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ANALYSIS OF EVENT LOCATION ERRORS USING ARRAYS IN SCANDINAVIA

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ABSTRACT

Systematical teleseismic P-wave slowness anomalies up to 1 sec/deg are observed at three array stations in Fennoscandia, NORSAR in Norway, Hagfors Observatory in central Sweden and the telerecording station at Helsinki in Finland. The slowness anomalies are smallest at Helsinki and largest at Hagfors. Resulting event mislocations would exceed ten degrees at certain regions if no corrections were used. It is shown that an integration over the horizontal velocity gradient along the ray path near the receiver gives the slowness anomaly caused by that gradient. A given horizontal gradient produces an order of magnitude smaller mislocation when situated near the source than when situated near the receiver. It is concluded that a major part of the observed slowness anomalies can be explained by lateral changes in structures in the upper mantle and the crust under the arrays. In particular, an increase of the average vertical velocity in the upper mantle under the Scandinavian peninsula in the direction from oceanic towards continental areas is required by the observed slowness anomalies and by the travel time residuals of the conventional stations in the same region.

BASIC CONSIDERATIONS

The travel time residual of the i-th recording point in an array relative to a selected reference point from the j-th event is

$$r_{ij} = (\text{absolute residual})_{ij} - (\text{absolute residual})_{oj}$$

where the index o refers to the reference point. The relative residual r_{ij} can be expressed as

$$r_{ij} = H_i \cdot \Delta S_j + c_{ij} \tag{1}$$

where H_i is the location vector of the i-th recording point relative to the reference point, ΔS_j is the slowness anomaly vector for the j-th event and c_{ij} is a particular time delay caused by the structure in the vicinity of the i-th recording point. For the NORSAR array the r_{ij} have a range of one second. If the ΔS_j are assumed to be zero, the c_{ij} also have a range of one second. To explain so large relative time residuals the depth of the major velocity contrast, the crust-mantle interface, should vary by as much as 25 km (Nuttli and Bolt, 1969). This is not true as shown by crustal refraction work in the NORSAR siting area (Kanestrøm and Haugland, 1971). The conclusion is that the P-waves have large slowness anomalies already before arriving to the crust.

To study how slowness anomalies might be produced, the ray slowness vector S together with its derivatives is used (Julian, 1970);

$$S_i = \frac{\cos\alpha_i}{v} \tag{2}$$

where v is the local wave velocity and the $\cos\alpha_i$ are the ray direction cosines. According to Julian the time derivative of slowness is

$$\dot{S} = \frac{\text{grad } v}{v} \tag{3}$$

from which a change of a slowness component can be computed using the integration

$$\Delta S_i = \int_{t_1}^{t_2} \frac{D_i v}{v} dt \tag{4}$$

Approximate computations were made using a model anomaly having a horizontal velocity gradient in the depth range 300 - 500 km with a total lateral velocity change of 10%. The resulting location errors were of the order of 50 km for the anomaly under a surface source and of the order of 10% of the event-receiver distance for the anomaly under the receiver. The use of an incorrect standard earth model without the presence of horizontal velocity gradients cannot produce systematical errors in observed azimuths, also errors as large as 1 sec/deg are contrary to the idea of the accuracy of the standard earth models.

It is concluded that appreciable horizontal velocity gradients not only close to the source are required by the large slowness anomalies observed.

OBSERVATIONS IN FENNOSCANDIA

Sites of the arrays used in this study are shown in Fig 1. The Norwegian Seismic Array or NORSAR, operated jointly by the Norwegian Council for Scientific and Industrial Research and the Advanced Research Projects Agency, USA, is situated in southeastern Norway.

