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RUSSIAN GEOLOGY AND GEOPHYSICS

Russian Geology and Geophysics 51 (2010) 1219–1226

www.elsevier.com/locate/rgg

TEM surveys for magnetic viscosity of rocks *in situ*

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Received 5 August 2009; accepted 9 April 2010

Abstract

We discuss the results of a field experiment in the Malaya Botuobiya area (West Yakutia) at a site where earlier surveys revealed slowly decaying transient responses. That time-dependent voltage decay indicated magnetic viscosity effects associated with magnetic relaxation of superparamagnetic grains in rocks. In this study, we have applied a high-resolution array TEM survey to contour the anomaly and parametric soundings with systems of different configurations to explore the vertical pattern of magnetic viscosity. The parametric data have been inverted, by means of manual and automated fitting, with a reference model of a layered magnetically viscous earth, using, respectively, analytical formulas and simulation based on a forward solution by separation of variables. According to both automated and manual inversion, the section at the center of the anomalous site fits a three-layer earth model with an intermediate magnetically viscous layer between two nonmagnetic layers. This model is consistent with a priori evidence of local geology and may provide more details of the latter. The inversion results have been further used to estimate the volumetric percentage of superparamagnetic grains in the magnetically viscous layer, assuming magnetite to be the main ferrimagnetic phase.

Keywords: TEM method; flood basalt; magnetic viscosity; superparamagnetic grains; inversion; Yakutia

Introduction

Magnetic viscosity is a property of ferromagnetism. Effects of magnetic viscosity in TEM data are associated with superparamagnetism, or magnetic relaxation of ultrafine grains in ferrimagnetic minerals (Buselli, 1982; Kozhevnikov and Snopkov, 1990, 1995).

According to evidence collected through two recent decades (Barsukov and Fainberg, 2001, 2002; Kozhevnikov and Snopkov, 1990, 1995; Neumann, 2006), magnetic viscosity effects have bearing on composition and structure of uppermost crust, as well as of manmade objects, and can record shallow geological processes. Nevertheless, magnetic viscosity is most often treated as geological noise that interferes with TEM responses to be interpreted in terms of "normal" electrical conductivity (Buselli, 1982; Lee, 1984; Zakharkin et al., 1988; Zakharkin and Bubnov, 1995). This may be the reason why induction survey reports never focus on magnetic viscosity of rocks as a subject of interest. Most often magnetic viscosity has been investigated just because of its casual appearance in TEM data.

Lee (1984), Pasion et al. (2002), and Kozhevnikov and Antonov (2008a, 2009b) specially explored TEM responses of a magnetically viscous earth by means of mathematical modeling. In this paper we develop the earlier approaches and discuss a field experiment performed by Vas.V. Stognii at a site of the Siberian Trap Province in order to investigate magnetic viscosity of igneous rocks *in situ*. We report the results of inversion of the collected magnetic relaxation-affected transients and the respective geological and geophysical implications.

Geological and historic background

Magnetic viscosity of rocks *in situ* was studied at a site in the Irelyakh River–Chuonalyr Brook interfluve in the Malaya Botuobiya area, 30 km west of Mirnyi city (Fig. 1). The superparamagnetic rocks consist of outsized basaltic tuff of the Triassic Chichikan Formation exposed in fault blocks and among dolerite sills. Less abundant rocks at the site are tuffaceous sediments, such as tuffite and tuffaceous sandstone,

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doi:10.1016/j.rgg.2010.10.007

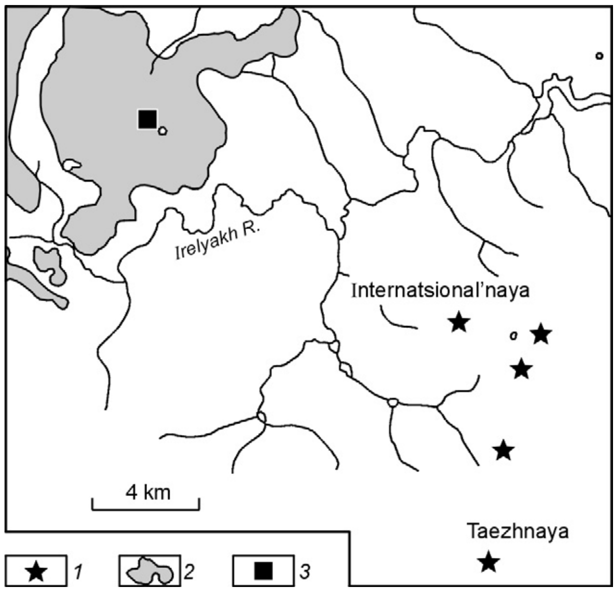

Fig. 1. Location map of study area. *1*, kimberlite, *2*, flood basalt, from aeromagnetic surveys, *3*, site of measurements.

siltstone, and mudstone. The tuff material ejected into the air during volcanic eruptions cooled down rapidly, and ferromagnetic grains crystallized when being of ultrafine superparamagnetic sizes (Worm, 1999). Magnetic relaxation shows up in TEM data as “long tails”, i.e., as an unusually slow voltage decay.

