
Estimating specific yield and transmissivity with magnetic resonance sounding in an unconfined sandstone aquifer (Niger)

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Abstract The unconfined aquifer of the Continental Terminal in Niger was investigated by magnetic resonance sounding (MRS) and by 14 pumping tests in order to improve calibration of MRS outputs at field scale. The reliability of the standard relationship used for estimating aquifer transmissivity by MRS was checked; it was found that the parametric factor can be estimated with an uncertainty $\leq 150\%$ by a single point of calibration. The MRS water content (θ_{MRS}) was shown to be positively correlated with the specific yield (Sy), and θ_{MRS} always displayed higher values than Sy. A conceptual model was subsequently developed, based on estimated changes of the total porosity, Sy, and the specific retention Sr as a function

of the median grain size. The resulting relationship between θ_{MRS} and Sy showed a reasonably good fit with the experimental dataset, considering the inherent heterogeneity of the aquifer matrix (residual error is $\sim 60\%$). Interpreted in terms of aquifer parameters, MRS data suggest a log-normal distribution of the permeability and a one-sided Gaussian distribution of Sy. These results demonstrate the efficiency of the MRS method for fast and low-cost prospection of hydraulic parameters for large unconfined aquifers.

Keywords Magnetic resonance sounding (MRS) · Pumping test · Hydraulic properties · Sedimentary rocks · Niger

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Introduction

Evaluation and management of groundwater resources require hydrogeological system parameterization at the aquifer scale. This parameterization is generally performed by some local measurements of hydrogeological properties, and estimating their distribution either by interpolation, by geostatistical methods or by numerical modelling (De Marsily et al. 1999). In this framework, non-invasive geophysical surveys can be carried out for improving the knowledge of hydrogeological properties and of their distribution (e.g. Ezzy et al. 2006; Singha et al. 2007).

The magnetic resonance sounding (MRS) method is particularly adapted to hydrogeological investigations, being sensitive almost exclusively to groundwater (Roy and Lubczynski 2003; Legchenko et al. 2004). In natural conditions, MRS was shown to provide consistent results on transmissivity, once calibrated using pumping tests data. However, the calibration process still needs to be improved with respect to the estimation of storativity (Lubczynski and Roy 2005; 2007). In a recent review, Vouillamoz et al. (2007a) reported only two datasets where the MRS water content could be compared to the storativity measured by pumping tests: one with six data in a mixed confined/unconfined weathered granite aquifer in Burkina Faso (Vouillamoz et al. 2005), the second one with seven data in a clayey to sandy confined aquifer in Myanmar (Vouillamoz et al. 2007b).

In the well-known unconfined aquifer of SW Niger, 23 MR soundings were recently carried out. Eleven of these

soundings were performed near boreholes where pumping tests were previously conducted. In November 2006, three piezometers (water table depth between 8 and 24 m) were drilled near pumping wells in order to obtain additional information on aquifer storativity. A total of 14 transmissivity values and six specific yield values were finally obtained for sites where MR soundings were also performed. To the best of our knowledge, this represents the largest MRS-pumping test dataset currently available for a single unconfined aquifer.

Preliminary results (15 MR soundings performed in 2005 with seven values of transmissivity but no specific yield data) were presented in Vouillamoz et al. (2008). This article presents additional measurements and complementary analysis of MRS data used for estimating aquifer parameters. The main goal of the study was to propose a conceptual model of calibration for estimating the specific yield using MRS. The second goal was to assess the reliability of transmissivity values estimated by MRS, using a reliable dataset of pumping test and MRS data obtained in a single unconfined aquifer.

Overview of hydrogeological parameters estimated by MRS

Storage parameters

Several usual hydrogeological parameters describe the quantity of water stored in aquifers: the total porosity, the effective porosity, the storativity and the specific yield. The total porosity (n) is the proportion of voids in relation to the total volume of rock. The effective porosity (n_e) is the ratio between the volume of mobile water in saturated aquifer (i.e. water that contributes to flow) and the total volume of rock. The storativity (S) is the percentage of water that can be extracted from 1 m of saturated aquifer thickness under a drop of 1 m of head. The specific yield (S_y) is the ratio between the volume of water that can be extracted by gravity and the total volume of drained rock. The specific yield can be expressed as follow:

$$S_y = n - S_r \quad (1)$$

where S_r is the specific retention, which includes bound water (i.e. water that is attached to the solid grain by molecular forces), water in unconnected pores and water retained by capillarity. For an unconfined aquifer, the specific yield and the storativity can be considered as equivalent, whereas in a confined aquifer, the storativity depends both on the compressibility of water and of porous media.

Various methods (either field experiments or laboratory measurements) are used to estimate storage parameters (e.g. Stephens et al. 1998; Healy and Cook 2002). Pumping tests are known to be difficult to interpret accurately in terms of specific yield (Meier et al. 1998; Kollet and Zlotnik 2005; Tartakovsky and Neuman 2007) but still represent one of the best available methodologies at field-scale. In this study, data obtained by pumping tests were used to calibrate MRS outputs, because the radius of

influence of pumping tests (typically ranging in distance from the pumping well from a few tens to hundreds of metres) corresponds approximately to the MRS investigation scale, which typically ranges from 500 to 15,000 m² at surface (Legchenko et al. 2004).

MRS parameters

Parameters obtained by magnetic resonance soundings are the water content (θ_{MRS}) and the relaxation times (T_1 and T_2^*) versus depth (Legchenko et al. 2004). θ_{MRS} is defined as the ratio between the volume of water detectable by MRS and the total sampled volume. The relaxation times T_1 and T_2^* (usually expressed in ms) are related to the exchange of energy, respectively between protons and their environment and between protons.

Up to present (2009), the existing instrumental implementation does not allow the measuring of signals with short relaxation time T_2^* (shorter than the instrumental dead time of 40 ms). As a consequence, part of the aquifer water content cannot be detected by MRS, and the MRS water content is always lower than the total porosity. The relaxation time T_2^* depends mostly on the distance between water molecules and the pore walls: the longer the relaxation time, the higher the distance of water molecules to pore walls (and conversely). The undetectable part of the water content corresponds to water molecules located close to solid grain, i.e. typically bound water. As a consequence, no signal is usually recorded from clayey formations. The MRS water content is assumed to be close to the effective porosity (n_e), as long as unconnected and dead-end pores can be neglected (Vouillamoz et al. 2007a). This assumption is reasonable for sandstones and quartz-rich clastics but is likely to be less valid for carbonates where $\theta_{\text{MRS}} > n_e$ or for volcanic rocks where $\theta_{\text{MRS}} < n_e$ (Legchenko et al. 2004). However, additional field validations of this assumption are needed to better constrain MRS interpretations (Lubczynski and Roy 2005, 2007).

