On the exploration of a marine aquifer offshore Israel by long-offset transient electromagnetics

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ABSTRACT

The existence of aquifers extending from land beneath the sea floor up to a distance of several kilometres has been observed and examined all over the world. The coastal aquifer of Israel is a heavily used groundwater reservoir which has to be constantly monitored to ensure the drinking water supply. Former land-based electromagnetic measurements show that it is, in several places, blocked to seawater intrusion and is consequently a candidate for submarine extension. Multicomponent long-offset transient electromagnetic measurements were carried out offshore on the coast of Israel. We deployed a 400-m-long grounded dipole as transmitter and several electric and magnetic receivers on the sea floor up to a distance of 4.8 km from the coast. Altogether, we deployed 8 transmitter positions and received data sets at 14 receiver stations onshore and offshore, with offsets of mostly 400 and 800 m. In this paper, we present the survey and 1D Occam and Marquardt inversions of the offshore horizontal electric components in the broadside and inline configuration. In addition, the vertical magnetic component in the broadside position is also considered. Only single inversions, both single offset and single component, were used to detect the aquifer under sea sediments. We prove the submarine existence of the Israeli coastal aquifer up to a distance to the coast of approximately 3.2 to 3.6 km using all measured long-offset transient electromagnetic components. In addition, we present modelling studies with synthetic data derived from a subsurface model adjusted to our survey area with very shallow water from 10 to 50 m. As part of the planning before the survey, a parameter study of the expected subsurface, the examination of the airwave phenomenon and the justification for our 1D inversion strategy are shown. More detailed eigen parameter analyses are added to explain the measured data.

Key words: Marine electromagnetics, Long-offset transient electromagnetics (LOTEM), Submarine aquifer, Israel coastal aquifer.

INTRODUCTION AND THE LONG-OFFSET TRANSIENT ELECTROMAGNETIC MEASURING SYSTEM FIELD SET-UP

The coastlines of the world are not necessarily a boundary for coastal groundwater systems: aquifers reaching from land to the sea below the sea floor are a well-known phenomenon and reported from all over the world (Post et al. 2013). Fresh groundwater may discharge into the sea, and saltwater can intrude the landside aquifer. This process is called submarine groundwater discharge or recharge, respectively (e.g. Burnett et al. 2003). The existence of submarine fresh groundwater bodies extending offshore to distances between a few metres to several tens of kilometres was reported and examined all over the world with several methods, for example, direct measurements of discharging groundwater offshore Japan (Taniguchi

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and Iwakawa 2004), airborne temperature measurements offshore Lebanon (Shaban et al. 2005) or Tracer measurements of isotopes offshore Israel (Yechieli et al. 2009) or offshore the eastern coast of the USA (Swazenski et al. 2001). Electromagnetic methods are very suitable to distinguish saltwater from freshwater, due to the close physical relationship between salinity and electrical resistivity Goldman et al. (1988) reports one of the first measurements to detect seawater intrusion into a coastal aquifer. Other examples of using electromagnetic methods to distinguish saltwater from freshwater are Land et al. (2004) on the eastern coast of the USA or Duque et al. (2008) in Spain.

With more than 40% of the global population living within 100 km from the coast (Post et al. 2013), the significance of coastal aquifers and possible extension under the sea floor is undoubted. Especially in areas like the East Mediterranean, the supply of drinking water has always been a complex task. The Mediterranean coastal aquifer of Israel is one of the main groundwater resources of the country and is monitored since decades by permanent boreholes as described, for example, by Nativ and Weisbrod (1994). Figure 1 shows a typical hydrogeological cross section of the Mediterranean coastal aquifer in Israel. It is well known that the aquifer is grouped into four subaquifers, which are nowadays managed separately (Nativ and Weisbrod 1994). The upper two subaquifers (A, B in Fig. 1) are known to be subjected to lateral seawater intrusion and to downward anthropogenic pollution. The lower ones (C, D in Fig. 1) are assumed to be, in

Figure 1 Typical hydrogeological cross section of the Mediterranean coastal aquifer in Israel (taken from Haroon et al. 2018). The coastal aquifer is divided into four subaquifers: the two upper (A, B) and the two lower subaquifers (C, D). The position of this cross section can be seen in Figure 3.

Figure 2 Merged 1D inversion results from onshore SHOTEM data (taken from Goldman et al. 2005). The profile is perpendicular to the coast and shows the lower and upper subaquifers. The position of this profile can be seen in Figure 3.
places, blocked to the sea or respectively to extend under the sea (Nativ and Weisbrod (1994).

