Geological Society of America Memoirs

CHAPTER 6 The 1960s (1960–1970)

Geological Society of America Memoirs 2012;208;113-198
doi: 10.1130/2012.2208(06)

Notes

CHAPTER 6

The 1960s (1960–1970)

6.1. CRUSTAL STRUCTURE RESEARCH BY QUARRY BLAST OBSERVATIONS AND BOREHOLE SHOOTING

The 1960s saw a major breakthrough in the study of the Earth's crust. During the first experimental phase of crustal studies in the 1950s, new types of instruments had been developed and tested. This experimental stage continued at the beginning of the 1960s and finally led to the production of powerful instrumentation in major numbers, in particular in western Europe, North America, and the former USSR. In parallel to the fieldwork, the art of interpretation was pushed forward when scientists at different places in the world started to make use of rapidly developing computer technology and searched both for improvements of the known methods and at the same time developed new methods to interpret the increasing number of recently observed data.

The majority of the seismic-refraction experiments in the 1960s were carried out on a national basis, but there was also major international cooperation involved. The investigation of the Alps involved scientists from all bordering nations as well as British scientists. Under the heading of the USSR, a series of international profiles was recorded throughout southeastern Europe, and the Lake Superior experiments involved all major North American research institutions.

For the Atlantic Ocean, a detailed overall picture of the oceanic crust was established, but significant deviations from this average were found on the flanks of the Mid-Atlantic Ridge. Numerous new data were gathered in the Indian and Pacific Oceans. For the first time, the existence of seismic anisotropy of the uppermost mantle was investigated by special anisotropy experiments in the Pacific Ocean (e.g., Raitt et al., 1971).

These efforts were strongly supported by the formulation of major research programs, which were financed by national research foundations and thus allowed the systematic study of specific tectonic regions of the Earth. In Europe, the European Seismological Commission (ESC) had been founded with meetings every two years held alternately in eastern or in western Europe that served as a most effective communication center between scientists in eastern and western Europe. A special subcommission of the ESC dealt in particular with experiments and results of explosion seismology investigations. In 1964, under the umbrella of the IASPEI (International Association of Seismology and Physics of the Earth's Interior) the Upper Mantle Project was started as an international program of geophysical, geochemical, and geological studies concerning the "upper mantle and its influence on the development of the earth's crust."

In 1968, a conference of experts on explosion seismology was organized in Leningrad, where leading scientists such as A.S. Alekseyev (Novosibirsk), M.J. Berry (Ottawa), P. Giese (München), A.L. Hales (Dallas, Texas), J.H. Healy (Menlo Park, California), I.W. Litvinenko (Leningrad), I.P. Kosminskaya (Moscow), R. Meissner (Frankfurt), R.P. Meyer (Madison, Wisconsin), P.N.S. O'Brien (Sunbury-on-Thames), N.I. Pavlenkova (Kiev), N.N. Puzyrev (Novosibirsk), V.Z. Ryaboy (Kiev), V.B. Sollogub (Kiev), A. Stein (Hannover), I.S. Volvovsky (Moscow), and S.M. Zverev (Moscow) exchanged their ideas and experiences and discussed in detail the current problems of explosion seismology.

By the end of the decade, a series of publications on explosion seismology, introduced by Kosminskaya (1969), was jointly published (Hart, 1969) presenting major results obtained within the Upper Mantle Project.

6.2. SEISMIC-REFRACTION INVESTIGATIONS IN WESTERN EUROPE IN THE 1960s

6.2.1. Instrumentation

In Germany, the priority program "Geophysical Investigation of Crustal Structure in Central Europe," funded by the German Research Society, was continued until 1964 and involved university institutions, state geological surveys, and exploration companies. Following the start of the Upper Mantle Project initiated in 1960 by the International Union of Geodesy and Geophysics, a national priority program with the same name was started and funded by the German Research Society in 1965.

The development of portable seismic equipment was vigorously continued, and the number of individual instruments was gradually increased. For example, at Munich, up to 15 field stations could be mobilized by 1965 which, due to their highly sensitive galvanometers, could be easily arranged as one, three, or multi-component stations and were at the same time very easy to handle (Figs. 6.2.1-01 and 6.2.1-02). However, due to the individual development at the various research institutions, the diversity of instrumentation was large, and the demand for a unique recording system became stronger and stronger.

The first step was accomplished in 1960 when a new portable highly sensitive electromagnetic seismograph (2 Hz) constructed by H. Berckhemer at Stuttgart (Fig. 6.2.1-03) and manufactured by the company Stroppe (Schwenningen, Germany) became available. This seismograph, named FS-60, could be used either as a vertical or as a horizontal sensor.

Figure 6.2.1-01. Mobile field station of the Institute of Applied Geophysics, University of Munich (photograph by C. Prodehl). The grey box below is the central recording unit with galvanometer and a photographic film device driven by a windshield wiper motor, top left: a radio receiver for time signal reception, top right: FS-60 seismometer.

Figure 6.2.1-02. Mobile three-component field station of the Institute of Applied Geophysics, University of Munich (photograph by C. Prodehl). The larger wooden box below is the central recording unit with three galvanometers and a photographic film device driven by a windshield wiper motor. Left and right of the central unit: two time signal receivers (left, a commercial radio; right, a special army radio for short-wave reception); top left: small box for carrying an FS-60 seismometer; in foreground left and to the right: batteries.

The second step toward obtaining unique equipment was the development of a new recording device. In close cooperation with H. Berckhemer and under the supervision of E. Wielandt, the company Lennartz at Tübingen finally developed new portable magnetic-tape recording equipment, which has been described in detail by Berckhemer (1970). In this system, frequency-multiplex modulation recording techniques on 1/4″ magnetic tape were applied (Fig. 6.2.1-04).

Figure 6.2.1-03. Electromagnetic seismograph constructed by H. Berckhemer, University of Frankfurt, Germany (photograph by C. Prodehl).

Each recording unit consisted of three 2-Hz field seismometers, with the corresponding amplifiers and modulators in one unit, a tape recorder, and a time-signal receiver (Figs. 6.2.1-05 and 6.2.1-06). The signal frequency ranged between 0.3 and 100 cps. The highly dynamic range of 60 dB was obtained by a very effective flutter and wow compensation. Special playback centers were established at selected institutions. By 1966, the MARS-66 system (Magnetic Apparatus for Recording Seismic sounding, established in 1966) was ready for use.

By a special grant from the Volkswagenstiftung issued to the FKPE (Forschungs-Kollegium für die Physik des Erdkörpers = Research Consortium for the Physics of the Earth, an organization in which members were the directors of all geophysical research institutions in Western Germany), it was possible to buy 40 calibrated instruments at once which were distributed evenly amongst the state and university research institutions. By the aid of further individual grants, the German

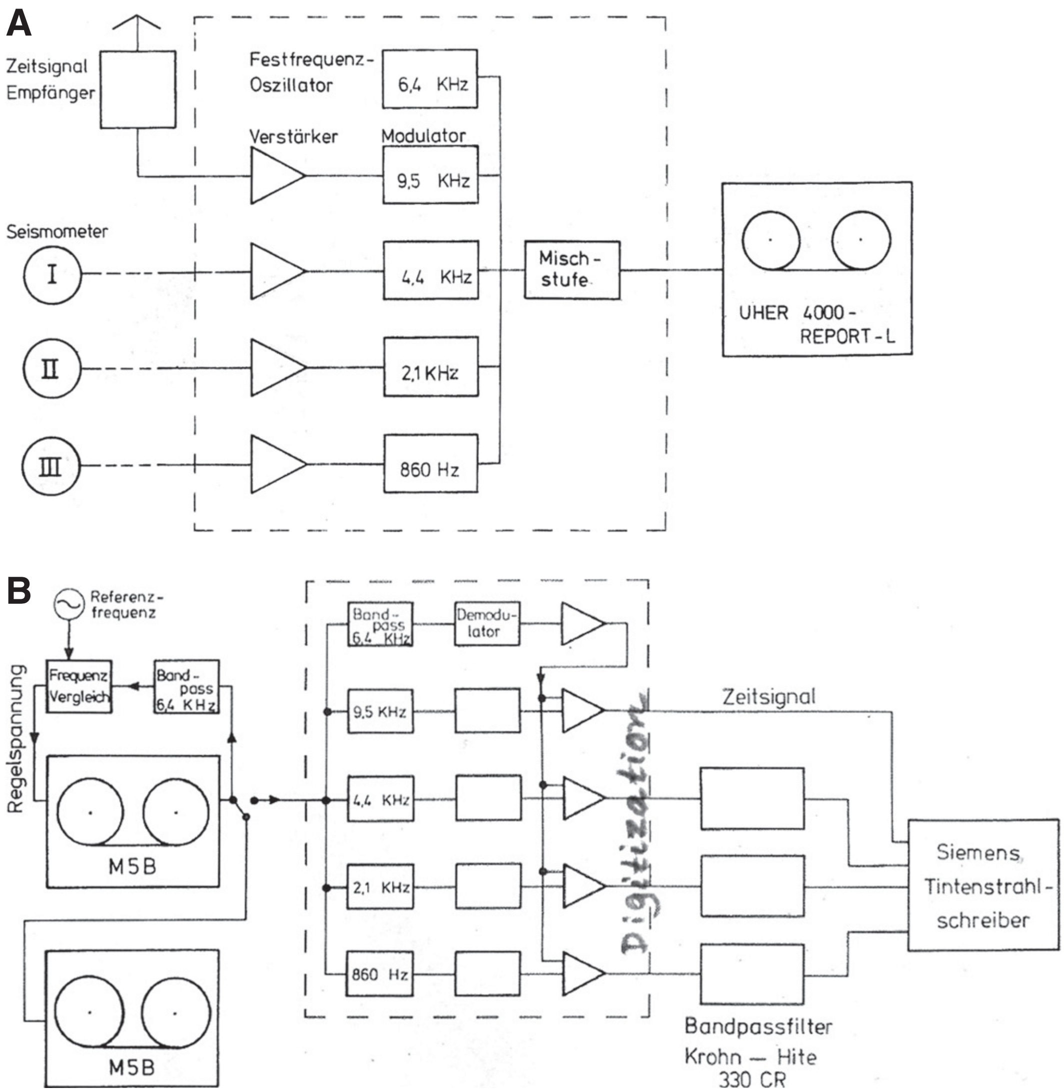


Figure 6.2.1-04. (A) block diagram of the recording unit MARS 66. (B) Block diagram of the playback unit (from Berckhemer, 1970, figs. 4 and 5). [Zeitschrift für Geophysik, v. 36, p. 501–518. Reproduced with kind permission of Springer Science+Business Media.]

scientific community could finally increase the number to over 70 instruments.

This new equipment not only enabled the recording of seismic profiles with a unique and calibrated system, but also allowed digital data preparation. The first digitizing system was developed and installed by W. Kaminski at Hamburg and later successfully reestablished at Karlsruhe. In the 1970s, MARS-66 equipment became widely distributed throughout Europe, in particular in France and Italy, so that in the last field project with this equipment in 1986, more than 200 MARS-66 units could be assembled for the seismic fieldwork along the central part of the European Geotraverse (EUGEMI Working Group, 1990).

A major problem in the early to mid-1960s, before MARS-66 became available, had been the timing of the seismic observations. Only in a few cases could commercial radio stations be convinced to broadcast a special time signal for a limited time. For example, the broadcast station Munich would, on request, interrupt its normal radio program and transmit a time signal for 2 minutes when quarry blasts at locations in Bavaria were scheduled. This middle-frequency radio station could be received by ordinary portable commercial radio receivers. Special short-wave stations like Pontoise in France had a very special schedule and required a very sophisticated planning of shooting, but through trial and error during the years, some other continuously operating short-wave transmitters were located, such as, e.g., Potsdam at 4525 kHz or Moscow at 10,000 kHz. It required, however, a special skill to receive these time signals with special army radios, which were difficult to find and expensive to buy

Figure 6.2.1-05. Recording unit MARS 66 (photograph by C. Prodehl). Left: modulator, lower right: tape recorder.

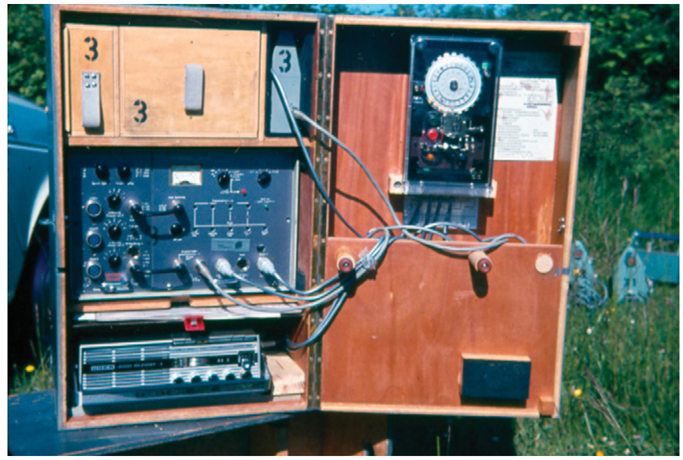


Figure 6.2.1-06. Recording unit MARS 66 equipped with a mechanical clock (upper right) (photograph by C. Prodehl). Upper left: T75A time-signal receiver; middle left: modulator; lower left: tape recorder. FS-60 seismometers in the background to the right.

in large quantities (see, e.g., Fig. 6.2.1-02). In the early 1960s, a time signal receiver was offered by Lennartz which was equipped with quartzes for special short-wave frequencies like those of Pontoise, Moscow, or Potsdam. A continuous time signal such as, e.g., WWV in North America, described in the next section, only became available in the mid-1960s when a continuous time signal broadcasted at 77.5 kHz was installed by the Federal Technical Office at Braunschweig which could be received by small specially designed receivers T75A almost everywhere in Europe (see, e.g., Fig. 6.2.1-06). This time signal receiver was incorporated into the MARS-66 recording unit. Also useful for projects in central and southern Europe since the mid-1960s were the Swiss time signal HBG, broadcasted continuously at 75 kHz, and in western and northern Europe, the British time signal MSF at 66 kHz could be well received.

6.2.2. Seismic-Refraction Experiments in Central and Western Europe

In the above-mentioned priority program "Geophysical Investigation of Crustal Structure in Central Europe," a systematic use of major quarry blasts had already been initiated in the late 1950s, offering the most economic seismic energy source. This program was continued in the 1960s, with quarry blasts serving as the major seismic energy source for crustal research work in Germany. Under the leadership of Prof. H. Closs, the state geological survey of Lower Saxony (Niedersächsisches Landesamt für Bodenforschung) overtook the task of contacting all major quarry companies in western Germany and surrounding countries and to inform all research institutions that had suitable equipment a few days before a major blast was scheduled. Thus a system of reversed and unreversed seismic-refraction lines of several hundred kilometers in length was gradually established (Fig. 6.2.2-01).

For the majority of these profiles, record sections were constructed and published on three maps in a special volume *Explosion Seismology in Central Europe* (Giese et al., 1976a). All record sections are reproduced in Appendix A6-1. Details of the shotpoints were published in Chapter 3.1 (Stein and Schröder, 1976) and details of the profiles in Chapter 3.3 (Giese et al., 1976b) of this special volume and are reprinted in Appendix A6-2.

In subsequent years, the organization of seismic fieldwork in Germany and adjacent countries was handed over to a special working group ASFA (Arbeitsgruppe für Seismische Feldmessungen und Auswertung = Working Group for Seismic Fieldwork and Interpretation) of the above mentioned FKPE, being responsible for a centralized organization of seismic fieldwork and interpretation which has remained active until 2006.

The exclusive use of quarry blasts in the 1960s had the disadvantage of a rather unequal distribution of available shotpoints inadequate to tectonic research goals. Most of the usable quarries were located in the Rhenish Massif and Hessisches Bergland, where Paleozoic rocks and Tertiary basalts are outcropping. Here, the location of several quarries allowed an overlapping arrangement of reversed profiles. Due to the uneven spacing and the large distances between such quarries, however, the term *reversed profile* had to be regarded with caution. Other shotpoints were concentrated in the Vosges, the Black Forest, and the Bohemian Massif. A very productive shotpoint with large explosions was a quarry near Eschenlohe (01 in Fig. 6.2.2-01), located near the northern margin of the Eastern Alps, which allowed successful recording up to distances of more than 200 km and also offered the chance for extending the observations into the Alpine region. Figure 6.2.2-02 shows a good example of data from a quarry blast near Hilders (Rhön), compiled into a record section.

First results of this cooperative research program were reported in 1964 (German Research Group for Explosion Seismol-

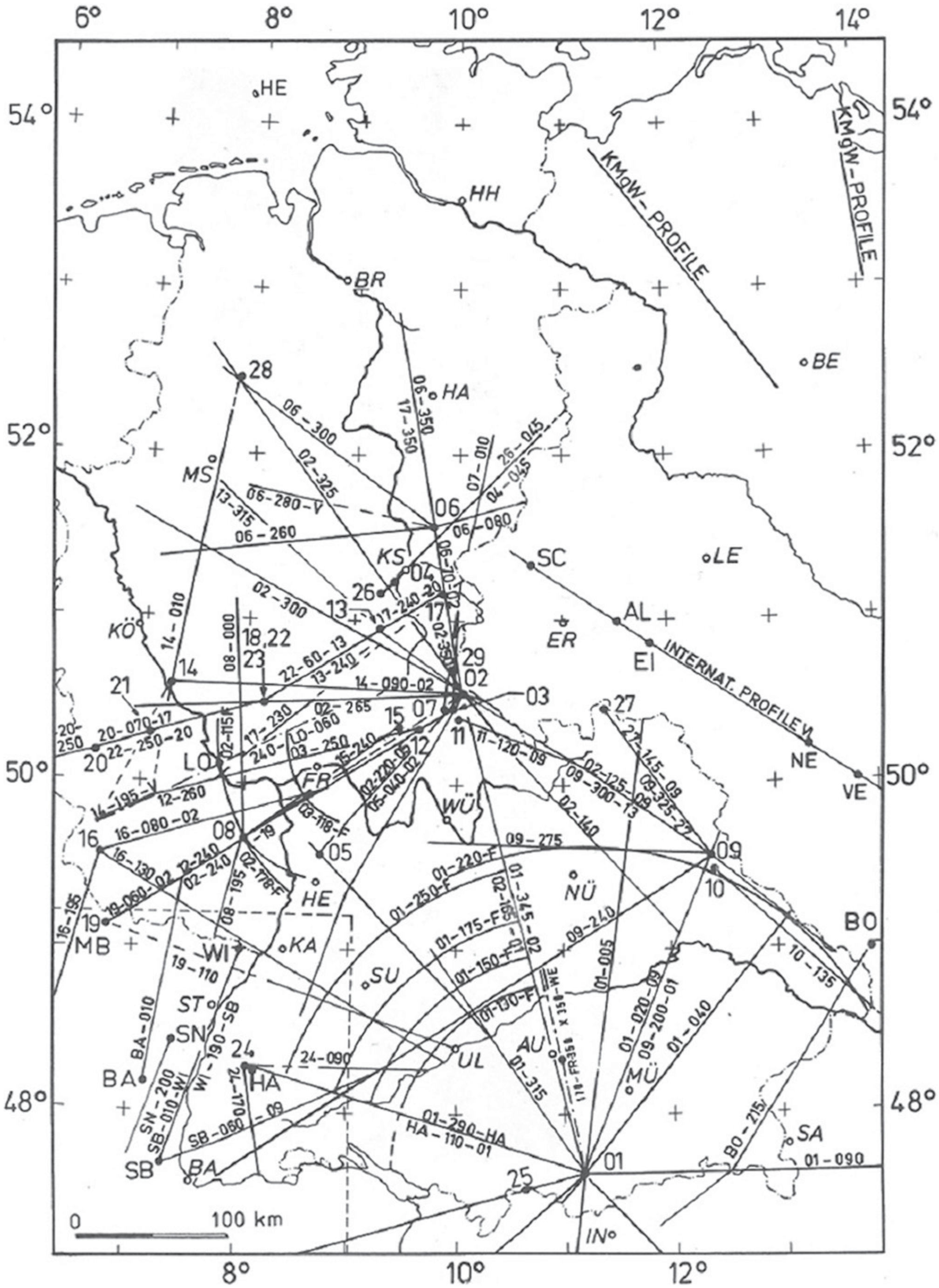


Figure 6.2.2-01. Location map of seismic-refraction observations in central Europe (from Giese et al., 1976a, fig. 1). Numbers are quarry blast locations active in the 1950s and 1960s, letters are centers of special surveys (FR, LO) and quarry and borehole locations outside Germany. Also included are the Heligoland (HE) and Haslach (HA) locations of 1947–1948. Italic letters designate major cities. Code of observations: 16-080-02—reversed profiles (shotpoint code–azimuth–reversing shotpoint code); 01-040—unreversed profiles (shotpoint code–azimuth); 14-195-V—non-completed lines (V = test); 01-150-F—fan observations (shotpoint code–average distance–fan). [*In* Giese, P., Prodehl, C., and Stein, A., eds., Explosion seismology in central Europe—data and results: Berlin-Heidelberg-New York, Springer, p. 73-112. Reproduced with kind permission of Springer Science+Business Media.]

ogy, 1964) presenting first contour maps for central Europe, such as the surface of the crystalline basement and the Conrad and the Moho discontinuities (Fig. 6.2.2-03).

These maps included an evaluation of deep reflections (Fig. 6.2.2-04) recorded during routine seismic exploration surveys of various oil companies. Already in the 1950s, Dohr (1959) had detected that commercial seismic prospecting offered the chance to record deep reflections from the Moho if the recording time was prolonged. So he had set up a research program to obtain seismograms recorded at least up to 12 seconds that could be evaluated statistically.

Liebscher (1962, 1964) applied this method to a number of seismic prospecting areas in southern Germany and was able to compile histograms, cross sections, and first contour maps for the Conrad and the Moho discontinuities in southern Germany.

Following the early successful recordings of near-vertical reflections from the Moho near Blaubeuren in 1952 (Reich, 1953),

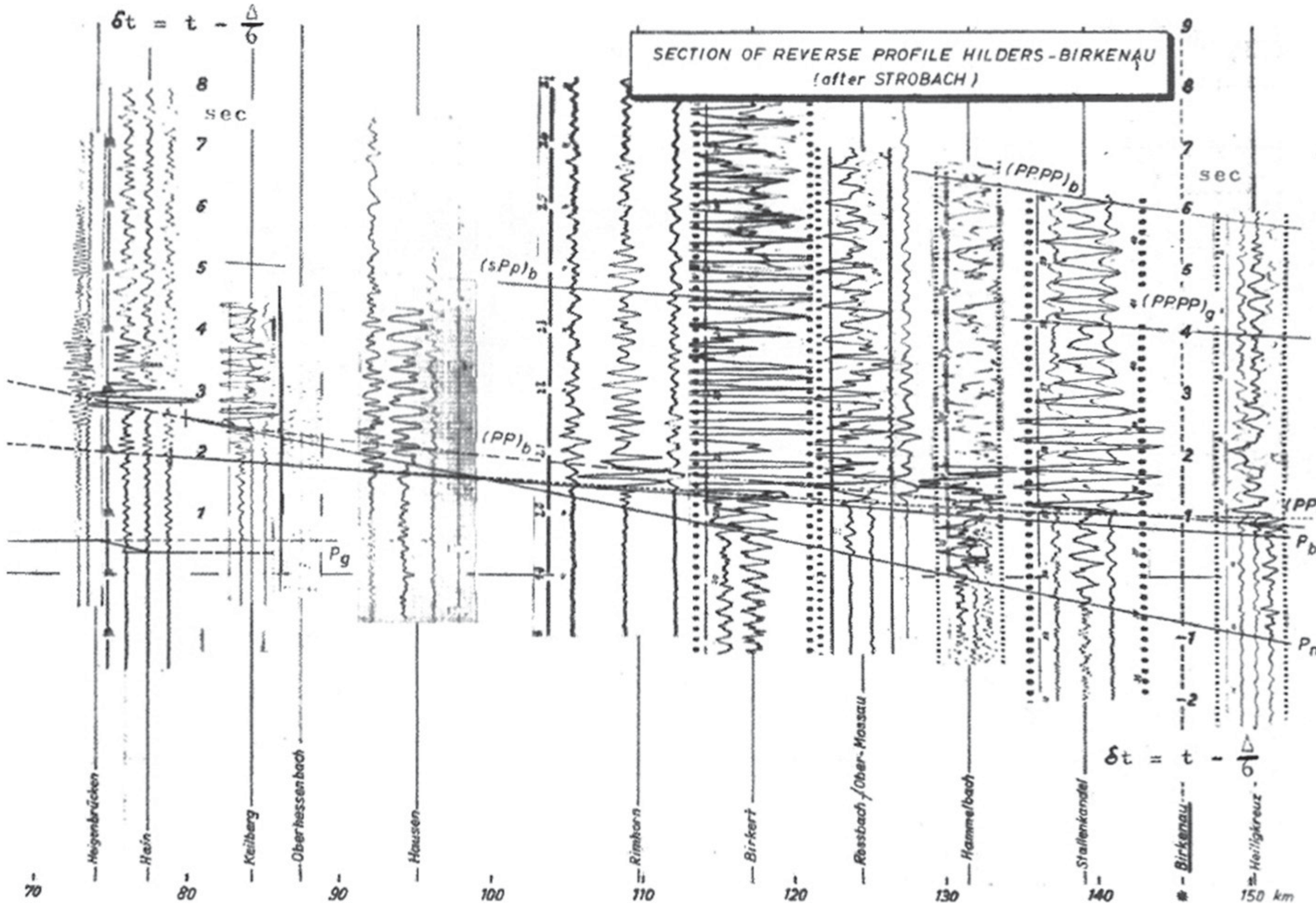


Figure 6.2.2-02. Part of a record section compiled from recordings of a blast in a basalt quarry near Hilders/Rhön (from German Research Group for Explosion Seismology, 1964, fig. 4). [Zeitschrift für Geophysik, v. 30, p. 209–234. Reproduced with kind permission of Springer Science+Business Media.]

other trials were made by university groups that had inherited older reflection instrumentation from exploration companies to record deep near-vertical reflections near multiple shots of large-scale refraction experiments, such as the Lago Lagorai experiments in 1961 and 1962. These attempts were not very successful, however.

It was R. Meissner, based in Frankfurt at that time, who first had the idea to apply the common depth point techniques of exploration seismics to the investigation of the whole crust. He designed and carried out an experiment in 1964 in the Bavarian Molasse basin (FR in Fig. 6.2.2-01) where shotpoints and a relatively small number of mobile recording units would systematically move apart from the central common depth point FR and record in the wide-angle distance range for the corresponding depth range (Meissner, 1966).

In 1968, R. Meissner arranged a second experiment of similar design in the Rhenish Massif (LO in Fig. 6.2.2-01) across its southern ranges Hunsrück and Westerwald. Both near-vertical (Glocke and Meissner, 1976) and wide-angle (Meissner et al., 1976a) reflections could successfully be recorded and evaluated (Fig. 6.2.2-05).

These special projects where wide-angle reflection profiles using noncommercial equipment were recorded using especially arranged borehole shots (Meissner, 1966; Meissner et al., 1976a) added to the success of the German crustal research program. Furthermore, in northern Germany, a commercial seismic-refraction program was made available (Dohr, 1983), and on a profile in southeast Bavaria, borehole shots (BO in Fig. 6.2.2-01) were recorded that were fired along the International Profile VII traversing Poland and the Bohemian Massif in Czechoslovakia.

Special attention was given to the Rhinegraben, where at a symposium in 1966 a special research program was initiated which started with quarry blast observations both in Germany and in France (shotpoints BA and MB in Fig. 6.2.2-01). It finally led to a special program in 1972 with borehole shots in Alsace (shotpoints SB and WI in Fig. 6.2.2-01), which were recorded on various lines through the Rhinegraben, Vosges, and Black Forest

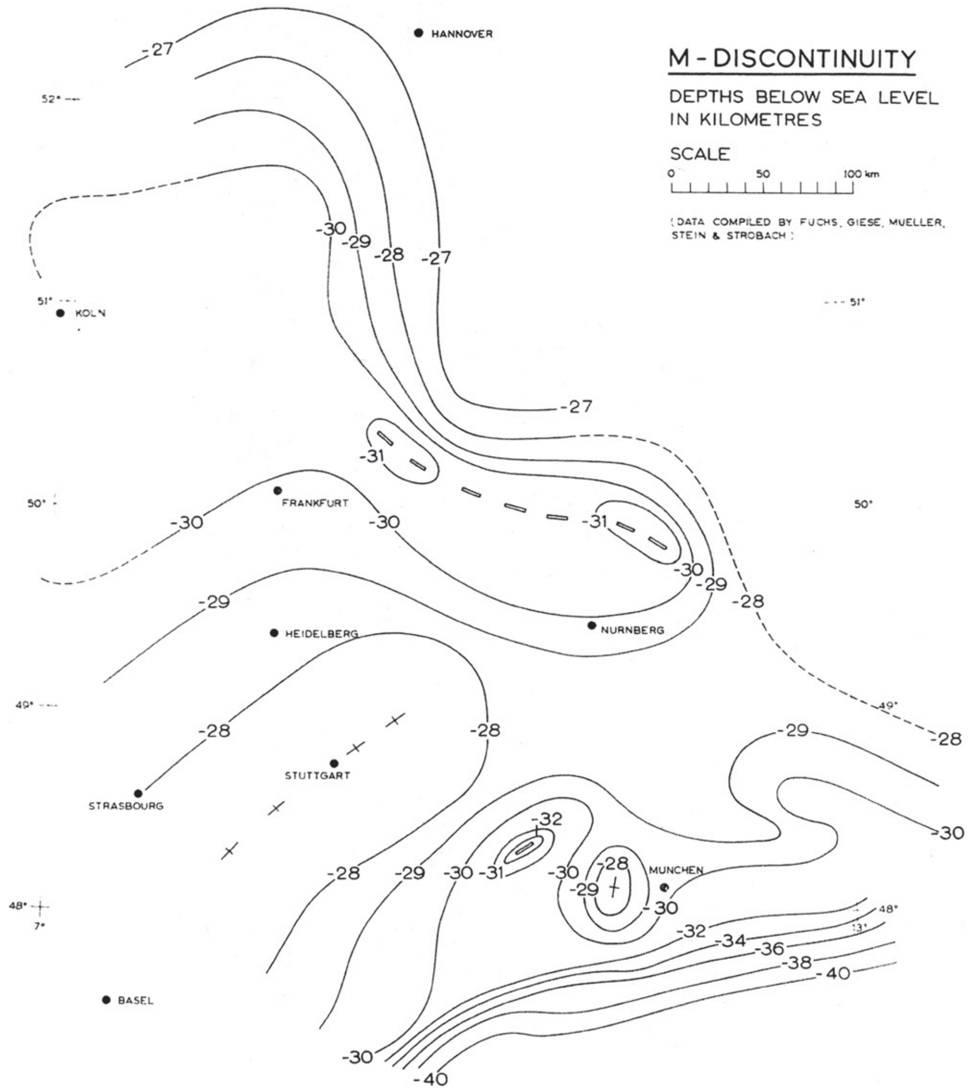


Figure 6.2.2-03. Depth contour map of the Mohorovičić discontinuity compiled until August 1963 by K. Fuchs, P. Giese, St. Mueller, A. Stein, and K. Strobach (from German Research Group for Explosion Seismology, 1964, fig. 10). [Zeitschrift für Geophysik, v. 30, p. 209–234. Reproduced with kind permission of Springer Science+Business Media.]

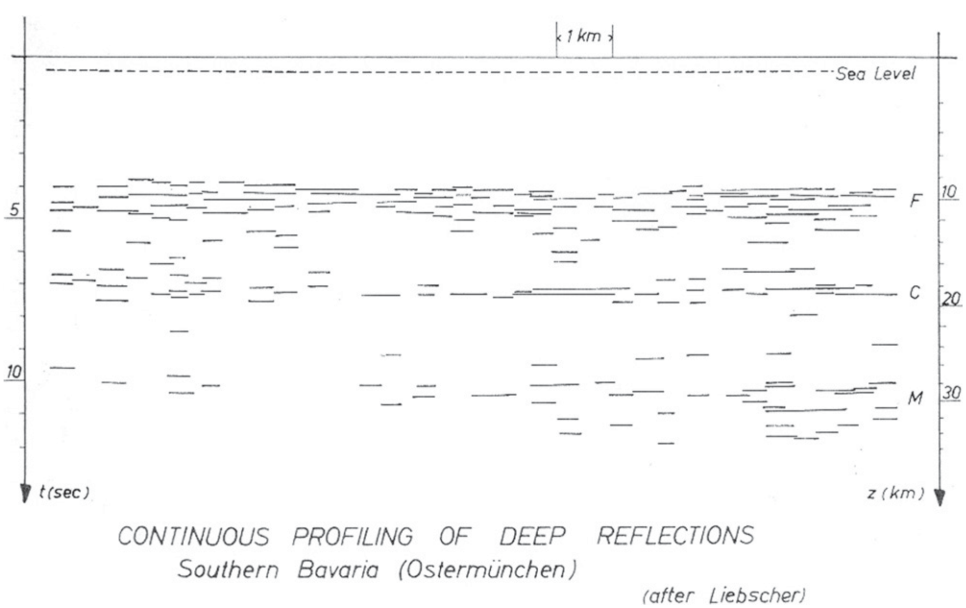


Figure 6.2.2-04. Profile of deep reflections, statistically evaluated from oil prospecting records, in the Bavarian Molasse Basin (from German Research Group for Explosion Seismology, 1964, fig. 7). [Zeitschrift für Geophysik, v. 30, p. 209–234. Reproduced with kind permission of Springer Science+Business Media.]

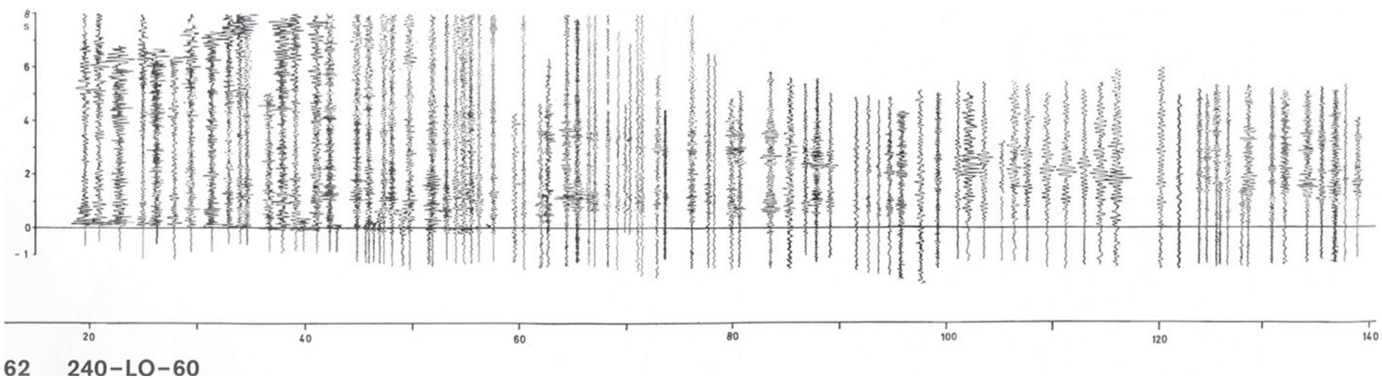


Figure 6.2.2-05. Record section of profile 240-LO-60 (from Giese et al., 1976, map 3, fig. 62, reproduced from Meissner et al., 1976) [*In* Giese, P., Prodehl, C., and Stein, A., eds., Explosion seismology in central Europe—data and results: Berlin-Heidelberg-New York, Springer, Map 3. Reproduced with kind permission of Springer Science+Business Media.]

(Prodehl et al., 1976a). This program will be discussed in more detail in Chapter 7.

For southern Germany, Prodehl (1964, 1965) compiled all available seismic-refraction data (Fig. 6.2.2-06). With increasing numbers of observations, more details could be obtained and more accurate maps could be produced (e.g., Prodehl, 1965). The large amount of data collected in central Europe by the systematic observation of numerous quarry blasts led P. Giese to a more generalized interpretation of seismic data by mapping characteristic parameters for phases which could be regionally correlated, such as, e.g., the critical distance for P_MP (crust-mantle boundary) reflections, the crossover distance of P_n and P_MP, average velocities for certain depth ranges, the P_n velocity, etc. This allowed already for the drawing of some conclusions on general features of the crustal structure before applying any depth computations (Giese and Stein, 1971; Giese, 1976a, 1976c).

6.2.3. Seismic Research in Britain and Scandinavia

The continental region of the British Isles adjoins the Northeast Atlantic margin and includes the onshore regions of the British Isles as well as the stretched continental crust beneath the seas to the south and west of Ireland and northern Norway and beneath the North Sea.

The first crustal research was undertaken on the continental shelf to the west of Britain in 1960 and reported by Bunce et al. (1964). They reported that the deepest horizon sampled may have been "an intermediate layer rather than the mantle."

Another offshore study involved a major crustal profile across the North Sea in 1967, linking Scandinavia and the British Isles (Sornes, 1968). However, Sornes (1968) deferred presenting profiles showing the crustal structure "until more data about the upper layers became available." Nevertheless it was the first such "trans–North Sea" work, which in time was followed by the major Barton and Wood (1984) work (see Section 8.3.2).

Onshore crustal research in Britain started in ca. 1962. From the very beginning, British scientists made use of the favorable position of the British Isles surrounded by water. Most activities involved marine work, with offshore underwater shots being recorded on land. For the interpretation of most surveys, the time-term method, developed by Willmore and Bancroft (1960) was applied. A comprehensive review of early crustal studies in the British Isles is given by Blundell and Parks (1969) and was republished by Willmore (1973) with additions of post-1969 work.

The first crustal investigation in the British Isles reaching the Moho was undertaken to calibrate the Eskdalemuir seismological array (Agger and Carpenter, 1965) with depth charges fired in the North Sea and the Irish Sea in July 1962, resulting in a mean Moho depth of ~25 km. However, Blundell and Parks (1969) noted that the "geological significance of the variations in crustal thickness in their model is very doubtful," considering that there was "no measure of any near surface structural complexities" available.

In 1965, 25 shots were fired again in the Irish Sea and recorded by permanent land stations in Ireland, Wales, and Scotland, including the Eskdalemuir seismological array (Blundell and Parks, 1969). This and the following experiments were especially designed to permit the time-term approach (Willmore and Bancroft, 1960; lines and models 2 and 3 in Fig. 6.2.3-01).

Another sea-to-land seismic-refraction experiment followed in 1966. Eleven seismic recording stations (including three arrays) in Cornwall, northwestern France (Brittany) near Brest, and southeast Ireland (Waterford) and on the Scilly isles recorded two lines of shots at sea, one extending from northwestern France near Brest across the westernmost tip of Cornwall to the Irish coast near Waterford, and the other line running from Cornwall (lands end) along the seaward extension of the granite batholith of southwest England (Holder and Bott, 1971; unmarked lines and model 1 in Fig. 6.2.3-01). In total, 45 × 136 kg depth charges were fired. The seismic recordings on a cross array, two linear arrays, and a long profile of stations allowed the use of phase correlation to obtain apparent velocities. A major development of the Holder and Bott (1971) survey was

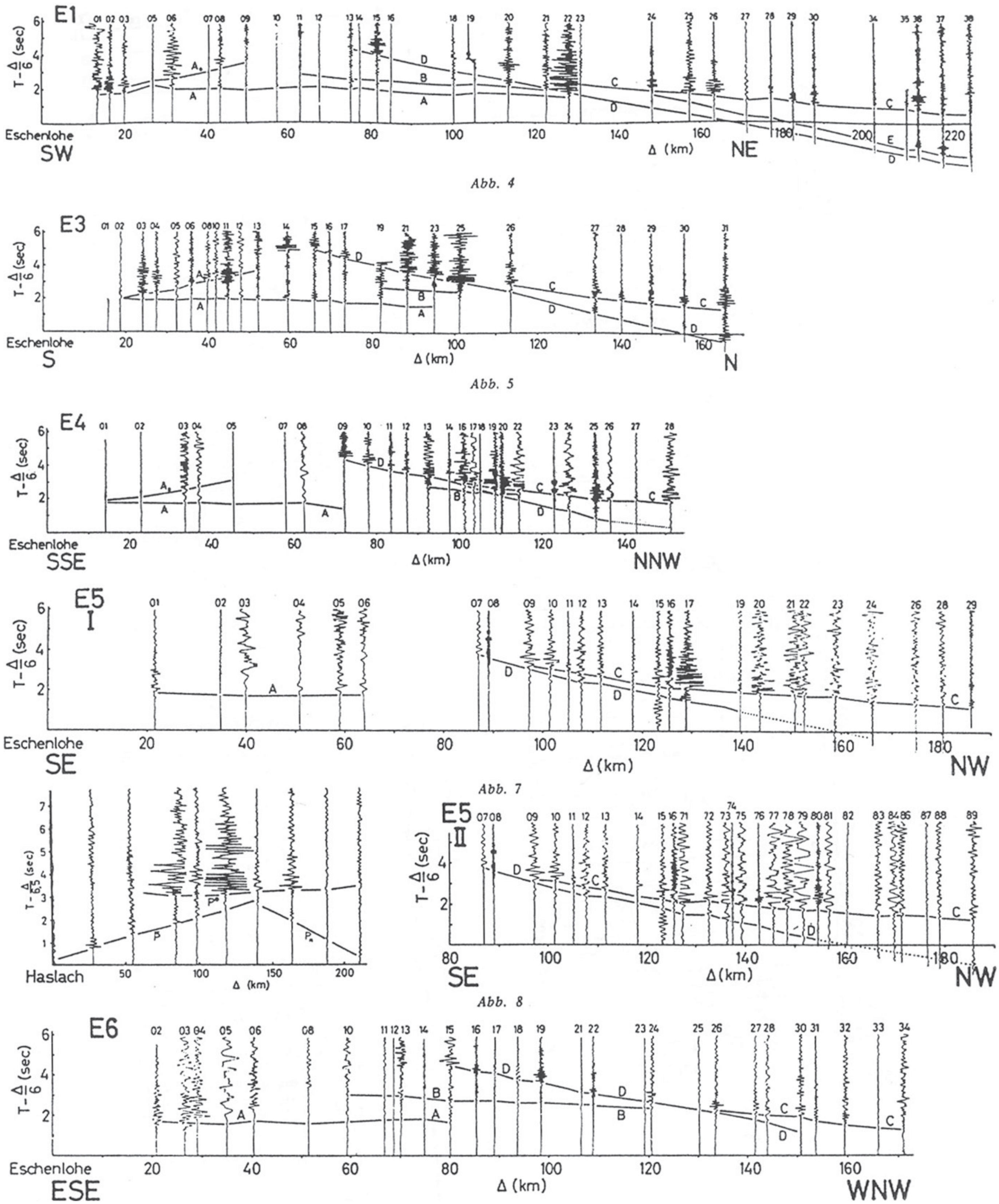


Figure 6.2.2-06. Record sections of data recorded from quarry blasts near Eschenlohe (01 in Fig. 6.2.2-01) across southern Germany (from Prodehl, 1965, figs. 4–9) [Bollettino di Geofisica Teorica e Applicata, v. 7, p. 35–88. Reproduced by permission of Bollettino di Geofisica, Trieste, Italy.]

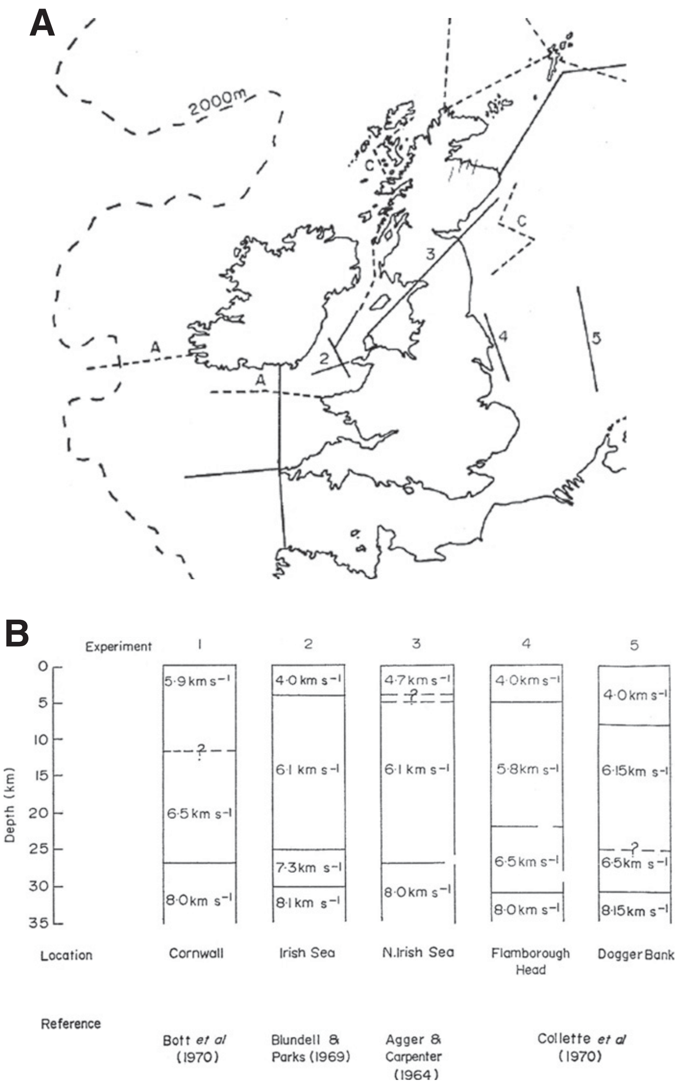


Figure 6.2.3-01. (A) Seismic-refraction profiles in the British Isles and surrounding sea areas (from Bamford, 1971, fig. 1). 1—Bott et al. (1970); 2—Blundell and Parks (1969); 3—Agger and Carpenter (1965); 4–5—Colette et al. (1970); dashed lines A, C and others—unpublished. (B) Associated crustal structure (after Bamford, 1971, fig. 2). Model 1 corresponds to the two unmarked perpendicular lines south of Ireland and west of Cornwall. [Geophysical Journal of the Royal Astronomical Society, v. 24, p. 213–229. Copyright John Wiley & Sons Ltd.]

the use of supercritical reflections to constrain the Moho depth. This was a significant introduction to the analysis of crustal refraction data.

Finally a Continental Margin Refraction Experiment (CMRE) was carried out in 1969 with 68 depth charges fired along lines south and southwest of Ireland and recorded by 11 temporary stations in Ireland, Wales, and England, and 5 sonobuoy stations at sea (A in Fig. 6.2.3-01). The seismic arrivals were recorded at distances up to 400 km. The CMRE shot-station network was designed for time-term analysis, although some of the data was also suited for a classical interpretation as a reversed profile (Bamford, 1971).

Further investigations also included two seismic-refraction profiles on the Dogger Bank in the North Sea (Collette et al., 1970; lines and models 4 and 5 in Fig. 6.2.3-01). The Moho depths of all these surveys were fairly uniform, lying between 27 and 31 km (Bamford, 1971, 1972).

By the end of the 1960s, these marine surveys were extended into the Atlantic to the west of Ireland. In 1969, data were collected across the Rockall Plateau, northwest of the British Isles between 15–20°W and 54–59°N. The velocities of 6.4 and 7.1 km/s for the main crustal layers and the Moho depths varying from 21 to 30 km were regarded as typical for a microcontinent (Scrutton, 1972).

Marine shots were also the prime source for the early crustal investigations of Scandinavia, resulting in a number of profiles recorded in Norway, Finland, and Denmark (Fig. 6.2.3-02).

In 1965, two lines were recorded across the Caledonian mountain range in southern Norway to determine the crustal structure underneath the seismological array NORSAR. These resulted in Moho depths of 36–38 km under the central mountains and 28–30 km under the coast of southwest Norway (Sellevoll and Warrick, 1971). Preliminary results based on various progress reports were summarized in a map for Scandinavia compiled and published in a review of crustal structure in western and northern Europe by Closs (1969).

More details were published by various authors in the proceedings of a colloquium held in Uppsala in December 1969 (Vogel, 1971) dealing in particular with preliminary results along a Trans-Scandinavian Deep Seismic Sounding project carried out in summer 1969. Nine explosions of 1000–4000 kg TNT at five shotpoints were recorded along five lines through Norway, Sweden, and Finland, resulting in 150 recordings (data examples in Figs. 6.2.3-03 to 6.2.3-05). The average station spacing on all profiles on land except for the Kola profile was between 7 and 15 km. P-wave velocities and Moho depths (Fig. 6.2.3-02) increasing from 30 km at the coast of Norway to more than 40 km under central Sweden, the northern Baltic Sea, and Finland were later published by Sellevoll (1973).

When by 1971 these results were published, clear seismic sections were being presented (as shown in the text) providing a major advance, the reader being able to assess the phase arrivals, rather than merely being presented with the time-term data as common for work undertaken in the 1960s.

6.2.4. Crustal Research in the Alps, Southern France, and Italy

Following the international cooperative investigation of the Western Alps of 1956, 1958, and 1960 described in the former chapter, the Central and Eastern Alps also became major targets of the international community in the 1960s (Fig. 6.2.4-01).

In 1961 and 1962, a series of shots was organized by C. Morelli in Lago Lagorai located in the Dolomites in northern

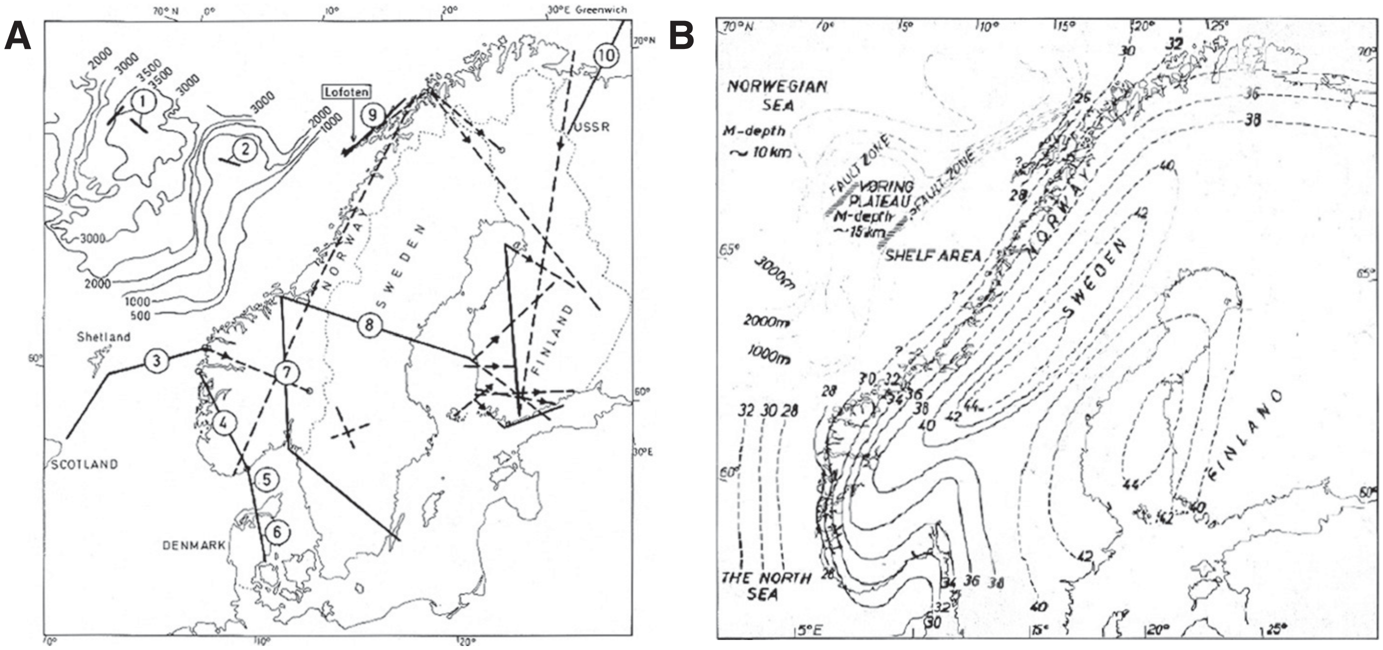


Figure 6.2.3-02. Seismic-refraction profiles (A) and crustal thickness (B, in km) in Fennoscandia (from Sellevoll, 1973, figs. 2 and 3). Full lines = reversed profiles (references in Sellevoll, 1973): 1–3—North Sea profiles; 4—southern Norway; 5–6—Northern Jutland–Skagerrak profiles; 7–8—Trans-Scandinavian project; 9—Lofoten; 10—Kola-Barents Sea. Dashed lines with arrows = unreversed profiles; short cross lines in south-central Sweden = Hagfors observatory array. [Tectonophysics, v. 20, p. 359–366. Copyright Elsevier.]

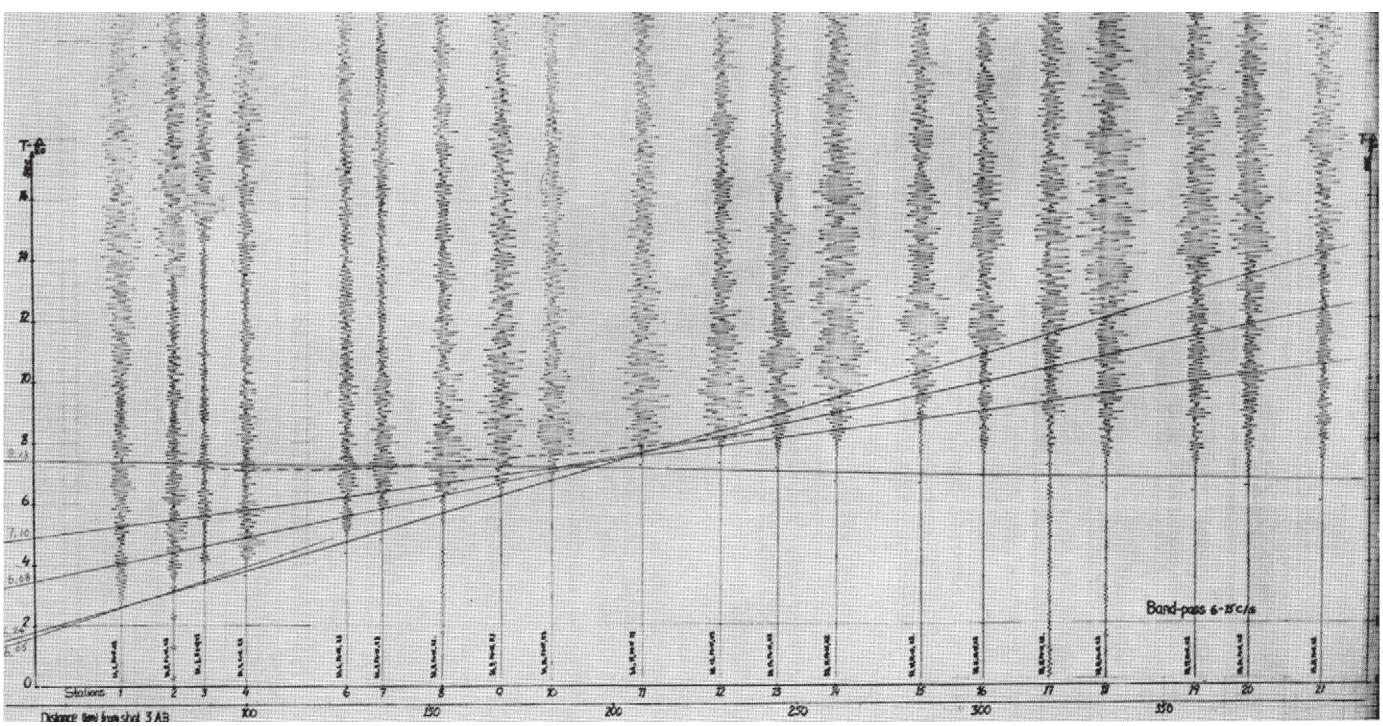


Figure 6.2.3-03. Seismogram section for line 8 (Fig. 6.2.3-02), shotpoint 3 across Norway and Sweden, band-pass filter 6-15 Hz, overlain by linear approximations correlating direct and refracted waves (solid lines) and Moho reflections (dashed lines) of a theoretical model (from Vogel and Lund, 1971, fig. 7). [*In* Vogel, A., ed., 1971, Deep seismic sounding in Northern Europe: Swedish Natural Science Research Council (NFR), Stockholm, 9 p. Permission granted by Swedish Research Council, Stockholm.]

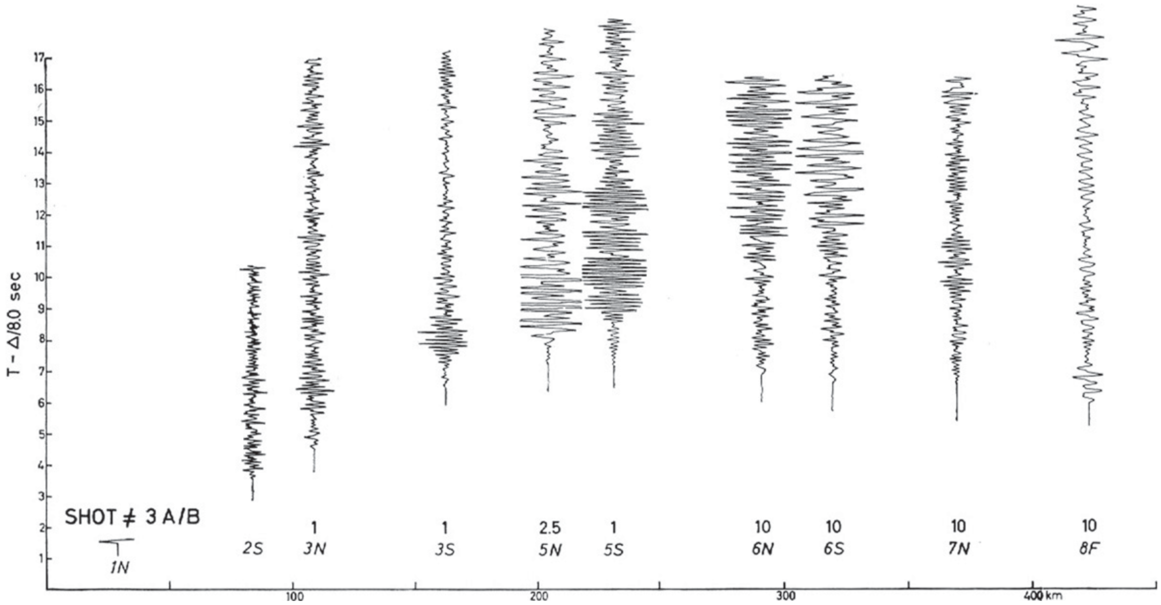


Figure 6.2.3-04. Seismogram section for line 7 (Fig. 6.2.3-02), northern shotpoint, along southeastern Norway, (from Kanestrom and Haugland, 1971, fig. 2). [*In* Vogel, A., ed., 1971, Deep seismic sounding in Northern Europe: Swedish Natural Science Research Council (NFR), Stockholm, 98 p. Permission granted by Swedish Research Council, Stockholm.]