Relationships between MRS and hydrogeological parameters

The relaxation times of the MRS signal (T_1 and T_2^* , in ms) are linked to the mean size of pores that contain groundwater and therefore provide a flow-related parameter. Considering oil industry experience, the followed relationship was established for estimating the aquifer transmissivity (Legchenko et al. 2002):

$$T_{\text{MRS}} = C_p \cdot \theta_{\text{MRS}} (T_1)^2 \Delta z \quad (2)$$

where T_{MRS} (in m²/s) and θ_{MRS} (dimensionless) are the transmissivity and the water content estimated from MRS, Δz is the thickness of the saturated layer (in m), and C_p (in m/s/ms²) is a parametric factor which ranges between $\sim 10^{-10}$ and $\sim 10^{-6}$ m/s/ms² and mostly depends on the geological context. Equation 2 can be used with the parametric factor proposed

by default in inversion softwares (7×10^{-9} m/s 2), but for a more accurate estimate of the aquifer transmissivity, calibration with some local pumping tests is required. This empirical relationship is now widely used (e.g. Vouillamoz et al. 2007b; Chalikakis et al. 2008) but the consistency of the parametric factor heavily depends on the dataset available for calibration. Once the parametric factor calibrated for a given aquifer, the equation can be applied to other sites in the same geological context in order to increase rapidly and at low cost the number of transmissivity estimates.

The parametric factor C_p in Eq. (2) is usually expressed for a temperature of 10°C (Vouillamoz et al. 2007b). The transmissivity T depends both on fluid and on rocks properties, as follows:

$$T = k \cdot \frac{g}{\nu} \cdot \Delta z \quad (3)$$

where k is the intrinsic permeability, g is the acceleration of gravity and ν is the kinematic viscosity of the fluid, that is strongly temperature-dependent. Thereby, a temperature correction is necessary for comparing parametric factor of aquifers in different geological or climatic contexts.

In a first attempt to correlate the MRS water content with the specific yield, Vouillamoz et al. (2005) proposed the following equation:

$$S_{yMRS} = C_y \cdot \theta_{MRS} \quad (4)$$

where S_{yMRS} is the MRS specific yield, θ_{MRS} is the MRS water content and C_y is a parametric factor depending on the geological context. The relevance of this equation will

be discussed below considering the new data obtained in southwestern Niger.

Field experiments

The Continental Terminal unconfined aquifer

The studied aquifer is located in semiarid southwestern Niger, eastward of the Niger River and at few tens of kilometres east of Niamey (Fig. 1). In this area, the unconfined aquifer belongs to the Continental Terminal (CT) formation, a large aquifer that spreads over 150,000 km 2 , and consists in loosely cemented sandstones of tertiary origin (Lang et al. 1990). Eastward, the valleys of the Dallol Bosso and the Dallol Maouri are paleo-rivers filled up with coarse Quaternary sands (Fig. 1, small inset). Numerous reliable field measurements of water table level and groundwater salinity were performed for the past two decades in this well-known part of the aquifer (Favreau et al. 2009). The water-table depth ranges from a few metres below the dry valleys to up to 75 m below the lateritic plateaus; the saturated aquifer thickness is a few tens of metres. The substratum of the unconfined CT aquifer is a continuous clayey layer of a few tens of metres in thickness that prevents any leakage from the underlying confined aquifers.

MRS implementation

Twenty-three MR soundings were performed in December 2005 and in November 2006 (Boucher 2007). The

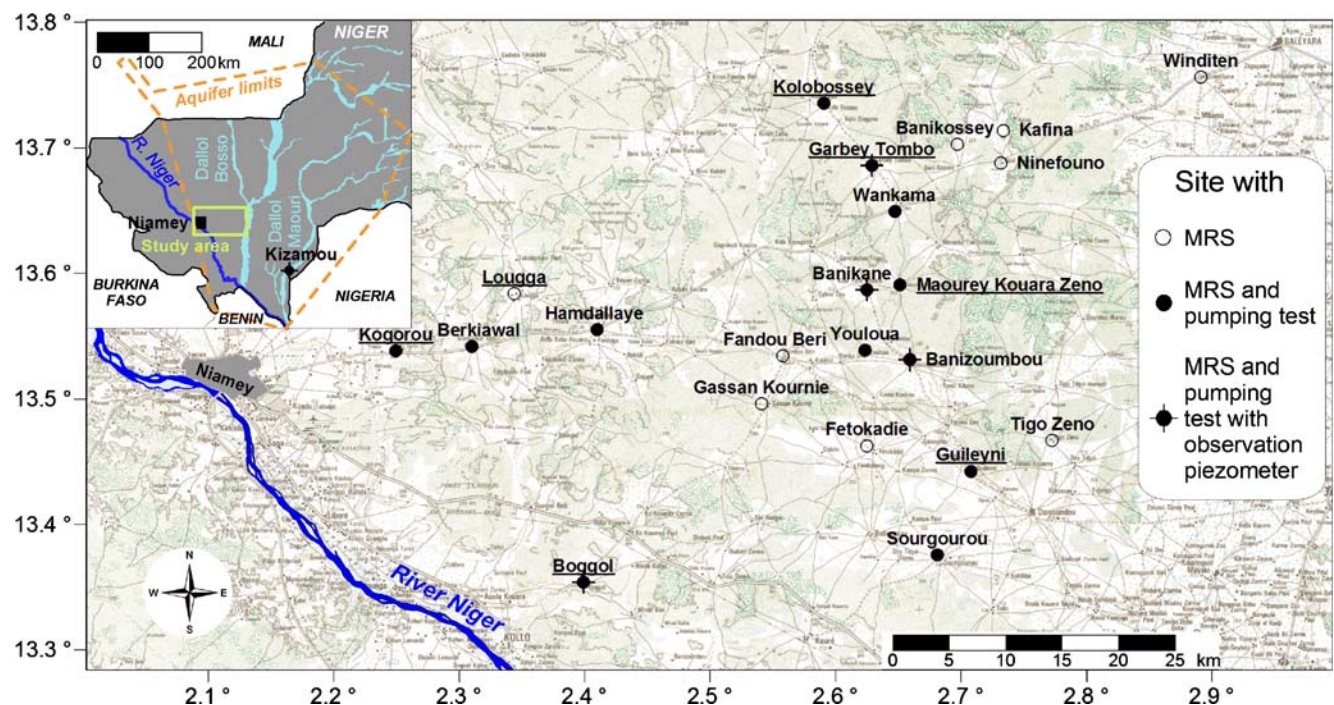


Fig. 1 Location of the surveyed sites. The *underlined sites* were surveyed with MRS in November 2006, while *non-underlined sites* were surveyed in December 2005 (Vouillamoz et al. 2008). *Small inset*: location of the study area in southwestern Niger and aquifer limits (orange dotted line); large paleovalleys (locally referred to as “Dallol”) are delineated in light blue

location of the prospected sites was chosen first to make possible a calibration of MRS parameters with existing pumping test data (Fig. 1), and to get a representative dataset of the changing properties of the aquifer (water-table depth, aquifer thickness). The observed electromagnetic noise, which is considered as the main source of error for MRS (Girard et al. 2005) was low and allowed for a high signal-to-noise ratio, most often > 3 , to be reached (Vouillamoz et al. 2008).

The methodology for data interpretation was presented in details in Vouillamoz et al. (2008). The data inversion was performed by considering a single layer for the saturated zone. The aquifer geometry (depth and thickness) was fixed in the inversion process using previous geological information, in order to solve the so-called issue of “equivalences” (i.e. the fact that different distributions of water content can result in similar MRS signal; Legchenko et al. 2004); this allowed increased accuracy on the estimate of the MRS water content. The effect of groundwater temperature (ranging from 28 to 33°C in the surveyed area) was taken into account for MRS interpretation.