The division into subaquifers can be shown with electromagnetic measurements: Figure 2 shows 1D inversion results of short offset transient electromagnetics (SHOTEM) measurements by Goldman et al. (2005). The SHOTEM data were observed perpendicular to the coast, and the shallow and deep subaquifers were clearly detected. The SHOTEM method uses the TE-transmission mode; therefore, the resolution of the resistivity value of high resistive bodies is poor, but the detection and distinction between salty, brackish and fresh water works perfectly (Goldman et al. 2005). The authors proved this fact by water samples from boreholes. In addition, Kafri and Goldman (2006) showed the distribution of resistivity values of the deeper subaquifers along the coast of Israel (Fig. 3). The area from the cities of Ashdod in the south to Bat Yam in the north, the so-called Palmahin disturbance, has to be highlighted: Here, resistivity values show constant high values, which argues against lateral seawater intrusion. Offshore measurements by Goldman et al. (2011) and Levi et al. (2018) prove the extension of the aquifer under the Mediterranean Sea in this region. For this survey, the authors used a grounded dipole emitting TE and TM mode and recorded the vertical magnetic field at a short offset (e.g. 50 m) in the broadside configuration.

For this work, the long-offset transient electromagnetic measuring system (LOTEM) of the University of Cologne was adapted to the marine environment. The system consists of a grounded transmitter dipole and several receivers for magnetic and electric components (Strack 1992). Transmitters and receivers were deployed on the sea floor; we did not use a towed system.

We present multicomponent LOTEM measurements on an offshore profile located at the Palmahin disturbance (Fig. 3). On a 4.8-km-long profile perpendicular to the coastline in the northern part of the Palmahin disturbance, 8 electric dipole transmitters were placed on the sea bottom and electric and magnetic field components were measured using inline and broadside configurations. Figure 4 shows the measured transmitter and receiver positions. By definition of the LOTEM method, the distance between the transmitter and the receiver (i.e. offset) is larger than the depth of the examined target. Justified by the onshore SHOTEM results (Fig. 2), the expected aquifer target depth is 100 to 200 m and our preferred offsets were 400 and 800 m. Data were collected in the broadside position, with the receivers on a line perpendicular to the middle of the transmitter, and in the inline position, with transmitter and receiver directions on one line.

![Figure 3 Resistivities of the two lower subaquifers (C and D in Figs. 1 and 2) in Israel derived from the 1D inversions from onshore SHOTEM measurements (taken from Kafri and Goldman 2006). Our measured LOTEM profile is additionally drafted as a red line; the cross section shown in Figure 1 is drafted as a blue line and the SHOTEM profile shown in Figure 2 is drafted as a black line.](image)
Transmitters were deployed perpendicular and parallel to the coastline. In this paper, we show the horizontal electric data in the broadside and inline configurations and the vertical magnetic field component in the broadside position and their interpretations by 1D conductivity models.

SYNTHETIC MODELLING STUDIES

This section deals with modelling studies done before and after the acquisition and inversion of the field data. Although not in chronologic order, the studies are grouped together before showing the field data to provide the reader with the results to understand and interpret the real data results.

In order to study the resolution of thin resistive layers, modelling studies in the marine environment have already been realized by, for example, Key (2009). The author concentrated on the canonical model as introduced by Constable and Weiss (2006). Key’s (2009) studies based on joint inversions of different offsets in the frequency domain, examining single and joint inversions of data sets for both configurations and multiple components. The transmitter was floating in the sea, and the receivers were placed on the sea floor. Key’s (2009) results favour the inline electric component for resolving the thin resistive layer in the canonical model. Connell and Key (2013) compared numerically time and frequency-domain data from the inline configuration regarding their resolution of resistive targets. They used, besides the canonical model, a model with 50-m water depth. Only the inline component was examined. For this shallow water, the transmitter was on the sea surface and the receivers were placed on the sea floor. Both mentioned studies used a deep resistive target with 1000 m of overburden sea sediments. We present studies for detecting and resolving a resistive layer for the hydrogeological situation at the Mediterranean coast of Israel: very shallow water with a depth of less than or up to 50 m and a shallow target with maximum 90 m beneath the overburden sea sediments. In contrast to the mentioned publications above, both transmitter and receivers were placed on the sea floor. We examine single-offset Occam inversions of the different configurations and single components in the time domain.

The modelling study with synthetic data was realized with the one-dimensional forward and inversion code MARTIN by Scholl and Edwards (2007). Figure 5 shows the...
Figure 6 Parameter study of the conductivity model shown in Figure 5 with $d_{H2O} = 30$ m water depth and 400 m offset between the transmitter and the receiver. The thickness, depth and resistivity of the marine groundwater layer were varied systematically and compared with a fixed three-layered model (see text). The variations of the parameter are arranged in the rows, the components in the columns. We show the ratio of the transients of the varied subsurface to a homogeneous half space of 1 $\Omega$ m. A threshold of 10% relative difference between varied and fixed model is plotted as dashed lines.

