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Quantifying aquifer properties and freshwater resource in coastal barriers: a hydrogeophysical approach applied at Sasihithlu (Karnataka state, India)

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

We applied the proposed methodology in an unconfined aquifer of a coastal sandy barrier in South-Western India. We jointly used magnetic resonance and transient electromagnetic soundings and we monitored rainfall, groundwater level and groundwater electrical conductivity. The combined interpretation of geophysical and hydrological results allowed estimating the aquifer properties and mapping the freshwater lens. Depending on the location and season, we estimate the freshwater reserve to range between 400 and 700 $\mathrm{L\,m^{-2}}$ of surface area ($\pm$ 50 %). We also estimate the recharge using time lapse geophysical measurements with hydrological monitoring. After a rainy event close to 100 % of the rain is reaching the water table, but the net recharge at the end of the monsoon is 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).

Small islands and coastal barriers in Africa and Asia are usually populated with fishermen communities. Not only do 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 freshwater lens lying on the top of saline groundwater. But thin freshwater 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 freshwater 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 aquifers 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 freshwater 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 succeed to provide accurate estimates of aquifer properties. Most of the previous works focused on the link between aquifer properties and electrical-based parameters (among others Slater, 2007; Soupios et al., 2007; Chandra et al., 2008; Buchanan and Triantafilis, 2009; Steuer et al., 2009). However electrical resistivity or conductivity, as the majority of geophysical parameters, results from several factors and the relationships between geophysical parameters and hydrogeological properties are usually site specific and valid only inside their calibration range (Vereecken et al., 2006). But a non-invasive geophysical method can provide a more close link to the presence of water. Indeed, magnetic resonance sounding (MRS) is selective with respect to groundwater (Legchenko and Valla, 2002) and the resulting MRS parameters are the distribution of groundwater content and decay time (related to pore-size) with depth.

In coastal areas, ground based geophysical tools already showed their capability to map freshwater and to estimate some aquifer properties (among others Goldman et al., 1991; Kafri and Goldman, 2005; Nielsen et al., 2007; Zarroca et al., 2011; Igel et al., 2012). Airborne geophysical surveys have also been carried out for mapping resistivity of large surface areas. Interpreted with ground based geophysical measurements and sometimes hydrological data, this resistivity information has been useful for mapping geological structures (e.g. Burschil et al., 2012; Gunnink et al., 2012; Jørgensen et al., 2012) and for setting up groundwater models (e.g. de Louw et al., 2011; Faneca Sànchez et al., 2012; Rasmussen et al, 2012; Sulzbacher et al., 2012). However, previous geophysical studies do not estimate all of the relevant aquifer properties since hydraulic conductivity or specific yield are not included. Vouillamoz et al. (2007) has proposed a more comprehensive approach to estimate the 1-D properties of coastal aquifers but it was applied in confined aquifer conditions, thus concerning elastic storage and not specific yield. More recently, Günther and Müller-Petke (2012) presented an improved methodology to estimate hydraulic properties of an island unconfined aquifer.

This paper presents a ready-to-use methodology for estimating not only hydraulic properties of coastal aquifer but also for mapping fresh groundwater reserve and estimating recharge in unconfined conditions. The methodology is based on the joint use of two geophysical tools (i.e. the magnetic resonance and the time domain electromagnetic soundings) together with common hydrological measurements (i.e. rainfall and groundwater monitoring). The proposed methodology is applied for estimating the groundwater resource on a strip barrier in South-West 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 (for a detailed description of the method see Legchenko and Valla, 2002). During measurements, 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 average conditions, but in salty water context (i.e. high electrical conductive medium) the maximum depth of investigation is drastically reduced (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 records, the MRS result is the depth related distribution of groundwater content θ_{MRS} and decay rate (usually T_2^*).

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, 2008, 2012). Specific yield S_{y} is an essential storage-related property of an unconfined aquifer because it quantifies the amount of water an aquifer releases by gravity forces when drained. Not only is S_{y} 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, 2012) showed that on average $\theta_{\mathrm{MRS}} \approx 3 \cdot S_{\mathrm{y}}$ 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}.

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 equations linking aquifer grain size and hydraulic conductivity (Kasenow, 2002), 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 Vouillamoz et al., 2002, 2005, 2008; Plata and Rubio, 2008; Ryom Nielsen et al., 2011). The common formulation used for calculating aquifer transmissivity T_{MRS} is

$$T_{\mathrm{MRS}} = K_{\mathrm{MRS}} \cdot \Delta z = C_T \cdot \theta_{\mathrm{MRS}} \cdot T_i^2 \cdot \Delta z \tag{2}$$

where K_{MRS} is the hydraulic conductivity calculated from MRS, Δz is the thickness of saturated layer as defined by MRS, T_i is the decay rate of MRS signal and C_T is a parametric factor. For sandy conditions $C_T \approx 10^{-2}\,\mathrm{m\,s^{-3}}$ (Vouillamoz et al., 2008).

