Geological Society of America Memoirs

CHAPTER 5 The 1950s (1950–1960)

Geological Society of America Memoirs 2012;208;71-111
doi: 10.1130/2012.2208(05)

Notes

CHAPTER 5

The 1950s (1950–1960)

5.1. SEISMIC-REFRACTION INVESTIGATIONS IN WESTERN EUROPE IN THE 1950s

Since the beginning of the 1950s, commercial quarry blasts have been increasingly used since they proved to be a powerful and low-cost energy source for a systematic investigation of the detailed structure of the Earth's crust (Reinhardt, 1954). Also, it was recognized early on, that underwater explosions are much more effective than other blasts due to better damming of the charge by water (Förtsch and Schulze, 1948; Schulze, 1974). The crustal investigations involved both the seismic-refraction and reflection methods. In 1950, Reich (1953) observed for the first time clear reflections from the Mohorovičić-discontinuity in southern Germany using a quarry blast.

Reinhardt (1954) also gave a detailed description how seismic measurements for crustal studies had to be organized and which technical requirements had to be fulfilled at that time, for example, the recording of the zero-time of an explosion, the organization of a central time signal, and the spacing of recording stations as well as minimum distances depending on the goal of an experiment. He also discussed the various seismometers available at that time.

The leading generation of post-war geophysicists in Germany had a very clear primary focus: to generate a climate of cooperation between the many small Earth science institutes or departments in Germany and to return to international cooperation. Heinz Menzel at Clausthal and Otto Rosenbach at Mainz belonged to the young driving forces in this cooperation of Earth sciences, including geophysics institutions in Germany. They had both studied mathematics and physics at Königsberg University before and during the war. After the war, they had begun their professional career with PRAKLA at Hannover in the exploration industry. In seismology and deep-seismic sounding, the other driving forces were located in Stuttgart with Wilhelm Hiller and the even younger generation with international contacts to Lamont: Hans Berckhemer and Stephan Mueller. In Munich, Hermann Reich attracted in particular scientists such as Otto Förtsch, a pioneer in the early deep-seismic sounding studies in Germany, and the young Peter Giese with a strong geological background.

In 1957, systematic seismic crustal research started in Germany when a priority program ("Geophysical Investigation of Crustal Structure in Central Europe") was initiated, funded by the German Research Society and involving all geophysical university institutes and geophysical departments of the German state geological surveys. At the beginning of these investigations of crustal structure, however, it was necessary to develop new recording systems with higher magnification than the mechanical-optical ones used to record the Heligoland and Haslach explosions in the late 1940s. Also, more flexibility in field operations was required. In Germany, two systems were constructed and applied, each in connection with highly sensitive electromagnetic and electrodynamic seismographs with a natural frequency of 1–2 cps. The one system was based on low-sensitive galvanometers operating with electronic amplifiers, of which various types were built in the laboratories of the individual institutions. The other system used only high-sensitive galvanometers of the Kipp and Zonen type. Both systems reached a maximum amplification of 10^6 in the frequency range of 5 to 20 cps (Giese et al., 1976a).

Geophysical companies in Germany showed particular interest in the scientific deep-seismic sounding programs and supported these efforts by recording up to 12 s two-way traveltime (TWT) (e.g., Dohr, 1957, 1959; Schulz, 1957; Liebscher, 1962, 1964). Due to their much higher frequency contents, these reflection seismic data offered completely new information than did the seismic-refraction surveys. In 1952, reflections from great depths were reported by Reich (1953) on recordings of quarry blasts in southern Germany by PRAKLA with commercial reflection instrumentation.

Schulz (1957) interpreted deep reflections at 4.0 and 5.4 s TWT as being caused by reflectors at 10.4 and 13.4 km depth, respectively. Liebscher (1962, 1964) prepared histograms, i.e., the number of reflections per time interval of 0.2 seconds, to derive the depth of the main crustal boundaries (e.g., the Förtsch, the Conrad, and the Mohorovičić-discontinuities; see Fig. 6.2.2-04) and published contour maps for the C- and M-boundaries in southern Germany (see Fig. 6.2.2-03).

In 1954, the Alps became a special target of crustal research following the founding of the Subcommission of Alpine Explosions under the umbrella of the International Union of Geodesy and Geophysics (IUGG). Under the leadership of Y. Labrouste (University of Paris) and H. Closs (BGR Hannover), a series of larger explosions was arranged in two lakes of the French Alps: in Lac Rond des Rochilles between Briançon and Modane in 1956, and in Lac Nègre north of Monaco in 1958 (Closs and Labrouste, 1963). Numerous British, French, German, Italian, and Swiss institutions cooperated in the observation of these explosions and set out recording stations on several profiles and fans throughout the Western Alps (Groupe d'Etudes des Explosions Alpines, 1963) which, due to topography, resulted in a more or less areal coverage by seismic stations (Fig. 5.1-01, see also Fig. 6.2.4-01). With the specific goal of studying the zone of Ivrea, which is

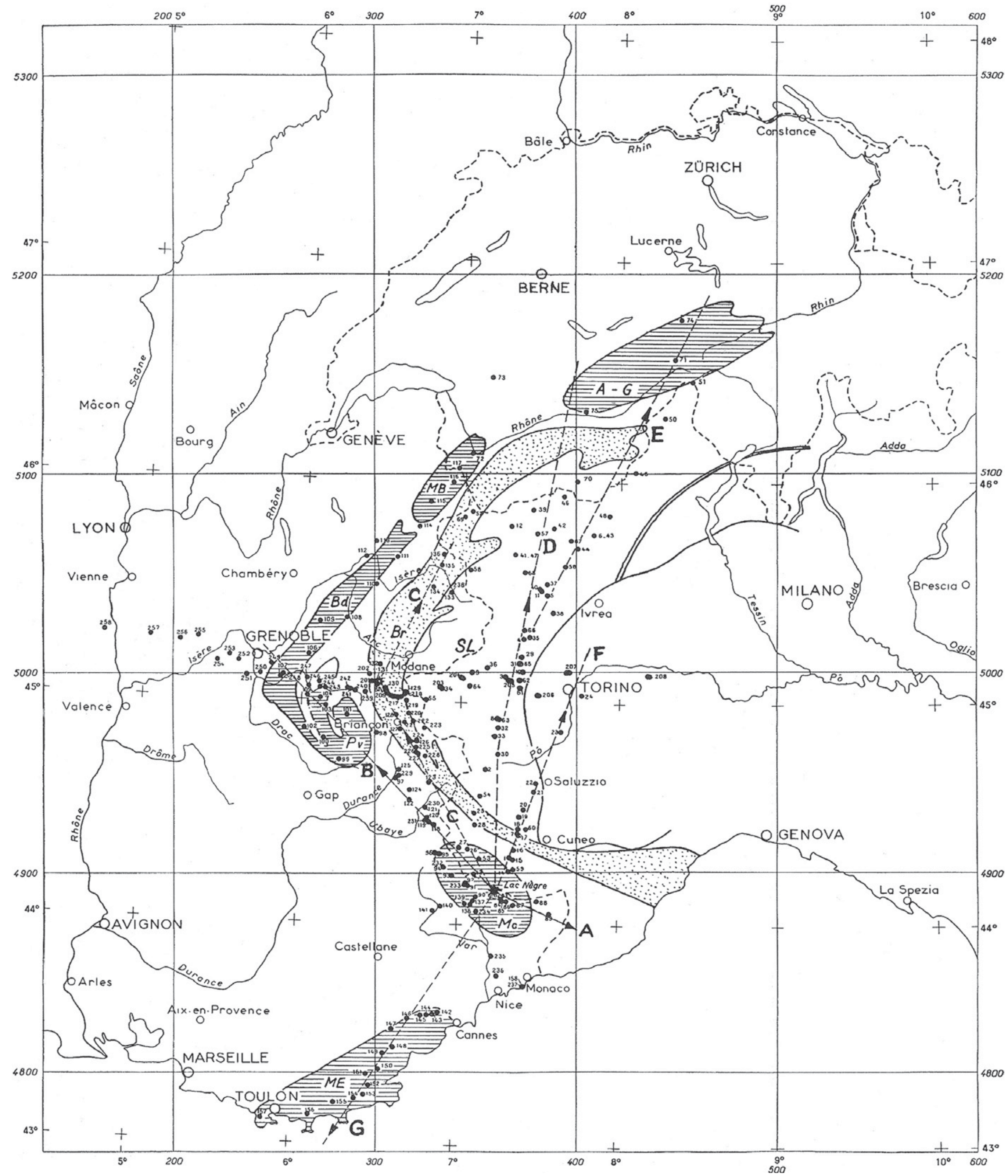


Figure 5.1-01. Location map of seismic observations in the Western Alps in 1956 and 1958. Hatched areas: crystalline massifs of the Western Alps. *A* to *G*: originally planned profiles (from Fuchs et al., 1963a, fig. 1). [*In* Closs, H., and Labrouste, Y., Séismologie: Recherches séismologiques dans les Alpes occidentals au moyen de grandes explosions en 1956, 1958 et 1960. Mémoir Collectif, Année Géophysique Internationale, Centre National de la Recherche Scientifique, Série XII, Fasc. 2, p. 118–176. Reproduced by permission of the author.]

characterized by an anomalous gravity high, some smaller explosions near Levone (SW of Ivrea) and at Monte Bavarione (near Lago Maggiore) were organized in 1960 by L. Solaini (Istituto di Geofisica Applicata del Politecnco, Milano) and observed by the same international community.

The correlation of observed phases was extremely difficult due to the topography, the areal distribution of the recording sites, and differing types of instruments. E. Peterschmitt used a mechanical affinograph which allowed tracing the seismograms in a normalized scale (unique time scale and normalized amplitudes) on Plexiglass. These Plexiglass seismograms were then placed at their proper location on a large topographic map which allowed a three-dimensional view of all phases. This procedure finally allowed Fuchs et al. (1963a) to recognize and correlate six wave types and to determine their velocities and intercept times. An example of seismograms and their correlation by Fuchs et al. (1963a) is shown in Figure 5.1-02.

While the phases *i* and *g* were interpreted as P-waves, the phase *m* was regarded as a converted phase sPp by Fuchs et al. (1963a), arguing that on 40% of the seismograms weak pPp arrivals could be recognized 4 seconds earlier which could be correlated as P_n given the Moho-depth calculated from phase *m* as sPp phase. Using this assumption, under the Western Alps the Moho depth increased from ~30 km under the outer crystalline massifs in the west to a maximum of 53 km in the east under the zone of Ivrea. Figure 5.1-03 shows a W-E cross section through the Western Alps. If the phase *m* had been interpreted as a P-wave, a maximum Moho depth of 70 km would have resulted, which seemed unlikely to the authors at that time.

In contrast to Fuchs et al. (1963a, 1963b), Labrouste et al. (1963) derived a quite different model assuming that only one crustal layer exists which drops from near 10 km depth under the Ivrea gravity high to 40 km depth under the western crystalline massifs, thus assuming that the surface of the zone of Ivrea is part of the Moho. The discrepancy was later solved and discussed during a symposium in 1968, which is discussed in Chapter 6.

At about the same time, when the Western Alps were investigated in great detail, the first seismic investigations of the Eastern Alps started (Reich, 1958, 1960). Quarry blasts and organized drillhole explosions at the northern margin of the Eastern Alps enabled the determination of the thickness of the Molasse sediments. Because of the high velocities in the Triassic limestones

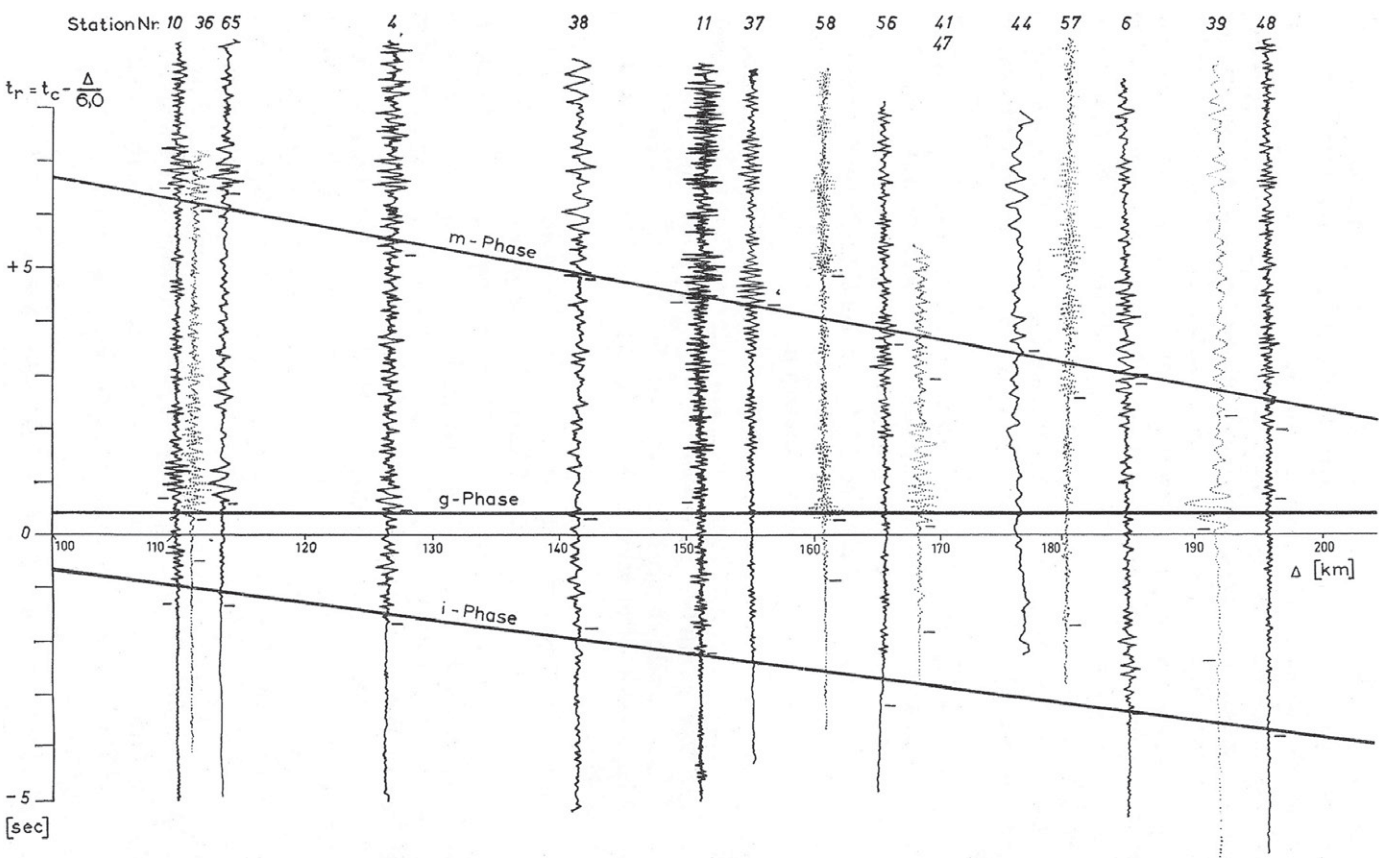


Figure 5.1-02. Profile-like arrangement of seismograms around the zone of Ivrea (line E on Figure 5.1-01). For clarification the onsets of the phases *i*, *g*, and *m* are connected by straight lines (Fuchs et al., 1963a, fig. 4). [*In* Closs, H., and Labrouste, Y., Séismologie: Recherches séismologiques dans les Alpes occidentals au moyen de grandes explosions een 1956, 1958 et 1960. Mémoir Collectif, Année Géophysique Internationale, Centre National de la Recherche Scientifique, Série XII, Fasc. 2, p. 118–176. Reproduced by permission of the author.]

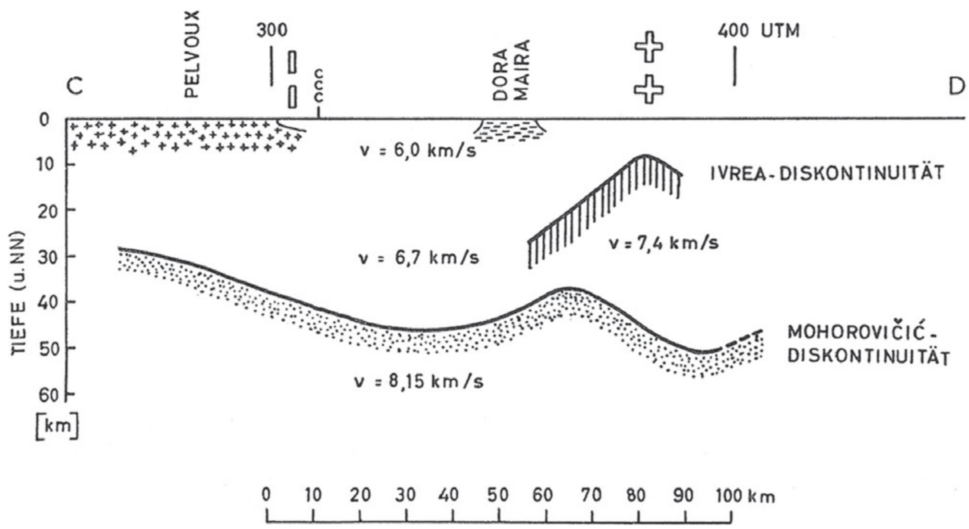


Figure 5.1-03. West-east cross section through the Western Alps from south of Grenoble to Torino (Fuchs et al., 1963a, fig. 8 below). [*In* Closs, H., and Labrouste, Y., Séismologie: Recherches séismologiques dans les Alpes occidentals au moyen de grandes explosions en 1956, 1958 et 1960. Mémoir Collectif, Année Géophysique Internationale, Centre National de la Recherche Scientifique, Série XII, Fasc. 2, p. 118–176. Reproduced by permission of the author.]

of the Northern Alps, however, it was problematic to observe the depth to the crystalline basement. Therefore the investigations there concentrated on the uppermost 10 km of the crust and touched only the northern margin of the Eastern Alps (see Fig. 6.2.4-01 for shotpoint locations).

5.2. THE BEGINNING OF SEISMIC-REFRACTION INVESTIGATIONS IN NORTH AMERICA

Since the late 1940s, studies of the Earth's crustal structure using blasts and rockbursts were also undertaken in North America (Gutenberg, 1952; Tuve et al., 1948, 1954) and Canada (Hodgson, 1953). Tatel and Tuve (1955) and Katz (1955) carried out seismic-refraction experiments in various regions of the United States. Tatel and Tuve (1955) have described in some detail the techniques of their time and the difficulties of choosing useful observation sites with a sufficiently low ground noise to detect the weak impulses of sometimes only a few Angstrom units in displacement amplitude. The portable seismometers used were electrodynamic: their output was electronically amplified and recorded on a pen-and-ink oscillograph. The earlier equipment had an extended frequency range of 0.3–400 cps, but experience showed that more effective records could be produced with a spread of 3–30 cps. This part of the spectrum was then employed in the newer equipment. Timing was obtained by recording second impulses of a chronometer simultaneously with the time signals from the National Bureau of Standards Radio Station WWV, obtaining times reliable to 0.03 seconds or better.

Some data examples are shown in Figure 5.2-01. Distances from shot to receiver positions were located on the best maps available and errors were estimated to be well below 0.5 km. The resulting travel times of correlated arrivals were plotted to 0.1 second accuracy, and true fluctuations were observed around an average line of approximately ± 0.5 seconds.

Tatel and Tuve (1955) concluded that the variations in their measurements were due to the Earth's characteristics and not to the measurements themselves. The data obtained were plotted on charts of time versus distance. To obtain better accuracy, reduced times were plotted; e.g., the difference between the actual time and the time required for a wave to traverse the distance at the mean compressional velocity of the waves. This mean velocity varied from 5.5 km/s in Utah to 6.2 km/s in the Shenandoah Valley (Fig. 5.2-02).

The experiments covered various geologic provinces. From 1948 to 1953, underwater shots by the U.S. Navy in the Patuxent River and in the Chesapeake Bay of 600–2400 lbs enabled the investigation of the Atlantic coastal region in Maryland and Virginia. Three profiles were recorded in 1950 in the Appalachians, along strike up to 1100 km toward NNE and 150 km toward SSW, and up to 1500 km perpendicular to the strike, enabled by several blasts of up to 650 tons during the construction of a dam in northern Tennessee. Open-pit mine blasts were recorded in 1951 on the southern Canadian Shield in Minnesota.

In California, a SW-trending profile was recorded from a blast at Corona, and a second set of shots was recorded from explosions set of off the California coast near Santa Barbara. In 1951, depth charges were exploded in the Puget Sound in Washington and recorded on two lines. Critical reflections, named PP, could not be observed, and the P_n-velocity (named P_2 by the authors) could not be accurately determined in either in California or in Washington. Recordings of explosions in the Bingham mine near Salt Lake City, Utah, caused some technical problems. Also here critical reflections were not observed, but the crossover distance of the wave P_2 (today's P_n) showed that the velocity transition was small and its depth of 29 km was similar to that found in Arizona and New Mexico in 1954. There, 9 explosions could be arranged and 30 good seismograms were obtained. Tatel and Tuve were surprised to "obtain a close-in set of second arrivals and cross-over distances" of P_n with a shallow depth of 30 km for

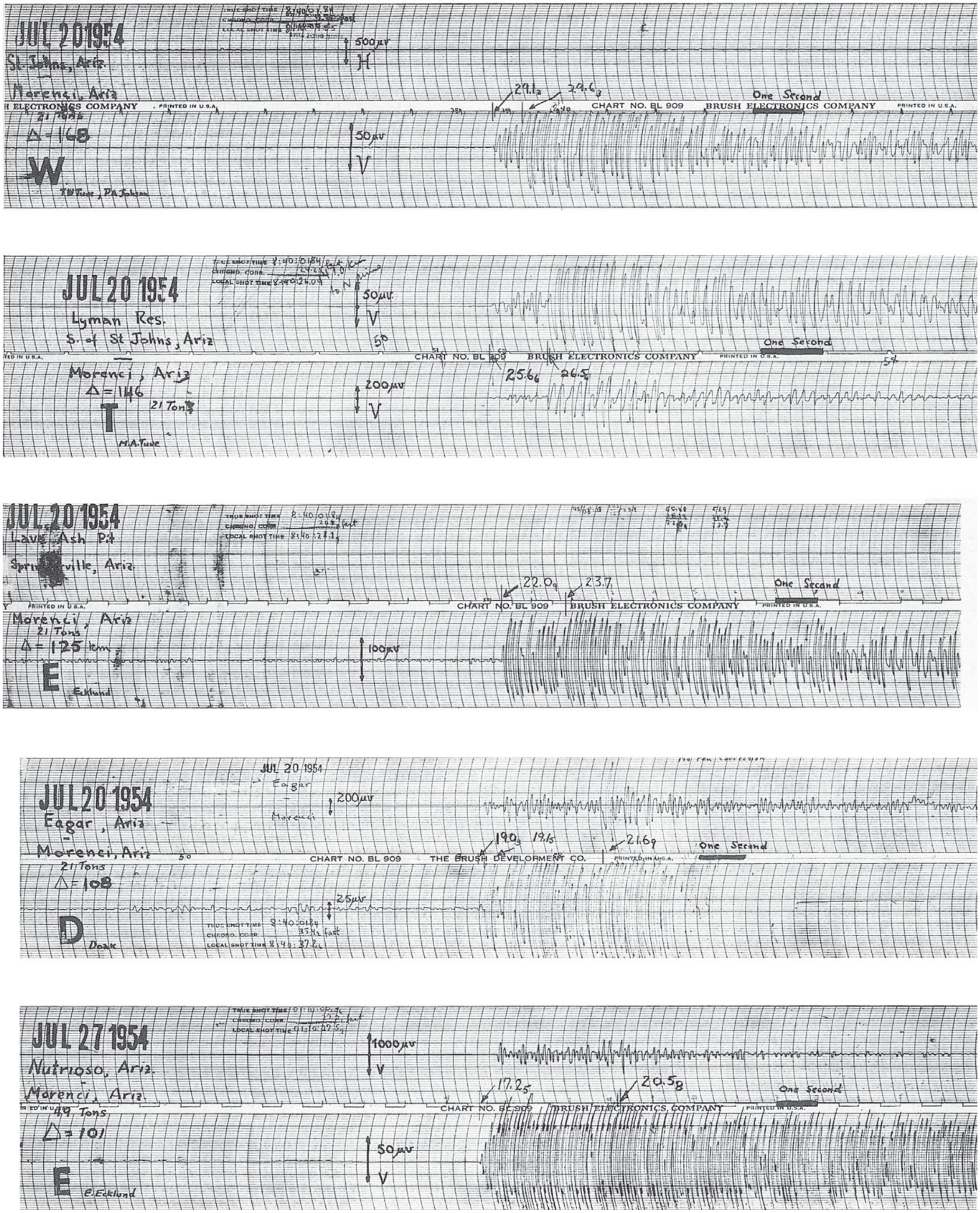


Figure 5.2-01. Field seismograms with critical reflections PP (3.2 to 0.5 seconds after first arrival from bottom to top seismogram) obtained in the Colorado Plateau in 1954 (from Tatel and Tuve, 1955, plate 5). [Geological Society of America Special Paper 62, p. 35–50. Reproduced by permission of the Geological Society of America.]

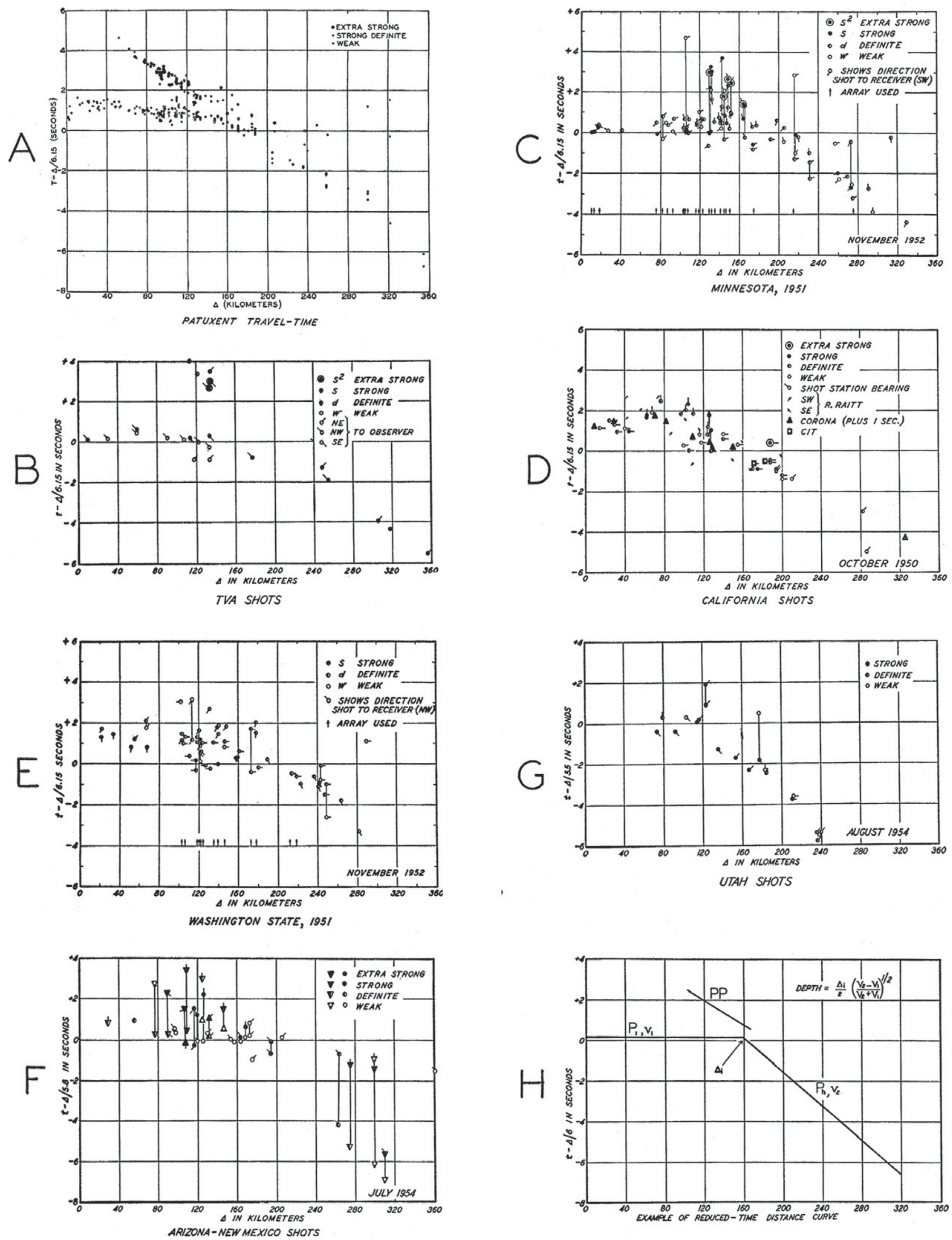


Figure 5.2-02. Time-distance plots for (A) Maryland-Virginia (coastal plain); (B) Tennessee-Virginia (Appalachians); (C) Minnesota (southern Canadian Shield); (D) southern California; (E) Washington (Puget Sound); (F) Arizona–New Mexico (southern Colorado Plateau); (G) Utah (central Rocky Mountains); (H) sketch of the prominent features (from Tatel and Tuve, 1955, fig. 4). [Geological Society of America Special Paper 62, p. 35–50. Reproduced by permission of the Geological Society of America.]

the velocity discontinuity. The results of their investigations were summarized in a table (Fig. 5.2-03).