Figure 7 1D conductivity models for the study of the airwave: (a) full space sea water, (b) double half space seawater/sediments and (c,d) realistic subsurfaces with water layer and with/without resistive target layer (grey). Used subsurface model in the modelling study. The resistivity of the seawater, measured with a conductivity, temperature and depth probe at the survey area, was nearly constant, and, therefore treated as a constant value of 0.2 $\Omega$ m over the complete water depth.

First, the detectability of a shallow resistive target with multicomponent electromagnetics in very shallow water is qualitatively examined by a parameter study. Afterwards, we examined the shallow-water problem in marine EM, the so-called airwave, and its influence on our measurements. Third, we justify the inversion of obvious 2D data sets with a 1D inversion code by comparing synthetic data from one- and two-dimensional offshore conductivity structures. Finally, we present an eigen parameter study and inversion of synthetic data to prove the detectability of the poor conductive target layer by each measured component.

Parameter studies for the resistive marine aquifer

As a first overview of the detectability of the target resistive layer, a parameter study was done and interpreted qualitatively. We derived synthetic data from the subsurface model
shown in Figure 5 using inline and broadside configurations. Perturbation of the following parameter of the resistive marine groundwater layer was made: the top boundary \(d_1\), the thickness \(d_2\) and the resistivity \(\rho_2\). The water depth \(d_{H2O}\) was 30 m, and 400 m was chosen as a distance between the transmitter and the receiver. If the parameters were not varied, they were kept at \(d_1 = 70\) m, \(d_2 = 100\) m, \(\rho_2 = 10\ \Omega\)m (fixed reference model). Figure 6 shows the ratios between switch-on transients of those models and switch-on transients of a homogeneous half space of 1 \(\Omega\)m. As a threshold of detection, we chose a difference of 10% between those transients.

Electric field components (Fig. 6, left and middle rows) are sensitive to changes in the model over the whole examined time range. For the inline component, the target resistive marine groundwater layer has to be thicker than 10 m (Fig 6, left top) and not deeper than 125 m (Fig. 6, left middle) to result in a ratio over the threshold. Lower target resistivities could be better resolved (Fig. 6, left bottom), values larger than 6 \(\Omega\)m give very similar ratios indicating high resistive values for the target layer and they cannot be distinguished from each other. Electric field values in the broadside position (Fig. 6, middle) are, for a layer thicker than 10 m, over the given threshold for all models. All parameters show clear different ratios indicating good resolution. To resolve the resistive layer using the vertical magnetic component \(dBz/dt\) in the broadside configuration, the layer has to be thicker than 50 m (Fig. 6, right top). For the other two parameters (Fig. 6, right middle and bottom), one can distinguish between half space and the existence of the resistive layer, but the varied parameter values have a rather similar ratio to the half space, that is, they are not well resolved.

Figure 8 Maximal amplitudes and signatures for the four models shown in Figure 7. Data derived in the broadside configuration for two different offsets. Upper part (a,b): maximal amplitudes of the electric fields for different depths of water. Lower part (c, d): normalized impulse responses for different depths of water.
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This qualitative parameter study shows that the detection of a resistive marine aquifer layer is possible using all examined LOTEM field components; however, the broadside electric field component seems to be the best in order to resolve the model parameters of the target layer for our specific survey area.

Influence of the airwave

According to Weidelt (2007), there are three pathways of the electromagnetic signal from the transmitter to the receiver: (1) directly through the water, (2) through the subsurface and (3) from the transmitter up to the air–water boundary, along and down to the receiver. This third pathway is called the ‘airwave’.

The airwave is an important feature for shallow marine electromagnetic measurements, as it can mask the target signal and lead to a wrong geological interpretation of the measured data. For frequency-domain CSEM systems, common methods have been developed to mitigate the parasitic airwave signal (Chen and Alumbaugh 2011). Time-domain CSEM offers a time window where no airwave signal is present (Weiss 2007). The interaction between water depth, transmitter to receiver offset and target depth is very complex and not comparable with other settings in the literature; we examined the effect of the airwave on the measured transients for our subsurface situation: very shallow water (10–50 m) and a target layer in 100 m depth beneath the sea level.

We followed the procedure of Weiss (2007). The author examined the phenomenon of the airwave for a minimum of 100 m water depth and a minimum of sediment thickness above the target of 500 m. He used the inline Ex component and examined the phenomenon for different offsets up to 10 km.
Figure 10  Comparison of one- and two-dimensional data sets. The upper subplot shows the 2D model, which is used to derive the 2D data sets of broadside electric field. One example pair of the transmitter and the receiver is given. The lower subplot shows the error between the 2D data and 1D models with different water depths between the depths of transmitter $d_{Tx}$ and receiver $d_{Rx}$. The set-up example from the upper subplot is marked in red. All positions of the real measured broadside Ex data sets were examined and shown.