Slowly decaying transients were first observed in the area in the early 1980s during detailed TEM surveys by the Irelyakh group of the *Yakutskgeologiya* Association in search for kimberlite buried under flood basalt. The survey was by a 100×100 m coincident-loop system with an *Impuls-C* instrument. The TEM measurements revealed seventeen local conductivity highs, out of which twelve were in flood basalts overlying Lower Paleozoic carbonates. The responses were the most intense in the vicinity of Lake Siellyakh, and the anomaly was attributed to a pipe-like mafic body, proceeding from the presence of tuff in the upper section in the lake surroundings.

The Irelyakh Geophysical Survey group of the *Soyuzpromgeofizika* Association, with V.A. Sidorov at the head, proposed to set up detailed TEM soundings over the lake itself using a *Kaskad* measurement system. At that time people were unaware that slow voltage decay could result from magnetic relaxation of supeparamagnetic grains. The measured transients were processed with a common approach, i.e., in terms of “normal” electrical conductivity. As a result, a zone of highly conducting carbonate rocks was inferred to occur beneath the lake, though V.A. Sidorov and his colleagues possibly felt that something was wrong about the conductivity-based explanation of the slowly decaying transients (Sidorov et al., 1990).

The following geophysical surveys at the site were run in the middle–late 1990s by a survey group of the *ALROSA* Association (YaNIGP TsNIGRI). The researchers (the team headed by V.A. Vanchugov) already knew that superparamagnetism of flood basalt was the most probable cause of slowly decaying transient responses. The surveys were mainly by a 100×100 m coincident-loop system, and additional measurements with a smaller 25×25 m loop were applied at several points of most intense responses. The latter reduced the eddy current effects and highlighted almost pure signals due to magnetic relaxation (Fig. 2). Yet, it remained unknown which loop configuration was required to identify magnetic viscosity as the cause of slow voltage decay, and, especially, to estimate its magnitude. In this paper we report the results of surveys in 2008, the third geophysical experiment at the site. Thus, about thirty years have elapsed from the discovery of slowly decaying transients to the time when their appropriate interpretation may appear.

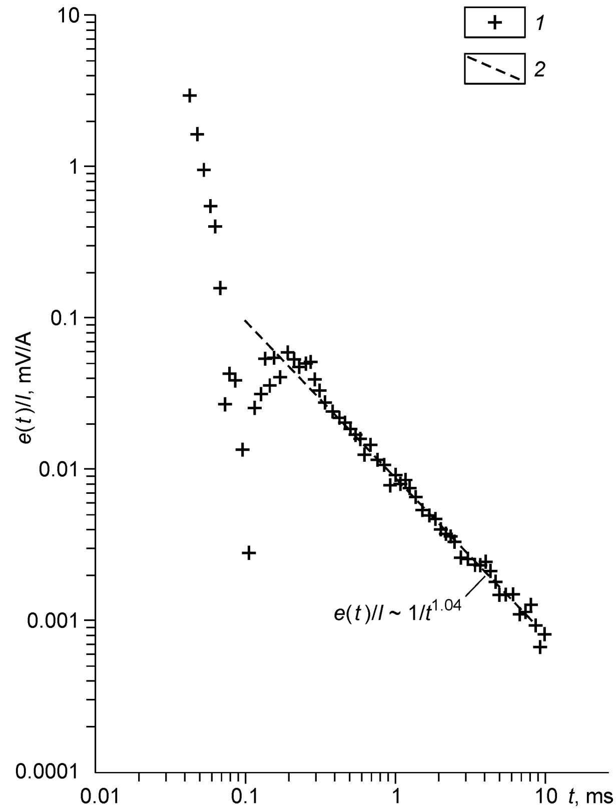

Fig. 2. Coincident-loop (25×25 m) transient response measured in area of Lake Siellyakh. *1*, data points; *2*, power relationship approximating “long tail”. Data courtesy of V.A. Vanchugov (Kozhevnikov and Vanchugov, 1998).

Methods and results

The measurements at the site of detailed surveys were preformed on a *Tsikl* (Russian for *Cycle*) system and consisted of two stages. First it was an areal survey, with a square 75×75 m transmitter and a receiver of an effective area of 10,000 m^2 (equivalent to a 100×100 m loop). There were five noncoincident-loop measurements at each transmitter position (Fig. 3): one with a central-loop configuration (the receiver placed inside the transmitter, at its center) and four with an offset-loop configuration (the receiver placed outside

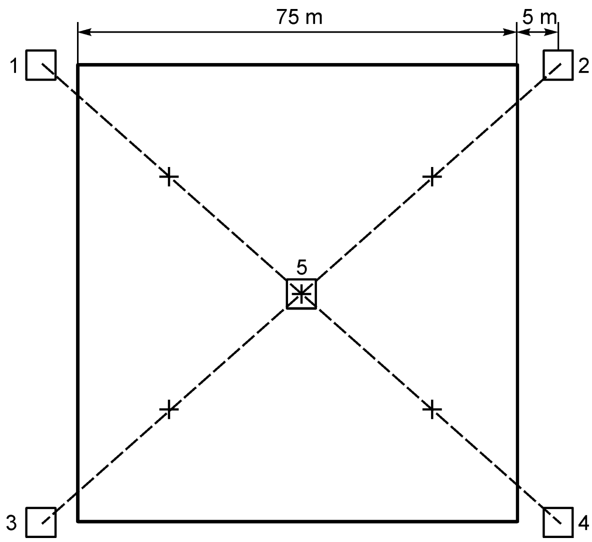

Fig. 3. Loop configuration (1–5 not to scale) in areal TEM measurements. Crosses mark measurement points used in transient voltage mapping.

the transmitter, successively at each of its angles). This is the approach the *ALROSA* Association practices usually in kimberlite exploration in West Yakutia (Stognii and Zhandalinov, 2006; Zhandalinov and Stognii, 2008). Altogether, measurements were run at thirty transmitter positions within an area of 225×300 m.

