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Tectonic process monitoring by variations of the geomagnetic field absolute intensity

Dmitry B. Avdeev, Alexei V. Kuvshinov and Oleg V. Pankratov
Institute of Geoelectromagnetic Research, Russian Academy of Sciences, Troitsk, Moscow Region, Russia

Abstract
We propose a novel technique for tectonic process monitoring. The technique includes: measurements of absolute intensity variations, at network sites, synchronously with measurements of horizontal geomagnetic field variations at a reference site; spectral analysis of measured time series and construction non-conventional transfer functions and interpretation of the transfer functions constructed in order to detect or/and to forecast the tectonic processes. Using numerical modelling we show the sensitivity of transfer functions obtained with respect to the temporal changes in the Earth's resistivity associated with tectonic processes. We also demonstrate that the components of the geomagnetic field are reconstructed in terms of spatial distribution of absolute intensity variations.

Key words *geomagnetic variations – absolute intensity – transfer function – monitoring – tectonic process*

1. Introduction

There are two principal techniques for monitoring of tectonomagnetic effects (Zlotnicki, 1995; Johnston, 1996). The first technique is based upon measurements of the geomagnetic field components B_z and $\boldsymbol{B}_\tau$, their time series analysis, and an interpretation of the transfer functions constructed. For example, induction vector $\boldsymbol{v}_\tau$ is used (Chen and Fung, 1993) to detect seismomagnetic effects. The induction vector, $\boldsymbol{v}_\tau = \{v_i\}$ $(i = x, y)$, being defined at a site $\boldsymbol{r}_b$ as

$$B_z(\boldsymbol{r}_b, \omega) = \boldsymbol{v}_\tau(\boldsymbol{r}_b, \omega) \cdot \boldsymbol{B}_\tau(\boldsymbol{r}_b, \omega) \quad (1.1)$$

connects variations of vertical B_z and horizontal $\boldsymbol{B}_\tau$ geomagnetic fields. Here the sign «·» denotes scalar product of the vectors; and ω is frequency of the variations. The basic supposition of the technique is that tectonic processes produce some changes in amplitude and in direction of the induction vector. By constructing the induction vectors for a number of network sites and for a number of frequencies one can obtain quantitative information about the location and magnitude of tectonogeneric sources. The weak point in the technique is that the network measurements require a number of three-component magnetovariational stations, which are permanently in deficit.

The second technique is based on measurements of absolute intensities T of geomagnetic field, $T = |\boldsymbol{B}|$, and on a subsequent interpreta-

Mailing address: Dr. Dmitry B. Avdeev, Institute of Geoelectromagnetic Research, Russian Academy of Sciences, 142092 Troitsk, Moscow Region, Russia; e-mail: avdeev@gemri.msk.ru

tion of differences of the absolute intensities. Since this technique uses just a magnetometer survey and does not require time series analysis, it is more simple than the first one. In a word, some site, say $\boldsymbol{r}_b$, should be chosen as a reference. Then, at each network site $\boldsymbol{r}_f$ intensity T, should be measured and the difference constructed

$$\Delta T(\boldsymbol{r}_f, \boldsymbol{r}_b, t) = T(\boldsymbol{r}_f, t) - T(\boldsymbol{r}_b, t). \quad (1.2)$$

A basic supposition of this technique is that temporal changes of the difference ΔT are correlated with tectonic processes. But a drawback of the technique is that some of magnetic field generated by ionospheric currents is not fully cancelled in the difference ΔT, and so temporal changes of ΔT are not totally of internal origin.

In this paper we propose a new technique of geomagnetic monitoring of the tectonic processes. It comprises an ingenious scheme of geomagnetic measurements and constructing non-conventional induction vectors. The technique combines the advantages of both techniques mentioned above without their weak points. We demonstrate the validity of the technique proposed by numerical modelling. Some of the results discussed here were previously presented in (Avdeev *et al.*, 1994).