We adopted Weiss’s (2007) procedure for very shallow water (depths up to 50 m) and a shallow resistive layer with overlaying sediments of maximal 90 m. The offsets were chosen according to the real measured offsets: 400 and 800 m as it was the case for our field measurements at the coast of Israel. We used the electric fields in the inline and broadside configuration.

Figure 7 presents the synthetic models for this study: Model A (Fig. 7a) is a uniform double half space with a resistivity of water of 0.2 $\Omega$m. Numerically, an infinite water depth is treated as 10 km of water. The signal can only travel through the water in this case, represented by blue in Figures 8 and 9. Figure 7(b) shows model B, a double half space with resistivities of water (0.2 $\Omega$m) and sediments (1 $\Omega$m). The signal can now also travel through the sediments, which is represented by grey in Figures 8 and 9. In model C (Fig. 7c), we considered real water depths between 10 and 50 m at the coast of Israel. The phenomenon of the airwave is represented by black in Figures 8 and 9. As the last model D (Fig. 7d), a resistive target layer (100 $\Omega$m) located 100 m under the water surface and with 100 m thickness was considered. The thickness of the sediments will be less if the water depth is varied in the model D in Figure 7. This model is represented by red in Figures 8 and 9.

We show the impulse response of the signal by calculating the time derivative of the electrical signal and normalized it to its largest value for a better presentation according to Weiss (2007). Figure 8 shows the results for broadside and Figure 9 for the inline configuration: In each case, the maximum amplitude of the electric fields is plotted as a function of varying water depths and for chosen conductivity models (Fig. 7) in the upper subplots. In the lower subplots, normalized impulse responses of each model are presented for different water depths and sediment thicknesses.

Maximum amplitudes of the models with target layer (and airwave) are larger than the amplitudes of models with sediments (and airwave) only (see Fig. 8a–c and Fig. 9a–c). Weiss (2007) suggested an observation window which is bordered by the signature of the airwave for early times and the signature of the seawater for late times. Inside this window, the signatures are not covered by the signature of the airwave.

In our models, the signature of the seawater also determines a border for the late times. For early times, we have a different situation compared with Weiss (2007): For a water depth of 10 m, the airwave clearly arrives at first. This arrival moves to later times for deeper water depths. In addition, the arrival of the signature of the model with the target layer moves towards early times and clearly arrives at first for water depths of 50 m. As a consequence, at medium water depths, the signature of the airwave arrives at the same time as the signature of the target layer: for the broadside configuration at a water depth of 35 m (at 400 m offset, see Fig. 8c) and
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Figure 11 Linear eigen parameter analysis for two different offsets (rows) for all existing components (columns) over an one-dimensional subsurface. The columns (EP: eigen parameters) in each subplot represent the linear combination of the transformed model parameters. Their contribution is represented by the radius of the circles. The normalized eigenvalues (EV/EV1) are also given under each subplot. An EV smaller than 3% of the biggest EV1 is considered as not resolved.

40 m water (at 800 m offset, see Fig. 8d) and for the inline configuration at a water depth of 25 m (at 400 m offset, see Fig. 9c) and 35 m (at 800 m offset, see Fig. 9d).

Fortunately, the maximal amplitude of the model with a target layer is, in all cases, more than double compared with the maximal amplitude of the airwave. We take the airwave effect into account by including the water layer as a fixed model parameter in the inversion process and, therefore, no disturbance during the inversion of data sets is expected where the signature of the target is overprinted by the airwave. Although the arrival time of the airwave depends strongly on the water depth, we assume that little errors in the depth do not lead to wrong inversion result because the amplitude of the airwave is only half of the amplitude of the subsurface signal. We realized 1D inversions using the synthetic data derived from the conductivity models of Figure 7. As expected, all the original models could be resolved by a good data fit (Lippert 2015).

Using a 1D inversion code for a 2D data set

Our next synthetic modelling study justifies the use of a 1D inversion for obvious 2D data. For this, we compare the broadside electrical field component with offsets of 400 and 800 m for synthetic data sets derived from a 2D subsurface with data sets derived from a 1D subsurface. The influence of multidimensional structures on the interpretation with 1D models was analysed before by Scholl (2005), but here we concentrate on our specific marine case. The 2D forward calculations for the presented study were realized using the finite-difference code SLDMEM3t by Druskin and Knizherman (1988). A typical grid check was done by comparing synthetic data of a homogeneous half space and 1D conductivity model with an analytical solution. The percentual errors of this grid check are used for calculating the misfit below (see equation (1)).

