Geological Society of America Memoirs

# CHAPTER 9 The 1990s (1990–2000)

*Geological Society of America Memoirs* 2012;208;443-652
doi: 10.1130/2012.2208(09)



**Notes**



THE
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CHAPTER 9

# *The 1990s (1990–2000)*

## 9.1. THE FIRST DECADE OF DIGITAL RECORDING

With advanced technology, controlled-source seismology had already reached a stage in the 1980s where its models were no longer schematic diagrams, but approached realistic cross sections and sometimes even 3-D models, which attracted geoscientists from other fields and led them to accept seismic models as a real picture of geologic-tectonic features. In particular, the recently obtained reflection seismograms of large dimensions covering the whole Earth's crust along hundreds of kilometers down to the Moho were simulating the complicated realistic features of the real Earth. At the same time, however, seismic projects had become more and more expensive, using multiple energy sources and a large number of sophisticated recording devices, with the change from analogue to digital recording at the turn of the decade causing a major breakthrough toward modern recording and interpretation techniques.

The combination of these factors meant that most projects were no longer those of individual researchers or individual research institutions, but became mostly embedded in large-scale research programs which involved a multitude of cooperating institutions and an interdisciplinary cooperation among scientists of various geoscientific fields. Starting in the 1980s and continuing in the 1990s, this was reflected, e.g., in the formulation of large reflection programs on national scales such as COCORP (Phinney and Roy-Chowdhury, 1989; Smithson and Johnson, 1989), BIRPS (Klemperer and Hobbs, 1991), ECORS (Bois and ECORS Scientific Parties, 1991), DEKORP (Meissner and Bortfeld, 1990), etc., which was followed by the foundation of IRIS/PASSCAL (Incorporated Research Institutions for Seismology/Program for Array Seismic Studies of the Continental Lithosphere) in the United States and LITHOPROBE in Canada, and by international and interdisciplinary geoscientific priority programs in Europe, dealing with large-scale tectonic topics.

Here, in 1992, supported by the European Science Foundation, EUROPROBE was founded, a Lithosphere Dynamics program, concerned with the origin and evolution of the continents (Gee and Stephenson, 2006). EUROPROBE's focus was mainly, but not restricted, on Europe and was particularly driven to emphasize and encourage East-Central-West European collaboration. It was dedicated to enable the realization of major projects, which aimed to investigate the whole lithosphere and which required close multinational cooperation of geologists, geophysicists and geochemists. Particular EUROPROBE projects, which involved major controlled-source seismic experiments, were the "TESZ" project, investigating the Trans-European fault zone Tornquist-Teysseire Zone (TTZ) in Scandinavia and Poland, the "Uralides" with a major seismic campaign in the Urals, "Georift," emphasizing the Dniepr-Donets basin, "Eurobridge," establishing a seismic crustal traverse from Lithuania to the Ukraine, and "Pancardi," an interdisciplinary geoscientific frame for large-scale investigations in the Carpathian area, in particular focusing on the deep-earthquake region of Vrancea in the Romanian Carpathians. Other large-scale interdisciplinary projects in Europe were the finalization of the deep-drilling project in Germany, a seismic-reflection traverse through the Eastern Alps, a German priority program focusing on the tectonics of the Central European Variscides, and a major research program studying the Variscan front in Ireland, to name only a few examples.

An important role for successful worldwide cooperation in seismic projects was the development of new digital equipment which had started by the end of the 1980s and continued in the 1990s both in North America and in Europe. In particular, the North American equipment not only stimulated and enabled new large-scale seismic-refraction/-reflection experiments in Canada and in the United States, but would also enable many large-scale projects in Europe. As will be described in detail in the individual subchapters of this chapter, e.g., the IRIS/PASSCAL instrument pool, starting with the U.S. Geological Survey–Stanford instrumentation Seismic Group Recorders and continuing with the powerful RefTek generation, was fundamental for successful joint European-U.S. projects in Kenya, France, and Spain in the early 1990s. The Canadian equipment PRS-1 would be the fundamental equipment for seismic-reflection work in Hungary in 1992. Finally, at the end of the 1990s and in the beginning of the 2000s, large-scale projects in Eastern Europe and in Ethiopia would not have been possible without the IRIS/PASSCAL instrument pool. Vice versa, European participation supported several seismic projects in North America.

The continuing rapid progress of data collection and interpretation required the continuation of continuous scientific exchange which was only partly met by the annual meetings of the various national and international geoscientific organizations, such as, e.g., the American Geophysical Union or the European Geophysical Union. Other special meetings of earth scientists were also required.

Since the 1960s, a special subcommission of the European Seismological Commission, which meets every two years, dealt in particular with exchange of knowledge concerning fieldwork, interpretation methods, and results of explosion seismic data. In the early 1980s, the series "International Symposia on Deep

Seismic Reflection Profiling of the Continental Lithosphere" was started, which continued into the 1990s and beyond, with meetings at Banff, Canada, in 1992 (Clowes and Green, 1994), Budapest, Hungary, in 1994 (White et al., 1996), Asilomar, California, in 1996 (Klemperer and Mooney, 1998a, 1998b), Platja d'Aro, Spain, in 1998 (Carbonell et al., 2000a), Ulvik, Norway, in 2000 (Thybo, 2002), and in Taupo, New Zealand in 2003 (Davey and Jones, 2004). Also the irregularly scheduled CCSS (Commission on Controlled-Source Seismology) workshops brought together scientists to discuss and practice interpretation methods on pre-distributed data sets. They were organized under the umbrella of IASPEI (International Association of Seismology and Physics of the Earth's Interior) since the late 1970s and continued in the 1990s, e.g., in 1993 in Moscow, Russia (Pavlenkova et al., 1993), in 1996 in Cambridge, UK (Snyder et al., 1997), and in Dublin, Ireland, in 1997 (Jacob et al., 1997) and 1999 (Jacob et al., 2000).

### 9.1.1. The North American Controlled-Source Seismology Community

The development of digital instrumentation for the controlled-source seismology community in North America had started in the mid-1980s. It had been initiated by increased crustal and upper mantle research activity of the U.S. Geological Survey (USGS) and the Geological Survey of Canada, and, involving all North American university groups, by the initiation of the LITHOPROBE project in Canada, and by the founding of IRIS in the United States in 1984 with its sub-organization PASSCAL, which was funded by the U.S. National Science Foundation. So, at the beginning of the 1990s a large set of digital instrumentation was available.

The Portable Refraction Seismograph (PRS-1), which, as part of LITHOPROBE, was developed and acquired by the Geological Survey of Canada and afterwards built by EDA Instruments Ltd., was the first digital recording device, which came into use in a large quantity in the late 1980s. Its involvement in large cooperative U.S.–Canadian projects was already described in Chapter 8. The PRS-1 was a single-channel instrument that used a Mark Products L-4A 2-Hz vertical-component geophone. Automatic gain-ranging from 1 to 1024 in binary steps allowed a total dynamic range of 132 dB. Seismic data were sampled at 120 samples per second by a 12-bit A/D board and stored in memory until the data were uploaded to a computer (Murphy et al., 1993; see also Appendix A8-3-8, OFR 93-265). This type of instrument was later complemented by a three-component version, the PRS-4 unit.

Toward the end of the 1980s, the oil industry (AMOCO) had donated 200 units of Seismic Group Recorders (SGR) to Stanford University (Figs. 9.1.1-01 and 9.1.1-02). This SGR III was a single channel, digital seismic recorder with a dynamic range of 156 dB. Data were sampled at 500 samples per second by a 12-bit A/D board with gain ranging from 0 to 90 dB in six steps. Originally these units, manufactured by Globe Universal

Figure 9.1.1-01. Seismic Group Recorders being prepared at a warehouse in Nairobi for the KRISP90 experiment (Prodehl et al., 1994b) in Kenya. (Photograph by C. Prodehl.)

Figure 9.1.1-02. Seismic Group Recorders at a field test during the KRISP90 experiment (Prodehl et al., 1994b) in Kenya. (Photograph by C. Prodehl.)

Sciences, Inc. (a division of Grant-Norpac, Inc.), were designed to record seismic data on cartridge tape under the control of a radio signal.

In partnership with the USGS and IRIS/PASSCAL, the Stanford SGRs were modified by adding a timer board that could trigger the recording of data in place of the radio control signal. Each timer board had a precision crystal clock and was loaded with a schedule of trigger times from a PC or laptop to allow programmable turn-on as well as radio-triggering. The clock, a temperature-compensated internal oscillator (TXCO), was synchronized to a master clock prior to deployment, so that the units could individually be installed in the field. The fixed sample rate of 2 ms, maximum record length of 99 s, and maximum recording capacity per cartridge of ~20 min was ideal for then-conventional refraction recording, but greatly limited applicability for roll-along reflection experiments, or recording of offshore airgun sources.

Each SGR III unit was connected to a single string of 6 modified Mark Products L-10B vertical-component geophones (8 Hz) connected in series. Later in the 1990s, 2-Hz Mark Product geophones were also used (Murphy et al., 1993; for more details, see Appendix A8-3-8, OFR 93-265).

In the 1990s, until 1998, the SGR became a workhorse instrument for the community thanks to cooperation between Stanford University, the USGS, and PASSCAL. Figures 9.1.1-01 and 9.1.1-02 show the SGRs at a survey in Kenya (Prodehl et al., 1994b).

The latest development in the late 1980s ready for use by 1990 was the RefTek 72A-02 instrument which was developed by RefTec Inc. and was being bought by PASSCAL. These instruments were six-channel digital seismic data acquisition systems with a 200 Mbyte data disk and an OMEGA clock. The instruments had a variable-gain preamplifier which could be set between a gain of 1 and a gain of 8192. During the 1990 Alaska campaign, for example, the gain was set to 256. Cables were available to be connected to six sets of Mark Products L-10B 8-Hz vertical-component geophone strings or to L-4 2-Hz vertical- or three-component geophones (Murphy et al., 1993; see also Appendix A8-3-8, OFR 93-265). For the fieldwork in Alaska in 1990, where, besides the analogue cassette recorders of the USGS, all three digital types of instruments described above were in the field, 35 RefTeks were available (see Chapter 8, section 8.5.4; Fuis et al., 1997).

Recognizing the controlled-source community's need for a simple instrument for their experiments, PASSCAL worked with RefTek to design a three-component version of the main RefTek instrument (Fig. 9.1.1-03). However, this instrument required heavy batteries, external GPS clocks for many applications, and was far from simple. To complement the PASSCAL passive-seismology instrument center at Lamont, an active-seismology PASSCAL center was set up at Stanford University, from 1991 (50 RefTeks, 1 staff member) to 1998 (320 RefTeks, 3 60-channel Geometrics, 4 staff members plus interns). Though not a simple instrument, the RefTeks were reliable, and data recovery was routinely >90%, in cases >98%, even on international, multi-deployment experiments. More than 20 RefTek experiments a year were supported from this center. In 1998, PASSCAL relocated both instrument centers to New Mexico Tech in a new purpose built facility that has grown to manage the Earthscope U.S. Array instrumentation (Keller et al., 2005b).

Figure 9.1.1-03. RefTek 3-component equipment in the field during the CD-ROM experiment (Rumpel et al., 2005; Snelson et al., 2005) in the Rocky Mountains. (Photograph by C. Prodehl.)

By pooling various combinations of the Canadian instruments, the RefTeks, the Seismic Group Recorders (SGR), and the Seismic Cassette Recorders (SCR), a number of large experiments were carried out. In the early 1990s, the typical large experiment was 200 SGR, 200 PRS-1, and 200 RefTek (e.g., Mendocino 1993). The SSCD (Southern Sierra Continental Dynamics), DELTA FORCE and DEEP PROBE experiments were the last to use the SGRs (Keller et al., 2005b).

A new generation of recorders designed primarily for use in controlled-source experiments became available in 1999. The RefTek 125 (Texan) instruments became available via a combination of grants to the University of Texas at El Paso. Though only a one-component instrument, due to its small size and weight it has become a very powerful tool for fast-moving large-scale experiments (Fig. 9.1.1-04). Its weight is less than 1.4 kg, the power consumption is small (150 mW in recording and 25 mW in sleeping mode), allowing to be operated by 2 D-cell batteries for 5–8 days, it has an adequate memory (the initial standard was 32 Mbyte) to record continuously for at least 32 h at a 10 ms sampling interval, and it has a high-precision clock (0.1 ppm accuracy temperature-compensated crystal oscillator) for accurate timing of reflection-seismic data. Experience showed that the clock drift rate was 5–7 ms per day.

The initial grant was from the Texas Higher Education coordinating board and funded collaboration between the University of Texas at El Paso and RefTek to design this instrument. Broad input from the international community and PASSCAL was sought during this process. Thanks to an MRI grant from NSF, 440 instruments were available by mid-1999, and during this year, 5 large experiments (3 with 400 instruments) were undertaken (Keller et al., 2005b).

Figure 9.1.1-04. RefTek 1-component equipment ("Texan") during the VRANCEA-2001 experiment (Hauser et al., 2002) in Romania. (Photograph by C. Prodehl.) The transport boxes (right) contain 15 recorders.

### 9.1.2. The European Controlled-Source Seismology Community

The European Geotraverse was the last large international experiment in Europe that had used MARS-66 analogue equipment on a large scale. However, also in Europe, controlled-source seismic projects as well as a gradually increasing number of teleseismic projects for studies of crust and upper-mantle structure called for new instrumentation which would also allow to record over a great length of time. However, in contrast to North America, and in spite of the previous success of the European-wide distributed MARS66 system, a new European initiative did not get established. Thus, individual national initiatives were started, but for many years with very limited success. For example, since the early 1980s, in Germany a working group with members from several German universities had started to design the needs for a modern recording device as a replacement for the worn-out analogue MARS66 system, and a first digital system was being developed, but never came to completion. Various companies offered the so-called PCM recording devices which had a great storage capacity and could be left in the field for several days and record earthquake activity. However, these instruments were complicated to use and expensive, so that only very limited numbers of instruments could be purchased, mainly for seismology purposes and teleseismic tomography projects. It was finally the North American initiative, which brought about various products becoming available for the European community by the beginning of the 1990s. In 1991, for example, hardware and software tests of different equipment, offered by the companies Lennartz (Germany) and EDA (Canada) as well as by the U.S. companies Geotech-Teledyne, Kinemetrics, RefTec Inc., and Sprengnether, were arranged at Munich, and a recommendation was made as to which instrument might best serve the German research community. However, it took another couple of years before a major number of new digital instruments became available.

After the unification of Germany, the former East German "Central Institute of Earth Science" at Potsdam had been overtaken by the Federal Republic of Germany and reorganized into the "GeoForschungsZentrum Potsdam" (GFZ, GeoScience-Center Potsdam). It was this new research facility, which finally obtained sufficient funding to establish a new German digital equipment pool which, similar to the PASSCAL pool, would become available for all research institutions in Germany. In 1993, GFZ had purchased a larger number of PDAS equipment from Teledyne and was going to purchase a larger set of RefTek equipment, so that the first research projects with new GFZ digital recording equipment could go ahead from 1993 onward.

In 1993, a first cooperative project INSTRUCT of GFZ, DEKORP, and the University of Kiel was undertaken around the deep continental drilling site KTB in Germany with 14 PDAS. However, the first major projects with the new pool were the PISCO experiment in South America (60 PDAS; see section 9.7.3 below), the FRAC project of the University of Bochum with 25 PDAS, both in 1994 and, in 1995, the GRANU-project with 126 stations (61 PDAS and 65 RefTeks; Fig. 9.1.2-01 and 9.1.2-02; see section 9.2.1 below).

In 1995, a working group was established consisting of GFZ and university delegates who would meet semiannually to deal with the use of the equipment on application by German and other researchers. At that time, 59 PDAS-100 (Fig. 9.1.2-01; Table 9.1.2-01), 66 RefTek-72A/07 (Table 9.1.2-01), and 5 MARS-88 digital recording devices had become available, and 107 short-period three-component 1Hz MARK-L4 seismometers and 21 broadband sensors had been purchased.

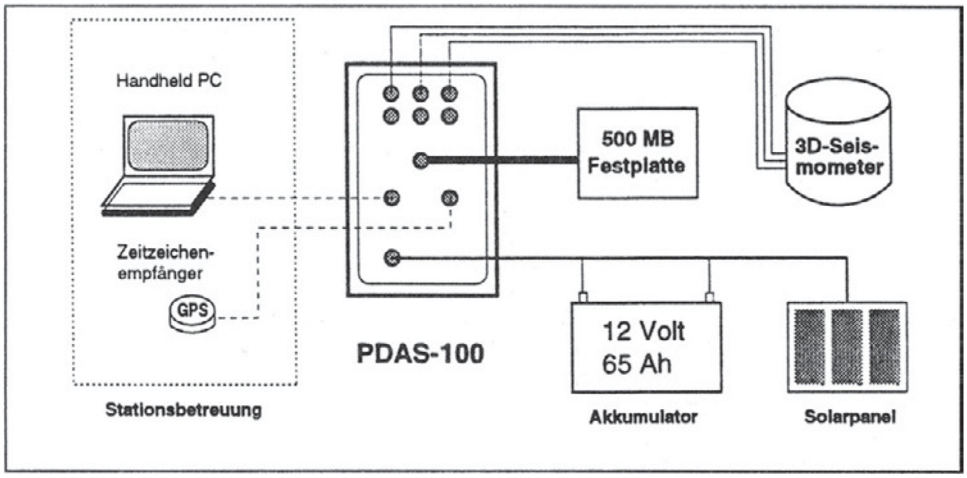



Figure 9.1.2-01. PDAS-100 digital recording device of GeoForschungsZentrum (GFZ) Potsdam (from Lessel, 1998, fig. 4.4), used both for controlled-source and seismology experiments. [Berliner Geowissenschaftliche Abhandlungen, Reihe B, Band 31, 185 p. Published by permission of Institut für Geologische Wissenschaften, Freie Universität Berlin.]

Figure 9.1.2-02. PDAS-100 digital recording device of GeoForschungs-Zentrum (GFZ) Potsdam, being prepared for fieldwork at the GRANU-95 experiment (Enderle et al., 1998) in eastern Germany. (Photograph by C. Prodehl.)

TABLE 9.1.2-01. TECHNICAL SPECIFICATIONS OF THE SEISMIC-REFRACTION EQUIPMENT FROM THE GEOPHYSICAL INSTRUMENT POOL OF THE GEOFORSCHUNGSZENTRUM (GFZ) POTSDAM

| Model: | RefTek 72A-07 | PDAS-100 |
|---|---|---|
| Description | Digital recording systems for standalone use, time window and trigger mode, final storage on hard disk | |
| Manufacturer | Refraction Technology, Dallas, Texas | Teledyne Brown Engineering, Dallas, Texas |
| Availability | Geophysical instrument pool of the GFZ Potsdam, available for academic use | |
| Sample rate | 20, 25, 40, 50, 100, 125, 200, 250, 500 or 1000 samples per second, adjustable | 0.1, 0.2, 0.5, 1, 2, 5, 10, 20, 50, 100, 200, 500, 1000 samples per second, adjustable |
| Max. recording | Variable length | |
| Anti-aliasing | 6pole Butterworth, 200 Hz cut-off | |
| Digital filter | Linear phase finite impulse response; pass band up to 82% Nyquist, –130 dB at Nyquist | Linear phase finite impulse response, pass band up to 80% Nyquist, –120 dB at Nyquist |
| Pre-amp | Unity and 32, adjustable settings | Unity, 10 and 100, adjustable |
| Programmability | 99 time windows of different length | 10 time windows of different length, expandable to 9999 repetitions each |
| Clock drift | Max. 5 ms due to sync. by GPS each hour, lOe-6 or about, 100 ms per day if GPS fails | 7*lOe-8 (DCXO type), 2*lOe-7 (TCXO type), <10 ms due to sync. by GPS every 4 hours |
| Power consumption | 2.7 to 3 Watts (3 channels, 100 sps each) | 3 to 3.6 Watts (3 channels, 100 sps each) |
| Disk capacity | 500 MB and 1 GB | 540 MB |
| Seismometers | Mark L-4C-3D, 1Hz, three components | |
| Operating temperature | –20 to +60 °C | –20 to +65 °C |

*Note:* From Geophysics Journal International, v. 133, p. 245–259. Copyright John Wiley & Sons Ltd.

Concerning the new Texan instruments, of which the PASSCAL pool and the University of Texas at El Paso in the United States had a total of 840 by 2000, the University of Copenhagen, Denmark, purchased 100, the Earth Science Research Institute TUBITAK-MAM in Gebze, Turkey, acquired 60, and other groups in Central Europe obtained another quantity of ~40 units (Guterch et al., 2003a).

Also in Britain, PDAS and RefTek instrumentation was acquired by the NERC Geophysical Equipment Pool and by various universities, e.g., Leicester University, University of East Anglia, and Cardiff University. Some years later, Leeds University purchased the Orion system. In 2000, finally, the British seismic community (Leicester, Cambridge, Leeds, and Royal Holloway) was awarded a grant to acquire large numbers of Guralp 6TDs and 40Ts, together with a number of the Guralp 3Ts, bringing the British scientists to the forefront of observational broadband seismology.

## 9.2. CONTROLLED-SOURCE SEISMOLOGY IN EUROPE

### 9.2.1 The British Isles and Ireland

The fruitful cooperation of the Dublin Institute for Advanced Sciences with German partners from Hamburg and Karlsruhe, which had led to major seismic surveys in Ireland and the adjacent seas continued in 1995 with a 3-year-program named VARNET (Variscan Front Network) The project was funded by the European Community and aimed to study the deep structure of the Variscan Front in southern Ireland by seismic, seismological and magnetotelluric studies. For this purpose, in 1996, a seismic-refraction project was designed in southwestern Ireland, which crossed the Variscan-Caledonian boundary, the supposed Variscan Front (Fig. 9.2.1-01).

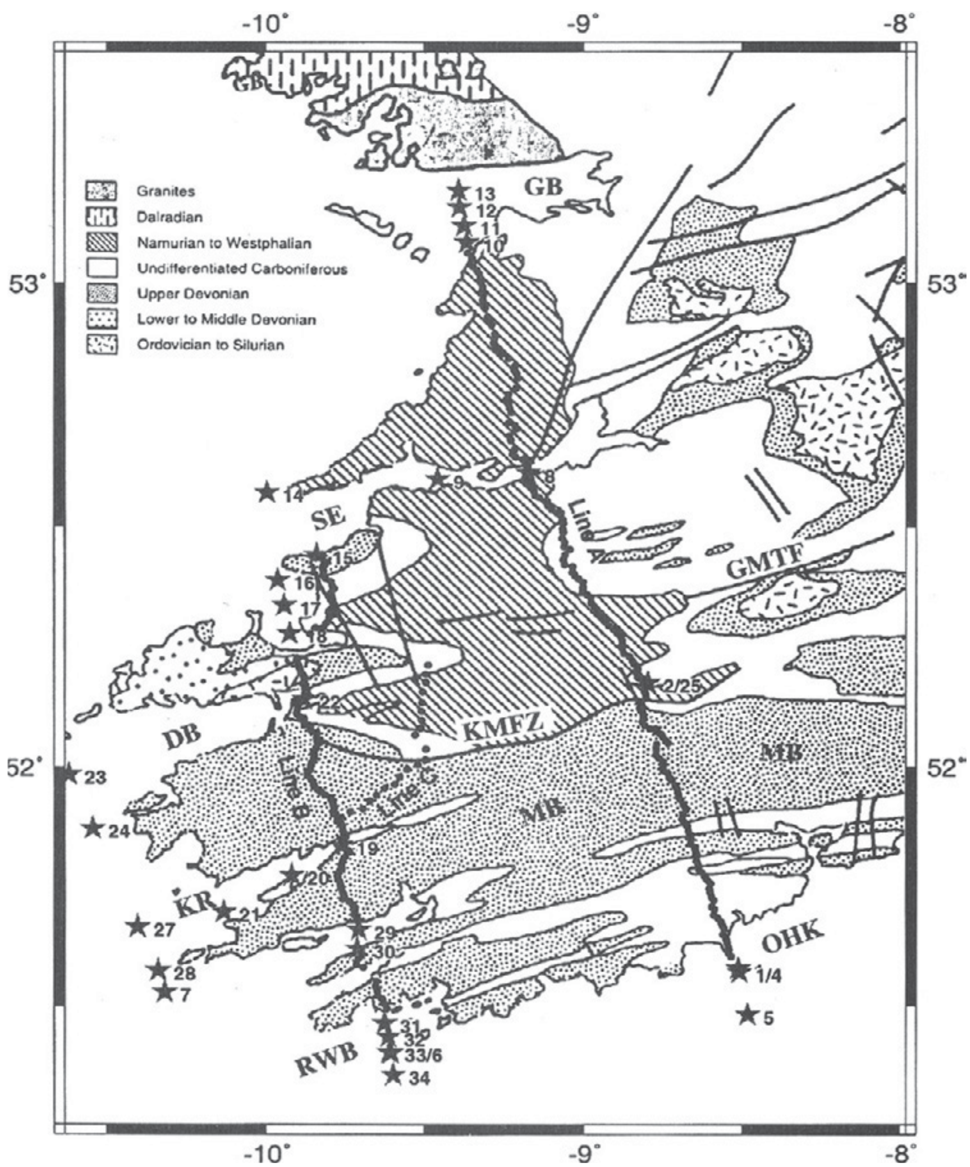



Figure 9.2.1-01. Simplified geological map of southwestern Ireland and location of the VARNET seismic-refraction profiles (from Landes et al., 1990, fig. 1). Stars—shotpoint locations. DB—Dingle Bay, GB—Galway Bay, SE—Shannon Estuary, KR—Kenmare River, MB—Munster Basin, OHK—Old Head of Kinsale, RWB—Roaringwater Bay, GMTF—Galtee Mountains Thrust Fault, KMFZ—Killarney-Mallow Fault Zone. [Terra Nova, v. 17, p. 11–120. Copyright Wiley-Blackwell.]

During this experiment, 34 shots were recorded by 170 stations deployed mainly along two approximately N-S–directed profiles of 140 and 200 km in length, extending from the southern coastline toward the Shannon Estuary and Galway Bay, respectively. The shotpoint geometry was designed to allow for both inline and offline fan shot recordings on the profiles to extend the 2-D interpretation of the inline data to three dimensions (Landes et al., 2000, 2003, 2005; Masson et al., 1998).

A series of shots at small intervals were fired at the northern and southern ends of lines A and B as well as on the western extension of line C. Along the western line B, a series of sea inlets made it possible to fire quite small shots (25 or 50 kg) in water in between the various land segments, so that interleaving of shots and stations provided a multicoverage of observations. An additional "land" shotpoint was added in the center of line A, using a lake in a drowned quarry. The P-wave data of the in-line shots recorded on lines A and B are shown in Appendix A2-1 (p. 64–68) and Appendix A9-1-1.

The large number of shots at the ends of line A allowed a well constrained model for this line (Landes et al., 2000), where particularly in the area of the Shannon estuary at the proposed site of the Iapetus Suture Zone (ISZ) a pronounced thickening of the middle crust resulted. For line B, a rather detailed picture of the upper and middle crust was established by Masson et al. (1998), while the data for the lower crust and upper mantle were rather scarce, due to the insufficient length of the profile. Therefore the Moho was not modeled for line B in that first interpretation. The models shown in Figures 9.2.1-02 and 9.2.1-04 are from a more recent compilation by Landes et al. (2005). The overall crustal thickness reached from 28 km near the coast in the south and west to 33 km in the center of southwestern Ireland (Landes et al., 2003).

In 1999, the latest onshore seismic-refraction experiment, LEGS (Leinster Granite Seismics), was carried out in the southeast of Ireland (Hodgson, 2001; Hodgson et al., 2000; for location, see Fig. 8.3.2-05 and inset of Fig. 9.2.1-04). Its main aim was the determination of the dimensions of a major granite batholith, outcropping as the Leinster Granite in this area, which according to gravity data should have a much larger subsurface extension toward the southwest. A total of 19 shots was recorded simultaneously along three profiles and recorded by 300 TEXAN stations of PASSCAL and the University of Texas at El Paso. The axial profile, which extended in the NE-SW direction, was, including its off-end shots, 230 km long. Two perpendicular profiles were each 110–140 km long. Station spacing was 1.0–1.5 km.

The resulting depth to Moho showed a decrease from 32 km to 28 km underneath the center of the main granite body, which itself appeared to be between 2 and 5 km thick (Figs. 9.2.1-03 and 9.2.1-04).

Landes et al. (2003, 2005) compiled all crustal investigations in Ireland and the surrounding seas into 3-D block diagrams (Fig. 9.2.1-04). The review of Landes et al. (2005) also includes seismic-tomography studies, the first of which was an outcome of the Varnet-96 program, as mentioned above, when some earthquakes and a nuclear event had also been recorded during the recording windows set for the controlled-source seismic experiment. These data could be used to model the probable trace of the Iapetus suture in the upper mantle just south of the Shannon river estuary between 30 and 110 km depth (Masson et al., 1999).

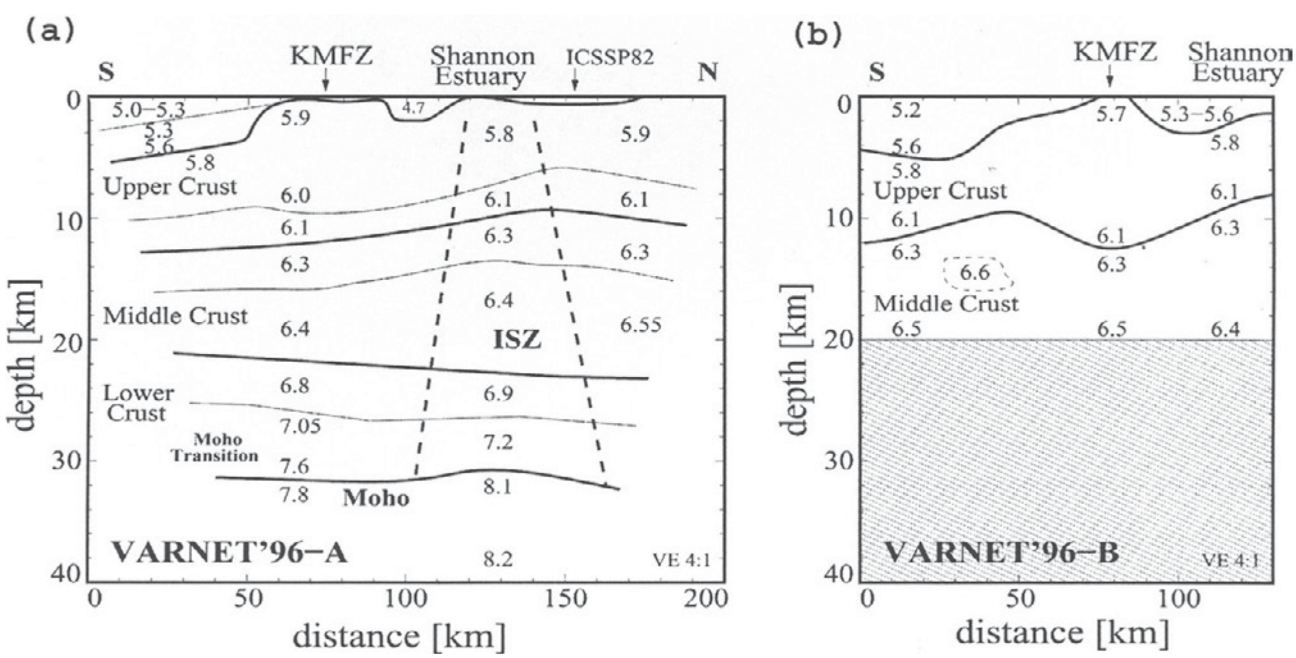



Figure 9.2.1-02. Crustal structure underneath southwestern Ireland (from Landes et al., 2005, fig. 6). KMFZ—Killarney-Mallow Fault Zone; ISZ—Iapetus Suture Zone. [Terra Nova, v. 17, p. 111–120. Copyright Wiley-Blackwell.]

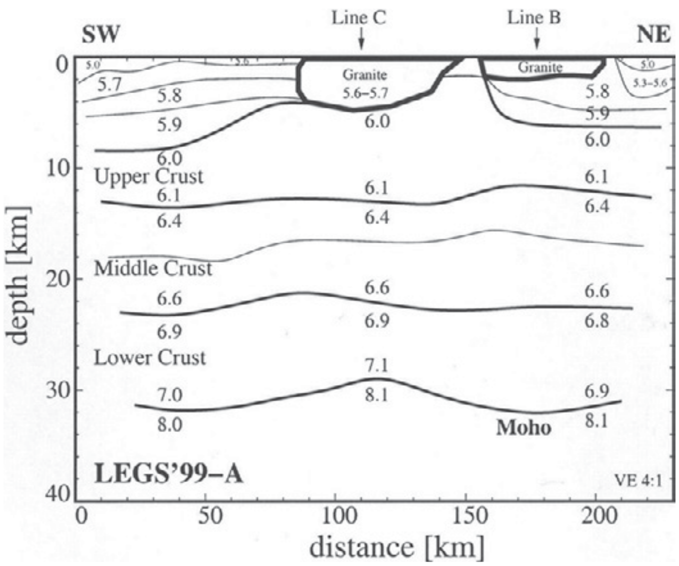



Figure 9.2.1-03. Crustal structure underneath the Leinster Granite region (from Landes et al., 2005, fig. 9): LEGS'99 profile A, modified after Hodgson (2001). [Terra Nova, v. 17, p. 111–120. Copyright Wiley-Blackwell.]

### 9.2.2. Central Europe

In Germany, research interest in the first half of the 1990s concentrated on two areas: the Variscicum in Saxony, eastern Germany, and the southern end of the Rhinegraben, in southwestern Germany. In the second half of the 1990s, major seismic research projects were undertaken in northeastern Germany and the adjacent Baltic Sea (DEKORP-BASIN) as well as in the Eastern Alps (TRANSALP).

In 1990, immediately following the approach of West and East Germany toward unification, geophysicists agreed to shoot a modern seismic-reflection line across the former border, exploring the Saxothuringian area of the central-European Hercynian mountain system from the Hessen depression in the west to the Erzgebirge in the east. The survey was named DEKORP 3/MVE-90 and covered a total of 600 km deep reflection profiling (Fig. 9.2.2-01). DEKORP 3B had already been recorded at an earlier stage of DEKORP (see Chapter 8.3.1.5), as one of three nearly parallel seismic profiles crossing the Rhenohercynian and the Saxothuringian Zone of the western central part of the Federal Republic of Germany nearly perpendicularly to the Variscan strike. DEKORP 3-A was planned to connect 3-B to oil industry

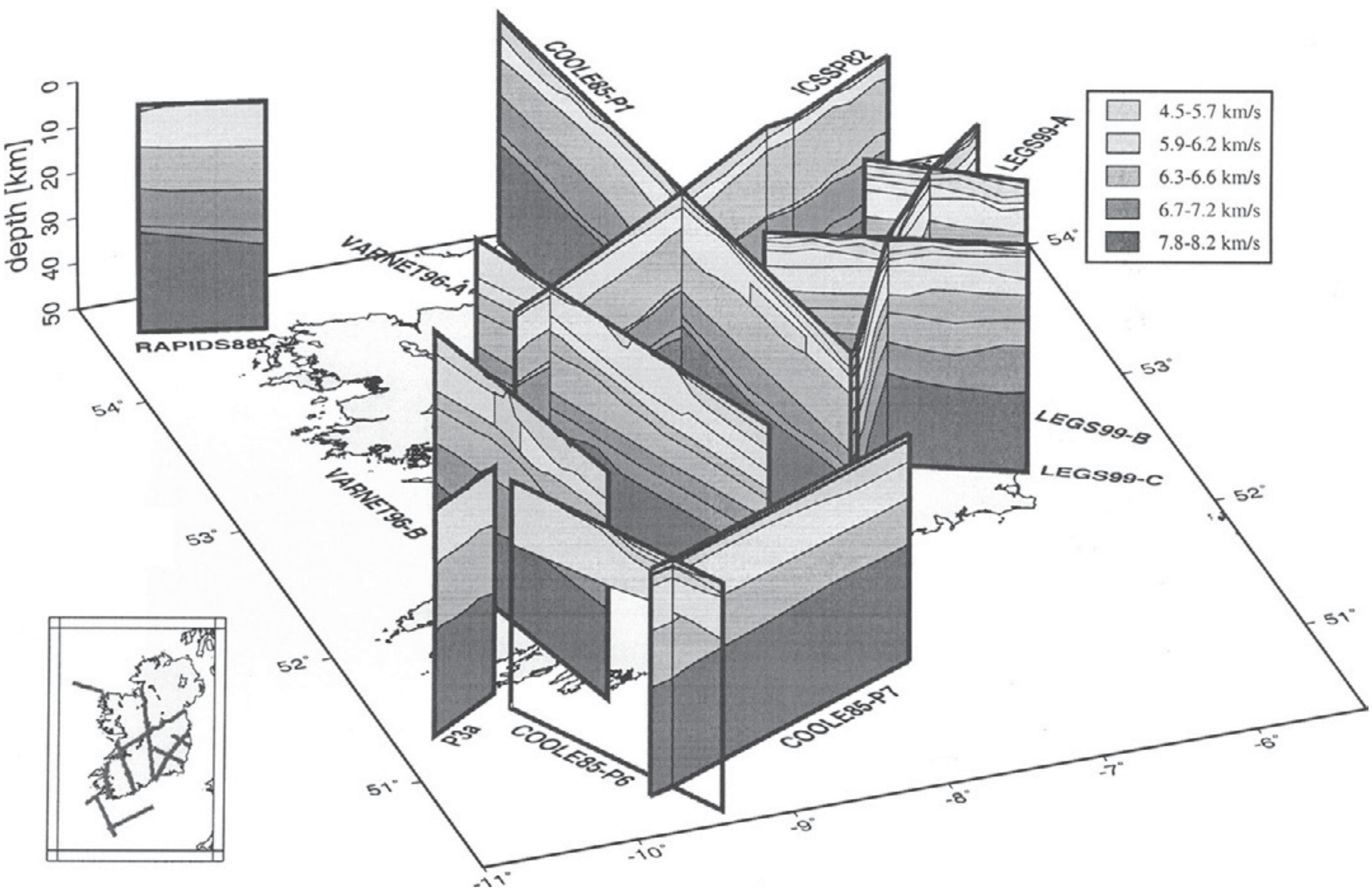



Figure 9.2.1-04. Crustal structure underneath Ireland and adjacent seas (from Landes et al., 2005, fig. 8). [Terra Nova, v. 17, p. 111–120. Copyright Wiley-Blackwell.]

seismic profiles which had been recorded in the Leinegraben area (DEKORP Research Group, 1994a).

DEKORP 3-A and MVE-90 were both carried out in late 1990. The same field parameters and processing procedures were applied as had been used for DEKORP 3-A and 3-B. For logistic reasons, the MVE part was divided into a western and eastern part and the results were published as two separate papers in the same volume (DEKORP Research Group, 1994a, 1994b).

In addition, wide-angle observations accompanied the reflection survey: expanding-spread surveys and a seismic-refraction line were performed simultaneously, while the recording equipment, the 16-km-long, 320-channel receiver spread of the contractor, moved along the line. To obtain reliable and detailed velocity information down to the Moho, two large-scale expanding-spread experiments with shot-to-receiver distances up to 110 km were incorporated into the program. They were centered at two locations with particular gravity anomalies. While the reflection unit moved gradually during the course of the standard Vibroseis reflection survey from SW to NE, two series of properly timed borehole shots were placed symmetrically NE of the sounding points. Charges of 90 kg provided satisfactory signal/noise ratios up to 110 km offset.

The twofold reversed and partly overlapping seismic-refraction line was observed along the 200-km-long eastern portion of MVE-90 (Fig. 9.2.2-01) and was designed in particular to provide wide-angle reflections. The experiment was performed by recording repeated and well-timed explosions at three shotpoints using the 16-km-long receiver spread of the standard reflection survey. The three shotpoints were equally spaced with mutual offsets of ~100 km. Shot charges varied with increasing distance to the receiver spread between 30 and 150 kg. The maximum observation distances were 100 km from the central shotpoint in both directions, and 120 km from the other two shotpoints.

Data and interpretation were published in a special edition of the *Zeitschrift für Geologische Wissenschaften* (*Journal for Geological Sciences*) including an appendix in which the final stacks of the deep reflection seismic data were reproduced (DEKORP Research Group, 1994). The interpretation of the seismic data also included existing gravity data and resulted in a 4-layer crustal model.

The NE-SW–striking section of the MVE-90 profile of 385.3 km length revealed different features in comparison with other, mainly NW-SE–trending DEKORP profiles. While in the NW-SE–striking seismic DEKORP lines, south- and north-dipping reflections appeared widespread, reflective elements of that scale were absent on the MVE-90 profile. This general pattern showed an upper crust down to 8 km depth, characterized by relatively poor reflectivity. In the middle crust, to a depth of

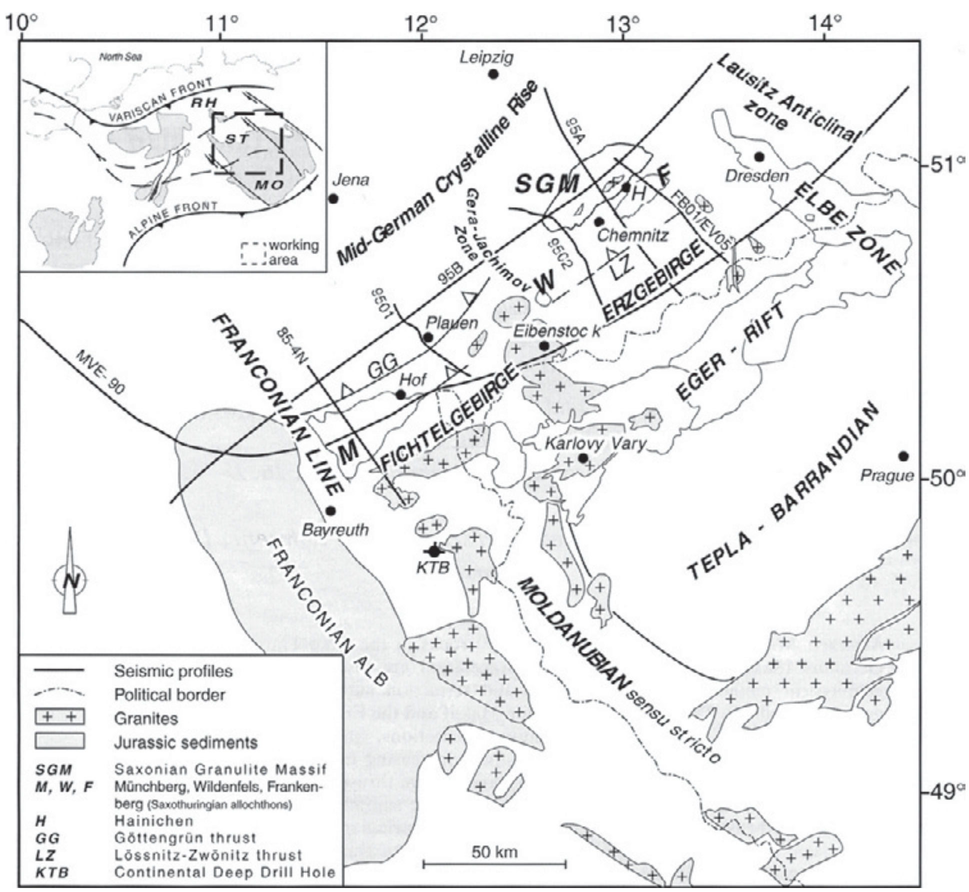



Figure 9.2.2-01. Simplified geological map of southeastern Germany, showing seismic profiles of 1990–1995, investigating the Saxuthuringian part of the Variscides in Saxony, Germany (from Krawczyk et al., 2000, fig. 1). Inset shows the major divisions of the German Variscides into Rhenohercynian, RH, Saxuthuringian, ST, and Modanubian, MO. MVE-90, 85-4N, 9501, 9502 are DEKORP seismic-reflection lines, 95A and 95B are seismic-refraction lines (project GRANU-95), FB01/EV05 are older seismic-reflection lines, reprocessed in the nineties. [*In* Franke, W., Haak, V., Oncken, O., and Tanner, D., eds., Orogenic processes: quantification and modelling in the Variscan belt: Geological Society of London Special Publication 179, p. 303–322, 2000. Reproduced by permission of Geological Society Publishing House, London, U.K.]

18 km, flat and lense-like structures became evident by higher reflectivity. Underneath, again, a poorly reflective crust was seen with sloping short reflective elements dipping to the SW. This general pattern along the MVE-90 line deviated substantially, however, where the line crossed the Münchberg Gneiss Complex, the Vogtland and the Elbe Zone. The lower crust between 24 km depth and the Moho emerged by high reflectivity and horizontal to scarcely dipping structures. The geometry of the Moho showed, different than seen on other DEKORP profiles, an uneven topography varying between 28 and 30 km depth. Under the eastern Erzgebirge an updoming of the Moho was interpreted from both the recording of near-vertical and wide-angle reflections (DEKORP Research Group (B), 1994b).

In 1992, the German Research Society established a new priority program, "Orogenic processes: their quantification and simulation at the example of the Variscides." One of the highlights was the investigation of the so-called Saxonian Granulite Mountains (SGM in Fig. 9.2.2-01), a metamorphic core complex within the Saxuthuringian section of the Variscides (see inset of Fig. 9.2.2-01), located 50–100 km north of the seismic-reflection line MVE-90 which had targeted the Erzgebirge.

In 1995, a major seismic survey targeted this area with both a seismic-refraction survey, GRANU-95 (Enderle et al., 1998a, 1998b) with two lines crossing each other at the center of the Saxonian Granulite Mountain metamorphic core complex, and two seismic-reflection lines, one, 9502, crossing the same structure, the other, 9501, outside to the southwest. Furthermore, the data of an older seismic-reflection survey, FB01/EV05, located to the northeast, and the data of DEKORP 4 (see Chapter 8.3.1.5) and the above described MVE-90 line (lines 85-04 and MVE-90 in Fig. 9.2.2-01), were incorporated in the final interpretation (Krawczyk et al., 2000).

The seismic-refraction project GRANU-95 (Fig. 9.2.2-02) involved 126 recording stations and two deployments. In the first deployment, stations were installed along the 90-km-long line 95-A, running from Leipzig to the Czech border, and along the perpendicular line 95-B around the crossing point D, and 4 shots, A1 to A4, along line 95-A, were recorded. In the second deployment each second station of line 95-A was removed and shifted to line 95-B, which extended over a length of 260 km from north of Dresden, Saxony, to Bamberg, Bavaria, and 8 shots, B to I, more or less equally spaced along line 95-B, were recorded.

Depending on the distance range to be covered, shot sizes varied from 100 kg to 900 kg. The P-wave data of the two seismic-refraction profiles are shown in Appendix A2-1 (p. 32–35) and Appendix A9-1-2 (Enderle et al., 1998b). Furthermore, seismic-reflection equipment with 96 channels, allowing geophone spacing of 80 m, was installed at shotpoint F and recorded all shots.

Southwest of a region with exposed granulites, a highly reflective zone at 1.5 s TWT (two-way traveltime) beneath the reflection lines 9501 and 9502 corresponded well with the top of a

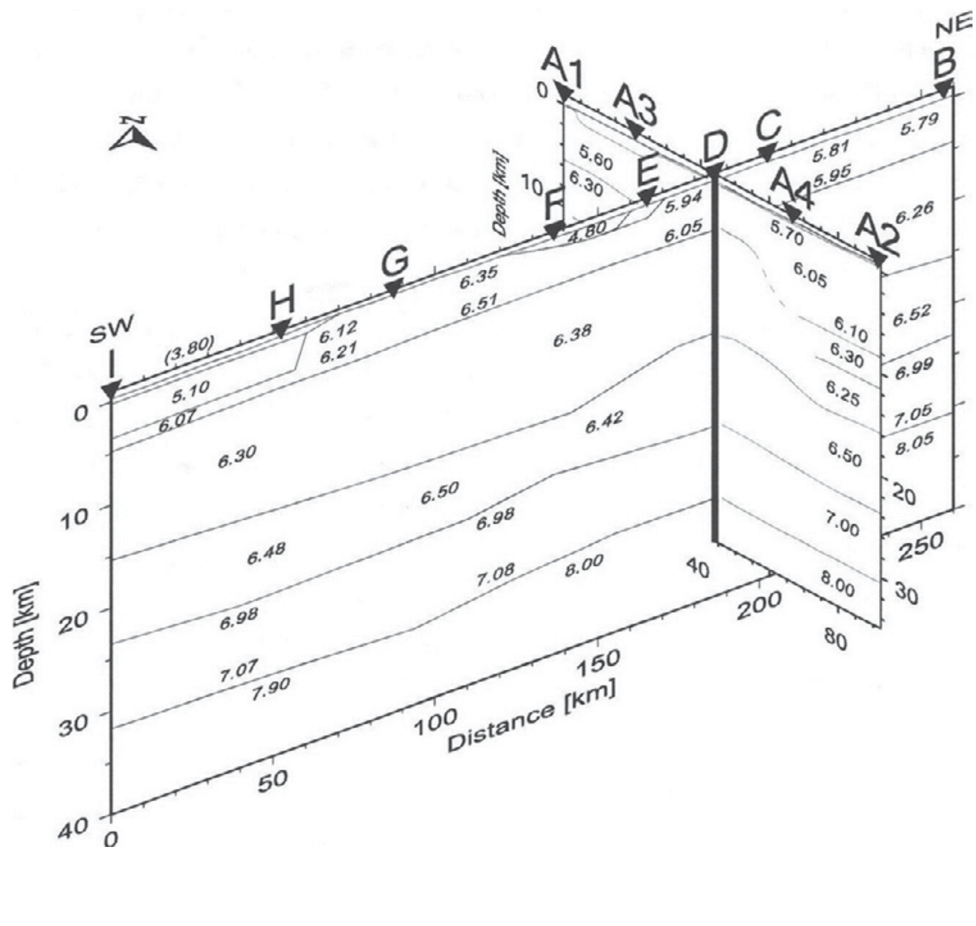



Figure 9.2.2-02. 3-D model of the seismic-refraction project GRANU-95 (from Enderle, 1998, fig. 2.28). [Ph.D. thesis, University of Karlsruhe. Published by permission of Geophysical Institute, University of Karlsruhe, Germany.]

high-velocity zone under line 95-A at ~4 km depth. In this region, high velocities of 6.5–6.6 km/s were also seen at 15–17 km depth (Enderle et al., 1998a; DEKORP and OROGENIC PROCESSES Working Groups, 1999). A lower crust with 7 km/s between 24 and 30 km depth could be well established. Figure 9.2.2-03 gives a composite picture of seismic-reflection boundaries, magnetotelluric anomalies, and Bouguer gravity.

In 1996, DEKORP continued its activities in northeast Germany. The DEKORP-BASIN Group, centered at GFZ Potsdam, undertook a major seismic-reflection survey in the southeastern Baltic Sea and in the lowlands of East Germany. In the Baltic Sea, a grid of marine seismic-reflection profiles was recorded (Fig. 9.2.2-04), covering the area between southern Sweden, Denmark, the island of Bornholm, and the region of northeastern Germany between the Bay of Kiel and the island of Ruegen (DEKORP/BASIN Research Group, 1998, 1999; Krawczyk et al., 1999, 2002; Bleibinhaus et al., 1999).

The marine survey was extended on land into northeastern Germany (Fig. 9.2.2-04) by a detailed seismic-reflection profile BASIN 9601 along a 330-km-long, SSW-directed line from the island of Ruegen to the Harz Mountains (Bayer et al., 1999) and a short cross line BASIN 9602. The land survey consisted of two identical reflection seismic profiles: First, a Vibroseis line was recorded for imaging the upper crust, aiming for depth coverage from the surface to 40 km depth, but concentrating on details of the sedimentary cover and the upper crystalline crust.

Second, an explosive seismic survey was added to ensure imaging of the lower crust and the Moho, aiming for the depth interval between 15 and 40 km (Fig. 9.2.2-05). The observations were complemented by wide-angle recordings for mapping velocities in the deeper crust and at Moho level.

In general, the reflection Moho in the Baltic Sea was found at 28–35 km depth. The data also showed that north of the island of Rügen, the ambiguous picture was characterized by a sequence of northward-dipping reflectors extending from the Moho into the upper mantle. On land, the Moho appeared as a broad band throughout the basin. The continuation of the Moho from its northern boundary into the Baltic Sea, however, was less obvious, and the authors discussed two possible solutions: either the Moho might rise to ~28 km or deepen to ~38 km. Below the basin the Moho appeared flat at a constant depth of 30–32 km depth, but somewhat uplifted south of the southern basin margin to 28 km depth and finally dipping to ~37 km below the Harz Mountains.

In summary, the crystalline crust was found to thin from 32 km at the basin margins to ~22 km below the basin center (derived by subtracting the sediments) and a high-velocity lower crust was detected below the basin. The thickness of the lower crust approached 18 km below the basin center, and thus the middle crust was thinned to only a few kilometers. Nevertheless, the Moho appeared extremely flat and continuous and was well defined as a first-order velocity contrast.

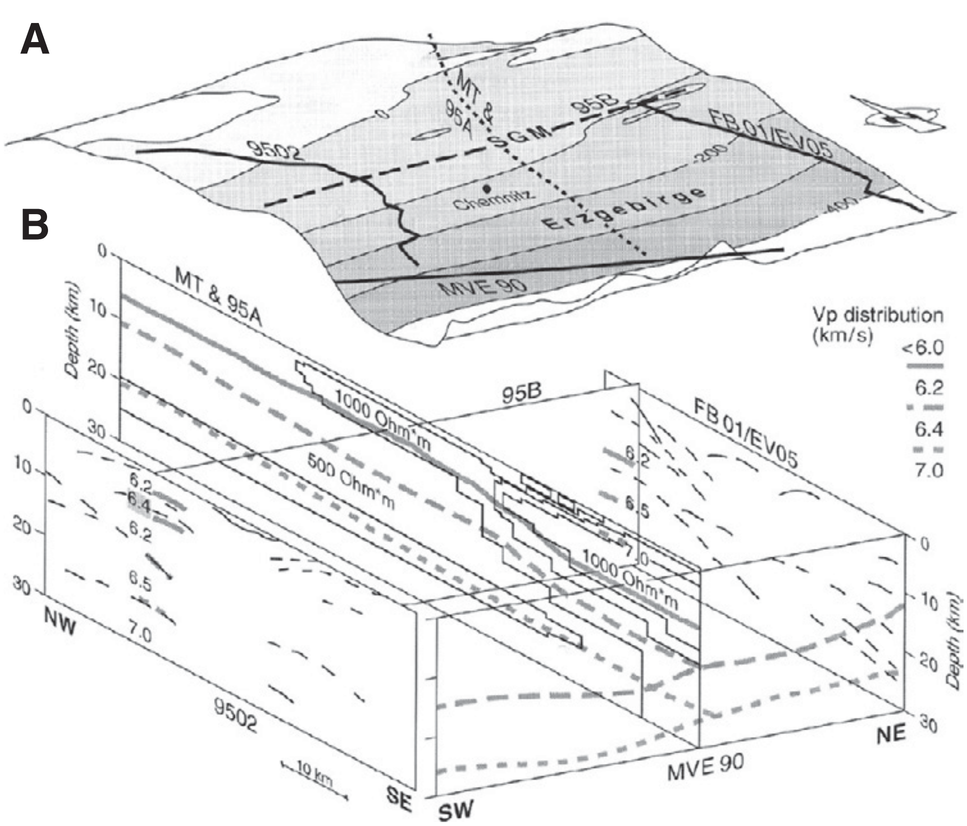



Figure 9.2.2-03. (A) Bouguer gravity contours. (B) 3-D model of the Saxonian Granulite Mountain area comprising seismic, magnetotelluric and gravity results (from Krawczyk et al., 2000, fig. 6). [*In* Franke, W., Haak, V., Oncken, O., and Tanner, D., eds., 2000, Orogenic processes: quantification and modelling in the Variscan belt: Geological Society of London Special Publication 179, p. 303–322. Reproduced by permission of Geological Society Publishing House, London, U.K.]

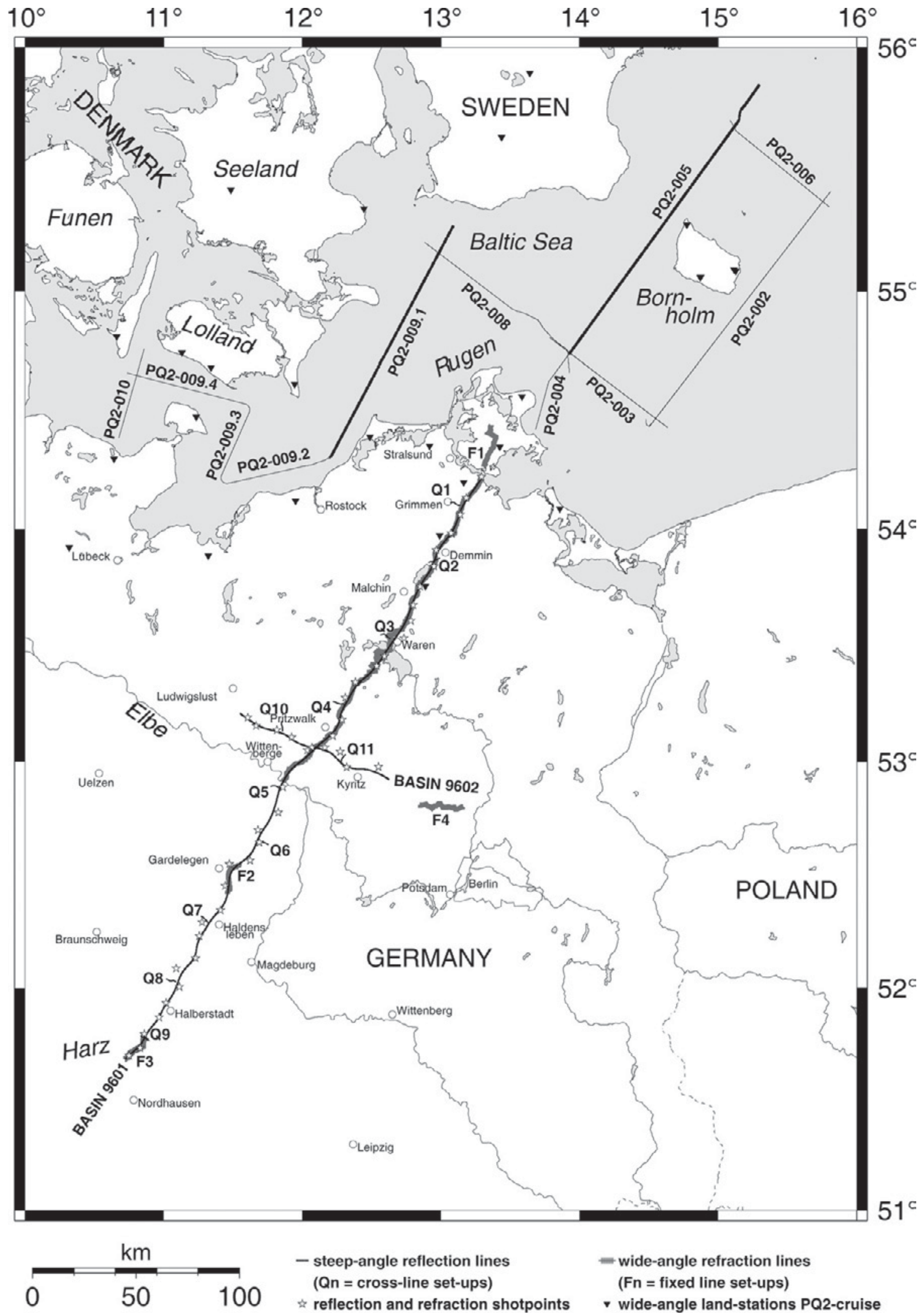



Figure 9.2.2-04. Marine (PQ2-lines) and land (BASIN 9601) seismic reflection lines of the DEKORP-BASIN campaign in 1996 (from Krawczyk et al., 1999, fig.1). [Tectonophysics, v. 314, p. 241–253. Copyright Elsevier.]

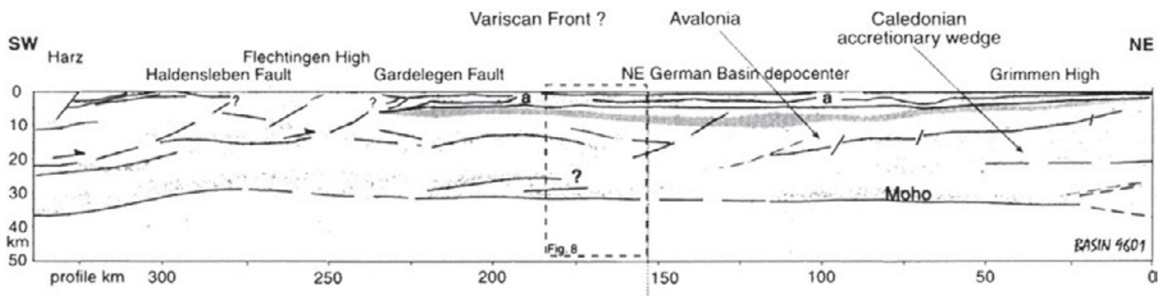



Figure 9.2.2-05. Depth-converted seismic reflection line BASIN 9601 (from Krawczyk et al., 1999, fig. 3). [Tectonophysics, v. 314, p. 241–253. Copyright Elsevier.]

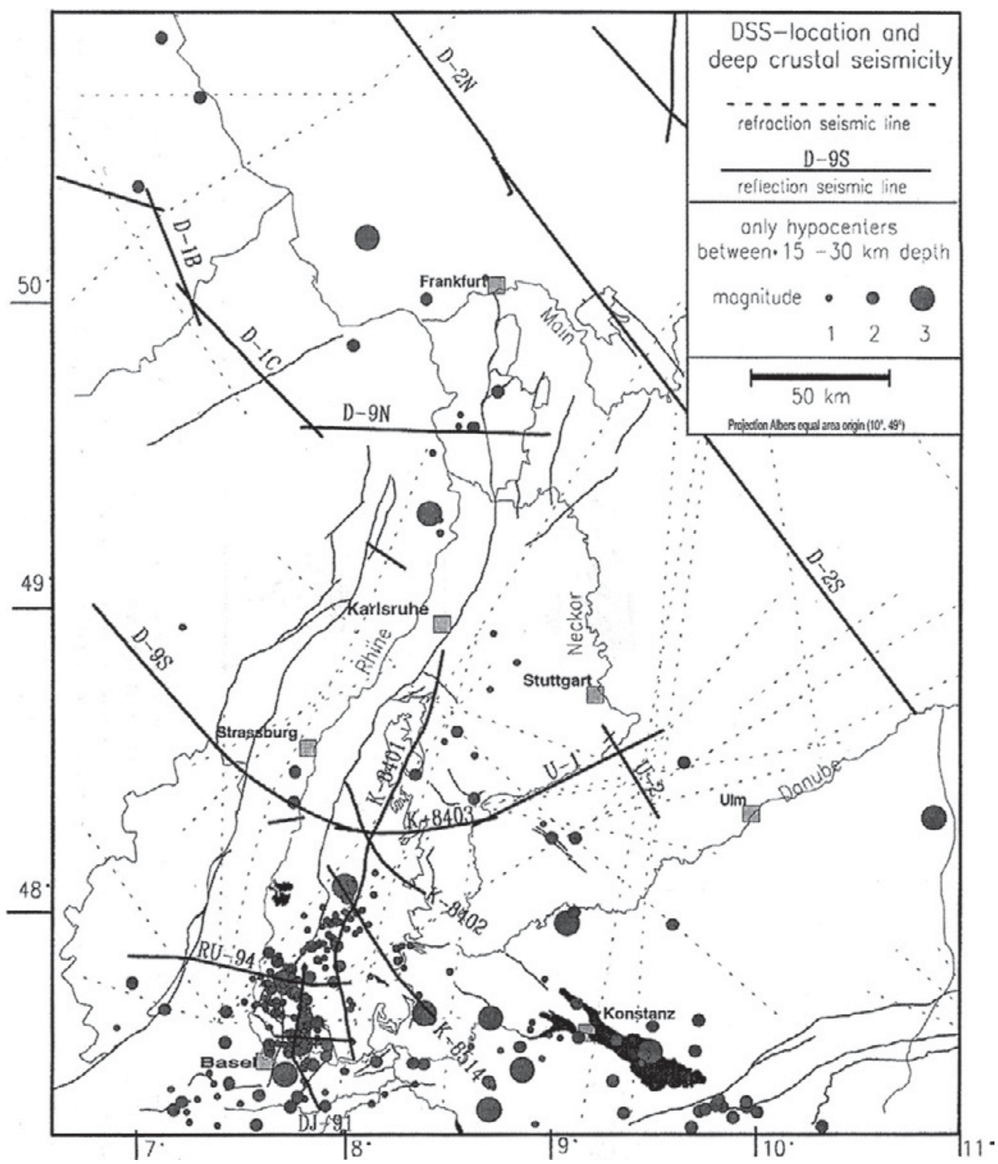



Figure 9.2.2-06. Location of seismic-reflection (solid lines) and refraction (thin dotted lines) profiles in the southern Rhinegraben (from Mayer et al., 1997, fig. 2). Also shown are deep crustal events with hypocenters between 15 and 30 km depth (after Bonjer, 1997). [Tectonophysics, v. 275, p. 15–40. Copyright Elsevier.]

Almost simultaneously with the DEKORP activity in eastern Germany in 1990, the University of Karlsruhe undertook a new initiative to explore the southernmost end of the Rhinegraben (Mayer et al., 1997). From seismology observations, it was long known that this part of the Rhinegraben had an unusual asymmetric distribution of local earthquakes which only on the eastern flank north of Basel occurred at depths of 15–30 km (Bonjer, 1997; Figs. 9.2.2-06 and 9.2.2-07).

Therefore, in 1990 and 1991, two seismic-reflection lines, D-90 and DJ-91, were recorded across the region of deep-earthquake activity, across the Dinkelberg fault block, which is located at the southeasternmost edge of the Rhinegraben (Fig. 9.2.2-07).

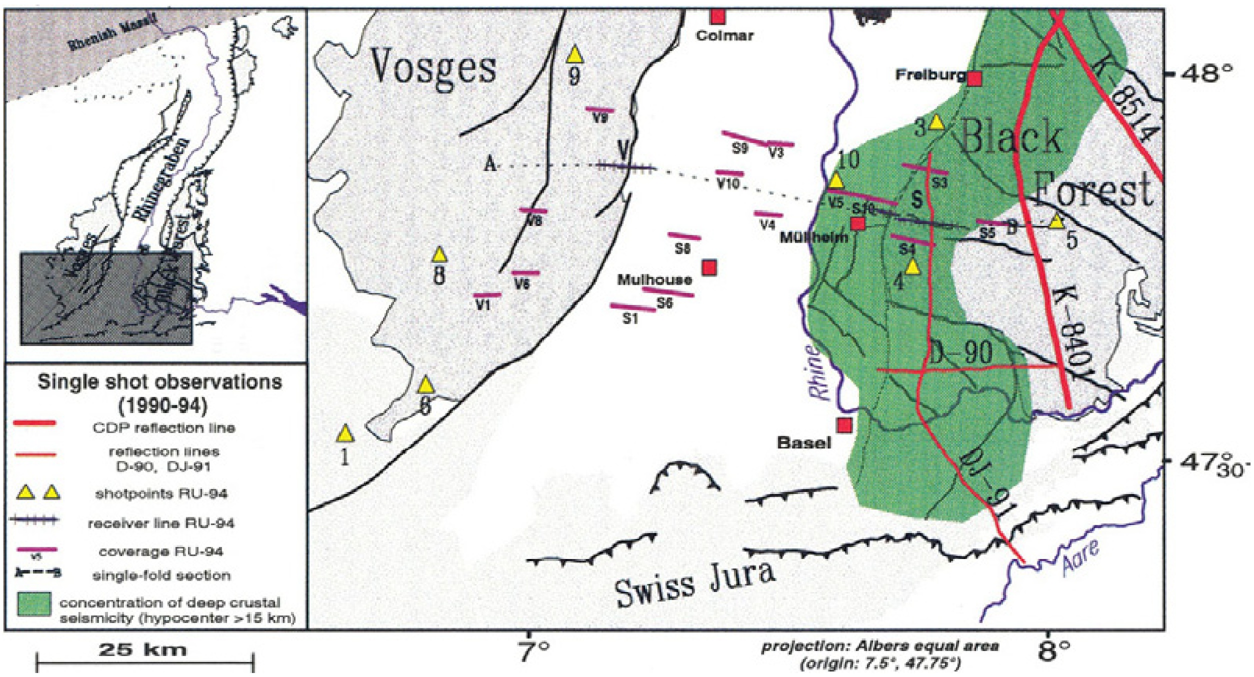



Figure 9.2.2-07. Location of reflection observations in the southern Rhinegraben (from Mayer et al., 1997, fig. 9). RU-94—reflection experiment of 1994 with single-shots (yellow triangles) and recordings (short violet lines) ranging from near-vertical to near-critical reflection distances. D—reflection lines of 1990 and 1991 across the Dinkelberg area. K (thick red lines)—multifold KTB reflection lines of 1984–85. Green area—region of deep earthquake locations. [Tectonophysics, v. 275, p. 15–40. Copyright Elsevier.]

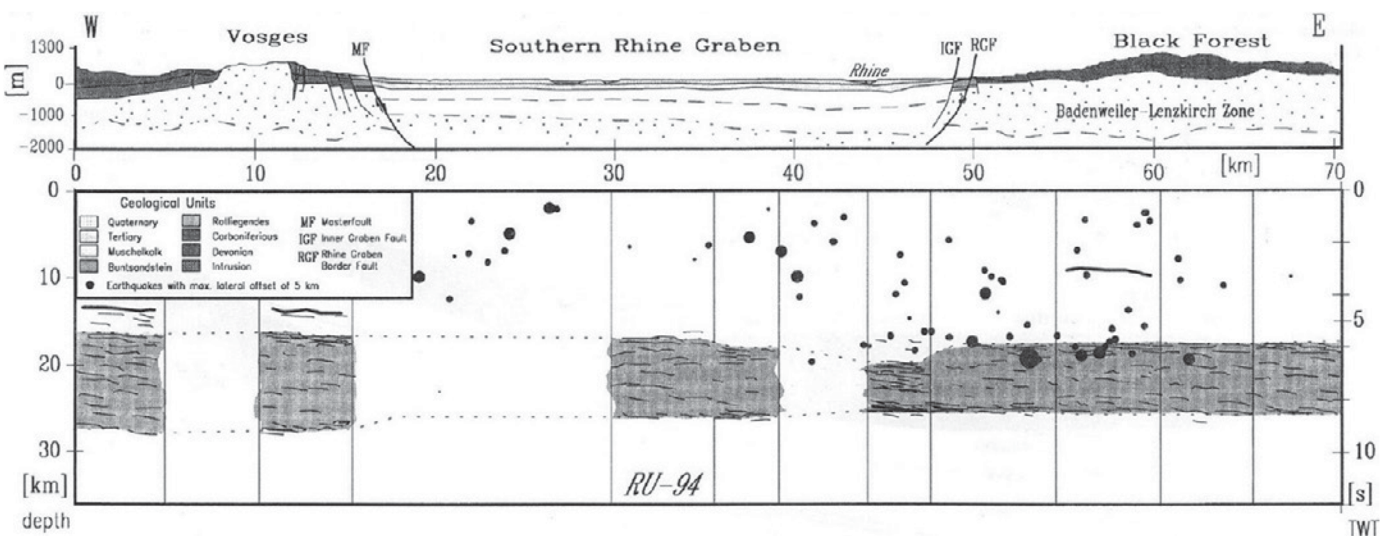



Figure 9.2.2-08. Crustal cross section through the southern end of the Rhinegraben, extending from the Vosges Mountains in the west to the Black Forest in the east, showing local seismic events and a reflective lower crust (from Mayer et al., 1997, fig. 12). [Tectonophysics, v. 275, p. 15–40. Copyright Elsevier.]

Furthermore, a near-vertical to wide-angle survey, partly undershooting the Rhinegraben and traversing it at its southern end, was added in 1994 (RU-94). It had source positions on both rift shoulders and receiver positions both on the rift shoulders and in the graben proper (Fig. 9.2.2-07).

The resulting cross section RU-94 (Fig. 9.2.2-08) was compiled from single-shot observations and shows the top of the lower crust at an almost constant depth of 5.7–6.0 s TWT, corresponding to 17 km depth. Only beneath the inner graben fault a local increase in depth to 19 km was observed.

The Moho was located at the lower end of the reflective lower crust, at 9 s TWT, which is ~27 km depth and, beneath the Vosges, dips slightly toward the west. The maximum depth of hypocenters of local events, projected into the cross section, was found to increase from west to east, reaching the deepest position close to the Rhinegraben border fault and seemed to coincide with the upper boundary of a thinned lower crust.

Also in 1991 and 1992, a second initiative to obtain more details about the crustal and upper-mantle structure of the Central European Rift System was undertaken in central France by a joint French-German experiment in the Limagne graben and surrounding Massif Central. The project consisted of two parts (Fig. 9.2.2-09). First, a teleseismic tomography study involved a six-month period in 1991 and 1992, during which 79 mobile short-period stations and 14 permanent stations recorded teleseismic events (Granet et al., 1995).

The second part was a special seismic-refraction survey in the central part of the Central Massif, observed in late 1992 (Zeyen et al., 1997). The seismic-refraction survey was performed with 150 mobile stations, which recorded seven borehole shots with charges of 400 and 800 kg along two ESE-WNW–trending traverses. As the new equipment pool of GFZ Potsdam had not yet materialized, the recording equipment was borrowed from PASSCAL: 150 SGR stations including a field computer were shipped from the United States to France, and personnel from PASSCAL and the University of Texas, El Paso, helped to run the equipment in the field and produce a complete data tape immediately at the end of the project.

One hundred and five stations were deployed on the 230-km-long southern line crossing two of the neogene volcanic fields and the southern end of the Limagne graben. The other 45 stations were placed on a parallel profile, 70 km farther north, traversing the Limagne graben at its central part, with a gap in the recording sites in the noisy graben proper. Station spacing was generally between 2 and 2.5 km. Due to the simultaneous recording on both profiles and the 70 km distance between the two lines (the expected critical distance for $P_MP$ reflections), reversed fan-type data were obtained, which aimed in particular for a 3-D coverage of the Moho. The P-wave data of the two seismic-refraction profiles are shown in Appendix A2-1 (p. 49–53).

Figure 9.2.2-10 gives an idea which phases were recorded at which locations, assuming horizontal reflectors, and Figure 9.2.2-11 shows the resulting cross section for the southern line.

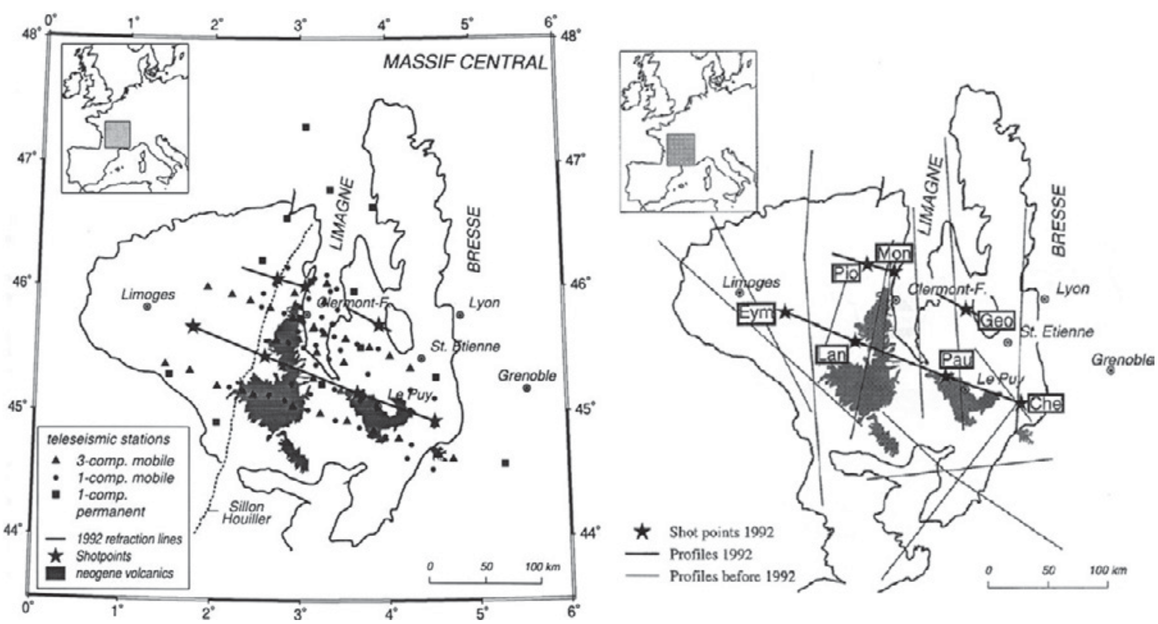



Figure 9.2.2-09. Left: Map of the Central Massif of France, showing the location of teleseismic stations (points and triangles) and the seismic-refraction lines (stars—shotpoints). Also shown is the location of the Limagnegraben and the distribution of neogene volcanics (from Prodehl et al., 1995, fig. 4-4).). Right: Map of Central Massif of France, showing the location of all seismic-refraction lines since 1969. Thick lines design project of 1992 (from Zeyen et al., 1995, fig. 2). [Left: Olsen, K.H., ed., Continental rifts: Evolution, structure, tectonics: Amsterdam, Elsevier, p. 133–212. Right: Tectonophysics, v. 275, p. 99–118. Copyright Elsevier.]

The position where the southern end of the Limagne graben is located clearly showed a drastically decreased velocity to 7.5–7.7 km/s below the general Moho depth range of 28 km.

The resulting crustal thickness varied between 25 km depth under the Limagne graben and 30 km depth under most of the Massif Central. The contemporary teleseismic tomography study resulted in a 3-D velocity structure model of the upper mantle down to 180 km depth, whereby the upper 60 km of the lithosphere displayed strong lateral heterogeneities and showed a remarkable correlation between the volcanic provinces and the observed negative velocity perturbations (Granet et al., 1995).

The deep seismic sounding (DSS) observations of the early 1990s had gathered a great wealth of new high-quality and dense data sets. A critical review of the then-utilized interpretation procedures led Karl Fuchs and co-workers to a new view of the dynamic properties of seismic wave propagation through the crust-mantle transition. In various papers (e.g., Enderle et al., 1997; Tittgemeyer et al., 1996), the wave propagation was tested by modeling the scattering in the upper mantle and by comparing the resulting synthetic seismograms with observed record sections.

The abrupt termination of near-vertical reflections at the Moho and the coincident presence of a strong supercritical $P_MP$ reflection coda in record sections of continental seismic wide-angle refraction experiments led to the vision of a hitherto unrec-

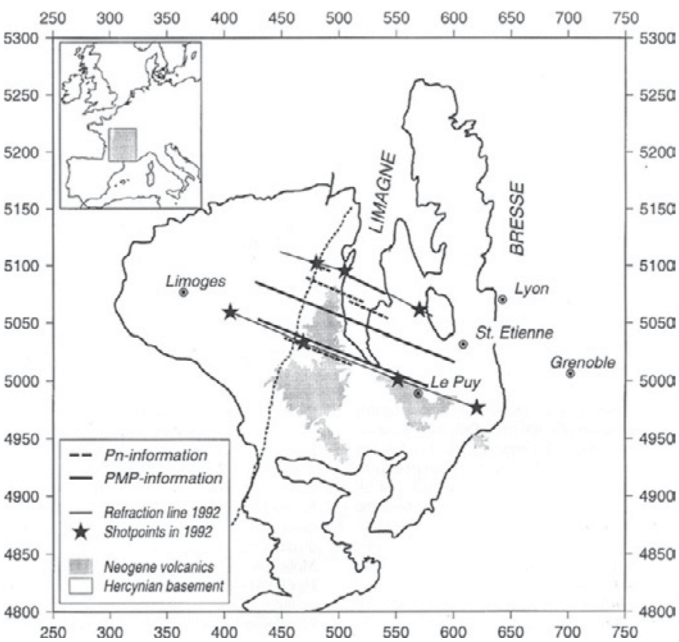



Figure 9.2.2-10. Map of the Central Massif of France, showing the location of the two refraction lines and the resulting positions where $P_MP$ and $P_n$ information is expected assuming horizontal reflectors (from Zeyen et al., 1997, fig. 8). [Tectonophysics, v. 275, p. 99–118. Copyright Elsevier.]

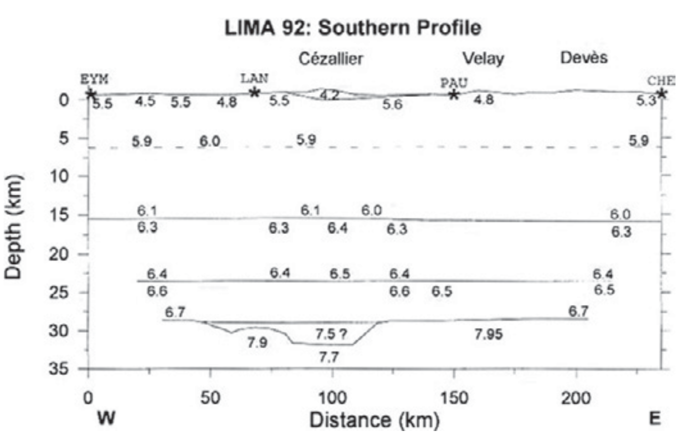



Figure 9.2.2-11. Crustal cross section through the neogene volcanic fields of Cantal (south of Clermont Ferrand) and le Deves (around Le Puy) and the southern end of the Limagne graben in between. (from Zeyen et al., 1997, fig. 7). [Tectonophysics, v. 275, p. 99–118. Copyright Elsevier.]

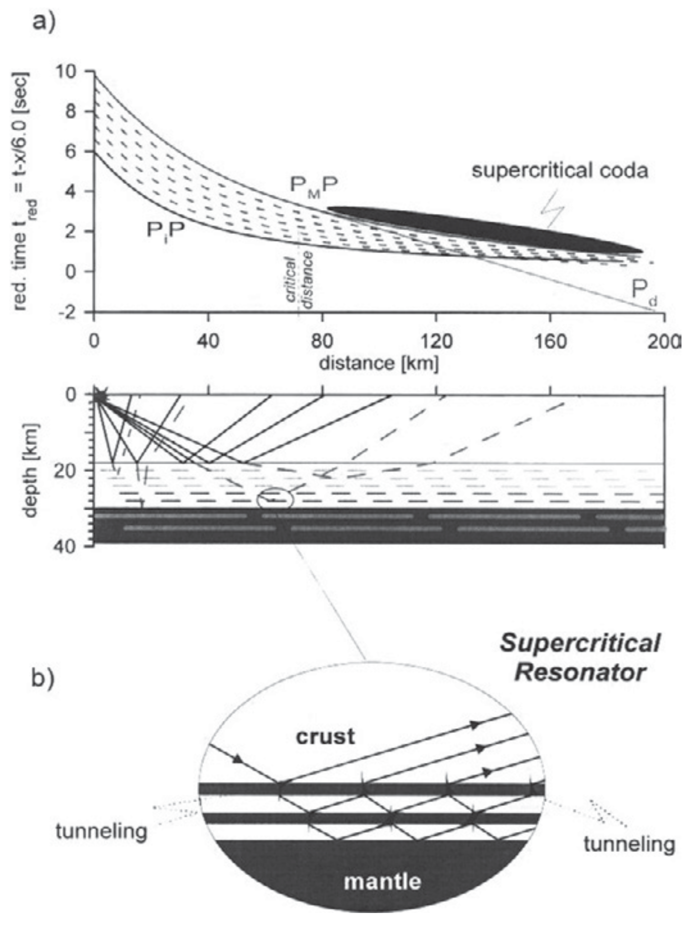



Figure 9.2.2-12. Scattering of energy between $P_IP$ and $P_MP$ phases, caused by internal structure of the lower crust from 18 to 30 km model-depth, is followed by supercritical reverberations following $P_MP$ (supercritical coda), caused by internal structure of the topmost mantle below 30 km model-depth, which has a different scale of structural dimensions (from Enderle et al., 1997, fig. 4a). [Tectonophysics, v. 275, p. 165–198. Copyright Elsevier.]

ognized property of the crust-mantle boundary and a new picture of the lithosphere. It was proposed (Enderle et al., 1997) that the Moho formed a sandwiched mix of crust-mantle material between the lower crust and uppermost mantle in combination with a stepwise increase in mean velocity, and that at Moho level a significant change occurred in the scale of the structural dimensions and of velocity variance (Fig. 9.2.2-12). A similar approach was later used by Mereu (2000a, 2000b) to interpret LITHOPROBE data obtained in the Abiti-Grenville province of Canada (see subchapter 9.4.1.3).

Another advance in interpreting the large amount of recent seismic data was developed at the ETH Zurich. Here, F. Waldhauser undertook a new approach to determine the 3-D topography and lateral continuity of seismic interfaces using 2-D–derived controlled-source seismic reflector data (Waldhauser et al., 1999). He applied his method on more than 250 controlled-source seismic-reflection and -refraction profiles in the greater Alpine region (Fig. 9.2.2-13) to obtain structural information of the crust-mantle boundary (Moho).

The resulting 3-D model of the Alpine crust-mantle boundary showed two offsets that divided the interface into a European, an Adriatic, and a Ligurian Moho, with the European Moho subducting below the Adriatic Moho, and with the Adriatic Moho underthrusting the Ligurian Moho (Fig. 9.2.2-14). Each subinterface depicted the smoothest possible (i.e., simplest) surface, fitting the reflector data within their assigned errors. The results were consistent with previous studies for those regions with dense and reliable controlled-source seismic data and confirmed earlier models proposed, e.g., by P. Giese and co-workers (see Chapter 8.3.4). The newly derived Alpine Moho interface, however, surpassed earlier studies by its lateral extent over an area of ~600 km by 600 km, by quantifying reliability estimates along the interface, and by obeying the principle of being consistently as simple as possible.

The great success of the Swiss NFP 20 (Swiss National-Fond Project) deep seismic-reflection program at the end of the 1980s, when a seismic-reflection survey crossed the entire Swiss Alps in the north-south direction (Fig. 8.3.1-05), had led to the foundation of an international Working Group aiming to establish a second north-south geotraverse through the Eastern Alps. By the end of the 1990s, from 1998 to 2001, a detailed seismic-reflection program, TRANSALP, targeted the Eastern Alps along a north-south–directed geotraverse between Munich, Germany,

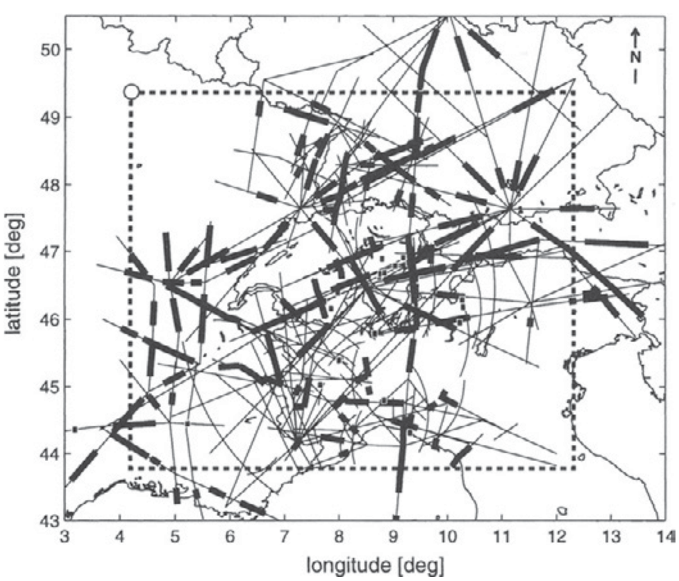



Figure 9.2.2-13. Seismic profiles observed in the greater Alpine region in the past decades (from Waldhauser et al., 1999, fig. 1). Superimposed as thick lines are locations of 2-D–migrated Moho elements from published interpretations of controlled-source seismic data. The dashed box shows the extent of Waldhauser's model in Figure 9.2.2-16. [Geophysical Journal International, v. 135, p. 264–278. Copyright John Wiley & Sons Ltd.]

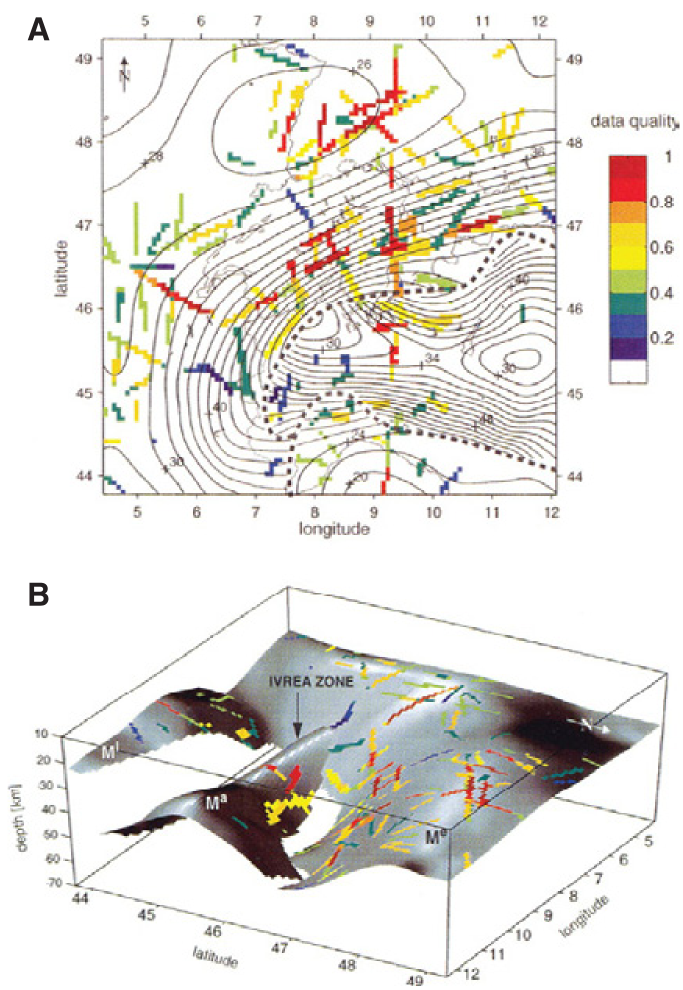



Figure 9.2.2-14. 3-D model of the Alpine Moho (from Waldhauser et al., 1999, fig. 13). (A) Alpine Moho interface contoured at 2-km intervals derived by the smoothest interpolation of the 3-D–migrated controlled-source data. Data quality of the 3-D–migrated database, i.e. reliability of the original models, is indicated by colors. (B) Perspective view toward the SW on the Alpine Moho. [Geophysical Journal International, v. 135, p. 264–278. Copyright John Wiley & Sons Ltd.]

and Venice, Italy (TRANSALP Working Group, 2001, 2002; Lueschen et al., 2004, 2006; Gebrande et al., 2006). It was a cooperative program of Germany, Austria, and Italy (Fig. 9.2.2-15).

The main data acquisition was divided into four different phases. The first phase of 1998 accomplished 120 km and reached from the northern foreland near Freising, Bavaria, across the Northern Alps to the Inn valley in Austria. The second phase, during the winter of 1998–1999, covered 50 km in northern Italy, between Belluno and Treviso, and was devoted to the southernmost Alps and the adjacent Po plain. In 1999, the main phase was realized, covering 170 km of the Central Alps between the Inn valley, Austria, and Belluno, Italy. Finally, in 2001, additional explosive measurements were undertaken in the northernmost Italian section because of noise problems and some bad shots in this particular area during the 1999 campaign.

For the first time, a continuous, 340-km-long seismic-reflection section could be achieved that included the complete orogen at its broadest width, as well as the two adjacent peripheral foreland basins (Fig. 9.2.2-15). The design of the experiment was very complex. Vibroseis near-vertical seismic profiling formed the core of the field-data acquisition. It was complemented by explosive near-vertical profiling and cross-line recording for 3-D control. Furthermore, wide-angle recording of Vibroseis and explosive sources were undertaken by a stationary array for velocity control and active-source tomography, and finally, for another 9–11 months, another stationary array recorded local and teleseismic events to enable passive tomography studies (Lueschen et al., 2004, 2006; Millahn et al., 2006).

The Vibroseis survey was designed to achieve high resolution and penetrate the upper and middle crust. The explosive seismic survey was to provide low-fold, but high-energy signals from the deeper parts of the crust. Shots of 90 kg charges, with 5 km spacing, were fired in 30-m-deep boreholes, when the Vibroseis rolling spread arrived at both the north and south off-

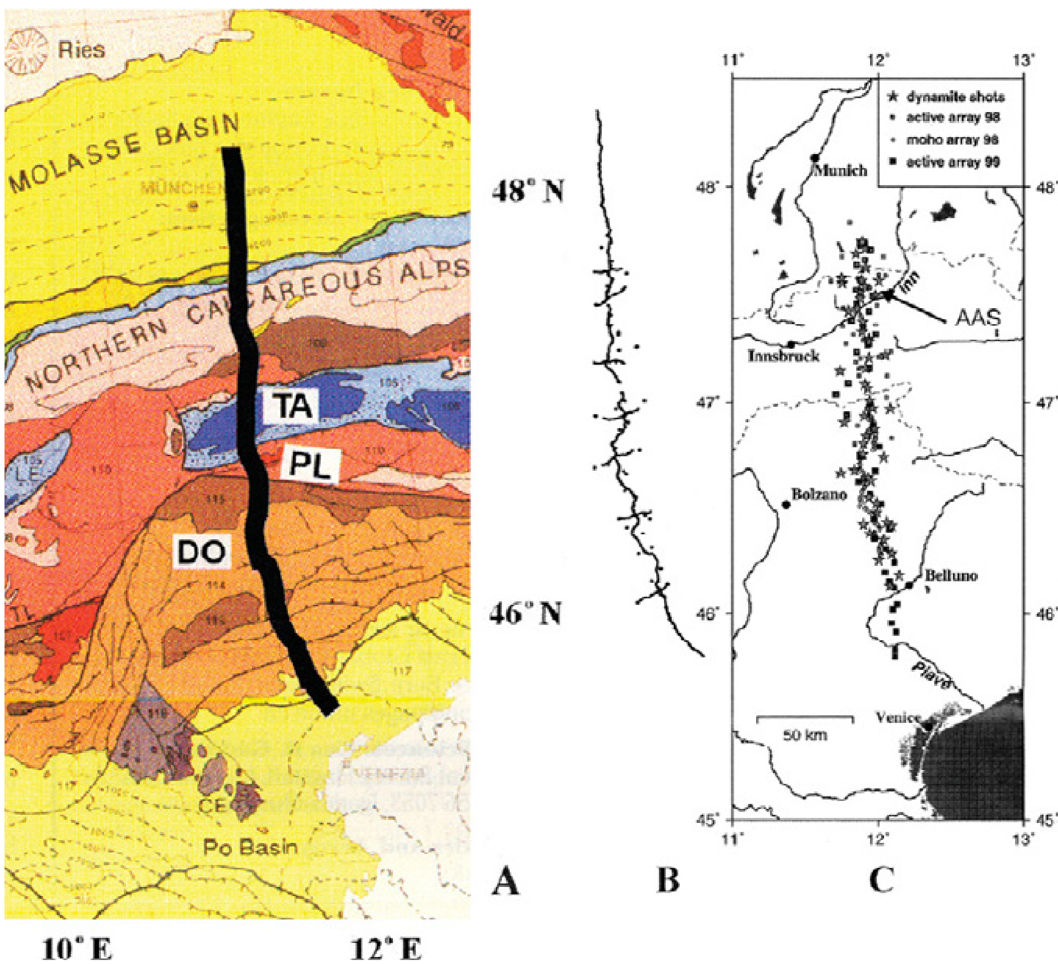



Figure 9.2.2-15. Location map of TRANSALP (from TRANSALP Working Group, 2001, figs. 1 and 3). (A) Tectonic map. (B) Location of explosive sources (charges of 90 kg in three 30-m-deep shot holes) recorded off-end with up to 30-km-long spreads. (C) Stationary, continuously running network of up to 50 three-component units recording all Vibroseis and explosive sources. [Eos (Transactions, American Geophysical Union), v. 82, p. 453, 460–461. Reproduced by permission of American Geophysical Union.]

end configuration. By this procedure, Vibroseis and explosive sources were recorded by the same recording unit and produced data simultaneously.

The resulting data were of excellent quality, with the different types of data complementing each other in depth penetration and resolution characteristics. The northernmost and southernmost parts of the 300-km-long Vibroseis section displayed the stratified Molasse basins, the Tertiary base in the Bavarian Molasse being the most prominent reflection. The section also showed the transition from unfolded Molasse sediments at maximum 6 km depth to the folded Molasse sediments and Flysch zone, characterized by a chaotic signature, and into the Northern Calcareous Alps at maximum 11 km depth. Several onlap structures beneath the Alpine front and a sudden displacement of the Tertiary base by 4–5 km were also visible. At ~9–10 km depth, the Northern Calcareous Alps and their substrata are bounded by a basal, almost horizontal reflection pattern above a relatively transparent upper crystalline crust. Farther to the south, a prominent south-dipping reflection pattern, outcropping at the Inn valley, was interpreted as a thrust fault system along which the Northern Calcareous Alps were overthrusted by their former basement, the Greywacke Zone. Beneath the axis of the Central Alps the profile shows a bivergent asymmetric structure of the crust. Here, the crust reaches a maximum thickness of 55 km, and two 80–100-km-long transcrustal ramps could be traced in the data: north of the axis a southward dipping "Sub-Tauern-Ramp" and south of the crest a northward-dipping "Sub-Dolomites-Ramp" (Fig. 9.2.2-16, Lueschen et al., 2004). The two models, shown in the center and bottom of Figure 9.2.2-16, illustrate clearly the main results of the TRANSALP seismic survey, but express different viewpoints of the tectonic interpretation. For comparison, the predicted model of TRANSALP Working Group (2001) is shown in Figure 9.2.2-16 (top cross section). A more refined interpretation of the wide-angle data was prepared by Bleibinhaus and Gebrande (2006).

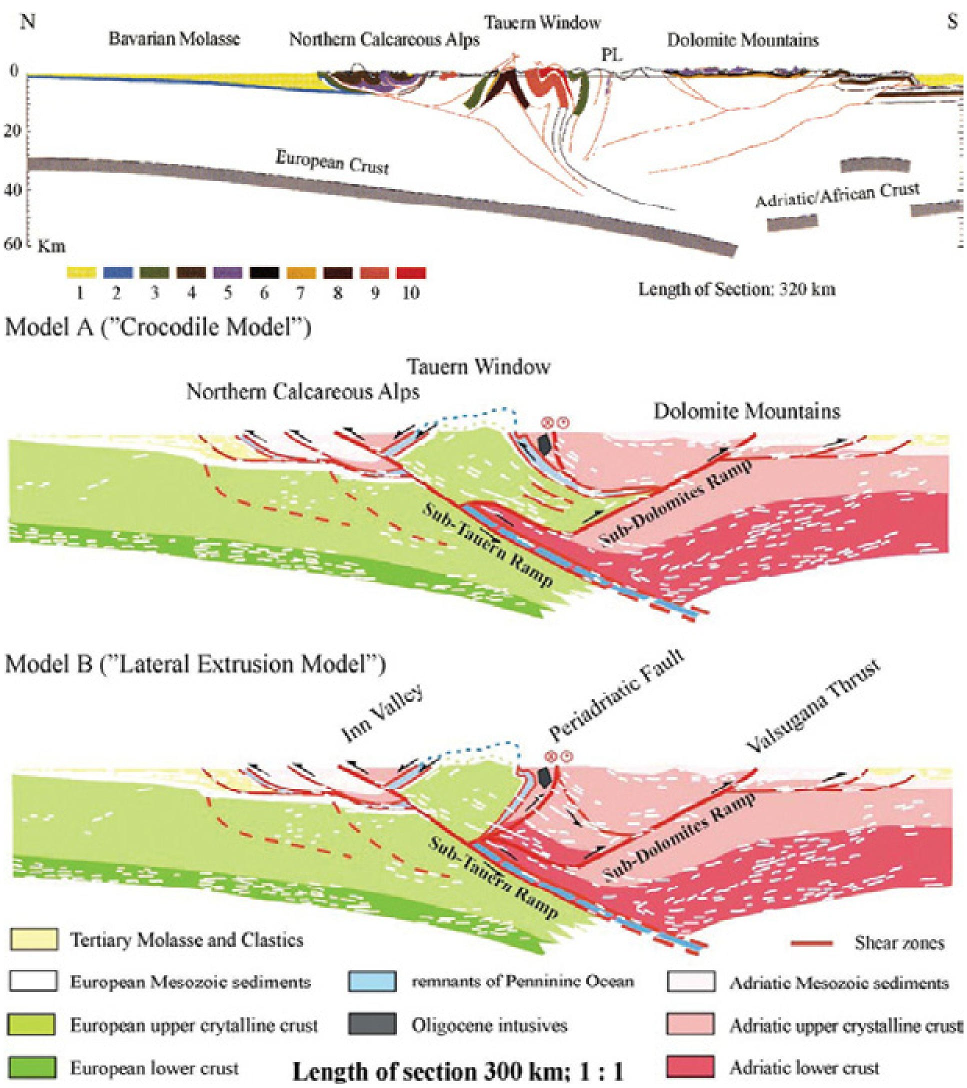



Figure 9.2.2-16. Top: Pre-experiment crustal cross-section along the TRANSALP traverse. The section is mainly based on geological mapping, deep seismic refraction velocity models of the 1970s, and inspired from "indenter tectonics" applied for the Western Alps (from TRANSALP Working Group, 2001, fig. 2). Center and Bottom: Cartoon of two alternative structure and evolutionary models, based on the data of TRANSALP (from Lueschen et al., 2004, fig. 10). [Tectonophysics, v. 388, p. 85–102. Copyright Elsevier.]

### 9.2.3. Southern Europe

Following the last stage of the European Geotraverse, the ILIHA project on the Iberian Peninsula in 1989, in the early 1990s the focus shifted to multichannel seismic-reflection profiles. In 1990 the Spanish Research and Development Plan initiated a deep vertical seismic-reflection program named ESCI (Estructura Sismica de la Corteza Iberica). Three key areas were selected: the northwest Hercynian Peninsula, the northeast Spain-Valencia trough, and the Betics. Subsequently 450 km of terrestrial and 1325 km of offshore profiles in northern, eastern, and southern Iberia were recorded (Díaz and Gallart, 2009).

Crustal research activities started in 1991 in northern Spain. In 1991 the project ESCIN (Estudio Sismico de la Corteza Iberica Norte) was launched. It consisted of several land and marine seismic-reflection profiles (dotted lines ESCIN-1 to -4 and IAM-12 in Fig. 9.2.3-01). Two of the ESCIN lines (1 in 1991, 2 in 1993) were shot and observed on land (Pulgar et al., 1996). The other two lines ESCIN-3 and -4 were marine profiles investigating the continental margin and heading into the abyssal plain of the Bay of Biscay (Alvarez-Marron et al., 1996, 1997; Ayarza et al., 1998).

In 1992, the scientists of Spain made use of the 150 PASSCAL recording units and its computing facility that had been shipped to France for the seismic investigation of the Central Massif before (see above) for a detailed land survey of the Cantabrian Mountains and the adjacent Duero Basin. Five seismic lines were recorded in the Cantabrian Mountains (bold lines in Fig. 9.2.3-01), one in the east-west direction along the strike, and four in the north-south direction into the Duero Basin to the south (Perez-Estaun et al., 1994; Fernandez-Viejo et al., 2000).

In 1993, the Iberian Atlantic Margins European project (IAM), a marine seismic project, focused on the Atlantic margin of Iberia (for location, see Fig. 9.8.4-15). Up to 3500 km of multichannel profiles were acquired in the North Iberian and Atlantic margins, the Gorringe bank and the Gulf of Cadiz (Banda et al., 1995). The shots were also recorded by ocean-bottom seismometers (OBSs) and by a variety of land stations (Fig. 9.2.3-01).

One profile, IAM12, was oriented in N-S direction north of La Coruna. Three profiles, IAM 3, 9, and 11, were continued on land by short in-line profiles providing images of the continent-ocean transition. IAM 11 ran in an east-west direction from the Galician coast across the Galicia margin, IAM 9 also ran east-west farther south in front of the central Portuguese coast, and IAM 3 ran NE-SW in front of Canbo San Vicente. In the follow-

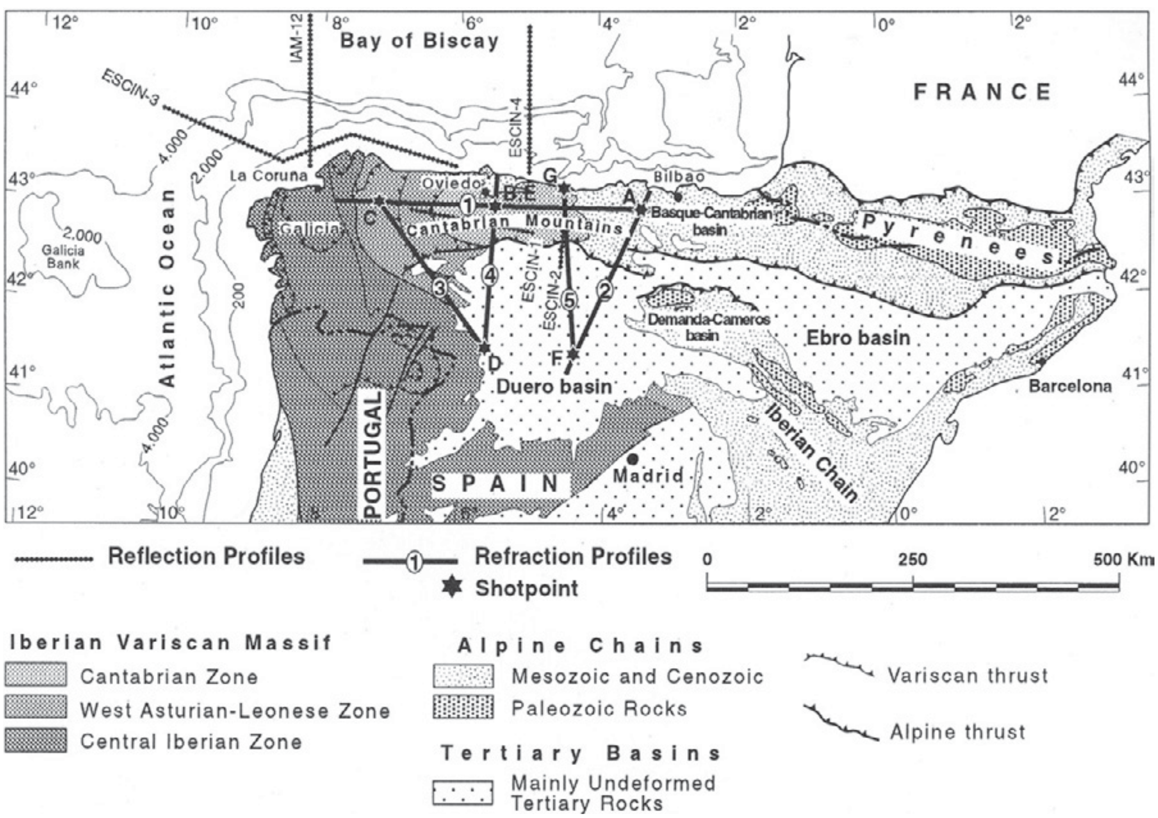



Figure 9.2.3-01. Simplified tectonic map of the northern Iberian Peninsula and adjacent North Iberian margin showing seismic profiles for crustal studies (from Fernandez-Viejo et al., 2000, fig. 1). Bold lines—crustal survey of 1992, dotted lines: projects ESCI and IAM, also recorded in 1992. [Journal of Geophysical Research, v. 105, p. 3001–3018. Reproduced by permission of American Geophysical Union.]

ing years, additional marine experiments were implemented in Atlantic Margins. More details will be discussed in subchapter 9.8.4.3 (North Atlantic).

Combining the data of ESCIN and one of the new lines of 1992, Pulgar et al. (1996) compiled a 350-km-long transect across the Cantabrian Mountains, which consisted of an onshore part in the south and an offshore part in the north (Fig. 9.2.3-02A). The onshore section showed gradual thickening of the crust from 30 km underneath the Duero Basin in the south to ~35 km at the southern edge of the Cantabrian Mountains. Underneath the mountains, the crust thickens drastically northward to almost 60 km at the shoreline. The northern marine cross section, viewed from north to south, started with 16 km crustal thickness under the abyssal plain of the Bay of Biscay. It showed a smooth gradual thickening across the northern Iberian continental margin to near 30 km at the shoreline, overlapping the root zone, seen from the south (Fig. 9.2.3-02A). Fernandez-Viejo et al. (2000) interpreted the whole set of 5 profiles recorded in 1992 and compiled a block diagram (Fig. 9.2.3-02B) showing the thickening of the crust from the Duero Basin into a pronounced crustal root zone underneath the Cantabrian Mountains.

For the second key area, northeastern Spain and the adjacent Valencia trough, a 50-km-long seismic-reflection profile was set up on land between the southern end of a 250-km-long north-south–striking ECORS profile, recorded in the late 1980s, and the Mediterranean coast, and continued toward the south-

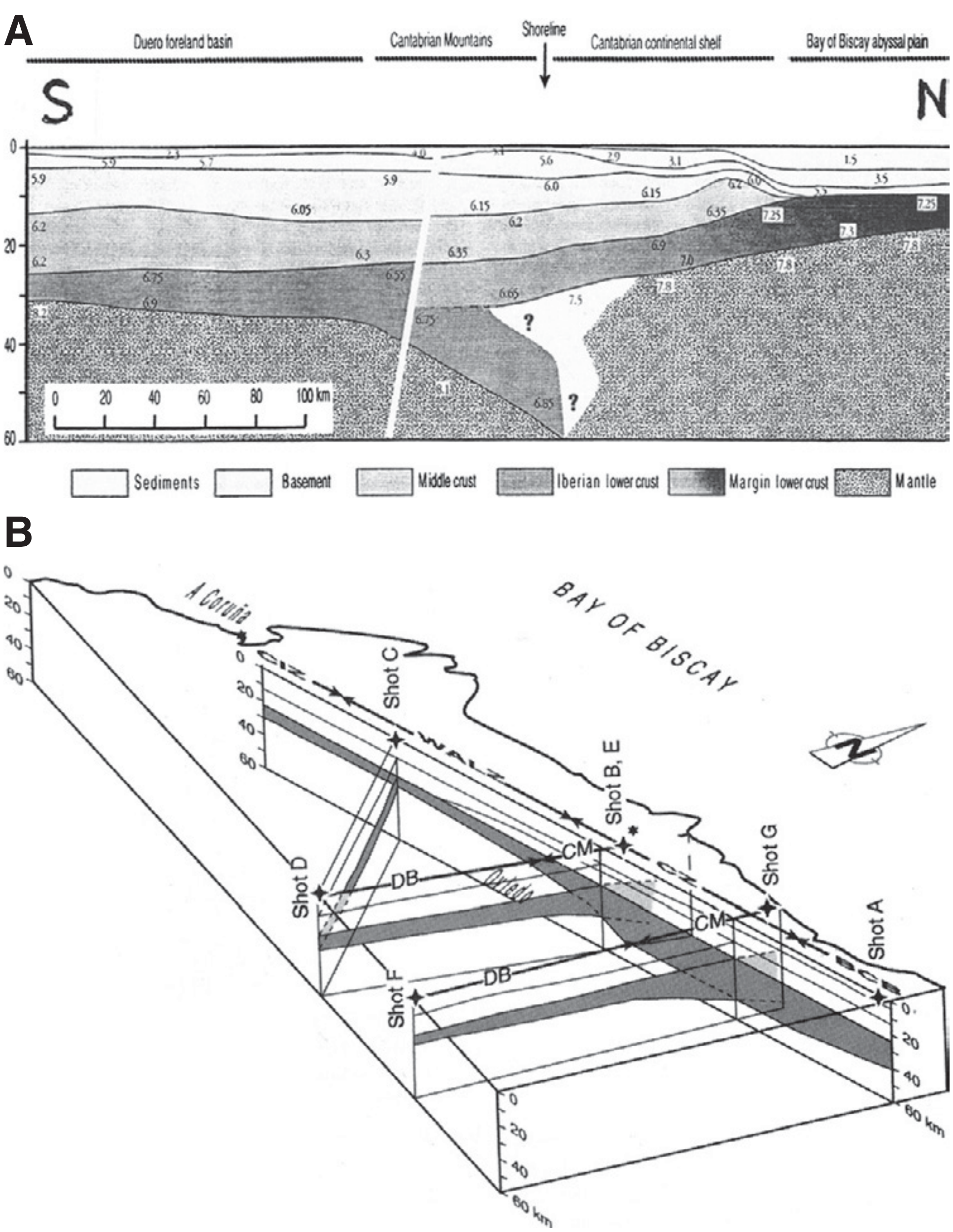



Figure 9.2.3-02. (A) Schematic cross section and (B) fence diagram illustrating the Alpine crustal thickening along the Cantabrian Mountains obtained from the seismic refraction data in the northern Iberian Peninsula (from Pulgar et al., 1996, fig. 11 [Tectonophysics, v. 264, p. 1–19. Copyright Elsevier] and Fernandez-Viejo et al., 2000, fig. 13). CM—Cantabrian Mountains; OB—Duero basin. [Journal of Geophysical Research, v. 105, p. 3001–3018. Reproduced by permission of American Geophysical Union.]

east with a 400-km-long marine profile extending in NW-SE direction across the Valencia trough and continuing south of Mallorca (Gallart et al., 1994, 1995a). The marine airgun shots were also recorded at far offsets by six stations on the Iberian mainland and by two stations on Mallorca (Fig. 9.2.3-03). Also shown on the map are two earlier marine seismic-reflection profiles (VALSIS) which were recorded parallel and ~30 km off the coast line. The ESCI deep seismic-reflection profile showed a highly reflective lower crust of ~15 km thickness overlying the Moho at 29 km depth. The far offset recordings of the marine shots indicated that a major crustal thinning into the Valencia trough of as much as 10 km over a distance of 60 km starts close to the shoreline.

In 1995, the French project LISA (Ligurian Sardinian margins project) sampled different areas of the western Mediterranean from seismic multichannel profiling (Nercessian et al., 2001). Five marine profiles were observed near the eastern end of the Pyrenees. The airgun shots were also recorded by 5 portable land stations (Gallart et al., 2001). This addition aimed to investigate the transition between the Pyrenees affected by the Alpine compression and the Mediterranean, an area of Neogene extension.

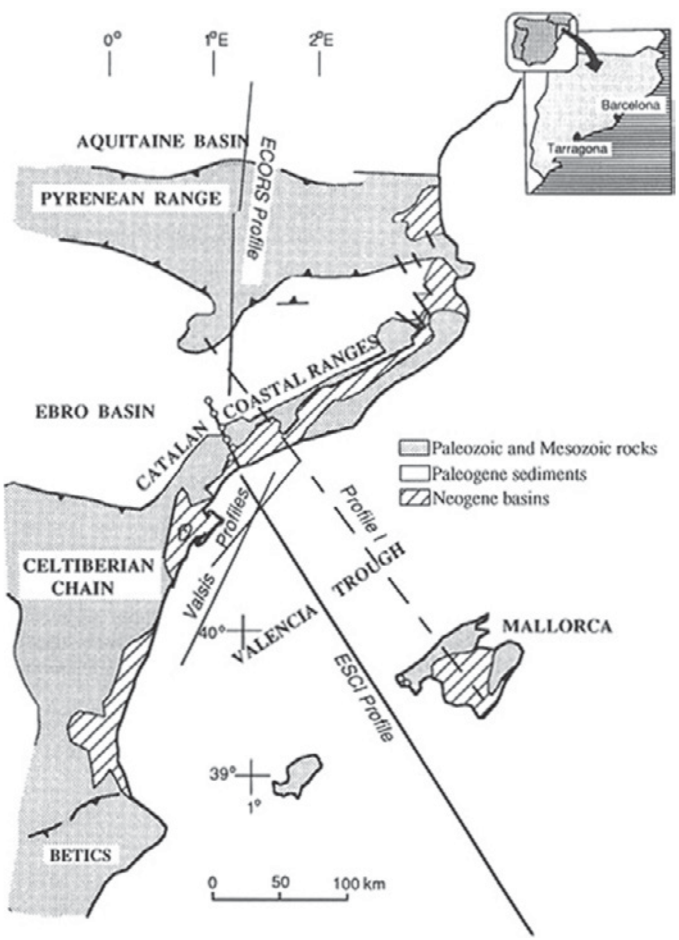



Figure 9.2.3-03. Location map of the ESCI profile extending from the northeastern Iberian Peninsula into Valencia trough and beyond (from Gallart et al., 1994, fig. 1). Open circles mark the land stations recording the marine shots. Profile 1: a seismic-refraction line not yet performed by 1994. [Tectonophysics, v. 232, p. 59–75. Copyright Elsevier.]

In southern Spain, two seismic-reflection profiles ESCI-B1 and ESCI-B2 (Fig. 9.2.3-04) were recorded in the course of the ESCI (Estudio Sismico de la Corteza Iberica) project, traversing the Betic Cordillera, with ESCI-B1 in NW-SE direction, sampling mostly the external Betics, and ESCI-B2 in NE-SW direction, through the inner Betics (Garcia-Duenas et al., 1994; Carbonell et al., 1998a). A third marine seismic-reflection profile, ESCI-A1, extended the second line ESCI-B2 into the Alboran Sea (Comas et al., 1995). The land and marine surveys were connected by a piggy-back wide-angle recording of the airgun shots at sea by portable land stations (Gallart et al., 1995b). Another offshore line, approximately W-W directed, connected the Alboran Sea and the South Balearic Basin (Booth-Rea et al., 2007).

On average the Iberian crust appeared to be seismically featureless. On ESCI-B1 a high-amplitude reflection sequence was seen between 20 and 30 km, partly down to 40 km, and the Moho was imaged as a 3–5-km-thick double band dipping southwards. ESCI-B2, providing a transect through the Sierra Nevada core, showed a highly reflective deep crust, overlying a subhorizontal Moho, but a fairly transparent upper crust and upper mantle (Martínez-Martínez et al., 1997). The highly reflective structure, identified as the crust-mantle boundary at 28–30 km depth differed from tomographic modeling, based on the 21 seismic stations of the Spanish National Seismic Network, also shown in Figure 9.2.3-04, which suggested the existence of a 34–36 km crustal root (Carbonell et al., 1998a).

In Italy, the deep seismic-reflection program CROP (Crosta Profonda) continued with its second phase from 1989 to 1997 (Bernabini and Manetti, 2003). Four seismic-reflection surveys were accomplished on land (red lines in Fig. 9.2.3-05).

CROP 04 covered the southern Apennines with an acquisition length of 154 km. Fieldwork lasted from December 1989 to April 1990. It was recorded using explosives or Vibroseis as energy sources. The acquisition parameters and processing sequence was the same as had been applied in the Alps. However, the results were less favorable, probably due to the different geological subsurface conditions. The first seismic sections showed almost no reflections. Therefore, the line was completely reprocessed in the time interval 0–10 s TWT. The reprocessed section is now comparable with high-quality commercial lines in the region (Scandone et al., 2003). But while commercial lines normally do not exceed 5–6 s TWT suitable for geologic interpretation, the reprocessed CROP 04 stacks show continuous and well-structured events up to 8–9 s TWT. Reflectors evident at 6–8 s TWT have been interpreted as important constrain for the sole thrust of the Apenninic tectonic wedge that in the region of the Alborni Mountains reaches a depth of 20–25 km. This depth coincides with the depth of the so-called “Tyrrhenian Moho” revealed by seismic-refraction experiments in the 1970s (e.g., Colombi et al., 1973; Giese et al., 1973).

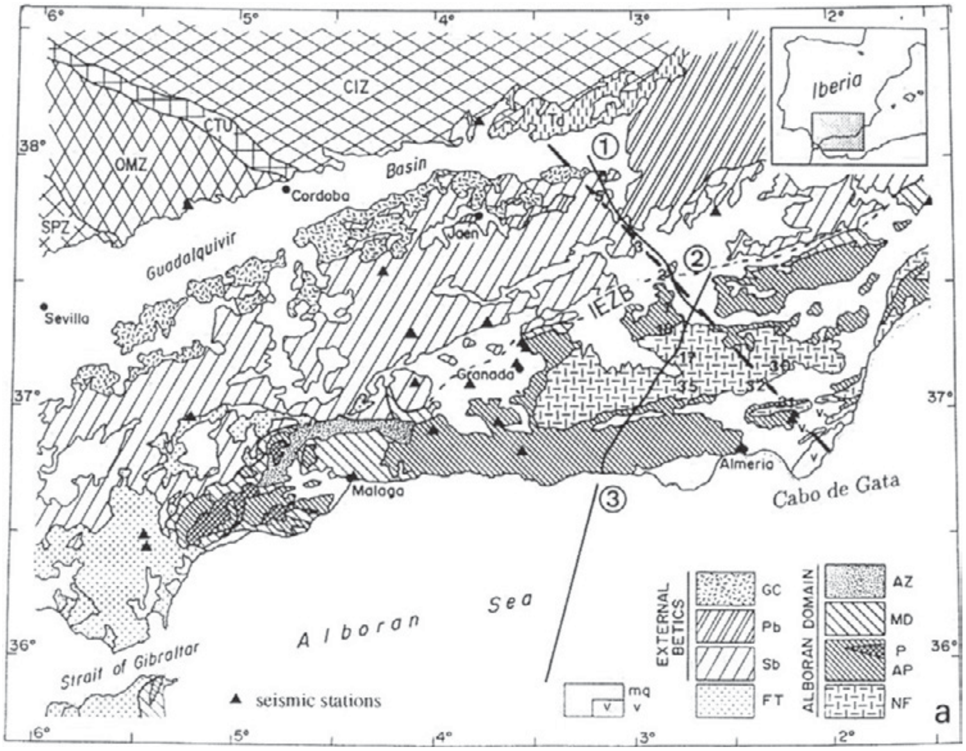



Figure 9.2.3-04. Geologic map of southern Spain, showing the location of the ESCI-profiles: 1—ESCI-B1, 2—ESCI-B2, 3—ESCI-A1, triangles—stations of the Spanish National Seismic Network (from Carbonell et al., 1998, fig. 1a). [Tectonophysics, v. 288, p. 137–152. Copyright Elsevier.]

In 1992–1993, the 220-km-long line CROP 03 was recorded across the northern Apennines, oriented in WSW-ENE direction between Punta Ala near Grosseto on the Tyrrhenian side, opposite of the island Elba, and Gabicce near Pesaro on the Adriatic side, and crossing the crest of the Apennines between the cities Siena-Arezzo to the north and Perugia to the south. The acquisition parameters were adjusted, and so events up to 15 s TWT are visible in the sections. Wide-angle reflections were also recorded along the line with large shots. The crustal structure could be well resolved, both for the extensional Tyrrhenian domain, where the Moho is supposed to be very young and is located at a depth of 22–25 km, and of the compressional Adriatic domain, where the compression is still active and where the well-stratified crust reaches a depth of ~35 km (Barchi et al., 2003).

CROP 18 was shot in 1995 and was recorded in a NW-SE direction in southern Tuscany between the geothermal areas of Ladarello and Mount Amiato which is an area with exceptionally high heat flow of 120 mW/m$^2$ and local peaks of up to 1000 mW/m$^2$. Line 18 crosses line 03 near its western end (Fig. 9.2.3-05). The area was affected first by the convergence and subsequent collision of the European and African margins in Late Cretaceous and early Miocene and second by the extensional tectonics in the inner part of the Apennines since the Miocene with the emplacement of mixed crustal and mantle magmas. The reflection line was acquired primarily for geothermal purposes. A regional mid-crustal reflector, referred to as K-horizon, and some bright spots are the dominant features in the reflection section. The crust-mantle boundary was located at 22–25 km depth (Batini et al., 2003).

CROP 11 crosses central Italy between Marina di Tarquinia at the Tyrrhenian side near the city of Civitavecchia and Vasto at the Adriatic coast. The line was arranged oblique, but as perpendicular as possible to the strike of the Apennines. It was acquired in two stages, the first 109 km in 1996, and the second 156 km in 1999. The deep crustal structures are well imaged in correspondence to the Apulian foreland. At around 7 s TWT, a reflector could be recognized, dipping gently toward the west. It could be easily followed to the Bomba ridge where complex structures and velocity changes near the surface and at depth were encountered. Other deeper and subhorizontal reflective sections were images at 10 s and at 12 s TWT. They were assigned to the 2.5 s thick lower crust of ~9 km thickness. The Moho was identified with 12.5 s reflection time at 32 km depth. The corresponding reflections deepened westward reaching 13–14 s TWT beneath Mount Maiella.

The final land project of CROP was the participation at the international TRANSALP project (see subchapter 9.2.2; TRANSALP Working Group, 2001, 2002; Lueschen et al., 2004, 2006; Gebrande et al., 2006), a cooperative program of Germany, Austria, and Italy (Fig. 9.2.2-15) carried out from 1998 to 2001. The Italian seismic-reflection part was accomplished in 1998 and 1999 and covered a distance range of 327 km. In addition to the main profile, six short profiles, orthogonal to the main line, were recorded. Both Vibroseis and explosive sources were used.

At the end of CROP Phase II, plans were under way to continue the seismic-reflection program in the 2000s with profiles through Calabria and Sicily.

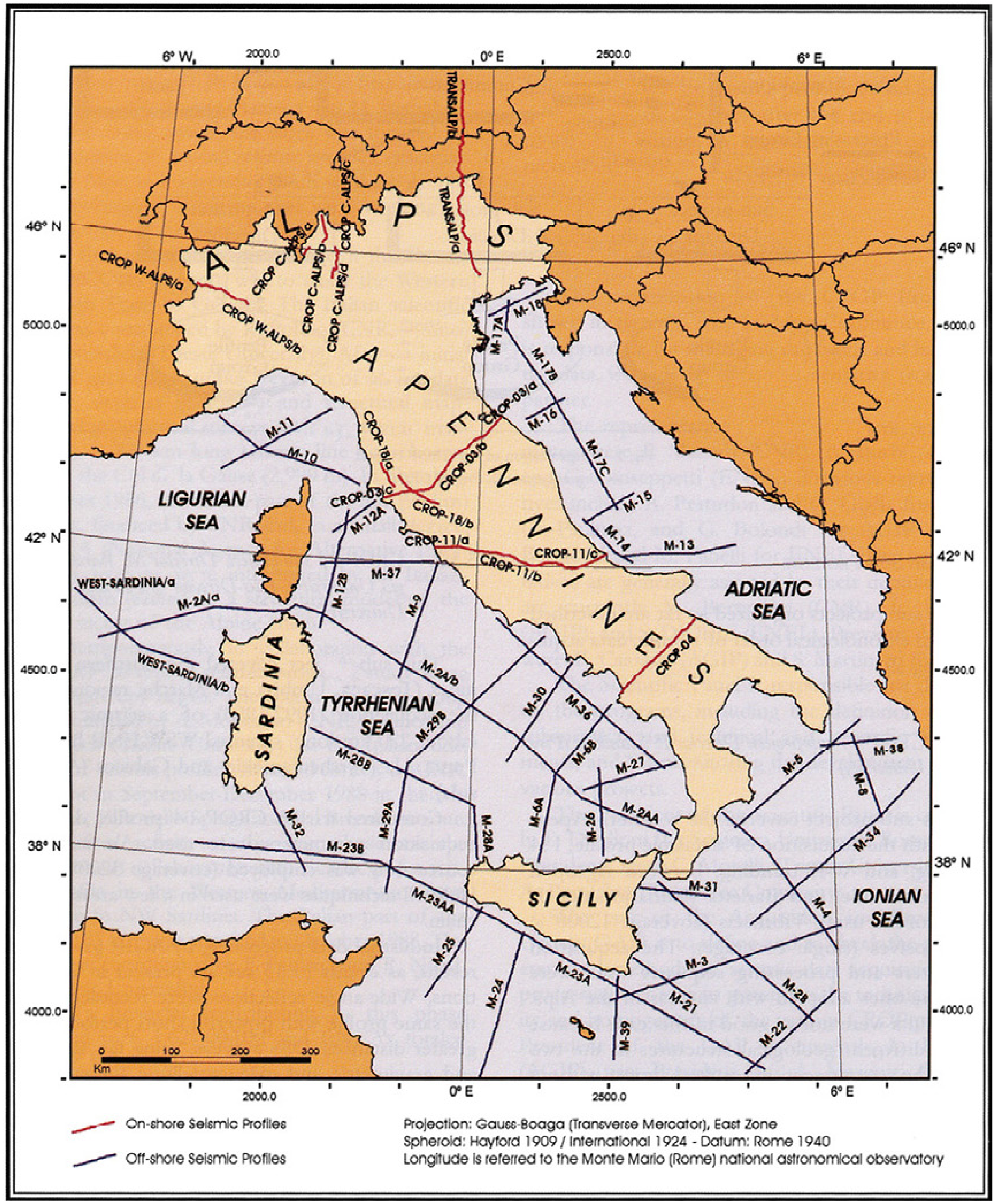



Figure 9.2.3-05. Location map of CROP seismic reflection lines on land and at sea (from Bernabini and Manetti, 2003, fig. 2) [*In* Scrocca et al., eds., 2003, CROP Atlas: seismic reflection profiles of the Italian crust: Memorie Descrittive della Carta Geologica d'Italia, Roma, v. 62, 194 p. Copyright CROP.]

Besides the onshore program, a large marine program was realized in two phases. During CROP MARE 1, in 1991 ~3400 km of seismic profiles were recorded in the Ligurian, Tyrrhenian, and Ionian Sea (Fig. 9.2.3-05). In the second phase, CROP MARE 2 in 1993–1994, over 5000 km marine seismic profiles were acquired in the southern Tyrrhenian Sea, in the Sardinia Channel, and in the Ionian and Adriatic Seas (Fig. 9.2.3-06). The profiles of both phases were acquired by the oceanographic vessel *Explora*, using airgun sources (Bernabini and Manetti, 2003; Finetti, 2003). The data quality was fair to good in foreland regions, such as the Adriatic and Pelagian Seas, and in parts of the deep basins, but in the areas with a more complex tectonic setting, such as the Tyrrhenian Sea margins, the Corso-Sardinian block, and the Calabrian arc, the data quality was strongly variable (Finetti, 2003)

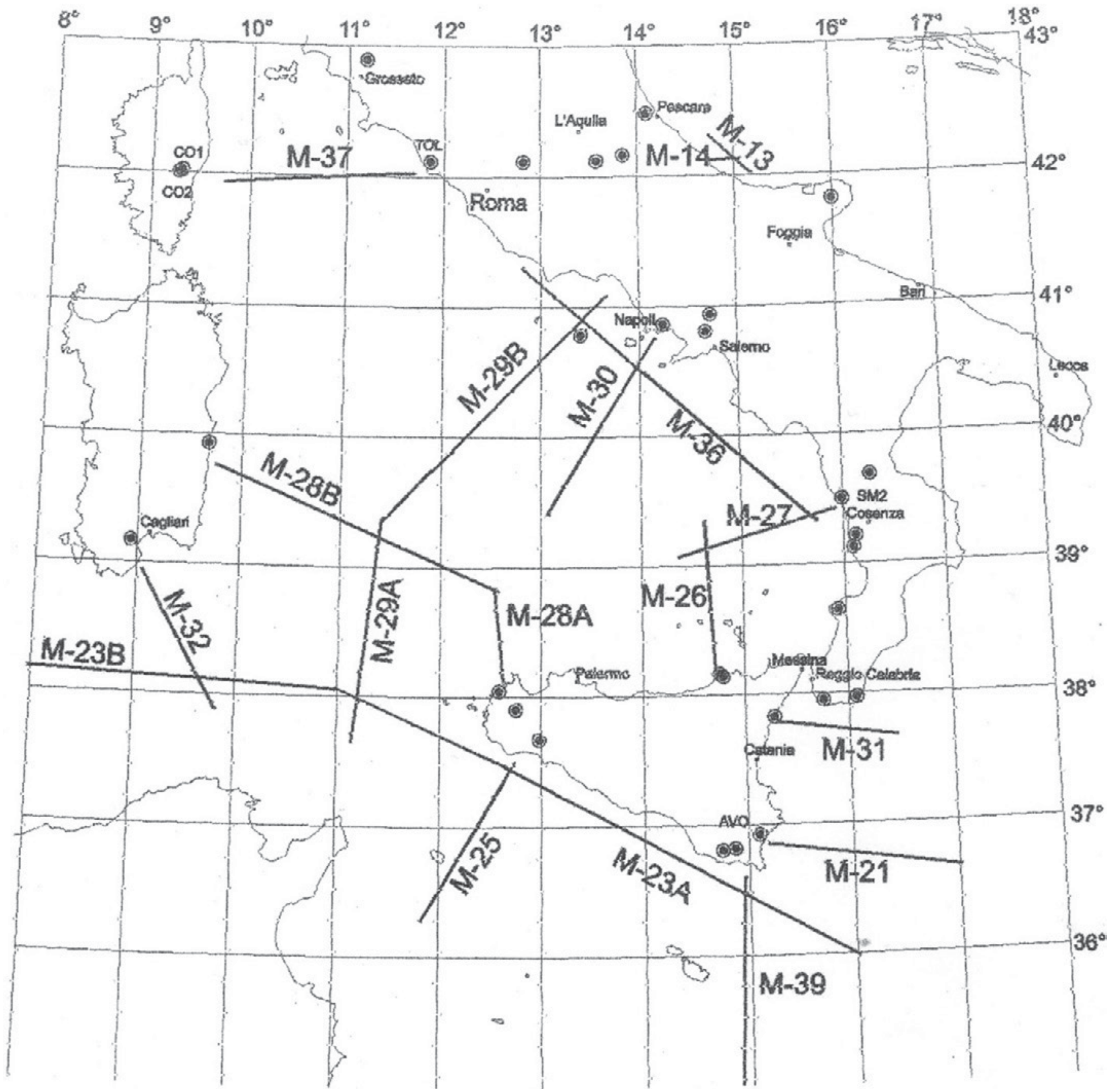



Figure 9.2.3-06 Location map of wide-angle seismic profiles recorded during the CROP MARE 2 phase (from Caiello et al., 2003, fig. 2) [*In* Scrocca et al., eds., 2003, CROP Atlas: seismic reflection profiles of the Italian crust: Memorie Descrittive della Carta Geologica d'Italia, Roma, v. 62, 194 p. Copyright CROP.]

During the acquisition of the marine data, low-frequency seismographs recorded the offshore shots on land (Fig. 9.2.3-06). These data provided additional high-density wide-angle reflection/refraction profiles in some offshore-onshore configurations along the Italian coast (Caielli et al., 2003). Depending on the local conditions, a shooting interval of 20 s provided at least 300 and up to 3000 traces of wide-angle data with trace spacing of 50 m, covering maximum recording distances between 60 and 200 km.

While the originally planned CROP lines in southern Italy could not be realized in the CROP Phases 1 and 2, in the early 1990s, southern Italy became the target of some other seismic projects, most of them combined onshore-offshore experiments, targeting in particular the Ionian Sea between southern Italy and Greece.

The first project was a small-scale land-seismic experiment. In southern Calabria, southern Italy, crystalline basement is exposed, which is interpreted as lower-crust material and therefore had been defined as a key area for an investigation of the structure, composition, and evolution of the Hercynian lower crust in Europe (Schenk, 1990).

Therefore, in early 1990, a particular experiment was carried out by an Italian-German collaboration (Lueschen et al., 1992). A small-scale seismic-reflection experiment aimed to compare in situ seismic properties with petrological data from field mapping and from laboratory measurements of seismic properties in the same rock units (Fig. 9.2.3-07, top). In this respect, the experiment was going to complement super-deep continental drilling programs, which aim for upper and mid-crustal levels. The seismic profile was located in the center of the Serre Mountains and crossed the main lithological units, following a plateau in a N-S direction. The basic principle used in the field layout was to deploy a fixed, densely spaced receiver spread with all available recording channels and fire all shotpoints along the complete line, then move the spread to the next of a total of 3 sectors and repeat all 9 shots (Fig. 9.2.3-07, bottom).

Additionally, four short E-W transverse profiles ~1 km long (with off-end shots of small 5 kg charges) were recorded within the main lithological units in order to detect azimuth-dependent seismic velocities. Recording was done with three-component geophones at 80 m spacing. The experiment gave reasonable results (Fig. 9.2.3-07, center), however, the reflectivity of the outcropping lower crustal units was lower than theoretically predicted, and the analysis of refracted-wave velocities revealed values systematically lower by up to 30% than laboratory data on rock samples or calculated data from modal analysis. The authors attributed the main discrepancy to large-scale alteration of the rocks due to Apennine tectogenetic events (Lueschen et al., 1992).

In 1992, the STREAMERS project was launched, comprising more than 500 km of seismic-reflection profiles throughout the Ionian Sea between southern Italy and Greece (Hirn et al., 1996).

A series of six offshore lines, using a 180-channel 4.5-km-long streamer and airgun sources was recorded to the south and east of Calabria into the western Ionian Sea (Cernobori et al., 1996). On land recording stations were positioned at the site TS-1 and along a profile named Serre (Fig. 9.2.3-08). The map

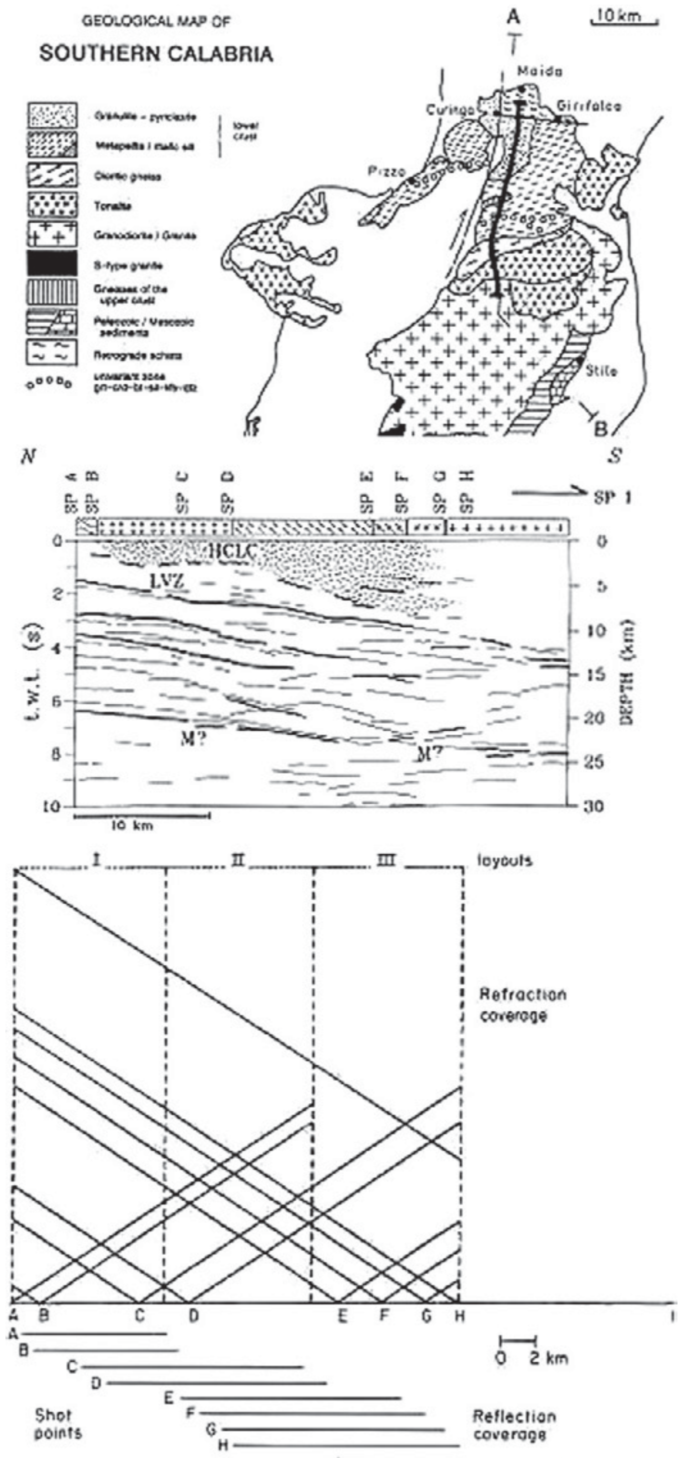



Figure 9.2.3-07. A lower-crust experiment in Calabria, southern Italy (from Lueschen et al., 1992, figs. 1, 4, and 6). Top: Geologic background. Center: Line drawing. Bottom: Layout. [Terra Nova, v. 4, p. 77–86. Copyright Wiley-Blackwell.]

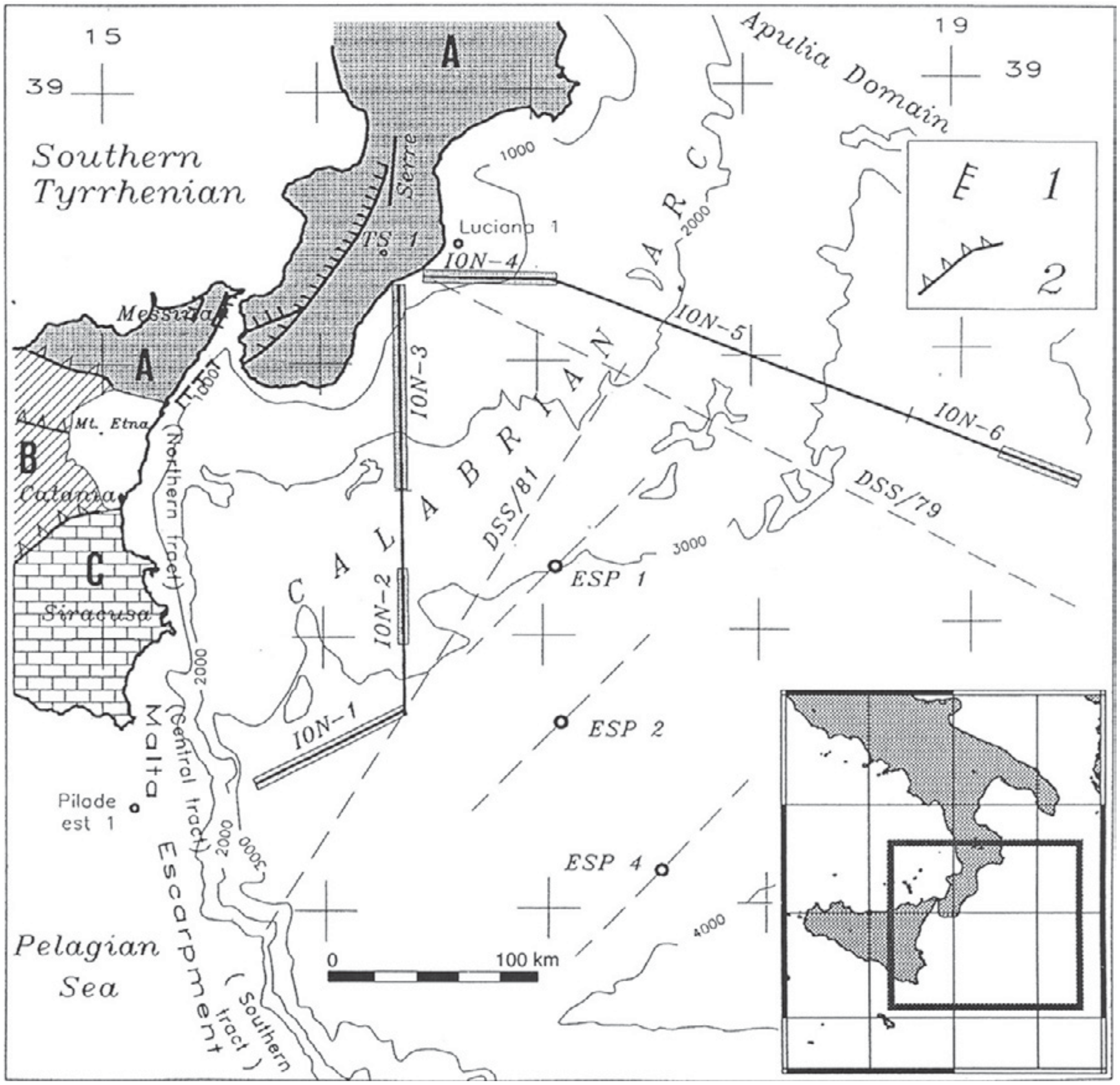



Figure 9.2.3-08. Location map of the STREAMERS profiles (ION-1 to -6) in the western Ionian Sea (from Cernobori et al., 1996, fig. 1). [Tectonophysics, v. 264, p. 175–189. Copyright Elsevier.]

also shows the sites of the earlier seismic investigations of the area (Ferrucci et al., 1991; Makris et al., 1986; de Voogd et al., 1992), carried out in 1979 and 1981 (DSS79, DSS81, and ESP [expanding spread profile] locations). One of the principal features of the multichannel reflection data beneath the Ionian basin was a band of "layered" high-amplitude reflections near the base of the crust, whereby the traveltimes increased gradually from the basin toward the southern and eastern margins of Calabria.

A more detailed investigation with additional shorter lines (Fig. 9.2.3-09) west and north of the long offshore lines ION-1, -2, and -3 of the 1992 survey followed in 1993, using industrial-grade reflection profiling with improved marine sources, and aiming to investigate the offshore region east of Sicily near Mount Etna (Nicolich et al., 2000). The inferred Moho depth (Fig. 9.2.3-09) decreases from near 25 km at the shore lines to 17–18 km in the Ionian basin.

In the eastern Ionian Sea, a 180-km-long seismic-reflection profile, augmented by a few wide-angle seismometer stations on land, has been recorded to the west of the Peleponnesus, Greece (Fig. 9.2.3-10), using a powerful airgun source (Hirn et al., 1996), in order to investigate the crust in the area where the Ionian Sea plate is being subducted beneath the western Hellenides. This profile, ION-7, extended from the deep Ionian basin into the Gulf of Patras. East of the Ionian Islands, the reflectivity pattern suggested a rather thin continental-type crust of ~23 km thickness.

Under the western slope of the islands, a major normal-incidence reflector dipping eastward was found. In particular, a bright reflector at 13 km depth was detected, which had

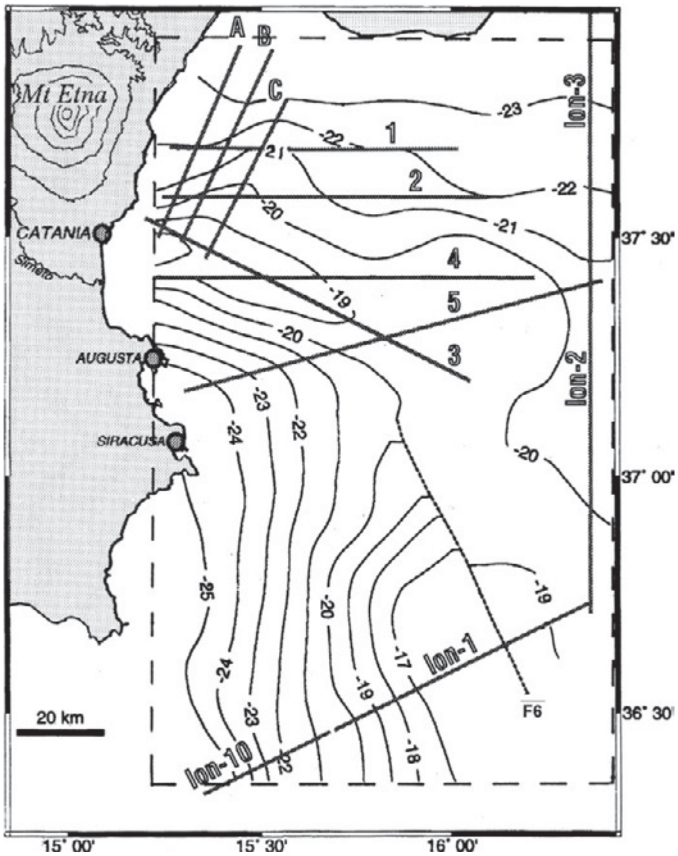



Figure 9.2.3-09. Location of seismic lines and Moho contours in the western Ionian Sea east of Sicily (from Nicolich et al., 2000, fig. 7). [Tectonophysics, v. 329, p. 121–139. Copyright Elsevier.]

been suggested as being the interface along which the western Hellenides override the African plate.

