Integrated geophysical survey for 3D modelling of a coastal aquifer polluted by seawater

R. Martorana, L. Lombardo, N. Messina and D. Luzio

Dipartimento di Scienze della Terra e del Mare, University of Palermo, Italy

Received May 2012, revision accepted January 2013

ABSTRACT
Geophysical surveys are carried out in the coastal area of Petrosino (south-western Sicily) to study the time evolution of seawater contamination of the coastal aquifer, probably increased due to human impact. The overexploitation of the aquifer, due to an intensive agricultural use has affected significantly the natural hydro-geochemical state of the basin. The study is based on a processing and integrated analysis of hydrogeological, geochemical and geophysical data. In particular in the last two years seasonal time-lapse electrical resistivity tomographies (ERT), new TDEM soundings and Multi-Analysis Surface Wave soundings (MASW) have been carried out. The interpretation of the total set of previously existing and new geophysical data made it possible to reconstruct a three-dimensional model of the electrical resistivity of the aquifer, aimed at defining the extent and geometry of the seawater intrusion. Furthermore, the execution of a series of high-resolution time-lapse electrical tomographies and a correlation analysis between geophysical measures and geochemical, geological and hydrogeological data allowed to discriminate the effects of the salt concentration in the groundwater and the porosity and saturation degree of the rock on the time variations of the measured electrical resistivity. Finally, the average porosity of the rocks forming the reservoir was determined.

INTRODUCTION
In coastal aquifers, an unsustainable pumping of groundwater leads to pollution of freshwater and salinization processes, modifying the characteristics of the rocks with which it is in contact. In clay soils, for instance, the phenomenon of sodium absorption may occur that causes the reduction of hydraulic conductivity and soil salinization.

Determining the pollution degree of coastal aquifers susceptible to seawater intrusion can be improved through the optimization of geophysical prospecting and monitoring procedures, aimed at the detailed geometrical reconstruction and physical characterization of the aquifer, with particular regard to the intrusion wedge (Deidda et al. 2006; Khalil et al. 2011; Trabelsi et al. 2012). To define guidelines for the management of coastal aquifers with critical characteristics, a sample study was carried out, based on the integration of geophysical techniques suitable for hydrogeological monitoring (electrical tomography and seismic and electromagnetic soundings).

The selected test-site is located between Marsala and Mazara del Vallo (Fig. 1) along the south-western coast of Sicily. Here the aquifer is subjected to intensive exploitation, which strongly interferes with the phenomenon of seawater intrusion. The overexploitation of groundwater has caused strong increases of underground water flows, resulting in a strong degradation of specific coastal wetlands locally called ‘Margi’. These very specific environments, characterized by an outcrop of the local piezometric surface, are sensitive ecosystems with a high-conservation value. In addition they are very valuable ecosystems, qualified as special protection areas of the Sicilian region and between the areas of intervention of the European project LIFE-Nature.

The geophysical surveys, integrated by hydrogeochemical data (Nguyen et al. 2009; Perttu et al. 2011) are aimed to reconstruct the geometric features and the distribution of the porosity of the main geological structures of the aquifer and the variations in time of the salt concentration and degree of saturation (Di Napoli et al. 2011). To this end a large-scale three-dimensional resistivity model was reconstructed and, subsequently, high-resolution time-lapse vertical sections of the same parameter were obtained for a coastal area that was found to be particularly vulnerable to intrusion.

For this study, electrical resistivity tomographies (ERT), TDEM soundings and Multi-Analysis Surface Wave soundings (MASW) were carried out with a spatial distribution suitable for 3D modelling of underground structures. New measurements were integrated by some TDEM measures and a chemical analysis of water samples, taken in many wells in the area, already used in a previous investigation (Capizzi et al. 2010). These enabled the spatial patterns of the concentration of the principal
the presence of wetlands locally known as ‘Margi’. Their origin can be attributed to the outcrop of the piezometric surface. The important role played by these wetlands is twofold: the Margi are points of artificial recharge of freshwater for the aquifer and natural barriers to the intrusion of seawater. Unfortunately some years ago many pools partially dried up due to the excessive exploitation of the groundwater. As a consequence the gradual extinction of Margi is enhancing the natural seawater intrusion.

