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Reykjanes Ridge Iceland Seismic Experiment (RRISP 77)

RRISP Working Group.

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Abstract. A long-range seismic refraction experiment is described which was realized in 1977 along an 800 km long line across Iceland and along the southeastern flank of Reykjanes Ridge. The main purpose of the experiment was to resolve the structure of the crust and upper mantle to greater depth than previously possible and to study the transition from the oceanic to the Icelandic structure. Shots, both on Iceland and at sea, were recorded by up to 90 stations from Iceland, Germany, and the Soviet Union and by 7 ocean-bottom stations from Germany and Canada. In addition, detailed marine seismic investigations of the Reykjanes Ridge have been carried out from board RV METEOR.

A well developed 10-km-thick oceanic crust and a stratified lower lithosphere with surprisingly high P-wave velocities (up to 8.6 km/s) has been revealed at the 10 Ma isochron on the eastern flank of Reykjanes Ridge. While the Icelandic crust differs from the oceanic one mainly by its greater thickness, the subcrustal structure is fundamentally different. The Icelandic crust is underlain by a low P-wave velocity (7.0 to 7.6 km/s) upper mantle in a state of partial fusion, which is interpreted as a diapiric updoming of the asthenosphere. No indications for continental fragments beneath Iceland could be found. Details of the interpretation are presented in two accompanying papers.

Key words: Iceland – Reykjanes Ridge – Deep seismic sounding – Crust – Lithosphere – Asthenosphere – Anomalous mantle – Partial fusion.

Introduction

As early as 1912, A. Wegener published the hypothesis 'that the Mid-Atlantic Ridge may be the place where during the still progressing expansion of the Atlantic the sea floor is continuously breaking and making room to relatively fluid and high-temperature simatic material rising from the depth' He thereby clearly anticipated a fundamental idea of the modern concepts of sea-floor spreading and plate tectonics.

For the development of these concepts, the North Atlantic and Iceland have again played an important role. The symmetric magnetic anomalies over Reykjanes Ridge (Heirtzler et al. 1966; see also Talwani et al. 1971, Vogt and Avery 1974) have been instrumental for the breakthrough of the sea-floor spreading concept. Radiometric and paleomagnetic analyses of lava successions in Iceland have become an important tool for the refinement and extension of the geomagnetic polarity time scale (McDougal et al. 1976) which is a basic requirement for paleogeographic reconstructions from geomagnetic lineations.

Anomalously low seismic velocities of 7.4 km/s in the topmost mantle first discovered below the Reykjanes ridge (Ewing and Ewing 1959) turned out to be a general property of mid-oceanic ridges and an important constraint for geodynamic and petrogenetic models. They led Ewing and Ewing (1959) to the conclusion that the mid-ocean ridges are built up by rising convection currents which supply basalt magma from the mantle and exert extensional forces on the ridges.

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Anomalous upper mantle velocities below Reykjanes Ridge were later corroborated by refraction seismic investigations of Talwani et al. (1971), Whitmarsh (1971), and Snoek and Goldflam (1978). Values between 7.2 and 7.4 km/s were also observed below Iceland by Båth (1960), Pálmason (1971) and, with a somewhat different interpretation, by Zverev et al. (1976). However, up to now very little information is available on the depth extent and internal structure of the anomalous mantle itself because of the limited range of the earlier seismic profiles.

The RRISP experiment (Reykjanes Ridge Iceland Seismic Project 1977) whose design, technical execution, and main results are described in this paper, was mainly devoted to this problem. Other important questions to which RRISP addresses itself is the structure of the crust and upper mantle beneath Reykjanes Ridge and the transition from the ridge to Iceland, about which nothing is known to date. Such an experiment must provide deep penetration of observable seismic rays and, at the same time, shallow control along the profile. For these reasons the experiment was amphibian with a terrestrial part on Iceland and a marine part on the southeast flank of the northern Reykjanes Ridge. A more detailed description of the two parts is given in the following papers of this volume, paper 2 on the land part, and paper 3 on the marine part. The present paper will be referred to as paper 1

Previous Investigations

Iceland is a unique object of earth science research because it is the world's largest well exposed and accessible segment of a mid-oceanic ridge. It is an excellent platform for the study of deep structure by methods which cannot be applied at all or only at much greater expense in the submerged parts of the ridges. But Iceland also deserves attention on it's own as the culmination of a large-scale anomaly in the North Atlantic. It is situated at the intersection of the spreading Mid-Atlantic Ridge and the aseismic transverse ridges between Baffin Island, Greenland, Iceland and the Faroes. It has been suggested that the transverse ridges have been produced since the break-up of the adjacent continents by a particularly vigorous basaltic volcanism such as is still active in Iceland (Bott et al. 1971, Bott 1974; Nilsen 1978). Spreading in Iceland, though questioned by some (Einarsson 1967; Beloussov 1970; Beloussov and Milanovsky 1976) is now convincingly demonstrated (Pálmason and Saemundson 1974; Saemundson 1978), particularly by the current rifting episode in the Krafla area (Björnsson et al. 1977, 1979; Gerke et al. 1978). Seemingly conflicting observations (Beloussov and Milanovsky 1976) are probably rather the expression of the anomalous situation of this part of the spreading ridge than truly contradicting the spreading concept.