To investigate this structure in more detail, in 1997 the French-Greek project SEISGRECE was performed around the Ionian islands west of the Peloponnesus, consisting of a number of marine multichannel reflection profiles and OBS and land-based stations to record refraction and wide-angle reflection data (Clément et al., 2000).

From three of the multichannel seismic profiles of the projects STREAMERS and SEISGRECE, Laigle et al. (2000) determined an active fault which they could image to 10 km depth.

During the SEISGRECE project of 1997, an airgun profile was also recorded with shots every 50 m in the part of the Gulf of Corinth to the east of Aigion (Sachpazi et al., 2003; Clément et al., 2004). Along the N 120° E axis of the Gulf of Corinth, a multichannel reflection seismic profile was shot in the 0.9 km deep marine basin and OBS refraction lines were recorded that penetrated the basement. Furthermore, multichannel seismic reflection profiles, which had been shot earlier with a Maxipulse explosive source at 25-m intervals into a 96-channel, 2.4-km-long streamer, were reprocessed for a joint interpretation. The latter two profiles were NNW-SSE transects which crossed

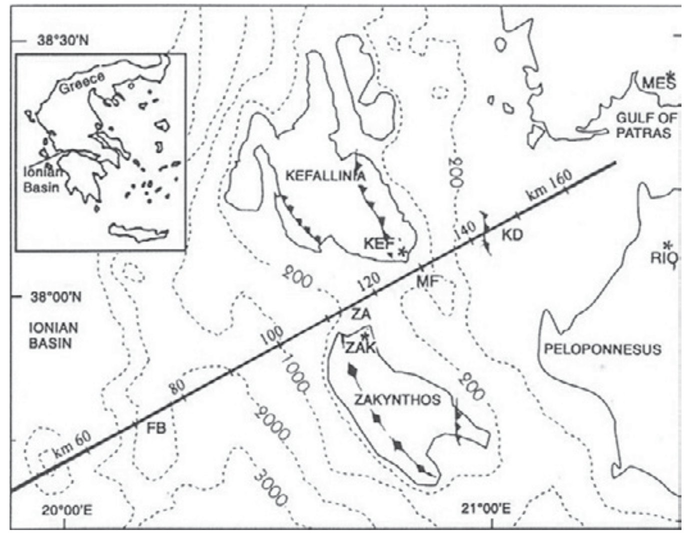



Figure 9.2.3-10. Location map of the STREAMERS profiles ION-7 in the eastern Ionian Sea west of the Peloponnesus, Greece (from Hirn et al., 1996, fig. 1). [Tectonophysics, v. 264, p. 35–49. Copyright Elsevier.]

almost orthogonally the deep marine basin of the Gulf (Sachpazi et al., 2003). Using an array of 14 airguns, a source with maximum efficiency became available, because a signal was provided that is peaked in the rather low-frequency 12–20 Hz but has a duration short enough for acceptable resolution for seismic normal-incidence reflections. Ocean bottom seismometers and land stations, offset at either end of the shot line along the axis of the Gulf of Corinth recorded clear seismic waves out to the maximum recording distance of 105 km. The Moho was found at 40 km depth under the western Gulf north of Aigion and 32 km under its north coast, north of Corinth (Clément et al., 2004).

Another shallow-marine seismic survey was performed in the eastern Mediterranean between south of Cyprus and the Syrian coast revealing a thick sedimentary sequence in the Levantine Basin with the base of the Mesozoic sedimentary unit at 7.8–8.0 s TWT, being interpreted as top of the crystalline basement (Vidal et al., 2000).

### 9.2.4. Eastern Europe

When the EUROPROBE program started in 1993, the investigation of the Trans-European Suture Zone, separating the mobile western Hercynian Europe with its predominantly 30-km-thick crust from the East European Platform and adjacent Baltic Shield with 40–50-km-thick crust, became one of its earliest international collaborative research projects (Thybo et al., 1999). While the transition from the Baltic Shield to Hercynian Europe had been one of the objectives of the European Geotraverse in the 1980s, the deep structure of the border zone between Hercynian Europe and the East European platform in Poland, which had always been a target of Polish DSS research since the late 1960s, was now, in the second half of the 1990s, addressed with renewed power.

In 1997, the Polish Academy of Sciences and the University of Warsaw, under the leadership of A. Guterch and M. Grad, organized a large seismic-refraction program, POLONAISE (Polish Lithosphere Onsets—An International Seismic Experiment), to investigate the deep structure of the lithosphere under the Trans-European Suture Zone in east-central Europe and the southwestern portion of the Precambrian East European craton in the area of northwestern Poland (Fig. 9.2.4-01).

The seismic-refraction wide-angle reflection experiment was conducted in May 1997 and included contributions from the geophysical and geological communities in Poland, Denmark, United States, Germany, Lithuania, Finland, Sweden, and Canada (Guterch et al., 1999). The project involved two deployments of 613 refraction instruments, the majority of which came from the University of Texas and the IRIS/PASSCAL pool, and five multichannel (90 or 120 channels) seismic-reflection stations to record shots along five interlocking profiles with a total length of ~2000 km, covering 800 sites, which, on average, were spaced 1.5 km on profile 4 and 3 km on the other lines. By this simultaneous recording of shots on several profiles, a good 3-D coverage was ensured. Sixty-four explosions were fired in 40 m deep boreholes, each loaded with 50 kg of TNT. The total charges varied from 100 to 1000 kg. Distances between shotpoints were 15–35 km. The shot efficiency was excellent. In general, clear first arrivals and later phases were obtained to the farthest offsets of ~600 km from the shotpoints. As the two sample record section demonstrate, the data showed striking differences between the timing and appearance of seismic phases, depending on whether the shots were located southwest (Fig. 9.2.4-02) or northeast (Fig. 9.2.4-03) of the Trans-European Suture Zone (Guterch et al., 1999; see also Appendix A9-1-3).

An example of the modeling procedure is shown in Figure 9.2.4-04 for profile 4. This line started in Germany, ran through the whole of Poland, and extended through most of Lithuania, covering a distance of ~900 km. The model was obtained by forward ray-tracing using the SEIS83 package of Červený and Pšenčik (1984c). It clearly shows the stepwise change of crustal thickness, from near 30 km underneath the Paleozoic platform of Hercynian Europe, from where the Moho increases drastically to almost 45 km under the Trans-European Suture Zone and farther to near 50 km under the Precambrian East European Platform. Figure 9.2.4-04 also demonstrates the enormous thickness

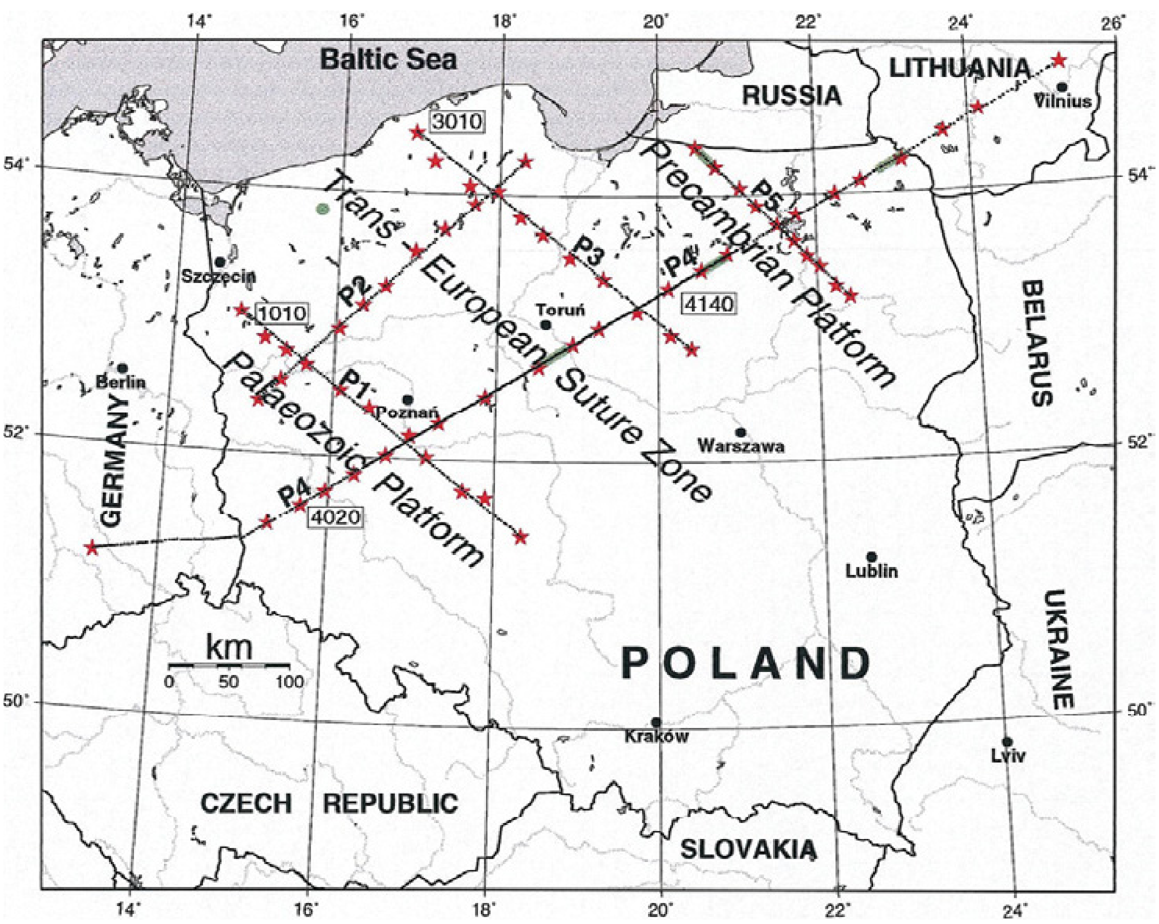



Figure 9.2.4-01. Location of POLONAISE '97 seismic refraction profiles P1 to P4 in northern Poland (from Guterch et al., 1999, fig. 2). [Tectonophysics, v. 314, p. 101–121. Copyright Elsevier.]

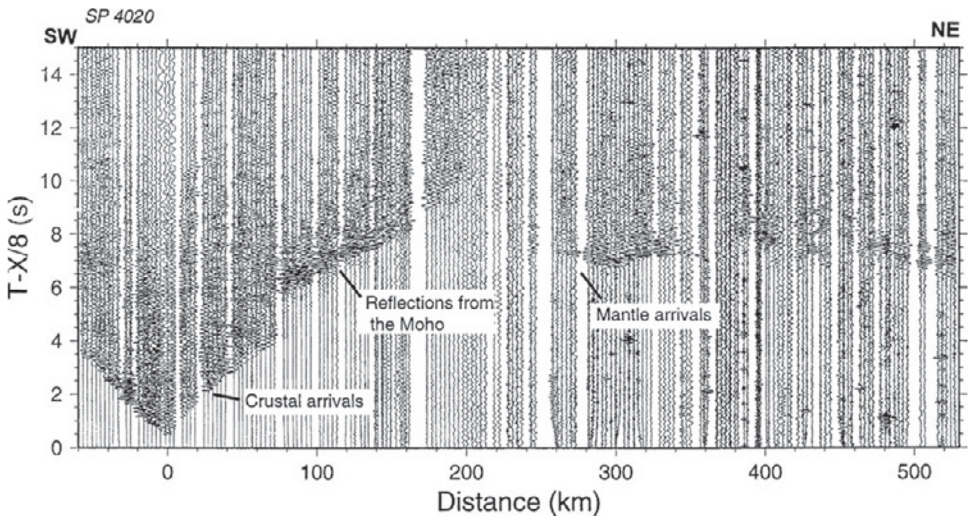



Figure 9.2.4-02. Data example of POLONAISE '97 seismic refraction profile P4, with shotpoint (SP 4010) location SW of Trans-European Suture Zone (from Guterch et al., 1999, fig. 11). [Tectonophysics, v. 314, p. 101–121. Copyright Elsevier.]

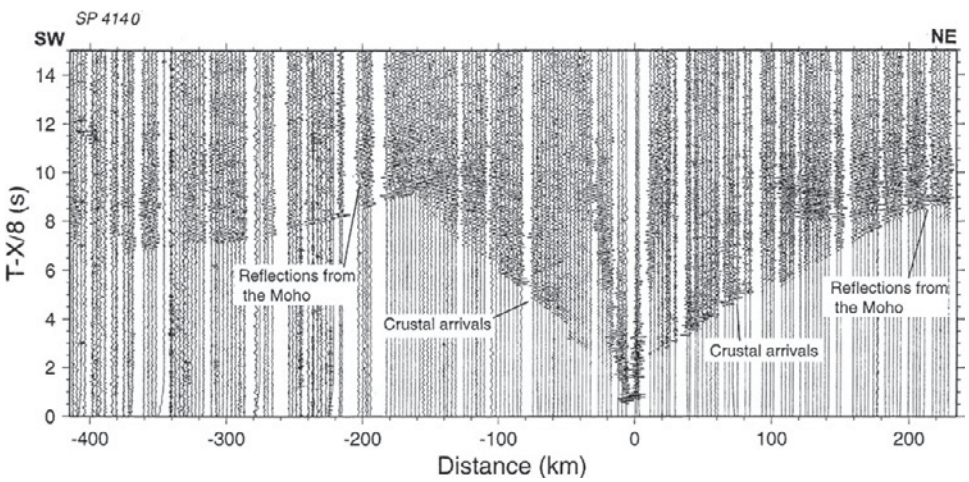



Figure 9.2.4-03. Data example of POLONAISE '97 seismic refraction profile P4, with shotpoint (SP 4140) location NE of Trans-European Suture Zone (from Guterch et al., 1999, fig. 12). [Tectonophysics, v. 314, p. 101–121. Copyright Elsevier.]

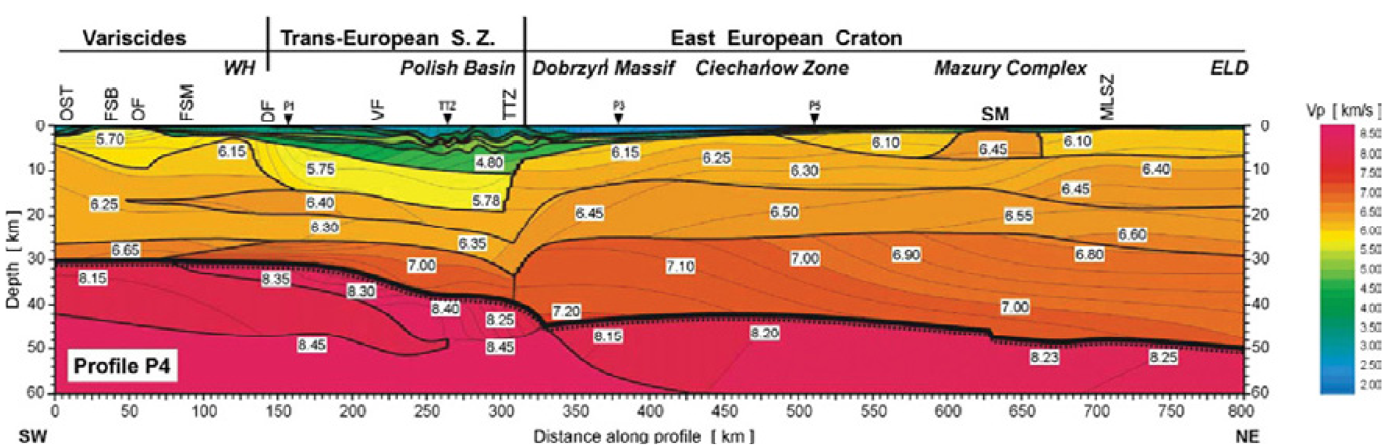



Figure 9.2.4-04. Two-dimensional P-wave crustal model along the POLONAISE profile P4 (from Dadlez et al., 2005, fig. 2). [Tectonophysics, v. 411, p. 111–128. Copyright Elsevier.]

of sediments, filling the Polish Basin above the Trans-European Suture Zone, with velocities of less than 6 km/s. The basin has maximum depths of 16–20 km.

In only another three years, the organizers of POLONAISE undertook a project of even larger size, the project CELEBRATION 2000 (Central European Lithospheric Experiment Based on Refraction). Scientific organizations in Poland, Hungary, the Czech Republic, Slovakia, Austria, Russia, Belarus, and Germany provided the sources in their countries. This project involved the participation of 28 institutions from Europe and North America and targeted the Trans-European Suture Zone region, as well as the southwestern portion of the East European craton (southern Baltica), the Carpathian Mountains, the Pannonian basin, and the Bohemian massif (Fig. 9.2.4-05). The profiles covered southeastern Poland, the Czech Republic, Slovakia, and Hungary, and reached into Austria in the southwest, into Germany in the west, and into Belarus and Russia in the northeast (Guterch et al., 2001).

The layout of the CELEBRATION 2000 experiment was organized similarly to POLONAISE 1997. The project comprised a network of interlocking seismic-refraction profiles (Fig. 9.2.4-05) with a total length of ~8900 km. Station spacing along the profiles was either 2.8 or 5.6 km. One hundred forty-seven shots, ranging in size from 90 to 15,000 kg, were fired along most of the profiles, so that ~5400 km of traditional profile data were obtained. In addition, a large array of 3-D coverage resulted. To accomplish this network of profiles, three deployments were required, which were completed within a period of one month (June 2000). In total, ~100,000 usable seismograms were obtained. About 840 of the 1230 instruments were RefTek

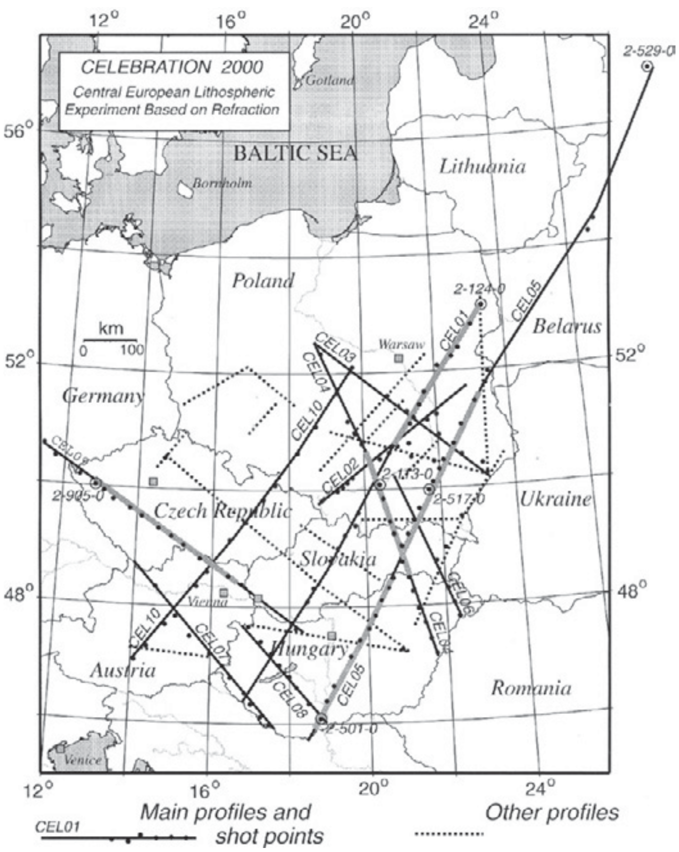



Figure 9.2.4-05. Location of CELEBRATION 2000 profiles (CEL) in central-eastern Europe (from Guterch et al., 2003a, fig. 2). [Studia Geophysica et Geodaetica, Academy of Science, Czech Republic, Prague, v. 47, p. 659–669. Permission granted by Studia Geophysica et Geodaetica, Prague, Czech Republic.]

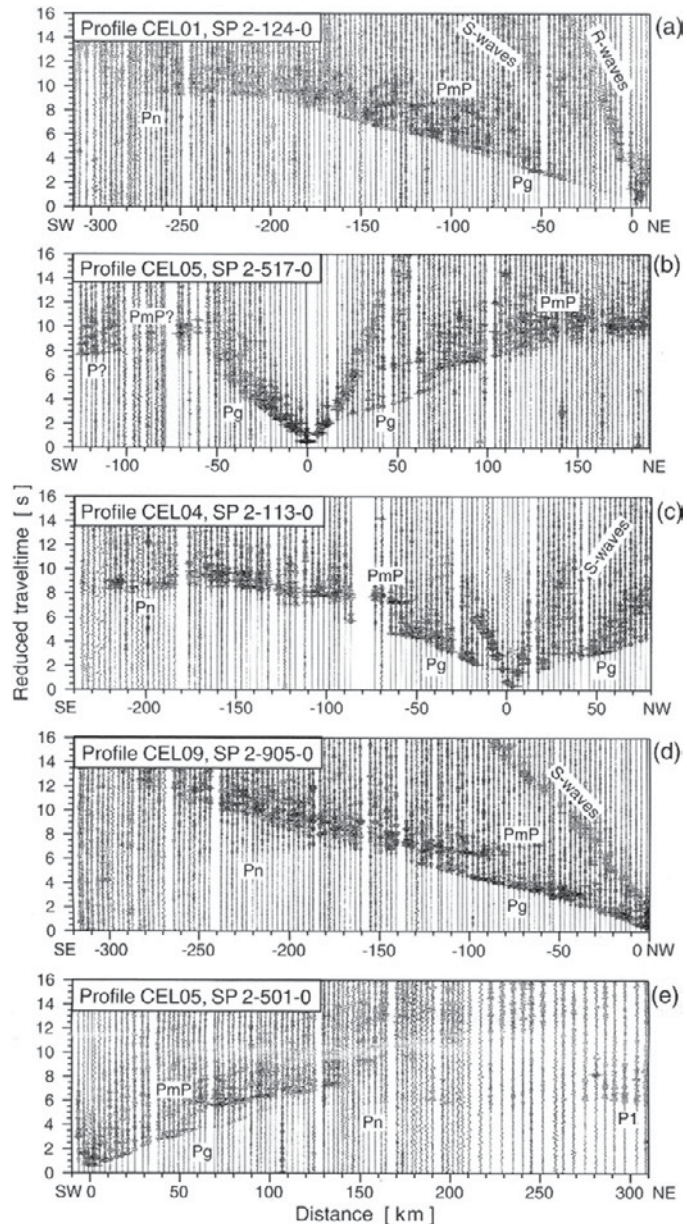



Figure 9.2.4-06. Record sections from different tectonic regions traversed by CELEBRATION 2000 profiles (CEL) in central-eastern Europe (from Guterch et al., 2001, fig. 3). [Eos (Transactions, American Geophysical Union), v. 82, p. 529, 534–535. Reproduced by permission of American Geophysical Union.]

125 (“Texan”) recorders, 640 of them from the joint pool of IRIS/PASSCAL and the University of Texas at El Paso, and 200 from European partners: Austria—15, Denmark—100, Finland—10, Poland—15, and Turkey—60 (Guterch et al. (2003b). In addition, 50 three-component RefTek 72 instruments of the IRIS/PASSCAL pool were deployed in the Czech Republic, and another 20 three-component stations in Poland (Polish MK-4P units) for one month to record shots and earthquakes. Figure 9.2.4-06 shows record sections recorded in different tectonic regimes: (a) East European craton, (b) Trans-European Suture Zone region, (c) Carpathian Mountains, (d) Bohemian Massif, and (e) Pannonian Basin (see also Appendix A9-1-4).

The longest line, CEL05, was 1400 km long, traversed the East European craton, the Trans-European Suture Zone, the Carpathians, and the Pannonian basin, and provided the key information on the lithosphere (Guterch et al., 2003b). For the first modeling, a tomographic approach after Hole (1992), using first arrivals only, was applied (Guterch et al., 2003b). Later forward ray-tracing modeling (Fig. 9.2.4-07) used the full P-wave field (Grad et al., 2006).

The resulting crustal thickness along this line varied from 43 to 50 km beneath the East European craton to more than 50 km under the southernmost unit of the Trans-European Suture Zone (crossing point of lines CEL03 and CEL05; see Fig. 9.2.4-08), and decreased to 30 km under the Carpathians and less than 23 km under the Pannonian Basin. While under the East European craton a three-layer crust prevailed, the crust under the Carpathians and the Pannonian basin appeared to consist of only two distinct layers with relatively low velocities (Fig. 9.2.4-07). The uppermost mantle velocity also varied, from 8.25 km/s and more under the craton to 8.0 km/s and less under the Carpathian-Pannonian area.

Another profile, CEL03, 700 km long, targeted central Poland along the Trans-European Suture Zone, a key target of the Polish crustal structure investigations already in earlier decades, also named TTZ after its discoverers Teisseyre and Tornquist in the nineteenth century (Teysseire and Teysseire, 2003). The detailed interpretation of profile CEL03 revealed that the Trans-European Suture Zone is not uniform, but has to be subdivided into several units (Fig. 9.2.4-08) with crustal thickness increases from near 30 km near the Baltic Sea in the northwest to 50 km and more near the Ukrainian border in the southeast (Janik et al., 2005).

The 450-km-long line CEL09 (Fig. 9.2.4-05) served to investigate the deep structure of the Bohemian Massif (Hrubcova et al., 2005). Its northwestern end in eastern Germany crosses the German seismic lines GRANU’95 and MVE’90 and runs

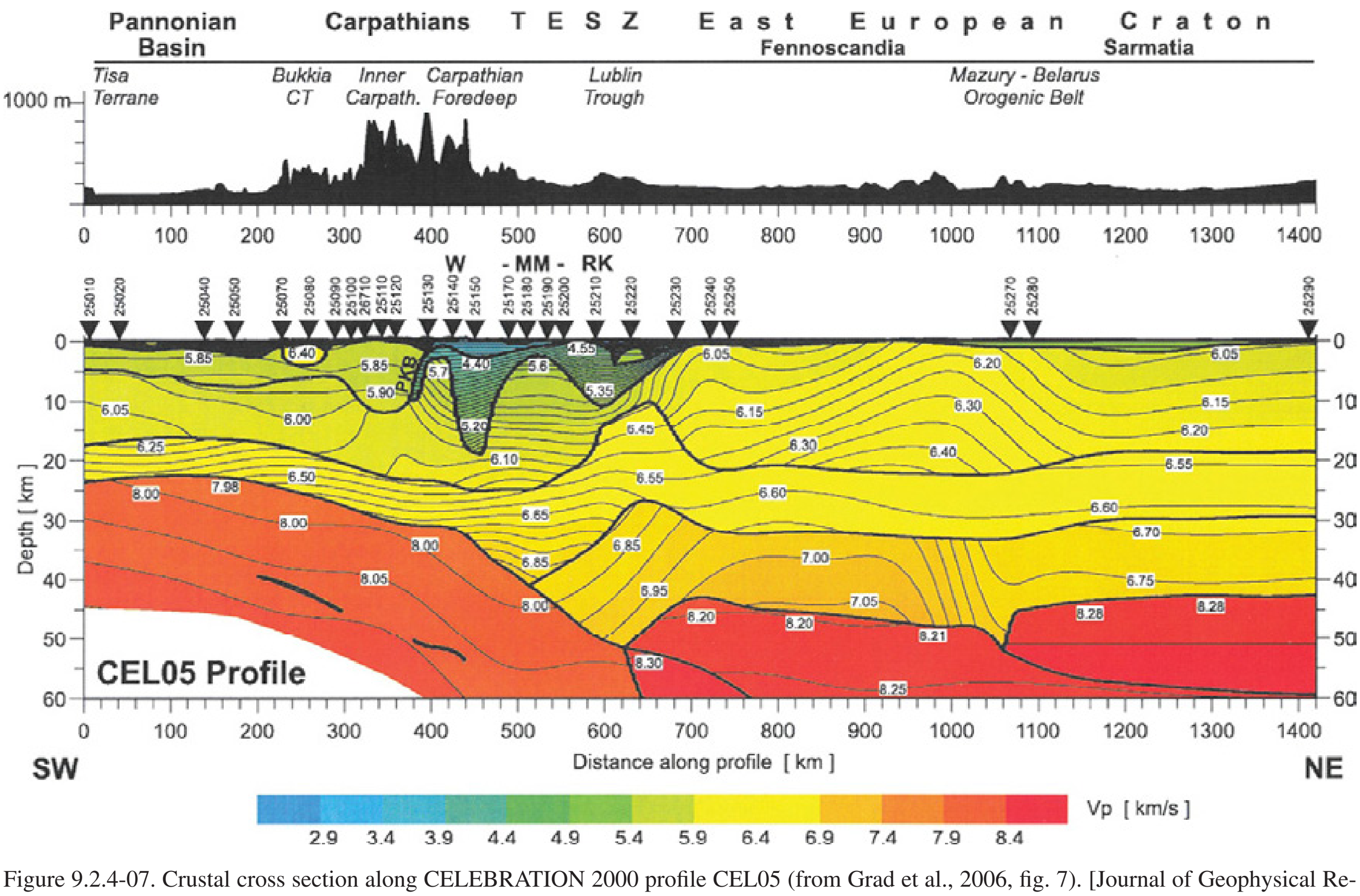



Figure 9.2.4-07. Crustal cross section along CELEBRATION 2000 profile CEL05 (from Grad et al., 2006, fig. 7). [Journal of Geophysical Research, v. 111: B03301, doi:10.1029/2005JB003647. Reproduced by permission of American Geophysical Union.]

almost parallel to the reflection line 9HR (Tomek et al., 1997) within the Czech Republic. The highly reflective lower crust under the Saxothuringian zone in the northwest terminates at the Moho at 26–35 km depth, while the central Moldanubian crust reaches up to 39 km in thickness and is underlain by a pronounced Moho with the strongest velocity contrast. To the southeast, a thick crust-mantle transition zone ranges from 23 to 40 km depth (Fig. 9.2.4-09).

One of the particular tectonic features in central-eastern Europe is the Pannonian basin, with an extremely thin crust of 20–25 km which is surrounded by orogenic regions of much thicker crust. Its thick Neogene cover has been the goal of an ongoing program for deep exploration. The PANCARDI project within the EUROPROBE program in the 1990s became a driving force for a detailed review of existing and new reflection seismic profiles within the Pannonian basin (Horvath et al., 2006).

Here, in 1992, with the support of Canadian scientists, the Pannonian Geotraverse (Fig. 9.2.4-10) was launched as an extension of previous investigations (Hajnal et al., 1996). Using a special observation scheme for recording units, normally used in large-scale refraction profiling, special seismic-reflection measurements were arranged. Expecting narrow-band, low-frequency signals, a 100-km-long seismic-reflection profile was designed, using 200 Canadian PRS units and 12 Swiss recorders. A mobile reflection spread comprised 195 of the PRS units, while the remaining stations were permanently placed at the ends of the line. Fifty-kg explosive sources, placed at 65 m depth, generated excellent record quality, in which the Moho reflections were generally visible already in the field monitors. The results confirmed the existence of a thin crust with 25–27 km thickness, with the crust being decoupled by well-defined detachment faults and the Moho dipping gently eastwards.

A more extended but less detailed coverage of the Pannonian basin followed in 2000 with CELEBRATION 2000, which has been described above, and in 2002 and 2003, during the long-range refraction profiling programs ALP 2002 and SUDETES 2003, which will be described in Chapter 10.

While CELEBRATION 2000, ALP 2002, and SUDETES 2003 targeted also the western Carpathians in Poland, Slovakia, and Hungary, the Carpathians of Romania became the goal of a joint German-Romanian research program, aiming to study the seismically high-risk Vrancea area at the center of the southeastern Carpathians and the causes of the strong and deep-seated (around 100 km depth) intermediate-upper-mantle earthquakes.

Seismic-refraction studies and a large-scale tomography study of southeastern Romania were part of an interdisciplinary

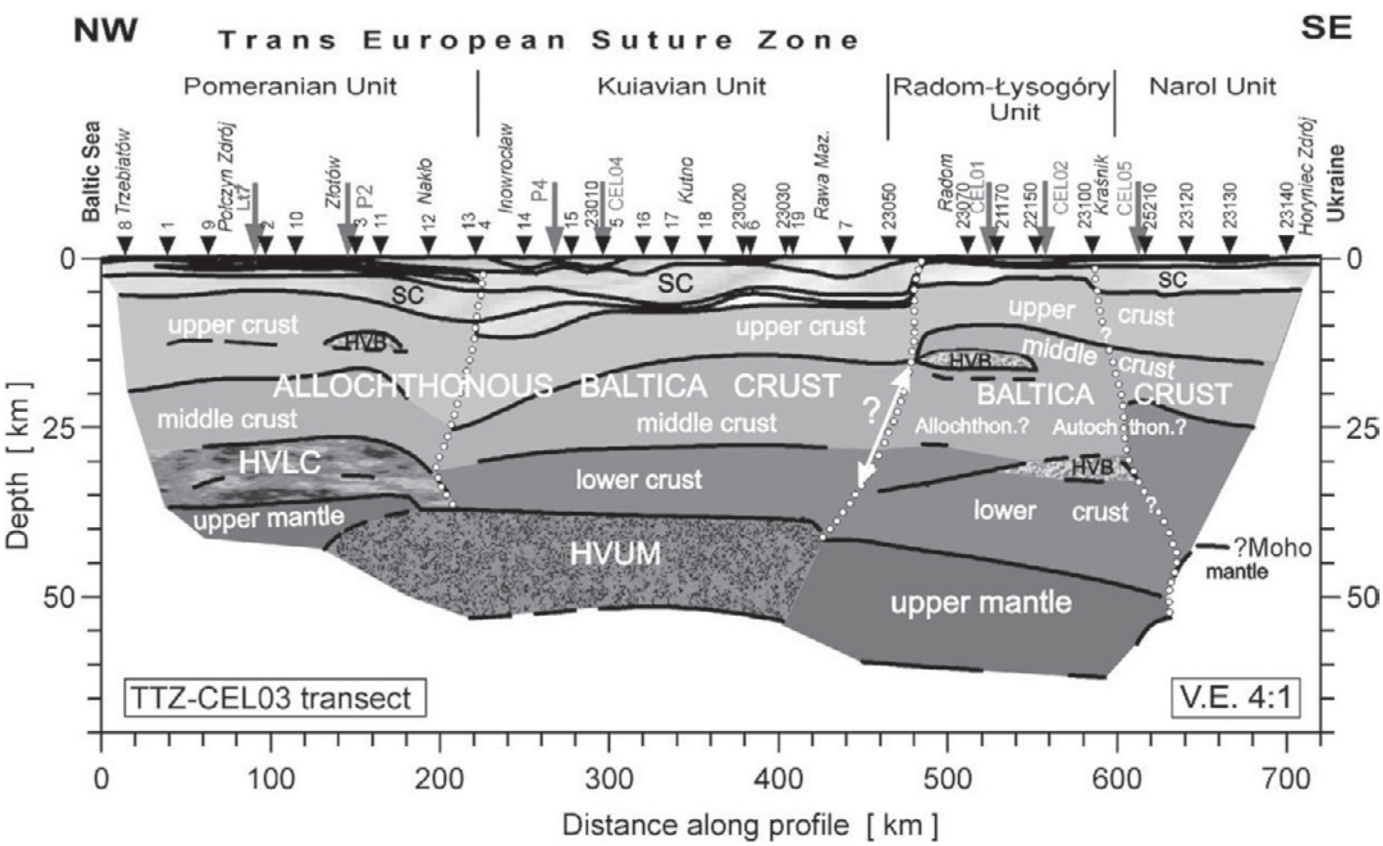



Figure 9.2.4-08. Crustal structure sketch of CELEBRATION 2000 profile CEL03 (from Janik et al., 2005, fig. 12). SC—sedimentary cover, HVB—high-velocity body, HVLC—high-velocity lower crust, HVUM—high-velocity upper mantle. [Tectonophysics, v. 411, p. 129–156. Copyright Elsevier.]

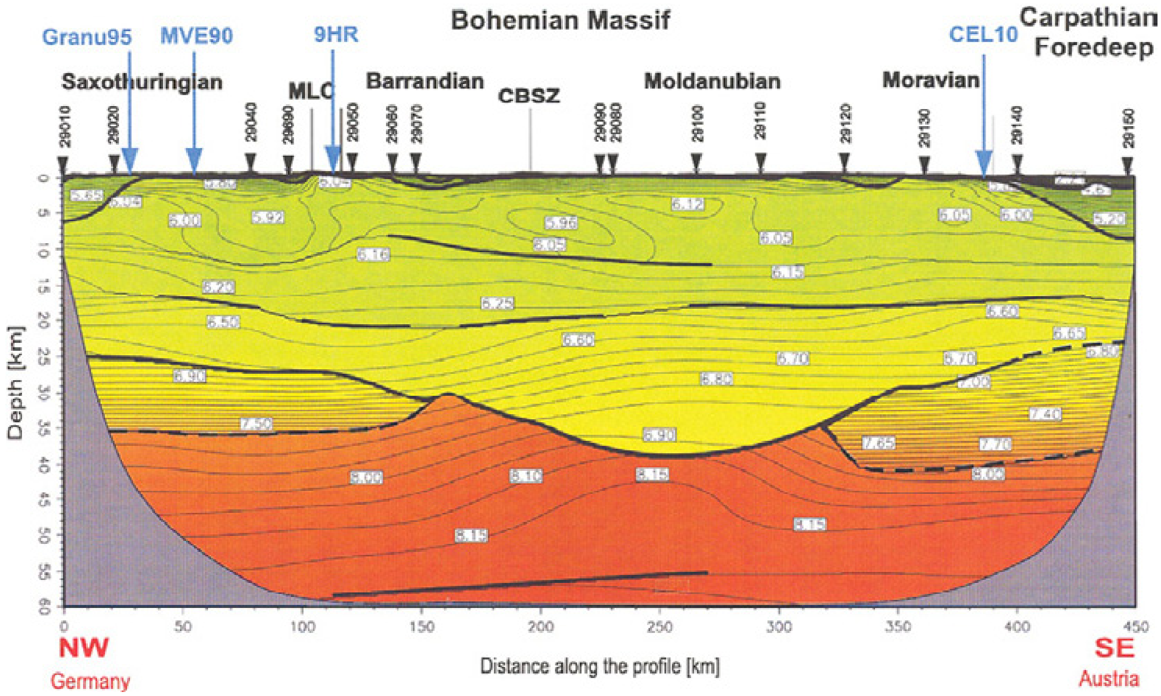



Figure 9.2.4-09. Crustal structure along CELEBRATION 2000 profile CEL09 traversing the Bohemian Massif (from Hrubcova et al., 2005, fig. 6) [Journal of Geophysical Research, v. 110: B11305, doi:10.1029/2004JB003080. Reproduced by permission of American Geophysical Union.]

research program, carried out by the Collaborative Research Center 461 (CRC 461) "Strong Earthquakes—a Challenge for Geosciences and Civil Engineering" at the University of Karlsruhe (Germany) and the Romanian Group for Vrancea Strong Earthquakes (RGVE) at the Romanian Academy in Bucharest (Wenzel, 1997; Wenzel et al., 1998a, 1998b; Martin et al., 2005, 2006).

Two major active-source seismic-refraction experiments were carried out in the eastern Carpathians of Romania in 1999 and 2001 (Fig. 9.2.4-11) as a contribution to this research program (Hauser et al., 2001, 2002, 2007a). They were designed to study the crustal and uppermost mantle structure to a depth of ~70 km underneath the Vrancea epicentral region and were jointly performed by the Geophysical and Geological Institutes of the University of Karlsruhe (Germany), the National Institute for Earth Physics in Bucharest (Romania) and the University of Bucharest (Romania).

The VRANCEA'99 seismic-refraction experiment (Hauser et al., 2001) was a 300-km-long refraction profile and was

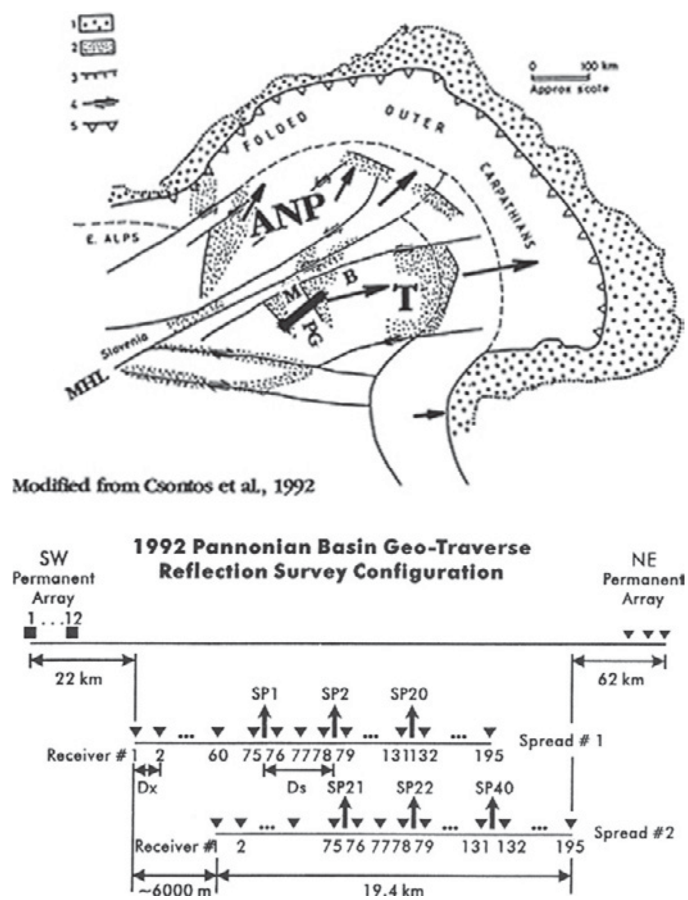



Figure 9.2.4-10. Top: Tectonic position of the Pannonian Basin, PG—location of the 1992 Pannonian Geotraverse (from Hajnal et al., 1996, fig. 3). ANP—Alpine North Pannonian unit, T—Tisza unit, MHL—Mid Hungarian lineament, M—Mako trough, B—Bekes basin. Bottom: Survey configurations of the 1992 Pannonian Geotraverse data acquisition program (from Hajnal et al., 1996, fig. 4). Dx—100 m, Ds—300 m. Total number of shots: 289, record length: 60 s, total coverage 95 km, nominal fold: 24. [Tectonophysics, v. 264, p. 191–204. Copyright Elsevier.]

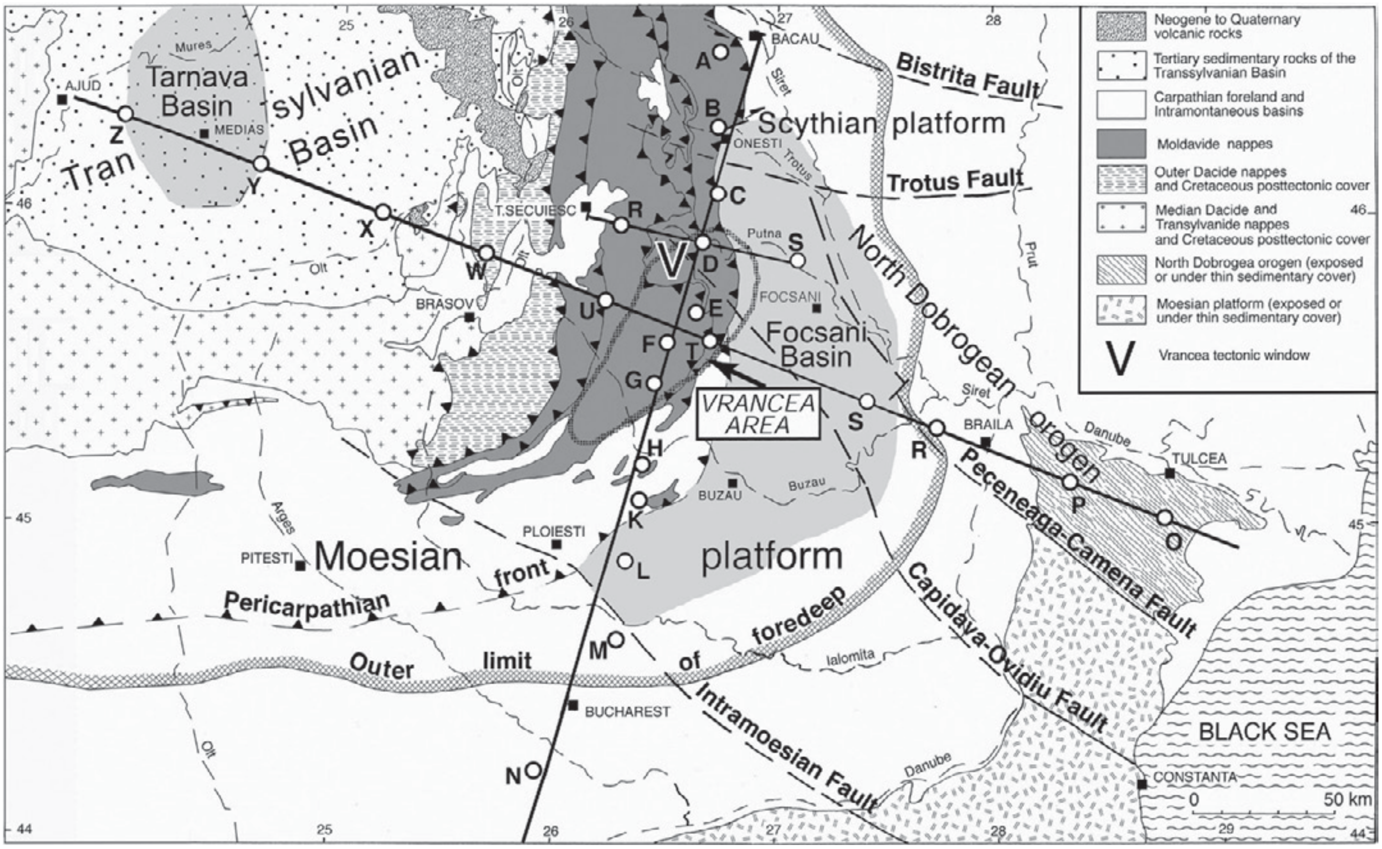



Figure 9.2.4-11. Location map of seismic-refraction VRANCEA 1999 (N-S line) and 2001 (W-E line) profiles through the seismogenic Vrancea zone in the eastern Carpathians of Romania (from Hauser et al., 2002). V—Vrancea seismogenic zone with strong intermediate-depth earthquakes between 70 and 160 km depth. [Eos (Transactions, American Geophysical Union), v. 83, p. 457, 462–463. Reproduced by permission of American Geophysical Union.]

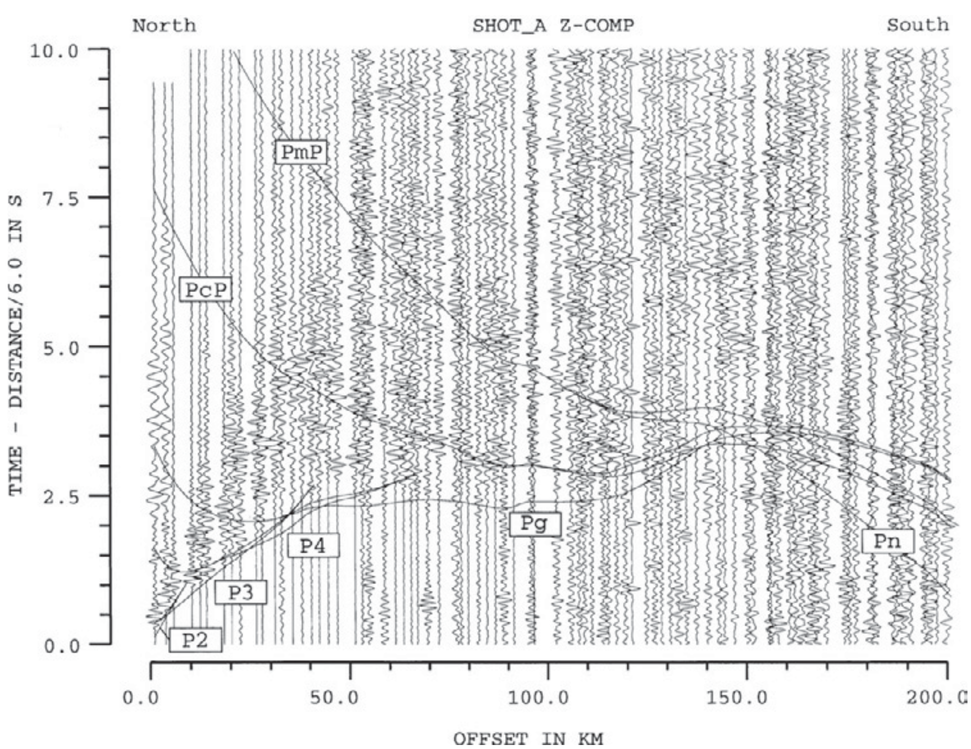



Figure 9.2.4-12. Record section from the northernmost shotpoint A along the north-south profile VRANCEA 1999 (from Hauser et al., 2001, fig. 5a). P1–P4—first arrivals from sedimentary cover; Pg, PcP, PmP—reflections from top basement, top lower crust and Moho; [Tectonophysics, v. 340, p. 233–256. Copyright Elsevier.]

recorded between the cities of Bacau and Bucharest with 12 shotpoints A to N, traversing the Vrancea epicentral region in a NNE-SSW direction (Fig. 9.2.4-11). It was complemented by a short perpendicular line, which ran transverse to the geological structures along the Putna Valley, north of the Vrancea seismogenic zone (V in Fig. 9.2.4-11), intersected the main refraction line at shotpoint D, and had two additional shotpoints R and S (north of V in Fig. 9.2.4-11).

Along these two segments, a total of 140 recording sites were occupied with an average station spacing of 2 km. Seismic recording equipment for the experiment was provided by the GeoForschungsZentrum Potsdam (Germany), by Leicester University (UK), and by the NERC geophysical equipment pool (UK). The spacing of shotpoints ranged from 12 to 30 km, with an average of 22 km. The charge sizes varied from 300 kg to 900 kg with the larger shots at the end points of the main line (Fig. 9.2.4-11). All shots were recorded simultaneously along the main line and the transverse line. Hauser et al. (2000; Appendix A9-1-5) and Landes et al. (2004b) compiled data and other technical details in data reports. A data example is shown in Figure 9.2.4-12 (for more data, see Appendix A9-1-5; Hauser et al., 2000, 2001).

The P-wave interpretation (Fig. 9.2.4-13) showed up to 10-km-thick sedimentary sequences. Both the sedimentary and crystalline upper-crustal layers appeared to thicken underneath the seismogenic Vrancea zone and the adjacent southern margin of the Carpathians. As a result, the intra-crustal discontinuity varies strongly in depth, between 18 km at both ends of the line and 31 km at its center. Strong wide-angle $P_MP$ reflections indicated the existence of a first-order Moho, the depth of which also varies from 30 km near the southern end of the line to 41 km near the center and to 39 km near the northern end. Within the uppermost mantle, a low-velocity zone was interpreted from a reflection, named $P_LP$, which defined the bottom of this low-velocity layer at a depth of 55 km (Hauser et al., 2001). The data also showed reasonable appearance of S-waves, which were incorporated into the P-wave model (corresponding velocities are shown in brackets in Fig. 9.2.4-13) and Poisson's ratios were calculated for the individual crustal layers (Raileanu et al., 2005).

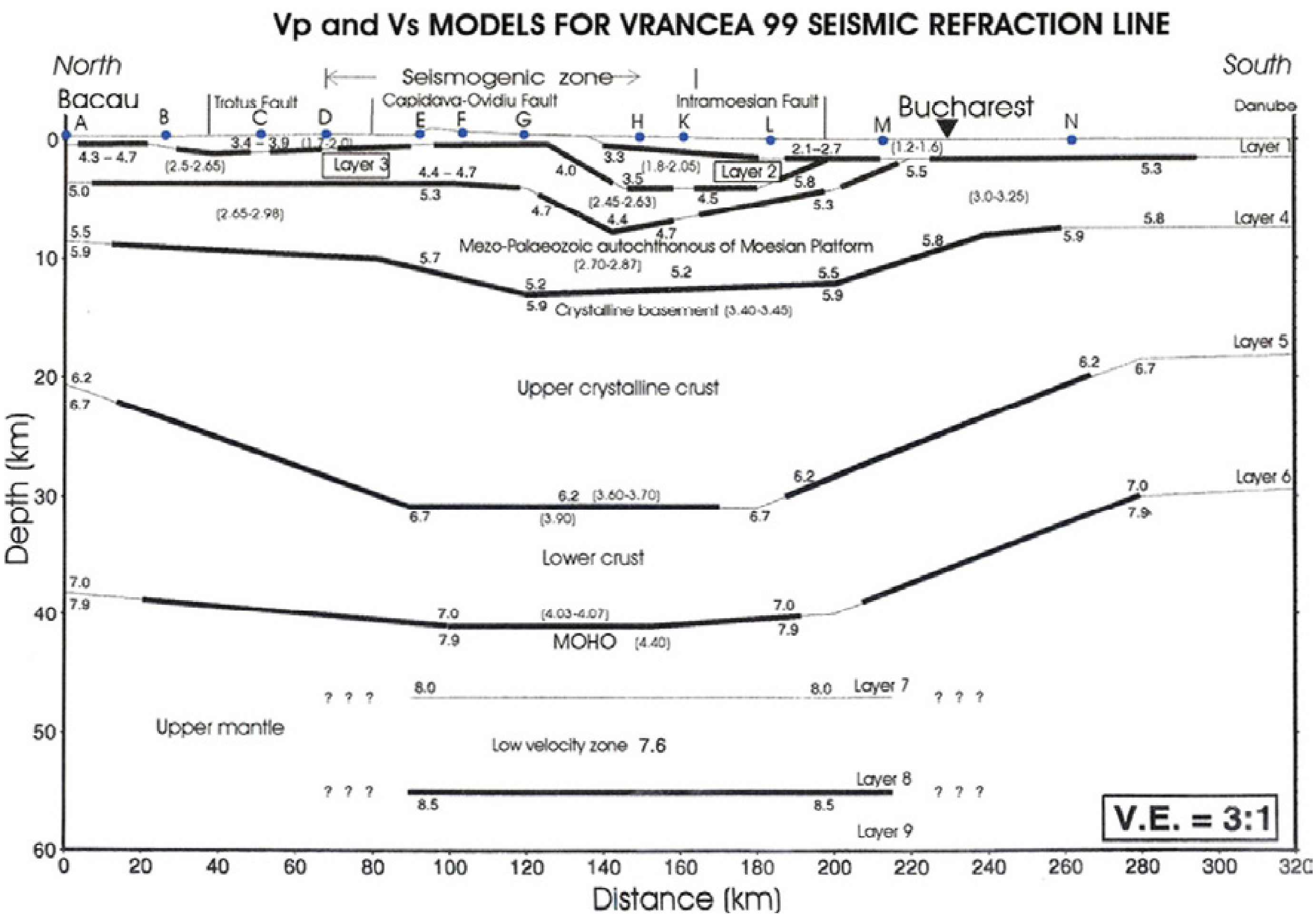



Figure 9.2.4-13. Crustal P-wave cross section along the north-south profile VRANCEA 1999 (from Hauser et al., 2001, fig. 9). Velocities in brackets are corresponding S-velocities. [Tectonophysics, v. 340, p. 233–256. Copyright Elsevier.]

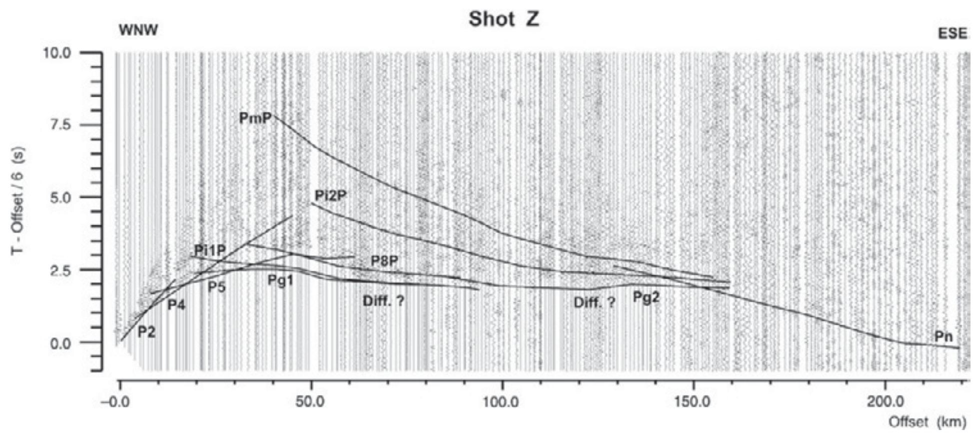



Figure 9.2.4-14. Record section from the westernmost shotpoint Z in Romania along the west-east profile VRANCEA 2001 (from Hauser et al., 2007a, fig. 9). P1–P5—first arrivals from sedimentary cover; Pgx, PxP—refracted/reflected phases from crustal discontinuities; PmP, Pn—reflected/refracted phases from Moho. [Tectonophysics, v. 430, p. 1–25. Copyright Elsevier.]

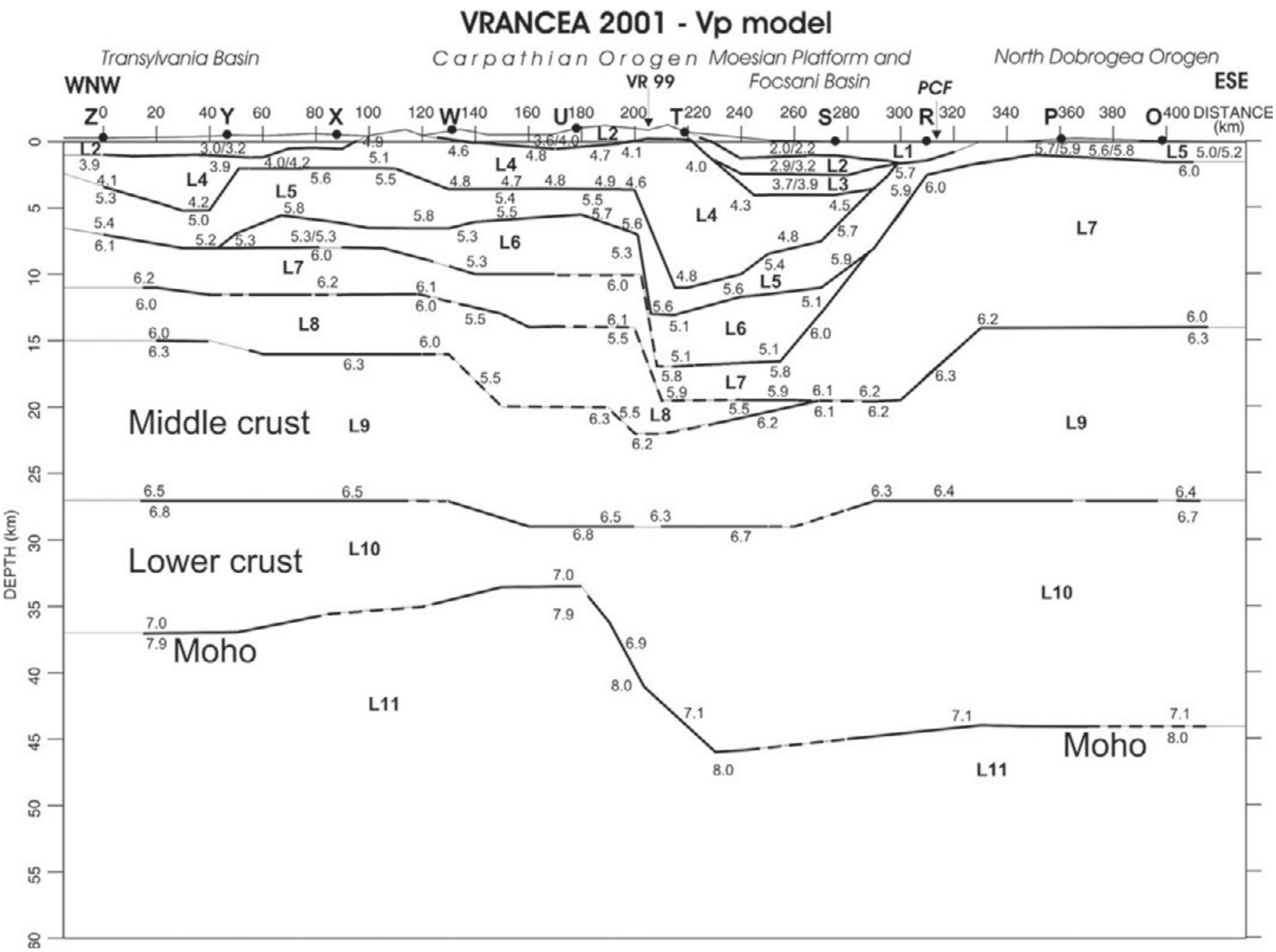



Figure 9.2.4-15. Crustal P-wave cross section along the west-east profile VRANCEA 2001 (from Hauser et al., 2007a, fig. 11). [Tectonophysics, v. 430, p. 1–25. Copyright Elsevier.]

The second experiment across the Vrancea zone, VRANCEA 2001, was recorded in August/September 2001. Because of its close relation to the VRANCEA'99 project, its details will be described here in Chapter 9. VRANCEA'2001 was a 700-km-long WNW-ESE–trending seismic-refraction line in Romania with a short extension into Hungary (Hauser et al., 2002, 2007a). The main part (Fig. 9.2.4-11) ran from the Transylvanian Basin across the East Carpathian Orogen and the Vrancea seismic region to the foreland areas with the very deep Neogene Focsani Basin and the North Dobrogea Orogen near the Black Sea. Ten large drillhole shots with 300–1500 kg charges (O to Z in Fig. 9.2.4-11) were fired in Romania between the town of Aiud at the western margin of the Transylvanian Basin and the Black Sea, which resulted in an average shotpoint spacing of 40 km. To the west an additional shot (500 kg) was fired in Hungary. These shots generated 11 seismogram sections each with almost 800 traces (data examples in Fig. 9.2.4-14 and Appendix A9-1-5; Hauser et al., 2007a). The spacing of the geophones was variable. It was around 1 km from the eastern end to Aiud (~450 km length), 6 km from Aiud to Oradea (Romanian-Hungarian border), and ~2 km on the Hungarian territory. Between shotpoints T and U, the geophones were deployed at a spacing of 100 m (Panea et al., 2005). In total, 790 recording instruments were available. The small 640 one-component geophones (Texan type) were mostly deployed in the open field outside of localities, while, for safety reasons, the 150 three-components geophones (RefTek and PDAS type) were deployed in guarded properties within towns and villages.

The experiment was jointly performed by the research institutes and universities of Germany, Romania, the Netherlands, and the United States. They included the University of Karlsruhe and The GeoForschungsZentrum Potsdam, Germany, the Free University of Amsterdam, Netherlands, the National Institute for Earth Physics and the University of Bucharest, Romania, and the Universities of El Paso and South Carolina, United States. Field recording instruments were provided by the University of El Paso and PASSCAL, United States (Texan one-component stations), and the GeoForschungsZentrum, Germany (three-component RefTek and PDAS stations).

The data interpretation for the eastern 450-km-long segment (Hauser et al., 2007a) indicated a multi-layered structure with variable thicknesses and velocities. The sedimentary stack comprised up to 6 layers (L1-L6 in Fig. 9.2.4-15) with seismic velocities of 2.0–5.9 km/s and reached a maximum thickness of more than 15 km within the Focsani Basin area. The underlying crystalline crust (L7–L10 in Fig. 9.2.4-15) showed considerable thickness variations in total as well as in its individual subdivisions. The model (Fig. 9.2.4-15) shows the lateral velocity structure of these blocks along the seismic line which remained almost constant with ~6.0 km/s along the basement top and with 7.0 km/s above the Moho. Under the Transylvanian basin, the crust appeared to be 34 km thick with low-velocity zones in its uppermost 15 km. Under the Carpathians, the Moho deepens to beyond 41 km, reaching 46 km under the Focsani basin. However, the crystalline crust does not exceed 25 km in thickness and is covered by up to 15 km of sedimentary rocks. The North

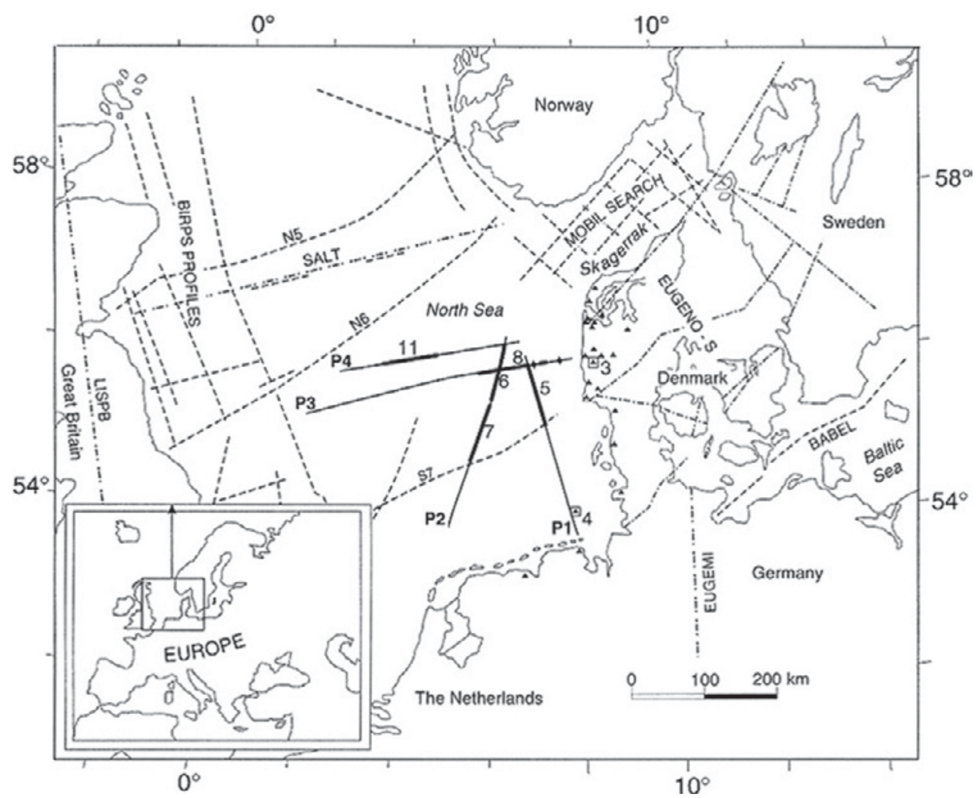



Figure 9.2.5-01. Location map of MONA LISA seismic normal incidence profiles and land recording sites (solid lines P1 to P4 and triangles) and former seismic investigations (dashed lines) in the North Sea area (from MONA LISA Working Group, 1997, fig. 1). [Tectonophysics, v. 269, p. 1–19. Copyright Elsevier.]

Dobrogea crust reaches a thickness of ~43 km and was interpreted as thick eastern European crust overthrusted by a thin 1–2-km-thick wedge of the North Dobrogea Orogen (Hauser et al., 2007a).

The two seismic-refraction campaigns and the teleseismic tomography survey of 1999 (Wenzel et al. 1998b; Martin et al., 2005, 2006) produced a wealth of data, concentrating on the Vrancea zone proper, and allowed for a 3-D tomographic interpretation of the crustal data, which were presented in contour maps at various depth levels (Landes et al., 2004a).

### 9.2.5. North Sea and Northern Europe

After many successful marine seismic-reflection and -refraction investigations of the North Sea and the adjacent Skagerrak in the 1980s (see Fig. 9.2.5-01), as, e.g., BIRPS, SALT, MOBIL SEARCH, and others (e.g., Klemperer and Hobbs, 1991; Barton and Wood, 1984; Husebye et al., 1988), two major seismic projects were added in the 1990s: in 1992, a BIRPS experiment and in 1993, the MONA LISA project (Marine and Onshore North Sea Acquisition for Lithospheric Seismic Analysis).

The BIRPS (British Institutions Reflection Profiling Syndicate) project of 1992 was a two-ship coincident near-vertical and wide-angle experiment to determine the continental structure of the central North Sea (Neves and Singh, 1996; Singh et al., 1998a). Its location was within the "BIRPS profiles" (Fig. 9.2.5-01) parallel to and halfway between the SALT line and the next WSW-ENE–oriented line to the south shown in Figure 9.2.5-01. It was a 106-km-long expanding spread profile. It was obtained by the steaming of two vessels at a constant speed toward a central midpoint, with one of them shooting every 75 m with simultaneous recording on a 114-channel 2.7-km dual streamer, and the other one receiving with a 240-channel 6-km streamer. In addition, seven ocean-bottom hydrophones (OBHs) were deployed at 20-km intervals along the profile. In a second stage, the two vessels steamed in the same direction at constant offsets corresponding to critical distances for the top of the lower crust and the crust-mantle transition. The principal results along the

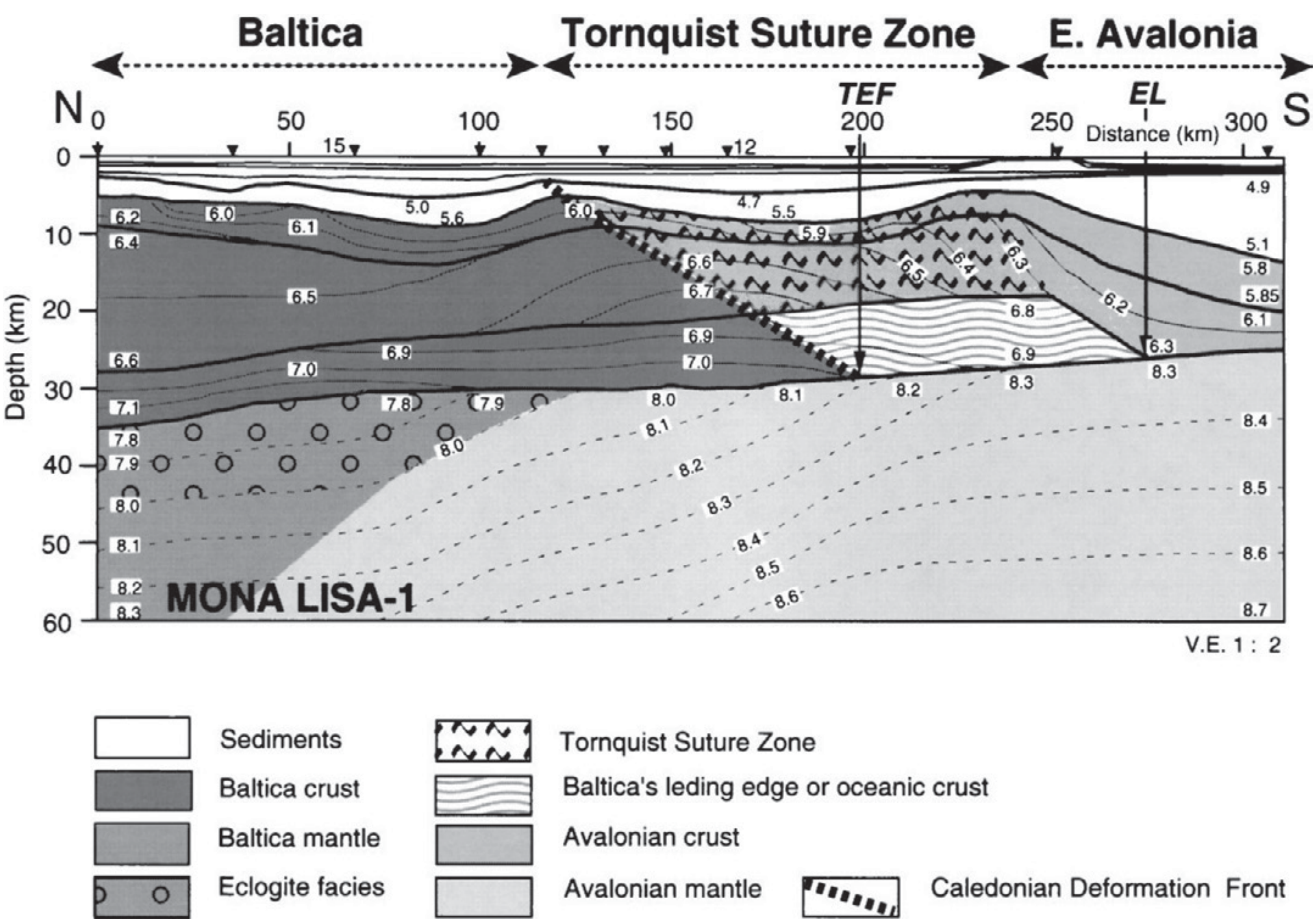



Figure 9.2.5-02. P-wave velocity model of MONA LISA profile 1 (from Abramovitz et al., 1999, fig. 2). Small triangles indicate the position of OBHs. TEF—Trans-European Fault, EL—Elbe Lineament. [Tectonophysics, v. 314, p. 69–82. Copyright Elsevier.]

135-km-long profile were an upper crustal thickness of 19 km and a Moho depth of 32–34 km (Singh et al., 1998a).

The main aim of the MONA LISA project (Fig. 9.2.5-01) was to image plate collision structures and reflectivity characteristics of the crust and uppermost mantle in order to map the Caledonian Deformation Front and possible remains of the Tornquist and Iapetus lithospheric plates (MONA LISA Working Group, 1997). Another aim was to identify rifting processes by profiles crossing the Central Graben and the Horn Graben. Here, attempts were made to avoid anticipated problems due to strong attenuation of seismic signals in the graben sediments by locating the seismic lines primarily across structural highs

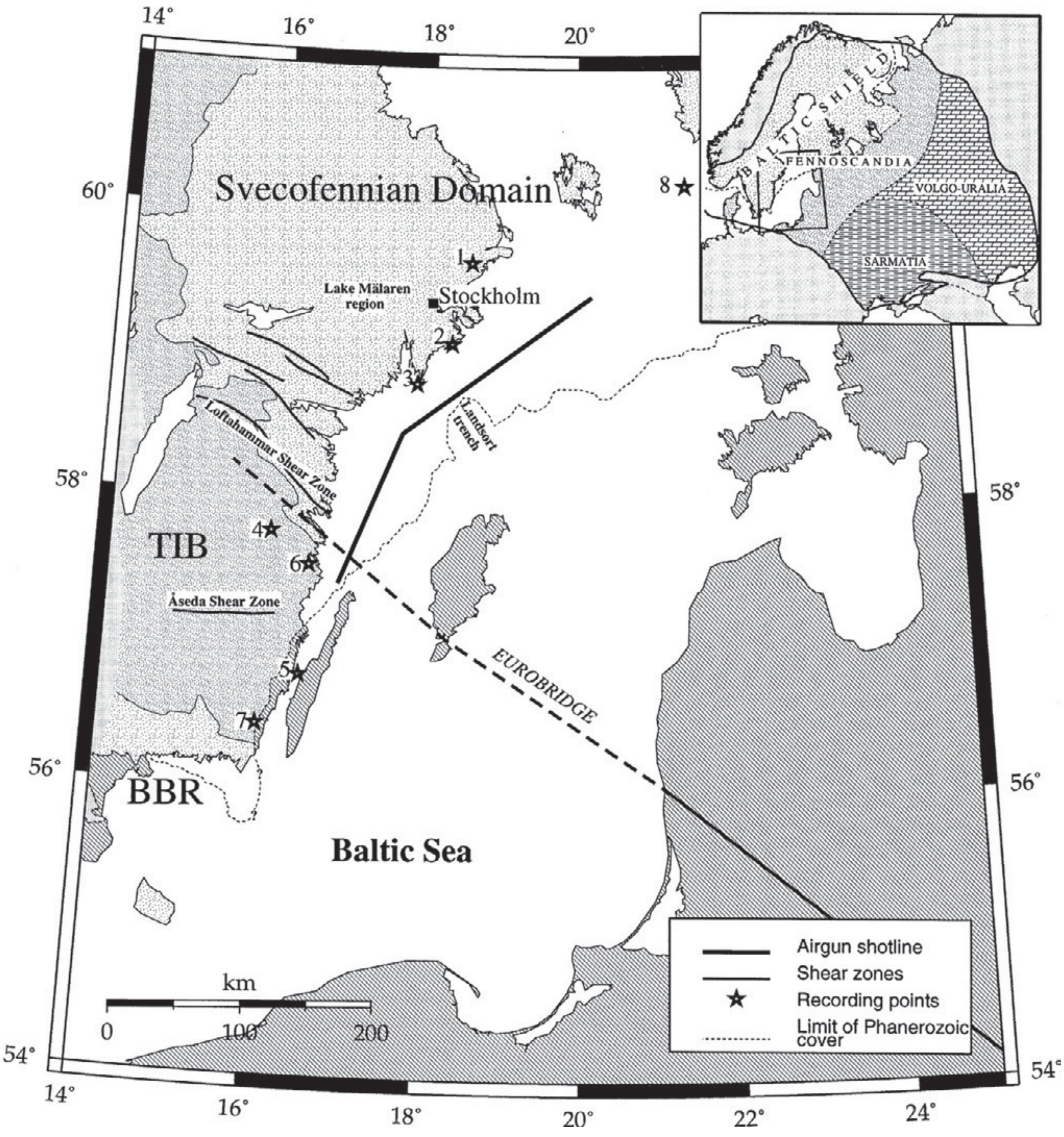



Figure 9.2.5-03. Location map of the Coast Profile transect along the southeastern coast of Sweden (from Lund et al., 2001, fig. 1). Dashed line—EUROBRIDGE line. TIB—Transscandinavian Igneous Belt; BBR—Blekinge-Bornholm region. [Tectonophysics, v. 339, p. 93–111. Copyright Elsevier.]

and by the use of a very powerful airgun array tuned to low-frequencies.

The fieldwork was achieved in 1993 by the cooperation of scientific and commercial institutions of Denmark, Germany, Great Britain, and Sweden. The four profiles comprised a total length of 1112 km of normal incidence seismic data and the recording of the airgun signals at 26 onshore and 2 offshore locations, allowing for successful recording distances of up to 400 km. Shots at 75 m intervals were recorded by a 180-channel streamer of 4500 m length, towed at 15 m depth which provided 30 fold data with 26 s record length and 12.5 m CMP spacing. In 1995, additional wide-angle profiling with OBH stations was carried out.

The normal-incidence reflection profiles showed an unusually low intracrustal reflectivity, in particular a poorly reflective lower crust. The Moho could be identified on all four profiles, but also, for the first time, dipping and subhorizontal reflections from the mantle up to 24 s TWT could be correlated in the normal-incidence reflection sections. For the Central Graben a reduction in crustal thickness by ~5 km was observed (MONA LISA Working Group, 1997). The following interpretation (Abramovitz et al., 1998) revealed a three-layered Baltica crust in the north thinning from 34–35 km to 29–30 km at the Tornquist Suture Zone and changing into a 24–25-km-thick two-layered East Avalonia crust (see, e.g., Fig. 9.2.5-02) in the south.

In Scandinavia, after the thorough investigation of the Scandinavian crust by a large series of seismic-refraction campaigns in the 1980s, which concentrated particularly on the crust underneath the Baltic Shield in Sweden, Finland, and under the Baltic Sea, only one crustal survey was added in the 1990s. This was a marine study in 1995 along the southeastern coast of Sweden named Coast Profile (Lund et al., 2001). The shooting vessel, employed from the Polar Geosurvey Expedition in St. Petersburg, Russia, fired airgun shots every two minutes for 38 h along the shooting line, resulting in distances of 270 m between consecutive shotpoints.

Eight recording stations had been deployed on land along the Swedish coast, of which six provided high-quality data ranging from 25 km to 327 km airgun-receiver offsets (Fig. 9.2.5-03). The data were interpreted together with data of the FENNOLORA project of 1979 resulting in a crustal model along the southeastern Swedish coast where Moho depths varied from 36 to 52 km (Fig. 9.2.5-04). The results were later discussed in the context of EUROBRIDGE which will be discussed in subchapter 9.3, "Deep Seismic Sounding in Eastern Europe and Adjacent Asia (Former USSR)."

Otherwise, seismic projects in Scandinavia in the 1990s were devoted primarily to upper mantle studies. Already in the 1980s, some of the long-range seismic-refraction profiles had enabled the detection of details of sub-Moho upper-mantle structure (see, e.g., Guggisberg and Berthelsen, 1987).

In Russia, the reinterpretation of super-long profiles of several 1000 km length, based on the use of nuclear explosions in the late 1970s, had revealed details of the upper-mantle structure to several 100 km depth. In Scandinavia, such strong energy sources were not available, therefore in the 1990s, several major projects were organized, using the methodology of teleseismic tomography. According to the scope of this publication, however, these projects will not be discussed in much detail.

Examples are the Teleseismic Tomography Tornquist (TOR) project, investigating the deep lithospheric structure of the Trans-European Suture Zone between Phanerozoic and Proterozoic Europe by a teleseismic array of 120 stations, recording for half a year in 1996 and 1997 and simulating a 100-km-wide and 900-km-long profile from northern Germany through Denmark to southern Sweden and reaching a depth range of 300 km (Gregersen et al., 2002). Some details of crustal S-wave structure were obtained by applying the time-domain inversion method to the receiver functions to compute S-wave velocities in the crust and uppermost mantle beneath each station down to 60 km depth (Wilde-Piorko et al., 2002).

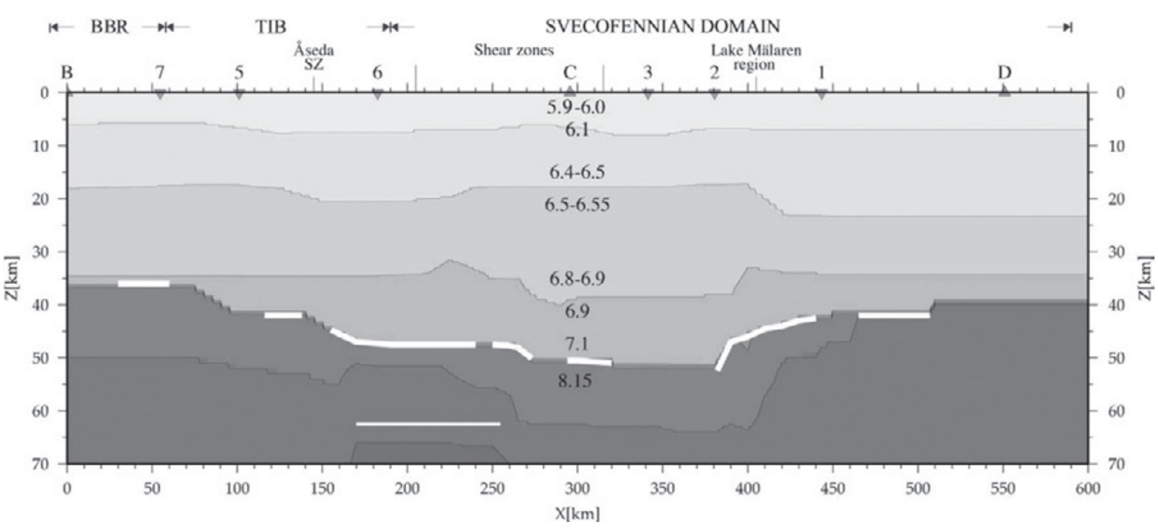



Figure 9.2.5-04. P-wave velocity model of the Coast profile 1 (from Lund et al., 2001, fig. 7). Numbers—Coast profile recording sites; letters—FENNOLORA shotpoints. [Tectonophysics, v. 339, p. 93–111. Copyright Elsevier.]

The second large project was the Svecofennian-Karelian-Lapland-Kola Transect (SVEKALAPKO) project, which included near-vertical and wide-angle seismic-reflection and refraction work, large-scale magnetotelluric studies, geothermal surveys, and an intensive study of teleseismic data with a network of 144 temporary and permanent seismic stations (Bock et al., 2001). The temporary seismic station array was operated between August 1998 and May 1999 covering the Proterozoic and Archean crust in Finland and Russian Karelia.

## 9.3. DEEP SEISMIC SOUNDING IN EASTERN EUROPE AND ADJACENT ASIA (FORMER USSR)

Also located on the Baltic Shield, in 1992, a seismic-reflection line was recorded across the Kola Superdeep Borehole (Fig. 9.3-01), which had been cored to a depth of 12.25 km and had attracted worldwide attention, because, due its large depth, hypotheses on the nature of crustal reflectivity could be tested against in situ geological control (Kozlovsky, 1988).

Single-fold reflection shooting and DSS had been conducted in the SG-3 region before, but it had lacked deep-crustal, common-depth-point (CDP) seismic profiling (Ganchin et al., 1998). Near-vertical reflection profiling, based on the "common midpoint" seismic method (CMP), began in the Kola Peninsula in the beginning of the 1990s. The Russian state geophysical company EGGI recorded two profiles in the northwestern part with a record length of 15s (Kostyuchenko et al., 2006).

In 1992, an international group acquired a 25s TWT line across the Kola super-deep borehole (Ganchin et al., 1998). The experiment of 1992 comprised the 38-km-long CDP profile across, and two deep (6 km) and four shallow (0.5 km) VSP (vertical seismic profiles) within the borehole. The CDP profiling was arranged along a SW-NE–directed line, starting ~38 km south of the borehole and ending just north of the borehole. It provided a 20-fold coverage and was recorded to 15 s TWT (Smythe et al., 1994; Ganchin et al., 1998). It resulted in an image of a highly reflective section: Dipping reflections could be correlated with lithological contrasts within the supracrustal series and at shear

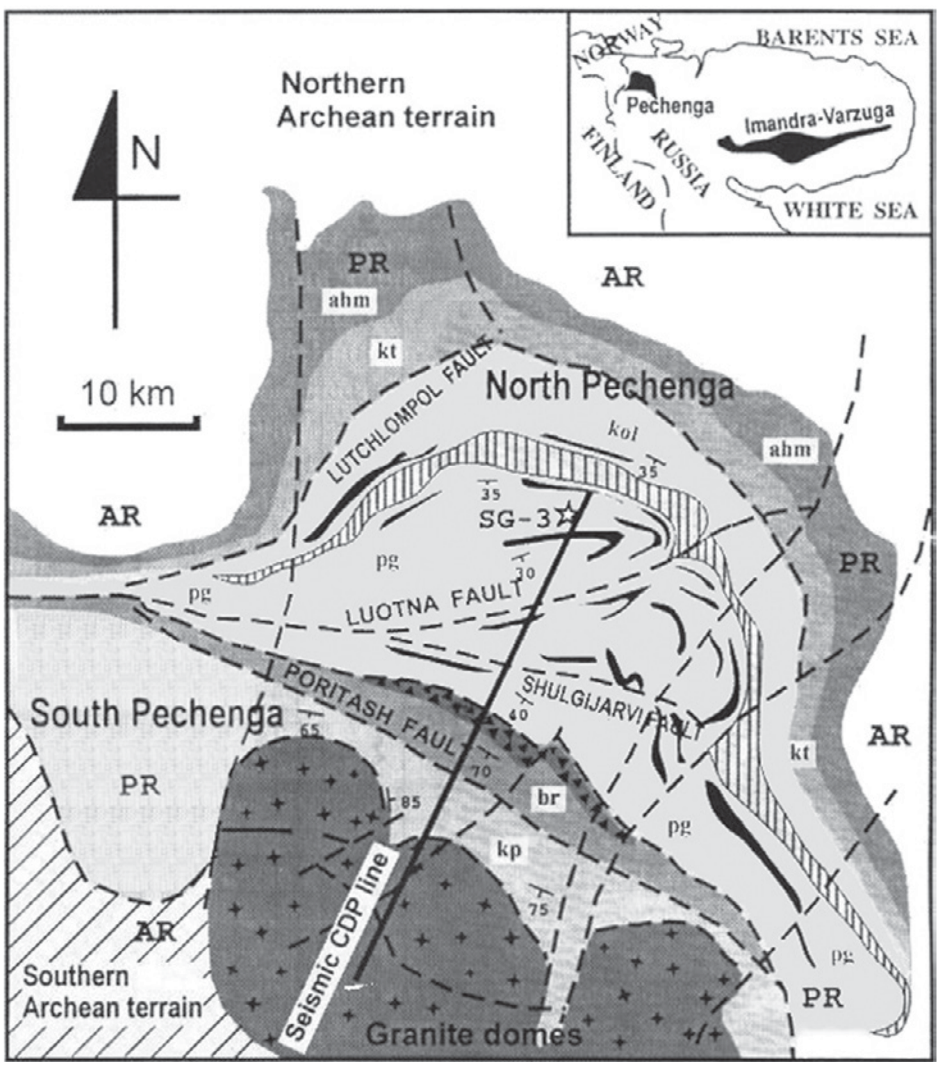



Figure 9.3-01. Geological sketch map of the Pechenga region on the Kola Peninsula, showing the location of the Kola-92 seismic-reflection line (seismic common depth point line) and the Kola Superdeep Borehole SG-3 (from Ganchin et al., 1998, fig. 1). [Tectonophysics, v. 288, p. 1–16. Copyright Elsevier.]

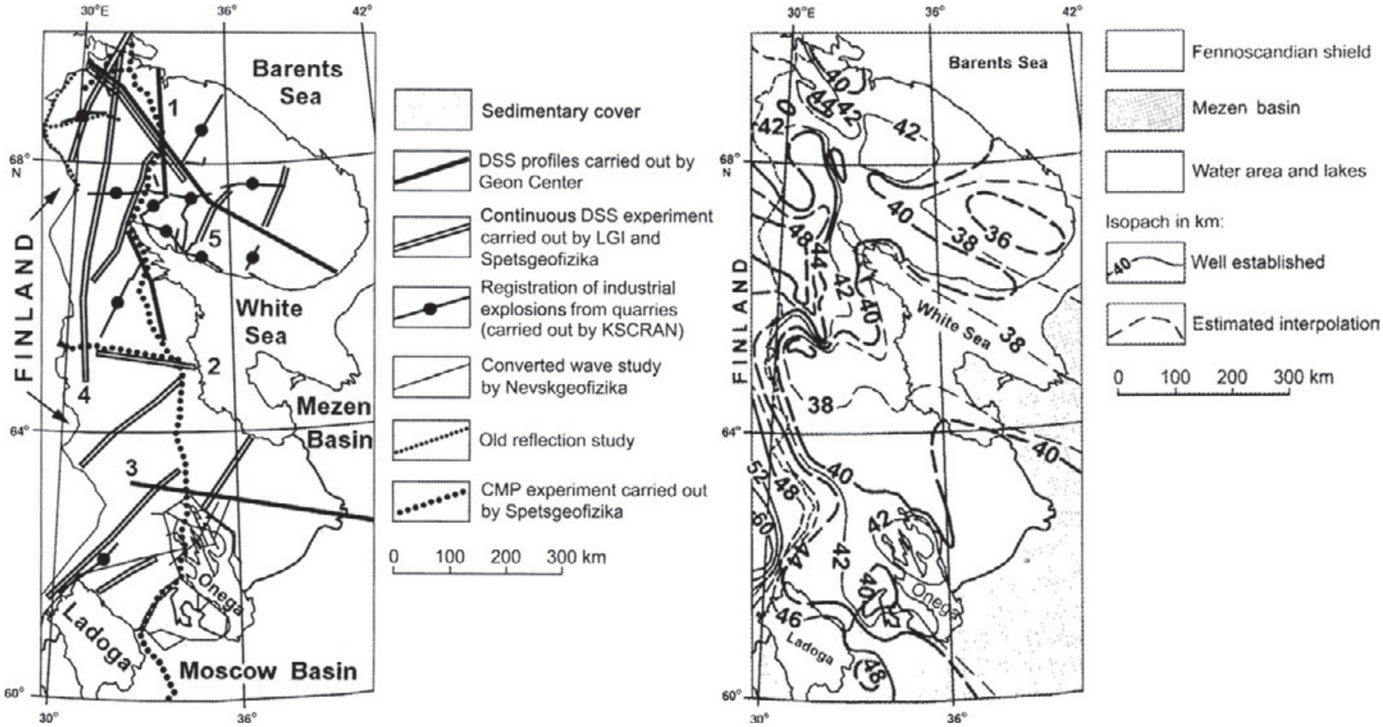



Figure 9.3-02. Left: Location of regional deep seismic sounding (DSS) profiles across the eastern part of the Fennoscandian Shield (from Kostyuchenko et al., 2006, fig. 4). Numbered single lines—DSS profiles performed by GEON; numbered double lines—continuous DSS profiles performed by Spetsgeofizika. Right: Depth to Moho under the northeastern part of the Fennoscandian Shield (from Kostyuchenko et al., 2006, fig. 10). [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 521–539. Reproduced by permission of Geological Society Publishing House, London, U.K.]

zones. The results suggested the presence of fluids down to a depth of at least 12 km in the upper crust.

From 1997 to 2000, Spetsgeofisika completed two profiles (dotted lines in Fig. 9.3-02), including a long CMP profile from the Kola borehole in the north to Petrozavodsk in the south and a second profile in W-E direction from the Finnish border to the White Sea (Berzin et al., 2002).

From 2000 to 2002, Spetsgeofisika conducted a Vibroseis CMP experiment across the western and central Mezen Basin and along a line across the Timan Range (dotted lines 3 and 4 in Fig. 9.3-03). The total length of available CMP profiles (both state and industry) amounted to ~2000 km. Figures 9.3-02 and 9.3-03 show only the state-financed CMP lines. The maximum source-receiver distance was 10 km. The energy was provided by 4 or 5 10-t vibrators spread over a base of 50 m. Groups of 12

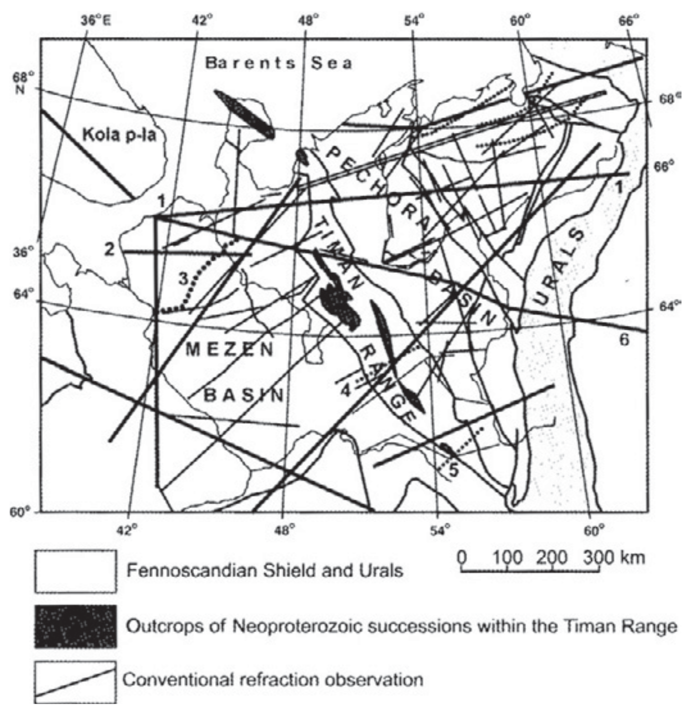



Figure 9.3-03. Location of deep seismic sounding (DSS) profiles across the Mezen Basin, Timan Range and Pechora Basin (from Kostyuchenko et al., 2006, fig. 5). Thick lines—CMP profiles; thin lines—older conventional DSS lines. [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 521–539. Reproduced by permission of Geological Society Publishing House, London, U.K.]

geophones were spaced every 44 m. The record length was 25 s TWT and a 50-fold stack was obtained. Figure 9.3-03 also shows older conventional DSS lines which had been recorded by the former Ministry of Geology.

During these studies, 48–60 channel recording instruments had been used, usually spaced every 100–200 m. Seismograms were recorded on paper by the wiggle-trace method. Shotpoint spacing was from 10–20 km to 50–70 km and shooting was direct and reversed. Refraction observations were shot with maximum offset up to 100–150 km (Kostyuchenko et al., 1999, 2006). The interpretation of these data resulted in several cross sections (Kostyuchenko et al., 2006, figs. 7–9) and two Moho maps (Figs. 9.3-02 and 9.3-04), showing on average an overall 38–42-km-thick crust. Only underneath the Mezen and Pechora basins, crustal thinning to less than 36 km was observed. The Moho contour maps also show a sudden crustal thickening along the Finnish border as well as to the east of the Pechora basin (Figs. 9.3-02 and 9.3-04).

A link between the super-deep SG-3 well on the Kola peninsula to the base Hayes-1 well on Franz-Joseph land was established in 1995, when geophysical investigations in the southern part of the AP-1 geotraverse across the Barents Sea (Fig. 9.3-05) were completed (Sakoulina et al., 2000). The integrated seismic experiment included offshore and onshore wide-angle reflection/refraction observations along a 700-km-long profile (Fig. 9.3-05). The offshore recording devices comprised 41 three-component OBSs with 5–20 km spacing and a multi-channel reflection study at different offset ranges. Onshore, eight seismometers had been installed. The seismic energy was provided by powerful 80-l array airgun shots with a 250 m interval. They produced clear arrivals to 150–180 km distances, and in some cases even up to 280 km. Intensive reflections from the Moho at 35–40 km depth were almost continuously recorded (Fig. 9.3-06). The seismic investigations continued into 2001 (Roslov et al., 2009). The AP-2 profile with a total length of 935 km crossed the center of the Barents Sea, Novaya Zemlya Island and the Kara Sea up to the Yamai Peninsula of western Siberia.

To investigate the deep lithospheric structure of the East European Craton between the exposed Proterozoic and Archean complexes of the Baltic and Ukrainian Shields, in 1994 the project EUROBRIDGE was established within the frame of EUROPROBE and aimed for an integrated geoscience research of a ~2000-km-long swath of the lithosphere, extending from Scandinavia to the Black Sea (Bogdanova et al., 2001, 2006). The main part of the EUROBRIDGE project was a major seismic-refraction and wide-angle reflection experiment. It was conducted in several legs from 1994 to 1997, commencing in the north and terminating in the south, in the Ukraine (Fig. 9.3-07).

The first leg was a marine project across the Baltic Sea (EUROBRIDGE'95 Seismic Working Group, 2001) from Västervik (Sweden) to Shventoji (Lithuania), recorded in 1994. For its northwestern end at the Swedish coast, the Coast Profile (Lund et al., 2001), discussed further above, added important information (Fig. 9.3-07; see also Figs. 9.2.5-03 and 9.2.5-04).

It was followed by three seismic-refraction surveys on land, carried out in 1995, 1996, and 1997, establishing a

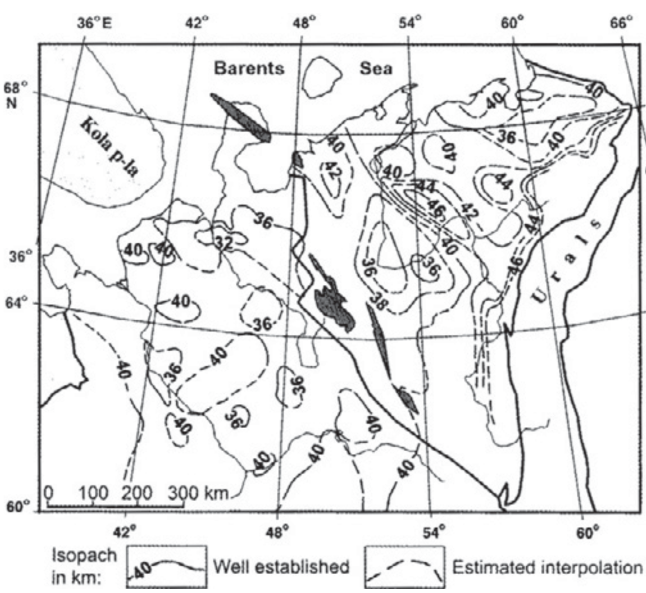



Figure 9.3-04. Depth to Moho under the Mezen Basin, Timan Range and Pechora Basin (from Kostyuchenko et al., 2006, fig. 10). [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 521–539. Reproduced by permission of Geological Society Publishing House, London, U.K.]

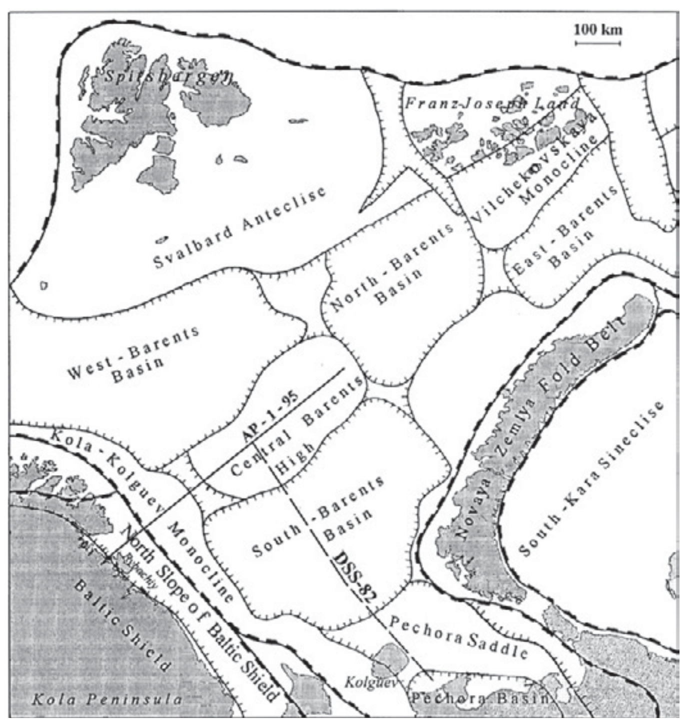



Figure 9.3-05. Tectonic sketch map of the Barents Sea showing the location of the AP-1-95 geotraverse (from Sakoulina et al., 2000, fig. 1). [Tectonophysics, v. 329, p. 319–331. Copyright Elsevier.]

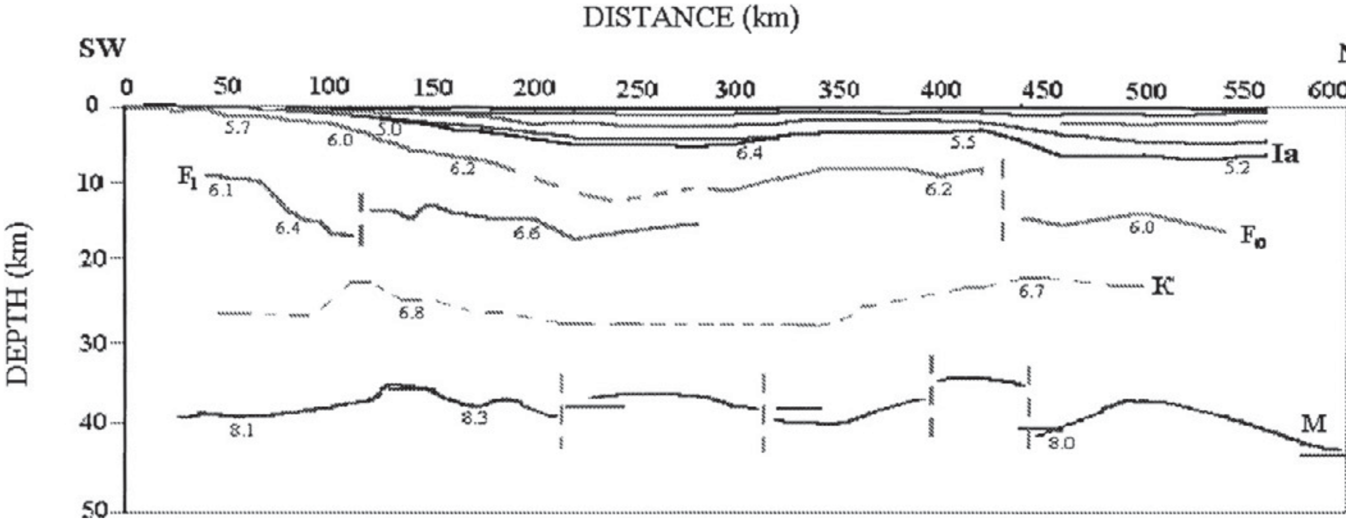



Figure 9.3-06. Crustal cross section through the Barents Sea along the AP-1-95 geotraverse (from Sakoulina et al., 2000, fig. 6). [Tectonophysics, v. 329, p. 319–331. Copyright Elsevier.]

line running from NW to SE, which extended from Sweden across the Baltic Sea and farther on land from the coast of Lithuania through Belarus into the Ukraine (Figs. 9.3-07 and 9.3-08). The 1995–1996 line crossed the northeastern end of POLONAISE profile P4 and ran parallel to P5. Interpretations of profiles P4 and P5 are included in the summary of Bogdanova et al. (2006, Fig. 8).

The first onshore stage, EUROBRIDGE'95, was 280 km long and was recorded on the Lithuanian part of the East European Platform, traversing the Proterozoic West Lithuanian Granulite Domain and East Lithuanian Belt terranes (Fig. 9.3-08). Explosive shots of up to 1000 kg TNT were detonated at 10 shotpoints (SP01-SP10) at intervals of ~30 km, and an offshore shot (SP00) was fired in the Baltic Sea close to Gotland. 76 three-component stations allowed an average station spacing of 3.5 km (EUROBRIDGE'95 Seismic Working Group, 2001).

The second onshore stage, EUROBRIDGE'96, was 544 km long and covered the area (Fig. 9.3-08) from the East Lithuanian Belt to the Ukrainian Shield. The southernmost part was recorded in the Pripyat Trough (see Fig. 9.3-10). During EUROBRIDGE'96, 14 shots were fired at 30 km intervals from new shotpoints 11–24, and two shots from shotpoints of 1995, nos. 08 and 10. The shots were recorded in two deployments by 114 three-component stations, allowing a station spacing between 3 and 4 km (EUROBRIDGE Seismic Working Group, 1999). A record section for shotpoint 24 is shown in Figure 9.3-09, demonstrating the energy distribution for crustal phases up to 300 km recording distance (more data are shown in Appendix A9-1-6).

The third onshore stage, EUROBRIDGE'97 (Fig. 9.3-10) covered Ukrainian territory and was recorded in 1997 (Thybo et al., 2003). The line was 530 km long and crossed the EUROBRIDGE'96 profile in southern Belarus. Data acquisition was obtained in two deployments with overlapping coverage in the center of the line. Seismic energy came from 18 borehole shots with charges ranging from 250 to 1000 kg, which were recorded by 120 mobile three-component seismographs with a typical station spacing of 3–4 km (for data, see Appendix A9-1-6).

The interpretation of EUROBRIDGE'95 and '96 (Fig. 9.3-11) revealed that the crust consists generally of three layers, but the middle crustal layer is much thinner in the northwest. The Moho in the northwest of Lithuania (West Lithuanian Domain in Fig. 9.3-11) and the adjacent Baltic Sea is ~44 km deep, but its depth increases toward southeast under the East Lithuanian Belt to 50 km and the crust remains 50 km thick below southern Lithuania and Belarus, with Moho elevations of a few kilometers in between. Uppermost-mantle velocities immediately beneath the Moho are generally 8.2–8.4 km/s. A lower lithosphere reflector was found at 65–70 km depth. High lower-crustal velocities and a crustal thickness of ~50 km were observed throughout the EUROBRIDGE' 96 profile. The boundary between the East Lithuanian Belt and West Lithuanian Granulite Domain was associated with pronounced crustal velocity changes, and a thinning of the crust toward the northwest (EUROBRIDGE Seismic Working Group, 1999; EUROBRIDGE'95 Seismic Working Group, 2001).

The interpretation of the EUROBRIDGE'97 data (Fig. 9.3-12) revealed for most of the line (Thybo et al., 2003) a similar velocity-structure model as shown above for the southernmost part of EUROBRIDGE'96, with the exception that the upper part of the lower crust had velocities less than 7.0 km/s and was separated from the lowermost crust by a distinct velocity jump near 35–40 km depth. In the model of Thybo et al. (2003) the lowermost crust, defined by velocities greater than 7 km/s, gradually dissolves toward the south underneath the Ukrainian Shield.

While EUROBRIDGE covered the westernmost part of the Dniepr-Donets rift system, DOBRE (Donbas Refraction and Reflection), a multinational study group, concentrated on the Donbas Foldbelt of the Ukraine (Fig. 9.3-13), the uplifted and

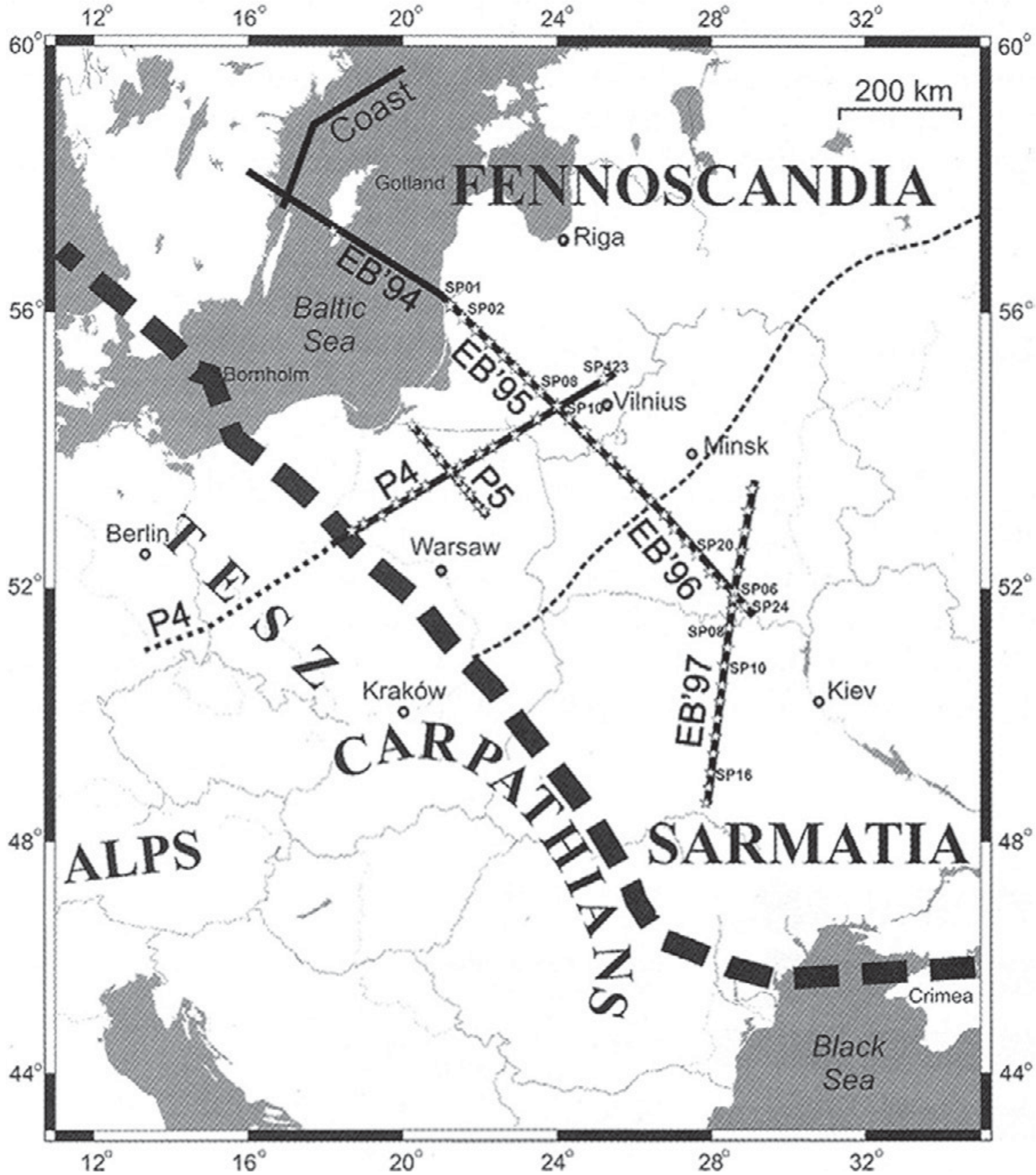



Figure 9.3-07. Location of deep seismic sounding (DSS) profiles in the area of the southwestern margin of the East European Craton (from Bogdanova et al., 2006, fig. 5). EB'94 to EB'97—EUROBRIDGE 1994–1997; Coast—Coast Profile 1995; P4, P5—Polonaise 1997; TESZ—Trans-European Suture Zone. [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 599–625. Reproduced by permission of Geological Society Publishing House, London, U.K.]

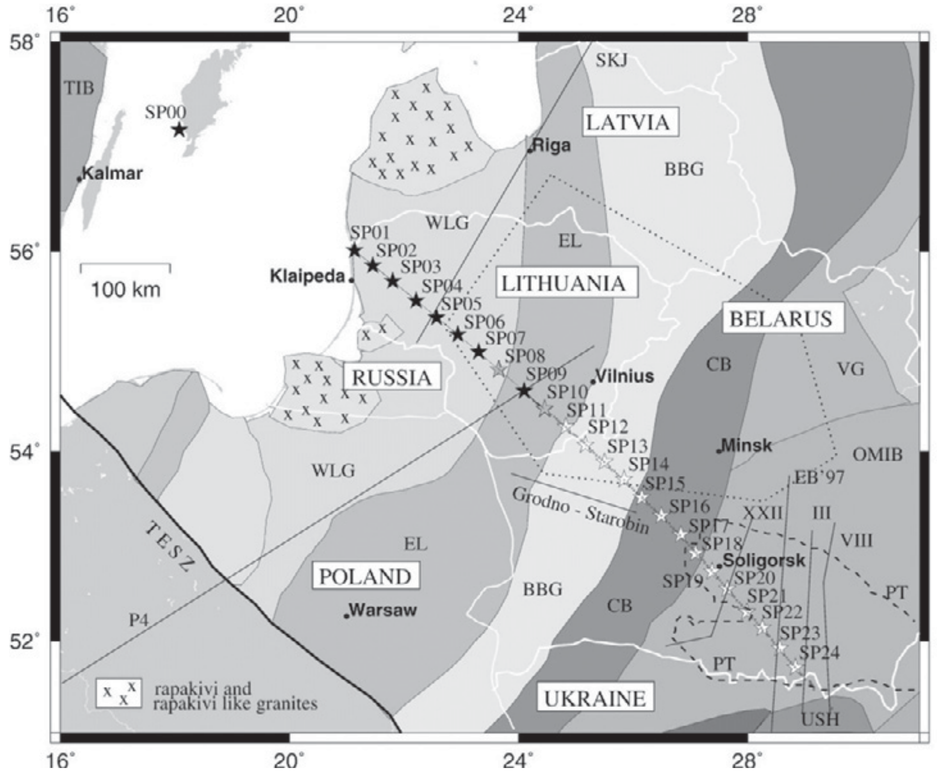



Figure 9.3-08. Location map of EUROBRIDGE '95 and '96 seismic-refraction campaigns (from EUROBRIDGE Working Group, 1999, fig. 2), showing major regional tectonic units. Black and grey stars—shotpoints of 1995; white and grey stars—shotpoints of 1996; thin black lines—other seismic refraction lines; Roman numbers—former lines of 1960s and 1970s; EB'97—EUROBRIDGE 1997; P4—Polonaise line 4; SKJ—seismic line Sovietsk-Kohtla-Jarve; TESZ—Trans-European Suture Zone; dashed area, PT—Pripyat Trough; USH—Ukrainian Shield; other signatures—regional tectonic units. [Tectonophysics, v. 314, p. 193–217. Copyright Elsevier.]

deformed part of the more than 20-km-thick Dniepr-Donets Basin that formed due to Late Devonian rifting of the East European Craton in the eastern Ukraine and in southern Russia (DOBREflection-2000 and DOBREfraction '99 Working Groups, 2002; DOBREfraction '99 Working Group, 2003; Stephenson et al., 2006).

In 1999 a seismic-refraction/wide-angle reflection survey was performed across the Donbas Foldbelt. A total of 261 one-component stations was supplied by the University of Texas at El Paso and the University of Copenhagen, and 20 three-component stations by the Polish Academy of Sciences. The project consisted of a main line with 245 recording stations and a 190-km-long subsidiary line with 36 stations (line DOBRE'99 in Fig. 9.3-13), and 11 shots were placed in line every 25–30 km along the main line. In 2000, a seismic-reflection survey, using Vibroseis and explosive sources, along a 133-km-long part of the 1999 refraction line was added between shotpoints 3 and 7, reaching from the Azov Massif of the Ukrainian Shield northward across the axial part of the Donbas Foldbelt (DOBREflection-2000 and DOBREfraction '99 Working Groups, 2002; DOBREfraction '99 Working Group, 2003; Appendix A9-1-7; Stephenson et al., 2006).

The resulting DOBREfraction '99 P-wave model (Fig. 9.3-14) showed a remarkable basin underneath the 110-km-wide Donbas Foldbelt with up to 20-km-thick sediments. The Moho showed some topography, but was found on average at ~40-km depth. Along the reflection profile this generally corresponded to a broad reflective band which could be traced along the whole reflection profile at ~14 s below the Ukrainian Shield and at 12 s

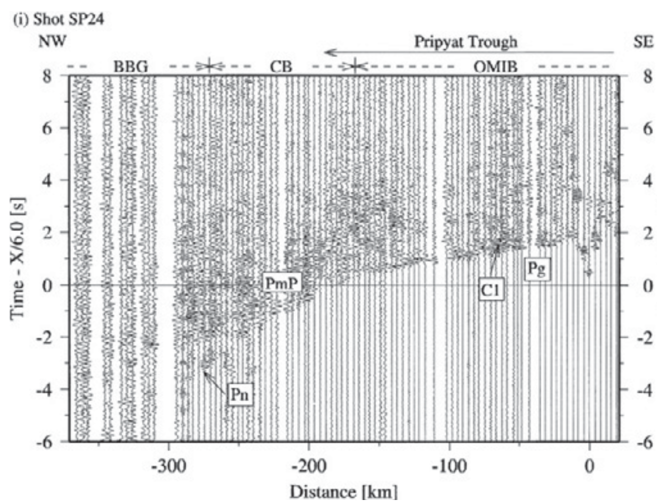



Figure 9.3-09. Record section of EUROBRIDGE '96 shotpoint 24 (from EUROBRIDGE Seismic Working Group, 1999, fig. 4i), showing Pg and Pn (refracted first arrivals from crystalline basement and Moho) and reflections: C1 from a mid-crust layer, PmP from Moho. For location, see Fig. 9.3.08). [Tectonophysics, v. 314, p. 193–217. Copyright Elsevier.]

below the Donbas Foldbelt (DOBREflection-2000 and DOBREfraction '99 Working Groups, 2002; Stephenson et al., 2006).

For the Peri-Caspian basin a N-S oriented crustal cross section along longitude 50.7°E (Figs. 9.3-15 and 9.3-16) was compiled from unpublished data (Brunet et al., 1999; Stephenson

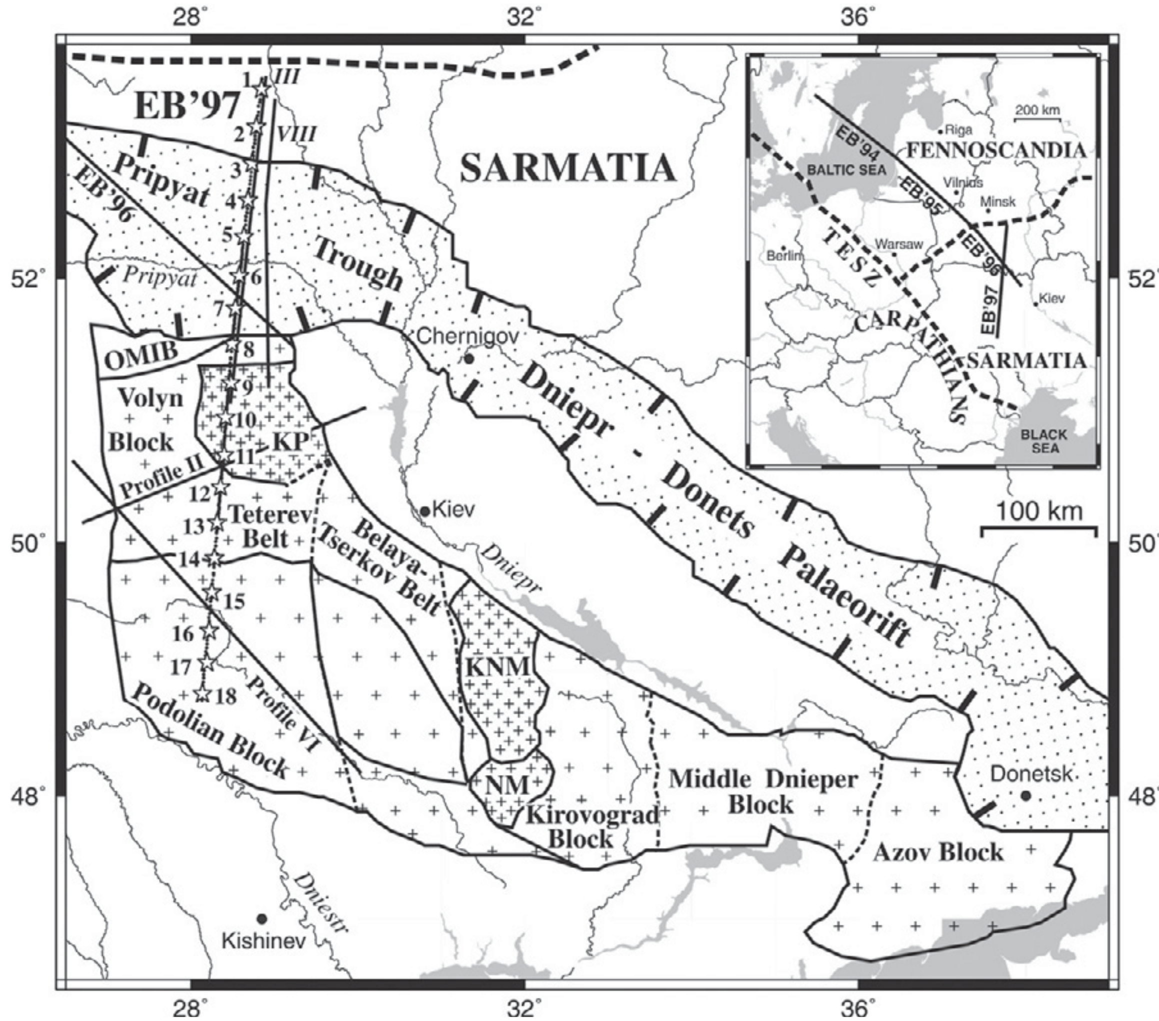



Figure 9.3-10. Location of EUROBRIDGE '97 seismic-refraction line (from Thybo et al., 2003, fig. 1). Nos. 1 to 18: shotpoint locations. Profiles III, VI, VIII: earlier deep seismic sounding profiles, but numbered differently by Sollogub et al. (1969, see Fig. 6.3.1-01 and 1972, see Fig. 6.4-04). [Tectonophysics, v. 371, p. 41–79. Copyright Elsevier.]

et al., 2006), showing a 35–40-km-thick crust, underlain at its center, where the sediments reach a thickness of 15–20 km, by a high-velocity layer of 8.1 km/s which was discussed with some controversy as to whether it is part of the crust or not.

When EUROPROBE started in 1990, a close cooperation between Russian and German scientists was immediately agreed upon. As a result, when in 1991 an ~3000-km-long DSS profile, named GRANIT (Fig. 9.3-17), was recorded by Russian scientists from eastern Ukraine to the West Siberian basin (GRANIT transect, 1992; Juhlin et al., 1996), a sub-experiment, aiming for a detailed anisotropy investigation was arranged. It was one of several sub-experiments of the project GRANIT and was a large-scale seismic anisotropy experiment, carried out in 1991 across the Voronesh Massif, named ASTRA (Fig. 9.3-17, left). It consisted of 280-km-long seismic-refraction profiles, which crossed each other in a central point. The ASTRA experiment was characterized by observations with dense multi-azimuthal more-component recordings in the near-vertical incidence to wide-angle distance range. This recording scheme (Fig. 9.3-17, right) allowed for investigation of a 5 million km$^2$ large block for signs of rock anisotropy, including shear-wave birefringence and azimuthal dependency (Lueschen, 1992).

The size of the borehole shots ranged between 200 and 300 kg. For recording the following devices were available: five 48-channel "Progress" systems and 10 three-component analog stations plus two digital PDAS from GFZ Potsdam. The operation shown in Fig. 9.3-17 was planned to be repeated for each line. However, for logistic reasons, only half of the experiment could finally be realized. Unfortunately, results of this project were never published in the international literature. The northeastern part of the profile GRANIT is shown in Fig. 9.3-18.

Being part of the international program EUROPROBE, and coincident with the 1991 GRANIT line across the Middle Urals,

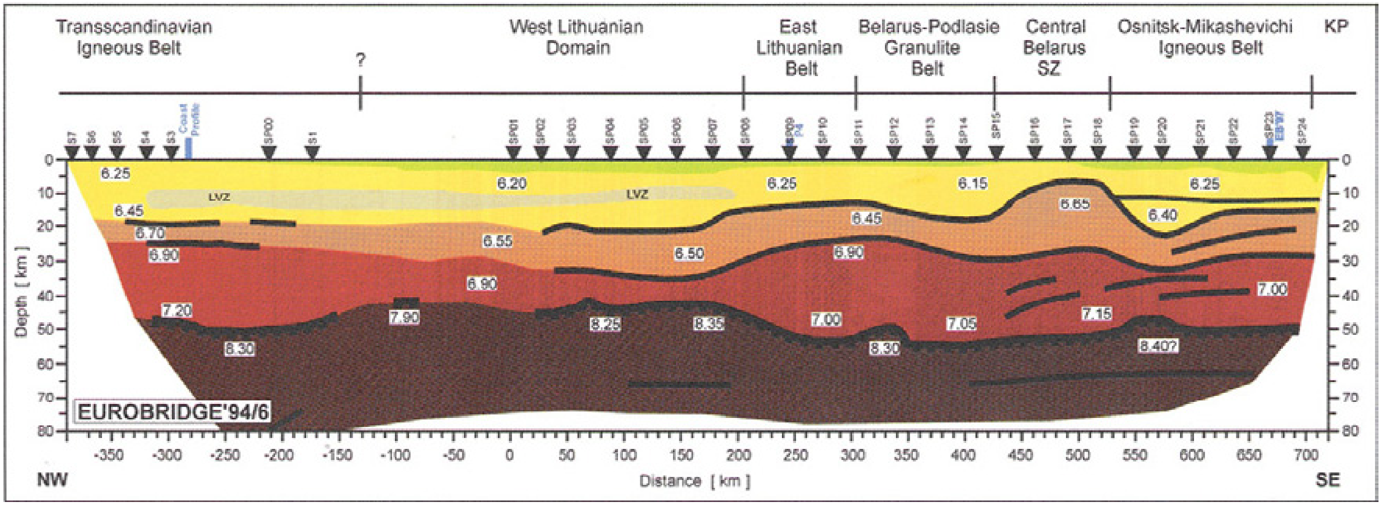



Figure 9.3-11. P-wave crustal model from Lithuania to Belarus from EUROBRIDGE '94 to '96 seismic-refraction campaigns (from Bogdanova et al., 2006, fig. 8). Triangles—shotpoint locations; KP—Korosten Pluton. [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 599–625. Reproduced by permission of Geological Society Publishing House, London, U.K.]

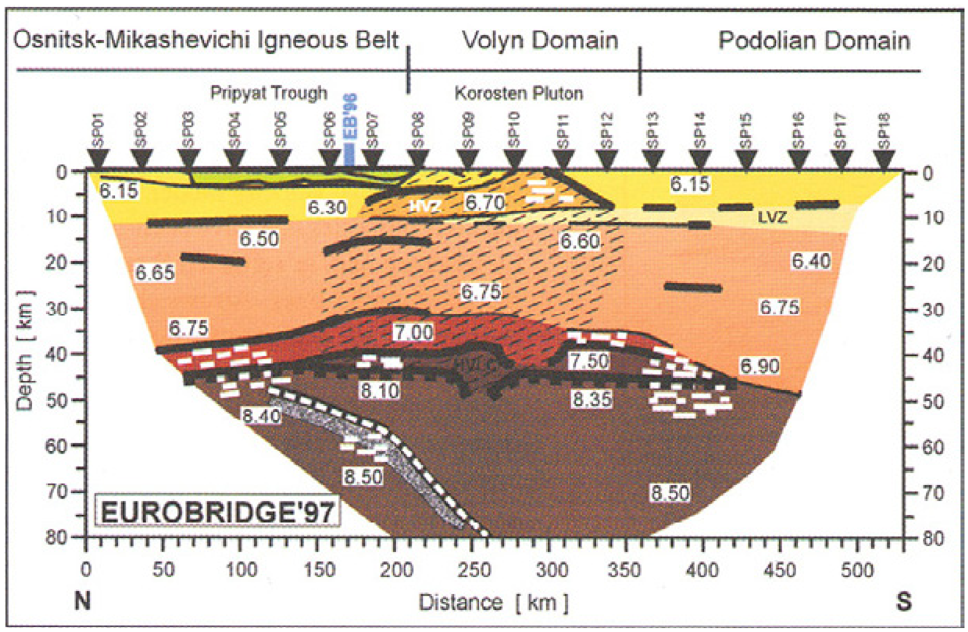



Figure 9.3-12. P-wave crustal model from southern Belarus to northwestern Ukraine from the EUROBRIDGE '97 seismic-refraction campaign (from Bogdanova et al., 2006, fig. 8). Triangles—shotpoint locations; dipping black lines—elements of seismic boundaries obtained from reflected and refracted waves; body with thin dashed lines—high-velocity zone; zones with white bars—zones of high reflectivity. [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 599–625. Reproduced by permission of Geological Society Publishing House, London, U.K.]

in 1993 the ESRU (EUROPROBE Seismic Reflection profiling in the Urals) project started in a joint effort of Russian, Swedish, U.S., and German scientists. Until 1999, it was conducted in five seismic-reflection experiments in the Middle Urals (Figs. 9.3-18 and 9.3-19), approximately along the latitude 58°N, and was in total ~335 km long (Juhlin et al., 1995, 1996; Kashubin et al., 2006). The ESRU line crossed the orogen close to the Urals Deep Borehole SG4.

ESRU93 was oriented in a SW-NE direction and was 60 km long. ESRU95 consisted of two profiles, running in W-E and in N-S directions, which crossed each other at the deep borehole SG-4. ESRU96 was an extension of the ESRU95-WE line. ESRU98 overlapped the eastern end of ESRU96 before running toward the NE over Mesozoic strata of the West Siberian Basin. A 2 km gap remained between ESRU98 and ESRU 99, which continued the whole ESRU line to the east. The line was extended to the west in 2001–2003 reaching a total length of 440 km (see Chapter 10.2.3).

In addition to the ESRU-SB and Michailovsky profiles, data of the R17 profile (Juhlin et al., 1996) in the northern part of the area and the Alapaevsk and Chernoistoichinsk profiles (Steer et al., 1995), just south of the ESRU profile, were relevant for the large-scale structural interpretation of the Middle Urals. The 70-km-long seismic-reflection profile R-17 in the northern Middle

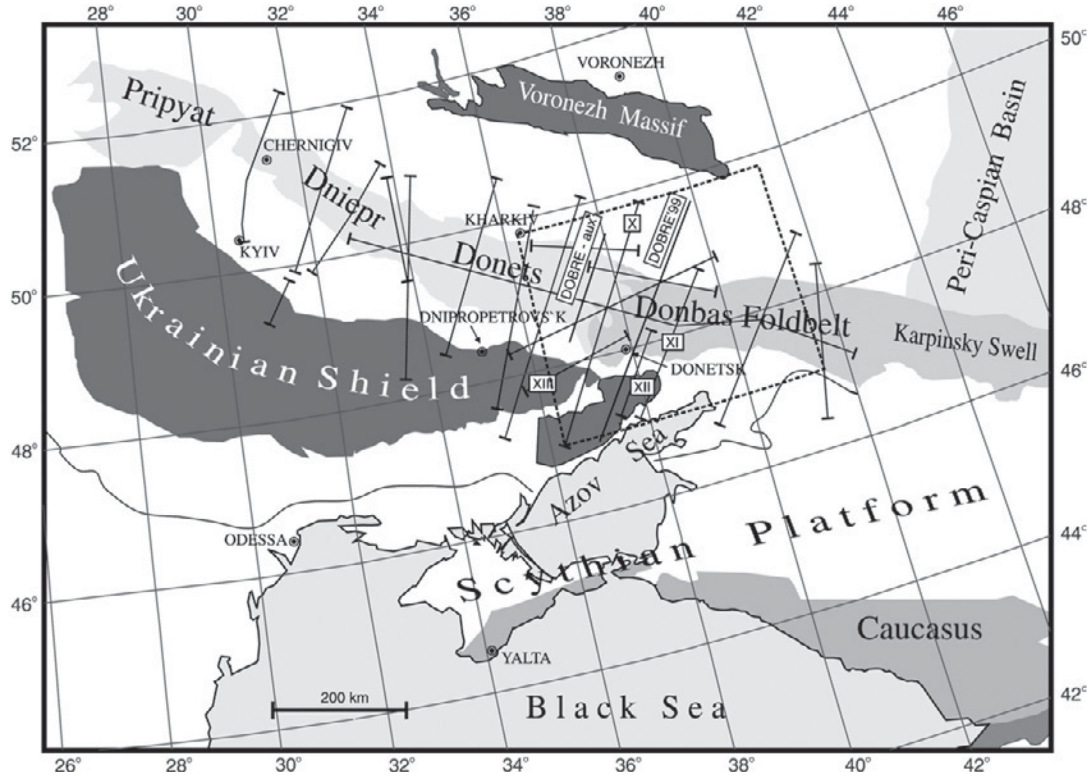



Figure 9.3-13. Regional tectonic setting of the Dniepr-Donets Basin and its contiguous inverted segment Donbass Foldbelt, showing location of DOBRE '99 (from DOBREfraction '99 Working Group 2003, fig. 1) and earlier seismic lines (see also Figure 6.4-04, corresponding profiles X, XI, XII re-numbered 10, 11, 12 by Sollogub et al., 1972). [Tectonophysics, v. 371, p. 81–110. Copyright Elsevier.]

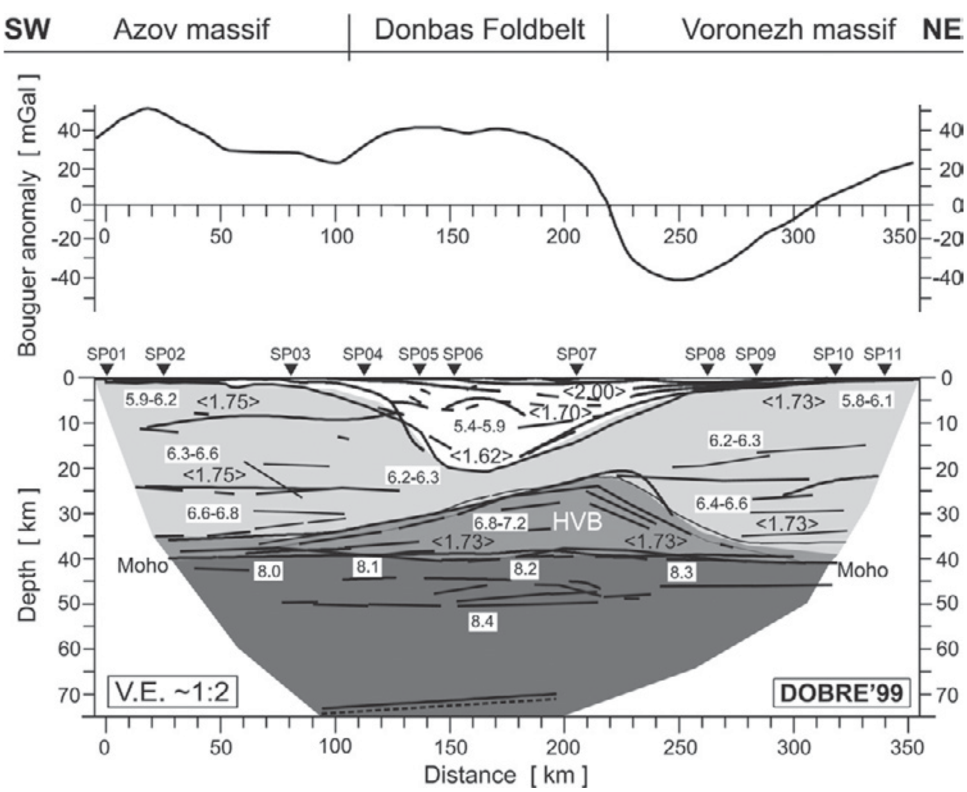



Figure 9.3-14. Top: Bouguer gravity anomaly along DOBREfraction '99 line. Bottom: DOBREfraction '99 P-wave model (from DOBREfraction '99 Working Group, 2003, fig. 11). Also shown are $V_p/V_s$ ratios (in brackets). HVB—high velocity body. [Tectonophysics, v. 371, p. 81–110. Copyright Elsevier.]

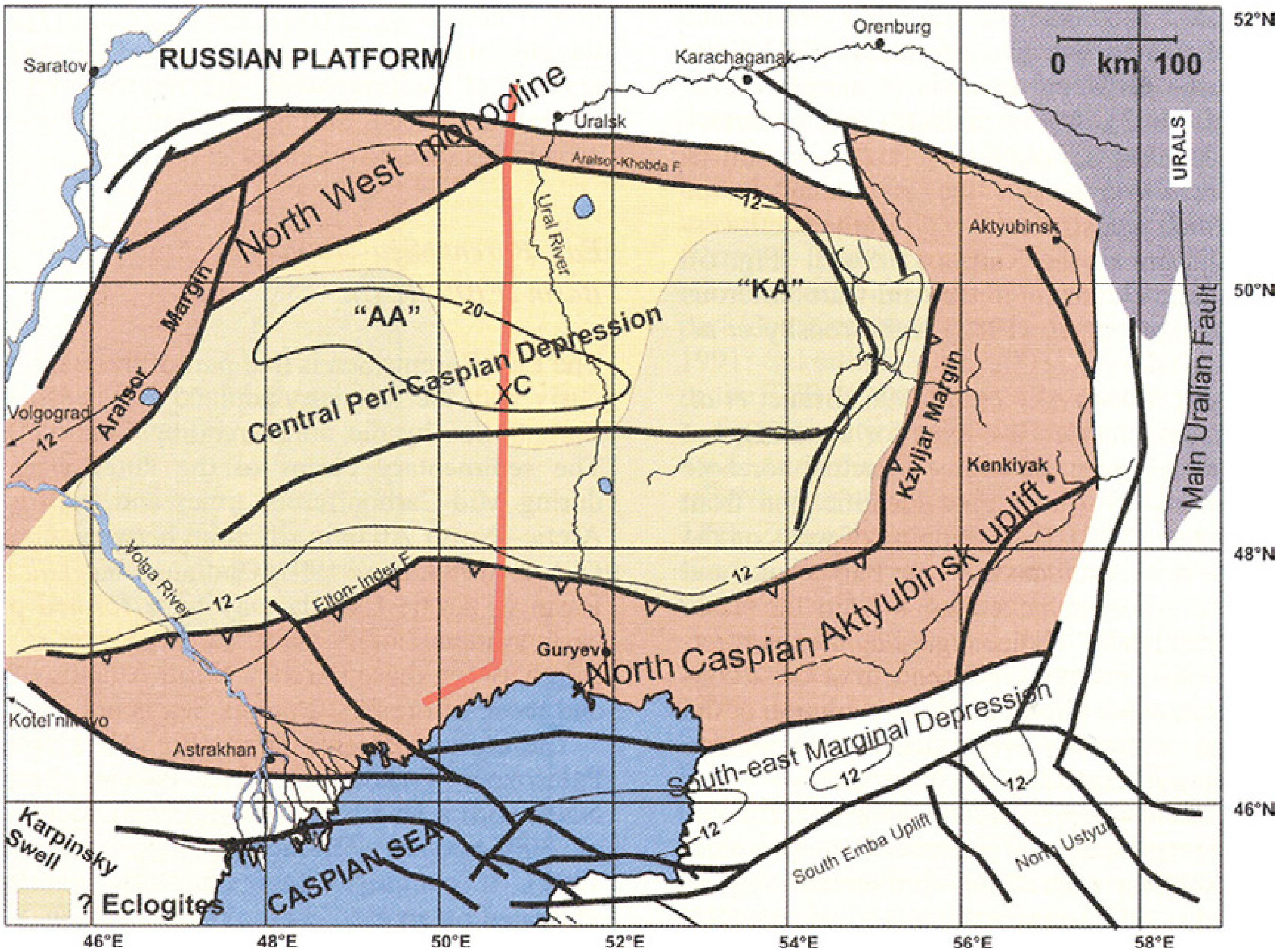



Figure 9.3-15. Crystalline basement contour map of the Pre-Caspian Basin showing the location of the cross section in Figure 9.3-16 (from Stephenson et al., 2006, fig. 6, *after* Brunet et al., 1999, fig. 2). [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 599–625. Reproduced by permission of Geological Society Publishing House, London, U.K.]

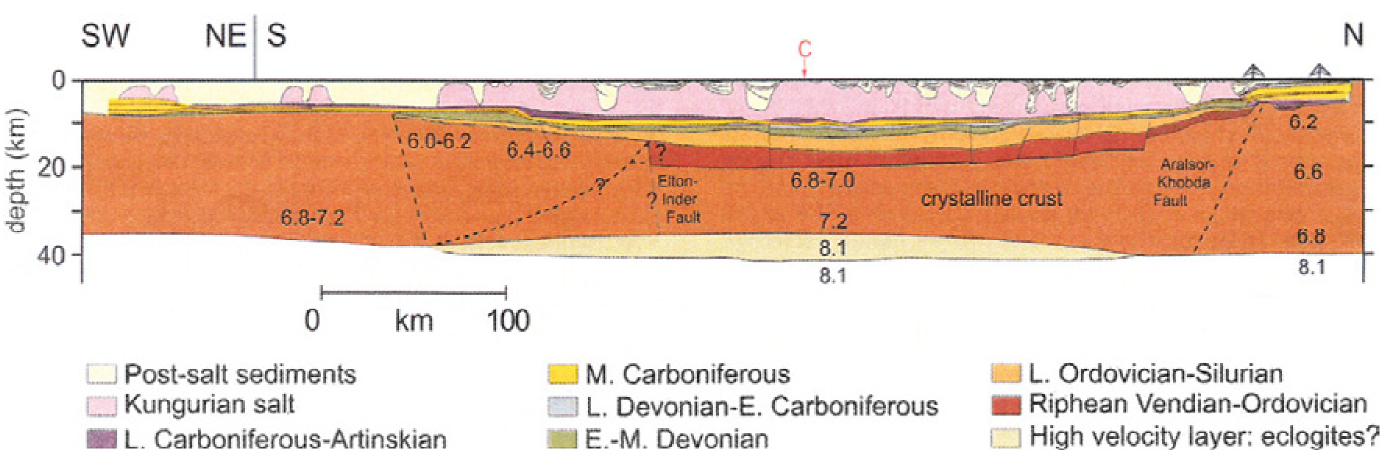



Figure 9.3-16. Simplified sketch along a N-S transect through the Peri-Caspian Basin (from Stephenson et al., 2006, fig. 7, *after* Brunet et al., 1999, fig. 5). [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 599–625. Reproduced by permission of Geological Society Publishing House, London, U.K.]

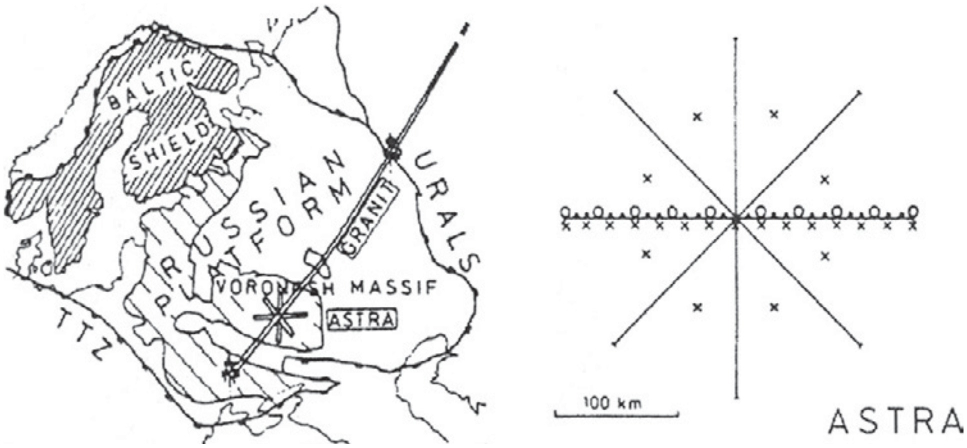



Figure 9.3-17. Left: Location of the ASTRA sub experiment (from Lueschen, 1992). Right: Observation scheme. Crosses are borehole shots with 200–300 kg charges, 20 km apart. Circles are 48-channel digital recording instruments of type <Progress>, which were spaced 30 km apart. Dots are 3-component analogue recording devices, which were 10 km apart. [DGG-Mitteilungen (News of the German Geophysical Society), v. 3, p. 24–39. Reprinted with permission of German Geophysical Society (DGG).]

Urals (Fig. 9.3-18) was recorded also in 1993 in conjunction with a standard oil prospecting survey in the West Siberian basin.

Every eighth shot (~400 m) along the profile was recorded up to 20 s TWT, which was also available for the reinterpretation of Juhlin et al. (1996), who interpreted the Moho to be at the base of a zone of continuous reflections in the lower crust between 11 and 14 s TWT. Finally, the data of the Michailovsky profile, recorded in the southern Middle Urals, were available for the interpretation (line M in Fig. 9.3-19).

Juhlin et al. (1995, 1996) have described and interpreted the first phase of ESRU93. The reinterpretation of the corresponding wide-angle part of the GRANIT line together with the ESRU data by Juhlin et al. (1996) showed comparatively high uppermost-mantle velocities of 8.4–8.5 km/s and a lower-crustal root with its deepest point along the ESRU profile and directly below the surface expression of the Main Uralian fault (Fig. 9.3-20).

Similar Moho depths were reported for the recent wide-angle UWARS experiment, which line is located 100 km farther south (Fig. 9.3-18), but the thickest crust there was interpreted to be farther to the east (Thouvenot, et al., 1994). Together with older DSS and peaceful nuclear explosions data, Juhlin et al. (1996) saw convincing evidence for the presence of an anomalously thick root under the Urals from ~45–55 km depth with low reflectivity below the Urals.

A presentation of the complete ESRU data set and its interpretation (Fig. 9.3-21) is described by Kashubin et al. (2006). A reflective zone of ~12–15 km thickness, deepening to the east, had been interpreted already by Juhlin et al. (1995) as East European Craton crystalline lower crust. This eastward thickening was also recognized farther south on the Michailovsky profile (line M in Fig. 9.3-19). Below the strong mid-crustal reflectivity, west-dipping reflections were seen, beginning at 60 km depth (km –30) and extending to ~25 km depth (km +100), indicating the deepest part of the Urals crustal root under the Kvarkush Anticline. Toward the east (east of km +50), underneath the hinterland east of the Main Uralian Thrust Zone in the Middle Urals, which according to Kashubin et al. (2006) appears to be dominated by Paleozoic island-arc terranes, the reflectivity of the lower crust is variable, but the reflective

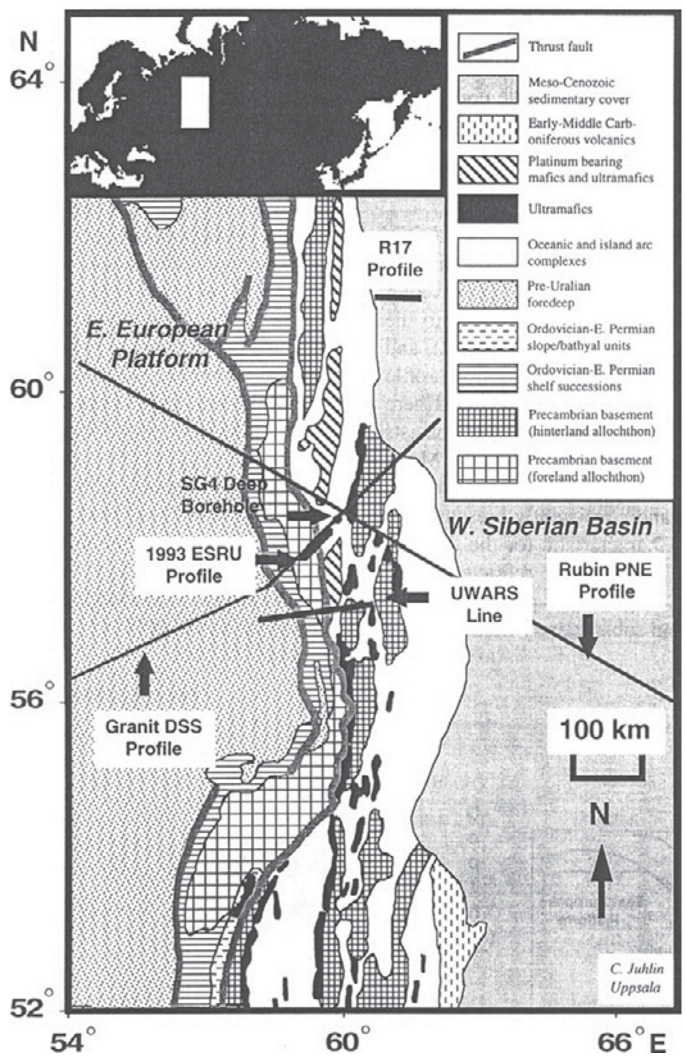



Figure 9.3-18. Seismic profiles in the Middle Urals (from Juhlin et al., 1996, fig.1). The RUBIN PNE profile was recorded in the 1970s. [Tectonophysics, v. 264, p. 21–34. Copyright Elsevier.]

