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CHAPTER 7 The 1970s (1970–1980)

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CHAPTER 7

The 1970s (1970–1980)

7.1. THE DECADE OF INTERNATIONAL COOPERATION

While it took almost 100 years from its beginnings in 1851 to lay the foundations of successful controlled-source seismic experiments, the development of this particular branch of seismology grew rapidly in the first 25 years following the end of the Second World War in 1945. In the two foregoing chapters we have tried to present an overview of all major developments and experiments carried out in the 1950s and 1960s. However, by the end of the 1960s, the number of controlled-source seismology experiments had reached a level such that a complete description of all major projects for the following decades would be forbidding. Therefore, we shall restrict ourselves to particular highlights providing major impact on our knowledge of crustal and uppermost-mantle structure.

The personnel situation in the geophysical departments of the universities and other research institutions of Western Europe had improved significantly owing to the favorable development in the 1960s. By the beginning of the 1970s, many young scientists had become professors and heads of departments. In Paris, L. Steinmetz, assisted by A. Hirn, G. Perrier, J.P. Ruegg, and many others, had started the Groupe Grands Profils Sismiques. In Great Britain, M.H.P. Bott and R.E. Long at Durham, D.H. Griffiths and R. King with their assistants and students D. Bamford, D.J. Blundell, M.A. Khan, M. Brooks and K. Nunn at Birmingham, as well as P.L. Willmore and B. Jacob in Edinburgh, pursued explosion seismology research. In Germany, explosion seismology was pushed forward in particular by H. Berckhemer and B. Baier (Frankfurt), J. Dürbaum and K. Hinz (BGR Hannover), K. Fuchs and C. Prodehl (Karlsruhe), P. Giese (Free University of Berlin), H. Gebrande and H. Miller (Munich), J. Makris and W. Weigel (Hamburg), and R. Meissner (Kiel). In Switzerland, it was S. Mueller, who in 1971, together with J. Ansorge, E. Wielandt, and others, moved from Karlsruhe to Zurich, Switzerland, and started a new period of active and passive seismology research at the Eidgenössische Technische Hochschule (ETH). In Italy, it was C. Morelli in particular who, since the early 1960s, had established a strong explosion seismology group, assisted by G. de Visintini and R. Nicolich and supported strongly by the Milan group with R. Cassinis and S. Scarascia. In northern Europe, M.A. Sellevoll at Oslo, Norway, C.-E. Lund and A. Vogel at Uppsala, Sweden, and M. Korhonen and U. Luosto in Finland were actively involved in explosion seismology projects.

In the former USSR, deep seismic sounding projects were continued in particular by I.P. Kosminskaya (Moscow), V.B. Sollogub (Kiev), and N.N. Puzyrev (Novosibirsk). Furthermore, in the USSR, L.P. Vinnik (Moscow), A.S. Alexeyev (Novosibirsk), S.M. Zverev (Moscow), and N.I. Pavlenkova (who had moved from Kiev to Moscow) pursued explosion seismology in active experiments and methodology studies. In Poland, A. Guterch (Warszaw) and in Czechoslovakia, V. Červený (Prag) became well-known scientists to the western explosion seismology community.

In the United States, wide-angle reflection and refraction seismology played a minor role throughout the first half of the 1970s, but in 1975, strong interest in exploring the crust by deep seismic-reflection surveys led to the foundation of COCORP (Consortium for Continental Reflection Profiling), pushed forward by, e.g., J.A. Brewer, L.D. Brown, S. Kaufman, J.E. Oliver (all at Cornell University, Ithaca, New York), R.A. Phinney (Princeton, New Jersey), and S.B. Smithson (Laramie, Wyoming). Seismic-refraction activities were concentrated at a few universities, e.g., in Wisconsin (R.P. Meyer and W.D. Mooney), Salt Lake City (R.B. Smith, assisted by L.W. Braile, G.R. Keller, and others), and Dallas (A. Hales. who soon moved to Australia). At the U.S. Geological Survey, while the main activity of the seismology group concentrated on earthquake research, J.H. Healy started to raise new interest by developing a new type of equipment. With one hundred pieces of the so-called cassette recorder, the U.S. Geological Survey started a new style of fieldwork, first tested in the Saudi Arabian long-range profile in 1978, which laid the foundation for major new activities in the following decade.

In Canada, it was, e.g., M.J. Berry (Ottawa), R.M. Clowes, D.A. Forsyth, Z. Hajnal, R.F. Mereu (London, Ontario), and others who continued controlled-source seismology projects, which finally, at the end of the 1970s, led to the foundation of the COCRUST project. This project was different to other national pure reflection programs, as it included both refraction and reflection surveys.

In the southern hemisphere, it was especially in Australia where an active group of scientists such as J.C. Dooley, D. Denham, D.M. Finlayson, and C.D.N. Collins (Bureau of Mineral Resources, Canberra), as well as K. Muirhead and B.J. Drummond (Australian National University, Canberra) pursued crustal and upper-mantle research by both active and passive source seismology. In South Africa, it was R. Green in the University of Witwatersrand at Johannesburg, South Africa, who pushed crustal and upper mantle research by developing instrumentation that enabled the performance of one-man seismic-refraction surveys.

The 1970s can be characterized by several highlights. The first was the detection of fine structures in the lithosphere below the Moho all over the world, using controlled-source seismic experiments with observation distances beyond 300 km.

The second highlight was advanced research and understanding of continental rifting in Europe, Asia, Africa, and North America. The consequent application of the time-term method widely used by British scientists on data from dense networks of seismic-refraction profiles led unexpectedly to the detection of uppermost mantle velocity anisotropy also under continents.

A third highlight was the increased investigation of the Mediterranean area, and special research efforts concentrated on details of the crustal structure above hot spots such as Hawaii and the Yellowstone–Snake River Plain area.

The 1970s also saw the opening of the first national large-scale seismic-reflection program: COCORP in the United States, which was soon followed by the Canadian equivalent, COCRUST.

The interpretation of the new data was guided by the improvement of lithosphere models using the reflectivity method, which allowed the calculation of amplitudes, frequency contents, and reverberations in laterally homogeneous media.

A similar rapid development was achieved in oceanic research. Numerous marine experiments were carried out, their character changing gradually with development of instrumentation, logistics, and theoretical background. Here, the reader is referred to Chapter 7.8.1.

During the second half of the decade, the ray method was developed, which led to the development of several ray tracing methods that could also handle data in complex tectonic structures and which would become the standard tool for interpreting seismic data in the 1980s and following decades up to the present day.

From the large amount of new data and their interpretations, Soller et al. (1981) have prepared a global thickness map, and Prodehl (1984) has prepared maps of the world showing the location of crustal and upper-mantle surveys available at that time and has summarized the main crustal features in tables (reprinted in Appendix A7-1).

7.2. CONTROLLED-SOURCE SEISMOLOGY IN EUROPE

7.2.1. Fine Structure of the Lower Lithosphere—The First West European Long-Range Profile

Though many seismic-refraction profiles observed in the 1960s reached recording distances of several hundred kilometers, the fine structure of the P_n phase had barely been noticed. A first report on systematic amplitude variations of the P_n phase was published in 1966 by a Soviet scientist (Ryaboi, 1966). Ansorge (1975) took up this idea and started to investigate the long-range observations of the Lake Superior experiments in North America, which at the same time were also more closely investigated by Massé (1973), and compared them with data recorded in the past 25 years in central and northern Europe with distances ranging from 400 to 2200 km (Heligoland-South 1947, Lac de l'Eychauda 1963, Norway 1965, Lac Nègre 1966, Folkestone 1967, Trans-Scandinavian Profile 1969, quarry blast observations at Böhmischbruck, Germany 1960–1971). He found that a similar fine structure described by Ryaboi (1966) for the P_n phase could also be seen in these sparse data, which indicated that a rather complex structure in the uppermost mantle down to 120 km depth was to be expected (Ansorge, 1975; Ansorge and Mueller, 1971).

So, the idea was born to systematically plan for an experiment out to 1000 km in western Europe to be placed within a tectonically unique surrounding. Experience from the Early Rise project and the theoretical investigations of Wielandt (1972, 1975) had shown that underwater shooting of relatively small depth charges (1000–3000 kg) at optimal water depths would produce sufficient energy to be recorded across a profile of possibly up to 1000 km long. A profile planned diagonally through France from the Bretagne across the Massif Central to the Mediterranean coast being located mainly on Variscan rocks (Fig. 7.2.1-01) would almost fulfill this requirement and would furthermore offer the chance to work with sufficiently large offshore shots at both ends.

Following the recommendation of their German colleagues, in 1969, the French geophysical community had bought 30 MARS-66 recording devices from the German company Lennartz. The purchase of the German equipment by the University of Paris continued the close cooperation between French and German geophysicists which had started with the Haslach explosion in 1948 and was continued in the investigation of the Western Alps in the 1950s and 1960s. This equipment was thoroughly tested and used in detailed research across the Massif Central, which will be described in more detail below, where the French colleagues also gained experience with organizing borehole shots on land (Perrier and Ruegg, 1973; Hirn and Perrier, 1974). Also the Swiss ETH and the Portuguese Meteorological Survey bought new MARS (magnetic tape recording system) equipment, thus increasing the total amount of seismic systems to be used in the subsequent international cooperative projects.

The joint effort of L. Steinmetz and A. Hirn at Paris, S. Mueller and J. Ansorge at Zurich, and K. Fuchs and C. Prodehl at Karlsruhe raised the necessary funding and so, with the German and the recently acquired French equipment, in 1971, the first long-range profile in Europe with a total length of 1200 km was realized (Groupe Grands Profils Sismiques and German Research Group for Explosion Seismology, 1972). In addition to the offshore sea shots crustal control became involved by placing borehole shots at 200–300 km intervals along the line (see record sections in Appendix A2-1, p. 36–37, France 1971). The station spacing was from 5 km up to 600 km distance and 10 km for distances between 600 and 1200 km. The shots in the Atlantic Ocean off Brest and the intervening land shots along the line carried energy as expected, even to the island of Corsica at 1400 km distance, but due to unfavorable locations as well as bad-weather conditions the reverse upper-mantle shots off St. Tropez in the Mediterranean caused a considerable decrease in seismic efficiency and could not be recorded along the entire line.

As expected and hoped for, the crustal profiles (Sapin and Prodehl, 1973; Fig. 7.2.1-02) revealed a rather homogeneous

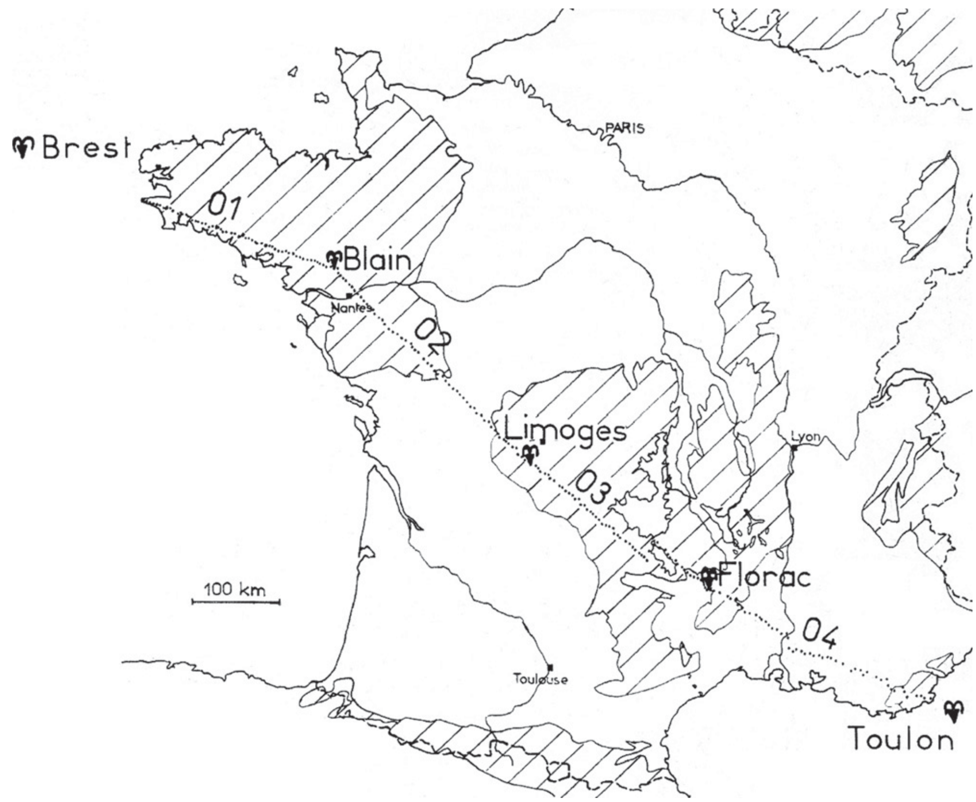


Figure 7.2.1-01. Location of shots and recording sites of the first long-range profile in 1971 through France. Dashed areas are basement outcrops (published also by Groupe Grands Profils Sismiques and German Research Group for Explosion Seismology, 1972, fig. 1). [Annales de Géophysique, v. 28, p. 247–256.]

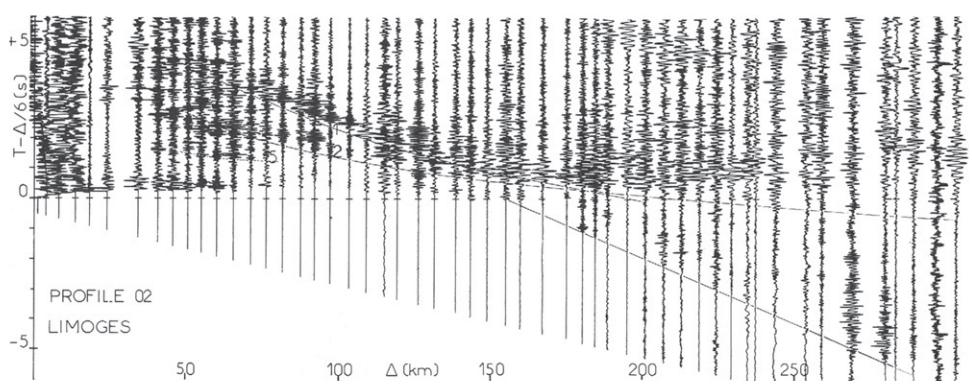


Figure 7.2.1-02. Record section of a crustal profile through France in 1971, data recorded from shotpoint Limoges along profile 02 toward northwest (published also by Sapin and Prodehl, 1973, fig. 4). [Annales de Géophysique, v. 29, p. 127–145.]

structure and a more or less flat Moho at 30–31 km depth, meaning that the mantle data could be interpreted without particular traveltime corrections.

At first glance, the first arrivals of the sea shot recordings beyond 300 km distance showed a rather uniform picture and the first-arrival energy propagated with an average velocity of ~8.1 km/s between 150 and more than 1000 km distance from the shotpoint. However, the close station separation of only 5 km revealed a fine structure as was known from Ansorge's previous investigations (Fig. 7.2.1-03).

The proper P_n phase with a velocity of 8.15 km/s disappeared beyond 210 km, and from 320 km onward, signals with strong amplitudes and a hardly noticeable delay in traveltime replaced the P_n phase.

A first interpretation of these mantle data correlated two mantle phases: P_I with 8.1–8.3 km/s between 320 and 480 km and P_{II} with 8.3–8.6 km/s between 450 and 620 km (Hirn et al., 1973). A reinterpretation of the same data shortly afterwards revealed a third upper-mantle phase P_{III} beyond 650 km distance (Kind, 1974). A similar interpretation as that of Kind (1974) was proposed by Ansorge (1975), but he adjusted the model by introducing first-order discontinuities to enable a direct comparison with the older less detailed long-range observations published earlier by Ansorge and Mueller (1971). On the basis of the 1971 long-range profile, Bottinga et al. (1973) searched the P_n-P_I range data of other long profiles obtained in France in 1970 and 1972. They detected a similar fine structure of the subcrustal lithosphere throughout France and speculated about its petrologic nature.

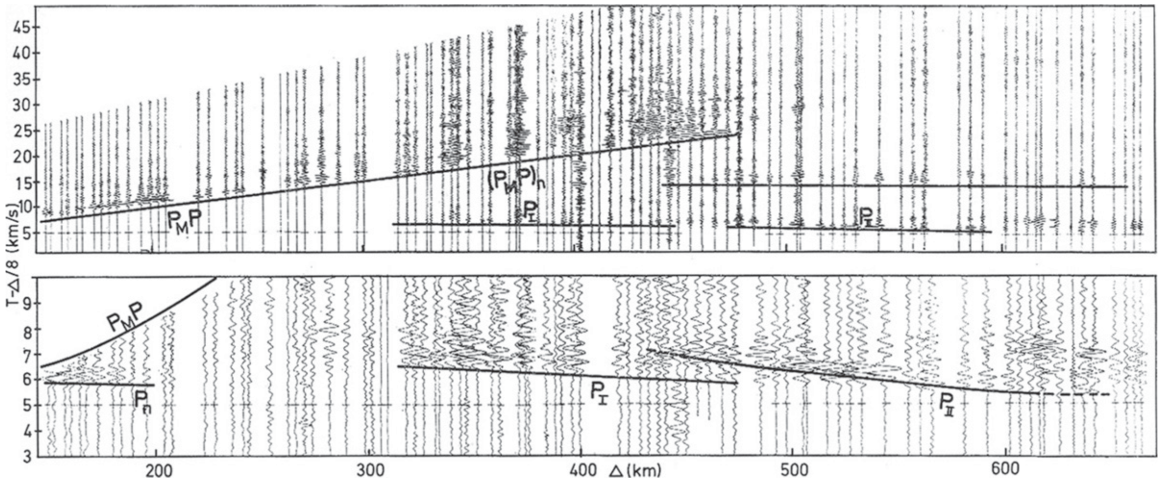


Figure 7.2.1-03. Record sections of long-range profiles through France in 1971 with different time and amplitude scales (from Hirn et al., 1973, fig. 3). P_MP reflection from Moho, P_n refracted head wave guided at Moho, P_I, P_{II} reflections from subcrustal interfaces at 55 km and 80 km depth. The lower section shows a clear separation of P_I and P_{II}, which is not clearly visible in the upper section. (See Fig. 7.2.1-06.) [Zeitschrift für Geophysik, v. 39, p. 363–384. Reproduced with kind permission of Springer Science+Business Media.]

In 1972, the long-range observations were extended to 1500 km distance. During the time of the Rhônegraben experiment jointly carried out by French, German, and Swiss institutions in southern France (see below in section 7.2.2) a large shot of 10 tons was organized off the coast of Scotland by B. Jacob and P.L. Willmore. They had already experimented with another 10-ton shot in the late 1960s and had successfully collected seismograms from permanent European stations which proved its energy efficiency (Jacob and Willmore, 1972). For this 10-ton shot off Scotland, the recording stations operating in France at that time were distributed throughout the French Massif Central over a range of 900 to 1500 km (Steinmetz et al., 1974). Simultaneously another profile was organized in western Germany, where stations recorded on a line into the Bavarian Forest between 800 and 1400 km distance (Bonjer et al., 1974). From the data recorded in France, Steinmetz et al. (1974) deduced a model to the base of the asthenosphere at 200 km depth.

In 1973, the experiment of 1971 was practically repeated (Hirn et al., 1975), but with slight variations. It was now known that the upper-mantle phases correlated in the 1971 data were not caused by near-surface irregularities at particular station locations, but that the traveltimes and amplitudes really varied with shotpoint distance, and thus they were directly dependent on the depth penetration of the corresponding P-waves. In the 1973 experiment, the first (northwestern) half of the 1971 line was reoccupied, i.e., for 650 km, and the shotpoint off Brest was shifted 95 km toward the coast (shotpoint Brest 2 in Fig. 7.2.1-04). In addition, fan profiles were set up in the distance range of 450–650 km where both P_I and P_{II} had been correlated and another 1-ton shot was fired at the same position as in the 1971 experiment (shotpoint Brest 1 in Fig. 7.2.1-04). Finally, a second profile was recorded 90 km to the north and parallel to the main line. The shotpoint Issoudun (Fig. 7.2.1-04) carried sufficient energy along the whole line, which was 500 km long. Unfortunately, the corresponding reversing sea shot failed.

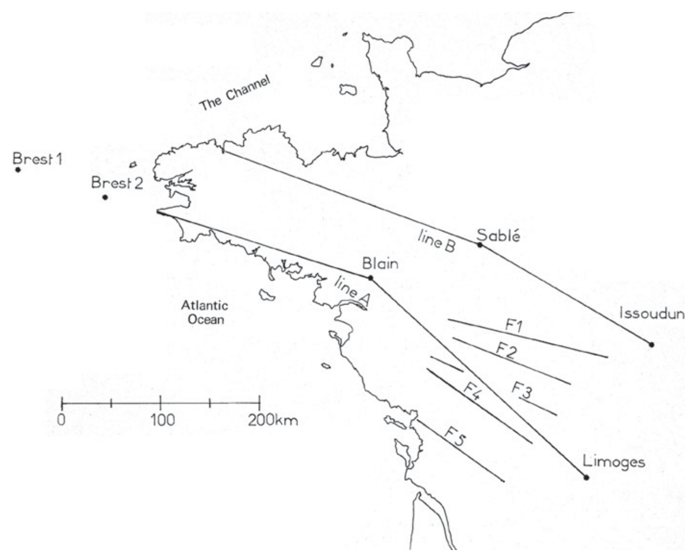


Figure 7.2.1-04. Location of upper-mantle profiles recorded in western France in 1971 and 1973. Shots at location Brest 1 were recorded on line A in 1971 and on fan profiles F1 to F5 in 1973, shot Brest 2 was recorded on line A in 1973. Blain, Limoges: borehole shots for reversed crustal profiles on line A, Issoudun, Sablé: borehole shots for upper mantle and crustal observations along line B in 1973 (published also by Hirn et al., 1975, fig. 1). [Annales de Géophysique, v. 31, p. 517–530.]

In all cases, a similar pattern of upper-mantle phases could be correlated (Figs. 7.2.1-03 and 7.2.1-05), proving that the model obtained for the sub-crustal lithosphere in 1971 is of more general nature. Figure 7.2.1-05 shows the observed fan data, and Figure 7.2.1-06 summarizes the results and also shows corresponding synthetic data calculated with the reflectivity method for the local

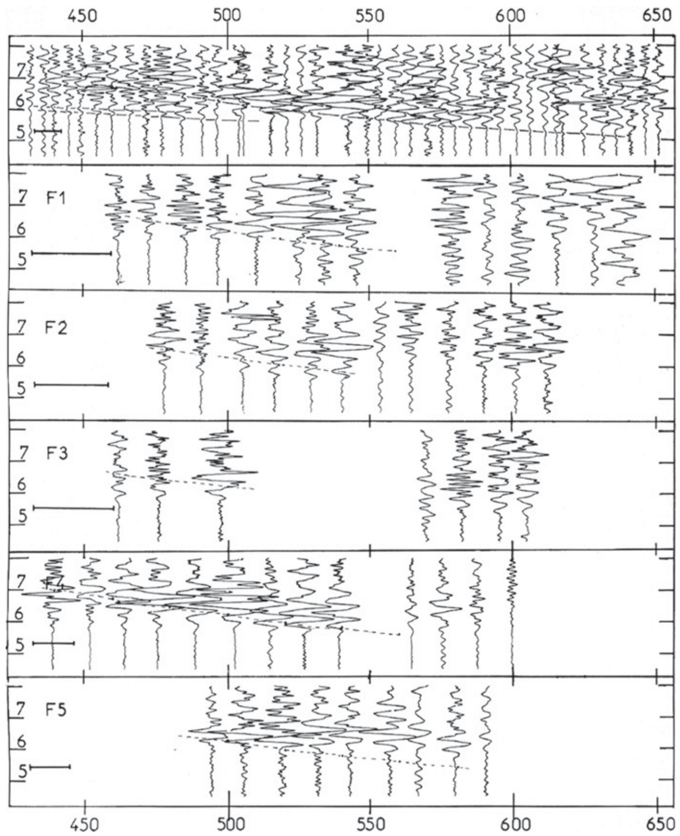
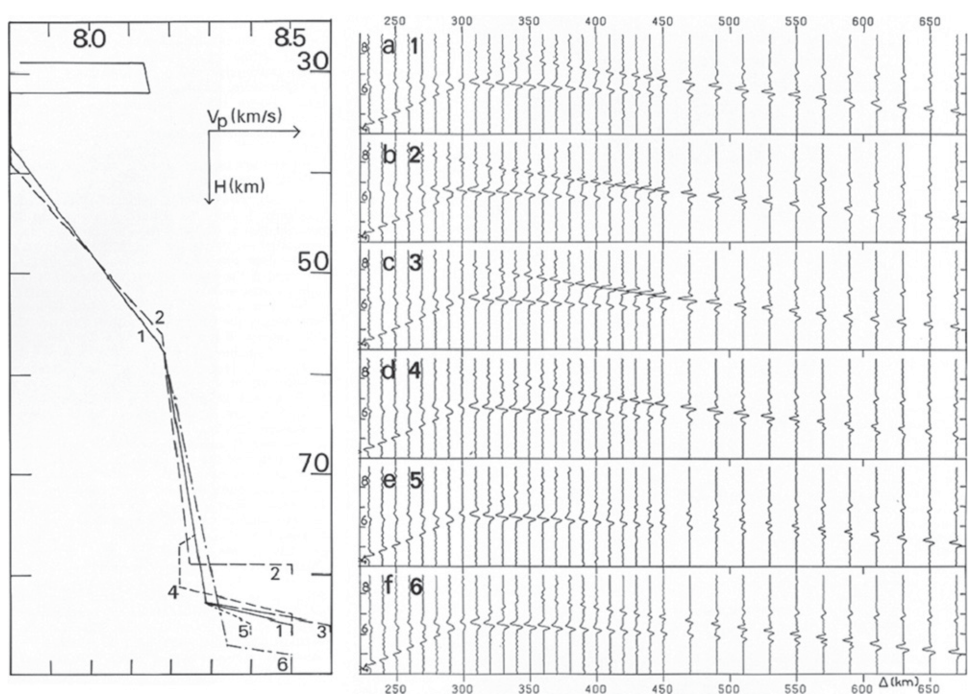


Figure 7.2.1-05. Observed data on the main profile Brest 1 (1971) and fan profiles F1 to F5 obtained in 1973 from a shot at the same position Brest 1 (published also by Hirn et al., 1975, fig. 4). Reduction velocity is 8 km/s. The dotted line on each section indicates the travel time curve correlated for P_{II} on Brest 1, line A (top section). Overall magnifications are uniform for all traces in one section, but note its variation from one section to another (0.5 µ/s amplitude bar in lower left corner). [Annales de Géophysique, v. 31, p. 517–530.]

Figure 7.2.1-06. Left: Velocity-depth models, corresponding to the long-range profiles in western France, show small possible variations in velocity and depth for slightly varying recording positions (published also by Hirn et al., 1975, fig. 7). Right: Synthetic record sections in the P_I, P_{II} distance range computed by the reflectivity method for the models to the left (from Hirn et al., 1975, fig. 8). Reduction velocity 8 km/s. Frequency of input signal 2.5–3 Hz is close to the main frequency of the observed data. [Annales de Géophysique, v. 31, p. 517–530.]



variations of the general model. Below the Moho, there is a distinct velocity inversion where the velocity drops from 8.15 km/s to ~7.8 km/s. Underlying the low-velocity zone, a reflector at 55–60 km depth causes the reflection P_I. At ~80–90 km, there is a second boundary producing phase P_{II}. The variations of some kilometers in depth for the reflectors or some tenths of km/s for the velocities are likely for the region (Hirn et al., 1975).

Compilations of many of these models and comparison with upper-mantle models worldwide down to 150 km depth, including Anderson's general model (Anderson, 1971), were prepared by Prodehl (Prodehl et al., 1976b; Prodehl, 1984). Figure 7.2.1-07 shows a diagram with models for western Europe as published until 1976.

7.2.2. Crustal Research of the Central European Rift System and the Variscan Basement

The first investigations of crustal structure of the Rhinegraben had already started in the mid-1960s but had exclusively been based on quarry blast observations. The results had been discussed and published by Mueller et al. (1967) and Ansorge et al. (1970) in two successive symposium volumes: Rhinegraben Progress Report 1967 (Rothé and Sauer, 1967) and Graben Problems (Illies and Mueller, 1970). The most famous result was the detection of a high-velocity rift cushion in the lowermost crust, located under the axis of the graben proper.

As mentioned above, with the new MARS-66 purchased in 1969, in 1970 the French group started a detailed investigation of the French Massif Central and its intervening rift structures, the Limagnegraben and the Bresse-Rhônegraben system (Perrier and Ruegg, 1973; Hirn and Perrier, 1974). At the same time, the German scientists at Karlsruhe started a major operation in Portugal (see below). Finally, from 1971 onwards, forces were put together, in particular pushed forward on the French side by A. Hirn, G. Perrier, E. Peterschmitt, J.P. Ruegg, and L. Steinmetz as well as on the German side by J. Ansorge, K. Fuchs, St. Mueller, and C. Prodehl.

Following the first highly successful joint effort of French and German scientists to unravel details of the subcrustal lithosphere under France by controlled-source seismic experiments as described above, the same group started a major crustal research of the Central European rift system with specially designed borehole shots. Two major experimental phases were carried out in 1972.

The first project was undertaken in May 1972 to investigate the Upper Rhinegraben in much detail (Rhinegraben Research Group for Explosion Seismology, 1974). For this project, borehole shots were planned along the axis of the graben. Due to logistical problems, these were arranged on the French side only, with the southernmost shotpoint being located near Steinbrunn (SB in Fig. 7.2.2-01), in the border region between the Rhinegraben and the Swiss Jura, and the northernmost shotpoint being close to the French-German border near Wissembourg (WI in Fig. 7.2.2-01).

In addition to the reversed axial profile, a series of fan profiles across the Vosges in the west and the Black Forest in the east was recorded (Rhinegraben Research Group for Explosion Seismology, 1974; solid lines in Fig. 7.2.2-01).

The data (reproduced in Appendix A2-1, p. 11–14) showed profound differences of phases to be correlated under the graben and under the flanks (Fig. 7.2.2-02), as was discussed in detail by Edel et al. (1975). As a result of the extended data set, the idea of a high-velocity lower-crust rift cushion was abandoned and replaced by a transitional crust-mantle boundary centered under the rift. Under the graben flanks the Moho appeared as a first-order discontinuity (Fig. 7.2.2-03).

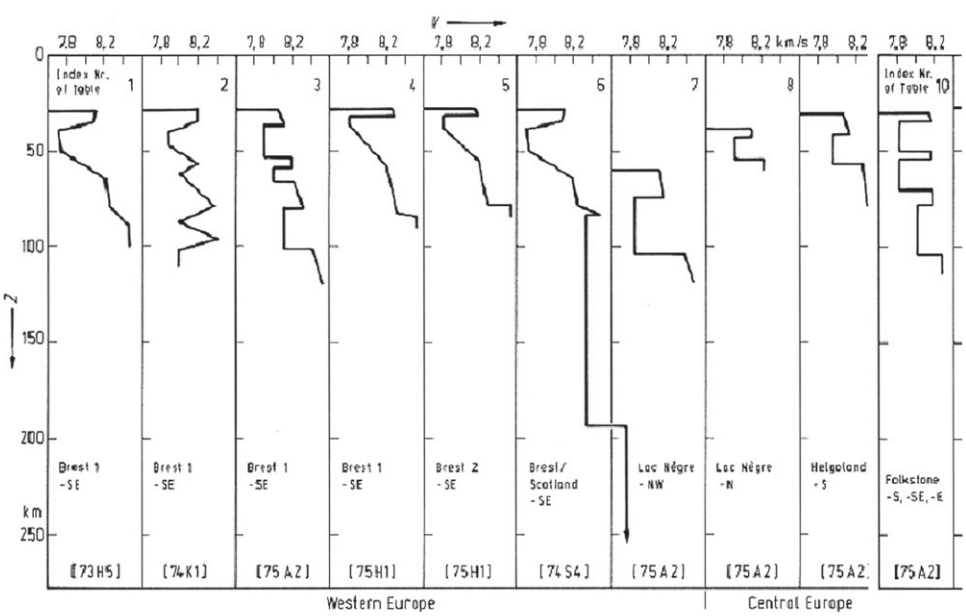


Figure 7.2.1-07. Velocity-depth models of the subcrustal lithosphere in western Europe for profiles recorded until 1973 (from Prodehl, 1984, fig. 15). 73H5—Hirn et al. (1973); 74K1—Kind (1974); 75A2—Ansorge (1975); 75H1—Hirn et al. (1975); 74S4—Steinmetz et al. (1974). [*In* K.-H. Hellwege, editor in chief, Landolt Börnstein New Series: Numerical data and functional relationships in science and technology. Group V, Volume 2a: K. Fuchs and H. Soffel, eds., Physical properties of the interior of the earth, the moon and the planets: Springer, Berlin-Heidelberg, p. 97–206. Reproduced with kind permission of Springer Science+Business Media.]

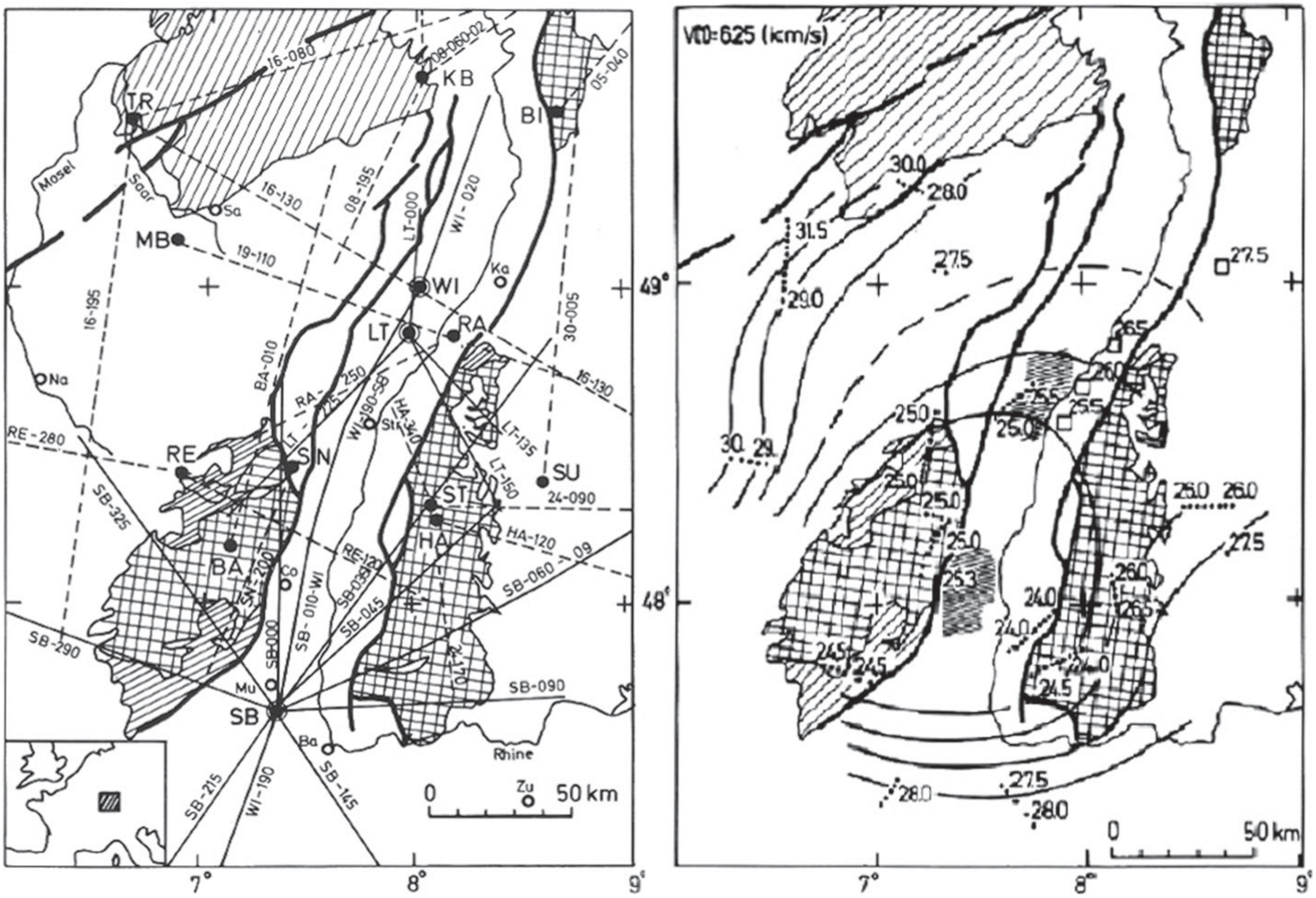


Figure 7.2.2-01. Left: Location map of seismic-refraction profiles in the surroundings of the southern Upper Rhine Graben (from Prodehl et al., 1976, fig. 3). Right: Moho contour map of the southern Upper Rhinegraben (from Prodehl et al., 1976, fig. 12). [*In* P. Giese, C. Prodehl, and A. Stein, eds., Explosion seismology in central Europe—data and results: Springer, Berlin-Heidelberg-New York, p. 313–331. Reproduced with kind permission of Springer Science+Business Media.]

The new data furthermore evidenced a broad warping in an upward direction of the Moho by up to 5 km and a normal P_n velocity of ~8 km/s (Edel et al., 1975; Prodehl et al., 1976a, 1995). The interpretation of Edel et al. (1975) incorporated both old and new data, which were all published in small-scale format.

During the Rhônegraben experiment carried out in July 1972 the international working force of French and German institutions was joined by Swiss and Portuguese participants and their equipment (Sapin and Hirn, 1974; Michel, 1978; Prodehl et al., 1995).

This survey did not only consist of a reversed profile along the rift axis, but also supplied reversed lines along the flanks to the west in the Massif Central and to the east in the sub-Alpine region. Furthermore intermediate shots were added in the middle of all lines, and transverse profiles across the Rhône valley were arranged. The data in the southern Rhône Valley was further complemented by the crustal data of the NW-SE long-range profile of 1971 (Fig. 7.2.2-04). This large amount of data was of extremely high quality and was split into a southern part (Sapin and Hirn, 1974) and a northern part (Michel, 1978). Most of the data were presented as record sections and are reproduced in Appendix A2-1 (p. 38–48).

7.2.3. The Mediterranean Region

Stimulated by German colleagues, the Portuguese scientists became interested in a detailed crustal research of Portugal. By 1970, the Portuguese community had acquired enough funding to buy MARS-66 equipment. Then, in 1970 and 1972, in close cooperation with the University of Karlsruhe, Germany, a concentrated effort resulted in a number of seismic-refraction profiles (Fig. 7.2.3-01) across the Variscan part of southern Portugal (Mueller et al., 1973, 1974; Moreira et al., 1977). In 1971, a line was added north of Cabo da Roca and reversed from Nazaré (Fig. 7.2.3-01). All shotpoints were at sea, which was enabled by the cooperation with a Portuguese navy vessel and a shooting crew of the German navy (Prodehl et al., 1975).

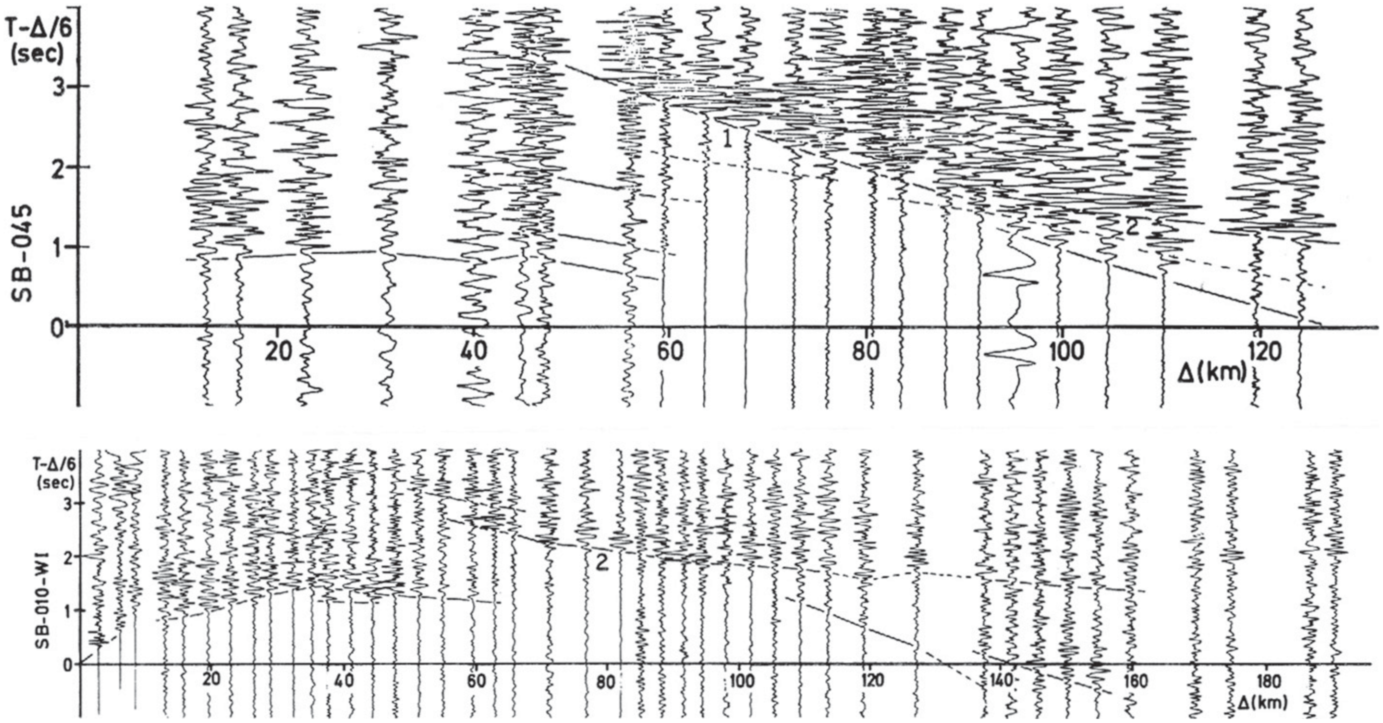


Figure 7.2.2-02. Record sections of the Rhinegraben area of two profiles from Steinbrunn . Top: Profile SB-045 across the flank of the graben, phase 1 = PMP, phase 2 = reflection from mid-crust boundary (from Prodehl et al., 1977, fig. 12). Bottom: Profile SB-010-WI along the graben axis, only phase 2 is visible (from Prodehl et al., 1977, fig. 13). [*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.]

The profiles were arranged so that all profiles, except the SW-NE line of 1972, could be reversed (Fig. 7.2.3-01). All data, recorded in Portugal from 1970 to 1972, were published as record sections (reproduced in Appendix A2-1, p. 55).

Two years later, from 1974 to 1975 (Working Group for Deep Seismic Sounding in Spain 1974–1975, 1977; Appendix A7-2-1), and continued later in 1977 and 1980 (Barranco et al., 1990; Appendix A7-2-2), the first deep seismic sounding experiments followed in southern Spain by the cooperation of Spanish, Swiss, Portuguese, French, and German research institutions and the Spanish navy. Profiles crossed the area of the Betic Cordillera with shotpoints both at sea and on land (Banda and Ansorge, 1980). The interpretation resulted in drastically varying crustal thickness, from near 25 km near the southeastern coast at Cartagena and near the southern coast near Adra to as much as 39 km under the central Cordillera. The experiments included a long-range line of 450 km length between Cartagena in the east and Cadiz in the west (Fig. 7.2.3-02). Record sections and detailed maps for these profiles and all following experiments in the Mediterranean area are reproduced in Appendix A7-2.

These first seismic investigations of southern Spain were soon followed by quarry blast profiling on the Meseta (Banda et al., 1981b; Appendix A7-2-3) and by deep seismic sounding experiments in eastern Spain in 1976 and 1981 (Zeyen et al., 1985; Appendix A7-2-4), resulting in 31 km crustal thickness under the Meseta, which near the southeastern coast gradually decreased to as low as 20 km thickness. Land-sea operations around both the Balearic islands in 1976 (Banda et al., 1980; Fig. 7.2.3-02; Appendix A7-2-5) and the Canary islands in 1977 and 1979 (Banda et al., 1981a; Appendix A7-2-6) gave average crustal thicknesses of 20 km under the Balearic Islands and 12–15 km under the Canary Islands.

In close cooperation with the French group, a detailed investigation of the Pyrenees was performed in 1978 with two E-W lines parallel to the strike of the mountain range (Fig. 7.2.3-02) and some additional lines into the northern and southern forelands (Appendix A7-2-7). Under the Pyrenees, for example, an abrupt thinning by 10 km from the Paleozoic axial zone with a 40–50-km-thick crust toward the North Pyrenean zone with 25–30 km crustal thickness was detected (Explosion Seismology Group Pyrenees, 1980; Gallart et al., 1980; Hirn et al., 1980; Daignières et al., 1982).

On the African side, in Morocco, the structure of the Moroccan Meseta, the High Atlas and the Anti-Atlas was the goal of a major seismic-refraction campaign in 1975 organized by J. Makris of the University of Hamburg, Germany, and A. Demnati of the Geological Survey of Morocco (Fig. 7.2.3-03; see also Fig. 8.3.4-17).

The seismic energy was generated by quarry blasts at several mines and at sea by shots fired from the German research vessel

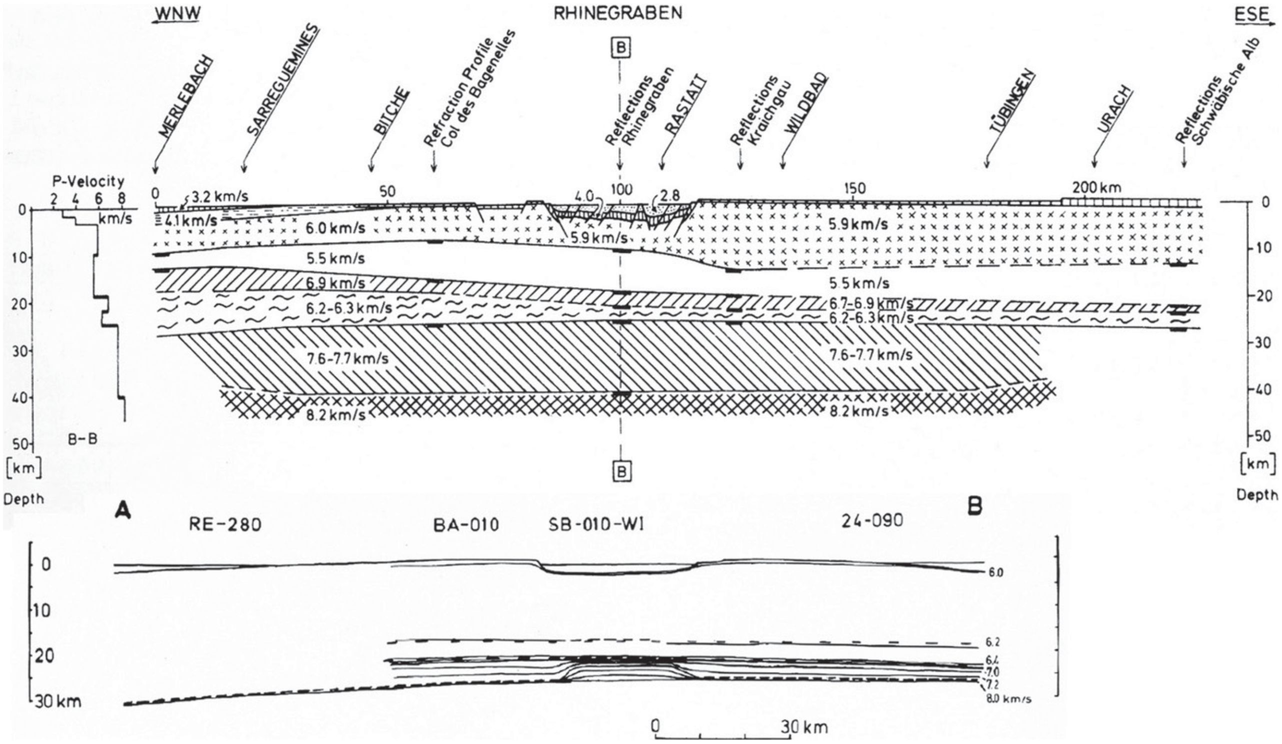


Figure 7.2.2-03. Crustal cross sections through the central part of the Rhinegraben area. Top: The depth range of 25–40 km is interpreted as a rift cushion, i.e. a high-velocity lower crust after Mueller et al., 1969 (from Prodehl et al., 1976a, fig. 6). Bottom: The Moho rises gradually from 30 km west of the graben to about 25 km east of the graben, but becomes a transitional gradient zone under the graben proper, with its top upwarping to less than 25 km depth (from Prodehl et al., 1976a, fig. 14, upper section). Thin lines show equal velocity contours, interval is 0.2 km/s. [*In* P. Giese, C. Prodehl, and A. Stein, eds., Explosion seismology in central Europe—data and results: Springer, Berlin-Heidelberg-New York, p. 313–331. Reproduced with kind permission of Springer Science+Business Media.]

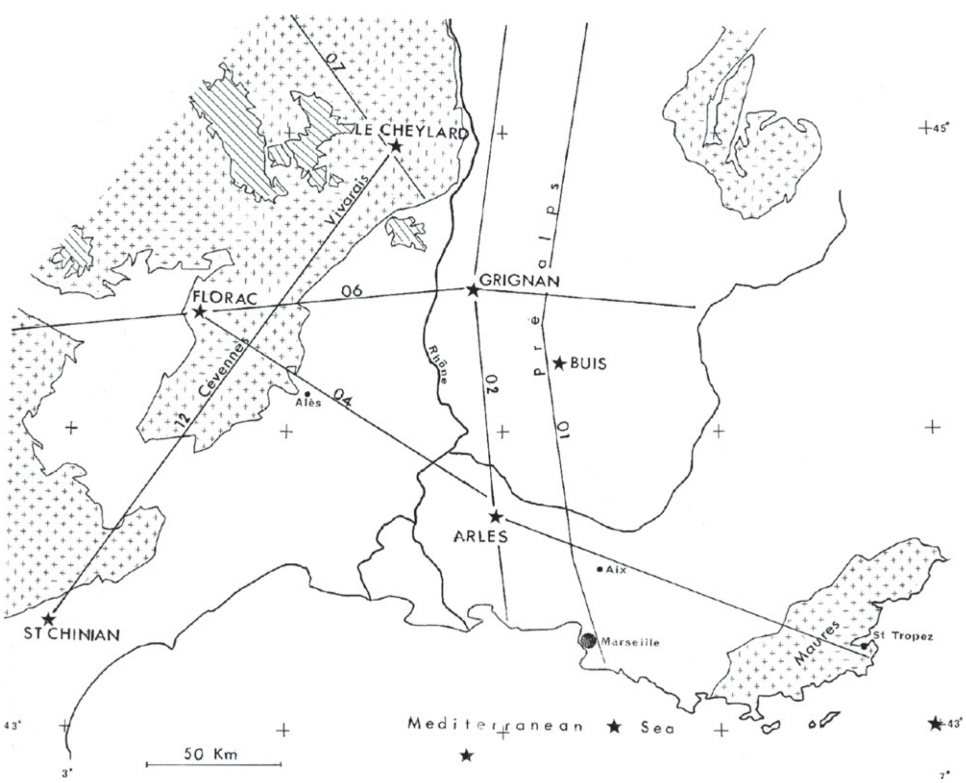


Figure 7.2.2-04. Location map of seismic-refraction profiles in the area of the southern Rhône valley (from Sapin and Hirn, 1974, fig. 1). [Annales de Géophysique, v. 30, p. 181–202. Reproduced by permission of the author.]

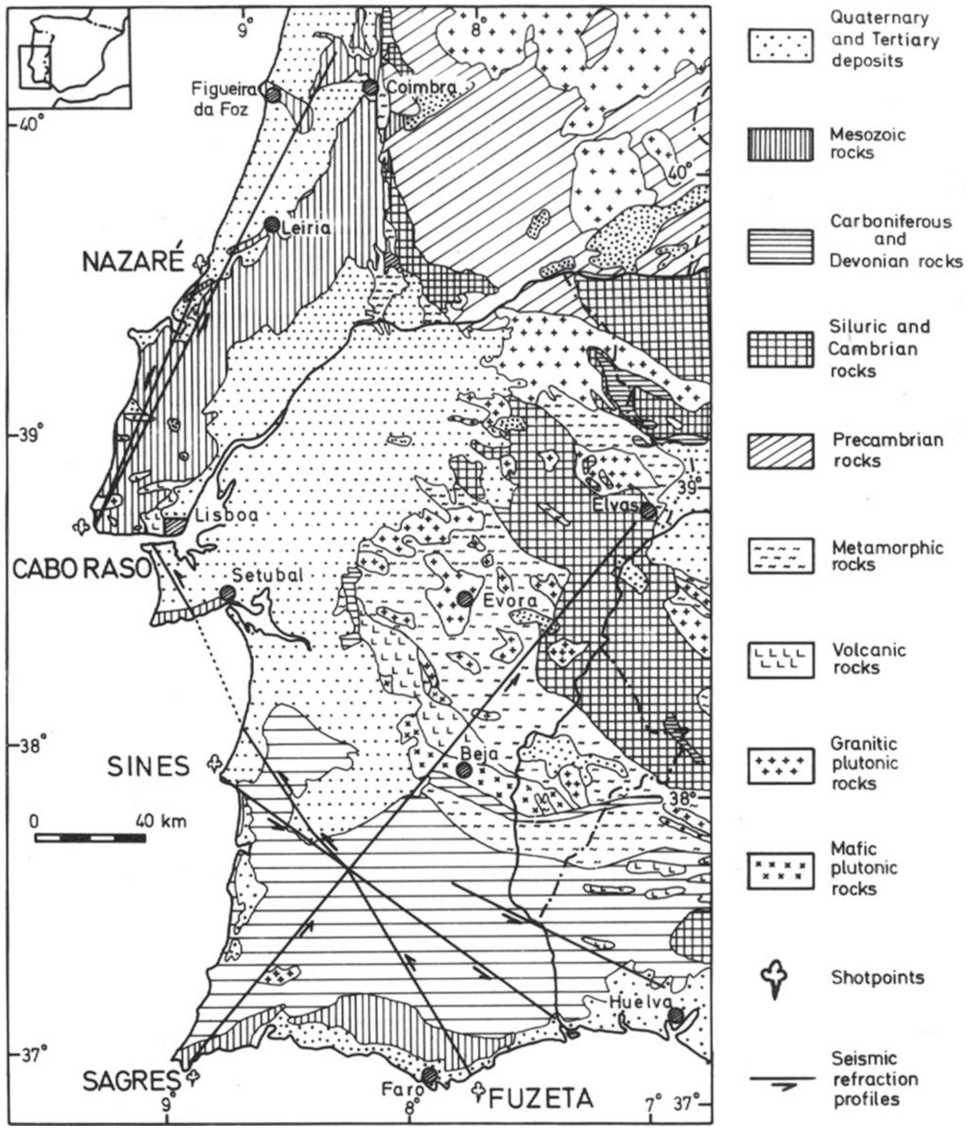


Figure 7.2.3-01. Crustal profiles in SW Portugal in 1970 to 1972 (from Prodehl et al., 1975, fig. 1). Proc. 14th General Assembly of the European Seismological Committee (Trieste, 1974) [Akad. Wiss. DDR, Berlin, p. 235–241.]

Meteor during cruise M39 (Sarnthein et al., 2008) between the Canary Islands and the Moroccan coast (Appendix A7-2-8). The resulting model shows a gradual thickening of the crust from 24 km near the coast, to 30 km under the Meseta and the Anti-Atlas, and to as deep as 38 km under the High Atlas. At the continent-ocean transition, a deep reflector below the Moho was identified at depth ranges between 60 and 80 km (Makris et al., 1985).

Mainly under the heading of the Trieste group, the Western Mediterranean was investigated by several sea-seismic and onshore-offshore campaigns (Hirn et al., 1977; Leenhardt et al., 1972; Morelli et al., 1977; Morelli and Nicolich, 1980; Steinmetz et al., 1983). In 1970, a joint cruise—then called the *Anna* cruise—was undertaken by French and German marine geophysicists, employing both reflection and refraction measurements along three lines located along a north-south–directed traverse from the Rhône delta across the Balearic islands to the African coast near Algiers (Leenhardt et al., 1972; Appendix A7-2-9). Profile 1, located in the Gulf of Lyon, was only recorded to 25 km distance. Profiles 2 and 3 were north-south lines, located north and south of the island of Mallorca (Fig. 7.2.3-04). They were successfully recorded up to 100 km distances and resulted in a crustal thicknesses of 14 km north (line 2) and 12 km south of Mallorca (Hinz, 1972).

In 1974, a large seismic-refraction program was carried out as a joint project of Italian, German, and French geophysical research institutions (Morelli et al., 1977; Appendix A7-2-10). It covered the Northern Apennines, the Ligurian Sea and Corsica. Only sea shots of 200 kg charges were involved fired at optimum depth at 90–100 m to obtain constructive interferences of the bubble pulse with the surface reflections (Wielandt, 1972, 1975). The shots were arranged on lines in the Corsica–Elba Channel, the Ligurian Sea, and the Provençal Basin and recorded on

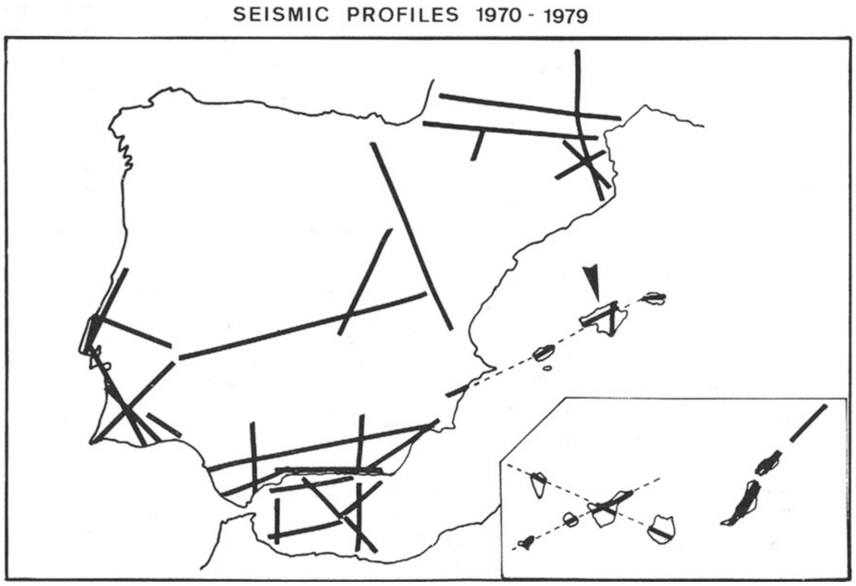


Figure 7.2.3-02. Location of deep seismic-sounding experiments on the Iberian Peninsula. Seismic lines in Portugal and Spain from 1970 to 1979 (modified from Cordoba and Banda, fig. 1). [Geofisica Internacional, Mexico, v. 19, p. 285–303.]

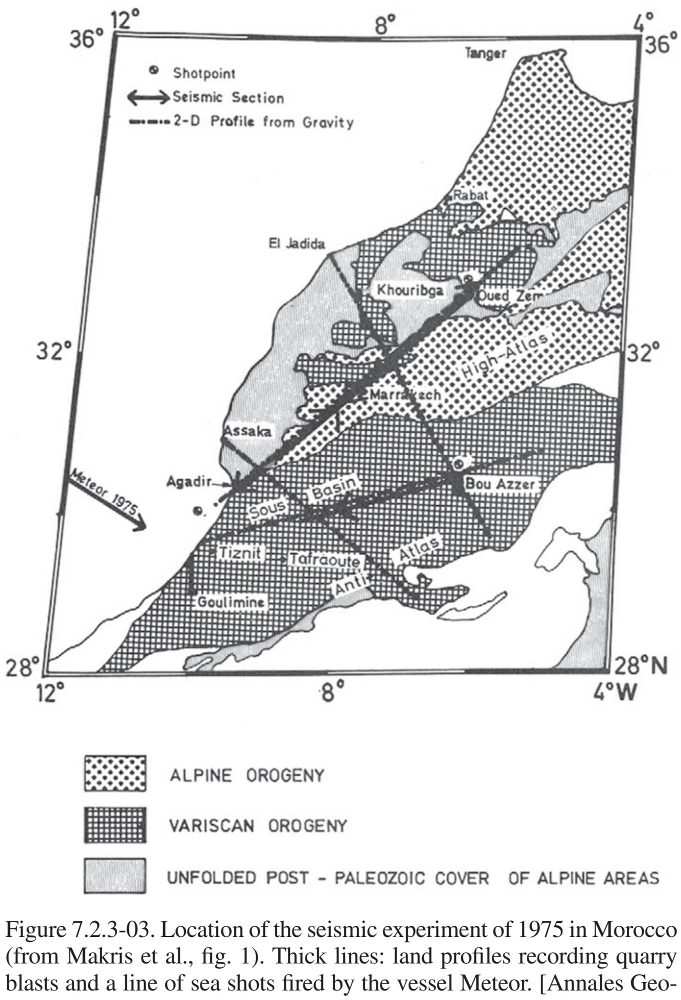


Figure 7.2.3-03. Location of the seismic experiment of 1975 in Morocco (from Makris et al., fig. 1). Thick lines: land profiles recording quarry blasts and a line of sea shots fired by the vessel Meteor. [Annales Geophysicae, 3, p. 369–380. Reproduced by permission of the author.]

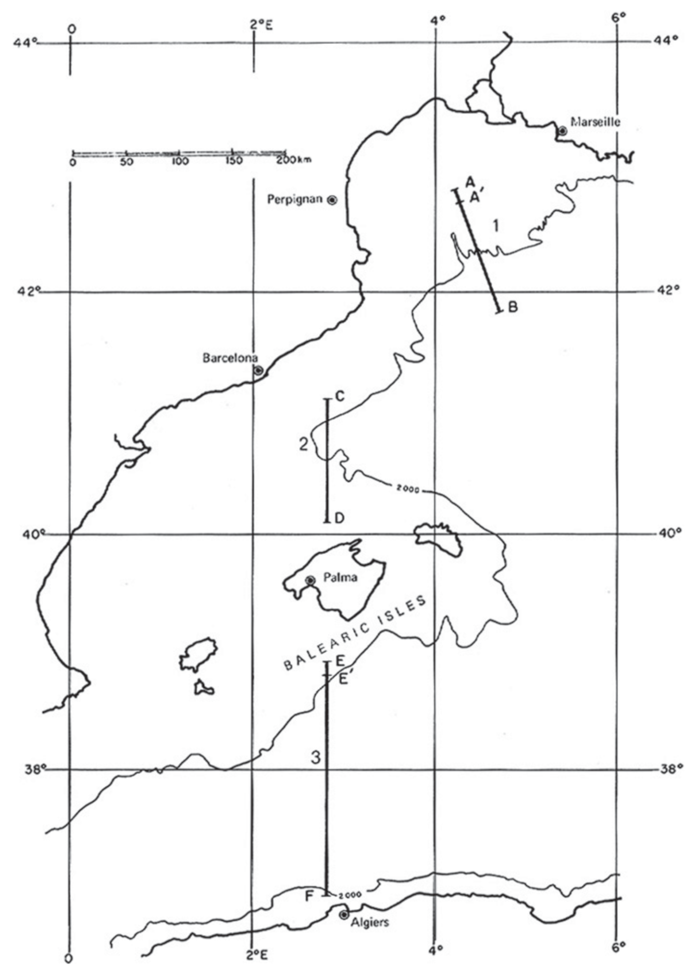


Figure 7.2.3-04. Location of Project Anna 1970 deep seismic-sounding experiments in the Ligurian Sea and surrounding land areas (from Leenhardt et al., 1972, Organization, fig. 1). [Bulletin of the Centre Recherche Pau–SNPA, v. 6, p. 373–381. Reproduced by permission of TOTAL–CSTJF.]

Corsica and along four profiles across the Apennines in northern and central Italy (Fig. 7.2.3-05).

For the interpretation of the data (data example in Fig. 7.2.3-06), various author groups have contributed to the publication of Morelli et al. (1977). The crust underneath the island of Corsica with 30 km thickness proved to be typically continental, updipping rapidly toward the west, thus coinciding with the earlier results of an 11–13-km-thick crust under the Provençal basin (Hirn and Sapin, 1976; Appendix A7-2-11). Toward northeast, the crustal thickness appeared to be at least 20 km under the island of Elba and 15–20 km under the northern Ligurian Sea, thickening to at least 35 km under the crest of the Apennines (Morelli et al., 1977).

This campaign was followed by two long-range experiments performed in the western Mediterranean (Appendix A7-2-12). The first was a 430-km-long line between the Balearic Islands and Corsica. It was recorded in 1974 and consisted of a line of shots spaced ~8–10 km, which were recorded by land stations both in NW Corsica and on the islands of Menorca and Mallorca. Unfortunately, the only ocean-bottom sensor, a prototype, failed, and shots started at only 70 km from the nearest land station. The result was a 6-km-thick low-velocity zone with 7.7 km/s beneath a Moho lid (8.25 km/s) at 28 km depth, followed by a 20-km-thick layer with 8.3 km/s, underlain by 8.7 km/s mantle material (Hirn et al., 1977). The second long-range profile was a 550-km-long traverse, recorded in 1979 in the Tyrrhenian Sea between Corsica and southern Italy. It involved recently developed ocean-bottom seismometers and thus enabled the investigation of gross features of the upper mantle (Steinmetz et al., 1983).

In southern Italy, in the early 1970s, profiles had been recorded along Puglia and Calabria with shots recorded from the adjacent Tyrrhenian Sea. In 1971, two profiles were recorded from shotpoint C (Fig. 7.2.3-07). One line ran along Puglia and was reversed from shotpoint B; the other line was unreversed and was recorded toward Napoli (Giese et al., 1973). Colombi et al. (1973) have described in much detail a second experiment of 1971, where recording stations in southern Italy, in Sicily, and on the island of Pantelleria recorded shots in the Tyrrhenian Sea (line E in Fig. 7.2.3-07) and in the Channel of Sicily (line F in

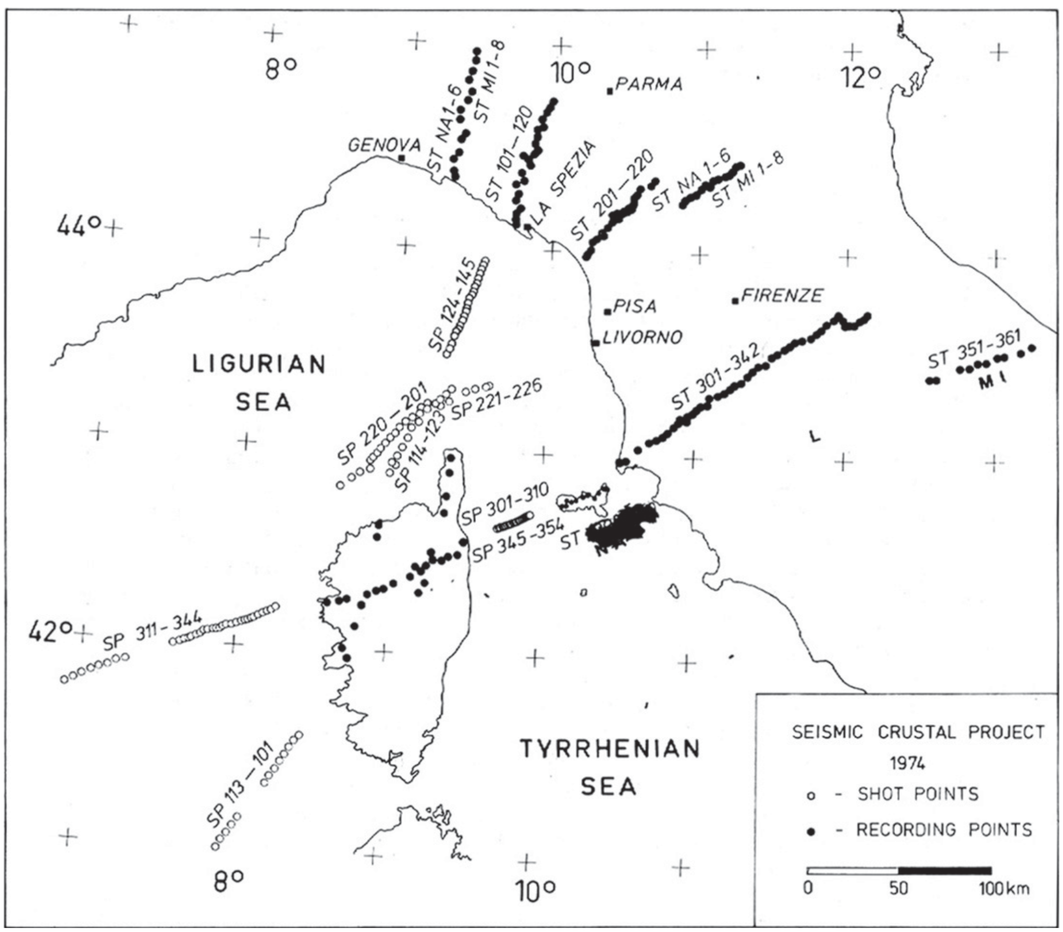


Figure 7.2.3-05. Location of deep seismic-sounding experiments in the Ligurian Sea and surrounding land areas (from Morelli et al., 1977, fig. 1.2). [Bollettino di Geofisica Teorica e Applicata, v. 7, p. 199–260. Reproduced by permission of Bollettino di Geofisica, Trieste, Italy.]

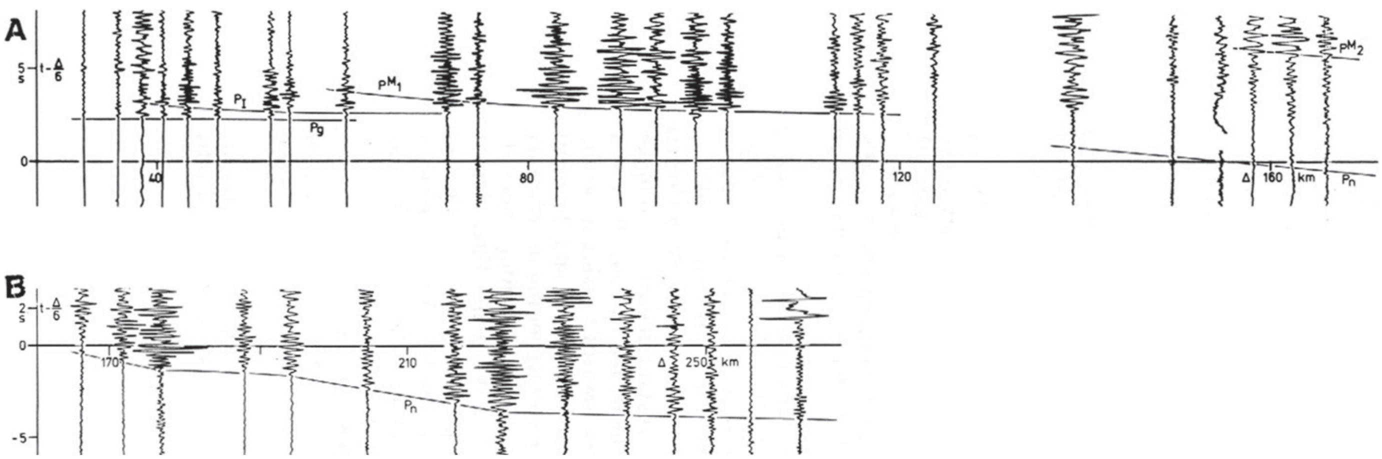


Figure 7.2.3-06. Data example recording along line 3 from the Ligurian Sea into the Appennines (from Morelli et al., 1977, fig. 4.6). [Bollettino di Geofisica Teorica e Applicata, v. 7, p. 199–260. Reproduced by permission of Bollettino di Geofisica, Trieste, Italy.]

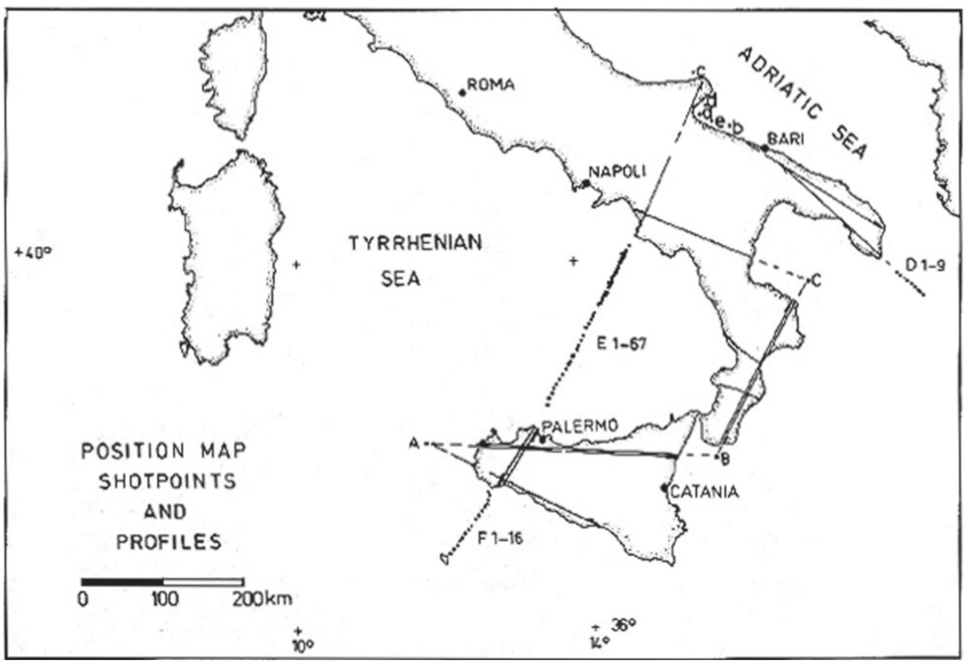


Figure 7.2.3-07. Location map of shotpoints, refraction lines and fans in southern Italy (from Giese et al., 1973, fig. 8). [Tectonophysics, v. 20, p. 367–379. Copyright Elsevier.]

Fig. 7.2.3-07). Record sections, maps, and models are reprinted in Appendix A7-2-13.

Combining all observations of the late 1960s (Cassinis et al., 1969) and early 1970s (Colombi et al., 1973), Giese et al. (1973) compiled a Moho depth map for the area of southern Italy, Sicily, and the adjacent Tyrrhenian Sea (Fig. 7.2.3-08) showing crustal thickness variations from oceanic to continental crust with less than 10 km in the center of the Tyrrhenian Sea to 35 km underneath central Sicily, Calabria, and Puglia.

By the end of the 1970s and beginning of 1980s, Italy was densely covered with seismic-refraction profiles (Morelli, 2003; Fig. 7.2.3-09). Most of the data had been acquired by the Osservatorio Geofisico Sperimentale (OGS) under the leadership of Prof. Carlo Morelli, by whose influence many international collaborations, e.g., with Prof. Peter Giese, Berlin, or Dr. A. Hirn, Paris, helped to fulfill this goal. In the years from 1956 to 1982 ~21,160 km of seismic profiles had been obtained. Many of the deep seismic sounding profiles were later digitized and reinterpreted by the Istituto di Ricerca del Rischio Sismico CNR, Milano, and were made available to the scientific community (Scarascia et al., 1994). Similar activities of OGS at sea—much thanks to the driving forces of C. Morelli and R. Nicolich—had accumulated a wealth of marine data in the Mediterranean Sea. In the period of 1969–1982, 39,500 km of regional marine seismic-reflection profiles were recorded by the CNR vessel *Marsili*, equipped with a seismic recorder DFS, a neutral buoyant streamer (24 traces,

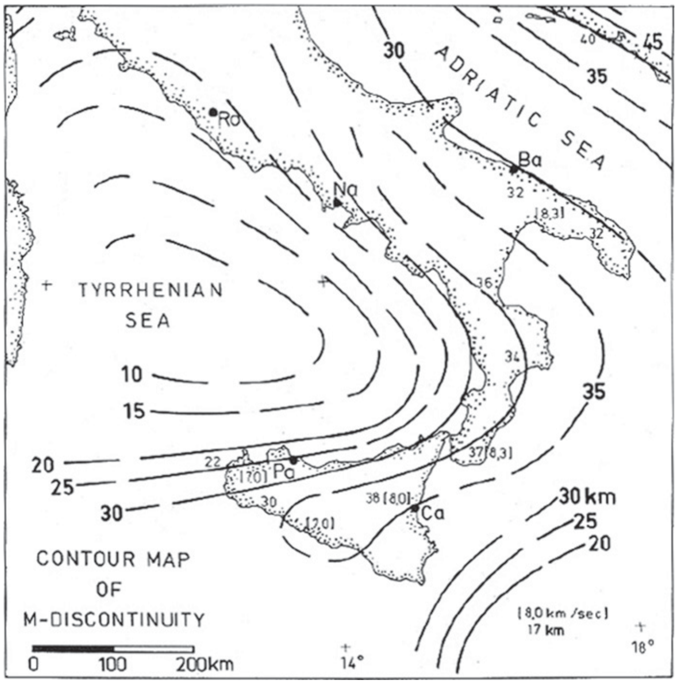


Figure 7.2.3-08. Crust-mantle boundary in southern Italy and surrounding seas, defined as depth of the strongest gradient in the velocity interval from 7.2–8.4 km/s (from Giese et al., 1973, fig. 9). [Tectonophysics, v. 20, p. 367–379. Copyright Elsevier.]

Figure 7.2.3-09. Deep seismic sounding profiles acquired from 1956 to 1982 in Italy by OGS, Trieste (from Morelli, 2003, fig. 1) [*In* Scrocca et al., eds., 2003, CROP Atlas: seismic reflection profiles of the Italian crust: Memorie Descrittive della Carta Geologica d'Italia, Roma, v. 62, p. 194 p. Copyright CROP.]

2400 m long) and two IFP Flexotir guns. Besides detailed sections for the sedimentary strata, also initial information on the crystalline parts of the crust were obtained in all areas (Finetti and Morelli, 1973). Other surveys added to the wealth of data. For example, the deep structure of the Ionian Sea was one of the goals to record a network of seismic-reflection profiles during cruise M50 of the German vessel *Meteor* (64) (Avedik and Hieke, 1981).

