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New Heat Flow Observations on the Reykjanes Ridge

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Abstract. During METEOR cruise 45 in August 1977 14 heat flow measurements were obtained along a profile east of the Reykjanes ridge and perpendicular to the ridge axis covering a distance range from 30 to 240 km. Closely spaced measurements were grouped together. The mean heat flow of all groups amounts to 109 ± 20 mW m^{-2}. The values do not reveal a distinct increase with decreasing distance from the ridge axis as may be expected from the theoretical heat flow distribution based upon a cooling plate model. Including earlier measurements a high and very uniform heat flow in the distance range from 170 to 340 km was observed with a mean value of about 100 mW m^{-2}. In order to explain this high heat flow a temperature of 930° C is required at the lower boundary of the lithosphere at a depth of 50 km, assuming a purely conductive heat transport. Compared with the results obtained from previous measurements west of the Reykjanes ridge, the data reveal an asymmetric thermal behaviour of the ridge area. The average heat flow east of the ridge amounts to 93 mW m^{-2} being nearly twice the heat flow of the region west of the ridge.

Key words: Geothermics – Heat flow – Reykjanes Ridge.

Introduction

The mid-Atlantic ridge and Iceland which forms part, and represents a remarkable anomaly of, this ridge are regarded to be the result of geodynamic processes in the interior of the earth. To get a better understanding of those processes, knowledge not only of the structure of the earth's crust is necessary but also of the terrestrial heat flow which reflects the thermal behaviour of the earth's interior.

A compilation of all heat flow values obtained until 1970 in the region of the Reykjanes ridge was given by Talwani et al. (1971). The mean heat flow of the stations, most of which lie west of the ridge, amounts to 53 mW m^{-2} which is still below the average heat flow of the Atlantic ocean. This result does not correlate with the theoretically expected heat flow distribution at an active sea-floor spreading center without additional assumptions about convective heat transport (e.g., Elder, 1965; Palmason, 1967; Williams et al., 1974). Later measurements in the same region are reported by Langseth and Zielinski (1974), yielding a somewhat higher value of 67 mW m^{-2}.

Within the scope of geophysical investigations of the deeper crustal structure of oceanic ridges, geothermal investigations were carried out during leg 2 of the RV METEOR cruise 45 in August 1977. The aim was to increase the number of heat flow data in the region of the Reykjanes ridge and to study the thermal state of the lithosphere.

Geothermal Stations

The Reykjanes ridge extends with a nearly constant strike of N 35° E from lattitude 56° N to the Reykjanes peninsula of Iceland. Geothermal investigations were performed along an observation line perpendicular to the ridge axis at the eastern part of the ridge (Fig. 1). This profile was selected according to known ocean bottom topography and sedimentary cover obtained from a high resolution sparker profile during leg 1 (Fig. 2). Topography and the approximate thickness of the sediments are shown in Fig. 3 together with the total intensity of the earth's magnetic field recorded along the profile.

The topography in the area of stations 316 through 318, situated outside the range covered by the sparker profile, was obtained from an ELAC narrow-beam echosounder. The positions of the stations were determined by satellite and Loran C navigation with an accuracy better than 1,000 m. The station next to the ridge axis lies just at the beginning of the eastern flank whereas the easternmost stations are situated on the flat und obviously undisturbed deep sea bottom. The age of the ocean bottom covered by the geothermal profile ranges from about 2.5 Ma to about 25 Ma. The magnetic anomaly 5 corresponding to an age of 9 Ma is clearly indicated at a distance of 90 km from the ridge axis (Fig. 3).

The Measurements

The measurements were carried out with a modified deep sea probe (Haenel, 1972). Five outriggers equally spaced and fixed on a piston core barrel measure the temperature of the sediment and the thermal conductivity in situ with the needle probe method (Von Herzen and Maxwell, 1959). A pressure vessel contains the electronic recording equipment. The length of the core barrels used was 3.5 and 5.0 m respectively.

According to the length of the core barrels the maximum depth penetration up to the base of the pressure vessel was 4 and 5.5 m, respectively. In addition to the in situ measurements the thermal conductivity of the sediments recovered in the core barrels was also measured by the needle probe method. Measurements were made every 0.2 m down the core sample. In general, the cores were about 1.5 m shorter in length than the depth of penetration. In order to correlate the conductivity values of the core with the positions of the measured temperature, a loss of the upper 1.5 m of sediments was assumed. The reason is probably that the piston may not have started to suck before the core barrel has penetrated to this depth.

