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VI.3 3-D seismic mapping of the Iceland hot spot

An efficient way of mapping upper mantle structural heterogeneities is that of using the ACH-method (Aki et al, 1977) for inversion of travel time observations from a seismic array or similar kinds of seismograph networks. In the latter cases, the ACH-method has been adapted to handle observations from large-scale networks (aperture ~ 10 deg) in which cases the necessary data easily can be retrieved from the ISC-bulletin tapes (e.g., see Hovland et al, 1981; Gubbins, 1981). This approach for mapping the extent of lateral heterogeneities in the upper mantle has proved to be successful as compared to surface wave dispersion analysis.

In this section we describe an attempt to map seismically the deep structure beneath Iceland using available P-travel time residuals from the local seismograph network (see Fig. VI.3.1). This problem is an interesting one from a tectonic point of view, as evidence for deep-seated roots (down to 300-400 km) of hypothesized hot spots like Iceland and Hawaii have not been presented to our knowledge. Indeed, recently Anderson (1981) argued that the depth extent of hot spots should be confined to the uppermost 200 km (lithosphere) of the mantle.

Data and method of analysis

The travel time observations used in this inversion experiment were taken (read) from original seismogram records of the Icelandic network comprising altogether 39 stations. In the time interval 1974-80 the total number of events available for analysis was about 160, out of which 61 were found useful. The event selection criteria imposed were that a minimum of 5 stations exhibited reasonably clear P-wave recordings, and besides that the azimuth/distance distribution was reasonably homogeneous. Excessive errors in P-wave onset readings were attempted avoided by waveform correlation between recording stations for each event subject to analysis. Also, for each event the network average residual was estimated and subtracted from the individual observations.

Travel time anomalies are caused by velocity variations within an a priori confined volume immediately beneath the station network (see Fig. VI.3.1), and thus are related to departures from standard earth models. The surface expression of this volume is marked in Fig. VI.3.2 and extends to a depth of 375 km. Also, the velocity structure is represented by a smooth cubic interpolatin between slowness values on a three-dimensional grid of 4x6x6 knots. This means that the upper mantle beneath the seismograph network is subdivided into 4 levels (0-75 km, 75-175 km, 175-275 km, 275-375 km) with slowness estimates at individual grids of 6x6 knots). Now the basis for linear inversion of travel time data is Fermat's principle stating that the variation in the travel time caused by small change in the ray path is zero to the first order. From this we may formulate a linearized relationship between travel time residuals (ΔT_{ij}) and velocity variations, namely:

$$\Delta T_{ij} = \int_{A_j}^{B_i} \delta s \, d\ell$$

where $s = 1/v$ is slowness or reciprocal velocity in s km^{-1}. A_j is the j-th receiver while B_i is the i-th source. This equation constitutes the very basis for linearized inversion of observed travel time residuals (the ACH-method and variants hereof) and for details here in this particular case, reference is made to Tryggvason (1981).

The validity of the linearity assumption above is based on the assumption that the non-linear terms are negligible, namely, the contributions due to change in velocity along the initial ray path and the effect of the change in ray path in the initial medium as detailed by Thomson and Gubbins (1981). We did indeed check the effects of these non-linear terms and found them ignorable.

Results

The anomaly maps of Figs. VI.3.2 represent the estimated P-wave fractional velocity anomalies for Iceland and adjacent areas. For details on the particular 'standard earth' model used, resolution and standard errors of the estimated velocity anomalies, etc., reference is made to Tryggvason (1981). It suffices to remark that for knots with resolution less than 0.4 and/or standard errors larger than the anomaly itself the corresponding knot fractional velocity estimates are not considered significant.

