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Hydrogeohysics techniques for the characterization of a heterogeneous aquifer

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ABSTRACT The target of this work is the determination of hydraulic conductivity spatial distribution in a real aquifer using a hydrogeophysical approach. This was made possible by analyzing geoelectrical data coming from "cross-hole" electrical resistivity tomography. The geophysical techniques most used in hydrogeological studies are the geoelectrical methods because electrical resistivity is very sensitive to the presence of water. Since the electrical resistivity strongly depends on some hydraulic parameters such as porosity, water content and hydraulic conductivity, several relationships have been developed in order to estimate these hydraulic parameters using the measured electrical resistivity values. In this work, we present an experiment carried out at the Montalto Uffugo test site (Cosenza, Italy) of University of Calabria, where we determined the spatial distribution of hydraulic conductivity using a geoelectrical technique with electrodes located in two boreholes. The cross-hole electrical resistivity tomography (ERT) allowed us to obtain the resistivity distribution in the subsoil and compare it with the geological-stratigraphic information obtained through the analysis of two cores. Then, the data acquisition with the ensuing elaboration led to the characterization of the main aquifer of Montalto Uffugo and the reconstruction of the hydraulic conductivity distribution in the subsoil applying Archie's Law and Purvance and Andricevic's Law. The estimated hydraulic conductivity compares favourably with values previously determined in situ by some hydraulic tests.

Key words: hydraulic conductivity, electrical resistivity, porous medium, Calabria.

1. Introduction

A primary objective in hydrogeological investigations is to obtain information on the hydraulic conductivity of rocks. The characterization of the spatial distribution of hydraulic properties of porous media is a necessary step toward high resolution predictions of water flow and contaminant transport in an aquifer (Straface *et al*., 2007a).

Hydraulic conductivity is usually obtained from pumping test experiments (Domenico and Schwartz, 1997) and down-hole measurements (Rabaute *et al*., 2003); however, these traditional test methods are thought to yield average hydraulic properties over a large volume of geological media (Butler and Liu, 1993). The use of geophysics in a hydrogeological context has several advantages; one of them is to obtain a great deal of information in a small amount of time and at a low cost. The geophysical techniques most used in hydrogeological studies are geoelectrical methods because electrical resistivity is very sensitive to water in the pores (Archie, 1942).

Electrical resistivity of the ground is a function of some hydraulic parameters (porosity, water content, and hydraulic conductivity); in particular, Archie's empirical Law defines the electrical resistivity as a function of porosity and water content while Purvance and Andricevic (2000) recently described a relation between hydraulic conductivity and electrical conductivity.

In this work, an experiment carried out at the Montalto Uffugo test site (Cosenza, Italy) of University of Calabria is described. Several hydraulic and geophysical tests have been conducted in the Montalto wellfield since the early 1990s until today (Troisi and Straface, 1996; Fallico *et al*., 2002; Rizzo *et al*., 2004; Straface *et al*., 2007b).

The aim of this study is to estimate the hydraulic conductivity spatial distribution in a real heterogeneous porous aquifer using a hydrogeophysical approach. This was made possible by analyzing geoelectrical data obtained through electrical resistivity tomography (ERT) of the "cross-hole" type, using electrodes located in two boreholes. Cross-borehole resistivity imaging is an extension of conventional surface resistivity imaging and it uses similar inverse modelling techniques. However, in comparison with conventional, surface-deployed surveys, this method has been shown to provide high-resolution images of hydrogeological structures at depth and, in some cases, detailed assessment of dynamic processes in the subsurface environment (Binley *et al*., 2002).

In cross-borehole ERT, quadrupole resistance measurements are made using electrodes in two or more boreholes. Often, surface electrodes are used to supplement the electrode array. Inversion of the resistance data is necessary in order to estimate an image of resistivity between the boreholes, by discretising the domain of interest in parameter cells: the objective of the inversion procedure is to compute the 'best' set of resistivity values, which satisfies both the measured data set and some a priori constraints, in order to stabilize the inversion and constrain the final image (deGroot-Hedlin and Constable, 1990).

Cross-borehole ERT has been employed in a wide range of environments. One of the earliest examples of hydrological applications of ERT is in Daily *et al*. (1992): they studied the vadose zone, moisture migration through the application of a tracer. Other examples of vadose zone studies using cross-hole ERT include Ramirez *et al*. (1996), Slater *et al*. (2000), and Binley *et al*. (2002).

