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Quantifying freshwater resource in coastal barriers: the joint use of transient electromagnetic and magnetic resonance soundings

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Abstract

Many human communities living in coastal areas in Africa and Asia rely on thin fresh water lenses for their domestic supply. Population growth together with change in rainfall patterns and sea level will probably impact these vulnerable groundwater resources. A spatial knowledge of the aquifer properties and the use of groundwater model are required for the sustainable management of the resource. This paper presents a ready-to-use methodology for estimating the key aquifer properties based on the joint use of two non-invasive geophysical tools together with common hydrological measurements.

We applied the proposed methodology on a coastal sandy barrier in South-Western India. We found that the joint use of magnetic resonance and transient electromagnetic soundings allows to map the fresh water lens and to estimate the specific yield, the hydraulic conductivity, the water salinity and the water table recharge. From the geophysical results, we estimate the fresh water reserve to range between 400 and 700 l m^{-2} of surface area according to the location and to the season. Using time lapse geophysical measurements and common groundwater monitoring, we also estimate the recharge of a rainy event to be about 100 % of the rain, and the net recharge at the end of the monsoon to be less than 10 % of the rain. Thus, we conclude that a change in rainfall patterns will probably not impact the groundwater resource since most of the rain water recharging the aquifer is flowing towards the sea and the river. However, a change in sea level will impact both the groundwater reserve and net recharge.

1 Introduction

In Africa and Asia, climate change along with rapid population growth will probably impact all water resources. The management of groundwater in coastal areas is already critical since the highest concentrations of human settlements occur along the coasts where the rate of water withdrawals has increased three times faster than the rate of population growth since the 1900s (Unesco, 2006).

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Small islands and coastal barriers in Africa and Asia are usually populated with fishermen communities. These communities rely on groundwater for their domestic supply but also for providing water to the small fishing-related industries. In such areas, the infiltration of rain supplies a fresh water lens lying on the top of saline groundwater. But thin fresh water lenses are highly vulnerable to sea water intrusion, and their management must be extremely careful to be sustainable. Groundwater numerical modelling is a powerful tool for simulating the behaviour of such fresh water lenses under different conditions but it requires a spatial distribution of input parameters such as the thickness of the aquifer, its storage-related properties, its hydraulic conductivity and the salinity of the water.

The preferred approach of hydrogeologists for estimating the properties of aquifer relies on drilling boreholes and carrying out hydraulic tests. However, these techniques are not always appropriate because of the risk of salty water intrusion. Moreover, drilling boreholes is costly and time-consuming and the spatial density of sampling is rarely sufficient for an effective characterisation. Thus, there is a need to develop new tools and methodologies to assess the fresh water resource in small islands and coastal barriers.

Non-invasive surface geophysical methods capable of providing rapid, dense and low cost data coverage can be very useful if they provide accurate estimates of aquifer properties. Most of the previous works focused on the link between aquifer properties and electrical-based parameters (among others Buchanan and Triantafilis, 2009; Chandra et al., 2008; Slater, 2007; Soupios et al., 2007; Steuer et al., 2009). However electrical resistivity or conductivity, as the majority of geophysical parameters, result from several factors including but not limited to groundwater, and the relationships between geophysical parameter and hydrogeological properties are usually site specific and valid only for their calibration range (Vereecken et al., 2006). But a non-invasive geophysical method can provide a more direct link to the presence of water. Indeed, magnetic resonance sounding (MRS) is selective with respect to groundwater

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(Legchenko and Valla, 2002) and MRS result is the distribution of groundwater content and pore-size related parameters with depth.

In coastal areas, geophysical tools already showed their interest to map fresh water and to estimate some aquifer properties when applied with complimentary hydrological tools (among others Albouy et al., 2001; Comte et al., 2010; de Louw et al., 2011; Ezerski et al., 2011; Goldman and Neubauer, 1994; Goldman et al., 1991, 1994b; Kafri and Goldman, 2005; Levi et al., 2008; Nielsen et al., 2007; Zarroca et al., 2011). However, previous studies do not estimate all of the relevant aquifer properties since hydraulic conductivity and specific yield are not included. Vouillamoz et al. (2007b) has proposed a more comprehensive approach to estimate the properties in 1-D of coastal aquifers in Myanmar but it was applied in confined aquifer conditions.

This paper presents a ready-to-use methodology for mapping the relevant parameters for characterizing freshwater lenses on small islands and coastal barriers. The methodology is based on the joint use of two geophysical tools together with common hydrogeological measurements. The magnetic resonance sounding (MRS) and the transient electromagnetic sounding (TEM) are used for mapping the fresh water reserve, for estimating the hydraulic properties of the aquifer and the salinity of the groundwater. Recharge is estimated by coupling geophysical and hydrological measurements. This proposed methodology is applied for estimating groundwater resource on a strip barrier in southwest of India.

2 Methodology

2.1 Estimating specific yield and hydraulic conductivity with MRS

Magnetic resonance sounding (MRS) is a non-invasive geophysical method designed for groundwater investigation (see Legchenko and Valla, 2002, for a detailed description of the method). To carry out a measurement, the nuclei of the hydrogen atoms of water molecules in the subsurface (i.e. protons) are energized with an electromagnetic pulse, and the signal response of the hydrogen nuclei is measured after the energizing pulse is switched off. The maximum investigation depth is about 100 m below ground surface in common conditions, but in salty water context (i.e. high electrical conductive medium) the maximum depth of investigation is reduced to about 50 m (Legchenko et al., 2008).

The main advantage of MRS as compared to other geophysical methods is that the recorded signal is generated by groundwater molecules. Two characteristics of the MRS signal are related to hydrogeological properties: (1) the initial amplitude of the signal is proportional to the number of hydrogen nuclei of the sampled aquifer, and (2) the decay rate of the signal is linked to the mean size of the pores that contain groundwater. After inversion of the records, the MRS result is the distribution of groundwater content θ_{MRS} and pore-size related parameters with depth.

