, as one might expect, superstitious and not very scientific. It was noted that earthquakes and volcanos tended to go together, and explanations for earthquakes involving underground explosions were common. In the early 1800s the theory of elastic wave propagation began to be developed by Cauchy, Poisson, Stokes, Rayleigh, and others who described the main wave types to be expected in solid materials. These include compressional and shear waves, termed *body waves* since they travel through solid volumes, and *surface waves*, which travel along free surfaces. Since compressional waves travel faster than shear waves and arrive first, they are often called primary or *P*-waves, whereas the later arriving shear waves are called secondary or *S*-waves. At this time theory was ahead of seismic observations, since these waves were not identified in Earth until much later.

In 1857 a large earthquake struck near Naples. Robert Mallet, an Irish engineer interested in quakes, traveled to Italy to study the destruction caused by the event. His work represented the first significant attempt at observational seismology and described the idea that earthquakes radiate seismic waves away from a focus point (now called the *hypocenter*) and that they can be located by projecting these waves backward to the source. Mallet's analysis was flawed since he assumed that earthquakes are explosive in origin and only generate compressional waves. Nevertheless, his general concept was sound, as were his suggestions that observatories be established to monitor earthquakes and his experiments on measuring seismic velocities using artificial sources.

Early seismic instrumentation was based on undamped pendulums, which did not continuously record time, although sometimes an onset time was measured. The first time-recording seismograph was built in Italy by Filippo Cecchi in 1875. Soon after this, higher-quality instruments were developed by British scientists in Japan, most notably including John Milne. The most successful of these was a horizontal pendulum design by James Ewing that recorded on a rotating disk of smoked glass. The first observation of a distant earthquake, or *teleseism*, was made in Potsdam in 1889 for a Japanese event. In 1897 the first North American seismograph was installed at Lick Observatory near San Jose in California; this device was later to record the 1906 San Francisco earthquake. These early instruments were undamped, and they could provide accurate estimates of ground motion only for a short time at the beginning of shaking. In 1898 E. Wiechert introduced the first seismometer with viscous damping, capable of producing useful records for the entire duration of an earthquake. The first electromagnetic seismographs, in which a moving pendulum is used to generate an electric current in a coil, were developed in the early 1900s, by B. B. Galitzen, who established a chain of stations across Russia. All modern seismographs are electromagnetic, since these instruments have numerous advantages over the purely mechanical designs of the earliest instruments.

The availability of seismograms recorded at a variety of ranges from earthquakes led to rapid progress in determining Earth's seismic velocity structure. By 1900 Richard Oldham reported the identification of *P*-, *S*-, and surface waves on seismograms, and later (1906) he detected the presence of Earth's core from the absence of direct *P* and *S* arrivals at source–receiver distances beyond about 100°. In 1909 Andrija Mohorovičić reported observations showing the existence of a velocity discontinuity separating the crust and mantle (this interface is now generally referred to, somewhat irreverently, as the "Moho"). Tabulations of arrival times led to the construction of travel time tables (arrival time as a function of distance from the earthquake); the first widely used tables were produced by Zöppritz in 1907. Beno Gutenberg published tables in 1914 with core phases (waves that penetrate or reflect off the core) and reported the first accurate estimate for the depth of Earth's fluid core (2900 km, very close to the modern value of 2889 km). In 1936, Inge Lehmann discovered the solid inner core, and in 1940 Harold Jeffreys and K. E. Bullen published the final version of their travel time tables for a large number of seismic phases. The JB tables are still in use today and contain times that differ by only a few seconds from current models.

The travel times of seismic arrivals can be used to determine Earth's average velocity versus depth structure, and this was largely accomplished by the 1930s. The crust varies from about 6 km in thickness under the oceans to 30–50 km beneath continents. The deep interior is divided into three main layers: the mantle, the outer core, and the inner core (Figure 1.1). The mantle is the solid rocky outer shell that makes up 84% of our planet's volume and 68% of the mass. It is characterized by a fairly rapid velocity increase in the upper mantle between about 300 and 700 km depth, a region termed the *transition zone*, where several mineralogical phase changes are believed to occur (including those at the 410- and 660-km seismic discontinuities, shown as the dashed arcs in Figure 1.1). Between about 700 km to near the core–mantle–boundary (CMB), velocities increase fairly gradually with depth; this increase is in general agreement with that expected from the changes in pressure and temperature on rocks of uniform composition and crystal structure.