Uncertainty in the amplitude of the MRS signal depends on both the electromagnetic (EM) noise and on T_2^* uncertainty. The mean EM noise was calculated from the amplitude of the stacked noise of each pulse moment, and uncertainties on the decay time T_2^* were computed using automatic inversion conducted with Samovar inversion software. Uncertainty in the initial amplitude of the MRS signal was estimated from the uncertainty in T_2^* value (used to extrapolate the initial amplitude of the signal) and within the range of the mean EM noise (Vouillamoz et al. 2008).

Pumping tests

Pumping tests performed in the Niger part of the Continental Terminal unconfined aquifer were discussed in Favreau (2000). For the scope of this study, this data set was supplemented by drilling three observation piezometers (of 8, 13 and 24 m in depth) near three pumping wells in order to obtain three additional values of the specific yield. All pumping test data were fully re-analysed, except for one site (Kizamou; Fig. 1, small inset) for which a detailed summary of the interpretation was available (Pallas 1970).

A first interpretation of the drawdown and recovery of pumping test data was performed using the simplified Cooper and Jacob (1946) approximations of the Theis (1935) equation, in order to get an approximate estimate of the unconfined aquifer parameters. In a second step, considering the characteristics of each pumping well, a more refined interpretation was performed using a commercial program (Aquifer Test Pro, Schlumberger Water Services) for solving equations. Most of the pumpings were performed in large diameter (1.80 m) hand-dug wells, with a few cubic metres of groundwater filling the wells. The analytical solution of Papadopoulos and Cooper (1967), which takes into account the capacity

effect in wells of large diameter, was chosen to interpret the drawdown in pumping wells. As drawdown was usually less than 10% when compared to the saturated thickness of the aquifer (Table 1), the drawdown in the observation piezometer (when existing) was interpreted with the Theis (1935) analytical solution. The recovery curves were interpreted with the Agarwal's (1980) solution which is not affected by well effects. The uncertainty on both transmissivity and specific yield was estimated by sensitivity testing of the different interpretations. The averaged uncertainty on transmissivity was between +70% and -50%, with maximal uncertainty between +330 and -96%. The uncertainty on the specific yield was on average of 40%, for a maximum value of 55%.

Investigated aquifer volumes

The radius of influence (R) of each pumping test was computed with Dupuit's formula, which gives the drawdown in permanent state, and from the Jacob-Cooper logarithmic approximation (in De Marsily 1986):

$$R = 1.5\sqrt{Tt/S} \quad (5)$$

where T and S are the estimated transmissivity and storativity (here, the specific yield), and t is the duration of the pumping test. The calculated radiuses of influence ranged between 74 and 280 m. Another estimate of R was obtained by simulating the drawdown (using Aquifer Test Pro) according to the analytical solution that was chosen for the pumping-test interpretation. This method provided similar results as those obtained by Eq. (5), with radiuses of influence ranging between 56 and 350 m. Direct observations confirmed this range of values, with drawdowns of respectively 0.14 and 0.20 m in piezometers located at 64 and 100 m from two different pumping wells. Comparatively, the radius of influence investigated by MRS ranged from ~ 100 to ~ 300 m, mostly dependent upon the size of the loops (75×75 m or 50×50 m). In depth, the thickness investigated by MRS corresponds to a mean value of the whole saturated thickness of the aquifer, whereas pumping tests mostly characterize the superficial layer that is dewatered during pumping of the unconfined aquifer.

Results

MRS transmissivity

MRS transmissivity was estimated using the standard conversion Eq. (2); the averaged parametric factor (C_p) value for the CT aquifer was of 1.4×10^{-8} m/s/ms²—which corresponds to 8.7×10^{-9} m/s/ms² at 10°C when taking into account the change of permeability with temperature; cf. Eq. (3). This value remains close to that obtained after the first field acquisition (1.0×10^{-8} m/s/ms² at 10°C), computed with seven points of calibration (Vouillamoz et al. 2008; Fig. 2). Considering the possible range of C_p values (from $\sim 10^{-10}$ up to $\sim 10^{-6}$ m/s/ms²; Vouillamoz

Table 1 Pumping tests and MRS characteristics

Site	Pumping test					MRS				
	Formation	Duration	s_{max} (m)	Δz (m)	T_{PI} (m ² /s)	$Sy_{(PI)}$ (%)	T_{MRS} (m ² /s)	Cp (10^{-8} m/s/ms ²)	θ_{MRS} (%)	Sy_{MRS} (%)
Banikane (BK)	CT	23 h	1.34	24	4.6×10^{-3} ($1.5-6.4 \times 10^{-3}$)	1.1±0.5	$4.6 \cdot 10^{-3}$ ($2.7-7.4 \cdot 10^{-3}$)	1.6	14.6 (12.7-17.2)	2.9
Banikossej	CT	-	-	-	-	-	$4.1 \cdot 10^{-3}$ ($1.8-7.9 \cdot 10^{-3}$)	-	16.1 (12.9-19.2)	3.7
Banizoumbou (BZ)	CT	46 h	12.91	29	$3.7 \cdot 10^{-3}$ ($1.5-16.10^{-3}$)	2.1±0.5	$1.7 \cdot 10^{-3}$ ($1.1-2.7 \cdot 10^{-3}$)	3.3	9.7 (8.3-11.3)	0.90
Berkiawal	CT	72 h	3.72	17	6.1×10^{-4} ($2.0-7.3 \times 10^{-4}$)	-	9.7×10^{-4} ($1.6-2.9 \times 10^{-4}$)	0.97	5.4 (3.7-7.2)	0.14
Boggol (BG)	CT	30 h	1.26	13	2.2×10^{-3} ($1.2-4.5 \times 10^{-3}$)	1.1±0.6	1.5×10^{-3} ($0.6-2.9 \times 10^{-3}$)	2.3	10.8 (7.5-14.0)	1.2
Fandou Beri	CT	-	-	-	-	-	2.4×10^{-4} ($0.5-6.4 \times 10^{-4}$)	-	7.6 (6.0-9.2)	0.43
Fetokadie	CT	-	-	-	-	-	1.8×10^{-3} ($0.8-3.6 \times 10^{-3}$)	-	8.8 (6.8-11.7)	0.68
Garbey Tombo (GT)	CT	19 h	1.29	35	3.2×10^{-3} ($1.0-4.8 \times 10^{-3}$)	9.0±3	9.7×10^{-3} ($3.2-2.2 \times 10^{-3}$)	0.46	19.9 (14.9-24.8)	6.4
Gassan Kournie	CT	-	-	-	-	-	2.9×10^{-3} ($1.4-5.1 \times 10^{-3}$)	-	14.9 (12.3-17.0)	3.0
Guileyni	CT	24 h	1.81	36	6.9×10^{-3} ($6.0-8.0 \times 10^{-3}$)	-	4.5×10^{-3} ($1.6-10 \times 10^{-3}$)	0.90	12.9 (9.7-17.2)	2.0
Hamdallaye	CT	26 h	2.52	23	1.9×10^{-3} ($0.3-4.2 \times 10^{-3}$)	-	1.5×10^{-3} ($0.7-2.6 \times 10^{-3}$)	2.0	10.2 (7.8-12.6)	1.0
Kafina	CT	-	-	-	-	-	7.3×10^{-3} ($2.9-15.10^{-3}$)	-	10.2 (7.7-12.6)	1.0
Kizamou (K1)	Q	≥ 72 h	-	6	1.4×10^{-2} ($1.0-2.0 \times 10^{-2}$)	20±8	2.0×10^{-2} ($0.9-3.6 \times 10^{-2}$)	-	43.0 (37.6-48.4)	18.3
(K2)	CT	-	-	54	6.9×10^{-3} ($6.0-8 \times 10^{-3}$)	1.2±0.5	1.6×10^{-2} ($0.6-3.2 \times 10^{-2}$)	0.67	15.6 (12.4-18.8)	3.4
Kogorou	CT	48 h	3.7	11	2.0×10^{-3} ($1.5-2.8 \times 10^{-3}$)	-	3.3×10^{-3} ($1.3-9.4 \times 10^{-3}$)	0.95	14 (9.8-26.0)	2.5
Kolobossej	CT	30 h	0.74	31	1.2×10^{-3} ($0.9-1.5 \times 10^{-3}$)	-	7.3×10^{-3} ($3.2-15 \times 10^{-3}$)	1.9	22.6 (18.5-28.2)	8.6
Lougga	CT	-	-	-	-	-	2.0×10^{-3} ($0.2-7.7 \times 10^{-3}$)	-	8.5 (4.8-13.0)	0.60
Maourey Kouara Zeno	CT	48 h	1.17	30	4.5×10^{-3} ($3.0-9.0 \times 10^{-3}$)	-	5.3×10^{-3} ($2.1-11 \times 10^{-3}$)	1.3	12.6 (9.4-16.6)	1.9
Ninefouno	CT	-	-	-	-	-	4.7×10^{-3} ($0.9-13 \times 10^{-3}$)	-	11.2 (8.1-14.3)	1.4
Sourgourou	CT	4 h	0.82	36	4.5×10^{-3} ($3.8-5.2 \times 10^{-3}$)	-	2.6×10^{-3} ($1.2-5.0 \times 10^{-3}$)	2.6	10.8 (8.4-13.3)	1.2
Tigo Zeno	CT	-	-	-	-	-	6.9×10^{-3} ($5.4-8.5 \times 10^{-3}$)	-	16.8 (15.8-17.5)	4.1
Wankama	CT	25 h	1.60	37	2.6×10^{-3} ($1.9-3.0 \times 10^{-3}$)	-	3.3×10^{-3} ($1.5-6.2 \times 10^{-3}$)	1.2	12.0 (10.2-13.9)	1.7
Winditen	Q	-	-	-	-	-	-	-	~30 ~17	-
Youloua	CT	4 h	0.94	25	6.1×10^{-3} ($6.1-6.7 \times 10^{-3}$)	-	6.3×10^{-3} ($4.0-9.9 \times 10^{-3}$)	1.5	18.1 (16.5-21.0)	5.0