The first detailed investigation in Greece and the surrounding maritime areas started in 1969 and 1971 with two onshore-offshore experiments from the Ionian Sea into Greece. In 1969, three seismic land stations near the coast of the Peloponnesus had recorded shots during the *Meteor* cruise M17. A second offshore survey followed in 1971 during the cruise M22 of the German research vessel *Meteor* (Weigel, 1974). This investigation included a series of large shots recorded by four sonobuoy stations and 25 stations on land across the Peloponnesian Peninsula by a number of German institutions headed by H. Berckhemer and J. Makris (Fig. 7.2.3-10). The profile line crossed the Mediterranean Ridge and the Hellenic Trench at sea and ended on land in the east near Aegina on the Peloponnesus (Weigel, 1974). The Moho was not clearly seen along the offshore line, but was inferred from gravity. Under the coast of the Peloponnesus, the Moho reached a depth of ~46 km; farther west, the crust-mantle transition zone, though not clearly marked by a discontinuity, was interpreted to rise to 18 km depth under the western part of the Mediterranean Ridge and the Ionian Abyssal Plain. Under the Malta marginal trough, the Moho was found at 17 km depth (Weigel, 1974).

The island of Crete was the goal of an axial east-west profile in 1973. A NW-SE line recorded in 1973 and 1974 was arranged from central Greece in the northwest across the island of Evia toward a series of smaller islands in the Aegean Sea in the southeast (Makris and Vees, 1977). Another survey of the German research vessel *Meteor* in 1974 concentrated on the substructure of the Cretan Sea north of the island of Crete (Dürbaum et al., 1977). Record sections, maps, and models of all projects in and around Greece are reprinted in Appendix A7-2-14.

Summarizing contributions by Makris (1976, 1985) show the results of these seismic (Fig. 7.2.3-10) and other geophysical surveys and discuss the geodynamic implications for the evolution of the Hellenides. The Moho map for Greece and surroundings (Fig. 7.2.3-11), constructed by Makris (1977, 1985), using seismic and gravity data, demonstrates the complicated crustal structure of the Eastern Mediterranean area.

7.2.4 Fine Structure of the Lower Lithosphere—The Second Stage of Long-Range Profiling

In 1973, a young scientist, David Bamford from Birmingham, received a grant to work for one year at the University of Karlsruhe. This visit was the starting point of a very active cooperation

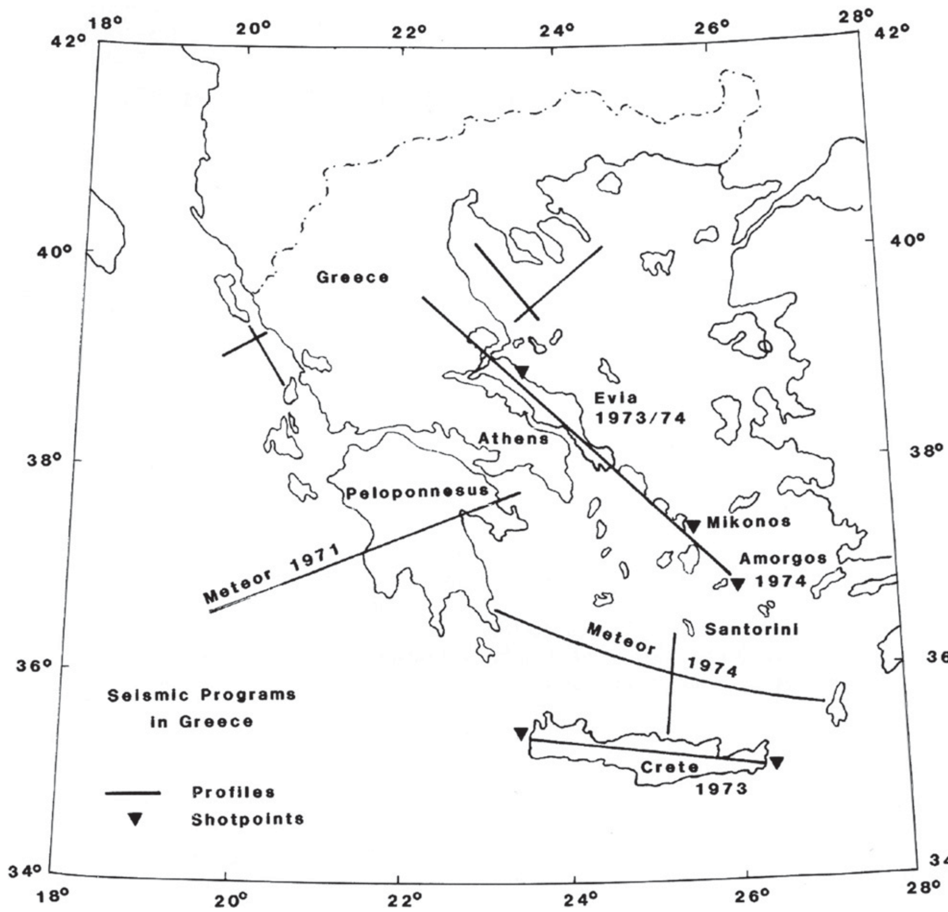


Figure 7.2.3-10. Location of deep seismic-sounding experiments in Greece and surrounding maritime areas (from Makris, 1985, fig. 11.1). [*In* Stanley, J., and Wezel, F-C, eds., Geological evolution of the Mediterranean Basin: Springer, New York-Berlin, p. 231–248. Reproduced with kind permission of Springer Science+Business Media.]

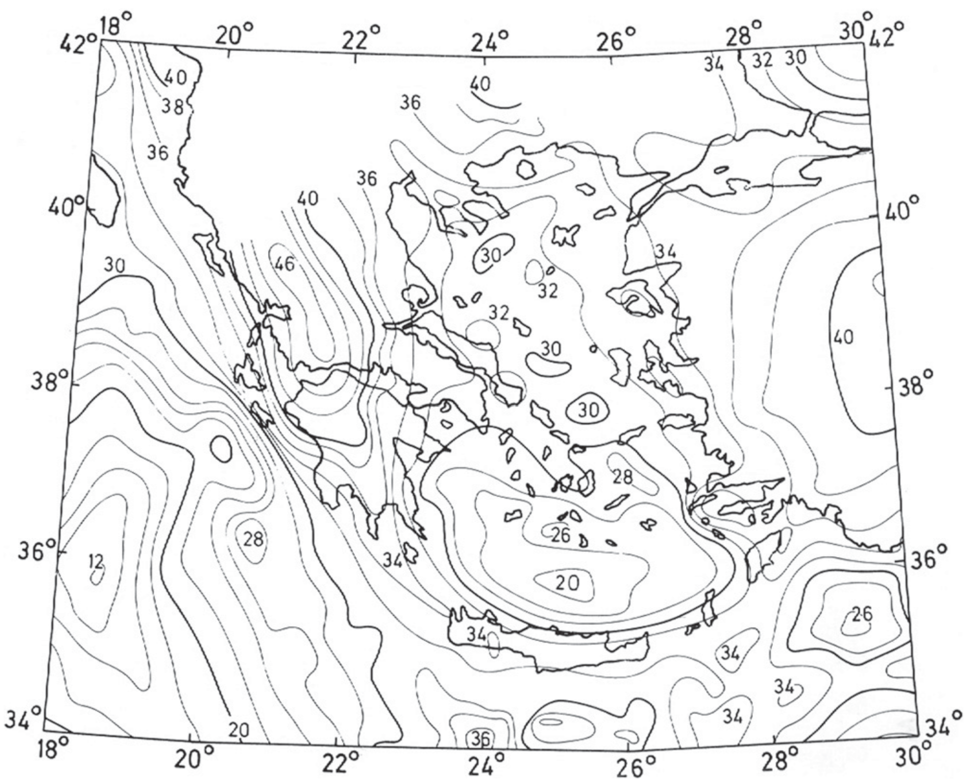


Figure 7.2.3-11. Moho map of Greece and surrounding maritime areas (from Makris, 1985, fig. 11.15). [*In* Stanley, J., and Wezel, F-C, eds., Geological evolution of the Mediterranean Basin: Springer, New York-Berlin, p. 231–248. Reproduced with kind permission of Springer Science+Business Media.]

between British and German scientists for the next 25 years and almost immediately stimulated two exciting projects.

The first project was the research for which the grant had been given: a time term analysis of the P_n data of the dense German network of quarry blast observations including the Rhinegraben (Fig. 7.2.4-01). This project resulted in an unexpected finding, namely the largest continental anisotropy of more than 3% in the uppermost mantle, with the fast velocity in the NNE direction and the slow velocity in the WNW direction (Bamford, 1973; Fig. 7.2.4-01). Bamford's results stimulated K. Fuchs to discuss the impact of the seismic uppermost-mantle anisotropy with respect to the petrological composition of the subcrustal mantle material (Fuchs, 1975, 1977) and to construct his famous "anvil" model (Fuchs, 1983; Fig. 7.2.4-02).

Some years later, D. Bamford, in cooperation with his Karlsruhe colleagues, extended his research to other continental areas where rather dense networks of seismic profiles existed. In northern England and in the eastern United States, the uppermost mantle appeared to be isotropic within the limits of measurement error. However, a similar result as that for southern Germany

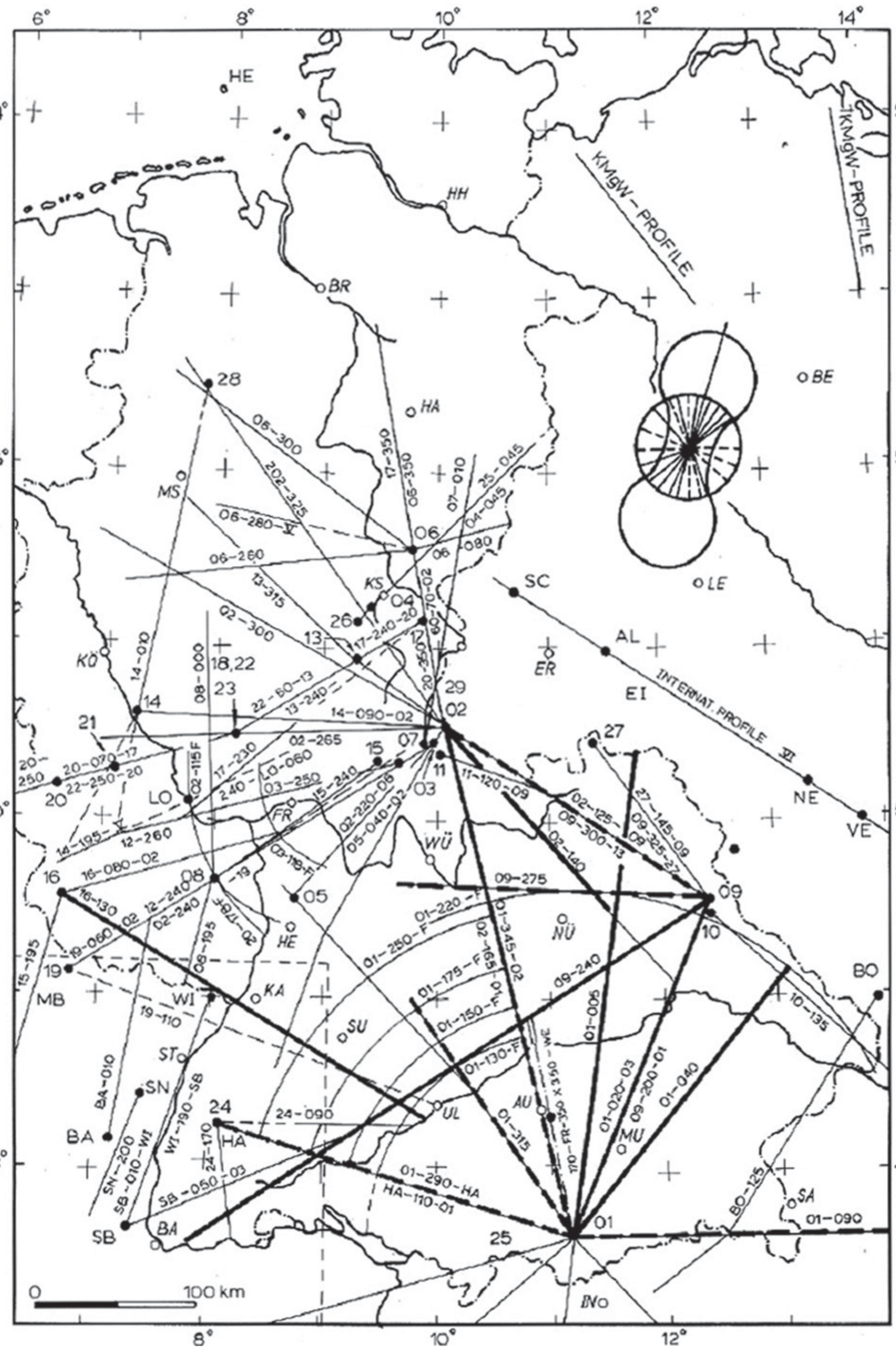


Figure 7.2.4-01. Azimuthal distribution of P_n/P_MP amplitude ratios on a location map of all seismic-refraction profiles in western Germany (from Fuchs, 1983, fig. 12). Thick solid lines: ratios = 1, thick dashed lines, ratios < 1. These amplitude ratios are projected into Bamford's (1973) velocity distribution plotted to the right above. The large-amplitude sector is centered on Bamford's fast direction, the small amplitudes around the slow direction. [Physics of the Earth and Planetary Interiors, v. 31, p. 93–118. Copyright Elsevier.]

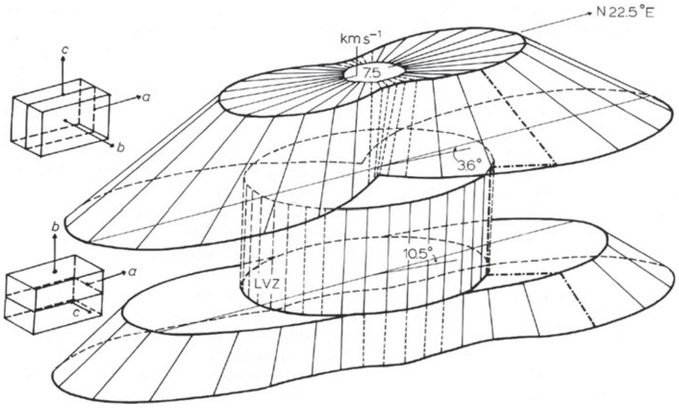


Figure 7.2.4-02. Three-dimensional "anvil" model of Fuchs (1983) showing the continental anisotropic velocity-depth distribution in the uppermost mantle under southern Germany found by Bamford's (1973) time-term analysis (from Fuchs, 1983, fig.13). [Physics of the Earth and Planetary Interiors, v. 31, p. 93–118. Copyright Elsevier.]

emerged from the crustal data of the western United States (Prodehl, 1970, 1979). Here a similarly large anisotropy of ~3% was found, with its high velocity direction 70–80° east of north, parallel to the spreading direction of the Basin and Range province (Bamford et al. (1979).

Both the amount and the direction were very similar to results found for the Pacific Ocean off California (Raitt et al., 1971; Fuchs, 1975, 1977). This coincidence led to the conclusion that this anisotropy is present as a consequence of the subduction of oceanic lithosphere beneath the western United States during Mesozoic times (Fig. 7.2.4-03).

The second project that was stimulated by D. Bamford's visit to Karlsruhe was the planning of the second West European long-range profile. This profile would be a 1000-km-long seismic line along the axis of the British Isles, from northern Scotland to the English Channel (LISPB 1974—Lithospheric Seismic Profile through Britain 1974). This project was to become an excellent collaboration of all major seismically oriented British and German institutions. On the British side the leading scientists were D. Bamford, K. Nunn, R. King, and D. Griffiths of the University of Birmingham, as well as B. Jacob and P. Willmore of the British Geological Survey at Edinburgh, while on the German side the project leaders were K. Fuchs and C. Prodehl.

The involvement of B. Jacob was particularly important for the shooting side. From previous experience, he knew that an optimum depth shot of ~5 tons would be adequate to reach the required distance range of 1000 km (e.g., Iyer et al., 1969). However, this would require a water depth of 180 m, which was not available at the ends of the LISPB line. Also, the main energy of such shots would be near 2 Hz, which was unfavorable for most of the available equipment. To overcome these problems, B. Jacob devised a simple, but effective system that used multiples of much smaller depth charges ("dispersed shots").

He had already experimented in a lake to see how seismic charges could be most efficiently detonated at shallower water

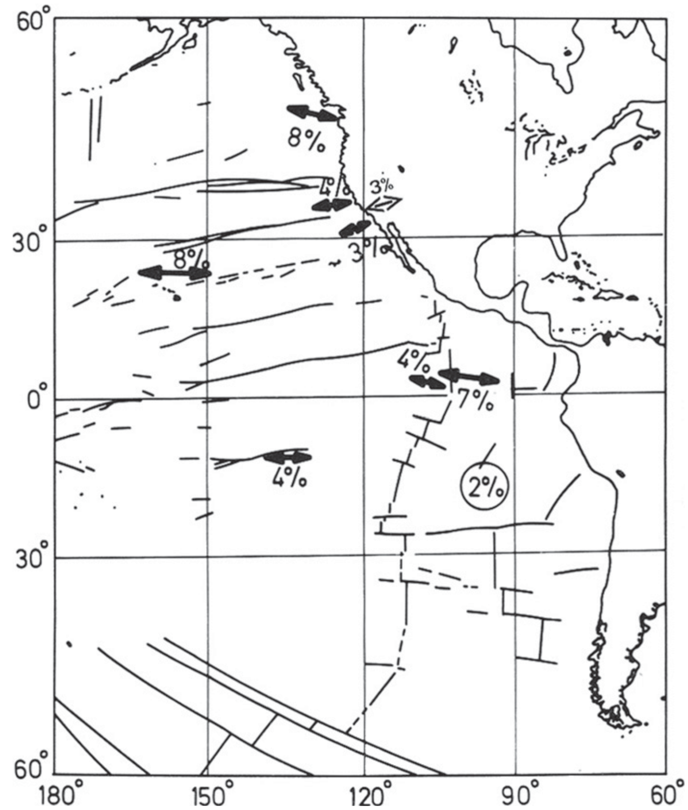


Figure 7.2.4-03. Anisotropy of uppermost-mantle velocity extending from western North America into the Pacific Ocean (from Bamford et al., 1979, fig. 17). [Geophysical Journal of the Royal Astronomical Society, v. 57, p. 397–430. Copyright John Wiley & Sons Ltd.]

depths, making use of the optimum depth which is directly dependent on the charge size. He found that splitting a charge into small units to be detonated simultaneously resulted in a much smaller optimum depth than would be necessary for a single larger charge (Jacob, 1975). As a result, 3–9 units of 200 kg were assigned for the northern offshore shots and 4–6 units for the southern offshore shots, which would be equivalent to unit charges of up to 5000 kg (Fig. 7.2.4-04).

The project was carried out from July to August 1974 (Bamford et al., 1976). The southern portion of the line was shifted

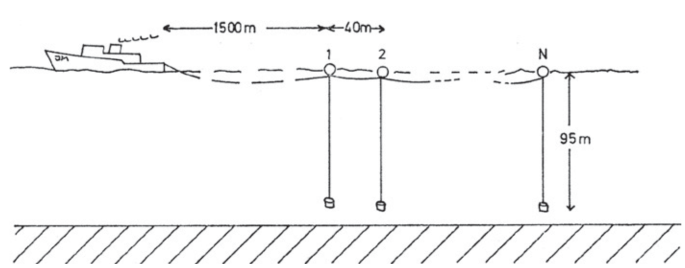


Figure 7.2.4-04. Scheme of sea shots with split charges off northern Scotland and in the English channel for LISPB 1974 (from Bamford et al., 1976, fig. 3). [Geophysical Journal of the Royal Astronomical Society, v. 44, p. 145–160. Copyright John Wiley & Sons Ltd.]

toward the west to run through Wales and Cornwall to avoid a profile line through the industrial centers of central England. The section through Wales was covered by two sea shots and an intermediate land shot for crustal control, while the line through northern England and Scotland contained three land shots and an underwater shotpoint in the Firth of Forth (Fig. 7.2.4-05; Appendix A2-1, p. 56–59).

This particular project was the first major effort to obtain a detailed picture of crustal and uppermost-mantle structure under the British Isles proper. It stimulated the participation of all seismological research institutions in Britain and obtained much public attention. A particularly lucky event was the local earthquake with a magnitude of 3 that occurred under western Scotland during the recording window of the northern half of the line. This small Kintail earthquake (Fig. 7.2.4-05) served as an additional "land shotpoint" (Kaminski et al., 1976), even though it was slightly off the main line.

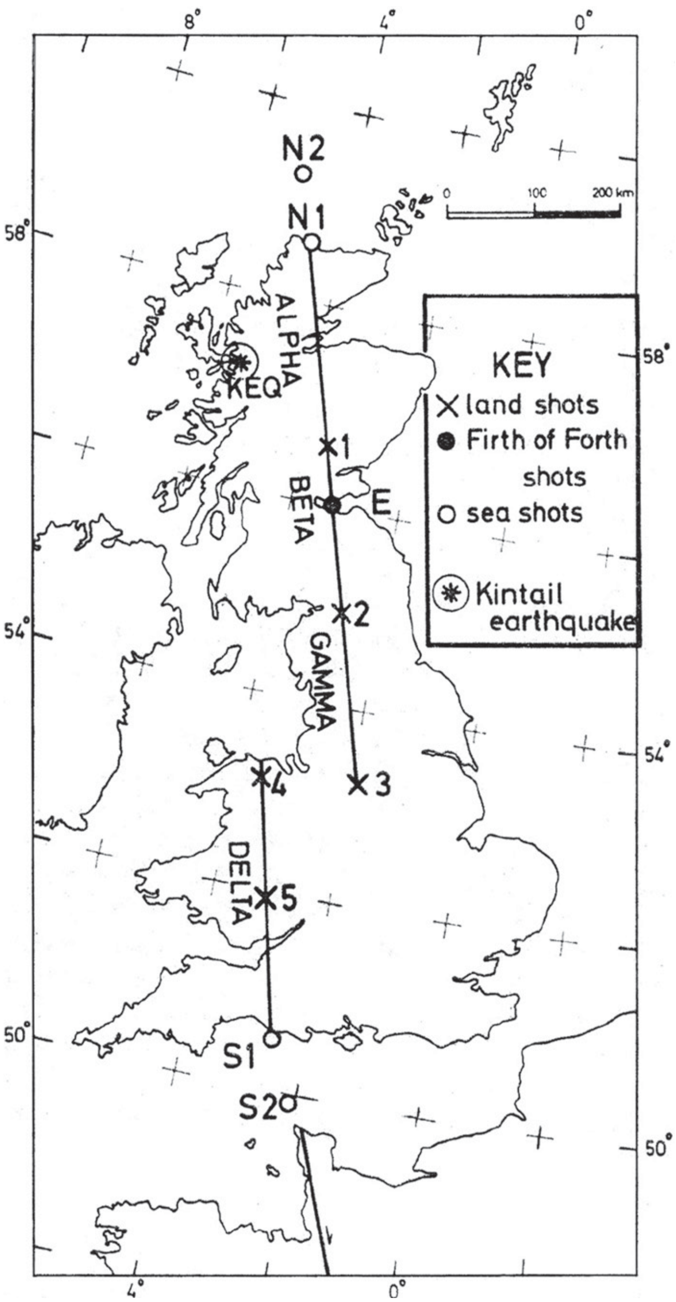


Figure 7.2.4-05. Location map of LISPB 1974 (from Bamford et al., 1976, fig. 1a). [Geophysical Journal of the Royal Astronomical Society, v. 44, p. 145–160. Copyright John Wiley & Sons Ltd.]

The interpretation, similar to the interpretation of the experiment in France, was divided into a crustal part and an upper-mantle part. The crustal model (Fig. 7.2.4-06) focuses on the northern section of LISPB (Bamford et al., 1978). The southern section DELTA was only subsequently published (Griffiths and Westbrook, 1992; Maguire et al., 2011). While the resulting crustal structure (Bamford et al., 1978) was of particular interest to the British earth scientists, the upper-mantle part was of high interest to the international community.

As in the French long-range profiles, mantle phases could be correlated on several parts of the line from different shotpoints (Fig. 7.2.4-07). The model proposed by Faber (1978) and Faber and Bamford (1979) showed two high-velocity layers with velocities near 8.15 and 8.4 km/s embedded in an upper mantle ranging from 28 to 85 km depth where the background velocity gradually increased from near 7.95 to near 8.25 km/s (Fig. 7.2.4-06, bottom). The model was similar to the previously obtained model in France, except for the depth range around the Moho.

The strong velocity contrast of the French model immediately below the Moho, consisting of a high-velocity lid underlain by a distinct velocity inversion, was missing in Britain. Also, some lateral variations of the sub-crustal structure could be determined which are evidently related to gross geological variations beneath northern Britain (Faber and Bamford, 1981).

The successful completion of LISPB had a large impact on the British earth science community. Jeremy Hall undertook two very important surveys, but concentrating on the upper crust. LUST (the Lewisian Units Seismic Traverse) was recorded in 1976 across the Ben Stack Line separating granulite and amphibolite facies blocks with the Archean basement of NW Scotland (Hall, 1978). The deep crust project SISSE (South Irish Sea Seismic Experiment) in 1977 (Ransome, 1979) concentrated on Wales, and the project WISE (Western Isles Seismic Experiment) was carried out offshore western Scotland in 1979 and 1981 (Summers et al., 1982). Explosive and airgun shots were recorded along two profiles with the aim to map the Lewisian basement from the foreland to the Caledonian mountain belt (Summers et al., 1982). The first profile was shot in 1979 between Barra and Ballantree, and the second one in 1981 between Mull and Kintrye. The shallow seismic-refraction project SWESE (SW England Seismic Experiment) was also carried out in 1979 between Lizard and Stait peninsulas in SW England (Doody and Brooks, 1982).

In 1979 the Deep Geology Unit of the Institute of Geological Sciences (IGS) started to acquire seismic-reflection data to study the continental crust beneath the UK landmass particularly at levels deeper than those examined on a routine basis by explora-

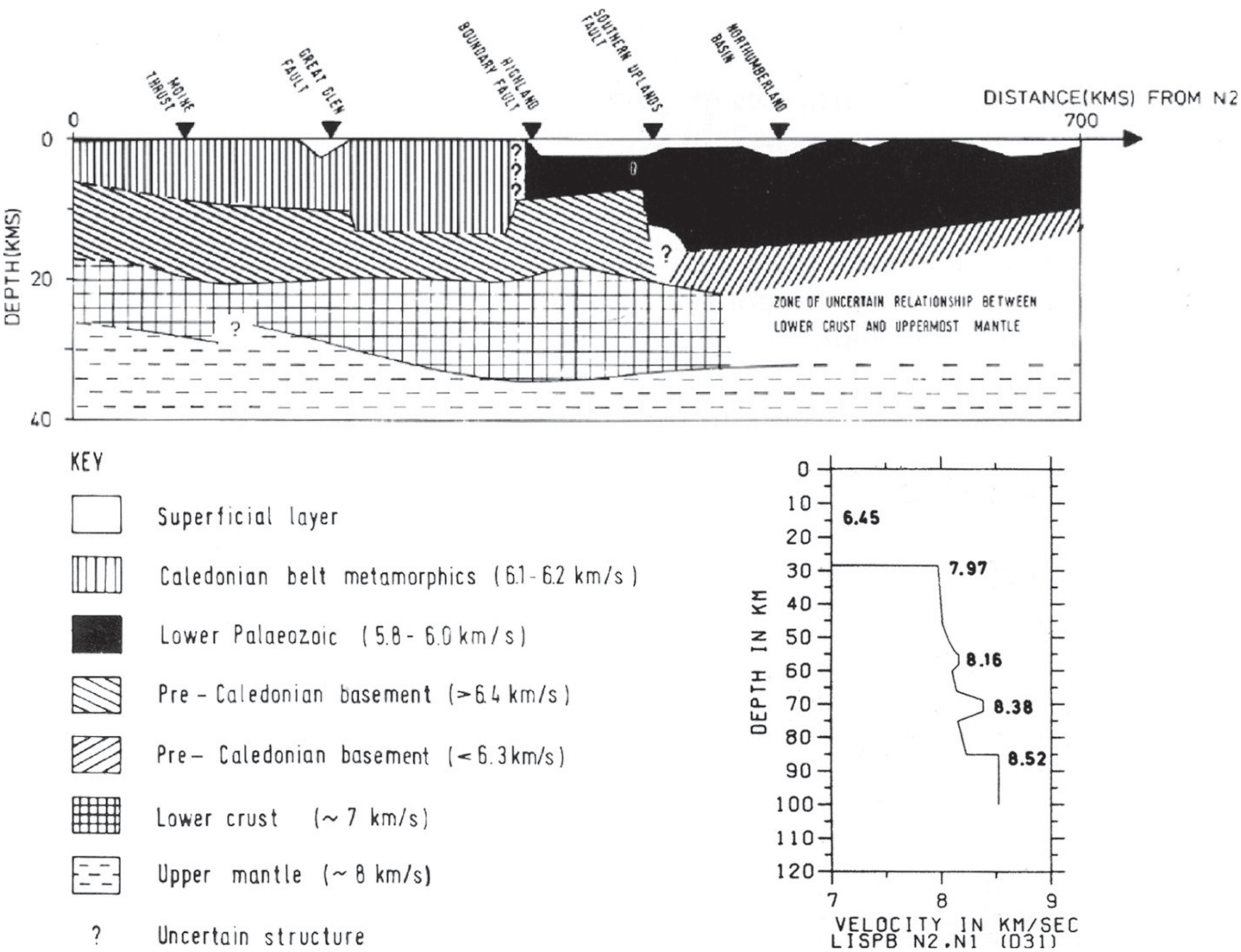


Figure 7.2.4-06. Top: Crustal cross section of LISPB 1974 (from Bamford et al., 1978, fig. 15). Bottom: Schematic cross section of subcrustal reflections (from Faber, 1978, fig. 10.4). [Ph.D. thesis, Universitat Karlsruhe, 132 p. Published by permission of Geophysical Institute, University of Karlsruhe, Germany.]

tion companies (Whittaker and Chadwick, 1983). For these experiments, either commercial companies were hired or lines were shot using IGS in-house equipment but processed commercially, or the record lengths of commercial survey lines were increased obtaining extra grant money. The first project was the IGS 1979 Wiltshire Survey. It comprised two intersecting lines located over the Variscan Front. Higher amplitude events of restricted lateral extent were seen between 8.8 and 9.4 s TWT, possibly reflected from the base of the crust.

Following the successful experiments to explore the lower lithosphere beneath the Variscides of France and the Caledonides of Great Britain, the European international community started to plan for a detailed investigation of the entire lithosphere underneath a young orogen, the Alps.

For this purpose a long-range line was planned that would follow the strike of the main geological units as much as possible. However, this aim could only be realized for the Central and Eastern Alps, but not for the Western Alps due to the strong curvature of the Alpine arc. A second guideline for the location of the main line was the Bouguer gravity map. It was aimed to position the line as close as possible through the center of the negative Bouguer anomaly (Fig. 7.2.4-08).

For the detailed planning and a first detailed interpretation an international working group was founded consisting of scientists from Austria (Vienna), England (Birmingham), France (Grenoble), Germany (Munich and Karlsruhe), Hungary (Budapest), Italy (Milano and Trieste), and Switzerland (ETH Zurich), headed by J. Ansorge (Zurich) and H. Miller (Munich). The final line was split. The eastern half within Austria followed the Eastern Alps in an E-W direction, and the western half within Switzerland and eastern France was slightly rotated into the ENE-WSW direction. Several side lines served both as fans and lines with

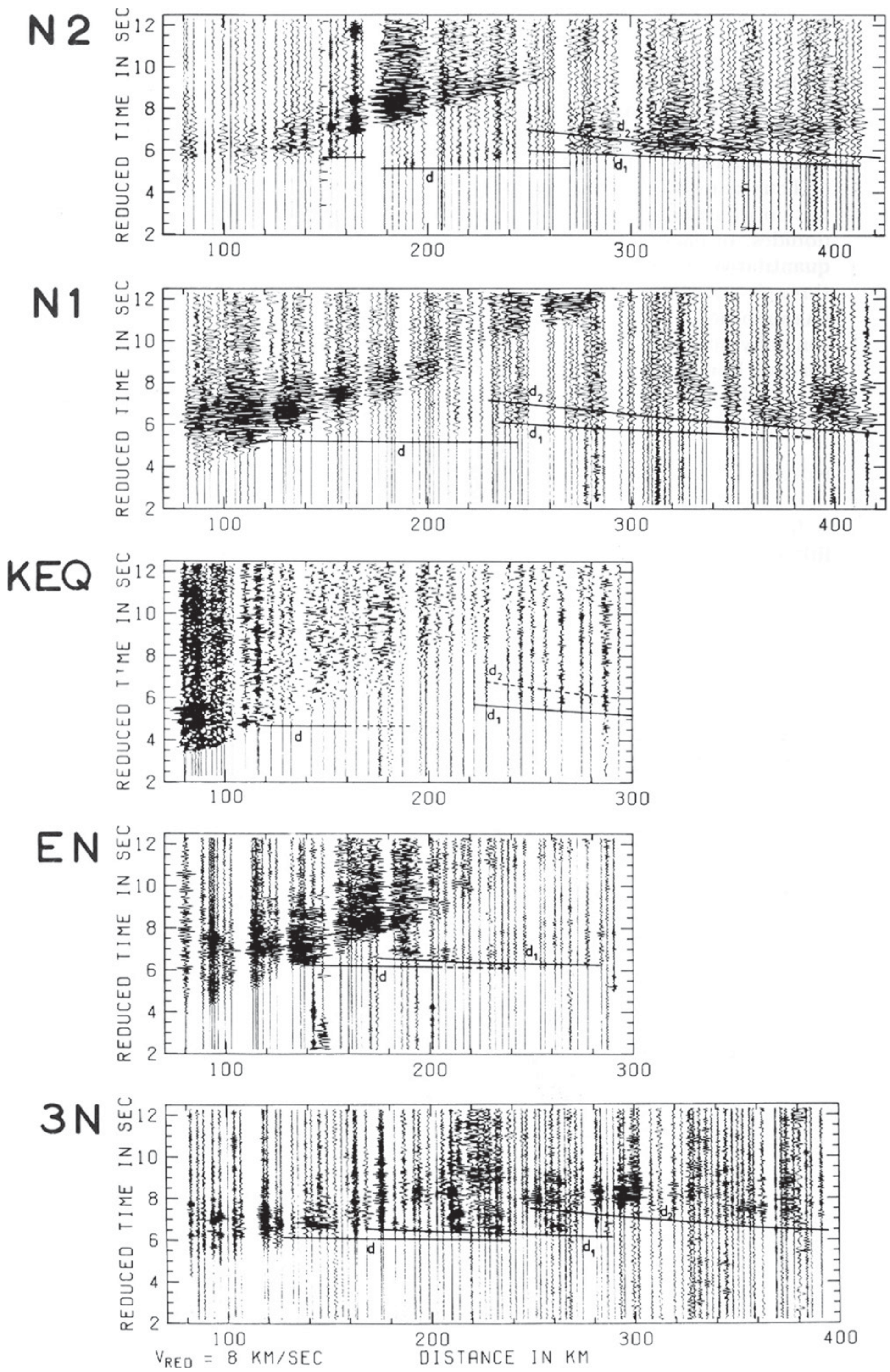


Figure 7.2.4-07. LISPB 1974 seismic refraction mantle data (from Faber and Bamford, 1979, fig.5). [Tectonophysics, 56, p. 17–30. Copyright Elsevier.]

additional shotpoints at their ends and one line between Innsbruck and Vienna tried to continue the direction of the western line toward the ENE in order to catch the ray paths of the deeper penetrating waves as close as possible to the axis of the Central Alps underneath Switzerland (Fig. 7.2.4-08).

In order to obtain a distance range up to 850 km, the line was extended beyond the Alps proper to the east, where colleagues from Hungary managed to organize two large borehole shots of 2 and 4 tons. The other shots within the Alps ranged between 0.6 and 3.3 tons and were mainly borehole shots as well, save for

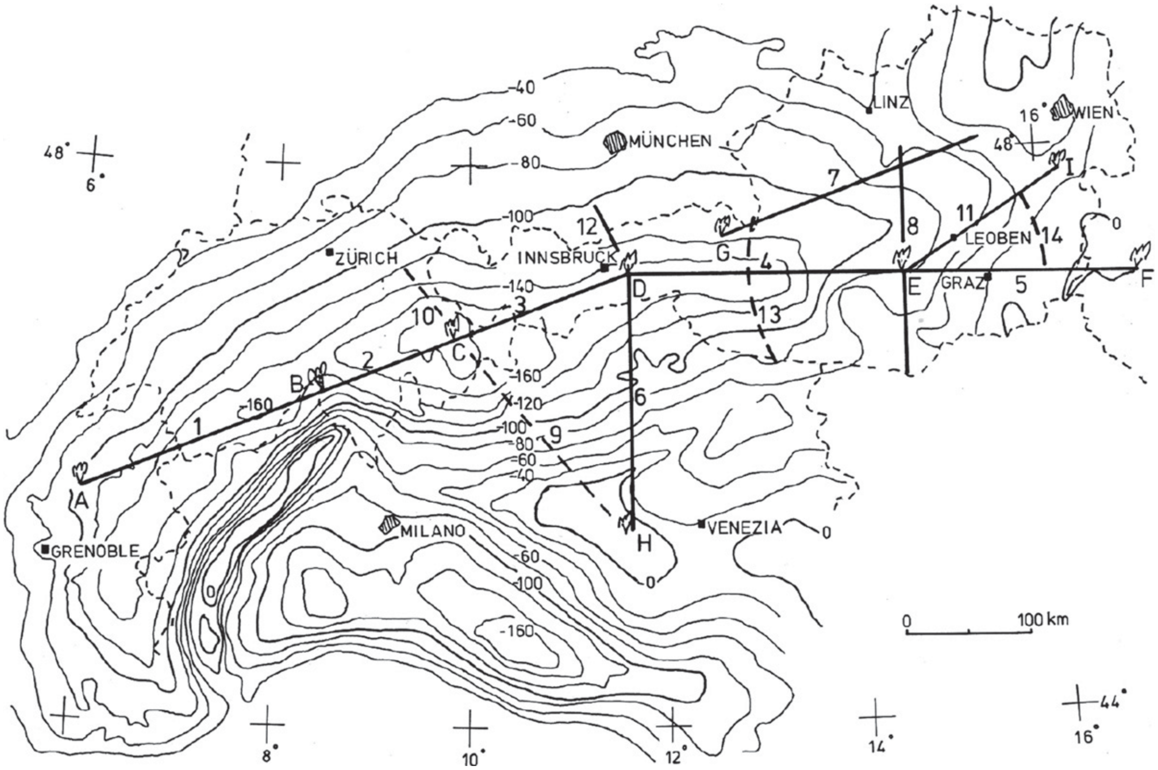


Figure 7.2.4-08. Simplified gravity map of the Alps and the location of ALP75 (from Alpine Explosion Seismology Group, 1975, fig. 2). [PAGEOPH, v. 114, p. 1109–1130. Reprinted by permission from Birkhaeuser Verlag, Basel, Switzerland.]

two locations. Alpine lakes at a high enough elevation were used for underwater shooting, and two shotpoints were quarry blasts. A total of 193 recording stations could be acquired, of which 158 were MARS-66 units. Within Hungary, four 24-trace reflection units were used. The maximum distance range to which energy was recorded was near 620 km.

The crustal data were of high quality (Appendix A2-1, p. 7–10). P_n phases could be observed to ~360 km distance, and two upper-mantle phases at 300–440 km and at 420–620 km distance could be correlated. Due to the complicated crustal structure of the Alps, the first interpretation by the Alpine Explosion Seismology Group (1975) did not go beyond a rough outline of the Moho along ALP75, but many other publications were to follow in the subsequent years as, e.g., two publications written by Miller et al. (1978, 1982).

The most recent interpretation was published by Yan and Mechie (1989; Appendix A2-1, p. 7–10, and Appendix A7-3-1) which confirmed a crustal root under the main part of the line between shotpoint locations B and E as well as several high-velocity lenses in the lower lithosphere beneath the Moho. These lenses, however, are not continuous features but differ in number, as well as lateral extension and depth under the Central and Eastern Alps. Yan and Mechie have also investigated the mantle phases, which could be correlated on several record sections at distances beyond 250 km, following the P_n phase, named d by Yan and Mechie (1989). An example is shown in Figure 7.2.4-09, other sections and the corresponding 1-D models can be viewed in Appendix A7-3-1.

Shortly after the completion of ALP75, in 1977 a second longitudinal profile was laid out through the Southern Alps, named SUDALP77 (Ansorge et al., 1979b). SUDALP77 was more or less parallel to ALP75, and was complemented in 1978 (Italian Explosion Seismology Group and Institute of Geophysics ETH Zurich, 1981). SUDALP77 involved the cooperation of Italian, Swiss, French, and German institutes, ran from Lago Maggiore to Gemona near the Italian-Yugoslavian border and was based on four shotpoints. The additional line ran southeastward from the Central Alps to the Adriatic Sea near Trieste (Fig. 7.2.4-10; Appendix 7-3-2).

Presenting a lithospheric cross section across the border region between the Central and Southern Alps, Ansorge et al. (1979b) demonstrated the significant difference in structure of the Central and Southern Alps as well as the termination of the anomalous Ivrea zone north of that border region.

Combining all available seismic data at the time being, Mueller et al. (1980) compiled a crustal cross section along the so-called "Swiss Geotraverse." That crustal cross section extended from the Rhinegraben near Basel, Switzerland, to the Po Plain near Como, Italy. It traversed the Swiss folded Jura Mountains, the Molasse basin, the Aar Massif, the Central or Penninic Alps between the Rhine–Rhône Line and the Insubric Line, and the Southern Alps (Fig. 7.2.4-11).

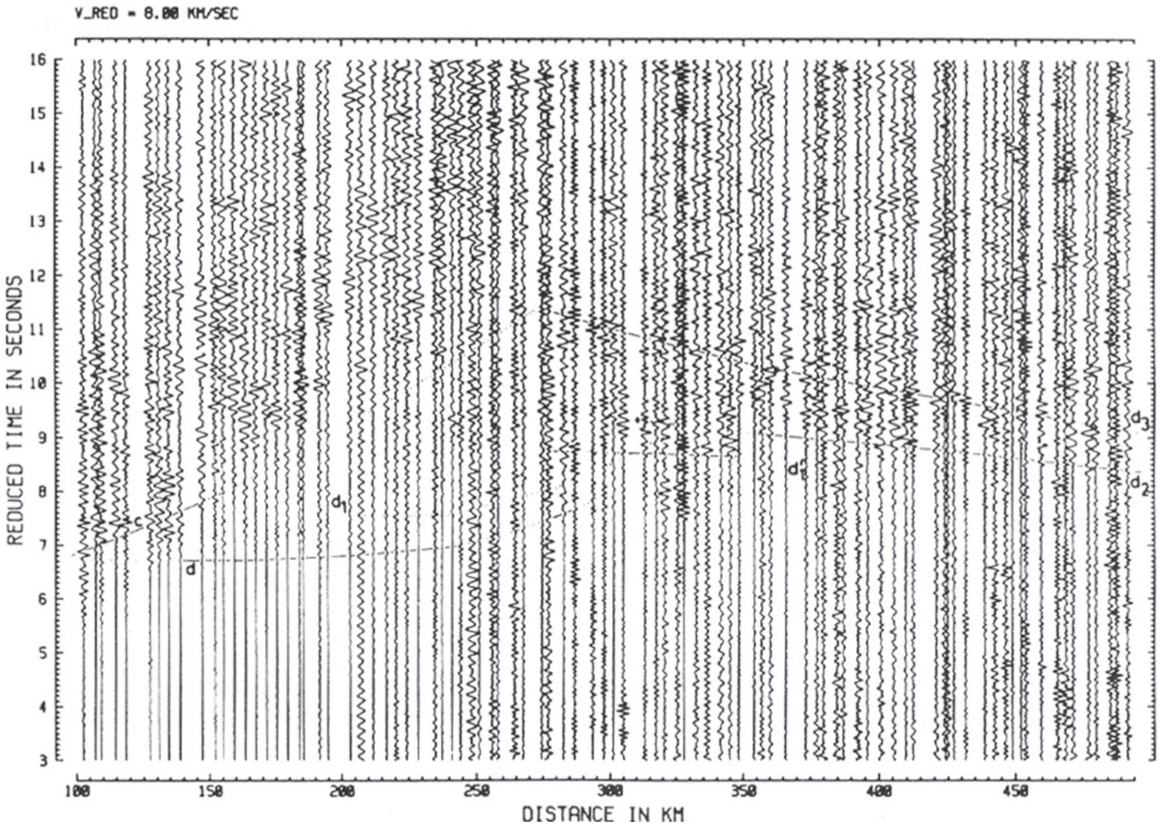


Figure 7.2.4-09. Example of mantle data recorded during ALP75 (shotpoint F to west) (from Yan and Mechie, 1989, fig. 17). Reduction velocity = 8 km/s. [Geophysical Journal International, v. 98, p. 465–488. Copyright John Wiley & Sons Ltd.]

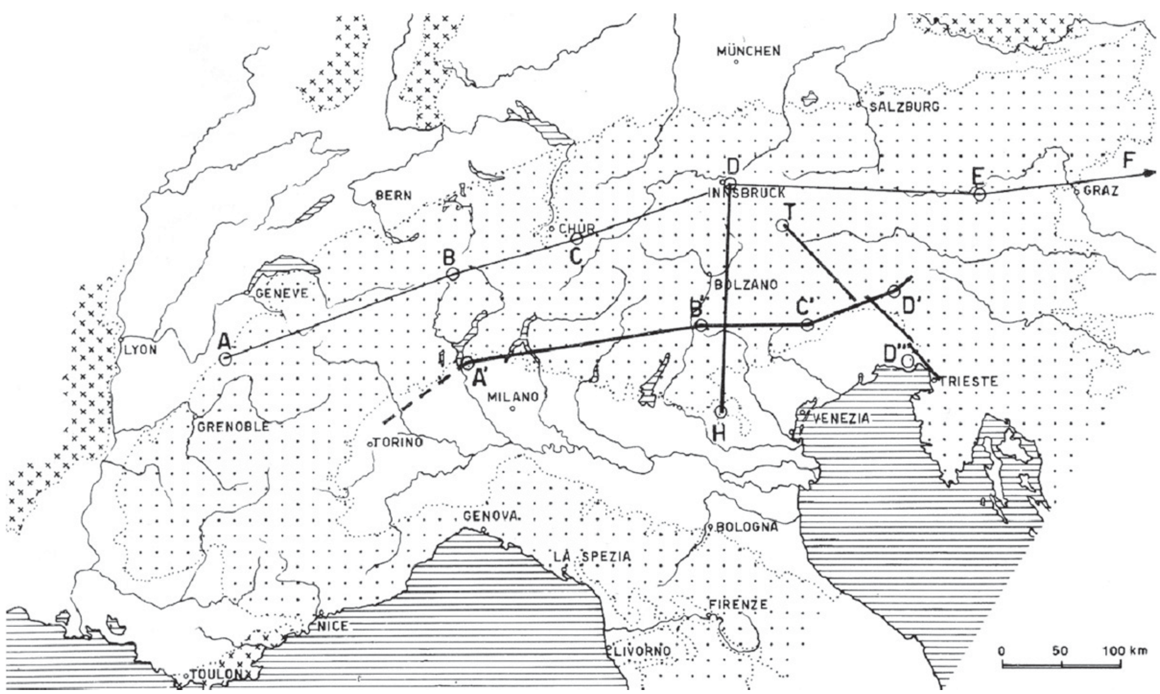


Figure 7.2.4-10. Simplified tectonic map of the Alps showing the location of ALP75 (shotpoints A to H) and SUDALP77 (shotpoints A′ to D′, D′, and T) (from Ansorge et al., 1979, fig. 1). [Bollettino di Geofisica Teorica e Applicata, v. 21, p. 149–157. Reproduced by permission of Bollettino di Geofisica, Trieste, Italy.]

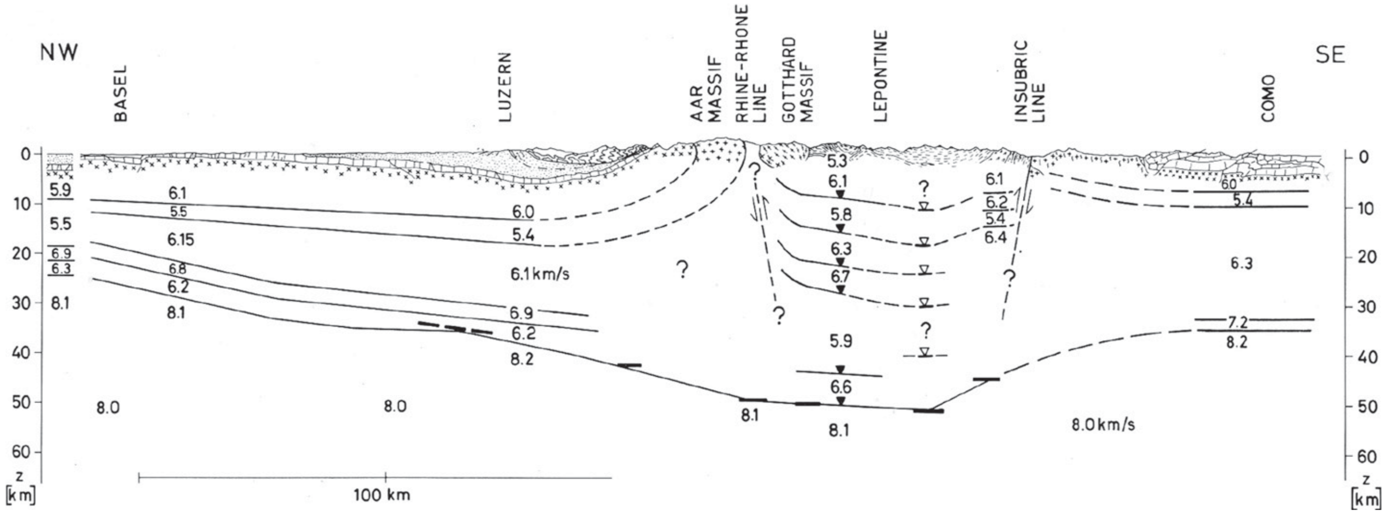


Figure 7.2.4-11. Compilation of seismic crustal data along the Swiss Geotraverse (from Mueller et al., 1980, fig. 7). [Eclogae Geologicae Helvetiae, v. 73, p. 463–483. Reproduced by permission of Swiss Geological Society, Neuchatel, Switzerland.]

7.2.5. Continued Explorations of Anomalous Areas and a Second Stage of Deep-Seismic Reflection Campaigns in Central Europe

In 1975 and 1976, a first attempt was undertaken to unravel the crustal structure of the North German basin by way of a large-scale seismic-refraction experiment from the North Sea northwest of Heligoland to the area near Goettingen (location is shown as line 1 in Fig. 8.3.4-08).

In spite of a high natural noise level, with dense station spacing and 10 shotpoints on land distributed over a distance of ~250 km plus 4 additional sea shots in the North Sea, a rather complex crustal structure model was obtained (Fig. 7.2.5-01). A thick sedimentary sequence of more than 15 km overlies a rather thin crystalline crust, and the depth to the Moho was found to decrease from 33 km in the north and to 27 km in the south. Below the Moho at 37–42 km depth, a velocity inversion was modeled (Reichert, 1993).

In 1974, and continuing into the 1980s (see Chapter 8.3.4.2) the hydrocarbon industry started to acquire an extensive seismic-reflection database in the North German Basin near the Elbe lineament between Hamburg to the south, Kiel to the north, the North Sea to the west, and the Baltic Sea to the east (Dohr et al., 1989; Yoon et al., 2008). The area of the Glückstadt graben, being

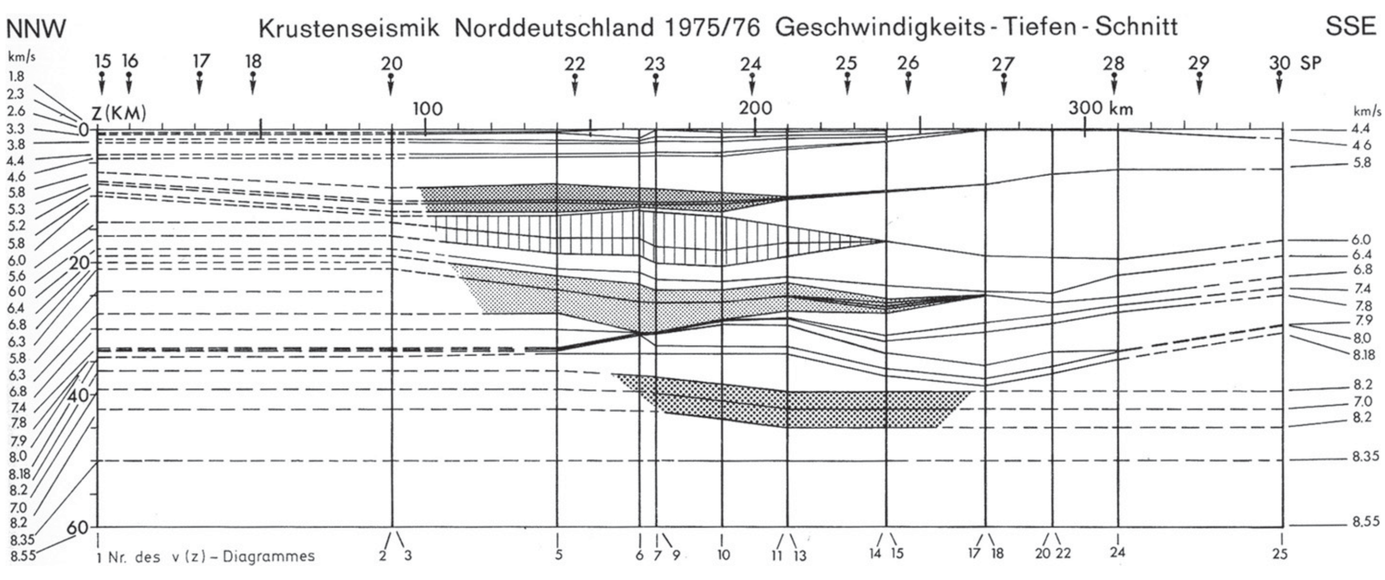


Figure 7.2.5-01. Crustal structure of the North German Basin between Heligoland and Goettingen (from Reichert, 1993, fig. 18). Location shown as line 1 in Fig. 8.3.4-08. [Geol. Jb., Hannover, E, 50, 87 p. Reproduced by permission of Schweizerbart, Stuttgart, Germany (www.schweizerbart.de).]

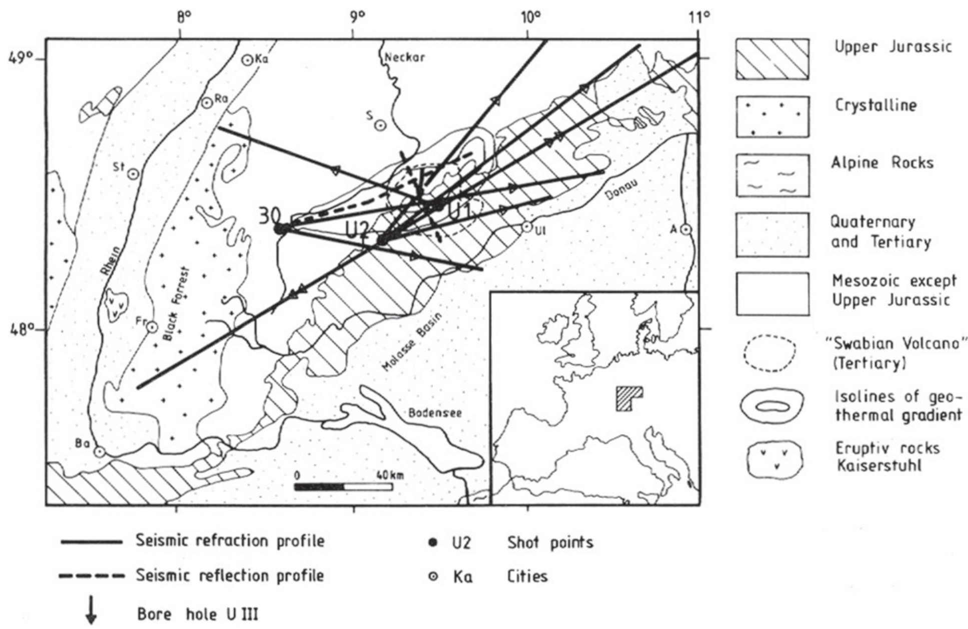


Figure 7.2.5-02. Location of the seismic refraction (full lines) and reflection (dashed lines) surveys (from Gajewski and Prodehl, 1985, fig. 1). [Journal of Geophysics, v. 56, p. 69–80. Reproduced with kind permission of Springer Science+Business Media.] Permit no. 16

the deepest Mesozoic graben structure dominated by strong salt tectonics, was of particular interest as it revealed the deep crustal structure along the transition zone from the Precambrian Fennoscandian Shield to the Caledonian accreted crust in NW Germany. The first N-S directed line 4, located west of Lübeck (see fig. 2 of Yoon et al., 2008) was 66 km long and consisted of 590 shot recordings using 48 receivers. Shotpoint and receiver spacings were 100 m, and the total recording time was 15 s TWT, resulting in an actual common midpoint fold between 15 and 20.

In 1976, the area between the Rhine and Rhône grabens was a renewed goal for French, German, and Swiss scientists. Some of the shotpoints of the 1972 Rhônegraben experiment were revived, and some profiles observed in a northerly direction, both along and across the Saône and Bresse depressions as well as into the Swiss Jura (Michel, 1978; Ansorge and Mueller, 1979). An assumed upward doming of the Moho in the area of the so-called "Burgundische Pforte," the lowland between the Rhine and Rhône valleys, was confirmed.

Another goal of investigation close to the Rhinegraben was the area of the so-called "Swabian Volcano" near Urach, east of Stuttgart, where a geothermal anomaly reaching temperatures of 130–140 °C at 3 km depth had been mapped. Under the auspices of the European Community, a multidisciplinary geoscientific research project of this anomaly had started to look for geothermal resources with the final goal to drill a geothermal hole which, when it was finally realized, reached a depth of 3333 m. To investigate the area of the geothermal anomaly prior to drilling in some detail, in 1978 and 1979, both seismic-refraction and reflection measurements were carried out (Bartelsen et al., 1982; Jentsch et al., 1982; Gajewski and Prodehl, 1985; Fig. 7.2.5-02; Appendix A2-1, p. 25–26). The initial goal was to map the topography and the velocity structure of the crystalline basement. Two different approaches were undertaken to accomplish this goal.

The wider surroundings were explored by a seismic-refraction survey (Jentsch et al., 1982). Two specially designed borehole shots were arranged along a WSW-ENE line along the axis of the Swabian Jura, (U1 and U2 in Fig. 7.2.5-02) from which, with a recording distance of up to 120 km, a model of the entire crust was to be obtained, while a number of nonreversed additional lines, crossing the area in various directions, were mainly aimed to map the basement (Jentsch et al., 1982). The main line was arranged so that it was located entirely on the limestone ridge of the Swabian Jura and also coincided with the prolongation of one of the fan profiles, which were recorded in 1972 from shotpoint Steinbrunn to the ENE (Edel et al., 1975), and was later interpreted by Gajewski and Prodehl (1985; Appendix A2-1, p. 25–26).

Since the 1950s, it had become evident that reflection seismology, as used for oil exploration studies, was a useful tool also for the investigation of the whole Earth's crust. In general, seismic-reflection surveys, working with much higher frequencies than refraction investigations, deliver completely different information than do refraction–wide-angle studies. Often they show structures and tectonics directly and with high accuracy while

Figure 7.2.5-03. From top to bottom: Stacked reflection section of profile U1 with correlatable events. Record section of wide-angle observations from shots along profile U1 with traveltime curves calculated from velocity model below. Velocity model with ray paths. Final velocity model along profile U1 showing a body of reduced velocity (hatched area). (From Bartelsen et al., 1982, figs. 9, 16, and 13). [*In* Haenel, R., ed., The Urach geothermal project (Swabian Alb, Germany): Schweizerbart, Stuttgart, p. 247–262. Reproduced by permission of Schweizerbart, Stuttgart, Germany (www.schweizertbart.de).]

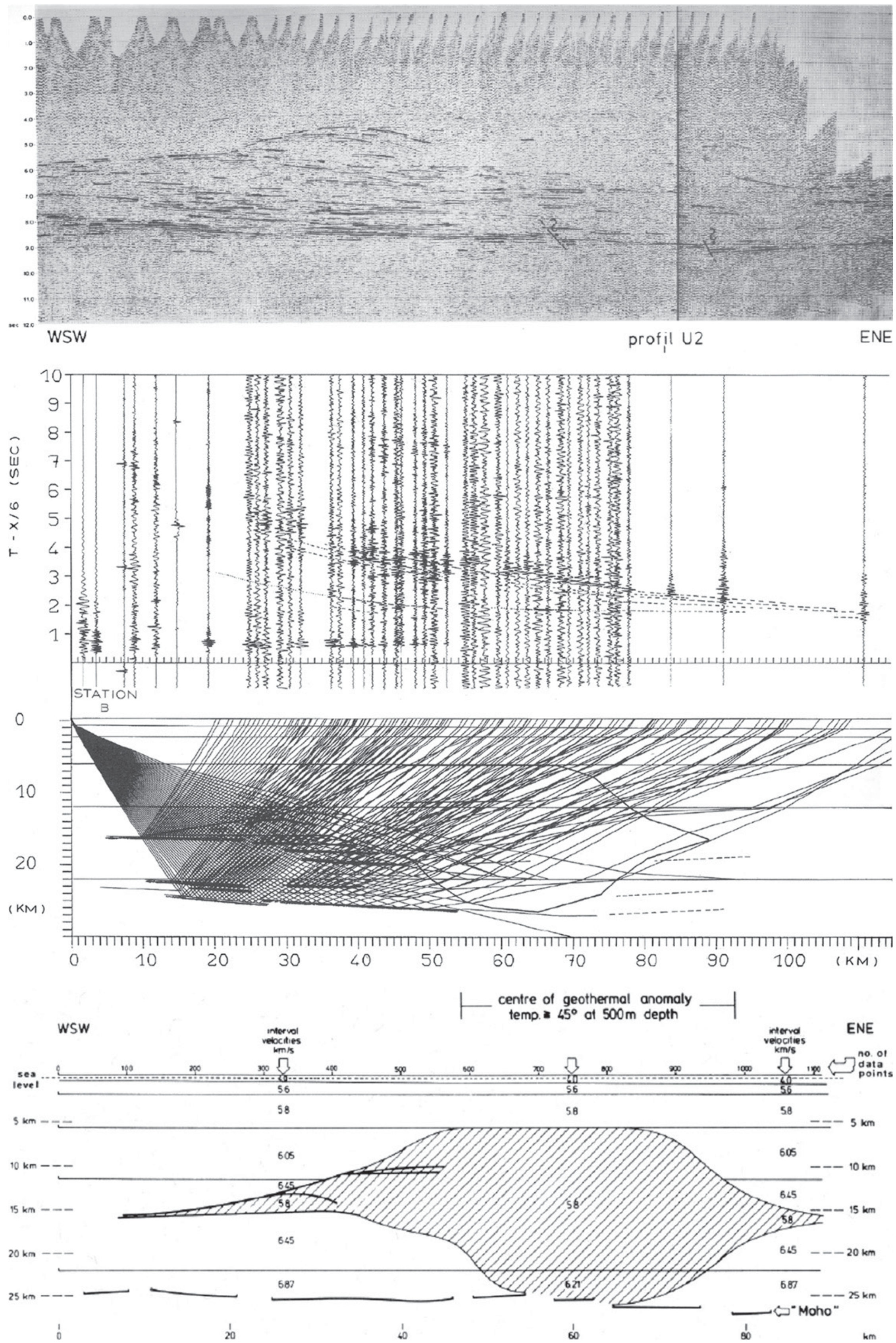


Figure 7.2.5-03.

refraction–wide-angle studies show sequences of velocities, important for petrological structure. It was R. Meissner who organized the first large-scale, purely scientific, seismic-reflection experiments in Germany.

For the detailed investigation of the geothermal anomaly of Urach, R. Meissner proposed a detailed seismic-reflection survey (Bartelsen et al., 1982). Parallel to the main seismic-refraction line but further to the north in the foreland of the Swabian Jura, a seismic-reflection profile of 27 km length was arranged and recorded by a commercial company (dashed lines in Fig. 7.2.5-02). The steep-angle observations were complemented by additional refraction stations recording all shots in the wide-angle distance range up to 100 km. Furthermore, a second shorter steep-angle reflection line was set up, running perpendicular to the first one (Bartelsen et al., 1982) and crossing the center of the anomaly near the town of Urach where the drill hole was finally placed.

This combined seismic reflection-refraction survey with extended spread lengths and an 8-fold coverage provided a high resolution result with significant velocity anomalies. As a result, a large body with velocity as low as 5.8 km/s could be identified to underlie the area at mid- and lower-crust levels down to the Moho at to 25 km depth, where the geothermal anomaly had been mapped (Fig. 7.2.5-03). Both thermal effects and a large-scale crustal alteration were suggested to explain this result (Bartelsen et al., 1982).

Following the wide-angle profile through the Hunsrück in 1968, as referred to in Chapter 6, in 1973, R. Meissner and H. Murawski started the project Geotraverse Rhenoherzynikum (Meissner et al., 1980).

They followed the idea to systematically fill in seismic-reflection surveys along a SSE-NNW–directed seismic-refraction line between Landau in the Upper Rhinegraben to Aachen at the border between the Rhenish Massif in the south and the Lower Rhine embayment in the north. In 1973 and in 1975, two short reflection seismic surveys were performed. The first one ran across the Hunsrück fault, which separates the Paleozoic Saar-Nahe trough from the Rhenish Massif. The second one was placed near Landau, crossing the western Rhinegraben border zone. Both

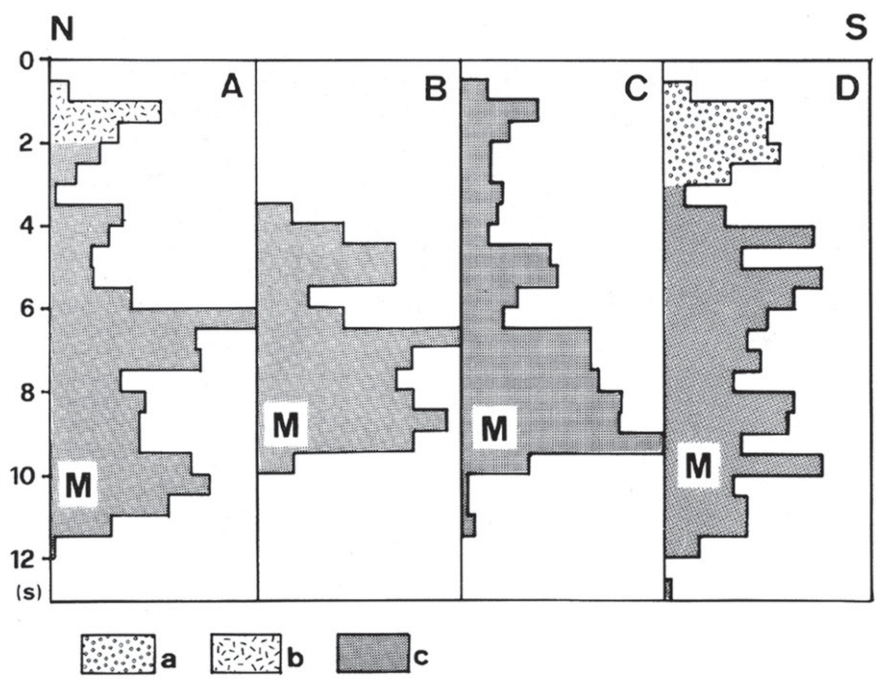
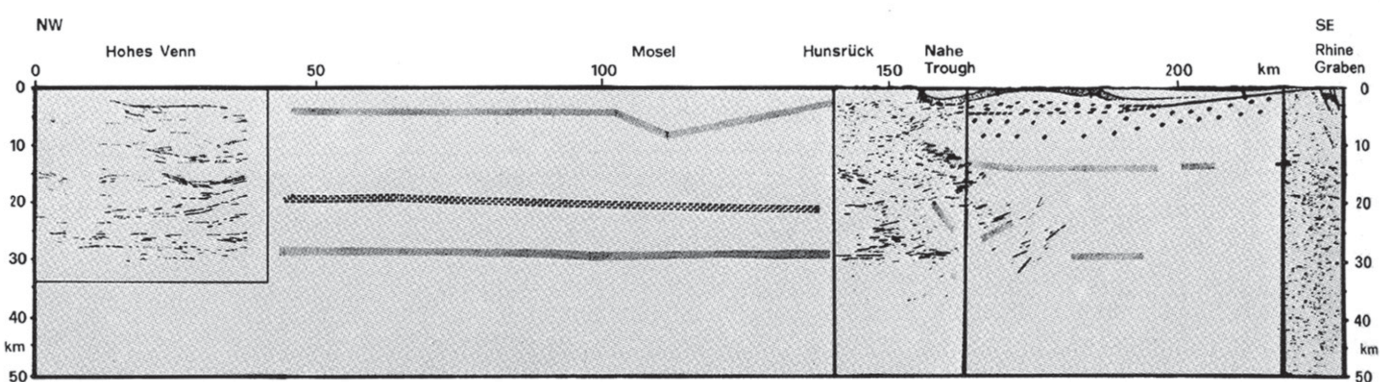


Figure 7.2.5-04. (Top) Comparison of the density of reflections for four areas: A—Hohes Venn near Aachen 1978, B—Hunsrück 1968, C—Hunsrück border zone fault 1973, D—Rhinegraben border fault 1975. M = Moho; (a) sedimentary basin; (b) thrust faults; (c) crystalline crust (from Meissner et al., 1983, fig. 7). (Bottom) Cross section along the Geotraverse Rhenoherzynikum (from Meissner et al., 1983, fig. 8) based on seismic-reflection and refraction studies of Meissner et al. (1980, 1983) and Mechie et al. (1983). [*In* Fuchs, K., von Gehlen, K., Mälzer, H., Murawski, H., and Semmel, A., eds., Plateau uplift: The Rhenish Massif—a case history: Springer, Berlin-Heidelberg, p. 276–287. Reproduced with kind permission of Springer Science+Business Media.]



surveys included seismic-refraction stations recording the reflection shots at wide-angle distance ranges.

Later, in 1978, R. Meissner organized another seismic-reflection survey combined with wide-angle observations (Meissner et al., 1980, 1983) in the northernmost part of the Rhenish Massif near its northern margin, close to the city of Aachen (see next section, line L1-L2 in Fig. 7.2.6-01). This time, a reversed seismic-refraction profile of ~100 km in length could be arranged as the wide-angle part with the reflection shots near Aachen in the north, and a number of shots fired at its southern end in the troop training area of Baumholder within the Saar-Nahe trough (shots M1-M2 in Fig. 7.2.6-01; see next section). Here, a few months earlier, the Karlsruhe colleagues, using their Rhinegraben seismic network, had located a strong event which turned out to be an unplanned non-delayed quarry blast within this troop training area.

This detection opened a very fruitful cooperation of German geophysicists with the German army, which in the following years would lead to a number of zero-cost shots for scientific purposes. The interpretation of the reflection survey L1-L2 (Fig. 7.2.6-01) by Meissner et al. (1983) led to the identification of the shape and depth pattern of prominent fault zones, suggesting that a huge thin-skinned nappe forms the northern Variscan deformation front (Fig. 7.2.5-04).

7.2.6. The Third Stage of Long-Range Profiling

In 1979, two major long-range projects were undertaken by the western European geophysical community. The first one in western central Europe targeted the crust and uppermost mantle under the Rhenish Massif and Ardennes, while the second one in Scandinavia targeted the lithosphere and asthenosphere.

The first project took place in May. It consisted of a 600-km-long line through Germany, Luxembourg, and northeastern France and was recorded from the Harz Mountains through the Rhenish Massif and Ardennes into the Paris Basin of northern France and ended south of Reims (Fig. 7.2.6-01; Appendix A2-1, p. 15–18; Appendix A7-4-1). In addition, two side profiles, up

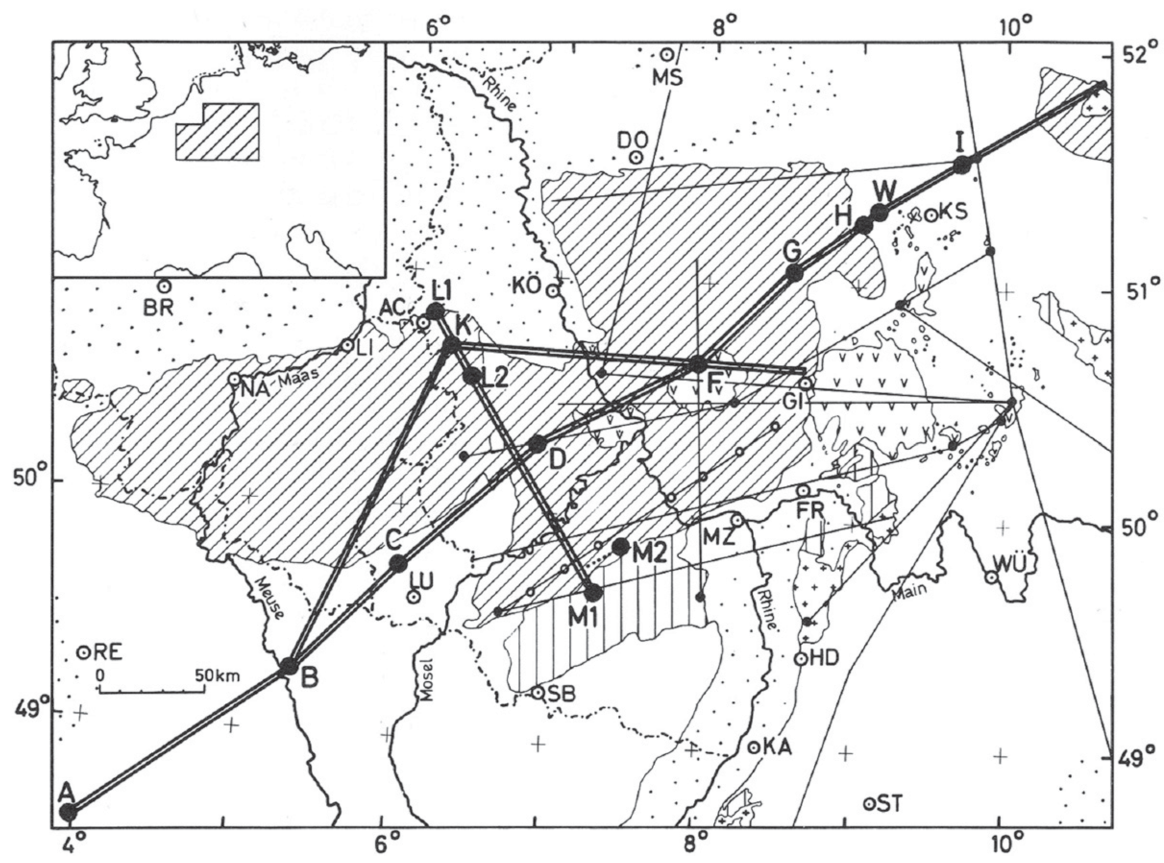


Figure 7.2.6-01. Recording lines through the Rhenish Massif and adjacent areas (from Mechie et al., 1983, fig. 1). Thick double lines: shotpoints and recording lines of 1978–1979. Thin lines: shotpoints and recording lines prior to 1978. Cities: AC—Aachen, BR—Brussels, DO—Dortmund, FR—Frankfurt, GI—Giessen, HD—Heidelberg, KA—Karlsruhe, KÖ—Cologne, KS—Kassel, LI—Liège, LU—Luxembourg, MS—Münster, MZ—Mainz, NA—Namur, RE—Reims, SB—Saarbrücken, ST—Stuttgart, WÜ—Würzburg. [*In* Fuchs, K., von Gehlen, K., Mälzer, H., Murawski, H., and Semmel, A., eds., Plateau uplift: The Rhenish Massif—a case history: Springer, Berlin-Heidelberg, p. 260–275. Reproduced with kind permission of Springer Science+Business Media.]

to 170 km long, were also recorded from Aachen toward the southwest through the Ardennes of eastern Belgium and toward the east into the volcanic area of the Vogelsberg (Mechie et al., 1983). The Geotraverse Rhenoherzynikum of 1978 (Meissner et al., 1983) served as the third side profile toward the SSE (Fig. 7.2.6-01; Appendix A7-4-1).

A reinterpretation of Mooney and Prodehl (1978) had revealed peculiarities of the Moho whenever seismic lines crossed volcanic areas or touched the northern end of the Rhinegraben. Here, the Moho, elsewhere a first-order discontinuity, seemed elevated and disrupted (Fig. 7.2.6-02). The new data showed very few P_n arrivals and, not in all cases, very strong secondary arrival phases. In particular, on the eastern half of the main line, under the Eifel and Westerwald, the secondary arrivals indicated the existence of two phases mixed into each other (phases d and e in Fig. 7.2.6-03).

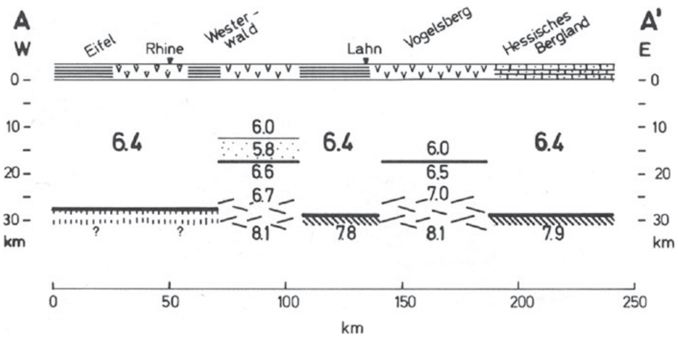


Figure 7.2.6-02. Crustal sketch through the Rhenish Massif, reinterpreted from data prior to 1978 (from Mooney and Prodehl, 1978, fig. 17). [Journal of Geophysics, v. 44, p. 573–601. Reproduced with kind permission of Springer Science+Business Media.]