At the CMB, the *P* velocity drops dramatically from almost 14 km/s to about 8 km/s and the *S* velocity goes from about 7 km/s to zero. This change (larger than the velocity contrast at Earth's surface!) occurs at a sharp interface that separates the solid mantle from the fluid outer core. Within the outer core, the *P* velocity again increases gradually, at a rate consistent with that expected for a

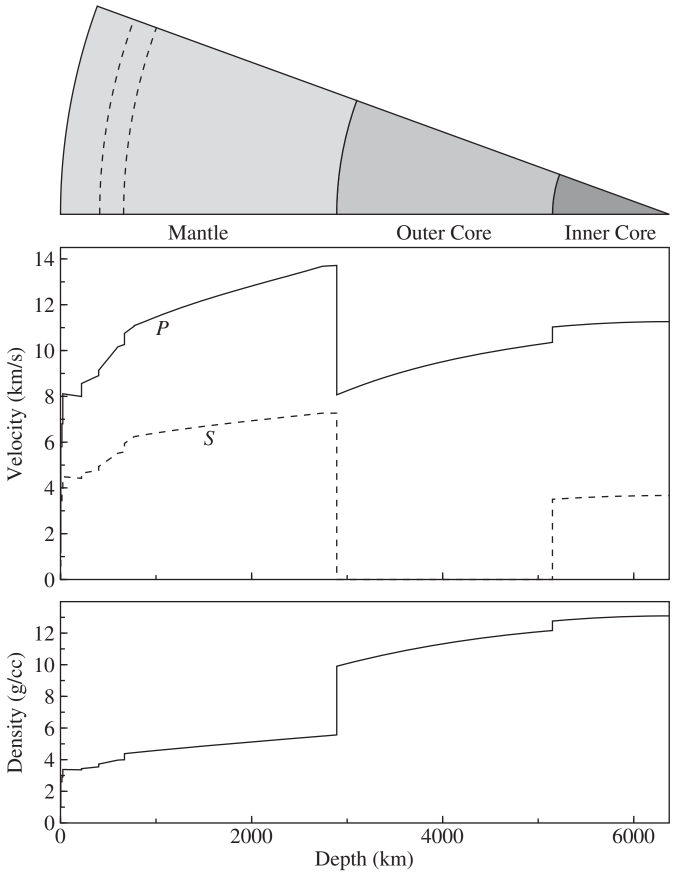

Figure 1.1 Earth's P velocity, S velocity, and density as a function of depth. Values are plotted from the Preliminary Reference Earth Model (PREM) of Dziewonski and Anderson (1981); except for some differences in the upper mantle, all modern Earth models are close to these values. PREM is listed as a table in Appendix A.

well-mixed fluid. However, at a radius of about 1221 km the core becomes solid, the P velocities increase slightly, and nonzero shear velocities are present. Earth's core is believed to be composed mainly of iron and the inner-core boundary (ICB) is thought to represent a phase change in iron to a different crystal structure.

Earth's internal density distribution is much more difficult to determine than the velocity structure since P and S travel times provide no direct constraints on density. However, by using probable velocity versus density scaling relationships and Earth's known mass and moment of inertia, K. E. Bullen showed that it is possible to infer a density profile similar to that shown in Figure 1.1. Modern results from normal mode seismology, which provides more direct constraints on density (although with limited vertical resolution), have generally proven consistent with the older density profiles.

Seismic surveying using explosions and other artificial sources was developed during the 1920s and 1930s for prospecting purposes in the oil-producing regions of Mexico and the United States. Early work involved measuring the travel time versus distance of P-waves to determine seismic velocity at depth. Later studies focused on reflections from subsurface layering (reflection seismology), which can achieve high resolution when instruments are closely spaced. The common-midpoint (CMP) stacking method for reflection seismic data was patented in 1956, leading to reduced noise levels and higher-quality profiles. The Vibroseis method, also developed in the 1950s, applies signal-processing techniques to data recorded using a long-duration, vibrating source.