As pointed out by Binley and Kemna (2005), the main advantage of cross-borehole resistivity imaging, in comparison to surface resistivity imaging, is the high resolution at depth. However, the method suffers from a number of disadvantages: 1) boreholes are required (which often need to be purpose drilled), 2) the images cover only the region between the boreholes, 3) the boreholes must not be too far apart otherwise sensitivity is reduced, 4) the borehole and the electrode characteristics cause data noise levels usually higher than those using surface electrodes, therefore, data processing techniques are more complex.

2. Hydrological and geophysical data

2.1. Geological setting and description of the experiment

The test site extends over a total area of 2100 m^2 and is located near the town of Montalto Uffugo, in the region of Calabria, in southern Italy (Troisi *et al*., 2000).

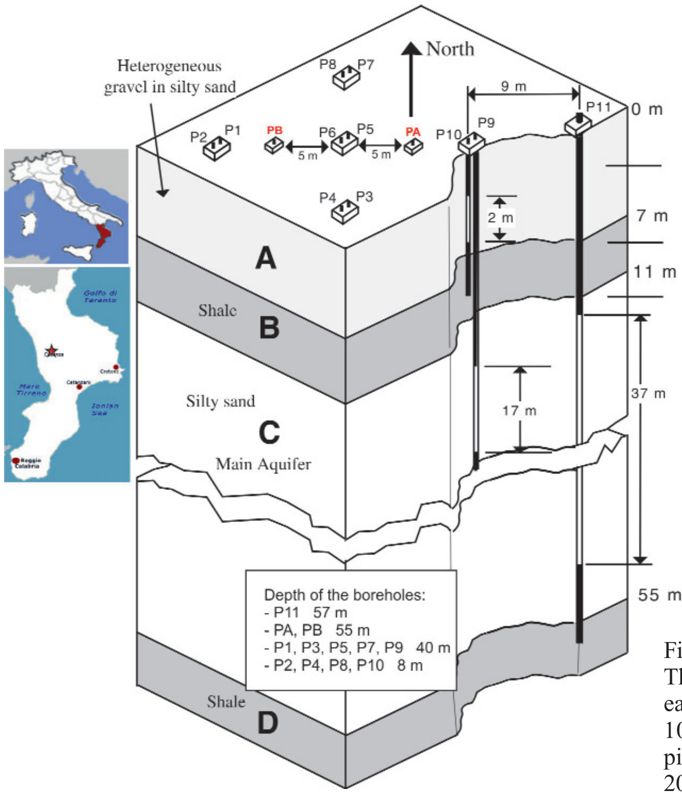

Fig. 1 - Sketch of the test site (Rizzo *et al*., 2004). The test site comprises five monitoring units, each unit is composed of a 40 m deep well plus 10 m deep piezometers. Two additional piezometers (PA and PB) were drilled in May 2009 to a depth of 52 m.

The geology of the test site, reconstructed by Rizzo *et al*. (2004), can be divided into four geological formations (Fig. 1). Heterogeneous gravels in a silty sand matrix (formation A) compose the first formation. This formation extends from the ground surface to a depth of 7 m. The second formation is a shale layer (formation B in Fig. 1, from 7 to 11 m). The third formation bears the main aquifer, investigated in this paper, and is composed of a silty sand layer (layer C, from 11 to 55 m). The deepest formation is the shale substratum (formation D, Fig. 1). A shallow, perched aquifer is present during part of the year in formation A. The water table is at approximately -6 m.

Ten boreholes (two boreholes per monitoring location) have been drilled at this test site. They are numbered P1 to P10. The locations of the boreholes are shown in Fig. 1. Each pair includes a borehole reaching a depth of 8 m (i.e., reaching the shallow perched aquifer) and a second borehole reaching a depth of 40 m and is, therefore, connected to the aquifer of interest. Moreover, there is an additional borehole coded P11 (pumping well), which is located 19 m from the central well P5; it reaches the impervious bottom of the main aquifer (Fig. 1). Troisi *et al*. (2000) estimated the hydraulic conductivity of Montalto Uffugo aquifer, whose values are reported in Table 1.