In layer 1 (Fig. VI.3.2a), which represents the uppermost 75 km, a broad and dominant low velocity zone is extending from the Tjörnes fracture zone (66.5°N, 16-19°W) north of Iceland, southward to the Krafla area (65.5°N, 17°W) and then west-southwestward in direction to the Snæfellsnes area (65°N, 19°W). This low coincides with a major part of the neovolcanic zone in northern and central Iceland together with late Quaternary and early Tertiary areas west of it. Pronounced low velocity values are tied to the Tjörnes shearing zone and the active Krafla volcanic area. Offshore there are only few significant anomalies and all of them represent a continuation of the pronounced high velocity regions in the southeastern and northwestern parts of the country. The grid points within Iceland itself are well resolved, that is, the resolution is around 80% and up to 90% in southern Iceland, which reflects a denser station network there. The standard error estimates for these most significant grid point values are around 1% relative velocity change. The capital letters A-A' and B-B' together with the heavy arrows indicate vertical cross-sections through the model box (Fig. VI.3.2).

In layer 2 (Fig. VI.3.2b), which ranges from 75 to 175 km depth, the low velocity zone is shifted southeastward compared to layer 1 and with the strongest anomalies south of Reykjanes (Keflavik), beneath the Hekla area (65°N, 16°W) and northeast of Kverkfjöll. A continuation of the high in the northwestern part of the country is obvious. In addition, two highs are prominent, namely, south of Iceland and northeast of Iceland. These areas are poorly sampled in layer 1, so a comparison between the two layers is not possible there.

For layer 3 (Fig. VI.3.2c), depth range 175-275 km, somewhat different features appear compared to the two uppermost layers. A significant low is covering an area south of the Tjörnes fracture zone and extending westward to the southwest of the Kolbeinsey ridge. Another low velocity zone, but a weaker one, is underneath central and southwest Iceland. The resolution values are even higher here than in layer two, most of them around 0.9 and with accompanying standard errors down to 0.8%.

Layer 4 (Fig. VI.3.2d) covering the depth range of 275-375 km, is marked by a prominent velocity low beneath central Iceland. This is a relatively broad area with maximal east-west and north-south extension around 300 and 200 km respectively. The resolution and standard error estimates are about the same as for layer 3. For the lower layers most of the edge points are sufficiently well resolved. Significant anomaly contours can therefore be drawn quite to the edge of the box, as is the case for the highs around the country in layers 3 and 4.

We made two vertical cross-sections (Figs. VI.3.2) of the above four-layer anomaly patterns. The sections visualize the main lows and highs of all four layers, indicating an anomalous mantle beneath Iceland. Below circa 250 km depth of central Iceland a dominant low velocity zone might indicate a mantle plume or a so-called hot spot. Another strong low in the uppermost 70-80 km is beneath almost the whole of Iceland except for the periphery of the oldest rocks mainly in the SE and NW parts of the country. A broad 'transition zone' of relatively low velocities including a few strong low velocity pockets (see layers 2,3 of the anomaly maps) is clearly connecting the two major lows.

Discussion

To our knowledge, this is the very first time a comprehensive 3-D mapping of the Icelandic rift zone has been undertaken. Of particular interest is that the surface rift manifestations are reflected at depths of the order of 350 km as discussed by Tryggvason (1981). It is here tempting to draw a parallel to the NORSAR siting area which coincides with an ancient (Permian) aborted rift zone. Also here there are strong indications that the relatively large time and amplitude anomalies observed are caused by heterogeneities of a predominantly vertical extent (e.g., see Christoffersson and Husebye, 1979; Haddon and Husebye, 1978; Thomson and Gubbins, 1982; and Troitskiy et al, 1981).

