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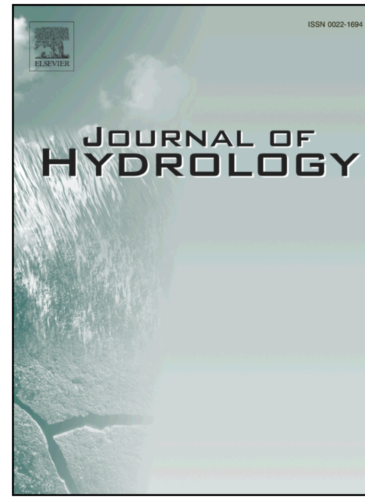

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Hydrodynamic parameters estimation from self-potential data in a controlled full scale site.

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Abstract

A multi-physical approach developed for the hydrodynamic characterization of porous media using hydrogeophysical information is presented. Several pumping tests were performed in the Hydrogeosite Laboratory, a controlled full-scale site designed and constructed at the CNR-IMAA (Consiglio Nazionale delle Ricerche – Istituto di Metodologia per l'Analisi Ambientale), in Marsico Nuovo (Basilicata Region, Southern Italy), in order to obtain an intermediate stage between laboratory experiments and field survey. The facility consists of a pool, used to study water infiltration processes, to simulate the space and time dynamics of subsurface contamination phenomena, to improve and to find new relationship between geophysical and hydrogeological parameters, to test and to calibrate new geophysical techniques and instruments. Therefore, the Hydrogeosite Laboratory has the advantage of carrying out controlled experiments, like in a flow cell or sandbox, but at field comparable scale. The data collected during the experiments have been used to estimate the saturated hydraulic conductivity k_s [ms^{-1}] using a coupled inversion model working in transient conditions, made up of the modified Richards equation describing the water flow in a variably saturated porous medium and the Poisson equation providing the self-potential ϕ [V], which naturally occurs at points of the soil surface owing to the presence of an electric field produced by the motion of underground electrolytic fluids through porous systems. The result obtained by this multi-physical numerical approach, which removes all the approximations adopted in previous works, makes a useful instrument for real heterogeneous aquifer characterization and for predictive analysis of its behavior.

Keywords: Porous medium characterization; hydrogephysics; multi-physical inverse modeling; self-potential; pumping test.

1. Introduction

A key problem in aquifers characterization is the requirement of a large number of direct and high resolution measurements of system parameters and state variables. Measurements of state variables typically include the drawdown induced in observation boreholes by pumping tests. This method is intrusive and consequently the hydrogeological system is perturbed by drillings. On the contrary, geophysical approaches aim at using non-intrusive techniques to obtain a large amount of information on the subsurface and with moderate costs.

In the last few decades hydrologists have increasingly resorted to hydro-geophysical information to estimate groundwater flow parameters (Cassiani and Medina, 1997; Cassiani et al.,1998; Jardani and Revil, 2009; Straface et al., 2010; Troisi et al., 2000). A key innovation in this field is based on the passive measurements of the self-potential generated by (1) ground water flow through a so-called electrokinetic coupling (Allègre et al., 2014; Aubert and Yéné Atangana, 1996; Birch, 1993, 1998; Darnet et al., 2003; Fournier, 1989; Ishido and Mizutani, 1981; Jouniaux et al., 1999; Pozzi and Jouniaux, 1994; Revil and Leroy, 2001) and (2) electro-chemical processes associated with gradients of the chemical potentials of charge carriers (ionic species and electrons) in the pore water (e.g., Maineult et al. (2006); Naudet et al., 2004; Revil et al., 2005; Straface and De Biase, 2013). A general theory of all these effects was developed by Pride (1994), Revil and Linde (2006) and Revil et al. (2009). In the self-potential approach, the measured response is a function of an unknown source of electrical currents and resistivity structure. Therefore, there is an inherent ambiguity when interpreting the self-potential data when the earth resistivity is unknown. This difficulty can be solved upon merging different kinds of measurements obtained from other non-intrusive techniques, i.e. DC and EM-based electrical resistivity tomography and (Jardani and Revil, 2009) or borehole data (Straface et al., 2010). Interpretation schemes that could model the self-potential field recorded during pumping tests were developed (Jardani et al., 2009; Malama et al., 2009a, b; Revil et al., 2003; Rizzo et al., 2004; Soueid Ahmed et al., 2014; Straface et al., 2007a; Titov et al., 2002; 2005). Data from several pumping tests can be combined and used in a tomographic fashion to characterize heterogeneity in the aquifer (for recent field examples see, Bohling et al., 2007; Cardiff et al., 2009; Li et al., 2007; Straface et al., 2007b). Straface et al. (2007a) used experimental hydraulic heads and self-potential signals associated with a pumping test in an inverse model based on the Successive Linear Estimation (SLE) (Yeh et al., 1996), to estimate the transmissivity distribution of a small-scale aquifer. Bianchi Jannetti et al. (2010) extended the moment equations-based inverse method of Hernandez et al. (2003, 2006) to quasi-steady state flow conditions and presented its first application by using hydraulic heads and self-potential signals collected during a pumping test at the Montalto Uffugo research site. In these works the authors used self-potential signals in a two-dimensional inverse modeling. Nevertheless, the self-potential method is able to locate the spatial distribution of electrical sources in the earth generated by the coupling electrokinetic mechanism. In other words, it provides a 3D information estimate of the free surface location unlike borehole readings which provides a depth-averaged hydraulic head. In fact, even though a pumping test generates a velocity field in the aquifer, water inside an observation well is approximately at rest (or the Dupuit assumption can be considered to be locally valid). This means that pseudo-hydrostatic pressure conditions are established in the water column contained by the well. Therefore, the hydraulic head does not depend on the transducer vertical position, but attains a constant value, representing a depth averaged quantity in that location. On the other hand, the self-potential source inversion problem is highly non-unique. There are several possible distributions of the sources that can fit the data equally. This dilemma is common to nearly all geophysical inverse problems, but it is even more crucial for passive potential field techniques (Jardani et al., 2008). Jardani et al. (2013) have recently proposed a remedy to this problem, adopting a fully coupled inversion approach. Moreover, as with all geophysical techniques, data errors degrade our ability to interpret the measured signal. Common sources of self-potential measurement noise can be associated with the degradation or drift of the measuring nonpolarizable electrodes, poor contact between the electrode and soil, cultural noise (Clerc et al., 1998; Corwin, 1973; Perrier et al., 1997; Petiau, 2000) and Haines jumps electrical disturbance, occurring during drainage experiments (Haas and Revil, 2009). As a consequence, joint inversion of self-potential and borehole-based head data requires estimating the relative weight of the different data types adopted (Bianchi Jannetti et al., 2010). In this context, a relevant question is how these two types of measurements can be combined within a three-dimensional inverse modeling approach. Some authors have recently worked on this issue, focusing on synthetic case studies operating in steady state conditions (Soueid et al.,2014), or on analytical solutions requiring simplifying assumptions to solve the involved equations (Malama, 2014). The innovation presented with this study, instead, is the characterization of a variably saturated, unconfined aquifer through the inversion of hydrogeophysical experimental data in transient flow conditions. The theory and the model we developed are both of general nature and therefore applicable in problems involving heterogeneous and anisotropic media, even though the setup adopted for the experimental apparatus is homogeneous and isotropic. The choice of this configuration, is only a starting point for the validation of such a complex modeling approach, as it also deals with real experimental data. The hydrogeophysical experiment was carried out at the Hydrogeosite Laboratory at the Istituto di Metodologie per l'Analisi Ambientale (IMAA) of the Consiglio Nazionale delle Ricerche (CNR) site in Marsico Nuovo (Potenza, Italy). We illustrate the first results of the characterization effected using the hydraulic and geophysical measurements recorded during the pumping tests in the Hydrogeosite Laboratory. The results open new perspectives, owing to the great increase in information, regarding the possibility of improving the characterization of the real heterogeneous porous media.

2. Theoretical background

The grain of a porous medium in contact with a fluid, develops an electrical charge on the interface between the grain and the fluid. As a result, of the proton exchange and the sorption of cations and anions onto its surface, a triple layer configuration is formed: 1) a fixed charge of density (q_0) occurring at the mineral surface, 2) the Stern layer produced by the sorption of ions onto the mineral surface (q_S) and finally 3) a diffuse layer of anions and cations of the bulk pore water attracted or repelled according to the sign of the charge (q_b) (Davis et al., 1978; Revil et al., 2003). When the water flows within a porous medium, the excess of ions in the diffuse electrical layer are transported downstream, producing an excess of net charge and then an electric field parallel to the flow direction. This effect is called electrokinetic coupling (Briggs, 1928) and was observed for the first time by Reuss (1809).

The water flow is due to two components, a main one owing to the hydraulic gradient and a secondary one, generally of lower order than the former, to the natural electric potential. The general constitutive equation of the groundwater flow in a variably saturated aquifer is (Straface and De Biase, 2013)

$$\boldsymbol{u} = -L(S_w)\nabla\phi - k_r(S_w)\,k_s\nabla h \tag{1}$$

where $\boldsymbol{u}$ is the Darcy velocity [ms^{-1}], L the coupling electrokinetic term [m^{2}V^{-1}s^{-1}], S_w represents the saturation degree, ϕ is the electric potential [V], k_r is the relative hydraulic conductivity [dimensionless], $k_s = k_0\rho g/\mu$ is the saturated hydraulic conductivity [ms^{-1}], in which k_0 is the intrinsic permeability [m^2], ρ is the density of the fluid, g is the gravitational acceleration [ms^{-2}], μ is the dynamic viscosity of the fluid [Pa m] and finally, h is the hydraulic head [m]. The term L is set equal to $\dfrac{\varepsilon_f\zeta}{\mu F}$ with ε_f the dielectric constant of the porous medium [Fm^{-1}], ζ the zeta potential [V] , i.e. the electrical potential located at the shear plane where the relative velocity between the deformable solid and the pore water is zero, and F the formation factor which can be defined, when the surface conductivity is negligible, by the ratio of the saturated rock resistivity to the resistivity of the saturating water. When the component, owing to the electric potential, is negligible compared to that due to the hydraulic potential, the average velocity $\boldsymbol{u}$ is only represented by the mechanical component of the Darcy velocity

$$\mathbf{u} = -k_r(S_w)\, k_s \nabla h \tag{2}$$

For $S_w=1$, equation (2) coincides with the classical Darcy's law. The relative hydraulic conductivity k_r and the saturation degree S_w are assumed to be represented by the Gardner model (Gardner, 1958; Russo, 1988)

$$k_r = \mathrm{e}^{-\alpha_G \psi} \tag{3}$$

$$S_e = \left[\left(1 + 0.5\alpha_\mathrm{G}|\psi|\right)\mathrm{e}^{-0.5\alpha_\mathrm{G}|\psi|}\right]^{\frac{2}{l+2}} \tag{4}$$

in which α_G is a pore-size distribution parameter [m^{-1}] and $\psi = p/(\rho\, g)$ is the capillary pressure head [m], l is a parameter that accounts for the dependence of the tortuosity and the correlation factors on the water content, estimated to be about 0.5 as an average for many soils (Mualem, 1976), $S_e = (S_w - S_r)/(1 - S_r)$ is the effective dimensionless reduced water content and S_r is the residual (irreducible) saturation degree.

