54

Seismic Velocity Structure of the Continental Lithosphere from Controlled Source Data

Walter D. Mooney
US Geological Survey, Menlo Park, CA, USA
Claus Prodehl
University of Karlsruhe, Karlsruhe, Germany
Nina I. Pavlenkova
RAS Institute of the Physics of the Earth, Moscow, Russia

1. Introduction

The purpose of this chapter is to provide a summary of the seismic velocity structure of the continental lithosphere, i.e., the crust and uppermost mantle. We define the crust as the outer layer of the Earth that is separated from the underlying mantle by the Mohorovičić discontinuity (Moho). We adopted the usual convention of defining the seismic Moho as the level in the Earth where the seismic compressional-wave (*P*-wave) velocity increases rapidly or gradually to a value greater than or equal to 7.6 km sec^{-1} (Steinhart, 1967), defined in the data by the so-called "*Pn*" phase (*P*-*n*ormal). Here we use the term uppermost mantle to refer to the 50–200+ km thick lithospheric mantle that forms the root of the continents and that is attached to the crust (i.e., moves with the continental plates).

This summary has been preceded by 90 y of intense scientific activity. Mohorovičić (1910) was the first to publish an

TABLE 1 Published summaries of crustal structure by region covered

Year	Authors	Areas covered	J/A/B[a]
1961	Closs and Behnke	World	J
1963	Hill	Oceans	B
1966	James and Steinhart	World	B
1966	McConnell *et al.*	World	J
1969	Hart	World	B
1970	Ludwig *et al.*	Oceans	A
1970	Maxwell	Oceans	B
1971	Heacock	N-America	B
1973	Meissner	World	J
1973	Mueller	World	B
1975	Makris	E-Africa, Iceland	A
1977	Bamford and Prodehl	Europe, N-America	J
1977	Heacock	Europe, N-America	B
1977	Mueller	Europe, N-America	A
1977	Prodehl	Europe, N-America	A
1978	Mueller	World	A
1980	Zverev and Kosminskaya	Europe, Asia	B
1982	Soller *et al.*	World	J
1984	Prodehl	World	A
1986	Meissner	Continents	B
1987	Orcutt	Oceans	J
1989	Mooney and Braile	N-America	A
1989	Pakiser and Mooney	N-America	B
1991	Beloussov *et al.*	Europe, Asia	B
1991	Collins	Australia	A
1992	Blundell *et al.*	Europe	B
1992	Holbrook *et al.*	Continents	A
1992	Mooney and Meissner	Continents	A
1993	Mueller and Kahle	Mediterranean	A
1994	Durrheim and Mooney	Continents	J
1995	Olsen	Continental rifts	B
1995	Christensen and Mooney	World	J
1995	Rudnick and Fountain	World	J
1995	Tanimoto	World	A
1996	Pavlenkova	Europe, Asia	A
1997	Prodehl *et al.*	Near East, Africa	J
1998	Li and Mooney	China	J
1998	Mooney *et al.*	World	J

[a]J, Article published in an international journal; A, Article published as part of a book; B, Book publication.

INTERNATIONAL HANDBOOK OF EARTHQUAKE AND ENGINEERING SEISMOLOGY, VOLUME 81A ISBN: 0-12-440652-1

estimate for crustal thickness (54 km near Zagreb, Croatia), and to describe the seismically defined boundary between the crust and mantle that now bears his name (often shortened to "the Moho"). The fact that the oceanic crust is significantly thinner than continental crust (5 km versus about 40 km) was documented 40 y later (e.g., Hersey *et al.*, 1952). Numerous later studies demonstrated that the continental crust varies in thickness from about 15 km to greater than 70 km beneath the Tibet plateau. Jarchow and Thompson (1989) provide a useful summary of early crustal studies, and Table 1 provides additional references. We emphasize results from active-source seismic refraction profiles that provide detailed *P*-wave velocity information. The shear-wave (*S*-wave) structure of the crust and uppermost mantle, as determined by surface waves, is discussed, for example, by Ekström *et al.* (1997) and Ritzwoller *et al.* (1998).

2. Data and Methods of Analysis

The properties of the Earth's lithosphere are particularly well suited to measurements using seismological data due to the existence of the pronounced variations in seismic wave speeds both laterally and with depth. Beneath continents, the Moho is characterized by a more or less rapid increase of *P*- and *S*-velocities with depth. However, because *S*-waves travel more slowly than *P*-waves and therefore cannot be detected as first arrivals, they are often hidden in the coda of the *P*-waves. Thus, the study of *S*-waves for crustal research has been a limited, but rapidly growing, area of research, and a worldwide mapping of crustal and uppermost mantle *S*-wave results from seismic refraction data is not yet possible.

The chief advantages of controlled-source seismic-refraction data are: (1) The exact time and position of the seismic sources are accurately known; (2) It is possible to plan, in relation to the specific geologic target, the number, spacing, and geometry of sources and receivers. The main disadvantages of these seismic-refraction data are: (1) The seismic sources are limited to the surface of the Earth, whereas natural seismicity extends to a depth of 10–20 km within the crust, and to a maximum depth of 670 km in subduction zones; (2) There is a practical limit to the strength of chemical or other controlled sources, thereby limiting the depth of investigation to the lithosphere.

It is important to note that vertical-incidence seismic reflection profiles, which are not reviewed here, provide the highest resolution of structural detail within the crust. These data provide an image of the entire crust that is similar to the pictures obtained within sedimentary basins by the oil exploration industry. The application of seismic reflection profiles to the study of the continents began in Germany (cf., discussion in Meissner, 1986), and was adopted in the United States of America in 1974 by COCORP (Barazangi and Brown, 1986a,b). Many other countries soon followed suit with their own national programs, such as BIRPS (UK, Matthews, 1986), DEKORP (Germany, Meissner *et al.*, 1991). Crustal reflection profiles show that the crust has many fine-scale impedance contrasts that are related to compositional layering, shear zone and fractures, and other geological features, such as dikes and sills (Brown *et al.*, 1983; Matthews, 1986; Meissner, 1986; Barazangi and Brown, 1986a,b; Mooney and Meissner, 1992; Klemperer and Mooney, 1998a,b). Deep reflections from the subcrustal lithosphere have also been recorded in many regions. Consequently, the advent of deep seismic reflection profiling has led to a major new understanding of the evolution of the continental lithosphere. Optimal information is obtained when coincident seismic refraction and vertical-incidence reflection profiles are collected, as has been the case for the Canadian LITHOPROBE program (Clowes *et al.*, 1999).

Since the early 1950s a large number of seismic profiles have been collected around the world. Advancing technology has led not only to better instrumentation, but also to improved data processing and interpretation techniques. Useful recent reviews of seismic methods for determining crust and upper mantle structure have been provided by Mooney (1989) and Braile *et al.* (1995), whereas Giese *et al.* (1976) summarize older methods of data analysis. Table 1 lists previously published summaries of crustal structure by region covered. The papers and books most comparable to the present work are Soller *et al.* (1982), Prodehl (1984), Meissner (1986), Tanimoto (1995), Pavlenkova (1996), Chuikova *et al.* (1996), and Mooney *et al.* (1998).

Resolution is a major consideration when discussing seismic data. Because of the variety in seismic techniques, however, general statements about resolution are difficult. In seismology, resolution is related both to data quality and physical laws. The quality of a data set is usually considered to be a function of the strength of the signal relative to noise (expressed as signal-to-noise ratio) and the number or density of measurements depending on the number of sources and receivers and their relative spacing. Generally, the more data available, the higher is the resolution. However, physical laws limit the resolution of even near-perfect data sets. The Earth strongly attenuates high frequencies so that signals penetrating deeper will contain relatively low frequencies. Typically, signals traversing the whole crust have peak frequencies of 5–20 Hz resulting in absolute accuracy of depth determination of not better than 2–5% of the depth (e.g., 1–2 km for a 40 km-thick crust).

It is also of interest to use seismic velocities to infer the composition of the lithosphere. This can be accomplished by comparison with laboratory measurements of the seismic velocity of rock specimens believed to have once resided at depth within the lithosphere. This research was pioneered by Francis Birch, who measured numerous igneous rock samples at confining pressures up to 100 MPa and compared these results with early field determinations of seismic velocities (Birch, 1960, 1961). Such measurements have been expanded by others to include metamorphic rocks, the measurement of shear-wave velocities, and the determination of temperature effects (Simmons, 1964; Christensen, 1965, 1966, 1979; Kern, 1978). These studies are reviewed in Holbrook *et al.* (1992), Rudnick and Fountain (1995), and Christensen and Mooney (1995). The interplay of temperature and pressure is worth noting here; a temperature increase of 100°C at constant pressure will decrease seismic velocity by 0.05 km sec^{-1}. However, the influence of increased temperature with depth is compensated for by increased pressure. Thus, seismic velocities will generally increase with depth only if there is a change in rock composition, or an unusually low temperature gradient. An exception is the uppermost (0–5 km) crust, where the closing of cracks in the crystalline crust will lead to an increase in both *P*- and *S*-wave velocities.

3. Main Features of Continental Crustal Structure

Over the past 50 y a vast amount of data has been accumulated regarding the seismic structure of the continental crust. A typical data example is shown in Figure 1. The basic features of continental crustal structure were recognized by the 1960s (Steinhart and Meyer, 1961; James and Steinhart, 1966; Ludwig *et al.*, 1970; Pavlenkova, 1973). The seismic velocity distributions vary widely in different geographic localities, and crustal models generally consist of two, three, or more layers separated by velocity discontinuities or gradients. In relatively stable continental regions the thickness of the crust is between 30 and 50 km. Seismic velocities in the upper crustal layer are usually 5.6–6.3 km sec^{-1}. At a depth of 10–15 km the seismic velocity commonly increases to 6.4–6.7 km sec^{-1}. This velocity increase is sometimes referred to as the Conrad discontinuity, a term that is now in disfavor since the velocity structure of the crust has been found to be sufficiently complex that the use and definition of this term is ambiguous. In many stable continental interiors there is a third, basal crustal layer with a velocity of 6.8–7.2 km sec^{-1}. The seismic velocity below the Moho (*Pn* velocity) is typically about 8 km sec^{-1}. The exact nature of the velocity–depth distribution is not always well defined. The evidence for distinct layers within the continental crust almost exclusively depends on the interpretation of second-arrival phases. In some regions, clear evidence of later arrivals confirms that the velocity increases discontinuously

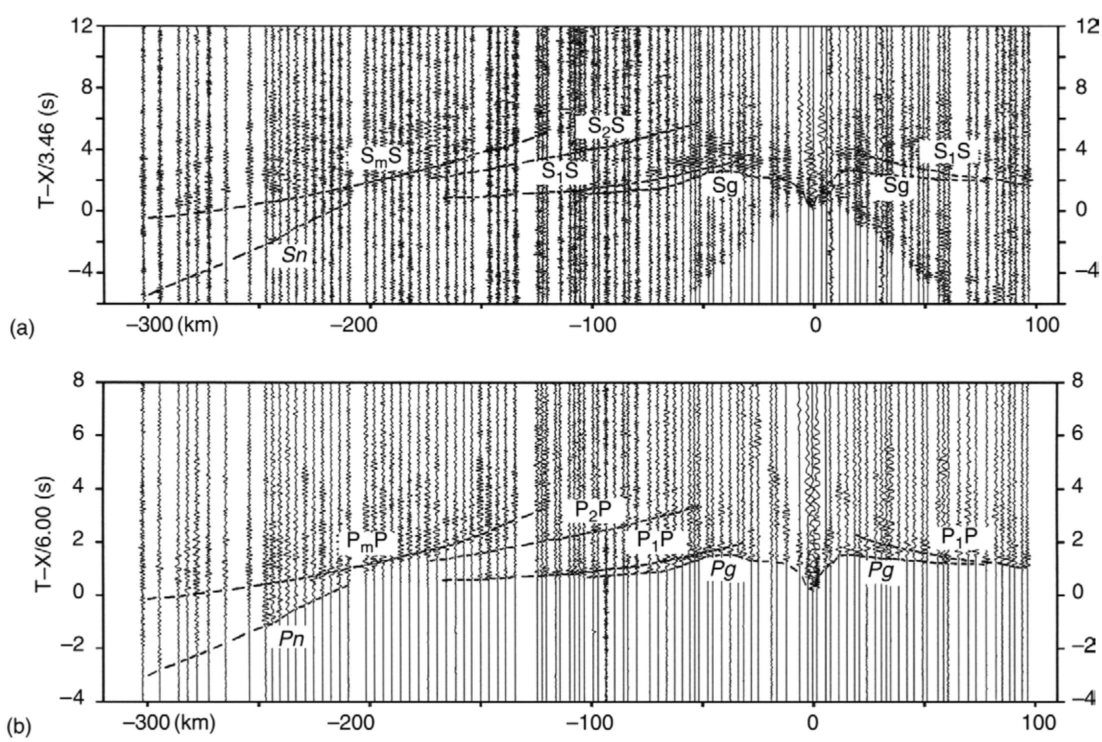

FIGURE 1 Sample record section from western China. (a) Shear (*S*) wave record section with a reduction velocity of 3.46 km sec^{-1}. Shot point corresponds to 0 km; *Sg*, direct arrival; S1S, reflection from boundary between upper and middle crust; S2S, reflection from middle and lower crust; SmS, reflection from Moho; *Sn*, refraction from uppermost mantle. (b) Compressional (*P*) wave record section with a reduction velocity of 6.0 km sec^{-1}; *Pg*, direct arrival; P1P, P2P, PmP, *Pn*, *P*-wave arrivals as defined for *S*-waves above. (Wang, Y.X. *et al.*, submitted 2002).

through intermediate layers within the crust, whereas in other regions velocity may increase gradually with increasing depth producing no distinct intracrustal reflection (Levander and Holliger, 1992).

The compilation of seismic crustal data around the world has defined the characteristic primary crustal types connected with specific tectonic settings (Fig. 2). Each primary crustal type is defined by the average seismic model of the crust of a specific age or tectonic setting (Fig. 3).

Despite the abundance of refraction-seismic experiments having been carried out throughout the world, the coverage is very selective (Fig. 4). For example the continents of the northern hemisphere are reasonably well covered, whereas in the southern hemisphere only selective seismic surveys dealing with particular target areas, such as the Andes of South America and the East African Rift System have been studied. Figure 3 shows a world map indicating the main tectonic units of the continents. The map also indicates the location of the seismic cross sections that are discussed later.

Figure 4 shows that Eurasia is densely covered with seismic profiles, as is the central portion of North America (the United States of America and southern Canada). Only limited data are available for northern Canada, Mexico, Central America, and Greenland. The South American and African continents have a concentration of data points in select locations, but vast regions remain without seismic control. Data are lacking for significant portions of the Middle East and Southeast Asia. Data coverage is good within Australia, New Zealand, and the coastal portions of Antarctica. These data points have been used to map the thickness of the Earth's crust (Fig. 5). Crustal thickness is estimated in regions lacking data by determining the geologic setting (Fig. 3) and assuming a typical crustal structural for that setting (Fig. 2). Short period (35–40 sec) Love-wave phase and group velocity maps (Ekström *et al.*, 1997; Ritzwoller *et al.*, 1998), which are mainly sensitive to crustal thickness and average crustal shear velocity, provide a useful guide to estimating crustal thickness in regions lacking active-source seismic profiles (cf., Mooney *et al.*, 1998).