The increasing number of seismic stations established in the early 1900s enabled large earthquakes to be routinely located, leading to the discovery that earthquakes are not randomly distributed but tend to occur along well-defined belts (Figure 1.2). However, the significance of these belts was not fully appreciated until the 1960s, as part of the plate tectonics revolution in the Earth sciences. At that time, it was recognized that Earth's surface features are largely determined by the motions of a small number of relatively rigid plates that drift slowly over

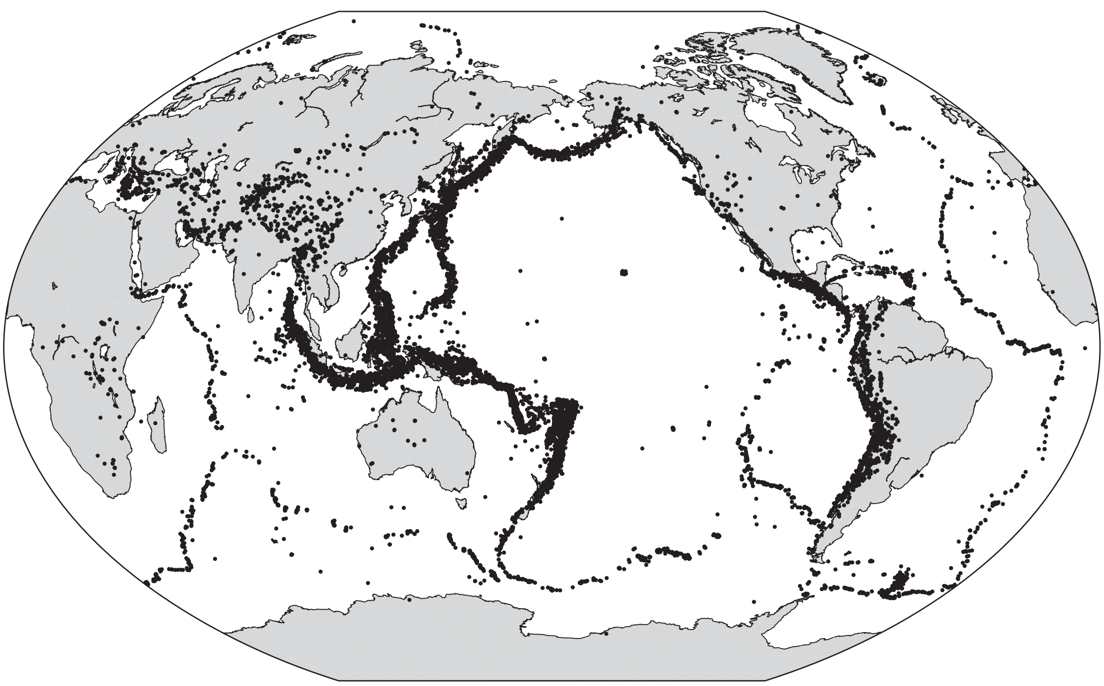

Figure 1.2 Global earthquake locations from 1977 to 2018. Earthquakes occur along well-defined belts of seismicity; these are particularly prominent around the Pacific rim and along mid-oceanic ridges. We now know that these belts define the edges of the tectonic plates within Earth's rigid outermost layer (see Figure 1.3). Data are $M > 5.5$ events from the PDE and ISC catalogs. Figure courtesy of Ian Bastow.