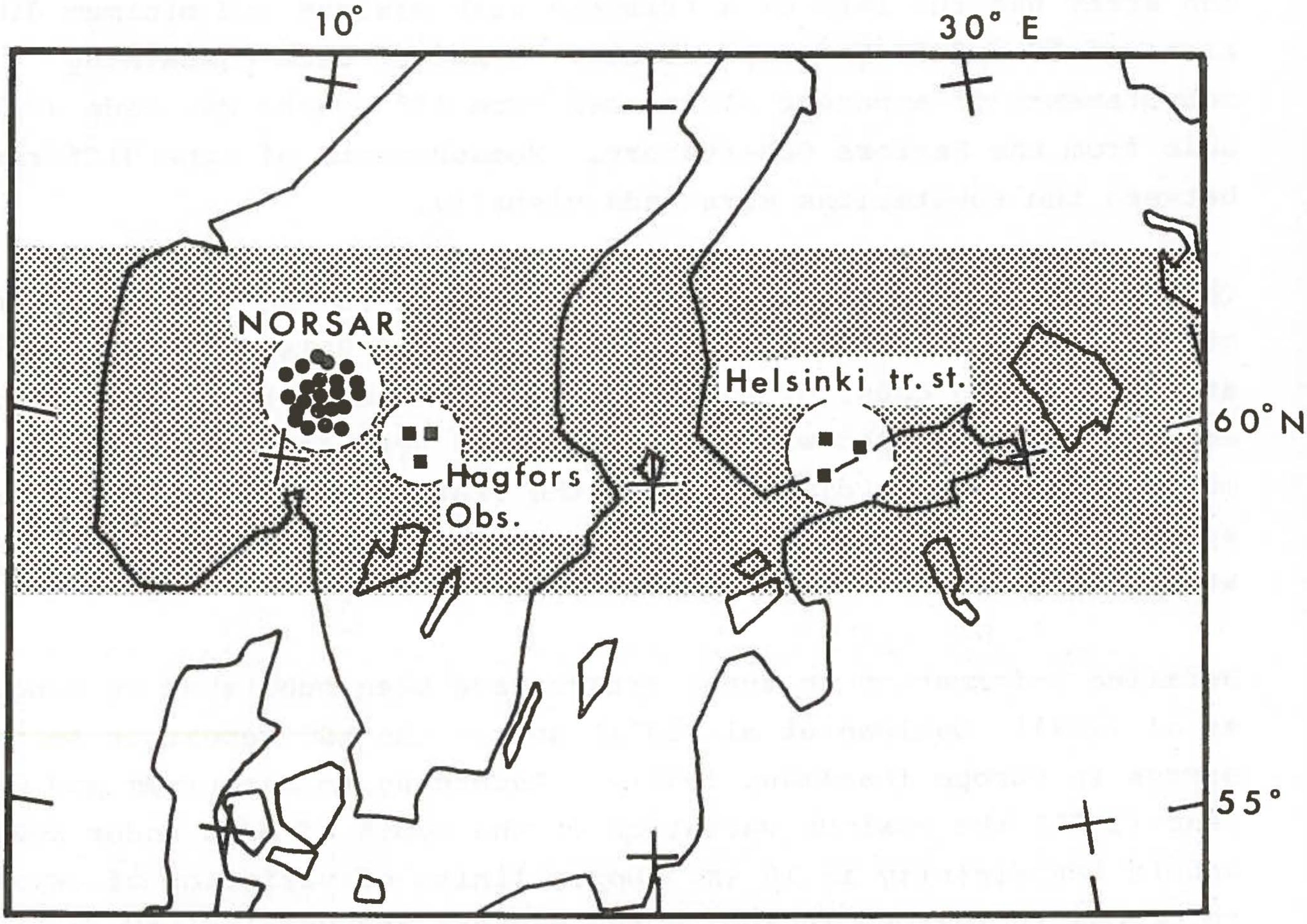

Fig 1. The three arrays NORSAR, Hagfors Observatory and Helsinki telerecording station. Circles denote subarrays and squares denote substations.

It has 22 subarrays each with six short period seismometers and a three component set of long period seismometers. The diameter of the subarrays is 7-8 km, while that of the array itself is around 110 km. Data used in this study was recorded during the spring and summer of 1971 when the array was already fully operational. The apparent slownesses were measured from a plane wave fit to the arrival times of the 22 subarray beams. The time differences between the signals from the subarrays were measured using an iterative crosscorrelation procedure. The data base comprises 132 events. Data from an earlier interim phase of operation, during which seismic signals were recorded at 18 center seismometers in different subarrays, was compared to the present data and the pattern of mislocation of events was found to be similar.

The Hagfors Observatory, operated by the Research Institute of National Defence, Sweden, is situated in western Sweden and consists of three substations having long and short period seismometers. The array has the form of a triangle with maximum and minimum dimensions of 50 and 30 km respectively. A set of data containing measurements of apparent slownesses from 158 events was made available from the Hagfors Observatory. Measurements of time differences between the substations were made visually.

The Helsinki telerecording station belonging to the station network of the Institute of Seismology, University of Helsinki, is situated at the southern coast of Finland. It consists of three substations, each of them equipped with a short period vertical seismometer. The maximum and minimum dimensions of the station are 70 and 40 km, respectively. Slownesses measured from 201 events are used in this study. Time differences were measured visually.

Detailed information on these arrays have been published by Bungum et al (1971), Dahlman et al (1971) and in the ESC report on seismic arrays in Europe (Dahlman, 1971). According to Kanestrøm and Haugland (1971) the maximum variation of the depth of Moho under NORSAR and in its vicinity is 10 km. Upper limits of variation of crustal thickness in regions including the Hagfors and Helsinki arrays have been estimated from the range of P_n residuals on refraction lines crossing these arrays. Results of Dahlman et al (1970) and Luosto (1967) show the peak-to-peak variations of the depth of Moho on refraction lines crossing the Hagfors and Helsinki arrays to be less than 5 and 8 km, respectively. These limits, being computed from the transit time of the P_n wave across the crust, are applicable for estimation of delays of teleseismic body waves due to variation of crustal structure. An increase of crustal thickness of 5 km produces delays of approximately 0.45 and 0.15 seconds for the P and the teleseismic P-wave (at the range of 60°) respectively.