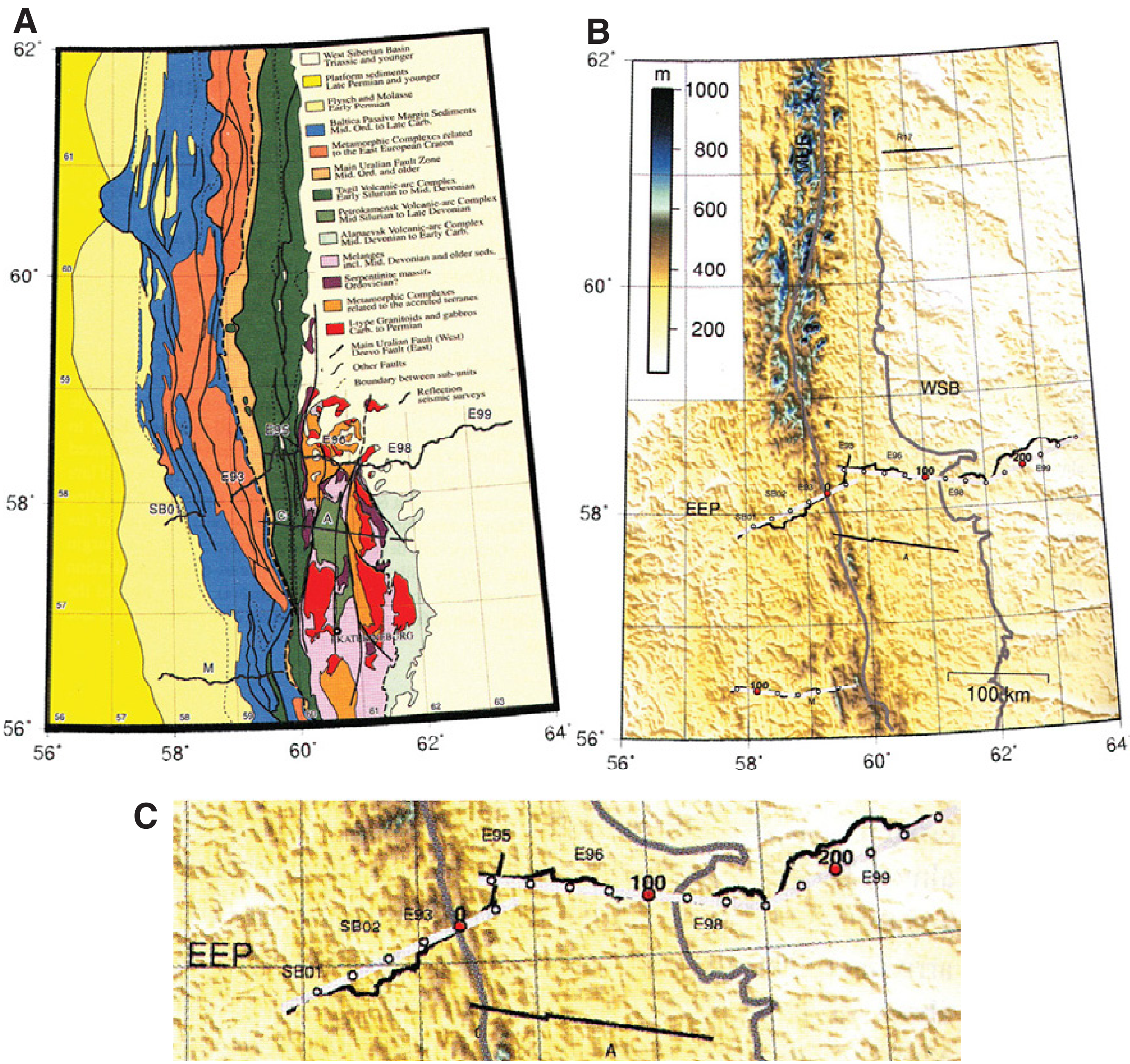



Figure 9.3-19. Location maps of ESRU (Europrobe Seismic Reflection profiling in the Urals) in the Middle Urals (from Kashubin et al., 2006, fig.1). (A) Geological map. (B) Physiographic map of the Middle Urals (EEP—East European Platform; WSB—West Siberian Basin). E93 to E99—ESRU; SB—Serebrianka-Beriozovka profile; M—Michailovsky profile. (C) Enlarged section of (B) showing location of ESRU. [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 427–442. Reproduced by permission of Geological Society Publishing House, London, U.K.]

Moho is still reasonably well defined, is generally flat and continuous at 40–45 km depth (Kashubin et al., 2006, 2009).

The largest international seismic project, carried out in Russia after 1990, was the URSEIS'95 project (Berzin et al., 1996, Carbonell et al., 1996, Knapp et al., 1998), which crossed the Urals at ~53°N, ~400 km farther south than the lines discussed above. URSEIS'95 was an integrated seismic experiment designed by scientists from Russia, Germany, the United States, and Spain to reveal the detailed crustal and upper-mantle structure of the Urals. The transect, which was chosen for this purpose, extended from the East European platform in the Uralian foreland to the West Siberian basin (Fig. 9.3-22). The seismic survey comprised three parts: (*i*) a 465-km-long near-vertical incidence, Vibroseis-source reflection survey, (*ii*) a coincident

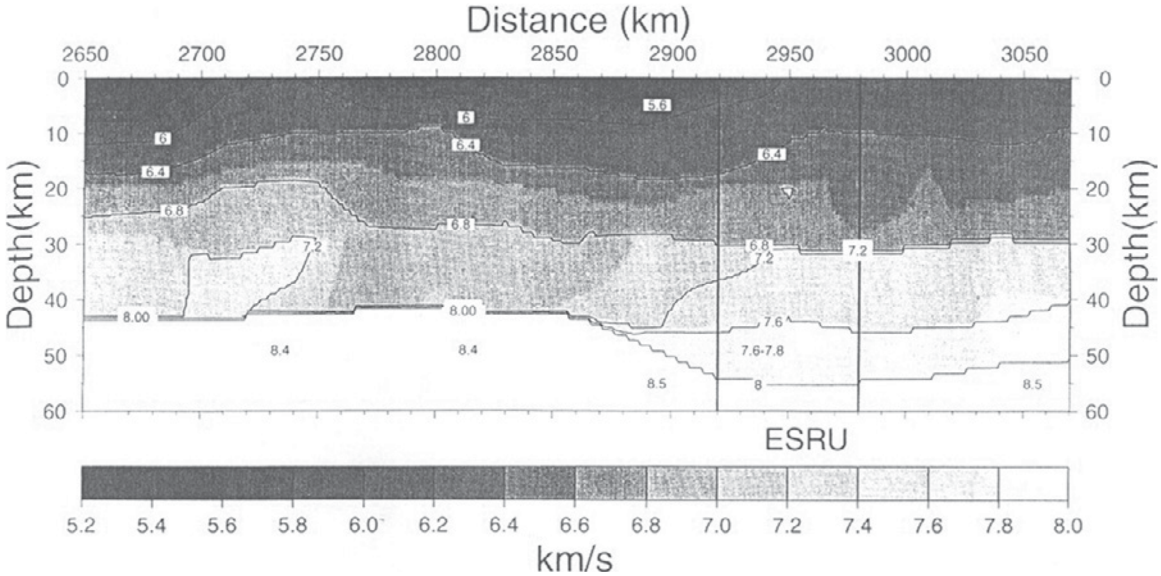



Figure 9.3-20. Velocity model based on a 400 km segment of the GRANIT deep seismic sounding profile (from Juhlin et al., 1996, fig.4). [Tectonophysics, v. 264, p. 21–34. Copyright Elsevier.]

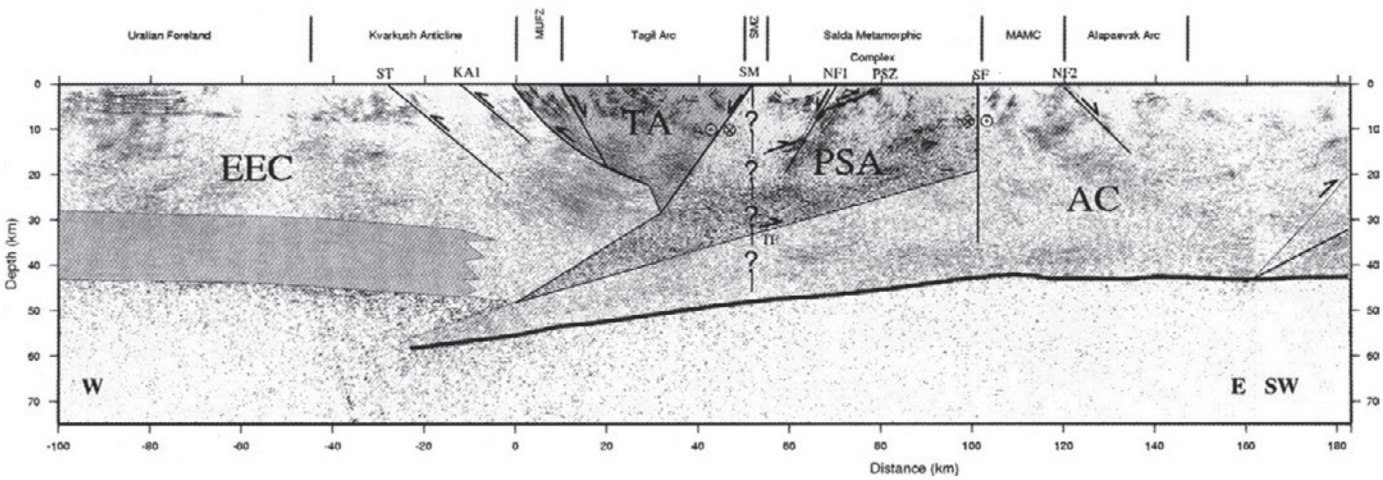



Figure 9.3-21. Merged migrated line drawing with interpretation along the entire ESRU profile from km –100 to km +200 (from Kashubin et al., 2006, fig.5). MTFZ—Main Uralian Thrust Fault (km 0); SMZ—Serov-Mauk Zone; MAMC—Murzinka-Adui Metamorphic Zone; EEC—East European Craton; TA—Tagil Arc; PSA—Petrokamensk-Salda crust; AC—Alapaevsk Crust. [Gee, D.G., and Stephenson, R.A., eds., European lithosphere dynamics: Geological Society of London Memoir 32, p. 427–442. Reproduced by permission of Geological Society Publishing House, London, U.K.]

near-vertical incidence, explosive-source survey, and (*iii*) a 340-km-long explosive-source, wide-angle reflection and refraction survey, including a cross line and two off-line fan-recording shots (Fig. 9.3-22).

From six shotpoints (Fig. 9.3-22), a total of 33,000 kg of explosives distributed among 15 shots, with charge sizes that ranged between 1500 and 3000 kg was used for the seismic-refraction/wide-angle reflection survey. The seismic energy was recorded by 50 three-component digital recording instruments (RefTek and Lennartz) in successive deployments. The difficult logistics forced station spacings that ranged from 1 to 2.5 km (Carbonell et al., 1996, 1998b, 2000b; Stadtlander et al., 1999).

The operation required six deployments. During the first deployment, a N-S line along the 60°E meridian was completed. A shot was fired at shotpoint S6 and recorded southwards from the cross-point with the E-W main line for ~120 km. With the second and third deployments, fan profiles were recorded from shotpoints S5 and S6, located 60 km north of the main line, and recorded along an E-W line, arranged 60 km south of the main line (Fig. 9.3-22).

The fans were designed so that the reflection points would lie below the E-W main line. The major effort of the wide-angle experiment involved a 335-km-long E-W main profile (Fig. 9.3-22). This profile was coincident with the eastern 335 km of

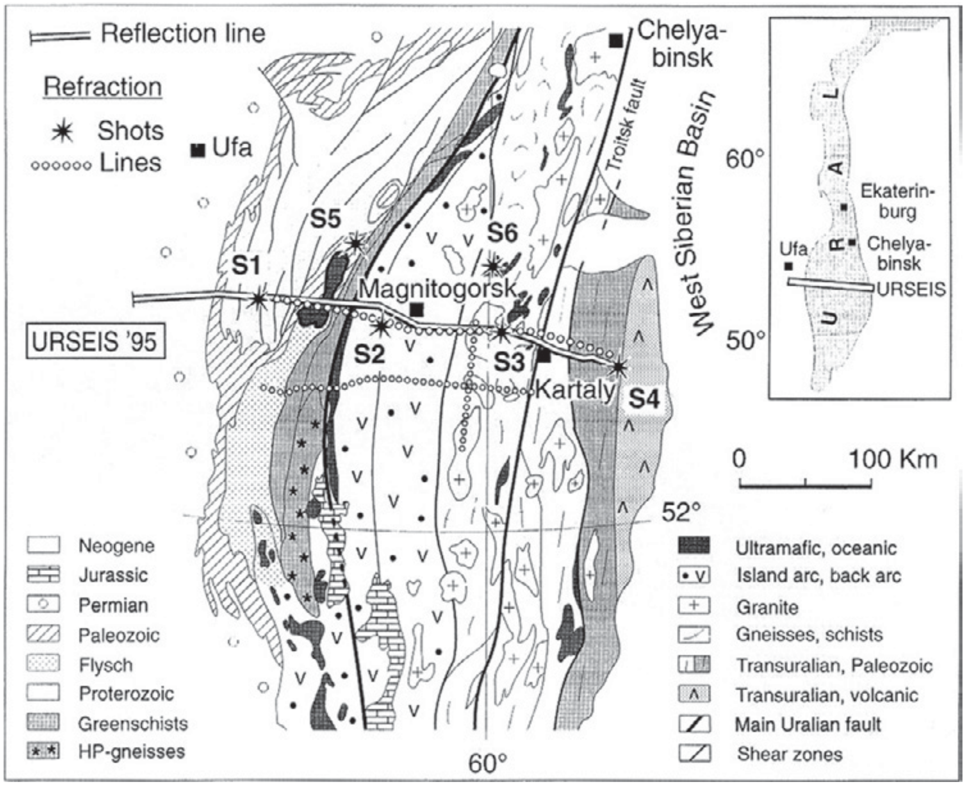



Figure 9.3-22. Generalized tectonic map of the southern Urals, showing the location of the seismic-reflection and refraction surveys (from Berzin et al., 1996, fig. 1). [Science, v. 274, p. 220–221. Reprinted with permission from AAAS.]

the 465-km-long URSEIS'95 near-vertical incidence reflection profile. The profile was completed in deployments four to six, and an average instrument spacing of 2.3 km was realized along the whole line for shotpoints S2, S3, and S4. For shotpoint S1 an average spacing of 2.3 km was achieved between shotpoints S1 and S2, while between shotpoints S2 and S4 the average spacing was 4.6 km. Carbonell et al. (2000b) show both reflection and refraction data sets displayed as CMP stacks (Figs. 9.3-23B and 9.3-23C). The P-wave data of shotpoints 1–4 are shown in detail in Appendix A9-1-8.

The major aim of the E-W main line was to delineate the Moho (crust-mantle boundary) structure beneath the line and thus identify the presence of a crustal root beneath this part of the mountain belt and, if present, to determine the depth of such a root. The recording with three-component instruments allowed to determine both P- and S-wave velocities and thus to place constraints on the rock types beneath the orogen and in particular in the crustal root, if present. At the same time the main profile served to provide velocity control for the near-vertical incidence reflection profile. The main purpose of the N-S cross-line was to determine crustal velocities and structure parallel to the trend of the mountain belt. The fan profiles were planned to detect major E- W variations in the main structural interfaces and in the Moho.

Carbonell et al. (2000b) and Stadtlander et al. (1999) have both interpreted the P- and S-wave data of the seismic-refraction/wide-angle reflection experiment of the URSEIS'95 seismic project (see, e.g., Fig. 9.3-23A). Both author groups found the presence of a crustal root at least 10 km thick beneath the central part of the orogen, where the crustal thickness increased from 43 to 45 km at the margins of the transect up to 53–56 km beneath the central part of the profile. However, the center of this crustal root appeared to be displaced by 50–80 km to the east of the present-day maximum topography. Also, an upper crustal body from 10 to 30 km depth on average, with a high P-wave velocity of 6.3 km/s, was found and interpreted as consisting of mafic and/or ultramafic rocks. Another major feature of the seismic models showed the presence of high P- and S-wave velocities (7.5 and 4.2 km/s, respectively) at the base of the crustal root. P-wave velocities of 8.0–8.2 km/s and S-wave velocities of 4.5–4.7 km/s were found as characteristic features of the upper mantle.

The Vibroseis reflection profiling component of URSEIS'95 provided a high-resolution crustal-scale image of the unextended southern Uralide orogen. The section showed a marked lateral and vertical variation of reflectivity throughout the entire crust which was interpreted to differentiate the former margin of the East European craton to the west from the accreted terranes to the east. Continuous reflections and reflective domains correlated with major tectonic features interpreted from surface geology and defined a bivergent orogen in which the Paleozoic collision structure was largely preserved. At depth a less reflective zone between the East European platform and the accreted terrane regions coincided with the location of the Magnitogorsk volcanic arc and corresponded to the crustal root zone visible in the refraction data.

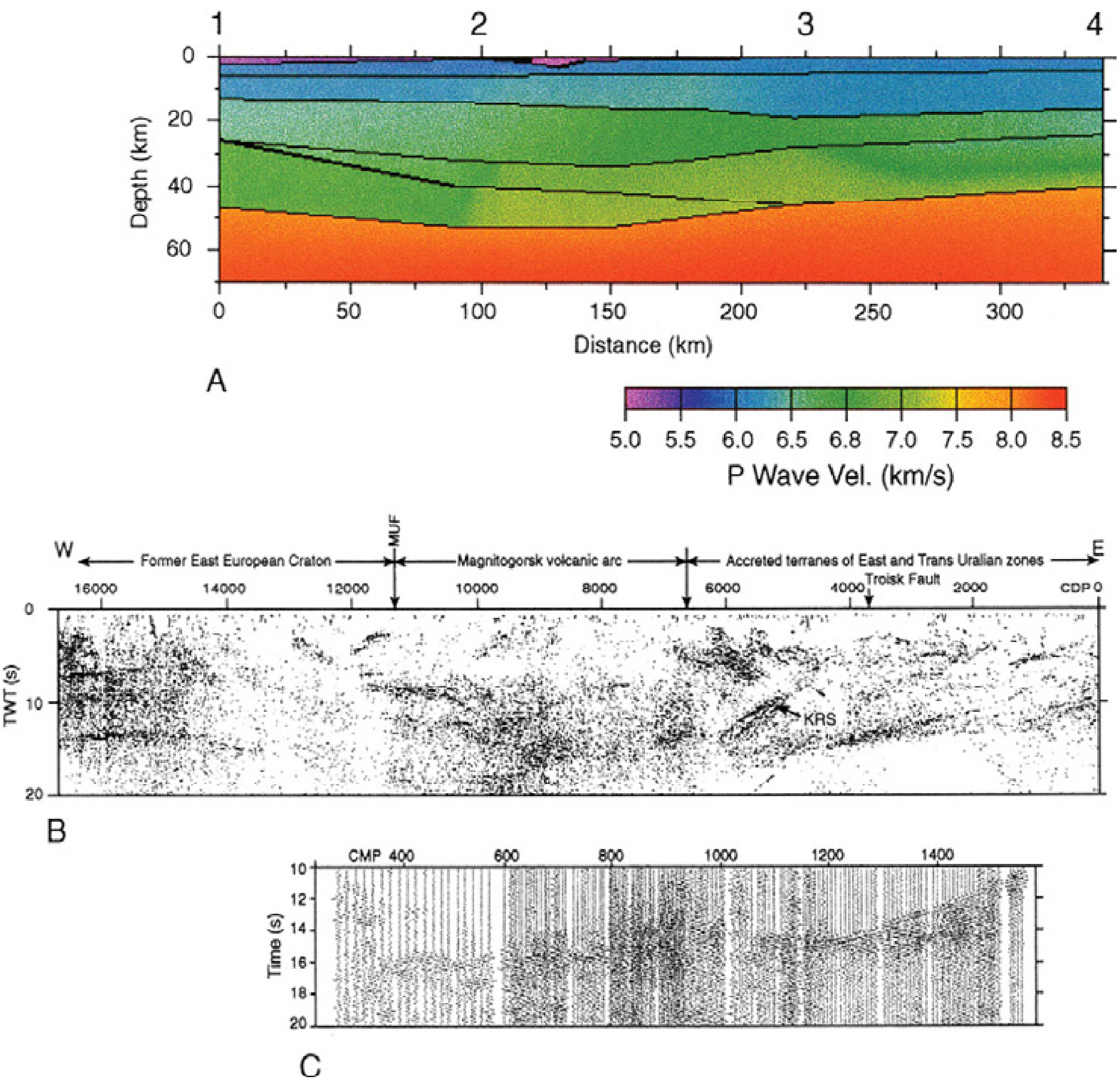



Figure 9.3-23. URSEIS '95 (from Carbonell et al., 2000, plates 1 and 2). (A) P-wave crustal cross section from seismic-refraction data. (B) Normal incidence stack of the explosion seismic-reflection common depth point data set. (C) Stack of the seismic wide-angle reflection data. [Journal of Geophysical Research, v. 105, p. 13,755–13,777. Reproduced by permission of the American Geophysical Union.]

## 9.4. CONTROLLED-SOURCE SEISMOLOGY IN NORTH AMERICA

### 9.4.1 Canada

First established in 1984, LITHOPROBE, Canada's national collaborative Earth science research program, was established to develop a comprehensive understanding of the evolution of the North American continent. It was active in the following 20 years with various phases, and aimed to present 3-D structure and the past geotectonic processes which were responsible for the present structure. For this purpose a series of ten study areas (named transects) was defined, each of which presented a representative geological key target. In total, the transect provided a nearly continuous coverage across the northern part of the North American continent, including an active margin on the west coast, the Cordillera, platform and shield regions, the Appalachians, and the passive margin on the east coast (Clowes et al., 1996, 1999). Phase I started in 1984, Phase II followed from 1987 to 1990. A summary of results to late 1990 was provided by Clowes et al. (1992) and Clowes (1993; see corresponding part of Chapter 8).

In the 1990s, from 1990 to 1998, Phases III and IV followed, and finally, LITHOPROBE's Phase V ran from 1998–1999 to 2002–2003, and, prolonged into 2004–2005, presented a concluding phase of an extensive, Canada-wide project. In Phase IV, four transects were completed (Fig. 9.4.1-01): Abitibi-Grenville

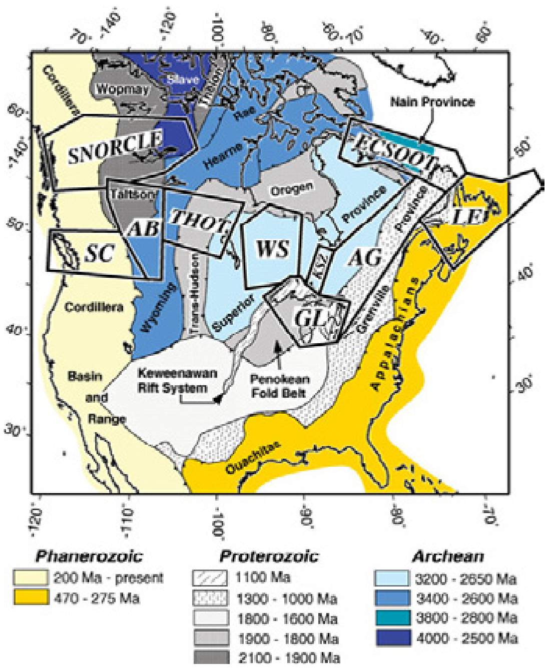



Figure 9.4.1-01. Location of LITHOPROBE transects (study areas) on a simplified tectonic map (from Clowes, 1999, fig. 1). SC—Southern Cordillera; AB—Alberta Basement; SNORCLE—Slave Northern Cordillera Lithospheric Evolution; THOT—Trans-Hudson Orogen Transect; WS—Western Superior; KSZ—Kapuskasing Structural Zone; GL—Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE); AG—Abitibi-Grenville; LE—LITHOPROBE East; ECSOOT—Eastern Canadian Shield Onshore-Offshore Transect. [Episodes, v. 22, p. 2–20. Reproduced by permission of IUGS Publications Committee, Calgary, Alberta, Canada.]

(AG; Hynes and Ludden, 2000), Trans-Hudson Orogen (THOT; Hajnal et al., 2005a), Alberta Basement (AB; Ross, 1999, 2002), and Eastern Canadian Shield Onshore-Offshore (ESCOOT; Wardle and Hall, 2002a, 2002b), while the transects WS (Western Superior) and SNORCLE (Slave Northern Cordillera Lithospheric Evolution) (Cook and Erdmer, 2005) were started in Phase IV, but completed in Phase V. For each transect (Fig. 9.4.1-01), data analyses and interpretations, preparation of syntheses and compilation of a digital database and archive were completed. The main purpose of Phase V, however, was to develop a pan-LITHOPROBE synthesis from all ten LITHOPROBE transects and to complete the LITHOPROBE data archives. In all transects, the primary geophysical data acquired by LITHOPROBE comprised seismic-reflection, seismic-refraction and magnetotelluric data.

Clowes et al. (1996, 1999) presented summaries for various LITHOPROBE transects, as far as available at the time being. Clowes et al. (1996) discussed in particular seismic-reflection lines located in Precambrian regions. Almost all the data was acquired in the early 1990s: a marine example of ECSOOT, recorded in 1992 along the Labrador coast, new reflection data along two corridors in the east-central Grenville province (A-G), recorded in 1993, and data of the western Grenville province (GLIMPCE), recorded in the 1980s (Fig. 9.4.1-02). Furthermore, they discussed seismic-reflection lines located in central to western Canada: profiles recorded in 1991 across the Trans-Hudson orogen (THOT) and profiles recorded in 1992 across AB—central Alberta (Fig. 9.4.1-03). On all data reflectivity persisted throughout the crust, and the Precambrian Moho could be well defined.

In another summary paper, Mandler and Clowes (1998) discussed a number of multichannel seismic-reflection surveys, which were carried out in western and central Canada (Fig. 9.4.1-03) and which showed bright extensive reflectors between 2 and 3 s TWT. Such data were recorded in a number of different locations, as, e.g., in the surveys CAT92, PRAISE94, THOT94, and SALT95, whereby the Winagami reflection sequence was the most extensive one, extending over an estimated area of 120,000 km$^2$ in the Paleoproterozoic basement of Alberta, and the Wollaston Lake Reflector stretched continuously over 160 km in the western hinterland of the Paleoproterozoic Trans-Hudson Orogen.

In the following, starting at the east coast of North America, only the major seismic-refraction and reflection studies in Canada will be discussed, as they were performed within the specific transect regions following the definition of LITHOPROBE's division (Fig. 9.4.1-01).

### *9.4.1.1. LITHOPROBE East*

Already in the 1980s several seismic-reflection and refraction surveys had been performed in the area of the LITHOPROBE East transect (Fig. 9.4.1-04) to investigate the structural architecture of the Northern Appalachians.

In 1991 another large survey, the LITHOPROBE East 91 (LE91) onshore-offshore seismic-refraction/wide-angle reflection experiment added a new wealth of data. The observations comprised closely spaced receivers on land and closely spaced shots offshore. The principal elements of the LE91 experiment were: (1) a marine component north, east, and south of Newfoundland consisting of five profiles using OBSs and an airgun source; (2) a land component within central Newfoundland consisting of three profiles using explosive shots and land seismographs; and (3) a series of combined offshore-onshore recordings (Marillier et al., 1994). More than 400,000 seismic traces were generated from ~800 land sites and 36 OBS stations (Fig. 9.4.1-04).

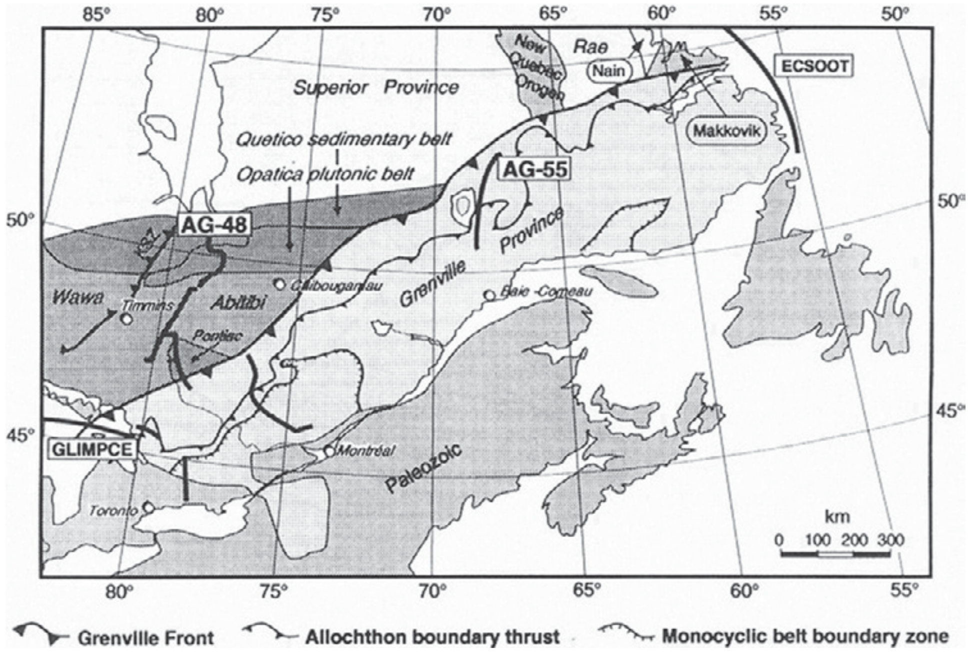



Figure 9.4.1-02. Simplified tectonic map of eastern Canada with location of LITHOPROBE seismic-reflection profiles (from Clowes, et al. 1996, fig. 2). [Tectonophysics, v. 264, p. 65–88. Copyright Elsevier.]

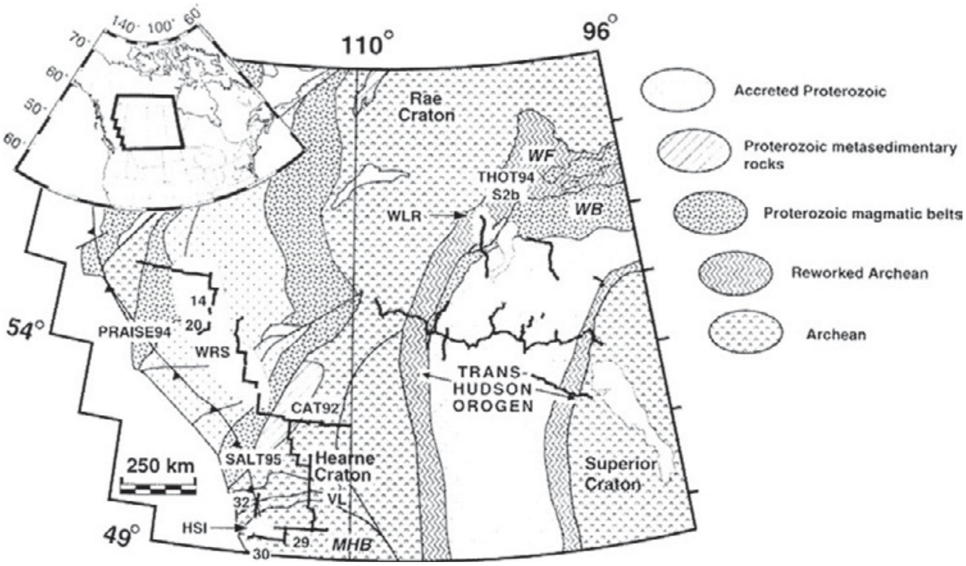



Figure 9.4.1-03. LITHOPROBE seismic reflection lines in western Canada (from Mandler and Clowes, 1998, fig. 1). THOT94 were recorded in the Trans-Hudson Orogen (see also Figure 9.4.1-08). CAT92, PRAISE94 and SALT95 were recorded in the Alberta Basement Transect region (AB; see also Figure 9.4.1-09). [Tectonophysics, v. 288, p. 71–81. Copyright Elsevier.]

One of the results is the thinning of a crustal unit in the west from the Laurentian plate margin toward the east from 44 to 40 km. In central Newfoundland, underneath the central mobile belt, the crust thins to a maximum of 35 km with average continental crustal velocities. Farther east, the crust thickens again to 40 km (Marillier et al., 1994).

### *9.4.1.2. LITHOPROBE Eastern Canadian Shield Onshore-Offshore Transect*

In 1992, 1250 km of normal incidence seismic profiles and, in 1996, seven wide-angle seismic profiles were recorded (Fig. 9.4.1-05) across Archean and Proterozoic rocks of Labrador, northern Quebec, and the surrounding marine areas (Wardle and Hall, 2002a, 2002b; Hall et al., 2002) to study the crustal and upper mantle structure within the Eastern Canadian Shield Onshore-Offshore Transect (ECSOOT). The seismic-refraction data were collected by variable combinations of marine shooting and recording and land recording.

The interpretation revealed a 33–44-km-thick Archean crust with P-wave velocities increasing downwards from 6.0 to 6.9 km/s. A moderate crustal reflectivity was observed, but the reflection Moho remained unclear (Hall et al., 2002). The Proterozoic crust appeared similar, partly with greater thickness and with variable lower crustal velocities, but the crustal reflectivity was strong and the Moho could be clearly defined.

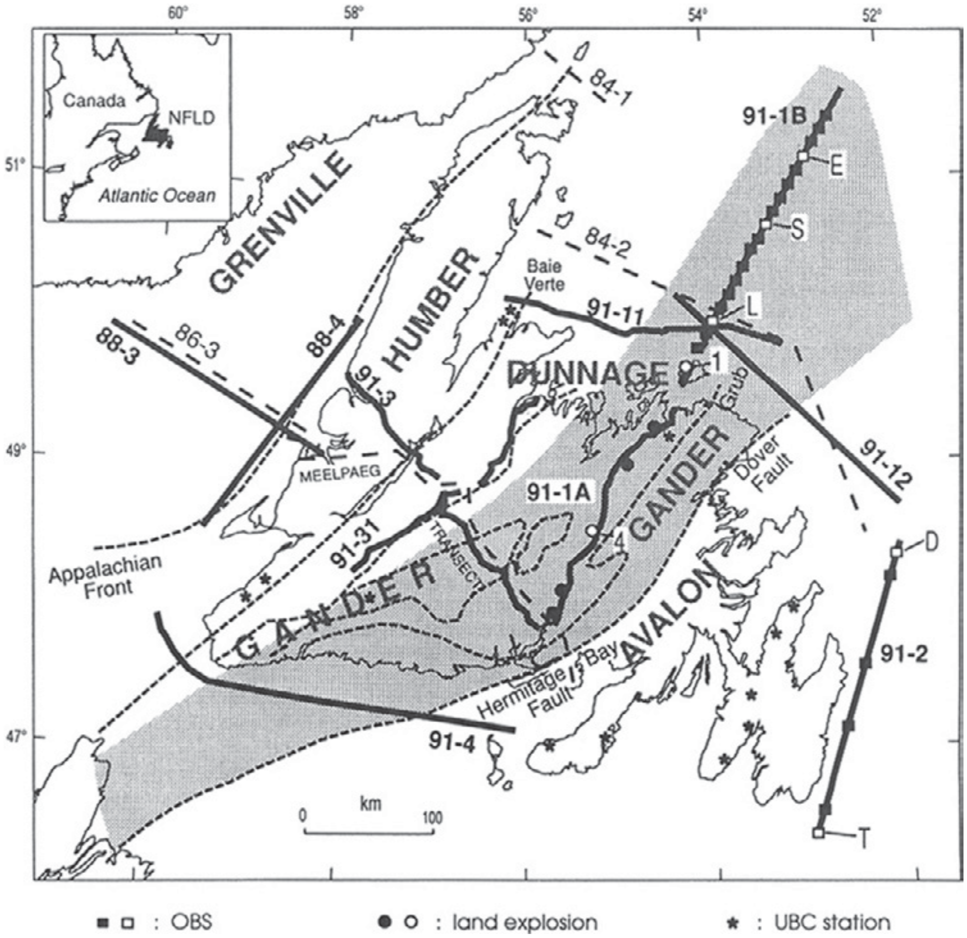



Figure 9.4.1-04. Location map for the Newfoundland Appalachians, showing LITHOPROBE East 1991 and 1988 wide-angle/refraction profiles (from Mariller et al., 1994, fig. 1). Continuous black lines: seismic refraction/wide-angle reflection profiles. Dashed lines: near-vertical incidence reflection profiles. [Tectonophysics, v. 232, p. 43–58. Copyright Elsevier.]

Strong crustal variations appeared in particular along line 5: While under the Ungara Bay the data revealed a 38–40-km-thick crust, under the Torngat Orogen (for location, see Fig. 9.4.1-05) a crustal root down to 55 km depth was found. Farther to the east, under line 5E, the Moho depth decreased in two steps; first to a plateau of 40 km depth under the onshore part of line 5E and then, offshore, almost discontinuously to ~20 km depth (Hall et al., 2002).

### *9.4.1.3. LITHOPROBE Abitibi-Grenville Transect*

The early 1990s saw major seismic activities in the area of the LITHOPROBE Abitibi-Grenville Transect (AG). Both in 1991 and 1993, high-resolution near-vertical incidence seismic-reflection lines were recorded (Clowes et al., 1992, 1996) and in 1992 a major seismic-refraction/wide-angle reflection survey took place (Fig. 9.4.1-06), the results of which were discussed together with previous observations carried out in the area in the 1980s (Winardhi and Mereu, 1997; Mereu, 2000a; White et al., 2000).

The seismic-reflection surveys of the LITHOPROBE studies in the Neoarchean southeastern Superior Province and the Mesoproterozoic Grenville Province in the southeastern Precambrian Canadian Shield included more than 1800 km of near-vertical incidence reflection data.

The 1992 LITHOPROBE Abitibi-Grenville seismic-refraction experiment comprised four profiles across the Grenville and Superior provinces of the southeastern Canadian Shield (Fig. 9.4.1-06). Several hundred kilometers coincided with or were parallel to the 1991 LITHOPROBE reflection lines (Clowes et al., 1992).

Three deployments of 417 instruments, 47 of those three-component seismographs with an average spacing of 1–1.5 km, provided more than 17000 seismograms produced by 44 shotpoints with drillhole shots of 400–800 kg charges. The shot spacing averaged 30 km. The longest line (G-M in Fig. 9.4.1-06) extended in N-S direction over 620 km across the Grenville Front, supported by a 210-km-long cross line (E-F) in strike direction of the Abitibi Greenstone Belt. Its southern end intersected one of the 1986 GLIMPCE experiment lines and the 1988 GRAP line (named "1988 Ontario–New York refraction experiment" in Fig. 9.4.1-06; Hughes and Luetgert, 1992). The two shorter lines to the west (A-B and X-Y) covered 280 km in the N-S direction and 170 km in the W-E direction and were dedicated to study in particular the Sudbury Impact Structure.

The detailed controlled-source seismic surveys were complemented by a seismic tomography study based on data from a teleseismic experiment which comprised a N-S array of 28

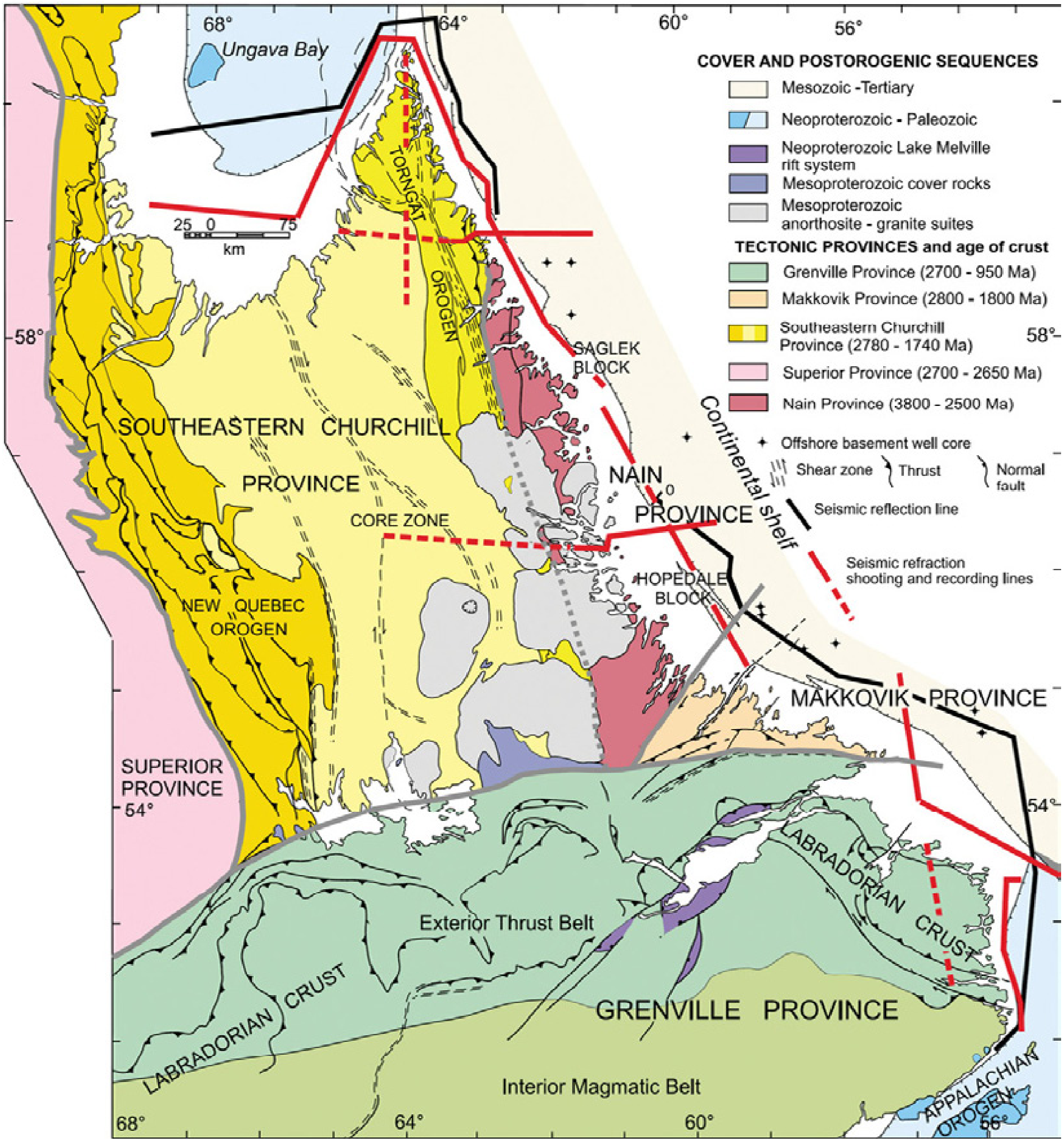



Figure 9.4.1-05. Seismic acquisition lines in the ECSOOT Transect area (from Wardle and Hall, 2002, fig. 2). Straight black lines: marine seismic lines. Straight dashed lines: onshore refraction recording lines. [Canadian Journal of Earth Science, v. 39, p. 563–567. Reproduced by permission of NCR Research Press/National Research Council, Canada.]

broadband seismographs deployed for half a year in 1996 along a 600-km-long line along the Quebec-Ontario border across the Grenville and Superior Provinces of the Canadian Shield (Rondenay et al., 2000).

Complete overviews of the geoscientific investigations including the seismic data and their interpretations were published in special review papers and special volumes (Ludden, 1995; Ludden and Hynes, 2000a, 2000b). In the summary volume on the Abitibi-Grenville transect, Ludden and Hynes (2000a, 2000b) have compiled two composite cross sections based on more than 1800 km of seismic-reflection data, complemented by a seismic-refraction experiment to constrain the velocity structure of the

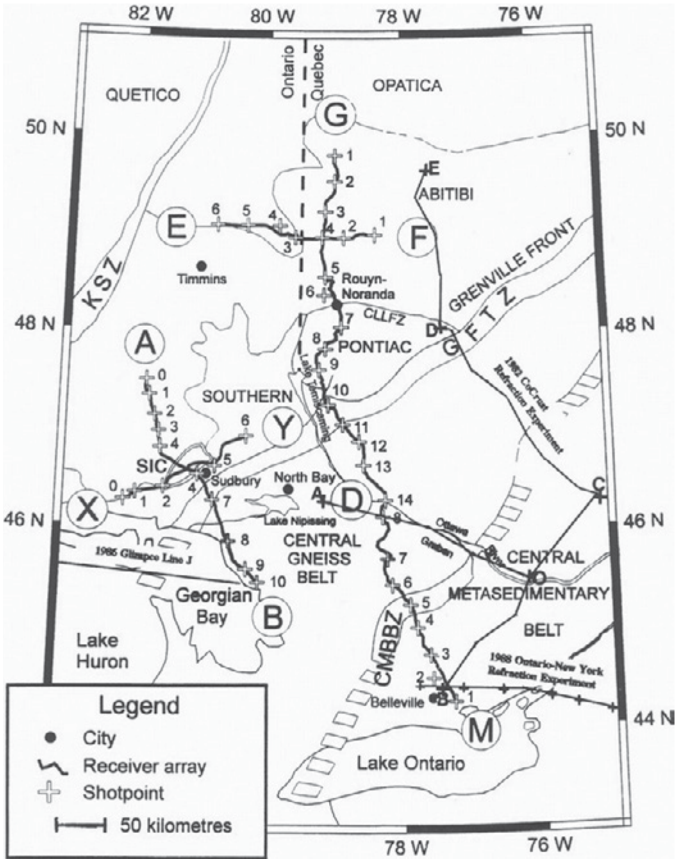



Figure 9.4.1-06. Seismic refraction experiments across the Abitibi-Grenville region (LITHOPROBE AG, from Mereu, 2000a, fig. 1). The location map includes observation lines recorded from 1982 to 1992. GFTZ—Grenville Front Tectonic Zone. [Canadian Journal of Earth Science, v. 37, p. 439–458. Reproduced by permission of NCR Research Press/National Research Council, Canada.]

crust and upper mantle. They have discussed how the different styles of crust formation and tectonic accretion in the Neoarchean and Mesoproterozoic have been defined by these data.

In their interpretation of both the reflection and refraction data, Winardhi and Mereu (1997) have noted the Grenville Front Tectonic Zone (GFTZ in Fig. 9.4.1-06) as a southeast-dipping region of anomalous velocity gradients extending to the Moho. From the wide-angle $P_MP$ reflections, they inferred that the Moho is 4–5 km deeper south of the Grenville Front and varies from a sharp discontinuity to a rather diffuse and flat boundary under the Abitibi-Greenstone Belt north of the Front.

This was essentially confirmed by White et al. (2000) who have seen additional structural boundary zones in the same data. Mereu (2000a, 2000b) has tried to reconcile the differences between seismic models which usually occur when interpreting either reflection or refraction data. He analyzed crustal coda waves that follow the first arrivals and explained the coda by a class of complex heterogeneous models in which sets of small-scale, high-contrast sloping seismic reflectors are "embedded" in a uniform seismic velocity gradient field. A similar approach had first been attempted by Sandmeier and Wenzel (1986) and later on by Enderle et al. (1997) and Tittgemeyer et al. (1996) when reinterpreting seismic-refraction data in Central Europe (see Fig. 9.2.2-12, subchapter 9.2.2).

#### *9.4.1.4. Western Superior Transect*

In 1996, LITHOPROBE acquired two ~600-km-long seismic-refraction/wide-angle reflection profiles oriented parallel (line 1, 600 km, running in an E-W direction north of latitude 50°N) and orthogonal (line 2, 500 km, running in a N-S direction west of longitude 90°W) to the regional geologic strike of the Archean western Superior Province of the Canadian Shield (Fig. 9.4.1-07).

Five hundred fifteen seismographs were deployed at intervals of 1–1.5 km. Shot sizes ranged from 1000 to 3000 kg, with shots being at 30–50 km intervals along each line. The 515 recording devices deployed consisted of: 170 PRS-1 seismographs with 2Hz vertical-component seismometers and 35 PRS-4 with three-component 4Hz seismometers from the Geological Survey of Canada; 170 SGR seismographs with 2Hz vertical-component seismometers or 8Hz vertical-component geophone strings from the USGS; and 140 RefTek seismographs with three-component geophones from IRIS/PASSCAL. The interpretation of the data by Musacchio et al. (2004) showed an intermediate middle crust with 6.6–6.7 km/s velocity, extending to 25–30 km depth and its top varying in depth from 12 to 20 km. Large variations were determined for the lower crust (6.7–7.5 km/s) and upper-mantle (8.0–8.8 km/s) velocities as well as for the Moho depths (32–45 km). From wide-angle reflections a deep boundary was modeled at 110–120 km depth along both profiles. For parts of the lower crust the authors saw a $V_P$ anisotropy of 8%, with the fast propagation normal to the strike, as well as in a 15–20-km-thick layer in the upper mantle (>6%, fast propagation parallel to strike), which dips northwards from 50 to 75 km depth at ~ 10°.

#### *9.4.1.5. Trans-Hudson Orogen Transect*

The investigation of the deep structure underneath the Trans-Hudson Orogen Transect (THOT) of Manitoba and Saskatchewan started in 1991 (Lucas et al., 1994; Hajnal et al., 2005b), when LITHOPROBE acquired ~1100 km of near-vertical incidence Vibroseis seismic-reflection data (gray lines in Fig. 9.4.1-08). The E-W profiles showed strong, E-dipping reflections extending throughout the crust, while the N-S profiles outlined antiformal and synformal structures (Lucas et al., 1994). A second phase was implemented in 1994 with acquisition of 590 km of regional seismic high-resolution reflection profiling with 32 km in the western part plus 370 km in the eastern part of the orogen. This Phase II was the first LITHOPROBE survey to use a 480-channel recording system (Nemeth and Hajnal, 1998; Hajnal et al., 2005b).

The Trans-Hudson Orogen seismic-refraction experiment (THORE) followed in 1993. It consisted of three long-range pro-

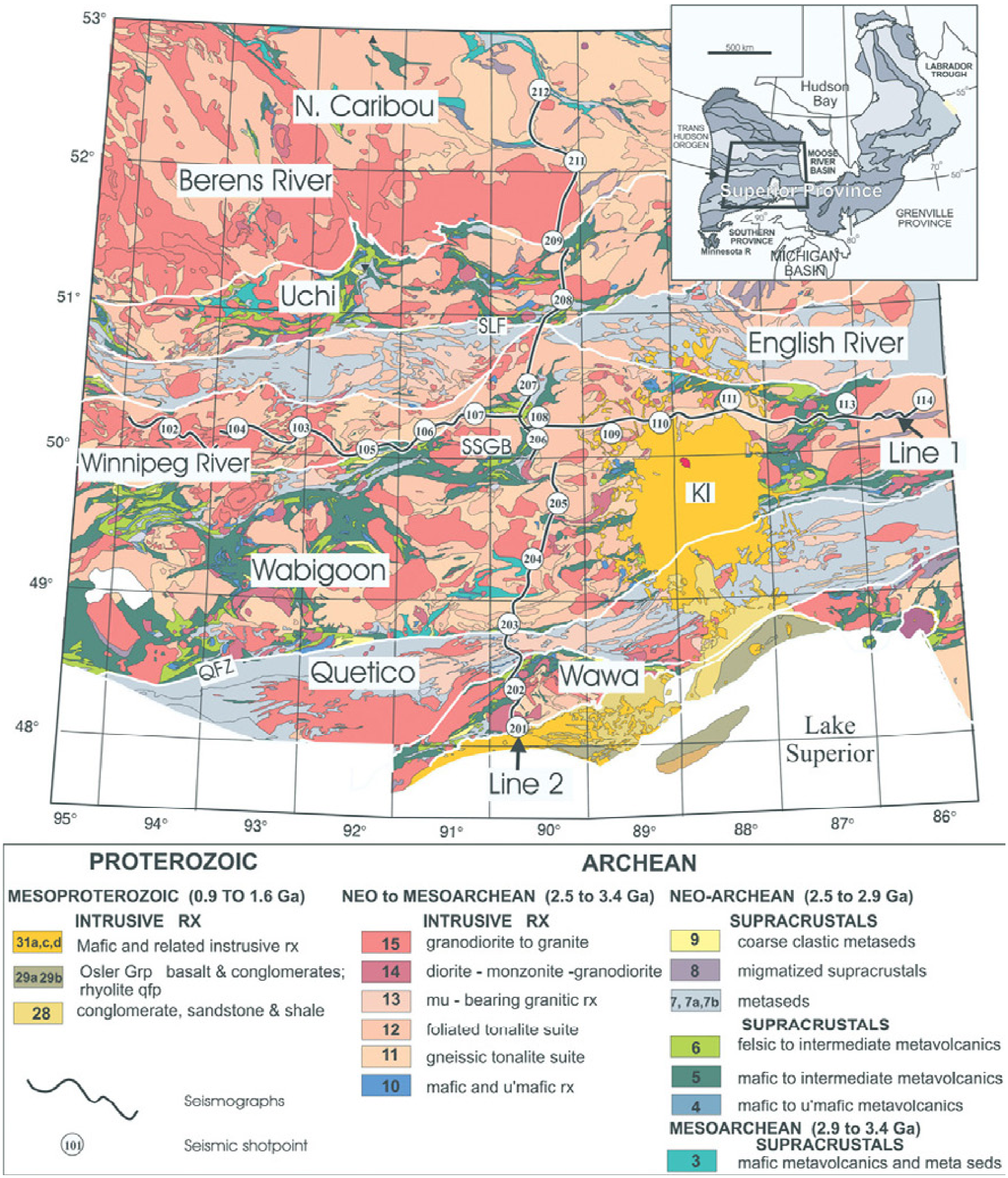



Figure 9.4.1-07. LITHOPROBE seismic survey of 1996 in the region of the Western Superior Transect (from Musacchio et al., 2004, fig. 1). [Journal of Geophysical Research, v. 109, B3, B03304, 10.1029/2003JB002427. Reproduced by permission of American Geophysical Union.]

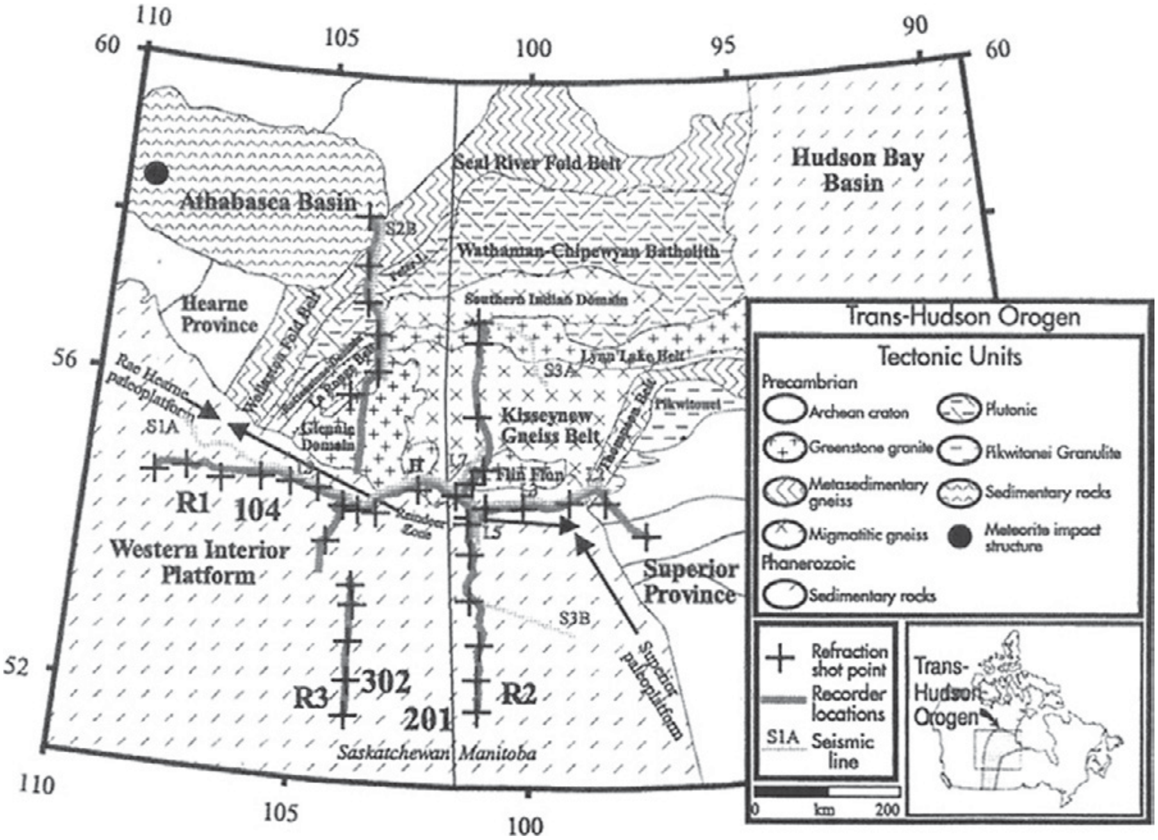



Figure 9.4.1-08. Seismic reflection and refraction lines and shotpoints in the Trans-Hudson Orogen Transect region (LITHOPROBE THOT, from Nemeth and Hajnal, 1998, fig. 1). [Tectonophysics, v. 288, p. 93–104. Copyright Elsevier.]

files, two of them 750 km long and one 500 km long (black lines in Fig. 9.4.1-08). A total of 525 seismographs was involved: 185 PRS-1 (vertical-component Portable Refraction Seismographs with 2Hz seismometers) and 35 three-component PRS-4 recorders with 4Hz seismometers, both from the Geological Survey of Canada; 185 USGS instruments (135 vertical-component Seismic Group Recorders [SGR] with 2Hz seismometers and 50 SGRs with strings of 8Hz geophones); and 120 three-component RefTek recorders with 2Hz seismometers from IRIS/PASSCAL. A maximum of 505 recorders with a spacing of 1.0–1.5 km was deployed along each line. The additional (three-component) systems were placed on a selected broadside profile. Thirty-nine borehole shots (crosses in Fig. 9.4.1-08), with charge sizes from 800 to 3000 kg and an average shot spacing of 50 km, were detonated during acquisition (Nemeth et al., 1996, 2005; Nemeth and Hajnal, 1998). All THOT seismic-reflection lines are also shown in Figure 9.4.1-03.

Generally, crustal velocity structure did not correlate with the location and extent of major geological domains. Crustal thickness variations were substantial. The generally thick crust, 40–45 km, thickened at some locations up to 55 km, and a small root down to ~50 km depth was observed below the Sask craton. However, the crustal thickness variations seemed not to be correlated with major crustal features or boundaries, such as suture zones or paleosubduction zones, suggesting post-collisional variations (Nemeth et al., 2005; White et al., 2005).

Bezdan and Hajnal (1998) discussed, in particular, the results and advantages of four expanding spread profiles that had been recorded in various tectonic domains along line R3, which had been selected on the basis of high crustal reflectivity identified during the regional survey and which provided more detailed images of the crust below 6 s TWT.

#### *9.4.1.6. Alberta Basement Transect and "Deep Probe"*

The Alberta Basement region was of particular relevance for the oil and gas exploration industry and many seismic-reflection surveys were performed which partly were also relevant for the crust underlying the sediments. Two special issues of the *Canadian Journal of Earth Sciences*, edited by G.M. Ross, provided a synthesis and new results for the sub-sedimentary region of the transect (Ross, 1999, 2000, 2002).

In 1992, an innovative 3-D seismic-reflection experiment (CAT92) was undertaken by LITHOPROBE (for location, see Fig. 9.4.1-03) as a feasibility study to study the influence of the basement structure beneath the Western Canada Sedimentary Basin on the sedimentary section above (Kanasewich et al.,

1995) and to study the crustal structure underneath the Alberta Basement Transect. Two-dimensional reflection data sets were recorded on N-S and E-W profiles and totalled 513 km. A 3-D recording was conducted in the region of the intersection of lines 2 and 3 (Kanasewich et al., 1995, fig. 1). The Moho was found at the base of a generally strong reflective crust at 12–14 s TWT, corresponding to ~35–40 km depth, but its sharpness varied.

In 1994 and 1995, LITHOPROBE added further reflection lines to the CAT92 survey (for location, see Fig. 9.4.1-03), including PRAISE94 toward the NW and SALT95 (Southern Alberta Lithospheric Transect) toward the S and SE (Mandler and Clowes, 1998). Furthermore, in 1995, the project VAULT (Vibroseis Augmented Listen Time) was conducted (Ross et al., 1997), another seismic-reflection experiment across the northern margin of the Medicine Hat Block, a region with anomalous sedimentation in southern Alberta and northern Montana, ~250 km south of CAT92 (no. 32 in Fig. 9.4.1-03).

Again, exceptional reflectivity in the crust was recorded, but also good reflectivity within the uppermost mantle down to ~20 s TWT (Ross et al., 1997). The SALT95 data in southern Alberta, ~100 km south of VAULT, were later re-investigated by Welford and Clowes (2004) in conjunction with an extended exploration survey to longer traveltimes, performed in 1999 by a partnership group of Canadian oil industry partners, with respect to an approach to investigate upper-middle crustal structure to depths of ~20 km using industry-style 3-D data.

All seismic-reflection lines, conducted between 1991 and 1995 in the region of the Alberta Basement Transect, are shown as dashed, thin full, or thick full lines in Figure 9.4.1-09 (Clowes et al., 2002) and will be partly included in the following discussion of the South Alberta Seismic Refraction Experiment.

SAREX, LITHOPROBE's Southern Alberta Refraction Experiment, was the corresponding seismic-refraction experiment to the seismic-reflection surveys discussed above. It was conducted in 1995 in connection with the long-range upper-mantle project "Deep Probe" and aimed to provide crustal velocity structure in southern Alberta as part of the Alberta Basement Transect (Clowes et al., 2002; Gorman et al., 2002). The line extended over 800 km from east-central Alberta to central Montana (Fig. 9.4.1-09). Charges, ranging from 800 to 1300 kg, were detonated at ten shotpoints in Canada at a spacing of 50 km and were recorded by 521 seismographs deployed at 1 km intervals. An additional 140 seismographs recorded these shots along the extension of the line into Montana with 1.25–2.50 km intervals. Data from the Deep Probe shotpoint 49 (large star in Fig. 9.4.1-09), a 5000 kg shot co-located within 300 m of shotpoint 1, supplemented the SAREX data for the region south of the Canadian-U.S. border.

The interpretation (Fig. 9.4.1-10) by Clowes et al. (2002) showed a strong relief of the Moho, with its depth within the Canadian territory gradually increasing from ~35 to 45 km near the Canadian-U.S. border and then suddenly increasing to almost 60 km under Montana. Also, the upper-mantle velocity increased in the same direction from 8.05 to 8.2 km/s. The most prominent feature in the crust was a thick (10–25 km) lower-crustal layer (LCL in Fig. 9.4.1-10) with high velocities ranging from 7.5 to 7.9 km/s, located under the Medicine Hat block of southern Alberta and the Wyoming block of Montana.

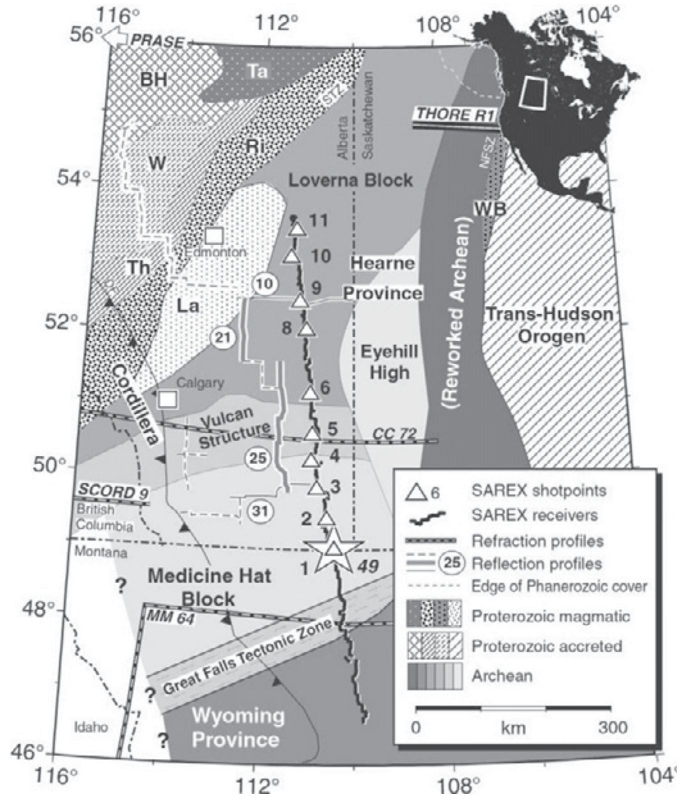



Figure 9.4.1-09. LITHOPROBE's Southern Alberta Refraction Experiment (SAREX) and location of seismic-reflection lines in the Alberta Basement Transect region (from Clowes et al., 2002, fig. 1). Also shown for comparison are the locations of the western end of the THOT-R1 line and SCORD9 (Southern Cordillera Refraction Experiment (Zelt and White, 1995). [Canadian Journal of Earth Science, v. 39, p. 351–373. Reproduced by permission of NCR Research Press/ National Research Council, Canada.]

The "Deep Probe" experiment (Fig. 9.4.1-11) followed immediately after the SAREX crustal investigation. It targeted the velocity structure from the base of the crust to depths as great as the mantle transition zone near 400 km, i.e., to expose the roots of western North America in terms of both lithospheric structure and constructional history (Deep Probe Working Group, 1998; Gorman et al., 2002; Snelson et al., 1998). Originally, it was planned only for the region of Alberta, but could be substantially enlarged by cooperation with the United States institutions and scientists, enlarging the number of seismographs and in particular providing a number of large shots throughout the United States. As a result, the project spanned a distance range of 3000 km from the southern Northwest Territories to southern New Mexico. The study corridor followed approximately the 110th meridian. Ten shots, which were up to 10 times larger than

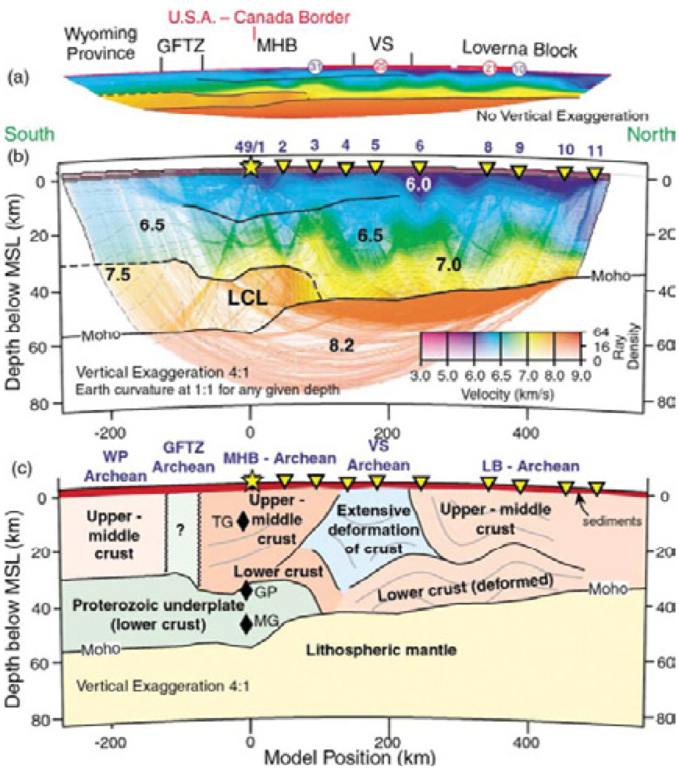



Figure 9.4.1-10. Crustal cross section along LITHOPROBE's SAREX line (from Clowes et al., 2002, fig. 8). LB—Loverna Block; GFTZ—Great Falls Tectonic Zone; MHB—Medicine Hat Block; WP—Wyoming Province; VS—Vulcan Structure. Numbers in circles: crossing of seismic reflection profiles. Diamonds identify locations of xenolith samples. [Canadian Journal of Earth Science, v. 39, p. 351–373. Reproduced by permission of NCR Research Press/National Research Council, Canada.]

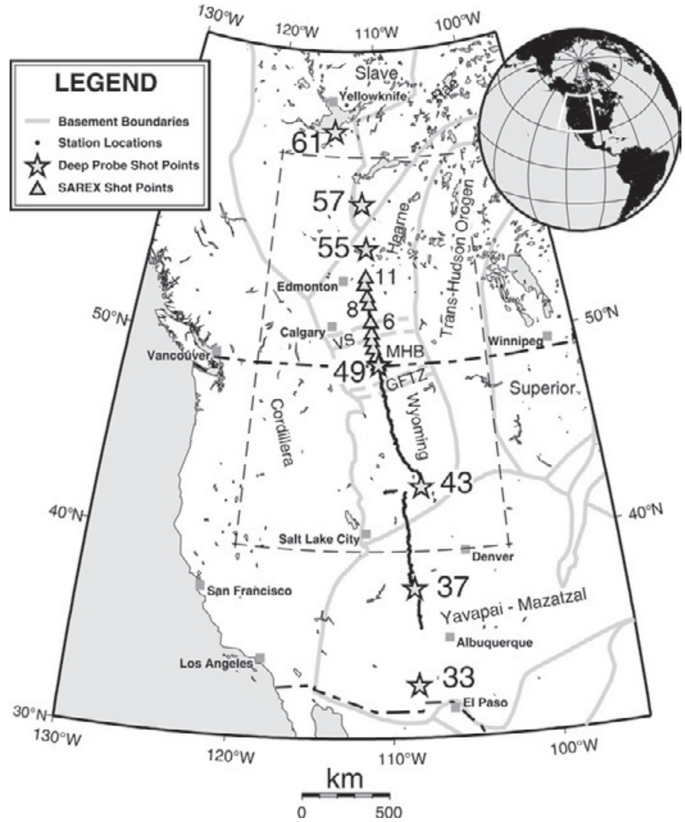



Figure 9.4.1-11. Location map showing the Deep Probe corridor in western North America (from Gorman et al., 2002, fig. 1). Stars: Deep Probe shotpoints. Triangles: SAREX shotpoints. Small circles (overlapping to a line): seismograph locations. [Canadian Journal of Earth Science, v. 39, p. 375–398. Reproduced by permission of NCR Research Press/National Research Council, Canada.]

those used for conventional crustal studies were recorded every 1–2 km to maximum source-receiver offsets of ~2500 km (Gorman et al., 2002). Three of these shotpoints had already served as shotpoints for the SAREX line.

A first lithospheric model displayed a colored mantle cross section reaching from Tyrone, southern New Mexico, to shotpoint 55 in the Alberta plains north of Edmonton (Deep Probe Working Group, 1998, fig. 3), while the analysis of Gorman et al. (2002) focused on the region between Deep Probe shots 43 and 55. Both interpretations resulted in a continental-scale model of the velocity structure of the lithosphere of platformal western Laurentia to a depth of 150 km.

A particular detail of the cross section of the Deep Probe Working Group (1998) is the indication of a low-velocity zone within the uppermost mantle between 50 and 80 km depth. At the top of this zone, the velocity decreased from 7.9–8.0 km/s to 7.6–7.7 km/s. This low-velocity zone gradually thins out to the north, disappearing in Wyoming north of the Cheyenne Belt, and is absent under Canada. A further interpretation (Snelson et al., 1998) concerns mainly the territory of the United States and will be discussed in section 9.4.2.5 (see Fig. 9.4.2-25).

### *9.4.1.7. Slave Northern Cordillera Lithospheric Evolution*

The Slave geological province is a relatively small craton in NW Canada and includes the five oldest rocks in the world. Its southern portion was already addressed in the 1980s during Phases I and II. Phase I in 1984 and 1985 dealt mainly with an investigation of Vancouver Island, and Phase II in 1987–1990 investigated the Southern Cordillera Transect. In particular, in 1989 and 1990, major seismic-refraction work had been carried out by SCoRE (Southern Cordillera Seismic Refraction Experiment) which was described in Chapter 8.5.2 (Clowes et al., 1995).

The ACCRETE seismic experiment in 1994 (thin lines south of line 22 in Fig. 9.4.1-12), in fact a U.S.-based project (Hammer et al., 2000; Morozov et al., 1998, 2001), was the first project executed in the area of the Northern Cordillera Transect. Using coastal fjords for access along Portland Inlet and adjoining waterways, 1700 km marine multichannel reflection data were acquired. The tuned airgun array also served as the seismic source for a seismic-refraction/wide-angle reflection survey, a land-based seismograph array deployed inline with Portland Inlet from north of Prince Rupert to the north around Stewart (see Fig. 9.4.1-12). Thirty-seven instruments were positioned along the

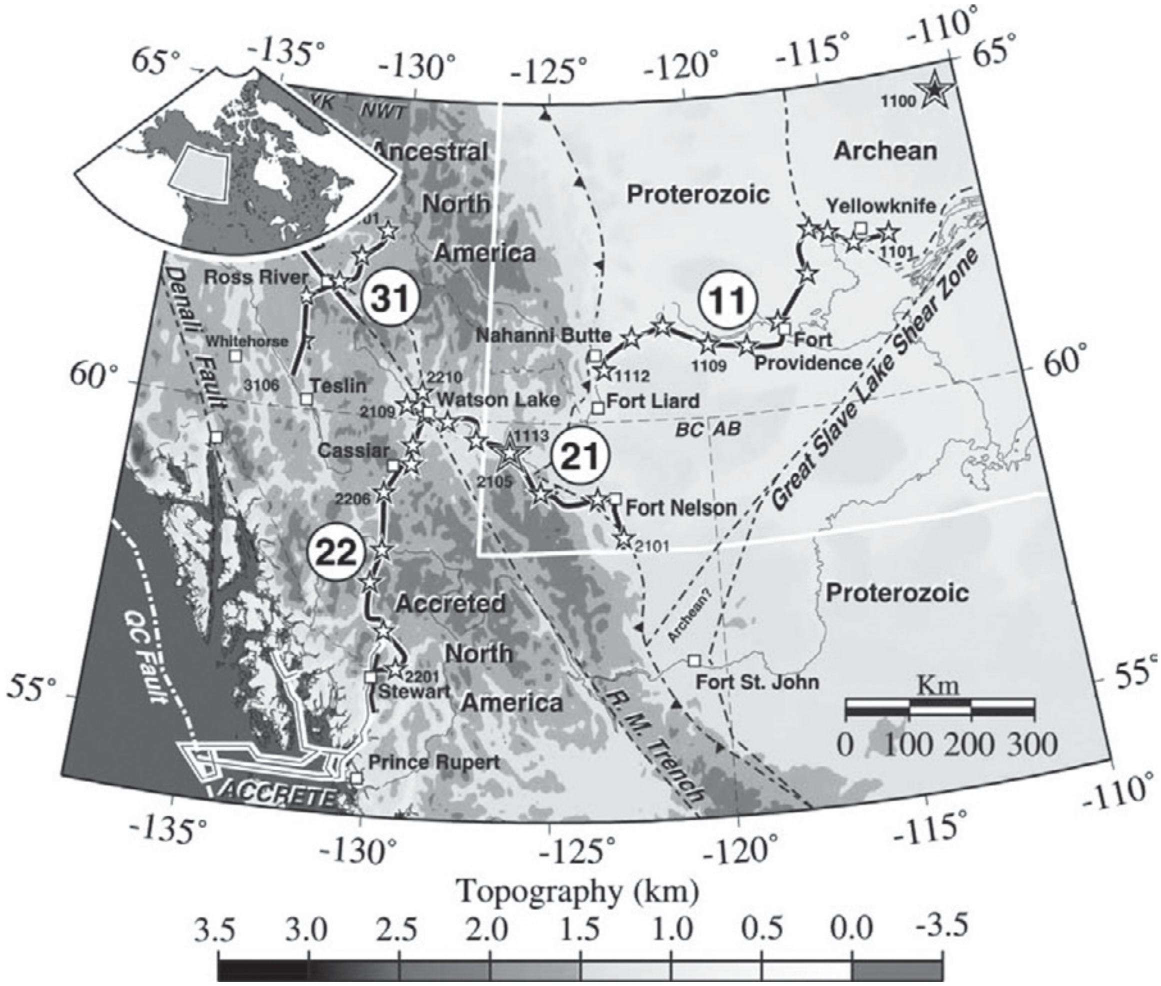



Figure 9.4.1-12. Location map of the SNoRE seismic refraction project within the SNORCLE transect (from Fernandez-Viejo and Clowes, 2003, fig. 1). [Geophysical Journal International, v. 153, p. 1–19. Copyright John Wiley & Sons Ltd.]

eastern shore of the 200-km-long fjord, and other instruments were deployed along a 105-km-long profile extending NE through the Coast Mountains.

The next project was a multichannel seismic-reflection survey, which was carried out in 1996 along line 11 (Fig. 9.4.1-12; Cook et al., 1998, 1999), following a sequence of roads from east of Yellowknife to west of Nahanni Butte in the southwestern part of the Northwest Territories.

In 1997, the SNoRE97 seismic-refraction/wide-angle reflection survey was initiated to investigate the lithosphere of the entire Slave Northern Cordillera Lithospheric Evolution (SNORCLE) Transect (Clowes et al., 2005; Fernández-Viejo and Clowes, 2003). It included four individual lines along three transect corridors (Fig. 9.4.1-12). A total of 37 shots, with a nominal spacing of 60 km and with charges between 800 and 3000 kg, was recorded by 594 recording stations, 226 of them being three-component seismographs. The resulting station spacing was 0.8–1.2 km for vertical and 1.6–2.4 km for three-component recordings.

The easternmost line 11 was 720 km long and ran along the same roads as the 1996 reflection survey. Only along line 11 were all available seismographs exclusively deployed in-line. In addition, two shots 200 km off either end of line 11, with charges of 5000 and 10,000 kg, were recorded along the line (stars in Fig. 9.4.1-12) to determine structure to the base of the lithosphere. Lines 21 and 22 were observed simultaneously, and during the recording of line 31, a number of stations were installed up to 100 km off-line along a fault running perpendicular to the main line.

Together with earlier studies, a 2000-km-long lithospheric velocity model resulted that extended from the Archean Slave craton to the present Pacific basin (Clowes et al., 2005). In Figure 9.4.1-13, we show the 800-km-long part of lines ACCRETE and SNORE 22. On a large scale, the seismic data identified a variety of orogenic styles. The Moho remained remarkably flat and shallow (33–36 km) across the majority of the transect, despite the variety of ages, orogenic styles, and tectonomagmatic deformations, which were crossed by the seismic corridors. Laterally variable crustal velocities were consistently slower

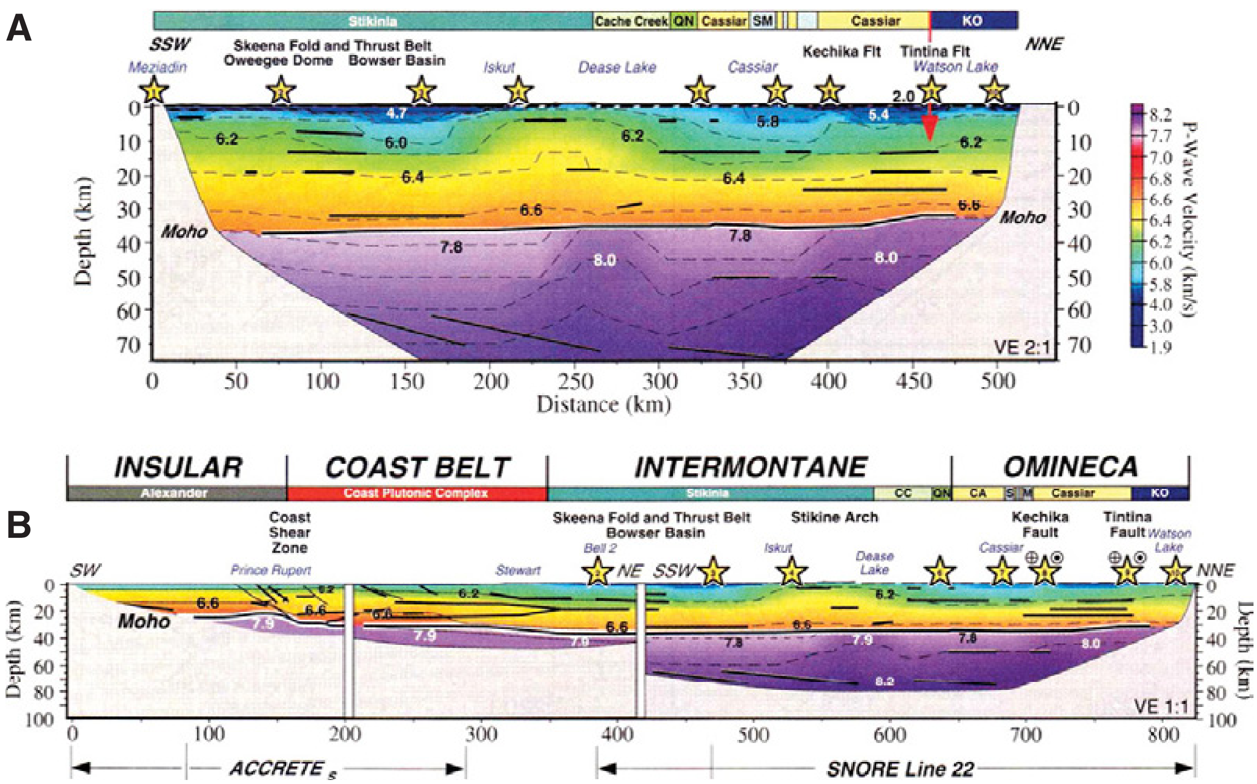



Figure 9.4.1-13. Crustal cross section (A) along SNoRE line 22 vertical exaggeration 2:1), (B) along lines AACRETE and SNoRE 22 (no vertical exaggeration). Stars—shotpoints (from Hammer and Clowes, 2004, fig. 5). [Journal of Geophysical Research, v. 109, B06305, doi:1029/2003JB002749, 19p.] Reproduced by permission of American Geophysical Union.]

beneath the Cordillera than beneath the cratonic crust. Beneath much of the Cordillera slow upper mantle velocities (7.8–7.9 km/s) were observed.

More details of the various lines were published separately. Fernández-Viejo and Clowes (2003) concentrated on line 11, including the reflection data. Lines 21 and 22 were interpreted by Welford et al. (2001) and Hammer and Clowes (2004), respectively. The latter tied their interpretation of the data of line 22, generated by explosive shots, to data of the earlier overlapping project ACCRETE of 1994 (Fig. 9.4.1-13). Creaser and Spence (2005) finally dealt with line 31. An alternative interpretation of part of the seismic-reflection data was prepared by Evenchick et al. (2005).

### 9.4.2. United States

The successful start of IRIS in the United States in the 1980s, with its sub-organization PASSCAL had led to a continuous funding by the U.S. National Science Foundation in the 1990s and the following decade until today. Today, PASSCAL is one of two major instrumentation programs of IRIS, with the second one being the Global Seismic Network. PASSCAL instruments and support are available to the academic research community according to the rules and policies set by the IRIS Executive Committee. It nowadays operates a pool of over 1000 portable seismic instruments to record active source reflection data, active source refraction data or natural sources, e.g., earthquakes. The instrumentation is supported by an instrument center which at present is based at the New Mexico Institute of Mining and Technology at Socorro, New Mexico.

With its large set of digital instrumentation, for which small research teams with limited funding can apply, it enables a large number of North American university large-scale research projects and also many major projects of the USGS. On request, IRIS/PASSCAL also provides technicians for individual fieldwork and other technical support, and it takes care that all instruments are properly maintained and replaced, when needed. A special study group also ensures that the equipment is properly modernized, whenever advanced technology provides new generations of equipment. The rules of IRIS/PASSCAL also request from their customers that copies of the data that are obtained by their equipment are provided to be stored at a joint data center.

#### *9.4.2.1. Alaska*

The experiments for deep seismic exploration in mainland Alaska had reached a preliminarily final stage with the northernmost segment of TACT (Trans-Alaska Crustal Transect), namely the seismic exploration of the Brooks Range in 1990 (Fuis et al.,

1997) which was already described in Chapter 8.5.4. A synthesis of all data obtained in the 1980s and 1990s along the 1350 long TACT corridor was finally compiled by Fuis et al. (2008). The most distinctive crustal structures and the deepest Moho along the transect were found near the Pacific and Arctic margins. Near the Pacific margin, a stack of tectonically underplated oceanic layers was inferred which were interpreted as remnants of the extinct Kula (or Resurrection) plate. Continental Moho just north of this underplated stack was found more than 55 km deep. Near the Arctic margin, the Brooks Range is underlain by large-scale duplex structures that overlie a tectonic wedge of North Slope crust and mantle. There, the Moho has been depressed to nearly 50 km depth. In contrast, the Moho of central Alaska is on average 32 km deep (Fuis et al., 2008).

Approximately parallel to the TACT survey, in 1989 the EDGE Alaska transect had been acquired over the shelf off Alaska using a 4-km streamer and airgun sources (Moore et al., 1991). Along this transect, the top of the subducting oceanic plate, representing the North America–Pacific plate boundary, had been traced to more than 30 km depth beneath the inner shelf 250 km landward of the trench. In 1994, a deep-crustal seismic-refraction experiment was carried out along the same transect using an airgun source with a high shot density and closely spaced OBHs. This experiment aimed to verify the previously observed deep crustal structure, to increase the depth range, and to obtain velocity information for an improved image reconstruction (Ye et al., 1997). In addition, across EDGE 302 several perpendicular profiles were shot. Eleven OBHs were used to record the shots, which had an average spacing of 90 m, along six single-line segments, with two each along EDGE 301 and 302 and two perpendicular crossing EDGE 302. The new experiment improved in particular the data along EDGE 302 (Fig. 9.4.2-01).

In the 1990s, experimental work continued in the marine areas around mainland Alaska. A large-scale active-source seismic experiment (Fig. 9.4.2-02) in the region of the Aleutian vol-

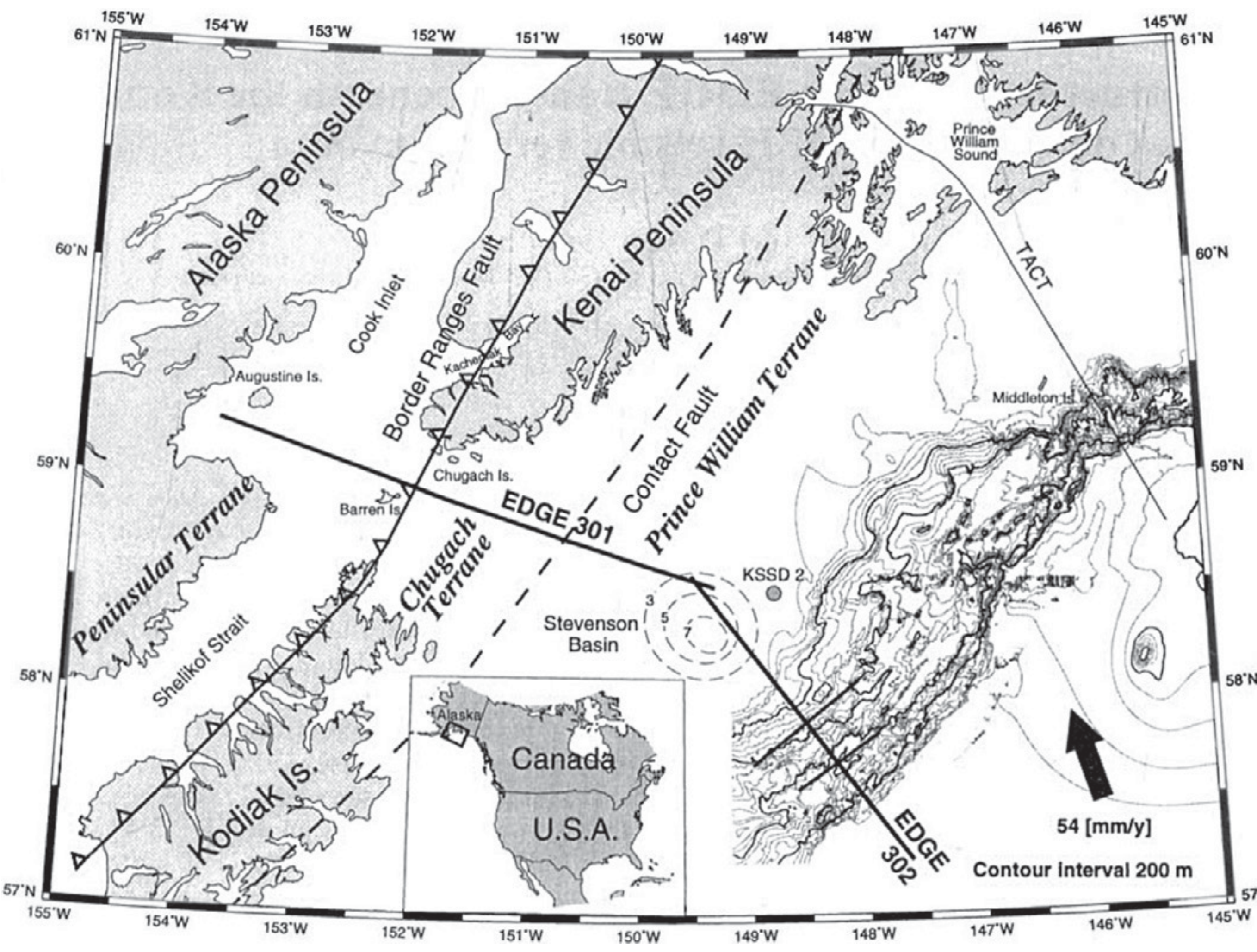



Figure 9.4.2-01. EDGE Alaska transect over the shelf off Alaska south of Kenai Peninsula (from Ye et al., 1997, fig. 1). [Geophysical Journal International, v. 130, p. 283–302. Copyright John Wiley & Sons Ltd.]

canic arc and the back-arc Bering shelf was carried out in July 1994 (Fliedner and Klemperer, 1999). The first component of the experiment was a multichannel reflection acquisition along the ship tracks shown in Figure 9.4.2-02, of which tracks A1, A2, and A3 were used by Fliedner and Klemperer (1999) for their interpretation.

The second component was a wide-angle reflection survey, for which OBS/OBH recorders and 27 portable three-component stations on islands and on the Alaskan peninsula were deployed (Fig. 9.4.2-02). These seismometers recorded the airgun shots of the reflection survey up to distances of 450 km (300 km in the island are section). Stations 1–15 recorded all shots along ship

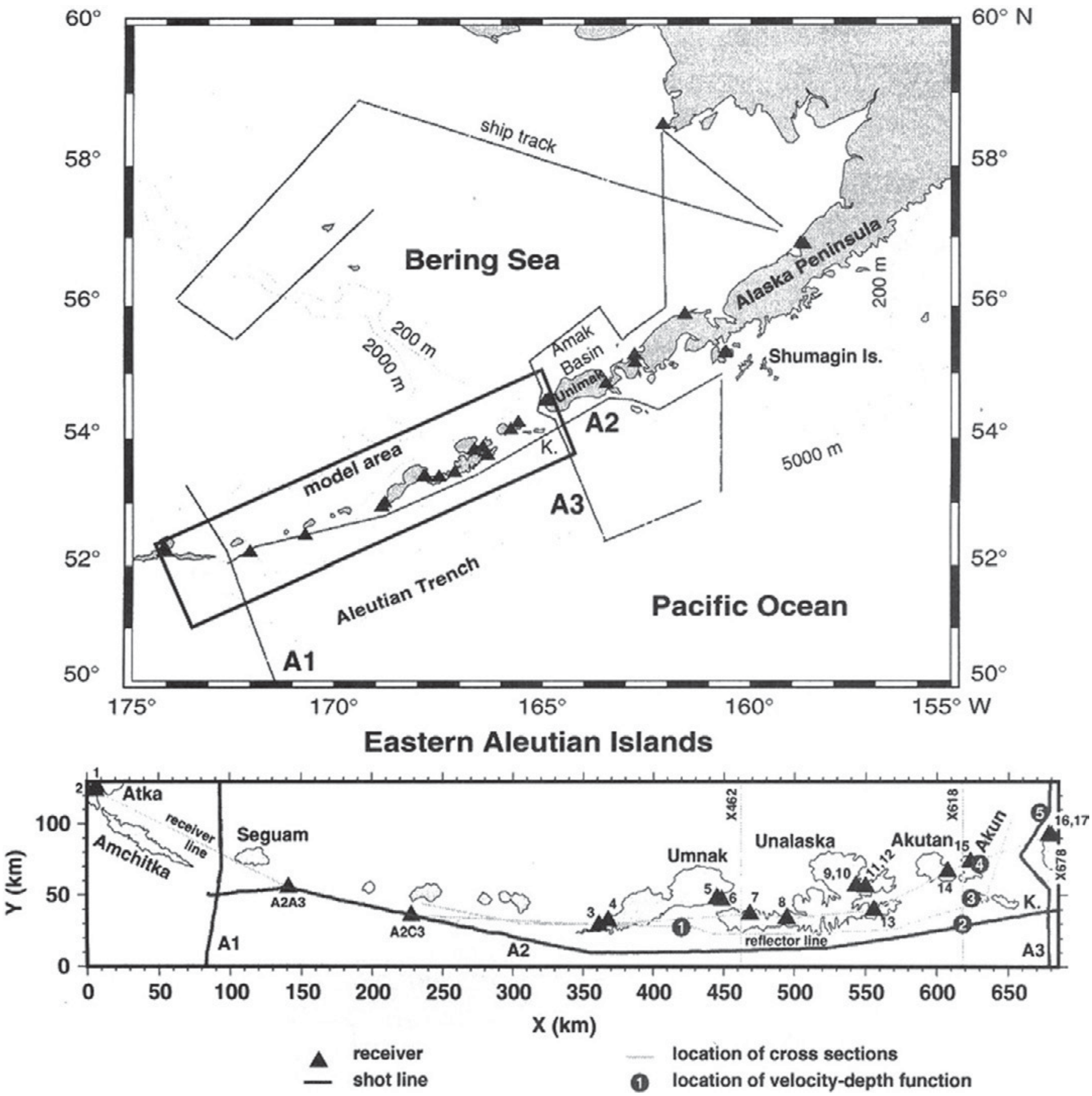



Figure 9.4.2-02. Seismic experiment in southwestern Alaska (from Fliedner and Klemperer, 1999, fig. 1) showing location of the stations and ocean-bottom seismometers (triangles). [Journal of Geophysical Research, v. 104, p. 10,667–10,694. Reproduced by permission of American Geophysical Union.]

lines A2 and A3, while usable data from shots of line A1 were only recorded by OBS/OBH stations and two land stations 1 and 2 on the island of Atka (Holbrook et al., 1994a). Stations to the east of track line A3 recorded only shots from this line but not from line A2.

From their interpretation, Fliedner and Klemperer (1999) concluded that the Aleutian crust, with the Moho at ~30 km depth, is almost continental in thickness and consists of ~20% preexisting oceanic crust (Fig. 9.4.2-03). The average crustal velocity of 6.7 km/s was substantially higher than usually found for continental crust, and the uppermost-mantle velocity beneath the Moho was generally 7.8 km/s, increasing gradually to 8 km/s near 50 km depth.

One month later, in August 1994, two long deep-crustal seismic-reflection profiles were acquired along the continental shelf with airgun shots at intervals of 50 m or 75 m (thick lines in Fig. 9.4.2-04). The profiles had a total of 3754 km and on each line ~40,000 shots were fired. The marine reflection data were obtained by a 4.2 km, 160-channel digital streamer, with record lengths up to 16–23 seconds. From these marine airgun sources in the Bering and Chukchi Seas, 14 deep-crustal wide-angle seismic-reflection and -refraction profiles were recorded (Brocher et al., 1995a; Appendix 9-2-1). This was achieved by an onland array of twelve three-component stations (quadrangles in Fig. 9.4.2-04) which continuously recorded the airgun shots on line EW94-10, where the ship steamed northward.

Seven of these stations also recorded the airgun shots fired during the southward cruise along the USGS/Stanford line. The map also shows the ship tracks and stations of the July 1994 survey (thin dashed lines and quadrangles in Fig. 9.4.2-04), part of which was interpreted by Fliedner and Klemperer (1999).

Airgun sources were recorded as far as 400 km, which enabled to image reflectors throughout the crust, to determine crustal and upper mantle refractor velocities, and to provide constraints on the geometry of the Moho. The recorded crustal and upper mantle refractions indicated substantial variations in the crustal thickness along the transect (Brocher et al., 1995a).

### 9.4.2.2. Pacific Northwest

As a consequence of its position along the Cascadia subduction zone, where the Juan de Fuca plate of the Pacific Ocean is being subducted underneath the North American continent, western Washington and adjacent British Columbia are regarded as having a potential for seismic hazards, as was recognized by earthquakes, which have been observed within the subducting Juan de Fuca plate since 1940, and by crustal faults within the Puget Lowlands capable of large earthquakes (Fisher et al., 1999; Parsons et al., 1998).

On the Canadian side, the deep crustal structure had been investigated intensely by LITHOPROBE already in the 1980s, the Southern Cordillera Transect being the first target of LITHOPROBE. Its Phase I had started in 1984 with primary scientific activity on Vancouver Island including a detailed seismic-refraction and -reflection survey (Clowes et al., 1986, 1987a) and was continued in Phase II from 1987 to 1990, involving SCoRE, the Southern Cordillera Seismic Refraction Experiment, which was carried out in 1989 and 1990 (Clowes et al., 1995) and which provided long profiles over the southern Canadian Cordillera, including lines across the Intermontane and Coast belts and Vancouver Island (see Chapter 8.5.2).

In the adjacent United States, the Pacific Northwest and the Cascadia margin were investigated in great detail in the 1990s to better understand crustal structure, tectonics, and earthquake and volcanic hazards. Five major projects were carried out. (1) The first project was conducted in 1991 and concentrated on the area south and east of Puget Sound. (2) Close to this survey in 1995 another offshore-onshore experiment investigated again the seismic properties of the subducting Juan de Fuca plate. (3) The

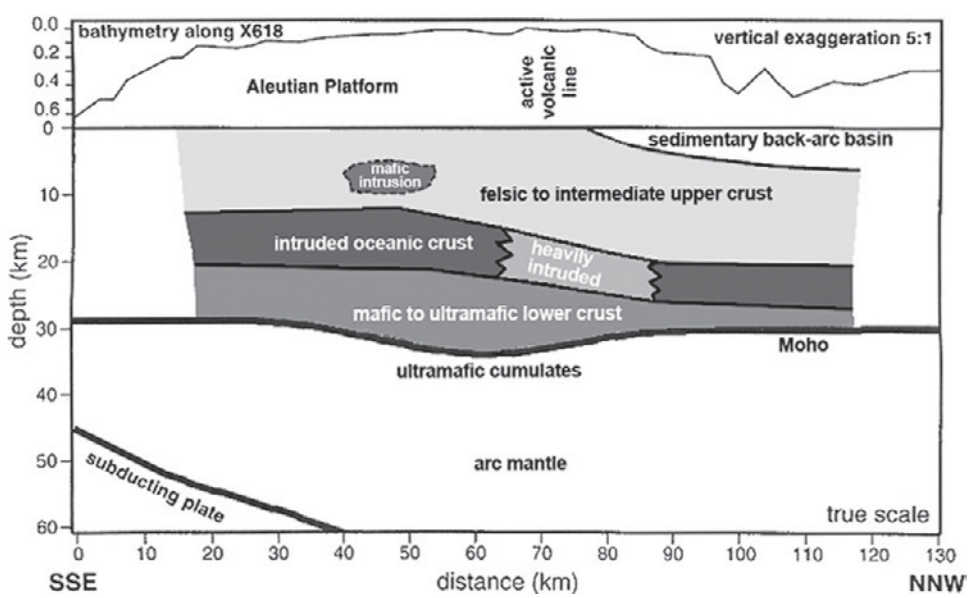



Figure 9.4.2-03. Cartoon cross section through the eastern Aleutian Arc (from Fliedner and Klemperer, 1999, fig. 13). [Journal of Geophysical Research, v. 104, p. 10,667–10,694. Reproduced by permission of American Geophysical Union.]

marine waterway system of Puget Sound and its outlets into the Pacific Ocean around Vancouver Island, the Strait of Juan de Fuca in the south, and the Strait of Georgia in the north, were the target of the first part of SHIPS 98 (Seismic Hazards Investigation of Puget Sound, also called "Wet SHIPS") in 1998 (Brocher et al., 1999). (4) SHIPS 99 (also called "Dry SHIPS") was an E-W–directed onshore-offshore survey across the Puget Sound targeting the Seattle Basin (Brocher et al., 2000). (5) The Cascadia margin of southern Oregon was the goal of an onshore-offshore seismic experiment in 1994.

The 1991 Pacific Northwest experiment investigated the Cascade Range and the adjacent Puget Lowland of Washington and was a major seismic-refraction/wide-angle reflection survey. It was preceded by an offshore-onshore seismic experiment in 1989 (Fig. 9.4.2-05) which collected both seismic-reflection and -refraction data along an east-west line in central, coastal Oregon (Brocher et al., 1993a, Trehu and Nakamura, 1993; Appendix 9-2-2). An airgun array served as a seismic source for 8 OBSs positioned on the continental shelf, slope, and adjacent abyssal plain. The airgun shots were also successfully recorded by 10 land stations, consisting of two one-component stations and 8 three-component five-day recorders which were deployed in coastal central Oregon and recorded continuously the airgun signals up to recording distances of 180 km.

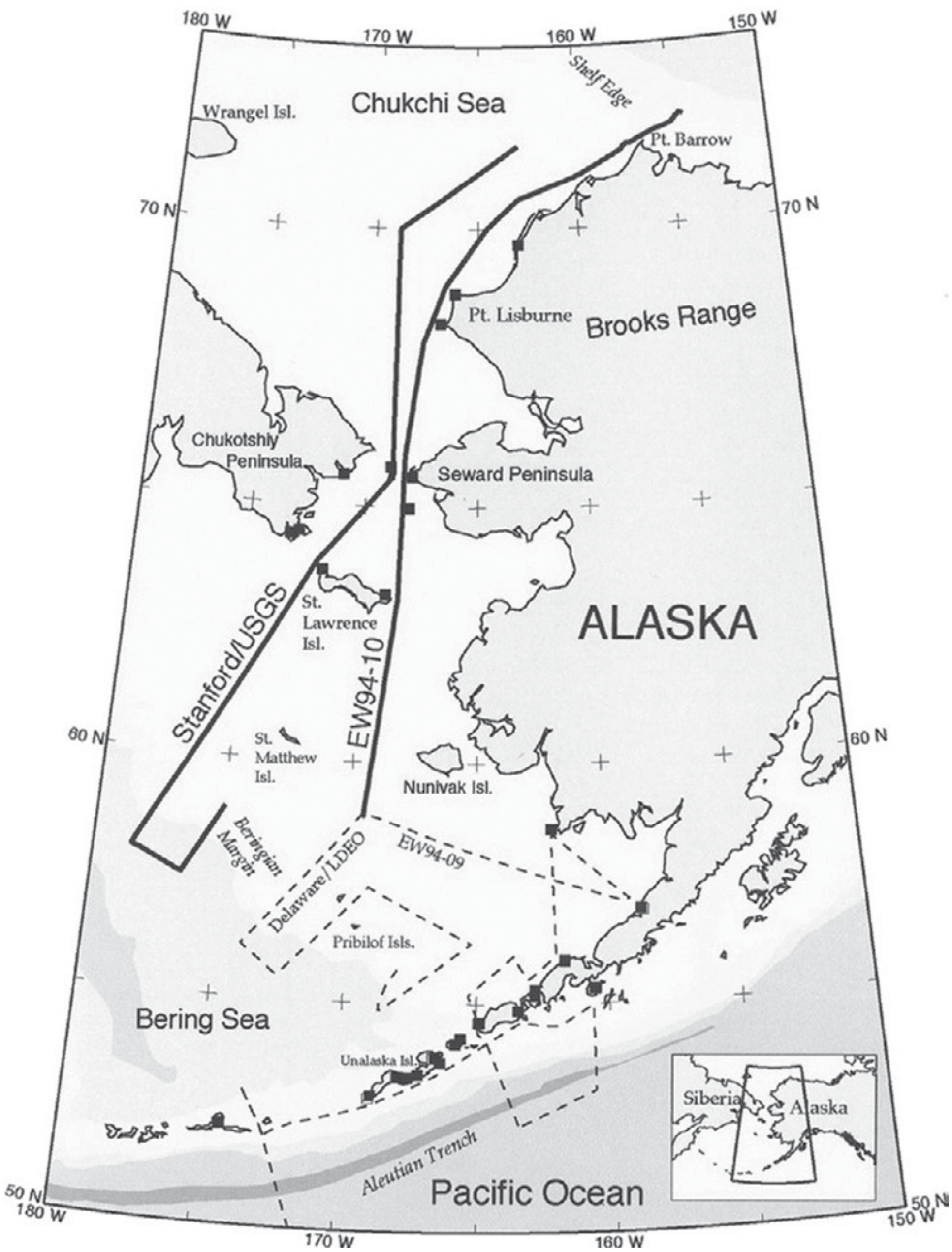



Figure 9.4.2-04. Seismic reflection and refraction survey in the Bering and Chukchi Seas and recording sites in Alaska and Siberia (from Brocher et al., 1995a, fig. 1). [U.S. Geological Survey Open-File Report 95-650, 57 p.]

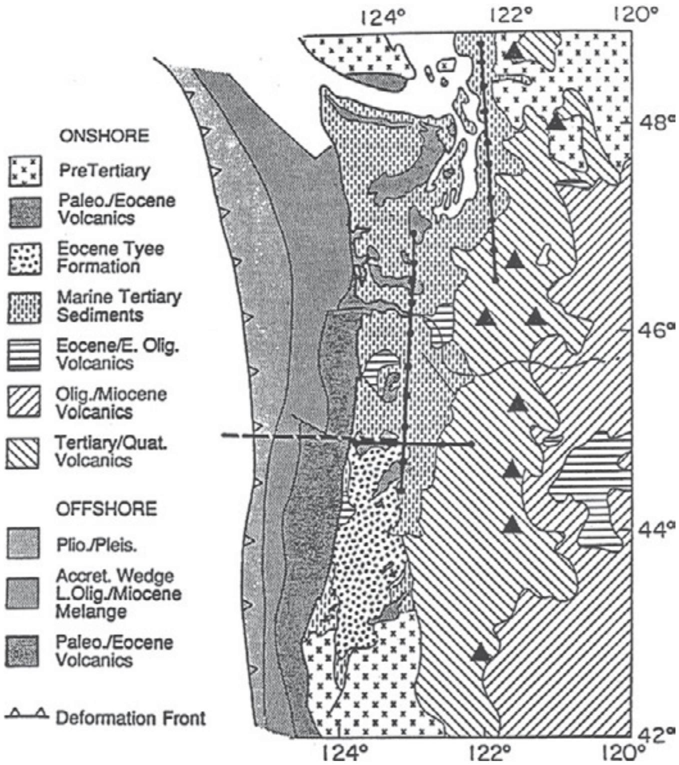



Figure 9.4.2-05. Seismic experiment in Washington and Oregon (from Luetgert et al., 1993, fig. 1) showing the location of the 1991 seismic refraction lines (dots—shotpoints) and the 1989 marine seismic reflection profile (dashed line). Triangles—Cascade volcanoes. [U.S. Geological Survey Open-File Report 93-347, 71 p.]

The 1991 Pacific Northwest experiment comprised two N-S directed profiles, 270 km and 330 km long, and a shorter E-W line, which were recorded in the Puget Basin and the Willamette Valley of western Washington (Fig. 9.4.2-05). Along each of the N-S lines 10 shots of 900–1800 kg charge size at intervals of ~30 km were recorded by ~500 seismic recorders, spaced 600–700 m (Luetgert et al., 1993). The E-W line, located at 44.8–44.9°N, crossed the western 330 km long N-S profile, here 7 shots of 1000 kg charge size were recorded by 420 stations at 350 m spacing (Trehu et al., 1993; Appendix 9-2-2). This line followed and complemented a marine seismic-reflection profile and complementary onshore-offshore recordings (dashed line in Fig. 9.4.2-05), collected in 1989, to the east to the Cascade foothills (Trehu et al., 1990; Trehu et al., 1993; Appendix 9-2-2). Prior to this study crustal studies had only been performed in the Columbia Plateau and the central Cascade Range of Oregon (Catchings and Mooney, 1988a, 1988b; Leaver et al., 1984), and Schultz and Crosson (1996) had investigated the seismic velocity structure across the central Washington Cascade Range from refraction interpretation with earthquake sources.

Interpretations of these data were published separately for the individual lines. Miller et al. (1997) investigated the eastern line which reached farther north and concentrated on the eastern part of the Puget Sound Basin and adjacent Cascade Range. Trehu et al. (1994) investigated the western N-S line in the Willamette Lowland and the W-E cross line across the Cascade Range (Fig. 9.4.2-05). Along the western N-S profile, a 45–50-km-thick crust was obtained, whereby the lower 5–8 km were interpreted as oceanic crust of the subducted Juan de Fuca plate (Trehu et al., 1994). For the east-west profile, the authors suggested a dip of the Moho from 22 km beneath the outer shelf to 30 km beneath the coast (Trehu et al., 1994). The resulting crustal thickness along the eastern line (Miller et al., 1997) ranged from 42 km in the north to 47 km in the south, where the lower-crustal layer appeared as a 2–8 km transitional layer with velocities of 7.3–7.4 km/s above a low-velocity upper mantle (7.6–7.8 km/s).

In 1995, wide-angle seismic data were collected along a 325-km-long E-W profile near 46.5°N (Fig. 9.4.2-06), which crossed southern Washington from Willapa Bay to the Columbia plateau (Parsons et al., 1998, 1999). About 1500 stations spaced at 200 m intervals recorded 17 large explosions. The wide-angle reflections of these explosion data showed images of the subducting slab and provided continuous first arrivals to 230 km offsets (Parsons et al., 1999).

In 1996, the German research vessel *Sonne* conducted an extensive investigation of the offshore Oregon and Washington margins (Flueh et al., 1997; Parsons et al., 1999). Offshore Washington (Fig. 9.4.2-06), a total of more than 14,500 airgun sources (50–150 m spacing) was fired, with just over 6000 detonated along lines instrumented with a cumulative total of 53 OBSs on the seafloor. The balance of the airgun sources was fired for marine multichannel profiles. On land, 44 stations distributed along three profiles (triangles in Fig. 9.4.2-06) continuously recorded all the airgun sources. The OBSs and on-land profiles provided continuous phase coverage across the margin and enabled a comparison of the structure from north to south.

The 3-D crustal and upper mantle model (Parsons et al., 1998; Fig. 9.4.2-07) was the basis for an extended 3-D modeling of coastal Washington (Parsons et al., 1999), in which surface geology, well data and seismic data were combined to resolve details of the individual crustal layers/terranes at the transition from ocean to continent.

The experiment offshore central Oregon (Fig. 9.4.2-08, Flueh et al., 1997; Gerdom et al., 2000) comprised two N-S directed offshore strike lines over the continental shelf, on each of which six OBHs and one OBS were deployed and recorded the airgun shots of both lines. On the E-W dip line, located at ~44.7°N, 14 OBHs and six OBSs were deployed. Additionally, the airgun shots were recorded by a linear array of 29 seismometers located across central western Oregon, located near 44.7°N, slightly south of the 1991 Pacific Northwest E-W line described by Trehu et al. (1993, 1994). Along the E-W line, the subducting oceanic crust of the Juan de Fuca plate could be traced from the trench to some 10 km landward of the coast with an increasing dip angle changing gradually eastward from 1.5° west of the trench to 16° under the coast. The Moho at the coast line was found at 30 km depth (Gerdom et al., 2000).

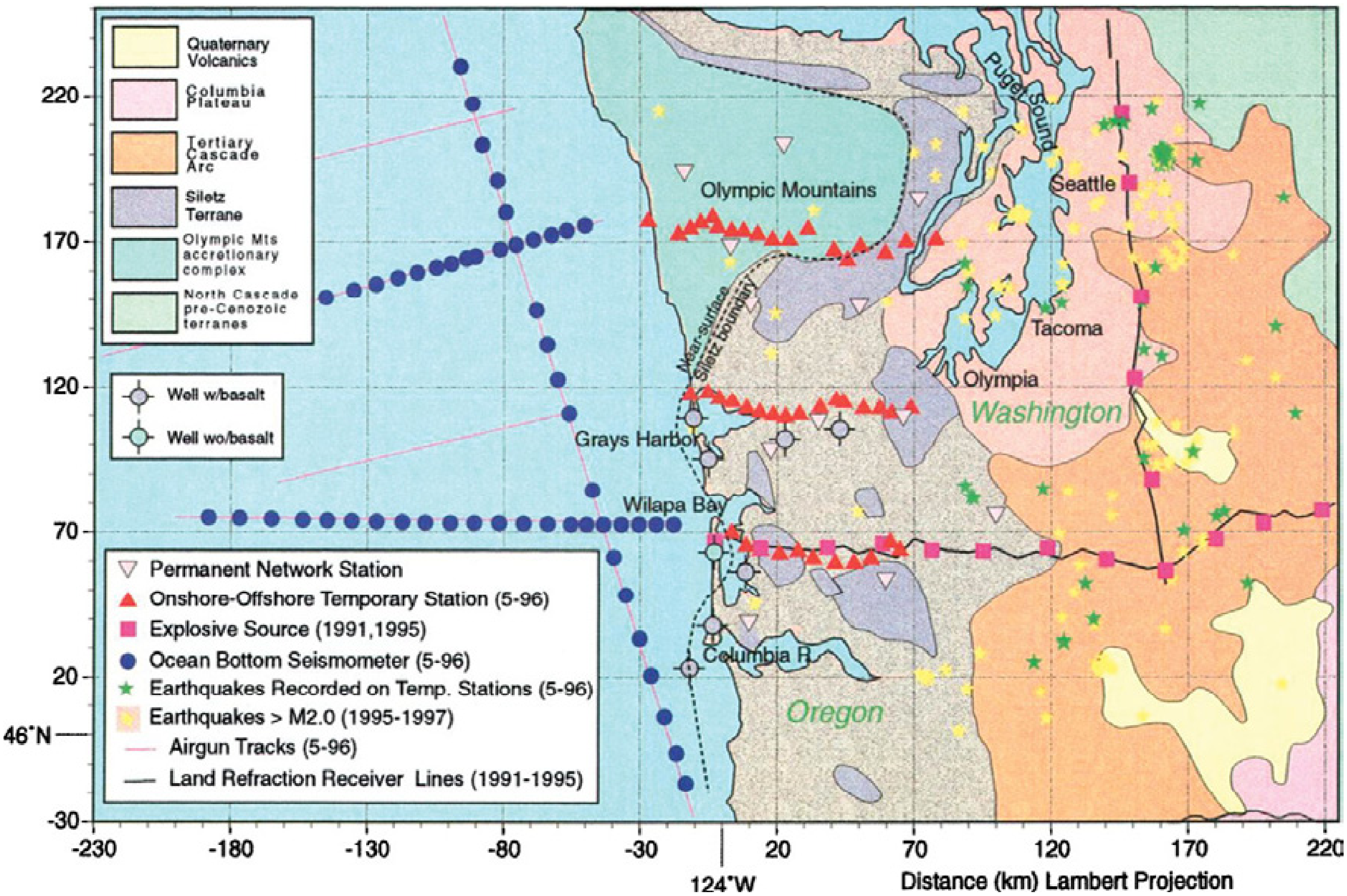



Figure 9.4.2-06. Locations of controlled and earthquake sources and recording stations in 1991–1996 in western Washington (from Parsons et al., 1999, plate 3). The southern E-W and the eastern N-S lines include the locations of the 1991 explosive sources of Figure 9.4.2-05. [Journal of Geophysical Research, v. 104, p. 18,015–18,093. Reproduced by permission of American Geophysical Union.]

The 1998 SHIPS seismic project (Fig. 9.4.2-09) used airgun shots and acquired 1000 km of deep-crustal multichannel reflection profiles and 1300 km of wide-angle lines in northwestern Washington and southwestern British Columbia (Brocher et al., 1999; Appendix 9-2-2). In total, nearly 33,300 airgun shots were released. The experiment was carried out in several stages. First, nearly 13,000 airgun shots were fired by one of the ships exclusively for wide-angle recording, while the ship, without towing a streamer, sailed from Lake Washington near Seattle up and down through the Puget Sound and the narrow Hood Canal and then continued into the Strait of Juan de Fuca, returned and went on through the Strait of Georgia to Texada Island, Canada.

For the airgun shot profiles 257 land stations, 95 of them equipped with three-component seismometers, and 15 OBSs were deployed which allowed recordings to far offsets. Most of the land stations were deployed along the shores of the waterways to provide quasi–2-D lines. Three-dimensional coverage with land stations was achieved east of Puget Sound in the Seattle area and in the southeastern edge of Vancouver Island. Linear arrays were deployed SSE-ward from Tacoma, Washington, and along the Fraser River north of Vancouver, British Columbia. In addition, the seismic stations of the permanent earthquake monitoring arrays in Washington and British Columbia recorded all the shots. During the deployment, 30 local earthquakes were also recorded.

In the second part of the experiment (Fig. 9.4.2-09), the ship towed a 2.4-km-long, 96-channel digital seismic streamer which reduced the size of the airgun array. With this equipment, reflection lines were acquired along the three main waterways. The near-trace of the seismic streamer was 200 m and the far-trace 2575 m behind the ship, resulting in a group interval of 25 m. Using a second ship, towing two short streamers of 300 m length, it was furthermore possible to record constant offset profiles (COPs), thus prolonging the recording aperture of the first ship. Also, 12 expanding spread profiles were acquired as the two ships sailed apart from each other in opposite directions at the same speed, maintaining a common midpoint.

Only a limited amount of wide-angle data could be collected. Most stations provided usable data to offsets of 40–50 km, although airgun shots can be recorded to much greater ranges if

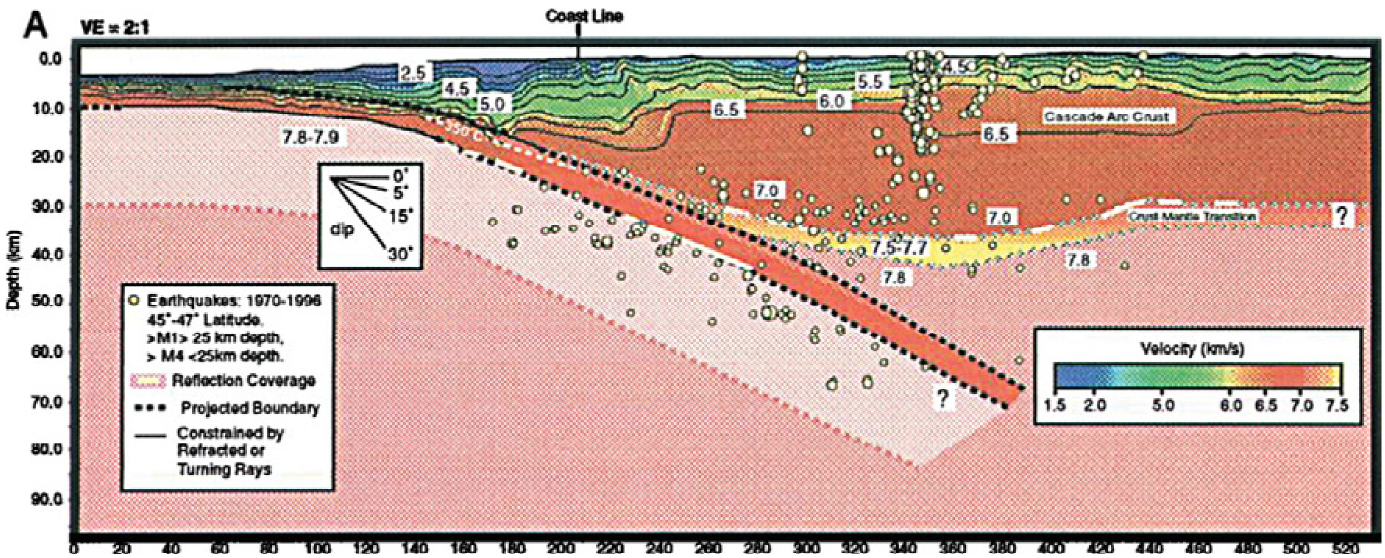



Figure 9.4.2-07. Crustal and upper mantle velocity cross section across southern Washington (from Parsons et al., 1999, plate 2A). [Journal of Geophysical Research, v. 104, p. 18,015–18,093. Reproduced by permission of American Geophysical Union.]

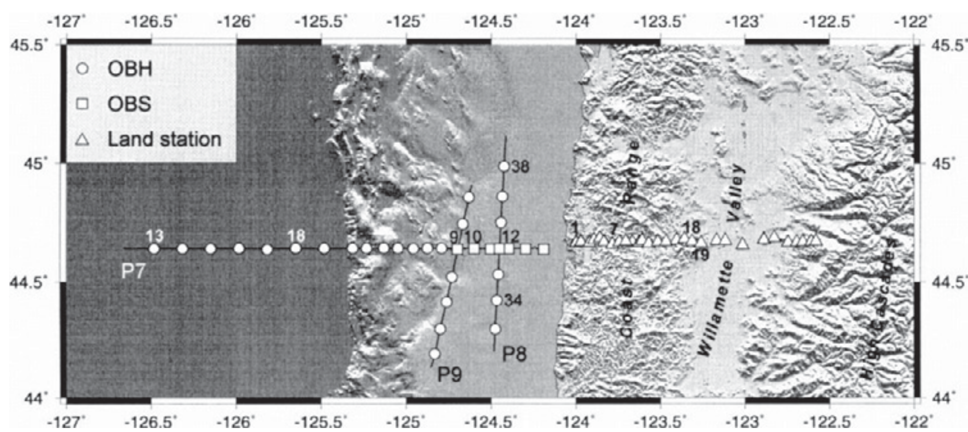



Figure 9.4.2-08. Location of seismic refraction lines of 1996 across the Oregon continental margin (from Gerdom et al., 2000, fig. 2). [Tectonophysics, v. 329, p. 79–97. Copyright Elsevier.]

recorded on remote bedrock sites. One of the results from this experiment was a tomographic model which imaged the upper crust of the Puget Lowland to a depth of ~11 km (Brocher et al., 2001).

As a sequel to the 1998 airgun experiment "Wet SHIPS," the 1999 SHIPS experiment, nicknamed "Dry SHIPS," acquired a 112-km-long E-W–trending multichannel seismic-reflection and -refraction line in the Seattle Basin (Fig. 9.4.2-10). A total of 1008 seismographs were deployed at an average spacing of 100 m, and 29 shotpoints were organized at ~4 km intervals along the seismic line. To bridge the gap in line 2 caused by the Puget Sound, eight of the seismographs were OBSs, deployed in the Sound at water depths ranging from 50 to 250 m. In addition to the active source data, 35 continuously recording three-component stations, deployed at 4 km intervals, and the OBSs sampled data from 26 regional earthquakes and blasts as well as 17 teleseismic events (Brocher et al., 2000; Appendix 9-2-2). The receiver geometry provided both a low-fold reflection coverage and a high-fold refraction coverage through the center of the basin. Due to the geometry of the local waterways and the availability of public lands, the line had to be "broken" into lines 1 (39 km long) and 2 (88 km long) which overlapped for 12 km on the Kitsup Peninsula (Fig. 9.4.2-10). Lines 3–6 (22–25 km long) are N-S–trending fan lines to provide 3-D control in the eastern part of the Seattle Basin. A total of 38 shots was detonated at the 29 shotpoints, all in drill holes, with charges ranging from 11 to 1136 kg. At nine shotpoints, shots were repeated to allow stacking of the data.

A subsequent experiment, the "Kingdome SHIPS" phase, followed in early 2000 with 206 stations deployed with a nominal spacing of 1 km in a hexagonal grid in the city of Seattle to look at the site response within the upper 2 km of the Seattle Basin. They recorded the implosion (100,000 kg delayed) of the Kingdome sports arena and four additional 68-kg shots at the corners of the grid, recording usable energy up to 10 km (Snelson, 2001).

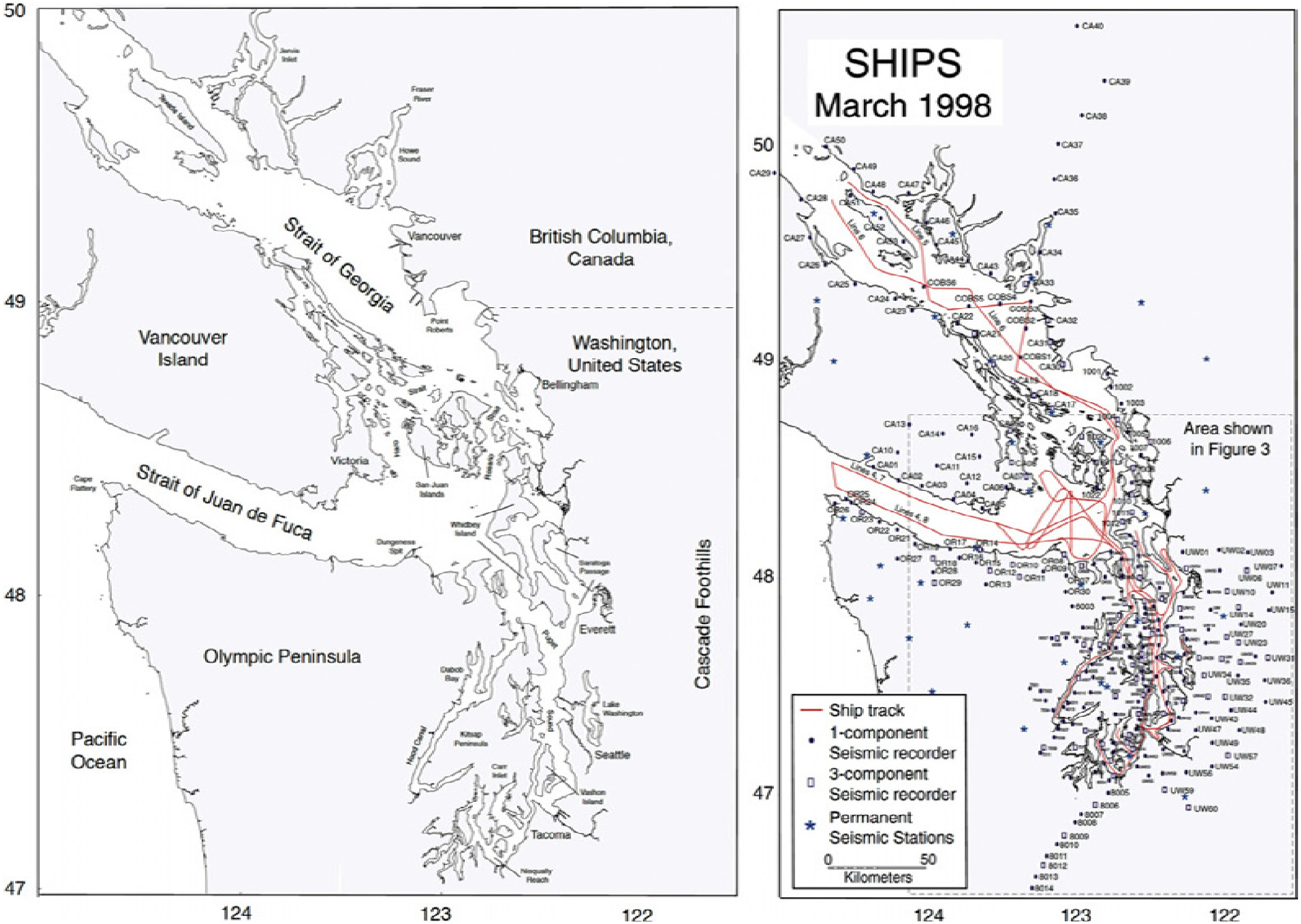



Figure 9.4.2-09. Left: Marine waterways of NW Washington and SW British Columbia (from Brocher et al., 1999, fig. 1). Right: Location of ship tracks (seismic lines) and recorders of SHIPS 98 seismic project (from Brocher et al., 1999, fig. 2). [U.S. Geological Survey Open-File Report 99-314, 123 p.]

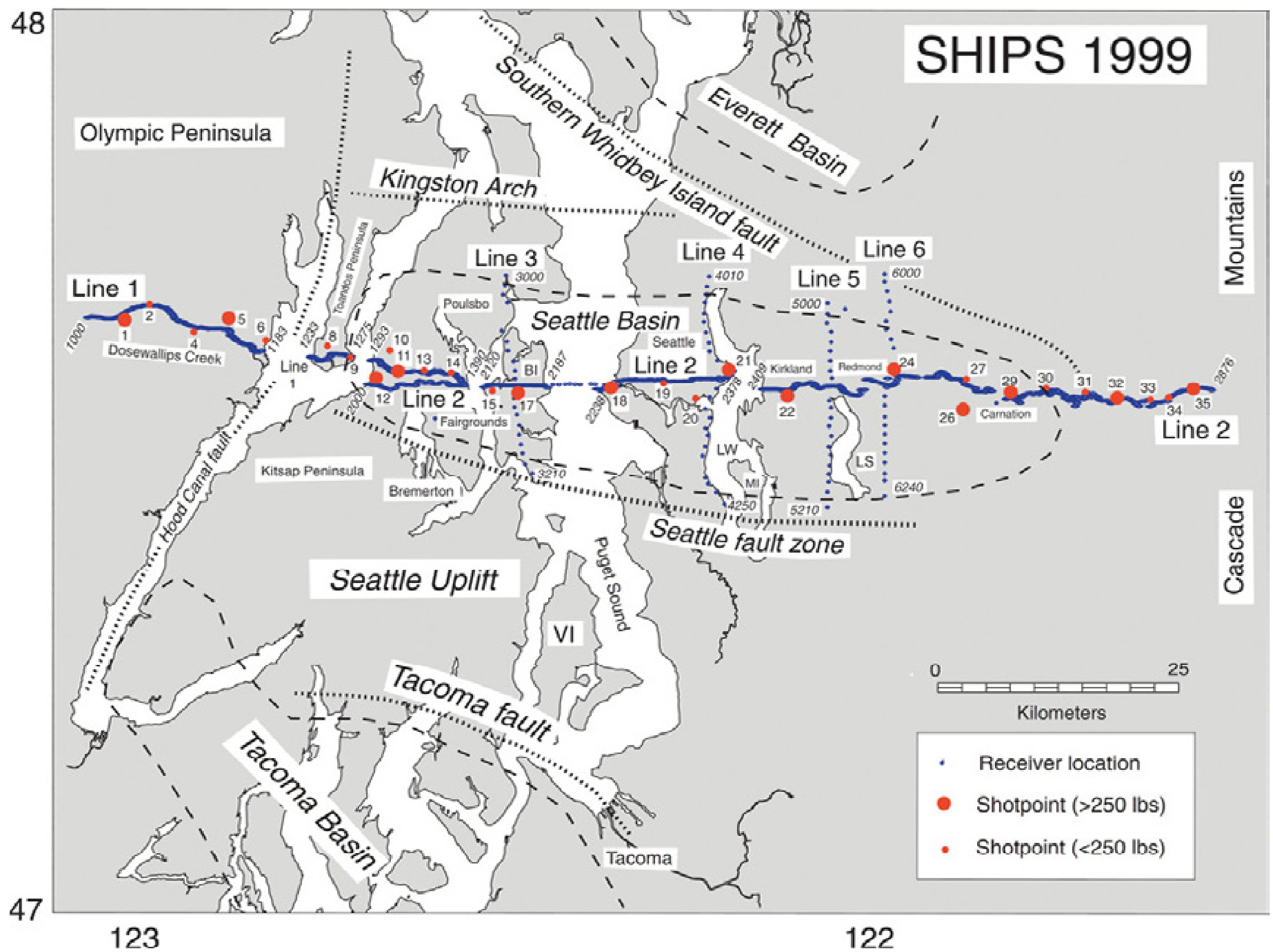



Figure 9.4.2-10. Location of Dry SHIPS 99 seismic shots and recorders in the Puget Lowland (from Brocher et al., 2000, fig. 1). [U.S. Geological Survey Open-File Report 00-318, 81 p.]

As a result of a 3-D P-wave tomography study of the data of the 1999 SHIPS experiment, the basin is ~70 km wide and contains sediments with velocities from 1.8–4.5 km/s. The basement rocks underneath the Seattle basin, characterized by a rapid increase in velocity from 4.5 to 5.0 km/s, are found at 6–7 km depth at the center of the seismic line, consistent with the 1998 N-S reflection line recorded along the Puget Sound (Snelson, 2001; Snelson et al., 2007).

The southernmost seismic project in the coastal area of Washington and Oregon was carried out in late 1994 in southern Oregon in the vicinity of Cape Blanco between 42°N and 43.5°N. It was a deep-crustal wide-angle seismic-reflection and -refraction experiment to study the lateral variations in crustal structure along the Cascadia margin (Brocher et al., 1995b; Appendix A9-2-2). The aim of the project was to image the lower crustal structure, observe significant crustal differences across a major E-W–trending shear zone near Cape Blanco, and to image the subducting Gorda and Juan de Fuca plates.

Two linear arrays of land stations were deployed along E-W transects across the Oregon Coast Range, each array having 10 three-component recording stations and stretching landward to 80 km from the coast (Fig. 9.4.2-11). Nearly 12,300 airgun shots were recorded by both arrays shot in-line along lines 5 (86 km long) and 6 (81 km long), while the shots of the other offshore lines were recorded onshore in a fan geometry. Airgun signals were recorded from 8 km to 160 km offsets on both deployments. The resulting data quality was extremely satisfying. Coherent arrivals could be traced as far as 170 km, $P_n$ arrivals, e.g., were visible over intervals of 30 km (Brocher et al., 1995b).

The marine data collection comprised 760 km of multichannel seismic-reflection profiles, collected on the continental margin of southern Oregon by recording airgun shots on a 4.2-km-long, 160-channel digital streamer. Seven profiles were obtained along strike-lines parallel and dip-lines perpendicular to the Oregon coast (Fig. 9.4.2-11).

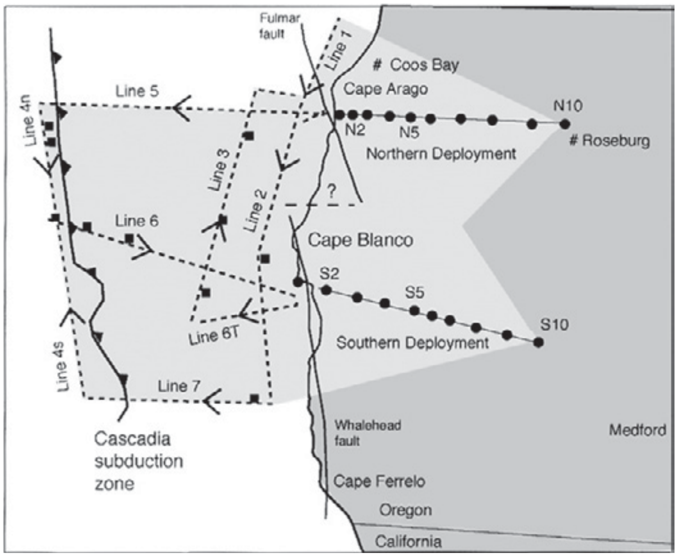



Figure 9.4.2-11. Location of onshore-offshore seismic wide-angle reflection lines near Cape Blanco, southern Oregon (from Brocher et al., 1995b, fig. 1). [U.S. Geological Survey Open-File Report 95-819, 69 p.]

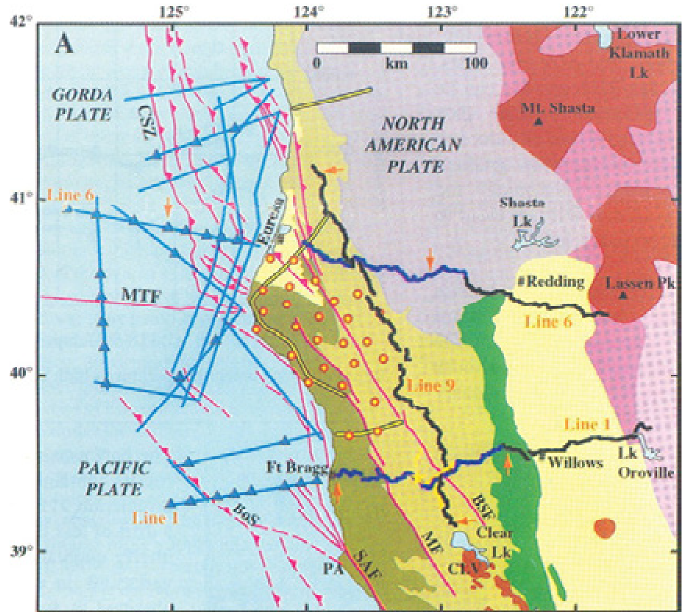



Figure 9.4.2-12. 1994 offshore multichannel seismic profiles (straight blue lines); 1994 OBS/OBH positions (triangles); and 1993 onshore profiles (crooked black and dark blue (reoccupied in 1994) lines) of the Mendocino Triple Junction experiment in northern California (from Trehu and Mendocino Working Group, 1995, fig. 1). Red lines: faults. [Eos (Transactions, American Geophysical Union), v. 76, no. 38, p. 369, 380–381. Reproduced by permission of American Geophysical Union.]

### *9.4.2.3. San Andreas Fault Region*

***9.4.2.3.1 Mendocino Triple Junction in northern California.*** The Mendocino Triple Junction in northern California has raised the particular interest of geoscientists, because here the active tectonic regime of northwestern California changes abruptly from transform motion along the San Andreas fault zone to subduction north of the Mendocino fault zone. Since Atwater (1970) proved the influence of plate tectonics on the geological history of western North America, it has become evident that three plates meet at the Mendocino Triple Junction (Fig. 9.4.2-12).

The North American and Pacific plates slide against each other onshore and offshore along the complex San Andreas fault system, while the Pacific and Gorda plates slide against each other offshore along the Mendocino transform fault. North of the Triple Junction, the Gorda plate is being subducted obliquely beneath the North American continent along the megathrust of the Cascade subduction zone (Trehu and Mendocino Working Group, 1995).

To study the crust and upper-mantle structure underneath this complex region of northern California and the adjacent continental margin, in 1993 and 1994 a network of large-aperture seismic profiles was arranged (Fig. 9.4.2-12). Approximately 650 km of onshore seismic-refraction and reflection data and 2000 km of offshore multichannel seismic (MCS) reflection data were collected during this survey. Furthermore, simultaneous onshore and offshore recordings of the MCS airgun shots were made (Godfrey et al., 1995; Appendix 9-2-3; Godfrey et al., 1998; Trehu and Mendocino Working Group, 1995).

The onshore survey of 1993 comprised two seismic lines of 200 km in length, where along each line 570 stations, spaced on average at 350 m, recorded 8 shots, spaced 25 km apart, with charges of 400–2000 kg (lines 1 and 6). The third profile (line 9) was 250 km long. Here 11 shotpoints were realized and the station separation was near 400 m. The offshore MCS data of 1994 were obtained by airgun shots, fired at 50 m intervals and recorded on a 4-km streamer up to 14 s TWT. The survey was complemented by OBS/OBH deployments at 34 seafloor sites and by the reoccupation of 477 onshore sites. In addition, a 1993 shotpoint near the intersection of lines 1 and 9 was revived (Godfrey et al., 1995, 1998; Trehu and Mendocino Working Group, 1995).

Interpretations of the individual data sets were published (e.g., Godfrey et al., 1998; Leitner et al., 1998; Henstock et al., 1997; Henstock and Levander, 2000). A cartoon of Leitner et al. (1998), interpreting MCS data in combination with gravity data, indicated a crustal thinning from 20 km near the coast to less than 11 km at 100 km distance from the mainland. Interpreting in particular the data of line 1 (Fig. 9.4.2-12), the southernmost offshore-onshore line, Henstock and Levander (2000) found that the top of the lower crust, a 5–10-km-thick high-velocity layer (6.4–7.2 km/s), deepened from 7 km beneath the Pacific Ocean basin at the west end to 23 km at the east end of line 1, being offset at the San Andreas and the Maacama faults (SAF and MF in Fig. 9.4.2-12) by up to 4 km, down to the east. The Moho was found to be similarly deformed, although by only 2 km.

***9.4.2.3.2. San Francisco Bay Area.*** While the San Andreas fault zone throughout most of northern California, with a few exceptions, runs offshore close to the coast, it enters the mainland in the area of the city of San Francisco. The San Francisco Bay Area is one of the most seismically active metropolitan areas in the United States. Here, the fault zone is spread over a broad, 100-km-wide zone, encompassing several major active faults, such as the San Gregorio, San Andreas, Hayward and other faults. Southwards, in Central California, the San Andreas fault juxtaposes the Cretaceous granitic Salinian terrane to the west and the Late Mesozoic/Early Tertiary Franciscan Complex to the east side (Fig. 9.4.2-13). On the San Francisco Peninsula,

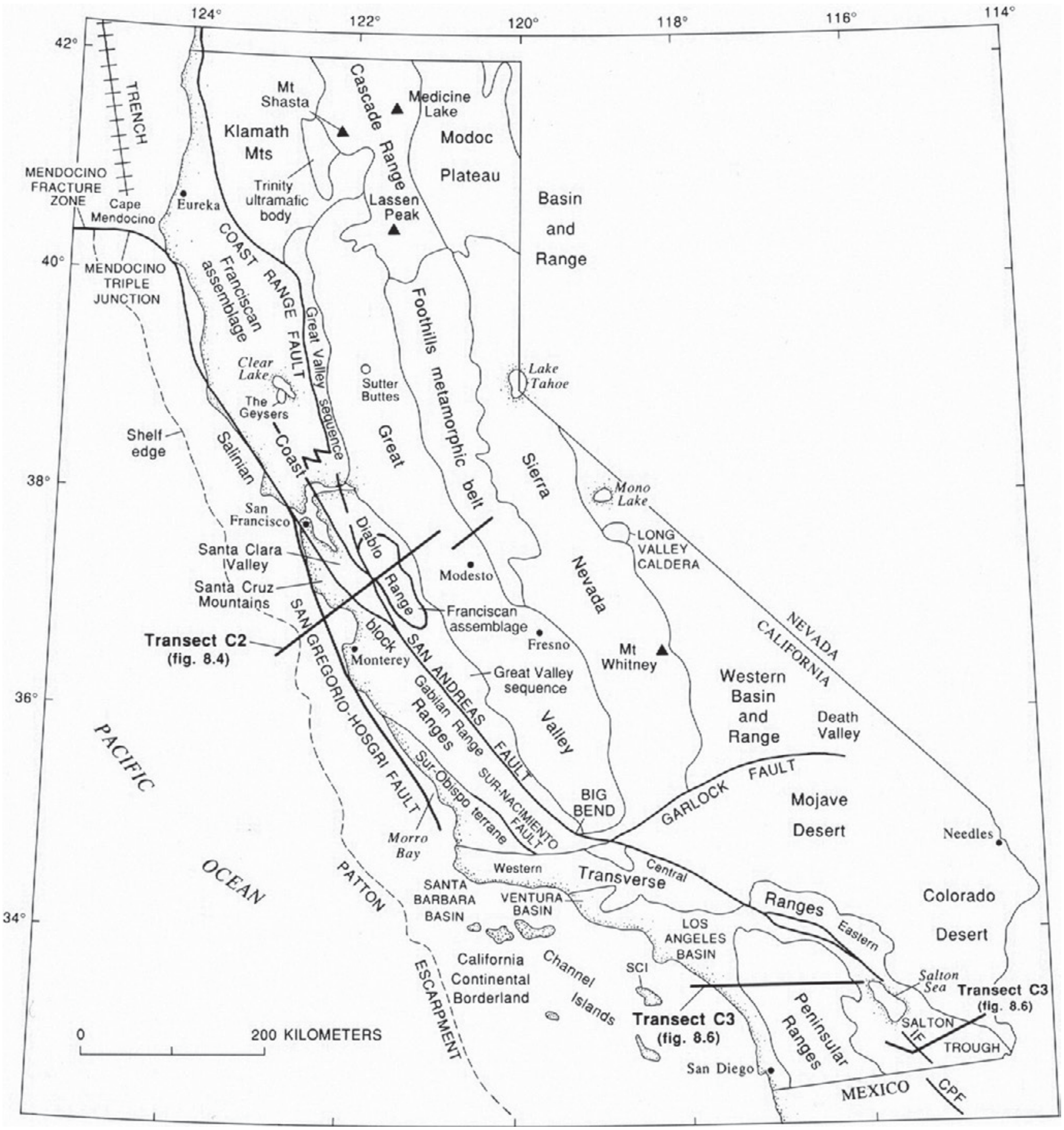



Figure 9.4.2-13. Map of California, showing place names, geologic provinces, selected geologic units, and locations of the crustal transects summarizing the deep structure of the crust known until 1990 (from Fuis and Mooney, 1990, fig. 8.2). [U.S. Geological Survey Professional Paper 1515, p. 207–236.]

however, the present-day San Andreas fault is completely within a Franciscan terrane, while the Salinian-Franciscan boundary is located southwest, marked by the subparallel Pilarcitos fault (Parsons and Zoback, 1997).

Due to the major earthquake hazard, the area contains one of the most densely occupied seismic networks in the world. However, until the early 1990s, little was known about the details of crustal structure which is important in understanding the tectonic processes that produce the seismicity and in accurately locating regional earthquakes. Sparsely sampled 1967 refraction lines south of the San Francisco Bay area (Walter and Mooney, 1982) and 1976 lines north of the Bay (Warren, 1981) provided first-order information on either side of the fault. Several tomography studies made use of the dense local network of the permanent Northern California Seismic Network (NCSN), which was started in 1966 and has since then been successively enlarged to its present-day status. However, none of the models focused on the entire Bay Area region. Fuis and Mooney (1990) have published two transects summarizing the deep structure of the crust along the San Andreas fault zone based on the knowledge available until 1989.

It was only in 1990, following the Loma Prieta magnitude 7.1 earthquake of 17 October 1989, that detailed crustal studies started. The first project was a marine seismic-reflection investigation of the central California margin and its prolongation on land. The marine seismic-reflection investigation of the central California margin was conducted along two 60-km-long parallel profiles running offshore from the San Francisco peninsula toward the SW. The airgun array of the offshore reflection lines was used as a sound source for recording with three-component seismic five-day recorders deployed onshore along two profiles, continuing the offshore lines on land (Brocher et al., 1992; Appendix A9-2-4).

The main recorder array consisted of 13 stations and stretched from the coast across the epicentral region of the Loma Prieta earthquake over a distance of 92 km (profile line identical with line FG in Fig. 9.4.2-14). The second array of six stations was located 25 km to the northeast over a distance of 70 km (Brocher et al., 1992, 2004a; Page and Brocher, 1993). On the basis of this data set, Brocher et al. (2004a) concluded that the hypocenter of the Loma Prieta earthquake was located at the base of a wedge of Salinian rocks beneath the continental margin, but above oceanic crust. The Loma Prieta line was reversed in 1993 (Brocher, 1994; Appendix A9-2-4) when 33 stations (line FG in Fig. 9.4.2-14) recorded some of the chemical explosions set off during the 1993 experiment which is described further below.

To obtain fundamental information on the crustal structure and fault geometries that underlie the Bay Area, in the fall of 1991 a major seismic-reflection investigation was conducted utilizing the marine waterway system that dissects the area (Figs. 9.4.2-15 and 9.4.2-16). The project BASIX (Bay Area Seismic Imaging Experiment) comprised three complementary seismic methods: high-resolution reflection profiling, wide-angle reflection/refraction profiling, and multichannel seismic profiling (McCarthy and Hart, 1993; Brocher et al., 1991b, 1994; Holbrook et al., 1996; Hole et al., 2000). The high-resolution shallow-penetration data were acquired to constrain near-surface faulting, while the wide-angle reflection/refraction data were essential to explore the deep-crustal velocity structure of the crust down to the Moho (McCarthy and Hart, 1993; Appendix A9-2-4).

