

# Fast-decaying IP in frozen unconsolidated rocks and potentialities for its use in permafrost-related TEM studies

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## ABSTRACT

We investigate the early time induced polarization (IP) phenomenon in frozen unconsolidated rocks and its association with transient electromagnetic (TEM) signals measured in northern regions. The distinguishing feature of these signals is the distortion of the monotony or sign reversals in the time range from a few tens to a few hundreds of microseconds. In simulating TEM data, the IP effects in frozen ground were attributed to the dielectric relaxation phenomenon rather than to the frequency-dependent conductivity. This enabled us to use laboratory experimental data available in the literature on dielectric spectroscopy of frozen rocks. In our studies we focused on simulating the transient response of a coincident-loop configuration in three simple models: (i) a homogeneous frozen earth (half-space); (ii) a two-layered earth with the upper layer frozen; (iii) a two-layered earth with the upper layer unfrozen. The conductivities of both frozen and unfrozen ground were assumed to exhibit no frequency dispersion, whereas the dielectric permittivity of frozen ground was assumed to be described by the Debye model. To simplify the presentation and the comparison analysis of the synthetic data, the TEM response of a frozen polarizable earth was normalized to that of a non-polarizable earth having the same structure and resistivities as the polarizable earth. The effect of the dielectric relaxation on a TEM signal is marked by a clearly defined minimum. Its time coordinate  $t_{\min}$  is approximately three times larger than the dielectric relaxation time constant  $\tau$ . This suggests the use of  $t_{\min}$  for direct estimation of  $\tau$ , which, in turn, is closely associated with the temperature of frozen unconsolidated rock. The ordinate of the minimum is directly proportional to the static dielectric permittivity of frozen earth. Increasing the resistivity of a frozen earth and/or decreasing the loop size results in a progressively stronger effect of the dielectric relaxation on the TEM signal. In the case of unfrozen earth, seasonal freezing is not likely to have an appreciable effect on the TEM signal. However, for the frozen earth, seasonal thawing of a near-surface layer may result in a noticeable attenuation of the TEM signal features associated with dielectric relaxation in a frozen half-space. Forward calculations show that the dielectric relaxation of frozen unconsolidated rocks may significantly affect the transient response of a horizontal loop laid on the ground. This conclusion is in agreement with a practical example of inversion of the TEM data measured over the permafrost.

## INTRODUCTION

Geophysical methods have been widely used in research on frozen ground and engineering studies in northern regions. Although in the last few decades significant progress has been

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made in permafrost geophysics (Scott, Sellmann and Hunter 1997), some problems in permafrost-related studies still remain unsolved.

The cryolithosphere, especially its upper part, is relatively inhomogeneous and to geophysical exploration it represents a kind of 'geological noise', hindering the search for and prospecting of such targets as ore deposits, kimberlite pipe, ground water, etc. At the same time, the investigation of the cryolithosphere is important in itself. In the research and the exploration of permafrost regions it is necessary to map the distribution of frozen ground and to estimate its temperature and ice content. Geoelectrical methods have an advantage in the study of the cryolithosphere due to the difference between the electrical properties of thawed and frozen rocks and due to the dependence of these properties on the temperature, water/ice content and other permafrost-related parameters. However, since such dependences are ambiguous, the solution of the above problems is quite complicated and not always possible. In particular, the interpretation of resistivity data in terms of geocryologic parameters may represent a challenge to a geophysicist, since the resistivity of frozen ground is affected by many variables (Scott *et al.* 1997).

The most promising geophysical methods in permafrost-related studies are those relating to the physical properties that are particularly characteristic of frozen ground. In this respect, the fast-decaying induced-polarization processes in frozen unconsolidated rocks (Kozhevnikov, Nikiforov and Snopkov 1995) deserve special consideration. These processes were found to affect TEM data measured in cold regions such as Yakutiya, Russia (Molchanov *et al.* 1984; Vanchugov and Kozhevnikov 1998), Alaska (Walker and Kawasaki 1988) and Northern Canada (Smith and Klein 1996).

#### FAST-DECAYING IP IN FROZEN GROUND

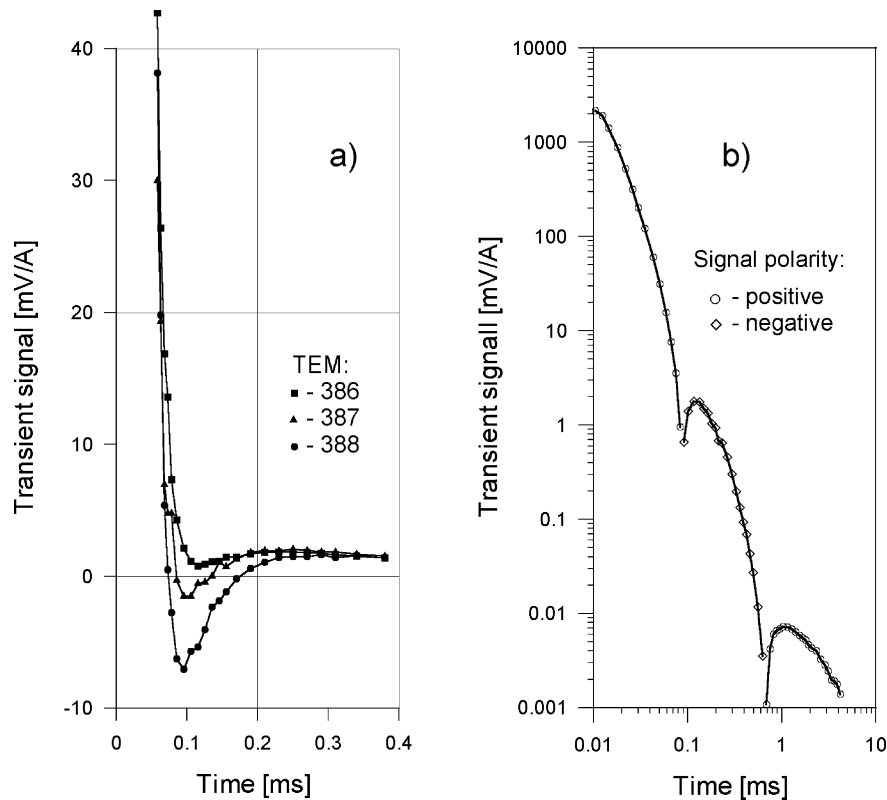
In the late 1960s and early 1970s, reports on non-monotonous TEM transients appeared in the geophysical literature. Distortions in the monotony, including sign reversals, were observed in the time interval 0.1–1 ms. These signals, especially when measured with single-loop or coincident-loop configurations, could not be explained in the context of the conventional theory of TEM prospecting methods (Weidelt 1983). The problem was solved by assuming the anomalous frequency-dependent conductivity  $\sigma(\omega)$  of the geological materials, which accounts for the well-known induced polarization (IP) phenomenon investigated by the electrical IP prospecting method (Wait 1982; Spies and Frischknecht 1991).

Since the early 1980s, numerous TEM measurements in Western Yakutiya, Russia, have shown transient signals with distortions in monotony and sign reversals, which occurred in the time range from a few tens to a few hundreds of microseconds. These early time distortions (Fig. 1) were initially considered to be noise, hampering the inversion of TEM signals in terms of the 'normal' conductivity. It has been proposed that these distortions were caused by the fast-decaying induced polarization of frozen ground (Molchanov *et al.* 1984; Sidorov 1987). The nature of this polarization remained unclear, resulting in much controversy. A variety of mechanisms which would account for the polarization of frozen earth were discussed: large-scale heterogeneity of frozen rocks resulting in the strong Maxwell–Wagner polarization (Sidorov 1987), dielectric relaxation of ice (Kozhevnikov 1991), fractal geometry of the frozen rock texture (Krylov and Bobrov 1996), and small-scale heterogeneity in combination with dielectric relaxation of ice and unfrozen water films (Artyomenko and Kozhevnikov 1999).

Thousands of TEM signals indicating fast-decaying IP effects were measured in Western Yakutiya during diamond prospecting. The areas where the IP-affected transients occurred account for about half of the total survey area. The geology of Western Yakutiya consists of Quaternary unconsolidated sediments, Jurassic sand and clay, Lower Palaeozoic terrigenous-carbonate sedimentary and other widely occurring rocks. Only a limited number of IP-affected TEM signals are reported to be occasionally measured in TEM surveys throughout the world. This gives rise to the question: what is the distinguishing feature of Western Yakutiya which causes the earth to be polarizable? In our opinion, one feature is beyond question: the cold. It would appear reasonable that upon freezing the ground becomes more polarizable at early times. This speculation is supported by TEM surveys in Alaska and in Northern Canada (Walker and Kawasaki 1988; Smith and Klein 1996), where IP-affected TEM signals were measured over the permafrost.

