An Inverse Procedure to Estimate Transmissivity from Heads and SP Signals

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Abstract

Experimental hydraulic heads and electrical (self-potential) signals associated with a pumping test were used in an inverse model to estimate the transmissivity distribution of a real aquifer. Several works reported in the literature show that there is a relatively good linear relationship between the hydraulic heads in the aquifer and electrical signals measured at the ground surface. In this experimental test field, first, the current coupling coefficient was determined by the best fit between experimental and modeled self-potential signals at the end of the pumping phase. Soon afterward, with the hydraulic heads obtained from the self-potential signals, the transmissivity distribution of the aquifer was conditioned by means an inverse model based on the successive linear estimator (SLE). To further substantiate the estimated T field from the SLE analysis, we analyzed the drawdown rate, the derivative of the drawdown with respect to the ln(t), because the drawdown rate is highly sensitive to the variability in the transmissivity field. In our opinion, these results show that self-potential signals allow the monitoring of subsurface flow in the course of pumping experiments, and that electrical potentials serve as a good complement to piezometric observations to condition and characterize the transmissivity distribution of an aquifer.

Introduction

Characterizing 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. Over the past few decades, the characterization of aquifers has relied on traditional aquifer test methods (i.e., cross-hole tests: pumping at one well and observing the response at another well, and the use of Theis’ or Jacob’s analysis) or slug tests. Traditional aquifer tests are thought to yield averaged hydraulic properties over a large volume of geologic media (Butler and Liu 1993). In reality, the classical analysis for aquifer tests yields spurious average transmissivity values that are difficult to interpret (Wu et al. 2005). On the other hand, they yield storage coefficient estimates that reflect the local geology between the pumping and the observation well. Similarly, Beckie and Harvey (2002) have questioned the validity of storage coefficients estimated from slug tests. In spite of these controversies, high-density measurements of the hydraulic properties of an aquifer using traditional aquifer test methods over a large basin are deemed cost-prohibitive and impractical. Generally speaking, measurements of water level responses (i.e., well hydrographs) of an aquifer are less costly and relatively abundant. Making use of the well hydrographs to characterize the spatial distribution of hydraulic properties of an aquifer is therefore rational. This approach is known as inverse modeling in subsurface hydrology. Nevertheless, without sufficient data to meet necessary and sufficient conditions of the inverse problem, the problem can be ill posed, and its solution will be nonunique. In this context, the interest of the hydrologist is to either increase the number of aquifer responses or improve geostatistical techniques to solve the inverse problem.
Two-dimensional (2D) hydraulic tomography can yield many useful sets of secondary information, namely head responses, which can be used to identify the heterogeneity of an aquifer. However, a reliable and efficient methodology is still required to decipher the information so that a reliable image of the transmissivity or the hydraulic conductivity field can be obtained. Yeh et al. (1996) and Zhang and Yeh (1997) developed an iterative geostatistical technique in which a linear estimator was used successively to incorporate the nonlinear relationship between hydraulic properties and pressure head. This method is referred to as a successive linear estimator (SLE). They demonstrated that with the same amount of information, the SLE revealed a more detailed conductivity field than cokriging. Hughson and Yeh (2000) showed that the SLE method is computationally efficient compared to the classical inverse method.

Several aquifer tests have been conducted in the Montalto wellfield from the early 1990s until today (Troisi and Straface 1996; Fallico et al. 2002; Straface et al. 2005). The aim of this study is the estimation of the spatial distribution of transmissivity by applying a geostatistical type approach based on the SLE to the self-potential (SP) signals measurements. This purpose has been achieved thanks to a new methodology able to determine the distribution of the hydraulic head in the aquifer around the pumping well by means of the SP signals measurements, and the application of an efficient geostatistical approach (SLE) able to condition the aquifer transmissivity distribution with the hydraulic heads obtained with the SP signals. SP signals are electrical potentials associated with ground water flow through the so-called electrokinetic coupling. In fact, in any natural porous material, there is an excess of electrical charge in the pore water due to electrochemical reactions at the surface of the minerals (Leroy and Revil 2004). The drag of this excess of electrical charge contained by ground water flow is responsible for a polarization of charge through the whole medium. The idea is naturally to use this electrical information to characterize the geometry of ground water flow. The application yielded a spatially varying transmissivity field. We then checked the validity of the estimated $T$ and $S$ fields using both a geological and a time drawdown derivative analysis.