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

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

$$\frac{\rho_{\mathrm{w}}}{\rho_{\mathrm{aq}}} = n^m \tag{3}$$

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, value of m depends on aquifer rocks and ranges between 1.8 and 2 for consolidated sandstone, with a value of approximately 1.3 for unconsolidated sand (Archie, 1942). Archie's equation has been empirically confirmed by numerous field experiments and is often reported as (Worthington, 1993):

$$\frac{\rho_{\mathrm{w}}}{\rho_{\mathrm{aq}}} = \frac{n^m}{a} \tag{4}$$

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 resistivity or aquifer porosity. To solve Archie's 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 electrical conductivity (EC) is known or can be easily measured, the only remaining unknown parameter of Archie's equation is the porosity that hence can be calculated for the sea water layer. Assuming the porosity to be the same for the entire saturated thickness, the calculated porosity value can be used to solve Archie's 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 freshwater 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's equation. We include measurements of not only electrical resistivity but of 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, 2012). Note that Hertrich and Yaramanci (2002) proposed to jointly interpret resistivity measurements and MRS based on Archie's equation. But their approach aimed at improving the interpretation of θ_{MRS} rather than solving Archie's equation and assumptions on values of m and water EC were still needed. More recently, Günther and Müller-Petke (2012) proposed an improved joint inversion of resistivity measurements and MRS. They deduced salt concentration from a modified Archie's equation which is parameterised using external data (i.e. direct push soundings). In our approach, MRS water content θ_{MRS} is used to estimate aquifer porosity and to solve Archie's 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 jointly used with MRS to calculate the electrical resistivity of rocks. Hertrich and Yamaranci (2002) chose the common vertical electrical sounding (VES) and they developed a joint inversion algorithm for VES and MRS. They demonstrated that their algorithm improves the model characterisation as compared to the characterisation obtained from the inversion of a single method, but their approach is limited in its application by the use of a simplified Archie's law. Vouillamoz et al. (2007) proposed a combined use of VES and MRS in the framework of a hydrogeological approach for quantifying hydraulic properties of a coastal aquifer. They demonstrated that inverting VES with a fixed geometry obtained from MRS significantly diminishes resistivity uncertainties and thereby improves the estimate of water EC. Moreover, from joint MRS, VES and pumping test interpretation, they proposed semi-empirical relationships for estimating aquifer elastic storage, transmissivity and water EC. Günther and Müller-Petke (2012) developed an improved joint inversion approach of VES and MRS for estimating hydraulic properties of an island aquifer. They demonstrated that their results presented slightly but generally lower uncertainties than the approach proposed by Vouillamoz et al. (2007).

Based on numerical modelling, Behroozmand et al. (2012) demonstrated that transient electromagnetic sounding (TEM, description of the method is given in numerous publications, e.g. Nabighian, 1991) is probably the best choice among complementary geophysical method to be used with MRS because of its superior resolution of conductive layers (as compared to direct current resistivity). Indeed, conductive layers affect the magnetic field values and thereby the MRS response, and the authors demonstrated the need for sufficiently deep and accurate resistivity information for MRS inversion. They developed a fast computation method for iteratively updating the MRS response in the framework of a joint MRS/TEM inversion. They found that their inversion approach improves the determination of aquifer characteristics in conductive environments, and that the use of MRS diminishes the equivalence of the resistivity model. Moreover, they also showed that laterally constrained inversion of joint MRS/TEM results in a reasonable accurate estimation of smooth structures.

In our study, we selected TEM method for complementing MRS measurements, and we used a sequential inversion scheme. In contrast to the common stepwise inversion where the results of one method is used to define the starting model for the non-constrained inversion of the other method (Behroozmand et al., 2012), the sequential inversion is an iterative interpretation where the result of one method is used to constrain the inversion of the other one and so on. In our approach, the sequential inversion uses the best capability of each method for constraining the inversion of the other one: the geometry of the salty water layer obtained from a non-constrained TEM inversion is used to constrain the MRS inversion. Then, the depth to the saturated layer obtained from the MRS is fixed in a new TEM inversion. Note that the MRS inversion is conducted based on the resistivity obtained from the first TEM result. As showed by Behroozmand et al. (2012), our sequential inversion introduces some errors in the resulting model because the magnetic fields are not updated as it is done in a joint inversion process. In our study, the resulting error in the MRS results is negligible because