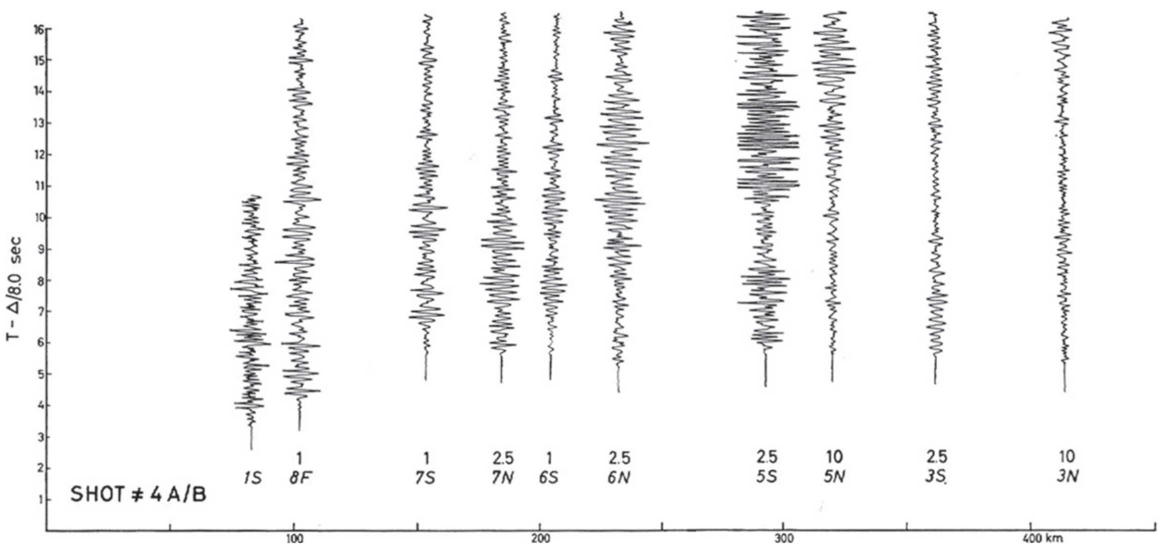


Figure 6.2.3-05. Seismogram section for line 7 (Fig. 6.2.3-02), southern shotpoint, along southeastern Norway, (from Kanestrom and Haugland, 1971, fig. 3). [*In* Vogel, A., ed., 1971, Deep seismic sounding in Northern Europe: Swedish Natural Science Research Council (NFR), Stockholm, 98 p. Permission granted by Swedish Research Council, Stockholm.]

Italy, which were recorded on several lines toward the east, west, south, and north, the northern line reversing the quarry blast shots of Eschenlohe (Prodehl, 1965; data examples in Fig. 6.2.4-02). In 1963 and 1964, similar experiments were organized by A.E. Süsstrunk, centered on a series of shots in Lago Bianco near the northern end of the anomalous zone of Ivrea (Behnke, 1971; Behnke and Giese, 1970; data examples in Fig. 6.2.4-03). All record sections published by Giese and Prodehl (1976) are reproduced in Appendix A6-3 with more details on shotpoints and record sections in Appendix A6-4.

The first models for crustal structure of the Eastern Alps showing the Conrad and Moho discontinuities were published by Behnke et al. (1962) and Peterschmitt et al. (1965) using the observations of the Lago Lagorai experiments. From the data of the

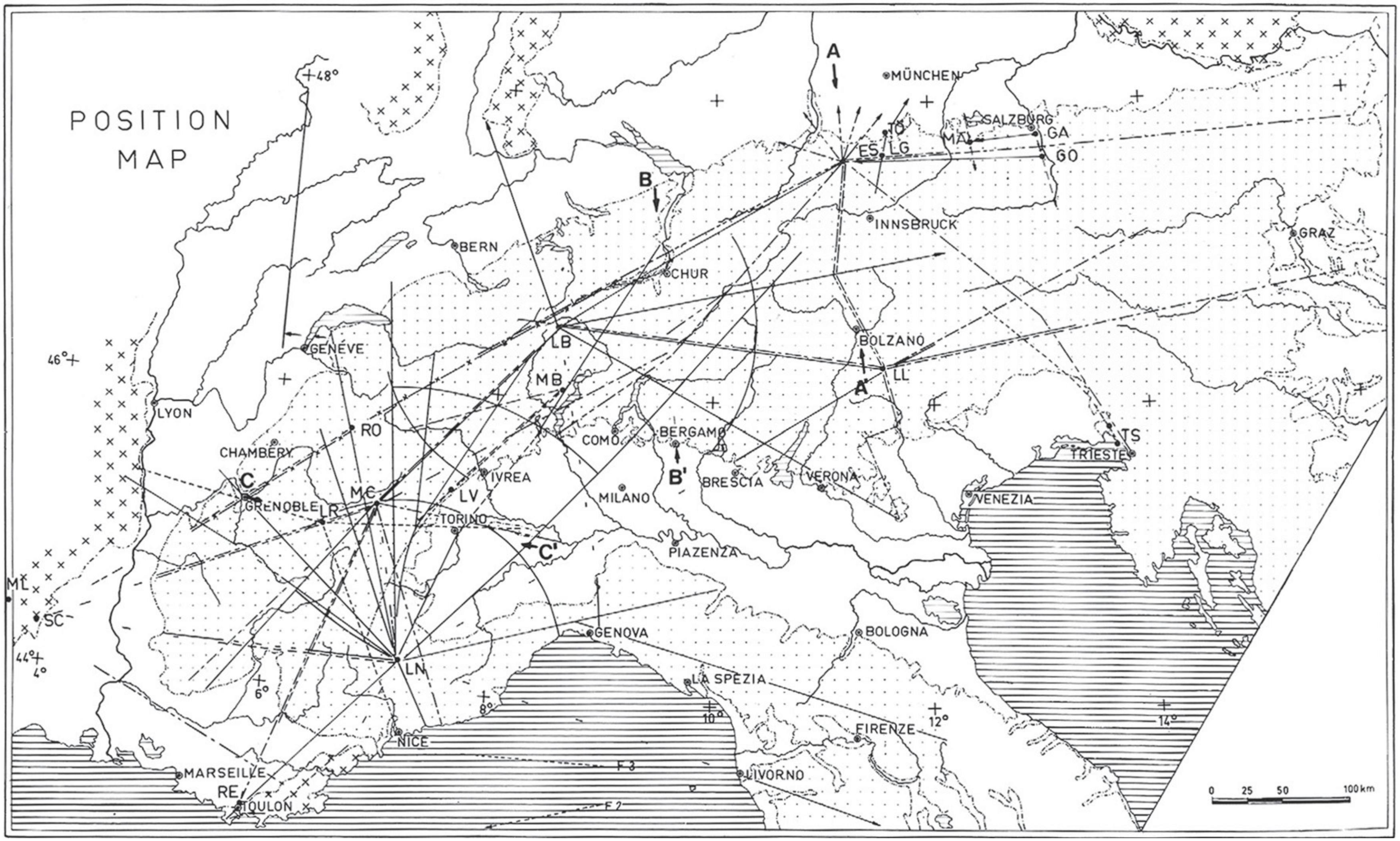


Figure 6.2.4-01. Location of shotpoints and seismic-refraction profiles in the Alps. Shotpoints: ES—Eschenlohe, GA—Gartenau, GO—Golling, LB—Lago Bianco, LG—Lenggries, LL—Lago Lagorai, LN—Lac Nègre, LR—Lac Rond, LV—Levone, MA—Marquartstein, MB—Monte Bavarione, MC—Mont Cenis, ML—Mont Lozère, RE—Le Revest, RO—Roselend, SC—Ste. Cécile d'Andorge, TÖ—Tölz, TS—Trieste (from Giese and Prodehl, 1976, fig. 1). [*In* Giese, P., Prodehl, C., and Stein, A., eds., Explosion seismology in central Europe—data and results: Berlin-Heidelberg-New York, Springer, p. 347–375. Reproduced with kind permission of Springer Science+Business Media.]

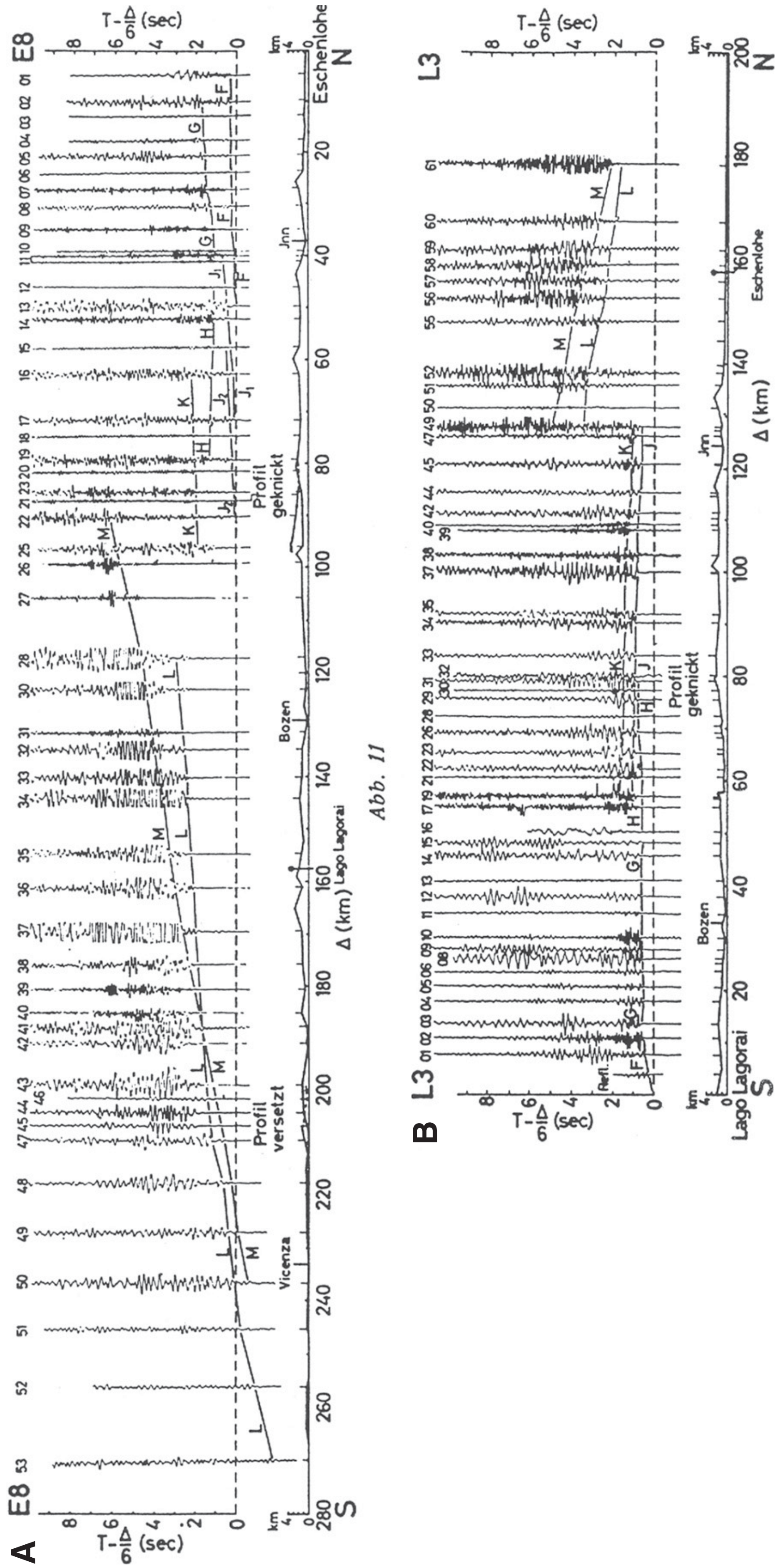


Figure 6.2.4-02. Record sections of profiles from Eschenlohe–Lago Lago Lagorai. (A) Eschenlohe–S; (B) Lago Lagorai–N (from Prodehl, 1965, figs. 11 and 12). [Bollettino di Geofisica Teorica e Applicata, v. 7, p. 35–88. Reproduced by permission of Bollettino di Geofisica, Trieste, Italy.]

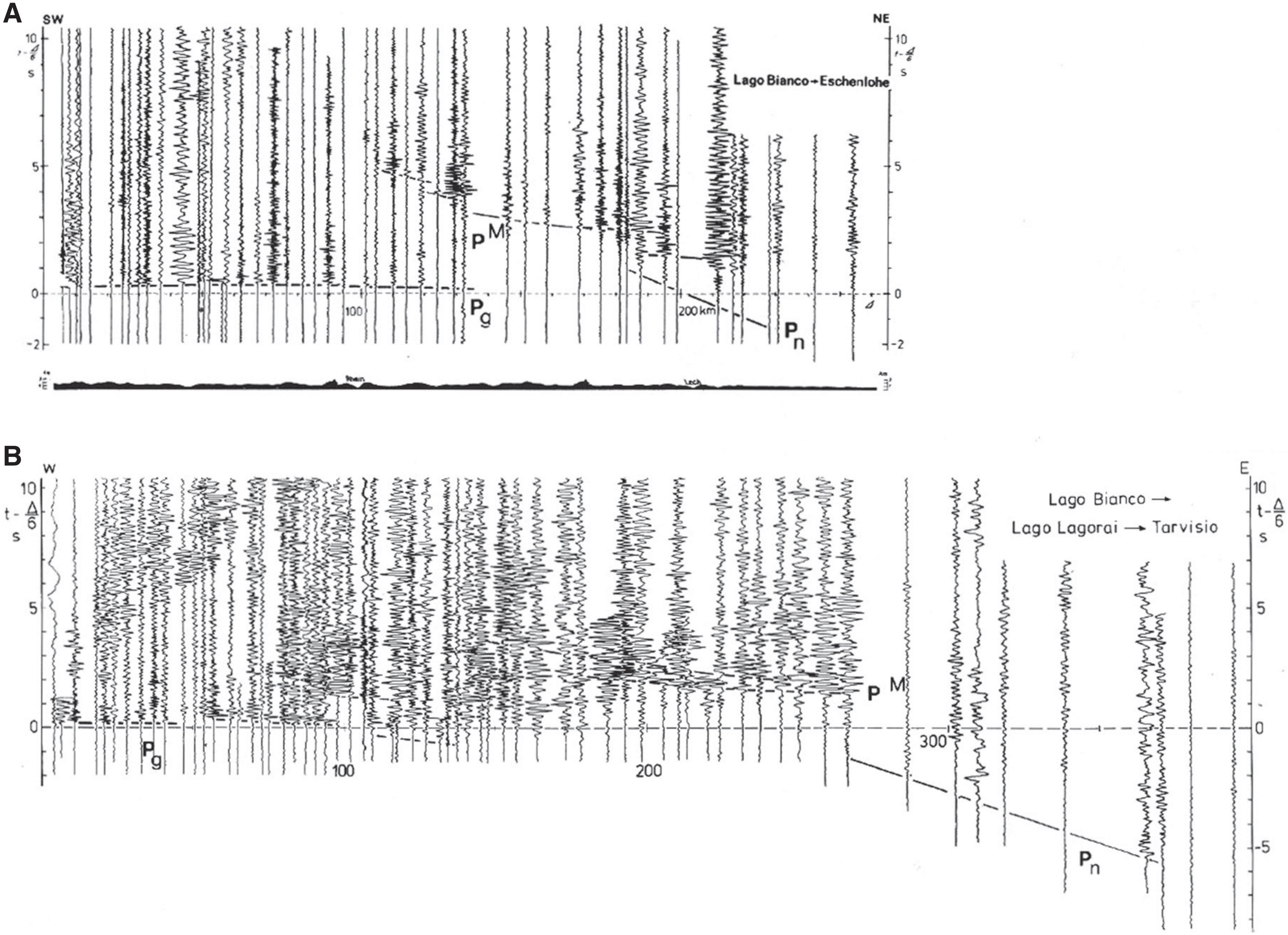


Figure 6.2.4-03. Record sections of profiles from Lago Bianco: top: towards NE, bottom: towards E (from Giese and Prodehl, 1976, figs. 20 and 21). [*In* Giese, P., Prodehl, C., and Stein, A., eds., Explosion seismology in central Europe—data and results: Berlin-Heidelberg-New York, Springer, p. 347–375. Reproduced with kind permission of Springer Science+Business Media.]

reversed profile Eschenlohe–Lago Lagorai, using a wave-front method, Prodehl (1965) constructed a cross section through the Eastern and Southern Alps, introducing a high-velocity layer at the base of the lower crust (7.2–7.4 km/s).

A new concept for Alpine crustal modeling was introduced by Giese (1968a). He interpreted the late appearance of high-amplitude reflections in most of the Alpine profiles as being caused by a strong velocity inversion within the crust. Furthermore, he interpreted the crust-mantle boundary as a transition zone of several kilometers width, where the velocity gradually increases from lower-crustal values to 8 or more km/s.

A special summary of all available data along the Geotraverse Ia through the Eastern Alps was published by Angenheister et al. (1972).

At the same time, from 1961 to 1967, the French seismic-refraction investigations concentrated on the French Massif Central and the Western Alps (Giese et al., 1967, 1973; Labrouste et al., 1968; Perrier and Ruegg, 1973, Appendix A6-4-6), which demonstrated that the crust outside the Alps does not show any anomalous behavior (Fig. 6.2.4-04).

In 1965, Y. Labrouste organized various shotpoints in the Western Alps (Roselend and Mont Cenis, RO and MC in Fig. 6.2.4-01) and in the Massif Central (Ste. Cécile d'Andorge and Mont Lozère, Fig. 6.2.4-04; SC and ML in Fig. 6.2.4-01) (Giese et al., 1967). In 1966, she revived the shotpoint Lac Nègre (LN in Fig. 6.2.4-01) allowing long-range recordings on various lines through the Alps and the Northern Apennines (Röwer et al., 1977).

The peculiarities of the structure of the Ivrea zone caused strong discrepancies in the models of Fuchs et al. (1963a, 1963b) and Labrouste et al. (1963) when the seismic data from 1956, 1958, and 1960 was interpreted. This was discussed in detail and resolved during a symposium in 1968, where several authors (Ansorge, 1968; Berckhemer, 1968; Giese, 1968b) introduced a strong upper crustal velocity inversion underlying the zone of Ivrea and thus explained the phase *m* of Fuchs et al. (1963a, 1963b) as a pure P-wave without excessive Moho depths.

Similar to the earlier interpretation of Labrouste et al. (1963), the anomalous body of Ivrea was now regarded as a mantle wedge gradually detaching from the mantle under the Po plain and intruding the crust under the Western Alps, its surface being at less than 10 km under the eastern edge of the Western Alps and underlain by upper to lower crustal rocks. Reinterpreting all available data, Giese et al. (1971) investigated the areal extension of this anomaly. A cross section through the Western Alps and the anomalous zone of Ivrea as compiled by Giese and Prodehl (1976) is shown in Figure 6.2.4-05.

Choudhury et al. (1971) compiled all data then available and elaborated a structural model for the Alps based on a unique concept of interpretation which was first presented at the General Assembly of the IUGG in 1967. Many record sections, selected cross sections, and contour maps published before 1971 can be viewed in a summary article compiled by Giese and Prodehl (1976) including a contour map of the crust-mantle boundary shown in Figure 6.2.4-06. Under the Western Alps north of Torino, the

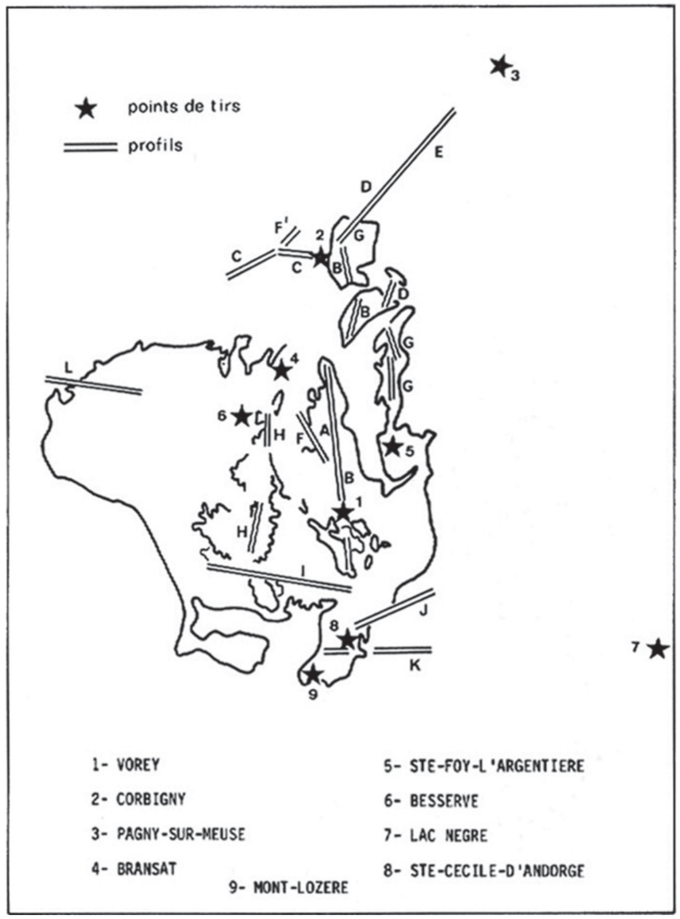


Figure 6.2.4-04. Seismic profiles observed in the Central Massif of France before 1968. 1—Vorey, 2—Corbigny, 3—Pagny-sur-Meuse, 4—Bransat, 5—Ste-Foy-L'Argentiére, 6—Besserve, 7—Lac Nègre, 8—Ste-Cecile-d'Andorge (modified after Perrier and Ruegg, 1973, fig. 1 [Annales de Géophysique, v. 29, p. 435–502.])

crust-mantle boundary of the Po Plain gradually turns over into the Ivrea body, while the deep crust-mantle boundary of the Western Alps under the low-velocity wedge under Ivrea gradually dissolves eastward as shown in detail in Figure 6.2.4-05.

The international cooperation in the Alps, in particular the close cooperation of German and Italian institutions also stimulated a major crustal seismic research program in southern Italy (Giese et al., 1973) in the late 1960s. The first profiles were observed in 1965 and 1966 in Puglia, and two seismic lines were recorded through Sicily in 1968 (Giese et al., 1973). The lines through Sicily were described in much detail by Cassinis et al. (1969). The crustal research was continued in the early 1970s (see Chapter 7, where the location of all projects in southern Italy and Sicily is shown in Fig. 7.2.3-05).

Of particular interest were the long-range profiles across the Apennines of central Italy, recorded in 1966 from the shotpoint Lac Nègre in the Western Alps (see Fig. 6.2.4-01). From the two prominent reflections P^M (A) and P^M (E), correlated between 70

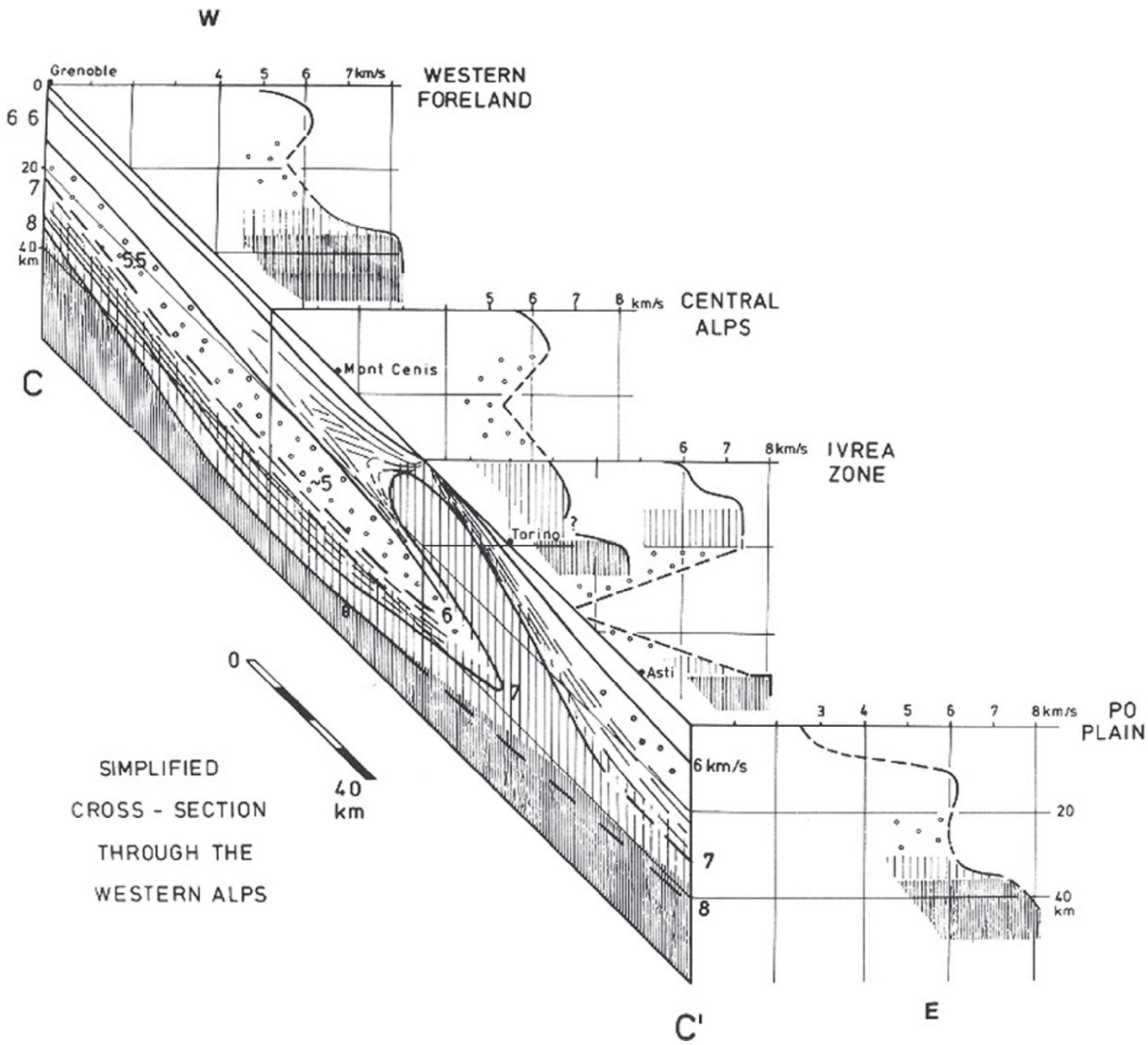


Figure 6.2.4-05. Simplified W-E cross section through the western part of the Alps (*CC′* in Fig. 6.2.4-01) showing velocity-isolines and average velocity-depth functions (from Giese and Prodehl, 1976, fig. 27). [*In* Giese, P., Prodehl, C., and Stein, A., eds., Explosion seismology in central Europe—data and results: Berlin-Heidelberg-New York, Springer, p. 347–375. Reproduced with kind permission of Springer Science+Business Media.]

and 170 km distance (Fig. 6.2.4-07; see also Appendix A7-2-10), the authors derived two overlapping crust-mantle boundaries: the upper crust mantle boundary is the shallow Adriatic crust at 25 km depth in the southeast which is being shifted across the Alpine Moho, and the lower crust-mantle boundary, at 46 km depth in the west and north. Together with crustal structure data of Fahlquist and Hersey (1969) for the adjacent sea areas, Giese and Morelli (1973) were able to construct a Moho map for all of Italy combining the previously published maps for northern and southern Italy.

6.2.5. Advances in Theory and Interpretation

Stimulated by the wealth of data obtained particularly during the 1960s, new methodologies were developed, most of them in close cooperation with or as a consequence of the numerous high-quality data obtained by deep seismic sounding fieldwork, and which have been described in some detail in the previous subchapters.

In Germany, the priority program “Geophysical Investigation of Crustal Structure in Central Europe” had generated refraction/wide-angle reflection observations of high quality and growing and dense quantities. The young generation of scientists, such as Peter Giese at Munich, Rolf Meissner at Frankfurt, Karl Fuchs at Clausthal, Gerhard Müller at Mainz, and others realized very soon, that the previous methods of interpretation, based primarily on traveltimes, were inadequate compared to the richness of information in the new data.

Mueller and Landisman (1966) demonstrated evidence for a world-wide existing crustal low-velocity zone (Fig. 6.2.5-01): the concept might bring crustal seismic-refraction and seismic-

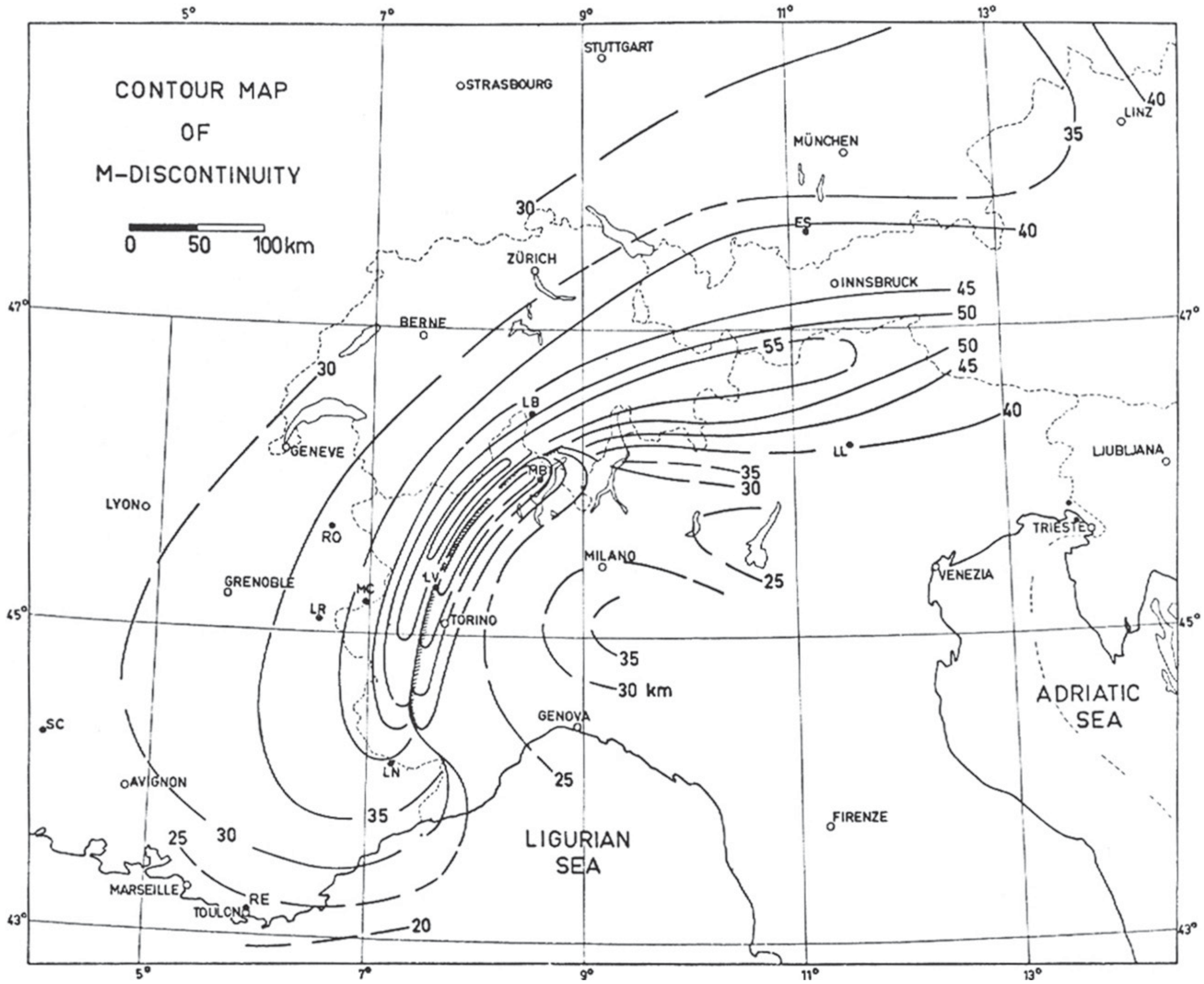


Figure 6.2.4-06. Contour map of the depth of the strongest velocity gradient of the Alpine area in the velocity range 7.5–8.2 km/s, which is regarded as the crust-mantle boundary (from Giese and Prodehl, 1976, fig. 35). [*In* Giese, P., Prodehl, C., and Stein, A., eds., Explosion seismology in central Europe—data and results: Berlin-Heidelberg-New York, Springer, p. 347–375. Reproduced with kind permission of Springer Science+Business Media.]

reflection data into better agreement, as Landisman and Mueller (1966) proved from examples from southern Germany.

Independently, Giese (1968a) investigated in detail the characteristics of reflected waves. In the ideal case of a sharp discontinuity, the reflection appears as a hyperbola of the corresponding traveltime curve in a time-distance diagram. The curvature of the traveltime curve and its position changes when the discontinuity is replaced by a transition zone with a strong, but not infinite, velocity gradient (Fig. 6.2.5-02).

At the same time, P. Giese undertook an approach to explain delays of phases versus the foregoing phases as could be seen in most of the record sections obtained in central Europe and the Alps (Giese, 1968a; Giese and Stein, 1971). He developed a particular approximation method to rapidly calculate the velocity-depth distribution directly from any given traveltime curve system, including options of low-velocity layers and velocity gradients instead of first-order discontinuities (Giese, 1976b). This method was later applied by Prodehl (1970, 1979) on data from the western United States, where the method is also explained in some detail (see Section 6.5.1).

Based both on wide- and near-angle observations, Meissner (1966, 1967a, 1967b, 1976) discussed the increased reflectivity of the lower crust. In a later, more detailed paper, Meissner (1973) described the "Moho" as a transition zone. Often sparse wide-angle reflection data in many continental areas were used to derive an overall velocity gradient of a few kilometers. From denser near-vertical and subcritical reflection data, Meissner deduced a stepwise character of the gradient zone. From wide-angle P_MP and P_CP arrivals in the Rhenish Massif (LO-profile in Fig. 6.2.2-05), showing a nearly perfect correlation and interpreted by Meissner et al. (1976a), Meissner has plotted vertical reflections versus the velocity-depth function (Fig. 6.2.5-03).

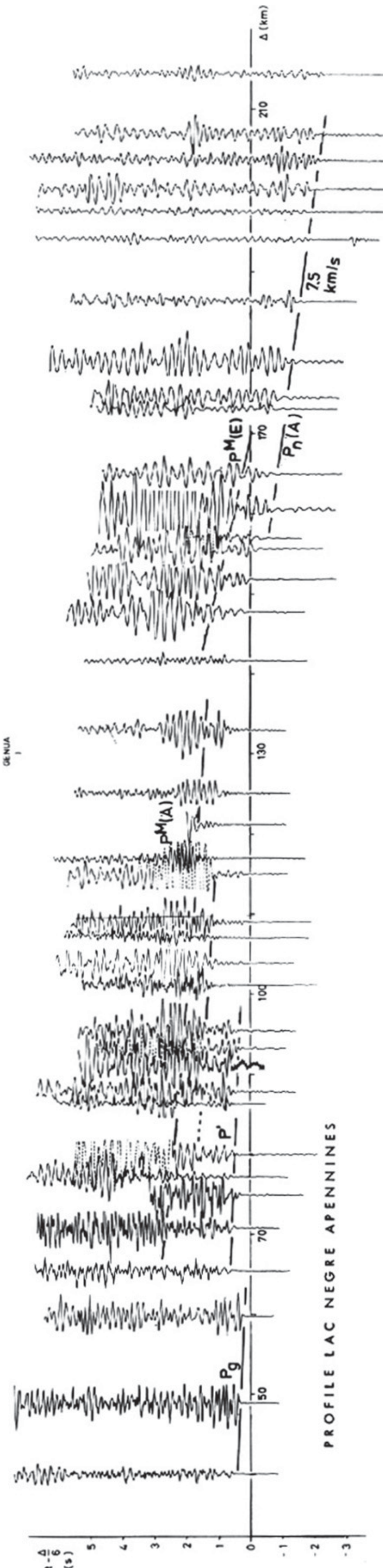


Figure 6.2.4-07. Record section of the profile from Lac Nègre, France, across the Apennines of central Italy (from Morelli et al., 1977, fig. 7.2). [Bollettino di Geofisica Teorica e Applicata, v. 19, p. 199–260. Reproduced by permission of Bollettino di Geofisica, Trieste, Italy.]

The Moho as well as the boundaries CD (Conrad discontinuity) and SCD (sub-Conrad discontinuity) all reflect energy in the near-vertical and the wide-angle range (Fig. 6.2.5-03). The overall picture from different seismic investigations consequently showed the Moho as a laminar transition zone of a few kilometers thick in which the general stepwise increase of the velocity may be interrupted by layers with lower velocities until values around 8 km/s for P-waves are reached in the uppermost part of the upper mantle. As possible petrological explanations, these features may deal with layers of partial melts, crystallization seams, intrusions, and some peeling of mantle matter (Meissner, 1973).

The new computer technology available since the mid-1960s enabled sophisticated traveltime routines by programming the formulas for multi-layered structures, so that models could be checked by recalculating the theoretical traveltimes and comparing them with the observations. Gerhard Müller and Karl Fuchs were not satisfied with traveltime calculations alone, but started to develop methods to calculate synthetic seismograms. A first approach was published by Fuchs (1968).

Initially, K. Fuchs and G. Müller, working separately and independently at different places and having different and common forerunners and academic teachers, developed the basic ideas for the reflectivity method (e.g., Fuchs, 1968, 1969; Müller, 1970), which successfully enabled the calculation of the dynamics of the seismic wave field, i.e., amplitudes, frequency contents and reverberations. Subsequently, they jointly programmed and published the method (Fuchs and Müller, 1971; Müller and Fuchs, 1976), which for the first time allowed the computation of whole synthetic seismograms allowing more severe testing of crustal models (Fig. 6.2.5-04).

A very different approach to interpreting seismic data was the time-term method. It was first suggested by Scheidegger and Willmore (1957) and elaborated further by Willmore and Bancroft (1960) to be applied to seismic-refraction data. This method was described in some detail both by Berry and West (1966) and Smith et al. (1966) and was successfully applied to various data sets in Britain and surroundings and for the interpretation of the first-arrival data of the 1963 Lake Superior experiment, which is described in some detail in Section 6.5.2.

6.3. THE INTERNATIONAL DEEP SEISMIC SOUNDING INVESTIGATION OF SOUTHEASTERN EUROPE

6.3.1. Introduction

In 1963, a large seismic crustal project was initiated in eastern and southeastern Europe (Fig. 6.3.1-01), and 13 international profiles were planned traversing southern Russia, the Ukraine, the Black Sea, Bulgaria, Romania, Yugoslavia, Hungary, Czechoslovakia, Poland, and East Germany (Sollogub, 1969). In 1971, five of the international profiles were more or less completed and ready for a first publication (Sollogub et al., 1972, 1973a, 1973b; Vogel, 1971; Radulescu and Pompilian, 1991).

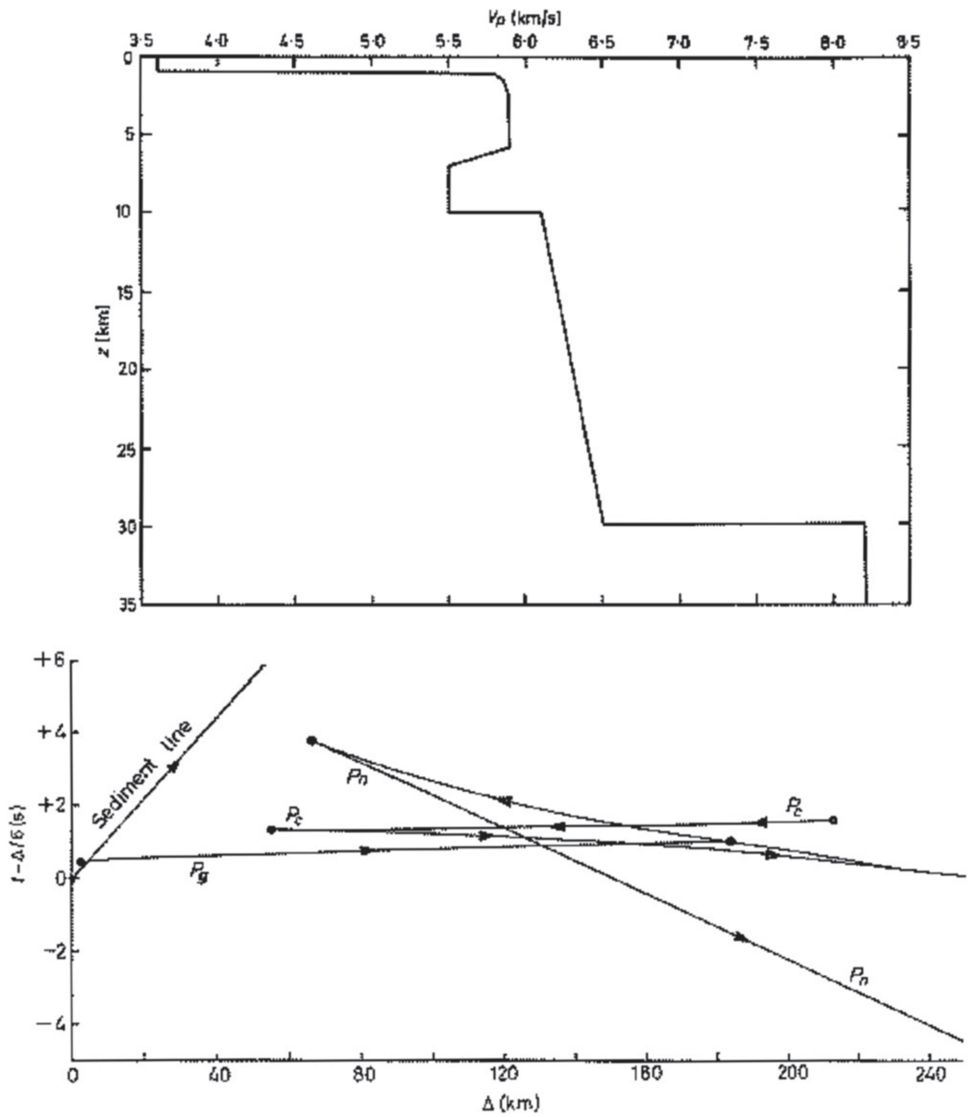


Figure 6.2.5-01. Top: New Crustal Model proposed by Mueller and Landisman (1966) illustrating a well-developed sialic low-velocity zone overlying an abrupt velocity increase. Bottom: Reduced traveltimes for New Crustal Model (from Mueller and Landisman, 1966, fig. 2). [Geophysical Journal of the Royal Astronomical Society, v. 10, p. 525–538. Copyright John Wiley & Sons Ltd.]

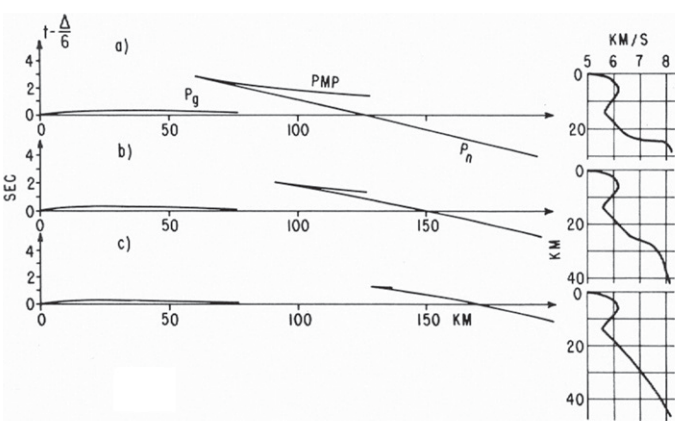


Figure 6.2.5-02. Traveltime diagrams and corresponding models for different structures of the crust-mantle boundary (after Giese, 1968, from Prodehl, 1977, fig. 1). [*In* Heacock, J.G., ed., The earth's crust: American Geophysical Union Geophysical Monograph 20, p. 349–369. Reproduced by permission of American Geophysical Union.]

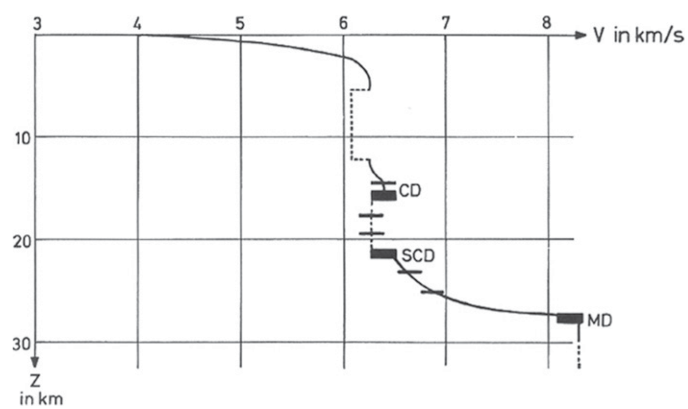


Figure 6.2.5-03. Near-vertical reflections (horizontal bars) in a velocity-depth curve from the Rhenish Massif (profile LO in Fig. 6.2.2-01) (from Meissner, 1973, fig. 10). Discontinuities: CD = Conrad, SCD = sub-Conrad, MD = Moho. [Geophysical Surveys, v. 1, p. 195–216. Reproduced by kind permission of the author.]

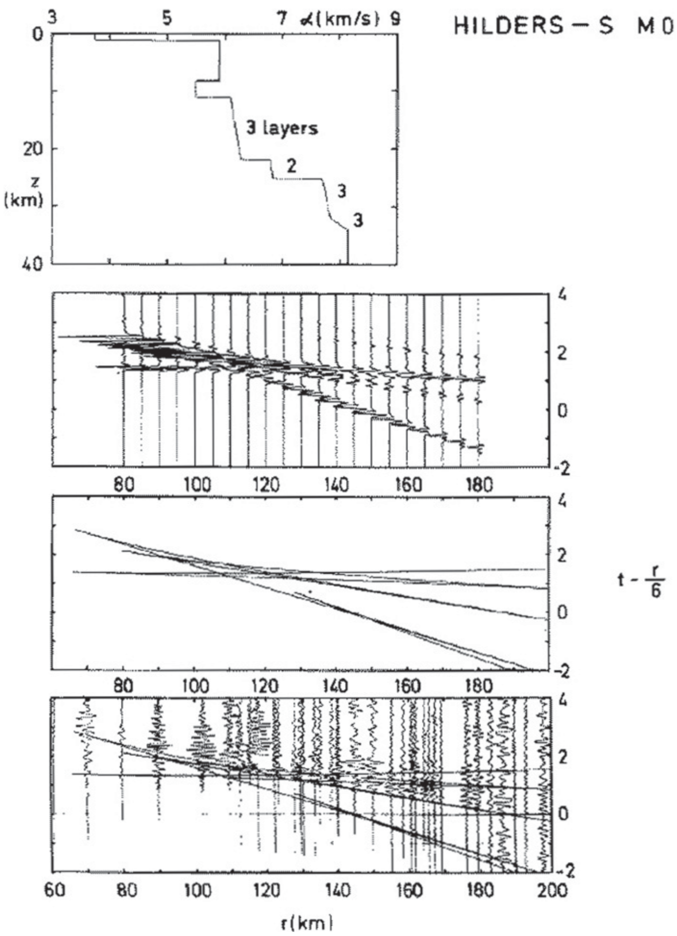


Figure 6.2.5-04. Velocity-depth function, synthetic seismograms (vertical displacement), traveltime curves and traveltime curves overlain on the observed seismograms shown for the example of the profile from Hilders-S (shotpoint 02 in Fig. 6.2.2-01). Only reflections from the bottom of the layers are included (from Müller and Fuchs, 1976, fig. 3). [*In* Giese, P., Prodehl, C., and Stein, A., eds., Explosion seismology in central Europe—data and results: Berlin-Heidelberg-New York, Springer, p. 178–188. Reproduced with kind permission of Springer Science+Business Media.]

The length of the individual profiles varied from 100–110 km to 160–180 km, with shotpoint separations of 20–25 km for the shorter profiles and 40–50 km for the longer lines (Subbotin et al., 1968; Sollogub, 1969; Sollogub et al., 1972). Figure 6.3.1-02 shows a typical seismogram, recorded at 233 km distance. The phase correlation allowed the deduction of detailed traveltime diagrams (see, e.g., Fig. 6.3.3-02), with P^S identifying sedimentary phases and P_g, with P^c and P^M, identifying intracrustal phases (intracrustal waves) from the Moho.

As can be seen in Figure 6.3.1-01, besides the southeastern USSR, other countries in central-eastern and southeastern Europe were also covered fairly well by these international profiles, performance and interpretation of which was executed by the scientists of those countries and discussed in individual chapters of a monograph *The Crustal Structure of Central and Southeastern Europe Based on the Results of Explosion Seismology*, edited by V.B. Sollogub, D. Prosen, and H. Militzer, and published in English in 1972 (Sollogub et al., 1972; see Appendix A6-5).

The fieldwork and the interpretation of the data followed in general a unique style, which was proposed by the Soviet scientists, but details of the organization and realization remained the responsibility of the scientists of each country. The equipment, usually national constructions, and the style of shooting also differed from country to country.

6.3.2. The 1960s in Poland

Jan Uchmann was the first to start explosion seismology in Poland. Aided by Alexander Guterch, he planned in the late 1950s a 485 long seismic N-S traverse through Poland, starting at the Baltic Sea in the Gdansk Bay and ending in the West Carpathians near Raciborz (line A in Fig. 6.3.2-01). The project was finally realized in 1960 and 1961 (Guterch et al., 1973). Eighteen prospecting multi-channel seismic stations allowed an average station spacing of 25 km by recording explosions from three shotpoints. In the north, underwater explosions were carried out in the Gdansk Bay, the southern shotpoint was an old water-filled quarry, and a middle shotpoint was placed in the Konin area (Wojtczak-Gadomska et al., 1964).

This first seismic reconnaissance survey was followed in 1963 and 1964 by a second line B, 242 km long, running in an E-W direction in southern Poland (line B in Fig. 6.3.2-01), again with three shotpoints, the western one of which being the same as used for the N-S line. This time, 15 stations were used. On both profiles, the charges ranged from 500 to 3000 kg (Wojtczak-Gadomska et al., 1964; Guterch et al., 1967).

Since 1965, continuous profiling was introduced using 60-channel recording equipment and geophone spacing of 100 m. This method was first applied on the northeastern section of profile A, and later on a third, 136-km-long profile (line C in Fig. 6.3.2-01). This profile was recorded south of Warsaw, applying continuous profiling with shots from the two end points. The shotpoints consisted of multiple boreholes with 60 kg per hole, totaling a 600–800 kg charge size (Betlej et al., 1967).

In the middle of the 1960s, a special investigation of the fore-Sudetic region was started, aiming primarily to study the sedimentary cover for location of possible oil reserves. A dense network of seismic-refraction lines was carried out with the goal of determining the morphology of the basement (Toporkiewicz, 1986). Most of the lines were recorded in a SW-NE direction, but some were also recorded parallel to the strike of the Tornquist-Teisseyre Tectonic Zone (TTZ). From 1967 onward, magnetic tape recording equipment was used, with distances of 100 m between channels. Shotpoints on the individual lines were ~12 km apart. Shots were fired in grouped boreholes at 40–45 m depth, with charge sizes from 30 to 600 kg. The so-called M-lines were the longest of this dense

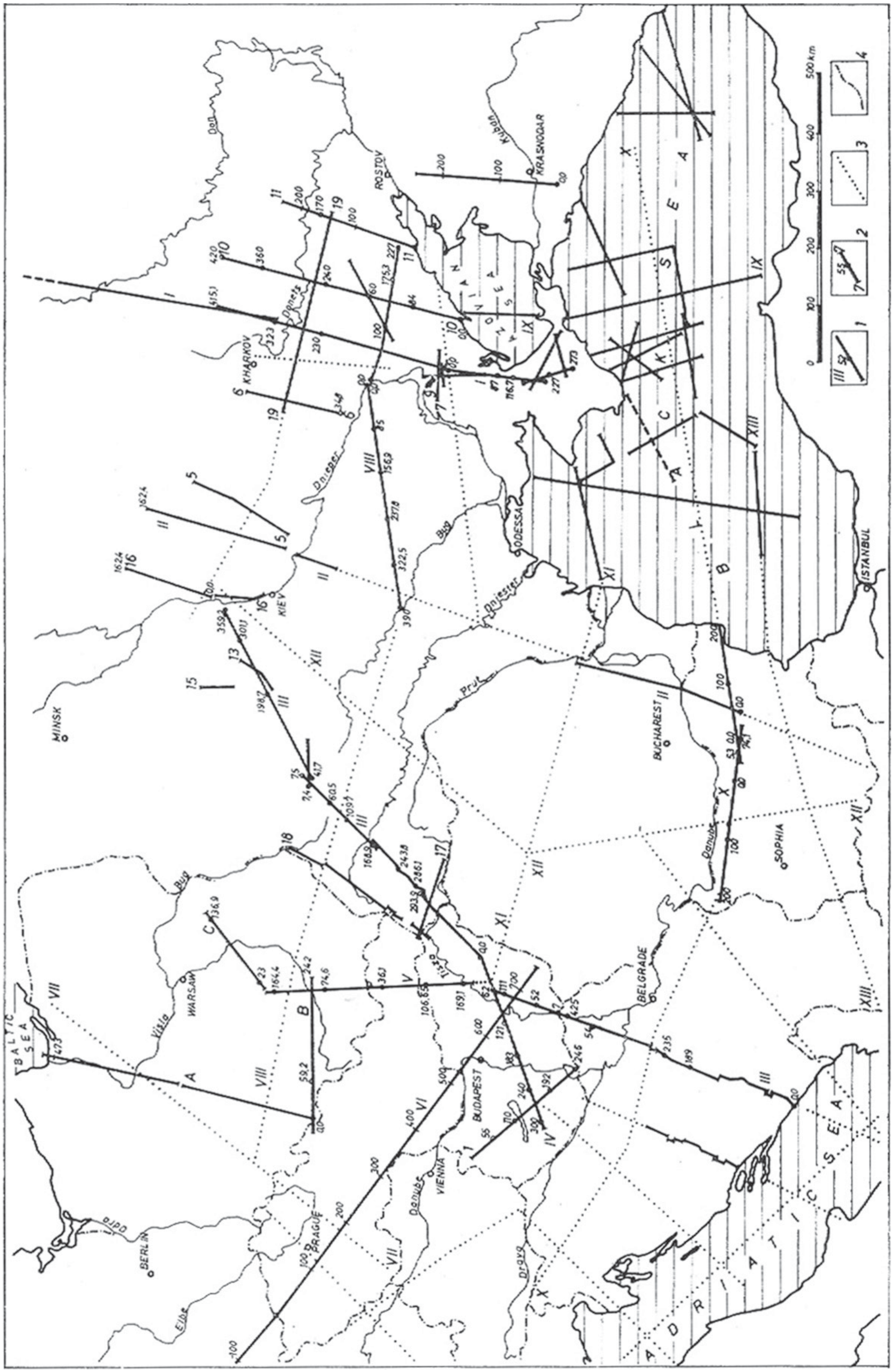


Figure 6.3.1-01. Deep seismic sounding profiles in southeastern Europe (from Sollogub et al., 1972, fig. 1). Solid lines: completed until 1969, dotted lines: planned. [*In* Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

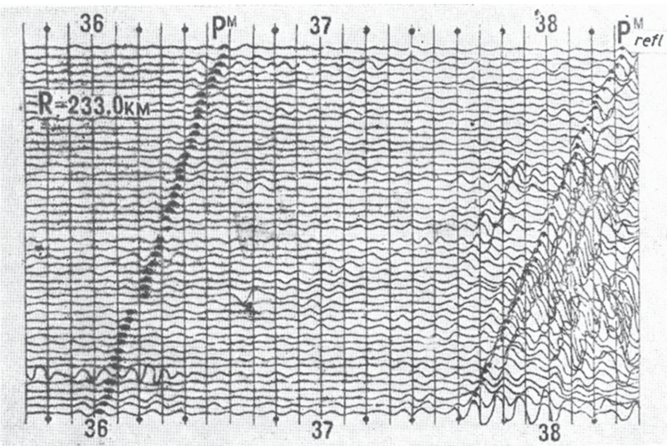


Figure 6.3.1-02. Sample recording showing the refracted P_n (here named P^M) and reflected P_MP phases, observed on the Ukrainian Shield (from Sollogub et al., 1972, fig. 11). [*In* Sollogub, V.B., Prosen, D., and Militzer, H., eds., 1971, Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

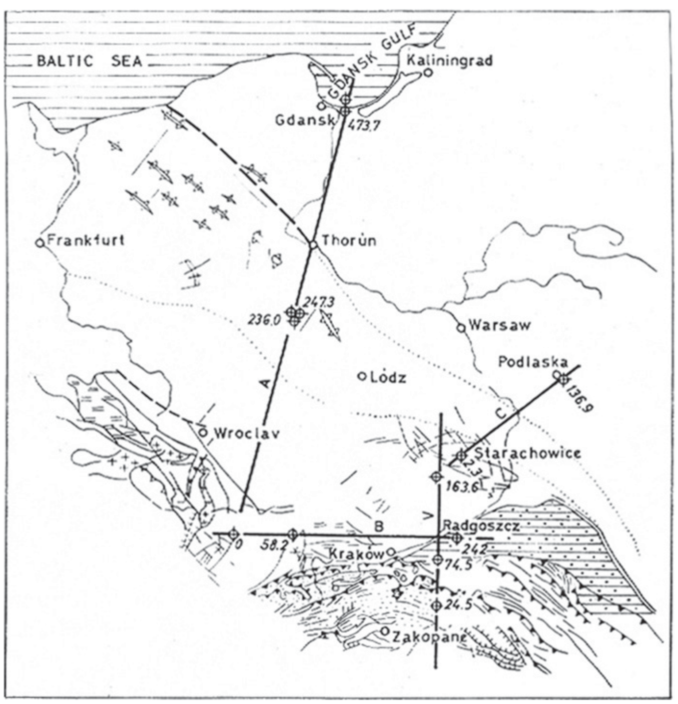


Figure 6.3.2-01. Deep seismic sounding profiles performed in Poland until 1969 (from Sollogub et al., 1972, fig. 29). Stars in circles: shotpoints. The unnamed N-S line in the southeast, leading into the Carpathians, is the International Profile V. [*In* Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

network of lines and served to investigate the fore-Sudetic Monocline, which is part of the Paleozoic platform (Toporkiewicz, 1986; Guterch et al., 1991a). The investigations extended far into the 1970s. This survey therefore will be discussed in more detail in Chapter 7.