The measurement results were used to map current-normalized transients (voltage $e(t)/I$) at fixed delay times (see Fig. 4 for the maps at 4.7 ms). The local conductivity pattern is such that at 4.7 ms eddy current becomes vanishing to zero when a 75×75 m transmitter is employed, and the voltage maps thus image the patterns of magnetic viscosity. The data points were at the center of the system in the case of the central-loop configuration and midway between the transmitter center and the center of the respective receiver in the offset-loop configuration (Fig. 3). Mind that, unlike the "normal" transient process (eddy current decay), the responses of a magnetically viscous earth have positive polarity in central-loop data and negative polarity in offset-loop data (Kozhevnikov and Antonov, 2008a).

The two maps based on central-loop and offset-loop data (Fig. 4, *a*, *b*) image an anomalous zone where the magnetic viscosity effects are the most prominent. In both cases the anomaly is located in the central part of the area, and its contours are more or less the same. Voltage decreases to the background on the flanks of the area, where magnetic relaxation effects are thus absent. A small difference in the position of points 1 and 1′ (background signal), 2 and 2′ (anomaly center) between the two maps is due to the fact that the measurement points are different in the central-loop and offset-loop configurations (see above).

Figure 5 shows typical central-loop and offset-loop transient responses from the area flanks and from the anomaly center. The transmitter center was at point 1 in the former case and at point 2 in the latter case (Fig. 4, *a*).

The central-loop response of the anomaly periphery (Fig. 5, *a*) is affected by IP polarization at early times (from 20 to 80 μs), which shows up as a monotony break, or even sign reversal. The offset-loop responses bear no visible IP effects. The reason is in a larger loop spacing in the offset-loop system than in the central-loop one, which makes the IP effects relatively weaker than the eddy current component (Kozhevnikov and Antonov, 2009a; Stognii, 2008; Zhandalinov and Stognii, 2008). At times later than 0.1 ms, the central-loop and offset-loop responses are identical, i.e., eddy current has the greatest contribution to the transients, and the earth is virtually layered. At late times, voltage decreases as t^{-2}, which corresponds to a section with a moderately conducting lower layer.

Unlike the pattern of Fig. 5, *a*, the central-loop response of the anomaly center (Fig. 5, *b*) bears no IP effects, and voltage decreases as $1/t$ at late times. This slow voltage decay is a diagnostic feature of magnetic relaxation (magnetic viscosity effect). The offset-loop response is positive at early times, then experiences a sign reversal about 0.25 ms, rapidly reaches the deepest minimum, and after that it decreases as $1/t$ remaining negative. According to modeling results, these are the patterns expected for offset-loop responses of a

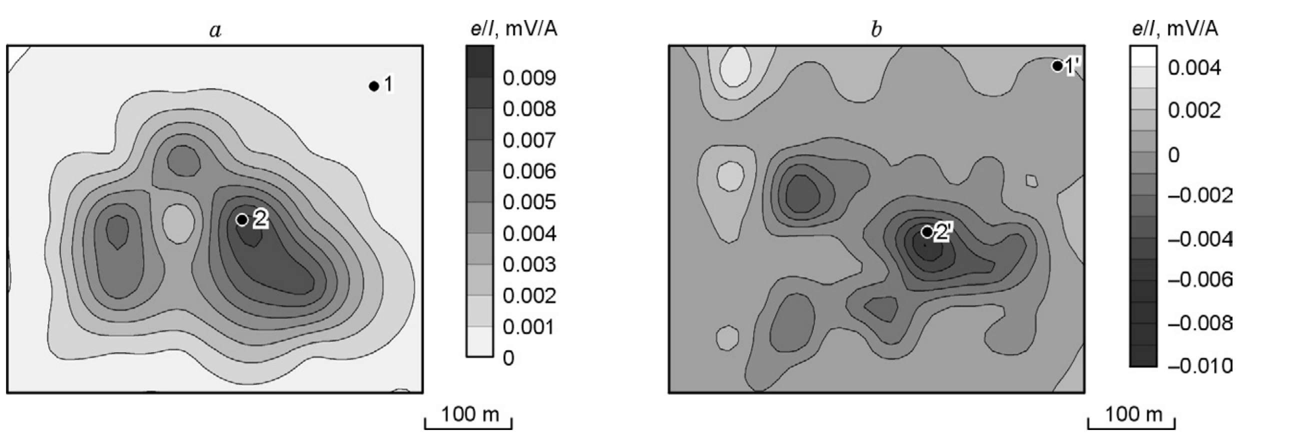

Fig. 4. Maps of current-normalized voltage measured at $t = 4.7$ ms, with a receiver inside (*a*) and outside (*b*) a 75×75 m transmitter loop. Points 1, 1′ and 2, 2′ correspond to background and anomalous signals, respectively.