The assumption that fast-decaying IP is an inherent feature of frozen rocks was demonstrated by a specially designed field experiment (Kozhevnikov *et al.* 1995). The main objective of their experiment was to answer the question: does the transient response of a particular rock depend on whether it is frozen or thawed, and if it does, in what manner?

The test site was located within a discontinuous permafrost area near the village of Taksimo in the Muya River valley in the north of Buryatiya, Russia. According to both borehole and integrated geophysical survey data, the site lithology to a



**Figure 1** Transients measured over permafrost in West Yakutiya, Russia: (a) Nakyn kimberlite field, central-loop configuration with 200 m transmitter loop and 100 m by 100 m receiver loop; (b) near a kimberlite pipe in the vicinity of the city of Mirny, 100 m by 100 m coincident-loop array.

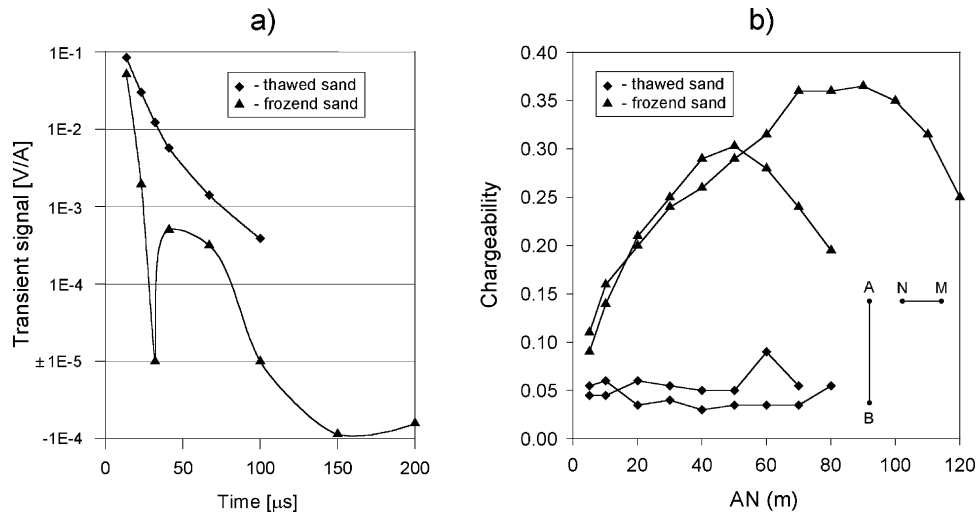
depth of 100–200 m consists of water-saturated quartz sand. The water content within the sand layer varies from a few percent to tens of percent. The absence of a clay fraction excluded membrane polarization on clay minerals. The mean temperature of frozen sand is  $-(1-2)^{\circ}\text{C}$ . The TEM responses of thawed and frozen sand, measured with a 100 m by 100 m coincident-loop configuration, are shown in Fig. 2(a). As can readily be seen, the early time TEM response of frozen sand is characterized by strong distortions of the monotonous behaviour.

In parallel with the TEM survey, IP measurements at early times (0.1–10 ms) were carried out with a grounded dipole-dipole array. Seguin and Frydecki (1990) found that the chargeability of frozen unconsolidated rocks measured in the time interval 200–1200 ms is very low. However, measurements at the delay time of 0.1 ms showed low chargeabilities for thawed sand and high chargeabilities (up to 37%) for the frozen sand (Fig. 2b). These data favour the assumption that the distortions of the early time TEM response result from the IP effects.

Subsequently, the strong galvanic IP response of frozen unconsolidated rocks at delay times of 0.1–10 ms was confirmed by Karasyov *et al.* (2004) on the basis of extensive laboratory and on-land measurements. Thus, there is a good reason to believe that measuring early time IP in frozen ground may result in substantial improvements in the efficiency of electrical methods when applied to permafrost-related studies. Since galvanic contact with frozen ground may represent a severe problem, the TEM method has an advantage over the traditional galvanic IP technique.

The effect of induced polarization on TEM signals has been reported and discussed by Spies (1980), Lee (1981), Raiche (1983), Molchanov *et al.* (1984), Raiche *et al.* (1985), Smith and West (1989), Smith and Klein (1996), El-Kaliouby *et al.* (1997), Descloitres *et al.* (2000), and in many other papers.

Unfortunately, there are only a few publications in which the effect of fast-decaying IP in frozen ground on the TEM signal has been discussed and estimated on the basis of computer simulations (Walker and Kawasaki 1988; Krylov and Bobrov 1995, 1996; Krylov, Bobrov and Soroka 1995; Krylov,



**Figure 2** TEM and galvanic IP field data from a test site in the Muya River valley in the north of Buryatia, Russia (Kozhevnikov *et al.* 1995): (a) TEM response of the 100 m by 100 m coincident loops; (b) chargeability sounding curves measured at a delay time of 0.1 ms. An orthogonal sounding configuration AB-MN was used to minimize the induction effects.

Bobrov and Wachter 1998). This paper aims to compensate for the lack of modelling results and should be considered as a first step towards understanding how the early time IP phenomenon in frozen ground influences TEM signals. It is hoped to provide a rough idea of the applicability and limitations of the TEM method in permafrost-related studies.

### ELECTRICAL MODEL FOR CALCULATING FAST-DECAYING IP IN FROZEN UNCONSOLIDATED ROCKS

Traditionally, the IP phenomenon is explained and simulated in terms of the frequency-dependent, complex conductivity  $\sigma^*(\omega)$  of geological materials. The dielectric permittivity  $\epsilon$  is assumed to be independent of frequency and to have no effect on IP signals measured in geoelectric studies. In simulating IP signals, the empirical Cole-Cole formula is used to describe  $\sigma^*(\omega)$  (Wait 1982), i.e.

$$\sigma^*(\omega) = \sigma_0 \left[ 1 + \frac{m(j\omega\tau)^c}{1 + (j\omega\tau)^c(1-m)} \right], \quad (1)$$

where  $j = \sqrt{-1}$ ,  $\omega$  denotes angular frequency,  $\sigma_0$  denotes dc conductivity,  $m$  denotes chargeability,  $\tau$  is the IP time constant, and  $c$  is the exponent with limits 1 (a single relaxation) and 0 (an infinitely broad, continuous distribution). The chargeability  $m$  can be expressed as

$$m = \frac{\sigma_\infty - \sigma_0}{\sigma_\infty}, \quad (2)$$

where  $\sigma_\infty$  is the high-frequency limit of conductivity (in practice,  $\sigma_\infty$  is the conductivity at a frequency that is much higher than the relaxation frequency  $\omega_\theta = 1/\tau$ ).

In our study, we used an alternative model to describe the earth's electrical properties. In this model, the conductivities of both frozen and unfrozen ground were assumed to exhibit no frequency dispersion, whereas the dielectric permittivity  $\epsilon^*(\omega)$  of frozen ground was assumed to be described by the Debye formula,

$$\epsilon^*(\omega) = \epsilon_0 \left( \epsilon_\infty + \frac{\epsilon_s - \epsilon_\infty}{1 + j\omega\tau} \right), \quad (3)$$

where  $\epsilon_0 = 8.85 \times 10^{-12}$  F/m is the dielectric permittivity of free space,  $\tau$  is the time constant of dielectric relaxation,  $\epsilon_s$  and  $\epsilon_\infty$  are the relative dielectric permittivities at frequencies that are, respectively, much higher and lower than the relaxation frequency  $\omega_\theta = 1/\tau$ . It is common practice to represent and to interpret the results of laboratory measurements of the electrical properties of frozen rocks in terms of non-dispersive conductivity and complex frequency-dependent dielectric permittivity. Therefore, in our investigations of IP effects in TEM measurements, we were able to use laboratory experimental data on dielectric spectroscopy of frozen materials.

The dielectric spectroscopy of frozen unconsolidated rocks has been reported by Olhoeft (1977), Bittelli, Flury and Roth (2004) and Frolov (1998). The latter reviews many theoretical considerations and the experimental data on the electrical properties of ice and frozen rocks.

The above literature shows that the dielectric permittivity of frozen unconsolidated rocks invariably decreases with increasing frequency; that is, dielectric relaxation is an inherent characteristic of frozen unconsolidated rocks. For all types of frozen unconsolidated rocks, decreasing temperature results in (i) decreasing low-frequency dielectric permittivities, and (ii) an increase in the effective time constant of dielectric relaxation.

The available data on the high-frequency dielectric permittivities of frozen soils and rocks agree well. Many experiments have shown that the high-frequency dielectric permittivity  $\varepsilon_\infty$  is usually in the range from 3 to 10. At high frequencies, the dielectric permittivity of frozen rocks is correctly predicted using mixing formulae based on the theory of composite dielectrics.