The oldest sequences of the catchment area belong to the ‘Cozzo Terravecchia’ formation (Flores 1959; Schmidt di Friedberg 1962). This is characterized by deltaic deposits consisting of clay and sandy clay in colour from brown to grey-green, containing lenses of sand and pebbles. The dating of these sediments ranges from upper Tortonian to lower Messinian (Ruggieri et al. 1977). In unconformity on Cozzo Terravecchia are reef deposits composed of porite limestones laterally passing to calcarenites and marls, with corals, bryozoans and mollusks, of the middle Messinian. The sequence proceeds with limited outcrops of gypsum in unconformity on the Cozzo Terravecchia formation. Above the chalk, there are deep-sea deposits called ‘Trubi’ consisting of Globigerina limestones and marly limestones of the lower Pliocene looking yellowish and intensely fractured. Above the Trubi there are terrigenous deposits composed of marl and clayey marl with interbedded sandstones attributable to the ‘Belice’ formation of the upper Pliocene (Ruggieri and Unti 1974). The largest outcrops are concentrated in the northern portion of the basin. The roof of the aforesaid land is the ‘Marsala Calcarenites’ formation of Emilian II-Sicilian (Ruggieri et al. 1977). The true thickness of this formation is not known but it is possible to distinguish a lower part with a thick-

FIGURE 1
Location of the study area.
ness of about 30 m consisting of poorly cemented calcarenites with lenses of clayey sands. The upper part, likely much more powerful, consists of well-cemented, smooth-grained calcarenites, still widely quarried to obtain blocks for construction. This formation lies according to a monoclinal that dips NE-SW with slopes that rarely exceed 5° and emerges along a strip extending from Marsala to Mazara del Vallo. Above the Marsala calcarenites, there is a system of marine terraces of the late Pleistocene, with a NE-SW direction and altitudes ranging from more than 150 m to the sea level. The highest terrace is outcropping in the north-eastern area and represents the ultimate expression of marine ingression on the shore during the Middle Pleistocene. At altitudes between sea level and 40 m a system of terraces, attributable to the Tyrrenian period, consist in a calcarenitic table, which stretches along the coast between Marsala and Mazara del Vallo. This coastal plate is characterized by the presence of Margi.

The relatively high-water potential of the area is the result of near-surface geology. The outcropping formations and ones just below are in fact calcarenitic structures (Marsala calcarenites and terraced deposits) that rest on an almost impermeable substrate mainly composed of clay and marl. More precisely, two zones are distinguishable: the first where the Marsala calcarenites outcrop and the second represented by sandy and calcarenitic terraced deposits that cover the Marsala calcarenites. We can therefore distinguish between a shallow aquifer, made up of the terraces and a deep aquifer, formed by the Marsala calcarenites (Cosentino et al. 2003).

The main way for water circulation in this aquifer consists in natural porosity, circulating along stratification joints and along a net of cracks and fractures. Furthermore horizons with low permeability are present in both aquifers, constituted by silty or clayey materials, which determine an aquifer of multi-layered or semi-confined type. Between different water layers there are vertical water exchanges as a function of the piezometric level of each. The thickness of the aquifer varies from a few metres up to about 60–70 m.

The area studied is intensively cultivated and is marked by the presence of several wells for agricultural and domestic use. The disappearance of several springs and Margi in the area highlights the serious environmental conditions of this aquifer. Data obtained by hydrogeological studies carried out in the area by the Sicilian Aqueducts from the late ’50s, clearly show a deterioration in the quality of groundwater in relation to the progressive lowering of dynamic levels and decrease in average flow rates. Subsequent studies (SOGESTA-ENI 1974) have pointed out that the annual pumping of water has passed the annual recharge, exploiting the reserves. Also a recent hydro-geophysical study of the aquifer (Cosentino et al. 2007) highlighted its condition of overexploitation.

Chemical analyses were performed on water samples taken from 45 wells (Fig. 3) to determine their quality (Capizzi et al. 2010). The sampling period was limited to a few days so as to avoid anomalies related to changes of climatic conditions. The water analysis allowed the concentration of the main dissolved salts, the ion concentration of chlorides, sulphates and nitrates and the Total Dissolved Solids (TDS) in the water (Fig. 4) at the piezometric surface to be assessed (Fig. 5).