Figs 2, 3 and 4 show the slowness anomalies for the three arrays NORSAR, Hagfors and Helsinki. Anomalies are represented by arrows, each starting from the slowness value computed using the location determined by NOAA and Herrin's tables (1968), and ending at the observed slowness. Outlines of continents are also shown. Coordinates are defined as x_1 positive to west, x_2 positive to south.

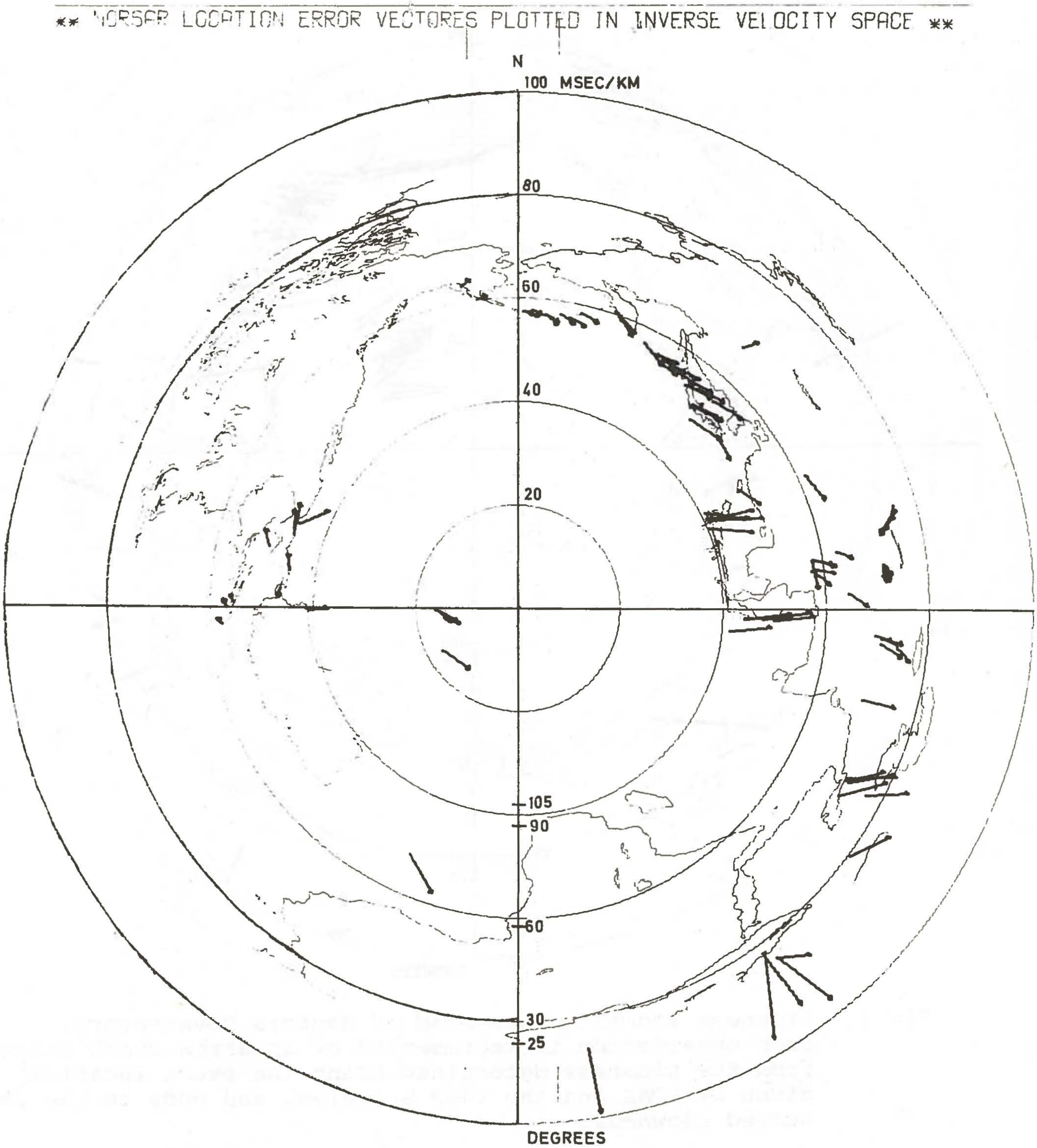

Fig 2. Slowness anomalies observed at NORSAR. Each observation is represented by an arrow which starts from the slowness determined using the event location given by NOAA and the 1968 P tables, and ends to the observed slowness.

