The links between MRS parameters and the hydrogeological parameters

Patrick Lachassagne1*, Jean-Michel Baltassat2, Anatoly Legchenko3 and Hubert Machard de Gramont4

1 BRGM, Water Division, 1039, rue de Pinville, 34000 Montpellier, France
2 BRGM, Development Planning and Natural Risks Division, BP 6009, 45060 Orléans Cedex 2, France
3 IRD, 32, Avenue Henri Varagnat, 93143, Bondy Cedex, France
4 BRGM, Water Division, BP 6009, 45060 Orléans Cedex 2, France

Received June 2004, revision accepted May 2005

ABSTRACT

Magnetic resonance sounding (MRS) is a non-invasive geophysical method that can be used to discriminate between the geophysical signals of pore water of the surrounding rock. This method is thus particularly attractive for hydrogeological applications.

In order to investigate MRS for the benefit of applied hydrogeology, a comprehensive interdisciplinary study was carried out. The aim of this project was to develop an optimal methodology which could be easily adopted by hydrogeologists. This paper summarizes the main results obtained, with the objective of translating the MRS parameters into parameters that can be directly used by hydrogeologists.

It was found that MRS is well adapted to the working scales of the hydrogeologist (i.e. field scale, well scale). It is able to provide significant information for the hydrogeologist, including:

- Reliable detection of the presence of water in the subsurface. This is the basic and most significant advantage of the method, which proved to be useful, particularly in arid to semi-arid environments.
- Locating water-saturated formations (top and bottom), situated at depths between 0 and 100 m approximately.
- Estimation of the hydrodynamic parameters of detected aquifers:
  - Through a rigorous inversion and calibration process, the MRS method enables the quantification of the specific yield (effective porosity) of aquifers and of the storage coefficient, the latter in unconfined aquifers only.
  - In the present state-of-the-art, MRS is able to estimate the hydraulic conductivity (or transmissivity) of aquifers in localized favourable configurations, when calibration is available; otherwise, aquifers can be compared qualitatively.

Thus the MRS method provides data that cannot be obtained with other non-invasive geophysical tools, and is already a valuable tool for applied hydrogeological projects.

INTRODUCTION

One of the most recent geophysical techniques, i.e. magnetic resonance sounding (MRS), is based on the application of nuclear magnetic resonance (NMR). Besides being described in the context of well-logging applications (Allen et al. 1997; Kenyon 1997; Dunn et al. 2002), it has recently been developed and used in non-invasive applications (Schirov et al. 1991; Goldman et al. 1994; Beauce et al. 1996; Roy et al. 1998, paper presented at EEGS meeting, Barcelona). The method and the main principles of the measurements have been well described (Weichman et al. 2000; Legchenko et al. 2002; Legchenko and Valla 2002; Lubczynski and Roy 2003, 2004; Roy and Lubiszynski 2003).

At present, two types of geophysical measurements are used in hydrogeology: precise and informative invasive geophysical methods (i.e. mainly well-logging), which are expensive and time consuming, and the less expensive non-invasive surface geophysical methods, which provide ambiguous information with respect to the discrimination between water and the rock medium. The MRS method appears to have a great future as a non-invasive method that can be used to differentiate between the geophysical signals from water and from the surrounding rock. More detailed information about the MRS method can be found in the literature (e.g. Weichman et al. 2000; Yaramanci et al. 2002; Roy and Lubiszynski 2003; Lubczynski and Roy 2004; Legchenko et al.)
The method is still being developed (Legchenko 2004) as there are significant limitations related to the measurements: in electrically conductive environments (Weichman et al. 2000), in regions with a low signal-to-noise ratio (Roy and Lubszynski 2000; Plata and Rubio 2002), in areas with an inhomogeneous magnetic field, e.g. volcanic rocks (Legchenko et al. 2002), etc. Nevertheless, convincing results have been obtained in various geological and hydrogeological contexts (Goldman et al. 1994; Vouillamoz 2003; Vouillamoz et al. 2003; Legchenko et al. 2004; Wyns et al. 2004), thus justifying the use of this methodology for hydrogeological applications.

