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GEOELECTRICS IN COMPREHENSIVE GROUND WATER CONTAMINATION STUDIES

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ABSTRACT: A suite of geoelectric techniques were used to characterize an aquifer for a pollution study. Vertical electrical soundings were used to define aquifer layer resistivities and thicknesses. Porosities, hydraulic conductivities, and transmissivities were then estimated from the resistivities and thicknesses. The mise-a-la-masse method was used to define direction of flow, tracer velocity, and dispersion coefficients. Finally, borehole logging was used to obtain additional information on porosities and velocities.

KEYWORDS: ground water, geophysics, electrical resistivity, pollution

Geoelectrics can be applied to many specific hydrogeologic problems from aquifer characterization to pollution detection. In general, a single hydrogeological measurement technique is not sufficient, and supplementary techniques are needed. Geoelectric methods may themselves be supplementary methods.

Mazac and others [1,2] have presented the principles of geoelectrics as applied in aquifer characterization studies, and protection and pollution studies respectively. In this paper a comprehensive case history is presented, in which a suite of borehole and surface geoelectric methods were employed to characterize an aquifer as part of a comprehensive pollution study.

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BACKGROUND

Vertical electrical soundings typically are a primary geoelectrical measurement technique. With proper appraisal, sounding results can provide a complete picture of subsurface electrical conditions subject to the interpretation limitations. Normally, soundings should be calibrated at boreholes. Maps of layer resistivities, layer thicknesses and depths, or the Dar Zarrouk parameters [3], and geoelectrical cross sections can be prepared for determining or estimating geometric and hydrogeologic characteristics of an aquifer.

If information on ground water and relevant aquifer resistivity is available, the effective porosity can be calculated [4].

Aquifer resistivity can sometimes be related to aquifer hydraulic conductivity [1], and there are several conditions in which relations may be obtained. One case is where the percent of clay varies systematically in an aquifer. In this environment, the resistivity of the matrix can be expected to decrease with clay content and a relationship between permeability and matrix resistivity can be obtained. A complicating factor is that field measurements of matrix resistivity require the measurement of layer resistivities at at least two ground water resistivities. With the aquifer resistivity known, and if aquifer thickness can be determined, transmissivity can be calculated.

Flow direction can be obtained by monitoring the movement of an injected tracer using the mise-a-la-masse method [5,6]. Tracer experiments or movement of contaminants can be interpreted to estimate dispersivities. This can be done using monitoring wells and measuring specific conductances or, under favorable conditions, surface geoelectrics [7] or the mise-a-la-masse method can be substituted for the measurement of specific conductances.

The characteristics of aquitards can also be defined by electrical measurements [3]. The vertical permeability or leakance (ratio of vertical permeability to thickness) can be determined and can be useful in both protection and pollution studies.

SETTING

A geophysical survey was conducted at a locality in southern Moravia (Czechoslovakia) to investigate the causes of pollution in two drainage areas designated areas DA1 and DA2 on Fig. 1. The aquifer is used as a potable water source and more than 40 l/s of water is pumped from the aquifer for drinking water. The aquifer consists of Quaternary sand and gravel terrace deposits along the Satava River overlying an impermeable Neocene clay.

To characterize the pollution and define the probable pollution sources, water quality measurements were made along with surveys using several geoelectrical methods. The first objective of the geoelectric surveys was to define the geometry of the aquifer and its hydrogeological characteristics [8].

Sources of agricultural pollution:

- ① farm activity ② chemical manure
- ③ natural manure

DA drainage areas 1,2

P19 boreholes with local direction of ground-water flow (Tab.)

I structural units I to IV according to the aquifer thicknesses d

→ regional direction of groundwater flow

FIG. 1 -- Boundaries of hydrogeological units I, II, III, and IV based on electrically determined thicknesses.

From the vertical electrical sounding results, four areas were distinguished on the basis of the thicknesses and resistivities of the sand and gravel layers (Figs. 1,2). Thicknesses range from about 5 to more than 8 meters along the river (Area IV) to less than 1 meter in the northern part of the study area (Area I). The aquifer resistivity varies from less than 200 ohm-meters to more than 700 ohm-meters.