Unlike high-frequency dielectric permittivity, the available experimental data on the low-frequency dielectric permittivity  $\varepsilon_s$  of frozen porous materials are not nearly so consistent, especially for materials with a high content of clay minerals. According to Olhoeft (1977) and Bittelli *et al.* (2004), dielectric permittivities of frozen clay, measured at frequencies of about 1 kHz, range from  $10^3$  to  $10^4$ . As the frequency decreased below 1 kHz, a further increase in the dielectric permittivity was observed without it tending towards an asymptote. Measurements carried out by Frolov (1998) indicated that at 1.5 kHz the dielectric permittivity of a frozen loam does not exceed 100; this is 1–2 orders of magnitude lower than the permittivities measured by Olhoeft (1977) and Bittelli *et al.* (2004). Frolov (1998) attributed the discrepancy between the clay-related results to the different features of the measurement methods and techniques used in the various experiments.

Excellent reviews of the dielectric properties of porous rocks have been given by Chelidze and Gueguen (1999) and Chelidze, Gueguen and Ruffet (1999). The nature of electrical polarization in frozen rocks has been discussed by Frolov (1998) and Bittelli *et al.* (2004). Low-frequency dielectric permittivities of frozen unconsolidated rocks and soils have been found to be dominated by the Maxwell–Wagner effect and surface conduction and polarization of the electrical double layer. The effective dielectric permittivity of frozen rocks increases with the surface-to-volume ratio of the system. Surface conduction and electrical double-layer polarization become more pronounced as particle size decreases, and they dominate the low-frequency dielectric permittivity in clay soils.

In a coarse-grained rock, such as gravel or sand, even at temperatures just below 0°C, practically all water turns to ice (King, Zimmerman and Corwin 1988). Because of this,

the low-frequency dielectric permittivity  $\varepsilon_s$  of a frozen coarse-grained rock has only a weak dependence on the temperature and is controlled mainly by the ice content  $W_i$ . Thus, for frozen sands (Frolov 1998), we have

$$\varepsilon_s \approx a + bW_i, \quad (4)$$

where  $a = 120 \pm 20$ ,  $b = 12 \pm 3$ .

The relaxation time constant  $\tau$  for fresh polycrystalline ice is given (in seconds) by King and Smith (1981) as

$$\log \tau = 2900/T - 15.3,$$

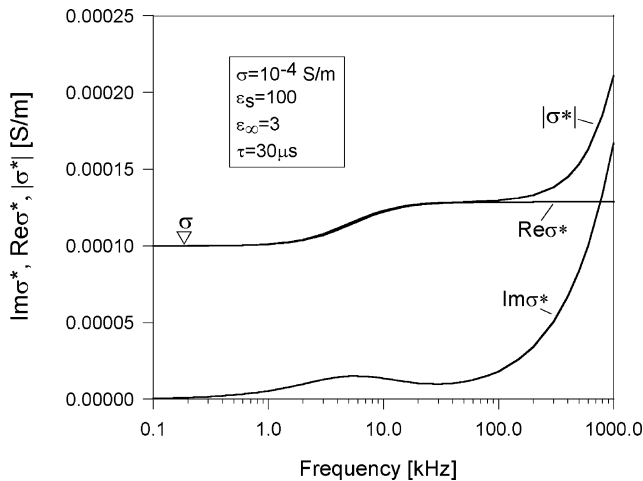
where  $T$  is the absolute temperature, in °K. At water freezing point,  $\tau$  is about 20  $\mu$ s, and at  $T = -60^\circ\text{C}$ , it is as large as 30 ms. Some examples of the practical application of this temperature dependence have been given by Reynolds (1997). A similar dependence of the dielectric-relaxation time constant  $\tau$  on temperature  $T$  has been found for frozen coarse-grained unconsolidated rocks. In rocks with high clay content, other factors in addition to temperature, may control the relaxation time constant. However, for a given rock, and all other factors being fixed, the relaxation time constant is always inversely proportional to the temperature.

In our studies, we decided to use sand as a model for simulating the TEM response of frozen ground. This choice was motivated by the following considerations: (i) a large amount of data on the dielectric spectroscopy of frozen sands is available in the literature; (ii) these data correlate well (unlike those on rocks with a high clay content); (iii) the amount of unfrozen water in sands is small, making it possible to avoid considering the strong polarization effects arising due to the migration of ions.

As a basis for modelling the TEM response of frozen unconsolidated ground, we used the laboratory data measured and reviewed by Frolov (1998). Most of our forward calculations were carried out with  $\varepsilon_s = 86$  and  $\varepsilon_\infty = 4$ . In most cases, calculations were made for  $\tau = 10 \mu$ s ('warm permafrost'), 30  $\mu$ s ('permafrost') and 100  $\mu$ s ('cold permafrost'). This allowed us to account for the dispersion in the experimental estimates of this parameter, and to obtain an idea of how its change (for example, in dependence on temperature) affects the TEM response of frozen ground.

The conductivity  $\sigma^*(\omega)$  of a rock with frequency-independent (ohmic) conductivity  $\sigma$  and complex dielectric permittivity described by the Debye formula (3) is given by

$$\sigma^*(\omega) = \sigma + j\omega\varepsilon_0 \left( \varepsilon_\infty + \frac{\varepsilon_s - \varepsilon_\infty}{1 + j\omega\tau} \right). \quad (5)$$



**Figure 3** Frequency dependence of the parameters,  $\text{Re } \sigma^*(\omega)$ ,  $\text{Im } \sigma^*(\omega)$ , and  $|\sigma^*(\omega)|$ , of frozen ground. The conductivity  $\sigma$  of the ground is independent of frequency, whereas its dielectric permittivity is frequency-dependent according to the Debye model of dielectric relaxation.

Separating  $\sigma^*$  into real  $\sigma'$  and imaginary  $\sigma''$  parts, results in

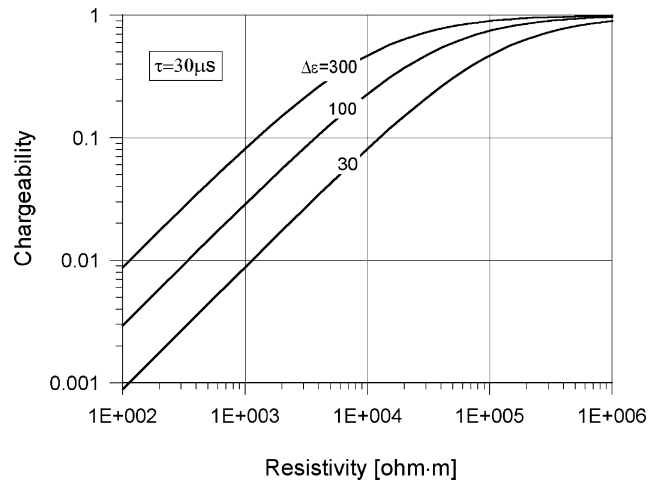
$$\text{Re } \sigma^* = \sigma' = \sigma + \frac{\omega^2 \epsilon_0 \tau (\epsilon_s - \epsilon_\infty)}{1 + \omega^2 \tau^2},$$

$$\text{Im } \sigma^* = \sigma'' = \omega \epsilon_0 \left( \epsilon_\infty + \frac{\epsilon_s - \epsilon_\infty}{1 + \omega^2 \tau^2} \right).$$

The frequency dependence of  $\sigma'$ ,  $\sigma''$  and the modulus  $|\sigma^*| = [(\sigma')^2 + (\sigma'')^2]^{1/2}$  is illustrated in Fig. 3. In these calculations, the following model parameters, typical of wet frozen sand at temperatures just below freezing, were used:  $\sigma = 10^{-4}$  Sm/m,  $\epsilon_s = 100$ ,  $\epsilon_\infty = 3$ ,  $\tau = 30 \mu\text{s}$ . The time constant of  $30 \mu\text{s}$  corresponds to the frequency of the Debye relaxation of about 5 kHz. At this frequency, the graph of  $\sigma'$  versus  $\omega$  exhibits an inflection, and  $\sigma''$  reaches its local maximum. At frequencies below 100 kHz, we have  $\sigma''(\omega) \ll \sigma'(\omega)$ , so we can assume  $\sigma^* = \sigma'$ , and (5) may be rewritten as

$$\sigma^*(\omega) = \sigma + \frac{\omega^2 \epsilon_0 \tau (\epsilon_s - \epsilon_\infty)}{1 + \omega^2 \tau^2}. \tag{6}$$

Figure 3 shows that at frequencies lower than 100 kHz,  $|\sigma^*(\omega)|$  and  $\sigma'(\omega)$  are practically coincident. Since the upper frequency limit in the signal spectra measured in inductive prospecting methods does not normally exceed a few tens of kHz, the frequencies above 100 kHz can be considered as practically infinitely high for determining  $\sigma_\infty$ .



**Figure 4** Chargeability versus resistivity plots for frozen ground with dielectric permittivity described by the Debye dielectric relaxation model.