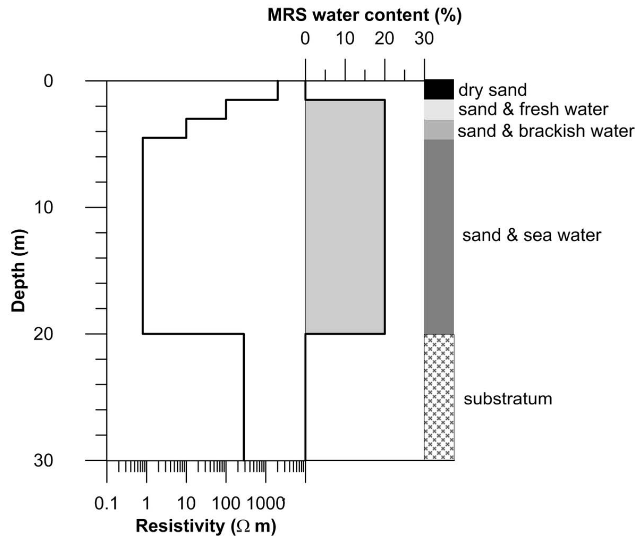

Fig. 1. Geoelectrical and water content models with a thickness of 1.5 m for the first 3 layers. Numerical modelling is conducted with a thickness of the first 3 layers ranging from 1.5 to 10 m and considering a water content of 20 % in medium- to coarse-grained sand ($T_2^* = 0.2$ s).

the salty water layer (which has the greatest impact on the magnetic fields) is reasonably defined by the first TEM inversion. Moreover, as showed by Legchenko et al. (2008) the uncertainty in TEM results has an insignificant effect on MRS.

To assess the capability of TEM and MRS methods to characterise shallow aquifers on strip barriers or small islands, we propose a similar hydrogeological model of a thin and shallow freshwater lens. A water content and geoelectrical models are then calculated according to the hydrogeological assumptions presented in 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.

We compute synthetic data according to the field conditions encountered in our survey. Concerning TEM, we consider low noise condition and we use a time range of 5 µS to 1.6 ms with a coincident square loop of 25 m side (shaded zone in Fig. 2a). We add 1 % of noise to the synthetic data. The starting model for inverting the synthetic data is a 5-layer model (chosen according to our hydrogeological a priori). Then, we reduce the number of layers assessing the impact on the RMS (i.e. the root mean square fitting error). The smallest number of layers which provides the smallest RMS comparable to the added noise (1 %) is selected as the best fit output model. Note that several output solutions can equally fit the input data (i.e. equivalent solutions with comparable RMS). We assess the range of acceptable solutions using the equivalence analysis tool of IX1D V3 package (Interpex).

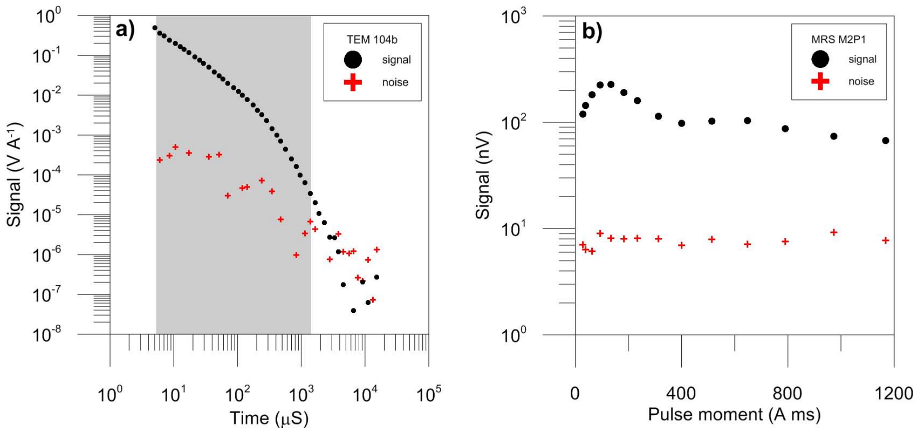

Fig. 2. Example of geophysical field measurments. **(a)** TEM measurements: the grey zone is the time window used for the synthetic modelling. **(b)** MRS measurements.

Concerning MRS, we compute synthetic data considering the low noise conditions of the surveyed area and the capabilities of the used device (Fig. 2b): the loop size (25 m side, 2 turns) and pulse duration (10 ms) are chosen to fit the shallow target, with a Larmor Frequency of 1720 Hz and an inclination of the geomagnetic field of 17° N. 10 nV of Gaussian noise is randomly added to compute synthetic data. Output solution and uncertainty in the MRS results (i.e. the equivalence analysis) are calculated with the Samovar V11 package (Legchenko et al., 2011).

On one hand, numerical modelling conducted with TEM alone shows that resistivity and depth of the salty water layer is determined with a low uncertainty: 15 cm on the depth to the layer and 0.1 Ωm on its resistivity value for the depth range encountered in our simulation. However, the dry sand and the fresh and brackish water layers cannot be separately resolved if the individual 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 better resolved. Then, the resolution of θ_{MRS} can be significantly improved if the geometry Δz of the saturated layers is known. The sequential inversion combines the best capability of each method and improves the determination of the model as compared to the results of each method interpreted independently. Indeed, our numerical modelling shows that (1) the water content and the static water level are estimated with a mean error of ± 8 % and ± 7 %, respectively, (2) the saline water layer is defined with a low uncertainty (less than ± 5 % on the resistivity value and ± 1 % on the layer boundaries), and (3) the fresh and the brackish water layers boundaries are estimated with an accuracy which is controlled by the depth and thickness of the layer. The thickness and the resistivity of the freshwater layer are resolved with an error of ± 3 % and ± 20 %, respectively, if the layer is 10 m thick. If the thickness of the freshwater lens is less than 10 m, the fresh and brackish layer cannot be separated and a single layer is obtained. Note that our modelling results are only valid for low noise and shallow layers conditions. Indeed, noisy data will generate a larger range of equivalence both in TEM and MRS results, and the results of the newly developed joint inversion approaches will probably be more accurate than our simplified sequential approach (Behroozmand et al., 2012; Günther and Müller-Petke, 2012).