In an isotropic material, the electrical current density $\boldsymbol{J}$ [Am^{-2}] is defined by the sum of an electrical current due to the electric field and another one due to the groundwater flow. The equation that describes the total electrical current density in porous media is the following (Straface et al., 2010)

$$\boldsymbol{J} = -\sigma(S_w)\nabla\phi - \mathrm{L}(S_w)\,\rho g \nabla h \tag{5}$$

Equation (5) indicates that the electrical current density $\boldsymbol{J}$ is furnished by two terms: the first is the electrical and the second is the electrokinetic term. In the first term σ indicates the electrical conductivity of the porous medium [Sm^{-1}] and ϕ the self-potential [V]. Using the relationship between the formation factor F and the water content θ proposed by Archie (1942) and extending it on the unsaturated case, it is possible to define the electrical conductivity as

$$\sigma(S_w) = \sigma_w \theta^m S_w^n = \sigma_{sat} S_w^n \tag{6}$$

with m and n the cementation and saturation Archie's exponent respectively and σ_w the water electrical conductivity [Sm^{-1}]. Using equation (2), the constitutive equation of the electrical flow becomes

$$\boldsymbol{J} = -\sigma(S_w)\nabla\phi - \frac{\mathrm{C}(S_w)\sigma(S_w)\rho g}{k_r(S_w)k_s}\boldsymbol{u} \tag{7}$$

knowing that C is the classical electrokinetic coefficient [VPa^{-1}] (named classical streaming potential coefficient too). In general, it is defined as

$$C(S_w) = \frac{\partial \phi}{\partial p} = -\frac{L(S_w)}{\sigma(S_w)} \tag{8}$$

Perrier & Morat (2000) proposed that $C(S_w)$ varies as a function of the relative hydraulic conductivity k_r as

$$C(S_w) = C_{sat} \frac{k_r(S_w)}{S_w^n} \tag{9}$$

where C_{sat} is the classical saturated electrokinetic coefficient.

Regarding the groundwater flow in transient conditions, when the component due to the electric potential is negligible compared to that due to the hydraulic head, the continuity equation of an incompressible fluid in a variably saturated medium is the Richards equation (Panday et al, 1993; Richards, 1931)

$$\left(S_w S_s + \frac{dS_w}{dh}\right)\frac{\partial h}{\partial t} - \nabla \cdot \left(k_r(S_w) k_s \nabla h\right) = Q_s \tag{10}$$

where S_s is the specific storage coefficient [m^{-1}] and Q_s is a specific source term [s^{-1}]. The term dS_w/dh is a function called specific moisture capacity and could be determined for different soil types using curve fitting and laboratory experiments measuring the rate of infiltration of water into the soil column.

Regarding the electrical field, taking in account equation (9), the continuity equation for the charge in the low frequency limit of the Maxwell equations valid for a variably saturated aquifer is

$$\nabla \cdot \left(\sigma(S_w) \nabla \phi\right) = -\nabla \cdot \left(\frac{C'_{sat}\sigma(S_w)}{k_s}\right)\frac{\boldsymbol{u}}{S_w^n} \tag{11}$$

where $C'(S_w) = C(S_w)\rho g$ is the streaming potential coefficient [$V\ m^{-1}$], which turns into the following expression in the saturated zone $C'_{sat} = C_{sat}\rho g$. Equation (11) is referred to as Poisson equation and its solution is the self-potential field.

Once the spatial distributions of the hydraulic head h and the self-potential signals ϕ are known it is possible to estimate the hydraulic conductivity and the storage coefficient field by inverting the coupled Richards-Poisson model.

3. Materials and methods

3.1. The Hydrogeosite Laboratory

The Hydrogeosite Laboratory is a controlled site designed and constructed at the research institute CNR-IMAA in Marsico Nuovo (Basilicata region, Southern Italy). It consists of a pool (10x7x3 m^3) completely covered with a steel shed. Figure 1 shows a close representation of the laboratory: the tank is externally examinable because it is completely surrounded by (A) a corridor placed under the pattering level. This arrangement, allows an uninterrupted supervising of the system behavior during experiments and testing. The interior of the tank is completely covered with a fine impermeable film and each vertical face is provided with access points making up an overall system of 64 close holes, in which it is possible to take samples and/or to install sensors without disturbing the soil surface. The large pool was designed to reproduce an artificial phreatic aquifer for hydrogeophysical experimentation. For this reason it was filled with homogeneous quartz-rich sand (95% of SiO2) with a high percentage (86.4%) of grains between 0.063mm and 0.125mm and a medium-high permeability (in the order of 10^{-5} m/s). A gravel filter, covered with geo-textile, was placed on the bottom to permit a controlled discharge of water. The filling operations were accompanied by the placement of (B) 17 boreholes with a 8 cm diameter and (C) a 15 cm well, created with corrugated drainpipes in PET (Polyethylene terephthalate), completely penetrating and covered with geo-textile to stop the sand entering. The key phase for hydrogeophysical experimentation was the creation of the hydraulic system transforming the pool into a phreatic aquifer. The aim of this hydraulic system is the maintenance of the constant hydraulic head boundary conditions. What we wanted to gain was the lateral recharge phenomenon that naturally happens within an aquifer, when this latter is subjected, for example, to a perturbation induced by a pumping well. An infinite water source was simulated, in the Hydrogesite laboratory, to obtain this kind of response. The constant hydraulic head, on pool boundaries, was obtained by (D) two loading reservoirs placed on two opposite sides of the pool itself. These reservoirs determine the saturation of the sand to a desired height through a connection to (E) a “ring” of draining pipes settled along the inside perimeter of the pool. The draining “ring” is divided into four parts, one along each side of the pool, and made up of the same type of PET corrugated drainpipes which constitute well and piezometers. Every loading reservoirs supplies water to two parts of the “ring”, one settled on a short side and one on a long side of the pool. Both these parts are supplied with a valve system which allows the supply of one or both the segments. The loading system is governed by a floating device, which maintains a constant water level inside the reservoir by allowing or not water entrance according to how the aquifer answers to the induced perturbations. This system, according to its characteristics, allows the execution of pumping tests or the generation of a natural water flow by imposing a hydraulic head difference between the reservoirs.

FIGURE_1_APPROXIMATELY_HERE

3.2. The experiments

Several pumping tests were performed, in the Hydrogeosite laboratory. This section presents the experiment whose data have been used in the multi-physical inversion. A peristaltic pump was used in order to produce head changes, in the well, that were well outside the range of instrument error. The peristaltic pump was chosen for many reasons: the large flow rates range (from 1 micro-litre/min to 18 litre/min) which allows steady state conditions to be reached with different settings of the loading system; the dimension of well and piezometers which did not allow the entrance of a submersible pump; the electrical engine which is external and far from the geophysical acquisition zone and so it did not put noise in the system. Flow rates, during pumping tests, were measured by an electromagnetic flow-meter provided with a micro-processor electronic signal converter. Hydraulic head variations were monitored through 7 high precision pressure transducers placed in as many boreholes. The facility was filled from the bottom to avoid the formation of air bubbles and using the drainage ring a (initial) hydraulic head was fixed at 2.80 m above the bottom of the pool. From a geophysical point of view, 63 nonpolarizable Petiau electrodes (SDEC, France) were installed on the surface of the model according to the grid with regular meshes shown in Figure 2.

FIGURE_2_APPROXIMATELY_HERE

A multichannel voltmeter (Keithley, model 2700) was adopted for self-potential measurements and acquisition. The flow rate pumped from the well was fixed at 0.085 l/sec. Figure 3 shows the hydraulic head variation, with respect to time, measured using the pressure transducers located in some boreholes. Figure 4 shows the self-potential curves acquired, with time, in two channels during the pumping test.

FIGURE_3_APPROXIMATELY_HERE

FIGURE_4_APPROXIMATELY_HERE

The water used during the experimental phase was tap water from the town water supply, its electric conductivity was continuously measured by means of a conductivimeter, and its value, practically constant, compensated to 25 °C (σ_w = 305 [µS/cm]). The sand saturated water content θ_S was determined using the liquid saturation method. A known volume of sand, previously dried in the oven at 105 °C for one hour, was weighted using a precision scale. The sample was then water saturated and weighted again, to determine, by difference, the weight of the water within the sand pores and so its volume. The ratio between the water volume and the sample volume provided a value for the saturated water content of 0.48. The cementation and saturation Archie's exponents were estimated by combining Electrical Resistivity Tomography (ERT) and Time-Domain Reflectometry (TDR) techniques. For this purpose, the tank was water saturated up to a given height, and an ERT, with an electrodes grid with a 50 cm mesh, was performed. Three TDR rods were then placed at different depths from the top soil, without changing the tank saturation conditions. The rods were arranged at a distance of 25 cm from each other, and the water content θ was measured at these points. The estimation of the Archie's exponents was then possible using equation (6) described in Section 2, which defines the electrical conductivity as a function of the water content and the saturation degree. Knowing the spatial distribution of the electrical conductivity from the ERT interpretation , the m exponent can be calculated by selecting the value of σ derived in the saturated zone. In this zone the electrical conductivity, for a homogeneous medium, is essentially constant and its variation law becomes $\sigma(S_w = 1) = \sigma_w \theta_S^m$, in which m is the only unknown (m =1.55). Moving in the unsaturated zone , a σ value can be read, from the tomographic map, at those points where the water content θ was measured by the TDR rods. Equation (6) assumes the general form $\sigma(S_w) = \sigma_w \theta^m S_w^n$, in which $S_w = \theta / \theta_S$ and n represents the only unknown. Three n values were calculated, one for each TDR measuring point, and the average value was finally adopted (n =1.82). Moreover in situ water infiltration tests were performed in three different locations of the soil surface of the Hydrogeosite laboratory, in order to find out the value of the Gardner parameter. The estimation of this parameter was obtained by means of a tension infiltrometer (Eijkelkamp) designed to measure the unsaturated flow of water into the soil. Several tensions were imposed by the infiltrometer on the topsoil and a corresponding number of infiltration rates q were measured. According to Wooding (1968), the α_G value was obtained, for each location, as the slope of the linear regression of the q natural logarithm versus the applied pressure ψ. Because of the homogeneity of the porous medium, an average value was finally determined ($\alpha_G = 0.45$ [m^{-1}]). Assuming the α_G estimates to be normally distributed and an upper/lower confidence limits of 95% the uncertainty in the evaluation of this parameter is ± 0.0848 [m^{-1}].The relative hydraulic conductivity and the capillary pressure head of the sand are shown, in Figure 5, as function of the saturation degree.