On the one hand, we have derived synthetic data of a realistic 2D model (Fig. 10, top) from the nautical map (Fig. 4).
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Figure 12 Occam 1D inversion of synthetic data derived using the conductivity model in Figure 5 with different standard deviations. The synthetic three-layer model and model parameters of the sea are also plotted in blue.

There is a poor conductive layer of 100 $\Omega$m with a thickness of 100 m starting at a depth of 100 m, which is embedded by 1 $\Omega$m sediments. The transmitter and receiver positions in this synthetic model and their water depths (i.e. $d_{TX}$ and $d_{RX}$) are the same as in the real measurements. Because we are close to the coast, $d_{TX}$ usually differs from $d_{RX}$ due to the topography of the sea bottom.

On the other hand, we have a synthetic 1D three-layer model where $d_{TX} = d_{RX}$ (comp. Fig. 5). Its parameters are: $\rho_{1,3} = 1 \Omega$m and $\rho_{2} = 100 \Omega$m; the resistive layer always starts in a depth of 100 m and its thickness ($d_2$) is 100 m. We varied the water depth of this 1D model between $d_{TX}$ and $d_{RX}$ and compared those transients with the 2D solution.

The misfits between the 1D and 2D forward data are calculated as

$$\chi = \sqrt{\frac{1}{N} \sum_{i=1}^{N} \left[ \frac{d_i (2D) - d_i (1D)}{\sigma (2D)^i} \right]^2}.$$  \hspace{1cm} (1)

Here, $N$ is the number of the data points $d_i$, which are the same for 1D and 2D and $\sigma_i (2D)$ are the data errors received from the grid check for the 2D forward code, as mentioned above. In Figure 10 (bottom), the misfits $\chi$ (Eq. (1)) between the 2D and 1D forward solutions are plotted as a function of the water depth of the 1D models. There is always a clear misfit $\chi$ minimum when using the mean water depth between transmitter $d_{TX}$ and receiver $d_{RX}$. When using a mean water depth, the forward calculated transients of 1D and 2D only differ to an error value of approximately 3 so that we consider a 1D inversion as reliable for the most broadside electric field data of the survey area. Near the coast, where the slope at the sea bottom and the influence of the coast is big, only misfit values of minimum 10 can be achieved by using mean water depths. The consequence is to handle 1D results with care when the data sets are measured close to the coast.

According to this synthetic data modelling, we use mean water depths between the transmitter and the receiver for the 1D inversion of the field data. Ji, Tezkan and Li (2018) carried out more detailed synthetic data modelling on the effect of a sloped sea floor including systematic studies on the influence of the slope angle up to a minimum of 50 m water depth. The authors came to the same conclusion of using the mean water depth between the transmitter and the receiver.

Detailed synthetic modelling after data acquisition and inversion of the LOTEM field data

After the 1D inversion of the LOTEM field components, the resolution of the model parameters of the resistive marine
aquifer layer at the coast of Israel is studied by synthetic modelling using single inline and broadside configurations of electric and magnetic field data.

For the quantitative examination of the resolution of a resistive target layer, we choose—in addition to the parameter study—a linear eigen parameter analysis of synthetic data sets as used by Scholl and Edwards (2007).

As synthetic model, we used the three-layered subsurface (see Fig. 5) with \(d_{\text{H2O}} = 30\) m of seawater above. The target layer starts at a depth of 100 m from the sea surface (i.e. \(d_1 = 70\) m) and is \(d_2 = 100\) m thick with a resistivity of \(\rho_2 = 100\ \Omega m\). The surrounding first and last layers represent sea sediments with a resistivity of \(\rho_1 = \rho_3 = 1\ \Omega m\).

Synthetic data were calculated for the inline and broadside horizontal electric fields with a switch-off process and for vertical magnetic components. As offsets, 400 and 800 m were chosen. The Jacobian matrices were set up via perturbation of the model parameters by 10%. The model parameters in the Jacobian are transformed logarithmically, and the data values are transformed by an area sine hyperbolic function as described in Scholl and Edwards (2007). This Jacobian \(J\) was now split in three matrices using a singular-value decomposition (SVD): \(J = USV^T\). \(U\) and \(V\) are \(M \times M\) and \(N \times N\) matrices, where \(M\) is the number of model parameters and \(N\) is the number of data points. \(S\) is an \(M \times N\) diagonal matrix containing the eigenvalues (EV) of \(J\). We analyse in detail the so-called eigen parameters (EPs) in the matrix \(V\). Eigen parameters are linear combinations of the transformed model parameters.