2.4 Estimating the recharge of the aquifer

The water table fluctuation method (WTFM) is widely used for calculating unconfined aquifer recharge (Healy, 2010). WTFM is based on the assumption that rise in groundwater level is 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 R can be calculated as

$$R = S_{\mathrm{y}} \cdot \Delta H / \Delta t \tag{5}$$

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} (Scanlon 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.

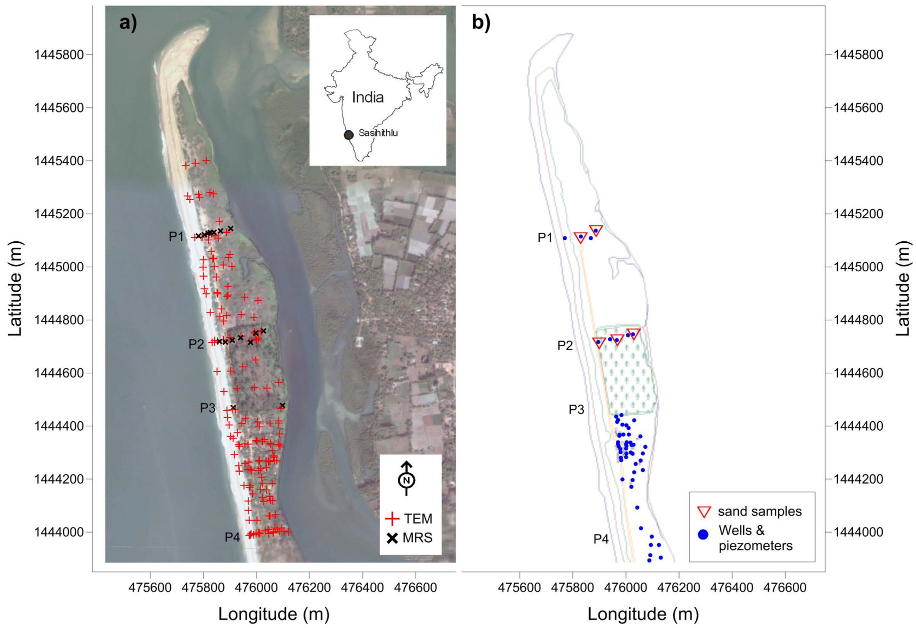

Fig. 3. Location of Sasihithlu barrier and measurements. **(a)** Geophysical measurements (including pre and post monsoon time lapse). **(b)** Hydrological measurements.

3 Experiments

3.1 Investigated site

The proposed methodology has been applied on a coastal barrier in South-Western India, Karnataka state (Fig. 3). The so-called 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 a few months per year and sea water invades 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 the sand deposit is unknown. The highest elevation of the barrier is a strip dune located on the backshore with a maximum elevation of 4 m a.m.s.l. (above mean sea level).

The annual rainfall is 3900 mm (average for the 1970 to 1985 period) which mainly occurs during the southwestern 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 freshwater lens morphology is not known.

3.2 Methods and material

Two geophysical surveys have been carried out: a first one before the monsoon (February/March) and a second one at the end of the monsoon (Octobre/November). Additional MRS and TEM measurements were carried out during the monsoon on two profiles (P1 and P2, Fig. 3) 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 per side was used (except at one location where a 50 m 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 using the known TEM resistivity for computing the magnetic fields (Legchenko et al., 2008).

A total of 140 TEM soundings have been carried out (60 soundings in February/March and 80 soundings in October/November) using the TEM-FAST 48HPC instrument from AEMR (Applied Electromagnetic Research). A coincident loop of square shape and 25 m per side has been used. The electromagnetic noise level was low and the signal to noise ratio remained high for all measurements within the selected time window (usually higher than 50 for the late times, Fig. 2a). TEM data were interpreted with IX1D V3 software (Interpex).

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.

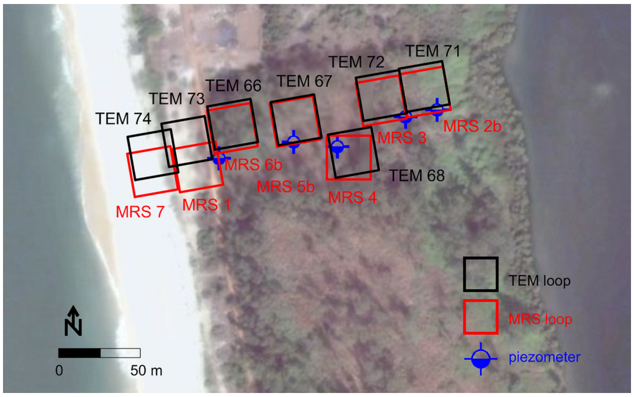

Fig. 4. Location of measurements, Profile 2.

