THE USE OF THE TIME DOMAIN ELECTROMAGNETIC (TDEM) METHOD TO EVALUATE POROSITY OF SALINE WATER SATURATED AQUIFERS

M. GOLDMAN[1,✉] and U. KAFRI[2]

[1] Geophysical Institute of Israel, Lod 71100. E-mail: mark@gii.co.il
[2] Geological Survey of Israel, Jerusalem 95501

Abstract

Geoelectric (GE) and geoelectromagnetic (GEM) methods are successfully used to delineate the geometry of different hydrogeological targets. Their use to determine porosity of freshwater saturated aquifers leads in many cases to significant inaccuracy. This is caused by uncertainties in determining both fluid (ρ_w) and bulk (ρ) resistivities required for calculating porosity using Archie's equation. In order to overcome the problem, it is proposed herein to significantly reduce both uncertainties by studying only the saline and, in particular, seawater saturated parts of the aquifer, where ρ_w is generally known and ρ can be determined very accurately by applying a properly selected GE or GEM method. Experience shows that the most suitable geophysical technique for the exploration of saline groundwater is the time domain electromagnetic (TDEM) method. Numerous measurements carried out in different granular clastic aquifers have proven that the bulk resistivity of saline water saturated portion of the aquifer is accurately detected by TDEM without any use of *a priory* information. Therefore, under favorable conditions, the porosity of the saline water saturated portion of the aquifer can be also accurately determined from the Archie's equation by using only geophysical data. The obtained porosity might be projected then to the freshwater saturated portion of the aquifer since it is reasonable to assume that hydraulic properties of the aquifer are essentially the same both below and above the freshwater/seawater interface. The method was tested in both granular and carbonate coastal aquifers of Israel. The porosities measured in granular coastal aquifers exhibit a good agreement with those calculated using TDEM results. In the case of carbonate aquifers, the situation is more complicated due to their heterogeneity and much wider range of porosities caused by their karstic and fractured nature.

Keywords: Saline groundwater, granular aquifers, carbonate aquifers, porosity, TDEM

✉ Corresponding author

Introduction

Geoelectric (GE) and geoelectromagnetic (GEM) methods are the leading geophysical techniques in the exploration of saline groundwater. This is due to more close connection between electrical conductivity/resistivity and groundwater salinity than that existing for all other physical properties (e.g. density, acoustic velocity, magnetic susceptibility, dielectric permittivity, etc.) measured in the rest of the geophysical techniques.

The most successful applications of the methods were focused on studying different geometrical features of the target such as the lateral distribution of seawater intrusion, the depth to fresh-saline groundwater interface, etc. Several attempts have also been made to quantitatively evaluate different hydraulic/hydrological parameters using either measured or interpreted geoelectrical parameters. The most extensive efforts were devoted to studying the relationship between porosity and ground water salinity, on one side, and electrical resistivity on the other. The theoretical basis for such investigations is being constituted by the empirical Archie's equation (Archie, 1942):

$$\rho_w / \rho = \alpha \Phi^m \tag{1}$$

where ρ_w is the resistivity of water within the pore space, ρ is the bulk resistivity of the rock, Φ is the porosity of the rock (approximately representing the volume of water filling the pore space), α and m are material-dependent empirical factors, which are introduced to force the equation to fit the behavior of a rock in question.

Since m increases with cementation, Archie (1942) named it *cementation index,* having the characteristic value of 1.3 for unconsolidated sands and varying between 1.8 and 2 for consolidated sandstones.

Equation (1) already represents a significant simplification of the real situation, as it is only valid for 100% saturated rocks, fixed temperature, no clay content, etc. We would like to further simplify it by letting α = 1 and m =2 and thus getting finally

$$\rho_w / \rho = \Phi^2, \tag{2}$$

Even in such an oversimplified case, under the assumption that the bulk resistivity of rocks, ρ, is provided by surface GE/GEM measurements, we still have one equation with two unknowns, ρ_w and Φ. Moreover, experience shows that in many cases the determination of ρ could be biased due to the so-called equivalence problem (discussed below in details). Thus, the solution to equation (2) is expected to be inherently non-unique.

All previous attempts to resolve both kinds of non-uniqueness were based on an extensive use of *a priori* information regarding ground water salinity (for fixing ρ_w) and regarding geoelectric structure (for fixing ρ). Experience shows that the amount of *a priori* information available, in most cases, is insufficient to quantitatively resolve non-uniqueness. Therefore most of the attempts to apply surface GE and GEM to determine hydrological parameters suffer from one common drawback: they are apparently "successful" in close proximity to boreholes, where the information is available, but might be fairly unreliable far away from them. Moreover, even in the proximity of boreholes, the required *a priori* information provided by well logs is inevitably scale dependent and, thus, does not necessarily coincide with the required bulk geoelectric parameters. This phenomenon, which is less pronounced in the case of granular clastic aquifers, becomes crucial for fractured/cavernous karstic or crystalline aquifers.