2. A new geomagnetic monitoring technique

The novel technique for tectonic process monitoring includes:

1) measurements of absolute intensity $T(\boldsymbol{r}_f, t)$, at network sites $\boldsymbol{r}_f$, in synchrony with measurements of horizontal magnetic field $\boldsymbol{B}_\tau(\boldsymbol{r}_b, t)$ at a reference site $\boldsymbol{r}_b$;

2) time series analysis of $T(\boldsymbol{r}_f, t)$, and $\boldsymbol{B}_\tau(\boldsymbol{r}_b, t)$ in order to obtain spectra $T(\boldsymbol{r}_f, \omega)$ and $\boldsymbol{B}_\tau(\boldsymbol{r}_b, \omega)$;

3) constructing non-conventional induction vectors $\boldsymbol{a}_\tau(\boldsymbol{r}_f, \boldsymbol{r}_b, \omega)$, so that $T(\boldsymbol{r}_f, \omega) = \boldsymbol{a}_\tau \cdot \boldsymbol{B}_\tau(\boldsymbol{r}_b, \omega)$;

4) interpretation of the vectors $\boldsymbol{a}_\tau$ in order to detect or/and to forecast the tectonic processes.

3. Non-conventional induction vectors

For any site $\boldsymbol{r}_f$ we decompose total geomagnetic field $\boldsymbol{B}$ as follows

$$\boldsymbol{B} = \boldsymbol{B}_0(\boldsymbol{r}_f) + \boldsymbol{B}(\boldsymbol{r}_f, t), \quad (3.1)$$

where $\boldsymbol{B}_0(\boldsymbol{r}_f)$ and $\boldsymbol{B}(\boldsymbol{r}_f, t)$ are temporally constant and variable parts of the total field respectively. Since $\boldsymbol{B}(\boldsymbol{r}_f, t)$ is by several orders of magnitude smaller than $\boldsymbol{B}_0(\boldsymbol{r}_f)$, we, as a first approximation, obtain the following decomposition of absolute intensity T,

$$T = T_0(\boldsymbol{r}_f) + T(\boldsymbol{r}_f, t). \quad (3.2)$$

Here $T_0(\boldsymbol{r}_f) = |\boldsymbol{B}_0(\boldsymbol{r}_f)|$ and

$$T(\boldsymbol{r}_f, t) = \boldsymbol{p} \cdot \boldsymbol{B}(\boldsymbol{r}_f, t), \quad (3.3)$$

where

$$\boldsymbol{p} = \frac{\boldsymbol{B}_0(\boldsymbol{r}_f)}{|\boldsymbol{B}_0(\boldsymbol{r}_f)|}. \quad (3.4)$$

Transforming equality (3.3) into frequency domain, we obtain

$$T(\boldsymbol{r}_f, \omega) = \boldsymbol{p} \cdot \boldsymbol{B}(\boldsymbol{r}_f, \omega), \quad (3.5)$$

but

$$\boldsymbol{B}(\boldsymbol{r}_f, \omega) = \hat{M}\boldsymbol{B}(\boldsymbol{r}_b, \omega), \quad (3.6)$$

where $\hat{M} = \{m_{ij}(\boldsymbol{r}_f, \boldsymbol{r}_b, \omega)\}$ $(i, j, = x, y, z)$ is induction matrix. Substituting expressions (1.1) and (3.6) into equality (3.5) we readily derive

$$T(\boldsymbol{r}_f, \omega) = \boldsymbol{a}_\tau \cdot \boldsymbol{B}_\tau(\boldsymbol{r}_b, \omega), \quad (3.7)$$

where vector $\boldsymbol{a}_\tau$ has the form

$$\boldsymbol{a}_\tau = (\hat{M}^+\boldsymbol{p})_\tau + (\hat{M}^+\boldsymbol{p} \cdot \boldsymbol{e}_z)\,\boldsymbol{v}_\tau. \quad (3.8)$$

Here the sign «+» denotes matrix $\hat{M}^+$ transposition, $m^+_{ij} = m_{ji}$; $\boldsymbol{e}_z$ is a downward unit vector. The equalities (3.7) and (3.8) show that: 1) intensity $T(\boldsymbol{r}_f, \omega)$ at field sites $\boldsymbol{r}_f$ can be expressed in terms of horizontal field $\boldsymbol{B}_\tau(\boldsymbol{r}_b, \omega)$ at a reference site $\boldsymbol{r}_b$; 2) vectors $\boldsymbol{a}_\tau$ are nothing but transfer functions since they are just some combination of induction vectors v_τ and induction matrices $\hat{M}$.