This may be true also for the possible continental affinity of the lower Icelandic crust suggested by Zverev et al. (1976). There is no reason to dismiss this view immediately, since indications for the existence of continental fragments have been found in other parts of the North Atlantic, e.g., at Rockall Plateau (Scrutton 1972; Roberts et al. 1973), the Faeroe Block (Bott et al. 1974) and Jan Mayen Ridge (Gardé 1978). However an opposite interpretation, which more easily fits the sea-floor spreading concept, has been given by Pálmason (1971) according to whom the crust has oceanic affinity with greater thicknesses of individual layers. According to Bott (1974) this type of crust should be termed "Icelandic" rather than "oceanic"

A decision between the conflicting ideas and interpretations may only be reached through knowledge of the upper mantle structure. Anomalous upper mantle velocities have been obtained by refraction studies to date only for the depth range of some 10 to 15 km (Pálmason 1971), but teleseismic travel-time residuals of 1.3 s (Trygvasson 1964; Long and Mitchell 1970) as well as apparent velocities of P-arrivals from Mid-Atlantic-Ridge earthquakes (Francis 1969a) indicate that low velocities may extend down to some 250 km. S-wave travel-time residuals of over 5 s found by Giradin and Poupinet (1974) indicate an even more pronounced S- than P-wave velocity anomaly in the upper mantle below Iceland. Of great importance as they are, travel-time residuals alone cannot be used to resolve details of structure because of their integral nature. Surface wave studies (Tryggvasson 1962; Giradin and Jacoby 1979; Jacoby and Girardin 1980) are subject to similar limitations. Only in combination with the boundary conditions provided by detailed and deeply penetrating refraction seismic investigations, can the above methods be put to full use in developing geodynamic models.

Layout of the Experiment

Figure 1 shows the layout of the whole experiment. The present paper deals mainly with Profile I. The investigations along the marine segment were the central objective of cruise 45/I of RV METEOR. During this cruise three additional lines at sea (Profiles III-V) were shot in order to investigate the evolution with age of the oceanic lithosphere between the ridge crest and our main line. Details of these lines will be published elsewhere.

The main line starts at shot-point A on the southeastern flank of Reykjanes Ridge and runs slightly oblique to magnetic anomaly 5 (approximately 10 Ma), but parallel to the depth contours. It enters Iceland north of Heimaey and lies mostly within the zone of active rifting and volcanism (eastern neovolcanic zone), which is thought to be the continuation of the active spreading axis offset by transform faults south and north of Iceland. North of the central volcano Askja the main line splits into three branches. One continues straight through the Quaternary and Tertiary basalts towards shot-point E (Vopnafjördur), while the second follows the strike of the neovolcanic zone towards shot-point F in the north (Axarfjördur). The third branch is perpendicular to the line A–E in order to provide information on possible lateral variations between the neovolcanic zone and the Tertiary region of the southeast coast. Since the main program could be carried out within 9 days as planned, the remaining days have been used for the observation of an additional profile along the southeast coast (Profile II), which stays completely within the older part of Iceland and which therefore can serve as a reference line.

Different types of recording instruments were used by the different participating groups. The Soviet team used continuously recording refraction stations equipped with three-component seismometers with 2 s natural period (Zverev et al. 1978) at the positions marked by solid inverted triangles. The solid circles in Fig. 1 show positions of recording sites occupied by the German group using a total of 40 seismic stations of MARS 66 type (Berckhemer 1970) and two digitally recording MARS-66-compatible PCM stations (Gebrande et al. 1977). At sea seven refraction instruments were launched and recovered: five anchored telemetric buoy systems of the Institut für Geophysik, Hamburg (Kebe 1971, Weigel et al. 1978) and 2 ocean bottom seismographs (OBS) of Bedford Institute of Oceanography, Dartmouth, Nova Scotia (Sutton et al. 1977; Heffler and Barrett 1979). Their positions are shown in Fig. 1. All systems were provided with bottom hydrophones. In addition each OBS contained one vertical and one horizontal