Finally, in close cooperation with Russian, Czech, and Hungarian scientists, under the heading of V.B. Sollogub, Kiev, the so-called International Profile V was planned and performed in 1969 (unnamed N-S line in Fig. 6.3.2-01). It started at the East European platform on Polish territory, crossed the western Carpathians, and finally ended in the Hungarian lowlands (Uchman, 1972, 1973). The combined interpretation of this line, as published by Beránek et al. (1972b), is shown in Figure 6.3.3-02.

6.3.3. International Profiles through Czechoslovakia

For Czechoslovakia, three international profiles—V, VI, and VII—had been planned, but in the 1960s, only the two lines V and VI were completed (Fig. 6.3.3-01). Profile VI traversed the southwestern part of the country and continued into eastern Germany. Profile V traversed the Carpathians in the Slovakian part, the interpretation of which is shown in Figure 6.3.3-02 indicating a deep crustal root under the Carpathians at least for southern Poland. Two institutions were responsible for the organization of the fieldwork: the Institute for Applied Geophysics at Brno and the Central Geological Institute at Prague (Beránek et al., 1972a).

6.3.4. Seismic Research in Eastern Germany in the 1960s

One of the Czech lines, profile VI, continued into East Germany, the German Democratic Republic (Knothe and Schröder, 1972). Here, from 1960 to 1967, quarry blasts had been used to study the Earth's crust, mainly along profiles, with scattered observations and different techniques. The work was mainly organized by the Technical University of Freiberg (School of Mines) and the National Geophysical Company at Leipzig. Between 1963 and 1967, this company recorded a few profiles in northeastern Germany, which mainly served the search for hydrocarbons but also obtained some random data from deeper crustal horizons. In 1964, a deep-seismic-reflection survey had also been integrated into the crustal study (Schumann and Oesberg, 1966). The fieldwork along the International Profile VI (Fig. 6.3.1-01) started in 1966 and was completed in 1968. In total, five shotpoints in Germany and in the western part of Czechoslovakia were used to prolong the line by 200 km into eastern Germany. The charges were between 800 and 1200 kg and were either detonated in 40–50 m deep boreholes (1966) or in 4–5 m deep water reservoirs (1968). The system of recording was continuous in-line profiling with 150 m spacing of seismometers, using 8-channel recorders. Responsible for the realization and a first interpretation of the data was H. Knothe and co-workers of the University of Freiberg (Knothe and Walther, 1968).

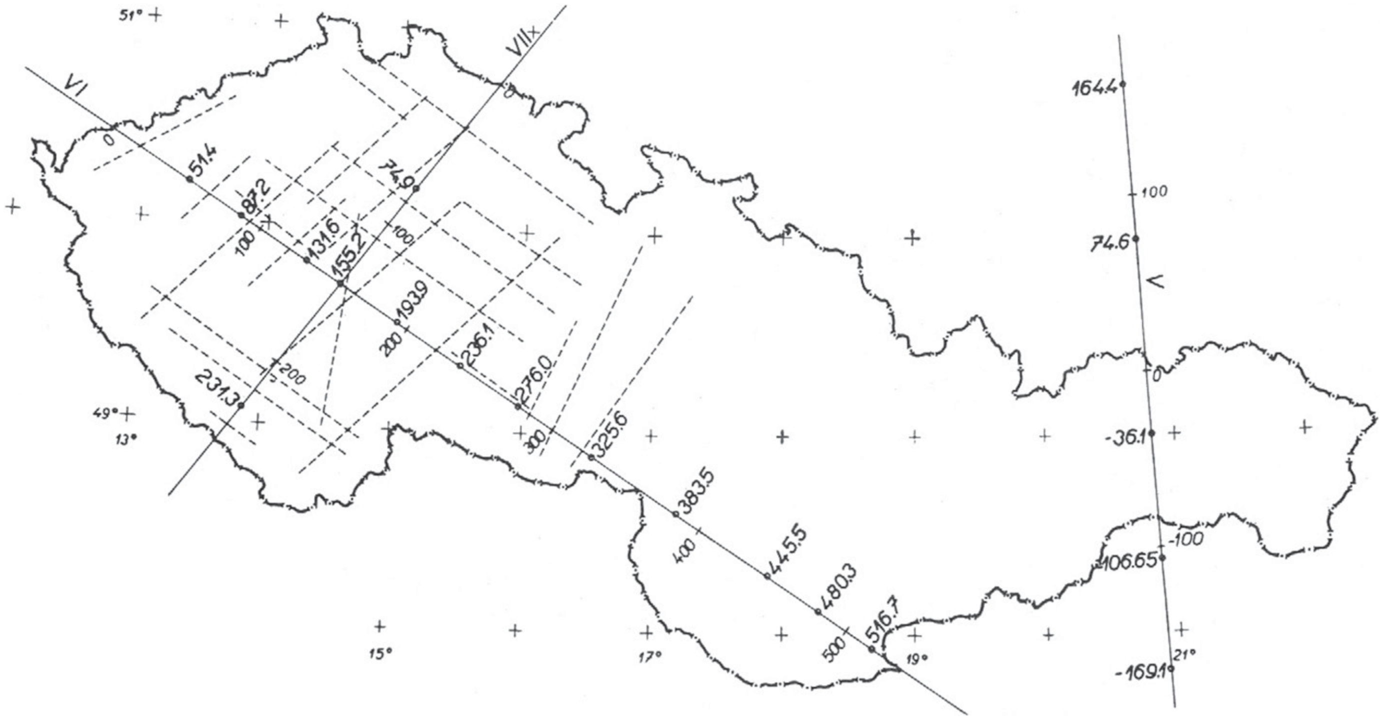


Figure 6.3.3-01. International Profiles recorded on the territory of the CSSR (from Sollogub et al., 1972, fig. 38). [*In* Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

6.3.5. International Profiles through Hungary

In Hungary, four of the International Profiles (III, IV, V, and VI; Fig. 6.3.1-01), crossed each other in order to investigate the anomalous thin crust in much detail (Mituch and Posgay, 1972). As an example of the interpretation of the data, the model along the International Profile V, recorded across the Carpathians from Poland into Hungary, is shown above (Fig. 6.3.3-02).

Already in 1958, a 120-km-long profile had been recorded along the later International Profile VI, but the isolated recordings had not allowed a continuous tracing of phases. Besides the four international profiles, a national NW-SE–running line was also added in western Hungary, traversing Transdanubia from Sopron via Lake Balaton toward the Danube River. Besides in-line continuous and point wise recording, broadside profiling also was applied where two parallel lines, spaced at 60 km, were covered with shotpoints and recording stations. Here, all shots from one line were recorded by four spreads of 2.2 km length each on the other line, resulting in a total length of 8.8 km per shotpoint. The imaginary intermediate profile corresponded to the International Profile. The leading scientist was K. Posgay of ELGI, Budapest.

A preliminary crustal thickness contour map of Hungary constructed from these data showed that the Moho depth ranged between less than 25 km to 28 km, with only one exception north of Lake Balaton in northwestern Hungary (Fig. 6.3.5-01).

6.3.6. International Profile III through the Dinarides of Yugoslavia

The International Profile III (Fig. 6.3.1-01) was continued through the Dinarides of central Yugoslavia to the Adriatic coast (Prosen et al., 1972). Preparations started in 1964 and the fieldwork was realized in 1968 with continuous in-line profiling (150 m geophone spacing) as far as geography allowed, from one single underwater shotpoint only, with charges detonated at 50–100 m depth ranging in size from 100 to 1400 kg, depending on the observation distance.

The project was continued in 1970 with more detailed observations from additional borehole shotpoints. The preliminary interpretation showed a 32-km-thick crust near the Adriatic coast and a crustal root of 52 km under the Outer Dinarides. Further inland, the crustal model showed a gradually thinning to near 30 km in thickness (Fig. 6.3.6-01).

6.3.7. International Profiles through Bulgaria

The territory of Bulgaria was investigated by two crossing lines, the International Profile X running for 550 km distance in an E-W direction from the Black Sea to the Yugoslavian border, and a Profile II, running from the Danube delta in Romania to the south (Fig. 6.3.1-01), which by the end of the 1960s terminated at the intersection with Profile X (Dachev et al., 1972).

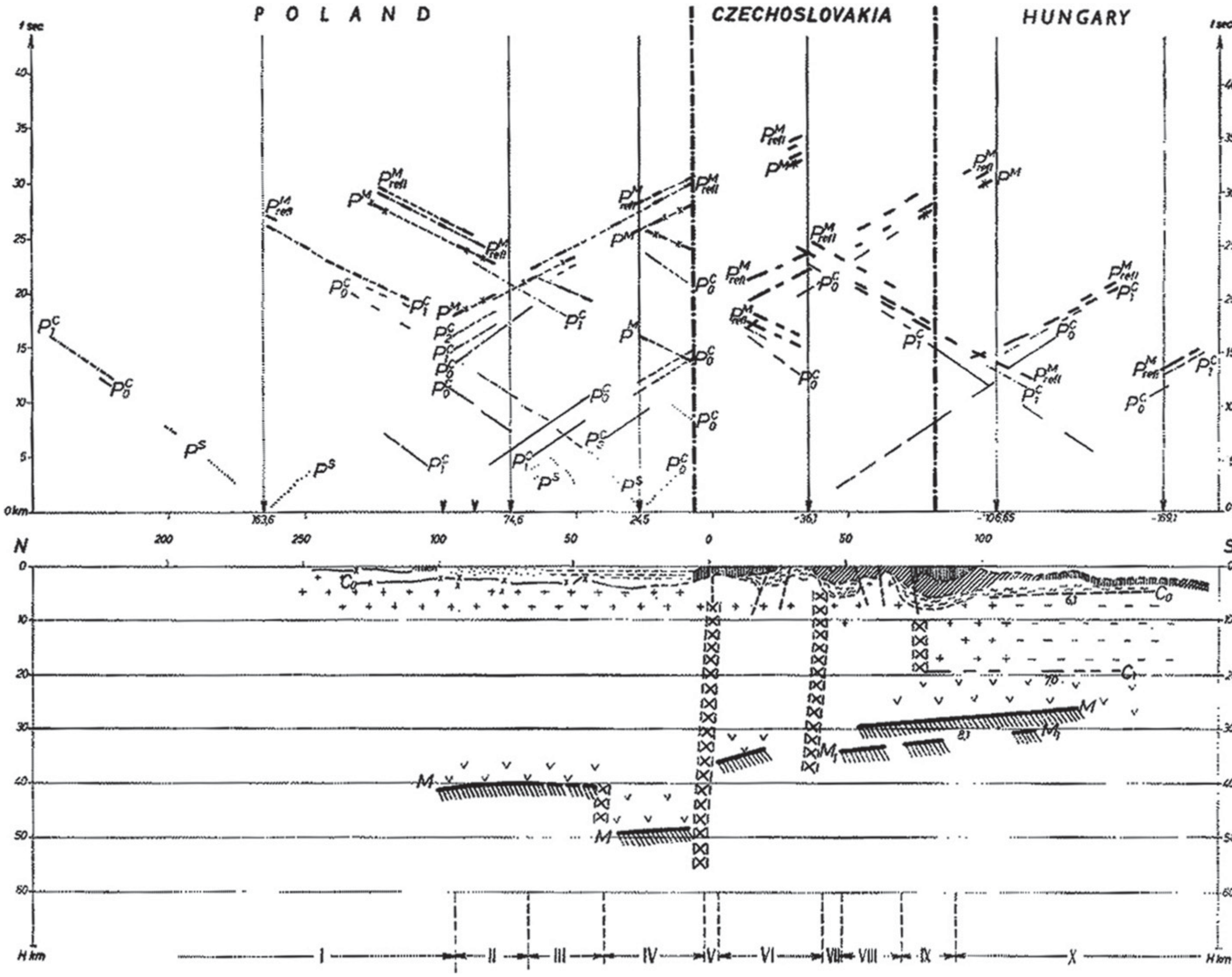


Figure 6.3.3-02. Interpretation of the DSS data of the International Profile V, crossing the Carpathians and reaching from Poland into Hungary (for location see Figs. 6.3.1-01 and 6.3.3-01, from Sollogub et al., 1972, fig. 72). [*In* Sollogub, V.B., Prosen, D., and Militzer, H., eds., 1971, Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

Three institutions were responsible for organizing the project: the Geophysical and Geological Mapping Company, the Physics Department of the University of Sofia, and the Geological Board at Sofia. Small shots spaced between 6.5 and 12 km served to continuously trace the basement, while seven unevenly spaced large shots supplied energy up to 140 km away to obtain seismic waves penetrating the whole crust. The resulting preliminary cross section of Profile X (Fig. 6.3.7-01) showed only scattered pieces of Moho at around 30–37 km depth.

6.3.8. International Profiles through Romania

Following some first crustal structure experiments in 1962 in the central Dobrogea of Romania, fieldwork on the International Profile II in Romania (Fig. 6.3.1-01) was performed in 1967–1970, in cooperation with Soviet and Bulgarian scientists (Constantinescu and Cornea, 1972). The line was mainly observed with continuous profiling, here and there with isolated (critical) soundings to obtain Conrad and Moho arrivals from shotpoints 40–45 km away. The preliminary results showed a 30–35-km-thick crust in the southern part of southeastern Romania, and a 40–45-km-thick crustal section in the northern part, separated by a deep fault (Fig. 6.3.8-01). First point wise observations were also made on Profile XI at the end of the 1960s, which was planned to traverse Romania in ESE-WNW direction.

6.4. SEISMIC CRUSTAL STUDIES IN THE USSR

In the USSR in 1948–1955, G.A. Gamburtsev had designed the deep seismic sounding method which is based on

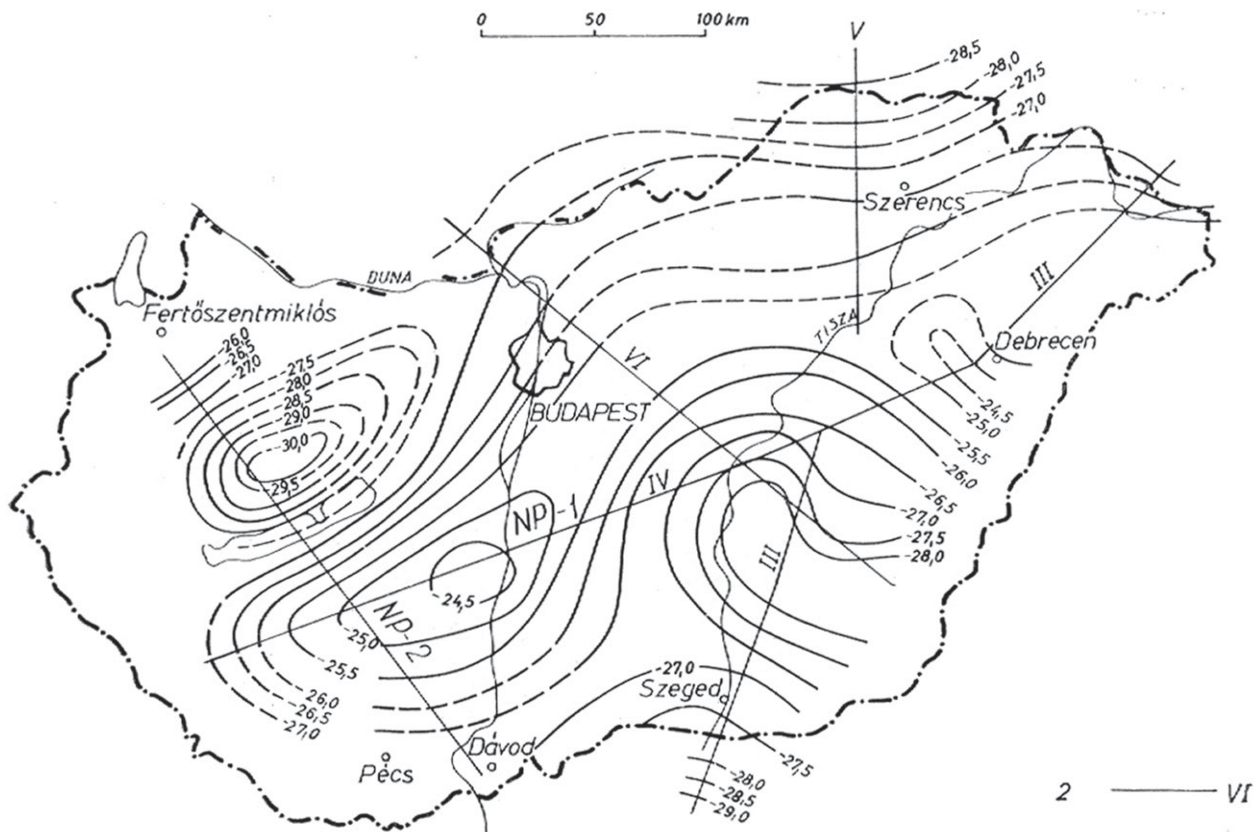


Figure 6.3.5-01. Moho contour map for Hungary and location of seismic DSS profiles, from Sollogub et al., 1972, fig. 70). [*In* Sollogub, V.B., Prosen, D., and Militzer, H., eds., 1971, Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

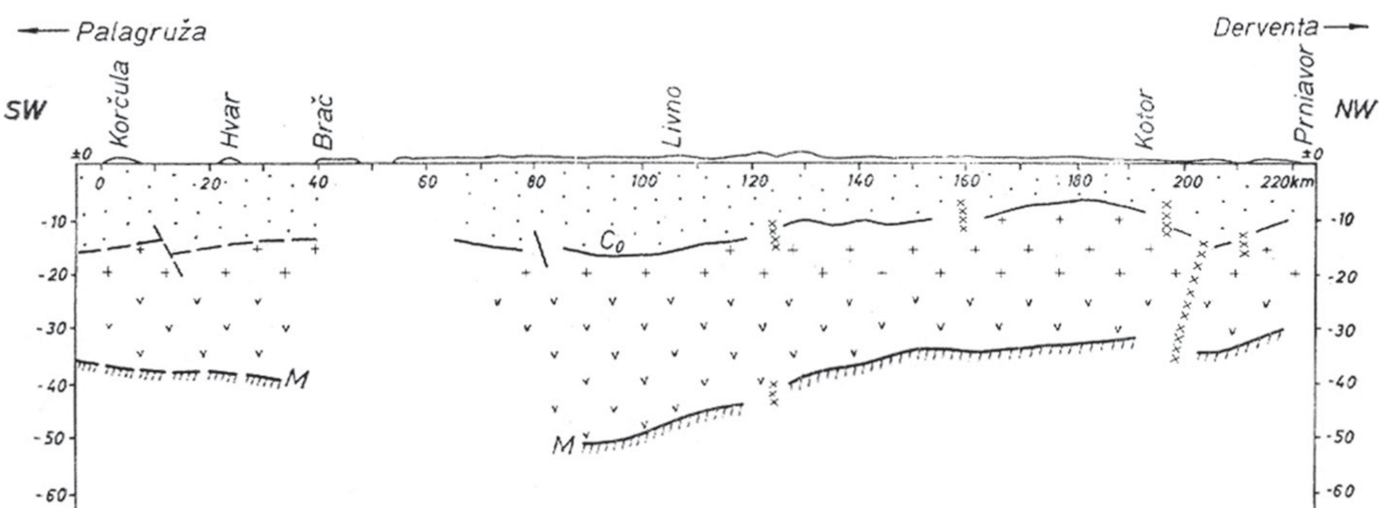


Figure 6.3.6-01. Interpretation of the DSS data of the Yugoslavian part of the International Profile III, across the Dinarides (for location see Fig. 6.3.1-01, from Sollogub et al., 1972, fig. 48). [*In* Sollogub, V.B., Prosen, D., and Militzer, H., eds., 1971, Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

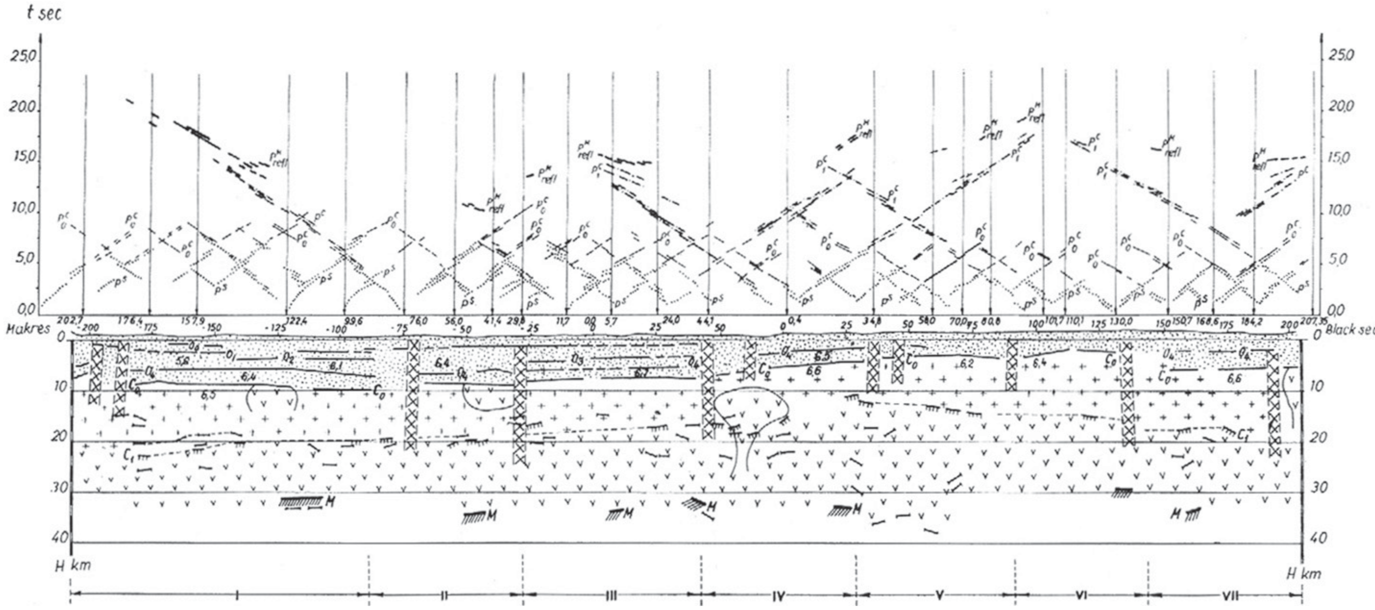


Figure 6.3.7-01. Interpretation of the DSS data of the E-W–directed International Profile X through northern Bulgaria (from Sollogub et al., 1972, fig. 52, for location see Fig. 6.3.1-01). [*In* Sollogub, V.B., Prosen, D., and Militzer, H., eds., 1971, Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

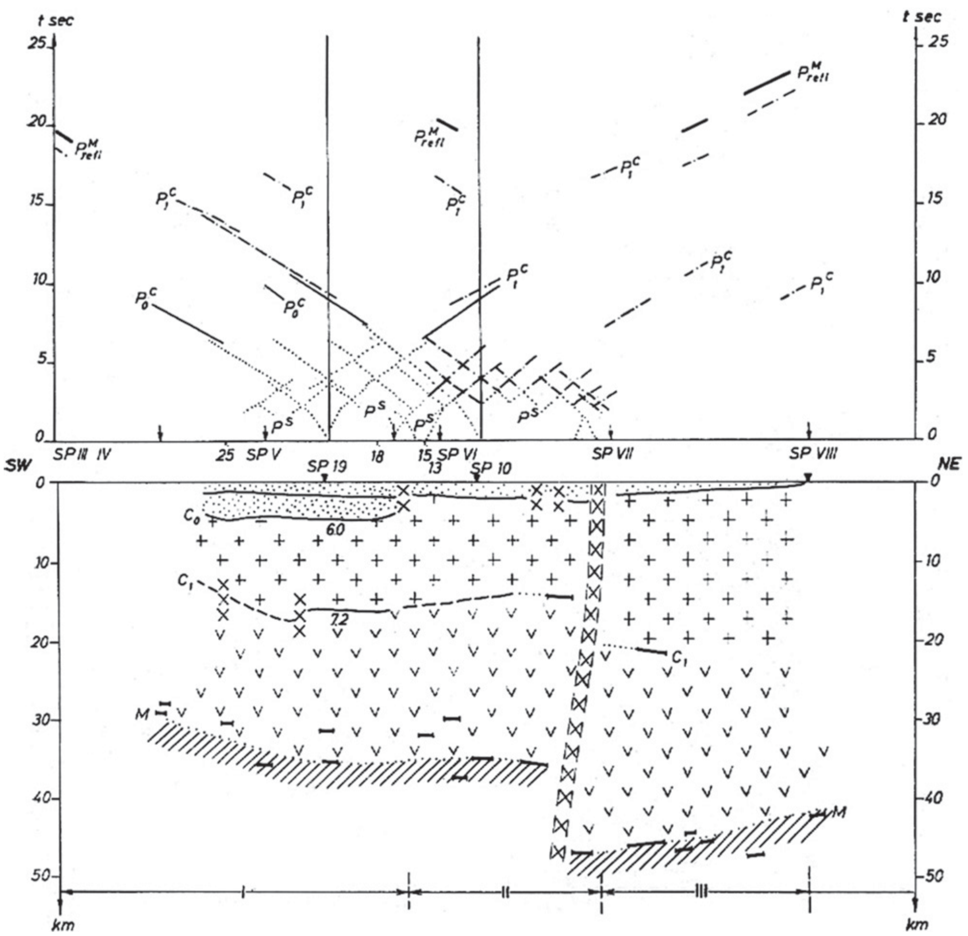


Figure 6.3.8-01. Interpretation of the DSS data of the Romanian part of the International Profile II, across the Dobrogea (for location see Fig. 6.3.1-01; from Sollogub et al., 1972, fig. 56). [*In* Sollogub, V.B., Prosen, D., and Militzer, H., eds., 1971, Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

the correlation of seismic waves generated by small explosions (100 kg charges). The optimal frequency range to produce a depth range of 40–50 km for land investigations that used 1–3 ton charges turned out to be 5–15 Hz. Applying this methodology, seismic profiles of several hundred kilometers in length were recorded across different tectonic zones (Fig. 6.4-01). The explosions were recorded by multichannel seismic prospecting stations with a frequency range of 5–30 Hz using 3–5 Hz seismometers (Kosminskaya et al., 1969). As far as possible, reversed and overlapping observations were obtained, with observation distances ranging between 200 and 300 km and shotpoint intervals of 50–100 km.

As already described in Section 6.3, the crustal profiles in southern Russia and the Ukraine were part of the large international seismic crustal project, which was initiated in 1963 by V.B. Sollogub and co-workers (Fig. 6.3.1-01) and which consisted of 13 international profiles traversing southern Russia, the Ukraine, the Black Sea, Bulgaria, Romania, Yugoslavia, Hungary, Czechoslovakia, Poland, and East Germany (Subbotin et al., 1968; Sollogub et al., 1972; Appendix 6.5), as discussed in some detail in Section 6.3.

As can be seen in Figure 6.3.1-01, the majority of profiles covered the Ukrainian Shield and the Black Sea and delivered a very detailed picture of crustal structure for this part of the USSR (Sollogub and Chekunov, 1972). The recording equipment used in the Ukraine and southern Russia consisted of geophones and low-frequency 60-channel seismic stations with amplifiers allowing the recording of a frequency range of 6–15 Hz (Sollogub, 1969). A typical data set was shown in Figure 6.3.1-02.

A good example, which also shows the complex crustal structure, as interpreted at this time, is the national profile 10, running parallel to International Profile I to the east (Figs. 6.3.1-01 and 6.4-02), crossing the Ukrainian Shield and the Dniepr-Don aulacogen north of the Azovian Sea. More data examples, to illustrate the recorded phases and their interpretations, were published by Subbotin et al. (1968). The corresponding figures are reproduced in Appendix A6-5-2.

In order to compare the data, usually obtained with a geophone spacing of 100 m, and their interpretation of the Russian deep seismic sounding method with the data and interpretations obtained in western Europe at the same time, but with a much wider sensor spacing of 3–5 km in average, Jentsch (1979) has digitized the data of the Russian national profile 10 and presented as record sections with differing sensor spacing: 3 km, 1 km, and 200 m (Fig. 6.4-02, Appendix A6-5-3).

It was obvious that the key phases (reflections from the Moho, for example) could be detected equally well, but that details such as deep crustal faults recognized in the dense data sets with 100 m spacing could not be resolved by station spacings of 1 km or larger. The main crustal boundaries were the Conrad and the Moho discontinuities, but due to the high observation density along parts of the recording lines, the main seismic interfaces were interpreted by Sollogub and Chekunov (1972) as unconformable, i.e., deep reaching faults were introduced which penetrated the whole crust (Fig. 6.4-03).

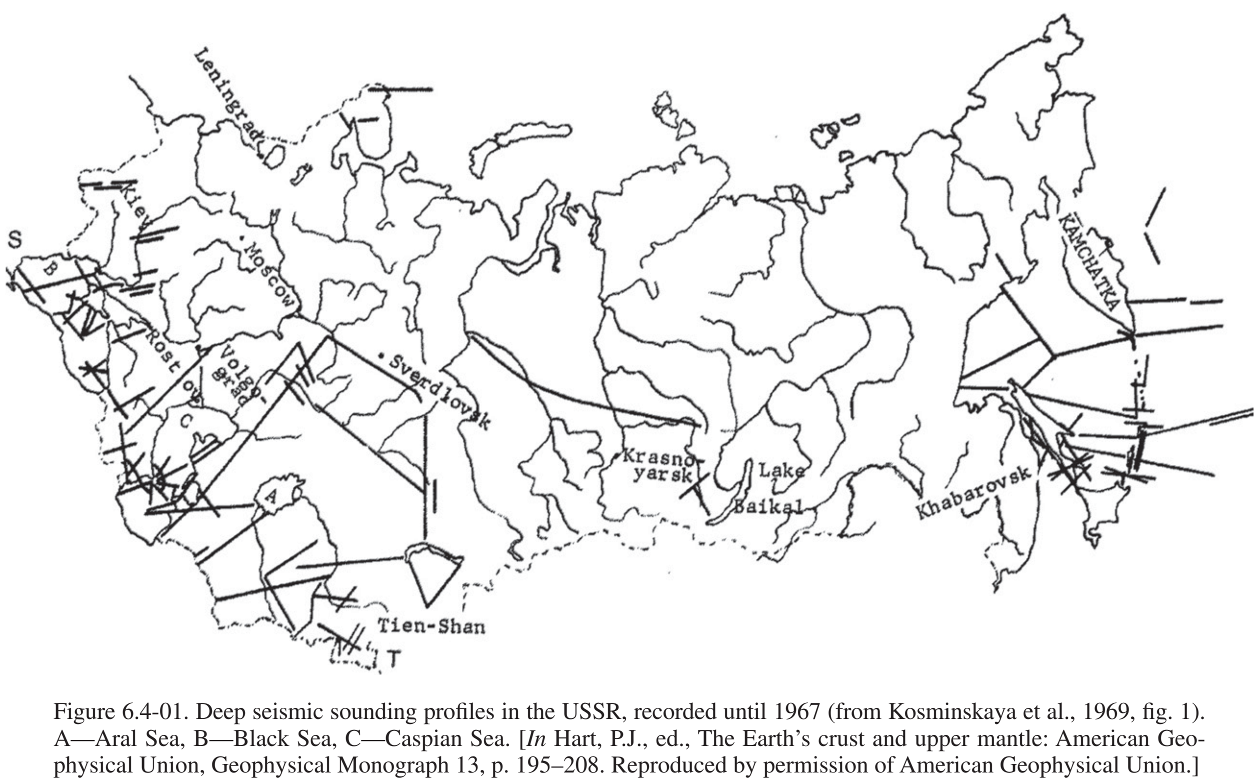


Figure 6.4-01. Deep seismic sounding profiles in the USSR, recorded until 1967 (from Kosminskaya et al., 1969, fig. 1). A—Aral Sea, B—Black Sea, C—Caspian Sea. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 195–208. Reproduced by permission of American Geophysical Union.]

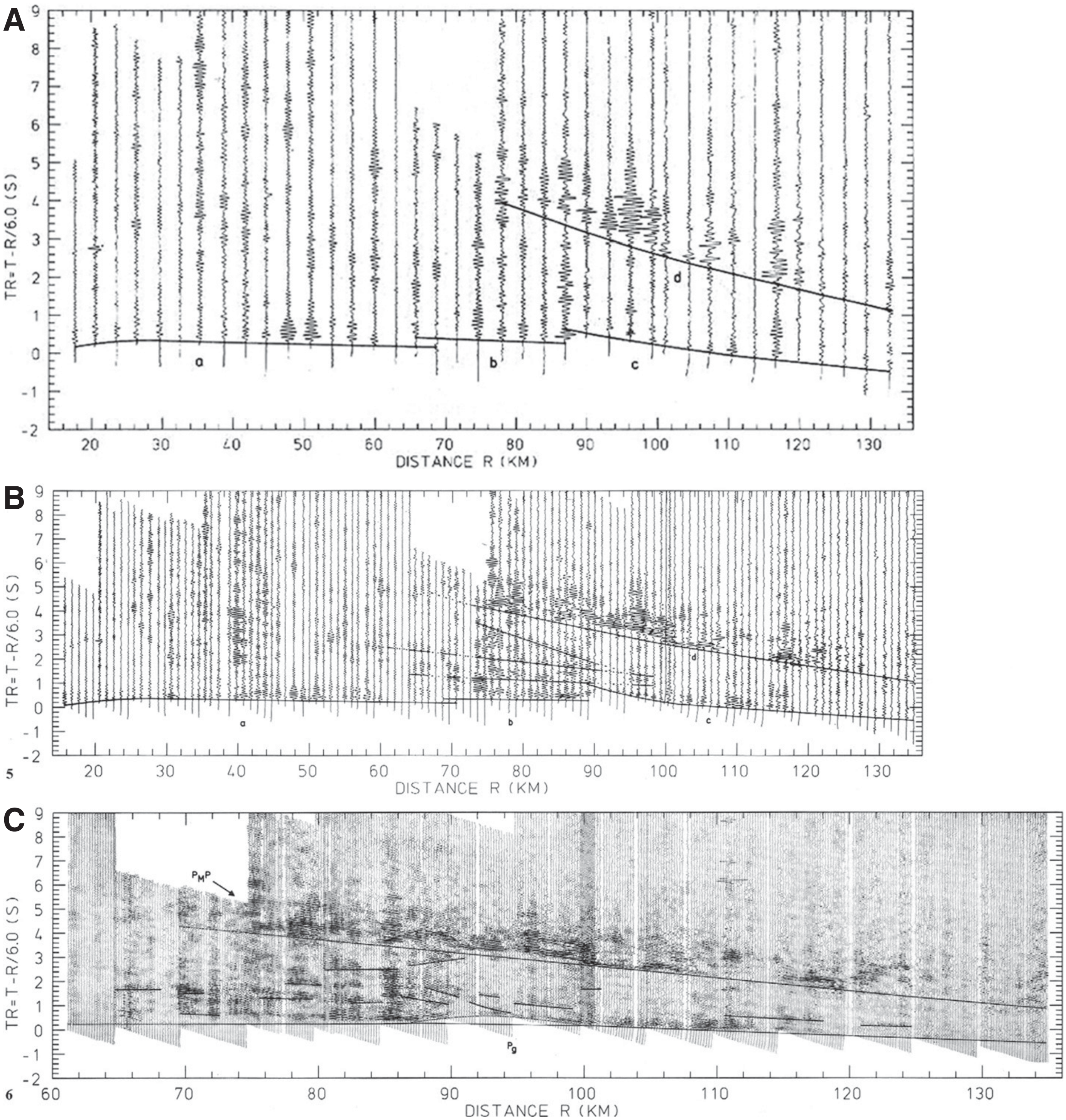


Figure 6.4-02. Record sections of the DSS National Profile 10 across the Ukrainian Shield (line to the east of Profile I in Fig. 6.3.1-01, with a mean spacing of 3 km (A), 1 km (B) and 200 m (C), from Jentsch, 1979, figs. 3, 5, and 6, selected from an original data set of Sollogub et al., 1972). [Journal of Geophysics, v. 45, p. 355–372. Reproduced with kind permission of Springer Science+Business Media.]

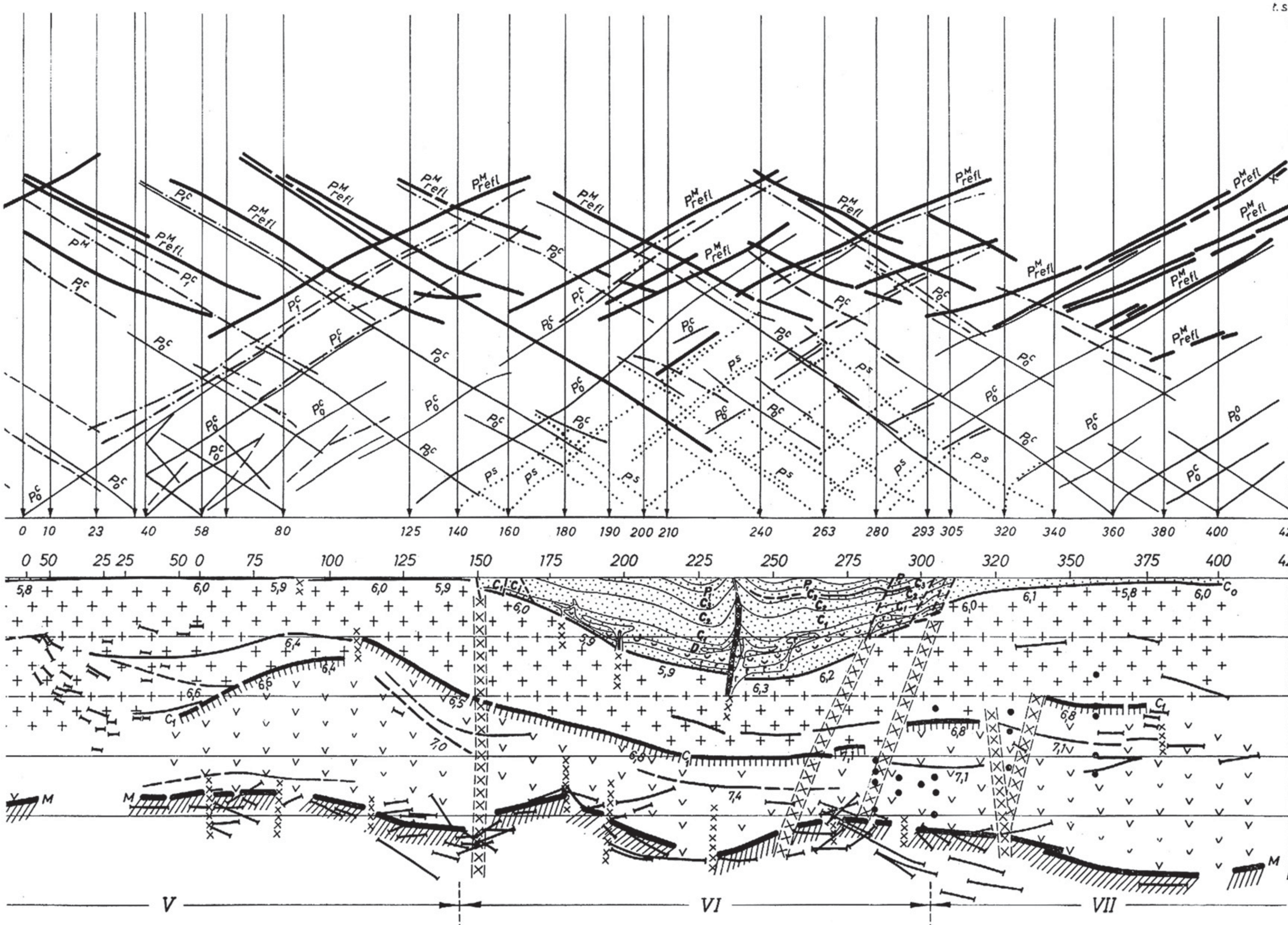


Figure 6.4-03. Interpretation of the DSS data of the part of the National Profile 10 across the Ukrainian Shield (line to the east of Profile I in Fig. 6.3.1-01, from Sollogub et al., 1972, fig. 24). [*In* Sollogub, V.B., Prosen, D., and Militzer, H., eds., 1971, Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

The Moho contour map (Fig. 6.4-04) of Sollogub and Chekunov (1972) displays best the complex structure of the whole area of southern Russia and the Ukraine and wider surroundings, as resulting in the interpretations of the Soviet scientists until the end of the 1960s.

As far as possible, the method of continuous profiling was also applied in other parts of the USSR. Seismic profiles of several 100 km in length were recorded across different tectonic zones, such as old shields, e.g., the Ukrainian (described in some detail above) and Baltic Shields, old and young platforms, mountain regions, sub-montane basins, graben structures, and continental margin areas. Figure 6.4-05 shows a data example recorded in the Fergana intermontane basin, and Figure 6.4-06 shows the observation system and the corresponding interpreted traveltime curves of one of the profiles recorded in the Kopetdag–Central Kyzulkum area. The corresponding distances and traveltimes are included as a table.

In some regions of the Soviet Union, where a complex relief prohibited continuous profiling, a piece-continuous profiling was applied, i.e., the observations were restricted to discrete intervals. In the almost inaccessible taiga regions of Siberia and in the Far East, a special point sounding method was used (Puzyrev et al., 1973, 1978). Here, one explosion was recorded at one point only at a distance optimal for observing a single or several prominent deep-penetrating waves.

Deep-sea investigations of Neprochnov et al. (1970) dealt with the structure of crust and upper mantle under the Black and Caspian Seas. The cross section from the Black Sea to the Caspian Sea (Fig. 6.4-07A) and its continuation to the Tien Shan (Fig. 6.4-07B), which also crosses the Caucasus mountains, shows a summary of all seismic data obtained for the southwestern USSR until the end of the 1960s.

The overall complex structure of the USSR also becomes apparent in Figures 6.4-08 and 6.4-09, where Figure 6.4-08 shows

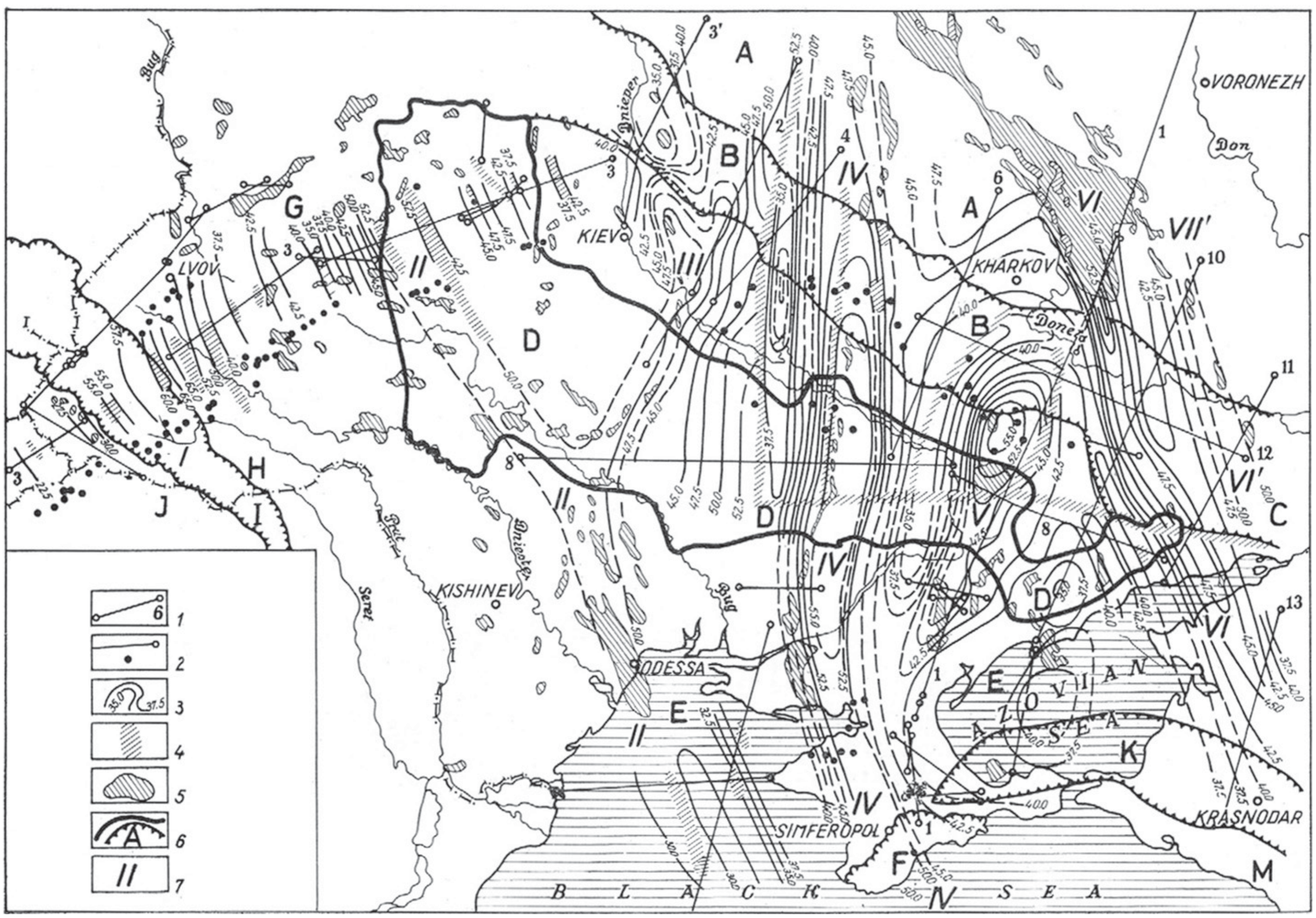


Figure 6.4-04. Crustal thickness map for the Ukraine and the European part of southern Russia and adjacent areas (from Sollogub et al., 1972, fig. 25). [*In* Sollogub, V.B., Prosen, D., and Militzer, H., eds., 1971, Crustal structure of central and southeastern Europe based on the results of explosion seismology (in Russian). English translation edited by György, S., 1972, Geophysical Transactions, spec. ed., Müszaki Könyvkiado, Budapest, 172 p. Reproduced by permission of Eötvös Lorand Geophysical Institute of Hungary.]

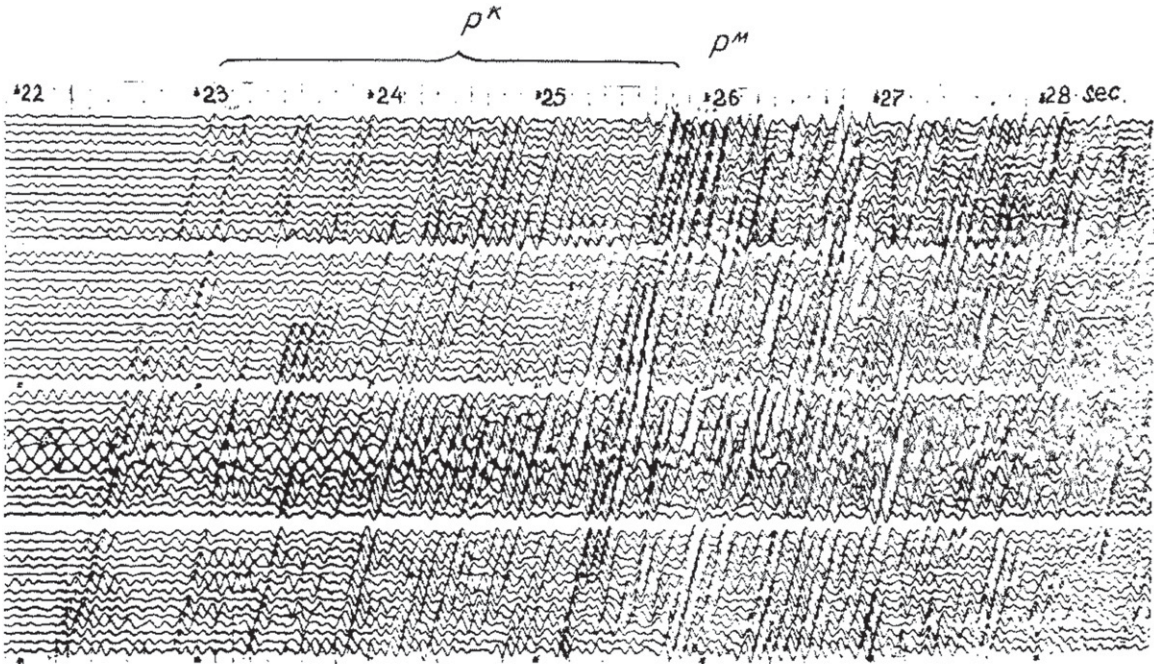


Figure 6.4-05. Seismogram observed in the Fergana intermontane basin at 160 km shotpoint distance (from Kosminskaya et al., 1969, fig. 2). P^k crustal phases, P^M mantle phases. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 195–208. Reproduced by permission of American Geophysical Union.]

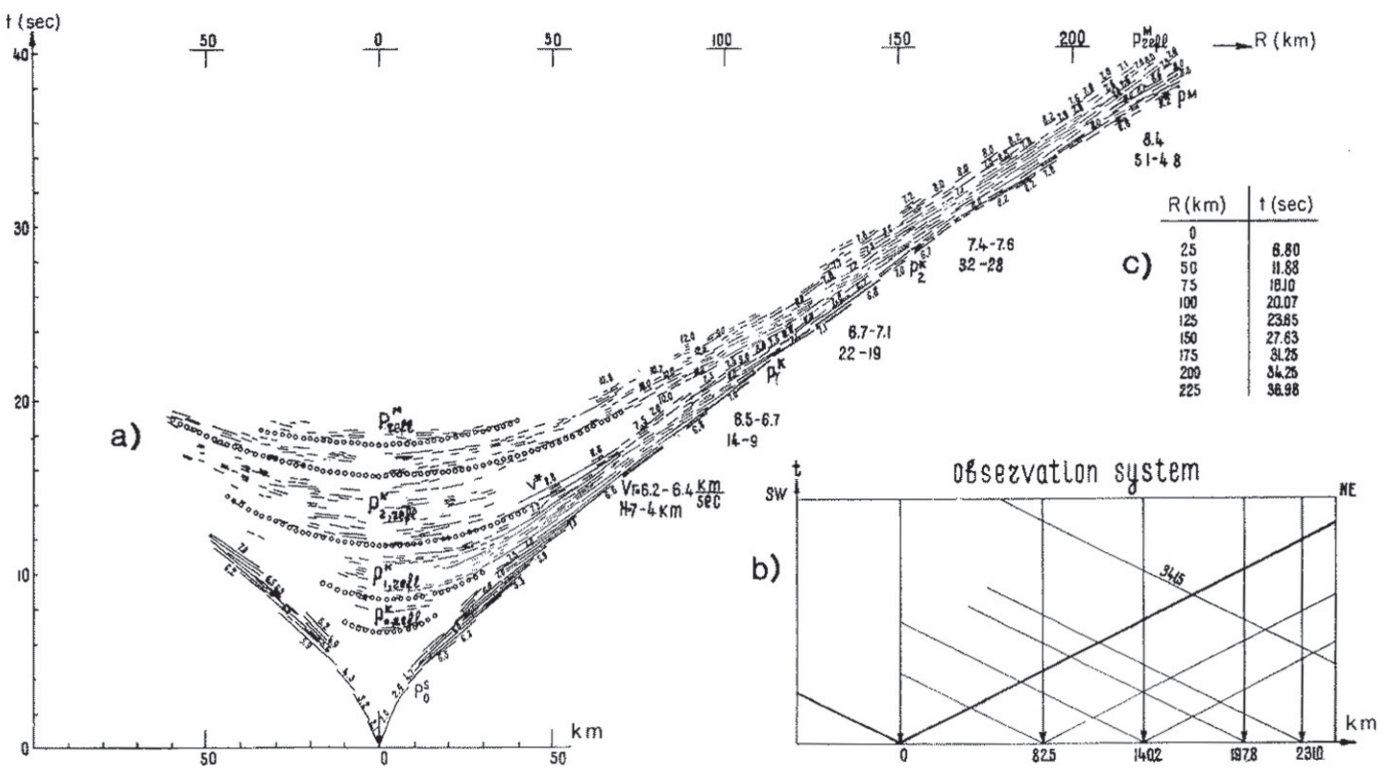


Figure 6.4-06. Interpretation of the DSS data of the profile from Kopetdag to central Kyzylkum (from Kosminskaya et al., 1969, fig. 3). (a) phase travel-time curves of deep waves, showing reflected (index: refl) and refracted P-waves (indices 1, 2 for intracrustal phases k, index M for Pn phase); (b) observation system; (c) average travel times for the first arrivals. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 195–208. Reproduced by permission of American Geophysical Union.]

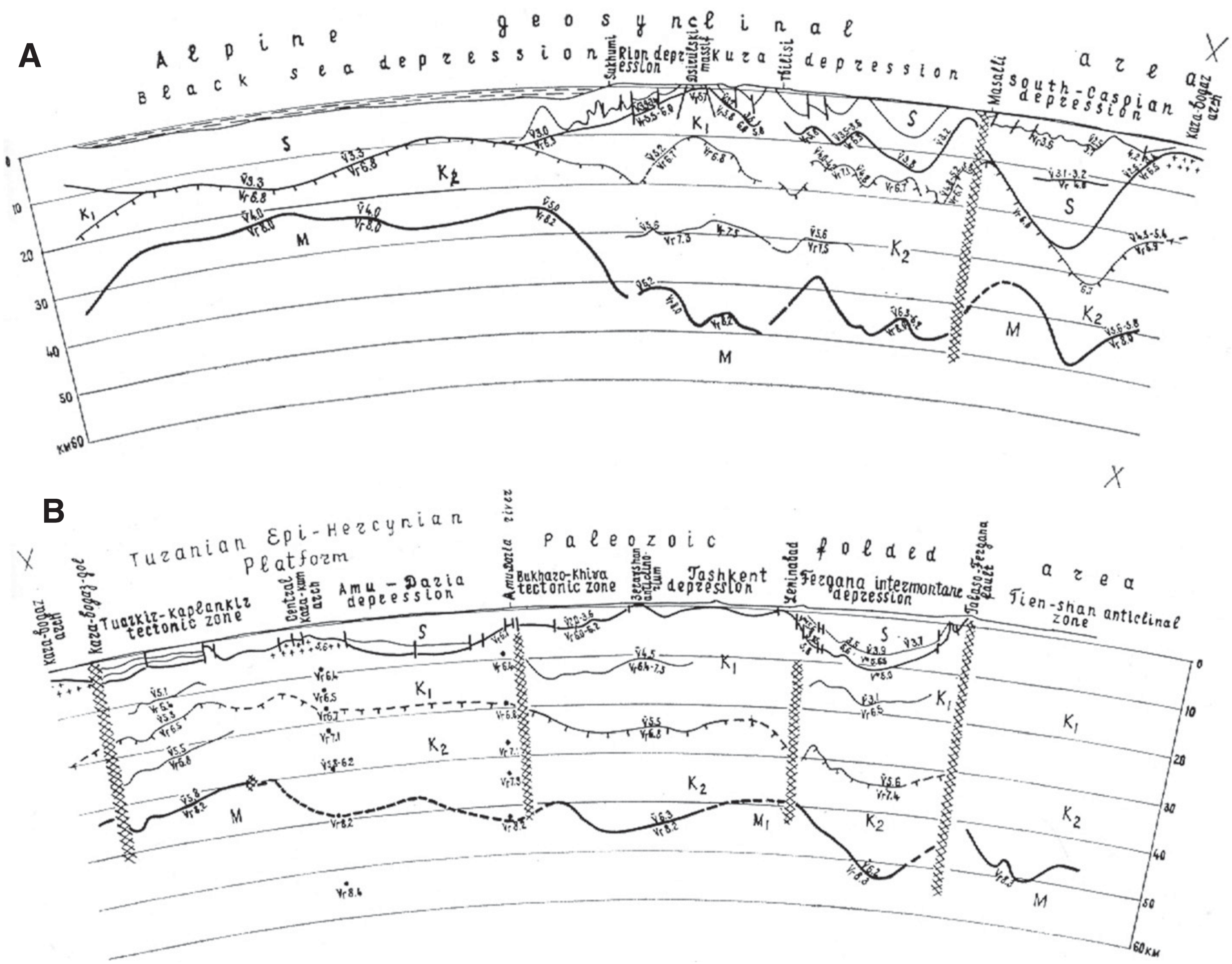


Figure 6.4-07. Section of the Earth's crust from the Black Sea through the Caucasus and the Caspian Sea to the central Kara-Kun-Tien Shan (from Kosminskaya et al., 1969, fig. 8). S—sedimentary layers, K—crustal layers, M—Moho. Velocities in km/s. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 195–208. Reproduced by permission of American Geophysical Union.]

generalized crustal models for different tectonic areas throughout the USSR, and Figure 6.4-09 shows a crustal thickness contour map for the USSR and adjacent areas. Other transcontinental crustal sections throughout the USSR and adjacent areas were also summarized and published, e.g., by Belyaevsky et al. (1968).