Due to the different degree of interaction of groundwater with seawater, chloride concentrations are between 82–1313 mg/l. The ionic distribution maps show that the maximum ion concentration is along the main directions of seawater ingression. The

FIGURE 2
Geological map of the area between Marsala and Mazara del Vallo (D’Angelo and Vernuccio 1992).
increasingly distant from the profile. The constrained inversion of the new measures was performed using the final model relative to the nearest measurement point as the initial model for the new inversion and searching the model, compatible with the new data, with minimal Euclidean distance from the initial model. This quasi-homogeneity constraint was applied only to sets of TDEM decay curves that were *a priori* assigned to the subareas, characterized by small lateral variability, shown in Fig. 6.

**TDEM: data acquisition and processing**

The measurements were performed using the equipment TEM-FAST 48 of Applied Electromagnetic Research (Ranieri 2000).

Several factors were considered in the design of the survey. The criteria that guided the deployment of measuring points in the area have already been discussed. The selected antennas are of ‘coincident antennas’ configuration, while the optimal size of the side of the square loop transmitting and receiving antennas was considered to be about 50 m, taking into account the required depth of investigation and the logistic problems induced by the area by intensive cultivations. The current in the transmitting antenna was set to 3 A and stacking for up to 260 experimental discharge curves was carried out to improve the signal-to-noise ratio. The acquisition time windows varied from about 2 ms to about 16 ms.

A preliminary analysis of the acquired data was carried out in the field, using a fast pseudo-inversion algorithm. This proce-
dure, by adopting Occam’s razor technique, calculates a solution model characterized by a continuous trend of the resistivity versus depth. This solution, even when characterized by a low misfit, does not look very realistic from a geological point of view. However it proves to be a useful tool, in order to evaluate in situ the quality of the data, their interpretability with 1D models and their compatibility among neighbouring acquisitions.

When the decay curve resulted in excessive noise in the late times, the measurement was generally repeated making small shifts to the antenna. This technique ensured a good quality set of data even in some areas where the signal/noise, in the late times of the decay curve, was degraded by the presence of strong high-frequency electromagnetic noise.

The next step consisted in recognition and attenuation by appropriate high-cut filters of the noise, for the most part found in the early times and in some cases even on the end of the decay curve. When necessary, smoothing was performed on these curves that presented scattered data, usually caused by electromagnetic noise due to the presence of steel structures to support the vines.

The decay curves of the full set of 150 TDEM soundings were transformed into apparent resistivity versus time curves. These were plotted in a log-log graph in order to identify similar curves. The high variability of the resistivity in the shallower portion of the subsoil, residual effects of the ramp time and mutual induction between the antennas cause a distortion of the trend of the experimental curves at low times. For this reason the curve likeness was tested only for times greater than 7 µs.

The full set of curves was clustered based on their spatial distribution (distance between neighbouring points relative to a similar measurement less than a threshold value and the convexity of the set of points) and on their similar shape (Fig. 6). An example of TDEM curves of apparent resistivity related to a homogeneous area is shown in Fig. 7. The subareas thus obtained were considered nearly homogeneous and the 1D resistivity model related to each sounding, carried out inside, was determined using the TEM-Researcher code (Barsukov et al. 2007) and the following sequence of steps. The first inverted TDEM sounding is for the largest amount of a priori information available, coming from another close geophysical survey or drilling.
After determining the inverse model, the nearest measurement point is considered for the subsequent inversion, by adopting as an initial model the previously found model. Therefore a resistivity model, compatible with the new experimental data, characterized by minimal Euclidean distance from the initial one is determined and again adopted as an initial model for another nearest measurement point. Continuing interpretation by this technique, minimal horizontal variations of resistivity and layer thickness were imposed upon the 3D model.

ERT: data acquisition and processing

Three 2D electrical resistivity tomographies (ERT) were performed along the profiles, perpendicular to the coastline, shown in Fig. 3, to minimize in advance the transversal changes of resistivity. The profiles AA’ and CC’, which respectively have a length of 870 m and 1280 m, were carried out with an inter-electrode step of 5 m, using Wenner-Schlumberger and linear grid (Martorana et al. 2009) electrode configurations. The longest among these (BB’) was selected for the design of a resistivity tomographic monitoring (Loke 1999), in order to construct a model for the seasonal evolution of the aquifer. This profile has a length of 960 m with an inter-electrode step of 5 m. With the objective of reaching an investigation depth of about 65 m and to improve the signal-to-noise ratio, an optimization of the measuring sequence was considered necessary. For constant theoretical investigation depth and measurement areal density in the pseudosection, the acquisition sequence that minimizes the average and maximum values of the geometric factor is considered optimal. The pole-dipole array was mainly used in this sequence, with a maximum dipole order of 5 and dipole length from one to six times the inter-electrode step. The use of a 48-channel Syscal Pro resistivity meter enabled the acquisition of seven lines of 240 m each, with double coverage, so as to obtain full coverage of the pseudosection down to a pseudo-depth of 40 m. Unfortunately, below the depth of 40 m, where the measures are quite noisy, the coverage is discontinuous. This makes it difficult to accurately identify the clayey basement of the aquifer without integration of the geoelectrical survey with the seismic one.