Tatel and Tuve (1956) also attempted to determine the crustal structure in central Alaska along lines from College Fjord to Dawson and Fairbanks; these data were later (in the 1960s) reinterpreted together with other data (see, e.g., Berg, 1973).

Using similar equipment as Tatel and Tuve and seismometers built at the Lamont Geological Observatory, Katz (1955) reported on quarry blasts in northern New York and central Pennsylvania recorded from 1950 to 1953 to a distance of 309 km (Fig. 5.2-04). Two sample seismograms recorded at 181.4 km distance around the P_n crossover distance and at 220.5 km with P_n as the first arrival are shown in Figure 5.2-05 and a typical travel time plot in Figure 5.2-06. He derived an essentially homogeneous crust with an average thickness of 34.4 km for both regions, "with elastic wave velocities possibly increasing with depth" observing P-velocities of 6.31–6.39 km/s for New York and 6.04 km/s for Pennsylvania. His P_n velocity was 8.14 km/s. He did not see any vertical reflections from the Mohorovičić discontinuity. Katz mentioned some abnormally small first arrivals between 80 and 140 km distance as well as a large arrival at 80–95 km distance following the P_1 phase one second later on all three profiles; since he could not interpret it at the time, he speculated on the possible existence of a low-velocity layer.

In 1958, researchers at the University of Wisconsin observed a non-reversed seismic-refraction profile in Arkansas and Missouri, using quarry blasts of 2000–24,000-lb charges near Little Rock (Steinhart et al., 1961a; Appendix A5-1). In 1958 and 1959, crustal profiles were recorded in Wisconsin and Upper Michigan using underwater shots in Lake Superior and a quarry blast to reverse one of the profiles (Steinhart et al., 1961b; Appendix A5-1).

The University of Wisconsin researchers also did some early studies of the Rocky Mountain area (Steinhart and Meyer, 1961). In 1959, several profiles were recorded in Montana (Meyer et al., 1961b; Appendix A5-1). Most of the observations in Montana were reversed profiles. The preliminary results were used to construct a fence diagram of the area investigated (McCamy and Meyer, 1964; Figure 5.2-07), but the authors pointed out that at several crossing points the results were inconsistent and would require further analysis.

In 1957, the University of Wisconsin undertook an expedition to measure the thickness of the Earth's crust under the Central Plateau of Mexico (Meyer et al., 1958, 1961a; Appendix A5-1). A series of shots, fired in connection with open pit mining operations near Durango, Mexico, was recorded to ~320 km distance. To obtain observations near the main shotpoint, a reverse profile of 30 km length was also arranged. Several models ranging from a one-layer crust to a three-layer crust resulted in Moho depths between 40 and 45 km.

West of the Rocky Mountains, from 1956 to 1959, Berg et al. (1960) recorded quarry blasts at Promontory and Lakeside in Utah with shot sizes from 50,000 to 2,138,000 lbs and two nuclear explosions near Mercury, Nevada, equivalent to 1.7 and 23 ktons, using 17 temporary and 15 permanent stations. Most stations were located around the quarries Promontory and Lakeside out to ~150 km distance with station spacings between 10 and 30 km, but single stations were placed, e.g., at Gold Hill, Utah, and at Wells, Elko, and Eureka, Nevada, up to distances of 355 km. Also, a few stations outside the Basin and Range province, e.g., in Wyoming and Montana, recorded these events (Fig. 5.2-08). A three-layer model was proposed for the eastern Basin and Range province assuming horizontal layering (Fig. 5.2-09): The upper crust with velocity of 5.73 km/s is 9 km thick; the lower crust with 6.33 km/s extends to 25 km depth, and is underlain by a layer with 7.59 km/s reaching to 72 km depth where a velocity of 7.97 km/s is obtained. The authors hesitated to name the 25-km-deep discontinuity the

Region	v_1, P_1 km/sec	v_2, P_2, km/sec	Critical reflection	Δ_i, km	h (km)	Approximate surface elevation (ft)	Topography
Maryland-Virginia	6.1	8.1	Yes	145–160	26–29	100	Flat
Minnesota	6.1	8.1	Yes	200 ± 10	37	1400	Flat
Tennessee-Virginia	6.0	8.1	Yes	210 ± 20	39	2500	Mountains—Deep Valleys
California							
Coast	6	8	No	120	~23	200	Coast—Mountains
Corona				120–170	23–32	1500	
Washington							
East–South	6	8	No	~160	~30	1200	Coast—Mountains
West				~100	~19	700	
Arizona-New Mexico							
North	5.8	8.1	Yes	175	34	7000	Colorado Plateau —Mountains
South				145	28	4500	
Utah	5.5	8	No	135	29	~6500	Plains—Mountains

$h = \frac{1}{2}\Delta_i\left(\frac{v_2 - v_1}{v_2 + v_1}\right)^{\frac{1}{2}}$ —*assumes no increase of velocity with depth—for purposes of comparison only*

Figure 5.2-03. Regional comparison of velocities of waves P_1 (crust) and P_2 (today's P_n), "being recorded," an estimate of the cross-over distance Δ_i, depth of discontinuity (Moho), approximate surface elevation (feet), and topography (from Tatel and Tuve, 1955, table 1). [Geological Society of America Special Paper 62, p. 35–50. Reproduced by permission of the Geological Society of America.]

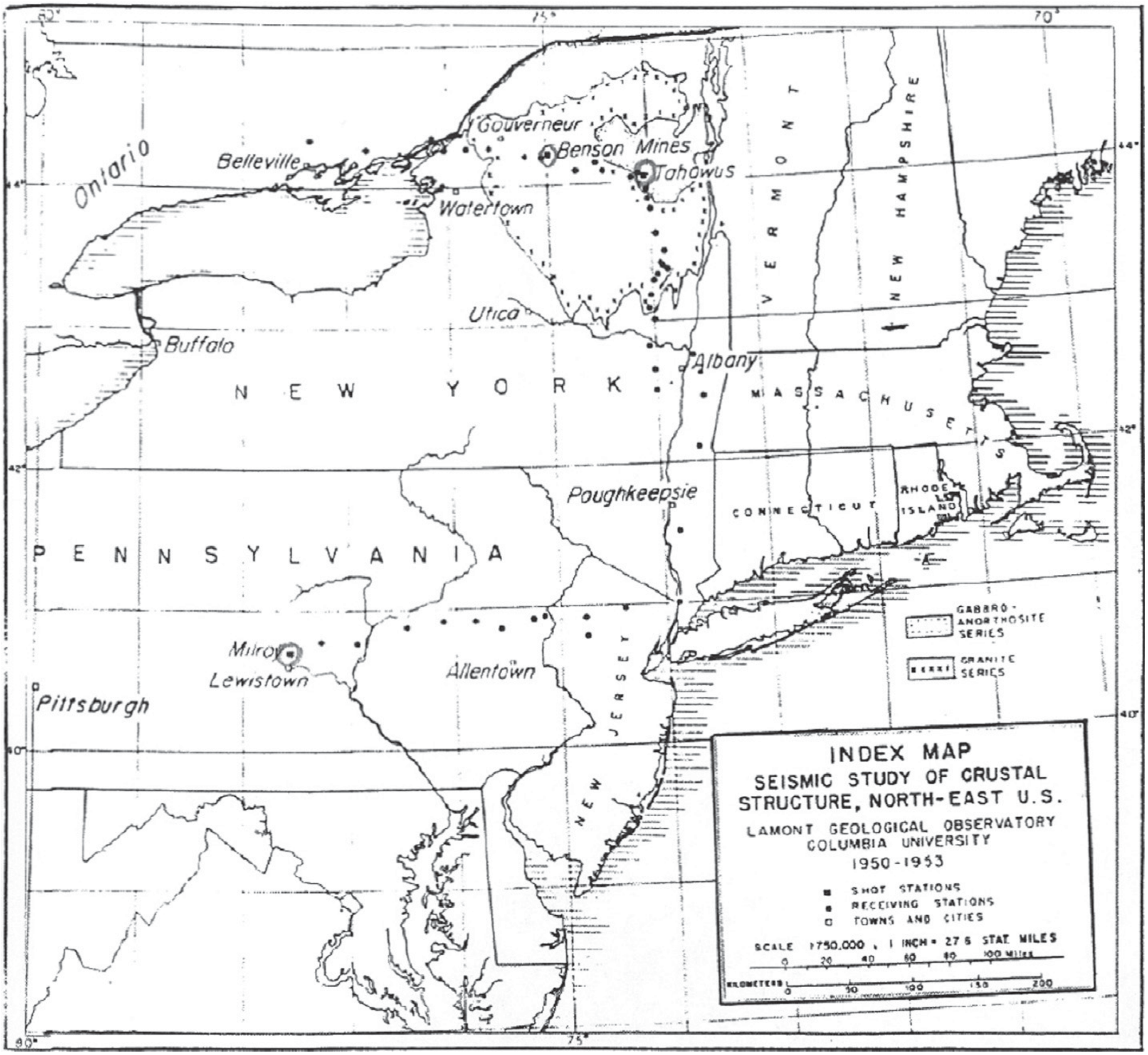


Figure 5.2-04. Index map of seismic-refraction lines in 1950–1953 in the northeastern United States (from Katz, 1955, fig. 1). [Bulletin of the Seismological Society of America, v. 45, p. 303–325. Reproduced by permission of the Seismological Society of America.]

Moho because of the low velocity of only 7.5 km/s underneath, but nevertheless they inferred that this material is "intimately associated with the mantle" and suggested that this mantle has undergone expansion and, referring to Kennedy (1959), discussed a possible phase change caused by increased heat.

Press (1960) has discussed a non-ideal but sufficient two-way coverage of data recorded in southern California from a nuclear blast in Nevada in the north and from quarry blasts near Corona and Hectorville, California, in the south, concluding that no significant crustal variations occur within the region. He obtained a P_2 velocity of 7.66 km/s and a P_n velocity of 8.11 km/s and discussed his favorite model of a 50-km-thick double-layered crust with an intermediate layer at a depth of 24 km with an unusual high velocity (Fig. 5.2-10).

Starting in ca. 1952, the U.S. Geological Survey conducted a modest program of crustal studies using gravity and seismic-refraction methods. The gravity studies of crustal structures were made in conjunction with the Geological Survey's regular program of regional gravity surveying. The immediate purpose of most of these regional gravity surveys was the elucidation of

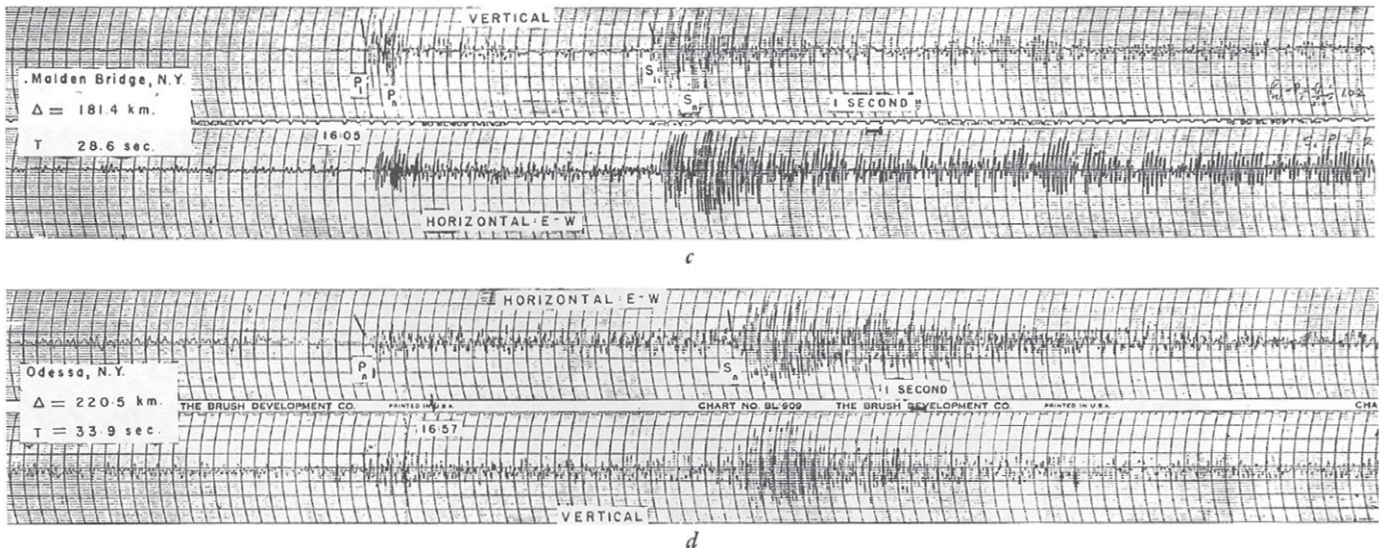


Figure 5.2-05. Seismograms recorded from quarry blasts in New York and Pennsylvania (from Katz, 1955, fig. 5c+d). [Bulletin of the Seismological Society of America, v. 45, p. 303–325. Reproduced by permission of the Seismological Society of America.]

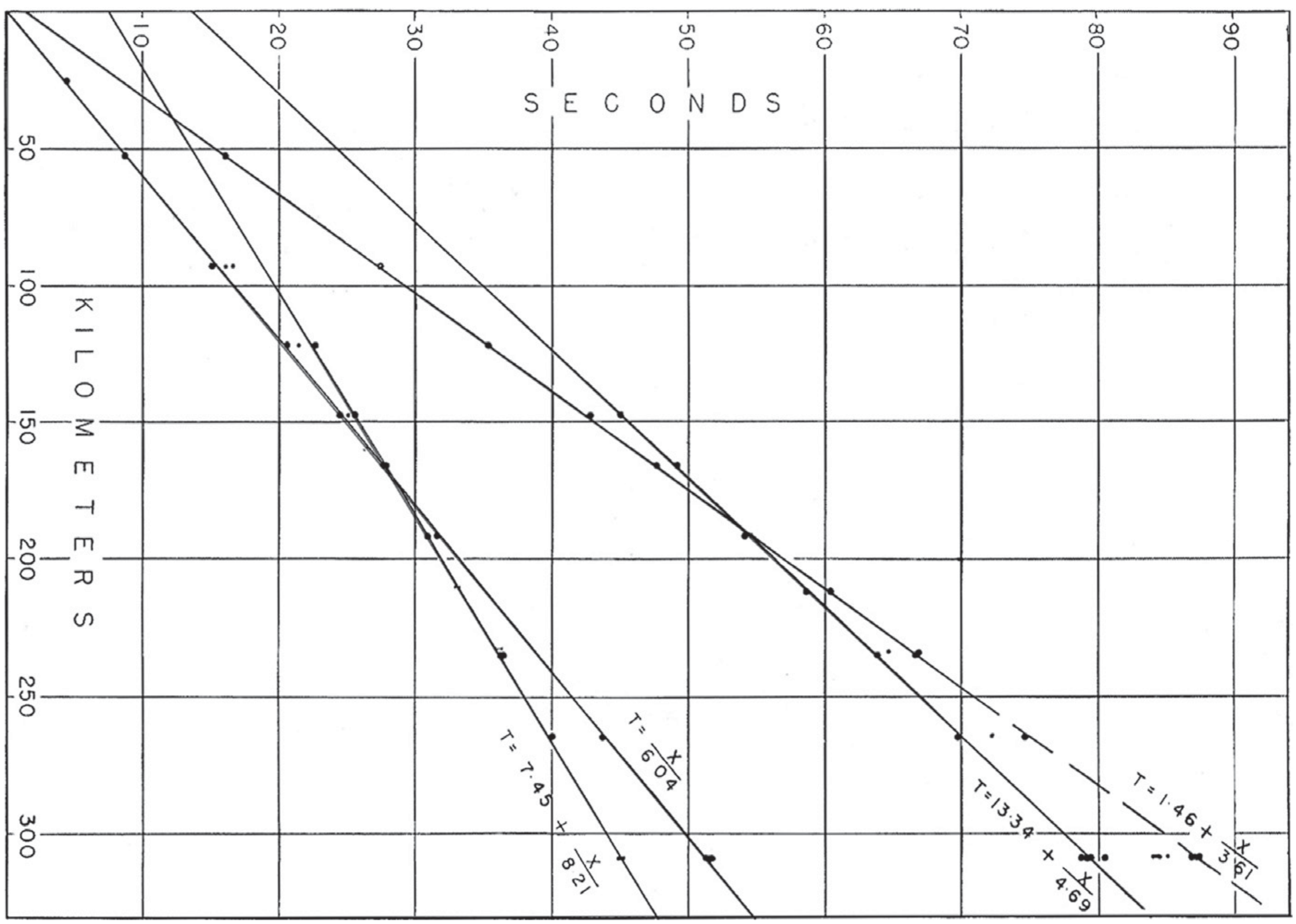


Figure 5.2-06. Travel-time distance correlation of seismic observations in 1950–1953 (from Katz, 1955, fig. 4). [Bulletin of the Seismological Society of America, v. 45, p. 303–325. Reproduced by permission of the Seismological Society of America.]

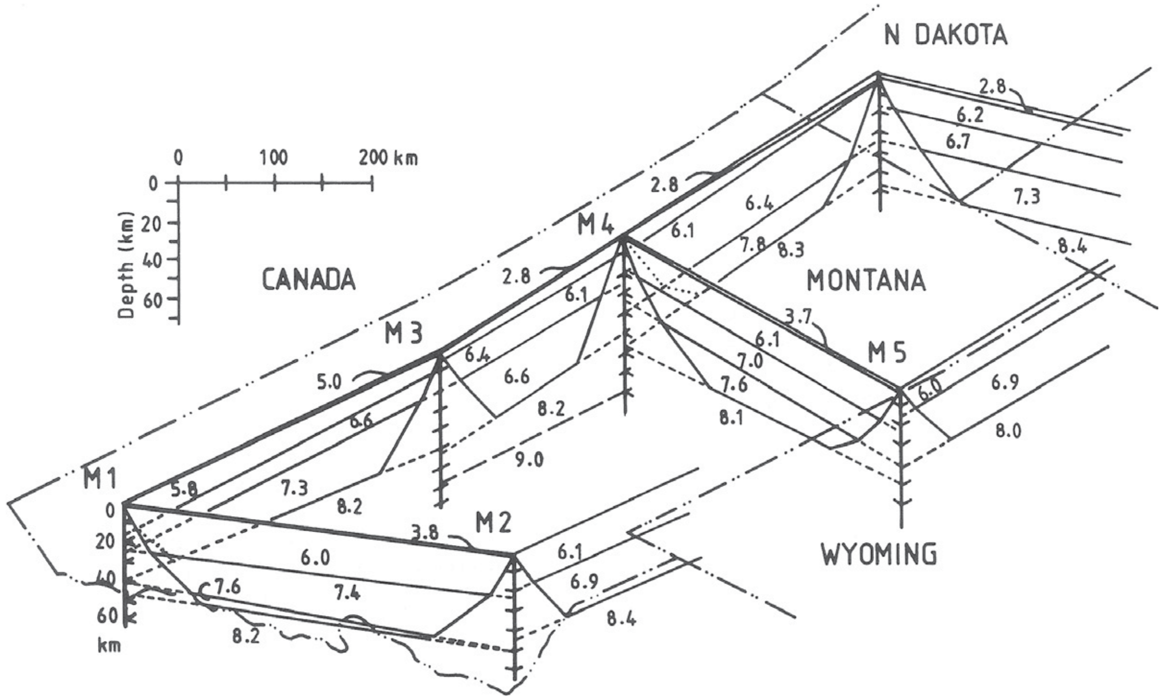


Figure 5.2-07. Fence diagram showing the crustal structure in Montana (McCamy and Meyer 1964, fig. 1, reprinted from Prodehl and Lipman, 1989, fig. 2). [*In* Pakiser, L.C., and Mooney, W.D., eds., Geophysical framework of the continental United States. Geological Society of America Memoir 172, p. 249–284. Reproduced by permission of the Geological Society of America.]

large-scale geologic features of interest in the Geological Survey's regular geologic mapping program. The gravity coverage, however, was sufficiently broad to permit interpretation of the gravity data in terms of major crustal features.

Regional gravity work had been done, for example, in the Sierra Nevada and adjoining areas of California (Pakiser et al., 1960; Pakiser, 1961; Kane and Pakiser, 1961; Oliver et al., 1961). Broad gravity coverage had also been obtained over a large part of the Basin and Range province in California and Nevada and the related Mojave block in southern California (Mabey, 1960a, 1960b) as well as in other parts of the western United States, resulting in new knowledge of the nature of crustal structure and isostatic compensation in the Sierra Nevada region as well as in the Basin and Range province.

In conjunction with the U.S. Geological Survey's geophysical investigations at the Nevada Test Site during the test programs of 1957 and 1958, a number of recordings of seismic waves generated by nuclear explosions were made along a line from the Nevada Test Site to the vicinity of Kingman, Arizona, using seismic instruments available to the Geological Survey at that time (Diment et al., 1961). As a result of these recordings and related work by Press (1960) and Berg et al. (1960), new information on the thickness and structure of the Earth's crust was obtained in the region surrounding the Nevada Test Site.

Alaska and adjacent northwestern Canada also became a target of seismic research in the 1950s. During an expedition of a small field party from the Department of Terrestrial Magnetism of the Carnegie Institution of Washington in 1955, a series of near-shore 1-ton underwater explosions in College Fjord, Alaska, was recorded on two 300-km-long lines toward the SW along the Kenai peninsula and north toward Fairbanks and on a line toward Dawson, Canada, with stations at distances between 200 and 450 km (Fig. 5.2-11). Also in 1955, during the same expedition, another series of near-shore 1-ton underwater explosions near Skagway, Alaska, was recorded on several lines up to 300 km distance (Fig. 5.2-12) ranging in a northerly direction into the Yukon Territory (Hales and Asada, 1966; Berg, 1973). One of the main results was the observation of clear differences in travel times for various azimuths in many of the areas studied. From the College Fjord shots to the northeast the crustal thickness appeared to be 48–53 km, while to the southwest in the region of the Kenai Peninsula it appeared 10–15 km thinner. Northwest of Skagway, the Moho was found at 36–42 km depth, but north of Skagway it was only 35 km.

5.3. SEISMIC-REFRACTION INVESTIGATIONS IN EASTERN EUROPE IN THE 1950s

The history of deep-seismic sounding experiments in the former USSR is described in detail by Pavlenkova (1996). The first period started at the end of the 1940s. The first deep-seismic research was conducted in 1948–1954 under the leadership of G.A. Gamburtzev, E. Galperin, and I.P. Kosmininskaya in central Asia and in the southern Caspian area. Since the middle of the

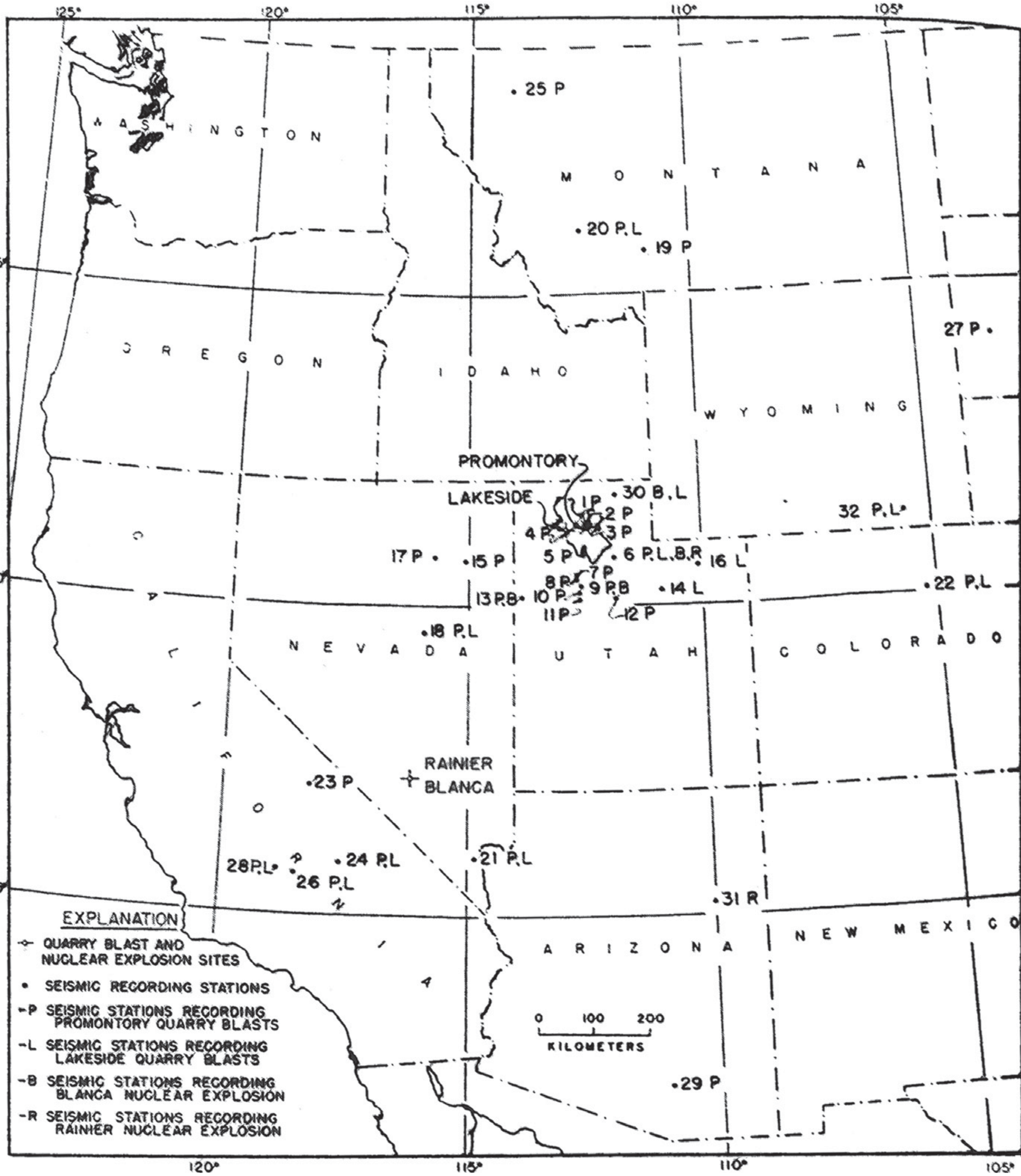


Figure 5.2-08. Index map of quarry blasts in Utah and nuclear blasts in Nevada and recording stations in the western United States (from Berg et al., 1960, fig. 1). [Bulletin of the Seismological Society of America, v. 50, p. 511–535. Reproduced by permission of the Seismological Society of America.]