It was interpreted as a Moho lid, at 30 km depth overlying an ~5–7-km-thick low-velocity zone where the velocity drastically decreases and then gradually increases to 8.4 km/s or more below 35 km depth (Fig. 7.2.6-04). As the long-range line did not cross the volcanic area of the Westerwald, the disrupted Moho of Mooney and Prodehl (1978) did not show up in the new model. West of the Eifel, under the Ardennes, this Moho lid had disappeared, as well as much of the intracrustal structure, modeled elsewhere, and only the lower velocity boundary, at 37 km depth, appeared as a deep-reaching transition between the crust and mantle. A normal crust, with a sharp Moho at 30 km depth, appears again under the Paris Basin (Mechie et al., 1983).

The previous successful experiments with observations up to and beyond 1000 km distance, which had enabled viewing fine structure of the subcrustal lithosphere to 80 km in depth and more, and the successful long-range lines in North America, created the desire to look deeper into the upper mantle beneath Europe. Only Scandinavia offered the chance of planning for a line up to 2000 km in length on relatively homogeneous geological subsurface structure. It was hoped that the homogeneity of crustal terrane would allow a uniform "window" into the underlying mantle.

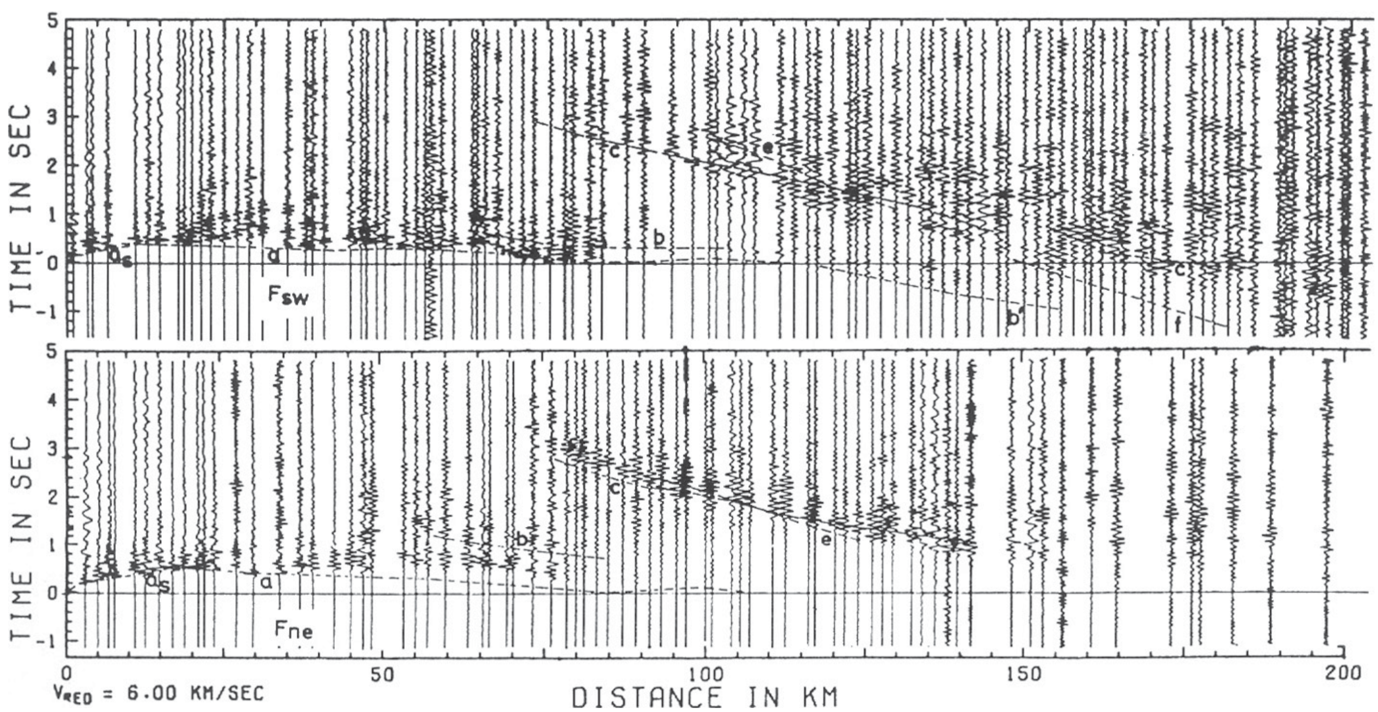


Figure 7.2.6-03. Crustal record sections of the 1979 Rhenish Massif project recorded from shotpoint F towards SW and NE (from Mechie et al., 1983, figs. 2h and 2i). [*In* Fuchs, K., von Gehlen, K., Mälzer, H., Murawski, H., and Semmel, A., eds., Plateau uplift: The Rhenish Massif—a case history: Springer, Berlin-Heidelberg, p. 260–275. Reproduced with kind permission of Springer Science+Business Media.]

In Scandinavia in 1972, following the 1969 observation of the Trans-Scandinavian profile (TSSP 69 in Fig. 7.2.6-05), another long profile had already been recorded through central Scandinavia along the so-called "Blue Road," starting at about the Arctic circle in Norway and traversing in a SSE direction the Caledonian mountain range and adjacent Baltic Shield of northern Sweden (BR 72 in Fig. 7.2.6-05).

Distance ranges of up to 550 km were reached, but in a first interpretation, the authors (Hirschleber et al., 1975) concluded that no fine structure of the P_n wave could be detected, possibly due to the rather thick crust of ~40 km underneath the whole transect. Later, Lund (1979) reinterpreted the data and found a fine structure of the subcrustal lithosphere similar to that found in France and Britain.

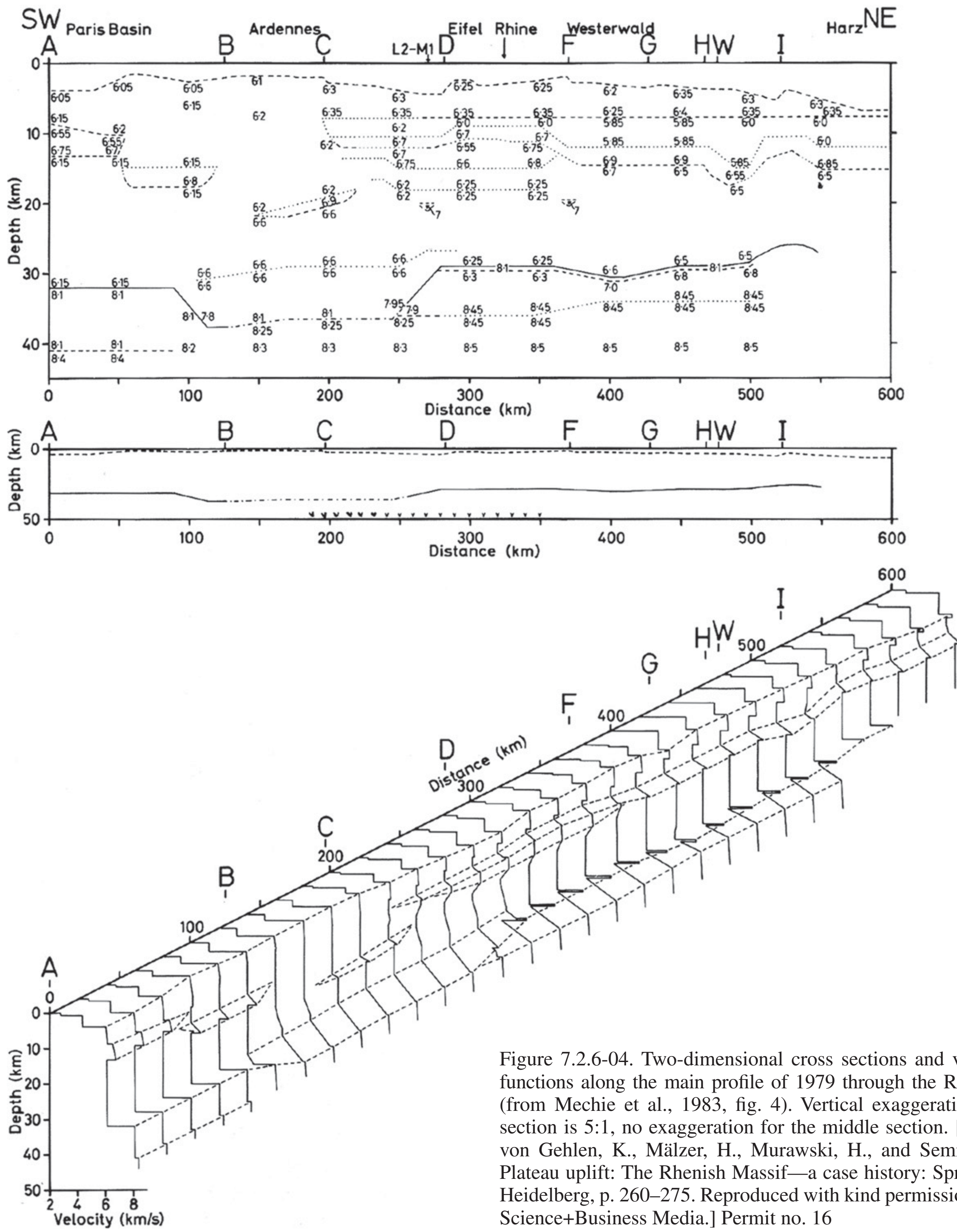


Figure 7.2.6-04. Two-dimensional cross sections and velocity depth functions along the main profile of 1979 through the Rhenish Massif (from Mechie et al., 1983, fig. 4). Vertical exaggeration of the top section is 5:1, no exaggeration for the middle section. [*In* Fuchs, K., von Gehlen, K., Mälzer, H., Murawski, H., and Semmel, A., eds., Plateau uplift: The Rhenish Massif—a case history: Springer, Berlin-Heidelberg, p. 260–275. Reproduced with kind permission of Springer Science+Business Media.] Permit no. 16

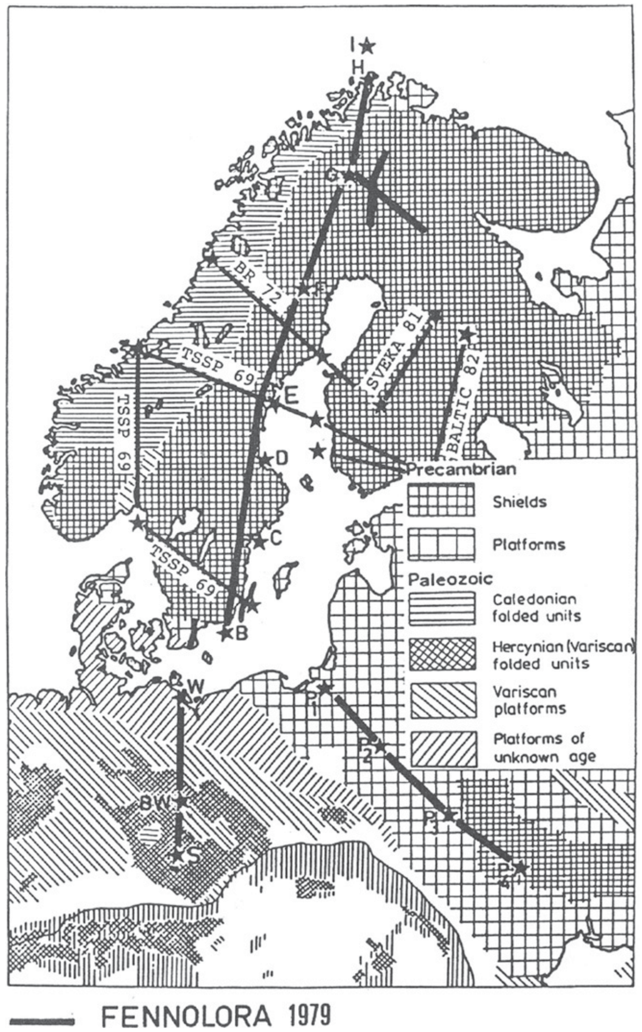


Figure 7.2.6-05. Seismic-refraction lines in Scandinavia. Thick lines: FENNOLORA 1979 including side branches through East Germany and Poland–Ukraine. Thin lines: TSSP 69 Trans-Scandinavian Deep Seismic Sounding project of 1969, BR 72 Blue Road project of 1972, SVEKA 81 and BALTIC 82 Finnish projects of 1981 and 1982 (from Guggisberg et al., 1991, fig. 1). [Tectonophysics, v. 195, p. 105–137. Copyright Elsevier.]

In 1979, the long-range project FENNOLORA (Fennoscandian Long Range Project) was launched in Scandinavia (Fig. 7.2.6-05), aiming for a much deeper penetration of the upper mantle than any previous long-range profile.

One of the goals of FENNOLORA was to reach the mantle transition zone at 400 km depth named by seismologists the 20° discontinuity. Therefore observation distances were laid out to almost 2000 km between shotpoints off the North Cape, and off the southern coast of Sweden (Fig. 7.2.6-05). In parallel, a series of intermediate shots, most of them detonated in the Baltic Sea close to the shore of Sweden, served to obtain all necessary details of crustal and uppermost mantle structure (Fig. 7.2.6-06; Appendix A2-1, p. 69–71 and Appendix A7-4-2).

The FENNOLORA shots were also recorded on two lines to the south through eastern Germany and to the southeast on a line through Poland and the Ukraine (Fig. 7.2.6-05). Furthermore, the map shows profiles in Finland, which were recorded in 1981 and 1982, in the framework of the European Geotraverse which will be discussed in detail in Chapter 8.

The interpretation of the crustal profiles (Fig. 7.2.6-06; Guggisberg et al., 1991; Appendix A2-1, p. 69–71; Appendix A7-4-2, p. A3–A58) showed that the crust is not uniform along the whole line. In the southernmost part of Sweden it was found

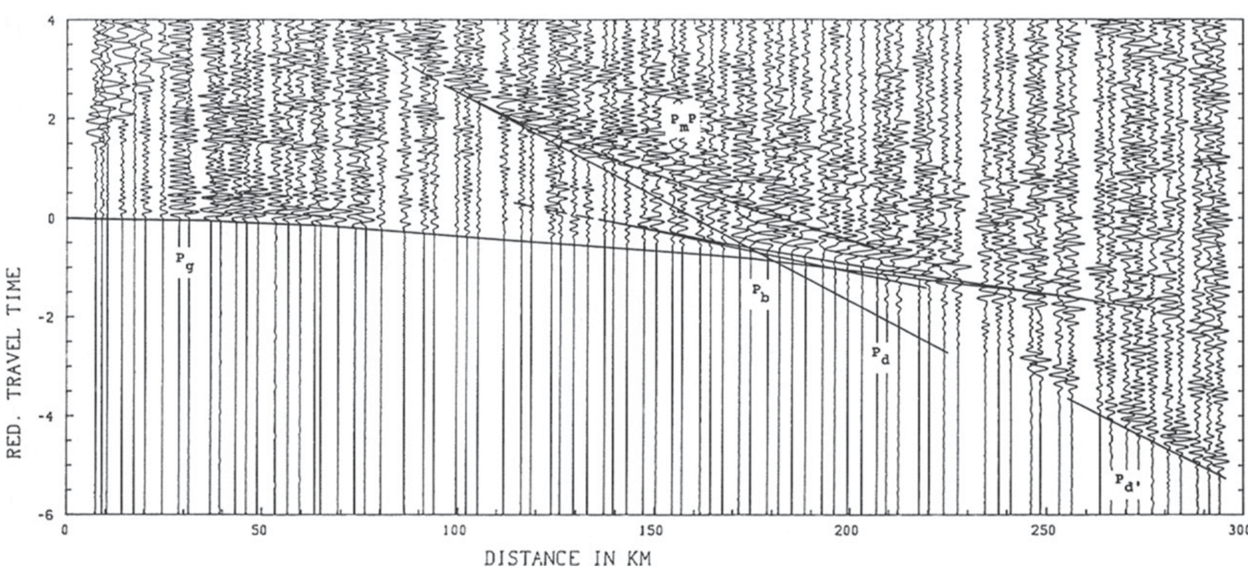


Figure 7.2.6-06. Record section of a FENNOLORA crustal profile, shotpoint B-north, distance range 0 to 300 km, reduction velocity 6 km/s (from Guggisberg et al., 1991, fig. 5). [Tectonophysics, v. 195, p. 105–137. Copyright Elsevier.]

to be less than 40 km thick, but for most of the line, its thickness ranged between 40 and 55 km. There were a few areas with an additional deep lowermost crustal layer with a velocity around 7.5 km/s, and major tectonic peculiarities were mirrored by anomalous features at Moho level (Fig. 7.2.6-07).

The lithosphere-asthenosphere data was interpreted by several authors (e.g., Guggisberg and Berthelsen, 1987; Stangl, 1990). As an appendix to his Ph.D. thesis, Stangl (1990) prepared an Open-File-Report containing all crustal and mantle P- and S-data as record sections (Appendix A7-4-2). Though varying in detail, the overall picture showed a stratified upper mantle where high-velocity areas are intercalated with low-velocity regions with substantial velocity inversions (Fig. 7.2.6-08; Appendix A7-4-2, p. A59–A128).

Of particular interest is a reflection at a distance range of 1650–1900 km reflected from the top of the mantle transition zone (Fig. 7.2.6-09A). Fuchs et al. (1987) produced one of the first velocity-depth models (Fig. 7.2.6-09B) showing a rather complicated velocity-depth structure with the mantle-transition zone at its base, and compared this result with models obtained from other seismological data.

7.2.7. Deep Seismic Sounding Projects in Eastern Europe

With the establishment of the international network of long and detailed seismic-refraction profiles throughout southeastern Europe, initiated in particular by V.B. Sollogub in the 1960s, considerable details of the crustal structure had been obtained. In the 1970s, investigations by deep seismic sounding were particularly continued in Poland due to the engagement of A. Guterch and his co-workers at the Geophysical Institute of the Polish Academy of Sciences in Warszaw. As a continuation and to complement the knowledge obtained, the international profiles VII and VIII were completed in the early 1970s (Fig. 7.2.7-01).

A special study of the Tornquist-Teisseyre tectonic zone was now the goal by observing a series of regional profiles named LT2 to LT5 (e.g., Guterch, 1977; Guterch et al., 1983; Fig. 7.2.7-02). These profiles crossed the Tornquist-Teisseyre tectonic zone perpendicular to its strike, and ran from the fore-Sudetic monocline, or Paleozoic platform, in the southwest to the East European platform in the northeast (Fig. 7.2.7-02). Seismic measurements along these lines were taken using a technique of continuous profiling with distances of 100 or 200 m

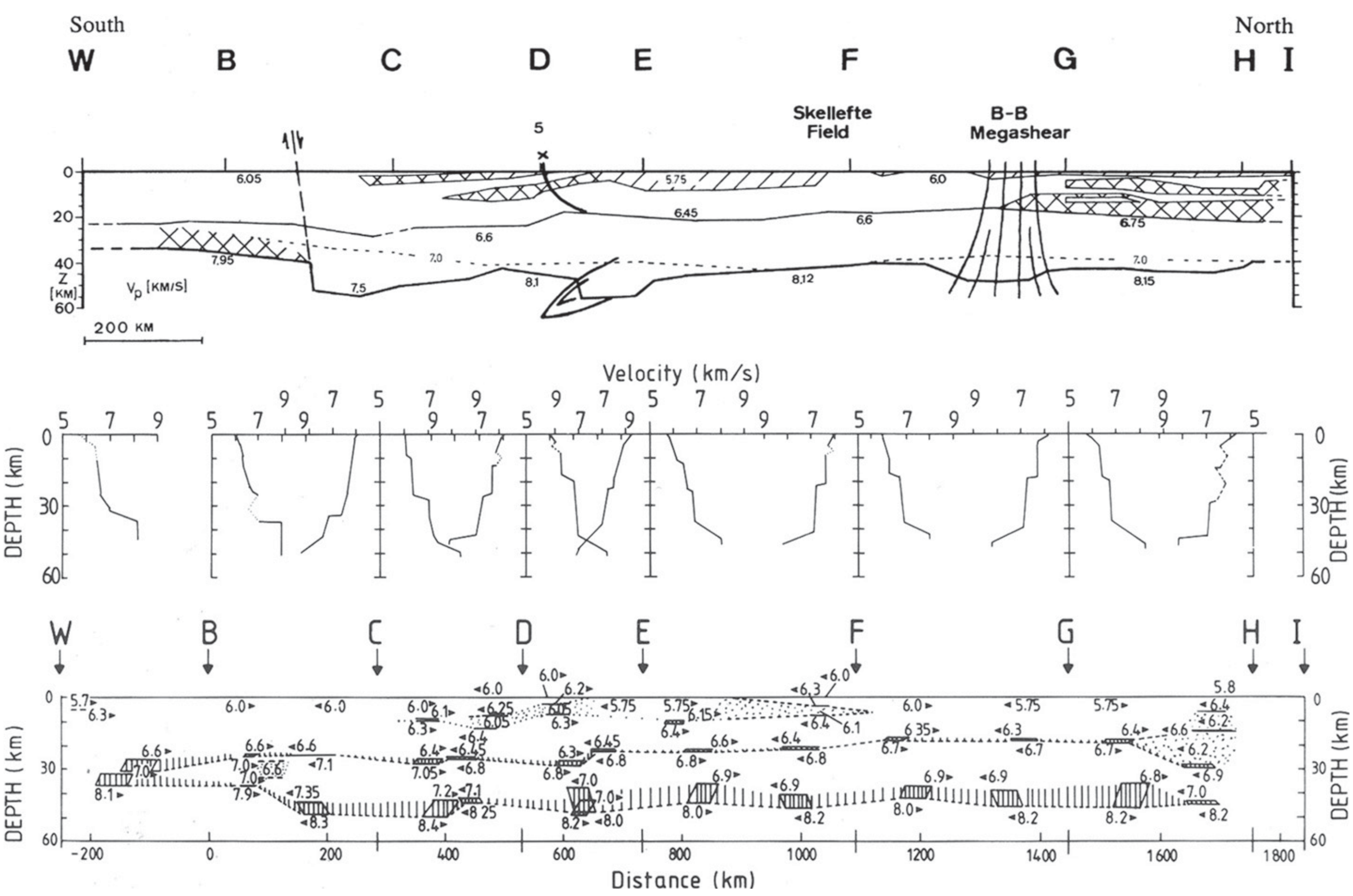


Figure 7.2.6-07. Top: Velocity-depth cross section of the crust underneath FENNOLORA. Center: 1-D velocity-depth functions. Bottom: Depth ranges which generate wide-angle reflections correlated in the data (from Guggisberg et al., 1991, figs. 31 and 18). [Tectonophysics, v. 195, p. 105–137. Copyright Elsevier.]

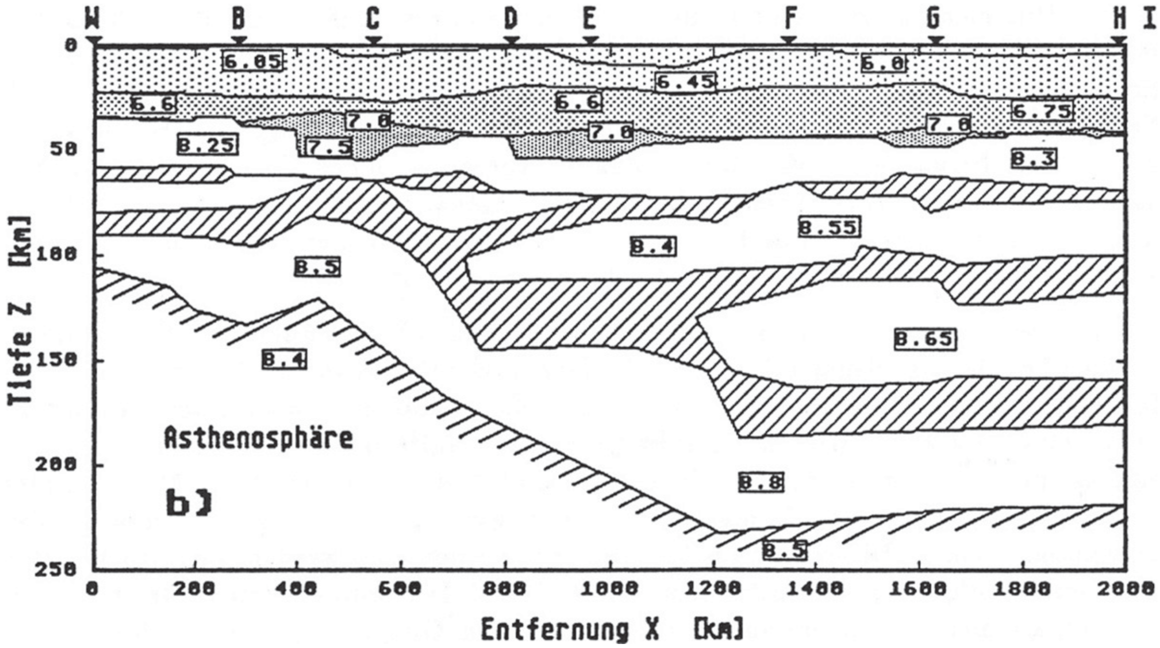


Figure 7.2.6-08. Velocity-depth cross section of the upper mantle underneath FENNOLORA (after Guggisberg and Berthelsen, 1987, fig. 6, from Stangl, 1990, fig. 7–11b). [Ph.D. thesis, University of Karlsruhe, 187 p. Published by permission of Geophysical Institute, University of Karlsruhe, Germany.]

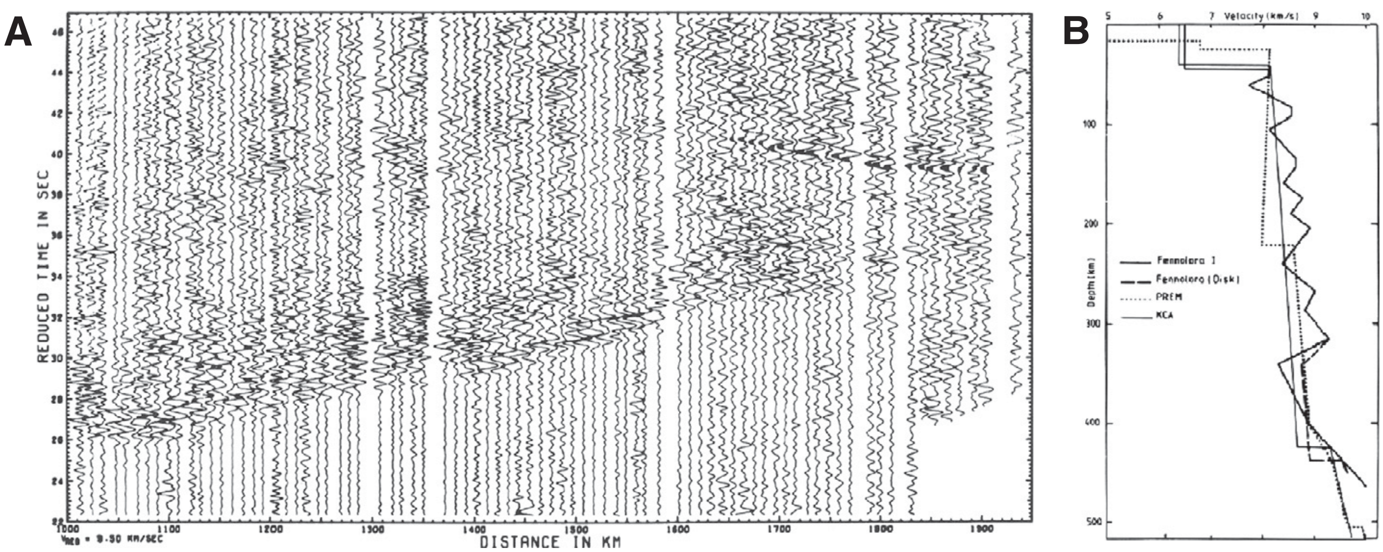


Figure 7.2.6-09. (A) Record section of the FENNOLORA profile, shotpoint I, distance range 1000–1950 km, reduction velocity 9.5 km/s. A later phase near 40 sec reduced traveltime at 1650 and 1900 km is interpreted as reflection from the mantle transition zone (from Fuchs et al., 1987, fig. 5). (B) Preliminary velocity-depth function deduced from the data of shotpoint I, in comparison with other earth models based on seismological data (from Fuchs et al., 1987, fig. 6). [*In* Fuchs, K., and Froidevaux, C., eds., Composition, structure and dynamics of the lithosphere-asthenosphere system: American Geophysical Union Geodynamics Series 16, p. 137–154. Reproduced by permission of American Geophysical Union.]

between recording points and shotpoint intervals of 50–80 km. The shots were recorded in the offset range between 50–70 km and 220–320 km (Guterch, 1974, 1977).

The interpretation of the new LT-profiles included the data of the recently completed international profiles VII and VIII. The densely spaced recordings allowed a reliable phase correlation of different waves, and resulted in a detailed picture of crustal structure in Poland (Fig. 7.2.7-03). The most important outcome of the interpretation was the detection of a 50 km deep Moho trough along the Tornquist-Teisseyre tectonic zone

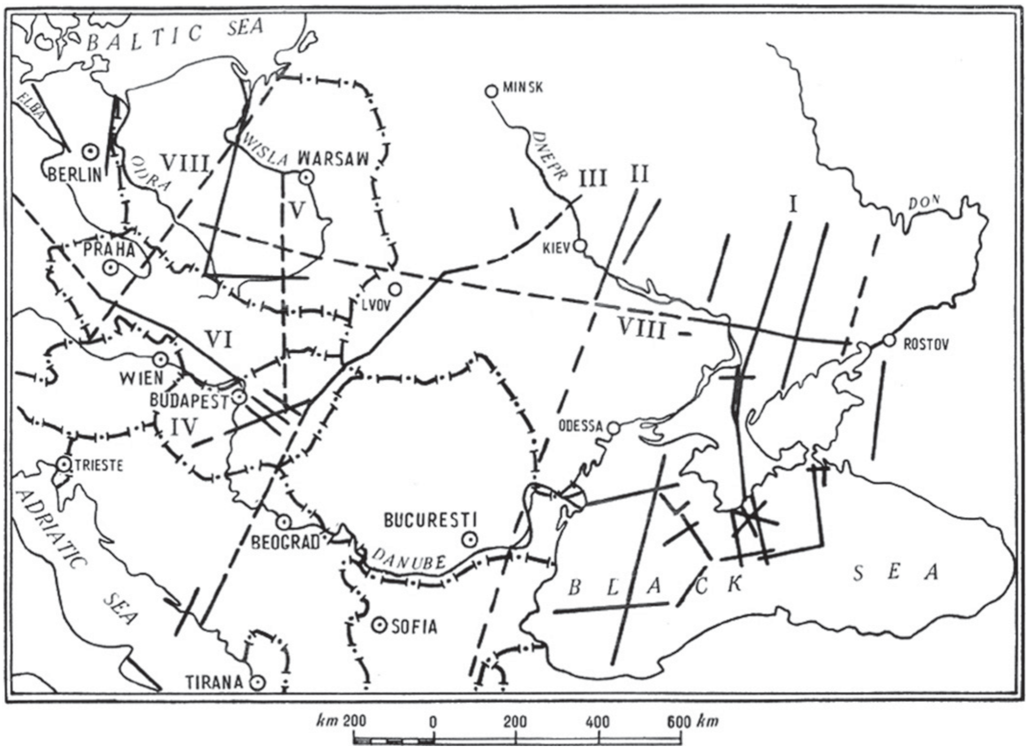
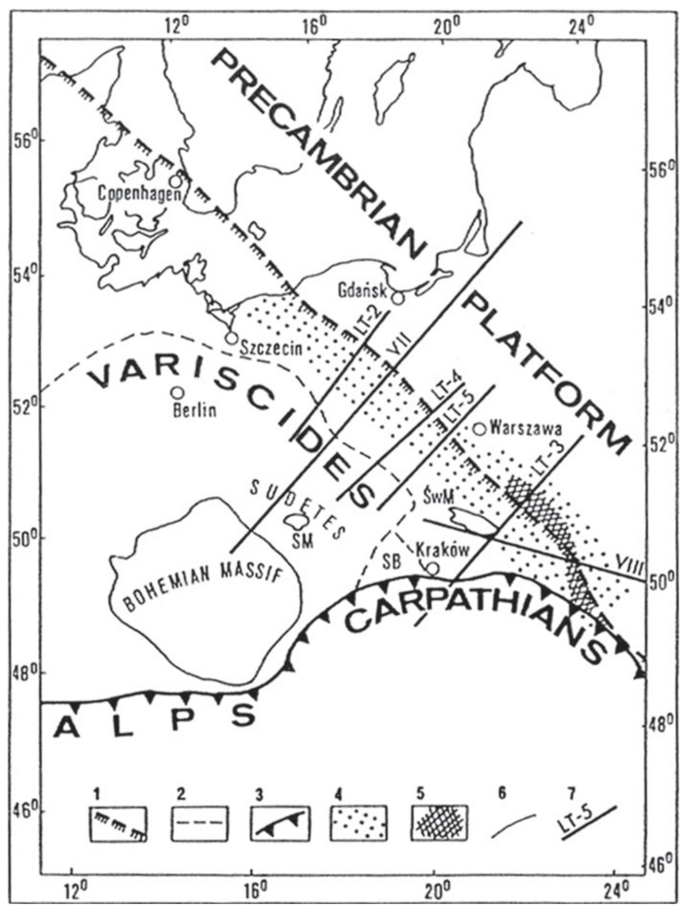


Figure 7.2.7-01. Location of deep seismic profiles and corresponding profiles in southeastern Europe (from Sollogub, 1969, fig. 1). Solid lines: completed in the 1960s, dashed lines: planned and partly completed in the 1970s. Remark by the authors: The NE-SW running profile through Poland and Czechoslovakia is Profile VII. It was erroneously named VIII. [*In* Hart, P.J., ed., The Earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 189–195. Reproduced by permission of American Geophysical Union.]

Figure 7.2.7-02. Location of deep-seismic sounding profiles in Poland around the Tornquist-Teysseire Zone (from Guterch et al., 1991a, fig. 1). [Publs. Institute of Geophysics, Polish Academy of Science, A-19 (236), p. 41–61. Reproduced by permission of Instytut Geofizyki Akademii Nauk, Warsaw, Poland.]



bordered by a 30-km-thick crust under the Hercynian terrain to the southwest and a 40-km-thick crust under the East European Platform to the northeast.

As already mentioned in Chapter 6, in the middle of the 1960s, a special investigation of the fore-Sudetic region had started, aiming primarily to study the sedimentary cover for location of possible oil reserves. The investigation extended over many years into the 1970s, when the interpretation of the complete network became possible (Toporkiewicz, 1986). On selected lines recordings were made up to 90 km distances, thus allowing to study the whole crust (Toporkiewicz, 1986), resulting in a network of deep seismic sounding profiles M1–M13 in Poland (Fig. 7.2.7-04) with a total length of 1500 km. This survey continued into the 1970s and its interpretation was not completed before 1978. The traveltime branches of the crustal phases could be traced up to 60–90 km distance ranges and the resulting models are shown in Figure 7.2.7-04 (Guterch et al., 1991a).

Finally, at the end of the decade, in 1979, the FENNOLORA project prompted the observation of two long-range profiles through eastern Europe (see Fig. 7.2.6-05); one of the lines ran in

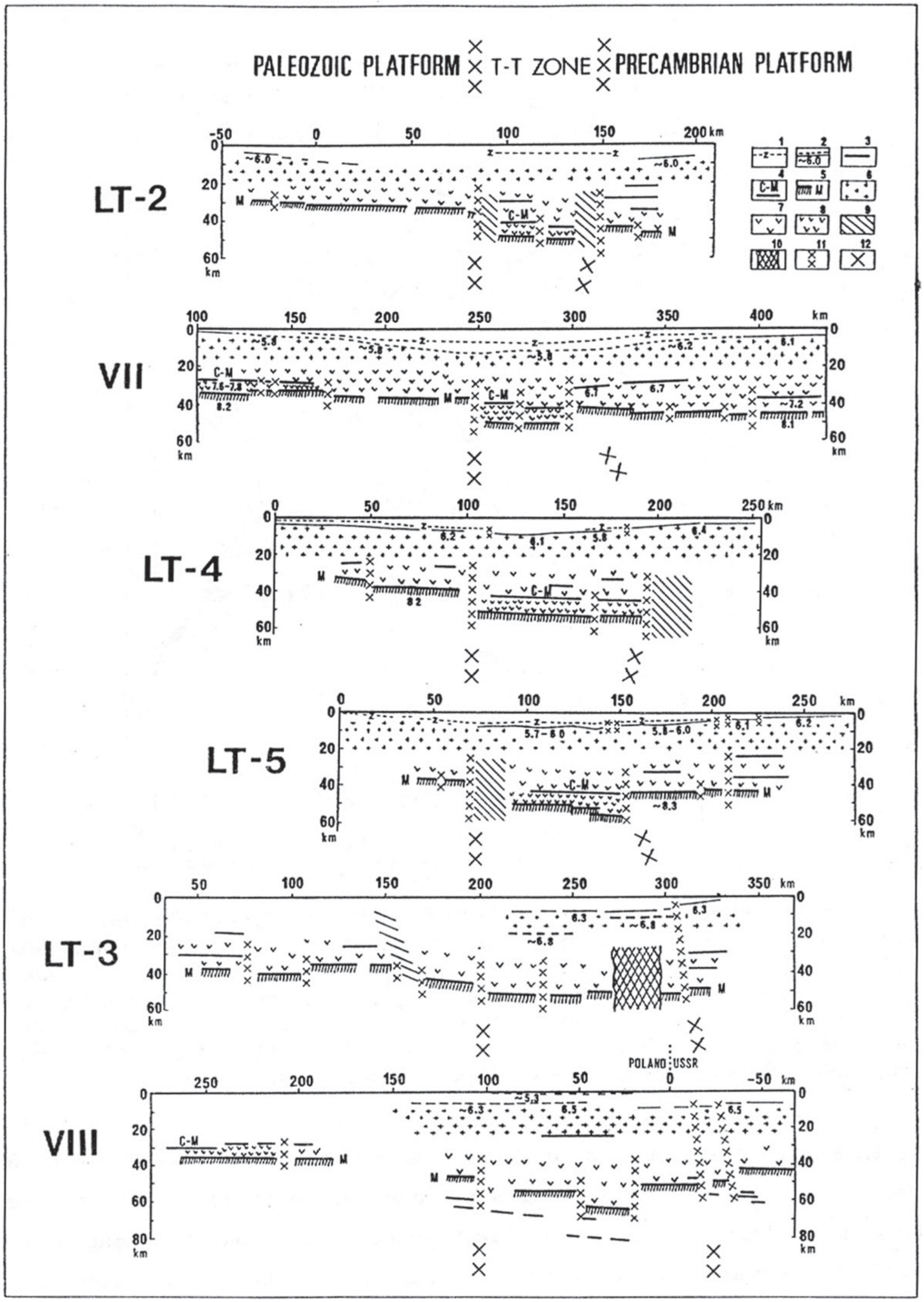


Figure 7.2.7-03. Crustal cross sections through Poland around the Tornquist-Teysseire Zone as interpreted in the 1970s 1Reproduced by permission of Instytut Geofizyki Akademii Nauk, Warsaw, Poland.]

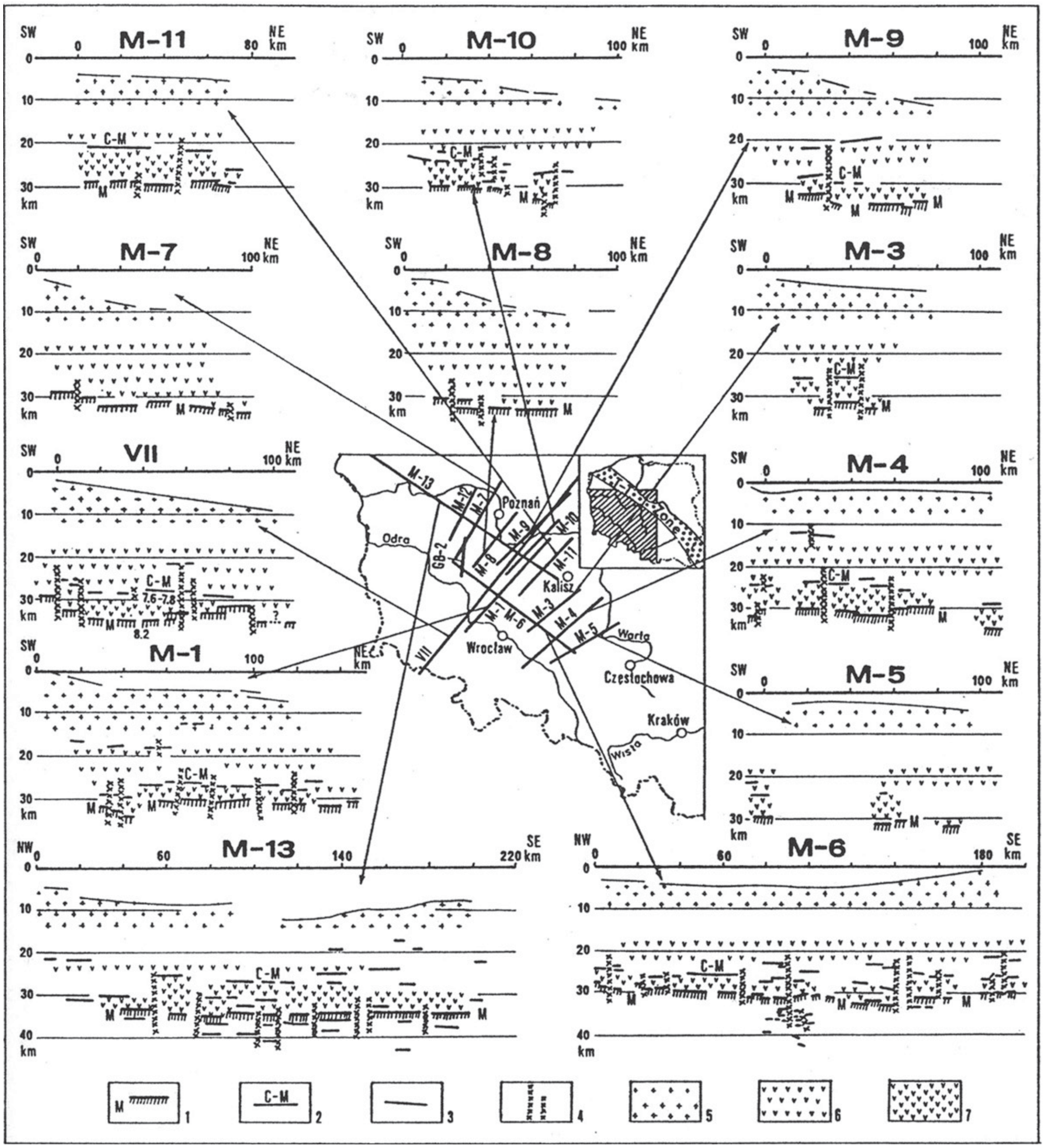


Figure 7.2.7-04. Location of seismic profiles and corresponding crustal cross sections in the fore-Sudetic region, southwestern Poland (from Guterch et al., 1991a, fig. 3). [Publs. Institute of Geophysics, Polish Academy of Science, A-19 (236), p. 41–61. Reproduced by permission of Instytut Geofizyki Akademii Nauk, Warsaw, Poland.]

a NW-SE direction through Poland into the Ukraine. Some of the FENNOLORA shots here were also successfully recorded, and for crustal control, additional shotpoints were organized.

In adjacent Czechoslovakia, after completion of the two International Profiles V and VI mentioned in Chapter 6 (for location see Figs. 6.3.1-01 and 7.2.7-01), the International Profile VII was completed in 1970 (Beránek et al., 1973). The previous profile VII was continued into southeastern Germany to the margin of the eastern Alps by the University of Munich, using the shots along profile VII as energy sources (Miller and Gebrande, 1976). Several seismic regional profiles were recorded between 1975 and 1980 investigating the deep structure of the Western Carpathians, and in 1979, shots of the international project FENNOLORA were also recorded on a N-S line through eastern Germany and through the westernmost Bohemian Massif of Czechoslovakia (Mayerova et al., 1994).

In Romania (Fig. 7.2.7-05), seismic investigations, along the 390-km-long International Profile XI, continued until 1974, and

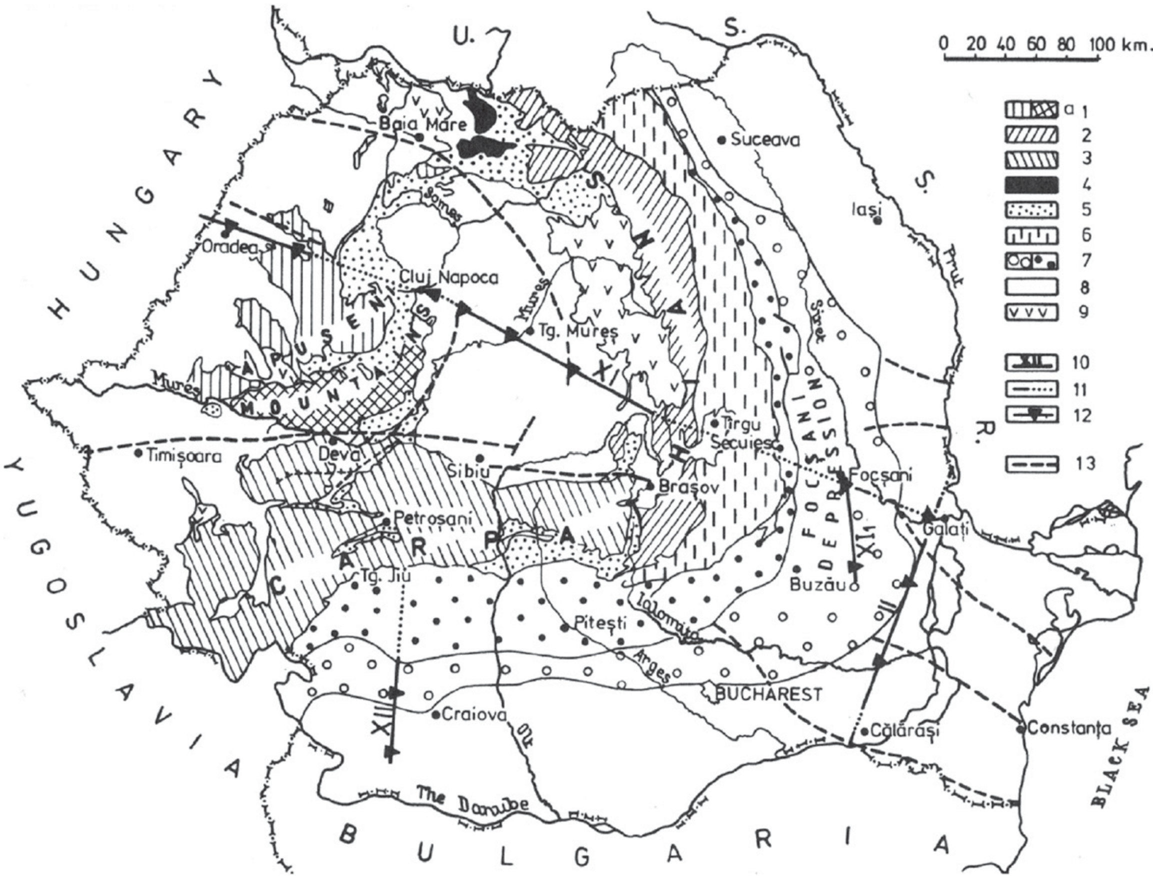


Figure 7.2.7-05. Seismic observations in Romania to the end of the 1970s (from Cornea et al., 1981, fig. 1). Legend: 1–9—geological units, 10—continuous seismic recording, 11—discontinuous seismic recording, 12—shotpoints, 13—tectonic fracture zones. [PAGEOPH, v. 119, p. 144–1156. Reprinted by permission from Birkhaeuser Verlag, Basel, Switzerland.]

were complemented by observations in the Focsani depression (profiles I_1 and II), as well as in the southern Carpathian foreland (line XII). Data examples and a detailed interpretation were published by Radulescu et al. (1976) and Cornea et al. (1981). The cross section along profile XI (Fig. 7.2.7-06) shows the crustal structure from the Pannonian basin in the west to the Moesian platform in the east.

Similar to the interpretation of the seismic lines through Poland by Guterch et al. (1991a), the Romanian crust was divided into crustal blocks, which were separated by deep-reaching fracture zones. The crust thickened to more than 40 km under the Eastern Carpathians and both the easternmost section of the Eastern Carpathians and the Focsani depression were underlain by a 20 km deep sedimentary trough.

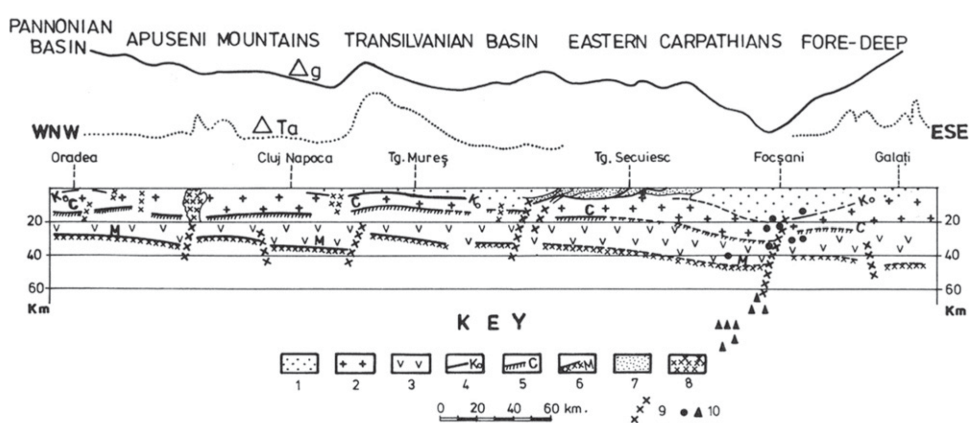


Figure 7.2.7-06. Bouguer gravity and topography (top lines) and seismic crustal cross section through Romania along the international profile XI (from Cornea et al., 1981, fig. 9). 1—sediments, 2—granitic layer, 3—basaltic layer, 4—sediment-basement boundary, 5—Conrad discontinuity, 6—Moho, 7—Eastern Carpathian nappes, 8—volcanic rocks, 9—fracture zones, 10—crustal and subcrustal earthquake hypocenters. [PAGEOPH, v. 119, p. 144–1156. Reprinted by permission from Birkhaeuser Verlag, Basel, Switzerland.]

7.3. DEEP-SEISMIC SOUNDING IN EASTERN EUROPE AND ADJACENT ASIA (USSR)

In the USSR, deep seismic sounding operations were drastically reduced in the 1970s. It was a decade when new methods and interpretation programs were developed, and when the methods of field experimentation were also changed.

In the southwestern USSR, the network of international and national profiles, which had been initiated and pushed forward by V.B. Sollogub in the 1960s, was supplemented in the early 1970s by some new measurements, and individual interpretations were prepared (see, e.g., Grad and Tripolsky, 1995).

Since 1972, the Voronezh State University conducted seismic investigations across the Voronezh shield which represents a 300–500-km-wide horst of Archean and Proterozoic highly dislocated supracrustal and magmatic rocks of complex tectonic nature (Tarkov and Basula, 1983). Commercial quarry blasts were used for the investigation. The seismic signals were tape-recorded using 6-channel telemetric arrays of type Taiga with 150 m geophone spacing at an average station spacing of 5–10 km. Several profiles were recorded throughout the area, all recording wide-angle reflections from the Moho at distances between 90 and 190 km. The interpretation revealed an upper crust of 12–15 km in thickness, the middle crust consisted of one or two layers with velocities from 6.35 to 6.85 km/s and the lower crustal velocities ranged from 6.8 to 7.95 km/s within a depth range from ~32 km to 43–45 km (Moho depth).

The Novosibirsk group led by N.N. Puzyrev applied, for example, the new method of differential time fields to explore almost inaccessible areas of Siberia and the Baikal rift zone. Here the source and several arrays of receivers moved along a profile, keeping the distance between them optimal for waves reflected from basic crustal boundaries (for references, see Pavlenkova,1996).

An overview of seismic observations made in Eastern Siberia around Lake Baikal in the late 1960s and early 1970s (Fig. 7.3-01) was compiled by Olsen (1983). The dashed lines in Figure 7.3-01

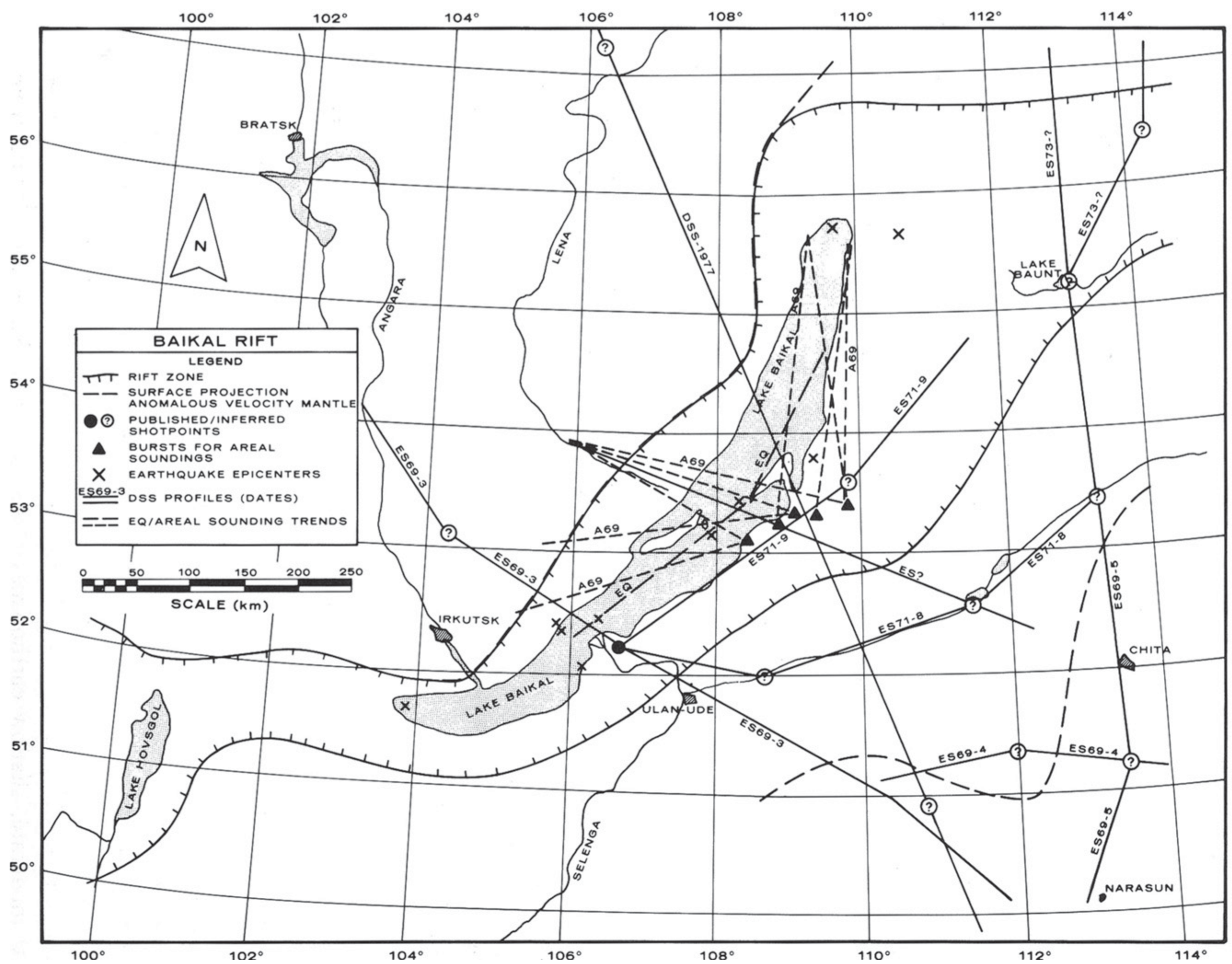


Figure 7.3-01. Map of the Lake Baikal area of eastern Siberia showing the approximate boundary of the Baikal rift zone, the surface projection of anomalous mantle material near Moho depths, and published seismic refraction lines employing both explosion and earthquake sources (from Olsen, 1983, fig. 4). [Tectonophysics, v. 94, p. 349–370. Copyright Elsevier.]

point from sources east of Lake Baikal to receivers which were distributed fan-like in order to study the three-dimensional structure beneath central and northern Lake Baikal. In this review by Olsen (1983), the original references can also be found.

Based on powerful explosions on land (nuclear devices, as became known after 1990), a number of profiles of 1000–2000 km in length were completed (Zverev and Kosminskaya, 1980; Vinnik and Ryaboi, 1981; Egorkin and Pavlenkova, 1981; see also Fuchs and Vinnik, 1982; Fuchs et al., 1987; Mechie et al., 1993), which were recorded by multichannel instruments with an average spacing of ~20 km.

A major part of these profiles was carried out throughout northwest Asia, across the West Siberian plate and the East Siberian platform as well as through central Turkmenia, central Kazakhstan, and other high-mountain areas of the USSR (Fig. 7.3-02). Both Kosminskaya and Pavlenkova (1979) and Zverev and Kosminskaya (1980) have published a representative collection of crustal and uppermost mantle cross sections for various

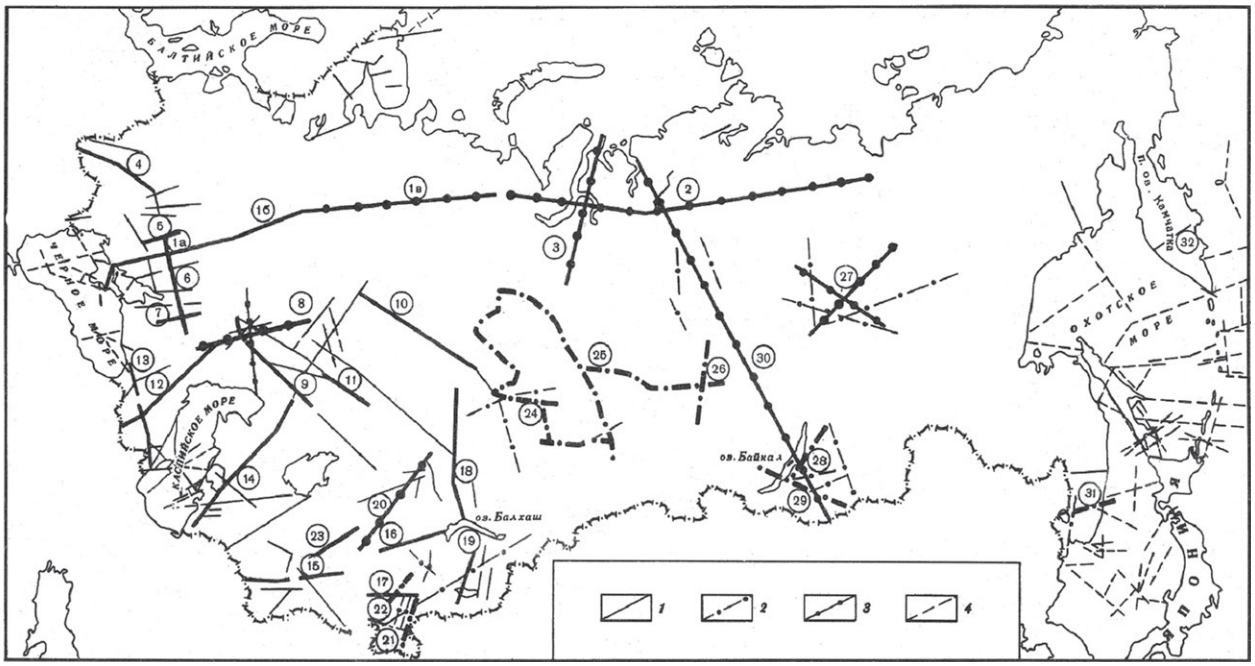


Figure 7.3-02. Location map of the main profiles recorded in the USSR between 1960 and 1978 (from Zverev and Kosminskaya, 1980, fig. 1). For references see Zverev and Kosminskaya (1980, table 1). [Publ. House NAUKA, Moscow, 180 p. Reproduced by permission of the Schmdt Institute of Physics of the Earth RAS.]

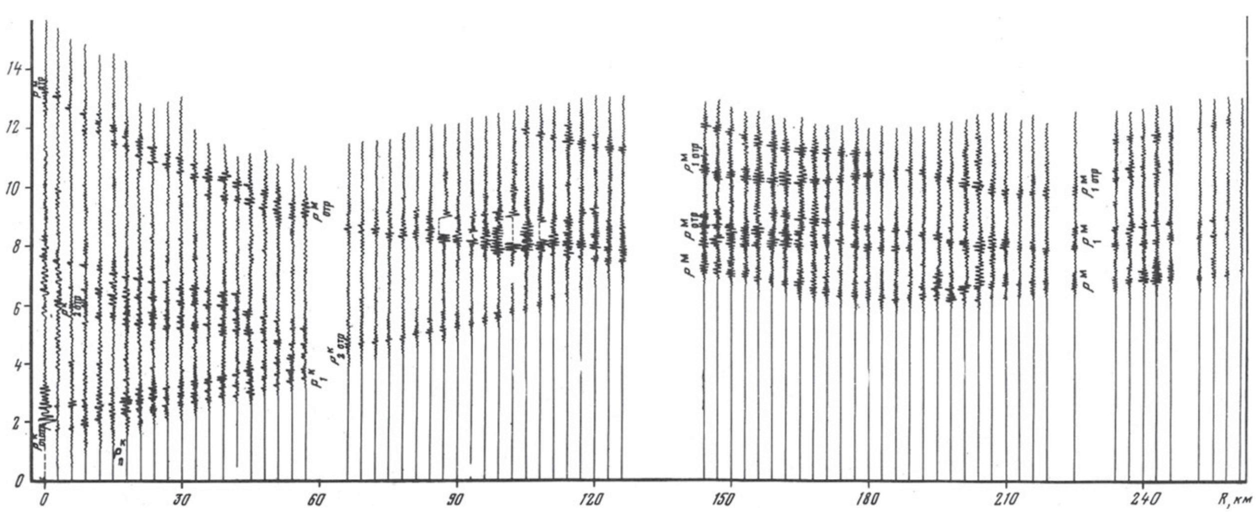


Figure 7.3-03. Data example of a profile recorded in 1970 through the Urals (from Zverev and Kosminskaya, 1980, fig. 50). [Publ. House NAUKA, Moscow, 180 p. Reproduced by permission of the Schmdt Institute of Physics of the Earth RAS.]

regions of the USSR. Data and a cross section of a profile recorded in the Urals in 1970 and published by Zverev and Kosminskaya (1980) are shown in Figures 7.3-03 and 7.3-04. In this context, a 700-km-long line across the Barents Sea also was recorded in 1976 (Davydova et al., 1985) to the northeast from Murmansk which consisted of a system of reversed and overlapping 200–400-km-long profiles.

The analysis of the long-range profiles not only revealed crustal structure information, but also showed pronounced differences in the structure of the lower lithosphere. Egorkin and Pavlenkova (1981) published a complete data set of a long-range profile across the East European platform with recordings up to 2800 km and discussed the inhomogeneous structure of the subcrustal lithosphere using the example of a profile through Kazakhstan (Fig. 7.3-05).

Examples of subcrustal velocity-depth functions for eastern Europe and northern Asia published until 1982 (Fig. 7.3-06) have been compiled, e.g., by Prodehl (1984; see also Fuchs et al., 1987).

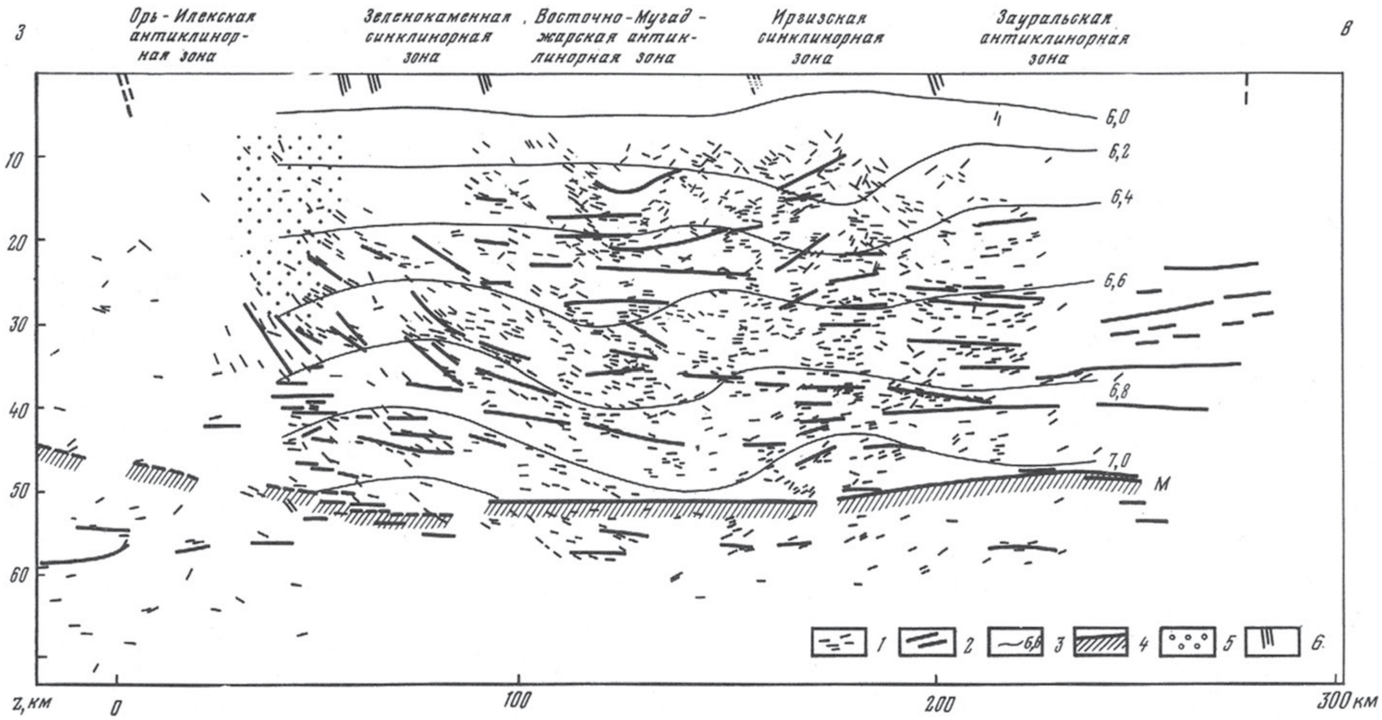


Figure 7.3-04. Crustal cross section along profile no 11 (Fig. 7.3-02) recorded in 1970 through the Urals (from Zverev and Kosminskaya, 1980, fig. 53). For references see Zverev and Kosminskaya (1980, Table 1). [Publ. House NAUKA, Moscow, 180 p. Reproduced by permission of the Schmdt Institute of Physics of the Earth RAS.]

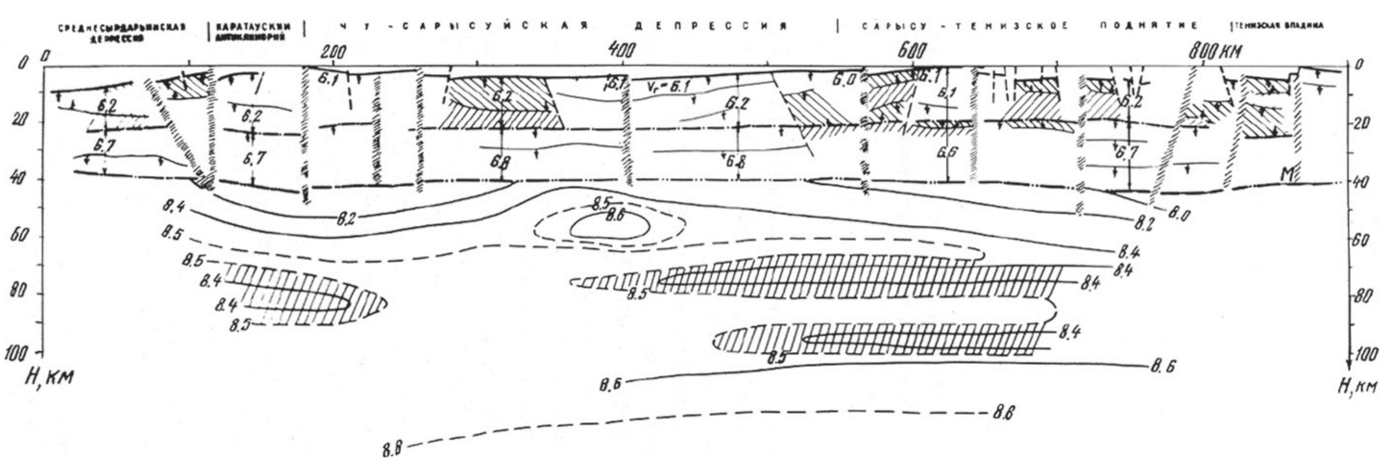


Figure 7.3-05. Seismic section of a long-range profile through Kazakhstan (from Egorkin and Pavlenkova, 1981, fig. 9). [Physics of the Earth and Planetary Interiors, v. 25, p. 12–26. Copyright Elsevier.]

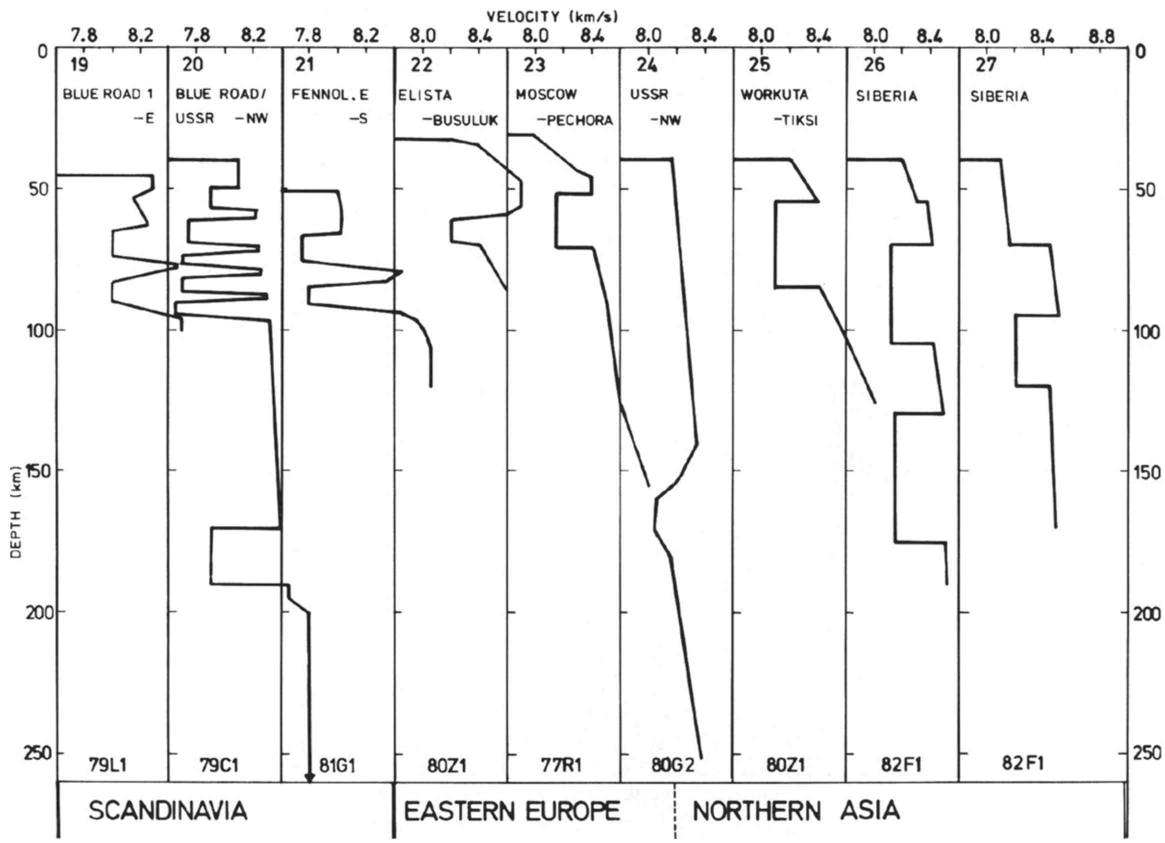


Figure 7.3-06. Selected velocity-depth models of upper-mantle structure under Scandinavia and the USSR (from Fuchs et al., 1987, fig. 2). [*In* Fuchs, K., and Froidevaux, C., eds., Composition, structure and dynamics of the lithosphere-asthenosphere system: American Geophysical Union Geodynamics Series 16, p. 137–154. Reproduced by permission of American Geophysical Union.]

7.4. CONTROLLED-SOURCE SEISMOLOGY IN NORTH AMERICA

7.4.1. Crustal Seismic-Refraction Work in the United States

Following the intensive seismic-refraction research activities in the 1960s in North America, the detection and development of plate tectonics turned much of the seismic and seismological activities at universities toward the oceans, and the interest in explosion seismology on continents in North America dropped. For example, the move of A.L. Hales from Dallas to Australia shifted the interest of the remaining scientists into other fields of geophysics. Furthermore, the major seismic activities of the U.S. Geological Survey, from 1966 onward, had moved into earthquake research with emphasis on the tectonically active west.

The interpretation of seismic data, which the U.S. Geological Survey had assembled throughout the United States in the 1960s, as described in Chapter 6 continued. Both for the Rocky Mountain and the Appalachian surveys, new models were prepared (Prodehl and Pakiser, 1980; Prodehl et al., 1984), first versions of which had been compiled (Fig. 7.4.1-01) in a foregoing publication as a synthesis of crustal data (Prodehl, 1977).

Only a few centers remained active. Here, however, active seismology was pursued with much emphasis. For example, at the University of Wisconsin, e.g., R.P. Meyer and his group continued their activities on controlled-source seismology in the Andean regions of South America. In Cornell University at Ithaca, New York, J.A. Oliver, S. Kaufman, J.A. Brewer, L. Brown, and others started a continent-wide deep seismic-reflection program, an approach which was also pursued by M.J. Berry and co-workers at the Earth Physics Branch of the Department of Energy, Mines and Resources at Ottawa, Canada.

At the University of Utah at Salt Lake City, R.B. Smith with his co-workers, L.W. Braile, G.R. Keller, and others had a particular interest in the exploration of the crust and mantle in the tectonically active western United States with emphasis on the Basin and Range province, the Colorado Plateau, and the middle Rocky Mountains. A major source for the investigation of the crustal structure of the Wasatch Mountains and Wasatch Front

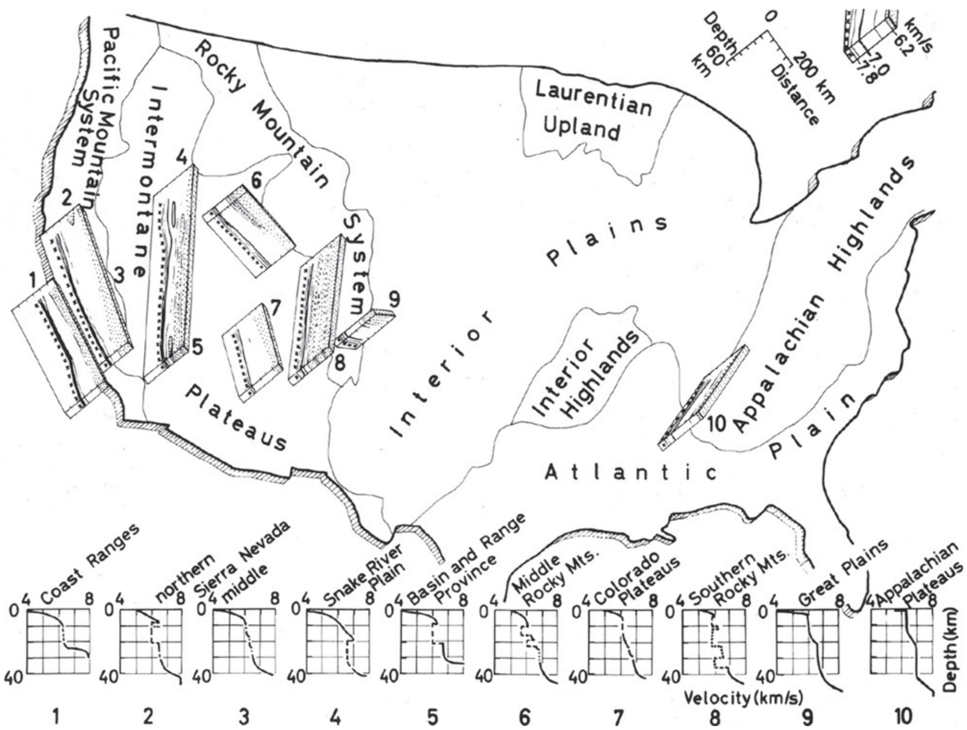


Figure 7.4.1-01. Compilation of selected crustal cross sections and representative velocity-depth functions for the U.S. Geological Survey seismic refraction data of the 1960s (from Prodehl, 1977, fig. 7). [*In* Heacock, J.G., ed., The Earth's crust: American Geophysical Union Geophysical Monograph 20, p. 349–369. Reproduced by permission of American Geophysical Union.]

was the regular quarry blasts in the Bingham copper mine (A in Fig. 7.4.1-02). In 1971 and 1972, several seismic-refraction profiles could be completed (lines A–C and A–B in Fig. 7.4.1-02; Braile et al., 1974; Keller et al., 1975; Appendix A7-5-1).

Finally, Yellowstone Park and the adjacent eastern Snake River Plain, which R.B. Smith claimed to be an active hot spot, became the focus of a major research project (e.g., Smith and Christiansen, 1980, Smith and Braile, 1994).

The seismic equipment available to universities in the late 1970s, throughout the United States, however, was not sufficient and not available in numbers great enough for large-scale experiments. Also, the powerful equipment of the U.S. Geological Survey of the 1960s was aging due to its extensive use.