The low values of resistivity that characterize the investigated subsoil almost from the surface and the low value of the maximum energization current (< 0.5 A) by the instrumentation have made the apparent resistivity measures relative to the lower part of the investigated volume affected by accidental errors with high variance and several outliers apparent. To improve the S/N, which prevents common filter algorithms altering excessively the actual relations between the values of apparent resistivity measures containing information spatially closely related, a special denoising algorithm was implemented. After having arranged the measures in a matrix compliant with the horizontal and vertical orders in the pseudosection, this algorithm, by comparison between the measures contained in a small moving window detects outliers and then replaces the remaining measures of the window with a weighted average. The results of applying this denoising technique show its effectiveness not only against the spikes but also in mitigating typical electrode effects that occur with linear anomalies in the pseudosections.

A $L_1$ norm, smoothness-constrained, optimization method (Claerbout and Muir 1973) has been preferred to the usual least-squares method, for the presence of a non-Gaussian noise component characterized by a high number of outliers.

The chosen algorithm is the iteratively reweighted least-squares method (Wolke and Schwetlick 1988), which implements an $L_1$ norm optimization method introducing in the
MASW: data acquisition and processing

The low-resistivity values (a few ohms per metre) that characterize the bottom of the aquifer across the investigated area make it impossible to determine the depth of the top of the clay basement by this parameter. Higher contrast is expected a priori between the body elastic wave velocity values in the aquifer and the underlying basement.

The MASW method (Park et al. 1997) determines a one-dimensional layered model of the S-wave velocity by the analysis of experimental dispersion curves of the Rayleigh waves. A series of 10 MASW profiles was carried out, parallel to the profile BB’ (Fig. 3) to mutually constrain layer thicknesses in the integrated inversion of electric, electromagnetic and seismic data. Each measurement array includes 24 geophones with a natural frequency of 4.5 Hz. Seven of them have geophonic spacing of 3 m and four source points were placed at the two extremes in order to obtain a maximum offset of 75 m, while the other three, placed in the area of the profile BB’ where there are major horizontal gradients of resistivity, have geophonic spacing of 2 m and six source points in order to obtain a maximum offset of 52 m. All signals were recorded with a sampling rate of 8000 sp/s. The processing of the recorded seismic sections and the determination of the experimental dispersion laws of the phase velocity of the first propagation mode of the Rayleigh waves (Fig. 8) were carried out by the software WINMASW (Dal Moro et al. 2006). The code DINVER of the software package SESARRAY, developed in the frame of the project Geopsy (Wathelet et al. 2004; Wathelet 2008) was used for the inversion of the dispersion curves related to each seismic section. This program adopts a constrained version of the Monte Carlo method to optimize the 1D models relative to each of the four or six seismic sections recorded for each profile. This is a heuristic technique of global analysis of model space and allows,
in a reasonable time, to move closer to the absolute minimum of the objective function. The starting model for each inversion is based on the available geological information for the area. The code provides a final model calculated by averaging over the models related to each source position. Given the small values of the velocity changes between neighbouring points inside the three layers of the models with respect to the contrasts between the layers, the depths of the two interfaces, assigned to the centres of the profiles, were interpolated with cubic spline functions in order to obtain the 2D layered $V_i$ section.

**INTERPRETATION OF RESULTS**

**3D resistivity model**

A topographic georeferentiation was made using the reference system WGS84 UTM 33N in order to assign to each TDEM model its correct location and altitude above sea level. The points $(x_i, y_i, z_i, \rho_i)$ near to the top and bottom of each layer of the 1D resistivity models were interpolated by the kriging algorithm, commonly used in spatial geostatistical analysis, to reconstruct the spatial pattern of resistivity in the survey area (Godio and Ranieri 2008). From the three-dimensional matrix of the resistivity of the whole basin, some horizontal and vertical sections were extracted and represented. The horizontal ones (Fig. 9) were extracted for $z = [+10 +5 0 -5 -10 -15 -20 -25 -30]$ metres a.s.l., the vertical ones (Fig. 10), however, at particular geological structures were highlighted by significant changes in resistivity.