6.5. SEISMIC-REFRACTION INVESTIGATIONS IN NORTH AMERICA IN THE 1960s

6.5.1. Western United States

In 1960, under the VELA UNIFORM program of the Advanced Research Projects Agency of the U.S. Department of Defense, the U.S. Geological Survey at Denver, Colorado, received major funding to start a large program of crustal studies, headed by L.C. Pakiser.

At first, staff had to be organized, largely by reassignment of available personnel, which involved scientists such as W.H. Diment, J.P. Eaton, J.H. Healy, D.P. Hill, W.H. Jackson, J.C. Roller, A. Ryall, S.W. Stewart, D.J. Stuart, R.W. Warrick, and others. In addition, the gravity studies of major crustal features in the western United States were continued and the organization of a new and more extensive program "Study of Seismic Propagation Paths and Regional Traveltimes in the California-Nevada Region" was planned.

Later on, in 1965–1966, the same group of the U.S. Geological Survey, again under the leadership of L.C. Pakiser, moved from Denver, Colorado, to Menlo Park, California, where the

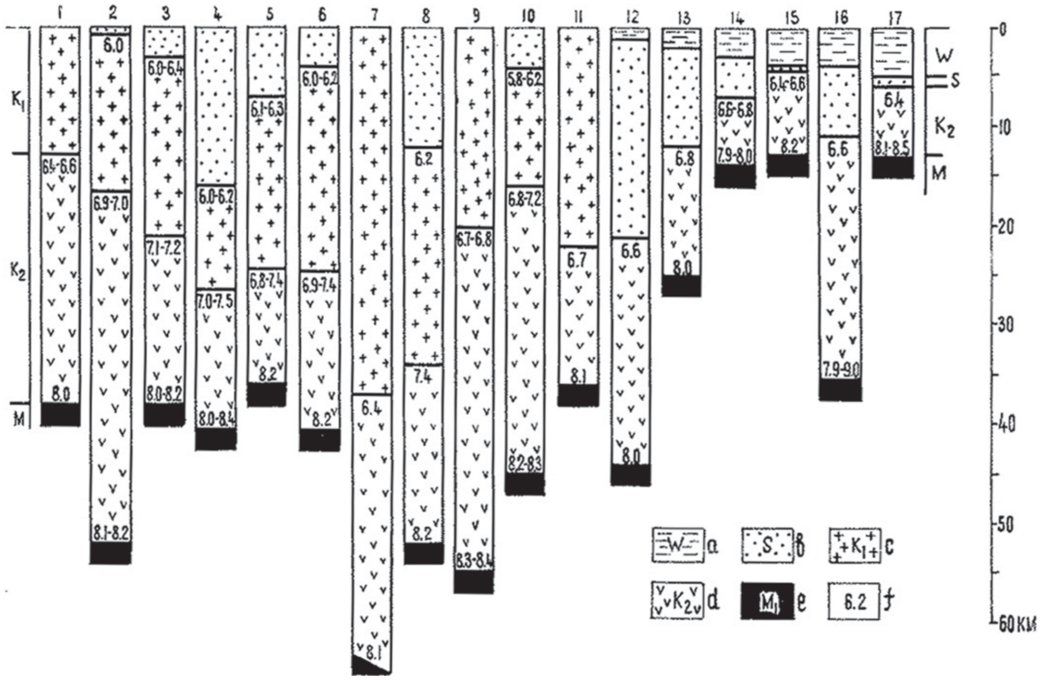


Figure 6.4-08. Generalized crustal models for different tectonic areas in the USSR (from Kosminskaya et al., 1969, fig. 9). (a) water; (b) unconsolidated crust (sediments); (c) consolidated upper crust; (d) consolidated lower crust; (e) upper mantle; (f) boundary velocities in km/s. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 195–208. Reproduced by permission of American Geophysical Union.]

Figure 6.4-09. Crustal thickness map for the territory of the USSR and adjacent areas (from Kosminskaya et al., 1969, fig. 11). [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 195–208. Reproduced by permission of American Geophysical Union.]

National Center of Earthquake Research was established to start a major earthquake research program.

Following detailed discussions with scientists and technicians of various laboratories in the United States to design the detailed performance specifications, in 1960 a new seismic-refraction system was developed and constructed by Dresser Electronics, SIE Division (Figs. 6.5.1-01 to 6.5.1-03).

The first seismic system was delivered, tested, and accepted in early 1961, and by the end of 1961, ten units had been placed in operation, each installed in a four-wheel-drive vehicle. In addition, mounted in two aluminum trailers and in a four-wheel-drive master vehicle for use at the shotpoints, communication and timing equipment was developed.

Compatible communications and timing systems were installed in each recording unit. The units (Figs. 6.5.1-02 and 6.5.1-03), recording simultaneously at two levels of amplification on magnetic tape (30 dB separation) and by an oscillograph on photographic paper (15 dB separation), were equipped with 8

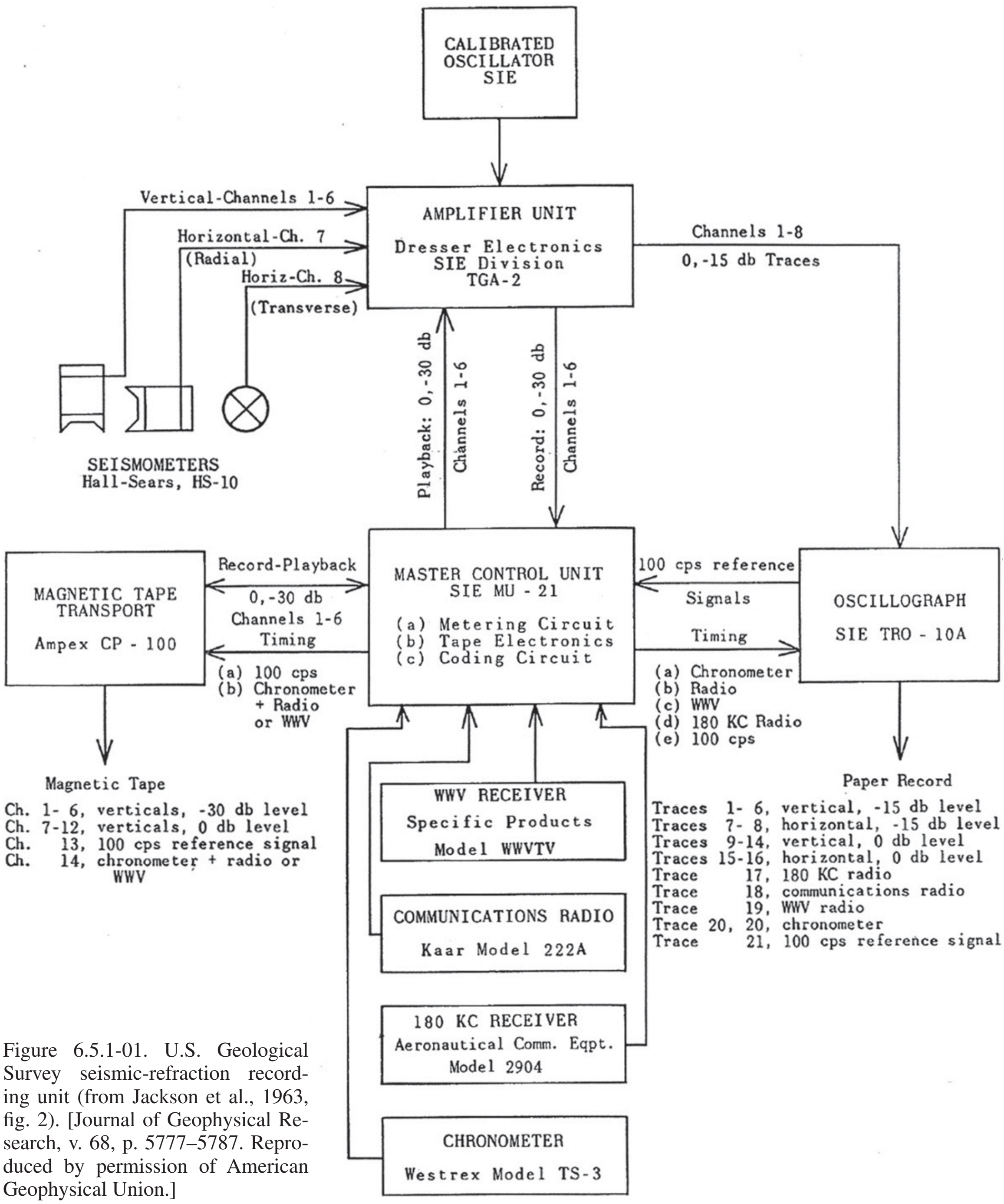


Figure 6.5.1-01. U.S. Geological Survey seismic-refraction recording unit (from Jackson et al., 1963, fig. 2). [Journal of Geophysical Research, v. 68, p. 5777–5787. Reproduced by permission of American Geophysical Union.]

A

B

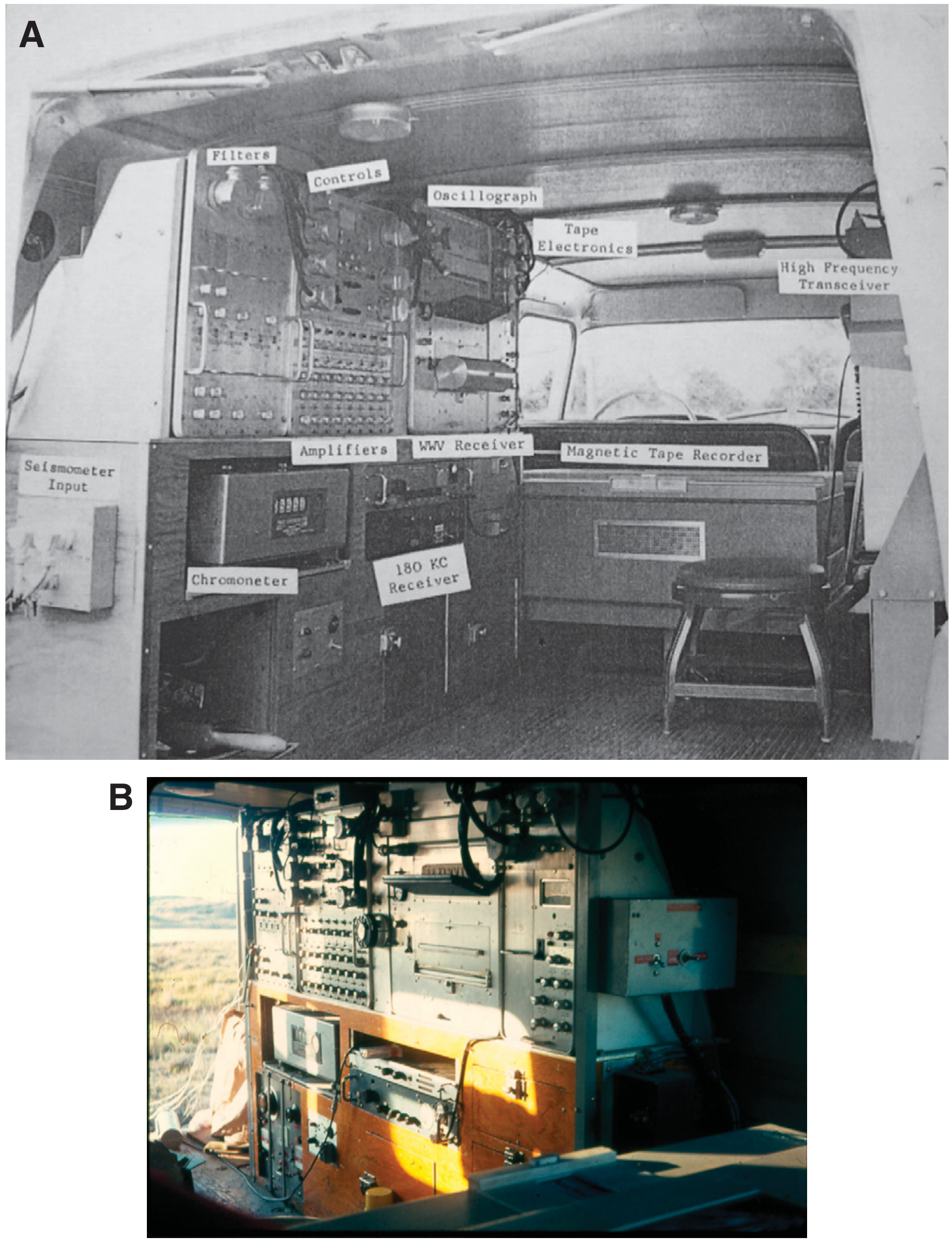


Figure 6.5.1-02. U.S. Geological Survey seismic-refraction equipment installed in a four-wheel drive recording truck. (A: from Warrick et al., 1961, fig. 2). [Geophysics, v. 26, p. 820–824. Permission granted by Society of Exploration Geophysicists, Tulsa, Oklahoma, USA.] (B: Photograph by C. Prodehl.)

seismic and two timing channels, which allowed a 2.5-km-wide array of 6 vertical seismometers (with natural frequencies of 1–2 Hz) with average spacing of 500 m where terrain permitted, while two horizontal seismometers were usually placed in the center. Additional traces were available for recording the time signals of the broadcast station WWV, a calibrated chronometer, and when possible the shot instant. The details of this new seismic system were discussed by Warrick et al. (1961).

The first seismic measurements were made from April to August 1961 along two unreversed profiles in eastern Colorado. Fieldwork was started with one recording unit, and six units were in operation at the end of the Colorado field program. One profile extended north from Nee Granda Reservoir, between Eads and Lamar, Colorado, to the vicinity of Sterling, Colorado. The profile was ~325 km long. Seismic recordings were made at 113 spread locations, and a total of 64 shots, ranging in size from a few to 1500 kg, were exploded in Nee Granda Reservoir and in drill holes along the shore of the reservoir (for location of Lamar, see Fig. 6.5.1-12).

A data example recorded at 227.5 km distance with P_n as first arrivals (P_7) is shown in Figure 6.5.1-04, showing the monitor seismogram (upper two sets) and magnetic-tape playback with filtering (lower two sets) with low (top) and high (bottom) gain traces in each set. Figure 6.5.1-05 shows parts of the result-

Figure 6.5.1-03. U.S. Geological Survey seismic-refraction recording truck, with Sam Stewart laying out an array of 6 seismometers (photograph by C. Prodehl).

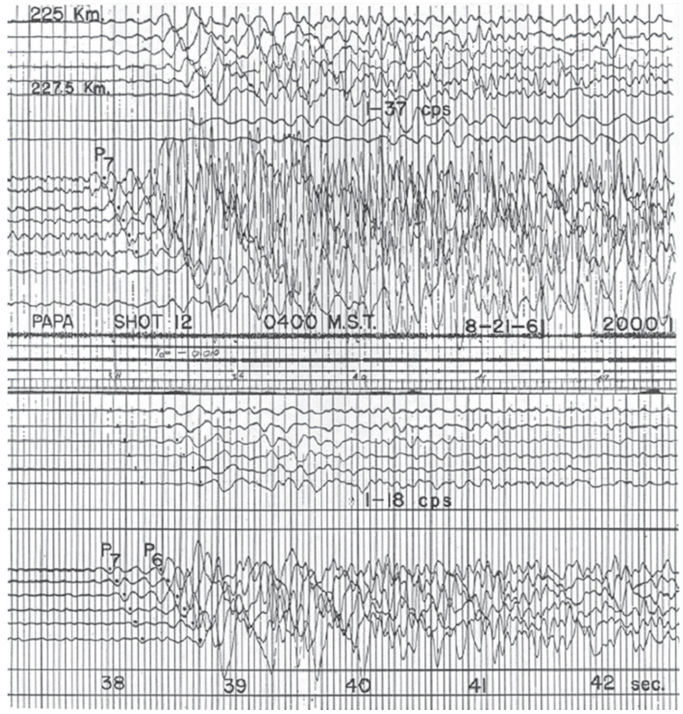


Figure 6.5.1-04. Monitor seismogram (upper two sets of traces) and magnetic-tape playback with filtering (lower two sets) recorded 227.5 km north of Nee Granda Reservoir (Lamar). The upper traces in each set are recorded with low gain, the lower traces with high gain. The 6 upper traces in each set are recordings from vertical seismometers, traces 7 and 8 on monitor only are in-line and transverse horizontal motion (from Jackson et al., 1963, fig. 6). [Journal of Geophysical Research, v. 68, p. 5777–5787. Reproduced by permission of American Geophysical Union.]

ing record section with traveltime curves correlated by Jackson et al. (1963). As seen in Figures 6.5.1-04 and 6.5.1-05, the data achieved by the new equipment at its first field experiment show an extremely good quality.

The second profile in Colorado extended west from Nee Granda Reservoir to the vicinity of Salida, Colorado, with a total length of 290 km. Along this profile, 13 shots in Nee Granda Reservoir were recorded at 54 spreads.

The data of both profiles were later digitized by Dave Warren (unpublished U.S. Geological Survey Open-File Report, data reproduced in Appendix A6-8, together with the 1965 Rocky Mountain profile data; see below).

The seismic-refraction recording of 1961 resulted in 45–50 km crustal thickness under eastern Colorado, as determined from interpretation of the profile from Nee Granda Reservoir to Sterling. However, the fieldwork served primarily to test thoroughly the new seismic-refraction recording, communications, and timing equipment and to train observers from the U.S. Geological Survey and United ElectroDynamics, Inc., in operation of the new seismic system. Furthermore, efficient field procedures had to be developed for the later major program of fieldwork in the California-Nevada region. In the remainder of 1961 and in the following two years, a network of 64 seismic-refraction profiles was recorded by the U.S. Geological Survey in Nevada and California and adjacent areas in Idaho, Wyoming, Utah, and Arizona (Fig. 6.5.1-06). The investigation concentrated mainly on the Great Basin of the Basin and Range Province and the Sierra Nevada. In addition, two recording lines extended into the western Snake River Plain of Idaho and the southern Cascade Range in California. Other profiles were recorded in the Coast Ranges of California, in the Colorado Plateau, and in the middle Rocky Mountains.

Nuclear test sites were also included in the program. From the first nuclear test site in southeastern New Mexico, GNOME, a line had been recorded in the Great Plains northwards, parallel to and east of the front of the southern Rocky Mountains (Stewart and Pakiser, 1962). From SHOAL, Nevada (no. 10 in Fig. 6.5.1-06), a line was recorded eastward through the Basin and Range province and reversed later by chemical shots near Delta, Utah. Finally, the repeated tests at the Nevada Test Site (no. 19 in Fig. 6.5.1-06) allowed the recording of long-range profiles of several 1000 km in length through the western United States in all directions.

Except for the nuclear tests, seismic energy was generated by chemical explosions fired in the Pacific Ocean, in lakes, or in drill holes. Ocean shots ranged from 2000 to 6000 lb and were placed on the sea bottom. The charges in lakes ranged from 2000 to 10,000 lb and were placed on the lake bottoms. At all the other shotpoints, the charges were fired in drill holes and ranged in size from 250 to 20,000 lb. Usually 10 recording units were available, and the average spacing of the recording units was 10 km, so that most of the lines required more than one layout. Only in a few instances were the units placed closer together.

The complete data set was later published in a U.S. Geological Survey Professional Paper in the form of record sections by Prodehl (1979). Data examples are shown for the Sierra Nevada (Fig. 6.5.1-07A), the Basin and Range province and western

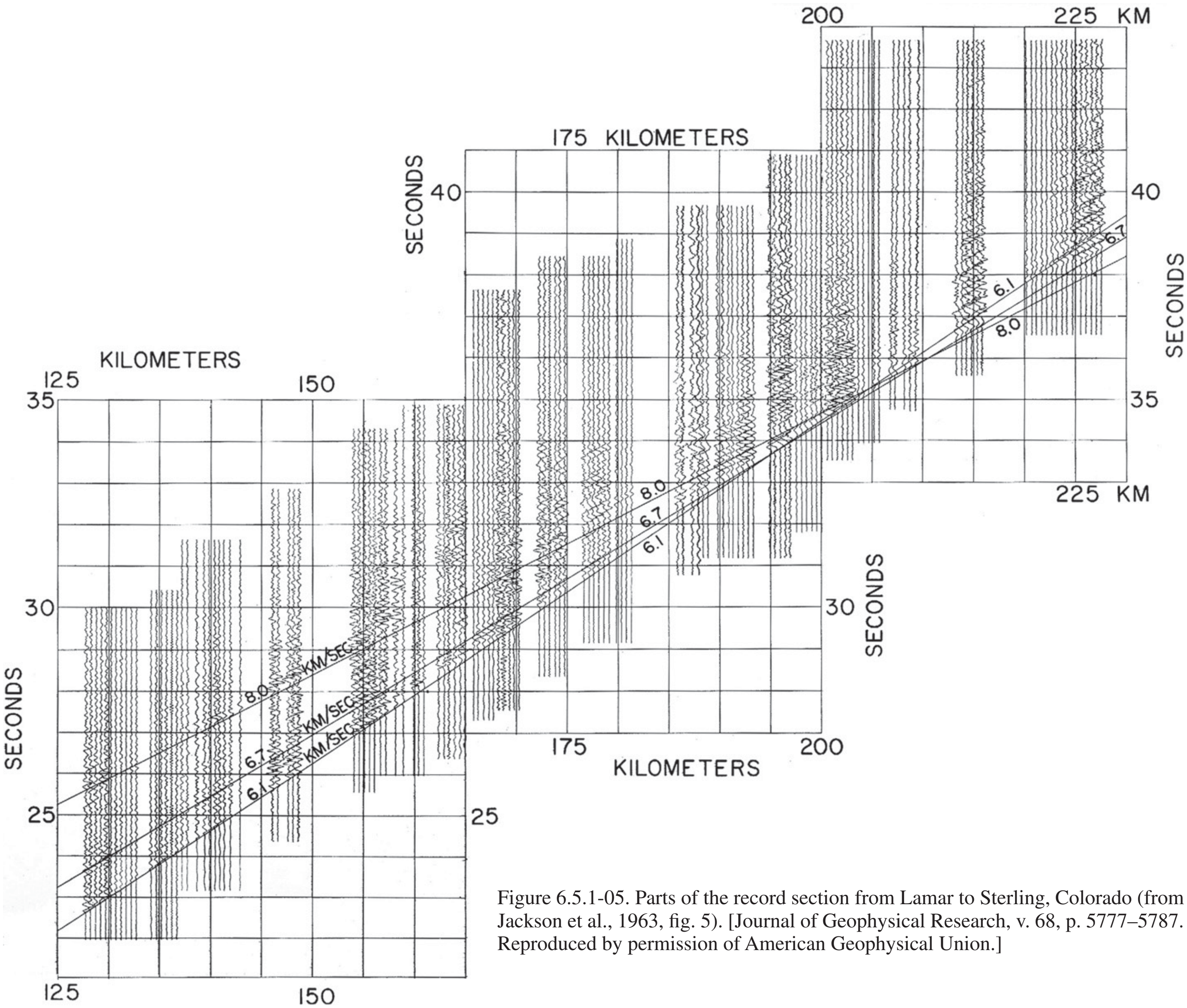


Figure 6.5.1-05. Parts of the record section from Lamar to Sterling, Colorado (from Jackson et al., 1963, fig. 5). [Journal of Geophysical Research, v. 68, p. 5777–5787. Reproduced by permission of American Geophysical Union.]

Snake River Plain (Fig. 6.5.1-07B), the middle Rocky Mountains and the Colorado Plateau (Fig. 6.5.1-07C). All record sections are reproduced in Appendix A6-6 and the corresponding basic data of shotpoint and recording site locations can be found in Appendix A6-7.

First interpretations of parts of this extensive fieldwork of 1961 and 1962, including the earlier line from the Nevada Test Site to Kingman, Arizona (Diment et al., 1961), were presented at a symposium of the American Geophysical Union and subsequently published in a joint volume in 1963 (Journal of Geophysical Research, vol. 68, no. 20, p. 5747–5849), introduced by an overview of Pakiser (1963). Over most of the Basin and Range province, including the Sierra Nevada, P_n velocities of 7.8–7.9 km/s had been obtained (Eaton, 1963; Pakiser and Hill, 1963; Roller and Healy, 1963; Ryall and Stuart, 1963). Along the California coastline (Healy, 1963), in the Colorado Plateau (Ryall and Stuart, 1963), and in the Great Plains province (Stewart and Pakiser, 1962), P_n velocities ranged between 8.0 and 8.2 km/s.

Two preliminary crustal cross sections in W-E and S-N directions with many question marks had been compiled for this symposium (Fig. 6.5.1-08). These sections showed a fairly thin crust but a questionable existence of an intracrustal boundary for the whole Basin and Range province. Toward the north, a thick intermediate layer underlying a very thin uppermost crust resulted for the Snake River Plain. Toward the west, a crustal root was indicated under the Sierra Nevada, while a thin crust again resulted for most of California.

In the frame of a series of maps published along a transcontinental survey covering the latitudes 35°–39°N (Hart, 1964; Pakiser and Zietz, 1965), Warren (1968a) compiled the most

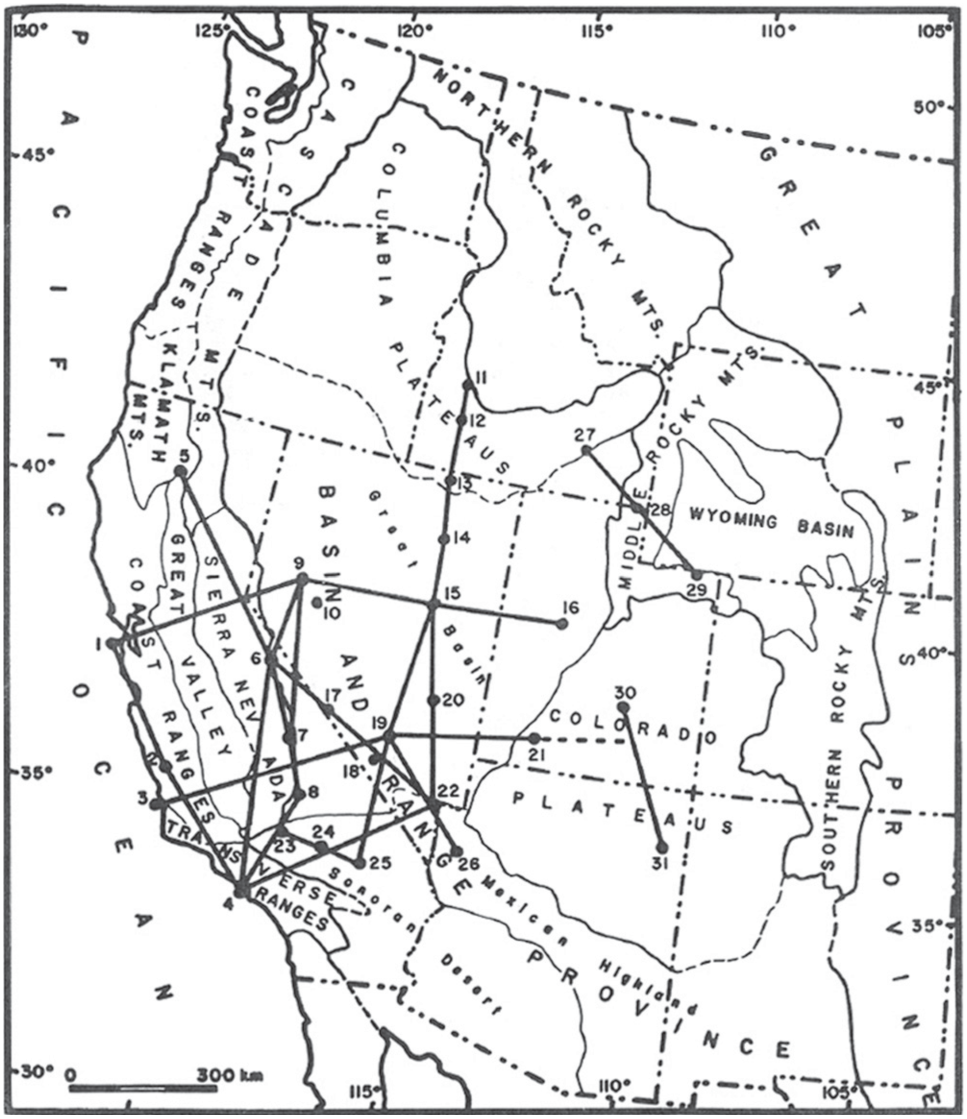


Figure 6.5.1-06. Location of seismic profiles recorded in 1961 to 1963 by the U.S. Geological Survey in the western United States (from Prodehl, 1970, fig. 1). MD—Mojave Desert, OV—Owens Valley, DV—Death Valley, LP—Lassen Peak National Park. Shotpoints: 1—San Francisco, 2—Camp Roberts, 3—San Luis Obispo, 4—Santa Monica Bay, 5—Shasta Lake, 6—Mono Lake, 7—Independence, 8—China Lake, 9—Fallon, 10—SHOAL, 11—Boise, 12—Strike Reservoir, 13—Mountain City, 14—Elko, 15—Eureka, 16—Delta, 17—Lida Junction, 18—Lathrop Wells, 19—NTS (Nevada Test Site), 20—Hiko, 21—Navajo Lake, 22—Lake Mead, 23—Mojave, 24—Barstow, 25—Ludlow, 26—Kingman, 27—American Falls Reservoir, 28—Bear Lake, 29—Flaming Gorge Reservoir, 30—Hanksville, 31—Chinle. [Geological Society of America Bulletin, v. 81, p. 2629–2646. Reproduced by permission of the Geological Society of America.]

prominent seismic-refraction lines within these latitudes and constructed preliminary cross sections, which he published in fence diagrams (see also Healy and Warren, 1969, fig. 1).

For a unified interpretation of the complete set of data recorded by the U.S. Geological Survey in 1961–1963 in the western United States, Prodehl (1970, 1979) used the same methodology as had been proposed and applied by Giese (1968a, 1976b) for the interpretation of many data in central Europe and the Alps.

Two crustal cross sections along the same lines as those shown by Pakiser (1963; Fig. 6.5.1-08) are reproduced in Figure 6.5.1-09 and the corresponding record sections for the N-S line Boise to Lake Mead are shown in Figure 6.5.1-07B. Figure 6.5.1-10 shows a fence diagram for the crustal structure under California and Nevada and adjacent areas obtained by Prodehl (1970, 1979) from the reinterpretation of the whole set of data.

In 1964, the U.S. Geological Survey fieldwork was extended to a network of profiles in Arizona across the Tonto Forest

A

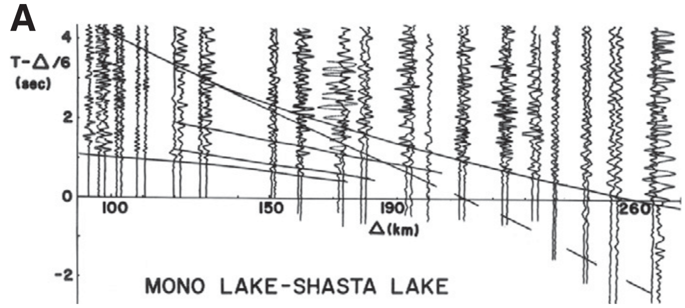
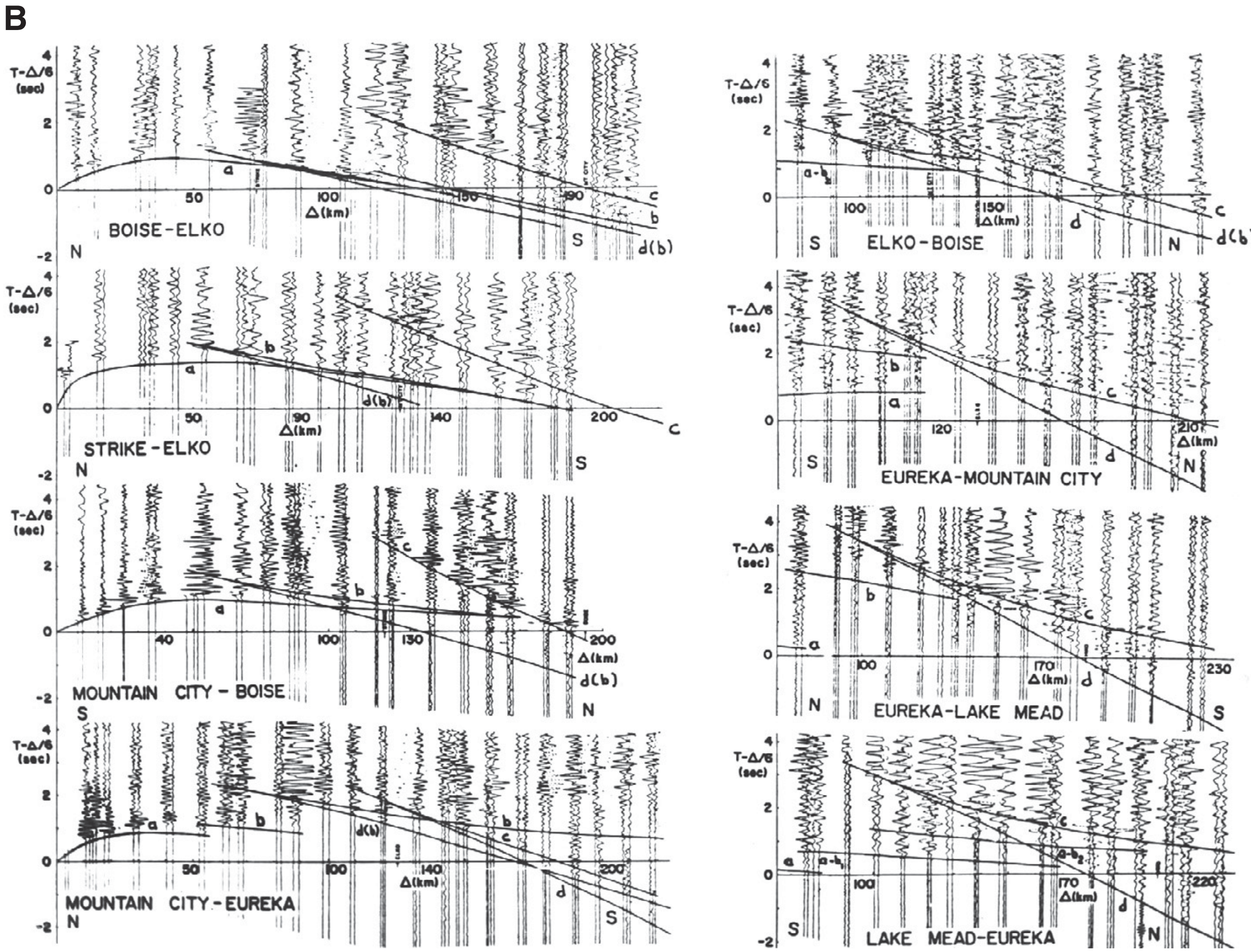


Figure 6.5.1-07. (A) Record section for the line Mono Lake to Shasta Lake, Sierra Nevada (from Prodehl, 1979, fig. 47). [U.S. Geological Survey Professional Paper 1034, 74 p. Copyright U.S. Geological Survey.] (*Continued on following page.*)



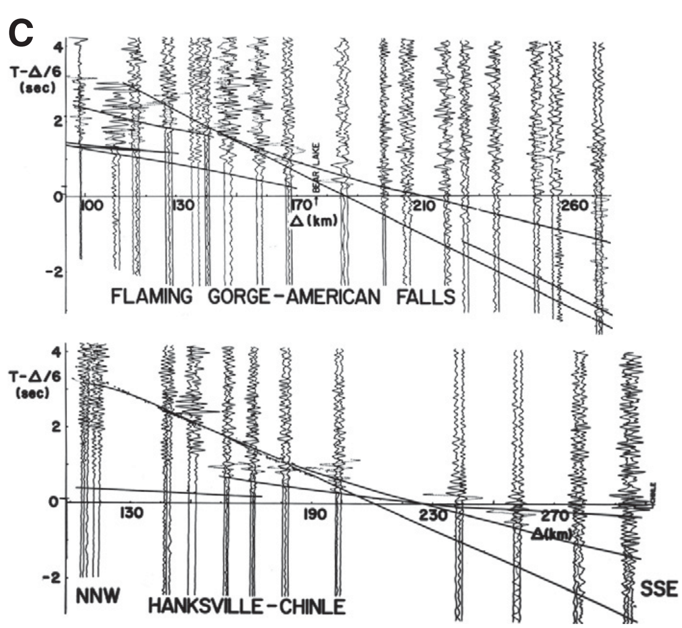


Figure 6.5.1-07, *continued*. (B) Record sections for the line Boise, Idaho, Columbia Plateaus, to Lake Mead, Nevada, Basin and Range province (from Prodehl, 1970, figs. 3 and 5). [Geological Society of America Bulletin, v. 81, p. 2629–2646. Reproduced by permission of the Geological Society of America.] (C) Record sections for the lines Flaming Gorge, Utah, to American Falls, Idaho, Middle Rocky Mountains and Hanksville, Utah, to Chinle, Arizona, Colorado Plateau (from Prodehl, 1979, figs. 78 and 74). [U.S. Geological Survey Professional Paper 1034, 74 p. Copyright U.S. Geological Survey.]

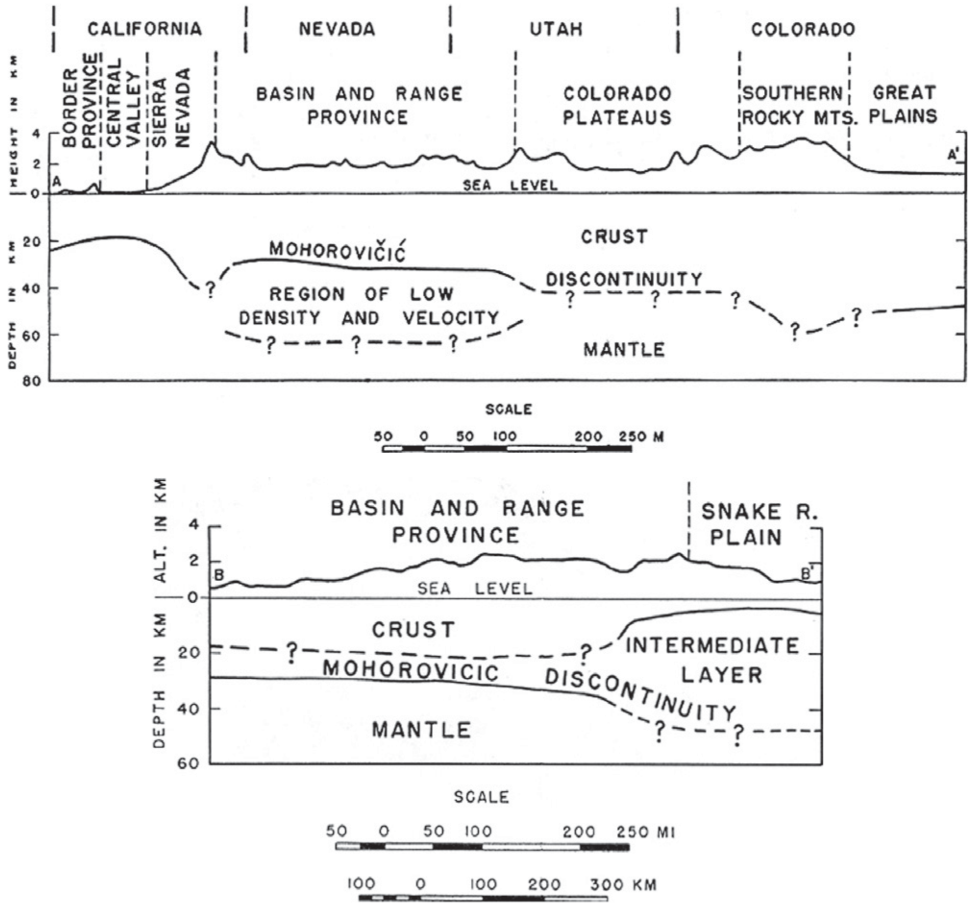


Figure 6.5.1-08. Variations in crustal thickness along two cross sections through the Basin and Range province (from Pakiser, 1963, figs. 3 and 4). Top: W-E section from San Francisco, California (A), to Lamar, Colorado (A′). Bottom: S-N section, including an intermediate crustal layer, from Ludlow, California (B), to Boise, Idaho (B′). [Journal of Geophysical Research, v. 68, p. 5747–5756. Reproduced by permission of American Geophysical Union.]

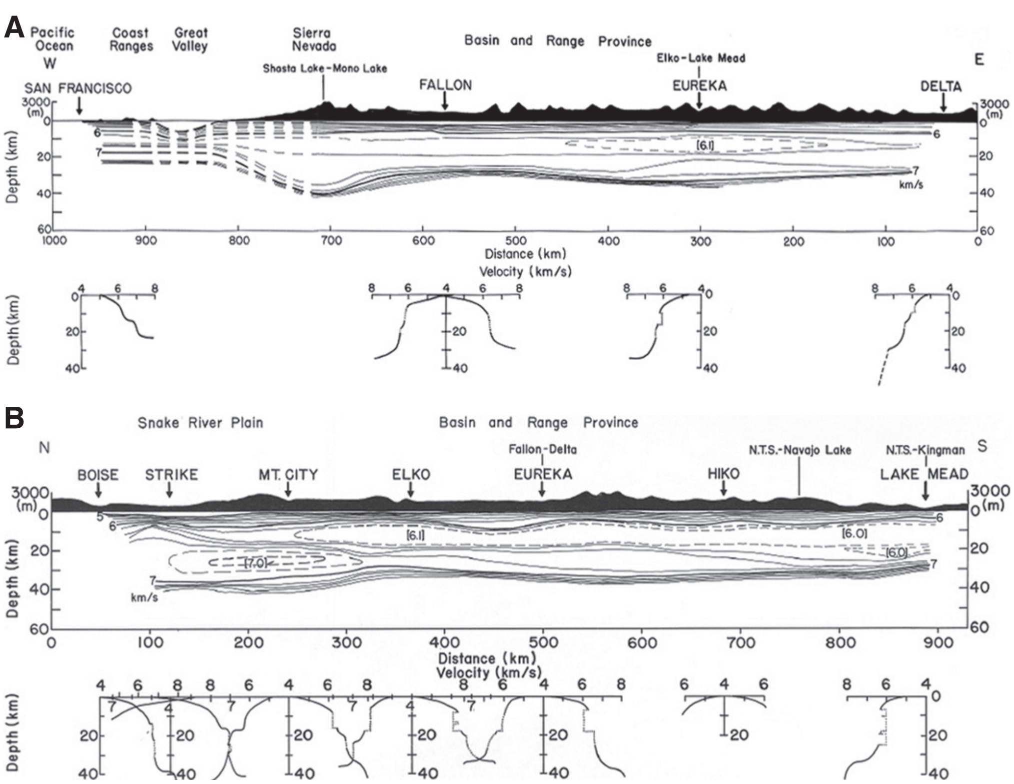


Figure 6.5.1-09. Crustal cross sections through the Basin and Range province and adjacent areas (from Prodehl, 1979, figs. 12 and 25). (A) W-E section from San Francisco, California, to Delta, Utah. (B) N-S section from Boise, Idaho, to Lake Mead, Nevada. [U.S. Geological Survey Professional Paper 1034, 74 p. Copyright U.S. Geological Survey.]

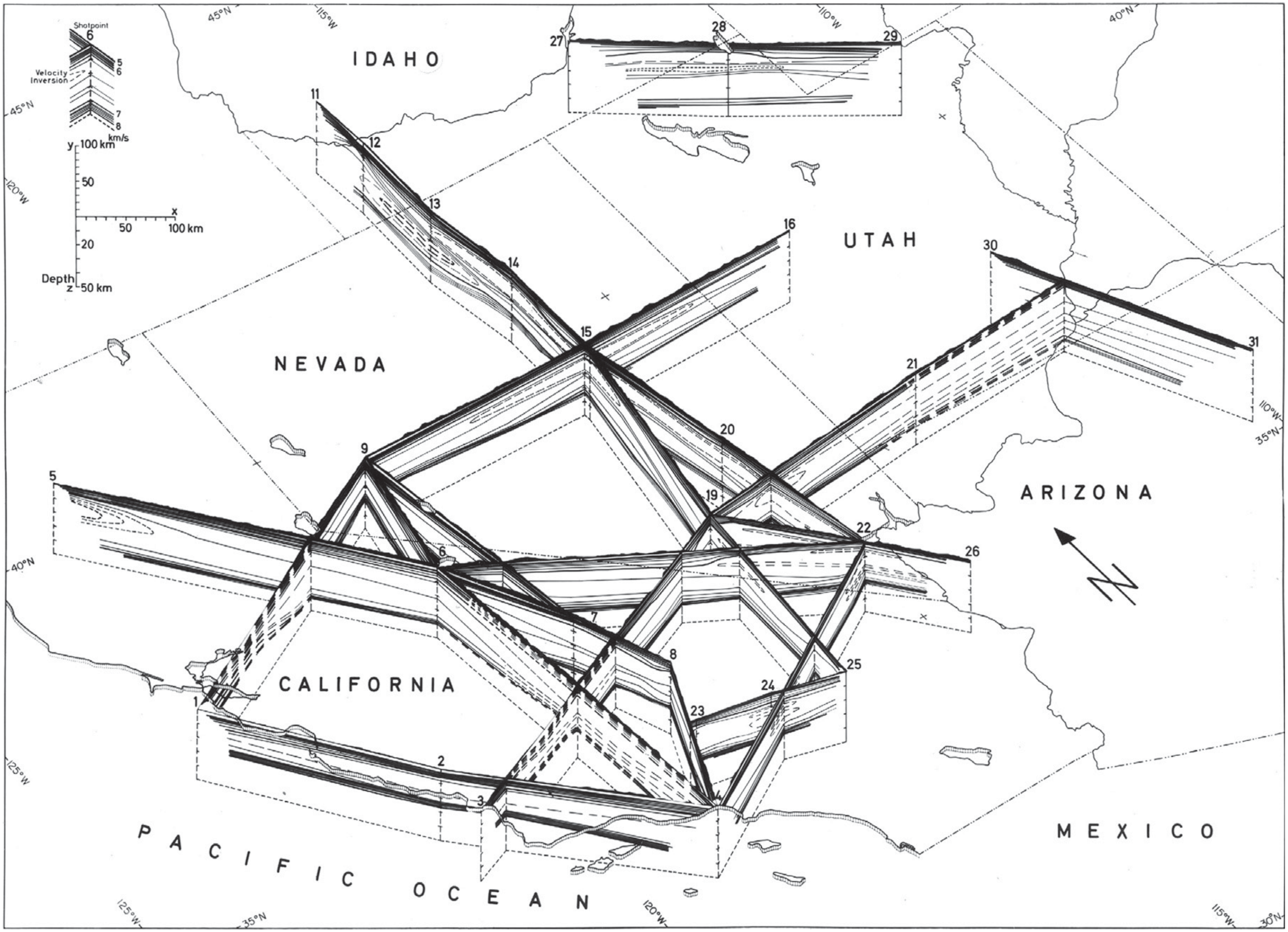


Figure 6.5.1-10. Fence diagram, showing the crustal structure under California and Nevada and adjacent areas obtained from the reinterpretation of the whole set of seismic-refraction data recorded by the U.S. Geological Survey in 1961–1963 (from Prodehl, 1970, fig. 10). [Geological Society of America Bulletin, v. 81, p. 2629–2646. Reproduced by permission of the Geological Society of America.]

Seismological Observatory, located 10 miles south of the Mogollon Rim near Payson in central Arizona (Warren, 1969). Two recording lines 400 km long intersected at the observatory (Fig. 6.5.1-11) and recorded chemical explosions from a total of 14 shotpoints. One line trended approximately parallel to the Mogollon Rim in SE-NW direction; the other one followed a SW-NE direction. The Moho contour map of Warren (1969) shows a rapid increase in depth from less than 25 km under the Basin and Range province in southwestern Arizona to ~40 km under the Colorado Plateau of northern Arizona (Fig. 6.5.1-11).

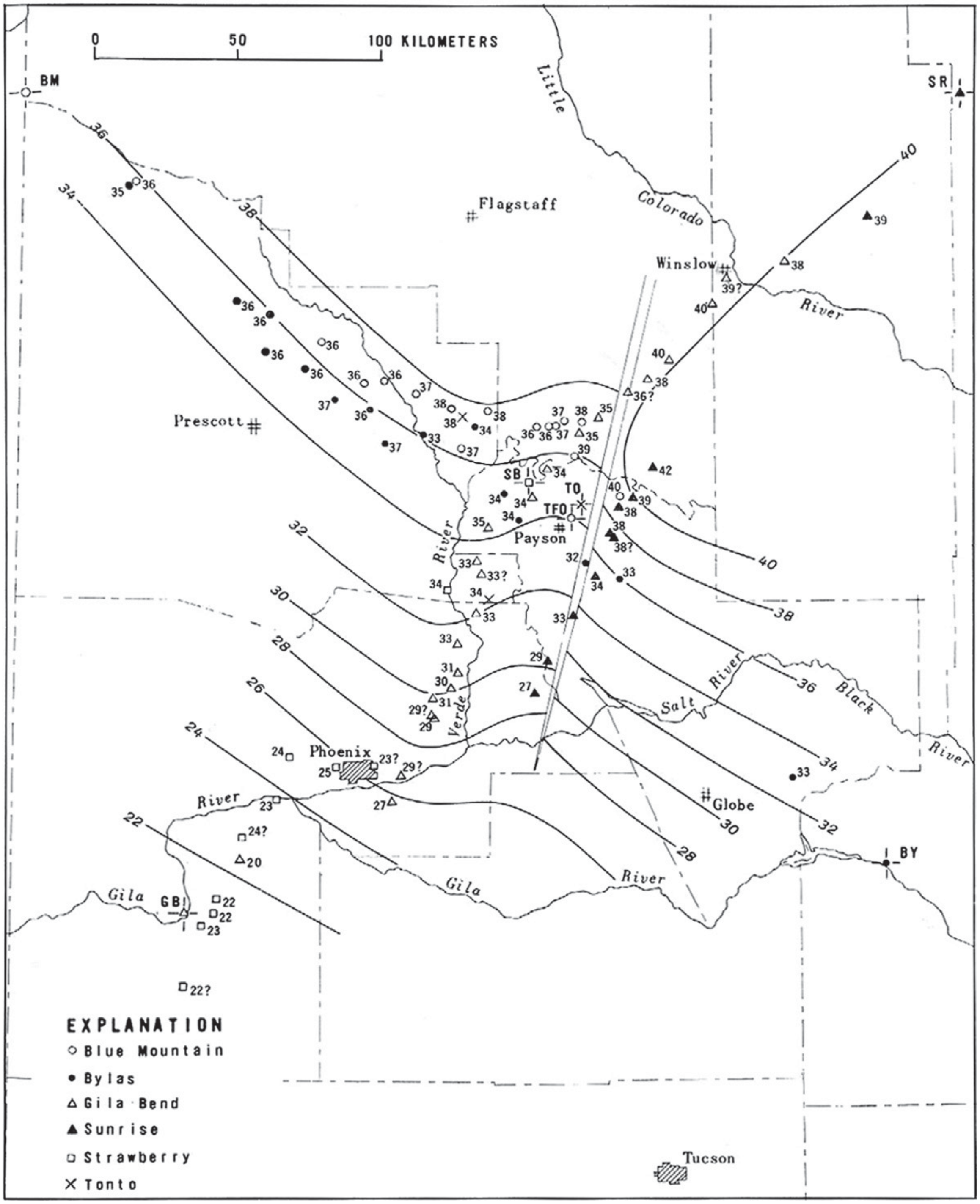


Figure 6.5.1-11. Moho configuration under Arizona (from Warren, 1969, fig. 13). TFO—Tonto Forest Seismological Observatory; stars named by two bold letters are shotpoint locations. [Geological Society of America Bulletin, v. 80, p. 257–282. Reproduced by permission of the Geological Society of America.]

In 1965, a network of profiles in the Southern Rocky Mountains and adjacent areas in the Great Plains of Colorado and Kansas was added. Some of the lines are contained in the compilation of Warren (1968b) along the transcontinental survey covering the latitudes 35°–39°N, showing preliminary crustal cross sections (see also Healy and Warren, 1969, fig. 2). Only one of the lines in the Rocky Mountains was interpreted in more detail at the time (Jackson and Pakiser, 1965). The remainder of the data was reinterpreted later (see, e.g., Prodehl, 1977; Prodehl and Pakiser, 1980; Prodehl and Lipman, 1989). Most of the data, recorded in the region of the Southern Rocky Mountains and adjacent Great Plains in the 1960s, were compiled as record sections and published in various publications (Prodehl and Pakiser, 1980; Prodehl and Lipman, 1989; Prodehl et al., 2005; Steeples and Miller, 1989) and, when the appropriate techniques had become available, were subsequently digitized by David Warren and compiled in internal U.S. Geological Survey Open-File Reports in the seventies. These reports and the corresponding record sections are reproduced in Appendix A6-8.

A more detailed crustal and upper mantle seismic investigation of the area followed only in the late 1990s (see Chapter 9; Karlstrom and Keller, 2005). The map in Figure 6.5.1-12 shows all profiles recorded in the 1960s but also includes more recent seismic-refraction projects Deep Probe of 1995 and CD-ROM (Continental Dynamics of the Rocky Mountain experiment) of 1999 (Prodehl et al., 2005).

The Moho contour map for Wyoming and Colorado (Fig. 6.5.1-13), which Keller et al. (1998) compiled before the CD-ROM data became available, is mainly based on data recorded in the 1950s and 1960s. Only the area of the Rio Grande rift in New Mexico is based on data recorded in the 1970s and 1980s (see Chapters 7 and 8).

In eastern Montana, a crustal study project of LASA (Large Aperture Seismic Array) (Fig. 6.5.1-14), located in the Great Plains between the earlier shotpoints Fort Peck (M4 in Fig. 5.2-07) and Acme Pond (M5 in Fig. 5.2-07), followed in 1966. The crustal structure study of Montana of 1956–1960 by various groups (see Chapter 5, Fig. 5.2-07) had indicated major changes

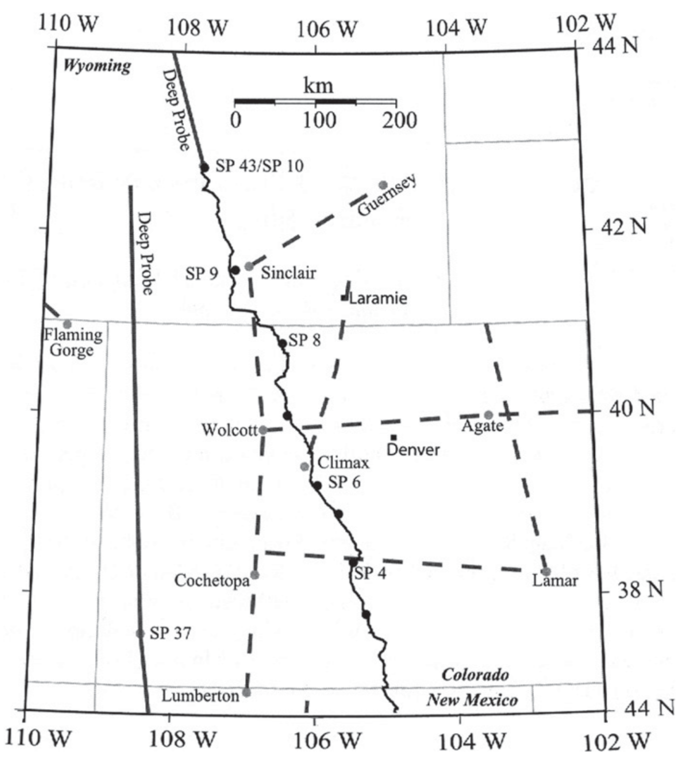


Figure 6.5.1-12. Seismic refraction lines, recorded in the Southern Rocky Mountains and adjacent Great Plains (from Prodehl et al, 2005, part of fig. 1). Dashed lines and grey circles—profiles recorded in the sixties; full crooked line—project CD-ROM of 1999; full straight lines—project Deep Probe of 1995. [The Rocky Mountain Region: An evolving lithosphere—tectonics, geochemistry, and geophysics: American Geophysical Union, Geophysical Monograph 154, p. 201–216. Reproduced by permission of American Geophysical Union.]