The results are presented in Figure 11. Each column (labelled EP\(_1\) to EP\(_5\)) is one eigen parameter and shows the combination of the transformed model parameter. The contribution of the particular model parameters is represented by the radius of the circles. The EPs are sorted by their decreasing eigenvalues (EV). The standardized EV (to the biggest EV\(_1\)) are also shown in Figure 11. We choose a ratio of 3% as a threshold value under which the EP will not be involved during the inversion process, that is, this combination of parameters is not detectable. This threshold is similar to the values used by Scholl and Edwards (2007).

In all six subplots of Figure 11, the first EP (with the biggest EV) is dominated by the model parameters of the first layer (\(d_1\) and \(\rho_1\)). For the broadside Ex configuration (left column in Fig. 11), all EP are resolved, that is, have a bigger EV than the threshold.

For an offset of 400 m, the smallest EV (of EP\(_3\)) is still \(1/4\) of the biggest EV. The parameters of the target (\(d_2\) and \(\rho_2\)) are well represented in various combinations. The first EP of the inline component (middle column in Fig. 11) is obviously the dominant factor. All the other EPs have, compared with the broadside electrical component, a very small EV. According to the chosen threshold, not all EP are detectable: for a 400-m offset only the two biggest ones, for an 800-m offset the first three EPs. For a 400-m offset, the target parameters...
Figure 14 Two examples of data fit and resulting models for inline electric data. On top a plan view shows all measured transmitter and receiver positions. The position of the presented transmitters and their corresponding receivers is drawn in red. On bottom left, a data set where the aquifer is not present is shown; on bottom right, a data set where the aquifer is present. CF is the calibration factor. Results are presented from the two Occam inversions with roughness 1 (black) and 2 (blue) and the best fit model from the Marquardt inversion (red).

$d_2$ and $\rho_2$ are not included in EP with EV over the threshold. For 800 m, only the thickness $d_1$ is present. For the broadside Hz component (right column in Fig. 11), only the thickness of the target $d_2$ is present in an acceptable EP. The smallest EP consists only of the target resistivity $\rho_2$ but its EV with less than 1% of the biggest EV leads to clear non-observance in the inversion process.

Eigen parameter analysis confirms the parameter study that the broadside Ex component seems to be the most suitable configuration for detecting the target aquifer at the coast of Israel and for correctly resolving its model parameters.

We have also studied the effect of data noise on the resolution of poor conductive aquifer layer using the inversion of inline and broadside electrical data.

Moghadas, Engels and Schwahlenberg (2015) tested different 1D inversion strategies for joint multi-offset data sets. To simulate our real measurements, we concentrated on the commonly used Occam single-offset inversion. These inversions were done with synthetic data sets, and only Occam inversion calculations with the regularization (roughness) 1, as described in Constable, Parker and Constable (1987) are shown in the following.

Again, a three-layered subsurface is used (comp. Fig. 5) as a conductivity model. This time with $d_{120} = 40$ m of seawater above. Again, the target layer starts at a depth of 100 m from the sea surface (i.e. $d_1 = 60$ m) and it is $d_2 = 100$ m thick with a resistivity of $\rho_2 = 100 \, \Omega\, \text{m}$.

Typically, some Gaussian distributed noise is added to the synthetic data before inversion. But here we want to examine the influence of the standard deviation (STD) on the resulting model. For real data, the STD is calculated during the stacking process of the single measurements. We added the same STD, that is, same % of error, to each data point before running the inversion. For each inversion run, we used different values for the STD, starting with 0.1% for nearly perfect data. One of the stopping criteria in the inversion code (Scholl and Edwards 2007) is a good data fit which is checked by the data misfit

$$
\chi = \sqrt{\frac{1}{N} \sum_{i=1}^{N} \left[ d_i - f_i \right]^2 \sigma_i^2}.
$$

(2)

Here, $N$ is the number of the “measured” data points $d_i$ in this case the data derived from the original synthetic model.
The forward data of the model \( m \) is \( f_i(m) \) and \( \sigma_i \) are the data errors.

Figure 12 shows the inversion results of the synthetic model. The original three-layer model and the water depth are also plotted. The broadside ex component (Fig. 12, left two subplots) shows for both offsets the resistive layer in the correct depth with an adequate resistivity value. The results are satisfying up to an STD of 6%. For a greater STD on each data point, the resistive target layer cannot be determined for a 400-m offset.

The result of the inline Ex component for a 400-m offset (Fig. 12, middle left subplot) indicates the target layer but with too small resistivity values. For perfect data (STD = 0.1 %), a value of approximately 10 \( \Omega \)m is given. The result for an STD of 0.5% or greater does not show any resistive target layer. The results for the 800-m offset (Fig. 12, middle right subplot) also do not show the correct model. The broadside \( dBz/dt \) component (Fig. 12, right two subplots) shows only for an offset of 800 m a correct result, but underestimates again the resistivity value. The results are stable up to an STD of 3%. The big \( \chi \) values result from very small data values close to zero at early times.