The horizontal distributions of resistivity at the topographic elevations between 5 m and –10 m allowed to locate the main ingestion points of seawater along the coast, which largely overlap to the coastal zones called Margi and to sketch the intrusion directions.

The comparison between the resistivity section near to the piezometric level (Fig. 9) and the geochemical maps (Fig. 4) shows significant correlations between total conductivity and salt concentration, especially chloride concentration. This observation means that the soil resistivity depends mainly on the salinity of the underground water.

The pattern of the horizontal sections at different depths provides some hypotheses regarding structural geology and morphotectonics.

Into the north-western sector of the studied area, down to about –30 m a.s.l., a conductive anomaly, elongated in a NW-SE direction, is evident between two resistive structures.

It can be ascribed to a buried anticlinal fold, in which the two resistive bodies would be the fold sides belonging to the limestone and the central conductive body, the marly-clayey core. The axis of this supposed fold is parallel to those of the folds recognized in the outcrop in the easternmost part of the area (D’Angelo and Vernuccio 1992). A further picture of the fold is given in the vertical section F-F” that cuts the 3D model perpendicular to the fold axis (Fig. 10).

The TDEM surveys were arranged according to preferential directions coinciding with those of marine ingressions. Six vertical sections of the 3D model (Fig. 4) are represented in Fig. 11. Sections A-A”, B-B” and C-C”, roughly perpendicular to the coastline, show a strong decrease of resistivity near the coast and provide useful information to deduce the shape and extent of the intrusion. The other three sections (D-D”, E-E” and F-F”) were extracted along directions roughly parallel to the coast in order to obtain information on the shape of the basin and on the resistivity of the geological formations involved.

As above mentioned, in the subsection B-B’, bordering the natural reserve of Capo Feto, the realization of a ERT profile, constrained through some MASW surveys, was used to study the details of the intrusion wedge.

A comparison of the B-B’ 2D sections, obtained with three different geophysical techniques, is shown in Fig. 11. The three-layered $v_s$ section (Fig. 11a) obtained by horizontally interpolating the MASW models shows a thin near-surface layer with low velocity, overlying a layer characterized by $v_s$ values ranging from 500–800 m/s. It extends to about –30 m a.s.l. Finally the underlying layer is characterized by $v_s$ values ranging from 800–900 m/s. Electrical resistivity tomography (black line) is shown in Fig. 11(b) and the corresponding coastal portion derived from the TDEM section B-B” is shown in Fig. 11(c).

The ERT model (Fig. 11b) obtained by geometrically constraining the inversion with the deeper layer boundary of the 2D $V_i$ section, shows that the resistivity values of the aquifer remain below 2 Ohm.m until about 1200 m from the coast. This result can be interpreted assuming that the seawater intrusion wedge reaches this distance.
A comparison between the ERT B-B’ section (Fig. 11b) and the part of the TDEM section B-B” nearest to the coast (Fig. 11c), highlights a strong correlation between the two models, especially in the middle and deeper portions where the resistivity shows similar trends and values. The near-surface portions also show similar trends but with substantially different resistivity values: this discrepancy is likely due to the higher resolving power of the ERT method in the shallower part of the model. A slight increase in resistivity can be recognized between –25 m and –35 m a.s.l. This is probably due to the lithological transition from calcarenites to clayey marls.

Estimation of the aquifer porosity

The relationship between the conductivity of fluid in pores and that of fully saturated rock is called formation factor, $F$ (Archie 1942).

$$F = \frac{\sigma_w}{\sigma},$$

(2)

where:

$\sigma_w$, $\rho_w = \text{conductivity and resistivity of fluid in pores}$

$\sigma$, $\rho = \text{bulk conductivity and bulk resistivity of fully saturated rock}$.

