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Basic theory of the Magnetic Resonance Sounding Method

J. L. Plata[(1)] and F. M. Rubio[(1)]

(1) Instituto Geológico y Minero de España. La Calera 1, 28760 Tres Cantos, Madrid (Spain)
jl.plata@igme.es - fm.rubio@igme.es

ABSTRACT

When an electromagnetic field of one specific frequency equal to the Larmor frequency of hydrogen protons is sent to the underground by a transmitter on the surface part of its energy is absorbed exclusively by the water molecules. When the excitation field is removed, the absorbed energy is released in the form of a new electromagnetic field which can be detected by a receiver at the surface. This response can only be produced by water, and has some identity characteristics: the released energy has the Larmor frequency for hydrogen protons and produces a voltage e(t) which amplitude decays exponentially with time until it vanishes, at a rate or decay time that depends on the mean size of the pores. The maximum voltage amplitude is directly proportional to the amount of water. When the excitation field is increased the signal comes from deeper parts of the subsurface, allowing making a sounding (Magnetic Resonance Sounding, MRS). The principal factors that affect the measured water signal are the depth and thickness of the water bearing layer, the electrical conductivity of the subsurface, the magnitude and inclination of the geomagnetic field, the type of water containing rocks, the size of the antenna and the electromagnetic noise. The general rule is that the maximum amplitude of the signal decreases when increasing the depth of the aquifer. The existence of rocks with high conductivity has a screening effect that lowers the amplitude of the water signal and causes the depth of investigation to be decreased. When magnetic rocks are present, the intensity and gradient of the geomagnetic field is different in different parts of the subsurface what causes the Larmor frequency be also different and hence MRS signal can not be detected. The best conditions are met in resistive-non magnetic environments, high latitudes, and coarse grained rocks. A mathematical model of the physical phenomenon and its measurement allows getting the theoretical signal produced by a defined configuration of the ground water and rocks distribution in the subsurface (modelling or direct problem), as well as deducing the aquifers configuration (distribution of the amount of water and time decay with depth) from the MRS signal actually measured in the field (inversion or inverse problem). Accuracy of the results depends on the signal quality and the hypothesis and assumptions made in the model.

Key words: Magnetic Resonance Sounding, mathematical model, MRS, time decay, water signal

Teoría básica de los Sondeos de Resonancia Magnética

RESUMEN

Cuando se transmite por el subsuelo un campo electromagnético generado en la superficie con una frecuencia específica, igual a la de Larmor de los protones de hidrógeno, parte de su energía es absorbida exclusivamente por las moléculas de agua. Finalizada la excitación, la energía absorbida es devuelta como un nuevo campo electromagnético que puede ser detectado por un receptor en la superficie. Esta respuesta sólo puede deberse a la existencia de agua, y tiene unas características distintivas: su frecuencia es la de Larmor de los protones de hidrógeno, y da lugar a una f.e.m. e(t) cuya amplitud disminuye exponencialmente con el tiempo, con una velocidad de decaimiento que depende del tamaño medio de los poros. La máxima amplitud de esta señal es directamente proporcional a la cantidad de agua. Aumentando la intensidad de la excitación se obtienen señales que provienen de mayor profundidad, lo que permite la realización de un sondeo (Sondeo de Resonancia Magnética, SRM). Los principales factores que afectan a la amplitud de la señal medida son la profundidad y potencia del acuífero, la conductividad eléctrica de las rocas, la amplitud e inclinación del campo magnético terrestre, el tipo de rocas que constituyen el acuífero, el tamaño de la antena utilizada y el ruido electromagnético existente. La norma general es que la máxima amplitud disminuye al aumentar la profundidad del acuífero. La alta conductividad de las rocas actúa como pantalla, disminuyendo la amplitud de la señal y disminuyendo la profundidad de investigación. Las rocas magnéticas provocan una variación espacial de la intensidad y gradiente del campo geomagnético, que da lugar a una variación espacial de la frecuencia de Larmor, lo que impide la medición. Las mejores condiciones se dan en rocas no magnéticas y resistivas, latitudes altas y rocas de grano grueso. A través de la formulación de un modelo matemático del fenómeno físico y de su medición, se puede calcular la señal teórica para un modelo dado del subsuelo (modelado o problema directo), así como deducir la distribución del contenido en agua y tiempo de decaimiento en función de la profundidad a partir de las mediciones SRM realmente efectuadas (inversión o problema inverso). La exactitud de los resultados depende de la calidad de la señal, así como de las hipótesis y simplificaciones establecidas en el modelo teórico utilizado.