According to the inversion results with 1D synthetic data, the broadside Ex component is the most suitable configuration for detecting the target aquifer: In the shallow sea on the coast of Israel, it is the only component which shows adequate resistivity values. Both broadside components, ex and \( dBz/dt \), are robust to data errors, whereas the inline Ex component needs an unrealistic perfect data quality of error bars smaller than 1% to at least dissolve a layer with a slightly higher resistivity.

**MARINE LONG-OFFSET TRANSIENT ELECTROMAGNETIC MEASURING SYSTEM FIELD SURVEY AT THE COAST OF ISRAEL**

Marine long-offset transient electromagnetic measuring system (LOTEM) measurements were carried out on the coast of Israel. Transmitters and receivers were put on the sea floor in offsets of 400 and 800 m. The parallel direction of the transmitter determines the \( x \)-direction (broadside). The magnetic receiver coils were checked manually by scuba divers with a magnetic compass to ensure correct orientation. We deployed a 400-m transmitter dipole on the sea floor and transmitted...
an alternating current signal of 20 A peak to peak with a 50% duty cycle. Figure 13 shows the transmitted current (Fig. 13a) and two examples of time series of received data (Fig. 13b,c).

Figure 4 shows all measured transmitter and receiver positions: Transmitters were deployed perpendicular and parallel to the coastline. Signals were received in the broadside and inline configurations. Although we collected more than 5/4 of the signal period (see Fig. 13) in each single measurement, we used switch-off processes exclusively because the behaviour of the transmitter (GGT30; Zonge Engineering) is more ideal at switch-offs. Those switch-offs were filtered by 50 Hz, and every data point was stacked individually to get the final transients presented here. Only 50-Hz periodic noise was present, and its energy percentage in the energy of the whole collected signal constantly decreased with distance to the coast (Lippert 2015). From the stacking of all measured single transients, we received the standard deviation (STD) of each data point.

A noise level was determined for every receiver position by recording the signals when the transmitter was turned off. We used an upper envelope function of the smoothed signals to define the noise level for a signal/noise (S/N) ratio of 1.

According to the amount N of the measured single transients, this noise level gets shifted by a factor \( \sqrt{N} \). In the data examples of Figures 14 to 16, these noise levels, normalized to receiver length or area, are also shown. Near the pipeline, parallel to the coast, which can be seen in Figure 4 as a non-anchor zone, no anomalies in the noise, neither periodic nor aperiodic, were observed (Lippert 2015).

Nearly all measured data points of the electric fields are above their corresponding noise levels. The data points of the magnetic fields have mostly an S/N ratio < 1 in the early time range (see as an example Fig. 15). We excluded those parts of the transients from the inversion.

For the inversion process, an additional inversion parameter is also considered in the inversion code: the calibration factor (CF in Figs 14–16). This linear time-constant factor absorbs effects of the so-called transmitter overprint, which is originated in small anomalies around the transmitter (Strack 1992; Hördt and Scholl 2004). Geometric errors from an imprecise set-up can be corrected by considering the CF in the inversion procedure (Strack 1992). This last fact is also valid for our relatively short offsets. Lippert (2015) examined this fact by 1D synthetic forward modelling.

Figure 16 Two examples of data fit and resulting models for broadside electric data. On top a plan view shows all measured transmitter and receiver positions. The position of the presented transmitters and their corresponding receivers is drawn in red. On bottom left, a data set where the aquifer is not present is shown; on bottom right, a data set where the aquifer is present. CF is the calibration factor. Results are presented from the two Occam inversions with roughness 1 (black) and 2 (blue) and the best fit model from the Marquardt inversion (red).
Figure 17 Resulting 1D models from Occam inversions with roughness 1 of the broadside horizontal electric fields. Transmitter positions are drawn as red squares on the sea floor. Each resulting model is plotted at the receiver position. Error values of the data fit are also given above each model. The black line is the depth where the two regularizations from the Occam inversion show different behaviour in the model and can be interpreted as the depth of investigation (DOI) (see text).

For the interpretation of the measured data, we started with 1D Occam inversions, which try to interpret the data by a smooth subsurface model using nearly as many subsurface layers as data points. Occam inversion uses a homogeneous half space as starting model and is therefore independent from a human interpreter. Out of the Occam results, we deduced three-layer starting models for the Marquardt inversion (Marquardt 1963). Marquardt inversion represents more realistically the subsurface with sharp layer boundaries and reproduces the true resistivity values much better than the Occam approach. Wrong starting models in the Marquardt approach can lead to wrong geological interpretations, especially in our geological situation for the electrical component in the inline configuration (Lippert 2015). Therefore, we chose the two-step inversion approach described here.