In May 2009, at the test site, two new boreholes (PA and PB, Fig. 1) were drilled at approximately 5 m from the central well P5 and installed at a depth of 55 m. These boreholes were drilled in order to estimate the hydraulic conductivity spatial distribution using cross-

Table 1 - Hydraulic conductivity values estimated by a Lefranc test (Troisi *et al*., 2000).

Depth (m)	Hydraulic conductivity (m s^{-1}) (main aquifer)
-18.00	1.65×10^{-6}
-28.00	6.42×10^{-6}

borehole ERT. Boreholes were completed with a 75 mm PVC casing to allow access for electrodes and cables. In each new borehole, 24 plate steel electrodes extended between depths of 6 to 52 m, with a 2 m spacing; the contact with the ground was achieved using a gravel grout.

The experiment was performed in July 2009 and consisted in taking geoelectrical measurements, in natural flow conditions, using the cross-borehole resistivity technique. Prior to the experiment, the piezometric levels and the water electrical conductivity were also measured directly in boreholes P1, P5, and P9 down to -35 m.

The 48 steel electrodes, installed along the two new piezometers, were used in order to perform the 2D cross-borehole resistivity measurements between a 6÷52 m depth. They were connected to a multi-electrode resistivity meter (SYSCAL Pro Switch resistivity instrument) by two multichannel cables.

A cross-borehole, azimuthal dipole–dipole array was adopted using a current and potential dipole separation (D) of 2 m up to 6 m (Fig. 2). A total of 2700 measurements were taken with reciprocal (swapped current and potential electrodes) measurements.

The collection of measurements in the reciprocal configuration, which took approximately 45 minutes, permitted the assessment of data errors (Binley *et al*., 2002).

2.2. Geophysical data

To obtain the true electrical resistivity values, apparent resistivity values, acquired during a data survey, have been inverted using the RES2DINV software (Geotomo Software) using a standard smoothness-constrained inversion method implemented by a quasi-Newton optimization technique.

After the inversion, electrical resistivity values were filtered in order to delete the noise due to the effect of borehole and electrodes. Then, analyzing the resistivity image (Fig. 3), we can observe a laterally heterogeneous distribution; nevertheless, we can identify an upper zone (to -17 m from ground level) at low resistivity ($\rho < 20$ Ωm), a central zone (from -17 to about -42 m) characterized by higher resistivity values ($\rho > 50$ Ωm) and the deepest zone characterized again by low resistivity values ($\rho < 10$ Ωm).

Comparing the resistivity tomography obtained with the test site stratigraphy (Troisi *et al*., 2000), we can see that the upper zone, characterized by low resistivity, corresponds to: the shallow aquifer (down to -7 m), the upper clay level (from -7 m to -11 m) and a silty-sand 3-m thick layer. Instead, the higher resistivity central zone corresponds to the test site's main aquifer, characterized by sand with crystalline fragments alternated with clay levels. Finally, the lower zone corresponds to the aquifer's lower portion where the alternation between sand and clay layers becomes denser until it reaches the clayey basement at about 50-51 m.

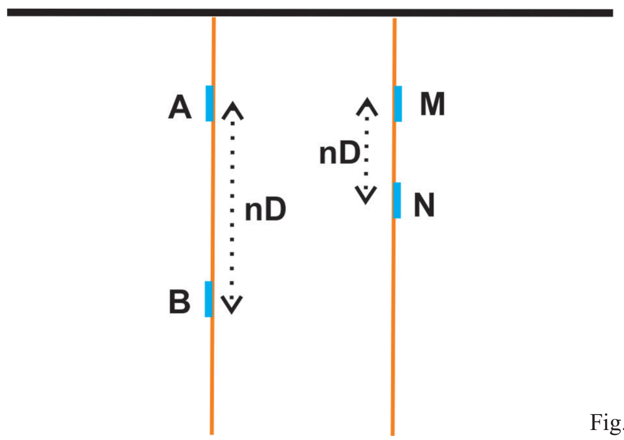

Fig. 2 - Sketch of the azimuthal dipole–dipole array used.