Palabras clave: modelo matemático, señal del agua, Sondeos de Resonancia Magnética, SRM, tiempo de decaimiento

The physical phenomenon and its measurement: the e(t), $E_0(q)$ and $T_d(q)$ functions

The Magnetic Resonance Sounding method is not based on a petrophysical property, as the rest of the geophysical methods referred to in Mejias and Plata (2007, this Issue) , but on a physical property of the hydrogen atom: the phenomenon of the Nuclear Magnetic Resonance (NMR) in geomagnetic field. Energy, in form of an electromagnetic field, is sent to the underground by a transmitter on the surface. In its travel through the rocks, part of this energy is absorbed exclusively by the water molecules. When the excitation field is removed, the absorbed energy is released in the form of a new electromagnetic field which can be detected by a receiver at the surface. This response can only be produced by water, and has some identity characteristics.

This physical phenomenon (NMR) is due because atom particles in the presence of the geomagnetic field can absorb energy ($\Delta\varepsilon$) only at a given frequency (f), and at specific amounts, multiple of a physical magnitude named Planck's constant (h):

$$\Delta\varepsilon = 2\Pi fh = \omega h$$

The atom components behave as small magnets (represented as vectors or magnetic moments) with two possible states of different energy, depending on its orientation in respect to the magnetic field of the Earth. In this case, the frequency mentioned before is known as Larmor frequency (f_L), and depends on two magnitudes: the amplitude of the geomagnetic field (B_0) and the gyromagnetic ratio (γ), which is a constant with a different value for each atom particle and element:

$$\omega_L = 2\Pi f_L = \gamma B_0$$

Atoms will be able to absorb energy only if the frequency of the excitation electromagnetic field is equal to the Larmor frequency for these atoms. To excite only the hydrogen protons the Larmor frequency calculated using the value of gyromagnetic ratio for hydrogen protons must be used

$$f_L = \gamma B_0 / 2\Pi = 0.04258\ B_0$$

The absorbed energy changes the orientation of the magnetic moments of these atoms, which originally are oriented in direction of the magnetic field of the Earth. As water is the most important component with hydrogen in the subsurface, the released energy when the excitation stops will come fundamentally from the water in the rocks.

Energy is released because the associated small magnets of hydrogen protons have to recover its previous state of energy, which is more stable, and then a movement of these magnets takes place in the space to become aligned again with the geomagnetic field. The energy is released in form of an electromagnetic field which has the Larmor frequency for hydrogen protons, and then it can be recognized. When the electromagnetic field associated to the protons reorientation reaches the surface, it induces a voltage e(t) measurable with a receiver loop or antenna. Amplitude of this voltage decays exponentially with time until it vanishes. Its maximum value is related to the amount of water, and the decay time depends on the mean size of the pores. The movement for the recovery of the initial orientation of magnets is three dimensional, in such a way that the rate of decay of energy is not the same in all orientations.

The original idea of surface prospecting using NMR in the Earth's magnetic field dates from 1962 (Varian, 1962). Until 1978 this method was not available. An instrument for the measurement of the nuclear magnetic resonance signal, HYDROSCOPE, constructed by the Institute of Chemical Kinetics and Combustion of the Russian Academy of Sciences, ICKC, and the first description of the method, as well as its mathematical inversion theory were developed (Semenov *et al.,* 1982, 1987, 1988, and 1989). In 1995 NUMIS instrument was developed by the French company IRIS in co-operation with the BRGM and ICKC. The selective characteristics of the NMR method allow the new equipments to be sensitive only to the water target. The effect of other minerals on the NMR signal is only indirect, such as screening and relaxation of the signal (Semenov, 1987).

To measure a MRS, the control instrument is connected to a loop of wire or antenna extended on the surface of the ground, and a current of intensity I_0 and frequency f_L

$$i(t) = I_0 \cos(\omega_L t)$$

is passed during a time τ, creating the excitation field. In Figure 1 a diagram for the process is shown.

