Monitoring of water seepage from a reservoir using resistivity and self polarization methods: case history of the Petergoph fountain water supply system

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Introduction

Petergoph, a suburb of Saint-Petersburg, was historically planned as a fountain centre similar to Versailles. The Petergoph fountain water supply system consists of artificial reservoirs formed by dams. Continued supply depends on the water level in the reservoir and the integrity of the dam. We used electrical methods for monitoring the stability of the reserve pond dam in this water supply system.

Ogilvy *et al*. (1969) and Bogoslowsky & Ogilvy (1970a, b) used geophysical methods to investigate water leakages from reservoirs but did not use continuous observations. Their results therefore give only snapshots of the spatial distribution of dam parameters and do not characterize their temporal variation. We use two complete series of geophysical observations to monitor changes in the hydrogeological regime over a spring month, at the time of annual flood and just after.

Geographical and geological setting

The dam studied separates the storage pond from a canal and river (Fig. 1). The subsurface consists of sand, gravel sand and sandy clay (Table 1). The dam is built of sand. The water table below the dam is situated at a depth of about 1 m.

There have been two incidents of water break-through producing the damage to the dam structure in the period of spring floods. Visual observation detected a spring outside the dam on the river bank (Fig. 1). We began our studies after the second incident. The observation grid was laid out on the isthmus between the pond and the river to investigate the dam state and the water pathways through the isthmus and their temporal variations.

Geophysical techniques

Resistivity and self polarization surveys were carried out twice, in April and May 1998, using similar techniques and observation grids. The resistivity sounding was conducted on a profile across the probable seepage area along the dam using a pole-dipole array (Fig. 1). The distance increment between current and potential electrodes (AO) was 2 m.

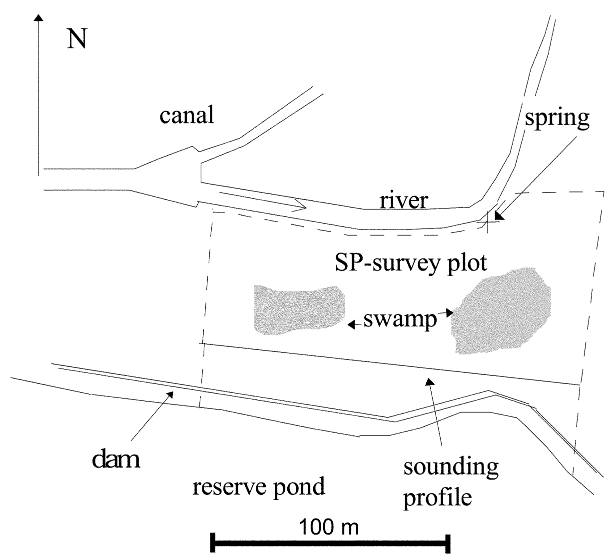

Figure 1 Sketch map of the dam area.

Table 1. Resistivity of sediments in the area of investigation.

Layers	Mean resistivity (electrical sounding data) Ohm-meter
Sand above water table	500
Sand and gravel sand	60
Sandy clay	145

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Both the dipole size and the distance between the sounding points were 2 m. The maximum distance between source and receiver was 20 m. Two-sided installation with two current electrodes installed left and right of the potential dipole was used. As shown by Khmelevskoi & Shevnin (1994), this technique allows reduction of perturbations to electrical sounding curves resulted caused by superficial heterogeneities. A high input impedance multifrequency receiver and 4.88 Hz generator (ERA-Limited Company production) were used for the survey.

A self polarization survey was carried out on the area between the pond and the river bank. We used Cl-Ag nonpolarized electrodes and the ERA receiver. The distances between the grid points were 4×8 m.

The data processing included: (i) the correction of SP values for electrode instability using the values obtained every 15 min at the reference station where one of the two electrodes was located, and (ii) the reduction of near-surface geological 'noise' in the sounding curves. This reduction of distortions (i.e. difference between the results of 1D data inversion and the real vertical distribution of resistivity for the sounding point) was done as shown in Fig. 2. The VES curves obtained during our field work contain peaks due to the passage of the current electrode across the superficial heterogeneities. Observations were made with distances between sounding points equal to the increment in the source-receiver distance. As a result, this type of distortion assumes a regular character that allowed us to remove the similar peaks on neighbouring curves.

The reduced curves obtained from each pole-dipole array were transformed to the Shlumberger type and then interpreted using a 1D program (Khmelevskoi & Shevnin 1994).