The experiment involved an airgun array towed at an average depth of 7–8 m. For the multichannel seismic-reflection profiling the seismic energy was recorded by hydrophones installed at the edge of the dredged shipping channel and radioed back to a 120-channel recording system on the ship. Each day the hydrophones were shifted to new positions along the profiles, and each night the ship steamed back and forth along the hydrophone array while firing the airguns. Thus, the nominal shot interval was 50 m on the inland and Golden Gate profiles and 75 m on the offshore profile. The extensive marine waterway system that dissects the San Francisco Bay Area allowed recording of a total of 140 km of multichannel reflection data (track lines in Fig. 9.4.2-15), which was achieved by acquiring the data in 13 separate, overlapping segments, each ~25 km long. More than 15,000 separate airgun shots were recorded during BASIX.

Furthermore, wide-angle reflection profiles were recorded (Brocher and Moses, 1993; Appendix A9-2-4) from the airgun shots by land stations, recording the airgun shots of the 13 reflection lines (Fig. 9.4.2-16). The majority of these stations were deployed along the E-W–trending line from the Sacramento River to the Golden Gate, resulting in offsets from 1 km to 160 km. The NW-SE line consisted of a mixture of wide-angle in-line and fan recording, spanning a length of 49 km. In addition to the 13 reflection lines, the airguns were shot along two other lines solely for wide-angle recording: along the NW-SE–trending line along the San Francisco Bay with a shot interval of 50 m and along the E-W–trending line on the continental margin with a shot interval of 100 m. Here, five OBSs were successfully deployed for each line separately (Fig. 9.4.2-16). The source-receiver offsets along the offshore line ranged from 11 to 259 km. As a result, the base of the upper crust with seismic velocities from 5.5 to 6.0 km/s was marked by high-amplitude reflections at a TWT of 6 s, corresponding to a depth of 15 km. The upper crust showed no internal discontinuities and was interpreted as the Franciscan assemblage, which is exposed throughout the Bay Area. For the lower crust high seismic velocities between 6.5 and 7.2 km/s and a thickness between 8 and 11 km east of the San Andreas fault resulted (Brocher et al., 1994).

A subsequent interpretation (Holbrook et al., 1996) confirmed the thickness of the upper crust to be 15–18 km, but divided the lower crust into two layers with 6.4–6.6 and 6.9–7.3 km/s velocities (Fig. 9.4.2-17). The authors saw a thickening of the crust from 12 km at the southwestern end of line OBS-2-25 km under the Bay Area. They also indicated a similar offset of the top and the bottom of the lower crust at the San Gregorio and San Andreas faults to the NE (SGF and SAF in Fig. 9.4.2-17, bottom section), as was noted farther north near the Mendocino Triple Junction by Henstock and Levander (2000).

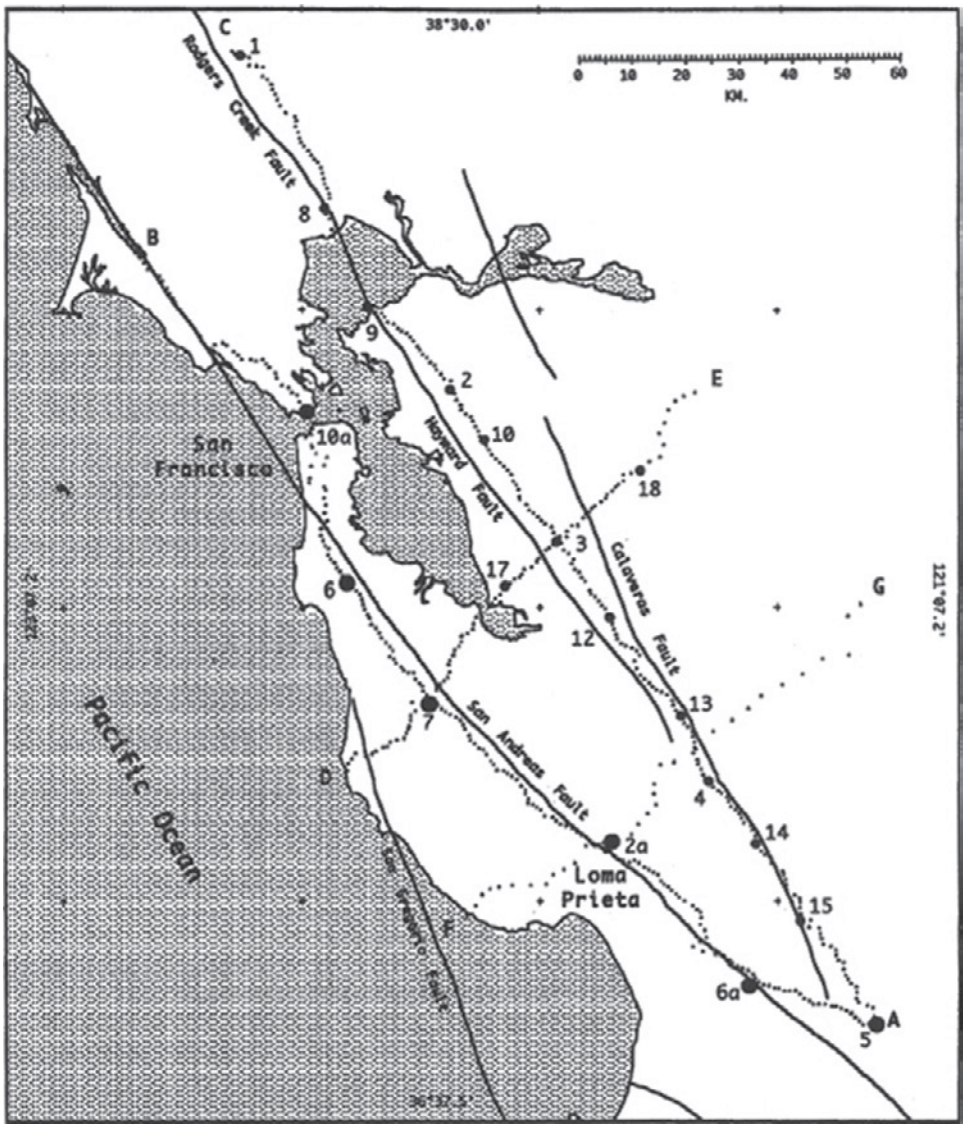



Figure 9.4.2-14. Map of the San Francisco Bay Area showing onshore seismic-refraction lines (from Kohler and Catchings, 1994, fig. 2). Line AB recorded along the San Andreas fault in 1991 (shotpoints 2a, 6a, 10a) and reshot in 1993; line AC recorded along the Hayward fault; line DE cross line; line FG Loma Prieta line reversing the 1991-airgun line-38 onshore-extension. [U.S. Geological Survey Open-File Report 94-241, 71 p.]

A tomography approach providing a 3-D model of the seismic velocity structure of the San Francisco Bay Area was applied by Hole et al. (2000), showing lateral variations of the velocity contrast from upper to lower crust which were correlated with surface faults, indicating changes in the depth of the mafic lower crustal layer. A comprehensive review was prepared by Parsons (2002) for seismic data obtained between 1991 and 1997, with the exception of the seismic-refraction survey of 1993, described further below.

In May 1991, a 180-km-long seismic-refraction line was recorded with 1 km spacing from five drillhole shotpoints (line AB in Fig. 9.4.2-14). The line started north of 38°N near Point Reyes, paralleled the San Andreas fault, passed through the San Francisco area (Murphy et al., 1992; Appendix A9-2-4) and ended south of 37°N ~40 km southeast of San Jose (Murphy et al., 1992). The line was extended by 20 km toward the south and re-shot in 1993 with the addition of a sixth shotpoint, when refraction profiling was carried out both along the Hayward and the San Andreas faults (Kohler and Catchings, 1994). The Moho was found at 30–32 km depth under the southeastern end of both lines and decreased toward the NW to 22 km under the city of San Francisco (Catchings and Kohler, 1996).

During this project, two more lines were recorded: the 220-km-long Hayward fault line (line AC in Fig. 9.4.2-14) and a 100-km-long cross line (line DE in Fig. 9.4.2-14) from the Pacific coast into the Diablo Range east of the Calaveras fault (Kohler and Catchings, 1994). The instrument spacing was 1.5 km. Along line AC, four OBSs provided recordings from the bottom of the San Pablo Bay. The 1993 experiment comprised a total of 17 shots. In the first deployment, lines AB and DE were occupied and five shots fired at shotpoints 5, 6, 7, 17, and 18. In the second deployment, lines AC and CD were occupied and 12

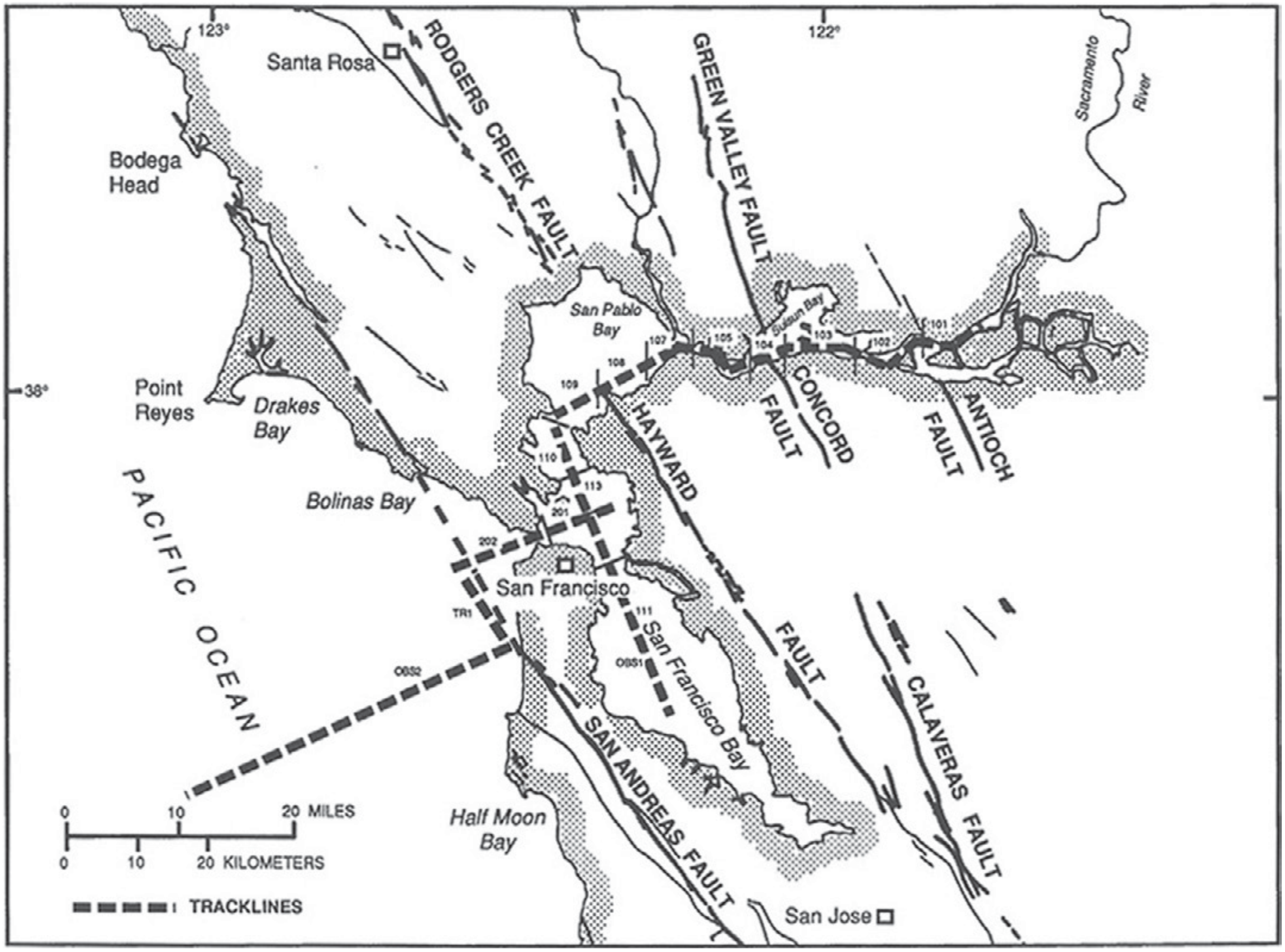



Figure 9.4.2-15. The Bay Area Seismic Imaging Experiment of 1991 (from McCarthy et al., 1993, fig. 4). Track lines of the multichannel survey. [U.S. Geological Survey Open-File Report 93-301, 26 p.]

shots at all remaining shotpoints were fired. A comprehensive interpretation of these data has not been published, but Boatwright et al. (2004) have collected the first-arrival traveltimes and geometry data and calculated tomographic velocity models for each line.

In 1995, 183 stations on the San Francisco peninsula recorded seven in-line and four fan shots along a SW-NE–trending linear array of 9 km length with an instrument spacing of 50 m. Reflections up to 14 s TWT could be correlated. Here the Moho was inferred at the lower end of middle and lower crustal reflectivity at 9s TWT, which was converted into 27–28 km depth (Parsons, 1998).

***9.4.2.3.3 Central California.*** The crustal investigation of the surroundings of the San Andreas fault was shifted southwards in 1995, when an array of 48 seismic instruments, which had been installed already by the end of 1994 to record local earthquakes for half a year, was used for an active-source experiment (Fig. 9.4.2-18). Thirteen shots of 200–300 kg charge size were fired into the array and additional stations, forming three profiles that paralleled and crossed the San Andreas fault (Thurber et al., 1996). An important result of this project was the confirmation that there was a sharp lateral velocity contrast at the surface trace of the San Andreas fault and that the San Andreas fault bounds the southwestern edge of a 6–7-km-wide trough of low P-wave velocity that extends to 3–4 km depth.

The most recent study of the San Andreas fault zone finally targeted the Parkfield area, halfway between the densely populated areas of San Francisco and Los Angeles. A series of magnitude-6 events with very similar characteristics (Brown et al., 1967) had led to a detailed geophysical investigation of this area (Bakun and Lindt, 1985) which should finally lead to the deep-drilling project SAFOD (San-Andreas-Fault-Observatory-at-Depth; Hickman et al., 1994). The main activities, the realization of a deep drillhole and a major seismic-refraction/-reflection project, however, were only started after 2000 and will therefore be described in Chapter 10.

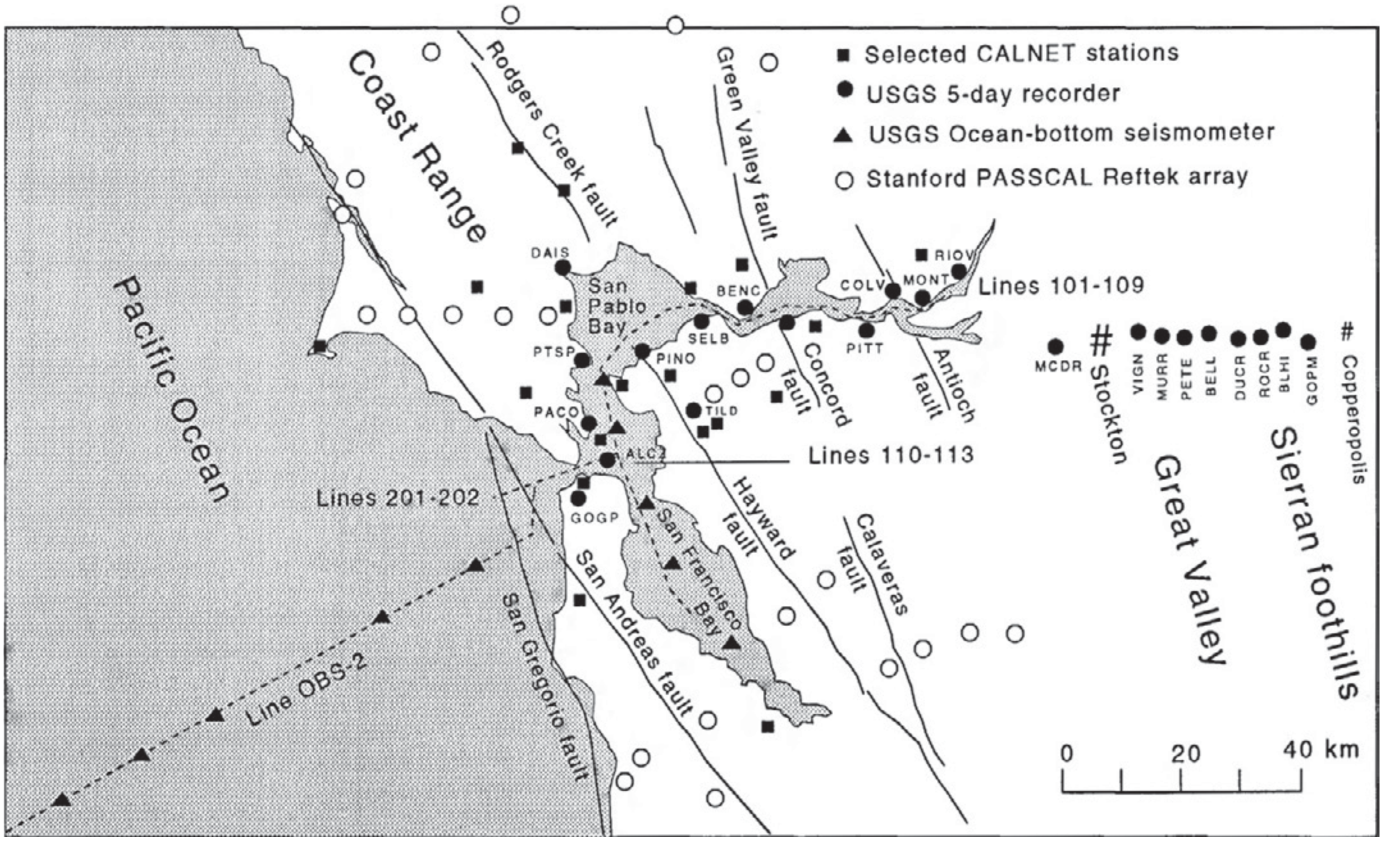



Figure 9.4.2-16. The Bay Area Seismic Imaging Experiment of 1991 (from Brocher and Moses, 1993, fig. 1): Basic seismic reflection lines (dashed lines) and location of wide-angle recorders. [U.S. Geological Survey Open-File Report 93-276, 89 p.]

***9.4.2.3.4. Southern California.*** Similar to the San Francisco Bay area in central California, also in the densely populated area of Los Angeles–San Diego in southern California, the strong earthquake activity in the immediate neighborhood of the active San Andreas fault system causes a severe threat. A dense network of permanent seismic stations is used to analyze the statistics of the earthquake activity, and major events are being systematically studied (e.g., Ellsworth, 1990). The lithospheric cross section through southern California by Fuis and Mooney (1990) indicates, however, how little was known about the crust and upper mantle up to 1990. The only detailed crustal seismic investigation had been carried out in the Imperial Valley in 1979 south of the Salton Sea (Fuis et al., 1984), described in Chapter 7.4.2.

Following several major earthquakes in the Los Angeles area which were associated with buried faults, the need for detailed knowledge of subsurface structures became obvious and the project LARSE (Los Angeles Region Seismic Experiment) was born with the aim to obtain a very detailed structural crustal model for the Greater Los Angeles Area along selected traverses reaching from the coastal regime across its basins and mountain ranges into the Mojave desert. The planning of LARSE-I concentrated mainly on line 1 (Fig. 9.4.2-19), but later on, following the Northridge event of 1991, in 1999 a detailed land survey (LARSE-II) was added on line 2 (Fig. 9.4.2-19).

LARSE-I started in 1993 with a passive study along line 1, with the goal to collect seismic waveform data from local and teleseismic earthquakes to refine 3-D images of the lower crust and upper mantle in southern California. In October 1994 an active seismic source experiment was conducted. It consisted of an onshore part along line 1 and an onshore-offshore survey. The onshore part, which followed the marine airgun shooting survey, concentrated on line 1 which started at Seal Beach, crossed the Los Angeles Basin and the San Gabriel Mountains and ended in the Mojave Desert (Fuis et al., 1996). Sixty explosions, with charges ranging from 5 kg to 3000 kg, were recorded by a linear array of 640 seismographs, with the majority of drillhole shots being set off in the San Gabriel Mountains (Fig. 9.4.2-19).

During LARSE-I, in October 1994, also deep-crustal multichannel seismic-reflection data were recorded offshore (Brocher et al., 1995c; ten Brink et al., 1996; Appendix A9-2-5). Three onshore-offshore wide-angle reflection lines, all centered on the Los Angeles basin, were recorded, each 200–250 km long in total, with the offshore portions being between 90 and 150 km long (Fig. 9.4.2-20). Using an airgun array as the seismic energy source, nearly 24,000 airgun shots were done. Wide-angle data

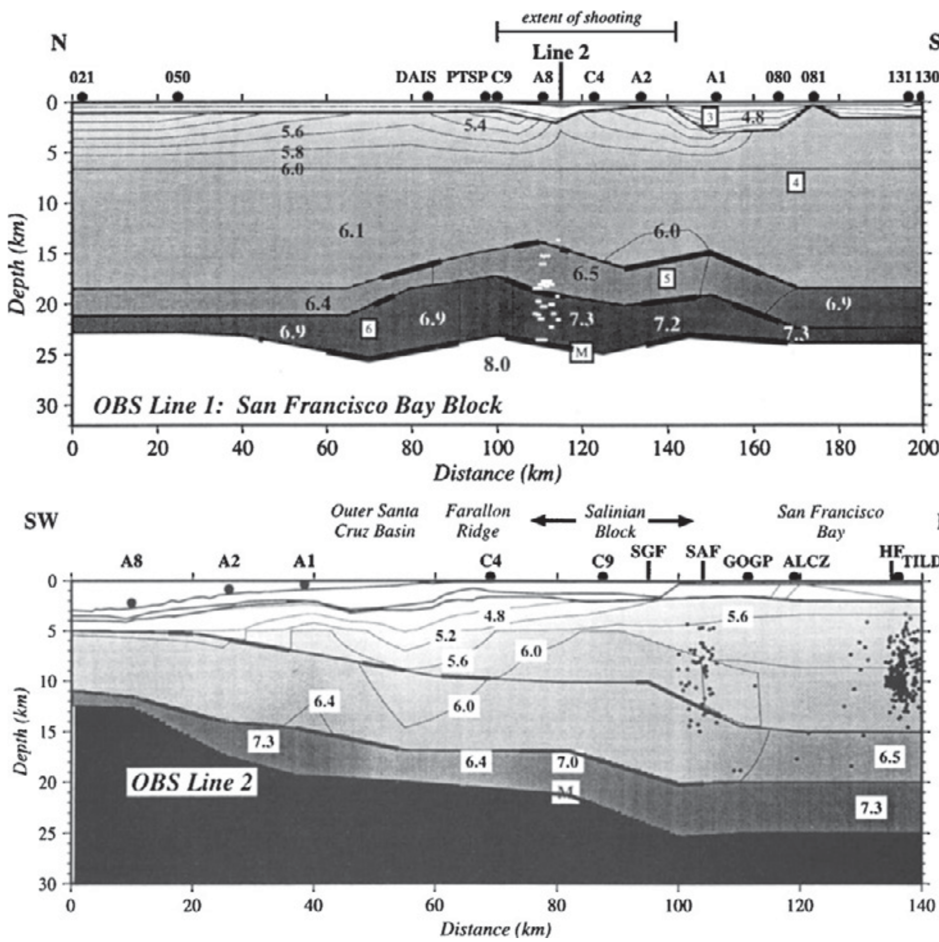



Figure 9.4.2-17. Crustal and upper mantle velocity cross sections across the San Francisco Bay Area (from Holbrook et al., 1996, figs. 2 and 16). Top: N-S section through the San Francisco Bay area. Bottom: SW-NE section across the continental margin to the Hayward fault (HF) east of the San Francisco Bay. [Journal of Geophysical Research, v. 101, p. 22,311–22,334. Reproduced by permission of American Geophysical Union.]

were recorded by 170 land stations deployed at 2 km intervals along all three profiles, and at sea by two sonobuoys and by 10 OBSs, deployed along lines 1 and 2 (details in ten Brink et al., 1996; Appendix A9-2-5). Three airgun lines TR1 to TR3 connected the main lines 1, 2, and 3 and were recorded by the onshore stations in an oblique fan geometry. Three more lines, 4–6, were solely multichannel seismic-reflection offshore lines. All multichannel reflection data were recorded by a 160-channel, 4.2 km-long digital streamer (Brocher et al., 1995c; Appendix A9-2-5). Airgun signals were visible at OBSs as far as 70 km, and in some cases also up to 140 km. At onshore sites the airgun energy carried over distances of 200 km and was even visible in the Mojave Desert.

In October 1999, LARSE-II was conducted along line 2 (Fuis et al., 2001a, 2001b; Appendix A9-2-5, Murphy et al., 2002). As on line 1, prior to the active-source experiment local and distant earthquakes were recorded along line 2 for a period of 6.5 months. The active-source survey complemented a 1994 onshore-offshore (airgun) survey along the same corridor. Chief imaging targets included the Santa Monica, San Gabriel, and San Andreas faults, blind thrust faults (including the Northridge fault), and the depths and shapes of the sedimentary basins in the San Fernando Valley and Santa Monica areas.

Ninety-three borehole explosions were detonated along the main line 2, which was 150 km long and extended from Santa Monica Bay northward into the western Mojave Desert, crossing the Santa Monica Mountains, the San Fernando Valley (through the Northridge epicentral area), the Santa Susana Mountains, and the western Transverse Ranges (Fig. 9.4.2-19), and along 5 auxiliary lines in the San Fernando Valley and Santa Monica areas, ranging in length from 11 to 22 km. The explosions were recorded by ~1400 seismographs, simultaneously on all lines. The goal of 1 km shotpoint spacing and 100 m geophone spacing for the main line was nearly achieved for the first 79 km from the coast to the southern Mojave Desert. Farther north, shotpoint spacing averaged 2.75 km from 79 to 101 km, and the remaining shotpoint was located at 136 km, while station spacing increased stepwise from 90 km to the end of the line at 150 km. The shot size, as far as open space was available, was planned to be a mix from small (near 100 kg) to large (~450 kg), in order to investigate reflective features with differing frequency return. However, in the densely populated areas the charge sizes depended on the location of the shotpoints and were determined from ground-shaking data obtained during LARSE-I and from calibration shots (Fuis et al., 2001a, 2001b).

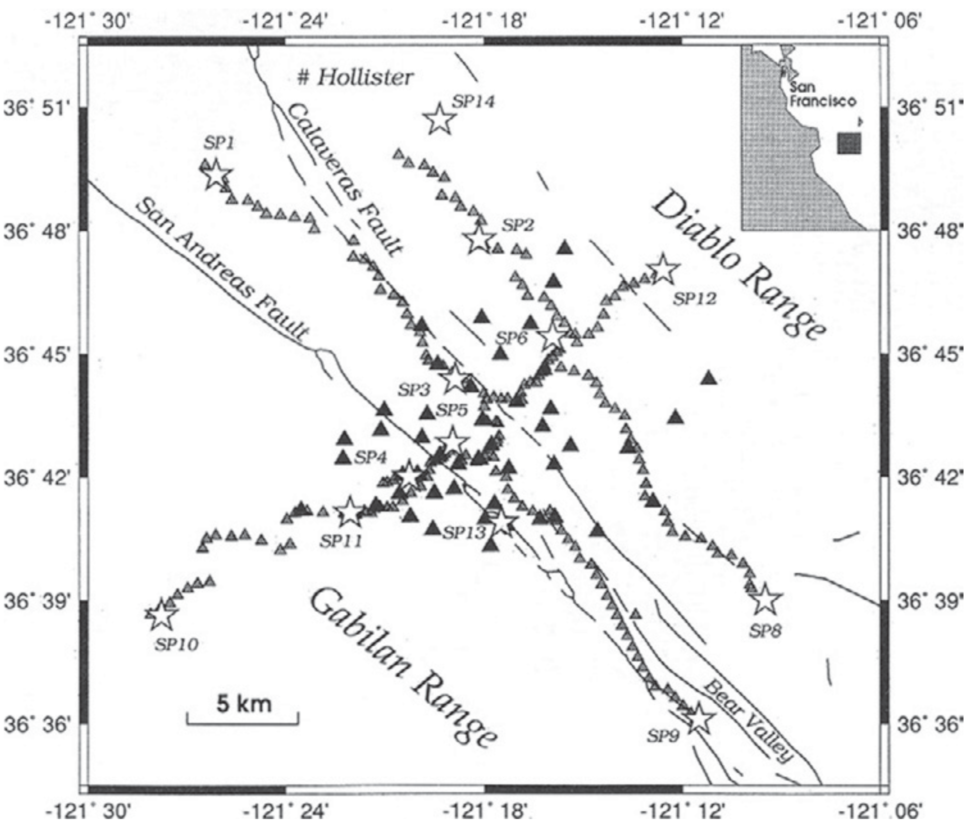



Figure 9.4.2-18. Location of the 1995 seismic experiment (from Thurber et al., 1996, fig. 2). Passive array (large triangles) and in-line active-source recording stations (small triangles) and shotpoints (stars). [Eos (Transactions, American Geophysical Union), v. 77, p. 45, 57–58. Reproduced by permission of American Geophysical Union.]

LARSE-I imaged the bottom of the Los Angeles basin for the first time (Fuis et al., 1996, 2001b). A velocity model (Lutter et al., 1999) combined with the line drawing of CMP (crustal mid-point) data by Ryberg and Fuis (1998) and a subsequently developed schematic block diagram of Fuis et al. (2001b) illustrate the tectonics which could be interpreted from the seismic results of the LARSE-I project (Fig. 9.4.2-21). One important first result was the discovery that areas of high seismicity correlated well with areas of increased velocities (Fuis et al. (1996). Beneath the San Gabriel Mountains, bright reflective zones (A and B in Fig. 9.4.2-21) were discovered along line 1 in the middle crust at ~20 km depth that appeared to connect with the San Andreas fault in the north and with compressional faults in the Los Angeles basin to the south (Ryberg and Fuis, 1998).

Furthermore, the explosion seismic data showed a dip of the Moho from 30 km under the Mojave Desert to a maximum of 38 km under the San Gabriel Mountains (bottom of the reflective lower crust in Fig. 9.4.2-21). Toward the south the data could not be continuously correlated.

From teleseismic observations a root zone thickened by 15 km under the San Gabriel Mountains was deduced (Kohler and Davis, 1997). Hauksson and Haase (1997) have obtained velocity-depth cross sections to 20 km depth by a 3-D inversion using all available explosion seismic and earthquake data. This model, however, was detailed enough only along line 1 where from the explosion seismic data the traveltimes between sources and receivers were sufficiently accurately known. Here an intrusion of basement rocks into the sediments of the basin was interpreted as a likely origin of the Whittier thrust fault (also seen in Fig. 9.4.2-21).

Tomographic modeling of the first arrival data of line 2 (Lutter et al., 2004) produced a similar model for line 2 as for line 1 (Lutter et al., 1999). Along line 2 prominent zones of dipping reflectors, which were superimposed onto the tomographic model, could be correlated. The most prominent feature is a north-dipping reflective zone which extends downward from near the hypocenter of the 1971 San Fernando earthquake at ~13 km depth (brown circle in Fig. 9.4.2-22) to ~30 km depth in the vicinity of the San Andreas fault in the central Transverse Ranges. A fainter south-dipping reflective zone extends from the 1994 Northridge hypocenter at ~16 km depth (blue circle in Fig. 9.4.2-22) to 20 km depth beneath the Santa Monica Mountains. These reflective bands could be correlated with aftershock series of the 1971 San Fernando and the 1994 Northridge earthquakes (brown and blue dots, respectively, in Fig. 9.4.2-22), which were projected from a zone 10 km east of line 2 (Fuis et al., 2001b, 2003).

#### *9.4.2.4. Sierra Nevada and Basin and Range Province*

In summer 1993, the Southern Sierra Nevada Continental Dynamics (SSCD) project recorded two seismic-refraction lines in central California, which addressed the Sierra Nevada root question (Wernicke et al., 1996; Ruppert et al., 1998; Fliedner et al., 2000). The first line was 400 km long and ran in a W-E


Figure 9.4.2-19. Shaded relief map of the Los Angeles region, southern California, showing location of the LARSE active source seismic profiles (from Fuis et al., 2003, fig. 1). Small circles—shotpoints; colored lines—seismic recording stations; yellow dashed lines—airgun pulses; large red circles—epicenters of large earthquakes (with focal mechanisms and magnitudes). [Geology, v. 31, p. 171–17. Reproduced by permission of the Geological Society of America.]

direction from the Coast Ranges across the Sierra Nevada into Death Valley, in the western Basin and Range province. The second line was 325 km long and ran in a N-S direction along Owens Valley east of the Sierra Nevada to the south across the Garlock fault (Fig. 9.4.2-23). A third line, unreversed, spanned 450 km from Beatty, Nevada, across the Sierra Nevada into the Coast Ranges and ran along the first line, recording an explosion of 2 million pounds of chemicals, the Department of Energy's Nuclear Non-Proliferation Experiment (NPE). Up to 675 seismic recorders had been installed for each seismic profile, with a nominal spacing of 500 m along the first two lines and 1500 m along the third line. Shot sizes ranged from 3600 kg for the end shots of the first two lines to 1360 kg at the profiles' intersection at Owens Lake. Near Death Valley, some additional 680-kg shots were added for increased resolution of upper-crustal features (Wernicke et al., 1996).

A small crustal root (40–42 km thick) centered ~80 km west of the Sierran topographic crest, low average $P_n$ velocities of 7.6–7.9 km/s under a laminated transitional Moho under the Sierra Nevada batholith, and slightly higher $P_n$ velocities of 7.9–8.0 km/s under a sharp first-order Moho under the Basin and Range province were the main results of a 3-D interpretation (Ruppert et al., 1998). Including earlier refraction and earthquake data, a subsequent interpretation revealed a 3-D velocity model of the crust (Fliedner et al., 2000). Crustal thickness increased from less than 25 km at the coast to 35–42 km in the western Sierra Nevada and decreased to 30–35 km in the Basin and Range province. The authors also saw a systematic thinning of the crust in the Basin and Range province from north to south in concordance with a decrease in elevation, and both the Basin and Range province and the Sierra Nevada batholith showed average crustal velocities of 6.0–6.2 km/s well below the worldwide continental average of 6.45 km/s.

In the Basin and Range Province of Nevada the USGS had first conducted seismic-refraction studies in 1980–1982 (Hoffman and Mooney, 1983) to aid an investigation of the regional crustal structure at a possible nuclear waste repository site near Yucca Mountain. In 1988, another test experiment had been carried out south of Yucca Mountain (Brocher et al., 1990, 1993b).

On the basis of the previous studies, in 1994 a detailed seismic-reflection survey along two lines of 11 and 26 km length followed, using hybrid seismic sources including vibrators, surface-mounted explosive charges, and conventional shallow minihole and deep shot holes for data collection. Besides a detailed image of the upper crust, the interpretation revealed a reflective lower crust between 15 km depth and the Moho between 27 and 30 km depth along the longer profile and excluded the existence of a lower-crustal magma chamber (Brocher et al., 1996; Appendix A9-2-6).

Another seismic-reflection survey, combining normal-incidence and wide-angle reflection data, was arranged by a joint project of the Universities of Arizona, Wyoming, and Georgia, and concentrated on the Ruby Mountains metamorphic core complex southeast of Elko, Nevada (Satarugsa and Johnson, 1998). The focus was to prove or disprove the statement that the Ruby Mountains had a flat Moho as had been postulated by the COCORP 40° transect (Allmendinger et al., 1987; Hauser et al., 1987b) which crossed the Ruby Mountains at its southern end, where topography is relatively low. The fieldwork was carried out in 1992 and 1993. The 1992 survey consisted of a 90-km-long N-S–trending line along the eastern flank of the mountains and three 3–9-km-long cross lines. The southern end of the long line coincided approximately with location 1000 of the COCORP 40° transect (see Fig. 8.5.3-02, eastern half, profile 4 at position 1000). Twelve shots had charges from 250 to 1080 kg and an average shot spacing of 20 km, and the in-line receiver spacing varied between 70 and 100 m with one-,

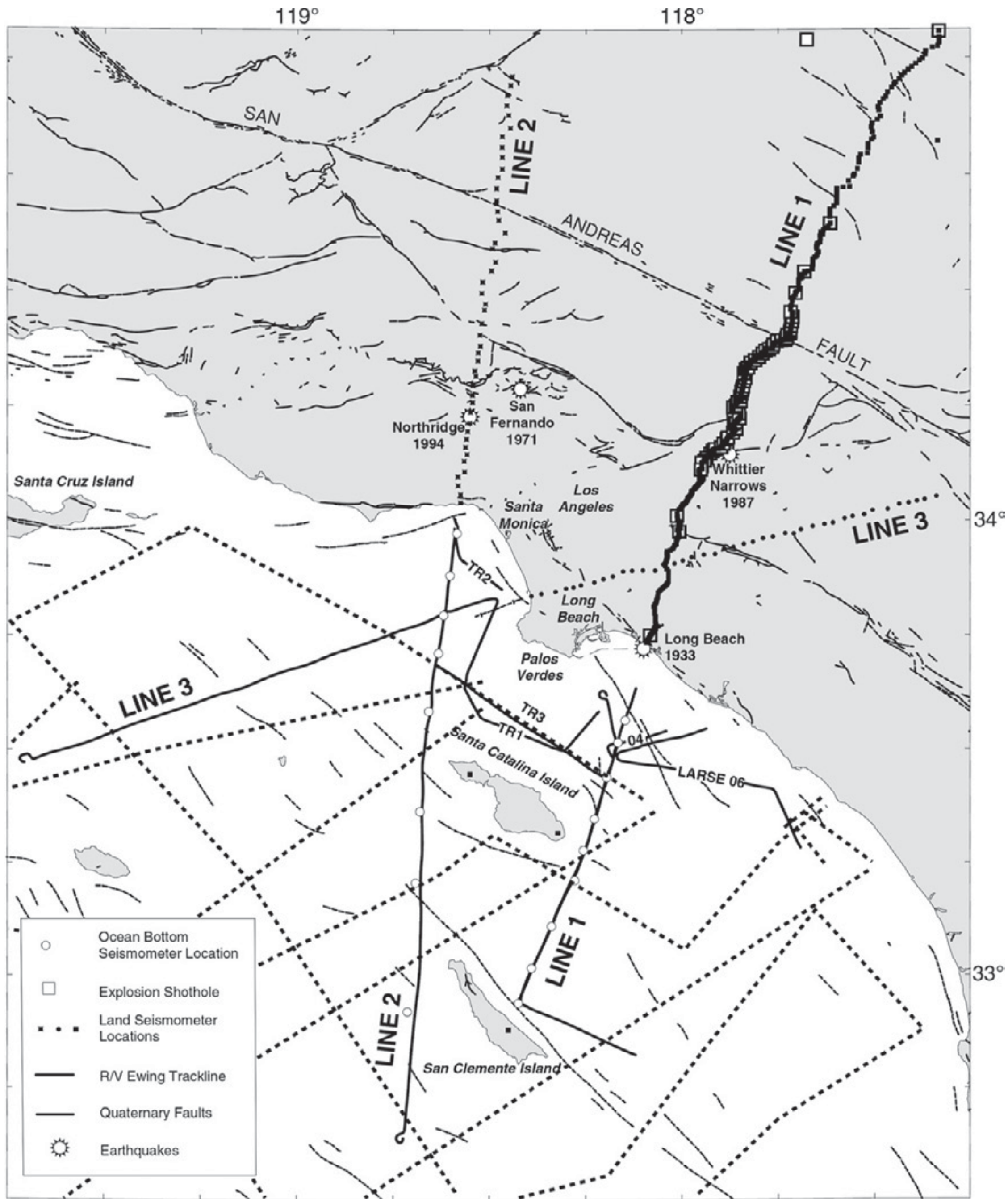
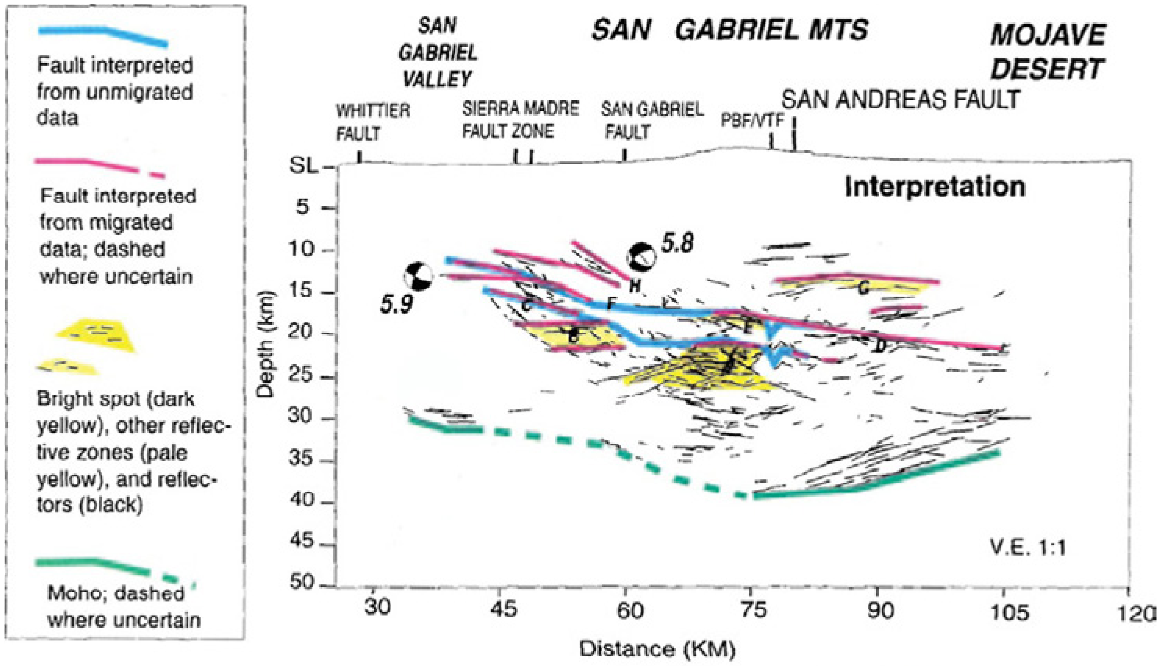



Figure 9.4.2-20. Offshore-onshore recording lines of LARSE-I in the Los Angeles region, southern California (from Brocher et al., 1995c, fig. 1), showing locations of the LARSE airgun lines and locations of recording stations (OBS and land stations). Also shotpoints of the LARSE-I land phase along line 1 are shown. [U.S. Geological Survey Open-File Report 94-241, 71 p.]

two-, and three-component geophones totaling 1622 recording channels. The line was divided into two 45-km-long segments which recorded the shots along the corresponding segment. The 1993 survey consisted of a 60-km-long in-line profile and two 8–9-km-long cross profiles which recorded 5 shots and provided infill subsurface coverage.

The data revealed that the depth to Moho along the eastern edge of the Ruby Mountains varied irregularly between 30.5 and 33.5 km and thus did not reflect local relief, concluding that isostatic balance is achieved by modifications of the crust during extension by intracrustal processes but not by a local crustal root (Satarugsa and Johnson, 1998, 2000).



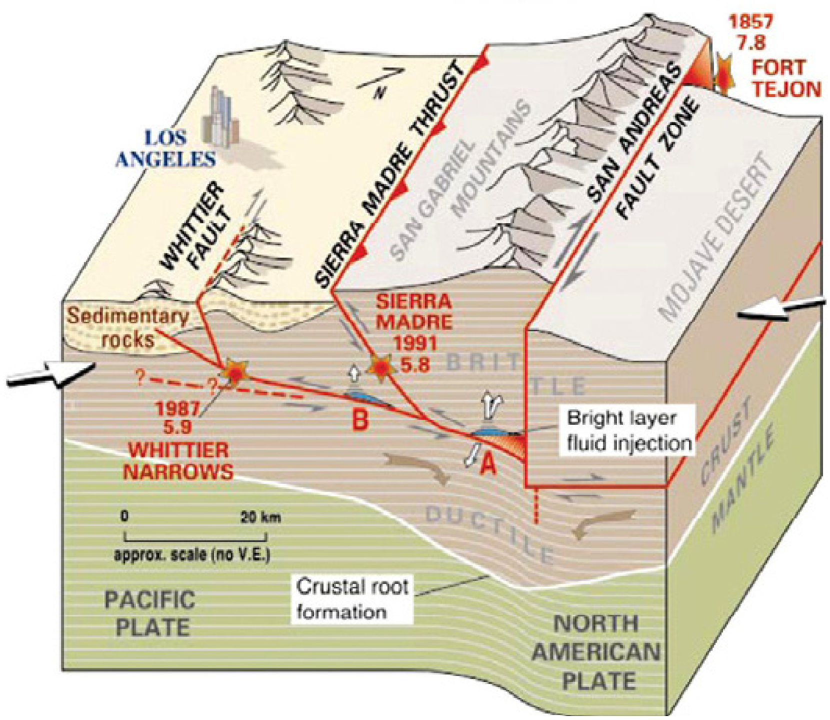




Figure 9.4.2-21. Top: Crustal structure based on line drawings after migration of common midpoint data along LARSE line 1 (from Ryberg and Fuis, 1998, fig. 8). Moderate earthquakes were projected into the cross section. A and B are bright reflectors. Bottom: Schematic block diagram (from Fuis et al., 2001b, fig. 3), showing interpreted tectonics in the vicinity of the location of the LARSE line 1. [Tectonophysics, v. 286, p. 31–46. Copyright Elsevier. Geology, v. 29, p. 15–18. Reproduced by permission of the Geological Society of America.]

The 1993 Delta Force seismic experiment aimed to develop a regional lithospheric structure model of southern Nevada, southwestern California, and western Arizona portions of the Basin and Range province and its transition into the Colorado Plateau. The project was designed as a cost effective, regional seismic experiment by taking advantage of the previous results from higher resolution seismic experiments such as the SSCD (Southern Sierra Continental Dynamics) and PACE (Pacific to Arizona Crustal Experiment) carried out in 1985, 1987, 1989, and 1992 (Hicks, 2001). The Delta Force experiment employed 4 seismic sources and 474 digital recording instruments laid out in a series of three intersecting recording lines, providing in-line and fan-type coverage of the study area.

### *9.4.2.5. The Rocky Mountain Region*

In spite of great geologic interest and abundant natural resources, relatively little was known about the crust and upper mantle structure of the Rocky Mountains until in the 1990s, when several major seismic experiments were started. Summaries of early surveys, all prior to 1980, were published by Keller et al. (1998), Prodehl and Lipman (1989), Smithson and Johnson (1989), and Prodehl et al. (2005). It was only in the 1990s, that three major seismic projects were undertaken: the Jemez Tomography Experiment dealing with the crustal structure of the Jemez Mountains in New Mexico (Baldridge et al., 1997); the Deep Probe experiment, with a major crustal component in Canada as described above, and dealing exclusively with the upper-mantle structure of the Rocky Mountain area in the United States (Deep Probe Working Group, 1998; Snelson et al., 1998; Henstock et al., 1998); and the CD-ROM experiment in 1999, a 1000-km-long lithospheric transect around 106°W, covering the Southern Rocky Mountains from Wyoming to New Mexico (Keller et al., 1999; Karlstrom and Keller, 2005).

The Jemez Tomography Experiment targeted the crustal structure beneath a large continental silicic magmatic system in the Jemez Mountains of New Mexico (Baldridge et al., 1997). The project centered upon the Valles Caldera, which at the same time was being geothermally explored, and followed an experiment conducted in 1981 (Olsen et al., 1986; Ankeny et al., 1986) which had elucidated upper crustal structure beneath the caldera providing evidence of a low-velocity body presumed to correlate with the so-called Bandelier magma chamber. The seismic component involved both active and passive sources and aimed to constrain the structure and composition of the lower crust and

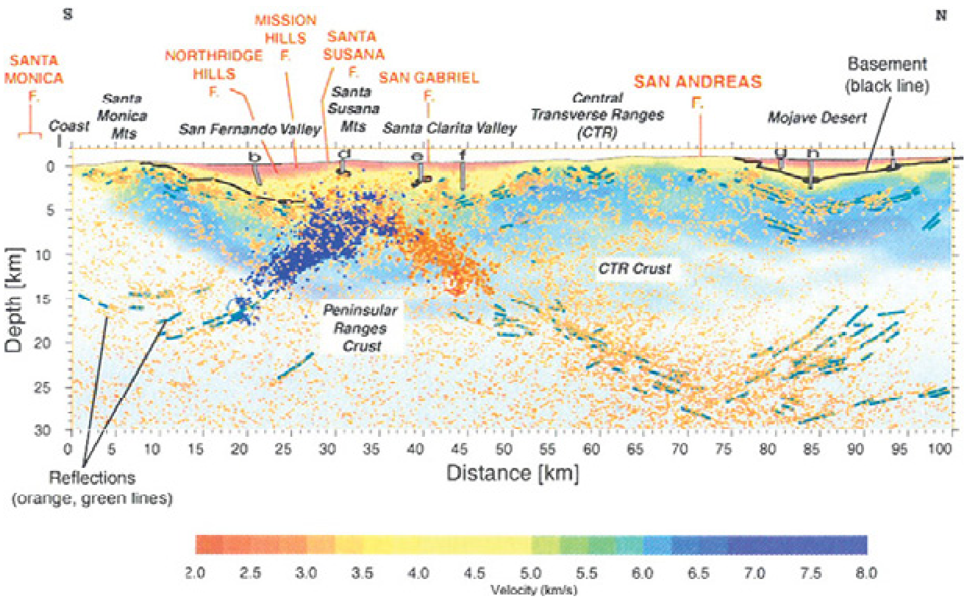



Figure 9.4.2-22. Tomographic velocity model and superimposed line drawings of CMP data along LARSE line 2 (from Ryberg and Fuis, 2004, written commun.). The 1971 San Fernando and 1994 Northridge main shocks (big circles) and relocated aftershocks (brown and blue dots, respectively) were projected into the cross section. [internal report, compiled from Fuis et al., 2001b, Tectonophysics, v. 286, p. 31–46. Copyright Elsevier. Geology, v. 29, p. 15–18. Reproduced by permission of the Geological Society of America.]

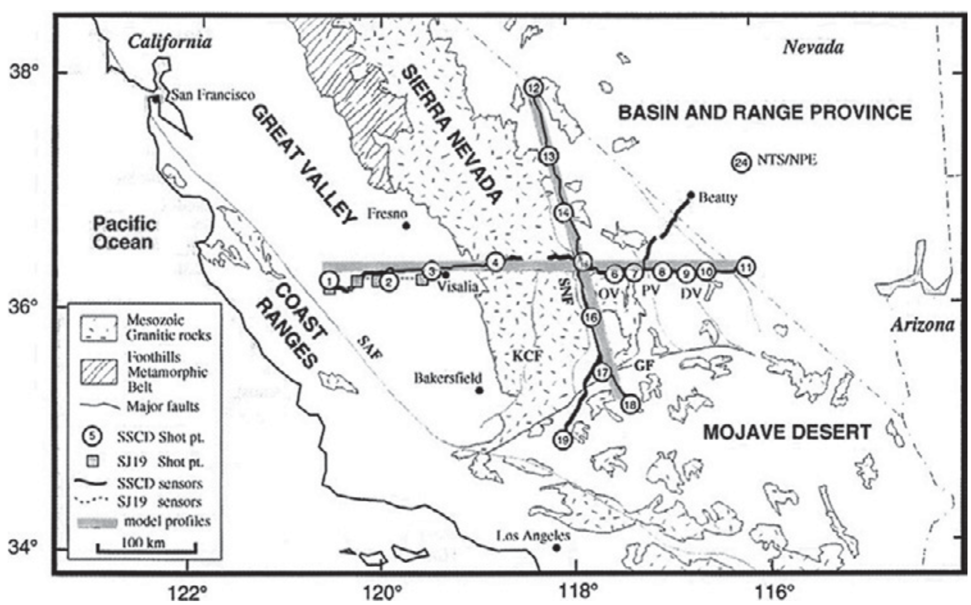



Figure 9.4.2-23. Location map of seismic profiles through the Sierra Nevada, recorded by SSCD (from Ruppert et al., 1998, fig. 1). NTS/NPE—Department of Energy's Nuclear Non-Proliferation experiment. [Tectonophysics, v. 286, p. 237–252. Copyright Elsevier.]

upper mantle, to delineate the geometry and physical state of the magma chamber, and to investigate the geometry of the caldera structure and its fill.

Though a comparatively modest experiment, the active component had to deal with unusual public oppositions and social issues in permitting which were described in much detail by Baldridge et al. (1997), before the three lines which crossed each other in the Valles caldera could be realized (Fig. 9.4.2-24).

The first line could be recorded in 1993, with the compromise that shotpoints in the caldera were dropped and replaced by Vibroseis trucks, while the other two lines, because of the severe permitting problems, had to be delayed until 1995. The three profiles finally realized were three nearly in-line arrays, 150–180 km long and incorporated 300 seismometers. The lines extended from the Colorado Plateau across the Rio Grande rift and Southern Rocky Mountains into the Great Plains.

The first line in 1993 was 180 km long and comprised 150 three-component and 135 one-component recorders with a nominal spacing of 300 m in the middle of the line and 1600 m near the ends. The simulation of the inner shotpoints was realized by three 20,000 kg truck-mounted vibrators with sweeps ranging from 5 to 30 Hz over 30-s intervals and repeated every two minutes, so that up to 150 sweeps could be stacked later on. The second phase in 1995 added the other two lines with similar instrument spacing, but with the addition of 23 seismographs placed in a 3-D array within the caldera and left in place for both lines. Again vibration points were added in the caldera to compensate for the loss of the originally planned shots.

The experiment confirmed the existence of a strong 12-km-thick and 10-km-wide mid-crustal low-velocity zone, interpreted as a magma chamber coincident with the above mentioned Bandelier magma chamber. The data also evidenced a deeper low-velocity zone extending from the lower crust into the uppermost mantle, which was broadest and strongest near the assumed Moho depth of 35 km.

The passive component extended over two years. It started also in 1993 with 22 instruments, placed along two profiles with a 3 km spacing, and recording for 3 months. A second phase in 1994 recorded for 2 months with a larger array of 49 stations deployed in and adjacent to the caldera (Lutter et al., 1995).

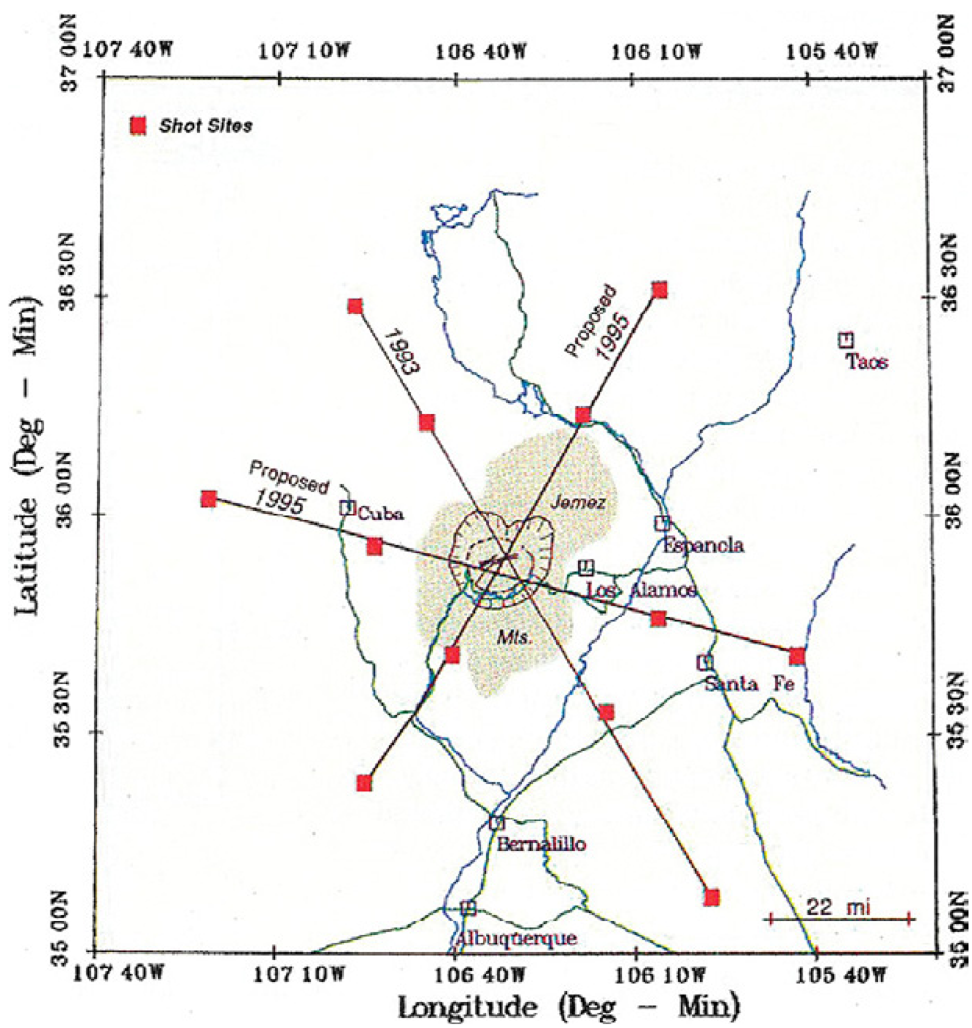



Figure 9.4.2-24. Layout of the active component of JTEX (Jemez Tomography Experiment). Seismic lines extend from the Colorado Plateau across the Rio Grande rift and Southern Rocky Mountains into the Great Plains (from Baldridge et al., 1997, fig.2). [Eos (Transactions, American Geophysical Union), v. 78, p. 417, 422–423. Reproduced by permission of American Geophysical Union.]

Some information on deep crustal structure farther north in the Southern Rocky Mountain region in southern Colorado was obtained by reprocessing data of a seismic-reflection survey, completed by Chevron USA Inc. in 1984 near latitude 38°N in the San Luis Valley, the northernmost part of the Rio Grande rift (Tandon et al., 1999). As a result, the trace of the main bounding fault of the basin against the Sangre de Cristo Range could be correlated to a wide zone of dipping reflections which soled out at lower crustal level at 26–28 km depth. However, unequivocal Moho reflections below the San Luis basin could not be identified, possibly due to a gradational nature of the Moho.

The "Deep Probe" experiment of 1995 followed approximately the 110th meridian (for location, see Figs. 9.4.1-11 and 9.4.2-26) and spanned a distance range of 3000 km from the southern Northwest Territories to southern New Mexico. It targeted the velocity structure from the base of the crust to depths as great as the mantle transition zone near 400 km depth. Details of the project were described in section 9.4.1.6, in particular those dealing with the Canadian side. On the U.S. side, only a few large shots were organized so that only the gross structure of the crust could be revealed.

The interpretation of the data resulted in a 2400-km-long lithospheric cross section (Fig. 9.4.2-25) reaching from the Colorado Plateau in southern New Mexico, United States, into the Alberta plains in Alberta, Canada (Snelson et al., 1998; Gorman et al., 2002). Two outstanding features were a high-velocity zone in the lower crust under Wyoming and Montana and indications for a deep-reaching suture zone connected with the Uinta uplift in Colorado. Under the Colorado Plateau, the crustal thickness increases gradually from ~35 km in southern New Mexico to ~45 km near the suture zone, where the crust thickens to 50 km.

Farther north beyond the Wind River uplift and basin with only 40 km thickness, the model shows a sudden increase to more than 50 km and then, underneath the Archean Wyoming Province, a northward gradual decrease to ~35 km under the Alberta Plains. Upper-mantle velocities are on average 7.9 km/s under the Proterozoic crust of the Colorado Plateau and 8.1 km/s under the Archean crust of the Wyoming Province. A low-velocity zone with 7.6–7.7 km/s between ~50 and 80 km depth in the uppermost mantle underneath a high-velocity lid of 7.9–8.0 km/s at Moho level was shown in the colored mantle cross section of Deep Probe Working Group (1998), but was not included in the more detailed interpretation of Snelson et al. (1998, Fig. 9.4.2-25).

In the same year (1995), when the Deep Probe experiment was conducted, 14 U.S. institutions combined their efforts and created the interdisciplinary geoscience project CD-ROM (Continental Dynamics–Rocky Mountains) with the aim of realizing a detailed crustal and upper mantle interdisciplinary investigation of the Southern Rocky Mountains from central Wyoming to central New Mexico (Fig. 9.4.2-26), involving tectonics, structure geology, regional geophysics, geochemistry, geochronology, xenoltih studies, and seismic studies (Keller et al., 1999, 2005a; Karlstrom and Keller, 2005).

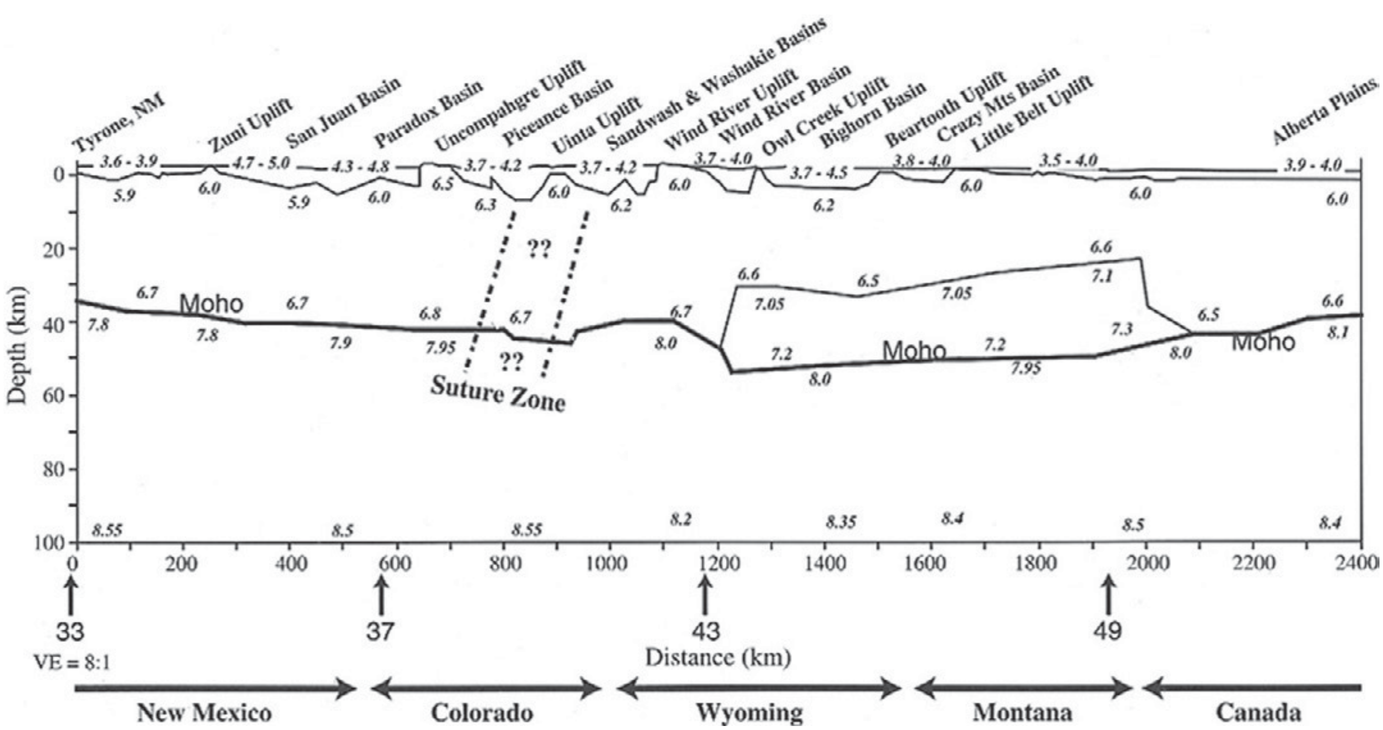



Figure 9.4.2-25. Velocity model for the Deep Probe transect (from Snelson et al., 1997, fig.2). For location, see Fig. 9.4.1-11. [Rocky Mountain Geology, v. 33, p. 181–198. Copyright Rocky Mountain Geology.]

Its main aim was to study the Proterozoic assembly of southwestern North America which has been overprinted by Phanerozoic orogenic processes such as the Ancestral Rocky Mountain orogeny (Kluth and Coney, 1981), the Laramide orogeny (e.g., Erslev, 2005), and the formation of the Rio Grande rift (Keller and Baldridge, 1999). After an initial workshop in 1995, CD-ROM was funded from 1997 to 2002 by the Continental Dynamics Program of the National Science Foundation, with supplemental funding from the German Research Society to the University of Karlsruhe.

The seismic investigations in particular comprised passive-source seismic studies (e.g., Sheehan et al., 2005, Burek and Dueker, 2005) and two active-source experiments (Figs. 9.4.2-26 and 9.4.2-27), consisting of two seismic-reflection profiles ~200 km long (Morozova et al., 2005, Magnani et al., 2005), and a 950-km-long seismic-refraction profile (Rumpel et al., 2005, Snelson et al., 2005, Levander et al., 2005).

The 950-km-long seismic-refraction/wide-angle reflection line was arranged perpendicular to the strike of the Proterozoic assembly of southwestern North America and therefore followed a NNW-SSE direction. It extended from central Wyoming into central New Mexico. Originally it had been planned to establish a seismic-reflection survey along the entire transect following exactly the seismic-refraction line (dashed line in Fig. 9.4.2-26), but for financial reasons only 400 km of seismic-reflection recordings could be realized, with the lines deviating partly from the refraction line for logistic reasons.

The northern reflection section crossed the Cheyenne belt (Morozova et al., 2005), an assumed Archean-Proterozoic suture zone, and the southern reflection section investigated the Yavapai-Mazatzal transition zone and the Jemez lineament in northern New Mexico (Magnani et al., 2005). To compensate at least in part for the central segment of the traverse in Colorado not being covered by reflection profiling, more shotpoints and reduced

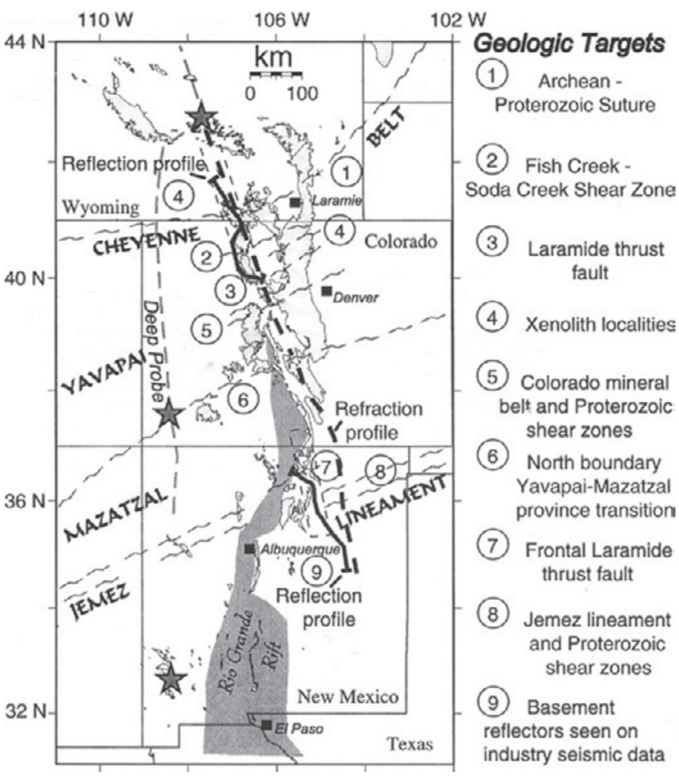



Figure 9.4.2-26. Targets of CD-ROM and preliminary location map of the CD-ROM transect (from Keller et al., 1999, fig.1). Thick full lines—seismic-reflection profiles; thick dashed line—seismic-refraction transect; thin dashed line—Deep Probe profile; stars—Deep Probe shotpoints 43, 37, and 33 (from N to S). [Rocky Mountain Geology, v. 33, p. 217–228. Copyright Rocky Mountain Geology.]

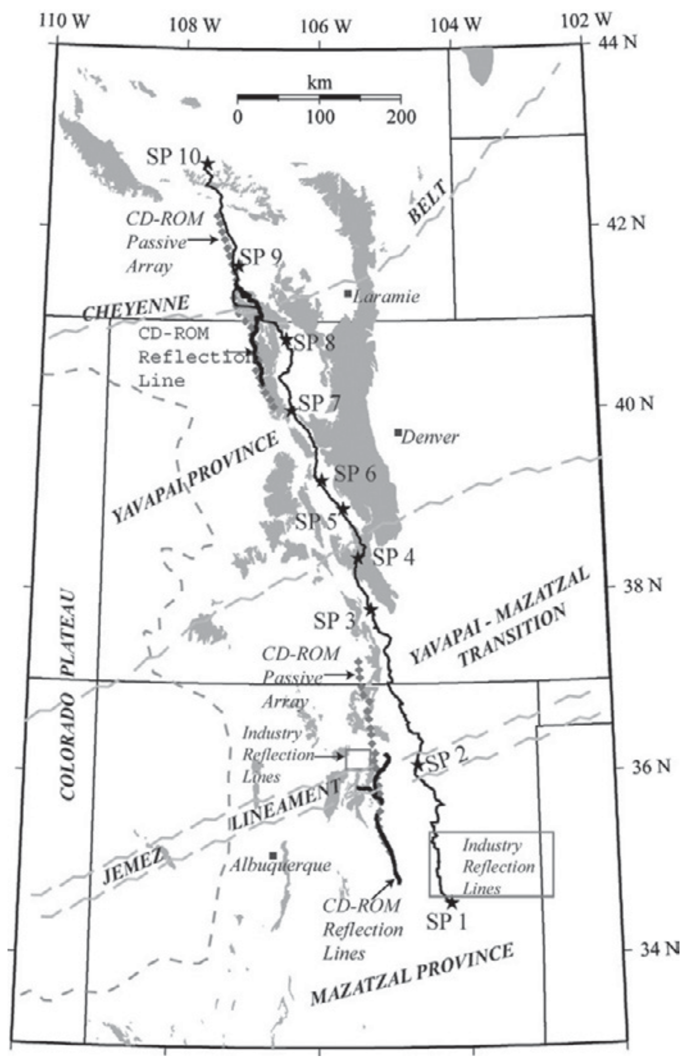



Figure 9.4.2-27. Location map of the CD-ROM seismic lines (from Rumpel et al., 2005, fig.1). Thick black lines—seismic-reflection profiles; thin black line—seismic-refraction profile; stars—CD-ROM refraction shotpoints; boxes—areas where industrial seismic reflection profiles were recorded; grey areas—exposed Precambrian basement. [The Rocky Mountain Region: An evolving lithosphere—tectonics, geochemistry, and geophysics: American Geophysical Union Monograph 154, Washington, D.C., p. 257–269. Reproduced by permission of American Geophysical Union.]

station spacing was established which became possible by the additional funding and participation from Germany.

The line started at the Deep Probe shotpoint 43 in the Gas Hills region of central Wyoming (shotpoint 10 in Fig. 9.4.2-27; see also Figs. 9.4.1-11 and 9.4.2-26), where a charge of 4530 kg could be detonated without involving ripple firing. From the next shotpoint (no. 9), coinciding with the northern end of the northern seismic-reflection profile, the seismic line continued through the Sierra Madre in Wyoming, crossed the North and South Parks, located to the west of the Front Range, and the Wet Mountains in SSE direction and left the Rocky Mountains south of the Wet Mountains near shotpoint 3 and terminated 350 km farther south in the Great Plains of central New Mexico. With the exception of shotpoint 10, the charge sizes ranged between 170 and 2720 kg. Shot spacing varied from 200 km to 41 km, with an overall average spacing of 106 km, and was smallest in the central section between shotpoints 3 and 8, where an average shot spacing of 60 km resulted. Approximately 600 seismographs were deployed twice along the entire profile, resulting in an average receiver spacing of 800 m for all shots. Three-component stations occupied the central section between shotpoints 3 and 8, and one-component stations were used for the northern and southern sections. Energy was well recorded up to 350 km offsets, but in some cases greater distances were reached.

The data preparation (Appendix A9-2-7) and first modeling was achieved simultaneously at El Paso, Texas, and Karlsruhe, Germany. A data example is shown in Figure 9.4.2-28, presenting the record section from the southernmost shotpoint near Gardner, with correlations by Snelson et al. (2005). The data were described and interpreted in detail by Rumpel (2003) and Snelson (2001), and the results were presented in three publications. Rumpel et al. (2005) and Snelson et al. (2005) primarily used a ray-tracing methodology and divided their attention in particular on details of the upper and middle crust (Rumpel et al., 2005) and on the lower crust and uppermost mantle (Snelson et al., 2005), while Levander et al. (2005) used an inversion and tomographic approach.

Crustal thickness varies substantially along the entire line (Figs. 9.4.2-29 and 9.4.2-30). Moho depth is near 50 km at the northern end, increases gradually toward the south, and reaches a maximum depth of ~57 km underneath the North and South Parks west of the Front Range in central Colorado, where the Colorado Mineral belt traverses the Southern Rocky Mountains. A local Moho high of ~50 km was interpreted under the Wet Mountains, followed by a crustal thickness increase of ~5 km around shotpoint 3 where the line enters the Great Plains. Southward from here Moho depth decreases gradually to near 40 km depth at the southern end of the line.

As mentioned above, the seismic-reflection investigations concentrated on the two major suture zones of the Proterozoic assembly of southwestern North America, the Cheyenne belt in southern Wyoming/northern Colorado and the Jemez lineament in north-central New Mexico, while the central Proterozoic boundary, coinciding with the present-day Colorado Mineral belt, had been covered by the more densely recorded central section of the seismic-refraction profile (Figs. 9.4.2-25 and 9.4.2-26).

The northern seismic-reflection line was recorded in 1999 under the leadership of scientists from the Universities of Wyoming and Arizona. It followed the axis of the Sierra Madre in northern Colorado and southern Wyoming in an overall N-S orientation, parallel to the strike of the Laramide uplifts, and was the first reflection survey recorded on land within the United

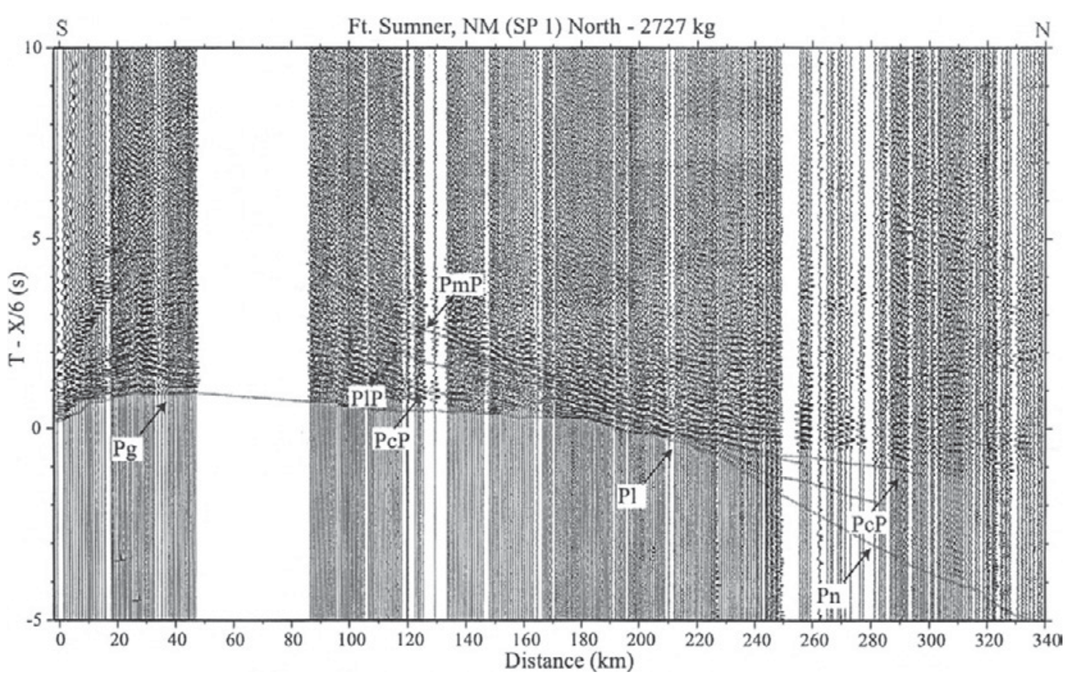



Figure 9.4.2-28. Record section of the CD-ROM seismic-refraction profile from SP1 Gardner with travel time correlations (from Snelson et al., 2005, fig.2). [The Rocky Mountain Region: An evolving lithosphere—tectonics, geochemistry, and geophysics: American Geophysical Union Monograph 154, Washington, D.C., p. 257–269. Reproduced by permission of American Geophysical Union.]

States since the last COCORP survey 8 years previously. It was a Vibroseis survey with four 25,000 kg-force vibrators, 25 m station and 100 m source intervals and a 4 ms sample interval into 1000 channels resulting in 25-km-long recording spreads, giving a nominal fold of 125. Record lengths were 25–30 s. The new reflection data, which were modeled together with the northern part of the refraction data, offered a dramatic new view of the geometry and tectonic history of the Cheyenne belt suture which was imaged as a complex series of wedges resulting in an interleaving of the Archean crust in the north and the Proterozoic crust in the south (Morozova et al., 2005).

The southern seismic-reflection line, organized by scientists from Rice University at Houston, Texas, and the University of Texas at El Paso, was also recorded in 1999. It extended for a total of 170 km north and south of the town of Las Vegas, New Mexico, which is located at the southern edge of the Jemez lineament. It was again a Vibroseis survey, and the data were acquired with 1001 active recording channels along a 25-km-long seismograph array, with 25 m group intervals and a 100 m source interval. The resulting record length was 25 s.

The data, crossing the Cenozoic Jemez lineament in northern New Mexico, showed impressive reflectivity throughout the crust. The middle-lower crust, from 10 to 45 km depth, evidenced oppositely dipping reflections that converged roughly at the center of the profile at 35–37 km depth. They were interpreted as Proterozoic-age crustal structures forming a doubly vergent suture zone that has survived subsequent tectonism (Magnani et al., 2005).

Keller et al. (2005a) compiled the results of pre-CD-ROM and CD-ROM results (Fig. 9.4.2-29) and summarized them in a Moho contour map (Fig. 9.4.2-30) which also included results obtained from receiver functions (Sheehan et al., 2005; Burek and Dueker, 2005). They also illustrated the geologic evolution of the crust-mantle boundary in the Southern Rocky Mountain region in time from the early Proterozoic until today and its connection in time with the evolution of surface elevation and petrologic crustal modifications (Keller et al., 2005a; Rumpfhuber and Keller, 2009; Rumpfhuber et al., 2009).

### *9.4.2.6. The Southern Appalachian–Atlantic Coast Region*

In a comparison of the tectonic history of the Appalachian-Ouachita orogen and the Trans-European suture region, Keller and Hatcher (1999) have examined all large-scale geophysical studies conducted in the U.S. portion of the Appalachians. As since the late 1980s, onshore geophysical studies in the southern Appalachians have been modest in scope, they had to rely

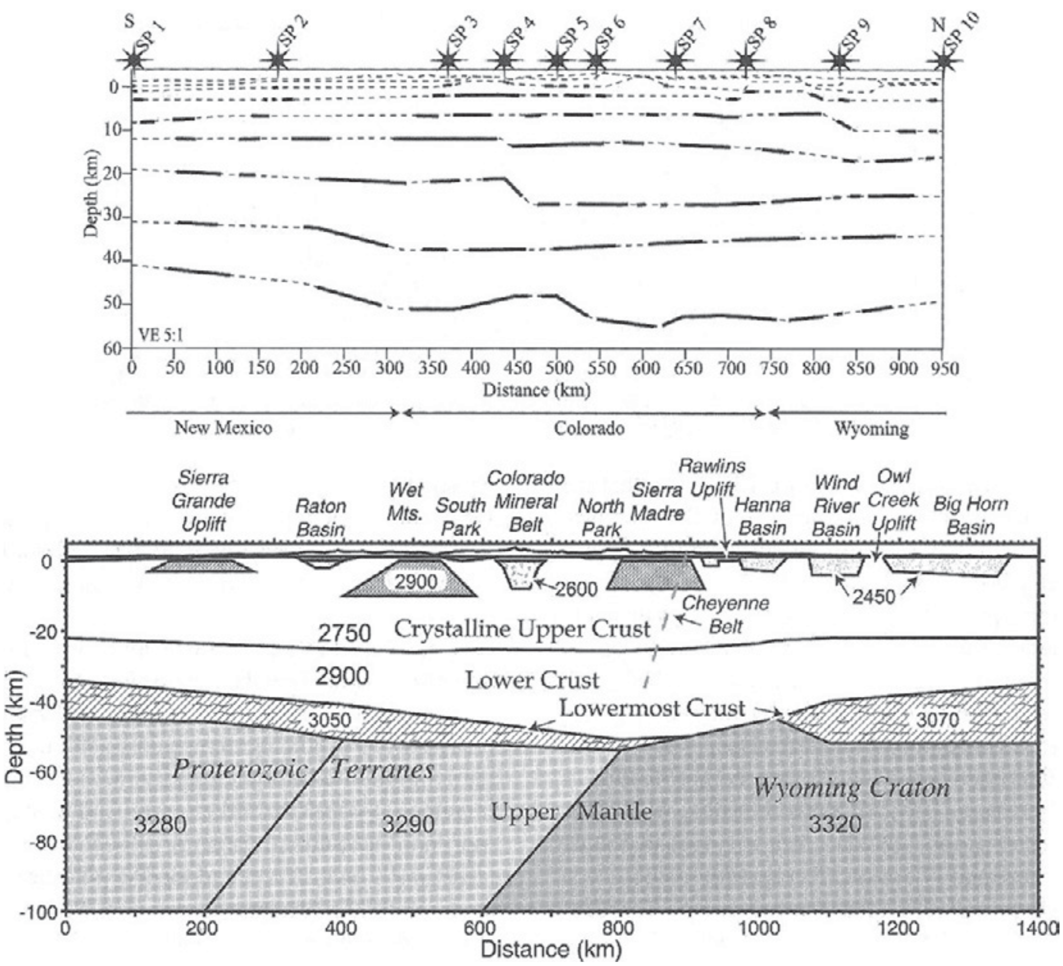



Figure 9.4.2-29. Interpretation of the CD-ROM seismic-refraction data (from Snelson et al., 2005, figs. 11c and 12d). Top: Model based on first arrivals and wide-angle reflections (from Snelson et al., 2005, fig. 11c). Bottom: 2.5-D density model along the CD-ROM transect. The distance scale corresponds to that of the seismic model (from Snelson et al., 2005, fig.12d). [The Rocky Mountain Region: An evolving lithosphere—tectonics, geochemistry, and geophysics: American Geophysical Union Monograph 154, Washington, D.C., p. 271–291. Reproduced by permission of American Geophysical Union.]

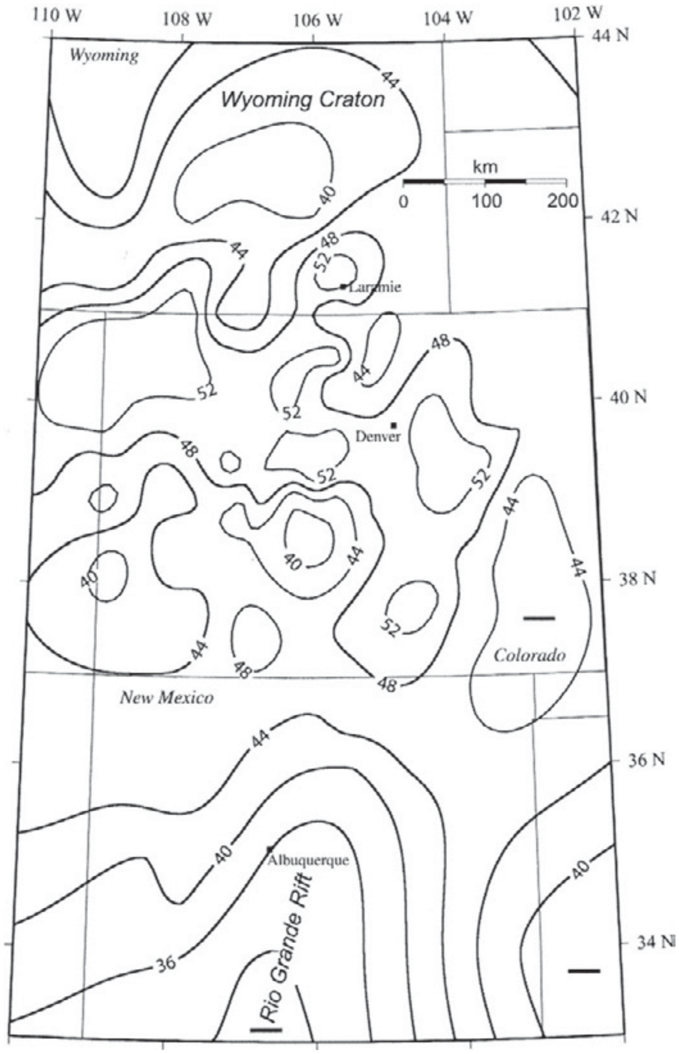



Figure 9.4.2-30. Moho contour map of the Southern Rocky Mountains derived by gridding and contouring the results of previous studies of seismic-refraction data and receiver functions in the area with the results of the CD-ROM seismic experiment (from Keller et al., 2005). [The Rocky Mountain Region: An evolving lithosphere—tectonics, geochemistry, and geophysics: American Geophysical Union Monograph 154, Washington, D.C., p. 271–291. Reproduced by permission of American Geophysical Union.]

on summaries prior to the late 1980s, provided, e.g., by Taylor (1989) and Phinney and Roy-Chowdhury (1989). Besides a local seismic-refraction survey in the New Madrid seismic zone, investigating the shallow upper crustal structure to 12 km depth (Hildenbrand et al., 1995), only one major active-source onland crustal investigation was carried out in the eastern United States in the 1990s in South Carolina (Luetgert et al., 1992, 1994). It was a seismic-refraction/wide angle reflection profile in the Atlantic coastal plain of South Carolina (Luetgert et al., 1992; Appendix 9-2-8).

The recording line of 1991 in South Carolina crossed the coastal plain from southeast to northwest over a length of 120 km. Portable vertical seismographs of the type SGR III were installed every 1 km, in addition horizontal seismographs were placed at 2 km spacing. Five shotpoints with 1000 kg charges were located at intervals of 30 km (Fig. 9.4.2-31).

The model of Luetgert et al. (1994) showed that the Coastal Plain sediments thicken from a few tens of meters to more than 1 km. Large lateral velocity variations were found for the upper 6–7 km of the crystalline crust. Around shotpoint 3 the velocity in the upper part of the basement (1–7 km depth) increased almost abruptly from 5.7 to 5.8 km/s in the east to 6.1–6.3 km/s in the west. Below 7 km depth no lateral velocity changes were evident. The mean crustal velocity increases only slightly from 6.4 to 6.5 km/s in the middle crust to 6.6–6.7 km/s in the lower crust. The Moho dips from ~32 km depth in the east to 37 km in the west (Luetgert et al., 1994).

In 1996 and 1997, Hawman et al. (2001) recorded quarry blasts in the area of the Cumberland Plateau Seismic Observatory, where the USGS had observed two 300-km-long crossing profiles in 1965 (Prodehl et al., 1984). The study was undertaken to evaluate the feasibility of using relatively small, delay-fired quarry blasts as energy sources for wide-angle reflection profiling of the middle and lower crust beneath the southern half of the eastern Tennessee seismic zone. Timed blasts at four crushed-stone quarries were recorded by a portable array of 19 stations with three-component geophones at roughly 200 m interval. Source-receiver offsets ranged from 16 to 67 km.

The blasts ranged from 75 to 542 ms duration and from 259 to 942 lbs explosives per delay. The first tests of the complicated interpretation procedure due to the extended source signatures produced by the ripple firing showed that usable signal levels at frequencies up to 40 Hz at distances to at least 65 km could be obtained and that also nowadays, since quarry blasts use exclusively ripple firing, useful results for deep crustal studies can be expected by a systematic observation of commercial quarry blasts.

The EDGE Mid-Atlantic seismic experiment along the east coast, which was performed in 1990 and collected multichannel and wide-angle seismic data offshore and onshore of Virginia (Holbrook et al., 1992b, 1994b, Lizarralde and Holbrook, 1997), was discussed in Chapter 8.9.4 in connection with marine seismic investigations in the Atlantic Ocean. The inclusion of the onshore data (Fig. 9.4.2-32) showed a gradual thickening of the crust from the continental margin westwards across the coastal plain, where a uniform thickness of 35 km was observed, to ~45 km under the Blue Ridge section of the southern Appalachians (Lizarralde and Holbrook, 1997).

## 9.5. THE AFRO-ARABIAN RIFT SYSTEM

### 9.5.1. The East African Rift System in Kenya

The Afro-Arabian rift system has attracted geoscientists since the early 1900s, but hostile environmental and problematic political conditions have set severe restrictions until today. Modest

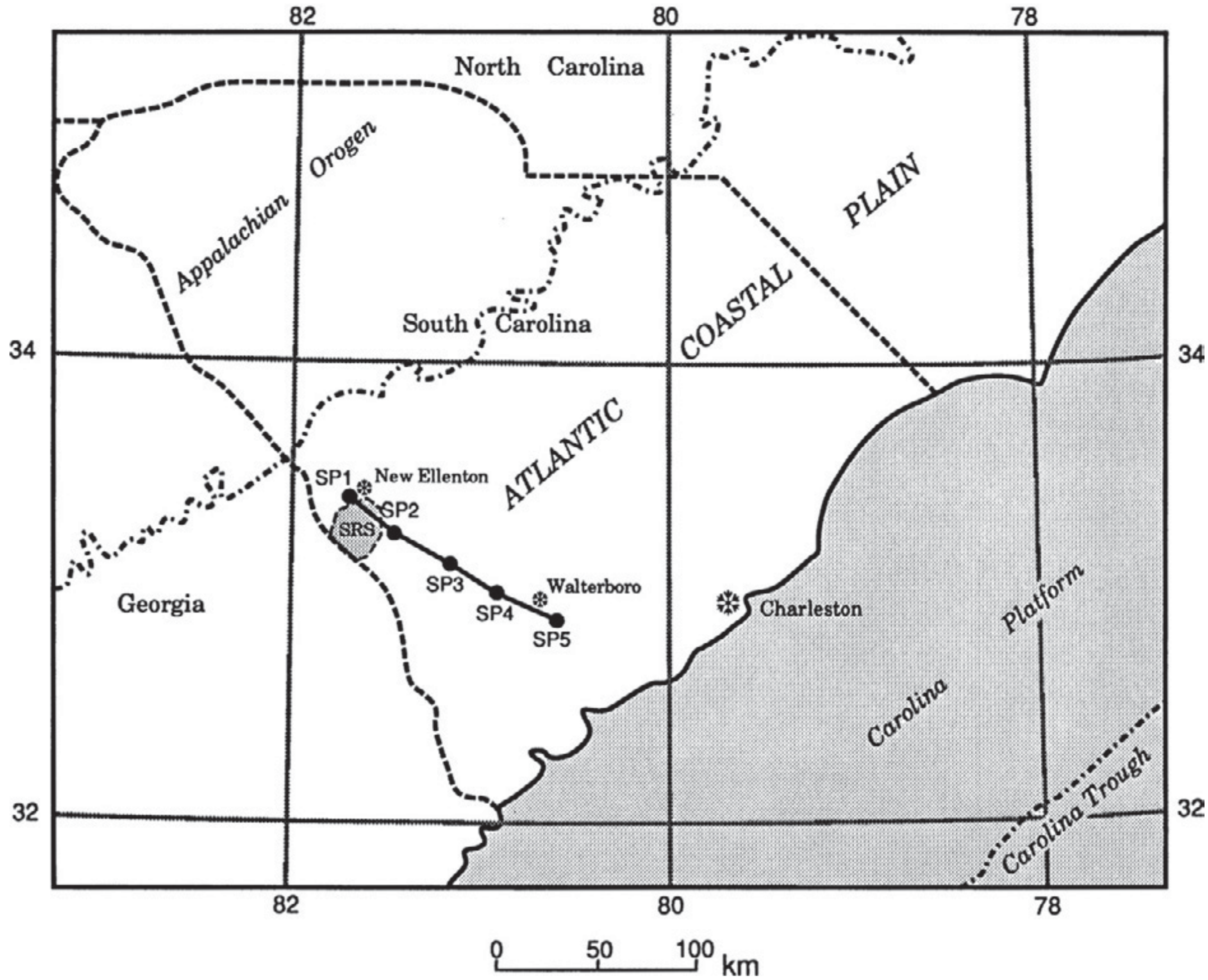



Figure 9.4.2-31. Location map of the 1991 South Carolina seismic experiment (from Luetgert et al., 1992, fig.1). SRS—Savanna River Site. [U.S. Geological Survey Open-File Report 92-723, 35 p.]

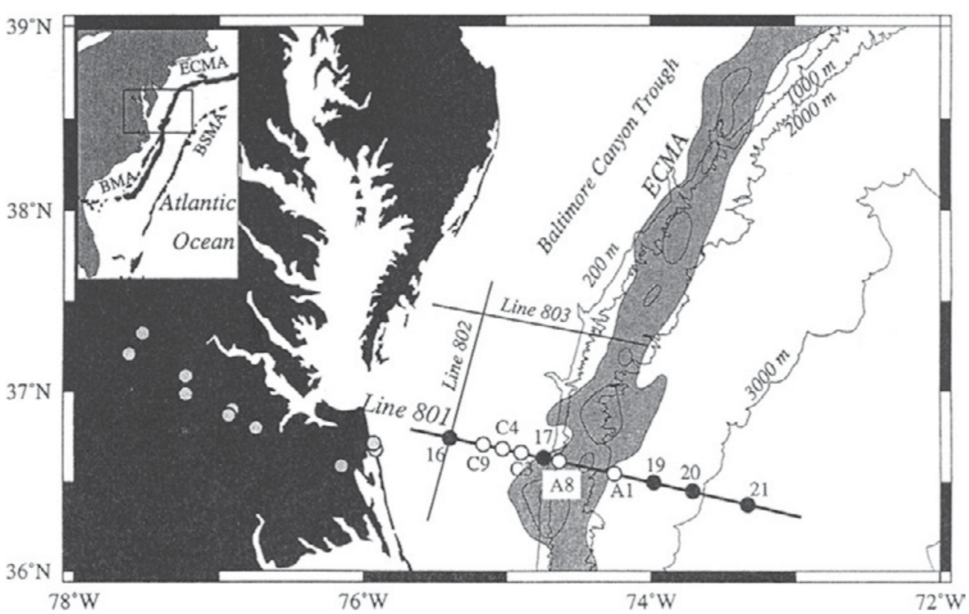



Figure 9.4.2-32. Location map of the EDGE Mid-Atlantic seismic experiment (from Holbrook et al., 1994b, fig.1). Black and white circles—locations of OBS; shaded circles—onshore seismometers. [Journal of Geophysical Research, v. 99, p. 17,871–17,891. Reproduced by permission of American Geophysical Union.]

attempts to study the crustal structure by controlled-source seismology had started in the late 1960s in the East African rift system of Kenya and continued in the 1970s, until a first major project was undertaken in 1985. The Ethiopian part of the East African rift system saw a major experiment in 1972, and the Dead Sea rift was studied in much detail at the end of the 1970s and in the mid-1980s (see corresponding sections in Chapters 6 to 8).

Based on the experiences collected by the earlier expeditions, major efforts were undertaken in the 1990s and in the first years of the twenty-first century to unravel the details of crustal structure of the East African rift system. In 1990 and 1994, two major international campaigns explored the crust and upper mantle under the East African rift of Kenya (KRISP Working Party, 1991; Keller et al., 1992; Prodehl et al., 1994a; KRISP Working Group, 1995a; Fuchs et al., 1997). In 2001, the northern end of the East African Rift system was the target of a major seismic experiment (Maguire et al., 2003; see Chapter 10), and the Dead Sea transform saw the beginning of a major international seismic campaign in 2000, covering both sides of the rift in Jordan and Israel (DESERT Team, 2000; DESERT Group, 2004), followed by a second cross-rift seismic profile in 2004 (ten Brink et al., 2006; see Chapter 10) and a third one in 2006 (Weber et al., 2007; see Chapter 10).

The seismic investigation of the East African Rift in Kenya of 1990 by KRISP (Kenya Rift International Seismic Project) contained, similar to the preceding project of 1985, two components (Prodehl et al., 1994b). An extended seismic-refraction/wide-angle reflection survey and a teleseismic tomography experiment (Fig. 9.5.1-01) were jointly undertaken to study the lithospheric structure of the Kenya rift down to depths of greater than 200 km.

During the teleseismic survey of 1989–1990, an array of 65 seismographs was deployed (circles in Fig. 9.5.1-01) to record teleseismic, regional and local events for a period of 7 months. The elliptical array spanned the central portion of the rift, with Nakuru at its center, and covered an area of ~300 × 200 km, with an average station spacing of 10–30 km (Green et al., 1991; Achauer et al., 1992, 1994; Slack et al., 1994).

The controlled-source seismic part consisted of three profiles. All P-wave data, except for line A, are presented as record sections in Appendix A2-1 (p. 91–96) and Appendix A9-3 (KRISP Working Group 1991). The axial profile ran in N-S direction from Lake Turkana in the north to Lake Magadi in the south (Fig. 9.5.1-01). A total of 206 mobile vertical-component seismographs were deployed three times and, with the exception of the first deployment, recorded the energy of underwater and borehole explosions, with an average spacing of 2 km, to distances of ~550 km.

The first deployment served as a test for instrumentation and shooting techniques in a hostile environment as well as to resolve fine structure of the crust for the part of the rift where a larger number of shots could most economically be arranged. It covered 140 km along the western shore of Lake Turkana. Average station spacing was 0.7 km and small underwater shots of 100–400 kg were recorded from five positions in Lake Turkana (LT1 to LT5). To decrease the recording interval to 350 m, at three positions two shots, separated by 350 m, were fired (Gajewski et al., 1994).

For the main part of the rift, based on the experience of the 1985 experiment, the majority of shots were also underwater shots, which were fired twice, once for each deployment, in Lake Turkana near its northern end (LTN, Fig. 9.5.1-02) and in the center (LTC), in Lake Baringo (BAR), in Lake Bogoria (BOG), and in Lake Naivasha (NAI) with charges ranging from 400 to 1200 kg. Only one drillhole shot was detonated near Lokoro (LKO), halfway between Lake Turkana and Lake Baringo (Mechie et al., 1994; Keller et al., 1994a).

The cross profile was 450 km long and traversed the rift in W-E direction from Lake Victoria in the west to Chanler's Falls on the Ewaso Ngiro River in the east, 50 km east of Archers Post. The line crossed the rift at Lake Baringo and was recorded with one deployment. Here only two underwater shot positions were available in Lake Victoria (VIC, Fig. 9.5.1-03) and Lake Baringo (BAR), where charges of ~1000 kg were set off. For security reasons, the easternmost shotpoint CHF (Chanler's Falls), planned as an underwater shot in the river, could not be realized for this line. Instead, several borehole shots were arranged which were located at positions with shallow groundwater table near the towns of Barsalingo (BAS), Tangulbei (TAN) and Kaptagat (KAP). Charges of 1000–2200 kg were set off here. Station spacing was 1 km within the rift proper, 2 km on both flanks to ~120 km east and 170 km west of Lake Baringo, and 5 km to both ends of the line (Maguire et al., 1994; Braile et al., 1994).

A third flank profile explored the eastern flank of the rift and ran in NW-SE direction for a distance of 300 km from Lake Turkana to Chanler's Falls. The purpose of this line was to obtain a model for an area which was not influenced by the processes shaping the Kenya rift. Except for the two shots in Lake Turkana (LTC and LTS), where underwater shots of 350 kg were set off, borehole shots with charges between 400 and 2000 kg had been organized along the remainder of the line near the villages of Illaut (ILA) and Laisamis (LAI), and at Chanler's Falls (CHF) on the Ewaso Ngiro River, where the only shot possible in this area was used for this flank line toward the northwest. Station spacing was 1.5 km with two major gaps at both ends (Prodehl et al., 1994c).

The KRISP90 field experiment required a careful organization and permissions, which lasted for several years and was finally accomplished by a close cooperation with Kenyan scientists who provided substantial support in all circumstances. Prodehl et al. (1994b) have summarized all technical details from the first planning to the final realization of the fieldwork. Jacob et al. (1994) had carefully investigated how to optimize wide-angle seismic signal-to-noise ratios and P-wave transmission. In total, 34 shots were fired from 19 shotpoints with charges ranging from 100 kg to 2000 kg. All lake shots had produced excellent recordings, limited in the rift to 400 km offsets and on the flanks to 450 km. Also the borehole shots, except LKO, provided suitable energy to at least 250 km distance. In total, 1063 recording

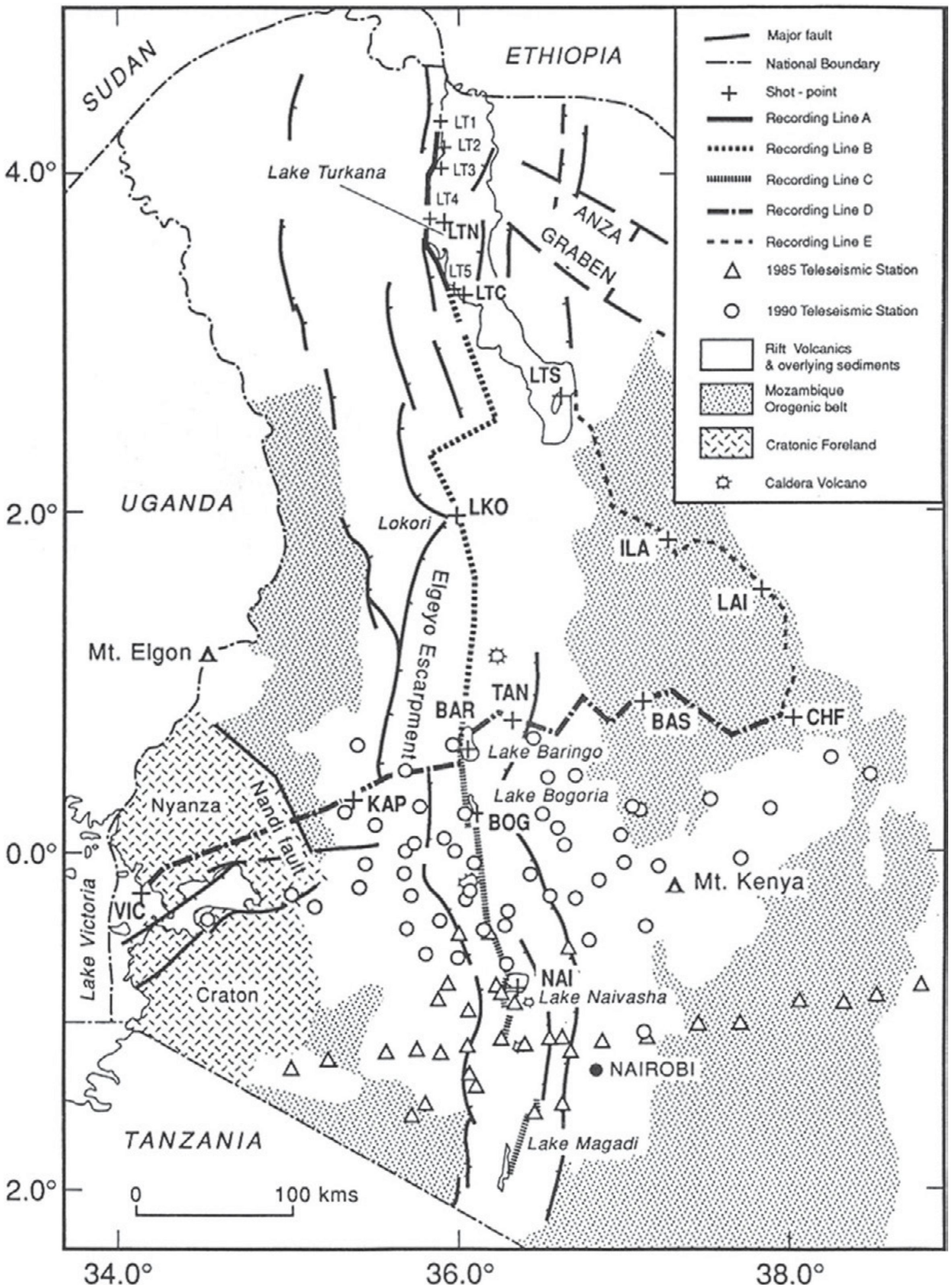



Figure 9.5.1-01. KRISP90 location map showing the seismic refraction/wide-angle reflection lines and the configuration of the teleseismic networks in 1985 and 1989–1990 (from Keller et al., 1994b, fig. 1). The 1985 refraction lines extended from Lake Baringo to Lake Magadi and across the rift valley south of Lake Naivasha, parallel to the 1985 teleseismic cross line (white triangles). [Tectonophysics, v. 236, p. 465–483. Copyright Elsevier.]

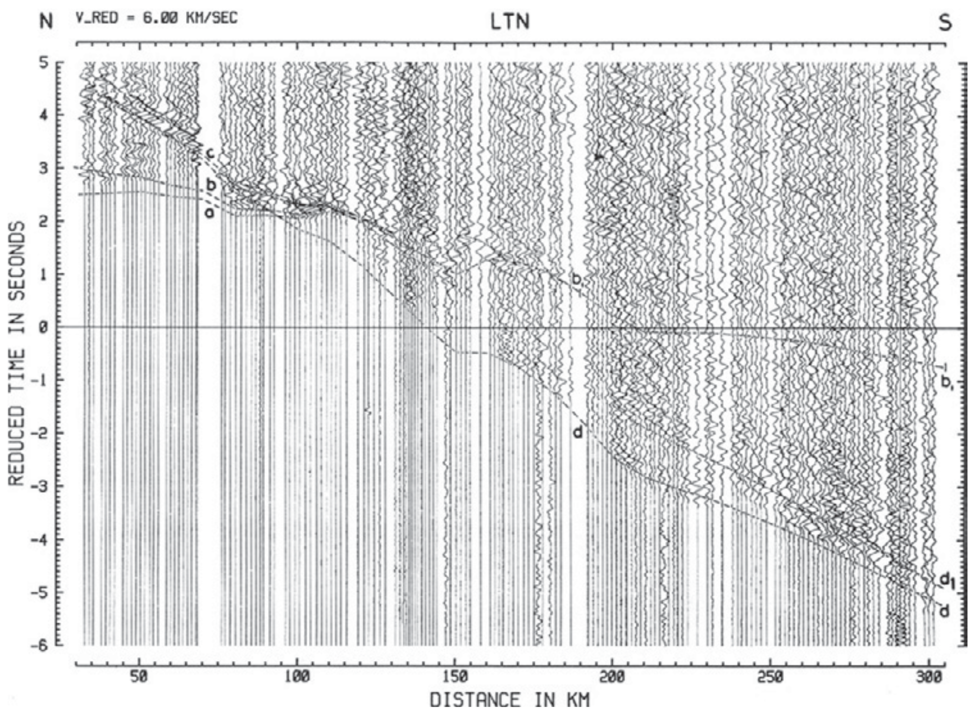



Figure 9.5.1-02. KRISP90 data example for the axial line. Record section of shotpoint LTN (from Mechie et al., 1994, fig. 3). [Tectonophysics, v. 236, p. 179–200. Copyright Elsevier.]

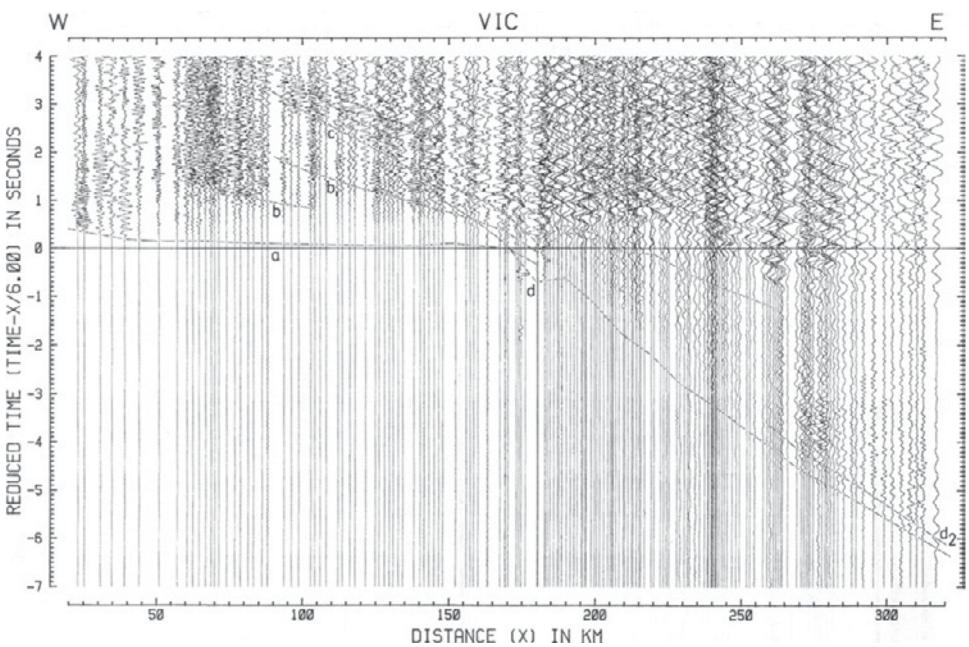



Figure 9.5.1-03. KRISP90 data example for the cross profile. Record section of shotpoint VIC (from Mechie et al., 1997, fig. 4). [Tectonophysics, v. 278, p. 95–119. Copyright Elsevier.]

sites were occupied. All sites had been located beforehand in pre-site surveys with many positions being determined by GPS measurements. Seventy-two scientists divided their tasks in self-sustaining recording and shooting parties and technical services at a mobile headquarters where batteries were recharged, instruments repaired and tapes collected to be transported to a temporary processing center at Egerton University near Nakuru.

The data were jointly processed, and their interpretation was coordinated by several workshops, leading to a special publication volume (Prodehl et al., 1994a). Concerning the controlled-source data, several contributions were prepared to deal with individual aspects and lines. The main results (Figs. 9.5.1-04 and 9.5.1-05) were summarized by Keller et al. (1994b) and Mechie et al. (1997) including a cross section through the southern part

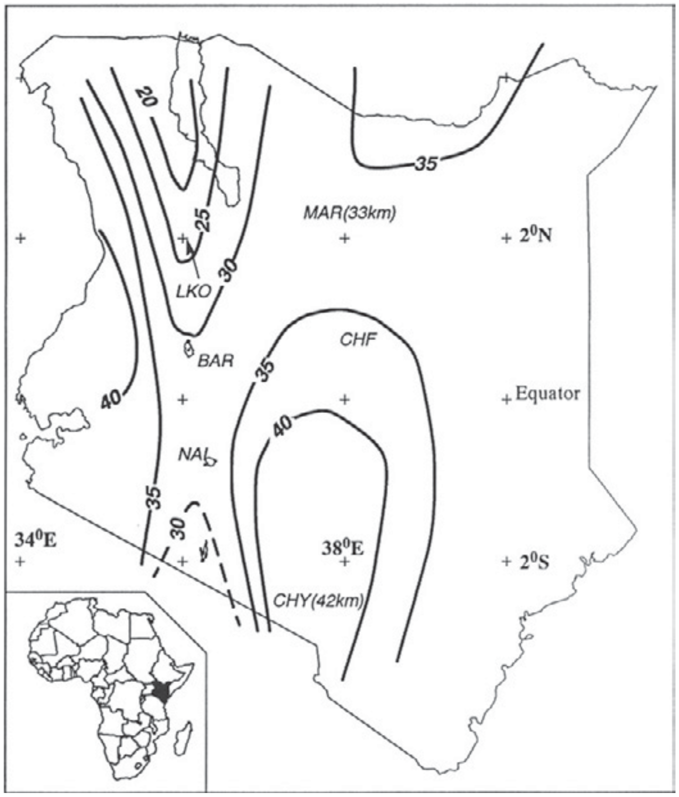



Figure 9.5.1-04. Crustal thickness map for the Kenya rift region (from Keller et al., 1994b, fig. 3). MAR and CHY indicate estimates of crustal thickness from xenoliths in the Marsabit and Chyulu Hills areas, respectively. [Tectonophysics, v. 236, p. 465–483. Copyright Elsevier.]

of the Kenya rift provided by teleseismic delay time studies (Achauer et al., 1992, 1994).

The most significant discovery from KRISP90 was that the crustal thickness along the rift varies from as little as 20 km beneath Lake Turkana to ~35 km under the culmination of the Kenya dome near Lake Naivasha. The transition between Lokori and Lake Baringo occurs over a horizontal distance of 150 km and resolves any discrepancies of earlier models. The limited data south of Lake Naivasha suggested a decrease in crustal thickness southwards toward Lake Magadi, where in 1990 no shotpoint could be realized. This was later accomplished in the KRISP94 experiment described further below. While under the rift uppermost mantle velocities near the Moho ranged from 7.5 to 7.7 km/s, under the flanks normal continental uppermost mantle velocities of at least 8.1 km/s were found. Here also the thickest crust of 40 km was found, but restricted to the surroundings of the southern portion of the rift (Keller et al., 1994a, Mechie et al., 1997).

Last but not least, during the active phase of KRISP90, the teleseismic array also recorded all shots. The data were used to interpret the P-wave first arrivals in order to determine lateral crustal inhomogeneities with a 3-D approach. Several high-velocity bodies found within the crust were attributed to possible mafic cumulates or magma chambers, which evolved during the major magmatic episodes and which are typical indicators for an active rift (Ritter and Achauer, 1994).

The KRISP89–90 project also unraveled details of the upper mantle structure beneath the Kenya rift and its flanks. The resolution of the controlled-source seismic data was restricted concerning the upper-mantle structure below the Moho. Only for the axial and the cross profiles some structure in the uppermost mantle could be revealed (Keller et al., 1994a; Maguire et al., 1994), indicating a low-velocity zone within the uppermost mantle below 55 km depth under the northern rift, and an overall velocity increase to 8.3–8.4 km/s under the northern rift and under both flanks at 60–65 km depth (Fig. 9.5.1-05).

In agreement with the model for the southern portion of the Kenya rift derived from the active-source seismic data of the axial profile, the teleseismic results (Achauer et al., 1994; Slack et al., 1994) indicated that below the southern part of the rift, a low velocity body exists which extends from the Moho to at least 165 km depth. While it was shown to be confined to the width of the rift proper in its upper portion, it widened at depths greater than 125 km. In a subsequent summary on the deep structure of the East African rift system, the teleseismic velocity perturbations in the upper mantle from 40 to 160 km depth, which had been published by Slack et al. (1994) as relative velocity variations in % (see Fig. 9.5.1-05), had been combined with the seismic-refraction cross section of Braile et al. (1994) and converted into absolute velocities (Braile et al., 1995; see unpublished report [IRIS, 2000] and figure by Braile [1997] in Appendix A9-3).

The continuation of the deep structure of the Kenya rift southward had remained an open question from KRISP90. Therefore, beginning in 1993 and continuing into early 1995, the KRISP Working Group began a new series of multidisciplinary field investigations in the rift in southern Kenya (KRISP Working Group, 1995a; Prodehl et al., 1997b). The research program involved six experimental efforts: (1) a teleseismic survey of the area of the Chyulu Hills located on the eastern flank of the rift, ~100 km to the southeast of Lake Magadi (Fig. 9.5.1-06; Ritter et al., 1995; Ritter and Kaspar, 1997); (2) a seismic-refraction/wide-angle reflection survey across southern Kenya extending from the Indian Ocean near Mombasa to Lake Victoria (Fig. 9.5.1-06); (3) the installation of a temporary seismological network in southern Kenya; (4) a local earthquake recording survey around Lake Magadi (Fig. 9.5.1-06) which was continued in 1997 and 1998 (Ibs-von Seht et al., 2001); (5) a detailed gravity survey along the active-source seismic lines (Birt et al., 1997); and (6) a magnetotelluric study at selected sites along the same seismic lines (Simpson et al., 1997).

The seismic-refraction/wide-angle reflection survey across southern Kenya consisted of two profiles which were in fact two deployments. Line F, the eastern line, extended from the eastern flank of the rift near Magadi to the Indian Ocean near Mombasa (Fig. 9.5.1-06). Line G, the western line, extended from the northwestern end of the Chyulu Hills across the southern rift and the game park Masai Mara to Lake Victoria near the border of Kenya and Tanzania. In total, 10 borehole shots were fired with

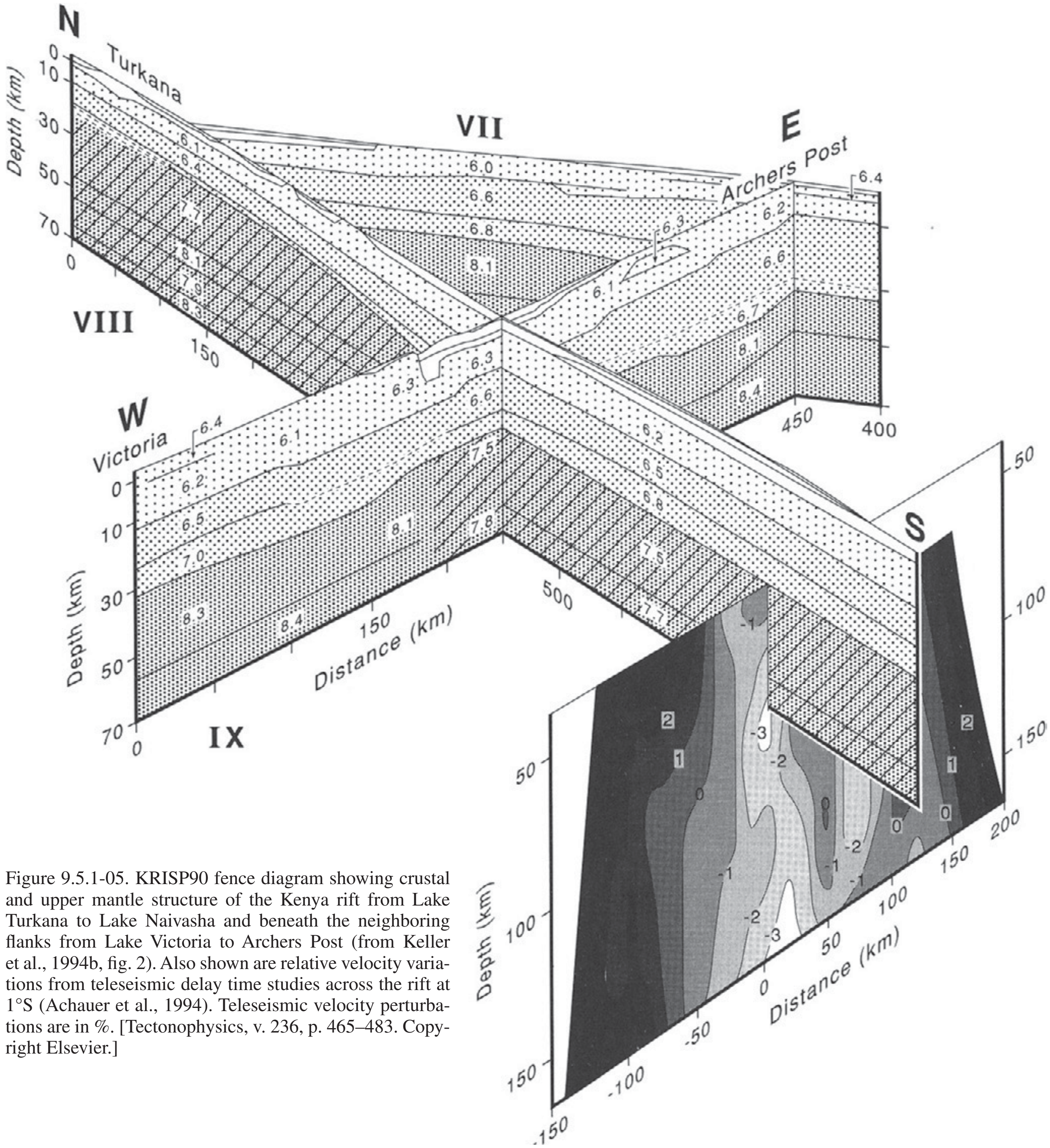



Figure 9.5.1-05. KRISP90 fence diagram showing crustal and upper mantle structure of the Kenya rift from Lake Turkana to Lake Naivasha and beneath the neighboring flanks from Lake Victoria to Archers Post (from Keller et al., 1994b, fig. 2). Also shown are relative velocity variations from teleseismic delay time studies across the rift at 1°S (Achauer et al., 1994). Teleseismic velocity perturbations are in %. [Tectonophysics, v. 236, p. 465–483. Copyright Elsevier.]

charges from 450 to 2080 kg, loaded in one or two boreholes. At the ends, underwater shots, where the charge was subdivided into smaller, separated units, could be arranged. Two shots in Lake Victoria had total charges of 900 kg each (VIS, Fig. 9.5.1-07). In the Indian Ocean, using the support by a Kenyan Navy vessel, only one small charge of 300 kg could be fired (IND). Furthermore, two quarry blasts (KIB) with 1000 kg charges near Kibini could be timed in accordance with the layout of the two lines and fired without delays between detonators.

To ensure an overlapping energy coverage along both lines, shots at CHN, MOR, and in Lake Victoria (VIS) were executed twice and recorded during both deployments. The energy distribution was almost perfect for all shots. In particular, the shots along the western part MOR and VIS) gave good arrivals to 450

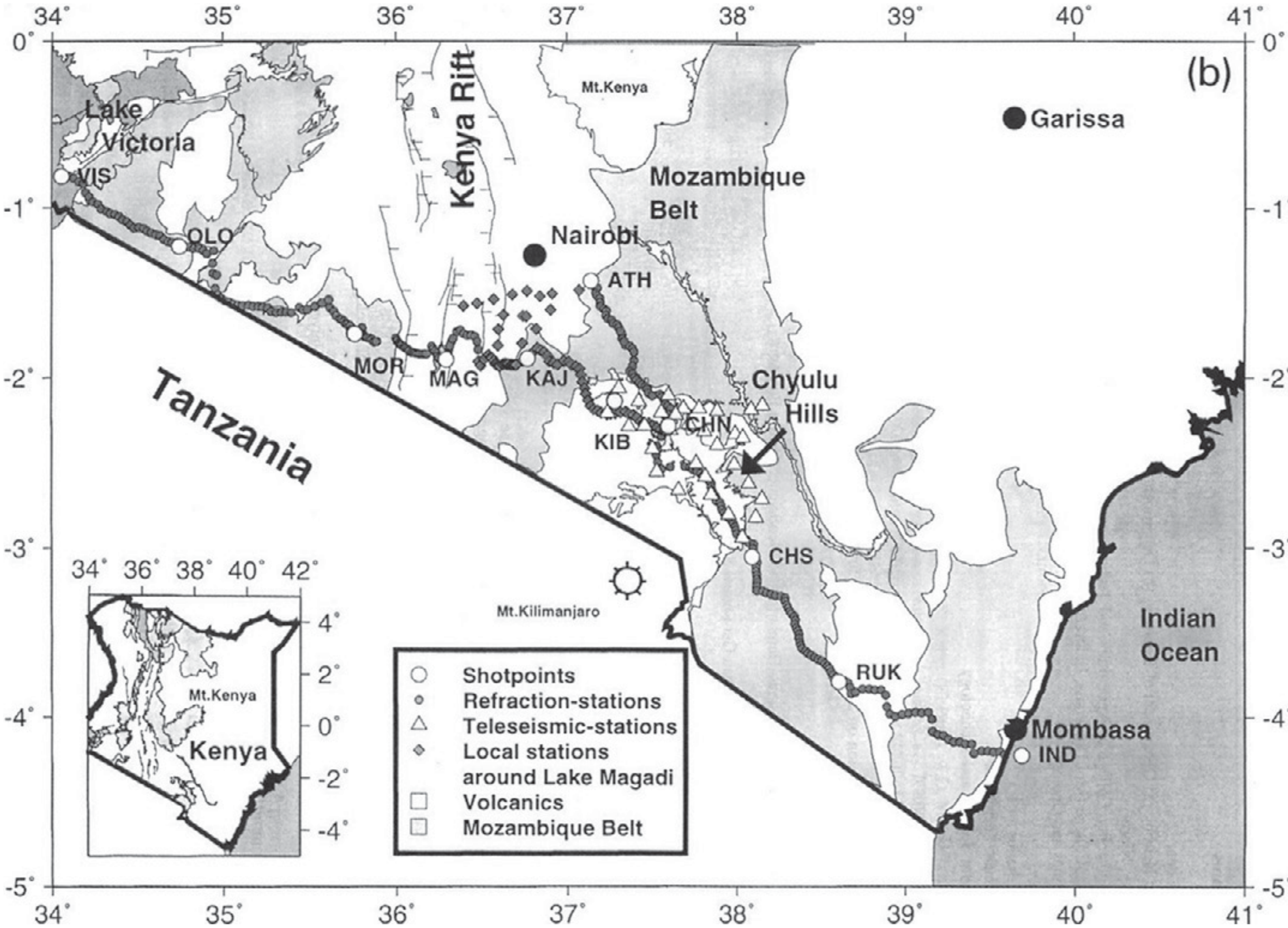



Figure 9.5.1-06. KRISP94 location map showing the seismic-refraction/wide-angle reflection lines (thick lines of black circles), the configuration of the Chyulu Hills teleseismic network of 1993–1994 (white triangles), and the local earthquake recording network (black quadrangles) around Lake Magadi (from Prodehl et al., 1997b, fig. 1b). [Tectonophysics, v. 278, p. 121–147. Copyright Elsevier.]

and 600 km distances, respectively (Fig. 9.5.1-07). All P-wave data, except for line A, are presented as record sections in Appendix A2-1 (p. 97–100) and Appendix A9-3 (KRISP Working Group, 1995b).

The details of crustal structures were interpreted separately for each line, the western line by Birt et al. (1997) and the eastern line by Novak et al. (1997a). Furthermore, Byrne et al. (1997) concentrated on the interpretation of mantle phases along both lines. A summary model for both lines F and G is included in a fence diagram covering the seismic-refraction/wide-angle reflection interpretations for both KRISP90 and KRISP94 (Khan et al., 1999; Fig. 9.5.1-08). Beneath both lines the crystalline crust could be divided into three principal layers with major boundaries at 10–15 km and 20–30 km depth. In the upper crust the P-wave velocities were found to gradually increase from 6.0 to 6.2 km/s under the western flank to 6.3–6.4 km/s underneath the area of the rift. The middle crust with velocities of around 6.5 km/s appeared thickest under the

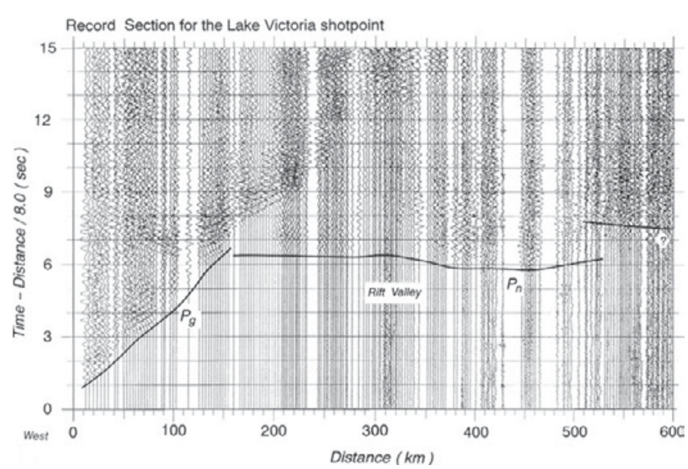



Figure 9.5.1-07. KRISP94 data example. Record section of shotpoint VIS (from Prodehl et al., 1997b, fig.5). [Tectonophysics, v. 278, p. 121–147.] Copyright Elsevier.]

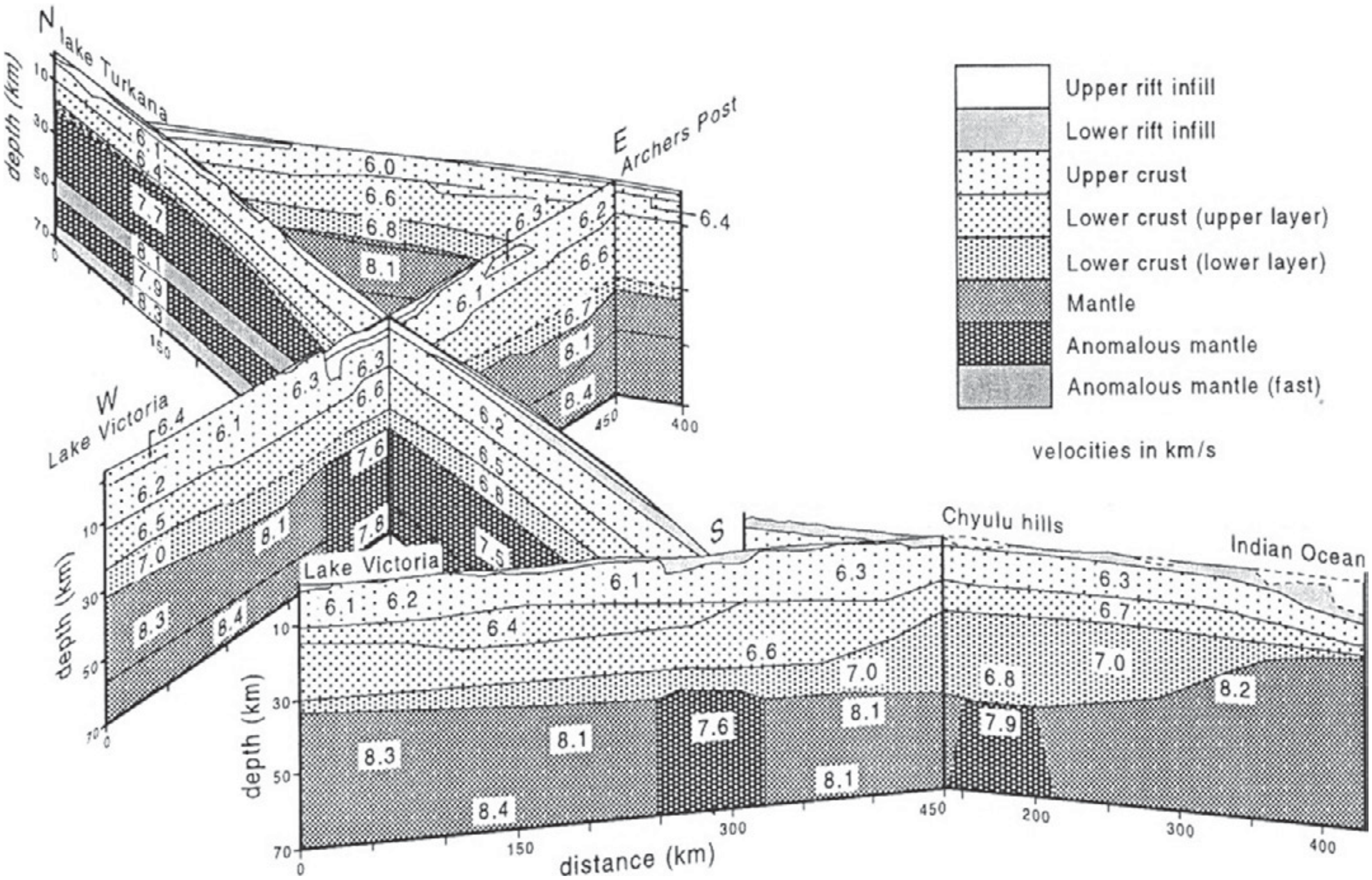



Figure 9.5.1-08. KRISP90/94 fence diagram showing crustal and upper mantle structure of the Kenya rift from Lake Turkana to Lake Magadi and beneath the neighboring flanks from Lake Victoria to the Indian Ocean (from Khan et al., 1999, fig. 4). [*In* Mac Niocaill, C., and Ryan, P.D., eds., Continental tectonics: Geological Society of London Special Publication 164, p. 257–269. Reproduced by permission of Geological Society Publishing House, London, U.K.]

western flank, where it might be subdivided into two layers. The lower mafic crust with velocities of around 7 km/s was thin near Lake Victoria, while under Lake Magadi the seismic velocity structure appeared similar to that found under the northern portion of the Kenya dome.

The dramatic thickening of the lower-crustal layer to more than 20 km eastward underneath the Chyulu Hills was unexpected, where it constitutes about half of the whole crust which here thickened to 44 km. However, the Moho here appeared as a diffuse crust-mantle transition zone, the lower limit of which could hardly be defined. Therefore, in agreement with the teleseismic modeling, a low-velocity zone was introduced into the lowermost section of the lower crust (Figs. 9.5.1-09 and 9.5.1-10). Furthermore, Novak et al. (1997b) interpreted the S-wave data and calculated $V_p/V_s$ ratios for all layers. Finally, Novak et al. (1997b) developed an integrated model for the deep structure of the Chyulu Hills volcanic field in southeastern Kenya, based on the seismic-refraction P- and S-data and on the tomographic modeling of the teleseismic data, as well as on available petrological data (Fig. 9.5.1-09).

The state of the art of crust and upper mantle research for the Dead Sea transform, Red Sea area, and the East African rift system was summarized in a compilation of all crustal cross sections published until 1994 (Prodehl et al., 1997a) from all controlled-source seismic experiments since the late 1960s touching the Afro-Arabian rift system. A collection of representative crustal columns indicated crustal thickness variations viewed as a function of rift development (Fig. 9.5.1-11).

## 9.5.2. The East African Rift System in Tanzania

Following the successful completion of crustal and upper mantle research in the eastern branch of the East African rift system in Kenya, the international Working Group discussed various possibilities how to continue their research work into Tanzania and Malawi.

This was partly stimulated by successful research work of the shallow subsurface structures in East African lakes by Duke University under the leadership of Bruce Rosendahl and co-workers in the 1980s (see, e.g., Rosendahl et al., 1989), but

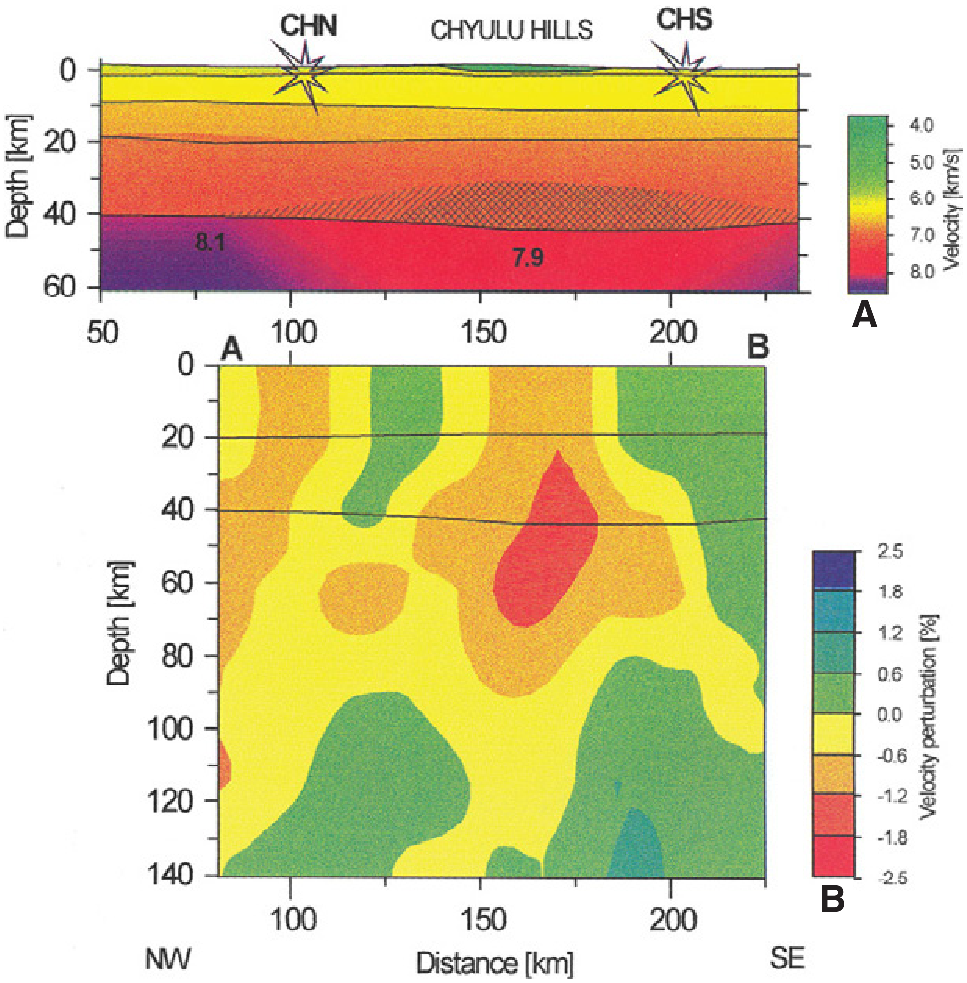



Figure 9.5.1-09. Crust and upper mantle structure underneath the Chyulu Hills volcanic field (from Novak et al., 1997, fig. 7). (A) Part of the crustal model along line F (see Fig. 9.5.1-06). (B) Cross section through the teleseismic tomography model of Ritter and Kaspar (1997). [Tectonophysics, v. 278, p. 187–209. Copyright Elsevier.]

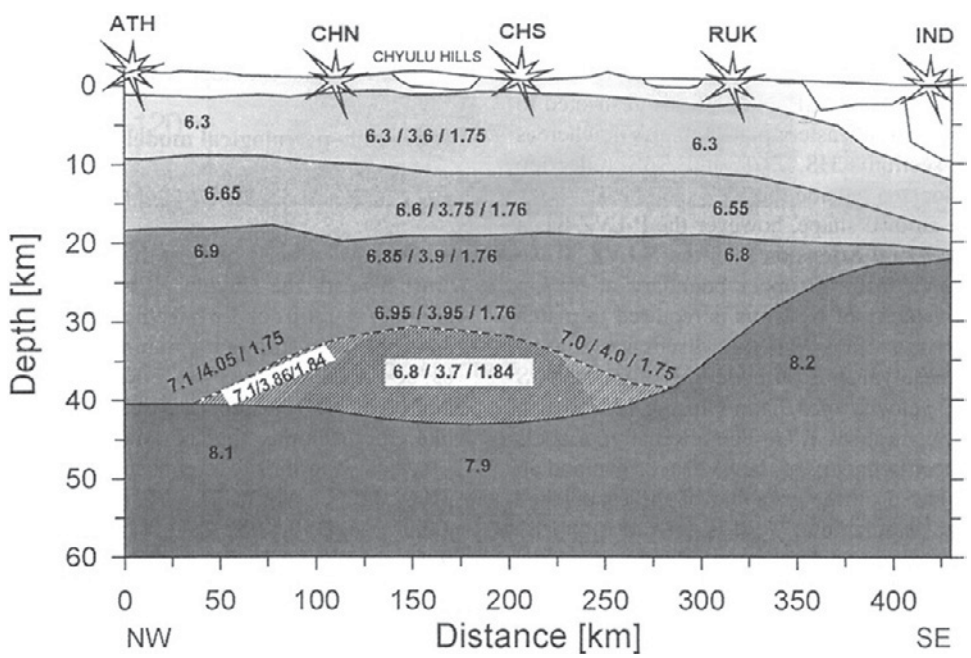



Figure 9.5.1-10. Crust and upper mantle structure underneath the eastern flank of the Kenya rift between Nairobi and Mombasa (from Novak et al., 1997, fig. 6). [Tectonophysics, v. 278, p. 187–209. Copyright Elsevier.]

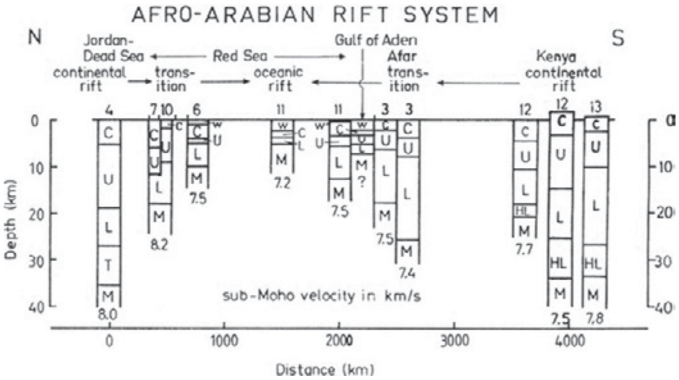



Figure 9.5.1-11. Evolutionary sequence illustrating the variation in crustal thickness under the Afro-Arabian rift system from the Dead Sea transform system through the Red Sea, Gulf of Aden and Afar triangle to the southern end of the Kenya rift (from Prodehl et al., 1997a, fig. 4). Depths refer to sea level. W—water; C—cover rocks; U—upper crust; L—lower crust; HL—high-velocity lower crust; T—crust-mantle transition rocks; M—Moho. [Tectonophysics, v. 278, p. 1–13. Copyright Elsevier.]

finally plans were dropped due to the lack of likely sources for funding and the nonexistence of local scientists or institutions as in Kenya. Nevertheless, experimental work in East Africa continued with large-scale teleseismic tomography surveys. Using the IRIS/PASSCAL broadband seismic array, in 1994 and 1995, teleseismic experiments were undertaken in Tanzania, in 2001–2002 in southern Kenya, and in 2000–2002 in Ethiopia (Fig. 9.5.2-01). These surveys, however, did not resolve any details of crustal and uppermost mantle structure but provided a telescoping view of the East African rift within this suspected plume province (Nyblade and Langston, 2002). In Tanzania, broadband stations were spread along two lines throughout the country (Nyblade et al., 1996). The array in southern Kenya followed approximately the KRISP94 crustal refraction survey (Nyblade and Langston, 2002; Fig. 9.5.2-01).

A small-scale experiment should also be mentioned, when from 1992 to 1994 a seismic network of five digital stations operated along the Rukwa trough in southwestern Tanzania near the border to Malawi, recording deep crustal earthquakes (Camelbeeck and Iranga, 1996). One hundred ninety-nine events were successfully recorded, and record sections were constructed for parts of the data. A crustal thickness of 42 km was determined from the arrivals identified as $P_MP$ reflections.

### 9.5.3. The Jordan–Dead Sea Transform

Crustal structure studies in the Jordan–Dead Sea transform, which had first been undertaken independently on the western side in Israel in 1978 (Ginzburg et al., 1979a) and on the eastern side in Jordan in 1984 (El-Isa et al., 1987a) and for which results for both sides had been summarized by Mechie and Prodehl (1988), were revived in 2000. For the first time a seismic profile could be established which covered both sides of the transform (Fig. 9.5.3-01).

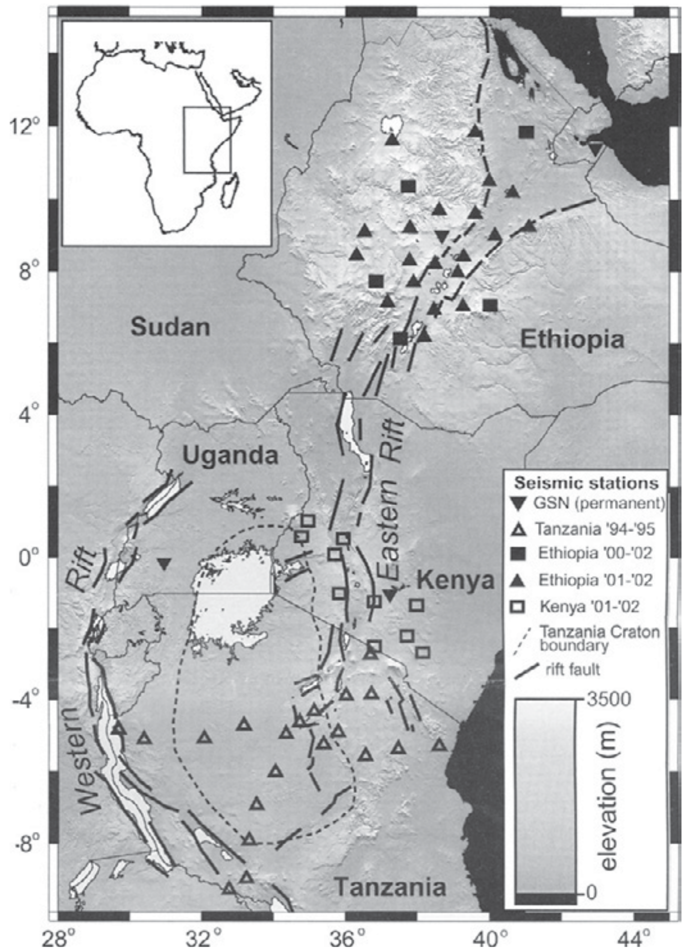



Figure 9.5.2-01. Topographic map of East Africa, showing the major rift faults of the East African rift system and the locations of broadband seismic stations (from Nyblade and Langston, 2002, fig. 1). [Eos (Transactions, American Geophysical Union), v. 83, no. 37, p. 405, 408–409. Reproduced by permission of American Geophysical Union.]

The seismic line started at the Mediterranean Sea south of Gaza, crossed the Jordan–Dead Sea transform in the Arava Valley south of the Dead Sea, and terminated in southern Jordan southeast of Ma'an (DESERT Team, 2000; DESERT Group, 2004; Mechie et al., 2005).

The project consisted of two parts: a seismic-refraction/wide-angle reflection line of 260 km length and a near-vertical incident seismic-reflection line of 100 km length. The project started with the refraction line. The 260-km-long NW-SE–trending line passed through Palestine, Israel, and Jordan and crossed the Arava fault ~70 km south of the Dead Sea (Figs. 9.5.3-01 and 9.5.3-02).

Thirteen shots, including two quarry blasts (nos. 10 and 31 in Fig. 9.5.3-01) were executed. The quarry blasts had charges of 8500 kg and 12000 kg. Five of the 11 borehole shots ranged from 720 to 950 kg, and the other six shots were smaller with charges of 30–80 kg. All shots were recorded by 99 three-component instruments, spaced at 1 km in the Arava Valley, at

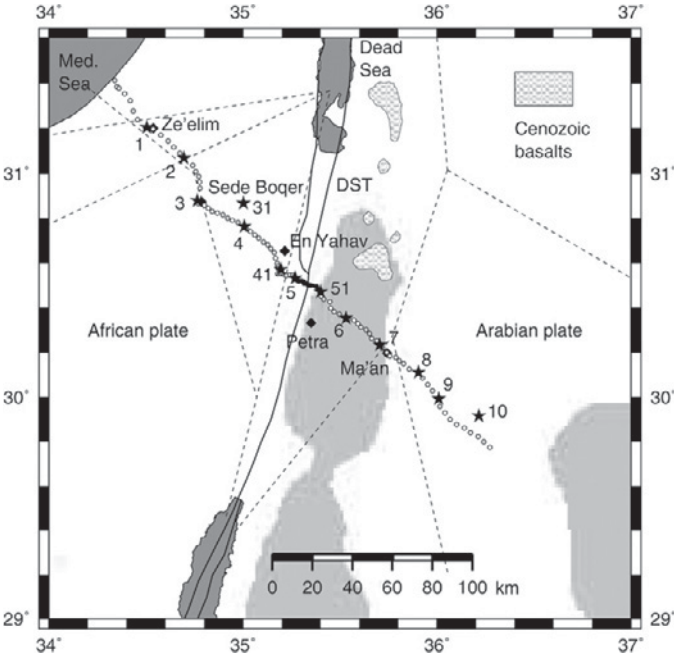



Figure 9.5.3-01. Location map of DESERT 2000 (from Mechie et al., 2005, fig. 1). Stars—shotpoints; circles—seismic recording sites. Dashed lines are earlier seismic-refraction lines (see Mechie and Prodehl, 1988). [Geophysical Journal International, v. 160, p. 910–924. Copyright John Wiley & Sons Ltd.]