Previous works already assessed the links between the field scale MRS parameters and hydrogeological properties of aquifers (e.g. Legchenko et al., 2002, 2004; Lubczynski and Roy, 2005, 2007; Plata and Rubio, 2011; Vouillamoz et al., 2002, 2005, 2007a, 2008). Specific yield S_{y} is an essential storage-related property of unconfined aquifer because it quantifies the amount of water an aquifer releases by gravity forces when drained. S_{y} is then used for calculating groundwater reserve (de Marsily, 1986) but also for estimating groundwater recharge (Healy, 2010). In fine grained-rocks, Boucher et al. (2009) and Vouillamoz et al. (2005) showed that on average $\theta_{\mathrm{MRS}} \approx 3 \cdot S_y$ because part of the groundwater measured by MRS cannot be drained by gravity (i.e. bound and capillary water). In coarse grained-rocks where bound and capillary water are negligible as compared to gravitational water, the MRS water content is probably close to the specific yield:

$$\theta_{\mathrm{MRS}} \approx S_{\mathrm{y}}. \tag{1}$$

To our knowledge, no field scale experiments have been carried out for confirming the relationship between θ_{MRS} and S_{y} in coarse-grained rocks. In this paper, we compare S_{y} measured on sand samples with θ_{MRS}.

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The decay rate of the MRS signal is linked to the mean size of the pores that contain groundwater (Schirov et al., 1991). Based on hydrogeological formulations linking aquifer grain size and hydraulic conductivity (Hazen and Kozeny-Carman), the relation between the MRS decay rate and the pore size has been successfully used for estimating transmissivity and hydraulic conductivity of saturated rocks (for example Plata and Rubio, 2008; Ryom Nielsen et al., 2011; Vouillamoz et al., 2002, 2005, 2008). The common formulation used for calculating aquifer transmissivity is:

$$T_{\mathrm{MRS}} = K_{\mathrm{MRS}} \cdot \Delta z = C_T \cdot \theta_{\mathrm{MRS}} \cdot T_1^2 \cdot \Delta z \quad (2)$$

where K_{MRS} is the hydraulic conductivity calculated from MRS, Δz is the thickness of saturated layer as defined by MRS, T_1 is the decay rate of MRS signal and C_T is a parametric factor. For sands $C_T \approx 10^{-8}$ (Vouillamoz et al., 2008).

2.2 Estimating groundwater salinity from the joint use of electrical resistivity and MRS

Geophysical methods that give access to the electrical resistivity of rocks are widely used for aquifers characterization because of the link that exists between electrical resistivity of rocks, rock water content and water salinity. This link is expressed by the so-named first Archie equation established with clean samples, i.e. free of clay minerals (Archie, 1942):

$$\frac{\rho_{\mathrm{w}}}{\rho_{\mathrm{aq}}} = n^m \quad (3)$$

where ρ_{w} and ρ_{aq} are the electrical resistivity of the water and of the saturated aquifer, respectively, n is the porosity and m is a parametric factor. For geophysicists, Eq. (3) is difficult to solve because it contains two unknown parameters: the resistivity of the water and the porosity. Moreover, as mentioned by Archie (1942), value of m depends on aquifer rocks and ranges between 2 and 1.8 for consolidated sandstone and is about 1.3 for unconsolidated sand. Archie equation has been empirically confirmed by numerous field experiments but it is often reported as (Worthington, 1993):

$$\frac{\rho_w}{\rho_{aq}} = \frac{n_a^m}{a} \tag{4}$$

where a is a reservoir constant. Values of a and m are reported to vary widely with rock type, what complicates the use of Eq. (4) without additional information on water electrical conductivity (EC) or aquifer porosity. To solve Archie equation in coastal aquifer, Kafri and Goldman (2005) proposed to first apply the equation to the sea water intruded portion of the aquifer. Because sea water EC is known or can be easily measured, the only remaining unknown parameter of Archie equation is the porosity that is then calculated for the sea water layer. Assuming the porosity to be the same for the entire saturated thickness, this porosity value is then used to solve Archie equation above the sea water saturated layer where the only remaining unknown parameter becomes the water EC. This methodology is relevant to estimate water EC and porosity in homogeneous aquifer, but cannot be used if the porosity of the fresh water layer is different from the porosity of the salty water layer. Moreover, the determination of parameters a and m is not possible, and values reported in the literature need to be used. Because the range of reported values is large, the choice of relevant values is not straightforward (Worthington, 1993).

This paper presents a methodology to overcome the difficulties in solving Archie equation. We measured not only electrical resistivity but a complementary geophysical parameter that gives access to another unknown parameter of the equation. MRS is relevant because it gives access to the MRS water content that is linked to the storage related parameters of saturated aquifer (Lubczinski and Roy, 2007; Vouillamoz et al., 2008). Note that Hertrich and Yaramanci (2002) proposed to jointly interpret resistivity measurements and MRS based on Archie equation. But their approach aimed at improving the interpretation of θ_{MRS} rather than solving Archie equation and assumptions on values of m and water EC still needed to be done. In our approach, MRS water content θ_{MRS} is used to estimate aquifer porosity and to solve Archie equation together with measured aquifer resistivity ρ_{aq}. We first applied our approach to a sea water saturated layer for determining the value of m factor in Eq. (3). Then, the only remaining unknown parameter is the water resistivity ρ_w when applying our approach to study the aquifer.

2.3 Sequential inversion of transient electromagnetic and magnetic resonance soundings

Several geophysical methods can be used to calculate the electrical resistivity of rocks. Transient electromagnetic sounding (TEM), also known as time domain electromagnetic (TDEM), has been extensively used in coastal areas to map sea water intrusion because of its high sensitivity to electrically conductive target as saline water saturated layers (for example Ezerski et al., 2011; Goldman et al., 1991; Nielsen et al., 2007). However, TEM is not effective for very shallow depths (i.e. a few meters) because of the limitation in the early time detection. Detailed description of the method is given in numerous publications (for example Nabighian, 1991).

In this paper, we propose to use the complementary effectiveness of MRS and TEM methods by applying a sequential inversion process. To assess the capability of TEM and MRS methods to characterize shallow aquifer on strip barriers or small islands, a likehood hydrogeological model of a thin and shallow freshwater lens is proposed. A water content and a geoelectrical models are then calculated according to the hydrogeological assumptions presented Fig. 1. Synthetic geophysical data are generated according to this input model with thickness of the first 3 layers ranging from 1.5 to 10 m. Then, layered model inversion is conducted to determine the accuracy for recovering the input model. The root mean square (RMS) fitting error is used as an indication of the fit between the input model and the output solution. Note that several output solutions can equally fit the input data (i.e. equivalent solutions with comparable RMS). For assessing the range of acceptable solutions we apply the approach proposed by

Goldman et al. (1994a) and we only consider solutions with RMS lower than ∼+20 % as compared to the best fit RMS (i.e. solution which has the lower RMS).