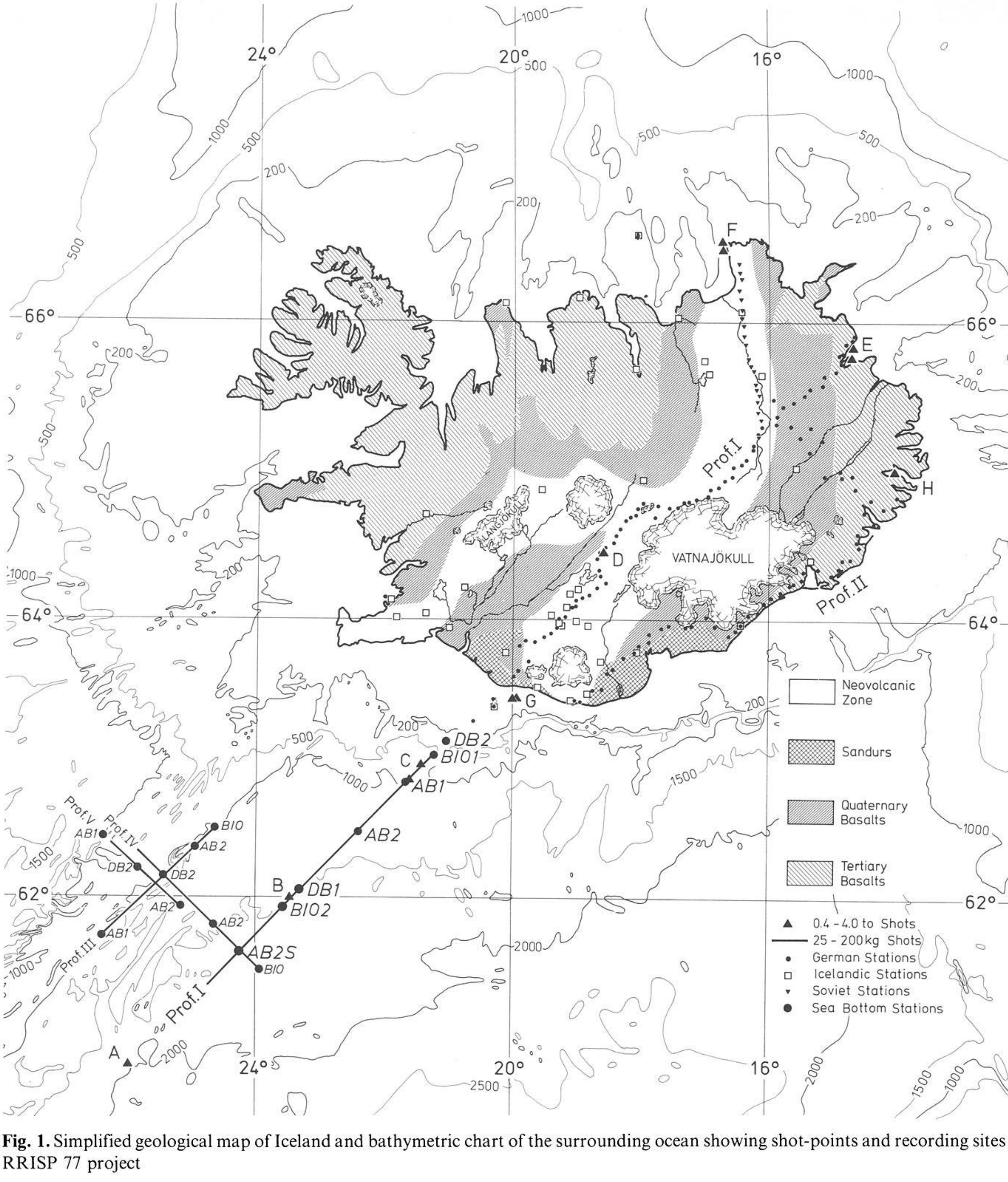

Fig. 1. Simplified geological map of Iceland and bathymetric chart of the surrounding ocean showing shot-points and recording sites of the RRISP 77 project

4.5 Hz seismometer. The principles of the shooting and recording technique at sea are summarized in Fig. 2. Furthermore most of the shots were observed on land by the continuously recording permanent and semipermanent seismic stations of the Icelandic seismological network (Einarsson 1979). Some of these stations were installed especially for the program. Positions are marked as open squares in Fig. 1.

Table 1 gives information on the schedule of the whole experiment and on the large shots fired. The shooting technique varied according to circumstances. At the only land-based shot-point (D) in a shallow lake (Thveraldavatn; depth 17 m) without drainage, the total charge was distributed over a rectangular grid of single 50 kg charges spaced 10 to 20 m apart at the lake bottom; they were fired simultaneously. This shooting technique (Burck-

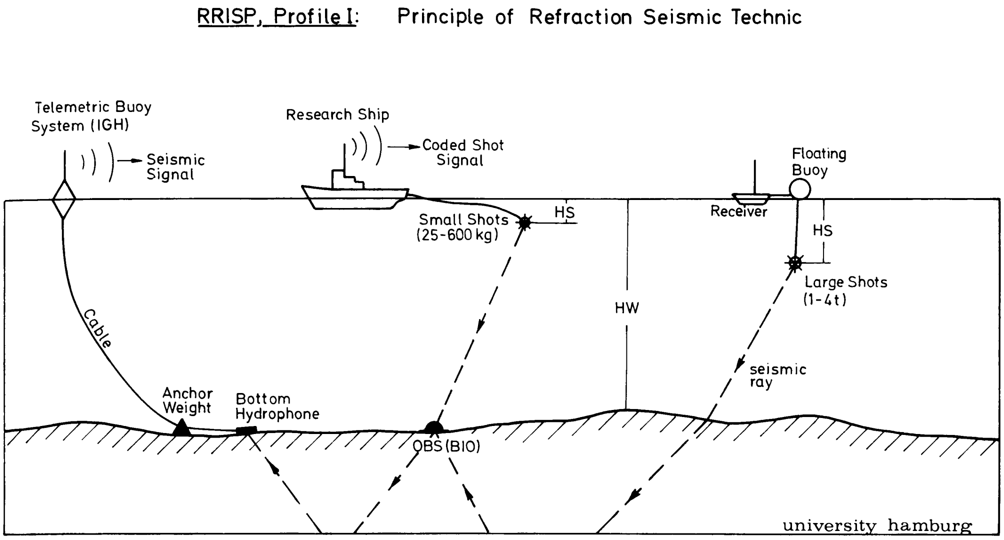

Fig. 2. Shooting and recording technique employed for the marine part of RRISP 77

Table 1. RRISP – Schedule of large shots

Shot no.	Date	Time (GMT)	Longitude west	Latitude north	Charge (kg)		Charge depth (m)	Water depth (m)
C0	12. 7. 77	16–30–00.65	21° 35.9′	62° 51.8′	400		270	1126
C1	14. 7. 77	19–58–02.86	21° 25.0′	62° 58.7′	1000		130	1025
C2	20. 7. 77	10–01–01.78	21° 37.1′	62° 52.0′	1000		130	1117
C3	25. 7. 77	17–01–02.92	21° 35.8′	62° 51.6′	400	single charge	110	1116
B1	15. 7. 77	12–30–36.85	23° 52.9′	61° 59.2′	600		120	1565
B2	17. 7. 77	17–06–22.78	23° 27.4′	61° 59.4′	2000		150	1462
A	18. 7. 77	13–01–04.38	25° 57.8′	60° 43.8′	4000		170	2005
D0	12. 7. 77	18–31–02.79	18° 35.78′	64° 28.33′	500		17	17
D1	15. 7. 77	11–31–02.98	18° 35.78′	64° 28.33′	1000		17	17
D2	18. 7. 77	11–31–02.44	18° 35.78′	64° 28.33′	1000		17	17
E1	19. 7. 77	14–02–10.83	14° 40.93′	65° 50.16′	500		104	104
E2	19. 7. 77	16–02–00.61	14° 42.06′	65° 46.22′	500		110	110
F1	15. 7. 77	15–00–58.42	16° 43.60′	66° 30.16′	500	distributed	115	115
F2	15. 7. 77	17–01–01.60	16° 43.10′	66° 28.15′	500		124	124
F3	17. 7. 77	16–01–02.12	16° 43.10′	66° 27.80′	500		124	124
F4	17. 7. 77	18–01–01.33	16° 43.60′	66° 30.16′	1000		115	115
G1	24. 7. 77	10–01–01.85	20° 00.57′	63° 27.49′	500		110	110
G2	24. 7. 77	11–01–02.85	19° 57.54′	63° 28.15′	500		110	110
H	24. 7. 77	19–01–09.26	14° 02.00′	65° 01.37′	500		113	113