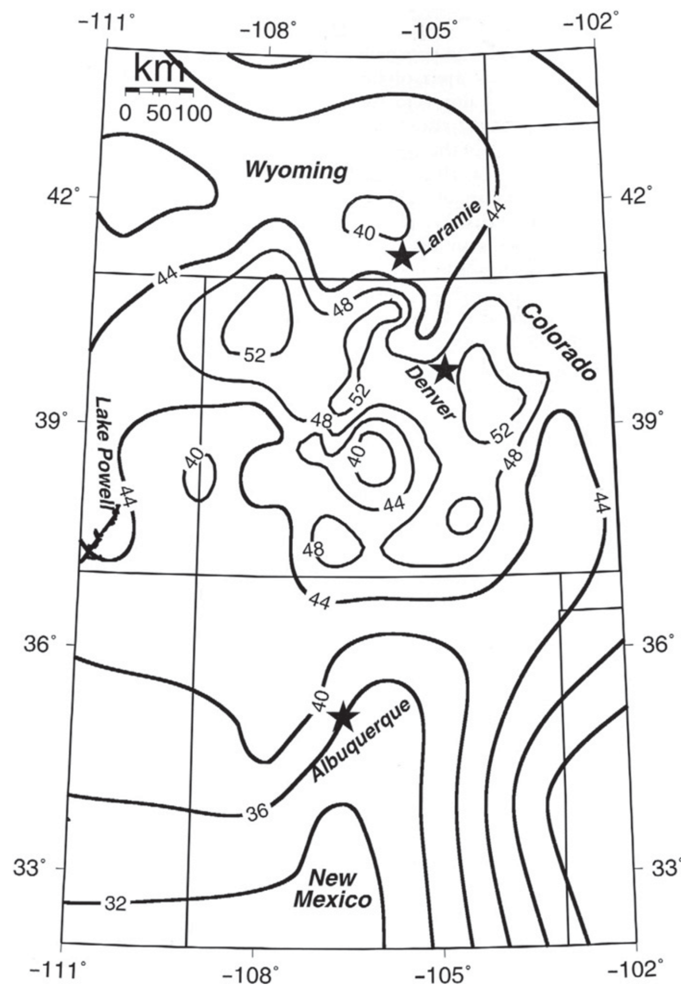


Figure 6.5.1-13. Moho contour map of the Southern Rocky Mountains Region and adjacent areas (from Keller et al, 1998, fig. 2). [Rocky Mountain Geology, v. 33, p. 217–228. Published by permission of Rocky Mountain Geology, Laramie, Wyoming, USA.]

in crustal structure throughout Montana, but had also left marked differences in the internal structure of the crust (McCamy and Meyer, 1964).

Therefore, in 1966, a crustal calibration seismic-refraction line with five shotpoints was recorded by the U.S. Geological Survey in a NE-SW direction across the center of the LASA array (Warren et al., 1972, 1973; Fig. 6.5.1-14), the analysis of which confirmed that marked changes of crustal structure existed within the array, calling for a more detailed study of the crust under LASA.

This was realized in 1968 (Warren et al., 1972, 1973). The USGS recording trucks (Figs. 6.5.1-01 to 6.5.1-03) were used to establish short arrays along profile lines; in addition, 20 portable 1- or 3-component seismic stations and the standard LASA array stations operated continuously for about two months recording both a number of teleseisms as well as the arrivals from the high-explosive shots fired in the calibration program.

Fifteen shots of either 4000 or 20,000 pounds were detonated and recorded by the expanded array of seismic stations (Fig. 6.5.1-15), resulting in a total of 325 seismic traces of the expanded LASA clusters plus 33 seismograms of the recording trucks and temporary seismic stations for each shot. The data were compiled as record sections and published as a U.S. Geological Survey Open-File report (Warren et al., 1972; reproduced in Appendix A6-9). The crustal thickness map resulting from a first analysis of the complex data set (Warren et al., 1972, 1973) shows depth variations from 45 to 55 km over relatively short horizontal distances (Fig. 6.5.1-16).

Finally, in 1967, two sparsely sampled seismic-refraction lines were recorded in the Coast Ranges south of the San Francisco Bay Area on both sides of the San Andreas fault system (Stewart, 1968a), with five shotpoints along each line (Fig. 6.5.1-17). The line in the Diablo Range to the northeast of the San Andreas fault was 200 km long, but only the central part between the shotpoints Cedar Mountain and Mount Stakes of 70 km length was well covered with recording stations. The Gabilan Range line in the Salinian block to the southwest of the San Andreas fault was 235 km long, but only the central

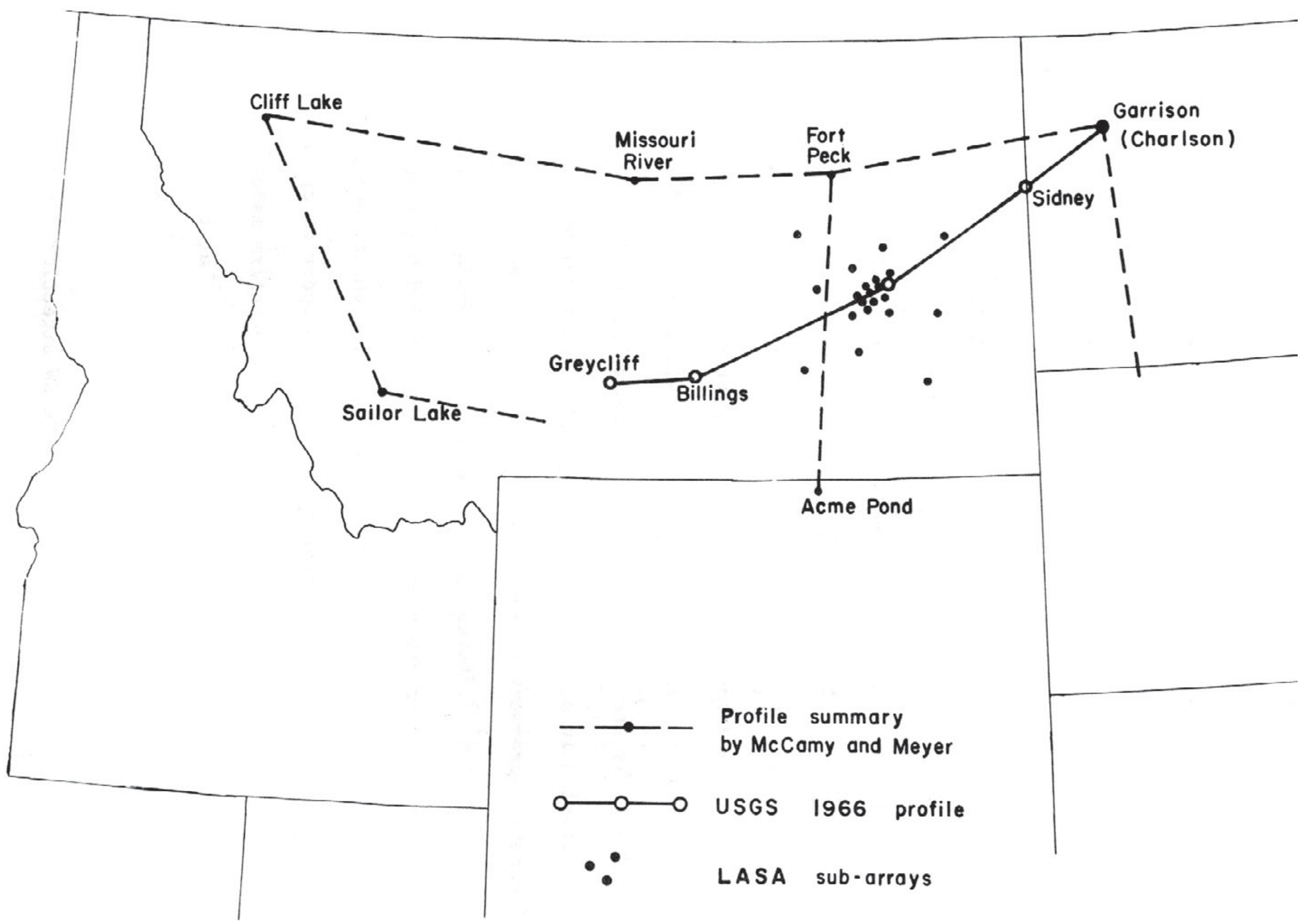


Figure 6.5.1-14. Seismic refraction, crustal calibration line, recorded across the LASA array in Montana in 1966 (from Warren et al, 1972, fig. 1). [U.S. Geological Survey Open-File Report, p. 1–58 and A1–A105.]

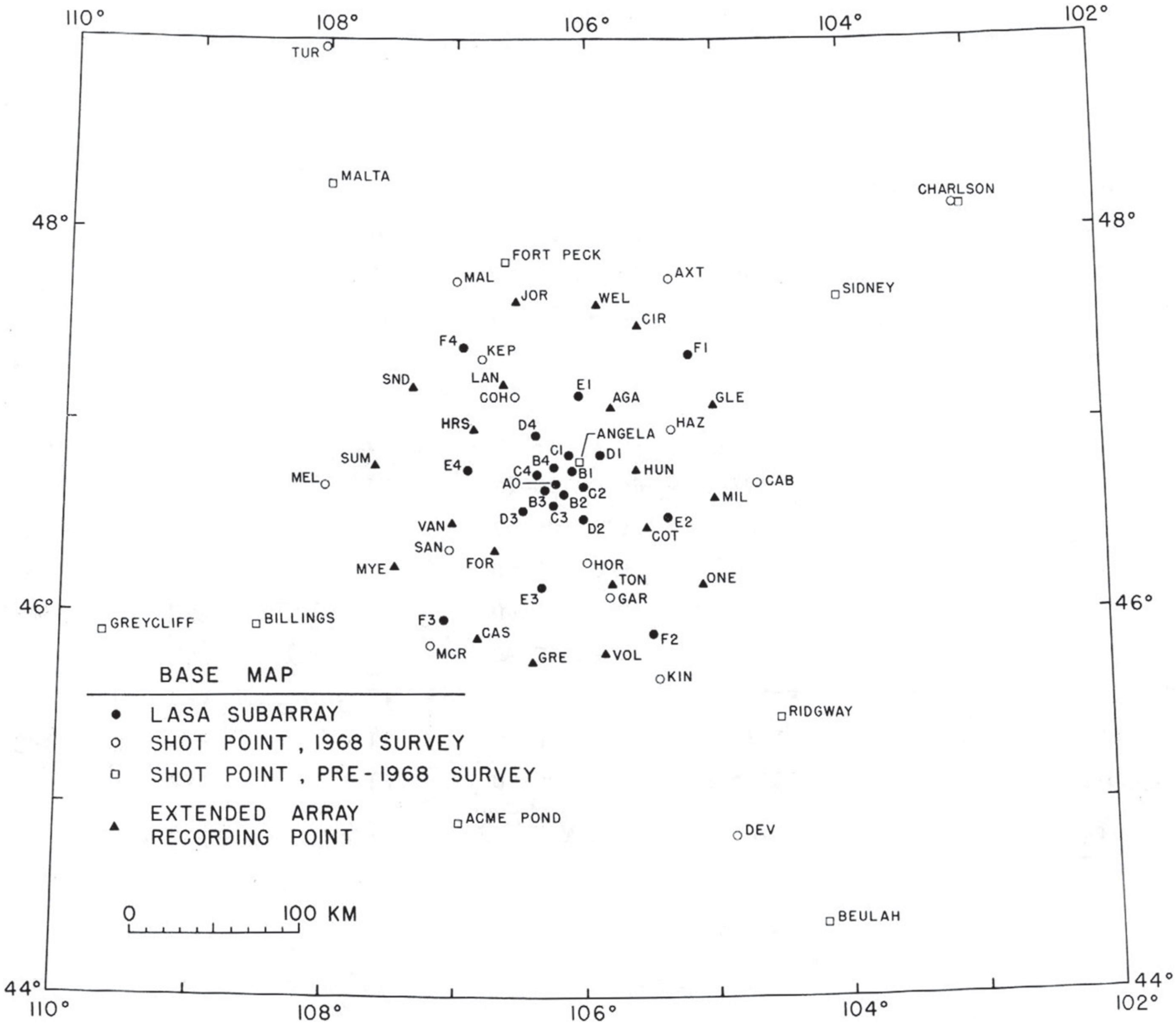


Figure 6.5.1-15. Shotpoint locations and temporary recording locations of the 1968 calibration project and LASA subarrays (from Warren et al, 1973, fig. 3). [Journal of Geophysical Research, v. 78, p. 8721–8734. Reproduced by permission of American Geophysical Union.]

part between shotpoints San Juan and Fernandez of 80 km length was sufficiently covered. The data are reproduced in Appendix A6-10.

Nevertheless, due to the overlapping data from five shotpoints along each line, the complete data could be successfully interpreted (Stewart, 1968a, Walter and Mooney, 1982). Another reinterpretation followed in conjunction with local earthquakes (Blümling and Prodehl, 1983; see also Chapter 7). The resulting overall crustal thicknesses, also including the investigations of other data sets by Steppe and Crosson (1978) for the northwestern side of the San Andreas fault system and of Healy and Peake (1975), and Prodehl (1979) for the Salinian side varied from author to author, as Walter and Mooney (1982) demonstrated, ranging between 26 and 30 km under the Diablo Range and between 24 and 26 km under the Gabilan Range. In earlier publications investigating the Salinian side, Hamilton et al. (1964), using quarry blasts for a profile between San Francisco and Salinas, had estimated 20–24 km, while McEvilly (1966) and Chuaqui and McEvilly (1968), using earthquake data, obtained 20 km total crustal thickness.

In their review of the first years of COCORP in the 1970s, Brewer and Oliver (1980) briefly review earlier seismic-reflection experiments. Besides the numerous large-scale seismic-refraction experiments to explore the Earth's

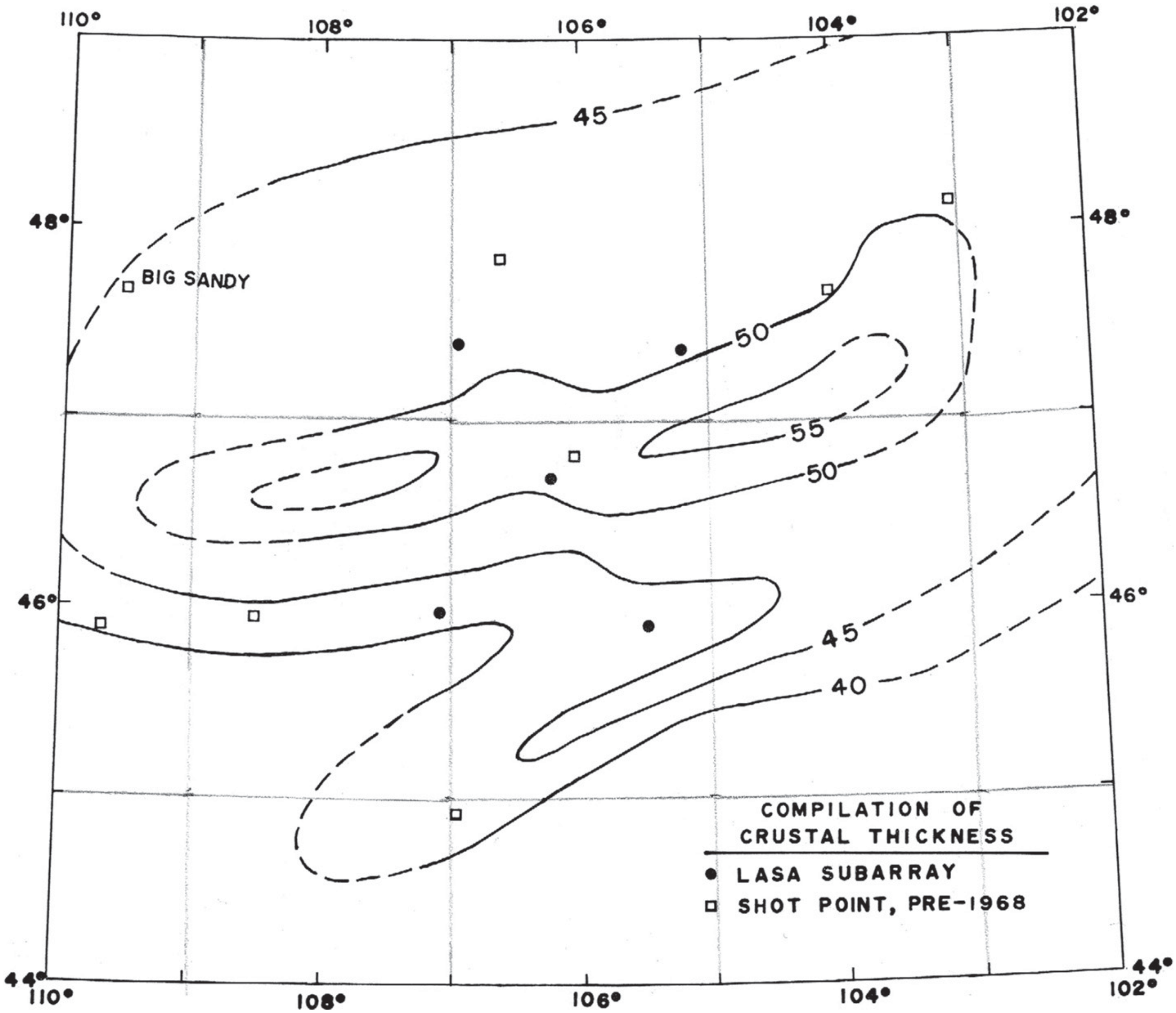


Figure 6.5.1-16. LASA Moho depth map derived from the 1968 calibration project (from Warren et al, 1973, fig. 18; Healy and Warren, 1974, fig. 5). [Journal of Geophysical Research, v. 78, p. 8721–8734; Tectonophysics, v. 20, p. 203–213. Reproduced by permission of American Geophysical Union.]

crust of the western United States, several attempts were undertaken already in the 1960s to study the crust using seismic-reflection techniques. One of the first attempts was a study by Narans et al. (1961) in Utah resulting in two reflecting horizons which roughly correlated with results from refraction profiles, but indicated a yet more complex structure. Another seismic-reflection study was undertaken in eastern Colorado. Hasbrouk (1964) tentatively identified the Moho and noted a complex crust and mantle structure. In an experiment in the Mojave Desert, Dix (1965) documented many reflectors of low dip between 8 and 35 km depth. In Wyoming in 1969, five deep crustal reflection profiles were obtained on a line crossing the Wind River Mountains. Here, deep crustal reflections at times as great as 14 s TWT (two-way traveltime) were recorded (Perkins and Phinney (1971).

6.5.2. Central and Eastern North America

In the south-central United States, several seismic-refraction studies concentrated on the structure of the Great Plains (Healy and Warren, 1969; Warren, 1968a, 1968b). Stewart and Pakiser (1962) interpreted an unreversed line, recorded northwards along the New Mexico–Texas border from the first nuclear test in southeastern New Mexico, GNOME. The model was republished

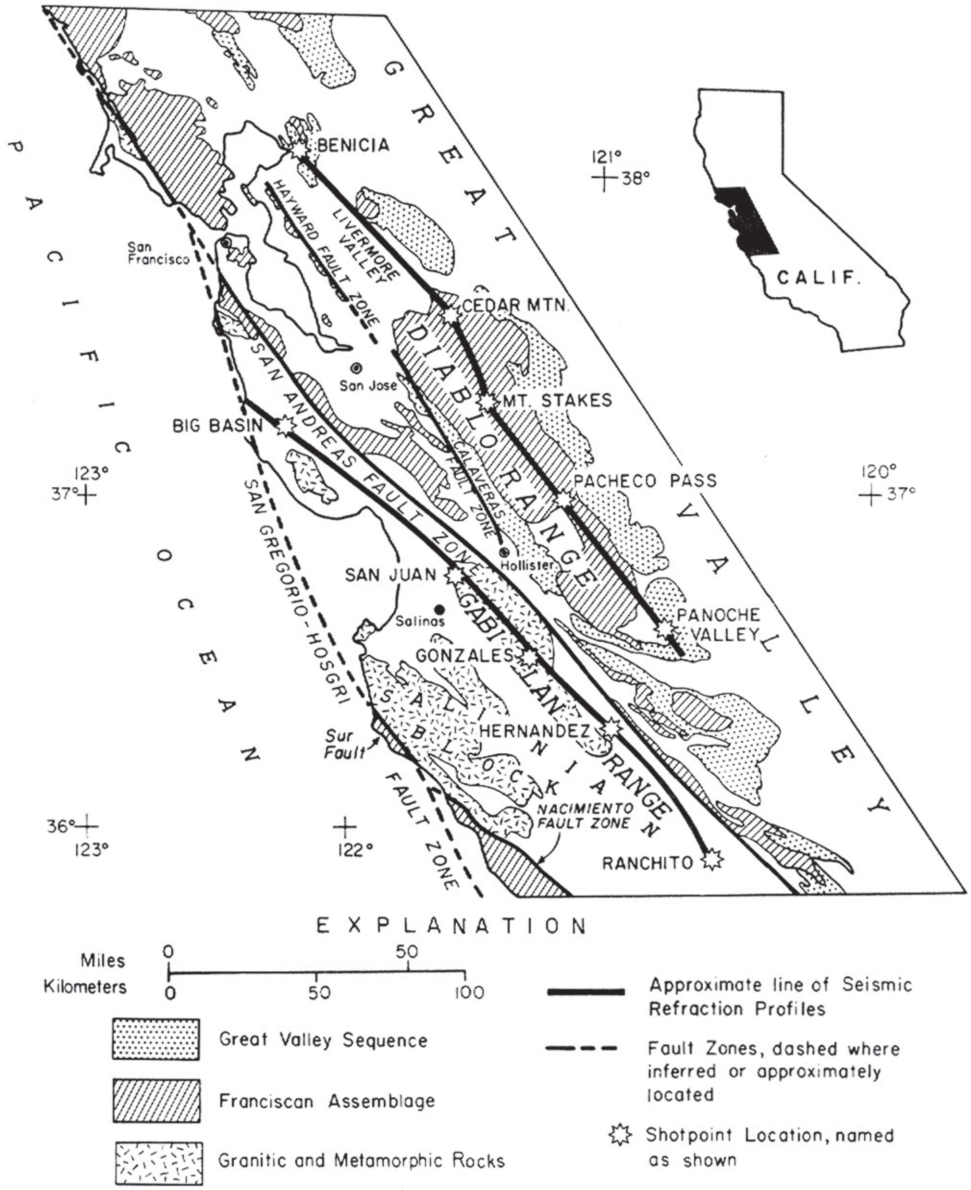


Figure 6.5.1-17. Seismic refraction lines with shotpoint locations in the Coast Ranges of central California along both sides of the San Andreas fault system (after Stewart, 1968a, from Walter and Mooney, 1982, fig. 1). [*In* Dickinson, W.R., and Grantz, A., eds., Proceedings of Conference on Geologic Problems of the San Andreas Fault System: Stanford University, School of Earth Science, p. 218–230; Bulletin of the Seismological Society of America, v. 72, p. 1567–1590. Reproduced by permission of the Seismological Society of America.]

in the compilation of Warren (1968b; see also Healy and Warren, 1969, fig. 2). The data were later reinterpreted by Mitchell and Landisman (1971) in conjunction with the interpretation of a 340-km-long reversed line through Oklahoma (Tryggvason and Qualls, 1967; see Fig. 6.5.2-01) with the shotpoints Manitou in the southwest and Chelsea in the northeast (Mitchell and Landisman (1971, data shown in Appendix A6-11). A >50-km-thick crust was the result of the data interpretations.

Stewart (1968b) interpreted a seismic-refraction network in Missouri, consisting of two 300-km-long lines, which were recorded in 1963 (Fig. 6.5.2-01; data shown in Appendix A6-11). Each line was reversed and contained a central shotpoint. South of the Missouri network, a 420-km-long line was recorded in 1962 from Cape Girardeau, Missouri, toward the southwest to Little Rock, Arkansas (Fig. 6.5.2-01), aiming to investigate the crustal structure of the northern Mississippi embayment. Five shotpoints were located in a bend of the Mississippi River and spaced such that they would be 1 km apart when projected on the recording line. Shots at each location were repeated when the recording stations had moved to new positions, until the largest distance of 420 km was reached (McCamy and Meyer, 1966). The average crustal thickness for Missouri resulting from these data was around 40 km. All lines, described above, were compiled into the fence diagram in Figure 6.5.2-01, which shows a unified simplified reinterpretation by Warren (1968c).

Finally, a detailed seismic-refraction study, located south of the Transcontinental Geophysical Survey and therefore not shown by Warren (1968c), concentrated on the structure of the southern part of the Mississippi embayment north of New Orleans (Warren et al., 1966). It was an almost 500-km-long line,

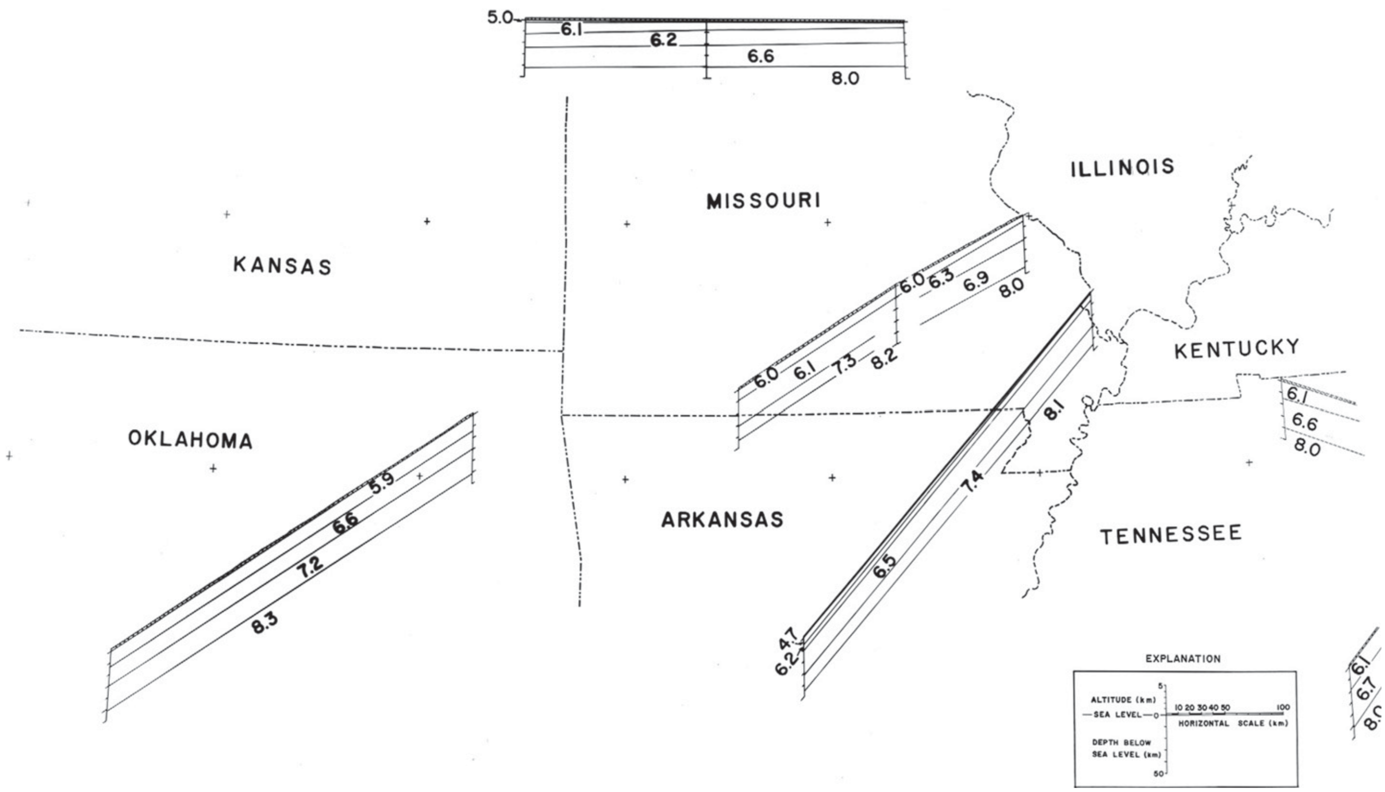


Figure 6.5.2-01. Fence diagram of crustal structure in the central-eastern portion of the Transcontinental Geophysical Survey (after Warren, 1968c, from Healy and Warren, 1969, fig. 3). Numbers are average velocities (km/s) of P-waves. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 208–220. Reproduced by permission of American Geophysical Union.]

trending north from Ansley, Louisiana, to Oxford, Mississippi, where ~200 seismograms were obtained. Along the southern 200 km section between Ansley and Raleigh, Mississippi, 22 shots were fired at 5 shotpoints spaced at intervals of ~50 km, creating an overlapping coverage. The section north of Raleigh was not reversed.

The main result was a crustal thickness of 41 km near Ansley at the Gulf Coast, decreasing toward north to near 29 km at Raleigh (Fig. 6.5.2-02). A location map and some of the data are shown in Appendix A6-11.

In the north-central United States and adjacent Canada, a target of special research became Lake Superior. As part of the

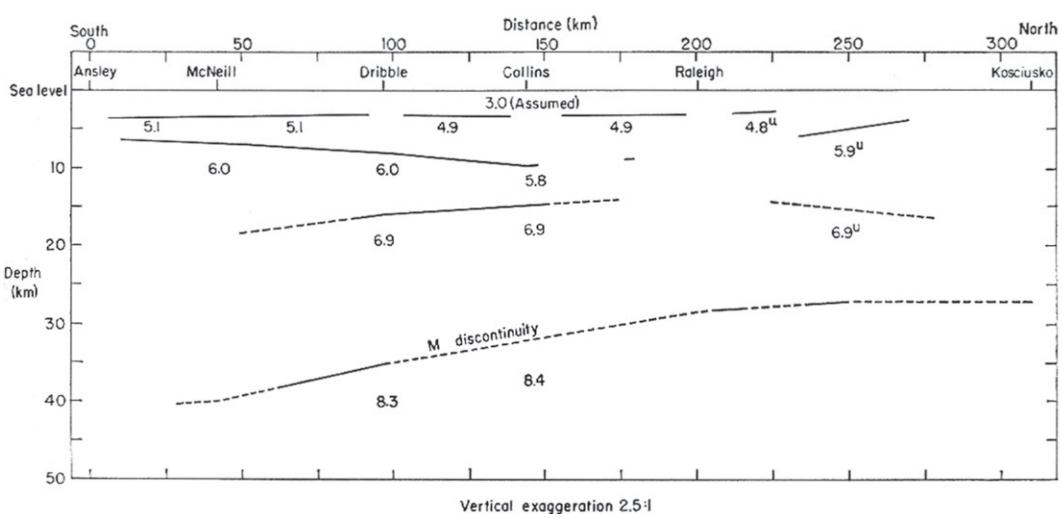


Figure 6.5.2-02. Crustal structure of the Mississippi embayment (after Warren et al., 1966, fig. 16). Numbers are average velocities (km/s) of P-waves. [Journal of Geophysical Research, v. 71, p. 3437–3458. Reproduced by permission of American Geophysical Union.]

VELA UNIFORM program in 1963 and 1964, three series of explosions were organized in the lake. In July 1963, 80 1-ton shots, in July 1964, 40 principally 1-ton shots, and in October 1964, 10 shots of 2–20 tons (Fig. 6.5.2-03) were recorded by 13 institutions from the United States and Canada at various azimuths and up to distances of more than 2500 km (Mansfield and Evernden, 1966).

The crust under Lake Superior was found to be unusual and rapidly varying in comparison with the Canadian Shield surrounding it (Smith et al., 1966). The depth of the Moho, for example, varies from 20 km west of the lake to 55 km and more in the eastern part of the lake. Cohen and Meyer (1966) used the explosions to study the Midcontinent gravity high along two reconnaissance profiles extending to the SSW through Minnesota and Wisconsin and found Moho depths of 46 km under the ridge of high gravity and 42 km under the low-gravity trough to the east.

In the southern Appalachians, a seismic-refraction survey with two 300-km-long crossing lines was carried out by the U.S. Geological Survey in 1965 across the Cumberland Plateau Seismic Observatory near Minville, Tennessee (Fig. 6.5.2-04). The data were preliminarily interpreted by Warren (1968d) and later reinterpreted (Prodehl, 1977; Prodehl et al., 1984; data shown in Appendix A6-11). A third but less densely recorded reconnaissance line with only 20–50 km spacing of the recording units ran through the eastern coastal plains to the Atlantic Ocean and was connected with the ECOOE (East Coast On-shore Off-shore Experiment) survey (profiles along the Atlantic coast line in Fig. 6.5.2-04; Hales et al., 1967, 1968; Warren, 1968d). For the majority of these lines, a preliminary interpretation was published in fence diagrams (Fig. 6.5.2-04; Warren, 1968d; Healy and Warren, 1969, fig. 4) in the frame of a series of maps published along a transcontinental survey covering the latitudes 35°–39°N (Hart, 1964; Pakiser and Zietz, 1965).

The more detailed interpretation of the network across the Cumberland Plateau Seismic Observatory near Minville, Tennes-

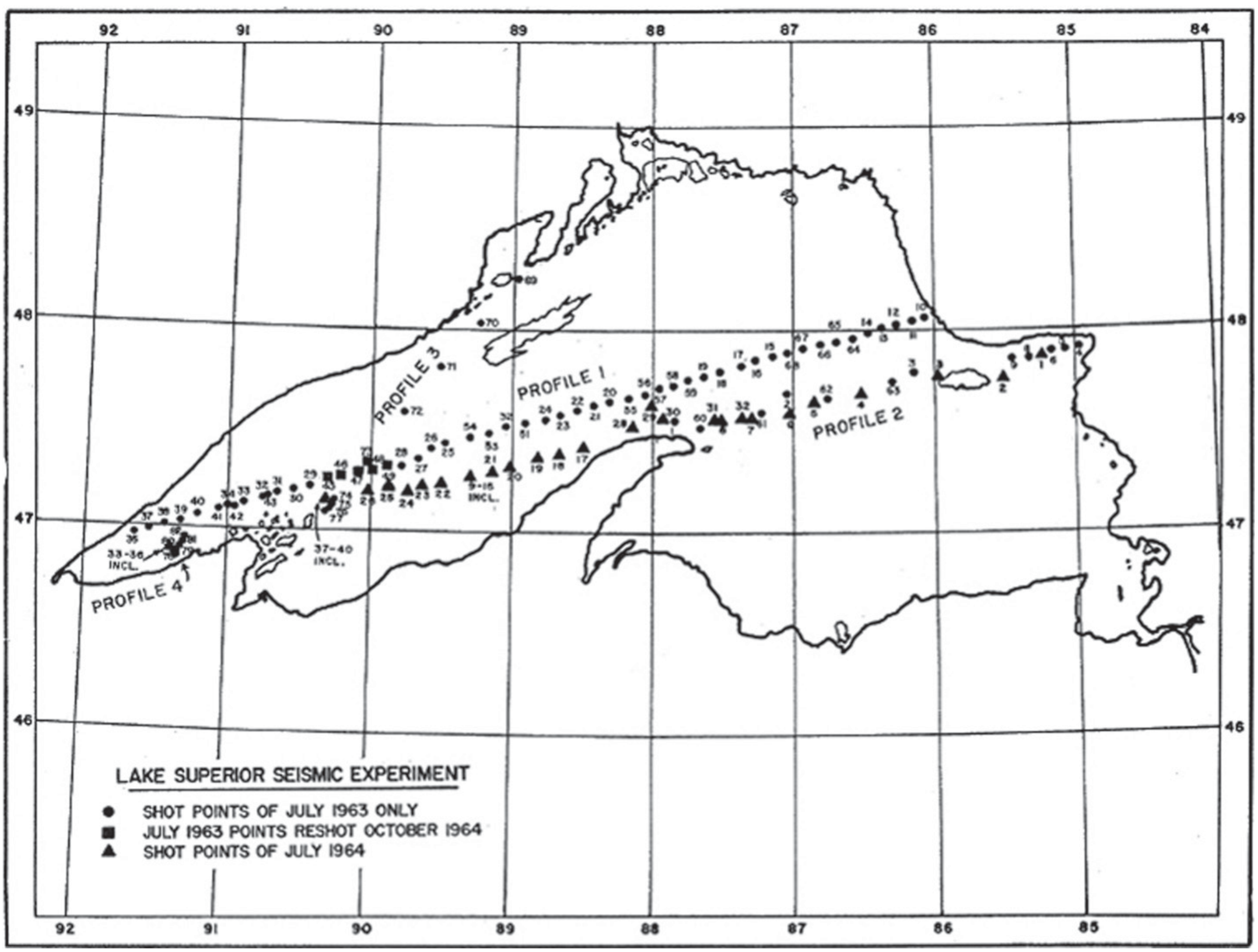


Figure 6.5.2-03. Lake Superior shot point map of 1963–1964 (from Mansfield and Evernden, 1966, fig. 1). [*In* Steinhart, J.S., and Smith, T.J., eds., The Earth beneath the continents: American Geophysical Union, Geophysical Monograph 10, p. 249–269. Reproduced by permission of American Geophysical Union.]

see (Fig. 6.5.2-04), where each profile had two end and two middle shotpoints, revealed that the lower crust was characterized by two distinct layers with velocities in the range of 6.7–6.8 km/s and 7.3–7.4 km/s. However, the boundaries between crustal layers as well as the crust-mantle boundary were not sharp but appeared as velocity transition zones ranging up to 11 km in thickness (Fig. 6.5.2-05). The high-velocity gradients between the various layers instead of distinct boundaries resulted from the fact that in the corresponding record sections clear reflected phases could hardly be detected. Instead, however, the first arrivals between 50 and 250 km shotpoint distance could be correlated as refracted phases resulting in unusual high velocities, and beyond 250 km distance clear P_n arrivals were present in the data (Fig. 6.5.2-06). Also, the total crustal thickness was rather high, between 49 and 55 km (Prodehl, 1977; Prodehl et al., 1984).

6.5.3. Achievements in the United States

The available interpretations on crustal structure in the United States of North America were reviewed from time to time (see, e.g., Pakiser and Steinhart, 1964; Healy and Warren, 1969) and in particular compiled for a transcontinental geophysical survey throughout the United States between 35° and 39°N latitude (Pakiser and Zietz, 1965).

A comprehensive summary on the state of the art of crustal structure interpretation as available in 1968, not including the reinterpretation of Prodehl (1970), can be found in Healy and Warren (1969). As already described above, it shows in maps fence diagrams of crustal cross sections with a unified simplified reinterpretation by D.H. Warren (Warren, 1968a, 1968b, 1968c, 1968d; see, e.g., Figs. 6.5.2-01 and 6.5.2-04), based on preliminary interpretations by various authors as, e.g., cited by Pakiser (1963), for most of the seismic-refraction profiles observed by the U.S. Geological Survey from 1961 to 1965.

The overview was complemented in 1970 by Warren and Healy (1973) with a map of all profiles recorded until 1970 within the United States and a table of references (Fig. 6.5.3-01 and Table 6.5.3-01) and a Moho contour map (Fig. 6.5.3-02).

The most common interpretation methods, which were used for the interpretation of seismic data by the end of the 1960s in North America, were described in much detail by Warren et al. (1972; Appendix A6-9) at the example of the interpretation of the detailed data set obtained during the seismic investigation of the deep structure under the LASA, Montana, array. A quite

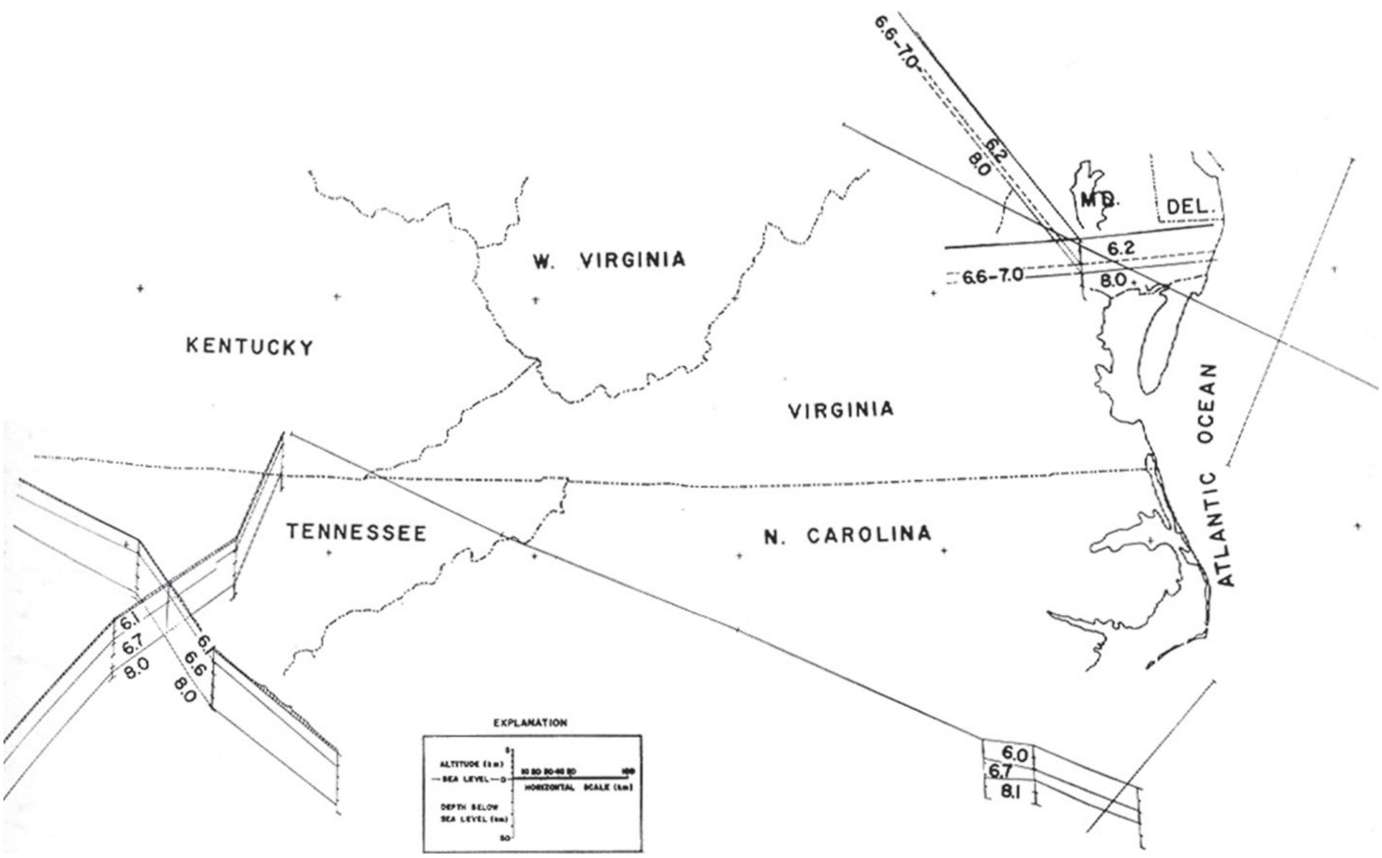


Figure 6.5.2-04. Fence diagram of crustal structure in the central-eastern portion of the Transcontinental Geophysical Survey (after Warren, 1968d, from Healy and Warren, 1969, fig. 4). Numbers are average velocities (km/s) of P-waves. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 208–220. Reproduced by permission of American Geophysical Union.]

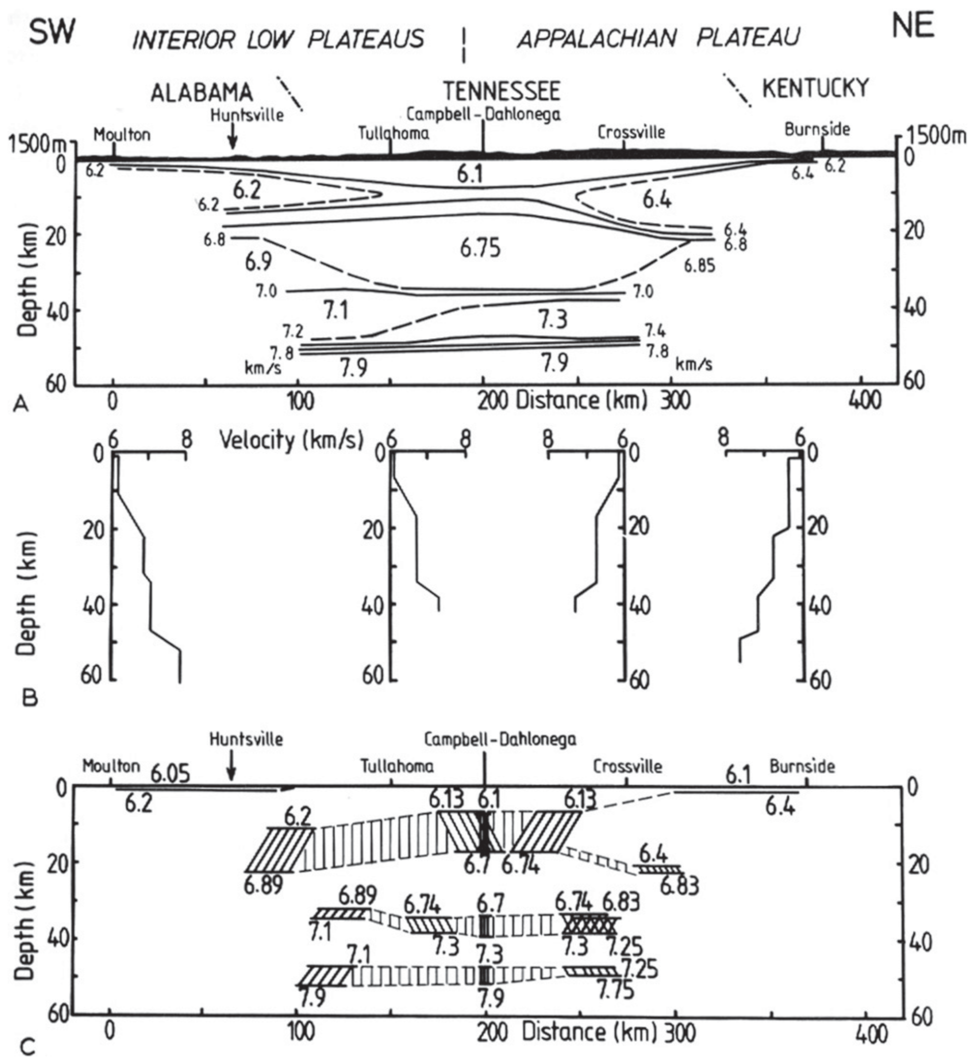


Figure 6.5.2-05. Crustal cross section between Moulton, Alabama and Burnside, Kentucky (from Prodehl et al., 1984, fig. 7). (A) Lines of equal velocity. (B) Individual velocity-depth profiles. (C) Position of velocity transition zones. [Tectonophysics, v. 109, p. 61–76. Copyright Elsevier.]

different approach, proposed by Giese (1968a, 1976b) for a rapid calculation of the velocity-depth distribution from any given travel-time curve system (see section 6.2.5) was described and applied by Prodehl (1979) on the majority of seismic-refraction profiles recorded by the U.S. Geological Survey in the western United States.

Based on a combination of data from deep seismic sounding, underground nuclear explosions and earthquakes, Herrin and Taggart (Herrin, 1969; Fig. 1) have compiled a map showing the estimated P_n velocity. The overall velocity was 8.0 km/s and above, except for the areas west of the Rocky Mountain front to the Pacific coast, where velocities smaller than 8.0 km/s prevailed.

6.5.4. Canada and Alaska

The achievements of crustal and uppermost mantle structure research until the end of the 1960s were summarized by Berry (1973). Figure 6.5.4-01 shows the seismic lines and average values of crustal thickness and upper-mantle P-wave velocities.

The Cordillera of western Canada had been studied by a kind of reconnaissance survey for over a decade (Berry, 1973). As a result, the Rocky Mountain trench appeared to be the western limit of thick crust and also of the North American craton (Berry et al., 1971). Under the central plateaus, between the Rocky and Coast Mountains, a crustal thickness of 30–35 km was obtained (White et al., 1968).

Detailed seismic-refraction experiments in southern Alberta and Saskatchewan revealed a 47-km-thick crust under the prairies of southern Alberta which decreased to the north with upper-mantle velocities of 8.2–8.3 km/s (Kanasewich, 1966).

A pioneering study was undertaken in 1964 by a group of seismologists of the University of Alberta under the leadership of Ernic Kanasewich. For the first time in North America, they used the near-vertical incidence seismic-reflection techniques, as used by the exploration industry, for an investigation of the whole crust and demonstrated that this technique was a viable tool of determining detailed crustal structure and imaging the Moho (Kanasewich and Cumming, 1965; Clowes et al., 1968;

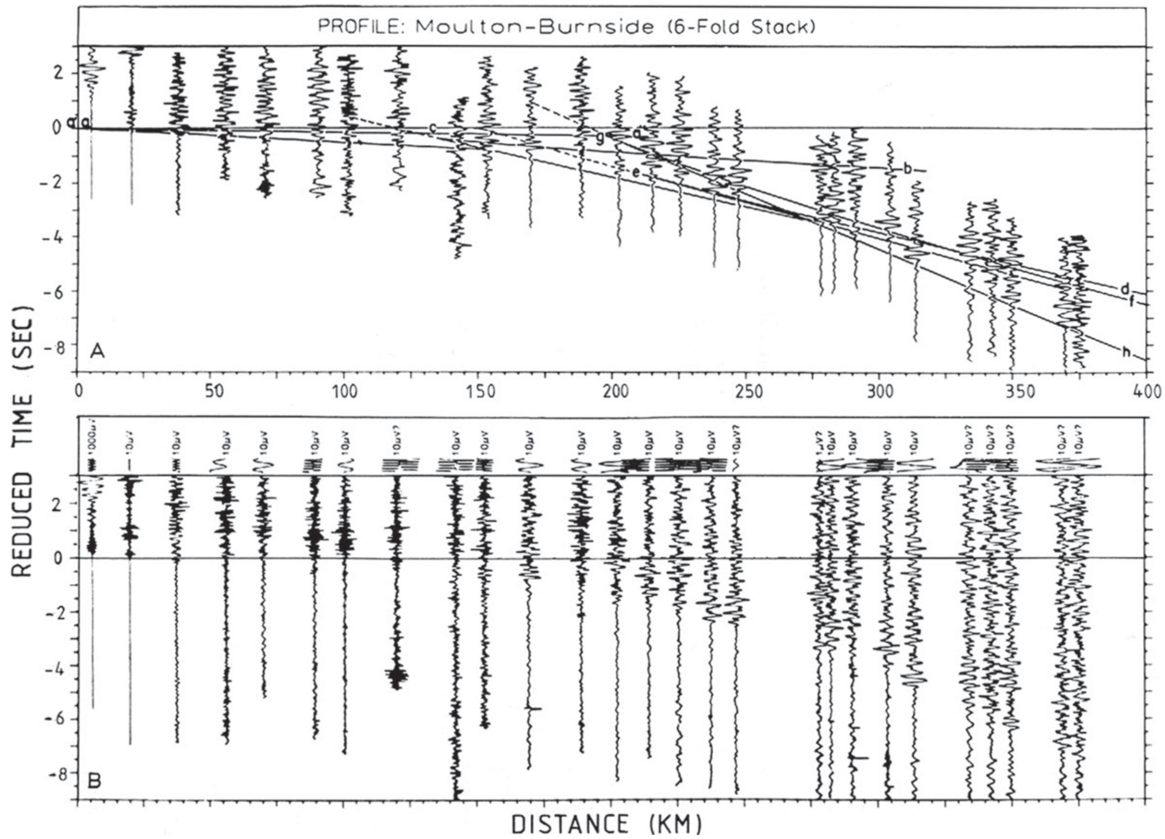


Figure 6.5.2-06. Record section of the profile Moulton–Burnside (from Prodehl et al., 1984, fig. 4). (A) Six-fold optimally stacked traces. (B) "Best" traces for each array. [Tectonophysics, v. 109, p. 61–76. Copyright Elsevier.]

Kanasewich et al., 1969). As outstanding results, they reported on reflected energy from the Moho "M" from 38 to 47 km depth and from the Reil boundary "R" at 26–34 km depth, the layer in between having velocity of 6.5–7.2 km/s. The seismic-reflection data revealed a surprisingly complex structure, including topography of 8 km over a horizontal distance of 25 km and dips of as great as 20° on some reflecting horizons at depth (Clowes et al., 1968).

In the same area, on the basis of combined reflection and refraction data, good evidence was found for major steeply dipping faults in the middle and lower crust which penetrated across Moho. Clowes and Kanasewich (1970) used the seismic near-vertical incidence reflection data and undertook detailed studies of an intra-crustal reflector (Conrad discontinuity), the Moho, and crustal Q. They concluded that there was significant fine structure at the base of the crust and proposed a lamellar model of alternating high and low velocity material at the base of the crust, as had been proposed by Fuchs (1969) for southern Germany. In another study, Clowes and Kanasewich (1972) investigated the special characteristics of the reflections by applying industry digital filtering techniques to enhance crustal reflections and enable a migrated "squiggle" section to be developed. Following the initial success of work in southern Alberta, Ernic Kanasewich helped to carry out a similar survey in Wyoming in 1969, in cooperation with Bob Phinney and co-workers (Perkins and Phinney, 1971; see subchapter 6.5.1; Clowes, 2010, personal commun.). In conjunction with the earlier work on deep reflections in Germany (Dohr, 1957, 1959; Krey et al., 1961; Liebscher, 1962, 1964; German Research Group for Explosion Seismology, 1964; Dohr and Fuchs, 1967) and Russia (Beloussov et al., 1962; Kosminskaya and Riznichenko, 1964; Zverev, 1967) the studies by the University of Alberta group paved the way for COCORP and subsequent deep reflection programs around the world in the 1970s and 1980s.

A large experiment was carried out in 1966 in northwestern Canada to calibrate the Yellowknife Seismic Array (Fig. 6.5.4-01). For this purpose, 17 explosions of 1–2 ton charges were detonated in lakes at distances ranging from 15 to 480 km. An average Moho depth of 37.5 km resulted (Weichert and Whitham, 1969). Another research project in 1969 investigated the fine crustal structure of the Yellowknife area by a combined reflection-refraction experiment on two lines with 250 m station spacing (Clee et al., 1974).

The Lake Superior experiments in 1963 and 1964 as well as the Early Rise experiment in 1965, which will be discussed in more detail below, also provided data for crustal structure studies

Figure 6.5.3-01. Map of seismic-refraction profiles in the conterminous United States, recorded to the end of the 1960s (from Warren and Healy, 1973, fig. 7). Numbers refer to the references in Table 6.5.3-01. [Tectonophysics, v. 20, p. 203–213. Copyright Elsevier.]

TABLE 6.5.3-01. REFERENCES TO THE MAP OF SEISMIC-REFRACTION PROFILES IN THE CONTERMINOUS UNITED STATES IN FIG. 6.5.3-01

1. Shor et al. (1968)	28. Merkel and Alexander (1969)
2. White and Savage (1965)	29. Asada et al. (1961)
3. Johnson and Couch (1970)	30. Steinhart and Meyer (1961)
4. Hill (1972)	31. Aldrich et al. (1960)
5. Berg et al. (1966)	32. McCamy and Meyer (1964)
6. Shor and Raitt (1958)	33. Warren et al. (1972)
7. Shor (1955)	34. Gibbs (1972)
8. Healy (1963)	35. Tuve (1953)
9. Eaton (1966)	36. Smith et al. (1966)
10. Eaton (1963)	37. O'Brien (1968)
11. Stewart (1968a)	38. Cohen and Meyer (1966)
12. Johnson (1965)	39. Slichter (1951)
13. Carder et al. (1970)	40. Roller and Jackson (1966)
14. Hill and Pakiser (1966)	41. Barrett et al. (1964)
15. Gibbs and Roller (1966)	42. Stewart (1968b)
16. Roller and Healy (1963)	43. Tryggvason and Qualls (1967)
17. Roller (1964)	44. McCamy and Meyer (1966)
18. Diment et al (1961)	45. Warren et al. (1966)
19. Steinhart et al. (1962)	46. Roller et al (1967)
20. Willden (1965)	47. Cram (1961)
21. Roller (1965)	48. Hales et al. (1970)
22. Warren (1969)	49. Ewing et al. (1966)
23. Warren and Jackson (1968)	50. Katz (1955)
24. Stewart and Pakiser (1962)	51. Hart (1954)
25. Jackson and Pakiser (1965)	52. Steinhart et al. (1963)
26. Jackson et al. (1963)	53. James et al. (1968)
27. Ryall and Stuart (1963)	54. Hales et al. (1968)

Note: From Warren and Healy (1973, table 1).

of the Churchill and Superior Precambrian provinces of Canada. They showed in particular that the crustal thickness under the Canadian Shield varies substantially. After Hodgson (1953) had derived a 35-km-thick crust from the observation of rockbursts in the Kirkland Lake area of northern Ontario, the new data and results of the Lake Superior experiments stimulated a variety of new experiments to study the Shield in more detail.

In the Hudson Bay experiment of 1965 (Hobson, 1967) and the Grenville Front seismic experiment in 1968 (Mereu and Jobidon, 1971; Berry and Fuchs, 1973), several Canadian research institutions cooperated. The Grenville Front seismic experiment, investigating the Grenville Front, was a large-scale reconnaissance refraction experiment with three parallel, SE-NW–trending profiles investigated the Grenville and Superior provinces of the Canadian Shield (Berry and Fuchs, 1973), north of the St. Lawrence River and immediately west of Appalachia (Fig. 6.5.4-01). Here, 26 shots of sizes between 2000 and 6000 lbs were fired by the Dominion Observatory in six different lakes and recorded up to 800 km distance traversing a major part of the Canadian Shield between Hudson Bay and the Gulf of St. Lawrence.