So, in 1977, R.B. Smith asked his colleagues in Europe for help and in 1978, the first U.S.-European cooperation started with a joint venture of the University of Utah, Salt Lake City, and Purdue University in West Lafayette, Indiana, with the ETH Zurich, Switzerland, and the University of Karlsruhe, Germany.

The Europeans shipped their MARS-66 equipment to Salt Lake City to increase the number of portable recording stations and thus to enable effective seismic-refraction work in the Yellowstone–Snake River Plain area. Fifteen shots were recorded that provided coverage to distances of 300 km (Fig. 7.4.1-03).

The fieldwork of 1978 was followed by a second stage in 1980 (Braile et al., 1982; Smith et al., 1982; Sparlin et al., 1982).

Most of the data showed extremely clear wide-angle reflections, both in the eastern Snake River Plain (Fig. 7.4.1-04) and in the Yellowstone Park area (Fig. 7.4.1-05). More data and other details of the experiment are provided in Appendix A7-5-2. The

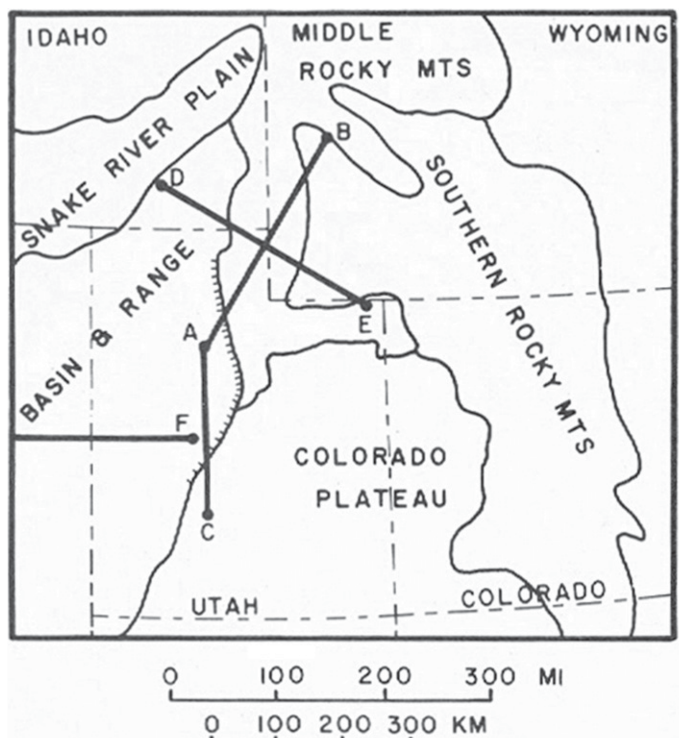


Figure 7.4.1-02. Location of seismic refraction profiles around the Wasatch Front, western United States (from Braile et al., 1974, fig. 1). A—Bingham Copper Mine, B—Fremont Lake, C—Piute Reservoir, D—American Falls Reservoir, E—Flaming Gorge Reservoir, F—Delta, Utah. [Journal of Geophysical Research, v. 79, p. 2669–2677. Reproduced by permission of American Geophysical Union.]

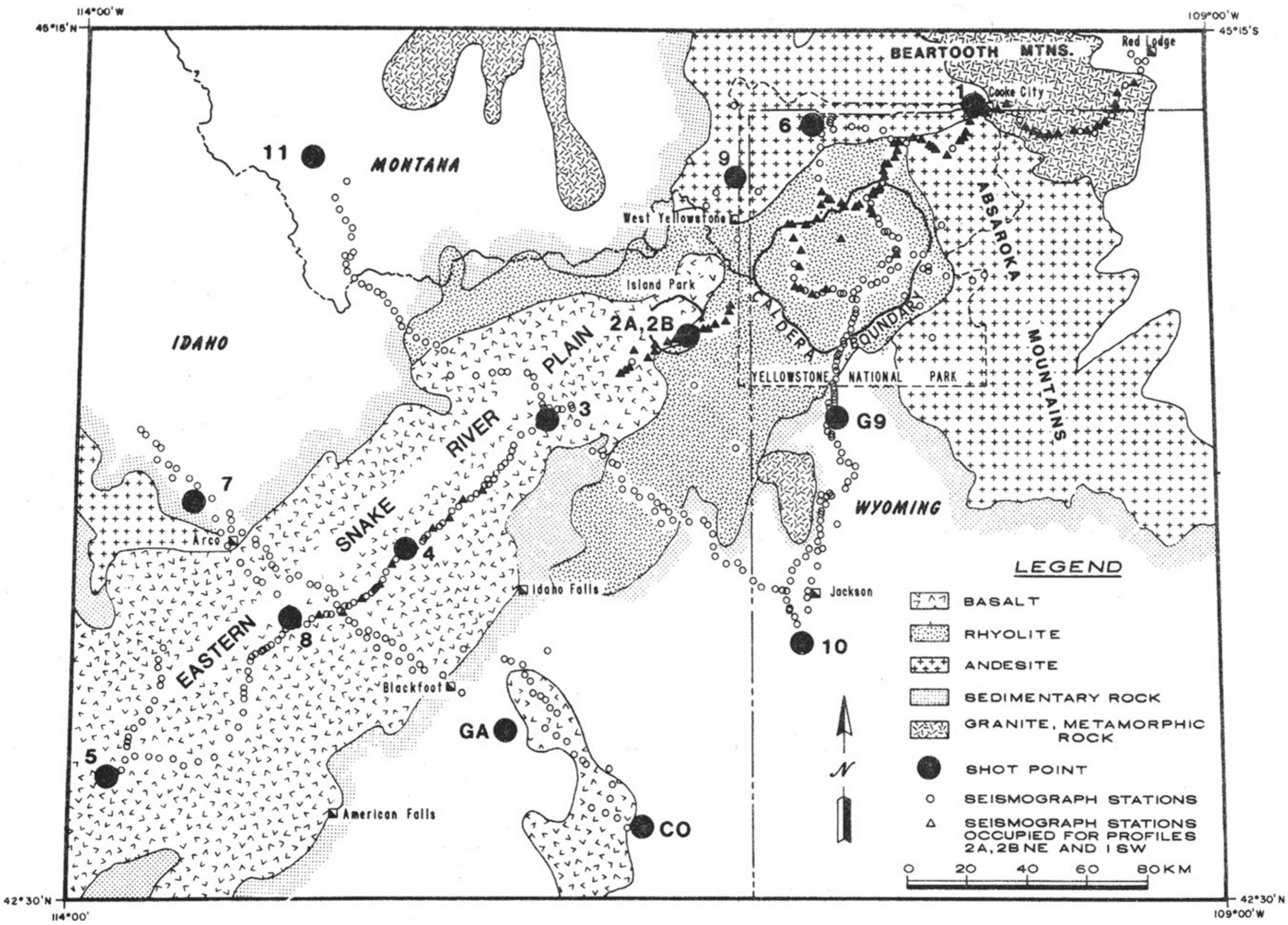


Figure 7.4.1-03. Location of the seismic refraction observations in the Yellowstone–Snake River Plain (from Smith et al., 1982, fig. 1). [Journal of Geophysical Research, v. 87, p. 2583–2596. Reproduced by permission of American Geophysical Union.]

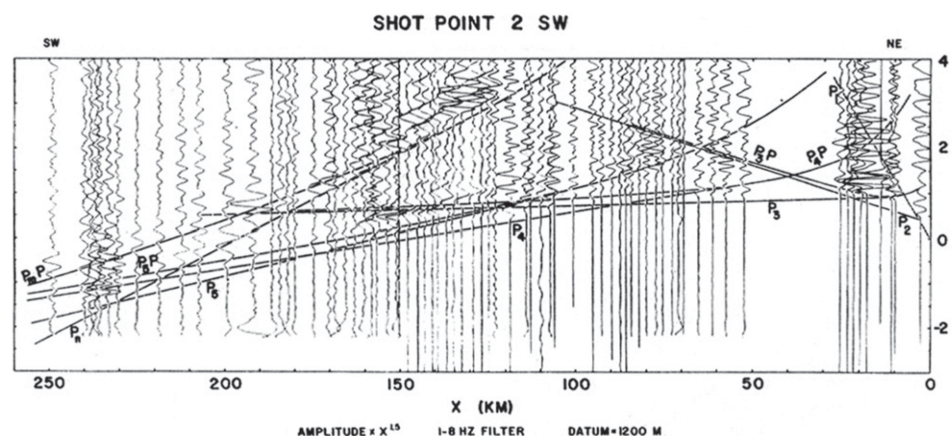


Figure 7.4.1-04. Record section of seismic refraction observations from shotpoint 2 towards southwest through the eastern Snake River Plain (from Braile et al., 1982, fig. 2a). [Journal of Geophysical Research, v. 87, p. 2597–2609. Reproduced by permission of American Geophysical Union.]

interpretation showed for the upper 10 km of the crust strong lateral inhomogeneities, probably reflecting the effects of a major lithospheric anomaly, which is evidenced by large volumes of volcanic rocks, and the systematic progression of the silicic volcanism along the eastern Snake River Plain to its present position beneath Yellowstone. The intermediate and lower crustal layers proved to be more homogeneous, and the total crustal thickness underneath the Island Park–Yellowstone Plateau–Beartooth Mountains was ~43 km, which increased slightly into the Snake River Plain to the southwest (Fig. 7.4.1-06).

Increased interest in the nature of the Rio Grande rift led to a first seismic-refraction study in 1976, organized and interpreted

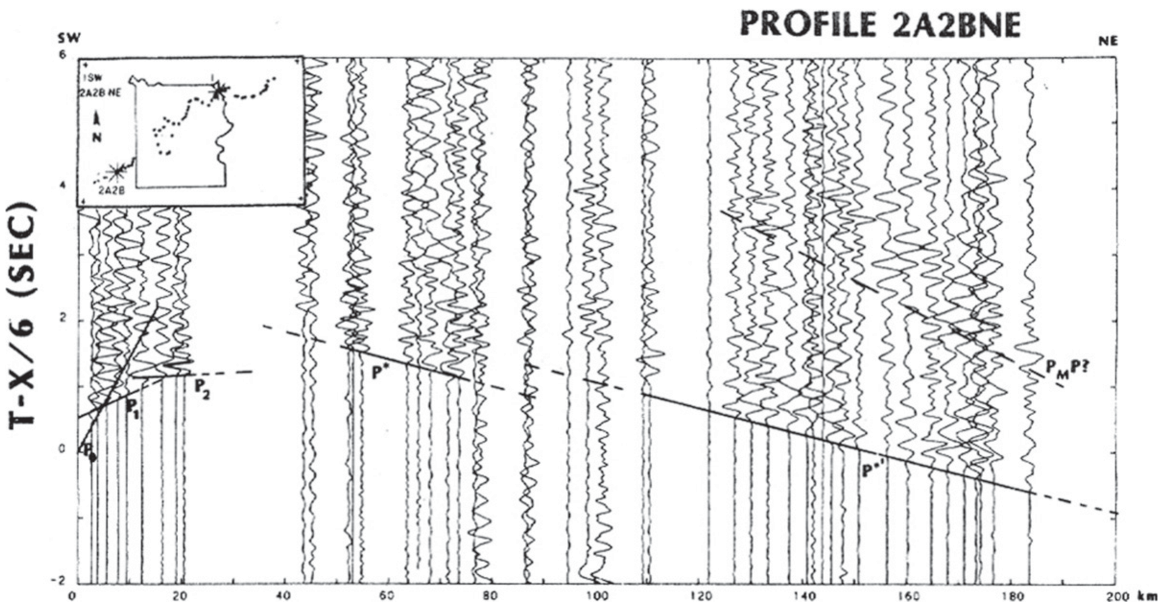


Figure 7.4.1-05. Record section of seismic refraction observations from shotpoint 2 towards northeast through the Yellowstone Park (from Smith et al., 1982, fig. 3a). [Journal of Geophysical Research, v. 87, p. 2583–2596. Reproduced by permission of American Geophysical Union.]

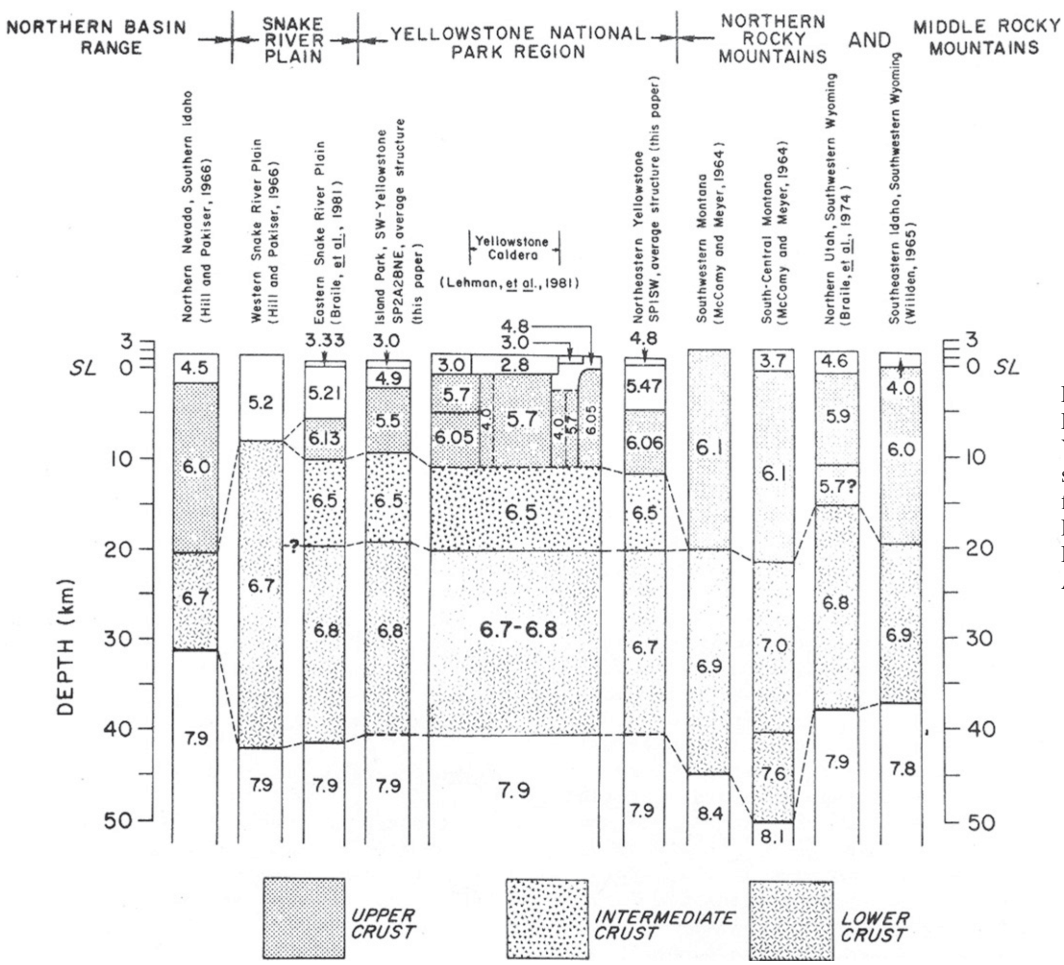


Figure 7.4.1-06. Generalized P wave crustal model for the Yellowstone–Snake River Plain survey (from Smith et al., 1982, fig. 8). [Journal of Geophysical Research, v. 87, p. 2583–2596. Reproduced by permission of American Geophysical Union.]

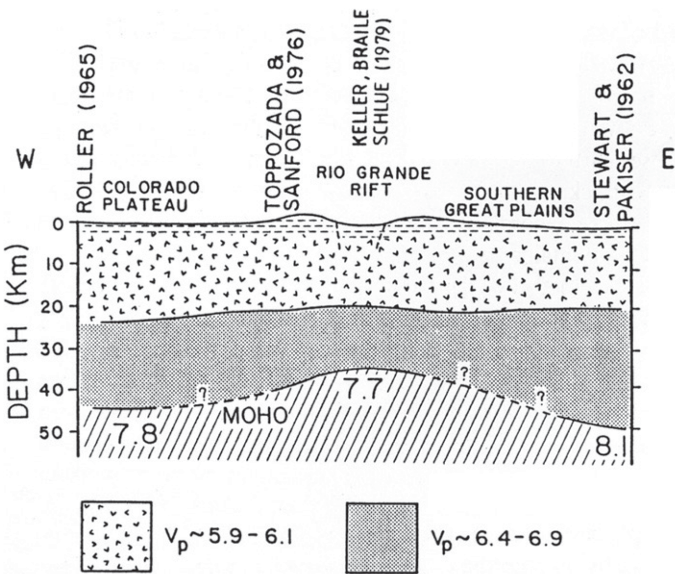


Figure 7.4.1-07. Schematic cross section across the Rio Grande rift in the Santa Fe–Albuquerque area (from Keller et al., 1979, fig. 6). [*In* Riecker, R.E., ed., Rio Grande rift; tectonics and magmatism: Washington, D.C., American Geophysical Union, p. 115–126. Reproduced by permission of American Geophysical Union.]

by K. Olsen of the Los Alamos National Laboratory, using the shotpoint DICE THROW north of Albuquerque, New Mexico, for a north-directed, single-ended profile along the eastern graben margin (Olsen et al., 1979, 1982; Olsen, 1983). Though only recorded with a large station separation of ~10 km, this profile (Appendix A7-5-3) provided information that the crust beneath the axis of the rift, with a thickness of 34 km, was ~10–15 km thinner than under the adjacent Great Plains and Colorado Plateau.

Also in the southern Rio Grande Rift, an unreversed seismic-refraction line was observed (Cook et al., 1979b; Appendix A7-5-3). It was recorded from mine blasts near Santa Rita, New Mexico, across the rift at ~32.9° latitude in a west-east direction and reached a maximum distance of 220 km.

From the data available, Keller et al. (1979) sketched an upward warping of the Moho underneath the Rio Grande rift (Fig. 7.4.1-07). In 1978, new data became available for the Basin and Range area in New Mexico and Arizona to the west of the Rio Grande rift (Gish et al., 1981; Sinno et al., 1981), which provided confirmation, e.g., of an earlier interpretation of crustal data around the Tonto Forest Seismological Observatory by Warren (1969), and showed that the crust thins rapidly from near 37 km along the Mogollon rim, the border of the Colorado Plateau toward the south, to less than 25 km south of Phoenix.

Involving all seismic-refraction and seismic-reflection data available by the early 1980s in the United States, including the data of the 1981 CARDEX project in the Rio Grande rift (Ankeny et al., 1986; Sinno et al., 1986; see Chapter 8), Prodehl and Lipman (1989) compiled a north-south section through the Rocky Mountains and Rio Grande rift (Fig. 7.4.1-08).

As mentioned above, scientists of the U.S. Geological Survey in the 1970s were mainly involved in earthquake studies with emphasis on active regions, California in particular. Over the years, the local seismic network along the San Andreas fault zone had become so dense that it became feasible to use these stations, in combination with suitable local earthquakes, as quasi-refraction lines (Fig. 7.4.1-09).

A successful approach in this direction was undertaken for the Coast Ranges south of San Francisco. Prodehl (1976) had compiled the recordings of a series of medium-size (magnitude 2.7–3.8) local earthquakes of 1975 as record sections. This collection was the base for a crustal study by Blümling and Prodehl (1983; Appendix A7-5-4) who used part of the existing seismic-refraction data (Fig. 7.4.1-10), which was recorded in 1967 (Stewart, 1968a) and which had been reinterpreted by Walter and Mooney (1982), and interpreted them together with some of the local earthquake data (Fig. 7.4.1-11).

A particular result of Blümling and Prodehl (1983) for both the refraction and the near-earthquake profiles was a velocity inversion in the lowermost crust between 21 and 26 km depth with velocities as low as 5.3. km/s overlying a 3–5-km-thick mantle transition zone with 7.6 km/s average velocity on top of the Moho at 29–30 km depth with 8.2 km/s.

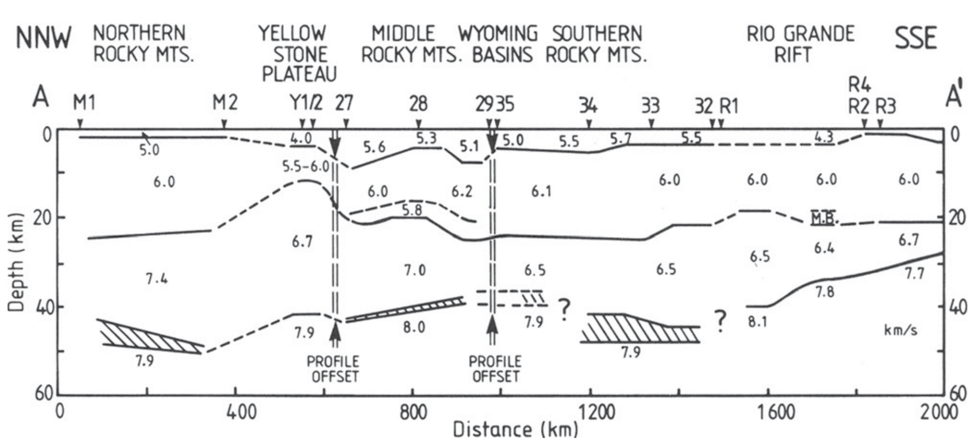


Figure 7.4.1-08. Simplified cross section through Rocky Mountains and Rio Grande rift between 33° to 49°N latitude and 105° to 115°W longitude (from Prodehl and Lipman, 1989, fig. 23). [*In* Pakiser, L.C., and Mooney, W.D., eds., Geophysical framework of the continental United States: Geological Society of America Memoir 172, p. 249–284. Reproduced by permission of the Geological Society of America.]

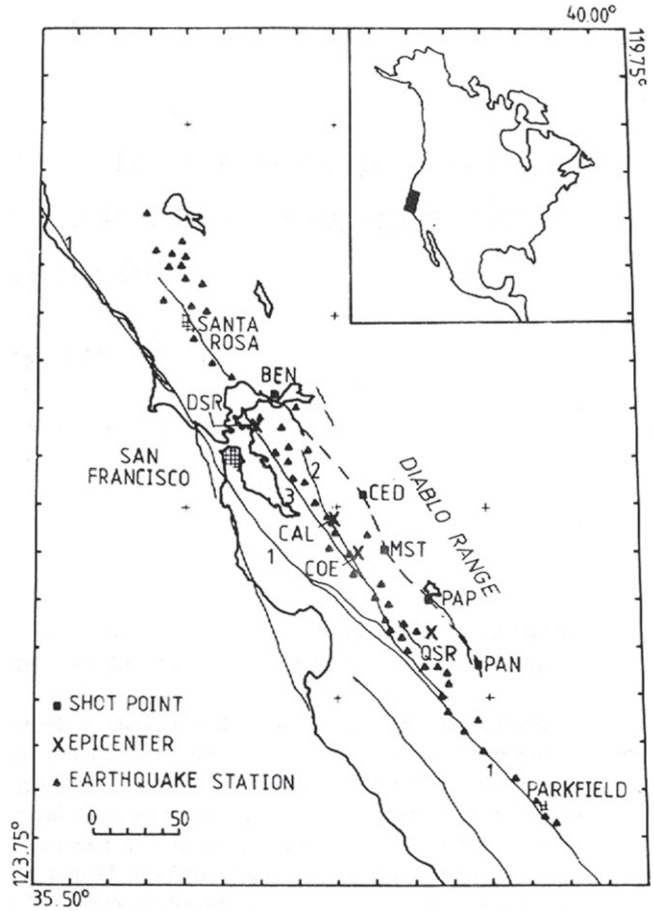


Figure 7.4.1-09. Location map of seismic refraction (dashed line) and near-earthquake (solid line) observations in the Diablo Range of central California (from Blümling and Prodehl, 1983, fig. 1). Squares: shotpoints of the 1967 refraction survey; triangles: permanent stations of the U.S. Geological Survey Central California Microearthquake Network. [Physics of the Earth and Planetary Interiors, v. 31, p. 313–326. Copyright Elsevier.]

Also north of the San Francisco Bay, a seismic-refraction survey was undertaken to study the Geysers–Clear Lake geothermal area in northern California (Warren, 1981).

Another project combining earthquake and active shot data dealt with the investigation of the crust in the western foothills of the Sierra Nevada (Spieth et al., 1981), where an earthquake near Oroville Dam, Oroville, California, prompted a very intense study. Record sections were compiled from both aftershock recordings up to 300 km and an active seismic-refraction experiment with distances up to 80 km (Appendix 7-5-4). The interpretation indicated a localized, shallow, high-velocity layer with 7 km/s at only 5 km depth, which had restricted extension and was underlain by normal 30-km-thick upper crust. The total crustal thickness was estimated at 39 km, based on a combined evaluation of Bouguer gravity and earthquake data recordings.

The surveillance of the Long Valley area, on the eastern side of the Sierra Nevada, employed seismic-refraction profiling in 1973. The goal was to define the upper 10 km of the crust in order to understand the nature and development of a resurgent caldron, and to investigate if the recorded wave forms might provide evidence for the presence of a geothermal reservoir within the caldera or a possible magma chamber at depth (Hill, 1976).

7.4.2. New Seismic-Recording Equipment Opens a New Stage of Crustal Research

A major step to reactivate active seismology was undertaken by J.H. Healy, whose approach led to major renewed research activity in controlled-source seismology in the 1980s, when W.D. Mooney became responsible for projects on crustal and upper-mantle studies. As the seismic equipment of the U.S. Geological Survey of the early 1960s was practically worn out, J.H. Healy started to develop a new seismic-refraction system, inspired by previous visits to Europe where he had watched the highly successful MARS-66 equipment.

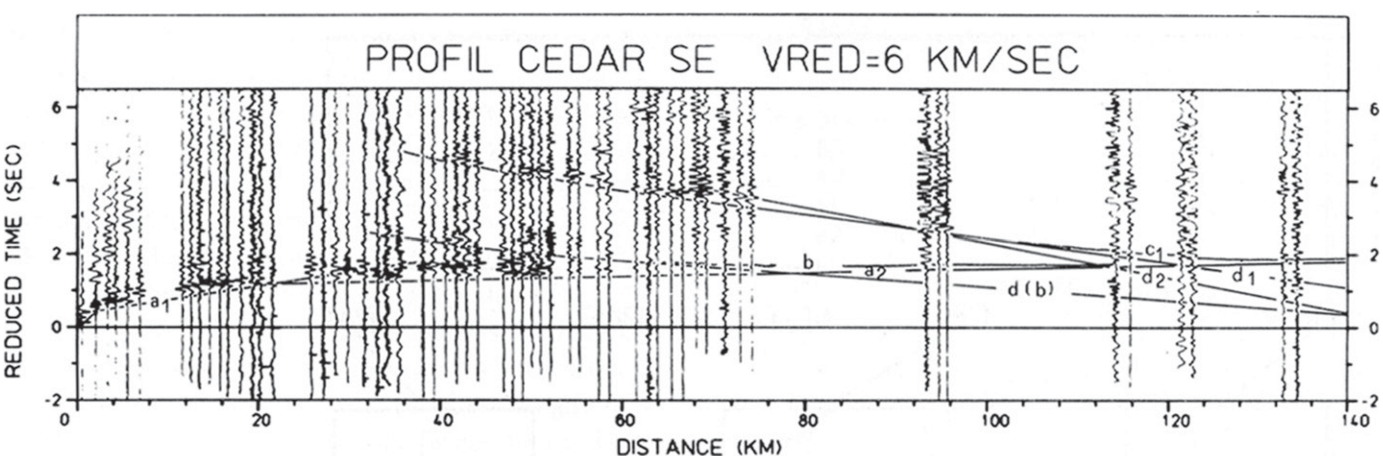


Figure 7.4.1-10. Example of seismic refraction observations in the Diablo Range of central California (from Blümling and Prodehl, 1983, fig. 5). [Physics of the Earth and Planetary Interiors, v. 31, p. 313–326. Copyright Elsevier.]

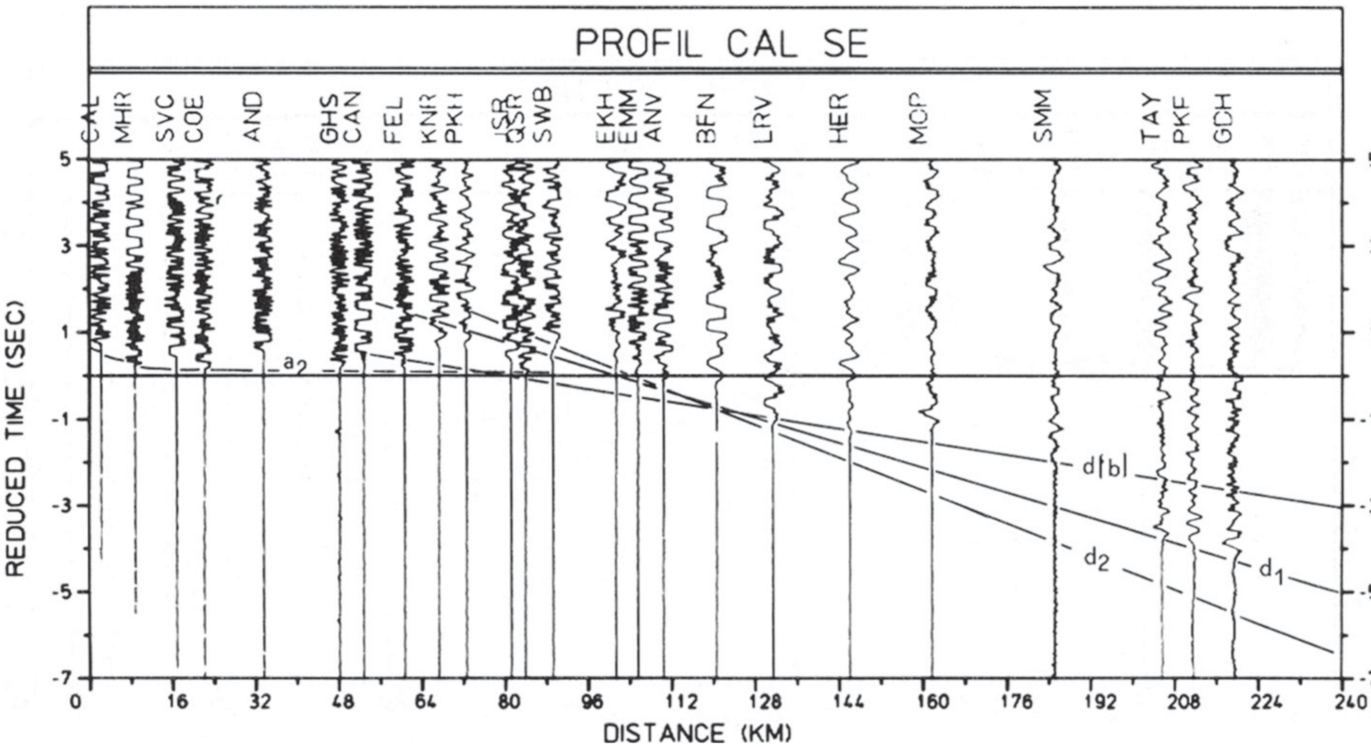


Figure 7.4.1-11. Examples of near-earthquake observations in the Diablo Range of central California (from Blümling and Prodehl, 1983, fig. 11). [Physics of the Earth and Planetary Interiors, v. 31, p. 313–326. Copyright Elsevier.]

This new system would fulfill several conditions which were not easily met. The instrument had to be small, with the ability to be built in large quantities, with at least 100 pieces. It would also have to be easily operated by untrained personnel, be able to record mostly automatically, i.e., to be switched on and off by a sophisticated clock of high accuracy, and be able to run unattended for several days.

In addition, the playback of the recorded data should be fast and easily gained for field control; i.e., within a few hours, the data of field tapes from 100 stations should be available on a field computer immediately after pick up, raw record sections be plotted on the spot, and the instruments would have to be ready for the next deployment within a very limited time period.

The rapid development of such instrumentation became possible by a grant and order from the Saudi Arabian government to perform a long-range seismic project in Saudi Arabia, which will be described in section 7.5, "The Afro-Arabian Rift System." By the beginning of 1978, 100 of these so-called cassette recorders were ready to be shipped to Saudi Arabia (Fig. 7.4.2-01). A detailed description of the recording system was prepared by Murphy (1988; Appendix A7-5-6).

Later in the same year, when R.B. Smith and colleagues accomplished their fieldwork in the Yellowstone–Snake River Plain area, the U.S. Geological Survey moved in and recorded an additional profile with these cassette recorders. It was this instrumentation which led to a renewed strong activity of the U.S. Geological Survey in the following decade.

Two more major projects using this new equipment followed at the end of the 1970s to investigate the crustal structure with much more detail than had been possible before. One project was the investigation of the northern Mississippi embayment (Hamilton, 1986; Mooney et al., 1983). In 1978 and 1979, thirteen deep seismic-reflection profiles with 3 s two-way traveltime had been recorded in the New Madrid, Missouri, seismic hazard zone. Subsequently, oil exploration in northern Arkansas provided 6400 km of reflection data, of which 350 km were acquired by the U.S. Geological Survey (Hamilton, 1986).

In 1979 and 1980, an extensive seismic-refraction program followed, providing velocity and structural information down to the crust-mantle boundary (Ginzburg et al., 1983; Mooney et al., 1983). The large number and the small size of the new cassette recorders now available and the instruments' refined clock system allowed a high station density as well as the layout of a series of profiles with a total of nine shotpoints across the embayment, within a reasonably short time frame (Fig. 7.4.2-02), and with a reasonably small number of people. Figure 7.4.2-03 shows the high data density and high data quality obtained by only one shot with the new equipment (Appendix A7-5-5). The important result of this study was the confirmation and delineation of a 7.3 km/s layer in the lower crust (see top section of

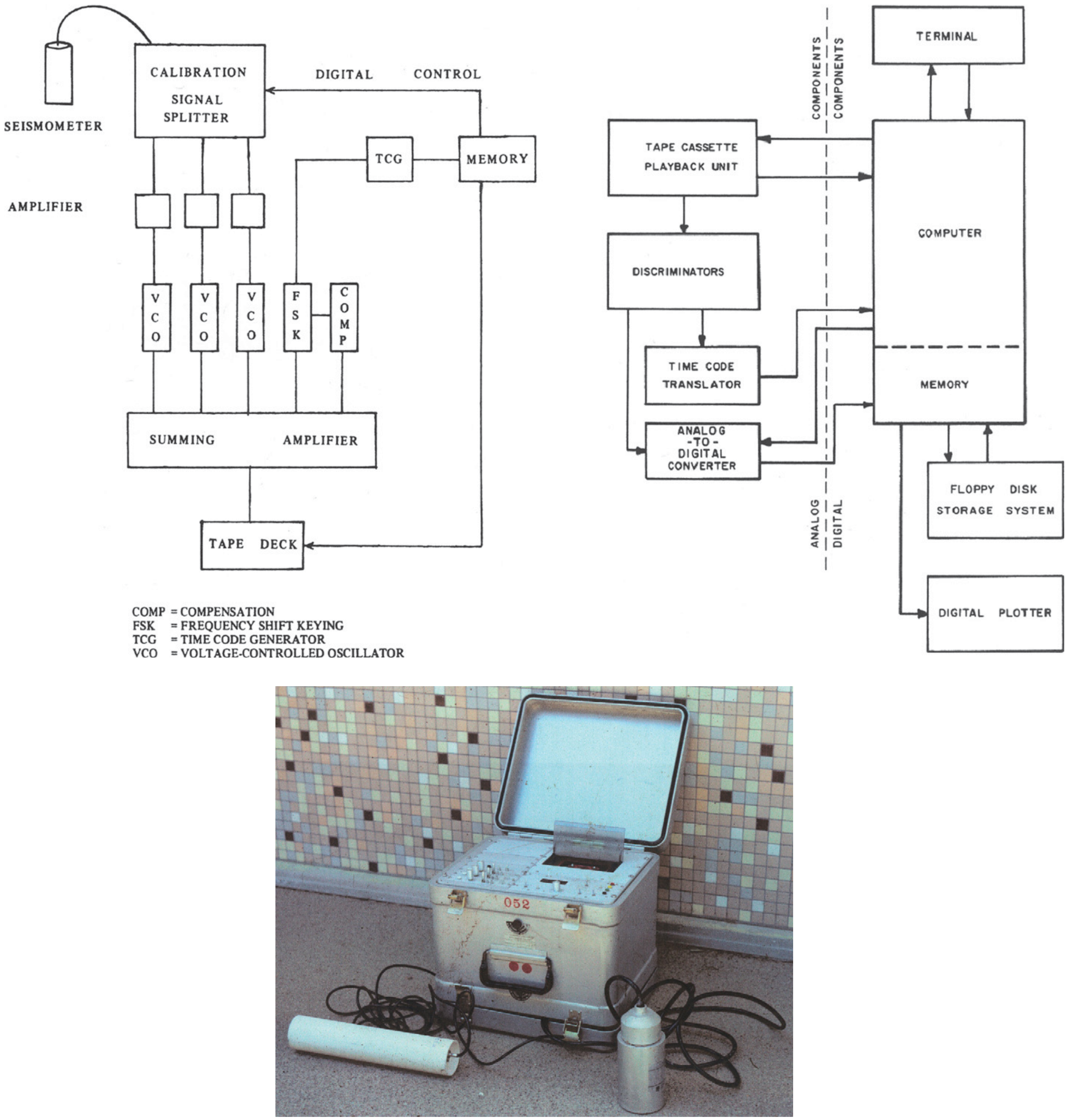


Figure 7.4.2-01. Cassette recorders developed by J.H. Healy and co-workers during 1976–1978. Schematic diagrams (from Healy et al., 1982, figs. 13 and 16). Left: Recording unit. Right: Seismic data processing system. [Final Project Report, Saudi Arabian Deputy Minister of Mineral Resources, Open-File Report, USGSOF-02-37, 429 p., including appendices; also U.S. Geological Survey, Open-File Report, 83-390.] Bottom: Cassette recorder with seismometer (courtesy of G. Fuis).

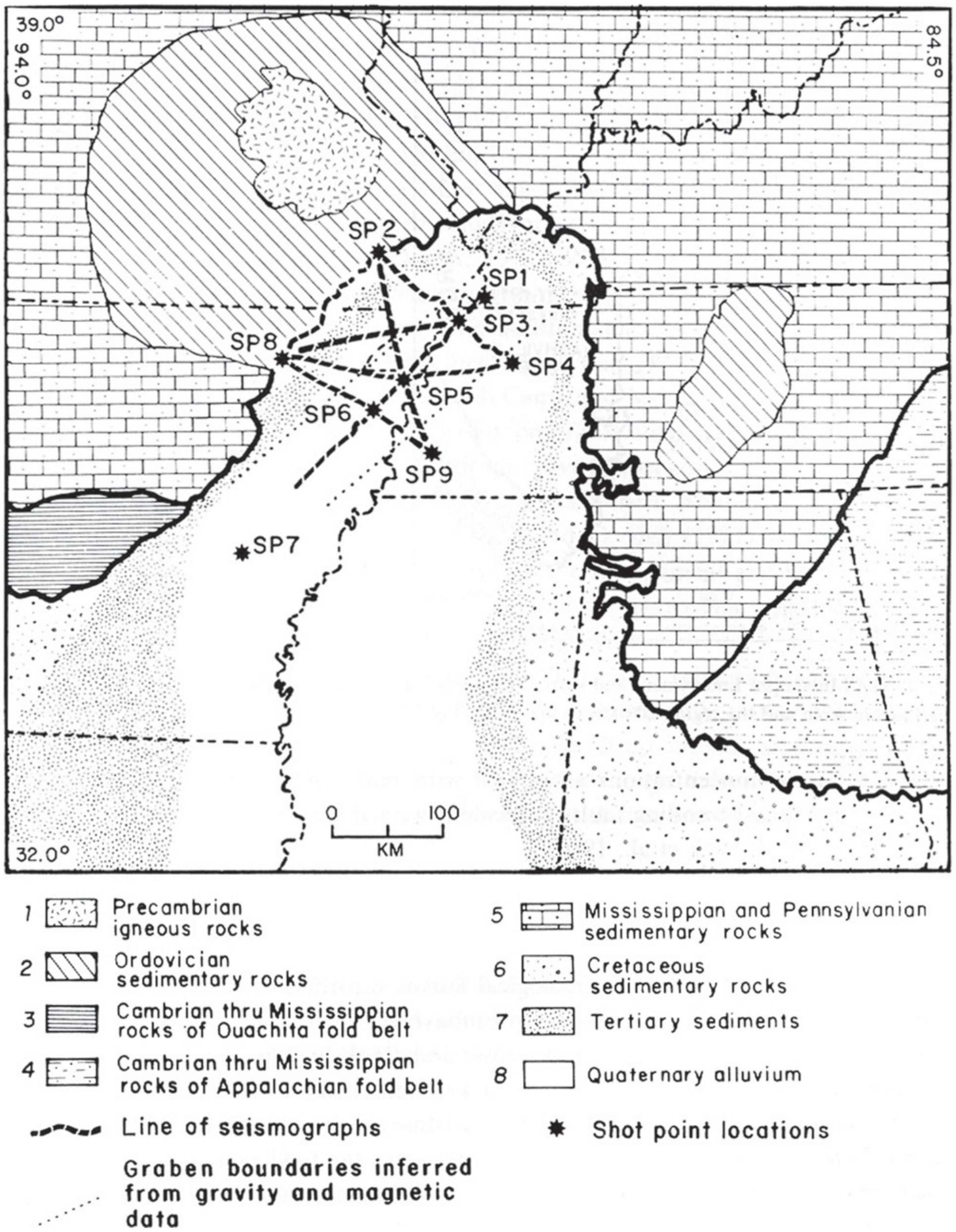


Figure 7.4.2-02. Location map of the seismic-refraction survey in the Mississippi embayment (from Mooney et al., 1983, fig. 1). [Tectonophysics, v. 94, p. 327–348. Copyright Elsevier.]

Fig. 7.4.2-05) which had already been suggested from previous studies, implying that the lower crust had been altered by injection of mantle material.

Another project aimed to make a detailed investigation of the Imperial Valley and its surroundings in southern California (Fuis et al., 1984; Kohler and Fuis, 1988; Appendix A7-5-6).

Forty shots at seven shotpoint sites, ranging in size from 400 to 900 kg, were detonated and recorded by 100 portable vertical seismic instruments, with a spacing of 0.5–1 km (Fig. 7.4.2-04). Usually in one night, three shots were recorded on a line. Data were primarily collected in January to March 1979, but, following the October 1979 Imperial Valley earthquake, additional data were collected (Fuis et al., 1984; Kohler and Fuis, 1988; Appendix A7-5-6).

Following the interpretation of the data in the Mississippi embayment in their publication, Mooney et al. (1983) added a discussion of young continental rift structures in central Europe, East Africa, and North America. All models show an anomalous transition zone above the Moho with velocities between 7 and 8 km/s. Their compilation of models, published by various authors, also contains velocity-depth cross sections through the Mississippi embayment (top section of Fig. 7.4.2-05) and through the Imperial Valley in the Salton Trough (bottom section of Fig. 7.4.2-05).

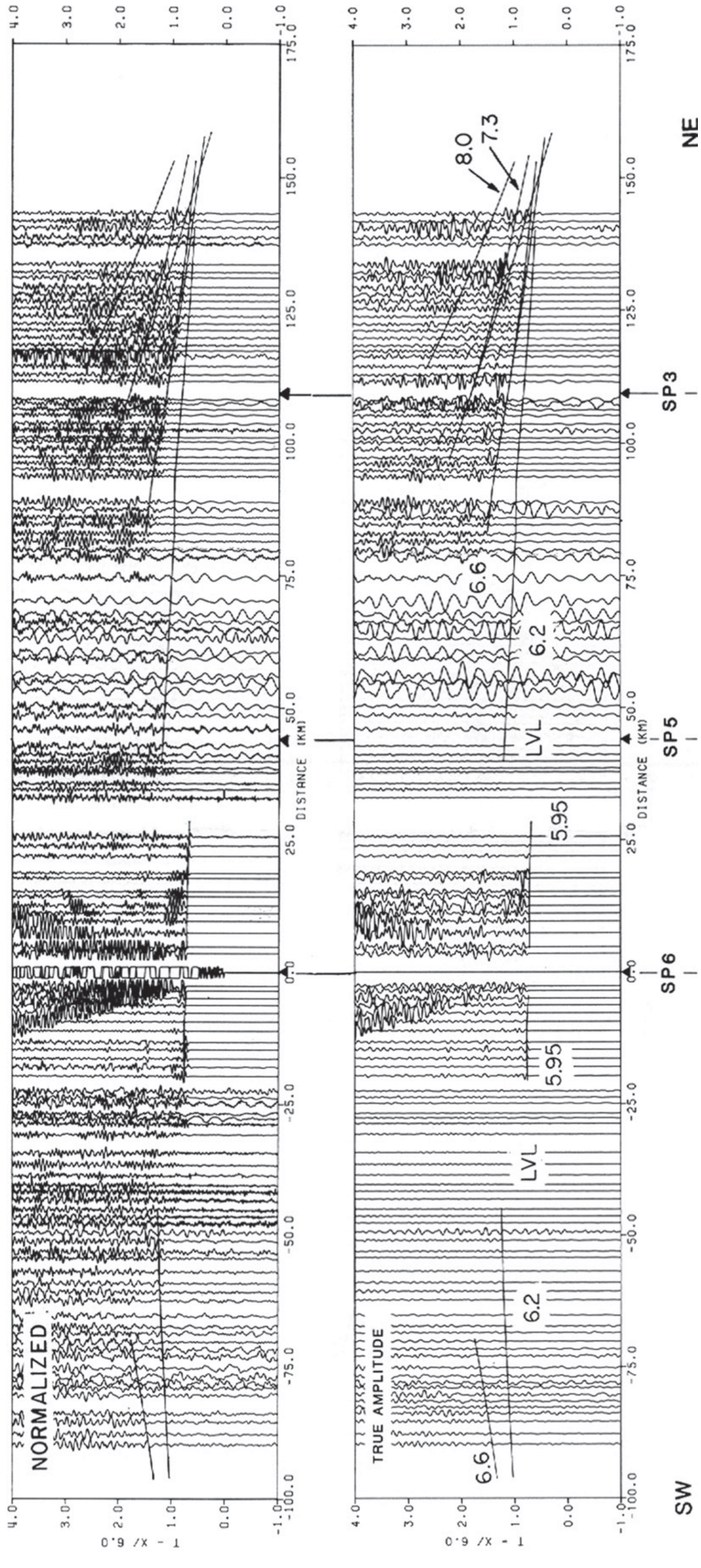


Figure 7.4.2-03. Data recorded from shotpoint 6 in SW and NE direction of the seismic-refraction survey in the Mississippi embayment (from Mooney et al., 1983, fig. 4). Top: record section plotted with normalized amplitudes. Bottom: with true amplitudes. [Tectonophysics, v. 94, p. 327–348. Copyright Elsevier.]

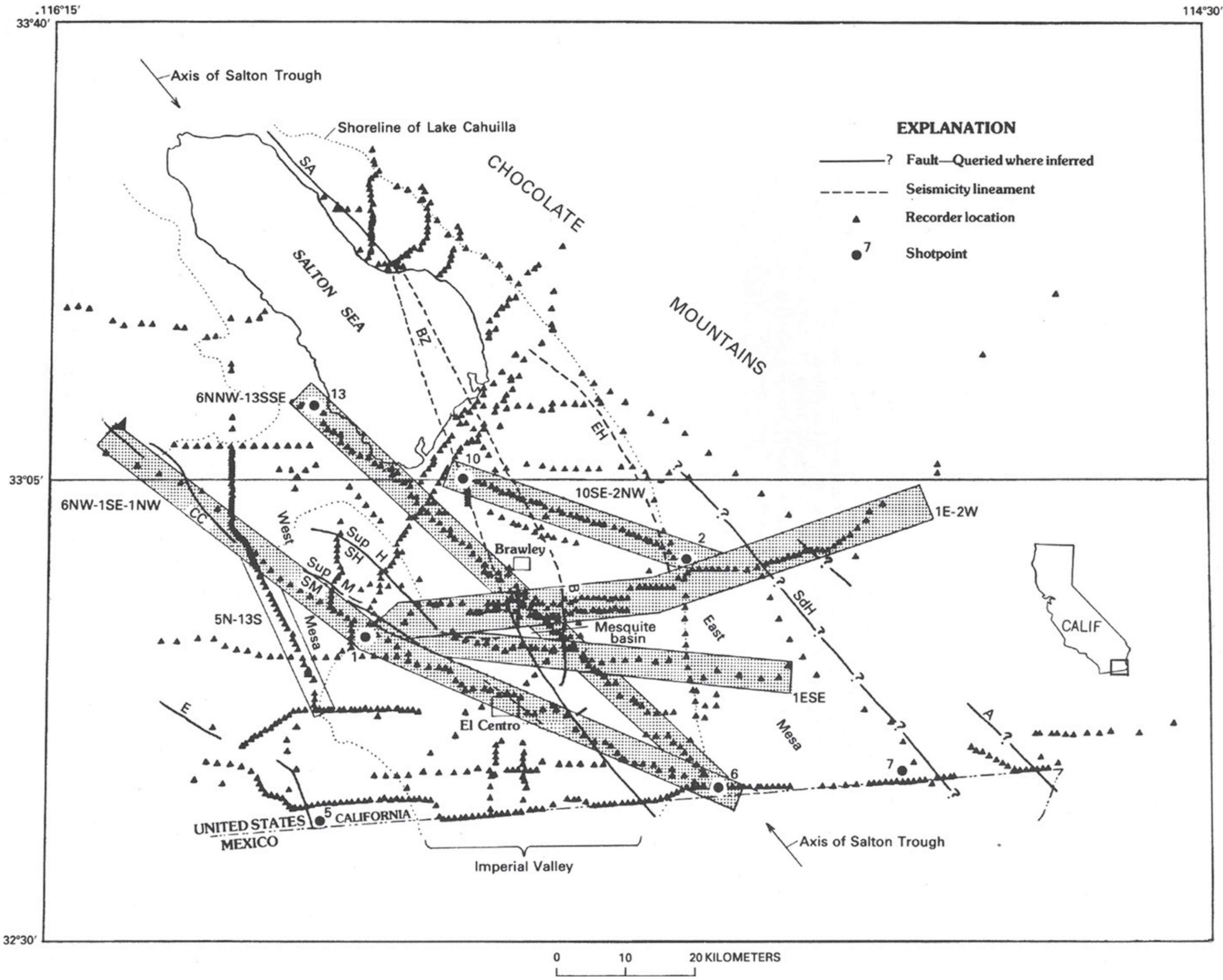


Figure 7.4.2-04. Location map of the seismic-refraction survey in the Imperial Valley, southern California (from Fuis et al., 1984, fig. 1). [Journal of Geophysical Research, v. 89, p. 1165–1189. Reproduced by permission of American Geophysical Union.]

7.4.3. The First Large-Scale Continental Reflection Profiling Project (COCORP and U.S. Geological Survey)

A completely different approach to study the Earth's crust in much more detail was started when at Cornell University (Ithaca, New York), a large-scale seismic-reflection program, the Consortium for Continental Reflection Profiling, or COCORP, was initiated. J.E. Oliver, who had already been one of the first promoters of plate tectonics in its early stages and was actually one of the first who discussed this theory in public, had thought about the complexity of the Earth's crust for a long time (Oliver, 1996). He realized how little science had hitherto contributed to the study of the continental crust in much more detail than large-scale seismic-refraction studies had been able to.

Pointing to successful deep seismic-reflection studies in the 1960s in Canada, the United States, Germany, the USSR, and Australia, he circulated the basic idea amongst the leading seismologists in the United States, which finally led to the establishment of COCORP. Together with S. Kaufman, whom he persuaded to leave the exploration industry, he developed the idea to study the Earth's crust by carrying out large-scale reflection seismology in applying commercially common methods of fieldwork and interpretation. In 1975, COCORP was created (Brewer and Oliver, 1980; Brown et al., 1986).

A commercial reflection company was hired, and the first fieldwork (Fig. 7.4.3-01) was started in 1975 in Hardeman County, Texas (Oliver et al., 1976). The Vibroseis system was used, because it offered high-quality data while maximizing

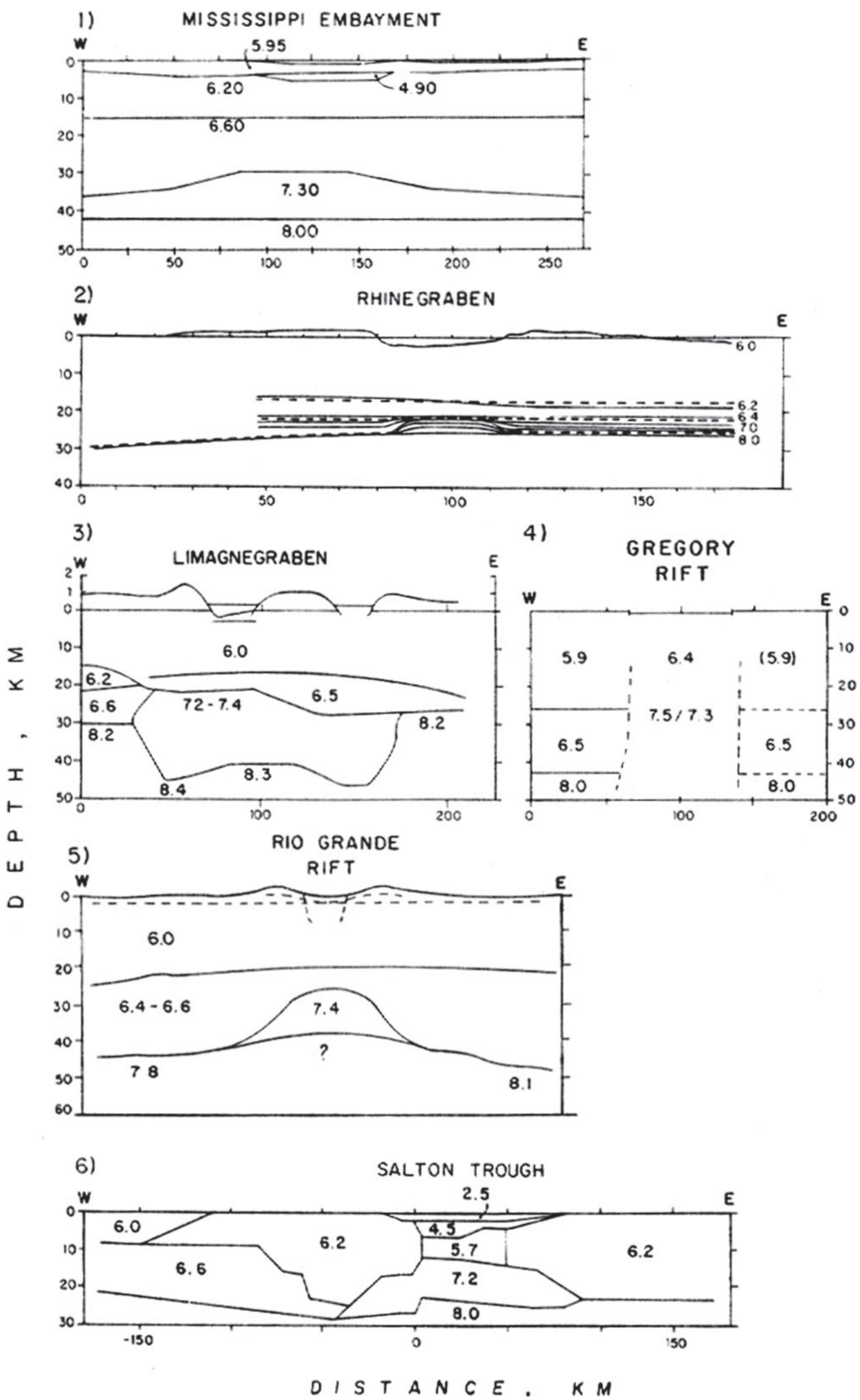


Figure 7.4.2-05. Continental rift structures (from Mooney et al., 1983, fig. 11). (1) Mississippi embayment (Mooney et al., 1983), (2) Rhinegraben, Germany (Edel et al., 1975), (3) Limagnegraben (Hirn and Perrier, 1974), (4) Gregory Rift, East Africa (Long et al., 1973), (5) Rio Grande Rift (Keller et al., 1979), (6) Imperial Valley (Fuis et al., 1984). [Tectonophysics, v. 94, p. 327–348. Copyright Elsevier.]

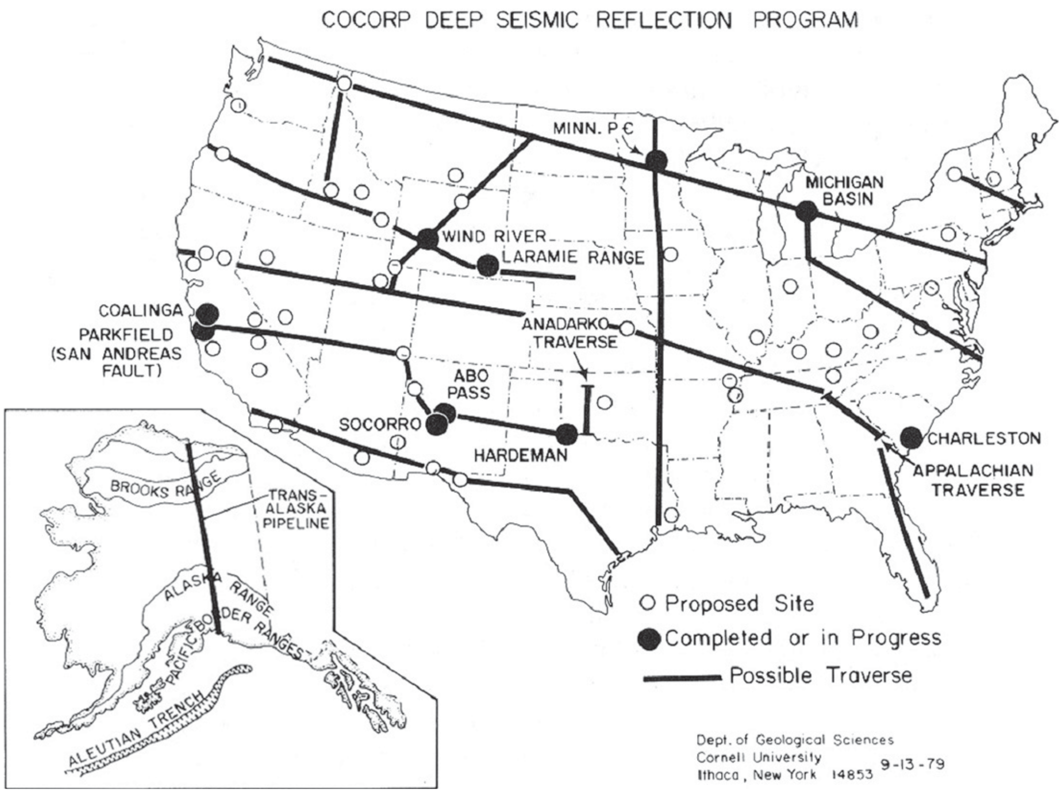


Figure 7.4.3-01. Location map of COCORP deep seismic reflection studies accomplished until 1979 (from Brewer and Oliver, 1980, fig. 2). [Annual Reviews of Earth and Planetary Science, v. 8, p. 205–230. Copyright Annual Reviews.]

efficiency and minimizing environmental impact. COCORP used five vibrating trucks which transmitted signals of 8–40 Hz bandwidth into the ground. The returning echoes were collected by 6–10-km-long linear arrays of geophones connected to a 96-channel truck-mounted recording system. About 24 geophones per channel were used. The data were collected and computer processed to produce a 24-fold common-depth point stacked seismic section, representing a cross section through the Earth's crust in terms of traveltime of the seismic waves. These sections served to provide important preliminary geological interpretations, before depth conversion and migration were applied for detailed studies (Brewer and Oliver, 1980). Phinney and Roy-Chowdhury (1989) have discussed in much detail the basics of seismic-reflection techniques and the long procedure leading from the raw data to the seismic-reflection sections used for final geological interpretation. It will be described in some detail at the beginning of Chapter 8.3.

Soon, other surveys followed, using the same techniques as applied for the Hardeman county project (Fig. 7.4.3-01). The Rio Grande rift, New Mexico, was the site of the first full-scale investigation of a specific geological problem (Brown et al., 1979; Oliver and Kaufman, 1976). The objective was to study the structure of an active rift, including the geometry of faulting, the nature of the crust-mantle boundary, and the possible existence of magma bodies at depth. Here, in the Socorro area of the Rio Grande rift, where a magma body had been inferred from local earthquake data, in 1975–1976 COCORP collected 155 km of 24-fold reflection data transverse and parallel to the rift. The most prominent coherent reflections were identified to correspond to a depth of ~20 km, corresponding to the top of the postulated magma body (Fig. 7.4.3-02).

In 1976, a 27-km-long seismic line was recorded across the San Andreas fault near Parkfield, California, where the San Andreas fault bifurcates to form a thin sliver. The line was short and crooked and the data quality was less good as elsewhere, but showed several interesting features indicating a zone of brittle fracturing limited to the upper 15 km of the crust, where generally the earthquakes occur, and a vertical blank zone of 4 km width beneath, possibly created by ductile flow (Long et al., 1978; Brewer and Oliver, 1980).

The seismic-reflection data, collected in 1977 in the Wind River Mountains and Laramide Range, Wyoming, on 158 km of seismic profiles, showed a pronounced set of dipping events originating at the surface at the position of the Wind River thrust and trending into the crust at a moderate angle (Brewer et al., 1980, 1982; Smithson et al., 1979), which could be traced to 24 km depth for certain and possibly further to 36 km depth.

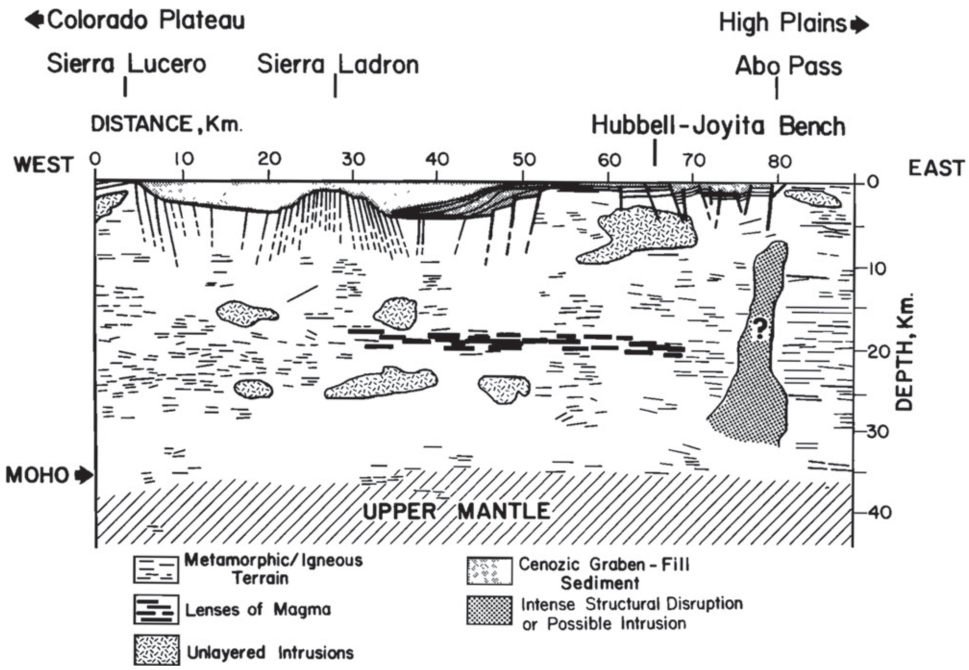


Figure 7.4.3-02. Schematic cross section of COCORP across the Rio Grande rift near Socorro, New Mexico (from Brown et al., 1980, fig. 14). [Journal of Geophysical Research, v. 85, p. 4773–4800. Reproduced by permission of American Geophysical Union.]

Other surveys were performed in the Midwest and eastern United States. In 1978, the Michigan Basin in Michigan was investigated (Jensen et al., 1979); reflection surveys in the Wichita uplift, Oklahoma (Brewer and Oliver, 1980) and in the Archean Province in Minnesota (Gibbs et al., 1984) followed in 1979. The seismic-reflection survey of 1980 in the Adirondack Mountains, New York, aimed to investigate an extensive dome of high-grade metamorphic rocks where seismic-reflection patterns against rocks once buried at mid-crustal depths but now exposed were to be calibrated (Brown et al., 1983a, 1986).

The southern Appalachian COCORP traverse of 1978 and 1979 was initiated to investigate the Brevard zone, an enigmatic fault zone in the Inner Piedmont. The interpretation of the seismic-reflection data implied a thin-skinned tectonics: that the crystalline southern Appalachians were allochthonous, vary in thickness between 6 and 15 km, and were thrust over a passive continental margin with little deformation of the underlying continental shelf and slope sediments (Cook et al., 1979a; Brewer and Oliver, 1980).

Besides intensifying its crustal seismic-refraction research by the new seismic cassette recorders, the U.S. Geological Survey also started a major seismic-reflection survey program which was to be continued in the 1980s. From 1973 to 1979, a major program of multichannel seismic exploration of the U.S. Atlantic continental shelf was carried out on the Long Island Platform (Phinney, 1986). It will be described in more detail in Chapter 7.8.3. In 1979, the first project on land was a survey conducted across the Grandfather Mountain, the line extending from Tennessee to North Carolina through the metamorphic sequences of the Appalachian orogen (Harris et al., 1981; for location, see line GMP in Fig. 8.5.3–24).

7.4.4. Canada

In 1970, geophysical studies were performed in the Baffin Bay, northern Canada (Keen et al., 1972).

In Canada, already in 1972, near-vertical reflection techniques were applied to record deep reflections along three profiles in the Canadian Cordillera (Mair and Lyons, 1976; Berry and Mair, 1977; Fig. 7.4.4-01). Up to six shotpoints were spaced ~1.5 km apart in each of three lakes, from where shots with charge sizes of 50 kg were recorded by closely spaced geophone arrays extended in a line along a road. The recording sites were spaced ~250 m apart, reaching maximum offsets of 20 km. This geometry allowed up to fivefold common reflection point (CDP) stacking. Only one relatively continuous event was recorded at ~11 s TWT, which, assuming an average crustal velocity of 6.3 km/s, was interpreted originating at a 34-km-deep reflector.

In 1973, a Vibroseis survey was added to obtain a roughly similar subsurface coverage. The experiment proved to be successful and reflections could be correlated throughout the crust and from the crust-mantle boundary.

In 1977, 1979, and 1981, experiments were conducted over the northern part of the Williston Basin as part of the COCRUST investigations (Fig. 7.4.4-02) which will be discussed in Chapter 8. The north-south and east-west inline profiles of 1977 and 1979 were supplemented in 1981 by three profiles arranged as a triangle which recorded all shots arranged at its edges. The data were interpreted several times (e.g., Hajnal et al., 1984; Kanasewich et al., 1987), indicating in particular a crustal thickening from 40 km in the Superior Province to 45–50 km under the Williston Basin.

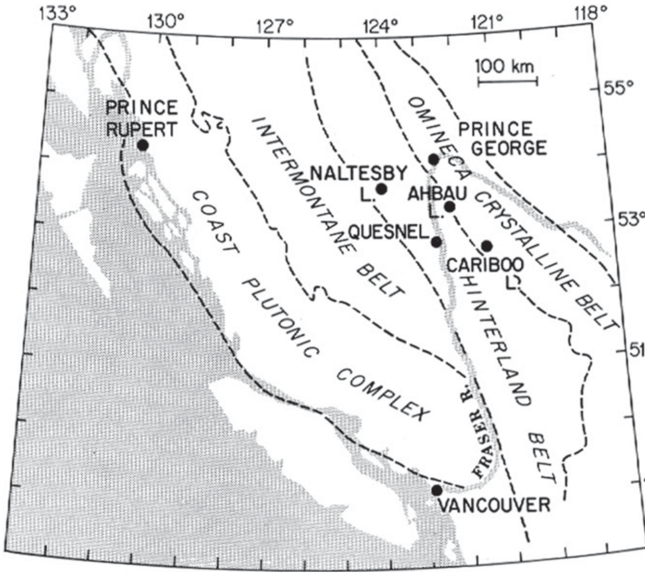


Figure 7.4.4-01. Location of Naltesby, Ahbau, and Cariboo Lakes in the Canadian Cordillera, shot sites for reflection profiles (from Berry and Mair, 1977, fig. 9) [American Geophysical Union, Geophysical Monograph 20, p. 319–348. Reproduced by permission of American Geophysical Union.]

7.5. THE AFRO-ARABIAN RIFT SYSTEM

7.5.1. The Northern End of the East-African Rift System

Following the joint effort of investigating the eastern Mediterranean area under Greece, a major expedition was planned and organized to investigate the northernmost part of the East African rift system in Ethiopia under the leadership of H. Berckhemer and J. Makris. The investigation aimed for the deep structure of the Afar region and part of the adjacent Ethiopian Highland by joint gravity and seismic-refraction measurements (Berckhemer et al., 1975).

In parallel, French scientists carried out a major crustal study of the French territory of Djibouti, adjacent to the Red Sea and Gulf of Aden (Ruegg, 1975). The investigations were embedded in an international research program on the Afar depression of Ethiopia within the Upper Mantle project mainly involving earth scientists from Britain, France, Germany, and Italy (Pilger and Rösler, 1975, 1976).

While quite conventional in principle, the actual field observations were rather unconventional due to the extreme and adverse environmental conditions. Five profiles, between 120 and 300 km long, were laid out in different parts of the Afar depression and one

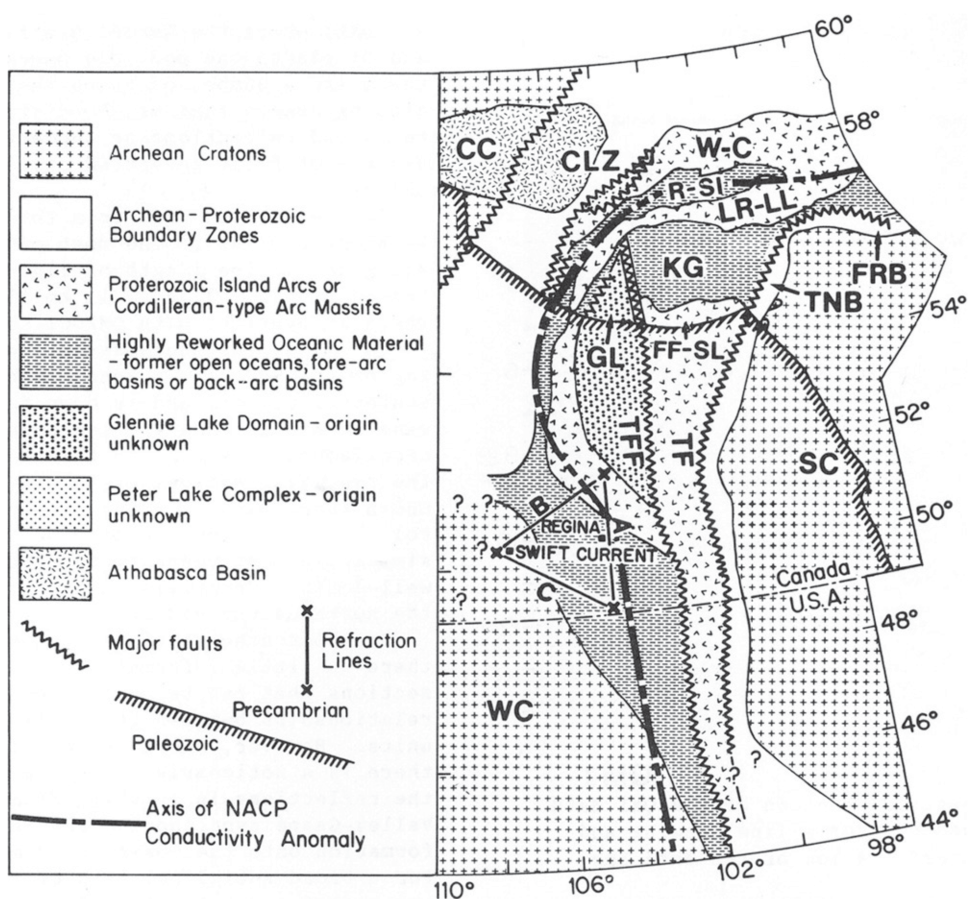


Figure 7.4.4-02. Location of a triangular array of receivers used for the first COCRUST refraction/wide-angle reflection survey in the Williston Basin (from Green et al., 1986, fig. 4). Letters refer to geological details which are of no relevance here. [*In* Barazangi, M., and Brown, L., eds., Reflection seismology: a global perspective. [American Geophysical Union, Geodynamics Series, v. 13, p. 85–97. Reproduced by permission of American Geophysical Union.]

profile was arranged on the Ethiopian Plateau (Fig. 7.5.1-01). All profiles were chosen along roads or tracks that could be reached under favorable weather conditions. All data of the Afar 1972 expedition were reproduced in Appendix A2-1 (p. 89–90).

Seismic shots were generated in lakes or rivers preferably with charges between 300 and 1600 kg. A particular shooting technique to yield high seismic efficiency even from explosions in very shallow water was applied (Burkhardt and Vees, 1975). Also, by chance, a local earthquake occurred during the time of recording along profile V, which produced excellent S-wave arrivals at distances from 70 to 120 km.

The record section of profile I (Fig. 7.5.1-02, bottom) recorded on the Ethiopian Plateau was similar to data recorded on 40-km-thick continental crust elsewhere, while the first arrivals

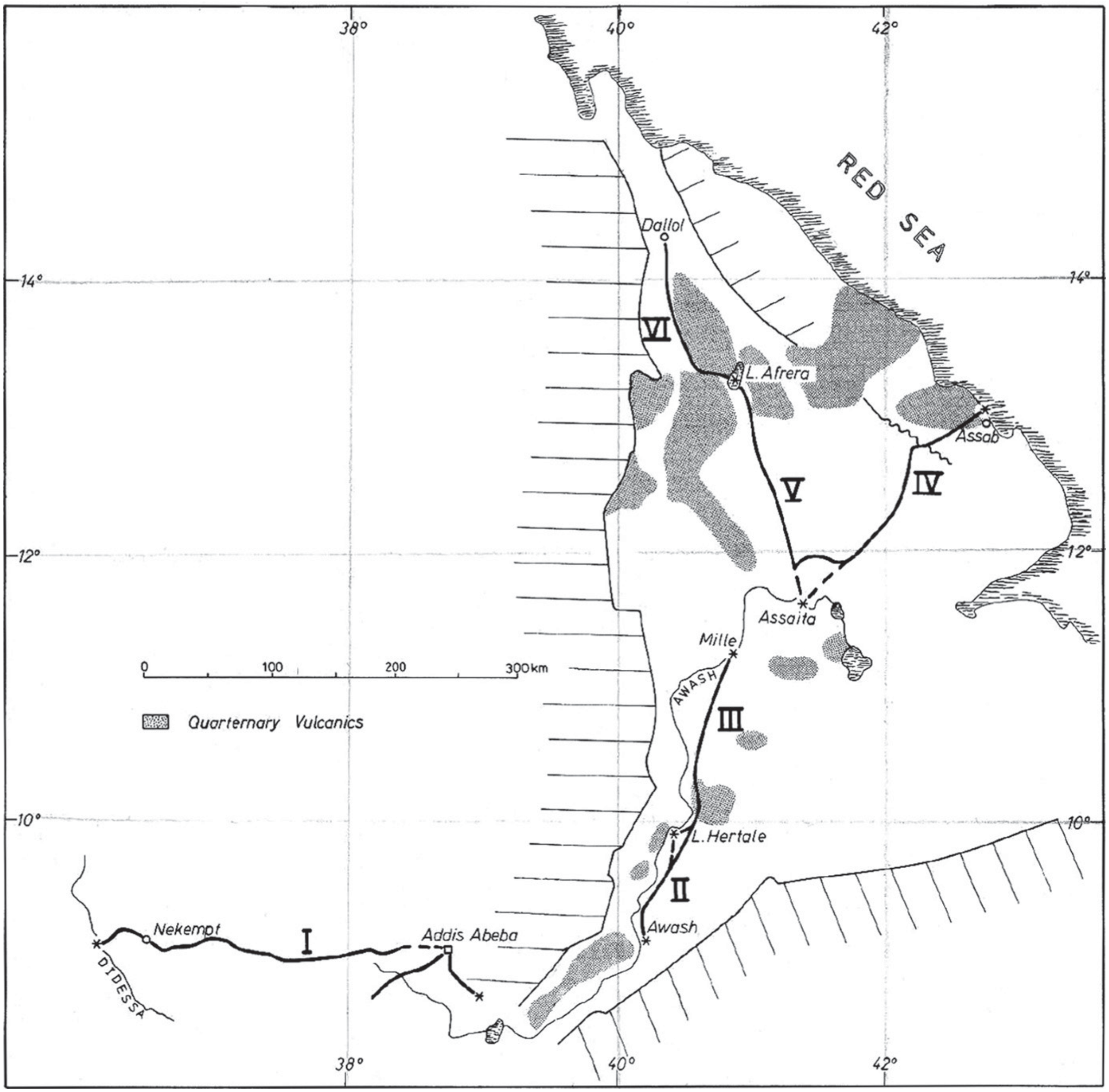


Figure 7.5.1-01. Location of deep seismic sounding profiles in Ethiopia (from Berckhemer et al., 1975, fig. 1). [*In* Pilger, A., and Rösler, A., eds., Afar Depression of Ethiopia: Stuttgart, Schweizerbart, p. 89–107. Reproduced by permission of Schweizerbart, Stuttgart, Germany (www.schweizertbart.de).]

