Near Surface Geophysics EAGE EUROPEAN ASSOCIATION OF GEOSCIENTISTS & ENGINEERS

Near Surface Geophysics, 2018 doi: 10.1002/nsg.12020

Estimation of hydraulic parameters using electrical resistivity tomography (ERT) and empirical laws in a semi-confined aquifer

Santiago Perdomo[1,2,3]*, Eduardo E. Kruse[2] and Jerónimo E. Ainchil[3]

[1]*Centro de Investigación y Transferencia del Noroeste de la Provincia de Buenos Aires, Buenos Aires, Argentina,* [2]*Consejo Nacional de Investigaciones Científicas y Técnicas (CONICET), Av. Rivadavia 1917 (C1033AAJ), Buenos Aires, Argentina, and* [3]*Facultad de Ciencias Astronómicas y Geofísicas, Universidad Nacional de La Plata, Buenos Aires, Argentina*

Received February 2017, revision accepted October 2018

ABSTRACT

The estimation of hydraulic parameters is critical for the rational use of water resources and the development of reliable hydrogeological models. However, the cost of such estimation can be very high and the data are limited to the area near the pumping well. For this reason, complementary methods for estimating hydraulic conductivity and transmissivity have become increasingly important in recent years, such as the adjustment of empirical relationships between geoelectrical and hydraulic parameters. In this paper, two linear relationships were tested, combining resistivity measurements from well logging profiles and hydraulic conductivity values from pumping test data, in a semi-confined fluvial aquifer in the province of Buenos Aires, Argentina. Furthermore, these relationships were used to obtain two-dimensional (2D) hydraulic conductivity and transmissivity sections from electrical resistivity tomography using a high-definition electrode array. Predicted values were compared with traditional pumping test in a near well showing very good agreement with both methods. Results showed that it would be possible to quantify the 2D variation of hydraulic parameters in aquifers and to identify high- or low-productivity areas. By knowing this information in advance, it is possible to reduce the number of failures or unexpected results when drilling a well. These 2D sections also provide additional information about hydraulic parameters and their lateral variability, and can improve hydrogeological models without drilling new wells.

Key words: ERT, Aquifer, Hydrogeology, Hydraulic conductivity, Hydraulic transmissivity.

INTRODUCTION

Drilling exploratory wells is the most reliable method to obtain direct and accurate information about the subsurface; however, it tends to be only locally representative and its regional validity is subject to assumptions about the spatial variation of a given quantity. Information from wells is inherently one-dimensional (1D) and the characterization of heterogeneous aquifers relies on their distribution (Soupios *et al.* 2007; Ruggeri *et al.* 2013; Ruggeri *et al.* 2014). Pumping tests that are used to determine hydraulic parameters, such as transmissivity (T in m^2/day) and hydraulic conductivity (K in m/day), are generally very limited in number and sparsely distributed. Besides, they tend to represent only an average or an effective property of the subsurface region tested (Rubin 2003; Ruggeri *et al.* 2013; Ruggeri *et al.* 2014). The number, depth and distribution of wells determine how precisely the subsurface variability is resolved. However, different assumptions must be made in order to study lateral variations of a certain property inferred with 1D data. Recent advances

*E-mail: perdomo.geofisica@gmail.com

in hydraulic tomography propose a sequential aquifer test, whereby the hydraulic head is measured in multiple wells (typically 5–10) for sequential pump tests on isolated vertical sections of an aquifer (Yeh and Liu 2000; Slater 2007; Bohling and Butler 2010; Hochstetler *et al.* 2016). This methodology is time-consuming and expensive, and may be ineffective in some environments. For these reasons a method is presented here to assess lateral variability of hydraulic parameters using two-dimensional (2D) geoelectrical data.

In this paper, we tested two empirical relationships between formation factor (FF) and K, and transverse resistance (TR) and T. Natural gamma-ray well logs were analysed in order to verify absence of clay sediments for the correct computation of FF. Therefore, resistivity values of the clean sand formation were considered for the calculation of FF and TR. Subsequently, these two relationships were applied to electrical resistivity tomography data, to identify the complexity of the alluvial sediments at a range of 30–70 m deep in a semiconfined aquifer in the Río de La Plata Craton, Argentina. In this aquifer, variations in hydraulic parameters were reported, and many wells were located in sites with low K. This procedure provides quantitative values of the variation of hydraulic parameters in a 2D section, which could be used to estimate effective hydraulic properties. Prior to choosing the location of a well, this method will assist hydrogeologists in selecting the best site according to a quantitative estimate of K or T.

PREVIOUS INVESTIGATIONS

Since in porous media porosity and tortuosity are supposed to control the electric current flow and also the flow of water, many authors have used this analogy to adjust empirical equations using electrical and hydraulic parameters (Freeze and Cherry 1979; Fitts 2002; Singh 2005; Milsch, Blöcher and Engelmann 2008).

The conduction of electricity in a porous media occurs by the movement of ions through the saturating fluid, and by movement of adsorbed ions along the surfaces of pores and cracks (Wildenschild, Roberts and Carlberg 2000). The first attempt to relate the bulk resistivity of a sedimentary rock and fluid resistivity was proposed by Archie (1942), using a capillary tubes model:

$$FF = \frac{\rho_b}{\rho_w}, \tag{1}$$

where FF mainly depends on tortuosity and porosity — ρ_b is the bulk resistivity and ρ_w is the water resistivity. This relationship shows conceptual simplicity, but also a strong relevance to sand and gravel aquifers (Mazáč, Kelly and Landa 1985; Niwas, Tezkan and Israil 2011; Niwas and Celik 2013; Ruggeri *et al.* 2013). A power relationship among bulk resistivity, water resistivity and porosity (ϕ) is also suggested for clean sands (Worthington, 1993):

$$\rho_b = a\rho_w\phi^{-m}, \tag{2}$$

where Superscript m is a cementation factor and a is a pore-geometry coefficient. Usually, m is interpreted as a grain size or pore shape indicator (Jackson, Taylor Smith and Stanford 1978; Worthington 1993). The a coefficient is considered constant and nearly one for uncemented soils and reservoir sands (Worthington 1993; Morin *et al.* 2006; Choo *et al.* 2016). Both parameters are variable and site specific, since it has been argued that equation (2) is a semi-empirical relationship (Worthington 1993). This variability is explained by different causes, such as cementation, sorting and packing of grains, tortuosity, pore geometry and clay content. When clay minerals are present, the surface conductivity increases significantly and cannot be neglected if freshwater aquifers are considered (Wildenschild *et al.* 2000). However, it is still valid to calculate formation factor (FF) from equation (1) and treat it as an apparent magnitude that depends on water resistivity (Worthington 1993). Archie's law was the first attempt to link hydraulic and electrical properties for sedimentary rocks. If the Kozeny (1953) equation is used, it is possible to estimate intrinsic permeability (k_f) from resistivity data:

$$k_f = \frac{d^2}{180}\frac{\phi^3}{(1-\phi)^2}, \tag{3}$$

where d is grain size [m]. Laboratory analysis of hydraulic permeability and electrical resistivity with rock samples is one of the approaches used by Jones and Buford (1951), Wildenschild *et al.* (2000), Milsch *et al.* (2008), Gomez, Dvorkin and Vanorio (2010) and Choo *et al.* (2016).

At a macroscopic scale, attempts have been made to adjust empirical relationships between hydraulic data from pumping tests and electrical resistivities derived from indirect methods (e.g. direct current methods, induced polarization). Some authors adjusted a linear regression (Kelly 1977, Urish 1981, Mazac *et al.* 1985, Chen, Hubbard and Rubin 2001), while others derived a potential equation (Heigold *et al.* 1979, Purvance and Andricevic 2000, Singh 2005; Dhakate and Singh 2005, Shevnin *et al.* 2006, Soupios *et al.* 2007, Sinha, Israil and Singhal 2009). Predictions concerning these relationships were consistent with the expected values in each area. Although these results were site specific, the methodology was important to improve the information at a low cost and to extend hydraulic parameters to other areas where bore wells were not available.