The seismic-refraction studies of the crust of the Appalachian region which had started as early as 1956 (Willmore and Scheidegger, 1956) were enforced in the early 1960s (Dainty et al., 1966, Ewing et al., 1966, Berry, 1973). At the flanks of the

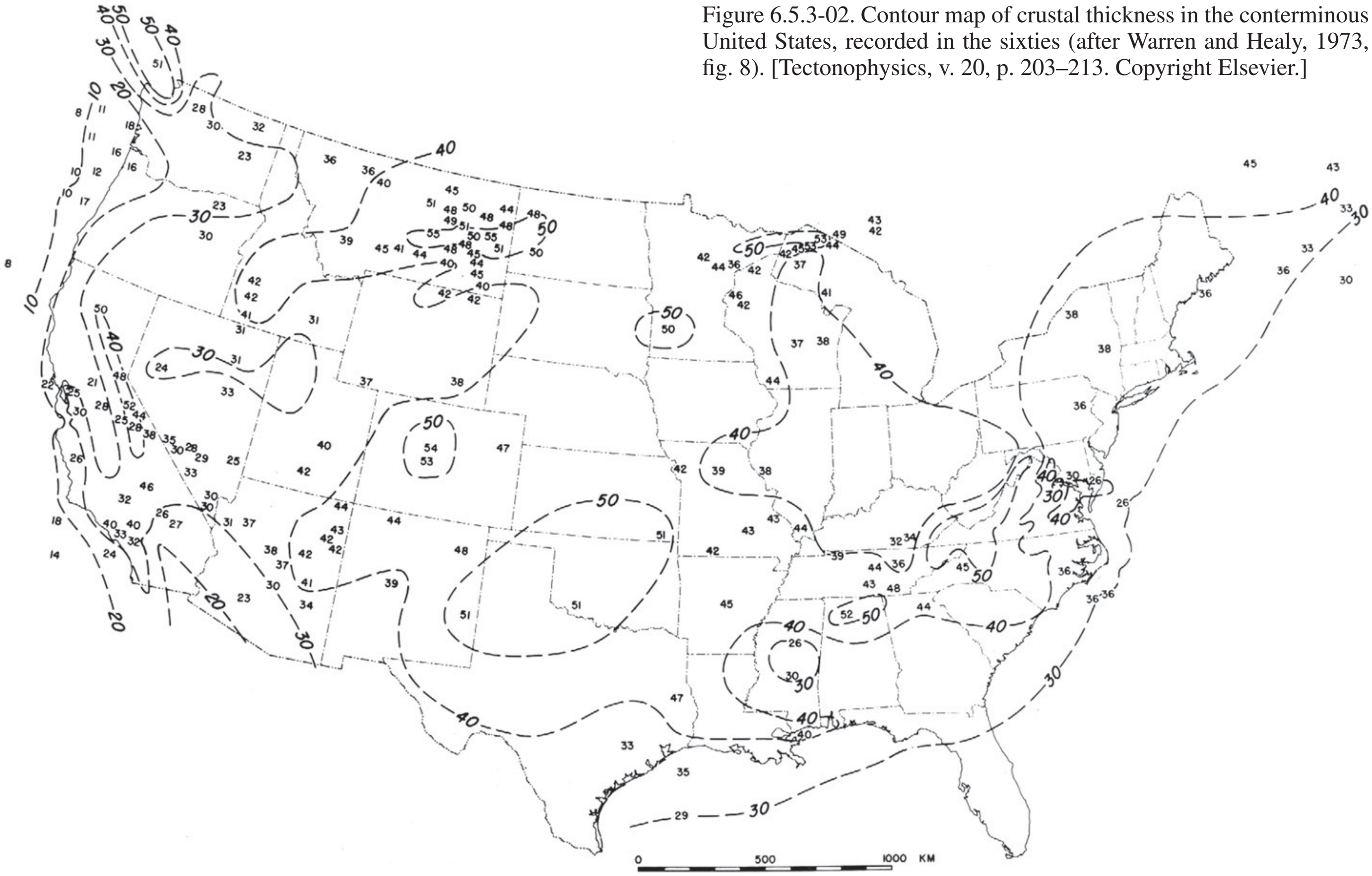


Figure 6.5.3-02. Contour map of crustal thickness in the conterminous United States, recorded in the sixties (after Warren and Healy, 1973, fig. 8). [Tectonophysics, v. 20, p. 203–213. Copyright Elsevier.]

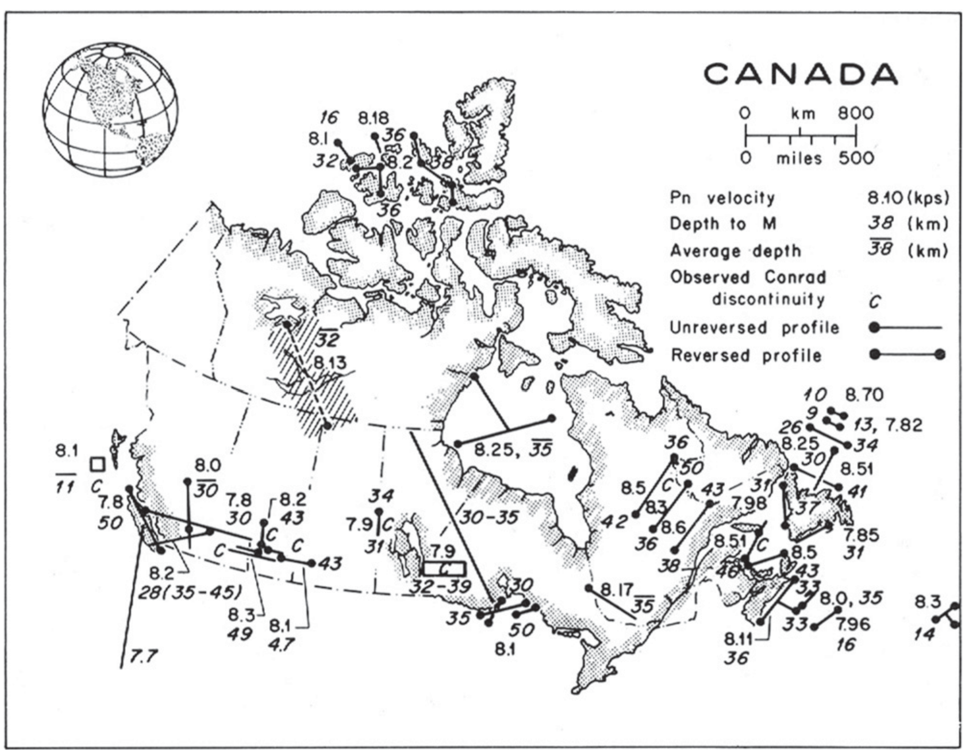


Figure 6.5.4-01. Map of seismic-refraction profiles in Canada, recorded to the end of the sixties (from Berry, 1973, fig. 1). [Tectonophysics, v. 20, p. 183–201 Copyright Elsevier.]

Appalachian system, crustal thicknesses between 30 and 40 km were obtained with an average crustal velocity of less than 6.5 km/s and an uppermost-mantle velocity close to 8.0 km/s. In the center of the system, a thick intermediate layer with 7.3–7.5 km/s velocity was seen causing the crustal thickness to increase to more than 40 km. Also, a much higher uppermost-mantle velocity of 8.5–8.7 km/s resulted (Figs. 6.5.4-01 and 6.5.4-02). Well-defined high uppermost-mantle velocities of 8.7–8.8 km/s at 52 km depth were also observed on data recorded from a number of shots from the north on a profile across the Gaspe Peninsula (Rankin et al., 1969).

In central Alaska during the winter of 1967–1968, an unreversed 217-km-long seismic-refraction profile was recorded from blasts in the Alaska Range toward north through the Tanana Basin, ending ~30 km north of Fairbanks (Fig. 6.5.4-03). Assuming a true P_n velocity of around 8 km/s, the interpretation of the observed 8.8.km/s apparent velocity results at a Moho depth of 31 km under Fairbanks dipping southward under the Alaska Range to ~48 km depth (Hanson et al., 1968; Berg, 1973). Carder et al. (1967) investigated the seismic data obtained from the LONGSHOT underground nuclear explosion on Amchitka Island. It was recorded at 27 sites in Alaska and 2 sites in Yukon Territory at distances from 2.6° to 26.6°. Upper-mantle traveltimes correlated to distances of 11° resulted in an average velocity of 7.85 km/s and 22–26 km Moho depth beneath the island chain.

6.5.5. Upper Mantle Projects in North America

The possibility of using lakes as shotpoints to record very long seismic-refraction profiles was recognized immediately. One of the first long-range lines was recorded in 1964 by the U.S. Geological Survey from Lake Superior to Denver at ~1400 km distance (Roller and Jackson, 1966). A special volume of the American Geophysical Union, edited by Steinhart and Smith (1966) and dedicated to M.A. Tuve, deals in particular with results from these surveys, but also contains other seismic, seismological and compositional studies related to the Earth's crust and mantle in various parts of North America and other continents, including a critical review of recent explosion studies by James and Steinhart (1966).

Following the experience from the previous Lake Superior experiments, in 1965 the Project Early Rise emerged with

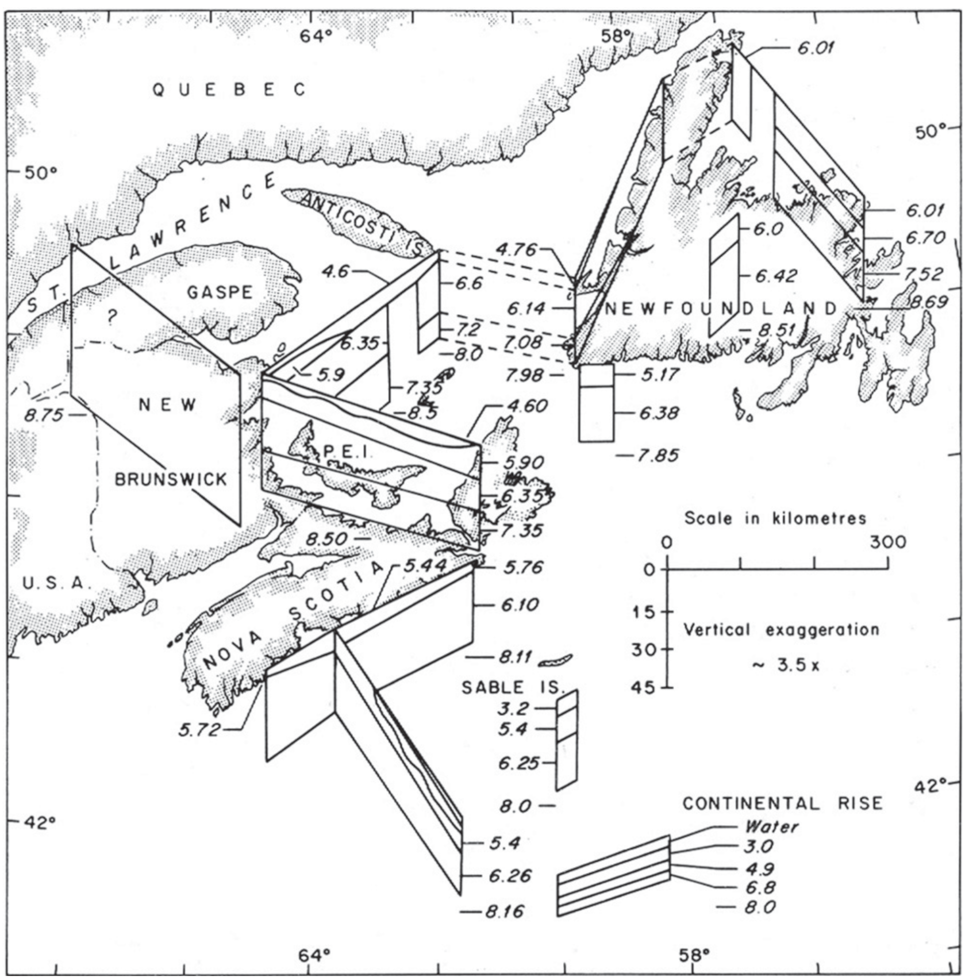


Figure 6.5.4-02. Crustal models through the Canadian Appalachian system (from Berry, 1973, fig. 2, after Ewing et al., 1966, with additions). [Tectonophysics, v. 20, p. 183–201. Copyright Elsevier.]

recording distances of 1000 km and longer. Thirty-eight 5-ton shots were detonated in the lake and recorded along lines in many directions from Lake Superior (Fig. 6.5.5-01) by groups from North American universities and government agencies. Due to its success in obtaining explosion seismic energy to large recording distances, this experiment, as with the Nevada Test Site observations, resulted in insight into the uppermost mantle below the Moho, due to the long recording distances.

A data example in Figure 6.5.5-02 shows the data recorded from Lake Superior into Colorado. The data of the Early Rise Project have been interpreted by many authors. Comprehensive interpretations were published, e.g., by Iyer et al. (1969), Mereu and Hunter (1969), Ansorge (1975), and Massé (1973).

A second large-scale crustal and upper-mantle seismic experiment, the Project Edzoe, was undertaken jointly by Canadian and American scientists in 1969 and aimed to study the structure of the North American Great Plains and the Rocky Mountains. Twenty explosions with charges between 5700 and 7700 kg were fired by the Dominion Observatory in Greenbush Lake near Revelstoke, British Columbia, and were recorded on several lines (up to 800–1000 km distance). The observations covered both Canada and the United States. Contrary to the Early Rise project of 1965, however, a joint publication evidently never exceeded an internal report status for project Edzoe of 1969, so a location map of all lines did not become publicly available. The interested reader has to rely on publications of the many individual lines.

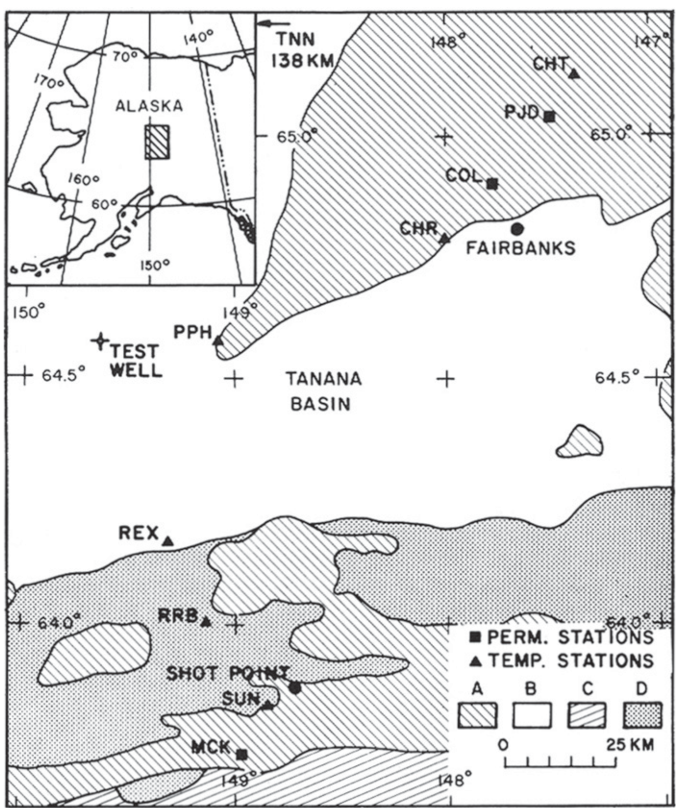


Figure 6.5.4-03. Location of a seismic-refraction line around Fairbanks, central Alaska (from Berg, 1973, fig. 8, after Hanson et al., 1968, fig. 1). [Tectonophysics, v. 20, p. 165–182. Copyright Elsevier.]

Individual interpretations of the long-range profiles recorded in Canada were published. These were profiles recorded across the Canadian Cordillera (Berry and Forsyth, 1975; Forsyth et al., 1974), through the North Cascade Mountains of Washington and British Columbia (Johnson and Couch, 1970), along the Rocky Mountain Front Range (Mereu et al., 1977), through the Great Plains of southern Alberta (Chandra and Cumming, 1972) and southern Saskatchewan and western Manitoba (Bates and Hall, 1975). In the United States, one line extended southwestward through the Columbia Plateau and was interpreted by Hill (1972). Two other lines were essentially north-south–oriented and traversed the central United States (Hales and Nation, 1973a). The first line was directed toward the Big Bend area in western Texas and traversed mainly the Rocky Mountains. The second line extended toward the SE corner of Kansas and covered the Great Plains area. Hales and Nation (1973a) correlated essentially one phase on both lines. On the Rocky Mountain line, they correlated a P_n phase in the first arrivals from 185 to 700 km distance, while on the Great Plains profile they observed P_n up to 1000 km distance. On both lines the average P_n velocity was ~8 km/s. Contrary to this interpretation, Mereu et al. (1977) correlated a high-velocity upper-mantle traveltime branch, beginning at distances beyond 300 km, and attributed it to a high-velocity layer with 8.5–8.8 km/s below 60 km depth, in correspondence to results found on European long-range profiles observed in the 1970s, which will be discussed in detail in Chapter 7.

6.6. EXPLOSION SEISMOLOGY RESEARCH IN SOUTHEAST ASIA IN THE 1960s

Following the start of the Upper Mantle Project, considerable progress was made in instrumentation and seismic fieldwork from 1964 onwards in Japan (Research Group for Explosion Seismology, 1966a). Several new lines on the island of Honshu were added to the early observations of the 1950s, described above (Aoki et al., 1972; Yoshii and Asano, 1972; Okada et al., 1973; Yoshii et al., 1974). For example, in southern Honshu, two reversed 300-km-long profiles were recorded, one along a north-south line at 139°E longitude, where at the same time an earthquake could also be recorded, and the other along an east-west line at ~35.5°N latitude. Furthermore, a series of 10 underwater shots off the eastern coast was recorded along a SE-NW line across northern Honshu and added new information to the very first quarry blast lines observed in the early 1950s.

A comprehensive map (Fig. 6.6-01) shows all seismic-refraction observations from 1950 to 1970. As a result, the one-layer crustal model of the 1950s could be replaced by a two-layer model, based on first arrivals with velocity greater than 6.4 km/s. In conclusion, the depth to the Moho is between 20 and 30 km, but increases by ~5 km in the central mountainous areas

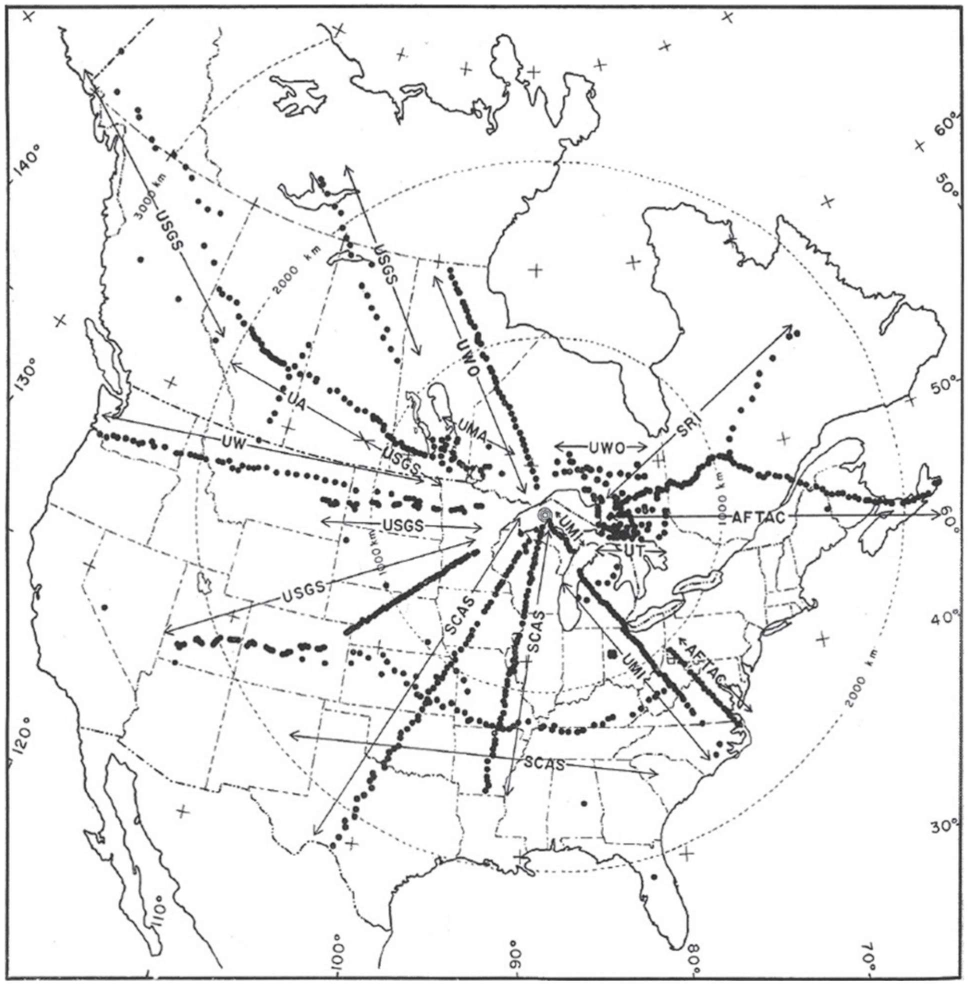


Figure 6.5.5-01. Location of recording stations for Project Early Rise of 1965. Shots were fired from the same location in Lake Superior. The codes for groups that recorded are indicated at the corresponding lines: AFTAC—Air Force Technical Applications Center; UA—University of Alberta; UMA—University of Manitoba; UMI—University of Michigan; USGS—U.S. Geological Survey; SCAS—Southwest Center for Advanced Studies; SRI—Stanford Research Institute; UT—University of Toronto; UW—University of Wisconsin; UWO—University of Western Ontario (from Healy and Warren, 1969, fig. 11). [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 208–220. Reproduced by permission of American Geophysical Union.]

of Honshu. The existence of an additional intermediate layer with 6.7–6.8 km/s is not unique, but if it is assumed, the Moho depths would increase by 10 km (Aoki et al., 1972; Yoshii and Asano, 1972; Research Group for Explosion Seismology, 1966a, 1966b, 1973; Yoshii et al., 1974). In general, the upper-mantle velocity is ~7.7 km/s under Japan. A crustal cross section through central Japan along line 3 is shown in Figure 6.6-02. A crustal profile was also recorded along the southern coast of Hokkaido (Fig. 6.6-01; Okada et al., 1973; Research Group for Explosion Seismology, 1973). The crustal cross section also includes marine seismic-refraction results (Ludwig et al., 1966).

6.7. EXPLOSION SEISMOLOGY RESEARCH IN THE SOUTHERN HEMISPHERE CONTINENTS IN THE 1960s

6.7.1. Australia

Crustal research in Australia using explosion seismology was already quite extensive in the 1960s due to the existence of a small but very active geophysical community.

Between 1959 and 1964, the structure of crust and upper mantle of southwest Western Australia was the goal of a variety of seismic-refraction investigations. The Mundaring Geophysical Observatory of the Bureau of Mineral Resources, Geology and Geophysics (BMR) acquired seismic data from a variety of quarry blasts and marine shots recorded at stations throughout southwestern Australia. In 1960, the Lamont Geological Observatory (USA) research vessel *Vema* conducted seismic investigations on the offshore Perth Basin. In 1962 the Scripps Oceanographic Institute (USA) vessel *Argo* exploded four charges off Perth. In 1963, the Royal Australian Navy ship HMAS *Diamantina* detonated six charges. The data resulted in crustal models both for southwestern Australia and the Western Australia margin (Everingham, 1965; Hawkins et al., 1965a, 1965b; Finlayson, 2010; Appendix 2-2). The crustal investigations using local earthquakes and quarry blasts were continued during 1963–1967 with a network of widely spaced stations in the southern part of Southern Australia (White, 1969).

Following the seismic observations of the Emu and Maralinga nuclear explosions of 1953 and 1956 and of some large explosions at Eaglehawk quarry in the Snowy Mountains in

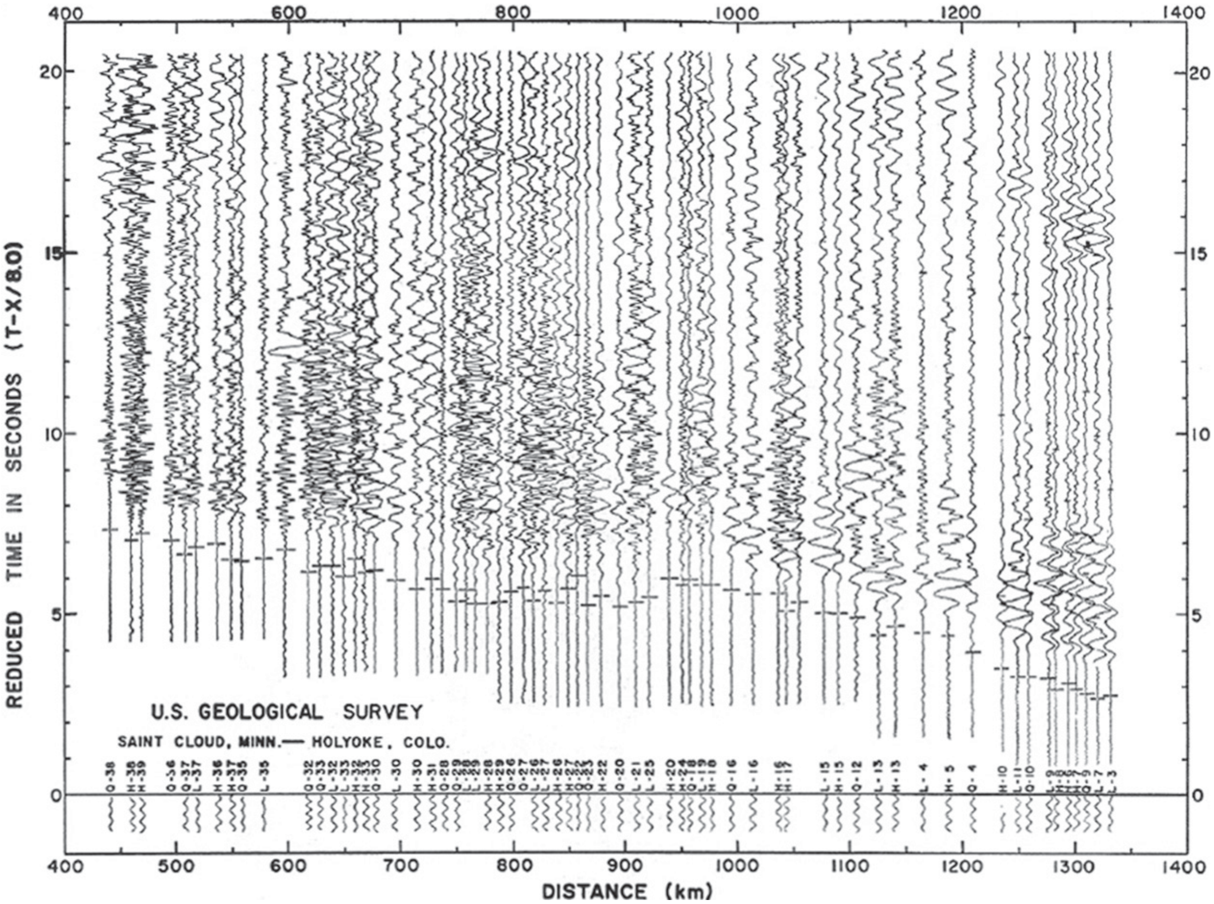


Figure 6.5.5-02. Record section of the Early Rise long-range seismic-refraction profile from Lake Superior, Wisconsin, to Denver, Colorado recorded by the U.S. Geological Survey in 1965 (from Healy and Warren, 1969, fig. 12). [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 208–220. Reproduced by permission of American Geophysical Union.]

1956–1957, several seismic experiments followed in the 1960s which are summarized in Figure 6.7.1-01 (Cleary, 1973). The Eaglehawk quarry observations were followed by several series of offshore shots, both off New South Wales in 1965 and those of the BUMP (Bass Strait Upper Mantle Project) experiment in the Bass Street in 1966.

The resulting crustal model (Fig. 6.7.1-02) shows a 40+ km thick crust under southeastern Australia, thickening by ~10 km under the Eastern Highlands. The data of the Maralinga explosions were supplemented in 1969 when a seismic Western Australia Geotraverse was conducted within the Archaean shield. In 1969, deep crustal refractions and reflections were recorded along the Geotraverse in southwestern Australia (Mathur, 1974). The line extended eastward from Perth to Coolgardie and continued in a southeast direction toward Point Culver on the Great Australian Bight. Figure 6.7.1-01 shows the location of the four large shots. In 1967, the Fremantle Region Upper Mantle Project (FRUMP) investigated the crust in the southwestern region of Western Australia (Finlayson, 2010; Appendix 2-2).

In northern Australia, several experiments were carried out: the CRUMP (Carpentaria Region Upper Mantle Project) experiment with offshore shots in 1966, the WRAMP (Warramunga Seismic Array Mantle Projects) experiment with some inland shots in 1966, and the ORD RIVER experiment with a series of shots in 1970–1971. Data recorded in northern and western Australia are relatively scarce but indicated crustal thicknesses of around 40+ km within the Australian continent and a gradual thinning toward the coastal areas. There appeared a tendency for a systematic increase of P_n velocities across Australia from east to west. There was also good evidence for the existence of a Conrad discontinuity at ~20 km depth (Denham et al., 1972; Cleary, 1973).

A historic overview on early seismic-reflection work in Australia was published by Dooley and Moss (1988) and Moss and Dooley (1988). In southeast Australia in November 1960, strong reflections from the Moho were recorded in the Murray Basin during extended recording time experiments. The first attempt to record continuous deep-reflection profiles was in 1960 over the Comet Ridge, Queensland. Records were obtained on 800 m split

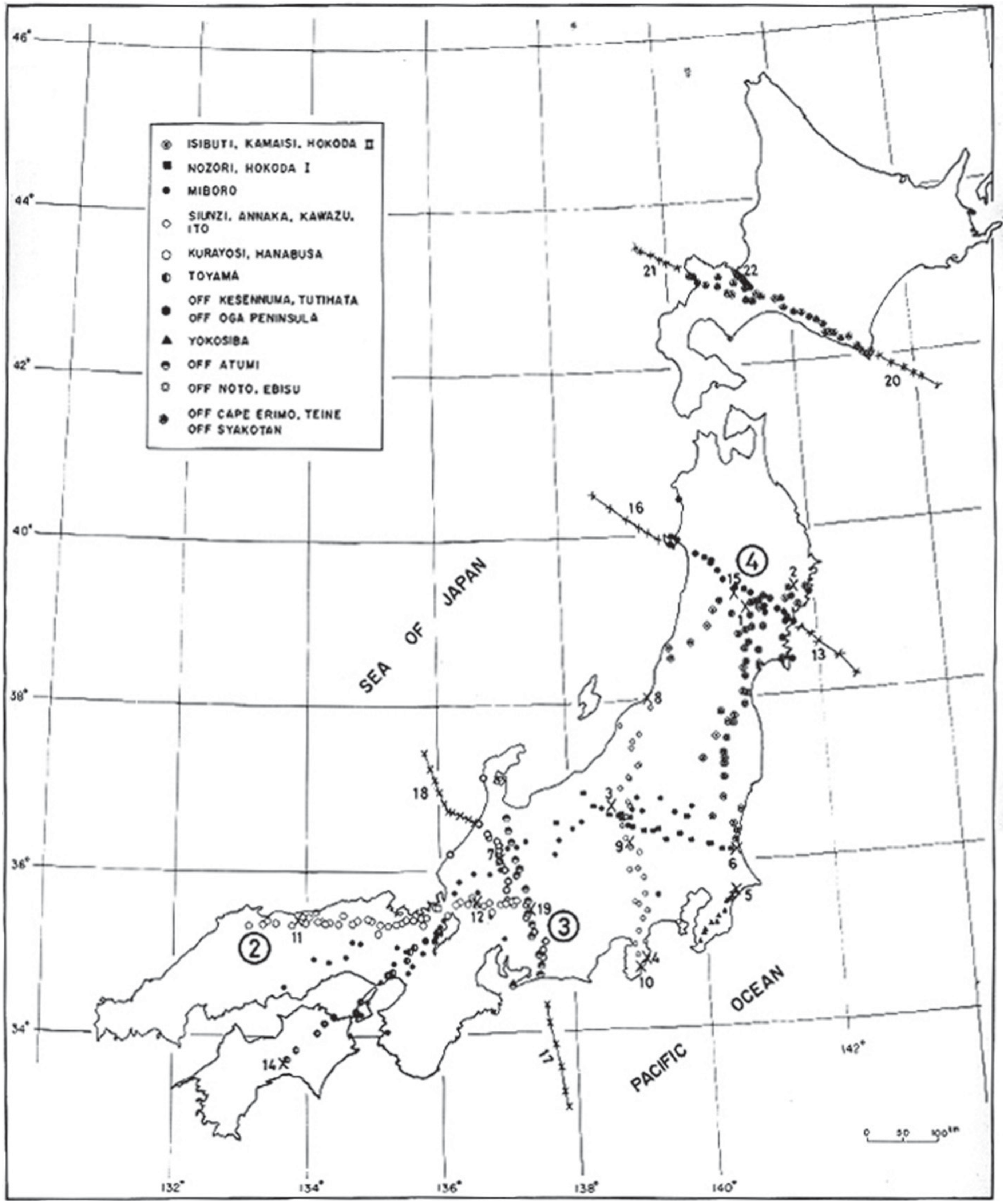


Figure 6.6-01. Shot and observation points in Japan from 1950 to 1970 (from Research Group for Explosion Seismology, 1973, fig. 1). [Tectonophysics, v. 20, p. 129–135. Copyright Elsevier.]

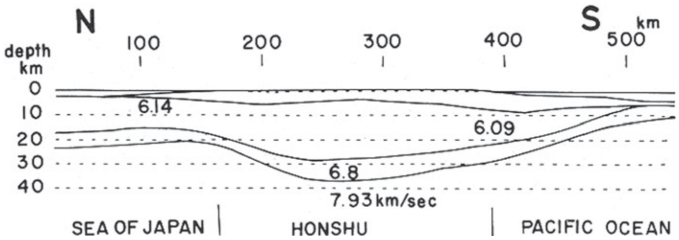


Figure 6.6-02. Crustal structure along the profile Atumi–Noto (line 3 in Fig. 6.6-01), central Japan (from Research Group for Explosion Seismology, 1973, fig. 3). [Tectonophysics, v. 20, p. 129–135. Copyright Elsevier.]

spreads along a 5.2 km traverse. Similar attempts were made at Bushbrook in the Perth Basin, Western Australia, in 1964, where recordings were made on 1100 m spreads along a 6.6 km traverse. Other experimental and offset recordings were made in the Roma shelf area of the Surat Basin, Queensland, in 1967. All data showed events between 5 and 12 s, but no complete continuity. Following this earlier work, seismic profiling trials near Mildura, Victoria, and Broken Hill, New South Wales, in 1968 and 1969 demonstrated convincingly that the Moho could be clearly imaged at ~10.2 s TWT using reflection profiling techniques (Branson et al., 1976). Additional tests were carried out in 1968 and 1969 in hard-rock areas at Tidbinbilla, Australian Capital Territory, and Braidwood, New South Wales, to complement the Mildura and Broken Hill experiments. Both large and small charges gave good results. During 1968, seismic-reflection profiling across the southern margin of the Ngalia Basin in central Australia imaged clearly the dipping boundary fault and underlying sedimentary sequences beneath overlying Precambrian basement rocks, thus clearly demonstrating the viability of seismic imaging in hard rock terrains. Data of good quality were also obtained in 1969 in the northern part of the Amadeus Basin, Northern Territory, using short cross section, a cross-spread, and a partial expanding spread (Brown, 1970).

6.7.2. Africa and the Afro-Arabian Rift System

While crustal research in Australia using explosion seismology was quite substantial already in the 1960s due to the existence of a small but very active geophysical community, crustal structure investigations in the two other continents of the southern hemisphere, Africa and South America, was mainly limited to earthquake-related research. Active-seismic projects required much more sophisticated logistics and were therefore very difficult to handle with the equipment available at that time.

The Afro-Arabian rift, extending from Lebanon through the Jordan–Dead Sea transform, the Red Sea, the Gulf of Aden, the Afar triangle of Ethiopia into the East African rift system was the goal of first seismic investigations already in the 1960s. Marine studies investigated the Red Sea (Drake and Girdler, 1964; Tramontini and Davies, 1969) and the Gulf of Aden (Laughton and Tramontini, 1970).

Only two seismic-refraction surveys were made in the Red Sea. The first one was a survey in 1958 which was reported upon by Drake and Girdler (1964), and the second survey was done in

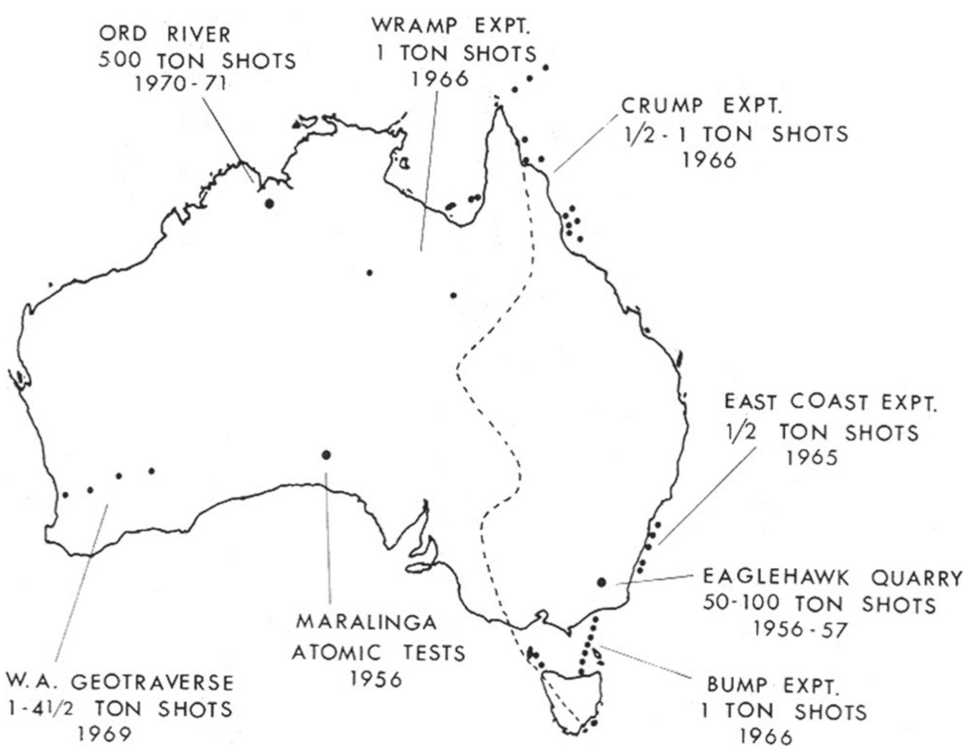


Figure 6.7.1-01. Seismic refraction experiments in Australia (from Cleary, 1973, fig. 1). The dashed line indicates the eastern limit of exposed Precambrian rocks. [Tectonophysics, v. 20, p. 241–248. Copyright Elsevier.]

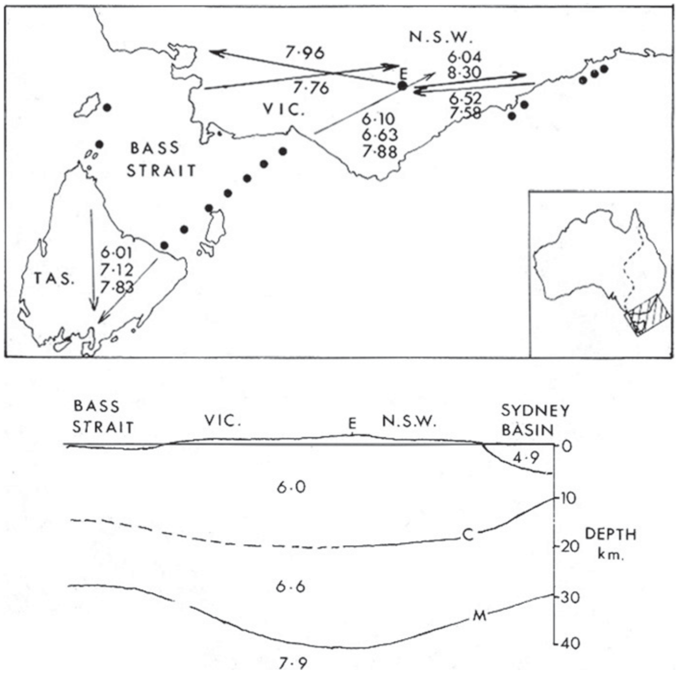


Figure 6.7.1-02. Top: Location of seismic-refraction experiments in Southeast Australia. Bottom: Crustal cross section through Southeast Australia (from Cleary, 1973, fig. 2). E—position of Eaglehawk quarry. [Tectonophysics, v. 20, p. 241–248. Copyright Elsevier.]

the late 1960s (Tramontini and Davies, 1969). Drake and Girdler had classified rock types attributed to velocities: 1.7–3.0 km/s unconsolidated sediments, 3.0–5.0 km/s evaporites (with possible volcanic material), 5.5–6.4 km/s shield rocks, 6.4–7.3 km/s basic intrusive rocks. The latter was referred to as layer 3, although the velocity range was wider than the average values for the world's oceans. Tramontini and Davies (1969) chose to survey the area at the northern limit of the large magnetic anomalies. They shot 20 seismic lines over the axial trough and the eastern flanks. No mantle velocities were detected at any of the stations. No shield rock was detected in the central trough but, without exception, the trough was found to be underlain by layer 3 material (6.4–7.3 km/s). The depth to the surface of layer 3 was ~4 km below sea level, i.e., 2 km below the ocean floor. Under the main trough, i.e., northern part of the Red Sea, where the axial trough has not yet developed, the depth to layer 3 increased to ~7 km below the sea surface. On the western flanks of the main trough shield rocks were detected and layer 3 was absent. There was some evidence that layer 3 was not only deeper on the eastern flank, but had also a lower velocity.

The seismic-refraction data in the Gulf of Aden showed that the whole of the Gulf of Aden between continental margins is oceanic in nature, leading to the conclusion that the Gulf of Aden had formed as a result of the continental separation of Arabia and Africa, driven by the same subcrustal forces that created the mid-ocean ridge systems. The crustal columns derived for the Gulf of Aden are also shown by Laughton et al. (1970, fig. 7) in their summary paper on the Indian Ocean and were later republished in a review of crustal studies along the Afro-Arabian rift system (Prodehl et al., 1997a). From the marine seismic-refraction and other geophysical studies in the Red Sea (Tramontini and Davies, 1969) the association with the Gulf of Aden was clear, but the opinion was divided on the amount of continental separation involved in the formation of the Red Sea.

In 1969, the University of Birmingham undertook the first seismic-refraction experiment in the East African rift (Griffiths et al., 1971; Long et al., 1973). With shots fired in Lake Turkana and in Lake Baringo recorded by 10 stations, a reversed line of 300 km length was established along the Gregory rift. Due to their limited number, however, the observations were contradictory (Long et al., 1973) and only fully understood when, 16–25 years later, the KRISP projects (Kenya Rift International Seismic Project) of 1985, 1990, and 1994 investigated the rift in much more detail (see Chapter 9; Prodehl et al., 1994a; Fuchs et al., 1997).

6.7.3. South America

The very first explosion seismic investigation in South America concentrated on the Andes (Fig. 6.7.3-01). In 1968, a reconnaissance survey of the Peru-Bolivia Altiplano involved recording distances of 320–400 km (Figs. 6.7.3-02 and 6.7.3-03). However, no P_n-first arrivals were obtained. From weak reflections, a total crustal thickness of more than 70 km but less than 80 km was estimated (Ocola and Meyer, 1972; Appendix A6-12).

6.7.4. Antarctica

Several Antarctica marine surveys were carried out in the 1960s and 1970s by Don Griffiths' group of the Birmingham University (Allen, 1966; Ashcrost, 1970; Harrington et al., 1972) investigating crust of the Scotia Sea and adjacent Antarctica peninsula region. It was either a two ship or a single ship–custom buoys operation using explosives as sources.

As was mentioned in Chapter 5, during a reconnaissance exploration of the Antarctic interior as part of the U.S. Antarctic research program during the 1950s and 1960s, some three dozen seismic-refraction profiles had been recorded which, however, provided only information about seismic velocities below the ice, but did not penetrate further than into the topmost upper crust (Bentley, 1973).

The very first time explosion seismic research on land in Antarctica reached through the ice into deeper levels of the crust, was in 1969, when two deep seismic sounding profiles were completed by members of the 14th Soviet Antarctic Expedition (Bentley, 1973; Kogan, 1972) in the vicinity of Novolazarevskaya near the northern coast of Queen Maud Land (Fig. 6.7.4-01). The interpretation resulted in an average thickness of the crust of 40 km, with a few kilometers thicker near the mountains and shallowing to only 27 km below sea level near the coast. The measured uppermost mantle velocity was 7.9 km/s.

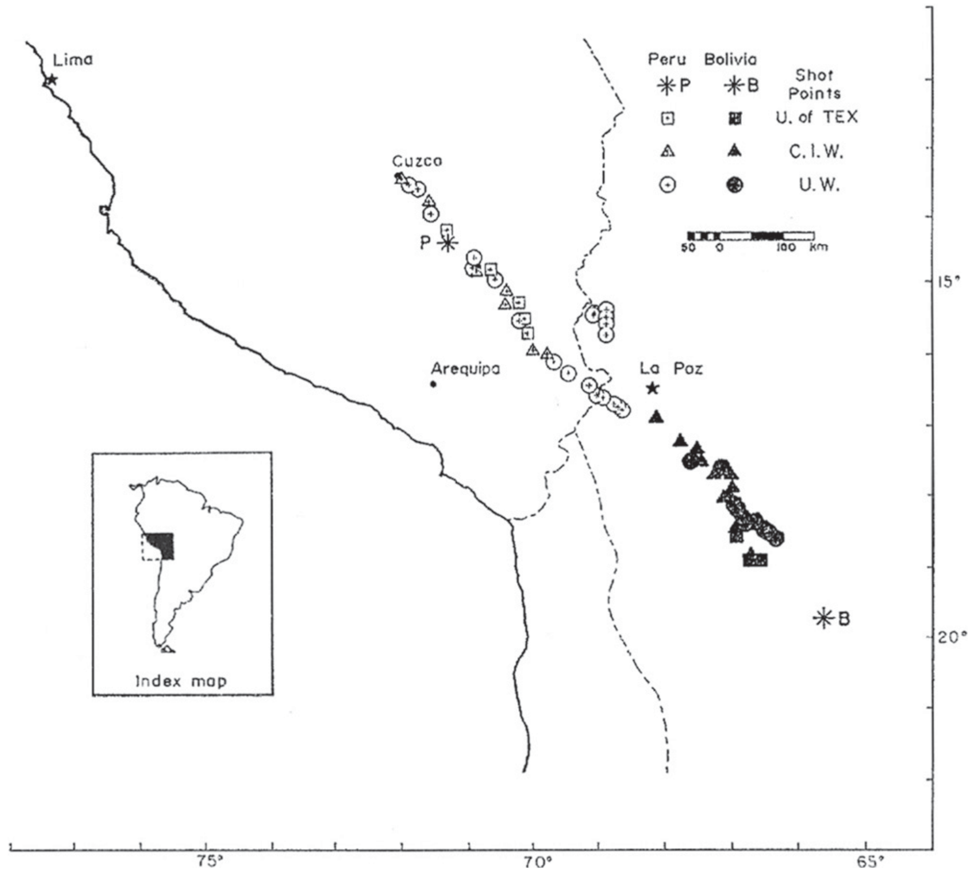


Figure 6.7.3-01. Seismic refraction experiment of 1968 in Peru and Bolivia (from Ocola and Meyer, 1972, fig. 1). [Geophysical Journal of the Royal Astronomical Society, v. 30, p. 199–209. Copyright John Wiley & Sons Ltd.]

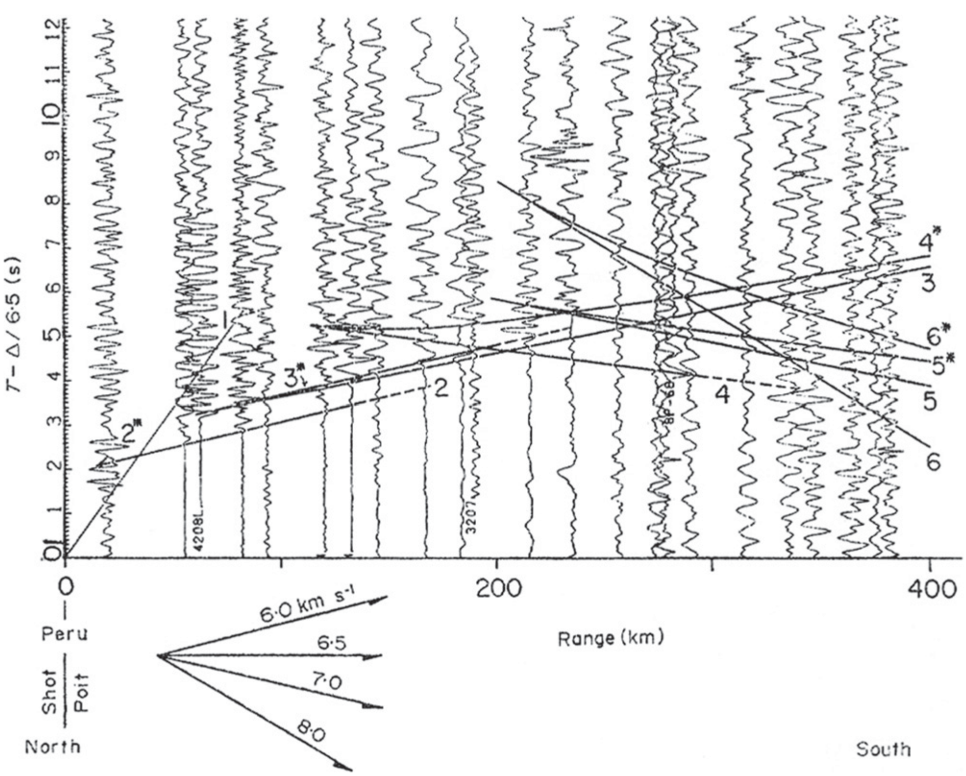


Figure 6.7.3-02. Record section for the 1968 Peru profile (from Ocola and Meyer, 1972, fig. 2). [Geophysical Journal of the Royal Astronomical Society, v. 30, p. 199–209. Copyright John Wiley & Sons Ltd.]

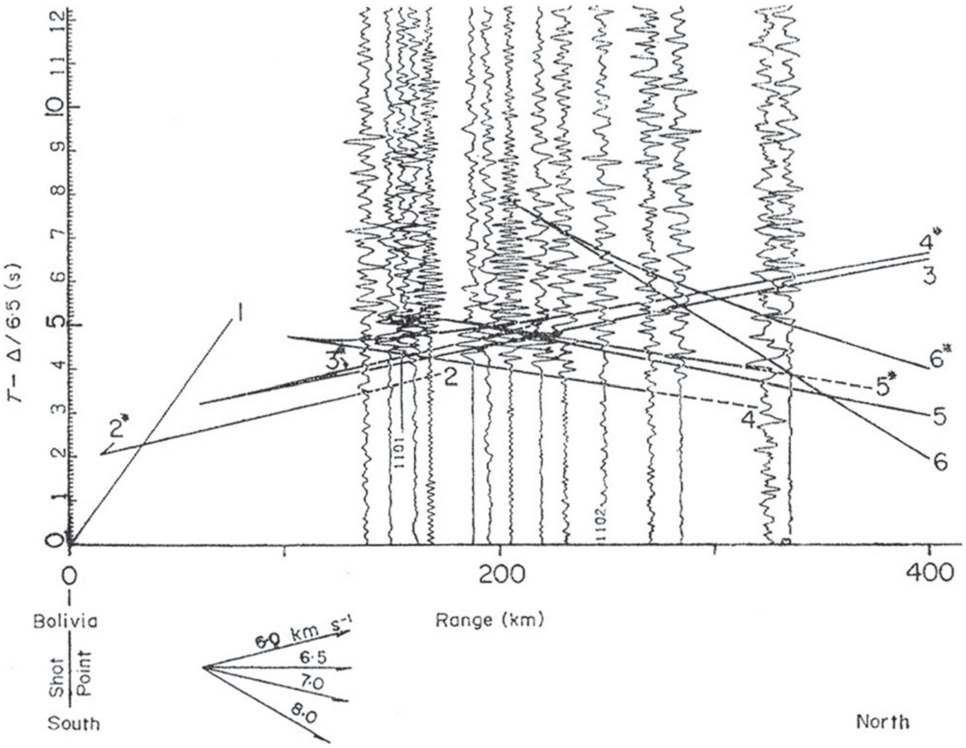


Figure 6.7.3-03. Record section for the 1968 Bolivia profile (from Ocola and Meyer, 1972, fig. 4). [Geophysical Journal of the Royal Astronomical Society, v. 30, p. 199–209. Copyright John Wiley & Sons Ltd.]

6.8. SEISMIC-REFRACTION EXPLORATIONS OF THE CRUST BENEATH THE OCEANS IN THE 1960s

6.8.1. Introduction

Following the reviews of Jones (1999) and Minshull (2002), the basic structure of the oceanic crust had been established and published in the comprehensive Volume 3 of *The Sea* (Hill, 1963a), before plate tectonics was introduced after the discovery of seafloor spreading (Hill, 1963a). In 1970, a second comprehensive publication followed, published as Volume 4 of *The Sea* and consisting of two voluminous parts edited by A.E. Maxwell: *The Sea, New Concepts of Sea Floor Evolution*, which comprised all new results assembled during the 1960s. Until the beginning of the 1960s, the averaging seismic-refraction method was applied which did not allow the measurement of local variations in sediment thickness or details of its stratification. The data were, however, of sufficient quality that they could be reinterpreted in the 1970s when the reflectivity method became available and synthetic seismograms could be computed (Orcutt et al., 1976; Spudich and Orcutt, 1980).

The 1960s were the decade when plate tectonics was born, and knowledge of the properties of the oceanic crust was critical. The proper interpretation of marine magnetic anomalies came at this time (Menard, 1964). The Lamont ships had surveyed a vast amount of profiles in all oceans (Ewing and Ewing, 1970, fig. 1), and the Scripps Institution also had accomplished extensive work. Other nations were active as well, for example, Canada (e.g., Dainty et al., 1966), Japan (e.g., Den et al., 1969; Murauchi et al., 1968; Okada et al., 1973), the former USSR (e.g., Galperin and Kosminskaya, 1964; Neprochnov et al., 1967; Neprochnov et al., 1970; Zverev and Tulina, 1973), and Germany. The German research vessel *Meteor* (1964) was the second research vessel carrying the name *Meteor*.

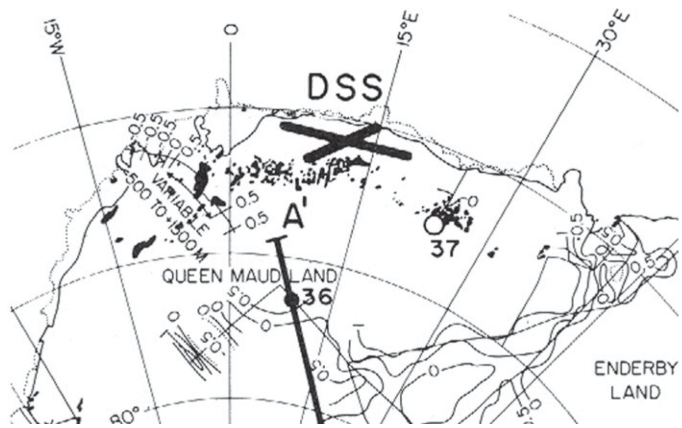


Figure 6.7.4-01. Location of the Soviet Deep Seismic Sounding (DSS) experiment in Antarctica (from Bentley, 1973, fig. 1). The line from A′ southward was part of shallow seismic investigations not discussed here. [Tectonophysics, v. 20, p. 229–240. Copyright Elsevier.]

Already between 1924 and 1939 the first *Meteor* (1924) had explored the oceans and had contributed to the detection of the existence of mid-ocean ridges. The second *Meteor* (1964) concentrated its research expeditions on the northeastern half of the Atlantic Ocean, the Mediterranean, and the Red Sea. Between 1964 and

1985, it undertook 73 expeditions and covered 650,000 nautical miles; at least 12 expeditions included major seismic investigations of the oceanic crust (Bröckel, 1994; see also http://www.dfg-ozean.de/berichte/fs_meteor, report M0; Sarnthein et al., 2008).

As was pointed out in Chapter 5, all instruments in use in the 1950s and 1960s 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. Techniques and equipment for continuous profiling were developed early in the 1960s and successfully applied in the following years, enabling studies of local variations in sediment thickness and details of its stratification. Ewing and Ewing (1970) have described in detail the instrumentation that was used in all their excursions to obtain detailed information on the sedimentary structure. Their figure 1 shows the seismic profiler traverses made by Lamont ships and personnel through 1967 (*in* Maxwell, A.E., ed., The Sea, Volume 4, New Concepts of Ocean Floor Evolution: New York, Wiley-Interscience, p. 53–84). Work on the application of ocean-bottom seismographs was also an important research aim (e.g., Neprochnov et al., 1968).

The use of internally recording sonobuoys allowed seismic-refraction experiments to conduct from a single ship. Since the late 1960s, low-cost expendable military sonobuoys for launching from ships became available for shooting wide-angle profiles. On entering the water, VHF transmitter aerials and hydrophones were deployed and salt-water batteries actuated (Jones, 1999). One or more hydrophones were suspended from the buoy to detect the seismic head waves which were then transmitted back to the firing ship, timed and recorded. The buoys were designed to sink after ~10 hours.

To overcome eventually the work with explosives, continuous research tried to develop efficient nonexplosive energy sources. High-voltage sparkers, boomers, gas exploders, and magnetostrictive transducers (Ewing and Zaunere, 1964), proved useful for shallow research. Working since 1961 on several types of pneumatic sources, Ewing and Zaunere finally found a pneumatic gun which worked highly satisfactorily in continual operation for several months and produced excellent profiler data both in shallow and deep ocean provinces. In connection with a towed receiving array of eight detectors, the pneumatic gun produced sufficient energy to penetrate deep-sea sediments of over a thousand meters while the ship proceeded with a speed at more than 10 knots. Ewing and Zaunere (1964) predicted that this gun could produce sufficient energy to penetrate the total sedimentary section in all deep ocean areas. Hinz (1969), for example, described in detail his experiences of seismic-reflection work around the Great Seamount in the North Atlantic Ocean using a pneumatic sound source. It would only be in the 1990s, however, that for deep crustal and upper-mantle research at sea, airguns became the almost exclusive powerful and reliable sources and finally replaced explosives almost completely.

The analysis methods, available in the 1950s and also during the 1960s (Ewing, 1963a), used least-squares slope-intersect solutions for picked first arrivals, but did not allow for velocity gradients. The igneous crust had been found to be subdivided into a 2–3-km-thick upper layer 2 and a 3–5-km-thick lower layer 3, while the overlying sediments were labeled as layer 1. The two-layer structure of the ocean basins which had been deduced from the many marine expeditions in the late 1940s and 1950s was basically confirmed in the 1960s.