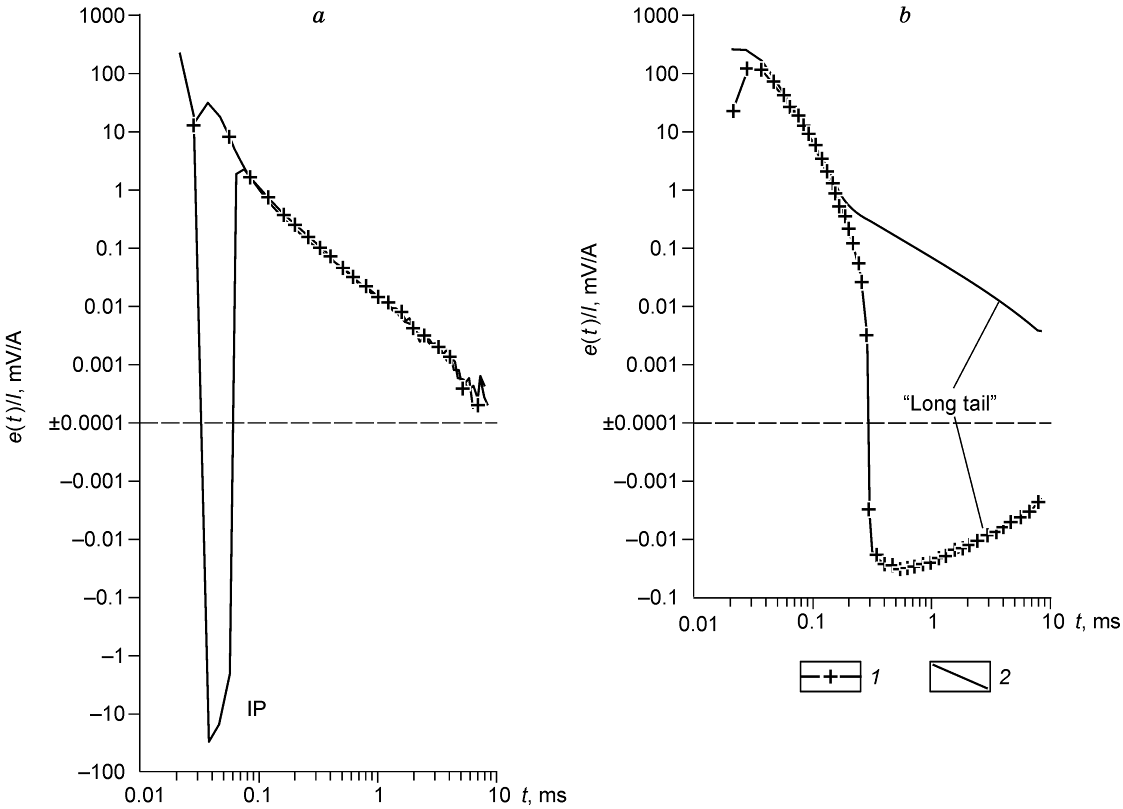

Fig. 5. Transient responses measured at site periphery (point 1 in Fig. 4) and anomaly center (point 2 in Fig. 4). *a*, Central-loop configuration, *b*, offset-loop configuration. *1*, *2*, receiver placed outside (*1*) and inside (*2*) transmitter loop.

magnetically viscous earth (Kozhevnikov and Antonov 2008a).

At the second stage of the experiment, parametric soundings were carried out at the point where the areal survey recorded the most prominent magnetic viscosity effects.

The TEM sounding principle fails with respect to magnetic viscosity because, unlike induced polarization, magnetic relaxation in rocks is independent of the "normal" transient process. That is why *geometrical* soundings (Kozhevnikov and Antonov, 2009b) are required to explore the depth dependence of magnetic viscosity. In the discussed case, geometrical soundings were performed with three central-loop systems (the transmitters 25×25 m, 50×50 m, and 75×75 m) at a single point. The centers of all transmitter-receiver configurations were at point 2 (Fig. 4, *a*). Additionally, TEM signals at the same point were sampled with offset-loop systems of 25×25 m and 50×50 m transmitters and a receiver moved successively to larger distances r from the transmitter center along a line between the system center and the loop side middle. The measurements were at four r lengths (0, 18, 23, and 28 m) with the 25×25 m loop and at three r lengths (0, 30, and 40 m) with the 50×50 m loop. The receiver effective areas were 10,000 m^2 in the case of the 75×75 m transmitter and 400 m^2 in all other cases.

As an illustration, Fig. 6 shows transients at different r values for a 25×25 m transmitter. See a sign reversal in the offset-loop transient at 0.15–0.2 ms, after which the signal decreases in its absolute value as $1/t$ but remains negative. At a fixed delay time, negative voltage is inversely proportional to the loop spacing r and increases as the receiver approaches the transmitter. This is easy to understand taking into account that, other things being equal, the magnetic relaxation-induced voltage is proportional to inductance between the transmitter and receiver loops (Kozhevnikov and Antonov, 2008a). As the loop spacing increases, the transmitter-receiver inductance and, hence, the signal, decrease rapidly thus leading to worse signal/noise ratios, which is clearly seen in the response at $r = 28$ m.