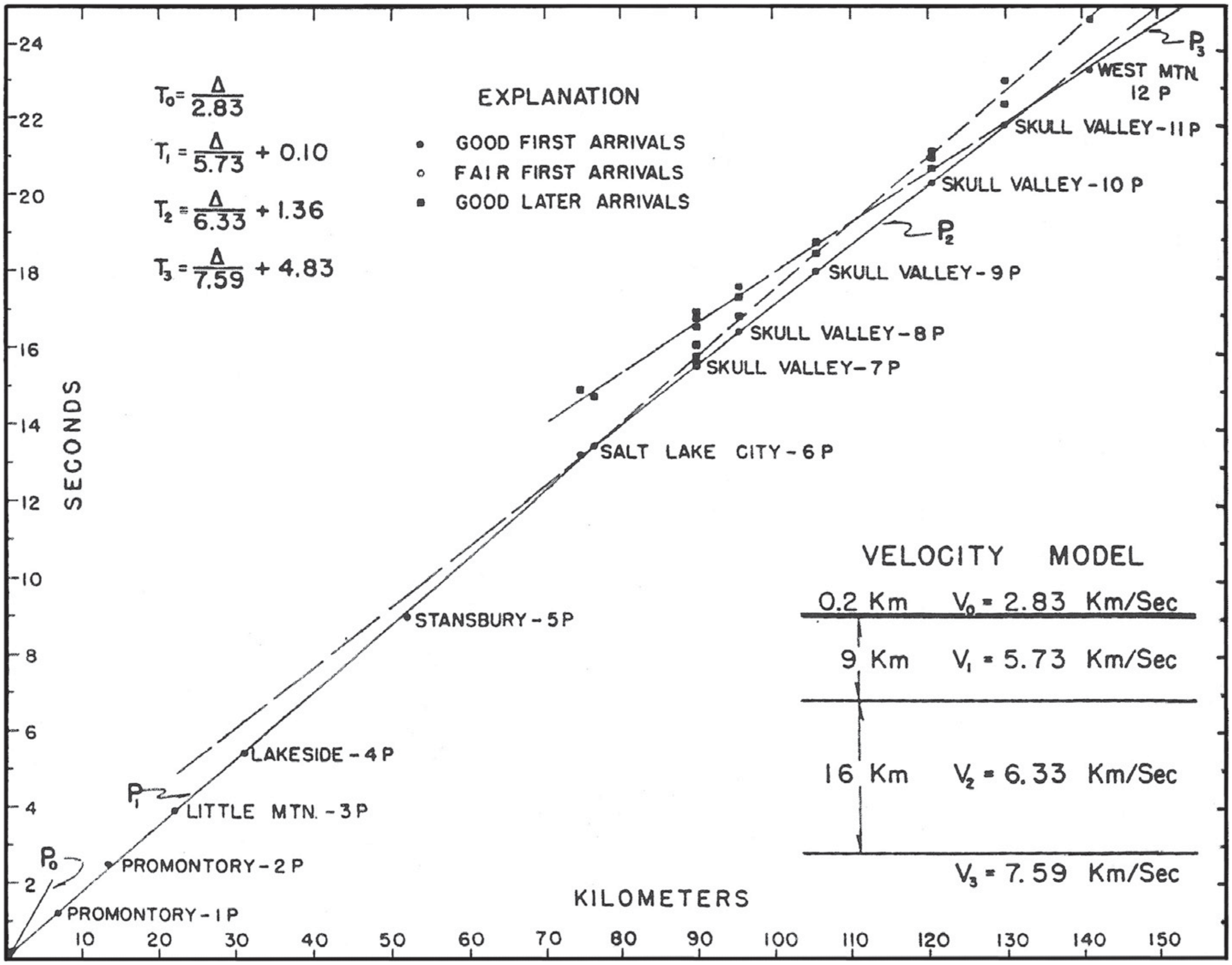


Figure 5.2-09. Travel-time distance correlation of seismic observations in Utah and a model for the eastern Basin and Range province (from Berg et al., 1960, fig. 5). [Bulletin of the Seismological Society of America, v. 50, p. 511–535. Reproduced by permission of the Seismological Society of America.]

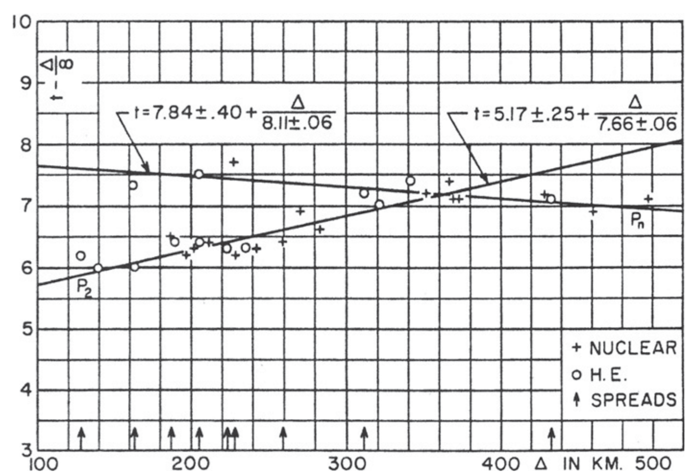


Figure 5.2-10. Experimental data from seismic refraction work in southern California (from Press, 1960, fig. 5). [Journal of Geophysical Research, v. 65, p. 1039–1051. Reproduced by permission of American Geophysical Union.]

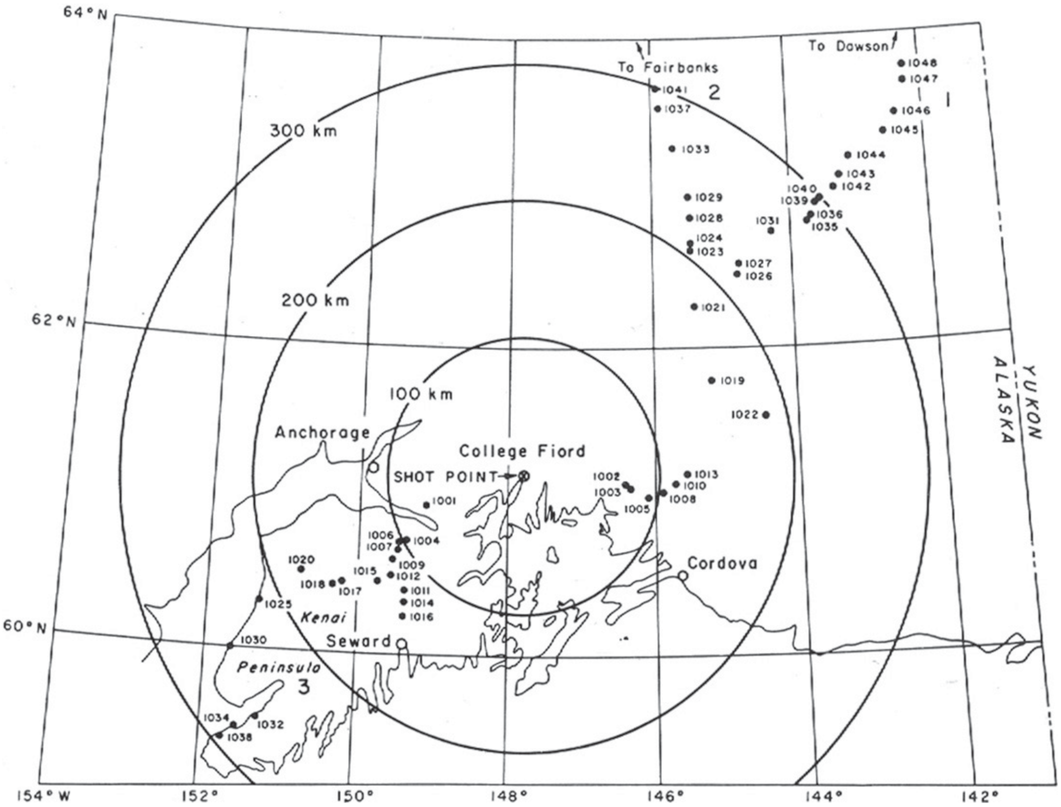


Figure 5.2-11. Location of the College Fjord shots and recording sites in Alaska (from Hales and Asada, 1966, fig. 8). [*In* Steinhart, J.S., and Smith, T.J., eds., The earth beneath the continents: American Geophysical Union, Washington, D.C., Geophysical Monograph 10, p. 420–432. Reproduced by permission of American Geophysical Union.]

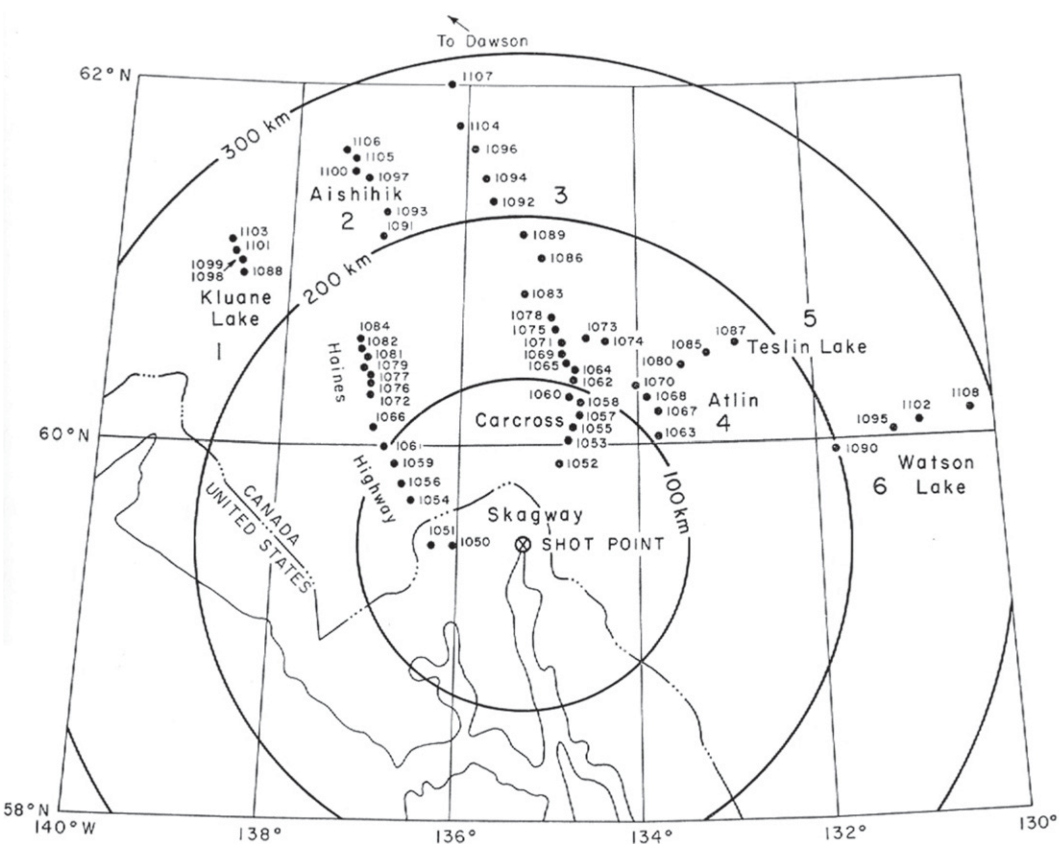


Figure 5.2-12. Location of the Skagway shots and recording sites in Yukon Territory, Canada (from Hales and Asada, 1966, fig. 1). [*In* Steinhart, J.S., and Smith, T.J., eds., The earth beneath the continents: American Geophysical Union, Washington, D.C., Geophysical Monograph 10, p. 420–432. Reproduced by permission of American Geophysical Union.]

1950s, deep-seismic sounding profiles were recorded on the Russian platform as well as on the Russian part of the Baltic Shield, in central Asia, in the Caucasus, and in the Urals covering thousands of kilometers (results published in Russian; for references see Pavlenkova, 1996).

In 1957 and 1958, a major research project was carried out in the transition zone from the Asian continent to the Pacific Ocean (Galperin and Kosminskaya, 1964). Nearly 30 profiles of several 100 km in length were laid out across the Sea of Ochotsk between Kamchatka and the islands of Sakhalin and Hokkaido and across the southern Kuril islands, mainly perpendicular to the strike of the continental margin (Fig. 5.3-01). Some of the data were reinterpreted 50 years later (Pavlenkova et al., 2009).

Figure 5.3-01. Location of deep-seismic sounding lines recorded in 1957–1958 in the Sea of Ochotsk (from Galperin and Kosminskaya, 1964, fig. 1.1). [Structure of the earth's crust in the transition zone between Asia and the Pacific: Moscow, Nauka (in Russian). Reproduced by permission of the Schmdt Institute of Physics of the Earth RAS.]

Data examples and a system of correlated travel-time curves for one of the profiles (1-M) are shown in Figure 5.3-02 and Figure 5.3-03. The resulting crustal thickness (Fig. 5.3-04) decreases from 30 km to less than 25 km in the center of the Sea of Ochotsk between Sachalin and Kamchatka, but also shows transitions to oceanic crust in the south near Hokkaido as shown, e.g., along line 1-M (Fig. 5.3-03 and Figure 5.3-04). Oceanic crust of 10 km thickness or less is encountered at the outermost southeastern end of the lines traversing the continental margin (Galperin and Kosminskaya, 1964).

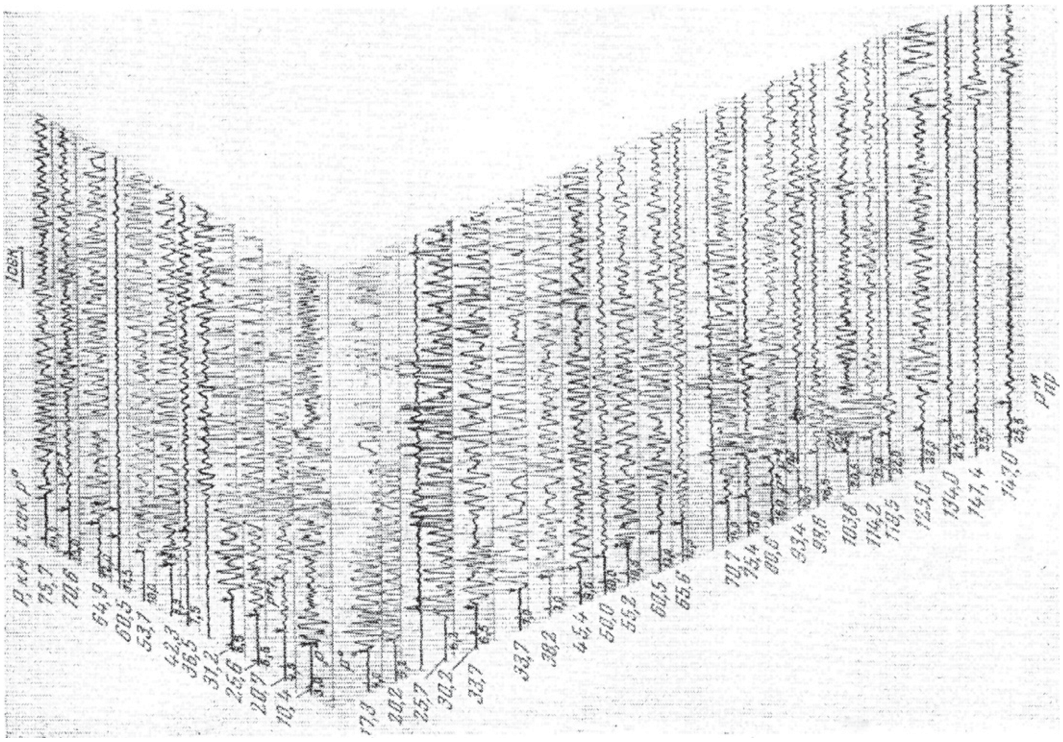


Figure 5.3-02. Seismograms of deep-seismic sounding observations in the Sea of Ochotsk from station 3 along profile 1-M (from Galperin and Kosminskaya, 1964, fig. 8.7). [Structure of the earth's crust in the transition zone between Asia and the Pacific: Moscow, Nauka (in Russian). Reproduced by permission of the Schmdt Institute of Physics of the Earth RAS.]

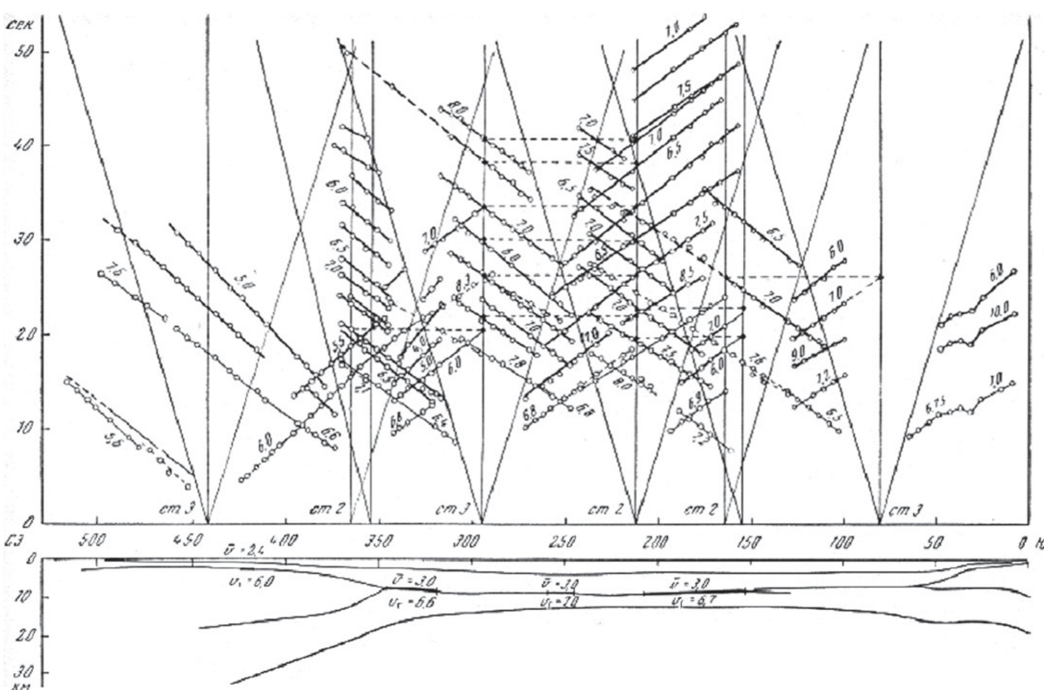


Figure 5.3-03. Travel-time diagram of deep-seismic sounding observations in the Sea of Ochotsk along profile 1-M (from Galperin and Kosminskaya, 1964, fig. 3.4). [Structure of the earth's crust in the transition zone between Asia and the Pacific: Moscow, Nauka (in Russian). Reproduced by permission of the Schmdt Institute of Physics of the Earth RAS.]

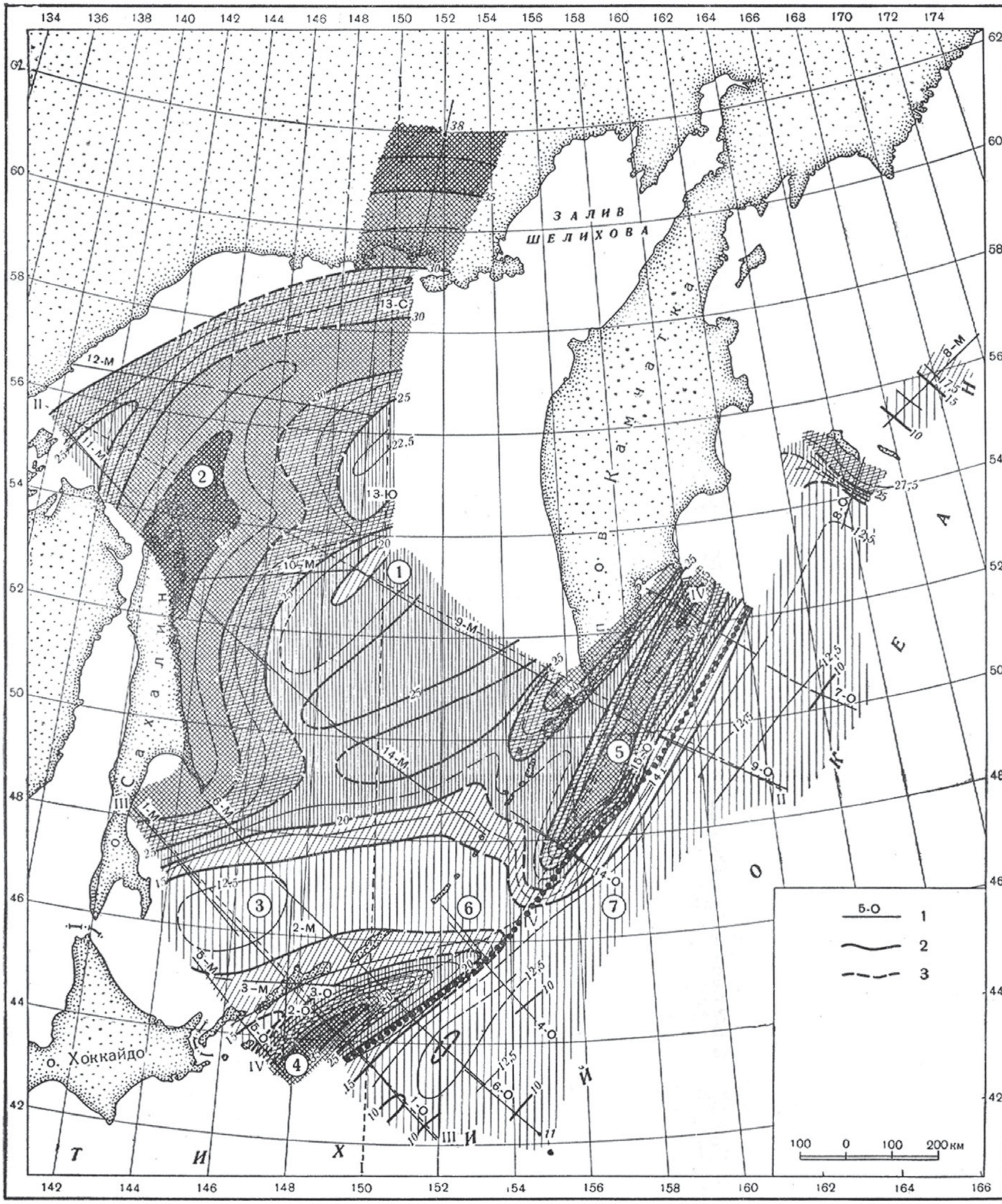


Figure 5.3-04. Crustal thickness map and seismic lines based of deep-seismic sounding observations in the Sea of Ochotsk (from Galperin and Kosminskaya, 1964, fig. 12.5). [Structure of the earth's crust in the transition zone between Asia and the Pacific: Moscow, Nauka (in Russian). Reproduced by permission of the Schmdt Institute of Physics of the Earth RAS.]

5.4. EARLY EXPLOSION-SEISMIC STUDIES IN JAPAN

In Japan, controlled-source seismology started on 25 October 1950, when the construction of the dam Isibuti in north-central Honshu required a large explosion of 57 tons of "carlit" to be detonated simultaneously (Fig. 5.4-01).

Within ten days, the Research Group for Explosion Seismology was established and eight temporary stations were placed at ~20 km intervals along the Tohuku railway line to a maximum distance of 120 km. The firing time was fixed such that the radio time signal JJY could be recorded. The recording equipment used was electromagnetic seismographs with 1, 3, and 10 cps, but also mechanical and mechanical-optical seismographs of 1 cps were used in combination with high-gain amplifiers and electromagnetic oscillographs. The speed of the recording paper or film varied between 2 and 40 mm/s. Sample records are shown in Figure 5.4-02. Two phases with velocities of 5.26 and 6.13 km/s were correlated and a depth of 1.3 km was determined for the second layer: this interpretation caused difficulties for the authors (Research Group for Explosion Seismology, 1951).

In the following years (1951 and 1952), two additional, but weaker explosions from the same source could be recorded, both with 18 stations on two profiles. Finally, in December 1952, a fourth explosion could be recorded from another source near Kamaisi near the coast in northeastern Honshu (Fig. 5.4-01). Unfortunately, the energy was not as strong as expected due to delayed firing of the 29.7 tons of dynamite (7 steps with 25 ms delay) (Research Group for Explosion Seismology, 1952, 1953). Nevertheless the shot was recorded up to 510 km distance on the southern profile. On this occasion three velocities were reported (Research Group for Explosion Seismology, 1954): 6.19, 7.37, and 8.20 km/s for which the authors determined layer thicknesses of 27.2 and 5.1 km, i.e., a total thickness of 32.3 km. However, the authors had doubts about the existence of layer 2 with a 7.37 km/s velocity.

The next series of explosions was fired in 1954 and 1955 at dam constructions near Lake Nozori, ~300 km farther south, of which two with 3.7 and 1.55 ton charges were recorded in a northwesterly direction up to 250 km distance. In addition, a reversing shot was organized by the Research Group for Explosion Seismology near Hokoda in 1956 and 1957, 180 km NE from Nozori,

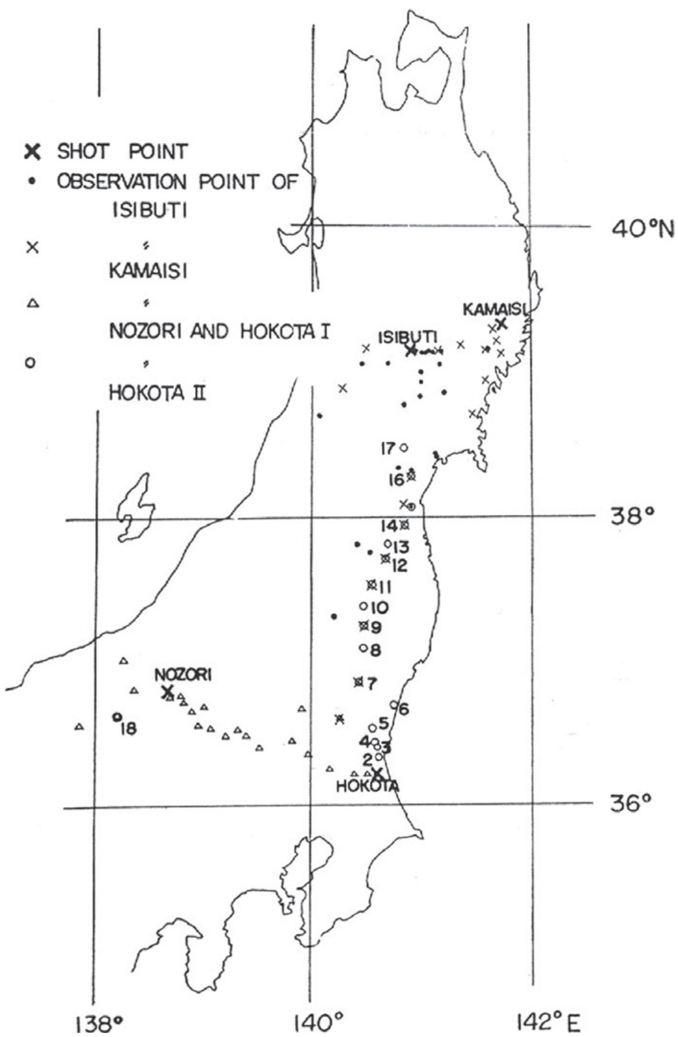


Figure 5.4-01. Shot and observation points of the first Japanese controlled-source seismology experiments (from Research Group for Explosion Seismology, 1959, fig. 1). [Bulletin of the Earthquake Research Institute, Tokyo, v. 37, p. 495–508. Published by permission of Earthquake Research Institute, University of Tokyo.]