Figure 16 shows two examples of measured electrical components Ex from the broadside configuration and their resulting models from Occam and Marquardt inversions. On the right is an example for an inversion result with a high resistive layer in a depth of approximately 100 m below sea level. Occam and Marquardt inversions produce a similar resistivity value of approximately 80 Ωm for this layer. This layer is associated with the marine groundwater layer and is in good agreement with the results of Levi et al. (2018) and Haroon et al. (2018). The left example in Figure 16 does not show a high resistive layer in the depth: the aquifer is not present at this station.

The results from inline electrical data sets (Ex) with STD on each data point larger than 1% did not show a poor conductor in the subsurface. These results lead to the conclusion that no aquifer is present at depth. As we showed by the modelling of synthetic data in the previous section, resolving the aquifer layer is nearly not possible for inline electric field data having an STD larger than 1%. Therefore, we discarded these results, and do not show them here. For the data sets with all STD smaller than 1%, we discover a resistivity of approximately 30 Ωm for the aquifer by the Occam inversion using the roughness 1 according to Constable et al. (1987), see an example in Figure 14 (right). Occam inversion with roughness 1 leads to a constant resistivity value in depths where no physical information is given by the data anymore. This resistivity value does not match the values given by the broadside electrical data sets, discussed earlier. In the previous section, we solved this discrepancy by using synthetic forward and inverse data modelling. Additionally, the depth of the poor conductor is larger than the depth derived by the broadside electrical data. This data set is close to the coast having a sea floor with a big slope. The Marquardt results are comparable with those from the broadside Ex data and show a resistivity value of approximately 100 Ωm. Nevertheless, we also detected the aquifer with this configuration. The other data example we show in Figure 14 (left) do does not show a poor conductor at all. Because of a good data quality (STD < 1%), the absence of the aquifer at this position can be regarded as valid.

The 1D inversion of the broadside vertical magnetic (dBz/dt) component data also shows a poor conductor at depth (Fig. 15). Here, we had to discard the early times because of a bad S/N ratio. This lack of early time data information leads to a very thick poor conducting layer in the inversion result. The resistivity values of the Occam inversions for the aquifer layer reach a maximum 30 Ωm. This value is in good agreement with the Occam results from the inline electrical field, but is too small compared with the broadside electrical
field. We have explained this discrepancy in the Modelling section by synthetic forward and inverse modelling. The Marquardt inversion results are comparable with the Marquardt results of the other components: The aquifer layer has a resistivity value of approximately 100 Ωm.

The high resistive groundwater layer at a depth of approximately 100 m under the sea level was detected by all examined components with a two-step 1D inversion approach using Occam and Marquardt inversions one after the other, as described above. In addition to the data misfit, we proved the Marquardt results by observing the residuals from forward calculations of models with or without a high resistive layer (Lippert 2015).

Finally, Figure 17 presents the Occam results of the broadside electric component (Ex) as a merged 2D profile section. The aquifer is clearly visible as a poor conductor of approximately 80 Ωm. It reaches from the coast to a distance of 3250 m. The next receiver position at 3650 m clearly shows the absence of the aquifer. Haroon et al. (2018) also carried out a survey of the same area using the differential electrical dipole as transmitter. Their results are in a very good agreement with the LOTEM survey concerning the depth of the marine aquifer and its extension from the coast. Two data examples of receiver positions – with and without aquifer – are given in Figure 16. Figure 17 also shows some kind of depth validation of the Occam results as proposed by Constable et al. (1987) (black horizontal lines beneath each receiver location): The depth at which the two Occam regularizations show different models, the models are derived only by regularization and not by data anymore. This depth can also be considered as the depth of investigation which is deeper than the depth of the aquifer at all stations (Fig. 17).

CONCLUSION

Our synthetic qualitative and quantitative modelling studies prefer the electric fields in the broadside configuration for resolving the parameters of a shallow poor conducting layer which is embedded in good conducting sea sediments for very shallow water depths up to 50 m. The electric field in the inline configuration has to be of excellent data quality to detect this target layer.

We also examined the phenomenon of the airwave for our situation with very shallow water: The signature of the airwave arrives for some water depths at the same point in time as the target signature. Our synthetic modelling shows no disturbance for the inversion results, as we took the airwave effect into account by including the water layer in the inversion process.

For the real measured data, the different results from the inline and broadside electric components were well explained by synthetic modelling. We detected and proved the existence of the submarine aquifer offshore Israel in the northern part of the Palmahin disturbance zone with different long-offset transient electromagnetic measuring system components. With 1D Occam inversions, we were able to pinpoint the aquifer edge at a distance of 3250 to 3650 m from the coastline. This result, the lateral dimension of the aquifer, can be very useful information for the future water management of the coastal aquifer in Israel.

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