FIGURE_5_APPROXIMATELY_HERE

3.3.The modeling

The coupled inversion approach, was obtained through the implementation of an iterative procedure aimed to minimize the deviation between experimental and simulated data. A three-dimensional finite element model of the pool working in transient conditions, was developed. The tetrahedral grid of the finite elements has the following specifications: a minimum element size of 0.1 m was used at the observation points, in which hydraulic and self-potential data were monitored and in correspondence to the pumping point, where a high calculation precision is needed. The size of the tetrahedral mesh increases, moving from the aforementioned points, with a maximum element growth rate of 1.45 up to a maximum element size of 0.8 m. The draining pipes ring was recreated and both constant head and zero electrical potential boundary conditions were set on it. No-hydraulic flow and electrical insulation conditions were adopted for the walls and the bottom of the pool. The artificial aquifer was modeled as a homogeneous and isotropic medium. The sand saturated water content, determined by means of laboratory tests, the initial hydraulic head distribution and the flowrate value recorded during the experiment, were provided, together with the beginning trial values of the parameters we want to estimate (i.e. a value of 1×10^{-5} [ms^{-1}] for the saturated hydraulic conductivity, 5×10^{-3} [m^{-1}] for the specific storage coefficient and -10 [mV/m] for the saturated streaming potential coefficient). The mathematical model, in which the Gardner pedofunctions were included (see eq. 3 and 4 in section 2), is then able to calculate the relative hydraulic conductivity k_r and the saturation degree S_w as functions of the pore-size distribution parameter α_G (average value experimentally determined), and the pressure head ψ (model calculated). The hydraulic head distribution and the field velocity are then calculated by means of Richards and Darcy equations respectively. Using the relation derived from the Archie's law(s) the electrical conductivity σ can be evaluated as a function of the saturation degree S_w, the water electric conductivity σ_w and m and n the cementation and saturation Archie's exponents. The self-potential field is then calculated throughout the pool by means of the Poisson equation. The hydraulic head and self-potential values evaluated in the model observation points, representing the monitored boreholes and electrodes of the real apparatus, are compared with the experimental values. The procedure changes the values of the parameters to be estimated until the deviation between calculated and observed values appears to be less or equal to a previously chosen threshold value. The threshold value represents the optimality tolerance used by the solver to determine whether optimality has been reached. The evaluation of the optimality tolerance parameter is related to possible convergence problems occurring during the optimization. As an example, if the optimization procedure reports a converged solution after just a few iterations, it is better to restart it with a tighter tolerance to make sure it has actually found the solution. If, on the contrary, it seems to iterate forever, despite the value of the objective function having converged, chances are that the tolerance value is too strict. The optimization algorithm is the Spars Nonlinear OPTimizer (SNOPT) (Gill et al., 2005). The SNOPT solver uses a gradient-based optimization technique to find optimal solutions to a very general class of optimization problems. It requires gradients of both the objective function and all constraints, which are computed using numeric differentiation (COMSOL, 2008). The coupled model, in conclusion, enables the conditioning of the hydraulic and/or the electrical parameters of the system, through the inversion of the observed hydraulic head and self-potential signals or by inverting just the latter. The approach adopted and discussed below, is the one regarding only the inversion of self-potential data, which is described after the presentation of the inversion results coming from a more "classical" methodology pertaining only to the hydraulic component of the experiment and the modeling.

3.4. Cross Validation Analysis

In order to validate the results obtained by the multi-physical modeling approach on the inversion of the self-potential data, a Cross Validation Analysis (CVA) is adopted. This model validation technique is used to assess how the results of a fitting process are affected or not by an estimation error showing a particular spatial trend. The way this analysis is performed is described below. The self-potential variation acquired with time in one electrode, is excluded from the experimental dataset and the inversion procedure is performed on these latter. The estimated parameters are then used to generate the self-potential curve in the point not included in the inversion. This curve is compared with the experimental one and the estimation error is evaluated. The procedure is then repeated by including again the data previously excluded, and by eliminating from the observed dataset the ones acquired on another electrode. The inversion procedure is executed several times in this way, one for each electrode of the complete dataset. The results of the CVA can be considered as validated when: (a) the distribution of estimation error is symmetrical, centered on a zero mean value and has minimum spread, (b) the plot of the estimation error versus the estimated value is centered around the zero error line (conditional unbiasedness property) and (c) the estimation error does not show any spatial correlation.

4. Results and discussion

4.1. Parameters estimation based on hydraulic head data

The first modeling stage is represented by the inversion of the hydraulic data only. In this phase the Richards model was performed in transient conditions, to estimate the values of both the saturated hydraulic conductivity and the storage coefficient of the homogeneous porous medium. The drawdown curves recorded in all the monitored boreholes were used to gain the aforementioned objective. A value of 3.25×10^{-5} [ms^{-1}] for k_s and 7.83×10^{-3} [m^{-1}] for S_S, was estimated (Fig. 6 and 7). These results will be used as a benchmark for the coupled inversion approach.

FIGURE_6_APPROXIMATELY_HERE

FIGURE_7_(WHOLE PAGE)_APPROXIMATELY_HERE

4.2. Parameters estimation based on self-potential data

This section shows the inversion of the self-potential data only through the coupled Richards-Poisson model. An optimum value of the saturated hydraulic conductivity, the storage coefficient and the saturated streaming potential coefficient were estimated. The first one, k_s, turned out to be equal to 2.90×10^{-5} [ms^{-1}], whereas S_S assumed a value of 7.99×10^{-3} [m^{-1}] and finally C'_{sat} = -8.21 [mV/m]. The values of the hydrodynamic parameters are very close to the ones previously estimated by the Richards model using only the hydraulic data. The streaming potential coefficient is also very close to the one estimated by Straface et al. (2010) for the same experiment, by using a methodology based on the Kriging with External Drift (KED). This approach, adopting the strong assumption of a linear relation between hydraulic head values and self-potential signals (Revil et al., 2002, 2003), returned a value of C'_{sat} = -7.4 [mV/m]. The good results comparison, obtained from the different modeling approaches, is a first validation stage for the coupled inversion model developed in this study. Figure 2 shows the position of the observation points (red circled) adopted for both the models: 7 boreholes for the Richards model and 16 electrodes for the fully coupled Richards-Poisson model. Not all the acquiring electrodes gave back a clear signal. The ones used for the inversion, showed a legible trend (a high "signal to noise" ratio) during the transient phase of the pumping test, the remaining ones are characterized, at this stage, by a noisy signal, so that it was decided to not include them in the modeling (Fig. 2). A better signal sharpness is required during the transient phase to perform the proper filtering operations to isolate exclusively the information coming from the controlled perturbation produced in the system. This issue is far less pronounced during the steady state condition owing to which there are several measures concerning the same perturbation state, resulting in an easier filtering operation, even for a noisy signal. In this case, the clear interpretation of the transient behavior of the system may have been affected by: 1) particular flow conditions for those electrodes close to the long sides of the pool (proximity to the draining pipes ring); 2) poor electrical contact between some electrodes and the soil and 3) wear and tear of some others electrodes employed in several field tests. The observation points used for the inversion are indeed represented by the newer electrodes placed, at least, one meter far from the pool edges and evidently creating a good electrical contact with the soil.

FIGURE_8_APPROXIMATELY_HERE

FIGURE_9_(WHOLE PAGE)_APPROXIMATELY_HERE

The results obtained by the multi-physical modeling approach were validated through a Cross Validation Analysis (CVA), described in Section 3.4, Figure 10 shows the good results obtained for the CVA and consequently for the inversion procedure, in fact, the distribution of the estimation error is symmetrical, centered on a zero mean value and has minimum spread (Figure 10a), the plot of the estimation error versus the estimated value is centered around the zero error line (Figure 10b) and the estimation error does not show any spatial correlation, i.e. the errors are independent of each other (Figure 10c).

FIGURE_10 – A,B and C_APPROXIMATELY_HERE

5. Conclusions

It is well known that measurable electrical signals can be recorded at the ground surface during a pumping test experiment including the relaxation phase of the phreatic surface after the pump shutdown. The electrical signals can be understood by solving the coupled hydroelectric problem. A new interpretative methodology is presented in this study for hydraulic and self-potential data coupling. The proposed methodology is based on the Richards-Poisson coupled model, which is able to return the estimation of the hydraulic parameters and the streaming potential coefficient of a porous medium by analyzing the experimental data acquired during a pumping/injection test. The value of this important coefficient is comparable with the one estimated, for the same experiment and a different modeling approach, by Straface et al. (2010). This result was obtained by a multi-physical numerical approach, which removes all the approximations adopted in previous works. The innovation presented here, resides in the design of a theory and a model, both of general nature and therefore applicable in problems involving variably saturated, heterogeneous and anisotropic media, capable of handling transient hydraulic and self-potential data. Although the setup adopted for the experimental apparatus is homogeneous and isotropic, this configuration is only a starting point for the validation of such a complex modeling approach. It can be stated that the implemented procedure can be a useful instrument for real heterogeneous aquifer characterization, predictive modeling and decision making. Considering the encouraging results obtained in this study, new experiments involving a heterogeneous porous medium are planned and also, regarding the modeling side, different assumptions for k_r and C', expressing their dependence on the saturation degree, will be tested in future works.

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Figure 1. Sketch of the facility of the Hydrogeosite Laboratory: (A) tank inspection corridor, (B) boreholes, (C) well, (D) loading reservoirs and (E) draining pipes ring.

Figure 2. Top view of the tank with the spatial disposition of the hydraulic and geophysical equipment in its complete configuration. The observation points adopted for the Richards model (boreholes) and for the Richards-Poisson model (electrodes), are red circled.

Figure 3. Hydraulic head variation observed in three boreholes placed at different distances from the pumping well.

Figure 4. Self-potential signals observed in channels 21 and 52 placed at different distances from the pumping well.

Figure 5. Relative hydraulic conductivity and capillary pressure head curves. The curves are plotted using the Gardner's pedofunctions (equations 3 and 4 in section 2), with a pore-size distribution parameter α_G = 0,45 [m^{-1}] (experimentally determined) and a *l*, accounting for the dependence of the tortuosity and the correlation factors on the water content, estimated to be 0.5 for many soils (Mualem, 1976)

Figure 6. Scattergraph between Observed and Calculated hydraulic heads.

Figure 7. Curve fitting between observed and calculated hydraulic head variations, at the end of the inversion procedure.

Figure 8. Scattergraph between Observed and Calculated self-potential signals.

Figure 9. Curve fitting between observed (red dots) and calculated (black line) self-potential variations in the electrodes, at the end of the inversion procedure.

Figure 10. Cross validation: a) Graph of the distribution of estimation error; b) Plot of the estimation error versus the estimated value; c) Plot of the covariance function of the estimation errors versus the distance.