The temperature of the sediments and their thermal conductivities are presented in Fig. 4. The temperatures are related to the bottom water temperature measured at the different stations. A

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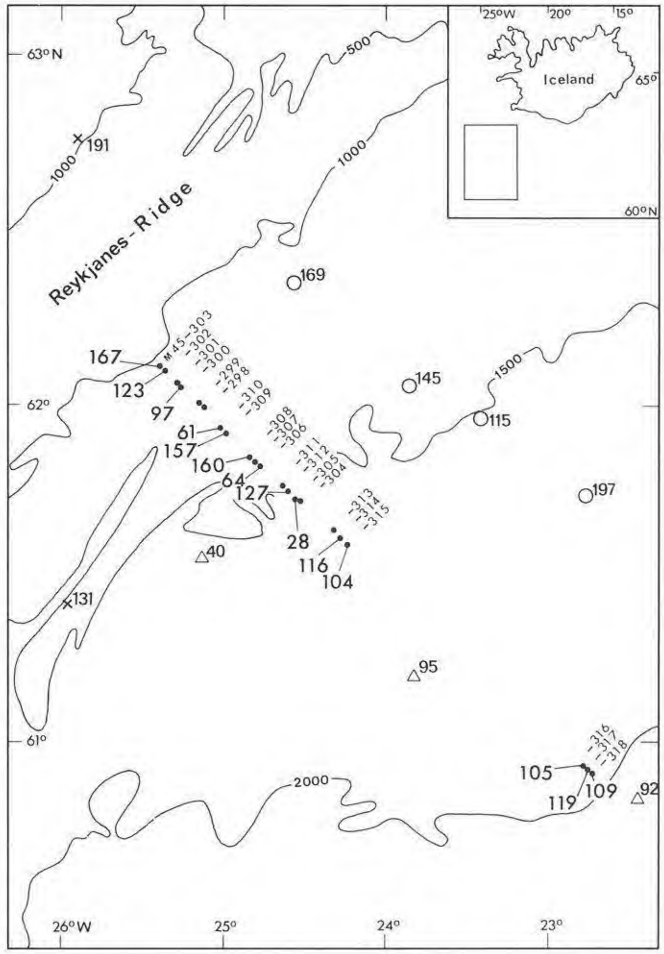

Fig. 1. Survey area and position of the geothermal stations. (*dots*). Stations of previous surveys are indicated by crosses (Talwani et al., 1971); *circles* (Scheljagin et al., 1973), and triangles (Langseth and Zielinski, 1974). Heat flow values are given in $mW\ m^{-2}$ and depth contours in m

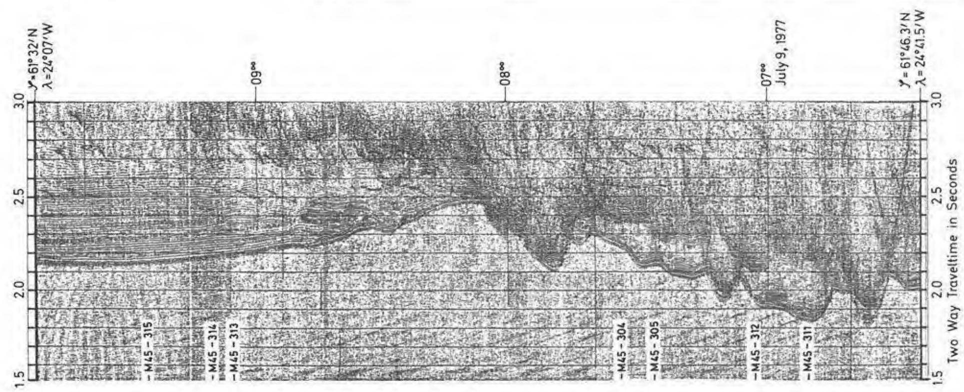

Fig. 2. Section of the sparker profile IV (by courtesy of the Institute of Geophysics, University of Kiel)