On the one hand, numerical modelling conducted with TEM alone shows that resistivity and depth of the salty water layer is well determined, but the dry sand and the fresh and brackish water layers cannot be separately resolved if the layers are less than 5 m thick. On the other hand, numerical modelling also shows that MRS inversion poorly resolves the water content θ_{MRS} and the saturated thickness Δz independently, but the product $\theta_{\mathrm{MRS}} \cdot \Delta_z$ is well resolved. Then, the resolution of θ_{MRS} can be significantly improved if the geometry Δz of the saturated layers is known.

Consequently, we propose to use a sequential inversion process which improves the resolution of both the water content and depth/thickness of the layered model. First, the depth and thickness of the salty water layer which are well defined by the inversion of TEM, are used to constrain the MRS inversion. It leads to an improved resolution of both MRS water content and saturated geometry. Second, the saturated geometry as resolved by MRS is used to constrain a new inversion carried out with TEM with an imposed depth to the fresh water layer. The result is a unique model that fits equally well the two data sets. Our numerical modelling shows that this sequential process allows (1) to estimate the water content and the static water level with a mean error of ±8 % and ±7 %, respectively, (2) to very well define the saline water layer (i.e. the exact resistivity is derived with about no equivalence, and the depth and thickness are derived with an accuracy of ±1 %), and (3) to define the fresh and the brackish water layers with an accuracy which is controlled by the depth and thickness of the layers. The thickness and the resistivity of the fresh water layer are resolved with an error of ±3 % and ±20 %, respectively if the layer is of 10 m thick. If the thickness of the fresh water lens is less than 10 m, the fresh and brackish layers cannot be separated and a single layer is obtained.

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2.4 Estimating the recharge of the aquifer

The water table fluctuation method (WTFM) is wildly used for calculating unconfined aquifer recharge (Healy, 2010). WTFM is based on the assumption that rises in groundwater level are due to inlets of water recharging the water table. If vertical flow is dominant (i.e. the rate at which the groundwater flows away from the measuring location is significantly slower than the rate at which the recharge water arrives at the water table), recharge can be calculated as:

$$R = S_y \cdot \frac{\Delta H}{\Delta t} \tag{5}$$

where ΔH is the change in water table over a time interval Δt.

The main difficulty in applying WTFM is the knowledge of specific yield S_y (Scalon et al., 2002). Since θ_{MRS} is related to S_y, Vouillamoz et al. (2008) proposed to use θ_{MRS} for estimating recharge in sandstones in Niger. In this study, we have collected 5 sand samples below the water table for laboratory analysis. From the comparison between the water content of samples and θ_{MRS} we have derived a relationship between θ_{MRS} and S_y, and we have calculated aquifer recharge from coupled water table monitoring, MRS and TEM measurements.

3 Experiments

3.1 Investigated site

The proposed methodology has been applied on a coastal barrier in South-Western India, Karnataka state (Fig. 2). The so-named Sasihithlu barrier is about 4 km long and 150 to 600 m wide. It is bounded on the west by the Arabian Sea and on the east by the Pavanje river. Pavanje river is seasonal, i.e. it flows only during few months per year and sea water ingresses into the river up to about 12 km inland (Chandrakantha, 1987). The barrier consists of sands of medium to coarse-grained size (Jayappa and Subramanaya, 1994). Sands lay on a granitic-gneisses basement of Archean age which outcrops in the southern part of the barrier. The thickness of sand deposit is unknown. The highest zone of the barrier is a strip dune located on the backshore with a maximum elevation of 4 m a.m.s.l.

The annual rainfall is 3900 mm (average for the 1970 to 1985 period) which mainly occurs during the south-western monsoon from June to September (Chandrakantha, 1987).

Most of the families living on Sasihithlu barrier have their own hand dug well which provides water all through the year. The average depth of 65 monitored wells is 3.3 m below ground surface, and the average static water lever is 0.2 m a.m.s.l. The fresh water lens morphology is not known.

3.2 Methods and material

Two geophysical surveys have been carried out: a first one before the monsoon (February and March) and a second one at the end of the monsoon (November). Additional MRS and TEM measurements were carried out during the monsoon on two profiles (P1 and P2, Fig. 2) every 30 to 60 days.

For each survey, 16 MRS have been implemented along 3 profiles (P1, P2 and P3) with the Numis^{plus} apparatus from Iris Instruments. A coincident square-shaped loop measuring 25 m long per side was used (except at one location where a 50 m long side loop was used). The high signal to noise ratio (average of 9) indicates the good quality of the measurements. MRS records were interpreted with Samovar software V11.3 and the appropriate routine for high electrical conductivity medium (Legchenko et al., 2008).

A total of 140 TEM soundings have been carried out (60 soundings in March and 80 soundings in November) using the TEM-FAST 48HPC instrument from AEMR. A coincident loop of square shape and 25 m long per side has been used. The electromagnetic noise level was low and the signal to noise ratio remained high for all measurements. TEM data were interpreted with IX1-D software (Interpex).

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Groundwater level and electrical conductivity (EC measured just below the water table and at the bottom of the wells) have been monitored in 65 wells on a weekly basis. Additionally, an automatic recorder was installed in a well for monitoring the water level and EC every hour. Two automatic rain gauges have also been installed for monitoring the rainfall.

Five sand samples were collected at different locations below the water table (1 to 2.3 m) for laboratory analysis. The samples were dried at 50 °C for 48 h, and then fully saturated. The volumetric water content was calculated by weight for comparison with the MRS water content.