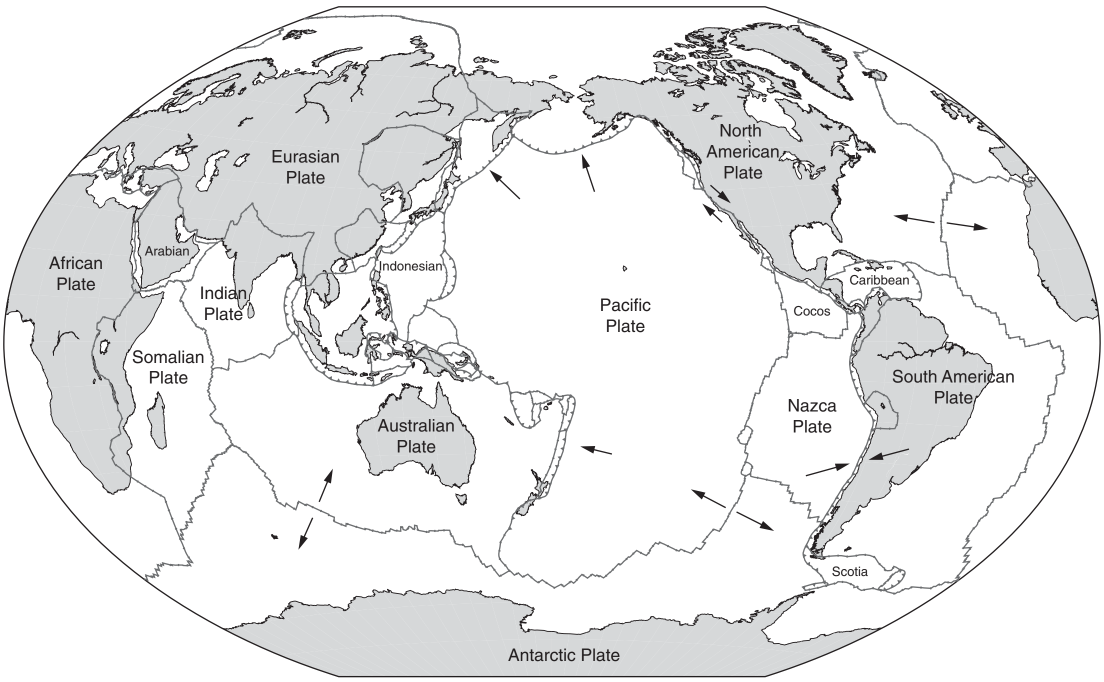

Figure 1.3 Earth's major tectonic plates. The arrows indicate relative plate motions at some of the plate boundaries. The plates are pulling apart along spreading centers, such as the Mid-Atlantic Ridge, where new crust is being formed. Along the subduction zones in the western Pacific, the Pacific Plate is sliding back down into the mantle. The San Andreas Fault in California is a result of shear between the Pacific and North American Plates. Plate boundaries from Bird (2003). Figure courtesy of Ian Bastow.

geological time (Figure 1.3). The relative motions between adjacent plates give rise to earthquakes along the plate boundaries. The plates are spreading apart along the mid-oceanic ridges, where new oceanic lithosphere is being formed. This has caused the splitting apart and separation of Europe and Africa from the Americas (the "continental drift" hypothesized by Alfred Wegener in 1912). The plates are recycled back into the mantle in the trenches and subduction zones around the Pacific margin. Large shear faults, such as the San Andreas Fault in California, are a result of transverse motion between plates. Plate boundaries across continents are often more diffuse and marked by distributed seismicity, e.g., the Himalayan region between the northward moving Indian Plate and the Eurasian Plate.

In the 1960s, seismologists were able to show that the *focal mechanisms* (the type of faulting as inferred from the radiated seismic energy) of most global earthquakes are consistent with that expected from plate tectonic theory, thus helping to validate the still emerging paradigm. However, considering the striking similarity between Figures 1.2 and 1.3, why didn't seismologists begin to develop the theory of plate tectonics much earlier? In part, this can be attributed to the lower resolution of the older earthquake locations compared to more modern results.

However, a more important reason was that seismologists, like most geophysicists at the time, did not feel that ideas of continental drift had a sound physical basis. Thus they were unable to fully appreciate the significance and implications of the earthquake locations, and tended to interpret their results in terms of local and regional tectonics, rather than a unifying global theory.

In 1923, H. Nakano introduced the theory for the seismic radiation from a double-couple source (for about the next 40 years, a controversy would rage over the question of whether a single- or double-couple source is most appropriate for earthquakes, despite the fact that theory shows that single-couple sources are physically impossible). In 1928, Kiyoo Wadati reported the first convincing evidence for deep focus earthquakes (below 100 km depth). A few years earlier, H. H. Turner had located some earthquakes at significant depth, but his analyses were not generally accepted (particularly since he also located some events in the air above the surface!). Deep focus events are typically observed along dipping planes of seismicity (often termed Wadati–Benioff zones) that can extend to almost 700 km depth; these mark the locations of subducting slabs of oceanic lithosphere that are found surrounding much of the Pacific Ocean. Figure 1.4 shows a cross section of the earthquake locations in the Tonga subduction zone in the southwest Pacific, the world's most active area of deep seismicity. The existence of deep events was a surprising discovery because the high pressures and temperatures that exist at these depths should make most materials deform ductilely, without the sudden