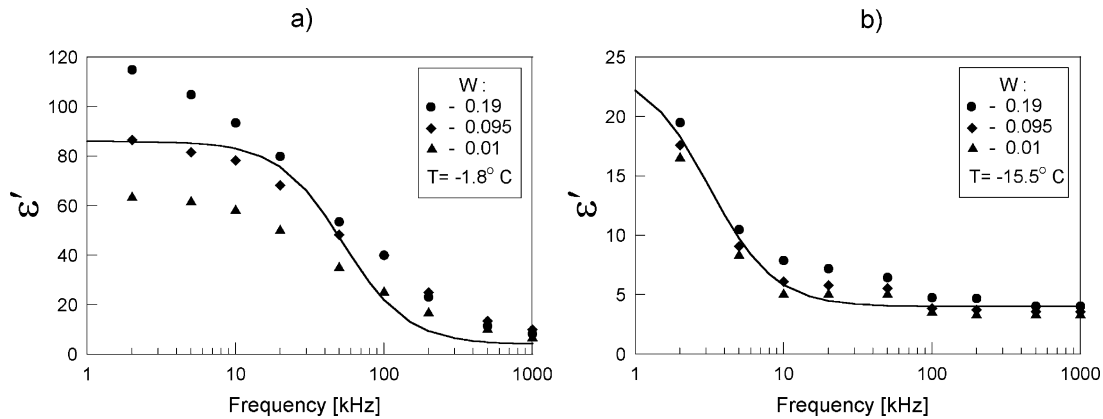
From (6), we obtain  $\sigma_0 = \sigma$ ,  $\sigma_\infty = \sigma + \epsilon_0(\epsilon_s - \epsilon_\infty)/\tau$ . Substituting the above expressions for  $\sigma_0$  and  $\sigma_\infty$  into (2) gives the formula for the chargeability of geological materials with frequency-independent conductivity  $\sigma$  and dielectric permittivity described by the Debye relaxation model:

$$m = \left[ 1 + \frac{\sigma \tau}{\epsilon_0(\epsilon_s - \epsilon_\infty)} \right]^{-1}. \tag{7}$$

According to (7),  $m$  is a function of the dimensionless parameter  $\beta = \sigma \tau / (\epsilon_0 \Delta \epsilon)$ , combining the ohmic conductivity  $\sigma$ , time constant of dielectric relaxation  $\tau$ , and the difference between low- and high-frequency dielectric permittivities  $\Delta \epsilon = \epsilon_s - \epsilon_\infty$ :  $m = 1/(1 + \beta)$ .

Figure 4 illustrates the dependence of  $m$  on the ohmic resistivity  $\rho = 1/\sigma$ . The graphs of  $m$  versus  $\rho$  suggest that there is a range of  $\rho$  (for the model under consideration it varies from very low resistivities up to  $10^4 \Omega\text{m}$ ) where the chargeability  $m$  is directly proportional to  $\rho$  and  $\Delta \epsilon$ . Hence the earth with a constant dielectric permittivity may exhibit effective chargeability that will change according to spatial variations in  $\rho$ . This means that during a combined resistivity and IP survey in a high-resistivity frozen environment, the apparent-resistivity and chargeability anomalies will correlate spatially.

Figure 4 also illustrates that for values of  $\Delta \epsilon$  and  $\tau$ , typical for frozen sands and other coarse-grained unconsolidated rocks, the Debye polarization can in practice only be measured if the rocks are resistive. For  $\Delta \epsilon = 100$ ,  $m$  is only about 2% if  $\rho = 10^3 \Omega\text{m}$ , and it does not reach 10% until  $\rho$  becomes as high as  $(3-5) \times 10^3 \Omega\text{m}$ . If  $\rho$  does not exceed a few



**Figure 5** Circles, diamonds and triangles represent the frequency of the dielectric response of frozen sand for different gravimetric ice contents. Solid lines show the dielectric response calculated using the Debye formula. A comparison between measured and calculated data suggests that, as a first approximation, the Debye formula is adequate for modelling dielectric relaxation in frozen ground: (a)  $\varepsilon_s = 86$ ,  $\varepsilon_\infty = 4$ ,  $\tau = 3 \mu\text{s}$ ; (b)  $\varepsilon_s = 24$ ,  $\varepsilon_\infty = 4$ ,  $\tau = 50 \mu\text{s}$ .

hundred  $\Omega\text{m}$ , the chargeability  $m$  is too small to be measured accurately and dielectric relaxation effects become ‘invisible’ for the IP method.

As shown above, in the time and frequency ranges in which induction methods usually operate, using the Debye model with parameters  $\sigma$ ,  $\varepsilon_s$ ,  $\varepsilon_\infty$ ,  $\tau$  is equivalent to using the Cole-Cole formula (1) with  $\sigma_0 = \sigma$ ,  $c = 1$ ,  $\tau$ , and  $m = 1/(1 + \beta)$ . In the general case, the assumption  $c = 1$  is somewhat too restrictive. However, in the case of frozen rocks this choice is justified (see discussion below).

Figure 5 shows the real part  $\varepsilon'$  of the measured dielectric permittivity of frozen sand as a function of frequency Frolov (1998). Measurements were made at temperatures of  $-1.8^\circ\text{C}$  (Fig. 5a) and  $-15.5^\circ\text{C}$  (Fig. 5b). The total weight water content  $W$  in the sand samples was 1, 9.5 and 19.5%. Along with the measured data shown by dots, Fig. 5 shows values of  $\varepsilon'$  calculated using the Debye formula (3). It is noticeable that both measured and modelled data exhibit similar behaviour of  $\varepsilon'$  as a function of frequency, showing a monotonous decrease from  $\varepsilon_s$  to  $\varepsilon_\infty$ . By lowering the temperature a better fit between the measured and calculated data is obtained. Figure 5 also indicates that the model parameters ( $\varepsilon_s = 86$ ,  $\varepsilon_\infty = 4$ ,  $\tau = 10\text{--}10^2 \mu\text{s}$ ), which were accepted as the basic parameters in our investigations, have proved to be reasonable in modelling dielectric relaxation in frozen unconsolidated rocks.

The validity of applying the Debye formula to simulating dielectric relaxation in frozen rocks is also based on a very specific aspect of the TEM response in frozen ground. In most cases, TEM signals are ‘distorted’ by a sharp local decrease, which may result in double sign reversals. Kormiltsev, Levchenko and Mesentsev (1990) investigated how the ex-

ponent  $c$  in the Cole-Cole equation (1) affects TEM signals measured over a homogeneous half-space using central-loop or coincident-loop configurations. It has been found that  $c > 0.5$  results in transient signals with a local minimum, including a double sign reversal. In the case of  $c < 0.5$ , only one sign reversal occurs. Both modelling and experimental TEM results obtained by Krylov and Bobrov (1995) and Krylov *et al.* (1995, 1998) show that the exponent  $c$  approaches 1 in a frozen environment. Similar results based on the TEM measurements were obtained by Walker and Kawasaki (1988).

## CHARACTERIZATION OF GEOELECTRICAL MODELS

The aim of the modelling experiment was to get an initial indication of the response of an ungrounded horizontal loop to frozen unconsolidated rock. We therefore focused on simulating the transient response of the following simple models only: (i) a homogeneous frozen half-space (ii) a two-layered earth with the upper layer frozen, and (iii) a two-layered earth with the upper layer unfrozen.

The first model, represented by a homogeneous low-conductivity half-space with dielectric permittivity described by the Debye formula (3), is a basic one. Using this model we were able to study the influence of the half-space parameters ( $\varepsilon_s$ ,  $\varepsilon_\infty$ ,  $\tau$ ,  $\rho$ ) and the loop size on TEM signals. This model simulates the permafrost in northern regions, as well as any volume of frozen ground with dimensions substantially larger than those of the loop.

The two other models investigated were represented by a two-layered earth with either the upper layer or the basement

being frozen. With a thick upper frozen layer, the first model approximates the permafrost in northern regions. If the frozen upper layer is thin, then the model represents a seasonally frozen layer underlain by unfrozen rocks. The second model may represent degrading permafrost or a seasonally thawed layer underlain by frozen rocks. For both two-layered models, it was assumed that only frozen rocks were polarizable.

## FORWARD-MODELLING DATA, RESULTS AND DISCUSSION

The transient response of frozen ground was calculated using the SIROTEST computer program, made available to us by its author, Prof P. Weidelt of the Technical University of Braunschweig, Germany. This program calculates the frequency response and converts it, through an inverse Fourier transform, to a transient response. The program calculates the TEM response of an ungrounded loop placed at height  $H$  above a horizontally layered ground. For our calculation,  $H = 0$ . The resistivity of the layers does not exhibit any frequency dispersion, whereas their dielectric permittivity is described by the Debye formula (3).