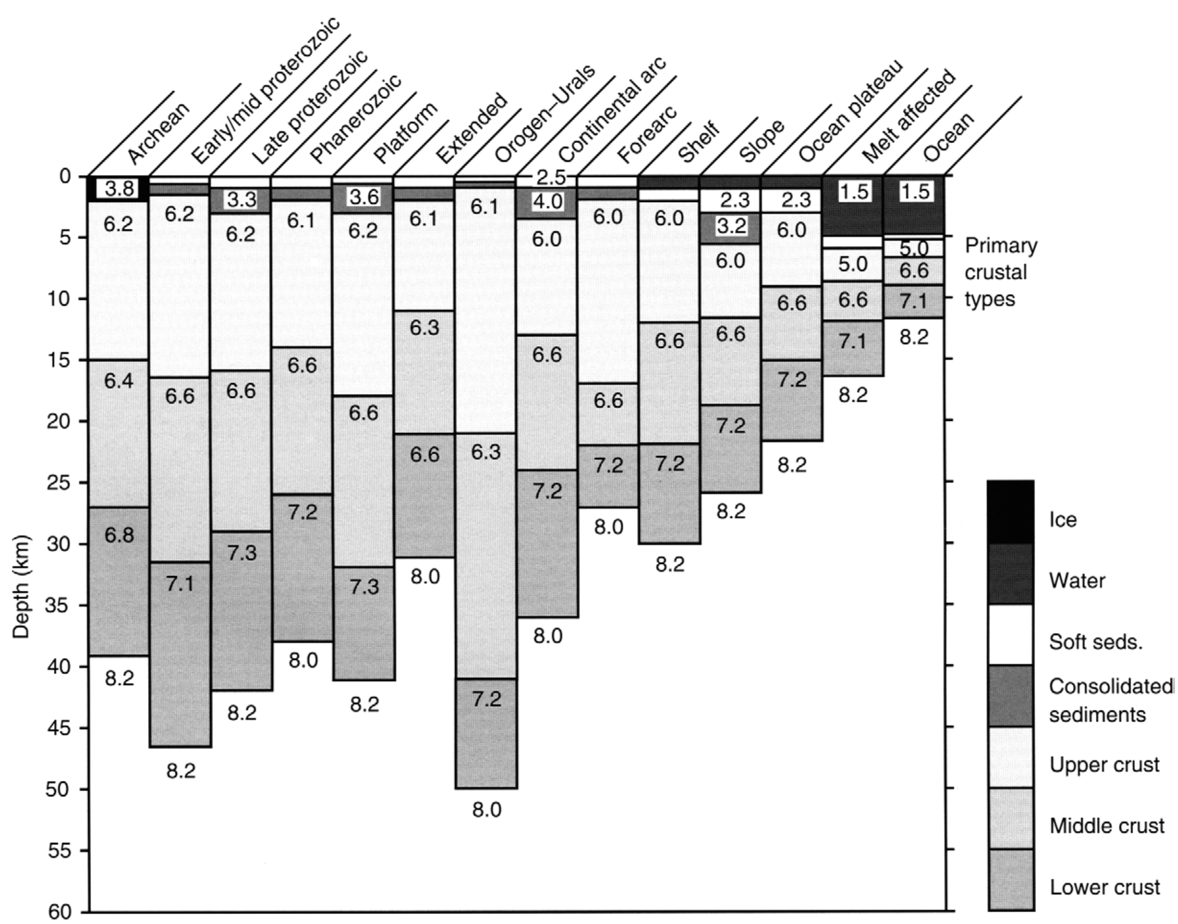

FIGURE 2 Fourteen common continental and oceanic crustal types (Mooney *et al.*, 1998). Typical *P*-wave velocities are indicated for the individual crustal layers and the uppermost mantle. Velocities refer to the top of each layer, and there is commonly a velocity gradient of 0.01–0.02 $km\,sec^{-1}\,km^{-1}$ within each layer. The crust thins from an average values of 40 km in continental interiors to 12 km beneath oceans.

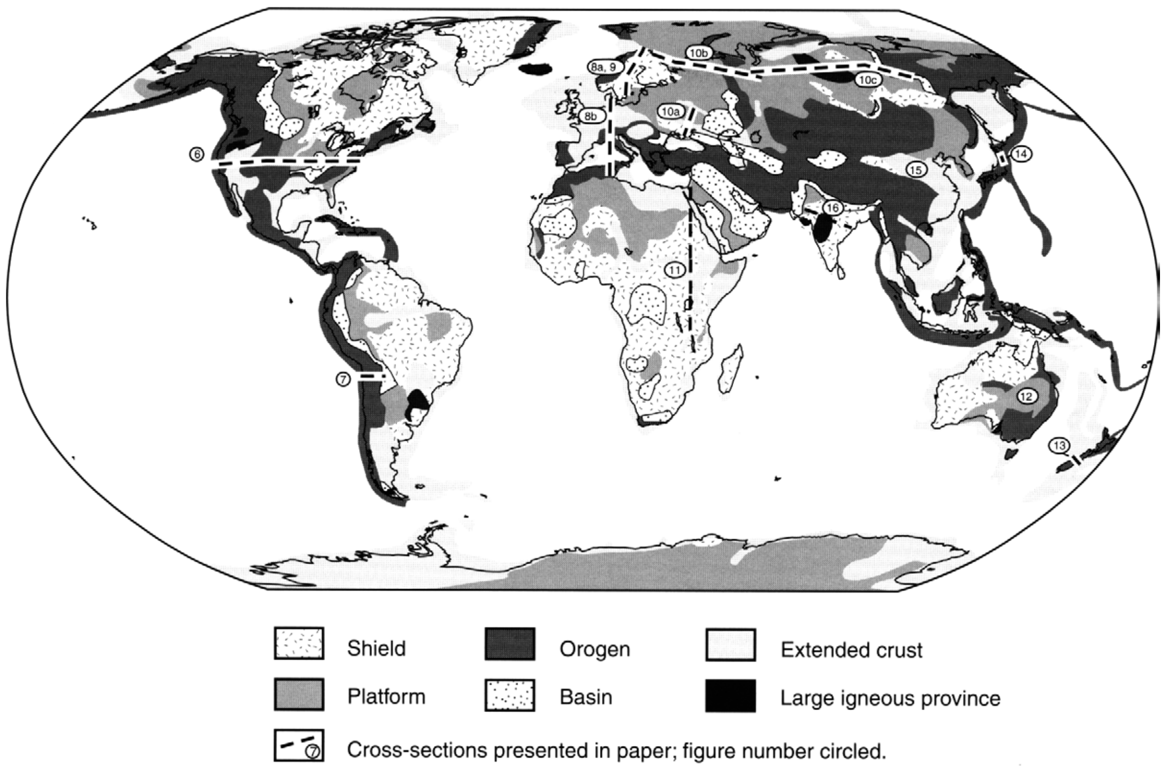

FIGURE 3 World map showing six primary Geological provinces and the Figure numbers of cross sections presented in this chapter. Index of locations shown on the map: N-America (Fig. 6) (38° N); N-Chile (Fig. 7); European Geo-Traverse (EGT; Figs. 8 and 9); Russia (Fig. 10); Afro-Arabian rift system (Fig. 11); Australia (Fig. 12); New Zealand (Fig. 13); Japan: northern Honshu (Fig. 14); China (Fig. 15); India (Fig. 16).

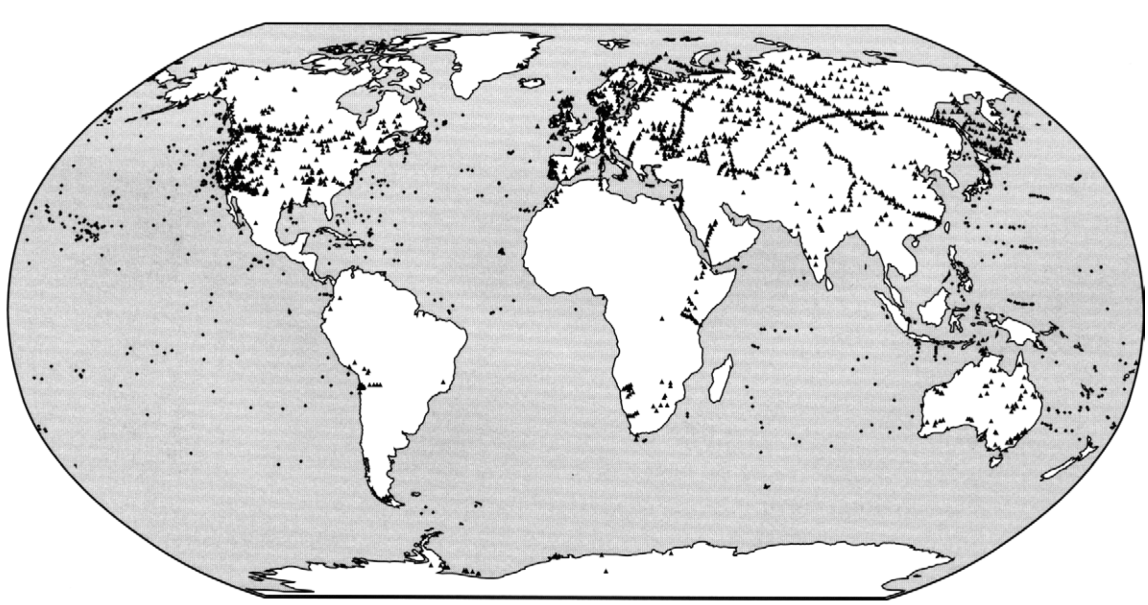

FIGURE 4 World map showing the location of seismic refraction profiles used to compile the maps of crustal thickness (Fig. 17) and mean crustal velocity (Fig. 18) (Mooney *et al.*, 1998). Each triangle corresponds to a point along a seismic refraction profile where a crustal column has been extracted.

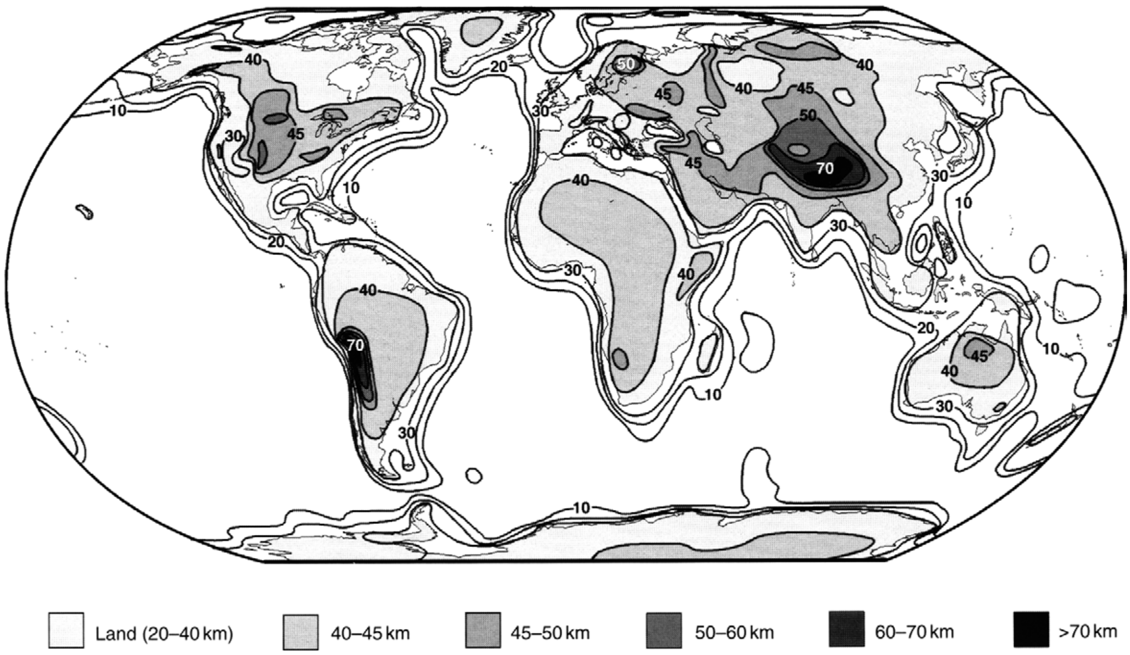

FIGURE 5 Global crustal thickness (Mooney *et al.*, 1998). The average thickness on continents is 40 km, and generally increases in thickness from 30 km near the margins toward the interior. In this figure, crustal thickness under narrow features (e.g., continental rifts, midocean ridges and oceanic fracture zones) is not visible.

The thickest crust (70+ km) occurs beneath the Tibetan Plateau and the South American Andes. The thickness of continental crust with an elevation above sea level is generally between 28 km and 52 km, and the average thickness (excluding continental margins) is 40 km. Aside from the two regions mentioned above, very little of the continental crust is thicker than 50 km. Likewise, continental crust thinner than 30 km is generally limited to rifts and highly extended crust including continental margins. It is important to bear in mind that the Earth's crust is very complex, and there are numerous local exceptions to these generalizations. In the following sections we provide a detailed description of the structure of the crust on a continent-by-continent basis.

3.1 North America

The deep structure of the North American lithosphere has been the subject of intense investigation since the pioneering work of M.A. Tatel and H.E. Tuve at the Carnegie Institution of Washington in the early 1950s (e.g., Tatel and Tuve, 1955). These and other early investigations established the main feature of a thick (~45 km) crust within the continental interior and thin (~30 km) crust at the margins. They also discovered unexpectedly thin (~30 km) crust beneath the Basin and Range Province, where a high average surface elevation of 1–2 km can be found.

Several summaries of the crustal structure of North America have been published over the past 25 y (Pakiser and Steinhart, 1964; Hart, 1969; Heacock, 1971, 1977; Prodehl, 1977; Pakiser and Mooney, 1989; Mooney and Braile, 1989). A great deal of crustal structure data has also been published by Canadian researchers over the past 15 y under the LITHOPROBE Program (Clowes *et al.*, 1999). These Canadian investigations have used coincident seismic reflection and refraction data to obtain higher-resolution data than previously was possible. A recent compilation of North American data contains more than 1400 measurements of crustal structure (Chulick and Mooney, 2002).

The contour map of crustal thickness of North America (Fig. 5) shows the previously mentioned pattern of increasing thickness toward the interior. The average thickness of crust is 36.7 km, with a standard deviation (SD) of 8.4 km. The thickest crust (50–55 km) occurs in four relatively isolated regions. The thinnest crust is associated with continental margins and regions that have undergone recent extension, including the Western Cordillera of Canada, the Rio Grande Rift, and the Basin and Range Province.

The upper mantle velocity, *Pn*, beneath North America is 8.03 km sec^{-1} with a standard deviation of 0.19 km sec^{-1}. Temperature plays a major role in determining *Pn* velocity, and thus the cold lithosphere of the continental interior typically has a *Pn* velocity of 8.1–8.2 km sec^{-1}, whereas velocities of 7.8–8.0 km sec^{-1} are measured beneath the warm crust of the Western Cordillera and the Basin and Range Province.

The average *P*-wave velocity of the crust is related to crustal composition. The global average is 6.45 km sec^{-1}

(SD 0.23), and corresponds to an average crustal composition equivalent to a diorite. The North American continental interior has an average crustal velocity of 6.5–6.7 km sec^{-1}, indicating that its bulk composition is denser and more mafic than the global average. The continental margins, Basin and Range Province, Sierra Nevada (California) and the Canadian Cordillera all have an average crustal velocity of 6.2–6.3 km sec^{-1}. This can be attributed to a missing, or very thin high-velocity (6.9–7.3 km sec^{-1}) lower crustal layer. Elevated temperature may also play a role in reducing the average crustal velocity in these regions.

Many of the features of the lithospheric structure of North America can be seen in a cross section from San Francisco, California, to Washington, DC at 38° N latitude (Fig. 6). This cross section shows that the crust is highly variable in structure. The western USA has a relatively thin crust, averaging 30 km, with a low average crustal velocity (6.2 km sec^{-1}). Surprisingly, a pronounced crustal root does not underlie the topographically high Rocky Mountains. Indeed, the thickest crust (45–50 km) underlies the low-lying Great Plains. The crust thins modestly to the east beneath the Appalachian Foreland, and then thins beneath the Atlantic Coastal Plain. The velocity of the uppermost mantle increases from 7.8–7.9 km sec^{-1} beneath the western USA to more typical continental values of 8.0–8.1 km sec^{-1} beneath the central and eastern USA. A laterally variable *S*-wave low-velocity zone exists at a depth of 60–180 km beneath North America, and the top of this zone is generally taken as the depth of the seismic lithosphere. The thermal lithosphere is commonly defined as the depth to the 1300°C isotherm and can be estimated from heat flow data. In Figure 6 we consider both the seismic and thermal lithosphere and estimate lithospheric thickness based on the summary of seismic results by Iyer and Hitchcock (1989) and the thermal estimates of Pollack and Chapman (1977) and Artemieva and Mooney (2001). The lithosphere thickens from about 60 km beneath the Basin and Range province to about 180 km beneath the Precambrian continental interior. We estimate a lithospheric thickness of 100 km at the Atlantic coastal margin.

3.2 South America

Few seismic refraction measurements have been made in South America, with the exception of the Andean region. East of the Andes the crust consists of a collage of Precambrian shields, platforms and paleorifts (Goodwin, 1991, 1996). Sparse refraction data indicate an average crustal thickness of 40 km, with typical shield velocity structure (Fig. 7). This estimate is consistent with phase velocity measurements of surface waves (Ekström *et al.*, 1997) and determinations of crustal structure from passive seismology (Myers *et al.*, 1998).