Geological Setting and Description of the Experiment

The test site is located near the town of Montalto Uffugo, in the region of Calabria in southern Italy. The precise location of the field is given by Troisi et al. (2000) and will not be repeated here. The geology of the test site can be divided into four lithological units (Figure 1). Heterogeneous gravels in a silty-sand matrix (unit A) compose the first formation. This unit extends from the ground surface to a depth of about 7 m. The second unit is a shale layer (unit B in Figure 1, from ~7 to ~11 m). The third unit bears the main aquifer investigated in this article and is composed of a silty-sand layer (layer C, from 11 to 55 m). The deepest unit is the shale substratum (unit D). A shallow perched aquifer is sometimes present during part of the year in unit A. This was the case at the time of our experiment, but the drawdown is below the shale layer. Ten boreholes (two boreholes per monitoring location) were drilled at this test site. They are numbered P1 to P10 and all have a metallic casing. The locations of the boreholes are shown in Figure 1. Each pair includes a borehole reaching to a depth of 8 m (i.e., reaching the shallow perched ground water) and a second borehole reaching to a depth of 40 m and therefore connected to the aquifer of interest. Moreover, an additional borehole, coded P11, is located 19 m from the central well P5 and reaches the impervious bottom of the main aquifer. The diameter of the pumping well is 20 cm. At this point, the bottom of the aquifer was found at around 55 m, from which we may consider a thickness of around 44 m for the aquifer. In this study, we estimate spatial distribution of transmissivity by applying a geostatistical type approach based on the SLE to the SP signals measurements. This purpose has been achieved by means of a new methodology able to determine the distribution of the hydraulic head in the aquifer around the pumping well by means of the SP signals measurements and the application of an efficient geostatistical approach (SLE) able to condition the aquifer transmissivity distribution with the hydraulic heads obtained with the SP signals. The experiment consisted in an aquifer test in which the piezometric response was recorded in a set of monitoring piezometers (see location in Figure 1). We also measured the electrical response at the ground surface with a monitoring network of 53 nonpolarizable Pb/PbCl₂ Petiau electrodes connected to a multimeter and a reference electrode (Rizzo et al. 2004).

The experiment was performed in July 2003. Prior to the experiment, pressure sensors were placed in the...
monitoring piezometers (except in P5) to record the hydraulic head every 5 min during the pumping test experiment. The piezometric levels were also measured directly during the course of the experiment. The pumping rate $Q$ was 2.7 L/s in the central well, achieving a pseudo–steady state condition in about 4 d. After 4 d, the pump was shut off, but the hydraulic head measurements still continued during the recovery phase for another 7 d. When the pump was shut off, recording of the SP signals was also begun, and this monitoring was performed for about 12 h in order to measure the behavior of the SP signals without the pumping forcing (Rizzo et al. 2004).

The monitoring of the SP signals was carried out using 53 nonpolarizable Pb/PbCl$_2$ Petiau electrodes manufactured by SDEC in France (Perrier et al. 1997; Petiau 2000). The electrodes were located along two profiles crossing each other in the vicinity of the pumping well (Figure 2). To have uniform ground contact conditions for all electrodes, we dug small holes (~10 cm deep) filled with a salty bentonite mud and covered with several small stones in order to keep the moisture conditions high during the whole monitoring experiment (the day of the SP measurement was very hot). All electrodes were connected with a high internal impedance (>10 Mohm) multimeter (Keithley 2701). The multimeter was interfaced to a notebook computer to observe the data acquisition in real time. The data were acquired automatically with a time step of 1 min. Note that if the electrodes do not experience the same temperature, a small potential drift can occur owing to a thermoelectric effect. According to the supplier (SDEC), the temperature drift of the Petiau electrodes is 0.2 mV/°C. So a temperature difference of 5°C can be responsible for a SP drift of 1 mV, which is a substantial value when compared with the strength of the electrokinetic component associated with ground water flow and described subsequently (<10 mV). This explains why great care should be taken when placing the electrodes in the field, especially regarding their exposure to sunlight. The electrode readings can also be affected by drying of the surrounding soil and by precipitation. Covering each of the electrodes with small, inverted piece of Styrofoam™, weighted down by small stones, helps reduce these effects.