According to the plots of Fig. 6, the sign reversal occurs at late times at a greater spacing. This is likewise consistent with the results reported in (Kozhevnikov and Antonov, 2008a). The time of sign reversal is defined by the interplay between eddy current and magnetic relaxation. The greater the loop spacing, the smaller the transmitter-receiver inductance and, on the other hand, the stronger the induction effect. That is why, as the spacing increases, the component due to magnetic relaxation becomes predominant at later times and, correspondingly, the sign reversal of the total transient occurs later.

Offset-loop transient responses are known to experience another sign reversal produced uniquely by eddy current decay (Kozhevnikov and Antonov, 2008a). For the measurement systems and conductivity patterns in point, the earlier sign reversal occurs at times less than 10 μs, i.e., in the domain where it is almost impossible to record because of the system and instrument inertia.

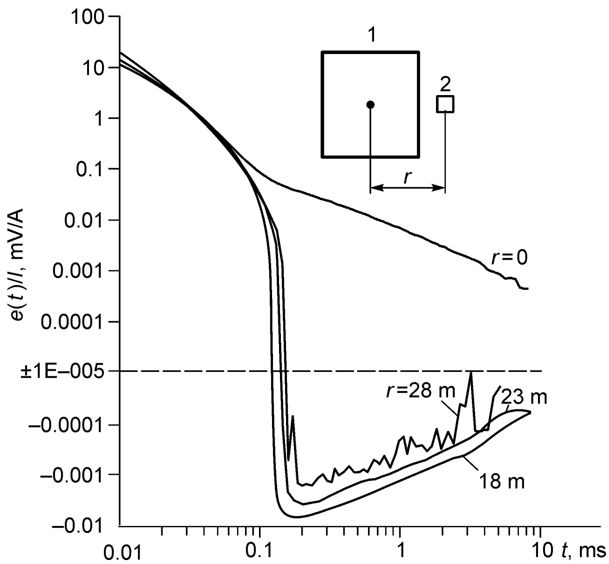

Fig. 6. Transient responses of a magnetically viscous earth measured at different loop spacing values. Center of 25×25 m transmitter was at anomaly center. Above is loop layout.

Inversion of parametric soundings

Procedure summary. Inversion of parametric soundings was made with a reference model of a layered earth with time/frequency-dependent magnetic susceptibility. Magnetic viscosity of rocks is normally due to magnetic relaxation of superparamagnetic grains. Then, the magnetic susceptibility of the ith layer is (Kozhevnikov and Antonov, 2008a, 2009b),

$$\kappa_i(t) = \frac{\kappa_{0i}}{\ln(\tau_{2i}/\tau_{1i})}(B + \ln t),$$

where κ_{0i} is the static magnetic susceptibility, τ_{1i} and τ_{2i} are the lower and upper bounds of the relaxation time for the ith layer, B is a constant, and t is the delay time. The time t of the measured transient response being most often within the gate $\tau_1 \ll t \ll \tau_2$, one may assume that the gate is the same for all layers, i.e., for each layer $\tau_{1i} = \tau_1$, $\tau_{2i} = \tau_2$.

In the frequency domain, the magnetic susceptibility of the ith layer is (Lee, 1984)

$$\kappa_i(\omega) = \kappa_{0i}\left[1 - \frac{1}{\ln(\tau_{2i}/\tau_{1i})}\ln\frac{(1 + j\omega\tau_{2i})}{(1 + j\omega\tau_{1i})}\right],$$

where $j = \sqrt{-1}$; ω is the angular frequency, in s^{-1}. In practice, the frequency ω is almost always within $1/\tau_1 \ll \omega \ll 1/\tau_2$. Then, as in the time-domain case, one may assume that the gate is the same for all layers, i.e., for each layer $\tau_{1i} = \tau_1$, $\tau_{2i} = \tau_2$.

Fitting in search for the best model was manual and automated. Manual fitting continued until the interpreter found the fit between the synthetic and field data reasonable. In the automated fitting, inversion of the measured transients consisted in search for the set of parameters **P** from the model data space **M** that corresponded to the minimum of the objective function $\varphi(\mathbf{P})$:

$$\varphi(\mathbf{P}) = \left\{\frac{1}{N-1}\sum_{j=1}^{N}\left[\frac{\varepsilon^{\text{meas}}(t_j) - F_{\mathbf{P}}(t_j)}{\delta(t_j)\,\varepsilon^{\text{meas}}(t_j)}\right]^2\right\}^{1/2}, \quad (1)$$

where t_j is the jth delay time, $\varepsilon^{\text{meas}}$ is measured transient voltage, N is the total number of delays, $F_{\mathbf{P}}$ is the forward operator, and $\delta(t_j)$ is the relative error of measurements for the time t_j. The function $\varphi(\mathbf{P})$ is a mean weighted sum of square relative differences between the computed and measured transients. The set of model parameters is the vector $\mathbf{P} = (\kappa_{0i}, h_i)\big|_{i=1,M}$, where M is the total number of layers, κ_{0i} is the static magnetic susceptibility, and h_i is the thickness of the ith layer. The minimization was by a modified method of Nelder and Mead (1965) which does not require calculating the derivatives of the forward function in the equation for the minimization functional.