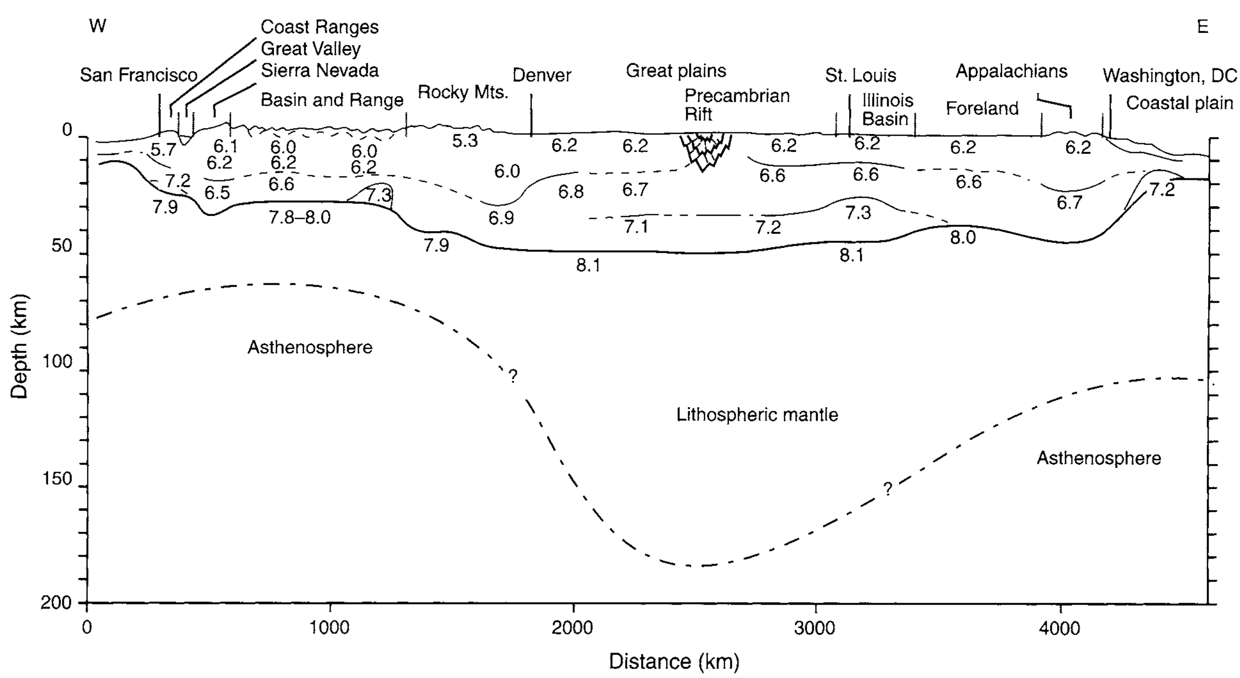

FIGURE 6 Lithospheric section through the conterminous United States. The section traverses the continent from east to west at approximately 38° N. Solid lines indicate velocity discontinuities where well-determined, dashed lines were inferred. The seismic velocity structure is generalized, and fine-scale structure is not included. The base of the lithosphere is from Iyer and Hitchcock (1989), Pollack and Chapman (1977) and Artemieva and Mooney (2001). Vertical exaggeration is 10 : 1.

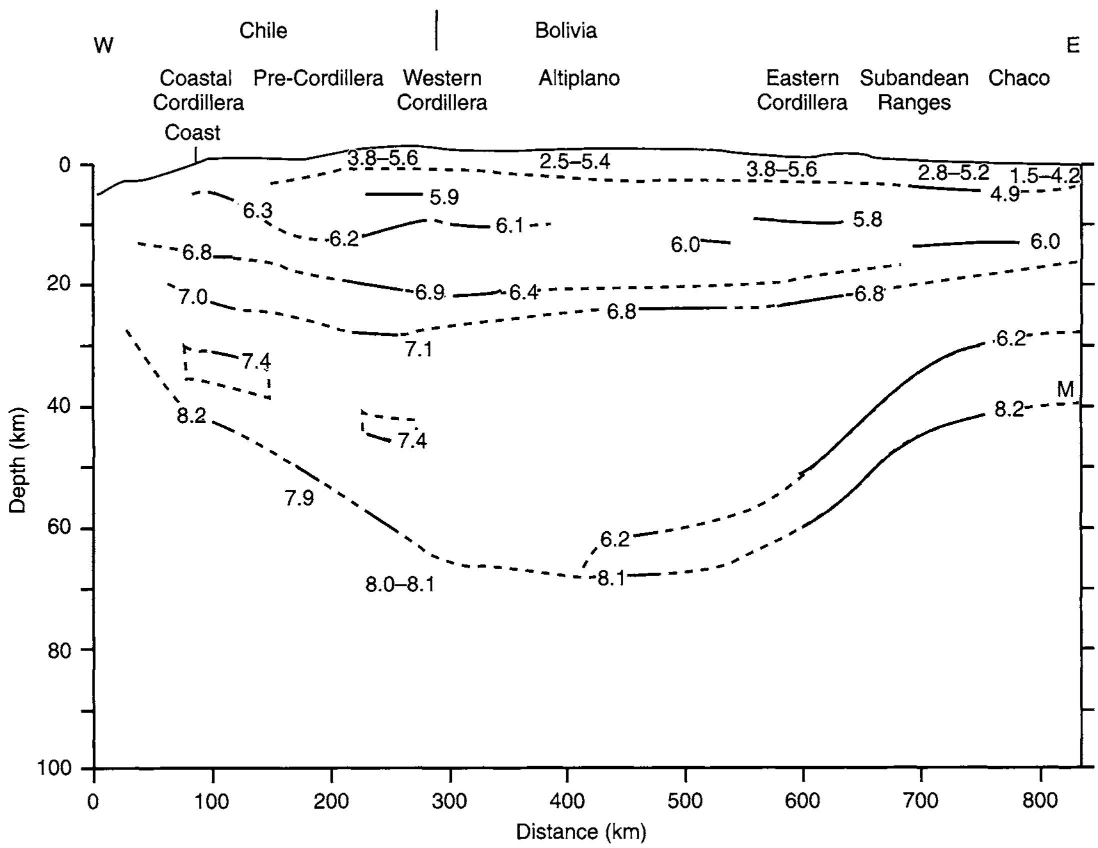

FIGURE 7 East–west lithospheric section through South America at approximately 21° S (see Fig. 4). Vertical exaggeration is 5 : 1. Source: Wigger *et al.* (1994), Giese *et al.* (1999), Patzwahl *et al.* (1999).

The structure of western South America is dominated by mountain building and volcanism associated with the subduction of the Pacific Plate beneath the continent. The structure of the central Andes was first determined from surface wave observations in a classic paper by James (1971) that demonstrated that the high Andes of Peru are underlain by a remarkably thick crust (~60 km). These results have been confirmed by more recent surface-wave phase-velocity measurements (Ekström *et al.*, 1997). In the northern Andes of Colombia, Meyer *et al.* (1976) present a cross section from the Pacific basin into the Western Cordillera. The plate dips at ~15°, and the velocities within the continental crust are higher than average due to the Cenozoic accretion of a basaltic oceanic plateau.

The southern Andes between 19° and 25° S were the subject of numerous seismic investigations during the 1990s by South American, German, and American researchers (e.g., Wigger *et al.*, 1994; Giese *et al.*, 1999; Patzwahl *et al.*, 1999). These studies have contributed a great deal of information regarding the deep structure, and confirm the existence of thick crust south of Peru (Fig. 7). Several detailed seismic offshore and onshore refraction and reflection profiling campaigns (e.g., Wigger *et al.*, 1994; Patzwahl *et al.*, 1999) were undertaken to investigate the subduction of the oceanic Nazca plate under the South American continent between the Peru–Chile trench and the Central Andes in northern Chile between 19° and 25° S. The active seismic campaigns provided details of crustal and upper mantle structure to depths of 70 km and more, and were supported by long-term passive seismological investigations that resolved structure to more than 100 km. The data clearly show the Nazca plate subducting under an increasing angle of 9–25° down to 30–50 km depth near the coast, and they show a portion of the Moho dipping eastward from 30–50 km near the coast to 55–64 km up to about 240 km inland.

Crustal thickness beneath the western Cordillera and the Altiplano is at least 70 km (Fig. 7). However, only weak Moho arrivals are observed by active seismic measurements beneath the Altiplano and the Western Cordillera. Broadband seismologic data are more reliable in these regions and indicate an approximately 60–70 km thick crust. In the Precordillera a pronounced Moho discontinuity is detected at a depth of 50 km. Along the coast, the oceanic Moho boundary can be identified at a depth of 40–45 km. The seismic measurements have revealed a clearly pronounced asymmetric structure and evolution of the Central Andean crust.

In the Andean backarc region, tectonic compression has produced pronounced crustal thickening and, locally, a doubling of the Moho discontinuity (Giese *et al.*, 1999). In the forearc region, dehydration of the lower plate may have caused hydration and metasomatism of the overlying mantle wedge that is associated with a decrease in density and seismic velocity to values typical of the crust. Such processes, if real, may reduce the velocity contrast at the Moho and result in a blurring of the crust–mantle boundary from a geophysical point of view. The paleo-Moho today appears as an intra-crustal boundary. Magmatic intrusions and underplating also have the potential to produce new seismic boundaries, including one that appears to be the Moho discontinuity. Giese *et al.* (1999) summarize their interpretation of the Central Andes by suggesting that there exists a variety of different Moho-discontinuities, which they refer to as a "Moho-menagerie". This term may be understood to refer to a highly complex seismic velocity structure in the lower crust and uppermost mantle.

3.3 Europe and Mediterranean

Europe becomes progressively younger in tectonic age from north to south. The northern Baltic Shield consists of Archean and Proterozoic rocks; to the west of the Baltic Shield the Caledonian orogeny has added crust of Paleozoic age. The basement rocks of central Europe are of Paleozoic age and formed during the Variscan orogeny, covering the area between Scotland and Portugal. The terranes that comprise this region have undergone strong orogenic collapse (extension), as evidenced by a relatively thin lithosphere and crust. The Tornquist Zone separates Paleozoic Western Europe from the extensive Precambrian platform areas of Eurasia. The Neogene Alpine orogeny shaped southern Europe, from the Spanish Pyrenees to the Carpathians, and provide a physiographic separation between central Europe and the Mediterranean.

The European Geotraverse (EGT) was a multinational geologic and geophysical lithospheric investigation carried out in the 1980s that reached from the northernmost tip of Scandinavia to Tunisia in Northern Africa. This investigation include very detailed seismic-refraction surveys, the results of which are summarized in Figure 9.

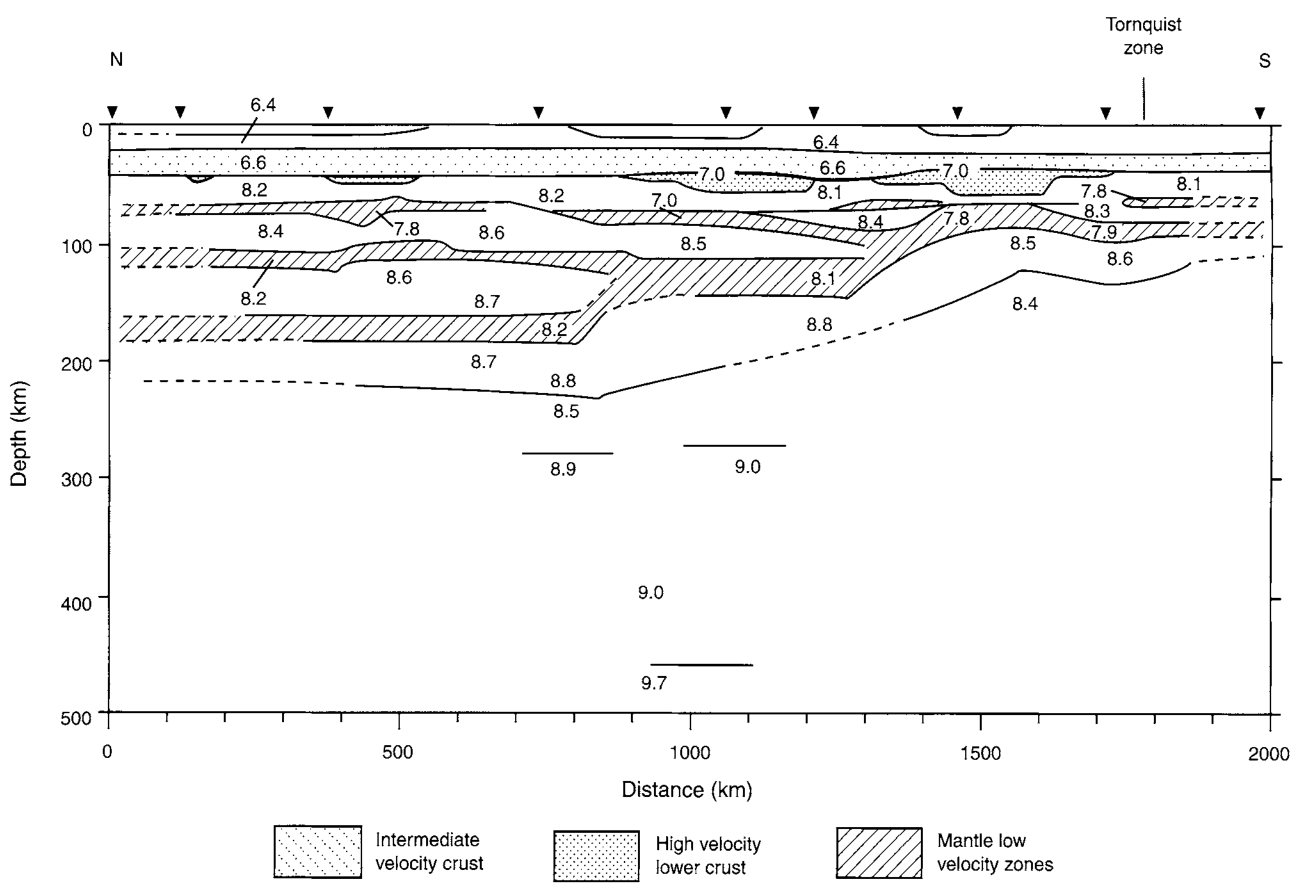

FIGURE 8 (a) Lithospheric cross section through northern Europe (EGT) (after Ansorge *et al.*, 1992, Figs. 3–18). The section traverses Scandinavia from north to south at approximately 20°E. Depth versus distance is 2:1.

The Baltic Shield has almost no sedimentary cover and is characterized by a generally thick crust with rather high *Pg*-velocities near the surface. Crustal thickness is 40–50 km and the lower crust shows rather high velocities (>7 km sec^{-1}). In southern Sweden, several 100 km north of the Tornquist Zone, the crustal thickness decreases to less than 35 km (Fig. 9a), a value that is typical of the Caledonian and Variscan regions of central Europe. Seismic velocities within the crust of the Baltic shield increase gradually from slightly over 6 km sec^{-1} near the surface to 7.0–7.5 km sec^{-1} at 40–50 km depth. It is notable that an intermediate crustal boundary, where the velocity increases by 0.1–0.2 km sec^{-1}, can be followed near 20 km depth throughout the Baltic Shield.

Sub-Moho velocities along the Baltic portion of the EGT are 8.0–8.2 km sec^{-1}, but velocities as high as 8.5 km sec^{-1} are measured on the adjacent Polar Profile (about 50 km further east). Due to very quiet noise conditions and efficient energy propagation (up to 2000 km), Scandinavia is the only place in Western Europe where seismic penetrations to a depth of 400 km by explosion sources has been achieved (Fig. 8). The model derived from upper mantle arrivals beneath Scandinavia shows several distinct velocity inversions between Moho and 200 km depth (Guggisberg and Berthelsen, 1987). Corresponding *S*-wave velocities were derived from surface wave studies that were calculated for all of western Europe and thus cover the whole area of the EGT (Ansorge *et al.*, 1992; Mueller and Panza, 1986)

South of the Tornquist Zone, the crust of the Danish and the North German Basins thins significantly to 26 km. Deducting the voluminous sediments, only 15–20 km of crystalline crust remain. A distinct feature underlying the southern end of the North German Basin is the Elbe line. It separates the lower crust into two regions with 6.9 km sec^{-1} average velocity in the north and 6.4 km sec^{-1} in the south (EUGENO-S Working Group, 1998; Prodehl and Aichroth, 1992).