4 Results and discussion

4.1 Parameterization of Archie equation

TEM and MRS soundings are implemented on the beach (for example TEM74 and MRS7, Fig. 3). The result of the sequential inversion reveals a layered model with a homogeneous water content of 27 % from 1 to 17 m below ground surface (Fig. 4a) and a 5 resistivity layers (Fig. 4b). Note that the resistivity of the first layer is not known because (1) the TEM measurement is blind from ground surface to about 1.5 m deep with the used configuration and equipment and (2) resistive targets are not well defined with TEM measurements (Albouy et al., 2001). Crossing both MRS and TEM results, one can propose an obvious hydrogeological interpretation (Fig. 4c): from ground surface to depth, we found a dry sand layer, a sandy layer saturated with probably brackish water, then a sandy layer saturated with sea water and at depth the gneissic substratum which is probably weathered in its upper part. To parameterize the Archie equation, we measure the electrical conductivity of the sea water ($EC_{sea} = 56\,\mathrm{mS\,cm^{-1}} \Leftrightarrow \rho_w = 0.18\,\Omega\mathrm{m}$) and we calculate the m factor of Archie first equation for the sea water saturated layer ($\theta_{MRS} = 0.27$ and $\rho_{aq} = 0.83\,\Omega\mathrm{m}$) as:

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$$m = -\frac{\log\left(\frac{\rho_{aq}}{\rho_w}\right)}{\log(\theta_{MRS})}. \tag{6}$$

The m values calculated from 16 MRS/TDEM soundings all over the investigated area range between 1.16 and 1.36, with an average of 1.27. Note that the value proposed by Archie (1942) for clean sand is 1.3, thus suggesting that θ_{MRS} is close to the porosity of the sand. Knowing the m value, we can now use the parameterized Archie equation for calculating the water EC whatever the depth and location. Because the TEM solution of the sea water layer is almost unique (about no equivalence, see Sect. 2.3) the use of the parameterized Archie equation should be robust.

4.2 Quantification of the fresh water reserve

An example of the application of the proposed methodology is presented Fig. 5. The TEM and MRS field data are first inverted using the sequential process presented Sect. 2.3 (Fig. 5a). The output geophysical model fits well both TEM and MRS data and proposes a water content of 30 % between 1 and 19.5 m below ground surface with a 4 layers resistivity model (Fig. 5b). Then, we proceed to the hydrogeological interpretation of the geophysical models (Fig. 5c):

1. The distribution of groundwater EC with depth is calculated from the TEM resistivity model using the parameterized Archie equation. Because the TEM does not differentiate the fresh water from the brackish water layers, we use a simplified model which sets that the water EC is linearly increasing with depth. This assumption is based on measurements of the groundwater EC carried out in numerous monitoring wells. For example the value of water EC measured in the piezometer adjacent to the MRS loop (Fig. 3) is $EC_{piezometer} = 630\,\mu S\,cm^{-1}$ which is close to the value calculated for the same depth from the TEM measurements $EC_{TEM} = 600\,\mu S\,cm^{-1}$ (Fig. 5c). Then, knowing the maximum value

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of water EC which is acceptable by the local communities for domestic supply ($EC_{\text{domestic water}} < 1500\,\mu S\,cm^{-1}$), one can estimate the thickness of the fresh water lens. Note that this thickness is calculated for three TEM solutions: the best fit solution which has the lowest RMS value, and two solutions issued from the equivalence analysis which indicate the maximum and the minimum resistivity and thickness of the layer.

2. To calculate the groundwater reserve from MRS results, one needs to know the relationship between the specific yield S_y and the MRS water content θ_{MRS}. The porosity of the sand samples collected below the water table ranges between 28 and 34 %. This porosity is less than the total porosity because part of the bound water attached to sand surface by molecular attraction cannot be removed by the used laboratory method. Total porosity less bound water is named effective porosity *ne* by hydrogeologists (if we neglect the unconnected and dead-end pores, Lubzinski and Roy, 2007). The MRS water content measured on the same location is ranging between 27 and 36 %. Even though a $213 \times 10^{-6}\,m^3$ sample cannot be rigorously compared to MRS measurements ($\approx 280\,m^3$ for the survey), θ_{MRS} is close to *ne* for the sampled sands. Moreover, the long MRS decay rates indicate coarse average grained-size of the sands (Fig. 5b) for which the amount of capillary water is probably negligible as compared with the amount of gravitational water. Thus $ne \approx S_y \approx \theta_{MRS}$. The distribution of θ_{MRS} with depth is then used to calculate the groundwater reserve as the product $\theta_{MRS} \cdot \Delta z$ where Δz is the saturated thickness. Note that the sequential inversion which sets the depth to the bedrock in the MRS inversion limits the uncertainty in the MRS results (low equivalence, Fig. 5a, b). The total water reserve at this location is $5.5\,m^3\,m^{-2}$ of surface area. However, only 600 l of water (±50 l according to the equivalent solutions) per square meter of surface area have an EC lower than the threshold value and are considered as fresh water.

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3. The MRS results are used for calculating the hydraulic conductivity and the transmissivity using Eq. (1). The hydraulic conductivity is $K_{\mathrm{MRS}} = 4 \times 10^{-4}\,\mathrm{m\,s^{-1}}$ and the transmissivity of the fresh water lens is $T_{\mathrm{MRS}} = 6.2 \times 10^{-4}\,\mathrm{m^2\,s^{-1}}$. Note that the hydraulic conductivity estimated in the monitored well is $3.6 \times 10^{-4}\,\mathrm{m\,s^{-1}} < K < 9 \times 10^{-4}\,\mathrm{m\,s^{-1}}$ thus suggesting that value of C_T used in Eq. (1) is appropriate ($C_T = 10^{-8}$).

Using rapid TEM/MRS measurements, one can apply the same 1-D methodology all over the targeted area. Then, 2-D sections can be calculated using simple interpolation in-between 1-D measurements (Fig. 6a) and 3-D interpolations techniques can be used for mapping the fresch water lens (Fig. 6b). Finally, from all the TEM and the MRS carried out before and after the monsoon, we are able to estimate the fresh water thickness and corresponding fresh water volume (Table 1).