The measurement sequence is composed of records of the voltage induced by the water signal, the e(t) decay curves, obtained for increasing values of the excitation energy, expressed by the moment

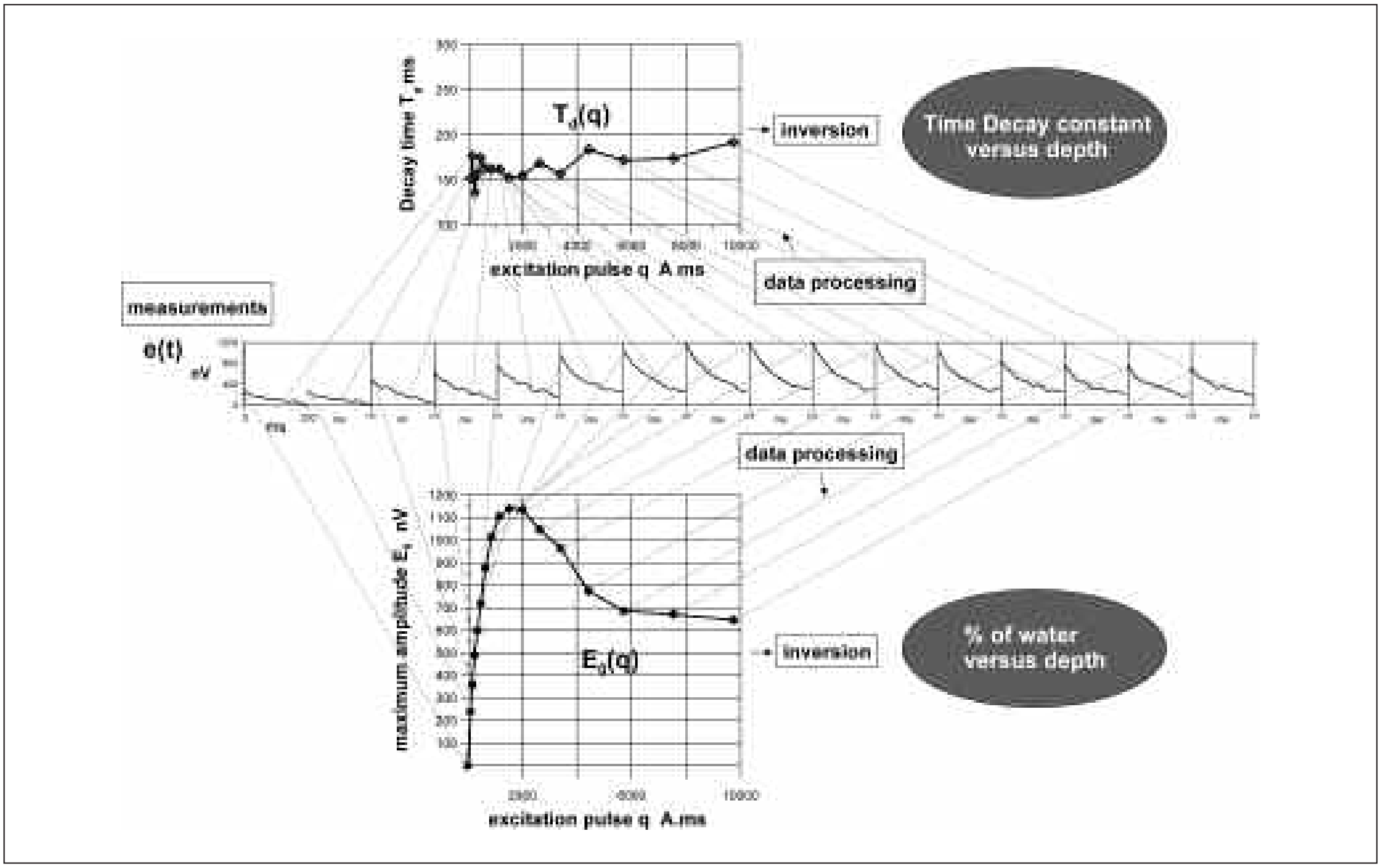

Fig. 1. Scheme of the performance of a Magnetic Resonance Sounding. Real data. (Modified from Plata and Rubio, 2001)
Fig. 1. Esquema de la realización de un Sondeo de Resonancia Magnética. Datos reales. (Modificado de Plata y Rubio, 2001)

$$q = I_0 \tau$$

which governs the depth of penetration. A more detailed description of the instruments and field work can be found in Bernard (2006) and Shushakov, (2006), and will be described in Bernard (2007, this Issue).