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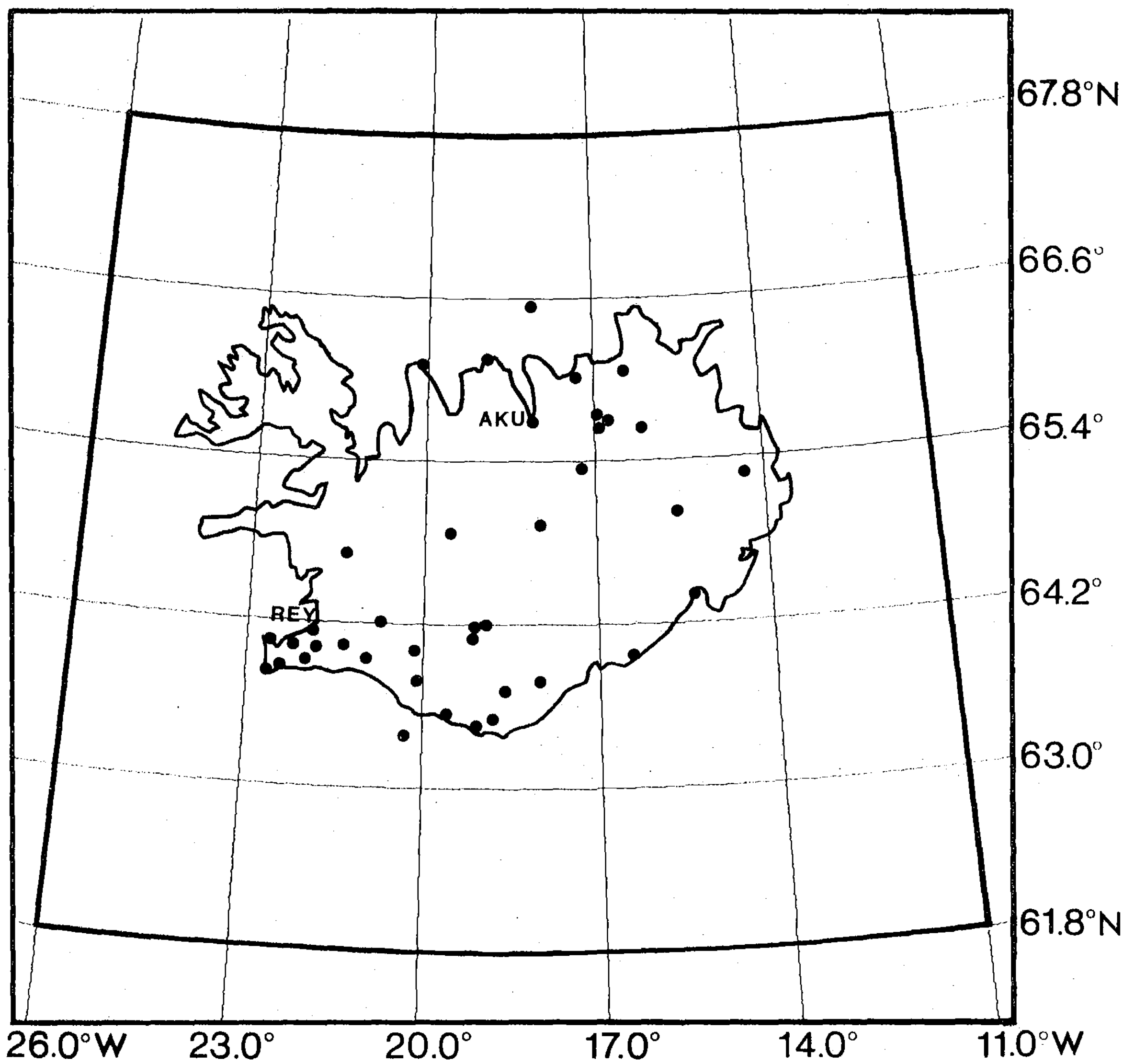

Fig. VI.3.1 The Icelandic seismograph network of 39 seismograph stations used in analysis. The model box is circumscribed with heavy lines. The latitude/longitude grid system intersections correspond to the model knots used in the time residual inversion as detailed in Tryggvason (1981).