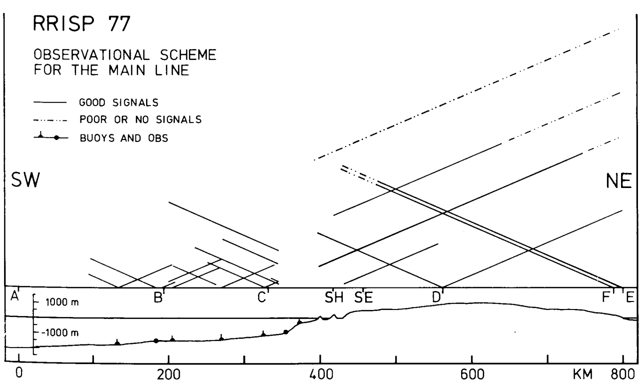

Fig. 3. Simplified diagram showing scheme of observations for the RRISP 77 main profile. Letters *A* to *F* denote shot-points. SH and SE positions of two stations of the Icelandic seismological network. *Solid* and *dot-dashed lines* indicate observational range of sufficient and insufficient signal to noise ratio

hardt and Vees 1975) yielded satisfactory results, even though no optimization with regard to water depth and charge size could be achieved. At the shotpoints close to the coast 500 kg single charges (Soviet TNT and GEOSIT II(TM)) were fired at the sea bottom at optimum depth (110 m), determined by the condition of constructive interference between the gas bubble pulsation and the reverberation of the water layer. Shooting at these points was carried out from small Icelandic fishing vessels rented for that purpose. These shots on the sea bottom were especially valuable, because they radiated not only the expected strong P-waves but also strong S-waves. The shots at points A, B and C were fired from RV METEOR. They were suspended below a buoy at optimum charge depth and detonated by means of a coded radio signal transmitted from the ship. In addition to the large shots, a total of 141 small shots (25–200 kg) spaced 1.8 km apart were fired along the line from receiver BIO1 to south of receiver AB2S (see Fig. 1). These charges were dropped directly from METEOR and detonated by means of a firing line. For the shots fired from METEOR, GEOSIT II(TM) was used as the explosive.

The positions of the buoys, OBS's, and shots at sea were determined by LORAN C and satellite navigation. At the shotpoints close to the coast positioning was done by radar tracking of the coast line, determination of water depth, and photofixes of the shore line. The positions of land stations were determined from aereal photographs and topographic maps on the scale of 1:50000 and 1:100000. In all cases, the accuracy of the positions is estimated to be ± 200 m or better.

Figure 3 gives the scheme of observations along the main line showing the system of reversed and overlapping profiles actually observed. Most of the shots were recorded up to the expected range, but for shot-points A and B energy propagation was not sufficient to cover the whole land line in spite of charges as large as 4 and 2 t. As will be shown later, this is most probably caused by structure and/or absorption rather than by poor shot energy. The two positions SH and SE are locations of two stations of the permanent Icelandic seismological network (Storhöfdi on Heimaey and Selkot at the south coast) which like others have recorded many of the smaller METEOR shots. These recordings were useful in closing the information gap between the marine and the land part.

A combined land-sea experiment of such an extent obviously calls for good communication links between the ship and the land stations. Shooting at sea is highly dependent on weather conditions especially when large charges are to be fired. Therefore the whole schedule is subject to changes and land stations must be informed immediately. During the experiment two headquarters were established, one at Sigalda in the south, the other at Krafla in the north. Both headquarters maintained a permanent shortwave communication link with each other and at Sigalda a strong transmitter (2.5 kW) was used to maintain contact with RV METEOR and also to broadcast news for the observers, who were all equipped with appropriate receivers. During shot-windows Sigalda also transmitted a quarz clock controlled time signal, which could be used as back-up in case the MSF time signal, which was normally recorded, could not be received. Besides our own news broadcasts, information for the observers was also broadcast three times daily by all stations of the Iceland State Broadcasting Service, as shortwave propagation varied very much, but medium and long wave propagation conditions remained fairly constant. This rather elaborate communications set-up was necessary because recording crews consisting of two persons in one four-wheel drive vehicle could not contact the headquarters frequently because of remote and rugged terrain. They were completely self-supporting with respect to food and shelter and were supplied with gasoline whenever tapes were relayed into headquarters. To gain quick information on travel-times and energy propagation, as well as to spot equipment problems, the tapes were played back at headquarters as they came in.

Data

All recordings were digitized for later processing. This was done at the Geophysical Institutes at Karlsruhe (MARS 66 data), Mu-