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.

To validate our approach, we drilled 9 observation piezometers and we used the 65 wells monitoring data. We did not carry out hydraulic tests in order to prevent the upcoming of saline water.

4 Results and discussion

4.1 Parameterisation of Archie's equation

TEM and MRS soundings are implemented on the beach (for example TEM74 and MRS7, Fig. 4). 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. 5a) and a 5 layered resistivity model (Fig. 5b, note that the inversion of TEM alone results in only 4-layer model). The resistivity of the first layer is not known because (1) the TEM measurement has poor resolution from ground surface to about 1.5 m deep with the used configuration and equipment and (2) resistive targets are not well resolved with TEM measurements (Albouy et al., 2001). Crossing both MRS and TEM results obtained with the sequential inversion, one can propose an obvious hydrogeological interpretation (Fig. 5c): from ground surface down, we interpret a dry sand layer, a sandy layer saturated with probably brackish water, then a sandy layer saturated with sea water. The 4th layer, which is more conductive than the sea water layer, is interpreted as a clayey weathered gneiss (saturated with sea water) as observed in deep boreholes drilled several kilometres north of Sasihithlu barrier. The last resistive layer is the gneissic substratum. To parameterise 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

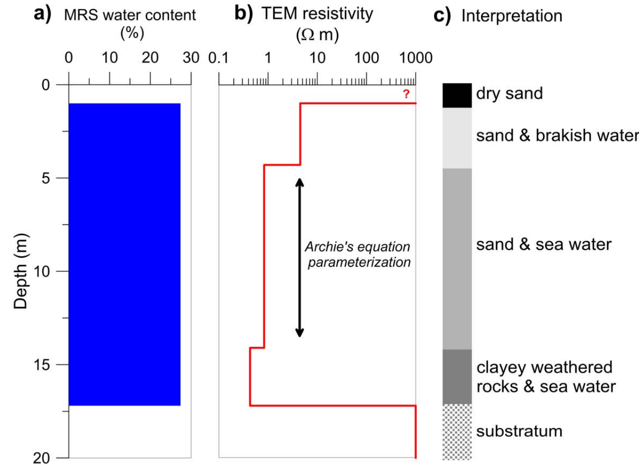

Fig. 5. Example of results of sequential inversion at the beach side. **(a)** MRS7, **(b)** TEM74 and **(c)** hydrogeological interpretation.

m factor of Archie's first equation for the sandy sea water saturated layer ($\theta_{MRS} = 0.27$ and $\rho_{aq} = 0.83\,\Omega\mathrm{m}$) as

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

The m-values calculated from 16 MRS/TEM 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. Knowing the m-value, we can now use the parameterised Archie equation for calculating the water EC whatever the depth and location.

4.2 Quantification of the freshwater reserve

An example of the application of the proposed methodology is presented in Fig. 6. The TEM and MRS field data are first inverted using the sequential process presented Sect. 2.3 (Fig. 6a). 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 layered resistivity model (Fig. 6b). Then, we proceed to the hydrogeological interpretation of the geophysical models (Fig. 6c):

1. The distribution of groundwater EC with depth is calculated from the TEM resistivity model using the parameterised Archie equation. Because the TEM does not differentiate the freshwater from the brackish water layers, we use a simplified model which sets that the water EC is linearly increasing with depth. This assumption is probably acceptable because it results in calculated water EC values close to measured values. For example the value of water EC measured in the piezometer adjacent to the MRS loop (Fig. 4) is $EC_{piezometer} = 630\,\mu\mathrm{S\,cm^{-1}}$ which is close to the value calculated for the same depth from the TEM measurements $EC_{TEM} = 600\,\mu\mathrm{S\,cm^{-1}}$