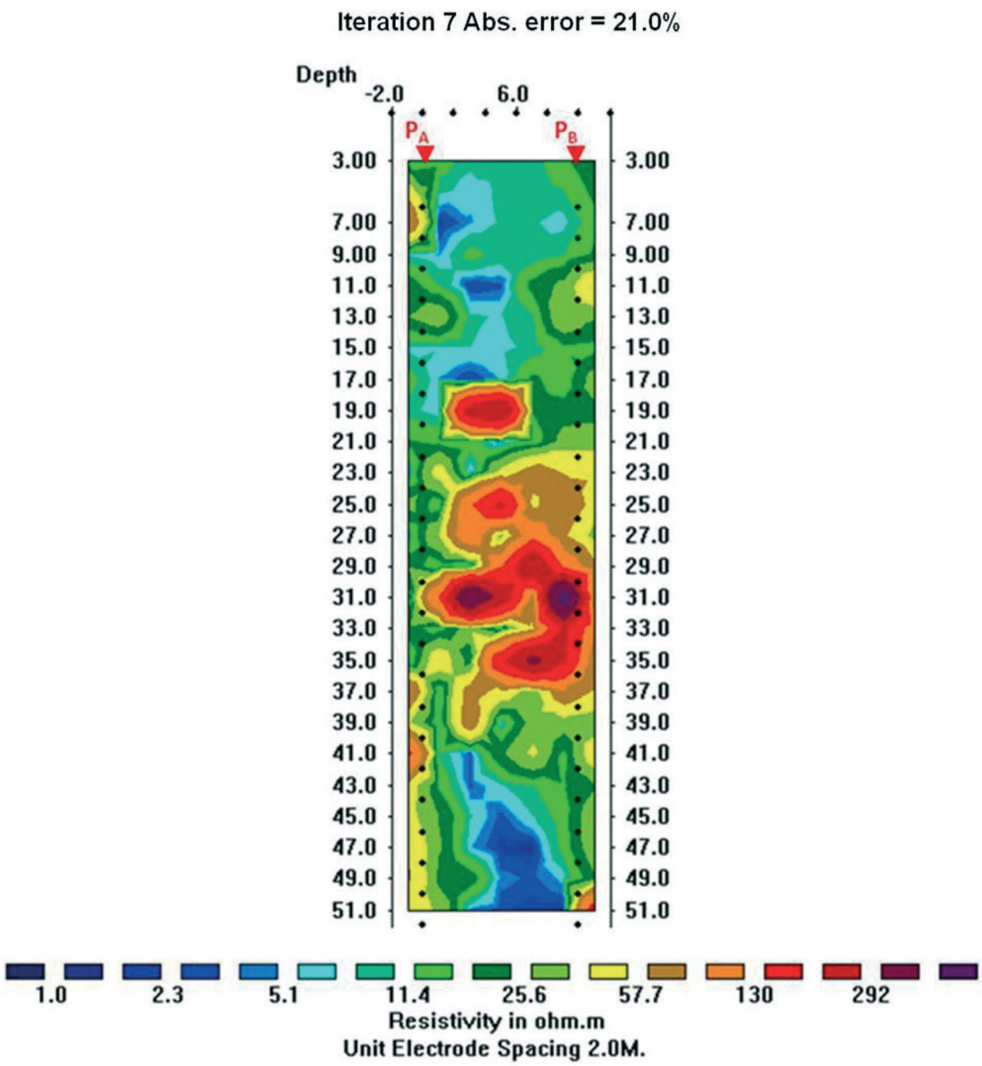

Fig. 3 - Electrical resistivity tomography obtained from inversion of apparent electrical resistivity values using RES2Dinv software.

3. Hydrogeophysical relationship

As described by Archie's (1942) Law and Purvance and Andricevic's (2000) Law, the electrical resistivity depends on some hydraulic parameters such as porosity, water content, and hydraulic conductivity.

According to Archie's Law, the resistivity of water-saturated, clay-free material can be described as:

$$\rho_r = a\rho_w F \tag{1}$$

where ρ_r is the electrical resistivity of the saturated clay-free aquifer (Ωm), ρ_w is electrical resistivity of water in the pores (Ωm), a depends on the tortuosity of interconnections between pores (usually $a = 1$) and F is the formation factor.

The formation factor F is a parameter linked to the mean porosity (φ), in fact:

$$F = \varphi^{-m} \tag{2}$$

where m represents the cementation degree that varies between 1.3 and 2.5 (Worthington, 1993; Schön, 1996).