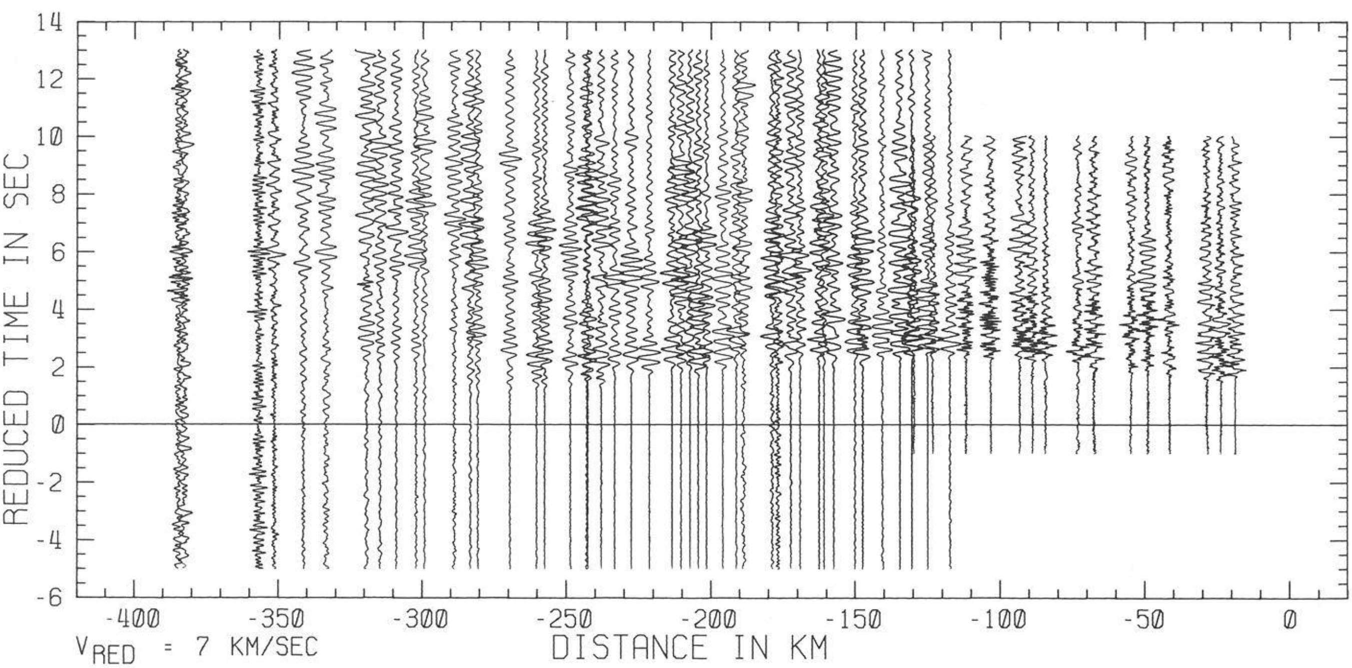

Fig. 4. Record section along the land segment of the main profile; shots F1 and F4, bandpass filtered 1–10 Hz. The section contains records obtained by the Soviet group (0 to 135 km) as well as records obtained by the German group (120 to 390 km). Reduction velocity is 7 km/s

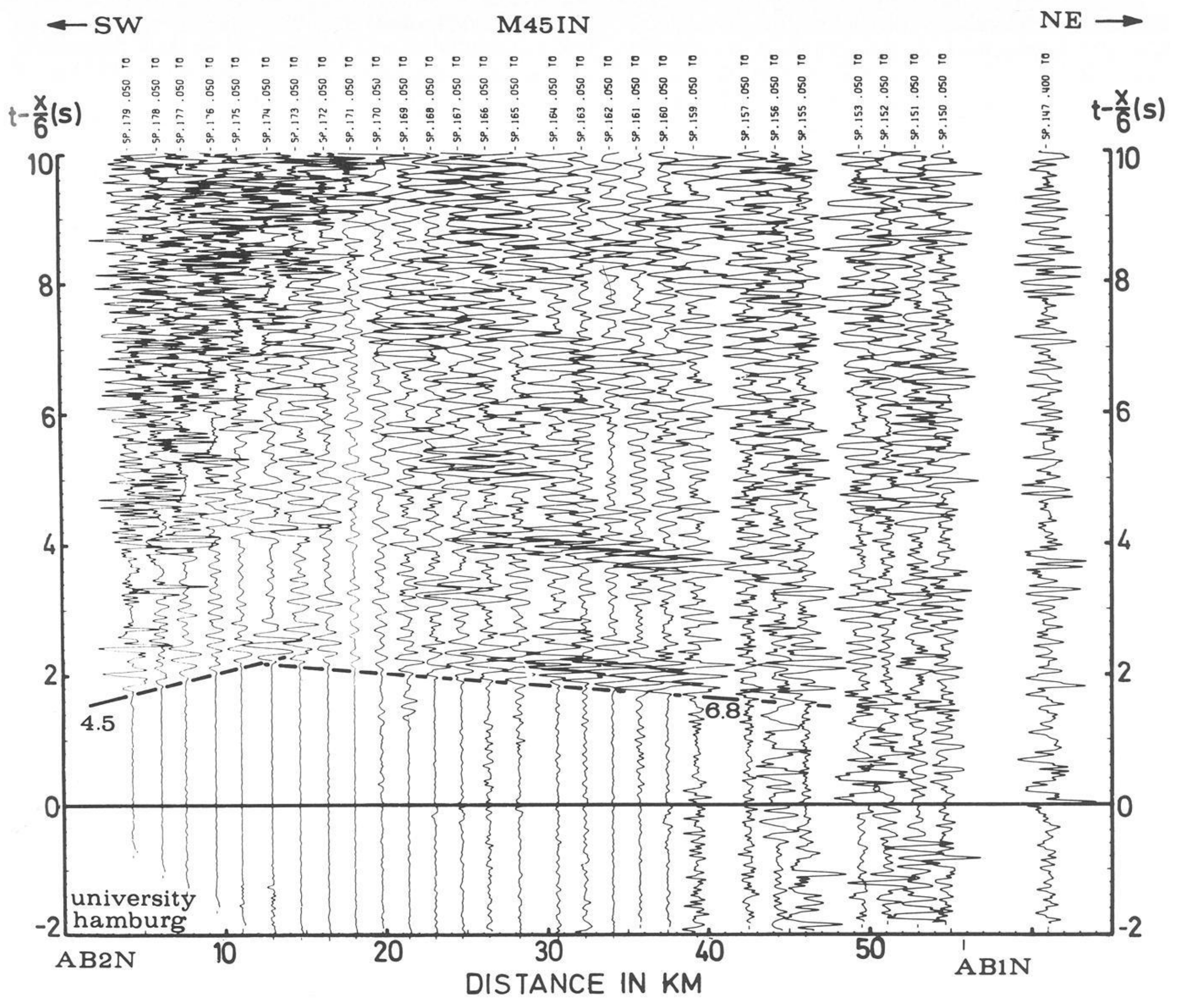

Fig. 5. Marine record section reduced by 6 km/s and bandpass filtered 2–20 Hz. Data were recorded by the ground hydrophone (water depth: 1389 m) of buoy AB2N. Shots were fired along the main line towards the northeast