Figure 1

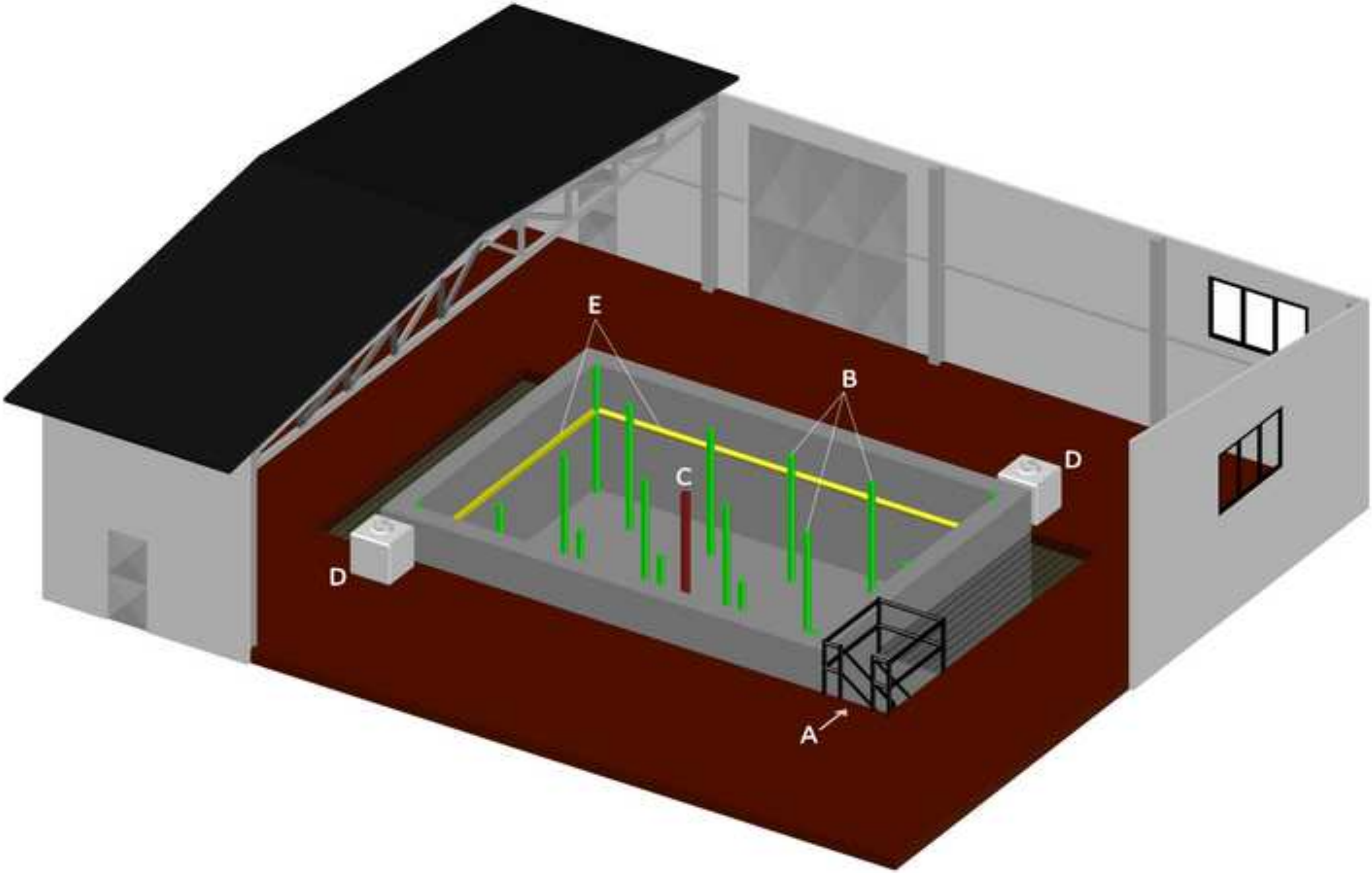
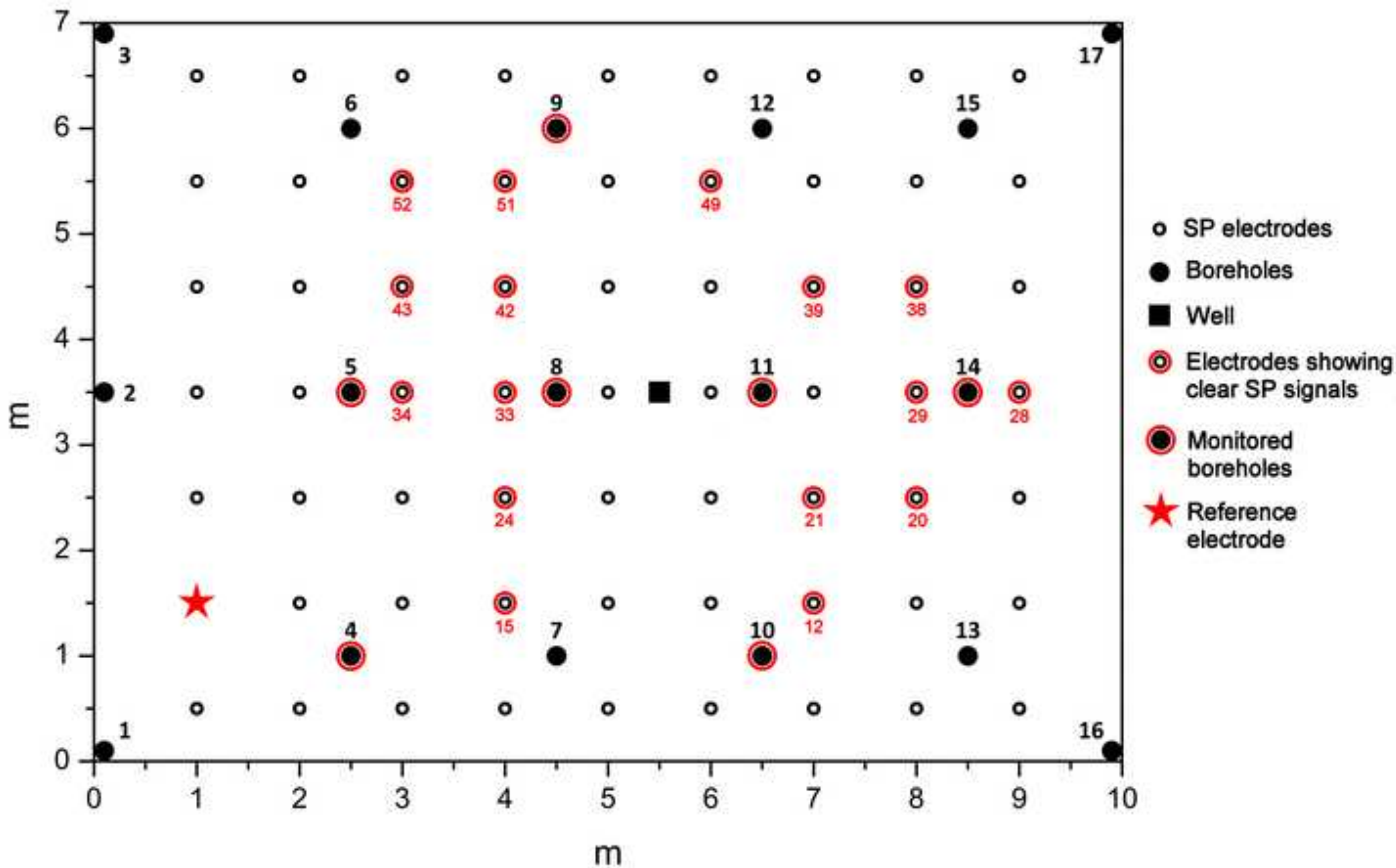


Figure 2



7
6
5
4
3
2
1
0
m
0
1
2
3
4
5
6
7
8
9
10
m
1
2
3
4
5
6
7
8
9
10
11
12
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14
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16
17
52
51
49
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42
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38
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24
21
20
15
12
SP electrodes
Boreholes
Well
Electrodes showing clear SP signals
Monitored boreholes
Reference electrode

Figure 3

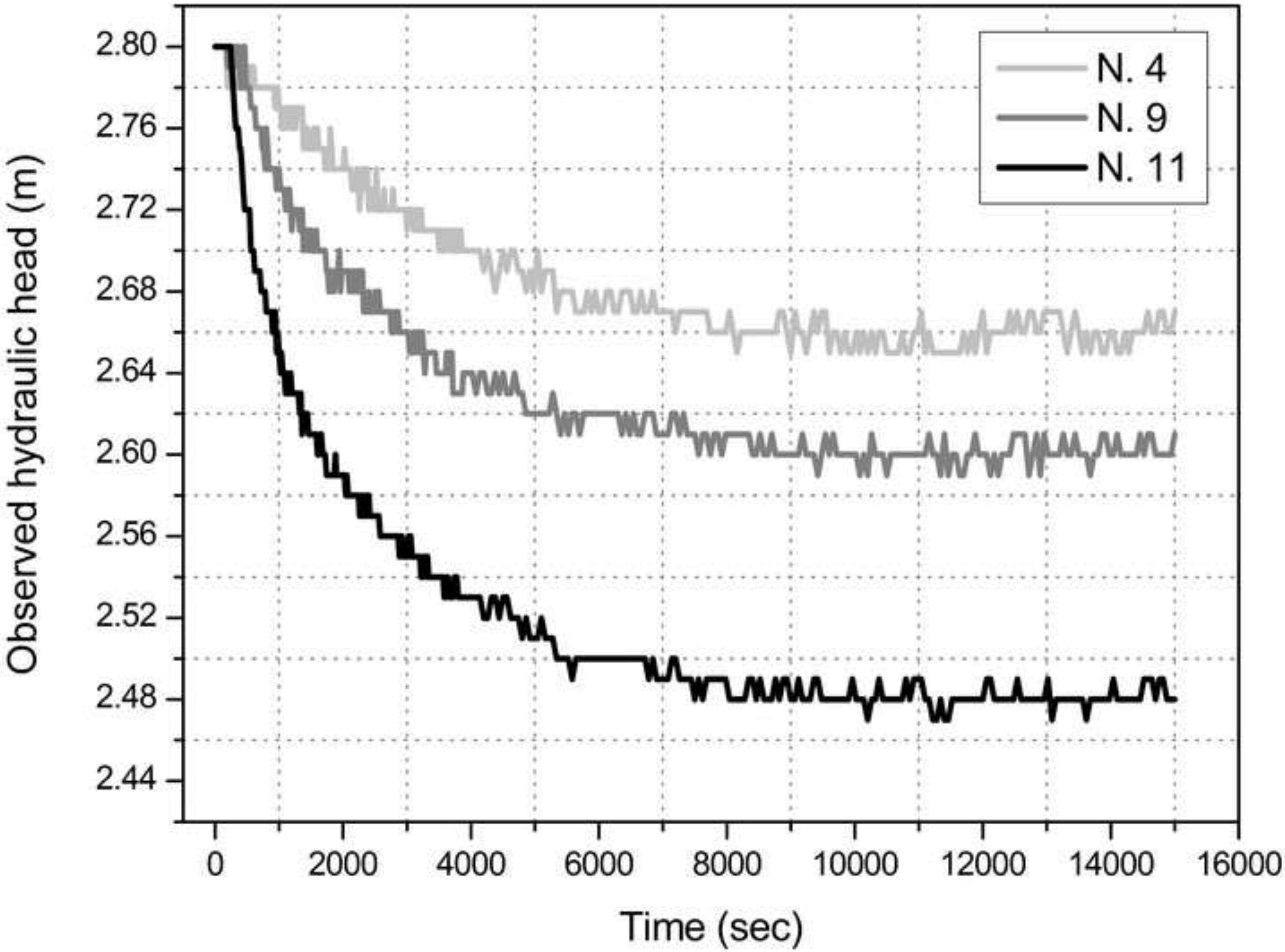

N. 4
N. 9
N. 11
Observed hydraulic head (m)
2.80
2.76
2.72
2.68
2.64
2.60
2.56
2.52
2.48
2.44
0
2000
4000
6000
8000
10000
12000
14000
16000
Time (sec)