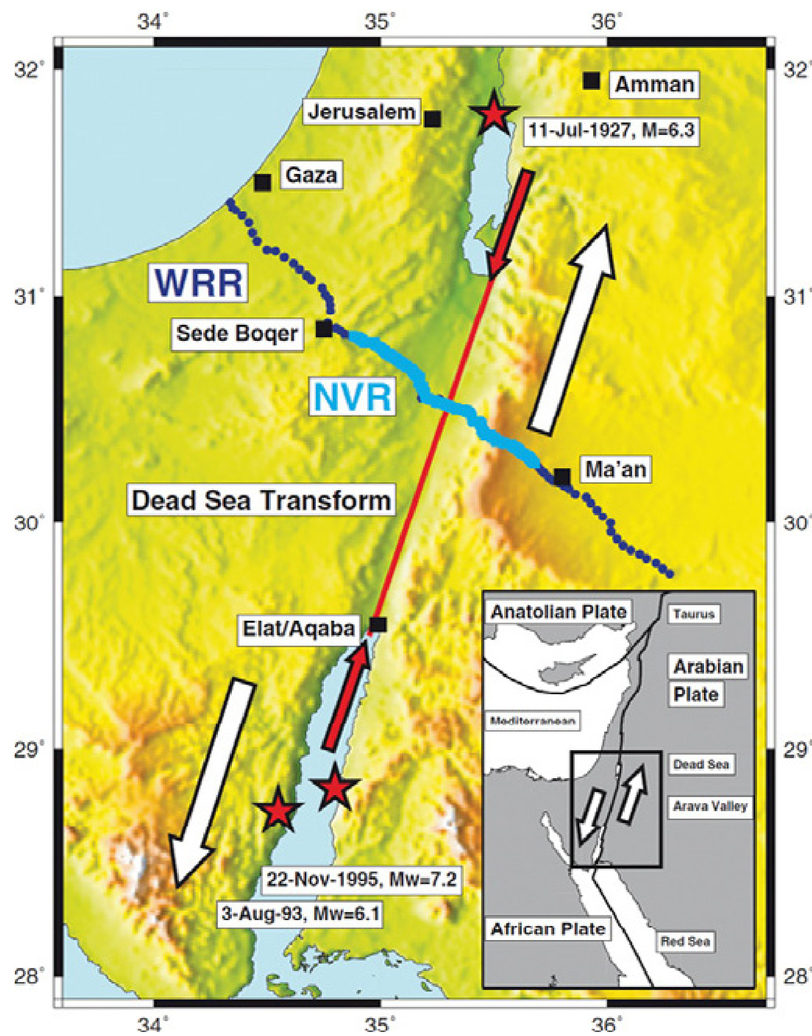



Figure 9.5.3-02. Topographic map of the Jordan–Dead Sea rift showing the location of DESERT 2000 (from DESERT Group, 2004, fig. 1a). Light blue dots—near-vertical seismic reflection line; dark blue dots—extension of reflection line for wide-angle reflection/refraction observations. White arrows indicate the relative movement of plates; red line and arrows indicate the Dead Sea Transform between Dead and Red Seas. [Geophysical Journal International, v. 156, p. 655–681. Copyright John Wiley & Sons Ltd.]

2.5 km on the adjacent flanks between Sede Boqer and Ma'an, and 4–4.5 km apart along the remaining parts of the line. All shots were recorded by 125 vertical-component geophone groups with 100 m spacing along a 12.5-km-long section of the profile on the Jordanian side of the Arava Valley between shots 5 and 51. A data example is shown in Figure 9.5.3-03 (for more data, see Appendix A9-3).

The study provided the first whole-crustal image across the Dead Sea transform (Fig. 9.5.3-04B). While for the seismic basement an offset of several kilometers was interpreted underneath the Arava fault, the main fault of the Dead Sea transform, the Moho depth increased steadily from ~26 km at the Mediterranean Sea to ~39 km under the Jordan highlands, except for a small asymmetry under the Arava valley (DESERT Group,

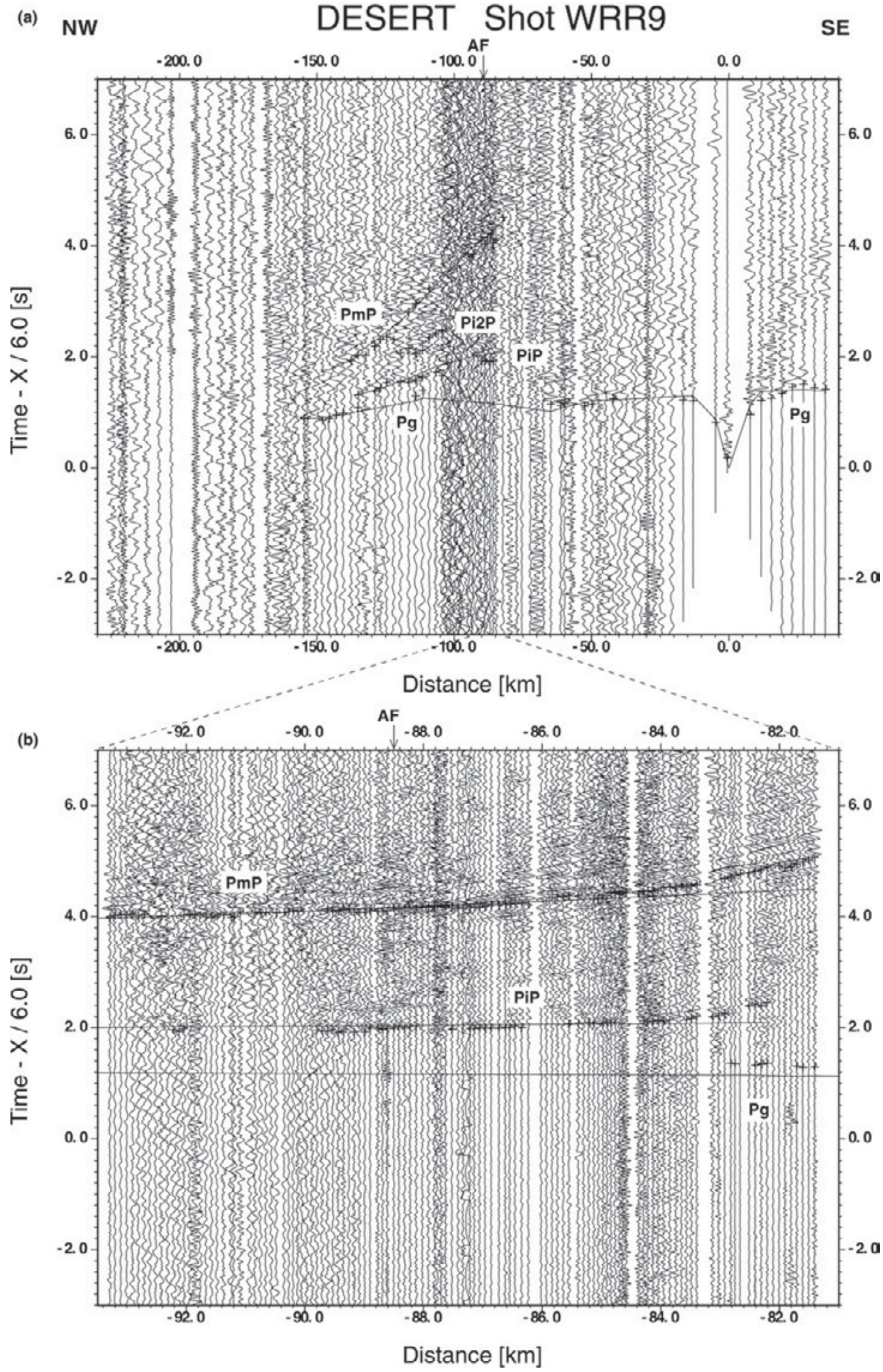



Figure 9.5.3-03. Record section of shot 9 along the DESERT WRR profile across the Jordan–Dead Sea rift (from DESERT Group, 2004, fig. 8). For location, see Fig. 9.5.3-01. Top: P-wave data along the whole length of the line. Bottom: P-wave data along the 12.5-km-long segment in the Arava valley with densely spaced vertical geophone groups. [Geophysical Journal International, v. 156, p. 655–681. Copyright John Wiley & Sons Ltd.]

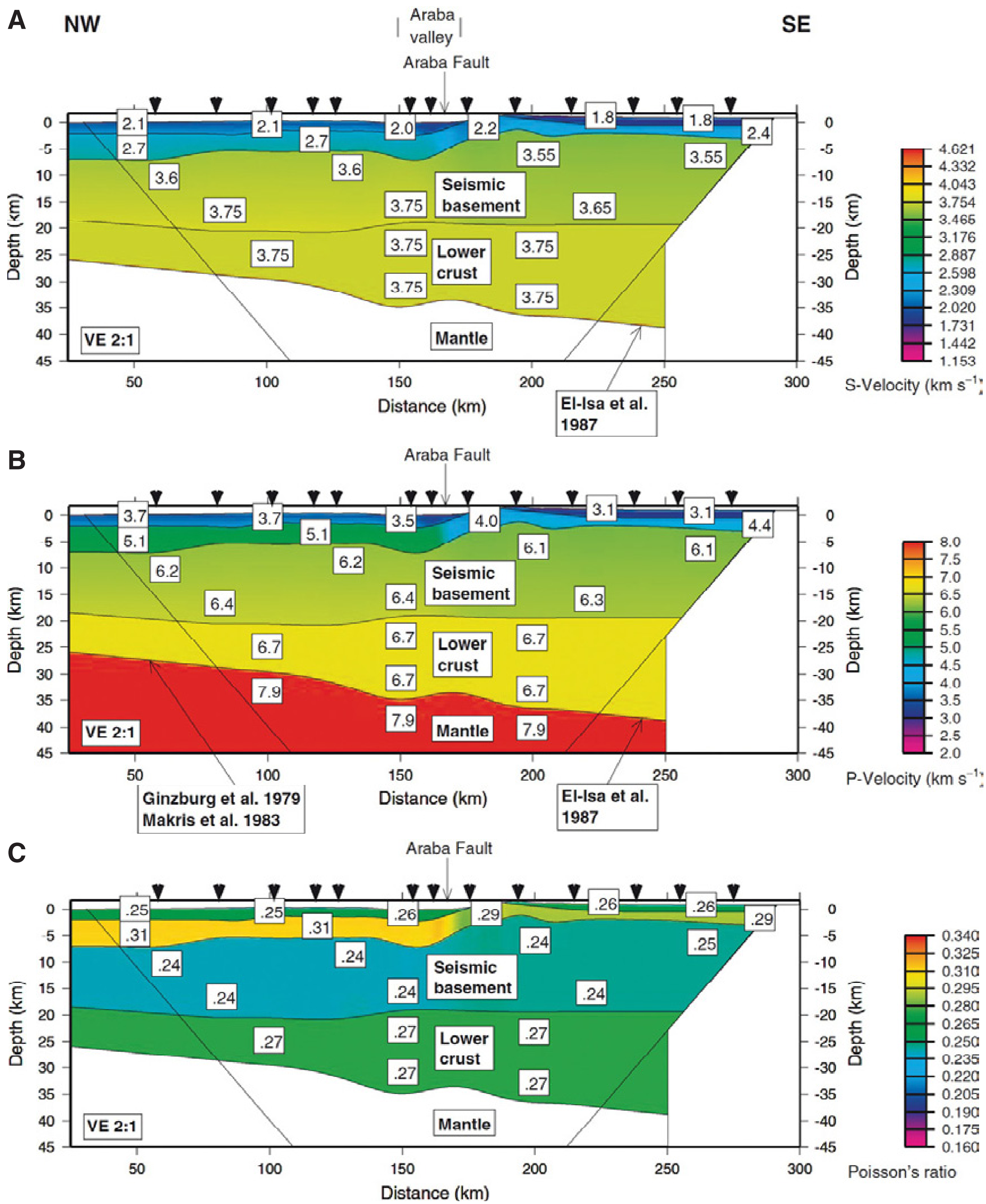



Figure 9.5.3-04. (A) S-wave model. (B) P- wave model. (C) Poisson's ratio model of the DESERT profile across the Jordan–Dead Sea rift (from Mechie et al., 2005, fig. 10). For location, see Fig. 9.5.3-02. [Geophysical Journal International, v. 160, p. 910–924. Copyright John Wiley & Sons Ltd.]

2004). Under the Arava Valley, no significant uplift of the Moho was noted. From the near-vertical data it could be inferred that the Arava fault cuts through the crust, develops into a broad zone in the lower crust and reaches down to Moho level.

Based on the excellent quality of the observed P- and S-wave data, a more detailed interpretation (Mechie et al., 2005) specialized on the S-wave data of the seismic-refraction/wide-angle profile (Fig. 9.5.3-04A), resulting in Poisson's ratios of 0.24 for most of the upper crust and 0.27 for the lower crust (Fig. 9.5.3-04C).

## 9.6. DEEP SEISMIC SOUNDING STUDIES IN SOUTHEAST ASIA

### 9.6.1. India and Indonesia

Crustal and upper mantle research in India has been almost exclusively performed by the National Geophysical Research Institute at Hyderabad. Scientists from this institution have visited all leading research institutions around the world and over a few decades a very powerful research team has evolved.

Following the example of COCORP in the United States and equivalent organizations in many West European countries in the 1980s promoting deep crustal reflection work in large dimensions, seismic-reflection investigations on a large scale were performed in India (Rajendra Prasat et al., 1998; Reddy et al., 1995; Tewari et al., 1995, 1997), aiming to unravel the structure and tectonics of the Aravalli-Delhi fold belt in northwestern India. For example, a 200-km-long deep seismic-reflection profile across the northwestern Indian shield is described in more detail (Fig. 9.6.1-01).

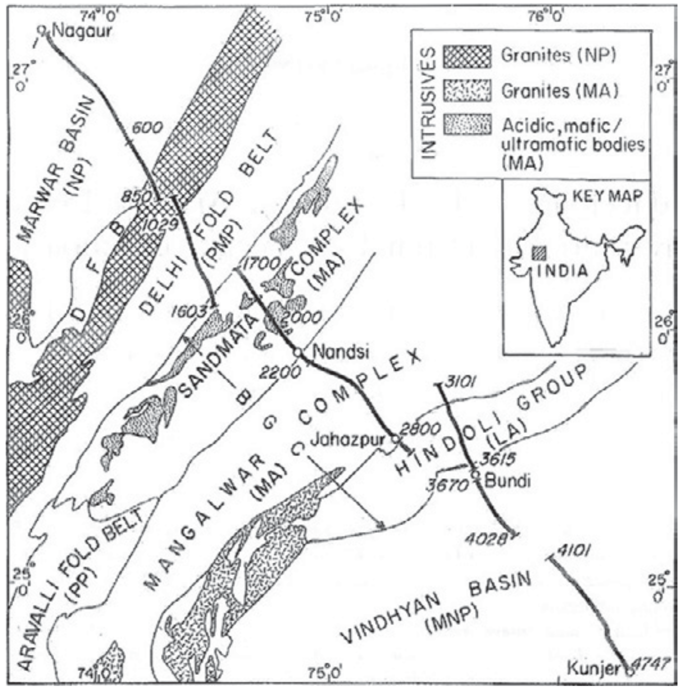



Figure 9.6.1-01. Location map of the Nagaur-Kunjer deep reflection profile, traversing the northwestern Indian shield (from Rajendra Prasat et al., 1998, fig. 1). [Tectonophysics, v. 288, p. 31–41. Copyright Elsevier.]

The seismic-reflection profile ran from the Marwar basin across the Proterozoic Delhi Fold Belt and gneissic complexes into the Vindhyan basin. Field parameters included a 20 s TWT record with 60-fold theoretical CDP coverage using 120-channel data at a 4 ms sampling rate and geophone interval/shot interval of 100 m. The actual fold coverage, however, varied between 40 and 45. The energy source was 30–35 kg of explosives loaded at 15–20 m depth. The Delhi Fold Belt appeared to be a zone of thick (45–50 km) crust that might have been generated due to the doubling of the lower crust during compressional Proterozoic tectonics. This resulted in three Moho reflection bands, two of which dip SE from 12.5 to 15.0 s TWT and from 14.5 to 16.0 s TWT. Another band of subhorizontal Moho reflections at ~12.5 s TWT was attributed to the post-Delhi tectonic orogeny. The signatures of the Proterozoic collision have been preserved in the form of strong SE-dipping reflections in the lower crust and Moho. The lower crust appeared fairly reflective in the Delhi Fold Belt, but poorly reflective under the gneissic complexes, where the base of the crust was estimated to be at 50 km, based primarily on gravity and not on reflection data (Tewari et al., 1997).

The deep-crustal reflection data recorded along the traverse were projected onto a cartoon (Fig. 9.6.1-02) imaging the two sets of oppositely dipping reflection bands from upper to lower

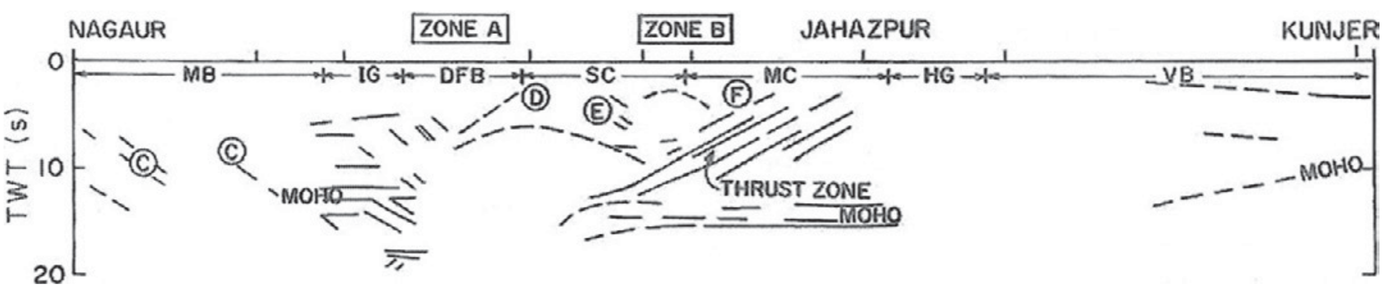



Figure 9.6.1-02. Cartoon showing the prominent reflections along the Nagaur-Kunjer deep reflection profile, traversing the northwestern Indian shield (from Rajendra Prasat et al., 1998, fig. 3a). [Tectonophysics, v. 288, p. 31–41. Copyright Elsevier.]

crustal levels correlated with the collision leading to the development of the Aravalli-Delhi fold belt in the Proterozoic (Rajendra Prasat et al., 1998).

The Indian crustal structure and Moho configuration beneath the Himalaya was investigated with teleseismic recording experiments (Mitra et al., 2005; Rai et al., 2006) the locations of which are shown in the next subchapter (Fig. 9.6.2-02). Mitra et al. (2005) reported on the project NE INDIA in northeastern India, while the NW Himalaya was studied by the project HIMPROBE extending from India into Tibet (Rai et al., 2006). A teleseismic receiver function analysis of seismograms recorded on a 700-km-long profile of 17 broadband seismographs traversing the NW Himalaya showed a progressive northward deepening of the Indian Moho from 40 km beneath Delhi south of the Himalayan foredeep to 75 km beneath Taksha at the Karakoram fault (Rai et al., 2006). Similar studies by Wittlinger et al. (2004) to the north of the Karakoram fault indicated that the Moho may continue to deepen to as much as 90 km depth beneath western Tibet before shallowing substantially to 50–60 km at the Altyn Tagh fault, which, however, is not shown in the Moho contour map of China (see Fig. 10.5.2-01), compiled by Li et al. (2006).

In southern India, new deep seismic-reflection and refraction data did not become available. However, following the devastating 1993 Latur earthquake in southern India (~18°N, 76°30′E), a mobile network of 45 seismic stations was operated for aftershock activity in cooperation with the GeoForschungs-Zentrum Potsdam, Germany. The data could be used to plot record sections of well-located aftershocks out to 80 km offsets and deduce details of crustal structure. The result was a sequence of alternating high- and low-velocity layers for P- and S-waves in the upper crust at depths of 6.5–9 km and 12.3–14.5 km with ~7% velocity reduction and a low-velocity layer at 24–26 km depth (Krishna et al., 1999).

In 1998, a small-scale seismic project was undertaken on Java, Indonesia, to investigate the structure of an active volcano. The German institution GFZ Potsdam studied the seismic structure of the stratovolcano Merapi (Java, Indonesia) by means of a controlled-source seismology experiment, using 90 three-component stations and a specially devised airgun source. The project concentrated on the heterogeneous structure of the volcano within a radius of 5 km from the active dome, where the sources of most of the natural volcanic seismic events are located, because the seismic properties of the propagation media in an active volcano is important to understand the natural seismic signals used for eruption forecasting. Three 3-km-long seismic profiles, each consisting of up to 30 three-component seismometers with a station interval of 100 m, were deployed in an altitude range between 1000 m and 2000 m above sea level. The distance from the highest stations to the active summit region at an altitude of almost 3000 m was ~5 km. As energy sources airgun sources were used at the lower end of each of the three profiles. The sources consisted of water basins equipped with a 2.5-l mudgun. Airgun shots were released every 2 min, so that at each seismometer the repeated shots were recorded between 20 and 120 times. The modeling of the data was described in detail by Wegler and Lühr (2001). The observed seismograms showed a spinlep-like amplitude increase following the direct P-phase. This shape of the envelope could be explained by the diffusion model. According to this model there were so many strong inhomogeneities that the direct wave could be neglected and all energy was regarded as being concentrated in multiple scattered waves. The natural seismic signals at Merapi volcano showed similar characteristics to the artificial shots. The first onsets had only small amplitudes and the energy maximum arrived delayed compared to the direct wave. Therefore it can be concluded that these signals are also strongly affected by multiple scattering.

### 9.6.2. China

The large-scale seismic investigations of the crust and upper mantle of China with explosion seismology of the 1980s were continued even more intensively in the 1990s. Summary papers representing the current state of the art were published from time to time (e.g., Li and Mooney, 1998; Li et al., 2006) presenting DSS profile locations in mainland China (Fig. 9.6.2-01), many of which were completed by the Research Center of Exploration Geophysics (RCEG) and China Earthquake Administration (CEA, previously known as the China Seismological Bureau or CSB).

Besides the location map, a list of all profiles, shown as red lines in Figure 9.6.2-01, contains details of the RCEG/CEA lines (Li et al., 2006), such as location names, length in km, azimuth, number of shots, and year. From this list it can be noted that the RCEG/CEA lines 1–25 were completed between 1980 and 1989, except for nos. 9–10 and 12–14 which were already observed from 1977 to 1979, while lines 26–46 were shot in the 1990s including the year 2000, and lines 47–50 in the early 2000s. The list also contains information on the seismic profiles completed by other institutions, such as location names and corresponding references for each profile A1 to A41 shown in Fig. 9.6.2-01 as green lines. The corresponding tectonic sketch map of the main geological provinces was also shown (see, Figs. 8.7.2-01 and 10.5.2.04). Furthermore, Li and Mooney (1998) and Li et al. (2006) have summarized the main results in crustal columns with average crustal structure for the main tectonic areas as well as in contour maps, showing crustal thickness (Moho depths) and uppermost-mantle velocities ($P_n$-velocities).

Figure 9.6.2-01 does not show, however, lines which had been organized by international collaborations in Tibet such as the Sino-French lines recorded in the 1980s or all the INDEPTH lines observed in the 1990s (Figs. 9.6.2-02 and 9.6.2-03).

It was Tibet in particular that attracted scientists from all over the world. In the 1980s, major collaborations of Chinese and French scientists had stimulated major seismic-refraction investigations (e.g., Hirn et al., 1984c; Sapin et al., 1985), but with modern technology, even major seismic-reflection work became possible in the 1990s.

For the area of Tibet and surroundings, Klemperer (2006) has investigated all major seismic and magnetotelluric experiments

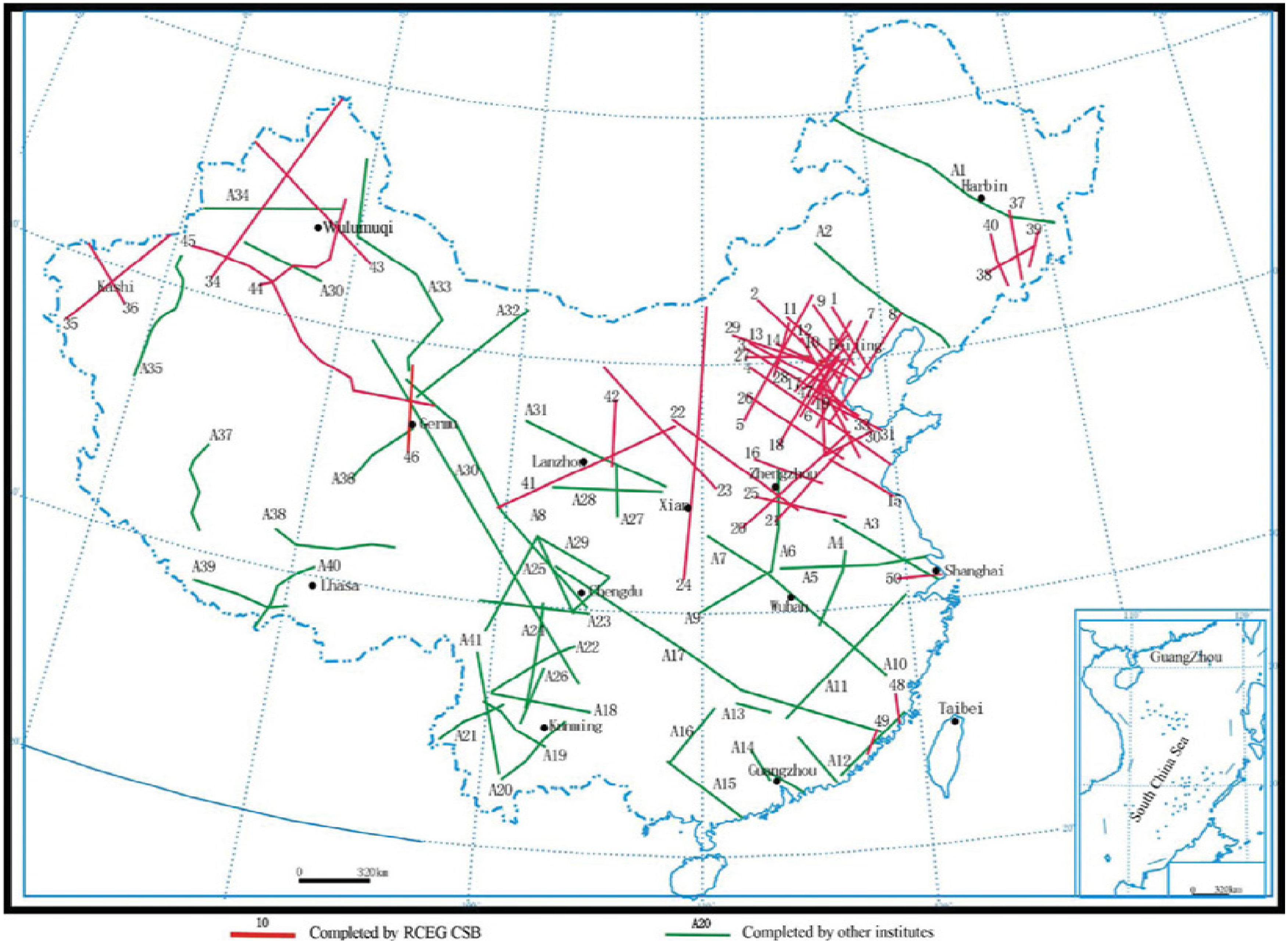



Figure 9.6.2-01. Location map of deep seismic sounding profiles in mainland China (from Li et al., 2006, fig. 1). Profiles 1–45 were completed by RCEG and CEA, profiles A1–A41 were completed by other institutions. [Tectonophysics, v. 420, p. 239–252. Copyright Elsevier.]

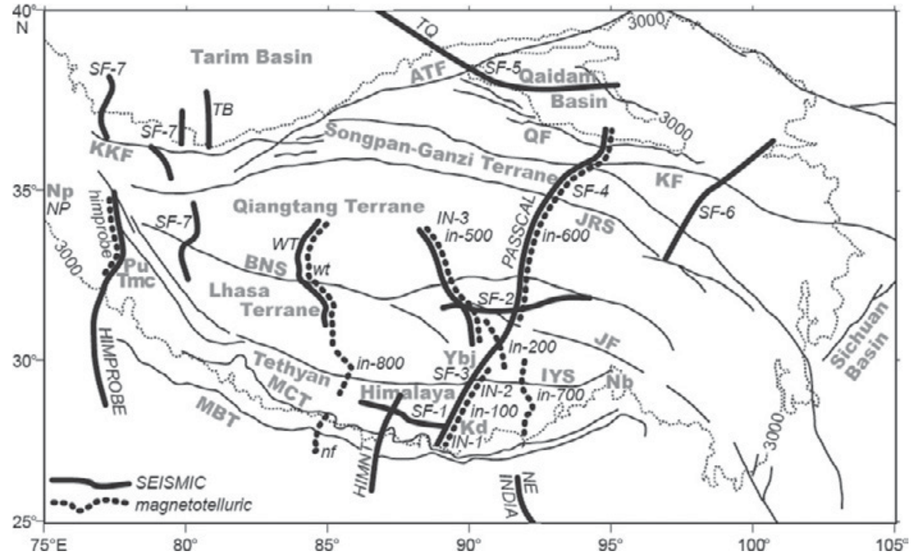


Figure 9.6.2-02. Location map of main geophysical experiments in and around Tibet (from Klemperer, 2006, fig. 1). Heavy black lines and upper-case names are generalized locations of seismic experiments. Heavy dotted lines are magnetotelluric profiles. Thin lines are major faults and sutures. Thin dotted line is 3000 m elevation contour. INDEPTH: IN-1 (Zhao et al., 1993; Makovsky et al., 1996), IN-2 (Nelson et al., 1996; Makovsky and Klemperer, 1999), IN-3 (Zhao et al., 2001; Ross et al., 2004). Sino-French: SF-1 (Hirn et al., 1984c), SF-2 (Zhang and Klemperer, 2005), SF-3 (Hirn et al., 1995; Galvé et al., 2002), SF-4 (Herquel et al., 1995; Vergne et al., 2003), SF-5 (Wittlinger et al., 1998), SF-6 ( Galvé et al., 2002; Vergne et al., 2003), SF-7 (Wittlinger et al., 2004). PASSCAL (e.g., McNamara et al., 1996): WT—West Tibet (Kong et al., 1996), TB—Tarim Basin (Gao et al., 2000; Kao et al., 2001), NB—Nanga Parbat (Meltzer et al., 2001). HIMPROBE (Rai et al., 2006): TQ—Tarim-Qaidam (Zhao et al., 2006), HIMNT—Himalaya-Nepal-Tibet experiment (Schulte-Pelkum et al., 2005), NE INDIA (Mitra et al., 2005). [*In* Law, R.D., Searle, M.P., and Godin, L., eds., Channel flow, ductile extrusion and exhumation in continental collision zones: Geological Society of London Special Publication 268, p. 39–70. Reproduced by permission of Geological Society Publishing House, London, U.K.]

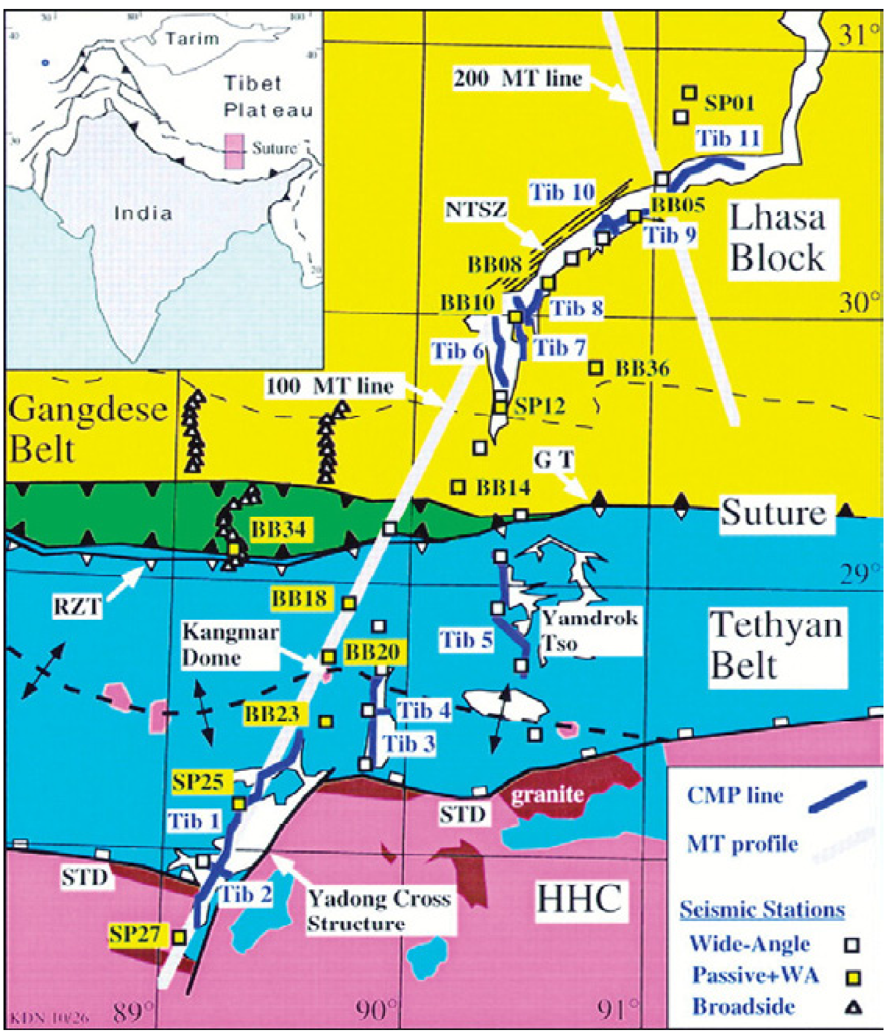



Figure 9.6.2-03. INDEPTH I-II location map (from Nelson et al., 1996, fig. 1). SP—short period stations, BB—broadband stations for both active-source wide-angle and passive-event recording. [Science, v. 274, p. 1684–1688. Reprinted with permission from AAAS.]

with respect to crustal flow in Tibet and has published a location map of all experiments in and around Tibet including the corresponding references (Fig. 9.6.2-02). From the experiments cited, IN-1, IN-2, and locally along IN-3 were near-vertical reflection profiles, while controlled-source wide-angle profiles were SF-1, SF-2, SF-3, SF-6, IN-1, IN-2, IN-3, WT, and TQ. Teleseismic recording experiments were PASSCAL, IN-2, IN-3, SF-3, SF-4, SF-5, SF-6, SF-7, TB, NP, HIMPROBE, HIMNT, and NE INDIA.

One of the largest cooperative seismic projects in Tibet was INDEPTH (Figs. 9.6.2-03 and 9.6.2-05), involving both near-vertical incident and wide-angle seismic-reflection methodology. This project was accomplished with major U.S. and German participation in several phases, starting in 1992 (e.g., Brown et al., 1996; Makovsky et al., 1996; Nelson et al., 1996). Its southern part (IN-1, IN-2) is shown by line A40 of Figure 9.6.2-01. The cooperation with French institutions was also continued in northeastern Tibet (e.g., Galvé et al., 2002). Major cooperative work of Chinese institutions was accomplished and results published together with scientists of the USGS, e.g., in the Dabie Shan orogenic belt in eastern China (Wang et al., 2000; A33 in Fig. 9.6.2-01), or at the northern and northeastern margin of the Tibetan plateau (Zhao et al., 2006; Liu et al., 2006).

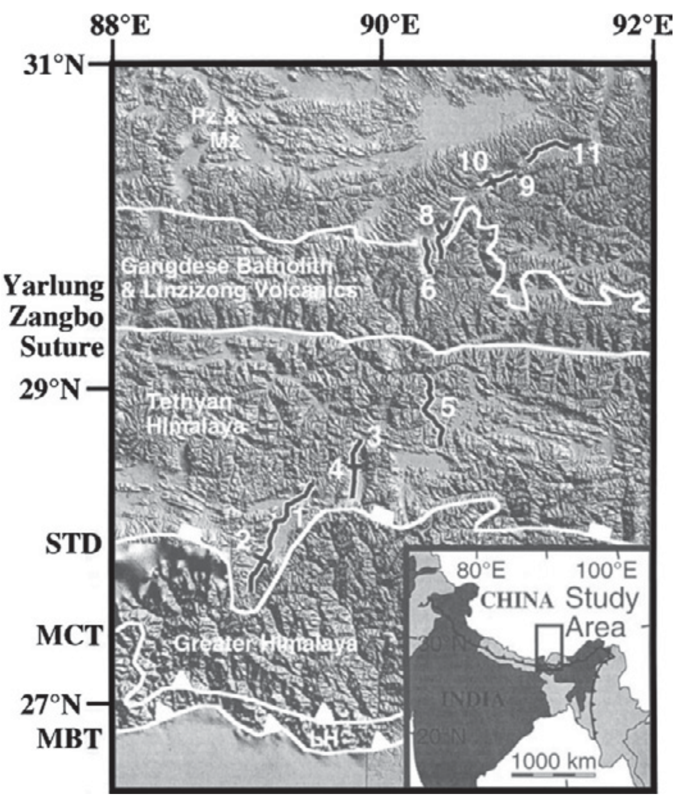



Figure 9.6.2-04. Topographic map showing the location of INDEPTH I and II seismic lines and major structural boundaries of the India-Tibet collision system (from Alsdorf et al., 1998, fig. 1). [Journal of Geophysical Research, v. 103, p. 26,993–26,999. Reproduced by permission of American Geophysical Union.]

The project INDEPTH (International Deep Profiling of Tibet and the Himalaya) was a geoscientific project to understand the crustal and upper mantle structure beneath the Himalaya and the Tibetan Plateau. It was carried out under the direction of a collaborative geoscience team involving scientists from the Chinese Academy of Geological Sciences and several North American and European institutions. The project acquired multichannel seismic-reflection (common midpoint or CMP data), wide-angle reflection/refraction, broadband earthquake, magnetotelluric, and surface geological data. In the first two phases, data were collected along a 400-km-long corridor, extending from the crest of the Himalaya to the center of the Lhasa block, and most of the data were collected within the Yadong-Gulu rift in southern Tibet (Figs. 9.6.2-03 and 9.6.2-04).

INDEPTH consisted of four phases. INDEPTH I was conducted in 1992 when pilot near-vertical incidence CMP reflection profiles of 100 km length in southern Tibet (Tib-1 and Tib-2) were recorded, with 50-m receiver-group spacing and 50-s listening time (Zhao et al., 1993). Along these profiles, companion wide-angle data were also collected.

INDEPTH II was carried out in 1994 and 1995 and involved a larger program with the acquisition of CMP reflection and wide-angle reflection profiles Tib-3 to Tib-11 in 1994 as well as collecting broadband earthquake, magnetotelluric, and surface geological data (Nelson et al., 1996; Alsdorf et al., 1998). As an addition, the densely spaced shots of the reflection lines were also recorded parallel to the main lines at offsets between 70 and 130 km (broadband stations in Fig. 9.6.2-03; GEDEPTH location in Fig. 9.6.2-05B).

During INDEPTH III, in 1998 a 400-km-long SSE-NNW wide-angle seismic profile was recorded in central Tibet (Fig. 9.6.2-05), extending from the central Lhasa block in NNW direction to the central Qiangtang terrane (Zhao et al., 2001; Ross et al., 2004). It consisted of a closely spaced (5–10 km) seismic array with 60 broadband and short-period stations which recorded 11 large shots with charges from 180 to 1160 kg (Zhao et al., 2001). In addition, a mobile 60-channel array also recorded at near-vertical angles of incidence four in-line groups and one cross-line group of 160 small shots with charges of 6–50 kg at variable offsets up to 16 km, two of them sampling the Lhasa block south of the Banggong-Nujiang suture, and two of them sampling the Qiangtang terrane farther north (Ross et al., 2004). Following the active-source experiment, 57 three-component stations, including eight offline stations, were left until 1999 to record local, regional, and far-distant events. This phase also included a magnetotelluric experiment.

INDEPTH IV was designed to study the northeastern boundary of the Tibetan Plateau as represented by the Kunlun Mountains and the Qaidam Basin. It started in spring 2007 and was planned to continue into 2009 (Brown, 2006).

The most prominent laterally extensive feature of the INDEPTH I seismic-reflection section was a "mid-crustal" reflection dipping gently to the north from 9 s to 13 s TWT. Using a range of velocities of 6.0–6.4 km/s as was provided by the

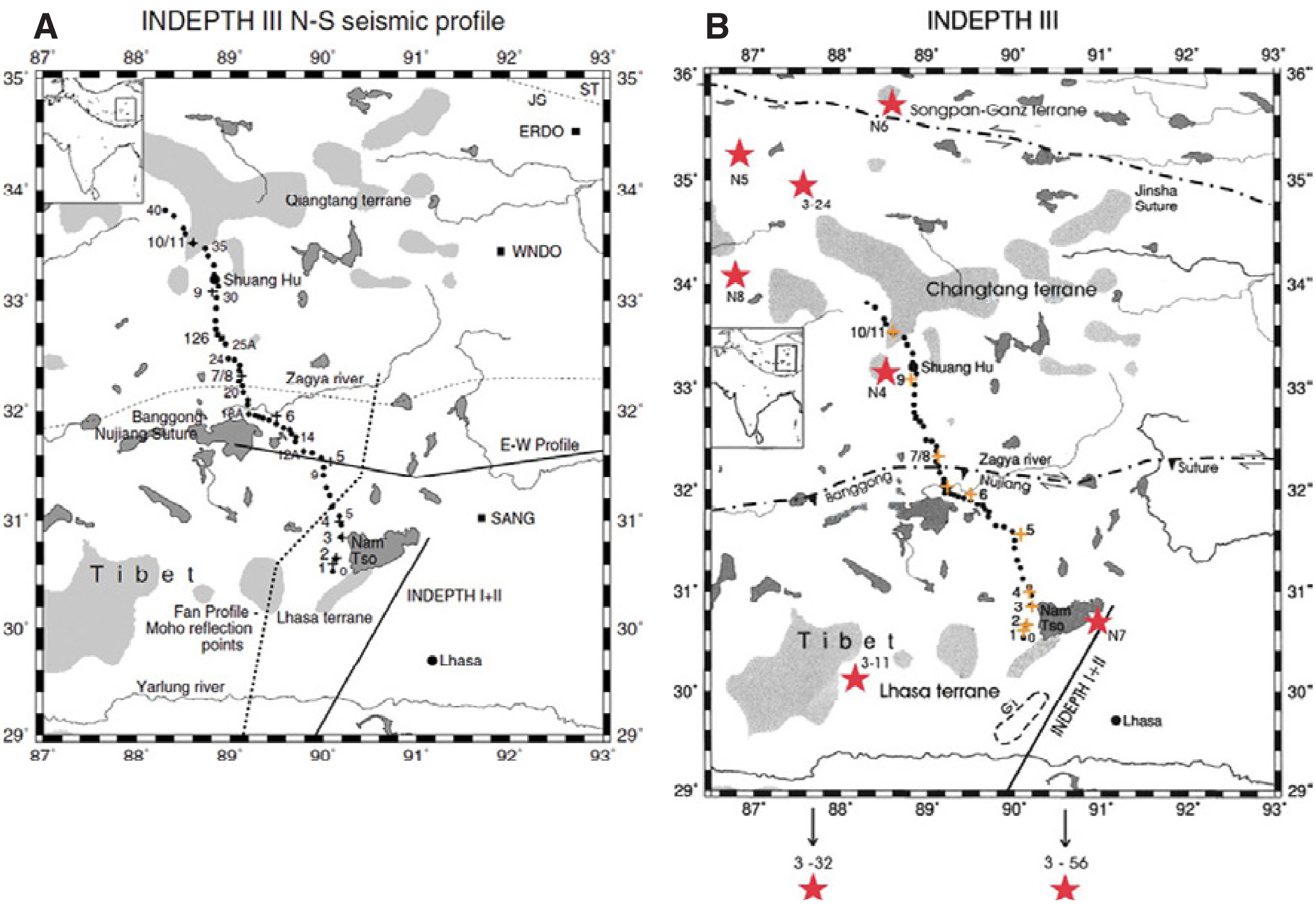



Figure 9.6.2-05. (A) INDEPTH III location map (from Zhao et al., 2001, fig. 1). Also shown are the locations of INDEPTH I and II and earlier profiles of Sino-French campaigns in the eighties: an E-W profile (Sapin et al., 1985) and Moho reflection points of a fan profile (dotted line, Hirn et al., 1984c; Fig. 9.6.2-04). [Geophysical Journal International, v. 145, p. 486–498. Copyright John Wiley & Sons Ltd.] (B) INDEPTH III and earthquake location map (from Meissner et al., 2004, fig. 1). G1—location of the wide-angle GEDEPTH recording area during INDEPTH I and II. [Tectonophysics, v. 380, p. 1–25. Copyright Elsevier.]

French-Sino refraction experiment in the Himalaya (Hirn et al., 1984b), for the southern end of the line a depth of 28 km and for the north a depth of 40 km was calculated. Wide-angle reflections recorded by RefTek stations deployed farther north, allowed to extend this horizon in the subsurface by at least 30 km beyond the north end of the reflection line. Comparison with existing earthquake and geological data let assume that this reflection could be interpreted as marking the thrust fault—termed the Main Himalayan Thrust—along which India is presently underthrusting Tibet. Another notable feature of this first INDEPTH project was the occurrence of coherent reflections at ~23 and 24 s TWT. These were interpreted as originating from the Moho at the base of the Indian continental crust. Using an average velocity of 6.3 km/s, a Moho depth of ~75 km results.

A particular emphasis of all INDEPTH projects was its near-vertical incidence reflection component. So far the most significant results of this component were summarized by Ross et al. (2004):

(1) The reflection Moho was found at 23 s TWT at the southern end of the transect (Zhao et al., 1993). (2) A prominent reflection marks the Main Himalayan Thrust, the active décollement along which the Indian plate is underthrusting Asia. (3) A series of extremely high-amplitude reflections or "bright spots" was detected at 15 km depth beneath the northern portion of the INDEPTH II profile (Brown et al., 1996; Alsdorf et al., 1998; Makovsky and Klemperer, 1999). (4) The short seismic-reflection sections from the four locations of INDEPTH III, spanning the Jurassic Banggong-Nujiang Suture, showed thin (<2 km) sedimentary cover sequences, an unreflective upper crust down to ~25 km depth, a strongly reflective lower crust down to 65 km depth (22 s TWT), and a distinct change in seismic character from layered reflectivity to an unreflective domain, that Ross et al. (2004) interpreted as marking the reflection Moho.

In order to collect wide-angle seismic reflections during INDEPTH I and II, the densely spaced shots of the reflection line were recorded at offsets up to 300 km along and parallel to the main line and thus provided an image of the crust straddling the Indus-Yarling suture. The major features were prominent reflections at ~20 km depth beneath and extending to ~20–30 km north and south of the surface expression of the suture, and north-dipping reflections north of the suture (Makovsky et al., 1999; Zhao et al., 1997).

The excellent data along the wide-angle seismic profile of INDEPTH III in central Tibet (Fig. 9.6.2-06; Appendix A9-4-1) revealed a crustal thickness of 65 km beneath the line (Zhao et al., 2001), in agreement with 70 km depth found by Sapin et al. (1985) just south of the Banggong-Nujiang suture along the E-W line shown in Figure 9.6.2-05A. However, the 20 km-step in the Moho postulated by Hirn et al. (1984c) from the fan-recordings, could not be confirmed by the new data.

Another major result of the INDEPTH III line was the evidence for a 25–30-km-thick lower crust with high velocities ranging from 6.5 to 6.6 km/s at its top to 7.2–7.3 km/s at its base and a velocity contrast of 0.7–0.8 km/s across the Moho (Fig. 9.6.2-07).

Zhao et al. (2001) have summarized all data available at that time in a sketch showing how Moho depths in Tibet decrease from south to north between 28°N and 36°N (Fig. 9.6.2-08). Further details of structural and tectonic features were detected by Meissner et al. (2004), who reinterpreted the INDEPTH III wide-angle data in conjunction with the observations of eight earthquakes (data examples in Appendix 9-4-1), recorded during the same time.

The Sino-French cooperation of the 1980s was continued at the end of the 1990s with a network of seismic-refraction profiles in northeastern Tibet on the Tibetan Plateau (Galvé et al., 2002; Appendix 9-4-2).

Six borehole shots with charges of 3–5 tons were recorded in three deployments of 250 seismometers with 5–10 km spacing on profiles reaching offsets of more than 250 km. The system of reversed and overlapping profiles extended from the North Kunlun crustal block to the Qang Tang terrane of central Tibet (Fig. 9.6.2-09).

The interpretation (Fig. 9.6.2-10) showed a drastic change of crustal structure when crossing the Kun Lun fault. In the south, a Moho-like high-velocity lid was obtained at 35 km depth, overlying a 37-km-thick lower crust with an average velocity of 6.5 km/s and a Moho at 72 km depth. Under the North Kunlun block the "35-km" reflection was absent. Here, a very low average velocity resulted for the more or less featureless crust and Moho depth was found at 64 km (Galvé et al., 2002). The data were further interpreted a few years later (Jiang et al., 2006; Galvé et al., 2006).

The northern margin of the Tibetan Plateau was investigated by an active-source seismic-refraction profile across the Altyn Tagh Range (Zhao et al., 2006). The line crossed the Qaidam Basin in E-W direction and continued in northwesterly direction into the Tarim Basin (Fig. 9.6.2-11). Ten shotpoints separated by 60–320 km provided the seismic energy by 2-ton-charges which were recorded by 240 three-component seismographs spaced at intervals between 1.5 and 3.0 km (Fig. 9.6.2-12; Appendix A9-4-2). The southeasternmost shotpoint (no. 10 in Fig. 9.6.2-11) was located ~200 km south of the southernmost shotpoint of the seismic-refraction survey of 1988 (Wang et al., 2003; see Chapter 8.7.2 and Appendix A9-4-2) which had been

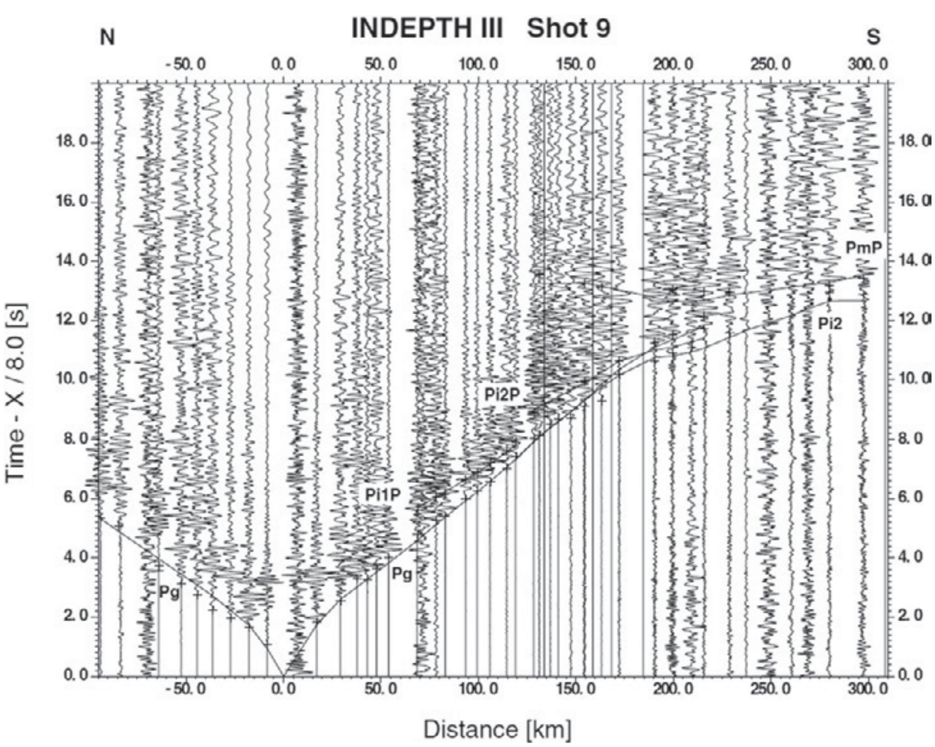



Figure 9.6.2-06. Data example of the INDEPTH-III profile across central Tibet, shot 9 (from Zhao et al., 2001, fig. 8). For location, see Fig. 9.6.2-05. [Geophysical Journal International, v. 145, p. 486–498. Copyright John Wiley & Sons Ltd.]

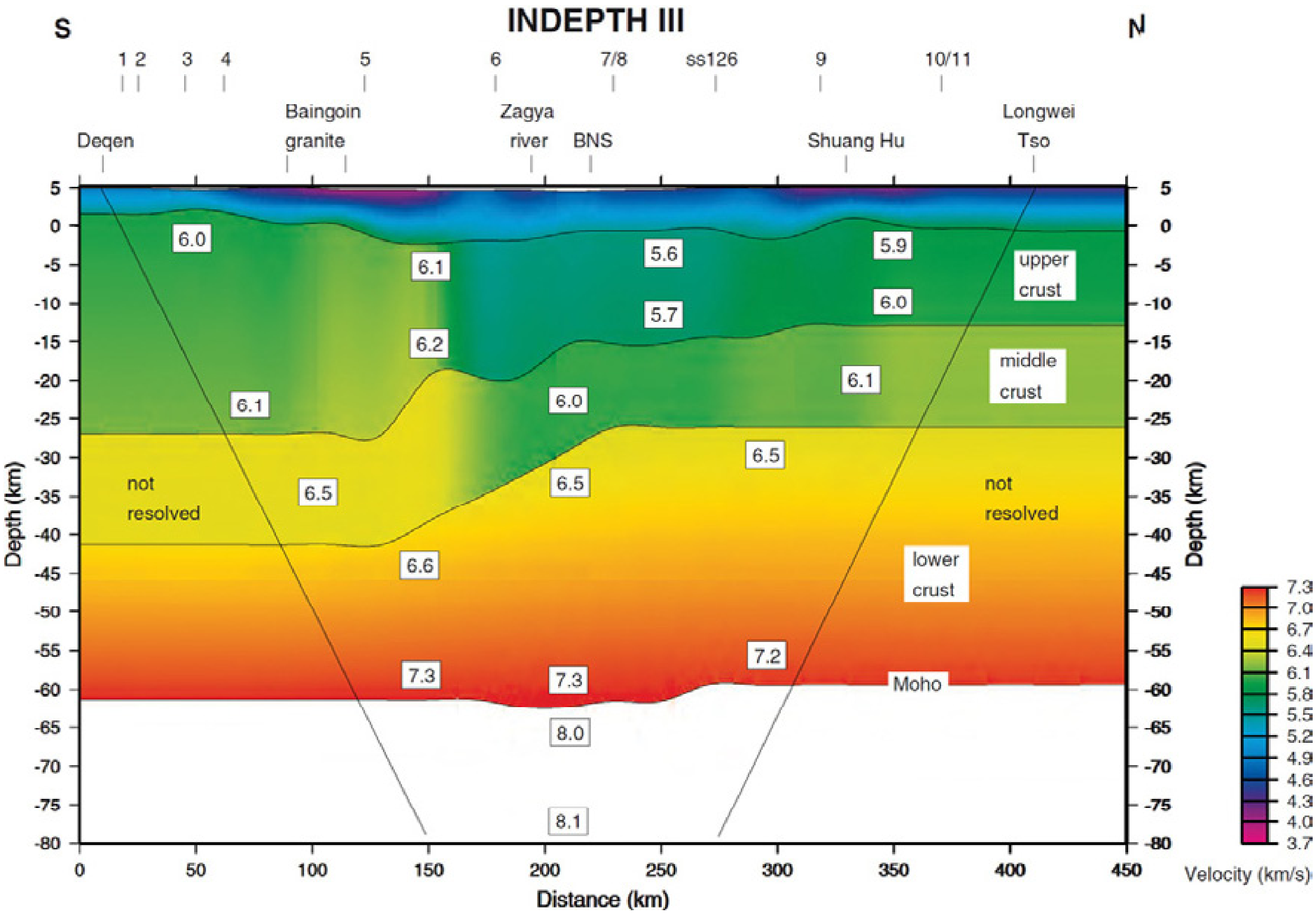
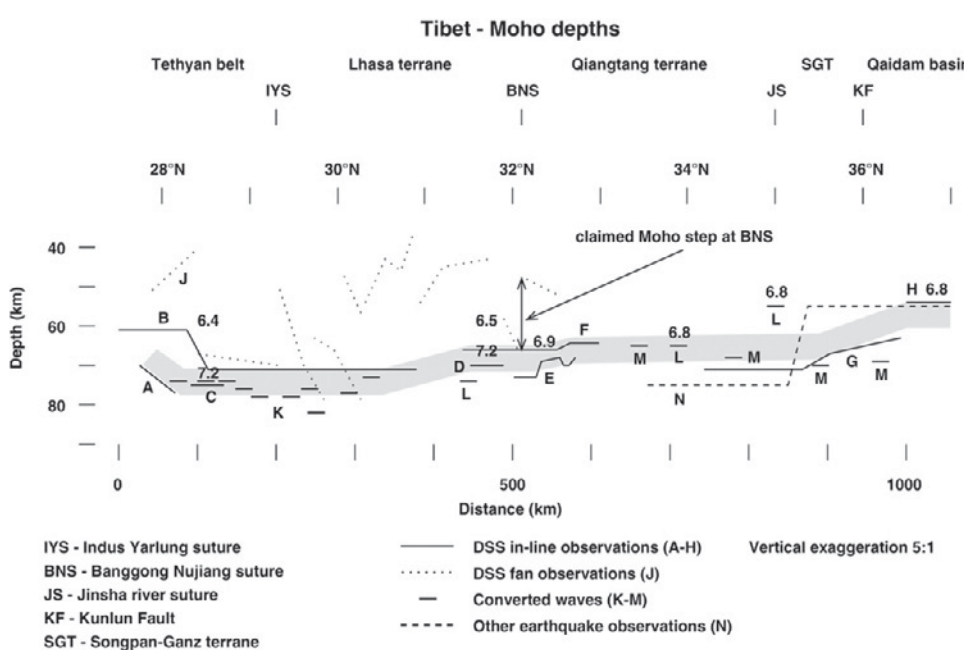



Figure 9.6.2-07. P-wave model of the INDEPTH-III profile across central Tibet (from Zhao et al., 2001, fig. 7). For location, see Fig. 9.6.2-05. [Geophysical Journal International, v. 145, p. 486–498. Copyright John Wiley & Sons Ltd.]

Figure 9.6.2-08. Cross section showing Moho depths across Tibet (from Zhao et al., 2001, fig. 10). Letters indicate the corresponding references: J—Hirn et al. (1984c); C—Sapin et al. (1985); D—Yin et al. (1990); B—Lu and Wang (1990); G—Zhao et al. (1993); A—Gao et al. (1995); H—Herquel et al. (1995); N—Kong et al. (1996); E—Zhao et al. (1996); M—Kind et al. (1996); K—Yuan et al. (1997); K—Owens and Zandt (1997); L—Zhao et al. (2001). [Geophysical Journal International, v. 145, p. 486–498. Copyright John Wiley & Sons Ltd.]



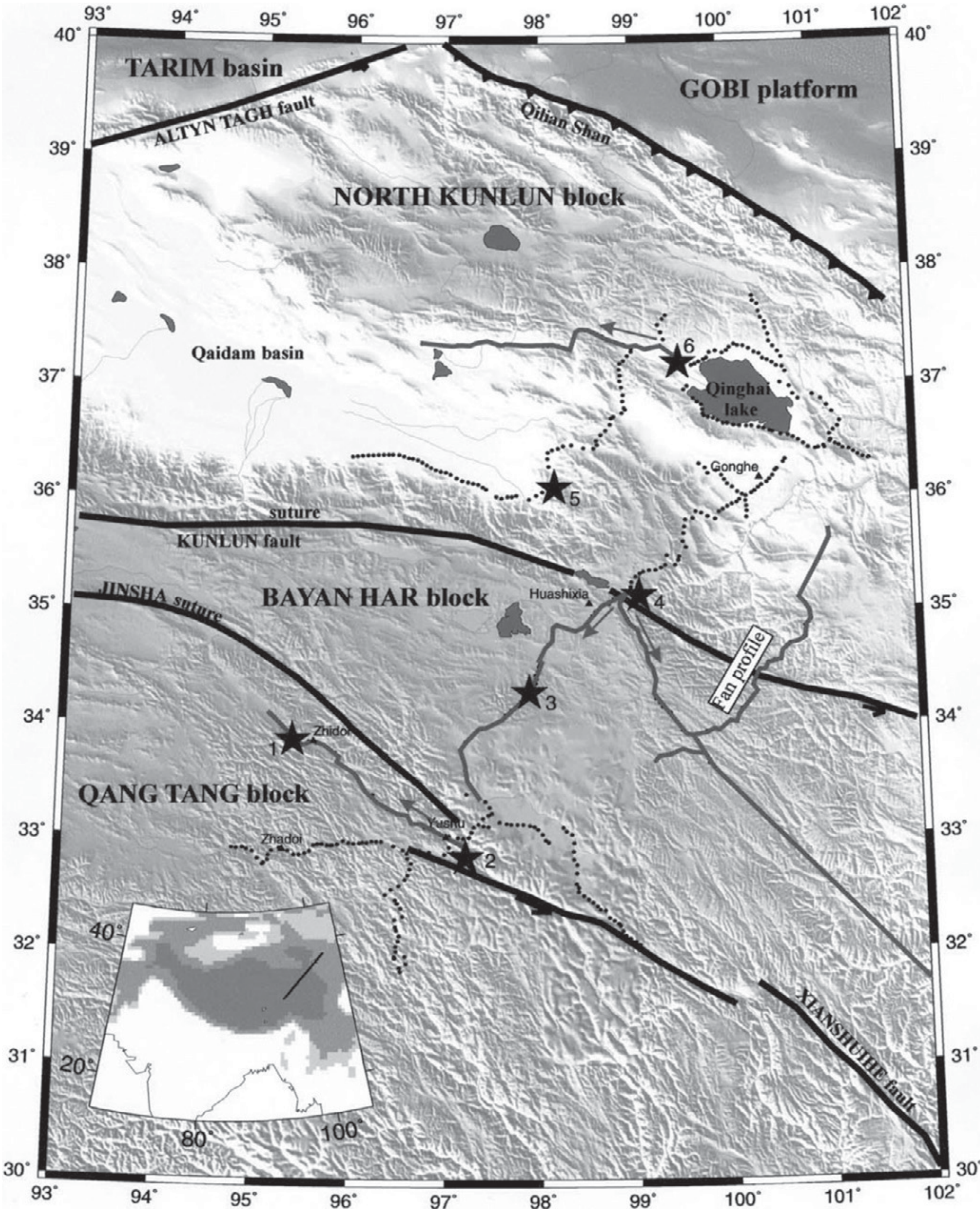



Figure 9.6.2-09. Location map of Sino-French seismic refraction profiles (from Galve et al., 2002, fig. 1). Stars—shotpoints; full and dotted lines—discussed and not discussed, respectively, in detail by Galvé et al. (2002). [Earth and Planetary Science Letters, v. 203, p. 35–43. Copyright Elsevier. .]

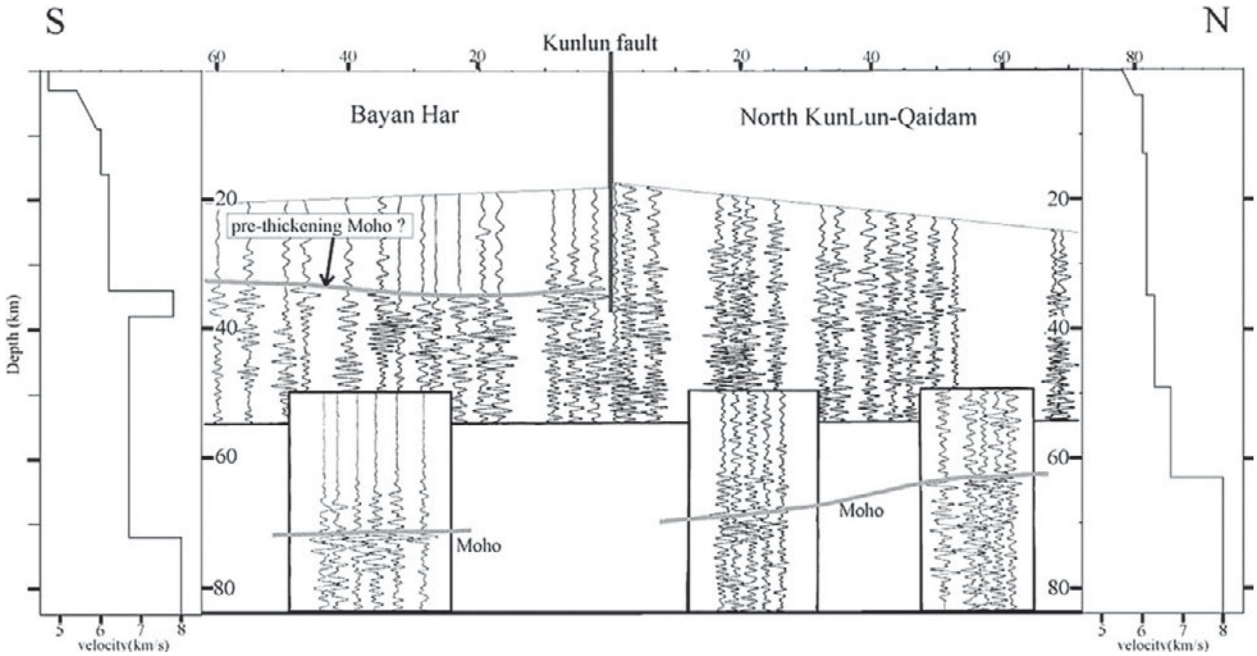



Figure 9.6.2-10. Depth-converted record section of the fan profile at 155 km offset from broadside shot 4 (for location, see Fig. 9.6.2-09) and depth section spanning 140 km across the Kun Lun fault (from Galvé et al., 2002, fig. 4). The insets show depth-converted seismograms from offsets of 190–240 km. [Earth and Planetary Science Letters, v. 203, p. 35–43. Copyright Elsevier.]

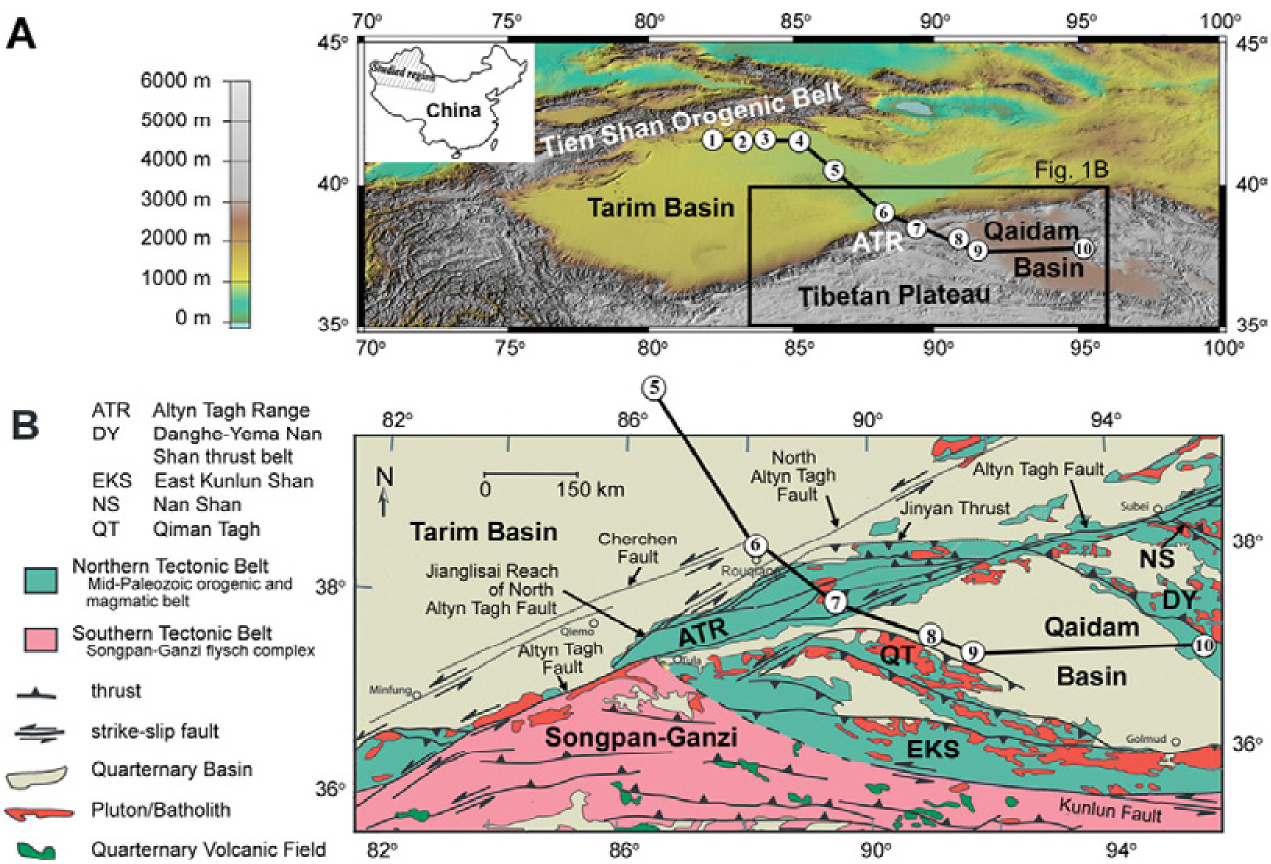



Figure 9.6.2-11. Location map of NW China showing a 1420-km-long seismic refraction profile extending from the Qaidam Basin to the northern margin of the Tarim Basin (from Zhao et al., 2006, fig. 1). [Earth and Planetary Science Letters, v. 241, p. 804–814. Copyright Elsevier.]

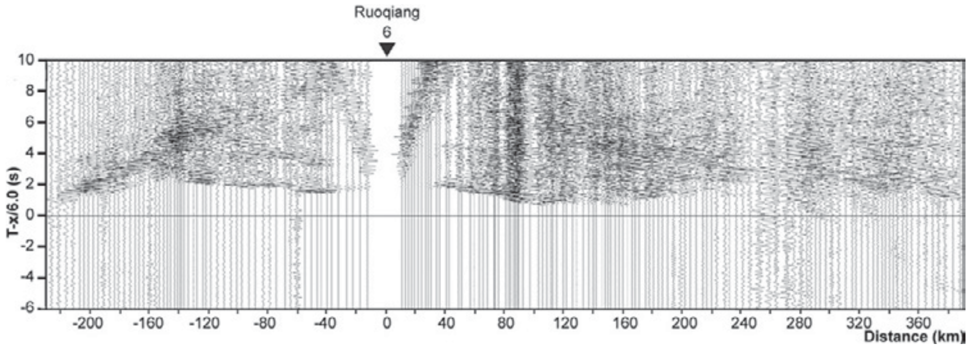



Figure 9.6.2-12. P-wave record section of shot 6 of the 1420-km-long seismic refraction profile across the Altyn Tagh Range, NW China (from Zhao et al., 2006, fig. 2A). For location, see Fig. 9.6.2-11. [Earth and Planetary Science Letters, v. 241, p. 804–814. Copyright Elsevier.]

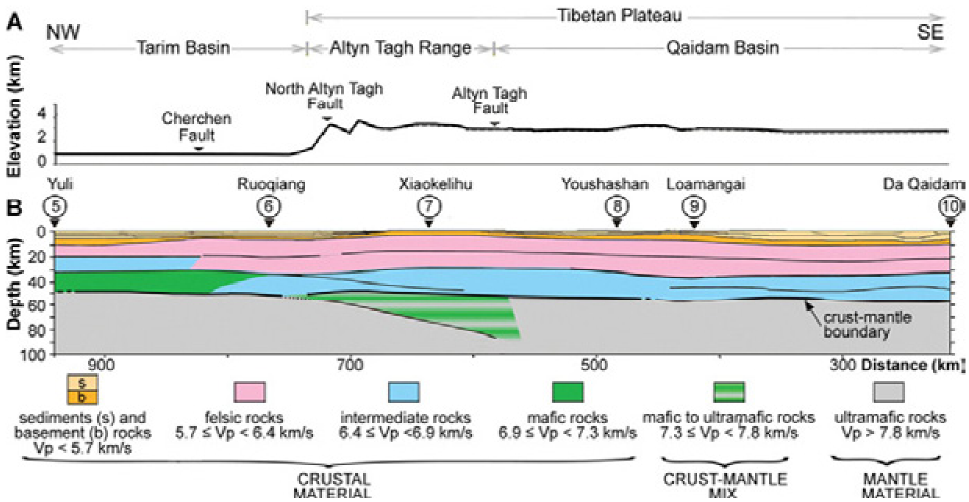



Figure 9.6.2-13. P-wave model across the Altyn Tagh Range, NW China (from Zhao et al., 2006, fig. 7). For location, see Fig. 9.6.2-11. [Earth and Planetary Science Letters, v. 241, p. 804–814. Copyright Elsevier.]

recorded in NNW direction from the Altyn Tagh fault to the Altai Mountains (see Fig. 8.7.2-08).

Similar to the Kun Lun fault farther south, the Cherchen fault seemed to mark a drastic change in crustal structure (Fig. 9.6.2-13). The interpretation was carried out in cooperation with the USGS (Zhao et al., 2006). The resulting model showed a platform-type crust with a high-velocity lower crust beneath the Tarim Basin north of the Cherchen fault (for location, see Fig. 9.6.2-09). This lower crust was not found south of the fault. In contrast, the crust beneath the northern Tibetan Plateau evidently lacks a high-velocity mafic lower crustal layer.

The velocity-depth structure was found to be similar under the Altyn Tagh Range and the Qaidam Basin to the southeast. However, the seismic model of Zhao et al. (2006) also indicated that the high topography of the Altyn Tagh Range of ~3000 m is supported by a wedge-shaped region underlying the crust with seismic velocities of 7.6–7.8 km/s that extends to 90 km depth and which the authors defined as crust-mantle mix.

Another 1000-km-long seismic profile (Liu et al., 2006) investigated the northeastern margin of the Tibetan Plateau to study the interaction between the Tibetan Plateau and the Sino-Korean platform. The joint China-U.S. project was a combination of seismic refraction, teleseismic observations and magnetotelluric soundings. The line started to the southeast of the Qaidam Basin in the Songpan-Ganzi-terrane and ended in the Ordos Basin, part of the Sino-Korean platform. Twelve charges of 2 tons were fired at nine shotpoints, eight of them in boreholes, while shotpoint no. 1 was an underwater shot (Fig. 9.6.2-14). The seismic energy was recorded by 200 portable digital seismographs, DAS-1, developed by the Geophysical Exploration Center of the China Earthquake Administration. The spacing of the three-component receivers was 1.5–2 km, enabling the interpretation of both P- and S-waves (Fig. 9.6.2-15; Appendix A9-4-2).

The interpretation revealed significant variations of crustal velocities along the profile and a decrease of average crustal thick-

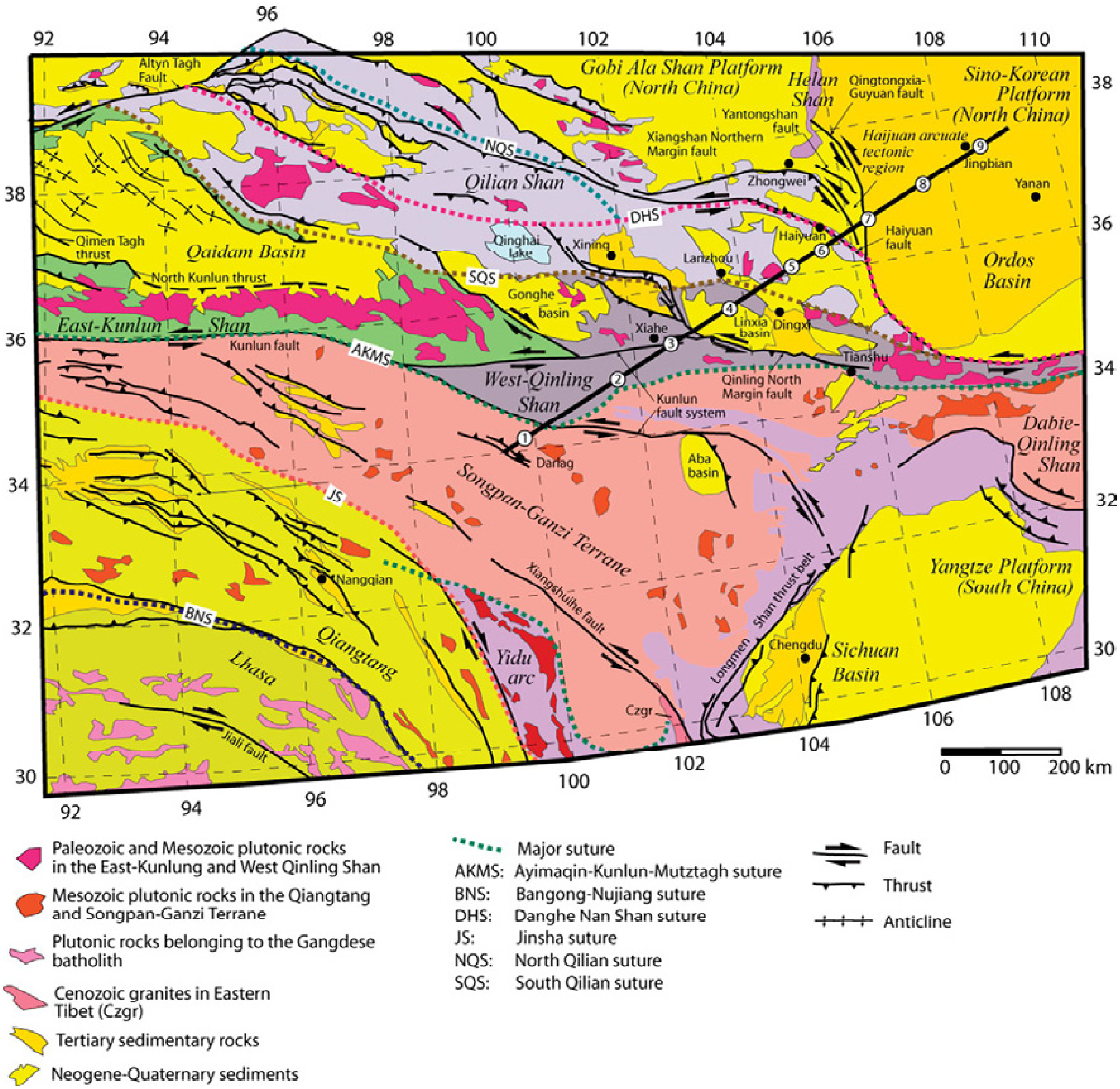



Figure 9.6.2-14. Location map of north central China showing a 1000-km-long seismic refraction profile extending from the Songpan-Ganzi terrane into the Sino-Korean platform (from Liu et al., 2006, fig. 1). [Tectonophysics, v. 420, p. 253–266. Copyright Elsevier.]

ness from around 60 km in the Songpan-Ganzi terrane at the northeastern margin of the Tibetan Plateau to 40–45 km under the Sino-Korean platform (Fig. 9.6.2-16). In particular, the lower crust varies in thickness from 38 km in the southwest to 22 km in the northeast, but in general, relatively low velocities are predominant in the lower crust (Liu et al., 2006).

Another example of a cooperative project between Chinese and U.S. scientists was a seismic-refraction project through the Dabie Shan in eastern central China (Wang et al., 2000), shown as project A33 on Fig. 9.6.2-01. The project was carried out in 1994 and consisted of a 400-km-long, N-S–directed line through the E-W–trending ultra-high pressure Dabie Shan orogenic belt in east central China (Fig. 9.6.2-17).

Eight shots (seven borehole shots; one underwater shot) of 1000–2000 kg, spaced ~50 km apart, were recorded by a total of 150 analog recorders (Fig. 9.6.2-18; Appendix A9-4-2).

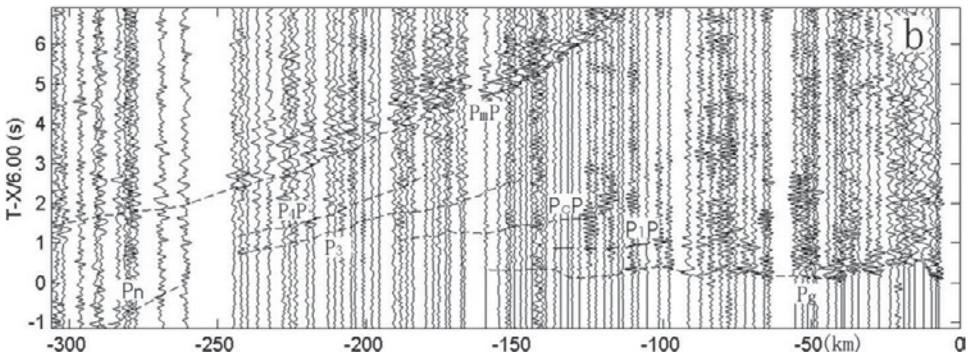



Figure 9.6.2-15. P-wave record section of shotpoint 4 of the seismic refraction profile extending from the Songpan-Ganzi terrane into the Sino-Korean platform (from Liu et al., 2006, fig. 3B). [Tectonophysics, v. 420, p. 253–266. Copyright Elsevier.]

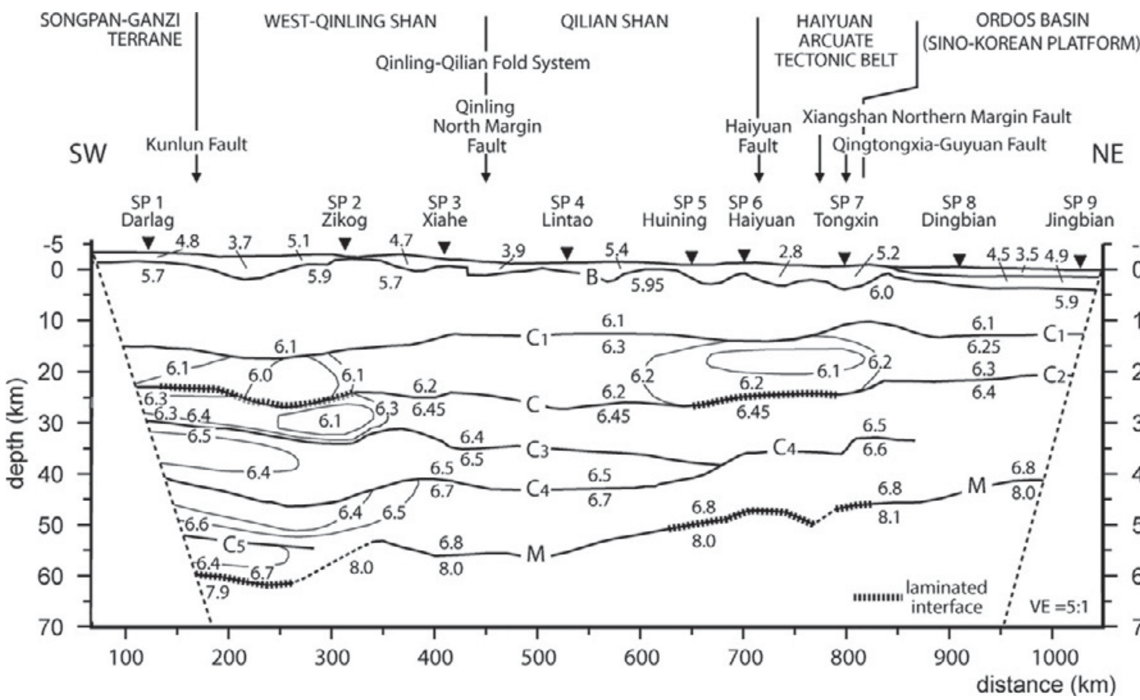



Figure 9.6.2-16. P-wave model between the Tibetan Plateau and the Sino-Korean platform, north central China (from Liu et al., 2000, fig. 6a). For location, see Fig. 9.6.2-14. [Tectonophysics, v. 420, p. 253–266. Copyright Elsevier.]

The recorders were spaced 1 km apart inside the orogen and 3–4 km outside.

The interpretation (Fig. 9.6.2-19) revealed that the cratonal blocks north and south of the orogen were composed of a 35-km-thick, three-layered crust with average seismic velocities of 6.0, 6.5, and 6.8 km/s. The crust thickened underneath the Dabie Shan orogenic belt and reached a maximum thickness of 41.5 km beneath the northern margin (right half of Fig. 9.6.2-19) of the orogen. Here, also a Moho offset was observed, caused possibly by an active strike-slip fault (Wang et al., 2000).

The recording with three-component seismographs allowed an interpretation of both P- and S-waves and the determination of Poisson's ratio resulting in values of 0.23–0.25 for the upper crust, 0.24–0.27 for the middle crust, and 0.27–0.29 for the lower crust. In the upper crust, the lowest Poisson's ratios were observed underneath the orogen, whereas in the middle and lower crust, Poisson's ratios were highest in the south and decreased northward (Wang et al., 2000).

Li and Mooney (1998) and Li et al. (2006) have summarized the main results of crustal studies in China in crustal columns with average crustal structure for the main tectonic areas (see Fig. 10.5.2-03) as well as in contour maps, showing crustal thickness (Moho depths, see Fig. 10.5.2-01) and uppermost-mantle velocities ($P_n$-velocities).

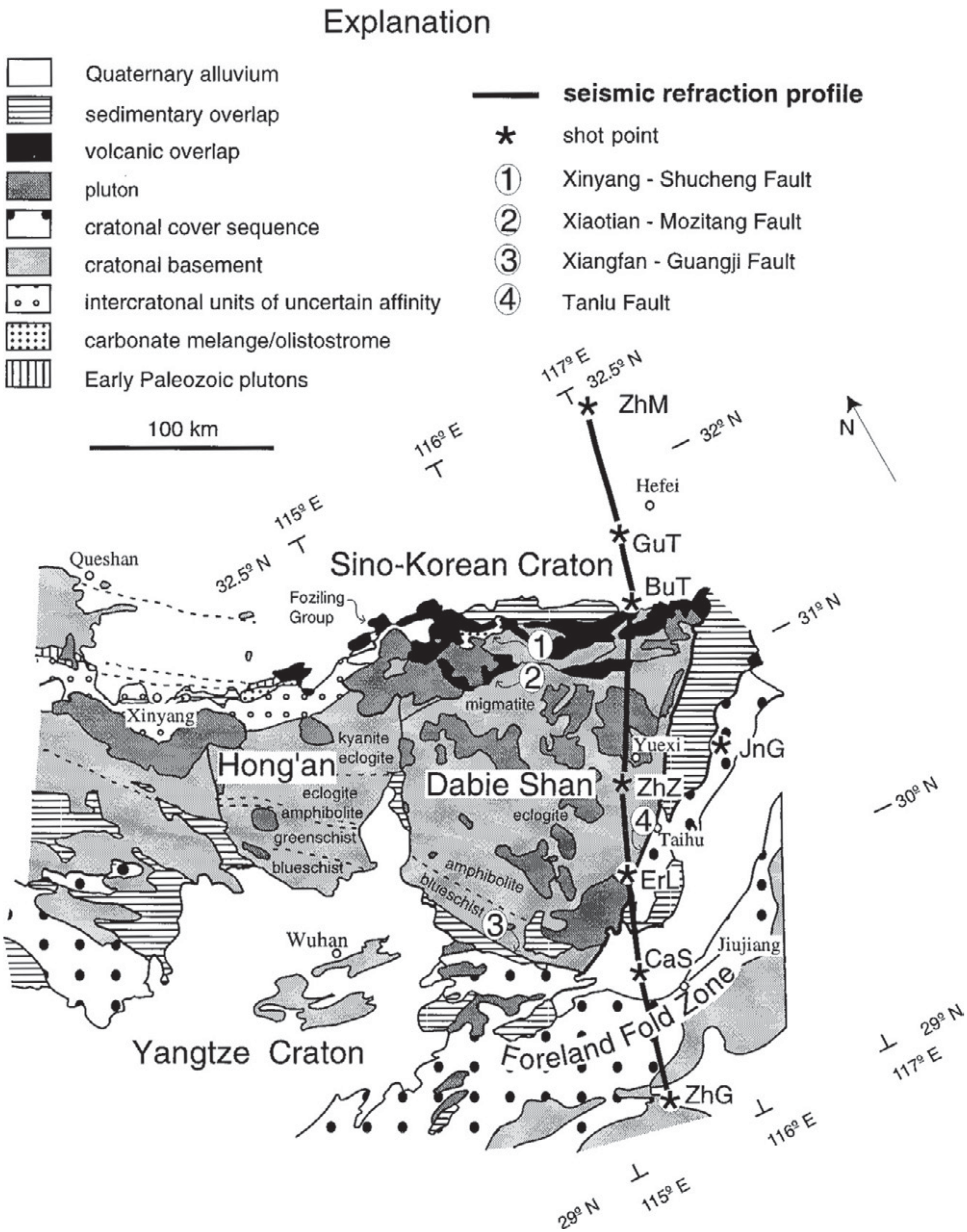



Figure 9.6.2-17. Location map of the Dabie Shan orogen, central eastern China, showing the location of the 400-km-long N-S–directed seismic refraction profile (from Wang et al., 2000, fig. 2). [Journal of Geophysical Research, v. 105, p. 10,857–10,869. Reproduced by permission of American Geophysical Union.]

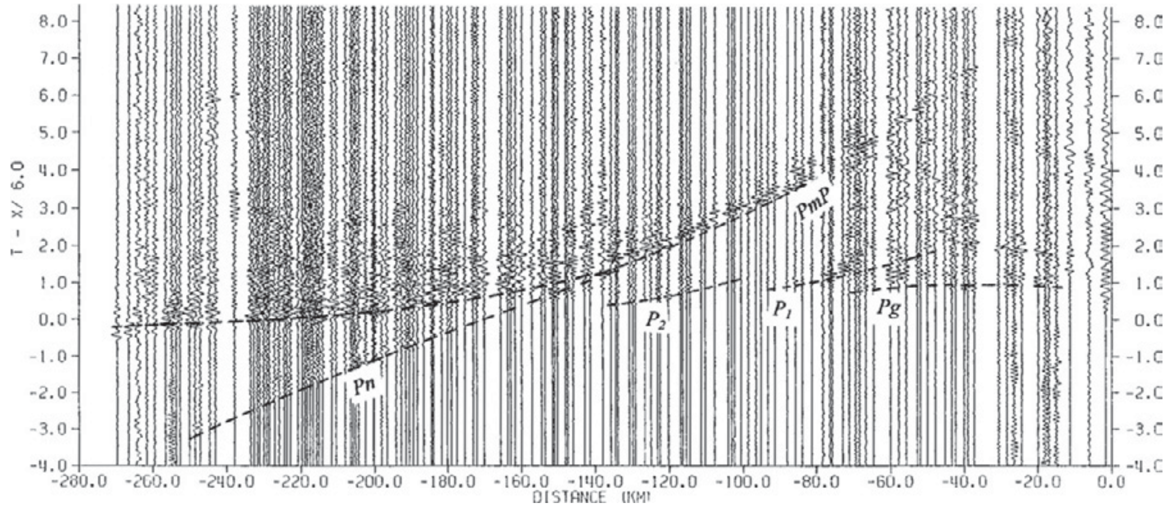



Figure 9.6.2-18. P-wave record section of shotpoint ZhM of the seismic refraction line across the Dabie Shan orogen, central eastern China (from Wang et al., 2000, fig. 3a). [Journal of Geophysical Research, v. 105, p. 10,857–10,869. Reproduced by permission of American Geophysical Union.]

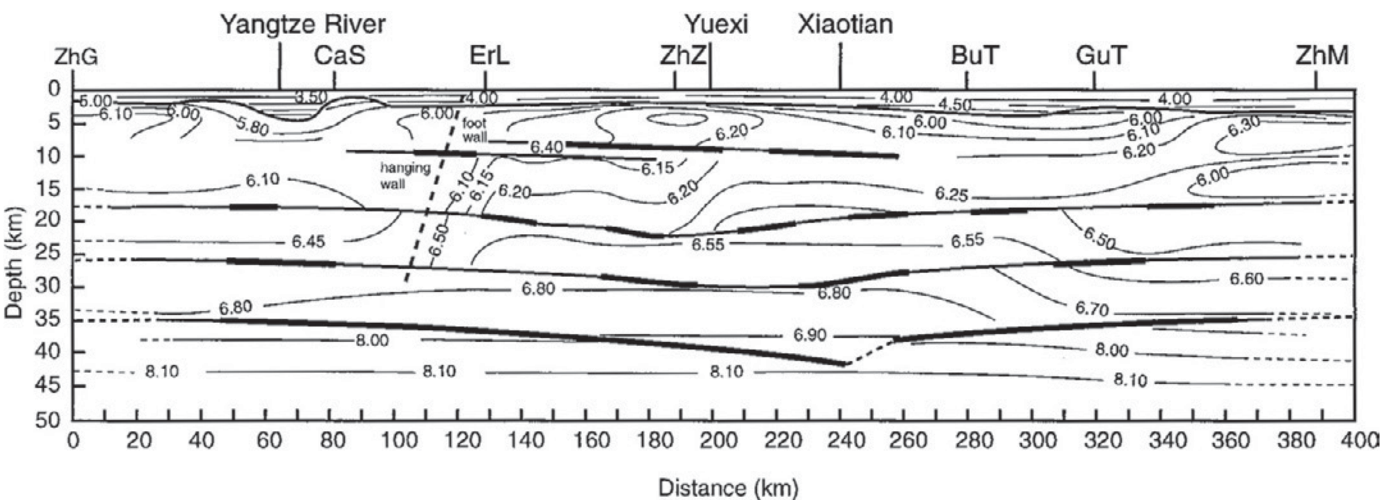



Figure 9.6.2-19. P-wave model across the Dabie Shan orogen, central eastern China (from Wang et al., 2000, fig. 9a). For location, see Fig. 9.6.2-17. [Journal of Geophysical Research, v. 105, p. 10,857–10,869. Reproduced by permission of American Geophysical Union.]

### 9.6.3. Japan

A summary of crustal studies in Japan performed in the 1990s also contained a review of results of the pre-1990s studies (Iwasaki et al., 2002). The tectonic map shows all seismic profiles performed in and around Japan until the end of the 1990s (Fig. 9.6 3-01). Lines A–A′, B–B′, and C–C′ are profiles which had already been obtained in the 1960s (see Chapter 6).

Controlled-source seismology in Japan was activated in the 1990s as part of the Japanese Earthquake Prediction Program. The first projects took place on the island of Honshu in 1989 and 1990. A 200-km-long profile was shot in 1989 with four shotpoints between Fujihashi and Kamigori in central-western Honshu (named "SW Japan" in Fig. 9.6 3-01), and a project in the Kitakami region was accomplished in 1990 (line *a* in Fig. 9.6 3-01). These projects were described in Chapter 8.7.3 (Iwasaki et al., 1993, 1994; Research Group for Explosion Seismology, 1992b, 1995).

Other seismic-refraction studies followed. Some of these projects were described in Japanese with only some technical details and local maps provided in the English abstracts. The general scheme of most projects, described in the following, involved four shotpoints and 170–190 recording sites along profiles which usually were ~180 km long.

In 1991 a 180-km-long, E-W–directed, line with 169 recording sites was observed in central Japan (*1* in Fig. 9.6.3-01) between Agatsuma and Kanazawa (Research Group for Explosion

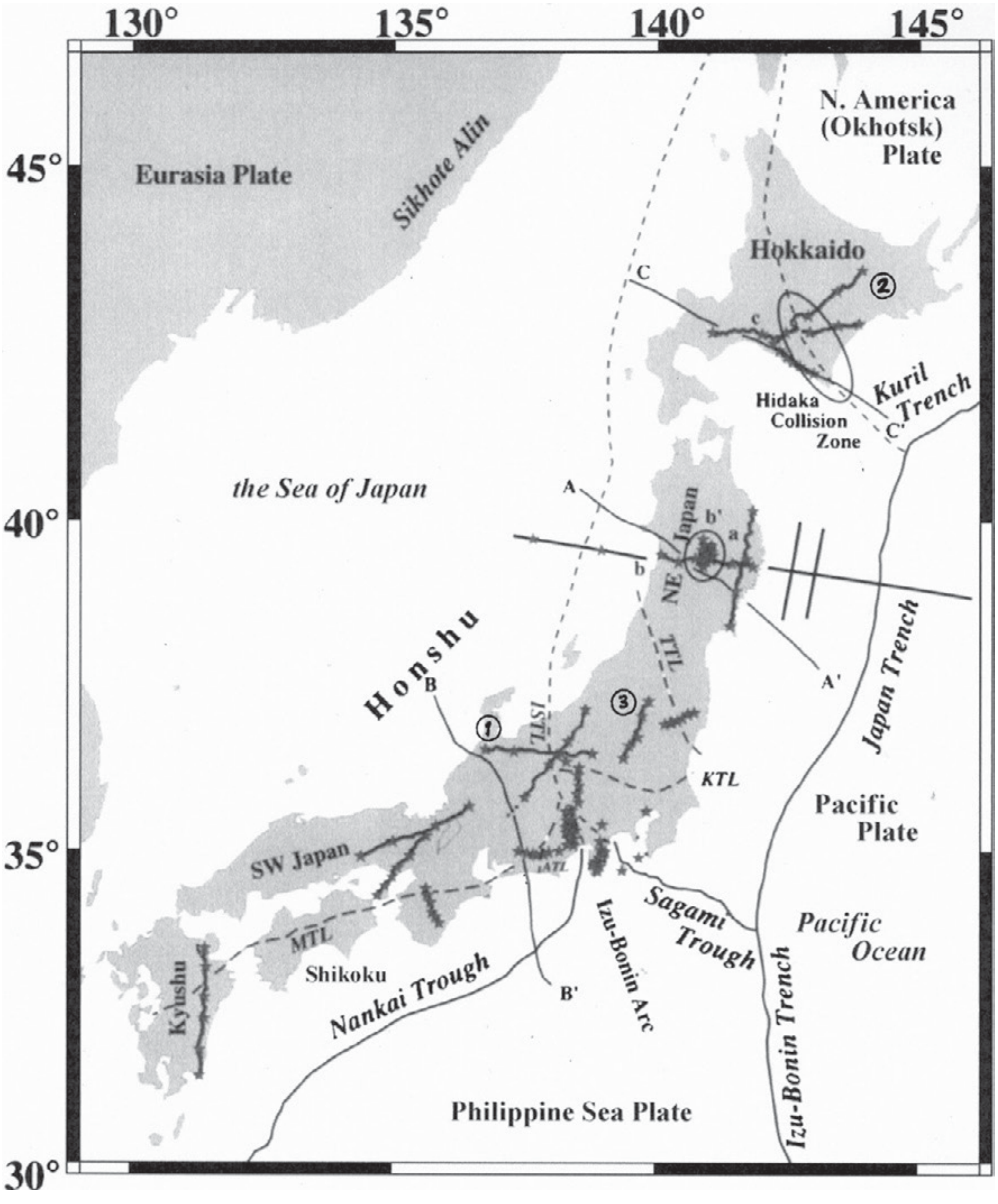



Figure 9.6.3-01. Seismic-refraction profiles in and around Japan (modified from Iwasaki et al., 2002, fig. 1 [Tectonophysics, v. 355, p. 53–66.]). A–A′: line of cross section by Asano et al. (1981) shown in Figure 7.6.3-01. B–B′: Atsumi-Noto profile, central Japan (Aoki et al., 1972). C-C': southern Hokkaido (Okada et al., 1973).

Seismology, 1994). In 1992, the central part of Hokkaido was the target of a 178-km-long profile (*2* in Fig. 9.6.3-01) with 184 sites between Tsubetsu and Hikada Monbetsu (Research Group for Explosion Seismology, 1993). In 1993 a 105-km-long profile (*3* in Fig. 9.6.3-01) with five shotpoints and 195 recording sites was recorded in central Japan in the northern part of the Kanto district between Shimogo and Kiryu (Research Group for Explosion Seismology, 1996; Ozel et al., 1999).

Following the destructive Kobe earthquake of 17 January 1995, seismic-refraction (Research Group on Underground Structure in the Kobe-Hanshin Area, 1997; Research Group for Explosion Seismology, 1997; Ohmura et al., 2001) and -reflection (Sato et al., 1998) measurements were carried out in the surroundings of Kobe. Four refraction lines covered the Kobe area (Fig. 9.6.3-02), and more than 150 recorders were deployed. Six major shots (S1 to S6, some outside the map area in Fig. 9.6.3-02), with charges between 350 and 700 kg, provided information on the whole crust. In addition, smaller shots were fired to study small-scale crustal heterogeneities in the area of Awaji Island (Demachi et al., 2004), while shot U1 served in particular to study the basement structure under the city of Kobe. To reveal the structure of the seismogenic and other associated active faults, a seismic-reflection line (Sato et al., 1998) was shot across the northern part of Awaji Island and the neighboring offshore region. On land, four Vibroseis trucks were in operation, while the offshore operation used airguns with a shot spacing of 50 m. The total length of the line was 42 km.

In the 1990s, the Nankai Trough to the south of Shikoku Island and Kii peninsula on Honshu Island as well as the Japan Trench east of Honshu (for geographic locations of the plate structure around Japan, see Fig. 9.8.2-13) were investigated by many marine deep seismic wide-angle reflection surveys, some of which also had a land component. These projects will be discussed in subchapter 9.8.2.

In 1997 a new interdisciplinary project was started to elucidate deformation processes of island arc crust in NE Japan, starting with an onshore-offshore wide-angle seismic experiment across the northern part of Honshu Island. 293 stations were placed along a 150-km-long E-W line (line *b* in Fig. 9.6.3-01), perpendicular to and crossing the 1990 N-S line through the Kitakami Mountains (line *a* in Fig. 9.6.3-01), and recorded 10 onshore (100-kg and 500-kg shots, data example in Fig. 9.6.3-03) and two offshore shots to the west in the Sea of Japan (Research Group for Explosion Seismology, 1997; Iwasaki et al., 2001a, 2002). The crustal cross section of the onshore profile showed a thickening of the crust from 27 km under its western end to 32 km under the Kitakami Lowland and a thinning again farther east to near 28 km under the Kitakami Mountains (Fig. 9.6.3-04).

From the offshore data, a further crustal thinning toward the east could be proved underneath the continental margin, until 80 km off the coast the continental upper-mantle wedge

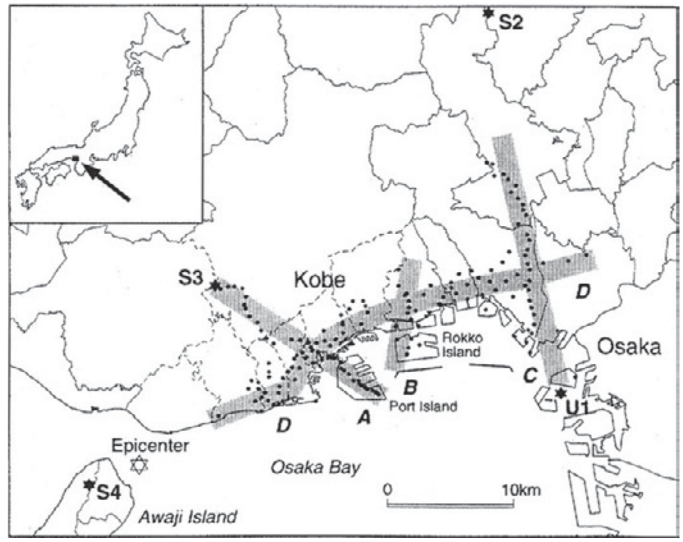



Figure 9.6.3-02. Seismic-refraction profiles in the Kobe-Hanshin area (from Research Group on Underground Structure in the Kobe-Hanshin Area, 1997, fig. 1). [Bulletin of the Earthquake Research Institute, University of Tokyo, v. 72, p. 1–18 (in Japanese). Published by permission of Earthquake Research Institute, University of Tokyo.]

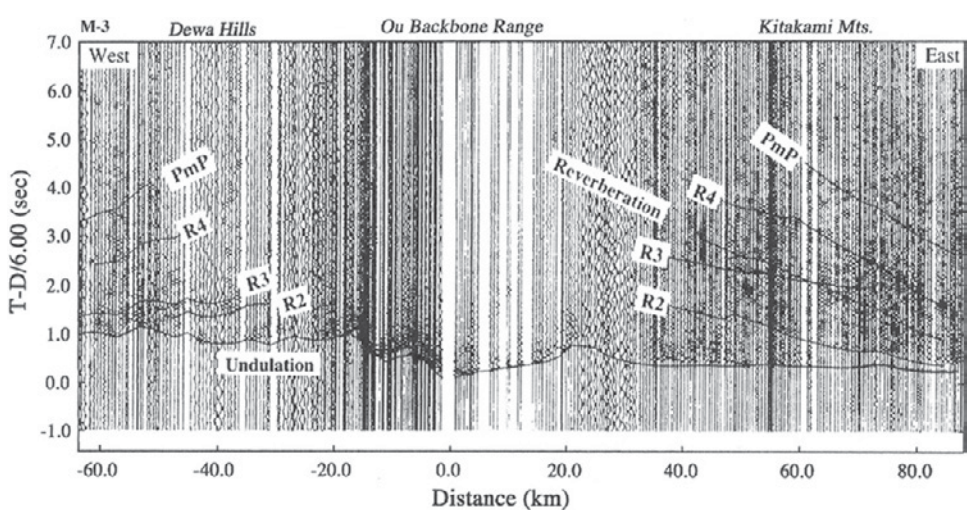



Figure 9.6.3-03. Onshore P-wave crustal data recorded across northern Honshu, Japan from M3 in the center of line *b* in Figure 9.6.3-01 (from Iwasaki et al., 2001a, fig. 4b). [Geophysical Research Letters, v. 28, no. 12, p. 2329–2332. Reproduced by permission of American Geophysical Union.]

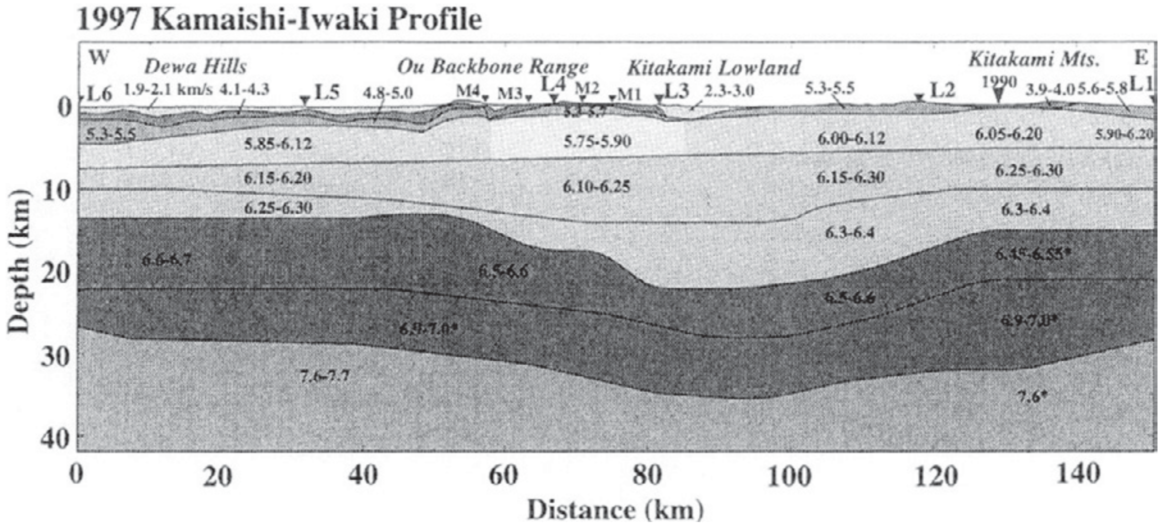



Figure 9.6.3-04. Onshore P-wave model across northern Honshu, Japan (from Iwasaki et al., 2001a, fig. 3; line *b* in Figure 9.6.3-01). [Geophysical Research Letters, v. 28, no. 12, p. 2329–2332. Reproduced by permission of American Geophysical Union.]

converges into the oceanic crust and mantle (Fig. 9.6.3-05). The data also allowed mapping the subduction of the oceanic crust and mantle underneath the Japan island arc, with the subduction angle changing rapidly in E-W direction from 2 to 4° near the trench axis to almost 20° under Japan.

As already mentioned, also Hokkaido Island, northern Japan, was seismically investigated. Following the first seismic-refraction experiment along the southern shore in 1984, in 1992 a 178-km-long line was recorded across the center of the island (Fig. 9.6.3-06), in order to investigate the crustal-scale structural variation and deformation associated with the accretion and collision tectonics in central Hokkaido (Iwasaki et al., 1998).

Here, from 1994 to 1997, the first deep seismic-reflection surveys in Japan were carried out in the southern part of the Hikada

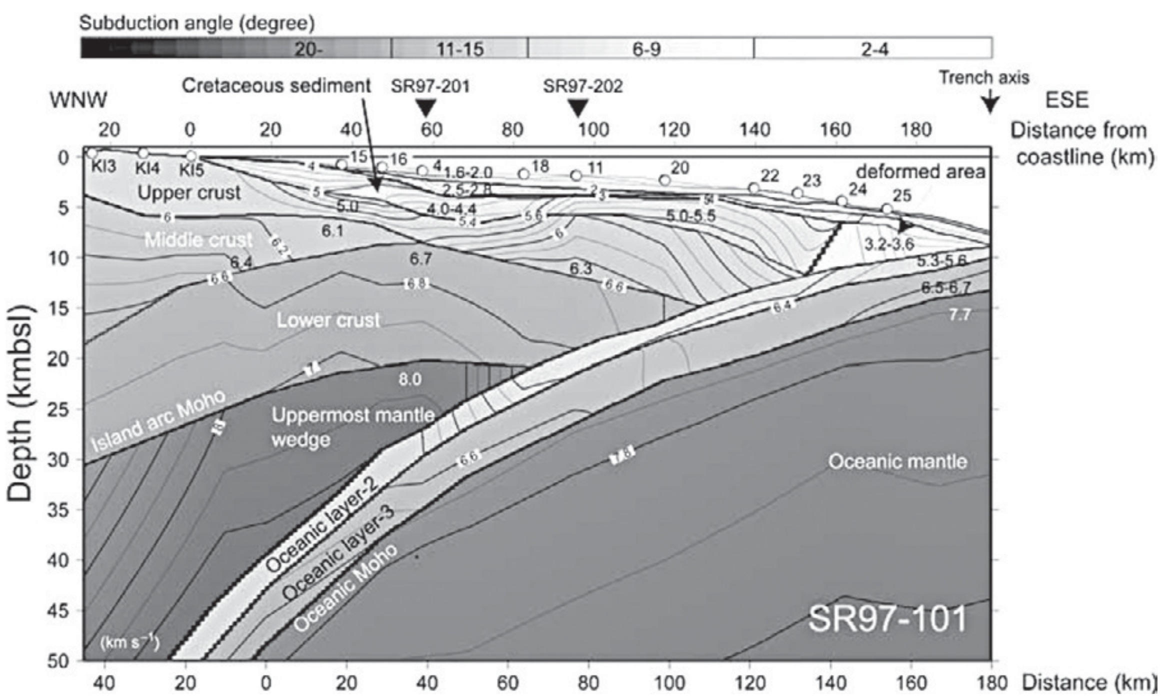



Figure 9.6.3-05. Continuation of the onshore P-wave model across northern Honshu, Japan, into the Pacific Ocean (from Takahashi et al., 2004, fig. 7a; line *b* in Figure 9.6.3-01). [Geophysical Journal International, v. 159, no. 1, p. 129–145. Copyright John Wiley & Sons Ltd.]

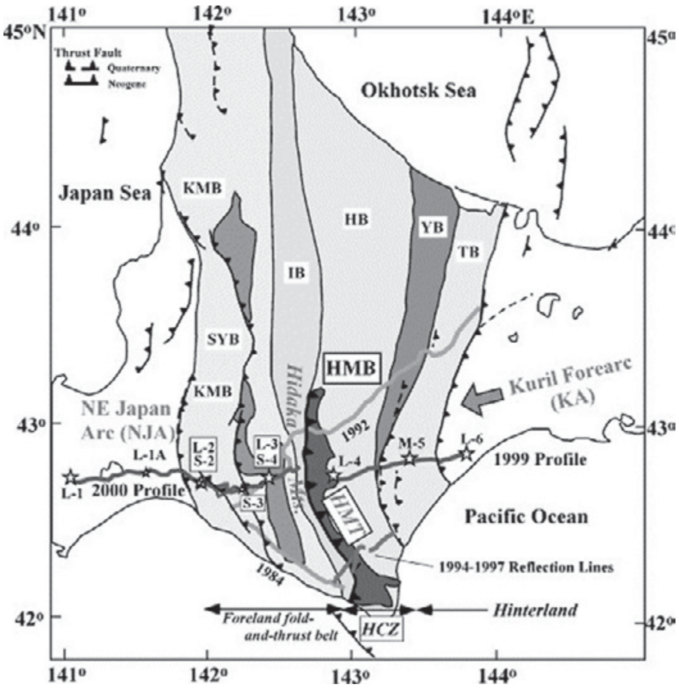



Figure 9.6.3-06. Tectonic map of central Hokkaido with all seismic refraction/reflection profiles and shotpoints (from Iwasaki et al., 2004, fig. 2). [Tectonophysics, v. 388, p. 59–73. Copyright Elsevier.]

collision zone (Arita et al., 1998, Ito, 2000, 2002; Tsumura et al., 1999), applying Vibroseis techniques. The crustal section based on the 1994–1997 Vibroseis reflection surveys imaged direct evidence for the delamination of the Kuril forearc, whereby the wedge of the NE Japan arc was seen as intruding eastward into the delaminated Kuril forearc (Iwasaki et al., 2002).

Three successive experiments followed from 1998 to 2000. In 1998 (Iwasaki et al., 2001b), two short profiles of 10–20 km length were shot, using Vibroseis and dynamite sources. Additional offline recorders mapped the deeper parts of the crust. In 1999, another reflection line was observed, which, together with the 1998 observations, resulted in 83 km total length of reflection observations, where 1567 receivers with spacings of 50–150 m recorded 74 Vibroseis and 30 explosive shots with charge sizes ranging from 40 to 500 kg (Iwasaki et al., 2002). These reflection experiments of 1998–1999 aimed to investigate the deeper structure of the northern part of the collision zone.

In 1999 (Research Group for Explosion Seismology, 2002a), a refraction experiment followed along a 227-km-long east-west–running profile with 297 stations and 6 shots with charge sizes of 100–700 kg. In 2000, the eastern 114 km of the 1999 profile were again occupied by 327 stations and additional 4 shots with charges of 100–300 kg to constrain the crustal structure which had not been resolved by the 1999 project (Research Group for Explosion Seismology, 2002b; Iwasaki et al., 2001b, 2002, 2004). The aim of the wide-angle seismic experiment was to obtain an entire crustal model from the hinterland to the forearc fold-and-thrust belt crossing the Hikada collision zone. Figure 9.6.3-01 and, more detailed, Figure 9.6.3-06 show the location of these experiments. A structural model of upper and middle crust across southern Hokkaido (Fig. 9.6.3-07), obtained from the refraction/wide-angle reflection data was compiled by Iwasaki et al. (2004). In this section, the most important results of the reflection studies were integrated, namely two remarkable reflections with an eastward dip at depths of 15–22 km and 17–26 km. However, the down-

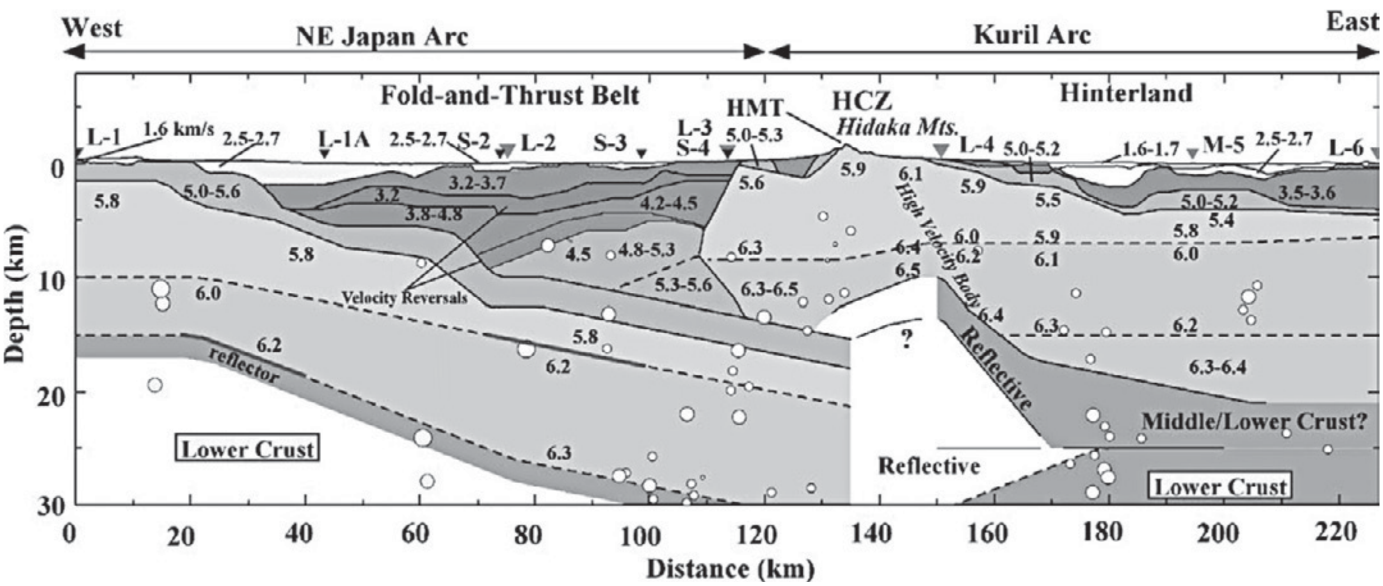



Figure 9.6.3-07. Upper and middle crustal structure model across southern Hokkaido (from Iwasaki et al., 2004, fig. 4). Open circles—hypocenters relocated within 10 km from the seismic profile. [Tectonophysics, v. 388, p. 59–73. Copyright Elsevier.]

going lower crust could not be imaged by the seismic data (Iwasaki et al., 2002, 2004).

In 2000, the Japan Marine Science and Technology Center (JAMSTEC), in cooperation with the University of Tokyo, performed an onshore-offshore experiment along a 492 km profile from the Sea of Okhotsk in the NNW to the Pacific Ocean in the SSE, crossing the northeastern coastal part of Hokkaido and the southwestern Kuril Arc-Trench system (Kurashimo et al., 2007; Nakanishi et al., 2009). In the Okhotsk Sea to the NNW of Hokkaido, 10 OBSs and in the Pacific to the SSE, 30 OBSs were deployed. Energy was supplied by a large airgun array and fired every 150 m. For the onshore part four explosive charges with 300 kg charge size (J1 to J4 in Fig. 9.6.3-08) were detonated and 74 seismic stations were deployed with 1.5 km spacing along the 96-km-long onshore line. The three types of observation—shot gathers airgun shots–OBS, airgun shots–land stations, and land shots–OBS—provided detailed data which also allowed revealing the structure around the onshore-offshore boundary. The interpretation revealed that the crust in the Kuril back-arc region has a thick accretionary complex with velocities of 5.8–6.2 km/s and was considered to continue into the Cretaceous accretionary belt observed in the Hikada Belt based on onshore studies (Iwasaki et al., 1998). The lower crust could not be clarified in this study, but it was concluded that an extremely thick middle/lower crust should occupy ~70% of the entire crust (Nakanishi et al., 2009).

In a piggyback experiment, the land shots of this experiment were also recorded along a 162-km-long line along the northeastern coast of Hokkaido (Taira et al., 2002), using 75 recorders (Fig. 9.6.3-08). On this line shots were recorded in one direction only. The profile was located above the Tokoro belt and aimed to investigate the deep structure of the back arc of the Kuril Arc. The high-quality data allowed estimating the depth to the top of the lower crust and of the Moho (Taira et al., 2002).

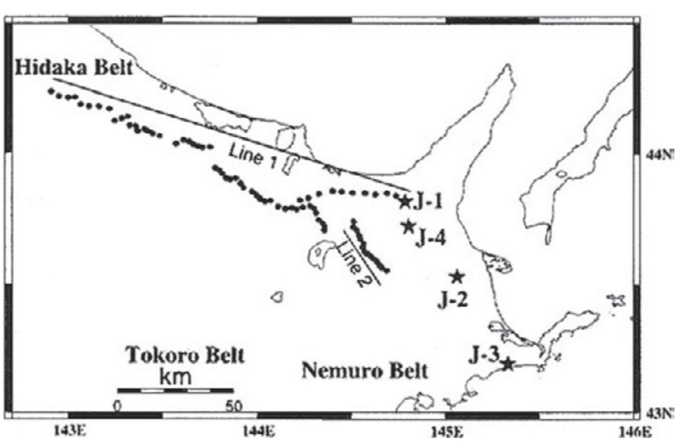



Figure 9.6.3-08. Location of the 2000 Okhotsk seismic experiment in northeastern Hokkaido showing recording sites and shotpoints (from Taira et al., 2002, fig. 1). [Bulletin of the Earthquake Research Institute, University of Tokyo, v. 77, p. 225–230 (in Japanese). Published by permission of Earthquake Research Institute, University of Tokyo.]

## 9.7. CONTROLLED-SOURCE SEISMOLOGY IN THE SOUTHERN HEMISPHERE

### 9.7.1. Australia and New Zealand

#### *9.7.1.1. Australia*

As in the decades before, in the 1990s and 2000s the main activities concerning deep crustal research in Australia with seismic-reflection and -refraction methods remained with the Bureau of Mineral Resources, Geology and Geophysics at Canberra, renamed in the 1990s to the Australian Geological Survey Organisation (AGSO). The Research School of Earth Sciences of the Australian National University at Canberra (ANU), in contrast, concentrated its deep earth research on the upper mantle and below, applying mainly passive seismological methods using earthquake sources and broadband recording systems. The 1992 SKIPPY project—a broadband recording program across the Australian continent—was initiated by ANU and was the first of many field deployments following between 1993 and 1996 (Van der Hilst et al., 1994, 1998; Zeilhuis and van der Hilst, 1996; Clitheroe et al., 2000). The main features of the recording systems used in Australia were summarized by Finlayson (2010; Appendix 2-2).

Since 1985, AGSO has run a large program of deep seismic-reflection surveys offshore and onshore Australia (Fig. 9.7.1-01), which until 1992 were summarized in a table by Goleby et al. (1994).

The onshore group targeted basins, fold belts or Archean basement with petroleum, coal or mineral deposits and operated a 120-channel telemetric acquisition system, using buried explosive charges as energy sources. The data were typically collected using a 40–60 m station interval and appropriate shot intervals to give an 8- to 18-fold data redundancy. The offsets were up to 4 km and record lengths were usually 20 s. Between 1989 and 1992, the yearly data collection amounted to 450–500 km (Goleby et al., 1994). Throughout the 1990s, onshore DSS investigations were focused on the major mineral provinces of Broken Hill, the eastern Lachland Orogen, the goldfields of western Victoria, the Yilgarn Craton, Tasmania, the Hamersley Basin, and the New England Orogen (Finlayson 2010; Appendix 2-2). Surveys which were connected with seismic-refraction/wide-angle reflection investigations will be discussed in some detail.

Together with the deep marine seismic-reflection profiling, wide-angle seismic data were obtained from OBSs along the Vulcan transect in the Timor Sea and the Petrel transect in the Bonaparte basin off northern Australia (Petkovic et al., 2000) in 1996 and 1998. Additionally, a few land stations recorded the airgun shots (Fig. 9.7.1-02). The Vulcan line, e.g., was 336 km long and shot spacing was 100 m.

From the data of the Vulcan transect, the crustal and upper mantle architecture between the Precambrian Australian craton and the Timor Trough was determined. Linking the data with earlier deep seismic marine profiling, the crustal architecture across the major boundary between the Australian and SE Asian

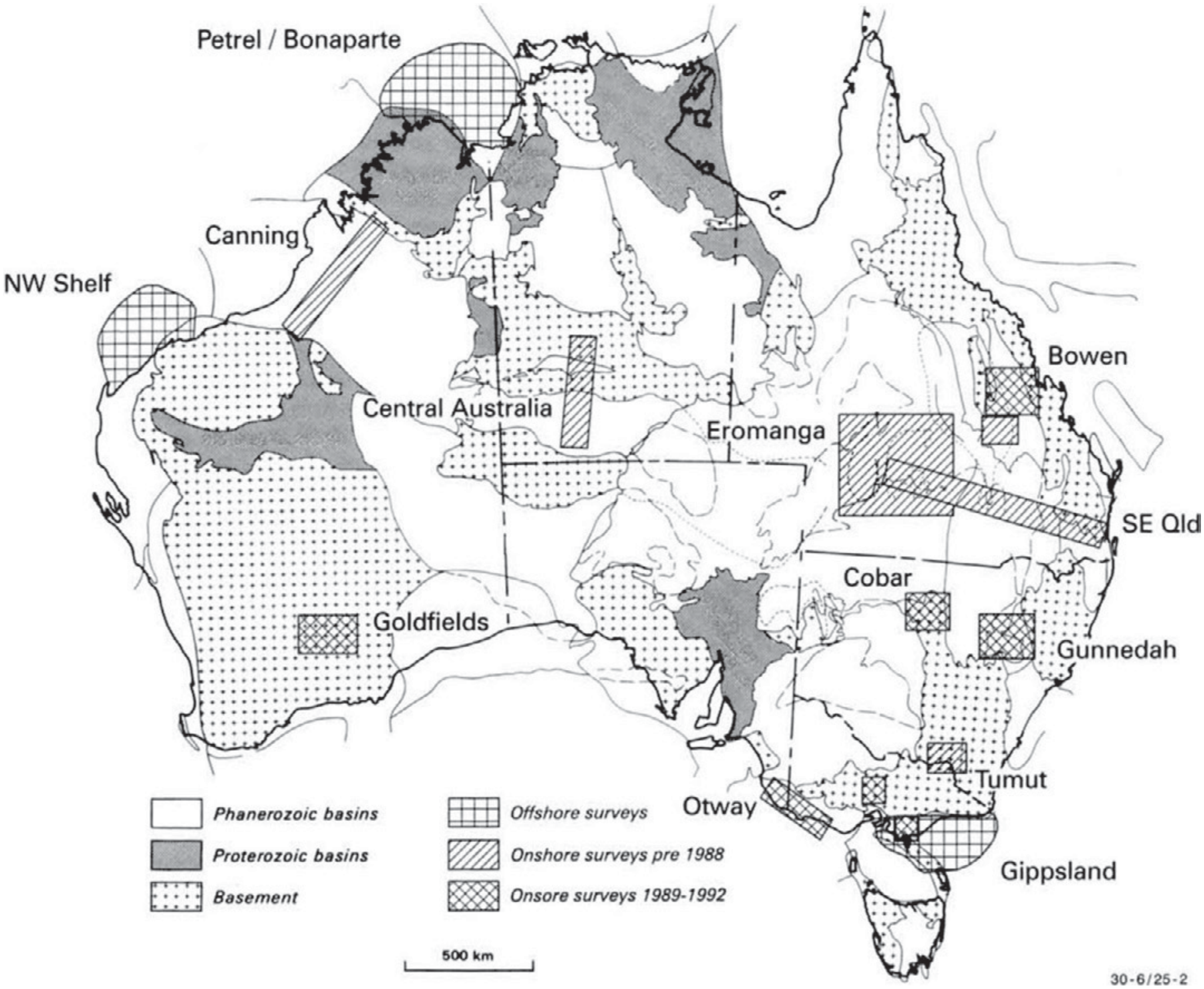



Figure 9.7.1-01. Location of Australian onshore and offshore deep seismic reflection surveys from 1979 to 1992 (from Goleby et al., 1994, fig. 1). [Tectonophysics, v. 232, p. 1–12. Copyright Elsevier.]

plates could be established (Petkovic et al., 2000). Near the Australian coast, a crustal thickness of 35 km resulted. The crust then thinned northwestward to 26 km under the outer shelf of the Timor Trough. While the Paleozoic/Mesozoic basin sequences thickened from almost zero to more than 10 km, the Precambrian basement rocks suffered attenuation from 35 km to 13–14 km across the margin.

Following the 1989 seismic-reflection traverse across the northern Bowen basin, discussed in Chapter 8.8.1, in 1991 a 253-km-long seismic-reflection profile was recorded along a transect on the western margins of the New England batholith from the Gunnedah basin (for location, see Fig. 9.7.1-01) across the Tamworth belt to the Bundarra pluton west of Armidale, Queensland (Korsch et al., 1997; Finlayson 2010; Appendix 2-2).

Another seismic transect was recorded farther north through Queensland, crossing the Mount Isa inlier in North Australia (Drummond et al., 1998). The seismic survey was commissioned in 1994 within a multidisciplinary study of the inlier and its mineral deposits. The reflection line was 250 km long and was overlapped and extended at each end by a refraction survey, 500 km long, to create sufficient offsets for energy penetrations into lower crust and Moho depth ranges (Fig. 9.7.1-03).

Underlain by a thin low-velocity zone, the interpretation (Fig. 9.7.1-04) showed a westward dipping high-velocity body with velocities of 6.9–7.3 km/s in the middle of the crust underneath the Eastern Fold Belt, which appeared to be interrupted toward the west but to continue west of the Mount Isa fault, and finally to merge at ~40 km depth, with the crust-mantle bound-

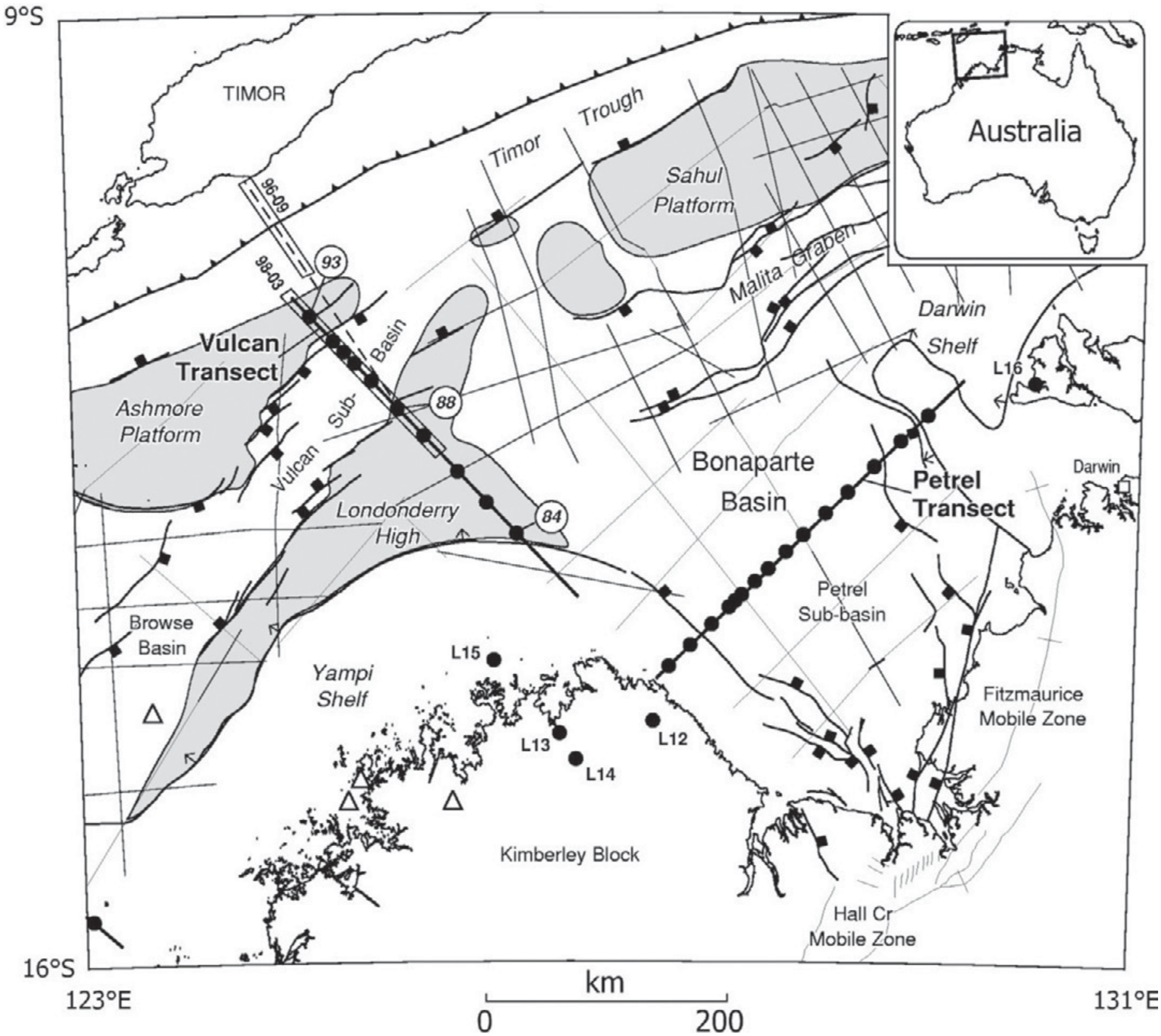



Figure 9.7.1-02. Location of deep seismic reflection surveys in the Timor Sea region (from Petkovic et al., 2000, fig. 1). Thick lines with dots—Vulcan and Petrel transects; black dots—OBS and land recording sites; thin lines—AGSO deep seismic profiling lines. [Tectonophysics, v. 329, p. 23–38. Copyright Elsevier.]

ary, which itself was modeled as a 15-km-thick transition zone between 40 and 55 km depth (Goncharov et al., 1998; Drummond et al., 1998).

In central Australia in 1985, 485 line kilometers of deep seismic-reflection profiling had already been acquired across the Arunta Block and Amadeus Basin, which were followed in 1993 by another 550 line kilometers across the southern Musgrave Block and adjacent Officer Basin (all located within the box labeled "Central Australia" in Fig. 9.7.1-01). All seismic sections were recorded to 20 s TWT. The resulting 900-km-long, N-S–directed, crustal cross section displayed a crustal thickness of ~45 km underneath the Officer Basin in the south and ~50 km underneath the Amadeus Basin farther north and a disruption of the Moho underneath the Musgrave Block in between (Leven and Lindsay, 1995; Korsch et al., 1998).

The Goldfields Province of Western Australia (Fig. 9.7.1-01) was investigated several times. In 1991, the Kalgoorlie traverse EGF01 (Fig. 9.7.1-05) was shot (Goleby et al., 1994, Drummond et al., 2000a). In 1997, several high- resolution seismic-reflection lines were recorded north and south of Kalgoorlie, and in 1999, a

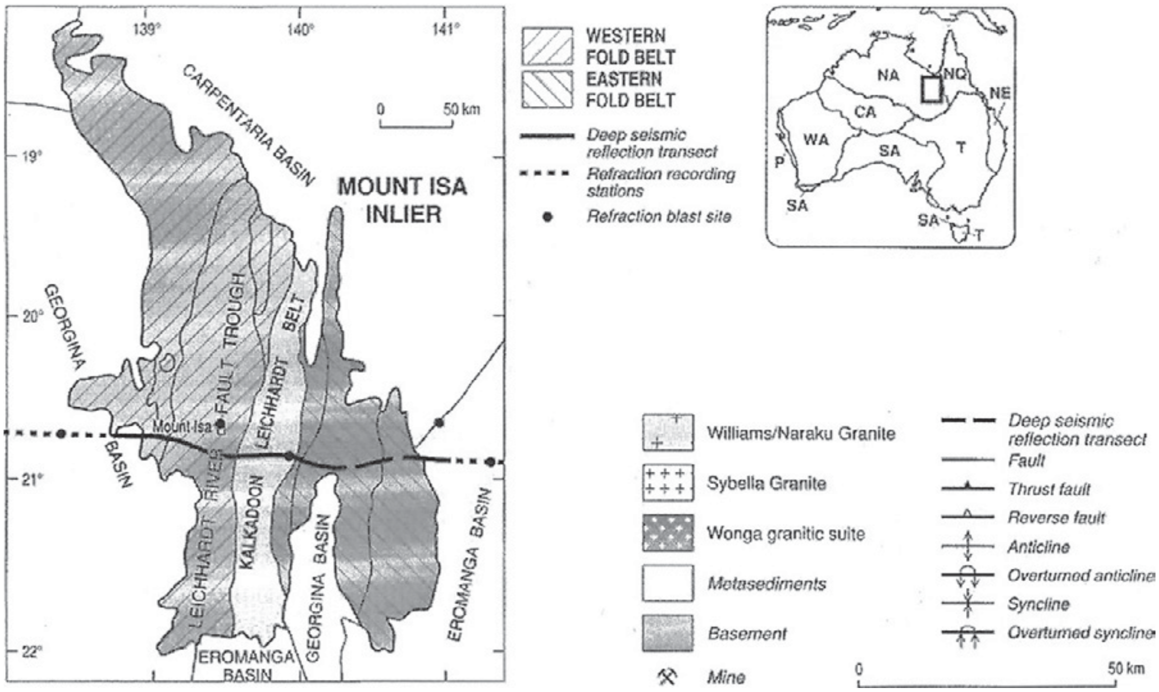



Figure 9.7.1-03. Location of the deep seismic reflection transect through the Mount Isa Inlier, North Australia (from Drummond et al., 1998, fig. 1 top). [Tectonophysics, v. 288, p. 43–56. Copyright Elsevier.]

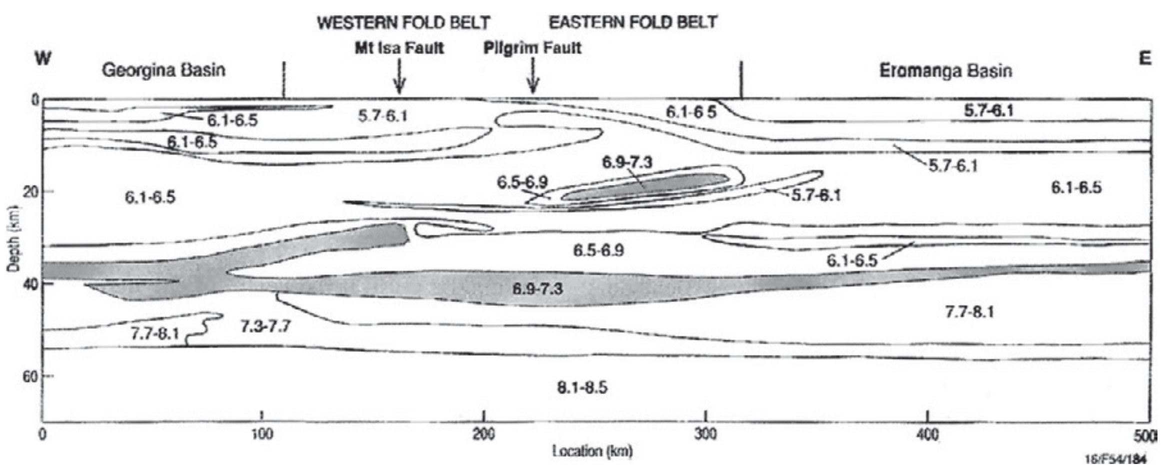



Figure 9.7.1-04. Crustal cross section through the Mount Isa Inlier, North Australia (from Drummond et al., 1998, fig. 4). [Tectonophysics, v. 288, p. 43–56. Copyright Elsevier.]

grid of deep seismic-reflection traverses was acquired to examine the 3-D geometry of the region. It was tied to the 1991 and 1997 profiles (Goleby et al., 2002; Finlayson 2010; Appendix 2-2). In 2001, another deep seismic-reflection survey 01AGSNY was recorded (Goleby et al., 2004), located ~200 km farther to the north-northeast (Fig. 9.7.1-06) using vibrator seismic sources. It will also be discussed briefly in Chapter 10.6.1.

The seismic images of the crust below the greenstone supracrustal rocks of the Eastern Goldfields Province of the Archean Yilgarn Craton showed structures that Drummond et al. (2000a) interpreted in terms of tectonic events. While the upper crust appeared to be largely unreflective, the geometry of the reflective middle crust implied thickening by west-directed thrust stacking, and the lower crust showed a fabric indicative of ductile deformation. Total crustal thickness was determined to be 32–34 km in the west and ~36–38 km in the east, deepening slightly east of the Ida fault (Goleby et al., 1994; Drummond et al. (2000a). The new data of the 2001 traverse confirmed the subdivision of the crust into three layers. A review of all investigations across the Yilgarn Craton considering wide-angle and reflection profiling as well as

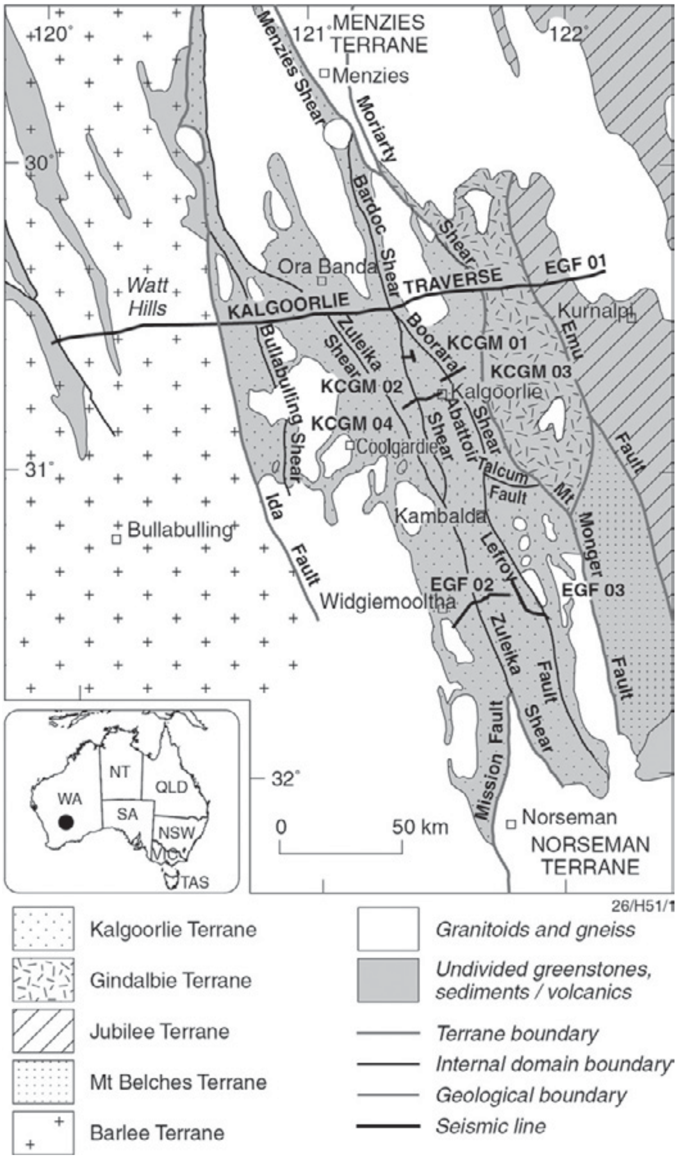



Figure 9.7.1-05. Location of the 1991 Kaloorlie Traverse EGF01 through the Eastern Goldfields Province, Western Australia (from Drummond et al., 2000a, fig. 1). [Tectonophysics, v. 329, p. 193–221. Copyright Elsevier.]

receiver function work included the compilation of a 3-D seismic model of crustal architecture (Goleby et al., 2006).

The seismic survey program of AGSO in 1992 in southeastern Australia targeted the Otway Basin along Australia's southern continental margin (Fig. 9.7.1-01). It consisted of seven regional deep seismic profiles which were shot across the onshore part of the basin (Fig. 9.7.1-07) using boreholes with 10-kg dynamite charges and recorded by 120-channel split spreads with group intervals of 50 m and 20-s record lengths (Finlayson et al., 1996). The longest line (AGSO line 5) was 101 km long and started in Paleozoic basement of the Delamerian Lachlan Orogen (Fig. 9.7.1-07). The crustal thickness was interpreted by Finlayson et al. (1996) to be 31 km (10.3–10.5 s TWT) near the northern basin margin, thinning to ~25 km (9 s TWT) seaward of the Tartwaup fault (Fig. 9.7.1-08; TWF in Fig. 9.7.1-07).

In late 1994 and early 1995, marine seismic profiling was added to explore the offshore region of the Otway Basin (Fig. 9.7.1-09) both by near-vertical and wide-angle incidence across the Otway Basin continental margin (Finlayson et al. (1998). For this survey, seismic stations were also established onshore to record the airgun signals to distances as far as 200 km. The right panel of Figure 9.7.1-09 shows the locations of lines and stations which were used for wide-angle recording. The composite result of the 1992 and 1994–1995 surveys as interpreted by Finlayson et al. (1998) shows that the Moho shallows from a depth of nearly 30 km onshore to 15 km depth at 120 km from the nearest shore, and farther to 12 km in the deep ocean at the end of the profile (Fig. 9.7.1-10).

The 1990s saw several other long-range seismic-reflection programs on continental Australia described in detail by Finlayson (2010; Appendix 2-2) which are only briefly mentioned here. In 1996 and 1997, a 295-km-long seismic-reflection line was recorded across the Broken Hill block in the west of New South Wales. In 1999, a deep seismic-reflection profile northeast of Broken Hill followed. In 1997, the Grampian Mountains and western Lachlan Orogen in western Victoria were studied by deep seismic-reflection lines (Korsch et al., 2002). In 1997, the Molong volcanic belt of the eastern Lachlan Orogen in New South Wales became the location of 105 km of deep seismic-reflection profiling (3 lines) and of a wide-angle seismic survey using explosive seismic sources aimed at determining possible crustal compositional variations across the region (Finlayson et al., 2002). In 1999, reflection profiling was performed farther west in the Junee-Naromine volcanic belt (Glen et al., 2002).

Finally, a seismic-reflection survey of AGSO in 1995 in and around Tasmania (Fig. 9.7.1-11), part of which was interpreted by Drummond et al. (2000b), should be mentioned. The program of deep seismic-reflection profiling was combined with a seismic tomography survey.

The offshore program covered the surrounding continental shelf areas of the entire island of Tasmania and was designed to map the large-scale structures of Tasmania at depth. The airgun seismic energy was also recorded by a number of onshore stations (dots in Fig. 9.7.1-11) to study structures which could not be imaged by the offshore profiling. A preliminary interpretation (Drummond et al., 2000b) revealed a crustal thickness of 33–35 km for data of the offshore line 1 along the eastern coast of Tasmania, which is about the same as derived from line 15 along the southern coastline (Fig. 9.7.1-12).

Several reviews have summarized the results of crust and upper-mantle studies for Australia (Collins, 1991; Collins et al., 2003; Clitheroe et al., 2000). Clitheroe et al. (2000) have based their crustal thickness map mainly on applying the receiver function techniques on events beyond 30° with sufficiently energetic P-wave arrivals recorded by an array of permanent and temporary

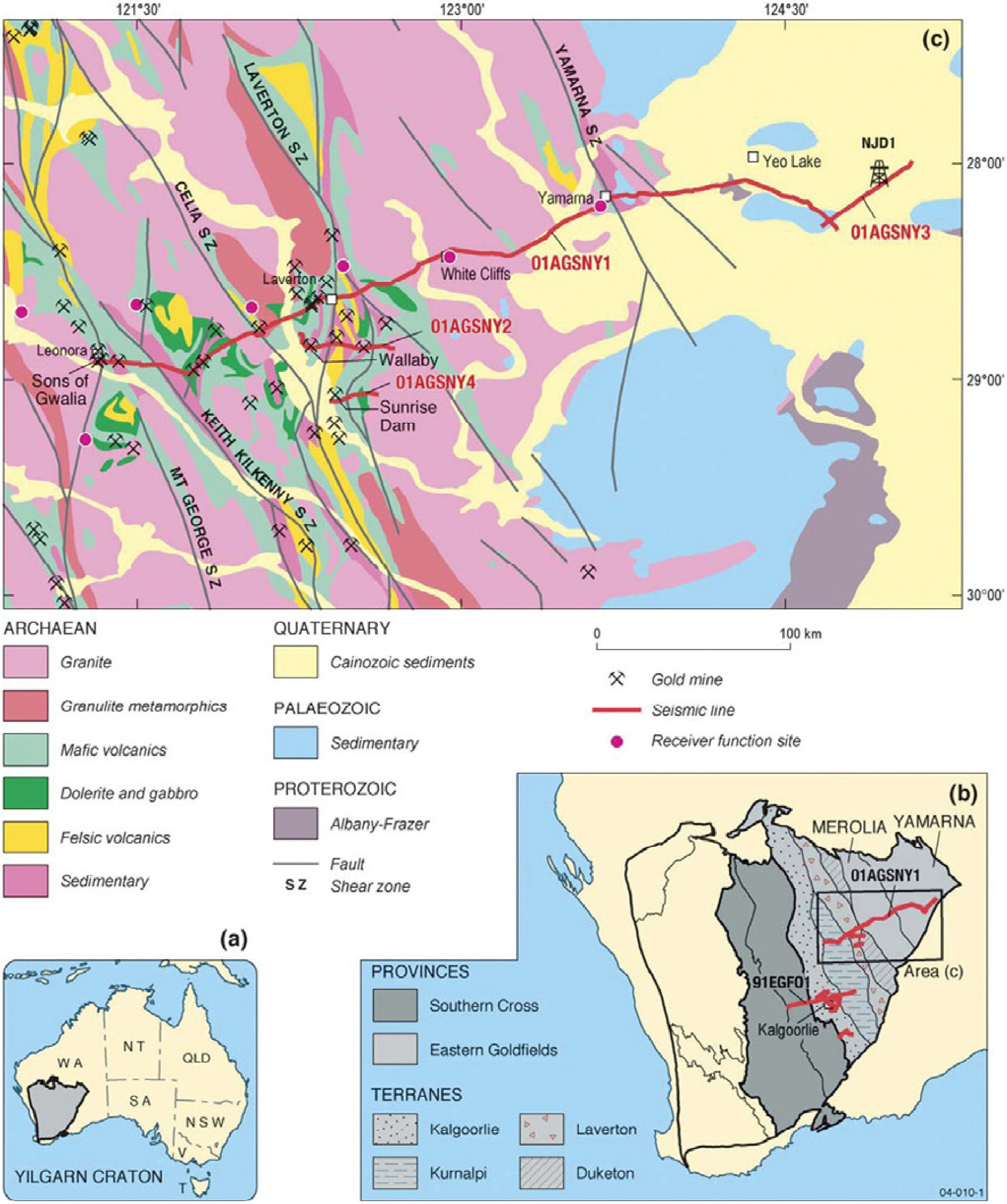



Figure 9.7.1-06. Location of the deep seismic reflection profiles through the Eastern Goldfields Province, Western Australia (from Goleby et al., 2004, fig. 1). (a) Index map. (b) Location of the 1991 and 2001 seismic surveys. (c) Detailed map of the 2001 reflection profiles. [Tectonophysics, v. 388, p. 119–133. Copyright Elsevier.]