The synthetic transients were generated likewise in two ways, correspondingly. In manual fitting, the approach was based on relation between magnetic relaxation of rocks and the magnetic flux it produces in the receiver. In the case of automated inversion, the Helmholtz equation in a boundary-value problem for a layered earth was solved using the Fourier transform with frequency-dependent magnetic permeability. For details of both ways see (Kozhevnikov and Antonov, 2008a, 2009b). The delay time being within 0.01–10 ms, we assumed that $\tau_1 = 10^{-6}$ s and $\tau_2 = 10^6$ s in the two cases, respectively.

Inasmuch as one of the objectives of the reported study was to find out whether TEM data are applicable, in principle, to estimate magnetic viscosity, obtaining independent estimates of magnetic viscosity patterns was especially important. The problem was solved due to the use of different forward solutions and inversion approaches. Furthermore, the automated and manual inversion procedures were carried out by different interpreters (E.Yu. Antonov and N.O. Kozhevnikov).

Inversion results: manual fitting. Curves in Fig. 7, *a* are central-loop transients measured by systems of 25×25 m, 50×50 m, or 75×75 m transmitters and a 0.8×0.8 m multicoil receiver (effective area 400 m^2). Voltage decreases as $1/t$ after 0.1–0.2 ms, and its values at $t \geq 0.1$–0.2 ms are applicable to inversion for time-dependent apparent susceptibility (Kozhevnikov and Antonov, 2008a, 2009b).

Thus, in manual fitting we used voltages measured at $t = 1$ ms, with loop sizes as above (diamonds in Fig. 7, *b*). The transmitter size in this case of a central-loop system is an analog of loop spacing as a control of effective penetration depth.

Solid line in Fig. 7, *b* is the model curve; Fig. 7, *c* shows the corresponding model of a layered earth, with a ~40 m thick magnetically viscous (κ_{02} = 0.03 SI units) layer between two nonmagnetic ones: a thin (H_1 = 3 m) nonmagnetic (κ_{01} = 0) layer above and a nonmagnetic (κ_{03} = 0) layer below.

Inversion results: automated fitting. Figure 8 illustrates the potentiality of automated inversion with an example of data measured at different positions of the receiver lying on

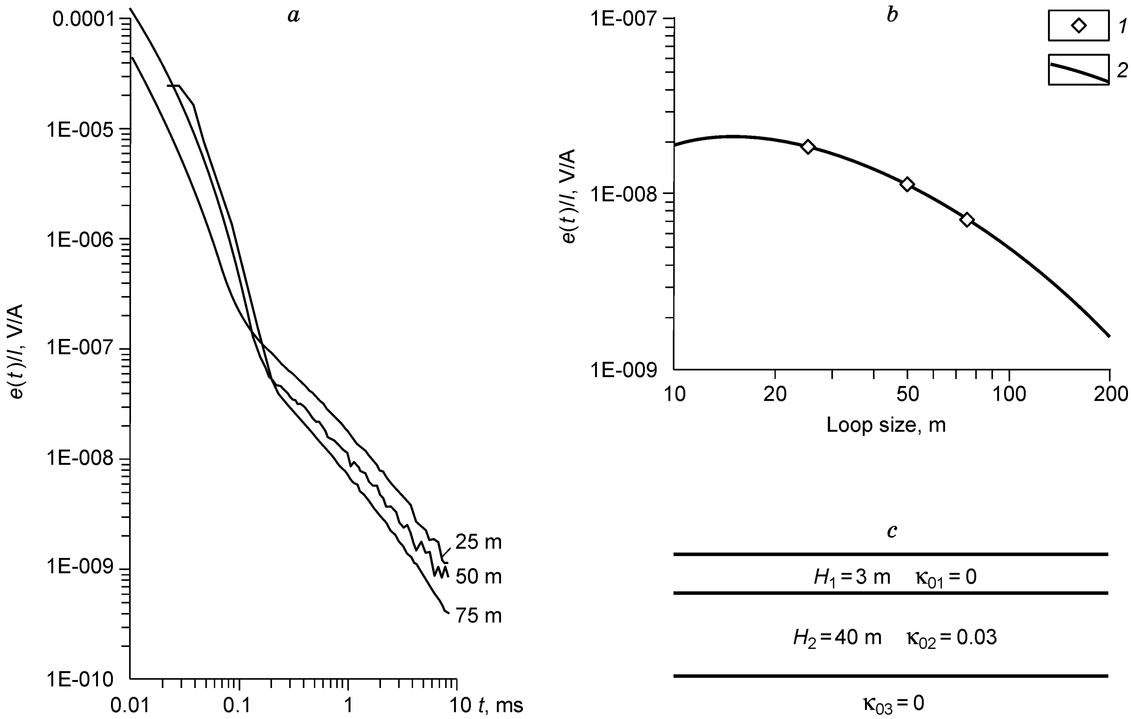

Fig. 7. Manual inversion of parametric sounding data. *a*: Measured central-loop transients, 25 m × 25 m, 50 m × 50 m, and 75 m × 75 m transmitters; *b*: measured and theoretical curves of voltage vs. loop size at t = 1 ms; *c*: earth model. Voltage in a and b is normalized to 1 m^2 receiver area. *1*, experiment, *2*, model.

a straight line between the center and the side middle of a 25 × 25 m transmitter.