For each curve e(t), the maximum value E_0 and the decay time constant T_d is calculated, forming the functions $E_0(q)$ and $T_d(q)$. The inversion of these data, made through a mathematical model, gives rise to the solution searched: the distribution of the water content θ_{MRS} and time decay T_d constant with depth (see Yaramanci and Hertrich (2007, this Issue) , and Legchenko (2007, this Issue). The reliability of the solution depends on the signal to noise ratio and on the simplifications and assumptions made in the theoretical model. The correspondence between the measured MRS parameters and hydrogeological parameters of aquifers has been described by Legchenko *et al.* (1990), Schirov *et al.* (1991), Legchenko and Shushakov (1998) among others. From empirical relations involving θ_{MRS} and T_d, storativity, hydraulic transmissivity and other aquifer parameters can be deduced, as is explained in Lubczynksi and Roy (2007, this Issue).

The principal factors that affect the measured amplitude and time decay of the water signal

In MRS, the existence or absence of the signal e(t) is directly linked to the existence or absence of the subsurface water. The maximum amplitude and the shape of the $E_0(q)$ curve depend on:

1. The depth, thickness and number of the aquifer layers:

In Figure 2a, the MRS $E_0(q)$ curves for a water layer with a 20 % volume fraction of water and 10 m thick

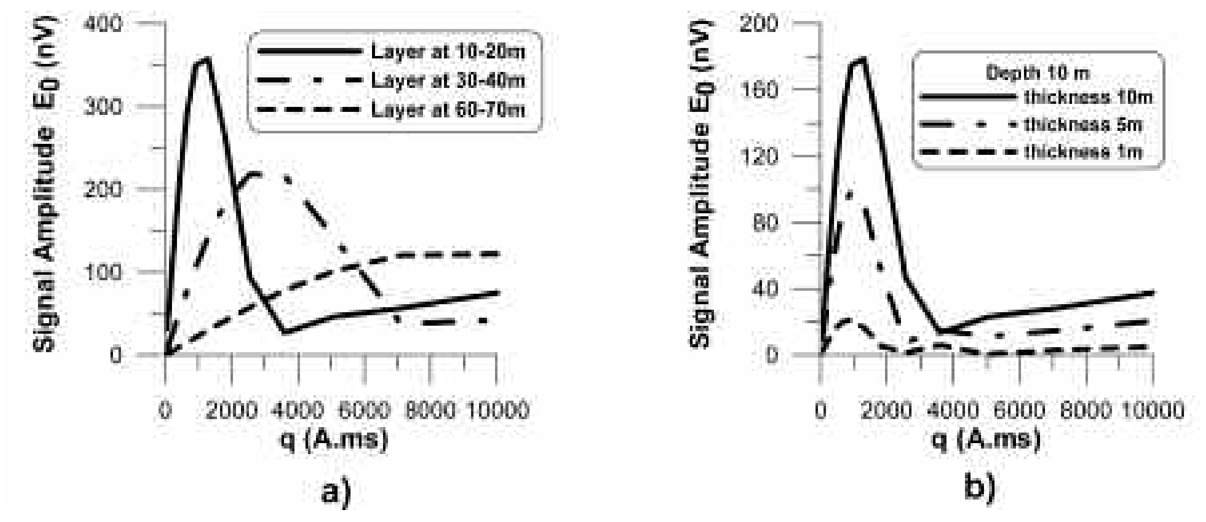

Fig. 2. a) Water signal amplitude curve $E_0(q)$ for different depths of the aquifer (modified from Lieblich *et al.,* 1994). b) Water signal amplitude curve $E_0(q)$ for different thickness of the aquifer (modified from Pusep *et al.,* 1991). Geomagnetic field dip 55° N; Larmor frequency 1890 Hz; antenna square loop 100 m side; water content: 20% for a) and 10 % for b); ground resistivity 50 ohm.m