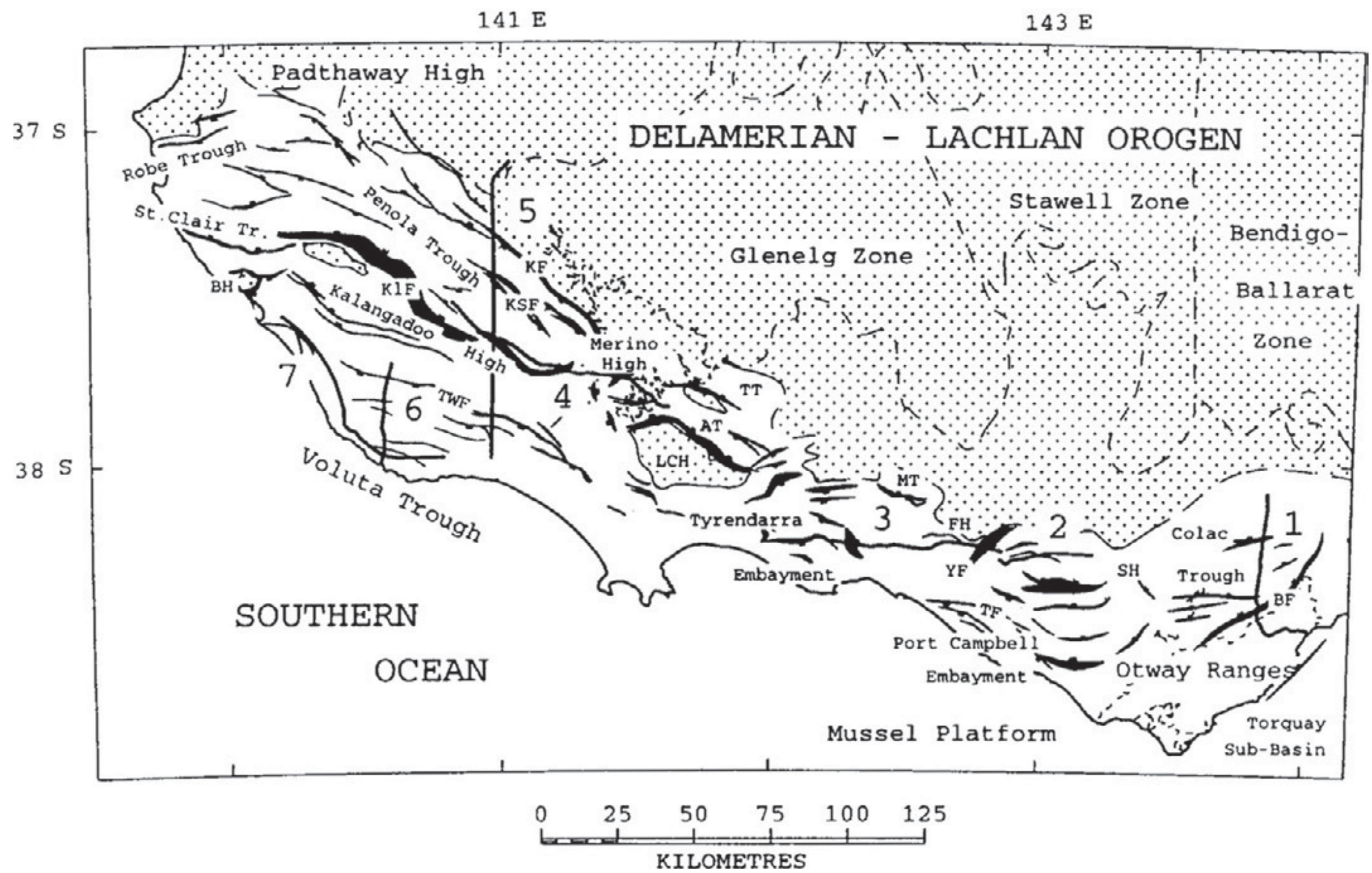



Figure 9.7.1-07. Location of the 1992 onshore seismic reflection profiles in SE Australia (from Finlayson et al., 1996, fig. 3). [Tectonophysics, v. 264, p. 137–152. Copyright Elsevier.]

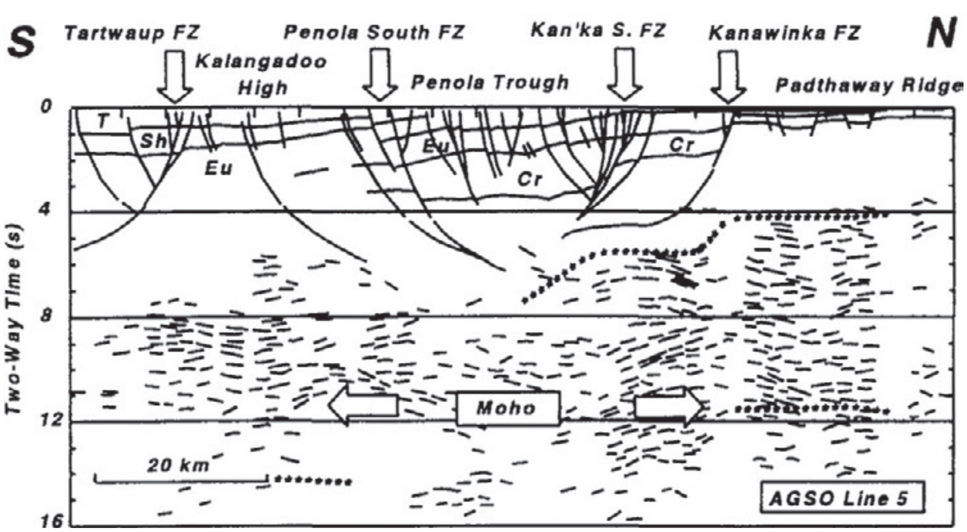
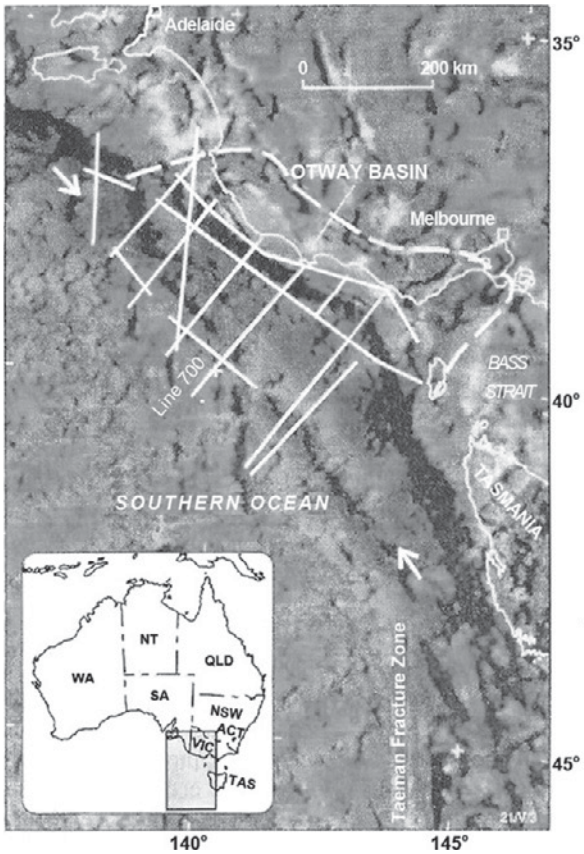



Figure 9.7.1-08. Line drawing and simplified basement features along 1992 AGSO line 5 in SE Australia (from Finlayson et al., 1996, fig. 8). [Tectonophysics, v. 264, p. 137–152. Copyright Elsevier.]



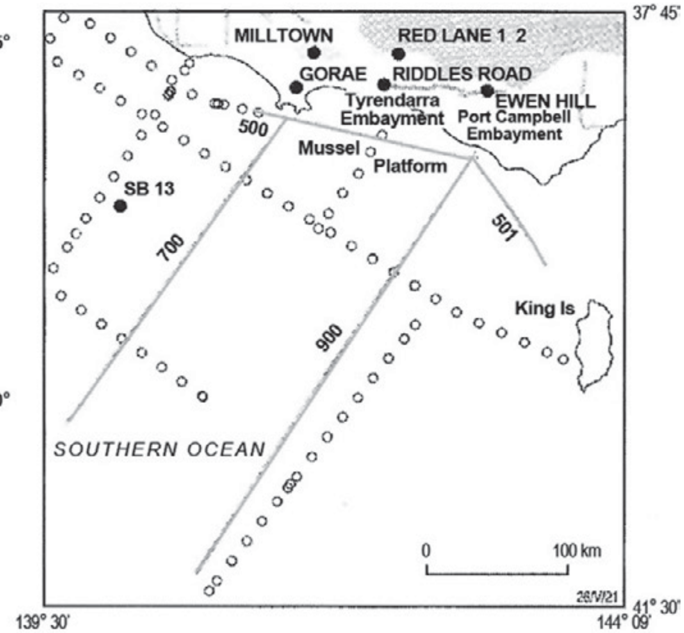




Figure 9.7.1-09. Location of the 1994–1995 offshore seismic reflection profiles in SE Australia (from Finlayson et al., 1998, figs. 1 and 4). Right: Full lines and black dots—offshore-onshore refraction recordings; open circles—other reflection lines. [Tectonophysics, v. 288, p. 177–189. Copyright Elsevier.]

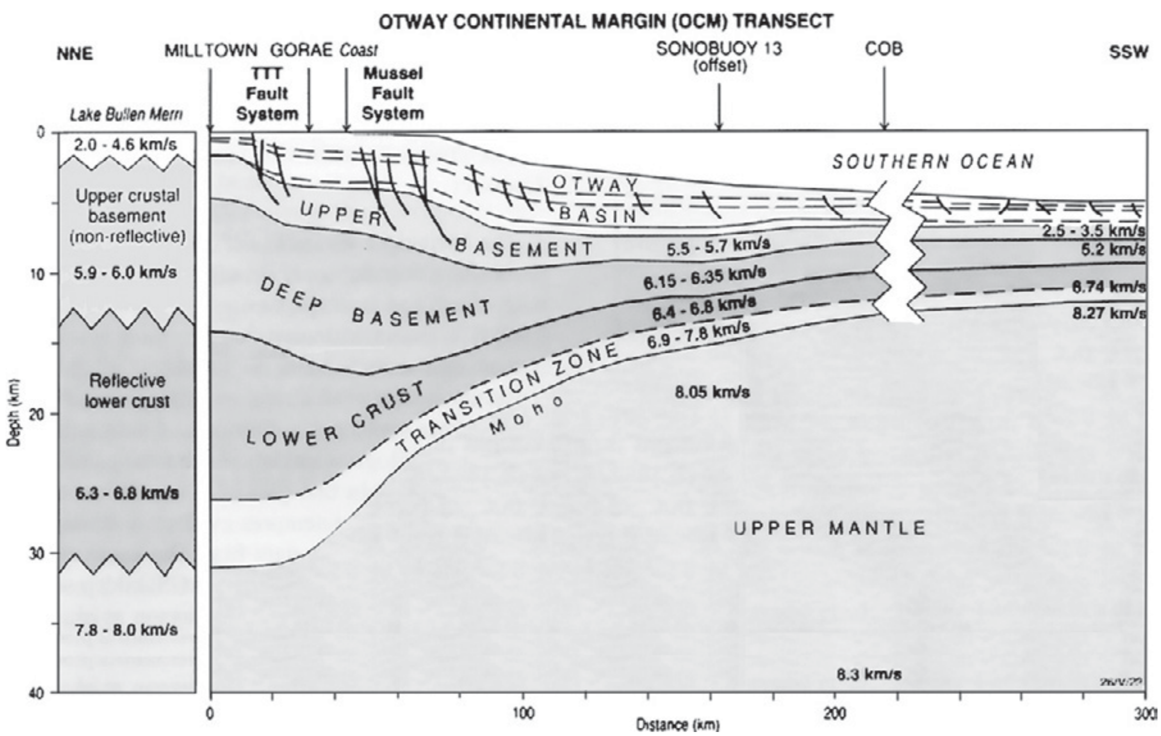



Figure 9.7.1-10. Cross section from the Otway Basin, SE Australia, across the continental margin based on the 1992 and 1994–1995 onshore and offshore seismic reflection profiles (from Finlayson et al., 1998, fig. 9). [Tectonophysics, v. 288, p. 177–189. Copyright Elsevier.]

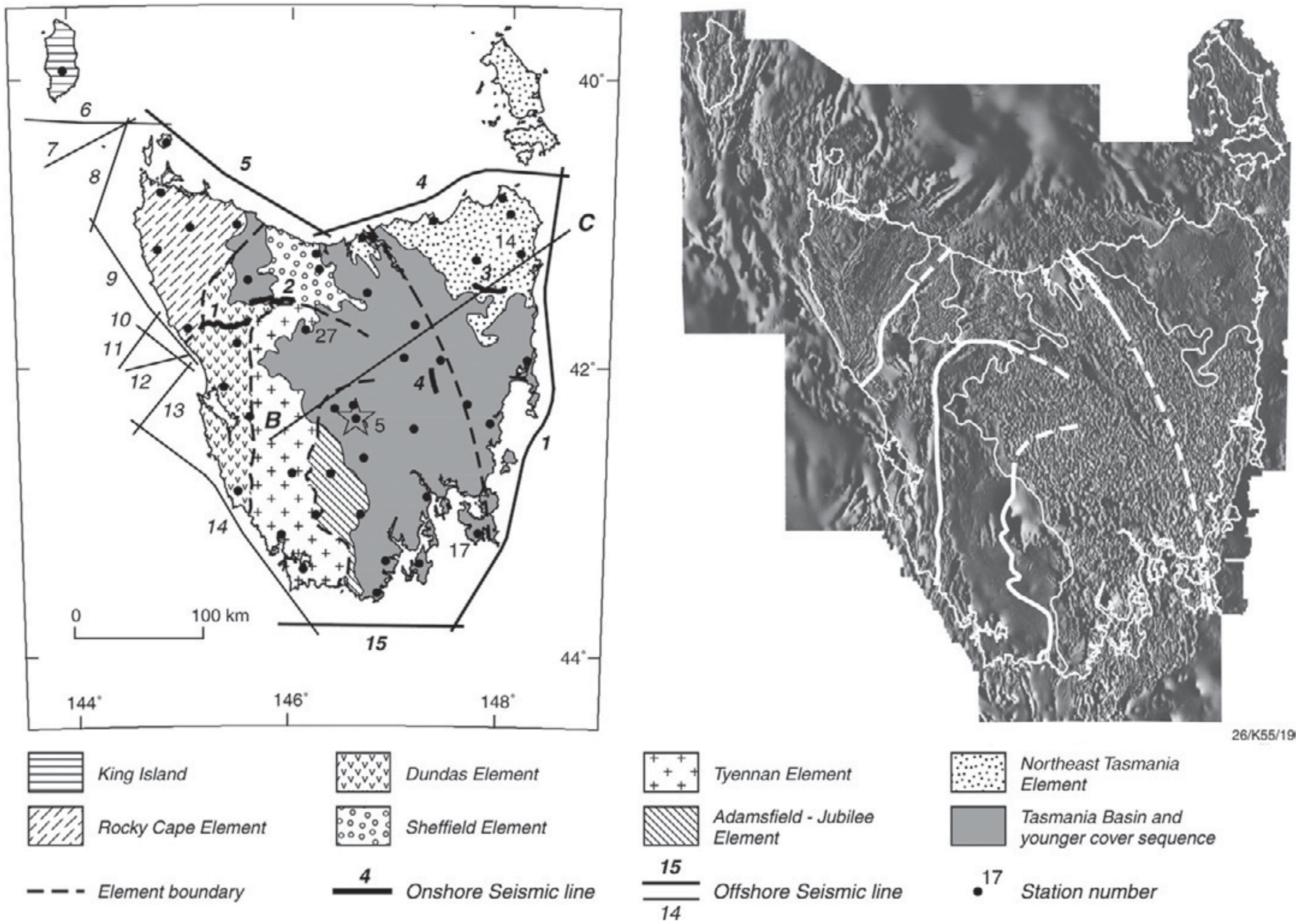



Figure 9.7.1-11. Location of the 1995 offshore and onshore seismic reflection profiles in and around Tasmania (from Drummond et al., 2000b, fig. 1a). Stars—locations of onshore blasts; thick short lines 1–4—onshore profiles; thin lines 1–15—offshore seismic lines; dots—stations of a seismic tomography survey. [Tectonophysics, v. 329, p. 1–21. Copyright Elsevier.]

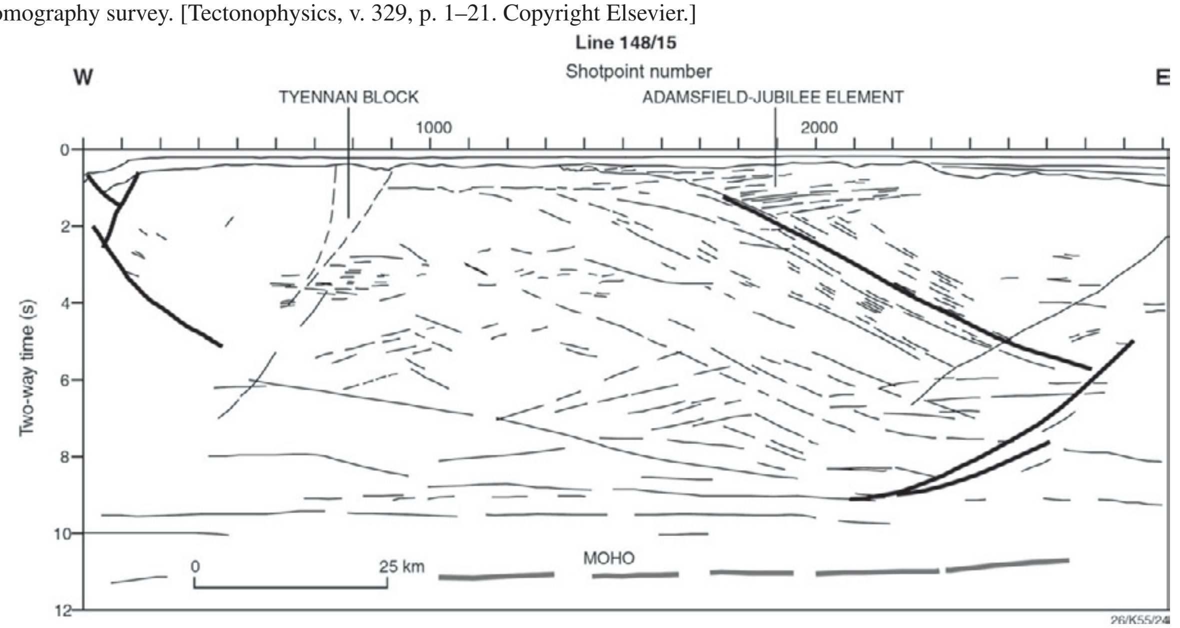



Figure 9.7.1-12. Line drawing of the 1995 offshore seismic reflection profile south of Tasmania (from Drummond et al., 2000b, fig. 8). [Tectonophysics, v. 329, p. 1–21. Copyright Elsevier.]

broadband stations. The latter ones were installed during two projects, where, from 1993 to 1996 during the SKIPPY project, a single array of 12 stations was moved across the continent in six deployments, and in the KIMBA project, an array of 10 stations was deployed in the Kimberley Craton in the northwest of Australia. Collins (1991) based his crustal thickness map primarily on the results of controlled-source seismology investigations, but the most recent, revised map (Collins et al., 2003) also included the results of Clitheroe et al. (2000). The depth to Moho under Australia shows significant variations (Fig. 9.7.1-13).

In general, the authors stated that within Archean regions of Western Australia the Moho appears to be relatively shallow with a large velocity contrast, while the Moho is significantly deeper under the Proterozoic North Australian platform, under central Australia and under Phanerozoic southeastern Australia. They also stated that there is a broad transition zone from crustal to mantle velocities where the Moho is deep (Collins et al., 2003). While usually the Moho is mapped as a continuous sub-horizontal boundary, based on averaging wide-angle data, from seismic-reflection profiling, due to the smaller wavelengths, offsets were observed where major crustal faulting reached down to the Moho or intersected the crust-mantle boundary, as was shown, e.g., for seismic surveys in central Australia or in Tasmania. In conclusion, the crustal thickness under Australia varied considerably, ranging from ~24 km along the coastlines and continental shelves to 56 km under central Australia and underneath the Lachlan Foldbelt in southeastern Australia.

### 9.7.1.2. New Zealand

In contrast to the Australian continent, New Zealand lies on the Australian/Pacific plate boundary with Pacific oceanic crust being subducted under eastern North Island, a transform boundary which transects the South Island, and oceanic crust of the Australian plate being subducted under southwestern South

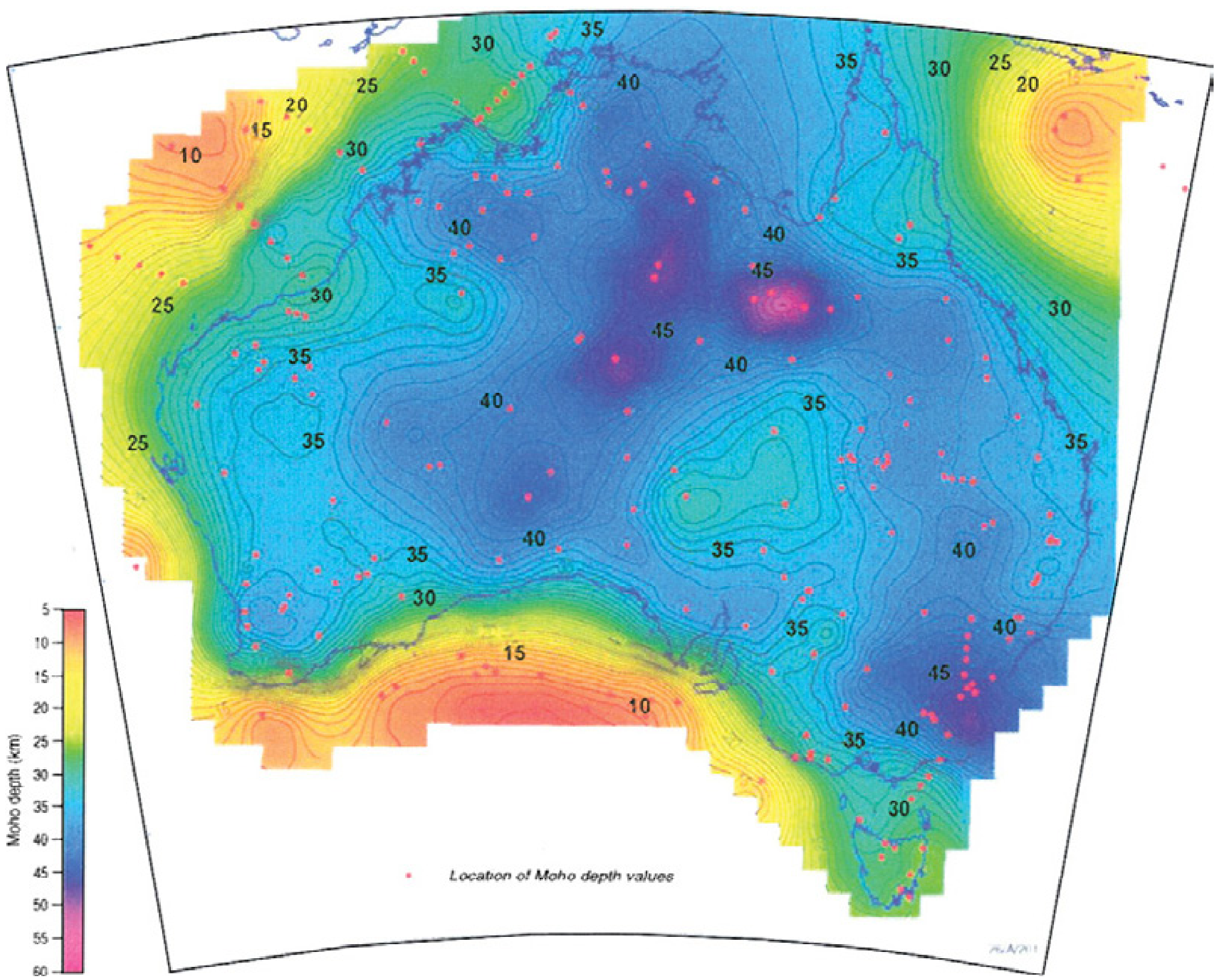



Figure 9.7.1-13. Crustal thickness map of Australia (from Collins et al., 2003, fig. 1). Contour interval—1 km. Red dots—location of depth values used to generate the contours. [Geological Society of Australia Special Paper 273, p. 121–128. Published with the permission of the Geological Society of Australia.]

Island (Sutherland, 1999). South Island is thus being deformed by the oblique collision of the oceanic Pacific plate, the eastern South Island and Campbell plateau to the southeast, with the continental Australian plate, the northwestern South Island and North Island and the Challenger Plateau to the northwest (Fig. 9.7.1-14). This deformation of South Island is marked by the rise of the Southern Alps. About 450 km of dextral strike-slip motion, 80 km of convergence and 20 km of uplift have occurred across this plate boundary since the Oligocene (e.g., Allis, 1986; Davey et al., 1998; Sutherland, 1999).

One of the main features of the North Island is the Central Volcanic Region which has been a zone of back-arc extension and arc volcanism associated with the subduction of the Pacific plate under the Australian plate along the Hikurangi margin. Its present activity is believed to be concentrated on the Taupo Volcanic Zone along its eastern margin (Davey and Lodolo, 1995).

As described in the previous chapters, detailed seismic surveys started in the 1970s and continued through the 1980s. In 1990 a crustal marine seismic-reflection profile of 360 km length was recorded along the north coast of North Island which crossed the Central Volcanic Region (Davey and Lodolo, 1995; Davey et al., 1995). The record length was 16 s TWT. The objective was to obtain a crustal section across the Hikurangi subduction zone and back-arc in the north of North Island. One major goal was

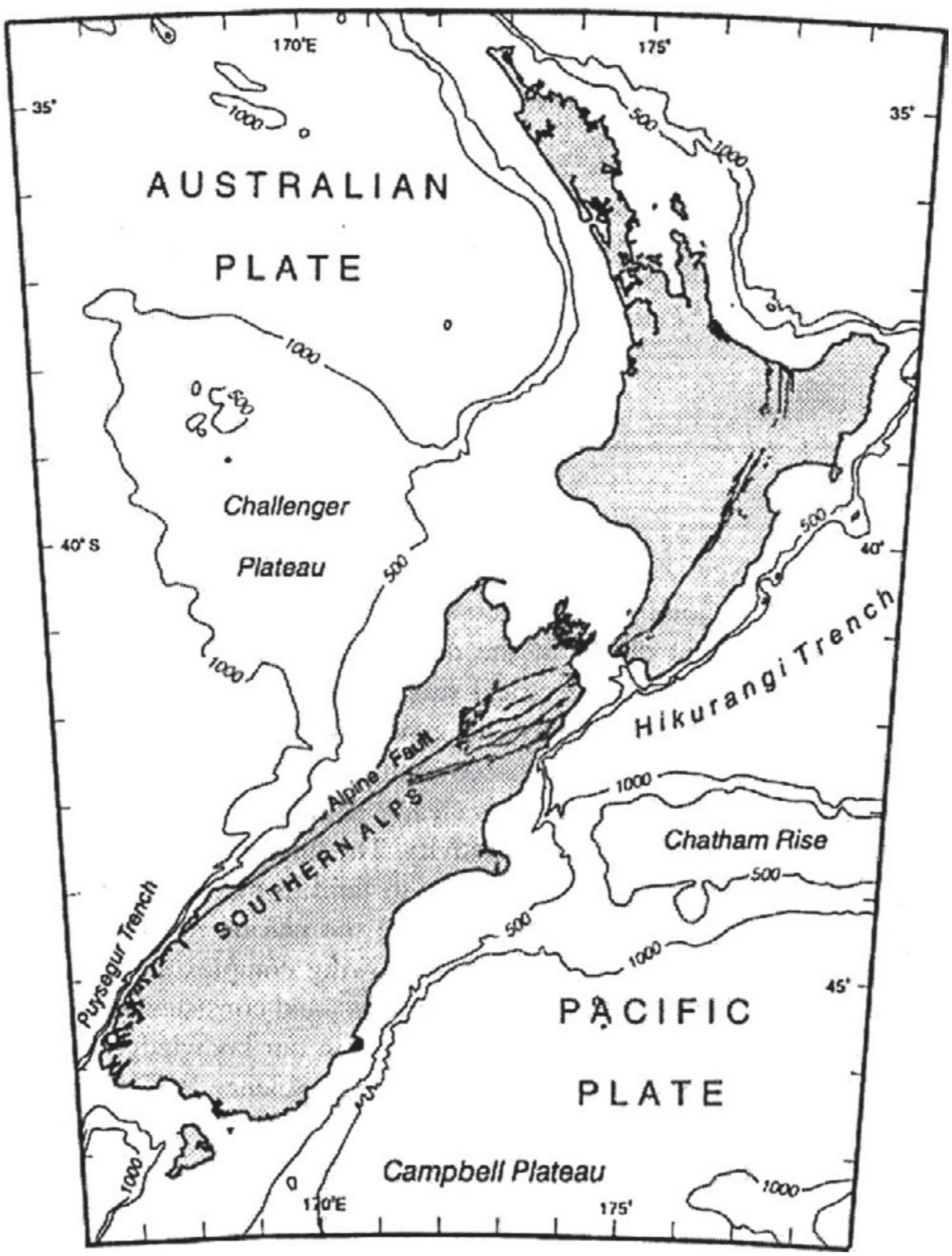



Figure 9.7.1-14. Location map of the New Zealand region (from Davey et al., 1998, fig. 1). [Tectonophysics, v. 288, p. 221–235. Copyright Elsevier.]

to delineate the near-surface structure and image the extensional tectonics inferred to form the Central Volcanic Region. A second goal was to image the lower crust–upper mantle structure, in particular to define the base of the crust which had been determined at 15 km depth by seismic-refraction work in the 1980s (Stern and Davey, 1987). The data have revealed mainly the complex tectonics of the near-surface structures. Under the presently active Taupo Volcanic Zone strong mid and lower crustal reflectors were found at depths of 4.5–5.5 s TWT under the volcanic ridge and western frontal graben which were associated with magma or volcanic sills. The deepest reflector was close to the refraction Moho of Stern and Davey (1987).

In 1994, another project, MOOSE (Marlborough Onshore Offshore Seismic Experiment), investigated the southern Hikurangi subduction zone at the northeastern part of South Island. It was a 300-km-long marine crustal reflection profile with 16 s record length, where normal-incidence and also sparse off- onshore wide-angle seismic data were collected (Henrys et al., 1995). A band of reflectors at 5–8 s TWT could be correlated with the base of a 12–15-km-thick upper crustal layer of velocity 5.0–5.5 km/s underlying part of South Island.

Also in 1994, a 27-km-long crustal seismic-reflection profile was recorded in the central South Island using explosive charges. Its aim was to obtain a crustal section over the eastern part of the Pacific-Australian plate boundary in central South Island (Kleffman et al., 1998).

A third project in 1994 was a marine seismic-reflection survey of 80 km in length in the area of Stewart Island, south of South Island. The recording times were 16 s TWT (Davey, 2005). These data were interpreted together in the context with the marine seismic-reflection profile along the eastern coast of South Island, which was part of SIGHT (see below) and a short third profile along the south coast of Stewart Island recorded in 1996. These seismic lines which partly overlapped provided a crustal seismic image from the southern end of South Island to south of Stewart Island (Davey, 2005). The data provided constraints on the evolution of the Gondwana margin in this region, because lower crustal and upper mantle seismic reflectivity off southeastern Stewart Island were interpreted in terms of a northeast-dipping paleosubduction zone.

The investigations in the area of Stewart Island were continued in 1996 (Melhuish et al., 1999). A marine seismic-reflection line of 610 km in length was recorded from the southern tip of Stewart Island toward WNW through the Solander Basin and across the Puysegur Bank and Trench south of South Island. In the Solander Basin, the seismic-reflection data showed well developed reflections to depths of 30–35 km (12–13 s TWT). On the basis of lower crustal reflectivity, beneath the adjacent Stewart Island shelf the base of the crust was determined at ~30 km depth (9 s TWT), rising westwards to ~20 km depth (8 s TWT) under the Solander basin. Prominent dipping reflections were interpreted as an active fault which could be traced to 30 km depth (~12 s TWT), where they merged into a zone of strong reflectivity in the upper mantle (Melhuish et al., 1999).

To study the characteristics of the Southern Alps orogen of New Zealand and to understand the collisional processes involved in continental collision, in 1995 and 1996, a joint U.S.–New Zealand geophysical project was undertaken which involved active source and passive seismology, magnetotelluric and geoelectric studies, petrophysics, geological mapping and gravity measurements (Okaya et al., 2007). The principal field activity during the southern summer was an integrated onshore and onshore-offshore seismic-refraction and wide-angle reflection experiment, carried out across the South Island and extending ~200 km offshore (Fig. 9.7.1-15) and named SIGHT (South Island Geophysical Transect). The advantage of the offshore-onshore geometry of SIGHT was the illumination of the lower crust and mantle by dense shots offshore into land stations from both sides of the orogen allowing high resolution of the crust and upper mantle structure (Davey et al., 1998; Godfrey et al., 2001; Okaya et al., 2002).

The experiment had two main components. The first was a wide-angle reflection-refraction experiment along the two land transects (profiles 1 and 2 in Fig. 9.7.1-15) across central South Island. Twenty-three chemical explosives with charges of 350–1200 kg were equally spaced, with 16 shots on the northern transect and 7 shots on the southern transect. Four hundred twenty recording instruments were deployed along each transect with a nominal spacing of 400 m. Data examples of line 2 are shown in Figure 9.7.1-16.

The second component consisted of three offshore-onshore transects, two along the profiles 1 and 2 and a third along southeastern South Island. This third transect supplied tie lines across the eastern parts of the two main transects. During this phase, data were recorded by 225 land stations, spaced at ~1.5 km intervals along the two profiles 1 and 2 and at ~10 km intervals along the tie lines of the third transect. Twenty ocean-bottom seismograph/hydrophone instruments (OBS/H) were deployed offshore, first on the west side of South Island, and airgun shots were fired at 50 m intervals along eight western profiles in total, each ~200 km long. The OBS/Hs were then redeployed on the east side of South Island when shooting the profiles on this side of the island. Energy from the onshore shots was clearly recorded to maximum distances of 160 km, while airgun shots reached to recording distances of 300 km. The vessel firing the shots also recorded near-vertical incidence multichannel seismic data to 16 s TWT along all eight profiles (Fig. 9.7.1-15).

To record both active (transect) and passive source data, SAPSE (Southern Alps Passive Seismic Experiment), in addition to 17 permanent New Zealand stations, broadly distributed 26 broadband and 15 short-period recording stations across the South Island, but centrally weighted toward the central Alpine fault and transect region (black and white circles in Fig. 9.7.1-15).

The preliminary modeling (Fig. 9.7.1-17) of both explosion and onshore-offshore data indicated a crust of fairly constant P-wave velocity overlying a lower crustal layer with seismic velocity of ~7 km/s and 5–10 km thick (Davey et al., 1998; Stern and McBride, 1998). The total crustal thickness varied from ~30 km


Figure 9.7.1-15. Location of the SIGHT transects, OBS/OBH locations and SAPSE recording sites across the South Island of New Zealand (from Davey et al., 1998, fig. 2). [Tectonophysics, v. 288, p. 221–235. Copyright Elsevier.]

at the east coast to ~42 km under the Southern Alps, whereby the deeper crustal layer was interpreted as possibly corresponding to old oceanic crust. The data also provided evidence of deep reflections from within the dipping Alpine fault zone and delayed lower crustal phases for west coast seismographs from shots east of the Southern Alps, interpreted as due to a low-velocity zone associated with the Alpine fault zone (Davey et al., 1998; Stern and McBride, 1998). In the marine data clear lower crustal-Moho reflections were imaged along most of the tracks, indicating that Moho lies at ~10 s TWT off the west coast and 8 s off the east coast.

The double-sided onshore-offshore seismic imaging of the South Island of New Zealand by "super-gathers" across a plate boundary was discussed in more detail by Okaya et al. (2002) with respect to its application in other areas around the world

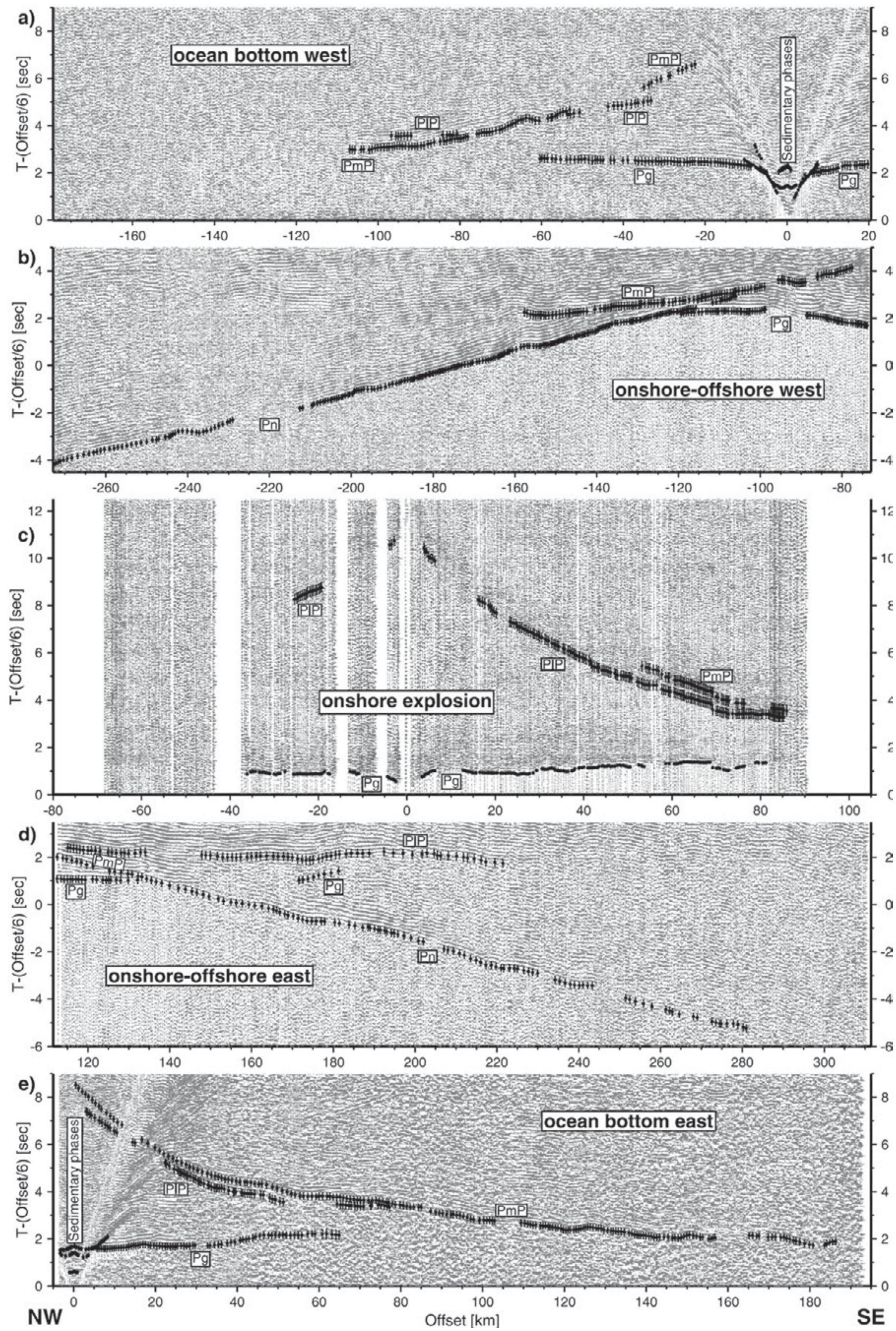



Figure 9.7.1-16. Examples of line 2 wide-angle reflection data across the South Island, New Zealand, derived from the onshore and offshore data (from Scherwath et al., fig. 2). [Journal of Geophysical Research, v. 108 (B12), 2566, doi: 10.1029/2002JB002286. Reproduced by permission of American Geophysical Union.]

where a transition between continental and oceanic environments coincides with the existence of islands or narrow landmasses allowing for a double-sided experiment.

A full interpretation of the two main transects (Scherwath et al., 2003; Van Avendonk et al., 2004) showed up to 17 km of crustal thickening to form a 44-km-deep crustal root that was offset 10–20 km to the southeast of the highest mountains, with a velocity reversal below the midcrust in the eastern part of the crustal root and a strong thickening of the 7.0 km/s lower crust within the root (Fig. 9.7.1-18).

In 1998 a 65 km seismic-reflection transect (SIGHT98) was shot onshore on South Island across the MacKenzie Basin to Mount Cook Village to provide a detailed crustal image in the central Southern Alps. Fifty kg shots at 1 km intervals were recorded by an 800 channel array, providing a CDP survey with vertical seismic profiling in central South Island (Long et al.,

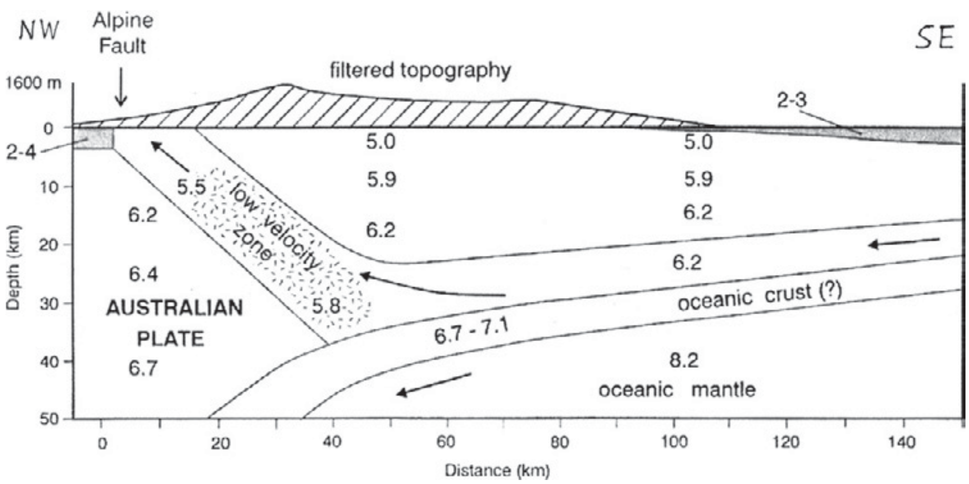



Figure 9.7.1-17. Preliminary crustal model across the South Island, New Zealand, derived from the onshore wide-angle data of profile 2 (from Davey et al., fig. 6). [Tectonophysics, v. 288, p. 221–235. Copyright Elsevier.]

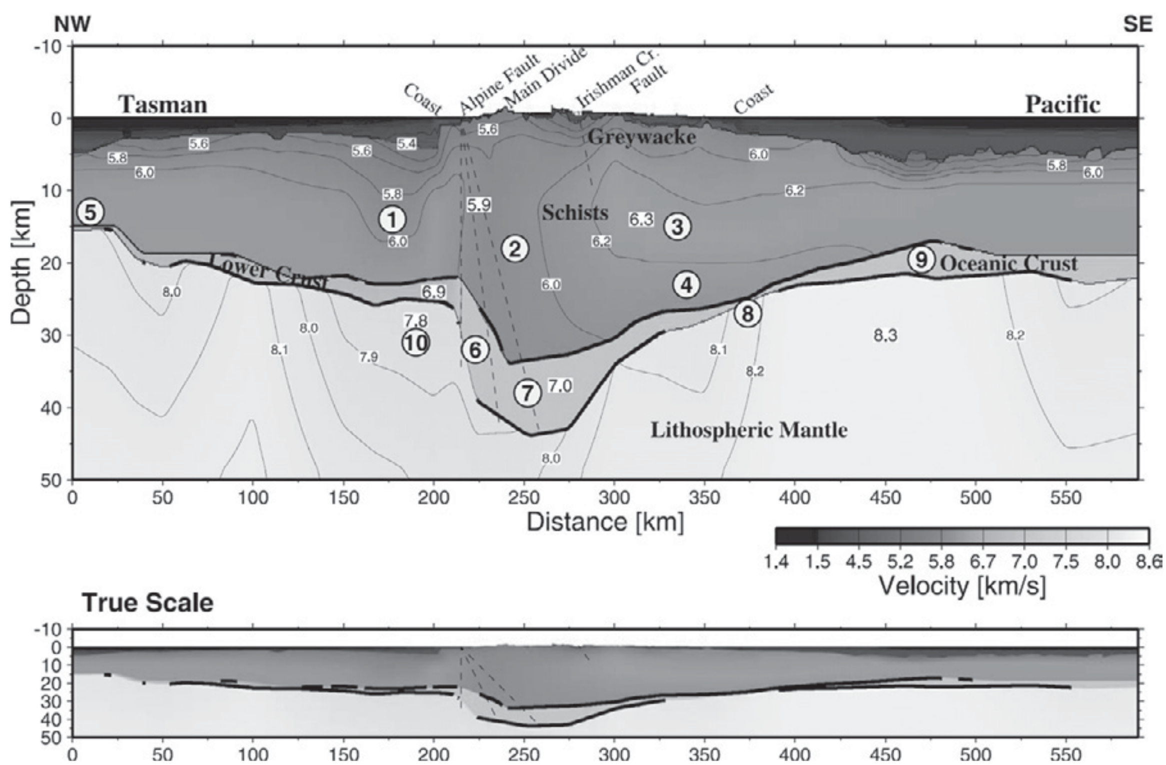



Figure 9.7.1-18. Final crustal model across the South Island, New Zealand, derived from the onshore and offshore data (from Scherwath et al., fig. 4). [Journal of Geophysical Research, v. 108 (B12), 2566, doi: 10.1029/2002JB002286. Reproduced by permission of American Geophysical Union.]

2003; Okaya et al., 2007). The maximum array aperture was nearly 35 km and the maximum shot-receiver offset ~30 km. The raw seismic data were recorded up to 30 s time and processed in two parts. For the first study, only the first 5 s TWT were processed with a maximum offset of 14 km and imaged the upper 12–15 km of the crust (Long et al., 2003). These data did not image any major regional-scale reflections or discontinuities in the uppermost 5 s. Instead, more subtle, smaller-scale features were present, interpreted as to mark faults, bedding, or lithological contacts. The deeper data imaged a band of reflectivity inferred to correspond to a lower crustal décollement that turned upwards at its western end to align with the Alpine fault at the surface (Stern et al., 2007).

### 9.7.2. South Africa

Besides the large-scale projects in the East African rift system (see subchapter 9.5.1), active-source seismic projects were also carried out in the 1990s in Namibia. In 1995, the project MAMBA (Geophysical Measurements Across the Continental Margin of Namibia experiment), a combined onshore-offshore multichannel and wide-angle seismic survey, included a major land component along two transects (for location, see Fig. 9.8.4-02) traversing deeply eroded basement rocks of the late Precambrian Damara Orogen onshore and extending offshore out across the ocean-continent transition. Airgun shots were recorded simultaneously on land and at sea. For the interpretation receiver gathers were compiled for each station (Bauer et al., 2000). More details are described in subchapter 9.8.4.

A second project, ORYX of the German GeoForschungs-Zentrum Potsdam, was carried out in 1999. The project aimed to investigate details of the crustal structure crossing the Waterberg fault zone south of the Erongo Mountains in central Namibia. The special aim was a combined near-incidence and wide-angle seismic-reflection survey using 35 mobile six-component stations of GFZ Potsdam, each supplied with 6 vertical geophones (180 seismic channels) along two 18-km-long lines across the Waterberg fault zone. Conventional CDP-style seismic processing was extended by P- and S-wave tomography of the shallow subsurface. Results were not yet published.

### 9.7.3. Central and South America

Since the first beginnings of crustal structure studies in the central plateau of Mexico in the 1950s (Meyer et al., 1958, 1961a) and in the Andean region of Peru in the 1950s (Tatel and Tuve, 1958) and 1960s (Ocola and Meyer, 1972), Central and South America have been the target of continued international research projects, some of which continue until today.

In the southeast of Mexico, the International Continental Deep Drilling Program (ICDP) initiated deep crustal studies of the Chicxulub meteorite impact crater. The continental margin of Central America became the target of a German project, with an onshore-offshore project at the coast of Costa Rica. In the north and northeast of South America, crustal structure studies were undertaken in Venezuela and Brazil. The majority of crustal investigations, however, concentrated on the region of the Andes, where since the 1960s the Free University of Berlin, Germany, had initiated detailed geological research work which in the 1980s and 1990s was finally complemented by intensive onshore and offshore geophysical crustal and upper-mantle studies.

#### *9.7.3.1. Mexico and Costa Rica*

The Chicxulub impact crater, located on the present coast of the Yucatan peninsula and expressed onshore as a semicircular ring of sinkholes with a diameter of ~160 km (Pope et al., 1996) was recognized as such in 1980 (Alvarez et al., 1980). Widely distributed drill holes sampled cores of the post-impact Tertiary sediments and showed that the sediments thicken to more than a kilometer in the interior of the basin (Ward et al., 1995).

In 1996, BIRPS (British Institutions Reflection Profiling Syndicate) acquired ~650 km of deep seismic-reflection profiles off the northern coast of the Yucatan peninsula in Mexico (Snyder et al., 1999; Christeson et al., 2001; Morgan et al., 2002).

The data were recorded to 18 s TWT on a 240-channel, 6-km-long streamer, using a 50 m shot spacing. The 13,000 airgun shots were also recorded at wide-angle on 34 OBSs, provided by the University of Texas at Austin, and by 91 land recorders supplied by PASSCAL (Fig. 9.7.3-01).

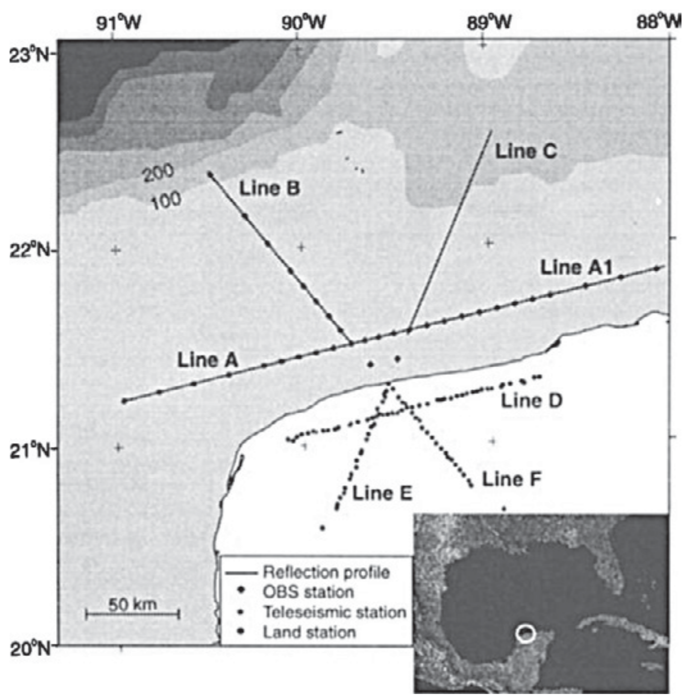



Figure 9.7.3-01. Location of the 1996 Chicxulub seismic experiment, showing BIRPS seismic reflection survey lines, OBS, and land stations (from Snyder et al., 1999, fig. 1). Inset—Gulf of Mexico, showing location of the Chicxulub impact crater on the Yucatan peninsula. [Journal of Geophysical Research, v. 104, p. 10,743–10,755. Reproduced by permission of American Geophysical Union.]

The shallow water limited the acquisition of marine seismic data to distances of not closer than 25 km from the crater center which is located close to the coastline.

The seismic-reflection data of the Chicxulub deep seismic-reflection profile A/A1 were converted to depth using the velocity information from wide-angle OBS data (Fig. 9.7.3-02), imaging the crater fill and impact lithologies and structures to the base of the crust at ~35 km depth (Christeson et al., 2001). Using in particular the higher-resolution data covering the uppermost 8 km (4-s record length), the new seismic offshore data revealed a ring around the peak and three annular zones where the style of deformation changed at discrete radial distances of 55–65 km and 85–98 km (Snyder et al.,1999).

Morgan et al. (2002) inverted the whole seismic-refraction data set in 3-D using first-arrival travel-time tomography and found a high-velocity body within the central crater region that most likely represents lower-crustal rocks which were stratigraphically uplifted during the formation of this complex crater. A suite of checkerboard tests confirmed that the observed velocity anomalies were real, in spite of the irregular experimental geometry of the refraction data.

A second seismic experiment took place in the beginning of 2005, organized by a team of scientists from Mexico, the United States, and the UK (Morgan et al., 2005). At this time, 1500 km of reflection profiles were acquired and 36,560 airgun shots were recorded on 28 ocean-bottom seismometers and 87 land seismometers (Fig. 9.7.3-03).

A subset of the seismic data constituted a site survey for two proposed Integrated Ocean Drilling Project (IODP) drill holes (Chicx-01A and Chicx-02A in Fig. 9.7.3-04).

The seismic data of the 1996 experiment had allowed the conclusion that the crater is 180–200 km in diameter, and in 2002 Chicxulub was drilled onshore at Yaxcopil-1 (Yax-1 in Fig. 9.7.3-04) under the auspices of the ICDP. Besides the proposals for the two offshore drill holes Chix-01A and Chix-02A, another onshore drill hole (ICDP-hole 2 in Fig. 9.7.3-04) has been proposed close to the crater center. Initial analyses of the new seismic data indicated that important information could be obtained of how large impact craters are formed. They also promise important constraints on changes in lithology across the crater (Morgan et al., 2005).

In Costa Rica, the convergent margin off the Nicoya Peninsula had been the site of single- and multichannel seismic-reflection surveys (von Huene and Flueh, 1995). Within the project TICOSECT (Trans Isthmus Costa Rica Scientific Exploration of a Crustal Transect) project in 1995 a major seismic-refraction survey was conducted offshore Costa Rica near the Nicoya

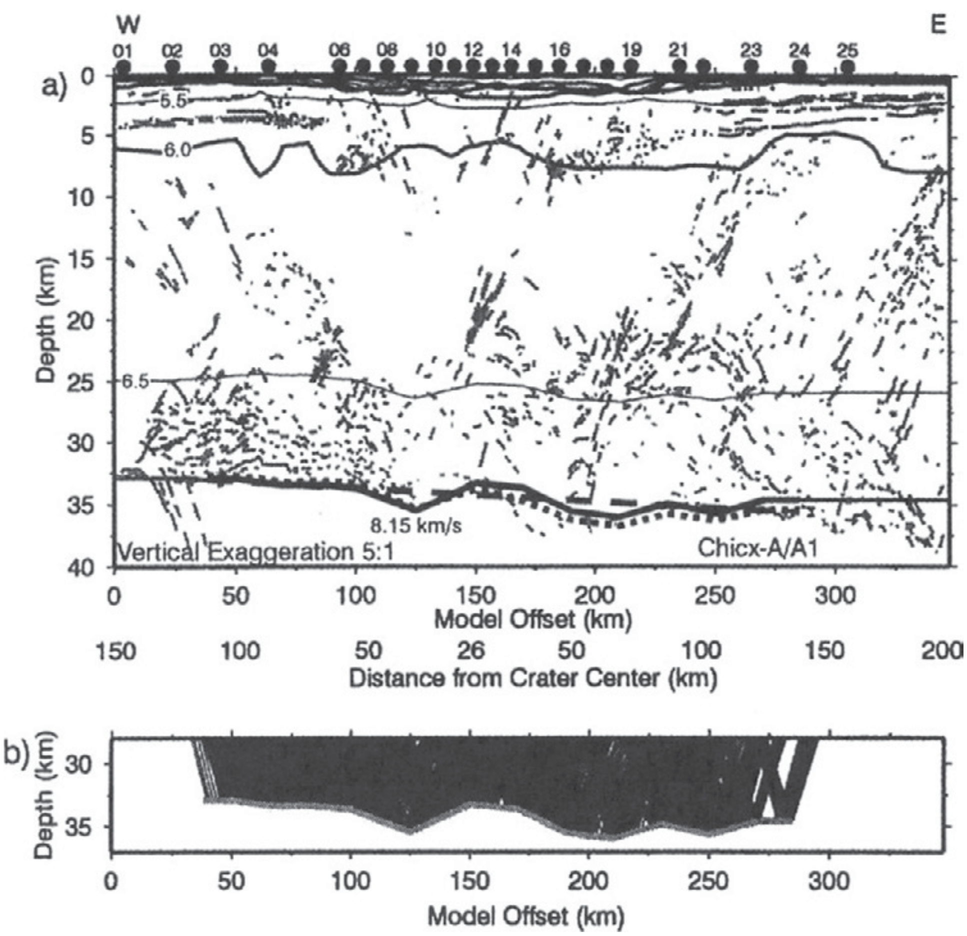



Figure 9.7.3-02. Line drawing for Chicxulub deep seismic reflection profile A/A1, converted to depth using the velocity information from wide-angle OBS data (from Christeson et al., 2001, fig. 9). [Journal of Geophysical Research, v. 106, p. 21,751–21,769. Reproduced by permission of American Geophysical Union.]

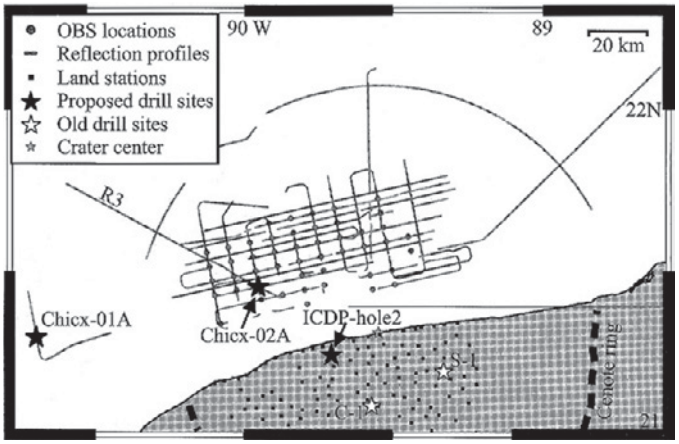



Figure 9.7.3-03. Location of the 2005 Chicxulub seismic experiment (from Morgan et al., 2005, fig. 1). [Eos (Transactions, American Geophysical Union), v. 86, no. 36, p. 325, 328. Reproduced by permission of American Geophysical Union.]

Peninsula. Using 22 ocean-bottom hydrophones and seismographs and a number of land stations as well, a dip profile and three strike profiles were recorded (Christeson et al., 1999). The land recorders extended the dip line for ~50–60 km onto the Nicoya Peninsula. The airgun profiles were carried out by R/V *Maurice Ewing* with a 20-gun, 138-L array and was recorded on a 100-m two-channel streamer. The dip line was shot twice, at both 50 m and 125 m shot spacing. The strike line closest to the shore, running approximately along the 1500 m bathymetric depth contour, was shot with 50 m shot spacing; the other two strike lines, running approximately above the 3000 m and 3500 m water depth contour lines, were shot with 125 m spacing.

The joint onshore-offshore data of the 1995 survey were used to construct a crustal structure model of the convergent margin from 20 km seaward of the Middle American Trench onto the Nicoya Peninsula (Christeson et al., 1999). At 60–80 km off the coastline the Moho was found at 11 km depth below sea level and then dipped gradually to ~25 km depth at ~10 km inland from the coastline. The top of the subducted slab was well resolved, it deepens from 5 km depth at the trench to 15–16 km depth at the Nicoya Peninsula coastline, the dip angle of the subducting plate increasing from 6° to 13° at a distance of ~30 km from the trench.

The project was continued in 1996 as part of the COTCOR (Comparative Transactions in Costa Rica) international collaboration project of GEOMAR (Kiel, Germany), ICE (San José, Costa Rica) and ICTJA-CSIC (Barcelona, Spain). Borehole shots with charges from 550 to 1230 kg TNT were fired twice at five locations and recorded by 60 portable stations, occupying alternatively one half of the 170-km-long profile, resulting in 120 recordings. The 1996 profile was located coincident with the 1995 onshore profile (Sallares et al., 1999). The resulting crustal model was combined with the 1995 results to a 250-km-long crustal section, from the outer rise to the back-arc basins. The oceanic Moho reaches 10 km depth west of the Middle American Trench, deepening progressively toward the isthmus to reach an angle of seduction of 35° down to 40 km depth. The obtained onshore model is characterized by significant changes in the surface velocity (2.0–5.0 km/s), an upper crust of 4 km with velocities of 5.3–5.7 km/s, a mid-crust with a mean thickness of 13 km and velocity variations of 6.2–6.5 km/s and an up to 20-km-thick lower crust with velocities between 6.9 and 7.3 km/s. The Moho is reached at 38–40 km depth.

### *9.7.3.2. Northeastern South America*

In Venezuela, two deep wide-angle seismic projects were undertaken to investigate the crust and upper-mantle structure of the Guayana Shield and the Oriental Basin to the north.

The field campaign ECOGUAY (Estudios de la Estructura Cortical del Escudo de Guayana) in 1998 targeting the Guayana Shield comprised nine seismic-refraction lines of up to 320 km distance (Fig. 9.7.3-05), using blasts of iron mines, gold mines, and a quarry with ripple firing as energy sources. The profiles were partly reversed and were obtained by choice scrolling the available 13 seismic stations in up to four deployments for each shotpoint. The spacing of recording points varied between 5 and 10 km (Schmitz et al., 2002).

In 2001, a second experiment followed, this time targeting the Oriental Basin north of the Guayana Shield. The study area of ECCO (Estudio Cortical de la Cuenca Oriental) comprised a region from the Coastal Cordillera and Interior Ranges in the north, crossing the entire Oriental Basin (eastern Venezuela basin) up to the Guayana Shield in the south (Fig. 9.7.3-06). Also this project used mine blasts and borehole blasts as energy sources.

The north-south line (squares in Fig. 9.7.3-06) was 300 km long and ran from the Caribbean Sea across the Oriental Basin

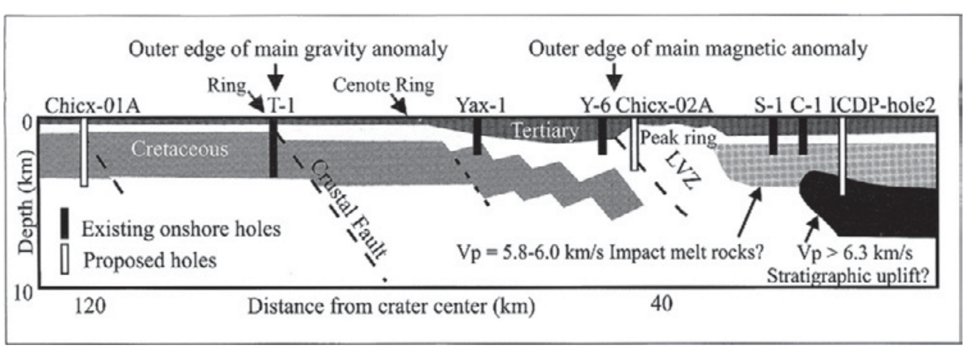



Figure 9.7.3-04. Cartoon of the Chicxulub crater based on marine seismic reflection, offshore-onshore refraction, and drill hole data (from Morgan et al., 2005, fig. 3). [Eos (Transactions, American Geophysical Union), v. 86, no. 36, p. 325, 328. Reproduced by permission of American Geophysical Union.]

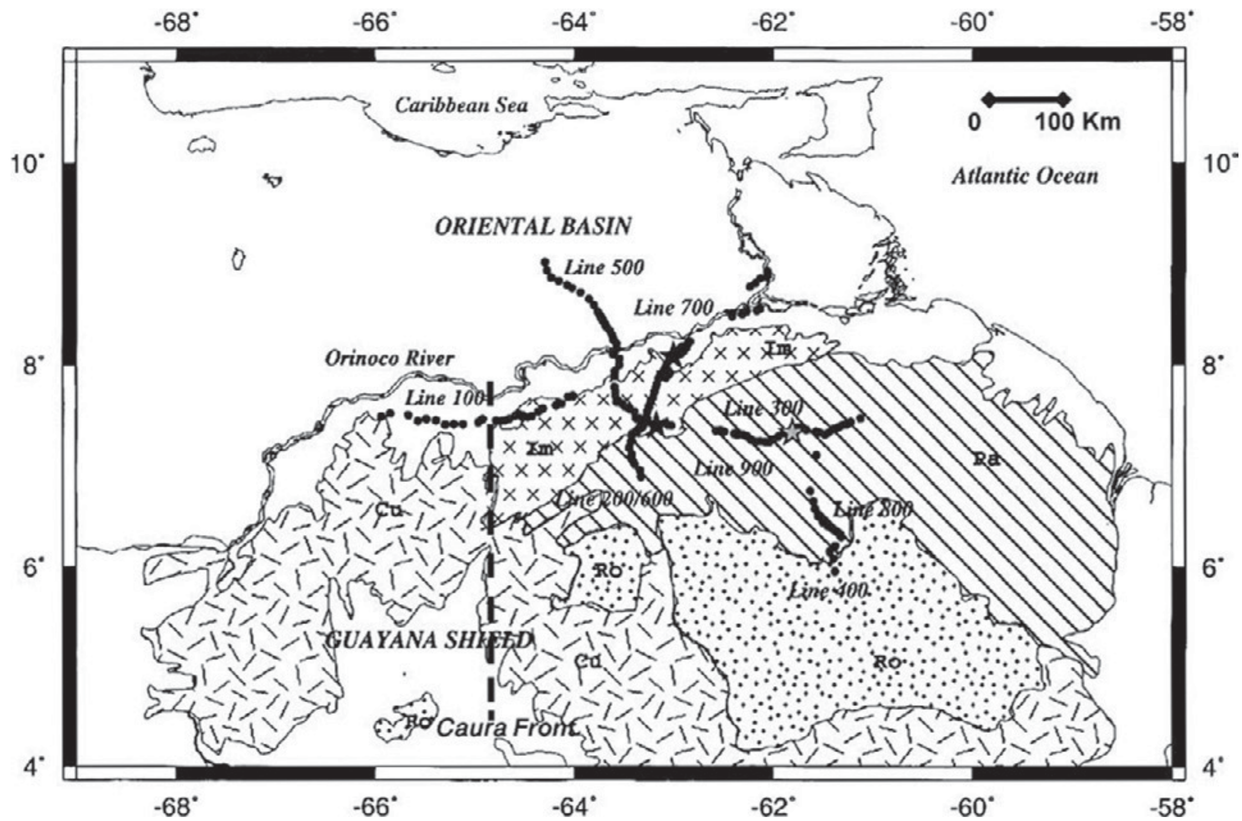



Figure 9.7.3-05. Location of the 1998 seismic experiment investigating the Guayana Shield in Venezuela (from Schmitz et al., 2002, fig. 1). [Tectonophysics, v. 345, p. 103–118. Copyright Elsevier.]

into the Guayana Shield. It comprised five shotpoints, spaced 50–100 km apart, with charges of 150–500 kg, which were recorded by 190 one-component stations (Texans) with a spacing between 1.5 and 2 km (Schmitz et al., 2005). Two other profiles of this project are shown by triangles and circles in Figure 9.7.3-06.

The interpretation of the 1998 data revealed a subdivision of the crust into a 20 km upper crust with velocities of 6.0–6.3 km/s, and a lower crust with velocities of 6.5–7.2 km/s below 20–25 km depth. The total crustal thickness of the Guayana Shield varied from west to east and ranged from 46 km in the west (Archean segment) to 43 km in the east (Proterozoic segment). The average crustal velocity was ~6.5 km/s (Schmitz et al., 2002).

Toward the north, underneath the Oriental Basin, a sedimentary thickness of up to 13 km was derived, while the total crustal thickness decreased from 45 km beneath the Guayana Shield to 39 km at the Orinoco River, and 36 km close to El Tigre in the center of the Oriental Basin (Fig. 9.7.3-07). The average crustal velocity decreased in the same direction from 6.5 to 5.95 km/s.

Starting in 1997, combined geological and geophysical studies aimed to investigate the deep structure of central Brazil. For this reason a seismic-refraction survey was planned. The project started with a trial experiment in late 1997 (Berrocal et al., 2004). It consisted of 20 digital recorders, deployed along a 260-km-long, N-S–oriented profile with two seismic sources with 800 kg

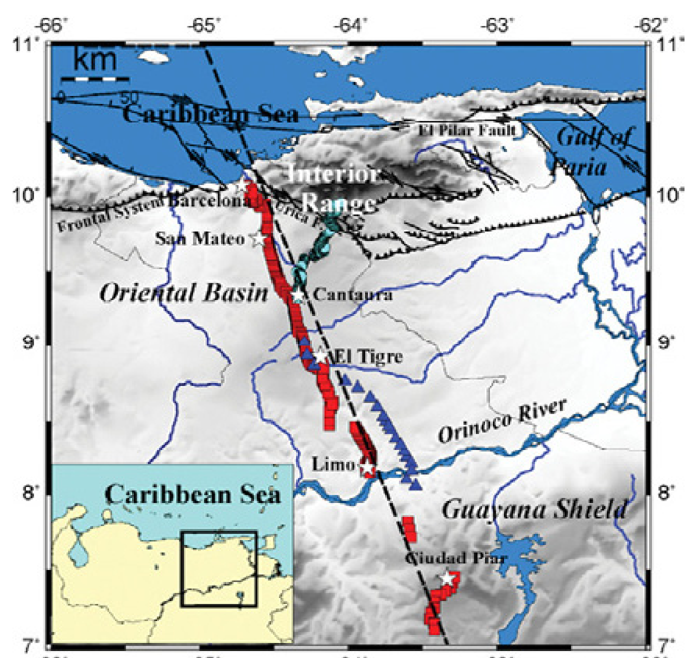



Figure 9.7.3-06. Location of the 2001 seismic experiment investigating the Oriental Basin in Venezuela (from Schmitz et al., 2005, fig. 1). [Tectonophysics, v. 399, p. 109–124. Copyright Elsevier.]

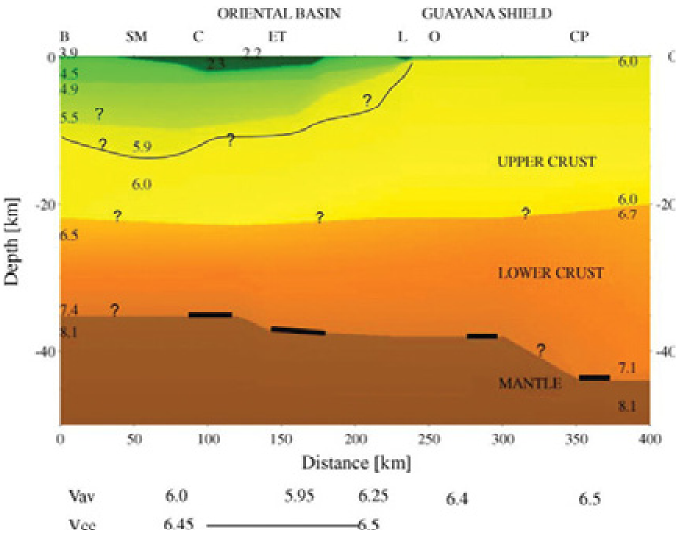



Figure 9.7.3-07. Crustal structure along a N-S section from the Oriental Basin to the Guayana Shield in Venezuela (from Schmitz et al., 2005, fig. 10). [Tectonophysics, v. 399, p. 109–124. Copyright Elsevier.]

charges at each end of the profile. Some of the stations were permanent seismographic stations of the Serra de Mesa network and Brasilia array. This experimental line served mainly to define the shot sizes for the main seismic-refraction lines. Following the successful test experiment, three main seismic profiles were recorded in central Brazil (L1, L2, L3 in Fig. 9.7.3-08).

The profiles L1 and L2, located in the central section of Tocantins Province, were each ~300 km long and were aligned in WNW-ESE direction. On each profile, 120 stations were deployed at 2.5 km spacing, and explosive sources were detonated in boreholes, spaced 50 km apart on each line. The charges used varied from 1000 kg at the end of the lines to 500 kg for the intermediate shots. The two lines overlapped each other by 50 km. Line L3 was located in the southeastern sector of Tocantins Province and ran in NE direction. It started in the SW in the Parana basin, crossed the Brasilia fold belt and ended on the Sao Francisco craton. On this line, only shots from the ends of the line and some middle shots were successful.

The interpreted model for lines L1 and L2 (Fig. 9.7.3-09) showed a crustal thickening from west to east, where the depth to the Moho varied from 32 to 43 km. The mean crustal velocity was around 6.3 km/s. For line L3 on the southeastern Tocantins Province a thick two-layer crust of 41–42 km was determined both under the Parana basin and the Sao Francisco craton, but a thinner crust of ~38 km was found under the Brasilia fold belt (Berrocal et al., 2004).

### *9.7.3.3. Southern Andean Region of South America*

The successful research work of the geoscientific interdisciplinary research group “Mobility of active continental margins” of the Free University of Berlin, Germany, under the leadership of Peter Giese from 1982 to 1989 was only the beginning of a large interdisciplinary research of the Andes and the adjacent

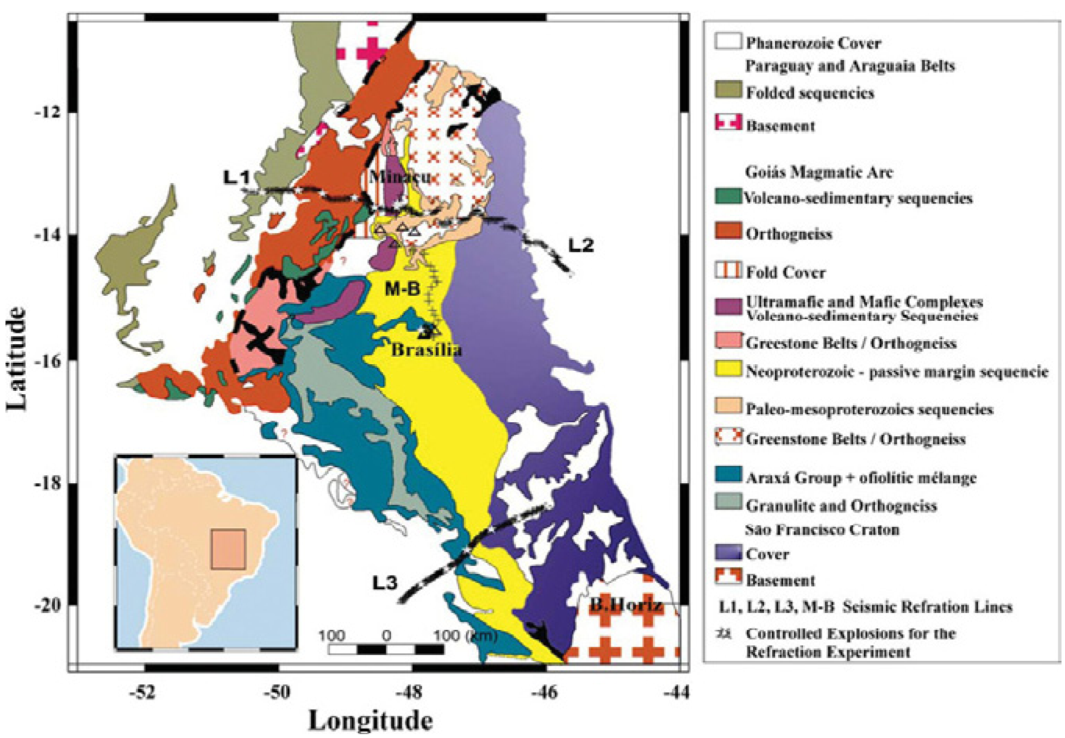



Figure 9.7.3-08. Geological map of Tocantins Province, central Brazil, with location of seismic-refraction lines (from Berrocal et al., 2004, fig. 1). [Tectonophysics, v. 388, p. 187–199. Copyright Elsevier.].

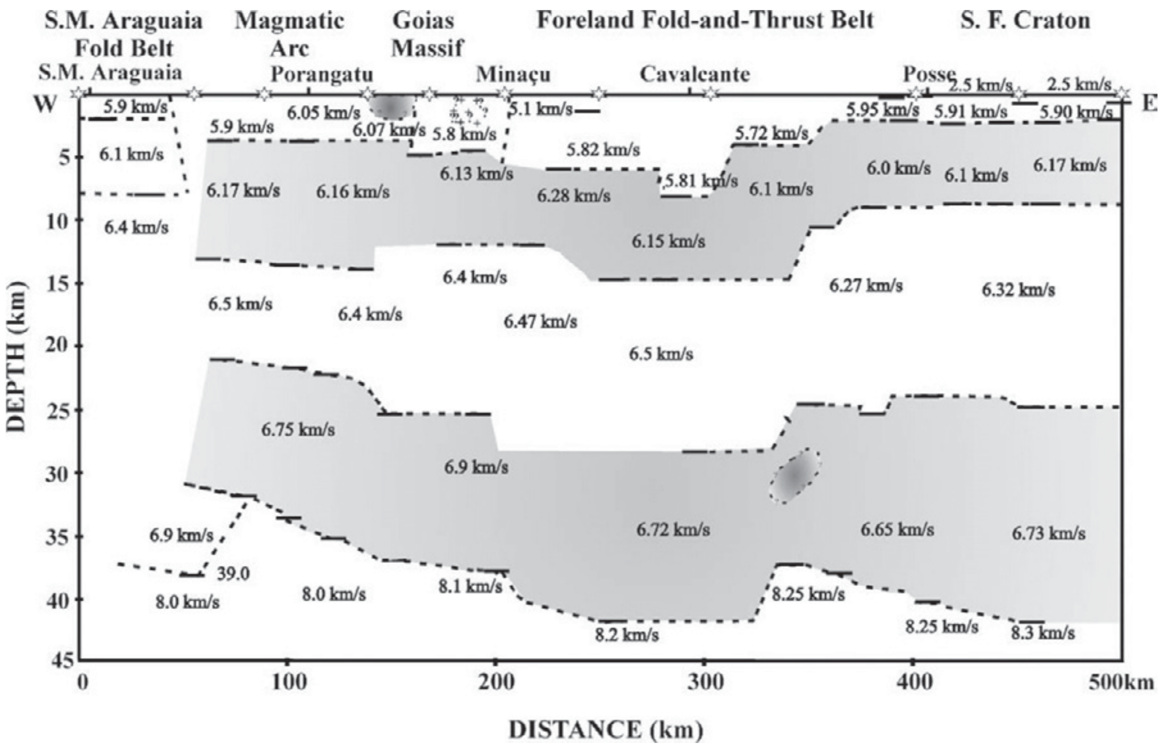



Figure 9.7.3-09. Seismic model of the central sector of Tocantins Province, central Brazil (from Berrocal et al., 2004, fig. 6). [Tectonophysics, v. 388, p. 187–199. Copyright Elsevier.].

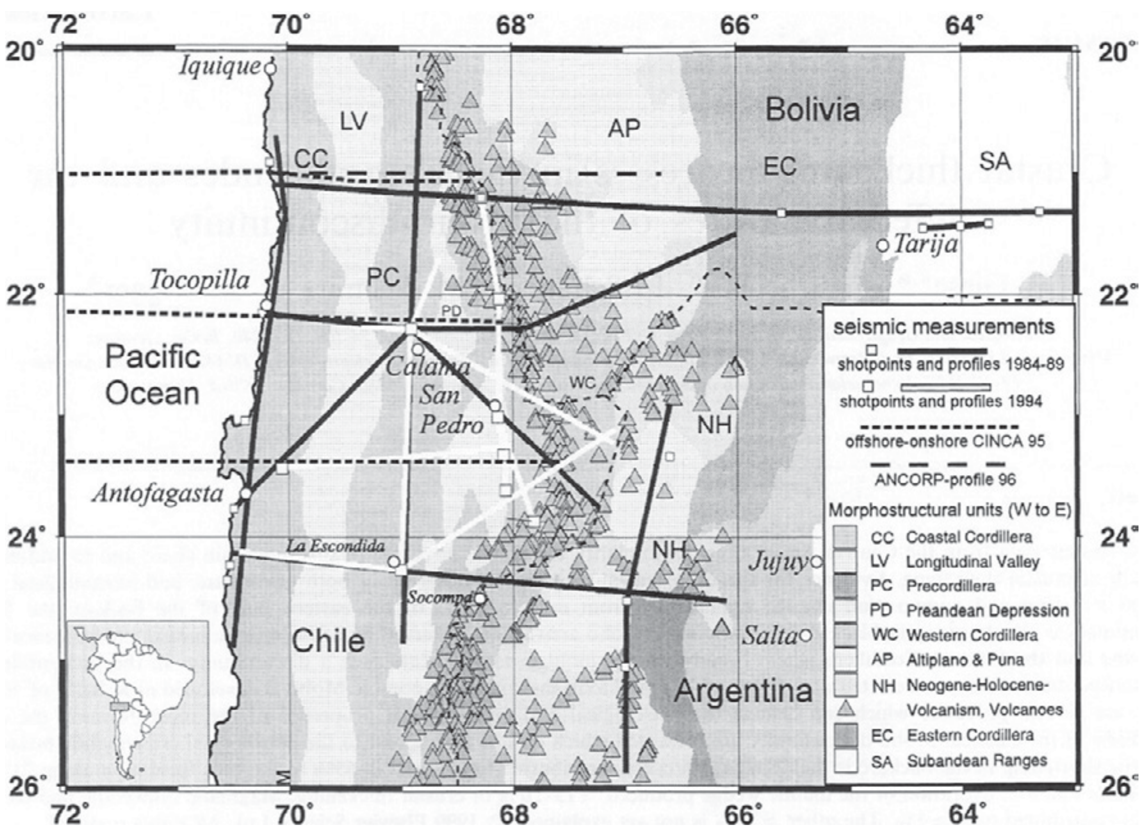



Figure 9.7.3-10. The onshore network of the main seismic refraction profiles in the Central Andes (from Giese et al., 1999, fig. 1). [Journal of South American Earth Science, 12, p. 201–220. Copyright Elsevier.]

Pacific Ocean in the 1990s which went on into the first years of the twenty-first century. The core of this research was formed by the foundation of the Collaborative Research Center 267 (CRC 267) "Deformation Processes in the Andes" at the Free University of Berlin and the University of Potsdam (Germany), which was funded by the German Research Society for 15 years. Soon after its foundation, the GeoScience Center at Potsdam, Germany, became a partner in this project and as such increased its research power. The new research center not only enabled extended seismic research, but involved all geoscientific branches available at the three research institutions and included furthermore a strong support from various South American research facilities.

In the years 1994–1996, three major seismic investigations were performed in northern Chile and adjacent parts of Bolivia and Argentina (Fig. 9.7.3-10): the project PISCO 94 (Proyecto de Investigaciones Sismologicas de la Cordillera Occidental 1994), the project CINCA 95 (Crustal Investigations off- and onshore Nazca plate/Central Andes 1995), and the project ANCORP 96 (Andean Continental Research Project 1996) after a successful test project PRECORP in 1995.

The project PISCO 94 (white lines in Fig. 9.7.3-10) was the first seismic campaign in the 1990s (Schmitz et al., 1999; Lessel 1998). The main profile of 320 km length extended in a N-S direction from shotpoints OLL to VAR (Fig. 9.7.3-11), approximately along the longitude 68°W, along the western border of the recent magmatic arc, the western Cordillera. In its southern part it crossed the Pre-Andean depression (Salar de Atacama). The station elevations ranged from 2300 m to 5000 m above NN; their spacing was on average 2.5 km, accomplished by 4 deployments recording shots from 5 shotpoints along the line. In the southern half, along the Salar de Atacama, with a sedimentary sequence of up to 8 km thickness, an additional N-S line with an additional shotpoint at its southern end was arranged where a station spacing of 1 km was achieved. In addition, the seismic profiles of the 1980s radiating from the mine Chucicamata (CHU in Fig. 9.7.3-11) were partly reoccupied, additional shotpoints were arranged, and a seismological network consisting of 31 stations distributed in an area of 200 × 200 km was in operation for 100 days recording more than 5300 local seismic events and also all the shots (Graeber and Asch, 1999). In total, 33 shots were recorded with charges ranging from 50 kg to 1300 kg, fired in patterns of up to 20 holes with depths from 10 to 50 m. The shotpoints were either located in salt lakes or in sedimentary basins. Data examples as published by Schmitz et al. (1999) are shown in Figure 9.7.3-12 and Appendix A9-5-1.

In 1995, a second project, CINCA 95, followed. It was a combined land-sea experiment (dashed lines in Fig. 9.7.3-10 and white lines in Fig. 9.7.3-14) and was jointly organized by the German Geological Survey (BGR Hannover), the German research institution GEOMAR (Kiel) and the CRC 267, the joint interdisciplinary research group of the Free University at Berlin and the GeoScience Center at Potsdam (Patzwahl, 1998; Patzwahl et al., 1999; Appendix A9-5-1). It comprised three west-east running lines crossing the Peru-Chile Trench offshore and

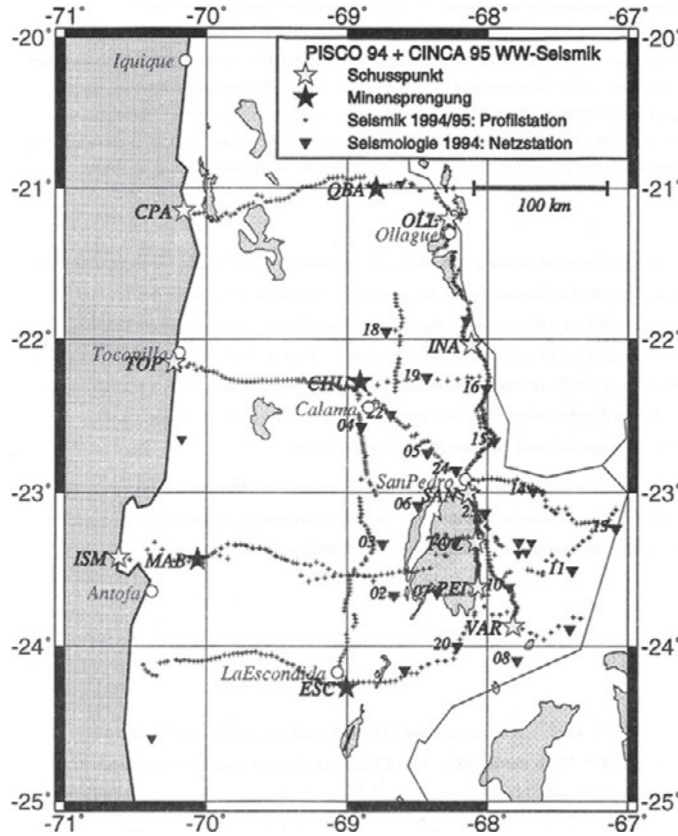



Figure 9.7.3-11. Location of the seismic refraction profiles of the projects PISCO'94 and CINCA'95 (stars—shotpoints, dots—recording sites) in the Central Andes onshore and offshore and the location of the seismological network PISCO'94 (triangles) in the Central Andes (from Lessel, 1998, fig. 4.3). [Berliner Geowissenschaftliche Abhandlungen, Reihe B, Band 31: 185 p. Published by permission of Institut für Geologische Wissenschaften, Freie Universität Berlin.]

reached to the magmatic arc of the western Cordillera in northern Chile. The BGR offshore reflection lines are shown as white lines in Figure 9.7.3-14.

Furthermore, in 1995, the PRECORP seismic-reflection experiment (Fig. 9.7.3-13) was carried out as a feasibility study for a longer deep seismic-reflection survey across the Central Andes. The 50-km-long profile was located at 22.5°S south of Calama, with shot and receiver locations extending in an E-W direction. Thirty-eight shot gathers were recorded using a maximum of 144 receivers with a spacing of 100 m. The maximum offset was 14.3 km. The shotpoint interval of the split-spread recording geometry was 2.4 km, providing twofold coverage (Yoon et al., 2003).

Its successful operation allowed a full-scale long seismic-reflection experiment of ~400 km length, the project ANCORP 96 (ANCORP Working Group, 1999, 2003). Its realization was enabled by a successful cooperation between the CRC 267 of Berlin-Potsdam, Germany and university and governmental research institutions of Chile and Bolivia. As shown in Figure 9.7.3-14, it continued the northern E-W offshore line of CINCA 95 at 21°S and extended across the Coastal Cordillera, the Longi-

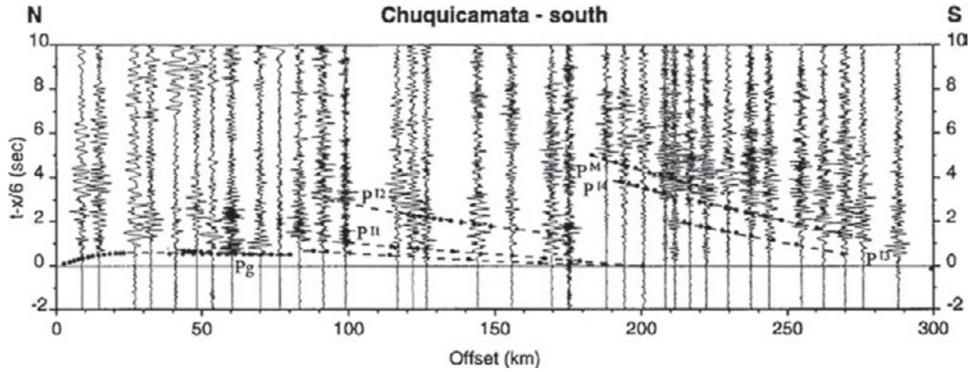



Figure 9.7.3-12. Record section of the profile Chuquicamata-S in the Precordillera (from Schmitz et al., 1999, fig. 6a). $P_g$, $P^{I1}$, $P^{I2}$ are phases from the upper and middle crust, $P^M$ branch is interpreted as reflection from the recent geophysical Moho. [Journal of South American Earth Science, 12, p. 237–260. Copyright Elsevier.]

tudinal Valley, the Precordillera, the western Cordillera, and the Altiplano (Figs. 9.7.3-13 and 9.7.3-14).

Recording was achieved by 56 independently operating 3- and 6-channel PDAS stations with strings of vertical geophones of 4.5 Hz natural frequency, resulting in 252 recording channels. With a spacing of 100 m a spread length of 25 km was achieved. During the roll-along operation, the back 6.25 km of the spread was removed daily and added to the front end.

The aim of the project was to provide an image of the deep lithosphere. Near-vertical profiling was performed with a relatively wide shot spacing of 6 km. Each shot was fired twice off-end into the 252 recording channels at 100 m spacing, resulting in a 50-km-long split spread when the spread had completely moved. The resulting subsurface coverage was fourfold. Shots of 90 kg were fired in 20 m deep boreholes, except for 5 shots in Bolivia that had a charge size of 300 kg. Figure 9.7.3-15 shows the automatic line drawing of the migrated ANCORP 96 reflection line data (see also Appendix A9-5-1).

From 10 locations along the ANCORP transect, repeated shots were fired into the 25-km-long geophone line to achieve seismic observations at large offsets. These shots for wide-angle recording, which were detonated each time, after the recording spread had obtained a full new position, were repeated up to 9 times with increasing charges of up to 450 kg and served in particular for velocity control. One of these wide-angle shotpoints was 6 km offshore, operated by the Chilean Navy, and one was at a mine using regular quarry blasts. The resulting 70 shot sections of 25 km length each reached maximum observation distances between 120 and 230 km.

The combined interpretation of the processed near-vertical incidence reflection data, the wide-angle refraction data, and the data of a subsequent seismological monitoring of local and teleseismic earthquakes by a widespread 30-station array for 3 months continuously allowed workers to derive a rather detailed cross section (Fig. 9.7.3-16a) and its geodynamic interpretation (Fig. 9.7.3-16b).

In an overview, Giese et al. (1999) summarized the seismic, seismological and other geophysical investigations and compiled a representative crustal and uppermost-mantle model for the Andean region of Chile and adjacent areas at latitude 21–22°S

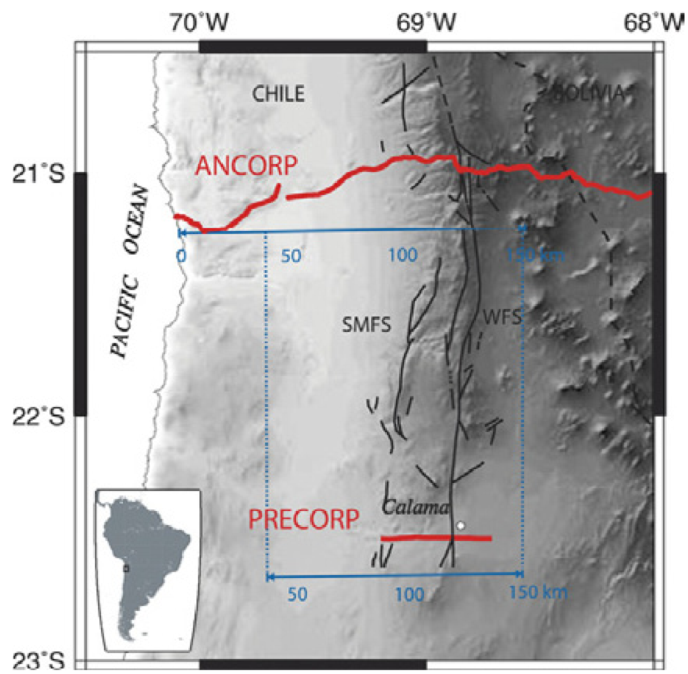



Figure 9.7.3-13. Location of the seismic reflection profiles PRECORP'95 and ANCORP'96 (from Yoon et al., 2003, fig. 1). [Geophysical Research Letters, v. 30: no. 4, 1160, doi: 10.1029/2002GL015848. Reproduced by permission of American Geophysical Union.]

(Fig. 9.7.3-17). In total, the cross section covered an area from the Pacific Ocean to the Brazilian craton, a distance range of more than 800 km. In the forearc and in the eastern backarc, the geophysical Moho was well defined by seismic-refraction data. Beneath the western Cordillera and the western Altiplano, the depth to the Moho transition zone could be derived from P to S converted waves (Giese et al., 1999; Scheuber and Giese, 1999).

Besides northern Chile, central Chile became the target of an expedition of the German *Sonne*. Also in 1995, within the multidisciplinary CONDOR (Chilean Offshore Natural Disaster and Ocean Environmental Research) project, an amphibic operation (Fig. 9.7.3-18) investigated the Valparaiso Basin offshore Valparaiso, central Chile, along two marine-seismic reflection and

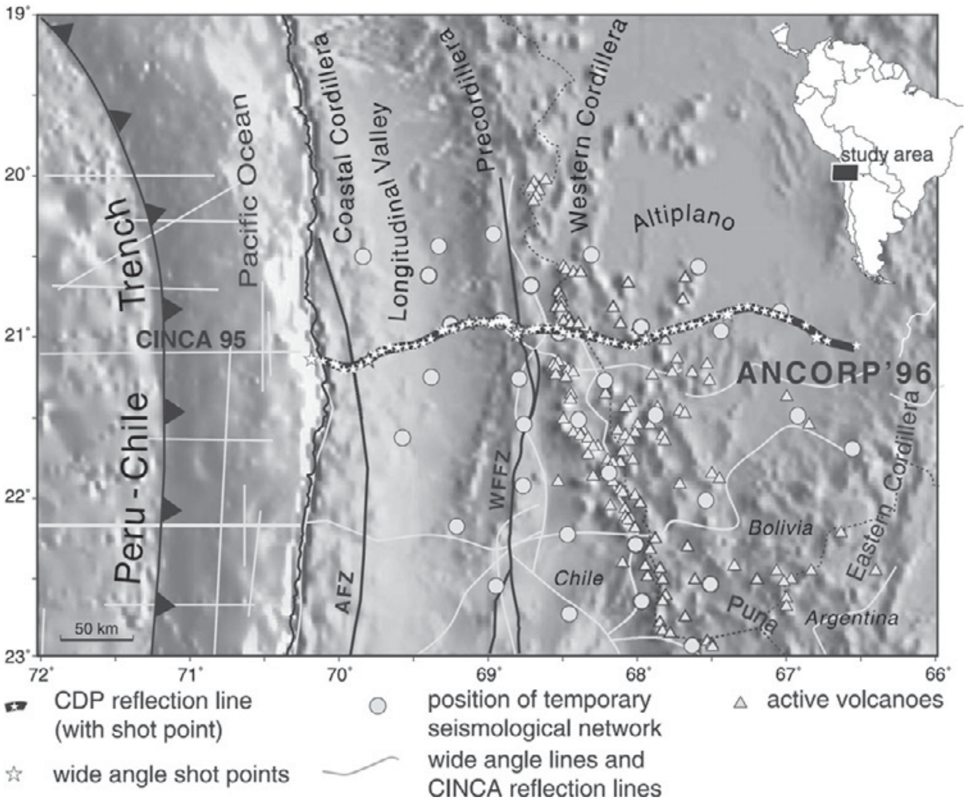



Figure 9.7.3-14. Location of the seismic reflection line ANCORP'96 (from ANCORP Working Group, 2003, fig. 1). Also shown are the CINCA'95 wide-angle and reflection lines onshore and offshore (white lines) and the ANCORP'96 seismological network in the Central Andes (white circles). [Journal of Geophysical Research, v. 108, B7, 2328, doi: 10.1029/2002JB001771. Reproduced by permission of American Geophysical Union.]

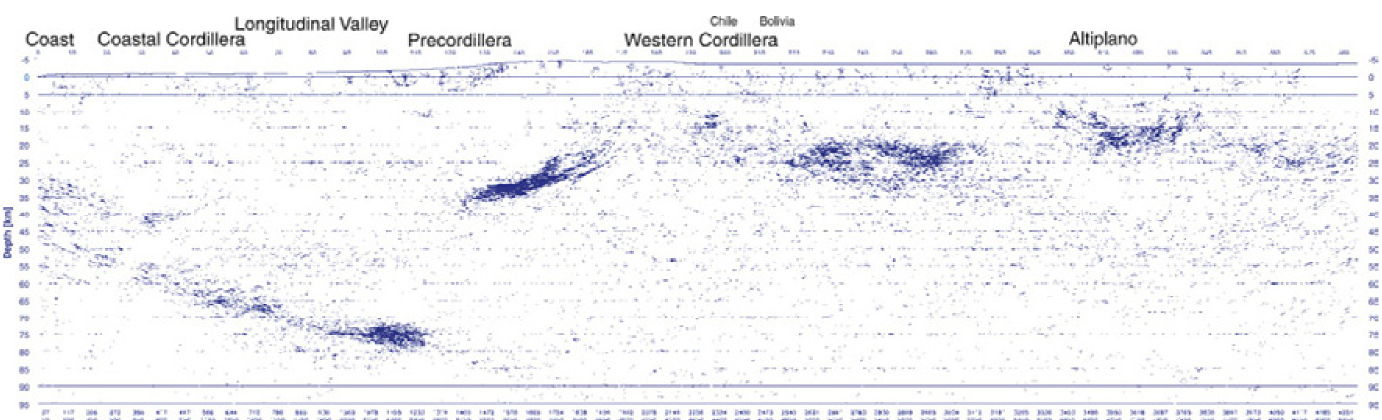



Figure 9.7.3-15. Automatic line drawing of the migrated ANCORP'96 reflection line data (from ANCORP Working Group, 2003, foldout 2b). Only laterally coherent phases were enhanced. [Journal of Geophysical Research, v. 108, B7, 2328, doi: 10.1029/2002JB001771. Reproduced by permission of American Geophysical Union.]

refraction profiles north and south of the latitude 33°S (Flueh et al., 1998a). The project was operated by the BGR (German Geological Survey, Hannover) and GEOMAR (Kiel).

The profiles crossed the continental margin and the Chile Trench, where thick trench sediments had been observed. Profile 1 crossed the rupture area of the 1985 Chile earthquake. Multichannel seismic-reflection data were obtained by a 3-km-long streamer recording airgun shots. Wide-angle measurements were performed by OBH recording devices (digital, high data capacity ocean-bottom recorders) and by land stations prolonging the 60–100-km-long marine profiles up to 200 km on land (data examples in Fig. 9.7.3-19 and Appendix A9-5-1). The crustal velocity models (Fig. 9.7.3-20) showed that the continental upper-plate crust extended to the middle-lower slope boundary and that the crustal structure of the oceanic plate appeared rather similar on both profiles (Flueh et al., 1998a).

In 2000, the Collaborative Research Center SFB 267 at Berlin and Potsdam, Germany, started a detailed research project in

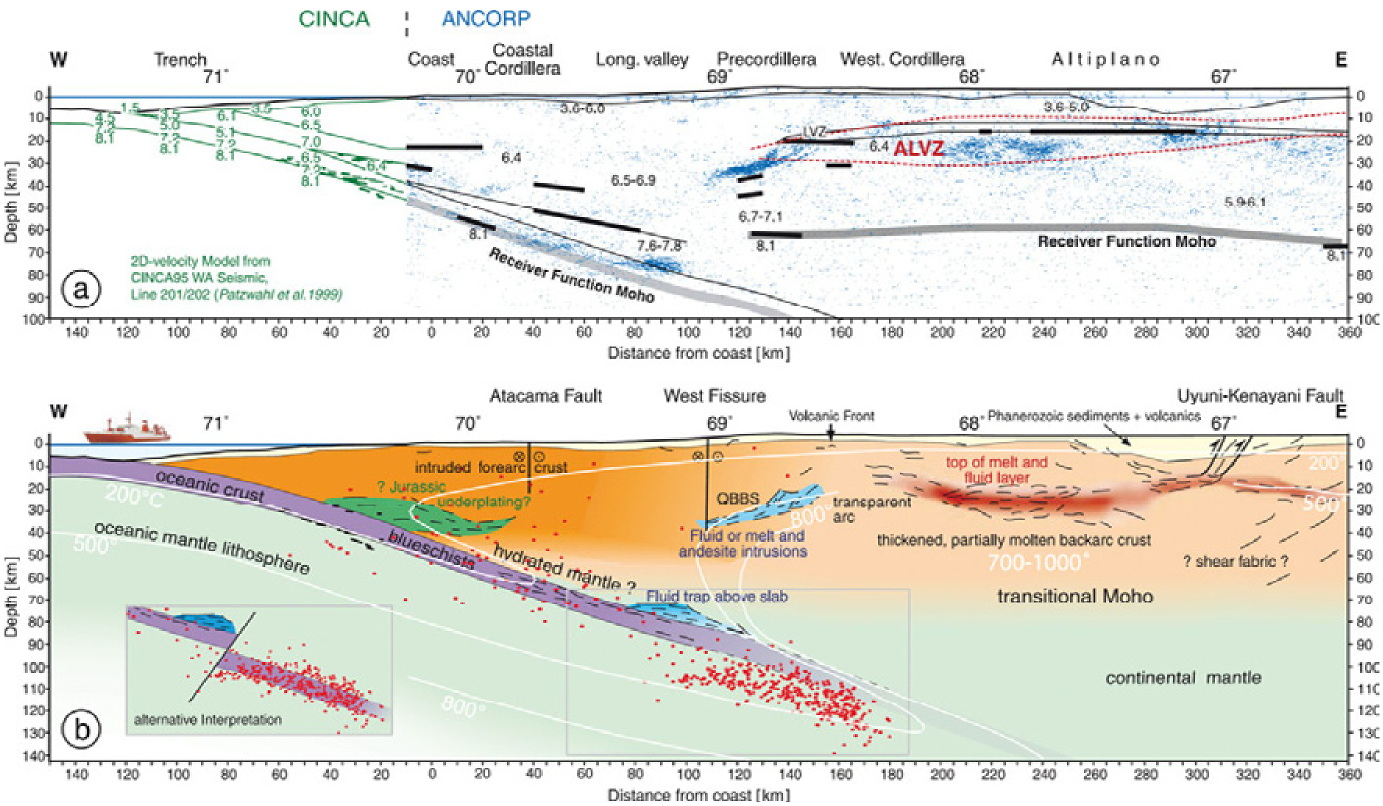



Figure 9.7.3-16. Seismic results and interpretation of CINCA'95 and ANCORP'96 along a cross section at 21°S (from ANCORP Working Group, 2003, fig. 7). (a) Automatic line drawing of depth-migrated ANCORP reflection data, including onshore wide-angle and receiver function results and merged with results from the offshore CINCA experiment and its onshore recordings. (b) Suggested geodynamic interpretation of the seismic results. [Journal of Geophysical Research, v. 108, B7, 2328, doi: 10.1029/2002JB001771. Reproduced by permission of American Geophysical Union.]

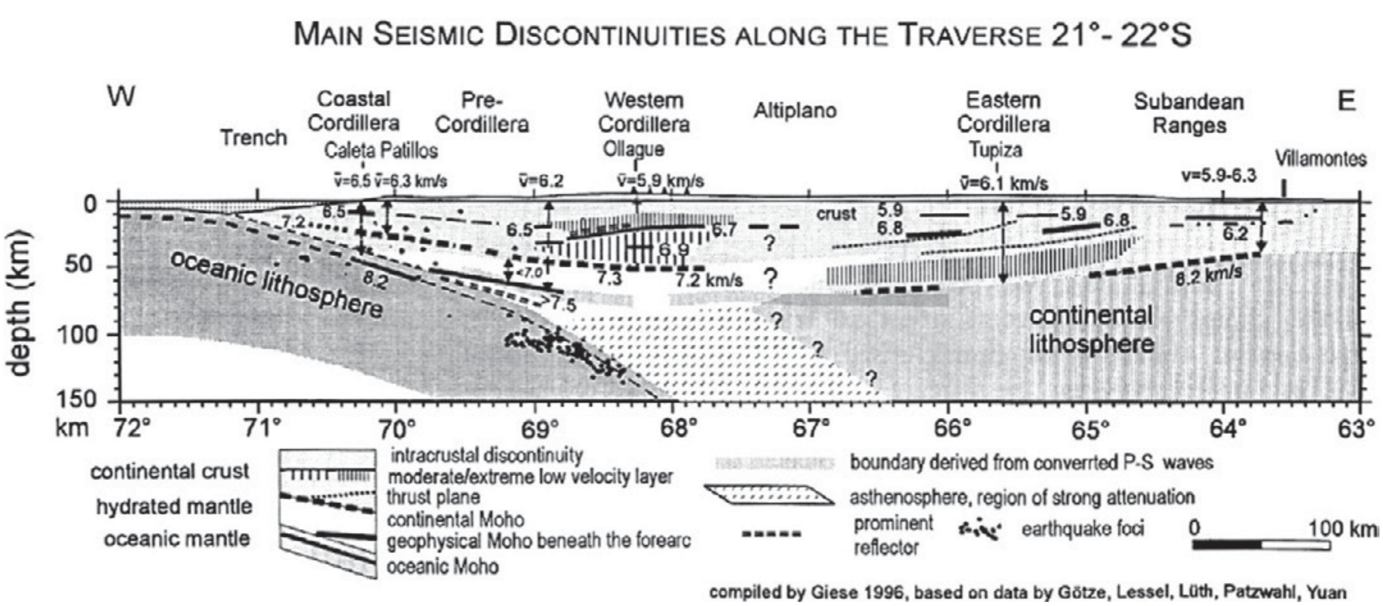



Figure 9.7.3-17. West-east cross section along 21°S through the Andean lithosphere (from Giese et al., 1999, fig. 2). The crustal structure is based on active and passive seismic measurements. [Journal of South American Earth Science, 12, p. 201–220. Copyright Elsevier.]

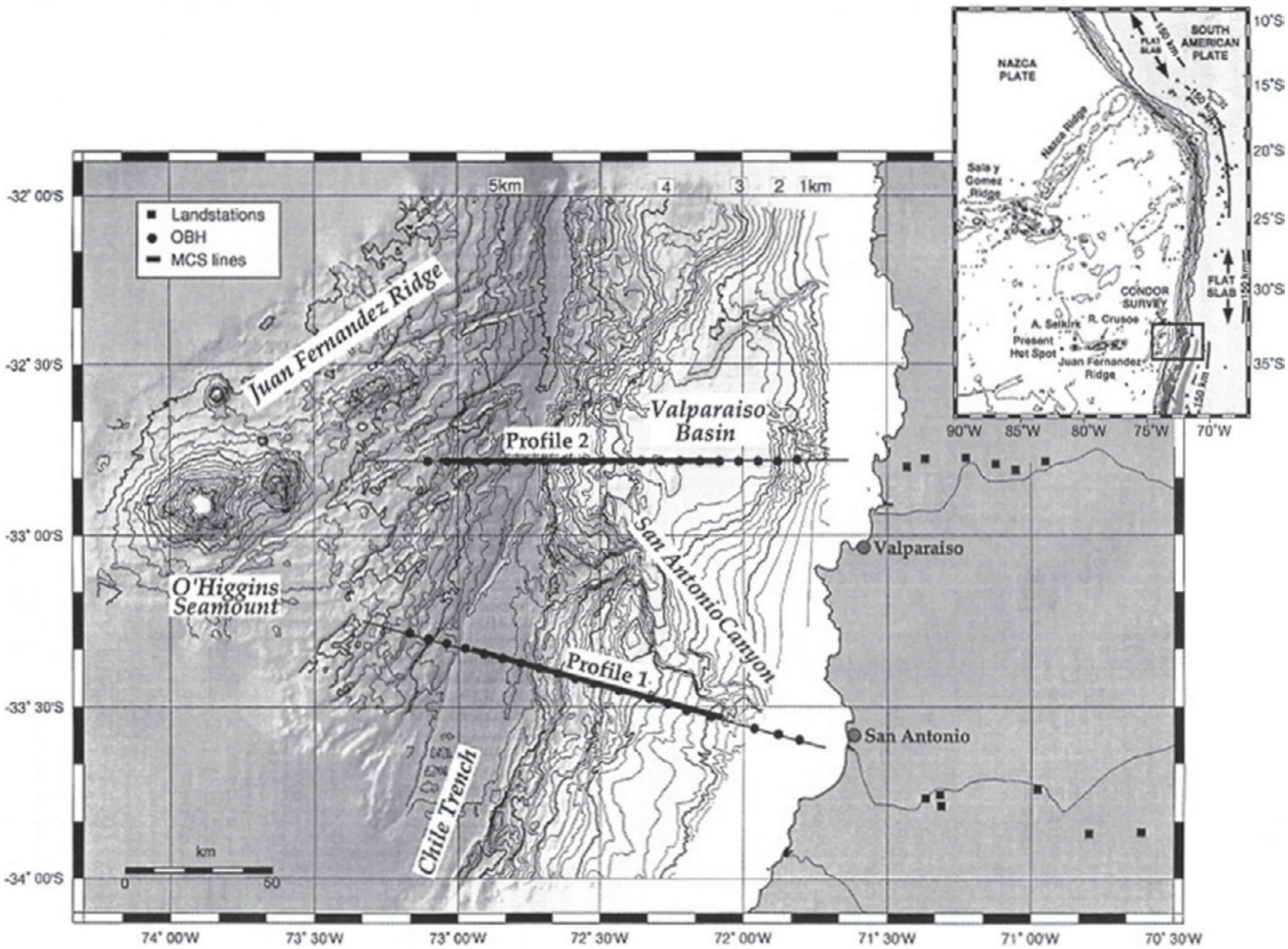



Figure 9.7.3-18. West-east offshore-onshore profiles near 33°S in the CONDOR study area in central Chile (from Flueh et al., 1998a, fig. 1). [Tectonophysics, v. 288, p. 251–263. Copyright Elsevier.]

southern Chile. The first component was the project ISSA 2000 (Integrated Seismological experiment in the Southern Andes) which consisted of a temporary seismological network and a seismic-refraction profile (Fig. 9.7.3-21). Before and after the installation of this network across the Southern Andes around 38°S, which recorded 333 local events over a period of 3 months, all 62 land stations were deployed along a profile at 39°S and five chemical explosions were recorded, one from a small lake in the Chilean Main Cordillera and four in the Pacific Ocean (Bohm et al., 2002; Lueth et al., 2003). Data examples are shown in Figure 9.7.3-22 and Appendix A9-5-1.

A typical continental Moho was not clearly observed in this region, crustal thickness was first inferred from Bouguer gravity to be ~40 km (Fig. 9.7.3-23). The oceanic Moho was observed down to ~45 km depth under the continental fore-arc.

This project was followed by the project SPOC (Subduction Processes off Chile) in 2001 (Krawczyk and the SPOC Team, 2003, 2006). It consisted of three components. The first one was a shipborne geophysical experiment with the R/V *Sonne* in fall and winter 2001/2002 and was operated by the BGR (German Geological Survey, Hannover), GEOMAR (Kiel) and the SFB 267 (Berlin and Potsdam). The second and third components were predominantly land-based onshore-offshore experiments.

The second component was a 54-km-long near-vertical incidence reflection line (NVR) which was recorded in three spreads and recorded 14 shots fired at 10 locations. Furthermore, the airgun shots along the offshore profile SO161-38/42 (Fig. 9.7.3-24) were recorded along the first 18 km. The third component was a wide-angle seismic experiment where simultaneously the airgun pulses from R/V *Sonne* were recorded by 30 three-component stations deployed in an array, and 52 stations along three consecutive W-E profiles and 22 OBH/OBS stations along the offshore part of the 38°15′S line. Besides the airgun shots, 8 chemical

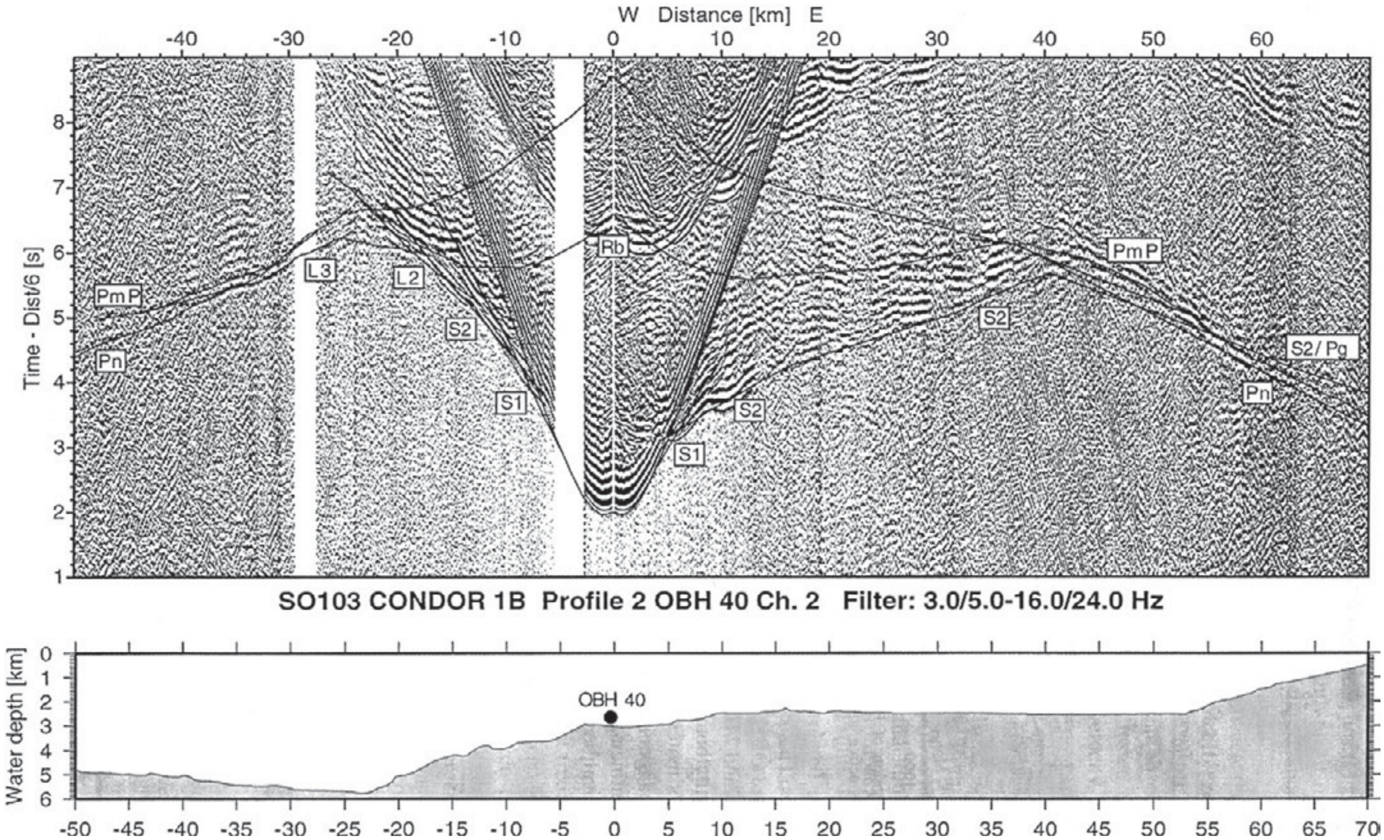



Figure 9.7.3-19. Record section from OBH 40 recorded from airgun shots along the west-east offshore-onshore profile 2 near 33°S in the CONDOR study area in central Chile (from Flueh et al., 1998a, fig. 5). [Tectonophysics, v. 288, p. 251–263. Copyright Elsevier.]

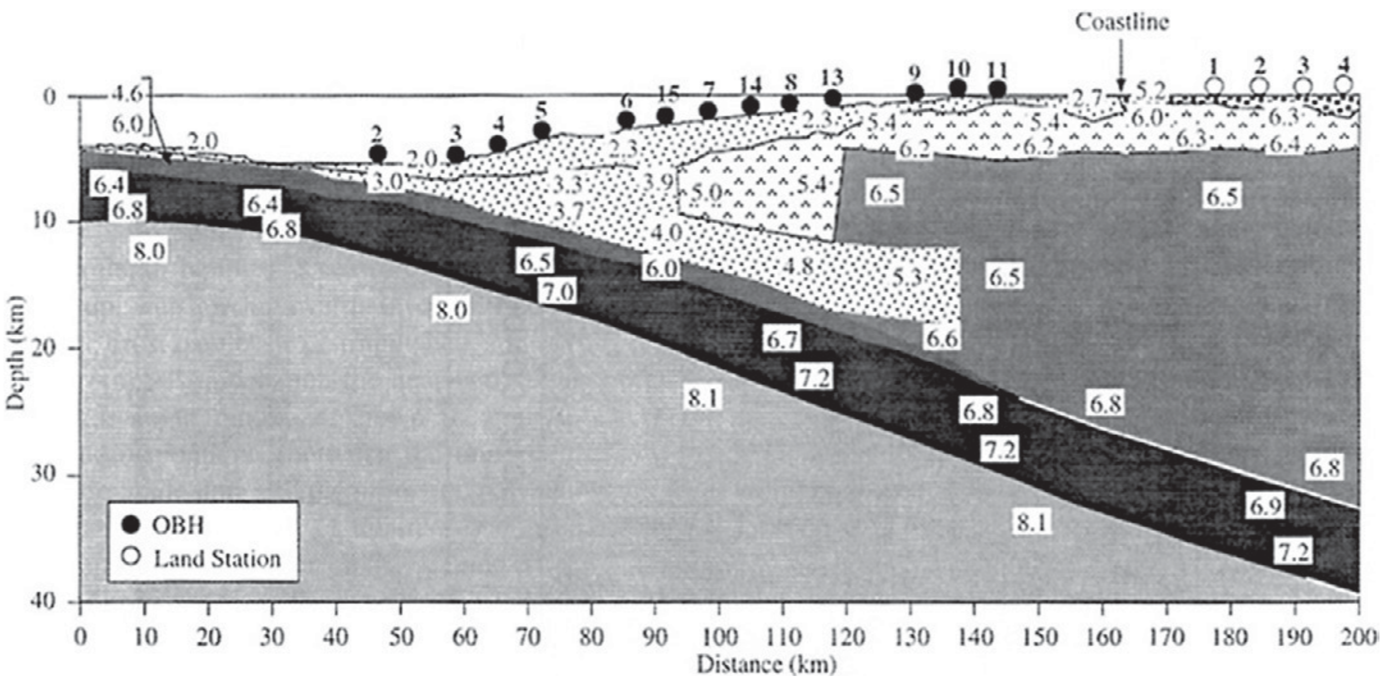



Figure 9.7.3-20. West-east offshore-onshore crustal and uppermost-mantle cross section near 33°S in the CONDOR study area in central Chile (from Flueh et al., 1998a, fig. 2, lower part). [Tectonophysics, v. 288, p. 251–263. Copyright Elsevier.]

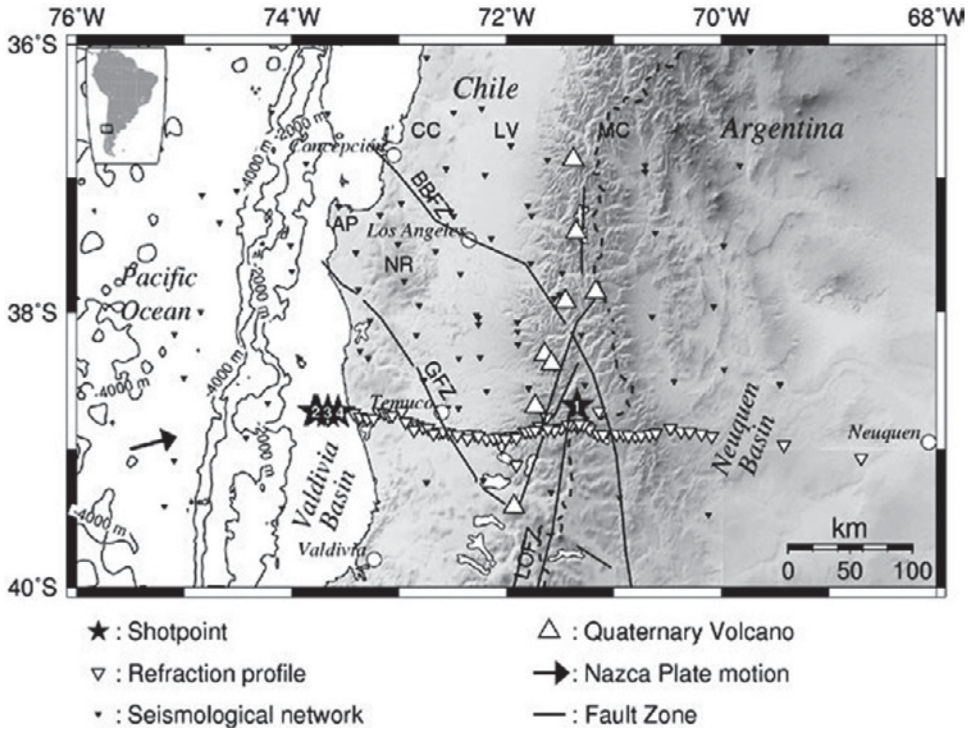



Figure 9.7.3-21. ISSA 2000 seismic measurements in southern Chile and Argentina (from Lueth et al., 2003, fig. 1). [Revista Geologica de Chile, v. 30, no. 1, p. 83–101. Reproduced with permission of Servicio Nacional de Geologia y Mineria, Chile, Andean Geology.]

shots were fired, mainly at the ends of the three profiles (Fig. 9.7.3-24 and Fig. 9.7.3-25).

The whole array also recorded teleseismic, regional and local events. Furthermore, broadband stations were deployed along the ISSA 2000 line. The most comprehensive data set of the SPOC onshore experiment came from the southern profile along 38.25°S (Sick et al., 2006). Here, a wide-angle OBS/OBH profile was acquired offshore. Simultaneously, the airgun pulses were recorded by onshore receivers.

The near-vertical reflection seismic land experiment consisted of a 54-km-long receiver set, with three 18-km-long deployments of 180 geophone groups at 100 m spacing. In total, a series of 2 offshore and 12 onshore chemical shots, fired at 10 different locations, was recorded. Thus, a 72-km-long common-depth point line resulted, with twofold coverage in the innermost 45 km and singlefold coverage at the margins, covering the offshore-onshore transition along the same transect (Krawczyk et al., 2006).

The images of the near-vertical profile provided a complex reflection pattern with several distinct reflectors, but did not reveal structures below 35 km depth. The whole data set covered the crustal and upper-mantle structure from ~200 km offshore to ~240 km onshore (Fig. 9.7.3-26). On the SPOC data, the oceanic Moho was observed down to ~40 km depth under the Coastal Cordillera, and the continental crust from the trench to the Main Cordillera. The continental crust was characterized by gradually increasing P-wave velocities from the trench to the Main Cordillera, whereby the upper crust was ~10–15 km thick and the lower crust ~25 km thick (Krawczyk et al., 2006; Sick et al., 2006). The continental Moho was imaged only along the northern SPOC profile. The combined results of ISSA 2000 and SPOC 2001 and teleseismic observations resulted in a Moho depth of 40 km beneath the Main Cordillera shallowing toward the east to ~35 km depth beneath the Neuquen basin.

The latest experiment in southern Chile was the TIPTEQ (from The Incoming Plate to Mega-Thrust EarthQuake Processes) array project of 2004 and 2005, which is described in Chapter 10.6.3.

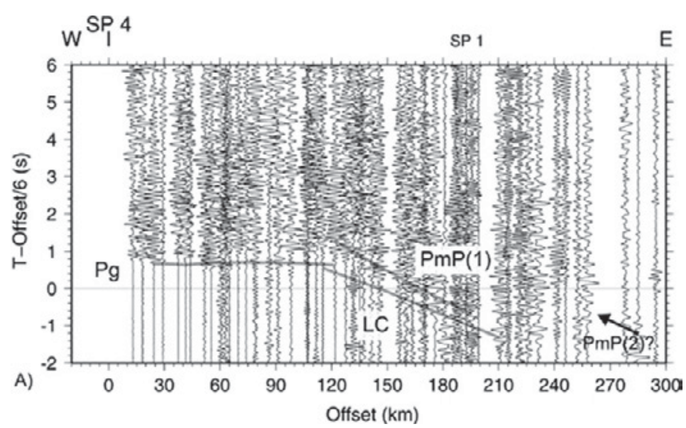



Figure 9.7.3-22. Record section of the ISSA 2000 seismic measurements in southern Chile and Argentina (from Lueth et al., 2003, appendix 1, fig. 4). [Revista Geologica de Chile, v. 30, no. 1, p. 83–101. Reproduced with permission of Servicio Nacional de Geologia y Mineria, Chile, Andean Geology.]

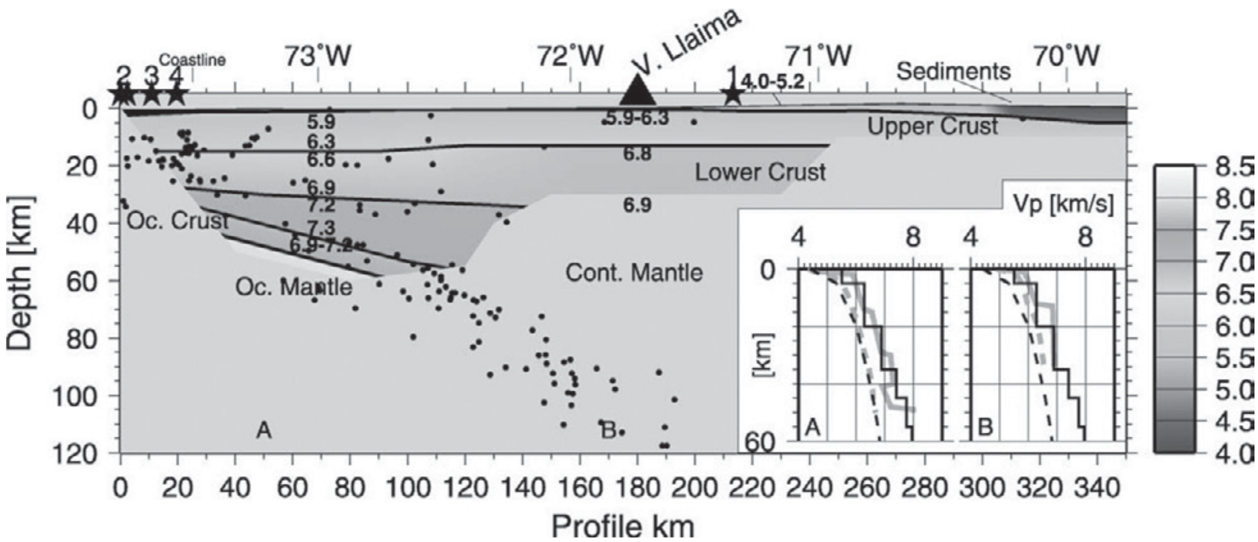



Figure 9.7.3-23. West-east onshore crustal and uppermost-mantle cross section near 39°S of the seismic-refraction line of the project ISSA 2000 in southern Chile (from Bohm et al., 2002, fig. 7). The thin black lines of the inset shows the detailed velocity-depth structure at the positions A and B of the cross section. Black dots indicate hypocenter locations of earthquakes observed by the ISSA seismological network. [Tectonophysics, v. 356, p. 275–289. Copyright Elsevier.].

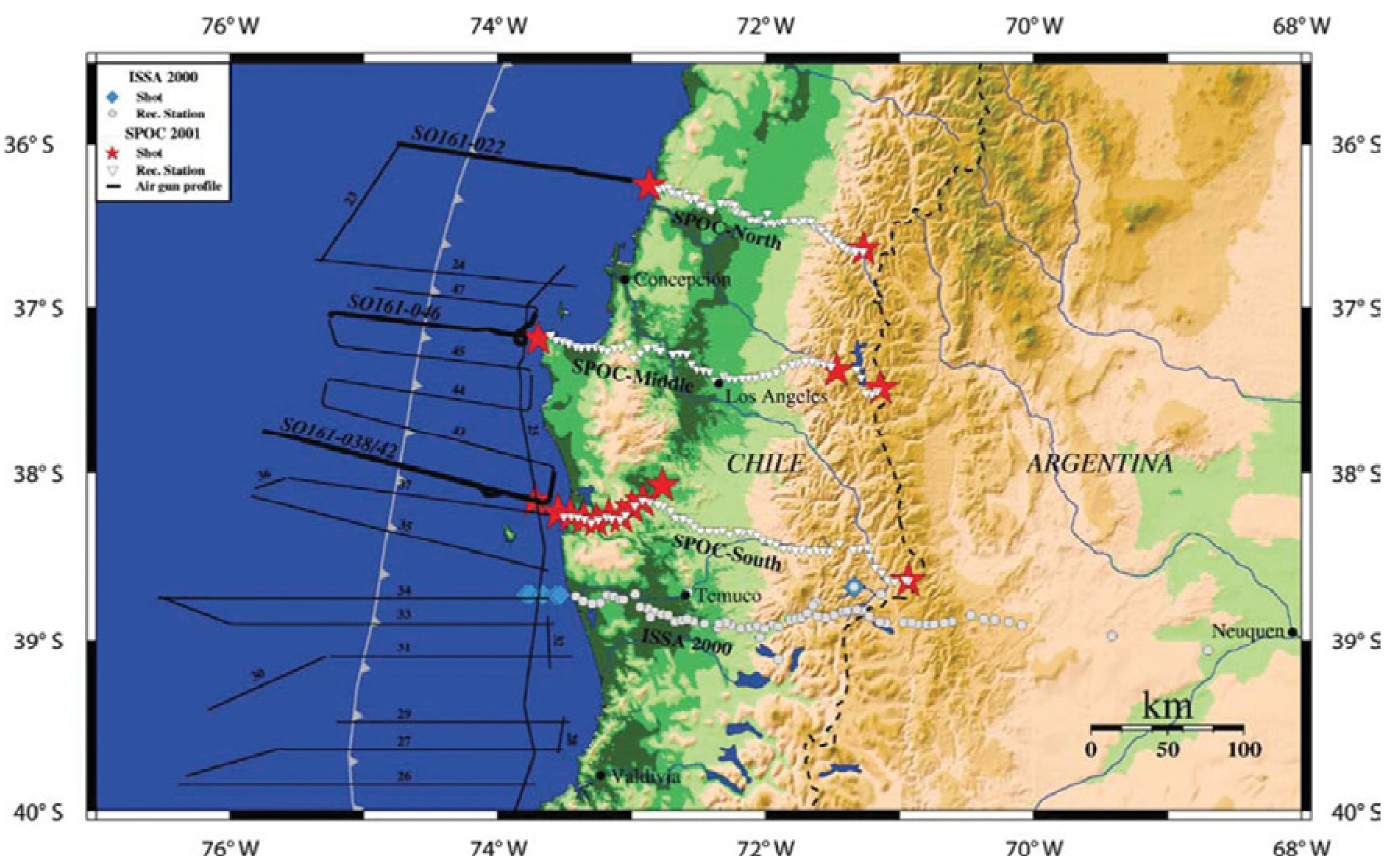



Figure 9.7.3-24. Map of SPOC 2001. White triangles and red stars—receiver locations and shotpoints of the three onshore SPOC 2001 seismic-refraction lines at latitudes 36°, 37°, and 38°S. The near-vertical reflection experiment covered the offshore-onshore transition at 38.25°S. Black lines—offshore airgun profiles. Grey circles and gray diamonds—receiver and shot locations of ISSA 2000 at 39°S. (from Sick et al., 2006, fig. 7–13). [Oncken et al., 2006, The Andes: Berlin-Heidelberg-New York, Springer, p. 147–169. Reproduced with kind permission of Springer Science+Business Media.].

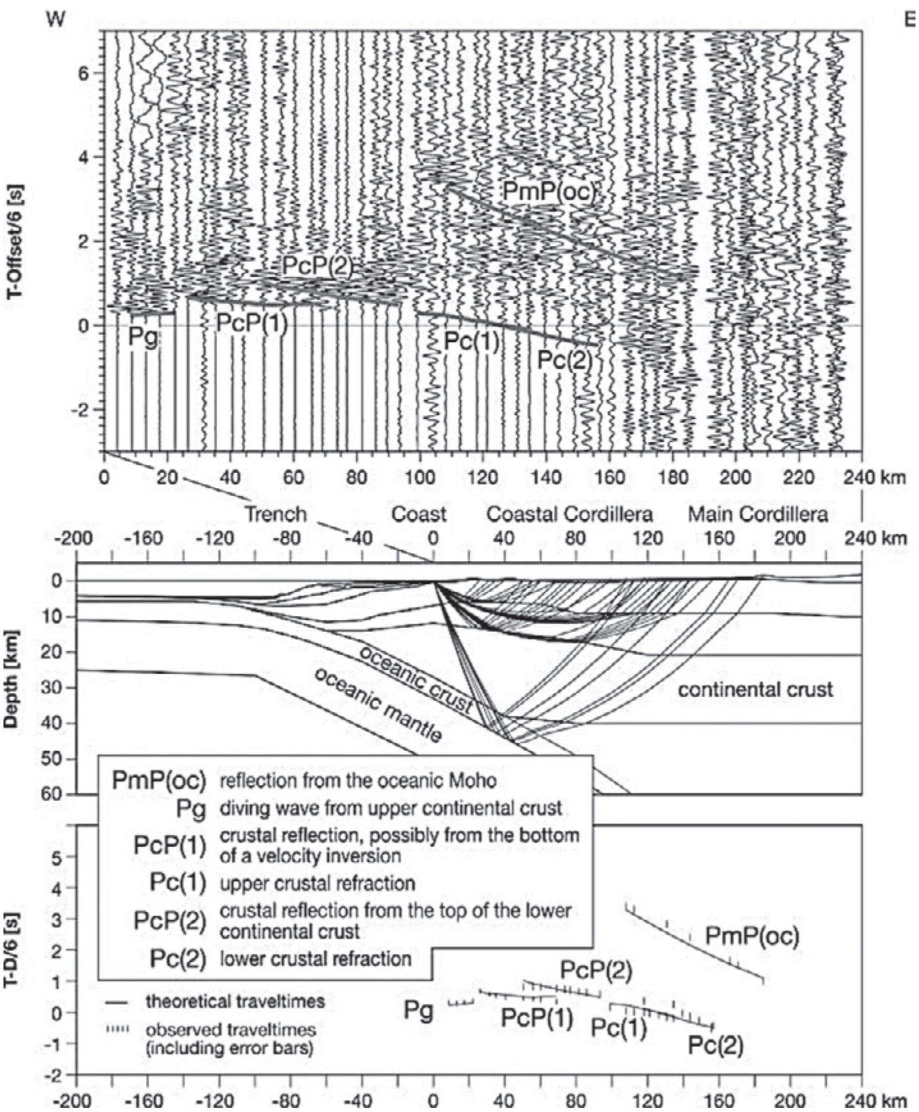



Figure 9.7.3-25. Seismic data, ray paths and traveltime curves for the western shotpoint of the south profile of the SPOC onshore survey (from Krawczyk et al., 2006, fig. 8-5). [Oncken et al., 2006, The Andes: Berlin-Heidelberg-New York, Springer, p. 171–192. Reproduced with kind permission of Springer Science+Business Media.]

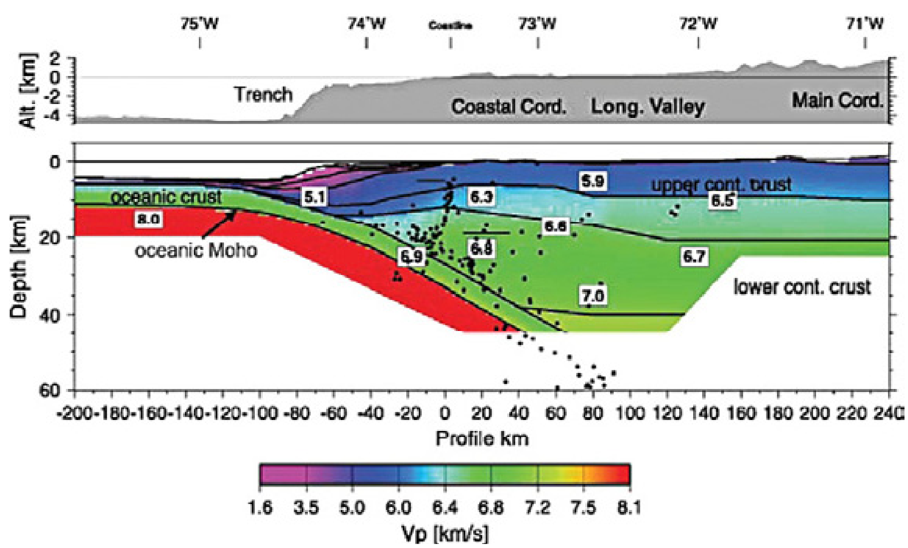



Figure 9.7.3-26. P wave model along the SPOC 2001 south profile (from Sick et al., 2006, fig. 7-15). [Oncken et al., 2006, The Andes: Berlin-Heidelberg-New York, Springer, p. 147–169. Reproduced with kind permission of Springer Science+Business Media.].

### 9.7.4. Antarctica

In 1990 and 1991, Antarctica again saw a Polish research investigation, in cooperation with Japanese scientists (Grad et al., 1997; Janik, 1997). While the previous surveys were primarily based on shot profiles and isolated land stations and had covered the whole northwestern margin of the Antarctic Peninsula, this new survey added a new profile along the axis of the Bransfield Strait (DSS-20) between the northernmost tip of the Antarctic Peninsula and the South Shetland Islands and involved OBSs (Fig. 9.7.4-01). The OBSs were supplied by Hokkaido University. They were equipped with three components and their operation depth was up to 6000 m.

In total, four expeditions were organized (Guterch et al., 1998; Janik, 1997). Some of them covered in particular the area of the Bransfield Strait between the South Shetland Islands and the Trinity Peninsula. Some of the lines were observed twice. The measurements of 1979 and 1980 were repeated in 1990 and 1991, e.g., along profiles DSS-1, -3 and -4, and other profiles were added in 1990 and 1991, when the OBSs became available. A data example from profile DSS-6 is shown in Figure 9.7.4-02 (for more data, see Appendix A9-5-2).

While the crust along the margin of the Antarctic Peninsula possesses a normal continental crustal thickness, the internal structure is quite complicated (Fig. 9.7.4-03). Only beyond the South Shetland Islands toward the northwest, the crust gradually changes into oceanic crust (Fig. 9.7.4-04). The interpretation of the whole Polish West Antarctica DSS data, obtained between 1979 and 1991, was summarized in Figure 9.7.4-04, illustrating the Moho depths along the various profiles. A recent reinterpretation was published by Janik et al. (2006; Appendix A9-5-2).

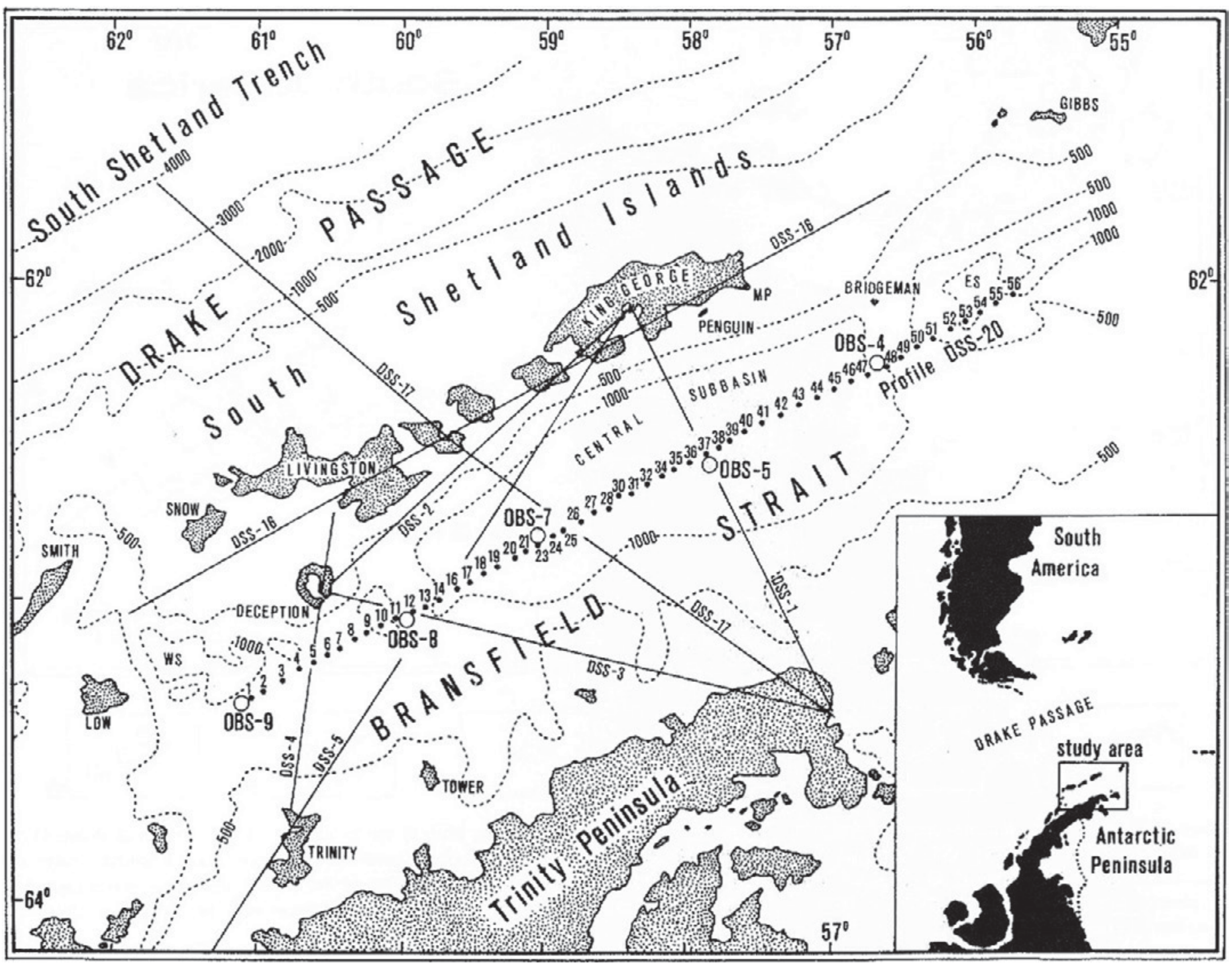



Figure 9.7.4-01. Location of deep seismic sounding lines (DSS) in the area of the Bransfield Strait, Antarctica (from Janik, 1997, fig. 11). [Polish Polar Research, v. 18, p. 171–225. Reproduced with permission of Polish Academy of Sciences, Warsaw, Poland.]

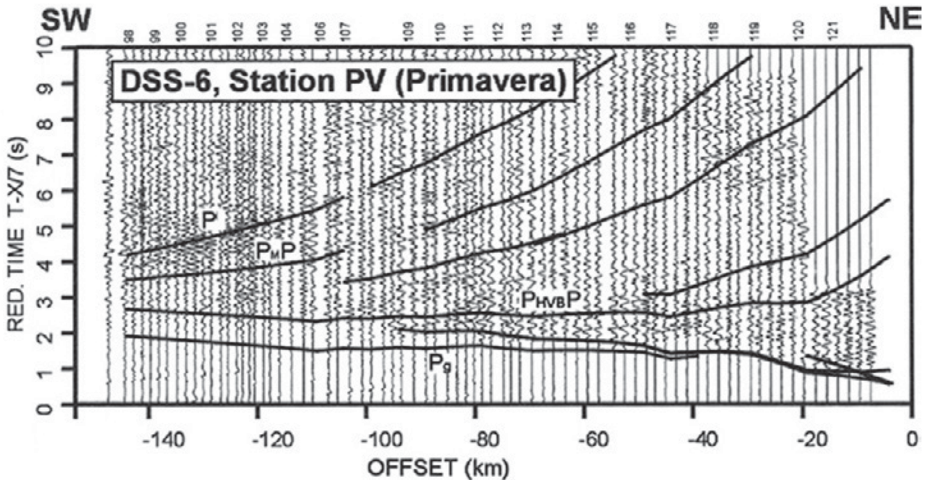



Figure 9.7.4-02. Record section of DSS-6 profile, Station PV, recorded along the Bransfield Strait, Antarctica (from Janik et al., 2006, fig. 5.2-2). [*In* Fütterer, D.K., Damaschke, D., Kleinschmidt, G., Miller, H., and Tessensohn, F., eds., Antarctica: contributions to global earth sciences: Berlin-Heidelberg-New York, Springer, p. 229–236. Reproduced with kind permission of Springer Science+Business Media.]

