Estimation of aquifer hydraulic parameters from surficial geophysical methods: A case study of Keritis Basin in Chania (Crete – Greece)

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KEYWORDS
Geophysical methods; Groundwater management; Hydraulic properties; Hydraulic testing; Formation factor

Summary
Knowledge of aquifer parameters is essential for the management of groundwater resources. Conventionally, these parameters are estimated through pumping tests carried out on water wells. Few boreholes may be available and carrying out pumping tests at a number of sites may be costly and time consuming. The application of geophysical methods in combination with pumping tests provides a cost-effective and efficient alternative to estimate aquifer parameters. A geophysical method is used to obtain aquifer characteristics that are estimated through the pumping tests. A correlation is established between these parameters at other sites where pumping has not been carried out. In this way, the entire investigation area could be covered to characterize an aquifer system. This study has been carried out in the Keritis basin in Chania, Crete – Greece, where the aquifer characteristics are required for the management of groundwater in the region. © 2007 Elsevier B.V. All rights reserved.

Introduction
During the last decades, the incremental need of accurate global groundwater resource assessment has led to a rapidly growing awareness in the field of groundwater management. Therefore, quantitative description of aquifers has

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become vital in order to address several hydrological and hydrogeological problems. Fluid transmissivity, transverse resistance, longitudinal conductance, hydraulic conductivity and aquifer depth are fundamental properties describing subsurface hydrology. As a result, many investigation techniques are commonly employed with the aim of the estimation of spatial distribution of the above mentioned hydraulic parameters. Field estimations of the above parameters are not always available. Hydraulic conductivity appears to be the most problematic to obtain because of either the great range of observed values or the unsatisfactory laboratory measurements (Zecharias and Brutsaert, 1988; Mendosa et al., 2003).

Traditionally, one of the most effective ways of hydraulic conductivity calculation are the pumping tests that are carried out on certain boreholes sites. Nevertheless, a probable sparse spatial distribution of the available boreholes gives rise to significant problems in modeling the hydrogeological systems. In such cases, drilling new boreholes has proved to be rather expensive as well as time consuming.

Geophysicists have realised that the integration of aquifer parameters calculated from the existed boreholes locations and surface resistivity parameters extracted from surface resistivity measurements can be highly effective, since a correlation between hydraulic and electrical aquifer properties can be possible, as both properties are related to the pore space structure and heterogeneity (Kelly, 1977; Mazac et al., 1985; Huntley, 1986; Mazac et al., 1988; Boerner et al., 1996; Christensen and Sorensen, 1998; Rubin and Hubbard, 2005; De Lima et al., 2005; Niwas et al., 2006).

Specifically, resistivity techniques are well-established and widely used to solve a variety of geotechnical, geological and environmental subsurface detection problems (Ward, 1990). The primary purpose of the resistivity method is to measure the potential differences on the surface due to the current flow within the ground. Since the mechanisms which control the fluid flow and electric current and conduction are generally governed by the same physical parameters and lithological attributes, the hydraulic and electric conductivities are dependent on each other. Of course, we should note that the factors which govern the current flow and conduction into the soil (lithology, size, shape, mineralogy, packing and orientation of grains, shape and geometry of pores and pore channels, magnitudes of porosity, tortuosity and permeability, compaction, consolidation and cementation and depth and water distribution) (Salem, 1999) are extremely variable. Thus, it should be born in mind that the measured resistivity values are not absolute but relative, and therefore only relative conclusions about the area’s hydraulic parameters could be made.