The ranges of possible values considering uncertainty are shown in *parentheses*. All data are given for an average groundwater temperature of 30°C (in Vouillamoz et al. (2008), MRS data were reported at 10°C). CT Continental Terminal; Q Quaternary sands. s_{max} maximal drawdown in the pumping well; Δz thickness of the saturated aquifer; T transmissivity; Sy specific yield; θ water content; Cp parametric factor of Eq. (2). The ranges of possible values considering uncertainty are shown in *parentheses*. All data are given for an average groundwater temperature of 30°C (in Vouillamoz et al. (2008), MRS data were reported at 10°C). The possible ranges of $T_{(PI)}$ and $Sy_{(PI)}$ values in Kizamou were computed using the mean relative uncertainty from the whole set of pumping tests

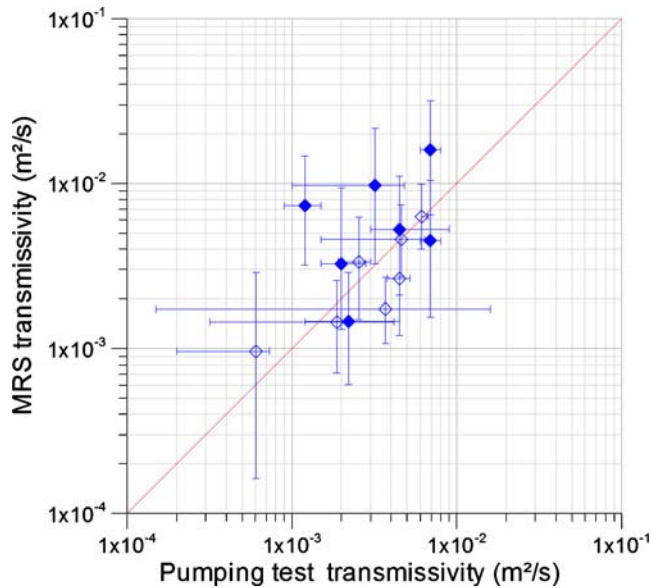


Fig. 2 Transmissivity estimated by MRS reported as a function of transmissivity estimated by pumping test. *White diamonds* represent data discussed in Vouillamoz et al. (2008), while *blue diamonds* correspond to additional data (this study)

et al. 2007a), the 13% change in the C_p value obtained when adding seven new transmissivity data in the calibration dataset is fairly minor. When using the seven more recent pumping test results as a validation dataset, the mean uncertainty on computed T_{MRS} was of 150%, compared to the transmissivity value estimated by pumping test.

MRS transmissivities remained mostly close to those estimated from pumping tests (Table 1; Fig. 2). The bulk uncertainty on MRS transmissivity was computed considering both the uncertainty on the MRS water content (θ_{MRS}), and on the relaxation time T_1 . The resulting uncertainty lied between -60 and +100% (Fig. 2). This range remains close to the mean uncertainty on transmissivity values obtained by pumping tests.

MRS water content and specific yield

The MRS water content appears positively correlated with the specific yield; sites with higher specific yield also present higher MRS water content (Fig. 3; coefficient of correlation is 96%). A second result is that MRS water contents are always higher than specific yield values.

The specific yield was first estimated from MRS water content using Eq. (4). The best fit, obtained with $C_y = 0.38$, achieves a residual error of 60% (Fig. 3). The higher misfits are observed for the lowest values of specific yield for which the difference can reach 400%. The main drawback of Eq. (4) is that the computed differences between the MRS water content and the specific yield increase for higher values, whereas the difference between the total porosity and S_y is known to decrease for higher values of S_y (Fetter 2001).