Resistivity-salinity calibration

During the last two decades, numerous TDEM measurements were carried out in Israel primarily to detect saline water bodies within different aquifers and to delineate the interface between these saline bodies and the fresh ones. The studies focused mainly on the Mediterranean coastal aquifer (more than 400 soundings). Some work was also done in the coastal aquifers of the Gulf of Elat and the Dead Sea, as well as in the regional carbonate Judea Group aquifer of Israel. The results of these investigations in the Mediterranean (granular) and the Judea Group (karstic) coastal aquifers are discussed below.

The Mediterranean coastal aquifer

The Pleistocene Mediterranean coastal aquifer consists of 100 to 200 m thick calcareous sandstone and sand, which form the aquiferous part of this system. Close to and within a few kilometers from the shore, the aquifer is subdivided by marine and continental clay and loam interlayers into a few (generally four) sub-aquifers customarily designated as A, B, C and D sub-aquifers (Figure 1).

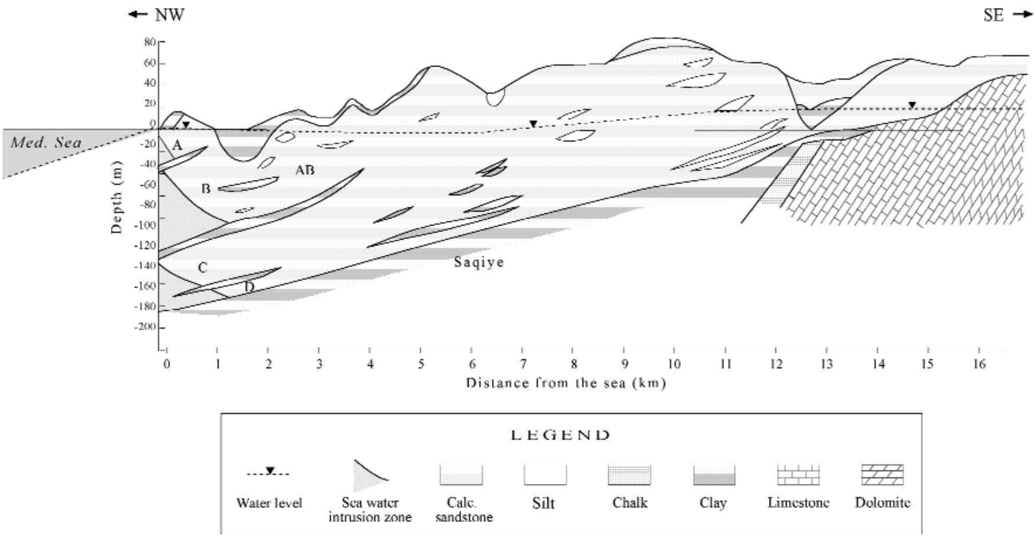

Figure 1. Typical hydrogeological cross-section of the Mediterranean coastal aquifer.

In order to quantitatively evaluate the ability of the TDEM method to both delineate geometrical features of seawater intrusion and characterize groundwater salinity, almost 50 of the TDEM measurements were performed in a close proximity of the existing observation wells. In all cases, the interpretation of the TDEM data was carried out without using any borehole information, sometimes prior to drilling the wells.

As a very typical example, Figure 2 shows the comparison of the borehole conductivity profile and the interpreted TDEM resistivity vs. depth section in the vicinity of observation well Ashdod-112. One can see that TDEM accurately delineates not only seawater intrusion within sub-aquifer B, but also the hydrological reversal within sub-aquifers C and D.

