ESTIMATION OF HYDRAULIC CONDUCTIVITY IN A LARGE SCALE LABORATORY MODEL BY MEANS OF THE SELF-POTENTIAL METHOD

S. Straface[1,2], E. Rizzo[2], V. Gianpaolo[2] and F. Chidichimo[1]

[1] Dipartimento Difesa del Suolo; Università della Calabria, Rende (CS), Italy,
E-mail: straface@unical.it; chidichimo@dds.unical.it
[2] Hydrogeophysics Laboratory; CNR-IMAA, Marsico Nuovo (PZ), Italy,
E-mail: rizzo@imaa.cnr.it, valeria.giampaolo@imaa.cnr.it

ABSTRACT

The characterization of a heterogeneous aquifer involves a wide spectrum of high resolution direct measurements of the characteristic parameters, defined in itinere according to the various approaches used. This information is usually obtained from the pumping tests and the relative measurements of the drawdown induced. Unfortunately, these methods are intrusive and the hydrological system is perturbed by the presence of perforations. An aim of geophysics is to use non-intrusive techniques obtaining a great deal of information on subsoil at low cost.

In this article a novel methodology to estimate the hydraulic conductivity from self-potential measurements is proposed. The hydrogeophysical experiments were carried out in the CNR-IMAA (Marsico Nuovo - Italy) Hydrogeosite Laboratory, a large scale model sized 10 × 7 × 3 m^3, filled with a homogeneous medium made up of a quartz-rich sand with a medium-high hydraulic conductivity in the order of 10^{-5} m/s.

Self-potential signals generated by the groundwater flow, during a pumping test, were measured on the ground surface, both on pumping and recovery phases. Firstly, hydraulic conductivity was calculated by means of Neuman Type-Curves method applied on the pumping test. As a result, hydraulic conductivity very close to that calculated from hydraulic data, by means of Neuman Type-Curves, has been obtained by solving numerically the electrical flow equation.

Keywords: Characterization; Hydro-geophysics; Self-potential

1. INTRODUCTION

The main problem in characterizing a real heterogeneous aquifer is the requirement of a large number of direct high resolution measurements of parameters, defined according to the various used approaches. Usually measurements concern the drawdown induced by pumping tests. Unfortunately, this method is intrusive since the hydrological system is strongly perturbed by the perforations. Geophysical approaches aim to use non-intrusive techniques to obtain, at low cost, a large amount of information on subsurface. In the last few decades, hydrologists have increasingly begun to use geophysical information to estimate groundwater flow parameters (Straface et al. [1], [2]; Cassiani and Medina [3]; Cassiani et al. [4]; Troisi et al. [5]). The correlation between self-potential signals and groundwater flow, which generate a detectable electrical field in the subsurface has been shown, in the pats, by several authors (Poldini [6]; Sill [7]). At present, the fundamental physical principles, describing the link between the water flow in a porous medium and the generation of an electrical field, are well-understood, on the minimum volume scale representative of the porous medium, by means of the electrokinetic theory (Bernabé [8]; Lorne et al. [9]). On the field scale, the forward electromagnetic problem can be defined as a potential field problem (Fitterman [10]) and, in the last few years, various algorithms have been developed to determine the source distribution responsible for the self-potential anomalies recorded at the ground surface. The inverse problem suffers from the non-uniqueness of the solution. To remove this non-uniqueness, it is necessary to merge different kind of measurements coming from non intrusive techniques or borehole data. Recently, several researchers have been developing interpretation schemes that could be applied to the natural electrical field recorded during pumping tests (Revil et al. [11]; Rizzo et al. [12]; Titov et al. [13], [14]). This approach enjoys two advantages: the economical use of sensors and the nonintrusive nature of the self-potential measurements. This allows the use of a large number of sensors at benefits of the stochastic treatments for the characterization of heterogeneous porous media.

Until a few years ago the results of the self-potential method were used only for qualitative analysis, and mainly to supply an indication of the presence of anomalies that generated the phenomenon. Recently various authors (Patella et al. [15]; Revil et al. [11], [16], [17]; Darnet et al. [18]; Rizzo et al. [12]; Suski et al. [19]; Straface et al. [1], [2]) have shown the possibility of using this electrical signal for the quantitative estimation of the hydrogeological parameters of a porous medium. They have demonstrated that there is a correlation between the self-potential signals and the values of the hydraulic head (Revil et al. [11]; Linde et al. [20]). Recently, Malama et al. ([21], [22]) have developed a semi-analytical solution for the transient streaming potential response of an unconfined and of a confined aquifer to continuous constant rate pumping. These solutions, fitted to field measurements of streaming potentials associated with a transient aquifer test, indicate that where observation wells are unavailable to provide more direct estimates, streaming potential data collected at land surface may, in principle, be used to provide preliminary estimates of aquifer hydraulic conductivity and specific storage or specific yield. Moreover, new experiments have been carried out using the combination of hydraulic and geophysical measurements during pumping tests (Rizzo et al. [12]) and hydraulic tomography (Barrash et al. [23]; Jardani et al. [24]).

This paper presents a hydrogeophysical experiment carried out at the Hydrogeosite Laboratory at the CNR-IMAA site in Marsico Nuovo (PZ). The following paragraphs illustrate the first results of the characterization effected using the hydraulic and geophysical measurements recorded during the pumping tests in the Hydrogeosite

Laboratory. The results open new perspectives regarding the possibility of monitoring the groundwater flow and therefore, owing to the great increase in information, of improving the characterization of the real heterogeneous porous media.