Within this perspective, a comprehensive interdisciplinary (geophysical, geological and hydrogeological) study, performed by BRGM, was aimed at developing an optimal methodology for applying MRS to hydrogeological studies, and at making this new geophysical method more accessible to hydrogeologists (Lachassagne et al. 2003a,b). This paper summarizes the main results obtained with the objective of translating the MRS parameters into parameters that can be used directly by hydrogeologists.

THE PARAMETERS DERIVED FROM MRS AND THOSE USED BY HYDROGEOLOGISTS

Three main parameters are derived from MRS measurements (see Fig. 1):

- the MRS water content \( w \), which is closely related to the amplitude of the MRS signal (Legchenko and Valla 2002);
- the MRS relaxation times \( T_2^* \) and \( T_1 \) or decay time (Dunn et al. 2002); and
- the geometry of the ‘detected aquifers’: i.e. the depth intervals or ‘layers’ to which \( T_2^* \) and \( T_1 \) are applicable (depths to the top and bottom of the aquifer).

An example of MRS results obtained in France, near a water supply well where pumping tests have been carried out, is shown in Fig. 1. The borehole log shows that the investigated aquifer is composed mainly of medium-to-coarse sand. The MRS inversion results (Fig. 1) indicate that the water content increases between 15 m and 20 m, this depth corresponding to the top of the confined aquifer. The relaxation time corresponding to the 15–30 m depth interval is short, which means that the water in this part of the aquifer is located close to the pore walls (capillary-bound water) and thus the hydraulic conductivity is low. This result is consistent with the geological data that show that the 15–30 m depth interval is a clay and chalk formation. Below, increases in both the relaxation time and MRS hydraulic conductivity for the 30–45 m depth interval correspond well with the aquifer tapped by the well. There is also good agreement between the transmissivity estimated by MRS \( (4 \times 10^{-3} \text{ m}^2/\text{s}) \), and that derived from pumping tests \( (5 \times 10^{-3} \text{ m}^2/\text{s}) \). It was not possible to calibrate the MRS water content as no effective porosity measurements were available for this site.

These MRS parameters can be linked more-or-less to the following hydrogeological parameters that are core parameters in the field of applied hydrogeology:

- The porosity \( \phi \) (dimensionless). Most rocks and soils naturally contain a certain percentage of empty space, which may be occupied by water or other fluids. This property is known as the porosity of a medium (in the present case, aquifers or impermeable or semi-permeable bodies), and it measures the volume of voids or pores, which may or may not be connected, in the medium (Castany and Margat 1977). It is expressed quantitatively as the ratio of the volume of the voids to the total volume of the medium (in practice, to a ‘representative elementary volume’ of this medium). This parameter is thus dimensionless. It is equivalent to the volumetric water content when the medium...
is saturated. Natural media display various kinds of porosity: interstice porosity, fissure porosity, fracture porosity, karstic porosity, etc. The MRS parameter ‘water content’ ($w$) appears to be related to this hydrogeological parameter.

• The permeability (or Darcy’s permeability), which describes the ability of a fluid to flow through a medium, under the influence of a hydraulic head gradient (Castany and Margat 1977). It is expressed quantitatively by the intrinsic permeability (dimensions: L$^2$) and, for water in standard conditions, by the coefficient of conductivity or hydraulic conductivity $K$ (dimensions: L/T). This latter coefficient is the one most commonly used by hydrogeologists. Hydrogeologists also use the transmissivity parameter $T$ (dimensions: L$^2$/T), which is equal to the coefficient of conductivity of an aquifer multiplied by its thickness (measured perpendicularly to the flow direction; in most cases this is along the Z-axis). The MRS relaxation times ($T_2^*$ and $T_1$) provide, to some extent and in combination with $w$, some similarities with the hydraulic conductivity or transmissivity (Legchenko and Shushakov 1998).