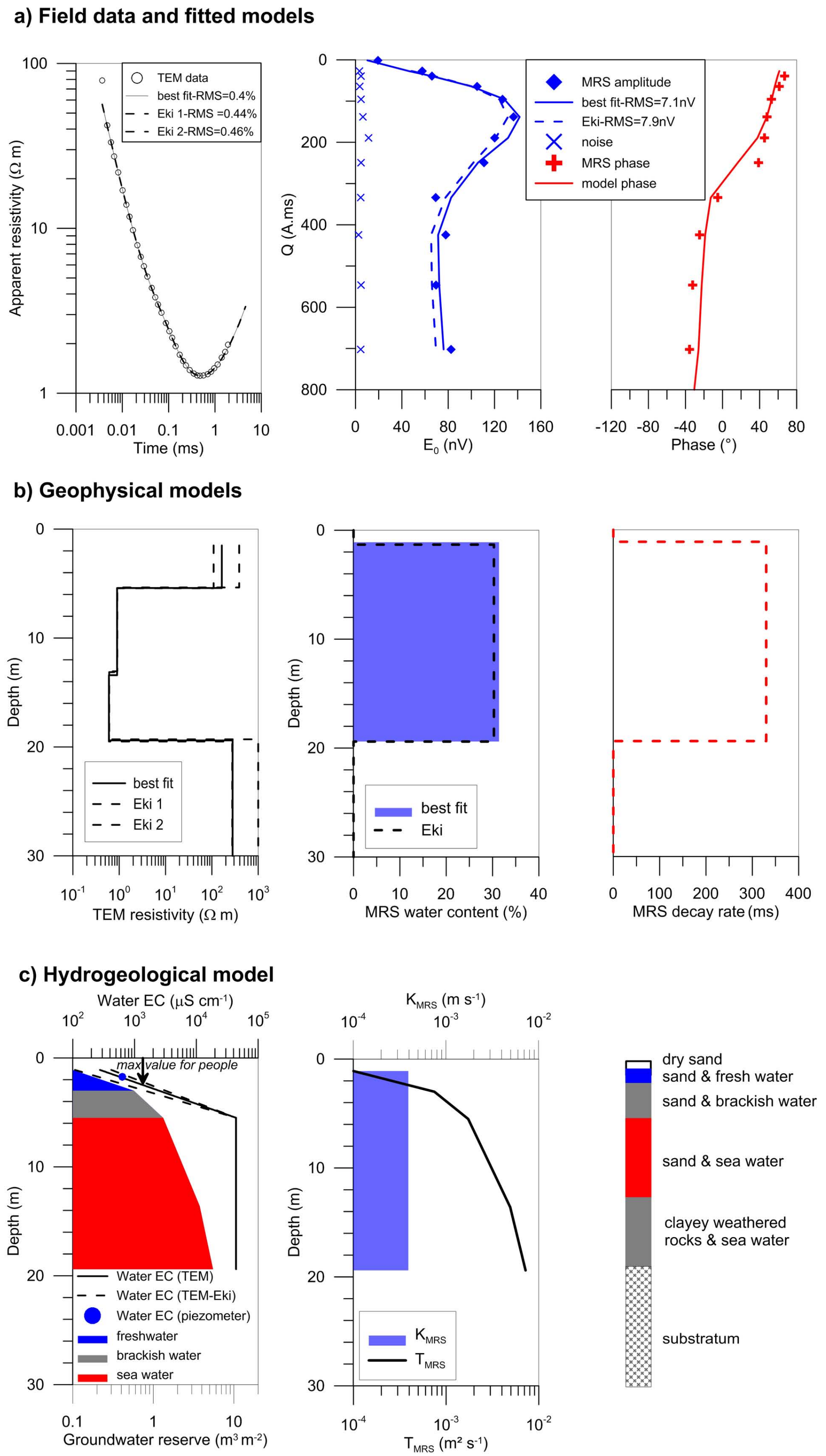

Fig. 6. 1-D application of the hydrogeophysical methodology. **(a)** MRS6 and TEM66 data and fitted models. **(b)** Geophysical models. **(c)** Hydrogeological interpretation. Eki are the calculated equivalent solutions.

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

	Dry season (February–March)		Rainy season (October–November)	
	Average	Uncertainty	Average	Uncertainty
North of the area (uninhabited)	2.2 m 500 L m^{-2}	43 %	2.9 m 670 L m^{-2}	34 %
South of the area (inhabited)	1.8 m 420 L m^{-2}	37 %	2.0 m 460 L m^{-2}	51 %

(Fig. 6c). Then, the maximum value of water EC which is acceptable by the local communities for domestic supply ($EC_{\text{domestic water}} < 1500\,\mu\text{S cm}^{-1}$) is used to estimate the thickness of the freshwater 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 28.3 and 33.6 % with an average relative uncertainty of 0.1 % (due to the uncertainty in weight measurements). 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 excluding 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 $213 \times 10^{-6}\,\text{m}^3$ sample cannot be rigorously compared to MRS measurements ($\approx 280\,\text{m}^3$ for the survey), θ_{MRS} is close to "ne" for the sampled sand. Moreover, the long MRS decay rates indicate coarse average grained-size of the sand (Fig. 6b) for which the amount of capillary water is probably negligible as compared with the amount of gravitational water. Thus $\text{ne} \approx S_y \approx \theta_{\text{MRS}}$. The distribution of θ_{MRS} with depth is then used to calculate the groundwater reserve as the product $\theta_{\text{MRS}} \cdot \Delta z$ 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. 6a and b). The total water reserve at this location is $5.5\,\text{m}^3\,\text{m}^{-2}$ of surface area. However, only 600 L of water ($\pm$ 50 L according to the equivalent solutions) per square metre of surface area have an EC lower than the threshold value and are considered as freshwater.