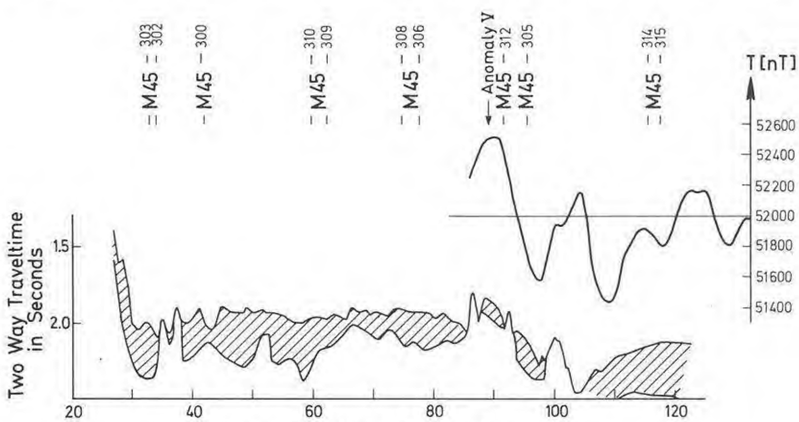

Fig. 3. Topography and sedimentary cover along the geothermal profile as obtained from the sparker profile IV and total intensity of the earth magnetic field recorded along the profile (by courtesy of Deutsches Hydrographisches Institut, Hamburg)

large positive temperature gradient is observed in the uppermost sediments. The gradient decreases rapidly to about half the value above a depth of 2 to 2.5 m and remains more or less constant below that depth at the stations 300 through 315. For the stations 316 through 318 the temperature gradient becomes nearly linear below a depth of about 3.5 m. Both the conductivity values measured in situ and from the core samples scatter considerably, but neither a significant difference nor an obvious depth dependence can be seen. Therefore it is very likely that the temperature distribution in the uppermost sediments is caused by variable bottom water temperatures.

It was shown, e.g., by Worthington and Volkmann (1965), Jones et al. (1970), Vogt and Johnson (1973) that the Norwegian Sea water crossing the Iceland-Faeroe ridge flows southwestward along the eastern flank of the Reykjanes ridge at a depth greater than 1,500 m. The temperature disturbances up to a depth of only 2.5 m, observed at the stations 300 through 312, may be explained, therefore, by a short-period emerging of cold water above the 1,500 m level. The mean bottom water temperature measured in the depth interval of 1,380 to 1,560 was 3.6° C. Extrapolating the linear segment of the temperature curve up to the sediment-water boundary, an average temperature variation of $\pm 0.3°$ C results. Based upon this value and assuming a temperature diffusivity of 0.003 cm^2 s^{-1}, the minimum temperature was probably reached during May or June. A similar result was obtained by Sclater and Crowe (1979), (John G. Sclater, personal communication) from the interpretation of geothermal measurements, carried out in July 1977 along the magnetic anomaly 13 east of the Reykjanes ridge.

