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Magnetometer array studies in Finland – determination of single station transfer functions

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Abstract. In 1981 and 1982 four arrays of 30 or 31 magnetometers were operated on the Baltic Shield in central and south-eastern Finland to measure the natural magnetic field variations. These measurements were used to deduce some information about the lateral variation of the electrical conductivity within the Earth's crust. The stations were situated between latitudes 56° and 64° geomagnetic north. As substorms often extend over this area, most magnetic disturbance events have strong external spatial gradients and are not suitable for determining the electrical conductivity distribution inside the Earth. Some magnetic disturbance events with only smooth external spatial gradients could be selected and used for further analysis. For 11 of these events (2-6 h long), the horizontal spatial wavenumber k has been calculated. The product of the wavenumber k and the inductive scale length C was then used as an acceptance criterion and as a weighting function in the calculation of single station transfer functions. Most of the data were not acceptable for the criterion $k \cdot |C| < 0.3$ for periods longer than 500 s. Because of the small number of acceptable data the statistical significance was not sufficient for all sites. Despite these problems induction vectors and conducted hypothetical vertical field maps could be used to locate conductivity anomalies. Intensive induction was found in three zones in the area under investigation.

Key words. Magnetometer arrays – Induction vectors – Source field effects – Baltic Shield

Introduction

In 1981 and 1982 four arrays of magnetometers were operated on the Baltic Shield (also known as the Fennoscandian Shield) in central and south-eastern Finland to locate local and regional induction anomalies in the Earth's crust. Each array recorded for about 2 months, producing simultaneous analog data from 30 or 31 stations on films. The recording interval for the three components of the magnetic field was 10 s and the amplitude resolution for the field variations was 2–3 nT. The magnetometers, which were on loan from the University of Münster, were of Gough-Reitzel type. The modification and characteristics of the instruments are described by Küppers and Post (1981).

First results of one of the four arrays were presented in a previous paper by Pajunpää et al. (1983) (hereafter referred to as paper 1). The authors presented magnetograms, polarization ellipses and induction vectors of two events (numbers 1 and 2 in this paper) and concluded that the induction vectors, especially the imaginary ones, were distorted by source fields. Correlation between significant normal vertical and horizontal fields was the probable cause of the errors.

The four arrays shown on Fig. 1 were located between latitudes 56° and 64° geomagnetic north. Often the source of a substorm extends over this area and most magnetic disturbance events have strong spatial gradients in amplitude and phase. As events with a plane wave source field and reasonable field amplitudes are rare, the choice of the events would be critical, if it were not possible to reduce the source field effect. One could attempt to remove the source field effect by a separation of the magnetic field into its external and internal parts (Porath et al., 1970; Mersmann et al., 1979) or normal and anomalous parts (e.g. reference station technique) or by smoothing over randomly distributed source field locations. In this study, the presence of both local and regional anomalies (see paper 1) and a rather small size of the arrays (some 300 × 200 km) make the separation of the vertical component difficult. Beamish (1979) has studied the source field effects at three stations at geomagnetic latitudes 60.03°, 56.47° and 54.22°. The data consisted of day-time and night-time 12 h records from disturbed days. He concludes that source field effects increase with both latitude and period. Only at the lowest latitude station in the period range 4-32 min, could the determination of a single station transfer function be considered independent of the external field characteristics.

The hard work of digitizing the analog records restricts the number and length of the analysed records and consequently the number of degrees of freedom. This makes it necessary to concentrate on events with slowly varying day-time source fields, which in turn limits the locations of the sources to the north. In this case the smoothing over events does not necessarily reduce the source field distortion, which is due to the correlation between normal vertical and horizontal fields. Thus, an attempt is made to classify the magnetic