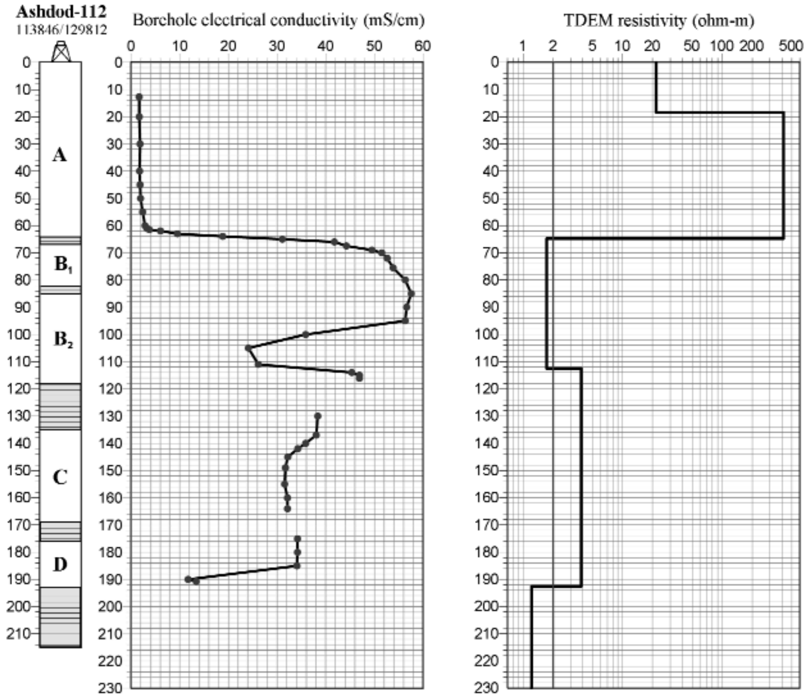

Figure 2. Comparison of the interpreted TDEM resistivity vs. depth model (in the right) with the borehole electrical conductivity profile (in the middle). The very left column shows the division of the aquifer into sub-aquifers A, B_1, B_2, C and D.

The attempts to characterize groundwater salinity in the salinity ranges typical for fresh and/or brackish groundwater using the interpreted TDEM resistivities are much less successful. Figure 3 shows the TDEM resistivity – groundwater chlorinity data based on 43 calibration measurements along the Mediterranean coastal aquifer of Israel. It can be seen that the ability of TDEM to quantitatively characterize groundwater salinity (chlorinity) dramatically decreases with decreasing the salinity of groundwater. Indeed, if the salinity exceeds 10,000 mg Cl/L (saline water), the TDEM resistivities vary in a fairly narrow range roughly between 1.5 to 2 ohm-m (excluding one outlier caused by a hydrogeological anomaly of a non-seawater origin). The resistivities characterizing the brackish water salinity range (roughly between 1,000 to 10,000 mg Cl/L)

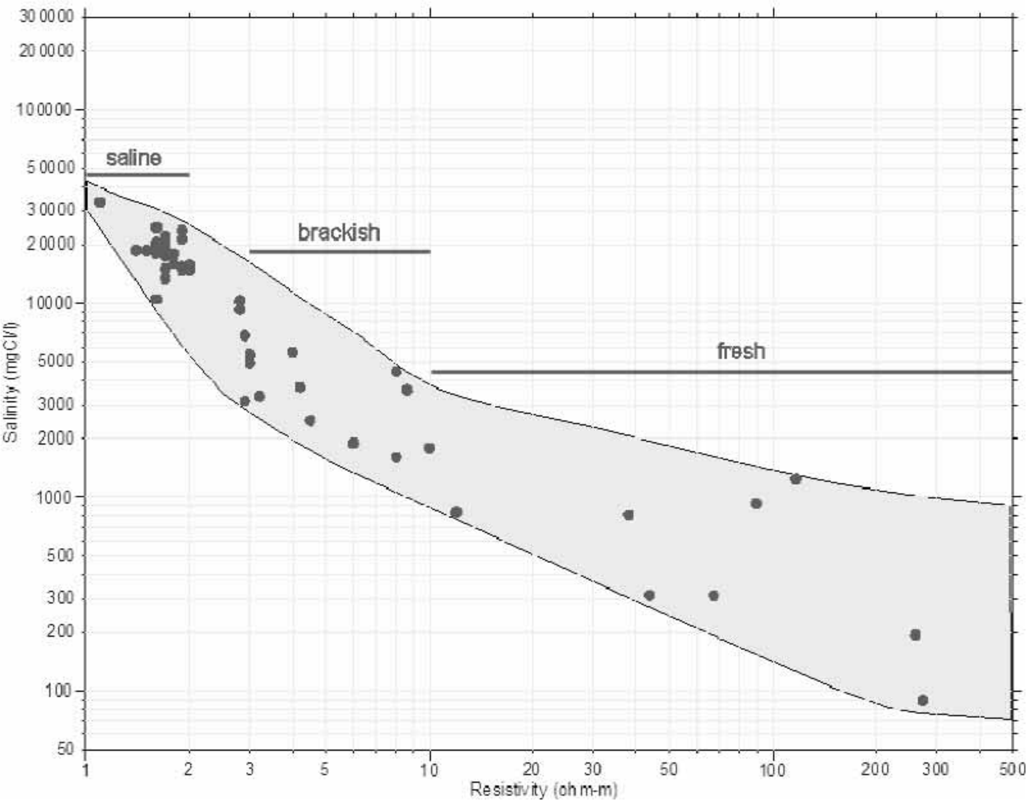

Figure 3. Resistivity-chlorinity calibration data based on 43 TDEM measurements carried out near observation wells along the Mediterranean coastal aquifer of Israel.

