# CHANGES IN TRANSFER FUNCTIONS WITH TIME


E. P. KHARIN

*Academy of Sciences of the U.S.S.R., Soviet Geophysical Committee, Molodezhnaya 3, Moscow 117 296, U.S.S.R.*



**Abstract.** The paper contains data on the transfer functions of the electromagnetic field of the Earth and their changes with time.

The results show that these changes are caused by the instability of parameters of the external sources of the geomagnetic field variations in high latitudes and by the changes in the physical properties of rocks and deep matter in the region of recordings of electromagnetic field components.

A statement is made that under certain conditions the observations of changes in transfer functions of the electromagnetic field of the Earth can be used as long-term forerunners of earthquakes.


## 1. Introduction

During the preparatory period before an earthquake the physical properties of rocks in its focus experience certain changes; in particular, an increase of electrical conductivity in the dilatancy zone (Scholz *et al.*, 1973). This increase of electrical conductivity is recorded by instruments either by the application of artificial electromagnetic fields, or from a study of the stability of transfer functions of the natural electromagnetic field of the Earth. This paper describes the analysis of changes in transfer functions and their causes.

As in the previous review (Niblett and Honkura, 1980), the coefficients of linear correlation between the vertical and horizontal components of geomagnetic field variations are formed as transfer functions:

$$\Delta Z_i = A \cdot \Delta H_i + B \cdot \Delta D_i$$

$$\Delta Z_i = A \cdot \Delta H_{xi} + B \cdot \Delta H_{yi} \, .$$

In the magnetotelluric method the impedance tensor relating horizontal electric and magnetic variation fields is formed:

$$\begin{pmatrix} E_x \\ E_y \end{pmatrix} = \begin{pmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{pmatrix} \begin{pmatrix} H_x \\ H_y \end{pmatrix} .$$

Firstly, the analysis of the stability of transfer functions should take into account that the actual observed fields of geomagnetic variations are different from theoretical fields.

In some cases these differences and the instability of the external geomagnetic field can cause changes in transfer functions. The paper discusses such situations.

Secondly, variations in the electrical conductivity of the dilatancy zone can be fixed only under definite conditions of variation-recording. The article gives the results of


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estimates of such conditions (period of variation, distance from earthquake hypocenter to observation point, etc.).

The analysis shows that the changes in transfer functions can be interpreted as forerunners of strong earthquakes, if their hypocenters are not far (a few tens of km) from the observation point of geomagnetic field variations.

Niblett and Honkura (1980) gave a detailed review of this topic at the Fourth Workshop (Murnau, 1978); this present review, therefore, includes only publications which appeared after 1978 or are not covered by the first review.

## 2. The Possibility of Monitoring an Electroconductivity of Dilatant zone by Observation of Geomagnetic Variations

The previous reviewers cited the work of Rikitake (1976) who numerically determined the period ($T$) of the geomagnetic variation field most sensitive to changes resulting from the dilatancy of crustal rocks before an earthquake. The results of Rikitake's calculation of the induction parameter $A$:

$$A = 4\pi\sigma\omega D^2 , \qquad D = 0.1r$$

are shown in Table I. The mean radius of the dilatant zone ($r$) was determined by Dambora's formula:

$$M = 1.96 \log_{10} r + 4.45 \tag{1}$$

It is assumed that $r$ in Equation (1) can be replaced by the half-width of the conductivity anomaly of linear dimension 10D, i.e. the assumption is made that the area of the dilatant zone cannot be greater than the area of the earthquake focus. Rikitake's (1976) calculations reveal that the changes in the components of geomagnetic field variations become marked at $A$ = 0.001. As seen from Table I, the tracing of anomalous conductivity changes associated with earthquakes of magnitude $M \leqslant 6$ calls for observations with periods of a second or less. Mijakoshi (1975), however, observed considerable changes in transfer functions connected with the Tashkent earthquake of April 26, 1966, for longer period variations, such as SSC's and SI's.