• The geometry of the ‘detected aquifers’, and also the geometry of the impervious or semi-pervious layers (respectively aquitards and aquicludes), that MRS is used to characterize, is also of primary interest to hydrogeologists. Since they are products of the inversion of MRS measurements, these parameters require calibration before they can be used. In addition, we see below that MRS is also able to provide valuable complementary information for hydrogeologists.

**POROSITY – STORATIVITY**

Below the surface, rocks comprise both a saturated zone and a vadose zone. The saturated zone is located below the piezometric level, where water completely fills the pores of the host rock. This water is composed of free (or flowing) water, including water in dead-end and non-connected pores, on one hand, and fixed (or bound) water on the other hand (Fig. 2). The saturated zone is overlain by a vadose, or unsaturated, zone, where the porosity contains both water (mainly fixed water; Fig. 2) and soil gases. Lubczynski and Roy (2003) proposed a complete description of the different types of ‘water’ present in soils and aquifers. In the saturated zone, the specific yield ($n_e$) quantifies the amount of water stored within the aquifer which is free to flow (including water in unconnected and dead-end pores). The term ‘effective porosity’ is often used as a synonym of specific yield. The two concepts are very similar, but the effective porosity is a ratio of speeds whereas the specific yield (or drainage porosity, de Marsily 1981) is a ratio of volumes: the effective porosity is equal to the ratio of the mean (flowing) water velocity in a permeable medium (measured experimentally by a tracer test, for instance) to the Darcy velocity.

The free water flows independently of the molecular forces of attraction of the solid particles and can be displaced by gravity or pressure gradients. This water can be extracted from the aquifer by pumping, i.e. under gravific drainage conditions. The specific yield parameter is of primary interest for the hydrogeologist, both for quantitative applications (prediction of the yield of a well in transient conditions, for instance) and qualitative applications (computation of the velocity of a pollutant transfer in an aquifer). Specific yield is always less than total porosity (Fig. 2), the difference between the two being mainly linked to the pore-size distribution, which is the main factor determining the particle surface area. It is also linked to mineral type.

---

![Figure 2](https://example.com/figure2.png)

*Groundwater storage concept (from Lubczynski and Roy 2003).*
Unconfined aquifers (Fig. 3, left) comprise both a vadose zone and a saturated zone, while confined aquifers (Fig. 3, right) comprise only a saturated zone. Thus, in the saturated zone of the subsurface, \( w \) can be related to the specific yield of the medium, if dead-end and unconnected porosity can be neglected, which is often the case in aquifers, and particularly in unconsolidated sediments. Indeed, it is assumed that the measured MRS signal after the 40 ms dead-time of the currently used instruments has a decay rate corresponding to free water. Although signals with a relaxation rate faster than 40 ms can be detected, it is widely and arbitrarily assumed that with the available instrumentation only free water is measured.

In fact, depending on the rock type, MRS measurements may either overestimate the specific yield by integrating both free, unconnected and dead-end water and, partly, fixed water, or underestimate the specific yield if losing fixed, but also partly free water, which is the case where some free water would be characterized by decay times below 40 ms. These uncertainties can be corrected by calibrations carried out for each geological formation.

Nevertheless, even if the theoretical reasons for this are not yet clearly understood and require further investigation, the results from numerous MRS measurements show that in many cases, i.e. porous aquifers (Legchenko et al. 2002), karstic aquifers (Vouillamoz et al. 2003) and hard-rock aquifers (Wyns et al. 2004), the difference between measured \( w \) and specific yield is less than the uncertainty in this last parameter (or the lack of much porosity data, if not a total lack of data). The MRS tendency to overestimate specific yield seems particularly important in chalk aquifers where it appears that MRS probably also detects fixed water stored in the primary porosity of the rock.