vary in a significantly wider range, between approximately 3 and 10 ohm-m. And finally, the freshwater salinities (below 1,000 mg Cl/L) are characterized by extremely wide resistivity range exceeding an order of magnitude (between approximately 10 and up to several hundreds of ohm-m). It is important to emphasize that the resistivity range characterizing saline water in saturated rocks is not only narrow, but also very different from that of any other lithologies whether saturated or not. This means, that, in case of saline groundwater, the quantitative hydrogeological interpretation of the TDEM results is generally single-valued, but it could be fairly non-unique in case of fresh/brackish water saturated rocks. Similar phenomenon is also observed in the geophysical interpretation and this important point is discussed below in great details.

Equivalence problem

The choice of a proper GE/GEM technique is crucial for the successful application of the proposed methodology. The main requirement to the method is the ability to accurately detect the fresh/saline interface and determine the bulk resistivity of the aquifer below the interface. Experience shows that, in most cases, the TDEM method is the right choice (Goldman et al., 1991). However, in some specific cases, where the application of TDEM is technically limited either due to a high level of ambient EM noise (e.g. within or close to industrial areas, etc.) or because of the difficulties to lay down appropriate transmitter loops (e.g. within urban areas or due to a rough topography), either 1-D resistivity sounding or 2-D resistivity imaging techniques can be applied instead. The 2-D resistivity imaging method seems particularly feasible either instead of, or better, in addition to TDEM in areas with an expected multi-dimensional distribution of the subsurface resistivity. Nevertheless, the following discussion is restricted to the TDEM method only.

The detailed description of the method is given in numerous publications, including those devoted to groundwater exploration (Fitterman and Stewart, 1986). The present paper discusses only those features of the method that are relevant to the determination of the porosity of the aquifer.

The most important feature that may significantly affect the ability of the method to quantitatively estimate the porosity, is the so-called equivalence range, i.e. the range, in which geoelectric parameters may vary without noticable variations in the measured signal. Let us consider this problem in some greater details, using a typical example of the TDEM measurement at the Mediterranean coastal aquifer.

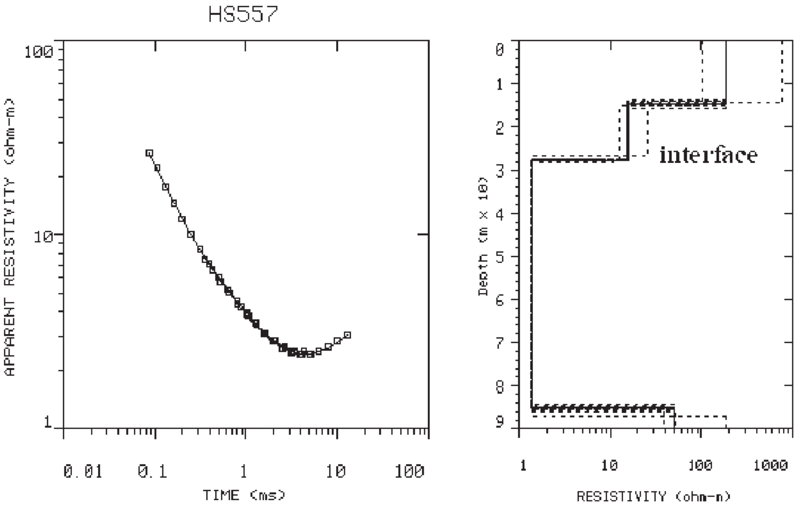

Figure 4. TDEM data collected at station HS557 in the Mediterranean coastal aquifer of Israel and the results of 1-D inversion including linear equivalence analysis by means of the TEMIX-XL interpretation software (Interpex, 1996).

Figure 4 shows the TDEM data collected using the Geonics EM-67 system (open squares at the left hand side) along with a family of the equivalent interpreted models at the right hand side of the figure. One of the models, shown as a solid line, represents the so called best-fit model, while the rest of the models, shown by dashed lines, are the equivalent models. The solid line connecting the data points at the left hand side is a family of synthetic responses for both the best-fit and equivalent models. The difference between the responses is so small that they all appear as a single curve. In this specific example, the best-fit model provides a misfit error of approximately 1.9 %, while the greatest misfit error for the equivalent models does not exceed 2 %. Thus, the maximum difference between all the synthetic responses does not exceed 0.1 %, i.e., less than the overall noise level of the measurement. Therefore, in practical terms all the interpreted models are equally consistent with the data and can not be distinguished without *a priori* information.