TABLE I

Periods for mesh dimensions when the values of induction parameter A are selected (by Rikitake, 1976)

| $A$ | $T$ (sec) | | | |
|---|---|---|---|---|
| | $D$ = 0.5 km ($M$ = 5.8) | $D$ = 0.1 km ($M$ = 6.4) | $D$ = 2 km ($M$ = 7.0) | $D$ = 5 km ($M$ = 7.8) |
| 0.001 | 19.8 | 79.0 | 316 | 1970 |
| 0.01 | 2.0 | 7.9 | 3.16 | 197 |
| 0.1 | 0.2 | 0.8 | 3.2 | 19.7 |

A more detailed description of this earthquake is given in the book 'The Tashkent Earthquake of April 26, 1966 (1971). The basic parameters of the earthquake focus are as follows:

area of basic rupture $50 \pm 15$ km$^2$
volume of focus $100 \pm 30$ km$^3$

There are other works, however, in which the areas of earthquake focii and of dilatant zones are estimated to be much larger than that postulated by Equation (1). The paper by Whitcomb *et al.* (1973), for example, determines the linear size of the effective rupture ($L$) by the formula:

$$L = 0.58\,(t)^{1/2}$$

where $L$ is given in km and $t$ is the precursor time interval in days. If for the Tashkent earthquake $t$ is taken to be 3 yrs then $L = 19$ km which for $A = 0.001$ gives the variation period perturbed by dilatancy effects to be $T = 5$ min. The paper by Scholz *et al.* (1973) uses $L = c(t)^{1/2}$, taking $c = 1$ the perturbed variation period would be estimated $T = 13$ min. The paper by Lukk and Yunga (1979) determines the size of the Tashkent focal zone to be 20–30 km, i.e. $T$ is estimated to be $\sim 13$ min. If for our calculations we use the formula suggested by Dobrovsky *et al.* (1979):

$$R = 10^{0.43\,\mathrm{M}} \qquad \mathrm{km}$$

where $R$ is the deformation radius of the zone of effective manifestation of earthquake precursors then $T = 130$ min. This last evaluation is obviously overestimated.

The above calculations are made on the assumption that the dilatant zone is equal in size to that of the earthquake focus. There is a point of view that the dilatant zone can be much greater than that of the earthquake focus (Scholz *et al.*, 1973; Whitcomb *et al.*, 1973; and others). If this is true then the period T obtained above will be greater still. There are other theories, besides the dilatancy theory, which explain the changes in physical properties of rocks in the earthquake focus region. Thus Barsukov (1979) indicates that the change in specific electrical resistivity of rock in the process of deformation can be caused not only by the appearance of dilatant fissures and fluid migration but also by the change in curvature of conducting channels without an increase in the porous solution volume. In the quasi-viscous approximation, the effect of differential compression results in the growth in cross-sectional area of a conductor in one fixed direction, and in the transfer of the conducting solution to the new area from pores and other water containing channels. The specific resistivity of rocks prior to failure can decrease before dilatancy occurs. In this case, there is no need to propose a hypothetical liquid reservoir and finite permeability of rocks to provide the filling of dilatant fissures. Within the proposed model the transfer of liquid is carried out by microdiffusion of solutions of conducting fluids from pores to fault fissures, the boundaries of which apparently coincide. The effect of dilatancy and liquid diffusions on the specific resistivity of rocks can be proposed only before those earthquakes for which the preparation zone is connected with the presence of considerable water masses, for example, near water basins, during pumping of water into holes and in irrigated areas, etc.

Miachkin (1978) has suggested a model of the process of earthquake preparation on the basis of avalanche-unstable fissuring. The model explains precursors both in the zone of preparation and outside that zone. It is interesting to note that the distance from the earthquake epicentre to the Tashkent geomagnetic observatory was about 30 km while the earthquake of July 30, 1970 (M = 6.6) had practically no manifestation at the Ashkhabad observatory the distance between epicentre and observatory being, in this case, some 190 km. In connection with the last comment it is of interest to reconsider some of the conclusions of previous works.