2. THEORY OF SELF-POTENTIAL METHOD

The self-potential method is a passive-type electrical method. It consists of measuring, by means of a pair of (nonpolarizable) electrodes pushed into the soil and connected to a multivoltmeter with high impedance, the potential drop among points of the surface owing to the presence of an electric field produced by natural sources distributed in the subsurface. Regarding the genesis of the self-potential, these are due to different physico-chemical mechanisms: i) potential of membrane or of diffusion that originates because of electrochemical reactions; ii) potential electrokinetics owing their existence to the motion of underground electrolytic fluids through porous systems. In the hydraulic case the phenomenon observed is electrokinetic, which is related to the drag of the excess charge found close to the pore water-mineral interface (Revil et al. [11]). In particular the fluid, subjected to a pressure gradient, circulates and transports cations, yielding, on one side, an excess of positive charges and on the other of negative ones. Until a few years ago the results of the self-potential method were qualitatively, that is, supplying an indication of the presence of anomalies that generated the phenomenon. Starting with the work of Revil et al. ([11], [17]), a relation has been developed between the hydraulic flow in the subsurface and the density of electrical current generated by it and measured at the surface. As shown in the Appendix, in an isotropic porous medium, the total electrical density J is given by the sum of a conductive current density, described by Ohm's law, and of a source current density described by the Darcy's law (e.g., Sill [6]):

$$J = -\sigma \nabla \varphi - L \rho g (\nabla h) \tag{1}$$

where J is the electrical current density (A m^{-2}), σ is the electrical conductivity (S m^{-1}), φ is the electrical potential (V), L is an electrokinetic coupling term (A m^{-1} Pa^{-1}), ρ the fluid density (kg m^{-3}), g the gravity acceleration (m s^{-2}) and h is the hydraulic head (m) given by $h = z + \psi$, where z is the elevation head and ψ is the pressure head. The principle of conservation of charge in the quasi-static limit, due to the rapid diffusivity of the electromagnetic disturbances, in a porous medium is given by the following continuity equation:

$$\nabla \cdot J = 0 \tag{2}$$

Substituting equation (1) into equation (2) leads to the following governing equation of the electrical flow for a heterogenous and anisotropic porous medium:

$$\nabla^2 \varphi = -\nabla \ln \sigma \nabla \varphi - C' \left(\nabla \ln L \nabla h + \nabla^2 h \right) \tag{3}$$

where $C' \equiv (\partial\varphi/\partial h)_{J=0} = -\rho g L/\sigma$ is the electrokinetic coupling coefficient associated with pressure head variations (V m^{-1}). If we assume the porous medium is homogeneous, from an electrical point of view (i.e. σ and L are constants), equation (3) reduces to

$$\nabla^2\varphi = -C'\nabla^2 h \tag{4}$$

Equation (4) takes different forms depending on the Laplacian of head, $\nabla^2 h$, and the hydraulic characteristics of the porous medium and flow conditions.

3. PUMPING TEST: HYDRAULIC CONDUCTIVITY ESTIMATION

Under steady state conditions, if we assume the porous medium is homogeneous in the electrical sense, the governing equation of the electrical flow is the (4). During a pumping test, under the Dupuit's assumption for a homogeneous unconfined aquifer, the hydraulic head $h(r)$ at distance r from the pumping well, for steady radial flow, is (de Marsily [25])

$$h(r) = \left[h_0{}^2 + \frac{Q}{\pi K}\ln\left(\frac{r}{R}\right)\right]^{1/2} \tag{5}$$

where Q is the volumetric pumping rate, h0 is the hydraulic head where the influence of the pumping is negligible and R is the influence radius. Combining the equations (4) and (5), and deriving the hydraulic head for steady radial flow in polar coordinates, $\nabla^2 h = 1/r(\partial/\partial r(r\,\partial h/\partial r))$, we obtain:

$$\nabla^2\varphi = \begin{cases} \Im(r) \quad r \in \Omega_i \quad \text{with} \quad \Im(r) = \dfrac{C'Q^2}{4\pi^2 K^2 r^2 h^3(r)} \\ \\ 0 \qquad r \in \Omega_e \end{cases} \tag{6}$$

where Ωi and Ωe are, respectively, the internal and external volumes of the source body in which fluid flow occurs (Figure 1). Application of Green's theorem to the previous boundary value problem yields (Sobolev [26], 1989)

$$\varphi(r) = -\frac{1}{\sigma_i}\int_{\Omega_i} G(r, r')\Im(r')dV \tag{7}$$

where G(r, r') is the Green's function and dV is a volume element of the ground.

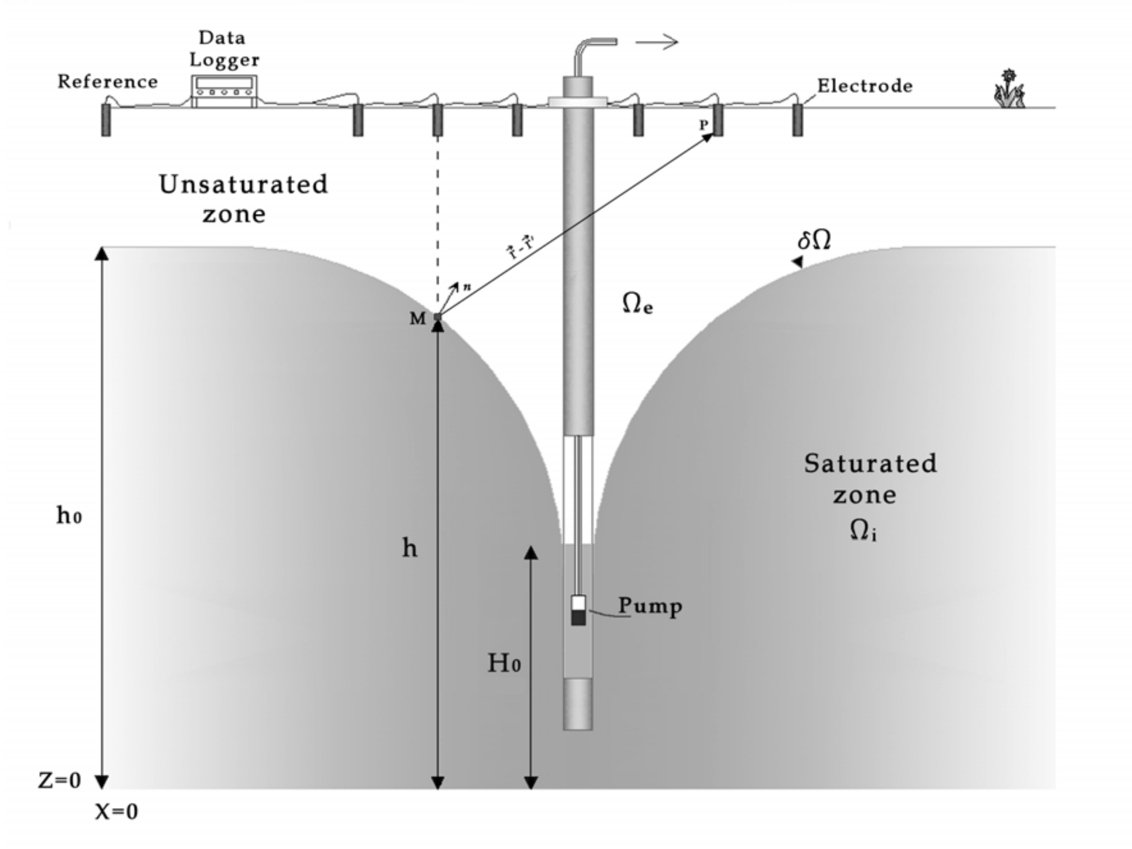

Fig. 1 Cross section of the depression cone of the water table in the vicinity of a pumping well in steady state conditions and monitoring of the self-potential signals. In this case each element of the water table behaves as a dipole with strength proportional to the piezometric head and directed along the normal to the water table. (Straface et al. [1])