It should be noted that both the data interpretation and the equivalence analysis have been carried out using TEMIX-XL interpretation software (Interpex, 1996). Strictly speaking, the software does not provide a full (non-linear) equivalence analysis, but only the one, which is based on the use of a parameter resolution matrix (linear equivalence). Nevertheless, for illustration purposes, it is enough to consider the partial equivalence analysis just remembering that the variations of some model parameters could be even wider than those shown in Figure 4. Some further details regarding the difference between linear and non-linear equivalence problems have been discussed by Goldman et al. (1994).

The equivalence analysis of the TDEM measurement (Figure. 4) exhibits the following resisitivity sequences:

Depth range (m)	Resistivity equivalence range (ohm-m)
0 – 14	80 – 600
14 – 26	11 – 27
26 – 87	1.35 ± 2 %

It should be noted that there is some insignificant depth equivalence range (Figure 4), considerably narrower than that of the resistivities. The depth values considered above are taken, therefore, for the best fit model only.

Porosity calculations

The above mentioned example of TDEM station HS557 (Figure. 4) is used to describe herein the methodology of calculating porosity of the saline part of the aquifer.

Based on the fact that bulk resistivity values below 2 ohm-m are solely indicative of seawater intrusion into the Mediterranean coastal aquifer (Goldman et al., 1991), one can fix the location of freshwater/seawater interface at the depth of approximately 26 m. The shallow sequence from the surface to the depth of 14 m exhibits high resistivities representing both the unsaturated zone and the underlying freshwater saturated aquifer. The thickness of the unsaturated zone at this site is only 2 m and therefore it is neglected in the further calculations. No resistivity drop is recognized at the depth of the water table, apparently due to the limited resolution of the TDEM method. The intermediate zone between 14 and 26 m exhibits resistivity values typical for either aquicludes/aquitards or aquifers saturated with brackish waters, salinity of which may vary in a fairly wide range.

The very low resistivity geolectric layer below 26 m depth, reflects seawater intrusion. Applying equation (2), one can calculate the porosity of the seawater intruded portion of the aquifer by using only the TDEM data. Indeed, the salinity of the Mediterranean water is known and the fluid resistivity (ρ_w) calculated using a salinity/resistivity nomogram (Repsold, 1989) is 0.18 ohm-m. This value resembles the one obtained from borehole conductivity logs (V. Friedman, 2003, personal communication). Thus, the calculated porosity is $\phi = 0.36$.

When dealing with the shallow resistivity sequence, porosity calculations lead to great uncertainty due to the wide equivalence range of resistivities (ρ). By using $\rho_w = 10$ ohm-m, typical of the fresh waters in the area (based on both the above mentioned nomogram and nearby borehole conductivity data), the calculated porosities vary between 0.13 and 0.35.

The intermediate resistivity sequence may represent a wide range of salinities and, thus, assumingly, a wide range of ρ_w values. Therefore, contrary to the upper geoelectric layer, here both ρ and ρ_w values may vary considerably, thus even increasing the uncertainty in the calculated porosity.

It is interesting to note that the suggested approach can even be used to resolve non-uniqueness in the upper resistivity sequence. Indeed, applying equation (2) for $\phi = 0.36$ and $\rho_w = 10$ ohm-m, one obtains $\rho = 77$ ohm-m, that is fairly similar to the lowest resistivity value in the equivalence range of this sequence.

Some additional semi-quantitative information can be also obtained regarding the groundwater salinity within the intermediate resistivity zone above the interface. Indeed, in order to obtain porosity equal to 0.36, ρ_w must vary in a certain range corresponding to the equivalence range of ρ in this zone. Applying the Archie equation, the obtained ρ_w values vary between 1.4 and 3.5 ohm-m. This resistivity range roughly corresponds to salinities varying between 300 and 3000 mg Cl/L (Repsold, 1989).

The Judea Group carbonate aquifer

The Judea Group Cretaceous carbonate aquifer is a regional, 600-700 m thick, partly karstic aquifer, which consists mainly of limestones and dolomites. The aquifer is in places hydrologically connected to the sea and in places may host trapped seawater or brines.

TDEM measurements in this aquifer are limited in number as compared to the coastal aquifers due to the scarcity of boreholes that can be used for correlation and to the depth of the target in several cases. Nevertheless, previous preliminary TDEM measurements carried out in the Mt. Carmel area, not far away from the sea shore, managed to detect resistivities close to 1 ohm-m just below the fresh/seawater interface calculated using the Ghyben-Herzberg approximation (Goldman et al., 1988).