Figure 4

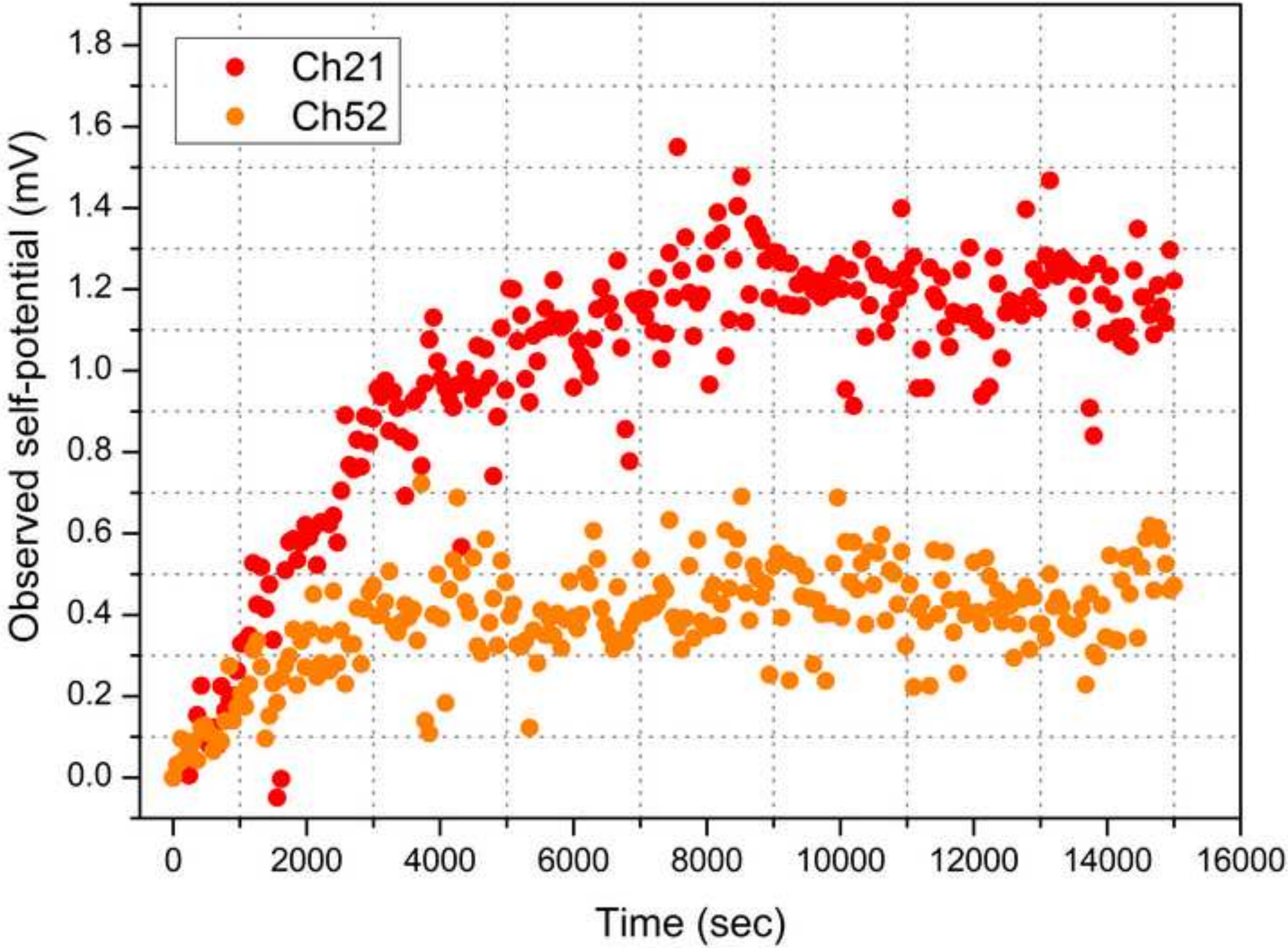
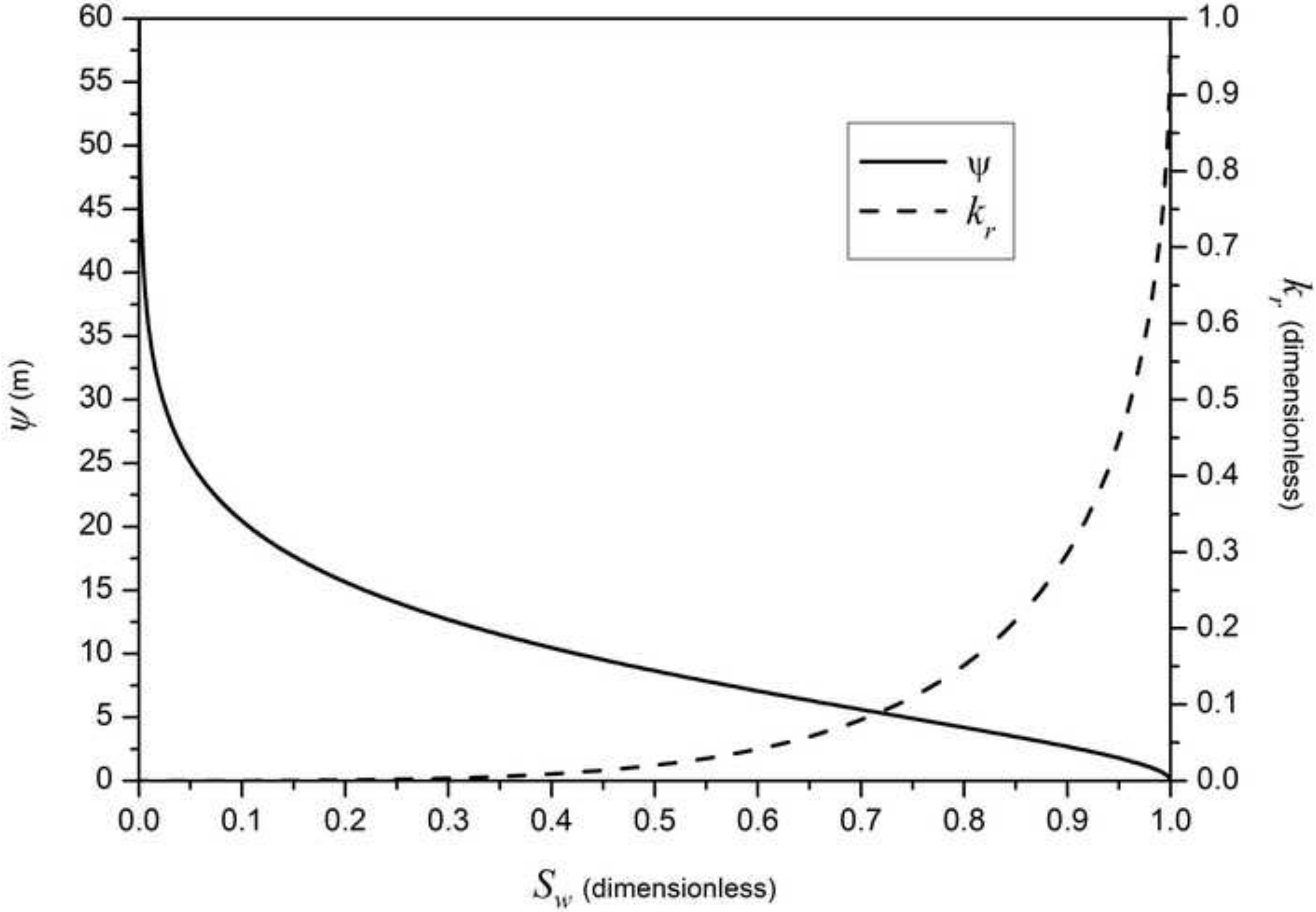

Ch21
Ch52
Observed self-potential (mV)
Time (sec)
1.8
1.6
1.4
1.2
1.0
0.8
0.6
0.4
0.2
0.0
0
2000
4000
6000
8000
10000
12000
14000
16000

Figure 5



ψ
k_r
ψ (m)
k_r (dimensionless)
S_w (dimensionless)
0
5
10
15
20
25
30
35
40
45
50
55
60
0.0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1.0

Figure 6

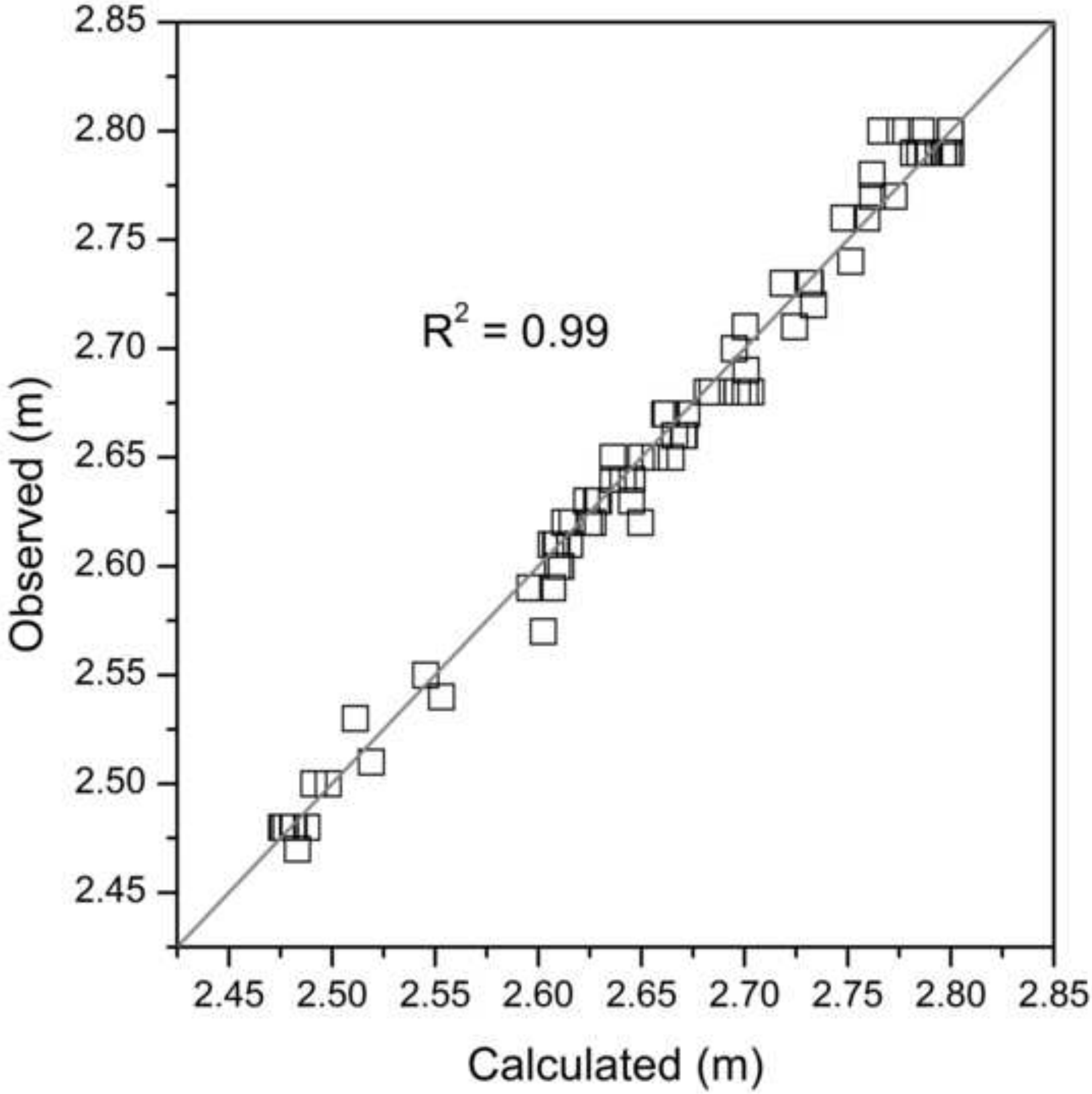

R^2 = 0.99
Observed (m)
Calculated (m)
2.45
2.50
2.55
2.60
2.65
2.70
2.75
2.80
2.85