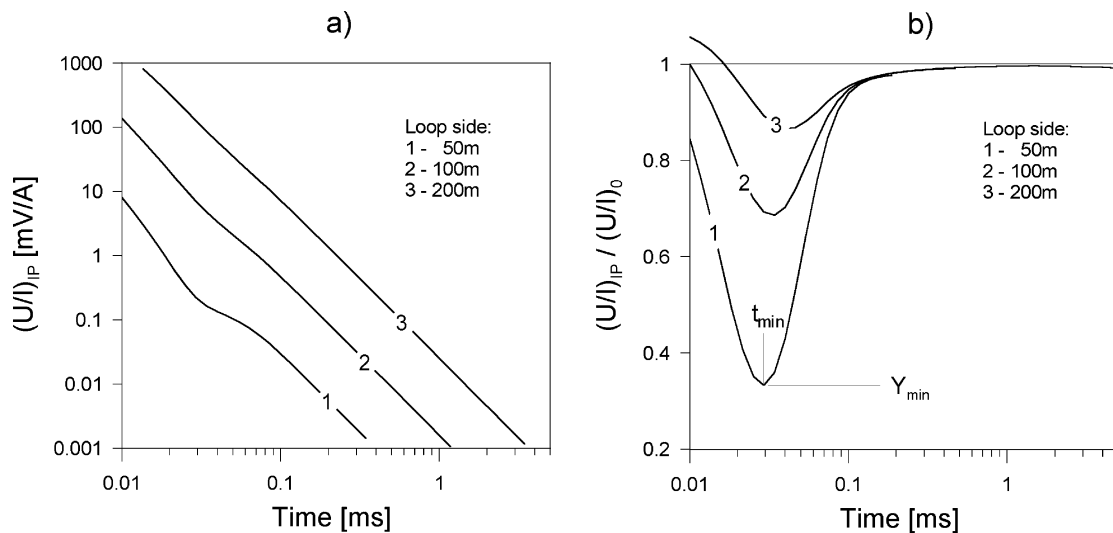
When representing and interpreting the model data, we encountered a problem with correlating heterogeneous data. Voltage and time ranges characteristic of transient signals, even if they are not influenced by IP, may vary over many orders of magnitude depending on the earth's resistivity and the loop size. The voltage TEM responses are usually trans-

formed to apparent-resistivity curves, which reflect predominately the electrical structure of the earth rather than the geometry of the TEM array. However, converting the voltage response to apparent resistivities in order to represent TEM data for a polarizable earth is senseless, because the use of apparent resistivities is based on an earth model with non-dispersive electrical properties; suffice it to say that in the case of polarizable earth, the coincident-loop TEM voltage may exhibit polarity changes which cannot be interpreted in terms of a non-polarizable earth model (Weidelt 1983).

The problem was solved by using special normalization of the model data (Nikiforov, Kozhevnikov and Artyomenko 1999; Kozhevnikov and Artyomenko 2004). We tested various methods of normalization and decided on one, which we call 'normalizing to a non-polarizable model'. In this method, the TEM response of a frozen polarizable earth  $(U/I)_{IP}$ , was normalized to that of a non-polarizable earth  $(U/I)_0$ , having the same structure and resistivities but with zero chargeability. So, instead of the transient response of a polarizable model, a normalized response  $Y(t)$  was used, where

$$Y(t) = \frac{[U(t)/I]_{IP}}{[U(t)/I]_0} \quad (8)$$

Figure 6(a) represents transient voltage-decay curves calculated for coincident square loops with side lengths of 50 m, 100 m and 200 m. The loops are laid down on the homogeneous half-space with  $\rho = 10^3 \Omega\text{m}$ ,  $\varepsilon_s = 86$ ,  $\varepsilon_\infty = 4$ ,  $\tau = 10 \mu\text{s}$ .



**Figure 6** The transient response of a polarizable half-space with  $\rho = 10^3 \Omega\text{m}$ , and  $\tau = 10 \mu\text{s}$ ,  $\varepsilon_s = 86$ ,  $\varepsilon_\infty = 4$ : (a) represented as graphs of  $U/I$  versus time, and (b) normalized to the transient response of a non-polarizable half-space. It is difficult, using conventional presentation, to estimate the IP effect on the TEM response. On the normalized TEM response graphs, even weak IP effects become clearly defined.



Comparison of the voltage decay curves (Fig. 6a) with the normalized transient responses (Fig. 6b) reveals that the latter show the effect of dielectric relaxation much more clearly than the former. It is also significant that normalization has made the TEM modelling data much more suitable for use in comparison studies.

Instead of inverting the field TEM data, we decided to investigate some characteristic features of the normalized transient response, namely the abscissa  $t_{\min}$  and the ordinate  $Y_{\min}$  of its minimum. These parameters turned out to be particularly useful in the overall estimation of the influence of dielectric relaxation of frozen ground on the TEM response.

### Homogeneous half-space

In most cases, the transient response of coincident square-loop arrays was calculated for typical loop sizes of 50 m by 50 m, 100 m by 100 m and 200 m by 200 m, commonly used in TEM surveys in Yakutiya and other northern regions of Russia (Vanchugov and Kozhevnikov 1998). It was assumed in all calculations that  $\varepsilon_s = 86$ ,  $\varepsilon_\infty = 4$ . In order to study the influence of temperature changes on the TEM signals, the value of  $\tau$  was varied in the range from 10 to 100  $\mu\text{s}$ . The earth resistivity  $\rho$  was varied from  $10^2$  to  $10^4$   $\Omega\text{m}$ .

As an example, Fig. 7(a) represents the calculated  $Y_{\min}$  versus  $\rho$  data for the 100 m by 100 m loop. Examination of the calculated data shows that the contribution of dielectric relaxation to the total TEM signal is most pronounced for small

loops, short  $\tau$  and resistive ground. It has been found for the above coincident-loop configuration that the  $Y_{\min}$  versus  $\rho$  dependence may be represented as

$$Y_{\min} = a_0 + a_1\rho + a_2\rho^2, \quad (9)$$

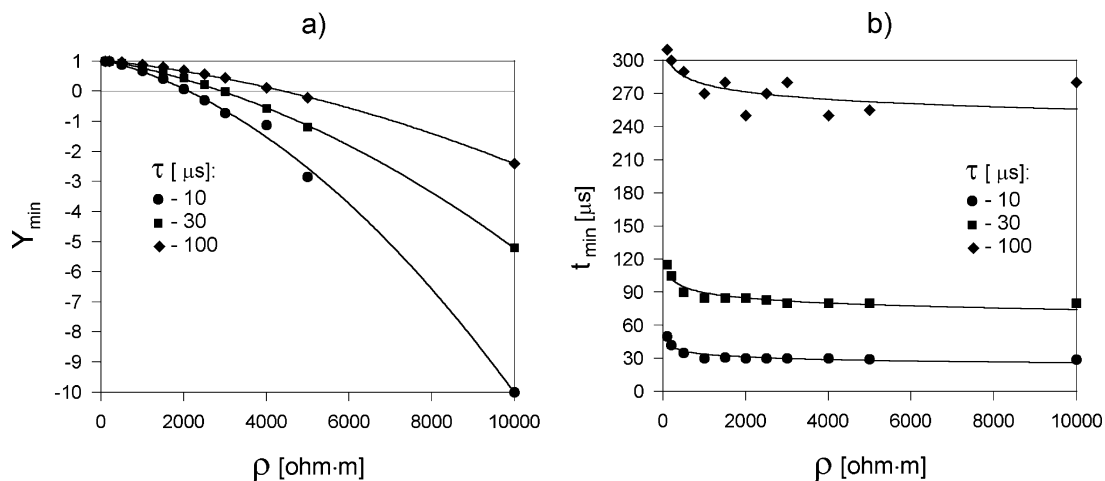
where the coefficients  $a_0$ ,  $a_1$  and  $a_2$  are found by the least-squares method. Examination of (8) shows that dielectric relaxation in a conductive earth has no appreciable effect on the total transient response. As  $\rho$  increases, the contribution of the dielectric relaxation increases, first proportionally to  $\rho$  and then as the square of  $\rho$ . Thus, increasing  $\rho$  results in a progressively stronger effect of dielectric relaxation on the TEM signal.

Significant results have been obtained by studying the relationship between  $t_{\min}$  and  $\rho$  (Fig. 7b). It has been found for loop sizes, 100 m by 100 m and 200 m by 200 m, that  $t_{\min}$  is slightly dependent on the half-space resistivity if it does not exceed approximately 500  $\Omega\text{m}$ . For higher resistivities,  $t_{\min}$  is practically independent of  $\rho$ . Within the limits of 10% error,  $t_{\min}$  is equal to three times the relaxation time constant, i.e.

$$t_{\min} = 3\tau. \quad (10)$$

The relationship (10) may have some potential for *in-situ* estimation of the temperature of frozen rocks, which is known to affect significantly the time constant of dielectric relaxation in frozen unconsolidated rocks.

Figure 8(a) shows the normalized coincident-loop transient response of a frozen half-space with resistivity  $\rho = 2 \times 10^3$   $\Omega\text{m}$ .