Recent deep TDEM measurements in this aquifer showed significantly higher resistivities varying between 5 and 10 ohm-m for roughly seawater salinities (Goldman and Kafri, 2002) either measured in boreholes or assumed according to available hydrogeological information. The Taninim deep well and the adjacent TDEM station N1 serve as a good example (Figure 5). The borehole penetrated the carbonate Upper Cretaceous aquifer that is known to be connected to the sea. The top of the aquifer is located at 580 m depth and, according to the Ghyben-Herzberg relationship, the fresh/sea water interface is expected at a depth of about 600 m. In fact, somewhat diluted seawater (17,000 mg Cl/L) was detected in the well at approximately the same depth and also appears as a resistivity drop in the TDEM resistivity/depth profile.

It should be noted that electric resistivity logs in boreholes (e.g. the Netanya I well), which penetrated saline water bodies, having roughly seawater salinities, also exhibited resistivities varying between 4 to 6 ohm-m in the relevant depth intervals.

The significant difference in the observed resistivities of seawater intruded granular clastic aquifers versus karstic carbonate aquifers is discussed below.

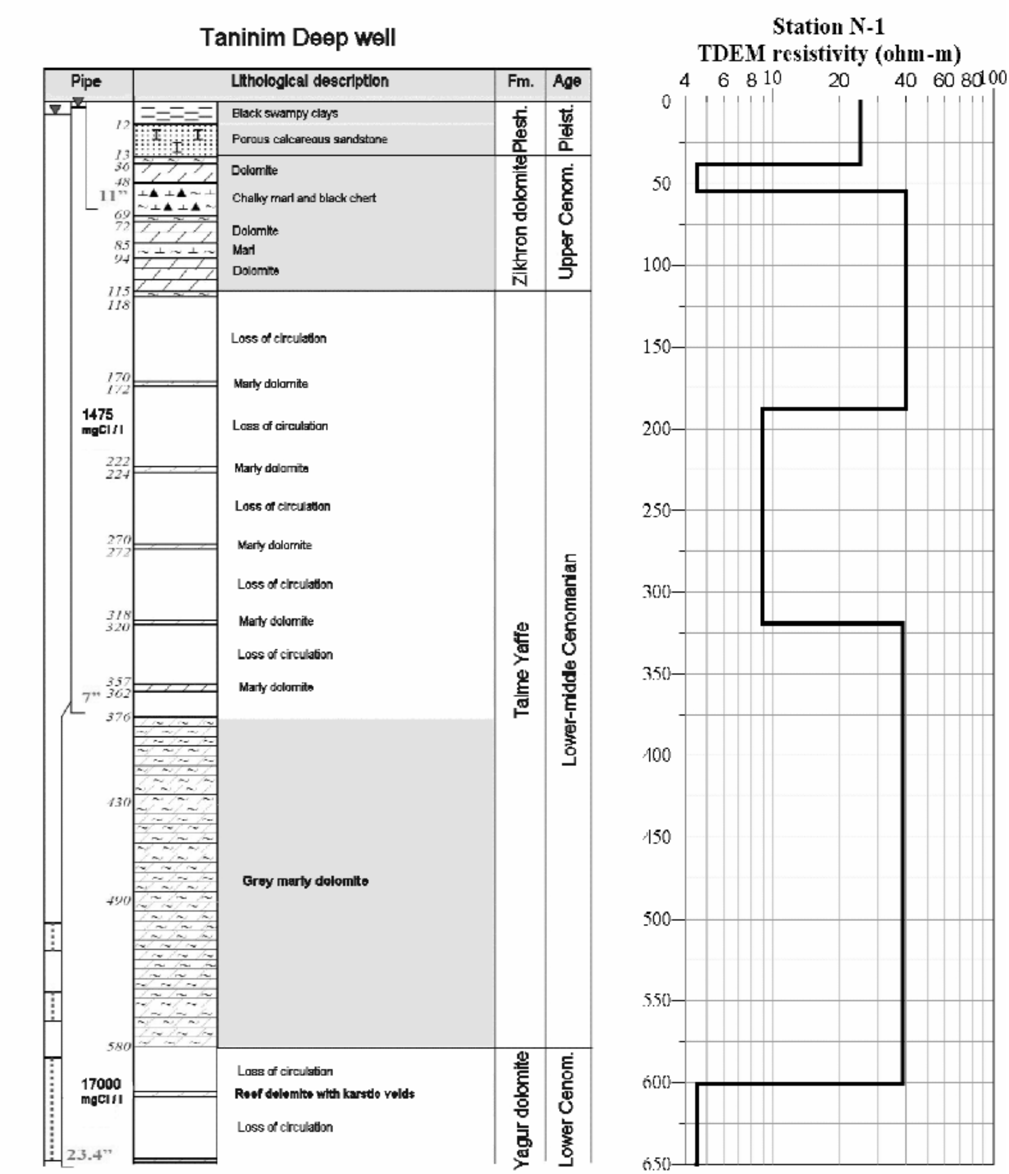

Figure 5. Comparison of TDEM results with borehole data from the Taninim Deep well. The low TDEM resistivities below 600 m depth are in agreement with the high groundwater salinity below the interface.