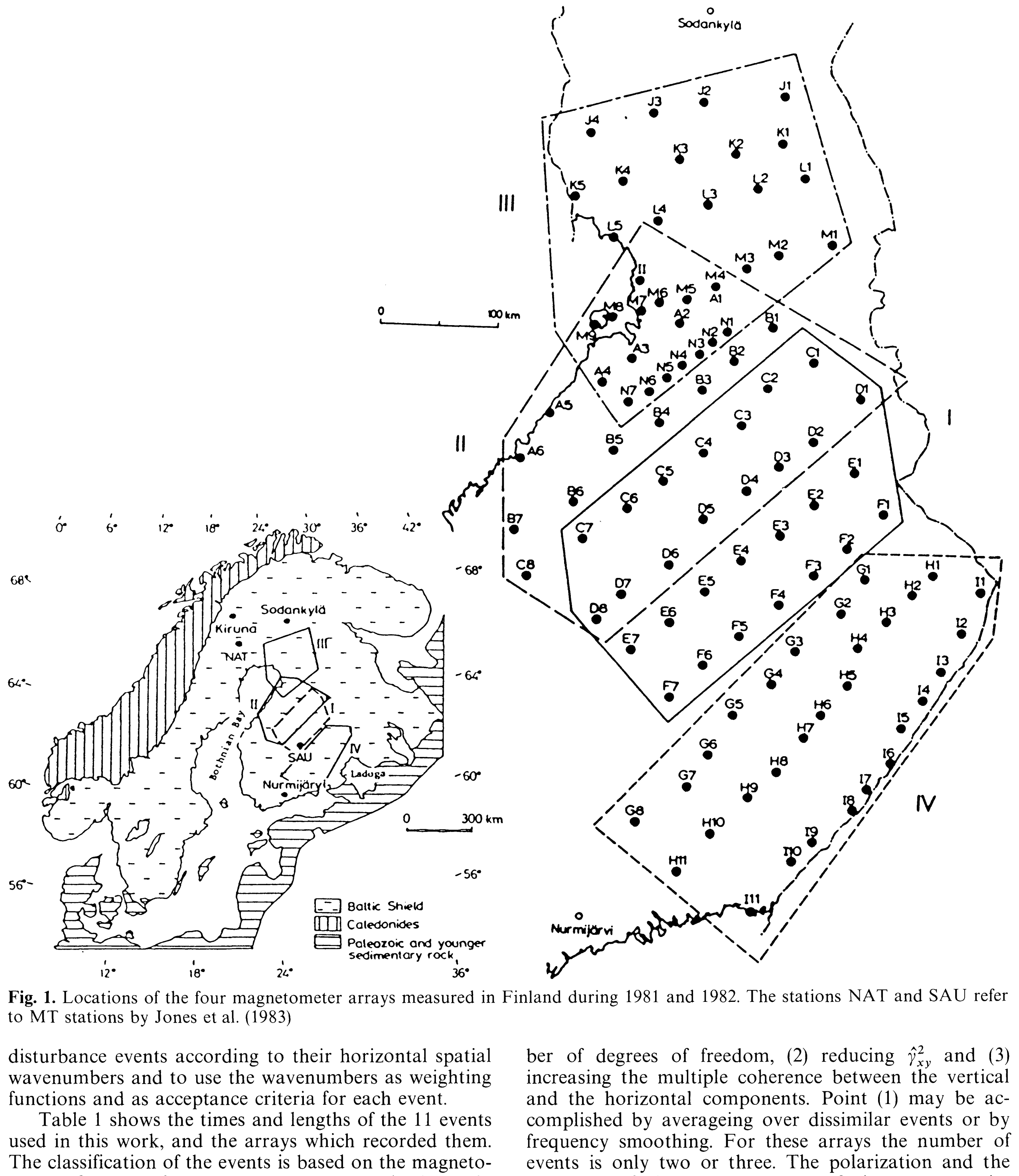

Fig. 1. Locations of the four magnetometer arrays measured in Finland during 1981 and 1982. The stations NAT and SAU refer to MT stations by Jones et al. (1983)

disturbance events according to their horizontal spatial wavenumbers and to use the wavenumbers as weighting functions and as acceptance criteria for each event.

Table 1 shows the times and lengths of the 11 events used in this work, and the arrays which recorded them. The classification of the events is based on the magnetograms. One set of magnetograms, event 5, is shown in Fig. 2. Moreover, Table 1 shows the polarization direction of the horizontal fields and $\hat{\gamma}^2_{xy}$, the estimate of the coherence between the horizontal components, in the period range 160–220 s. Provided that the bias related to the correlation between normal vertical and horizontal components is insignificant, three demands can be stated to reduce the bias and scatter of the transfer function estimates (Jones, 1981): (1) increasing the number of degrees of freedom, (2) reducing $\hat{\gamma}^2_{xy}$ and (3) increasing the multiple coherence between the vertical and the horizontal components. Point (1) may be accomplished by averageing over dissimilar events or by frequency smoothing. For these arrays the number of events is only two or three. The polarization and the character of the magnetograms of the events varies considerably, except the polarization for array III. $\hat{\gamma}^2_{xy}$ is usually less than 0.5 and uncorrelated energy is available in east-west and north-south directions.

Calculation of horizontal spatial wavenumbers

The digitizing interval for ten of the 11 events was 10 s and for one, number 3, 20 s. After calibration and ro-

Table 1. Analysed events. Source fields: A = pulsations, B = substorms with exceptionally uniform source fields and C = substorms with less uniform source fields. $\hat{\gamma}^2_{xy}$ is the estimate of the coherence between the horizontal components at a period band close to 200 s. Polarization direction is the angle, positive to the east of north, of the main axis of the horizontal field polarization ellipse at the same period