Fig. 2. a) Amplitud de la señal $E_0(q)$ para diferentes profundidades del acuífero (modificado de Leiblich et al., *1994). b) Amplitud de la señal $E_0(q)$ para diferentes potencias del acuífero (modificado de Pussep* et al., *1991). Inclinación de campo geomagnético 55° N; frecuencia de Larmor de 1980 Hz; antena cuadrada de 100 m de lado; contenido en agua del 20% para a) y 10% para b); resistividad del terreno de 50 ohm.m*

at different depths are shown. The general rule is that the maximum amplitude of the signal decreases for increasing depth of the aquifer, being necessary higher amplitudes of the excitation moments to reach the maximum of $E_0(q)$. In Figure 2b, the curves for a water layer with its top at 10 m of depth are shown for different thickness of the layer; the maximum amplitude is reached for about the same amplitude of the excitation moment for all the cases, but the signal amplitude decreases with decreasing thickness (amount of water) (Pusep *et al.,* 1991; Lieblich *et al.,* 1994).

According to the former rules, and from a qualitative point of view, the shape of the MRS $E_0(q)$ curve gives a certain indication of the relative depth and thickness of the aquifer. In Figure 3 several theoretical models are shown. In appearance there are noticeable differences among the situations, but the distinction of the models in a real field curve is not always obvious, as it happens for instance with Vertical Electrical Sounding curves. A multilayer aquifer is not distinguishable from a thick aquifer. The main usefulness of the shape analysis is for the foreseeing of the evolution of an aquifer when several MRS are measure over it.

2. The electrical conductivity of the subsurface:

The water signal amplitude and the viability of the signal detection depend also on several other factors, which affect the amplitude of the excitation field. There is a depth limit for the penetration of the excitation electromagnetic field into the ground, as well as for the distance of the way back to surface of the secondary response field. For a given geographical coordinates (fixed value of the Larmor frequency $f_L = 0.04258\ B_0$), the penetration of an electromagnetic field is controlled by the resistivity of the rocks, according to the denominated skin depth ([resistivity/frequency]$^{1/2}$), and electromagnetic waves can travel further before attenuating in resistive rocks.

For each particular case (geographical coordinates, antenna size, ground conductivity, water content), the depth at which no amount of water contributes significantly to E_0 depends on the thickness of the aquifer. For example, at geomagnetic field inclination of 56° S, Larmor frequency of 2300 Hz and using a square antenna of 100x100 m, a one meter thick layer of bulk water will be detected down to 50 m of depth in a conductive terrain of 1 ohm.m.

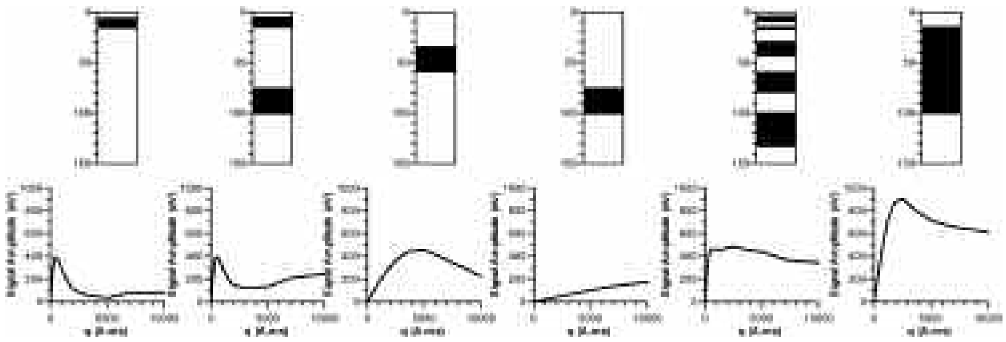

Fig. 3. Shape of the MRS curve $E_0(q)$ for different aquifer formations. Geomagnetic field dip 55° N; Larmor frequency 1890 Hz; antenna square loop 100 m side; water content: 20%; ground resistivity 50 ohm.m
Fig. 3. Formas de la curva $E_0(q)$ del SRM para diferentes formaciones acuíferas. Inclinación de campo geomagnético 55° N; frecuencia de Larmor de 1980 Hz; antena cuadrada de 100 m de lado; contenido en agua del 20%; resistividad del terreno de 50 ohm.m

However, for a 50 meter thick layer of bulk water, the detection depth is increased down to 65 m considering the top of the layer. For a resistivity of 100 ohm.m, this penetration increases down to 150 and 200 m respectively (Hunter and Kepic, 2003). At geomagnetic field inclination of 90° and Larmor frequency of 2500 Hz, for a 1000 ohm.m half space the detection depth of the signal from a one meter thick layer of bulk water is 150 m, but this depth is reduced to only 50 m for 2 ohm.m (Legchenko *et al.,*1997b).