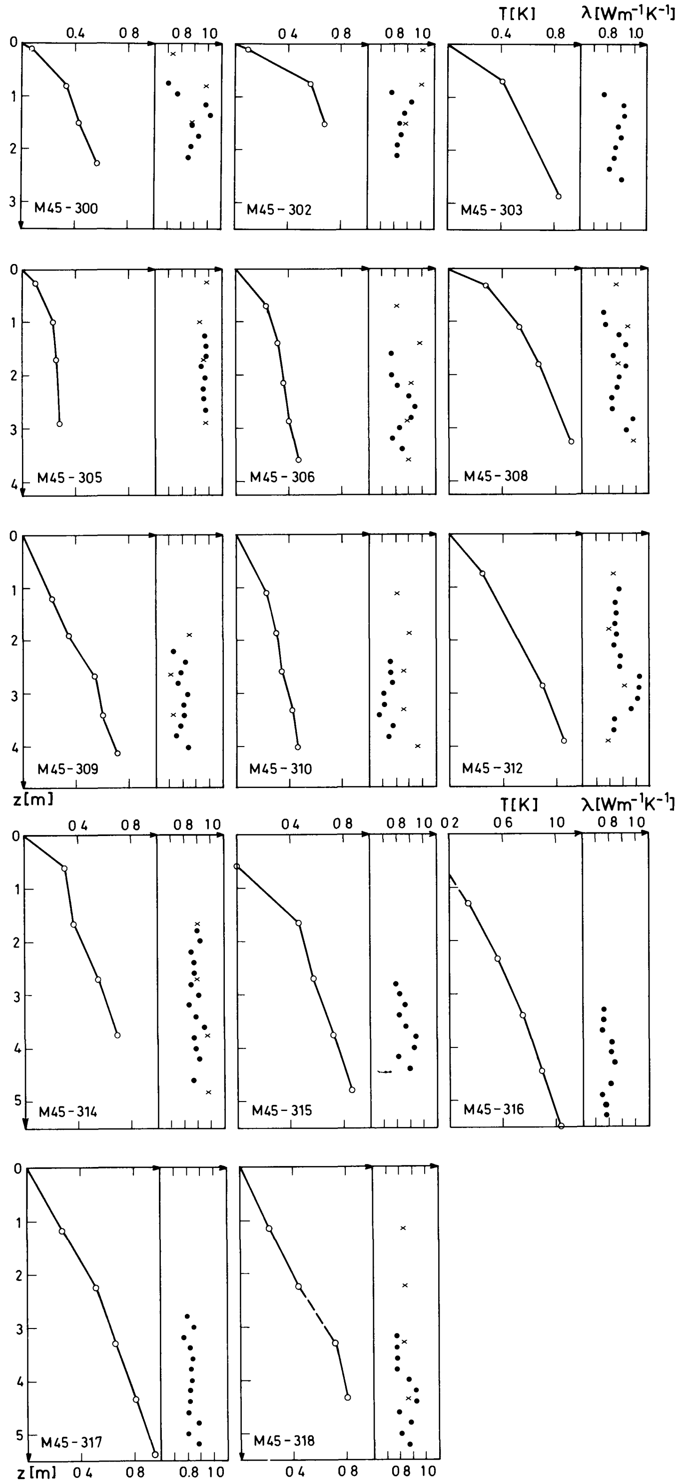

Fig. 4. Sediment temperatures and thermal conductivities obtained from in situ measurements (*crosses*) and from core samples (*dots*)

Discussion of Heat Flow Data

The heat flow values were calculated by multiplying the mean of the temperature differences between each pair of thermistors from the nearly linear segment of the temperature curve and the mean thermal conductivity. In the error range given for each gradient an error of 0.01° C is taken into account for the temperature measurements. In general, the instrumental error does not exceed 10% of the actual value. No corrections of the temperature gradient were applied with respect to the influence of topography and sedimentation. The estimated sedimentation rate is about 0.05 cm/a which would reduce the heat flow at least by 5% (Kappelmeyer and Haenel, 1974). In Table 1 the computed heat flow values are listed together with the station positions, bottom water temperature, conductivity, temperature gradient, and the deviation from vertical of the probe in the sediments. An estimation of the local environment according to the classification given by Sclater et al. (1976) is shown in the last column of Table 1.

The heat flow values vary between 27.9 and 167.2 mW m^{-2} This large scatter is not very surprising in the vicinity of active sea floor spreading centers. Today it is generally accepted that the variation of heat flow in those regions is due to hydrothermal circulation within a porous basement and/or through fissures and faults as was shown by investigations of, e.g., Pálmason (1967), Lister (1972), Williams et al. (1974). There is a high probability that the extremely low heat flow of 27.9 mW m^{-2} at station 305 is strongly affected by thin sedimentary cover and nearby outcropping basement. Obviously, a similar situation cannot account for the relatively low values at stations 306 and 310. It is likely that the small gradients result from downward migrating water into the sediments, but a reliable answer to this question is difficult to obtain in the absence of a temperature log down to at least several tenths of meter, showing a clear deviation from linearity. It is worth noting that the values at stations 314 through 318 are very uniform (110 ± 7 mW m^{-2}). Two heat flow values, reported from Langseth and Zielinski (1974) for the same distance range and close to the profile (s. Fig. 1), support this result. This may be explained by the rapidly increasing thickness of the sediments beyond the distance of 105 km (Fig. 3). Lister (1972) and