4.3 Recharge and groundwater resource estimate

For estimating the recharge, we first compare TEM and MRS measurements carried out at the same location but at different period of time. An example of the so-named time lapse TEM (El-Kaliouby et al., 2005) carried out at TEM73 location (Fig. 3) is presented in Fig. 7. Since the depth to the sea water layer is defined by TEM interpretation with a high accuracy (see Sect. 2.3) the comparison between TEM results is possible. The first TEM sounding is carried out in the dry season (February) and indicates a salty water interface at 6 m below ground surface. A second TEM is carried out in July after a total of 1090 mm of rain and reveals a deepening of the salty water interface of 0.7 m. Using the WTFM (Eq. 5) and $\theta_{\mathrm{MRS}} = 30\,\%$, the amount of water which reaches the water table and stays at the surveyed location is 66 % of the rain. A third TEM is carried out in August after a new 271 mm of rainfall. Surprisingly, the TEM indicates that the sea water interface rises of 0.1 m. This observation is confirmed by a last TEM carried out in October: after 603 mm of new rainfall the salty water interface continues to rise of 0.15 m. Finally, only 7 % of the total rain recorded between February and October

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remain in the aquifer. This observation is obviously explained by rainwater reaching the water table and then flowing towards the sea and the river (boundary conditions), and probably being taken by some evapotranspiration. Considering the 60 TEM measurements carried out in March and again in October all over the targeted area, the average net aquifer recharge for this period is 9 % of the rain. Note that the MRS time lapse carried out at the same date than the TEM does not indicate any change of the depth to the water table within the uncertainty of ±8 cm (see Sect. 2.3).

To confirm the TEM/MRS time lapse result, we applied the WFTM using the record of a water table monitoring (Fig. 8). We observe that the groundwater level increases rapidly in response to rainy events. Instantaneous recharge (i.e. the amount of rain which enters the water table at an event scale) calculated over short rainy periods (a couple of days) ranges between 85 and 100 % of the rain. This result is not surprising since there is no surface runoff on the sandy barrier and since evaporation and transpiration are probably low at that time scale. Moreover, Fig. 8 also indicates that groundwater level decreases quickly after the rainy events. Hence, no low frequency variation in groundwater level can be observed between the dry and rainy seasons: the depth to the water table remains approximately constant at the year scale. But groundwater EC measured at a constant depth below ground surface presents some season-scale variations: it increases during the dry season as the result of the upcoming of mineralized groundwater and decreases during the rainy season with the infiltration of low-mineralized rainwater. These observations are consistent with the TEM and MRS time lapse which indicates that the recharge process provokes the thickening of the freshwater lens by deepening the salty water interface.

Finally, joint analysis of TEM/MRS time lapse and water table monitoring indicates that about 100 % of the rain infiltrates and recharges the water table, but only few percents (less than 10 % according to TEM) of this instantaneous recharge remain in the aquifer at the end of the monsoon. Thus, a change in rainfall patterns will probably not impact the fresh water resource of the sand barrier, but a change in sea level will have a strong impact because any rise of mean sea level will modify the boundary

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head conditions: it will reduce the thickness of the fresh water lens together with the net recharge.

5 Limitations

This study outlines some of the difficulties encountered while assessing freshwater resources in coastal areas with geophysical tools, and more specifically for thin lenses.

TEM alone but also joint use of TEM and MRS do not allow the differentiation between fresh and brackish water layers. This limitation is well-known by geophysicists as suppression phenomenon. Suppression is common when a layer of intermediate resistivity lies between two layers with higher and lower resistivity, respectively (Albouy et al., 2001). Logically, our numerical study shows that fresh and brackish layers can be replaced by a single layer without impacting the fitting of the model (Sect. 2.3). Moreover, measurements of groundwater electrical conductivity in wells indicate that there are no sharp boundaries between fresh-brackish-salty waters, but rather a mixing layer with EC increasing with depth from fresh to sea water EC. Thus we propose to use a linear gradient of electrical resistivity with depth and we calculated the depth corresponding to the 1500 $\mu S\,cm^{-1}$ threshold thanks to the parameterized Archie equation. This simplified approach enabled us to quantify an approximate thickness of freshwater, but a more comprehensive model could also be proposed based on observations carried out in fully screened and deep enough observation boreholes (which were not available for our study).

The characteristics of the TEM-FAST device and the configuration used in this study are well adapted to shallow targets. However, a first 1.5 m thick resistive layer can be suppressed without impacting the RMS of the obtained model because the first time of measurement (5 µs) does not allow the detection of such shallow depth. Thus, we looked for other common non-invasive geophysical methods for improving the resolution of shallow targets. First, we carried out frequency conductivity sounding (CS) using 2 light multi-turn coils. We implemented the CS with the EM-34 device (Geonics)

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using 3 coil separations corresponding to 3 frequencies, and setting the coils coplanar in a horizontal plane and in a vertical plane. Thus, a total of 6 measurements were performed over a single location using all the available configurations. Both numerical modelling and field measurements indicate that the 5 layers model presented Fig. 1 can be solved with a 3 layers solution that exhibits a low RMS of 1 %. Although the range of equivalence of the first layers is narrower as compared to TEM, the CS cannot either differentiate fresh and brackish water layer. Moreover, the sampling of CS is very poor (i.e. 6 data and only a decade in penetration depth) as compared to TEM (i.e. average of 34 data over 3 decades of time). Thus, we conclude that CS does not improve the characterization of thin lenses as compared to TEM.

The most appropriate method to use jointly with TEM is certainly direct current methods (Albouy et al., 2001). Roughly, TEM is highly sensitive to conductive target while DC resolution is better for resistive target, and their joint interpretation produces a more constrained model as compared to the output of each method applied separately. To carry out a DC sounding, two pairs of electrodes providing galvanic contact with ground have to be used. However, ensuring a good contact in dry sandy cover as frequently met on small barriers and islands is very difficult. To achieve acceptable contact, the common method is to increase the surface area of contact between electrodes and sand (e.g. using several electrodes connected together) and to pour salty water on the electrodes. However, pouring salty water creates conductive anomalies around the electrodes which finally leads to poor data quality when focusing on shallow targets (Fig. 9). Consequently, DC measurements are not appropriate in coastal area covered by dry sand when shallow layers (i.e. few meters deep) are targeted.

Concerning the MRS method, the output parameters still need to be parameterized with known hydraulic conductivity and storage-related parameters. Today, the methodology developed in this paper for quantifying groundwater resource can not rely on geophysical tools alone but needs a coupled hydrological and geophysical approach. However, the potential of MRS for characterizing aquifer properties is still under development, and robust and perhaps universal relationships between MRS parameters and hydrogeological properties will be proposed (Plata and Rubio, 2008).

Note also that we parameterized Archie equation thanks to a joint TDEM-MRS approach. However, the proposed methodology is only possible because the underground medium does not contain any clay. In presence of clay, the use of the Archie equation might be impossible.