Figure 7

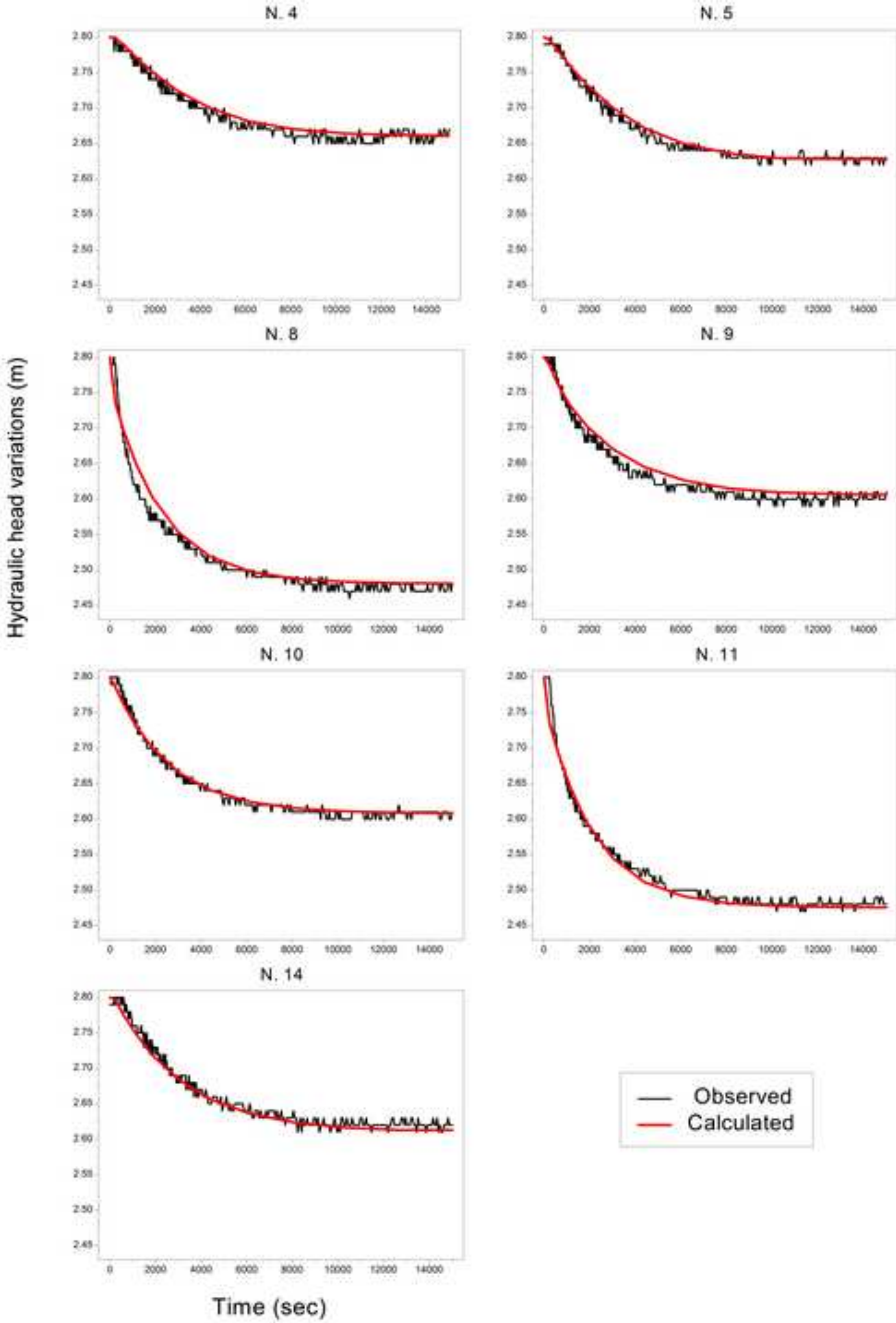

N. 4
N. 5
N. 8
N. 9
N. 10
N. 11
N. 14
Hydraulic head variations (m)
Time (sec)
Observed
Calculated

Figure 8

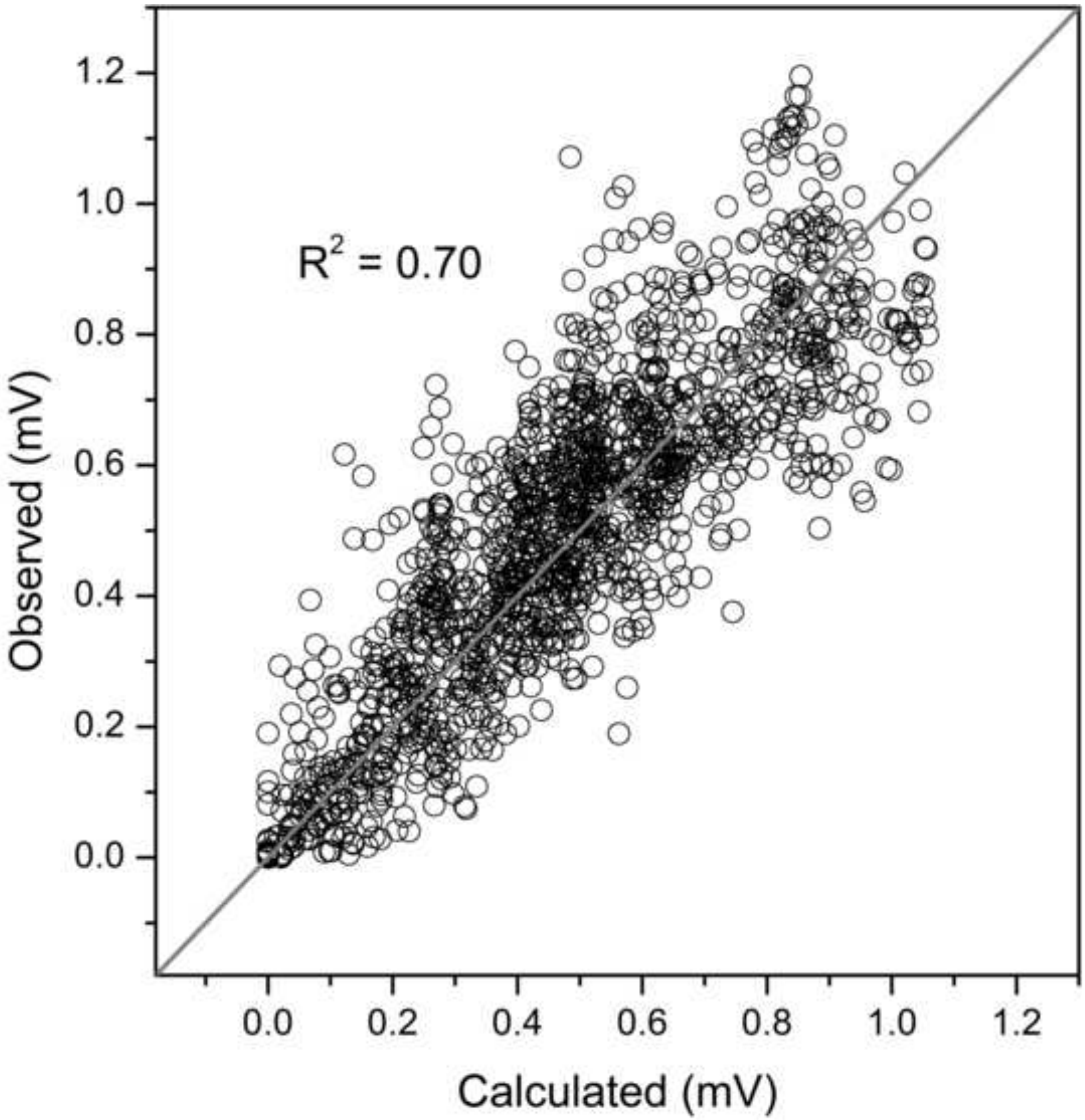


Figure 9

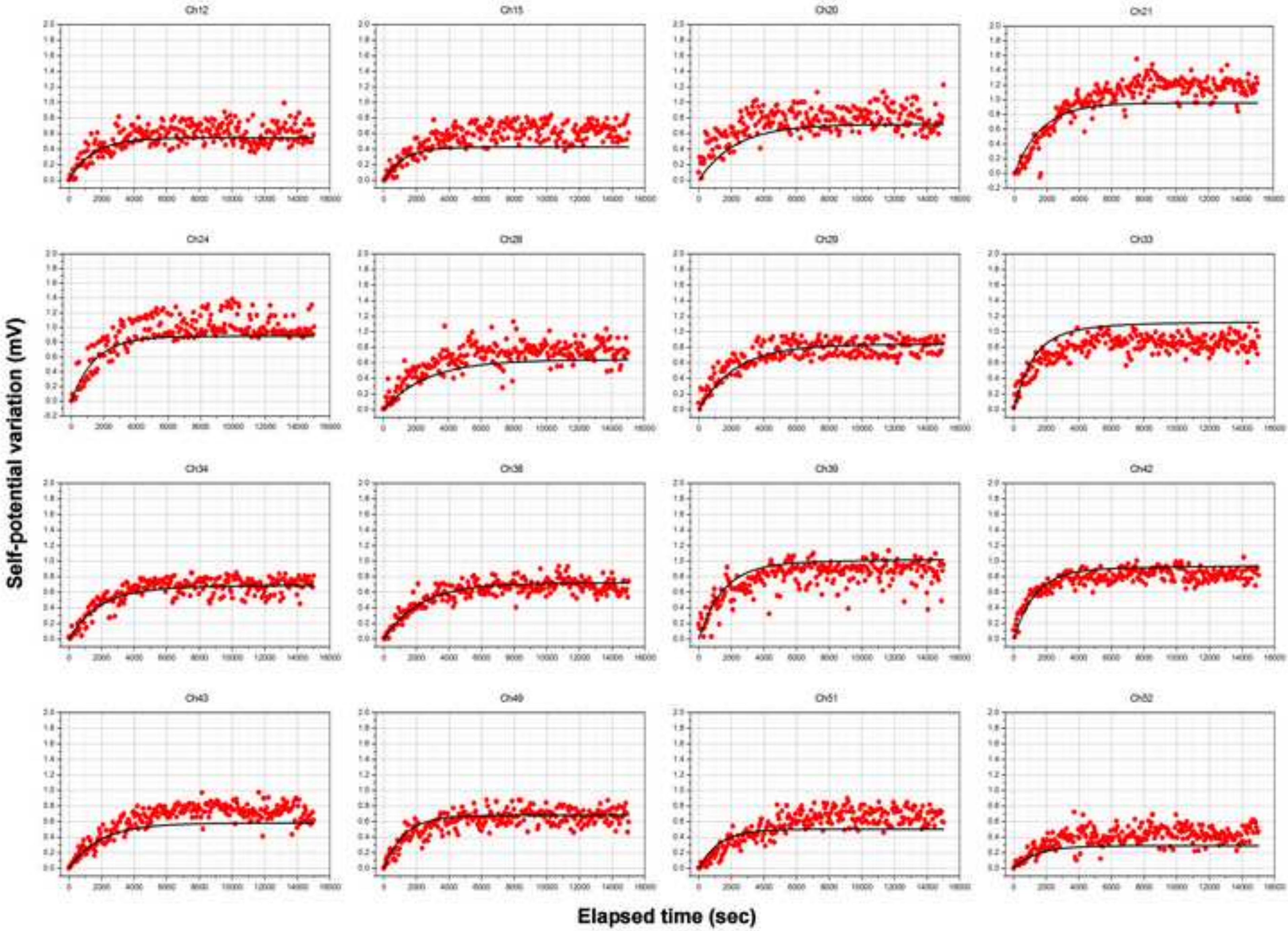
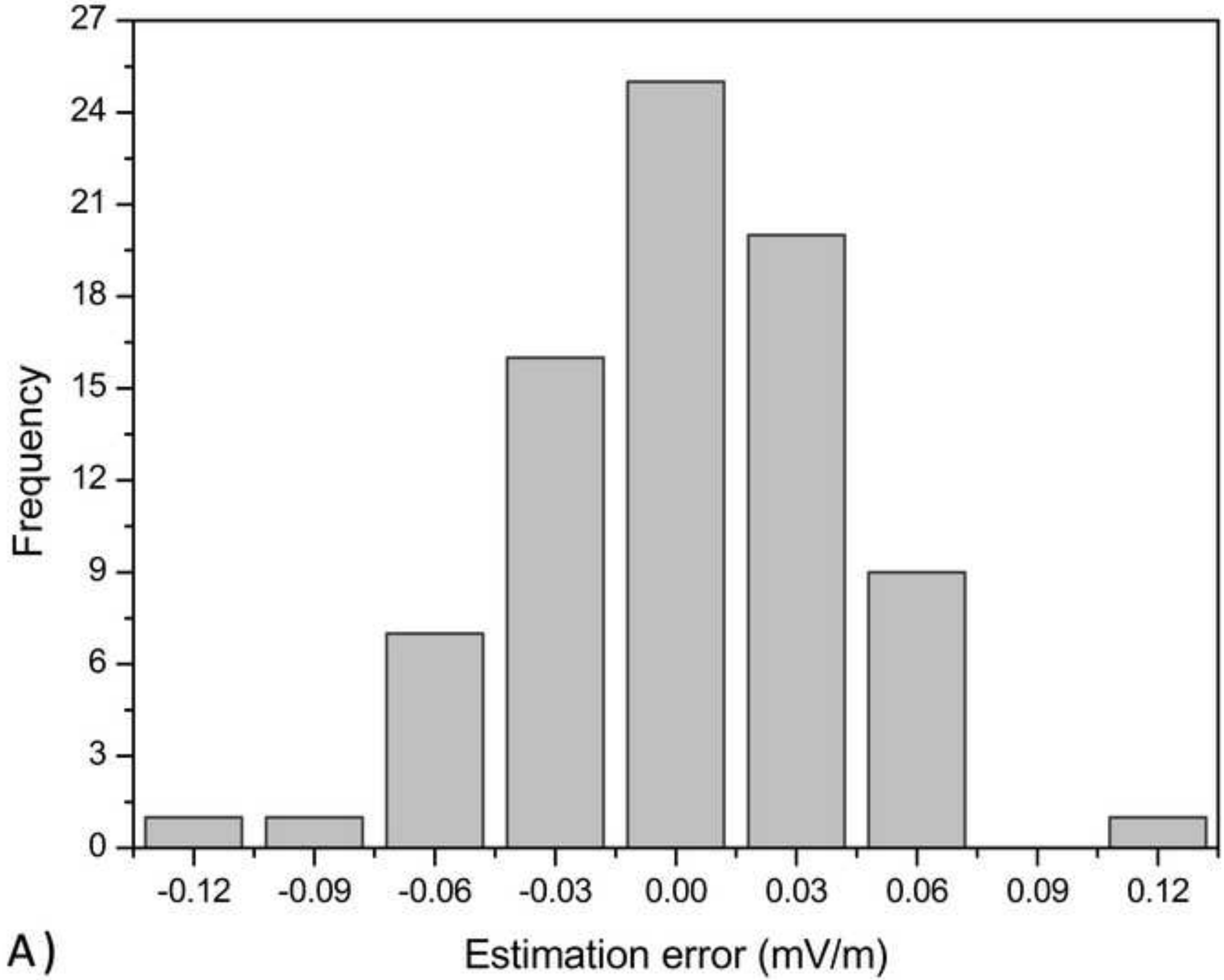
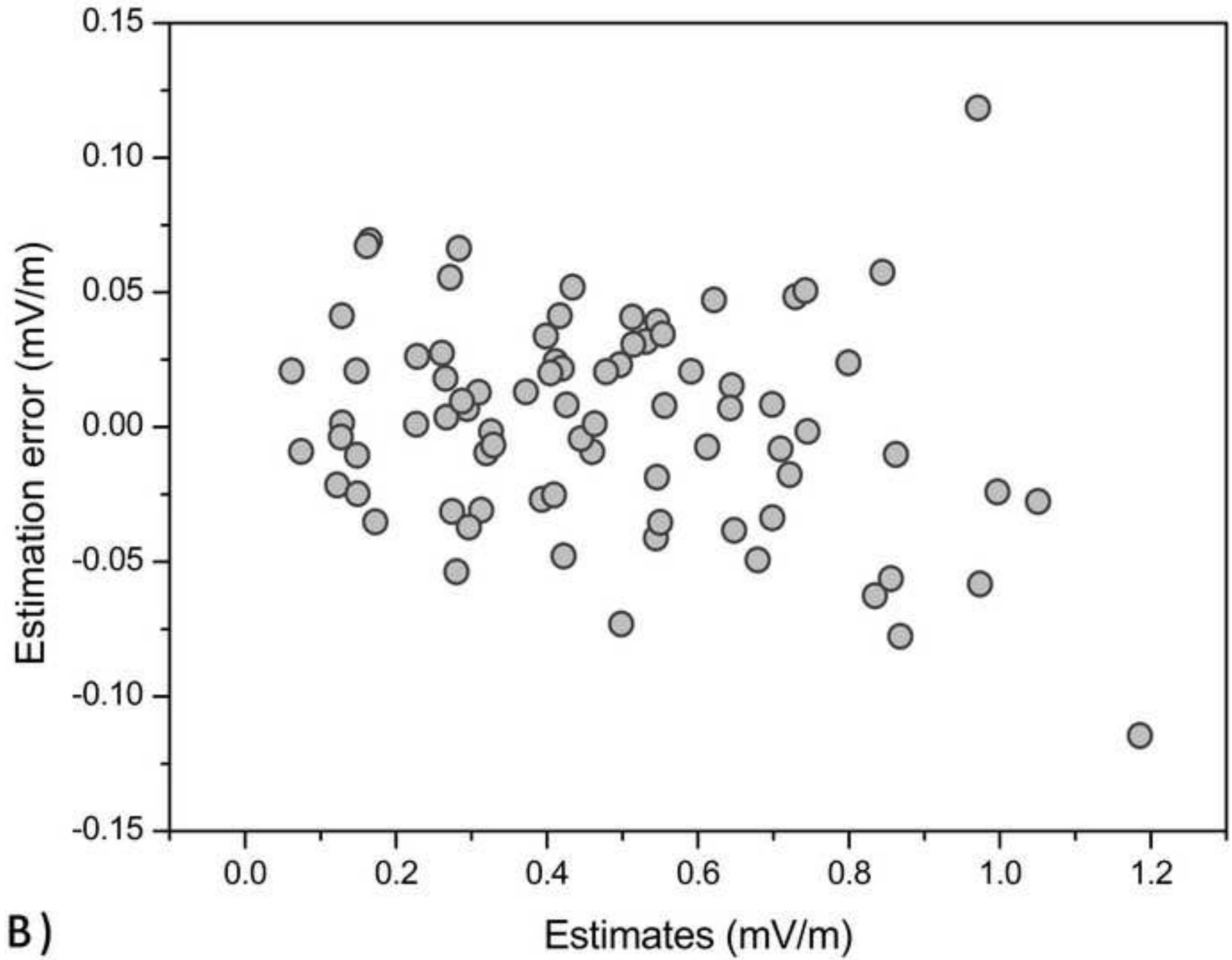