AQUIFER PROPERTIES

Porosity

The effective porosity $P_{eff}(\%)$ was calculated from the estimated resistivities using the relation (Worthington, 1976).

$$P_{eff} = \left[a \frac{\rho_{w,i}}{\rho_{o,i}} - a \frac{\rho_{w,i}}{\rho_s} \right]^{1/n} \quad (1)$$

where $\rho_{w,i}$ is the groundwater resistivity (contaminated $\rho_{w,c} = 3-10$ ohm-m, uncontaminated $\rho_{w,u} = 12$ ohm-m). The aquifer resistivity determined by VES is $\rho_{o,i}$ (contaminated $\rho_{o,c} = 170-550$ ohm-m, uncontaminated $\rho_{o,u} = 550$ ohm-m), and the constants $a=1$ and $n=2$ (Table 1) and ρ_s is the matrix resistivity. Matrix resistivity can be directly related to permeability where permeability is proportional to clay content; matrix resistivities can be calculated if measurements are made at the same point before and after contamination. The matrix resistivity is then

$$\rho_s = \frac{\rho_{o,u}\rho_{w,u}\rho_{o,c} - \rho_{o,u}\rho_{w,c}\rho_{o,c}}{\rho_{w,u}\rho_{o,c} - \rho_{o,u}\rho_{w,c}} \quad (2)$$

The effective $P_{eff}(\%)$ can also be calculated as

$$P_{eff} = 100 \frac{V_f}{V_r} \quad (3)$$

where the filtration (V_f) and the actual (V_r) velocities are determined by the dilution and mise-a-la-masse geophysical methods as shown later.

TABLE 1 -- Observed data $\rho_{o,c}$, $\rho_{o,u}$, $\rho_{w,u}$, $\rho_{w,c}$ and interpreted values of ρ_s and P_{eff} .

Borehole no.	$\rho_{o,c}$ (ohm-m)	$\rho_{w,c}$ (ohm-m)	$\rho_s \cdot 10^{-3}$ (ohm-m)	$P_{eff}(\%)$ $n = 2$	Remarks ρ (ohm-m)
P 1a	300	7	7.50	15.0	$\rho_{o,u} = 500$
P 1	350	8	3.50	14.3	...
P 6	200	4	2.00	13.4	$\rho_{w,u} = 12$
P 5	170	3	1.42	12.7	...

Flow direction and velocity

Both the direction and the magnitude of the groundwater flow velocity can be determined by a hydrogeological variant of the mise-a-la-masse method. The mise-a-la-masse method was originally used for detecting and determining the extent of conductive ore bodies.

The hydrogeological variant of the method [9,10] involves the continuous injection of a tracer (e.g., NaCl, KCl) which forms a conductive body in an aquifer which increases in size with time in the flow direction mainly due to convective motion, partly to dispersion. The body (plume) is charged by an electrical current and the distribution in time and space of equipotential lines are observed on the earth's surface (Fig. 3).

The actual or tracer flow velocity V_r in the mise-a-la-masse method is usually determined from the velocity of the movement $V_e = \frac{\Delta a}{\Delta t}$ (Fig. 4) of the equipotential lines as defined by the maximum slope of the $a_i(t_i)$ curve where a_i is the time displacement of the equipotentials in the direction of A of the maximum change. The main advantage of the mise-a-la-masse method is that only a single borehole is needed for determining the flow velocity and direction.

FIG. 3 -- Determination of the flow direction from the displacement of equipotential lines a_i (cm) by the mise-a-la-masse method.

FIG. 4 -- Displacement of equipotential lines a_i (cm) in the mise-a-la-masse method (derived from Fig. 3).

Due to the finite resistivity of the charged body (plume) the velocity V_e does not correspond to the true or tracer velocity V_r . The relation between the two velocities in the direction A is determined by Eq. (4):

$$V_r = C_n V_e \quad (4)$$

where the coefficient C_n ($1 \leq C_n \leq \infty$) depends on resistivity contrasts, depth to the water level and on other geometric factors [6]. If the resistivity ρ_o of an aquifer within the plume approaches zero ($\rho_o \rightarrow 0$) the constant $C_n = 1$ and $V_r = V_e$ which is a very rare case. At the surveyed locality $C_n = 1.3$ to 1.5 (Table 2).