Event number	Time	Length (hours)	Array	Lines	Source field	$\hat{\gamma}^2_{xy}$	Polarization direction
1	1981-11-11 16:00–19:00 UT	3	II	A, B, C, D	C	0.08	−15°
2	1981-11-14 14:00–19:00 UT	4	II	A, B, C, D	C	0.31	−39°
3	1981-07-23 11:00–17:00 UT	6	I	C, D, E, F	C	0.37	−43°
4	1982-07-16 15:00–18:00 UT	3	III	J, K, L, M, N	B	0.40	6°
5	1982-10-26 10:00–13:00 UT	3	IV	G, H, I	B	0.70	−25°
6	1982-11-25 5:00– 7:00 UT	2	IV	G, H, I	A	0.66	16°
7	1982-07-17 10:00–12:00 UT	2	III	J, K, L, M, N	A	0.31	10°
11	1982-07-11 9:00–13:00 UT	4	III	J, K, L, M, N	B	0.10	5°
12	1981-11-11 12:00–15:00 UT	3	II	A, B, C, D	C	0.27	−62°
13	1982-11-24 10:00–13:00 UT	3	IV	G, H, I	C	0.76	3°
14	1981-07-26 9:00–11:00 UT	2	I	C, D, E, F	A	0.45	10°

tation to geographic north, the data at each station have been analysed by the method desribed by Jones (1981) to get raw Fourier spectra and smoothed auto- and cross-spectra. The raw Fourier spectra of seven or ten stations were then used to calculate the horizontal spatial wavenumbers. The smoothed spectra were used in the combination of events to get the transfer functions.

The horizontal spatial wavenumber k is derived from the formulae given by Schmucker (1970).

$$k_x = \frac{\partial X}{\partial x} \Big/ X, \quad k_y = \frac{\partial Y}{\partial y} \Big/ Y \tag{1}$$

where X and Y are the northward and eastward directed components of the magnetic field variations, in the frequency domain, x and y are the corresponding coordinates and $k = (k_x^2 + k_y^2)^{1/2}$. To determine k_x and k_y, it is necessary to derive the spatial gradients $\partial X/\partial x$ and $\partial Y/\partial y$. This was accomplished by fitting second order polynomials (see Jones, 1980; Woods and Lilley, 1979)

$$\begin{aligned} X(x, y) &= h_0 + h_1 x + h_2 y + h_3 x^2 + h_4 y^2 + h_5 xy + \delta X \\ Y(x, y) &= d_0 + d_1 x + d_2 y + d_3 x^2 + d_4 y^2 + d_5 xy + \delta Y \end{aligned} \tag{2}$$

to the real and imaginary parts of the components X and Y.

To get proper estimates for k the horizontal fields must be free of anomalies or the anomalous fields must be smoothed out by the polynomial fitting. The following stations have been chosen for this purpose: C2, C7, D1, D5, D8, E1, E4, E7, F2 and F7 of the first array; A1, A5, B1, B6, C1, C5, C7, D1, D5 and D8 of the second array; J1, J3, K1, K4, L2, L4, M1, M4, M9 and N1 of the third array; and H1, H10, H11, I1, I2, I10 and I11 (see Fig. 2) of the fourth array. The basis of this selection was that the horizontal field magnetograms showed no anomalous behaviour and that the whole area of each array was reasonably represented. For the fourth array the second condition could not be fulfiled and only seven stations satisfied the first condition.

The determination of k was performed by first fitting, for each raw Fourier harmonic, the second order polynomials to calculate the ratios h_1/h_0 and d_2/d_0. These raw k values were then smoothed by the same constant Q box-car frequency windows, with $Q = 0.3$, as in calculating smoothed auto- and cross-spectral estimates.

Figure 3 shows the wavenumber estimates as a function of period for each array and event. Most k values fall between 0.002 and 0.006 km^{-1}. These correspond to spatial wavelengths 3,100 km and 1,000 km, respectively. The scatter of these k estimates is large but some features can be noticed. Events 11 and 5 have the lowest mean values of the k estimates, 0.0027 and 0.0033 km^{-1} respectively. The highest mean values are for events 2, 13 and 6; 0.0077, 0.0056 and 0.0055 km^{-1}. However, those estimates of the events 6, 13, and also 5, from the fourth array decrease as the period increases. The choice of the normal (horizontal field) stations was least clear for this array (see Fig. 2) and only 7 stations in its corners could be used. One explanation for the slope, assuming that its not a characteristic of the source itself, is that the cosen stations are not all normal, either for short periods or for long periods.

Acceptance criteria and weighting functions

The final averaging over events and frequencies was performed by a program originally written by Jones (1977) to average MT and GDS events. It was modified to take into account the wavenumber k as an acceptance criterion and as a weighting function. The program uses smoothed auto- and cross-spectral estimates of all available events at one station. The data are normalized to the geometric mean of the horizontal field powers. The transfer functions are then calculated as weighted means in three frequency bands per decade.