The necessary and sufficient condition for this problem to have a solution is

$$\int_{\Omega_i} \Im(r')dV = \int_{\Omega_i} \nabla^2 h(r')dV = 0 \tag{8}$$

obtained by the continuity equation for the mass of water under steady state conditions. The electrical potential can only be measured relative to a reference electrode (where $\varphi_0 = 0$ by definition). We assume that this reference electrode is outside the "electrical radius of influence" of the pumping well. Thus, the reference electrode can be considered to be at infinity. Assuming the porous medium is homogeneous, the electrical potential depends only on the distance r from the pumping well. Integrating equation (7) yields

$$\varphi(r) = \left(\frac{C'Q^2}{4\pi^2 K^2}\right)\int_R^r \frac{dr'}{r'^2 h^2(r')} \tag{9}$$

By solving of equation (9) numerically, it is possible to use it as the forward model in the inversion of the self-potential signals registered at different distances r from the pumping well, for hydraulic conductivity. This method is more precise than that proposed by Rizzo et al. [12], because in the latter the integral is solved analytically assuming hydraulic head as a constant (i.e. the Boussinesq linearization).

4. THE HYDROGEOSITE AND EXPERIMENTAL TESTS

4.1 The laboratory

In 2006 at the Hydrogeosite Laboratory of the CNR-IMAA-Polo in Marsico Nuovo (PZ) a large scale sand tank was built by economic support of Italian Ministry of Scientific Research. It is sized 10×7×3 m^3, filled with homogeneous quartz-rich sand (95% of SiO_2) and equipped with a granulometry with a high percentage (86.4 %) of grains between 0.063mm and 0.125mm and a medium-high permeability (in the order of 10^{-5} m/s). 17 boreholes were installed (see Figure 2), a pumping well and drainage ring around its edge. The pumping well and all the boreholes are fully penetrating and consist of corrugated drainpipes in PET (Polyethylene terephthalate) and covered with geo-textile to stop the sand entering. The drainage ring, connected to two inlet/outlet tanks, enables fixed head boundary conditions to be obtained (fixed or variable with time) or no flow condition depending on the configurations required during the experiment. The pumping well is equipped with an external peristaltic-type pump, while the 17 boreholes are fitted with pressure transducers that enable continuous monitoring (with high precision) of the variations in the hydraulic head owing to pumping. To measure the SP response, 63 nonpolarizable Petiau electrodes (SDEC, Francia) are installed on the surface of the model, along 7 parallel lines 1m apart, each with 9 electrodes placed 1m apart (Figure 2). SP signals are measured by means of a multichannel system realized ad hoc and made up of a Keithley Instruments datalogger on which each nonpolarizable electrode was connected to the reference electrode placed in the area least influenced by the variation of the hydraulic head (close to boundary of the pool). The experiments were preceded by a preliminary study phase by means of a "numerical experiment" or rather a mathematical model capable of simulating the behaviour of the physical model and therefore of calculating the optimum pumping flow, the boundary conditions, the length of the transitory in the pumping and recovery phase, as well as the response of the physical model in terms of hydraulic head and self-potential. The pumping test was simulated using COMSOL Multiphysics, a generalized finite-element modeling environment (COMSOL [26]). Once the preliminary study phase was concluded, the facility was filled from the bottom to avoid the entrapment of air bubbles and, using the drainage ring, the (initial) hydraulic head was fixed at 2.80 m above the bottom of the facility.