In resistive rocks, as a general rule, protons at a distance from the antenna more than twice the anten-

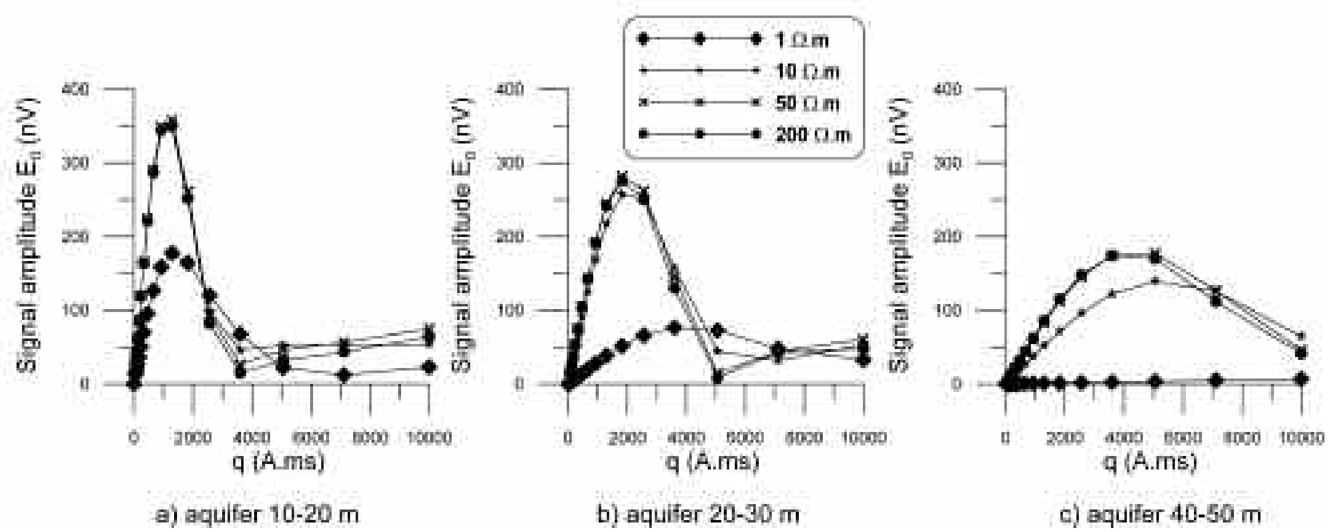

Fig. 4. The MRS $E_0(q)$ curves for a 20% saturated water layer 10 m thick, at different depths and for different conductive half-spaces. (Modified from Shushakov, 1996). Geomagnetic field dip 55° N; Larmor frequency 1890 Hz; antenna square loop 100 m side
Fig. 4. Curvas $E_0(q)$ del SRM para un acuífero con el 20% de agua y 10 m de potencia, situado a diferentes profundidades y para diferentes conductividades del medio. (Modificado de Shushakov, 1996). Inclinación de campo geomagnético 55° N; frecuencia de Larmor de 1980 Hz; antena cuadrada de 100 m de lado