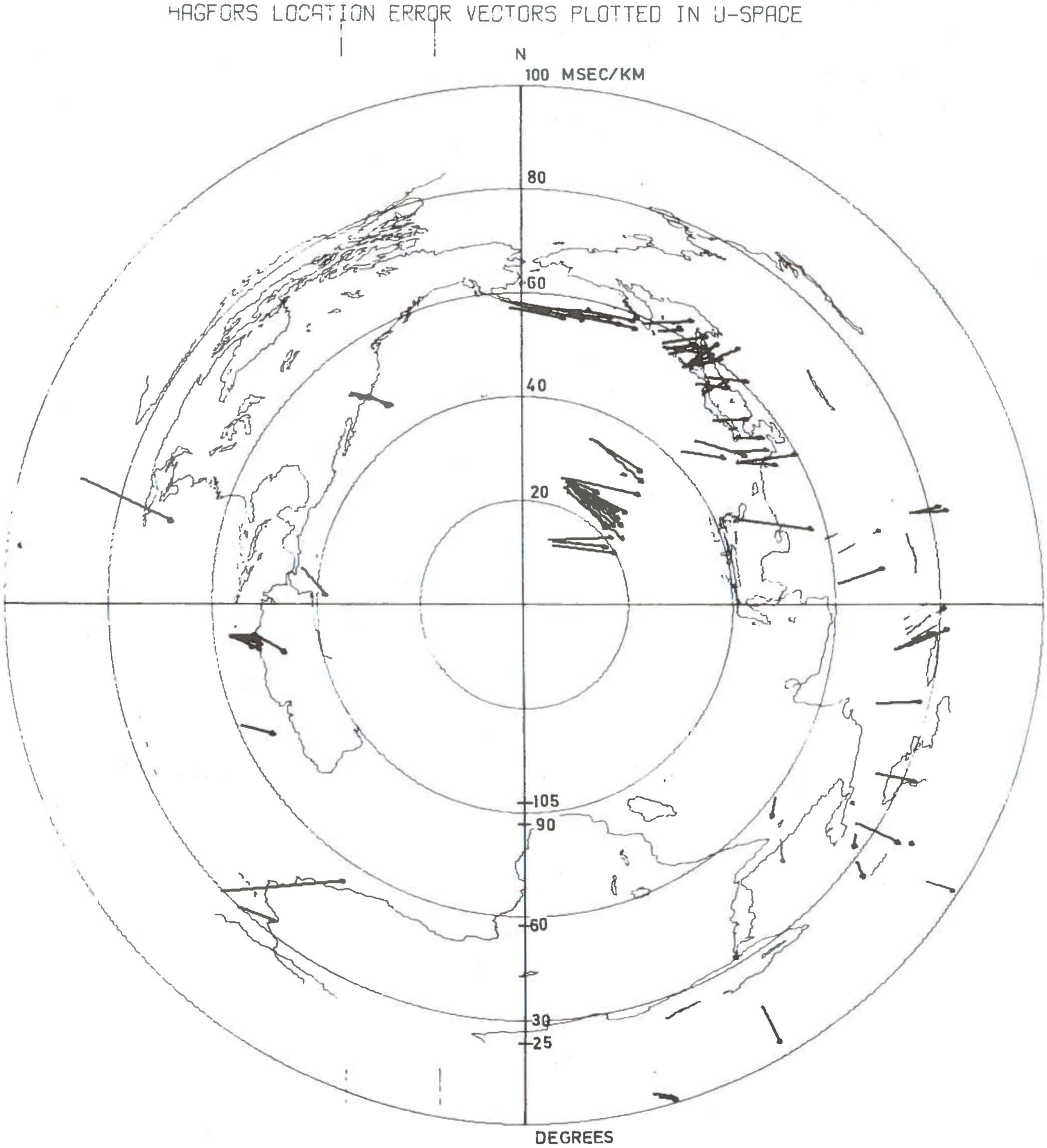

Fig 3. Slowness anomalies observed at Hagfors Observatory. Each observation is represented by an arrow which starts from the slowness determined using the event location given by NOAA and the 1968 P tables, and ends to the observed slowness.