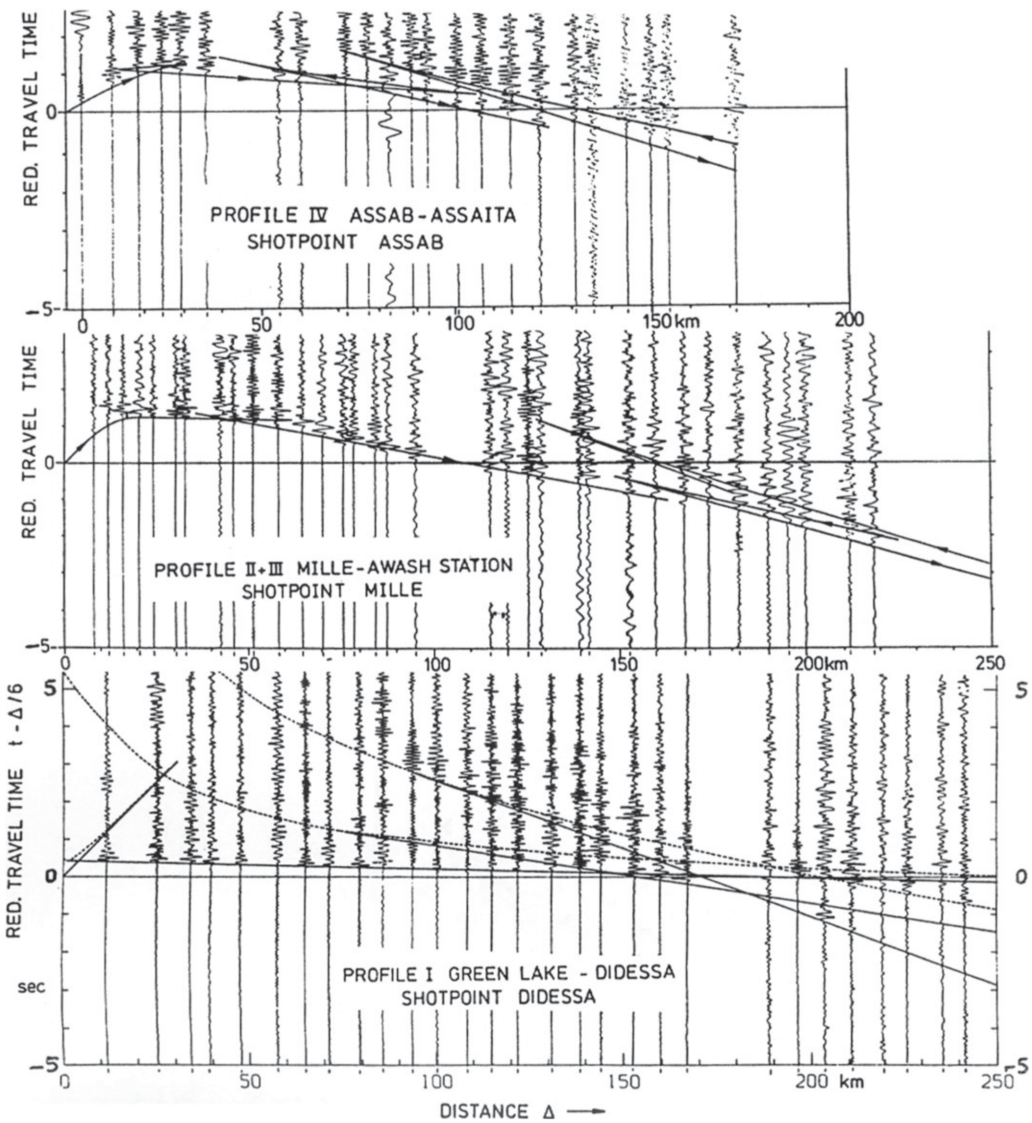


Figure 7.5.1-02. Record sections of deep seismic sounding profiles in Ethiopia (from Berckhemer et al., 1975, figs. 2b, 2a, 2e). For location, see Fig. 7.5.1-01. [*In* Pilger, A., and Rösler, A., eds., Afar Depression of Ethiopia: Stuttgart, Schweizerbart, p. 89–107. Reproduced by permission of Schweizerbart, Stuttgart, Germany (www.schweizertbart.de).]

of the other profiles (Fig. 7.5.1-02, top and center) reflected basic material with velocities of 6.6–6.8 km/s near the top of the crust.

The seismic investigation of Djibouti was carried out in 1971 by the Operation Grands Profils Sismiques de l'I.N.A.G. (Institut de Physique du Globe of Paris). Profiles were recorded along the coast of the Gulf of Tadjura making use of sea shots, while the inland profiles recorded sea shots, shots in shallow lakes, and borehole shots (Fig. 7.5.1-03).

The main result was that the bulk of the Afar crust, whose thickness changed from almost 26 km in the south (Fig. 7.5.1-02, center) to almost 16 km in the north (Fig. 7.5.1-02, top), appeared to have average P-wave velocities of 6.6–6.8 km/s and was underlain by a gradual transition into upper-mantle material with anomalously low velocities between 7.3 and 7.7 km/s under all of Afar (Fig. 7.5.1-04).

The authors saw evidence that this anomalous mantle was limited in depth to an extent of nearly 40 km (Berckhemer et al., 1975). Under the adjacent Djibouti region to the east, Ruegg (1975) postulated an oceanic crust of thickness ~6–10 km, which was underlain by an anomalous upper mantle with P-wave velocities as low as 6.7–6.9 km/s increasing gradually to 7.4 km/s at a depth of 20 km.

7.5.2. First Crustal Studies of the Dead Sea Rift, around the Red Sea, and in Iran

In 1977, the first deep seismic sounding experiment was carried out in the Jordan–Dead Sea rift. Initiated by A. Ginzburg and promoted by geologists such as Z. Garfunkel, B. Freund, and others, the Geophysical Departments of the universities of Tel Aviv, Israel, as well as of Hamburg and Karlsruhe, Germany, carried out a seismic-refraction survey on the western side of the rift (Fig. 7.5.2-01) with side profiles toward the Mediterranean Sea (Ginzburg et al., 1979a, 1979b).

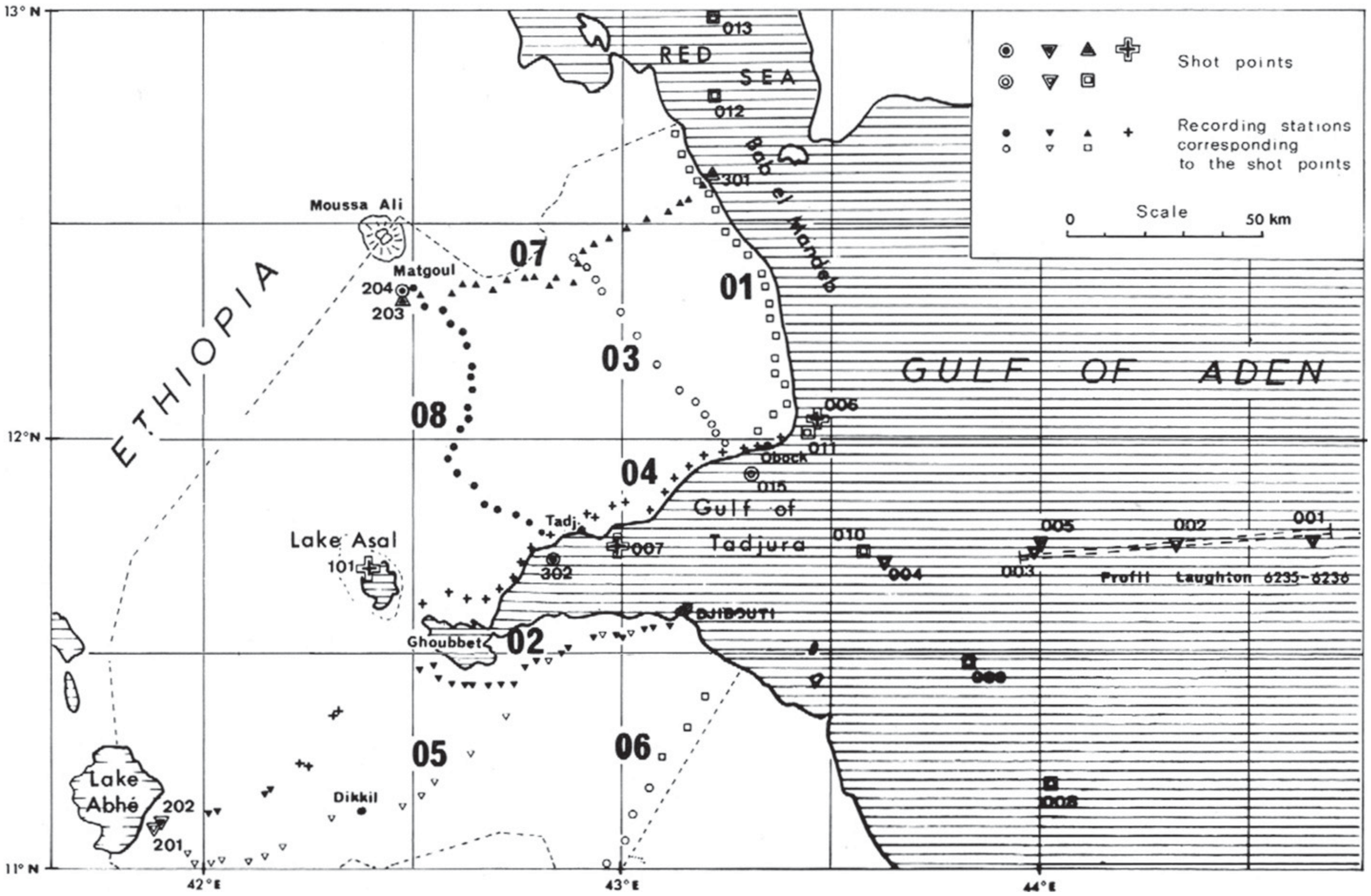


Figure 7.5.1-03. Location of deep seismic sounding profiles of 1971 in Djibouti (from Ruegg, 1975, fig. 1). [*In* Pilger, A., and Rösler, A., eds., Afar Depression of Ethiopia: Stuttgart, Schweizerbart, p. 120–134. Reproduced by permission of Schweizerbart, Stuttgart, Germany (www.schweizertbart.de).]

Shotpoints were in the Dead Sea, the Gulf of Aqaba, and the Mediterranean Sea (Fig. 7.5.2-01). Recording distances along most of the lines were up to 240 km, but along the rift a maximum distance of 400 km could be reached, due to the optimal damming of the shots in the Dead Sea (Fig. 7.5.2-03). The data (see Appendix A2-1, p. 82–83) suggested a transition zone between the lower crust and the upper mantle, which evidently was restricted to the rift area proper. The crust thickened from 20 km near the Mediterranean coast to ~30 km under the rift. The side profiles indicated a further crustal thickening toward 40 km under the Sinai Peninsula.

In 1978, a seismic-refraction line from Cyprus to Israel was added (Fig. 7.5.2-02). Thirty-three sea shots were recorded both by ocean-bottom seismometers along the line and by land stations in Cyprus and Israel (Makris et al., 1983a; Appendix A7-6-1).

Also in 1978, and in 1981, seismic profiles were recorded from the northernmost Red Sea into Egypt (Rihm et al., 1991) and a seismic-refraction profile was recorded from the Red Sea into northern Saudi Arabia (Makris et al., 1983b). Mechie and Prodehl (1988) have compiled these observations together with other surveys accomplished in the early 1980s in Jordan and the Red Sea area and have published the corresponding crustal cross sections and references. Locations and a summary of results will be jointly discussed in Chapter 8, where data of the 1980s in this area is described.

Farther south, another long-range profile was realized in 1978. It traversed the Arabian Shield of Saudi Arabia over a distance of nearly 1000 km from the southern Red Sea to the neighborhood of Riad (Fig. 7.5.2-04). This survey was carried out by the U.S. Geological Survey, using for the first time the newly built mobile array of 100 cassette recorders mentioned above which had been designed by J.H. Healy and co-workers (Healy et al., 1982; Murphy, 1988; see above; Fig. 7.4.2-01; Appendix A7-5-6). In total, 100 portable seismographs were available; they were successively deployed in each of five 200-km recording spreads. A portable computer center, including field playback, digitizer, and plotting system, was moved along the profile to provide rapid feedback of data quality and content, and to allow for preliminary assessment of scientific results as the experiment

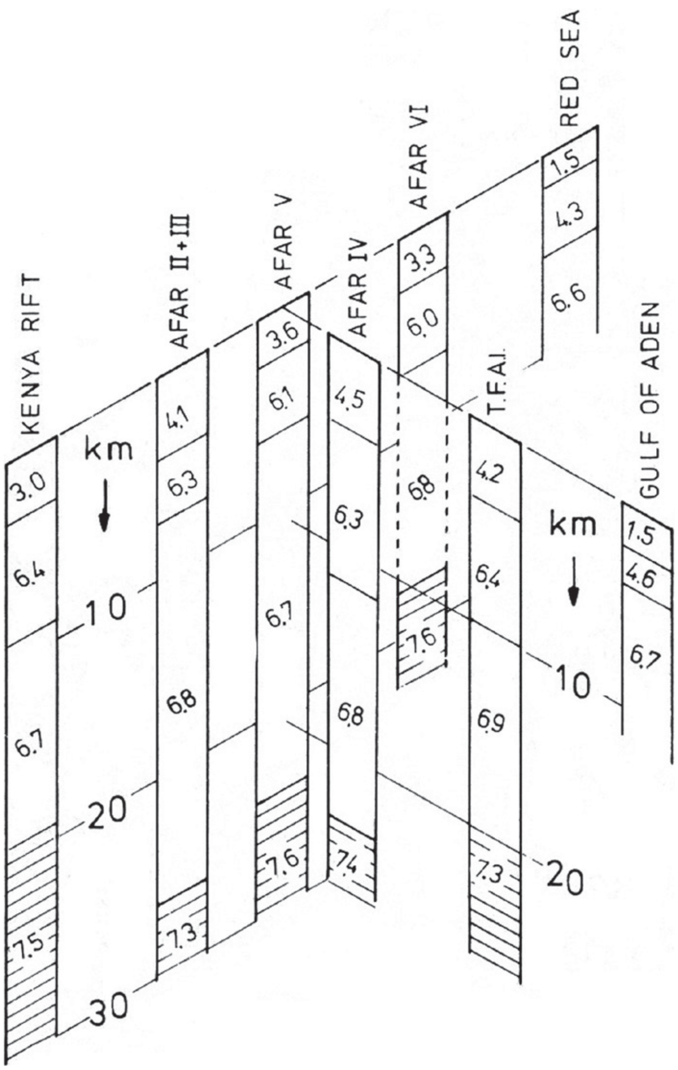


Figure 7.5.1-04. Simplified crustal models for Afar and adjacent rift zones (from Berckhemer et al., 1975, fig. 6). T.F.A.I. = French territory of Djibouti. [*In* Pilger, A., and Rösler, A., eds., Afar Depression of Ethiopia: Stuttgart, Schweizerbart, p. 89–107. Reproduced by permission of Schweizerbart, Stuttgart, Germany (www.schweizertbart.de).]

progressed (Gettings et al., 1986; Mooney et al., 1985; Healy et al., 1982; Appendix A7-6-3).

The data was of extremely good quality (Fig. 7.5.2-05; see Appendix A2-1, p. 86–88; Appendix A7-6-2 and A7-6-3), and was, in the course of a seismic workshop, interpreted by several other authors (e.g., Milkereit and Flueh, 1985; Prodehl, 1985; Mechie et al., 1986). This workshop will be discussed in some detail at the end of this chapter in the section "Advances of interpretation methodology." The overall structure (Fig. 7.5.2-06) shows a rather uniform 40-km-thick crust under the Arabian Shield, which thins more or less abruptly west of the escarpment under the coastal plain to less than 10 km in the Red Sea (Mooney et al., 1985; Gettings et al., 1986).

Also in 1978, a reconnaissance seismic-refraction survey was undertaken in southern Iran to study the deep structure of the Zagros Belt (Fig. 7.5.2-07). The project was a joint venture of the Geophysical Institutes of the Tehran University of Iran and the Free University of Berlin and the University Hamburg of Germany (Giese et al., 1984). Using 12 MARS66 seismic recording stations and a PCM equipment, commercial explosions at the mines Bafq and Sar-Chesmeh were used and recorded along two N-S directed profiles, the first one of 180 km length within the Lut block between the two mines, and the second one of 215 km length, running southward into the Zagros belt. The quality of the data was sufficient, but the density of the data obtained was relatively poor. Therefore only a preliminary interpretation with many open questions could be provided. The political circumstances prohibited any further research. Data and results, as published by Giese et al. (1984) are shown in Appendix A7-6-4.

7.6. DEEP SEISMIC SOUNDING STUDIES IN SOUTHEAST ASIA

7.6.1. India

Explosion seismology started in India at around 1972, when a three-year contract was signed by India and the USSR, and the first deep seismic sounding profiles were recorded. Field technology and interpretation was guided by Russian scientists, such as V.B. Sollogub, who was one of the initiators of the joint research program (e.g., Kaila et al., 1978, 1979, 1981a, 1981b; Mahadevan, 1994). A map of India showing all deep seismic lines shot between 1972 and 1992 is shown in the next Chapter 8 (Fig. 8.7.1-01).

The first 600-km-long profile crossed the Indian Shield in the southern part of India. It was carried out from 1972 to 1975. The profile started near Kavali, at the Bay of Bengal in the east, and ended near Udupi, at the Arabian Sea in the west, and resulted in a 40-km-thick crust (Kaila et al., 1979; Sarkar et al., 2001).

In the Deccan trap area of the Indian Shield south of Bombay, two 200-km-long profiles were recorded near the Koyna dam, and 60 km further north (Kaila et al., 1981b). A third N-S–trending profile was recorded in the Cambay basin, north of Bombay and east of the Gulf of Cambay, where 5-km-thick sediments overlie the Deccan traps (Kaila et al., 1981a). Later an E-W profile was added to the west of the Gulf of Cambay. Detailed crustal cross sections indicate deep faults and fracture zones throughout the crust down to a depth of ~50–60 km. Moho depths vary along these lines between 35 and 40 km, but the crust thins to 30 km and less toward coastal regions. The data were later reinterpreted by Rao and Tewari (2005).

Finally, an international Pamir-Himalaya project revealed, on average, a 60-km-thick crust in the Kashmir valley, but with strong local variations ranging from as low as 45 km to as deep as 76 km (Kaila et al., 1978, 1979).

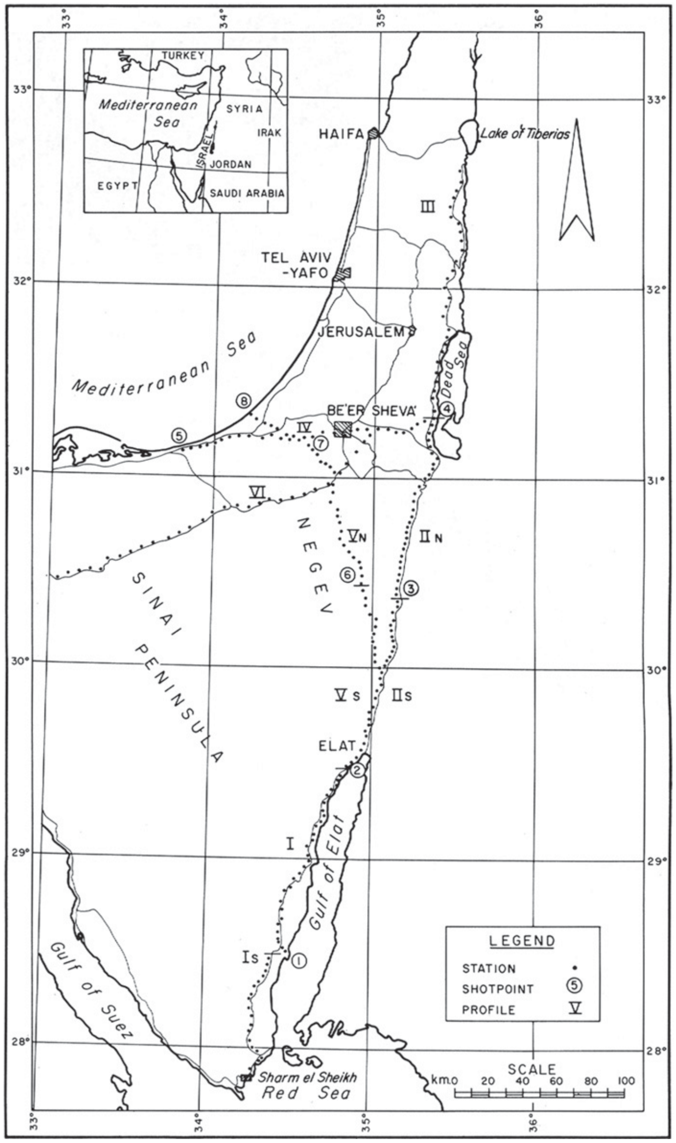


Figure 7.5.2-01. Location map of the seismic refraction survey of 1977 in Israel (from Ginzburg et al., 1979a, fig. 1). [Journal of Geophysical Research, v. 84, p. 1569–1582. Reproduced by permission of American Geophysical Union.]

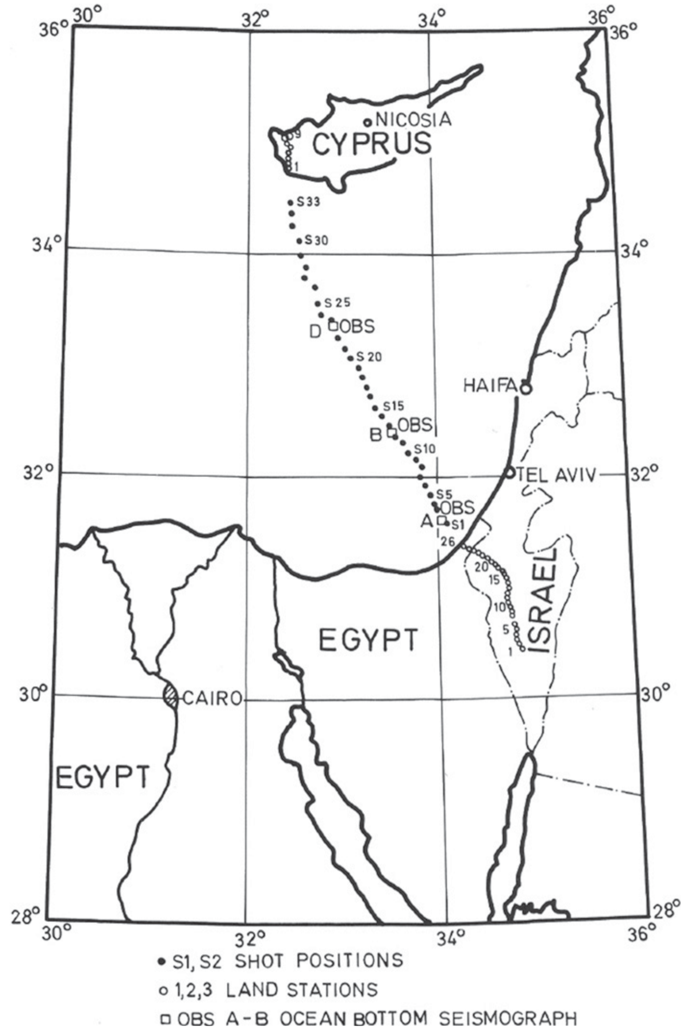


Figure 7.5.2-02. Location map of the seismic refraction survey of 1978 between Cyprus and Israel (from Makris et al., 1983a, fig. 1). [Geophysical Journal of the Royal Astronomical Society, v. 75, p. 575–591. Copyright John Wiley & Sons Ltd.]

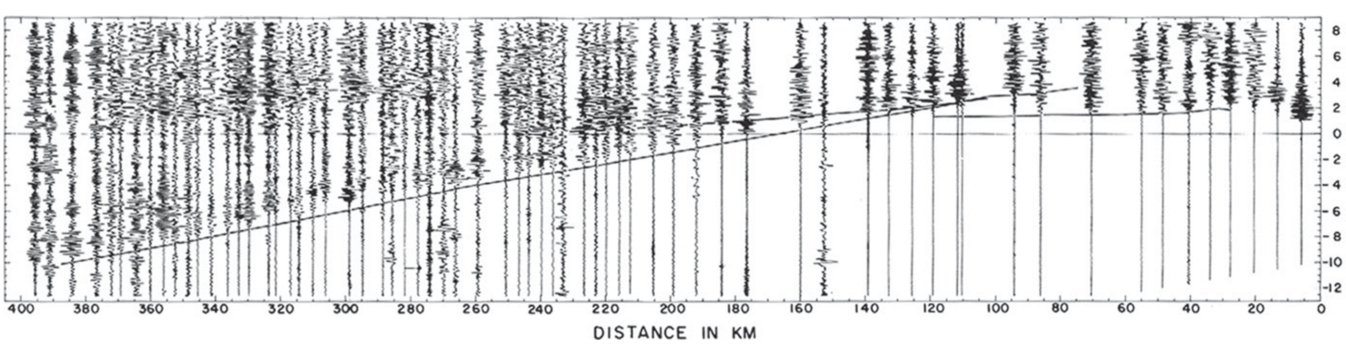


Figure 7.5.2-03. Record section, observed from the Dead Sea towards south to the southern end of the Sinai Peninsula (from Ginzburg et al., 1979a, fig. 4b). For location, see Fig. 7.5.2-01. [Journal of Geophysical Research, v. 84, p. 1569–1582. Reproduced by permission of American Geophysical Union.]

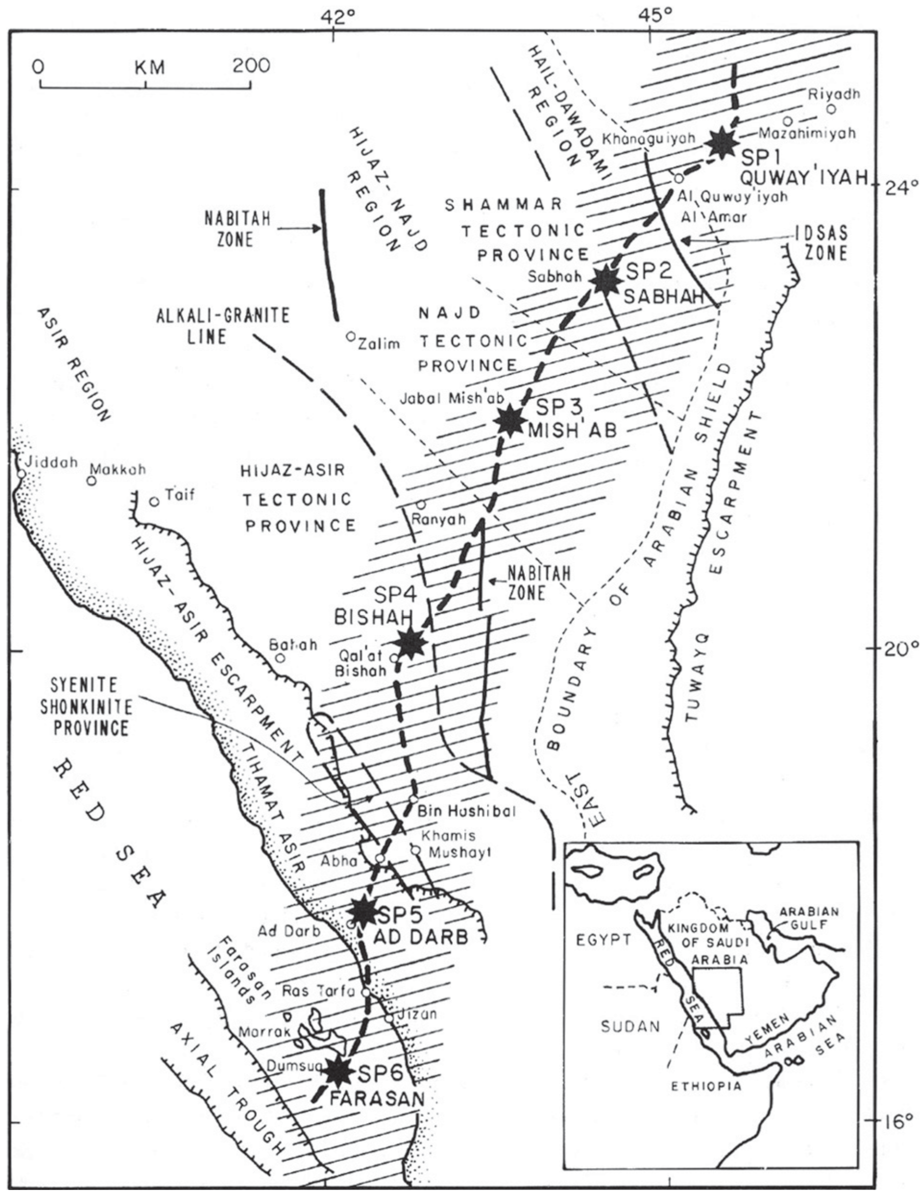


Figure 7.5.2-04. Location map of the long-range seismic refraction survey of 1978 through Saudi Arabia (from Mooney et al., 1985, fig. 1). [Tectonophysics, v. 111, p. 173–246. Copyright Elsevier.]

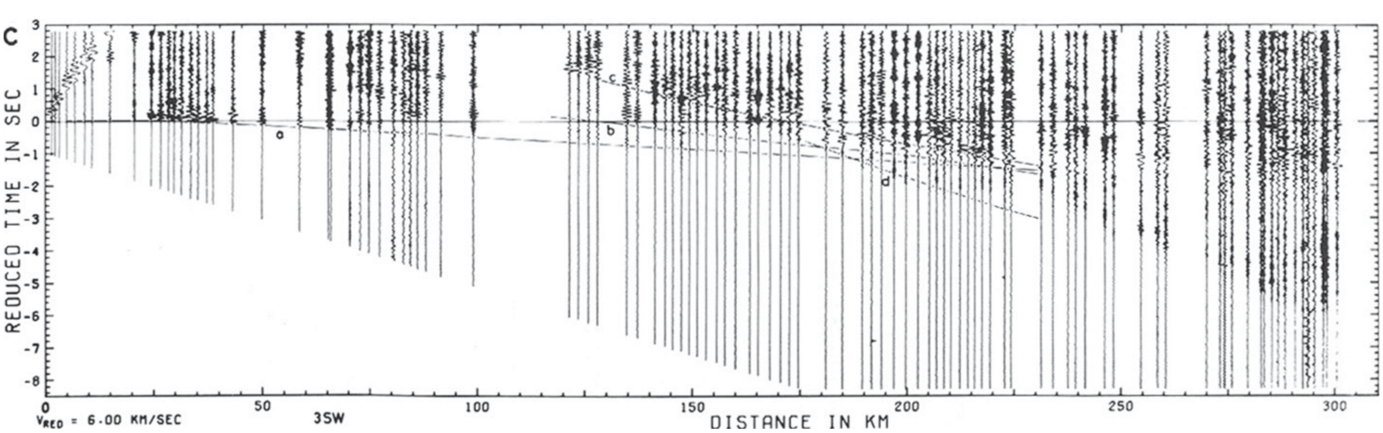


Figure 7.5.2-05. Record section of the Saudi Arabia 1978 experiment, observed across the Arabian Shield (from Mechie et al., 1986, fig. 3c). For location, see Fig. 7.5.2-04. [Tectonophysics, v. 131, p. 333–352. Copyright Elsevier.]

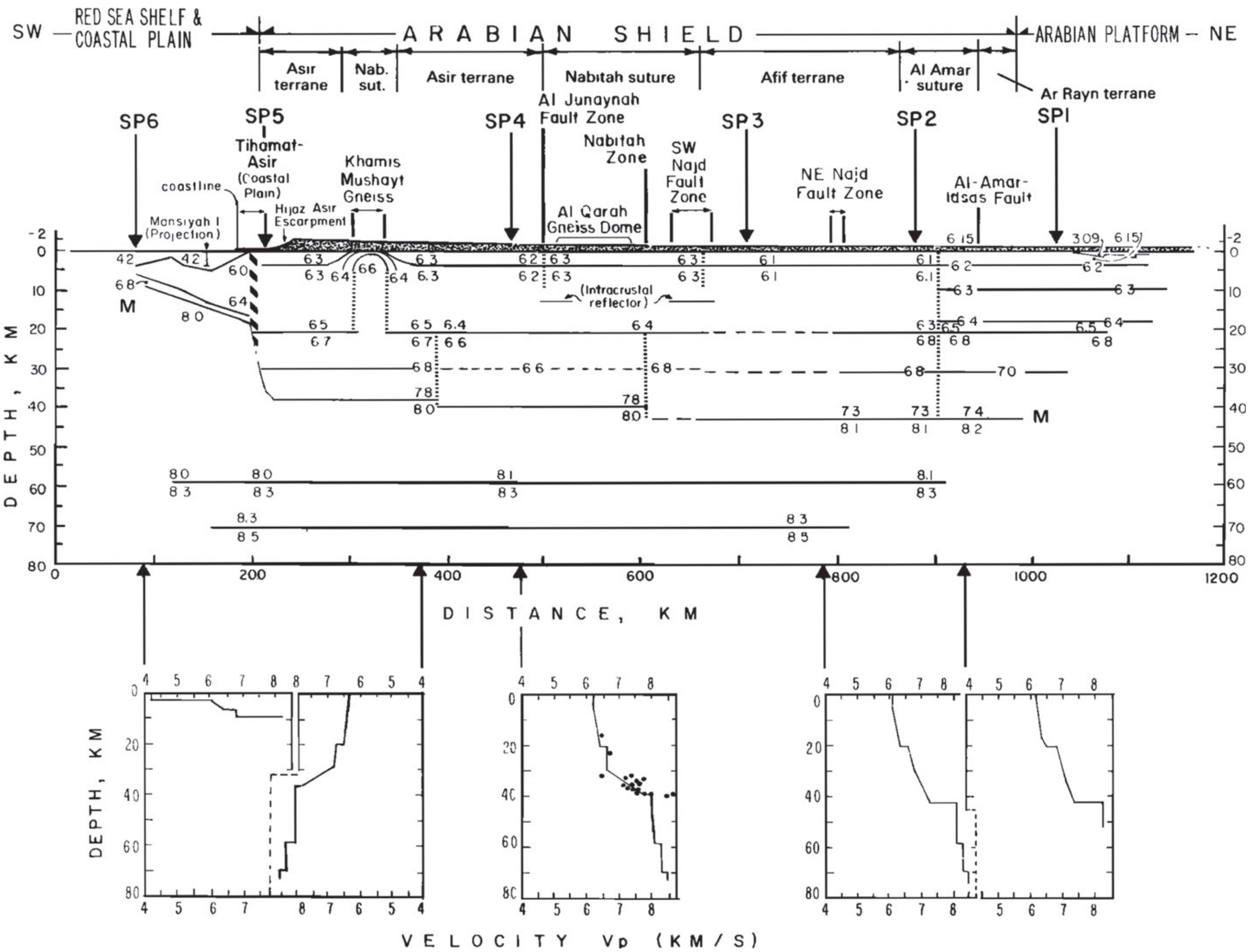


Figure 7.5.2-06. Crustal and upper mantle structure for southwestern Saudi Arabia under the Arabian Shield between the Red Sea and the Arabian platform. Velocity-depth functions for the major tectonic provinces are shown below (from Gettings et al., 1986, fig. 3). [Journal of Geophysical Research, v. 91, p. 6491–6512. Reproduced by permission of American Geophysical Union.]

7.6.2. China

In China, crustal studies were started moderately in the late 1970s and were connected with oil prospecting research in deep basins. Only one of these seismic surveys, carried out in the Qaidam basin in the 1970s, reached Moho level at 51 km depth (Teng et al., 1974; Teng, 1979). Explosion seismic studies of 1975–1977 in Tibet were continued in 1981 and 1982 (see Chapter 8.7.2) and the results, revealing a 75 km maximum thickness of the crust, were summarized by Teng (1987). In southeastern China in 1978, a 1000 ton explosion (Yongping explosion) ~500 km SW of Shanghai enabled the recording of a network of unreversed profiles which extended for 340–380 km toward the SE, NE, and SW and to a distance of 560 km toward the NW. A fence diagram shows a 3-layer crust of ~35 km thickness (United Observing Group of the Yongping Explosion, SSB, 1988; Yuan et al., 1986).

7.6.3. Japan

Japanese research activities in the 1970s were characterized by two directions. Big offshore shot experiments were carried out in 1974–1976 in the Pacific Ocean and the Sea of Japan, and onshore seismic-refraction surveys started in 1979 under the framework of the national project of earthquake prediction. One of these profiles, the Misima-Shimoda profile, was recorded across the Izu peninsula in central Japan (Asano et al., 1982; Yoshii et al., 1985). Another line was recorded on Shikoku Island (Ikami et al., 1982). The onshore activities were continued up to 2003, providing various crustal-scale heterogeneities existing within NE and SW Japan arcs (e.g., Asano et al., 1982; Yoshii, 1994).

The offshore research, carried out under the International Geodynamics Project, elucidated lateral variations in the upper-

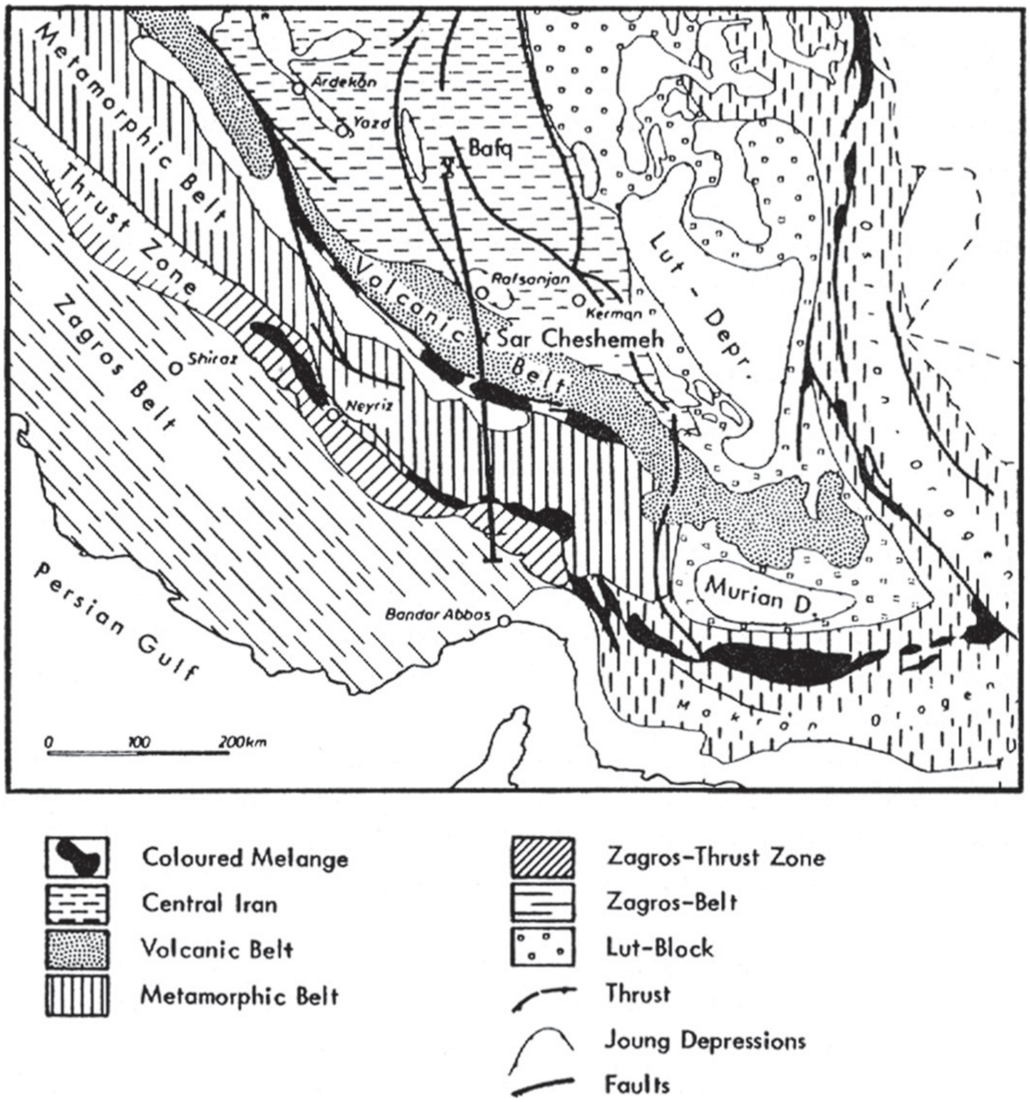


Figure 7.5.2-07. Tectonic map of Iran showing the position of the mines Bafq and Sar Cheshmeh and the layout of the seismic-refraction lines (from Giese et al., 1984, fig. 1). [Neues Jahrbuch für Geologie und Paläontologie Abhandlungen, v. 168, p. 230–243. Reproduced by permission of Schweizerbart, Stuttgart, Germany (www.schweizerbart.de).]

most mantle velocity from the trench area to the back arc basin (Research Group for Explosion Seismology, 1975; Okada et al., 1978, 1979; Yoshii et al., 1981). It is discussed in more detail in subchapter 7.8.1 (Pacific Ocean).

The seismic-refraction observations carried out in the 1970s in Japan, confirmed the basic results of the previous work summarized by the Research Group for Explosion Seismology (1973). These observations showed that the crust thickens from the south coast into the center of the island of Honshu, but that the total crustal thickness, as well as the thickness of the lower crust, proved to be substantially greater than in the previous models (Ikami, 1978; Kaneda et al., 1979).

From data obtained offshore, with ocean-bottom seismometers (Okada et al., 1979), and from previous results (Yoshii and Asano, 1972), e.g., compiled by Research Group for Explosion Seismology (1977), Asano et al. (1981) published a cross section from the Sea of Japan to the Pacific Ocean through northeastern Honshu (Fig. 7.6.3-01). For northeastern Japan (Asano et al., 1979), the upper-mantle velocity is considerably lower (7.5 km/s) than in the other models (7.9 km/s).

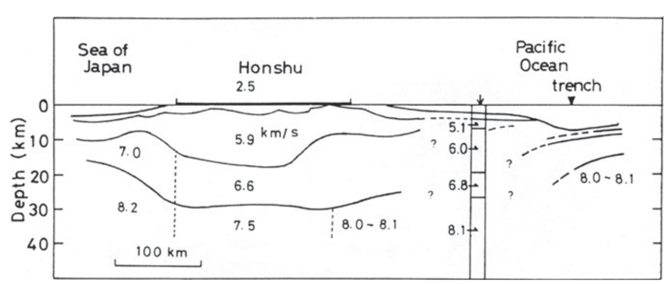


Figure 7.6.3-01. Crust and upper mantle structure from the Sea of Japan to the Pacific Ocean through northeastern Honshu (from Asano et al., 1981, fig. 9). Location of line is shown in Fig. 9.6.3-01, line A–A′. [Journal of the Physics of the Earth, v. 29, p. 267–281. Published by permission of the Seismological Society of Japan.]

7.7. CONTROLLED-SOURCE SEISMOLOGY IN THE SOUTHERN HEMISPHERE

7.7.1. Australia and New Zealand

The deep seismic-reflection experiments of the Bureau of Mineral Resources of the 1960s were continued in the 1970s. Extended records were successfully obtained at projects in the Galilee Basin, Queensland, in 1971 and in the Officer Basin, West Australia, and Murray Basin, New South Wales, in 1972 (Moss and Dooley, 1988), using large charges and in some cases an equipment which was specially modified for 24 s recording. Seismic-reflection investigations continued throughout the 1970s (e.g., Mathur, 1983, 1984). Where more comprehensive surveys were made, the events were consistent with deep reflections, though they lacked continuity in some cases. Finally, in 1976, the Bureau of Mineral Resources introduced digital recording and successfully carried out several deep crustal surveys between 1976 and 1978. Good quality reflection data up to 16 s were obtained at several sites in the southeastern Georgina, western Galilee and Bowen Basins (for locations, see Fig. 8.8.1-03) on short continuous profiles less than 15 km long (Dooley and Moss, 1988; Moss and Mathur, 1986; Finlayson, 2010; Appendix 2-2).

Long-range seismic probing of the crust and upper mantle in Australia, which was started in the 1960s, was continued. The ORD RIVER experiment with a series of shots in 1970 and 1971 (Denham et al., 1972) was already mentioned in Chapter 6.7.1.

In the following years, an increasing number of recording devices allowed the reduction of the spacing of stations considerably. It was particularly the group at the Bureau of Mineral Resources with D. Denham and D.M. Finlayson which took the lead in crustal and upper-mantle research. For investigations of the inland areas of Australia, the sources were mainly quarry blasts or drill-hole shots for mining exploration, but there was also marine work that was involved, e.g., around Tasmania and in the Papua New Guinea area and adjacent island chains.

With the move of A.L. Hales from Dallas, Texas, to the Australian National University at Canberra, active seismology research in Australia became a new component at the university. In particular, upper-mantle studies were the focus of this research. In subsequent years, K.J. Muirhead (Muirhead and Simpson, 1972) constructed long-recording devices (Figs. 7.7.1-01 and 7.7.1-02; Appendix A7-7-1; see also Finlayson 2010; Appendix A2-2) which were positioned on long-range profiles throughout the whole continent to record large explosions and the numerous earthquakes, which occurred frequently and with considerable magnitudes in the Southeast Asian island arcs.

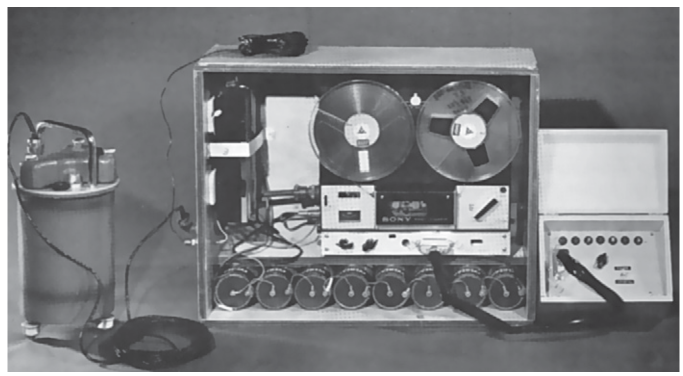


Figure 7.7.1-01. Photograph of three-quarter watt seismic station, designed by Muirhead and Simpson (1972, fig. 3): complete field system with clock set-up box connected at right. [Bulletin of the Seismological Society of America, v. 62, p. 985–990. Reproduced by permission of the Seismological Society of America.]

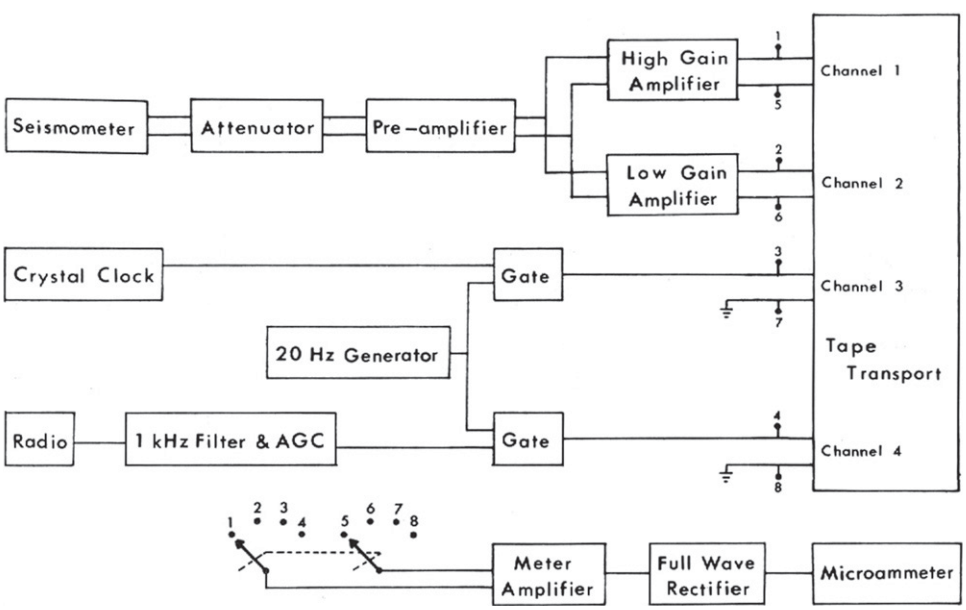


Figure 7.7.1-02. Three-quarter watt seismic station, designed by Muirhead and Simpson (1972, fig. 1): complete field system with clock set-up box connected at right. [Bulletin of the Seismological Society of America, v. 62, p. 985–990. Reproduced by permission of the Seismological Society of America.]

By these means, over the years, N-S–trending long-range lines were assembled (Fig. 7.7.1-03) which allowed scientists to derive refined models of the upper mantle down to several hundred kilometers in depth, showing, for example, discontinuities at 75 km and 200 km and below (e.g., Hales and Rynn, 1978; Hales et al., 1980; Leven, 1980). Data examples for the long-distance ranges of 500–2000 km and 1500–3100 km are shown in Figures 7.7.1-04 and 7.7.1-05 (see also Appendix A7-7-2).

A major survey designed to obtain upper-mantle arrivals was the Trans-Australia seismic survey (TASS) in 1972 was made possible by the Australian mining industry, which provided large explosions at Kunanalling, West Australia (80 tonnes), and Mount Fitton, South Australia (80 tonnes), as well as in the Bass Strait where a 10 ton underwater shot was detonated at 250 m depth (Finlayson et al. (1974). Most of the data was recorded in the range from 200 to 1400 km. According to the position of the shotpoints, three main lines were established (Fig. 7.7.1-06): an east-west line through the Nullarbor Plain, a north-south line toward Alice Springs and a NW-SE line across the Snowy Mountains. For the interpretation (Finlayson et al., 1974; Muirhead et al., 1977; Appendix A7-7-2), the data of earlier explosions such as those from the 1956 Maralinga atomic bomb explosion and the 1970–1971 Ord River explosions were also included. In contrast to similar experiments in Europe, the existence of a low-velocity zone in the subcrustal lithosphere was not evident. Dooley (1979)

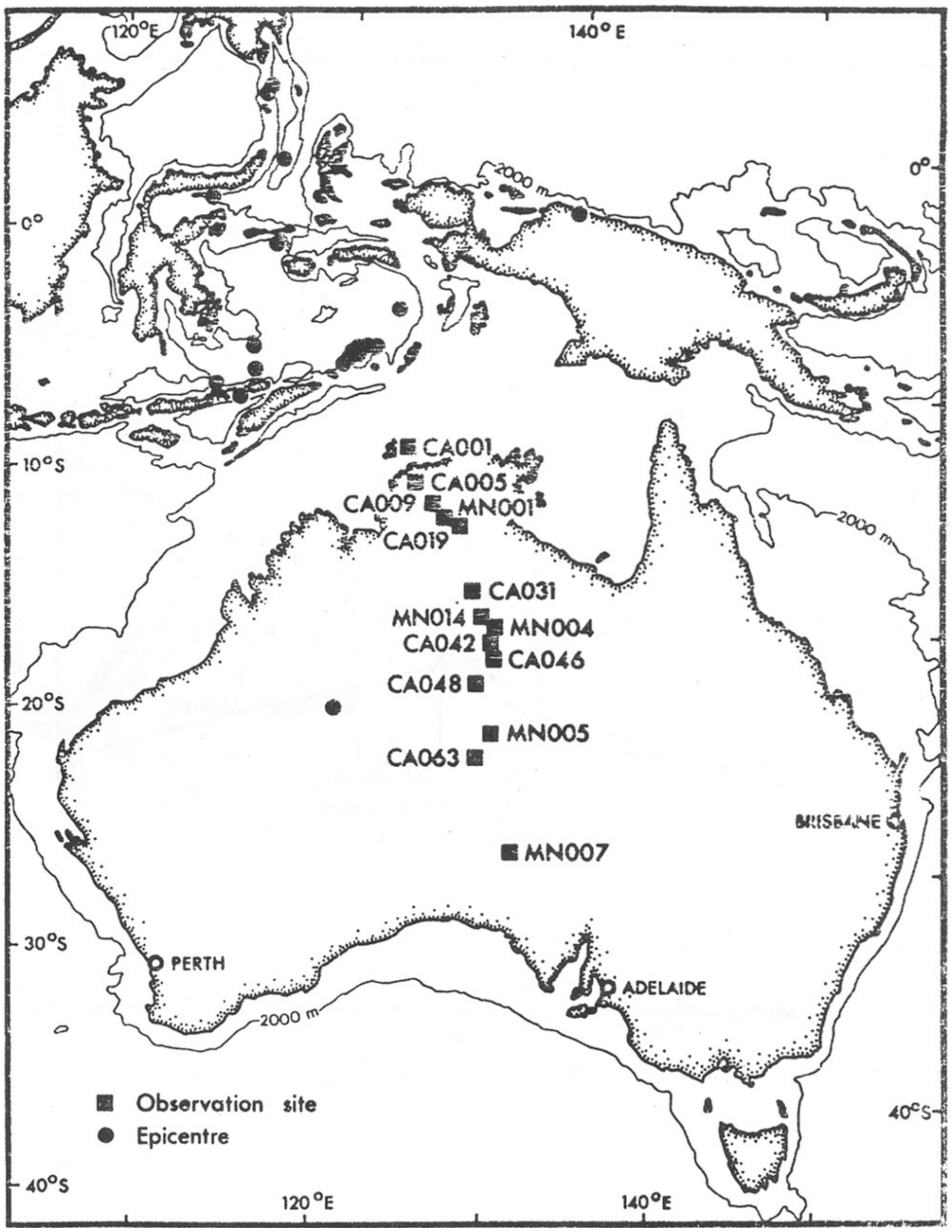


Figure 7.7.1-03. Location of recording stations in and of epicenters north of Australia (from Hales et al., 1980, fig. 1). [Research School of Earth Sciences, Australian National University Publication no. 1280. Published by permission of the Australian National University.]

compiled all available geophysical data along a geotraverse at 29°S (see Appendix A7-7-3); his crustal model from seismic data is shown in Figure 7.7.1-07.

A major effort was undertaken to study the crust under the Precambrian shield of western Australia with the Pilbara shield in the north and the Yilgarn shield in the south separated by the Capricorn orogenic belt. Two projects in 1977 and 1979 were carried out using quarry blasts of open-cut iron-ore mines (Drummond, 1979, 1981). A data example is shown in Figure 7.7.1-08 (for location and more data, see Appendix A7-7-4). A fairly thin crust of 28 km thickness resulted for the northernmost Pilbara craton thickening gradually southwards to 33 km and more underneath the Capricorn orogen, and finally reached a crustal thickness of between 35 and 40 km under the Yilgarn craton in the south (Drummond, 1988; Dentith et al., 2000).

In 1973, a crustal study of the Bowen Basin in Queensland with its extensive coal measures was conducted (Collins, 1978) where a 35-km-thick crust, with surprisingly high velocities both

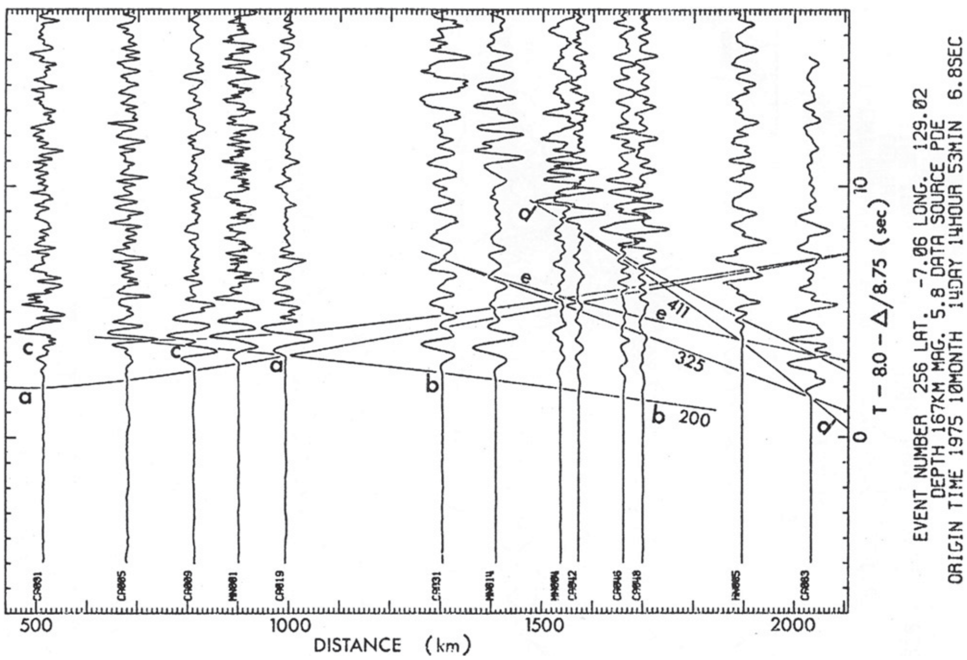


Figure 7.7.1-04. Record section of an event with focal depth of 167 km north of Australia (from Hales et al., 1980, fig. 14). The numbers denote the depth of the corresponding reflector in model 8 (see Appendix 7-7-2, fig. 9). [Research School of Earth Sciences, Australian National University Publication no. 1280. Published by permission of the Australian National University.]

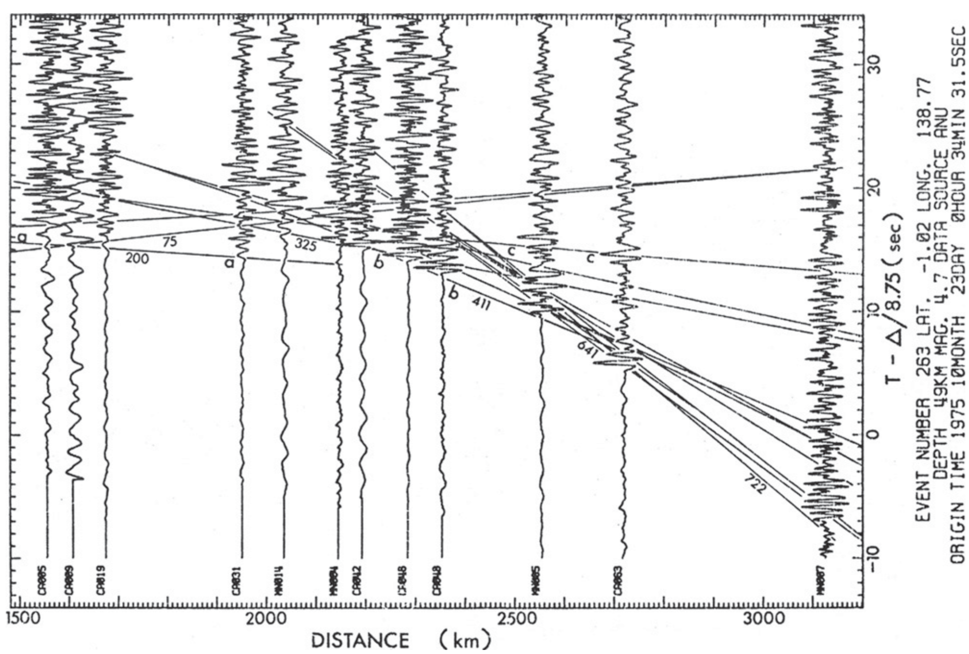


Figure 7.7.1-05. Record section of an event with focal depth of 49 km north of Australia (from Hales et al., 1980, fig. 20). The numbers denote the depth of the corresponding reflector in model 8 (see Appendix 7-7-2, fig. 9). [Research School of Earth Sciences, Australian National University Publication no. 1280. Published by permission of the Australian National University.]

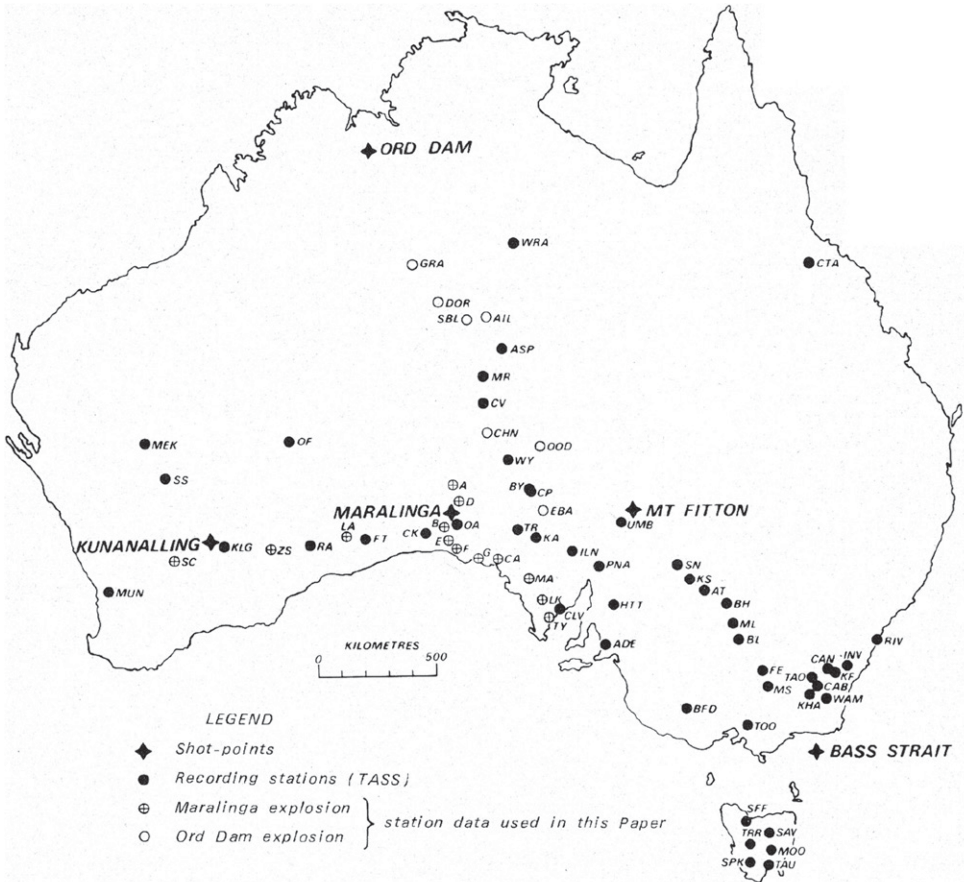


Figure 7.7.1-06. Location of the Trans-Australia seismic survey (from Finlayson et al., 1974, fig. 1). [Journal of the Geological Society of Australia, v. 21, p. 447–458. Published with the permission of the Geological Society of Australia.]

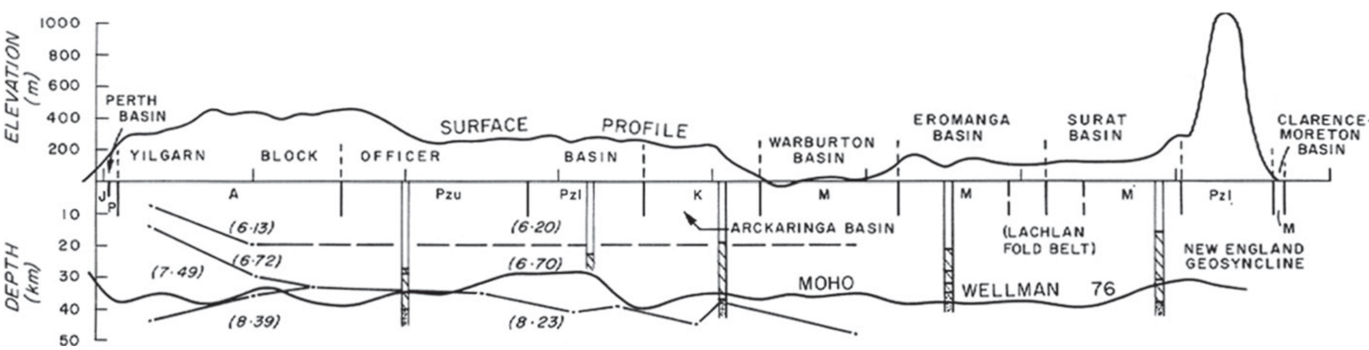


Figure 7.7.1-07. Crustal model along the Trans-Australia seismic survey (from Dooley, 1979, fig. 2c). [BMR Journal of Australian Geology and Geophysics, v. 4, p. 353–359. Copyright by Commonwealth of Australia (Geoscience Australia).]

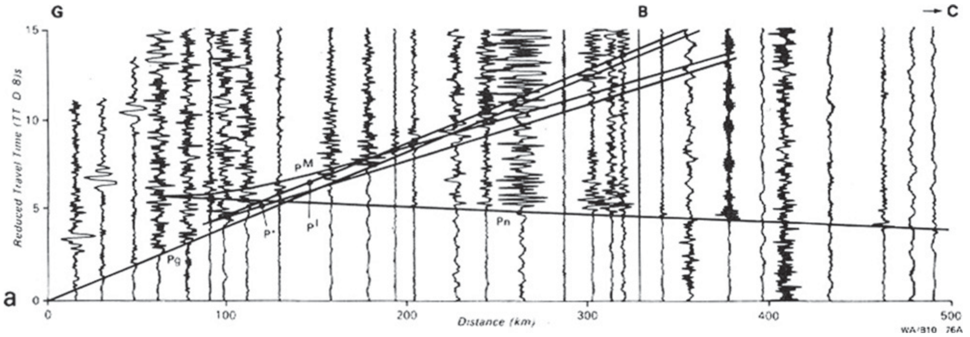


Figure 7.7.1-08. Record section of a profile through the Pilbara-Yilgarn cratons in northwestern Australia (from Drummond, 1979, fig. 3a). [BMR Journal of Australian Geology and Geophysics, v. 4, p. 171–180. Copyright by Commonwealth of Australia (Geoscience Australia).]

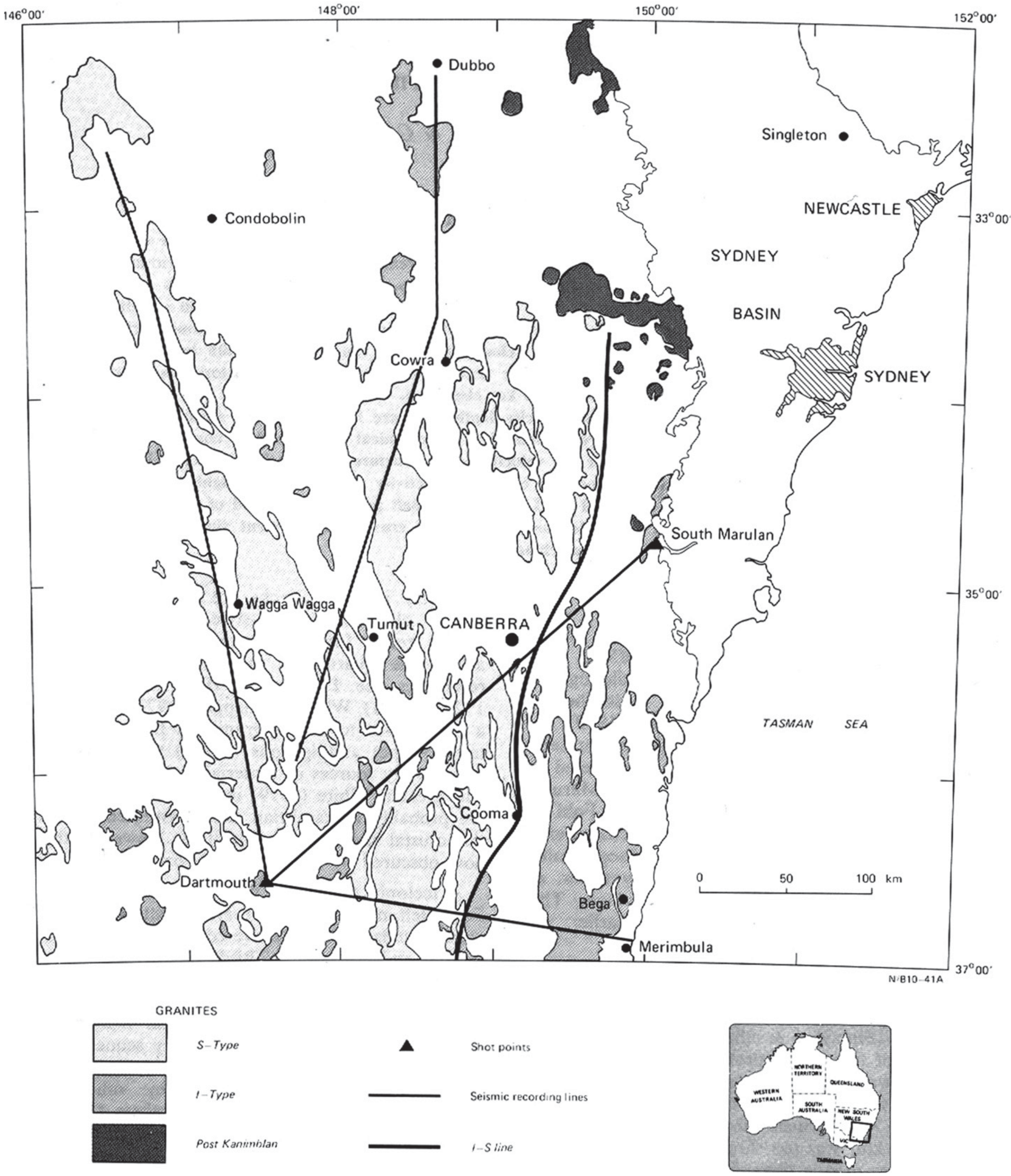


Figure 7.7.1-09. Location of seismic profiles through the Lachlan foldbelt in southeastern Australia (from Finlayson et al., 1979, fig. 1). [BMR Journal of Australian Geology and Geophysics, v. 4, p. 243–252. Copyright by Commonwealth of Australia (Geoscience Australia).]

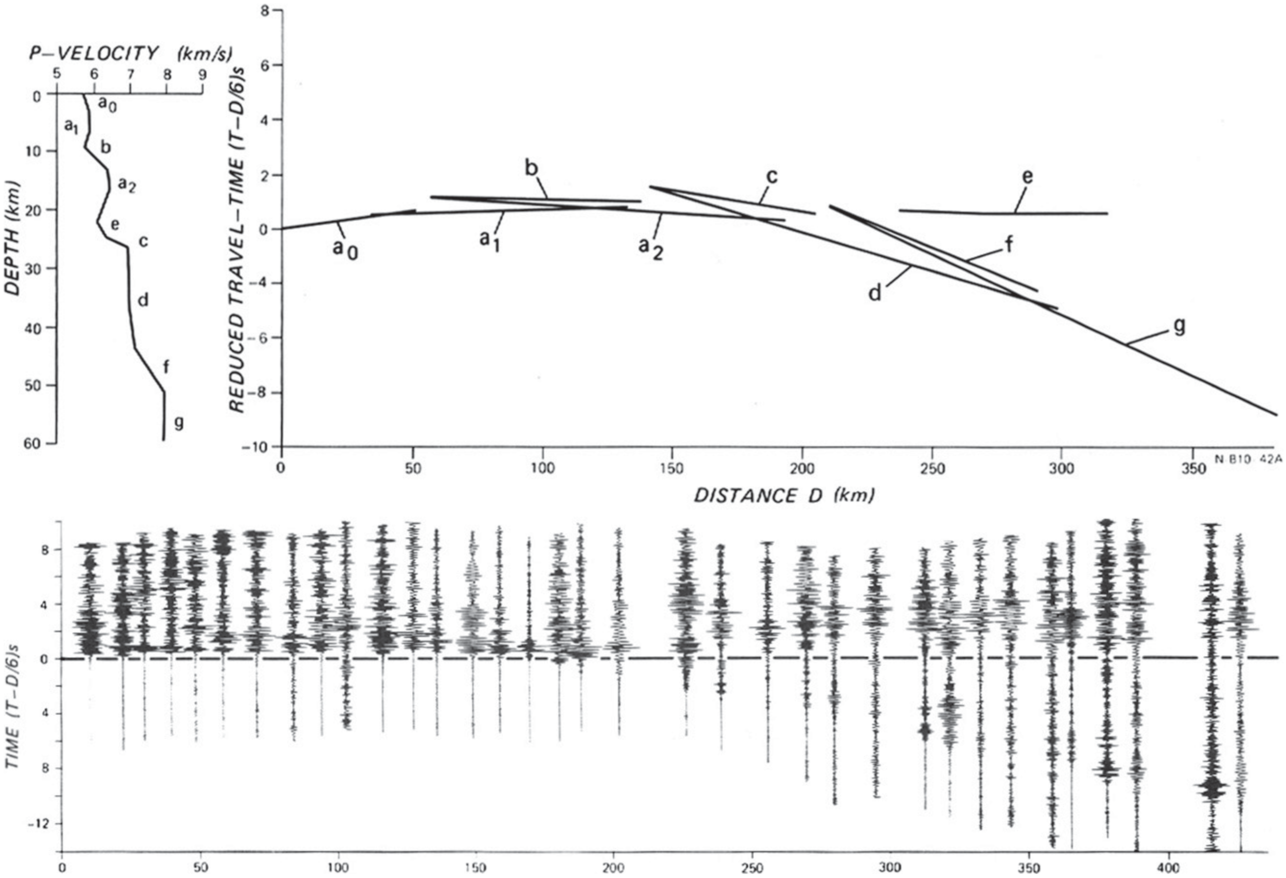


Figure 7.7.1-10. Top: Basic traveltime curve arrangement and model (from Finlayson et al., 1979, fig. 2). Bottom: corresponding data, here record section from Dartmouth towards NNW to Condobolin (from Finlayson et al., 1979, fig. 3a). [BMR Journal of Australian Geology and Geophysics, v. 4, p. 243–252. Copyright by Commonwealth of Australia (Geoscience Australia).]

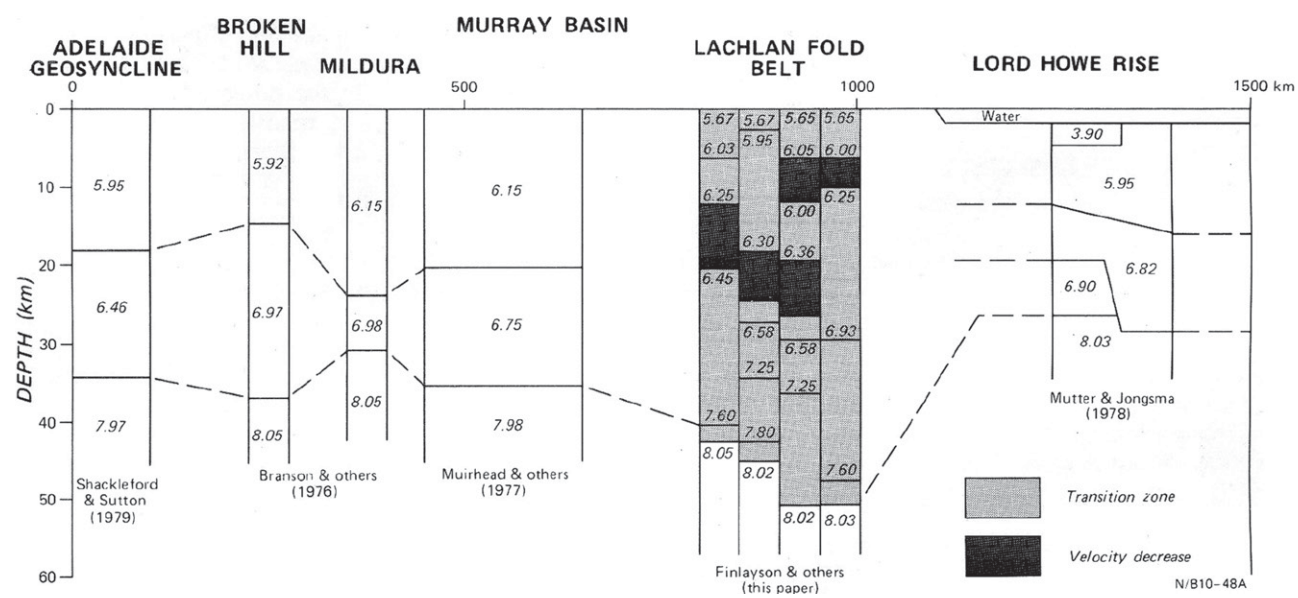


Figure 7.7.1-11. Crustal structure across southeastern Australia and its eastern continental margin (from Finlayson et al., 1979, fig. 8). [BMR Journal of Australian Geology and Geophysics, v. 4, p. 243–252. Copyright by Commonwealth of Australia (Geoscience Australia).]

in the upper and lower crust, was modeled (see Appendix A7-7-5 for data and model).

During the 1970s, a more detailed crustal research project was accomplished in New South Wales, southeastern Australia, in the Lachlan Fold Belt. Here, several lines, all based on quarry blast observations, were assembled between 1976 and 1978 (Finlayson et al., 1979; Fig. 7.7.1-09). In 1978, the survey was extended northward (Finlayson and McCracken, 1981). The interpretation of the data resulted in a thick continental crust averaging 40 km in thickness, but exceeding 50 km under the region of highest topography (Finlayson et al., 1979; Finlayson and McCracken, 1981; Appendix A2-1, p. 104–105, and Appendix A7-7-6). Also, the average crustal velocity was high and the interfaces were best modeled by thick transition zones (Fig. 7.7.1-10).

The cross section through southeastern Australia (Fig. 7.7.1-11) assembled by Finlayson et al. (1979) comprises earlier work in South Australia in the west (Branson et al., 1976; Muirhead et al., 1977; Shackleford and Sutton, 1979) and includes the Lord Howe Rise (Mutter and Jongsma, 1978) to the east of the southeastern Australian margin.

In 1979, a 500-km-long east-west line was recorded between the Mount Isa Mine, Queensland, using a 40 t rock breaking shot as the eastern source and the abandoned Skipper Mine near Tennant Creek, Northern Territory, shooting a 5 t charge at 40 m depth at the western end. In addition, on 9 August 1979, the stations along the profile recorded unexpectedly the New Britain earthquake (Finlayson, 1982).

In New Zealand in the 1970s, two offshore-onland seismic-refraction experiments were carried out. In 1974 and 1975, marine shots in fjords on the South Island were recorded by stations on land along several lines of up to 100 km length (for location, see Fig. 8.8.1-05). The objective was to verify an inferred shallow Moho under a high gravity anomaly (Davey and Broadbent, 1980; Priestley and Davey, 1983). The structure derived from the traveltime data was markedly different from the nearby stable areas of South Island. They gave no indication of a typical 6.0 km/s upper crustal layer, but substantiated the existence of high-velocity material typical of lower crust at shallow depths beneath western Fjordland and supported the idea that western Fjordlands represents a section of uplifted lower crust. The data did not succeed in defining the position of the Moho. At least there was no evidence that the Moho was shallower than 15 km.