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

Using rapid MRS/TEM 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. 7a) and 3-D interpolations techniques can be used for mapping the freshwater lens (Fig. 7b). Finally, from all the TEM and the MRS carried out before and after the monsoon, we are able to estimate the freshwater thickness and corresponding freshwater volume (Table 1). We computed the uncertainty of our results based on uncertainties in geophysical inversion (both MRS and TEM). We conclude that our approach allows estimating groundwater reserve with a relative uncertainty of 30 to 50 % (Table 1).

4.3 Recharge and groundwater resource estimate

For estimating the recharge, we use both hydrological monitoring (Fig. 8) and geophysical time lapse results (Fig. 9). From the water table monitoring, 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 using the WTFM 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. Figure 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 of the water table remains approximately constant at the year scale. However, 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

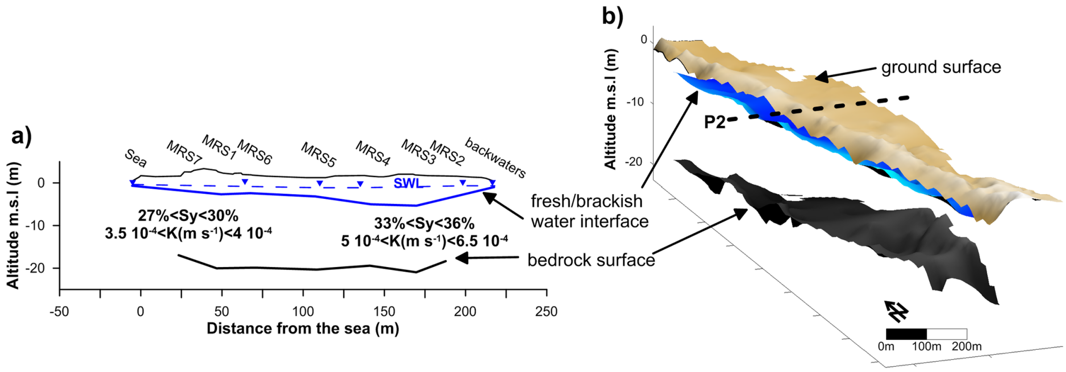

Fig. 7. 2–3-D representation of geophysical results (before the monsoon). **(a)** 2-D section of Profile P2. SWL is the static water level measured in piezometers. **(b)** 3-D map of the aquifer.

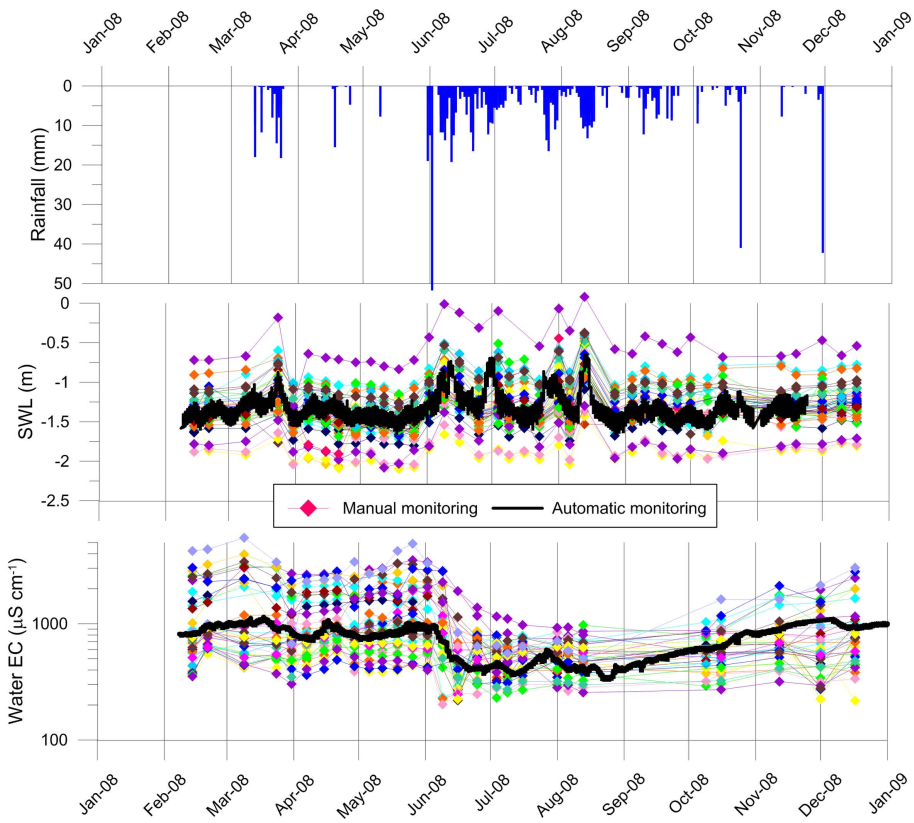

Fig. 8. Rainfall, water level and water EC monitoring. SWL is the static water level measured in piezometers.

mineralised groundwater and decreases during the rainy season with the infiltration of low-mineralised rainwater. Hence, at the monsoon time scale, the rain water which infiltrates the aquifer is deepening the salty water interface more than raising the water level.