When dealing with porosity, hydrogeologists often use the concept of storativity or the storage coefficient of a medium (\( S \); dimensionless). The storage coefficient of an aquifer is equal to the ratio of the volume of water produced (or stored) by an unit surface of the considered aquifer, to the corresponding variation in hydraulic head, without time reference (or for an unlimited duration) (Castany and Margat 1977). It is important to stress the fact that this coefficient refers to the total thickness (\( e \)) of the aquifer.

As a consequence, in unconfined aquifers (Fig. 3) \( w \) may also characterize, with the limitations described above, the storage coefficient of the aquifer, in which case storativity is equal to specific yield.

In confined aquifers (Fig. 3), storativity is mainly linked to the compressibility of both the hostrock and water, therefore \( w \) does not characterize the storativity of the aquifer.

Thus, as specific yield is a relatively expensive parameter to acquire at field scale (requiring at least a pumping test with two boreholes (a well and a piezometer) or the completion of tracer tests), MRS appears to be a valuable method for hydrogeologists, particularly when dealing with pollutant transport.

**HYDRAULIC HEAD**

The hydraulic head (\( h \); dimensions: L) is the altitude of a piezometric level, compared to a reference plane (Castany and Margat 1977). It corresponds to the sum of the hydrostatic and the hydrodynamic heads. It is a measure of the (groundwater) fluid potential to which it is proportional. It thus measures the available energy of the groundwater to flow within the aquifer.

In unconfined aquifers, MRS measurements can provide data on the hydraulic head within the aquifer as the piezometric level merges with the top of the saturated zone of the aquifer (Fig. 3), which is detected by MRS.

However, the MRS technique also measures signals from water in the capillary fringe within the aquifer (cf. Fig. 2) and can thus underestimate the depth of the piezometric level. Capillary fringes can be particularly thick (over 1 m) in porous fine-grained aquifers. In chalk, the thickness of the capillary fringe can reach some 10 m within the matrix of the aquifer (Price 1993). It could be one of the explanations for the fact that an MRS signal is also often measured within the vadose zone of chalk aquifers (see previous section).

In addition, small perched aquifers, not measured with the commonly available piezometers but that can be revealed by careful observations during drilling for example, can be detected by MRS measurements, providing their size is sufficiently large.
compared to that of the MRS antenna loop. As they are not connected to the main aquifer, their piezometric level must not be taken in account for the details of a general piezometric map of the main aquifer.

In the case where groundwater shows a significant vertical flow component (Fig. 4), the water level measured in a well can vary considerably (from a few centimetres to decimetres, and locally to a few metres) from the depth to the top of the saturated zone of the aquifer. This difference depends on the vertical hydraulic gradient, thus on the location of the observation well along a flow line, but also on the depth and length of the piezometer screen. Thus, the piezometric level as deduced from MRS measurements is only equal to the hydraulic head that would be measured at the aquifer top, and not to the hydraulic head that can be measured within the aquifer.

In confined aquifers, MRS measurements cannot provide information about the hydraulic head. The increase in water content measured by the MRS log corresponds only to water-saturated rocks. For example, in the case where a sandy aquifer overlain by clay is located at a depth of between 20 and 30 m and the piezometric level is at 5 m, then MRS will only locate the top and bottom of the aquifer, between 20 and 30 m.

Therefore, MRS does not measure any characteristics related to piezometric level or hydraulic head, but this kind of information can be indirectly inferred from MRS data in the case of unconfined aquifers, if some care is taken.

**GEOMETRY OF THE AQUIFER**

Apart from the above limitations in both confined and unconfined aquifers, the top and the bottom of an aquifer, and thus its thickness (Figs 3 and 4), can be determined though MRS measurements.

As the vertical resolution of MRS is limited (it is known that the vertical resolution of MRS is dependent on the loop size and, in general, decreases with depth (Legchenko and Shushakov 1998), the available inversion software provides a better accuracy for the shallowest aquifer than for multi-aquifer systems. Deeper aquifers can be partly masked by shallower water-saturated layers. The accuracy is also better for the top of an aquifer than for the bottom (Legchenko et al. 2002; Legchenko and Shushakov 1998).