Table 1. Heat flow data from Reykjanes Ridge

Station No.	Latitude (N)	Longitude (W)	Depth (m)	Bottom Water (Temperature °C)	Mean Cond. (W m^{-1} K^{-1})	Mean Temp. Grad. (K m^{-1})	Mean Heat Flow (mW m^{-2})	Incl.	Station evaluation
M45–300	62° 3.5′	25° 15.2′	1,418	3.63	0.88 ± 0.12	0.11 ± 0.03	96.8 ± 37.4	4°	C
302	62° 6.2′	25° 21.0′	1,544	3.63	0.88 ± 0.08	0.14 ± 0.04	123.2 ± 47.0	3°	C
303	62° 7.1′	25° 23.1′	1,392	3.63	0.88 ± 0.07	0.19 ± 0.05	167.2 ± 21	5°	C
305	61° 43.2′	24° 32.5′	1,557	(3.0)	0.93 ± 0.06	0.03 ± 0.01	27.9 ± 12.8	6°	C
306	61° 49.4′	24° 46.5′	1,438	3.43	0.87 ± 0.07	0.07 ± 0.01	63.6 ± 18.4	7°	B
308	61° 51.0′	24° 49.2′	1,468	3.43	0.89 ± 0.07	0.18 ± 0.01	160.2 ± 24.9	5°	B
309	61° 55.2′	24° 59.0′	1,498	3.50	0.78 ± 0.06	0.20 ± 0.06	157.1 ± 41.7	4°	B
310	61° 56.2′	25° 0.9′	1,495	3.50	0.81 ± 0.06	0.05 ± 0.03	60.7 ± 40.2	4°	B
312	61° 45.0′	24° 36.1′	1,379	3.65	0.91 ± 0.09	0.14 ± 0.03	127.4 ± 40.0	3°	C
314	61° 36.2′	24° 16.8′	1,660	(2.7)	0.90 ± 0.04	0.13 ± 0.05	116-4 ± 49.4	3°	B
315	61° 35.3′	24° 13.8′	1,615	(2.7)	0.86 ± 0.05	0.12 ± 0.02	103.8 ± 12.4	4°	A
316	60° 55.6′	22° 46.4′	1,934	2.30	0.79 ± 0.03	0.13 ± 0.01	105.3 ± 7.5	3°	A
317	60° 55.2′	22° 45.5′	1,943	2.28	0.83 ± 0.03	0.14 ± 0.01	118.6 ± 4.6	5°	A
318	60° 54.3′	22° 43.4′	1,949	2.33	0.84 ± 0.04	0.13 ± 0.08	104.2 ± 67	3°	A

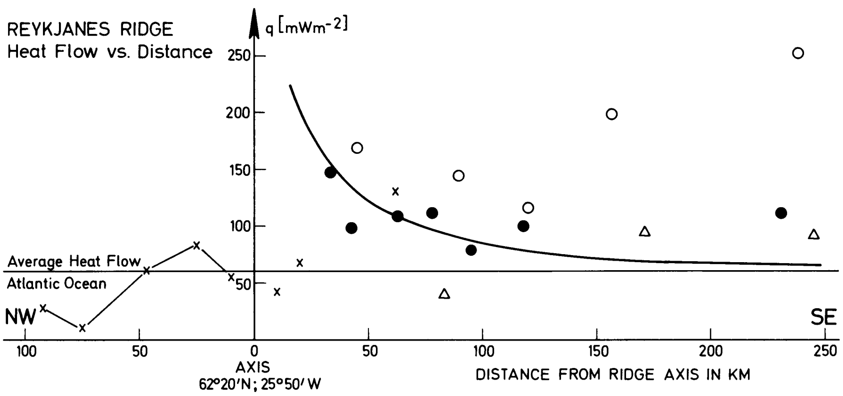

Fig. 5. Heat flow distribution versus distance. *Heavy dots* represent mean values of a group of stations, for the other symbols refer to Fig. 1. The heat flow calculated for a cooling plate model described in the text is indicated by the *solid curve*

Sclater et al. (1974) suggested that a thick and impermeable sedimentary cover seals the basement decreasing the influence of hydrothermal circulation.

A more representative heat flow for a small area is given if the values of closely spaced stations are averaged. Up to three stations, only 1.5 to 3 km apart (s. Fig. 1), are grouped together and their mean heat flow value is plotted versus distance of the ridge axis in Fig. 5 (heavy dots). Within the variation of 77 to 145 mW m^{-2} the values are in agreement with previously published results (Talwani et al., 1971, Langseth and Zielinski, 1974) in the same region. Considerably higher heat flow values from stations closer to Iceland (s. Fig. 1) are reported by Scheljagin et al. (1973). A mean heat flow of about 83 mW m^{-2} is given by Sclater and Crowe (1979), (John G. Sclater, personal communication) along magnetic anomaly 13.