For the most usual sedimentary rocks (not clays) Archie determined the following empirical relationship between the formation factor and porosity (second Archie’s law):

$$F = \alpha \rho^{-m},$$

(3)
were the cementation factor $m$ is generally between 1.3–2.5 and is about 2 for calcarenites; $\alpha$ is an empirical constant, usually set to 1 (Mavko et al. 1998). A value greater than 1 can be obtained when applying Archie’s law to rocks that do not follow closely Archie’s conditions.

Archie’s law, in fact, best describes the case of clean sand, well sorted, with electrical conduction to occur only by ion diffusion in interstitial fluids. Instead clayey sand rocks with secondary porosity by dissolution and rocks with isolated micro-porosity cannot be described by Archie’s relationships (Herrick 1988).

To estimate the mean formation factor of this calcarenitic aquifer, a correlation analysis was carried out of the electrical resistivity values of the water samples taken in the wells, at the aquifer top, with the corresponding ones determined for the same coordinates and depth from the 3D resistivity model. The measured values of the two quantities were reported in a scatter plot (Fig. 12).

The least squares linear regression with a straight line through the axis origin estimated the formation factor, $F = \rho_t/\rho_w = 6.39$, with a correlation coefficient between the two data sets of $R = 0.73$.

To carry out a second correlation analysis, the piezometric levels were estimated at the nodes of a square mesh net with a step of 100 m, $\rho_t$ values to the top of the aquifer were extracted for the same nodes from the 3D model (Fig. 13, top) and $\rho_w$ measures were interpolated into the same points (Fig. 13, bottom). Correlating the numerical grid of these two maps (Fig. 14a) we can note that in some areas close to the coastline the interpolation law generates values not in agreement with Archie’s law. This can probably be attributed to the difference between the typical wavelength of the real phenomenon in this coastal area subject to strong horizontal salinity gradients and the considerably greater differences imposed by spatial distribution of the measurement points. Removing data clearly not compatible with Archie’s law (Fig. 14b), the least square estimate of the formation factor resulted in $F = 7.13$ with a correlation coefficient between the two data sets of $R^2 = 0.80$.

There are several empirical relationships linking $F$ to rock porosity. Considering the Hashin-Shtrikman lower bound of the factor $F$ for a rock with porosity $\phi$, given by Berryman (1995):

$$F^{HS} = 1 + \frac{3}{2} \frac{1 - \phi}{\phi},$$

a minimum value of the average porosity for the investigated area of 22% or 20% of the total volume is obtained for the first and second estimate of $F$ respectively.

Considering the second Archie’s law (equation (2)) for clay-free saturated rocks and setting $m = 2$ and $\alpha = 1$, the average porosity estimates $\phi = 39\%$ and $\phi = 37\%$ are achieved for the two $F$ values. These values are consistent with those of calcarenites with mean porosity.

Considering the relationship:

$$\sigma = \sigma_e \phi^m,$$

or

$$F_{ORM} = \phi^m,$$

FIGURE 12
Statistical distribution and linear regression correlation of the electrical resistivity values of the water samples taken in the wells, at the aquifer top, with the corresponding ones determined for the same coordinates and depth from the 3D resistivity model.

FIGURE 13
Using the piezometric level estimates at the nodes of a square mesh net (represented with contour lines), a 3D surface of $\rho_t$ at the groundwater level (top) was obtained by extracting resistivity values from the 3D model in the same nodes and similarly $\rho_w$ measures were interpolated in the same points (bottom).
suggested by the differential effective medium (DEM) theoretical model proposed by Sen et al. (1981) for dispersive media (as $\omega \to \infty$) but sometimes also used for non-dispersive media, average porosity estimates of $\phi = 29\%$ and $\phi = 27\%$ are achieved for the two $F$ values.

Since water samples were collected from wells in which the piezometric surface should lie within the Marsala calcarenites, it can be concluded that this rock formation is characterized by a porosity value certainly greater than 20% having a mean value about 29%.

A map of minimum porosity (equation (4)) of the aquifer at the piezometric level (Fig. 15) was obtained by the punctual values of $\rho_1/\rho_w$ using equation (2). This map highlights a smooth trend of porosity that remains almost constant in most of the investigated area with values around 20%, except in the coastal areas where the anomalous porosity values are likely due to the incompatibility of the interpolated data with Archie’s law.

**TIME-LAPSE ELECTRICAL TOMOGRAPHY**

To monitor the seasonal aquifer variation in the part of the investigated area closest to the coast, the last phase of the research consisted in executing a 2D time-lapse electrical tomography in a central portion of the profile B-B'' (from about 380–745 m from the coast) by repeating the data acquisition every three months during 2010 (De Franco et al. 2009). The first data set acquired, related to March 2010, was used to construct the 3D model described in the preceding paragraph.