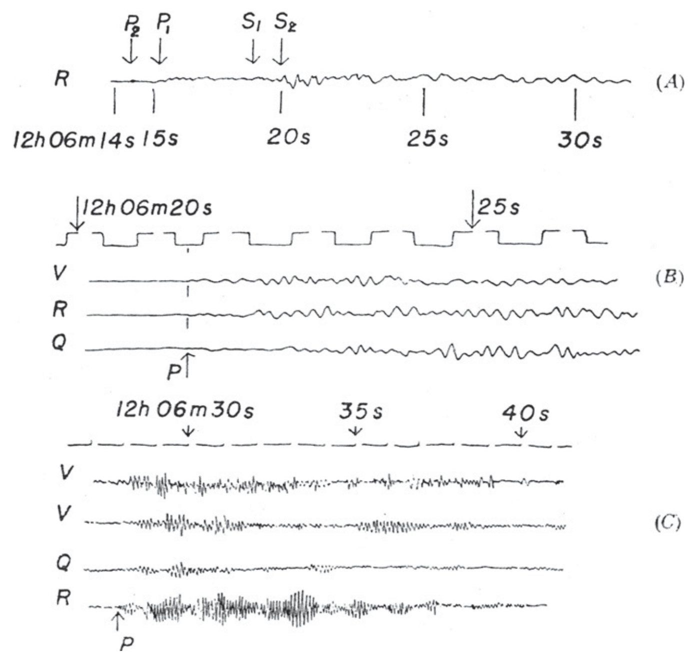


Figure 5.4-02. Seismograms of the first Japanese controlled-source seismology experiment. Distance from shotpoint: (A) 38.42 km, (B) 81.37 km, (C) 121.91 km (from Research Group for Exploration Seismology, 1951, fig. 7). [Bulletin of the Earthquake Research Institute, Tokyo, v. 29, p. 97–105. Published by permission of Earthquake Research Institute, University of Tokyo.]

where 1 ton of dynamite was distributed in 6 boreholes of 60–70 m depth and blasted simultaneously. In total, 33 temporary observation stations, with paper running speeds of 3 cm/s and more, recorded the shots (Fig. 5.4-01). The station separation varied from 5 to 25 km (Research Group for Explosion Seismology, 1958). Velocities of 5.5, 6.1, and 7.7 km/s were derived from the travel-time plots (Usami et al., 1958) and five differing models were discussed. The favorite model shows varying existence of layer 1 (5.5 km/s) along the profile, up to 5 km thick where it exists and an average depth of 25 km of the 6.1 km/s layer. The authors point out that the Moho is not 50 km deep as used before in earthquake studies as the standard model for the crust, but is between only 20 and 30 km depth (Usami et al., 1958). Finally, a second explosion was fired in August 1957 by the Research Group for Explosion Seismology near Hokoda to record a profile to the north and thus to reverse the S-profile from Kamaisi observed in 1952 (Research Group for Explosion Seismology, 1959). The resulting model by Matuzawa et al. (1959) shows the Moho dipping from 20 km in the north to 27.5 km in the south, the average velocity of the crust being 6.2 km/s and the P_n velocity 7.7 km/s.

In 1959, the first Japanese deep sea expedition (JEDS) was undertaken in close cooperation with U.S. institutions investigating the Japan Trench (Ludwig et al., 1966). Results were interpreted together with a second expedition undertaken in 1961 (see Chapter 6).

5.5. EARLY EXPLOSION-SEISMIC STUDIES IN THE SOUTHERN HEMISPHERE

Since the mid 1950s, studies of the Earth's crustal structure using blasts, rock bursts, and atomic explosions were also undertaken in Australia (Bolt et al., 1958), South Africa (Willmore et al., 1952), and South America (Tatel and Tuve, 1958; Steinhart and Meyer, 1961).

5.5.1. Australia and New Zealand

It was nuclear explosions which started seismic crustal studies in Australia (Bolt et al., 1958; Cleary, 1967; Doyle, 1957; Finlayson 2010, Appendix 2-2). British atomic explosions at Emu in 1953 and at Maralinga in 1956, both in South Australia, were recorded along the trans-Australian railway from South Australia to Perth up to 700 km distance and enabled the interpretation of P and S phases. From the recordings of the Maralinga explosion westwards across the Nullarbor Plain the first definitive measurement of Moho depth on continental Australia was interpreted (Bolt et al., 1958). Assuming a one-layer crust, because intermediate arrivals could not be recognized, the interpretation of the P data resulted in a crustal thickness of 32 ± 3 km and 39 ± 3 km being interpreted from P_n and S_n data respectively. Upper mantle P_n and S_n velocities of 8.21 km/s and 4.75 km/s were interpreted. Furthermore, large explosions for dam construction at Eaglehawk quarry in the Snowy Mountains in New South Wales with charges between 50 and 100 tons were recorded up to 375 km distance toward Sydney and Melbourne (Doyle et al., 1959). The locations of these early shotpoints are included in Figure 6.7.1-01 (Cleary, 1973) in Chapter 6.7. Systematic recording of seven large blasts of the Prospect blue metal quarry near Sydney in 1959 ranging in size from 2000 to 6260 kg at permanent seismographs resulted in P-wave velocities of 6.16 km/s for crustal rocks of the Southern Highlands of New South Wales and 5.88 km/s within the Sydney Basin (Bolt, 1962).

It is worthwhile noting that, similar to the observations in Germany (e.g., Dohr, 1959), since 1957 the Bureau of Mineral Resources undertook efforts to obtain near-incident vertical seismic reflections from the lower crust and upper mantle in at least one location on most seismic surveys by running near-vertical reflection records up to 38 s or by using large explosions at the end of surveys. Techniques for recording deep reflections improved with experience, but it was not until 1960 that the first definite result was obtained (Moss and Dooley, 1988).

In New Zealand, first preliminary explosion-seismic studies were performed near Auckland in 1951 and Clutha River, South Island, where surplus military explosives had been detonated (Davey, 2010, personal commun.). In 1952, the first crustal seismic-refraction profile was recorded in New Zealand in the Wellington area (for location, see Figure 8.8.1-05). The profile was ~175 km long and unreversed with shots fired at its southern end in Wellington Harbour. Strong reflections at 20 km depth were interpreted by Officer (1955b) as crustal thickness, but a subsequent reinterpretation gave 36 km crustal thickness and no refraction horizon associated with the strong reflection (Eiby, 1955, 1957; Garrick, 1968; Stern et al., 1986).

5.5.2. South Africa

The investigations in South Africa in the late forties have been discussed in Chapter 4.2.3. Further attempts of deep crustal structure in the 1950s are not known to the authors.

5.5.3. South America and Antarctica

In the 1950s, the Carnegie Institution, Washington, D.C., USA, undertook a seismic expedition into the Andean region of Peru and Chile (Tatel and Tuve, 1958). They found 65–70 km depths to Moho under the Altiplano of Peru and Chile and 51–56 km depths at the flanks of the plateau. Location and results were also included in the overview of Steinhart and Meyer (1961; Fig. 10.5 and Table 10.1; see Appendix A5-1-10).

In the 1950s, expeditions were also undertaken into the Antarctica to study the Earth's crust. In 1959 and 1960, a Belgian Antarctic expedition was organized to perform seismic-refraction and other geophysical measurements near the Belgian station in the Breidvika Bay (~70°S, 24°E). A 12-channel seismic recording device was positioned at 14 points on the ice shelf located along a 125-km-long line running southwards from the Soer-Rondane Mountains and shots were detonated around each station up to 20 km distance. In total, 68 shots were fired. The expedition achieved ice thickness results, but unfortunately, due to bad weather conditions, shots at larger distances to achieve a long-distance refraction profile could not be realized (Dieterle and Peterschmitt, 1964).

As part of the U.S. Antarctic research program, during the 1950s and continuing into the 1960s, some three dozen seismic-refraction profiles were completed during reconnaissance exploration of the Antarctic interior. Most of them were long enough to provide information about seismic velocities below the ice, but did not penetrate further than into the topmost upper crust (Bentley, 1973).

Not with active seismics, but with surface wave dispersion studies, the crustal thickness underneath Antarctica was determined by Evison et al. (1959).

5.6. SEISMIC-REFRACTION OCEANIC EXPLORATIONS OF THE CRUST IN THE 1950s

5.6.1. Introduction

Oceanic seismic-refraction experiments, which had started immediately after the end of World War II (see Chapter 4), were continued both in the Atlantic (Ewing et al., 1954) and in the Pacific and Indian Oceans (e.g., Raitt, 1956; Gaskell et al., 1958). The initial investigations concentrated particularly on the ocean basins. As described in some detail in Chapter 4, usually two ships were used for firing and recording where hydrophones floating in the sea water at ~50 feet depth and ~100 feet behind the ship recorded the vibrations of explosive charges of 4–300 lbs. The recording system had several channels to record both high-frequency water waves and lower-frequency ground waves (Ewing et al., 1950, 1954). The shot-receiver distance was calculated from the velocity of the water wave. More information on the techniques of seismic work at sea in the 1940s and 1950s has been given by Officer et al. (1952) and Shor (1963). There were also expeditions with one ship only. In this case, the ship was used for recording and a 20-foot whaleboat for shooting. At the same time, Russian scientists also undertook great efforts to investigate the deep structure under the oceans (e.g., Galperin and Kosminskaya, 1958, 1964; Neprochnov, 1960).

Ewing and Ewing (1959) described the experimental details of two-ship operations briefly as follows.

> Throughout the work, the seismic detectors were crystal hydrophones suspended at a depth of 30–50 m in the water. The output of each hydrophone was amplified, filtered, and recorded by a photographic oscillograph on the receiving ship [*see example in Figure 5.6.3-03*]. During the operations, the receiving ship remained hove to while the shooting ship fired a line of shots out to the required distance. The functions of the ships were then reversed, and the former receiving ship fired shots on a line up to the new receiving position. This procedure was referred to as "reversing the station." … All shots were fired with burning fuse, and the shot instant or "time break" was transmitted to the receiving ship by radio. In addition, events on both ships were related to absolute time by chronometers which were periodically checked against the radio station WWV.

There was no time to shoot short reverse profiles to determine the upper layers in addition to long profiles to determine the deeper layers. Additional deviation from an ideal reversed station resulted from the drift of the receiving ship.

Although the instrumentation of various research groups working in marine seismic experiments varied, all instruments were essentially echo sounders utilizing pulses of low frequency sonic energy and a graphic recording system that displayed the data in the form of cross sections. The sources were most commonly high-voltage sparkers, boomers, gas exploders, magnetostrictive transducers and explosives (Ewing and Zaunere, 1964). However, the majority of the sources was restricted to small offsets and could therefore only be applied for sedimentary structures. Only explosive charges were powerful enough to penetrate into the crystalline crust to Moho and therefore were used exclusively until the mid-1960s for scientific research.

To detect water-borne seismic signals, the hydrophone became the exclusively used receiver after WWII. They are generally made from a piezoelectrical material such as barium titanite, lead zirconate, or lead metaniobate. Small currents are generated in the transducer circuit by pressure changes in the water caused by a seismic source. These are amplified and transmitted by conventional conductor cable or radio link to display units or a data storage device. As hydrophones are small, they are essentially omnidirectional. In the 1940s and 1950s, single hydrophones were used and their signals directed to various channels, as described in detail by Hersey et al. (1952, see Chapter 4.3). Later hydrophones were grouped into array, which have important noise-reducing properties, including a directional response to seismic energy (Jones, 1999).

Already in 1937–1940, tests had been made to place seismographs on the ocean floor at depths exceeding 2000 m in several attempts and connected by an electric cable or wireless to a recording system, as was described in some detail in Chapter 3 (Ewing and Vine, 1938; Ewing et al., 1946; Ewing and Ewing, 1961). These tests were interrupted by World War II. The renewal of this effort was started in 1951 but proceeded slowly, primarily due to the lack of funding. Successful tests with the telemetering ocean-bottom seismograph could not be made before 1959 and were then continued into 1961. By modulation of a supersonic signal sent out from the ocean-bottom seismograph, the information was telemetered to a surface ship. The unit used on the ocean-bottom seismograph was a short-to-medium–period seismometer and an amplifier whose signals modulated the frequency of a 12 Hz acoustical source that sent signals in a broad beam toward the sea surface. The battery was adequate to operate the unit for 7–8 days. Successful tests were made at four positions at water depths between 3700 and 5600 m, shooting at various distances ranging from 15 to 55 km (Ewing and Ewing, 1961).

For the interpretation, first arrivals were picked and plotted into a time-distance graph. Arrivals were then connected by a straight line using a least-squares fit. Examples of travel-time plots have been reproduced for some experiments described in the following subchapters (see Figs. 5.6.3-02 and 5.6.3-05).

The reverse slope of the line would give the apparent velocity, and its backward prolongation to distance 0 would give the intercept time of the refracted wave, from which the depth of the "refracting" layer could be calculated. Ewing et al. (1954) and Ewing (1963a) have described in detail the theory for multilayer cases including inclined interfaces. Figure 5.6.1-01 shows the basic ray diagram, a sketch of the assumed structure.

Nafe and Drake (1957, 1963) have studied the dependence of the P-wave velocity in marine sediments from the thickness of the overburden and have derived corresponding velocity-depth functions using the results from the seismic-refraction measurements from the coastal shelf of eastern North America and the adjacent ocean basins. Having plotted average velocity-depth curves for the sediments in both environments, they found a pronounced difference in the dependence of the P-wave velocity in marine sediments from the thickness of the overburden; i.e., the velocity-depth relationship in shelf sediments differed distinctly from the one in deep-water sediments. They have also shown that, at the same depth of overburden, porosity is much greater in deep-water sediments than in shallow.

During the 1940s and 1950s, oceanographic research had made considerable progress, and publications had appeared in healthy numbers throughout the journals of the world. Therefore, by the end of the 1950s, it was suggested by scientists of the Scripps Institution of Oceanography to sum up the present knowledge in a comprehensive work. In 1963, three volumes of *The Sea* were published, all edited by M.N. Hill. Volume 1 dealt with "new thoughts and ideas on Physical Oceanography," volume 2 dealt with the "composition of sea-water and comparative and descriptive oceanography," and volume 3 contained "the Earth beneath the sea and history." The achievements of marine controlled-source seismic work were described in much detail in the first seven chapters of section 1, "Geophysical exploration," of volume 3 (p. 3–109): "Elementary theory of seismic refraction and reflection measurements" (Chapter 1 by J.I. Ewing); "Refraction and reflection techniques and procedure" (Chapter 2 by G.G. Shor Jr.), "Single ship seismic refraction shooting" (Chapter 3 by M.N. Hill), "Continuous reflection profiling" (Chapter 4 by J.B. Hersey), "The unconsolidated sediments" (Chapter 5 by J.I. Ewing and J.E. Nafe), "The crustal rocks" (Chapter 6 by R.W. Raitt), and "The mantle rocks" (Chapter 7 by J.I. Ewing). In 1970, a second comprehensive publication followed, published as volume 4 of *The Sea* and consisting of two voluminous parts, both edited by A.E. Maxwell: *The Sea, New Concepts of Sea Floor Evolution*.

In the following subchapters, we will discuss various marine expeditions in the Atlantic and Pacific Oceans, carried out in the 1950s. The term *station* was used in the corresponding publications as stationary position of the receiving ship, where it remained hove, while a second ship sailed off and fired a line of shots.

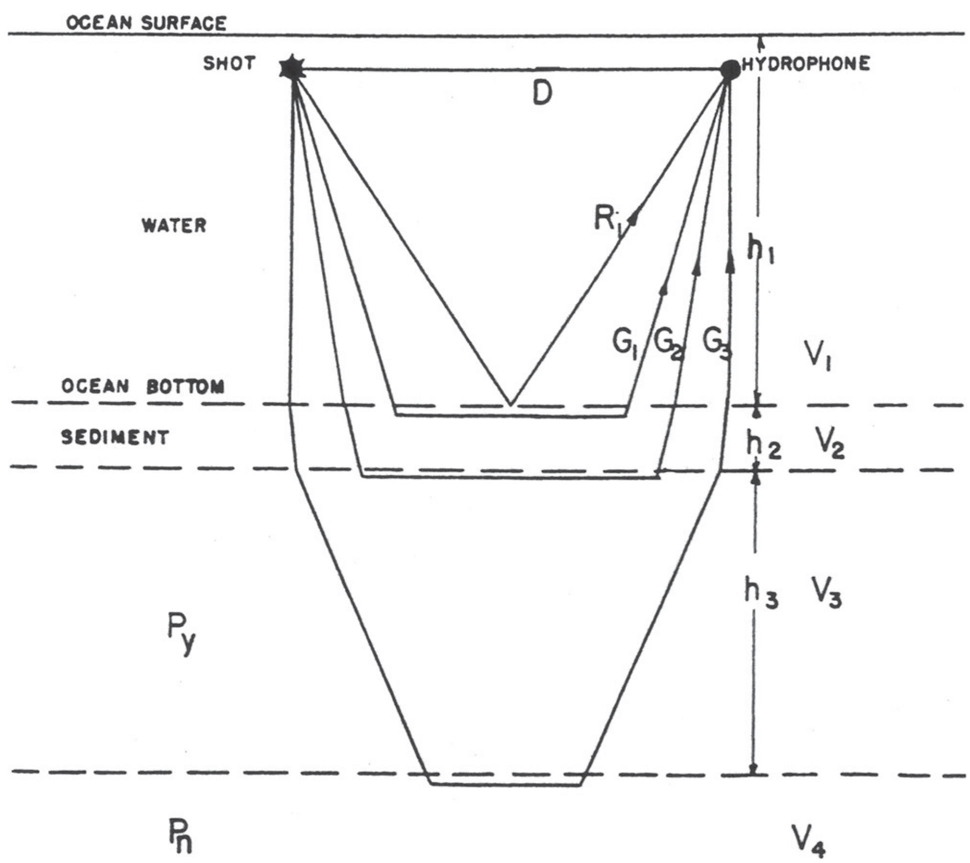


Figure 5.6.1-01. Ray diagram and basic structure derived from the seismic refraction measurements (from Ewing et al., 1954, fig. 2). [Bulletin of the Seismological Society of America, v. 44, p. 21–38. Reproduced by permission of the Seismological Society of America.]

5.6.2. Atlantic Ocean

Already at the end of the 1930s and again since the second half of the 1940s and in the following 1950s, seismic-refraction investigations of the Atlantic Ocean were undertaken in great numbers. Numbered publications were published in the *Bulletins* of the Geological and Seismological Societies of America, entitled "Seismic refraction measurements in the Atlantic Ocean" (e.g., Ewing et al., 1950, Part I; Tolstoy et al., 1953, Part III; Ewing et al., 1952, Part IV; Ewing et al., 1954, Part VI; Katz and Ewing, 1956, Part VII), "Geophysical investigations in the emerged and submerged Atlantic coastal plain" (e.g., Ewing et al., 1939, Part III; Ewing et al., 1940, Part IV; Officer and Ewing, 1954, Part VII; Press and Beckmann, 1954, Part VIII; Bentley and Worzel, 1956, Part X), or "Crustal structure and surface-wave dispersion" (e.g., Ewing and Press, 1952, Part II). British scientists were also heavily involved in marine projects to investigate the oceanic crustal structure applying the seismic-refraction method. Only a few of these many projects can be discussed in the following.

Officer et al. (1952) described an extensive two-ship refraction operation made in 1950. A total of 70 refraction profiles were recorded on tracks from Bermuda to Charleston, South Carolina; Norfolk, Virginia, to Bermuda; Bermuda to the Nares Deep, the deep basin south of the Bermuda rise, extending to the ridge just north of Puerto Rico trough; Bermuda to Halifax, Nova Scotia; Halifax to Sable Island; and Halifax to Woods Hole, Massachusetts. During 1951, another cruise obtained 33 profiles covering the Caribbean, the Puerto Rico trough, and tracks from Puerto Rico to Bermuda, and Bermuda to Woods Hole. The data were plotted on travel time graphs showing direct water-wave travel time in seconds as distance axis versus recorded travel time in seconds. On the profiles near the Nares Basin, four seismic layers were determined, unconsolidated sediments with 1.7 km/s, consolidated volcanics or sediments with 4.51 km/s, a basement with 6.63 km/s, and "a second basement" with 8.03 km/s average velocities, the Moho being at an average depth of 10 km. Over the Bermuda Rise only one high-velocity basement could be determined. The thickness of the volcanics decreased away from Bermuda from 4.5 km to 0.5 km.

The interpretation of six profiles in the North American Basin (Fig. 5.6.2-01), gathered by Ewing et al. (1954) in 1951, resulted in 3 layers: the unconsolidated sediments of 0.5–2 km thickness at the bottom of the ocean, the intermediate layer of 3–5 km thickness with an average velocity of 6.43 km/s and the mantle with an average velocity of 7.89 km/s. The Moho is ~10 km deep below water level (Ewing et al., 1954).

Katz and Ewing (1956) described the results of 25 seismic-refraction stations in the Atlantic Ocean west of Bermuda. They were occupied in 1950 and 1952 and their recordings were incorporated with other available data into four seismic cross sections through the oceanic crust and continental margins along tracts from Bermuda (65°W, 32°N) to the North American continent. From land measurements, Katz and Ewing (1956) inferred, that under the northeastern American continent, the velocity of near-surface rocks was close to 6 km/s and might increase with depth to near 7 km/s at ~35 km depth, with a mantle velocity of 7.7 km/s below. The upper continental crust (6.0–6.3 km/s) was inferred to thin seawards, disappearing near the foot of the continental slope. For the deep stations of their investigation (mean depth 4.9 km), the mean thickness of sediments was 1.3 km. The underlying crustal rocks had a mean thickness of 5.1 km, with velocities up to 7.1 km/s, but significantly lower near Bermuda. Arrivals from the mantle were observed on five of the longer stations, giving velocities of 7.7–8.5 km/s at 9.4–13.4 km below sea level. Subtracting the water column, the oceanic crust turned out to be less than one-fifth as thick as the continental crust. Toward Bermuda, the mean crustal velocity decreased, observed values ranged from 5.6 to 6.5 km/s. This decrease was associated with the structure of the Bermuda volcanoes.

Also in the early 1950s, during 1951, 1952, and 1953, seismic-refraction measurements were made on several cruises in the Atlantic Ocean as well as in the Mediterranean and Norwegian Seas (Fig. 5.6.2-02). They provided the first opportunity to make measurements in many parts of the ocean, and the data allowed a comparison of the crustal structure in these yet unexplored areas with better known regions. Ewing and Ewing (1959) reported on the results from 71 stations, of which 24 were in the western basins of the Atlantic, 6 in the eastern basins, 5 in the Mediterranean Sea, 19 on the Mid-Atlantic Ridge, 5 in the Norwegian and Greenland seas, 8 on the continental shelves of Britain and Norway and 4 were on the continental margin of eastern North America. As a general remark, the authors stated that both in western and the eastern ocean basins, they might have missed a masked 4.5–5.5 km/s layer. Figure 5.6.2-03 shows a typical travel time plot

Ewing and Ewing (1959) have published their results in a table showing for each station coordinates for the two reversing positions, water depth, and velocities and thicknesses for one to three sedimentary layers as well as for one or two layers of "high-velocity rocks" and the velocity for the Moho. For the western Atlantic Ocean basins, their average result was 5.02 km of water, 1.07 km of sediments with velocities ranging from 1.7 to 4 km/s, 6.15 km depth to and 5.16 km thickness of the oceanic layer with a mean velocity of 6.71 km/s, and 11.25 km depth to the mantle with 8.08 km/s mean velocity. Data example records from a station in the western Atlantic Ocean are shown in Figure 5.6.2-04.

In the eastern North Atlantic Ocean Basin, it was difficult to obtain a good average crustal section because of the scarcity of long reversed stations. Ewing and Ewing (1959) have therefore combined their results with those obtained in earlier operations (Hill, 1952; Hill and Laughton, 1954) and thus came to the following average result: 4.56 km of water, 1.18 km of sediments, 4.87 km thickness of oceanic layer with a mean velocity of 6.52 km/s, and a mantle velocity of 7.81 km/s. The only difference Ewing and Ewing (1959) observed versus the larger basins in the west was that the mantle velocity of 7.81 km/s was less than elsewhere.