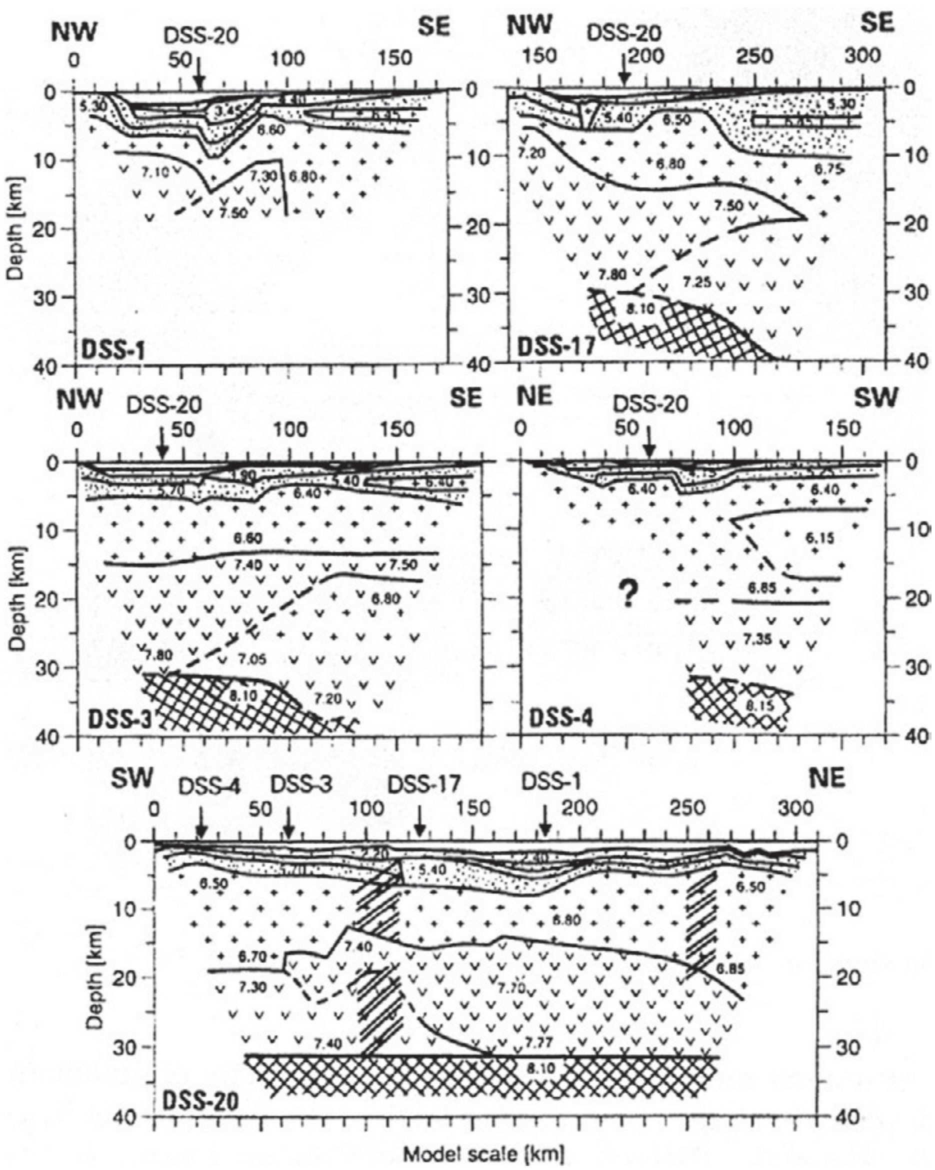



Figure 9.7.4-03. Crustal cross sections of DSS-1, -3, -4, -17 and -20 profiles through the Bransfield Strait, Antarctica (from Janik, 1997, fig. 31). [Polish Polar Research, v. 18, p. 171–225. Reproduced with permission of Polish Academy of Sciences, Warsaw, Poland.]

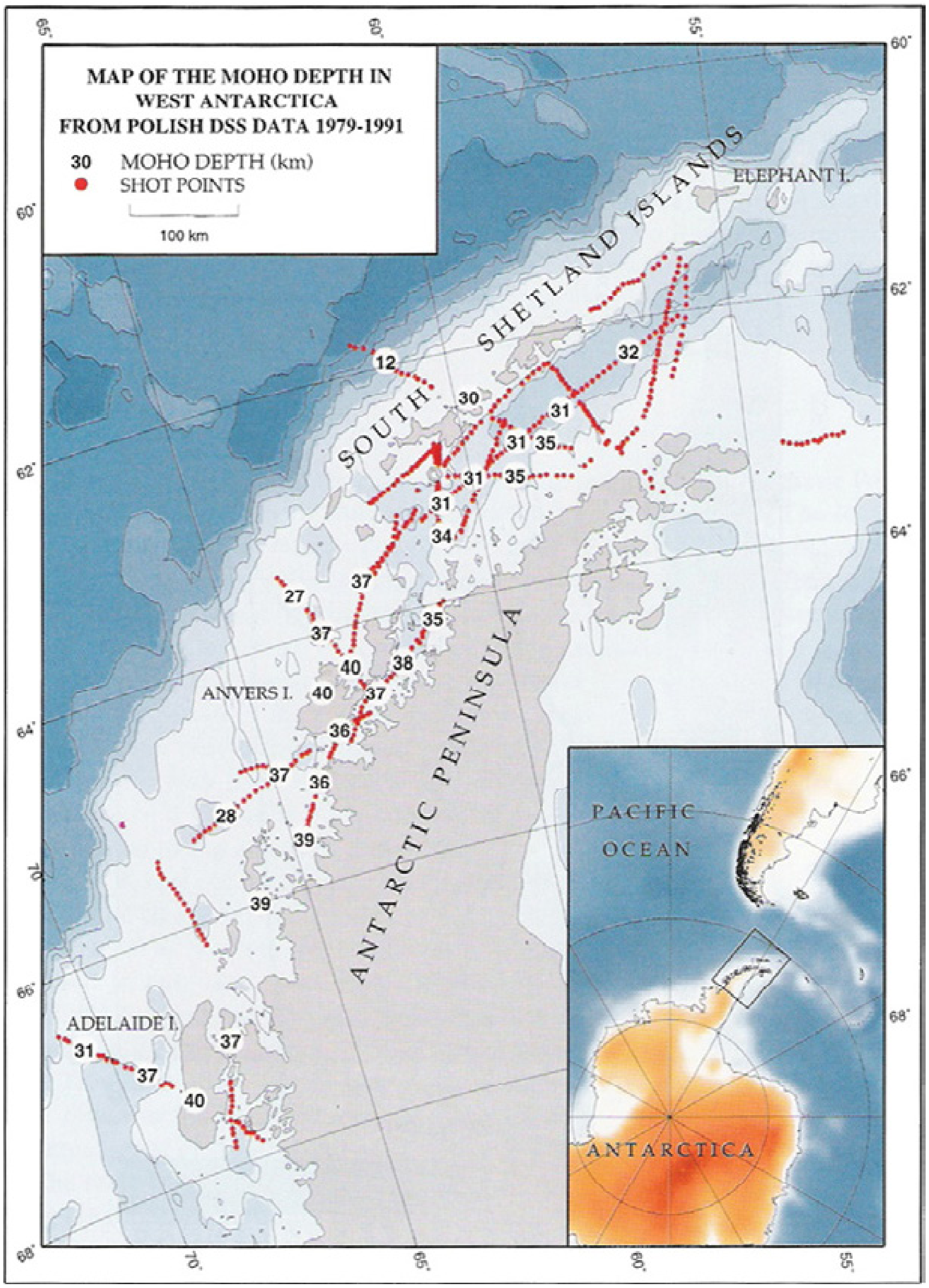



Figure 9.7.4-04. Moho depths in West-Antarctica (from Guterch et al., 1998, fig. 7). [Polish Polar Research, v. 19, p. 113–123. Reproduced with permission of Polish Academy of Sciences, Warsaw, Poland.]

SERIS (Seismic Experiment Ross Ice Shelf) was a large-scale modern multichannel seismic-reflection and -refraction experiment in the south summer of 1990–1991 which was carried out over the snow in Antarctica and aimed to examine the geometry of the boundary of the West Antarctica rift system, which includes the Ross Embayment and the East Antarctica shield as well as the causes for the uplift of the 3500-km-long and 4500-m-high Transantarctic Mountains (ten Brink et al., 1993). As access of East Antarctica from the Ross Ice Shelf, the Robb Glacier (about halfway between the McMurdo station and the South Pole) was chosen and a 141-km-long line was laid out across the Ross Ice Shelf over the Robb Glacier into the Transantarctic Mountains, from ~171°E, 82°S to 166°E, 83°S. Two types of seismic data were collected along the traverse. Multichannel seismic-reflection data were recorded with a 1.5-km-long streamer, which was a towed seismic cable to which 60 groups of gimbaled geophones were connected at 25 m intervals. Wide-angle seismic-reflection and -refraction data were recorded up to 90 km maximum distance. The energy was provided by dynamite charges, shot in 17 m deep holes. In this way, images of the subsurface layers and

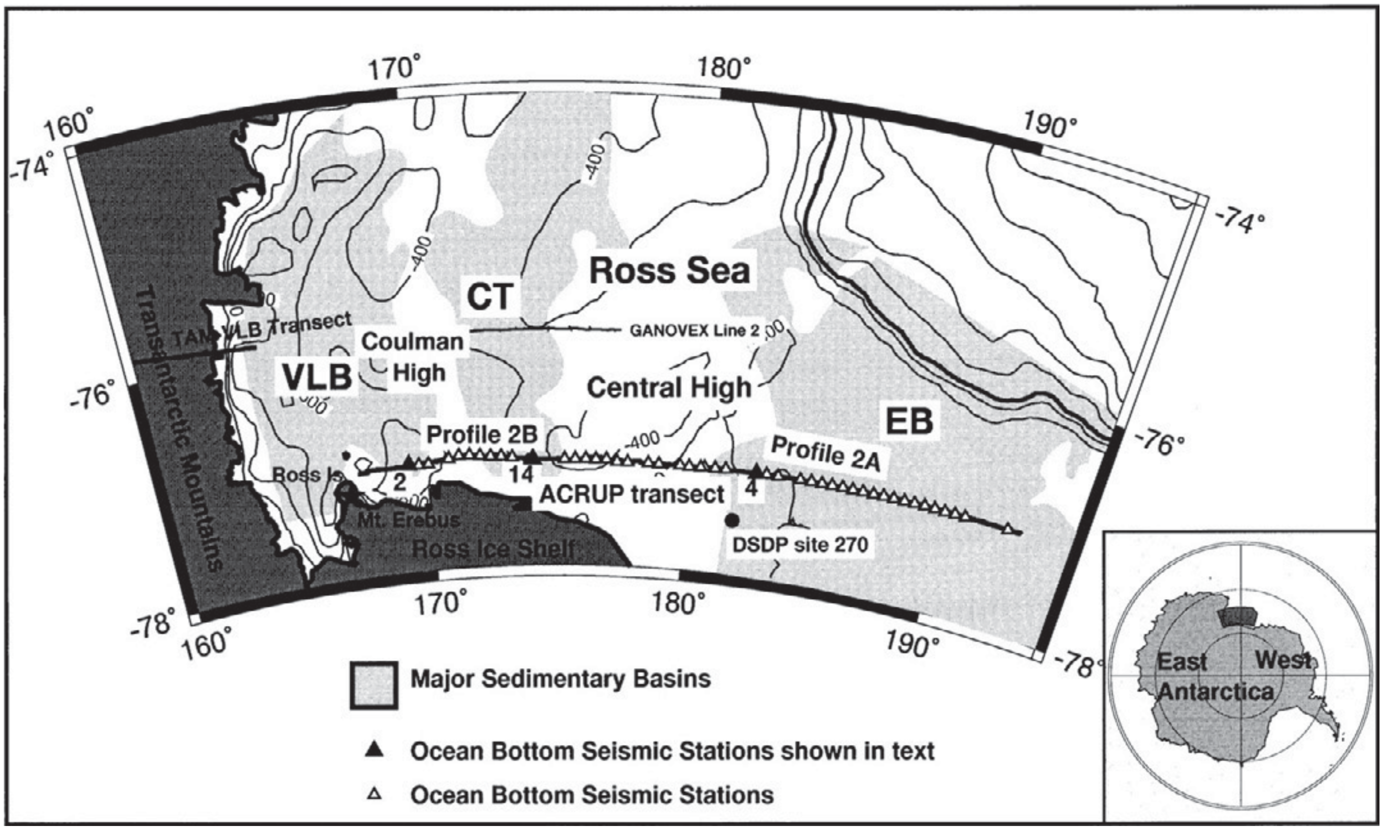



Figure 9.7.4-05. Location of the ACRUP transect in the Ross Sea, West Antarctica (from Trey et al., 1999, fig. 1). [Tectonophysics, v. 301, p. 61–74. Copyright Elsevier.].

images of the deep crust and velocities were provided. Because of logistical constraints, the wide-angle seismic part of the profile was split into a partially reversed 90-km-long profile within the Ross Embayment and a fully reversed 51-km-long part within the Transantarctic Mountains.

Underneath the ice shelf, an ~1-km-thick sedimentary layer was seen to overlie a low-grade basement or high-velocity sedimentary layer. A strong mid-crustal reflector was seen at 15 km depth. The crustal structure under the mountains was found to differ from that under the Ross Ice Shelf. Under the ice, no sedimentary layer was noticed and the upper crust appeared with a high velocity gradient. The midcrustal reflection could hardly be seen, despite recording good-quality Moho arrivals. Lower crustal velocities appeared to be typical for continental crust and lower than under the Ross Ice Shelf. The seismically determined Moho was found at similar depth under the Ross Ice Shelf as under the Ross Sea, dipping smoothly from the Ross Embayment toward the Transantarctic Mountains. Entering the mountains, where also the base of the crust dips from 15° to 20°, the crust gradually thickens from 22 to 30–35 km.

During the Australian summer of 1993–1994, the University of Hamburg, Germany, the USGS, and the Osservatorio Geofisico Sperimentale of Trieste, Italy, undertook the ACRUP (Antarctica Crustal Profile) project, during which a 670-km-long transect was recorded across the southern Ross Sea (Fig. 9.7.4-05) to study the velocity and density structure of the crust and uppermost mantle of the West Antarctic rift system. Forty-two OBSs, consisting of 34 analog three-component systems and 8 digital one-component systems (single vertical seismometer plus hydrophone), recorded strong airgun signals to distances of 120 km, which were generated by the Italian research vessel *Explora* at ~250 m intervals along the entire profile. The shots were also recorded by a 1200-m-long, 48-channel streamer to provide near-vertical incidence reflection data for mapping sedimentary sequences.

The interpretation of the data (Trey et al., 1999) revealed that the three major sedimentary basins (early-rift grabens), the Victoria Land Basin, the Central Trough and Eastern Basin, are underlain by highly extended crust and shallow mantle (minimum depth of ~16 km). Beneath the adjacent basement highs, the Coulman High and Central High, the Moho was found to deepen and to be at a depth of 21 and 24 km, respectively. For the crustal layers velocities ranging from 5.8 to 7.0 km/s were found. A distinct reflection was observed on numerous OBSs from an intracrustal boundary between the upper and lower crust at 10–12 km depth. In summary, the Moho depths and crustal thicknesses were found to vary significantly across the West Antarctic rift system (Fig. 9.7.4-06).

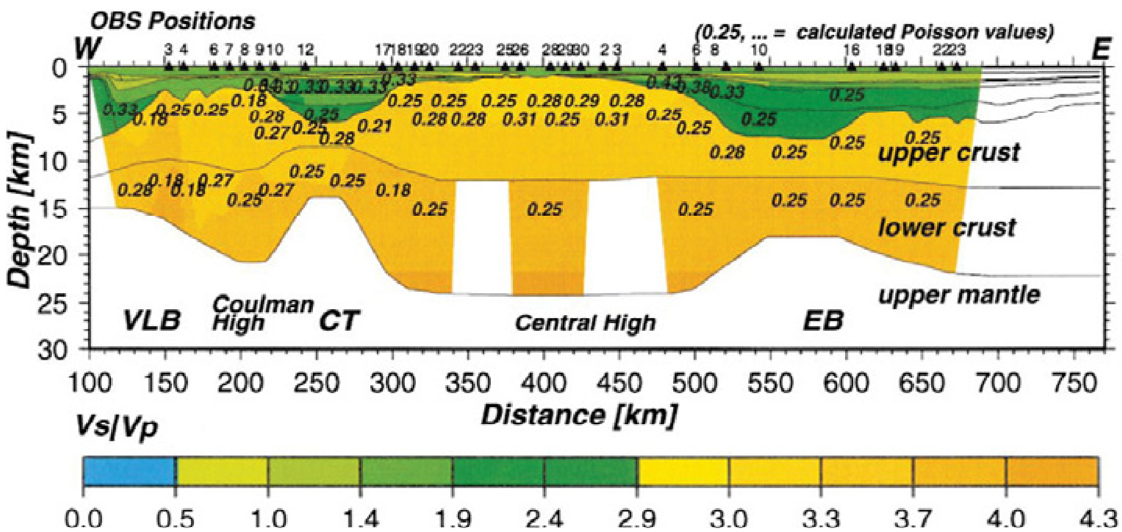



Figure 9.7.4-06. Crustal structure along the ACRUP transect in the Ross Sea, West Antarctica, showing Poisson's ratios (from Trey et al., 1999, fig. 6). [Tectonophysics, v. 301, p. 61–74. Copyright Elsevier.].

## 9.8. OCEANIC DEEP STRUCTURE RESEARCH

### 9.8.1. Introduction

Oceanic research in the 1990s and 2000s is characterized by two facts. In many projects, powerful airgun arrays replaced the conventional shooting technique with explosives and the analogue recording techniques was gradually replaced by digital equipment. The length of streamers, involving up to 300 channels, has reached several kilometers and for deep crustal research the use of large numbers of ocean-bottom instruments equipped with seismometers (OBSs) and/or hydrophones (OBHs) has become the rule. Since the introduction of digital equipment, storage of data is no longer a problem.

In particular, to concentrate the efforts of mid-oceanic ridge research worldwide, "InterRidge" was founded in 1993 to promote interdisciplinary, international studies of oceanic spreading centers through scientific exchange and the sharing of new technologies and facilities among international partners. Its members are research institutions dealing with marine research around the world. InterRidge is dedicated to sharing knowledge amongst the public, scientists, and governments and to provide a unified voice for ocean ridge researchers worldwide (www.interridge.org).

Many seaborne projects were conducted along the continental margins and were combined with land-based recording devices. Many of these projects were therefore included, as we have pointed out in Chapter 1, in the previous subchapters on onshore seismic projects, in particular those exploring the North Sea and Mediterranean Sea (e.g., many Spanish projects, all CROP surveys and surveys around Greece) and the continental margins of the British Isles, the Americas, Japan, Australia, New Zealand, and Antarctica.

The growing number of seaborne seismic projects which study crust and upper-mantle structure in oceanic areas makes it impossible, to include them all in the framework of this volume. As for the previous decades, only a limited number of examples can be presented with some detail.

### 9.8.2. Pacific Ocean

A purely marine wide-angle seismic study by three German institutions (GEOMAR, Kiel, GFZ Potsdam, and University of Hamburg) was carried out during the GEOPECO cruise of R/V *Sonne* in 2000 off the coast of Peru (Hampel et al., 2004; Krabbenhoeft et al., 2004). The aim of the survey was to study the area where the oceanic Nazca plate which originates at the East Pacific Rise is obliquely subducting underneath the continental South America plate (Figs. 9.8.2-01 and 9.8.2-02) along the Peru-Chile trench with water depths of more than 6000 m. The Nazca plate is segmented by fracture zones, e.g., the Viru Fracture Zone or the Mendana Fracture Zone (VFR and MFR in Fig. 9.8.2-01), leading to different crustal ages along the plate boundary.

The survey comprised six seismic wide-angle profiles (Figs. 9.8.2-01 and 9.8.2-02) with shots fired by an airgun array with intervals of 60 seconds and a ship speed of 4 knots, resulting in an average shot spacing of 120 m. The seismic-reflection and -refraction data were recorded by a total of 97 stations using the OBSs and OBHs of GEOMAR, Kiel, Germany (Krabbenhoeft et al., 2004). Simultaneously, multichannel seismic data were recorded with 150 m shotpoint distances by a 24-channel streamer, towed 50 m behind the research vessel.

As a result, the oceanic crust of the Nazca plate, divided into an upper and a lower crustal layer, was determined to be 6.4 km thick on average, with a pelagic sediment cover ranging from 0 to 200 m. In the area of the Trujillo Trough (TT in Fig. 9.8.2-01),

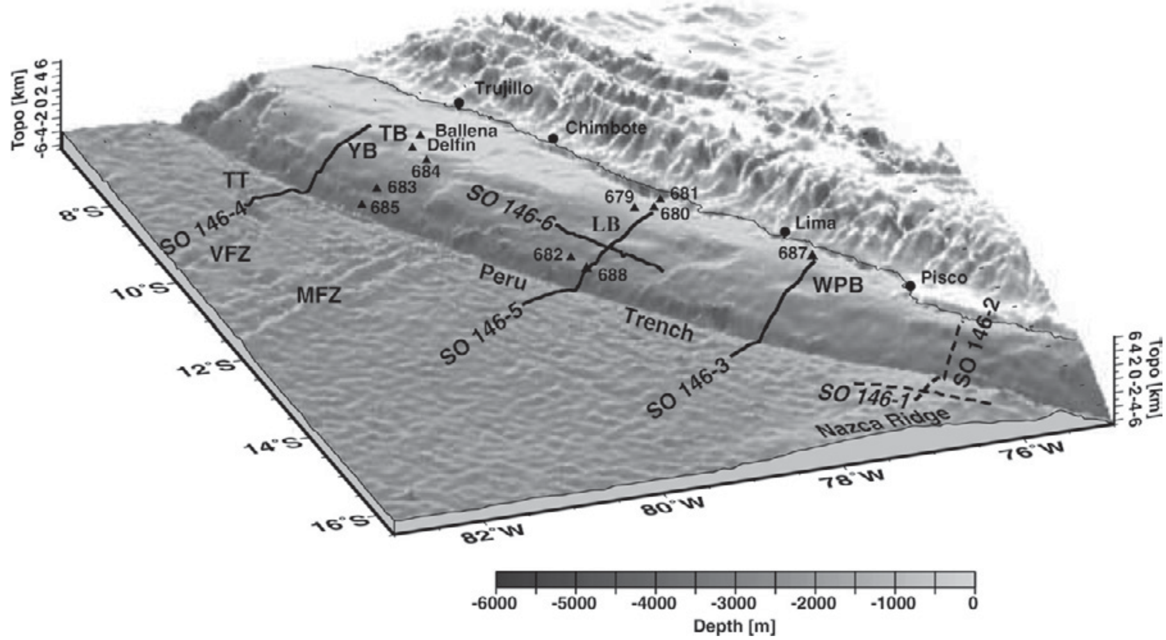



Figure 9.8.2-01. Location of the GEOPECO seismic survey off Peru, illuminated from the northeast (from Krabbenhoeft et al., 2004, fig. 2). [Geophysical Journal International, v. 159, p. 749–764. Copyright John Wiley & Sons Ltd.]

at 8°S, the crust thinned to 4 km. The underlying mantle had an average velocity of 7.9 km/s. Across the margin, the plate boundary, where the South American plate overrides the Nazca plate, could be traced down to 25 km depth (Fig. 9.8.2-03). Adjacent to the trench axis, a frontal prism, associated with the steep lower slope, was inferred from the velocity models (Krabbenhoeft et al., 2004). The data farther south, of profiles 146-1 and 146-2, were interpreted by Hampel et al. (2004).

Another project dealt with the seismic structure of the Galapagos volcanic province, located between 78°W and 95°W longitude and 3°S and 9°N latitude (Fig. 9.8.2-04). Three seismic experiments were carried out in this area, PAGANINI in 1999, G-PRIME in 2000, and SALIERI in 2001. Other seismic projects in this area had investigated the Cocos-Nazca Spreading Center (CNCS in Fig. 9.8.2-04) and the Galapagos Hot Spot (GHS in Fig. 9.8.2-04), described, e.g., by Canales et al. (2002) and Toomey et al. (2001), respectively.

The PAGANINI (Panama Basin and Galapagos Plume–New Investigations of Intraplate Magmatism) seismic project dealt in particular with detailed investigations of the Cocos and Malpelo volcanic ridges. The thickened oceanic crust found here was thought to be the result of an interaction between the Galapagos hot spot and the Cocos-Nazca Spreading Center during the last 20 m.y. (Sallares et al., 2003). To obtain accurate 2-D velocity models, three wide-angle seismic profiles were acquired across the two ridges (nos. 1–3 in Fig. 9.8.2-04). The airgun shots along the three lines had a shot spacing of ~140 m and were recorded by OBS and OBH over a length of 275 km, 260 km, and 245 km, respectively, with a receiver spacing ranging from 1 to 5 km. In total, 21 German OBSs and OBHs and 13 French OBSs were available. Clear arrivals could be recorded on all profiles up to 150 km offsets, but $P_n$ was only identified on profile 3. The main result (Sallares et al., 2003) was that the maximum thickness of the crust ranged between 16.5 km (southern Cocos) and 18.5–19.0 km (northern Cocos and Malpelo). While the upper layer of the crust (layer 2) was characterized by high-velocity gradients, the lower layer appeared in having nearly uniform velocities (layer 3).

The SALIERI (South American Lithosphere transects across volcanic Ridges) seismic experiment followed in 2001 (Fig. 9.8.2-04). The seismic data set obtained and used by Sallares et al. (2005) included 52 seismic sections recorded along two wide-angle profiles acquired by the German research vessel *Sonne*. Airgun shots were spaced 120 m apart, and 24 GEOMAR OBSs and OBHs and 13 French OBSs were available. The first profile 1 included 30 OBSs and OBHs with a receiver spacing between 7 and 15 km, which were deployed along a 350 km N-S–trending transect, which crossed the Carnegie Ridge at 85°W. Profile 2 comprised 22 instruments deployed along a 230 km N-S transect with a similar receiver spacing covering the northern flank of the ridge near the subduction zone at 82°W. From the interpretation (Sallares et al., 2005), a maximum crustal thickness of 19 km was obtained for the central Carnegie-1 profile, located over a 19–20-m.y.-old oceanic crust at about 85°W, while for the eastern Carnegie-2 profile, located over a 11–12-m.y.-old oceanic crust at about 82°W, a crustal thickness of only 13 km resulted. In comparison with the profiles of the PAGANINI-1999 project, the thickness of the oceanic layer 2 proved to be quite uniform along

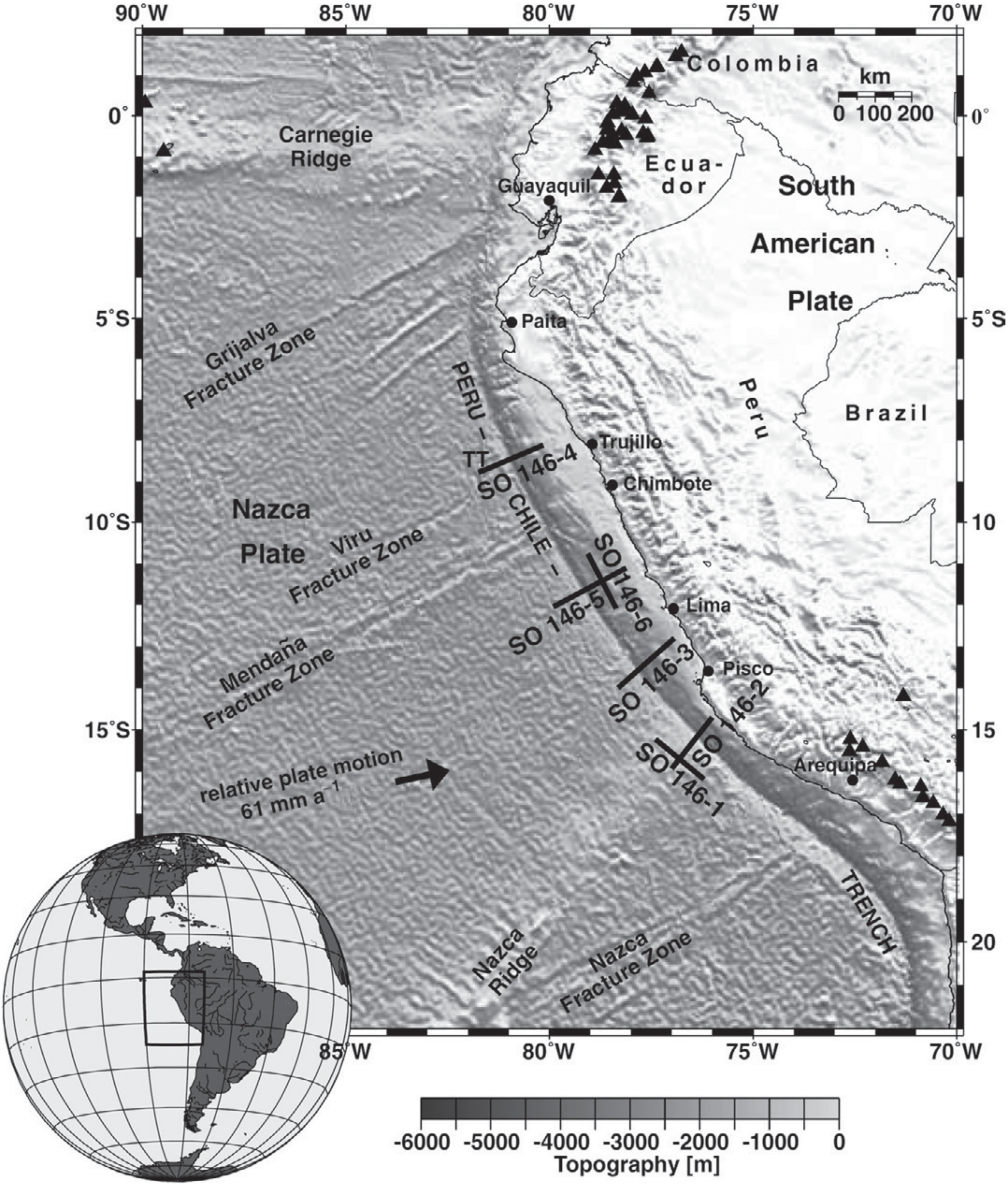



Figure 9.8.2-02. Location of the GEOPECO study area off Peru (from Krabbenhoeft et al., 2004, fig. 1). [Geophysical Journal International, v. 159, p. 749–764. Copyright John Wiley & Sons Ltd.]

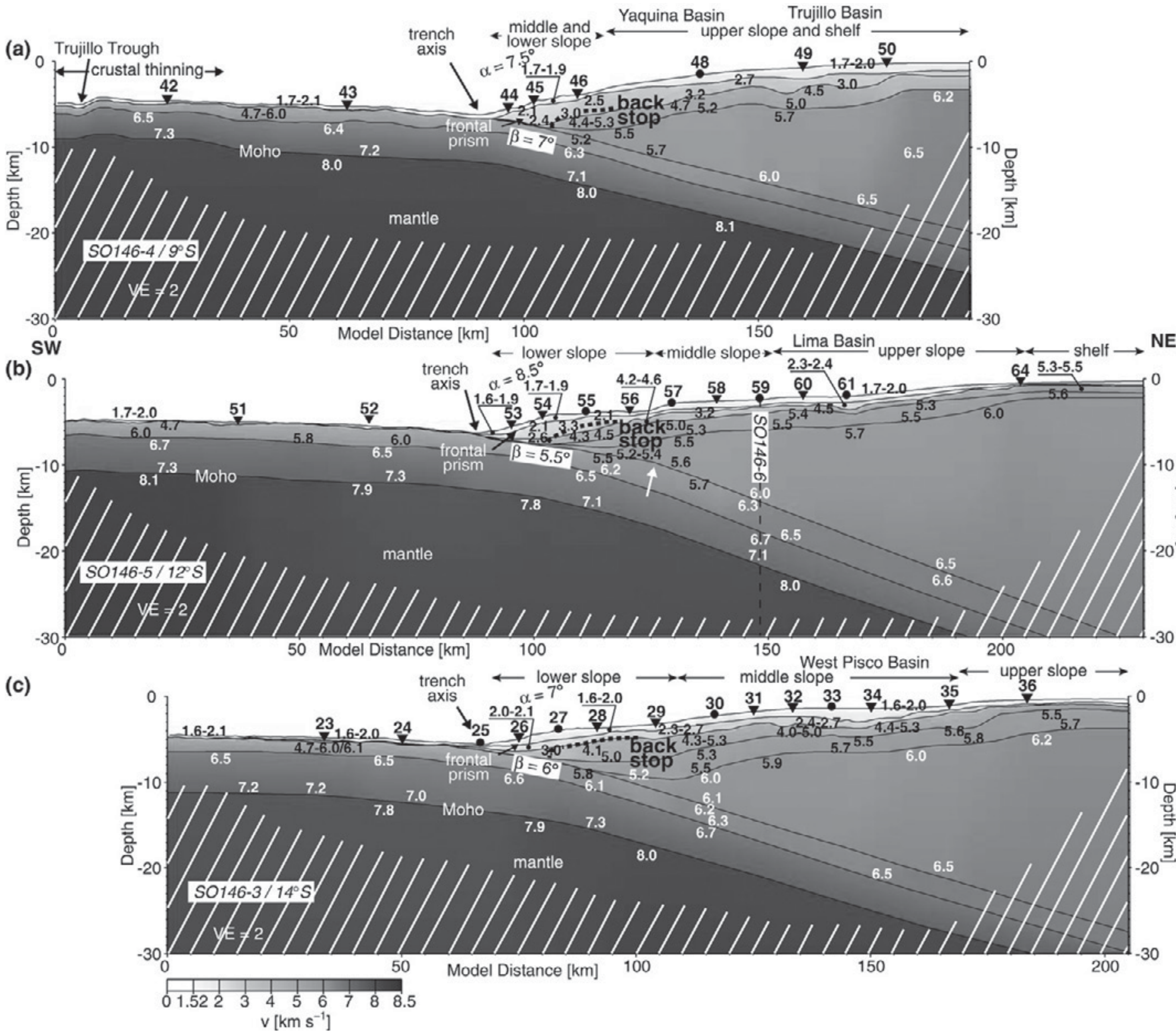



Figure 9.8.2-03. Models from N to S, obtained from the three seismic wide-angle profiles SO 160-4, -5, and -3, shown by thick black lines in Figure 9.8.2-01 (from Krabbenhoeft et al., 2004, fig. 10). [Geophysical Journal International, v. 159, p. 749–764. Copyright John Wiley & Sons Ltd.]

the five profiles, while crustal thickening was mainly accommodated by layer 3, the velocities of which appeared systematically lower where the crust thickened.

Farther west in the equatorial Pacific, at the western end of the Cocos-Nazca axial ridge and east of the Galapagos microplate (Fig. 9.8.2-05), the Hess Deep rift valley structure around 101.5°W, 2.1°N was investigated by a seismic-refraction experiment conducted in 1993 by Scripps Institution of Oceanography at San Diego, California (Wiggins et al., 1996). This experiment did not use energy sources near the sea surface, but seafloor shots of 30 kg charge size whose energy was coupled to the seafloor in a relatively small area. The shots were recorded by a linear array of digital OBSs which contained three-component seismometers and a hydrophone (Fig. 9.8.2-06). The resulting velocity model differed from typical young fast-spreading East Pacific Rise crust by ~1 km in depth with slow velocities beneath the valley of the Hess Deep and a fast region forming the intrarift ridge.

In 1997, a 3-D MCS reflection and tomography experiment was carried out over the 9°03′N overlapping spreading center of the East Pacific Rise (Singh et al., 2006a). It has an offset of 8 km and the two limbs overlap for ~27 km. The data were acquired over a 20 km by 20 km area of seafloor covering both limbs of the overlapping spreading center. Over a distance of less than 7 km along the ridge crest, a rapid increase in two-way traveltime of

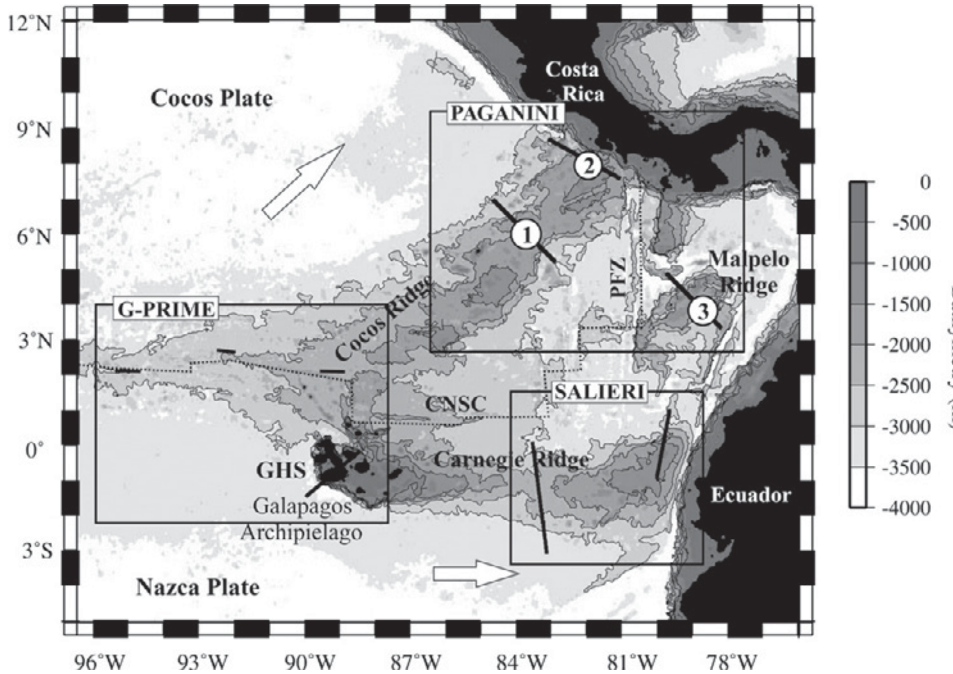



Figure 9.8.2-04. Location of the PAGANINI-1999, G-PRIME-2000, and SALIERI-2001 seismic projects in the Galapagos volcanic area (from Sallares et al., 2003, fig. 1). [Journal of Geophysical Research, v. 108, B12, 2564, doi: 10.1029/2003JB002431. Reproduced by permission of American Geophysical Union.]

seismic waves between the magma chamber and Moho reflections was observed

The Garrett Fracture Zone, located along a flow line on the eastern flank of the East Pacific Rise at 14°S, was the goal of the wide-angle reflection/refraction survey EXCO (Exchange between Crust and Ocean) of the Universities of Hamburg and Bremen, Germany, at the end of 1995 (Fig. 9.8.2-07). Six profiles were shot on 0.5–8.3-m.y.-old seafloor, using 17 OBHs, digital recording systems of GEOMAR, Kiel, and a single airgun source, towed at 10 m depth behind the ship. Shot spacing was 180 m and 250 m on the 8.3 my-line, and the maximum recording distances were 70 km (Grevemeyer et al., 1998). The result was a classical model of oceanic crust of ~6 km thickness. It comprised a 1.4–1.8-km-thick layer 2, with a low-velocity (2.9–4.6 km/s) surficial layer 2A at its top and bounded at its base by a sharp velocity increase to 5.5 km/s at the top of layer 2B, and a lower crustal layer 3 with velocities ranging from 6.8 to 7.2 km/s with a moderate velocity gradient.

The nearby East Pacific Rise, south of the Garrett Fracture Zone, had been investigated earlier, in 1991, by a U.S. team from Woods Hole Oceanographic Institution in Massachusetts, Scripps Institution of Oceanography at San Diego, California, and Lamont-Doherty Geological Observatory at Palisades, New York (Detrick et al., 1993a; Hussenoeder et al., 1996; Hooft et al., 1997). The seismic survey TERA was concentrated in three areas centered at 14°15′S, 15°55′S and 17°20′S (Fig. 9.8.2-08).

Three types of seismic data were obtained: common midpoint (CMP) reflection data, two-ship expanding spread profiles and ocean-bottom seismic refraction data. In each of the three areas, six OBSs were deployed. The reflection data were acquired by a 20-gun airgun array and a 4-km-long, 160-channel digital streamer. In the 14°15′S and 17°20′S areas, 23 expanding profiles, oriented parallel to the rise axis, were shot. A series of reconnaissance reflection profiles was also obtained along and across the rise axis over a distance of more than 800 km from ~20.7°S to the Garrett transform (Fig. 9.8.2-08). One of the main results of the interpretation of the seismic data from the ultrafast-spreading southern part of the East Pacific Rise was that the rise was underlain by a thin extrusive volcanic layer (seismic layer 2A of less than 200 m thickness) that thickened rapidly off the ridge axis. A relatively continuous reflection at 5.5–6.0 s TWT and at 1.8 s below the sea floor, which could be traced within 2–3 km off the rise axis on some profiles, was interpreted as Moho. A more or less continuous reflection from the axial magma chamber (Fig. 9.8.2-09) could be traced along the entire reconnaissance survey, resulting in an average depth of 2.6 km below the seafloor along the rise axis between 14°S and 18°S, but dipping farther south to nearly 3 km depth at the end of the survey line (Detrick et al., 1993a). Parts of the data were later further interpreted in more detail (Hussenoeder et al., 1996; Hooft et al., 1997; Singh et al., 1998b).

A more detailed description of the experiment in the 14°15′S area and its interpretation was published by Kent et al. (1994). They stated that, where the Moho could be imaged, it lies at a nearly constant traveltime below the base of layer 2A which suggested that layers 2B and 3 were, to first order, of uniform thickness. Aside from the Moho reflections at an approximate traveltime of 1.4 s below the base of layer 2A, the lower crust appeared acoustically transparent indicative of a seismically homogeneous layer 3.

The East Pacific Rise at 9°–10°N which had been the focus of an extended multi-channel experiment in 1985 (Detrick et al.,

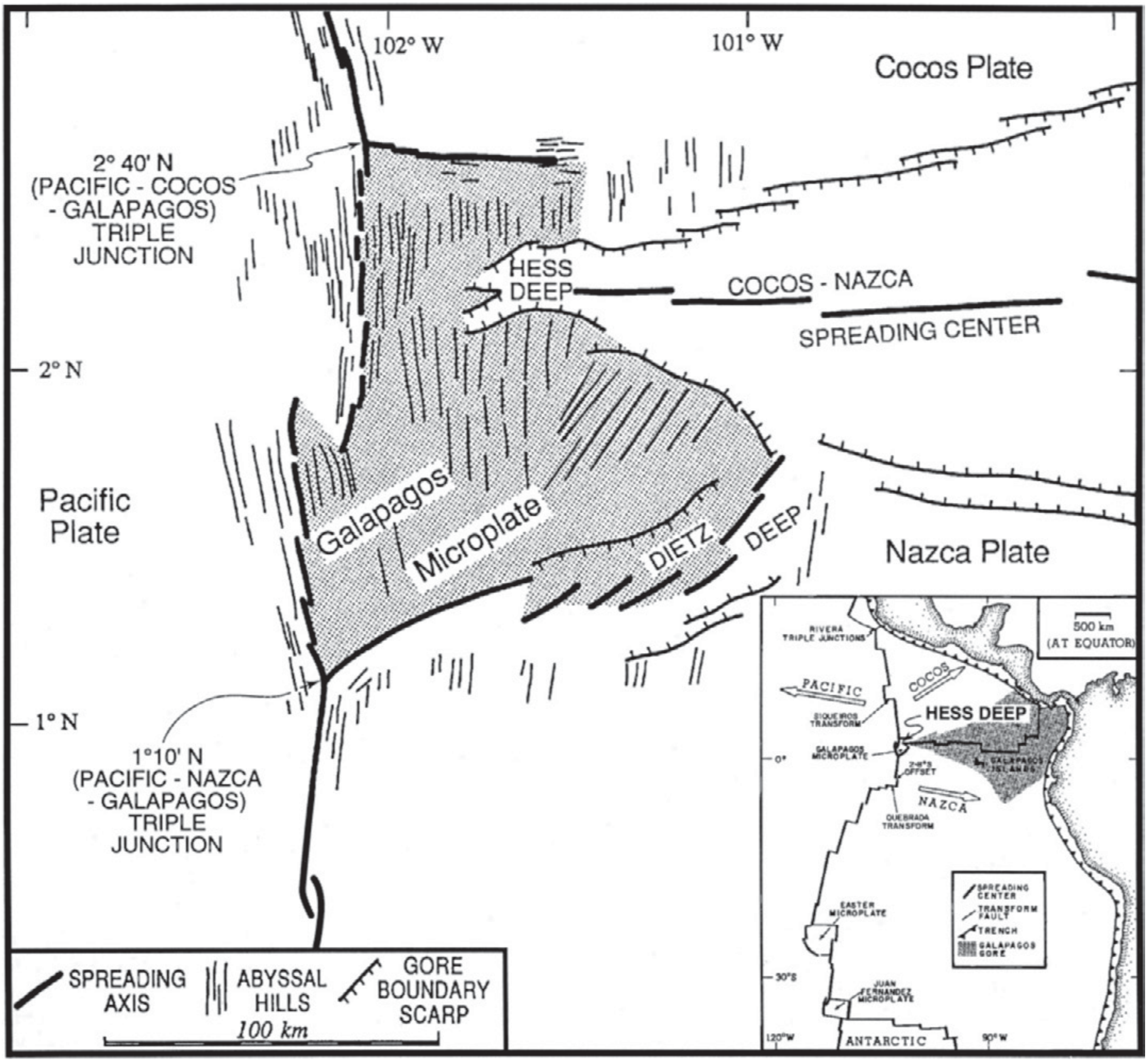



Figure 9.8.2-05. Location of the Hess Deep rift valley and its relation to the East Pacific Rise (from Wiggins et al., 1996, fig. 1). [Journal of Geophysical Research, v. 101, p. 22,335–22,353. Reproduced by permission of American Geophysical Union.]

1987), was again the focus of a large-scale experiment in 1993 (Christeson et al., 1997), this time using 59 OBSs deployments along six lines shot with a large airgun shot spacing of 250 m. The large number of OBSs allowed the study of regional compressional (P) and shear (S) velocity structure of seismic layer 2B and layer 3.

The MELT (Mantle Electromagnetic and Tomography) experiment in 1996 targeted the East Pacific Rise between 15° and 18°S and was designed to distinguish between competing models of magma generation beneath mid-ocean ridges (MELT Seismic Team, 1998). The project comprised seismic and electromagnetic observations. Besides the 800-km-long two seismic-passive linear arrays of 51 OBSs, an active-source component was also involved to study the crustal structure in detail (Canales et al., 1998). Airgun shots, fired at a 100-s repetition rate, were recorded by 15 OBSs deployed along three seismic-refraction lines. The northern line along the second MELT array was 360 km long and crossed the East Pacific Rise a few kilometers north of the 15°55′S overlapping spreading center and included five OBSs spaced 40–90 km apart. The remaining ten OBSs were deployed

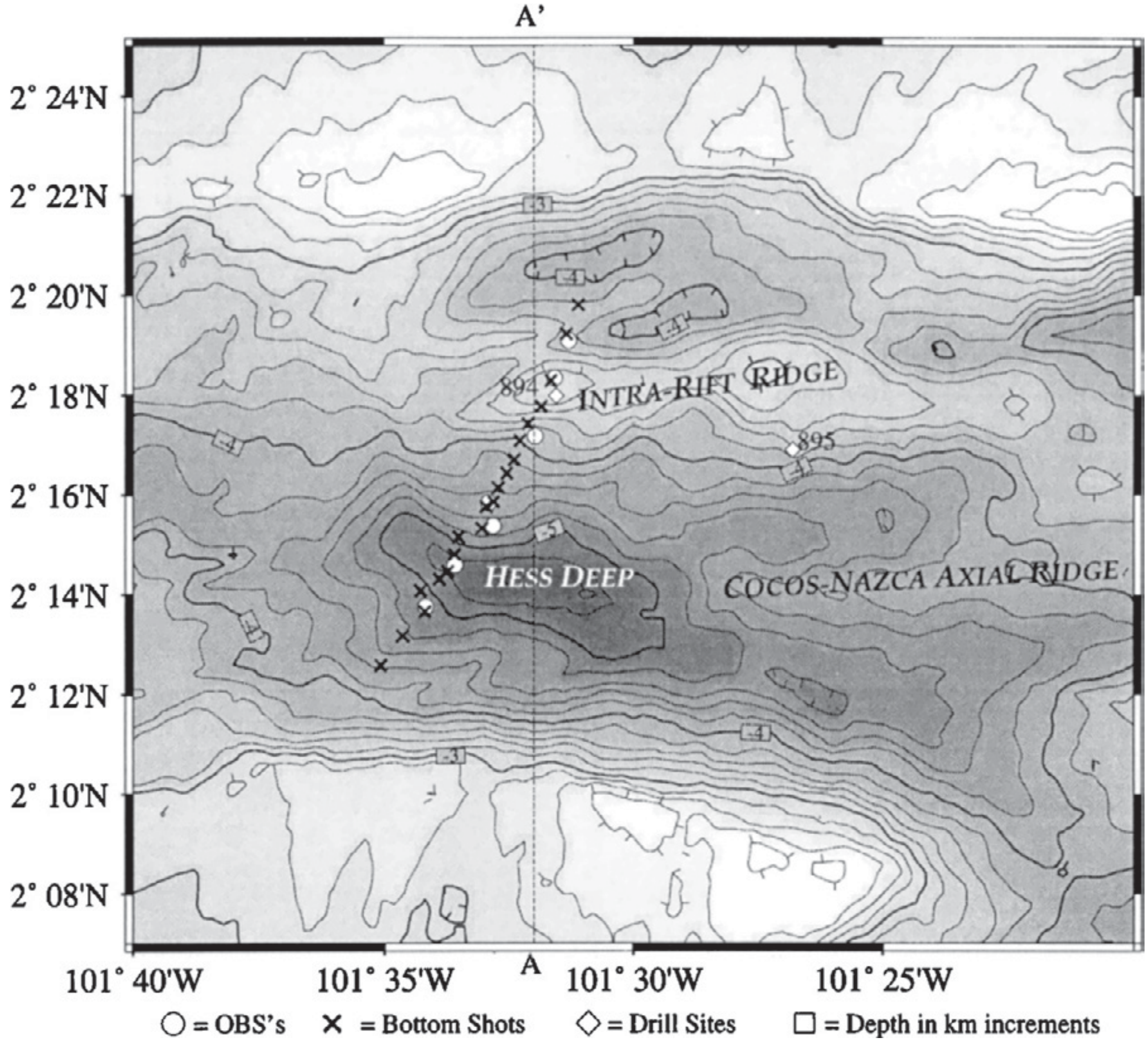



Figure 9.8.2-06. Location of a seismic-refraction profile through the Hess Deep rift valley (from Wiggins et al., 1996, fig. 2). [Journal of Geophysical Research, v. 101, p. 22,335–22,353. Reproduced by permission of American Geophysical Union.]

along the primary MELT array which intersected the rise axis near 17°15′S where a shallow axis was present. Four of these instruments, spaced 15–50 km apart, were located on the Nazca Plate to the east along a 160-km-long line. The other six instruments, spaced 15–80 km apart, were employed on the Pacific Plate along a 230-km-long line. The records showed clear crustal refractions ($P_g$) and Moho reflections ($P_MP$), but on only some of the receivers were upper-mantle refractions ($P_n$) observed. Below the top layer with velocities of 2.5–4.0 km/s, crustal velocities increased from 4.0 to 7.0 km/s at a sub-seafloor depth of 3.0–3.5 km. The lower crust was modeled by a 2.5-km-thick layer with P-velocities of 7.0–7.1 km/s on the Nazca Plate and up to 7.2 km/s on the Pacific Plate. Crustal thickness was 5.1–5.7 km at 17°15′S on both sides of the rift axis and was not resolvably different, but a small along-axis difference was found for similarly aged crust on the Nazca Plate, crustal thickness increasing slightly to 5.8–6.3 km at 15°55′S (Canales et al., 1998).

As example of a seismic project in the south central Pacific around 150°W, 17°S the investigation of the Society Island hotspot chain is described here (Fig. 9.8.2-10). The oceanic crust under the Society Islands volcanic chain, which is ~400 km long, ranges from 65 Ma in the south to 80 Ma in the north. The island chain is composed of five atolls and nine islands, one of them being Tahiti. The progressive decrease in age of volcanics eastwards implies that the chain was created by the passage of the Pacific plate over a hotspot source (Grevemeyer et al., 2001b).

The experimental part of the project had already been carried out in 1989. One profile approached Tahiti and covered the abyssal hill terrain to the east of Tahiti, whereas the other profiles covered oceanic crust north of the hotspot region. Twenty-six OBHs were deployed along four profiles. However, only 16 instruments provided seismic recordings useful for geophysical data analysis (Grevemeyer et al., 2001b). The instruments were

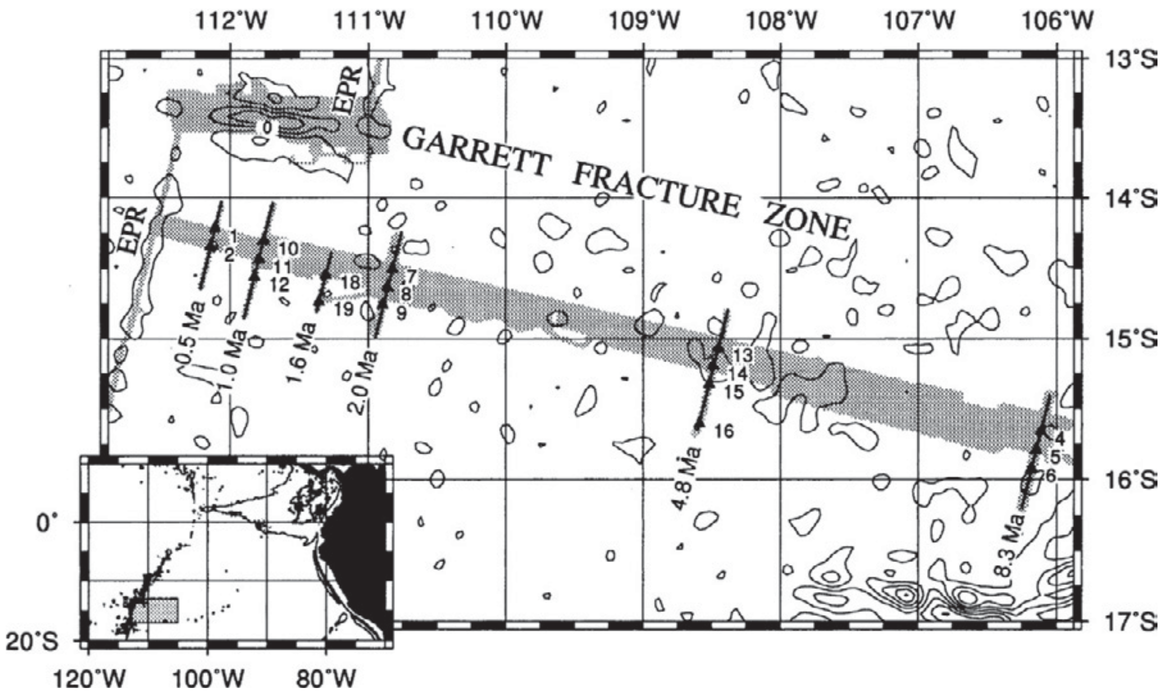



Figure 9.8.2-07. Location of the EXCO-1995 seismic project on the southern East Pacific Rise at 14°S. (from Grevemeyer et al., 1998, fig. 1). [Geophysical Journal International, v. 135, p. 573–584. Copyright John Wiley & Sons Ltd.]

analog pop-up systems from the University of Hamburg and digital systems from IFREMER (Institut Francais de Recherche et d'Exploitation de la Mer) in Brest. An airgun array towed at 10 m depth provided the energy source with a shot spacing of 250 m. No onshore shots had been planned, but in addition to the OBS data, the airgun shots were also recorded at the seismic stations of the Polynesian Seismic Network onshore Tahiti and Moorea, from which picks of the onset times of the first breaks were provided which could be used for the interpretation of profiles 2 and 3. Profile 2, for example, located offshore Tahiti, was ~280 km in length and airgun shots were recorded in-line by 9 OBHs and one land station.

The traveltime modeling provided pronounced variations in crustal structure along the hotspot chain (Grevemeyer et al., 2001b). After removal of the contribution of the uppermost low-velocity layer representing the volcanic edifices and debris, it was suggested that the islands and atolls sit on normal oceanic crust. In general, off the islands a nearly flat crust-mantle boundary at around 14 km depth below sea level was obtained, while under the islands, in response to loading, the crust has bent down and thickened. From amplitude modeling, Grevemeyer et al. (2001b) concluded that the crust on the hotspot trail had been modified with respect to the pre-hotspot crust and that the lower crust had been intruded by sills during the ongoing late-stage volcanic activity and that some melts were trapped in the crust-mantle boundary region.

The next project described in this subchapter was conducted off Taiwan on the Ryukyu margin, forming the boundary of the Philippine Sea plate and the Chinese continental margin at 22°N between 122°30′E and 123°30′E, where a north-south–trending oceanic ridge, the Gagua ridge, is entering the subduction zone (Fig. 9.8.2-11). Here the TAICRUST project was initiated as a seismic imaging program, which utilized deep seismic-reflection profiles and wide-angle reflection and refraction data recorded by both ocean-bottom seismometers and onshore seismic stations. The aim of the project was to investigate the deep structure and geodynamic processes of the south Ryukyu margin and the Taiwan arc-continent collision zone (Liu et al., 1995).

During this project, in 1995 three multichannel deep seismic-reflection profiles, located to the east, west, and north of the Gagua ridge (Fig. 9.8.2-11) were recorded by the R/V *Maurice Ewing* (Schnuerle et al., 1998). A 20-airgun array was fired at 20-s intervals and recorded up to 16 s by a 160-channel digital streamer of 4 km length. A sub-basement reflection was recorded on two of the profiles (EW9509-1 and EW9509-18) at ~1.5 s below the top of the oceanic basement (see, e.g., Fig. 9.8.2-12) as a high-amplitude long-wavelength reflector which could have been the Moho (Schnuerle et al., 1998).

However, if it were the Moho, the igneous oceanic crust between the basement and the above-mentioned sub-basement reflector would be extremely thin, i.e., less than 5 km, and even thinner toward the trench. Previously conducted seismic-refraction surveys offshore south Ryukyu showed a strong velocity contrast between layer 2 and layer 3 (5.6/7.0 km/s) and an abnormally thin layer 3. Therefore, Schnuerle et al. (1998) tended toward the interpretation that their continuous sub-basement re-

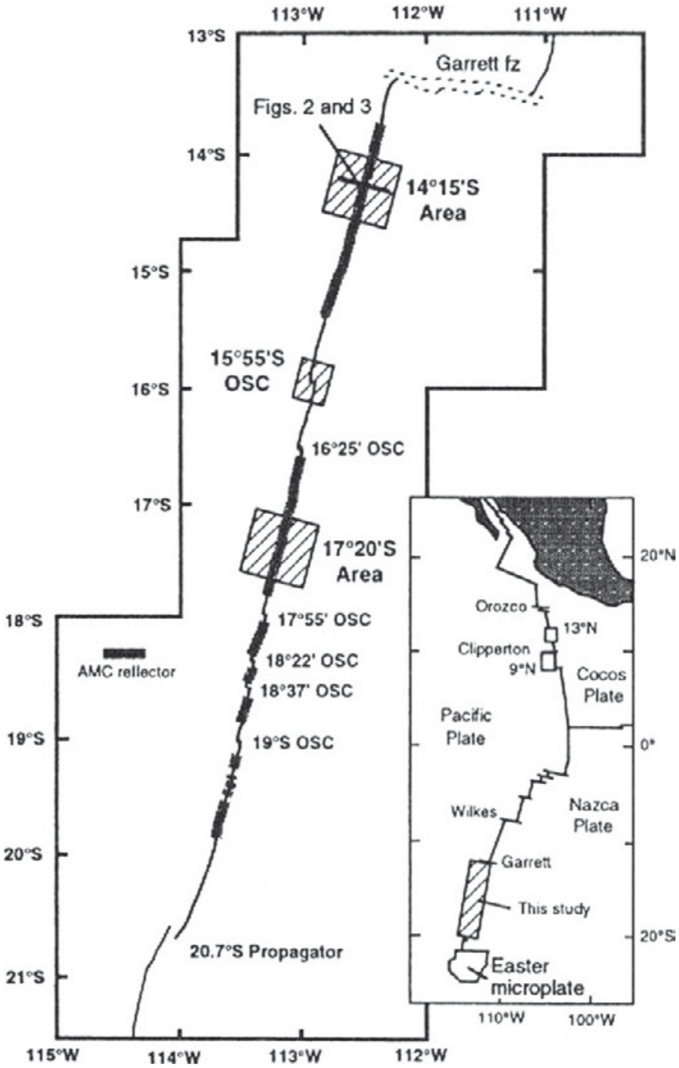



Figure 9.8.2-08. Location of the TERA-1991 seismic projects on the southern Pacific Rise axis south of the Cocos-Nazca-Pacific triple junction (from Detrick et al., 1993, fig. 1). Hatched boxes show areas of extensive reflection/refraction/tomographic surveys. Also shown are reconnaissance profiles and portions of the rise axis associated with an axial magma chamber (AMC) reflector. [Science, v. 259, p. 499–503. Reprinted with permission from AAAS.]

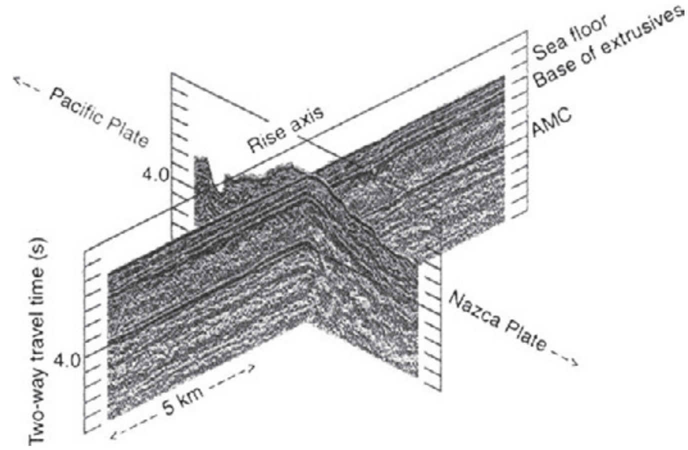



Figure 9.8.2-09. Two intersecting multichannel reflection profiles across and along the crest of the East Pacific Rise (from Detrick et al., 1993, fig. 2). AMC—reflection from top of axial magma chamber, Moho reflection at 5.5–6.0 s not shown. [Science, v. 259, p. 499–503. Reprinted with permission from AAAS.]

flector was rather the top of oceanic layer 3, whereas the Moho, which should be less than 1 s deeper, was not well imaged.

In continuation of the Ryuku Trench toward the northeast follow the Nankei Trough south of Shikoku Island, Japan, the Japan Trench east of Honshu, Japan, and the NE-trending Kuril Trench, east of Hokkaido, Japan (Fig. 9.8.2-13). Several projects were conducted in Japanese waters around Japan.

The instrumentation used in the 1990s and following 2000s was the same for most of the Japanese marine expeditions. The OBSs, designed originally by Shinohara et al. (1993) and converted into a digital system, had gimbal-mounted 4.5-Hz geophones and a hydrophone. Data were recorded by a digital recorder in each OBS. The digital recorder had a 16-bit A/D converter and was able to record data for ~17 days on 4 channels with 100 Hz sampling. Experiments and the airgun array for large distances had a volume of up to 207 l. The design of the onland recording systems in onshore-offshore experiments was similar to the OBS system, with 4.5-Hz geophones and a digital recorder (Kodaira et al., 2002).

In 1994, 1995, and 1997, wide-angle OBS studies investigated the crustal structure of the adjacent Nankai Trough seismogenic zone (Kodaira et al., 2000). The Nankai Trough (Fig. 9.8.2-13; see also Fig. 9.6.3-01), southwestern Japan, is known as a vigorous seismogenic zone with well-studied historic earthquakes. Several OBS profiles were recorded from the shelf areas south of Shikoku Island and Kii peninsula toward the SSW across the Nankai trough into the Philippine Sea (Fig. 9.8.2-14). The 250-km-long line MO104 crossed the presumed seismic slip zone of the 1946 Nankaido earthquake ($M_S$ = 8.2) and was recorded in conjunction with a network of multichannel reflection profiles. The resulting crustal model showed a gentle sloping subducting oceanic crust and a thick overlying sedimentary wedge. The normal oceanic crust was interpreted to consist of two oceanic layers 2 and 3 with velocities of 5.0–5.6 km/s and 6.6–6.8 km/s. The maximum of the sedimentary wedge was 9 km at 70 km to the north from the trench axis with velocities of 3.4–4.6 km/s in its deepest parts. The base of the ~5-km-thick subducting oceanic crust, seen at 10 km depth below sea level under the Philippine Sea, could be traced landward to a depth of 25 km. Under Shikoku Island it is estimated to be at 40 km depth.

In 1999, the Japan Marine Science and Technology Center (JAMSTEC), in cooperation with several universities, undertook another extensive experiment in the western Nankai Trough seismogenic zone and acquired onshore-offshore wide-angle as well as offshore multichannel seismic data (Kodaira et al., 2002).

Along the 185-km-long offshore part of the wide-angle seismic profile, 98 OBSs were deployed into which an airgun array (207 l) fired shots every 200 m. The onland part contained 93 land stations which were deployed at 1–2 km spacing and extended across the eastern side of Shikoku Island and farther to the north onto Honshu Island. Two explosive 500-kg charges were fired at both ends of Shikoku Island and recorded both on- and offshore. Also the airgun shots were successfully recorded onland by the stations on Shikoku Island. The multichannel seismic-reflection data which covered the southernmost 135-km-long section of the profile provided a shallow image of the offshore profile, but also saw clear reflection events at 11–12 km depth which were interpreted as the Moho of the subducting slab. From the wide-angle seismic data, a subducting seamount was found at the center of the proposed rupture zone (see Fig. 9.8.2-14) with dimensions of 13 km thick by 50 km width at 10 km depth and was interpreted as now colliding with the Japanese island arc crust. Beneath the subducted seamount a low P-wave velocity of 7.5 km/s was found (Kodaira et al., 2002).

Park et al. (2002) have interpreted additional multichannel seismic-reflection lines farther east across the Nankai Trough south and southeast of Kii Peninsula which revealed steeply landward-dipping splay faults in the rupture area of the 8.1 Tonankai earthquake of 1944. They were found to branch upward from the plate boundary interface, the subduction zone, at a depth of 10 km ~50–55 km landward of the trench axis, breaking through the upper crustal plate.

Already in 1992, an extensive marine seismic-reflection and -refraction survey was conducted at 32°15′N between 138° and 143° across the northern Izu-Ogasawara island arc system, which extends for more than 1000 km off the southern coast of Japan to the south, between the Izu-Ogasawara Trench and the Shikoku Basin south of Honshu Island (Suyehiro et al., 1996). Controlled sources and natural earthquakes were used to model the 3-D crustal and uppermost mantle structure and subduction slab geometry. The crust was thickest with ~20 km beneath the presently active rift zone. The middle crust with ~6 km/s was confined beneath the arc and comprised ~25% of the crustal volume, the lower crust with 7.1–7.3 km/s velocities comprised another 30% of the crust and the Moho became obscure east of a north-south–trending zone of volcanoes.

In 2000, the Tsushima Basin, located in the southwestern Japan Sea north of southern Honshu, was targeted by a seismic velocity structure survey using OBSs and airguns (Sato et al.,

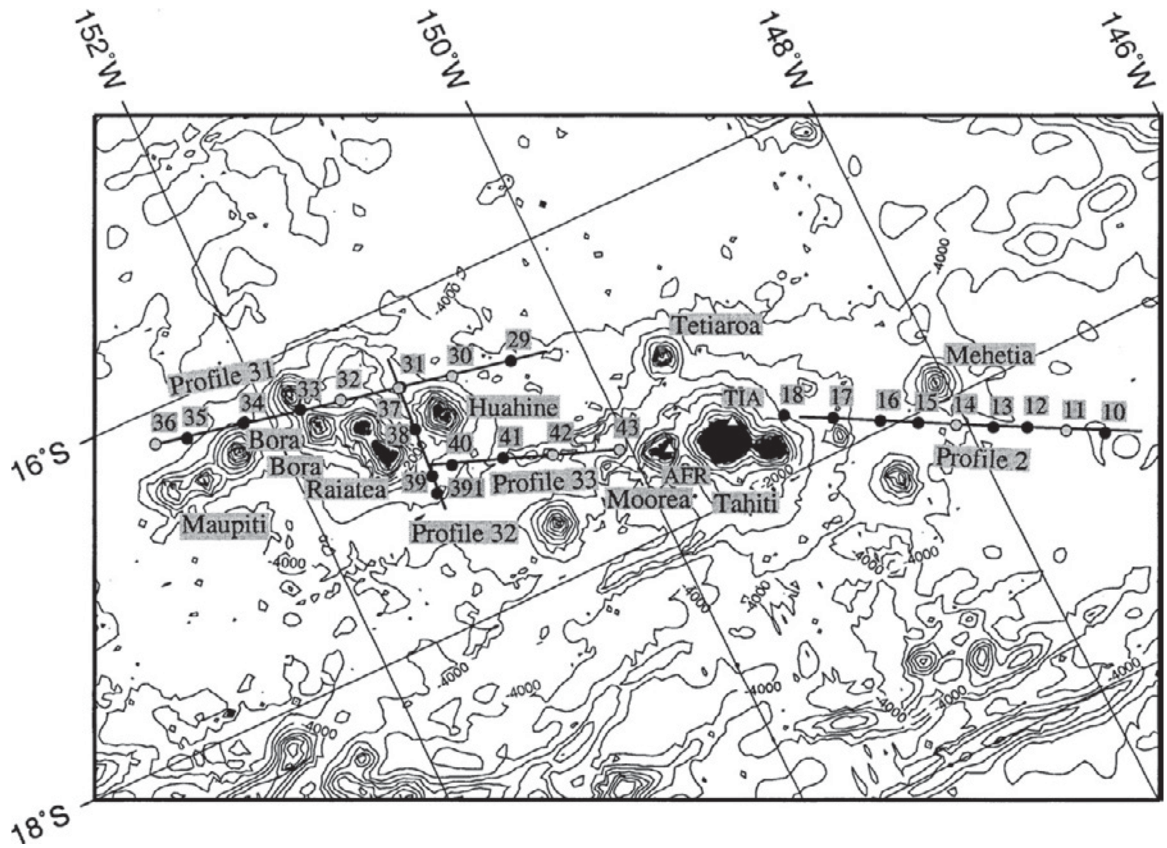



Figure 9.8.2-10. Location of the seismic investigation of the Society Islands hotspot chain in the south central Pacific (from Grevemeyer et al., 2001b, fig. 1). Solid lines are profile segments covered by airgun shots, black dots are OBH positions, and white triangles are land stations. [Geophysical Journal International, v. 147, p. 123–140. Copyright John Wiley & Sons Ltd.]

2006). The survey was an 89-km-long line extending from the southeastern Tsushima Basin toward the southwestern Japan Island Arc. Six OBSs were deployed with 15–20 km spacing, the shot interval was 80–100 s, corresponding to a spacing of ~200–270 m. The resulting crustal thickness was 17 km under the southeastern Tsushima Basin including a 5-km-thick sedimentary layer, and 20 km including 1.5-km-thick sediments under its margin.

Off the coast of northern Honshu, from 36° (latitude of Tokyo) to 41°N and at 142° to 143°E, a series of MCS reflection surveys was performed in the years 1996–2001, all lines crossing the Japan Trench. In total, 3120 km MCS lines were obtained.

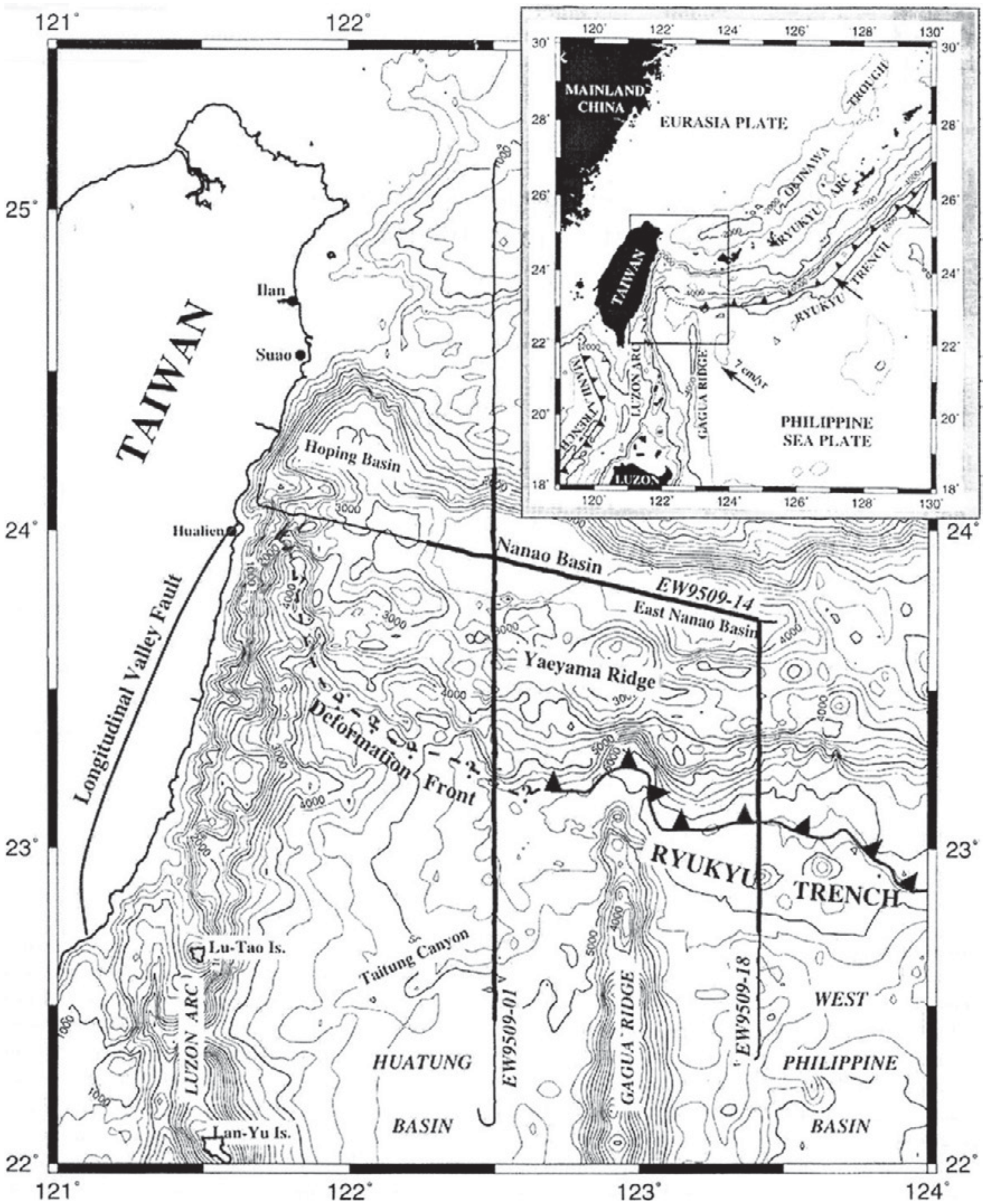



Figure 9.8.2-11. Location of the TAICRUST multichannel seismic profiles EW9509-1 (profile 1), EW9509-18 (profile 2) and EW9509-14 (profile 3) in the southern Ryukyu arc-trench system in the vicinity of the Gagua ridge (from Schnuerle et al., 1998, fig. 1). [Tectonophysics, v. 288, p. 237–250. Copyright Elsevier.]

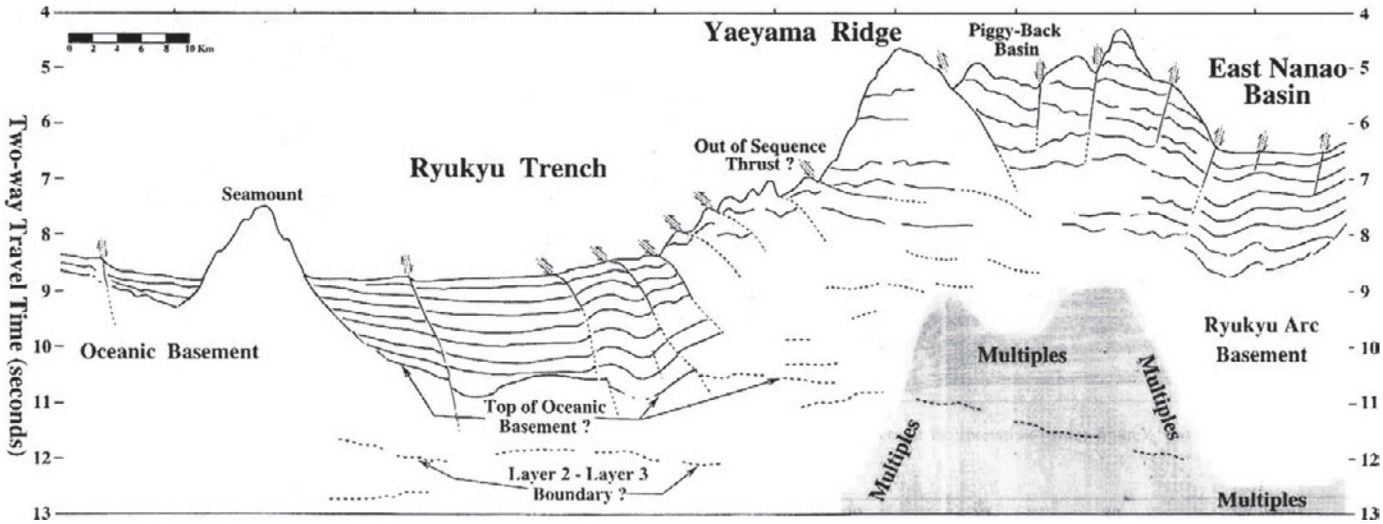



Figure 9.8.2-12. Seismic image along the TAICRUST multichannel seismic profile 2 (EW9509-18) showing at depth the top-basement and the sub-basement reflectors (from Schnuerle et al., 1998, fig. 4). [Tectonophysics, v. 288, p. 237–250. Copyright Elsevier.]

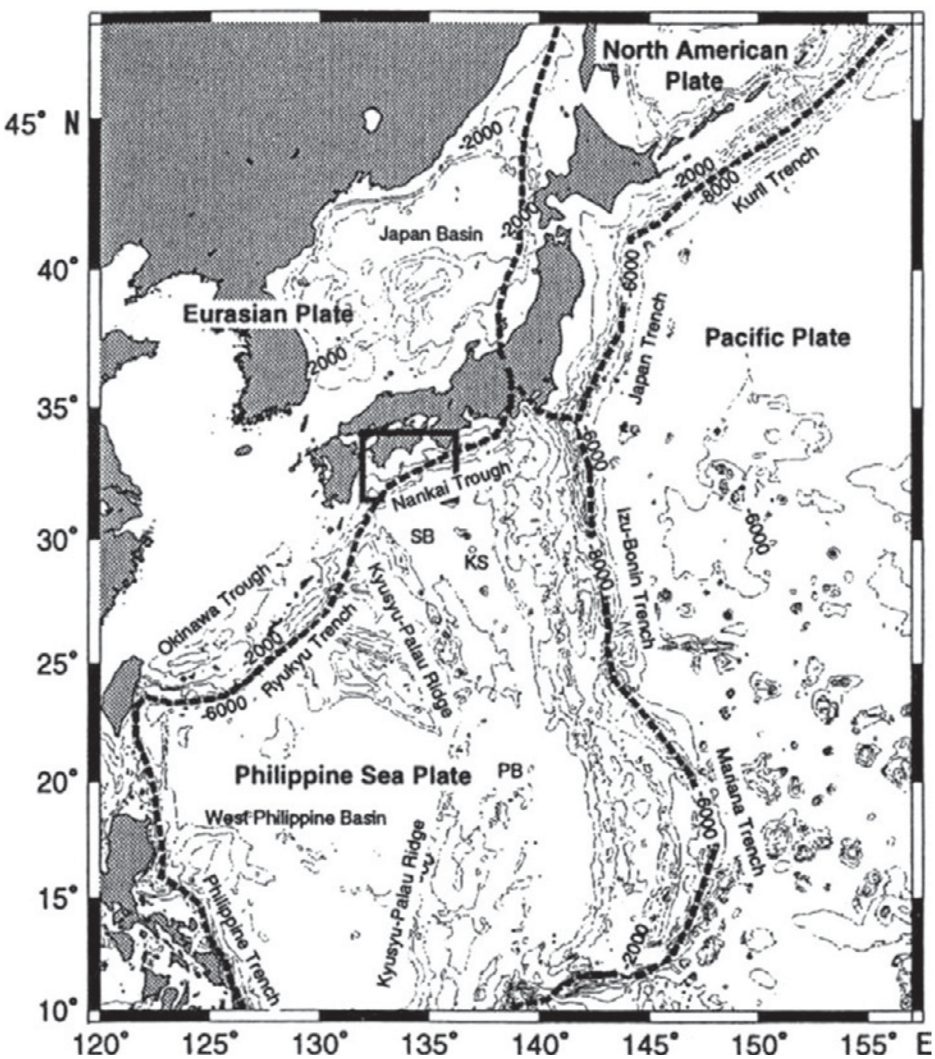



Figure 9.8.2-13. Map of the western Pacific and Philippine Sea showing the geographic location of oceanic features east of the Philippines and China and around Japan (from Kodeira et al., 2000, fig. 1). SB—Shikoku Basin, KS—Kinan Sea Mounts, PB—Parece Vera Basin. [Journal of Geophysical Research, v. 105, p. 5887–5905. Reproduced by permission of American Geophysical Union.]

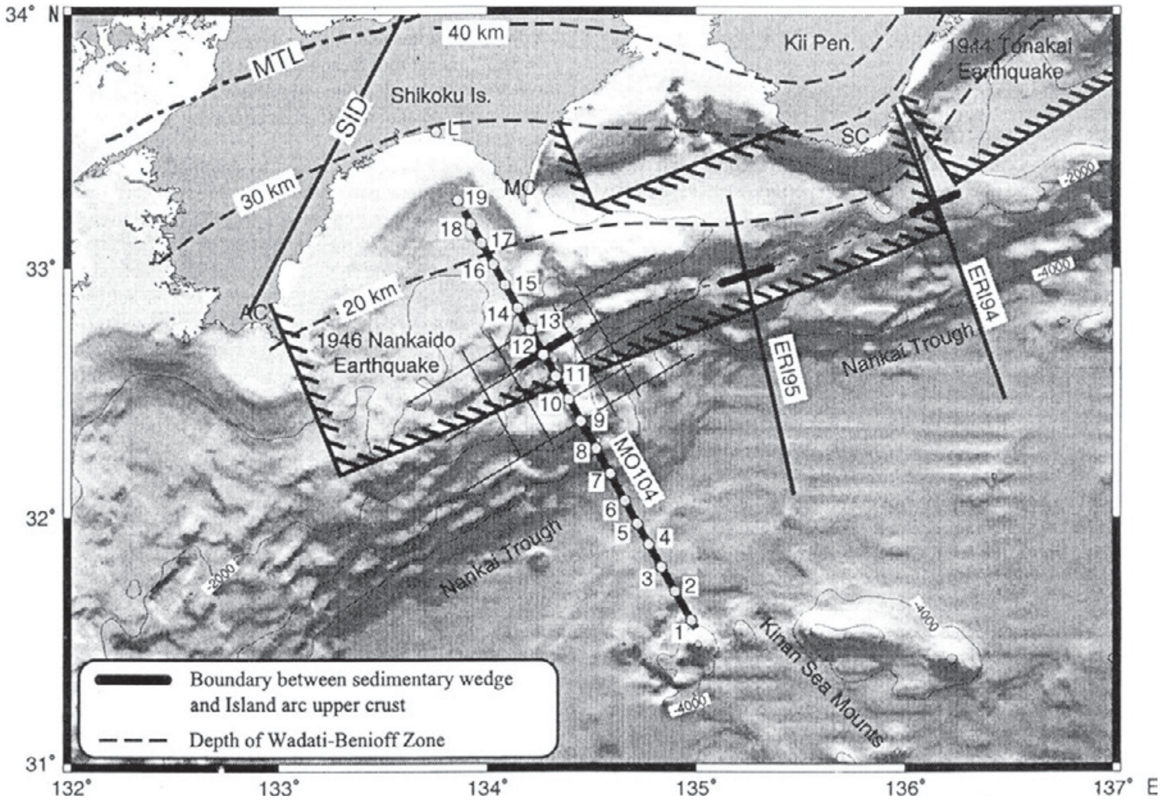



Figure 9.8.2-14. OBS seismic-refraction profiles MO108, ERI95 and ERI94 across the western Nankai Trough seismogenic zone south of Shikoku Island and Kii peninsula (from Kodeira et al., 2000, fig. 2). Also shown are the locations a network of multichannel seismic reflection lines (thin lines) and of the onshore wide-angle seismic profile on Shikoku Island. Thick dashed lines—isodepth contours of the Wadati-Benioff zone (after Hyndman et al., 1995). Short thick marks on the OBS profiles—boundary between sedimentary wedge and island arc upper crust. Rectangular areas—coseismic slip zones of the 1944 and 1946 8.2-earthquakes (after Hyndman et al., 1995). [Journal of Geophysical Research, v. 105, p. 5887–5905. Reproduced by permission of American Geophysical Union.]

For recording, a 3200–4100 m streamer cable and an airgun array of 66-197 l were used (Tsuru et al., 2002). The surveys revealed the tectonic pattern of the sedimentary layers and the top of the igneous crust, but did not penetrate into lower crustal levels.

In 1997, an offshore experiment was carried out in the Pacific Ocean to the east of northeastern Honshu, consisting of two parallel N-S–directed lines of 150–160 km length parallel to the Japan Trench axis farther east and furnished with seven OBSs each (Takahashi et al., 2000) and an E-W–directed line of 180 km length between the coast and the trench axis, along which another eight OBSs and three land stations were deployed (Takahashi et al., 2004). The shot interval of the airgun array along all lines was ~50 m. The OBSs were equipped with a hydrophone sensor and a three-component geophone. The location of these lines is shown in Figure 9.6.3-01 (north of line A–A′). Resulting from the interpretation of these data, the subduction angle increases from east to west, from 3° to 8° to 11°, with the top of the plate located at depths of 11 km at 40 km, 21 km at 120 km, and at 28 km at 140 km distance west from the trench. The Moho beneath the fore-arc was found at ~20 km depth, the velocity of the island arc upper mantle is ~8 km/s. While the velocities in the Cretaceous sedimentary layer and in the upper crust showed large variations in E-W direction, the velocity in the lower crust with 6.7–7.0 km/s was relatively stable. A middle crustal layer, which was found between upper and lower crust on land in some locations, extended 40 km seaward beyond the coastline (Takahashi et al., 2004).

In 1998, a wide-angle survey was recorded in the southern Japan Trench fore-arc region, consisting of two cross lines, one line being 290 km long and running perpendicular to the trench axis (from ~37°N, 141°E to 36°N, 144°E), the other one being 195 km long and running parallel to the trench. OBSs were deployed at 20 km intervals along the trench-parallel profile, on the perpendicular line nine OBSs were deployed at 10 km intervals symmetrically to the crossing point. The airgun shooting interval was 50 m for the 120-channel streamer and the land stations prolonging the marine line into Honshu. The interpretation resulted in a 4-layer crust consisting of Tertiary and Cretaceous sedimentary rocks and the island arc upper and lower crust, underlain by the island-arc Moho at 18–20 km depth. The uppermost mantle of the island arc (mantle wedge) extends to 100 km landward of the trench axis, its velocity increasing from 7.4 km/s at the ocean-side corner to 7.9 km/s below the coastline (Miura et al., 2003).

In 1999, a seismic experiment was conducted farther north in the Miyagi fore-arc region. 36 OBSs were deployed at 3–5 km intervals along a 270-km-long east-west profile at ~38°N which reached from the coast across the Japan Trench to the Pacific plate. During the shooting with a 50 m shot interval a 12-channel streamer for multichannel seismic surveying was towed to determine the shallow structure (Miura et al., 2005). The depth to the Moho was determined to be 21 km at 115 km distance from the trench axis which increased progressively landwards. The P-wave velocity of the mantle wedge was 7.9–8.1 km/s, which the authors regarded as typical for uppermost mantle without large serpentinization. The dip angle increased from 5 to 6° near the trench axis to 23° 150 km landward from the trench axis. The P-wave of the oceanic mantle was as low as 7.7 km/s.

Finally, farther to the north, the Kuril Island Arc and Trench were investigated in 2000 (Kurashimo et al., 2007, Nakanishi et al., 2009) by an onshore-offshore experiment. Because of its large onshore component, it is described in some detail in subchapter 9.6.3.

As noted, since 1985, the Australian Geological Survey Organisation (AGSO) ran a large program of deep seismic-reflection surveys offshore and onshore Australia (Fig. 9.7.1-01), which until 1992 were summarized in a table by Goleby et al. (1994). The offshore program targeted, in particular, Mesozoic basins with hydrocarbon potential. It had started in 1985 with a seismic research vessel operating a 4800 m, 240 channel streamer, which was towed at depths of 10–12 m and a 3000 cubic inch, tuned, sleeve airgun array towed at 9 m depth. The data which were usually sixfold were recorded up to 16 s. The continental margin surveys off Australia continued in the 1990s (Goleby et al., 1994; Finlayson, 2010). A total of 64 research cruises were conducted in the period 1985–1998. A table showing all cruises of the AGSO vessel *Rig Seismic* was also published by Finlayson (2010; Appendix 2-2).

The projects concerning the margins of NW and S Australia will be described in the next subchapter 9.8.3 (Indian Ocean). On the Pacific side of Australia, an investigation of the Macquarie Ridge and the Australian/Pacific plate boundary, southwest of New Zealand, took place in 1994. In cooperation with French and U.S. agencies, the AGSO vessel *Rig Seismic* acquired ~600 km of seismic-reflection data across the ridge (Finlayson 2010; Appendix 2-2).

### 9.8.3. Indian Ocean

In this subchapter, a few projects are described as examples for seismic investigations of the Indian Ocean. Detailed investigations covered the margins of Australia and the Banda Sea east of Timor. Other projects dealt with the center and the western parts of the Indian Ocean.

The offshore program of AGSO in the 1990s mainly covered areas off NW and S Australia. From 1990 to 1993 along the continental margin of northwestern and northern Australia ~35,000 km of mostly deep—commonly 15 s record lengths—seismic-reflection data were acquired. Maps and other details were presented by Finlayson (2010; Appendix 2-2). These surveys included 1894 km of seismic profiling recorded in 1990 in and across the Vulcan graben, two sets of data recorded in 1990 and 1991 in the Arafura Sea, several surveys on the Exmouth Plateau in 1990–1992; another 2100 km of seismic-reflection data were recorded in 1991 and 1993 in the Bonaparte basin–Timor Sea area, ~3460 km of data across the Browse basin were acquired in 1993, and 4085 km of data across the offshore Canning basin were also recorded in 1993.

The deep marine seismic-reflection surveys in 1996 and 1998 along the Vulcan transect in the Timor Sea and the Petrel transect in the Bonaparte basin off northern Australia have already been discussed in subchapter 9.7.1.1 (for location, see Figs. 9.7.1-01 and 9.7.1-02; Petkovic et al., 2000). These surveys had also supplied wide-angle seismic data from OBSs and from a few land stations recording the airgun shots.

The marine seismic profiling of the offshore region of the Otway Basin in late 1994 and early 1995 (for location, see Fig. 9.7.1-09) was already discussed in subchapter 9.7.1 in connection with the near-vertical and wide-angle incidence observations on land across the Otway Basin continental margin (Finlayson et al. (1998).

During 1992, the British Institutions Reflection Profiling Syndicate (BIRPS) and the Indonesian Marine Geological Institute (MGI) jointly investigated the crustal architecture of the Banda Arc. The project involved two long seismic-reflection profiles east of Timor and were 350 km long (Finlayson 2010; Appendix 2-2).

In 1998, a large ocean-bottom seismometer and hydrophone experiment was conducted near 17°S near ODP site 757 on Ninetyeast Ridge, a major Indian Ocean hotspot track which was created in Late Cretaceous and Early Tertiary times by the Kerguelen hotspot. The ridge can be traced for ~5000 km from 30°S latitude, where it joins the Broken Ridge, northward to the Bay of Bengal, where it is buried beneath the Bengal fan (Fig. 9.8.3-01). The aim of the project was to study the regional structure of the hotspot trail and the small-scale structure of the volcanic edifice.

In total, ~2500 km of seismic-refraction/wide-angle and near-vertical incidence data were acquired by R/V *Sonne* in a joint experiment of three German institutions—University of Bremen, GEOMAR at Kiel, and BGR at Hannover (Flueh et al., 1998b, 1999; Grevemeyer et al., 2001a). More than 100 OBSs and OBHs had been deployed and successfully recovered. In addition, a three-channel ministreamer was used. The kernel was a network of profiles in an area of ~55 × 110 km around ODP site 757 (17°S, 88°E) and a roughly 550-km-long east-west transect at 17°S through its center (Fig. 9.8.3-02).

Along the 550-km transect, 60 ocean-bottom receivers provided data used for traveltime inversion and modeling. The transect extended from the Indian Ocean basin across the Ninetyeast Ridge into the Wharton basin. It was composed of three profiles,

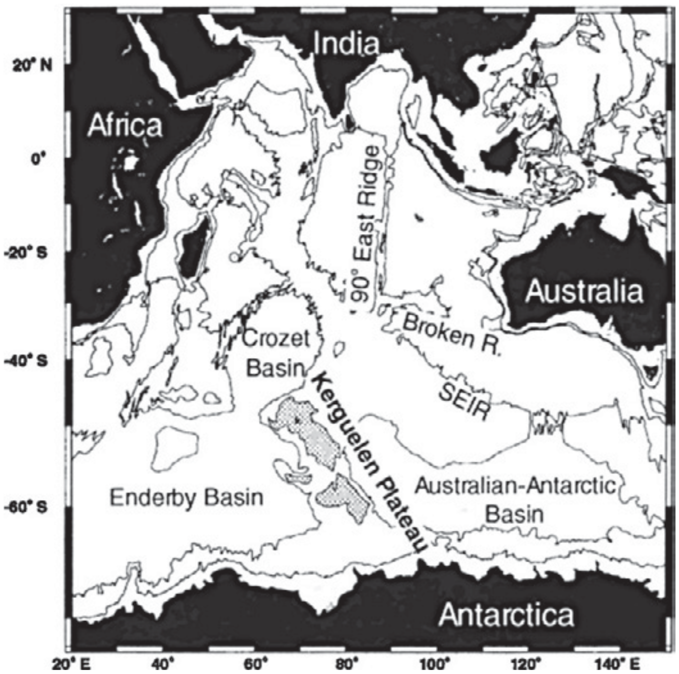



Figure 9.8.3-01. Map of the Indian Ocean, showing the bathymetric contours 2000 m and 4000 m (from Operto and Charvis, 1996, fig. 1). SEIR—South East Indian Ridge. [Journal of Geophysical Research, v. 101, B11, p. 25,077–25,103. Reproduced by permission of American Geophysical Union.]

each being ~300 km long. To simulate a single refraction line, the profiles were shot with an overlap of 100–150 km; 25, 25, and 15 ocean-bottom stations were deployed on profiles 05, 06, and 31, respectively (Fig. 9.8.3-02). The seismic source was a tuned array of 20 airguns providing a total volume of 51.2 l. Shot interval was 60 seconds, resulting in an average shot spacing of 110 m.

The resulting models indicated normal oceanic crust to the west and east of the edifice. Crustal thickness was on average 6.5–7 km. Based on wide-angle reflections from both the pre-hotspot and the post-hotspot crust–mantle boundary, the authors suggest that the crust under the ridge was bent downward, and hotspot volcanism has underplated the preexisting crust, reaching a total crustal thickness of ~24 km (Fig. 9.8.3-03). This underplating also occurred to the east of the ridge under the Wharton basin (Grevemeyer et al., 2001a).

The Southwest Indian Ridge saw two seismic projects in the 1990s during the RRS *Discovery* cruise 208. A wide-angle seismic experiment investigated the Atlantis II Fracture Zone and was interpreted together with geochemical analyses of dredged basalt glass samples from a site conjugate to Ocean Drilling Program hole 735B and thus allowed the determination of the thickness and the most likely lithological composition of the crust beneath hole 735B (Muller et al., 1997). The wide-angle seismic data were recorded by OBH deployed in a grid pattern and with an instrument spacing of 20 km. Shot spacing from an airgun source was ~100 m. The largest offsets from which seismic arrivals from crust and upper mantle were recorded were 25–35 km. The Moho depth was well constrained by $P_MP$ reflections along most of the N-S line. A seismic crustal thickness of 4 km was determined to the north and south of the Atlantis Platform on which hole 735B was located, while beneath the hole the Moho was found at 5 km depth beneath the seafloor. The base of the magmatic crust was estimated to be 2 km beneath the seafloor, underlain by a 2–3-km-thick layer of serpentinized mantle peridotite with a P-wave velocity of 6.9 km/s.

The E-striking Southwest Indian Ridge between 65°20′ and 66°30′E at latitude 27°40′S was the goal of an E-W–directed seismic profile recorded during the RRS *Discovery* cruise 208 (Muller et al., 1999). The line was directed approximately perpendicular to the spreading direction. By large changes in the crustal and seismic layer 3 thicknesses, three segments were defined, anomalously thin crust of 2–3 km being observed beneath the nontransform segment boundaries and 5–6 km beneath the segment midpoints. It was inferred that melt generation is focused on segment midpoints and varies between segments.

Another project investigated the Kerguelen Plateau to the south of the Ninetyeast Ridge (Fig. 9.8.3-01). Here, during

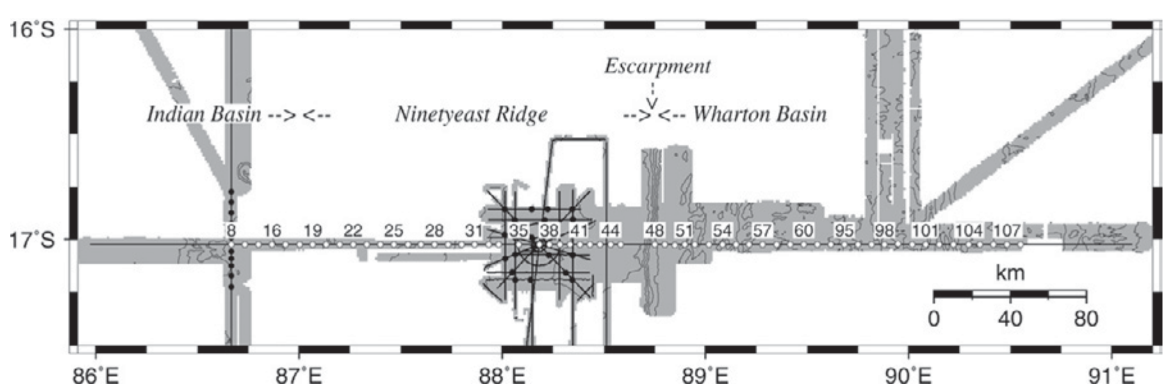



Figure 9.8.3-02. Location of the seismic survey lines along a 550-km-long transect across Ninetyeast Ridge (from Grevemeyer et al., 2001a, fig. 2). [Geophysical Journal International, v. 144, p. 414–431. Copyright John Wiley & Sons Ltd.]

the Kerguelen OBS experiment in early 1991, seven wide-angle seismic lines were shot at sea during the cruise of the M/V *Marion Dufresne* on the Kerguelen Plateau and in adjacent oceanic basins. Of those, two lines, lines 4 and 5, were shot in the Raggatt basin, along NNW-SSE and E-W directions, respectively (Fig. 9.8.3-04).

Operto and Charvis (1996) concentrated their discussion on lines 4 and 5 (lower part of Fig. 9.8.3-04). More than 960 shots were recorded along each seismic line with a maximum source-receiver offset of 165 km and an average spacing of shots of ~185 m. Five OBSs were evenly deployed along each line. The central OBS (OBS 3) was located at the cross point of lines 4

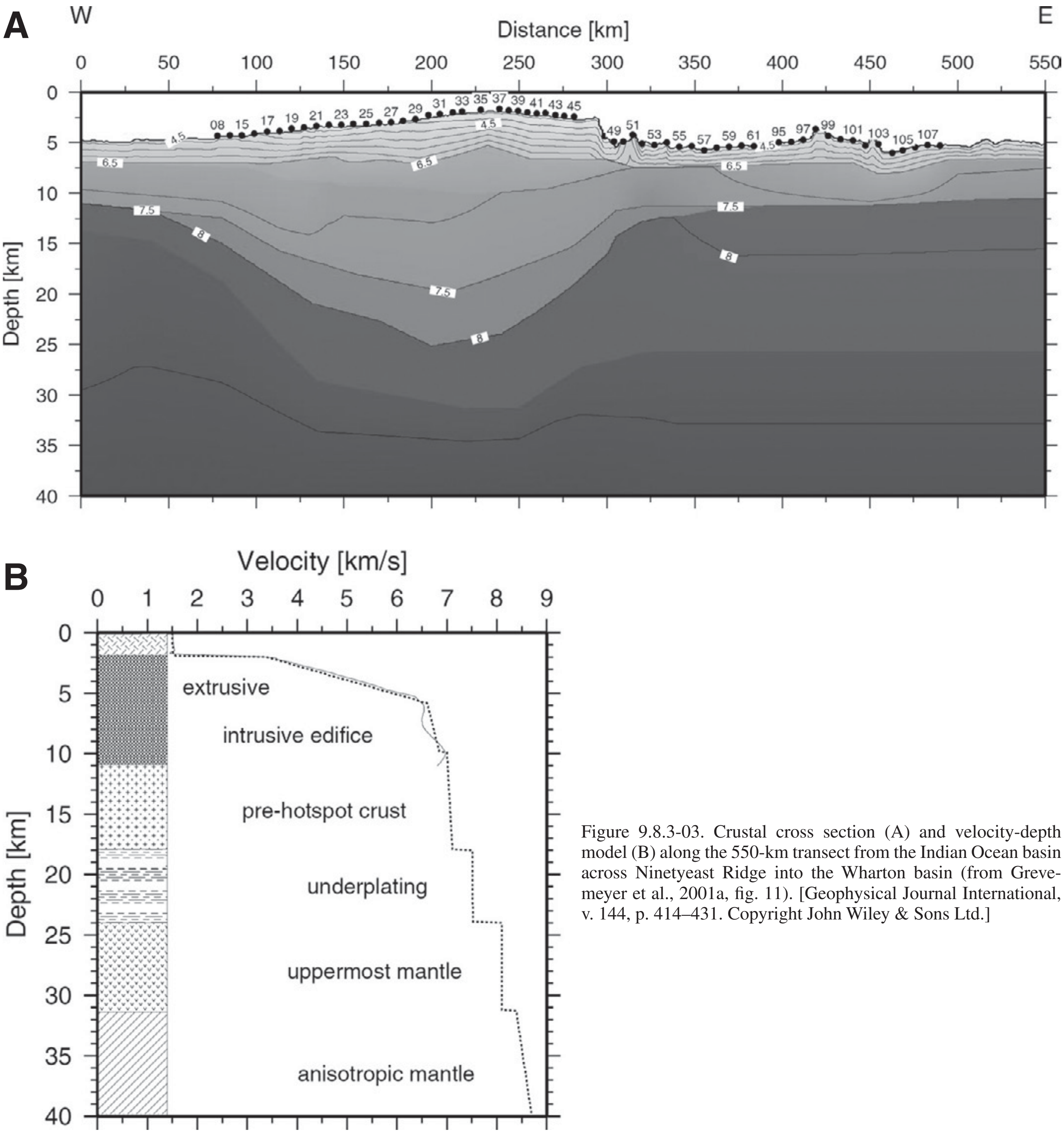



Figure 9.8.3-03. Crustal cross section (A) and velocity-depth model (B) along the 550-km transect from the Indian Ocean basin across Ninetyeast Ridge into the Wharton basin (from Grevemeyer et al., 2001a, fig. 11). [Geophysical Journal International, v. 144, p. 414–431. Copyright John Wiley & Sons Ltd.]

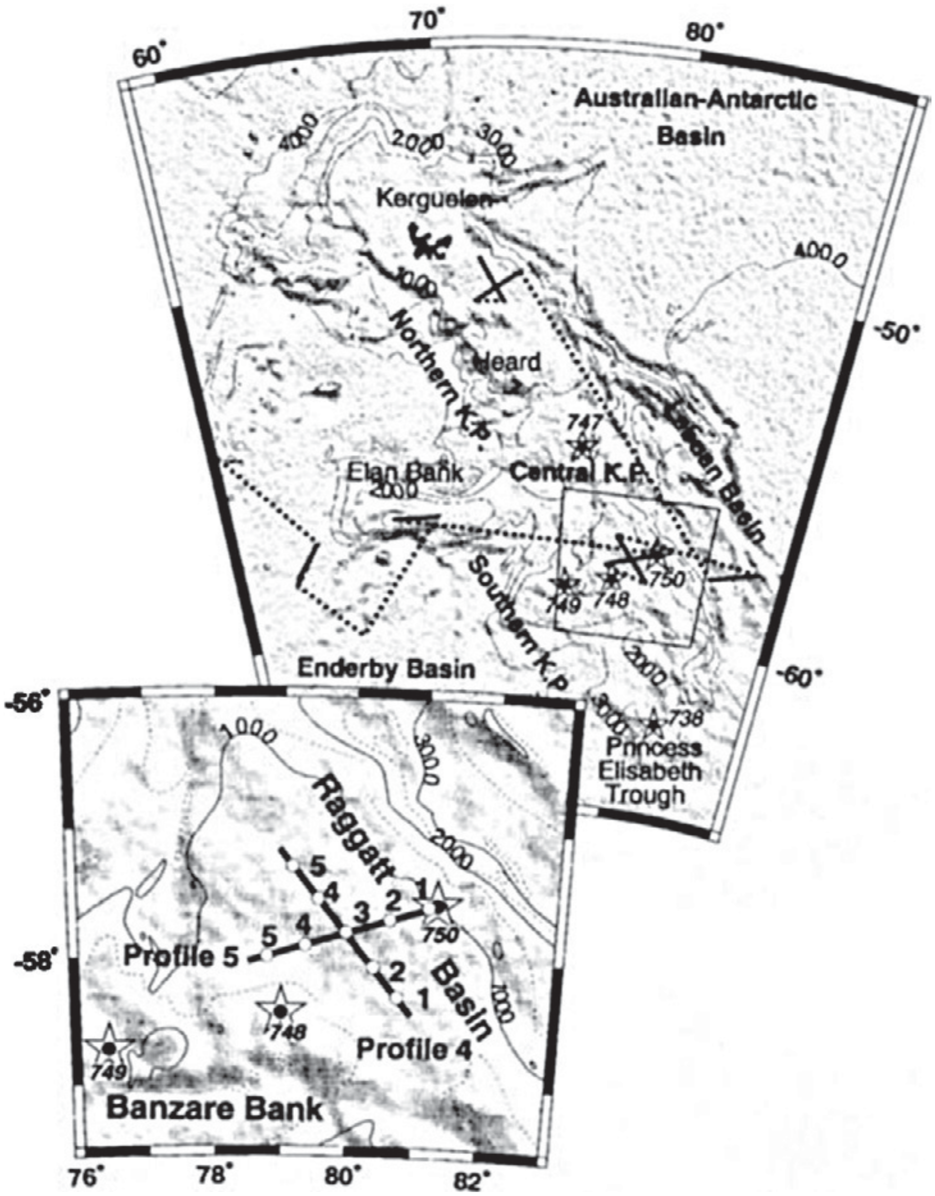



Figure 9.8.3-04. Location of the Kerguelen ocean bottom seismometer experiment (from Operto and Charvis, 1996, fig. 2). [Journal of Geophysical Research, v. 101, B11, p. 25,077–25,103. Reproduced by permission of American Geophysical Union.]

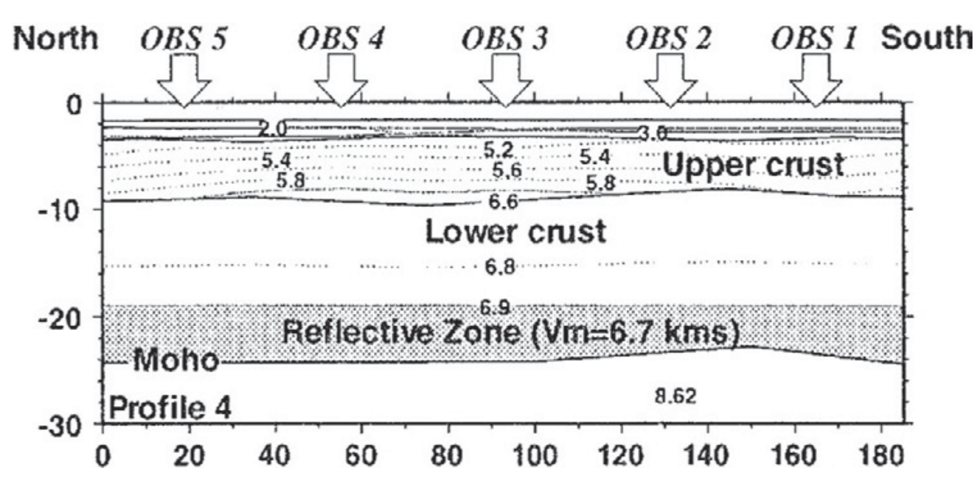



Figure 9.8.3-05. Crustal cross section along profile 4 of the Kerguelen ocean bottom seismometer experiment (from Operto and Charvis, 1996, fig. 4). [Journal of Geophysical Research, v. 101, B11, p. 25,077–25,103. Reproduced by permission of American Geophysical Union.]

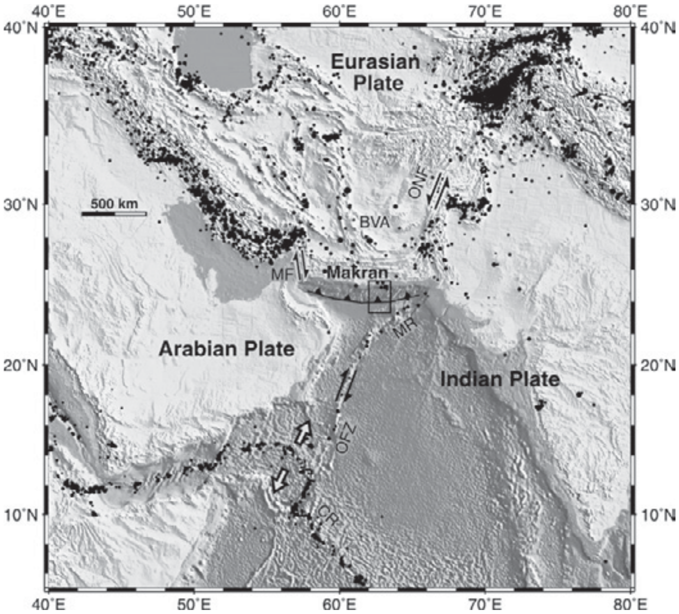



Figure 9.8.3-06. Location of the Makran accretionary wedge off Pakistan (from Kopp et al., 2000, fig. 1). MR—Murray Ridge; rectangle—research area. [Tectonophysics, v. 329, p. 171–191. Copyright Elsevier.]

and 5. The seismic source was an untuned array of eight airguns, 16 l, each fired simultaneously every 100 seconds. Receivers were three-component digital OBSs.

The result of the data interpretation of profile 4 was a 22-km-thick crust, subdivided into a 1.6-km-thick sedimentary cover, an upper crust of 5.3 km, a lower crust of 11 km thickness and a 4–6-km-thick reflective transition zone immediately above the Moho (Fig. 9.8.3-05). The structure along profile 5 was almost identical down to 19 km depth, but with some more relief. However, the transition zone appeared to be seismically transparent (Operto and Charvis, 1996).

The data of lines 1, 2 (the northern two cross lines at 50–51°S, 72–73°E), and 7 (the southwestern line at 57–58°S, 62–63°E) (Fig. 9.8.3-04) were interpreted by Charvis and Operto (1999). The crustal structure under the northern Kerguelen Plateau is almost identical to that shown in Figure 9.8.3-05 for profile 4. In the Enderby basin to the west, however, on line 7, the Moho was found at a shallower depth, dipping from ~15.5 km depth in the SW to 18 km in the NE, while the lower crust with substantially higher velocities of 6.8–7.2 km/s with an undulating upper boundary at 8.5 km depth in the SW and 6 km in the NE thickens in the same direction.

In the northwestern Indian Ocean, the detailed crustal structure of the Makran subduction zone off Pakistan (Fig. 9.8.3-06) was the goal of a cruise of the German R/V *Sonne* in 1997 in a cooperation of GEOMAR, Kiel, and BGR, Hannover in Germany, Cambridge University in Great Britain and the National Institute of Oceanography of Pakistan. The seismic experiment, performed in 1997, was part of a greater project MAMUT (Makran-Murray Traverse) with the objective to study the evolution of recent processes at the Murray Ridge and the Makran accretionary wedge (Kopp et al., 2000). Five wide-angle seismic lines were shot across and along the Makran accretionary wedge (Fig. 9.8.3-07). The 160-km-long, N-S–trending, dip line 8 was chosen to be coincident with an MCS line collected by Cambridge University in 1986. Four E-W–trending strike-lines, 100–125 km long, were shot parallel to the deformation front. Between seven and nine OBHs were deployed along each profile, located between and parallel to adjacent ridges (Kopp et al., 2000).

The interpretation (Fig. 9.8.3-08) revealed that the Moho dipped gently from ~14 km depth to ~22 km depth below sea level along the dip line. A bright reflector was interpreted as a décollement within the turbiditic sedimentary sequence, where a sequence of sediments more than 3 km thick sediments have bypassed the first accretionary ridges and have been transported to greater depth. In the strike lines, low-velocity layers were seen landward of the deformation front. The oceanic crust was identified by a strong velocity contrast between layers 2 and 3 (Kopp et al., 2000).

## 9.8.4. Atlantic Ocean

### *9.8.4.1. South Atlantic Ocean*

The seismic surveys we have collected for the South Atlantic deal mainly with surveys near the continental margins. Continental flood basalts on the conjugate margins of southern Africa and South America and the existence of paired hotspot tracks on the South Atlantic seafloor had been recognized as evidence that the South Atlantic is flanked by plume-related volcanic margins (Fig. 9.8.4-01). Shallow marine seismic data supported this interpretation with the recognition of seaward dipping reflector sequences on both sides of the South Atlantic (Light et al., 1993; Hinz et al., 1999). We first discuss projects exploring margin areas of Africa (Bauer et al., 2000; Edwards et al., 1997), and then projects investigating the offshore region of South America (Franke et al., 2006; Mohriak et al., 1998).

To obtain the hitherto missing deep-crustal seismic data, in 1995 the project MAMBA (Geophysical Measurements across the Continental Margin of Namibia experiment), a combined onshore-offshore multichannel and wide-angle seismic survey, was performed jointly by three German institutions, the BGR (Bundesanstalt fuer Geowissenschaften und Rohstoffe, Hannover), AWI (Alfred-Wegener-Institut, Bremerhaven) and GFZ (GeoForschungsZentrum, Potsdam). The seismic measurements were obtained along two transects (Fig. 9.8.4-02) offshore and onshore of Namibia, traversing deeply eroded basement rocks of the late Precambrian Damara Orogen onshore and extending offshore out across the ocean-continent transition. The experiment consisted of a two-ship marine survey using the Russian vessel R/V *Akademik Nemcinov* and the Norwegian vessel R/V *Polar Queen*. Airgun shots were fired at 60 s intervals, corresponding to a 140 m source spacing, and were recorded simultaneously by OBHs of AWI and by onshore seismometers of GFZ. Distances

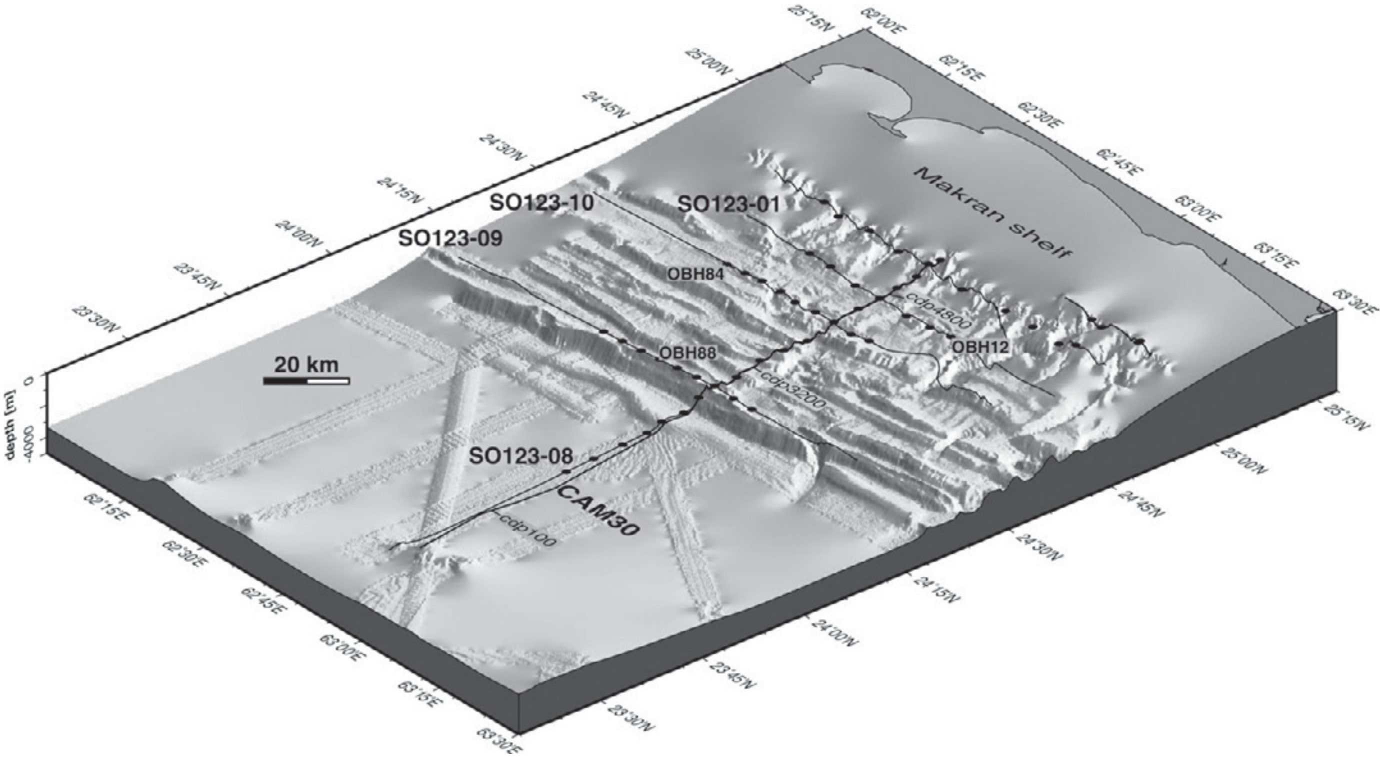



Figure 9.8.3-07. Bathymetric map of the Makran accretionary wedge with locations of seismic profiles and OBH locations (from Kopp et al., 2000, fig. 2). [Tectonophysics, v. 329, p. 171–191. Copyright Elsevier.]

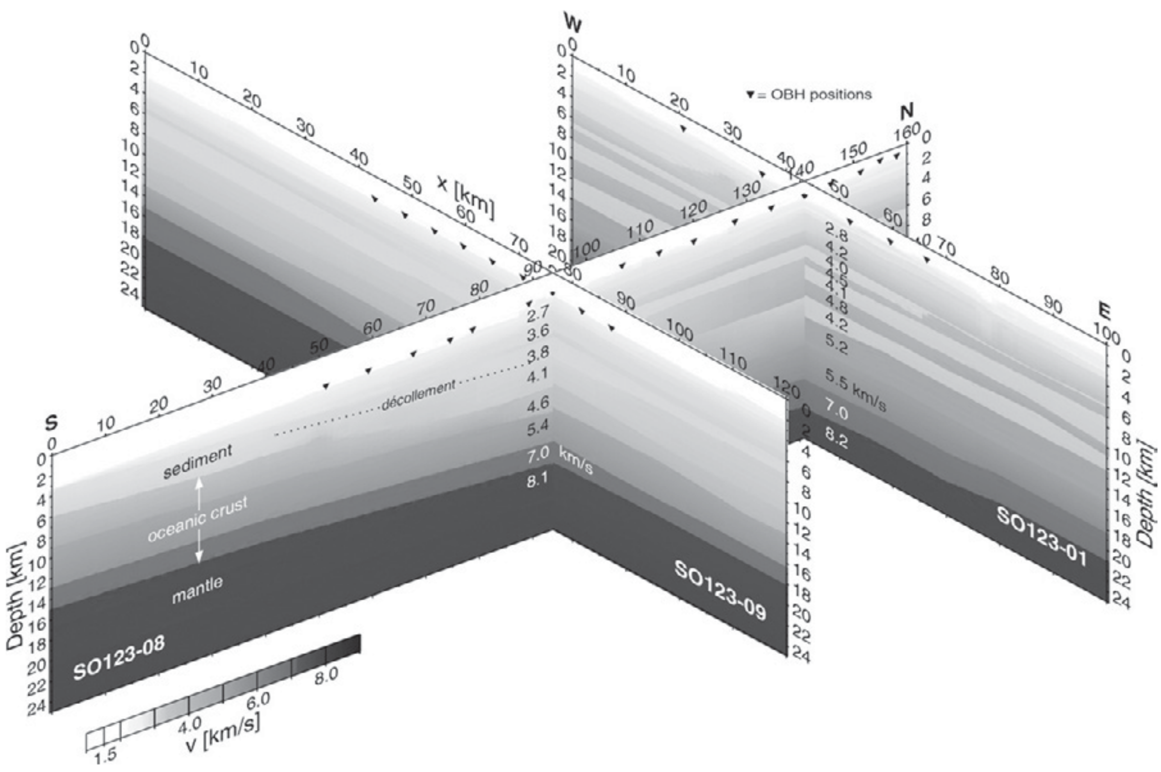



Figure 9.8.3-08. Fence diagram of the wide-angle seismic velocity models displaying shallow subduction beneath the Makran accretionary wedge (from Kopp et al., 2000, fig. 7). [Tectonophysics, v. 329, p. 171–191. Copyright Elsevier.]

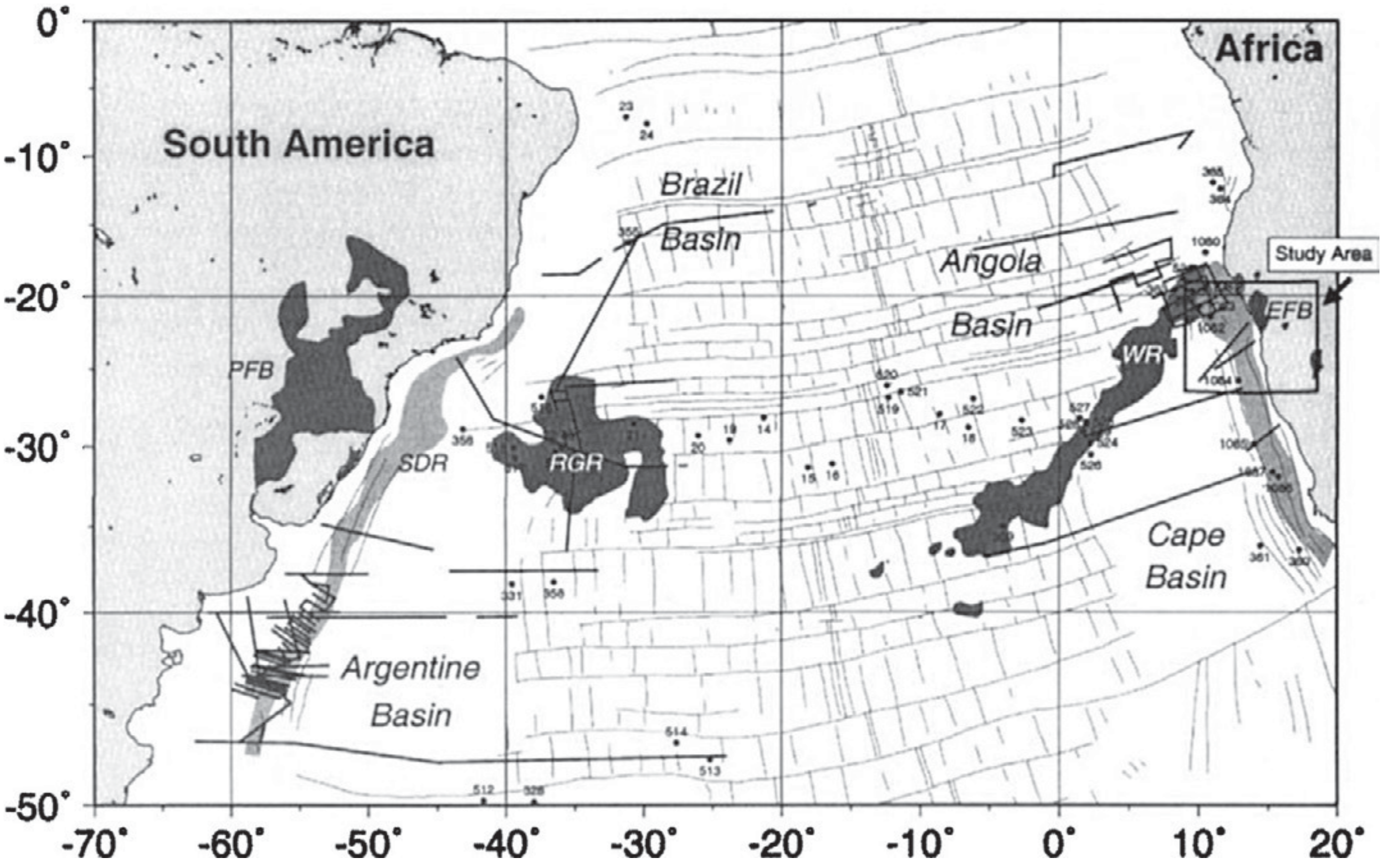



Figure 9.8.4-01. Reflection seismic lines of BGR in the South Atlantic Ocean (from Bauer et al., 2000, fig. 1) including transects 1 and 2 of Figure 9.8.4-02 (box "Study Area"). Grey areas—appearance of seaward dipping reflectors (SDR); black areas—magmatic features (PFP Parana flood basalts; RGR—Rio Grande Rise; WR—Walvis Ridge; EFB—Etendeka flood basalts) related to the Mesozoic opening of the South Atlantic. Numbers indicate DSDP/ODP drill sites. [Journal of Geophysical Research, v. 105, p. 25,829–25,853. Reproduced by permission of American Geophysical Union.]

between the OBHs varied between 40 and 70 km. On land, 25 recorders were spaced from 2.5 km (coastal region) to 15 km (NE end of the lines) apart. The refraction/wide-angle reflection data were compiled as receiver gathers for each OBH and land station (Bauer et al., 2000).

The interpretation (Fig. 9.8.4-03) shows a crustal thickness of ~10 km at the SW end of both lines. The Moho depths indicated a typically oceanic crust at the SW end of both lines to ~320 km distance from the shoreline. It was followed by a 150–200-km-wide zone of igneous crust of up to 25 km thickness off Namibia. The continental crust along the transects reflected differences in the regional geology. The uniform velocity structure of the southern transect showed similar values as derived by the 1975 Damara project (see Fig. 7.7.2-03) for the southern central zone of the Damara Belt (Baier et al., 1983).

In 1993, Edwards et al. (1997) explored the transform margin off Ghana. Six lines, between 80 and 120 km long, were shot parallel to the margin in SW-NE direction with ~10 km spacing, from the continental shelf down to the Guinea basin. A seventh line was normal to and bisected these lines. Airgun shots with 170 m and 300 m spacing as well as explosive charges with 1.0–1.5 km spacing were recorded by digital OBS. The modeling of this suite of wide-angle seismic lines across the continental margin at the eastern end of the Romanche fracture zone off Ghana showed that the transition from continental to oceanic crust is evidently confined to a narrow, 6–11-km-wide zone, located at the foot of the steep continental slope. The crust of the ocean-continent transition proved to be characterized by high velocities (5.8–7.3 km/s). The continental crust appeared unaffected by the proximity of the adjacent oceanic crust, and there was no evidence for underplating of continental crust adjacent to the transition zone.

On the South American side of the South Atlantic Ocean, one project dealt with the broad shelf off Argentina, where in 1998, a 665-km-long onshore-offshore refraction profile was recorded along latitude 40°S across the Colorado Basin, a sedimentary basin separated from the deep-sea basin by a basement high paralleling the continental margin (Franke et al., 2006). Along the 457-km-long marine wide-angle seismic profile and accompanying MCS lines, nine OBHs systems were deployed at an average spacing of 40 km. In addition, five three-component

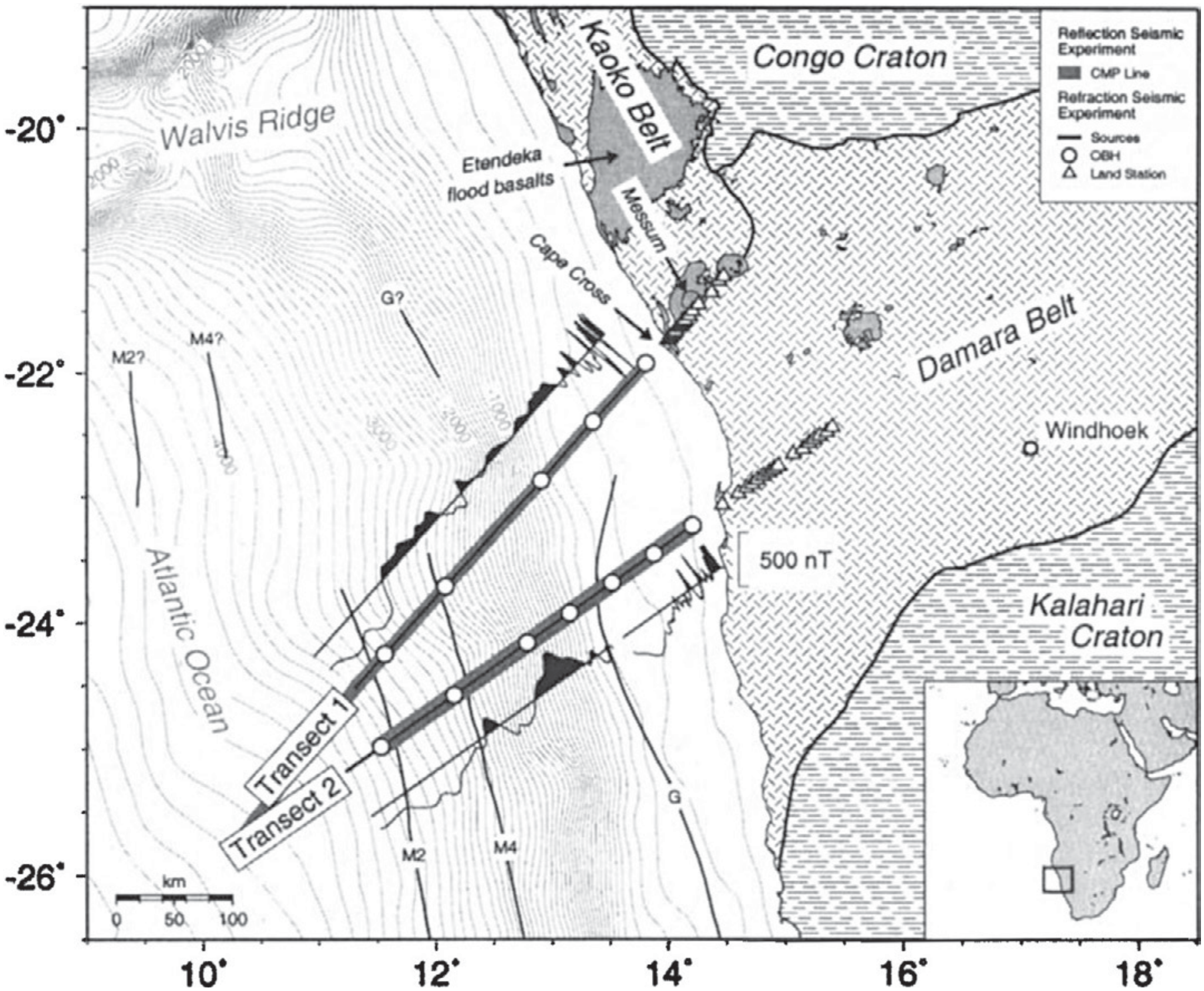



Figure 9.8.4-02. Location map of the offshore/onshore seismic reflection and refraction experiment off Namibia (from Bauer et al., 2000, fig. 2). The wiggle traces beside the seismic lines are simultaneously measured magnetic data. [Journal of Geophysical Research, v. 105, p. 25,829–25,853. Reproduced by permission of American Geophysical Union.]

seismometers were operated on land along the western prolongation of the refraction line BGR98-01 (Fig. 9.8.4-04).

The land stations were six-channel digital recorders (Teledyne PDAS 100), each with a three-component 1-Hz geophone and a six-4.5-Hz-geophones string sensor (SM-6). The seismic source was a tuned set of four linear subarrays with 32 airguns, fired at equidistant shot intervals of 175 m. The resulting velocity-depth cross section (Fig. 9.8.4-05) showed a relatively homogeneous three-layered crystalline crust with Moho depths around 30 km along the whole line.

Farther north, another seismic project concentrated on divergent margin basins located in the South Atlantic Ocean at the northeastern Brazilian margin aiming for petroleum exploration. It consisted of the acquisition of regional seismic profiles in ultradeep waters, close to the boundary between continental and oceanic crust. Here, during an extensive program of deep reflection profiling, high-quality 18 s (two-way traveltime) seismic-reflection profiles were acquired in the Sergipe-Alagoas and the Jacuipe basins off NE Brazil in the period between 1989 and 1991 by Petrobras (Petroleo Brazileiro S.A.; Rio de Janeiro, Brazil). Three lines, 80–110 km long, extended from the coastline toward the boundary with the oceanic crust (profiles A to C in Fig. 9.8.4-06). Along these lines special seismic acquisition techniques were applied. The cable lengths varied between 4 and 6 km, serving 160–240 channels. Airgun shots were spaced at 50 m, and the sampling rate of the digital recording was 2 ms.

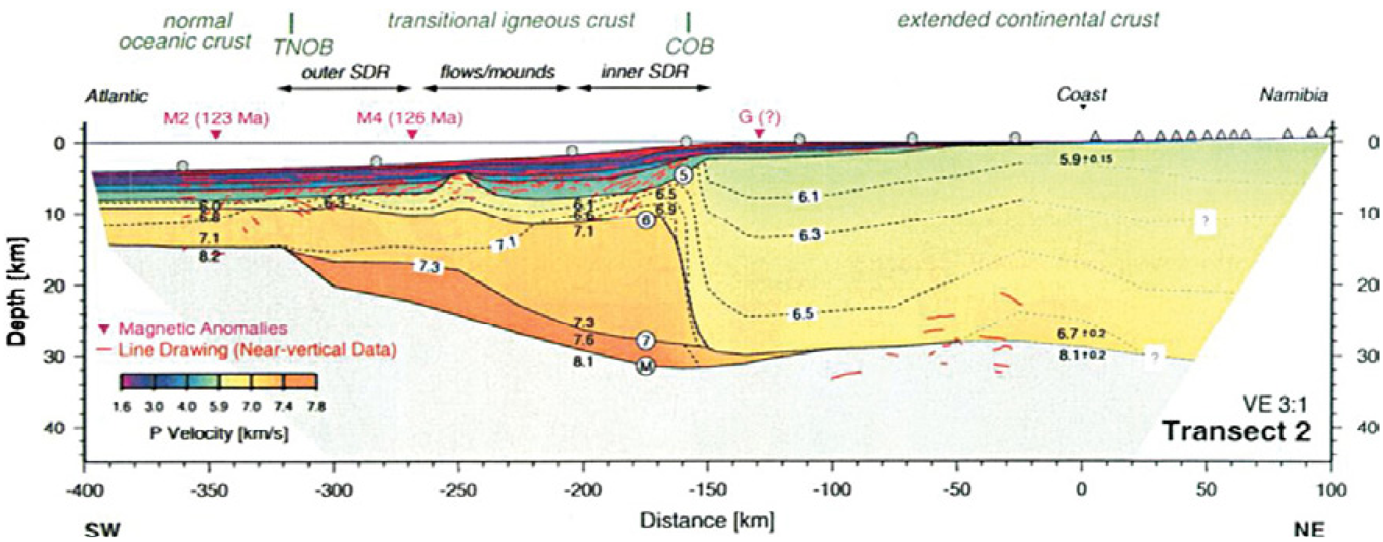



Figure 9.8.4-03. Crustal cross section along the southern Transect 2 of the MAMBA experiment (from Bauer et al., 2000, plate 1). [Journal of Geophysical Research, v. 105, p. 25,829–25,853. Reproduced by permission of American Geophysical Union.]

Other marine data with record lengths of 12 s were acquired in 1992 by the Brazilian Navy in cooperation with PETROBAS at a geophysical survey along the continental margin off northeastern Brazil (Gomes et al., 2000) to map the location of the ocean-continent transition. Known as EPLAC project, over 45,000 line-kilometers of multichannel seismic, gravity, magnetic and bathymetric data were collected along the Brazilian Continental Margin. Around the transition, ~50–100 km off the coast line, the Moho depth was found at a rather uniform depth of 15 km below sea level, the crystalline crust being ~8 km thick, overlain by 2–3-km-thick sediments.

Combined with potential field data and calibrated with the results from several exploratory boreholes, the seismic data provided evidence that the deep seismic reflectors in the deep-water region were related to crustal architecture rather than to sedimentary features. The record section of profile C is shown as an example (Fig. 9.8.4-07), showing the rapid decrease of crustal thickness from near 10 s to ~5 s over a distance range of 80 km. The section also shows a seaward dipping reflector wedge, which Bauer et al. (2000) also found on the Namibian side. Mohriak et al. (1998) pointed out that the deep-water province is characterized by wedges of reflectors that dip seaward marking the transition to a pure oceanic crust. A schematic model (Fig. 9.8.4-08) shows the gradational thinning of the crust (including sediments) from nearly 27 km thickness at the coast to ~10 km thickness at 100 km seaward distance and 4 km water depth, corresponding to a thinning of the crystalline crust from 18 to 9 km thickness.

A review of the major continental margin surveys performed in the 1980s and 1990s along the western borders of the Atlantic Ocean, both for South and North America, was prepared by Talwani and Abreu (2000) discussing the marine seismic-refraction and -reflection data in the context with other geophysical observations with the aim to compare and contrast the U.S. coast volcanic continental margin with the volcanic continental margins of the South Atlantic.

### *9.8.4.2. Mid-Atlantic Ridge*

The Mid-Atlantic Ridge was repeatedly studied since the 1960s. In each following decade, new experiments followed with modernized instrumentation. A particular goal was to map the seismic properties of the slow spreading Mid-Atlantic Ridge in contrast to the fast-spreading East Pacific Rise.

In 2000, a grid of four wide-angle seismic profiles were shot during the second leg of the *Meteor* (1986) cruise M47, using airguns and OBS or OBH across interesting structures of a segment boundary at 5°S on the Mid-Atlantic Ridge. This region had been selected because satellite gravity data showed that this region was relatively undisturbed: The 5°S Fracture Zone separates relatively wide spreading segments and is itself characterized by a single structure with an offset of about 50 km. Two profiles crossed the median valley obliquely, the other two crossed the 5° transform fault and fracture zone.

In total, 77 successful deployments of the GEOMAR ocean-bottom hydrophones and ocean-bottom seismometers were carried out and provided a solid database to derive a detailed picture of the seismic velocity field. Shooting along the profiles was made using up to three 32-l airguns. Velocity models showed a strong velocity gradient in the first 2–3 km below the seafloor, underlain by a lower velocity gradient in the lower crust/upper mantle. In the Median Valley, the velocities were significantly reduced. The crust mantle boundary was not well defined in all regions, but crustal thickness varied between 3 and 6 km. In general, morphological highs were marked by elevated velocities below the seafloor, coincident with results from dredging that recovered gabbros (Borus et al., 2001).

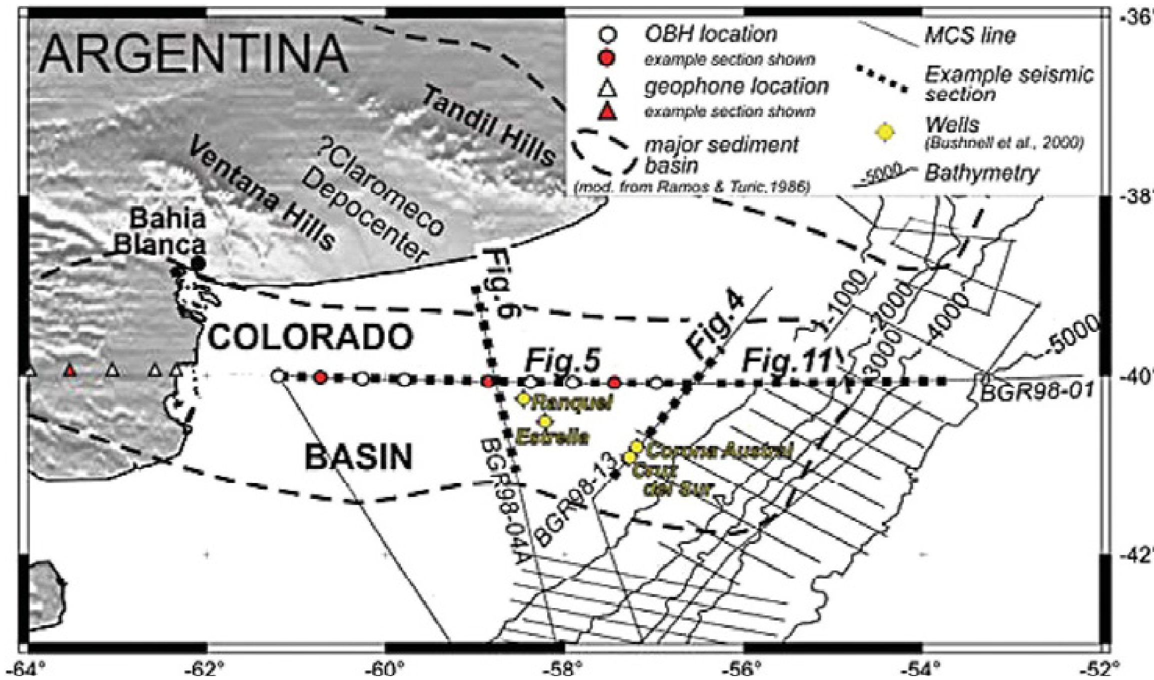



Figure 9.8.4-04. Location map of the offshore/onshore seismic reflection and refraction experiment off Argentina across the Colorado Basin (from Franke et al., 2006, fig. 2b). [Geophysical Journal International, v. 165, p. 850–864. Copyright John Wiley & Sons Ltd.]

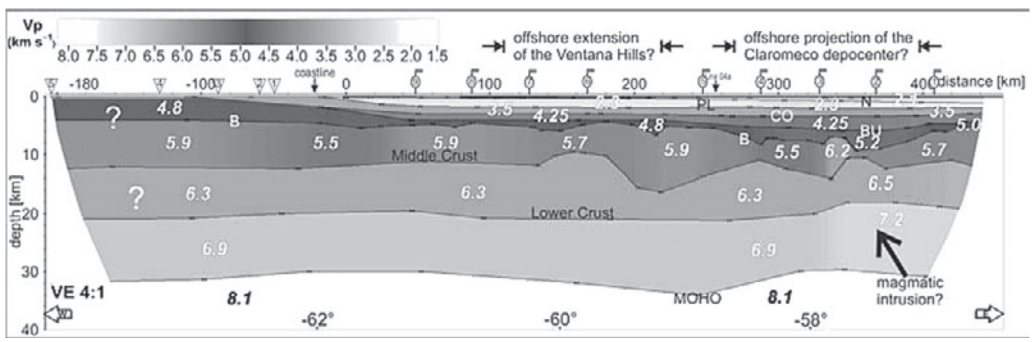



Figure 9.8.4-05. Crustal cross section along latitude 40°S across the Colorado Basin off Argentina (from Franke et al., 2006, fig. 9). [Geophysical Journal International, v. 165, p. 850–864. Copyright John Wiley & Sons Ltd.]

A group of projects (MARBE, Mid-Atlantic Ridge Bull's Eye) covered the Mid-Atlantic Ridge south of the triple junction west of the Azores. Another area of interest was a segment of the Reykjanes Ridge, investigated under the umbrella of BRIDGE (British Mid-Ocean Ridge Initiative). In 1996, the MARBE Experiment covered three spreading segments (Fig. 9.8.4-09) from ~33°30′ N to 35°30′ N longitude at around 37° W latitude (Canales et al., 2000; Hooft et al., 2000; Hosford et al., 2001). The 1996 cruise of R/V *Maurice Ewing* was the first major controlled-source seismic experiment on the northern Mid Atlantic Ridge south of the Reykjanes Ridge in over 15 years and was located ~450 km southwest of the Azores triple junction (Fig. 9.8.4-10A). Three seismic-refraction profiles followed the rift valleys (OH-1, OH-2 and OH-3 in Fig. 9.8.4-09), which are separated from each other by non-transform offsets (NTO-1 to NTO-3 in Fig. 9.8.4-09).

The shot interval was 90 seconds, resulting in an average shot spacing of ~200 m. The 90-km-long refraction profile along the northern segment, OH-1, comprised ten receivers, spaced 7–15 km apart, and 485 shots. Along OH-2 five stations, spaced 12–15 km apart, recorded 301 shots, and six stations along OH-3 also recorded 301 shots (Hooft et al., 2000). Furthermore, two off-axis refraction profiles were shot in segment OH-1. The western line was located in the western rift mountains (Canales et al., 2000) and crossed the non-transform offset NT01, following the eastern rift mountains of segment OH-2 (numbered line in Fig. 9.8.4-10). The line was 155 km long and comprised ten OBHs which recorded 615 airgun shots, fired at an interval of 120 seconds, providing a seismic trace spacing of ~250 m. The eastern line was 70 km long and was located in the eastern rift mountains of segment OH-1, 20 km away from the rift valley (Hosford et al., 2001). Here, eight OBHs were deployed at an

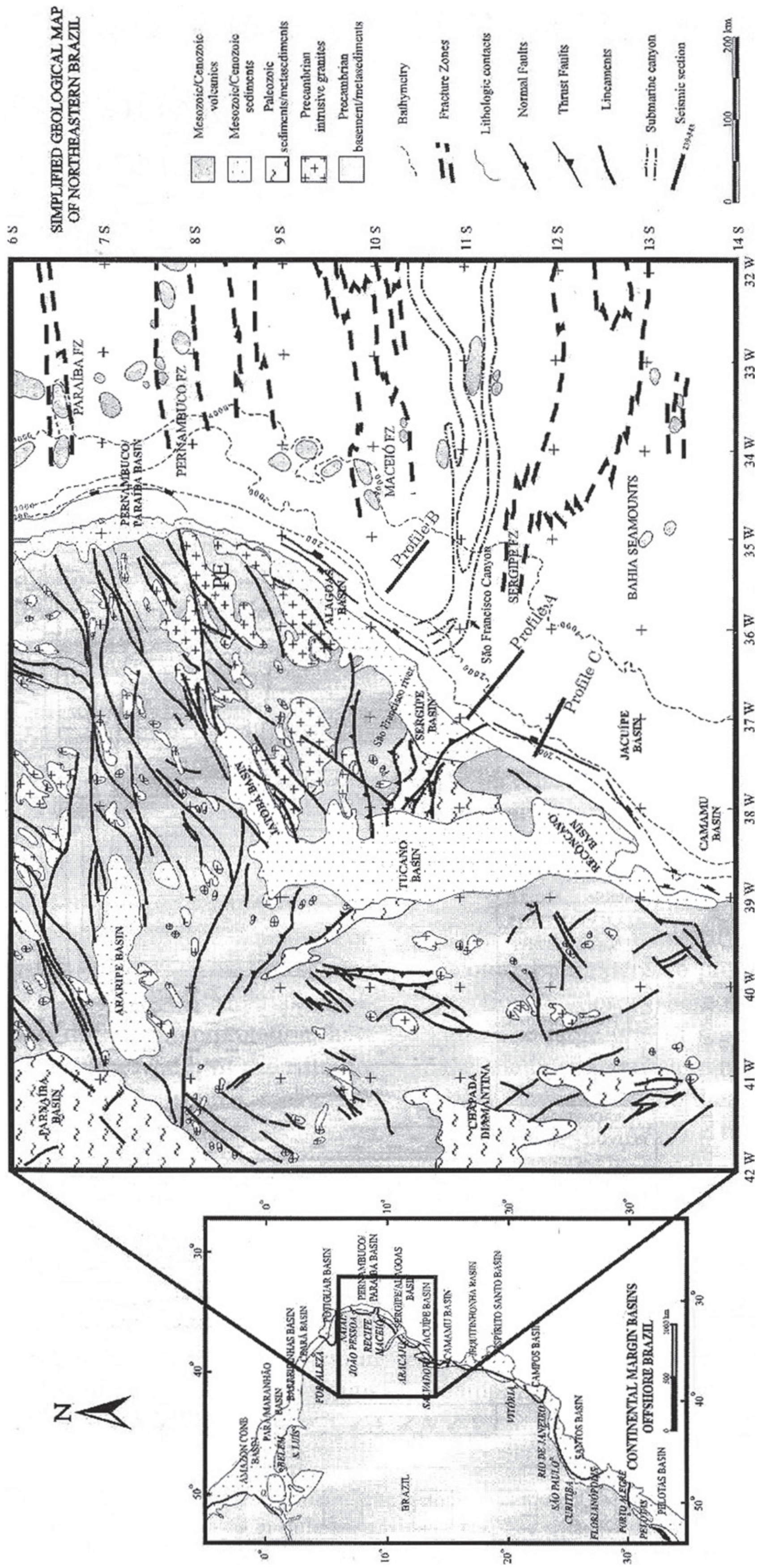



Figure 9.8.4-06. Location of three regional deep seismic profiles A, B, and C on the northeastern Brazilian margin on a regional map with tectonic elements along the South Atlantic Ocean (from Mohriak et al., 1998, fig. 1). [Tectonophysics, v. 288, p. 199–220. Copyright Elsevier.].