The internal crustal structure throughout the area of the European Variscides (Fig. 9b) is complex and shows considerable lateral variations. Distinct crustal blocks differing in their internal velocity structure can be assigned to geologically defined terranes of the Variscan orogeny. A subdivision of upper and lower crust by a well-defined boundary (Conrad discontinuity) is often, but not always seen. Towards the Alps the average velocity of the lower crust is as low as 6.2.km sec^{-1}, in contrast to the area north of the Swabian Jura where the velocities above the Moho vary between 6.8 and 7.2 km sec^{-1}. The total crustal thickness under the Variscan part of Germany is fairly constant between 28 and 30 km, except under the Rhine Graben area with 25–26 km and beneath the central part of the Rhenish Massif where an anomalous crustal thickening to 37 km is observed. An important point is that the crust consists of only two layers, in comparison with shield crust that is thicker and has three layers, including a high-velocity (6.9–7.3 km sec^{-1}) lower crustal layer. A joint consideration of seismic velocity structure and petrological information from xenolith studies indicates an intermediate-to-mafic composition for the deeper levels of the crust, at least for the central part of the Variscan crust along the European Geotraverse in Central Europe (Prodehl and Aichroth, 1992). More-recent combined reflection/refraction seismic studies have concentrated on the detailed crustal structure of the Saxothuringium terrane of the Variscan orogen in Saxony, southeast Germany (Enderle *et al.*, 1998; Krawczyk *et al.*, 1999) and on seismic signatures of the Variscan Front in southeast Ireland (Masson *et al.*, 1998, 1999).

The southern segment of the European Geotraverse crosses the Swiss Alps, the Po Plain, the Ligurian Sea and the continental fragments of Corsica and Sardinia, and thus covers units with extremely varying lithospheric structure (Fig. 9b) reflecting the collision of Africa and Europe. In combination with seismic-reflection measurements through Switzerland, the EGT data resulted in a complex and detailed model from the Alps to the Apennines, and demonstrates their asymmetric crustal structure (Valasek *et al.*, 1991, Blundell *et al.*, 1992, Pfiffner *et al.*, 1997). The major element in the upper crust of the central Alps is the relatively low velocity beneath the Aar Massif compared with the well-documented higher velocities in the metamorphic crystalline nappes of the Penninic zone (Pfiffner *et al.*, 1997). From the northern margin of the Alps, the crust steadily thickens southwards to about 60 km underneath the central Alps. The image obtained from combined reflection and refraction data indicates that the European lower crust is a pre-Alpine, possibly Variscan, feature and extends relatively undisturbed from the northern Molasse basin to about 120 km south of the northern margin of the Alps where it is subducted beneath the Adriatic microplate. A fault-like step of the Moho is indicated underneath the Insubric line, which is the tectonic boundary between the central and southern Alps, and as such is possibly the surface expression of the boundary between the European and African crust. Seismic reflection and refraction data indicate that the Adriatic lower crust has been driven northward as a wedge between the European upper and lower crust (Ansorge *et al.*, 1992; Mueller and Kahle, 1993; Valasek *et al.*, 1991; Waldhauser *et al.*, 1998). The complicated arrangement of the various plates in the Southern Alps can best be viewed in enlarged Moho contour maps and 3D block diagrams for this area (Ansorge *et al.*, 1992; Waldhauser *et al.*, 1998).

Egger (1992) provides a detailed discussion of the seismic refraction data of the southern segment of the EGT from northern Apennines into Tunisia. The results support the hypothesis that the opening of the Provencal basin, the rotation of Corsica and Sardinia, and the opening of the Tyrrhenian Sea belong to a system of back-arc spreading, with subduction as a counterpart that is active beneath the Calabrian arc in southern Italy. Complete crustal surveys across the margins and deep basins have complimented the EGT seismic survey to elucidate the nature of the crust at the incipient and ongoing

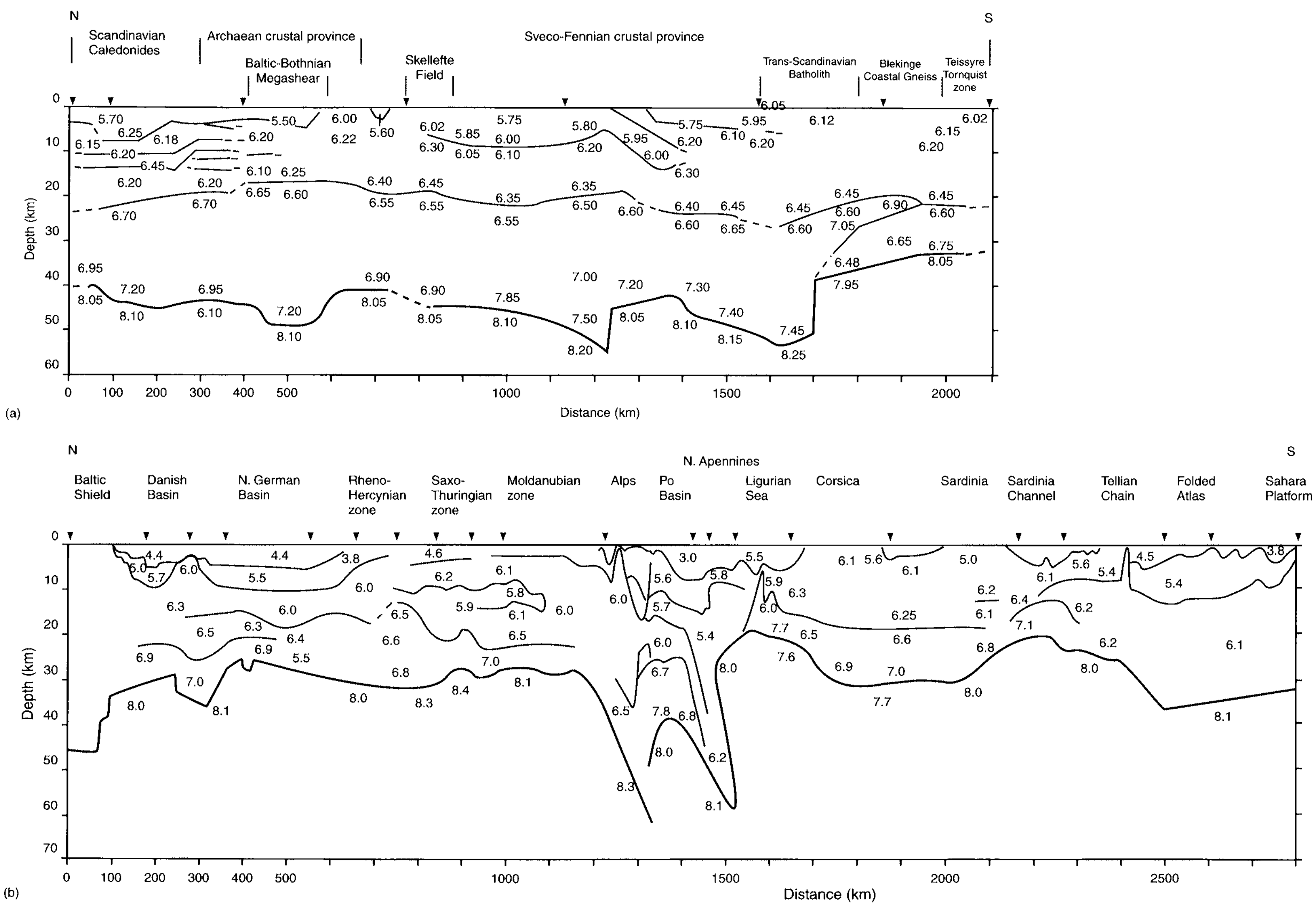

FIGURE 9 Crustal cross section through Europe along the European Geotraverse (EGT) (after Ansorge *et al.*, 1992, Figs. 3–4, 6, 9, 11, 12, 15, 16). Depth versus horizontal distance is 10 : 1. (a) Section through Scandinavia at approximately 20° E, (b) Section through central and southern Europe traversing central Germany, the Alps, the Appenines of northern Italy, Corsica, Sardinia and Tunisia at approximately 10° E.

rifting in the Ligurian Sea and the structure of the Sardinia Channel north of Tunisia. Corsica and Sardinia show a typical, 30-km thick Variscan crust (two layers with average velocities of 6.2 and 6.8 km sec^{-1}), underlain by *Pn*-velocities increasing gradually from 7.6 km sec^{-1} under northern Corsica to 8.0 km sec^{-1} under southern Sardinia and the Sardinia Channel (Fig. 9b). The crust thins to 20 km under the Ligurian Sea and Sardinia Channel.

The EGT reaches African continental crust in Tunisia. The crustal thickness increases steadily from 21 km under the center of the Sardinian Channel to 38 km in the central folded Atlas, from where it decreases southward to 33 km under the southern end of the traverse (Fig. 9b). The velocity of the uppermost mantle is 8.0–8.1 km sec^{-1}. There are thick post-Paleozoic sediments varying in thickness from 8 km in the north to 3 km in the south. A velocity of 6 km sec^{-1} is found everywhere at unusually great depths of 8–14 km. Other striking features are the low average velocity of the entire crust of 6.05 km sec^{-1} and the lack of any clear intracrustal layering (Ansorge *et al.*, 1992; Buness *et al.*, 1992).

Several large-scale seismic investigations have provided further details on the crustal structure of Europe. To the east of the EGT profile, Abramovitz *et al.*, (1999) describe the crustal structure of the North Sea (that portion due north of the Netherlands) and western Ireland. Both regions have thin (30–32 km) crust with three layers. To the east of the EGT profile, the POLONAISE and EUROBRIDGE projects investigated the crust of Precambrian and Variscan Europe, as described by Guterch *et al.* (1999), Grad *et al.* (1999), and the EUROBRIDGE Seismic Working Group (1999). These data document that the crust thickness from 32 km beneath the Paleozoic Variscan Europe to 42 km beneath the Precambrian East European Platform. The Precambrian crust has a very well-defined three-layer structure with seismic velocities in the upper, middle, and lower crust of 6.2, 6.6, and 7.1 km sec^{-1} respectively (Sroda *et al.*, 1999). This is in contrast to the two-layer crust of central Europe, as described above (Fig. 8).

3.4 Northern Eurasia

Seismic refraction profiles cover all of the major geologic structures of northern Eurasia, including Precambrian and younger platforms and shields, deep depressions (e.g., Caspian Sea), orogenic belts (e.g., Urals and Tien Shan Mountains), rifts, and marginal and inner seas. The crustal structure for three long-range profiles is summarized in Figure 10: (a) Black Sea to Dnjeper-Donetz basin, (b) Baltic Shield to West Siberian Plate (passing through the Urals); and (c) West Siberian Plate to Siberian Craton. The first profile was carried out in the 1960s on a closely spaced profile with a 100 m distance between seismometers and chemical explosions at 50–60 km intervals ("continuous profiles"; Pavlenkova, 1973). The long-range profiles from the Baltic Shield across the Urals to the West Siberian Plate and from the West Siberian Plate to the Siberian Craton were recorded with a 10 km interval between seismometers and two types of explosions: chemical and Peaceful Nuclear Explosions (PNEs). The latter profiles provided arrivals from the upper mantle to a depth of 700 km (Egorkin *et al.*, 1987; Beloussov *et al.*, 1991; Mechie *et al.*, 1993). For all of these profiles, the crustal structure can be described by four layers with the following seismic velocities: 2.0–5.5 km sec^{-1} (sediments), 5.8–6.4 km sec^{-1} (upper crust), 6.5–6.7 km sec^{-1} (middle crust) and 6.8–7.4 km sec^{-1} (lower crust).

The first profile (Fig. 10a) crosses the Black Sea basin, Crimean Mountains, Sivash Basin, Ukrainian Shield, and the Dnieper-Donetz Basin (Fig. 3). The Ukrainian Shield has a crustal thickness of more than 40 km and consists of three crustal layers, each with a thickness of 10–15 km. Beneath the basins the Moho is uplifted and the upper crustal layer thins. Such changes in crustal structure become even more dramatic in the Black Sea area where the consolidated crust is only 20 km thick and where the upper and lower crustal layers are absent. The adjacent Crimean Mountains have a deep crustal root, and the crustal average velocity is slightly lower than in the Ukrainian Shield.

The second profile (Fig. 10b) crosses the Baltic Shield, Russian Platform, Timan–Pechora plate, Ural Mountains, and West Siberian Platform. These tectonic units have similar crustal structure, an exception being the Timan–Pechora plate. This plate is characterized by considerably lower average crustal velocity, because a lower-crustal high-velocity layer ($Vp > 7$ km sec^{-1}) is absent. The Urals have a pronounced crustal root, whereas the Timan ridge is compensated by an increased thickness of the upper crust. Below the crust, strong reflections are recorded from a velocity discontinuity (referred to in Russian literature as the "N discontinuity") at depths of 70–90 km. The relief on this boundary appears to be opposite to that of the Moho. The seismic velocity between the Moho and the N discontinuity varies from 8.0 beneath the Pechora and West Siberian plates to 8.4 beneath the Urals. These changes clearly correlate with measured heat flow, with the lower velocities corresponding to higher heat flow areas. An anomalous high-velocity zone was revealed near the boundary between the old Russian plate and the younger Timan–Pechora plate. The dotted line in Figure 10b traces this zone from a high-velocity intrusion in the crust to the north where it is subducted beneath the Pechora Plate.

The third profile (Fig. 10c) crosses the West Siberian plate and the Siberian Platform. Though of differing age, these platforms have similar crustal thickness. Lateral variations in crustal structure are mainly noted beneath deep sedimentary basins. For example, beneath the West Siberian basin and the Tunguss basin the thickness of the middle crustal layer is nearly constant, whereas the thickness of the lower crust increases. The Vilui basin, which is younger than the Tunguss basin, has a clear Moho uplift. The upper mantle velocities hardly change along this profile. The only characteristic feature is the region with lower velocity (8.0 km sec^{-1}) in the

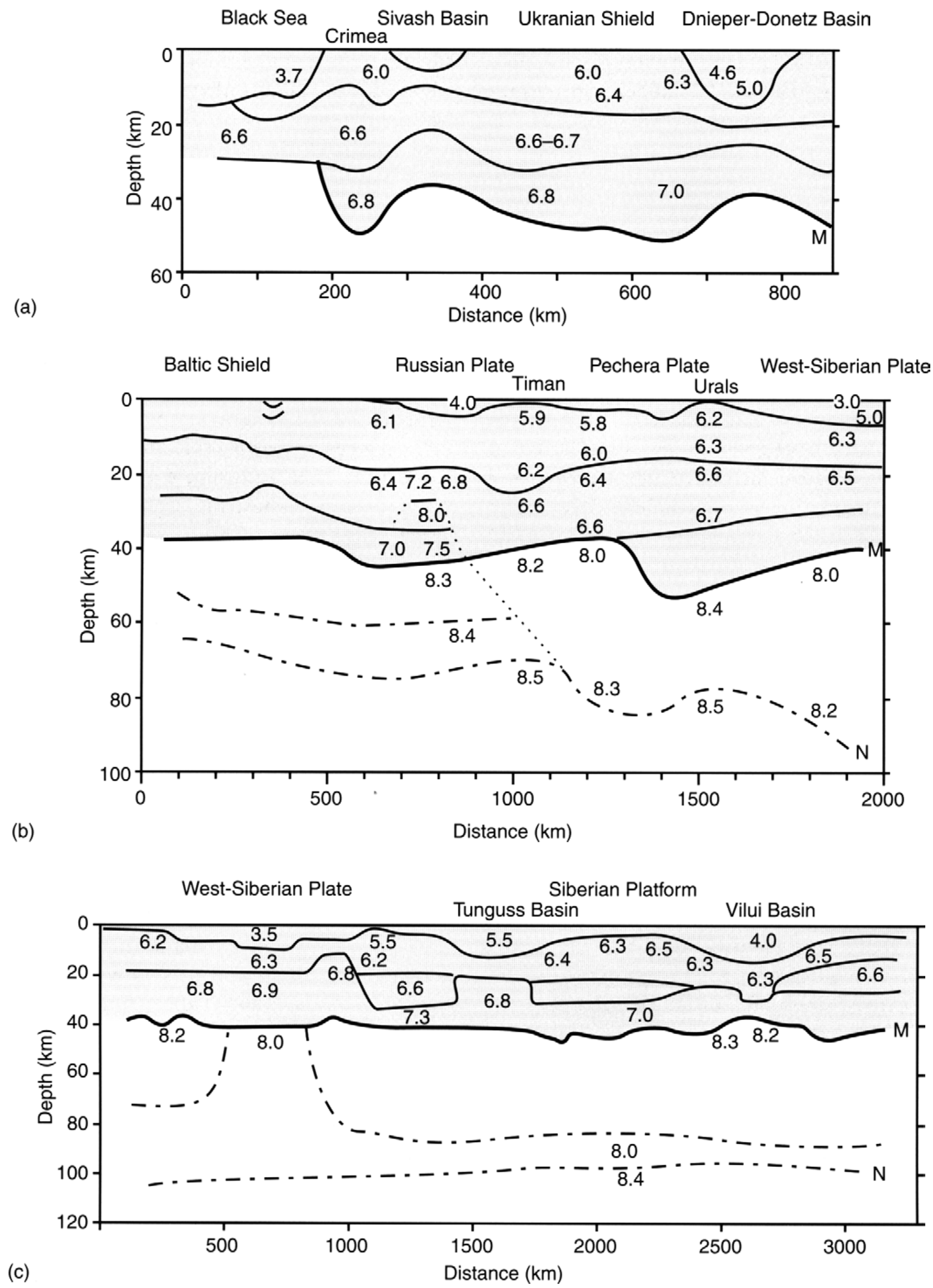

FIGURE 10 Crustal cross sections through Eurasia (Egorkin *et al.*, 1987; Beloussev *et al.*, 1991). (a) Black Sea–Dnjeper-Donetz basin. The section traverses southeastern Russia in a SSW–NNE direction from about 45 to 53° N and 34 to 40° E. Depth versus horizontal distance is 5 : 1. (b) Baltic Shield – Urals – West-Siberian Plate. The section traverses northwestern Russia in a WNW–ESE direction from about 69 to 62° N and 30 to 72° E. A seismic discontinuity ("N") within the uppermost mantle is often reported beneath northern Eurasia (e.g., Pavlenkova, 1996) but has not been well determined elsewhere. Depth versus horizontal distance is 6 : 1. (c) West-Siberian Plate–Siberian Craton. The section traverses northeastern Russia in a W–E direction from about 65 to 62° N and 66 to 132° E. Depth versus horizontal distance is 10 : 1.

central West Siberian plate. This location corresponds to a prominent rift zone that crosses the plate from north to south. Below the crust, an important feature is an upper mantle low-velocity zone ($8.0\,\mathrm{km\,sec^{-1}}$) immediately above the N boundary, which is at a depth of 100 km.