If we consider a clay amount negligible when using Archie's Law and the Kozeny-Carman Law (Bear, 1972), it is possible to obtain the aquifer hydraulic conductivity (K). In fact K (m s^{-1}) will be:

$$K = \left(\frac{\rho_w g}{\mu_w}\right)\frac{d_{10}^2 F^{-3/m}}{180\left(1-F^{-1/m}\right)^2} \tag{3}$$

where ρ_w is water density (kg m^{-3}), μ_w is the water dynamic viscosity (Pa s), g is the acceleration due to gravity (m s^{-2}), d_{10} is the equivalent diameter where 10% of the particles' mass has a smaller diameter (mm), and F is the formation factor.

Starting from two empirical relationships that are Archie's Law and the Kozeny-type Law, Purvance and Andricevic (2000) described a log-log relationship between electrical (S cm^{-1}) and hydraulic conductivity (σ-K), in the form:

$$Y = A \pm BX \tag{4}$$

where $Y = \log(K)$, $X = \log(\sigma)$, and A and B are experimentally determined constants, which depend on the medium and fluid characteristics. In particular, materials characterized by a low clay content and/or high salinity have a positive correlation, while materials characterized by a

Table 2 - Average hydraulic conductivity for each depth, estimated using Archie's Law and Kozeny-Carman Law (Bear, 1972).

Depth (m)	Hydraulic conductivity (m s^{-1}) (main aquifer)
-23.00	6.23 x 10^{-6}
-25.00	1.26 x 10^{-6}
-27.00	5.72 x 10^{-6}
-29.00	6.41 x 10^{-6}
-31.00	3.00 x 10^{-6}
-33.00	4.81 x 10^{-6}
-35.00	3.28 x 10^{-6}

high clay content and/or low salinity have a negative correlation.

4. Results

After the inversion of apparent resistivity values, we passed to the study of their correlation with hydraulic parameters, using the law proposed by Archie (1942).

Since the clay content of the main aquifer of Montalto Uffugo is very low, the formation factor (F) has been estimated by the ratio between ground resistivity values (ρr), obtained from inversion, and measured water electrical resistivity ρ_w. The average value is $\rho_w = 14.98 \pm 0.55$ Ωm.

The F average values calculated showed the presence of two different zones: the first one (from -13 m to -21 m) is characterized by an average $F = 2$, while the second one (from -23 m to -35 m) is characterized by an average $F = 11$. Therefore, keeping $m = 1.5$ (low cementation degree), the porosity values for each depth have been calculated by applying Eq. (2). The average porosity values obtained are 0.59 for the main aquifer portion, which is included between -13 and -21 m, and 0.34 for the under portion; the average values for each depth are shown in Table 2.

When the F value equals 2, it pertains to the range of silt. In this range, Archie's Law and the Kozeny-Carman Law no longer hold, therefore we decided to analyse only the main aquifer portion between -23 to -35 m.

Subsequently, the porosity values obtained were used to calculate hydraulic conductivity (K), applying the Kozeny-Carman Eq. (3). In fact, considering $d_{10} = 0.03$ mm (the average value obtained from the granulometrical analysis of samples coming from the two new drilled piezometers PA and PB), the hydraulic conductivity was estimated starting from resistivity values of cross-hole tomography in the main aquifer portions included between -23 to -35 m; Fig. 4 shows the hydraulic conductivity tomography obtained.

The average hydraulic conductivity value obtained is $K = 2.19 \cdot 10^{-6}$ m s^{-1}; average values for each depth are shown in Table 2.

In addition, the methodology allows us to observe the hydraulic conductivity variability in the x direction too: estimated values are shown in Table 3.