**Results**

To obtain the drawdown distribution over the aquifer, the linear correlation was exploited between the SP data measured close to the piezometers and the hydraulic heads measured in the piezometers (Rizzo et al. 2004). Afterward, these drawdown-time data for the pumping test were analyzed by two methods, which are based on either homogeneous or heterogeneous assumptions. First, assuming that the aquifer is homogeneous, we analyzed the well hydrographs of the aquifer tests using a drawdown-distance/time analysis; afterward, without invoking the aquifer homogeneity assumption, the data set was analyzed by the SLE algorithm developed for hydraulic tomography analysis to obtain spatially distributed $T$ and $S$ estimates of the field site.

Regarding the experimental setup shown in Figure 2, the pumping was conducted in well W5 with a pumping rate of 2.7 L/s during 4 d. The monitoring of the hydraulic heads in the main aquifer was performed on the five piezometers (one is 19 m from pumping well W5 and is not shown in Figure 2). All the wells were equipped with metallic casings. The SP signals were observed on the two lines where nonpolarizable Petiau (Pb/PbCl$_2$) electrodes were located. The reference electrode was located 25 m to the south of the pumping well and is not shown in Figure 2. The piezometric levels were observed during both the pumping and the relaxation phases, and the SP signals were monitored in the course of the relaxation phase within 12 h of pump shut-down. At the end of the experiment, the piezometric levels were considered nearly constant in the vicinity of the pumping well, and only the SP values obtained at 3 h were used as the temporal reference for each electrode. In addition to the SP monitoring, a resistivity tomography was performed along a profile, which allowed the distribution of the electrical conductivity of the formations to be assessed. In Rizzo et al. (2004), it was shown that there are two contributions to the SP signals. The first component is an the electrokinetic component associated with ground water flow. However, there is also a redox component associated with corrosion of the well metallic casing. The second contribution is 1 order of magnitude higher than the former but is stable over time. Consequently, by working with a reference time, the redox contribution can easily be removed. A similar approach was used by Naudet et al. (2004) to separate these two SP contributions over a contaminant plume in a shallow aquifer to determine the distribution of the redox potential. In Revil et al. (2002, 2003), it was shown how the SP is related to the depth of the water

![Figure 2. Position of the electrodes and wells during the pumping experiment. Only the central part of the test site is shown.](image-url)
table using the potential field theory. The transfer function between the SP data and the piezometric level depends on the intrinsic electrokinetic coupling coefficient, so a key question is to know if this coupling coefficient depends on the texture or the permeability of the porous materials. When the ground water is mineralized, so that the so-called surface at the grain/pore water interface conductivity (Leroy and Revil 2004) can be neglected, then the electrokinetic coupling coefficient is given by the Helmholtz-Schmoluchowski equation, which does not depend on the texture of the porous material. For the sand-gravel aquifer of Montalto Uffugo, surface conductivity around the grains can be safely neglected because the grains are coarse; therefore, our assumption to consider the electrokinetic coupling coefficient as constant is valid. In this condition, the transfer function between the hydraulic head distribution and the measured SP is independent of permeability. However, the SP distribution can also be affected by the distribution of the electrical resistivity that, as illustrated by Rizzo et al. (2004), shows a layered structure. In this case, the distribution of the electrical resistivity affects the strength of the SP signals (through the apparent coupling coefficient defined in the one-dimensional [1D]-linear approximation) but not on the relative variation of the SPs. A strong correlation was noted by Rizzo et al. (2004, figure 9) between SP change vs. piezometric change during the relaxation of the phreatic surface. They used the electrodes close to each piezometer including the pumping well. The 2D finite-difference modeling made by Titov et al. (2005) shows very clearly how the linearity between the SP changes and the piezometric level changes (1D approximation) is preserved despite heterogeneity in the hydraulic conductivity distribution. The validity of this equation was tested by Titov et al. (2005) who benchmark this 1D relationship with the finite-difference solution of the coupled hydroelectric flow problem. They show clearly that the 1D approximation is a good approximation for the geometry of the test field investigated by Rizzo et al. (2004). The validity of the SP/hydraulic head relationship is validated by figure 9 of Rizzo et al. (2004) with a high correlation coefficient ($r = 0.90$, seven points). To obtain the drawdown distribution over the aquifer (Figure 3), the linear correlation was exploited between the SP data measured close to the piezometers and the hydraulic heads measured in the piezometers (Rizzo et al. 2004). In essence, this linear relationship was applied to all the SP data by means of a kriging with external drift (Troisi et al. 2000).