For this purpose, surface geophysical methods have been used for aquifer zone delineation and evaluation of the geographical character of the aquifer zone in several locations (Ungemach et al., 1969; Kelly, 1977; Heigold et al., 1979; Kosinki and Kelly, 1981; Niwas and Singhal, 1981, 1985; Frohlich and Kelly, 1985; Huntley, 1986; Shakeel et al., 1988; Worthington, 1993; Frohlich et al., 1996; Yadav and Abolfazli, 1998; De Lima and Niwas, 2000; Salem, 1999, 2001a,b; Hubbard and Rubin, 2002; Niwas and de Lima, 2003; Dhakate and Singh, 2005; Khalil, 2006).}

The aim of our study is to demonstrate the use of aquifer parameters in the assessment and management of the shallow (sedimentary) groundwater resources of Keritis Basin, which is located in the eastern part of Chania Municipality, Crete Island, Greece. Keritis Basin is the main source of irrigation water for the whole plain land of Chania Basin, one of the most developed agricultural areas in Crete. The inadequate number of boreholes has been overcome using the correlation between geophysical parameters extracted from surface electrical measurements interpretation and aquifer hydraulic parameters extracted from pumping tests carried out on the existed boreholes. This approach has been utilized in order to estimate aquifer parameters in numerous locations providing effective and inexpensive characterization of the study area aquifer system. The calculated aquifer parameters have been interpolated using the radial basis function (RBF) interpolator. Compared with the rest of interpolation methods, RBF gives the best cross-validation results and root-mean-squared (RMS) errors for the available data.

**Study area**

Crete is considered to be a semi-arid region. The average annual precipitation is estimated to be 900 mm, the potential renewable water resources 2650 m³/yr and the real water used about 485 million m³/yr (Chartzoulakis et al., 2001). The major water use in Crete is in irrigation for agriculture (84.5% of the total consumption), while domestic use is 12% and other uses 3.5% (Chartzoulakis et al., 2001; Tsagarakis et al., 2004).

The study area is situated from 35°24’00”N to 35°30’00”N, and 23°49’50”E to 23°58’00”E (Fig. 1). The total county area is 137 km² and is located in the central part of Keritis river drainage basin, 3.5 km west of the city of Chania. The central study area is characterized by a rather gentle topography. It is bordered by the villages Vryses and Galatas to the north, Skines and Fournes to the south, Psathogiannos to the west and Varypetron to the east (Fig. 1). The area is drained by the Keritis river which is considered to be the main river of the area.

The climate of the study area is sub-humid Mediterranean with humid and relatively cold winters and dry and warm summers. During winter that starts in November, the weather is unstable due to frequent changes from low to high pressures.

The annual rainfall for the broader Chania area has been estimated to be 665 mm (Chartzoulakis et al., 2001). About 65% of the annual precipitation is lost to evapotranspiration, 21% as runoff to sea and only 14% recharges the groundwater (Chartzoulakis et al., 2001). The rainfall is mainly concentrated in the winter months while the drought period extends to more than 6 months (May–October). The monthly evaporation ranges from 140 mm to more than 310 mm in the peak month. As a result, the water resources availability is limited due to the spatio-temporal variations of precipitation (Tsagarakis et al., 2004). The demand for irrigation water is high, while at the same time only 31.0% of the available agricultural land is irrigated (Tsagarakis et al., 2004). The growing water demands make the water resources management extremely important for sustainable development.
Geo-tectonic and hydrolithological context

The surficial geology is composed of Quaternary deposits that form depositional plains oriented from north to south at an elevation of 20–200 m above msl. Miocene to Pliocene sediments crop out in the central and the northwestern part of the study area and Late Triassic carbonates (Tripolis nappe) in the northeastern part. Dissected hills of phyllites and quartzites, a Late Carboniferous to Late Triassic package of sedimentary rocks composed mostly of quartz-rich siliciclastic sediments, with minor limestone, gypsum, and volcanic rocks (Krahl et al., 1983) are observed mainly in the southwestern and southeastern part of the study area at an elevation ranging from 200 to 580 m above msl. The Triassic to Early Cretaceous carbonates of the Trypalion nappe are exposed in the central-eastern part of the study area.

In the present work the geological units have been classified in the sense of permeability into four hydrolithological units (Fig. 2): high permeability rocks which comprise the karstic Triassic limestones of Tripolis and Trypalion nappes, medium permeability rocks which consist of the Quaternary deposits as well as the Miocene to Pliocene conglomerates and marly limestones, low permeability rocks which consists of the Pliocene to Miocene marl and impervious rocks which mainly consist of the phyllites–quartzites unit.