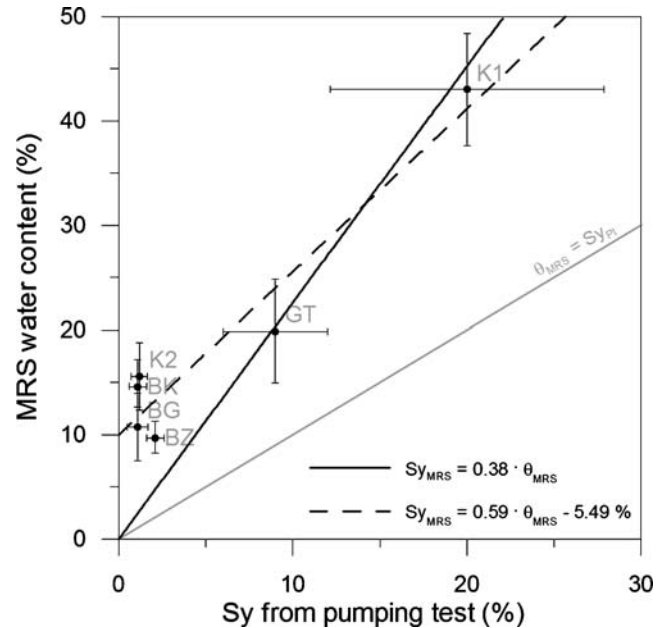


Fig. 3 MRS water content in the saturated part of the aquifer as a function of the specific yield obtained by pumping tests. *Data labels* refer to site names as reported in Table 1

In order to improve the fit, the conversion equation was modified in:

$$S_{y_{MRS}} = C_y' \cdot \theta_{MRS} - \zeta \quad (6)$$

where C_y' and ζ are two parametric factors. Using this equation, the residual error is 30%, with the best parametric factors ($C_y' = 0.59$ and $\zeta = 5.49$). However, this relationship provides negative values of $S_{y_{MRS}}$ for 5 out of the 23 surveyed sites, where the MRS water content is lower than 10% (Table 1). As the MRS water content ranges between S_y and the total porosity, a linear relationship between $S_{y_{MRS}}$ and θ_{MRS} can not fit the data; a more conceptual approach should be considered.

Discussion

Reliability of the parametric factor C_p

Fourteen values of transmissivity were used for computing the averaged parametric factor C_p ; this represents one of the largest dataset for an unconfined aquifer (Vouillamoz et al. 2007a). The C_p values ranged from 2.5×10^{-9} up to 3.3×10^{-8} m/s/ms² with an average value of 1.4×10^{-8} m/s/ms² (Fig. 4). With a single point of calibration, the uncertainty on the C_p value was between -80 and +130%. With an increasing number of calibration data points, the uncertainty on the C_p value decreased (for instance, with three points of calibration, uncertainty lies between -70 and +90%). All combinations with at least seven points of calibration would provide an acceptable C_p value differing by less than 50% from the average C_p value (Fig. 4). Given the possible range of C_p values reported in the literature (from $\sim 10^{-10}$ up to $\sim 10^{-6}$ m/s/ms²; Vouillamoz et al. 2007a), a single point of

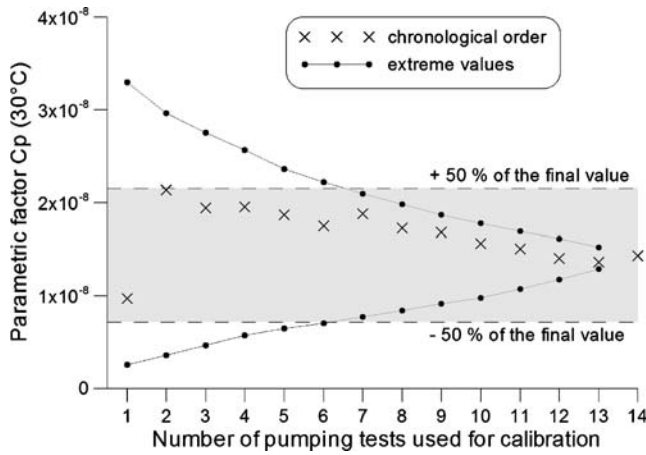


Fig. 4 Change of the computed parametric factor Cp as a function of the number of calibration data. The range of acceptable values stands for a deviation of ±50% around the mean value

calibration was shown to significantly improve the estimate of T_{MRS} compared to the use of the Cp value proposed by default in inversion software.

A complementary parametric function for determining T_{MRS} was suggested by Plata and Rubio (2008):

$$Cp = m F^{-n} \tag{7}$$

where Cp is the parametric factor of Eq. (2), $F = \theta_{MRS} T_1^2 \Delta z$, and where m and n are fixed values determined by a “universal” (i.e. valid for all aquifers) function. The computed values of the parametric factor Cp were reported as a function of F ; a negative trend was observed, with no clear grouping of data as a function of MRS site characteristics (Fig. 5). The coefficients m and n proposed as universal by Plata and Rubio (2008; i.e. $m = 4.05 \times 10^{-7}$ and $n = 0.4186$ for high values of F) did not allow for data fitting. Consequently, these coefficients probably depend also on the geological context. In the case of the CT aquifer, the adjusted

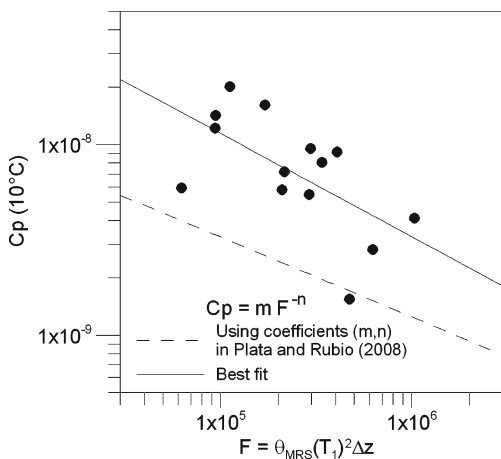


Fig. 5 Parametric factor Cp as a function of factor $F = \theta_{MRS}(T_1)^2 \Delta z$ (After Plata and Rubio 2008)

coefficients ($m = 5.89 \times 10^{-6}$ and $n=0.5423$) were used to compute T_{MRS} from Eqs. (2) and (7). Results showed only a slight improvement when compared with the use of a constant Cp value (residual mean square between T_{MRS} and T_{Pt} of 75% instead of of 86%). This improvement is poorly significant considering the requirement of two additional parametric factors (m and n).

Calibration of the specific yield

A review of the expected relationship between the MRS water content (θ_{MRS}) and the specific yield (Sy) was proposed by Lubczynski and Roy (2007). For an unconfined aquifer, key remarks are as follows: (1), θ_{MRS} should be higher than Sy, and (2), θ_{MRS} could be considered as roughly equal to the effective porosity (n_e). However, in situ measurements at site scales were lacking to validate these assumptions.

Results of the MRS surveys in SW Niger provided confirmation of the first of these two hypotheses. In addition, it was shown that θ_{MRS} for silty to fine sand formations was in a range from 5 up to 25%, whereas coarser sands from the Quaternary valleys showed θ_{MRS} of up to 43% (Table 1). It was therefore assumed that for sandstone aquifers, θ_{MRS} should increase with grain size.

The general relationship between the median grain size and storativity parameters indicate that both the total porosity (n) and the specific retention (Sr) decrease when grains become coarser (e.g. De Marsily 1986; Stephens et al. 1998). For the smallest grain sizes, the specific retention changes more rapidly than the total porosity, and the specific yield reaches a maximum for intermediate grain size (Fig. 6a). At the scale of a grain, the relationship between the specific yield and the grain size is due to the fact that the amount of bound water is directly linked to the specific surface, while the amount of capillary water is linked to the pore size distribution. Both specific surface and pore size distribution depend on the median grain size. However, as outlined by De Marsily (1986), this relationship is a simplified view and should not be applied when estimating the specific yield from the median grain size. An illustration of this statement is provided in Figure 6b; although the shape of this relationship is unique, the value of the specific yield for a given median grain size may change by more than 20% depending on the considered relationship. Similar conclusions can be drawn for the total porosity, with discrepancies of up to 15% for a given range of grain size.