The project WHUMP (Wellington–Hikurangi Upper Mantle Profile) of 1979, a SW-NE–directed crustal seismic-reflection–seismic-refraction experiment, was centered on Wellington (for location see Fig. 8.8.1-05). The project aimed to trace the extent of a strong reflector occurring at a depth of ~20 km beneath Wellington and to investigate if this reflector coincides with the interface between the lithosphere of the Indian plate and the underlying lithosphere of the Pacific plate, which is currently being subducted at the Hikurangi Trough in the east and dips under North Island, New Zealand, to the west (Davey and Smith, 1983). The measurements were made along a 50-km-long profile along the southwestern edge of the North Island between the Palliser Bay in the southeast and the Cook Strait in the northwest recording shots from two shotpoints. A series of shots fired 500 m offshore at 10 m water depth on the seafloor near Wellington were recorded at 1 km intervals in northwest and southeast direction. The northwest sector of the profile was partly reversed by a series of shots at sea at the northwestern end of the line.

7.7.2. South Africa

In the University of Witwatersrand at Johannesburg, South Africa, R.W. Green had developed instrumentation which allowed him to carry out one-man seismic-refraction surveys (Green, 1973). He would install his recording devices (BPI) and let them run in a continuous mode for up to a week, and in between he would organize shotpoints, load them, and detonate the charges whenever ready. This worked very well for small-scale surveys.

At the same time, in southwest Africa, present-day Namibia, a special geological research project was carried out to study the Damara orogen with stratigraphic, tectonic, and petrographic

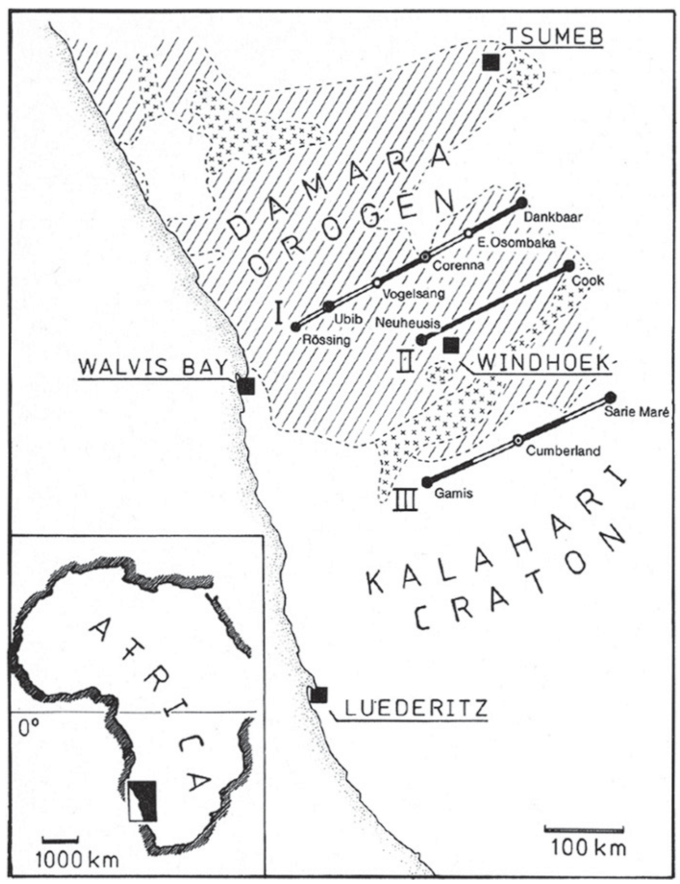


Figure 7.7.2-01. Simplified map of the Damara orogen in Namibia and location of the seismic-refraction lines (from Baier et al., 1983, fig. 1). [*In* Martin, H., and Eder, F.W., eds., Intracontinental Fold Belts: Berlin-Heidelberg, Springer, p. 885–900. Reproduced with kind permission of Springer Science+Business Media.]

investigations undertaken by the University of Goettingen, Germany, and funded by the German Research Society.

Within this research project, the crustal structure was also to be studied, for which purpose a cooperation of German geophysical institutes with the University of Witwatersrand at Johannesburg was initiated (Baier et al., 1983; Appendix A2-1, p. 101–103, and Appendix A7-8). The deep seismic sounding research was carried out in 1975. It involved three profiles of 225–350 km length, with shotpoints at the ends, and with some in between (Fig. 7.7.2-01). All three of these profiles ran in a WSW-ENE direction. Two of them traversed the Damara orogen (Fig. 7.7.2-02A) and the third traversed the Kalahari craton in the

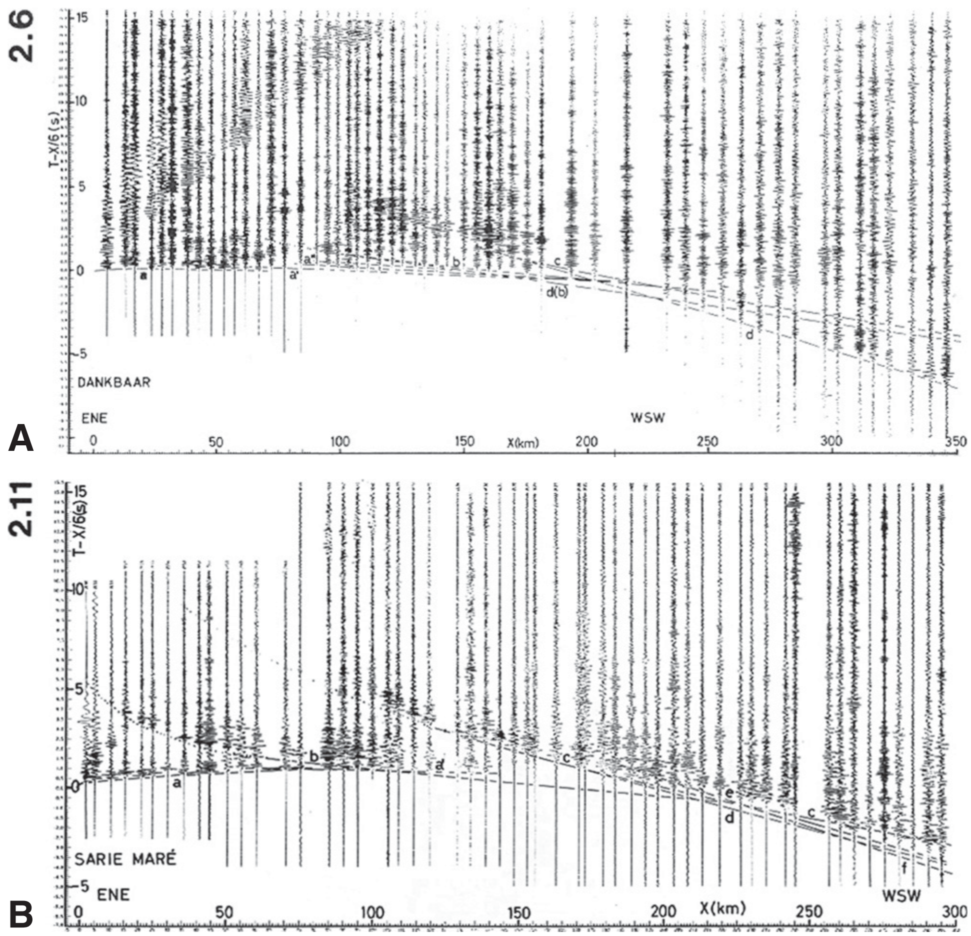


Figure 7.7.2-02. Record sections of the Namibia 1975 project. (A) Profile through the Damara orogen. (B) Profile through the Kalahari craton (from Baier et al., 1983, fig. 2). [*In* Martin, H., and Eder, F.W., eds., Intracontinental Fold Belts: Berlin-Heidelberg, Springer, p. 885–900. Reproduced with kind permission of Springer Science+Business Media.]

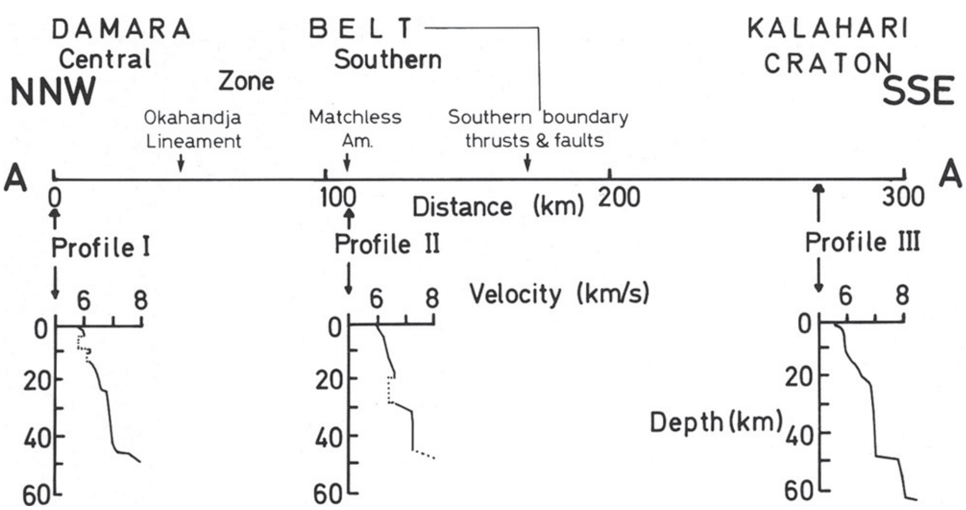


Figure 7.7.2-03. Velocity-depth sections through the Damara orogen and adjacent Kalahari craton, approximately 380 km NNE of the Atlantic coast (from Baier et al., 1983, fig. 4). [*In* Martin, H., and Eder, F.W., eds., Intracontinental Fold Belts: Berlin-Heidelberg, Springer, p. 885–900. Reproduced with kind permission of Springer Science+Business Media.]

south (Fig. 7.7.2-02B). With the exception of the westernmost shot of the northern profile I, which was a commercial mine blast, all shots were drill-hole shots arranged by teams of the participating institutions. Fifteen MARS-66 stations and 15 BPI stations were available for recording. The results were interpreted as a 47-km-thick crust, both underneath the Damara orogen and the adjacent Kalahari craton (Fig. 7.7.2-03).

7.7.3. South America

Crustal research in South America in the 1970s concentrated on the Andean region of Colombia and Ecuador, as well as the adjacent Pacific Ocean. In 1973, the Narino III project was launched between latitudes 2° and 4°N and was jointly organized by the University of Wisconsin under the direction of R.P. Meyer and the University of Texas at Dallas under the direction of A.L. Hales. Furthermore, the Carnegie Institution of Washington and the University of Kiel, Germany, joined in, arranging and recording quarry blast shots. The aim of the Narino III project was to investigate the crustal and upper-mantle structure in the area of the subducting Pacific plate of oceanic character underneath the South American continent (Meyer et al., 1976). The project involved a series of sea shots along lines away from the Pacific coast and up to 400 km long, as well as parallel to the coast, which were recorded by stations on land. The additional quarry blasts were recorded along a north-south line along the eastern edge of the western Cordillera (Fig. 7.7.3-01). Some of the results are shown in Figures 7.7.3-02 and 7.7.3-03.

As part of the Narino project, the University of Kiel, Germany, also was involved. The crustal cross sections derived for the southern part of Colombia by Meissner et al. (1976b) revealed an average thickness of the crust of 20 km in the oceanic and western part of the line and of 30–40 km underneath the continental part.

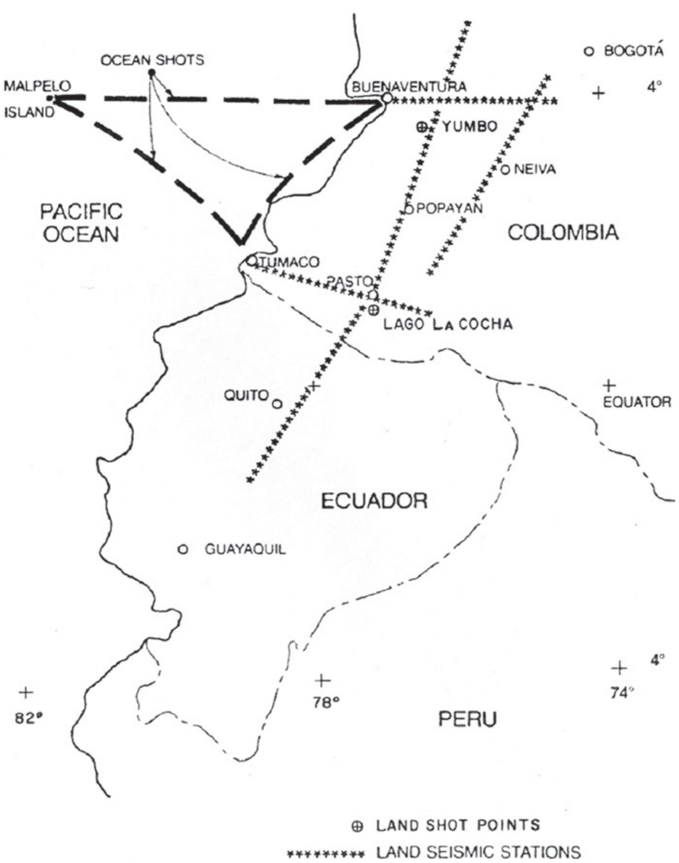


Figure 7.7.3-01. Sketch map of the Narino project in Colombia and Ecuador (from Meyer et al., 1976, fig. 1). [*In* Sutton, G.H., Manghnani, M.H., Moberly, R., and McAffe, E.U., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union Geophysical Monograph 19, p. 105–132. Reproduced by permission of American Geophysical Union.]

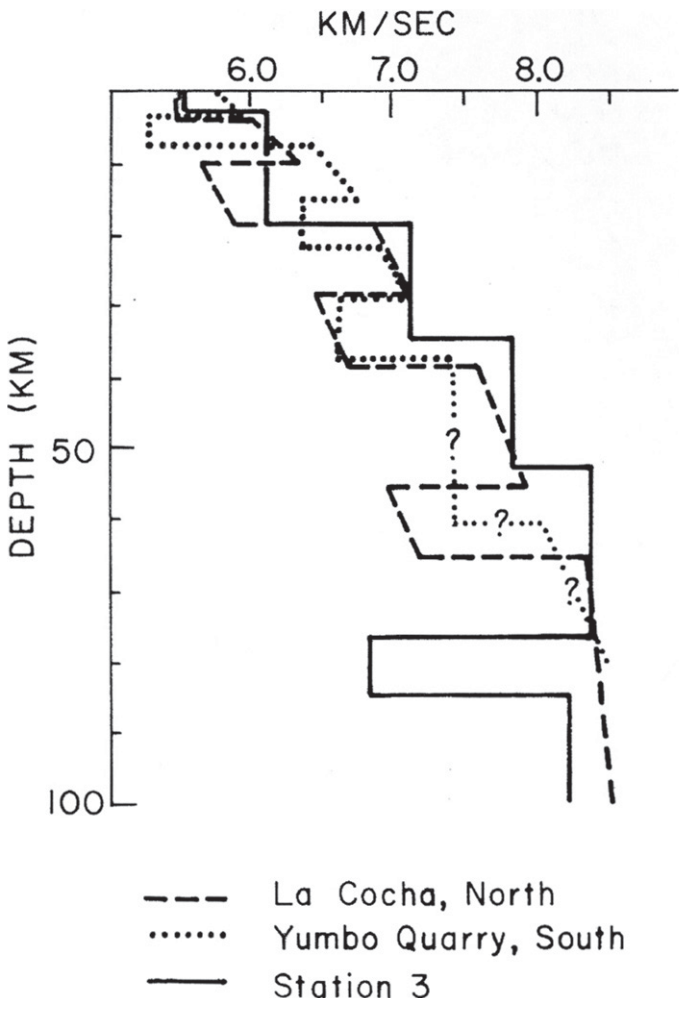


Figure 7.7.3-02. Velocity-depth functions for quarry-blast profiles running north-south along the eastern edge of the western Cordillera in comparison with Narino results (station 3) (from Meyer et al., 1976, fig. 24).). [*In* Sutton, G.H., Manghnani, M.H., Moberly, R., and McAffe, E.U., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union Geophysical Monograph 19, p. 105–132. Reproduced by permission of American Geophysical Union.]

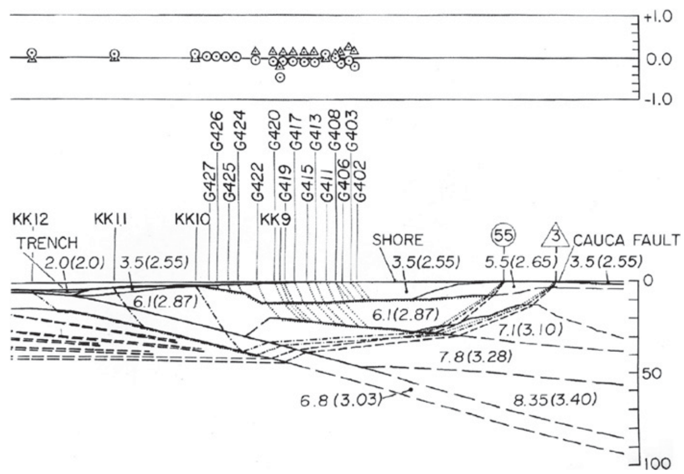


Figure 7.7.3-03. True-scale model derived from Narino data and gravity with shotpoints G402-G427 and KK12–KK9 (KK19–KK13 are farther west), ray paths and stations 55 and 3. Dotted and dash-dotted paths are critically refracted along the continental and oceanic mantle; dashed paths simulate diving waves (from Meyer et al., 1976, fig. 25). The numbers in the model are P-wave velocities; those in brackets densities. [*In* Sutton, G.H., Manghnani, M.H., Moberly, R., and McAffe, E.U., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union Geophysical Monograph 19, p. 105–132. Reproduced by permission of American Geophysical Union.]

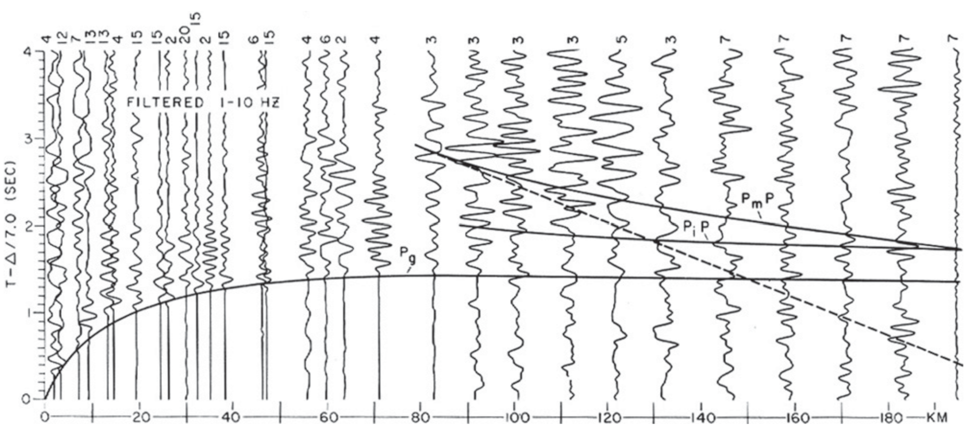


Figure 7.7.3-04. Record sections of the Colombia 1976 project. Yumbo-North profile (from Mooney et al., 1979, fig. 7a). [Bulletin of the Seismological Society of America, v. 69, p. 1745–1761. Reproduced by permission of the Seismological Society of America.]

In 1976, additional quarry blasts at one of the shotpoints of 1973 were arranged and recorded in the Western Cordillera of the Colombian Andes up to 200 km along strike as well as 60 km toward the west. In particular, the data of the quarry Yumbo, recorded to the north, were of very good quality (Fig. 7.7.3-04). A crustal thickness of 30 km resulted for this area (Mooney et al., 1979; Appendix A7-9-1).

In 1978, an experiment was carried out farther north, around latitude 5.5°N, with a series of shots at sea and a 350-km-long line crossing the Serra de Baudo near the coast and the Western and Central Cordilleras of northwestern Colombia (Flueh et al., 1981; Appendix A7-9-2). Again there was an international cooperation, involving U.S., German, Spanish, and Colombian scientists. Here, the Moho depth increases from 17 km near the coast to 30 km under the Serra de Baudo, and remains at this depth under the Western Cordillera and finally increases further to 45 km under the Central Cordillera.

7.7.4. Antarctica

A very detailed seismic survey by the Institute of Geophysics of the Polish Academy of Sciences was carried out in 1979 and 1980 in the southernmost end of the Pacific Ocean. The survey was to explore the structure underneath West Antarctica, and it covered the area from 61° to 65°S and 56° to 66°W.

The seismic measurements reached from the southern Shetland Islands to the Antarctic Peninsula (Fig. 7.7.4-01) and included reflection seismic profiling along shelf geotraverses, shallow seismic-refraction soundings on shelves, deep seismic soundings of the whole crust, and magnetic profiling (Guterch et al., 1985).

The deep seismic sounding profiles reached a total length of 2000 km. The interpretation of their results was as follows: the crust is 32 km thick in the region of the South Shetland Islands, and thickens to 40–45 km in the region of the Antarctic Peninsula. A preliminary geodynamic model by Guterch et al. (1985) is shown in Figure 7.7.4-02.

7.8. OCEANIC DEEP STRUCTURE RESEARCH

7.8.1. Introduction

Until the mid-1960s, all offshore seismic exploration work used exclusively chemical explosions as energy sources. For deep crustal studies, this remained so in the 1970s. In particular for offshore-onshore projects, when shots at sea were both recorded

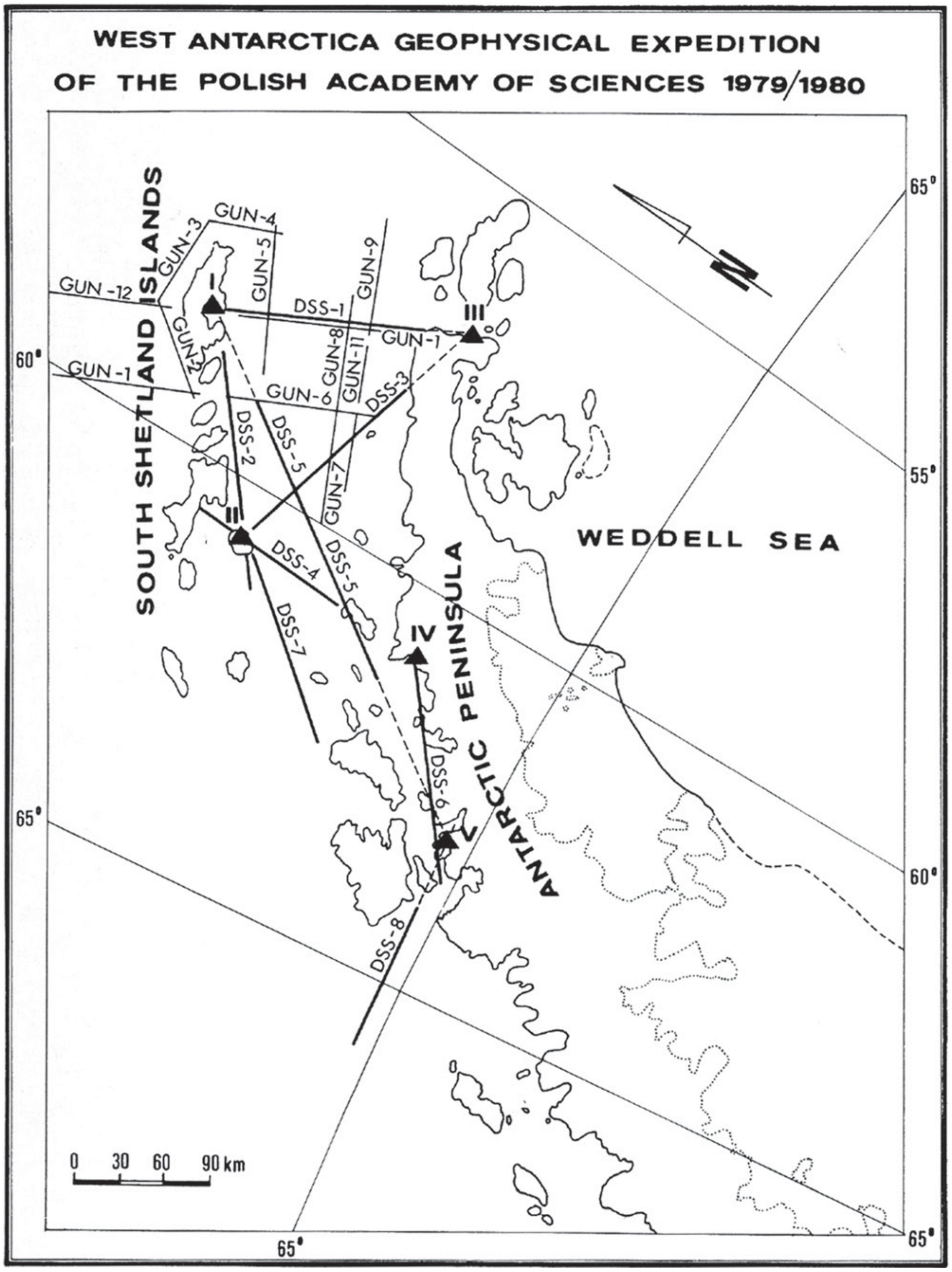


Figure 7.7.4-01. West Antarctica Polish expedition 1979–1980: Location of seismic profiles. DSS—deep seismic sounding profiles; GUN—reflection profiles; I, II, III, IV, V—position of seismic stations (from Guterch et al., 1985, fig. 1). [*In* Husebye, E.S., Johnson, G.L., and Kristoffersen, Y., eds., Geophysics of the polar regions: Tectonophysics, v. 114, p. 411–429. Copyright Elsevier.]

by recording devices on research vessels and on land, as well as for the 1000–2000-km-long land profiles in western Europe, small to large depth charges up to several tons of dynamite were detonated in shallow shelf waters wherever available. For the LISPB project of 1974, for example, Brian Jacob had carried out pre-studies to optimize the shots by splitting the single charges into smaller units and obtaining both a smaller quantity of explosives and a shallower optimum depth for underwater shots carrying sufficient energy over 1000 km distances (see subchapter 7.2.4). The development of airguns offered the chance to repeat shots much more rapidly than was possible with chemical explosions. Furthermore, the shallow depth of Moho under the deep oceans did not require profile lengths of several 100 km as was needed for land profiles. So, from the mid-1970s onward, marine crustal data could very efficiently and at a fast speed be collected in large quantities and the oceanic structure could now be unraveled with a much higher precision. But in many deep-reaching scientific research projects explosives were still preferred as the most powerful energy source.

Dobrin (1976) provided background information on basic multichannel seismic theory and techniques of computer processing. Stoffa and Buhl (1979) discussed in detail the value of expanding spread profiling and constant offset profiling for deep crustal studies. They acquired expanding spread profile data

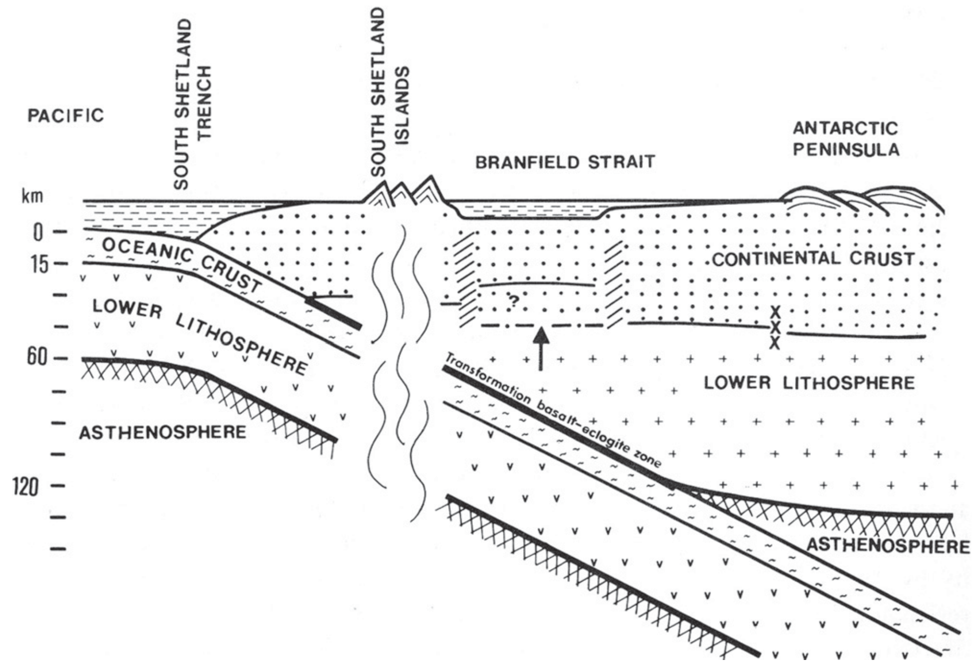


Figure 7.7.4-02. Preliminary geodynamic model of West Antarctica (from Guterch et al., 1985, fig. 12). [*In* Husebye, E.S., Johnson, G.L., and Kristoffersen, Y., eds., Geophysics of the polar regions: Tectonophysics, v. 114, p. 411–429. Copyright Elsevier.]

in the western Pacific, in the Caribbean, and in the Norwegian Sea and discussed in detail the accuracies and advantages of this specific techniques. They concluded that in the late 1970s, these experimental techniques were the only method available for a detailed continuous mapping of lateral crustal variations from nonreflecting horizons.

The 1970s were also the decade, when ocean-bottom seismometers (OBS) and ocean-bottom hydrophones were developed in various institutions and successfully tested in experiments around the world (Asada and Shimamura, 1976, 1979; Avedik et al., 1978; Carmichael et al., 1973; Rykunov and Sedov, 1967; Whitmarsh and Lilwall, 1983). Another example is "a new and inexpensive pop-up ocean-bottom hydrophone recorder" which had been developed in 1978 at the Bullard Laboratories at Cambridge, UK, for use in seismic-refraction experiments, in order to supplement or to replace the existing recording sonobuoys. The first "field" test at sea could be started in 1979 (Sinha et al., 1981a); in 1980 it was regarded to be ready.

These ocean-bottom instruments are self-contained recording units which are brought back to the surface on a line or after ballast weights are discarded by an acoustic command from the retrieving vessel (pop-up systems). They offered important advantages over free-floating sonobuoys. They could be placed at ranges beyond the maximum attainable with telemetering sonobuoys and could therefore be used for long offset seismic probing of several tens of kilometers below the seafloor. P-waves from shallow structural levels were recorded as first arrivals. The noise level was considerably lower, because the high-energy environment of the upper-water column was removed. The uncertainties in interpreting seismic traveltimes were considerably reduced, because the instruments were fixed on the ocean bottom. In an ocean-bottom seismograph both vertical and horizontal component seismometers and additional hydrophones could be incorporated.

From the numerous new investigations carried out in the 1970s, using airguns and expendable sonobuoys, a more detailed picture of the crust under the ocean basins could be obtained. Figure 7.8.1-01 presents a world map showing all locations of the Lamont-Doherty Laboratory where recordings with sonobuoys were made until the mid-1970s and which were used for a special investigation for the upper crustal structure (Houtz and Ewing, 1976). Following the summaries of Ewing and Houtz (1979), Jones (1999) and Minshull (2002), due to new and improved data, in many areas the seismic layer 2 had now to be split into two units, 2A and 2B, with average velocities of 3.5 km/s and 5.2 km/s. Between layers 2B and 3, refracted waves revealed a zone 2C with velocities of 6.0–6.2 km/s. In some regions, basement with velocities of 7.2–7.4 km/s was seen to occur just above Moho. This more refined crustal structure had the result, that Moho depth was ~1 km deeper than calculated from the earlier deep-sea refraction data. A correlation of these finer subdivisions of the oceanic basement with the P-wave velocity structure of ophiolites on land appeared satisfying.

The subdivision of layer 2 into layers 2A and 2B was originally based on slope-intersect interpretations of sonobuoy profiles, where layer 2A was recognized as a thin layer with low velocities at the top of the crust which did not generate refracted arrivals. The introduction of such a layer became necessary, because the layer-2 arrivals were not tangential to the top-basement reflections, while layer 2B was the deeper higher-velocity layer which generated the observed refracted arrivals (Houtz, 1976; Houtz and Ewing, 1976; Carlson, 1998; Minshull, 2002).

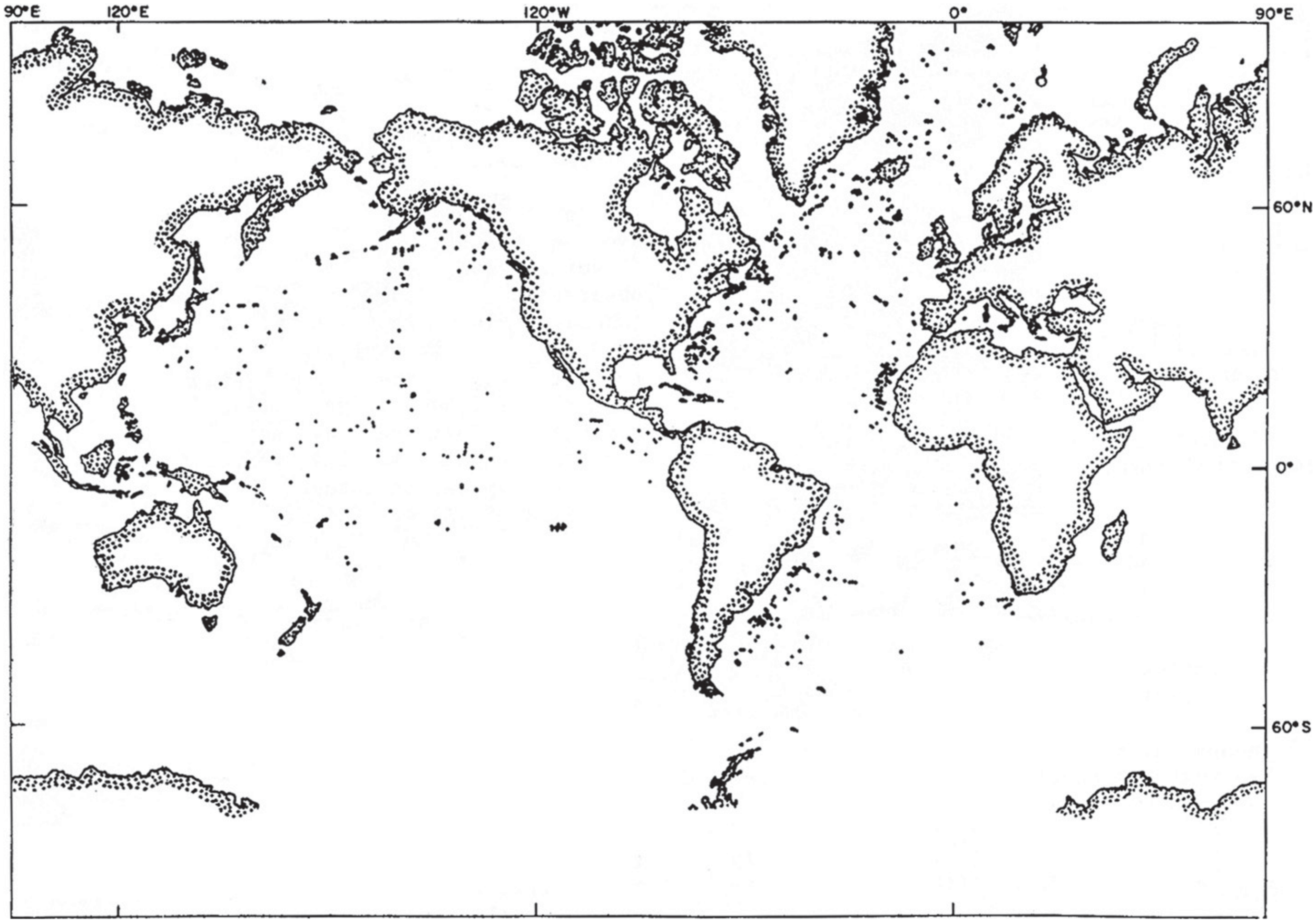


Figure 7.8.1-01. Location of Lamont-Doherty sonobuoy stations (from Houtz and Ewing, 1976, fig. 1). [Journal of Geophysical Research, v. 81, p. 2490–2498. Reproduced by permission of American Geophysical Union.]

Layer 2A has attracted geoscientists in connection with plate movements because of its supposed age-dependent properties. Therefore Houtz and Ewing (1976) and later in a subsequent study Houtz (1976) have analyzed all Lamont-Doherty sonobuoy data around the world as far as they were located on well-dated crust (Fig. 7.8.1-01). As a result, layer 2A was thicker on the Mid-Atlantic Ridge crest (1.5 km) than on the East-Pacific Ridge crest (0.7 km). In both oceans, layer 2A thinned gradually away from the ridge to ~100 m thickness, which was reached in the Atlantic at 60 m.y. old crust, in the Pacific at 30 m.y. old crust. This was attributed to the much faster spreading rate in the Pacific. Furthermore, regressions on refraction velocities in layer 2A as a function of age showed that its velocity increased from ~3.3 km/s at the ridge crests to that of layer 2B on crust ~40 m.y. old. In none of the deeper layers could a corresponding increase of velocity with age be detected.

The application of the reflectivity method to calculate synthetic seismograms for deep-ocean data proved to be most powerful to unravel details of the basement structure (Spudich and Orcutt, 1980). It was shown that the structure was better represented by velocity gradient zones than by discrete constant-velocity layers. Figure 7.8.1-02 shows a record section from observed data. Depending on the gradient, reflections were strong or weak. Figure 7.8.1-03 shows synthetic seismograms for an ocean-crust model with a velocity gradient above Moho, instead of Moho being a first-order discontinuity, which would not explain the observed data. The use of record sections and the reanalysis of the early data by modern synthetic seismogram analysis (White et al., 1992) showed that the slope-intercept time method systematically underestimated the thickness of the oceanic crustal layers, because secondary arrivals had not been recognized as reflections, but, if at all, had been interpreted as strong refracted arrivals. As Minshull (2002) points out, on modern record sections plotted from the early data, a strong layer-2 arrival, a weak layer-3 arrival, a strong Moho reflection, and weak mantle arrivals could be correlated and compared with synthetic seismograms.

There was also research work going on to determine if seismic anisotropy could be observed in the crust. In the 1960s,

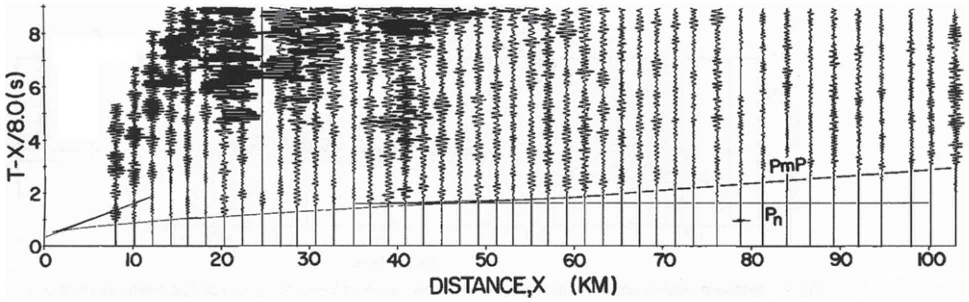


Figure 7.8.1-02. Observed data highlighting the P_MP phase, reflected from Moho, and the P_n phase, refracted in the uppermost mantle layer (from Spudich and Orcutt, 1980, fig. 4). [Reviews of Geophysics and Space Physics, v. 18, p. 627–645. Reproduced by permission of American Geophysical Union.]

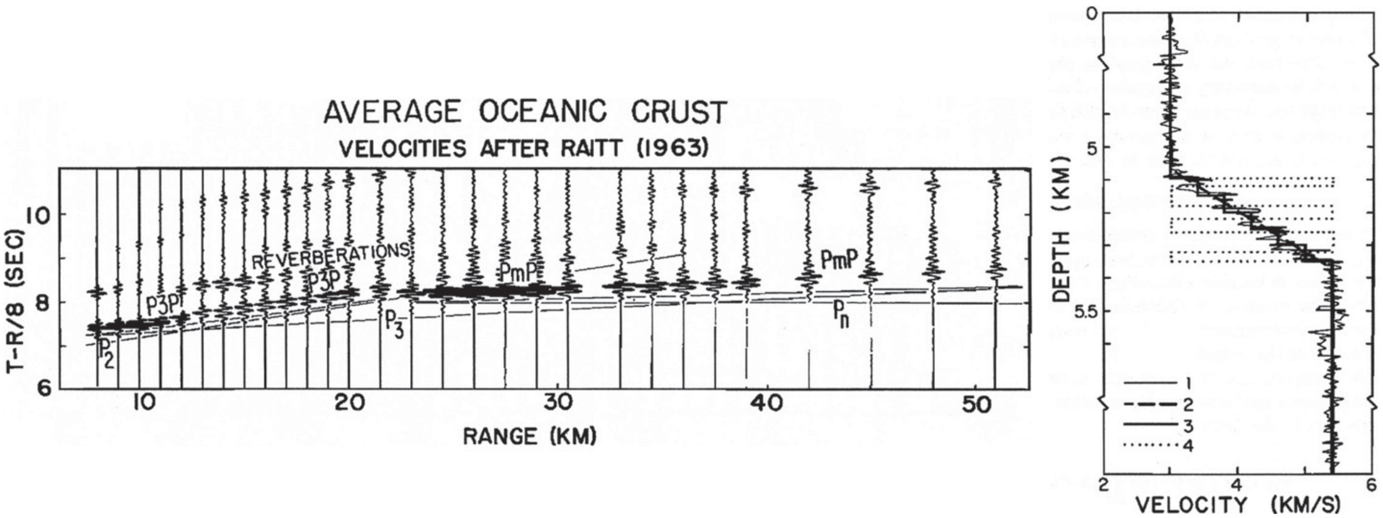


Figure 7.8.1-03. Left: Synthetic seismograms from the average oceanic crust model after Raitt (1963). P_2, P_3, P_n = head waves in layers 2, 3, mantle (discontinuities at 5 km (bottom of water column), 7 km (layer 2–3 border) and 13 km (Moho) depth). P3P, PmP = reflections from tops of layer 3 and mantle (from Spudich and Orcutt, 1980, fig. 6). Right: Four possible models, highlighting a gradient or lamellar transition above Moho (from Spudich and Orcutt, 1980, figs. 6 and 8). [Reviews of Geophysics and Space Physics, v. 18, p. 627–645. Reproduced by permission of American Geophysical Union.]

special deep seismic sounding projects in the Pacific Ocean had investigated the question of anisotropy in the subcrustal lithosphere (Keen and Barrett, 1971; Morris et al., 1969; Raitt et al., 1969, 1971), but they concentrated on possible anisotropy in the uppermost mantle. In the 1970s, also on land, evidence of anisotropy was detected in the uppermost mantle (Bamford et al., 1979; Fuchs, 1975, 1977, 1983), but only in areas of extensional tectonics. Investigations of anisotropy in the crust on land, however, were contradictory and did not allow researchers to convincingly conclude that anisotropy was present in the crystalline crust. Bibee and Shor (1976) investigated a large number of standard marine-refraction studies to determine the general relationship of crustal and mantle velocity to spreading direction and age in the Pacific. They concluded that anisotropy in the crust is insignificant, as is variation of lower crustal velocity with age. Mantle velocities, however, exhibited a high correlation with both age and azimuth, indicating an increase of velocity with age and ~5% anisotropy with highest velocity in the direction perpendicular to the local magnetic anomalies. The situation seemed to be different to Stephen and co-workers for the oceanic layers (Stephen et al., 1980; Stephen, 1981, 1985) and therefore in 1977, the specially designed Oblique Seismic Experiment was carried out in a borehole in the western Pacific 400 miles north of Puerto Rico. The experiment had been proposed to increase the usefulness of IPOD (International Phase of Oceanic Drilling) crustal holes as a means to investigate layer 2 of oceanic crust. However, the effect was quite small and it seems that for the usual deep seismic sounding experiments which were collected in our history, the possible neglecting of anisotropy in anyone of the crustal layers has not negatively influenced the interpretations of crustal structure and conclusions drawn from their results.

Christensen and co-workers studied rock samples from the Bay of Islands ophiolite complex of western Newfoundland (Salisbury and Christensen, 1978; Christensen and Salisbury, 1982) and from the Oman Mountains, where a continuous ophiolite succession is exposed (Christensen and Smewing, 1981). Based on their measurements of compressional and shear wave velocities to confining pressures of 6 kbar, they concluded that the seismic structure of the ophiolite sequences is remarkably similar to that of the oceanic crust and upper mantle. Kempner and Gettrust (1982a, 1982b) have calculated synthetic seismograms both for observed marine seismic data and for velocity profiles derived from laboratory measurements of rock samples from ophiolite sequences and proved that the composition of oceanic crust and upper mantle as derived from marine seismic data can directly be compared with exposed ophiolite sequences. A similar investigation was repeated in the 1980s by Collins et al. (1986).

Already in the 1960s and continuing in the 1970s, the main interest of oceanic crust and upper mantle research had shifted from the investigation of crustal and upper mantle structure under ocean basins to special structures with anomalous properties such as mid-ocean ridges, hotspots, ocean islands, and ocean-continent transition zones. Ewing and Meyer (1982) emphasized how the rapid evolvement of experimental and analytical techniques in marine seismology permitted a more flexible view of the variation of velocity with depth due to the employment of OBSs and due to the development of methods of analysis (e.g., Helmberger, 1968; Fuchs and Müller, 1971; Bessonova et al., 1974; Kennett, 1977). This greatly altered the conception of the structure of the oceanic crust and upper mantle and emphasized the need for detailed studies of specific areas and problems rather than well directed surveillance expeditions.

For example, in OBS refraction experiments in the 1970s on the northern East Pacific Rise, a fast-spreading ridge, a localized low-velocity zone was detected for the first time (Orcutt et al., 1976; Minshull, 2002) which was interpreted as resulting from the presence of a magma chamber containing partially molten rocks. Many seismic studies in the 1970s were concentrated on the Mid-Atlantic Ridge, a slow-spreading ridge. Early OBS refraction experiments on the Mid-Atlantic Ridge found neither a low-velocity zone corresponding to magma chamber nor a strong velocity contrast at Moho depths. Instead, the crust-mantle boundary was defined by a gradual increase in velocities (Fowler, 1976; Minshull, 2002). By the end of the 1960s, the traditional model of the hotspot Iceland consisted of four layers (Palmason, 1971): layer 0—unconsolidated material; layers 1 and 2—upper crust; layer 4—anomalous upper mantle with velocity of 7.4 km/s. At the end of the 1970s, a large-scale seismic experiment created a new model with a 15-km-thick two-layer crust and a mantle with gradual increase of velocity from 7.0 to 7.6 km/s at 50–60 km depth, with intermittent high-velocity sills near 30 km depth (Gebrande et al., 1980; see Fig. 7.8.4-08). Several aseismic ridges which are widespread in the oceans and oceanic plateaus representing vast outpourings of magma, but on a larger scale than aseismic ridges (Minshull, 2002) became the focus of seismic projects in the 1970s, such as the Iceland-Faeroe Ridge (Bott and Gunnarson, 1980), the Azores-Biscay Rise (Whitmarsh et al., 1982), or the Madagascar Ridge (Goslin et al., 1981; Sinha et al., 1981b) and the Agulhas Plateau off Cape Horn (Tucholke et al., 1981). Finally, continental shelves and adjacent oceanic areas were investigated in numerous projects along all continents.

An enormous amount of shallow seismic data was also collected in the 1970s in connection with the Deep Sea Drilling Project (DSDP). By 1968 the Joint Oceanographic Institutions for Deep Earth Sampling (JOIDES) had secured long-term funding from the National Science Foundation and leased a custom drillship from Global Marine Development Inc. named *GLOMAR Challenger*. Using powerful thrusters to stay in one spot even in heavy seas, the ship could lower a drill string through 7 km of water, then drill 1700 m and more into the sediment and rock of the seabed. For the next 15 years the *Challenger* crisscrossed the seas, making 96 separate voyages or "legs" and drilling at 624 sites on the seafloor. During that time, five other nations joined the program (Alden, 2010).

While the number of land projects of crustal and upper-mantle dimensions in a decade is limited due to the long precursory times of preparation, organization, and final fieldwork, marine projects, once ship time was available, could be achieved in much shorter time spans. Therefore, numerous oceanic research projects could be carried out in the 1970s and the following decades, many of them dealing exclusively with upper crustal studies. It would be beyond the scope of this history to mention them all. Therefore we have also not included the DSDP research work. Also, many projects were not published in easily accessible geophysical journals, but exclusively in special report volumes or journals specializing on general oceanic research topics other than geophysics and are therefore nearly impossible to find. We therefore restricted ourselves to a few and hopefully typical marine deep seismic projects in the following subchapters. Most projects we describe dealt with the crust and underlying uppermost mantle and were carried out in particular areas of tectonic unrest, such as along the mid-ocean ridges and along continental margins. Furthermore, similar to the continental efforts, some experiments aimed specifically to penetrate the subcrustal lithosphere by long-range seismic-refraction profiling. In the following description of active-source experiments carried out in the 1970s in the Pacific, Indian, and Atlantic Oceans, we will follow a more or less geographic order.

7.8.2. The Pacific Ocean

The knowledge on the crustal structure of the Pacific Ocean in the early 1970s was summarized, for example, by Woollard (1975), who had examined ~400 seismic-refraction measurements in the Pacific basin and presented them as block diagrams along selected lines (Fig. 7.8.2-01).

On a line through the northern Pacific from the Japan trench to the Gulf of Alaska (Fig. 7.8.2-02A), the crustal thickening beneath the Shatsky Rise was the dominant feature. On another line through the central Pacific from the Bismarck Archipelago to the Juan de Fuca Rise, the crust is 35–40 km thick underneath the Bismarck archipelago and adjacent Ontong-Java Rise, then thins to normal oceanic crustal thickness along the profile through the Gilbert islands, Darwin Rise, and Line Islands, then thickens substantially beneath the Hawaiian Islands, and thins again when traversing the Murray and Mendocino fracture

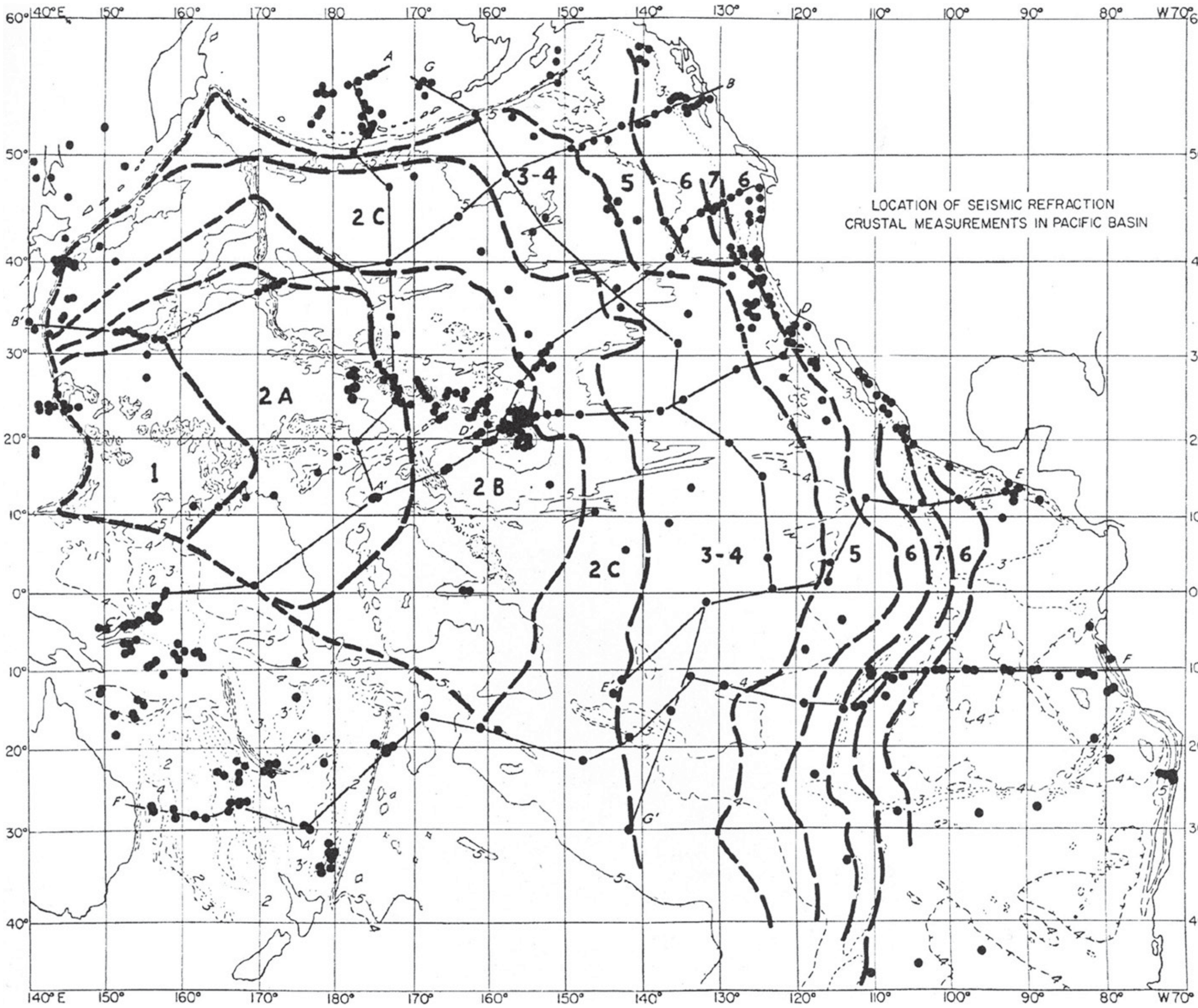


Figure 7.8.2-01. Location map of seismic refraction data measurements in the Pacific Ocean and location of profiles compiled by Woollard (from Woollard, 1975, fig. 2). [Reviews of Geophysics and Space Physics, v. 13, p. 87–137. Reproduced by permission of American Geophysical Union.]

zones as well as the Juan de Fuca Rise. A third line from the Queensland basin eastward (Fig. 7.8.2-02B) shows crustal thickening beneath the Lord Howe Rise (25 km), a slight thickening beneath the Tonga Trench and then a generally thin oceanic crust crossing the Tuamotu Islands and the East Pacific Rise until reaching the Peru-Chile trench.

Numerous investigations were performed by cruises of Soviet research vessels. In 1970 and 1971, the 49th cruise of the oceanographic research vessel *Vitiaz* covered a track length of 3500 miles of continuous reflection profiling using a 1-km-long multichannel streamer. The average penetration depth was 5–7 km; in some places up to 10 km. Deep seismic sounding observations were made along two polygons, which were located in the western part of the Pacific platform (151°W 29°S) and on the Eauripik Upland (Micronesia) near drill site 62 of the JOIDES Deep Sea Drilling Project.

At the first location—in cooperation with the U.S. research vessel *Mahi* from Honolulu, Hawaii—velocity anisotropy at the top of the mantle was investigated. At the second site—in cooperation with the Japanese vessel *Hokuku Maru*—subcritical reflections from deep boundaries were observed with three OBSs.

In the late 1970s, a large-scale investigation of the Pacific Ocean was undertaken by the Shirshov Institute of Oceanology (Neprochnov, 1989). From 1977 to 1979, basins and rises of the central Pacific were the goal of several cruises of their research vessel. Many deep seismic sounding profiles were recorded using OBSs and big airguns. Figure 7.8.2-03 shows numbers and locations, and Figure 7.8.2-04 shows averaged crustal columns.

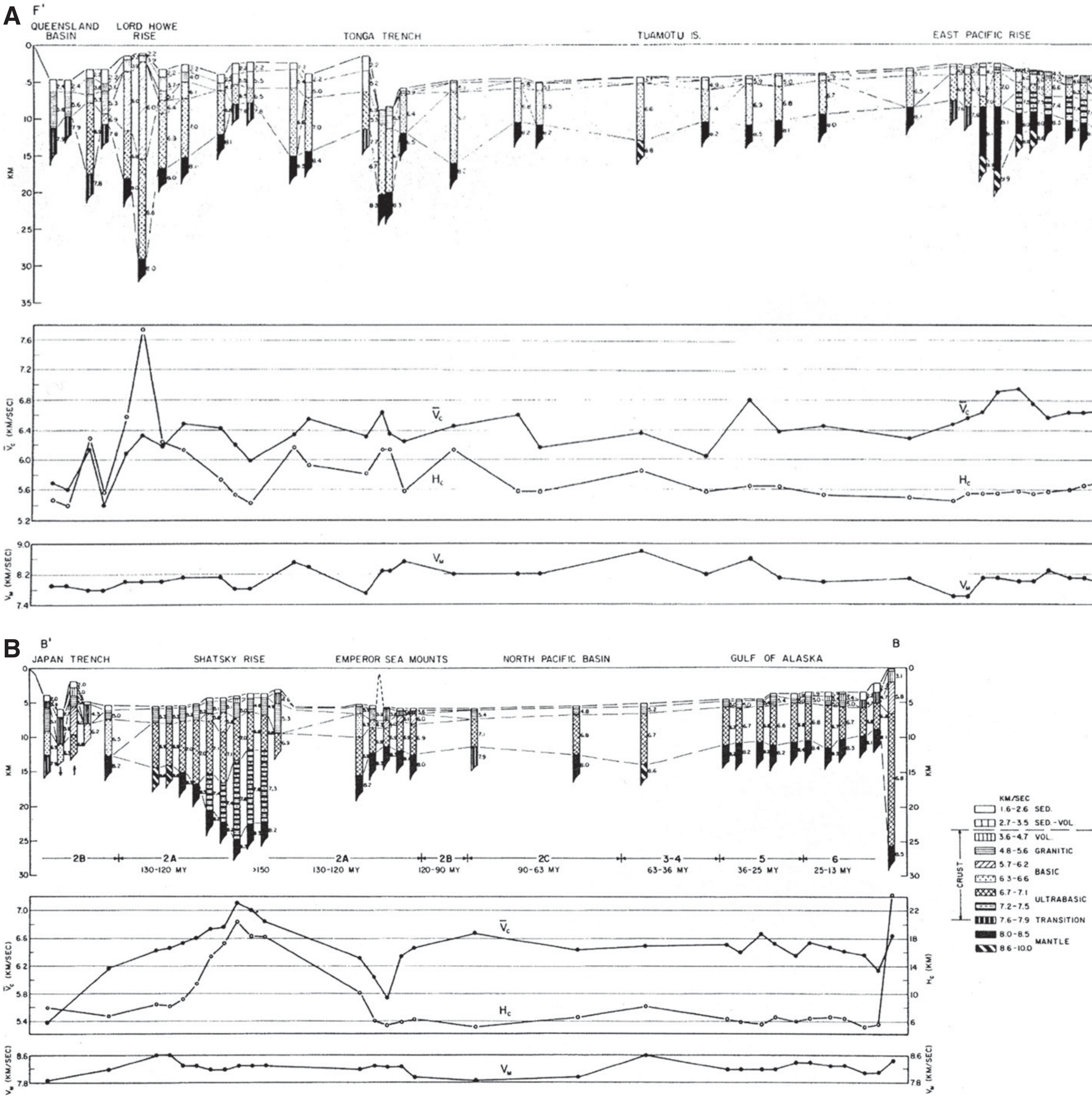


Figure 7.8.2-02. Crustal sections of seismic measurements through the Pacific Ocean compiled by G.P. Woollard. V_C—mean velocity of inferred crystalline rock crust, H_C—thickness of crystalline rock crust, V_M—upper mantle velocity. Geological identification of velocity values is based on laboratory studies of rocks (from Woollard, 1975, fig. 3). Section BB′ (A): WSW-ENE trans-Pacific section from Japan trench to Gulf of Alaska. Section FF′ (B): W-E trans-Pacific section from Queensland basin to Peru-Chile trench (from Woollard, 1975, fig. 5). [Reviews of Geophysics and Space Physics, v. 13, p. 87–137. Reproduced by permission of American Geophysical Union.]

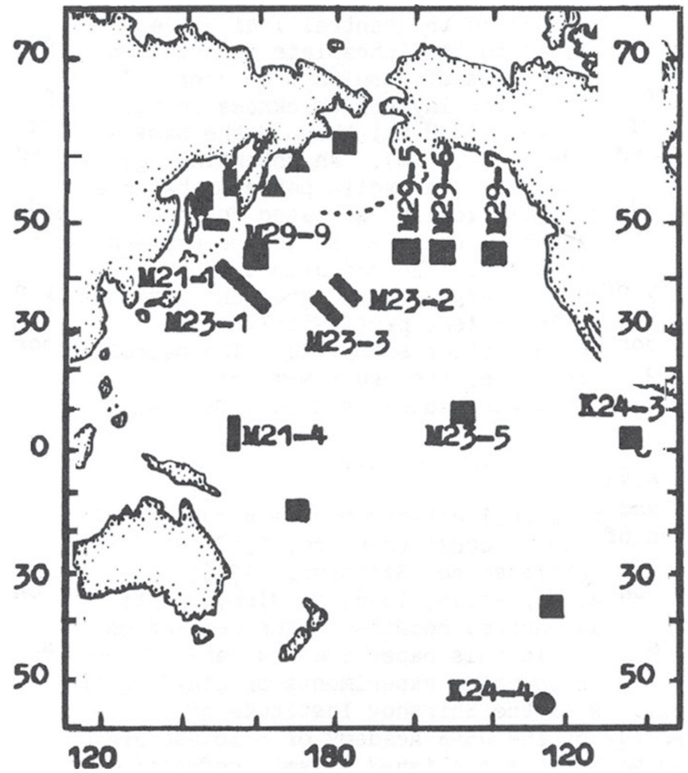


Figure 7.8.2-03. Location of seismic surveys during the various cruises of R/V *Kana Keoki* and R/V *Dmitry Mendeleev* from 1977 to 1979 (from Neprochnov, 1989, fig. 2). [*In* Mereu, R.F., Mueller, St., and Fountain, D.M., eds., Properties and processes of earth's lower crust: American Geophysical Union, Geophysical Monograph 51, p. 159–168. Reproduced by permission of American Geophysical Union.]

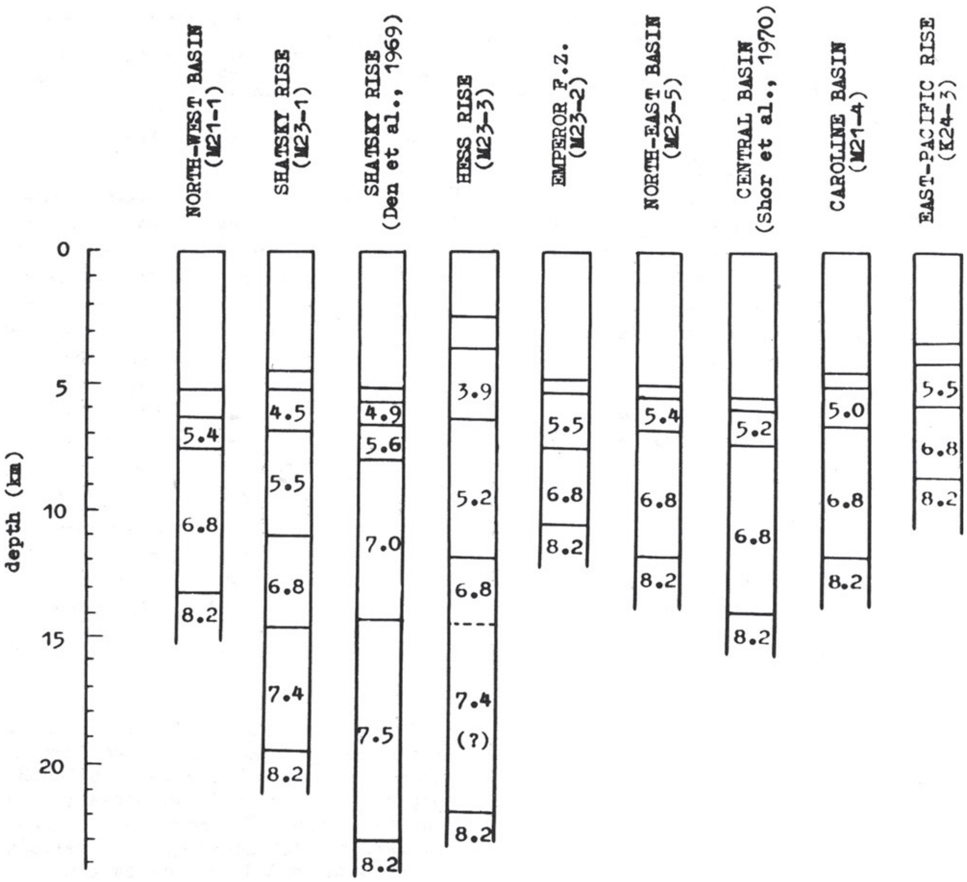


Figure 7.8.2-04. Averaged crustal columns for the Pacific region (from Neprochnov, 1989, fig. 3). [*In* Mereu, R.F., Mueller, St., and Fountain, D.M., eds., Properties and processes of earth's lower crust: American Geophysical Union, Geophysical Monograph 51, p. 159–168. Reproduced by permission of American Geophysical Union.]

In the following, we have selected a few projects to demonstrate the wealth of new data which became available with advanced technology gaining in particular from the development of OBSs and their gradual improvement both in quality and quantity. The majority of these marine experiments aimed for selected problems in tectonically anomalous areas.

In 1976, a marine geophysical survey was conducted in the region of the Explorer spreading center located west of Vancouver Island on Canada's west coast (Cheung and Clowes, 1981). It was an 80-km-long seismic-refraction line using an array of three OBSs and was located northwest of the northern tip of Explorer Ridge and parallel to Revere-Dellwood Fracture Zone in the northeast Pacific Ocean (Fig. 7.8.2-05). Eighty-one explosive charges, alternating large and small shots, were fired, resulting in one reversed profile and two split-spread profiles which overlapped the longer profiles. In addition, airgun shots were fired at 2 minute intervals to run 30-km profiles centered over each OBS in direction and across the explosion line, to add more detailed upper-crustal information. Figure 7.8.2-06 presents the record section obtained from the recordings of OBS1 (see also Appendix A7-10-1). Basically, a normal crustal thickness of 6.5 km resulted, but with an anomalously low P_n velocity of 7.4 km/s. A particular result was the absence of distinct structural discontinuities, even at the Moho.

During June 1974, a seismic-refraction and gravity survey of the East Pacific Rise was carried out near the Siqueiros Fracture Zone at 8°N. Three OBSs recorded shots along two N-S directed profiles west of (crossing the Fracture Zone at the 2.9–5 m.y. boundary) and along the crest of the East Pacific Rise (Orcutt et al., 1976). The charge sizes varied from 0.45 to 436 kg of TOVEX explosives. For the profile along the crest of the East Pacific Rise, the P-velocity increased quite

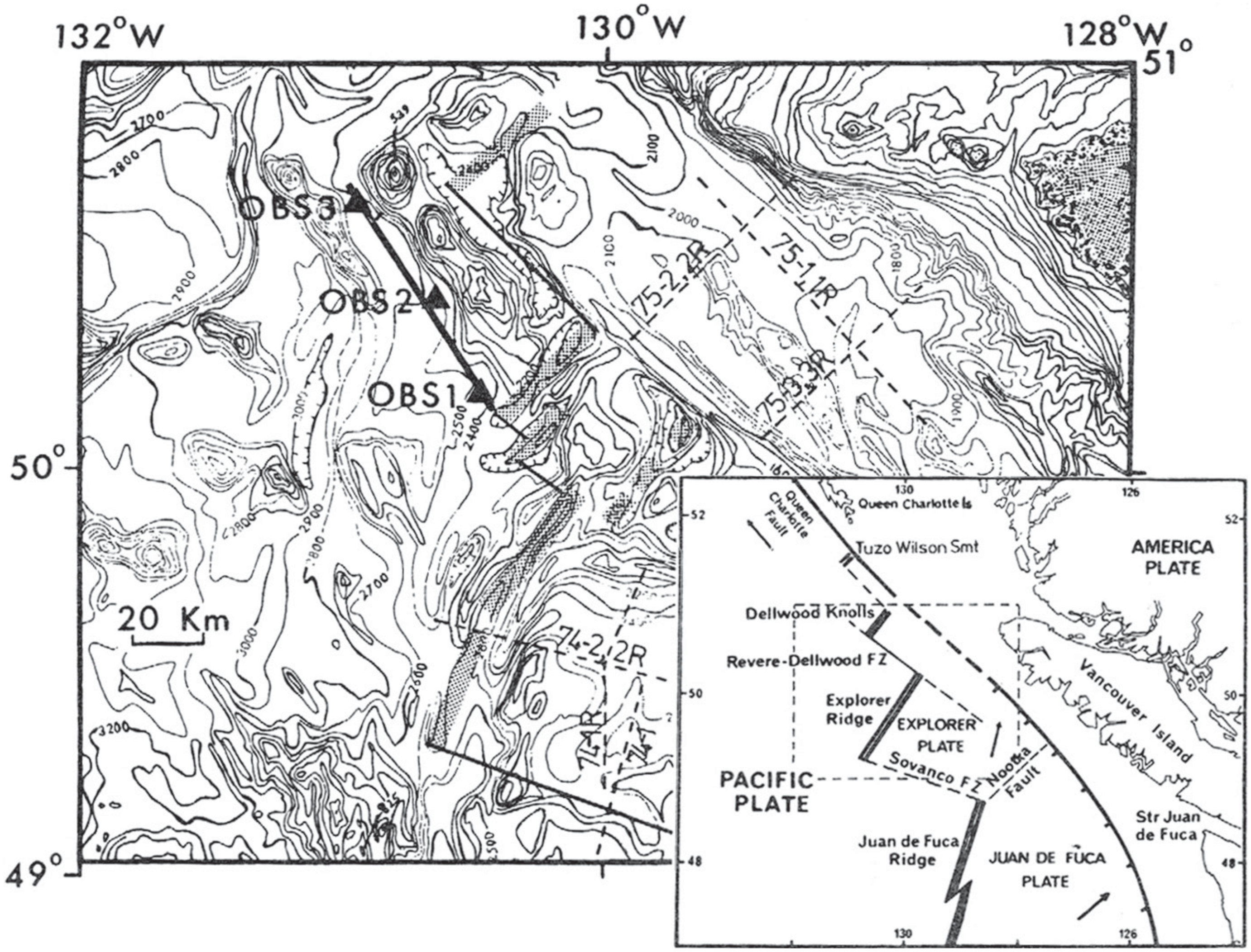


Figure 7.8.2-05. Location map of the seismic investigation of the Explorer Ridge and Revere-Dellwood Fracture Zone of the Northeast Pacific (from Cheung and Clowes, 1981, fig. 1). [Geophysical Journal of the Royal Astronomical Society, v. 65, p. 47–73. Copyright John Wiley & Sons Ltd.]

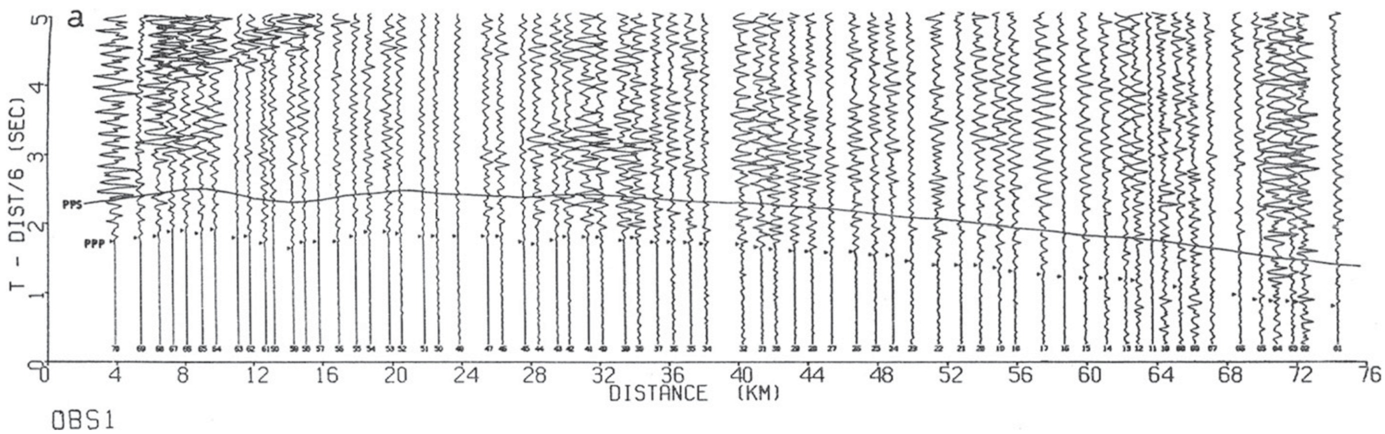


Figure 7.8.2-06. Record section of OBS1 of the Northeast Pacific experiment (from Cheung and Clowes, 1981, fig. 7). [Geophysical Journal of the Royal Astronomical Society, v. 65, p. 47–73. Copyright John Wiley & Sons Ltd.]

rapidly from 5 km/s at the seafloor to 6.7 km/s at the lid of a low-velocity zone. Mantle arrivals from below this zone indicated low P-velocities of 7.6 km/s. A second profile on 2.9 m.y. old crust showed a shallow region of strong velocity gradients which graded into velocities typical of the "oceanic" layer without any clear stratification and again a low mantle velocity of 7.6 km/s. The third profile along 5 m.y. old crust showed more distinct layering, but also velocity gradients within the layers. Here the velocities were more typical for Pacific refraction profiles and a mantle velocity of 8 km/s resulted. Orcutt et al. (1976) predicted that the reflector at 2 km depth beneath the crest of the East Pacific Rise might correspond to the top of a low-velocity magma chamber.

This was also postulated by Rosendahl et al. (1976), who from the same two-ship sonobuoy refraction study in this area concluded that this low-velocity zone is thickest under the summit of the rise, and then thins outward. Radical changes in the structure of the crust would occur within a few million years in this region, and stratification would become more pronounced with age.

In 1976, Lamont-Doherty Geological Observatory performed another investigation of the East Pacific Rise near the Siqueiros Fracture Zone near 9°N (Herron et al., 1978). They acquired ~900 km of digital 24-channel seismic-reflection data with large airguns on three ships crossings of the East Pacific Rise. The multichannel profiles were made aboard the vessel *Conrad* through the use of a 24-group hydrophone streamer, 2400 m long, and four airguns towed behind the ship. The airguns were fired every 50 m. The interpretation revealed three distinct crustal layers emerging as coherent events across the rise crest. The base of the third layer appeared at a depth of 2 km below the seafloor and was interpreted as possible top of a low-velocity magma chamber, as postulated by Orcutt et al. (1976). This layer seemed to thicken away from the rise crest toward the edge of the raised axial block that characterizes the East Pacific Rise according to Anderson and Noltimier (1973).

From a study of the northern Cocos plate, between the East Pacific Rise and the Middle America trench at 13°–15° N latitude, Lewis and Snydsman (1979) concluded that the crust thickens systematically with increasing age, caused by a gradual transformation of the top 2 km of the upper mantle into material having crustal velocities. The corresponding velocity-depth cross section showed a more or less constant crustal body of 5 km thickness slightly dipping toward the east underlain by a transformed mantle wedge changing gradually from 0 km thickness at 100 km distance from the rise axis, where Moho depth is 8 km, to 2 km thickness at ~600 km distance, resulting in a Moho depth of 10.5 km. The experiment was conducted in 1973 and 1974 on the Cocos Plate between the Orozco and Clipperton fracture zones and extended from the East Pacific Rise to the Middle America Trench (12.5°–15°N, 99°–104.5°W). Most of the data were obtained on five long-range telemetering buoys which were generally deployed as free-floating units ~2 km long. In addition, OBS were deployed twice, first near the trench and later near the rise axis.

In 1975, Orcutt and Dorman (1977) performed a long-range profile in the Pacific Ocean. The line was 600 km long (Fig. 7.8.2-07 and Appendix A7-10-1) and was laid out in the northeastern Pacific between the Clarion and Molokai fracture zones. In contrast to similar long-range observations in the western Pacific (see below), Orcutt and Dorman (1977) could not establish any velocities higher than 8.4 km/s down to 60 km depth, but they indicated the possible existence of a velocity inversion, with no less than 8.0 km/s, at around 50 km depth, similarly as had been detected by continental long-range investigations in western Europe (e.g., Hirn et al., 1973, 1975; see above).

In 1979, the Rivera Ocean Seismic Experiment (ROSE; Fig. 7.8.2-08) investigated the crust on the Orozco Fracture Zone of the East Pacific Rise at ~12° N (Ewing and Meyer, 1982; Lewis and Garmany, 1982). It was intended to become a combined sea and land seismic program to study a number of features of the structure and evolution of a mid-ocean ridge, a major oceanic

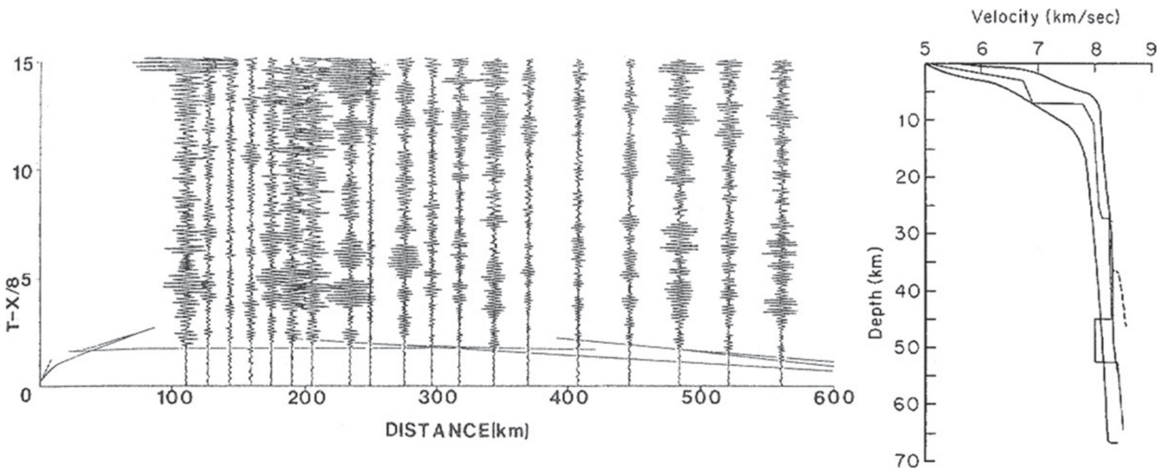


Figure 7.8.2-07. Left: Record section of a long-range seismic refraction profile in the Pacific Ocean (from Orcutt and Dorman, 1977, fig. 5). Right: Velocity-depth function derived from the long-range seismic observations on the Pacific Ocean (from Orcutt and Dorman, 1977, fig. 4). [Journal of Geophysics, v. 43, p. 257–263. Reproduced with kind permission of Springer Science+Business Media.]