The seismic measurements over the Mid-Atlantic Ridge were difficult on account of the great topographic relief, but definite

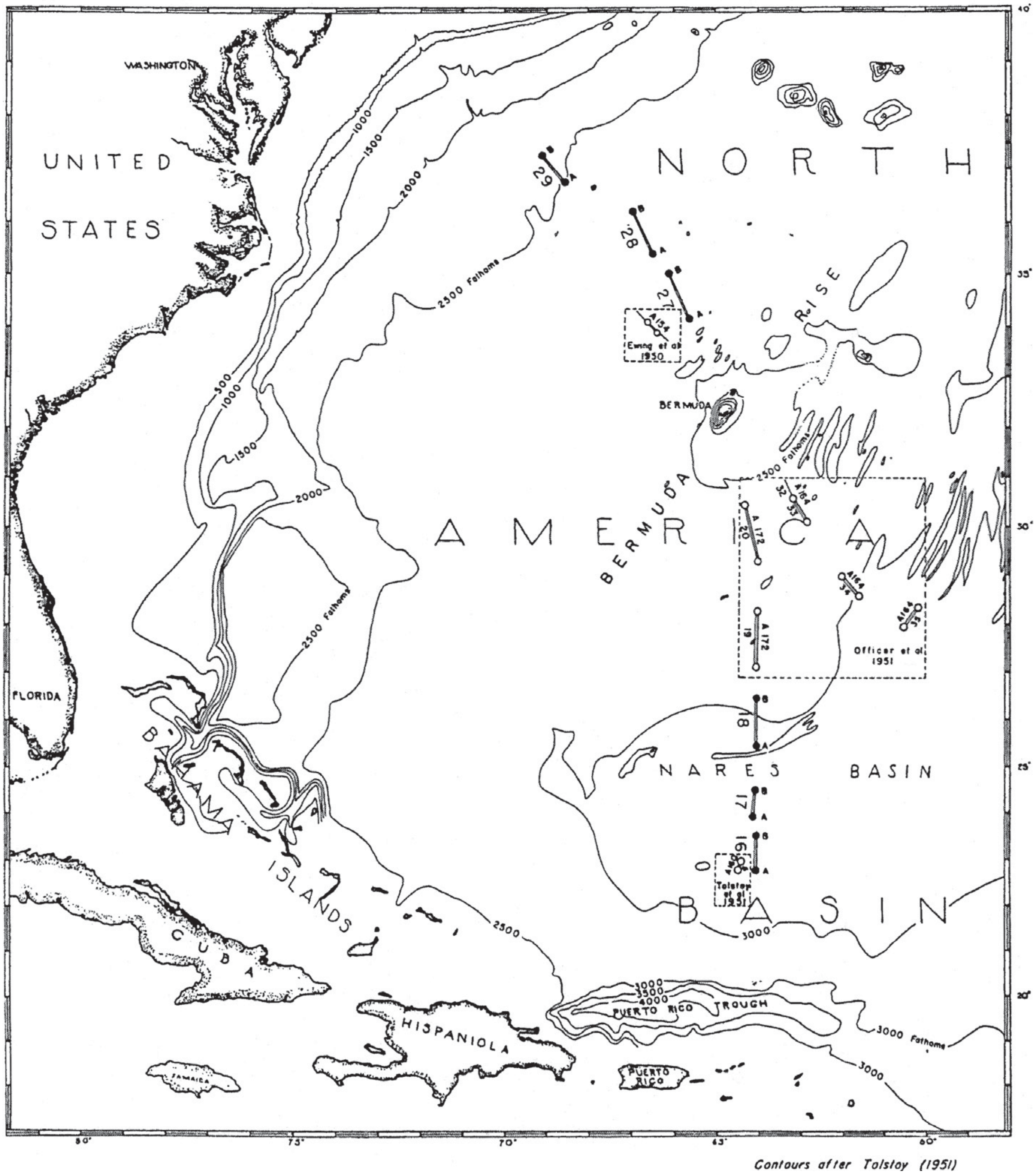


Figure 5.6.2-01. Geographic location of seismic profiles recorded in 1951 in the North American Basin (from Ewing et al., 1954, fig. 1). [Bulletin of the Seismological Society of America, v. 44, p. 21–38. Reproduced by permission of the Seismological Society of America.]

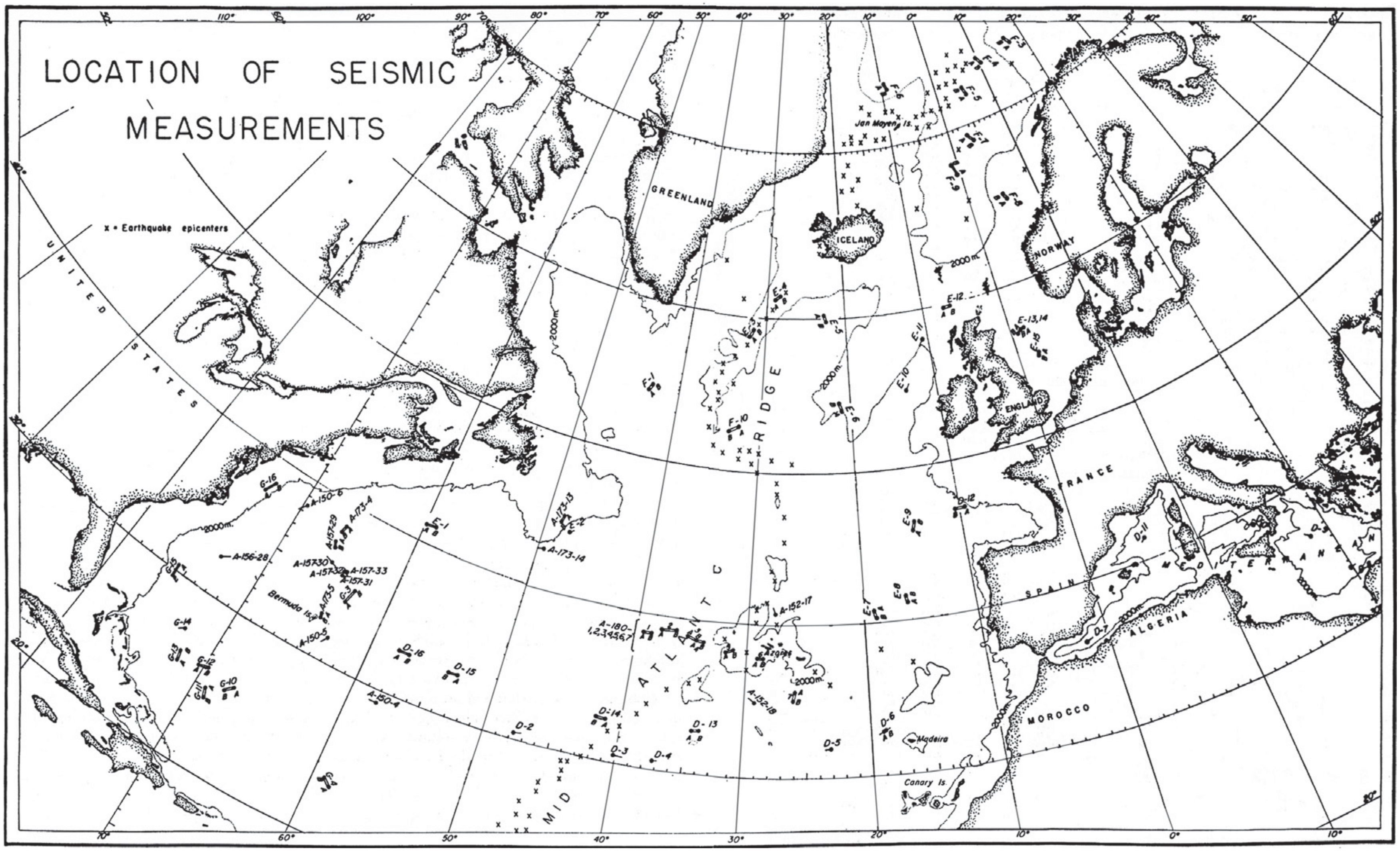


Figure 5.6.2-02. Geographic location of seismic profiles recorded from 1951 to 1953 in the North Atlantic Ocean, Mediterranean and Norwegian Sea (from Ewing and Ewing, 1959, fig. 1). [Bulletin of the Geological Society of America, v. 70, p. 291–318. Reproduced by permission of the Geological Society of America.]

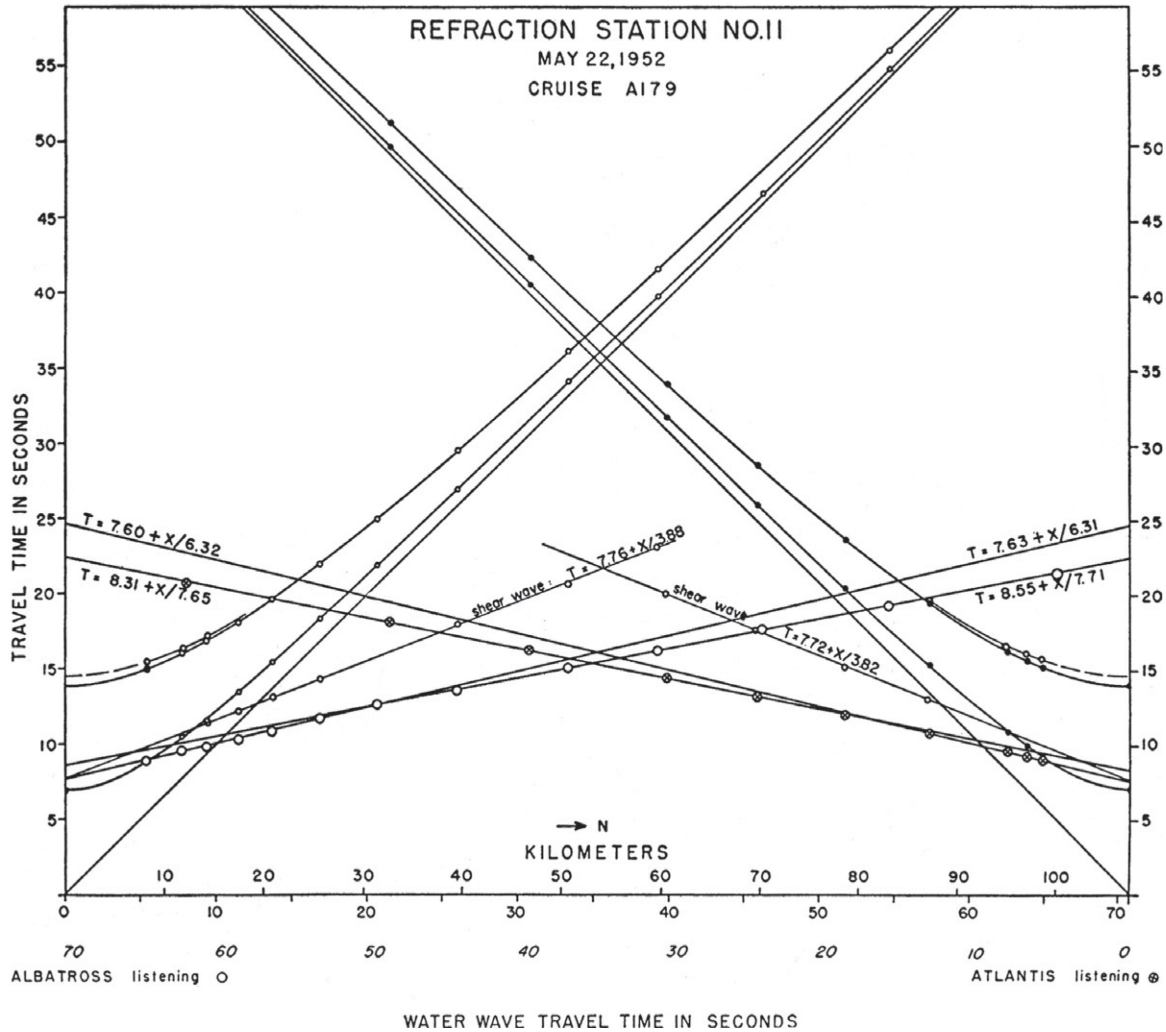


Figure 5.6.2-03. Travel-time plot of a seismic profile recorded in the North Atlantic Ocean (from Katz and Ewing, 1958, fig. 20). The distance is shown as water wave traveltime in seconds. [Bulletin of the Geological Society of America, v. 67, p. 475–510. Reproduced by permission of the Geological Society of America.]

results could nevertheless be obtained. The crustal layering turned out to differ significantly from the ocean basins (Figs. 5.6.2-05 to Fig. 5.6.2-07). The upper layer of the ridge with a velocity of 5.15 km/s was identified as basaltic volcanic rock, whose value agreed well with values given by Birch et al. (1942) and which also had been measured on Bermuda (Officer et al., 1952, see above). The deeper layer with average velocity of 7.21 km/s was tentatively identified as mixture of mantle and oceanic layer. Ewing and Ewing (1959) speculated that the formation of the ridge had required the addition of a great quantity of basalt magma which again raised the question of its source. As already described by Hess (1954), this was attributed to a rising convection current in the mantle which caused extensional forces to the crust to produce the axial rift.

Eight stations could be occupied in the Mediterranean Sea: two near the coast of Cyprus, two near Malta, three in deep water in the eastern basin, and one in shallow water southwest of Malta. At the deep-water stations, only one refracting layer (4.42–4.57 km/s) was found underlying low-velocity sediments. The shallow station showed a wider range of refracting velocities from 2.96 to 6.70 km/s. The other stations corresponded to the values found in this area by Gaskell and Swallow (1953) in their cruise in 1952.

In the Norwegian and Greenland Seas, only five deep-water and two shallow-water stations were occupied. In spite of the small number of experiments, the results were reasonably consistent and therefore appeared justified. The Norwegian Sea (Fig. 5.6.2-08) differed from that found usually under deeper oceanic areas. The upper high-velocity layer was fairly consistent in velocity (4.96–5.37 km/s) and thickness (2.5–3.0 km). The velocity in the deepest layer varied more (6.97–8.04 km/s). The depth of this layer below sea level was quite consistently 7 km. On the continental shelf and slope of Britain and Scandinavia, five stations were occupied with another three in the North Sea. Previous seismic work on the continental shelf and slope of the British

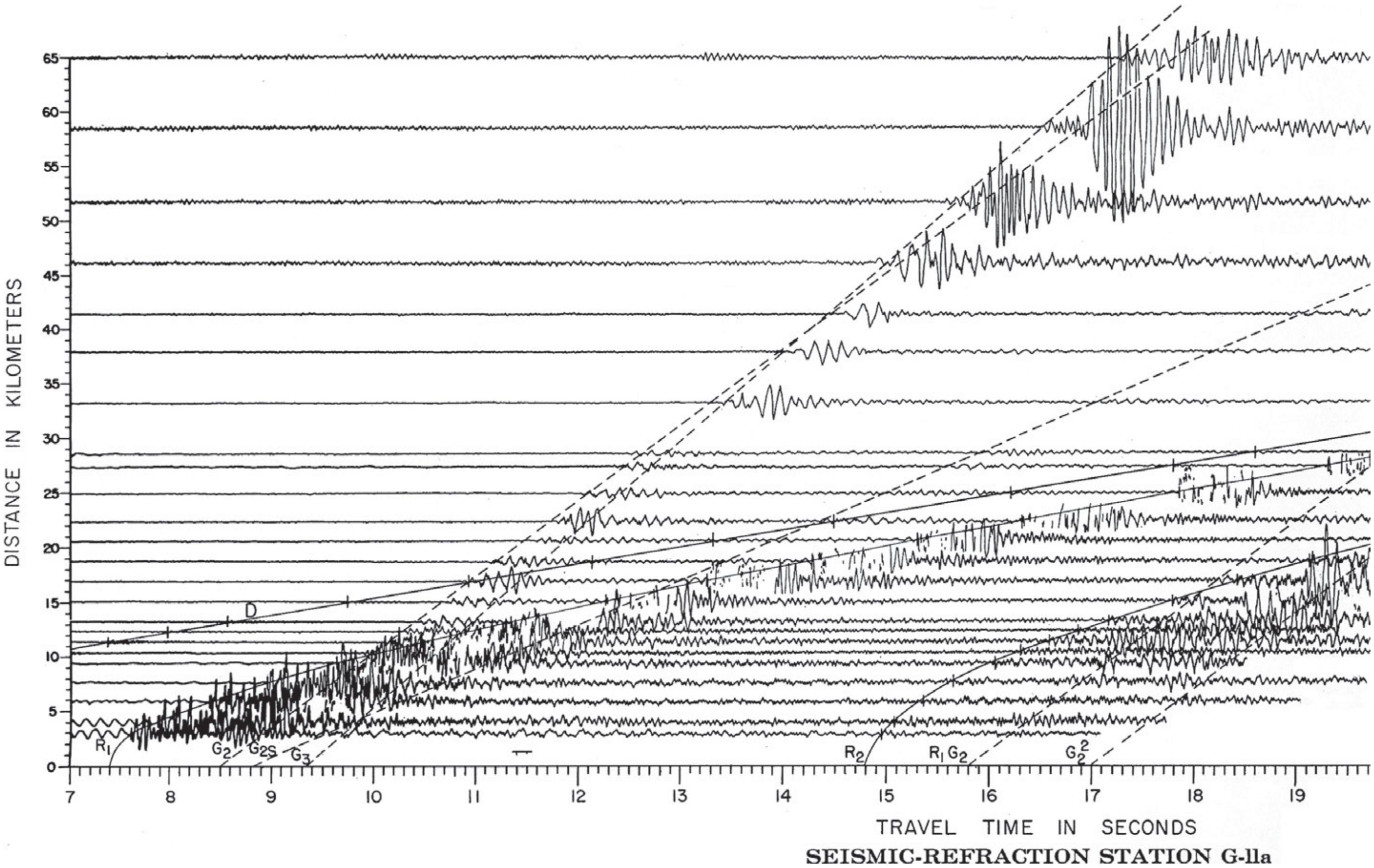


Figure 5.6.2-04. Record section of a seismic profile recorded in the North Atlantic Ocean (from Ewing and Ewing, 1959, pl. 1) [Bulletin of the Geological Society of America, v. 70, p. 291–318. Reproduced by permission of the Geological Society of America.]

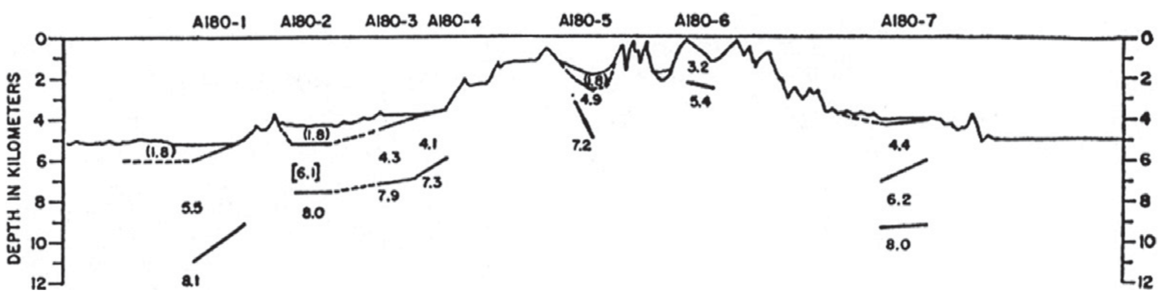


Figure 5.6.2-05. Topographic profile and seismic section across the Mid-Atlantic Ridge south of the Azores (from Ewing and Ewing, 1959, fig. 2). [Bulletin of the Geological Society of America, v. 70, p. 291–318. Reproduced by permission of the Geological Society of America.]

Isles had been reported in earlier years by Bullard et al. (1940), Bullard and Gaskell (1941), Hill and King (1953), and Hill and Laughton (1954). Ewing and Ewing (1959) reported that the velocities in unconsolidated sediments could be reasonably well determined, averaging 1.7 km/s. In all stations, they also found a semi-consolidated and a consolidated layer with velocities of 2.2 and 3.3 km/s, respectively. For the basement, they found an average of 5.11 km/s. The velocities in the central North Sea were about the same as at the shelf stations, but basement with depths greater than 4 km was not reached.

Finally Ewing and Ewing (1959) made measurements on the continental slope of North America, southeast of Cape May, New Jersey (Fig. 5.6.2-02), and near the southeastern tip of the Grand Banks. Their results were in general agreement with those of other investigations in this area. Their table again distinguishes unconsolidated, semi-consolidated and consolidated sediments

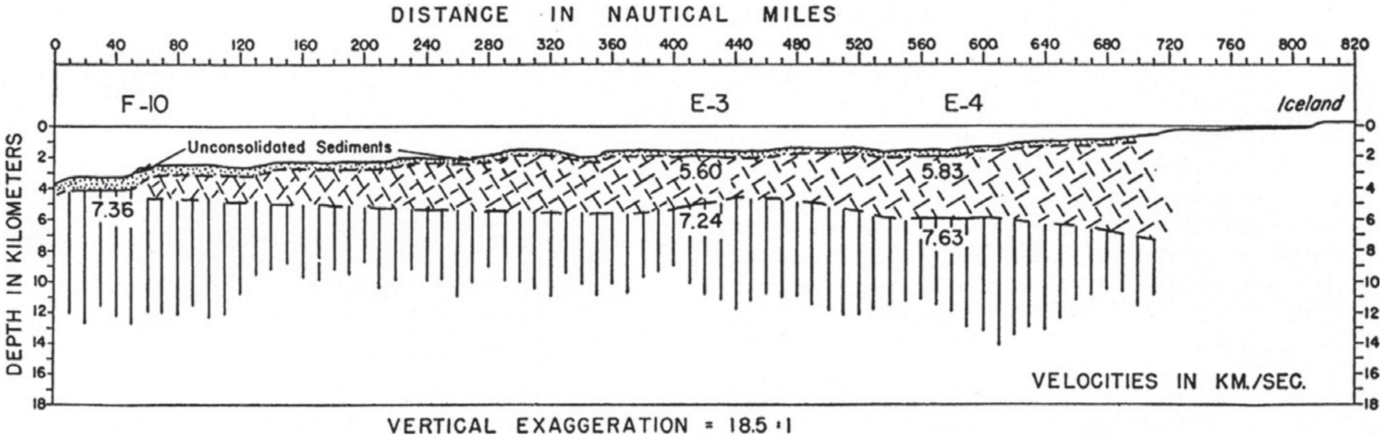


Figure 5.6.2-06. Structure section along the Mid-Atlantic Ridge south of Iceland (from Ewing and Ewing, 1959, fig. 4). [Bulletin of the Geological Society of America, v. 70, p. 291–318. Reproduced by permission of the Geological Society of America.]

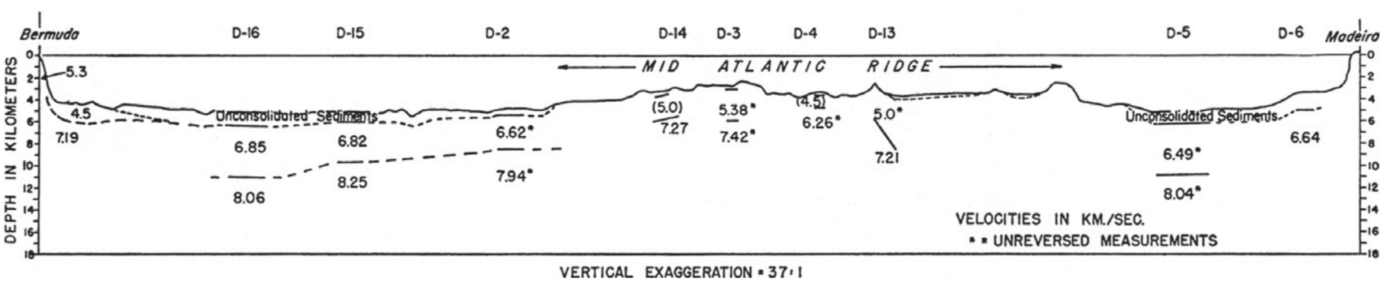


Figure 5.6.2-07. Structure section between Bermuda and Madeira (from Ewing and Ewing, 1959, fig. 5). [Bulletin of the Geological Society of America, v. 70, p. 291–318. Reproduced by permission of the Geological Society of America.]

with average velocities of 1.7, 2.4, and 3.8 km/s and thicknesses of 0.6, 1.9, and 2.1 km. For the basement, they determined a velocity of 5.9 km/s.

Using data from 22 seismic-refraction stations over the northern Mid-Atlantic Ridge and combining them with earlier data, Le Pichon et al. (1965) deduced a refined model of the crustal structure of the Mid-Atlantic Ridge in the North Atlantic Ocean. In the axial zone, they found a 5.8 km/s layer overlying a 7.3 km/s layer. They concluded that the seismic crustal structure is the same at equal water depths and does not thicken at the flanks of the northern Mid-Atlantic Ridge when water depth decreases, but is an uplifted ocean basin crust. In order for gravity and seismic results to agree, Talwani et al. (1965) computed fitting models postulating that the anomalous upper mantle under the ridge extends to the flanks below the normal mantle. They compared three different areas of the North Atlantic Ocean with sections from the East Pacific Rise and concluded that the northern Mid-Atlantic Ridge could be assumed to be in a later stage of evolution, the thickness and velocity of the basement layers being considered the best indicators of the degree of maturation of the mid-ocean ridges.

The first experiments of Ewing and coworkers were followed by many others. From the large amount of investigations in the Atlantic Ocean (e.g., Katz and Ewing, 1956; Ewing and Ewing, 1959; Officer et al., 1959), an overall structure of the consolidated oceanic crust could be derived. The crust under the ocean basin consisted on average of a 1.7-km-thick layer 2 with velocity 5.07 km/s and a 4.9-km-thick layer 3 with velocity 6.7 km/s, underlain by mantle rocks with an average velocity of 8.13 km/s at the top. A detailed overview and summary of the results has been given by Raitt (1963) for the crust obtained by the application of the seismic-refraction method in the 1950s, listing the individual experiments in tables.

In 1953, Officer (1955a) reported on a series of seismic-reflection measurements on the abyssal plain in the NW Atlantic Ocean north of Bermuda, made on a cruise from Woods Hole, Massachusetts, to Bermuda and return. Similar experiments were also reported by Ewing and Nafe (1963). The method of continuous reflection profiling was discussed by Hersey (1963) in detail and compared with the seismic-refraction method, but at that time, the reflection method did not result in data reaching into the crystalline crust below the sedimentary cover.

From 1954 to 1956, the eastern continental margin of the United States, the intervening Blake plateau and adjacent deep-water areas of the Atlantic Ocean was investigated by 40 seismic-refraction profiles (Hersey et al., 1959), the sur-

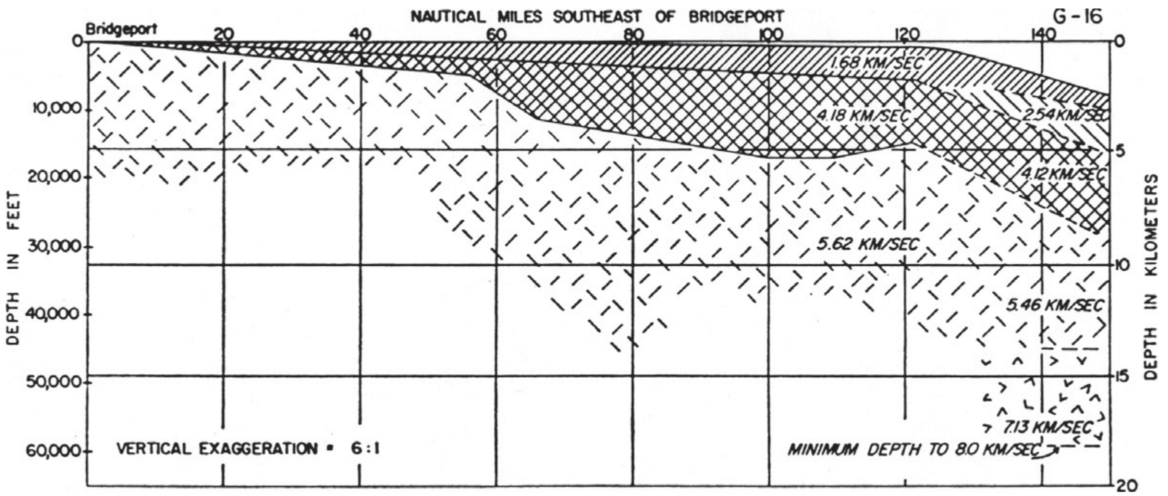


Figure 5.6.2-08. Structure section in the Norwegian Sea (from Ewing and Ewing, 1959, fig. 6). [Bulletin of the Geological Society of America, v. 70, p. 291–318. Reproduced by permission of the Geological Society of America.]

veys extending from 29°39′ to 36°30′N latitude and 73°30′ to 81°10′W longitude (Fig. 5.6.2-09). The results on the continental shelf could be correlated with adjacent continental geology. The deepest horizon traced along the shelf was interpreted as granitic basement with P-wave velocities of 5.82–6.1 km/s. At the southernmost profile near Jacksonville the greatest depth of 6 km to basement was found, toward north its surface rose to as low as 0.9 km near Cape Fear and deepened again north of Cape Hatteras to 3 km depth. North of Charleston, South Carolina, the measured depth could well be correlated with the granitic basement in a coastal borehole. On the Blake Plateau, a smooth and nearly flat area extending north from the Bahamas (located south of point D′ in Figure 5.6.2-09) to Cape Hatteras, several sedimentary layers with velocities from 1.83 to 4.5 km/s were found to overlie the crystalline crust with a 5.5 km/s and 6.2 km/s layer. Higher velocities, 8.0 km/s, as well as 7.3 km/s, were found at markedly different depths. The deep-water area is a continental slope and rise modified by the Blake Plateau and by a ridge trending southeastward from Cape Fear and deepening from ~2700 m to more than 3700 m (1500–2000 fathoms contours in Figure 5.6.2-09).