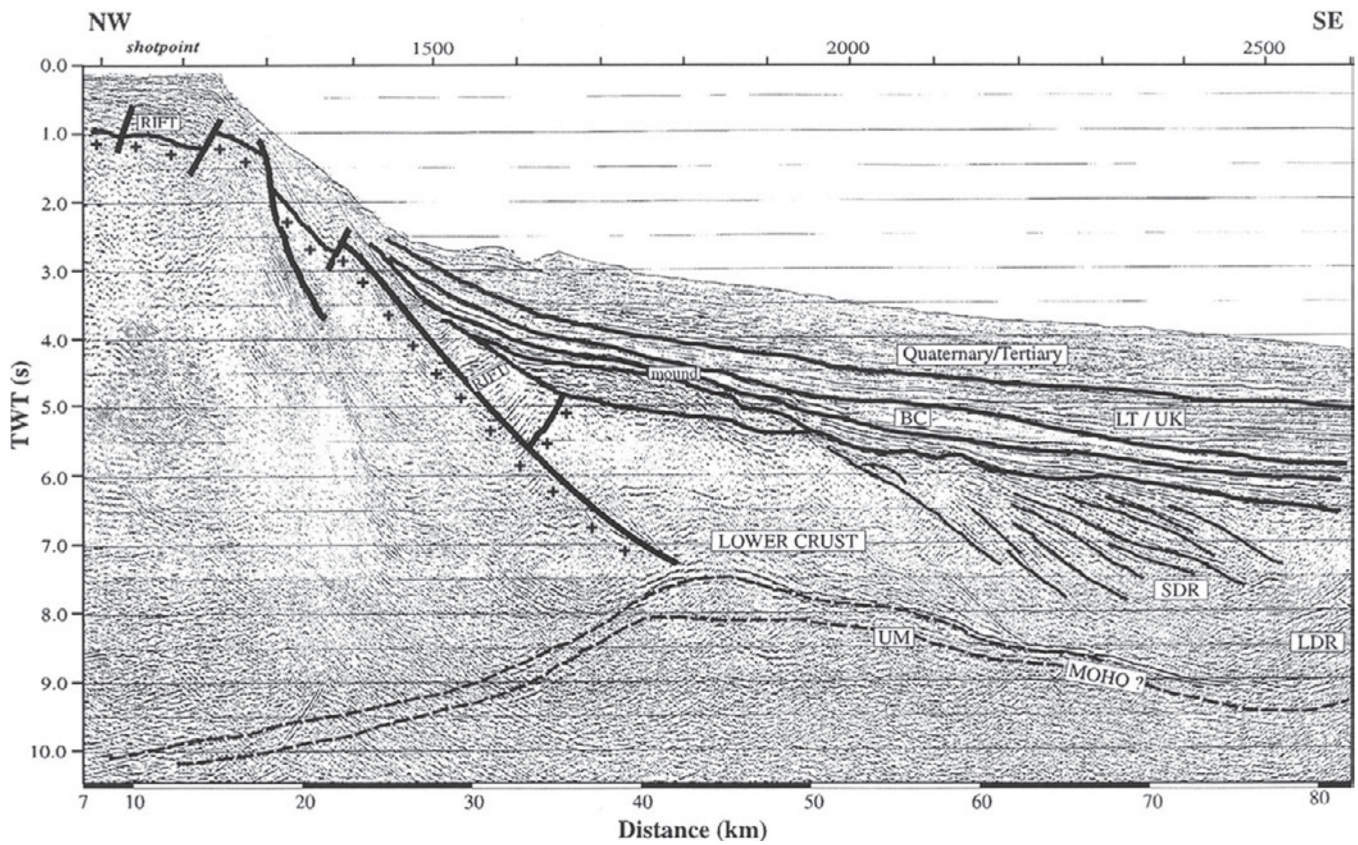



Figure 9.8.4-07. Deep seismic profile C in the Jacuipe basin on the northeastern Brazilian margin in the South Atlantic Ocean with interpretation and tectonic elements (from Mohriak et al., 1998, fig. 8). UM—upper mantle–lower crust transition; SDR—seaward dipping reflector wedge. [Tectonophysics, v. 288, p. 199–220. Copyright Elsevier.].

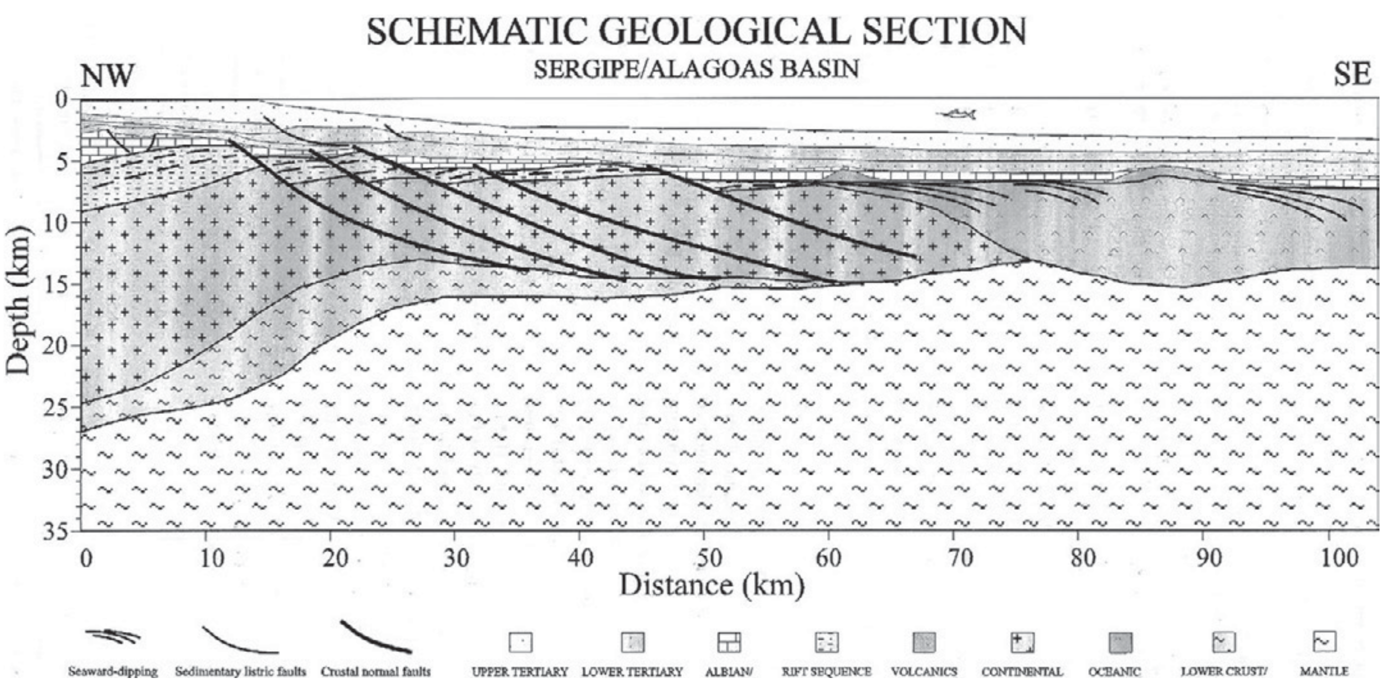



Figure 9.8.4-08. Schematic geological model of the Sergipe sub-basin on the Brazilian margin in the South Atlantic Ocean (from Mohriak et al., 1998, fig. 10). [Tectonophysics, v. 288, p. 199–220. Copyright Elsevier.].

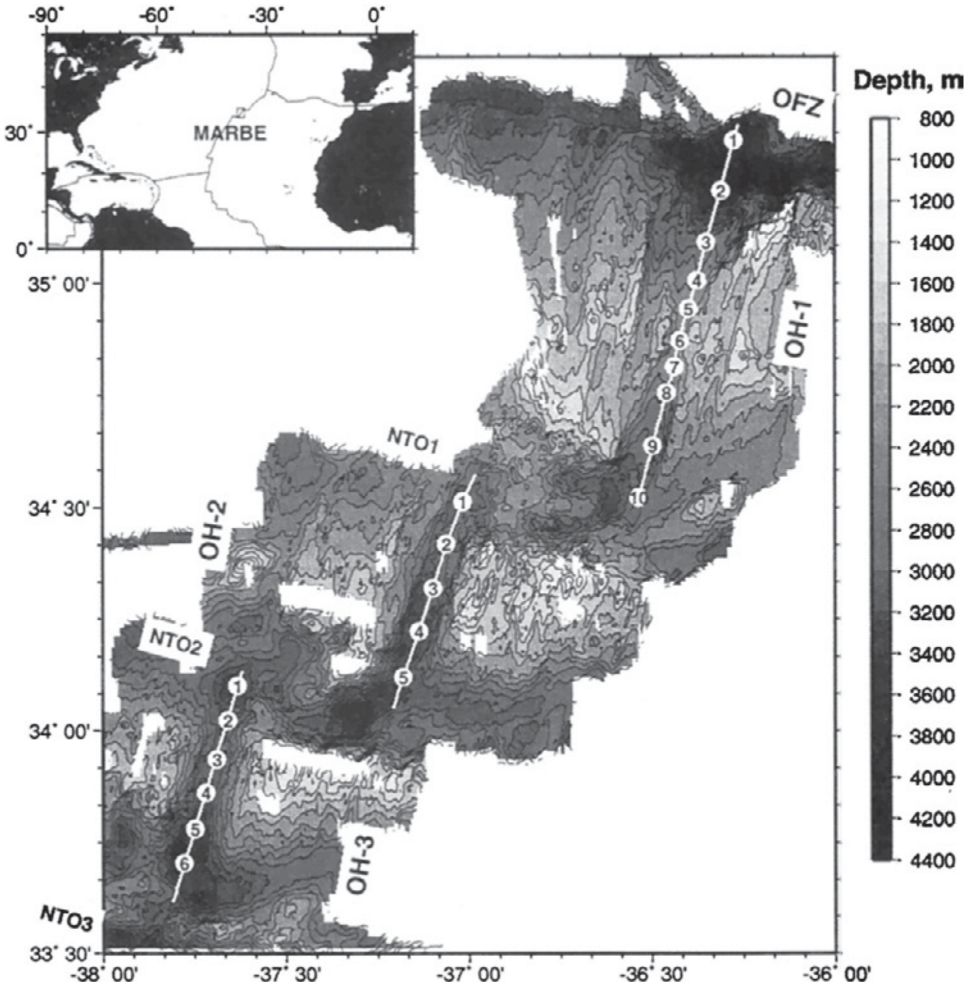



Figure 9.8.4-09. Bathymetric map along the Mid-Atlantic Ridge from 33°30′ N to 35°30′ N with the location of the three rift valley deep seismic profiles of the project MARBE (from Hooft et al., fig. 1). Lines—location of airgun shots; numbered circles—locations of ocean-bottom hydrophones; OH-1 to OH-3—spreading segments; NTO1 to NTO3—nontransform offsets; OFZ—oceanographer fracture zone. [Journal of Geophysical Research, v. 105, p. 8205–8226. Reproduced by permission of American Geophysical Union.]

average spacing of 9 km. They recorded 265 airgun shots with a 120 s firing rate, yielding an average shot spacing of 280 m (line C in Fig. 9.8.4-10).

Hosford et al. (2001) compiled P-wave velocity models for segment OH-1, both for the two rift mountain lines, imaging 2-m.y.-old crust, and the axial rift valley line. The seismic Moho along the lines was defined by the $P_MP$ reflection points (white stars in Fig. 9.8.4-11). The main results were summarized by Hosford et al. (2001) as follows: the mean upper crustal velocity (layer 2) increases by 16% from 4.5 km/s at the axis to 5.2 km/s at the 2-m.y.-old crust on each flank. The maximum crustal thickness is 8–9 km beneath the centers of all three lines and the crust thins by up to 5 km toward the ends of the segment OH-1. A mid-crustal high-velocity body was spatially associated with clusters of large seamounts in the western and eastern rift mountains (Fig. 9.8.4-11).

Rather little was evidently hitherto known about the internal velocity structure of ocean islands and of seamounts which have not reached the sea surface. Good constraints were obtained (Minshull, 2002) for only a few oceanic intraplate volcanoes, such as the Jasper Seamount in the northeast Pacific (Hammer et al., 1994) and the Great Meteor Seamount in the central Atlantic (Weigel and Grevemeyer, 1999).

Here the exploration of the Great Meteor shall be discussed in some detail. West from the triple junction of the Mid-Atlantic Ridge and the Azores-Gibraltar zone the Atlantis-Meteor seamount complex is located at 28°–30°W, 30°–34°N, a group of submarine volcanoes and guyots, ~700 km to the south of the Azores. As a continuation of the "Atlantische Kuppelfahrten" in the 1960s (Closs et al., 1968), the University of Hamburg, Germany, carried out a detailed geophysical survey in 1990, sampling the crustal structure of the four seamounts Meteor, Cruiser, Plato, and Atlantis. Weigel and Grevemeyer (1999) interpreted the data of the largest seamount, the Great Meteor. It had been discovered in 1938 and is a large-size, flat-topped guyot, rising from a depth of more than 4700 m to a crestal depth of 275 m. The top is covered by 200–900 m thick sediments which superimpose basement rocks with P-velocities of 5.9–6.0 km/s. Five OBHs were deployed along a 200 km line across the seamount. An airgun array of 76 l capacity was towed along the line and shot every 2 min. Two of the OBHs were positioned at the western and eastern flanks at 2930 m and 2700 m depth, respectively, two on the top

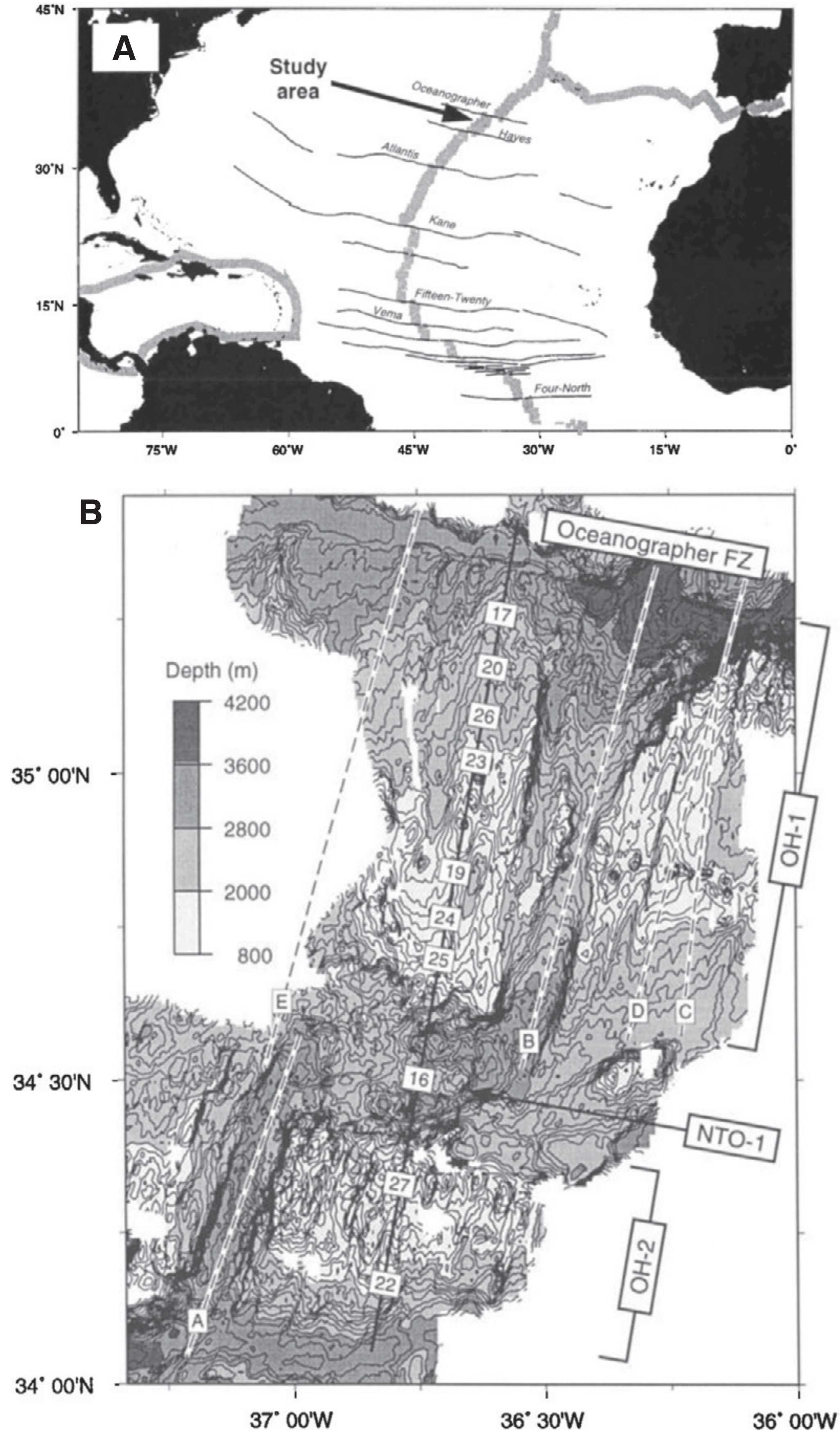



Figure 9.8.4-10. (A) Central part of the Atlantic Ocean showing the location of the Mid-Atlantic Ridge and other plate boundaries (shaded bands), the main Atlantic fracture zones (solid lines) and the location of the project MARBE (from Canales et al., 2000, fig. 2a). (B) Bathymetric map along the Mid-Atlantic Ridge from 34°00′ N to 35°30′ N with the location of all deep seismic profiles of the project MARBE along segment OH-1 (from Canales et al., 2000, fig. 2). C—eastern rift mountains line; B—rift valley line (corresponds to the seismic line along segment OH-1 in Figure 9.8.4-09); line with numbers—western rift mountains line (Canales et al., 2000; number denote locations of OBHs). E and D—earlier refraction lines of Sinha and Louden (1983, lines C1 and B); A—rift valley line along segment OH-2. [Journal of Geophysical Research, v. 105, p. 2699–2719. Reproduced by permission of American Geophysical Union.]

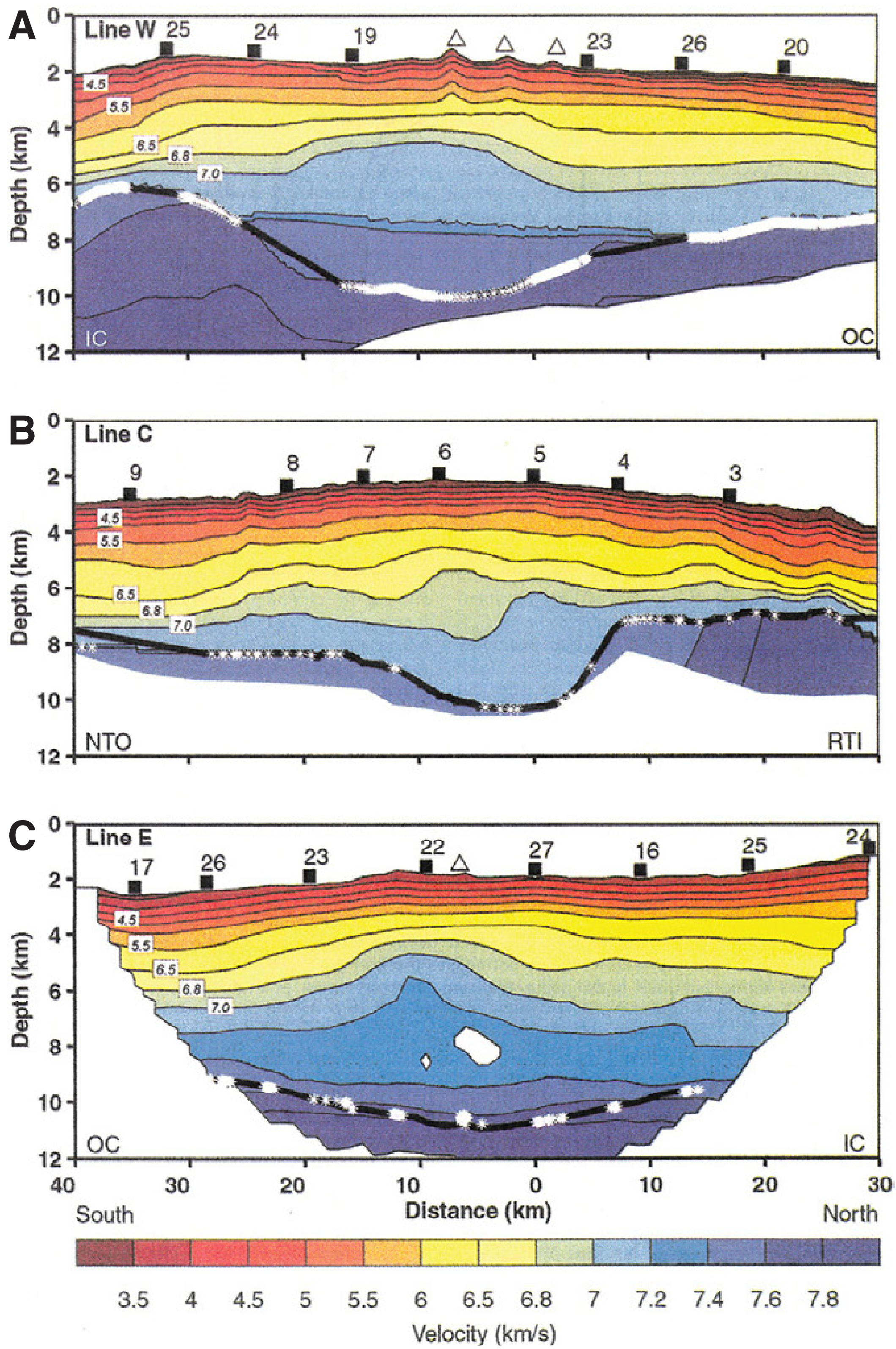



Figure 9.8.4-11. Final P-wave velocity models for segment OH-1 along the Mid-Atlantic Ridge from 34°15′ N to 35°30′ N (from Hosford et al., 2001, Plate 1). (A) Western rift mountains line W (= numbered line in Fig. 9.8.4-10). (B) Rift valley line C (= line B in Fig. 9.8.4-10). (C) Eastern rift mountains line E (= line C in Fig. 9.8.4-10). White stars—$P_MP$ reflection points; full squares—OBH locations; open triangles—seamount clusters. [Journal of Geophysical Research, v. 106, p. 13,269–13,285. Reproduced by permission of American Geophysical Union.]

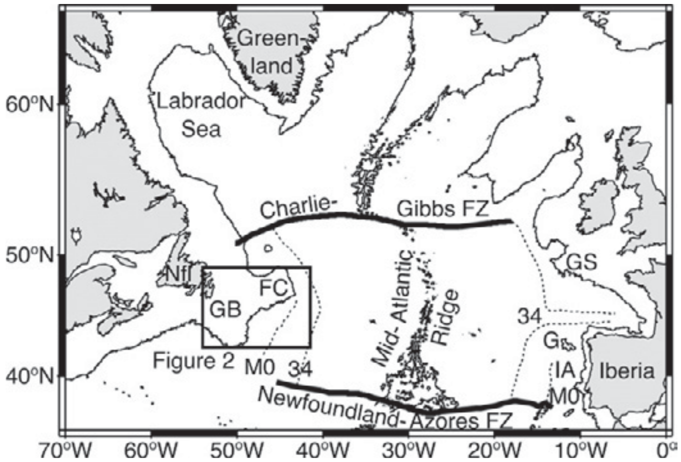



Figure 9.8.4-12. Location map of the mid–North Atlantic showing the Newfoundland, Iberia, and Irish margins (from Funck et al., 2003, fig. 1). Thick solid lines—fractures zones (FZ); thin solid line—2000-m bathymetric contour; dashed lines—M0 and 34 magnetic anomalies; FC—Flemish Cap; GB—Grand Banks; Nfl—Newfoundland; G—Galicia Bank; IA—Iberia Abyssal Plain; GS—Goban Spur. [Journal of Geophysical Research, v. 108 (B11), 2531, doi: 10.1029/2003JB002434. Reproduced by permission of American Geophysical Union.]

of the guyot at 300 and 440 m depth and the fifth OBH located 55 km off to the SE at 4700 m depth.

The interpretation was performed with a 2-D ray-tracing procedure. Starting from the 7-km-thick oceanic crust to the NW, with layer 2 at 5.0–8.5 km depth (velocities increasing gradually from top to bottom from 4.5 to 5.5 km/s) and layer 3 at 8.5–12.0 km depth (velocity of 6.5 km/s) crustal thickness increases underneath the edifice to a total thickness of ~17 km, with the top of the lower crustal layer upwarping to 2 km depth and the bottom downwarping to 17 km with a gradual velocity increase from 6.0 at the top to 7.5 km/s at the bottom. The upper-mantle velocity in the model was kept within 8.0 km/s. The upper crustal layer remains more or less constant in thickness, but its velocity is heavily decreased being upwarped along the flank of the sea mountain over the uprising and thickening lower crustal body (Weigel and Grevemeyer, 1999).

During the same expedition in April 1990 of the German vessel *Meteor* (1986), the Atlantis and Plato seamounts as well as the Great Meteor bank (27°–32°W, 28°–35°N) were targeted with 28 reflection seismic measurements and 40,000 shot positions deploying eight OBSs in two arrays of four along each line and using an airgun array (eight airguns deployed as two simul-

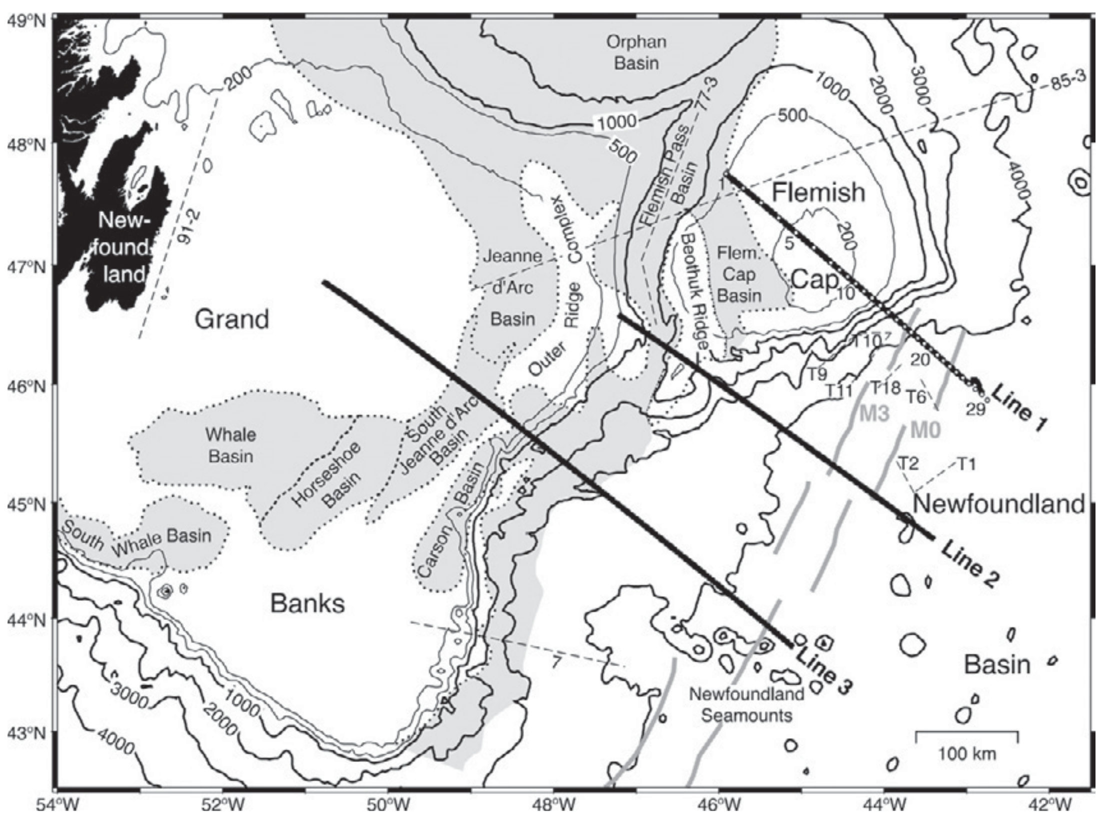



Figure 9.8.4-13. Bathymetric map of the Grand Banks and Newfoundland Basin east of Newfoundland, showing the location of the SCREECH experiment (from Funck et al., 2003, fig. 2). Thick solid lines—the three main lines; thin solid line—bathymetric contour lines; thin dashed lines—other seismic lines (e.g., for location of line 91-2 see also Figure 9.4.1-04); grey lines—M0 and M3 magnetic anomalies. [Journal of Geophysical Research, v. 108 (B11), 2531, doi: 10.1029/2003JB002434. Reproduced by permission of American Geophysical Union.]

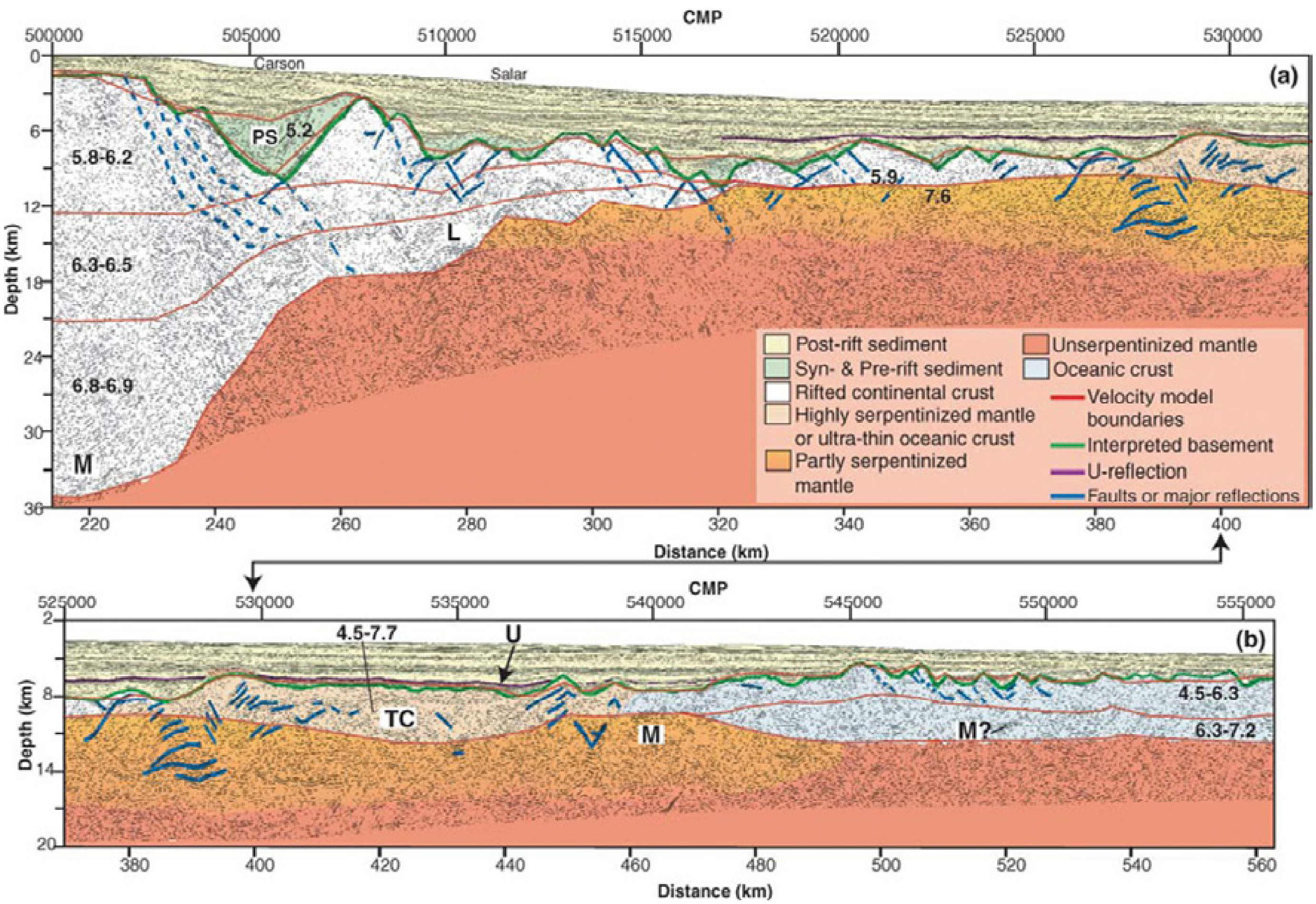



Figure 9.8.4-14. Velocity-depth model of SCREECH line 3: Comparison between reflectivity and the coincident velocity model derived from wide-angle data (from Lau et al., 2006a, fig. 8). L—landward-dipping reflections; TC—transitional crust; M—Moho reflections that coincide with the velocity model Moho; U—U-reflection; PS—package of pre-rift/syn-rift sediment within Carson basin. [Geophysical Journal International, v. 167, p. 157–170. Copyright John Wiley & Sons Ltd.]

taneously shot independent arrays of four airguns each with an airgun interval of 1 m at 5 m water depth) as energy source. Furthermore, along five refraction profiles 4000 shots were recorded from an airgun array of four 19-l-airguns. In total, 18 sites were occupied by five analogue OBSs, while for 11 sites, four digital OBSs were available (Wefer et al., 1991).

The Canary Islands, located at 14°–18°W, 28°–29°N to the west of the North African coast are an ocean island complex with anomalous structure of the oceanic lithosphere. Here a deep seismic experiment was carried out around the island Gran Canaria in 1993 during a cruise of the German research vessel *Meteor* (1986) (Schmincke and Rihm, 1994; Ye et al., 1999). Previous studies had been carried out on these islands in the 1960s and 1970s (e.g., Bosshard and MacFarlane, 1970; Banda et al., 1981a; see Chapter 7.8.4). The field work comprised three profiles radiating from Gran Canaria toward N, NNE, and NE. In addition, several transverse profiles and short profiles around the island were surveyed. Airguns were fired on each line at 1 or 2 min intervals and were recorded by eight onshore land stations and eight OBHs covering the N and NE sector off Gran Canaria up to a distance of 60 km from the coast. The interpretation showed that under the ocean basin adjacent to the volcanic edifice a 4-km-thick sedimentary sequence (average velocity 3.4 km/s) overlies the 7-km-thick igneous crust (velocity of 4.5 km/s). Beneath the island a pronounced lateral velocity variation takes place. Within the central volcanic edifice a 4–5-km-thick low-velocity zone was found, at roughly 4–12 km depth south of the island center. The massive volcanic island flank thins rapidly away from the island with velocities decreasing from 5.0 km/s near the coast to 3.5 km/s in the outermost part of the profiles, 50–60 km off the coast. A clear doming of lower crust material (>6.6 km/s) to 8–10 km depth was attributed to young mafic plutonic rocks. The Moho depth varies only slightly from 14 km north and northeast of Gran Canaria to 16 km depth under the center of the island (Ye et al., 1999).

In 1993, the RAMESSES study (Reykjanes Axial Melt Experiment: Structural Synthesis from Electromagnetics and Seismics) targeted an apparently magmatically active axial volcanic

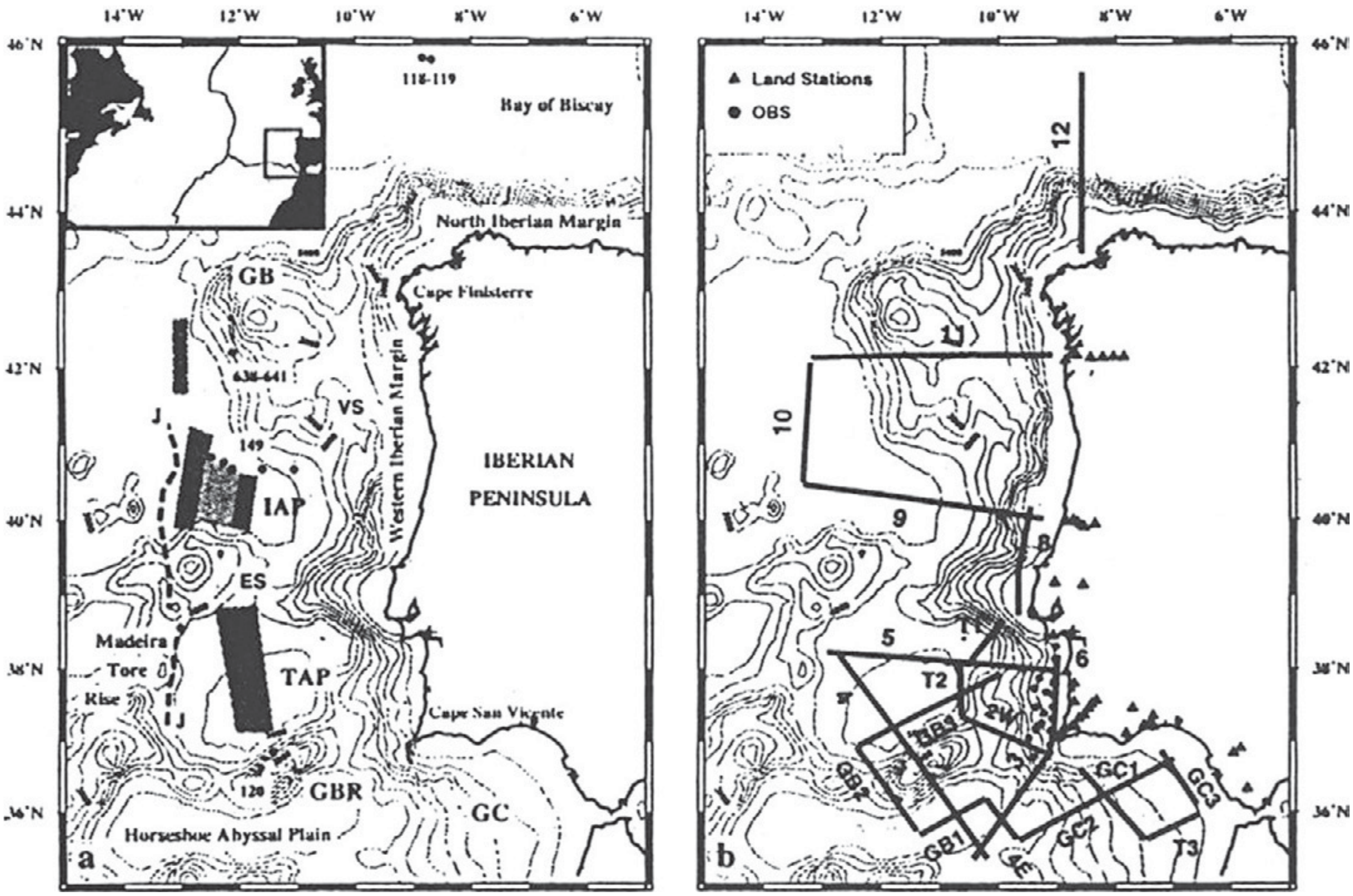



Figure 9.8.4-15. Bathymetry (500 m intervals) map of Iberian Atlantic margin (from: Banda et al., 1995, fig. 1). Left: Map of the IAM (Iberian Atlantic Margins European project) study area. Numbered solid circles—DSDP and ODP drill sites; ES—Extremadura Spur; GB—Galicia Bank; GBR—Gorringe Bank Region; GC—Gulf of Cadiz; IAP—Iberian Abyssal Plain; TAP—Tagus Abyssal Plain; VS—Vigo Seamount. Right: Location map of the IAM seismic reflection profiles. Triangles—land stations; circles—ocean-bottom seismometers. [Eos (Transactions, American Geophysical Union), v. 76, no. 3, p. 25, 28–29. Reproduced by permission of American Geophysical Union.]

ridge, centered on 57°45′N at the Reykjanes Ridge, with the aim of investigating the processes of crustal accretion at a slow-spreading mid-ocean ridge. As part of the project, airgun and explosive wide-angle seismic data were recorded by ten digital OBSs along profiles oriented both across- and along-axis. Coincident normal-incidence and other geophysical data were also collected (Navin et al., 1998). At 57°45′N, the Reykjanes Ridge appeared to have a crustal thickness of ~7.5 km on-axis. Slightly off-axis, the crust was modeled to decrease in thickness, implying off-axis extension. Modeling also indicated that the axial volcanic ridge was underlain by a thin (~100 m), narrow (~4 km) melt lens some 2.5 km beneath the seafloor, which overlay a broader zone of partial melt ~8 km in width.

Under the auspices of the British Mid-Ocean Ridge Initiative (BRIDGE) a multicomponent geophysical experiment was conducted on a magmatically active, axial volcanic ridge segment of the Reykjanes Ridge, centered on 57°43′N (Sinha et al., 1999). The site of the experiment was approximately located 840 km south of Reykjanes peninsula and over 1000 km off the center of the Iceland plume. At the site a clear median valley is present which extends for more than 100 km northward toward Iceland. The project involved wide-angle seismic profiling with OBSs, seismic-reflection profiling, controlled-source electromagnetic sounding and magnetotelluric sounding. The seismic part consisted of two lines where an array of eleven OBSs was deployed. The WNW-ESE line was 100 km long and line 2, 35 km long, ran perpendicular to line 1 along the axial volcanic ridge. From the seismic data it could be established that at distances of 10 km away from the axis, a normal Atlantic oceanic crust of 6–8 km thickness existed. There the sedimentary cover was thin (up to 200 m) and discontinuous and the seismic velocity at the top of the crystalline basement varied from 2.8–3.5 km/s. Following a steep velocity gradient at 2.0–2.5 km below the seafloor, a velocity of 6.5 km/s was reached, which depth was regarded as boundary between oceanic layers 2 and 3. With increasing depth the P-velocity increases more slowly, and just above the crust-mantle boundary at 6–8 km below the seabed a velocity of 7 km/s was seen. The uppermost mantle velocity was 7.9 km/s. The axial region had a quite different structure. The axial volcanic ridge lies within the shallow 30-km-wide and 800 m deep rift valley. Here, the P-velocities were between 1.0 and 1.5 km/s lower than the off-axis velocities at equivalent depths throughout most of layers 2 and 3. In agreement with the other data sets, the result was explained by the existence of an anomalous body, which was characterized by anomalously low P-velocity and low resistivity and was associated with a seismic reflector. This body was subsequently interpreted as melt body, i.e., a shallow crustal magma chamber could be identified beneath a slow spreading ridge segment.

In 1995, a 310-km-long seismic-refraction profile, ICEMELT, was recorded on land through central Iceland (Darbyshire et al., 1998). It traversed Iceland from the Skagi Peninsula on the north coast to the southeast coast, crossing central Iceland over the glacier Vatnajökull, below which the locus of the Iceland mantle plume is supposed to currently center. Up to 60 land-based instruments recorded each of six explosive shots, ranging

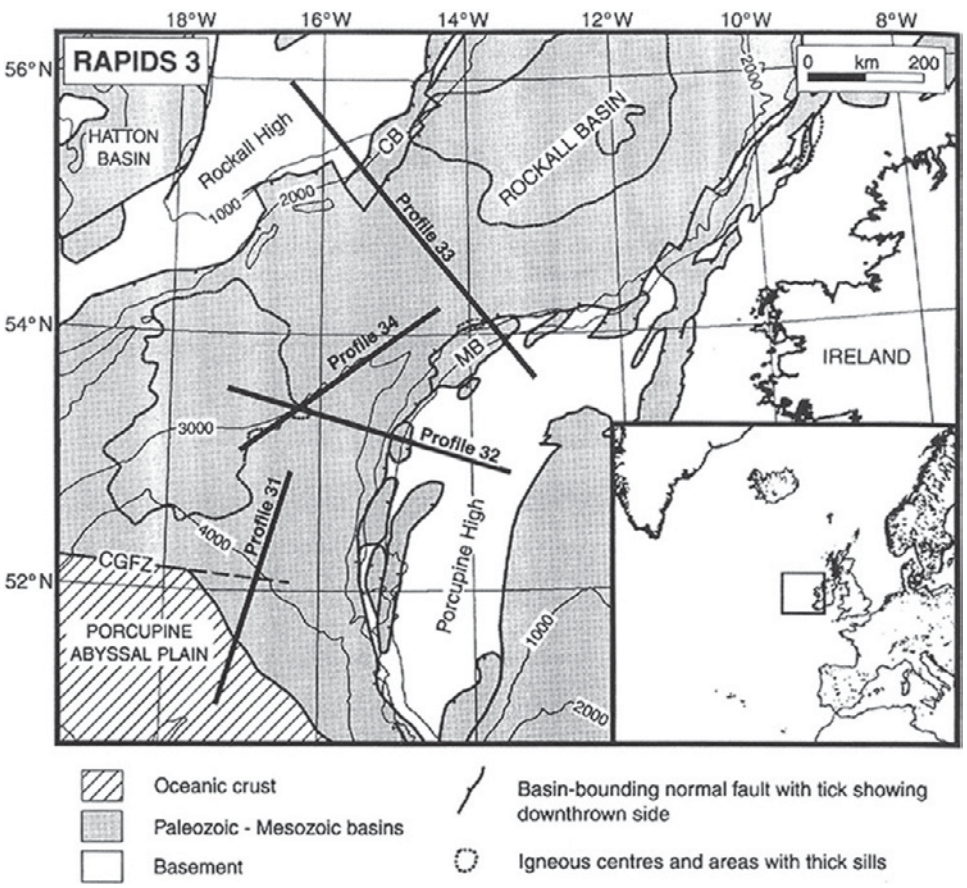



Figure 9.8.4-16. Location map of the RAPIDS 3 seismic profiles in the Hatton-Rockall area of the Northeast Atlantic margin (from Morewood et al., 2003, fig. 1). CGFZ—Charlie-Gibbs Fracture Zone; CB—Conall basin; MB—Macdara basin. [Eos (Transactions, American Geophysical Union), v. 84, no. 24, p. 225, 228. Reproduced by permission of American Geophysical Union.]

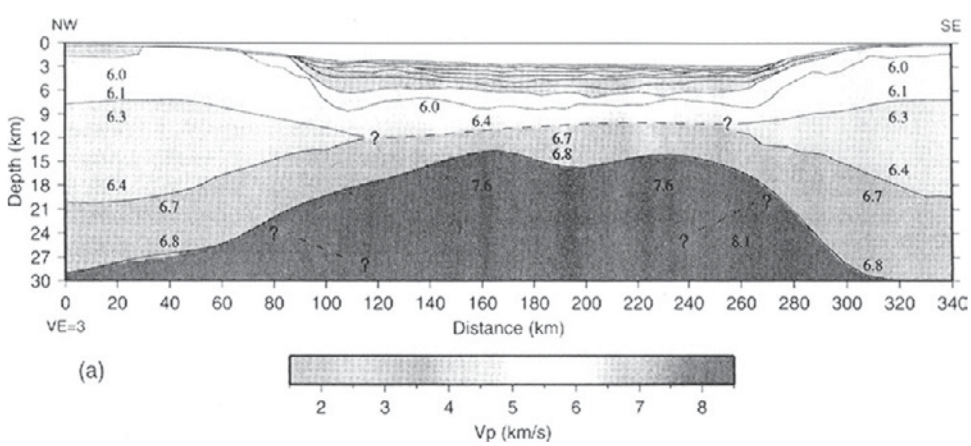



Figure 9.8.4-17. Preliminary P-wave velocity-depth model of RAPIDS 3 profile 33 across the Rockall basin (from Morewood et al., 2003, fig. 3a). [Eos (Transactions, American Geophysical Union), v. 84, no. 24, p. 225, 228. Reproduced by permission of American Geophysical Union.]

in size from 25 to 400 kg. Three shots were fired offshore (one in the north, two in the south) and three shots were in lakes in the central highlands of Iceland. Velocities in the upper crust ranged from 3.2 to 6.4 km/s. At both ends of the seismic line, the upper crust could be subdivided into two layers with a total thickness of 5–6 km. For the central highlands, only a single unit of upper crust was modeled, with seismic velocity increasing continuously with depth to almost 10 km below the surface. Below the central volcanoes of northern Vatnajökull, the upper crust was only 3 km thick. Below the base of the upper crust, velocities of 6.6–6.9 km/s were found and apparently increased with depth to 7.2 km/s. The seismic Moho was determined from the two large offshore shots at the profile ends. The resulting crustal thickness was 25 km at the north end and increased to 38–40 km beneath southern central Iceland. Mantle velocities could not be constrained (Darbyshire et al., 1998).

A second project (named B96), ~100 km farther east, was carried out in 1996 in northern Iceland (Menke et al., 1998). Along a north-south–striking array near the northern coast of 80 km length, three chemical underwater explosions were recorded, one

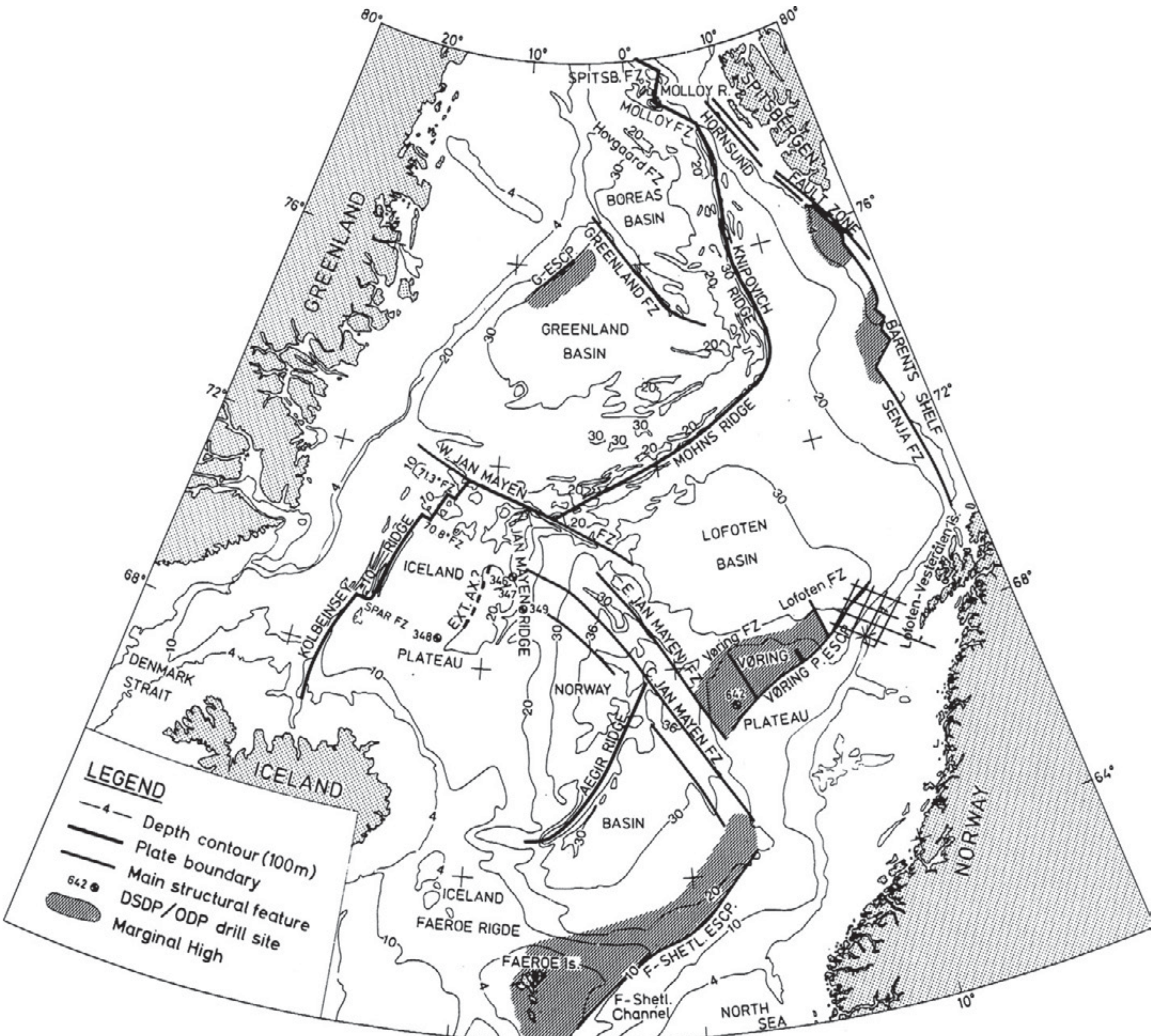



Figure 9.8.4-18. Regional physiography and structural basins and continental margins in the Norwegian-Greenland Sea. Also indicated are ODP drill sites and the 1988 survey off the Lofoten Islands, Norway (from Mjelde et al., 1992, fig. 2). [Tectonophysics, v. 212, p. 269–288. Copyright Elsevier.].

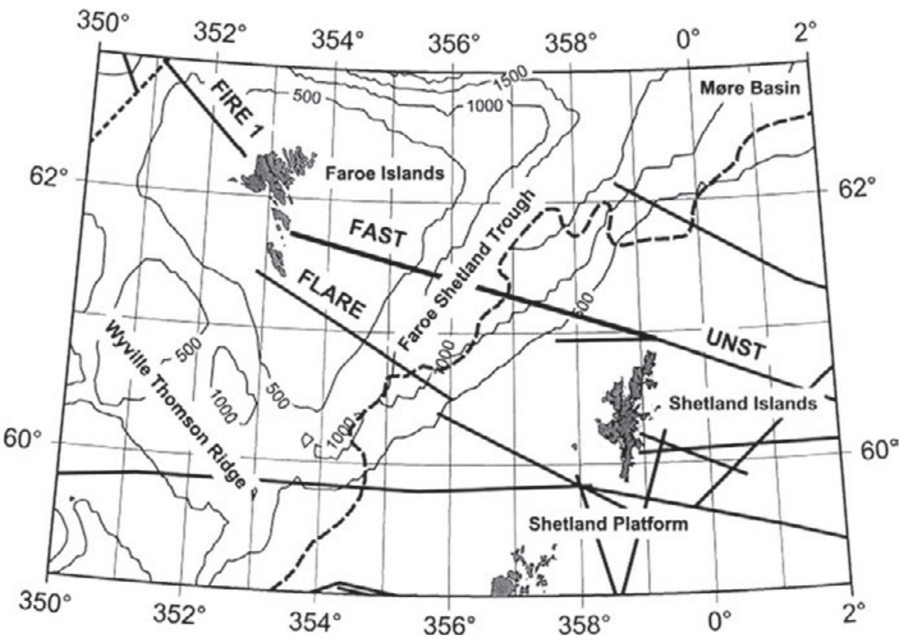



Figure 9.8.4-19. Location of FIRE 1, FAST-UNST and FLARE across the Faeroe-Shetland Trough (from England et al., 2005, fig. 1). [Journal of the Geological Society of London, v. 162, p. 661–673. Reproduced by permission of Geological Society Publishing House, London, U.K.].

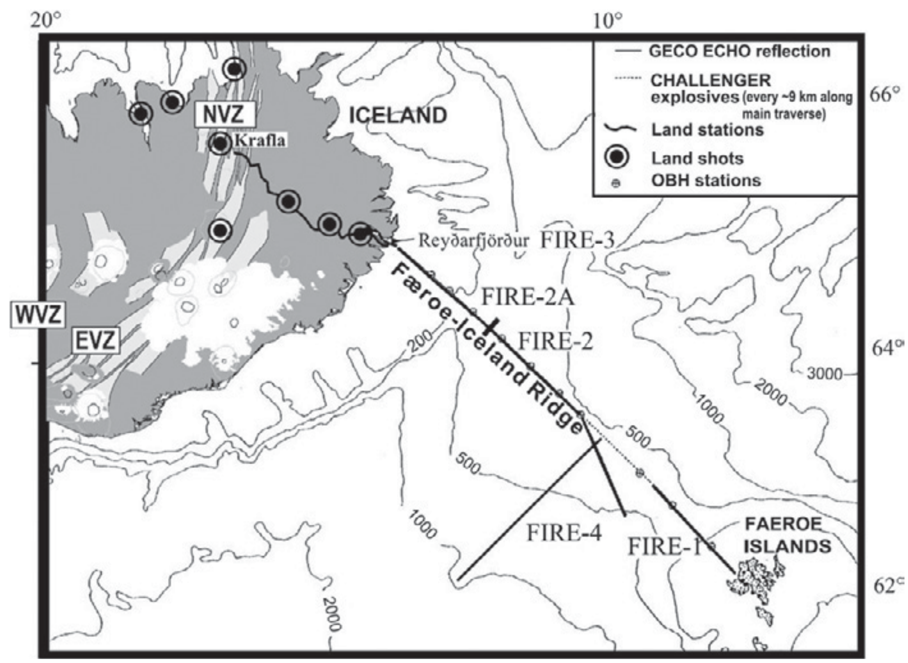



Figure 9.8.4-20. Location of FIRE, the Faeroe-Iceland Ridge experiment (from McBride et al., 2004, fig. 1). EVZ, WVZ, NVZ—eastern, western, northern volcanic zones of Iceland. [Tectonophysics, v. 388, p. 271–297. Copyright Elsevier.].

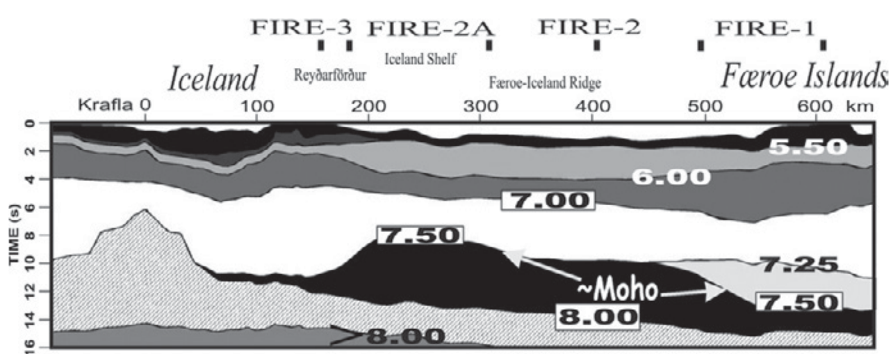



Figure 9.8.4-21. Velocity-depth model of the project FIRE between the Faeroe Islands and north-central Iceland (from McBride et al., 2004, fig. 2). [Tectonophysics, v. 388, p. 271–297. Copyright Elsevier.]

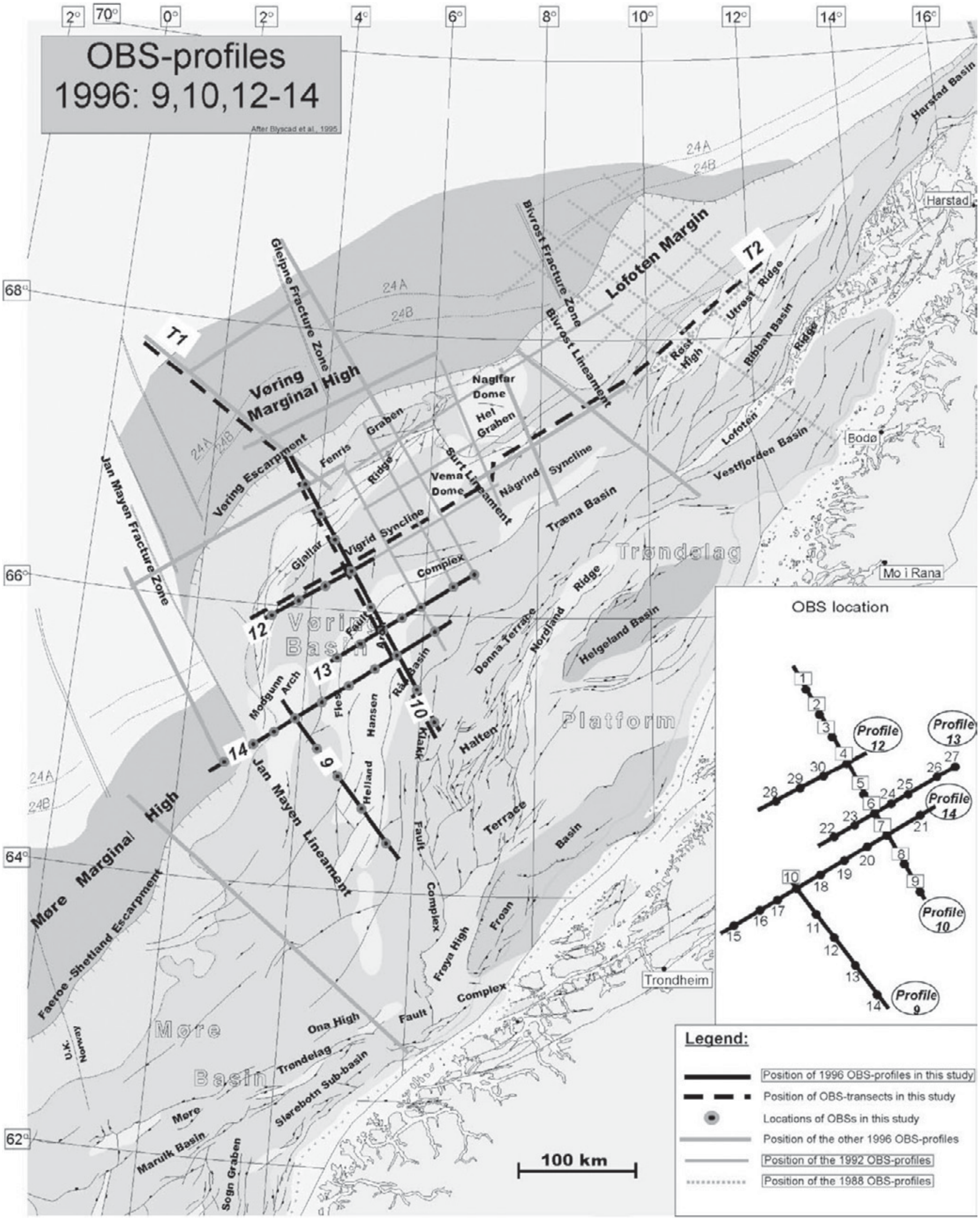



Figure 9.8.4-22. Location map of regional profiles at mid-Norwegian continental margin (from Raum et al., 2002, fig. 1). [Tectonophysics, v. 355, p. 99–126. Copyright Elsevier.].

shot at each end, and the third fan shot more than 100 km offline to the east. The array was operational for two months and also recorded a number of microearthquakes. The resulting crustal thickness was 25–31 km, with a southward thickening occurring in an abrupt step. An apparent rather high $P_n$ velocity of 8.0 km/s was observed from recordings of an earthquake in southern Iceland, which the authors explained to be consistent with a mantle lid, a layer of subsolidus mantle separating the Moho from a deeper partial melt zone.

### *9.8.4.3. North Atlantic Ocean*

The North Atlantic Ocean was the goal of many crustal structure investigations in the 1990s. We will start with the experiment SCREECH dealing with the margin areas off Canada (Fig. 9.8.4-12). Then we will jump to the other side of the Atlantic Ocean and continue with the project IAM (Iberian Atlantic Margin) and continue from there northwards to the microcontinents off Ireland and the margin areas off Norway. Finally, several projects will be discussed dealing with the northernmost Atlantic Ocean from the Shetland Island via the Faeroe Islands to Iceland and Greenland and finally close with an Arctic Transect.

The margin areas of the North Atlantic Ocean were investigated in much detail between 40° and 50°N latitude (Fig. 9.8.4-12). The Canadian margin in the west was the goal of a major experiment (SCREECH, Study of Continental Rifting and Extension on the Eastern Canadian Shelf) in 2000 across the Flemish Cap and into the deep basin east of Newfoundland (Funck et al., 2003; Hopper et al., 2006; Lau et al., 2006a, 2006b). In the east, major experiments (RAPIDS, Rockall and Porcupine Irish Deep Seismic) in 1988, 1990, and 1999 covered the margin off Ireland (Mackenzie et al., 2002; Morewood et al., 2003).

The SCREECH experiment, performed in 2000, consisted of three NW-SE–directed transects reaching into the Newfoundland Basin (Fig. 9.8.4-13). Line 1 was 320 km long and ran across the Flemish Cap, line 2 started at the Outer Ridge Complex, and line 3 covered parts of the Grand Banks (Funck et al., 2003; Hopper et al., 2006; Lau et al., 2006a, 2006b). The refraction/wide-angle reflection seismic data were obtained in a two-ship operation, one ship deploying OBSs and the other one towing the airgun array. In total, 29 OBSs were deployed, with 14 of them being equipped with three-component 4.5 Hz geophones and one hydrophone, and the other 15 having a hydrophone component only. On line 1 the station spacing varied from 21 km in the shallow water of Flemish Cap to 8 km in the Newfoundland Basin. Shot spacing was 200 m. For line 3, only 24 OBSs were available, of which 11 were equipped with three-component 4.5 Hz geophones. Along this line, the instrument spacing ranged from 10 to 40 km. While the OBSs were recovered by the first vessel, the second vessel shot the line a second time with a denser shot interval, mostly 50 m, to collect coincident multichannel seismic data by a 6-km streamer with 480 channels, resulting in 12.5-m group spacing. Furthermore, supplementary multichannel data were collected along additional parallel and perpendicular lines.

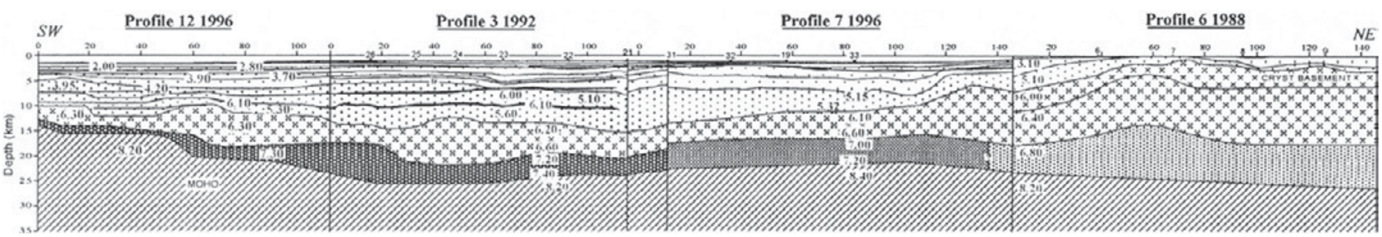



Figure 9.8.4-23. Crustal transect of the Voring margin off mid-Norway along the strike direction (from Raum et al., 2002, fig. 16). Explanation of symbols in Figure 9.8.4–24. [Tectonophysics, v. 355, p. 99–126. Copyright Elsevier.].

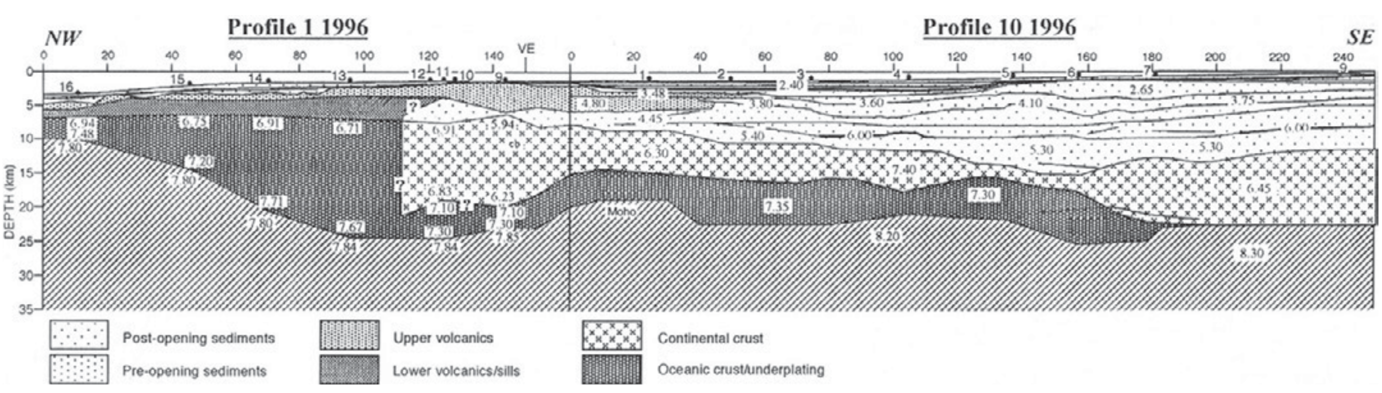



Figure 9.8.4-24. Crustal transect of the Voring margin off mid-Norway along the dip direction (from Raum et al., 2002, fig. 17). [Tectonophysics, v. 355, p. 99–126. Copyright Elsevier.].

The model for line 3 (Lau et al., 2006b) shows both the velocity-depth model, derived from the wide-angle data, and the coincident reflectivity pattern (Lau et al., 2006a, 2006b). The result (Fig. 9.8.4-14) was a 160–170-km-wide zone of continental rifted crust extending 120 km into the Newfoundland basin with abrupt thinning beneath the Carson basin from ~12–4 s TWT. It is separated from the oceanic crust in the east and southeast by a 70-km-wide transitional zone with a flat and unreflective basement in its central part (Lau et al., 2006b).

On the eastern side of the North Atlantic Ocean, many marine seismic projects were performed. They involved continental margin investigations off the west coast of Iberia, detailed crustal studies of the stretched continental crust west of Ireland, marine studies off the coast of Norway, and investigations of the northernmost areas around Spitsbergen, the Faeroes, Iceland and Greenland.

In 1993, the Iberian Atlantic Margins European project (IAM), a marine seismic project, focused on the Atlantic margin of Iberia (Fig. 9.8.4-15). Up to 3500 km of multichannel profiles were acquired in the North Iberian and Atlantic margins, the Gorringe bank and the Gulf of Cadiz (Banda et al., 1995). The shots were also recorded by OBS and by a variety of land stations. Three

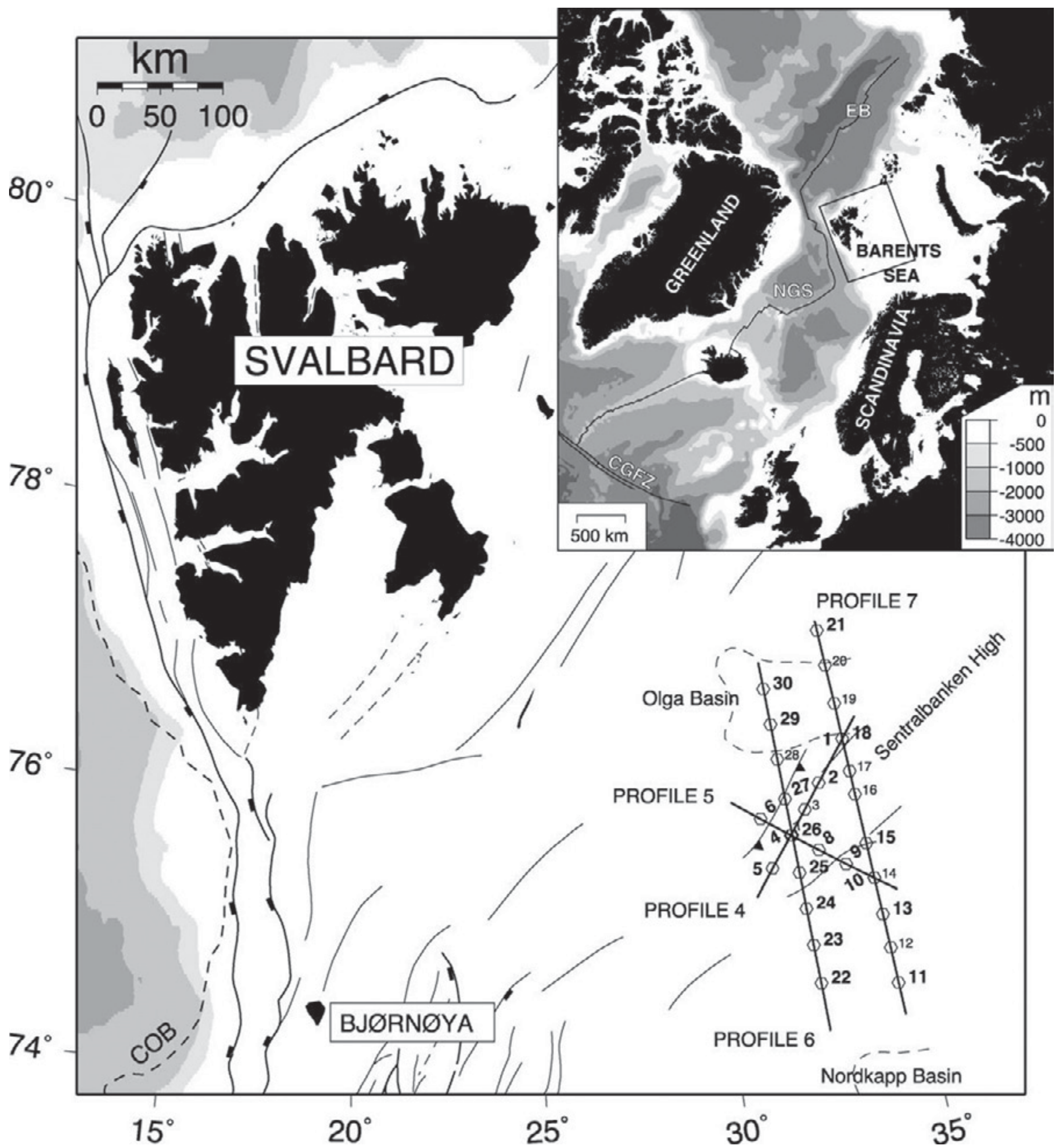



Figure 9.8.4-25. Location map of OBS profiles in the northwestern Barents Sea (from Breivik et al., 2002, fig. 1). COB—continent-ocean boundary west of Spitsbergen. [Tectonophysics, v. 355, p. 67–97. Copyright Elsevier.].

profiles, IAM 3, 9, and 11, were continued on land by short in-line profiles providing images of the continent-ocean transition.

The northernmost profile, IAM 12, was oriented in a N-S direction north of La Coruna (see also Fig. 9.2.3-01). IAM 11 ran in an E-W direction along latitude 42°N from the Galician coast across the Galicia margin, the E-W line IAM 9 ran approximately along latitude 41°N, and the N-S line IAM 10 surveyed the Iberian Abyssal Plain approximately along longitude 13.5°W. Lines IAM 8 and IAM 6 ran N-S parallel to the central and southern Portuguese coast and investigated the continental margin. The Tagus Abyssal Plain and the southwestern Iberian margin were surveyed by lines T1, 5, 2W, T2, and GB3. IAM 4 and some other lines investigated the Gorringe Bank region. IAM 3 ran NE-SW in front of Canbo San Vicente into the Horseshoe Abyssal Plain. Finally, four lines were recorded off southwestern Spain in the Gulf of Cadiz (GC1 to GC3 and T3).

The Gorringe Bank between the Tagus Abyssal Plain and the Gulf of Cadiz was of interest because of features resulting from the collision of the African and Eurasian plates including large-amplitude gravity anomalies. Line IAM 4 showed clear structural differences. To the north, a 1.5–2.0 s TWT thick sedimentary sequence overlay the oceanic crust of the Tagus Abyssal Plain; to the south at the Horseshoe Abyssal Plain the sediments are deformed by faulting and folding. Furthermore, a multicycle high-amplitude low-frequency reflector was located at ~8–9 s TWT. Along the IAM 9 seismic-reflection line in the southern Iberia Abyssal Plain between 40°N and 41°N, a wide transition zone was found, while the region of extended continental crust was relatively narrow. The transition zone along the IAM-9 profile was found to be a peridotite ridge, characterized by a thin upper layer, less than 3 km thick, and with velocities between 4.0 and 6.5 km/s with a high-velocity gradient. At depth followed a high-velocity layer with ~7.6 km/s with a low velocity gradient. The Moho was absent or very weak.

In the following years, additional marine experiments were implemented in Atlantic margins. In 1995, a wide-angle profile was shot along IAM-9 which included the deployment of up to 16 OBSs (Chian et al., 1999; Dean et al., 2000) with the aim to investigate the transition zone in more detail. The resulting velocity structure allowed the distinction of four zones. From west to east followed each other a 6.5–7-km-thick oceanic crust, an overlap of two peridotite ridges, transition zone, and thinned continental crust. The peridotite ridges, which cover a distance of ~50 km width, differ from the rest of the transition zone by a reduced top basement velocity and an elevated middle crust velocity. The transition zone is 170 km wide and has a velocity structure which is neither oceanic nor continental. The continental crust with velocities ranging from 5.5 to 6.8 km/s thickens from 7 km in the west across the lower continental slope over a distance of 80 km to 28 km in the east.

The stretched continental crust west of Ireland was the goal of continuing research. Following the RAPIDS project, shot in 1988 and 1990 (see, e.g., Fig. 8.9.4-05), in 1999 RAPIDS 3 followed (Fig. 9.8.4-16), involving the acquisition of four wide-angle seismic-reflection/-refraction lines within the Irish sector of the Rockall Basin area of the northeast Atlantic margin, offshore of the Republic of Ireland. It involved 1038 km of seismic profiling (Mackenzie et al., 2002; Morewood et al., 2003).

A total of 197 ocean-bottom seismometers were deployed and successfully recovered by the vessel RN *Akademik Boris Petrov*, operated by GeoPro GmbH of Hamburg. A 50-l airgun array was used as a seismic source. Firing every minute this gave a shot spacing of 120 m. A second vessel, the RN *Celtic Voyager*, operated by the Irish Marine Institute, was employed

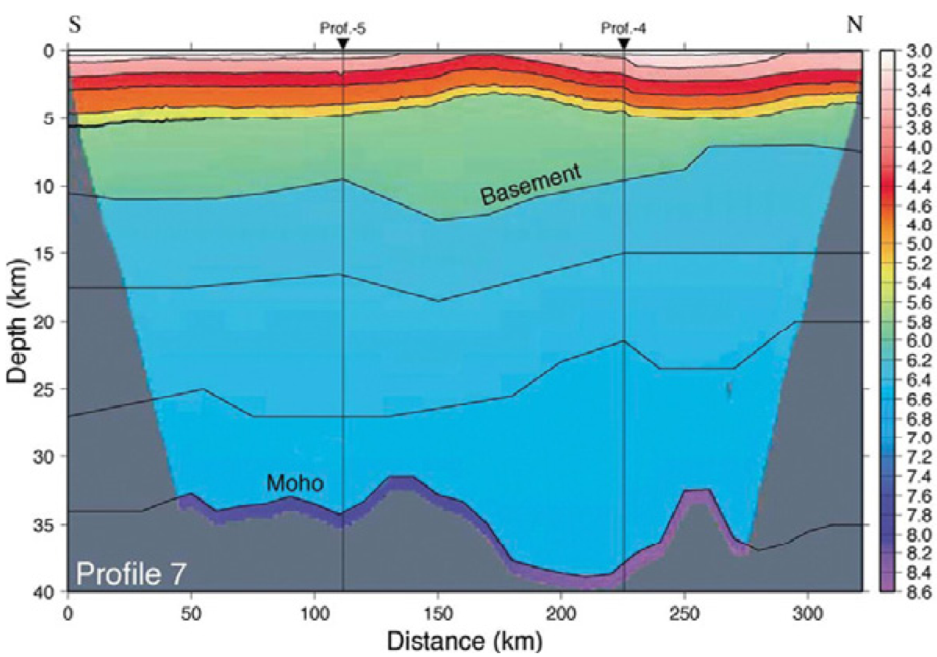



Figure 9.8.4-26. Model of OBS profile 7 in the northwestern Barents Sea (from Breivik et al., 2002, fig. 9b). [Tectonophysics, v. 355, p. 67–97. Copyright Elsevier.].

to fire 25-kg explosive shots as a secondary source. These were used as a backup to the airgun source, to ensure deep penetration of seismic energy into the crust and upper mantle. The quality of the airgun data was extremely good, with energy visible up to 140 km offset. Signal-to-noise ratios were high, and clear crustal and mantle phase arrivals were observed on the seismic sections.

Two profiles (32 and 33 in Fig. 9.8.4-16) were shot transverse to the trend of the Rockall Basin, thereby allowing the nature of the crustal thinning to be examined. Another profile (31) ran from the Rockall Basin and across the extension of the Charlie-Gibbs Fracture Zone. This profile was designed to allow examination of the crustal transition as the highly thinned continental crust of the Rockall Basin passes into the oceanic crust of the Porcupine Abyssal Plain. A fourth profile (34) extended approximately along the axis of the Rockall Basin.

The modeling resulted in a three-layer crust on the structural highs which flank the Rockall Basin. The crust appeared to be similar to that documented beneath onshore Ireland and the UK. A dramatic thinning of the crust was observed beneath the Rockall Basin. On profile 33 (Fig. 9.8.4-17), the crust thins from over 25 km beneath the Porcupine High to ~7 km beneath the Macdara Basin over a distance of ~60 km.

A similar thinning gradient was observed on profile 32. Crustal thinning at the northwestern end of profile 33 is more gradual, decreasing from over 25 km to ~6 km over a distance

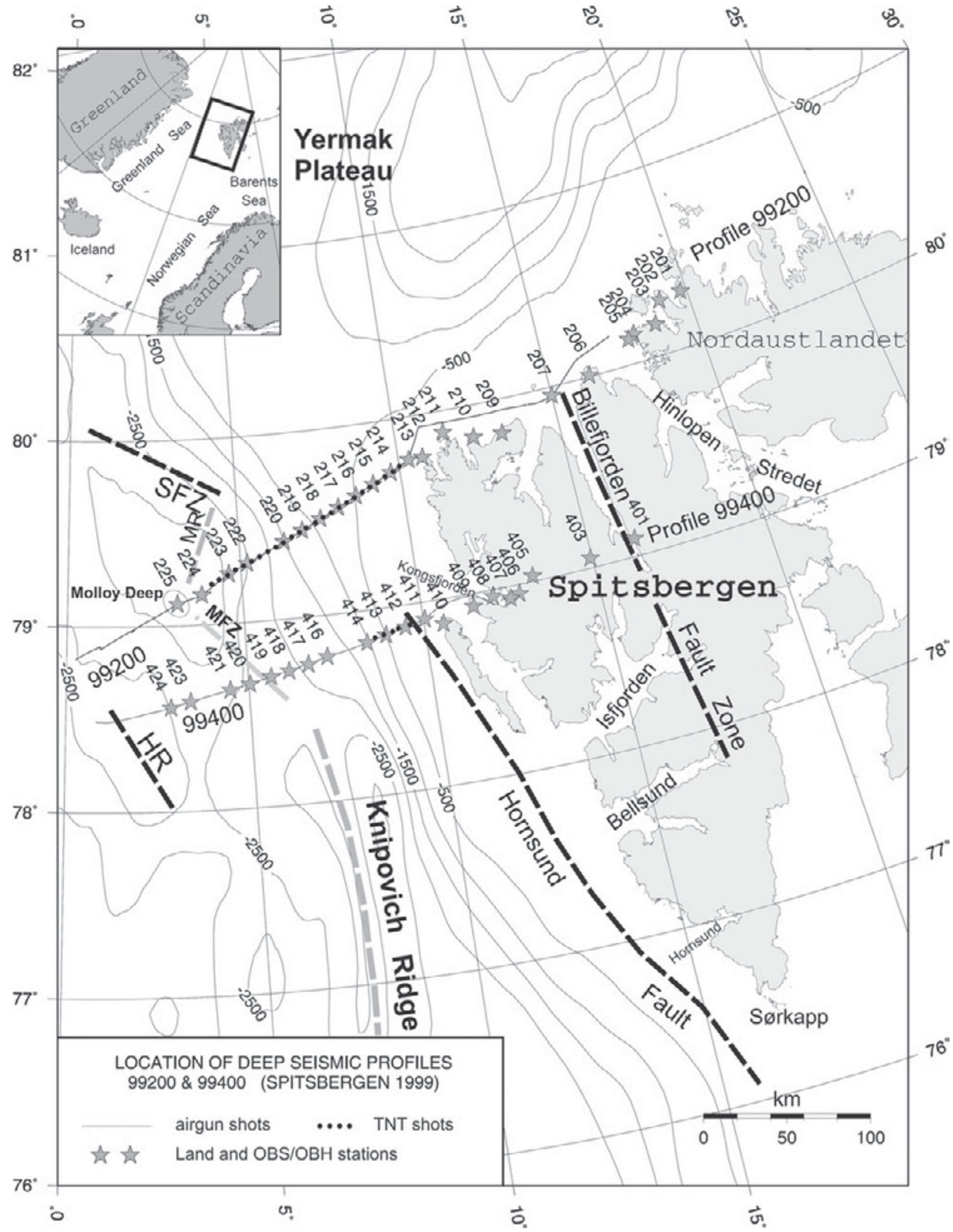



Figure 9.8.4-27. Location map of shot profiles and recording stations (stars: ocean-bottom seismometers and land stations) in the transition zone between the North Atlantic Ocean and Spitsbergen (from Czuba et al., 2004, fig. 1). [Polish Polar Research, v. 25, p. 205–221. Reproduced with permission of Polish Academy of Sciences, Warsaw, Poland.]

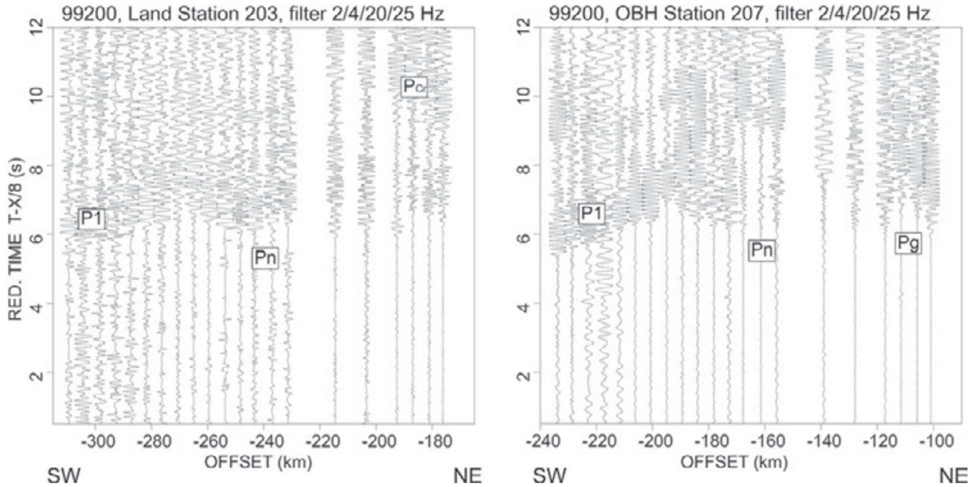



Figure 9.8.4-28. Examples of amplitude-normalized TNT seismic record sections (from Czuba et al., 2004, fig. 6) along line 99200, for stations RefTek 203 (left) and OBH 207 (right). Pg—first arrivals of crustal P-waves; Pcr—later crustal refracted P-waves; Pn—refracted P-waves beneath the Moho; P1—lower lithosphere reflections. [Polish Polar Research, v. 25, p. 205–221. Reproduced with permission of Polish Academy of Sciences, Warsaw, Poland]

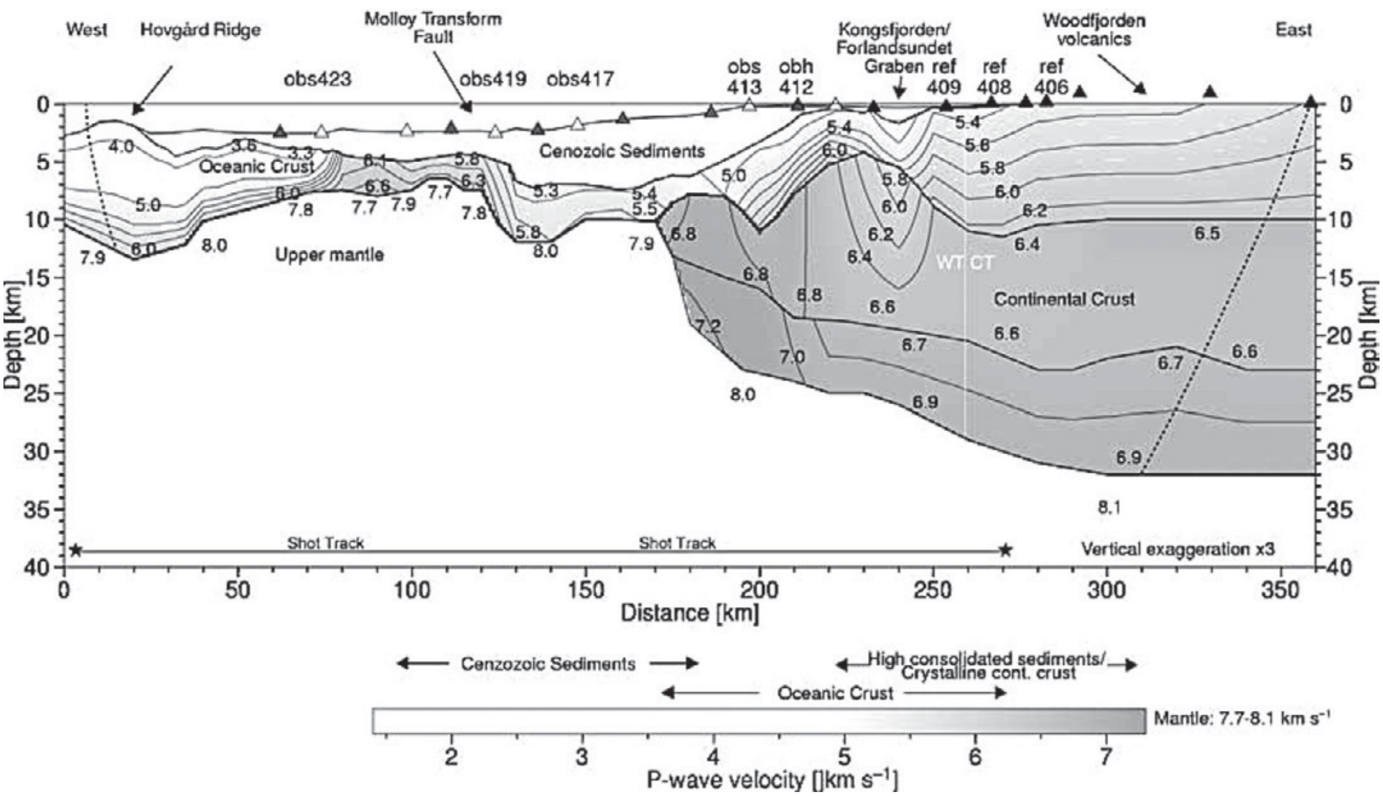



Figure 9.8.4-29. Crustal structure of the transition zone between the North Atlantic Ocean and Spitsbergen along line 99200 (from Ritzmann et al., 2004, fig. 7). [Geophysical Journal International, v. 157, p. 683–702. Copyright John Wiley & Sons Ltd.]

of ~120 km. Beneath the center of the Rockall Basin, a 6–7-km-thick sub-crustal layer was modeled, which thins toward the basin margins, with P-wave velocities lower than expected for normal upper mantle (Morewood et al., 2003).

South of the RAPIDS project, in 1997, data were obtained in the southern Porcupine Basin between 51° and 52°N. A 6-km-long, 240 channel digital streamer and a tuned sleeve-gun array were used with record length up to 9 s TWT (Reston et al., 2004). Five lines traversed the Porcupine Basin in an E-W direction; one line was shot along-strike in a N-S direction. Most prominent in the data was a bright reflection that was seen to cut down to the west from the base of the sedimentary section and in part to follow the top of a partially serpentinized mantle.

Various projects dealt with detailed seismic investigations of the northernmost part of the Atlantic Ocean, reaching from the Faeroe-Iceland area and the offshore area of Norway in the south to the Barents Sea, Spitsbergen and the offshore areas of Greenland in the north (Fig. 9.8.4-18).

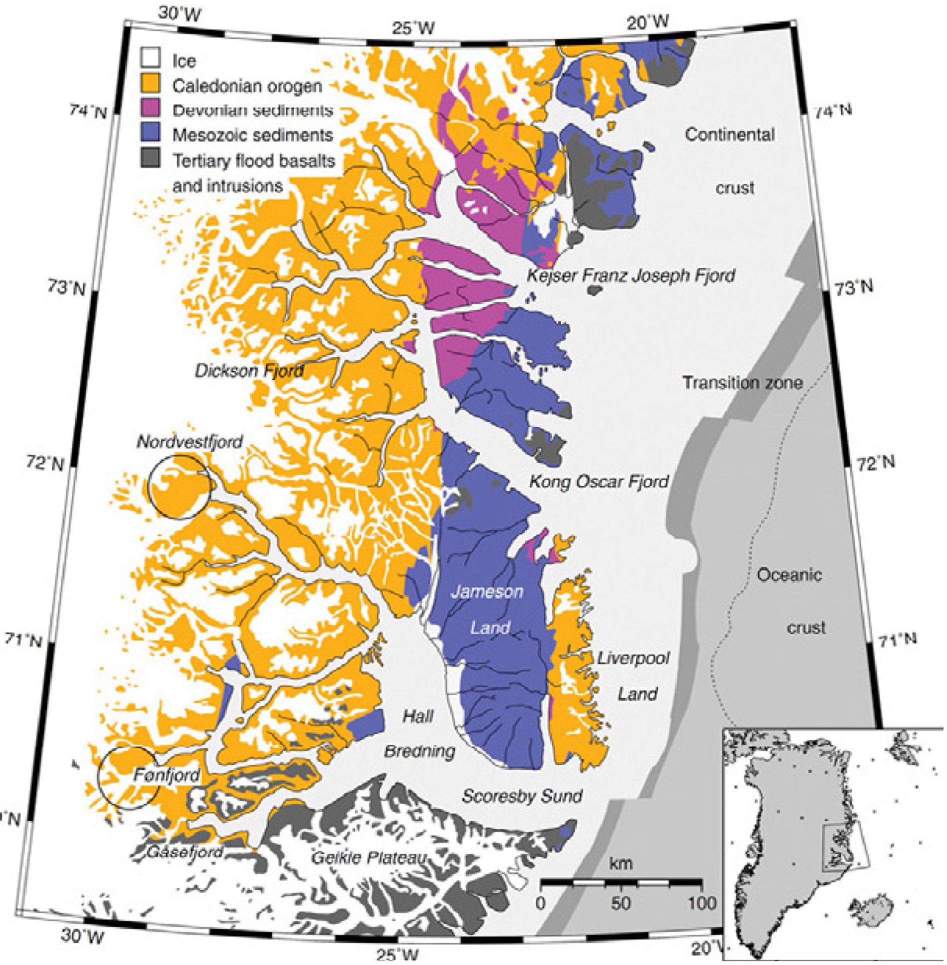



Figure 9.8.4-30. Simplified geological map of the East Greenland fjord region (from Schmidt-Aursch and Jokat, 2005, fig. 1). [Geophysical Journal International, v. 160, p. 736–752. Copyright John Wiley & Sons Ltd.]

Between the Faeroe Islands and the Shetland Islands, GeoPrakla shot several seismic-reflection lines FLARE and FAST-UNST across the Faeroe-Shetland Trough (Fig. 9.8.4-19) using the same acquisition configuration which had previously been used by BIRPS (England et al., 2005). The aim was to study the crust beneath the Faeroe basalts. Bright reflections were seen at 7–9 s beneath the basalts, dipping westward in the opposite direction than the dip of the basalts and reflections within the basalts. The sub-basalt reflections were interpreted to originate from near the top of the basement. The Moho was not imaged beneath the basalts. The authors explained this as being due to missing impedance contrast at the base of the crust. The main result was a thinning of the basement to ~10 km thickness beneath the center of the trough, which was explained as being caused by more than one extension period.

The Faeroe–Iceland Ridge to the northwest of the Faeroe Islands was investigated in 1994 by the project FIRE (Faeroe–Iceland Ridge Experiment), a combined onshore-offshore seismic-reflection and wide-angle acquisition program (Fig. 9.8.4-20) which extended from the Faeroe Islands continental block to the mid-Atlantic spreading center of Iceland (Smallwood et al., 1999; McBride et al., 2004).

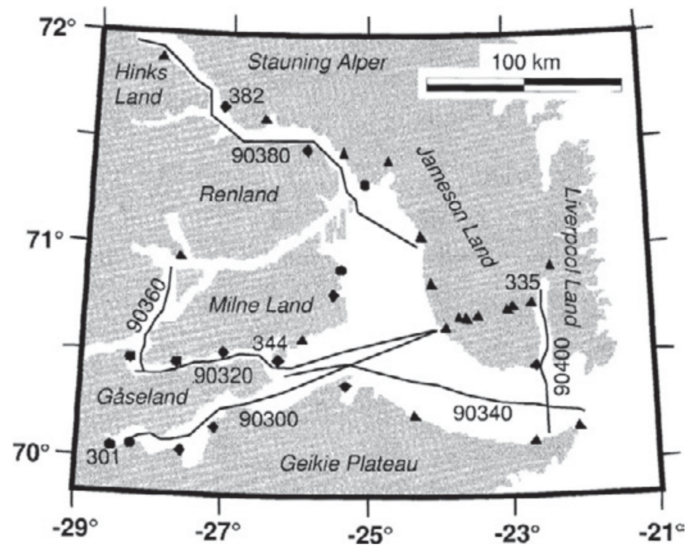



Figure 9.8.4-31. Location map of the 1990 seismic lines in the narrow fjords west of Scoresby Sund, to the east and south of Jameson Land, eastern Greenland (from Mandler and Jokat, 1998, fig. 2). [Geophysical Journal International, v. 135, p. 63–76. Copyright Wiley & Sons Ltd.]

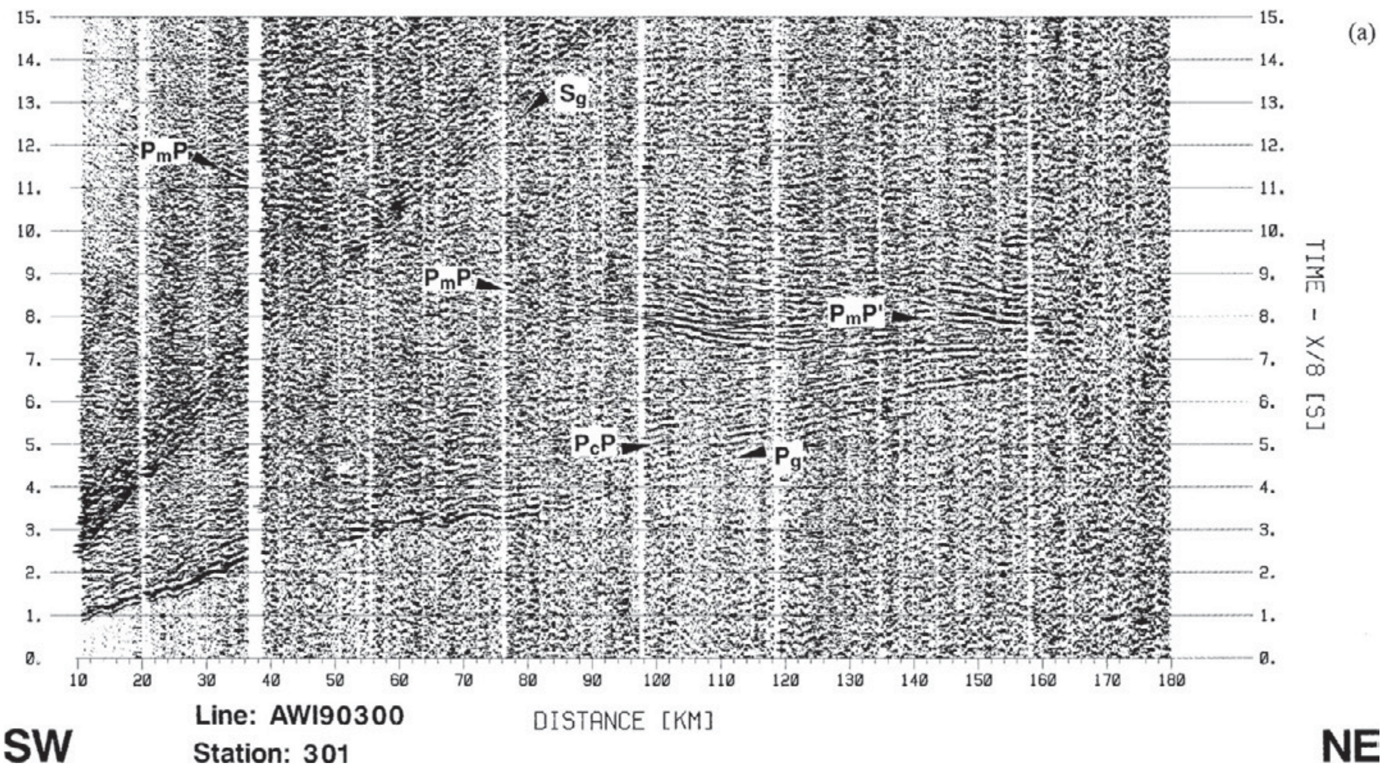



Figure 9.8.4-32. Seismic record section of station 301, line 90300, of the 1990 seismic survey in the Gaose fjord west of Scoresby Sund, eastern Greenland (from from Mandler and Jokat, 1998, fig. 3a). [Geophysical Journal International, v. 135, p. 63–76. Copyright John Wiley & Sons Ltd.]

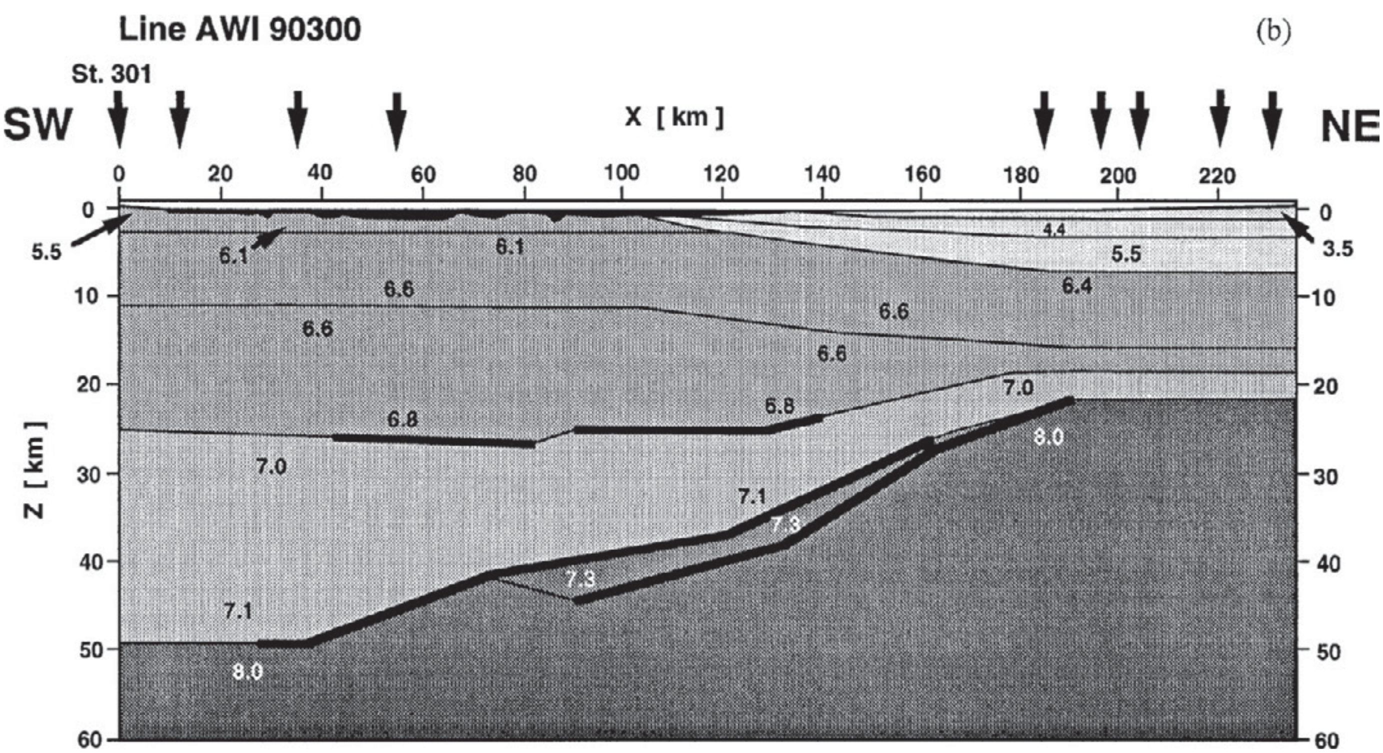



Figure 9.8.4-33. Crustal structure along line 90300 of the 1990 seismic survey in the Gaose fjord west of Scoresby Sund, eastern Greenland (from Mandler and Jokat, 1998, fig. 3b). [Geophysical Journal International, v. 135, p. 63–76. Copyright John Wiley & Sons Ltd.]

The project included a 440-km-long, conventional near-normal incidence multichannel seismic profile. The airgun shots were also recorded by land seismometers on both sides and by OBH. Over 6000 shots were fired at 75 m intervals. In addition, 55 explosive charges of 50 or 200 kg were detonated and recorded on the onshore seismometers, on the OBH and on a 6-km-long digital multichannel streamer with 240 receiver groups of hydrophones.

The land recording array on Iceland comprised 104 seismometers between the eastern coast and the axis of the present-day spreading center. The onshore array on Iceland also recorded explosive shots in lakes and fjords in Iceland. On the Faroe Islands seismometers were deployed at six sites.

The model presented in Figure 9.8.4-21 is based on interpretations of Richardson et al. (1998) for the Faeroes platform, Staples et al. (1997) for Iceland and the simultaneous modeling of the recordings from the FIRE shots along the arrays on Iceland and the Faeroe Islands and from the offshore OBH (Smallwood et al., 1999). The crust beneath the Faeroe Islands was estimated by McBride et al. (2004) to be ~46 km thick, and the Moho in Figure 9.8.4-21 was interpreted as the depth where velocity reached 7.5 km/s in the middle of a transition from 7.3 to 7.7 km/s.

From 1988 to 1996, a large number of seismic-refraction/wide-angle reflection profiles was recorded in the offshore area of central Norway off the Lofoten Islands (Figs. 9.8.4-18 and 9.8.4-22) by a consortium of Norwegian institutions, supported by Japanese scientists.

By introducing the OBS method, long-offset refraction studies were enabled, mapping the whole crust from upper sedimentary layers to Moho, because in small-offset reflection seismic surveys extrusive/intrusive rocks in the sedimentary layers caused seismic-imaging problems (Mjelde et al., 1992).

Following a first survey in 1988 (Mjelde et al., 1992), in 1992, an extensive OBS survey was performed in the central and northern part of the Voering Margin (Mjelde et al., 1997, fig. 1), comprising several experiments from regional to local-scaled acquisition. The success of the 1992 survey suggested a similar regional OBS survey to map larger areas of the margin.

In 1996, 15 profiles were acquired investigating the northern and southern part of the Voering Basin, seaward of the escarpment in the central part of the Voering Margin, as well as across the Jan Mayen Fracture Zone into the More Margin (Mjelde et al., 1998, 2001). Three profiles in the northern part of the Voering Basin were interpreted by Mjelde et al. (1998), while Raum

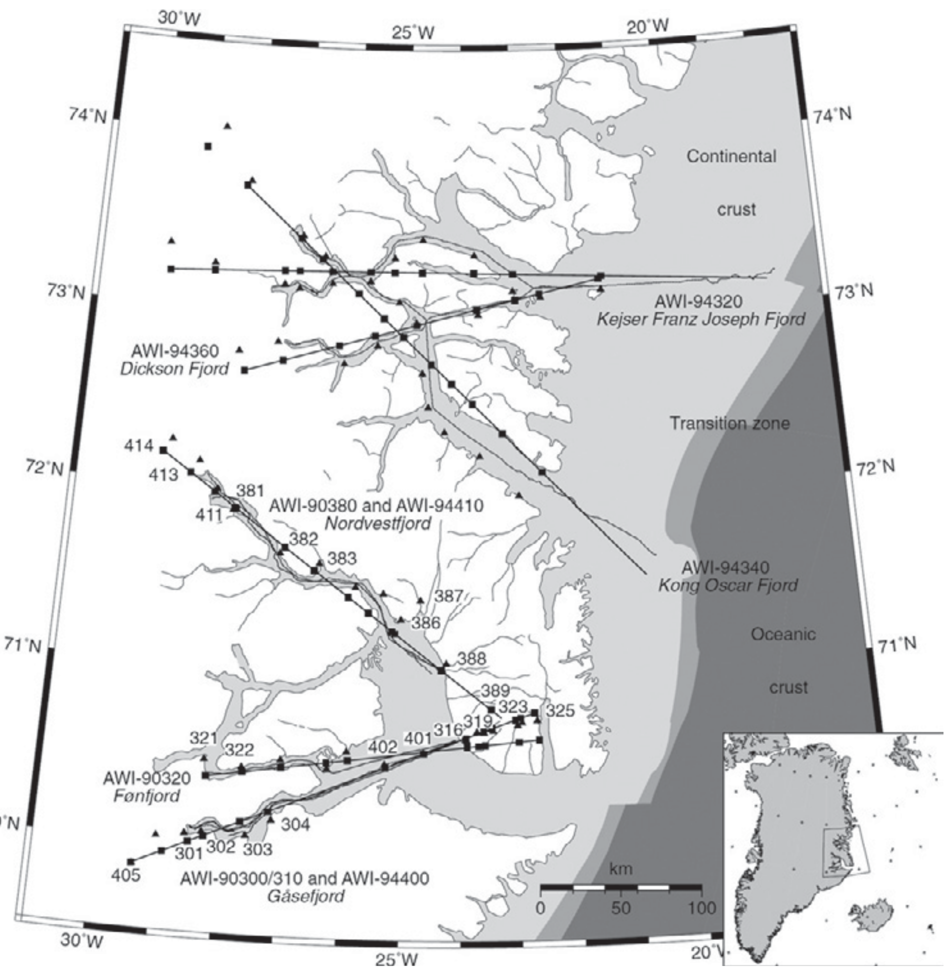



Figure 9.8.4-34. Location map of the 1990 and 1994 deep seismic lines in East Greenland (from Schmidt-Aursch and Jokat, 2005, fig. 2). [Geophysical Journal International, v. 160, p. 736–752. Copyright John Wiley & Sons Ltd.]

et al. (2002) modeled five profiles with a total length of 892 km in the southern part, which comprised the data of 30 recovered OBSs. The profiles were oriented in both strike and dip directions (lines 9, 10, 12–14 in Fig. 9.8.4-22).

The crystalline basement was modeled as a 6+ km/s refractor. Along one of the strike profiles, the transition from continental to oceanic crust was modeled where P-velocity increased from 6.2 to 7.3 km/s. Finally, Raum et al. (2002) compiled summary models of the 1988, 1992 and 1996 surveys along two transects, in both the strike and dip directions (Figs. 9.8.4-23 and 9.8.4-24).

The areas around Spitsbergen in the northernmost Atlantic Ocean had already seen several Polish-Norwegian activities in the past decades. The research was continued in the 1990s. In 1998, the northwestern Barents Sea southeast of Spitsbergen was the target of a wide-angle survey with OBSs (Breivik et al., 2002). Four lines were shot as a network (Fig. 9.8.4-25), with airgun shots every 200 m along each OBS profile. Two 160-km-long profiles extended in a SW-NE direction, and two 300-km-long profiles were recorded in N-S direction. The experiment was carried out by the vessel R/V *Hakon Mosby* of the University of Bergen.

The observed data enabled the identification and mapping of the depths to the crystalline basement and to the Moho by ray tracing and inversion. Refractions from the top of the basement together with reflections from the Moho constrained the basement P-velocity to increase from 6.3 to 6.6 km/s from the top to the bottom of the crust for which an average thickness of 35 km was obtained. On two profiles, the Moho deepened locally from near 33 to ~38 km depth in root structures (Fig. 9.8.4-26), which could be associated with high top-mantle velocities of 8.5 km/s (Breivik et al., 2002).

In 1999, the continent-ocean transition of northwestern Spitsbergen became again the target of a Polish expedition, ARKTIS VV/2 (Czuba et al., 2004). Seismic energy from airgun and TNT shots was recorded by onshore land stations, ocean-bottom seismometers and hydrophone systems on two profiles (Fig. 9.8.4-27). The northern profile 99200 was 430 km long and ran from the Molley Deep in the northern Atlantic to Nordaustlandet in northeastern Spitsbergen. The southern line 99400 was 360 km long and ran from the Hovgard Ridge to Billefjorden, Spitsbergen. Good quality records from the airgun shots were observed up to 200 km distance at land stations and 50 km at OBSs, while TNT shots reached to 300 km distance (Fig. 9.8.4-28 and Appendix A9-6-1).

The crustal structure of the transition zone proved to be very complex (Fig. 9.8.4-29), with Moho depths in the Atlantic Ocean as shallow as 6 km and Moho depths around 30 km under the center and the northern coast of Spitsbergen, thus confirming the result of the 1985 expedition (Czuba et al., 1999; Ritzmann et al., 2004).

Eastern Greenland and its margin on the western side of the northernmost Atlantic Ocean had been the goal of deep seismic investigations since the end of the 1980s (see Chapter 8; Mandler and Jokat, 1998). The research in the fjords of the Scoresby Sund (Fig. 9.8.4-30) was continued in 1990 by an expedition of the

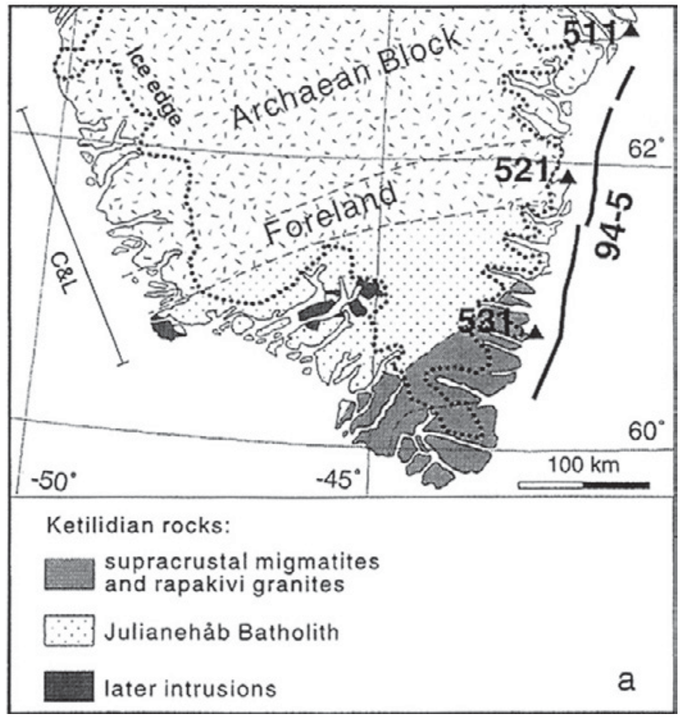



Figure 9.8.4-35. Location map of the offshore line 94-5 and three land-based recording stations (triangles) along the southeastern coast of Greenland (from Dahl-Jensen et al., 1998, fig. 1a). [Tectonophysics, v. 288, p. 191–198. Copyright Elsevier.].

Alfred-Wegener-Institute for Polar and Marine Research (AWI, Bremerhaven, Germany), offering the unique opportunity of applying marine seismic methods for the investigation of Greenland's continental crust up to 300 km west of the continent-ocean boundary. In total, more than 2300 km of seismic-refraction profile-km and 4000 km of normal-incidence reflection data were collected during the cruise. Mandler and Jokat (1998) discussed in particular data and results of profiles, obtained along the narrow western fjords crossing the southern part of East Greenland's Caledonian province (Fig. 9.8.4-31).

Here, the lines totalled ~1000 km and had maximum shot-receiver offsets of 230 km. The land-based recording stations were deployed by helicopter with between 5 and 10 stations per line, giving an average spacing of 25 km. The seismic source was provided by a single powerful airgun, shooting at an average spacing of 75 m. The record section shown in Figure 9.8.4-32 (see also Appendix A9-6-1) demonstrates the energy distribution and possible phase correlations.

In the western area, which is part of the Caledonian mountains of East Greenland, the seismic velocities of crystalline rocks increased continuously with depth from ~5.5 km/s at the surface to 6.6 km/s at 12 km. In the southwestern part of the region (28°W), the seismic measurements revealed high values of up to 48 km for the total thickness of the crust. Toward the east, the crustal thickness decreased rapidly, reaching a minimum of 22 km under the late Paleozoic–Mesozoic sedimentary basin of Jameson

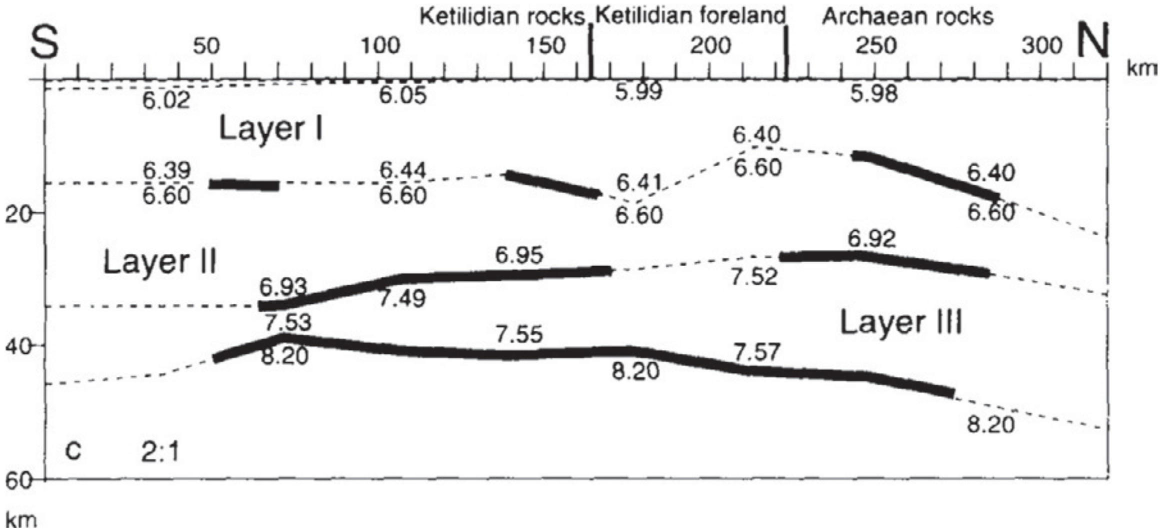



Figure 9.8.4-36. Crustal structure along the southeastern coast of Greenland (from Dahl-Jensen et al., 1998, fig. 3c). [Tectonophysics, v. 288, p. 191–198. Copyright Elsevier]

Land (Fig. 9.8.3-33). In the area of transition from thick to thin crust, the seismic data indicated a layered structure at the Moho (Mandler and Jokat, 1998).

The experiment of 1990 was continued in 1994 (Schmidt-Aursch and Jokat, 2005), re-shooting and adding to the East Greenland survey of 1990, and adding lines farther north around 73°N, some of which were recorded up to maximum distances of 325–375 km (Fig. 9.8.3-34).

In 1994, the Danish Lithosphere Center collected 1094 km of deep multichannel reflection seismic data along the SE Greenland coast south of 63°30′N, ~20 km offshore (Dahl-Jensen et al., 1998). The data were recorded on a 368-channel 4600-m-long streamer with a shotpoint interval of 75 m and 27 s recording time. Three land-based recording stations were set up along the coast and recorded wide-angle data during the shooting of line 94-5 (Fig. 9.8.4-35) and a margin-crossing line.

The data of the land station recordings were modeled by a three-layer crystalline crust, with layer III having high velocities of 7.4–7.5 km/s. The depth to Moho varies from 39 km in the south to 49 km in the north (Fig. 9.8.4-36).

In 1996, the R/V *Maurice Ewing* was used to conduct the SIGMA (Seismic Investigation of the Greenland Margin) experiment, a combined MCS reflection and wide-angle onshore/offshore seismic experiment across the southeast Greenland continental margin, organized jointly by the Woods Hole Oceanographic Institution (WHOI) and the Danish Lithosphere Center (Korenaga et al., 2000). Four transects were located at ~68°N, 66°N, 63°N and 59°N to systematically sample the structure of the margin with increasing distance from the Greenland-Iceland Ridge, the presumed Iceland hotspot track (Fig. 9.8.4-37).

Korenaga et al. (2000) focused their description on line 2, whose data comprised ~300 km of multichannel seismic-reflection data and coincident wide-angle seismic-reflection and

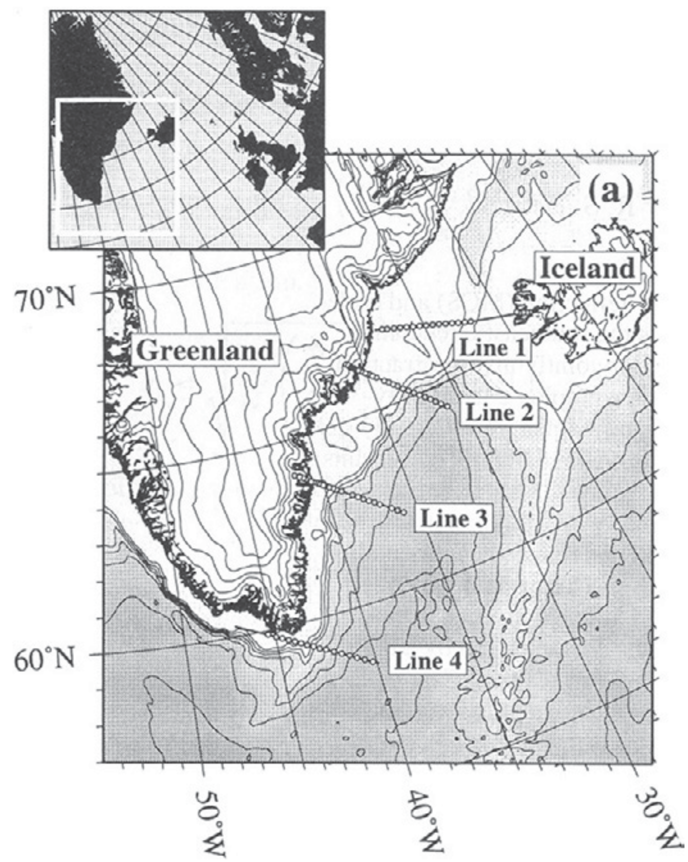



Figure 9.8.4-37. Location map of the 1996 SIGMA seismic experiment along the southeastern Greenland coast (from Korenaga et al., 2000, fig. 1a). Circles denote the locations of the onshore and offshore seismic instruments. Bathymetry is shown with 500-m contour interval. [Journal of Geophysical Research, v. 105, p. 21591–21,614. Reproduced by permission of American Geophysical Union.]

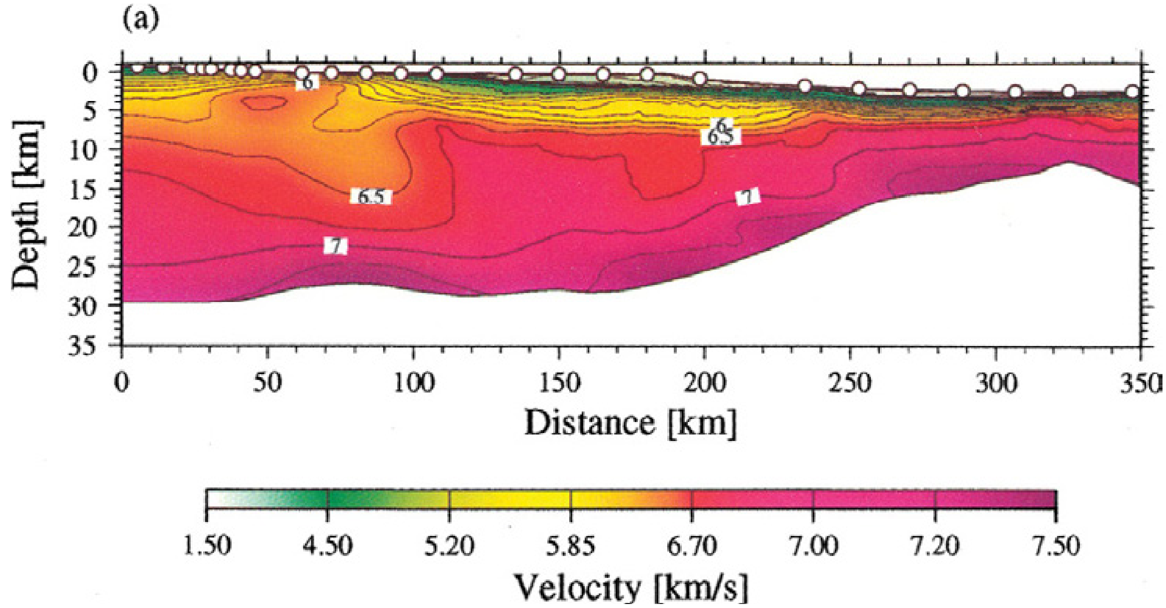



Figure 9.8.4-38. Model of transect 2 of the 1996 SIGMA seismic experiment along the southeastern Greenland coast (from Korenaga et al., 2000, plate 3a). [Journal of Geophysical Research, v. 105, p. 21591–21,614.] Reproduced by permission of American Geophysical Union.]

-refraction data recorded on 17 ocean-bottom and eight onshore seismometers. The multi-channel profiling was conducted using a 20-element, 8460 $in^3$ airgun array fired every 50 m (20 s) and a 4.0-km hydrophone streamer with 160 channels. Wide-angle seismic data were recorded on 10 WHOI ocean-bottom hydrophones, eight USGS ocean-bottom seismometers, and eight onland seismic recorders from PASSCAL, deployed along the transect. The total model distance of transect 2 was 350 km.

Most of the wide-angle data on transect 2 were of high quality. Airgun firing at a 20-s interval at 4.7 knots yielded densely spaced seismic traces at 50 m. The sampling interval was 10 ms, and a 5–14 Hz bandpass filter was applied to the data followed by predictive deconvolution and coherency weighting.

For the interpretation, a new seismic tomographic method was developed to jointly invert refraction and reflection traveltimes for a 3-D velocity structure. The resulting P-wave velocity-depth model for the 350-km-long transect (Fig. 9.8.4-38) showed strong lateral heterogeneity and included (1) a 30-km-thick, undeformed continental crust with a velocity of 6.0–7.0 km/s near the landward end, (2) a 30–15-km-thick igneous crust within a 150-km-wide continent-ocean transition zone, and (3) a 15–9-km-thick oceanic crust toward the seaward end. The thickness of the igneous upper crust varied from 6 km within the transition zone to 3 km seaward and was characterized by a high-velocity gradient. The bottom half of the lower crust generally had a velocity higher than 7.0 km/s, reaching a maximum of 7.2–7.5 km/s at the Moho.

A very sophisticated project covered the Mendeleev Ridge at 82°N along the Arctic-2000 transect in the Arctic Ocean between 164°W and 165°E (Lebedeva-Ivanova et al., 2006), a distance of 485 km (Fig. 9.8.4-39). It was the first ship-based Arctic expedition performing seismic investigations in the high Arctic Basin and was carried out by the ice-class vessel *Academian Fedorov*, accompanied through the pack-ice by the nuclear icebreaker *Russia*. The seismic acquisition included a comprehensive DSS survey along most of the line and a simple reflection profiling along the entire line. The experiment was carried out in three long segments, each 125 km long, and a fourth shorter segment in the west, 32 km long. The digital three-component seismometers were deployed by two helicopters along each segment and 7–8 shots per segment with charges ranging from 100 to 1000 kg were detonated in water at water depths of 70–100 m.

The shot spacing was 40 km. For each segment, four shots were within the deployment range, while the other shotpoints were offset by up to 75 km on each side. Data, as published by Lebedeva-Ivanova et al. (2006; Appendix 1-4), are shown in Appendix A9-6-1.

Simultaneously, a reflection survey was performed, using the same equipment and the same deployment, by detonating 0.4 kg shots at each recording site, i.e., at 5 km spacing. The measurements were single-point recordings of up to 15 s TWT. They were made while picking up the instruments and were designed to acquire data on the depth to the seafloor and the structure and thickness of the sedimentary cover in the uppermost crust at each recording site.

The velocity-depth structural cross section (Fig. 9.8.4-40) revealed seven layers. The uppermost three layers represent 3.5-km-thick sediments. The character of layer IV was debated as to whether it was to be interpreted as carbonate sediments or crystalline basement rocks. Layers V and VI constituted the

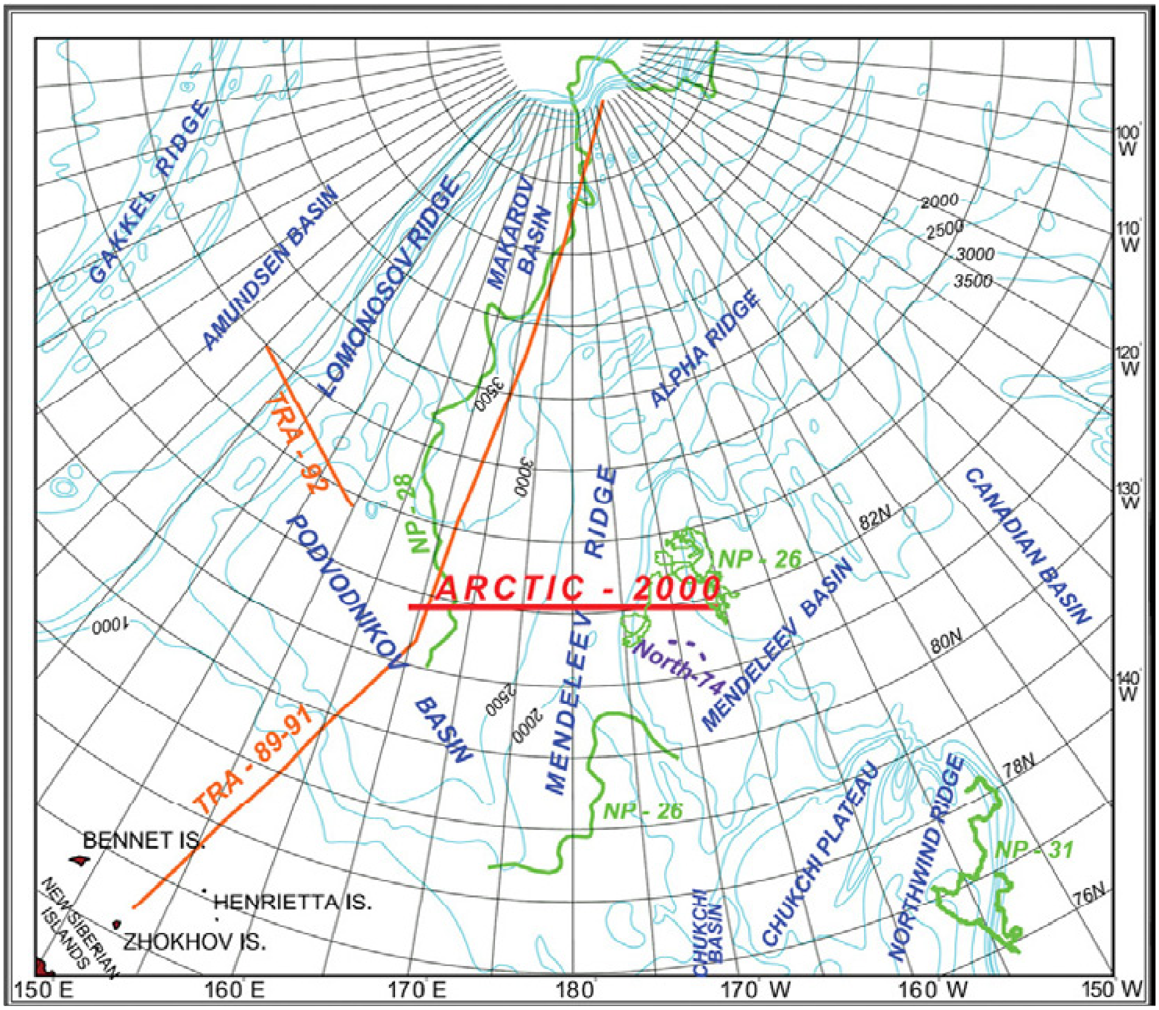



Figure 9.8.4-39. Location of the geotraverse "Arctic Transect-2000" and other combined deep seismic sounding and reflection geotraverses (TRA-92, TRA89-91) and reflection profiles (NP-26, NP-28, NP-31 and North-74) in the Arctic Sea (from Lebedeva-Ivanova et al., 2006, fig. 1). [Geophysical Journal International, v. 165, p. 527–544. Copyright John Wiley & Sons Ltd.]

crystalline crust, with the lower crust (layer VI) being 19–20 km thick. Layer VII with velocities of 7.4–7.8 km/s was interpreted as a crust-mantle mix, possibly resulting from underplating. The depth to the Moho along the Arctic Transect varied considerably, from 20 km under the Mendeleev Basin in the east and 13 km under the Podvodnikov Basin in the west to a maximum depth of 32 km in the center of the Arctic Transect. In conclusion, the authors favored the interpretation that the Mendeleev Ridge is composed of an attenuated underplated continental crust.

### 9.8.5. Summary of Seismic Observations in the Oceans in the 1990s

Most of the experiments we have described in more detail for the 1990s were largely devoted to the investigation of details of the mid-ocean rises, such as the East Pacific Rise. Here, for example, projects around the Galapagos hot spot and the Cocos-Nazca Spreading Center (e.g., (Sallares et al., 2003) and the Garrett Fracture Zone (Grevemeyer et al., 1998) were undertaken. A large number of Japanese marine surveys, some of which also had land components, targeted the trench system in the Philippine Sea and northwestern Pacific Ocean around Japan. In particular, the Nankei Trough and the Japan Trench south and east of Honshu was intensively investigated (e.g., Kodaira et al., 2000, Miura et al., 2003, 2005, Takahashi et al., 2004, Tsuru et al., 2002).

In the Indian Ocean, the Ninetyeast Ridge was investigated (Grevemeyer et al., 2001a). In the Atlantic, the Mid-Atlantic Ridge (e.g., Hooft et al., 2000) but also the continent-ocean transitions were major targets. In particular, in the northern Atlantic Ocean, projects targeted the microcontinents and intervening basins off Ireland (e.g., Mackenzie et al., 2002), the Norwegian

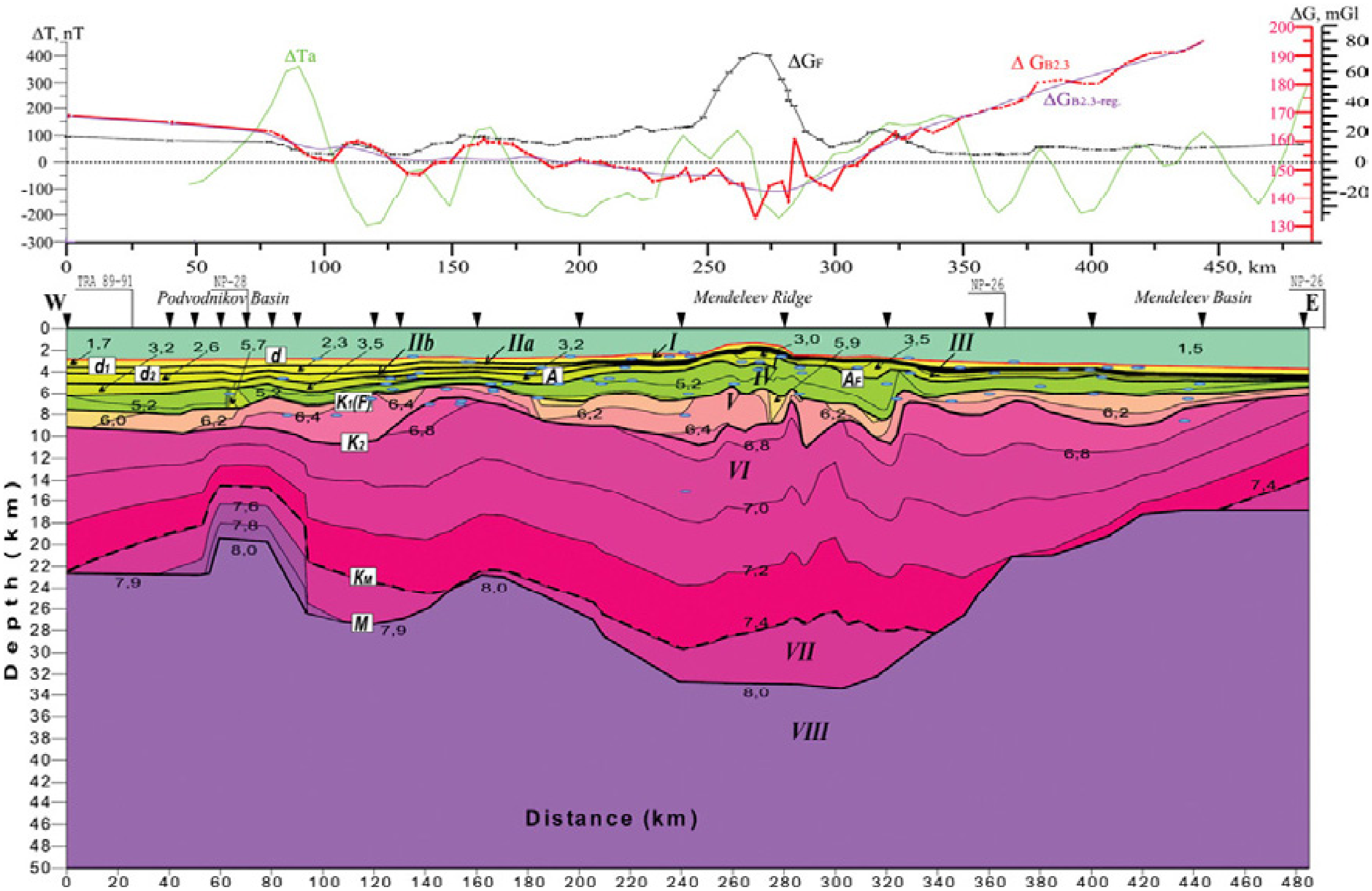


Figure 9.8.4-40. Crustal structure model along the Arctic Transect 2000 across the Mendeleev Ridge in the Arctic Ocean at 82°N (from Lebedeva-Ivanova et al., 2006, fig. 10). [Geophysical Journal International, v. 165, p. 527–544. Copyright John Wiley & Sons Ltd.]

margin (e.g., Raum et al., 2002), the Faeroe–Iceland Ridge (e.g., Smallwood et al., 1999), the northwestern Barents Sea southeast of Spitsbergen (Breivik et al., 2002), and eastern Greenland (e.g., Korenaga et al., 2000; Schmidt-Aursch and Jokat, 2005). The Arctic-2000 transect in the Arctic Ocean between 164°W and 165°E (Lebedeva-Ivanova et al., 2006) and the investigation of the North American margin off Canada (e.g., Funck et al., 2003) were other large marine projects.

The map of locations of experiments in the 1990s is based on data points of the USGS database which were published until 1999 in easily accessible journals and books (Fig. 9.8.5-01). It shows a concentration of projects on the Atlantic Ocean and the Polar Regions. The continental margins and adjacent deep-ocean parts of South America, South Africa, southern Australia, and New Zealand were covered by many new experiments. The map also indicates the activities along the East Pacific Rise and the Galapagos hot spot and special features in the southern Indian Ocean. The numerous marine projects of CROP MARE in the Mediterranean Sea, however, did not find their way into the database yet, as they were only published in connection with the CROP Atlas, which is little known.

## 9.9. ADVANCES IN INTERPRETATION METHODOLOGY

As mentioned at the beginning of this chapter, the early 1990s saw the gradual replacement of analogue recorders by digital recording instrumentation and the establishment of instrument pools, which allowed organizations to buy and to service large quantities of individual recording sets, and from which individual groups of scientists and/or institutions could loan any number. At a later stage, these instrument pools also started to request copies of the recordings and to archive large amounts of data for future use.

At sea, powerful airgun systems, replacing in many projects the destructive underwater explosions, as well as reliable ocean-bottom seismometer/hydrophone equipment and large-scale digital streamers of several kilometers length were developed and widely used in the following decade, supplying a vast amount of data which allows studying details of the oceanic crust and upper mantle structure.

Since large-scale international projects often needed more equipment than one instrument pool could provide, the equip-

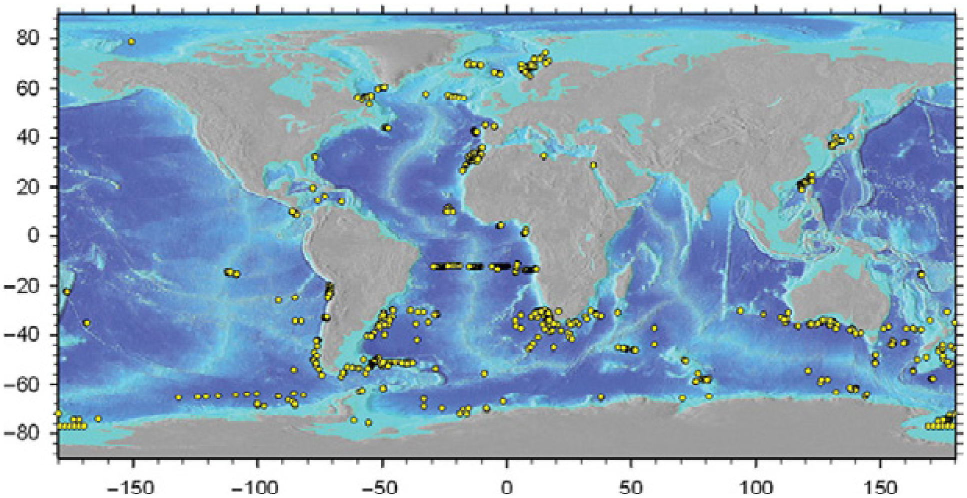



Figure 9.8.5-01. Seismic refraction measurements in the Atlantic, Pacific and Indian Oceans performed between 1990 and 2000 (data points from papers published until 1999 in easily accessible journals and books).

ment from various sources had to be handled. Here the problem of differing data recording formats soon became a time-consuming problem when trying to compile the various sub-data sets into a uniform format. To discuss this problem and seek a satisfactory solution, the International Lithosphere Program (ILP) and the International Association for Seismology and Physics of the Earth (IASPEI) formed the Mega Earth Mobile Seismic Array Consortium (MEMSAC), combining members of government agencies, universities and industry, which discussed the relevant problems at two meetings held in Germany and in California (MEMSAC Working Group, 1993).

In connection with large-scale seismic-refraction programs (so-called active studies, i.e., recording controlled sources, such as quarry blasts, borehole and underwater explosions, vibrators, or airguns), large-scale teleseismic tomography projects (so-called passive studies, i.e., long-term recording of earthquakes) were carried out since the early 1990s extending the crustal and uppermost mantle research to greater depth ranges. Examples, to name a few, are the KRISP and EAGLE investigations of the East African Rift System in Kenya (Prodehl et al., 1994a; Fuchs et al., 1997) and Ethiopia (Maguire et al., 2003), the onshore-offshore investigations of the Andean region in South America (e.g., ANCORP Working Group, 1999, 2003; Flueh et al., 1998a; Giese et al., 1999; Krawczyk and the SPOC Team, 2003, 2006; Rietbrock et al., 2005), the INDEPTH expeditions to Tibet (e.g., Brown et al., 1996; Nelson et al., 1996; Zhao et al., 1993, 1997, 2001), or the CD-ROM project in the southern Rocky Mountains (Karlstrom and Keller, 2005). For the interpretation of these data special inversion methods, used originally in passive experiments, were applied which in the course of time proved to be practical to apply to controlled-source seismology data also. In particular, the method for studies of crust and upper mantle with P-receiver functions, which was being developed since the early 1980s, and later S-receiver function studies in the late 1990s proved to be a very efficient and relatively cheap interpretation tool of the large-scale teleseismic data.

An overview on interpretation methods used in the 1990s to interpret in particular combined active and passive studies, was given at one of the latest CCSS (Commission on Controlled Source Seismology) workshop meetings, held in Dublin in 1999 (Jacob et al., 2000). Here, for example, Mechie (2000) referred to popular tools commonly used for interpreting seismic-refraction/wide-angle reflection data. Introduced in the 1970s, the ray-tracing method has remained an almost universal method for data interpretation (Červený and Pšenčik, 1984c), and the most commonly used programs include ray-theoretical and Gaussian-beam seismograms (Červený, 1985). Based on the theory of the finite-differences method for calculating the full wavefield in laterally inhomogeneous media, which was formulated already in the 1970s (e.g., Kelly et al., 1976; Reynolds, 1978), in the 1990s it became possible in practice to make such calculations for refraction/wide-angle reflection data on a crustal/lithosphere scale over several hundreds of kilometers and at realistic frequencies (e.g., Mechie et al., 1994). Since Vidale (1988) introduced ray-tracing using a finite-difference approximation of the eikonal equation, this method was improved in the 1990s (e.g., Podvin and Lecomte, 1991; Schneider et al., 1992) and also extended in a way that reflected arrivals and second arrival refractions from prograde traveltime branches could be calculated (Hole and Zelt, 1995). Also, commonly used methods in the processing of near-vertical incidence seismic-reflection data, such as normal moveout correction and migration, were applied to refraction/wide-angle reflection data (e.g., Lafond and Levander, 1995; Patzwahl, 1998).

Since the early 1990s, with the introduction of partial derivatives for the traveltimes with respect to the velocity field and layer boundaries, inversion of traveltime data in laterally heterogeneous media became possible (Lutter et al., 1990; Zelt and Smith, 1992; Hole, 1992). By the end of the 1990s, various tomographic methods existed (Hole, 1992; Zelt and Barton, 1998). Traveltime tomography became popular and evolved to a method widely applied as a first approach to model large amounts of data as well as a last check of the validity of a model (Zelt, 1998, 1999; Zelt and Barton, 1998; Zelt and White, 1995; Zelt et al., 1996, 1999, 2003; Hole et al., 2000, 2001). Nowadays there is hardly a publication on crustal and upper mantle interpretation which does not first apply a tomographic approach to the seismic data before a refined ray-tracing modeling is applied (for example, interpretations of the Polonaise and CELEBRATION 2000 data; see, e.g., Guterch et al., 2003a, 2003b; Grad et al., 2006; Janik et al., 2005). The irregularly scheduled CCSS workshops mentioned in the beginning of this chapter, which brought together scientists to discuss and practice interpretation methods on pre-distributed data sets, were continued in the 2000s (Hole et al., 2005).

## 9.10. THE STATE OF THE ART AROUND 2000

The large number of recording devices available by the year 2000 as well as the ability to record continuously over long time periods opened a new dimension. Projects could now be planned which extended seismic surveys into three dimensions. The availability of recording over long time periods has opened the possibility of avoiding organizing man-made sources such as quarry blasts, borehole or underwater explosions, but, at least for projects on land, to plan for seismic tomography surveys with teleseismic and/or local events as energy sources. As has been proved in particular by the examples of the Kenya Rift in East Africa and the CD-ROM project in North America, the combination of controlled-source seismology and teleseismic tomography has enlarged the areal coverage as well as the depth range to be explored.

Based on the incredible wealth of data gathered during the past 50 years on seismic crustal structure studies, Walter Mooney started to build up a database, which will be described in more detail in Chapter 10. Already its preliminary versions with data collected to the mid- and end-1990s enabled construction of a worldwide syntheses of crustal parameters (e.g., Christensen and Mooney, 1995; Mooney et al., 1998; Mooney, 2002) or, e.g., a synthesis of the seismic structure of the crust and uppermost mantle of North America and adjacent oceanic basins (Chulick and Mooney, 2002).