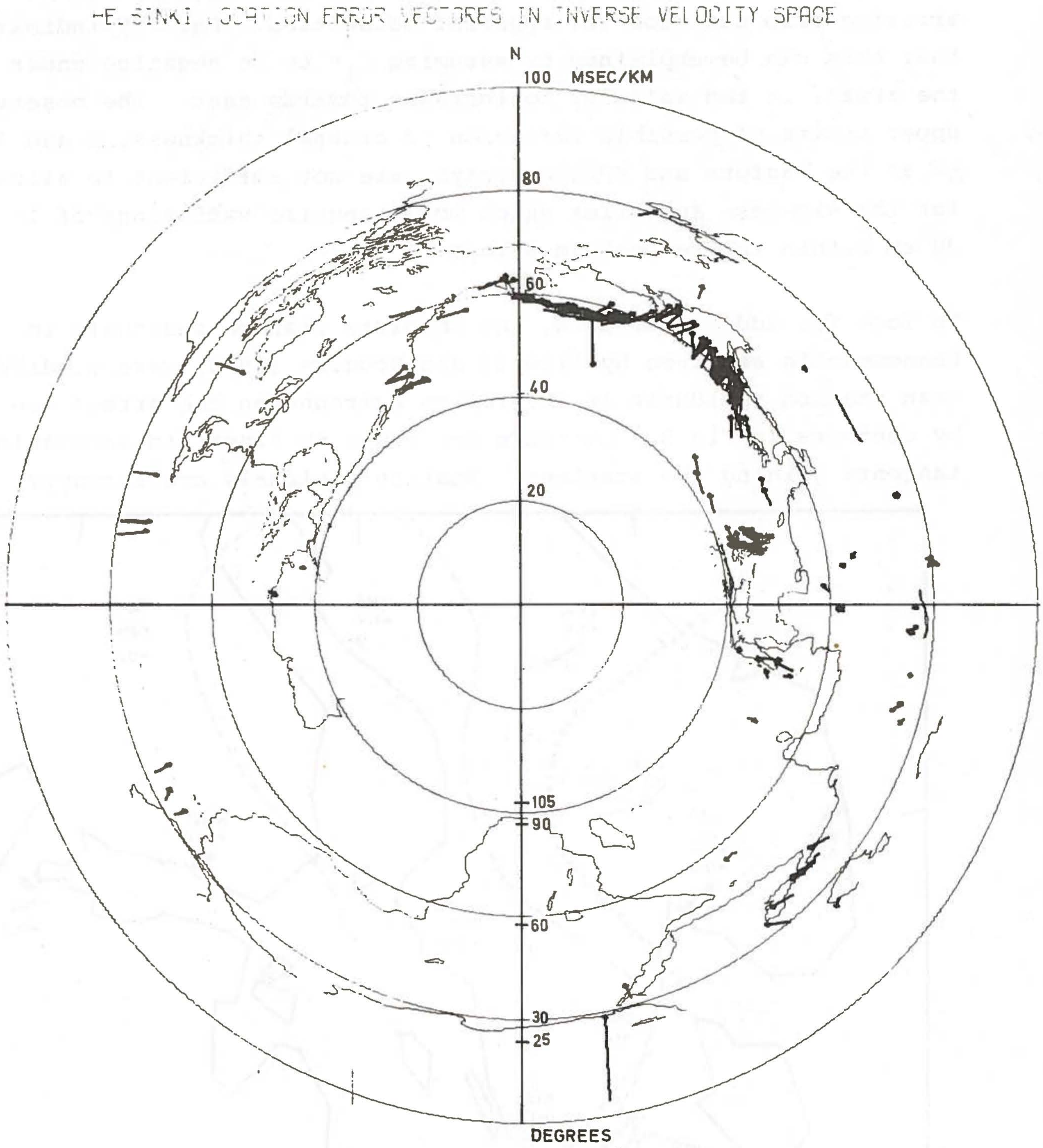

Fig 4. Slowness anomalies observed at Helsinki telerecording station. Each observation is represented by an arrow which starts from the slowness determined using the event location given by NOAA and the 1968 P tables, and ends to the observed slowness.

A dominating feature of the slowness anomalies displayed are the large negative values of the ΔS_1 component at the arrays NORSAR and Hagfors. Waves arriving from west have too high and those arriving from east too low apparent velocities. Eq. (4) indicates that this can be explained by assuming $D_1 v$ to be negative under the array, or the velocity to increase towards east. The observed upper limits of possible variation of crustal thickness, 5 and 10 km at the Hagfors and NORSAR arrays, are not sufficient to allow for the slowness anomalies which would require variations of 15 and 30 km within the respective arrays.

To look for additional data, the absolute station residuals in Fennoscandia as given by Lilwall and Douglas (1970) were studied. Mean station residuals in the region surrounding the arrays are shown by contours in Fig 5. Contours are drawn by linear interpolation on tangents joining the stations. Station residuals are strongly