Ch12
Ch15
Ch20
Ch21
Ch24
Ch28
Ch29
Ch33
Ch34
Ch38
Ch39
Ch42
Ch43
Ch49
Ch51
Ch52
Self-potential variation (mV)
Elapsed time (sec)

Figure 10a



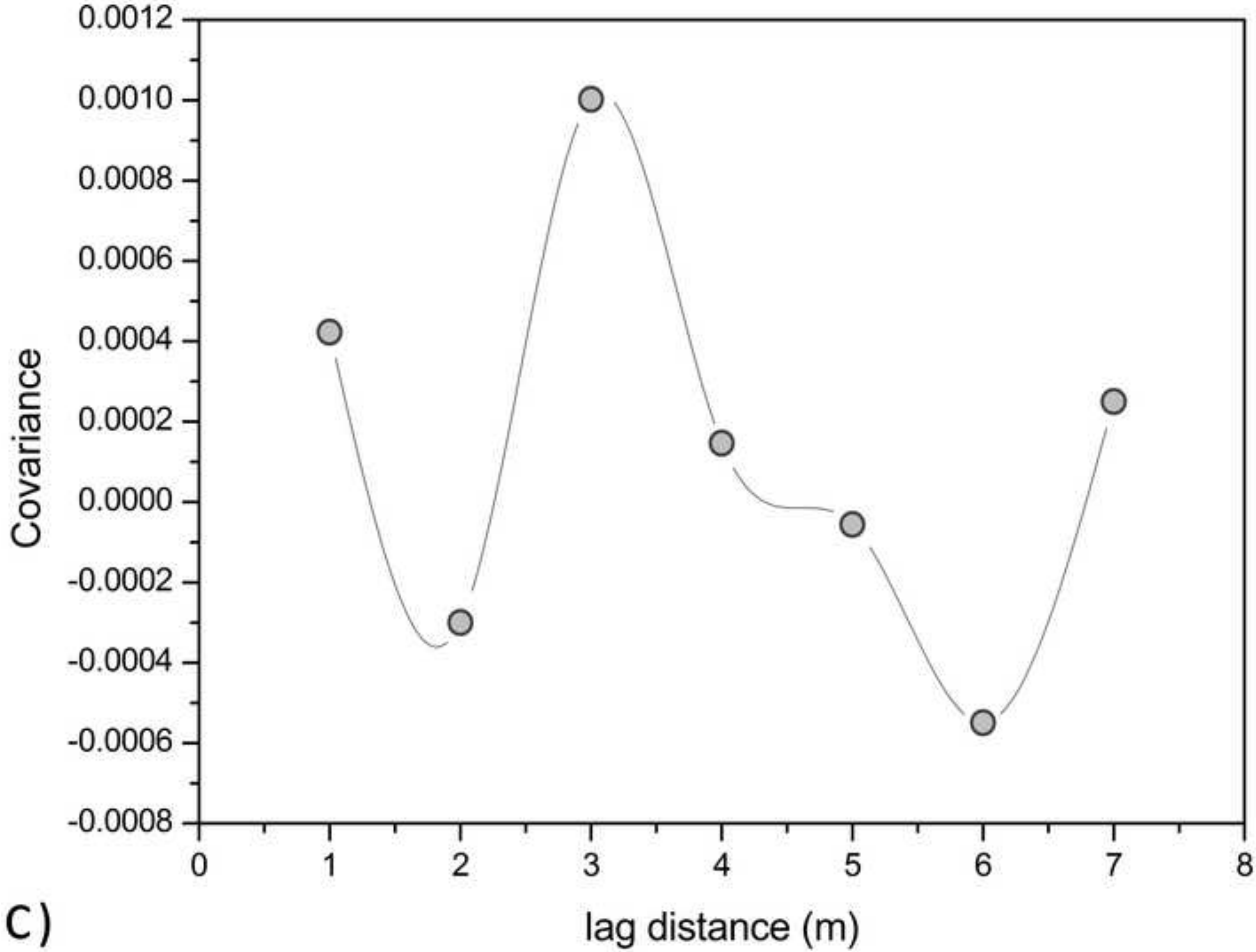
27
24
21
18
15
12
9
6
3
0
Frequency
-0.12
-0.09
-0.06
-0.03
0.00
0.03
0.06
0.09
0.12
A)
Estimation error (mV/m)