In addition to the wavenumber k, bias-reduced multiple coherence functions are used as a statistical acceptance criterion and weighting function (see Jones, 1983):

$$(\gamma^2_{zxy})_b = \gamma^2_{zxy} - \left(\frac{4}{\nu - 2}\right)(1 - \gamma^2_{zxy})^2 \, (1 + 4\gamma^2_{zxy}/\nu), \tag{3}$$

where ν is the number of degrees of freedom associated with the estimate. This estimate had to be greater than 0.5 for acceptance.

The formula given by Schmucker (1973) states the

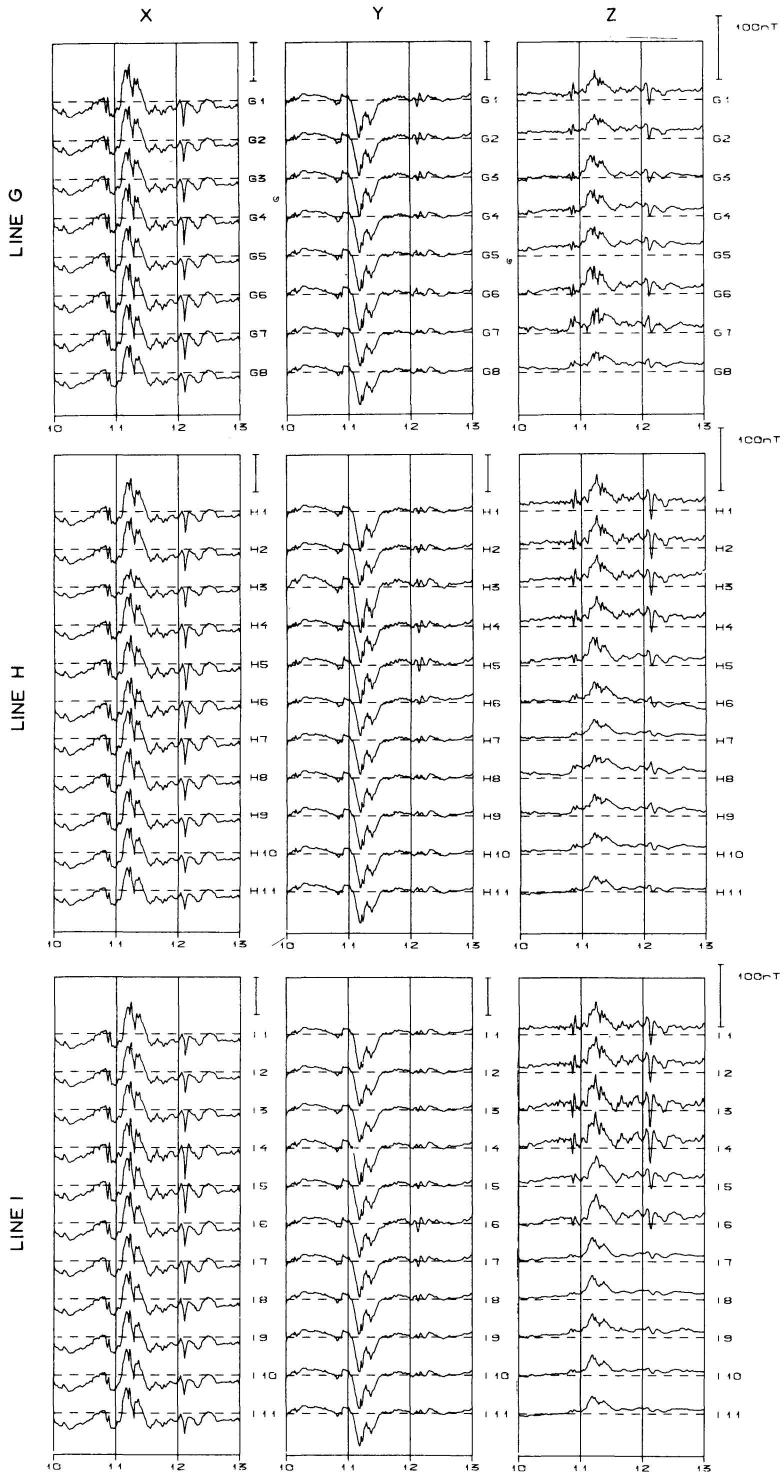

Fig. 2. Magnetograms of a substorm 10.00–13.00 UT on October 26th 1982, for lines G, H and I

ratio between the normal vertical component Z_n and the normal horizontal component H_n in a one-dimensional case:

$$\frac{Z_n}{H_n} = ik\,C(\omega, k), \tag{4}$$

where $C(\omega, k)$ is the inductive scale length. Schmucker states that when only a small sinusoidal modulation of the source field is allowed with the constraint that $k|C(\omega, 0)| \ll 1$, Eq. (4) is approximated by

$$\frac{Z_n}{H_n} = ik\,C(\omega, 0). \tag{5}$$

This formula has been used to calculate the norm Z_n/H_n by approximating the norm of the inductive scale length $C(\omega, 0)$ from available data. For this purpose, apparent resistivity curves given by Jones (1983), for stations SAU in southern Finland and NAT in