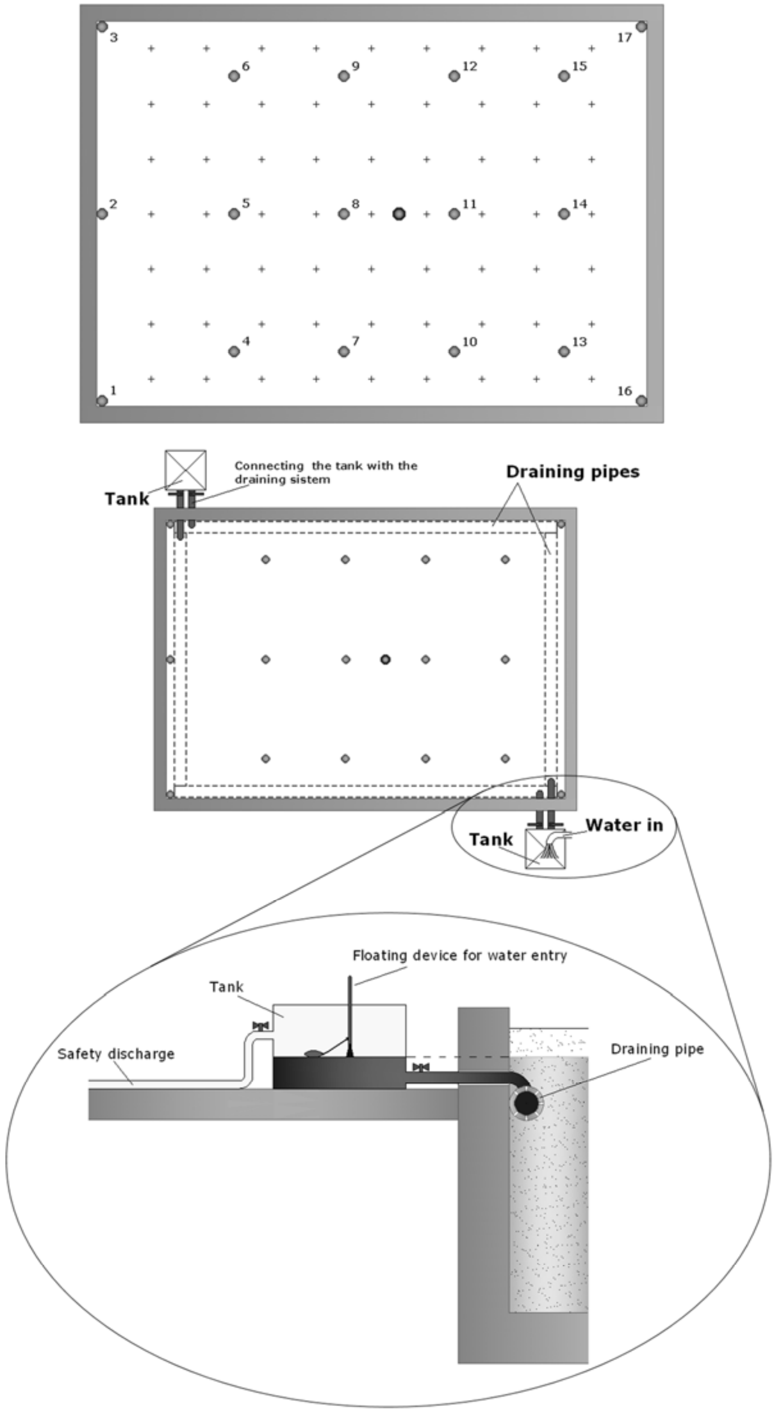

Fig. 2 Sketch of the Hydrogeosite laboratory facility; (a) it is sized 10×7×3 m^3, filled with homogeneous quartz-rich sand and equipped with 17 boreholes and one pumping well (circles), 63 nonpolarizable Petiau electrodes (crosses) installed on the surface of the model, along 7 parallel lines 1m apart, each of which with 9 electrodes with an electrodic distance of 1m; (b) the drainage ring, connected to two inlet/outlet tanks, enables fixed head boundary conditions or no flow condition depending on the configurations required during the experiment. (Straface et al. [1])

4.1 The laboratory

The flow rate pumped from the well was fixed at 0.085 l/sec, according to what had been calculated with the mathematical model, and monitored using an electromagnetic flowmeter with digital converter. The pumping lasted about 8 hours, while the subsequent recovery phase lasted about 12 hours. Figure 3 shows plots of the drawdowns, measured using the pressure transducers in eight piezometers, with respect to time. Figure 4 shows the SP curve acquired with time in a channel (ch. 43) during the pumping test, after a correction of the drift effect due negligible phenomena (i.e. electrode drift). The drift correction was achieved by use of a detrending statistic tool which involved fitting a polynomial function to pre-pumping data (usually it was linear) and then subtracting it from all the data. As can be easily seen, the different hydraulic phases of the experiment are distinctly evidenced by the SP electrical signal and its shape is a mirror image of that obtained from the drawdown measured in the boreholes. In fact, during the pumping phase an increasing trend of the SP signal is observed until it becomes constant at the moment that it reaches stationary state. After the shutdown of the pump, in the piezometric recovery phase, the electrical signal is almost restored to its initial conditions. Moreover, only 45 SP data were used in the following analysis, because there were problems during data acquisition with some of the SP sensors.

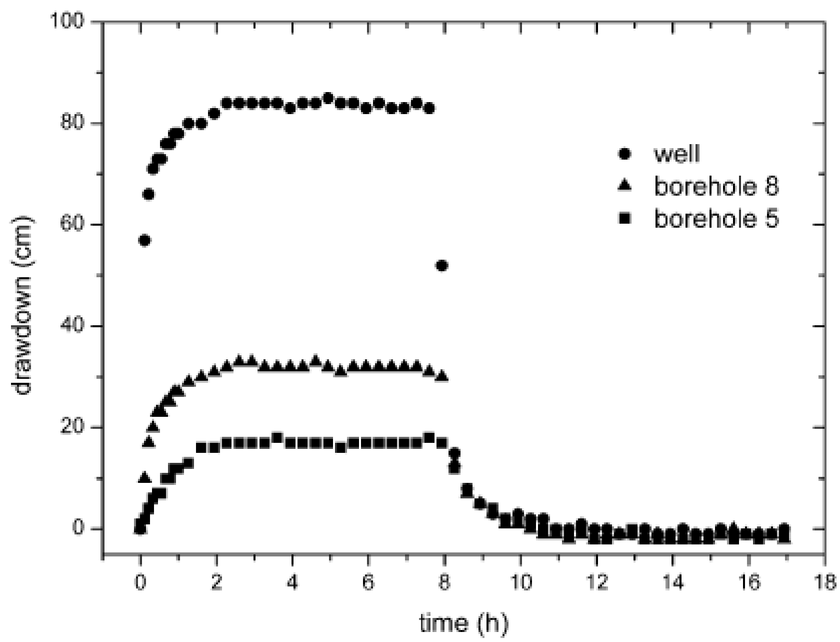

Fig. 3 Drawdown time curves in a few monitored boreholes and in the well. The steady state condition is reached after about 3 h from the start of the pumping, and after about 8 h the pump has been shut down and the recovery phase begins. (Straface et al. [1])