A drawdown-time-distance analysis was performed to estimate $T$ and $S$ values of the aquifer using a traditional method for the aquifer test. The approach used to estimate the effective $T$ and $S$ is a simultaneous regression of all the drawdown, induced by pumping at the central well and observed at remaining wells, from time zero to any given time. So, to obtain the transmissivity and storage coefficient, we used a nonlinear least square method to invert the parameters of the Theis equation (Theis 1935; Rizzo et al. 2004). This method is based on the minimization of the following merit function:

$$
\chi^2 = \sum_{i=1}^{n} \sum_{j=1}^{\max t} \left[ \frac{h(r_i, t_j) - \hat{h}(r_i, t_j)}{\sigma(r_i, t_j)} \right]^2 = \min
$$

Figure 3. Determination of the piezometric surface in steady-state conditions. Comparison between (a) the piezometric surface determined with only the piezometric data and (b) the piezometric surface determined using the piezometric data plus the SP data (Rizzo et al. 2004).
where the index \( i \in \{1, \ldots, n\} \) represents the observation well number (\( n = 5 \) here), \( j \) denotes the time index of the drawdown data, and:

\[
\hat{h}(r_i, t_j) = h^0(r_i) - \hat{s}(r_i, t_j)
\]

(2)

\[
h(r_i, t_j) = h^0(r_i) - s(r_i, t_j)
\]

(3)

\( h(r_i, t_j) \) are the observed hydraulic heads during the pumping test in the observation well \( i \) at time \( j \). \( \hat{h}(r_i, t_j) \) are the hydraulic heads determined by means of Theis solution, \( s(r_i, t_j) \) and \( \hat{s}(r_i, t_j) \) are the associated drawdowns, and \( \sigma(r_i, t_j) \) the measurement errors (standard deviation) of the \( i \)th data point at time \( j \), presumed to be known. To calculate \( \hat{s}(r_i, t_j) \), we use the Theis solution:

\[
\hat{s}(r_i, t_j) = \frac{Q}{4\pi T} \exp\left\{-\frac{r_i^2 S}{4Tt_j}\right\}
\]

(4)

In Figure 4, we report the plot of the drawdowns vs. time, during the pumping test. This analysis produces evolving and different \( T \) and \( S \) values for each observation well, during the pumping test. At long time, \( T \) estimates stabilize but do not converge to a single value (i.e., effective \( T \) ) (Figure 5). Estimated \( S \) values diverge but stabilize to different values for each observation well (Figure 6). These are consistent with the findings by Wu et al. (2005). Applying this numerical procedure to the present pumping test, the final transmissivities and storage coefficients were obtained, reported in Table 1.

After assuming the aquifer homogeneity, as in the previous analysis, we now try to estimate the spatial distribution of the transmissivity, assuming the aquifer as heterogeneous, through a geostatistical approach based on the SLE (Yeh et al. 1996). To condition the transmissivity distribution, the hydraulic heads measured by means of direct measurements in the observation wells and SP measurements in the 53 locations were used. Since the SP measurements regard only the steady-state condition of the pumping test, the equation used to characterize the aquifer is the steady-state flow equation that permits only the transmissivity parameter to be determined rather than transmissivity and storativity as in the transient case. To conduct this characterization analysis, a stochastic estimator (SLE) by Yeh et al. (1996) was used, which is briefly described subsequently. To characterize the heterogeneity of geologic formations, the SLE algorithm considers the natural logs of transmissivity as a spatial stochastic process. Therefore, \( T(x) = \exp[F(x) + f(x)] \), where \( x \) is the location vector, \( F \) is the mean of the log of \( T \) values, and \( f \) denotes the perturbation. The hydraulic head response to an aquifer test is represented by \( H(x) = \bar{H}(x) + h(x) \), where \( \bar{H} \) is the mean and \( h \) is the perturbation. Substituting these stochastic variables into a 2D depth-averaged equation, taking the conditional expectation, and conditioning with some observations of...
head and parameters generates the mean flow equation as follows:

$$\nabla [T_c(x) \nabla H_c(x)] + Q(x_p) = 0$$

(5)

where $T_c(x)$ and $H_c(x)$ are the conditional effective transmissivity and the hydraulic head, respectively (Yeh et al. 1996). According to Equation 5, the SLE method then seeks the conditional effective fields of transmissivity conditioned on the information from 2D hydraulic tomography surveys (from SP signals) and some direct measurements of $T$; if any. Specifically, the estimation procedure starts with a weighted linear combination of direct measurements of the hydraulic properties, if any, and draw-down data at different observation locations due to the pumping test to obtain the first estimate of the parameters. The weights are calculated based on statistical moments (namely, means and covariances) of parameters, the covariances of heads in space, and the cross-covariances between heads and parameters. After the estimate, the covariances are updated to reflect the effects of conditioning and they are called residual (or conditional) covariances, representing the uncertainty associated with the estimate. This estimation procedure is identical to cokriging. Afterward, the first estimate is used in the mean flow Equation 5 to calculate the new heads at observation locations (i.e., forward simulation). Differences between the observed and simulated heads are subsequently determined. A weighted linear combination of these differences is then used to update the previous estimates. During the update, the weights are evaluated using residual (conditioned) covariances and cross-covariances that are updated at each iteration. Iterations between the forward simulation and estimation continue until the improvement in the estimates diminishes to a prescribed value. Detailed algorithm and procedures are given in Yeh et al. (1996). To apply the SLE to the field site, a square area of $400 \times 400$ m was selected and discretized into $160 \times 160$ elements, with an element size of $2.5 \times 2.5$ m. The dimensions of the domain take into account the influence of radius measured during previous aquifer tests (Troisi and Straface 1996). The element size of the mesh has an important rule because the greater the size the fewer the number of SP locations in that it is a possible use to condition the transmissivity distribution. So the element dimension is a correct compromise between the necessity to use all SP information and to reduce the computational cost of the inverse model. In this case, the optimal element was of $2.5$ m, and for this size, only 17 averaged hydraulic heads from SP locations were used. Application of the SLE model, apart from the hydraulic head information, requires spatial variability statistics of the aquifer hydraulic properties, i.e., the mean, variance, correlation scales, and theoretical model of covariance. A mean value of $2.57 \times 10^{-4}$ m$^2$/s for $T$ was used, based on the results from the aquifer test in the previous section. To estimate the transmissivity distribution, we have to guess an initial value for the variance and the correlation scale. In this case, and for the first estimation, a previously determined variance and a correlation scale were used, from a structural analysis on transverse resistance, by Straface et al. (2005); i.e., $\sigma^2 = 0.75$ and $\lambda_s = \lambda_r = 13$ m. It should be emphasized that first statistical parameters are estimations; therefore, they are not defined exactly, and moreover, during head data collection, measurements errors are inevitable. For this reason, a sensitivity analysis was carried out to verify the effects of uncertainty in the statistical parameters and of errors in estimation measurement obtained with our inverse model. Many numerical experiments were carried out for verifying the average effect on estimation of the transmissivity value by means of the SLE (Yeh and Liu 2000). In short, it can be seen that the average uncertainty causes only a change in average of transmissivity field estimated, leaving the heterogeneity structure almost unaltered. Note that the uncertainty associated with the variance of transmissivity has no influence on the final estimate. This is attributed to the fact that the SLE inverse approach relies on correlation and cross correlation, which do not involve the variance (Yeh and Liu 2000). Uncertainty tests on the correlation scale in the horizontal and vertical directions were carried out under- and overestimating to 50%. The value of MSE of estimation was calculated for each test. As seen in Figure 7, the SLE inverse method is not very sensitive to the correlation scale uncertainty, unless this is not very large. At this point, the SLE inverse model was launched obtaining the transmissivity estimation by means of the SP hydraulic heads. As can be seen in Figure 8, the SLE results show a high degree of transmissivity heterogeneity close to the wellfield: a low permeable zone among the wells and a high permeable zone far from the wellfield. This means that, probably, our wellfield is located in a pocket of a high permeability zone as illustrated in our estimated $T$ field, and the less permeable zones are likely to present a pinchout of the aquifer.