The next section presents a model study of sensitivity of the vectors $\boldsymbol{a}_\tau$ with respect to the temporal changes in the Earth's resisitivity associated with tectonic processes.

4. Numerical modelling example

The geoelectric model we constructed to evaluate the tectonomagnetic effects, consists of a layered Earth with surface and deep inhomogeneous thin sheets. The model reflects the resistivity distributions in a vicinity of the Tchujski Depression in Kirghizia. The resistivities and thicknesses of five-layered Earth are as follows: $\rho_1 = 1000\ \Omega \cdot$ m and $h_1 = 30$ km; $\rho_2 = 300\ \Omega \cdot$ m and $h_2 = 8$ km; $\rho_3 = 30\ \Omega \cdot$ m and $h_3 = 12$ km; $\rho_4 = 300\ \Omega \cdot$ m and $h_4 = 170$ km; $\rho_5 = 10\ \Omega \cdot$ m. The surface thin sheet represents a conductance distribution of the Depression sediments (fig. 1, upper panel). The conductance varies from 1 S, in Tien Shan mountains surrounding the Depression, up to 300 S inside it. The deep thin sheet, embedded in stratified Earth at a depth of 7 km, contains an isometric crustal conductor (fig. 1, lower panel). The conductor has a diameter of 35 km and simulates a seismogeneric zone disposed at a south fringe of the Tchujski Depression. In order to evaluate the geomagnetic effects due to possible changes in the seismogeneric zone

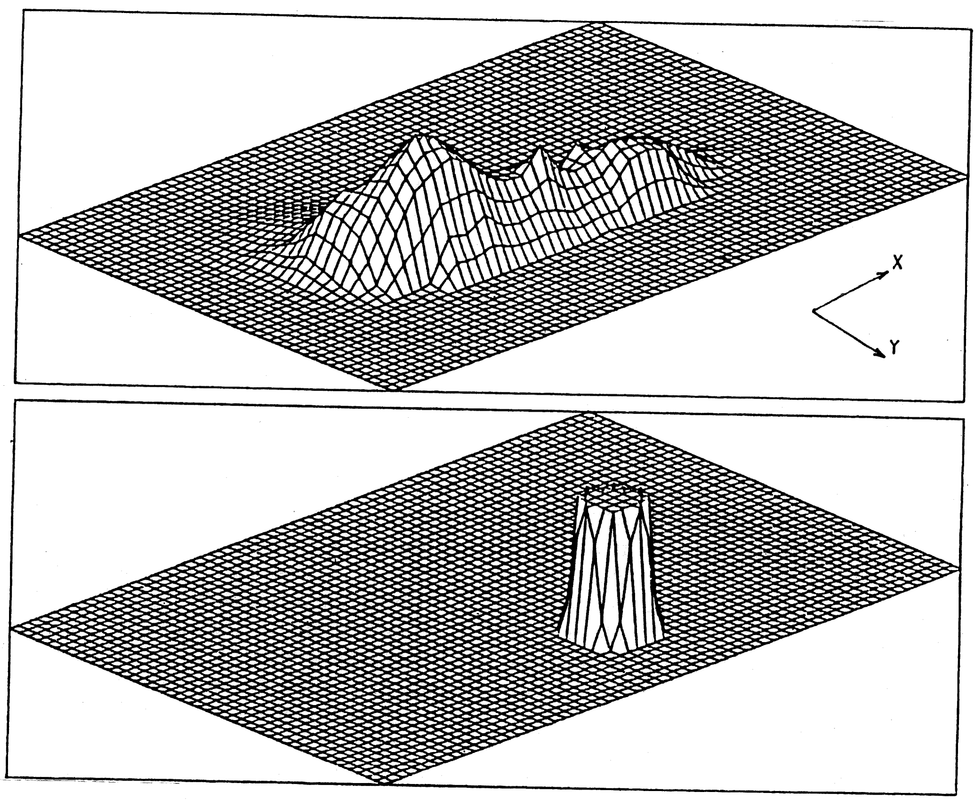
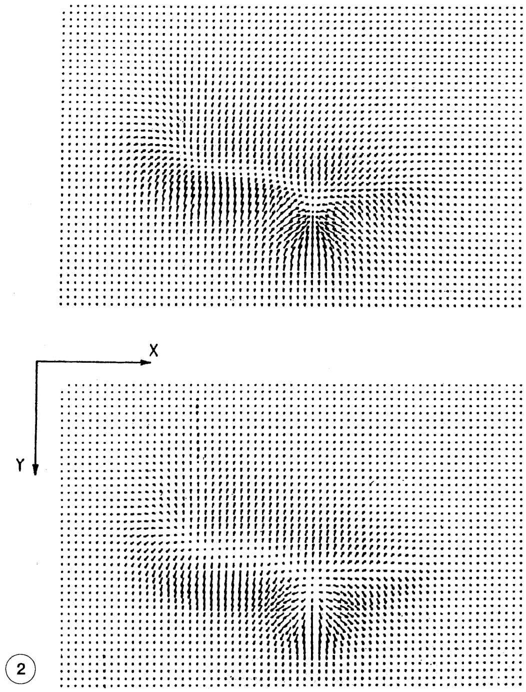