The proposed relationship between the MRS water content and the specific yield is based on the assumption that the specific retention (Sr), and the amount of water undetectable by MRS ($\theta_u = n - \theta_{MRS}$) follow a similar pattern as a function of grain size. This hypothesis is built on the fact that the amount of undetectable water depends first on the mean relaxation time T_2^* , which itself depends on the pore size; the standard deviation of T_2^* could also have a minor influence on the amount of undetectable water, but this was not taken into account. In order to get a pattern that follows the relationships available in literature (cf. Fig. 6), a model of Sr and θ_u changes as a function of

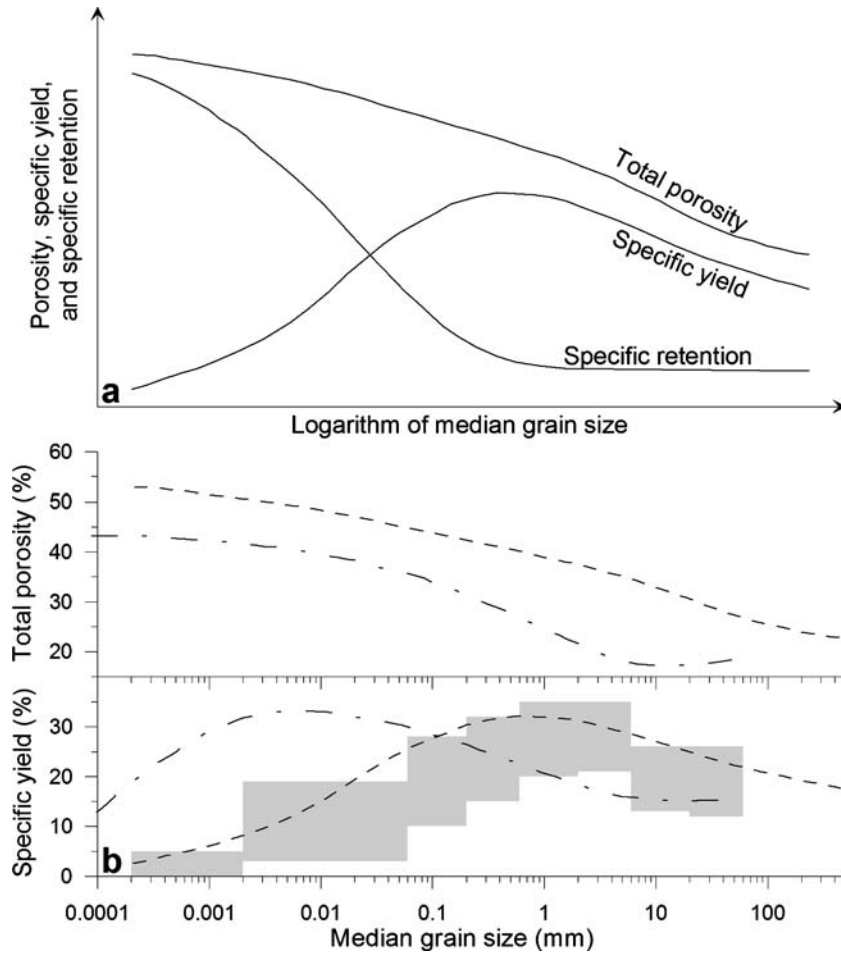


Fig. 6 Total porosity, specific retention, and specific yield versus the median grain size of the aquifer matrix. **a** General shape of the relationships; **b** Different graphs of the total porosity and of the specific yield as a function of grain size: Grey blocks represent data from Johnson (1967; in Fetter 2001); the dashed-dotted lines represent data from De Marsily (1986); the dashed lines represent data from Stephens et al. (1998)

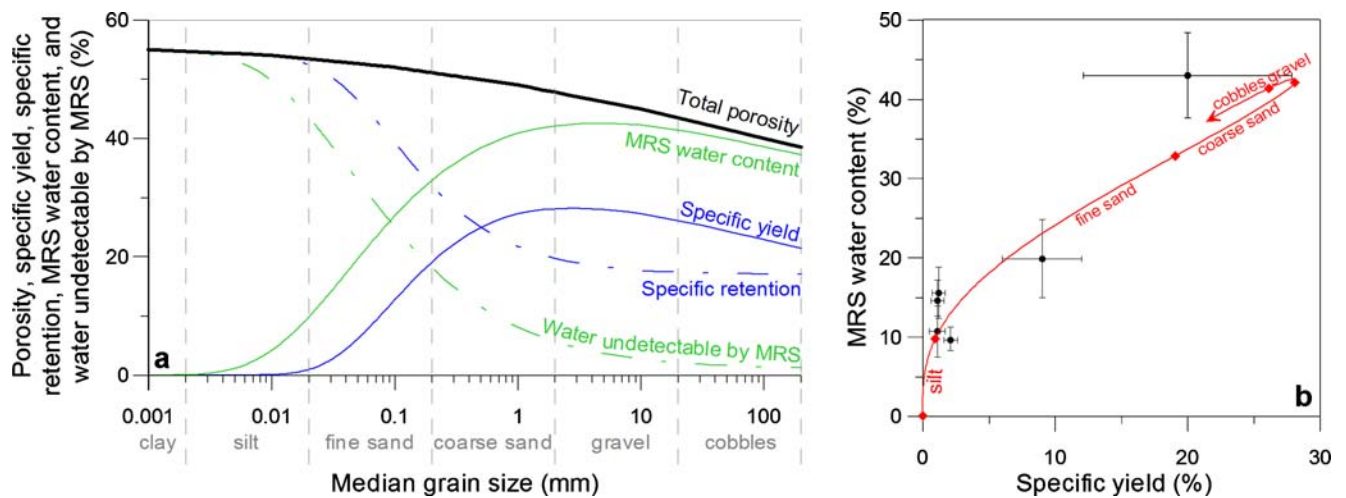


Fig. 7 Calibration of the MRS water content in the Continental Terminal unconfined aquifer. **a** Model of total porosity, specific yield, specific retention (Eq. 8), MRS water content and water undetectable by MRS (Eq. 9) as a function of grain size. **b** Corresponding diagram of the MRS water content as a function of the specific yield; the proposed relationship between the MRS water content and the specific yield is based upon the model of changes in the MRS water content and the specific yield as a function of the median grain size reported in Fig. 7a

grain size was built (Fig. 7a) using the following equations:

$$Sr(r) = (n(r) - Sr_{\min}) \cdot [1 - \exp(-A_1 \cdot r^{-B_1})] + Sr_{\min} \quad (8)$$

$$\theta_u(r) = (n(r) - \theta_{u_{\min}}) \cdot [1 - \exp(-A_2 \cdot r^{-B_2})] + \theta_{u_{\min}} \quad (9)$$

where r is the median grain size, $A_{1,2}$ and $B_{1,2}$ are parametric factors; and Sr_{\min} and $\theta_{u_{\min}}$ are the lower values of, respectively, the specific retention and the amount of water undetectable by MRS (corresponding to the coarsest grain sizes). The total porosity was arbitrary chosen at a maximum of 55% for grain sizes of 0.001 mm, and then linearly decreased by 1–5% per log cycle (Fig. 7a).