Results and discussion

Spatial distribution of parameters

Based on 1D interpretation, we obtained a geoelectrical cross-section containing three layers (Fig. 3). The first layer corresponds to nonsaturated sand, the second to relatively low resistivity saturated sand and gravel sand, and the third layer–to impermeable sandy clay. The close correspondence of values of mean resistivity obtained for the second layer (60 Ohm-m) and for the pond water (22 Ohm-m) suggests that sand and gravel sand have high porosity and a high permeability. Small lateral variations in resistivity of the second layer resulting from porosity variations cannot be determined because of the relatively small thickness of this layer. In our 1D interpretation, we assumed therefore a constant value for the resistivity (60 Ohm-m) along the whole profile. The interpretation of the VES profile allows us to detect four regions of higher thickness in the permeable layer (A-D) whose presence is potentially dangerous for dam stability (Fig. 3).

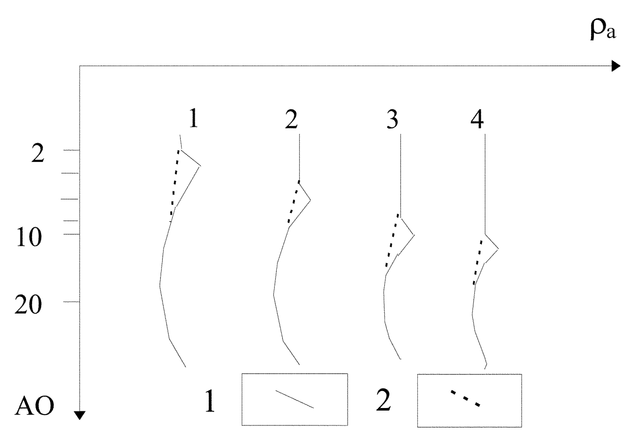

Figure 2 Schematic of sounding curve reduction. curves: 1–field, 2–reduced.

As can be seen on the self polarization map, a strong bipolar anomaly is present in only one of the four regions characterized by maximum thickness of the permeable layer (Fig. 4a). The negative part of anomaly is located near the pond, and the positive over the isthmus and near the river. We suggest therefore that this anomaly resulted from the great volume of water seepage from the pond to the river. The local character of the anomaly reflects the limited areas extent of water flow. We take a tube-like water flow as a simplified geometrical model for this seepage. The geometry of this model is similar to that used in the classical laboratory measurement of streaming potential properties (Ishido & Mizutani 1981; Sharma *et al*. 1987; Morgan *et al*. 1989). This similarity allows us to estimate the electrokinetic voltage cross-coupling coefficient given by the equation:

$$a = \Delta U/\Delta H \quad (1)$$

where ΔH is the water head drop obtained from measurements of the water levels in the pond and river, and ΔU is the corresponding electrical potential drop estimated as the difference between the average positive and negative values of the anomaly (Fig. 4a). The value of the cross-coupling coefficient obtained for the investigated anomaly is then about -2.5 mV/m (-2.5×10^{-7} V/Pa).

The cross-coupling coefficient represents a thermodynamic characteristic of a coupled process such as the generation of electrical potential by the water flow, and strongly depends on the water salinity. In contrast, the ζ-potential, the classical interface term, is less influenced by the salinity of the water and thus is more representative of the SP characteristics of the rocks. We therefore used the ζ-potential to estimate the SP properties of sediments in area studied. The calculated value of the cross-coupling coefficient and the resistivity of second layer, obtained from the VES interpretation, permits us to estimate the ζ-potential of the sands and gravel sand. The ζ-potential is given by the Helmholtz-Smoluchowski equation:

$$\zeta = \eta \Delta U/\rho \varepsilon \Delta H \quad (2)$$

where η, ρ, ε are the water viscosity, resistivity and dielectric constant, respectively. Using the resistivity value obtained