**Figure 7** Plots showing the coordinates ( $Y_{\min}$ ,  $t_{\min}$ ) of a minimum of the normalized TEM response versus resistivity  $\rho$  of a frozen half-space (100 m by 100 m coincident-loop configuration,  $\varepsilon_s = 86$ ,  $\varepsilon_\infty = 4$ ). Data on the upper ( $t_{\min} - \rho$ ) plot are scattered because for  $\tau = 100$   $\mu\text{s}$  the minimum on the normalized TEM response is smooth and its abscissa cannot be found with the same accuracy as in the case of smaller  $\tau$  (see Figure 9a).

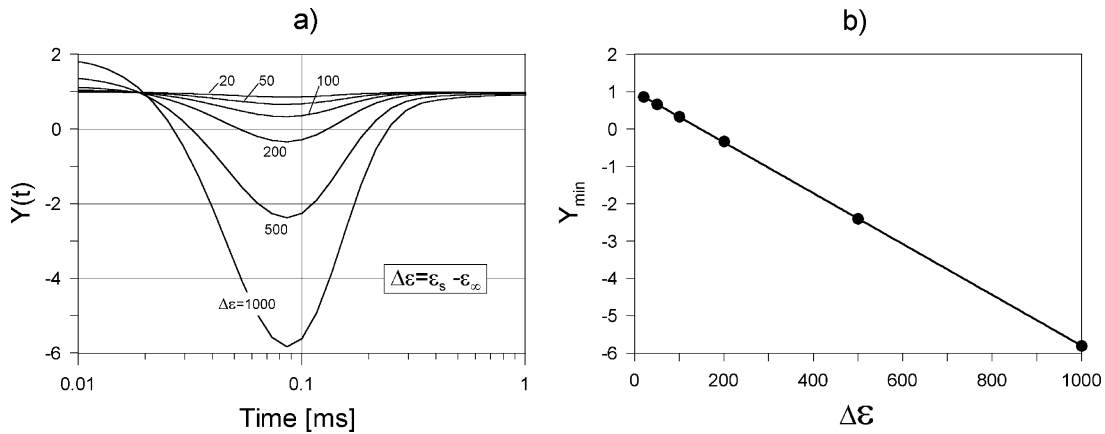


Figure 8 Normalized transient responses of a 100 m by 100 m coincident-loop configuration to frozen ground with  $\rho = 2 \times 10^3 \Omega\text{m}$ ,  $\tau = 30 \mu\text{s}$ ,  $\Delta\epsilon$  in the range 20–1000 (a), and  $Y_{\min}$  versus  $\Delta\epsilon$ (b).

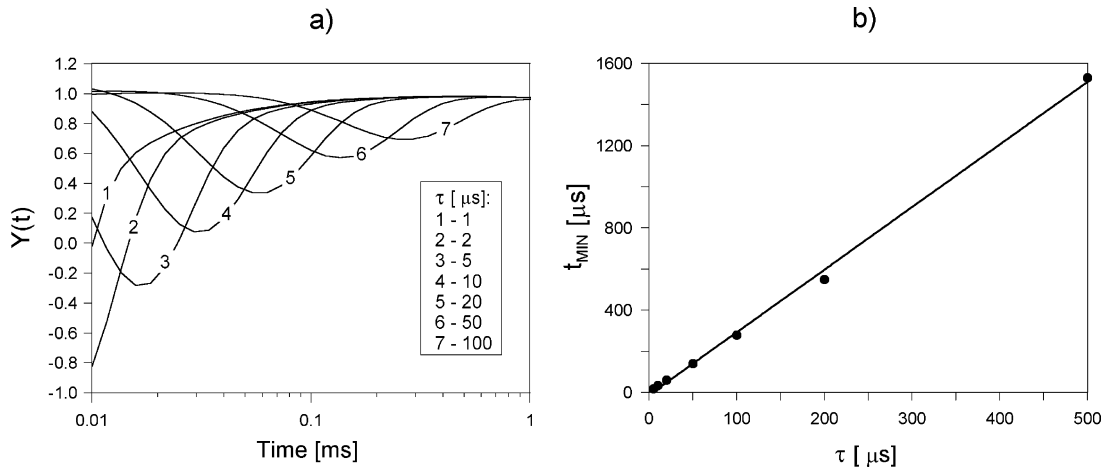


Figure 9 The normalized TEM response as a function of the dielectric-relaxation time constant: (a) normalized transient response of 100 m by 100 m coincident-loop configuration to the frozen half-space with  $\rho = 2 \times 10^3 \Omega\text{m}$ ,  $\epsilon_s = 86$ ,  $\epsilon_\infty = 4$ ; (b) graph of  $t_{\min}$  versus  $\tau$ , indicating that the abscissa of the minimum is directly proportional to the relaxation time constant.

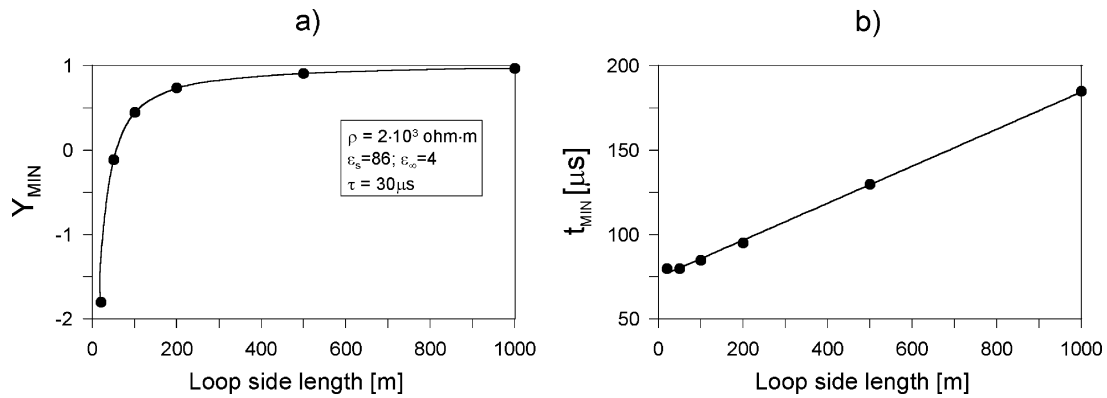
The TEM response was calculated for  $\tau = 30 \mu\text{s}$  and  $\Delta\epsilon = \epsilon_s - \epsilon_\infty$  with  $\epsilon_\infty = 4$  and  $\epsilon_s$  varying between 24 and 1004. It can be seen that, independently of  $\Delta\epsilon$ ,  $t_{\min} = 85\text{--}90 \mu\text{s}$ . Thus, the relationship (10) holds for a wide range of  $\Delta\epsilon$ .

Figure 8(b) illustrates how the amplitude of the minimum  $Y_{\min}$  is affected by  $\Delta\epsilon$ . The graph of  $Y_{\min}$  versus  $\Delta\epsilon$  is a straight line. Thus  $Y_{\min}$  can be used to estimate  $\Delta\epsilon$ . Since normally  $\epsilon_\infty \ll \epsilon_s$ , we can assume that  $\Delta\epsilon \approx \epsilon_s$ . This suggests, on the basis of (4), that TEM data can be used for *in-situ* estimation of the ice content in frozen sand and other coarse-grained unconsolidated rocks.

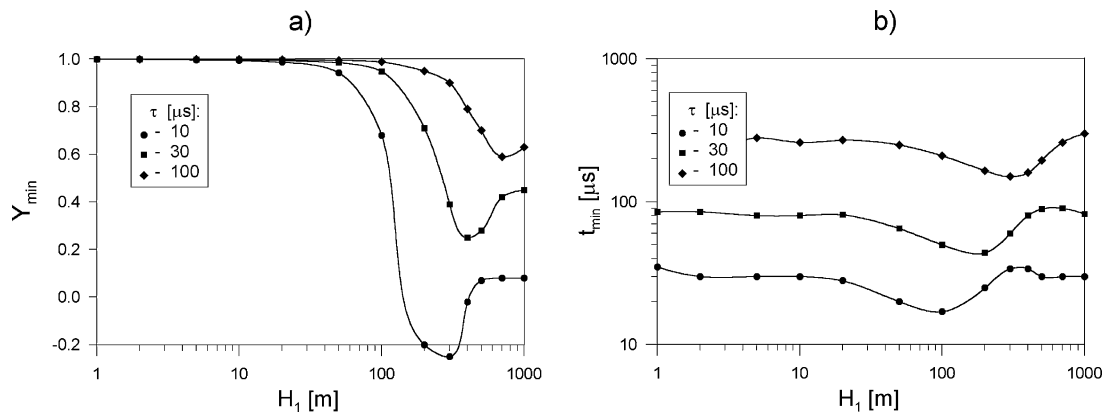
Figure 9 illustrates the effect of the dielectric-relaxation time constant on the TEM response. Figure 9(a) shows the normalized TEM response of a frozen half-space (100 m by 100 m

coincident loop configuration,  $\rho = 2 \times 10^3 \Omega\text{m}$ ,  $\epsilon_s = 86$ ,  $\epsilon_\infty = 4$ ) for  $\tau$  varying between 1 and  $10^2 \mu\text{s}$ . As  $\tau$  increases,  $Y_{\min}$  decreases as  $1/\sqrt{\tau}$ . Figure 9(b) shows that the graph of  $t_{\min}$  versus  $\tau$  is almost linear within the entire range of  $\tau$  considered:  $t_{\min} = 3\tau - 11.7$ , where both  $t_{\min}$  and  $\tau$  are in  $\mu\text{s}$ .