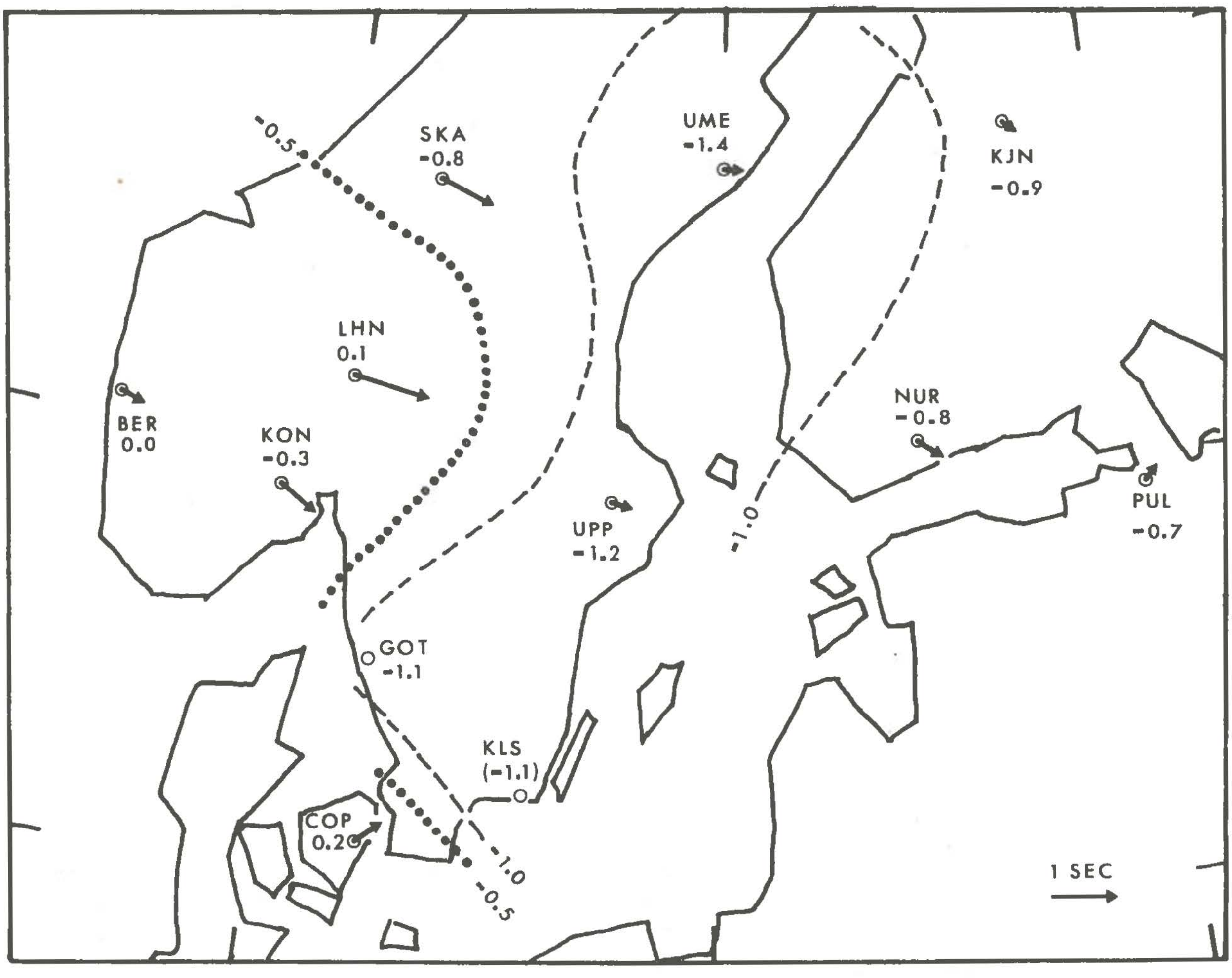

Fig 5. Observations (from Lilwall and Douglas, 1969) and contours of mean station residual in central Fennoscandia. "Earliest" direction of approach is shown for each station by a vector, length of which is proportional to the amplitude of the sine function approximating the azimuthal variation of the station residual.

negative at the coast of the Baltic sea, less negative or slightly positive in southwestern Norway and Denmark. Vectors in Fig 5 show the direction of approach of the earliest arrivals for each station. The largest azimuthal dependency of the station residual is observed at stations near the border of Sweden and southern Norway.

For all shown stations the "early" direction vectors point eastwards even though the mean station residuals have a minimum in the central region of Fennoscandia. This can be explained by the P-velocity anomalies in lower mantle observed by Husebye et al (1971).

Comparing the slowness anomalies in Figs 2, 3 and 4 to the station residuals shown in Fig 5 it is observed that:

a) The dominant direction of the slowness anomaly vectors for the arrays NORSAR and Hagfors is the same (appr. 100°) as the dominant direction of approach of the earliest arrivals to the adjacent conventional stations. According to eq. (3) a slowness anomaly vector points in the direction of the increasing velocity.

b) The mean station residual changes by one second between stations LHN and UPP, i.e. in the region covered by the NORSAR and Hagfors arrays. This will produce a slowness anomaly directed roughly towards east and with a magnitude depending on the width of the zone of change. If the zone width is 250 km a slowness anomaly of 0.004 sec/km is to be expected. Averages of the slowness anomaly components towards 100° are 0.004 and 0.007 sec/km for the NORSAR and Hagfors arrays, respectively.

In Fig 6 the observed slowness anomalies at the three arrays are given. The event base used is fairly similar for all arrays, even though it is not the same set of events. Only high-quality P-waves from events in the distance range 30 to 90 degrees are used. For the Helsinki array 90% of events produce slowness anomalies less than 0.006 sec/km, for NORSAR and Hagfors the corresponding figures are 0.009 and 0.011 sec/km, respectively. The larger slowness anomalies for arrays in Scandinavia imply the existence of larger horizontal velocity gradients there than in the vicinity of the Finnish Gulf.