As an attempt to develop a more theoretical insight, Niwas and Singhal (1981) investigated and proposed a relationship between T and Dar Zarrouk (DZ) parameters (Maillet 1947). They suggest adjusting a linear equation between the T and transverse resistance of an aquifer, considering the hydraulic and electrical resistivity ratio (K-ρ) at a well with hydraulic data:

$$T = (K\rho)\, S, \tag{4}$$

$$T = \left(\frac{K}{\rho}\right) TR, \tag{5}$$

where S is the longitudinal conductance.

These relationships could then be used to extend hydraulic information to other areas with geoelectrical data. It is only necessary to adjust a value for the (K-ρ) constant. Satisfactory results were obtained by Massoud *et al.* (2010), Taheri Tizro, Voudouris and Basami (2012) and Niwas *et al.* (2011) when using DZ parameters.

Petrophysical models that relate electrical and hydraulic conductivities use tortuosity as a link between both properties. A detailed revision of different tortuosity definitions is presented in Clennell (1997), as well as an analysis of the derivation of the Kozeny (1927) semi-empirical equation and the modification proposed by Carman (1937). The Kozeny–Carman equation is also a semi-empirical equation that shows very good results and agreements between experiments and simulations (Costa 2006; Soupios *et al.* 2007, Khalil, Ramalho and Monteiro Santos 2011, Niwas and Celik 2012; Choo *et al.* 2016). According to Clennell (1997), considering tortuosity as a retardation factor makes it possible to find relationships between transport properties and FF. Steady-state tortuosity represents an average of transport through all available pathways, whereas details of pore structure are only resolved considering temporal variations, e.g. concentration of a solute (Irving and Singha 2010; Crestani, Camporese and Salandin 2015). However, how to upscale results from petrophysical models to a regional scale is not discussed.

The content of clay minerals in shaly sands is very important, because in sand/shale mixtures, porosity is not a simple function of the shale fraction (Revil and Cathles 1999). The porosity of pure sand or shale is higher than the porosity of a sand/shale mixture. As clay minerals fill the pores of sand, the permeability decreases sharply with the shale content. It is reasonable to find a negative correlation between K and porosity under certain grain-size distributions and packing arrangements (Morin 2006). Revil and Cathles (1999) developed a model for the permeability of clean sand as a function of grain diameter, porosity and electrical cementation factor. The proposed methodology needs a grain-size distribution obtained from core samples, determination of the clay mineralogy and shale content (gamma-ray well logging), and also the state of compaction of sand and shale. Although it was considered a first step towards developing a more rigorous model, the need for these specific parameters limits the applicability of this methodology, mainly because these represent a high cost for hydrogeological studies.

Ruggeri *et al.* (2013) make an interesting study on the possibility of integrating resistivity and hydrological data at a regional scale. Through a synthetic example, they demonstrate that it is possible to generate a regional-scale K structure from surface-based electrical resistivity tomography (ERT) using a Bayesian sequential simulation approach. In order to simulate the spatial distribution of porosity, they used a linear regression between porosity and the logarithm of K (Heinz *et al.* 2003). Then, using Archie's law for saturated media, they simulate a spatial distribution of the electrical conductivity. The authors agreed that this idealized petrophysical scenario is conceptually very simple, but with a strong relevance to the case of surficial alluvial aquifers. As with many other methods, the quality of the final results is strongly dependent on the adequacy and reliability of the prior information and this accuracy obviously increases with the amount of borehole data available. Besides, their effectiveness depends critically on the statistical representativeness of the data. The method proposed requires an assumption about the relationship between electrical and hydraulic conductivity. This should be done using petrophysical models or empirical relationships, as proposed and used in this paper.

Previous researchers had different approaches that focus on the estimation of hydraulic properties from geophysical methods. In the end, the main discussion is about how to deal with heterogeneous media. It is possible to consider a spatial distribution of properties or to use a patchwork of homogeneous sub-units (Mesgouez *et al.* 2014). Each sub-unit is described by macroscopic laws (e.g. Darcy), and it is possible to study the structure of these sub-units using two-dimensional ERT data.

BACKGROUND OF THE STUDY AREA

The study area is located on the west bank of the Río de la Plata, at the northeastern edge of a flat plain with rich soil in Argentina. The Río de la Plata is an estuary of the Paraná

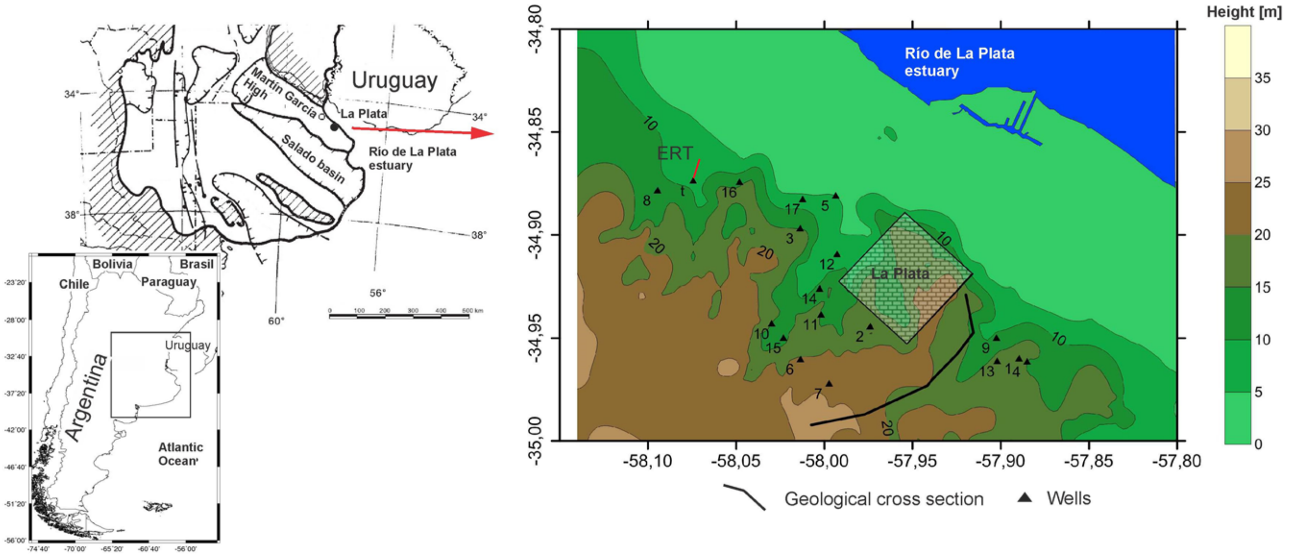

Figure 1 Map of the study area. La Plata is located on the Martín García High, on the northeast edge of the Salado Basin (modified from Caminos 1999).

and Uruguay rivers, which come together to form a broad, shallow marine inlet with muddy sediments between Uruguay and Argentina (Fig. 1). The study area covers 250 km^2 and it is located in an urban environment near the city of La Plata, in the province of Buenos Aires, Argentina. The city is 10–15 km away from the coast of the Río de La Plata.

The area is situated in the Río de La Plata Craton, on the northern border of the Salado Basin (Fig. 1). This region is known as the Martín García or Plata High because the hard rock basement is relatively shallow (300–500 m depth), while the thickest part of the Salado basin has more than 7000 m of sediments. The terrain is a wide plain, slightly undulating due to fluvial erosion. It has two distinguishing features: a low coastal land with heights below 5 m — which is poorly drained, causing an increase in groundwater salinity — and a high plain area, oriented NW to SE, with a gentle slope towards the river ($\approx$1.2 m/km) and underlying freshwater. The streams, creeks and canals in the high plain area are intermittent, whereas in the low coastal land, they are permanent.

Regional and local geological and hydrogeological contexts

The most important freshwater reserve in the region is a semi-confined aquifer called Puelche, but there are other aquifer units that are not as important due to their chemical quality or low permeability (Auge 2005).

This aquifer is composed of fine- to medium-grained quartz sands of fluvial origin (Plio-Pleistocene), with very good quality, low salinity groundwater (less than 1000 mg/L). For these reasons, it is one of the main sources of water for human consumption, but it is also used for irrigation and industry. Towards the low coastal area of the estuary, it becomes saltier (up to 20 000 mg/L). Its thickness ranges from 15 to 30 m.