6 Conclusion

We propose a methodology which enables to estimate the key parameters required for quantifying the groundwater reserve and for managing the resource. This methodology is based on the joint use of MRS and TEM, including time lapse measurements, together with common hydrological monitoring. It is ready-to-use at an affordable cost.

We applied the methodology in a coastal barrier of South-Western India, and we found that the geometry of the fresh water lens, the specific yield and the hydraulic conductivity of the aquifer, but also the recharge and the behavior of the lens can be estimated. We estimate that the fresh water reserve ranges between 400 and 700 liters per square meter of surface area, and that about 100 % of the rain infiltrates and reaches the aquifer. However, more than 90 % of this infiltrated rain water does not increase the fresh water reserve since it flows outwards, towards the sea and to the river, and probably also evaporates. We conclude that a change in rainfall patterns will probably not impact the fresh water resource but a rise of the mean sea level will both reduce the fresh water reserve and the net recharge.

This study also pointed out some limitations of the use of geophysical tools for assessing shallow and thin fresh water lens. Mainly, the comprehensive delineation of the freshwater lens can only be obtained for lenses thicker than 10 m. However, this paper shows that the joint use of MRS and TEM in the framework of a hydrogeological approach is already an appropriate methodology for quantifying freshwater

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resource. Moreover, new developments in the MRS method and in joint use of time lapse MRS/TEM are promising for improving aquifer characterization.

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Table 1. Average thickness of the freshwater lens and corresponding freshwater volume.

	Dry season (Mar)	Rainy season (Oct)
North of the area (uninhabited)	2.2 m $500\,\mathrm{l\,m^{-2}}$	2.9 m $670\,\mathrm{l\,m^{-2}}$
South of the area (inhabited)	1.8 m $420\,\mathrm{l\,m^{-2}}$	2.0 m $460\,\mathrm{l\,m^{-2}}$

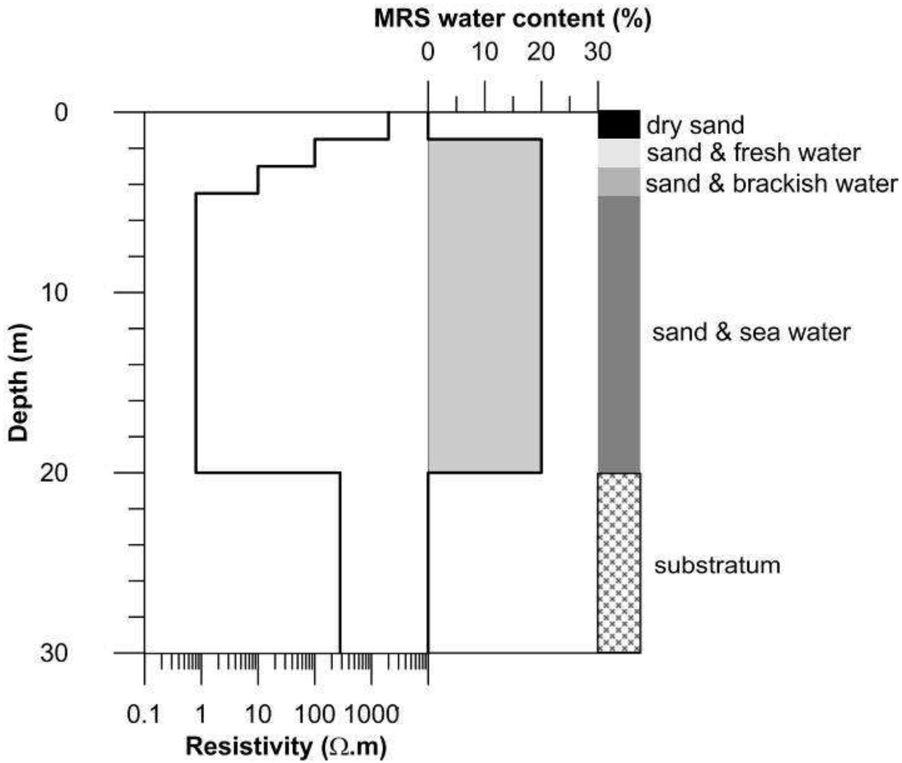

Fig. 1. Geoelectrical and water content models with a thickness of 1.5 m for the first 3 layers. Numerical modelling are conducted with a thickness of the first 3 layers ranging from 1.5 to 10 m and considering a water content of 20 % in medium to coarse-grained sand ($T_2^* = 200$ ms)

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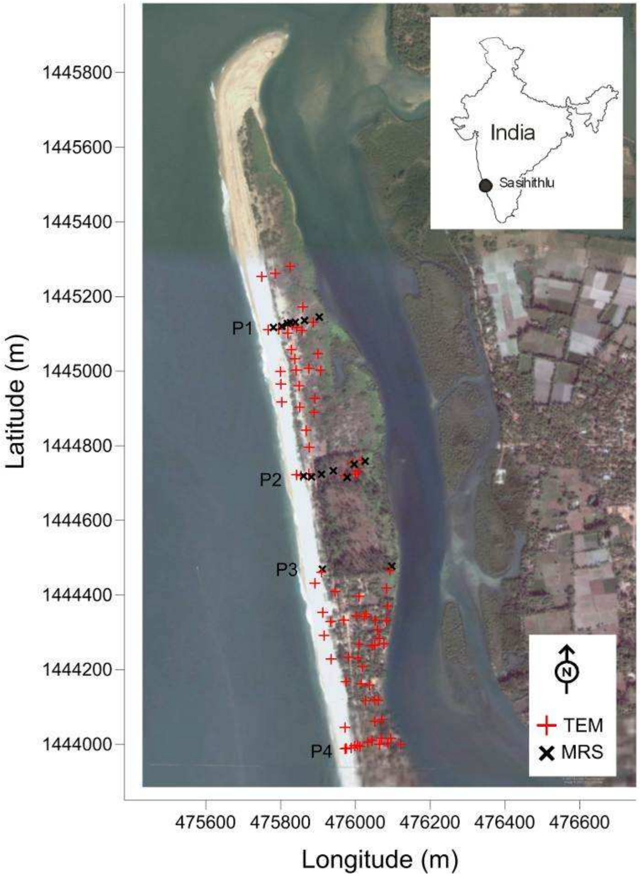

Fig. 2. Location of Sasihithlu barrier and measurements.