The data acquired in June, September and December were interpreted, in order of time, by an inversion technique that uses the model obtained from the previous data set as an initial reference model to constrain the inversion of the new data set (Loke 1999). The optimization is carried out by minimizing a weighted norm of the vector whose components are: differences between the experimental and calculated apparent resistivities, deviations of the current model parameters from the corresponding ones of the initial model and the same parameters of the current model, whose Euclidean norm is a measure of its roughness you wish to limit.

For the time-lapse inversion (Barker and Moore 1998) the equation proposed by Kim et al. (2009) was adopted:

$$\left[J^T J + \lambda (F + \alpha M^T R M) \right] \Delta \theta = J^T g - \lambda (F + \alpha M^T R M) \theta,$$

where the same symbols of equation (1) are used and the relative weight of the different objective functions is determined by the parameters $\alpha$ and $\lambda$. In this case $\alpha$ was set to 1, in order to give similar weight to reducing the difference between the models and to the other two targets (Rucker et al. 2011).

In the case of an aquifer affected by seawater intrusion, the time variations of the total resistivity may be caused by changes of salt concentration or by variation of water content in the rock pores or generally by a combination of the two events. To interpret the obtained time-lapse models, not having more independ-

![](image.png)

**FIGURE 14**

Statistical distribution and correlation by linear regression between the $\rho_i$ and $\rho_w$ numerical grids represented in Fig. 13: a) complete diagram of interpolated resistivities; b) diagram of interpolated resistivities compatible with Archie’s law.

ent data to constrain this interpretative issue, two extreme options were preliminarily taken into account.

Assuming that the aquifer saturation condition is maintained over time, the amount of total dissolved solids in the water (TDS) can be derived from measures of water conductivity or a known formation factor of bulk conductivity. The more accurate correlation laws between TDS and $\sigma$, for $\sigma$ values ranging from 0–16 mS/cm, are in the form $TDS = k \sigma^q$ (Chang et al. 2004), where $q$ is near to 1 if the TDS is expressed in mg/l. Linear approximations of this relationship are often used when the variability intervals of the two parameters are not very extended. In this case the conversion factor dependent on the chemical composition of the TDS and on the temperature, can vary between 5.4–9.6 (kg/m$^3$).

The linear regression analysis between the TDS and $\sigma$ measures on the water samples collected in the 45 wells in the area has led to a conversion factor estimate of $k = 6.97$ with a correlation coefficient of $R^2 = 0.987$ (Fig. 16).

Based on the above statements one can say that the relative time variations of the TDS can be determined by the equation:

$$\delta TDS = \frac{TDS_{\text{final}} - TDS_{\text{initial}}}{TDS_{\text{initial}}} = \frac{\rho_{\text{final}} - \rho_{\text{initial}}}{\rho_{\text{initial}}},$$

December there appears also a clear horizontal discontinuity for abscissae lower than 470 m. However no credit was given to the corresponding resistivity values since after the first survey in that portion of the profile a vineyard had been implanted, whose rows are supported by metal wires that form a dense distribution of surface short circuits that substantially perturbed the measures. From June–September, in the part of the model located below sea level, a decrease of resistivity is observed, while in the upper part a growth of resistivity still continues. During the period from September–December small changes in resistivity are observed, positive in some areas and negative in others, with almost zero average. Since no data were acquired in March 2011 the variations in winter (December–March) were determined using March 2010 data and assuming the absence of a long-term trend not related to seasonal cycles.

The resistivity variations described could be a priori attributed to either variations in the parameters TDS and $S_w$ (Figs 18 and 19), on the basis of equations (7) and (8). The amount of the observed resistivity variations suggest that probably both causes acted simultaneously.

Between March–June in a near-surface layer with thickness of about 10 m a slight decrease in resistivity is observed. This may be explained by the rising of the water table, consistent with the seasonal variation in rainfall, with the distance between the observed area and the main charge zones and by the consequent increase of the saturation percentage $S_w$ up to the surface. In this volume it seems unlikely that effects of salinization will prevail, due to the low pumping, in spring, in the area with vineyards. In the part below, the observed increase in resistivity is probably justified by a decrease in the salinity connected with the supply of freshwater that also justifies the superficial effect.