Overlying the sand, there is silt with carbonate intercalations (Sedimentos Pampeanos [Pampean Sediments]), containing the water table (Pleistocene). These silts have a variable thickness ranging from 25 to 45 m, with fine-grained interbedded sands. In certain sites, it was found that the base of the Sedimentos Pampeanos becomes more clayey, separating the underlying Puelche aquifer. These silty sediments behave as an aquifer of medium productivity with low permeability layers that behave as aquitards. Although water is of good chemical quality, it is subjected to contamination in some areas from urban wastes and agrochemicals, and its use is not recommended for human consumption.

At the base of this important aquifer occurs plastic green clay, which is reached in almost every exploration borehole for water supply. This clay is at the top of the Paraná Formation (Miocene) and it is possible to find fine-grained sands with saline water underneath. In the area, there is only one complete borehole that describes a sequence of aquitards, aquifers and aquicludes of marine origin for the Paraná Formation. Below this sequence, there are continental sands and clays of Oligocene age from the Olivos Formation. At a depth of 490 m, the hard rock basement is composed of granites and gneiss.

Table 1 Typical values for hydraulic parameters of the Puelche aquifer in the study area

Parameter	Range	Average Value
EC (μs/cm)	500–2000	970
T (m^2/d)	150–1200	500
K (m/d)	8–40	25
Thickness (m)	15–30	25

Hydraulic characteristics of the Puelche aquifer

The Sedimentos Pampeanos is a low to medium productivity aquifer with a K ranging from 1 to 10 m/d and a porosity that varies between 5% and 10%.

The groundwater of the Puelche aquifer has very low salinity in the high plain, water electrical conductivity values ranging from 500 to 2000 μS/cm, and a geometric average of 970 μS/cm (Auge, Hirata and López Vera 2004). In the low coastal land, it becomes slightly salty. The aquifer is generally classified as a homogeneous unit with mean T and K values of 500 m^2/d and 25 m/d, respectively, and a thickness of 25 m. However, when analysing data from each pumping well, there is a wide range of values for each parameter, which shows the heterogeneity of the unit (Table 1).

Figure 2 shows a cross section based on the analysis of well logging profiles and their lithological interpretation. In this figure, only the gamma-ray profile is shown because it is the most distinctive property among silt, sand and clay. The section presented here shows the occurrence of clay layers in the Puelche aquifer, and also the absence of the overlying aquitard layer.

METHODOLOGY

Electrical method

Measurements in a resistivity survey are obtained after circulating a direct current through an emission circuit and measuring a potential difference (ΔV) that is generated between two receiver electrodes. The current (I) and voltage measurements are then used to calculate an apparent resistivity (ρ_a) value by using equation (6):

$$\rho_a = G\frac{\Delta V}{I}, \tag{6}$$

where G is a geometric constant that takes into account electrode spacing and measurement settings. Measurements give 'apparent' resistivities because they depend on the geometry of the array and the heterogeneity of the medium. Apparent resistivity represents a weighted average of the true resistivity distribution, which is obtained after inversion of electrical data (Revil *et al.* 2012).

In a cell-based inversion model, the subsurface is divided into a large number of cells and an inversion algorithm is used

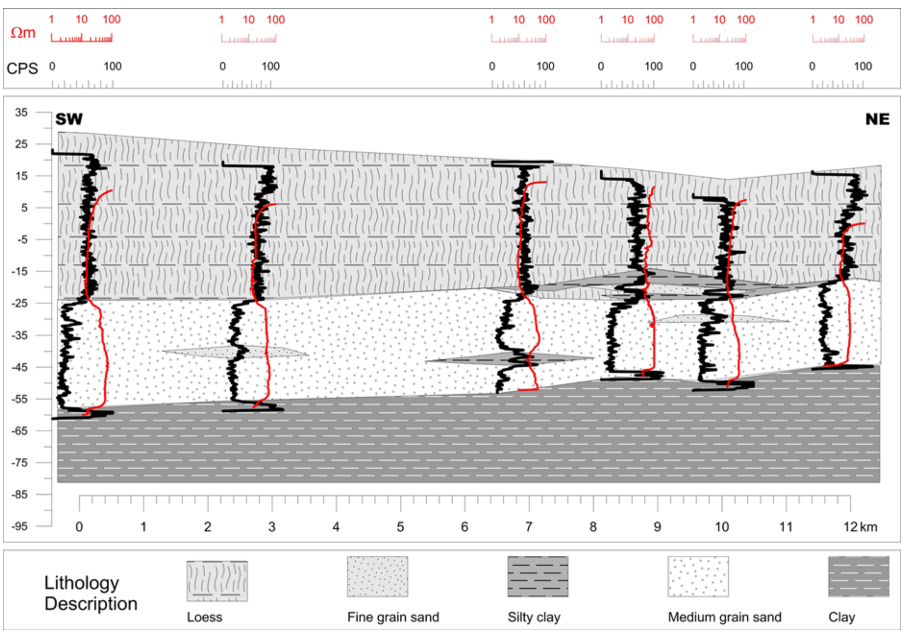

Figure 2 Geological cross section based on well descriptions and geophysical profiling (see Fig. 1). The gamma-ray profile, in counts per second (CPS), distinguishes between loess (Sedimentos Pampeanos), fine- and medium-grained size sands (Puelche aquifer) and clay (Paraná Formation).

to determine the resistivity of the cells. The inversion of resistivity data gives non-unique results. The inversion process departs from an initial model (e.g. homogeneous half-space), and a non-linear least-square optimization method is used to iteratively change the resistivity of the model cells (deGroot-Hedlin and Constable 1990; Loke and Barker 1996). The method minimizes the difference between measured data and calculated apparent resistivity values by adjusting the resistivity of the model blocks. To compare the closeness between calculated values and data a percentage root mean square error is calculated.

A discrete two-dimensional resistivity survey, i.e. 2D electrical resistivity tomography, combines measurements with increasing electrode separation and lateral shifts of the array. There are different arrays of electrodes that differ on many factors, such as horizontal/vertical sensitivity, signal to noise ratio and strength of the electric field. The most common ones are the Wenner, dipole–dipole and Wenner–Schlumberger arrays (Loke 2015).

Transverse resistance and longitudinal conductance

Maillet (1947) defined the Dar Zarrouk (DZ) parameters considering the dependence of electric current flow within a layered medium. Transverse resistance (TR) is defined as the resistance normal (Ωm^2) to the bedding face and S is the conductance (Ω^{-1}) in the parallel direction. If the resistivity model consists of a sequence of a finite number of horizontal, homogeneous and isotropic layers with constant thickness (h) and resistivity (ρ), the DZ parameters are defined as:

$$TR = h\rho, \tag{7}$$

$$S = \frac{h}{\rho}. \tag{8}$$

During an inversion process, the interpreter aims at obtaining two fundamental parameters for a multilayer resistivity model: thickness and resistivity. The concept of equivalence proposed by Maillet (1947) states that it is not possible to know both the resistivity and thickness of each layer, but each model may have similar values for TR and S. For this reason, the use of DZ parameters reduces the ambiguity related to an accurate interpretation of resistivity or thickness.

Long-term pumping test

Pumping tests were conducted and observation wells were available for only a few of the tests. A step-drawdown test consisting of 3–4 steps was conducted in each well for 12 hours.

Then, a constant-rate test was run for 24 hours and recovery was measured for the same duration as pumping. Step-test data were analysed by the Theis (1935) graphical method. Constant-rate tests were analysed by Jacob's straight-line method (Driscoll 1989):

$$T = \frac{0.183 Q_b}{\Delta s}, \tag{9}$$

where Q_b is the pumping rate [m^3/d] and Δs is the measured drawdown [m].

Empirical laws between hydraulic and geoelectrical parameters for the Puelche aquifer

In this paper, two linear empirical equations are suggested, using hydraulic properties obtained from 17 pumping tests in the high plain area (Fig. 3), as well as geoelectrical parameters derived from geophysical well logging profiles (Table 2). Geophysical measurements include a resistivity profile and natural gamma-ray well logs at a sampling interval of 0.02 m. The error was calculated taking logarithm values of transverse resistance (TR).