**HYDRAULIC CONDUCTIVITY**

The hydraulic conductivity ($K$) is a vectorial parameter that not only relies on the physical properties of the medium (anisotropy, heterogeneity, etc.), but also on other parameters such as:

- the flow direction;
- the scale of the measurement, depending on the type of hydraulic conductivity measurement (on a sample, in a well, on the scale of large parts of the aquifer through the calibration of an hydrogeological model, etc.).

In the case of a pumping test, the scale of the measurement also depends on the duration of the pumping.

Compared to porosity, $K$ is highly variable. It ranges over several orders of magnitude (de Marsily 1981).

The volume investigated by MRS (often a cylinder, 100 m in diameter and 100 m in depth) is similar to that investigated by a pumping test, which is the most common method used by hydrogeologists to measure the hydraulic conductivity of an aquifer. This scale of investigation is the one most commonly chosen by hydrogeologists, as one of the main purposes of determining the hydraulic conductivity of an aquifer is to estimate the yield that can be exploited from a well, and thus there is no significant difference in scale between the kind of measurement performed and the objective of the measurement. In fact, up-scaling from small-scale measurements is not straightforward and would thus provide very uncertain results.

However, the quest for a relationship between the decay time of the MRS signal and the hydraulic conductivity of the host rock is based mainly on the relationship between the MRS signal relaxation characteristics and the size of the rock pores (Timur 1968). Thus, there is an important difference in scale between the motion of water used for the determination of the hydraulic conductivity of an aquifer from a pumping test and the microscopic motion of nuclei used for the same purpose by MRS. There is also a difference between the two types of physical phenomenon involved. During a pumping test the flow direction is well constrained and organized (radially towards the pumped well, on average); this is not the case of the nuclei motion during the decay time of the MRS signal, which may be considered as Brownian (Sen et al. 1990; Kenyon 1997).

Thus, following the example of NMR logging, attempts have been made to find a correlation between a combination of the MRS parameters (water content and relaxation time, with some constants and exponents to be adjusted) and the hydraulic conductivity (or the transmissivity $T$), as deduced from pumping tests. The formula that has been proposed (Fig. 1) (Timur 1969a,b) is

$$T = k \cdot \Delta z,$$

where $T$ is the transmissivity of the aquifer (dimensions: L²/T) as deduced from MRS measurements, $\Delta z$ is the aquifer thickness.
(dimensions: L), \( k \) is the hydraulic conductivity (dimensions: \( L/T \)) as deduced from MRS measurements, \( w \) is the MRS water content (%), \( T_1 \) is the relaxation time (dimensions: \( T \)) and \( C_p \) is an empirical constant.

The interpretation of NMR geophysical borewell logging relies on a similar correlation (e.g. Timur 1969a,b; Sen et al. 1990; Kenyon 1997). This approach is consistent with the results of the considerable efforts made over previous decades (supported by the petroleum industry in particular), which have shown that only empirical links, to be assessed and calibrated for each kind of geological formation, can be established between porosity (or pore-size distribution) and hydraulic conductivity (de Marsily 1981).

Thus, this search for adjustments between pumping test results and MRS parameters seems to be promising, even if convincing results are not assured, considering the conceptual difficulties involved. A comparison between the MRS and borehole pumping test results is shown in Fig. 5 (Legchenko et al. 2004); the error bars were calculated taking into account the accuracy of the MRS data and possible equivalent solutions. Thus, the hydraulic conductivity of aquifers can be estimated as

\[
K_{\text{MRS}}(z) = C_p w(z) T_1(z),
\]

where \( w \) and \( T_1 \) are the water content (%) and relaxation time (in ms) derived from MRS measurements. For sandy aquifers, as well as for aquifers composed of weathered and fractured rock (granite, chalk, limestone), the hydraulic conductivity can be estimated using the same value of the empirical constant, \( C_p = 7.0 \times 10^{-10} \).