The local tectonic regime of the study area is characterized by faults of NW-SE and E-W directions, which define the boundaries between the existing hydrolithological units (Fig. 2) as well as the groundwater flow direction. Thus, these tectonic structures probably act as underground dams bounding the underground water movement.

Geoelectrical investigation

During the current work, twenty-one (21) of the total fifty-four (54) geoelectrical soundings (CH# and V#) (Fig. 2) with a maximum half current electrode separation of 500 m have been used. All the data (54 VES measurements) were collected in four phases of field work. The first data set was acquired by CGG Company (HydroGaia, 1972), the second data
set was collected by Technical University of Crete (Vafidis et al., 1991), the third data set was gathered by Apostolopoulos et al. (2001), and the last one was collected by the Technological Educational Institute of Crete in 2004. Twenty-one of the total 54 available VES measurements were selected as the most appropriate for the final modeling since they combine the low RMS error (best in quality) with the good coverage of the area of interest. The Schlumberger electrode array was selected because it is the most time effective in terms of field work.

The measurements were planned with the intention of covering the whole area while at the same time we wanted to be as close to the existing boreholes as possible in order to use them for calibration.

All resistivity soundings were inverted using the IPI2Win software. This package performs an automated approximation of initial resistivity model using the observed data (Bobachev, 2002). It works in an iterative mode by calculating at the end of each step: (a) an updated model of layer thickness and resistivity and (b) the misfit function between observed and calculated data. All resulting models produced a low RMS relative error of the order of 3%.

The starting model used during the inversion for each of the measured VES locations, consisted of four layers over a half-space and all depths were constrained again according to the nearest borehole information.

Table 1  Estimation of formation factor and other hydraulic parameters from the geophysical data obtained from locations CH-#, V#, L5 and BA-ND

<table>
<thead>
<tr>
<th>Location</th>
<th>Bulk resistivity (Ω-m)</th>
<th>Aquifer resistivity (Ω-m)</th>
<th>Aquifer thickness (m)</th>
<th>Formation factor (Fa)</th>
<th>1/Fa</th>
<th>Longitudinal unit conductance (S)</th>
<th>Transverse resistance (TR)</th>
<th>Transmissivity (m²/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH-1</td>
<td>122</td>
<td>30</td>
<td>50.2</td>
<td>4.0667</td>
<td>0.2459</td>
<td>1.673</td>
<td>1506</td>
<td>1.34E-02</td>
</tr>
<tr>
<td>CH-3</td>
<td>74</td>
<td>8.5</td>
<td>48</td>
<td>8.7059</td>
<td>0.1149</td>
<td>5.647</td>
<td>408</td>
<td>1.28E-02</td>
</tr>
<tr>
<td>CH-4</td>
<td>60.4</td>
<td>7.9</td>
<td>27.6</td>
<td>7.6458</td>
<td>0.1308</td>
<td>3.494</td>
<td>218.04</td>
<td>7.37E-02</td>
</tr>
<tr>
<td>CH-5</td>
<td>60</td>
<td>7.4</td>
<td>115.6</td>
<td>8.1081</td>
<td>0.1233</td>
<td>15.622</td>
<td>855.44</td>
<td>3.09E-02</td>
</tr>
<tr>
<td>CH-6</td>
<td>72.2</td>
<td>7.6</td>
<td>79.8</td>
<td>9.5000</td>
<td>0.1053</td>
<td>10.500</td>
<td>606.48</td>
<td>2.13E-02</td>
</tr>
<tr>
<td>CH-7</td>
<td>75</td>
<td>15.35</td>
<td>48.3</td>
<td>4.8860</td>
<td>0.2047</td>
<td>3.147</td>
<td>741.405</td>
<td>1.29E-02</td>
</tr>
<tr>
<td>CH-8</td>
<td>82.4</td>
<td>18.3</td>
<td>88.2</td>
<td>4.5027</td>
<td>0.2221</td>
<td>4.820</td>
<td>1614.06</td>
<td>2.35E-02</td>
</tr>
<tr>
<td>CH-9</td>
<td>185</td>
<td>22.72</td>
<td>61.3</td>
<td>8.1426</td>
<td>0.1228</td>
<td>2.698</td>
<td>1392.736</td>
<td>1.64E-02</td>
</tr>
<tr>
<td>CH-10</td>
<td>131</td>
<td>17.15</td>
<td>113</td>
<td>7.6385</td>
<td>0.1309</td>
<td>6.589</td>
<td>1937.95</td>
<td>3.02E-02</td>
</tr>
<tr>
<td>CH-11</td>
<td>240</td>
<td>26.31</td>
<td>111.2</td>
<td>9.1220</td>
<td>0.1096</td>
<td>4.227</td>
<td>2925.672</td>
<td>2.97E-02</td>
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<tr>
<td>CH-12</td>
<td>126</td>
<td>28.38</td>
<td>90</td>
<td>4.4997</td>
<td>0.2252</td>
<td>3.171</td>
<td>2554.2</td>
<td>2.40E-02</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>porosity</th>
<th>Low bound</th>
<th>Upper bound</th>
<th>Average</th>
<th>Hydraulic conductivity (m/s)</th>
<th>Hydraulic conductivity (m/day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>1.04</td>
<td>0.406</td>
<td>0.368</td>
<td>0.346</td>
<td>0.373</td>
</tr>
<tr>
<td></td>
<td>0.5</td>
<td>0.287</td>
<td>0.269</td>
<td>0.253</td>
<td>0.273</td>
</tr>
<tr>
<td></td>
<td>1.1</td>
<td>0.286</td>
<td>0.248</td>
<td>0.227</td>
<td>0.254</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>0.197</td>
<td>0.165</td>
<td>0.148</td>
<td>0.170</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>0.245</td>
<td>0.210</td>
<td>0.191</td>
<td>0.215</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>0.289</td>
<td>0.252</td>
<td>0.232</td>
<td>0.258</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>0.329</td>
<td>0.292</td>
<td>0.267</td>
<td>0.297</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>0.366</td>
<td>0.328</td>
<td>0.307</td>
<td>0.333</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>0.399</td>
<td>0.361</td>
<td>0.340</td>
<td>0.367</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>0.430</td>
<td>0.392</td>
<td>0.371</td>
<td>0.397</td>
</tr>
<tr>
<td>diameter d2</td>
<td>0.01 (0.0001)</td>
<td>0.324</td>
<td>0.289</td>
<td>0.269</td>
<td>2.67E-04</td>
</tr>
</tbody>
</table>