It follows from the papers of Vadkovsky and Kharin (1978) and Kharin (1980), that there is no direct correlation between the changes of Viese vectors and seismicity of the region of *Surlari* geomagnetic observatory and, on the other hand, some kinds of variations show correlation between changes in mean annual values of Viese vectors and the changes in mean annual values of geomagnetic indices. The mean annual values of Viese vectors were calculated for the interval 1957–1976 at Odessa (Stepanovka) observatory (Figure 1) and at Moscow (Krasnaya Pakhra) observatory (Figure 2) with the purpose of finding the effect of helio- and geo-magnetic activity in changes of the values of Viese vectors. A more detailed investigation of the changes in the Viese vectors at Surlari observatory throughout the same period is shown in Figure 3.

The Odessa geomagnetic observatory is the nearest observatory to Surlari and it is located in a more stable tectonic zone than Surlari. The geomagnetic latitudes of both these observatories are very close ($\phi_s = 42.5°$ and $\phi_0 = 43.7°$) thereby ensuring a similar effect of variations in geomagnetic activity on the changes of Viese vectors calculated at

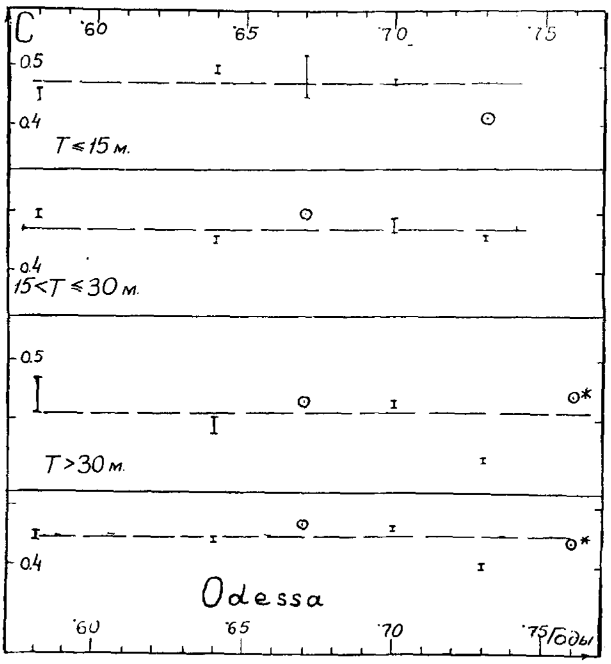


Fig. 1. Changes in values of Viese vector at Odessa observatory.

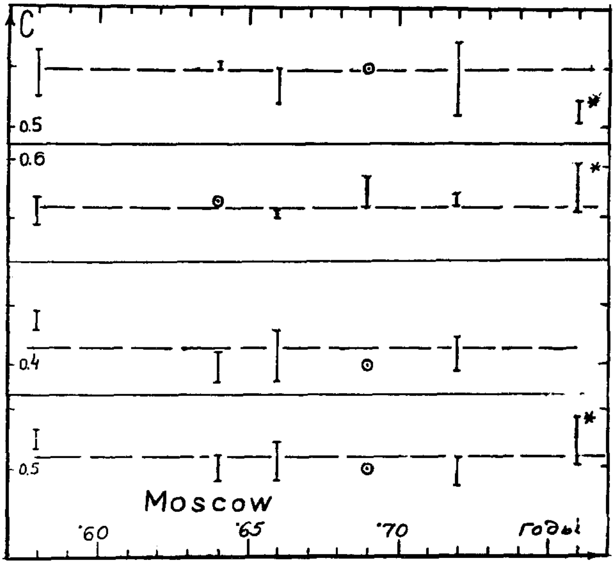


Fig. 2. Changes in values of the Viese vector at Moscow observatory.

the two observatories. The Moscow observatory is located very much to the North of Surlari observatory ($\phi_M$ = 50.9) and in an aseismic and tectonically more stable zone. Therefore the effect of variations of geomagnetic activity on Viese vectors at Moscow should be different from those at Surlari and Odessa observatories.