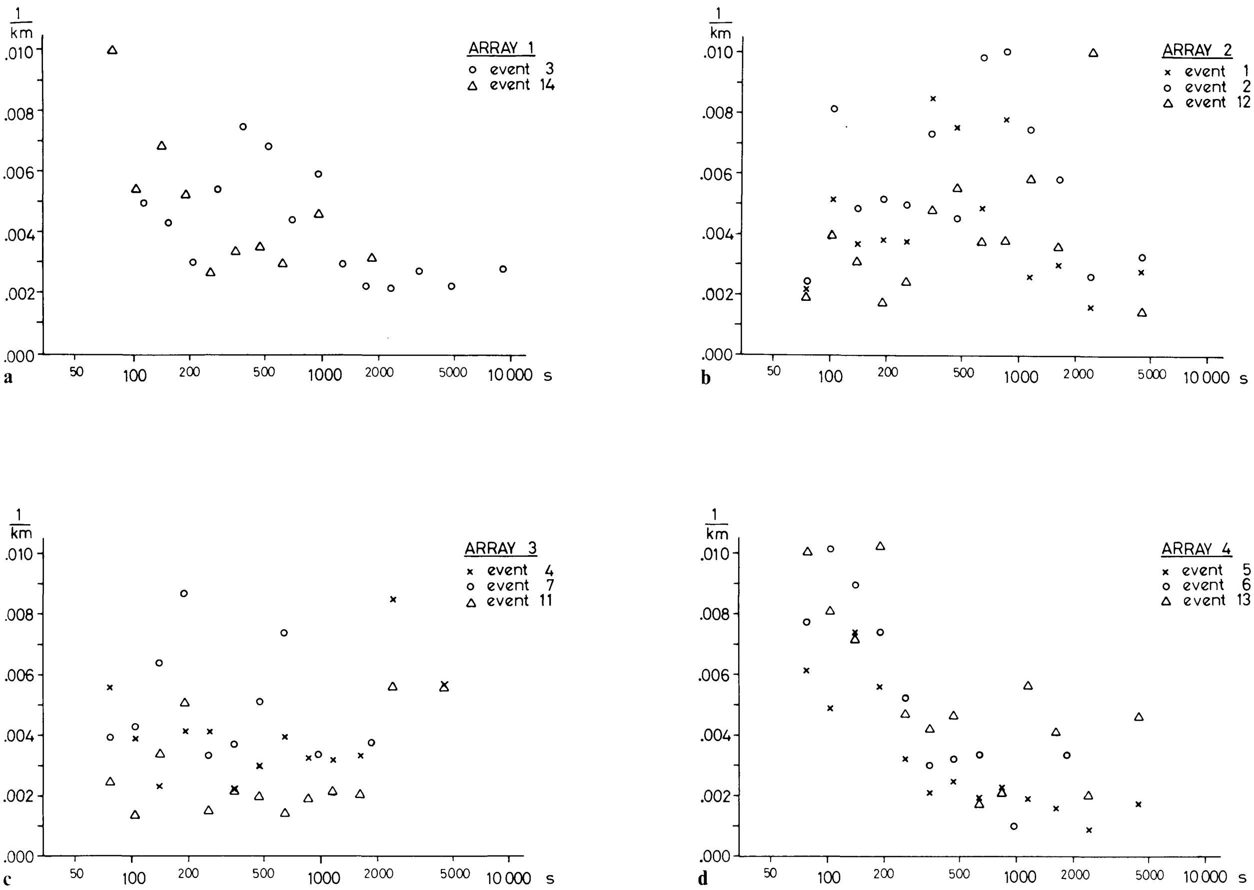

Fig. 3a–d. Horizontal spatial wavenumbers k as a function of period for the analysed events

northern Sweden, have been used (see Fig. 1). The formula

$$\rho_a(\omega) = \omega \mu_0 |C(\omega, 0)|^2 \tag{6}$$

where μ_0 is the magnetic permeability of free space, determines the relation between apparent resistivity ρ_a and the norm of the transfer function C.

The ratio $|Z_n/H_n|$ can be interpreted as a maximum error for the norm of the sum of real and imaginary induction vectors. As the correlation between the anomalous vertical component Z_a and the horizontal component H_n can be assumed to be high, whereas the correlation between the normal Z_n and the normal H_n is probably lower, the error related to the ratio Z_n/H_n is smaller than the calculated $|Z_n/H_n|$.

The apparent resistivity curve for station SAU has been used to calculate $k|C|$ for arrays 1, 2 and 4. The curve for station NAT has been used for array 3. The product $k|C|$ had to be less than 0.3 for acceptance in this study. Most of the data at periods longer than 500 s do not fulfil this criterion. The number of accepted estimates decreases as the period increases, which is due to the increase of the inductive scale length.