Fig. 1. The surface sheet conductance (upper panel) and the deep sheet conductance (lower panel). A full description is provided in the text.



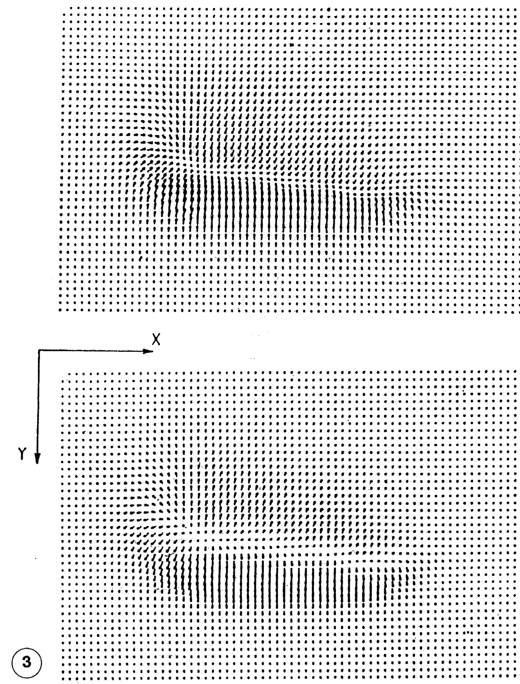


Fig. 2. The real part (upper panel) and imaginary part (lower panel) of induction vectors $\boldsymbol{a}_\tau$ for the case with isometric crustal conductor. On the panels, the scale of the vectors is 0.07 per 10 km. The vectors are shown for the central area of 65×43 cells. The reference site is taken at the upper left corner, cell ($i_x = 16$; $i_y = 27$), of the area shown.

Fig. 3. The same as fig. 2, but for the case without an isometric crustal conductor.

resistivity during earthquake preparation, we varied conductance of the crustal conductor from 250 S up to 1250 S. Background conductance of deep sheet was taken as 250 S.

To calculate the geomagnetic field within the model constructed we used the multi-sheet code (Pankratov, 1991; Singer *et al.*, 1992). We adopted a numerical mesh of 96×96 cells for each thin sheet, with a cell size of 5 km×5 km. The directions of the rectangular axes x and y were chosen as the geographic north and east respectively. The model was excited by x- and y-polarized plane waves of frequency $f = .00083$ Hz. The geomagnetic fields $\boldsymbol{B}(\boldsymbol{r}_f, \omega)$ were calculated at the Earth's surface in the nodes, $\boldsymbol{r}_f$, of the mesh; $\omega = 2\pi f$. Having chosen vector $\boldsymbol{p}$ as for Alma-Ata observatory, $\boldsymbol{p} = (.036, .464, .884)$, from equality (3.5) we derived $T(\boldsymbol{r}_f, \omega)$, for both polarizations. Finally, from equality (3.7) we calculated the induction vector $\boldsymbol{a}_\tau$. The reference site r_b was chosen 125 km to the north and 50 km to the west apart from the Tchujski Depression.