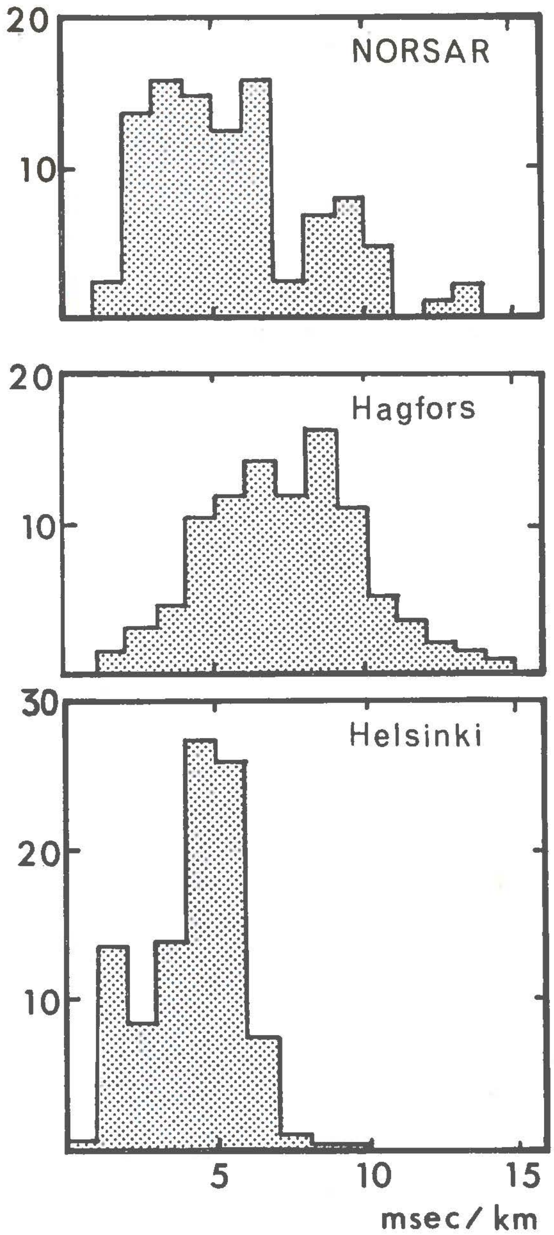

Fig 6. Size distribution of slowness anomalies of the three arrays. High quality P-waves from the distance range 30 to 90 degrees are used. Sample sizes are 88 events from NORSAR, 132 from Hagfors Observatory and 199 from the Helsinki telerecording station.

It is concluded that the increase of the average P-wave velocity from ocean towards continent in the upper mantle under the central part of the Scandinavian peninsula which manifests itself in the variation of station residuals also explains the main direction and partially the magnitudes of the slowness anomalies observed at NORSAR and Hagfors. This large-scale gradient is no longer present in the region east of Baltic sea, at distances larger than 1000 km from the Atlantic coast. However, it is recognized that additional structural irregularities are required to account for the variation of the magnitude and direction of the observed slowness anomalies. The slowness anomaly is an integration over the horizontal velocity gradient sampled by the ray path, while the station residual is an integration over the velocity anomaly along the ray path. It is reasonable that in an inhomogeneous earth the local velocity gradients indicated by the slowness anomaly can be more variable and occasionally considerably larger than the average velocity gradient over a large horizontal distance.

A MODEL OF SLOPING TRANSITION LAYER

Because of the evidence suggesting a single dominant direction of the horizontal velocity gradient in the central Scandinavian peninsula, a two-dimensional model is used to explain the observations in the region shown in Fig 8a. As the model is not unique, the purpose is only to arrive to a formal agreement between the sets of data available.

The horizontal velocity gradient is assumed to be due to a slope of the transition layer (in the general sense of Bullen's (1963) layer C) in the upper mantle. The transition layer is modeled as a single velocity step of 1.6 km/sec close to the depth of 500 km as shown in Fig 7. The updip direction is taken to be around azimuth 110°. From the slowness anomaly components perpendicular to the strike of the discontinuity, the slope of it can be computed using Snell's law at a number of points, corresponding to the crossing points of the observed rays and the base depth level of 500 km. The effect of the curvature of the earth is neglected, which produces errors less than 20%. From the station residuals the depth deviation of the discontinuity from the base level can be estimated at some points. Station residuals according to Lilwall and Douglas in the region of study are given in Table 1. Residuals have a form A + Bsin (Θ + ϕ) where Θ is the station azimuth. The early direction is ($3/4\pi$ - ϕ).

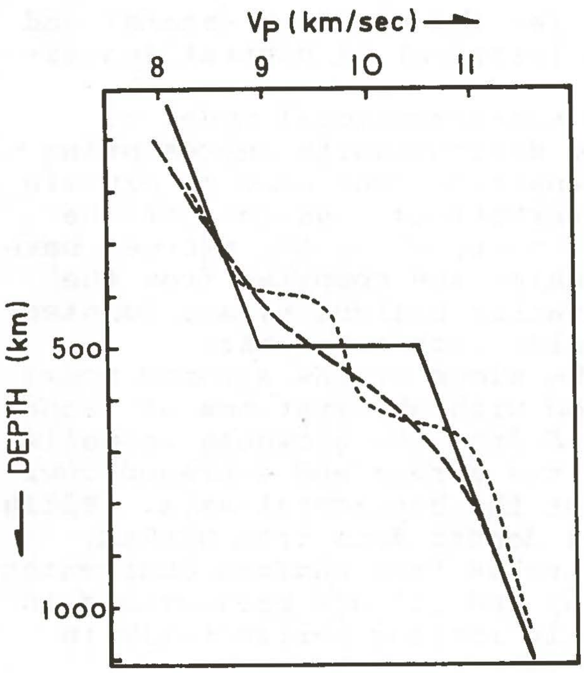

Fig 7. The simplified upper mantle velocity distribution used in model computations (continuous line). Velocity distributions of the Jeffreys-Bullen model (broken line) and model CIT 204 (stippled line) from Johnson (1967) are included for comparison. Units are km and sec.