The product $k|C|$ was also used to construct a second weighting function w_2 which varies from the acceptance value $A(=0.3)$ to one:

$$w_2(\omega) = 1 - \frac{(1-A)}{A} k_\omega |C(\omega)|. \tag{7}$$

The final weighting function was a geometric mean of the bias-reduced multiple coherence functions w_1 and w_2:

$$w = (w_1 \cdot w_2)^{1/2}. \tag{8}$$

Induction vectors

Figure 4 presents the derived single station induction vectors of period 200 s (period range 150–320 s). Real vectors have been reversed. The dots indicate stations with no accepted data. The method described by Jones (1981) was used to calculate the confidence limits. A low number of accepted estimates, and an analogously low number of degrees of freedom, made it impossible to calculate the confidence circles for many stations even at this short period, so that the circles are not shown here. The mean number of accepted estimates at 200 s was 3.2, for the lines A and B, 6.6 for the lines C and D, 2.2 for the lines E and F 3.9 for the lines G, H and I and 5.3 for the lines J, K, L, M and N.

According to the number of accepted estimates, the worst lines are A, B, E and F. An evident error in the induction vectors can be found at station A1, which is the same as M4. The two independently determined real and imaginary vectors differ significantly from each other. The longer vectors have been determined using the second array and only two estimates, and the shorter ones using the third array. Obviously the vectors at station A1, and also at station II, are strongly biased.