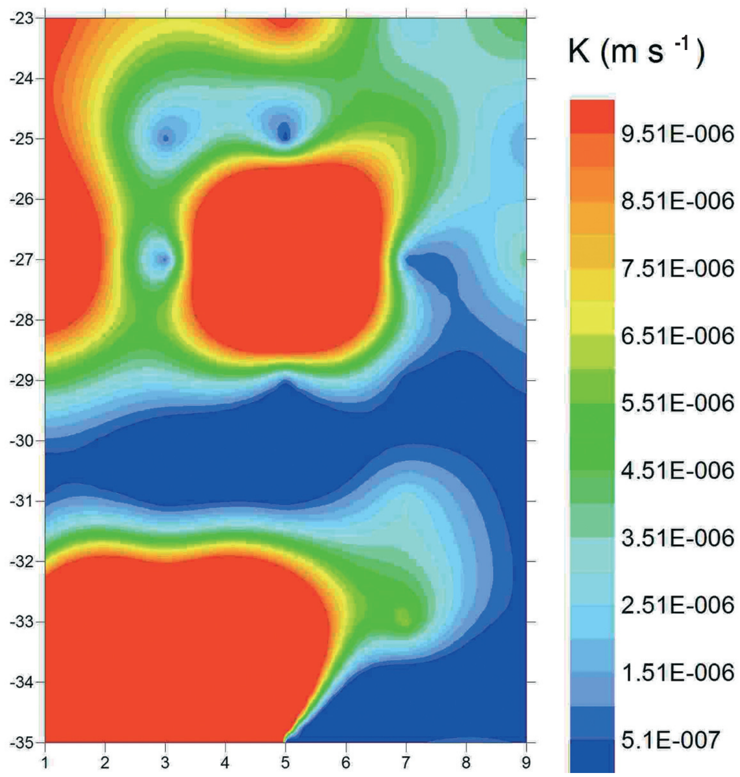

Fig. 4 - Hydraulic conductivity tomography of the aquifer portion between -23 to -35 m, obtained using Archie's Law and the Kozeny-Carman Law.

To obtain the hydraulic conductivity distribution over the aquifer, using Purvance and Andricevic's method, the linear correlation was exploited between the conductivity values (σ) extracted from the inverted resistivity data, and hydraulic conductivity data (K) measured by Troisi *et al*. (2000) in the piezometer 5 (Table 1).

Therefore, using the logarithmic correlation method giving a straight line of the type $Y = A + B * X$, proposed by Purvance and Andricevic's method (2000), we estimate $A = -11.03$,

Table 3 - Average hydraulic conductivity distribution along x direction, estimated in the aquifer portion between -23 to -35 m, using Archie's Law and Kozeny-Carman Law (Bear, 1972).

x (m)	Hydraulic conductivity (m s^{-1}) (main aquifer)
1.00	6.81×10^{-6}
3.00	7.74×10^{-6}
5.00	1.84×10^{-6}
7.00	1.44×10^{-6}
9.00	8.06×10^{-6}

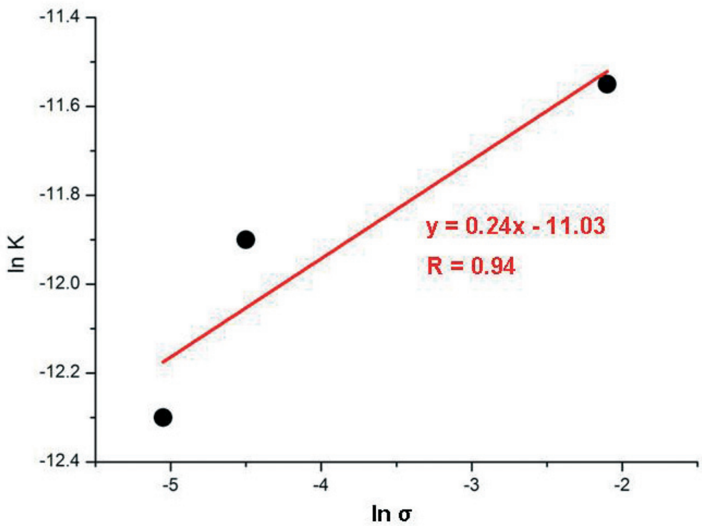

Fig. 5 - σ-K correlation between electrical resistivity values and hydraulic conductivity measured by Troisi *et al*. (2000). The red line is the straight line correlation for all data.

B = 0.24, and a correlation factor R = 0.94 (Fig. 5).

From a first analysis, we can observe that the ρ-K correlation is positive: this condition is typical of predominantly sandy materials, as the authors described in their paper.

Even though the linear correlation purposed is not consistent, because it has been achieved using a small number of points, we decided to use the calculated A and B coefficient, to obtain the hydraulic conductivity distribution starting from the electrical conductivity values. The hydraulic conductivity tomography obtained is shown in Fig. 6.