A comparison of Figures 1, 2, and 3 shows that the character of changes in the Viese vectors at the Surlari, Odessa and Moscow observatories is different. Though $C$ ($C = (A^2 + B^2)^{1/2}$) values at Surlari show considerable variation throughout the period of study, at Odessa and Moscow they are practically stable when an allowance is made for the errors in their determination. Only in 1973 did the $C$ values decrease at Odessa. In Figure 1 and Figure 2 the C values obtained by Logvinov and Rokitiansky (1980) and Rokitiansky (1977) are indicated by an *asterisk.* The variations as a function of frequency shown in Figure 1, 2, and 3 also confirm the stability of the $C$ values at Odessa and Moscow when compared to the variability at all periods exhibited at Surlari. An important conclusion of the above studies is that the change in Viese vectors at Surlari is connected with changes of electrical conductivity of the crust under Surlari. In view of this the changes in the Viese vectors at Surlari are now studied in greater detail and a more differentiated analysis is made with regard to the quantitative characteristics of seismicity by separately grouping near and far earthquakes of different magnitudes. In Figure 4, the curve shows the total number of earthquakes in the vicinity (not more than 150 km and not less than 30 km) of Surlari observatory during the period of study. Arrows indicate the occurrence of earthquakes of $M \geqslant 5$ (Richter scale). Seismicity plots are given below curve $N$ for different epicentral distances ($R$). The circles in these plots show the number of earthquakes with hypocentral depths of not more than 80 km. The lowest plot ($R$ = 40 km) shows only one earthquake, occurring in 1967, but with $M = 5$ and an epicentral distance of 40 km. Comparison of this plot with the variation of $C$ at Surlari (from Figure

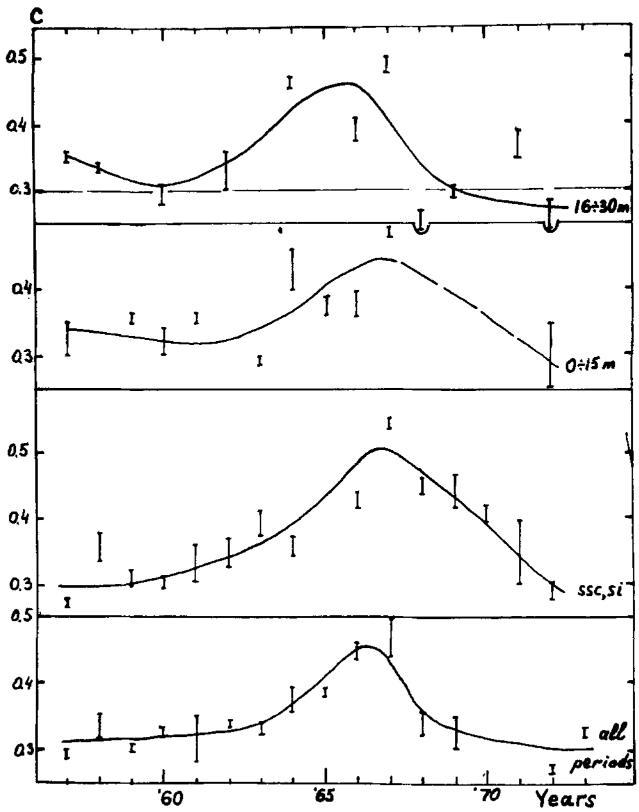


Fig. 3. Changes in values of the Viese vector at Surlari observatory.

3) reveals a long ($T \sim 3$ yrs) process of earthquake preparation, as in Tashkent. The analysis of all the plots in Figure 4 shows that temporal changes in the Viese vector connected with tectonic processes participating in earthquake preparation becomes marked only during strong ($M \geqslant 5$) shallow-foci earthquakes, provided that the epicentral distance is within the first few tens of kilometres. Strong earthquakes in 1959–1961 were not manifest since they occurred more than 100 km from Surlari observatory.