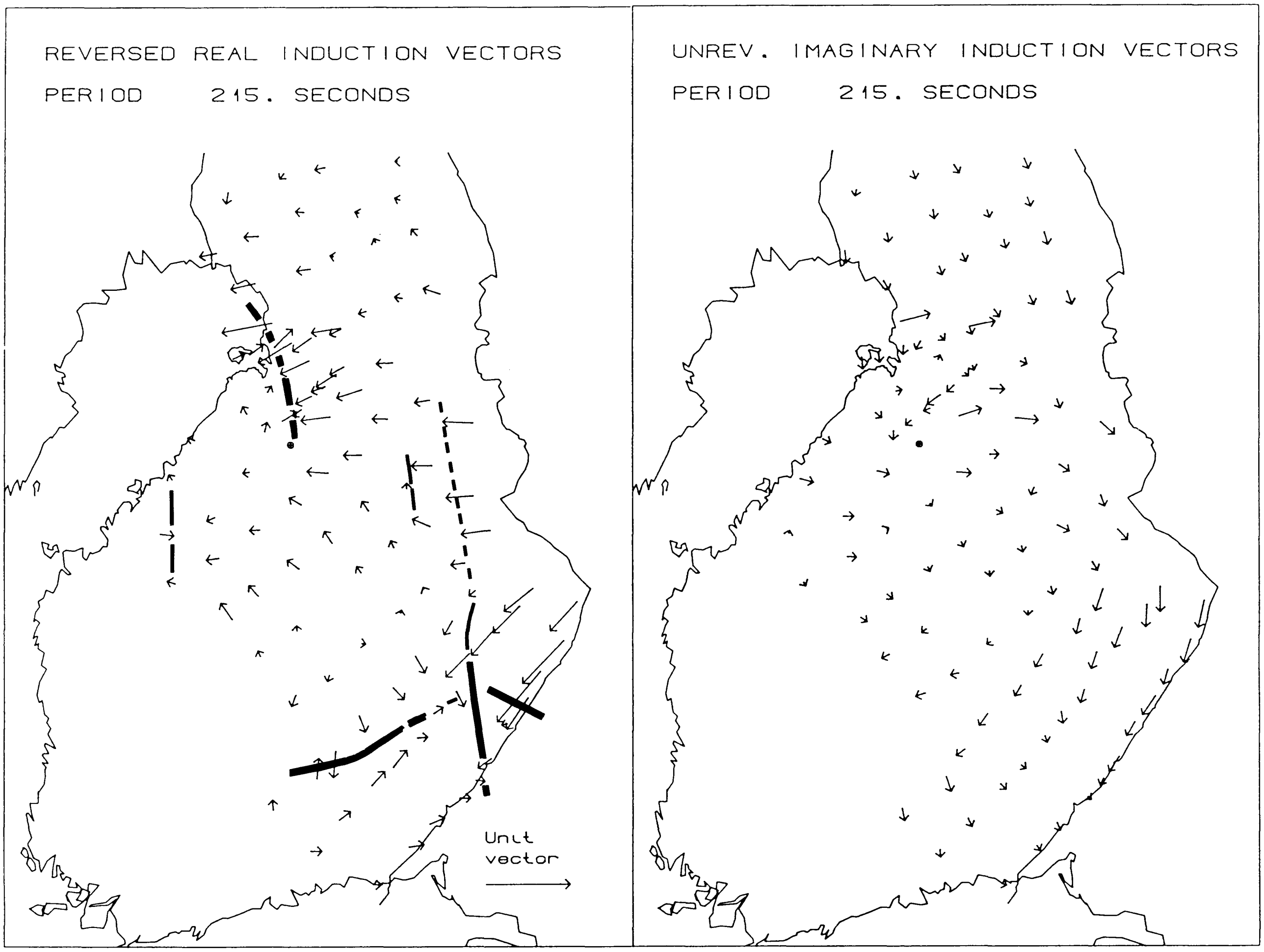

Fig. 4. Real reversed and imaginary induction vectors for period 200 s. The *dashed lines* show the main anomalous zones

The real vectors, except on the lines A and B, show qualitatively the anomalous features very well, which have also been found from magnetograms, polarization ellipses and Fourier maps. The main anomalous zones are drawn on the map of the real vectors. The imaginary vectors are mainly short showing that the source distortion is, in any case, rather small.

Hypothetical event analysis

An alternative way of presenting the transfer functions is to calculate the hypothetical Z field at all stations, which would occur in a geomagnetic variation whose external horizontal field is uniform with a specified polarization (Bailey et al., 1974). Figure 5 shows two maps of the in-phase vertical field that would occur if the horizontal field was linearly polarized and directed southwards and westwards, respectively.

There are three anomalous zones with strong gradients on the maps of the in-phase vertical field. The first one, on the north-eastern coast of the Bothnian Bay, is clearly observable for both polarizations striking slightly west of north. This good conducting zone, called the "Oulu" anomaly (paper 1), can be followed from the coast southwards for about 100 km. First MT soundings close to the N line reveal a conductor at depths of more than 5 km and with conductivity of more than 1 Sm^{-1} (Zhang et al., 1983). This conductivity is common for the black schists which are often encountered in the surface rocks of the Baltic Shield.

Another intense anomaly is observed in southern Finland. It causes an anomalous Z field only if the inducing field is polarized in the north-south direction. This zone, called the "Mikkeli" anomaly, runs east-north-east and meets another, north-south striking zone which is clearly observable in the map of the east-west polarization. This third zone is called the "Outokumpu" anomaly and, in its northern part, it probably forms a boundary in the conductivity of the crust, as the in-phase vertical field does not change sign when crossing the zone.

The easternmost corner of Finland shows anomalous behaviour on both of the hypothetical Z field maps. The reason for the complexity is that there are two current concentrations crossing the I line and the south-eastern boundary of Finland. The first one is the southern part of the above-mentioned Outokumpu anomaly which runs in the locality of stations I6 and I7. The other is close to station I4 and it may be connected with the Mikkeli anomaly in the west. This strongly three-dimensional current system causes a Z field magnitude equal to the horizontal field magnitude, for