3.5 Near East and Africa

The Near East and Africa are largely underlain by Precambrian cratons, with, in comparison to other continents, a very small fraction of the crust being of younger age. The younger crust occupies the Cape, Mauritanide, and Atlas fold belts, located at the south, northwest, and north margins, respectively. The geologic evolution of the African continent began in the Early Archean Eon, 3.2–3.65 Ga with the formation of the Kaapvaal craton of southern Africa. The entire landmass was assembled and stabilized by 0.6 Ga. The main tectonic events since the end of the Precambrian have been: (1) the Mesozoic and Cenozoic rifting at the west margin during the opening of the south Atlantic Ocean; and (2) rifting in East Africa and the Red Sea in the late Cenozoic.

As is evident in Figure 4, large portions of Africa have not been investigated by seismic refraction profiles. Major crustal seismic research has been carried out in the Afro-Arabian rift system and its immediate surroundings of eastern Africa (Prodehl *et al.*, 1997). The crustal and uppermost-mantle structure has been investigated by seismic refraction surveys in the Jordan–Dead Sea rift, the Red Sea, the Afar depression and the East African rift of Kenya (Fig. 11). With the exception of the Jordan–Dead Sea transform, the entire Afro-Arabian rift system is underlain by anomalous mantle with Pn-velocities less than $8\,\mathrm{km\,sec^{-1}}$, while under the rift flanks the Pn-velocity is clearly equal to or above $8.0\,\mathrm{km\,sec^{-1}}$. Various styles of rifting have been found. Oceanic crust floors the axial trough of the southern Red Sea rift, thinned continental crust underlies the margins of the Red Sea, Afar depression, and northern Kenya rift. In contrast, 30–35 km thick continental crust is found both under the Jordan–Dead Sea rift, an area of strike-slip rifting, and under the central Kenya rift, where updoming driven by a buoyant mantle is apparently the controlling feature. The transition from thinned continental crust to 5–6 km thick oceanic crust in the center of the Red Sea appears to be gradual. In contrast, the transition from the thin crust of the East Africa rift to undisturbed continental crust of 40 ± 5 km thickness is mostly rather abrupt. The seismic data indicate various stages of rifting, and imply the presence of hot uppermost mantle under most parts of the rift system, possibly related to plume activity. Volcanism may disrupt and/or underplate the crust in places, altering particularly the lower crust.

Portions of northern Africa have been investigated with seismic profiles. Measurements in Tunisia were completed as part of the European Geotraverse of 1982–1985, and are summarized in Section 2.3 and Figure 9. In addition, a geological and geophysical transect was completed through the High Atlas Mountains in Morocco. These mountains were formed over a major intracontinental rift system that had extended from what is now the Atlantic margin of Morocco to the Mediterranean coast of Tunisia (~2000 km) during the convergence of the African and European plates.

Crustal thickness variations in the region highlight one of the more remarkable aspects of this orogen. Despite topography that is locally in excess of 4 km, there does not seem to be crust that is thicker than 40 km. Beauchamp *et al.* (1999) report thin crust (18–20 km) beneath the Moroccan shelf and a thickness of 30 km south of the High Atlas Mountains. The crust thins to 24 km toward the Atlantic margin, and attains an average crustal thickness of 35 km to the north of the mountains. The crust is 39 km thick beneath the mountains. There is also evidence for a low-velocity zone at depths of 10–15 km beneath the High and Middle Atlas Mountains. This may be due to a crustal detachment located at the base of the Paleozoic (Beauchamp *et al.*, 1999).

3.6 Australia, New Zealand, and Antarctica

Australia is a stable continent that is largely isolated from the immediate effects of plate tectonics. Within its interior it contains no active rift systems, and only at its offshore northern domain does an active convergent margin exist. The central and western portions of the Australian continent are very ancient. Its western Archean cratons (Pilbara and Yilgarn) have yielded some of the oldest geochronologic ages (>4.0 Ga) determined so far, and the midcontinent is composed of Proterozoic crust. The eastern third of the continent and the Tasmanides, are Paleozoic and Mesozoic in age (Drummond, 1991).

Since the early 1950s, a considerable body of data has accumulated on the crustal structure of the Australian continent. Cleary (1973), Finlayson and Mathur (1984), Drummond and Collins (1986) and Collins (1988, 1991) summarize this work. These active-source results cluster at specific areas (Fig. 12). However, a recent study based on receiver functions has made possible the determination of a comprehensive Moho map (Clitheroe *et al.*, 1999).

The crust is thickest beneath the Proterozoic North Australian Craton (45–50 km), Central Australia (>50 km), and under the Paleozoic Lachlan Fold Belt (40–50 km). It is shallowest under the Archean Pilbara and Yilgarn Cratons of Western Australia (28–37 km), and possibly Tasmania (25–35 km). Pn-velocities are generally greater than $8\,\mathrm{km\,sec^{-1}}$. They are around $8.2\,\mathrm{km\,sec^{-1}}$ and higher in central Australia, slightly lower under the thinner western Australian Pilbara and Yilgarn cratons, and seem to be lowest under eastern Australia along the Great Eastern Divide and the Tasman Geosyncline.

New Zealand lies across the Australian/Pacific plate boundary that transects the South Island. Davey *et al.* (1998)

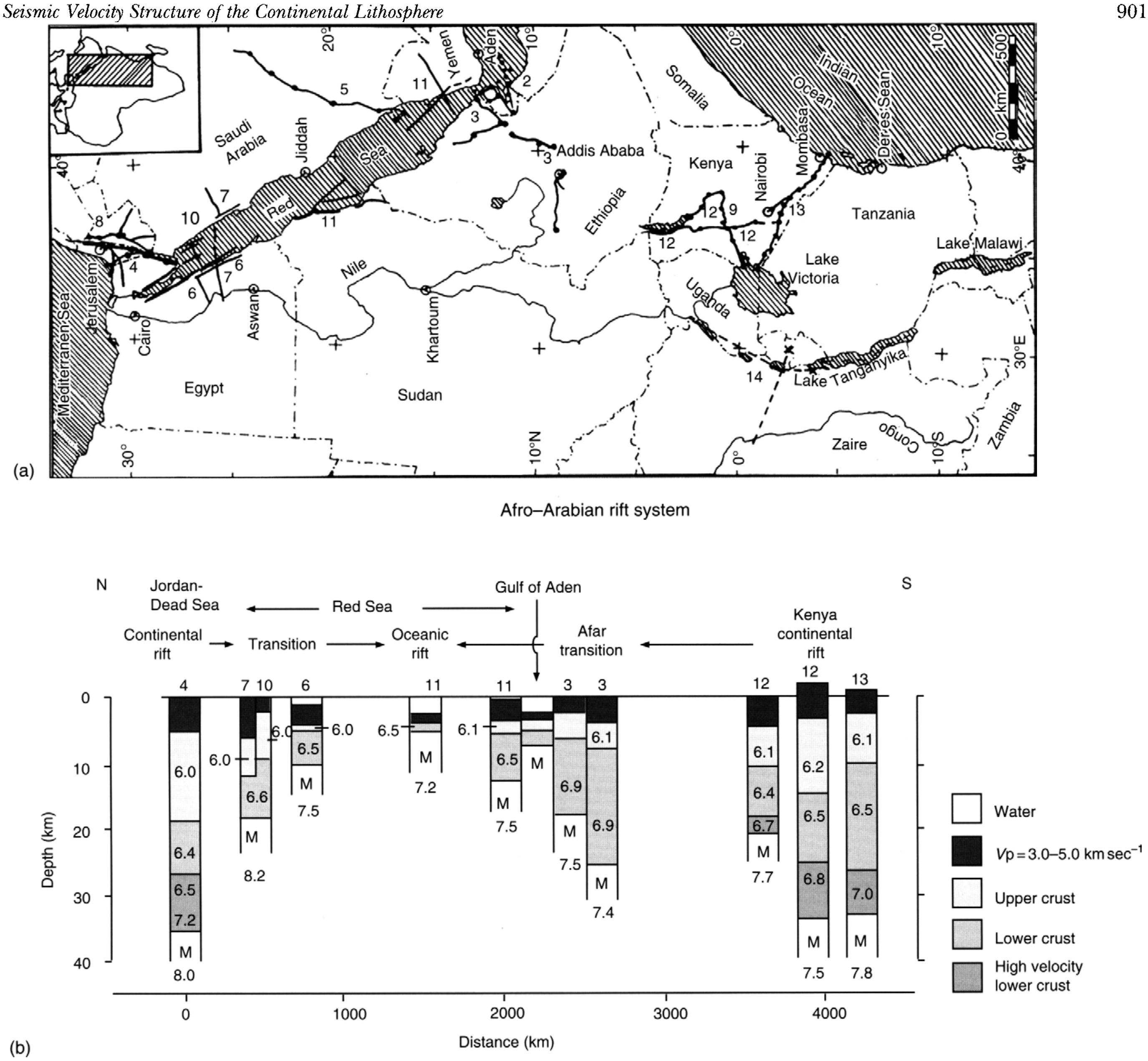

FIGURE 11 (a) Location of long-range seismic lines in the Afro-Arabian rift system (Prodehl *et al.*, 1997). Continuous lines: seismic-refraction surveys, full circles mark shotpoints where locations were published. Dashed lines: approximate lines through epicenters of local earthquakes, crosses: IRSAC (Institut pour la Recherche Scientifique en Afrique Centrale) network at Zaire. (b) Velocity–depth columns through the Afro-Arabian rift system. The evolutionary sequence [for locations see Fig. (a)] illustrating the variation in crustal thickness under the Afro-Arabian rift system from the Jordan–Dead Sea transform system through the Red Sea, Gulf of Aden and Afar triangle to the southern end of the Kenya rift. M, Moho. Depths refer to sea level. Numbers of crustal columns refer to (a) (Prodehl *et al.*, 1997).

and Stern and McBride (1998) present the crustal structure of the South Island based on recently recorded seismic refraction and near-vertical reflection data (Fig. 13). These results indicate that the crust has been formed by the accumulation of low-grade metasedimentary rocks in a convergent margin setting. Modeling of both onshore and offshore data indicate a crust of fairly constant *P*-wave velocity (5.9–6.2 km sec^{-1}) overlying a 5–10 km-thick lower crust with seismic velocity of about 7 km sec^{-1}. The crustal structure (Fig. 13) is remarkable for the unusual thickness of low-velocity materials that extend along the Alpine fault zone. The total crustal thickness varies from about 30 km at the east coast to about 42 km under the western mountains, the Southern Alps.

Antarctica is divided by the Trans-antarctic Mountains into East Antarctica, a Precambrian craton, and West Antarctica, which has experienced significant Cenozoic tectonic activity, including crustal extension. Relatively few measurements of

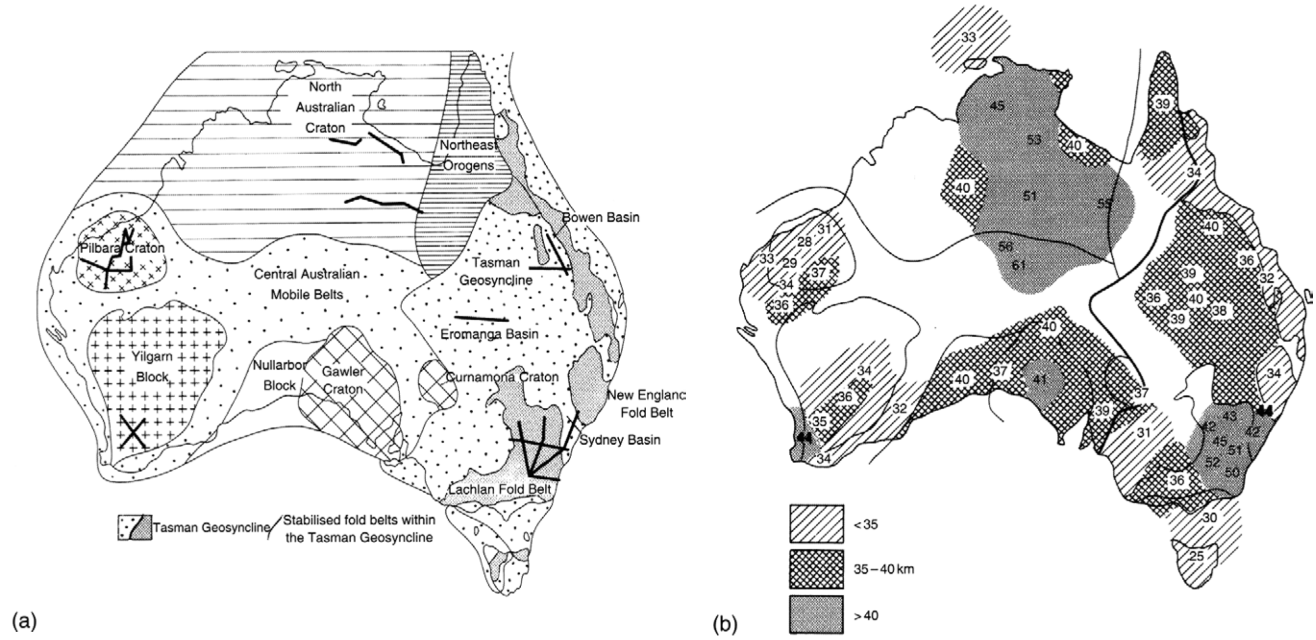

FIGURE 12 Crustal structure of Australia (Collins, 1991): (a) Location of major geological provinces and selected seismic profiles (solid lines); (b) Moho depths in Australia; patterns show regions with similar values, numbers indicate point values. Western Australia is composed of Precambrian crust, including some of the most ancient cratons on Earth (Archean Pilbara and Yilgarn blocks). These Archean cratons have a crustal thickness of 28–37 km, which is less than the global continental average of 40 km.

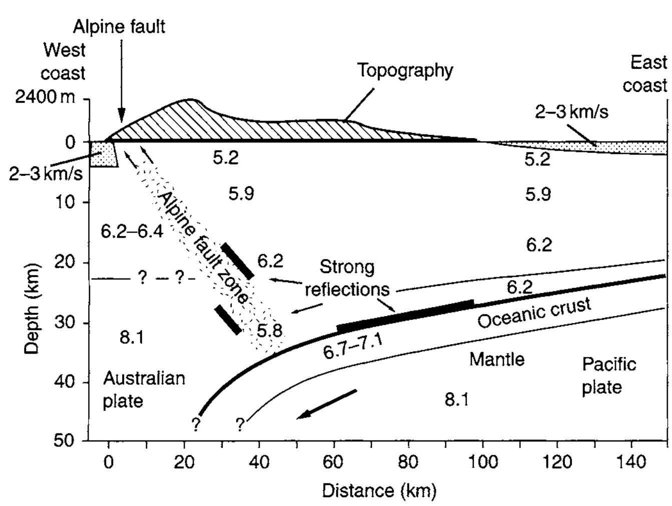

FIGURE 13 Seismic section across the South Island of New Zealand (Stern and McBride, 1998) showing western-thickening crust and east-dipping Alpine fault zone. The crustal structure is unusual in showing low seismic velocities (5.2–6.2 km sec^{-1}) throughout the crust. This indicates that the entire crust, above the hypothesized oceanic crustal layer (6.7–7.1 km sec^{-1}), is composed of low-grade metasedimentary rocks.

crustal structure have been reported for the continental crust of Antarctica. The top of basement in western Antarctica lies at least 2 km below sea level and drops an additional 2 km beneath the deeper parts of the Byrd Subglacial Basin. In the east, the basement is found very close to sea level, and the overlying section is typically no more than a few hundred meters thick. Seismic velocities in the upper crust range from 5.7 to 6.4 km sec^{-1} (Bentley, 1991).