For estimating the recharge at the monsoon time scale, we compare TEM and MRS measurements carried out at the same location but at different period of time. An example of the geophysical time lapse measurements is presented in Fig. 9. Since the depth to the sea water layer is defined by TEM interpretation with a high accuracy ($\pm 1\,\%$ of uncertainty on the layer boundary, 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 1.090 m of rain and reveals a deepening of the salty water interface of 2.4 m. Using the WTFM (Eq. 5) the amount of water which reaches the water table and stays at the surveyed location is 0.7 m or 66 % of the rain. A third TEM is carried out in October

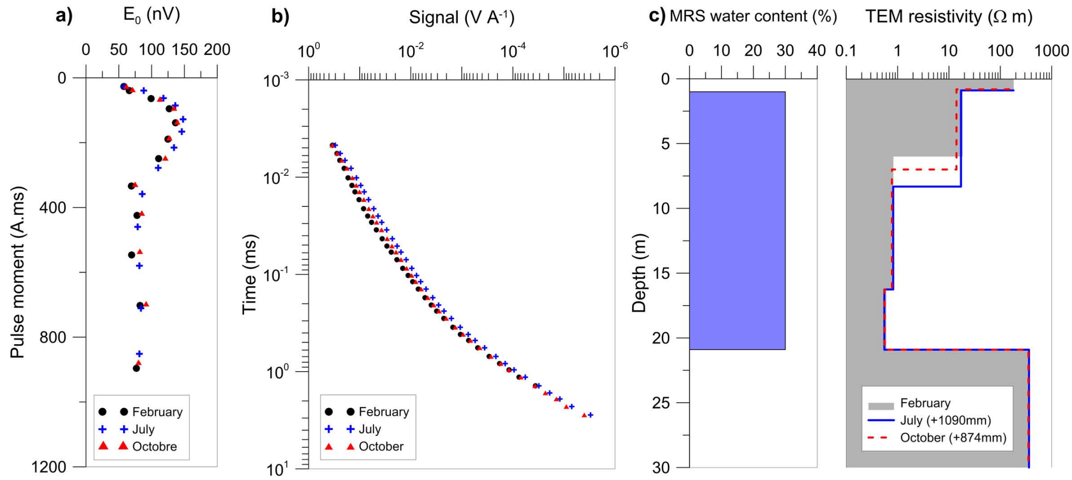

Fig. 9. TEM73 and MRS 1 time lapse. **(a)** MRS signals, **(b)** TEM signals and **(c)** geophysical models.

after a new 0.874 m of rainfall. Surprisingly, the TEM indicates that the sea water interface rises 1.5 m. This observation is consistent with the groundwater monitoring and it can be 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 before and after the monsoon all over the targeted area, the average net aquifer recharge is 9 % of the rain. Note that the MRS time lapse carried out on the same dates as the TEM does not indicate any change of the depth to the water table within the uncertainty of ± 8 cm (Fig. 9).

Finally, joint analysis of MRS/TEM 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 freshwater 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 head conditions: it will reduce the thickness of the freshwater lens together with the net recharge.

5 Discussion

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

Not only 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 proposed to use a linear gradient of electrical conductivity with depth and we calculated the depth corresponding to the 1500 µS cm^{-1} threshold thanks to the parameterised Archie's 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 another common electromagnetic method for improving the resolution of shallow targets. We carried out frequency conductivity sounding (CS) using 2 light multi-turn coils. We implemented the CS with the EM-34 device (Geonics) 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-layer model presented in Fig. 1 can be solved with a 3-layer 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 characterisation of thin lenses as compared to TEM.

Concerning the MRS method, the output parameters still need to be parameterised with known hydraulic conductivity and storage-related parameters. Today, the methodology developed in this paper for quantifying groundwater resource cannot rely on geophysical tools alone but needs a coupled hydrological and geophysical approach. However, the potential of MRS for characterising 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; Vouillamoz et al., 2012).

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

Finally, we have shown that our sequential inversion approach leads to acceptable results in favourable cases (i.e. low noise level, shallow salty water layer and few layer model) with uncertainty in the estimate of freshwater reserve ranging in-between 30 and 50 %. However, as showed by Güther and Müller-Petke (2012) uncertainties of sequential inversion are slightly but generally higher than uncertainties of joint inversion.

6 Conclusions

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 not only found that the geometry of the freshwater lens, the specific yield and the hydraulic conductivity of the aquifer, but also the recharge and the behaviour of the lens can be estimated. We estimate that the freshwater reserve ranges between 400 and 700 $\mathrm{L\,m^{-2}}$ of surface area, and that about 100 % of the rain infiltrates and reaches the water table. However, more than 90 % of this infiltrated rain water does not increase the freshwater 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 freshwater resource but a rise of the mean sea level will both reduce the freshwater reserve and the net recharge.

This study also pointed out some limitations of the use of geophysical tools for assessing shallow and thin freshwater 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 resource. Moreover, new developments in the MRS method, joint use of time lapse MRS/TEM and joint inversion of MRS/resistivity measurements are promising for improving aquifer characterisation.

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