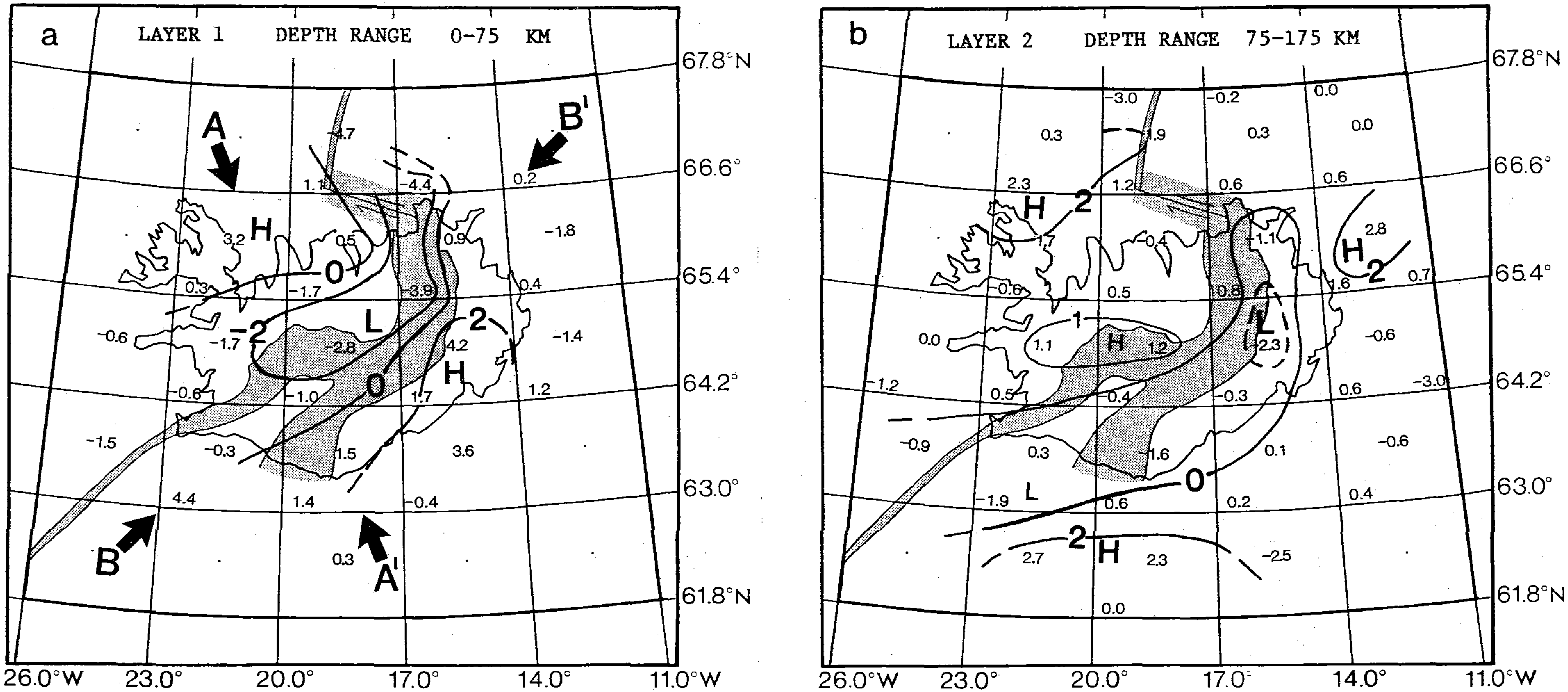

Fig. VI.3.2 (a) Velocity perturbations (in per cent) for layer 1. Areas of high and low velocities are indicated by capital letters H and L. The captial letters A-A' and B-B' together with the heavy arrows indicate vertical cross-sections through the model box as shown in Fig. VI.3.3. Resolution and standard errors for all knots are detailed in Tryggvason (1981). Figs. VI.3.1b,c & d captions as in (a) applied to the respective layers.