The best information on deeper crustal structure comes from seismic refraction profiles. In the east, two profiles in Maud Land show a depth to Moho of 38–40 km inland, which decreases to about 30 km near the coast. On one of the profiles, a pronounced intracrustal (Conrad?) discontinuity was found at a depth of 18 km between layers with seismic velocities 6.0–6.2 km sec^{-1} above and 6.7–6.9 km sec^{-1} below. The apparent *P*-wave velocity in the upper crust ranges from 6.0 to 6.4 km sec^{-1} and appears to gradually increase with depth (Bentley, 1991).

In the Amery ice shelf, Bentley (1991) describes a layered continental crust with thickness ranging from 30 to 40 km. The upper crustal layer is about 20 km thick. The seismic velocity seen in this section of upper crust is somewhat higher than expected, and may be due to the great age of the crystalline rocks in the basement.

In western Antarctica, a deep sounding in front of the Weddel Sea found the Moho at 33–36 km depth, but it shallows to about 30 km beneath the Crary Trough, then quickly drops to 40 km beneath Coats Land. The lower crustal layer is about 10 km thick. In the east, however, 15 km from the shoreline, the crust–mantle boundary is found to lie 25 km below sea level. From this point, the Moho depth increases to 28 km toward the west (beneath Ross island), and to 35 km eastward (at the shoreline). There is only one additional unpublished seismic profile available for western Antarctica, and this profile shows Moho depth at 24 km below sea level.

In summary, the mean thickness of the crust in East Antarctica is approximately 40 km compared to 30 km in West Antarctica, and the boundary between the two is abrupt. There is a precipitous drop in topography at the Trans-Antarctic Mountains, and geophysical evidence also implies that this change in crustal thickness is sudden.

3.7 Japan and Southeast Asia

The opening of the Sea of Japan in the early Miocene rifted Japan from the Asiatic margin (Finn *et al.*, 1994). Present-day Japan has a composite structure composed of Mesozoic metamorphic and plutonic rocks, and Cenozoic volcanic and plutonic rocks associated with active subduction and magmatism. The seismic structure of Japan has been investigated for many years (e.g., Matsuzawa, 1959; Asano *et al.*, 1979), and abundant seismicity within the crust and upper mantle has permitted detailed modeling of lithospheric structure (Zhao *et al.*, 1990). Iwasaki *et al.* (1994) provide a detailed interpretation of the *P*- and *S*-wave structure of the crust of northern Honshu (Fig. 14). The crust is about 33 km thick, and shows a nearly uniform increase in velocity with depth from 6.0 km sec^{-1} (beneath a 1-km thick layer of sediments and/or volcanic rocks) to 7.0 km sec^{-1} at the base of the crust. The seismic velocity of the uppermost mantle is unusually low, 7.5–7.6 km sec^{-1}, a value that is found only in regions of very high heat flow and/or active volcanism, such as the Afro-Arabian rift system (Fig. 11).

Indonesia has had a complex geologic history, and the physical properties of the crust are largely undetermined. Most likely the area consists of a continental or borderland type of crust of intermediate (30 km) thickness. It has been proposed that cratonization of mobile shelves occurs by sedimentary and calcalkaline accretion, following in the wake of island arc migration by backarc extension (Curray *et al.*, 1977).

The Gulf of Papua has been the focus of study by a number of researchers. Drummond *et al.* (1979) found sediments to be 5 km thick on the Papuan Plateau, but 10 km to the west and northwest along the axes of the Moresby and Aure Troughs. Offshore, sedimentary basins as thick as 11 km have been measured. Beneath these sediments, a layer with *P*-wave velocity of 6.1 km sec^{-1} was inferred over the region, but was thought to be underlain to the south by a layer with a velocity of 6.9 km sec^{-1}. This second, faster, layer is also thought to be present beneath the Eastern Plateau. This lower-crustal layer is noticeably absent from the Aure Trough in the north, though velocity increases are expected with depth.

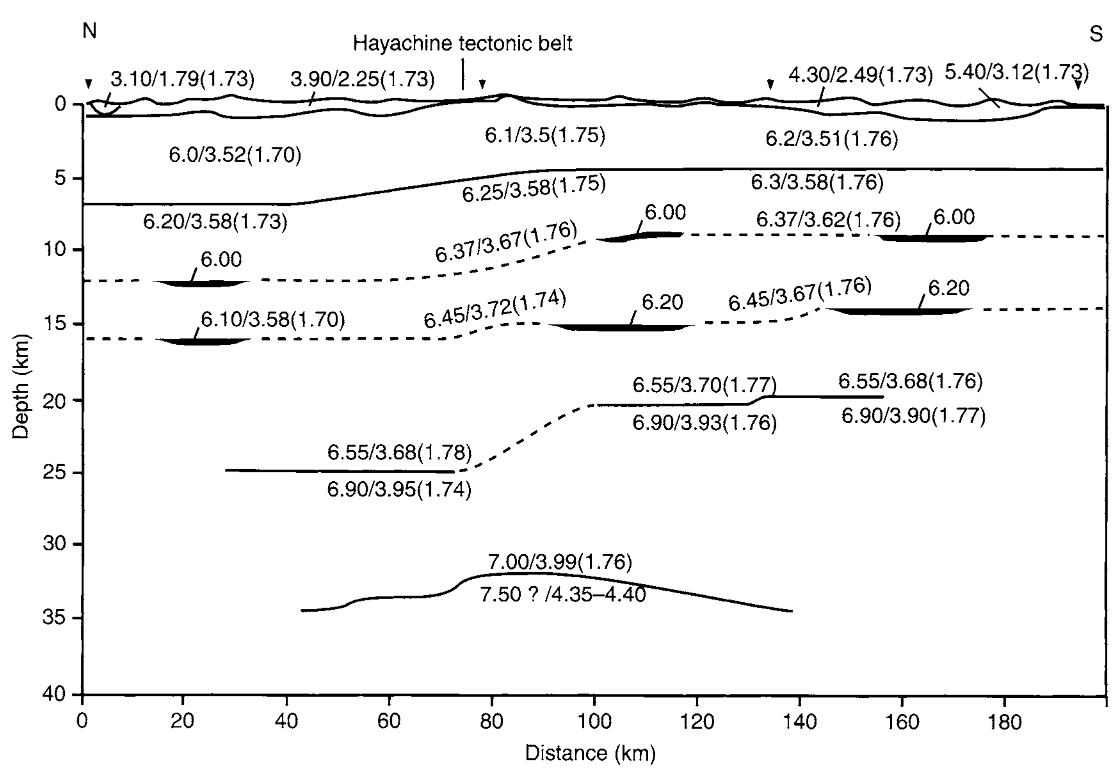

FIGURE 14 Crustal *P*-wave and *S*-wave velocity structure through a typical island arc: the Kitakami region, northern Honshu, Japan. Indicated are *P*-velocity/*S*-velocity, and their ratio. Modified from Iwasaki *et al.* (1994).

The Moho is 27–29 km deep along the southwestern coast of the peninsula and shallows beneath the Moresby Trough to 19 km. This depth is greater than normal for the Moho beneath oceanic crust, but it is not unreasonable when compared to other measurements in the area. Curray *et al.* (1977) found the Moho to be at 18 km depth south of the Island of Bali and 21–25 km south of the Island of Java.

Beneath the Eastern Plateau, the Moho again deepens to 25 km (Drummond *et al.*, 1979). With a lower crustal layer of 9 km in thickness, the crust beneath the Papuan Peninsula is

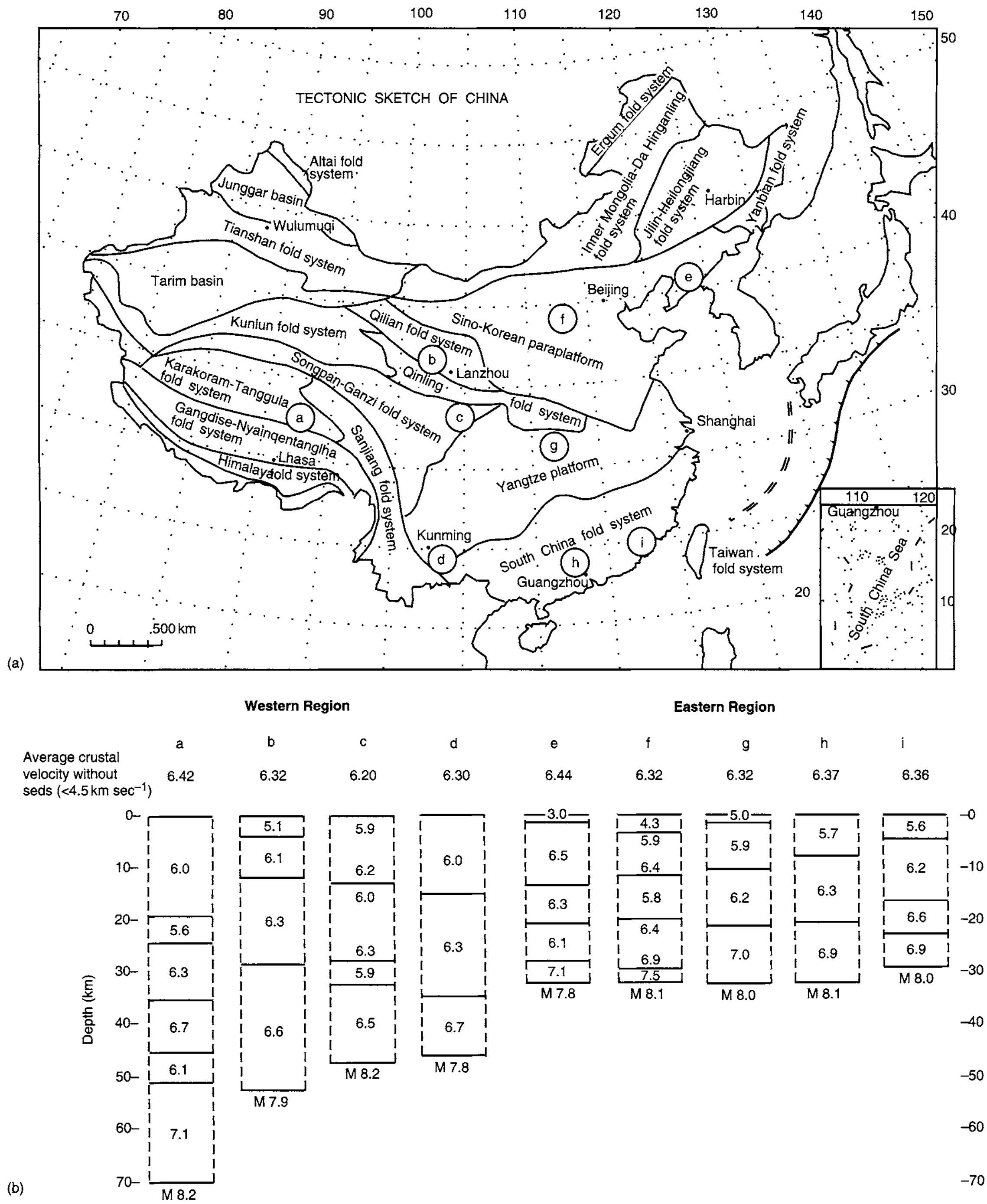

FIGURE 15 Crustal structure of China (Li and Mooney, 1998): (a) Tectonic sketch map of China with location of seismic crustal surveys; (b) Representative seismic velocity–depth functions for nine regions of China. The crust thickens from about 35 km in the east to greater than 70 km beneath the Tibetan Plateau to the west.

therefore continental. Offshore, the crust in the Moresby Trough is about 19 km thick, with sediments comprising about the top 10 km. Therefore, the crust below the sediments has a thickness typical of oceanic crust. The northern extent of the thin crust beneath the Moresby Trough is uncertain. Similar measurements of crustal thickness to those of Drummond *et al.* (1979) were made in Papua by Finlayson *et al.* (1976), who detected a low-velocity zone below the Moho and a return to normal upper-mantle velocities at a depth of about 50 km.

3.8 China and India

China has undergone a long geologic evolution, from the formation of the Archean Sino-Korean platform in the north to the active continent–continent collision in Tibet. More than 36 000 km of active-source seismic refraction profiles (referred to as Deep Seismic Sounding, DSS, profiles, in the Russian, Chinese, and Indian literature) have been collected in China since 1958. The most remarkable aspect of China's crustal structure is the >70 km thick crust of the Tibetan Plateau (Fig. 15). All of western China is underlain by crust that is 45–70 km thick and is separated from 30–45 km thick crust of Eastern China by a seismic belt which trends roughly north–south at 105° E longitude. The crustal structure of western China (Fig. 15b, columns a–d) show thick crust that is characteristic of young orogenic belts worldwide. Three of the four columns differ from the thick crust of shields in that they lack a high-velocity (7.0–7.5 km sec^{-1}) lower crustal layer (cf. Christensen and Mooney, 1995). A high velocity (7.1 km sec^{-1}) lower crustal layer has been reported in western China beneath the southernmost portion of the Tibetan Plateau. Several investigators have identified this layer as the cold lower crust of the subducting Indian plate (cf. Li and Mooney, 1998).

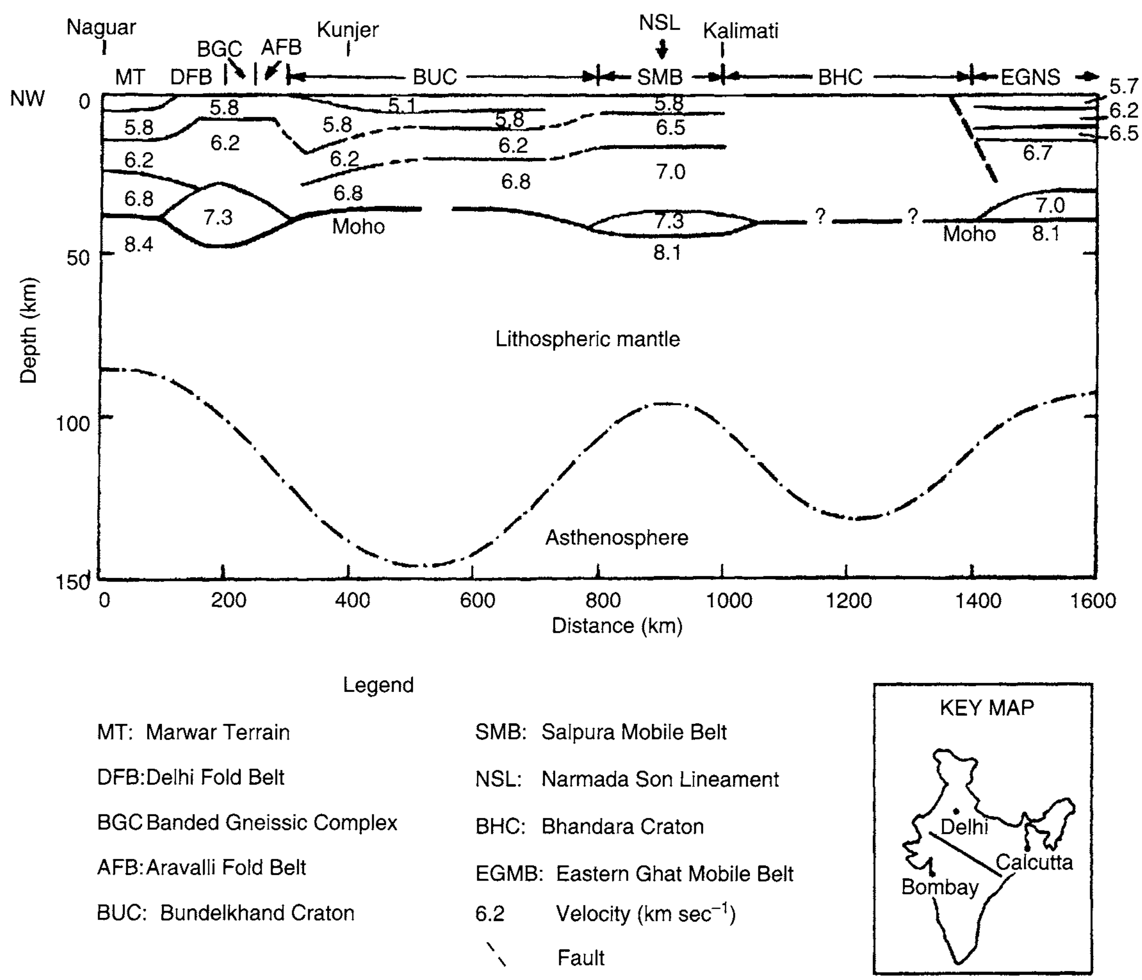

FIGURE 16 Lithospheric cross-section of the Indian subcontinent compiled from sources listed in the text. Inset shows the location of the cross section. The crustal structure is highly variable laterally, but the crustal thickness, where measured, is relatively uniform at about 40 km, equal to the global average for Precambrian cratonic crust. Estimated lithospheric thickness ranges from 80 to 150 km, and is greatest beneath the Bundelkhand (BUC) and the Bhandara (BHC) cratons.