Figure 10 shows the influence of the side  $l$  of the coincident loops on the TEM response to the frozen half-space ( $\rho = 2 \times 10^3 \Omega\text{m}$ ,  $\epsilon_s = 86$ ,  $\epsilon_\infty = 4$ ,  $\tau = 30 \mu\text{s}$ ). The length of the loop side varied from 20 to 1000 m. As it can be seen, the modulus of  $Y_{\min}$  decreases with increasing  $l$ . Additionally, increasing  $l$  results in shifting  $t_{\min}$  towards values larger than  $3\tau$ , but for the loop dimensions that are used in TEM surveys in northern regions of Russia ( $20 \text{ m} \leq l \leq 200 \text{ m}$ ), the relationship (10) is fairly accurate ( $\pm 10\%$ ).



**Figure 10** Dependence of the coordinates of the minimum on the loop side length  $l$ : (a) as may be inferred from the graph of  $[(U/I)_{IP}/(U/I)_0]$  versus  $l$ , the dielectric relaxation effect on the transient response of the coincident-loop configuration is inversely proportional to its size; (b) the abscissa of the minimum increases directly with  $l$ , but for the loops more frequently used (i.e.  $l = 20\text{--}200$  m),  $t_{\min}$  is still equal to  $3\tau$ , within the limit of 10% error.



**Figure 11** Characteristic features of a 100 m by 100 m coincident-loop transient response of the two-layered ground with the upper layer frozen: (a)  $Y_{\min}$  versus  $H_1$ ; (b)  $t_{\min}$  versus  $H_1$ . The model parameters are:  $\rho_1 = 2 \times 10^3 \Omega\text{m}$ ,  $\varepsilon_{s1} = 86$ ,  $\varepsilon_{\infty 1} = 4$ ,  $\rho_2 = 10^2 \Omega\text{m}$ .

### Two-layered earth with the upper layer frozen

Figure 11(a) shows the ordinate  $Y_{\min}$  of the minimum of the normalized TEM response versus the frozen layer thickness  $H_1$ . The response was calculated for a 100 m by 100 m coincident-loop configuration on the two-layered earth with  $\rho_1 = 2 \times 10^3 \Omega\text{m}$ ,  $\varepsilon_{s1} = 86$ ,  $\varepsilon_{\infty 1} = 4$ ,  $\tau_1 = 10, 30, 100 \mu\text{s}$ , and  $\rho_2 = 10^2 \Omega\text{m}$ .

Figure 11(a) shows that dielectric relaxation in the frozen layer has no appreciable effect on the transient response until  $H_1$  exceeds some minimum value. This threshold thickness can be defined as the thickness that results in, for example, a 20% decrease in the TEM signal as compared to that of a non-polarizable earth. Depending on  $\tau$ , this threshold thickness varies from 80 to 300 m. If  $H_1$  does not exceed the threshold value, the transient response is essentially the same as that which would be measured in a non-polarizable two-layered

earth with  $\rho_1 = 2 \times 10^3 \Omega\text{m}$  and  $\rho_2 = 10^2 \Omega\text{m}$ . This suggests that seasonal freezing of the ground should not have an appreciable effect on the inductively induced IP response. With increasing  $H_1$ , the effect of the dielectric relaxation on the transient response becomes more pronounced, reaching a maximum in the range of  $H_1$  between 300 m and 700 m, and then it normally decreases. When  $H_1$  is large, the transient response is essentially the same as it would be measured on a polarizable half-space with a resistivity of 2000  $\Omega\text{m}$ .

Figure 11(b) shows  $t_{\min}$  versus  $H_1$ : these graphs exhibit smooth minima located at  $H_1 \approx 100$  m ( $\tau = 10 \mu\text{s}$ ),  $H_1 \approx 200$  m ( $\tau = 30 \mu\text{s}$ ) and  $H_1 \approx 300$  m ( $\tau = 100 \mu\text{s}$ ). The average  $t_{\min}/\tau$  ratios are 2.9, 2.5 and 2.5 for  $\tau = 10, 30$  and  $100 \mu\text{s}$ , respectively. These ratio values are similar to those found in the above case of a frozen half-space. For most of the  $t_{\min}$  versus  $H_1$  graphs, relationship (10) still holds, whereas in the minima  $t_{\min} \approx 1.5\tau$ .

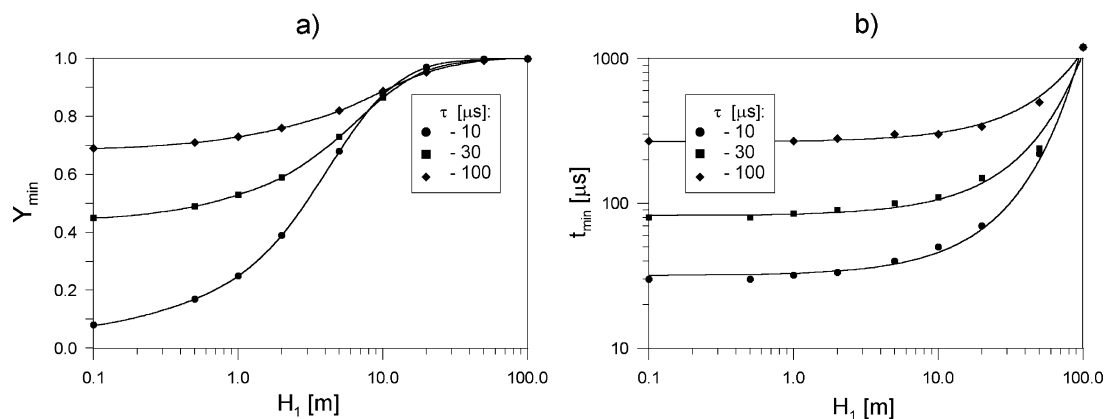


Figure 12 Characteristic features of a 100 m by 100 m coincident-loop transient response of the two-layered ground with the basement frozen: (a)  $Y_{\min}$  versus  $H_1$ ; (b)  $t_{\min}$  versus  $H_1$ . The model parameters are:  $\rho_1 = 10^2 \Omega\text{m}$ ,  $\varepsilon_{s2} = 86$ ,  $\varepsilon_{\infty 2} = 4$ ,  $\rho_2 = 2 \cdot 10^2 \Omega\text{m}$ .

### Two-layered earth with the upper layer thawed

Although the  $Y_{\min}$  versus  $H_1$  graphs shown in Fig. 12(a) could not be approximated by a simple equation, they are very simple in shape: as  $H_1$  increases, a smooth transition of  $Y_{\min}$  from the left-hand side asymptotic values to the common right-hand side asymptote takes place. Starting from  $H_1 \approx 10$  m, this model is equivalent to a two-layer non-polarizable earth. According to the approximate convolution approach (Smith *et al.* 1988), a conductive thawed layer will attenuate and smooth the vortex electrical field impulse in the frozen basement. This will result in attenuation of the polarization currents which are caused by the vortex electrical field impulse. Thus, due to seasonal thawing of the top layer, the underlying frozen basement may become practically undetectable by the TEM method.

Figure 12(b) shows the  $t_{\min}$  versus  $H_1$  graphs, which can be closely approximated by the exponential relationship,

$$t_{\min} = t_0 \exp(H_1/d), \quad (11)$$

where  $t_0$  is the  $t_{\min}$  value for a half-space, and  $d$  ranges from 27 m ( $\tau = 10 \mu\text{s}$ ) through 37 m ( $\tau = 30 \mu\text{s}$ ) to 69 m ( $\tau = 100 \mu\text{s}$ ). According to (11), increasing  $H_1$  results in an exponential increase of  $t_{\min}$ . However, if the thickness of the thawed layer is small ( $H_1 \leq 10$  m),  $t_{\min}$  does not depend on  $H_1$  being approximately equal to  $3\tau$ . It should be noted that for these values of  $H_1$ , the dielectric relaxation of the frozen basement becomes detectable by the TEM method.