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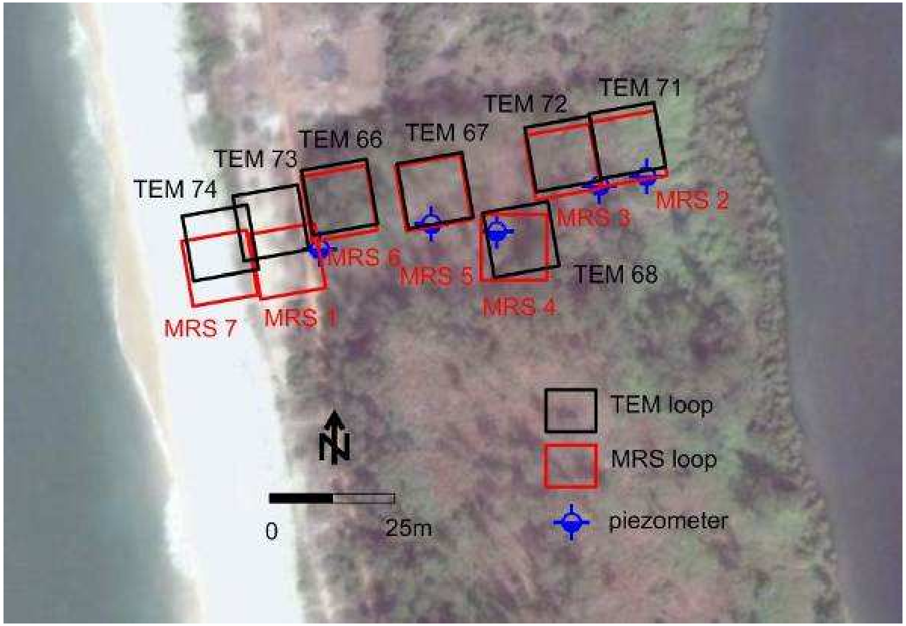

Fig. 3. Location of measurements, Profile 2.

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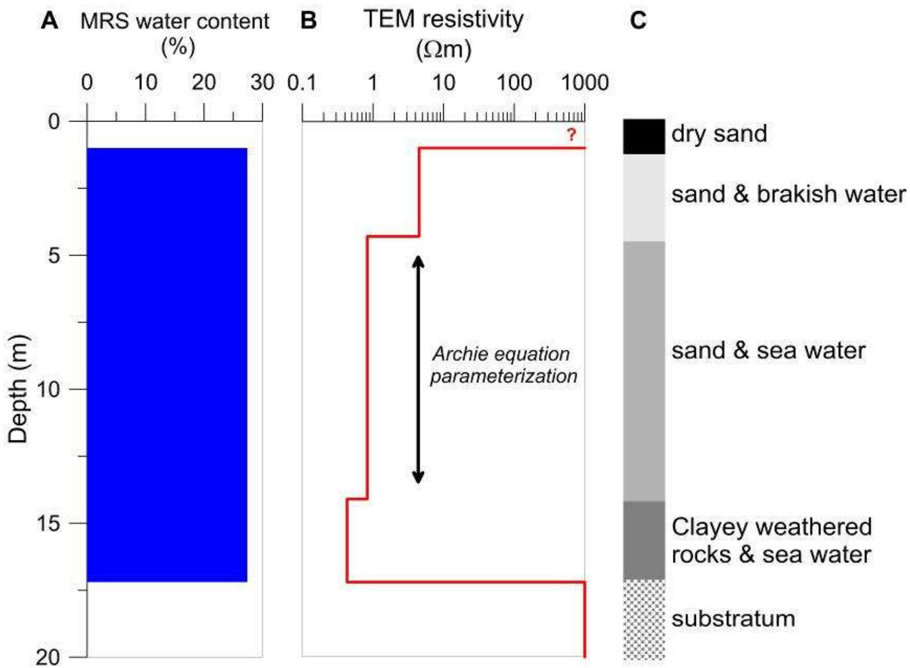

Fig. 4. Example of results of sequential inversion of measurements at the beach side. **(A)**: MRS7, **(B)**: TEM74, **(C)**: hydrogeological interpretation.

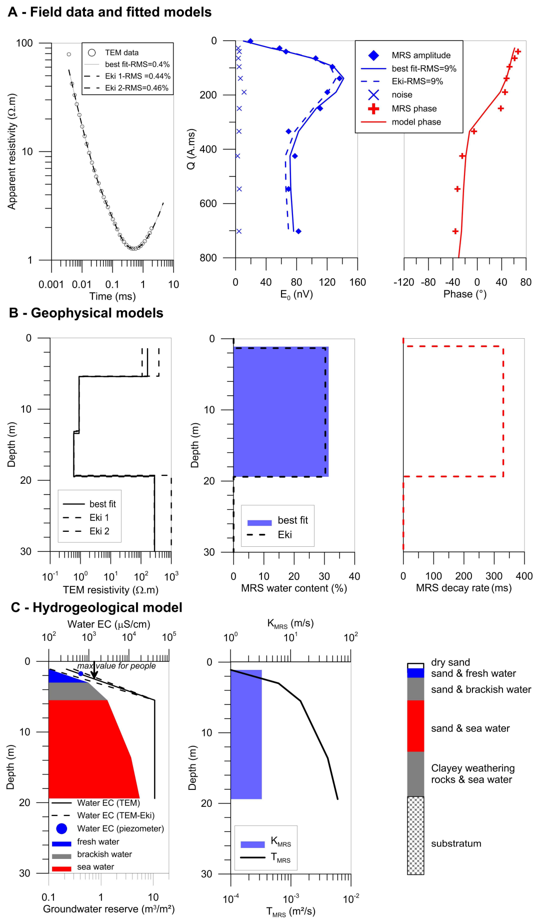

Fig. 5. 1-D application of the hydrogeophysical methodology. **(A)**: MRS6 and TEM66 data and fitted models. **(B)**: geophysical models, **(C)**: hydrogeological interpretation. Eki are the calculated equivalent solutions.