Parametric factors $A_{1,2}$, $B_{1,2}$, Sr_{\min} and $\theta_{u_{\min}}$ were adjusted considering the following criteria: (1) the model should reasonably fit field data (Fig. 7b); (2) MRS is unable to detect water in clay (Legchenko et al. 2002); (3) granulometry of the studied aquifer corresponds mainly to fine sand (Massuel et al., 2006). Considering these conditions, the best set of parametric factor was found to be (cf. Fig. 7): $A_1=0.16$; $B_1=0.8$; $Sr_{\min}=17\%$; $A_2=0.16$; $B_2=0.6$; and $\theta_{u_{\min}}=1\%$. The resulting relationship between the specific yield and the MRS water content is clearly non-linear; the fit with the data showed a residual error of 60% (Fig. 7b). This mean residual error is similar to those obtained with linear equations—Eqs. (4) and (6). Although this conceptual model calls for more parameters, it is more satisfactory, because errors are more scattered and do not concern only the lowest specific yields; in addition, the non-linear behaviour is well reproduced for lower values. Compared to Eq. (4), this conceptual model also shows the expected lack of increase in the difference between the MRS water content and the specific yield for higher values (Fig. 3). The main misfit is observed for the

highest Sy value, obtained in a slightly different context of coarse Quaternary sands (Table 1).

Limitations of the calibration model

The proposed calibration model is limited in its application by several key assumptions. First, the model assumes that the specific yield mainly depends on the median grain size, which may depend on the geological context. For instance, unconnected and dead-end porosity can be significant in volume in fractured aquifers, depending on the opening of fissures; in porous media, the specific retention mainly depends on the median grain size, but depends also on grain sorting (Stephens et al. 1998). Secondly, the water undetectable by MRS is assumed to be influenced mainly by mean pore size, via the relaxation time T_2^* . As T_2^* is also dependent on the magnetic properties of rocks, the calibration procedure should also consider this factor.

The validity of these two assumptions was examined for the study area. The storage of grains could be estimated through grading curves, which usually indicate relatively well-sorted sands (Boeckh 1965); however, the clay fraction can significantly vary with depth, with local values up to 40% (Massuel et al. 2006). The Quaternary alluvial sediments are known for being better sorted than the CT sandstones (Pallas 1970). This could partly explain the difficulty in fitting the highest Sy value (Fig. 7b).

The possible influence of the magnetic properties was estimated by comparing T_1 and T_2^* time parameters obtained by MRS. The relaxation time T_2^* was shown to be positively correlated (coefficient of correlation of 64%) with the relaxation time T_1 , which itself is poorly influenced by magnetic properties (Legchenko et al. 2002). The relaxation time T_2^* , and thus the water content undetectable by MRS, were therefore considered as mostly dependent on the pore size. As the influence of the magnetic properties of rocks decreases when the distance between water molecule and solid grain increases, smaller pores are more affected by magnetic effects. This could partly explain the observed variability

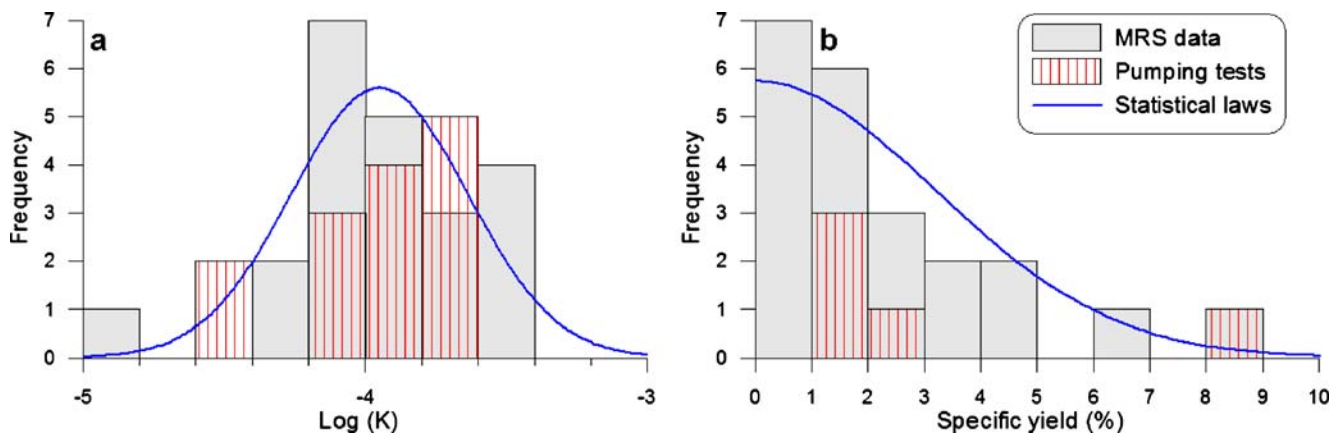


Fig. 8 Statistical distribution for the CT aquifer of a permeability values fitted by a lognormal law and b specific yield values fitted by one-sided Gaussian distribution

of the MRS water content for the lowest specific yield values (Fig. 7b).

Aid for aquifer parameterization

The permeability (K) was computed for each site surveyed by MRS, considering T_{MRS} and the estimated thickness of the aquifer. In addition, the proposed calibration model (Fig. 7b) allowed for the estimation of the specific yield for each site investigated with MRS (Table 1). For the whole CT aquifer, only 14 values of K and 5 values of S_y were obtained by pumping tests, whereas 22 values were available by MRS. The K and S_y values estimated by MRS therefore allowed a much more robust statistical analysis of their distribution within the aquifer. Histograms of K and S_y estimated by MRS are displayed in Fig. 8. The distribution of permeability values followed a lognormal law (Fig. 8a), in good agreement with the statistical distribution commonly considered in groundwater modelling (e.g. Journel et al. 1998; Deutsch 1999). Specific yield values followed a one-sided Gaussian distribution (Fig. 8b). These statistical distributions appear as useful pieces of information for better constraining groundwater simulation results.

Conclusion

The specific yield was estimated by pumping tests at six different sites. The MRS water content appeared to be systematically higher than the specific yield, both values being positively correlated. A conceptual model, based on the empirical relationship between grain-size distribution, total porosity, specific yield and specific retention was proposed. The observed fit between the MRS and pumping test data was reasonably satisfactory, considering the natural geological conditions of the experimental dataset (residual error of 60%). In order to better constrain the relationship between the specific yield with MRS outputs parameters, validation on samples in controlled condition would be relevant.

The large dataset (14 values) obtained by MRS and pumping tests data in the unconfined Continental Terminal aquifer in Niger allowed for the checking of the efficiency of the standard relationship commonly used for estimating aquifer transmissivity using MRS. An increase in the number of calibration data was shown to improve the accuracy of the parametric factor C_p , which depends on the aquifer characteristics. A single point of calibration allows for the estimation of the parametric factor with an uncertainty $\leq 150\%$, whereas with seven points or more, uncertainty on the final value is $< 50\%$. When interpreting MRS in terms of aquifer transmissivity, a mean discrepancy of 150% from the transmissivity value estimated by the pumping test was achieved.

MRS appears as a promising non-invasive method for obtaining a large dataset of transmissivity and specific yield parameters for shallow unconfined aquifers. For groundwater modelling, a consistent statistical distribution of T and S_y

parameters could be obtained at the aquifer-scale at low cost compared to other field-scale methods (e.g. pumping tests).