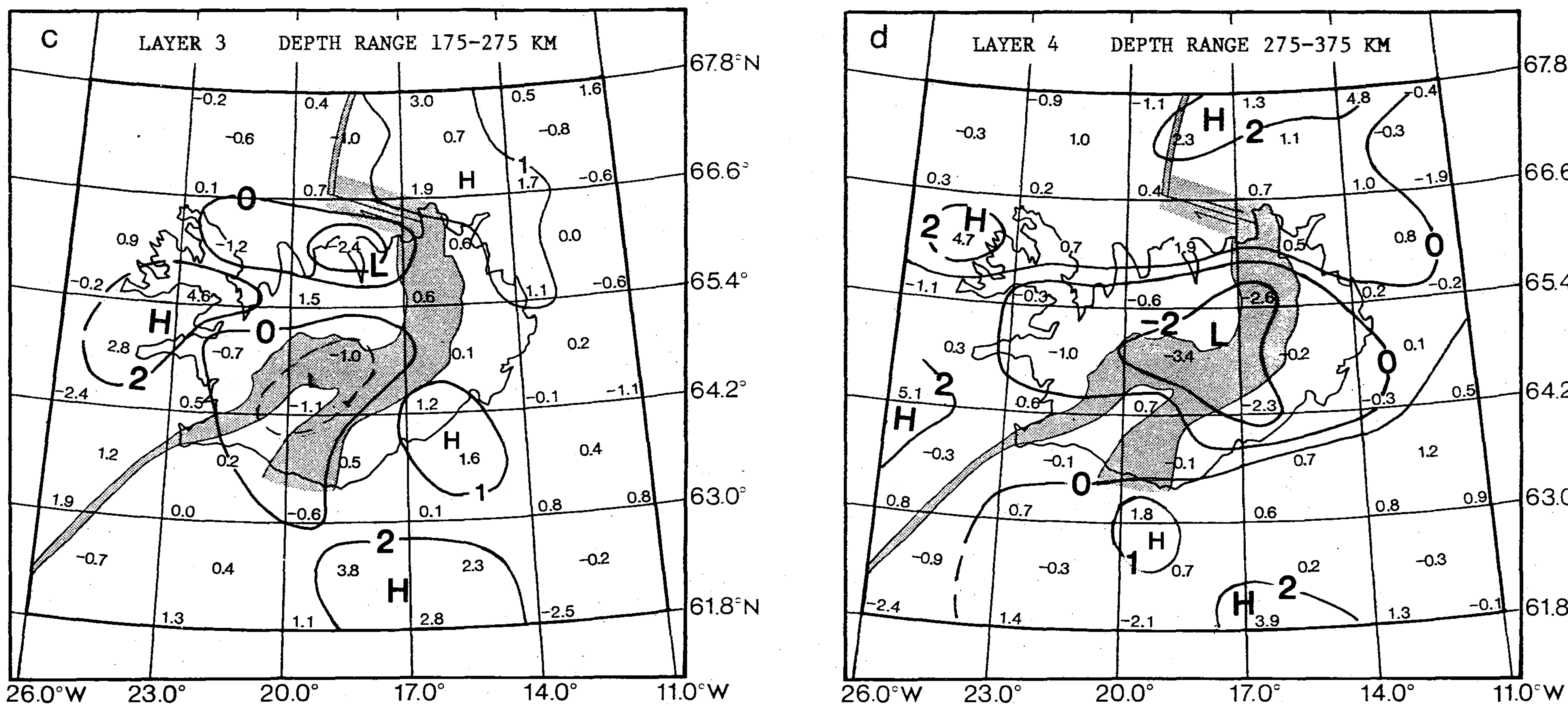

Fig. VI. 3.2 (continued).