The ridge was found to be underlain by thick low-velocity layers, interpreted as sediments, and higher-velocity layers which formed a linear structure and had the same trend as the ridge. South of the ridge, the profiles were similar to those in ocean basins (Hersey et al., 1959). The cross section E–E′ shown in Figure 5.6.2-10 starts near Jacksonville, Florida, and crosses continental shelf and Blake Plateau and ends in the deep-water area near 77°W. The cross section H–H′ of Figure 5.6.2-11 follows the approximate continuation of cross section E–E′ to the northeast into the deep-water area and runs about parallel to the continental slope.

Geophysical measurements were also made in the western Caribbean Sea and in the Gulf of Mexico (Fig. 5.6.2-12; Ewing et al., 1960; Antoine and Ewing, 1963). For the center of the Gulf of Mexico, a typical oceanic crust was obtained with Moho depths between 16 and 19 km below sea level. Toward the western Caribbean Sea, Moho depths varied considerably, indicating an interfering of oceanic and continental crust (Fig. 5.6.2-13).

Another detailed survey to determine the crustal structure of the Caribbean Sea was undertaken in 1957 (Edgar et al., 1971). In total, 30 seismic-refraction profiles were measured. Some of them were partly aligned to longer lines, one across the ridge between the Colombia Basin to the west and the Venezuela Basin to the east, while another line was laid out parallel to coast of Venezuela and various scattered profiles were measured in the Venezuela Basin and around the Lesser Antilles islands. The results were summarized in five cross sections. Particular attention was given to areas with complex structures such as Tobago Island, the Curacao Ridge, the Smaller Antilles (called "Netherlands West Indies" by the authors) and others. The main result of the detailed survey was to map the velocity structure of the sedimentary layers and the depth to basement.

A cross section drawn from Cuba to Curacao shows, underneath several sedimentary layers, an ~2–5-km-thick basement at depths varying between 5 and 10 km, a rather thick lower crust with velocities of 6.5–6.8 km/s and a Moho depth near 20 km. The measured upper-mantle velocities varied from 7.9 to 8.5 km/s. On other profiles the crust-mantle boundary was rarely seen. Along the Venezuelan coast only the uppermost basement was reached. Other short sections in complex areas show complex structures. For example under the Curacao Ridge a deeper discontinuity was found near 15 km separating a 4.0–4.4 km/s basement from 6.7 to 7.0 km/s lower crust.

Based on gravity and seismic results, Talwani et al. (1959) published a cross section across the Puerto Rico Trench. The section started in the Nares Basin, 450 km north of San Juan, to the Venezuela Basin, 250 km south of San Juan. The depth to

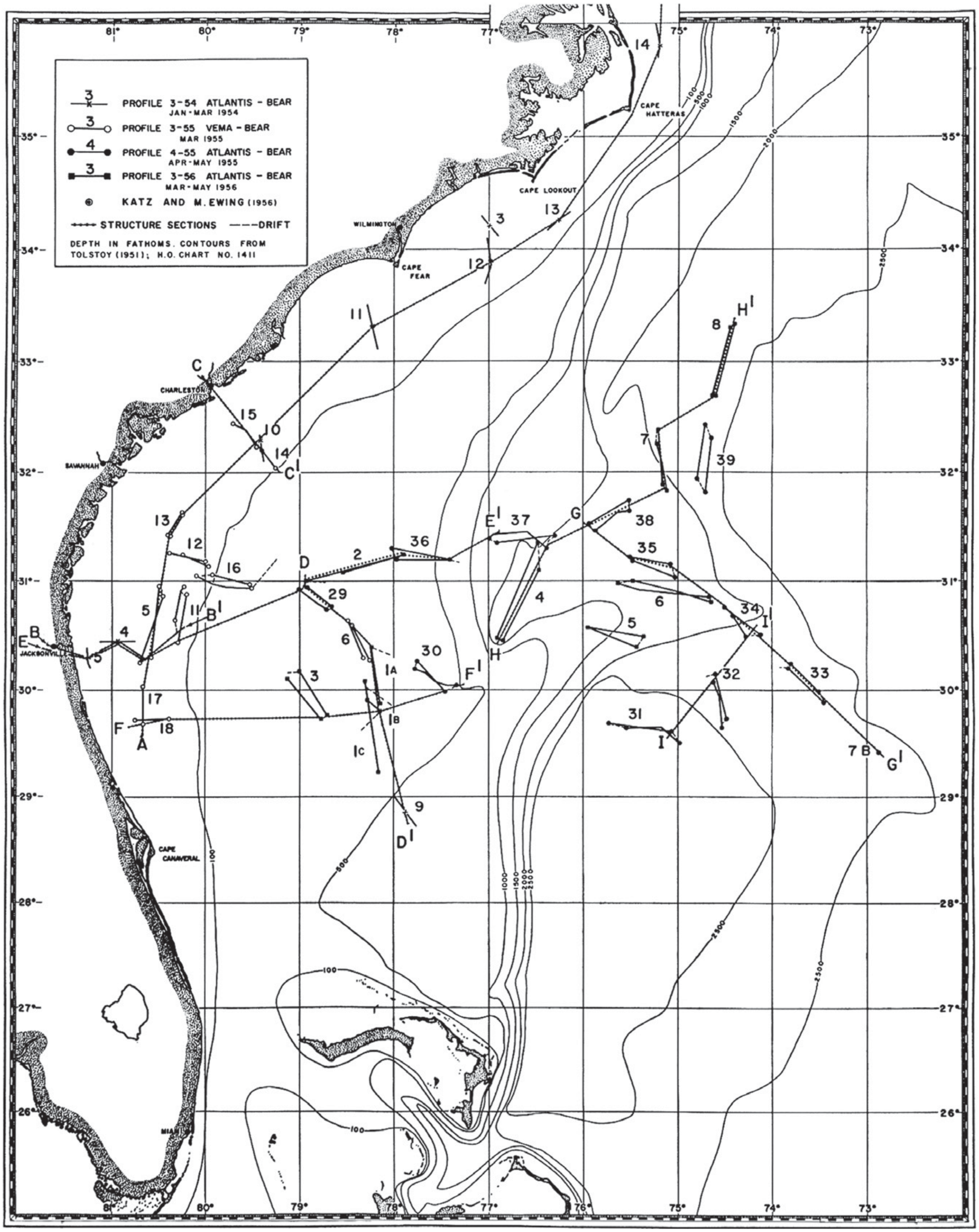


Figure 5.6.2-09. Location of seismic profiles recorded from 1954 to 1956 in the western Atlantic Ocean to the southeast of the United States (from Hersey et al., 1959, pl. 1). [Bulletin of the Geological Society of America, v. 70, p. 437–466. Reproduced by permission of the Geological Society of America.]

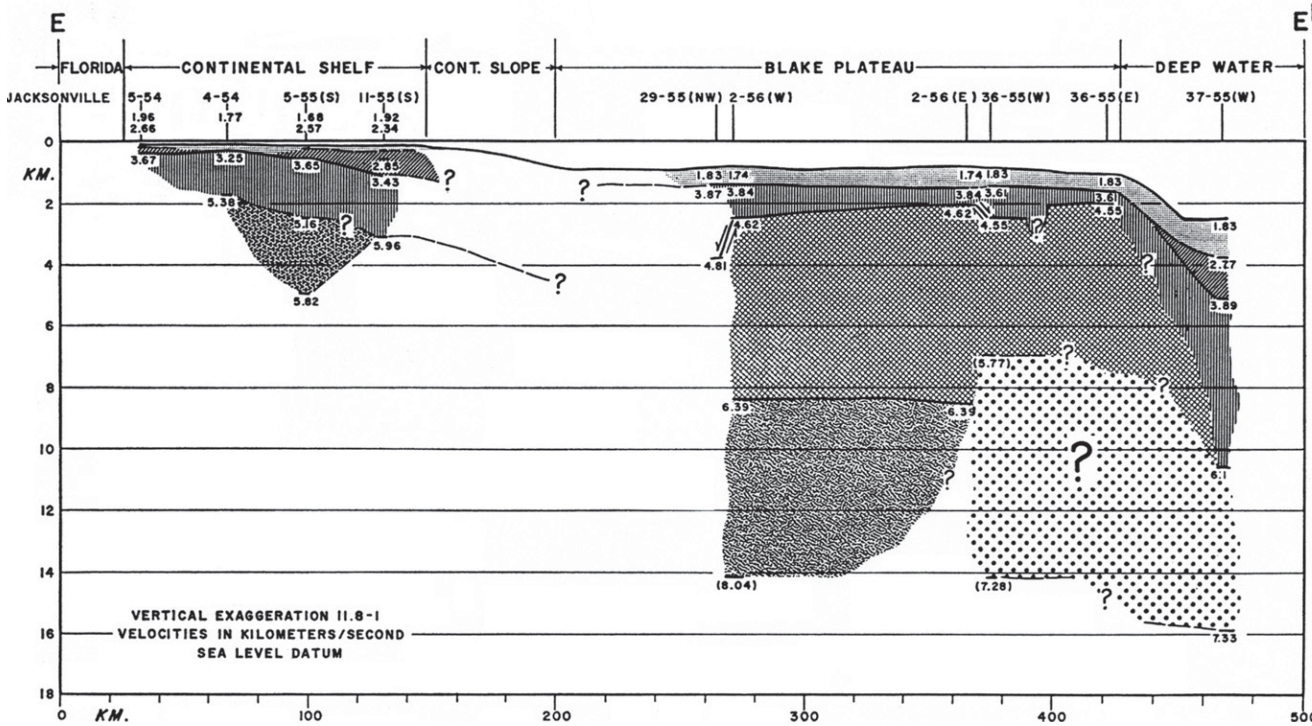


Figure 5.6.2-10. Cross section (E–E′ in Figure 5.6.2-09) from Jacksonville, Florida (81°30′W), across the continental slope and Blake Plateau into the deep-water area near 77°W (from Hersey et al., 1959, fig. 7). [Bulletin of the Geological Society of America, v. 70, p. 437–466. Reproduced by permission of the Geological Society of America.]

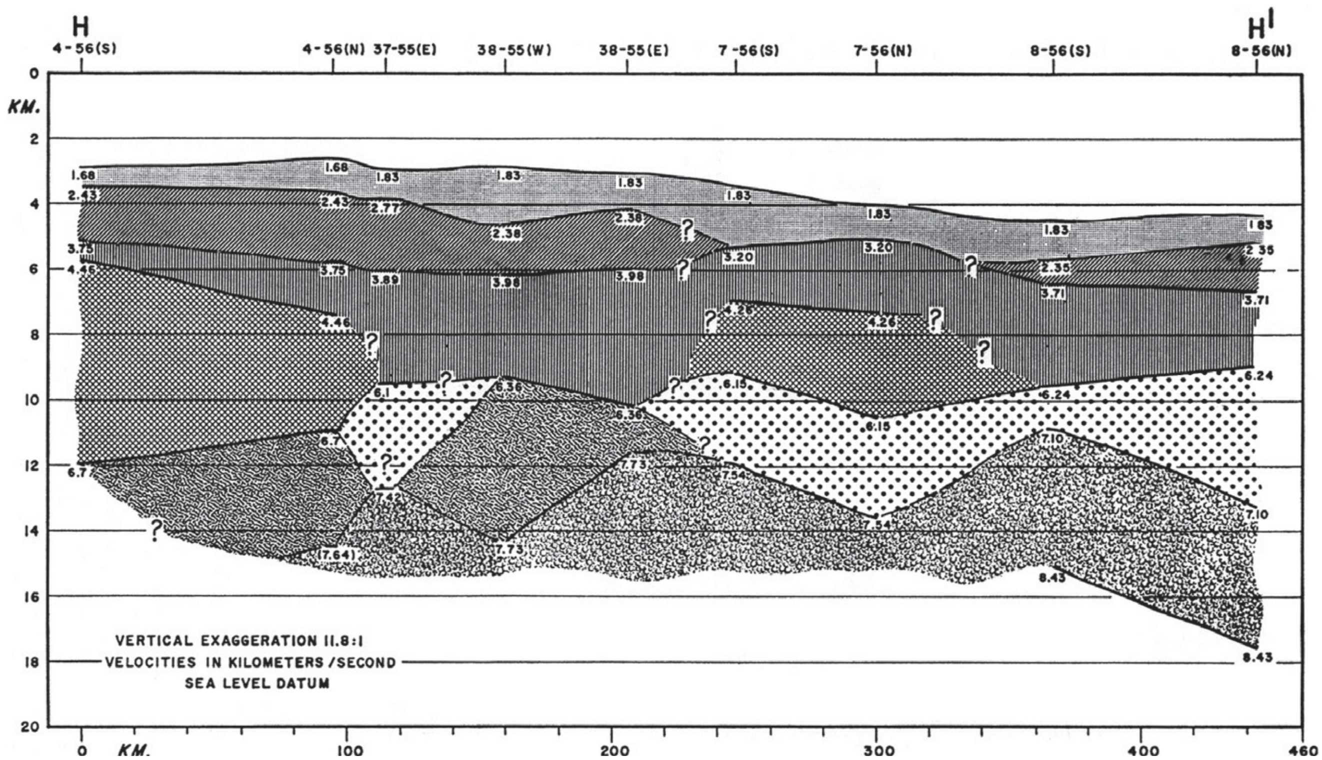


Figure 5.6.2-11. Cross section (H–H′ in Figure 5.6.2-09) through the deep-water area parallel to the continental slope in SW-NE direction (from Hersey et al., 1959, fig. 10). [Bulletin of the Geological Society of America, v. 70, p. 437–466. Reproduced by permission of the Geological Society of America.]

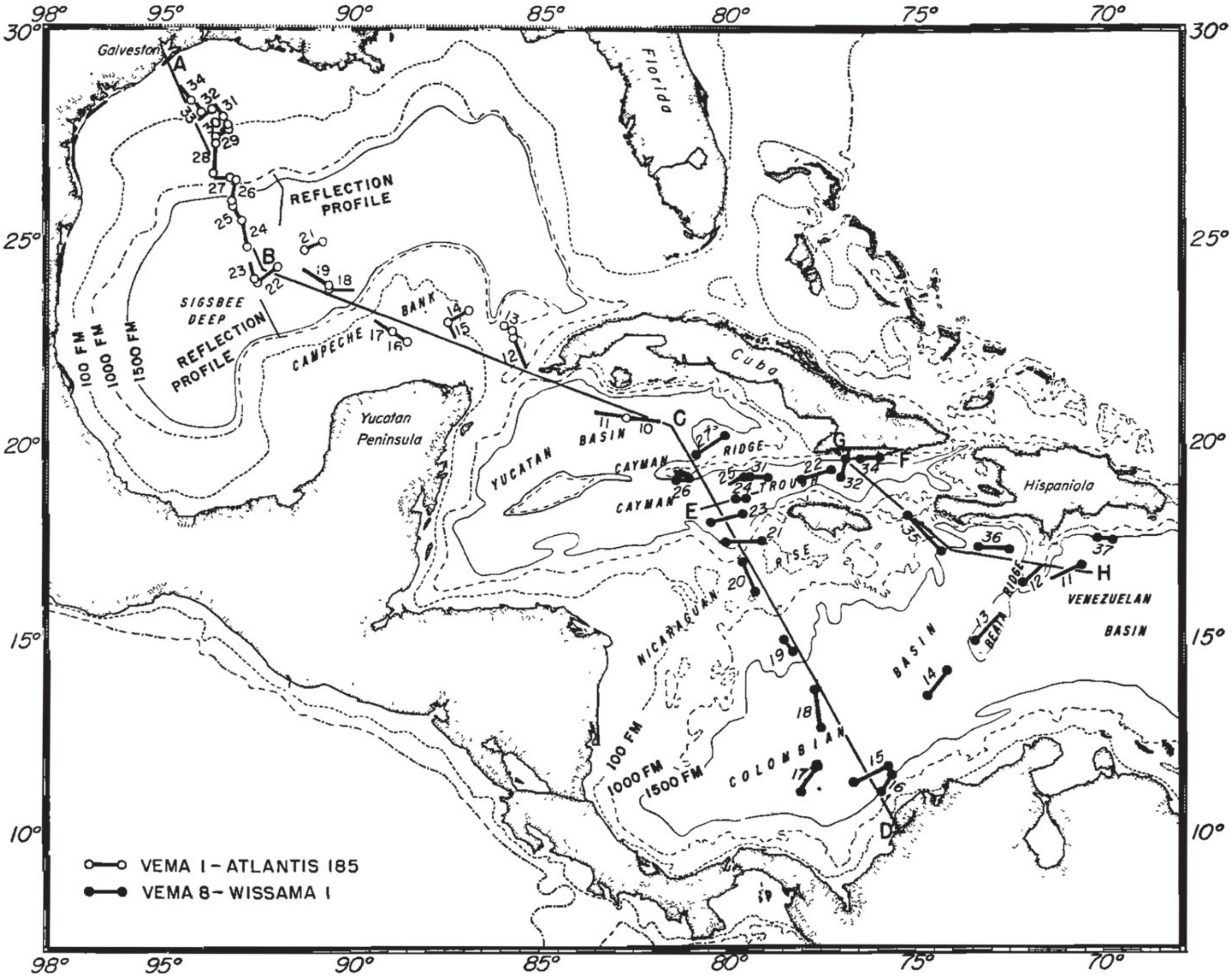


Figure 5.6.2-12. Location of seismic profiles recorded in the late 1950s in the western Caribbean Sea and the Gulf of Mexico (from Ewing et al., 1960, fig. 1) [Journal of Geophysical Research, v. 65, p. 4087–4126. Reproduced by permission of American Geophysical Union.]

Moho was ~20 km under the trench and decreased sharply to both sides. Northwards, Moho reached a minimum depth of 10 km under the Outer Ridge and then deepened gradually to ~13 km under the Nares Basin. To the south, Moho reached a depth of 17 km under the Puerto Rico shelf and then increased to 30 km beneath Puerto Rico. South of Puerto Rico under the Venezuela Basin, a depth of 14 km was obtained.

The first seismic-refraction investigation around the Mid-Atlantic Ridge in Iceland was undertaken by the Seismological Laboratory of Uppsala, Sweden, in cooperation with Icelandic earth scientists (Båth, 1960). The instrument used was a complete 12-channel refraction apparatus of Swedish construction with electromagnetic seismometers of 6 cps. The seismic signals were recorded via amplifier and oscillograph on paper of 150 mm width. The time marks were produced by a synchronous motor driven by an oscillator which was controlled by a temperature-compensated tuning fork. Fourteen galvanometers recorded in parallel. A series of explosions was set off in a lake with 45 m water depth in the southeast of Iceland. The break off of a tone when firing the shot was transmitted by telephone to the radio station of Reykjavik and from there broadcasted in the radio between or after their programs.

Two lines were recorded: a NNE line of 253 km length and a NE line of 337 km length, the distance between stations increasing with increasing distance from the shotpoint. In total, 19 shots were fired resulting in a total of 19 spreads. Due to strong absorption of energy in central Iceland, however, 150 km was the maximum distance on the NE line with recordable energy. A three-layer model resulted from these observations (Fig. 5.6.2-14) with a total crustal thickness of 27.8 km.

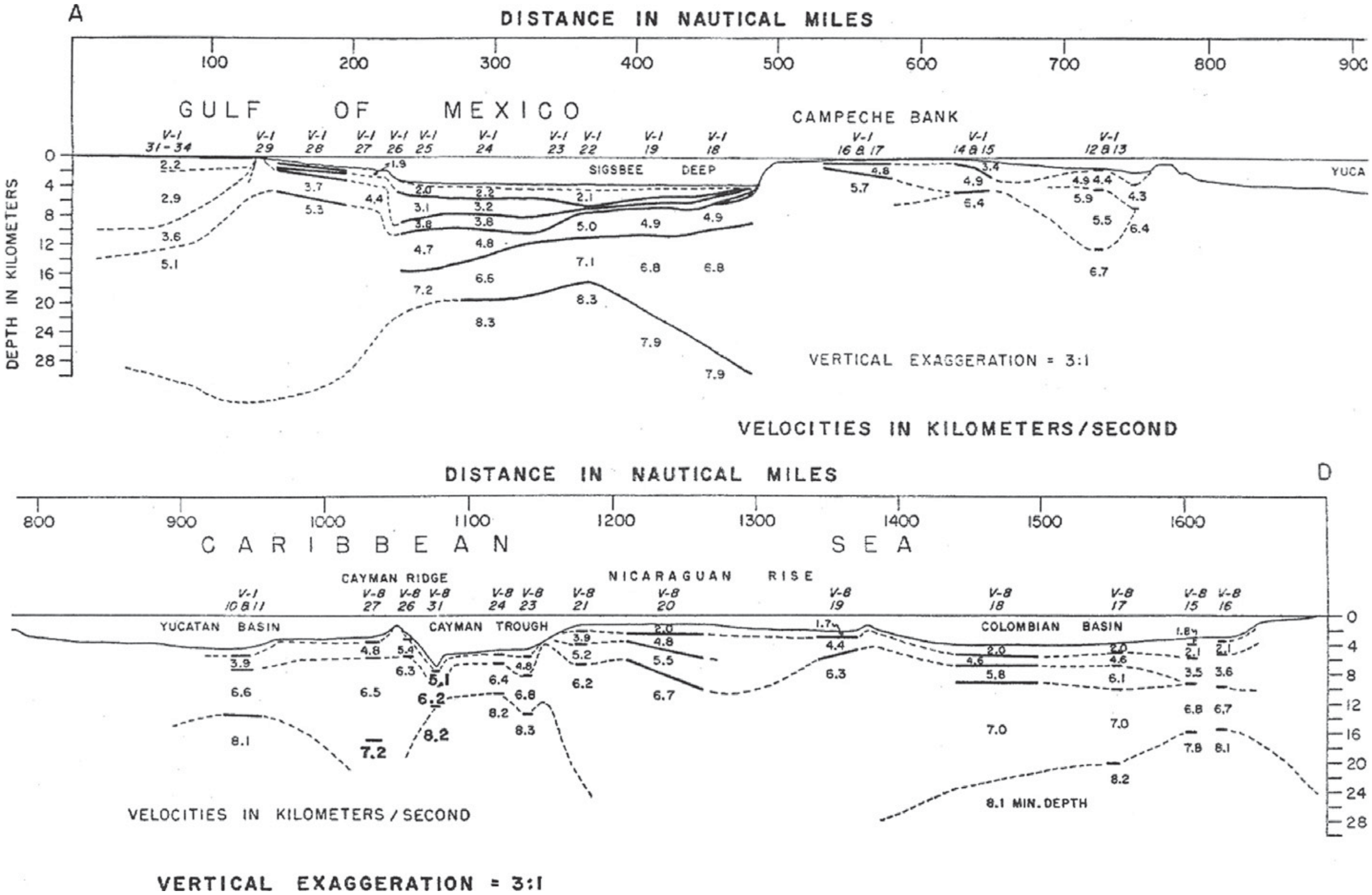
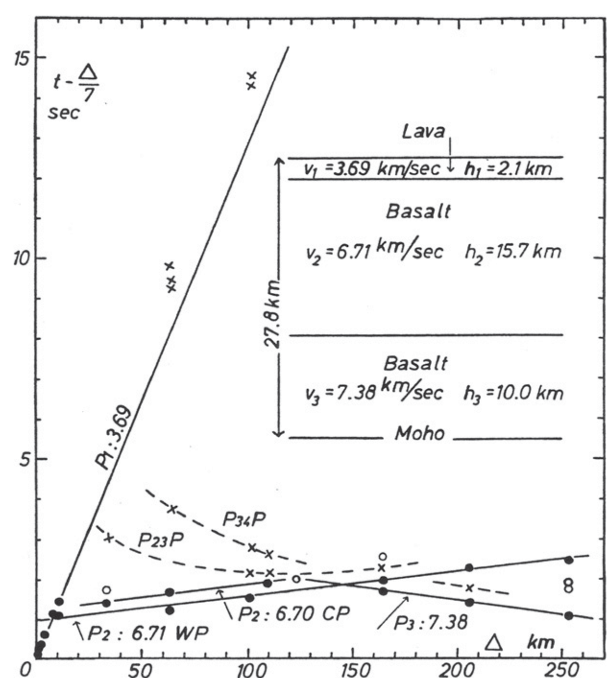


Figure 5.6.2-13. Crustal cross section through the Gulf of Mexico and western Caribbean Sea along line A–D (see Figure 5.6.2-12) recorded in 1951 (from Ewing et al., 1960, fig. 1). [Journal of Geophysical Research, v. 65, p. 4087–4126. Reproduced by permission of American Geophysical Union.]

Figure 5.6.2-14. Traveltime curves and crustal model of western Iceland (from Båth, 1960, fig. 6). [Journal of Geophysical Research, v. 65, p. 1793–1807. Reproduced by permission of American Geophysical Union.]



In the 1950s, expeditions into the Arctic were also undertaken to study the Earth's crust. The French Polar expeditions of 1949 and 1951 (Joset and Holtzscherer, 1953) had the purpose of investigating the inland ice of Greenland. Of particular interest were the dependence of the velocity in ice on the temperature and the velocity increase within the uppermost ice layers.

Katz and Ewing (1956) have made a statistical evaluation of velocities measured in the Atlantic Ocean crust at water depths exceeding 300 m (Fig. 5.6.2-15). Their frequency diagram clearly separated sediments with velocities around 2.4 km/s and transitional rocks with 4.4 km/s and showed peaks for crustal rocks at 5.5 and 6.3–6.8 km/s. Mantle rocks were defined from velocities of 7.4 km/s upward with an average of 8.1 km/s.

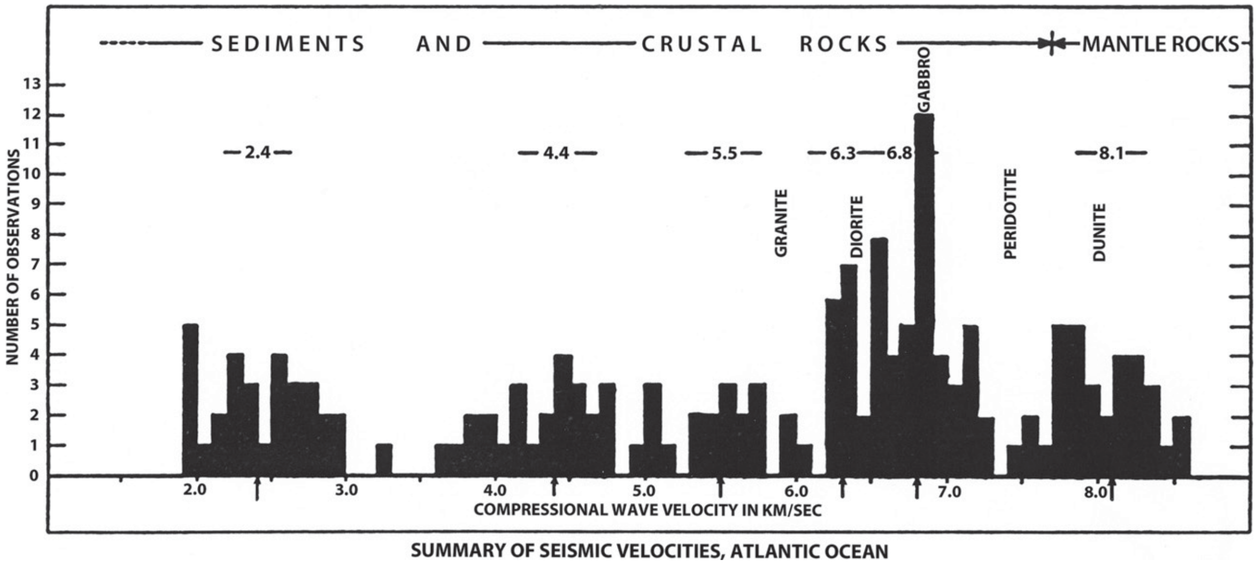


Figure 5.6.2-15. Frequency diagram of seismic velocities in the Atlantic Ocean, exceeding water depths of 300 m (from Katz and Ewing, 1956, fig. 25). [Bulletin of the Geological Society of America, v. 67, p. 475–510. Reproduced by permission of the Geological Society of America.]