Sensitivity analysis on porosity and hydraulic conductivity estimates using different Archie’s coefficients (grain size = 0.01 m).
Concerning our work, since bulk and water resistivities were available at several locations, it was desirable to examine the possibility to obtain hydraulic conductivity values using the Kozeny–Carman–Bear equation (Domenico and Schwartz, 1990). The porosity ($\phi$) required in this equation was estimated using Archie’s law, with its empirical nature and dependence on Archie’s parameters. Moreover, we calculated the range of hydraulic conductivities produced by the plausible values of these porosities.

Archie’s law (Archie, 1942) relates the bulk resistivity of a fully saturated granular medium to its porosity and the resistivity of the fluid within the pores according to Eq. (1):

$$\rho_o = \frac{\phi}{C_1} = \frac{1}{\frac{\rho_w}{C_1}} = \frac{\phi}{F_q \cdot \phi^{-m}}$$

where $\rho_o$ is the bulk resistivity, $\rho_w$ is the fluid resistivity, $\phi$ is the porosity of the medium, $C_1$ is known as the cementation factor, although it is also interpreted as grain-shape or pore-shape factor, and the coefficient $\alpha$ is associated with the medium and its value in many cases departs from the commonly assumed value of one.

For a clay-free medium, the $\rho_o/\rho_w$ ratio is known as the intrinsic formation factor, $F_i$. Thus, Eq. (1) could be easily reformulated in the following form, Eq. (2),

$$\phi = e^{\ln\phi - \ln\left(\frac{\rho_o}{C_1}\right)}$$

(2)

The values of the coefficients $a$ and $m$ should, ideally, be determined for each site under investigation. However, due to lack of core samples in our test area, this was not possible and an alternative approach was adopted whereby a wide range of values for $a$ and $m$ reported in the literature was used to obtain porosity estimations. Worthington (1993) reported three different expressions for the intrinsic formation factor in relation to the porosity of samples from different locations. A fourth expression in which the coefficient $a$ has the value of one while $m$ is allowed to vary from 1.3 to 2.5 was suggested by Jackson et al. (1978) and De Lima and Sharma (1990).