Each well was analysed individually; natural gamma-ray well logs were used to determine a more precise thickness and to estimate the clay volume in the alluvial sands. In general, the clay volume was estimated to be below 10% and only four wells showed a layer with a maximum of 50% clay, which was not considered for the computation of resistivity.

In the high plain area, 27 water samples were collected for water electrical conductivity (EC) measurement and only four samples showed relatively high EC values (1590, 1797, 1968 and 2080 μS/cm). The formation factor (FF) (equation (1)) was calculated using resistivity measurements and a mean value for the water EC (970 μS/cm or 11.5 Ωm). Besides, the TR in equation (7) was calculated using the thickness from natural gamma-ray well logs and the electrical resistivity for the clean sand formation.

A linear regression was fitted using the least-squares approach between FF and K (Fig. 4a). A correlation factor of 0.87 was calculated, showing that there is a strong positive linear relationship between these two parameters, and it might be possible to obtain satisfactory K estimates from the FF using bulk and water resistivity measurements in the Puelche aquifer.

$$K = 5.19FF. \tag{10}$$

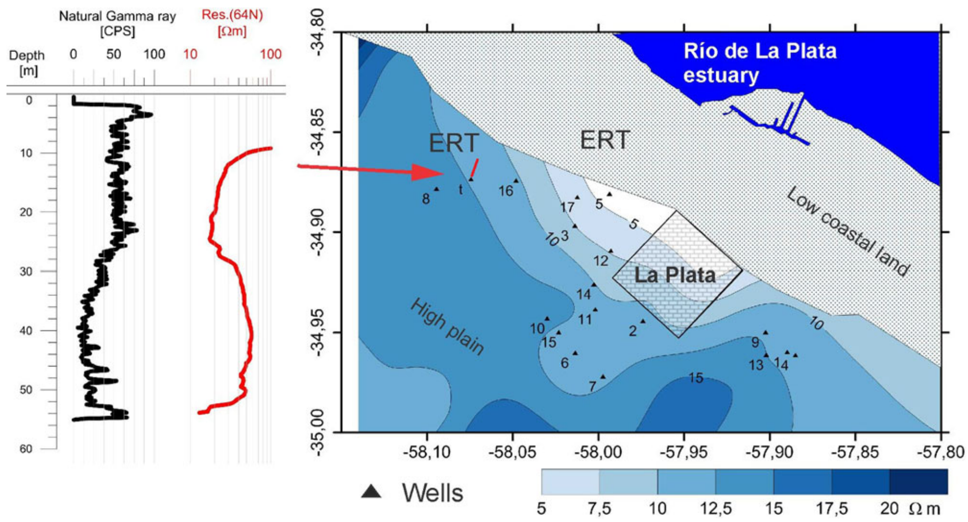

Figure 3 Distribution map of the water electrical resistivity of the Puelche aquifer is shown. Superimposed is the location of wells and ERT profiles in the high plain area. On the left, it is shown the geophysical logging used to compare with ERT results. The Puelche aquifer is at a depth between 30 m and 55 m.

Table 2 Geophysical logging data and pumping test parameters used to fit the regression line

ID	Depth (m)	Thickness (m)	Resistivity (Ωm)	FF –	TR (Ωm^2)	Error (%)	T (m^2/d)	K (m/d)
PW 1	−47	14	48 ± 6	4.1 ± 0.52	677	1.7	389	22
PW 2	−45	20	33 ± 7	2.8 ± 0.61	651	3.2	266	13
PW 3	−34	20	54 ± 6	4.7 ± 0.61	1077	1.5	412	21
PW 4	−41	18	41 ± 7	3.6 ± 0.52	729	2.6	400	23
PW 5	−42	20	39 ± 8	3.4 ± 0.61	784	2.7	424	21
PW 6	−44	25	67 ± 13	5.8 ± 0.69	1655	2.6	901	28
PW 7	−44	28	67 ± 13	5.8 ± 1.13	1882	2.3	1180	42
PW 8	−32	19	50 ± 5	4.4 ± 1.13	953	1.3	462	20
PW 9	−32	22	24 ± 6	2.1 ± 0.43	520	3.8	291	12
PW 10	−41	21	41 ± 10	3.6 ± 0.52	864	3.2	584	17
PW 11	−45	16	37 ± 5	3.2 ± 0.87	591	2.0	215	13
PW 12	−33	14	15 ± 2	1.3 ± 0.17	215	1.9	189	9
PW 13	−40	20	33 ± 8	2.8 ± 0.69	657	3.4	226	13
PW 14	−39	19	27 ± 4	2.4 ± 0.35	517	2.1	264	14
PW 15	−42	26	60 ± 4	5.2 ± 0.35	1586	0.6	623	24
PW 16	−47	11	28 ± 4	2.4 ± 0.35	292	3.3	151	14
PW 17	−31	22	30 ± 3	2.6 ± 0.24	669	1.3	363	17

A similar approach has also been adopted, using TR calculated from the resistivity profile and T estimated by pumping tests (Fig. 4b). These data showed a linear relationship with a very strong positive correlation ($R^2 = 0.93$):

$$T = 0.53TR. \tag{11}$$

It is interesting to note that a coefficient of 0.51 was calculated using equation (5) proposed by Niwas and Singhal (1981). The relationship between K and ρ was calculated in each well, obtaining this median value that is very similar to the one obtained by the linear fitting used here (0.51 and 0.53, respectively).

The relationships between FF-K and T-TR showed a strong correlation of the data in this semi-confined aquifer. They are both equally valid and could be satisfactorily applied to obtain hydraulic parameter estimates from surface electrical methods in the area.

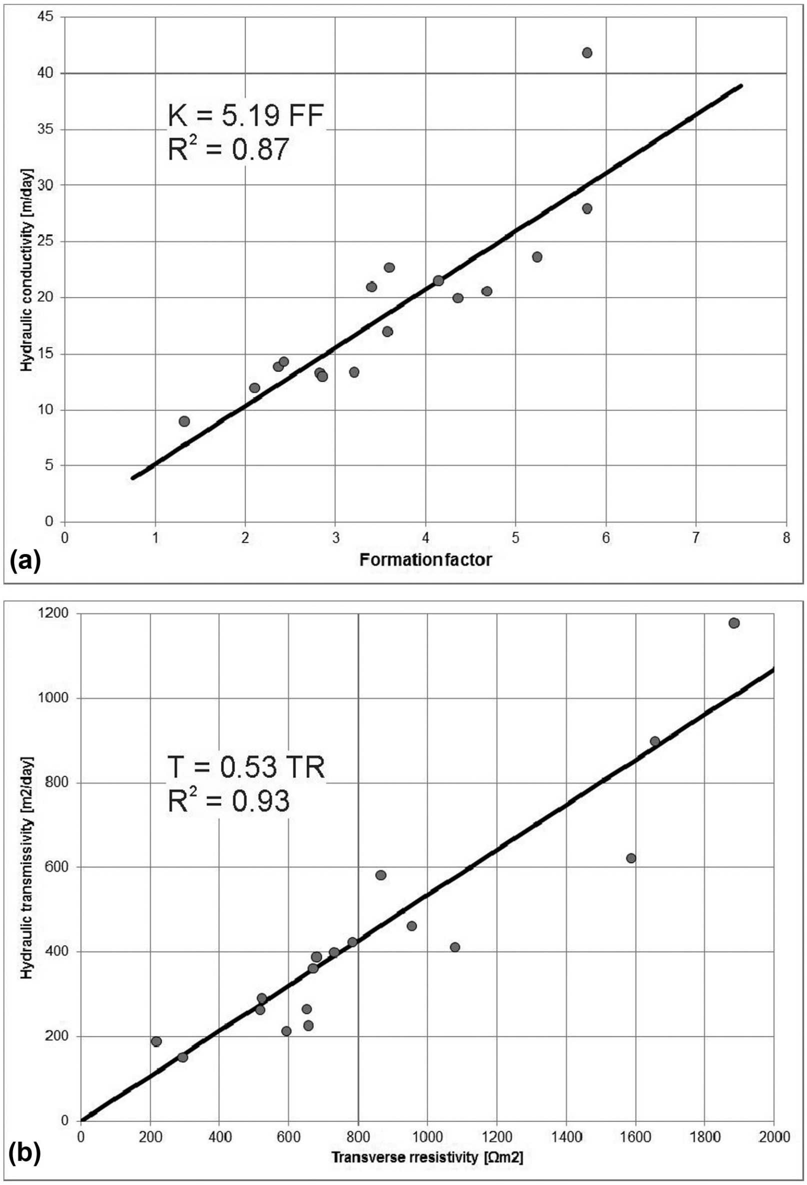

Figure 4 (a) Linear regression estimated using the least-squares method between FF from resistivity logging and *K* from pumping test data in the same well. (b) A straight line is also fitted between TR and T.