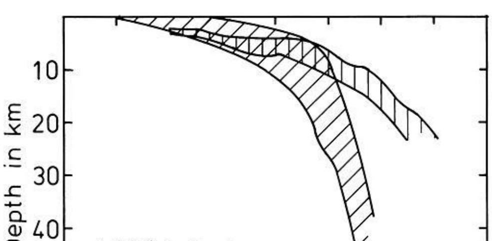

Fig. 6. Extremal bounds on possible velocity-depth distributions beneath Iceland and the southeastern flank of Reykjanes Ridge. The broader region of possible solutions for the land part is indicative of a more heterogeneous velocity distribution

nich (PCM and Soviet data), Hamburg (buoy data), and at Bedford institute (OBS data). The Soviet data were digitized by hand from analog playbacks. The record-sections of Figs. 4 and 5 show examples of the data after some basic processing (digital band-pass filtering). Fig. 4 shows a typical land record-section. It is a compilation of records obtained by the Soviet and German groups from shots F1 and F4, which were fired at the same location but recorded at different recording sites. Reduction velocity is 7 km/s. In the distance range of 120 to 135 km both data sets overlap and one may notice how well they fit.

Each seismogram is normalized with respect to its maximum amplitude and it may be seen that at distances beyond some 240 km the maximum energy within the seismograms is shifted from the first arrival wave train to later times. One may clearly distinguish two different travel-time segments as delineated by first arrivals. The first one with an intercept time of 1.3 s, indicating a thick low-velocity surface layer shows an apparent velocity of 6.8 km/s. At distances greater 120 km, the arrivals can be attributed to another segment with an apparent velocity of 7.2 km/s.

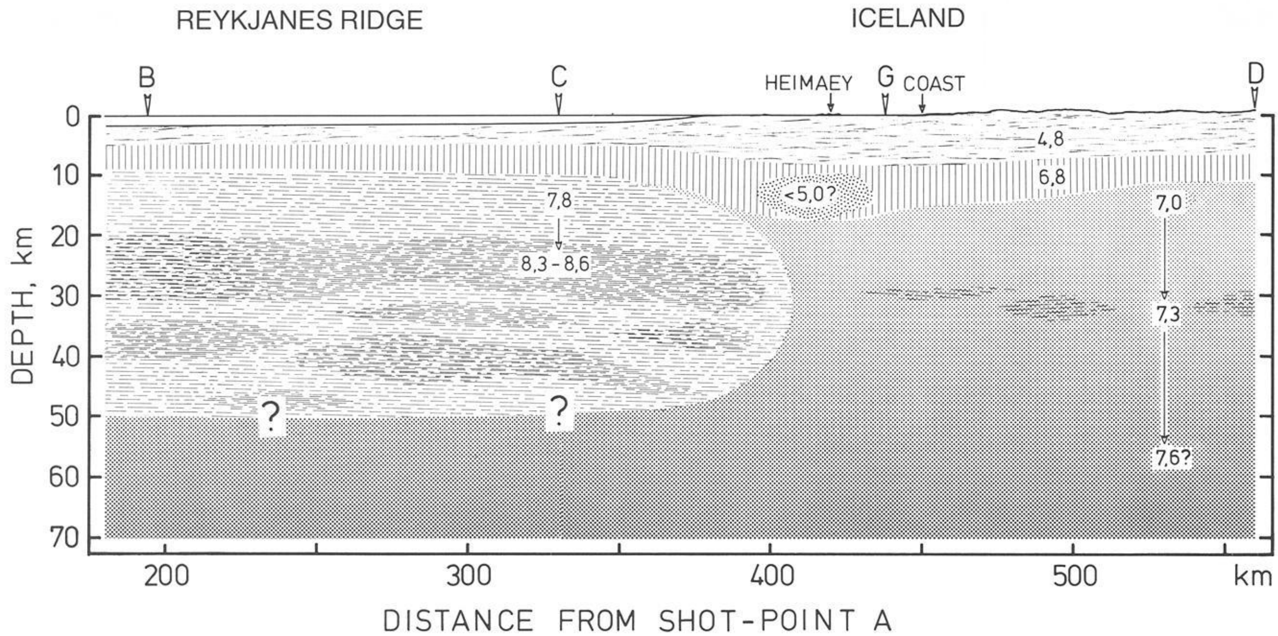

Fig. 7. Generalized crustal and upper-mantle cross section of the central part of RRISP 77 main profile. Letters indicate positions of large shots. Numbers give P-wave velocities in km/s. Crustal layers are continuous across the transition from the ridge to Iceland, whereas there is a drastic change in upper-mantle structure close to the shelf slope, where a well developed oceanic lithosphere beneath Reykjanes Ridge abuts on anomalous upper-mantle material in the state of partial fusion

Both segments can be alternatively correlated as one continuous travel-time curve, indicating velocity gradients rather than discontinuities. The existence of velocity gradients is also suggested by the absence of prominent later arrivals, which could be correlated over some distance and interpreted as reflections or diving waves. This is a salient distinction from seismogram sections obtained in continental areas.

Figure 5 is an example of a record section from the marine part, which (contrary to Fig. 4) has been reduced by 6 km/s. All records were obtained at buoy AB2N from shots along the line towards buoy AB1N. All seismograms were filtered with a passband of 2–20 Hz, and are normalized individually with respect to their maximum amplitudes. The section is uncorrected for shot and receiver depths, but nevertheless it gives a good idea of the general character of the marine data. True velocities obtained in combination with the reversed profile (see Fig. 6 in Goldflam et al. 1980) are rather typical for a well developed oceanic crust. The prominent later arrivals parallel to the first ones are multiple reflections from the water surface at the receiver side.

Results

The results presented below are based on most of the data, but only some of the record-sections are shown here, as well as in Gebrande et al. 1980 and Goldflam et al. 1980. As we concentrate here on the general structure, small-scale regional differences will not be discussed.

As a first stage in interpretation, smoothed travel-time curves of the land and marine part were inverted using the tau-p method (Bessonova et al. 1974) and a generalized Wiechert-Herglotz-method. The results are summarized in Fig. 6 with the hatched regions giving the range of possible velocity-depth distributions. In spite of broad bands, a clear difference between the Reykjanes-Ridge flank and Iceland is evident, the marine crust showing a stronger velocity increase with depth. To evaluate the differences in more detail and to obtain the horizontal velocity variations along the main line, extensive ray tracing was performed (see Gebrande et al. 1980; Goldflam et al. 1980).