However, for field data a complication arises due to the fact that Archie’s formula (Eqs. (1) and (2)) is valid only for clay-free, clean, consolidated sediments. Any deviations from these assumptions make the equation invalid as discussed by Worthington (1993). In the case of unclean, clayey and shaley sands and a mixture of sand/rubble/gravels, an additional corrective step for clay conductivity is required. A large number of such models are currently in use and the majority of them are either shale-fraction or cation-exchange models, derived empirically using the concept of parallel conductor (Patnode and Wyllie, 1950; Winsauer and McCardell, 1953; Waxman and Smits, 1968; Clavier et al., 1984; Sen et al., 1988).

As our aquifer consists of clayey/silty sand material enhanced with rubbles and gravels, a modification of the Archie’s equation was required. For this reason, the Waxman–Smits model was considered (Vinegar and Waxman, 1984) as it relates the apparent and intrinsic formation factors, $F_a$ (the ratio of bulk resistivity to fluid resistivity) and $F_i$, after taking into account the shale effects. According to Worthington (1993),

$$F_a = F_i \cdot (1 + BQ_o \rho_w)^{-1}$$

(3)

where the $BQ_o$ term is related to the effects of surface conduction, mainly due to clay particles. In case surface conduction effects are non-existent, the apparent formation factor becomes equal to the intrinsic one.

Re-arranging the terms of Eq. (3), we obtained a linear relationship between $1/F_a$ and $\rho_w$,

$$\frac{1}{F_a} = \frac{1}{F_i} + \left(\frac{BQ_o}{F_i}\right)\rho_w$$

(4)

where $1/F_i$ is the intercept of the straight line and $BQ_o/F_i$ represents the gradient (Huntley, 1986; Worthington, 1993). Thus, by plotting $1/F_a$ versus fluid resistivity $\rho_w$, we should, in principle, obtain a value for the intrinsic formation factor, which will subsequently enable us to estimate porosity using Eq. (2) as is shown in Table 1.

To follow the above approach, we used bulk resistivities $\rho_o$, as resulted from 1D resistivity inversion coupled with the measured fluid electrical resistivities, $\rho_w$, obtained using the wells nearest to the VES locations. These values were used to calculate the apparent formation factor ($F_a = \rho_o/\rho_w$) of the saturated top aquifer. From Eq. (4), it is clear that a major source of error would be the wrong estimation of the apparent formation factor which depends on the bulk resistivity as estimated from the inversion models. Thus, assuming that the fluid resistivities were measured correctly in situ, we considered by how much the resistivities resulting from the inversion could vary. This uncertainty was mainly due to the fact that some of the boreholes (sample locations) were at some distance away from the region represented in the resistivity model.

The resistivity values obtained from the inversions and the estimated apparent formation factors are shown in Table 1.

Fig. 3 shows $1/F_a$ plotted versus fluid resistivity, $\rho_w$. The data could be easily separated in two individual groups (depending on the geological complexity of the study area) with common characteristics as is shown by the two fitted
lines. This is obtained by applying least squares best fit of the individual subgroups of the data, and the range of the inverse of the intrinsic formation factor $F_i$ is calculated. For our data set, the $F_i$ varies between 8.27 and 11.96 as is estimated from Fig. 3. The porosities could now be determined through Eq. (2) for the reported values of a and m as shown in Table 1 applying an upper and a lower bound for the determined porosity based on the division of the data into the groups as mentioned above.