A first approach for a changed view of crustal structure assuming transitional rather than sharp boundaries between layers was proposed by Helmberger (1968). With an example of marine data obtained in the Bering Sea, he developed the theory to treat discontinuities as layered transition zones and calculated corresponding synthetic seismograms.

Until 1967, the oceans and their side branches, such as the Mediterranean Sea or the Gulf of Mexico, had been covered by cruises of the Lamont ships (Ewing and Ewing, 1970, fig. 1), and an enormous wealth of deep crustal seismic data had been obtained as was summarized in great detail in eight summary volumes of *The Sea*, of which Volumes 3 (Hill, 1963a) and 4 (Maxwell, 1970) specialized in seismic research. Introductions to the results of the 1960s in Volume 4 were given by Ewing and Ewing (1970) for the reflection method and Ludwig et al. (1970) for the refraction method. On all cruises aiming for deep seismic research of the whole oceanic crust, for which the refraction seismics was the only available tool at that time, detailed reflection surveys were always a major component. So, a very detailed picture on the sedimentary structure overlying the basement resulted from these measurements. In their sediment map of the North Pacific, Ewing and Ewing (1970, fig. 5) have shown an impressive summary what was known by 1967 on the thickness of the sediments for example in the northern Pacific Ocean.

A few ocean-floor samples had become available during the 1960s and confirmed the suggestion of Cann (1970) on the possible composition of the oceanic crust. Layer 2 consisted of basaltic lavas and their feeder dikes which were unmetamorphosed in their upper parts. The underlying layer 3 consisted of dense dike swarms of basic composition and massive plutonic gabbros. The two-layer crust was underlain by mantle peridotites. In places, serpentinite diapirs from the upper mantle would cut through the oceanic basement. Such a structure also conformed with land-based evidence from ophiolite sequences which had been interpreted as uplifted sections of oceanic crust. An example of such a structure was the Troodos Massif in Cyprus, which had been studied by Vine and Moores (1972).

6.8.2. The Atlantic Ocean

Searching through the literature for seismic-refraction work in the 1960s, there are not very many publications which report on new marine investigations in the Atlantic Ocean and adjacent seas, such as the Gulf of Mexico and Caribbean Sea. It appears that after the enormous effort of the 1950s, the researchers of the large laboratories in North America, which concentrated on deep sea research, regarded the experimental work in the Atlantic Ocean as being "done." Rather, the experimental activities in the 1960s shifted into the Pacific and Indian Oceans. Based on hundreds of kilometers of traverse recorded in the Atlantic Ocean in the 1950s, Ewing (1969) published a generalized structural section from the North American continental shelf to the Mid-Atlantic Ridge (Fig. 6.8.2-01). Ewing and Ewing (1970) have compiled a map showing all seismic profiler traverses accomplished by Lamont ships around the world until 1967.

Ewing (1969) described the results of numerous individual investigations and summarized them as follows: The main crustal layer, named oceanic layer or layer 3, is very uniform in wave velocity and thickness. It is ~5 km thick and has velocities between 6.5 and 7.1 km/s, while the velocities of the basement layer or layer 2 are between 4.5 and 5.5 km/s. This layer 2 was often not detected when thick sediments accumulated above. Velocity histograms showed that 80% of mantle velocities fell between 7.7 and 8.3 km/s, but no obvious relationship between upper-mantle velocity and depth or thickness of overlying crustal material could be detected. However, an inverse relationship did appear to exist between mantle depth and water depth. On average, the depth from the water surface to the Moho is 12 km, but significant deviations from this average were found on the flanks of the Mid-Atlantic Ridge. In the crestal zone the Moho disappears as a distinct boundary and mantle velocities were found at shallow depths of 9–10 km, and a thick layer of intermediate velocity exists instead of normal oceanic crust and upper mantle.

A brief review on crustal structure research of the Gulf of Mexico until the end of the 1960s was published by Hales (1973) based on profiles observed in the 1950s (see section 5.6; Ewing et al.,1960; Antoine and Ewing, 1963) and a profile recorded by the Texas A&M University in 1965 (Fig. 6.8.2-02).

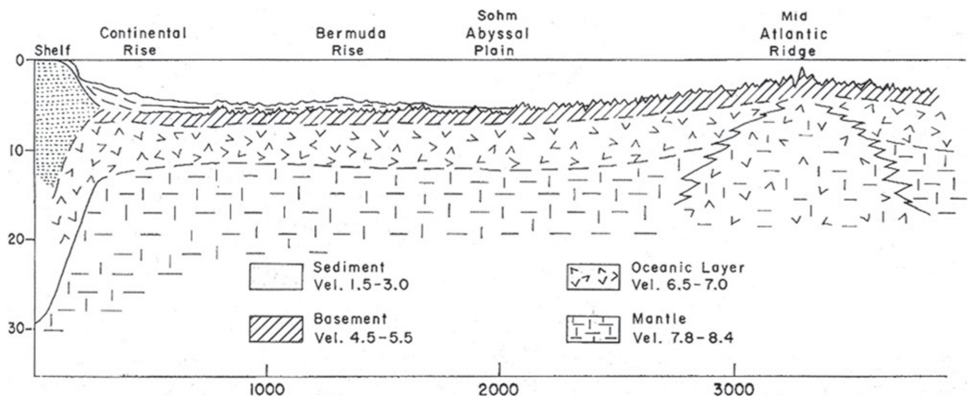


Figure 6.8.2-01. Generalized structural section from the North American continental shelf to the Mid-Atlantic ridge (from Ewing, 1969, fig.1). Horizontal and vertical scales in km, velocities in km/s. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 220–225. Reproduced by permission of American Geophysical Union.]

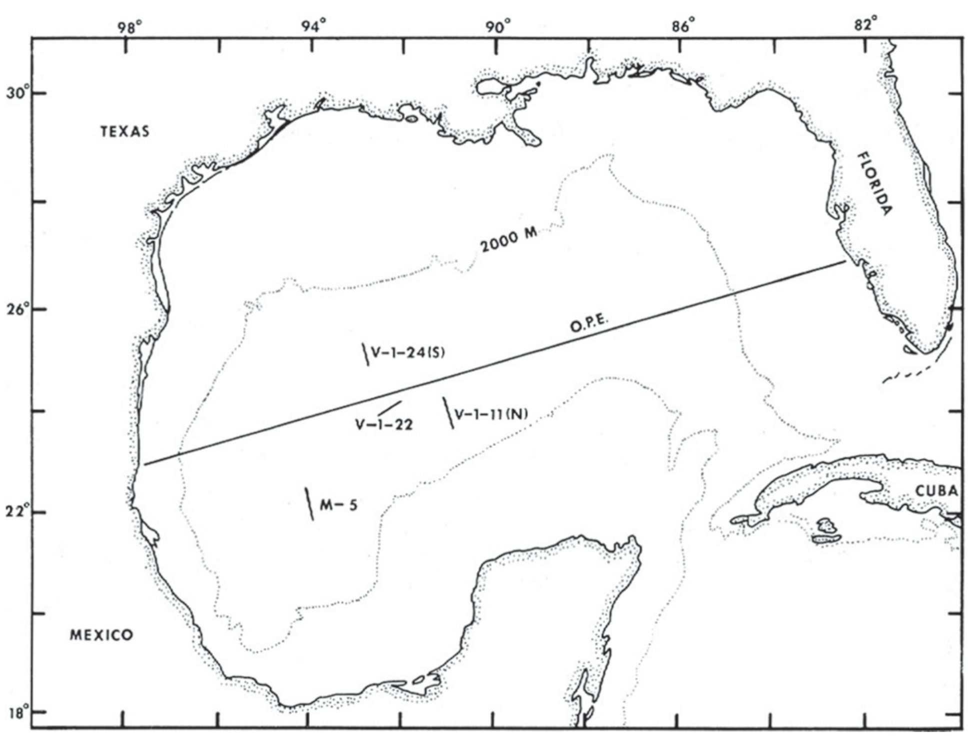


Figure 6.8.2-02. Location of seismic refraction profiles in the Gulf of Mexico (from Hales, 1973, fig. 1). *V-1* lines recorded in the fifties, *M-5* line of 1965, *O.P.E.* large-scale seismic refraction line of 1969. [Tectonophysics, v. 20, p. 217–225. Copyright Elsevier.]

A key result was that the crust underneath the deep part of the Gulf was similar to that of typical ocean basins. Depth to Moho beneath sea level on the crustal profiles shown in Figure 6.8.2-02 was between 16 and 19 km (Hales, 1973). Also from the interpretation of the observations in Mexico from the long line of shooting (O.P.E. in Fig. 6.8.2-02), Hales et al. (1970) concluded that only a typical oceanic crust in both the eastern and western deep basins of the Gulf could satisfy the observations. In addition, the long-range observations of the 1500-km-long O.P.E. line showed a phase beyond 1200 km distance (Fig. 6.8.2-03) which indicated a seismic discontinuity in the mantle at 57 km depth (Hales et al., 1970; Hales, 1973).

Other new activities in the Atlantic Ocean focused less on the general structure of the deep ocean basins than on anomalous features. In Germany in 1964, the second research vessel named *Meteor* (1964) had started its service. The expedition M04 of 1966 investigated the Reykjanes Ridge. In 1967, German researchers undertook a reconnaissance survey (expedition M09) with the German vessel *Meteor* (Aric et al., 1970; Closs et al., 1968, 1969b; Hinz, 1969) to the chain of seamounts (Meteor, Cruiser, Plato, and Atlantis), located at 28°–30°W, 30°–34°N to the west of the triple junction Mid-Atlantic Ridge and the Azores–Gibraltar zone. The area was revisited with the third *Meteor* (1986) in 1990 (Weigel and Grevemeyer, 1999, see Chapter 9.8.2). In 1969, the expedition M17 investigated the Mediterranean and the northern Arabian Sea (Closs et al., 1969a, 1969c). The last expedition of 1969 with a seismic component, M18, was active in the Norwegian Sea (Closs, 1972).

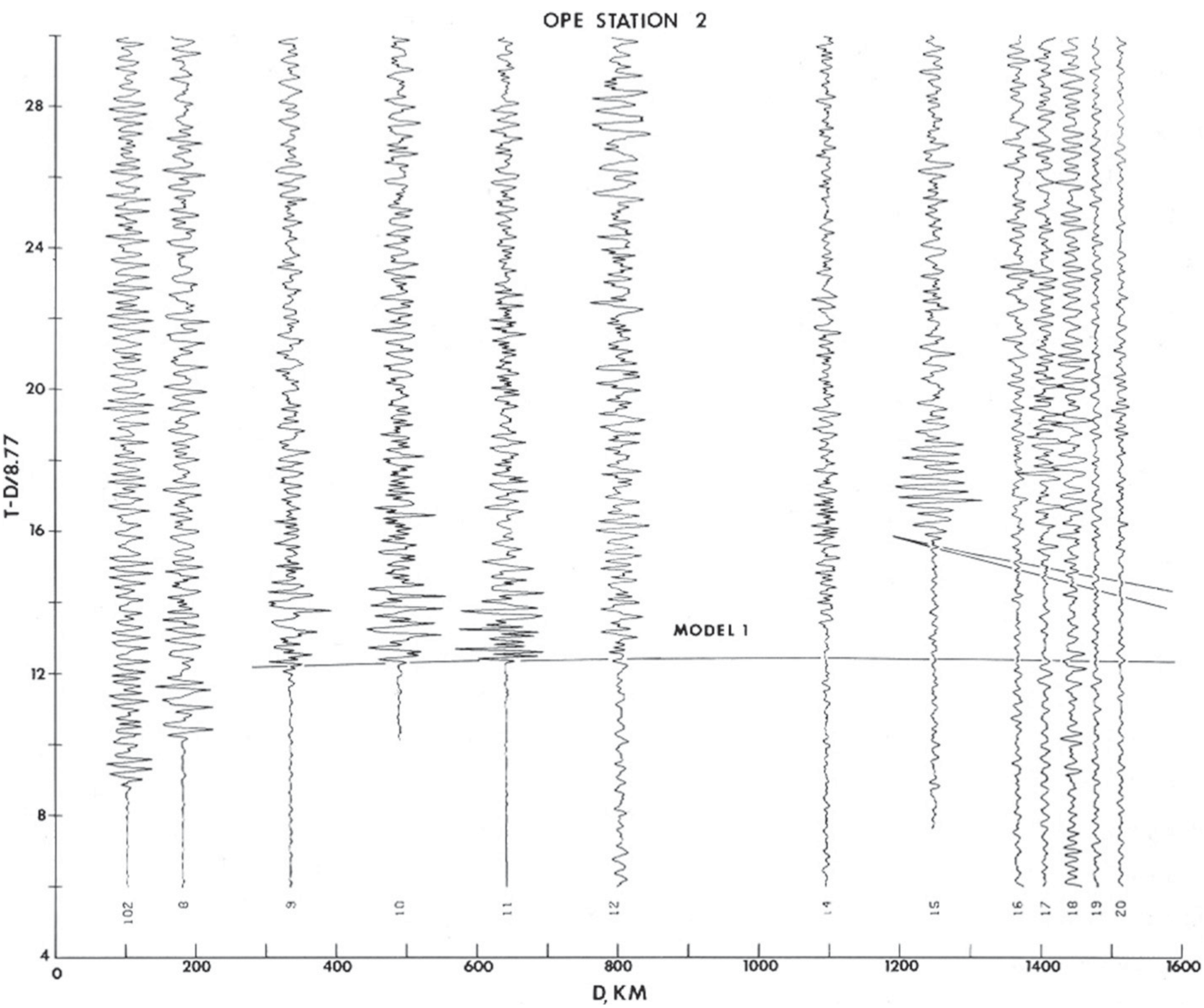


Figure 6.8.2-03. Record section of observations along the *O.P.E.* large-scale seismic refraction line of 1969 in the Gulf of Mexico (from Hales, 1973, fig. 4). [Tectonophysics, v. 20, p. 217–225. Copyright Elsevier.]

In 1968, Keen and Tramontini (1970) shot a seismic-refraction experiment on the western flank of the Mid-Atlantic Ridge between 45° and 46°N. The object was to obtain a reliable determination of the crustal structure near the axis of the ridge. Two ships, one shooting and one receiving, and six sonobuoys, whose positions could be made almost stationary, were used. The seismic arrivals observed by the sonobuoys were transmitted to the recording vessel which also towed a hydrophone array. Charges of up to 150 kg were fired at intervals of 5 or 10 minutes. In total, four profiles were completed The result was a "normal" oceanic crust yielding velocities of 4.6, 6.6, and 8.1 km/s for the first P-wave arrivals within ~30 km of the rift axis, but at 10 km from the western edge of the median valley, the first arrivals resulted in a 7.5 km/s velocity. Moho occurred at a mean depth of 7.5 km.

In 1969, a series of seismic-refraction lines was recorded between Iceland and the Faeroe Islands (Bott et al., 1971). The presence of a 6.8 km/s refractor was seen underneath the Iceland–Faeroe Ridge at a depth of ~7 km and a short unreversed segment with apparent velocity of 7.84 km/s was interpreted as head wave from the Moho at ~16 km depth. The results were later reinterpreted in a different manner (see Chapter 7.8.4; Bott and Gunnarson, 1980).

First surveys also explored the Arctic Ocean (Kutschale, 1966, Hutchins, 1969). Kutschale (1966) describes an experiment of 1962, when an ice island (Arlis II) drifted over a portion of the southern half of the Siberian basin. Amongst other geophysical studies, scientists from the Lamont Geological Observatory of Columbia University also carried out seismic-reflection measurements while aboard Arlis II using special ocean-bottom and ice-surface seismographs. At least 3.5 km of stratified sediments were found to overlie the crust, composed of basement and presumably the "oceanic layer," 6–8 km south of the Alpha ridge and thickening to ~22 km below the ridge.

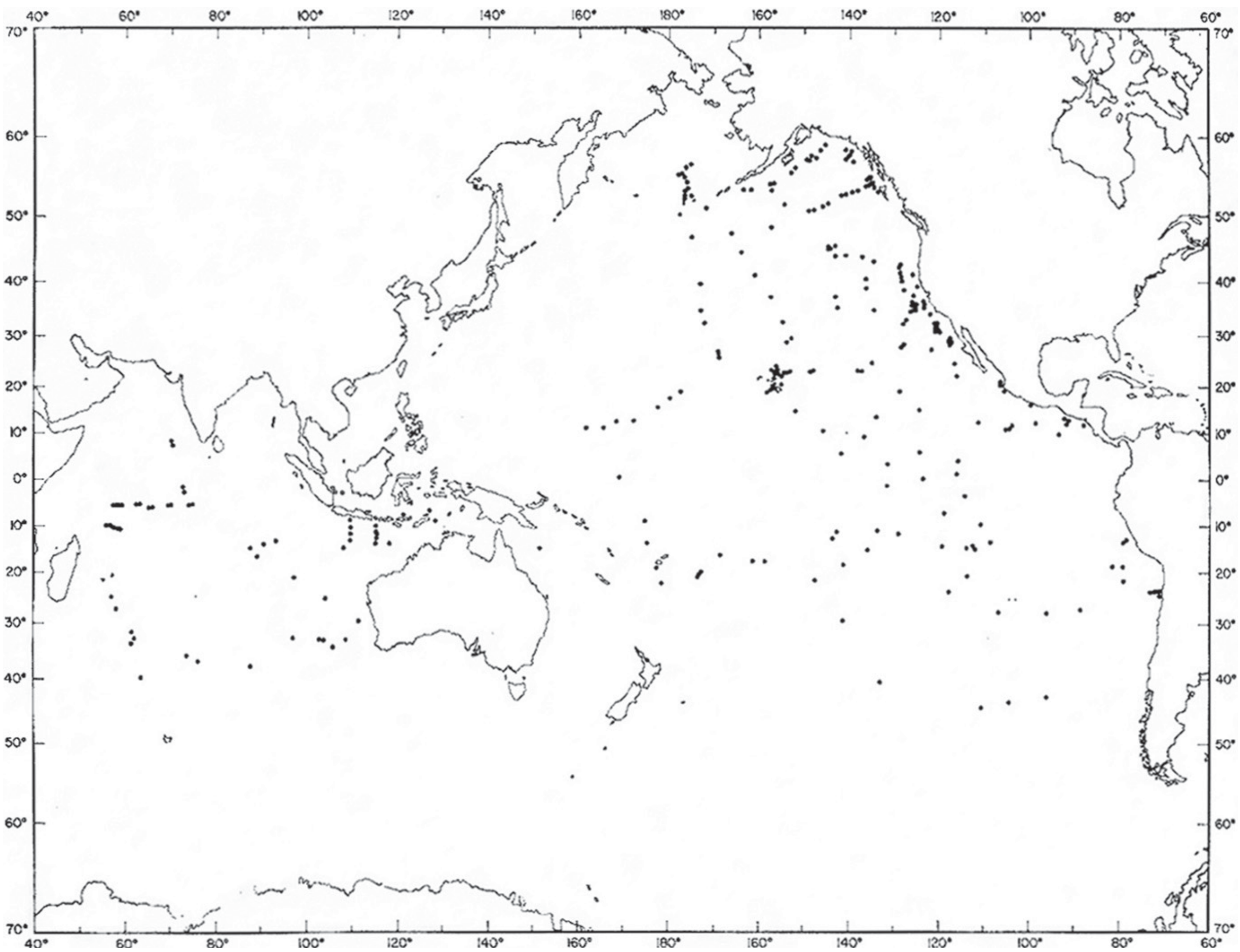


Figure 6.8.3-01. Location of Scripps refraction stations in water depths greater than 2000 m (from Shor and Raitt, 1969, fig. 1). [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 225–230. Reproduced by permission of American Geophysical Union.]

6.8.3. The Pacific Ocean

Shor and Raitt (1969) reviewed the results in the Pacific and Indian Oceans. Since the time when Ewing (1963b) and Raitt (1963) published summaries on crustal and upper-mantle velocities obtained from seismic-refraction data in the oceans, considerable additional data had been obtained. To demonstrate the amount of data collected, the surveys which were performed by the Scripps Institution of Oceanography in the Pacific and Indian Oceans are shown in Figure 6.8.3-01.

In a subsequent summary, Shor et al. (1970) summarized the results of more than 200 individual stations recorded in the Pacific Basin in a table, showing coordinates, azimuth, water column, velocities and thicknesses of sediments, transitional and oceanic crustal layers, mantle velocity, and depth to Moho, and they plotted frequency diagrams on velocities and thicknesses for the whole crust and for individual layers.

The many new data confirmed that the averages published by Raitt (1963) for the oceans continued to be valid (Shor and Raitt, 1969; Shor et al., 1970). The velocity of the sediments varied between 1.5 and 3.4 km/s, with their median thickness at 0.34 km. The velocities in layer 2, the transition, varied between 3.4 and 6.0 km/s, with a median of 1.21 km in thickness and a median velocity of 5.15 km/s. The oceanic layer 3 of 4.57 km average thickness showed a median velocity of 6.82 km/s, and the median mantle velocity was 8.15 km/s.

Consistent regional variations were observed for sediments where turbidite flows can be received from continents. Mantle velocities and depths were frequently less than normal along the mid-ocean rift system, and mantle depths appeared to become greater than normal beneath trenches and near the foot of the continental slope where no trenches were visible.

Shor et al. (1968) reported on a series of profiles made in 1965 across the Juan de Fuca Ridge, west of the coast of Oregon. They interpreted these data to mean that the ridge was the surface expression of an upraised oceanic crust and mantle. Near the ridge, the upper-mantle velocity appeared to be below normal and the Moho was found as shallow as 7 km below sea level. Profiles in the Mendocino escarpment in the vicinity of the Gorda Ridge off northern California gave evidence that normal oceanic depths near 11 km of the mantle and high velocities of 8.4–8.5 km/s exist on the south side of the escarpment. They were the same on the north side except near the Gorda Ridge, which seemed to have a similar structure as the Juan de Fuca Ridge.

In 1967, the Hawaii Institute of Geophysics started a program of single-ship seismic wide-angle reflection and refraction measurements aimed primarily at defining sediments and upper crustal layers. A 7000 J sparker was used as the energy source. A year later, in 1968, an airgun source was added to the system. In 1969, with sonobuoys as receivers and a 20-cubic-inch airgun as a repetitive source, a series of ASPER (Airgun and Sonobuoy Precision Echo Recorder) refraction stations were run in waters exceeding 5 km at sites well away from any obvious structural transition or topographic high (Fig. 6.8.3-02).

For the first time, the entire crust could be penetrated and the mantle be reached. In particular, a high-velocity basal crustal layer was detected with velocities between 7.1 and 7.7 km/s at several locations in the Pacific Ocean basin. Sutton et al. (1971)

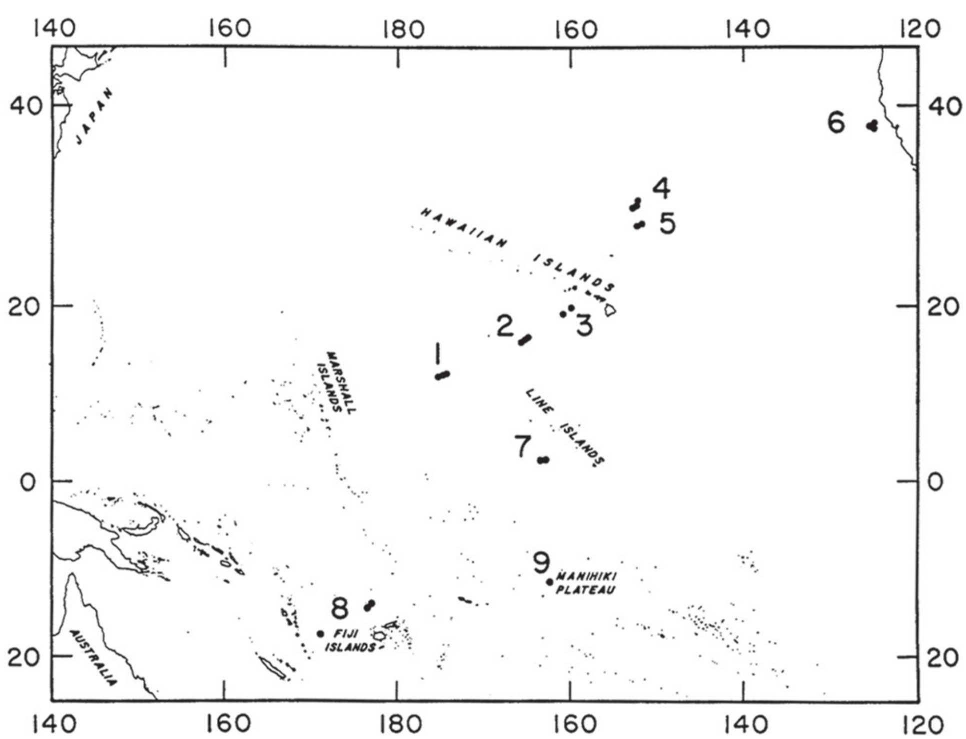


Figure 6.8.3-02. Locations of the ASPER stations conducted in 1969–1970 (from Sutton et al., 1971, fig. 1). The Solomon Islands referred to below and discussed in detail by Furumoto et al. (1973) are located to the NE of no. 8 (Fiji). [*In* Heacock, J.G., ed., 1971, The structure and physical properties of the Earth's crust: American Geophysical Union Geophysical Monograph 14, p. 193–209. Reproduced by permission of American Geophysical Union.]

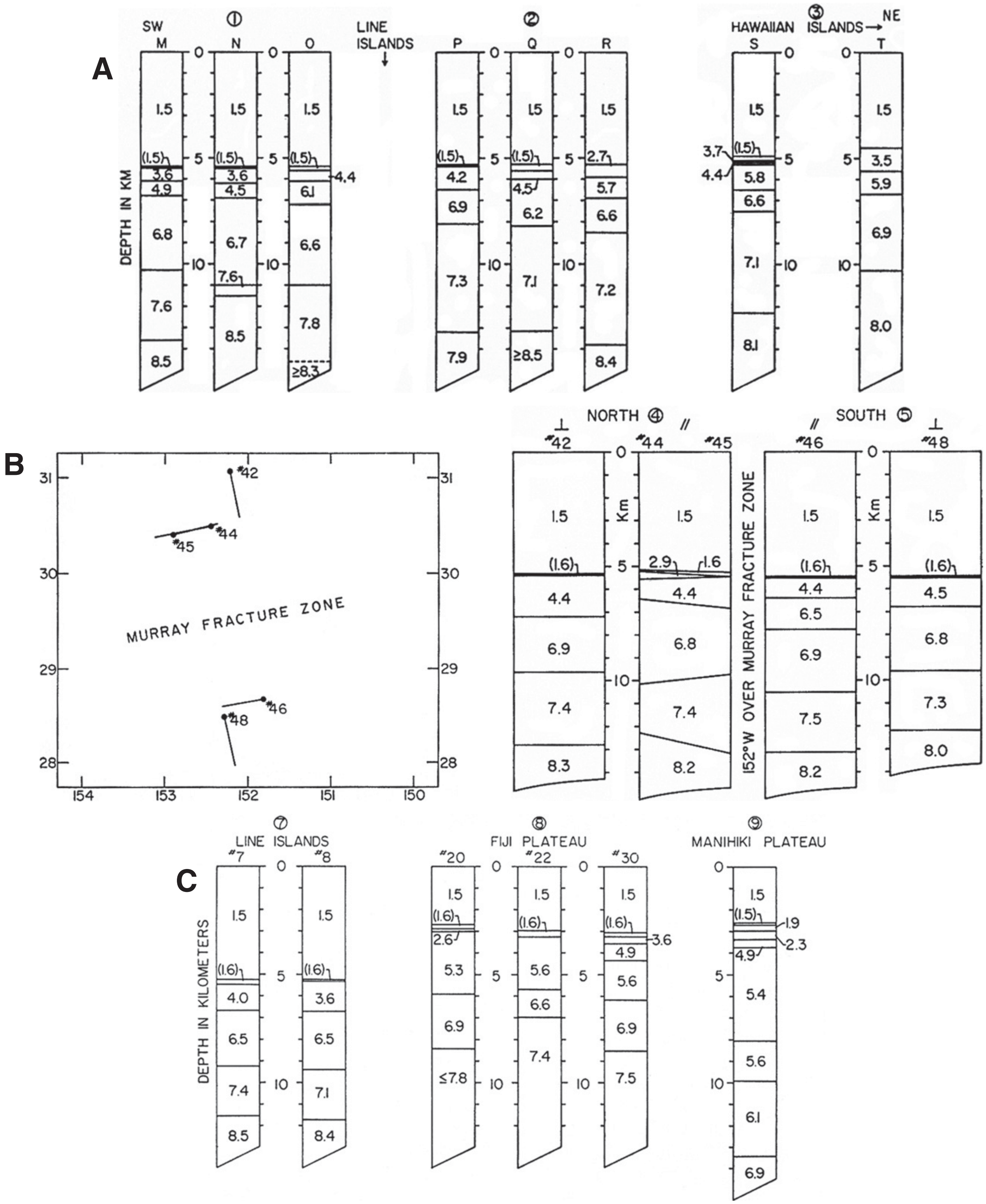


Figure 6.8.3-03. (A) Velocity-depth profiles of the crust in oceanic basin south of Hawaii (from Sutton et al., 1971, fig. 6). Location nos. 1–3 in Fig. 6.8.3-02. (B) Left: Location of ASPER stations on either side of the Murray fracture zone. Right: Velocity-depth profiles of the crust around the Murray fracture zone (from Sutton et al., 1971, figs. 7 and 8). Location nos. 4 and 5 in Fig. 6.8.3-02. (C) Velocity-depth profiles of the crust in oceanic basin south of Hawaii (from Sutton et al., 1971, fig. 10). For location see Fig. 6.8.3-02. [*In* Heacock, J.G., ed., 1971, The structure and physical properties of the Earth's crust: American Geophysical Union Geophysical Monograph 14, p. 193–209. Reproduced by permission of American Geophysical Union.]

could show that this layer, which in other data evidently was masked because it is seen primarily in secondary arrivals, occurred widely in the Pacific Ocean. On all sites, except on the Fiji and Manihiki Plateaus (sites 8 and 9 in Figs. 6.8.3-02 and 6.8.3-03C), the upper mantle velocity underlying the basal crustal layer was near or above 8.0 km/s

In total, Sutton et al. (1971) constructed 21 crustal columns from these measurements. The results of stations 1–3 (Fig. 6.8.3-03A), obtained in 1969, were spaced along a line extending almost over 1000 km over a deep ocean basin between the Marshall Islands in the SW and the Hawaiian Islands in the NE. Other data (Fig. 6.8.3-03B) were obtained, also in 1969, north and south of the Murray Fracture Zone (stations 4 and 5 in Fig. 6.8.3-02). In 1970, more data (Fig. 6.8.3-03C) could be recorded near the Line Islands, on the Fiji Plateau, and on the Manihiki Plateau (stations 7–9 in Fig. 6.8.3-02).

The results of Sutton et al. (1971) were included in the publication of Furumoto et al. (1973), who summarized in particular the many seismic surveys around the Hawaiian archipelago (Figs. 6.8.3-04 and 6.8.3-05), as well as those around northern Melanesia (Figs. 6.8.3-08 and 6.8.3-09). The crustal thickness of the island of Hawaii varies from 12 to 17 km, the crust of Maui and Lanai is ~15 km thick, and at 19–20 km, the crust underneath Oahu is the thickest (Fig. 6.8.3-04). The crust under Kauai and Nihoa is ~17 km thick, but under the smaller islands to the northwest, crustal thickness is on average only 12 km (Fig. 6.8.3-05).

A special crustal survey of the main island of Hawaii was also performed and published by Hill (1969). In the experiment of 1964, shots with charges between 150 and 200 kg were fired at sea within 2 km of the shore at 10-km intervals along each of the three coasts. The shots fired along a particular coastline were recorded by 5 seismic-refraction units spaced at ~25 km intervals along each coast and by 13 short-period seismograph stations of the permanent HVO (Hawaiian Volcano Observatory) seismic network.

Furthermore, in 1965, three 500-ton charges of high explosives were detonated on the ground surface on Kahoolawe Island south of Maui and recorded by the HVO network stations on Hawaii. For the interpretation, the position of shotpoints and recording stations were projected onto straight lines implying that variations in subsurface structure perpendicular to the "profiles" could be corrected by special treatment. Figure 6.8.3-06 shows an example of the record section of a station located at the northern edge of the island of Hawaii (corresponding to position 5 of Fig. 6.8.3-04) and its recordings of shots along the northeastern coast. The interpretation gave an 11–12-km-thick crust along the

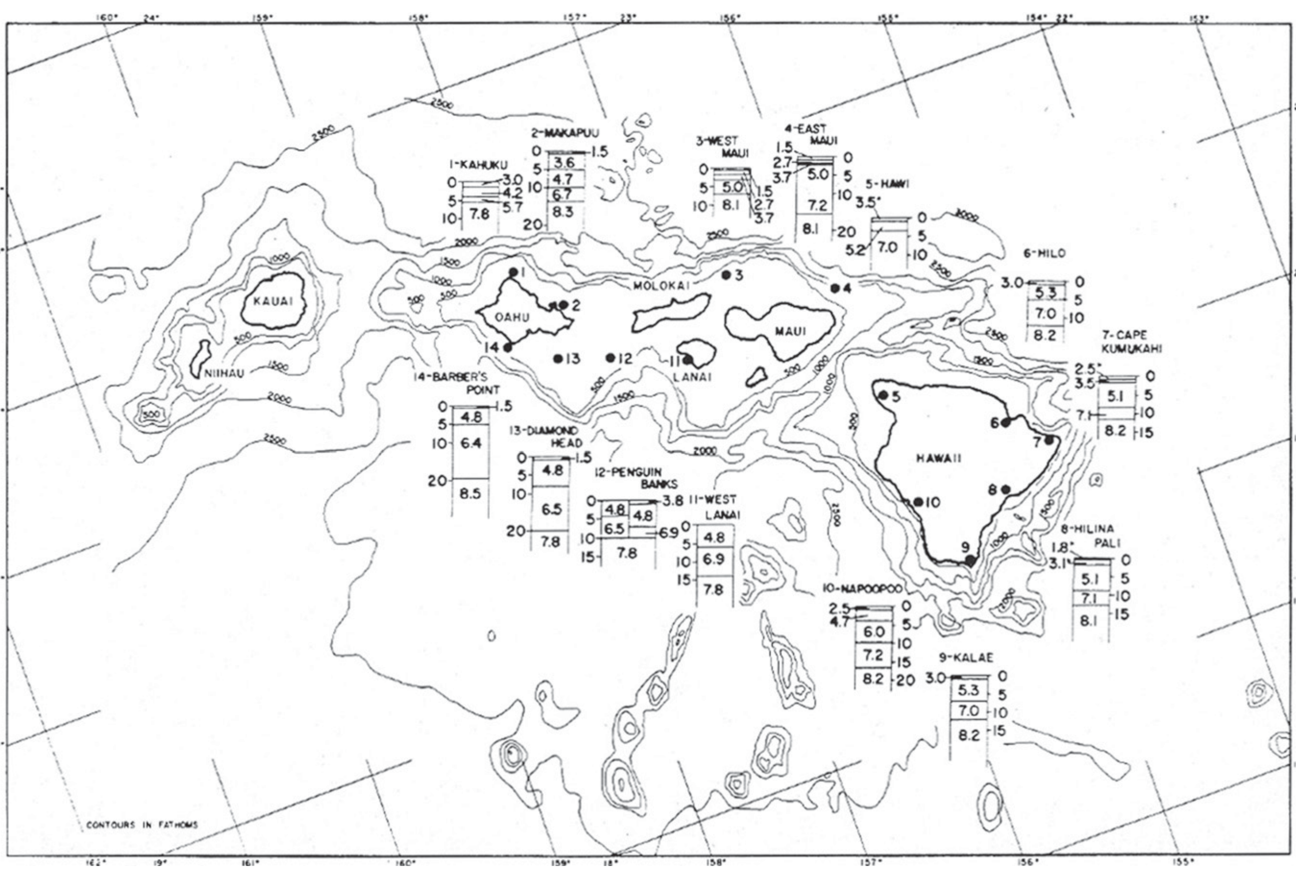


Figure 6.8.3-04. Velocity-depth profiles of the crust for the major Hawaiian islands (from Furumoto, 1973, fig. 2). [Tectonophysics, v. 20, p. 153–164. Copyright Elsevier.]

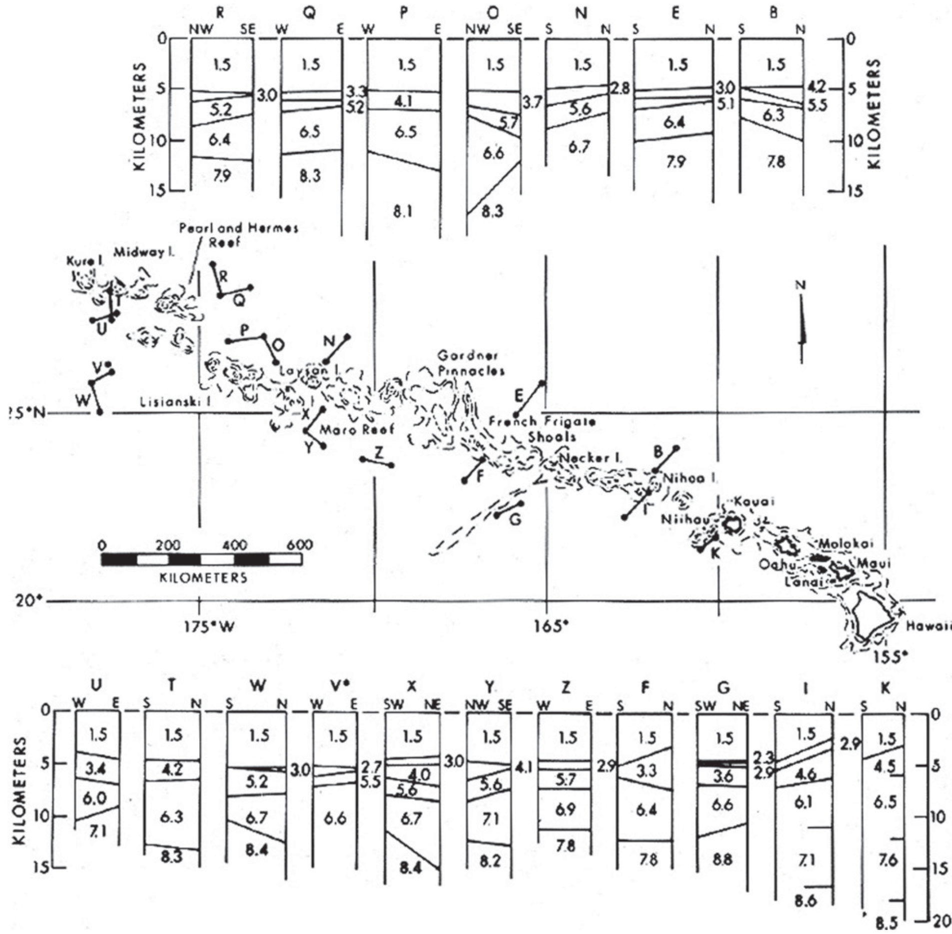


Figure 6.8.3-05. Velocity-depth profiles and locations of the seismic traverses around the Hawaiian archipelago west of the main islands (from Furumoto, 1973, fig. 3). [Tectonophysics, v. 20, p. 153–164. Copyright Elsevier.]

northeast and southwest flanks of the volcano Kilauea, but indicated an increase in thickness from north to south along the west coast from 14 km to ~18 km. The upper-mantle velocity under most of the island was 8.2 km/s.

The Lamont-Doherty Geological Observatory collected a large set of 190 sonobuoy stations seismic data in the northern Pacific, also in 1969 (Houtz et al., 1970), north of latitude 10°S and east of longitude 150°E (Fig. 6.8.3-07). In contrast to those of Sutton et al. (1971), these measurements had not aimed to study the whole crust and reached neither the previously discussed basal layer nor the Moho, but aimed instead for detailed research of the behavior of layer 2, the upper basement layer beneath the sediments only. On average, a layer-2 thickness of 1.11 km resulted with an average velocity of 5.52 km/s, overlying layer 3 with an average velocity of 6.63 km/s.

The Bismarck archipelago is the northwestern part of northern Melanesia and includes the islands of New Britain and New Ireland bordered by the Bismarck Sea to the northwest and the Solomon Sea in the southeast. Sea experiments were carried out here by the Australian Bureau of Mineral Resources, Canberra, in the area of the Bismarck and Solomon Seas near the continental margin north of Australia and New Guinea. In 1967 and 1969, a total of ~100 shots of up to 1 ton of TNT were recorded

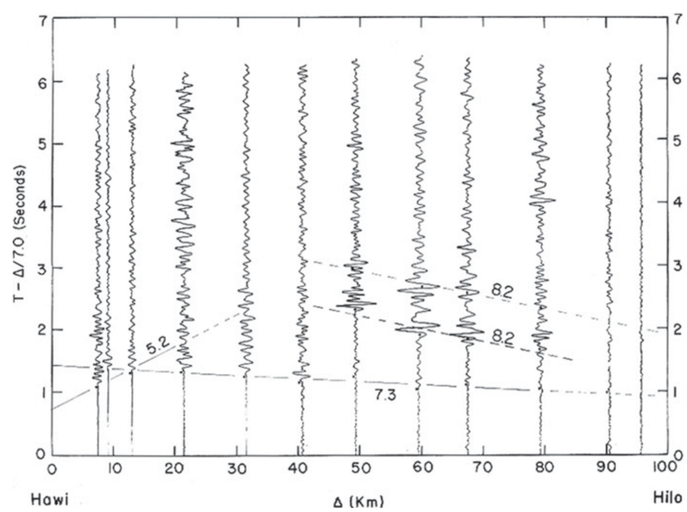


Figure 6.8.3-06. Record section of observations of a station located at the northern edge of the island of Hawaii (corresponding to position 5 of Fig. 6.8.3-04) and its recordings of shots along the northeastern coast of the main island of Hawaii (from Hill, 1969, fig. 5). [Bulletin of the Seismological Society of America, v. 59, p. 101–130. Reproduced by permission of the Seismological Society of America.]

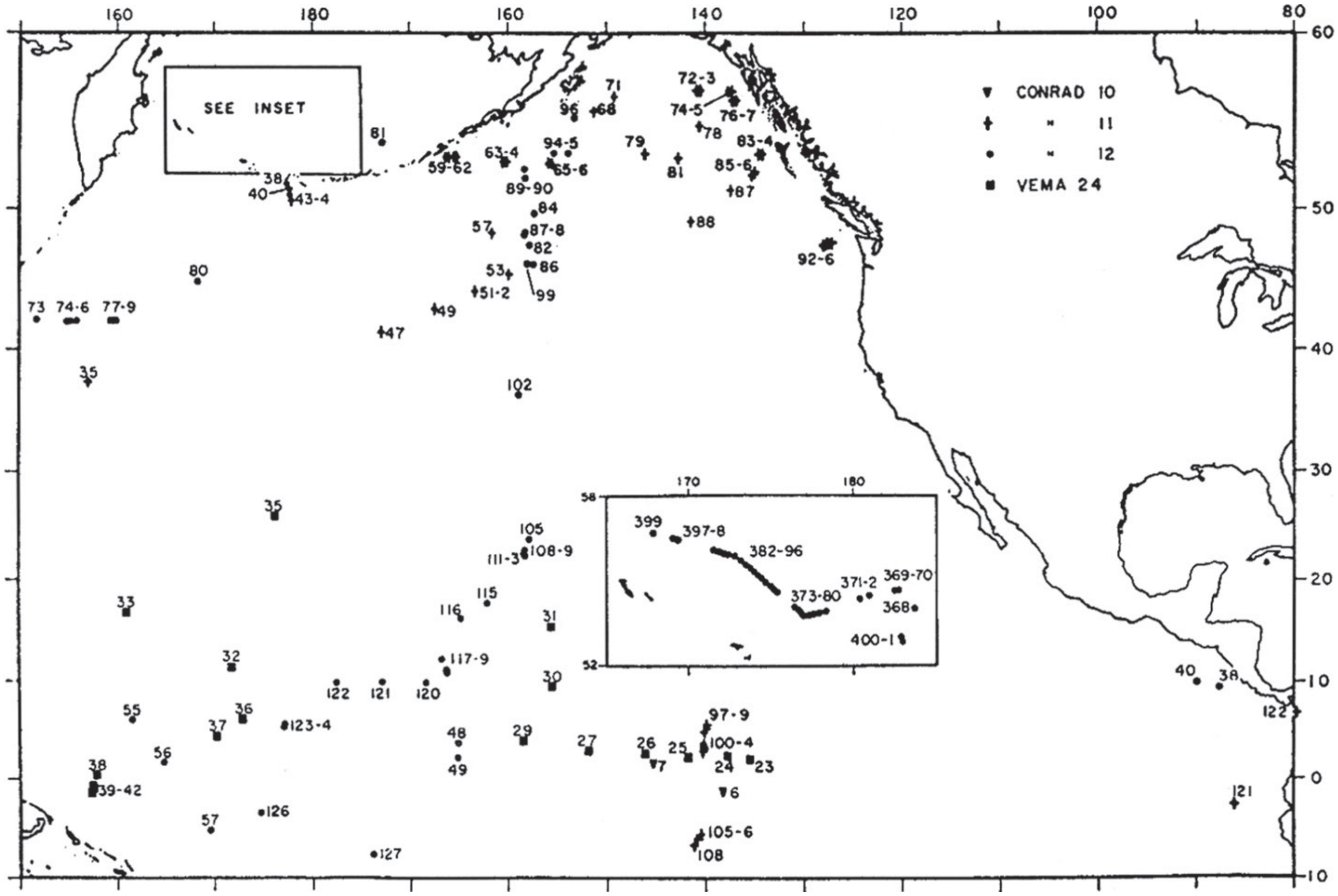


Figure 6.8.3-07. Chart of the North Pacific Ocean showing location of sonobuoy stations recorded in 1969 (from Houtz et al., 1970, fig. 1). [Journal of Geophysical Research, v. 75, no. 26, p. 5093–5111. Reproduced by permission of American Geophysical Union.]

by up to 47 stations. The recording stations were moved during the course of the survey to give wider coverage of New Britain and New Ireland (Fig. 6.8.3-08). Maximum distances were 300 km (Finlayson et al., 1972). A time-term analysis applied to the data by Finlayson and Cull (1973) resulted in a 10-km-thick crust including 2 km of water for the Bismarck Sea and a 12-km-thick crust including 5 km of water for the Solomon Sea to the southeast, with a crustal thickening to 20 km under the intervening New Britain. East of New Ireland, the crust of the Pacific plate appeared to be 35–40 km thick (Finlayson and Cull, 1973). The results have been summarized together with other surveys around the Solomon Islands and the southeastern part of Northern Melanesia by Furumoto et al. (1973) and are shown in Figure 6.8.3-09.

Seismic investigations were also performed in the area of the Cook Islands. In 1965, during the New Zealand Eclipse Expedition, seismic-refraction studies were made to investigate the structure of some of these volcanic islands. Near Aitutaki, two reversed profiles with a total length of 120 km determined a downwarp of the crust-mantle boundary by 1.6 km, the crustal thickness being 6.4 km and the upper-mantle velocity 7.7 km/s under the islands (Hochstein, 1968). The Lamont group was also active in this area in the 1960s and 1970s (F.J. Davey, 2010, personal commun.).

In 1967, the Scripps Institution investigated most of the Melanesian Borderland (Shor et al., 1971) bounded on the west by Australia and on the east by New Zealand and the Tonga-Kermadec trench (Fig. 6.8.3-10). Two ships were used for the seismic measurements, using the techniques described by Shor (1963) with only a few changes. The electronic equipment had been slightly modified, and reflection profiling was attempted throughout much of the refraction survey work. A shipboard computer enabled rapid computation of preliminary results. Explosives of up to 200 kg per shot continued to be the source of energy. Almost all profiles could be reversed and the profile directions were laid along strike of topographic features as much as possible.

The data analysis followed the conventional method by fitting lines to the traveltime data by least squares with the requirement that reversed times must agree for each layer. An example is shown in Figure 6.8.3-11. The results are shown as velocity-depth cross sections in Figure 6.8.3-12. The Lord Howe Rise

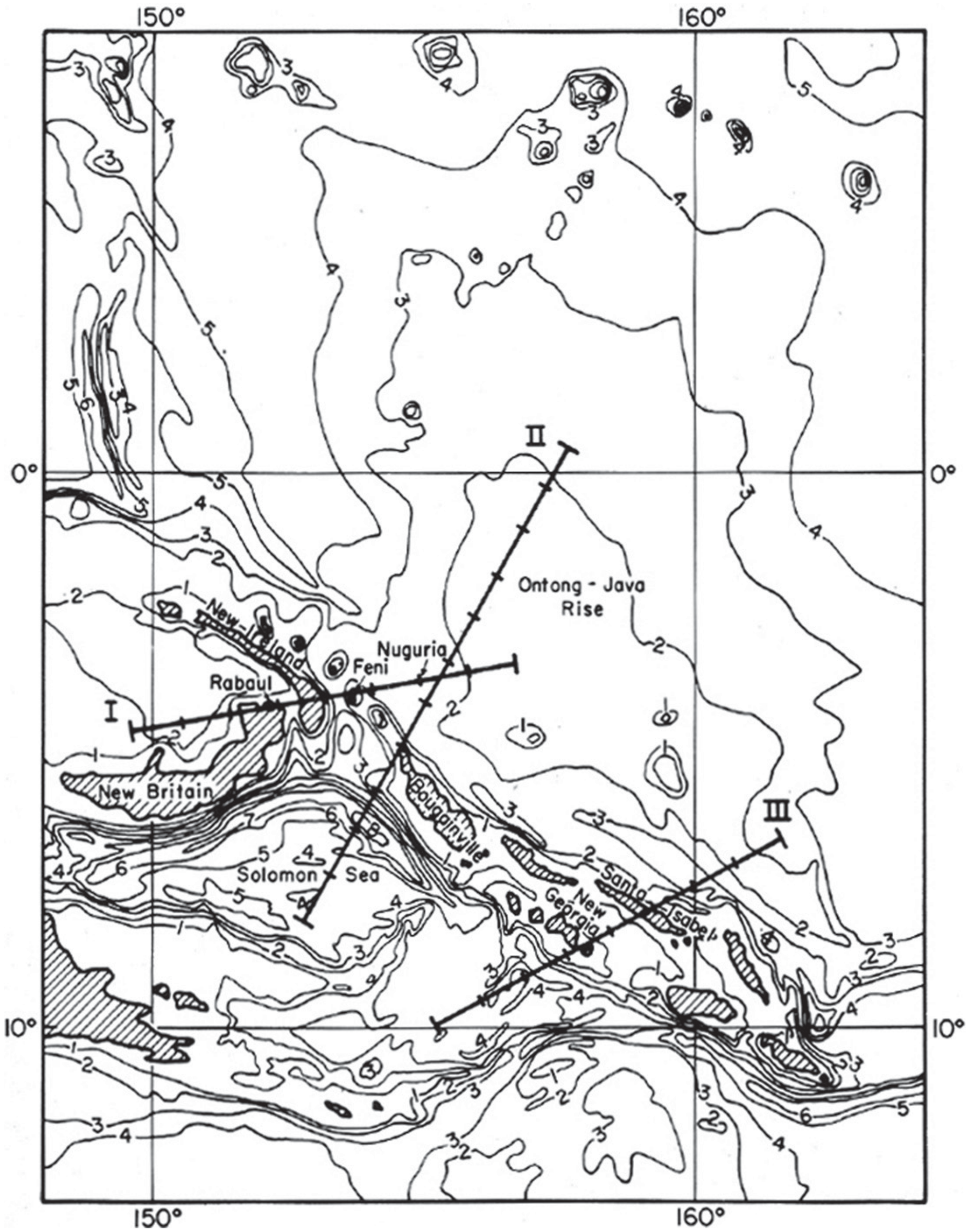


Figure 6.8.3-08. Map of northern Melanesia showing the locations of the crustal cross sections of Fig. 6.8.3-09 (from Furumoto et al., 1973, fig. 5). [Tectonophysics, v. 20, p. 153–164. Copyright Elsevier.]

and the Norfolk Ridge are topped by thick sediments and have deep crustal roots and thick layers with velocities measured on the Australian continent. The Kermadec and Lau ridges, on the other hand, have structure and velocities typical for island arcs. The South Fiji Basin resembles an oceanic area with additional sedimentary fill. The western part of the Fiji Plateau has thin sediments and the structure of a normal deep-ocean basin that has been uplifted by 2 km (Shor et al., 1971).

Also in 1967, the continental margin off northern Queensland was investigated by a two-ship wide-angle seismic profiling program, where 19 profiles were shot predominantly along a traverse from Cairns, Queensland, to Milne Bay, Papua New Guinea, and from Townsville across the Queensland Plateau to the deep Coral Sea Basin and Lousiade Rise (Ewing et al., 1970; Mutter and Karner, 1979; Finlayson 2010, Appendix 2-2).

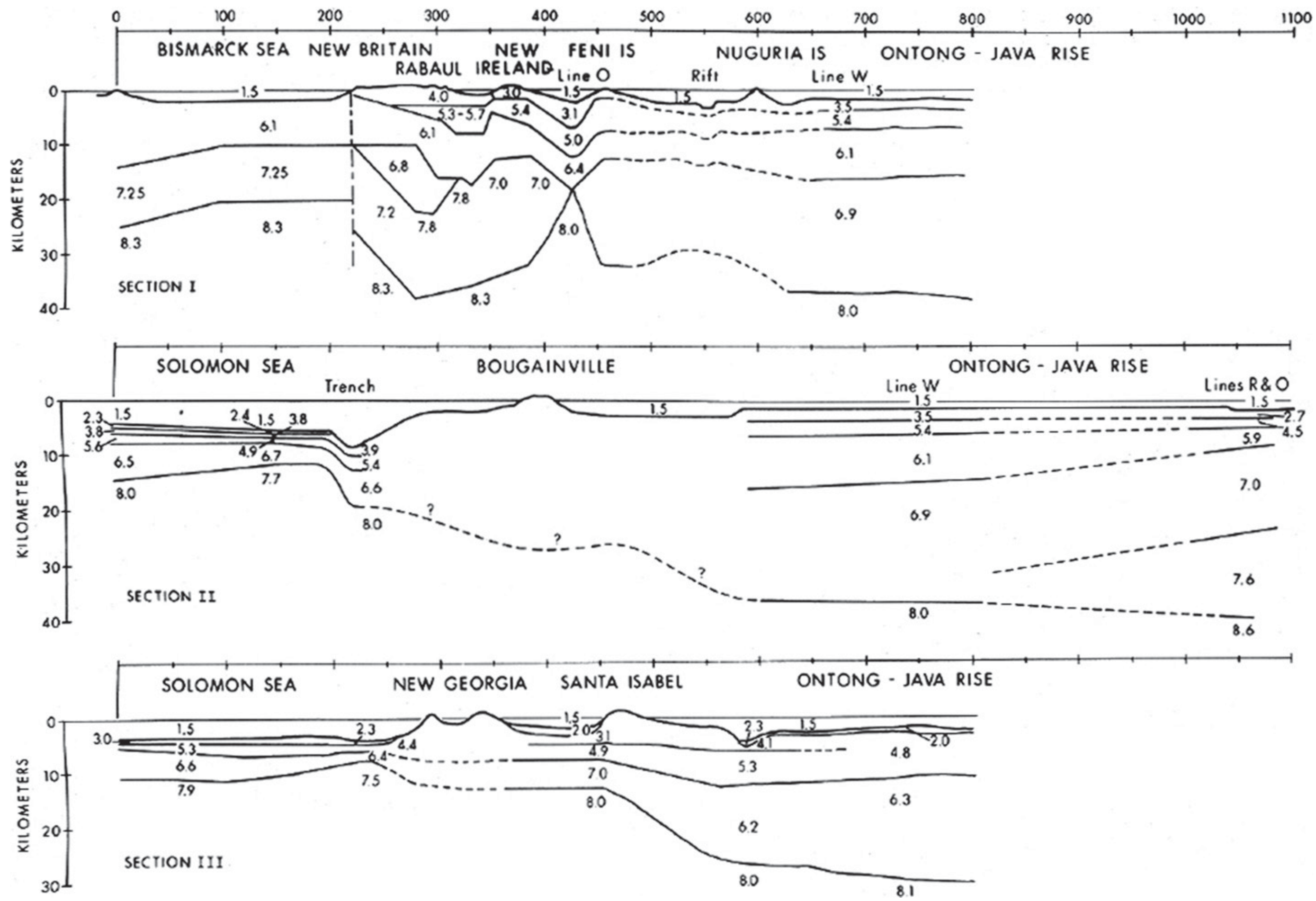


Figure 6.8.3-09. Velocity-depth profiles of the crust of northern Melanesia (from Furumoto et al., 1973, fig. 6). Locations are shown in Fig. 6.8.3 08. [Tectonophysics, v. 20, p. 153–164. Copyright Elsevier.]