RESULTS

Synthetic data model

The reliability of the electrical resistivity tomography (ERT) method for quantitative estimations depends on the accuracy of the determination of geoelectrical parameters (thickness and resistivity). There are different tools to evaluate the reliability of the resistivity model (Caterina *et al.* 2013), however, they strongly depend on the resistivity distribution and a unique estimator is not completely accepted yet. For this reason, a forward modelling is used to evaluate the ERT resolution for a given resistivity contrast and a specific geometry of the alluvial aquifer.

In this section, the electrical response of a simple synthetic model is analysed to yield information about image resolution and accuracy. This analysis provides a proof that ERT is able to recover the true depth and true resistivity of the sediments in the area. The true resistivity model is based on the analysis of natural gamma-ray well logging and resistivity logging. The first layer, with an average resistivity of 10 Ωm and a thickness of 40 m, represents the Sedimentos Pampeanos. The Puelche aquifer is represented by a 30-m thick layer and a

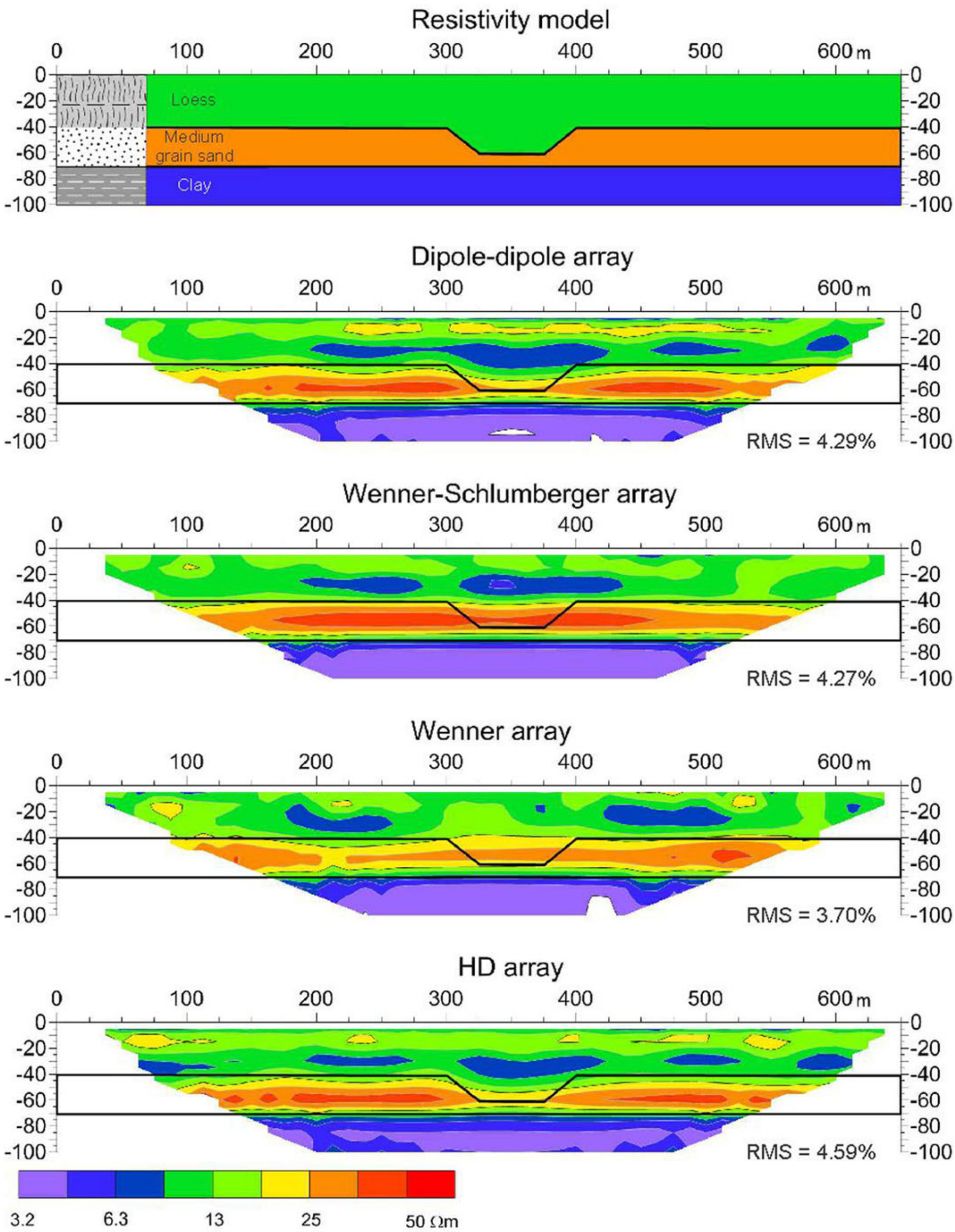

Figure 5 Synthetic resistivity model. (a) The first layer represents the Sedimentos Pampeanos with a resistivity of 10 Ωm. A thickness reduction occurs in the Puelche aquifer in the middle of the line (10 m) with a constant resistivity of 30 Ωm. The conductive base is an infinite layer of 5 Ωm. After the inversion process, the best models are presented: (b) Dipole–dipole, (c) Wenner-Schlumberger, (d) Wenner and (e) High-definition array.

30 Ωm resistivity layer, overlying a conductive base layer of 5 Ωm (Paraná Formation). The model considers a reduction in thickness of the Puelche aquifer (Fig. 5). This kind of lateral change is expected in alluvial sediments and it is important to test which array will be able to detect it. The apparent resistivity section of a simple three-layer model is calculated using the RES2DMOD software (Loke 1999), adding a typical Gaussian noise value of 5%., and then it is inverted using

RES2DINV (Loke 2006). For the inversion process, an initial model was added considering a flat surface at a depth of 70 m.

Apparent resistivity sections of four arrays were tested with a 25-m minimum distance between electrodes and a maximum depth of investigation of approximately 100 m. The array configurations (electrode spacing and measurement setting) were as follows:

- Dipole–dipole (DD), eight data levels with electrode separations of 25 m and 50 m
- Wenner-Schlumberger (WS), 10 data levels, with electrode separations of 25 m and 50 m
- Wenner (W), eight data levels
- High-definition (HD), a combination of all the previous arrays with overlapping data levels (Loke 2015).

Figure 5 showed that with the DD array, the top and base of the resistive layer were correctly identified, but the resistivity value was slightly lower than it should be. The WS and W arrays failed to resolve the geometry of the aquifer and the resistivity values were lower than expected for the resistive layer. The HD array adequately recovered the top and bottom parts of the aquifer layer and also the resistivity value. It also showed a low resistivity zone in the central portion of the section, as in the DD array, but it was located outside of the target layer (Fig. 5).

In the topmost 25 m of the inverted models, many anomalies appeared in the 8–16-Ωm range. At this stage, these blobs were interpreted as artifacts from the inversion process, but inversion parameters for real data should be chosen to minimize this effect, for example optimize damping factor, apply flatness filter and fix resistivity in the initial model.

ERT imaging results showed that none of the arrays individually recovered neither the true resistivity nor the thickness of the resistive layer (Puelche aquifer). However, a combination of overlapping data levels improved the resistivity models significantly.

Following this analysis, a combination of arrays with overlapping data levels is the best strategy to acquire data when trying to make reliable quantifications based on ERT inverted models. Therefore, a high-definition array was used for the field application.