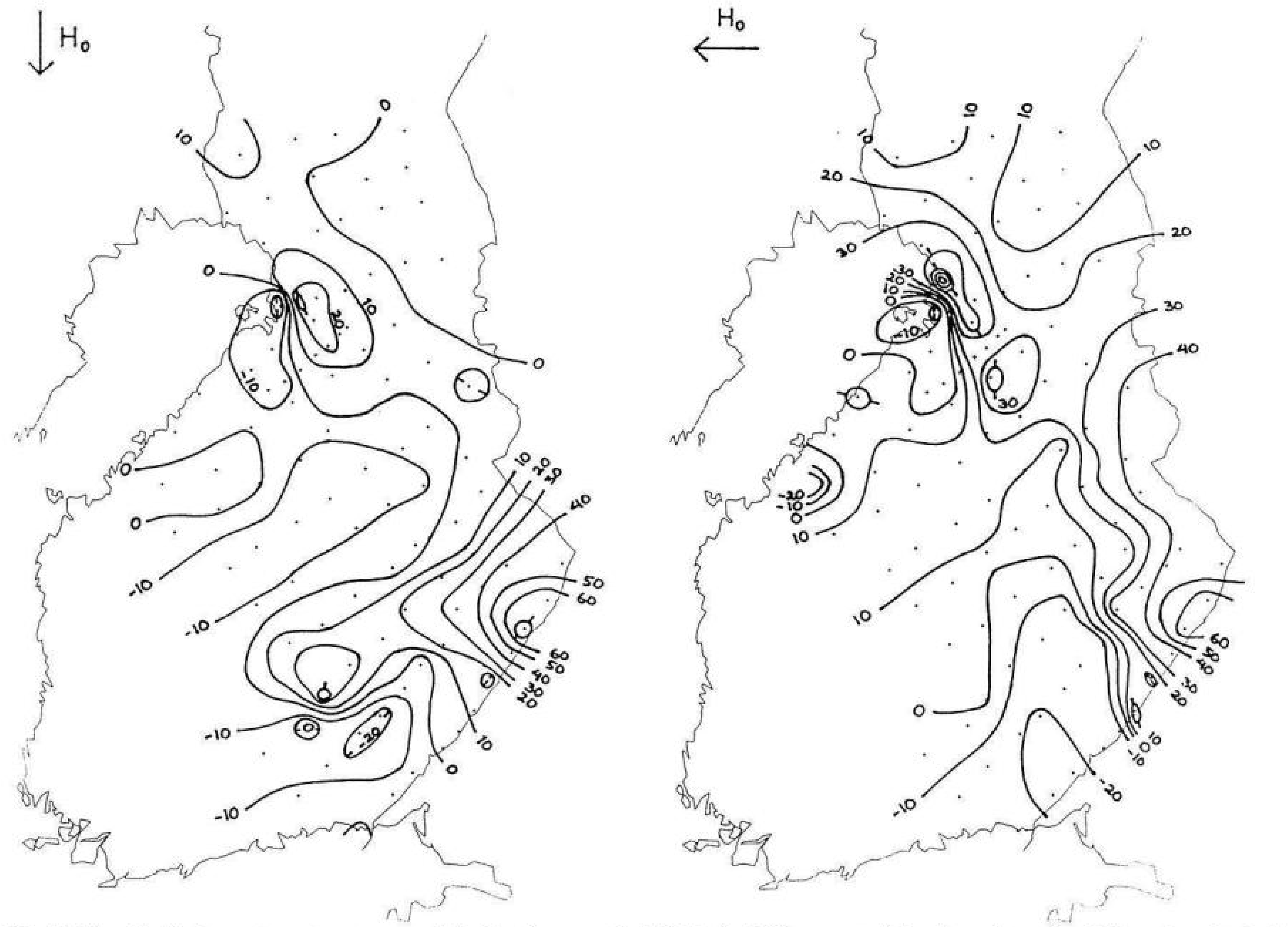

Fig. 5. Hypothetical event contour maps of the in-phase vertical fields (× 100) generated by the unit regional linearly polarized horizontal fields. The direction of the horizontal field is as shown on the maps

a suitable polarization, in the resistive eastern corner of Finland. This illustration of the anomalies in south-eastern Finland is probably too simple and the conducting zones may be composed of several good conducting formations like schists. Moreover, a part of the induced currents may diverge into central Finland where Jones (1983) has reported a conducting (18–36 Ωm) layer in the lower crust.

The east-west polarization also reveals a weak anomaly in the western part of the measured region. This anomaly was observed at only three stations and is reported also in Paper 1 together with a local anomaly close to station D3.

Discussion

In this paper the horizontal spatial wavenumbers of 11 magnetic disturbance events have been determined to obtain a measure of the normal field ratio Z_n/H_n. The chosen Earth model is not valid for the whole research area, but this may cause Z_n/H_n ratios which are too small only if the earth is more resistive than the model. This is obviously the case in eastern Finland. The correlation between the normal vertical field Z_n and the horizontal field H should be studied for each event to get the real error. Thereafter, one could obtain a suitable acceptance criterion.

The wavenumbers k have a rather large scatter and they would probably need further smoothing before calculating the $k|C|$ product. The number of accepted estimates decreases as the period increases, due to the increasing inductive scale length, and it is hard to get undistorted long period transfer functions from these arrays by this method. Also at period 200 s the number of acceptable data was low, especially on lines A and B. Therefore, more data will be needed to increase the statistical significance. However, reliable estimates of the real single station transfer functions can be determined by a careful selection of events at geomagnetic latitudes 56°–64°. Induction vectors and conducted hypothetical Z field maps can be used effectively to study conductivity structures in this sub-auroral region covered by several magnetometer arrays.

Finland is located in the central part of the Baltic, or Fennoscandian, Shield and the subsurface consists almost solely of crystalline rocks of Precambrian age. The conductivity anomalies which have been detected in this study are rather local, some of them giving rise to very intensive induction. Two other magnetic variation studies by Rokityansky et al. (1979) and by Jones (1981) on this shield have also detected anomalous regions; Jones in northern Sweden and in northern Norway, and Rokityansky in the south-eastern part of the shield in the Soviet Union. Rokityansky et al. have found an anomaly running under lake Ladoga from south-east to north-west. That anomaly is obviously connected with the conducting zones in south-eastern Finland.

As mentioned before, the conductive material of the Oulu anomaly is in the upper crust. This, depth together with the high conductivity ($>1\,\mathrm{Sm}^{-1}$), in an ancient shield suggest that the cause is graphite, which is common in the surface rocks of the schist belts. The nature of the other anomalies is probably analogous to that of the Oulu anomaly. However, a more quantitative analysis of the data and a greater number of magnetotelluric soundings are needed to get the final idea of the observed anomalies.

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