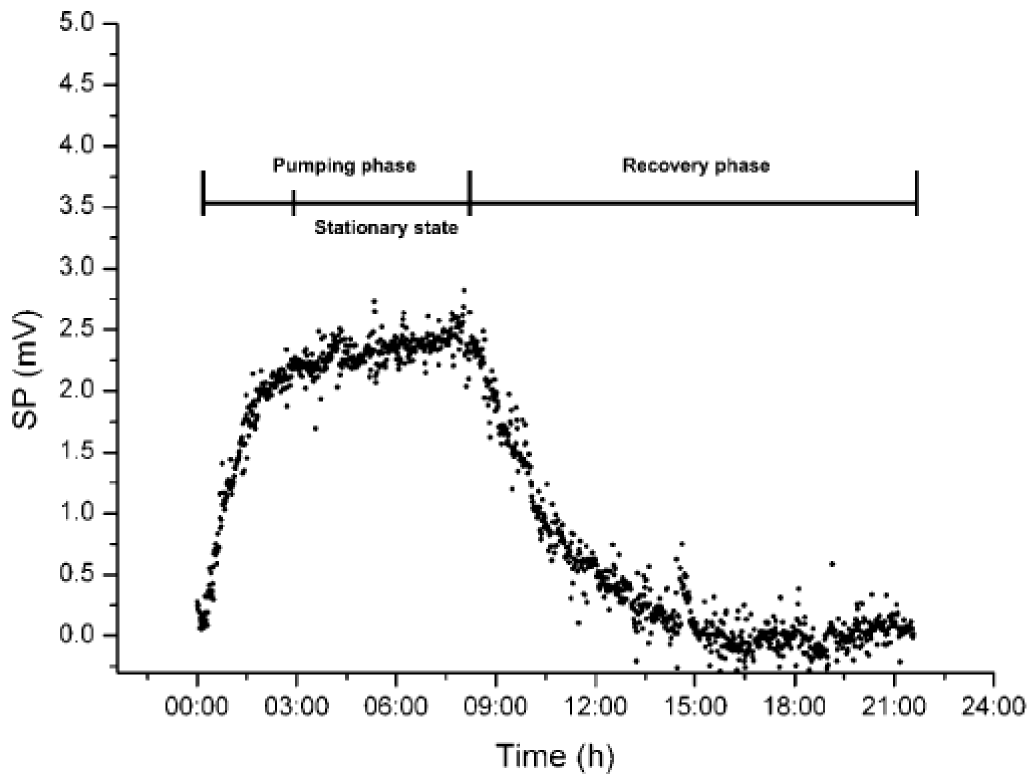

Fig. 4 Self-potential curve acquired with time in a channel (ch. 43) during the pumping test. The different hydraulic phases of the experiment are distinctly evidenced by the self-potential signal, and its shape is a mirror image of that obtained from the drawdown measured in the boreholes. During the pumping phase an increasing trend of the self-potential signal is observed until it becomes quite constant at the moment that it reaches stationary state (about 3 h). After the shutdown of the pump (about 7 h), in the piezometric recovery phase, the electrical signal is restored to its initial conditions. (Straface et al. [1])

5. THE HYDROGEOSITE AND EXPERIMENTAL TESTS

Hydraulic conductivity, K, and specific yield, Sy, of the porous medium were determined from the transient drawdown data. Since the laboratory setup simulates an unconfined aquifer, the Neuman [27] Type-Curve method (1975) was used in the analysis. It yields the two parameters, K and Sy, for a homogeneous and anisotropic unconfined aquifer of infinite radial extent and fully penetrating well and with delayed drainage. According to this theory, the drawdown-time dimensionless curve shows a steep initial phase, corresponding to release of water from elastic storage, a flatter second length at the intermediate times due to an additional source of water released form storage with some delay in time, and finally, when most of water is derived, a steeper segment at later times. This curve trend differs from the Theis solution for confined aquifers in the presence of the intermediate phase that interrupts the monotonic trend of the curve. This phenomenon has been traditionally referred in groundwater literature as “delayed yield”. Observing the experimental drawdown-time data on the Neuman Type-Curve it is found that they are monotonic or rather, that they do not have the intermediate length (Figure 5). The main reason for this behaviour can be ascribed to the thickness of the saturated zone (about 2.80 m), much thicker than the unsaturated zone (about 0.20 m). This implies that the quantity of stored water in the unsaturated zone is small and therefore the delayed drainage is negligible (Tartakovsky and Neuman [28]).

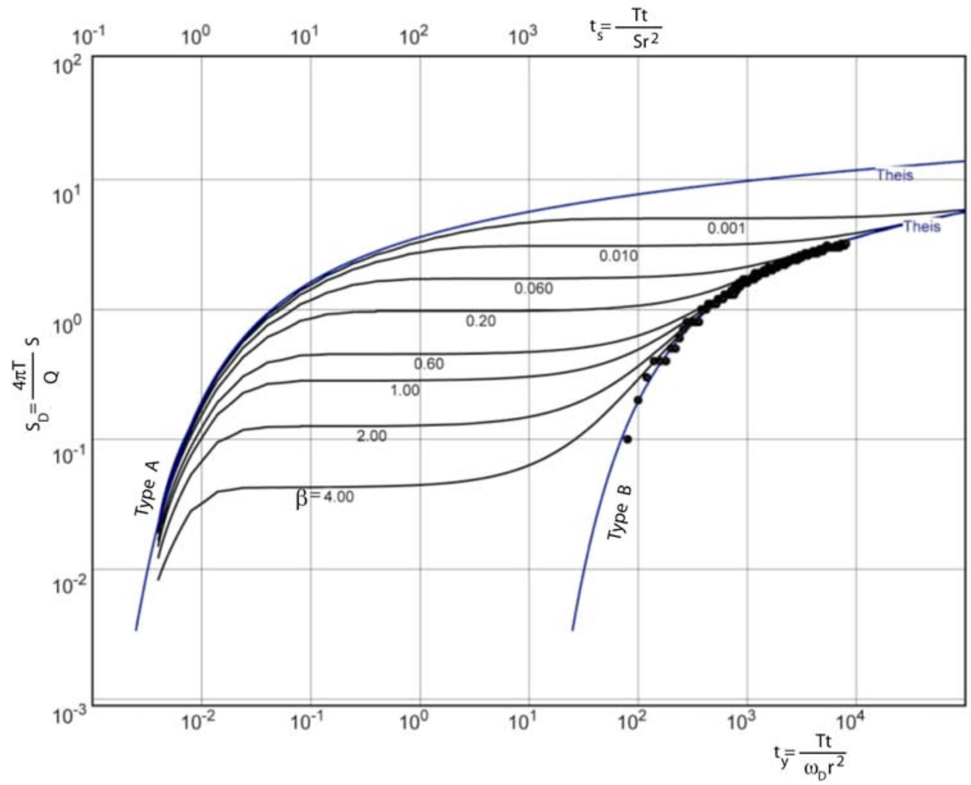

Fig. 5 Analysis of the drawdown for borehole 11 using the Neuman type curve method. The monotonic behavior and the absence of intermediate length can be ascribed mainly to the thickness of the saturated zone (about 2.80 m), much thicker than the unsaturated zone. This implies that the quantity of stored water in the unsaturated zone is small; therefore, the delayed drainage is negligible [Tartakovsky and Neuman, 2007]. The same behavior was obtained for each drawdown time curve of the monitored boreholes. (Straface et al. [1])