Porosity of granular and carbonate aquifers

The total porosity of the aquifer includes both the connected pores (or the aquifer storativity) and the unconnected pores. The total porosity is determined in a laboratory from core samples, whereas storativitivity is obtained from hydrological pumping interference tests, which are not available in most cases.

Two main types of aquifers exist, namely granular (Figure 6A) and those where porosity is mostly gained by solution (Figure 6B) and fracturing (Figure. 6C).

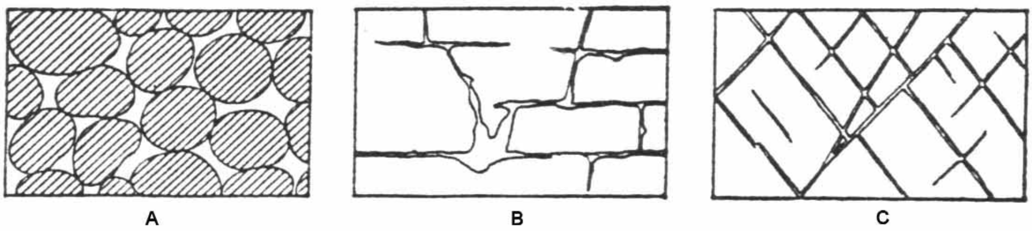

Figure 6. Diagram showing different types of rock interstices and the relation of rock texture to porosity. (A) Well-sorted sedimentary deposit having high porosity; (B) rock rendered porous by solution; (C) rock rendered porous by fracturing (after O.E. Meinzer, 1942).

In granular aquifers the porosity is mostly intergranular and the aquifers are not extremely heterogeneous. Thus, a porosity value obtained from a core sample might approximately represent larger volumes of the aquifer. As far as carbonate aquifers are concerned, the conditions are usually different. Most of the hydraulic conductivity, as well as storativitivity, are related to open fissures and solution channels and the aquifer is extremely heterogeneous. Very often an aquifer might have a double or even triple porosity system, which includes the initial matrix porosity and the secondary fissured and solution porosity. As a result, the hydraulic parameters obtained are scale dependent and often show higher parameters with increasing test radii, from a core sample to a pumping test and toward a regional study (Rover and Cherkauer, 1995; Zuber and Motyka, 1998; White, 2002).

It is thus essential to know, mainly for carbonate aquifers, whether a given porosity value represents a small or a large aquifer volume. Furthermore, when a resistivity value is used, it should be taken into account whether it is taken from a borehole log representing a small radius and thus lower porosities, or from TDEM measurements that integrate larger aquifer volumes and thus representing assumingly higher porosities.

Storativitivity values from pumping tests and hydrological models, as well as porosity values for both granular and carbonate aquifers, discussed herein, are summarized in Table 1.

Storativities in the coastal aquifers vary between 0.25 and 0.4 and they reflect the aquiferous portion of the sequence. It is assumed that in this portion the values are not quite similar to those of the total porosity. The situation, however, is different in the clayey aquiclude interlayers, which might exhibit a high non-effective porosity that contains trapped, unflushed fluids.

Table 1. Storativitivity and porosity of granular and karstic aquifers. The shaded locations are those, where TDEM data are available

Aquifer	Location/basin	Storativity	Reference
Coastal (granular)	Shiqma	0.3 – 0.4	Sellinger and Kapuler (1972)
	Nizzanim	0.3 – 0.4	Sellinger and Kapuler (1973)
	Western Galilee	0.25–0.35	Sellinger and Levin (1977)
	General	0.25	Mercado (2000)
Judea Group (carbonate)	Naaman	0.03– 0.08	Michaeli et al. (1973)
	Naaman	0.06	Mercado and Ben Zvi (1983)
	Kabri	0.04	Shachnai and Goldshtoff (1978)
	Yarkon-Taninim	0.05	Baida et al. (1978)
	Bet Shean	0.03-0.07	Goldshtoff and Shaliv (1979)
	Mt. Carmel Na'aman	0.03	Bar Yosef et al. (1997)
	Atlit 1 (Carmel)	0.013	Mercado and Ben Zvi (1983)
	N. Oren Carmel)	0.15	Mercado and Ben Zvi (1983)
		Porosity	
	Mt. Carmel reefs	Up to 0.15	Bein (1971)
	Ashqelon 3	0.15–0.20	Bein (1971)

The carbonate aquifer exhibits mostly storativities between 0.03 and 0.08, since the initial sedimentary (matrix) porosity is small and most of the storativity is related to secondary porosity gained by fracturing and solution. In places, like Mt. Carmel area, where a reef facies is known in both surface and subsurface, the initial porosity and, as a result, the storativity, are high, attaining a value of 0.15. In the case of the discussed carbonate aquifers one should take into account, in addition, the large aquifer volume with a small initial non-effective porosity.