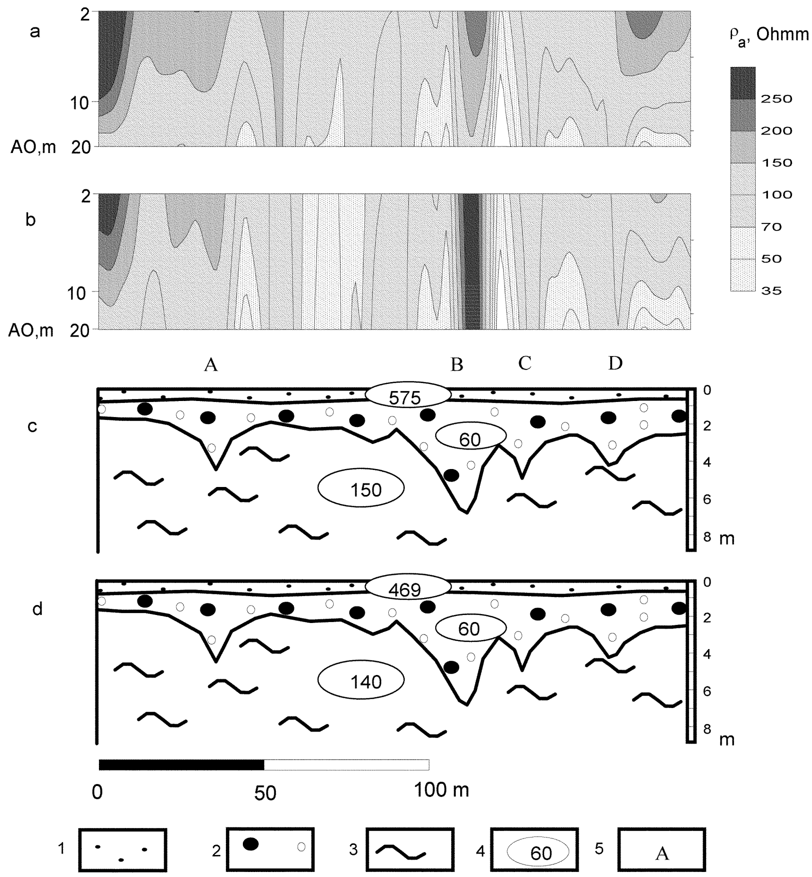

Figure 3 Resistivity surveys data. a, b–pseudo-sections, obtained in April and May, respectively. c, d–results of data inversion, obtained in April and May, respectively; 1–nonsaturated sand, 2–water saturated sand and gravel sand, 3–sandy clay, 4–resistivity values in Ohm-meters, 5–regions of high sand thickness.

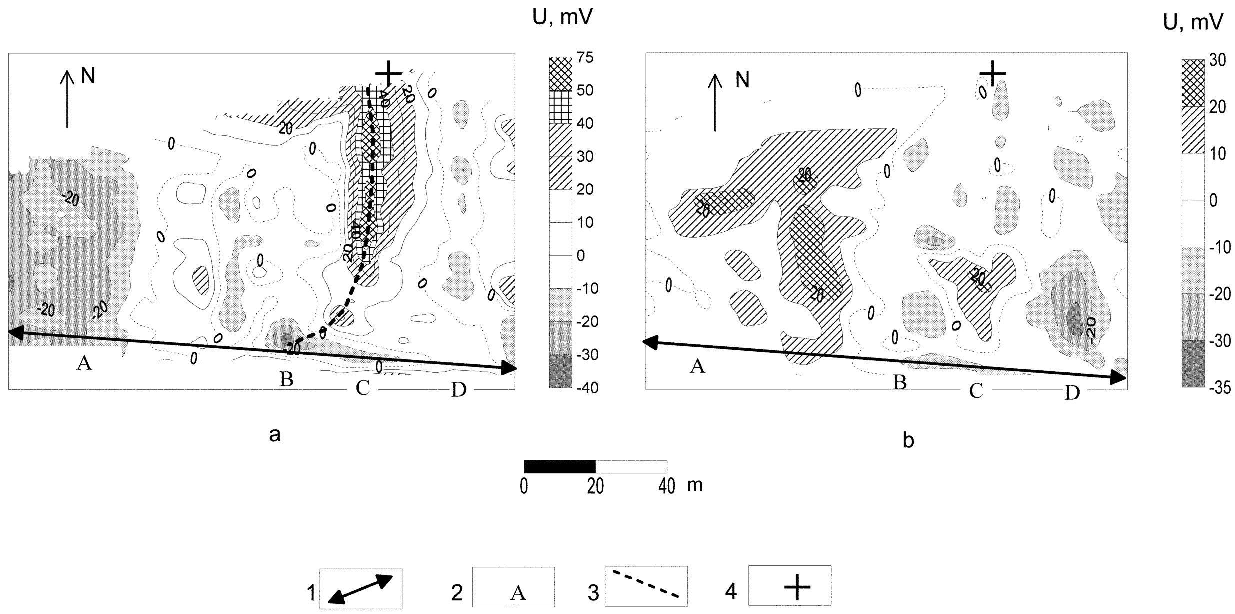

Figure 4 SP maps of two surveys: a–obtained in April, b–obtained in May 1–sounding profile, 2–regions of high sand thickness, 3–anomaly due to filtration, 4–spring.

from VES interpretation for the second layer, and the standard viscosity and dielectric constant values for the neutral water solution, we estimated the ζ-potential of sand and gravel sand to be −100 mV.