Rikitake (1979b) found a considerable change in the A coefficient transfer function before the earthquake of June 30, 1972 in Sitka, Alaska. It is interesting to note that the time of earthquake preparation does not correlate with the time scales established by Rikitake (1979a, c), Scholz *et al.* (1973) and others. An earthquake of $M = 7.2$ should be preceded by the development of a precursor over some 3 yrs. It should also be emphasised that the epicentral distance of the Sitka earthquake was some 40 km from the observatory.

The enumerated papers have shown that the behaviour of transfer functions in the period range 5–30 min can serve as precursors of strong earthquakes with shallow focii if the distance between focus and observation point is a few tens of kilometres.

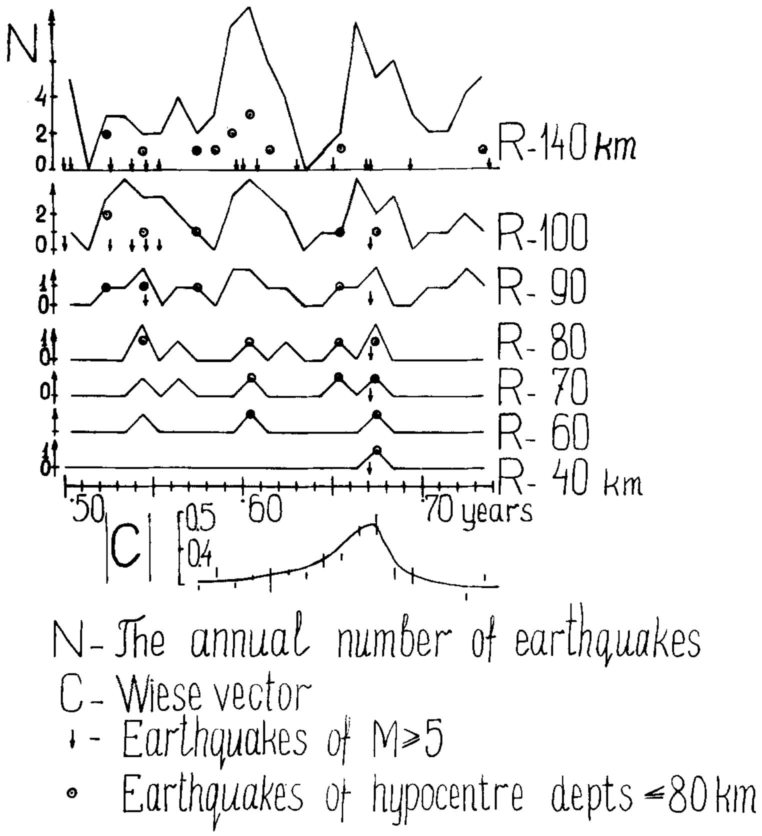


Fig. 4. Comparison changes of Viese vector with seismicity of region of geomagnetic observatory Surlari.

## 2. Changes of Electrical and Magnetic Properties with Time

In the following, the papers containing information on the changes in electrical and magnetic properties, caused by earthquake associated effects, are noted separately.

(1) After reinterpretation of the data of Yamazaki (1975), from which it appeared that an imminent precursor was observed before the earthquake of May 9, 1974 at Izu-Hanto-Oki, Rikitake and Yamazaki (1979) demonstrated that a further change in resistivity, measured by the Yamazaki variometer, was observed a few days before the earthquake. The epicentral distance was estimated to be 100 km.

(2) Kurtz and Niblett (1978) indicate a possible influence of seasonal changes of the hydro-regime in the region of their observations (adjacent to the St. Lawrence river in Eastern Canada) in magnetotelluric response functions. Due to low seismic activity during the observational period (there were only 2 earthquakes with $M \geqslant 3$ in 3 yrs) direct correlation between resistivity changes and seismic activity was not observed. The variation of the apparent resistivity sounding curve at one site (Charlevoix) between July 1975 and

September 1976 was about 20% for 1000 s period.