Table 3 Main parameters set for the inversion

Inversion Parameter	Value
Number of iterations	4
Initial damping factor	0.16
Damping factor	Optimized for each iteration
Mesh	Normal
Model refinement	Half width cells
Vertical to horizontal filter ratio	0.5

Field data analysis

The resistivity survey was conducted in an area in the high plain that considering Fig. 3, the water conductivity of the Puelche aquifer is assumed to be constant (Fig. 3). At the beginning of the profile, there was a borehole used to compare resistivity values and lithological description. A high-definition array was used with 36 electrodes and a minimum separation of 25 m. Resistivity data were inverted using a least-square method and a root mean square 2.23% was obtained. Main inversion parameters and settings are presented in the Table 3. Since some information from the resistivity log was known, it was incorporated as fixed regions of the initial model. A rectangular region was set to represent a conductive base layer 70 m thick with a resistivity of 3 Ωm. This value was allowed to change between iterations.

The ERT final model shows a good correlation with the lithological description of the well '*t*' (Fig. 6). The resistive anomaly occurring approximately between depths of 35 m and 60 m corresponds to the sandy sediments of the Puelche aquifer. The top of this resistive anomaly is fairly horizontal, but there is a thinning area in the centre of the section from the 400 m distance to 550 m in the NE direction.

High-definition electrical resistivity tomography (ERT) shows resistivity values ranging from 6.3 Ωm to 13 Ωm, from the surface to a depth of 30 m, typical values for the silty sediments of the Pampeano aquifer. Between depths of 30 and 60 m, there is a zone with intermediate resistivity values (20–60 Ωm), representing the electrical properties of the Puelche aquifer. Below depths of 60–70 m, there is a drop in resistivity, with values below 6.3 Ωm. This low resistivity zone is characteristic of the clayey sediments from the Paraná formation.

The reliability of the inverted model was evaluated using the percentage uncertainty with smoothness constrain. The uncertainty two-dimensional section in Fig. 6 showed the standard deviation of each parameter. In this sense, a variation of 8–9% in the resistivity values could be expected for the target depth (20–70 m).

The ERT was used to estimate K and T values using equations (10) and (11) for each cell of the discrete model. Fomation factor and Dar Zorrouk parameters were calculated using the bulk resistivity of each cell and a mean water resistivity value (11.5 Ωm).

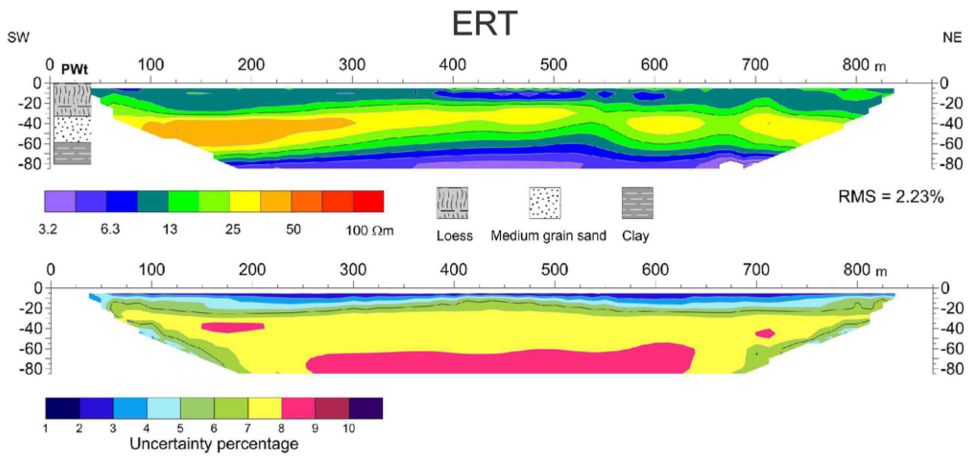

Figure 6 High-definition ERT model (RMS = 2.23%) shows a resistivity range of 25–50 Ωm for the sandy sediments of the Puelche aquifer. The uncertainty of the model is below 9%.

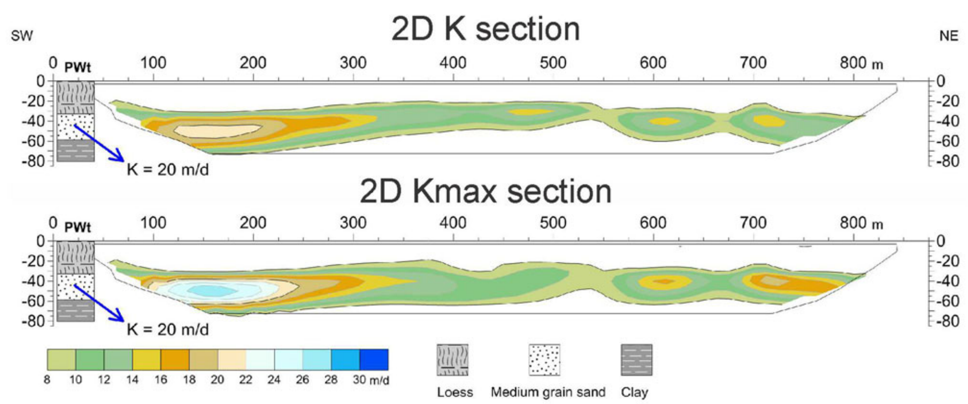

Figure 7 2D hydraulic conductivity section derived from an empirical law. The inverted model and maximum resistivity model (using uncertainty percentage) were analysed. The *K* values for the Puelche aquifer range from 10 m/d to 20 m/d. A *K* value of 20 m/d was calculated in the pumping well (PWt) at the beginning of the section.

Hydraulic conductivity 2D section

Figure 7 shows two two-dimensional *K* sections using the final resistivity model and the maximum resistivity model that result from adding the percentage uncertainty value for each region. These models differed mainly in the *K* estimated values. The Puelche aquifer showed values in the range of 8–20 m/d, with maximum values near the well '*t*'. Long termed pumping test estimated a value of 20m/d. For the maximum resistivity model, the area near PWt showed values in the range of 16–30 m/d. This two-dimensional section showed that there is a lateral variation of *K* values that contributes to an effective *K* estimate of 20 m/d in the pumping well. In this case it could be interpreted that *K* is underestimate near the well, but the range of *K* is still comprised within the uncertainty.

Hydraulic transmissivity 2D section

The calculated distribution of *T* using equation (11) for the Puelche aquifer shows values ranging from 100 m^2/d to 400 m^2/d (Fig. 8). A 440 m^2/d value was calculated in the reference pumping well. A maximum value of 300 m^2/d was derived using the final resistivity model. The lowest *T* values are found between distance 400 m and 600 m in the NE direction, where a thinning of the aquifer was interpreted. Again, the maximum resistivity model showed a better estimator for the *T* with a value 400 m^2/d near the well.

DISCUSSION

Quantitative analysis of synthetic data by forward modelling have shown that none of the typical electrode arrays would

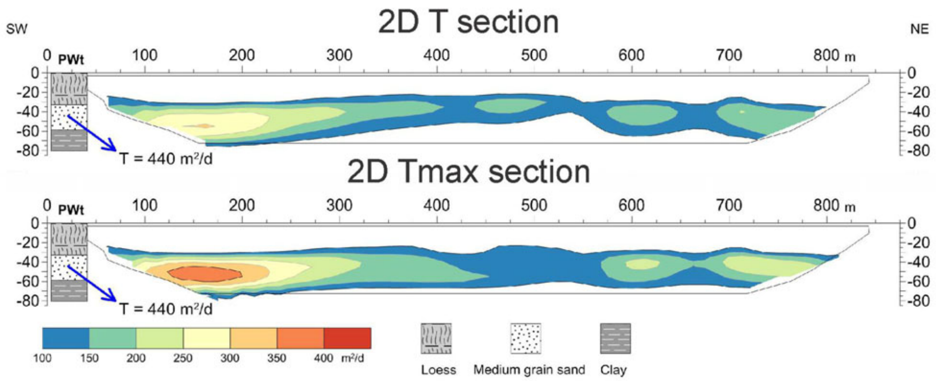

Figure 8 2D hydraulic transmissivity section derived from an empirical law. The inverted model and maximum resistivity model (using uncertainty percentage) were analysed. T values increase from the beginning to the end of the section. Near the reference well a maximum value of 400 m^2/d was estimated.

recover the true resistivity or thickness, and consequently the estimation of K or T would not be correct. But when combining data from these arrays, in an attempt to improve data coverage, a more accurate inverted model was obtained even using the default inversion parameters.