Moreover, the effect of the vertical flow in the unsaturated zone on the drawdown in a borehole decreases with an increase of the degree of penetration of the well and of the observation borehole - in the Hydrogeosite both boreholes and pumping well are fully penetrating. On the other hand, the effect of unsaturated flow is most pronounced at the water table, more so closer to the pumping well than farther from it, and least near the aquifer bottom. This is so because the average is less sensitive than are point drawdown to vertical fluxes and hence to drainage from above the water table. This was also found by Tartakovsky and Neuman [28] analyzing a series of pumping tests carried out at Cape Code (Massachusetts). To confirm this interpretation two further pumping tests were carried out in the Hydrogeosite using a flow rate lower and greater than the very first of 0.085 l/s. So we have realized a pumping test with a flow rate of 0.06 l/s and another of 0.12 l/s. We have chosen this two flow rates in order to investigate on the possibility of attributing the "non appearance" of the intermediate phase to the influence of the boundaries, i.e. to the vicinity of the Dirichlet boundary condition imposed by means of the drainage ring. Moreover, the closeness of the boundaries is in contrast with the hypothesis of "infinite extension" of the Neuman type curve method. Nevertheless, as we can see in Figure 6, the drawdown time curves for these different pumping tests show, again, a monotonic behaviour confirming the previous interpretation.

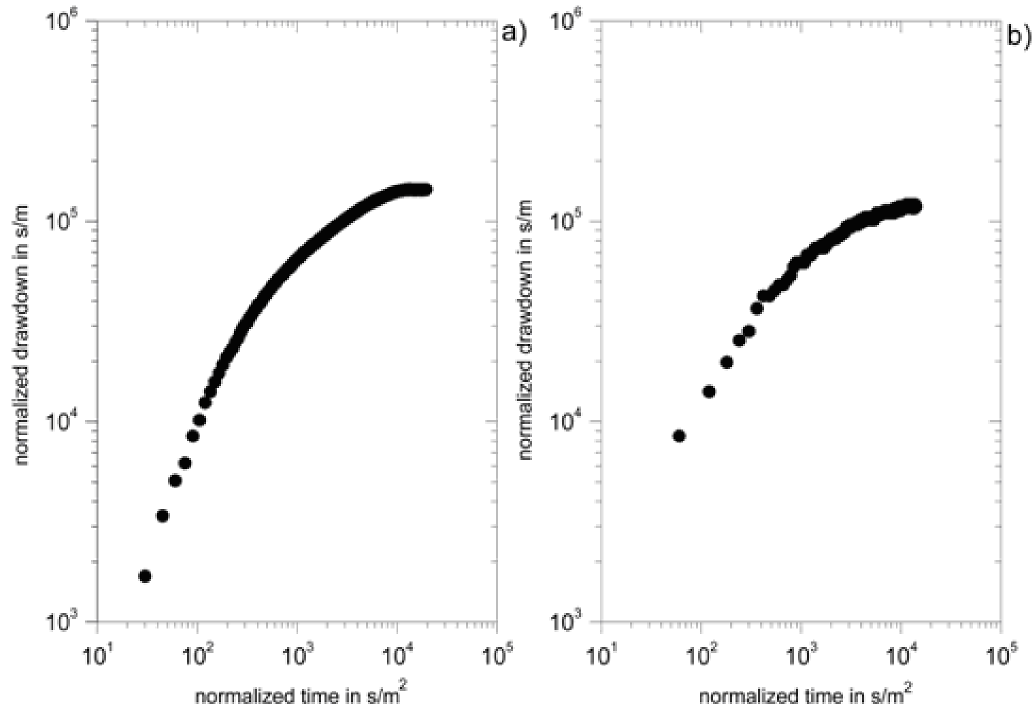

Fig. 6 Bi-logarithmic graphs of the normalized drawdown ($w_n = 4\pi Bw/Q$) versus normalized time ($t_n = t/r^2$) for borehole 11. The graphs show a comparison between two pumping tests with different flow rate (Q=0.12 l/s on the left and Q=0.06 l/s on the right) in order to investigate on the possibility to attribute the "non appearance" of the intermediate phase to the vicinity of the Dirichlet boundary condition imposed by means of the drainage ring. The same behaviour was obtained for all the other monitored boreholes. (Straface et al. [1])

Applying the Neuman Type-Curve of (1975) to the pumping test in order to determine the hydraulic conductivity K and the specific yield S_y (Figure 5) it is easy to see the points are aligned on the Theis-type curve confirming what was previously supposed. Working out for all boreholes monitored the results shown in Table 1 were obtained. The permeability values obtained fall within the order of magnitude suggested by the sand suppliers and this is a further confirmation of the consistency of the interpretation used in this study of characterization.

Table 1 Hydraulic conductivity K and specific yield Sy values obtained with the Neuman-type Curve method (1975) in all the monitored boreholes. (Straface et al. [1])

Borehole	K (m/s)	S_y
4	3.82×10^{-5}	0.69×10^{-2}
5	3.41×10^{-5}	0.71×10^{-2}
8	3.41×10^{-5}	0.67×10^{-2}
9	3.41×10^{-5}	0.89×10^{-2}
10	3.41×10^{-5}	0.79×10^{-2}
11	2.71×10^{-5}	1.12×10^{-2}
14	3.04×10^{-5}	0.70×10^{-2}
mean	$\mathbf{3.31 \times 10^{-5}}$	$\mathbf{0.79 \times 10^{-2}}$

6. ESTIMATION OF HYDRAULIC CONDUCTIVITY USING SP DATA

In order to estimate the hydraulic conductivity of the porous medium filling the Hydrogeosite, by means of the pumping test, the self-potential signals measured in pseudo-steady state condition were used. When assuming the homogeneity of the porous medium, the electrical potential depends only on the distance r from the pumping well. So, the self-potential signals generated by the groundwater flow can be used to estimate the hydraulic conductivity of an unconfined aquifer, by inverting equation (9). Rizzo et al. [12] proposed a graphical method, based on the Boussinesq linearization, to calculate the hydraulic conductivity. The integral in the equation (9) can be analytically solved if the steady state condition is achieved, and, consequently, the electrical potential can be expressed by means of a power law relationship (i.e. $\varphi(r) = Ar^{-1}$). Then, it becomes possible to estimate the mean hydraulic conductivity of an unconfined aquifer, by calculating the value of A from a best fitting of the electrical potential data in terms of the distance from the pumping well. In this paper, we propose a new method that overcomes the approximation of Rizzo et al. [12] who assumed constant head. To invert hydraulic conductivity a nonlinear least square method is proposed. It is based on the minimization of the following cost function

$$F = (\varphi(r) - \hat{\varphi}(r))^T S_P^{-1} (\varphi(r) - \hat{\varphi}(r)) \tag{10}$$