(3) The results of Morinson *et al.* (1979) are also worth attention. Whereas during an earthquake ($M$ = 4.2, epicentral depth = 9 km) occurring on June 22, 1973 on the San Andreas fault, resistivity was observed to vary by 15%, an earthquake occurring on December 15, 1977 ($M$ = 4.0, epicentral depth = 7 km) caused no resistivity variations on the same array of receiver dipoles even though the earthquake foci was about 1 km distant.

(4) Sidorin *et al.* (1979) measured electrical resistivity in the Garm seismoactive region by dipole sounding with powerful impulse magnetohydrodynamic generators. The dipole moment of the source makes it possible to carry out soundings with spacings of up to 40 km. During the period of observation, apparent resistivity variations were noted to be coincident with the strongest earthquakes in that region.

(5) Idarmachev *et al.* (1978) noted a change in electrical resistivity of up to 6% measured by a dipole installation located at distances of between 5 and 15 km from earthquake epicentres in *the North Caucasus.*

(6) Makhkmadjanov (1979) notes anomalous changes in effective specific electrical resistivity caused by tectonic processes on the Kyzylkum geomagnetic polygon.

(7) Ohchi *et al.* (1979) note the possible connection between changes of up to 5 nT in midnight values of the total magnetic field and the Izu-Oslima-Kinkai earthquake of 1978. They believe that the anomaly may be caused by piezomagnetic effects in the crust.

(8) Skovorodkin and Bezuglaya (1980) analysed the results of their observations in the Garm polygon and arrived at the conclusion that the appearance of local geomagnetic variations with amplitudes of several nT is associated with the seasonal changes of the hydro-regime in near surface layers. The most likely places for this to be observed are in river valleys containing alluvial deposits. Such a possibility should be considered in the interpratation of magnetic results in seismically active regions since if water filtration is present in the dilatancy zone during earthquake preparation then both piezomagnetic and electrokinetic effects may appear.

(9) Finkelshtein and Kambarov (1979) have compared magnetovariational data obtained at three points in the after-shock period of the Sarykamysh earthquake of 1970 ($M$ = 6.8). One of the points was 200 km from the earthquake focus. The variations of the field vectors at all points were identical; the difference amounting to fractions of or several nT.

(10) Poehls and Jackson (1978) when processing magnetic variations recorded in the San Andreas fault region, did not observe significant tectonomagnetic effects.

## 3. Changes in Transfer Functions Caused by the External Geomagnetic Field Variations

Of related interest are the papers that study the stability of transfer functions as a function of external field parameters.

(1) Beamish (1979) analysed the effects of changes in external field parameters on transfer functions as a function of the latitude of the observation point and the period of

the variations. The analysis was carried out in the 54–60° range of geomagnetic latitude. The analysis shows that $A$ and $B$ coefficients are determined independently of the external field only at the shortest periods and at the lowest latitude station. The effect of the external source field was found to increase with both geomagnetic latitude and period of the variation fields.

(2) Anderson *et al.* (1976, 1978) studied the variations of geomagnetic induction vectors at four stations in Canada at geomagnetic latitudes from 54.53 to 60.42°. Considerable variations of the induction vectors were revealed in their dependence on local time, ie the magnitudes and directions of the vectors were connected not only with the geoelectric structure of the region but with the structure of the external field as well. It was found necessary to average the vectors for several days in order to obtain a stable pattern.