During a survey the electrode spacing is increased in order to achieve a deeper depth of investigation. Also, the cell-size model in an electrical resistivity tomography (ERT) depends on the minimum distance between electrodes, typically tens of metres if the target is at depth between 50 m and 100 m. Thus, it is possible to determine heterogeneities of this order of magnitude in alluvial aquifers. In this context, the lateral resistivity variations observed with ERT are interpreted as hydraulic variations. Thus, all possible heterogeneities contribute to determine an average property in the area of the cone of depression. Estimated values of K and T obtained through long termed pumping tests are considered 'effective properties' of a homogenized media (Mesgouez *et al.* 2014). This statement does not contradict the fact that pumping test data reflect the average characteristics of the media, and that the ERT could be used to assess the complexity of the subsurface.

Also, it should be considered that a pumping test is an integrated measurement in three dimensions, and the ERT models only consider a two-dimensional (2D) geometry. This simplification could also lead to a misinterpretation of the derived values with the procedure proposed in this work. However, it is a significant improvement when comparing to other methods, such as geostatistical techniques that relies on interpolation and data coverage, or electrical soundings that only consider one-dimensional geometries.

The procedure proposed makes use of two empirical laws between FF-K and TR-T. There are more complex petrophysical models accepted (Soupios *et al.* 2007; Niwas and Celik 2012; Di Maio *et al.* 2015; Farzamian, Monteiro Santos and Khalil 2015; Choo *et al.* 2016), but in general, they were used to evaluate shallow aquifers or the vadose zone. However, for deeper aquifers (50–100m) it is difficult and expensive to acquire basic data for these models, such as granulometric distribution, type of packaging and even clay content.

In the scale of an ERT, there are no significant changes in the water conductivity of the Puelche aquifer (Fig. 3). In the case of an aquifer with significant variations of the water conductivity, this must be taken into account to adjust the empirical law and to apply it, at least as a scale factor (Chandra *et al.* 2008). Also the presence of clay changes resistivity values significantly, since it produces an increase of surface conductivity (Wildenschild *et al.* 2000). In this situation, equation (1) is used to calculate an apparent value for FF (Worthington 1993).

Finally, simulation of a 2D groundwater flow in the vicinity of a pumping well using the hydraulic conductivity field obtained through ERT, would be the next approach to validate these results.

CONCLUSIONS

The general objective of this work was to provide quantitative estimates for hydrogeological parameters from two-dimensional (2D) electrical resistivity measurements. Two empirical relationships between electrical and hydraulic properties were derived from geophysical logging data and pumping test analysis in a semi-confined aquifer in the area of La Plata, in the province of Buenos Aires, Argentina. Before going to the field and acquiring resistivity data with surface electrical resistivity tomography (ERT), a modelling stage was performed. Synthetic data models were tested in order to select the best electrode array that would accurately recover the geometry and true resistivity of the aquifer. A combination of three array types was used to acquire ERT data with overlapping levels to improve data coverage and increase resolution. The resistivity survey was conducted near a well in a high plain area, where the water conductivity of the Puelche aquifer is fairly constant. A high-definition array with 36 electrodes and a minimum separation of 25 m was used. By applying these empirical laws to each cell in the 2D models, it was possible to generate 2D K and T sections. The estimated values in the sections were within the expected range for the aquifer (10–30 m/d for K and 100–400 m^2/d for T). There is a good agreement between values estimated with the ERT near the well. This result sustains the concept that pumping tests estimate an effective hydraulic property in a media with heterogeneities. It was also possible to identify low K or T zones, revealing the hydraulic complexity of the subsurface.

This procedure could be successfully applied to other sedimentary areas with aquifers in which it would be possible to find lateral variations in the hydraulic properties. Finally, the 2D hydraulic sections would improve the hydrogeological models, incorporating a hydraulic parameter estimator as complementary data.

ACKNOWLEDGEMENTS

This work was supported by the Comisión de Investigaciones Científicas de la Provincia de Buenos Aires, Argentina (CIC). The authors would like to thank Martín Pensa, Soledad Lagos, Cecilia Dip, Virginia Teiper, José L. Molina and Joaquín Nigro for their help in the field work. We are grateful to Jean Beaujean for his helpful comments on the manuscript. Authors are acknowledging anonymous reviewers for their constructive suggestions.

REFERENCES

Archie G.E. 1942. The electrical resistivity log as an aid in determining some reservoir characteristics. *Transactions of the American Institute Mining, Metallurgical and Petroleum Engineers* **146**, 54–62.

Auge M. 2005. Hidrogeología de La Plata, Provincia de Buenos Aires. In: *Relatorio del XVI Congreso Geológico Argentino* (eds R.E. de Barrio, R.O. Etcheverry, M.F. Caballé and E. Llambías) pp.293–312. Quick Press, La Plata.

Auge M., Hirata R. and López Vera F. 2004. Vulnerabilidad a la contaminación por nitratos del acuífero Puelche en La Plata Argentina. *Centro de Estudios de América Latina*. Buenos Aires. http://www.bfa.fcnym.unlp.edu.ar/catalogo/doc_num.php?explnum_id=239

Bohling G.C. and Butler J.J. Jr. 2010. Inherent limitations of hydraulic tomography. *Groundwater* **48**, 809–824.

Caminos R. 1999. Geología Argentina. Buenos Aires. Instituto de Geología y Recursos Minerales. SEGEMAR. ISSN 0328–2325.

Carman P.C. 1937. Fluid flow through granular beds. *Transactions of the Institution of Chemical Engineers* **15**, 415–421.

Caterina D., Beaujean J., Robert T. and Nguyen F. 2013. A comparison study of image different appraisal tools for electrical resistivity tomography. *Near Surface Geophysics* **11**, 639–657.

Chandra S., Ahmed S., Ram A. and Dewandel B. 2008. Estimation of hard rock aquifers hydraulic conductivity from geoelectrical measurements: a theoretical development with field application. *Journal of Hydrology* **357**, 218–227.

Chen J., Hubbard S.S. and Rubin Y. 2001. Estimating the hydraulic conductivity at the South Oyster Site from geophysical tomographic data using Bayesian techniques based on the normal linear regression model. *Water Resources Research* **37**, 1603–1613.

Choo H., Kim J., Lee W. and Lee C. 2016. Relationship between hydraulic conductivity and formation factor of coarse-grained soils as a function of particle size. *Journal of Applied Geophysics* **127**, 91–101.

Clennell M.B. 1997. Tortuosity: a guide through the maze. In: *Developments in Petrophysics: Geological Society*, Vol. **122** (eds M.A. Lovell and P.K. Harvey) pp. 299–344. Geological Society, London.

Costa A. 2006. Permeability-porosity relationship: a reexamination of the Kozeny-Carman equation based on a fractal pore-space geometry assumption. *Geophysical Research Letters* **33**, L02318.

Crestani E., Camporese M and Salandin P. 2015. Assessment of hydraulic conductivity distributions through assimilation of travel time data from ERT-monitored tracer tests. *Advances in Water Resources* **84**, 23–36.

deGroot-Hedlin C. and Constable S. 1990. Occam's inversion to generate smooth, two-dimensional models for magnetotelluric data. *Geophysics* **55**, 1613–1624.

Dhakate R. and Singh V.S. 2005. Estimation of hydraulic parameters from surface geophysical methods, Kaliapani Ultramafic complex, Orissa, India. *Journal of Enviromental Hydrology* **13**, 1–11.

Di Maio R., Piegari E., Todero G., and Fabbrocino S. 2015. A combined use of Archie and van Genuchten models for predicting hydraulic conductivity of unsaturated pyroclastic soils. *Journal of Applied Geophysics* **112**, 249–255.

Driscoll F.G. 1989.*Groundwater and Wells*, 2nd edn. Johnson Screens.

Farzamian M., Monteiro Santos F.A. and Khalil, M.A. 2015. Estimation of unsaturated hydraulic parameters in sandstone using electrical resistivity tomography under a water injection test. *Journal of Applied Geophysics* **121**, 71–83.