In addition to the attention that must be paid to the MRS data inversion (see below), care is also required when dealing with hydrogeological data. The direct use of hydraulic conductivity/transmissivity values from the literature can lead to significant bias. For existing data, it imposes the systematic re-interpretation of the pumping tests in order to:

(i) check the adequacy between the scale of the MRS measurement and the volume investigated by the pumping test (and thus, if necessary, choose the part of the pumping test curve to be considered);
(ii) build a realistic conceptual geological and hydrogeological model of the studied site;
(iii) check the validity of the required hypothesis (type of porosity, homogeneity, isotropy, location and thickness of the well screen in the aquifer, etc.) and thus choose an appropriate analytical solution of the diffusion equation for the interpretation of the pumping test.

As MRS measurements provide discretized data along the vertical axis, it is also important to acquire geological and hydrogeological data that allow the vertical distribution of the hydrodynamic parameters to be determined.

**IMPORTANCE OF THE CALIBRATION**

As with most geophysical methods, MRS is subject to the principle of equivalence. The interpretation of MRS data requires the simultaneous analysis of all MRS parameters (water content, relaxation time, thickness of the different layers, etc.). In the determination of hydrogeological parameters from MRS measurements, a careful calibration process is essential. The knowledge of one parameter, or even better two parameters, makes it possible to determine the second and third parameters, or the third parameter, with a much higher accuracy.

Thus the MRS inversion process must comprise, after a careful analysis of the geophysical, geological and hydrogeological data, at least the following three steps:

1. (automatic) inversion of the data;
2. calibration of MRS parameters, on the basis of existing data, when available, or on the experience of the team of geologists/hydrogeologists/geophysicists;
3. estimation of the unknown parameter(s).

These three steps can include certain iterative procedures. For instance, step 2 would allow the precise identification of the depth of the top of a confined aquifer, or the piezometric level in an unconfined aquifer, whereas step 1 would only provide a progressive variation along \( Z \) of the medium properties.

We consider that the most valuable and significant contribution of the MRS method (in terms of added value/cost ratio) is the estimation of the hydrodynamic parameters (porosity, hydraulic conductivity) of aquifers or impervious or semi-pervious layers. Thus, it is important:

(i) in the second step, to use for calibration the parameter that is easiest (and thus cheapest) to acquire, i.e. the thickness of the various ‘layers’. This means that knowledge of a good geological and hydrogeological conceptual model of the studied site would be a definite advantage.

(ii) Then to estimate the hydrodynamic parameters (specific yield, hydraulic conductivity, transmissivity) on the basis of empirical relationships linking them with the MRS data.
parameters, established for the studied site or for similar hydrogeological contexts. Further R&D programs must then focus on such compilations.

CONCLUSIONS

The MRS method is able to provide useful data for hydrogeologists, including:

- Reliable direct detection, with few ambiguities, of the presence of free water in the subsurface (aquifers). This is the basic and most important advantage of this geophysical method, which could thus prove extremely useful, particularly in arid to semi-arid areas.

- The location of saturated formations (top and bottom), situated at depths between 0 and 100 m, approximately.

- Estimation of hydrodynamic parameters of detected aquifers when calibration is available; otherwise, aquifers within the same geological setting can be compared qualitatively.

MRS allows the quantification of the specific yield of aquifers through a rigorous inversion and calibration process. Various experiences have demonstrated that in localized favorable configurations, the MRS method also enables the estimation of aquifer hydraulic conductivity.

Thus the MRS method provides data that cannot be obtained through other non-invasive geophysical tools. In addition, it is well adapted to the working scale used in hydrogeology (i.e. field scale, well scale).

Nevertheless, the MRS method still has some limitations: it cannot be applied to magnetic rocks; it cannot be applied in areas affected by strong man-made and natural electromagnetic noise; the maximum depth of water detection is less than 100 m; the threshold of water detection is 0.5%, which limits its performance in fractured rocks; etc.

REFERENCES