The hydraulic conductivity estimation was achieved through the use of the Kozeny—Carman—Bear equation, given by Domenico and Schwartz (1990) as:

$$k = \left( \frac{d \phi^3}{180} \right) \left( 1 - \phi \right)^2$$

where $d$ is the grain size, $\phi$ is the fluid density (taken to be 1000 kg/m$^3$), and $\mu$ is the dynamic viscosity taken to be 0.0014 kg/m s (Fetter, 1994). The estimated hydraulic conductivity values (in m/s and in m/day) using Eq. (5) are shown in Table 1.

The average geometrical hydraulic conductivity value for the aquifer under investigation is $2.67 \times 10^{-4}$ m/s or 23.1 m/day.

Two of the most important parameters in electrical prospecting are the longitudinal unit conductance ($S$, layer thickness times resistivity) and the transverse unit resistance ($TR$, layer thickness times times resistivity), which define the Dar Zarrouk Parameters (DZP) (Maillot, 1947). The DZP were calculated for interpreted sounding layer parameters after taking into account only aquifer resistivities and its thicknesses:

$$S = \frac{h}{r} \quad \text{and} \quad TR = r \cdot h$$

where $h$ is the thickness of the aquifer (m) and $r$ is the resistivity of the aquifer ($\Omega$-m).

The transmissivity (ability of the layer with permeability $k$ to transmit fluids through its entire thickness $h$, see Eq. (7)), using the estimated hydraulic conductivity ($k = 2.67 \times 10^{-4}$ m/s), was also calculated using Eq. (7),

$$T = k \cdot h$$

where $T$ is the transmissivity (m$^3$/s), $k$ is the hydraulic conductivity (m/s) and $h$ is the aquifer thickness (m). The extracted parameters for each VES location are shown in Table 1. The transverse resistance ($\Omega$-m$^2$) is directly correlated to the transmissivity (m$^2$/day$^{-1}$) as is shown in Fig. 4. The progressive increase in both parameters suggests that the fluid potential (indicated by transmissivity) of the aquifer in the study area, increases considerably as the transverse resistance increases. The aforementioned simultaneous change is attributed to the influence of the hydraulic and electric anisotropies, as well as to the variations in lithology, mineralogy of the grains and size and shape of the pores and pore channels (Salem, 1999).

Correlation of the observed and estimated hydraulic parameters

The most common in situ test for the calculation of real water supply and the indirect estimation of hydraulic conductivity in a borehole is the pumping test performed on wells, and involves the measurement of the rise and fall of water level with respect to time. The water level fluctuations with time, are then interpreted to arrive at aquifer parameters. The availability of pre-existing wells makes the pumping test cost-effective.

More than 20 water wells were operated throughout the study area and one of them was selected as the most representative for the indirect estimation of the hydraulic conductivity through the use of pumping test (both pumping and recovery). The available information, such as, estimated water supply (m$^3$/h), pumping water level (m), and relative lithology were collected from the registration forms of the available water wells in the study area.

Using the estimated water supply ($Q = 100$ m$^3$/h), the diameter ($d$) of the screen, as well as the thickness ($h$) of the aquifer, the hydraulic conductivity was estimated using the following formula:

$$Q (\text{m}^3/\text{h}) = \pi \cdot d (\text{m}) \cdot h (\text{m}) \cdot k (\text{m}/\text{h})$$

Applying the above equation, the hydraulic conductivity $k$, is estimated equal to $6.8 \times 10^{-4}$ m/s (coarse grained sand). This indirect observed value of hydraulic conductivity, is in agreement with the estimated aquifer parameter ($k = 2.67 \times 10^{-4}$ m/s, medium grained sand) as calculated from the surficial geophysical measurements. Both values are within the range of $10^{-5} - 10^{-2}$ m/s which is the characteristic range of a pure sands and gravel aquifer (Kallergis, 1999).

Results and discussion

The calculated hydraulic parameters were interpolated with geostatistical as well as deterministic techniques. After
applying and testing several geospatial prediction techniques, the radial basis function (RBF) interpolator was chosen (spline with tension) for interpolating aquifer thickness, transverse resistance as well as transmissivity values. Compared with the rest of interpolation methods, RBF gave the best cross-validation results and root-mean-squared (RMS) errors (28.74 m for aquifer thickness, 473.6 for transverse resistance and 325 m²/day for transmissivity). RMS is given by the equation:

\[ \text{RMS} = \sqrt{\frac{1}{n} \sum_{i=1}^{n} (D_{ip} - D_{io})^2}, \]

where \( D_{ip} \) is the predicted value for site \( i \), \( D_{io} \) is the observed value for site \( i \), and \( n \) the total number of sites.