Calculated versus actual resistivities

The expected resistivities for different aquifers and related porosities and hosted saline waters could be calculated using the Archie equation.

The ρ_w values for the different saline potential end members can be obtained from a resistivity/salinity chart (Repsold, 1989). According to this chart, the resistivity of normal seawater ρ_w = 0.18 ohm-m. The ranges of porosities (ϕ) were those described above for the both granular and carbonate aquifers (Table 1). The resultant calculated ρ values for different porosities are given in Table 2. For convenience sake, the data relevant to actual TDEM measurements are shaded in both tables.

It can be seen that in the Mediterranean coastal aquifer, subjected to seawater intrusion, the TDEM measurements yielded resistivities between 1 and 2 ohm-m, similar to the ones shown in Table 2 for the porosity values greater than 0.3. It should be noted that in the lower sub-aquifers of the Mediterranean coastal aquifer the proportion of the clayey interlayers is higher as compared to the aquiferous ones. As a result, in some cases where fresh waters were sampled in the aquiferous parts, TDEM resistivities were below the expected values. It is assumed that in these cases the TDEM measurements integrate both aquiferous and clayey layers with apparently host trapped and unflushed saline waters, resulting in the lowering of resistivities of the entire sequence. The same phenomenon was encountered in the Gulf of Elat coastal aquifer.

Table 2. Bulk resistivities calculated using Archie's equation for different porosities and water resistivities and actual resistivities measured by TDEM in different aquifers. In the calculations for granular aquifers, parameter α in Eq. (1) was assigned a value of 1.23, normally used for sands.

Aquifer and saline water body	ϕ	ρ_w (ohm-m)	ρ (ohm-m)	Calculated TDEM
Coastal, granular aquifer saturated with normal seawater.	0.2 0.3 0.4	0.18 0.18 0.18	3.6 1.6 0.91	Between 1 and 2; in most cases 1.7
Judea Group carbonate aquifer saturated with normal seawater.	0.03 0.05 0.07 0.08 0.15 0.20	0.18 0.18 0.18 0.18 0.18 0.18	200 72 37 28 8 4.5	Between 4 and 8

The situation seems to be more complicated regarding the Judea Group carbonate aquifer when intruded by seawater. The calculated resistivities for relatively low storativities (0.03-0.08) are significantly higher (between 28 and 200 ohm-m) than those measured by TDEM, and only for higher storativities (0.15-0.20) resistivities are closer to actual TDEM or well log resistivities.

Measurements in Mt. Carmel, near N. Oren, where high storativities were measured, yielded resistivities around 1 ohm-m and in other places approximately between 5 and 10 ohm-m (Goldman and Kafri, 2002) assumingly due to lower storativities. It is noteworthy that in a few wildcats (Gaash 2, Netania 1 and Cesarea 3), carbonate sequences saturated with waters of seawater salinity (18,000 – 20,000 mg Cl/L) exhibited well log resistivities between 4 and 6 ohm-m, similar to the TDEM resistivities.

Conclusions

TDEM resistivity measurements allow accurate detection of the geometry and bulk resistivity of seawater intrusion into both granular and apparently carbonate coastal aquifers. This, in turn, facilitates more accurate estimation of porosity of seawater intruded parts of the aquifers, since both the bulk and the fluid resistivities are accurately determined in these parts. The obtained porosity, in most cases, adequately represents the entire aquifer or its major parts, since it is generally similar both above and below the freshwater/seawater interface.

Comparison of the calculated porosities and storativities obtained from pumping tests and hydrological models show generally good agreement in the granular coastal aquifers of Israel. The situation is more complicated in the carbonate karstic aquifers where both geophysical and borehole data are scarce and storativities vary over an extremely wide range. For these aquifers, an extensive and more detailed TDEM survey is proposed to provide a sufficient amount of the resistivity data.

It is essential to calibrate the method in coastal granular and, more important, carbonate aquifers in regions of available hydrological data (water levels, salinities, porosities). Suitable study areas, among others, exist along the Mediterranean coasts in southern Spain, southern France, Greece and the Adriatic coasts, as well as in the southern Atlantic coasts of the United States.

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