The main results obtained are shown in Fig. 7. Although generalized, it does show the principal structural trends rather accurately. The different surficial layers have been combined into one with an average velocity of 4.8 km/s. Velocity gradients in this layer are, at least in Iceland, very likely (Flóvenz 1980) but could not be resolved accurately with the receiver spacing of RRISP 77, which was chosen with the aim to obtain structure at greater depth. The necessary crustal control is provided by detailed small-scale seismic refraction work of Pálmason (1971).

Layer 3 is rather homogeneous and continuous from the ocean through Iceland; thus it exists also beneath the neovolcanic zone. The thickness of the crustal layers is generally greater underneath Iceland than under the ocean and their top and bottom show greater relief. Total crustal thickness including the water layer is some 10 km beneath the segment along Reykjanes Ridge, but varies between 15 km in the southeastern and northern, and 10 km in the central section beneath Iceland. Large travel-time residuals for the more distant shots at the recording sites on Heimaey, but normal arrival times for shots G1 and G2, are an indication for the possible existence of a magma chamber in the lower crust beneath Heimaey. This body is shown schematically in Fig. 7 and velocities lower than 5 km/s are attributed to it. A similar body may exist in the area of Askja in northern Iceland (Gebrande et al. 1980).

Below layer 3, fundamental differences in structure exist between the ocean and Iceland. On the marine side we find a velocity of 7.7 to 7.8 km/s which, if interpreted as layer 3b, would be anomalously thick; it is rather attributed to the upper mantle (Golflam et al. 1980). Similar upper-mantle velocities have earlier been reported by Whitmarsh (1971) for an area 200 km southeast of shot-point A and are also observed beneath the Iceland-Faeroe Ridge (Bott and Gunnarsson 1980). As there is no evidence for velocities of 7.2 to 7.4 km/s as found by Ewing and Ewing (1959) and Talwani et al. (1971) at or near the ridge crest, we must assume a rather well developed oceanic lithosphere. The sub-Moho velocity remains nearly constant down to 15 km but increases continuously at greater depth and reaches values of 8.3 to 8.6 km/s at a depth of about 20 km. This is an important new feature in that area. The 8.3 km/s velocity was observed at OBS BIO1 in the distance range of 90 to 140 km as first arrivals. In the reversed direction even higher velocities were observed at OBS BIO2 in the same distance range. These high velocities are corroborated by records of the Icelandic seismological network at distance ranges of 100 to 180 km. Apparent velocities at stations near the south coast observed from marine shots between points B and C reach values as high as 9 km/s after correction for water depth. This can only partially be explained by interfaces dipping towards Iceland. The data therefore require the existence of P-wave velocities of up to 8.6 km/s or so. These P-wave velocities, however, cannot be constant in the lower lithosphere, because the apparent velocities vary rather irregularly and because the mean S-wave velocities deduced from the inversion of surface waves along the same line are rather low (Jacoby and Girardin 1980). This inhomogeneous velocity distribution is schematically depicted by the darker hatched areas in Fig. 7.

No direct information was obtained on the depth of the lower boundary of the lithosphere below Reykjanes Ridge. However, evidence from surface waves and missing signals from shotpoint A, which must be explained by downward refraction of rays originating from A due to a negative velocity gradient, give a rough idea of the thickness of the lithosphere. The boundary to the asthenosphere was put, somewhat arbitrarily, at a depth of 50 km.

There is no evidence for the continuation of the marine high-velocity lithosphere underneath Iceland. Instead we find velocities increasing from 7 to 7.4 km/s in the same depth range. Only locally, slightly higher velocities may exist (see Gebrande et al. 1980), as symbolized in Fig. 7 by the darker-shaded areas at 30 km depth beneath Iceland. The transition from the high-velocity oceanic lithosphere to the low-velocity upper mantle beneath Iceland is rather abrupt and must occur close to the shelf slope. Although shown here only along the main line, the anomalous upper mantle is not confined to the neovolcanic zone, but extends under all of Iceland. It is also present beneath the older parts of Iceland such as the Tertiary in the east (as shown by Profile II) and west (Båth 1960).

This may account for the rather homogeneous travel-time anomalies (Long and Mitchell 1970). The residuals can be quantitatively explained by our model only if the anomalous mantle extends to depths greater than 100 to 200 km depending upon the velocity gradient assumed below 30 km.

More light on the nature of the anomalous mantle (layer 4 of Pálmason 1971) is shed by the analysis of S-wave propagation through Iceland. As shown by Gebrande et al. (1980) the P to S velocity ratio is 1 76 in the Icelandic crust, which is rather normal for basic material, but reaches values between 1.96 to 2.2 at greater depth. While gabbroic material could marginally account for the observed values at moderate temperatures and pressures (Christensen 1978), this explanation fails in view of the high temperatures (1000–1100°C) at the base of the crust as derived by magnetotelluric (Beblo and Björnsson 1978) and geothermal investigations (Pálmason and Saemundson 1974, Pálmason et al. 1979). The only remaining possibility is partially molten ultrabasic material, as has been suggested earlier by Bott (1965), Pálmason (1971) and Pálmason and Saemundson (1974). As shown by Gebrande et al. (1980) the observed P and S velocities allow a quantitative estimation of the degree of partial fusion. Assuming a mixture of solid peridotite and basaltic melt, we infer a melt concentration greater than 17% at the base of the crust decreasing with increasing depth (see also Beblo and Björnsson 1980). The predominantly ultrabasic composition is the main justification for calling the layer with P-wave velocities from 7.0 to 7.4 km/s 'anomalous mantle' Relying only on the P-wave velocities, it would have been more natural to attribute this depth range to the crust, since similar velocities are widespread in the lowermost crust. In this case the thickness of the Icelandic crust would be about 30 km, the inhomogeneities at this depth range (see Fig. 7) may give some support for such an alternative interpretation. The interpretation as anomalous mantle is largely based on the comparison of the observed seismic wave velocities with laboratory data. Additional laboratory investigations of seismic wave transmission in partially molten rocks would be very important to corroborate or possibly modify our conclusions.