TABLE 2 -- Observed or interpreted values of A, V_e , V_r , D_e , and D_{er} .

Borehole no.	A (degrees)	$V_e \cdot 10^{-5}$ (m/s)	$V_r \cdot 10^{-5}$ (m/s)	D_e (m^2/day)	D_{er} (m^2/day)
P 5	75	0.95	1.24	0.42	3.55
P 11	200	4.17	5.42	0.75	5.76
P 14	195	0.16	0.24	...	...
P 19	220	4.58	6.87	0.62	5.43

The depth penetration of the method is limited to about ten meters. The accuracy of the direction determination is higher than for the velocity determination. It is desirable to observe the movement of at least two different equipotentials at a distance from the borehole greater than 1.5 times the depth to the aquifer with a non-conductive casing and more than 3 times the depth in case of a conductive casing but in both cases less than seven times the depth to the aquifer. Equipotential movement should be observed for a continuous tracer input until their displacement $a_i(t_i)$ reaches a maximum (Figs. 3,4). Tracer concentrations should be in the range of 2 to 5 g/l, since higher concentrations can cause vertical segregation due to density contrast within the injected body and invalidate the

geometric and resistivity assumptions. The continuous concentration is maintained constant by monitoring the groundwater resistivity

Dispersion

The electrically determined dispersion coefficient $D_e = D_{er} P_{eff} 10^{-2}$ and the actual dispersion coefficient D_{er} are, together with the parameters V_r and P_{eff} , the most important transport parameters. Traditionally, determination of dispersion coefficients is based on monitoring variations of a tracer concentration $C(x, t)$, at a fixed point x , with time t . The point (e.g., an observation well) should be situated in the direction A of the groundwater flow which can be advantageously determined by the hydrogeologic variant of the mise-a-la-masse method described above (Fig. 3). Concentration variations at the point can be determined either in an observation well by chemical, electrical, photo- and thermometric methods or geophysically using two approaches: surface vertical electrical soundings at a point situated in the direction of groundwater flow or with the mise-a-la-masse method if only one borehole is available. The former method uses an impulse method and the latter a continuous input.

Relations for determining the dispersion coefficient D_e by the mise-a-la-masse method are based on the very rough assumption that the $C(t)$ curve corresponds to the $a_1(t)$ curve (Fig. 4).

This assumption is approximate, but is analogous to the assumption that the tracer concentration C can be determined from changes in aquifer resistivity ρ_0^+ interpreted from surface geoelectrical measurements [6]. Present interpretation techniques do not enable reliable determination of the distribution curve $\rho_0^+(x, t)$ and consequently, the concentration curve $C(x, t)$.

The relation for calculating D_e (Table 2) is taken from classical hydrogeological methods [7, 11, 12] using $a_1(t)$ for $C(t)$, with the notation as defined in Fig. 4. Here it is assumed that the dispersion is one-dimensional.

Filtration Velocity

The apparent $V_{a,f}$ and actual V_r filtration velocities were determined by the borehole resistivity dilution method (Fig. 5) based on monitoring groundwater resistivity with the depth to the water level.

The principle of the method was developed in the late fifties [13, 14]. Typically, salts (NaCl, KCl), organic dyes or radioactive elements are used as tracers. The apparent filtration velocity $V_{a,f}$ (m/s) can be calculated using equation (5):

$$V_{a,f} = - \frac{\pi r}{2t} \ln \left[\frac{C_t - C_0}{C_1 - C_0} \right] \quad (5)$$

where r is the well diameter in meters, t is the time since injection of the tracer in seconds, C_0 , C_1 , and C_t are concentrations in gl^{-1} of the tracer in the groundwater, immediately after introduction of the tracer and at time t .

FIG. 5 -- Determination of $V_{a,f}$ by the dilution method.