Recently recorded near-vertical and wide-angle reflection data indicate a low-velocity, perhaps fluid-rich or partially molten zone in the middle crust of the Tibetean Plateau (Zhao *et al.*, 1993; Nelson *et al.*, 1996; Makovsky *et al.*, 1996a,b).

The crustal columns for eastern China (Fig. 15b, columns e–i) all indicate relatively thin (29–33 km) crust, close to the continental global average for extended crust (30 km), but significantly thinner than the global average of 39 km (Christensen and Mooney, 1995).

The average crustal velocity of China (6.15–6.45 km sec^{-1}, Fig. 15b) indicates a felsic to intermediate bulk crustal composition. Upper mantle *Pn* velocities (about 8.0 km sec^{-1}) are equal to the continental global average. These results have been interpreted in terms of the most recent thermo-tectonic events that have modified the crust (Li and Mooney, 1998). In eastern China, Cenozoic extension has created a thin crust with low average crustal velocities while in western China, Mesozoic and Cenozoic arc–continent and continent–continent collisions have led to crustal growth and thickening.

India consists of Archean cratons and associated Archean to Proterozoic high-grade mobile belts, Proterozoic sedimentary basins, horst and graben belts associated with Gondwana sedimentation, Deccan volcanism, and the Himalayas. The investigation of the crustal structure of India by seismic refraction profiling was begun in 1972 as an Indo-Soviet cooperation. Since then, more than 20 long-range refraction, near-vertical reflection and wide-angle reflection profiles have been carried out. Results are reviewed by Kaila (1982), Kaila and Krishna (1992), Mahadevan (1994), Reddy (1999) and Reddy and Rao (2000). Estimates of crustal thickness are in the 35–40 km range, with the exception of the Himalayas that are known to have thicknesses of about 70 km (Mahadevan, 1994). A lithospheric cross section compiled from several seismic profiles illustrates the main properties of the seismic structure of the crust and uppermost mantle of India (Fig. 16).

4. Summary Discussion

4.1 Histograms of Crustal Thickness and Average Crust Velocity

We have divided the crust into six geologic provinces (Fig. 3) and present histograms for two basic properties: crustal thickness (Fig. 17) and mean crustal velocity (Fig. 18; Mooney *et al.*, 1998). These histograms reinforce many of the points made earlier regarding the relationship between the seismic structure of the crust and geologic setting. Shields and platforms have an average crustal thickness of about 40 km, with a standard deviation (SD) of 7 km. Shields have the highest average *P* velocity of any geologic province, 6.49 km sec^{-1} (SD 0.23 km sec^{-1}). Orogenic crust has a wide range of thickness of 20–80 km, and an average thickness of 43 km (SD 10 km). Isostatic forces uplift crust thicker than about 50 km, leading to erosion and crustal thinning. Most thick (>50 km) crust is young orogenic crust that is undergoing active tectonic compression. The average *P* velocity of orogenic crust is 0.13 km sec^{-1} lower than shield crust. This may be due to thickening of low-velocity upper crustal layers during compression, as seen in the European Alps, intrusion by silicic magma, such as are seen in the Andes, or by the delamination of the high-velocity lower crust, as found beneath the southern Sierra Nevada, California.

Extended crust refers to crust that has experienced localized rifting and/or regional extension. Examples include the Basin and Range province of the western USA (Fig. 6) and much of Western Europe. The average thickness of extended crust is about 30 km, and the mean *P*-wave velocity (6.16 km sec^{-1}) is 0.33 km sec^{-1} lower than shield crust. Forearc crust refers to those regions that were formed in front (oceanward) of a volcanic arc, such as many parts of the west coast of North America. Only 26 seismic refraction measurements are presently available for forearcs. These regions have an average crustal thickness of 27 km (SD 8 km) and a very low mean crustal velocity (6.09 km sec^{-1}). These values reflect the abundant sedimentary accumulations in forearc crust. Magmatic (volcanic) arcs, such as Japan (Fig. 14) and the Aleutian Islands, Alaska, have an average crustal thickness of 31 km (SD 8 km) and a mean *P*-wave velocity of 6.14 km sec^{-1} (SD 0.23 km sec^{-1}).

4.2 General Characteristics of Continental Crustal Structure

The continental lithosphere shows several important regularities. (1) Crustal thickness and average velocities increase from continental margins toward the interior; (2) The crust of old, stable shields and platforms is 35–45 km thick, whereas young, nonorogenic crust is significantly thinner (25–35 km); (3) An inverse correlation is clearly evident between the depths to the basement and to the M boundary: the crust is thicker under orogenic belts, and thinner under basins; (4) An average petrological model may be presented for stable continental interiors in terms of three layers. The two upper layers have silicic-to-intermediate composition, but the middle crust is composed of rocks with a higher degree of metamorphism. The third layer is composed of mafic rocks of granulite-grade metamorphism. The composition of the continental crust is discussed in detail by Rudnick and Fountain (1995) and Christensen and Mooney (1995); (5) The upper mantle, including the subcrustal lithosphere and possibly the asthenosphere, is characterized with fine-scale seismic and rheological stratification. Rheologically weak layers appear in the middle and lower crust, and the crust/mantle boundary. In northern Eurasia, a thin low-velocity zone is reported at a depth of 80–100 km based by a clear "N" reflector

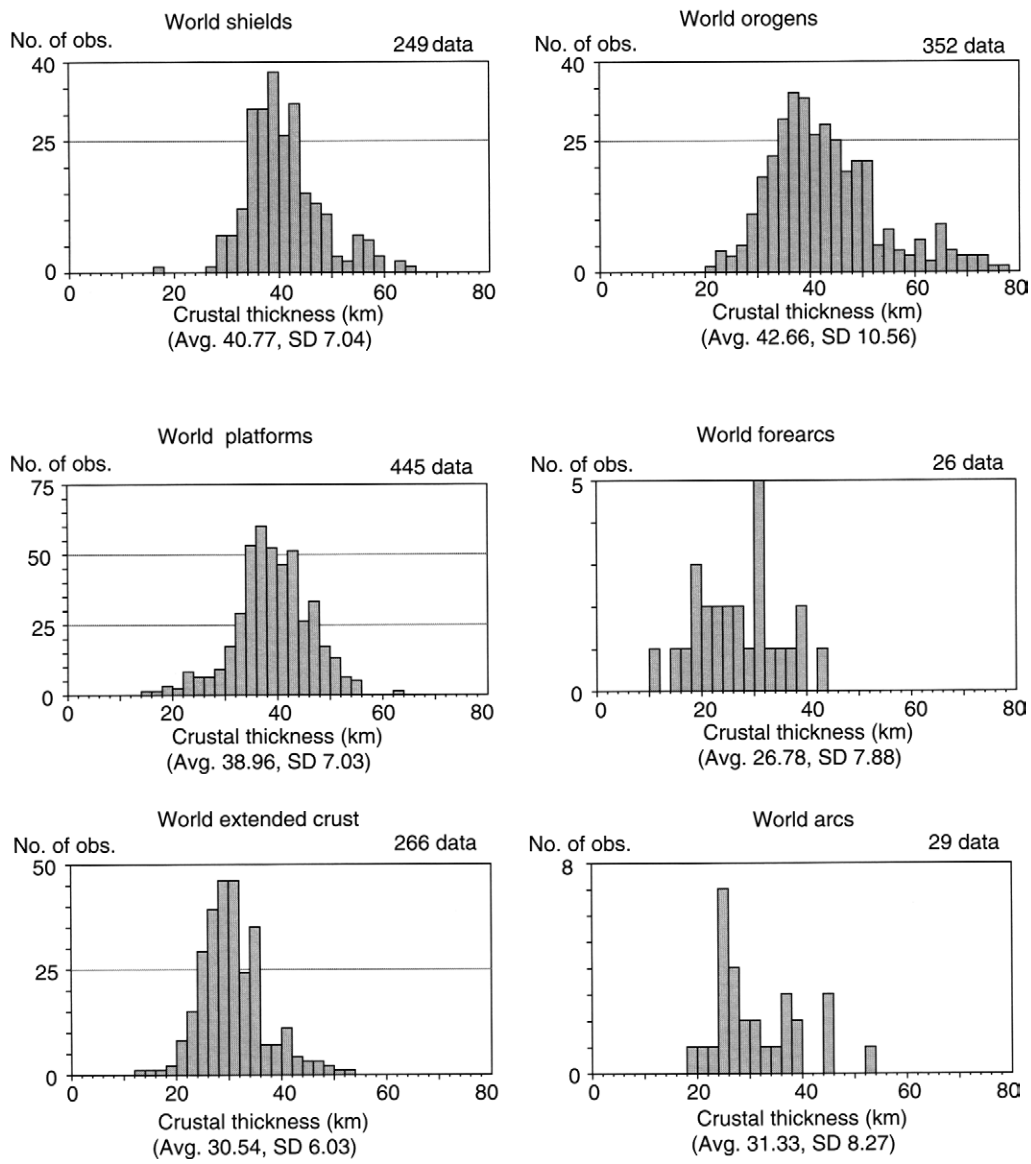

FIGURE 17 Histograms of crustal thickness for six continental tectonic provinces calculated from the individual point measurements (triangles) of Figure 4 (Mooney *et al.*, 1998). Average and SD are indicated. These histograms indicate systematic differences among tectonic provinces, and provide a basis for extrapolating crustal thickness into unmeasured regions.

(Pavlenkova, 1973, 1996; Zverev and Kosminskaja, 1980; Beloussov and Pavlenkova, 1984; Kozlovsky, 1987; Beloussov *et al.*, 1991; Krilov *et al.*, 1991, Puzyrev, 1993); (6) The thickness of the crust usually increases under orogenic belts. The depth of crustal roots varies in different kinds of orogens. The thickest crust is under young mountains, such as those situated on Alpine geosynclines (Caucasus, Pamir). In contrast, orogens developed on ancient platforms have smaller roots (e.g., the northern Tian-Shan and the Altai Mountains). There are also orogens that lack roots, such as the Canadian Cordillera (Clowes *et al.*, 1999), and that are held up by low density, buoyant mantle.

In most cases there is an inverse correlation between sediment thickness and depth to the Moho. For example, crustal thickness decreases under platform depressions, the basins of the inner and marginal seas, and under rifts. There is an inverse correlation between crustal thickness and heat flow. This correlation is observed both in active regions and on stable platforms.

As noted above, stable continental crust is 40 ± 7 km thick with three main crustal layers, each 10–15 km thick. This crustal type covers the interior of Eurasia, North America, Africa, and Australia. Thin (25–35 km) two-layer continental crust is observed on the marginal parts of Eurasia, in Western Europe, in the Pechora Plate of north central Russia, in the Far

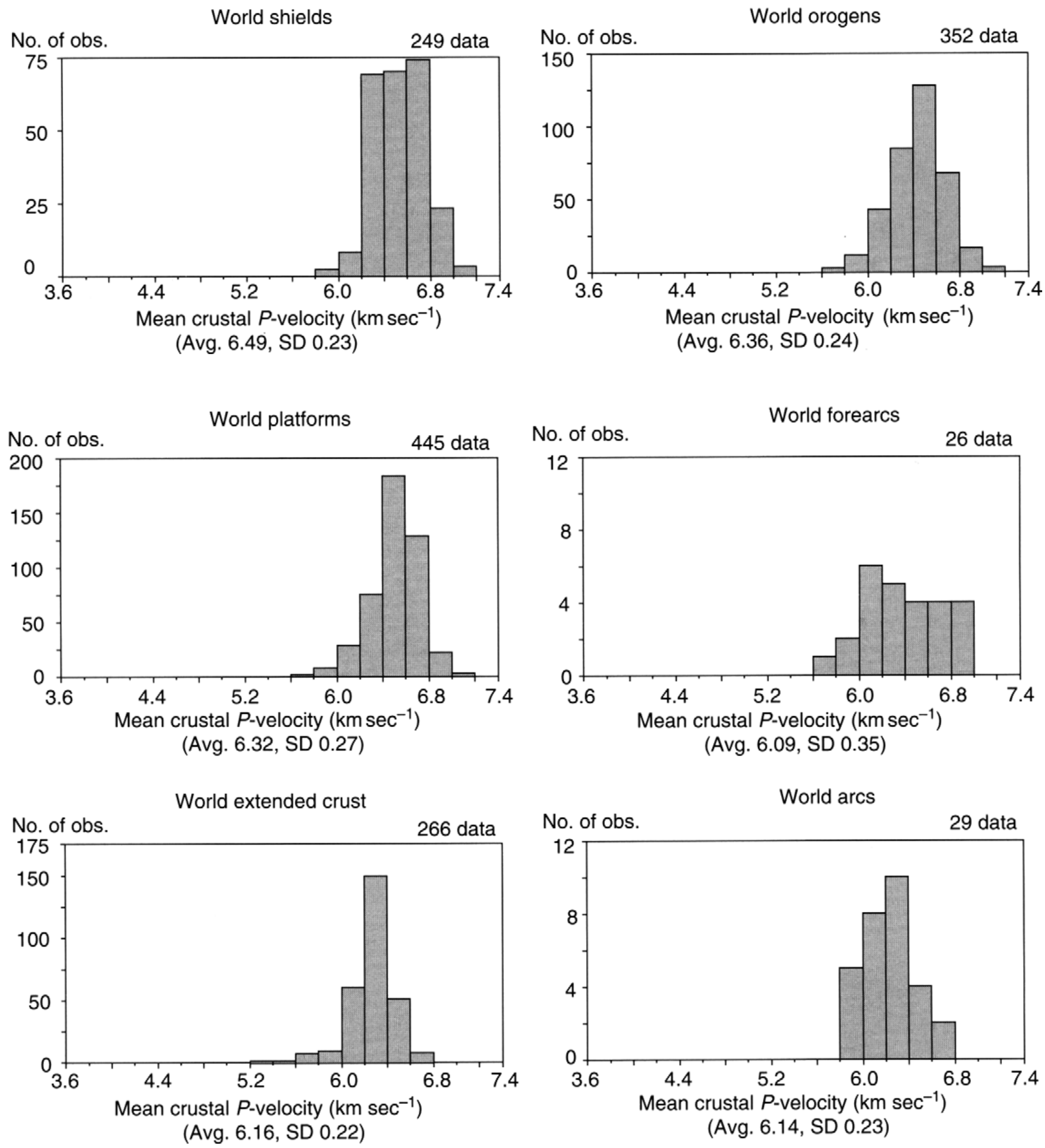

FIGURE 18 Histograms of mean crustal velocity for six continental tectonic provinces calculated from the individual point measurements (triangles) of Figure 4 (Mooney *et al.*, 1998). Average and SD are indicated. These histograms indicate systematic differences among tectonic provinces, and provide a basis for extrapolating mean crustal velocity into unmeasured regions.

East, Eastern China, and the Basin and Range Province, western United States of America. It differs from the stable crust not only in thickness but also by virtue of a lower mean crustal velocity (6.16 km sec^{-1} versus 6.49 km sec^{-1}). An important characteristic feature of this crustal type is the absence of the high-velocity layer (7 km sec^{-1}) in the lower crust.