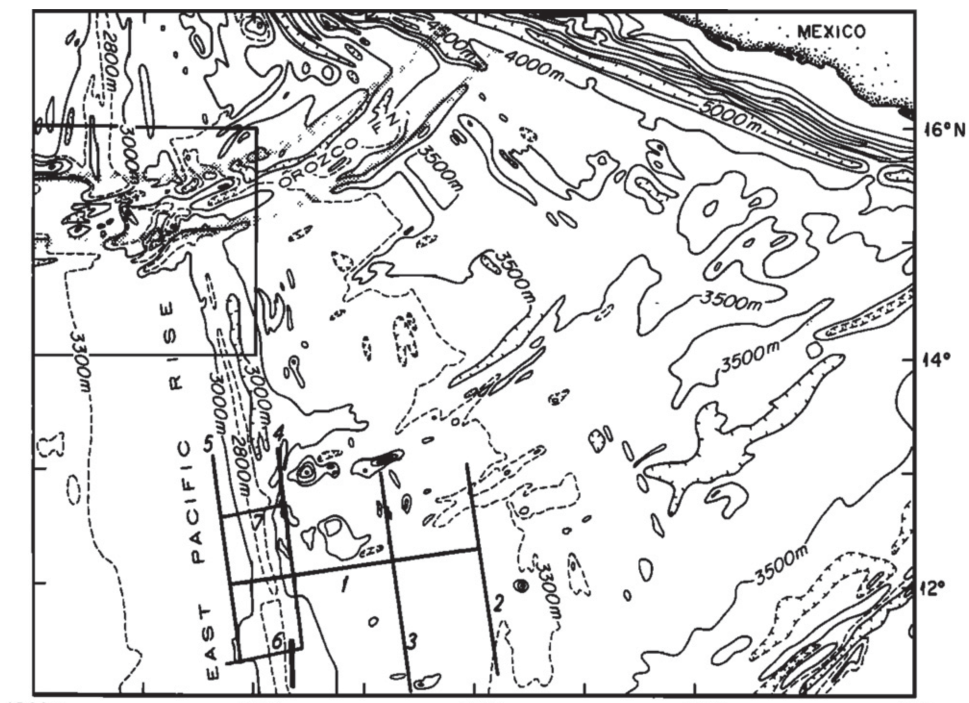


Figure 7.8.2-08. Location map of the seismic Rivera Ocean Seismic Experiment (ROSE) centered on the Orozco Fracture Zone of the East Pacific Rise (from Ewing and Meyer, 1982, fig. 1). [Journal of Geophysical Research, v. 87, p. 8345–8357. Reproduced by permission of American Geophysical Union.]

fracture zone, and the transition region between ocean and continent. Due to the lack of a permit to work within Mexico's 200-mile limit, the project had to be separated into three independent parts: an experiment to study the East Pacific Rise south of the Orozco Fracture Zone primarily using ocean bottom recording and explosive sources, a seismicity program at the Orozco, and a land-based program of recording natural events along the coastal region of Mexico. Due to the need to shift the marine part away from Mexican waters, only one strong event, the 7.6 Petatlan earthquake of 14 March 1979, at 17.5°N, 101.5°W, and some of its fore- and aftershocks could be recorded jointly by both land and marine instruments.

As part of the project, a series of multichannel seismic-reflection profiles were run, and along these lines, military sonobuoys were routinely deployed and recorded on a digital seismic recording system along the reflection profiling. The seismic energy was produced by an array of airguns fired every 20 s (ca. 50 m). These typically provided seismogram data to a range of ~20 km from which crustal structure to a subseafloor depth of 2.5 km could be deduced. In addition to the sonobuoy data sets, three expanded spread profiles were collected. In these experiments involving three ships, which were alternatively shooting or receiving, explosive charges were fired on a 5 minute schedule (ca. 750 m). Data were recorded up to 50 km distance allowing researchers to deduce crustal structure to a depth of 5 km below the seafloor. The data were analyzed separately for details of the upper crust (Ewing and Purdy, 1982; Purdy, 1982) and for variations in crust and upper mantle structure (Gettrust et al., 1982). A detailed reinterpretation followed in 1988 (Vera and Mutter, 1988). In general, the overall structure of the crust consists of a rapid increase in velocity from ~2.5–6 km/s within the uppermost 1 km of crust (layer 2A), a zone of maximum 1 km thickness with little or no velocity gradient (layer 2B), a thin transition layer with a relatively rapid increase of velocity from ~6.2–7 km/s (layer 2C), and a thick nearly homogeneous basal layer with velocities around 7.2 km/s (layer 3). The reanalysis of these data (Vera and Mutter, 1988) allowed researchers to visualize aspects of structural variability of the lower oceanic crust which were not apparent in earlier interpretations (e.g., Gettrust et al., 1982).

Spudich and Orcutt (1980) presented a new picture of the oceanic crust based on progress in instrumentation and experimental design, as well as on advances in mathematical and computational seismology. In their new picture, layer 2 appeared as a region where velocity increases rapidly with depth. Layer 3 seemed more homogeneous in the vertical direction than layer 2, with gentle velocity gradients and occasional velocity inversions. Although it was observed at several sites, they doubted the widespread existence of a high-velocity basal crustal layer. The seismic picture of the crust appeared in good agreement with seismic velocities of rocks gained from direct drilling and laboratory velocity studies on both oceanic and ophiolitic samples.

An international conference, held in 1974 in Honolulu, Hawaii, to honor G.P. Woollard, led to the publication of a special volume on the geophysics of the Pacific Ocean basin and its margins (Sutton et al., 1976). The papers presented there dealt in particular with the continent-ocean transition which, both in the east under South America and in the west under Japan, appeared as the subduction of an oceanic plate underneath a continent.

In the South Pacific, a major project dealing with the subduction of the Pacific Ocean under the Colombian Andes of the South American continent was Narino III, an onshore-offshore experiment launched in 1973 between latitudes 2° and 4°N (Meyer et al., 1976). Another land-sea operation across the coast of Colombia followed in 1978 between the latitudes 5° and 6°N (Flueh et al., 1981). Details of both projects were discussed above in subchapter 7.7.3 on South America.

Hussong et al. (1976) dealt with the crustal structure of the Peru-Chile trench. Five standard two-ship reversed explosion lines defined the crustal velocity structure, and also served as end points for a 375-km-long refraction line comprised of 11 extended overlapping split profiles. Furthermore, a commercial reflection line was available, and twelve Airgun Sonobuoy Precision Echo Recorders (ASPER) supplemented the survey carried out from 1972 to 1973 (Fig. 7.8.1-09). Typical Moho depths near 10 km were obtained west of the trench, while under the continental shelf a crustal thickness of 30 km was found. The data were used by the authors to develop a complex model of the convergence of the Nazca plate with the South American continental margin (Fig. 7.8.1-10), with the Nazca plate appearing with higher velocity crust than typically found for other oceanic plates.

Two crustal investigations dealt with the Hawaiian archipelago, where the island of Hawaii with its active volcano Kilauea was investigated in detail. In 1976 and 1978, the U.S. Geological Survey established two seismic-refraction profiles on and off the Hawaii Islands with offshore shots of 100 km length each (Fig. 7.8.2-11). The first profile was directed southeastward normal to

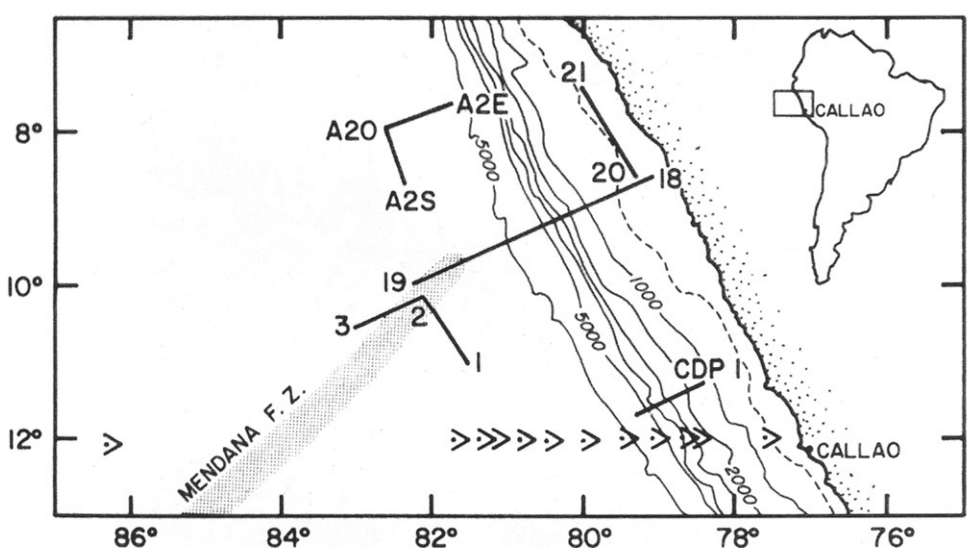


Figure 7.8.2-09. Location of seismic profiles around the Peru-Chile trench between 8° and 12°S latitude (from Hussong et al., 1976, fig. 1). The dot and arrow symbols at 12°S latitude assign the ASPER stations. [*In* Sutton, G.H., Manghnani, M.H., Moberly, R., and McAffe, E.U., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union, Geophysical Monograph 19, p. 71–85. Reproduced by permission of American Geophysical Union.]

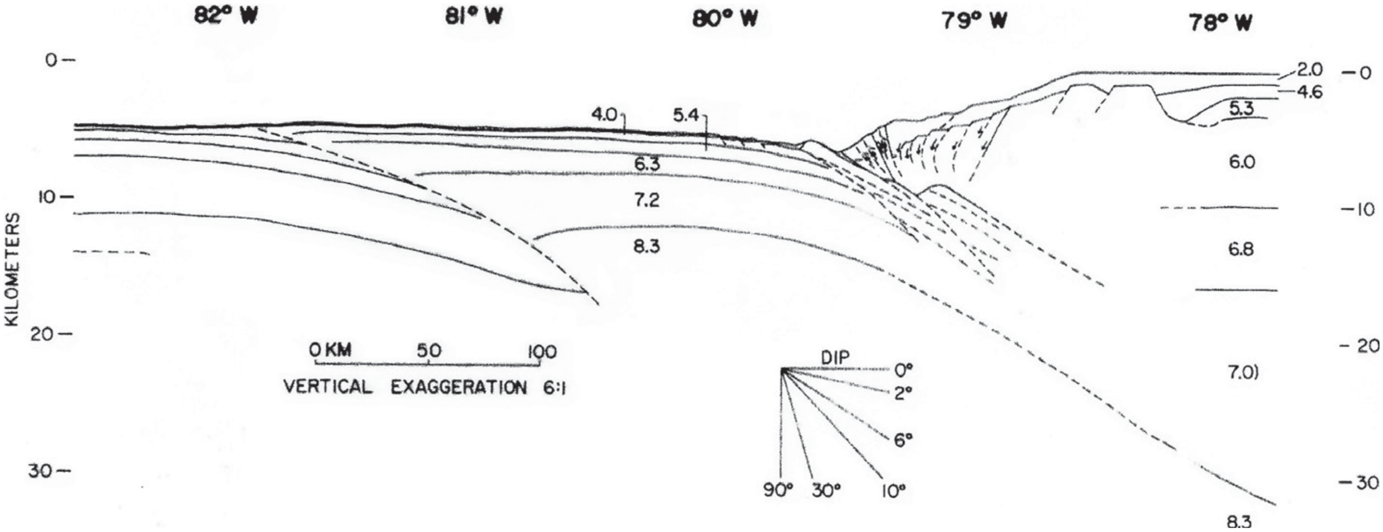


Figure 7.8.2-10. Composite crustal section across the Peru-Chile trench (from Hussong et al., 1976, fig. 12). [*In* Sutton, G.H., Manghnani, M.H., Moberly, R., and McAffe, E.U., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union, Geophysical Monograph 19, p. 71–85. Reproduced by permission of American Geophysical Union.]

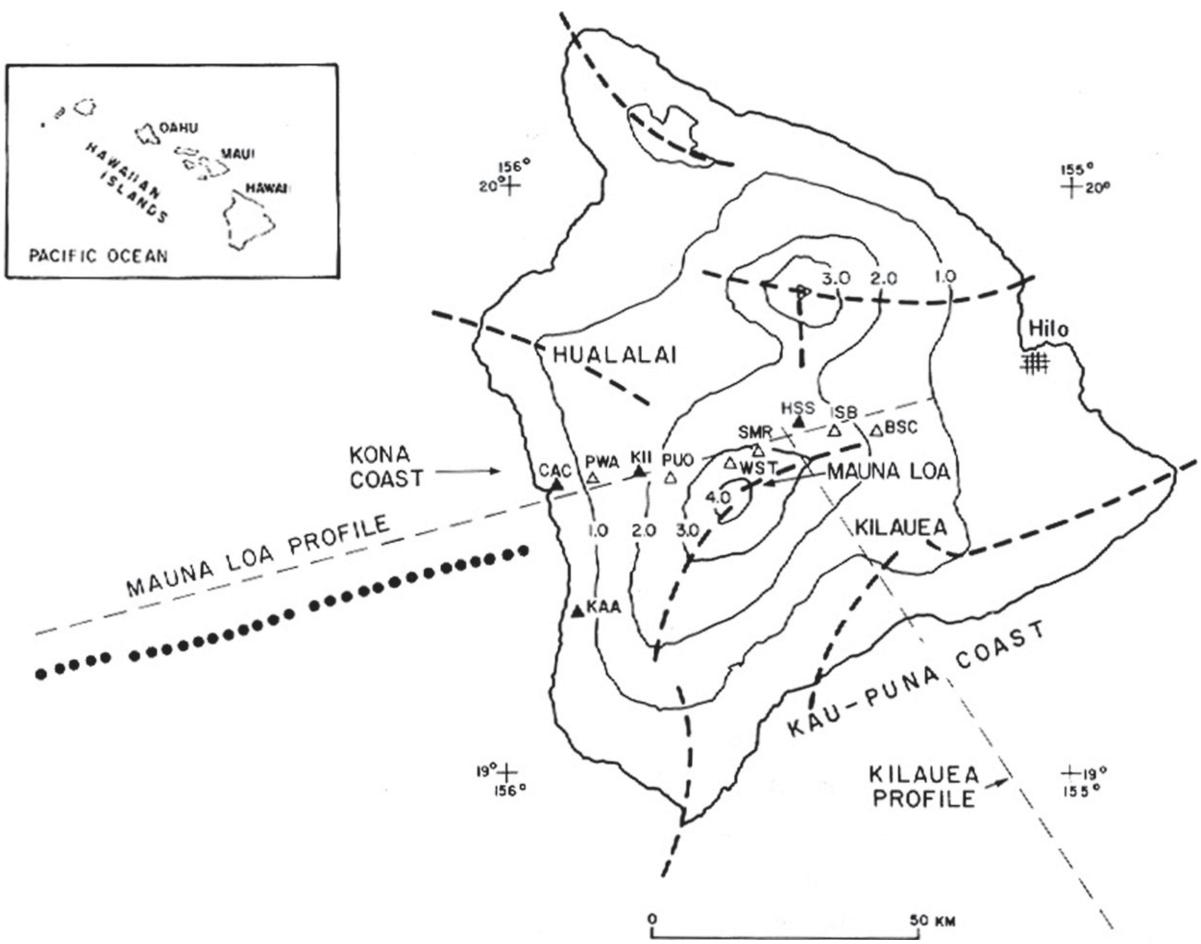


Figure 7.8.2-11. Location of the 1976 Kilauea and 1978 Mauna Loa profiles (from Zucca et al., 1982, fig. 1). Dots and triangles are shots at stations of the 1978 experiment. [Bulletin of the Seismological Society of America, v. 72, p. 1535–1550. Reproduced by permission of the Seismological Society of America.]

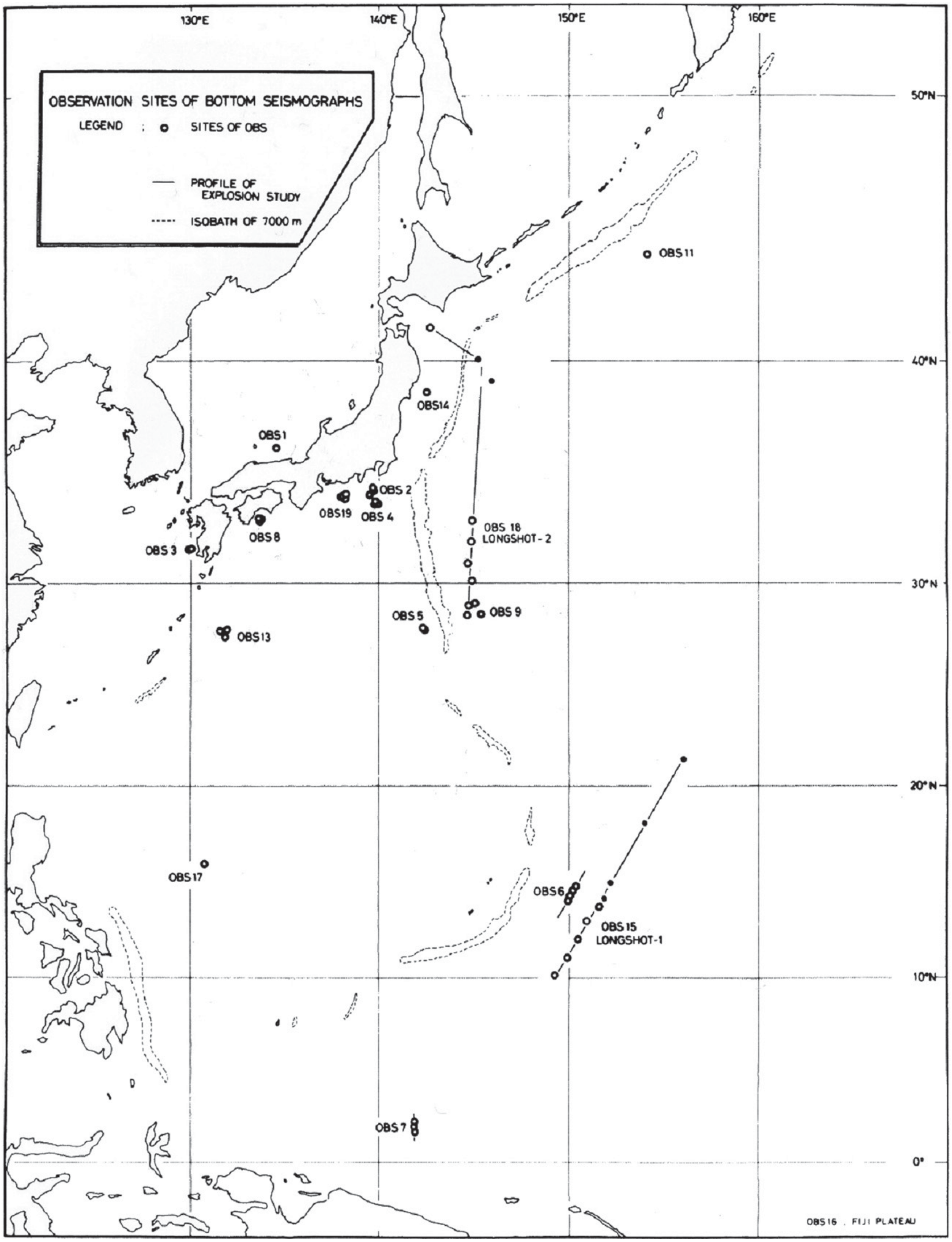


Figure 7.8.2-12. Observation sites of ocean-bottom seismographs and shot sites of the longshot experiments (from Asada and Shimamura, 1976, fig. 1). [*In* Sutton, G.H., Manghnani, M.H., Moberly, R., and McAffe, E.U., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union, Geophysical Monograph 19, p. 135–154. Reproduced by permission of American Geophysical Union.]

the southeast coast across the submarine flank of Kilauea volcano (Zucca and Hill, 1980), with 43 explosive charges oriented along an offshore line roughly 100 km long and oriented normal to the southeast flank of Kilauea.(Kilauea profile in Fig. 7.8.2-11).

The second profile (Mauna Loa profile in Fig. 7.8.2-11) was directed toward WSW normal to the Kona coast and contained 30 explosive charges evenly distributed along the 100-km-long offshore part (Zucca et al., 1982). Both profiles were recorded by the permanent seismic network, by portable 5-day recorders, and by OBSs (for details, see Appendix A7-10-2).

The objectives were to provide an improved resolution of the structure of the Mauna Loa and Kilauea volcanoes with emphasis on the nature of the transition from the "normal" oceanic crust to the massive volcanic pile. On both profiles a similar pattern

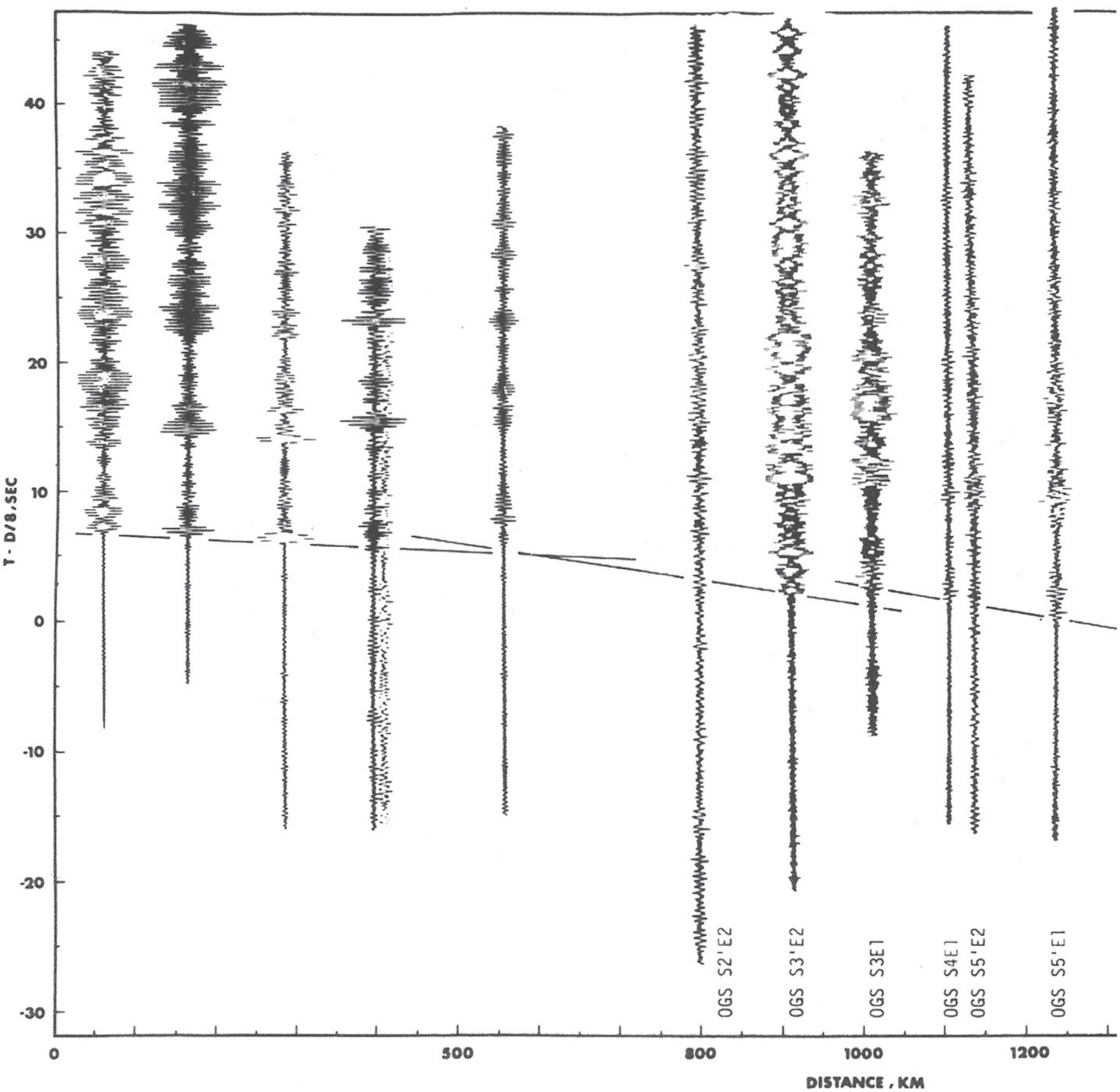


Figure 7.8.2-13. Record section of the longshot experiment (from Asada and Shimamura, 1976, fig. 9). [*In* Sutton, G.H., Manghnani, M.H., Moberly, R., and McAffe, E.U., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union, Geophysical Monograph 19, p. 135–154. Reproduced by permission of American Geophysical Union.]

could be seen: The oceanic crust begins to thicken landwards at the break of the slope between the volcanic pile and the ancient sea floor. From here, the Moho dips gently toward the land. At the coast, the Moho is 14 km deep and it reaches a maximum depth of 17–18.5 km under the center of the volcanoes.

Japanese research activities in the 1970s involved large offshore shot experiments in the western Pacific Ocean and in the Sea of Japan, carried out within the International Geodynamic Project. The aim was to elucidate lateral velocity variations in the upper mantle from the trench area to the back arc basin (Research Group for Explosion Seismology, 1975; Okada et al., 1978, 1979; Yoshii et al., 1981).

Two Longshot experiments were carried out in 1973 and in 1974 in the western Pacific basin. They aimed mainly for the structure of the subcrustal lithosphere and less for details of crustal structure (Asada and Shimamura, 1976, 1979).

For their seismic investigation of the structure of the western Pacific and its relation to Japan, Asada and Shimamura (1976, 1979) used large underwater explosions which were recorded by OBSs up to distances of 1300 km (Figs. 7.8.2-12 and 7.8.2-13). Since the ambient noise in deep oceanic basins turned out to be greater in the optimum frequency range (below 10Hz), 1.5-ton shots proved to be too small for obtaining clear signals, but 5- to 7-ton shots proved to be adequate for the Longshot experiments. The ocean-bottom seismographs, being used by the Japanese institutions since 1969, were designed to be self-contained in a small capsule and to withstand a depth of 9000 m.

Their preferred model 115 (Fig. 7.8.1-14) shows, underneath a 17-km-thick crust, a structure of the subcrustal lithosphere, similar to models obtained for long-range profiles in western Europe. From 1974 to 1980, further intensive long range experiments were undertaken in the northwestern Pacific, which detected large-scale lithospheric structure and anisotropy in the upper mantle (Asada and Shimamura, 1979; Asada et al., 1983; Shimamura et al., 1983).

Two other long-range explosion experiments using OBSs were carried out in the East Mariana Basin in 1973 and 1976 (Fig. 7.8.2-15). Twenty-five OBS stations were deployed along

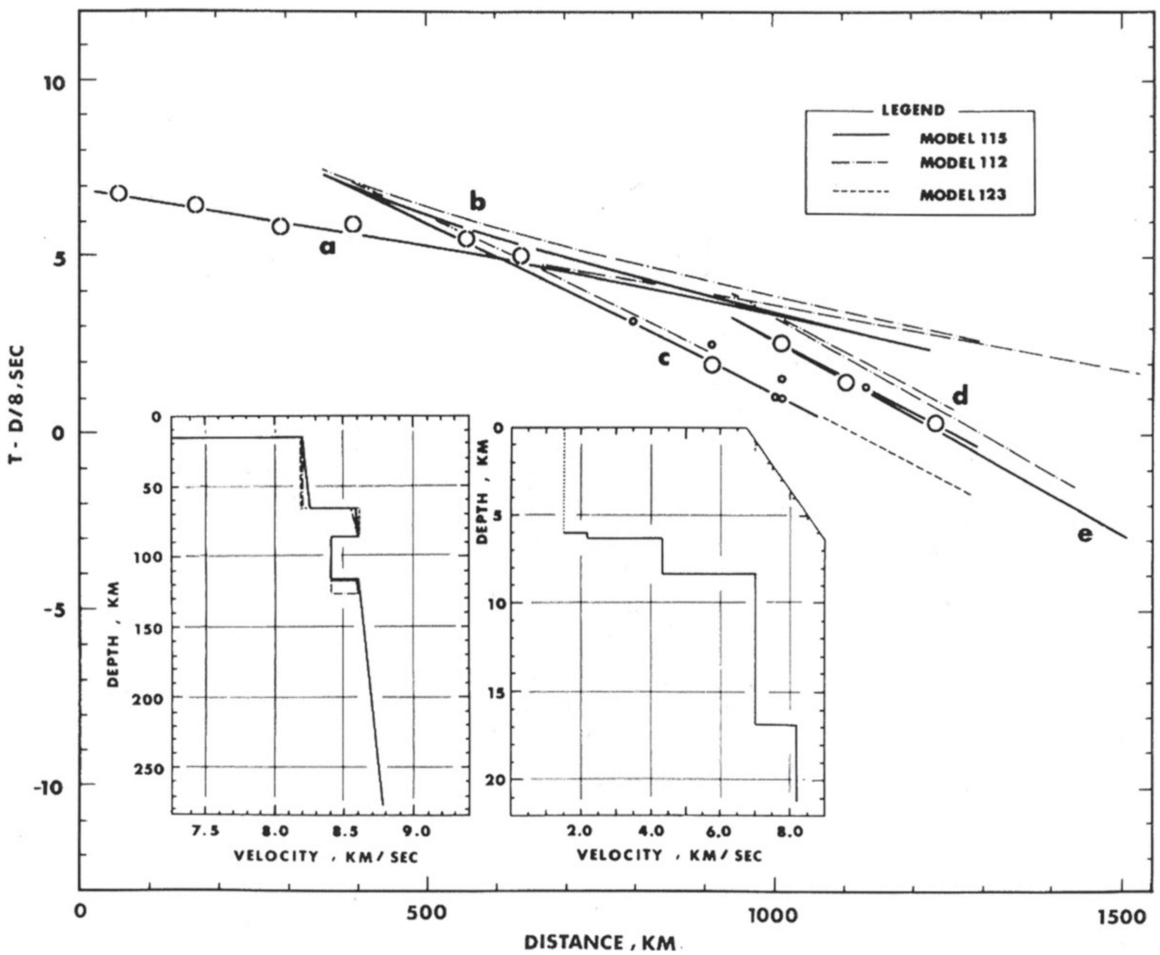


Figure 7.8.2-14. Reduced travel time curves of observed and calculated values for models shown below (from Asada and Shimamura, 1976, fig. 9). [*In* Sutton, G.H., Manghnani, M.H., Moberly, R., and McAffe, E.U., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union, Geophysical Monograph 19, p. 135–154. Reproduced by permission of American Geophysical Union.]

an 800-km-long line and seven large (1.5–8.5 tons) shots and several tens of small shots were fired. The maximum recording range was 1900 km.

The interpretation of the data indicated significant differences in the structure of the lower lithosphere. The attenuation of seismic waves appeared to be larger than east of Honshu and the low-velocity layer below the Moho seemed more accentuated. Nagumo et al. (1981) confirmed the result that the sub-Moho velocity layer, with 8.4 km/s velocity, existed, but it appeared to be extremely thin. Alternating high-velocity and low-velocity layers down to 90 km depth were modeled. Possible models for the upper low-velocity zone and the corresponding data are shown in Figure 7.8.2-16 (see also Appendix A7-10-1).

In the southwestern Pacific in 1973, the seismic investigation of the major structures of the Papua Ultramafic Belt, which had started in the 1960s, was continued. The ophiolite suite of rocks was widely considered to be a classic example of fragment of oceanic crust and upper mantle, obducted on to a continental mass during orogenesis in an area of continent/island arc convergence (e.g., Moores, 1973).

The intensive seismic investigation of the Papuan peninsula and its adjacent seas consisted of offshore-onshore operations in the Gulf of Papua and both the Coral Sea and Solomon Sea with recording distances up to 300 km and more (Drummond et al., 1979; Finlayson et al., 1976, 1977; Fig. 7.8.2-18; Appendix A7-10-3). Forty-two seismic stations were deployed in Papua New Guinea, and 111 shots were fired from the surrounding seas (Fig. 7.8.2-17). The shots had charges of 180 kg or 1000 kg and were detonated 100 m below the water surface, wherever water depth permitted. During the 1973 survey, a nuclear explosion on

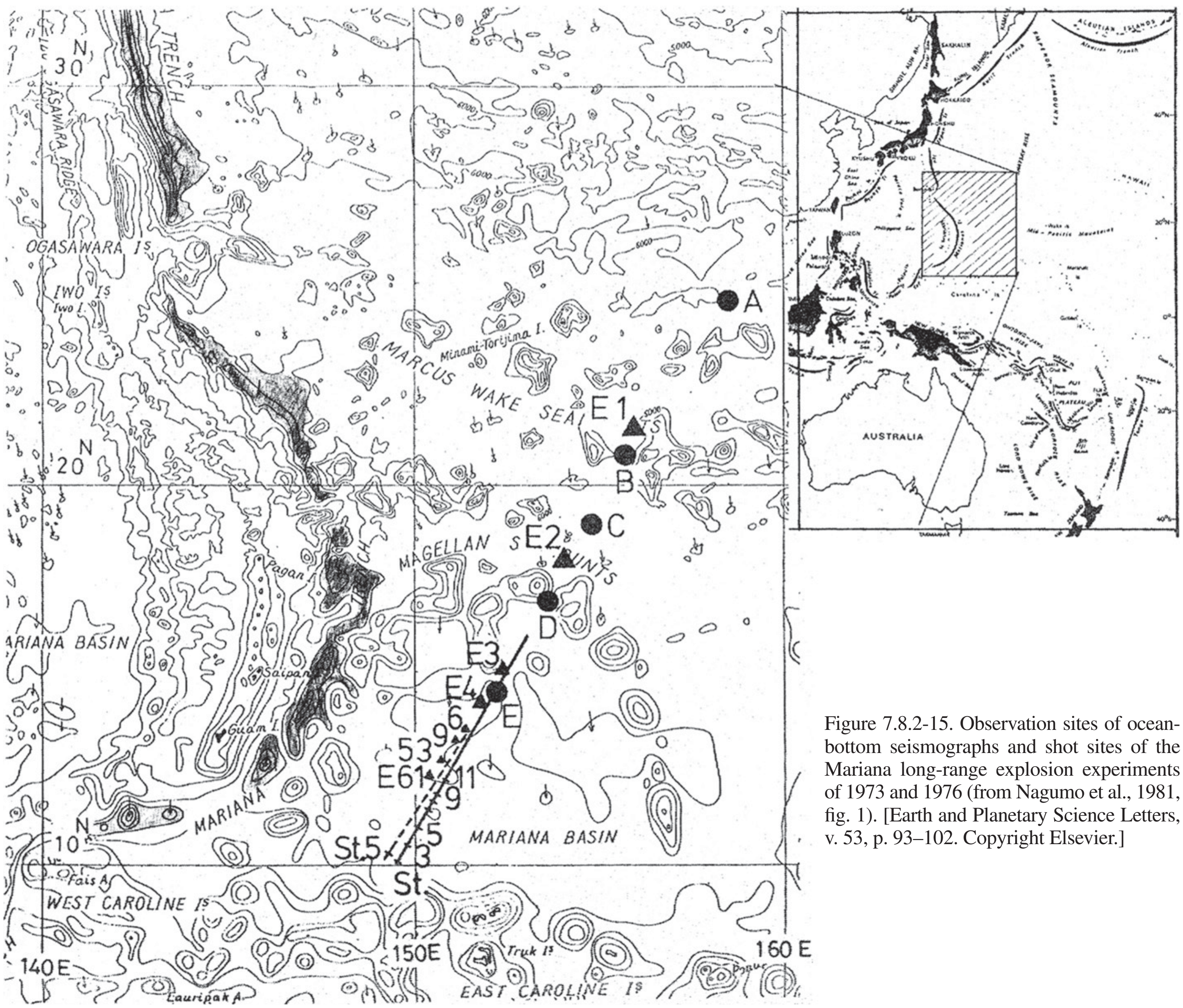


Figure 7.8.2-15. Observation sites of ocean-bottom seismographs and shot sites of the Mariana long-range explosion experiments of 1973 and 1976 (from Nagumo et al., 1981, fig. 1). [Earth and Planetary Science Letters, v. 53, p. 93–102. Copyright Elsevier.]

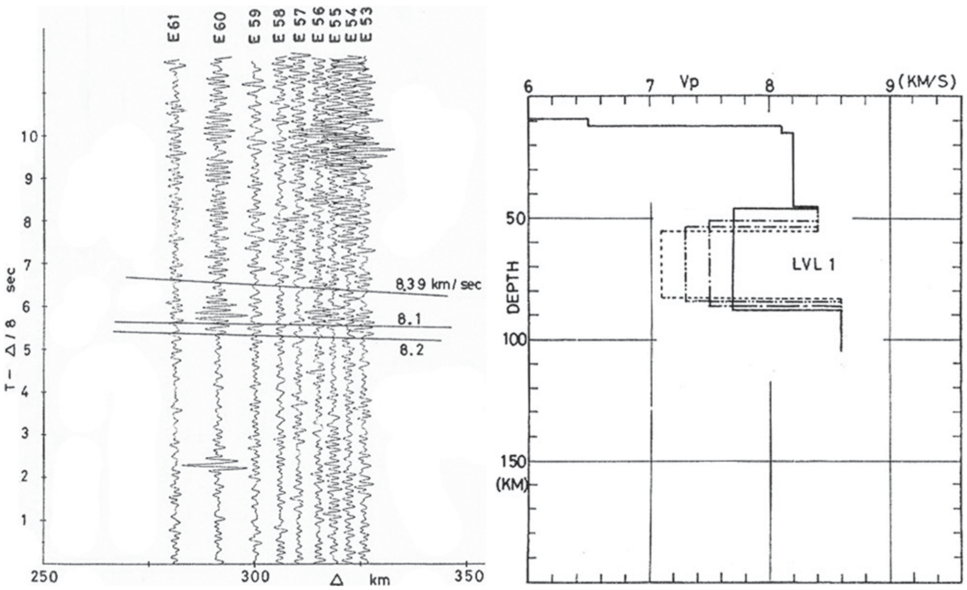


Figure 7.8.2-16. Part of record section (left) and models for upper low-velocity zone in the subcrustal lithosphere (right) of the Mariana long-range explosion experiments (from Nagumo et al, 1981, figs. 3 and 4a). [Earth and Planetary Science Letters, v. 53, p. 93–102. Copyright Elsevier.]

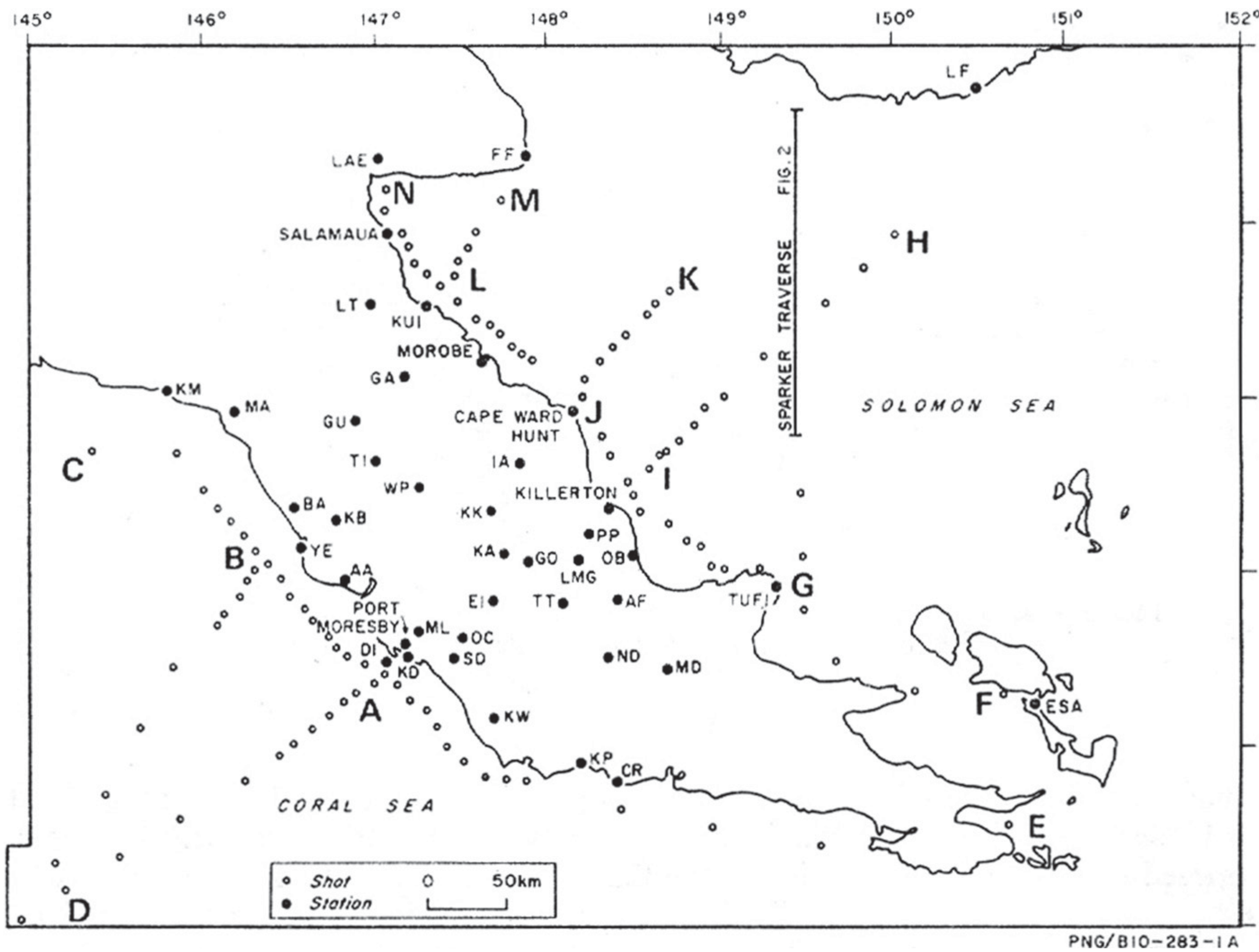


Figure 7.8.2-17. Location of shots (open circles) and recording sites (full circles) of the 1973 East Papua crustal survey (from Finlayson et al., 1976, fig. 3). [Geophysical Journal of the Royal Astronomical Society, v. 44, p. 45–60. Copyright John Wiley & Sons Ltd.]

Novaya Zemlya also was successfully recorded by 26 stations in east Papua (Finlayson, 1977).

The thickness of crustal material offshore ranges from 33 km in the western Solomon Sea (area L-M-N in Fig. 7.8.2-17) to 13 km in the central Solomon Sea (line H–I in Fig. 7.8.2-17). Along the southwest Papuan peninsula coast, a crustal thickness of 27–29 km was interpreted, which increased only slightly inland by 3–5 km until underthrusting takes place in the area of the Papuan Ultramafic Belt. The Moho appeared to shallow under the Moresby Trough of the Coral Sea (southwest of line A–B in Fig. 7.8.2-17) to 19 km and to deepen to 25 km under the Eastern Plateau in the center of the Coral Sea (around point D in Fig. 7.8.2-17). The results suggested a continental crust under the Papuan peninsula and the Eastern Plateau of the central Coral Sea, but an oceanic crust in between, underneath the Moresby trough (Fig. 7.8.2-19).

In 1978, a map was published showing the locations of marine crustal studies throughout the East and Southeast Asian Seas and the velocity-depth columns obtained. The map covered the area between latitudes 15°S and 45°N and longitudes 90°E and 150°E. The velocity columns contain velocity and thickness of both sedimentary layers and the layers of the crystalline crust and uppermost-mantle velocities as far as reached by the individual surveys. The map was compiled by Hayes et al. (1978).

Two major crustal surveys investigated the continental margins and adjacent seas of Australia. In 1976, the Scripps Institute of Oceanography, USA, conducted a seismic-reflection and wide-angle seismic survey in the Banda Sea north of Darwin using explosive seismic sources (Jacobson et al., 1979; Finlayson 2010; Appendix 2-2). The interpretation showed a crust of more than 30 km thickness north of Melville Island, decreasing slightly toward north to 30 km near the Timor Trough. The survey included observations across the Australian continental shelf between Darwin and Timor. In addition, Rynn and Reid (1983) recorded the shots fired in the Timor Sea along a 300-km-long offshore line on the continental shelf using OBSs. They determined a crustal thickness of 34 km. The offshore seismic shots were also recorded along a land profile between Darwin and Tennant Creek providing an upper-mantle velocity of 8.2 km/s and a crustal thickness of 45 km under the Pine Creek Geosyncline of northern Australia (Hales and Rynn, 1978; Hales et al., 1980).

The second marine survey of 1976 touched the margin of southern Australia, investigating the crustal structure within the Magnetic Quiet Zone of the Great Australian Bight. The

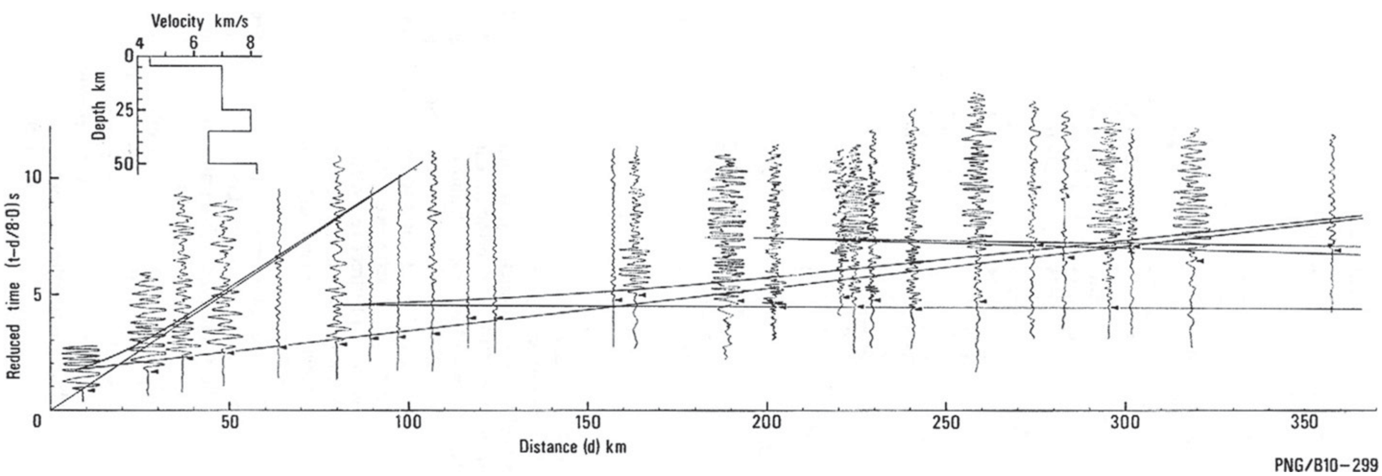
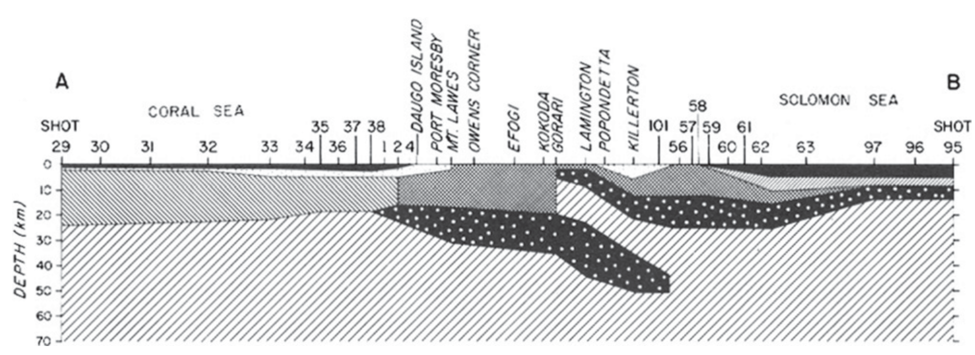


Figure 7.8.2-18. Record section of shots along the northeastern coast of the Papuan peninsula recorded between stations LAE and TUFI during the East Papua crustal survey (from Finlayson et al., 1976, fig. 6). [Geophysical Journal of the Royal Astronomical Society, v. 44, p. 45–60. Copyright John Wiley & Sons Ltd.]

Figure 7.8.2-19. Seismic model of the Papuan ultramafic belt along line D-H of Figure 7.8.2-17 (from Finlayson et al., 1977, fig. 9a). The location of the line D–H of Fig. 7.8.2-17 was renamed by Finlayson et al. (1977) into A–B. [Physics of the Earth and Planetary Interiors, v. 14, p. 13–29. Copyright Elsevier.]



marine seismic work was carried out by the Lamont-Doherty research vessel RV *Vema* (Talwani et al., 1979; Finlayson 2010; Appendix 2-2). Using sonobuoys, both refraction and reflection data were recorded applying both air gun and explosive sources with charges up to 30 kg. The resulting crustal thickness was 12–13 km, but the velocities were typical of oceanic crust.

7.8.3. The Indian Ocean

From the available and known literature, it appears that in the 1970s, only a limited amount of research was carried out in the Indian Ocean. Most of this work concentrated on the southwestern part adjacent to Africa. Some work had been carried out by Hales and Nation (1973b) in 1968 beneath the Transkei–Natal plateau and the Mozambique plateau (see Chapter 6.8.3).

In the eastern Indian Ocean, Kieckhefer et al. (1980) did a detailed crustal structure survey in 1977 which extended from the Sunda trench axis, across Nias island ridge, to the forearc basin near the coast of Sumatra. Six refraction lines were shot parallel to the regional topography (Fig. 7.8.3-01), with one line in the trench axis, two lines on the inner slope, and one line each on the Nias Ridge, the Nias Basin, and close to the coast of Sumatra. Both explosive charges and airguns provided energy recorded by arrays of sonobuoys.

A data example of long-range shots, recorded at station 1308 (Fig. 7.8.3-01) is shown in Figure 7.8.3-02. More data are shown in Appendix 7-10-4. The main feature of the interpretation of Kieckhefer et al. (1980) is a wedge of 4.7–4.9 km/s material seaward of the Nias Island ridge which underlies low-velocity sediments and thickens from 1.2 km beneath the trench axis to 13.5 km immediately seaward of the ridge (Fig. 7.8.3-03).

Tucholke et al. (1981) reported on new wide-angle reflection and refraction measurements in 1978 on the southern Agulhas Plateau 500 km off Cape Horn in the southwestern Indian Ocean, using an ocean bottom hydrophone, commercial long-range sonobuoys, and short-range military sonobuoys. Both explosives ranging from 2.5 to 100 kg and airgun shots were recorded up to distances of 72 km. Two reversed and six unreversed profiles were recorded over both smooth and irregular basement areas. Velocities of the deepest crustal layer

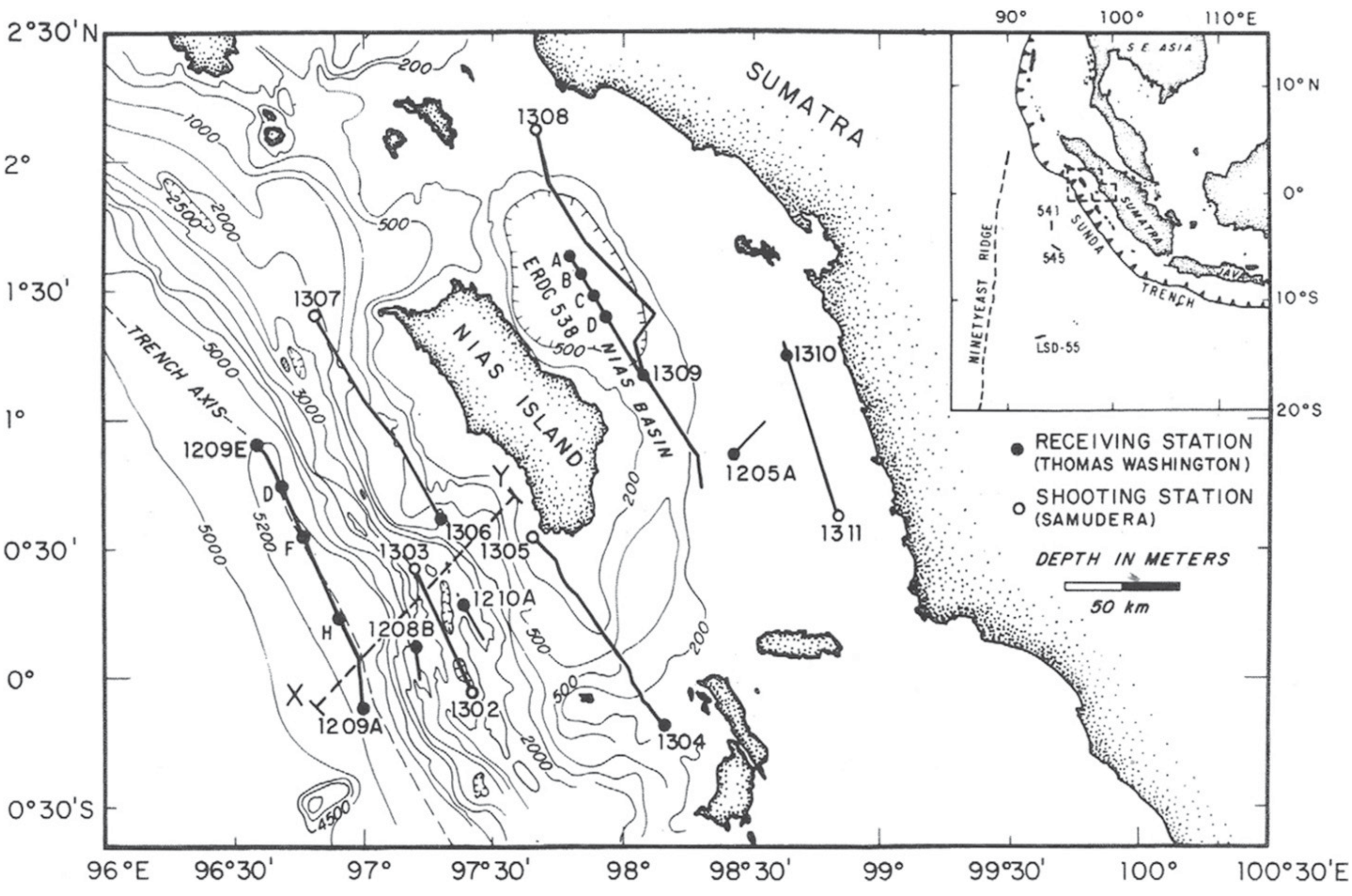


Figure 7.8.3-01. Location of seismic lines southwest of Sumatra between 0° and 2°N around Nias Island (from Kiekhefer et al., 1980, fig. 1). The contour interval of the bathymetric map is 500 nominal meters. [Journal of Geophysical Research, v. 85, p. 863–889. Reproduced by permission of American Geophysical Union.]

beneath the southern plateau ranged from 6.9 to 7.1 km/s and agreed with the basal crustal layer (6.6–7.0 km/s) of earlier models of the plateau and the adjacent Transkei Basin (e.g., Hales and Nation, 1973b). The thickness of this oceanic layer averaged 4 km or less in the Transkei Basin, but was at least 7–9 km thick under the Agulhas Plateau.

Goslin et al. (1981) investigated the anomalous crust of the Madagascar ridge south of Madagascar and the submarine Crozet plateau in 1978, located at ~45°S beyond the southwest Indian Ridge. Charges shot at intervals of 1–5 km and ranging from 25 to 400 kg were recorded on two seismic-refraction profiles of ~160 km length along the longitude of 45°E and north and south of latitude 32°S. On the Crozet Plateau, located to the south near 44°S, a 40-km-long refraction profile was recorded with shot intervals of 400 m. The deep structure of the Madagascar ridge resembled that of the submarine *Crozet Plareau*, corresponding to a strongly anomalous oceanic crust, Moho depths were 22–25 km under the northern and 14.5 km under the southern profile along the Madagascar ridge and 17.5 km under the Crozet Plateau; the P_n velocity was 8.0–8.1 km/s.

7.8.4. The Atlantic Ocean

As mentioned in the Introduction (section 7.8.1), seismic research in the 1970s concentrated mainly on particular structures, such as ridges, island arcs, continental margins, etc. This is evidently true also for the Atlantic Ocean.

Several surveys were carried out to study island chains off the western coast of Africa. The Cape Verde islands and emerged portions of a Mesozoic-Cenozoic volcanic accretion along fracture zones converging from the mid-Atlantic ridge toward Africa were both investigated by Dash et al. (1976).

The continental slope and the Cape Verde rise off Mauretania, an inactive margin with a sedimentary cover of 6 km and more, were the targets of a marine crustal survey by the University of Hamburg (Weigel and Wissmann, 1977).

The structure of the Canary Islands was investigated by a joint Spanish–Swiss research project (Banda et al., 1981a). Several *Meteor* (1964) expeditions—M33 in 1974 (Hinz et al., 1974), M39 in 1975 (see also Chapter 7.2.3; Makris et al., 1985), and M46 in 1977 (Sarnthein et al., 2008)—explored the continent-ocean transition west off Africa.

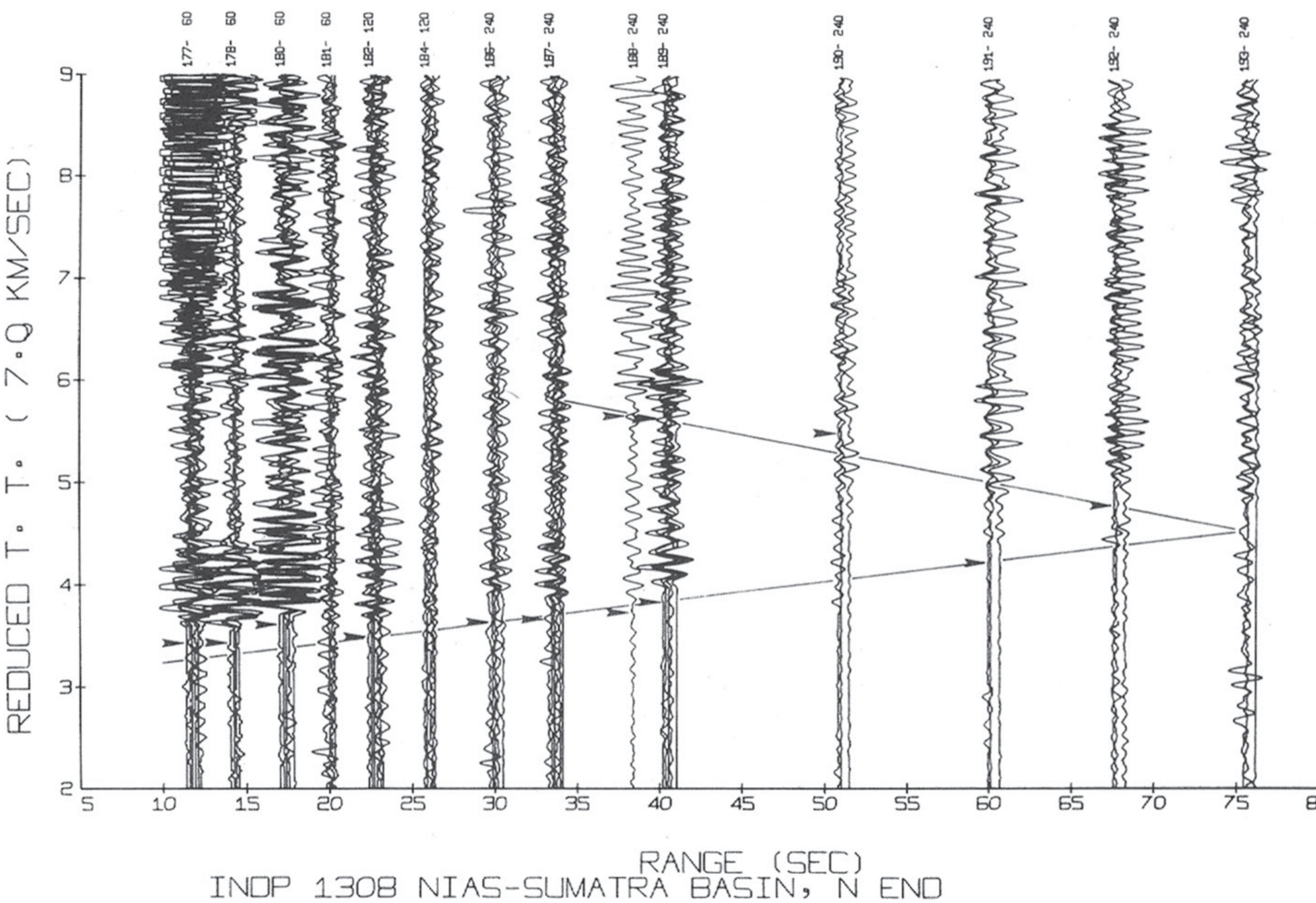


Figure 7.8.3-02. Record section for long-range shots (from Kiekhefer et al., 1980, fig. 4), recorded by array station 1308 (location shown in Figure 7.8.3-01). [Journal of Geophysical Research, v. 85, p. 863–889. Reproduced by permission of American Geophysical Union.]

The Mid-Atlantic Ridge was the goal of various seismic surveys, only a few of which are mentioned here. In 1977, a seismic-refraction experiment studied the crustal structure of the Kane Fracture Zone near 24°N, 44°W (Detrick and Purdy, 1980). Shooting lines and receivers formed a T configuration across the fracture zone, with two receivers located ~50 km apart in the fracture zone trough and the remaining eight receivers positioned 25–30 km apart on either side of the fracture zone (dashed lines in Fig. 8.9.4-01, showing both the 1977 and 1982 surveys; see Chapter 8.9.4). The data did not reveal any significant difference in crustal thickness or velocities on both sides of the fracture zone and were in the range typically observed for oceanic crust. Anomalously thin crust of only 2–3 km thickness, however, was found beneath the Kane fracture zone trough proper, extending over a width of at least 10 km. Further investigations were added in 1982 (see Fig. 8.9.4-01), when three more lines were shot across the Kane fracture zone (Cormier et al., 1984; see Chapter 8.9.4).

In 1973 within the Franco-American Mid-Ocean Undersea Study (FAMOUS) during cruise 54 of RRS *Discovery*, four seismic-refraction profiles were recorded along, across, and parallel to the median valley at 37°N (Whitmarsh, 1975; Fig. 7.8.4-01;

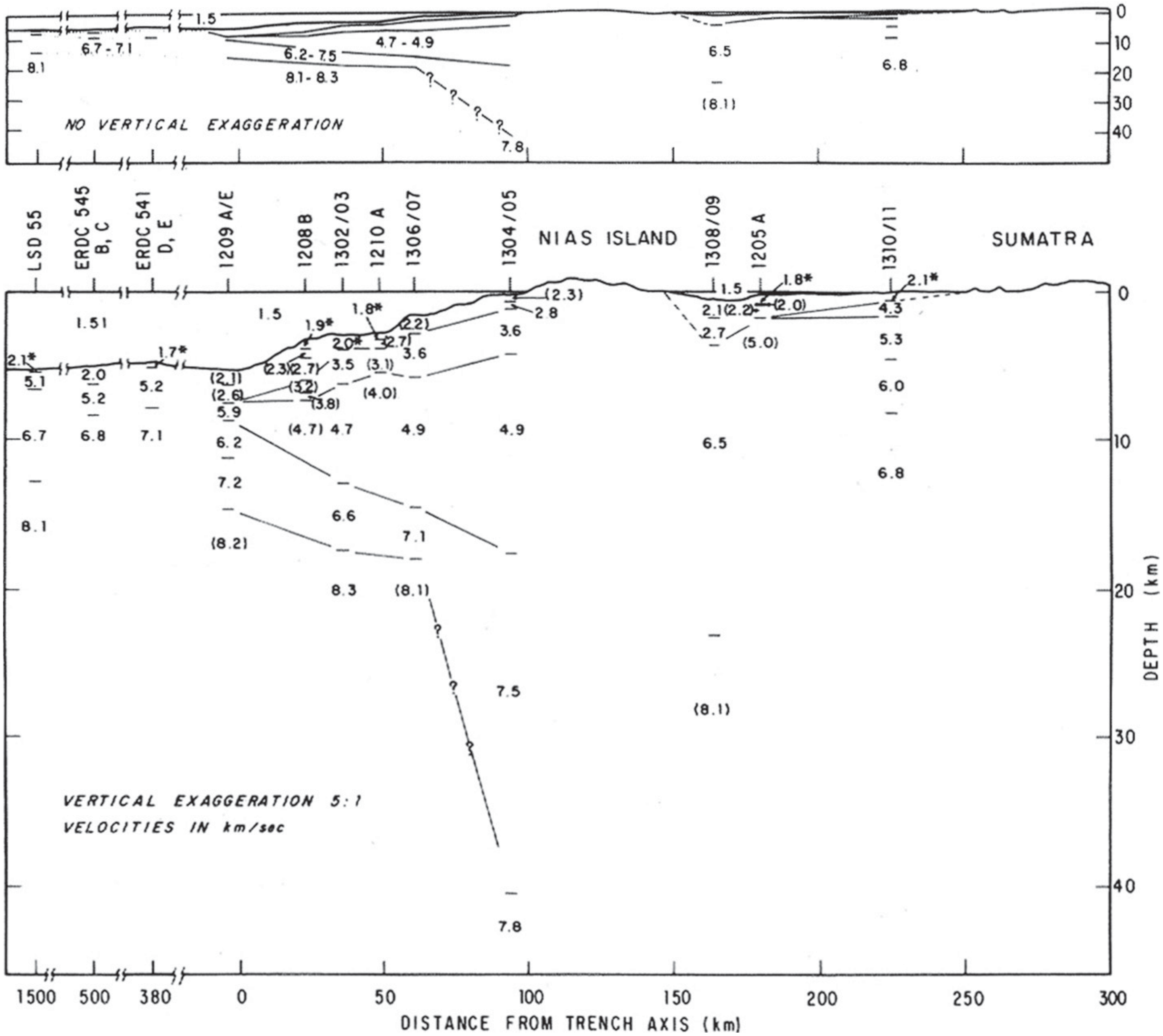


Figure 7.8.3-03. Summary cross section from the trench axis to Sumatra (from Kiekhefer et al., 1980, fig. 8), using depths of the midpoints of the reversed refraction profiles. [Journal of Geophysical Research, v. 85, p. 863–889. Reproduced by permission of American Geophysical Union.]

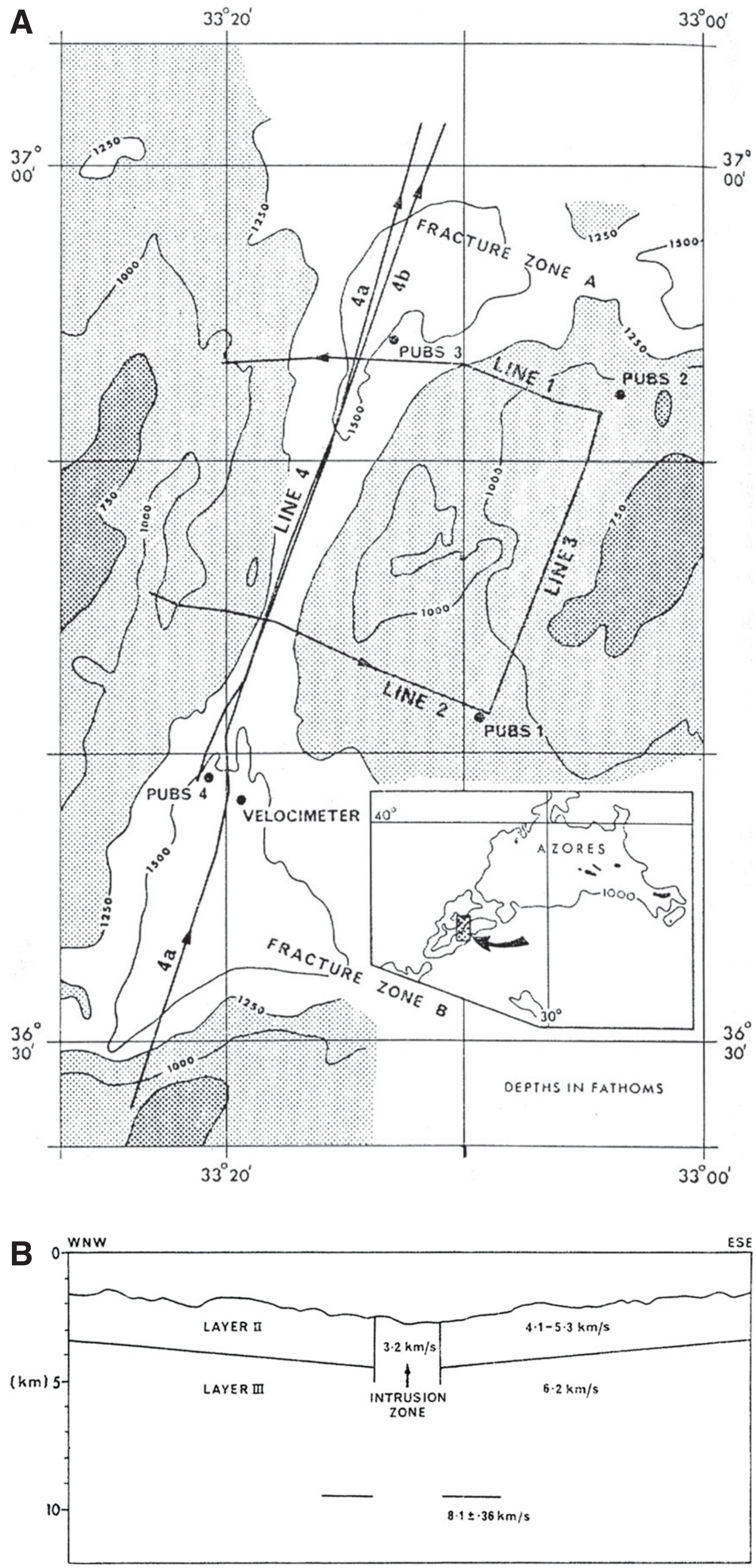


Figure 7.8.4-01. Location of seismic lines across the Median Valley of the Mid-Atlantic Ridge at 37°N (A) and model (B) along line 1 (from Whitmarsh, 1975, figs. 1 and 18). [Geophysical Journal of the Royal Astronomical Society, v. 42, p. 189–215. Copyright John Wiley & Sons Ltd.]

Appendix 7-10-5). Three of the lines had shots every 250 m. The shots were either explosive charges ranging from 2.5 to 100 kg, or were provided by a 1000 cubic inch airgun fired at 15 m depth every 2 minutes. As receivers, pop-up bottom seismic recorders were used, which recorded by means of a hydrophone and a tape recorder. Most arrivals were from a main refraction of apparent velocity of 5.4–6.3 km/s, only beyond a 35 km shotpoint distance were faster arrivals observed from an 8.1 km/s refractor. The main refractor corresponded in depth to approximately layer 3 of ocean basins, but its velocity was significantly less, possibly due to dip. A low-velocity layer with 3.2 km/s was found by lines 1 and 2 to underlie the whole width of the median valley floor. A problem remained as to how to reconcile the observation of the 6.3 km/s refractor along line 4 with the presence of the low-velocity zone as indicated by lines 1 and 2 (Fig. 7.8.4-01). Details of this survey were also described by other authors (Poehls, 1974; Fowler, 1976).

From 40° to 43°N near the Azores, four lithospheric profiles were recorded in 1974 and 1975 to unravel the structure of the lithosphere near the Mid-Atlantic Ridge (Steinmetz et al., 1977; Fig. 7.8.4-02). Crustal structure was obtained from an east-west profile which crossed the ridge axis.

Although energy was propagated across the ridge axis within the crust, the axial region marked a clear barrier to waves propagating in the mantle. Energy was successfully recorded up to 450 km distance in the strike direction of the ridge on a profile

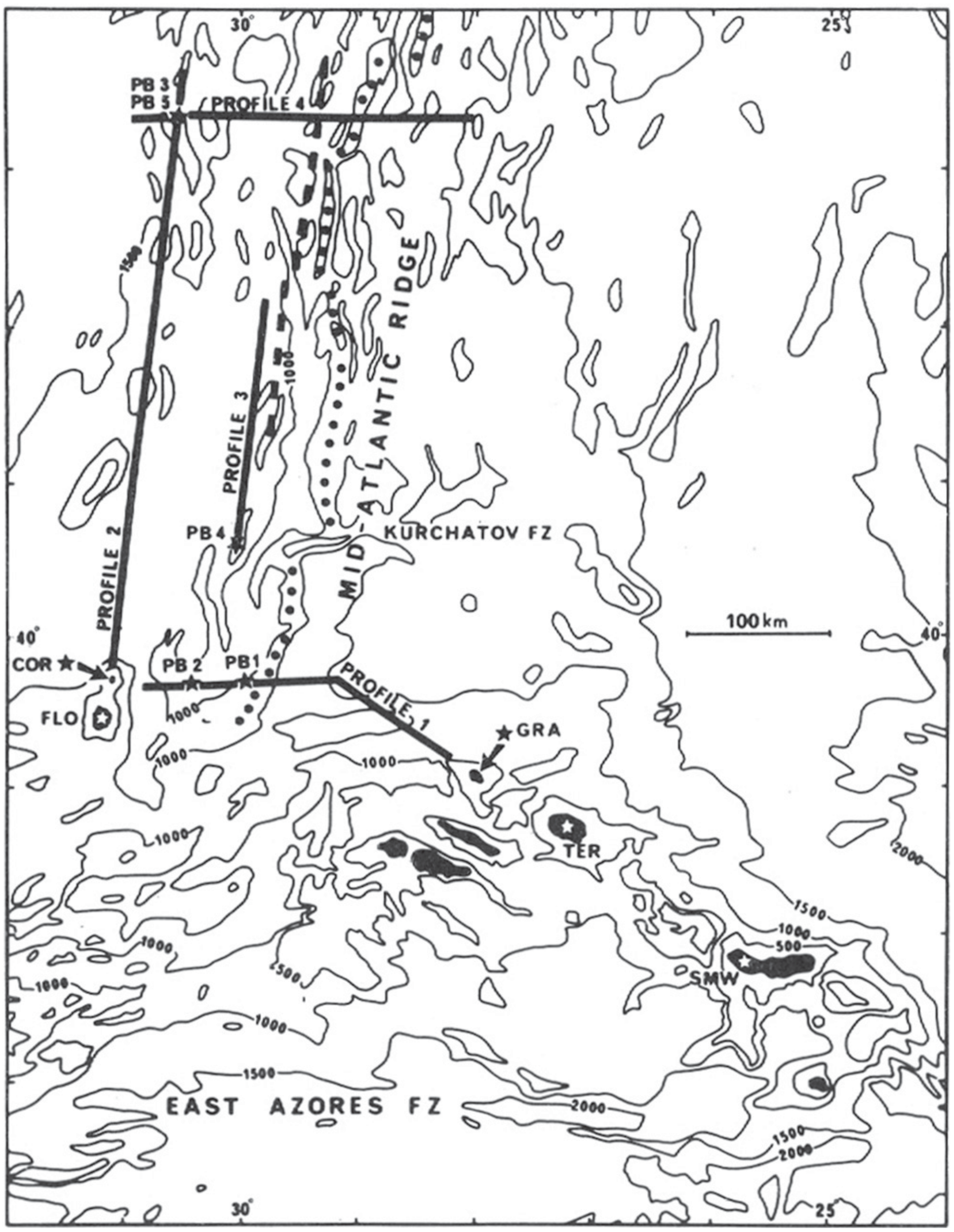


Figure 7.8.4-02. Location of seismic lines around the Mid-Atlantic Ridge at 40° to 43°N to the northwest of Azores islands (from Steinmetz et al., 1977, fig. 1). [Geophysical Journal of the Royal Astronomical Society, v. 50, p. 353–380. Copyright John Wiley & Sons Ltd.]

parallel to the axis along the 4 m.y. isochron. Here, below a layer with velocity 7.6 km/s, a low-velocity zone was found which was underlain by an 8.3 km/s refractor at 9 km depth below the sea bed (Fig. 7.8.4-03; see also Appendix 7-10-5). On another profile parallel to the rift axis, recorded along the 9 m.y. isochron, a velocity of 8.3 km/s was found at a much greater depth of 30 km.

Another survey investigated the crust along the Azores–Biscay rise (Whitmarsh et al., 1982). In 1978, a set of seismic-reflection and -refraction data was obtained in the vicinity of the rise. The seismic-refraction profiles were shot using pop-up bottom seismographs. Out to a range of 25 km, airgun firing provided shots every 2 minutes. At greater ranges, explosive charges of 11, 45, and 136 kg were used and recorded up to ~85 km distance. On all profiles, layer 2 with velocities of 5.7–6.0 and layer 3 with an average velocity of 6.7 km/s were underlain by an upper mantle with velocities of 7.8–8.2 km/s, but under one profile, a low-velocity zone on top of the mantle was found. Moho depth averaged around 13–14 km.

Fowler (1978) presented models for the crust beneath the Mid-Atlantic Ridge near 45°N. The above-mentioned 1973 seismic study of the FAMOUS area (37°N; Fowler, 1976) was continued in 1975 on RRS *Discovery* cruise 73 at 45°N. The purpose was to investigate the deep structure of the ridge crest. Seismographs were deployed west of the median valley, and shots were fired up to 120 km east of the valley. The experiment followed a refraction experiment shot by Keen and Tramontini (1970) in 1968 on the western flanks of the Mid-Atlantic Ridge at 45°N. They had obtained 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 (see Chapter 6.8.2). During the experiment of 1975, shots of 50, 100, and 150 kg were shot along three lines and recorded by five OBSs deployed along and west of the median valley. The first line was shot along the median valley and was 75 km long. A second line, also 75 km long, was recorded diagonally to the other two lines. For both lines, at the northern end of the median valley four internally recording sonobuoys were laid at 2 km intervals. The third line crossed the rift perpendicular to the rift axis; the majority of the shots were located east of the rift axis out to distances of near 120 km. On this line, an additional three sonobuoys were deployed at 2 km intervals at the eastern end of the line. The resulting traveltime data obtained from this experiment agreed well with those of Keen and Tramontini (1970), confirmed the presence of a normal oceanic upper-mantle velocity within a few million years of the ridge axis, and also seemed to confirm the presence of an anomalous upper mantle with lower velocities close to the ridge axis. The crustal structure appeared to be similar to the FAMOUS area at 37°N. At the ridge axis, an absorptive zone in the upper mantle was apparent; the depth below the seafloor to the top of this zone was determined to be ~6 km. Away from the rift axis, a positive velocity gradient of 0.04–0.05 km/(skm) was seen in the top 5–8 km of the upper mantle.