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References

- Agarwal RG (1980) A new method to account for producing time effects when drawdown type curves are used to analyze pressure buildup and other test data. Proceedings of the 55th Annual Fall Technical Conference and Exhibition of the Society of Petroleum Engineers. Paper SPE 9289, SPE, Richardson, TX
- Boeckh E (1965) Contribution à l'étude hydrogéologique de la zone sédentaire de la république du Niger [Contribution to hydrogeological knowledge of the sedentary zone of the Niger Republic]. Ministère des Travaux publics et des Mines de la république du Niger. Rapport technique DAK 64-A 20, BRGM/BFBH, Dakar, Sénégal
- Boucher M (2007) Estimation des propriétés hydrodynamiques des aquifères par Résonance Magnétique des Protons dans différents contextes géologiques, de l'échantillon à l'échelle hydrogéologique [Estimation of hydrodynamics properties of aquifers with magnetic resonance soundings in different geological contexts from sample to hydrogeological scale]. PhD Thesis, University of Orléans, France
- Chalikakis K, Nielsen MR, Legchenko A (2008) MRS applicability for a study of glacial sedimentary aquifers in Central Jutland, Denmark. *J Appl Geophys* 66:176–187
- Cooper HH, Jacob CE (1946) A generalized graphical method for evaluating formation constants and summarizing well-field history. *Trans Am Geophys Union* 27:526–534
- De Marsily G (1986) Quantitative hydrogeology. Academic Press, London
- De Marsily G, Delhomme JP, Delay F, Buoro A (1999) 40 years of inverse problems in hydrogeology. *C R Acad Sci Paris II* 329:73–87
- Deutsch CV (1999) Reservoir modeling with publicly available software. *Comput Geosci* 25:355–363
- Ezzy TR, Cox ME, O'Rourke AJ, Huftile GJ (2006) Groundwater flow modelling within a coastal alluvial plain setting using a high-resolution hydrofacies approach: Bells Creek plain, Australia. *Hydrogeol J* 14:675–688
- Favreau G (2000) Caractérisation et modélisation d'une nappe phréatique en hausse au Sahel: dynamique et géochimie de la dépression piézométrique naturelle du kori de Dantiandou (sud-ouest du Niger) [Characterization and modelling of a rising water table in the Sahel: dynamic and geochemistry of the Dantiandou kori natural hollow aquifer (southwest Niger)]. PhD Thesis, University of Paris-XI, France
- Favreau G, Cappelaere B, Massuel S, Leblanc M, Boucher M, Boulain N, Leduc C (2009) Land clearing, climate variability and water resources increase in semiarid southwest Niger: a review. *Water Resour Res* 45:W00A16. doi:10.1029/2007WR006785
- Fetter CW (2001) Applied hydrogeology. Prentice Hall, Upper Saddle River, NJ, USA
- Girard JF, Legchenko A, Boucher M (2005) Stability of MRS signal and estimation of data quality. *Near Surf Geophys* 3:187–194
- Healy RW, Cook PG (2002) Using groundwater levels to estimate recharge. *Hydrogeol J* 10:91–109

- Johnson AI (1967) Specific yield: compilation of specific yields for various materials. US Geol Surv Water Suppl Pap 1662-D
- Journel AG, Gunderso R, Gringarten E, Yao T (1998) Stochastic modelling of a fluvial reservoir: a comparative review of algorithms. *J Pet Sci Eng* 21:95–121
- Kollet SJ, Zlotnik VA (2005) Influence of aquifer heterogeneity and return flow on pumping test data interpretation. *J Hydrol* 300: 267–285
- Lang J, Kogbe C, Alidou S, Alzouma KA, Bellion G, Dubois D, Durand A, Guiraud R, Houessou A, De Klasz I, Romann E, Salard-Chebodaeff M, Trichet J (1990) The Continental Terminal in West Africa. *J Afr Earth Sci* 10:79–99
- Legchenko A, Baltassat JM, Beauce A, Bernard J (2002) Nuclear resonance as a geophysical tool for hydrogeologists. *J Appl Geophys* 50:21–46
- Legchenko A, Baltassat JM, Bobachev A, Martin C, Robain H, Vouillamoz JM (2004) Magnetic resonance sounding applied to aquifer characterization. *Ground Water* 42:363–373
- Lubczynski M, Roy J (2005) MRS contribution to hydrogeological system parameterization. *Near Surf Geophys* 3:131–139
- Lubczynski M, Roy J (2007) Use of MRS for hydrogeological system parameterization and modeling. *Bol Geol Miner* 118:509–530
- Massuel S, Favreau G, Descloitres M, Le Troquer Y, Albouy Y, Cappelaere B (2006) Deep infiltration through a sandy alluvial fan in semiarid Niger inferred from electrical conductivity survey, vadose zone chemistry and hydrological modelling. *Catena* 67:105–118
- Meier PM, Carrera J, Sánchez-Vila X (1998) An evaluation of Jacob's method for the interpretation of pumping tests in heterogeneous formations. *Water Resour Res* 34:1011–1025
- Pallas P (1970) Etude en vue de la mise en valeur du Dallol Maouri, Niger [Research for developing the Dallol Maouri, Niger]. Les eaux souterraines. Technical report, AGS: SF/NER 8, FAO, Rome, 162 pp
- Papadopoulos IS, Cooper HHJ (1967) Drawdown in a well of large diameter. *Water Resour Res* 3:241–244
- Plata J, Rubio F (2008) The use of MRS in the determination of hydraulic transmissivity: the case of alluvial aquifers. *J Appl Geophys* 66:128–139
- Roy J, Lubczynski M (2003) The magnetic resonance sounding technique and its use for groundwater investigations. *Hydrogeol J* 11:455–465
- Singha K, Day-Lewis FD, Moysey S (2007) Accounting for tomographic resolution in estimating hydrologic properties from geophysical data. *AGU Geophys Monogr Ser* 171:227–241
- Stephens DB, Hsu KC, Priksat MA, Ankeny MD, Blandford N, Roth TL, Kesley JA, Whitworth JR (1998) A comparison of estimated and calculated effective porosity. *Hydrogeol J* 317:156–165
- Tartakovsky GD, Neuman SP (2007) Three-dimensional saturated-unsaturated flow with axial symmetry to a partially penetrating well in a compressible unconfined aquifer. *Water Resour Res* 43, W01410. doi:10.1029/2006WR005153
- Theis CV (1935) The relation between the lowering of piezometric surface and the rate and duration of discharge of a well using groundwater storage. *Trans Am Geophys Union* 2:519–524
- Vouillamoz JM, Descloitres M, Toe G, Legchenko A (2005) Characterization of crystalline basement aquifers with MRS: comparison with boreholes and pumping tests data in Burkina Faso. *Near Surf Geophys* 3:193–201
- Vouillamoz JM, Baltassat JM, Girard JF, Plata J, Legchenko A (2007a) Hydrogeological experience in the use of MRS. *Bol Geol Miner* 118:531–550
- Vouillamoz JM, Chatenoux B, Mathieu F, Baltassat JM, Legchenko A (2007b) Efficiency of joint use of MRS and VES to characterize coastal Aquifer in Myanmar. *J Appl Geophys* 61:142–154
- Vouillamoz JM, Favreau G, Massuel S, Boucher M, Nazoumou Y, Legchenko A (2008) Contribution of magnetic resonance sounding to aquifer characterization and recharge estimate in semiarid Niger. *J Appl Geophys* 64:99–108