The value of the cross-coupling coefficient for sand and gravel sand is in good agreement with values obtained for sandstone and Westerly granites (Wurmstich & Morgan 1994; Morgan *et al*. 1989; Table 2). However, the ζ-potential of the sand and gravel sand may be slightly overestimated. This overestimation is probably caused by the use for a porous medium with different grain sizes of the Helmholtz–Smoluchowski equation deduced for the case of a cylindrical capillary. We thus demonstrate that estimation of streaming potential properties is possible for this simple case based on the resistivity and SP field data.

The general increase in SP trend from west to east corresponds to the river flow direction and can be explained by the general water flow feeding the river during the flood period.

Temporal variances of parameters

The second survey was carried out one month after the first survey during a period of low water level in the pond (Figs 3d, 4b). The average temperature was five degrees higher than the previous survey so that the temporarily frozen soils were completely thawed out. The results of the second electrical sounding were similar to those obtained during the first survey. We found a small difference in the average resistivity of the first layer which was probably due to the different water content of the soils. However, there was one strong local difference between the results of the first and the second surveys (stations 110–125, Fig. 3a,b). We interpreted this type of apparent resistivity change as a result of the electrodes not occupying precisely the same locations in a region of fast change in electrical potential drop. To avoid this false (measurement) difference during monitoring, fixed electrodes should be used.

The second SP survey demonstrates a dramatic change in the potential distribution (Fig. 4b). The anomaly observed during the first survey corresponding to the water seepage was found to have decreased in magnitude and area. These changes can be ascribed to the influence of the temporarily frozen soils which formed an impermeable superficial layer at the time of the first survey. These soils had thawed out by May, so that the near surface water flow was partially able to discharge to swampy areas near the river. This period corresponds to a strong decrease of flow in the spring on the river bank.

Conclusions

1 Resistivity sounding allowed us to detect the areas of maximal hydraulic conductance which corresponded to the greatest thickness of the permeable layer. However, the VES resolution was not sufficient to determine small lateral resistivity variations in the permeable (second) layer that would be caused by changes in sediment porosity and, as a consequence, its permeability. The areas of high hydraulic conductance are potentially dangerous for water seepage, and the SP anomaly reveals the actual water seepage route.

2 Repeating the survey on the same grid allowed us to observe the following changes: the resistivity spatial distribution is relatively stable vs. time, whereas the SP anomaly changes dramatically. This variation in SP allows us to monitor the temporal variations in the configuration of the groundwater pathway through the investigated area.

3 The supposed configuration of the water flow is similar to the geometry of classical laboratory measurements of streaming potential properties, which therefore allowed us to estimate the electrokinetic cross-coupling coefficient and ζ-potential on the basis of resistivity and SP field data.

4 Water pathways determined in April showed that the temporarily frozen soils represented an impermeable superficial screen so that the water flow completely discharged to the river. In contrast, the flow partially discharged to near-surface swampy areas when the soils were thawed out in May.

5 We consider that the high water level during the spring flood, while the adjacent soils are still frozen, is the most dangerous period for dam stability in the climatic conditions of Northern Europe.

Table 2. Cross-coupling coefficients and ζ-potential of rocks.

Rocks	ζ-potential (mV)	*a* (V/Pa)	Reference
Sandstone	–	-1.5×10^{-7}	Wurmstich & Morgan (1994)
Westerly granites	–70 – –80	-3×10^{-7}	Morgan *et al.* (1989)
Sand	–	$<-6\times10^{-7}$	Ahmad (1964)
Berea sandstone (pH = 7)	–58 – –30	–	Sharma *et al.* (1987)
Ottawa sandstone	–30	–	Sharma *et al.* (1987)
Quartz sand	–80	–	Rokitianski (1957)
Quartz (pH = 5.5)	–90	–	Ishido & Mizutani (1981)

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