A consideration of more than 1000 seismic refraction profiles worldwide lends strong support to the definition of the seismic Moho as that depth where seismic velocity first exceeds 7.6 km sec^{-1}. The Moho is nearly always evident as a refraction horizon, and can be clearly distinguished from the lower crust. The uppermost mantle, including the lithosphere and asthenosphere, is sometimes characterized by fine stratification; high velocities (up to 8.4+ km sec^{-1}) alternate with lower ones (7.8–8.0 km sec^{-1}). There is a close correlation between the properties of the crust and upper mantle; stable shield and platform crust exists above cold, high-velocity uppermost mantle, and thin, extended or young orogenic crust exists above low velocity regions. This correlation indicates that mantle processes play an important role in crustal evolution, such as providing a driving force for crustal extension.

This brief review of the seismic velocity structure of the continental lithosphere highlights only the general features of its structure. The focus has been on the structure of the upper portion of the lithosphere, the continental crust. A great deal of new, high-quality data are recorded each year, and these show that the lithosphere to be highly complex and subject to multiple processes, including compression accompanied by brittle faulting and dutile deformation, extension, displacement by strike-slip faulting, and modification by igneous processes. Controlled source seismic data have been important in developing an understanding of the structure, composition, and evolution of the lithosphere, but future progress will depend on investigations that combine geophysical, geologic, and geochemical constraints.

Acknowledgments

It is a pleasure to acknowledge our colleagues at the US Geological Survey, numerous US academic institutions, the Lawrence Livermore and Los Alamos National Laboratories, the Universities in Karlsruhe (Germany), Paris, and Strasbourg, the Institute of the Physics of the Earth, and the Russian Ministry of Natural Resources, and Russian Academy of Sciences. Review comments by R. Meissner, H. Kanamori, P.R. Reddy, I. Artemieva and an anonymous reviewer are appreciated. Special thanks to W.H.K. Lee for his encouragement, and to K. Favret (USGS) for making the figures. S. Detweiler (USGS) assisted with preparation of the final manuscript.

References

Abramovitz, T., *et al.*, (1999). *Tectonophysics* **314**, 69.
Ansorge, J., *et al.* (1992). In: "A Continent Revealed – the European Geotraverse," (D. Blundell *et al.*, Eds.) Cambridge University Press p. 33.
Artemieva, I.M. and W.D. Mooney (2001). *J. Geophys. Res.* **106**, 16387–16414.
Asano, S., *et al.* (1979). *J. Phys. Earth*, **27**, supplement, S1.
Bamford, D. and C. Prodehl (1977). *J. Geol. Soc. London* **134**, 139.
Barazangi, M. and L. Brown (Ed.) (1986a). *Am. Geophys. Un. Geodynam. Series 13*, Washington, DC.
Barazangi, M. and L. Brown (Ed.) (1986b). *Am. Geophys. Un. Geodynam. Series 14*, Washington, DC.
Beauchamp, W., *et al.* (1999). *Tectonics* **18**, 163.
Beloussov, V.V. and N.I. Pavlenkova (1984). *J. Geodynam.* **1**, 167.
Beloussov, V.V. *et al.* (Ed.) (1991). "Deep Structure of the USSR Territory." Nauka, Moscow (in Russian).
Bentley, C.R. (1991). In: "The Geology of Antarctica," (R.J. Tingey, Ed.), pp. 335. Oxford University Press.
Birch, F. (1960). *J. Geophys. Res.* **65**, 1083.
Birch, F. (1961). *J. Geophys. Res.* **66**, 2199.
Blundell, D., *et al.* (Ed.) (1992). "A Continent Revealed – the European Geotraverse," Cambridge University Press.
Braile, L.W., *et al.* (1995). In: "Continental Rifts," (K.H. Olsen, Ed.), pp. 61. Elsevier, Amsterdam.
Brown, L.D., *et al.* (1983). *Geol. Soc. Am. Bull.* **94**, 1173.
Buness, H., *et al.* (1992). *Tectonophysics* **207**, 245.
Christensen, N.I. (1965). *J. Geophys. Res.* **70**, 6147.
Christensen, N.I. (1966). *J. Geophys. Res.* **71**, 3549.
Christensen, N.I. (1979). *J. Geophys. Res.* **84**, 6849.
Christensen, N.I. and W.D. Mooney (1995). *J. Geophys. Res.* **100**, 9761.
Chuikova, N.A., *et al.* (1996). *Izv. Phys. Solid Earth* **34**, 701.
Chulick, G.S. and W.D. Mooney (2002). *Bull. Seism. Soc. Am.*, **92**, in press.
Cleary, J. (1973). *Tectonophysics* **20**, 241.
Clitheroe, G.M., *et al.* (1999). *J. Geophys. Res.* **105**, 13697.
Closs, H. and C. Behnke (1961). *Geol. Rundschau* **51**, 315.
Clowes, R., *et al.* (1999). *Episodes* **22**, 3.
Collins, C.D.N. (1988). *Australian Bureau of Mineral Resources, Geol. Geophys. Report 277.*
Collins, C.D.N. (1991). *Geol. Surv. Aust.* **17**, 67–80.
Curray, J.R., *et al.* (1977). *J. Geophys. Res.* **82**, 2479.
Davey, F.J., *et al.* (1998). *Tectonophysics* **288**, 221.
Drummond, B. (Ed.) (1991). *Geol. Soc. Aust. Special Publ. 17.*
Drummond, B. and C. Collins (1986). *Earth Planet. Sci. Lett.* **79**, 361.
Drummond, B., *et al.* (1979). *J. Aust. Geol. Geophys.* **4**, 341.
Durrheim, R.J. and W.D. Mooney (1994). *J. Geophys. Res.* **99**, 15359.
Egorkin, A.V., *et al.* (1987). *Tectonophysics* **140**, 29.
Egger, A.P. (1992). PhD Thesis, ETH Zurich.
Ekström G., *et al.* (1997). *J. Geophys. Res.* **102**, 8137.
Enderle, U., *et al.* (1998). *Geophys. J. Int.* **133**, 245.
EUGENO-S Working Group (1988). *Tectonophysics* **150**, 253.
EUROBRIDGE Seismic Working Group (1999). *Tectonophysics* **314**, 193.
Finlayson, D. and S.P. Mathur (1984). *Ann. Geophys.* **2**, 711.
Finlayson, D. *et al.* (1976). *Geophys. J. R. Astron. Soc.* **44**, 45.
Finn, C., *et al.* (Ed.) (1994). *J. Geophys. Res.* **99**, 22135.
Giese, P., *et al.* (Ed.) (1976). "Explosion Seismology in Central Europe." Springer, Berlin.
Giese, P., *et al.* (1999). *South Am. J. Geosci.* **12**, 201–220.
Goodwin, A.M. (1991). "Precambrian Geology," Academic Press, San Diego.
Goodwin, A.M. (1996). "Principles of Precambrian Geology," Academic Press, San Diego.
Grad, M., *et al.* (1999). *Tectonophysics* **314**, 145.
Guggisberg, B. and A. Berthelsen (1987). *Terra Cognita* **7**, 631.
Guterch, A., *et al.* (1999). *Tectonophysics* **314**, 101.
Hart, P.J. (Ed.) (1969). *Geophys. Monograph 13*, American Geophysics Union Washington, DC.
Heacock, J.G. (Ed.) (1971). *Geophys. Monograph 14*, American Geophysics Union Washington, DC.
Heacock, J.G. (Ed.) (1977). *Geophys. Monograph 20*, American Geophysics Union Washington, DC.
Hersey, J., *et al.* (1952). *Bull. Seismol. Soc. Am.* **42**, 291.
Hill, M.N. (Ed.) (1963). "The Sea," vol. 3, Interscience Publishers, New York.
Holbrook, W.S., *et al.* (1992). In: "Continental Lower Crust," (D.M. Fountain *et al.*, Eds.), pp. 1. Elsevier, Amsterdam.
Iwasaki, T., *et al.* (1994). *J. Geophys. Res.* **99**, 22187.
Iyer, H.M. and T. Hitchcock (1989). *Geol. Soc. Am. Memoir 172.*
James, D.E. (1971). *J. Geophys. Res.* **76**, 3246.
James, D.E. and J.S. Steinhart (1966). *Am. Geophys. Un. Geophys. Monograph* **10**, 293.

Jarchow, C.M. and G.A. Thompson (1989). *Annu. Rev. Earth Planet. Sci.* **17**, 475.
Kaila, K.L. (1982). *Geophys. Res. Bull.* **20**, 309–328.
Kaila, K.L. and V.G. Krishna (1992). *Curr. Sci.* **62**.
Kern, H. (1978). *Tectonophysics* **44**, 185.
Klemperer, S.L. and W.D. Mooney (Ed.) (1998a), *Tectonophysics* **286**, 1.
Klemperer, S.L. and W.D. Mooney (Ed.) (1998b). *Tectonophysics* **288**, 1.
Kozlovsky, Ye.A. (Ed.) (1987). "The Superdeep Well of the Kola Peninsula," Springer-Verlag, Berlin.
Krawczyk, C.M., *et al.* (1999). *Tectonophysics* **314**, 241–253.
Krilov, S.V. *et al.* (1991). *Geodynamics* **13**, 87.
Levander, A.R. and K. Holliger (1992). *J. Geophys. Res.* **97**, 8797.
Li, S.L. and W.D. Mooney (1998). *Tectonophysics* **288**, 105.
Ludwig, W.J., *et al.* (1970). In: "The Sea," (A.E. Maxwell, Ed.), vol. 4, part I, pp. 53. Wiley-Interscience, New York.
Mahadevan, T.M. (1994). *Geol. Soc. India Memoir 28*, Bangalore.
Makovsky, Y., *et al.* (1996a). *Tectonics* **15**, 997.
Makovsky, Y., *et al.* (1996b). *Science* **274**, 1690.
Makris, J. (1975). In: "Afar Depression of Ethiopia," (A. Pilger and Rösler, Eds.), pp. 379. Schweizerbart, Stuttgart.
Masson, F., *et al.* (1998). *Geophys. J. Int.* **134**, 689.
Masson, F., *et al.* (1999). *Tectonophysics* **302**, 83–98.
Matsuzawa, T. (1959). *Bull. Earthq. Res. Inst. Univ. Tokyo* **37**, 123.
Matthews, D.H. (1986). *Geol. Soc. London Special Publ.* **24**, 11.
Maxwell, A.E. (Ed.) (1970). "The Sea – New Concepts of Sea Floor Evolution," vol 4, Wiley-Interscience, New York.
McConnell, R.K., *et al.* (1966). *Rev. Geophys.* **4**, 41–100.
Mechie, J., *et al.* (1993). *Phys. Earth Planet. Inter.* **79**, 269.
Meissner, R. (1973). *Geophys. Surv.* **1**, 195–216.
Meissner, R. (1986). "The Continental Crust," Academic Press, London.
Meissner, R., and the DEKORP Research Group, 1991. The DEKORP surveys: Major achievements for tectonical and reflective styles. In: Meissner, R., Brown, L., Dürbaum, H.-J., Franke, W., Fuchs, K., and Seifert, F. (Eds.), Continental Lithosphere: Deep Seismic Reflections. *Am. Geophys. Un., Geodyn. Ser.*, **22**, 69–76.
Meyer, R.P., *et al.* (1976). *Am. Geophys. Un. Geophys. Monograph* **19**, 105.
Mohorovicic, A. (1910). *Jahrbuch Meteorol. Obs. Zagreb für 1909* **9**(4), 1.
Mooney, W.D. (1989). *Geol. Soc. Am. Mem.* **172**, 11.
Mooney, W.D. and L.W. Braile (1989). In: "The Geology of North America – An Overview." (A.W. Bally and A.R. Palmer, Eds.), pp. 39. Geological Society of America, Boulder, CO.
Mooney, W.D., *et al.* (1998). *J. Geophys. Res.* **103**, 727.
Mooney, W.D. and R. Meissner (1992). In: "Continental Lower Crust," (D.M. Fountain *et al.*, Eds.), pp. 45. Elsevier, Amsterdam.
Mueller, S. (Ed.) (1973). *Tectonophysics* **20**, 1.
Mueller, S. (1977). *Am. Geophys. Un. Geophys. Monograph* **20**, 289.
Mueller, S. (1978). In: "Tectonics and Geophysics of Continental Rifts," (I. Ramberg and E. Nuemann, Eds.), vol. 2, pp. 11. Reidel, Dordrecht.
Mueller, S. and H.-G. Kahle (1993). *Am. Geophys. Un. Geodynam. Series* **23**, 249.
Mueller, S. and G.F. Panza (1986). *Dev. Geotecton.* **21**, 93.
Myers, S.C., *et al.* (1998). *J. Geophys. Res.* **103**, 21233.
Nelson K.D. *et al.* (1996). *Science* **274**, 1684.
Olsen, K.H. (Ed.) (1995). "Continental Rifts." Elsevier, Amsterdam.
Orcutt, J. (1987). "Scattering and Q within the Anza Array: A Final technical report," University of California, San Diego.
Pakiser, L.C. and W.D. Mooney (Ed.) (1989). "Geophysical Framework of the Continental United States," *Geol. Soc. Am. Memoir 172*.
Pakiser, L.C. and J.S. Steinhart (1964). In: "Research in Geophysics," (J. Odishaw, Ed.), 2, pp. 123. M.I.T. Press, Cambridge, MA.
Patzwahl, R., *et al.* (1999). *J. Geophys. Res.* **104**, 7293.
Pavlenkova, N.I. (1973). "Wave Fields and Crustal Models of Continental Type." Naukova Dumka, Kiev, (in Russian).
Pavlenkova, N.I. (1996). *Adv. Geophys.* **37**, 1.
Pfiffner, O.A., *et al.* (Ed.) (1997). "Deep Structure of the Swiss Alps: Results of NRP 20." Birkhäuser, Basel.
Pollack, H.N. and D.S. Chapman (1977). *Tectonophysics* **38**, 279.
Prodehl, C. (1977). *Am. Geophys. Un. Geophys. Monograph* **20**.
Prodehl, C. (1984). In: "Physical Properties of the Interior of the Earth, the Moon and the Planets," (K. Fuchs and H. Soffel Eds.), pp. 97. Springer, Berlin.
Prodehl, C. and B. Aichroth (1992). *Terra Nova* **4**, 14.
Prodehl, C., *et al.* (1997). *Tectonophysics* **278**, 1.
Puzyrev, N.N. (Ed.) (1993). "Detailed Seismic Studies of the Lithosphere with P- and S-Waves," Nauka, Novosibirsk, (in Russian).
Reddy, P.R. (1999). *Curr. Sci.* **77**, 1606.
Reddy, P.R. and V. Vijaya Rao (2000). *Curr. Sci.* **78**, 899.
Ritzwoller, M.H., *et al.* (1998). *Geophys. J. Int.* **134**, 315.
Rudnick, R. and D.M. Fountain (1995). *Rev. Geophys.* **33**, 267.
Simmons, G. (1964). *J. Geophys. Res.* **69**, 1123.
Soller, D.R., R.D. Ray, and R.D. Brown (1982). *Tectonics* **1**, 125.
Sroda, P. and the POLONAISE Profile P3 Working Group (1999). *Tectonophysics* **314**, 175.
Steinhart, J.S. (1967). In: "International Dictionary of Geophysics," (K. Runcorn, Ed.),vol. 2, pp. 991.
Steinhart, J.S. and R.P. Meyer (1961). *Carnegie Inst. Washington, DC, Publ. 622*.
Stern, T.A. and J.H. McBride (1998). *Tectonophysics* **286**, 63.
Tanimoto, T. (1995). In: "Global Earth Physics: A Handbook of Physical Constants," (T. J. Ahrens, Ed.), pp. 214. American Geophysics Union, Washington, DC.
Tatel, H.E. and M.A. Tuve (1955). *Geol. Soc. Am. Special Paper* **62**, 35.
Valasek, P., *et al.* (1991). *Geophys. J. Int.* **105**, 85.
Waldhauser, F., *et al.* (1998). *Geophys. J. Int.* **135**, 264.
Wang, Y.X., *et al.* (2002). *J. Geophys. Res.* (in Press).
Wigger, P.J., *et al.* (1994). In: "Tectonics of the Southern Central Andes," (K.J. Reutterr, *et al.*, Eds.), pp. 23. Springer, New York.
Zhao, D., *et al.* (1990). *Tectonophysics* **181**, 135.
Zhao, W.J., *et al.* (1993). *Nature* **366**, 557.
Zverev, S.M. and I.P. Kosminskaya (Ed.) (1980). "Seismic Models of the Lithosphere for the Major Geostructures on the Territory of the USSR," Nauka, Moscow (in Russian).

Editor's Note

Due to space limitations, references with full citation are given in MooneyFullReferences.pdf on the attached Handbook CD, under the directory \54Mooney. See also Chapter 55, Seismic structure of the oceanic crust and passive continental margins, by Minshull.