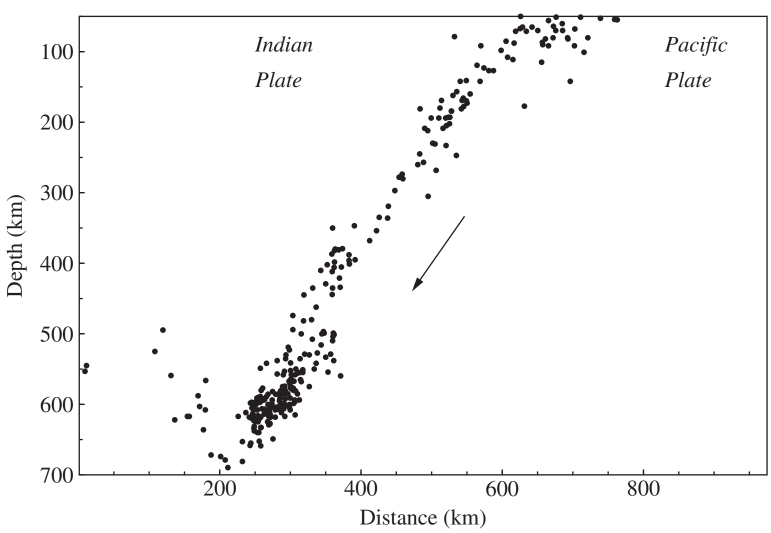

Figure 1.4 A vertical west–east cross section of the deep seismicity in the Tonga subduction zone, showing selected earthquakes from the PDE and ISC catalogs between 1977 and 1994. The seismicity marks where the lithosphere of the Pacific Plate is sinking down into the mantle.

brittle failure that causes shallow earthquakes in the crust. Even today the physical mechanism for deep events is not well understood and is a continuing source of controversy.

In 1946, an underwater nuclear explosion near Bikini Atoll led to the first detailed seismic recordings of a nuclear bomb. Perhaps a more significant development, at least for western government funding for seismology, was the 1949 testing of a Soviet nuclear bomb. This led to an intense interest by the US military in the ability of seismology to detect nuclear explosions, estimate yields, and discriminate between explosions and earthquakes. A surge in funding for seismology resulted, helping to improve seismic instrumentation and expand government and university seismology programs. In 1961 the Worldwide Standardized Seismograph Network (WWSSN) was established, consisting of well-calibrated instruments with both short- and long-period seismometers. The ready availability of records from these seismographs led to rapid improvements in many areas of seismology, including the production of much more complete and accurate catalogs of earthquake locations and the long overdue recognition that earthquake radiation patterns are consistent with double-couple sources.

Records obtained from the great Chilean earthquake of 1960 were the first to provide definitive observations of Earth's *free oscillations*. Any finite solid will resonate only at certain vibration frequencies, and these *normal modes* provide an alternative to the traveling wave representation for characterizing the deformations in the solid. Earth "rings" for several days following large earthquakes, and its normal modes are seen as peaks in the power spectrum of seismograms. The 1960s and 1970s saw the development of the field of normal mode seismology, which gives some of the best constraints on the large-scale structure, particularly in density, of Earth's interior. Analyses of normal mode data also led to the development of many important ideas in geophysical inverse theory, providing techniques for evaluating the uniqueness and resolution of Earth models obtained from indirect observations.

Between 1969 and 1972, seismometers were placed on the Moon by the Apollo astronauts and the first lunar quakes were recorded. These include surface impacts, shallow quakes within the top 100 km, and deeper quakes at roughly 800 to 1000 km depth. Lunar seismograms appear very different from those on Earth, with lengthy wavetrains of high-frequency scattered energy. This has complicated their interpretation, but a lunar crust and mantle have been identified, with a crustal thickness of about 60 km (see Figure 1.5). A seismometer placed on Mars by the *Viking 2* probe in 1976 was hampered by wind noise and only one possible Mars quake was identified. The InSight Mission landed a probe on Mars in late 2018 and deployed a seismometer, but results were not available before this book went to press.

Although it is not practical to place seismometers on the Sun, it is possible to detect oscillations of the solar surface by measuring the Doppler shift of spectral