Solid lines (Fig. 8) are synthetic transients computed for a model obtained by minimization of functional (1). Note that the model is a result of *joint* inversion in which measured data from each of the four systems (with r = 0, 18, 23, and 28 m, Fig. 6) were joined into a *single* data vector and the subsequent fitting was to the functional minimum for that vector. See Table 1 for the model parameters and (Kozhevnikov and Antonov, 2008b, 2009a) for details of the procedure and its application to joint inversion of transients affected by fast-decaying IP.

There is an additional explanation needed in comments to Fig. 8. First, we used only the right-hand branches of TEM curves in the inversion as they bear signature of magnetic viscosity which is the subject of the study. Second, the forward problem was solved with regard to the finite duration of transmitter current pulses, which allowed us to explain the voltage decay faster than 1/t on the curve “tails”, i.e., at latest times.

The result of joint automated inversion in Fig. 8 is not the only one. In the same way we processed data collected with a 50 × 50 m transmitter at r = 0, 30, and 40 m (Fig. 3), as well as transients measured at point 2 by three central-loop systems. The parameters of the respective models are listed in Table 1.

Discussion

The results of manual and automated inversion for different loop configurations (Table 1) indicate a local structure around point 2 corresponding to a three-layer earth: nonmagnetic upper and lower layers (κ_{01} = κ_{03} = 0), the upper one as thin as ~3 m, and a magnetically viscous layer in the middle. For

Table 1. Model parameters found by manual and automated inversion

Parameter	Manual inversion	Automated inversion		
	Transmitter 25 m × 25 m, 50 m × 50 m, 75 m × 75 m; r = 0	Transmitter 25 m × 25 m, 50 m × 50 m, 75 × 75 m; r = 0	Transmitter 25 m × 25 m; r = 0, 18, 23, 28 m	Transmitter 50 m × 50 m; r = 0, 30, 40 m
κ_{01}, SI unit	0	0	0	0
H_1, m	3	3.2	2.6	3.1
κ_{02}, SI unit	3×10^{-2}	3.1×10^{-2}	2.6×10^{-2}	2.5×10^{-2}
H_2, m	40	41	93	121
κ_{03}, SI unit	0	0	0	0