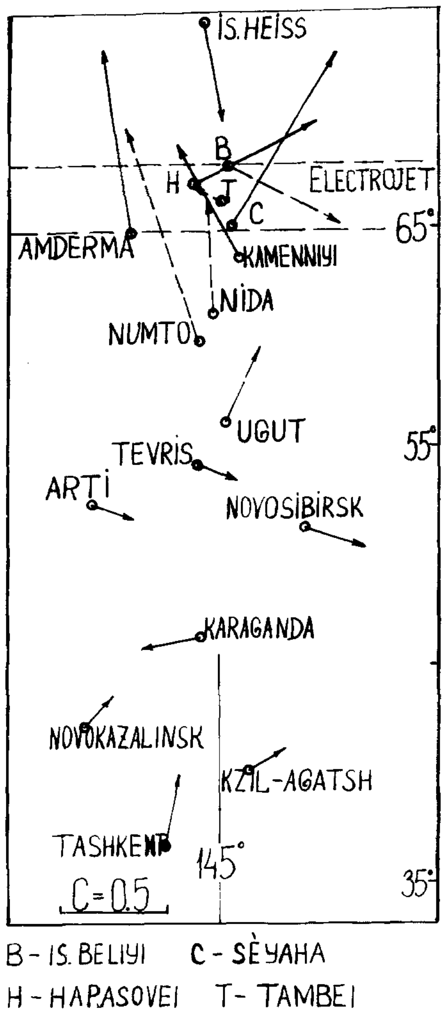


Fig. 5. Distribution of Viese vectors along geomagnetic meridean. Vectors were calculated by the method of least squares: dashes show vectors with Student criterion ⩽5.

(3) Kharin *et al.* (1977) have constructed Viese vectors from a network of observatories and temporary stations located along the geomagnetic meridian 145° (Figure 5). Their studies show that, in the latitude range from 35–55°, Viese vectors reflect the distribution of geoelectric inhomogeneities and that their direction is determined with stability at periods corresponding to bay fields. At high latitudes, however, the Viese vectors are found to change depending on the period of the bay field ($T$) and on the number of bays used in their determination as shown in Figure 6. It appears from the results at Andamera (Figures 5 and 6) that an averaged Viese vector at high latitude reflects the changes in the morphology of the polar electrojet.

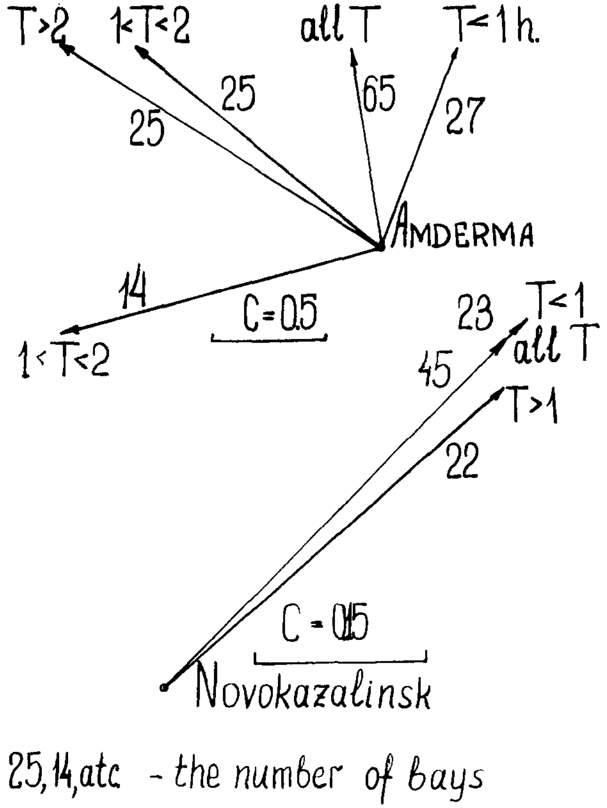


Fig. 6. Viese vectors for different periods of bays (in hours): in upper part – Amderma observatory; in lower part – Novokazalinsk observatory.

## Summary

The results show that changes of the transfer functions are caused by the instability of parameters of the outer sources of the geomagnetic field variations in high latitudes and by the changes in the physical properties of rock and deep matter in the region of recording of electromagnetic field components.

A statement is made that under certain conditions the observations of changes in transfer functions of the electromagnetic field of the Earth can be used as long-term forerunners of earthquakes.

During recent years the idea of using changes in transfer functions in the prediction of strong earthquakes has approached realisation. Rikitake in his first classification of earthquake precursors does not even mention this phenomenon whereas his classification of 1979 (Rikitake, 1979a) includes an additional new kind of earthquake precursor, that of conductivity anomalies and of changes in transfer functions.

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