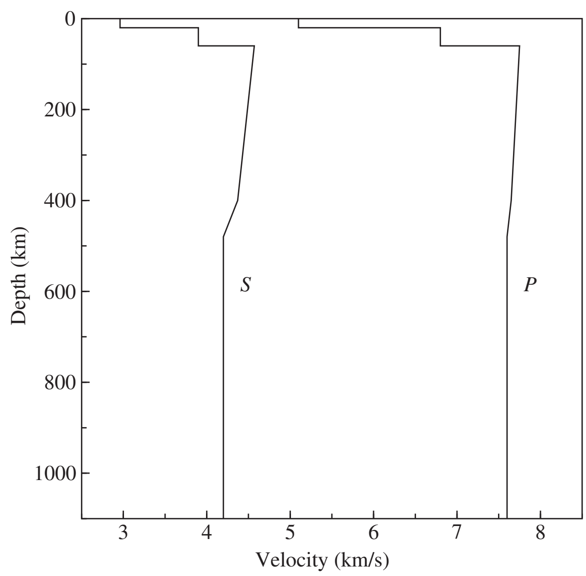

Figure 1.5 An approximate seismic velocity model derived for the Moon from observations of quakes and surface impacts (plot based on model from Goins et al., 1981). Velocities at greater depths (the lunar radius is 1737 km) are largely unconstrained owing to a lack of deep seismic waves in the *Apollo* data set.

lines. Such oscillations were first observed in 1960 by Robert Leighton, who discovered that the Sun's surface vibrates continually at a period of about five minutes and is incoherent over small spatial wavelengths. These oscillations were initially interpreted as resulting from localized gas movements near the solar surface, but in the late 1960s several researchers proposed that the oscillations resulted from acoustic waves trapped within the Sun. This idea was confirmed in 1975, when it was shown that the pattern of observed vibrations is consistent with that predicted for the free oscillations of the Sun, and the field of *helioseismology* was born. Analysis is complicated by the fact that, unlike Earth, impulsive sources analogous to earthquakes are rarely observed; the excitation of acoustic energy is a continuous process. However, many of the analysis techniques developed for normal mode seismology can be applied, and the radial velocity structure of the Sun is now well constrained (Figure 1.6). Continuing improvements in instrumentation and dedicated experiments promise further breakthroughs, including resolution of spatial and temporal variations in solar velocity structure. In only a few decades, helioseismology has become one of the most important tools for examining the structure of the Sun.

The advent of computers in the 1960s changed the nature of terrestrial seismology by enabling analyses of large data sets and more complicated problems and led to the routine calculation of earthquake locations. The first complete theoretical seismograms for complicated velocity structures began to be computed at this time. At first these calculations were restricted to 1-D models,

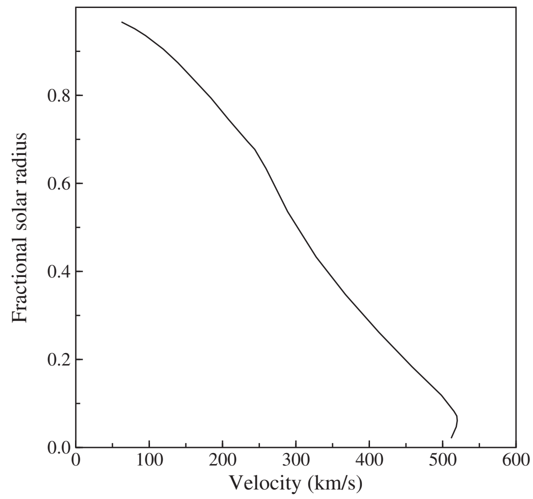

Figure 1.6 The velocity of sound within the Sun (plot based on model from Harvey, 1995).

in which properties vary only with depth, but continued advances in computer power now make possible exact calculations for 3-D models of arbitrary complexity. The computer era also has seen the rapid expansion of seismic imaging techniques using artificial sources that have been applied extensively by the oil industry to map shallow crustal structure. Beginning in 1976, data started to become available from global seismographs in digital form, greatly facilitating quantitative waveform comparisons. In the 1980s and 1990s, many global seismic stations were upgraded to broadband, high dynamic range seismometers, and new instruments were deployed to fill in gaps in the global coverage. Large numbers of portable instruments have also become available for specialized experiments in particular regions. Seismic records are now far easier to obtain than in the predigital era, with centralized archives providing online data access in standard formats.

Earth's average radial velocity and density structures were well established by 1970, including the existence of minor velocity discontinuities near 410- and 660-km depth in the upper mantle. Attention then shifted to resolving lateral differences in velocity structure, first by producing different velocity versus depth profiles for different regions, and more recently by inverting seismic data directly for three-dimensional velocity structures. The latter methods have been given the name "tomography" by analogy to medical imaging techniques. During recent years, tomographic methods of increasing resolution have begun to provide spectacular images of the structure of Earth's crust and mantle at a variety of scale lengths. Local earthquake tomography at scales from tens of hundreds of kilometers has imaged details of crustal structure in many different regions,