With the purpose to compare our approach with that proposed by Rizzo et al. [12], pumping test data were used on both methodologies, to characterize the above mentioned aquifer. Regarding the graphical approach, the electrical potential data are plotted, as a function of the distance from the pumping well, in Figure 7. The application of a power law relationship to these data yields $\varphi(r) = 2.923r^{-1}$, where the electrical potential φ is expressed in mV and the distance r is in meters. A Standard Error around 0.19 and an Adjusted R^2 (Everitt [29]) of 0.85 were obtained. The hydraulic conductivity K can be calculated through the analytical solution of equation (9): $A = -C'Q^2 / 4\pi^2 K^2 H_0^2$, where: H_0 is the hydraulic head in the pumping well (equal to 1.96); C' represents the electrokinetic coupling coefficient (equal to 7.4); A is the coefficient of the power law above determined (equal to 2.923); $Q = 0.085$ l/s is the pumping rate. The corresponding hydraulic conductivity is equal to $K = 5.8\times10^{-6}$ m/s.

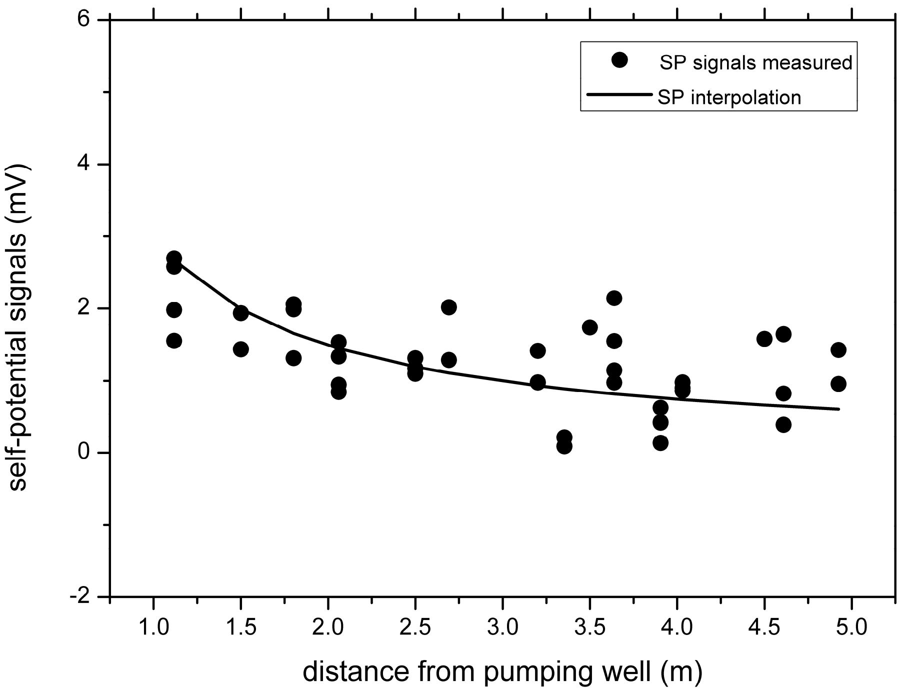

Fig. 7 Variation of the electrical potential from the distance from the pumping well. In steady state condition, the self-potential signal decreases as the inverse of the distance from the pumping well. The application of a power law relationship to these data yields $\varphi(r) = 2.923r^{-1}$, with the electrical potential φ expressed in mV and the distance r expressed in m. (Straface et al. [1])

In the new proposed methodology the integral of the equation (9) was solved numerically by means of a recursive adaptive Simpson quadrature algorithm (MATLAB® [30]). After the integral calculation, the non linear least square fit of φ was applied to get K. In this case K was equal to 1.52×10^{-5} m/s. This value is greater with respect to that previously estimated, but it is closer to the hydraulic conductivity value obtained by means of the Neuman type-curve method.

7. CONCLUSIONS

This paper reports a hydrogeophysical experiment in which self-potential signals generated by the groundwater flow have been measured on the ground surface in a large scale facility (Hydrogeosite Laboratory) both during pumping and recovery phases. This work points out that the hydraulic test can be well replicated in the used laboratory. As a matter of fact, the drawdown, measured during the hydraulic test, coincides with data calculated by the mathematical model. Moreover, the paper remarks the usefulness to use a hydrogeophysical approach to obtain the average hydraulic conductivity. A value of the hydraulic conductivity $K= 1.52 \times 10^{-5}$ m/s, very close to that derived by means of the Neuman type-curve method ($K_{ave} = 3.31 \times 10^{-5}$ m/s) has been obtained by solving numerically the integral of the equation (9), and using the non linear least square method. The value of K furnished by this new methodology is much greater with respect to that obtained by Boussinesq linearization ($K = 5.8 \times 10^{-6}$ m/s), proposed by Rizzo et al. [12]. The calculated hydraulic conductivities all fall within the order of magnitude given by the sand suppliers. Once more it has been shown that the nature of the problem here investigated necessarily requires an interdisciplinary approach, able to involve hydraulic and geophysical skills, in the planning and execution of the experiments, and in the conceptual and numerical modelling of the underground flow and transport processes. In conclusion it can be stated that, the methodology here presented, even though in its initial stage, promises important progress in the development of new hydrogeophysical techniques and, more generally, in the different approaches to inverse problem.

ACKNOWLEDGMENTS

We thank Prof. Vincenzo Cuomo, director of CNR-IMAA, and Dr. Vincenzo Lapenna, responsible for geophysical group, for their support and for facilitating the use of Hydrogeosite Laboratory.

REFERENCES

[1] Straface, S., Rizzo, E., and Chidichimo, F., (2010), Estimation of water table map and hydraulic conductivity in a large scale model by means of the SP Method, *J. Geophys. Res.*, 115, doi: 10.1029/2009JB007053.

[2] Straface S., Fallico, C., Troisi, S., Rizzo, E., and Revil, A., (2007), An inverse procedure to estimate transmissivity from heads and Self-Potential signals, *Ground water*, *45*(4), 420-428.