In the same area, at 45°N of the Mid-Atlantic Ridge, White and Whitmarsh (1984) reported on an investigation of seismic anisotropy due to cracks in the upper oceanic crust. The experiment was conducted in 1978 in an area centered ~28 km east of the median valley of the Mid-Atlantic Ridge at 45°25′N. The receivers were three 3-component ocean-bottom seismographs with hydrophones. The seismic source, two 16-l airguns, were towed along a shooting track, which gave an even azimuthal coverage at ranges of 4–10 km and 13–17 km, producing P-waves from layer 2 and from the lower crust, respectively. The result was a normal oceanic crustal velocity structure which could be explained by decreasing numbers of open cracks with increasing depth through the upper 2 km of the crust.

Seismic-refraction profiles across the continent-ocean transition, at the northern margin of the Bay of Biscay, identified an

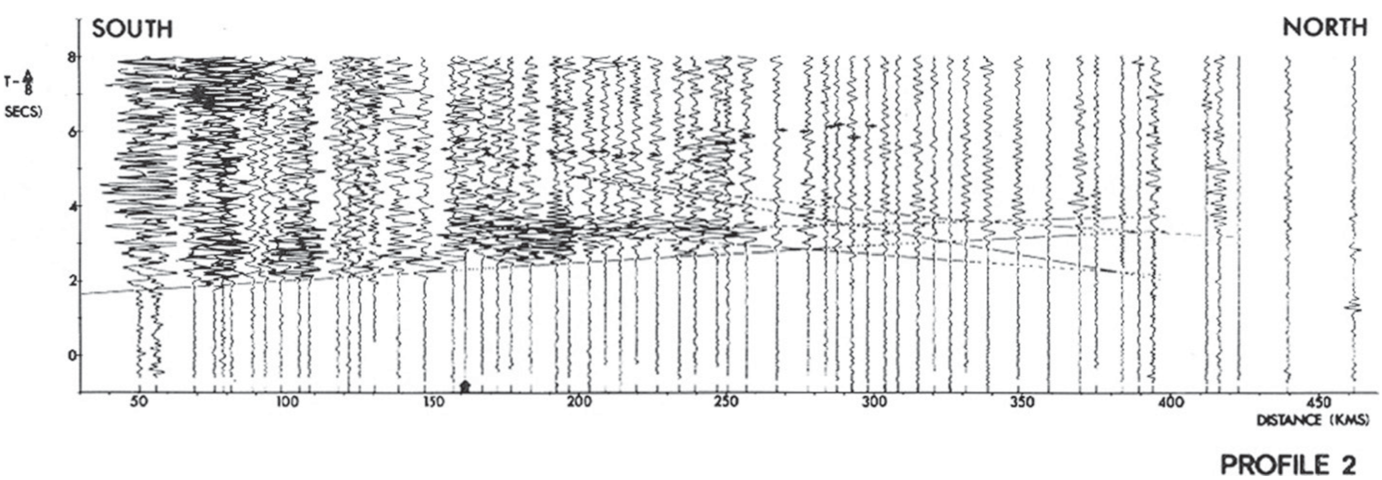


Figure 7.8.4-03. Record section of the north-south directed profile 2 west of the Mid-Atlantic Ridge at 40° to 43°N (from Steinmetz et al.,1977, fig. 5a). Data were observed on the island of Flores (FLO in Fig. 7.8.3-02). [Geophysical Journal of the Royal Astronomical Society, v. 50, p. 353–380. Copyright John Wiley & Sons Ltd.]

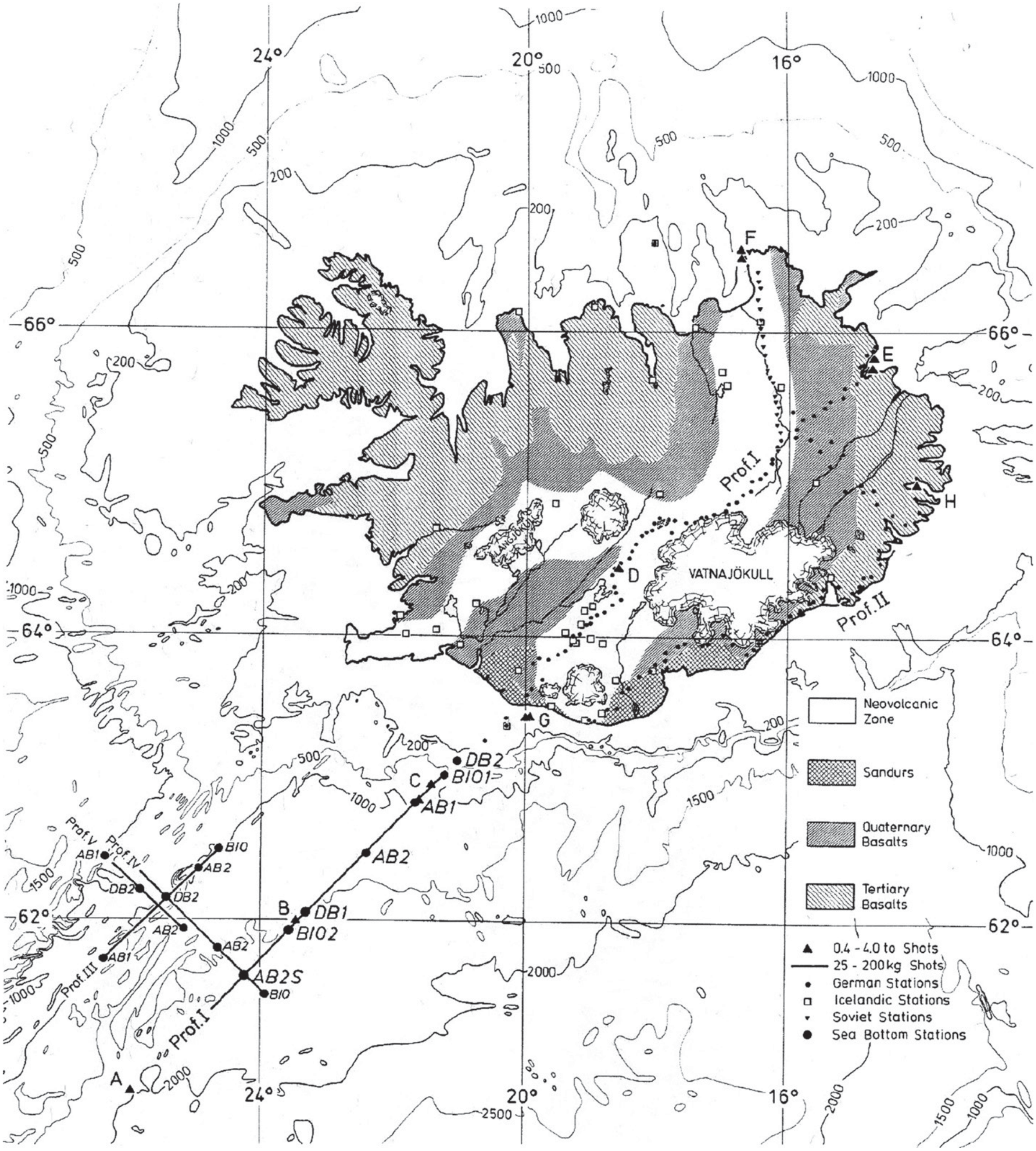


Figure 7.8.4-04. Geologic sketch of Iceland and bathymetric chart of the surrounding ocean showing shotpoints and recording sites of the RRISP project (from RRISP Working Group, 1980, fig. 1). [Journal of Geophysics, v. 47, p. 228–238. Reproduced with kind permission of Springer Science+Business Media.]

8-km-wide progressive thinning of the continental crust from 33 km thickness to ~5 km. In 1975 and 1976, several cruises were undertaken into the Bay of Biscay to shoot profiles, successfully using the newly constructed ocean-bottom seismographs under rough field conditions. More tests followed in 1976 and 1977 in the Norwegian Sea (Avedik et al., 1978).

The Hebridean Margin Seismic Project (Bott et al., 1979) targeted the crust west of northern Scotland and north of Ireland. During summer 1975, a line of large shots was fired across the continental margin between the Rockall Trough and the Hebridean shelf along latitude 58°N. Another line of large shots extended in an approximately N-S direction around longitude 8°W from the 58°N line southward to northern Ireland. The shots were recorded both across Scotland and in northwestern Ireland. The data indicated a 27-km-thick continental crust beneath the Hebridean shelf and a depth to Moho of 18 km beneath the Rockall Trough at 58°N.

Of particular interest to earth scientists in the 1970s was evidently the northernmost part of the Mid-Atlantic Ridge area around Iceland. Following Talwani's (Talwani et al., 1971) investigations of the Reykjanes Ridge between 60° and 63°N, several expeditions have dealt with the ridge between 59°N and Iceland.

In 1977, for example, between 60° and 63°N, three new along-strike profiles were recorded over crust 0.3–9 · 10^6 years old (Bunch, 1980). The profiles were 120 km long and were located at the southwestern limit of the section of the Reykjanes Ridge with no median valley where the ridge breaks up into sections whose axes lie obliquely to the overall spreading axis of the ridge. The detailed structures obtained indicated that with increasing age, the 4.6 km/s layer thins, and the crust with velocities 6.6–7.1 km/s thickens and increases its mean velocity. The velocity of the deepest layer seen in this experiment increased with age from 7.1 km/s at 0 m.y. to 8.2 km/s at 9 m.y.

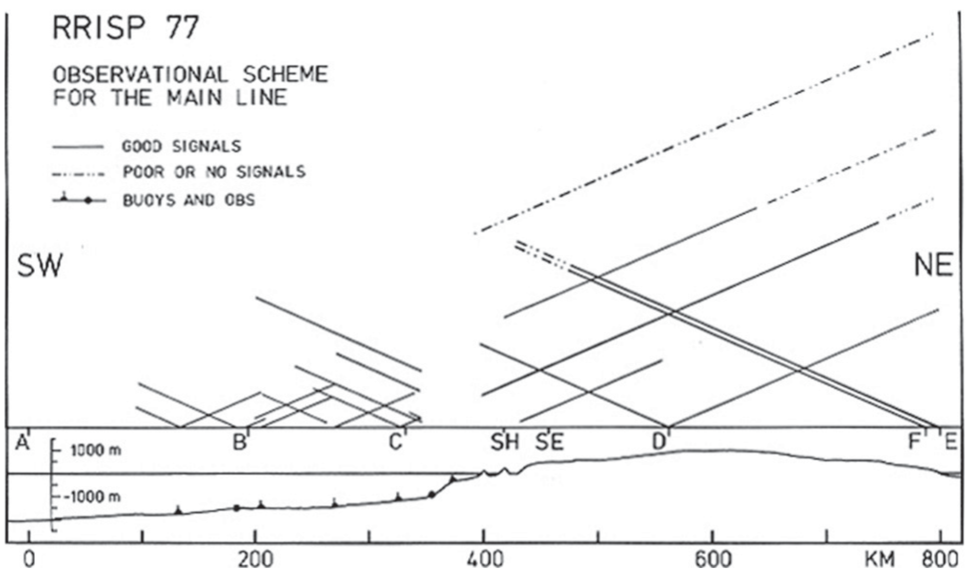


Figure 7.8.4-05. Layout of the RRISP project (from RRISP Working Group, 1980, fig. 3). [Journal of Geophysics, v. 47, p. 228–238. Reproduced with kind permission of Springer Science+Business Media.]

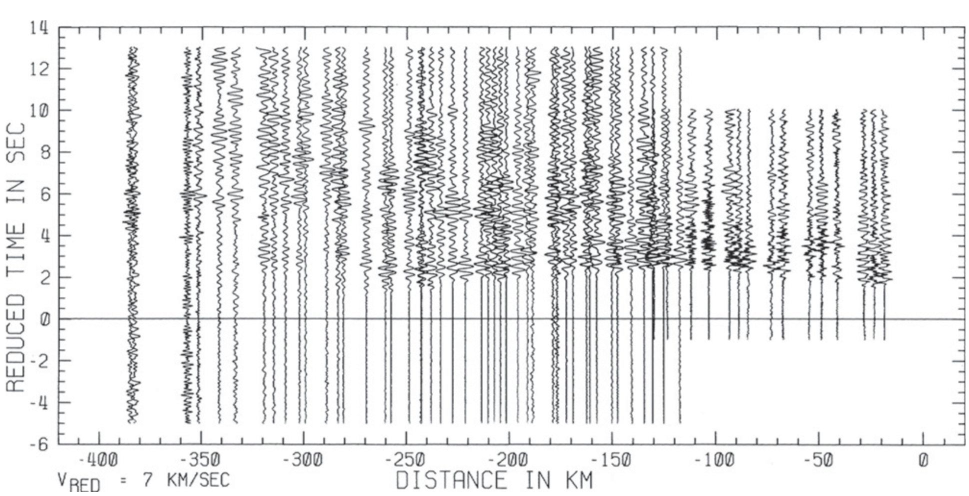


Figure 7.8.4-06. Record section along the land segment of the main profile, shots F1 to F4 (from RRISP Working Group, 1980, fig. 4). [Journal of Geophysics, v. 47, p. 228–238. Reproduced with kind permission of Springer Science+Business Media.]

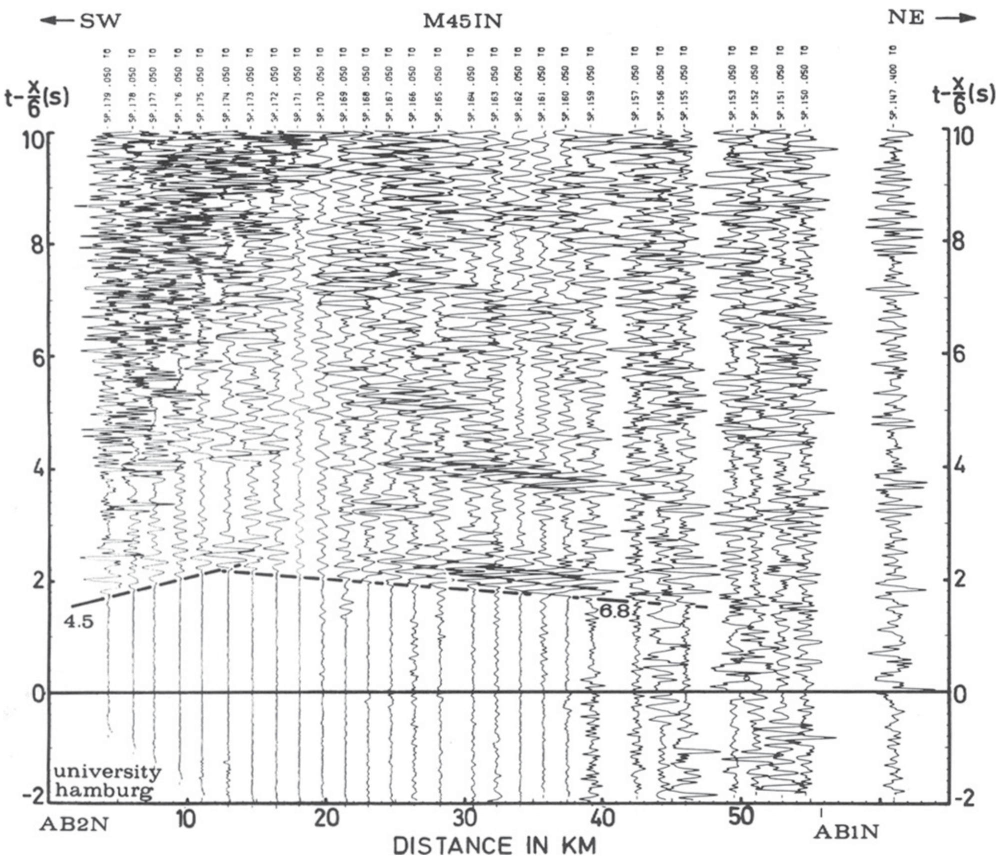


Figure 7.8.4-07. Marine record section (from RRISP Working Group, 1980, fig. 5), recorded by the ground hydrophone of buoy ABS2N (water depth 1389 m). [Journal of Geophysics, v. 47, p. 228–238. Reproduced with kind permission of Springer Science+Business Media.]

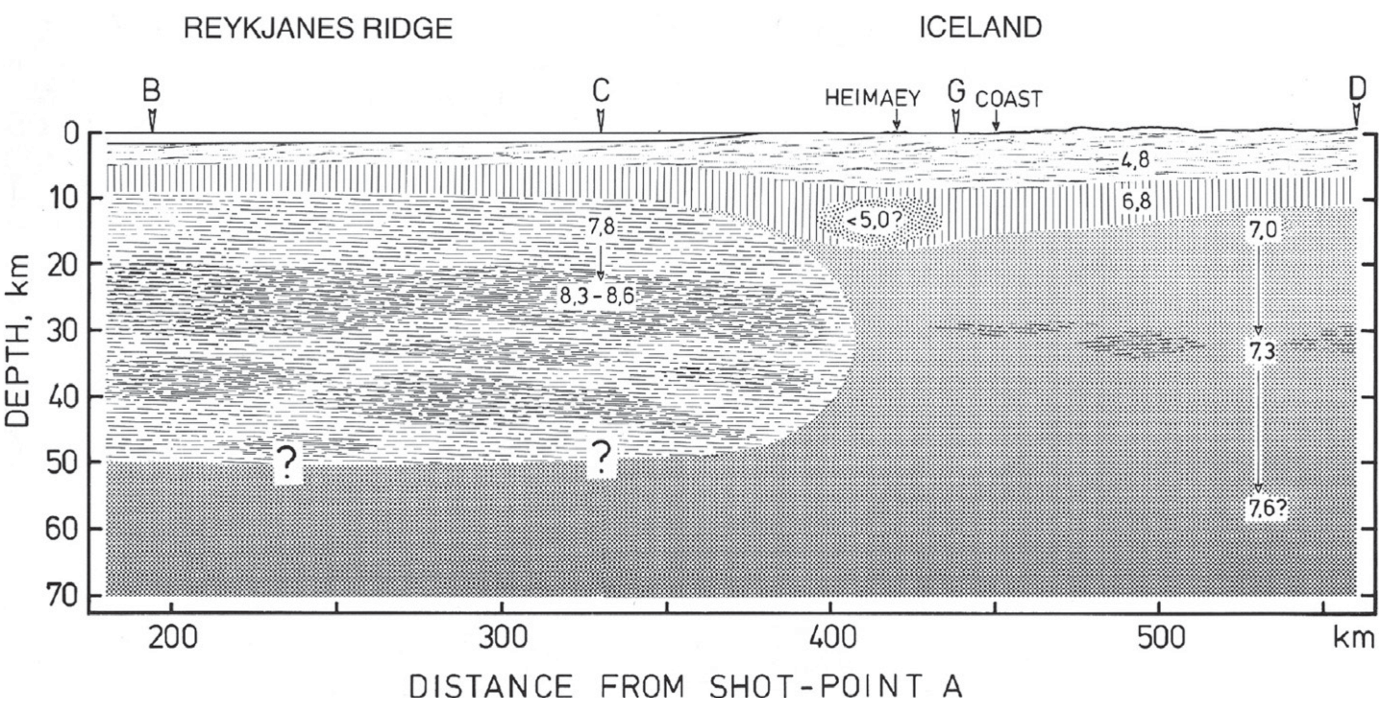


Figure 7.8.4-08. Generalized crustal and upper mantle cross section of the central part of the RRISP77 main profile (from RRISP Working Group, 1980, fig. 7). Letters indicate positions of large shots. [Journal of Geophysics, v. 47, p. 228–238. Reproduced with kind permission of Springer Science+Business Media.]

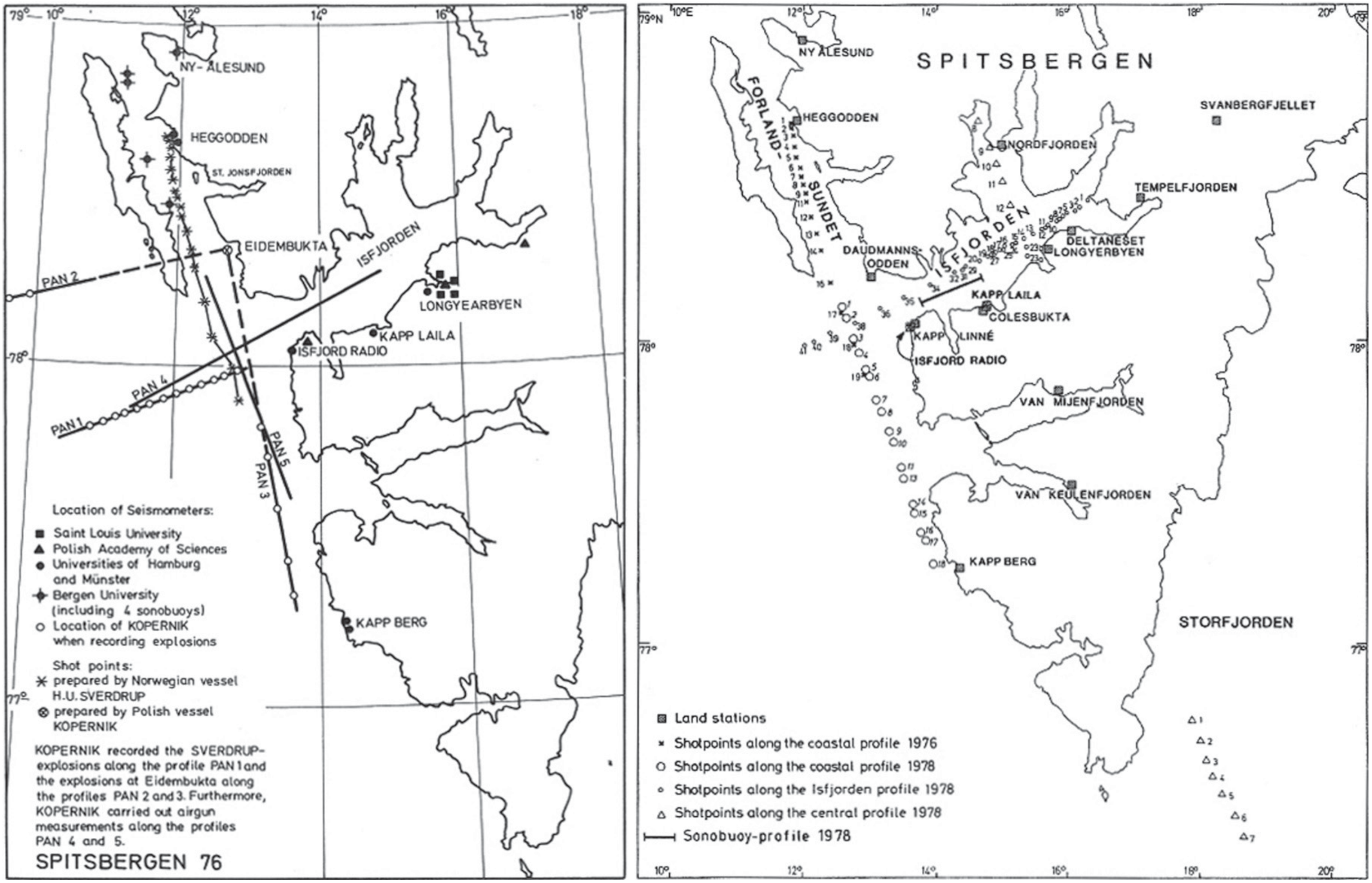


Figure 7.8.4-09. Location of shot profiles, sonobuoys and land stations seismic observations in and around Spitsbergen. Left: 1976 survey (from Guterch et al., 1978, fig.IV-2). Right: 1978 survey (from Sellevoll, 1982, fig. 4.1). [University of Bergen, Seismology Observatory, Bergen, 62 p. Published by permission of the author.]

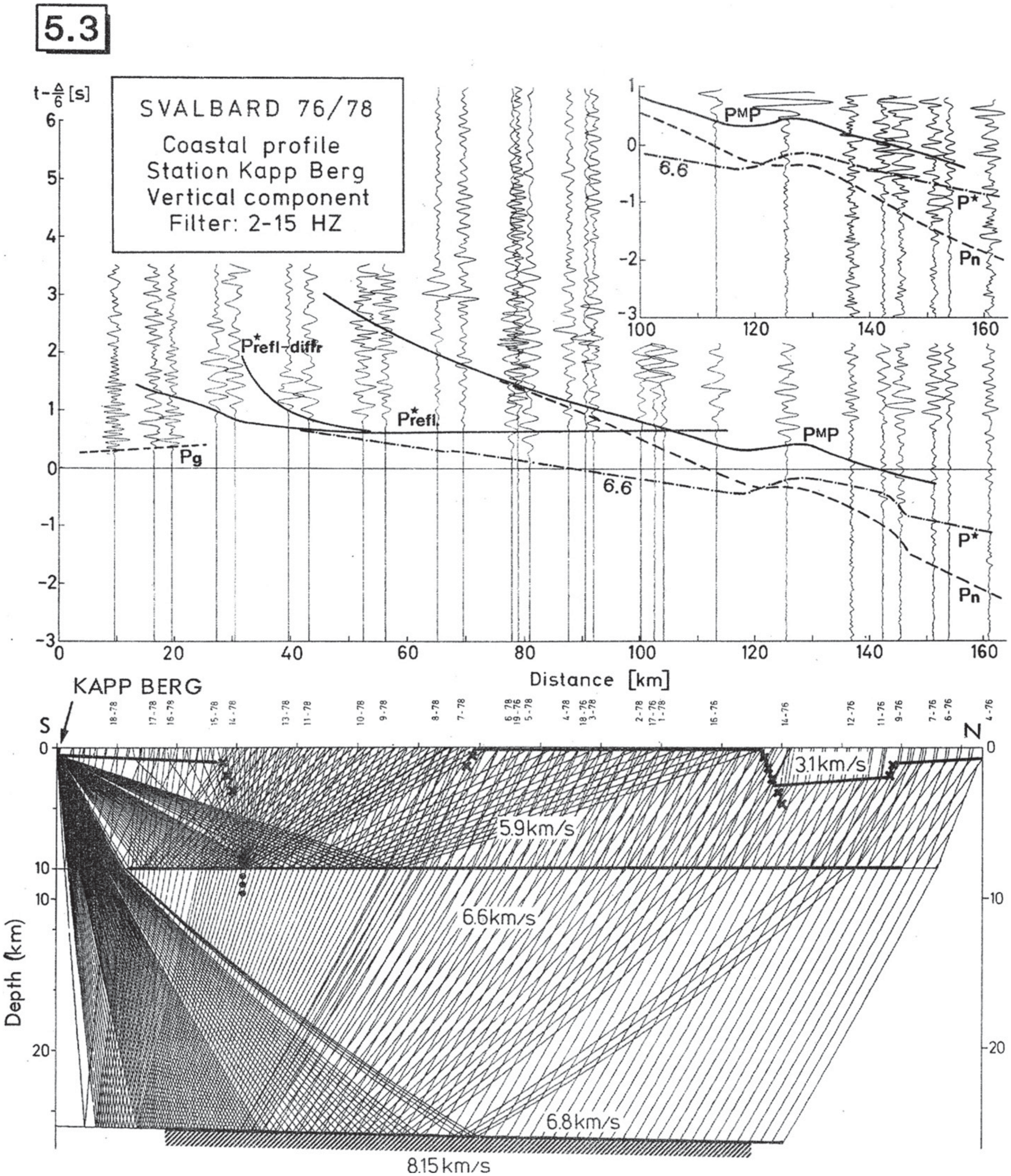


Figure 7.8.4-10. Record section of the coastal profile for station Kapp Berg on Spitsbergen and corresponding crustal model (from Sellevoll, 1982, fig. 5.3). [University of Bergen, Seismology Observatory, Bergen, 62 p. Published by permission of the author.]

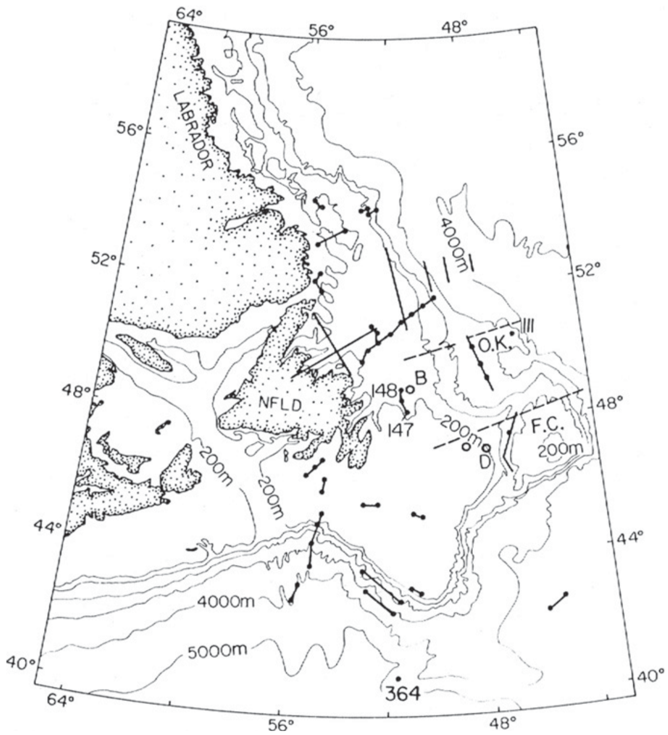


Figure 7.8.4-11. Location of crustal investigations of the continental margin off eastern Canada (from Keen and Barrett, 1981, fig. 1). Solid lines—earlier and the 1977–1978 seismic refraction lines; dashed lines—multichannel seismic reflection lines; open circles—deep exploratory wells; 111—DSDP drilling site; O.K.—Orphan Knoll; F.C.—Flemish Gap. [Geophysical Journal of the Royal Astronomical Society, v. 65, p. 443–465. Copyright John Wiley & Sons Ltd.]

The *Meteor* (1964) expedition M45 in 1977 aimed in particular to investigate the Reykjanes Ridge and to support the onshore-offshore Reykjanes Ridge Seismic Project (RRISP Working Group, 1980).

This was a major long-range seismic-refraction experiment which was realized in 1977 along an 800-km-long line along the southeastern flank of the Reykjanes Ridge, with the aim of resolving both crust and upper mantle to greater depths than previously possible, and of studying the transition from oceanic to the Icelandic structure (Fig. 7.8.4-04). Figure 7.8.4-05 presents the layout of the project. Shots on Iceland and at sea were recorded by up to 90 stations from Iceland, Germany, and the USSR, and by seven ocean-bottom seismographs from Germany and Canada (RRISP Working Group, 1980). In addition, detailed marine seismic investigations were carried out during cruise M45 of the German research vessel *Meteor* (1964). Data examples are shown in Figures 7.8.4-06 and 7.8.4-07 (see also Appendix A7-10-6).

Goldflam et al. (1980) reported normal, but thickened (8 km), oceanic crust for the flanks of the Reykjanes Ridge south of Iceland, thicker than elsewhere on mid-oceanic ridges. The unconsolidated sedimentary layer thickens when approaching the Iceland plateau, while the deeper crustal layers show an almost horizontal layering and no along-strike changes in crustal thickness were observed. Below the Moho, at 9 km depth below sea level, the upper-mantle velocity of 7.8 km/s was shown to increase gradually and reach 8.2 km/s at ~16 km depth and more.

Similar results were shown by Gebrande et al. (1980) and the RRISP Working Group (1980) for the Reykjanes Ridge flank, as well as for Iceland itself, which also sits on top of the active

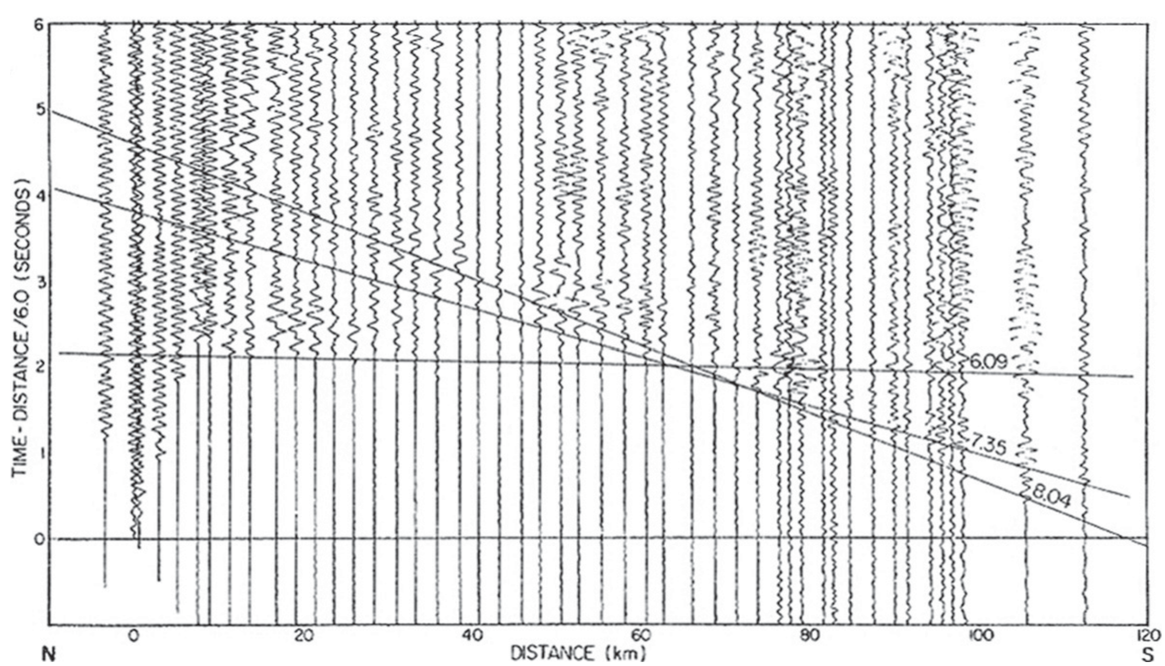


Figure 7.8.4-12. Record section of OBS 7-8-3, line 1, near Orphan Knoll (from Keen and Barrett, 1981, fig. 5b). [Geophysical Journal of the Royal Astronomical Society, v. 65, p. 443–465. Copyright John Wiley & Sons Ltd.]

spreading center of the Atlantic Ocean. Their structural model of Iceland beneath the neo-volcanic zone shows a generalized two-layered crust of variable (10–14 km) thickness underlain by an anomalous mantle with P-wave velocities of 7.0 km/s at the base of the crust increasing to 7.4 km/s at 30 km depth and 7.6 km/s at 50–60 km depth (Fig. 7.8.4-08). This anomalous mantle is evidently confined to Iceland, and a sharp transition resulted for the area of the shelf edge where normal oceanic lithosphere replaces the upward-doming asthenosphere (Gebrande et al., 1980).

In addition to the seismic-refraction measurements, reflection surveys were accomplished in the axial zone of southwestern Iceland, as well as in the flood basalts of northern Iceland (Zverev et al., 1980a, 1980b), to study the upper crust in detail. Reflectors with steep dip could be traced to depths of 15 km, while for northern Iceland, a seismically homogeneous body without reflectors seems to be present at 10–15 km depth.

Bott and Gunnarson (1980) interpreted the data of a project on the Iceland-Faeroe Ridge, which connects the Iceland block on the active spreading ridge with the Faeroe block which may be of continental character. They interpreted the data to demonstrate a thick oceanic crust of 30–35 km under the ridge to be separated by a true continental margin from the Faeroe block, and underlain by continental crust beneath the lava cover (Bott et al., 1974, 1976). Also a *Meteor* (1964) expedition (cruise M48) in 1978 explored details along the Iceland-Faeroe Ridge (Sarnthein et al., 2008).

Spitsbergen (also named Svalbard) located close to the northern end of the Mid-Atlantic became the target of an inter-

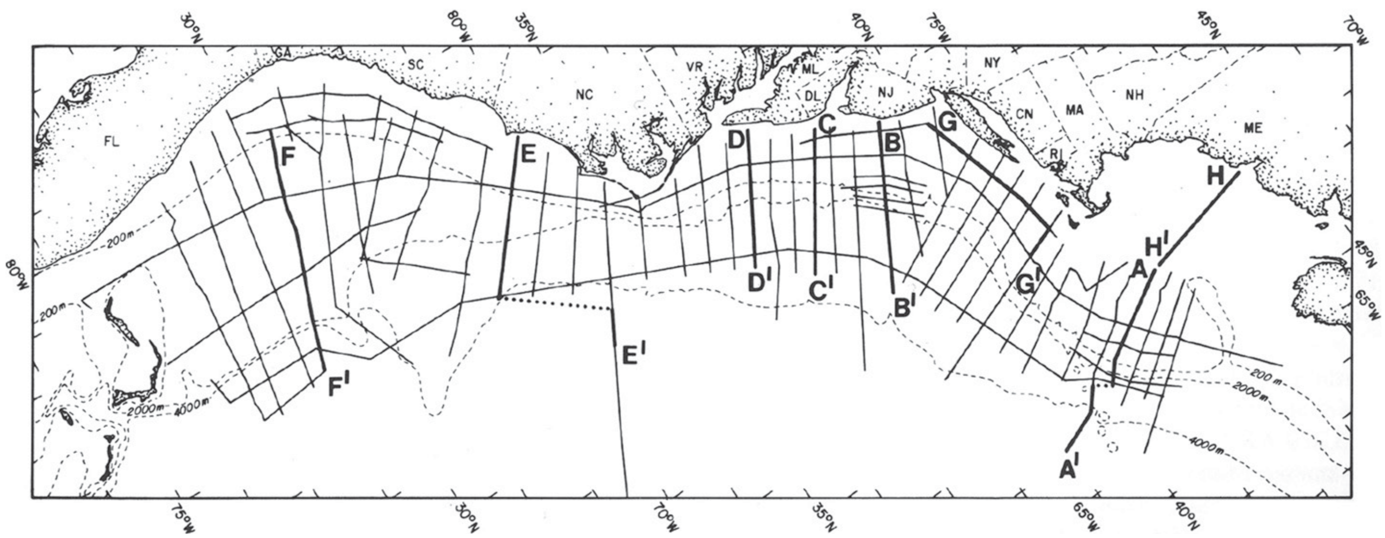


Figure 7.8.4-13. Location of multichannel seismic reflection profiles collected by the U.S. Geological Survey during the 1970s on the Atlantic continental margin (from Trehu et al., 1989, fig.7). [*In* Pakiser, L.C., and Mooney, W.D., eds., Geophysical framework of the continental United States: Geological Society of America Memoir 172, p. 349–382. Reproduced by permission of the Geological Society of America.]

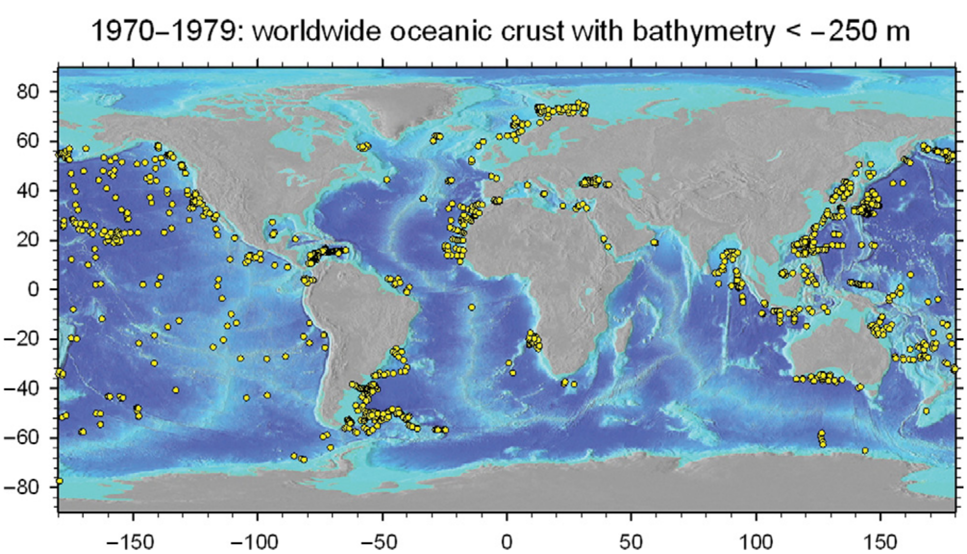


Figure 7.8.5-01. Seismic refraction measurements in the Atlantic, Pacific, and Indian Oceans performed between 1970 and 1979 (data points from papers published until 1979 in easily accessible journals and books).

national research project of institutions from Germany, Norway, Poland, and the United States under Polish leadership (Guterch and Sellevoll, 1981). Two expeditions were performed in 1976 and 1978. Shot profiles, sonobuoys, and stations on land provided a sophisticated system of observations along the western half of Spitsbergen and its adjacent oceanic margin (Fig. 7.8.4-09). The data were of good quality, but were only published in internal reports (Guterch et al., 1978; Sellevoll, 1982; Appendix 7-10-7). In 1976, the maximum observation distance along the N-S line was 180 km, allowing to record P_MP and P_n arrivals from the Moho (Fig. 7.8.4-10), with a preliminary interpretation giving a crustal thickness of ~25 km. Further inland crustal depths up to 35 km resulted (Faleide et al., 1991; Sellevoll et al., 1991).

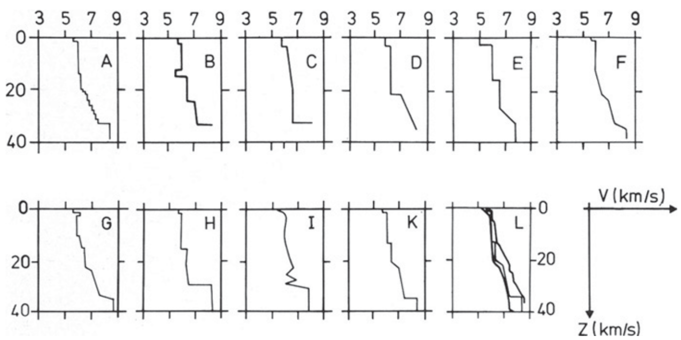


Figure 7.9-01. Comparison of interpretations of data set 2 by various authors (from Ansorge et al., 1982, fig. 24). [Journal of Geophysics, v. 51, p. 69–84. Reproduced with kind permission of Springer Science+Business Media.]

The sedimentary and crustal structure of the western margin of the Atlantic Ocean bordering the coast of North America was the target of detailed seismic-reflection campaigns.

The margin of eastern Canada (Fig. 7.8.4-11) had been intensively studied in the 1960s (Dainty et al., 1966). In the 1970s, multichannel seismic-reflection lines were added (Grant, 1975) and in 1977 and 1978, two deep crustal seismic-refraction measurements were carried out (Fig. 7.8.4-11; lines close to O.K. and F.C.), using OBSs (Keen and Barrett, 1981). The aim of this particular project was to determine the nature of the crust and the depth to Moho beneath Orphan Basin and Flemish Gap which region was characterized by the existence of continental fragments according to results of DSDP drilling at site 111 on Orphan Knoll (Ruffman and van Hinte, 1973) and other deep exploratory wells (Keen, 1979), for which locations are also shown in Figure 7.8.4-11. Data were recorded up to 120 km distance, (Fig. 7.8.4-12; Appendix A7-10-8) and the resulting crustal thickness ranged around 20 km both under the Flemish Gap and under the Orphan Basin (Keen and Barrett, 1981).

From 1973 to 1979, a major seismic-reflection program was carried out by the U.S. Geological Survey along the U.S. Atlantic

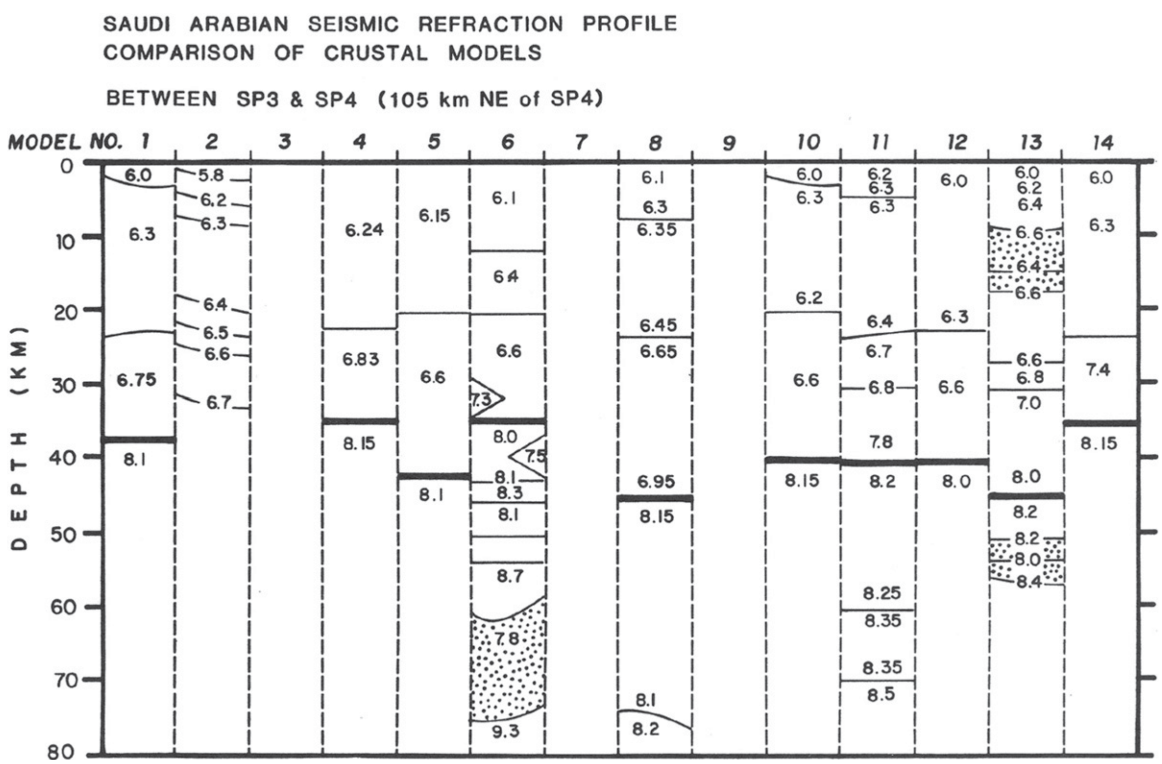


Figure 7.9-02. Comparison of interpretations of the same reversed data sets by various authors, here for shotpoints 3 and 4 (from Mooney and Prodehl, 1984, fig. 72). [Proceedings of the 1980 Workshop of the IASPEI on the seismic modeling of laterally varying structures: Contributions based on data from the 1978 Saudi Arabian refraction profile. U.S. Geological Survey Circular, 937, p. 140–153. .]

continental margin to evaluate the sedimentary basins of the shelf, slope, and rise for their hydrocarbon potential.

Within this program, a major portion was run on the Long Island Platform (Fig. 7.8.4-13, line LIP on Figure 8.5.3-24). The resultant CDP sections were processed to 12 seconds two-way traveltime and provided seismic sections through the cratonized crust of the southern New England Appalachians and its transition from continental thickness on the Long Island Platform to oceanic crust across the passive rifted margin (Hutchinson et al., 1986; Phinney, 1986; Phinney and Roy-Chowdhury, 1989).

Also in the 1970s, the surroundings of the site of the 1886 Charleston earthquake were investigated by the U.S. Geological Survey with regard to the likelihood of future earthquakes of similar size. Within this program, multichannel reflection work was started in 1979 on land and offshore in the Charleston area. The survey was continued in 1981 and was discussed in detail by Behrendt (1986), in conjunction with the 1978–1979 COCORP lines in Georgia and in the Charleston area (Cook et al., 1979a, 1981). A complete overview (Fig. 7.8.3-13) was published by Trehu et al. (1989b).

7.8.5. Summary of Seismic Observations in the Oceans in the 1970s

One of the highlights of the 1970s was the attempt to record super-long profiles not only on the continents, but also in the oceans, that penetrated well below Moho. In the Pacific Ocean, e.g., a 600-km-long line was laid out in the northeastern Pacific between the Clarion and Molokai fracture zones (Orcutt and Dorman, 1977). In the western Pacific basin, explosions were recorded by OBSs up to distances of 1300 km during two Longshot experiments in 1973 in the East Mariana basin (Asada and Shimamura, 1976). From 1974 to 1980, further intensive long-range experiments were undertaken in the northwestern Pacific, which detected large-scale lithospheric structure and anisotropy in the upper-mantle (Asada and Shimamura, 1979; Asada et al., 1983; Shimamura et al., 1983).

An excellent summary of marine crustal studies for the east and southeast Asian Seas was published as a map by Hayes et al. (1978) compiling all data available at that time between Japan in the northeast and Australia in the south, also covering part of the Indian Ocean west of Thailand and Indonesia.

Dealing also with the lower lithosphere, from 40° to 43°N a long-range profile was recorded near the Azores (Steinmetz et al., 1977). Another long-range seismic-refraction experiment in 1977 aimed to resolve both crust and upper mantle to greater depths than previously possible by establishing an 800-km-long line along the southeastern flank of the Reykjanes Ridge (RRISP Working Group, 1980).

An overview of seismic measurements in the oceans, compiled from the database assembled at the U.S. Geological Survey in Menlo Park, California, shows all surveys that were published until 1979 in generally accessible journals and books for the 1970s (Fig. 7.8.5-01).

The map shows, similar to the map for the 1960s, numerous observations in continental shelf and adjacent deep-ocean areas of North America and eastern Asia. There are only a few data points in the center of the Atlantic Ocean, and none in the Indian Ocean, but much research was conducted in the Pacific Ocean. A few points also touch the Antarctica. New are numerous proj-

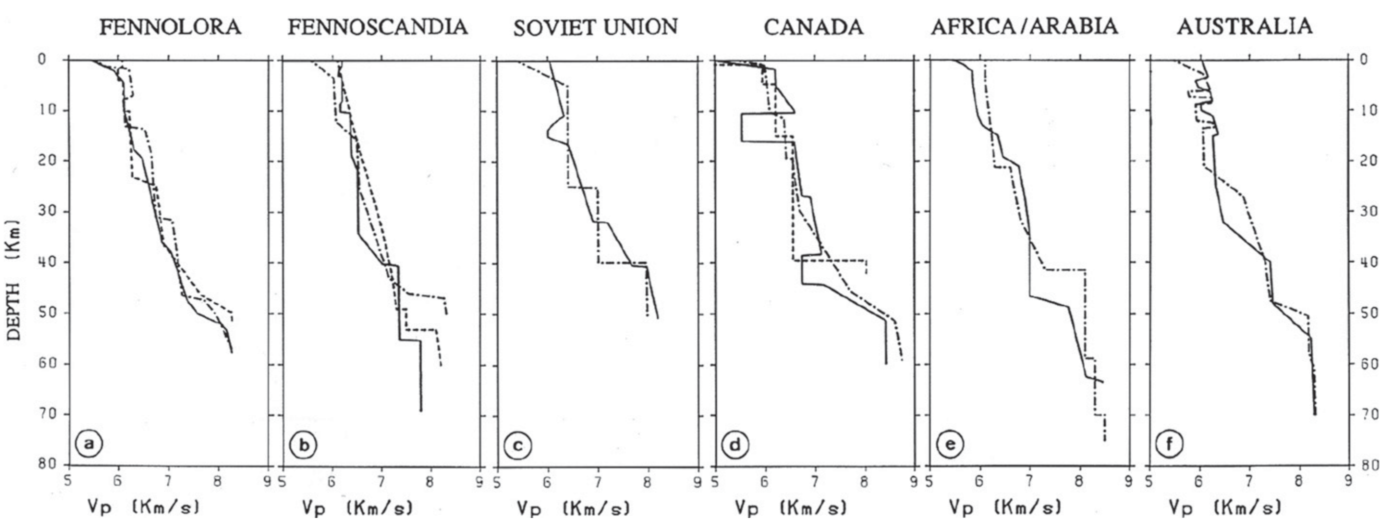


Figure 7.10-01. Representative velocity-depth functions of the crust of different shield areas (from Guggisberg et al., 1991, fig. 26). In corresponding order: solid line, dashed line, dashed-dotted line: (a) FENNOLORA 1979: E-south, C-north, G-south; (b) SVEKA 1981, block III (Luosto et al., 1984), BALTIC 82, block III (Luosto et al., 1985), Blue Road 1972, 5-west (Lund, 1979); (c) Russian platform (Pavlenkova, 1979), West Siberian Craton (Puzyrev and Krylov, 1971), (d) Grenville province, 2-west (Berry and Fuchs, 1973), Churchill province (Green et al., 1980), Superior province, 3-east (Berry and Fuchs, 1973); (e) Kalahari Craton, profile III (Baier et al., 1983), Arabian Shield, shotpoint 3 (Mooney et al., 1985); (f) North Australian Craton TCMI-3 west and TCMI-2 east (Finlayson, 1982). [Tectonophysics, v. 195, p. 105–137. Copyright Elsevier.]

ects which explored the continental shelves and adjacent ocean areas along eastern South America, northwestern Africa, the Norwegian coast and the Rockall trough, the Melanesian and Indonesian archipelagos, and southern Australia. The Mid-Atlantic Ridge in the northern Atlantic Ocean and the Hawaii Islands in the Pacific Ocean were of special interest.

7.9. ADVANCES IN INTERPRETATION METHODOLOGY

Throughout the 1970s, the interpretations of the many seismic-refraction experiments concerning the crust and upper mantle were mainly performed using traveltime routines for which many different computer programs had been developed over the years. In 1968, Helmberger (1968) had published the theory to calculate synthetic seismograms for layered transition zones. In addition, the reflectivity method published by Fuchs and Müller (1971) was applied all over the world. There is hardly an interpretation published during the 1970s in which the model has not been checked and verified by calculating synthetic seismograms and comparing them with the main observed phases. Braile and Smith (1975) published a whole series of synthetic record sections calculated for typical crustal models. Bessonova et al. (1974) had derived the tau method to invert seismic wide-angle reflection traveltimes.

However, a serious drawback of this methodology was that a horizontally homogeneous layered structure was required. Several, and not always successful, attempts were undertaken

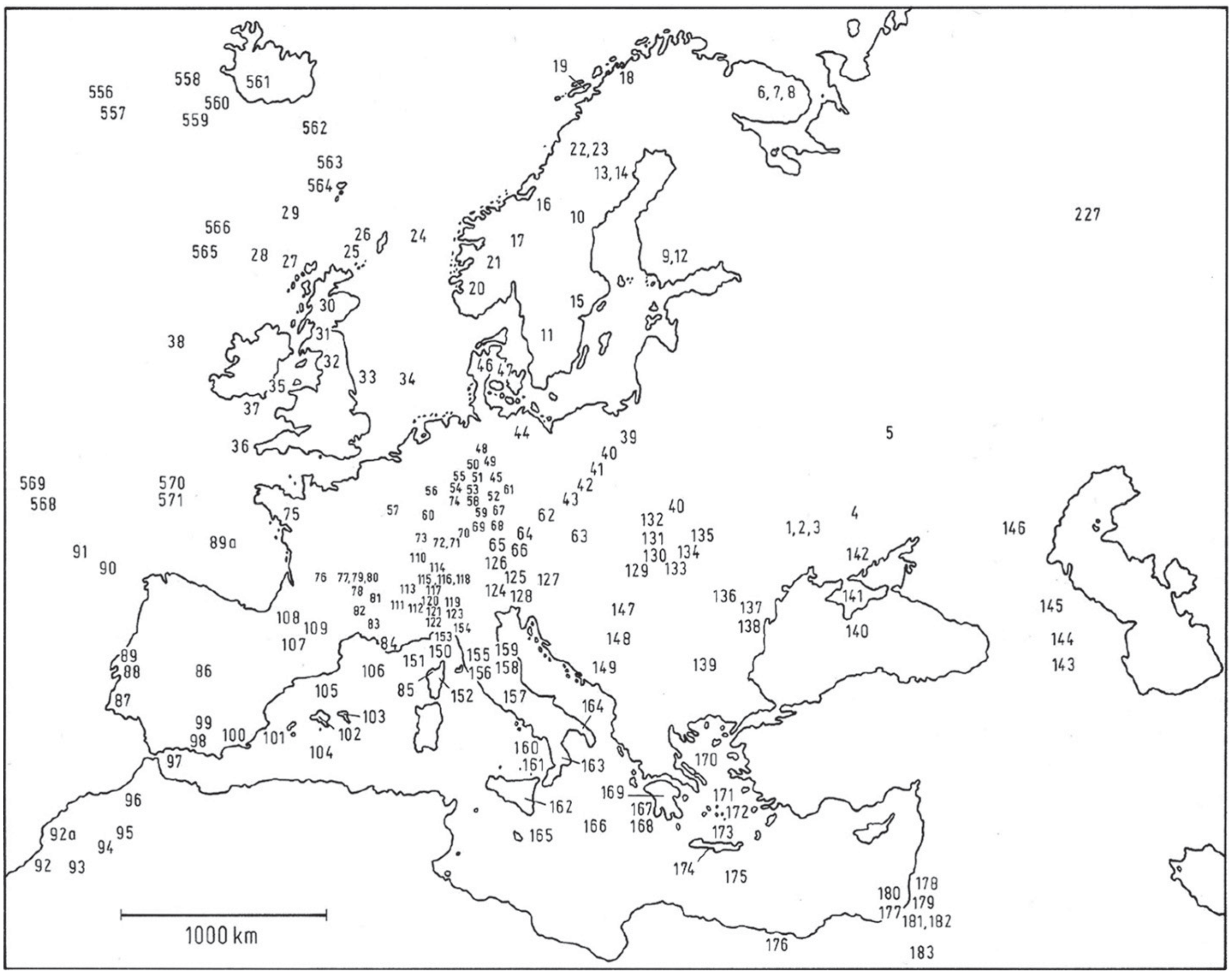


Figure 7.10-02. Map of Europe showing points where seismic data on crustal structure were available by 1980–1981 (from Prodehl, 1984, fig. 11). [*In* K.-H. Hellwege, editor in chief, Landolt Börnstein New Series: Numerical data and functional relationships in science and technology. Group V, Volume 2a: K. Fuchs and H. Soffel, eds., Physical properties of the interior of the earth, the moon and the planets: Berlin-Heidelberg, Springer, p. 97–206. Reproduced with kind permission of Springer Science+Business Media.]

to overcome this problem. For example, B. Kennett, amongst others, modified the reflectivity method to allow slight variations in structure on the shot side (Kennett, 1974, 1983).

The problem of laterally homogeneous structures was only solved when the first ray tracing approaches were published. In Munich, H. Gebrande was amongst the first who developed this approach which finally was successfully applied on Alpine data (Gebrande, 1976; Will, 1976). A breakthrough of this approach was made when Červený and co-authors published their book and later results on ray theory (Červený et al., 1977; Červený, 1979; Červený and Horn, 1980). A corresponding computer program for ray tracing had been prepared and made available for all interested users around the world, before it was finally published by Červený and Pšenčik (1984c). This procedure would become the major interpretation tool of the following decades. Červený's method and the computer routines developed on its basis also allowed researchers to compute synthetic seismograms. Other groups were also writing synthetic-seismogram programs on the basis of ray theory which were widely applied, for example, the program by McMechan and Mooney (1980) in the United States or the program by Spence et al. (1984) in Canada and elsewhere.

In the USSR, when deep seismic sounding operations were drastically reduced in the 1970s, the method of interpretation of deep seismic data underwent a period of heated discussions (Pavlenkova, 1996). Up until the 1960s, the correlated later-arrival phases had been interpreted as head waves, and only gradually were the phases correlated and recognized to be reflected waves. In the 1970s, new methods and interpretation programs were developed, and the methods of field experimentation were also changed. Numerous papers on wave dynamics and the nature of the recorded waves had already appeared in the 1960s, and the discussion continued in the 1970s. I.P. Kosminskaya had published a book on seismic methods in 1968, which was translated into English by G.V. Keller in 1971 (Kosminskaya, 1971). In the 1970s, new facilities for processing wave fields were applied to reinterpret a large part of the old deep seismic sounding material.

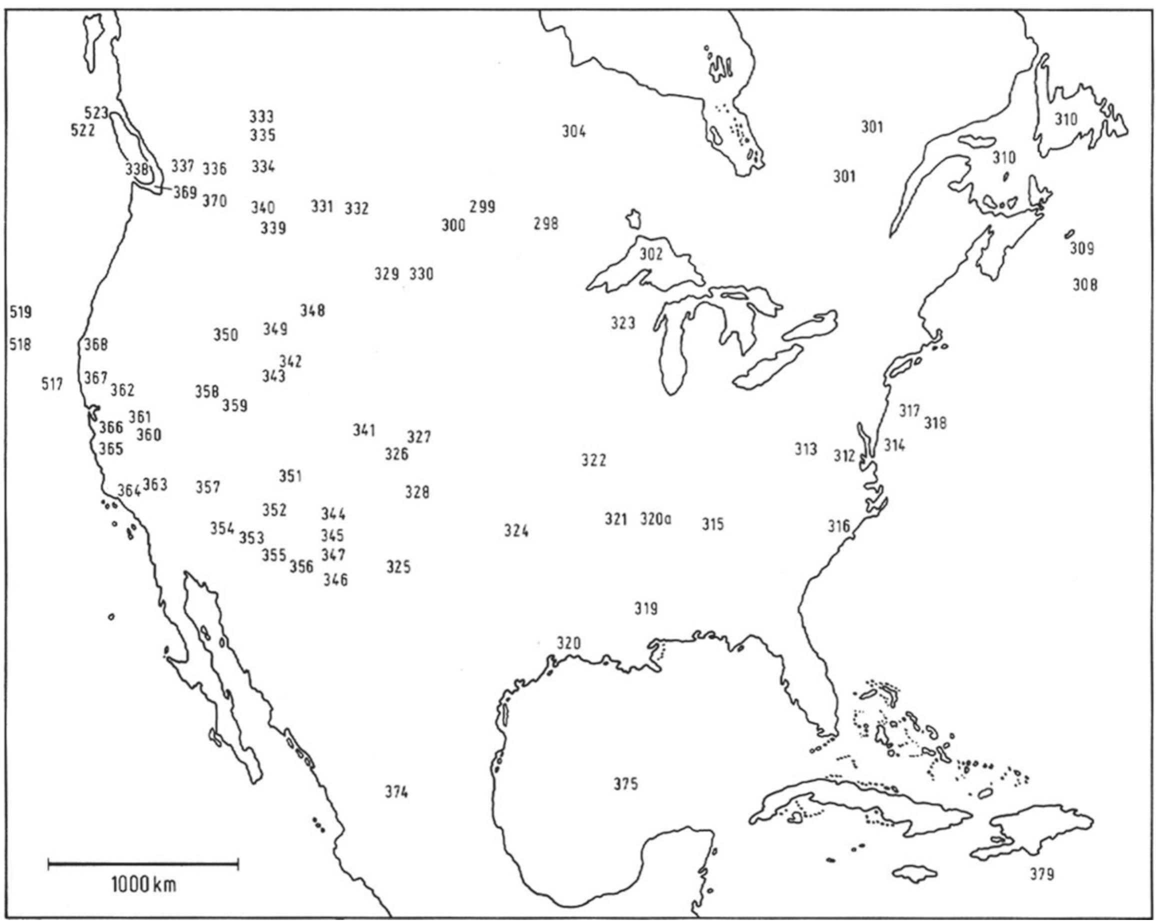


Figure 7.10-03. Map of the United States of America showing points where seismic data on crustal structure were available by 1980–1981 (from Prodehl, 1984, fig. 12). [*In* K.-H. Hellwege, editor in chief, Landolt Börnstein New Series: Numerical data and functional relationships in science and technology. Group V, Volume 2a: K. Fuchs and H. Soffel, eds., Physical properties of the interior of the earth, the moon and the planets: Berlin-Heidelberg, Springer, p. 97–206. Reproduced with kind permission of Springer Science+Business Media.]

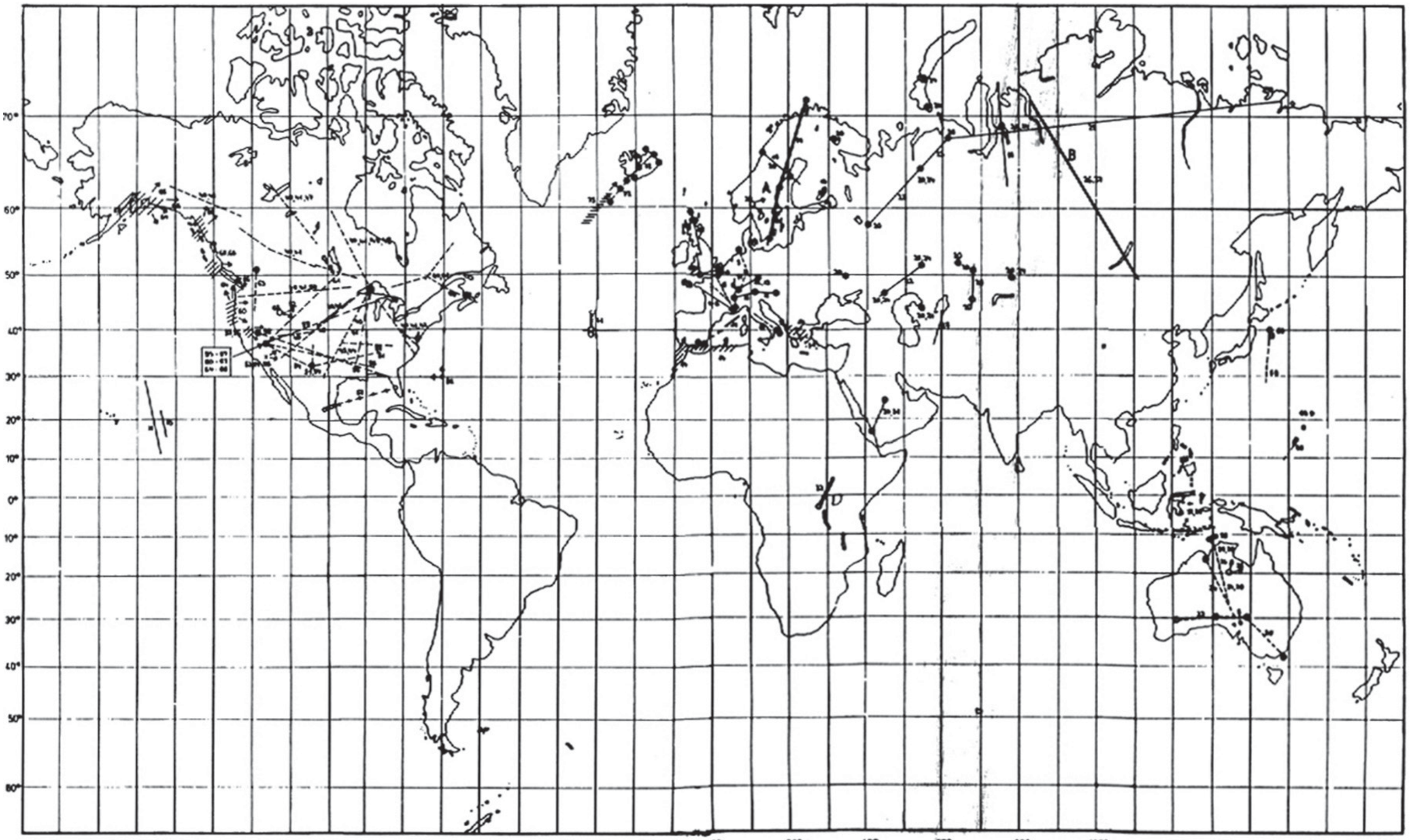


Figure 7.10-04. Enlarged section of the world map of Figure 2.7-01 showing long-range profiles until 1980 (from Fuchs et al., 1987, figure 1). [*In* Fuchs, K., and Froidevaux, C., eds., Composition, structure and dynamics of the lithosphere-asthenosphere system: American Geophysical Union Geodynamics Series 16, p. 137–154. Reproduced by permission of American Geophysical Union.]

In their book *Seismic Models of Basic Geostructures of the Territory of the U.S.S.R.*, Zverev and Kosminskaya (1980) presented a comprehensive set of reinterpreted—and representative—cross sections for various tectonic provinces of the USSR. As the scientists of the USSR at that time had very few opportunities to exchange ideas with their western colleagues, many approaches, algorithms, programs and methods of ray and wave field calculations in two- or three-dimensional media, as well as inversion methods for one- or two-dimensional media were developed in parallel and sometimes synchronously to those developed in the western world by, e.g., Alekseev, Bessonova, Matveeva, Michailenko, and others (for references, see Pavlenkova, 1996).

The problem of interpreting seismic-refraction data was internationally recognized. A number of workshops tried to bring together experimental and theoretical seismologists to discuss the various interpretation procedures and compare their results. Here, we mention only a few such workshops, where scientists of the whole world met to view the data and discuss the interpretation methods and their problems. So, in 1977, J.G. Heacock arranged a workshop on the Earth's crust at Vail, Colorado (Heacock, 1977). Czech scientists organized a series of workshops of differing character in 1977 and 1978 on problems of crustal studies and interpretation methods at the castle of Liblice near Prague. A year later, in 1978, I.P. Kosminskaya, N.I. Pavlenkova, and others arranged a workshop at Yalta on deep seismic sounding data and interpretation methods.

A special series of workshops starting in 1977 under the auspices of the IASPEI, specialized on problems of controlled-source seismology. These so-called CCSS workshops continue at non-regular intervals. For these workshops, data are distributed to the participants several months beforehand, and the scientists are requested to interpret the data with their presently used methodology. Thus, the various methods can be tested and verified during the discussions at these workshops.

The first CCSS workshop of this kind was organized in Karlsruhe, Germany, in 1977. Here, data from some unreversed seismic-refraction crustal profiles had been distributed to the participants, who were asked to bring their modeling results to the workshop (Fig. 7.9-01). One of the results was that the main differences between the models were not due to the interpretation method, but to phase correlations. The varying approaches and resulting models were finally published by Ansorge et al. (1982).

As data and modeling were tested using an one-dimensional approach only (not using any reversed shots), a second CCSS workshop followed three years later, organized by R.B. Smith at Park City, Utah, in August 1980. As a database for this workshop, the data of the 1978 refraction campaign of the U.S. Geological Survey in Saudi Arabia were distributed to the participants. This data allowed a two-dimensional approach to the interpretation. The project was already discussed in more detail above (see Figs. 7.5.2-04 to 7.5.2-06 and Appendix A7-6-3). Fourteen different models were presented at the workshop and were published in a U.S. Geological Survey Circular (Mooney and Prodehl, 1984). As an example, the interpretation between shotpoints 3 and 4 is shown (Fig. 7.9-02). Several of the models were also published independently in *Tectonophysics* (e.g., Mooney et al., 1985; Milkereit and Flueh, 1985; Prodehl, 1985).

7.10. THE STATE OF THE ART AT THE BEGINNING OF THE 1980s

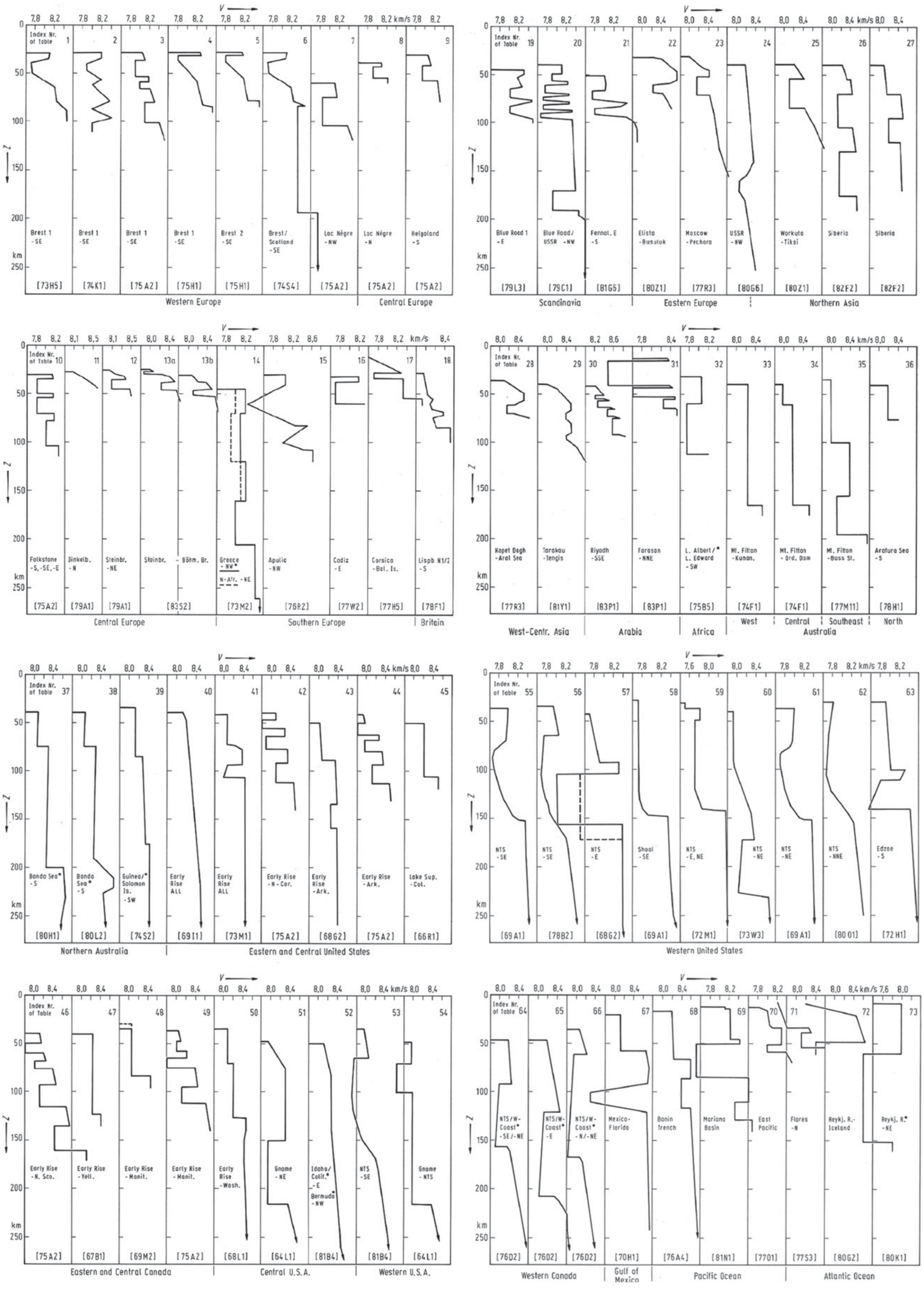
For the first time, in the 1970s, seismic near-vertical incidence reflection experiments were organized in large scales covering distances of 100 km or more which brought new insight into details of crustal structure and composition of the Earth's crust down to Moho. Sedimentary basins on the continents and ocean basins at sea had already been studied in much detail, the former by petroleum seekers since the 1930s and the latter by marine scientists largely during the decades following World War II. For the study of sedimentary basins, the seismic-reflection profiling method had been highly developed over the years by the petroleum industry. Consequently, academic researchers took advantage of the expertise, techniques, and equipment of the industry to study the whole crystalline crust and obtain not only geophysical models with velocity, density, and attenuation, but also determine the extent and the configuration of reflecting horizons (Oliver, 1986).

So, in the 1970s, the first national large-scale seismic-reflection programs were initiated. COCORP in the United States (e.g., Oliver et al., 1976) was soon followed by the Canadian equivalent COCRUST (Mereu et al., 1989), a combination of large-scale crustal reflection and refraction studies. Similar large-scale seismic near-vertical incidence reflection profiles, accompanied by wide-angle reflection observations, were recorded in Germany (Bartelsen et al., 1982; Meissner et al., 1980).

Seismic-refraction work in the 1970s was for a large part devoted to research projects dealing with the subcrustal lithosphere, but also much emphasis was laid on crustal research in tectonically anomalous regions. For western Europe, Mostaanpour (1984) compiled the record sections of seismic-refraction profiles of the past 30 years, mapped the characteristic parameters, and plotted a series of contour maps covering mainly Germany, the Alps, France, and Italy, following a working scheme applied by Giese and Stein (1971) for western Germany.

A representative selection of lithospheric velocity-depth sections from data obtained during the 1970s was compiled by Guggisberg et al. (1991) when they were discussing the results of the 2000-km-long FENNOLORA profile through Scandinavia (Fig. 7.10-01).

Figure 7.10-05. Selected velocity-depth models of upper-mantle structure below Moho for Europe and Asia (from Prodehl, 1984, fig. 15). [*In* K.-H. Hellwege, editor in chief, Landolt Börnstein New Series: Numerical data and functional relationships in science and technology. Group V, Volume 2a: K. Fuchs and H. Soffel, eds., Physical properties of the interior of the earth, the moon and the planets: Berlin-Heidelberg, Springer, p. 97–206. Reproduced with kind permission of Springer Science+Business Media.]


Western Europe
Central Europe
Scandinavia
Eastern Europe
Northern Asia
Central Europe
Southern Europe
Britain
West-Centr. Asia
Arabia
Africa
West
Central
Southeast
North
Australia
Northern Australia
Eastern and Central United States
Western United States
Eastern and Central Canada
Central U.S.A.
Western U.S.A.
Western Canada
Gulf of Mexico
Pacific Ocean
Atlantic Ocean

In 1981, a global crustal thickness map was published by Soller et al. (1981), based on the results of 273 publications.

An overview of crustal and upper-mantle studies by seismic-refraction experiments has been compiled by Prodehl (1984; see Appendix A7-1), essentially representing the state of the art by 1980–1981. The summary shows representative velocity-depth functions for the crust in tabular form and the corresponding locations on maps for Europe, North America, and the rest of the world. The map for the whole world (see Fig. 2.7-03 in Chapter 2) as well as the more detailed maps for Europe (Fig. 7.10-02) and the United States of North America (Fig. 7.10-03) show very well where data on crustal structure was available at the time. The second part of Prodehl's (1984) compilation also dealt with the subcrustal data (Fig. 7.10-04) and the models (Fig. 7.10-05), as they were published by 1980–1981.