Analyzing the hydraulic conductivity tomography obtained, we can see that values are included between $1.43 \cdot 10^{-6}$ and $9.36 \cdot 10^{-6}$ m s^{-1}, with an average value of $6.60 \cdot 10^{-6}$ m s^{-1}. In particular, the hydraulic conductivity tomography obtained can be divided into two portions: the upper part (from -19 to -40 m) is characterized by an average $K = 5.10 \cdot 10^{-6}$ m s^{-1} and the lower part (from -40 to -52 m) is characterized by an average $K = 8.10 \cdot 10^{-6}$ m s^{-1}.

5. Discussions and conclusion

The cross-borehole resistivity measurements allowed us to obtain the resistivity distribution in the subsoil and compare it with the geological-stratigraphic information obtained by the analysis

Table 4 - Comparison between estimated average values of hydraulic conductivity.

Depth (m)	Hydraulic conductivity (m s^{-1}) (main aquifer)		
	Troisi et al. (2000)	**Archie**	**Purvance and Andricevic**
-23<z<-35	6.42×10^{-6}	2.19×10^{-6}	6.60×10^{-6}

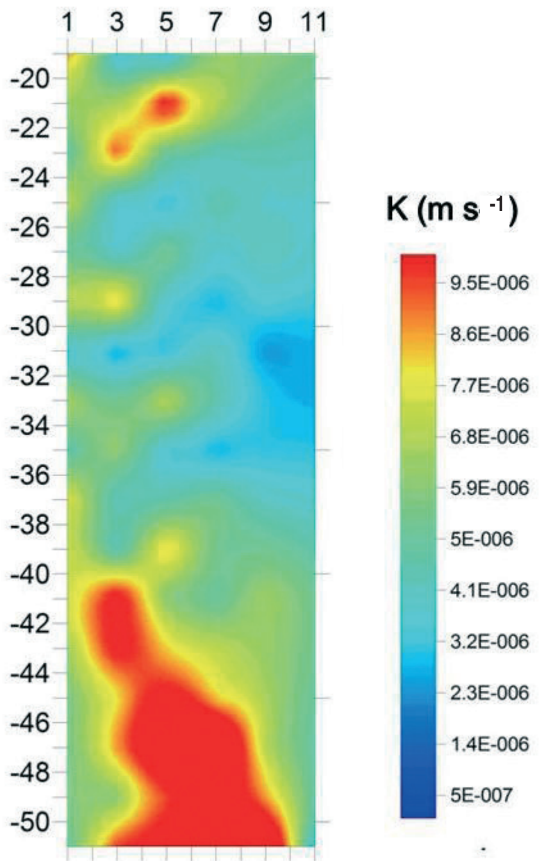

Fig. 6 - Hydraulic conductivity tomography of the aquifer, obtained using Purvance and Andricevic's Law.

of the two cores which derive from the new drilled piezometers. Then, data acquisition (ERT "cross-hole") and their elaboration led to the characterization of the main aquifer of Montalto Uffugo and the reconstruction of the hydraulic conductivity distribution in the subsoil by applying Eqs. (3) and (4). The magnitude of an estimated average hydraulic conductivity compares favourably with values previously determined in situ by some hydraulic tests (Troisi *et al.*, 2000).

Moreover, the adopted methodology allowed us to reconstruct the average spatial distribution of aquifer hydraulic conductivity with greater detail rather than a single measurement point.

However, the methodologies used have clear limitations. Niether Archie's Law nor Kozeny-Carman's Law can be applied to the entire aquifer because they no longer hold, in the ranges of silt and clay. In addition, both relationships depend on some parameters that describe porous medium geometry and that are not easy to determine.

Instead, the results obtained using Purvance and Andricevic's method, strongly depend on the availability of hydraulic conductivity data of the aquifer. In the case of Montalto Uffugo's main aquifer, we have only three values reported by Troisi *et al.* (2000); therefore, estimated coefficient A and B may not be valid.

Therefore, for further verification of the results, it might be useful to apply the same methodology, proposed in this article to a series of samples taken from two new cores. In this way, it can be possible to also examine the scale effect that determines the estimation of hydraulic conductivity.

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