[3] Cassiani, G., and Medina, M.A., (1997), Incorporating auxiliary geophysical data into ground-water flow parameter estimation, *Ground Water*, *35*, 79– 91.

[4] Cassiani, G., Böhm, G., Vesnaver, A., and Nicolich, R., (1998), A geostatistical framework for incorporating seismic tomography auxiliary data into hydraulic conductivity estimation, *J. Hydrol.*, *206*, 58– 74.

[5] Troisi, S., Fallico, C., Straface, S., and Migliari, E., (2000), Application of kriging with external drift to estimate hydraulic conductivity from electrical resistivity data in unconsolidated deposits near Montalto Uffugo, Italy, *Hydrogeol. J.*, *8*, 356– 367.

[6] Poldini, E., (1938), Geophysical exploration by spontaneous polarization methods, *Mineral. Mag.*, *59*, 278–282.

[7] Sill, W.R., (1938), Self-potential modeling from primary flows, *Geophysics*, *48*, 76– 86.

[8] Bernabé, Y., (1998), Streaming potential in heterogeneous networks, *J. Geophys. Res.*, *103*, 20827–20841.

[9] Lorne, B., Perrier, F., and Avouac, J.P., (1999), Streaming potential measurements: 1. Properties of the electrical double layer from crushed rock samples, *J. Geophys. Res.*, *104*, 17857–17877.

[10] Fitterman, D.V., (1978), Electrokinetic and magnetic anomalies associated with dilatant regions in a layered Earth, *J. Geophys. Res.*, *83*, 5923–5928.

[11] Revil, A., Naudet, V., Nouzaret, J., and Pessel, M., (2003), Principles of electrography applied to self-potential electrokinetic sources and hydrogeological applications, *Water Resour. Res.*, *39*(5), 1114, doi:10.1029/2001WR000916.

[12] Rizzo, E., Suski, B., Revil, A., Straface, S., and Troisi, S., (2004), Self-potential signals associated with pumping-tests experiments, *J. Geophys. Res.*, *109* (B10203), doi:10.1029/2004JB003049.

[13] Titov, K., Ilyin, Y., Konosavski, P., and Levitski, A., (2002), Electrokinetic spontaneous polarization in porous media: Petrophysics and numerical modeling, *J. Hydrol.*, *267*, 207–216.

[14] Titov, K., Revil, A., Konasovsky, P., Straface, S., and Troisi, S., (2005), Numerical modeling of selfpotential signals associated with a pumping test experiment. *Geophys. J. Int.*, *162*, 641-650.

[15] Patella, D., (1997), Introduction to ground surface self-potential tomography. *Geophysical Prospecting*, *45*, 653-681.

[16] Revil, A., Cathles, L.M., Losh, S., and Nunn, J.A., (1998), Electrical conductivity in shaly sands with geophysical applications, *J. Geophys. Res.*, *103*, 23925–23936.

[17] Revil, A., Hermitte, D., Voltz, M., Moussa, R., Lacas, J.G., Bourrié, G., and Trolard, F., (2002), Self-potential signals associated with variations of the hydraulic head during an infiltration experiment, *Geophys. Res. Lett.*, *29*(7), 1106, doi:10.1029/2001GL014294.

[18] Darnet, M., Marquis, G., and Saillac, P., (2003), Estimating aquifer hydraulic properties from the inversion of surface streaming potential (SP) anomalies, *Geophys. Res. Lett.*, *30*(13), 1679, doi:10.1029/ 2003GL017631.

[19] Suski, B., Revil, A., Titov, K., Konosavsky, P., Voltz, M., Dage`s, C., and Huttel, O., (2006), Monitoring of an infiltration experiment using the self-potential method, *Water Resour. Res.*, *42*(W08418), doi:10.1029/2005WR004840.

[20] Linde, N., Revil, A., Bolève, A., Dagès, C., Castermant, J., Suski, B., Voltz, M., (2007), Estimation of the water table throughout a catchment using self-potential and piezometric data in a Bayesian framework. *J. Hydrol.*, 334, 88-98.

[21] Malama, B., Revil, A., and Kuhlman, K.L., (2009a), A semi-analytical solution for transient streaming potentials associated with confined aquifer pumping tests, *Geophys. J. Int.*, 176, 1007–1016.

[22] Malama, B., Kuhlman, K.L., Revil, A., (2009b), Theory of transient streaming potentials associated with axial-symmetric flow in unconfined aquifers, *Geophys. J. Int.*, 179, 990-1003, doi: 10.1111/j.1365-246X.2009.04336.x.

[23] Barrash, W., Revil, A., Cardiff, M., Johnson, T., Malama, B., Rizzo, E., Miller, C., and Straface, S., (2009), Combined Hydraulic Tomography – Self Potential – Electrical Resistivity Tomography Experimentation for characterization and imaging of the unconfined aquifer at the Boise Hydrogeophysical Research Site, *Technical Report BSU CGISS 08-0X*.

[24] Jardani, A., Revil, A., Barrash, W., Crespy, A., Rizzo, E., Straface, S., Cardiff, M., Malama, B., Miller, C., and Johnson, T., (2008), Reconstruction of the Water Table from Self-Potential Data: A Bayesian Approach. *Ground Water*, doi: 10.1111/j.1745-6584.2008.00513.x.

[25] Marsily, G., De, *Quantitative Hydrogeology - Groundwater Hydrology for Engineers,* Accademic Press., 440 pp., 1986.

[26] COMSOL, *COMSOL: Multiphysics User's Guide*, Version 3.5, COMSOL, Stockholm, Sweden., 2008.

[27] Neuman, S.P., (1975), Analysis of pumping test data from anisotropic unconfined aquifers considering delayed gravity response, *Water Resour. Res.*, *11*(2), pp. 329-342.

[28] Tartakovsky, G.D., and Neuman, S.P., (2007), Three-dimensional saturated-unsaturated flow with axial symmetry to a partially penetrating well in a compressible unconfined acquifer, *Water Resour. Res.*, *43*(W01410), doi:10.1029/2006WR005153.

[29] Everitt, B.S., (2002) *The Cambridge Dictionary of Statistics*, King's College London, CUP.

[30] MATLAB®, *The Language of Technical Computing,* © 1994-2008 The MathWorks, Inc.