Figure 2 shows the real part (upper panel) and imaginary part (lower panel) of the induction vectors $\boldsymbol{a}_\tau$ for the case when conductance of isometric conductor is 1250 S. Figure 3 presents the vectors for the case when the isometric conductor blends into the background, *i.e.* for the case when its conductance is 250 S. A comparison of the induction vector $\boldsymbol{a}_\tau$ on figs. 2 and 3 demonstrates an impressive geomagnetic effect associated with a five-fold decrease in seismogeneric zone resistivity.

5. Discussion

The proposed technique for monitoring tectonic processes requires the measurements of absolute intensities rather than geomagnetic field components. Let us demonstrate that the horizontal $\boldsymbol{B}_\tau(\boldsymbol{r}_f, \omega)$ and vertical $B_z(\boldsymbol{r}_f, \omega)$ components can be reconstructed in terms of spatial distribution of absolute intensity $T(\boldsymbol{r}_f, \omega)$. Indeed, variations $\boldsymbol{B}(\boldsymbol{r}_f, \omega)$ are decomposed into its normal $\boldsymbol{B}^n$ and anomalous $\boldsymbol{B}^a$ parts as follows

$$\boldsymbol{B}(\boldsymbol{r}_f, \omega) = \boldsymbol{B}^a(\boldsymbol{r}_f, \omega) + \boldsymbol{B}^n(\boldsymbol{r}_f, \omega). \quad (5.1)$$

The normal field $\boldsymbol{B}^n(\boldsymbol{r}_f, \omega)$ is assumed to be known; in practice, this supposition is fulfilled as a rule. Substituting decomposition (5.1) into expression (3.5) we obtain the following equality

$$T(\boldsymbol{r}_f, \omega) - \boldsymbol{p} \cdot \boldsymbol{B}^n(\boldsymbol{r}_f, \omega) =$$
$$= \boldsymbol{p}_\tau \cdot \boldsymbol{B}^a_\tau(\boldsymbol{r}_{\mathrm{f}}, \omega) + p_z B^a_z(\boldsymbol{r}_f, \omega). \quad (5.2)$$

But it is known that the anomalous vertical B^a_z and horizontal $\boldsymbol{B}^a_\tau$ components are connected as follows

$$\boldsymbol{B}^a_\tau(\boldsymbol{r}_f, \omega) = \frac{1}{2\pi}\nabla_\tau \int \frac{B^a_z(\boldsymbol{r}, \omega)}{|\boldsymbol{r}_f - \boldsymbol{r}|} d\boldsymbol{r}. \quad (5.3)$$

Here ∇_τ stands for horizontal gradient with respect to $\boldsymbol{r}_f$; $\boldsymbol{r} = (x, y)$, and $d\boldsymbol{r} = dxdy$. The integral in (5.3) is taken over part of the Earth's surface, where B^a_z is non-zero. Finally, substituting (5.3) into (5.2), we derive integral equation with respect to unknown B^a_z

$$p_z B^a_z(\boldsymbol{r}_f, \omega) + \frac{1}{2\pi}\boldsymbol{p}_\tau \cdot \nabla_\tau \int \frac{B^a_z(\boldsymbol{r}, \omega)}{|\boldsymbol{r}_f - \boldsymbol{r}|} d\boldsymbol{r} =$$
$$= T(\boldsymbol{r}_f, \omega) - \boldsymbol{p} \cdot \boldsymbol{B}^n(\boldsymbol{r}_f, \omega). \quad (5.4)$$

When B^a_z is found from (5.4), horizontal field $\boldsymbol{B}^a_z(\boldsymbol{r}_f, \omega)$ is readily determined from (5.3).

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