When resistivity is used to measure dilution of the tracer, NaCl is commonly used as the tracer. The NaCl concentration (C) can then be determined from the resistivity R (ohm-m) values using the simple relationship

$$C_t = \frac{\alpha_t}{R_t^n}; \quad C_0 = \frac{\alpha_0}{R_0^n}; \quad C_1 = \frac{\alpha_1}{R_1^n}$$

where $\alpha_{t,0,1}$ are constants depending on groundwater temperature, $R_{t,0,1}$ are measured groundwater resistivities and $n = 1.04 \pm 0.02$ is a constant determined experimentally and valid for $0.001 \text{ g/l} \leq C \leq 10 \text{ g/e}$ [15].

Substituting the expressions for C_t , C_0 , and C_1 and provided that $\alpha_t = \alpha_0 = \alpha_1$ the expression for the apparent filtration velocity can be rewritten in the following form:

$$V_{a,f} = 3.62 \left[\frac{x}{t} \right] \log \left[\frac{\left(\frac{R_0}{R_1} \right)^{1.04} - 1}{\left(\frac{R_0}{R_t} \right)^{1.04} - 1} \right] \quad (6)$$

The true filtration velocity, V_f (Table 3) is calculated from Eq. 7

$$V_f = \frac{V_{a,f}}{\beta} \quad (7)$$

where β is the drainage coefficient [15] depending on the well configuration ($0 < \beta < 8$). In the study locality $\beta = 3$.

TABLE 3 -- Calculation of $V_{a,f}$ and V_f .

Borehole no.	$V_{a,f} \cdot 10^{-5} \text{ (m/s)}$	$V_f \cdot 10^{-5} \text{ (m/s)}$
P 1	2.41	0.80
P 2	3.50	1.17
P 3	6.65	2.22
P 4	2.02	0.67

Hydraulic Conductivity

The hydraulic conductivity K_f (m/s) (Fig. 6) was calculated from the aquifer resistivity ρ_0 using the following relation based on 8 parametric measurements

$$K_f (10^{-5} \text{ m/s}) = 165150 \rho_0^{-1.78} (\text{ohm-m}) \quad (8)$$

The inverse relationship between K_f and ρ_0 usually occurs in lithologically homogenous sediments and indicates K_f is primarily determined by variations in aquifer porosity [1]. Once we know the thickness b (m) of the aquifer (Fig. 1) and its resistivity ρ_0 (Fig. 2), and hence K_f , the transmissivity T (m^2/s) = bK_f can be easily calculated.

FIG. 6 -- Dependence of $K_f = f(\rho_0)$ based on borehole parametric measurements.

Leakage Coefficient

The leakage coefficient K_n ($1/\text{s}$) = $K_f / m = f(1/t)$ where K_f is the vertical hydraulic conductivity of an overburden and m is its thickness. It expresses the potential of water to infiltrate through the layers above an aquifer. It is inversely proportional to the protective capacity of the overburden [3] which is considered to be directly proportional to travel time. For a clayey overburden, as in the study area, the protective capacity could be considered as being proportional to the longitudinal unit

conductance $S = \sum_{i=1}^n \left[\frac{m_i}{\rho_i} \right]$ where m_i (m) and ρ_i (ohm-m) are the thicknesses and resistivities of the overburden sediments (Table 4).

TABLE 4 -- Calculation of leakage parameters.

Borehole no.	m_1 (m)	ρ_1 (ohm-m)	m_2 (m)	ρ_2 (ohm-m)	$s_i \sim f(t)$ (1/ohm)	$1/s_i \sim f(K_n)$ (ohm)
P 1	0.8	58	0.6	350	0.01551	64.48
P 6	0.5	85	0.85	200	0.01013	98.69
P 8	0.4	120	2.9	254	0.01476	67.73
P 11	0.3	90	2.9	450	0.00977	102.27
P 19	1.0	292	2.4	480	0.00842	118.70

For ground water pollution, this parameter can indicate attenuation of some contaminants as they move through the overburden layers to the water table [3].

CONCLUSIONS

Geoelectric methods can be used in all phases of ground-water contamination investigations. Their information content can be maximized by using them in complementary ways. Supplementary borehole information is required to calibrate the electrical methods. Borehole logging can complement measurements from the surface "mise a la masse" method.

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