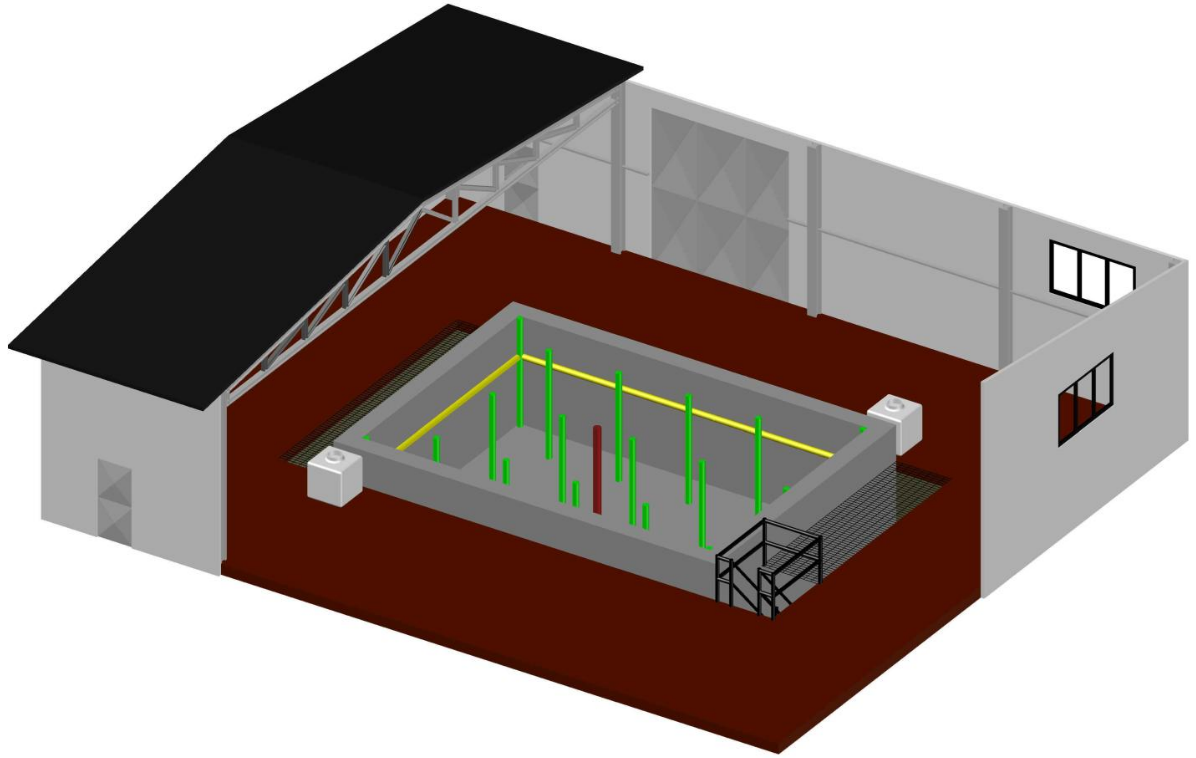
Figure 10b



0.15
0.10
0.05
0.00
-0.05
-0.10
-0.15
Estimation error (mV/m)
0.0
0.2
0.4
0.6
0.8
1.0
1.2
Estimates (mV/m)
B)

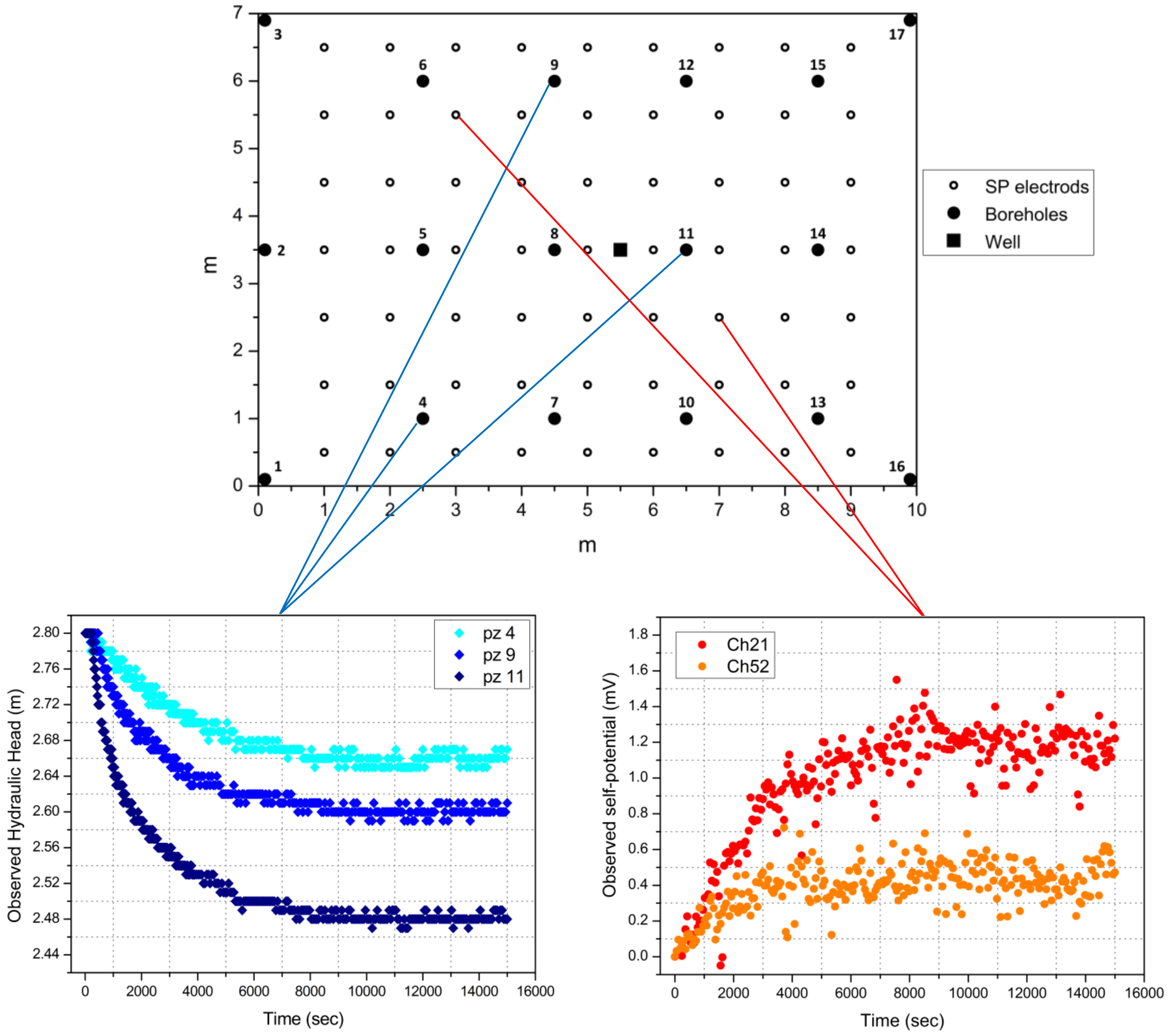
Figure 10c


0.0012
0.0010
0.0008
0.0006
0.0004
0.0002
0.0000
-0.0002
-0.0004
-0.0006
-0.0008
Covariance
0
1
2
3
4
5
6
7
8
c)
lag distance (m)

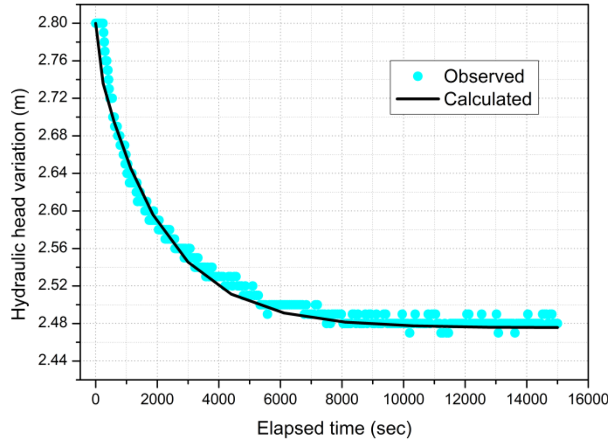


Large scale facility for hydrogeophysical experiments

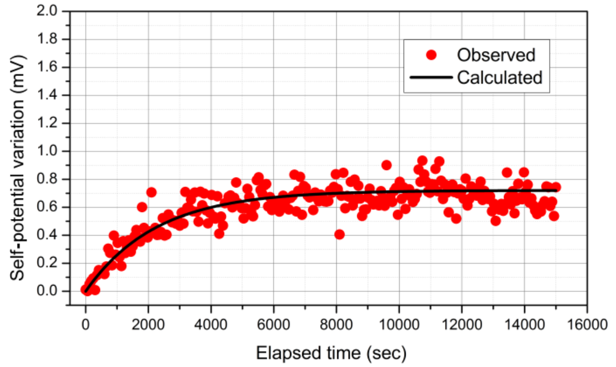
Pumping tests with hydraulic head and self-potential monitoring



Curve fitting in one of the monitored boreholes



Curve fitting in one of the monitored electrodes



Richards-Poisson coupled inversion modeling for porous media characterization

Highlights

- A multi-physical approach has been developed for real porous media characterization
- Hydrogeophysical experiments have been performed in a controlled full-scale facility
- Hydraulic head and self-potential variations have been acquired during a pumping test
- The collected data have been used into the coupled Richards-Poisson inversion model
- The obtained results remove all the approximations adopted in previous works