5.6.3. Pacific and Indian Oceans

Other marine expeditions investigated the Pacific and Indian Oceans (e.g., Raitt, 1956; Gaskell et al., 1958). In the early 1950s, two major expeditions were undertaken into the Central Equatorial Pacific (Fig. 5.6.3-01): in 1950 the Mid-Pacific expedition explored the Pacific north of the equator, while in 1952 and 1953, the Capricorn expedition aimed for the Pacific between 10° and 20°S. At all stations two ships were used, following the methodology described by Officer et al. (1952) and briefly repeated in Chapter 4. Examples of a time-distance plot and of an oscillogram are shown in Figures 5.6.3-02 and 5.6.3-03.

In total, seismic-refraction observations were made at 42 positions, scattered widely, extending from latitudes 22°S to 28°N and longitudes 162°E to 112°W. At 29 of these stations, velocities of ~8 km/s or more were reached at the depth of greatest penetration of the refracted waves. The mean velocity was determined to 8.24 km/s. The crustal thickness was defined as the depth below the seafloor at which 8 km/s velocity was reached. It ranged from 4.8 to 13.0 km. The group of stations which was regarded as typical of the deep Pacific Basin had an average thickness of 6.31 km with a standard deviation of 1.01 km (Raitt, 1956).

One of the special goals of the Capricorn expedition of 1952–1953 was two weeks of research work in the Tonga Trench and archipelago (Raitt et al., 1955), making echo-sounding, coring, seismic-refraction and towed-magnetometer studies. Seismic-refraction profiles were laid out parallel to the axis of the trench (Fig. 5.6.3-04). The interpretation of the data (Fig. 5.6.3-05) indicated that the basement beneath the sediments had P-wave velocities of 5.2 km/s throughout the area. The sediments thickness was ~2 km in the Tofua trough to the west, thinner than 200 m on average in the inner gorge of the trench, and ~400 m thick on the east flank, 50 km away from the trench axis. Beneath the 5.2 km/s basement, the crustal velocity was 7 km/s beneath the Tofua Trench and 6.5 km/s at the trench axis and on the east flank. The Moho, characterized by a velocity of 8.1 km/s, was estimated to be at 20 km depth below sea level beneath the trench axis, and 12 km below the eastern flank. Under the Tofua Trench, the highest observed velocity was 7.6 km/s, estimated to occur at 12 km depth below sea level.

At the same time, another series of experiments, lasting from 1950 to 1952, explored various sites in the Pacific Ocean (Gaskell and Swallow, 1952). Sixteen stations were occupied during this time and explored special types of geological structures at very different geographic locations. Four stations explored deep ocean basins: one was placed in the center of the northern Pacific at ~35°N, 143°W, one was located to the northwest of Hawaii, a third one was near the Kuril arch half-way between Kamchatka and Japan, and the fourth one was to the north of New Guinea. All stations showed a thin sedimentary layer overlying material with high P-wave velocity near 6.15 km/s. Other positions of seismic-refraction recordings were in deep ocean waters near islands. One station near Hawaii and another one near Funafuti showed thin layers of sediments followed by 2–3-km-thick material with P-velocities of 3.3–4.5 km/s, overlying a thin high-velocity layer (6.6–6.9 km/s apparent velocities, not reversed). In the western Pacific, two stations were placed in deep water southwest of Formosa (Taiwan). They showed a smaller velocity for the main layer under 600 m of sediments. Continental-type observations finally were made at stations off the North American coast (Southern California), but there were no reversing shots. Nevertheless, two distinct layers were indicated at each station, and the velocity in the deeper layer was substantially lower than at the deep-ocean stations. Some stations were placed near deep

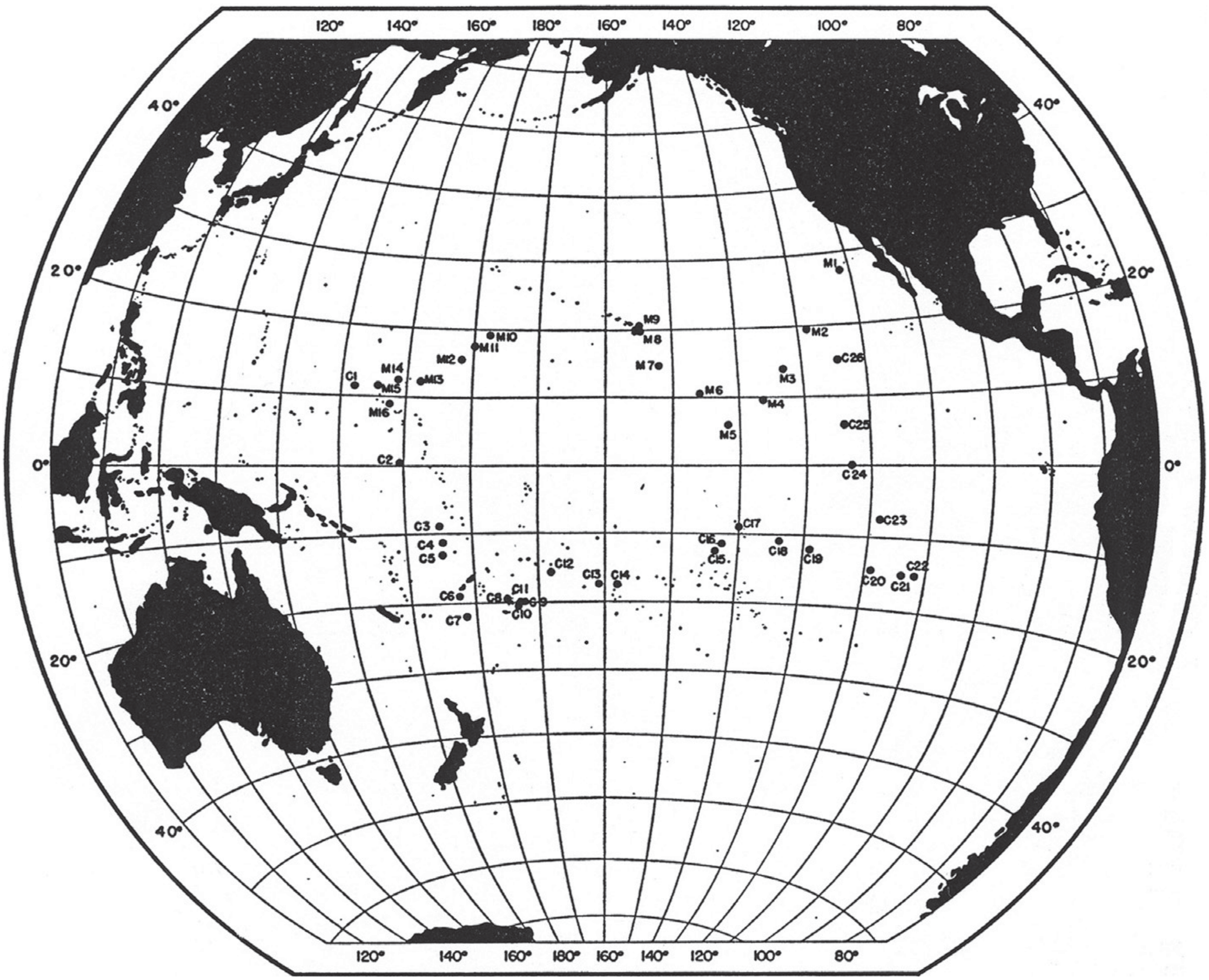


Figure 5.6.3-01. Seismic refraction stations of the Mid-Pacific and Capricorn expeditions (from Raitt, 1956, fig. 1). [Bulletin of the Geological Society of America, v. 67, p. 1623–1640. Reproduced by permission of the Geological Society of America.]

trenches, two, for example, southwest of Guam, but the results were not reliable.

Gaskell and Swallow (1953) made a third voyage into the Indian Ocean, passing through the Red Sea (reported upon by Drake and Girdler, 1964) and the Mediterranean Sea where they also did some measurements. On the basis of their worldwide study, they defined different structural types. In the ocean basins, the basement velocity was found to be higher than 6 km/s, sometimes as high as 7 km/s, and they assumed the presence of basic rocks of at least 2.4 km thickness. At two stations they could determine the peridotitic mantle layer at 8–9 km depth below sea level. Near island arcs layers with velocities of 3.0–4.5 km/s were found above the crystalline basement which Gaskell and Swallow interpreted as volcanics or limestones. Near the continents, they determined sediments with velocities of 2.5–3.3 km/s to be consolidated, and they concluded that the basement velocities between 5 and 6 km/s infer that the basement is of continental granitic character.

In 1957, the IGY (International Geophysical Year) made it possible to send two ships on the Expedition Downwind into the large unexplored region east of the Capricorn Expedition (Raitt, 1964). Exploration of two major features was planned: the broad East Pacific Rise and the great Peru-Chile Trench forming the western boundary of most of South America. On the travel time plots from 90-km-long seismic profiles across the Peru-Chile Trench, three significant velocities could be identified: 4.5 km/s, 7.0 km/s, and 8.4 km/s. Two cross sections were constructed by Raitt (1964), one across the continental margin of Chile and one of Peru. On the Chile section, the results of four profiles, at 260 km, 170 km, 80 km (above the trench axis), and 20 km west

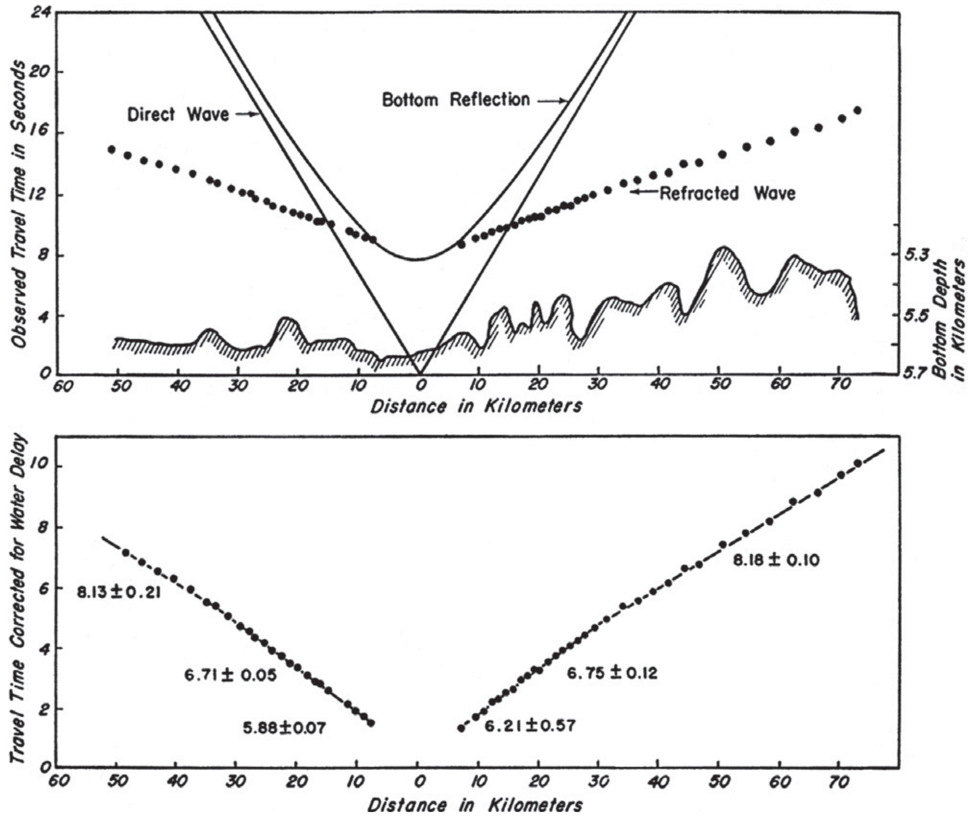


Figure 5.6.3-02. Example of travel-time plots, here for station M7 of the Mid-Pacific expedition (from Raitt, 1956, fig. 7). [Bulletin of the Geological Society of America, v. 67, p. 1623–1640. Reproduced by permission of the Geological Society of America.]

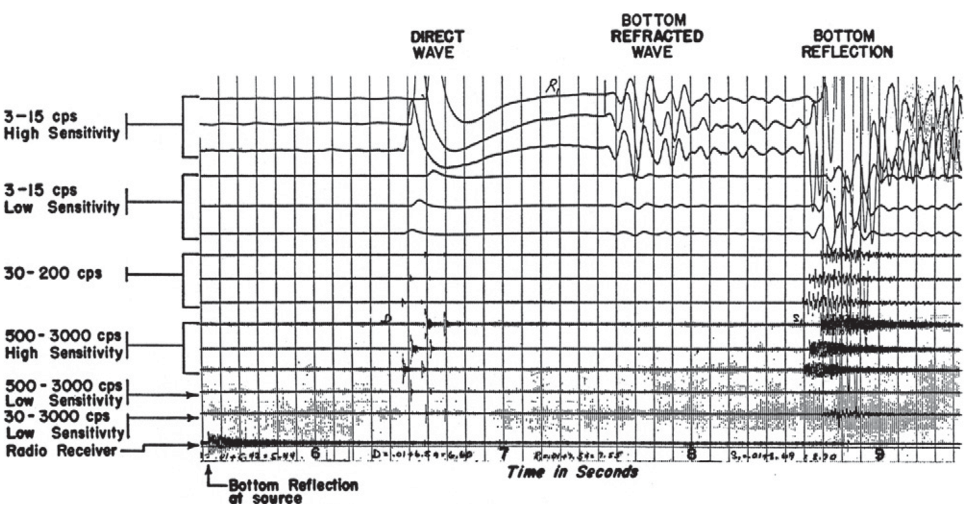


Figure 5.6.3-03. Sample of a complete oscillogram of seismic waves recorded at short range on the Mid-Pacific expedition in 1950 (from Raitt, 1956, fig. 4). [Bulletin of the Geological Society of America, v. 67, p. 1623–1640. Reproduced by permission of the Geological Society of America.]

of the coast, were compiled, showing a downwarp of Moho from 13 km depth at the 260-km station to ~23 km at the 20-km station. Both the 6.6–7.0 km/s refractor and the Moho were about parallel to the sea floor from the trench axis seawards. At the Peru continental margin, only two of the four measured stations gave a result for Moho: at 310 km off the coast, Moho depth was near 11 km, and under the trench axis, Moho depth was near 17 km. The 6.8–6.9 km/s refraction could be followed to 70 km off the coast, showing a generally increasing depth toward the coast. In summary, the crust west of the Chile Trench was 2 km thicker than west of the Peru Trench. Under both trench axes, the same crustal thickness of 10.5 km was found, though the sea floor under the Chile Trench is 2 km deeper than under the Peru Trench (Raitt, 1964).

A summarizing cross section across the East Pacific Rise by Raitt (1964) showed a slight decrease of the total crustal thickness under the axis of the Rise and a decrease of mantle velocities to 7.5 km/s. However, the mantle velocity decrease appeared to be concentrated on a narrow section across the axis of the Rise only, associated with high heat flow values. Outside

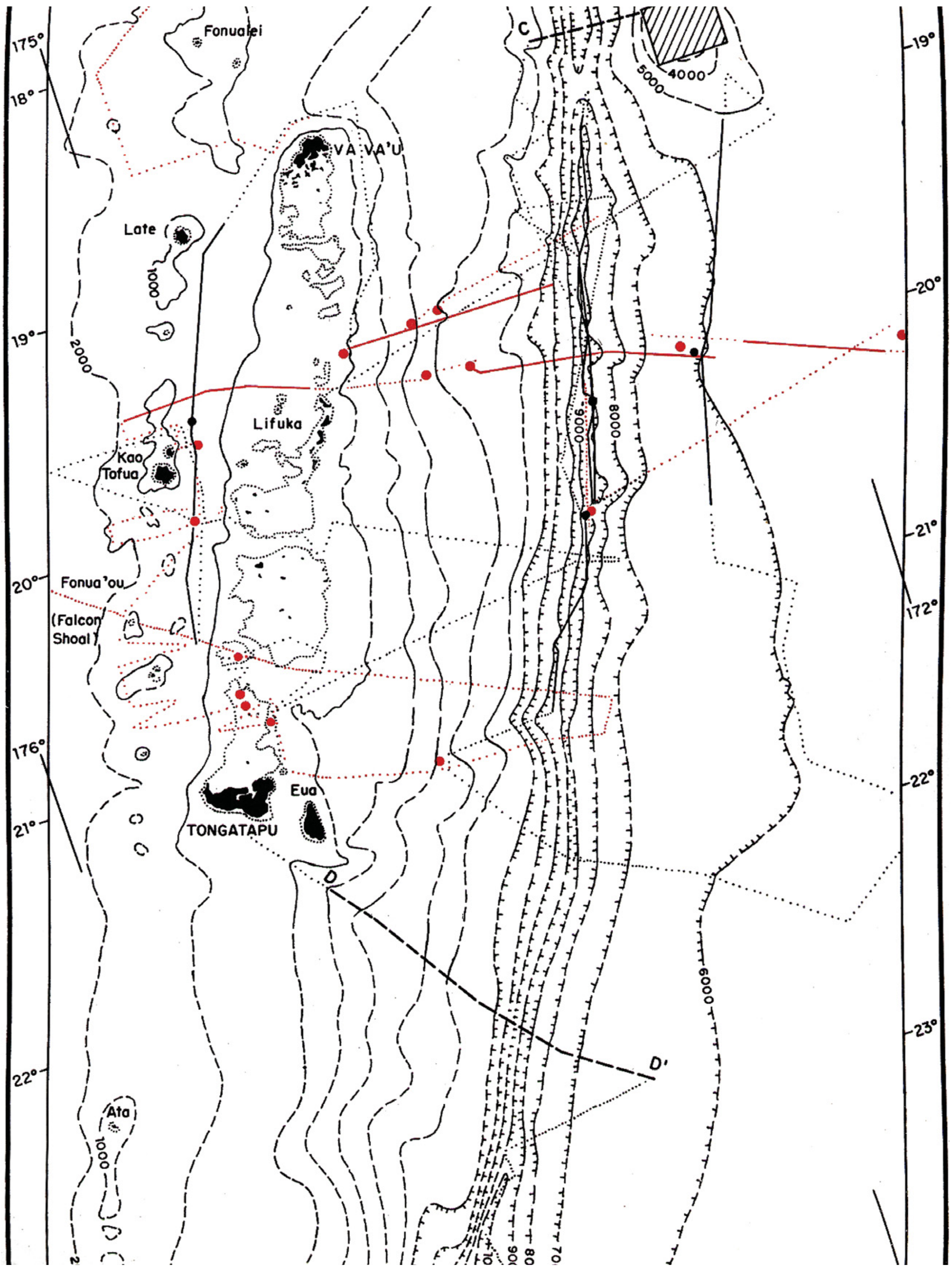


Figure 5.6.3-04. Location of seismic-refraction profiles in the Tonga area (N-S running straight lines) (from Raitt et al., 1955, pl. 1). [Geological Society of America Special Paper 62, p. 237–254. Reproduced by permission of the Geological Society of America.]

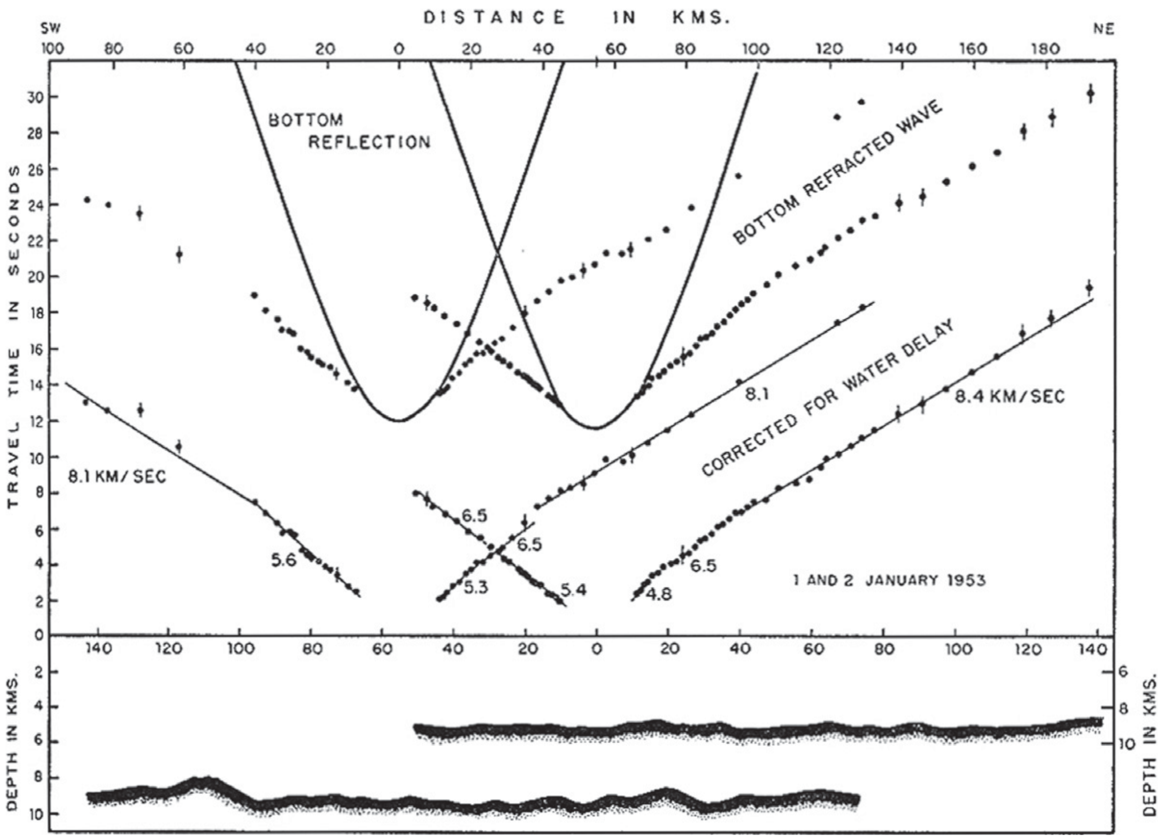


Figure 5.6.3-05. Travel-time plots the of seismic refraction profiles along the axis and the eastern flank of the Tonga trough (from Raitt et al., 1955, fig. 2). Axial profile to the left. Note that the depth axis of the flank profile is shifted upwards to avoid overlap. [Geological Society of America Special Paper 62, p. 237–254. Reproduced by permission of the Geological Society of America.]

the narrow central region, the mantle velocities were normal: equal or above 8 km/s.

Other information on the general crustal thickness in the Pacific area came from the study of surface waves (Officer, 1955b), in particular on the southwestern part including New Zealand and the Antarctica.

5.6.4. Summary of Seismic Observations in the Oceans in the 1950s

An overview of seismic measurements in the oceans was compiled from the database assembled by one of the authors (W.D.M.) and his seismic-refraction group at the U.S. Geological Survey in Menlo Park, California (Fig. 5.6.4-01). It shows all surveys that were published until 1959 in generally accessible journals and books. The map shows the numerous observations in the northwestern Atlantic Ocean, mainly organized through the initiative of William Maurice Ewing and his colleagues and performed by the Woods Hole Oceanographic Institution in Massachusetts and the Lamont Geological Observatory of the Columbia University at New York. The Mid-Atlantic Ridge and the northeastern Atlantic Ocean saw a considerable number of marine expeditions. Also in the Pacific Ocean, a substantial number of seismic measurements were already achieved in the 1950s. Few measurements were reported from the Indian Ocean. Ewing and Ewing (1970) have compiled a map showing all seismic profiler traverses accomplished by Lamont ships around the world until 1967.

The many seismic-refraction profiles which had been shot in the 1950s, confirmed that the crustal structure of ocean basins is quite different from that of continents. But, as Raitt (1963) had concluded, the results revealed that the seismic structure was remarkably similar from ocean to ocean. The approximately linear segments on the travel time plots (see, e.g., Fig. 5.6.3-02) showed that the oceanic basement is covered by a sedimentary blanket over much of the sea floor and that the basement is composed of two main units: layer 2 with P-velocities of 4–6 km/s and layer 3 with P-velocities of ~6.7 km/s (Jones, 1999). The mantle structure under the oceans as derived from the marine seismic profiles could only be revealed for the uppermost kilometers, because the length of the profiles did not exceed 200 km, but had generally less than 100 km length, therefore the velocity measurements could not be extended into any appreciable depth (Ewing, 1963b). Ewing (1963b) plotted ~180 recorded seismic P-velocities greater than 7 km/s from profiles in the Atlantic, Pacific, and Indian Oceans, the Caribbean Sea, and the Gulf of Mexico. He selected only those profiles where an overlying layer

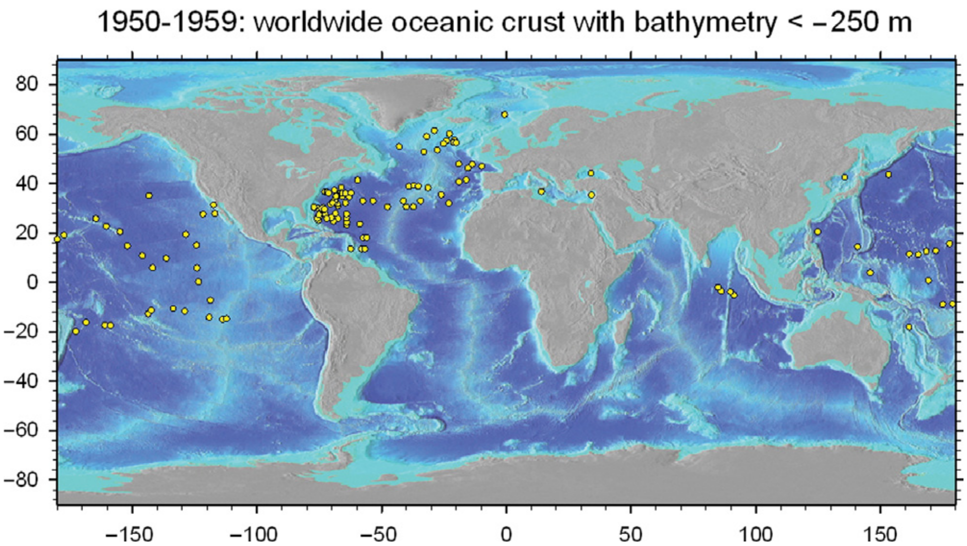


Figure 5.6.4-01. Seismic refraction measurements in the Atlantic, Pacific and Indian Oceans performed between 1950 and 1959 (data points from papers published until 1959 in easily accessible journals and books).

of high-velocity crustal rock (6–7 km/s) was present. This eliminated profiles from the Mid-Atlantic Ridge and some others on continental margins. The peak in the frequency was undoubtedly at 8.1–8.2 km/s, indicating that this is the average value of compressional waves in the upper mantle in deep-ocean areas.