### FIELD DATA INVERSION EXAMPLE

Figure 13 illustrates the inversion results for the TEM data measured along the 50–54–58 survey line on the Nakyn kim-

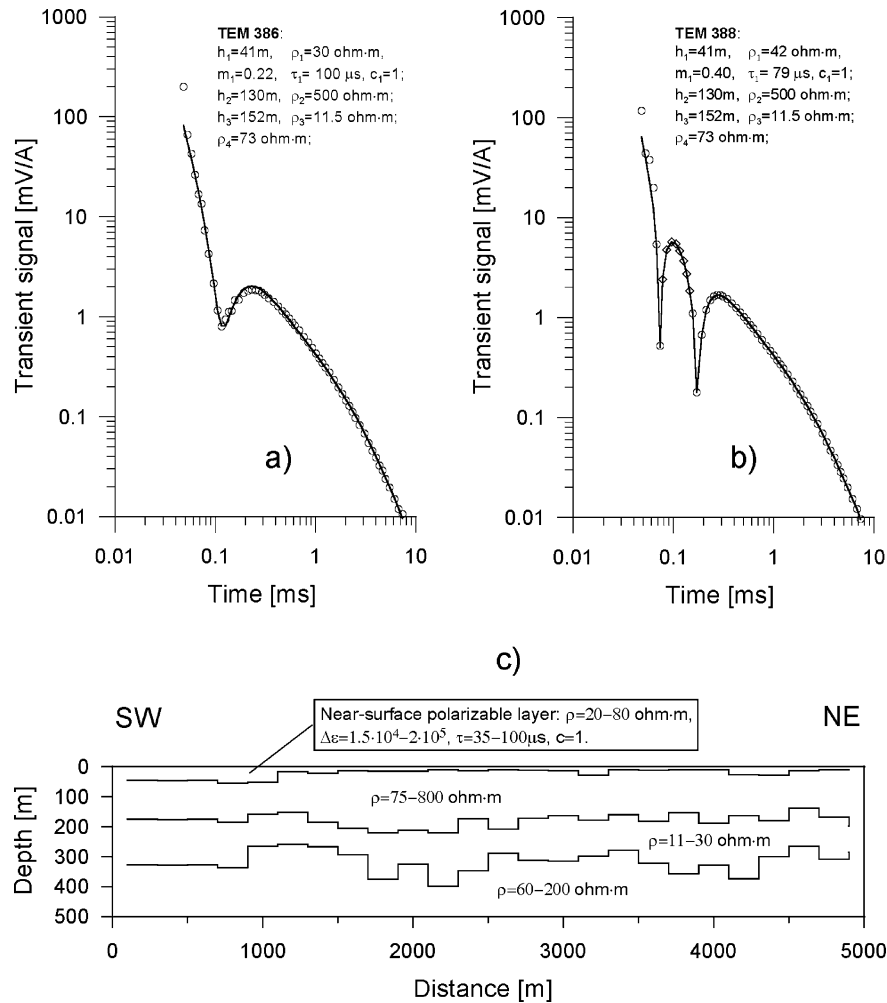
berlite field in Western Yakutiya. The site geology is represented by a thin layer of Quaternary unconsolidated sediments covering Jurassic sand and clay that are underlain by Lower Palaeozoic terrigenous-carbonate sedimentary rocks.

The TEM survey consisted of 25 central-loop soundings arranged on the 5 km long surveying line. 200 m by 200 m transmitter loops and 100 m by 100 m receiver loops were used in this survey. A commercial instrument, Tzicl-Micro, was used to produce and measure transients. The distance between the soundings was 200 m.

All TEM signals measured on this survey line were influenced by the fast-decaying IP (Figs 1a and 13a.b). Inversion of these data in terms of the Cole-Cole conductivity model (1) was carried out with a computer program developed at the Institute of Geophysics of SB RAS, Novosibirsk. The program computes the forward response using a digital filtering technique to carry out the Hankel and Fourier transforms. The inversion process uses the modified simplex method (Nelder and Mead 1965). Testing the inversion algorithm on synthetic data has shown its high effectiveness in recovering the complex multiparameter models (Yeltsov, Epov and Antonov 2002).

Open circles in the top of Fig. 13 indicate the measured TEM signals and the solid line shows the 1D inversion results, which were then assembled into a cross-section (Fig. 13c). Geoelectrically, the earth here is represented by a four-layer model with only the top layer polarizable. The thickness of this layer varies from 10 to 50 m, and the Cole-Cole parameters are:  $\rho = 20\text{--}80 \Omega\text{m}$ ,  $m = 0.2\text{--}0.46$ ,  $\tau = 35\text{--}100 \mu\text{s}$ ,  $c = 1$ . The fact that  $c = 1$  allowed (7) to be used to calculate  $\Delta\varepsilon$ , which was found to be in the range  $1.5 \times 10^4\text{--}2 \times 10^5$ .

Geologically, the top layer consists of frozen Quaternary unconsolidated sediments and underlying them is the upper part of the Jurassic sedimentary rocks. The frozen Quaternary



**Figure 13** Inversion of TEM data measured on the Nakyn kimberlite field in Western Yakutiya, Russia. (a) Measured and modelled TEM signals exhibiting distortion in the monotony without polarity change; (b) measured and modelled TEM signals exhibiting double polarity change; (c) resistivity and dielectric permittivity cross-section assembled from 1D inversion results. Open circles and diamonds in (a) and (b) indicate positive and negative measured data, respectively, and the solid line indicates the inversion data.

sediments have a thickness of up to 10 m and are known to have a very high (up to 55% and more) ice content (Klimovsky and Gotovtsev 1994). The high ice content and the electrical heterogeneity of these sediments may result in significant low-frequency dispersion of electrical properties, due to the dielectric relaxation in ice and the Maxwell–Wagner effect. The upper part of the Jurassic sedimentary rocks is also known to be fairly inhomogeneous. Due to the significant amount of clay, this part of the Jurassic rocks may exhibit a very high low-frequency dielectric constant (Olhoeft 1977; Bittelli *et al.* 2004), consisting of the  $\Delta\varepsilon$  values obtained from the inversion of the TEM data. The time constant  $\tau$  found from the inversion is typical of dielectric relaxation in frozen porous materials.

It has been found from the inversion data that the  $t_{\min}/\tau$  ratio varies in the range 1.1–3.9, averaging 2.25. This  $t_{\min}/\tau$  ratio is 25% less than that predicted by (10). It should be noted that the real geoelectrical structure of the Nakyn kimberlite field is more complicated than that of the models used in the forward calculations. Also, significant equivalence would be expected when inverting TEM data in terms of a multi-parameter polarizable model. Taking the above factors into consideration, the inverted results agree satisfactorily with the forward-calculations data.

The other geoelectrical layers were found to be non-polarizable. Geologically, these layers are represented (with the exception of the upper part of the second layer consisting of the lower part of a Jurassic sequence) by the

terrigenous-carbonate rocks of Lower Palaeozoic age. The low resistivities (11–30  $\Omega\text{m}$ ) of the third layer are due to it being saturated with saline water, which remains unfrozen, despite the fact that the permafrost base is situated here at a depth of about 400 m (Kozhevnikov and Vanchugov 1998).

## CONCLUSIONS

1 To simulate rapidly decaying IP processes in frozen ground, it was assumed that its conductivity is not dispersive whereas its dielectric permittivity is described by the Debye formula. This approach allowed the use of corresponding laboratory experimental results, obtained from dielectric spectroscopy of frozen rocks, as a basis for modelling the TEM response.

2 The proposed normalization of the TEM response due to a polarizable frozen earth by that of a non-polarizable ground provided an efficient method of presentation and comparative analysis of the calculated data. It was shown that the most informative parameters characterizing the normalized TEM response are the abscissa ( $t_{\min}$ ) and the ordinate ( $Y_{\min}$ ) of its minimum.

3 In the case of a frozen homogeneous half-space, the effect of the dielectric relaxation on the TEM response is directly proportional to the difference between static ( $\epsilon_s$ ) and dynamic ( $\epsilon_\infty$ ) dielectric permittivities, inversely proportional to the relaxation time constant ( $\tau$ ), directly proportional to the square of the earth's resistivity ( $\rho$ ), and inversely proportional to the loop side ( $l$ ). If the loop size does not exceed 200 m by 200 m,  $t_{\min}$  is equal to three times the relaxation time constant. Increasing the loop side above 200 m results in shifting  $t_{\min}$  towards later times.

4 The above features are also characteristic of the two-layered earth. Dielectric relaxation of a frozen layer underlain by a thawed half-space has an appreciable effect on the TEM response only when the frozen layer is more than approximately 100 m thick. Vice versa, even a thin ( $H_1 \approx 10$  m) thawed layer overlying a frozen half-space results in a significant reduction of dielectric relaxation effects.

5 Forward calculations and inversion of the field TEM data show that the dielectric relaxation of frozen unconsolidated rocks may significantly affect the transient response of a horizontal loop laid on the ground. Taking this phenomenon into account can improve the reliability and accuracy of the TEM measurements.

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