Table 1. Station residuals in central Scandinavia (Lilwall & Douglas, 1969)

Station	A	B	the "early" direction	number of observations
BER	0.01 ± 0.38	0.36 ± 0.64	105°	21
SKA	-0.79 ± 0.43	0.86 ± 0.59	109°	21
KON	-0.29 ± 0.29	0.61 ± 0.40	126°	44
LHN	0.08 ± 0.39	1.18 ± 0.39	100°	24
UPP	-1.17 ± 0.24	0.34 ± 0.26	116°	57

The total change of the station mean residual across the peninsula, which is practically independent of possible P-velocity anomalies in the lower mantle, is 1.2 ± 0.4 sec. For the simplified upper mantle model, this means a change of depth of the discontinuity of 70 ± 23 km. In Fig 8b the shape of the discontinuity inferred mainly from the slowness anomaly data is shown. The depth relative to the base level is determined by the mean station residuals. In Fig 8c the slope of this model is compared to the slopes estimated from the observed slowness anomalies.

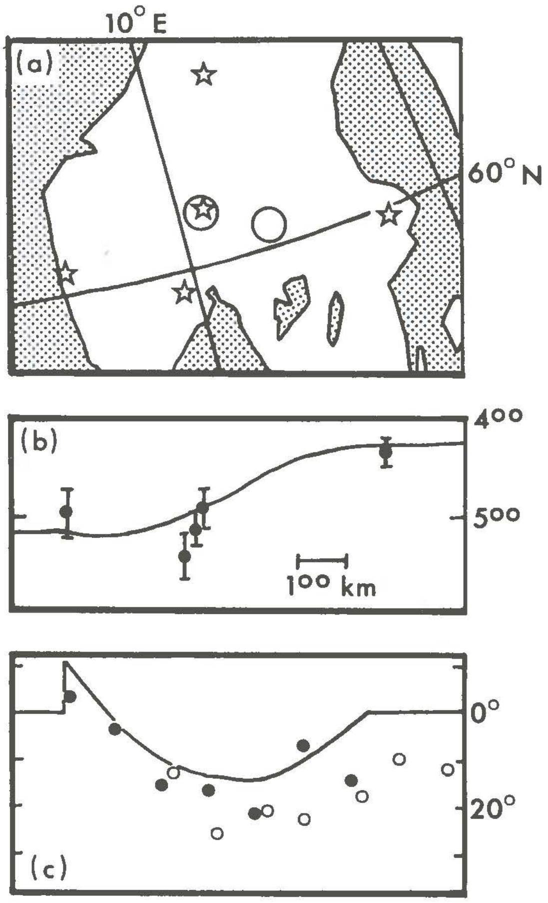

Fig 8. (a) The stations (stars) and arrays (circles) in central Scandinavia.
(b) A two-dimensional model of sloping discontinuity representing the transition zone used to explain the observations. Heights of the discontinuity above the assumed base level which are computed from the mean station residuals, are denoted by circles with error bars.
(c) The slope of the assumed model compared with observations of slope computed from the slowness anomalies at the two arrays and averaged over cells of the horizontal axis. Filled circles denote data from NORSAR, open circles from Hagfors Observatory. Figs (b) and (c) are projections to a profile running horizontally in figure (a).

The largest slowness anomalies predicted by this model occur for rays crossing the transition layer between the NORSAR and Hagfors arrays. The maximum anomaly should be 0.007 sec/km. The predicted station residuals are given in Table 2.

Table 2. Predicted station residuals

Station	A	B	"early" direction
BER	+0.3	0.1	110°
SKA, KON, LHN	-0.1	0.9	110°
UPP	-1.0	0.3	110°

The model is able to explain satisfactorily the general behaviour of the slowness anomalies observed at NORSAR and the station residuals. It is not quite sufficient for explaining the anomalies observed at Hagfors. Also it cannot and does not attempt to explain the irregular variation of magnitude and direction of the anomalies superposed on the general trend. However, after this trend is accounted for by this model, the remaining portion is much easier to explain e.g. by local variations of structure.

Some comments can be made on the uniqueness of the model. Because of the unidirectionality of the slowness anomalies and the azimuthal terms of station residuals in stations distributed over a large region, the reason behind the anomalies cannot be a local feature, e.g. a vertical velocity discontinuity, under one or both arrays. Accepting it to be a more gradual horizontal change of velocity, it cannot be restricted to a depth less than 100 km, as it is not enough space for it. Trying to push it much deeper than 500 km makes it impossible to find a simple geometry to account for both the observed mean and azimuthal station residuals.

It is concluded that the horizontal velocity increase, which not necessarily takes a form of a sloping discontinuity but could equally well be a horizontal velocity gradient in some depth range, occurs between the depths of 100 and 700 kilometers. Assuming the usual positive correlation between seismic wave velocity and density some form of compensation has to be implied to counterbalance the effects on gravity of the lateral density change.

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