Parker and Oldenburg (1973) presented a thermal model of ocean ridges which overcame the problem of infinite heat flow at the ridge crest as inherent in the cooling plate model proposed by McKenzie (1967). Yet the predicted heat flow does not differ markedly from that of McKenzie's model except very near the ridge crest. Therefore, and with regard to the scatter of the presented heat flow values, these data are compared with a calculated heat flow distribution from the simple cooling plate model.

The thickness of the lithosphere increases as the square root of crustal age (e.g., Parker and Oldenburg, 1973; Parsons and Sclater, 1977). Using the expression given by Parker and Oldenburg (1973) a value of 47 km results for the thickness of the lithosphere at a distance of 240 km (about 25 Ma) from the ridge crest. From a half-width of 60 km of the geothermal anomaly of the Mid-Atlantic Ridge (Lee and Uyeda, 1965), McKenzie (1967) deduced a lithosphere thickness of about 60 km. Therefore, the value of 50 km may be taken as an approximate thickness of the lithosphere in the observed range of distances. The average heat flow of the Atlantic ocean amounts to about 60 mW m^{-2} (Von Herzen and Lee, 1969). Based upon this value and on assuming a thermal conductivity of 4.19 W m^{-1} K^{-1} a temperature of 715° C results for the depth of 50 km. The boundary condition both at the lower side and the side where new material is accreted was taken to be of a constant temperature of 715° C. The spreading rate was taken to be 1 cm/a (Vine, 1966; Fleischer, 1974). The computed heat flow distribution as a function of distance from the ridge axis is shown by the solid curve in Fig. 5. The tendency of the few data to increase with decreasing distance from the ridge axis, as suggested by the theoretical heat flow curve, should be considered very carefully. The uniform values beyond the distance of 170 km, except those reported by Scheljagin et al. (1973), are approximately 30% above the computed heat flow. As discussed above, it is not likely that the observed gradients are too low (because of the relatively low depth of penetration) and, therefore, smaller heat flow values can be excluded. If the transport of heat is governed by pure conductivity only, a high heat flow had to be explained by a less thick lithosphere and/or high temperature at the lower boundary of the lithosphere, neglecting any radioactive heat sources within the lithosphere.

More reliable temperatures at the lower boundary of the lithosphere are obtained if the observations of heat flow are made at distances several times greater than the halfwidth of the geothermal anomaly (McKenzie, 1967), since it is in this region that the heat flow depends strongly on the lower boundary condition. The length of 240 km of the profile hardly satisfies this condition. But taking into account two previously reported values of 92 and 95 mW m^{-2} (Langseth and Zielinski, 1974) the profile is extended to a length of 340 km, thus permitting a reasonable estimate of the temperature at the lower plate boundary. If the average heat flow is increased by 30% to 78 mW m^{-2} and the thermal conductivity is that given above, a temperature of 930° C is required at a depth of 50 km. This temperature is somewhat lower than the temperature of 1,000° to 1,100° C deduced from a low-resistivity layer in a depth range of 12 to 22 km beneath Iceland (Beblo and Björnsson, 1978). According to the results of geochemical investigations along the Reykjanes ridge, Schilling (1973) proposed a hot mantle plume mixing model in the Iceland-Reykjanes ridge region. In the model temperatures vary between 1,200° C and 1,300° C at a depth range of 50 to 60 km, above which partial melting rapidly increases. Therefore, the value of 930° C may be regarded as the minimum temperature at a depth of 50 km in the region east of the Reykjanes ridge.

Mid-oceanic ridges show in several ways a more or less symmetric structure. A comparison of the heat flow values of the regions east and west to the ridge indicates an asymmetric thermal behaviour (Fig. 5). An average heat flow of 93 mW m^{-2} results for the region east of the ridge being nearly twice the heat flow of the western region. The values given by Scheljagin et al. (1973) are not included.

Vogt (1971) suggested a southwestward asthenosphere flow away from the hot spot Iceland. This suggestion received some support from the hot mantle plume mixing model, proposed by Schilling (1973). Although such a flow would plausibly explain high temperatures in the upper mantle south of Iceland, the available heat flow observations cannot resolve any details or even uniquely determine the sources.

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