Cross-validation (CV) uses all of the data to estimate the trend. Then it removes each data location, one at a time, and predicts the associated data value. The predicted and actual values at the location of the omitted point are compared. For all points, cross-validation compares the measured and predicted values. Generally, the best model is the one that has the mean nearest to zero and the smallest root-mean-squared prediction error. If the prediction errors are unbiased, the mean prediction error should be near zero. However, this value depends on the scale of the data (2.254 m for the aquifer thickness, 76.97 for transverse resistance and 27.44 m²/day for transmissivity). In addition, the use of Kriging interpolators led to a great number of artifacts, possibly because of the limited number as well as the random distribution of the available sample sites.

The radial basis function (RBF) is commonly used for interpolating small surfaces and has been found to be appropriate for interpolating electrical conductivity values (Robinson and Metternicht, 2006). For each point, an RBF is defined, which depends on the Euclidean distance between the prediction location and each sample location. The "roughness" of the surface is reduced by the action of forming a polynomial function through the data points.

The spline with tension method controls the stiffness of the surface while it creates a less-smooth surface with values more closely constrained by the sample data range. Due to the small amount of data, the produced maps (Figs. 5–7) still have significant errors. More data will lead to better refinement of the mapped parameters.

Using the above mentioned techniques, the aquifer thickness (m), transverse resistance and transmissivity (m² per day) thematic maps were created as is shown in Figs. 5–7, respectively. Moreover, uncertainty analysis was applied to demonstrate the cross validation of the created maps as is shown in the Table 2.

The spatial maps that were derived from the interpolation of the calculated hydraulic parameters show increased aquifer thickness as we move towards rocks of high permeability, which represent karstic limestone formations.

**Figure 5** Shaded relief of the Keritis basin area with aquifer thickness map and tectonic faults overlain. A spatial correlation between aquifer thickness and hydrogeological units is obvious, as aquifer thickness tends to decrease with the transition from permeable to impervious rocks.

**Figure 6** Transverse resistance (TR) map for the Keritis Basin area with tectonic faults overlay. The TR tends to increase from the northwestern (impervious rocks) to the southeastern part (rocks of high permeability, i.e., karstic limestones) of the study area as is shown in this figure. The topography is also shown by the shaded relief.
Aquifer transverse resistance and transmissivity also seem to be spatially related with the high permeability formations mentioned above (Figs. 6 and 7). Moreover, the increased transmissivity in the south-east part of the study area identifies the zones of high potential within the water-bearing formations. Furthermore, a relation between the estimated and presented hydraulic parameters as shown in Figs. 5–7 and tectonic structures seems to exist with faults acting as boundaries even between the same hydrological unit, and define the place where aquifer parameters varies.

Conclusions

Since drilling of wells to determine hydraulic parameters is often prohibitively expensive, determining the aquifer parameters from VES is a cost-effective alternative. Based on our results, the contribution of VES coupled with the available pumping test data proved to be significant to the quantitative estimation of aquifer parameters.

The estimated transmissivity of the geological formations in the study area shows a wide range due to the high inhomogeneity of the sedimentary formations and the complexity of the karstic formations.

Based on the calculated hydraulic parameters of the shallow aquifer in Keritis basin, several hydrogeological maps have been created, which are very useful for further studies of the groundwater regime in the study area. The maps could also be used to derive input parameters for contaminant migration modeling and to improve the quality of model. Finally, we should mention that the calculated aquifer parameters are well defined within the range of observed aquifer parameters.

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<th>Table 2 Cross-validation results of the calculated hydraulic parameters</th>
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Figure 7 Shaded relief of the Keritis basin area with transmissivity (m²/day) map and tectonic faults overlain. The range of transmissivity values shows a pattern similar with that of aquifer thickness (Fig. 5).
References


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