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<table>
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<tr>
<th>Observation Wells</th>
<th>Well 1</th>
<th>Well 3</th>
<th>Well 7</th>
<th>Well 9</th>
<th>Well 11</th>
<th>Effective Parameters</th>
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<td>3.52</td>
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<td>3.50</td>
<td>2.84</td>
<td>5.29</td>
<td>1.29</td>
<td>2.26</td>
</tr>
</tbody>
</table>

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**Table 1**

Transmissivity and Storage Coefficient Obtained with the Nonlinear Least Squares Method
Discussions

To prove, or otherwise, the consistency of the results shown in Figure 8, the drawdown rate can be analyzed and, specifically, the derivative of the drawdown with respect to the ln(t): \( d/d \ln t = t d/dt \), rather than the actual drawdown, because, as several authors have shown, the drawdown rate is more sensitive to the variability in the transmissivity field (Copty and Findikakis 2003; Wu et al. 2005). Regarding the drawdown rate, i.e., the derivative of the drawdown with respect to the ln(t), it is a very important quantity that defines the transmissivity and storativity heterogeneity and the boundary effects. Specifically, if the drawdown-log(t) curve is concave, its derivative grows, while if the drawdown curve is convex, its derivative decreases. Where this curve goes from a concave to a convex behavior, there is a flex point and, in the derivative curve, a relative minimum or maximum. When a boundary effect is met, a flex point appears in the drawdown-log(t), and the drawdown rate there is again a relative maximum that indicates this effect on the drawdown. So by analyzing the drawdown rate, the heterogeneity and the boundary effects of the porous media can be estimated. When a relative minimum or maximum is found, the position of the transmissivity variation or the boundary effect can be calculated using the relation of Oliver (1990) that is valid for a small permeability variation from an average value:

\[
 r = 0.015 \left( \frac{K_l}{\phi \mu c_t} \right)^{1/2} 
\]

where \( r \) is the radial distance between the pumping well and the boundary or heterogeneity (m), \( K_l \) is the average intrinsic permeability (\( \mu m \)), \( t \) is the time (h), \( \phi \) is the porosity (\( . \)), \( \mu \) is the dynamic viscosity (Pa.s), and \( c_t \) is the total compressibility (Pa\(^{-1}\)). Figure 9 shows the behavior of the derivative of the drawdown with respect to the log(t) for every observation well. If the behavior of all the drawdown rates is observed, three flex points can be seen on the curves, approximately, at 1, 33, and 80 h. For these times, the radius of influence was calculated by means of the Oliver relation. These results are shown in Table 2.

In Figure 8, it can be seen that the SLE model results are consistent with the radii of influence estimated by
means the drawdown rate and Oliver relation. In fact, drawing these influence radii on the transmissivity field obtained by means the SLE, it can be seen that first radius corresponds to a change of aquifer transmissivity. This analysis validates the SLE model results; thus, the methodology used to estimate the transmissivity distribution in the Montalto Uffugo Aquifer. Moreover, we have calculated the hydraulic heads using transmissivity distribution obtained by means of SLE and compared them with the measured ones, as we can see in the scattergraph of Figure 10, there is a good agreement between them.

Conclusions

The flow of ground water during a pumping test experiment is responsible for a measurable electrical field at the ground surface owing to the electrokinetic coupling between the Darcy velocity and the electrical current density. The SP method is a geophysical technique capable of measuring this electric potential, which spontaneously develops within the aquifer due to the velocity field led by pumping. This electric potential is correlated with the velocity and consequently, using the potential field theory, with the water table depth (Revil et al. 2002, 2004), so these measurements allow us to estimate the aquifer drawdown under the electrode placed on the soil surface (53 in the examined case) with a good degree of accuracy. On the other hand, the SLE is an iterative geostatistical method where a linear estimator was used successively to incorporate the nonlinear relationship between hydraulic properties and pressure head. By means of the SLE used on both transmissivity and head data provided through SP experiments, a transmissivity field conditioned by measurements of hydraulic head (Figure 8) was estimated. These results prove that this methodology can be used for the transmissivity estimation of a 2D aquifer. This opens up very interesting scenarios for aquifer characterization because a high number of measurement points can be used without a significant increment in the costs and change in the properties of the aquifer.

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Figure 10. Comparison of the calculated hydraulic heads obtained using SLE transmissivity distribution vs. the measured hydraulic heads obtained by means of SP method.


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