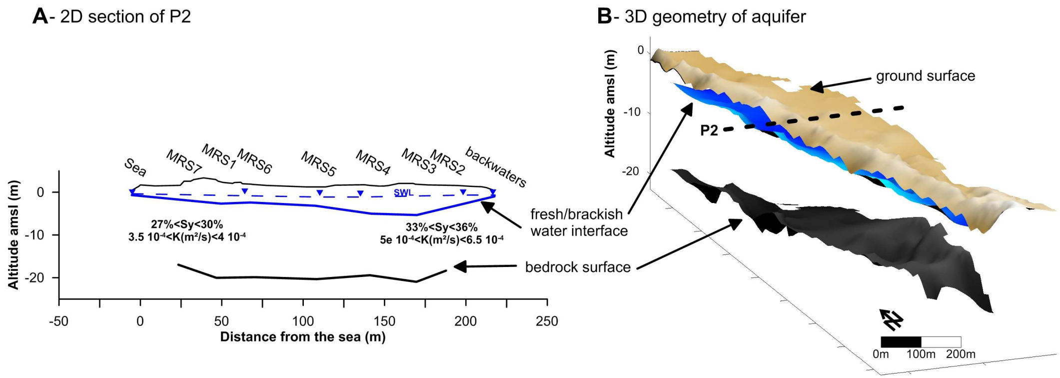

Fig. 6. 2–3-D representation of geophysical results (before the monsoon). **(A)**: 2-D section of Profile P2. SWL is the static water level measured in piezometers. **(B)**: 3-D map of the aquifer.

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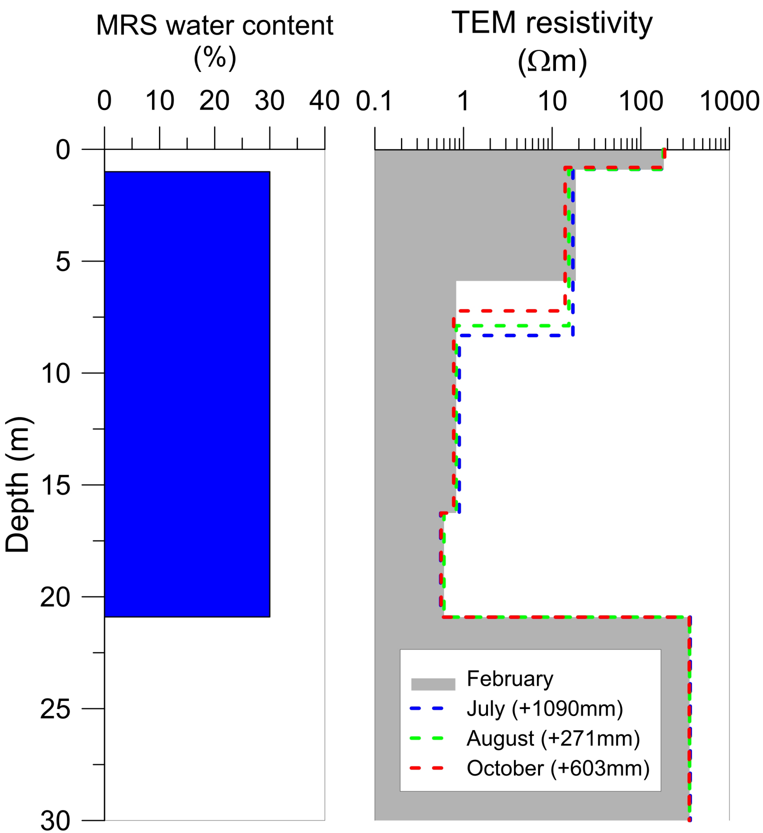

Fig. 7. TEM73 and MRS1 time lapse.

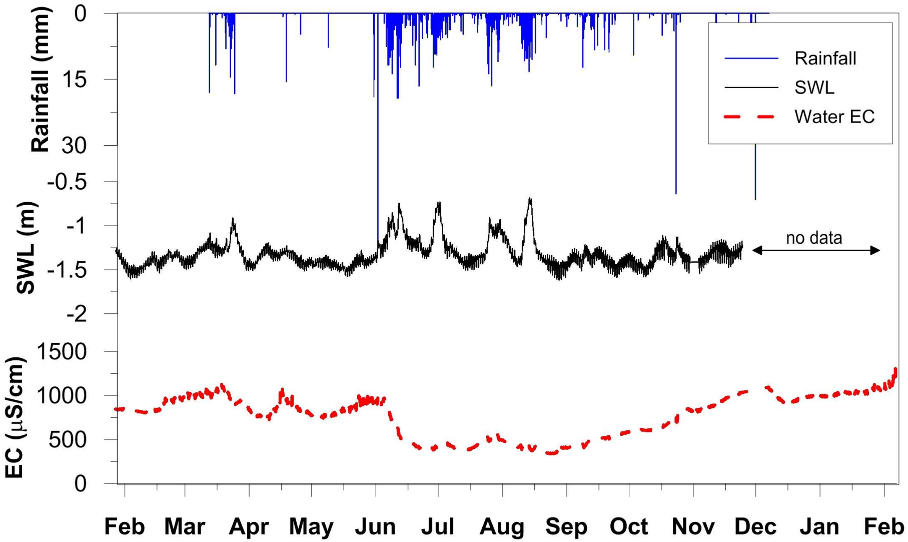

Fig. 8. Rainfall and groundwater monitoring (Well D13).

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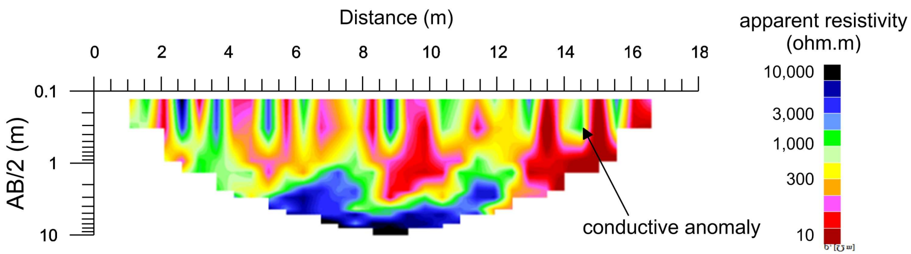

Fig. 9. Example of poor data quality of DC measurement (located on P2 section). Inter-electrode spacing of 0.5 m. Conductive anomalies are created by salty water poured on electrodes to improve contact in dry sand.

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