From June–September, the predominant effect is a decrease in resistivity below the surface layer, probably due to an increase in salinity produced by the intense pumping in summer for agricultural purposes.

FIGURE 15
Map of minimum porosity of the aquifer estimated at the groundwater level, obtained considering the correlation between the porosity and the Hashin-Shtrikman lower bound of the factor $F$ (equation (4)), this latter estimated at each node of the grid by the ratio $\rho/\rho_s$ (equation (2)). The grid indicates areas where the interpolated resistivities are considered compatible with Archie’s law.

On the contrary, assuming that in the examined part of the aquifer there is partial saturation whose degree changes over time as a result of variations of the water table, it would be possible to determine the spatial distribution of the saturation factor $S_w$, defined by Archie’s second law (Keller and Frischknecht 1966), if the spatial distribution of the bulk resistivity of the aquifer in saturation condition is known. This links the bulk resistivity $\rho$ of a partially saturated rock, the fraction of total pores filled with fluid $S_w$ (saturation factor), the resistivity of pore fluid $\rho_w$ and the porosity $\phi$ by the relation:

$$ (S_w)^n = \frac{\rho_w}{\rho} = \alpha \phi^n \frac{\rho_w}{\rho_s}, $$

where $\alpha$ is a constant related to sediment type, $\rho$ is the resistivity of the rock when fully saturated ($S_w = 1$) and the exponent $n$ of the saturation factor, empirically determined, is generally about 2.

Based on the above assumptions one can say that the relative time variations of $S_w$ can be determined by the equation:

$$ \delta S_w = \frac{S_{w, \text{after}} - S_{w, \text{before}}}{S_{w, \text{before}}} = \sqrt{\frac{\rho_{s, \text{after}}}{\rho_{s, \text{before}}}} - \sqrt{\frac{\rho_{s, \text{before}}}{\rho_{s, \text{before}}}} $$

In the time-lapse resistivity models of 2010 (Fig. 17), limited to the aquifer portion located above the interface of the MASW model separating the calcarenitic cover from the underlying marly-clayey formation, a sharp increase of resistivity is evident from March–June. In the models for the months of June, September and
their both low value and high-spatial heterogeneity.

Finally, from December–March, the clear decrease of resistivity that is observed below the surface layer is probably attributable to the increase of water content in the pores.

CONCLUSIONS

The water supply in the studied area is polluted by seawater intrusion that reaches up to great distances from the coast. This phenomenon has been caused by an overexploitation of the

FIGURE 17

Time-lapse electrical resistivity tomographies performed seasonally in 2010, in a central portion of the profile B-B", limited to the aquifer portion located above the interface of the MASW model separating the calcarenitic cover from the underlying marly-clayey formation (in grey).

FIGURE 18

2D sections of the relative time variation of the TDS, seasonally evaluated in 2010 on the basis of equation (7).
A comparison between the geophysical model and the water resistivity measures carried out in several wells located in the area, allowed to determine an average formation factor for the Marsala calcarenites and different porosity estimates were obtained using some empirical relationship.

Finally, a time-lapse resistivity tomography, carried out in a zone strongly affected by intrusion, according to the 3D model, enabled a description of the seasonal variation of the interaction between sea- and freshwater, linked both to natural and anthropogenic processes. The time-lapse tomography sequence is going to continue in the already investigated line to better define the seasonal average variations and a long-term trend not yet defined, linked to the efforts made to restore optimal environmental conditions.

The results obtained may enable tracking of guidelines for the evaluation of salinization risk and the consequent environmental protection of coastal aquifers with similar hydrogeological characteristics.

REFERENCES
**NEW FEATURES**

- **200W - 2A - 700Vpp**
- Up to 40000 measures for single reading
- Up to 6 measures/sec
- Unlimited electrodes
- Ideal also for 3D surveys
- Accessories for VES and any ABMN configuration

**AUTOMATED ELECTRICAL RESISTIVITY TOMOGRAPHY IN A.C.**

**POLARES 2.0**

sales@pasisrl.it - tel. 011 6507033  fax 011 658646

---

**EAGE**

**Amsterdam ’14**

Register now!

Experience the Energy

www.eage.org

76th EAGE Conference & Exhibition 2014 | 16-19 June 2014 | Amsterdam RAI