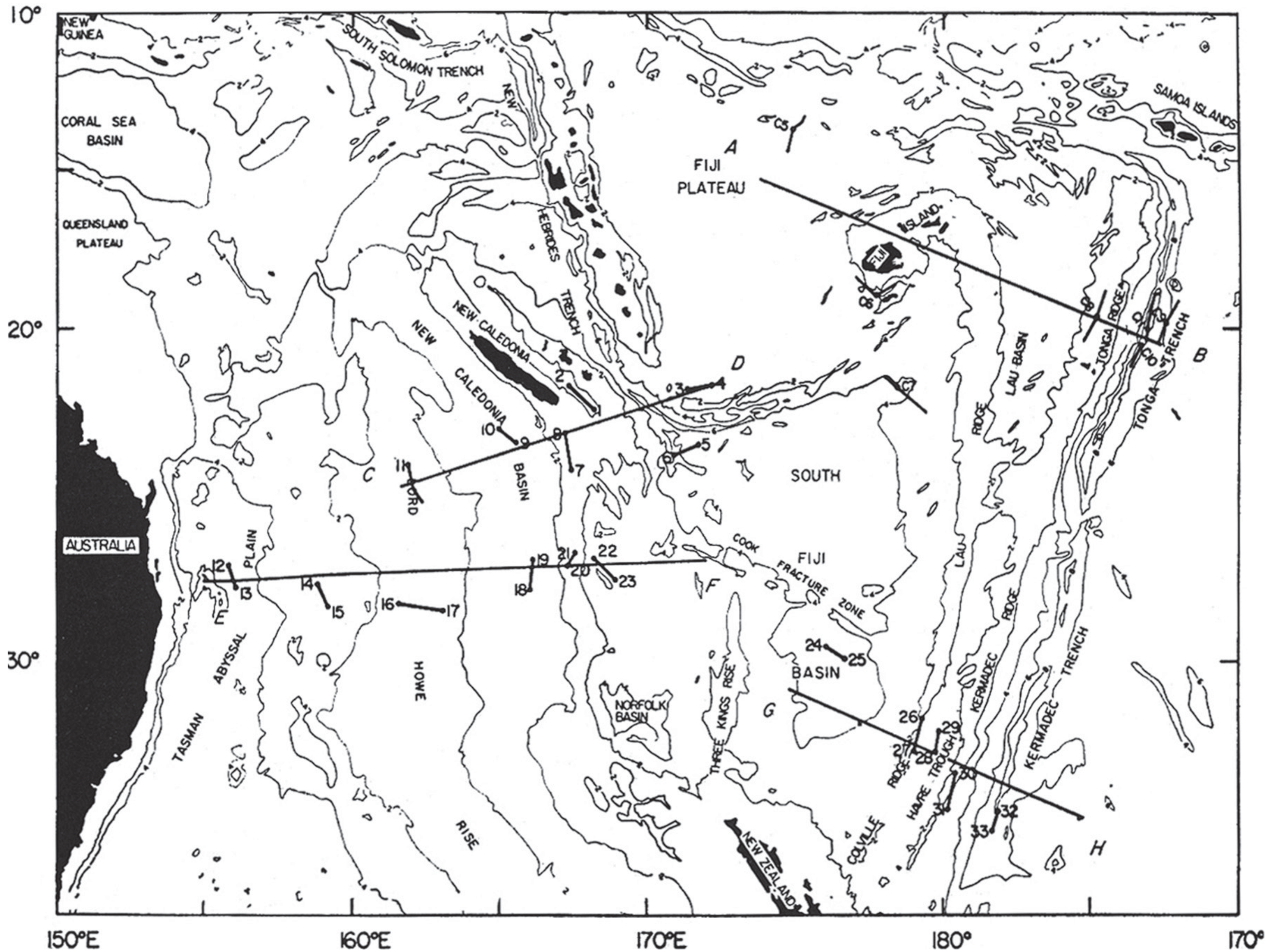


Figure 6.8.3-10. Location of seismic lines of the Nova expedition in 1967 in the Melanesian borderland (from Shor et al., 1971, fig. 1). Stations with prefix C are from the 1952 Capricorn expedition. [Journal of Geophysical Research, v. 76, no. 11, p. 2562–2586. Reproduced by permission of American Geophysical Union.]

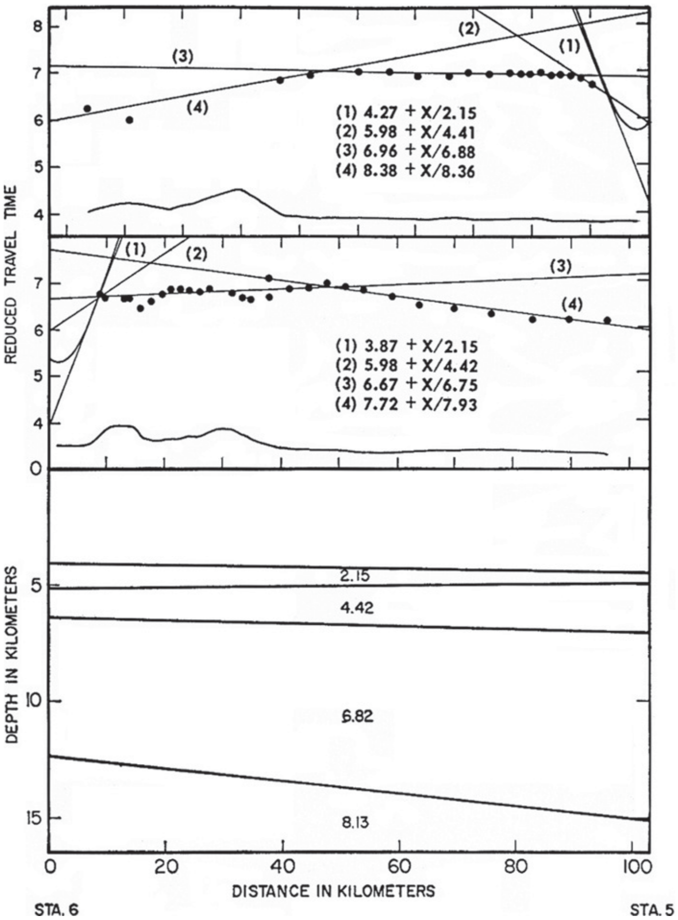


Figure 6.8.3-11. Traveltime plots and layer solutions for two of the Nova expedition reversing seismic refraction stations (from Shor et al., 1971, fig. 4). [Journal of Geophysical Research, v. 76, no. 11, p. 2562–2586. Reproduced by permission of American Geophysical Union.]

From various profiles in the Philippine Sea near latitude 25°N (Fig. 6.8.3-13), a crustal cross section (Fig. 6.8.3-14) was constructed by Murauchi et al. (1968) between the continental shelf of the eastern China Sea and the northwest Pacific Basin showing distinct crustal thickenings under the Nansei Shoto Ridge, the Oki-Daito Ridge, and the Honshu-Mariana Ridge, intervened by thin oceanic crust with Moho depths of 11–14 km below sea level. Only under the Honshu-Mariana Ridge, Moho was found with 8.0 km/s upper mantle velocity at 16 km depth; under the other ridges, the upper mantle was not reached.

Another interesting anomalous feature, found in the 1960s in the Northwest Pacific Basin along latitude 32°N and first published by Den et al. (1969), was the Shatsky Rise (Ludwig et al., 1970, fig. 13). Layer 3 with a velocity near 6.9 km/s at 8–14 km depth underneath the adjacent oceanic crust was found to be subdivided into two layers with average velocities of 6.9 and 7.5 km/s and reaching a total thickness of near 15 km, its lowest depth being near 24 km below sea level, its root being slightly shifted to the west versus the highest topographic elevation (Fig. 6.8.3-15). For comparison, Den et al. (1969) included to the left of their crustal section two crustal columns which had been obtained by Ludwig et al. (1966) in the westernmost part of the northwest Pacific Basin.

Russian scientists continued their exploration of the transition zone from the Asian continent to the Pacific Ocean with a second phase starting in 1963. The seismic investigations concentrated on the western part of the Pacific Ocean between the southern Kuril Islands and South Kamchatka (Kosminskaya et al., 1973). The experiments focused in particular on the Sakhalin-Hokkaido marginal zone, the Yamato Bank, the southern Kuril Islands and the earthquake focal zone east of Kamchatka. A particularly detailed profile was recorded across Iturup Island and further into the Pacific Ocean (Zverev and Tulina, 1971).

In 1964, Hess published an article on seismic anisotropy of the upper mantle under the oceans (Hess, 1964). To exclude the possibility that the observed directional dependence of seismic velocity was caused by lateral velocity changes, a series of special anisotropy experiments was subsequently carried out in various areas of the Pacific Ocean (Keen and Barrett, 1971; Morris et al., 1969; Raitt et al., 1969, 1971).With a special scheme of shot and receiver positions (Fig. 6.8.3-16), it was ensured that P-waves traversed the same area of the upper mantle in various azimuths.

In the subsequently applied time-term analysis of the P-traveltimes, both topography of the crust-mantle boundary and directional dependence of the velocities were taken into account. In the areas of Hilo and Show near Hawaii (Morris et al., 1969), this dependency became very clear: the anisotropy here was near 8%. In the areas of Flora and Quaret near the Californian coast (Raitt et al., 1969), the resulting anisotropy was 3% and 4%. Fuchs (1975, 1977) published a compilation of all anisotropy values obtained in the Pacific Ocean (Fig. 6.8.3-17). The compilation clearly shows that the directions of maximum velocity are almost perpendicular to the mid-ocean rises and run parallel to the major fault zones. In the Atlantic Ocean, only one experiment had been carried out by the end of the 1960s, resulting in 8% anisotropy (Keen and Tramontini, 1970).

6.8.4. The Indian Ocean

The crustal structure of the Indian Ocean was also investigated in the 1960s by several seismic refraction measurements, distributed to almost all directions. Laughton et al. (1970) have compiled a synthesis of the present knowledge on the Indian Ocean, as available up to January 1969, from seismic, magnetic, geothermal, topographic, earthquake epicenter, paleontological, and other geophysical and geological information, obtained during numerous cruises of research vessels mainly in the 1960s. In this context, we have extracted the information on the deep crustal structure as revealed from seismic data measured in the 1960s.

The Central Indian Ridge striking in N-S direction to the west of the Central Indian Basin was traversed in 1962 by a seismic-refraction profile at 6°S, 68°E (Fig. 6.8.4-01). However,

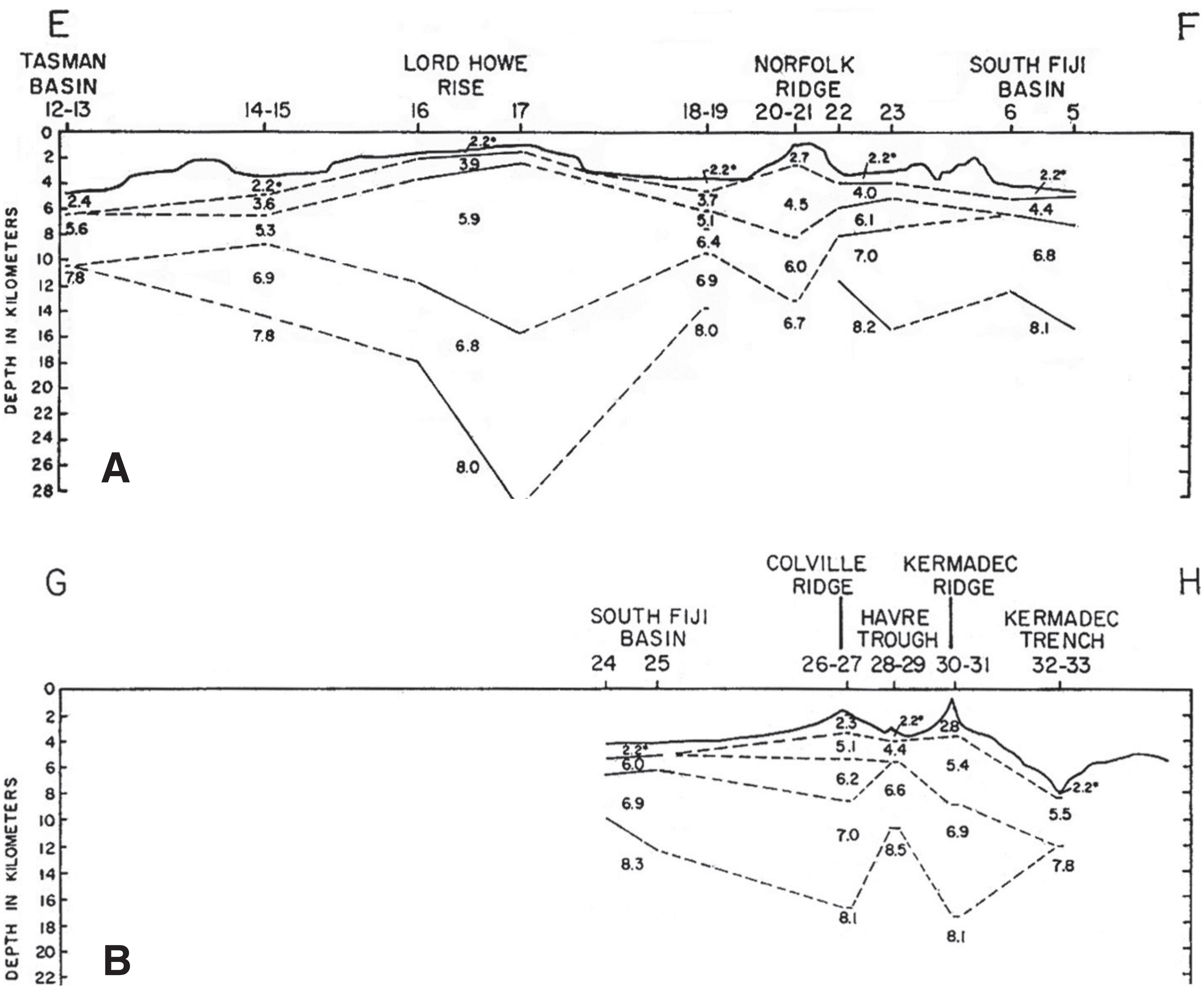


Figure 6.8.3-12. (A) Structures sections across the Melanesian borderland from the Tasman Basin to the Fiji Basin (for location see Fig. 6.8.3-10) (from Shor et al., 1971, figs. 2 and 3). (B) Structures sections across the Melanesian borderland from the South Fiji Basin to the Kermadec Trench (for location see Fig. 6.8.3-10) (from Shor et al., 1971, figs. 2 and 3). [Journal of Geophysical Research, v. 76, no. 11, p. 2562–2586. Reproduced by permission of American Geophysical Union.]

no station was recorded over the central 235 km (Francis and Shor, 1966). On the flanks, normal oceanic crust was found with crustal thicknesses of 4–7 km and subcrustal velocities of 8–8.4 km/s. Russian seismic-refraction data showed in their survey area, which lay in the center of the profile of Francis and Shor, a high-velocity layer with 7–7.5 km/s beneath 2.5 km of layer 2 with velocities of 4.5–5 km/s (Fig. 6.8.4-02). They found, however, no indication of subcrustal velocities (Neprochnov et al., 1967).

The Ninetyeast Ridge striking in a N-S direction to the east of the Central Indian Basin (Fig. 6.8.4-03; for location, see also Fig. 9.8.3-01) was first studied by a seismic survey in 1952 when it was thought to be an isolated seamount (Gaskell et al., 1958). Further seismic work followed in 1962, and it was then believed that the Ninetyeast Ridge at 10°S, 89°E was a horst-type feature in which oceanic crust had been upthrust (Francis and Raitt, 1967).

The Seychelles Bank at 5°S, 56°E was shown to be underlain by continental crust, based on seismic-refraction and gravity data obtained in 1962, while further south, the structure of the Saya de Malha Bank appeared as a coral capping, 1.5 km thick, which covers an oceanic type crust (Shor and Pollard, 1963).

In the eastern half of the Indian Ocean, three possibly continental-type aseismic plateaus were defined (Laughton et al., 1970). From seismic studies in 1962, Francis and Raitt (1967)

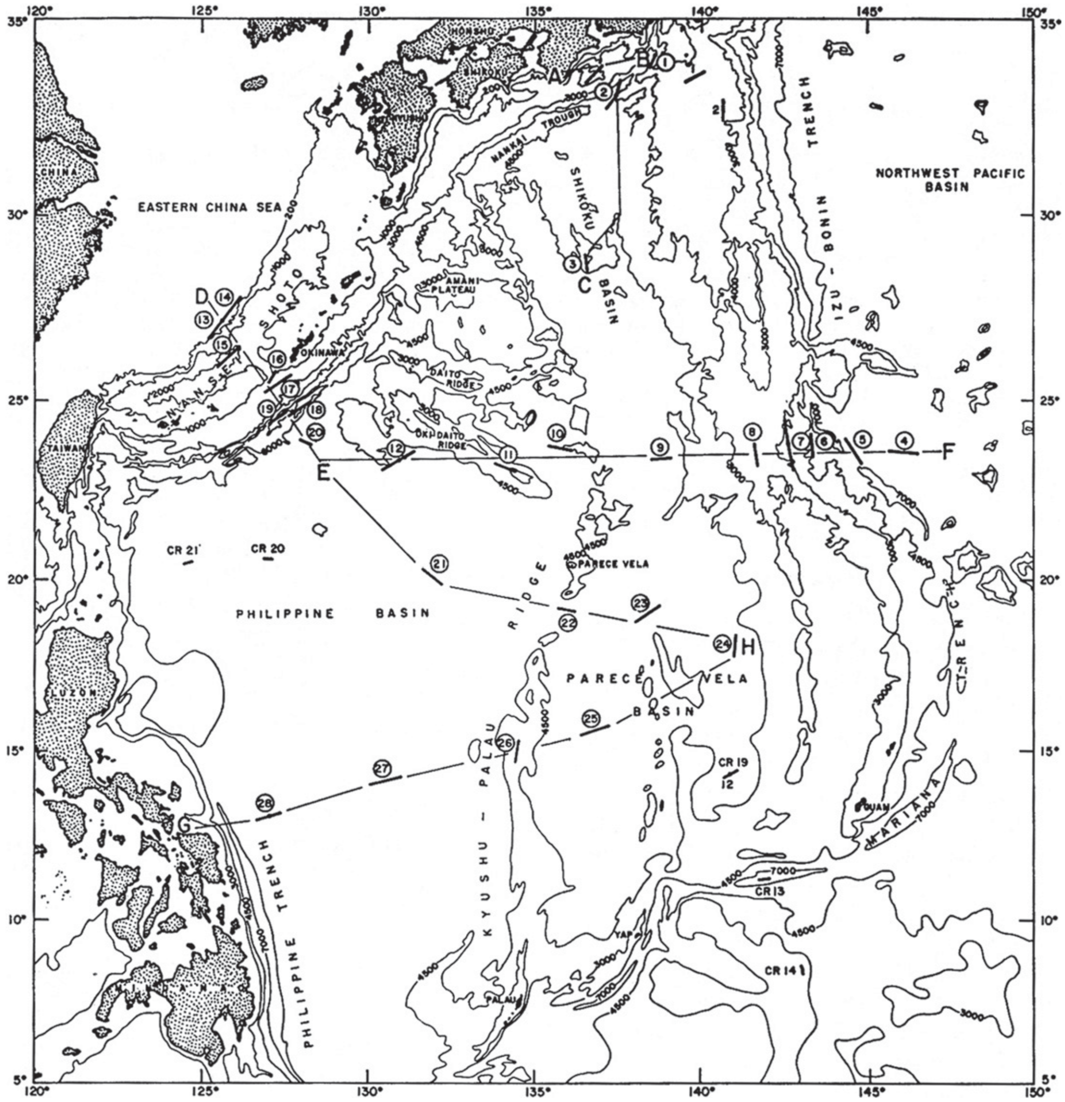


Figure 6.8.3-13. Geographic location of seismic profiles in the Philippine Sea (from Murauchi et al., 1968, fig. 1). [Journal of Geophysical Research, v. 73, no. 10, p. 3143–3171. Reproduced by permission of American Geophysical Union.]

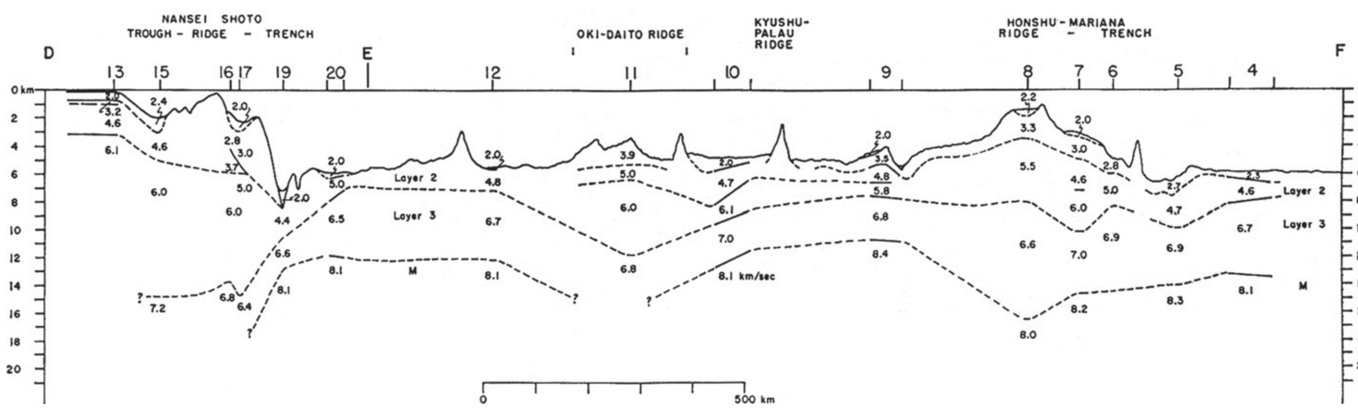


Figure 6.8.3-14. Structure section from the continental shelf of the Eastern China Sea into the Philippine Basin in NW-SE direction and continued into the NW Pacific basin in W-E direction (for location see Fig. 6.8.3-13) (from Murauchi et al., 1968, fig. 3). [Journal of Geophysical Research, v. 73, no. 10, p. 3143–3171. Reproduced by permission of American Geophysical Union.]

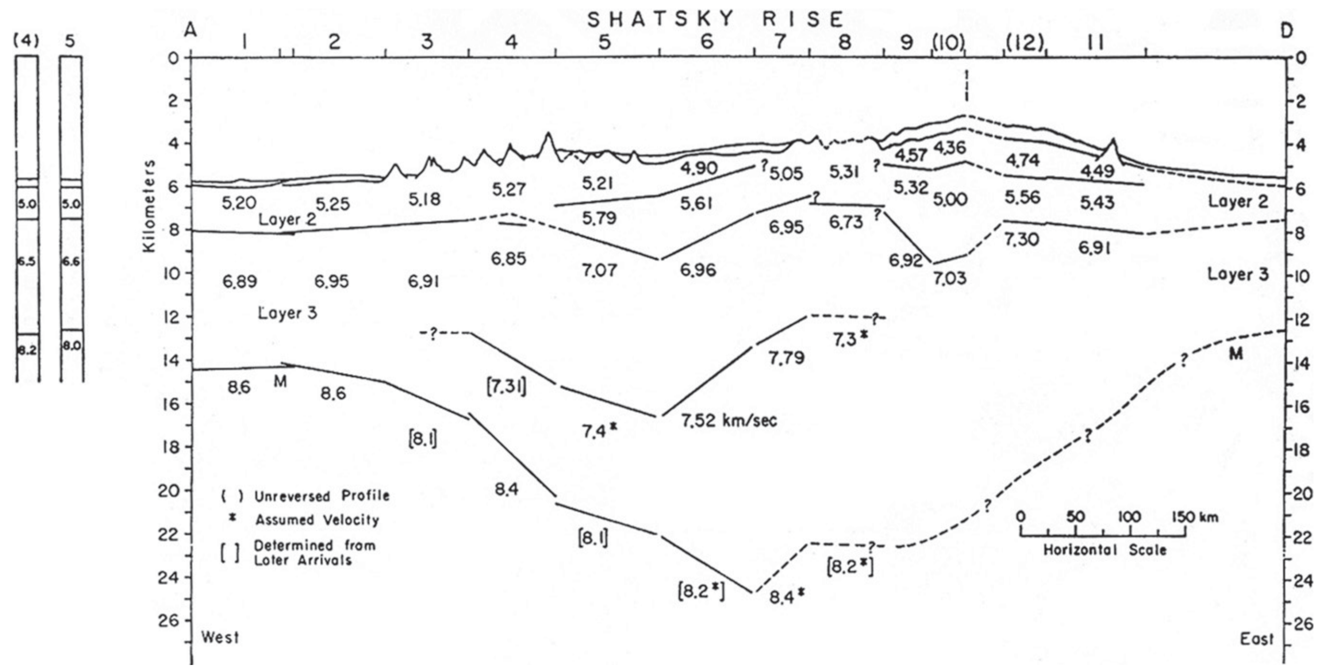


Figure 6.8.3-15. West-east structure section of the Shatsky Rise at approximately 32°N, 151°–160°E (from Den et al., 1969, fig. 3). Left: two crustal columns of Ludwig et al. (1966) from the Northwest Pacific basin. [Journal of Geophysical Research, v. 74, no. 6, p. 1421–1434. Reproduced by permission of American Geophysical Union.]

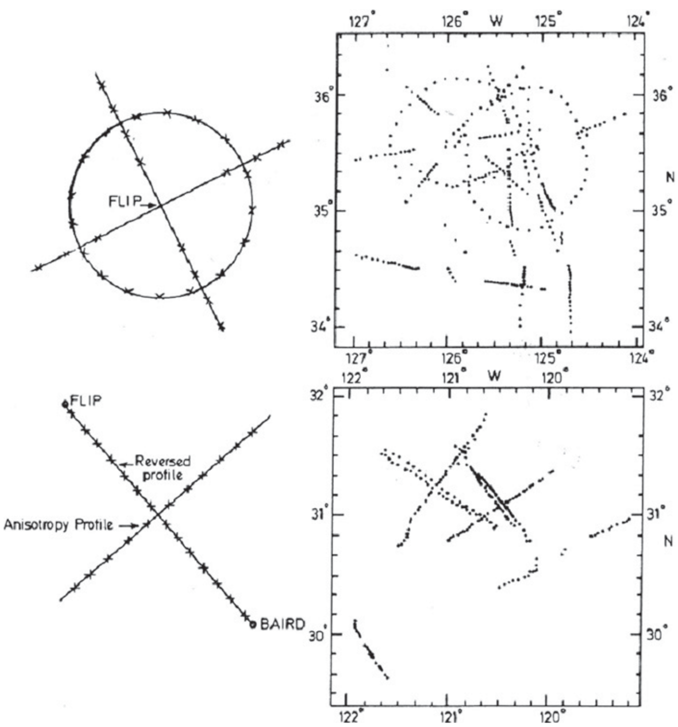


Figure 6.8.3-16. Shot and receiver arrangements in specific areas of the Pacific Ocean (from Fuchs, 1975, fig. 3). [Geologische Rundschau, v. 64, p. 700–716. Reproduced with kind permission of Springer Science+Business Media.]

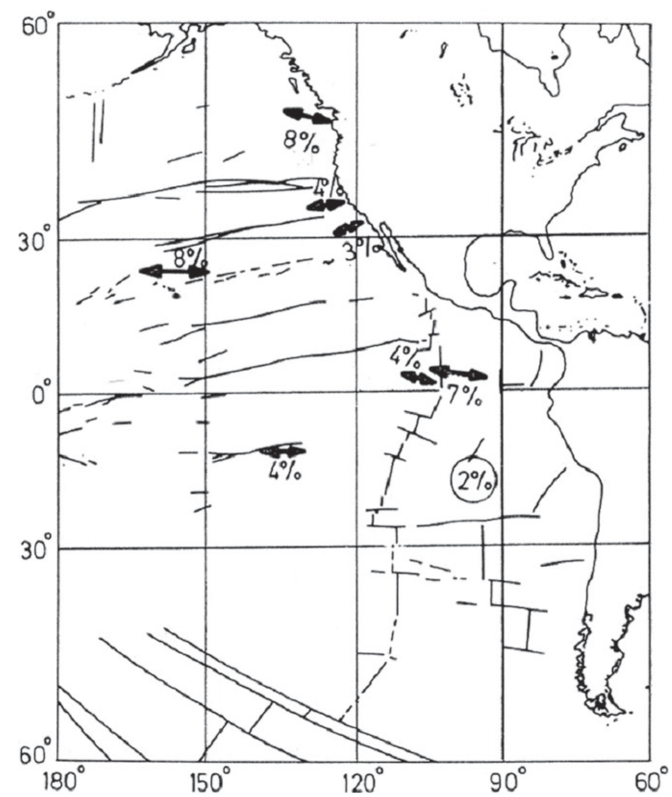


Figure 6.8.3-17. Distribution of anisotropy-values in the Pacific Ocean in direction and percentage (from Fuchs, 1975, fig. 6). [Geologische Rundschau, v. 64, p. 700–716. Reproduced with kind permission of Springer Science+Business Media.]

Figure 6.8.4-01. Station positions in the Northwest Indian Ocean, the line of stations 6 to 20 crossing the Central Indian Ridge (from Francis and Shor, 1966, fig. 1). [Journal of Geophysical Research, v. 71, p. 427–449. Reproduced by permission of American Geophysical Union.]

had suggested that the Broken Ridge and the Naturaliste Plateau were once part of the Western Australian Shield and that "the crust has either been thinned tectonically or the mantle has risen at the expense of the crust by some process at the Moho" (Fig. 6.8.4-04). The third Wallaby Plateau was not surveyed. In their summary, Laughton et al. (1970) concluded that the majority of the aseismic ridges might well be called "microcontinents."

Other seismic investigations in the Indian Ocean in the 1960s concentrated on the continental margins. On the Agulhas Bank, located south of Africa immediately to the southeast of the

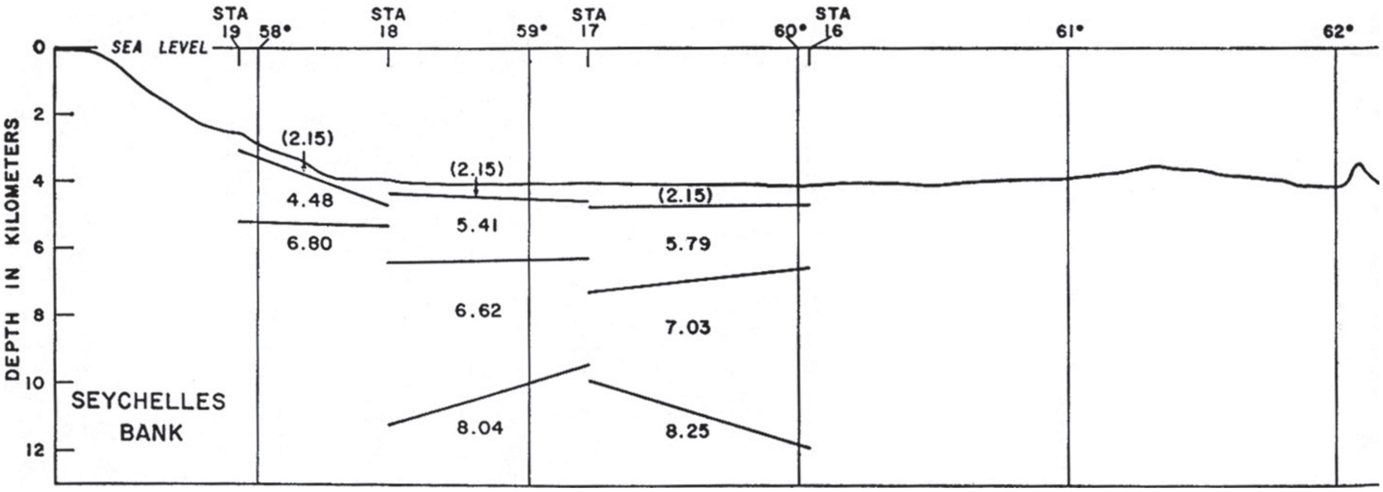


Figure 6.8.4-02 (*on this and following page*). Seismic profile between 57° and 67°E along a line between 5° and 6°S crossing the Central Indian Ridge (from Francis and Shor, 1966, fig. 4).). [Journal of Geophysical Research, v. 71, p. 427–449. Reproduced by permission of American Geophysical Union.]

Cape, the shelf extends 200 km from the coast (Fig. 6.8.4-05). Seismic-refraction studies were made in 1962 by Green and Hales (1966) and others were added later in the same year by Ludwig et al. (1968). A series of stations on the shelf showed continental crustal velocities that could be correlated with known lithologic structures (Fig. 6.8.4-06). An E-W section between the continental shelf off Durban and the Mozambique Basin along latitude 29°30′ S (Fig. 6.8.4-07) by Ludwig et al. (1968) suggested that there is oceanic crust on both sides of the ridge and that between the ridge and the continent is considerable cover. The Transkei Basin further south was established as being oceanic.

Additional seismic-refraction studies were carried out in the southern Indian Ocean in 1968 (Hales and Nation, 1972, 1973b) on the Agulhas Bank and Plateau and in deep water in the Transkei Basin adding to the previous work of Green and Hales (1966) described above. Free-floating recording buoys, built in 1963 and 1964 in the Southwest Center for Advanced Studies at Dallas, Texas, recorded shots which were alternatively floating and sinking or suspended at comparatively shallow depths (5–10 m) to produce bubble-pulse periods of 0.5–1 s. Several large shots of 900 kg, shot on a 190-km-long north-south profile along the Agulhas Bank, were also recorded by two land stations in South Africa. Arrival times beyond 120 km distance showed an anomalously high apparent velocity of 8.9 km/s. Hales and Nation (1972) did not believe that this high velocity was caused only by a dipping Moho.

Apart from the entrance to the Gulf of Aden, where the mid-oceanic Shaba Ridge penetrates the continent (crustal experiments see subchapter 6.7.2), the continental margins of Africa and Arabia as far north as Pakistan showed certain characteristics that distinguished them from the North Atlantic–type margins (Laughton et al., 1970). The question was raised as to whether the true edge of the African continent lay halfway between the coast and the Seychelles. To investigate this problem, a seismic-refraction section was laid out between Kenya and the Seychelles (Francis et al., 1966, reproduced by Laughton et al., 1970, fig. 10c). The western half of the section showed an abnormally thick crust (9–19 km) that comprised 4–12 km of sediments lying above a basement which, although having the appropriate velocity, lacked the magnetic properties of an oceanic crust. In contrast, the eastern half of the section showed an abnormally thin crust of 3 km in thickness with typical oceanic magnetic anomalies.

In the northern Arabian Sea, 17 seismic-refraction profiles were shot at the continental margins of Somalia and western India during the expedition M01 of the German research vessel *Meteor* (64) (Bungenstock et al., 1966; Closs et al., 1969a). These lines served to construct structural profiles, one from the Oman Basin across the Murray Ridge and three across the continental margin between Karachi and Bombay. In all the profiles, which extended 220 km seaward from the continental shelf edge, the basement was seen to be covered by 5–8 km of sediments. Basement velocities varied from 6 to 6.9 km/s, which seemed low for the oceanic crustal layer that was expected at a depth of 8 km. No base of this crustal layer was found.

From the various seismic profiles discussed above, several profiles also touched the Indian Ocean basins (Francis et al., 1966; Francis and Shor, 1966; Francis and Raitt, 1967; Neprochnov et al., 1967) and composite sections of the available seismic data were compiled and discussed by Laughton et al. (1970, fig. 11). They indicated that the structure under the Indian Ocean basins did not differ appreciably from that in other basins of the world. Apparent anomalous regions could be explained by extraordinary locations of the corresponding seismic stations close to disturbed structures. For example, huge sediment thicknesses near the Indus and Ganges cones were derived from erosion of the Himalayan mountain chain and extend far into the Central Indian Ocean Basin as far south as the equator with a thickness of greater than 2 km.

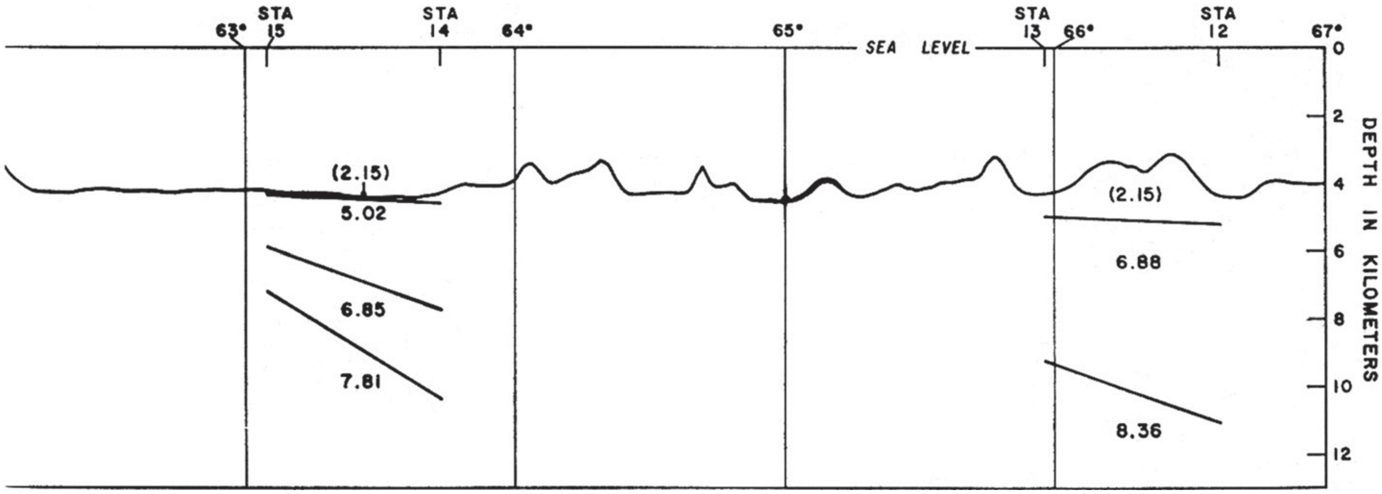


Figure 6.8.4-02 (*continued*).

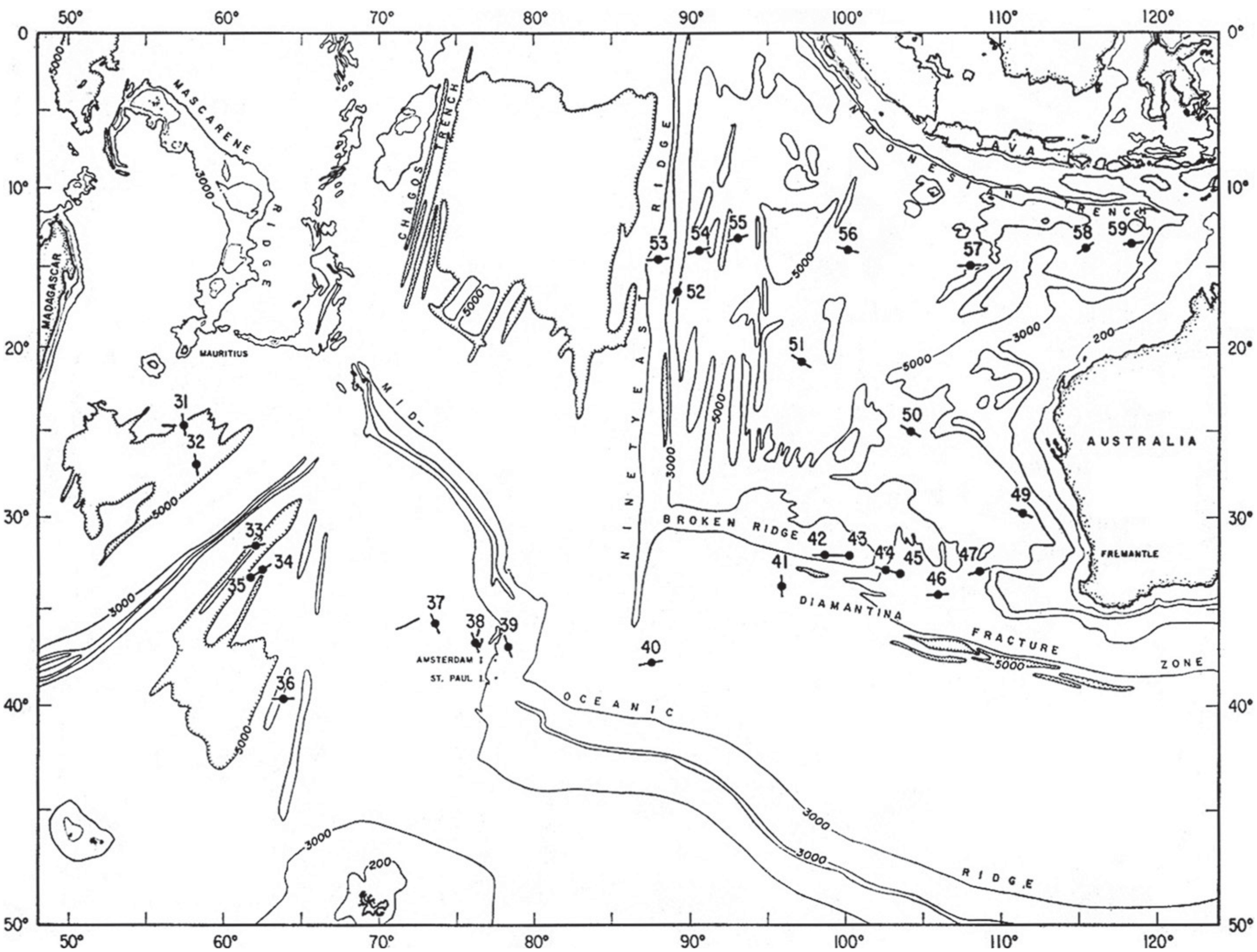


Figure 6.8.4-03. Station positions in the Southern Indian Ocean, the line of stations 42 to 47 crossing the Broken Ridge (from Francis and Raitt, 1967, fig. 1). [Journal of Geophysical Research, v. 72, p. 3015–3041. Reproduced by permission of American Geophysical Union.]

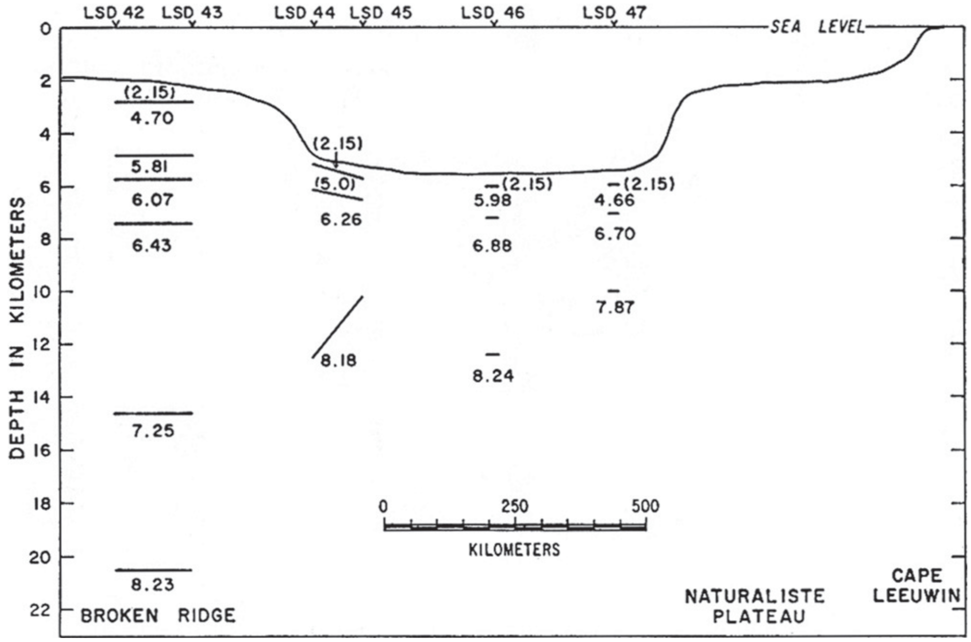


Figure 6.8.4-04. Seismic section on a line between station 42 and the tip of Australia (Cape Leeuwin) (from Francis and Raitt, 1967, fig. 3). [Journal of Geophysical Research, v. 72, p. 3015–3041. Reproduced by permission of American Geophysical Union.]

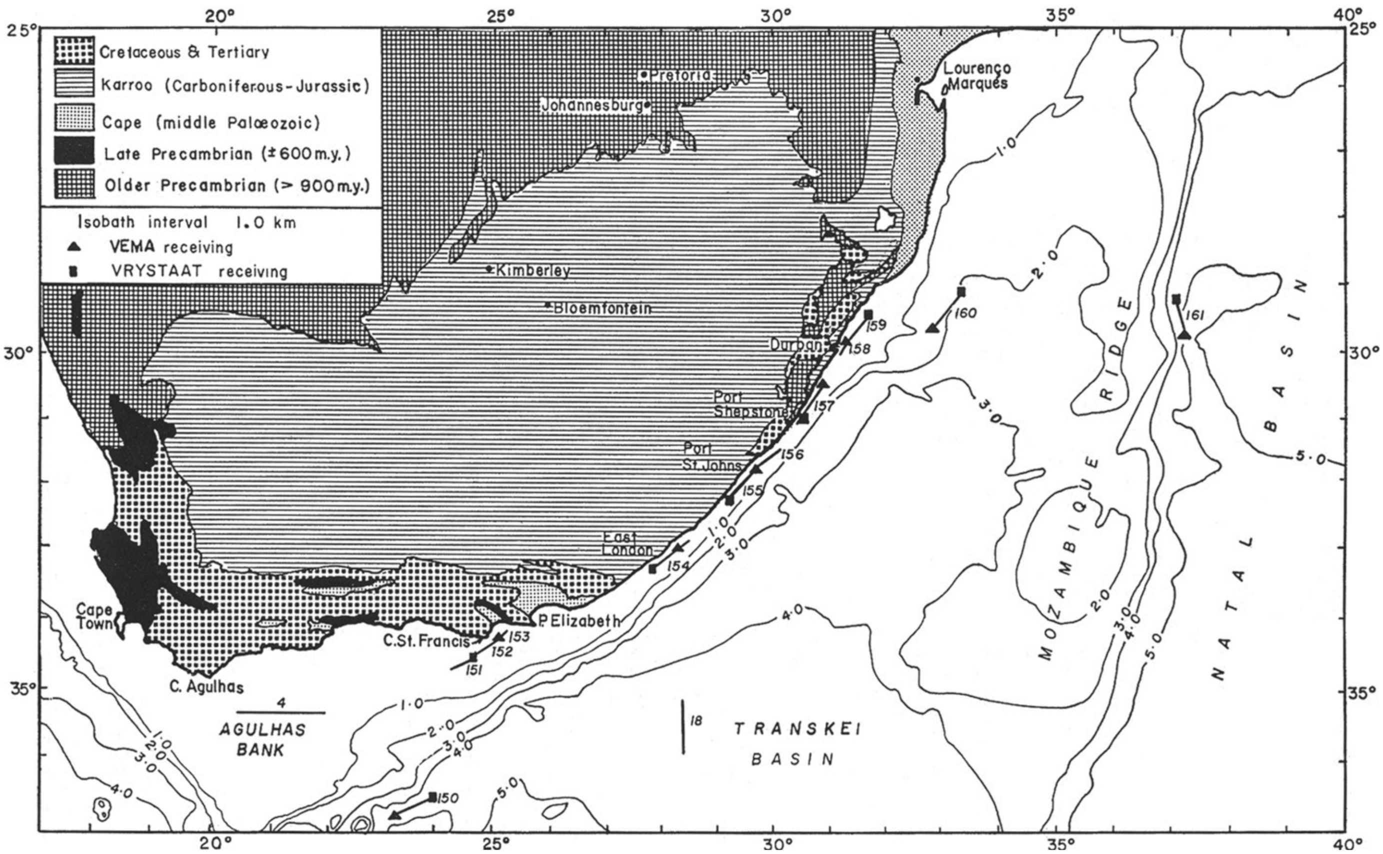


Figure 6.8.4-05. Simplified geological map of southeastern Africa showing the locations of offshore seismic refraction profiles (from Ludwig et al., 1968, fig. 1). [Journal of Geophysical Research, v. 73, p. 3707–3719. Reproduced by permission of American Geophysical Union.]

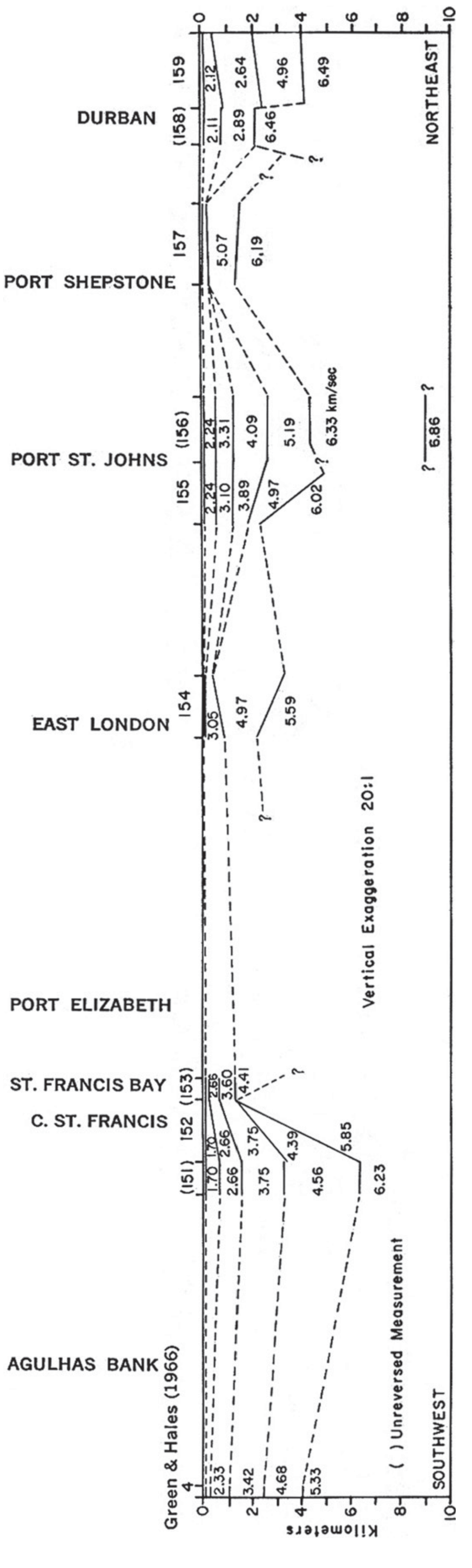


Figure 6.8.4-06. Seismic refraction section along the continental shelf between Agulhas Bank and Durban (from Ludwig et al., 1968, fig. 2). [Journal of Geophysical Research, v. 73, p. 3707–3719. Reproduced by permission of American Geophysical Union.]

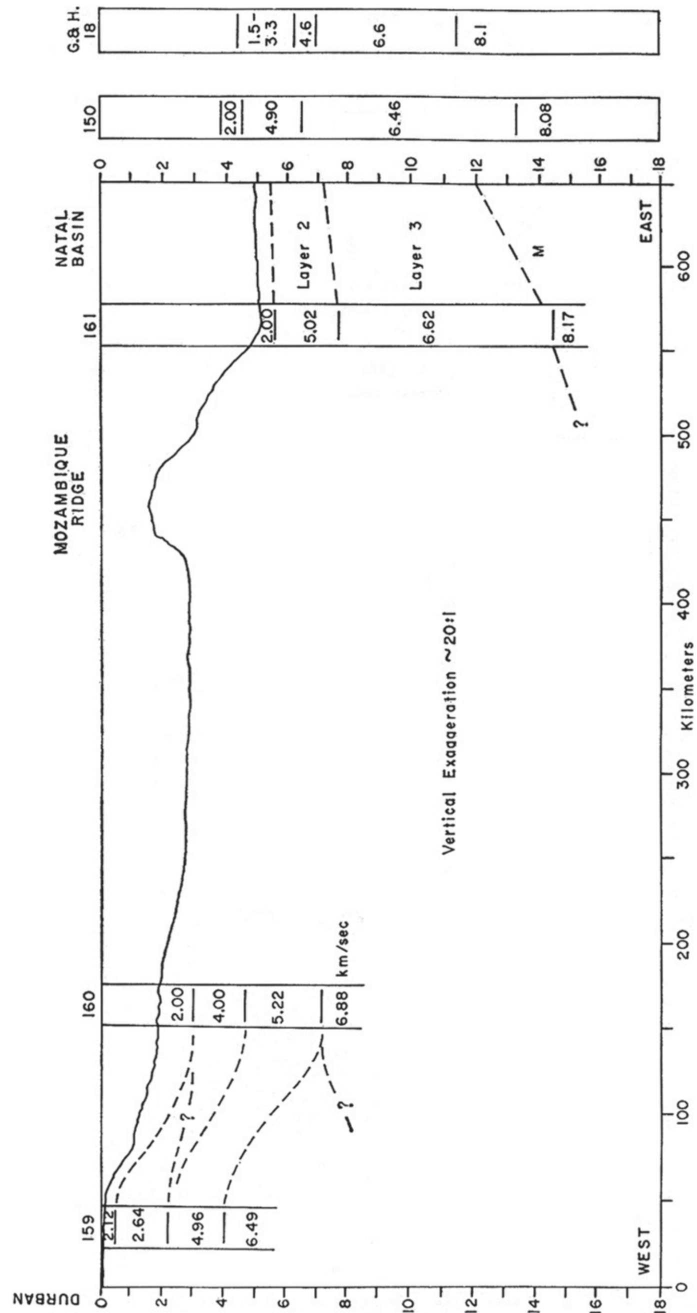


Figure 6.8.4-07. Seismic refraction section between the continental shelf off Durban and the Natal basin along latitude 29°30′ S (from Ludwig et al., 1968, fig. 3). [Journal of Geophysical Research, v. 73, p. 3707–3719. Reproduced by permission of American Geophysical Union.]

6.8.5. Summary of Seismic Observations in the Oceans in the 1960s

An overview of seismic measurements in the oceans, compiled from the database assembled at the U.S. Geological Survey in Menlo Park, California, for the 1960s shows all surveys that were published until 1969 in generally accessible journals and books (Fig. 6.8.5-01). The map shows the numerous observations in continental shelf and adjacent deep ocean areas of North America and eastern Asia. Also in the Red Sea, Gulf of Aden, as well as in the entire Indian Ocean, a large number of observations were made in the 1960s. In the northern Atlantic Ocean the Mid-Atlantic Ridge and in the Pacific Ocean the Hawaii islands were of special interest.

On the basis of many observations in the Atlantic Ocean, a detailed overall picture of the oceanic crust was established, but significant deviations from this average were found on the flanks of the Mid-Atlantic Ridge. Numerous new data were also gathered in the Indian and Pacific Oceans. Detailed surveys involved, for example, the surroundings of the islands of Hawaii and archipelagos to the northeast of Australia (not shown on the map). For the first time, the existence of seismic anisotropy of the uppermost mantle was discussed and investigated by a series of special anisotropy experiments in various areas of the Pacific Ocean (e.g., Raitt et al., 1971).

6.9. THE STATE OF THE ART AT THE END OF THE 1960s

The 1960s accumulated a wealth of data and the art of seismic-refraction techniques was gradually improved in both instrumental development and in the art of interpretation. In many experiments, seismic energy produced by effective explosion sources was recorded at large distances of up to several 100 km. Moho contour maps published for Europe (e.g., Morelli et al., 1967), the USSR (e.g., Belyaevsky et al., 1973), and the United States (Warren and Healy, 1973) gave evidence for the large amount of published models, available by the end of the 1960s. They not only resulted in a quite complete picture of the gross velocity-depth structure of the Earth's crust underneath the northern hemisphere around the world, but also detected specific properties of different tectonic areas such as shields, platforms, orogens, basins, rift zones, etc. However, the long-range data had not yet been interpreted in terms of subcrustal fine structure.

On the other hand, groups in Germany (e.g., Dohr, 1957, 1959; Krey et al., 1961; Liebscher, 1962, 1964; German Research Group for Explosion Seismology, 1964; Dohr and Fuchs, 1967), the USSR (e.g., Beloussov et al., 1962; Kosminskaya and Riznichenko, 1964; Zverev, 1967), and Canada (e.g., Kanasewich and Cumming, 1965; Clowes et al., 1968; Kanasewich et al., 1969) had demonstrated the feasibility of determining fine crustal structure using near-vertical incidence reflection methods, and such techniques have been used extensively in subsequent decades.

In order to understand the seismic wave field, various groups had worked on the theoretical background so that by the end of the 1960s, the art of computing synthetic seismograms was ready for application. Other groups had concentrated on the character and basic features of seismic phases so that gradually a fine structure of the hitherto homogeneously layered crust was detected. In particular the character of the crust-mantle boundary was attracting increased interest.

Furthermore, two volumes on oceanic research edited by Maxwell (1970) summarized all achievements and subsequent concepts on seafloor evolution, based mainly on the numerous marine seismic investigations in the oceans collected since the end of World War II until the end of the 1960s.

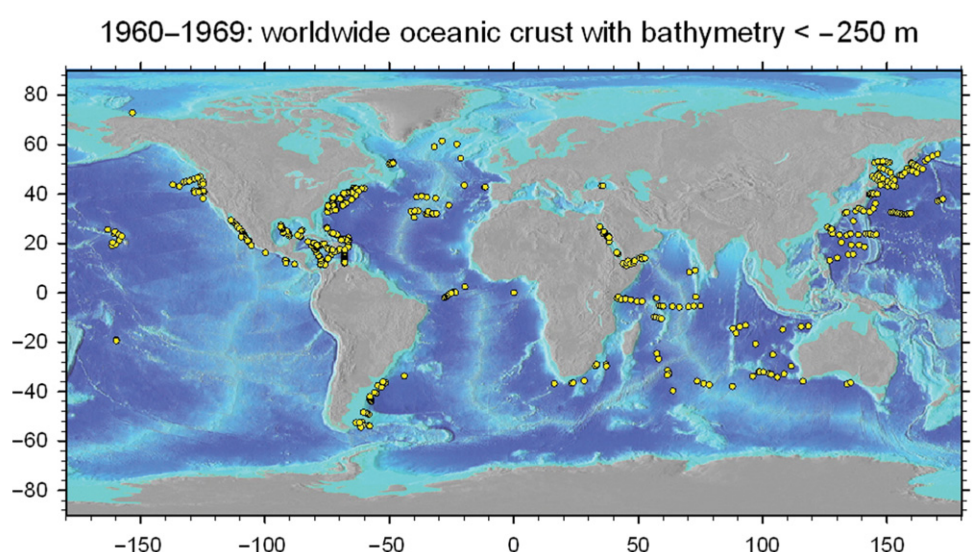


Figure 6.8.5-01. Seismic refraction measurements in the Atlantic, Pacific and Indian Oceans performed between 1960 and 1969 (data points from papers published until 1969 in easily accessible journals and books).