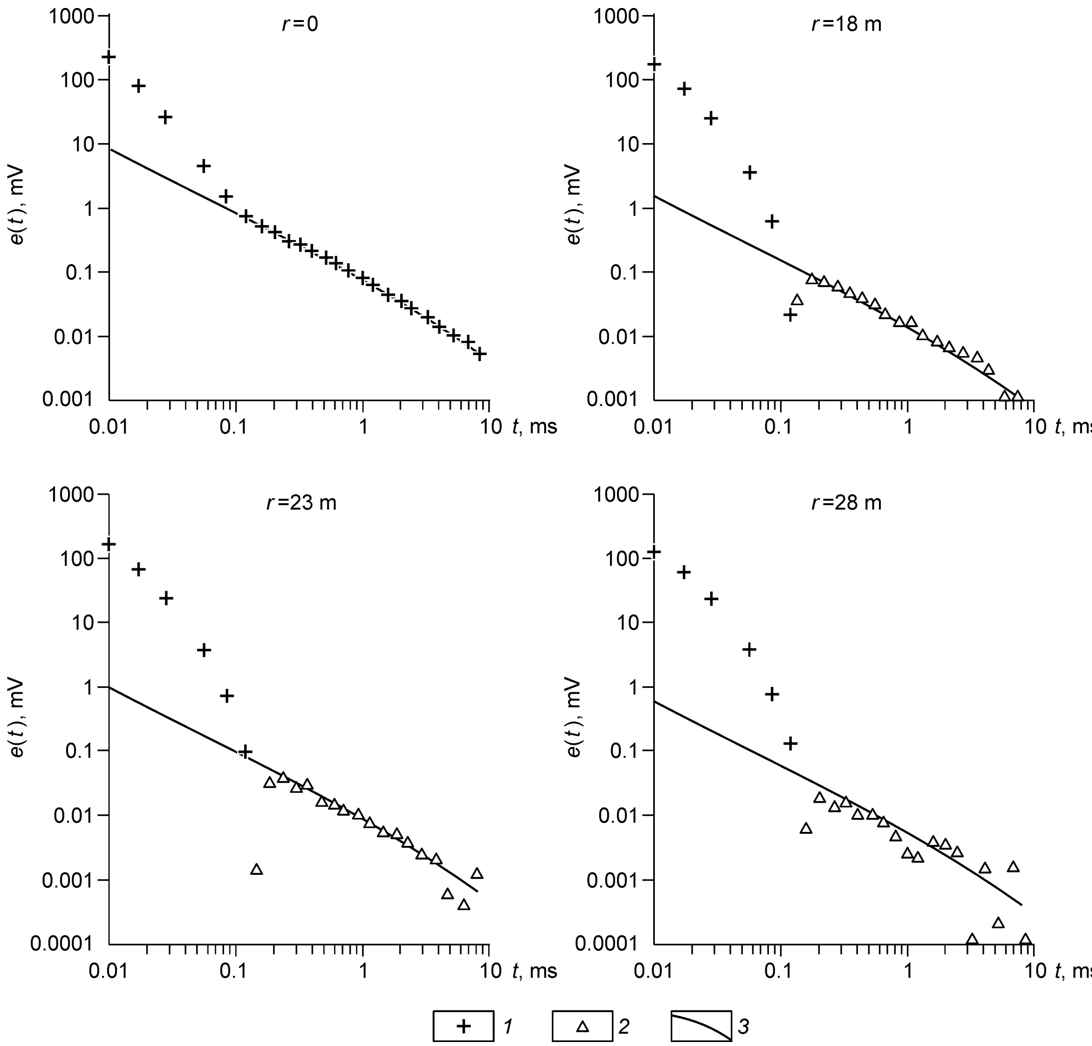

Fig. 8. Automated inversion of parametric sounding data. 25×25 m transmitter. *1*, *2*, measured positive (*1*) and negative (*2*) transients; *3*, computed transient. Effective receiver area 400 m^2, transmitter current 11.7 A.

the latter, both manual and automated fitting procedures give a stable static susceptibility of κ_{02} = 0.03 SI units, at any loop configuration, but the thickness H_2 varying from 40 to 120 m in different models.

In the course of inversion we found out that even a minor difference in the thickness H_1 of the upper layer and in the susceptibility κ_{02} of the intermediate layer from the respective values in Table 1 led to a large misfit between the measured and computed data. Therefore, transients are highly sensitive to these model parameters, which makes us expect our estimates to be rather accurate. As for the thickness of the magnetically viscous layer H_2, its considerable uncertainty may be due to the fact that even the largest loop size of 75×75 m is too small to resolve well the layer bottom. The model curve (Fig. 7, *b*) will change very little if one assumes, for instance, H_2 = 50 m instead of H_2 = 40 m. According to a numerical experiment, the inversion quality for H_2 can improve if additional measurements with a 200×200 m transmitter are applied.

Generally, the similarity of the models resulting from inversion of central-loop and offset-loop transients is evidence that a layered earth is a reasonable approximation of the site structure. The model appears quite realistic being consistent with what is known of the local geology: the 3 m thick upper nonmagnetic layer corresponds to Quaternary sediments, the intermediate magnetically viscous layer consists of Triassic basaltic tuff and volcanic-sedimentary rocks, and the lower layer represents weakly magnetic Paleozoic carbonates.

TEM data can be further used to infer the content of superparamagnetic particles assuming them to be fine grains of magnetite. Dividing the TEM-derived magnetic susceptibility (κ_{02} = 0.03 SI units) by the volumetric magnetic susceptibility of a superparamagnetic magnetite grain (230 SI units) (Emerson, 1980) gives the volumetric concentration 1.3×10^{-4}, or 0.013% magnetite. Correspondingly, the weight percent of ultrafine grains of magnetite (which has a density of 4.7 g/cm^3, the density of tuff being 2 g/cm^3 on average (Kobranova, 1986)) is of the order of 0.03%. Thus, the TEM method is a good tool for detecting superparamagnetic grains in basaltic rock *in situ* and for estimating their content.

In this paper we have applied the term “nonmagnetic” to rocks that are not magnetically viscous, for the sake of simplicity. However, this is not true in the general case because no rock is really nonmagnetic. Besides viscous magnetization, all rocks possess “normal” induced magnetization, which appears almost instantaneously on the time scale of the experiment. The normal induced component influences low-frequency induction data (Blokh et al., 1986) but is mute in the TEM method.

Conclusions

Studying magnetic viscosity of rocks *in situ* is among priority objectives of transient electromagnetic surveys and may be an essential element of rock magnetism research. Above we have reported the results of a field experiment at a site where earlier surveys revealed slowly decaying transient responses indicating strong magnetic viscosity effects. In this study, we have applied high-resolution array TEM soundings to contour the anomaly and parametric soundings with systems of different configurations to explore the vertical pattern of magnetic viscosity.

The parametric data were inverted with a reference model of a layered magnetically viscous earth, by means of manual and automated fitting. For this we have made use of analytical formulas in the former case and a numerical Fourier solution to the Helmholtz equation in the latter case. According to both automated and manual inversion, the section at the center of the anomalous site fits a three-layer earth model with an intermediate magnetically viscous layer between two nonmagnetic layers. This model is consistent with available evidence of local geology and may provide more details of the latter. The inversion results have been used to estimate the volumetric percentage of superparamagnetic grains in the magnetically viscous layer, with regard to magnetite being the main carrier of ferromagnetism.

The discussed results present an early experience in applying the TEM method to purposeful investigation into magnetic relaxation of rocks *in situ*. There are still a number of important issues that remain beyond this consideration and will be a subject of a special study, namely, equivalence, resolution, and influence of measurement errors on the inversion quality.

The study was supported by grant 10-05-00263 from the Russian Foundation for Basic Research.

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Editorial responsibility: M.I. Epov