Fitts C.R. 2002. *Groundwater Science*. Academic Press.

Freeze R.A. and Cherry J.A. 1979. *Groundwater*. Prentice-Hall Inc.

Gomez C.T., Dvorkin J. and Vanorio T. 2010. Laboratory measurements of porosity, permeability, resistivity and velocity on Fontainebleau sandstones. *Geophysics* **75**, E191–E204.

Heigold P.C., Gilkeson R.H., Cartwright K. and Reed P.C. 1979. Aquifer transmissivity from surficial electrical methods. *Groundwater* **17**, 338–345.

Heinz J., Kleinedam S., Teutsch G., and Aigner T. 2003. Heterogeineity patterns of Quaternary glaciofluvial gravel bodies (SW Germany): application to hydrogeology, *Sedimentary Geology* **158**, 1–23.

Hochstetler D.L., Barrash W., Leven C., Cardiff M., Chidichimo F. and Kitanidis P.K. 2016. Hydraulic tomography: continuity and discontinuity of high-K and low-K zones. *Groundwater* **54**, 171–185.

Irving J and Singha K. 2010. Stochastic inversion of tracer test and electrical geophysical data to estimate hydraulic conductivities. *Water Resources Research* **46**, W11514.

Jackson P.D., Taylor Smith D. and Stanford P.N. 1978. Resistivity-porosity-particle shape relationships for marine sands. *Geophysics* **43**, 1250–1268.

Jones P.H. and Buford T.B. 1951. Electric logging applied to groundwater exploration. *Geophysics* **16**, 115–139.

Kelly W.E. 1977. Geoelectric sounding for estimating aquifer hydraulic conductivity. *Groundwater* **15**, 420–425.

Khalil M.A., Ramalho E.C. and Monteiro Santos F.A. 2011. Using resistivity logs to estimate hydraulic conductivity of a Nubian sandstone aquifer in southern Egypt. *Near Surface Geophysics* **9**, 349–355.

Kozeny J. 1927. Ueber kapillare Leitung des Wassers im Boden. *Sitzungsber Akademie der Wissenschaften* **136**, 271–306.

Kozeny J. 1953. *Hydraulik*. Springer-Verlag.

Loke M.H. 1999. RES2DMOD ver 2.2: Rapid 2D resistivity forward modelling using the finite-difference and finite-element methods, Geotomo Software, Penang, Malaysia. (https://www.geotomosoft.com).

Loke M.H. 2006. RES2DINV ver. 3.50, Rapid 2-D resistivity and IP inversion using the least square method, Geotomo Software, Penang, Malaysia. (https://www.geotomosoft.com).

Loke M.H. 2015. Tutorial: 2D and 3D electrical imaging surveys, Geotomo Software, Penang, Malaysia. (https://www.geotomosoft.com).

Loke M.H. and Barker R.D. 1996. Rapid least-squares inversion of apparent resistivity pseudosections by a quasi-Newton method. *Geophysical Prospecting* **44**, 131–152.

Maillet R. 1947. The fundamental equation of electrical prospecting. *Geophysics* **12**, 529–556.

Massoud U., Santos F., Khalil M.A., Taha A. and Abbas A.M. 2010. Estimation of aquifer hydraulic parameters from surface geophysical measurements: a case study of the Upper Cretaceous aquifer, central Sinai, Egypt. *Hydrogeology Journal* **18**, 699–710.

Mazáč O., Kelly W.E. and Landa I. 1985. A hydrogeophysical model for relations between electrical and hydraulic properties of aquifers. *Journal of Hydrology* **79**, 1–19.

Mesgouez A., Buis S., Ruy S. and Lefeuve-Mesgouez G. 2014. Uncertainty analysis and validation of the estimation of effective hydraulic properties at the Darcy scale. *Journal of Hydrology* **512**, 303–314.

Milsch H., Blöcher G. and Engelmann S. 2008. The relationship between hydraulic and electrical transport properties in sandstone: An experimental evaluation of several scaling models. *Earth and Planetary Science Letters*, **275**, 355–363.

Morin R.H. 2006 Negative correlation between porosity and hydraulic conductivity in sand-and-gravel aquifers at Cape Cod, Massachusetts, USA. *Journal of Hydrology* **316**, 43–52. https://doi.org.10.1016/j.jhydrol.2005.04.013

Niwas S. and Celik M. 2012. Equation estimation of porosity and hydraulic conductivity of Ruhrtal aquifer in Germany using near surface geophysics. *Journal of Applied Geophysics* **84**, 77–85. https://doi.org.10.1016/j.jappgeo.2012.06.001

Niwas S. and Singhal D.C. 1981. Estimation of aquifer transmissivity from Dar-Zarrouk parameters in porous media. *Journal of Hydrology* **50**, 393–399.

Niwas S., Tezkan B. and Israil M. 2011. Aquifer hydraulic conductivity estimation from surface geoelectrical measurements for Krauthausen test site, Germany. *Hydrogeology Journal* **19**, 307–315.

Purvance D.T. and Andricevic R. 2000. On the electrical-hydraulic conductivity correlation in aquifers. *Water Resources Research* **36**, 2905–2913.

Revil A. and Cathles L.M. III 1999. Permeability of shaly sands. *Water Resources Research* **35**, 651–662.

Revil A., Karaoulis M., Johnson T. and Kemna A. 2012. Review: Some low-frequency electrical methods for subsurface characterization and monitoring in hydrogeology. *Hydrogeology Journal* **20**, 617–658.

Rubin Y. 2003. *Applied Stochastic Hydrogeology*. Oxford University Press.

Ruggeri P., Gloaguen E., Lefebvre R., Irving J. and Holliger K. 2014. Integration of hydrological and geophysical data beyond the local scale: application of Bayesian sequential simulation to field data from the Saint-Lambert-de-Lauzon site, Québec, Canada. *Journal of Hydrology* **514**, 271–280.

Ruggeri P., Irving J., Gloaguen E. and Holliger K. 2013. Regional–scale integration of multiresolution hydrological and geophysical data using a two-step Bayesian sequential simulation approach. *Geophysical Journal International* **194**, 289–303.

Shevnin V., Delgado-Rodriguez O., Mousatov A. and Ryjov A. 2006. Estimation of hydraulic conductivity on clay content in soil determined from resistivity data. *Geofísica Internacional* **45**, 195–207.

Singh K.P. 2005. Nonlinear estimation of aquifer parameters from surfficial resistivity measurements. *Hydrology and Earth System Sciences Discussions* **2**, 917–938.

Sinha R., Israil M. and Singhal D. C. 2009. A hydrogeophysical model of the relationship between geoelectric and hydraulic parameters of anisotropic aquifers. *Hydrogeology Journal* **17**, 495–503.

Slater L. 2007. Near surface electrical characterization of hydraulic conductivity: from petrophysical properties to aquifer geometries - A review. *Surveys in Geophysics* **28**, 169–197.

Soupios P.M., Kouli M., Valianatos F., Vafidis A. and Stavroulakis G. 2007. Estimation of aquifer hydraulic parameters from surficial geophysical methods: a case study of Keritis Basin in Chania (Crete-Greece). *Journal of Hydrology* **338**, 122–131.

Taheri Tizro A., Voudouris K. and Basami Y. 2012. Estimation of porosity and specific yield by application of geoelectrical method

– A case study in western Iran. *Journal of Hydrology* **454–455**, 160–172.

Theis C.V. 1935 The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using groundwater storage. *Transactions American Geophysical Union* **16**, 519–524.

Urish D.W. 1981. Electrical resistivity-hydraulic conductivity relationships in glacial outwash aquifers. *Water Resources Research* **17**, 1401–1408.

Wildenschild D., Roberts J.J., and Carlberg E.D. 2000. On the relationship between microstructure and electrical and hydraulic properties of sand-clay mixtures. *Geophysical Research Letters* **27**, 3085–3088.

Worthington P.F. 1993. The uses and abuses of the Archie equations, 1: the formation factor-porosity relationship. *Journal of Applied Geophysics* **30**, 215–228.

Yeh T.-C. J. and Liu S. 2000. Hydraulic tomography: development of a new aquifer test method. *Water Resources Research* **36**, 2095–2105.