From the extent of the anomalous mantle it may be deduced that partial melting is not confined to the narrow neovolcanic zone but may exist, most likely with local variations in concentration, below the whole island and may be responsible for the high mobility of the Icelandic crust and the existence of active volcanism outside the neovolcanic zones, e.g., at Snaefellsnes and Oraefi. The same conclusion has been reached by Beblo and Björnsson (1978) from the distribution of electrical conductivity derived from magneto-telluric measurements.

Under the seaward continuation of the profile similar partially molten material is obviously absent at shallow depth. However, one may speculate whether the 7.8 km layer might be the marine counterpart of the Icelandic anomalous mantle, but cooler and correspondingly with a smaller melt content. Another possibility would be that this layer is composed of perodotite enriched in a solidified low-melting-point fraction as postulated by the petrogenetic model of Bottinga and Allègre (1976) for the depth range of 10 to 20 km. Unfortunately, since S-waves were not observed on the marine part a decision between both alternatives is not possible.

Another open question is the petrological interpretation of the high upper-mantle velocities (up to 8.6 km/s) underlying the 7.8 km/s layer. Similarly high velocities have been reported for the lower oceanic lithosphere from different areas and by different authors, as has recently been reviewed by Asada and Shimamura (1979). It appears impossible to understand these high velocities in terms of todays knowledge about upper mantle petrology without invoking anisotropy (Green and Liebermann, 1976, Bottinga and Allègre, 1976). The model most favoured for the generation of upper mantle anisotropy is the one proposed by Francis (1969b). According to this model, anisotropy is due to the preferred orientation of the *a* crystallographic axis of olivine parallel to the upper mantle flow lines, induced by plastic deformation, the direction of maximum velocity is therefore expected to be perpendicular to the ridge axis and to the magnetic lineations. This model would require lower velocities in the direction of our main line, contrary to observation. Our results therefore suggest another explanation, namely that the high velocity layer behaves as a transversely isotropic medium with minimum velocity in the vertical axis of symmetry and maximum velocity in horizontal directions. This type of anisotropy is to be expected from the preferred orientation of the *b* crystallographic axis of olivine in the vertical direction and from randomly distributed *a* and *c* axes in the horizontal plane as is common in cumulate rocks (Christensen 1978). This would imply that the high velocity layer is the product of segregation of olivine from partially molten anomalous mantle rather than being a residual layer formed by depletion in low melting point fraction by rising magma. Another explanation of the upper-mantle velocity distribution, involving Francis-type anisotropy, is discussed by Goldflam et al. (1980). The additional marine profiles (Fig. 1) may resolve this question.

Conclusions

Combined land-sea refraction seismic investigations along an 800 km long line from Reykjanes Ridge to northeast Iceland have revealed the following:

1 Contrary to current opinion, we find a well developed oceanic crust and stratified subcrustal lithosphere already at the 10 Ma isochron (anomaly 5) on the southeastern flank of Reykjanes Ridge.

2. High P-wave velocities of up to 8.6 km/s have been measured parallel to the ridge axis at depths greater than 20 km bearing on possible models of lithospheric evolution.

3. Except for greater thickness and more pronounced lateral variations, the Icelandic crust is very similar to the crust beneath Reykjanes Ridge. Layer 3 can be traced without interruption from the 10 Ma old oceanic crust through Iceland and it exists also beneath the eastern neovolcanic zone, which is supposed to be the present-day active spreading axis.

4. The subcrustal structure below Iceland is fundamentally different from the one below the southeastern flank of Reykjanes Ridge. The Icelandic crust is underlain by low P-wave velocity (7.0 to 7.6 km/s) material down to depths greater than 50 km, which must be interpreted as mantle material in the state of partial fusion. This anomalous mantle seems to form a diapiric updoming of the asthenosphere, much broader than is likely beneath mid-ocean ridges proper. The formation of such an anomaly seems hardly possible without a locally enhanced upwelling convection current, which some might like to call a plume.

5. The transition of the subcrustal structure from Reykjanes Ridge to Iceland takes place within a very narrow zone below the shelf slope. One might speculate whether this sharp boundary is present everywhere around Iceland or whether it is a special feature connected with the Reykjanes Fracture Zone.

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Appendix

Participating Institutions (on board RV METEOR)

Institut für Geophysik, Universität Hamburg (IGH)
Deutsches Hydrographisches Institut, Hamburg (DHI)
Institut für Meteorologie und Geophysik, Universität Frankfurt (IGF)
Bedford Institute of Oceanography, Canada (BIO)
Institut für Geophysik, Universität Kiel (IGK)
Massachusetts Institute of Technology, USA (MIT)
Freie und Hansestadt Hamburg, Baubehörde (HHB)
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Participating Scientists and Technicians

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IGF:	W Jacoby
BIO:	B.D. Loncarevic and D. Heffler
IGK:	H.J. Brink
MIT:	S. Solomon and S. Richard
HHB:	H. Gäbler
E 71	E. Kuthning

Participating Institutions (on Iceland)

Institut für Geophysik, Freie Universität Berlin (B)
Institut für Geophysik, Technische Universität Clausthal (CLZ)
Institut für Meteorologie und Geophysik, Universität Frankfurt (IGF)
Institut für Geophysik, Universität Göttingen (G)
Niedersächsisches Landesamt für Bodenforschung, Hannover (NLfB)
Institut für Geophysik, Universität Karlsruhe (KA)
Institut für Allgemeine und Angewandte Geophysik, Universität München (M)
National Energy Authority (Orkustofnun), Reykjavik (ORK)
Science Institute, University of Iceland, Reykjavik (SCI)
Institute Physics of the Earth, Moscow (IPEM)
Mining Institute, University of Leningrad (MIL)

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