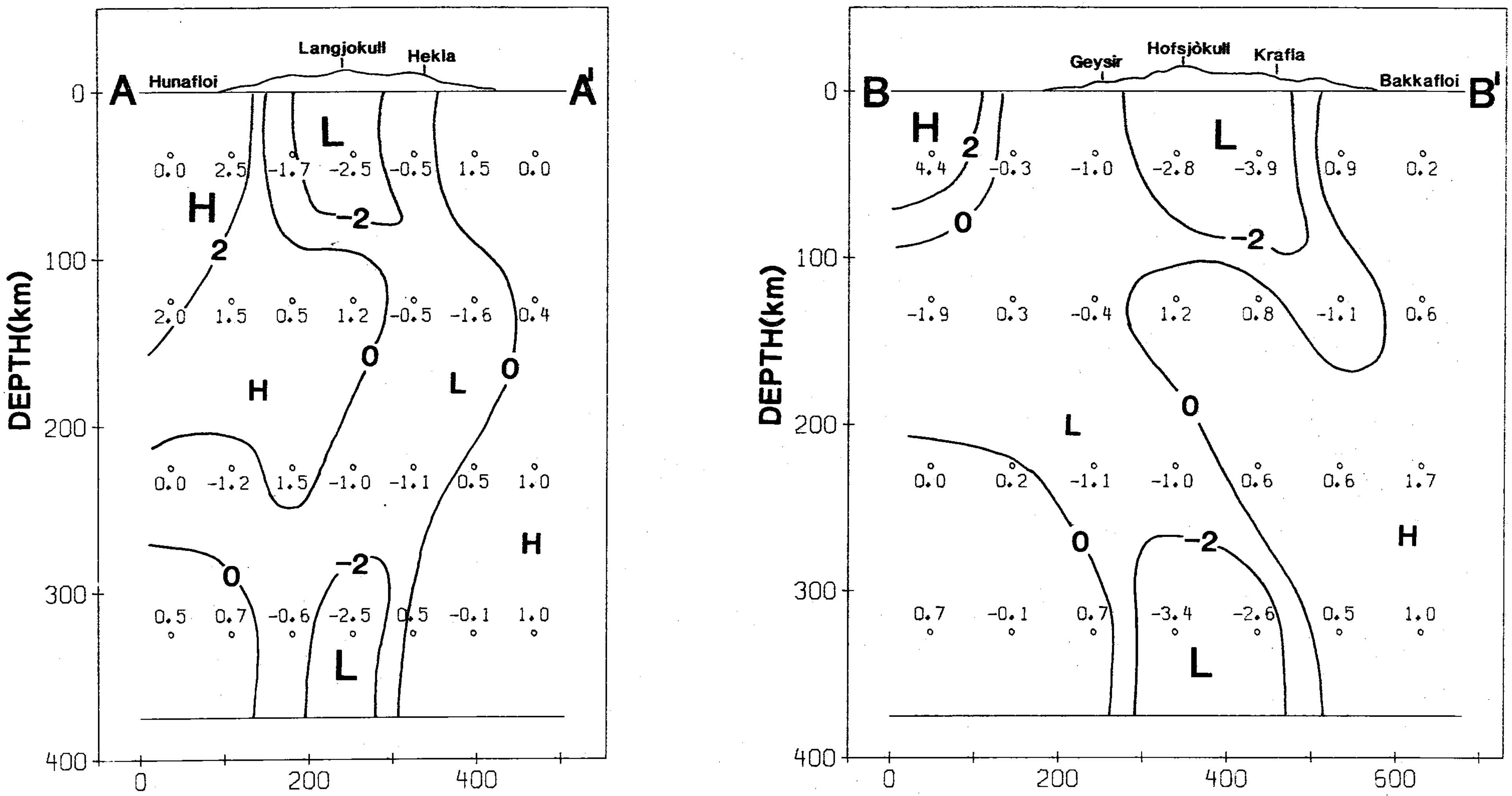

Fig. VI.3.3 The cross-sections span the latitude range from 63.°N to 66.0°N, from NW to SE (A-A') and from SW to NE (B-B'). The depth to distance scale is 2:1.