5.7. RESULTS AND ACHIEVEMENTS

The progress in research was due to the efforts of many individual scientists who tried to organize sophisticated fieldwork under usually commonly difficult conditions. Young scientists, who later would become heads of university departments or research centers and whose names are authors in the reference list, started their career in the late 1950s and pushed their experiments with much vigor, thus setting the starting point for centers of excellence, which in the coming decades would achieve worldwide fame. In North America, the early crustal studies were pushed forward mainly by scientists of the Seismological Laboratory of the California Institute of Technology at Pasadena, California, the Dominium Observatory in Ottawa, Canada, the Department of Terrestrial Magnetism of the Carnegie Institution at Washington, D.C., the Lamont Geological Observatory at Columbia University, New York, the U.S. Geological Survey at Denver, Colorado, the University of Wisconsin at Madison, and the University of Utah at Salt Lake City. In Germany in particular, the State Geological Survey of Lower Saxony and the geophysical departments of the Universities of Clausthal, Freiberg, Goettingen, Kiel, Munich, and Stuttgart were combining their efforts on crustal research, supported also by German exploration companies, and partly in cooperation with the French Universities of Paris and Strasbourg. In Russia, crustal research was mainly concentrated in Moscow, Kiev, and Novosibirsk. Crustal research was also started in Sweden by the Seismological Laboratory of Uppsala and in Japan by the Earthquake Research Institute at Tokyo, and British scientists supported ongoing research in Australia and South Africa.

By the end of the 1950s, a basic knowledge on crustal structure around the Earth had been established, based on a considerable number of seismic-refraction investigations recording man-made explosions out to several hundred kilometers on land and recording a vast amount of marine seismic profiles. A comprehensive state of the art summary was compiled by Steinhart and Meyer (1961; Appendix A5-1). In Chapter 2, J.S. Steinhart presents a very detailed review on the development of crustal structure research of the continents, with particular emphasis of the achievements in the United States. In Chapter 10, J.S. Steinhart and G.P. Woollard show detailed location maps of seismic-refraction profiles around the world together with the corresponding references and a table (Table 5.7-01) summarizing the main results of the individual investigations (see also Appendix A5-1-10). An enormous effort had in particular been undertaken to explore the deep ocean basins. Detailed descriptions of the numerous expeditions in various parts of the oceans around the world in the 1950s were published in a comprehensive volume, *The Sea*, volume 3, edited by Hill (1963a). By the end of the 1950s, the basic knowledge of the structure of the oceanic crust was obtained, which would be modified only a little in the following decades (Minshull, 2002).

Not only scientific results, but also instrumentation, management of fieldwork, data presentation, and interpretation of seismic-refraction work on land in the1950s was presented in detail by Steinhart and Meyer (1961). The individual chapters include discussion of the methods in the 1950s for interpreting seismic-refraction measurements and their uncertainties as well as the discussion of instrumentation and field problems. The formulas for a multi-layered medium assuming horizontal flat layering and also inclined layers published by Steinhart et al. (1961c; see Appendix A5-1) became the basis for refined computer programs when appropriate computers became available in the 1960s. For the seismic-refraction work at sea, Ewing (1963a) published a similar detailed overview on the elementary theory

TABLE 5.7-01. CONTINENTAL CRUSTAL STRUCTURE, A REVIEW OF THE STATE OF THE ART AT THE END OF THE 1950s (FROM STEINHART AND MEYER, 1961, TABLE 10.1, SEE ALSO APPENDIX A5-1-10).

	P_1		P_2		P_3		P_4		V_n	D	$\overline{V}$	B	I_1	I_2	R	E	
Profile no. & note	V_1, Km/Sec	h_1, Thickness, Km	V_2, Km/Sec	h_2, Thickness, Km	V_3, Km/Sec	h_3, Thickness, Km	V_4, Km/Sec	h_4, Thickness, Km	Velocity below M	Depth to M	Mean Velocity above M, Km/Sec	Bouguer Anomalies, mgals	Isostatic Anomalies, local mgals	Isostatic Anomalies, 1° squares (Tanni) mgals	Isostatic Anomalies, 5° squares	Elevation, meters	Location
										Europe							
1	3.6	6.0	5.4	7.5	6.5	13.9			8.2	27.4	5.564	–5	+5	+6	+10	50	Northern Germany
2	5.63	2.4	5.97	17.7	6.54	10.1			8.15	30.2	6.134	–25	+25	+20	+10	610	Haslach, Germany
3	5.62	10.9	6.35	19.9					8.15	30.8	6.092	–20	+10	+11	+13	305	Prague, Czechoslovakia
4	2.4	2.0	5.90	17.1	6.65	4.6			8.10	23.7	5.750	+10	+20	+27	+15	90	Hungary
										Central Asia							
5	4.5	10.0	[6.0]	13.0					[8.1]	23.0	5.348	+50?	0?	0	+15	–500?	Black Sea
6	4.3	7.0	5.6	17.0	6.5	[24.0]			[8.0]	[48.0]	5.860	–30?		+40	+25	400	Caucasus
7	4.8	15.0	6.0	10.0	6.6	15.0			8.0	40.0	5.775	–50?		–5	–20	–150	Caspian Sea
8	3.5	8.3	5.0	10.0	5.5	9.0	6.3	19.0	8.0	46.3	5.362	–75		–30	–20	0	Transcaspian Depression
9	3.5	3.9	5.5	9.9	6.3	16.0			8.0	28.8	5.865	–10		–30	–20	300	Great Balkhan Range
10	4.0	6.4	5.5	13.6	6.3	19.0			8.0	39.0	5.645	–30		–15	–20	450	Margin of Kopet Dag Mts.
11	5.8	30.0	6.4	20.0					7.9	50.0	6.040	–110		–35	–25	1400	Stalinabad
12	5.5	33.0	6.4	18.0					8.1	51.0	5.818	–300		–80	–50	3000	38°30′–70°20′
13	5.5	43.0	6.4	29.0					8.1	72.0	5.863	–410		–90	–50	5000	Academy Sci. Mts.
14	5.5	32.0	6.4	25.0					8.1	57.0	5.895	–390		–65	–50	4000	Zaalaisk Mts.
15	5.5	20.0	6.4	24.0					8.1	44.0	5.991	–190		–35	–20	1150	Osh
16	5.5	10.5	6.4	41.0					8.1	51.5	6.217	–210		–40	–35	1830	Lake Issk Kul, Tien Shan
17	5.5	15.0	6.4	23.0					8.1	38.0	6.045	–160		–50	–35	3300	Zailisk Mts.
18	5.5	11.0	6.4	36.0					8.1	47.0	6.189	–80		–30	–15	600	East Lake Balkhash
19	5.5	16.0	6.4	27.0					8.1	43.0	6.065	–30		–50	–15	400	West Lake Balkhash
20	(9 layers, increasing from 2.1–7.6 km/sec)								8.0	41.2	6.411	–75?		+20	0	500	Ural Mts.
										Southern Africa							
21	5.65	4.5	6.09	22.7	6.83	15.5			8.27	38.2	6.312	–130	–10		+20	1600	South Africa
22	5.4	1.3	6.20	36.6					8.21	37.9	6.173	–70	0		+20	1400	South Africa
23	6.03	28.2	7.19	8.4					7.96	36.6	6.296	–130	+20		+20	1500	South Africa
24	5.4	1.3	6.09	34.7					8.42	36.0	6.065	–140	+10		+20	1300	South Africa
										Australia–New Zealand							
25	6.03	37.4							8.32	37.4	6.03	–45	–30		–35	300	Western Australia
26	[4.5]	1.5	6.04	34.8					8.03	37.0	5.863	–40	–15		–20	625	Eastern Australia
27	3.5	0.6	5.5	3.1	6.07	3.3	6.22	10.6	8.02	18.1	5.807					100	Wellington NE, New Zealand
										South America							
28	5.3	4.1	6.2	21.2	6.7	39.6			8.0	64.9	6.448				+10	4300	Peru, Altiplano
29	5.3	4.1	6.2	21.2	6.7	26.4			8.0	51.7	6.384				+30	1520	Peru, (flank of plateau)
30	5.5	6.0	6.35	28.4	7.0	35.9			8.0	70.3	6.609					4500	Altiplano, Chile
31	5.5	6.0	6.35	28.4	7.0	22.2			8.0	56.5	6.526					1520	Chile (flank of plateau)
										Japan							
32	1.74	1.0	5.5	4.3	6.2	21.7			7.7	27	5.923	+125		+15	+35	25	
33	2.51	0.53	5.8	9.0	6.15	15.5			7.75	25	5.951	+50		+50	0	200	
34	5.8	1.5	6.15	22.5					7. 75	24	6.128	+125		–47	0	360	
35	5.5	12.0	6.3–6.7	23.0					8.0	35.0	6.157	+140	0		–20	0	Sakhalin Is.
36	1.5	3.0	6.3–6.7	17.0					8.0	20.0	5.750	+170	+30		0	–3000	Sea of Okhotsk
37	1.5	5.0	6.3–6.7	9.5					8.0	14.5	4.776	+410	+50			–5000	Pacific Margin

(continued)

TABLE 5.7-01. CONTINENTAL CRUSTAL STRUCTURE, A REVIEW OF THE STATE OF THE ART AT THE END OF THE 1950s (FROM STEINHART AND MEYER, 1961, TABLE 10.1, SEE ALSO APPENDIX A5-1-10). (*continued*)

	P_1		P_2		P_3		P_4		V_n	D	$\overline{V}$	B	I_1	I_2	R	E	
Profile no. & note	V_1, Km/Sec	h_1, Thickness, Km	V_2, Km/Sec	h_2, Thickness, Km	V_3, Km/Sec	h_3, Thickness, Km	V_4, Km/Sec	h_4, Thickness, Km	Velocity below M	Depth to M	Mean Velocity above M, Km/Sec	Bouguer Anomalies, mgals	Isostatic Anomalies, local mgals	Isostatic Anomalies, 1° squares (Tanni) mgals	Isostatic Anomalies, 5° squares	Elevation, meters	Location
									Iceland								
38	3.69	2.12	6.71	15.70	7.38	10.02			[8.0]	27.8	6.731	–10			+15	150	Western Iceland
									North America								
39	5.7	9.1	6.6	15.0	7.3	22.2			8.3	46.1	6.756	+10	+25		+20	0	Alaska: Prince William Sd.
40	5.7	6.1	6.6	10.6	7.3	32.2			8.3	48.9	6.948	–70	0		+20	850	Southern Alaska
41	5.5	4.0	6.1	10.0	6.7	24.0			[8.1]	38.0	6.416	–75	0		0	1000	Alaska: Skagway area
42	3.6	2.0	6.1	1.0	6.2	26	7.2	14	8.2	43.0	6.402	–115	–12		0	900	Alberta, Canada
43	6.0	increasing to 7.0 km/sec at base of crust							8.0	30.0	6.5	–45	–18		–15	300	Washington: Puget Sound
44	5.0	3.72	5.95	26.16	7.44	5.48			7.94	35.36	6.081	–160	+10		+10	1500	Western Montana North
45	5.0	2.29	5.95	19.89	7.44	23.91			7.94	46.09	6.676	–200	+3		+10	1850	Western Montana South
46	[3.6]	.7	5.63	14.77	6.70	13.51	7.24	25.72	7.87	54.7	6.691	–135	+5		+10	1500	Northern Montana
47	[3.6]	2.78	6.08	12.93	6.88	24.61			8.15	40.32	6.397	–170	+15		+20	1700	Southern Montana
48	3.50	2.09	6.08	15.13	6.97	16.91	7.58	22.82	8.07	56.92	6.854	–105	+8		+20	900	East Montana North
49	3.58	3.38	6.08	19.62	6.97	16.86	7.58	10.09	8.07	49.97	6.512	–105	+20		+20	950	East Montana South
50	[3.6]	3.06	6.14	13.62	6.87	11.02	7.25	18.32	7.97	46.02	6.588	–100	+20		+25	800	Northeast WyomIng
51	5.6	1.0	6.0	increasing to 7.0 km/sec					8.1	42.5	6.5	–8	+27		+10	430	Central Minn.
52	5.40	3.67	6.11	12.17	6.51	26.50			8.03	42.34	6.300	–53	–15		+10	300	N.W. Wisconsin
53	4.75	1.85	6.44	13.84	6.67	22.57			8.15	38.27	6.492	+3	+10		0	200	Keweenaw
54	5.40	.98	6.11	11.25	6.51	25.17			8.03	37.40	6.361	–63	–23		–10	300	Central Wisconsin
55	4.16	2.7	6.03	increasing at (1+.0038Z) to			6.9	37.95	8.17	40.65	6.317	–45	–15		0	200	Central Wisconsin
56	4.58	1.44	5.74	6.05	6.22	30.0			8.17	37.5	6.078	–45	–26		0	195	Central Wisconsin
57	4.5	.63	5.94	increasing at (1+.0049Z) to			7.09	42.95	[8.17]	43.6	6.488	–37	–11		–10	170	Central Wisconsin
58	6.23	28.1	7.09	12.1					8.17	40.2	6.489	–45	–18		–20	305	Ontario
59	5.9	?	6.08	32.5					[8.17]	32.5					–5	–40	St. Lawrence River Basin
60	6.3	(increasing by 2.5 × 10^{-4} km/sec/km to 7.2 krn/sec)							8.07	36.0	6.750	–40	–10		–10	315	Adirondacks
61	5.6	1.4	6.01	31.3					8.21	32.7	5.992	–50	–20		–10	270	Pennsylvania
62	5.6	0.7	6.15	increasing to 7.0 at base of crust					8.1	32.5	6.5	–5	–10		010	30	Chesapeake Bay
63	6.01	5.3	6.33	8.42	6.73	31.58			8.06	45.3	6.572	–75	–25		–15	760	East Tennessee
64	4.64	2.03	5.18	8.19	6.64	30.98			8.16	41.2	6.251	+13	+22		0	87	Arkansas
65	4.8	4.3	6.1	21.7	7.36	22.1	7.34	8.8	8.15	48.1	6.563	–210	–20		–10	2000	Arizona–New Mexico
66	5.2	5.9	5.8	11.7	6.26	21.3	7.34	8.8	8.18	47.7	6.215	–193	0		+10	1000	Utah
67	3.8	2.5	6.15	33.5					8.15	36.0	5.987	–130	–20		–10	1830	Nevada
68	6.11	23.0	7.66	26.0					8.11	49.0	6.932	–105	–25		–20	920	Southern Calif. (inland)
69	5.68	11.3	7.18	32.6					8.10	43.9	6.794	–50	–55		–20	450	Southern Calif.
70	5.9	1.0	6.1	5.0	6.5	10.0	6.85	24.0	8.20	40.0	6.645	–50	–55		–20	200	Southern Calif.
71	5.0	0.5	5.9	5.5	6.07	20.0	7.0	6.0	8.2	32.0	6.198	–20	–60		–20	100	Southern Calif. (near coast)
72	5.5	10.0	6.5	21.5					7.98	31.5	6.183	–3	–27		–15	200	Central Calif.
73	3.0	.80	4.95	3.40	6.01	28.46	7.63	10.69	8.38	43.36	6.269	–210	–20		–20	2200	Mexico

of seismic-refraction and seismic-reflection measurements in the introductory chapter of the comprehensive volume *The Sea*, volume 3 (Hill (1963).

Closs and Behnke (1961, 1963) reviewed the publications on crustal structure available by the end of the 1950s and compiled useful summarizing crustal cross sections of results obtained until 1960, representing the state of the art at that time. The profound difference between continental and oceanic crustal structure is demonstrated in Figure 5.7-01, showing a simplified east-west profile around the northern hemisphere. Some more details are shown in Figures 5.7-02 to 5.7-04. A summarizing cross section for North America is shown in Figure 5.7-02. For the southern hemisphere, some selected crustal columns are shown for the individual continents (Fig. 5.7-03), and finally for Europe and

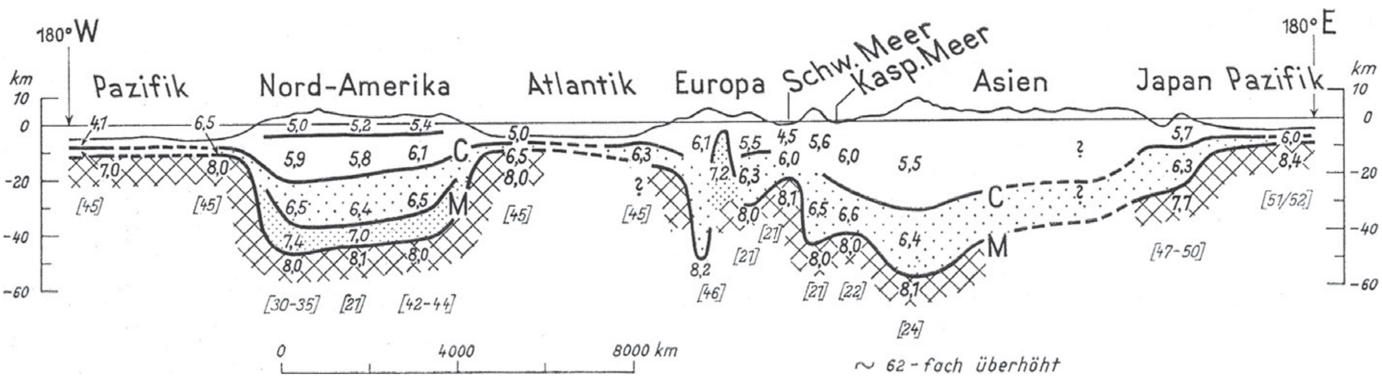


Figure 5.7-01. West-east profile across the earth at 45°N, compiled after individual results (for detailed references see Closs and Behnke, 1961). [Geologische Rundschau, v. 51, p. 315–330. Reproduced with kind permission of Springer Science+Business Media.]

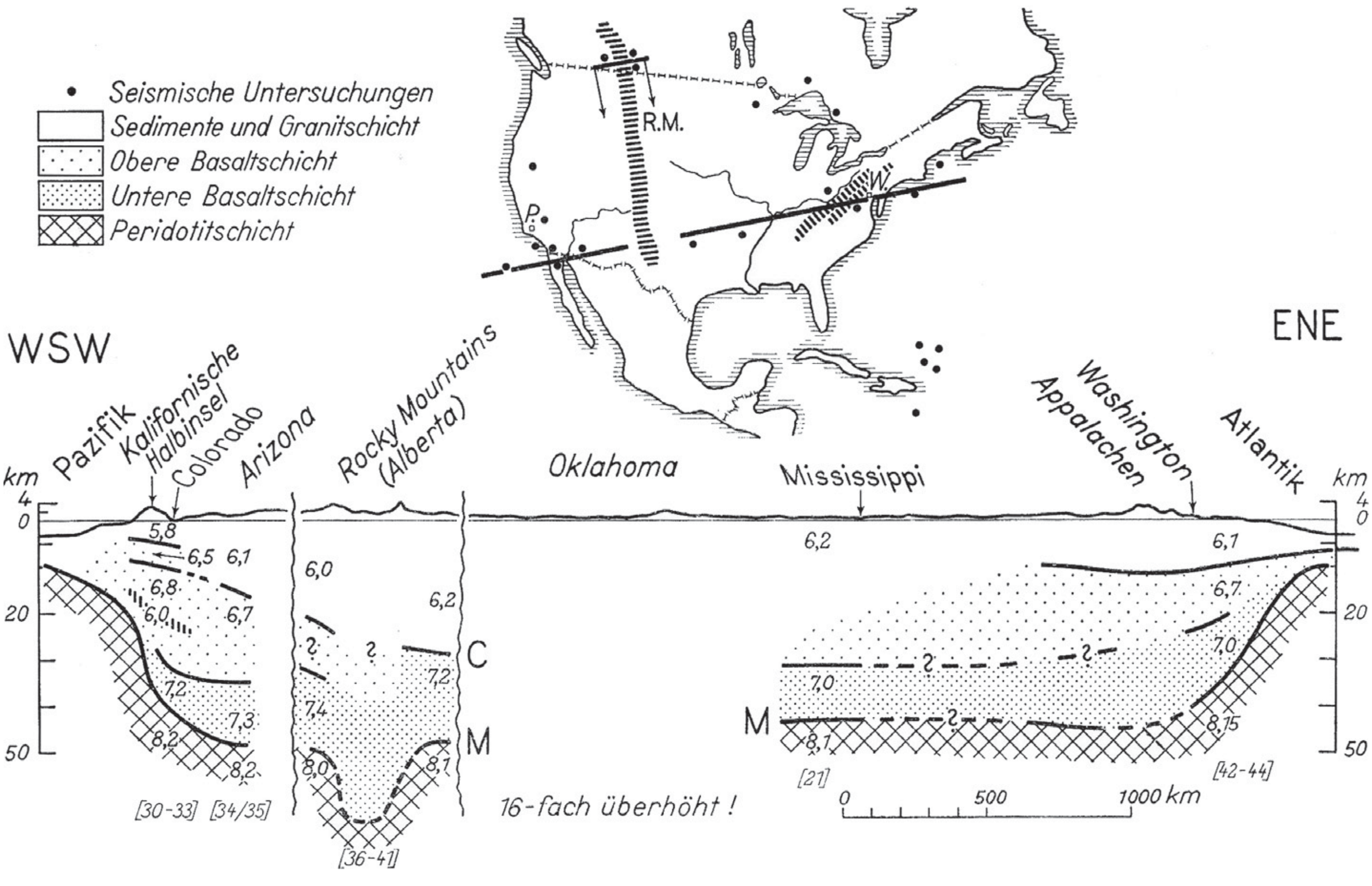


Figure 5.7-02. West-east profile through North America, compiled after individual results (for detailed references see Closs and Behnke, 1961). [Geologische Rundschau, v. 51, p. 315–330. Reproduced with kind permission of Springer Science+Business Media.]

Asia, a cross section was compiled reaching from central Europe into central Asia (Fig. 5.7-04).

The great interest in obtaining more detailed information on the Earth's crust became evident in general recommendations of the two geoscientific international unions: At its meeting in 1960 the International Union for Geodesy and Geophysics (IUGG) recommended exploring the Earth's crust more intensely rather than to bring it up-to-date. At about the same time, the International Congress for Geology established a commission which should recommend and advise major geophysical projects.

Table 5.7-01 shows the main results of continental crustal structure, as known by the end of the 1950s and summarized by Steinhart and Meyer (1961, Chapter 10), subdivided into continents. The table also includes for comparison average gravity anomalies and elevation of the areas investigated.

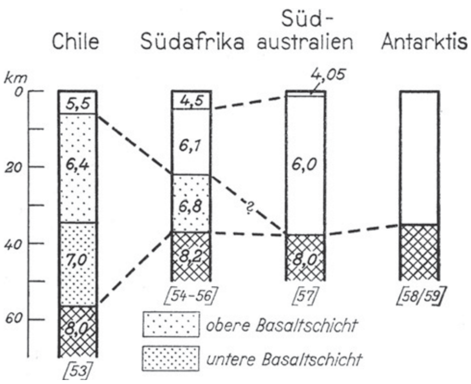


Figure 5.7-03. Velocity-depth columns for singular projects on the southern hemisphere, compiled after individual results (for detailed references see Closs and Behnke, 1961). [Geologische Rundschau, v. 51, p. 315–330. Reproduced with kind permission of Springer Science+Business Media.]

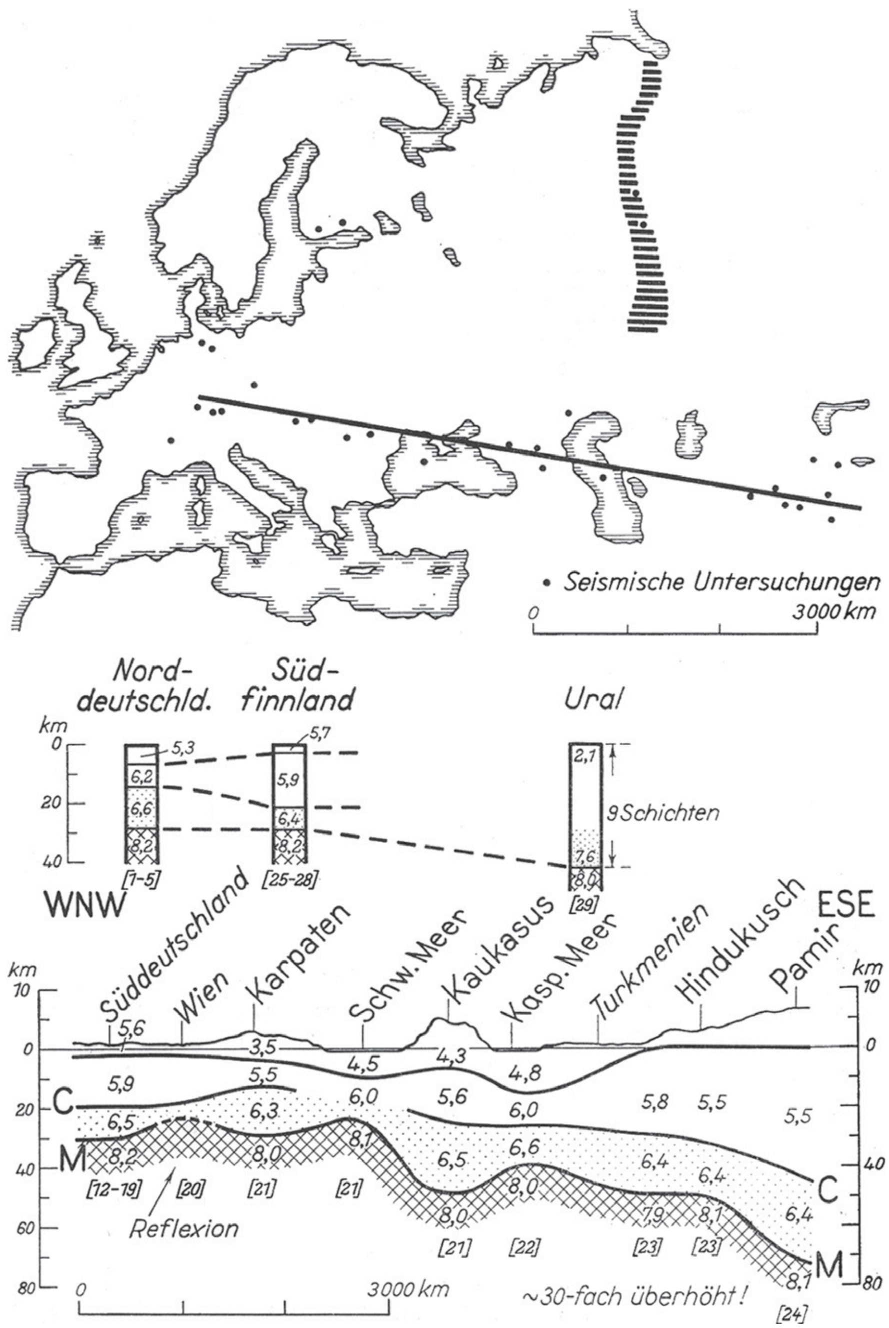


Figure 5.7-04. West-east profile from southern Germany to Pamir, compiled after individual results (for detailed references see Closs and Behnke, 1961). [Geologische Rundschau, v. 51, p. 315–330. Reproduced with kind permission of Springer Science+Business Media.]