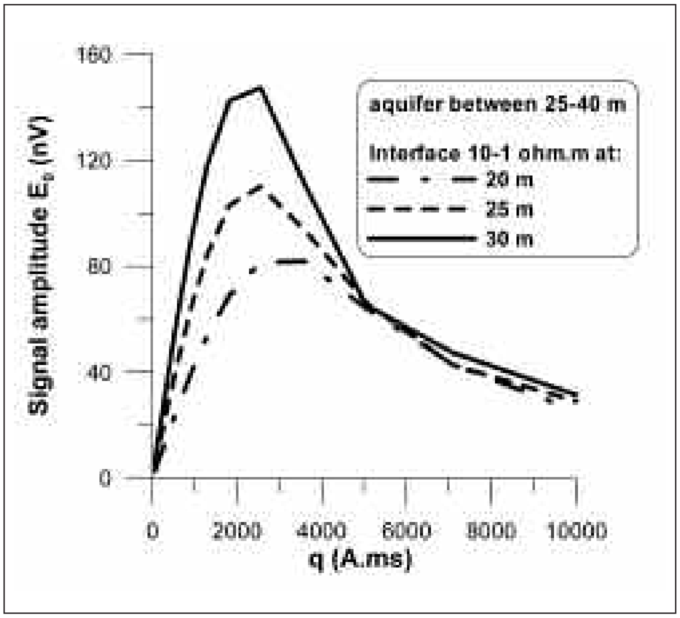

Fig. 5. The MRS E_0(q) curves for an aquifer layer between 25 m and 40 m of depth, for three different positions of a geoelectrical interface between a 10 ohm.m medium and a 1 ohm.m one. (Modified from Legchenko, 2006). Geomagnetic field dip 55° N; Larmor frequency 1890 Hz; square loop 100 m side; water content: 20%

Fig. 5. Curvas E_0(q) para SRM producidos por un modelo de acuífero formado por una capa entre 25 m y 40 m de profundidad, y tres posiciones diferentes de una interfase geoeléctrica entre un medio de 10 ohm.m y otro de 1 ohm.m (Modificado de Legchenko, 2006). Inclinación de campo geomagnético 55° N; frecuencia de Larmor de 1980 Hz; antena cuadrada de 100 m de lado; contenido en agua del 20%

na diameter produce negligible small part of the total signal, and for calculations the range of depth may be restricted to this distance. In Figure 4, an example of the attenuation produced by rocks conductivity is shown. A layer 10 m thick and with a 20 % of water is considered. In Figure 4a, the top of the layer is at 10 m of depth and E_0(q) functions are shown for a ground of 1, 10, 50 and 200 ohm.m of resistivity. The amplitude is higher with the resistivity, but no difference was found until very conductive space (1 ohm.m); the maximum value is reached for the same amplitude of the excitation pulse. As the top of the layer gets deeper (Figures 4b and 4c) the amplitude gets smaller and the maximum is reached for higher excitation moments. At 40 m of depth, no signal is registered when the ground has 1 ohm.m.

The effect of the geoelectrical model on the shape of the E_0(q) curve is shown in Figure 5 for an aquifer 15 m thick at 25 m depth, for three different depths of a geoelectrical interface between a 10 ohm.m layer and a 1 ohm.m one. Despite the constant depth of the aquifer interface at the three models, the position of the geoelectrical interface governs the shape of the curve: the shallower is the electrical interface, the smaller is the signal amplitude, and bigger excitation moments are needed to reach the maximum, because of the more important screening effect (Legchenko, 2006).

Other question related with the electrical conductivity of the subsurface is that not only the amplitude but also the phase of the signal is modified. The electrical conductivity of rocks makes the excitation field complex (this modifies its components along the geomagnetic field and transversal to it). The oscillating magnetic field generated by the excited hydrogen protons is also affected by conductive layers (Shushakov and Legchenko, 1994a,b; Trushkin *et al.,* 1995; Shushakov, 1996). In case of multilayer aquifers separated by a conductive medium the total received signal is not equal to the sum of signals from separate layers, due to the phase shift. Quite strong effects of phase shifts occur between signals from two aquifers separated by an electrically conductive layer, which has an influence on the apparent amplitude and decay time of the measured signal (Schirov and Rojkowski, 2002). In Figure 6 an example is shown for two layers 10 m and 20 m thick at 10 and 50 m of depth for several ground conductivities; for a resistive medium of 100 ohm.m the two aquifers are clearly identify in the E_0(q) curve, but for 2 ohm.m only the shallower layer is visible. The existence of rocks with high conductivity has a screening effect that lowers the amplitude of the water signal in such a way that the presence of deeper aquifers can be not seen in the E_0(q) curve, what causes the depth of investigation to be decreased. Erroneous interpretations can be made when the inversion scheme assuming a resistive ground is used when this is not the case.

In some places, the ground water salinity can be the reason for a low conductivity of rocks. It is not yet possible to discriminate between the existence of salt water or conductive rocks by measuring only the MRS signal amplitude and phase. As a consequence, in this case the geoelectrical information is strongly necessary and recommended to perform a MRS survey.

3. The phase shift between the Larmor frequency and the excitation pulse frequency:

In presence of shallow aquifers the magnetic resonance signal is complex, even for a resistive environment, and at large pulse moments, when the excita-

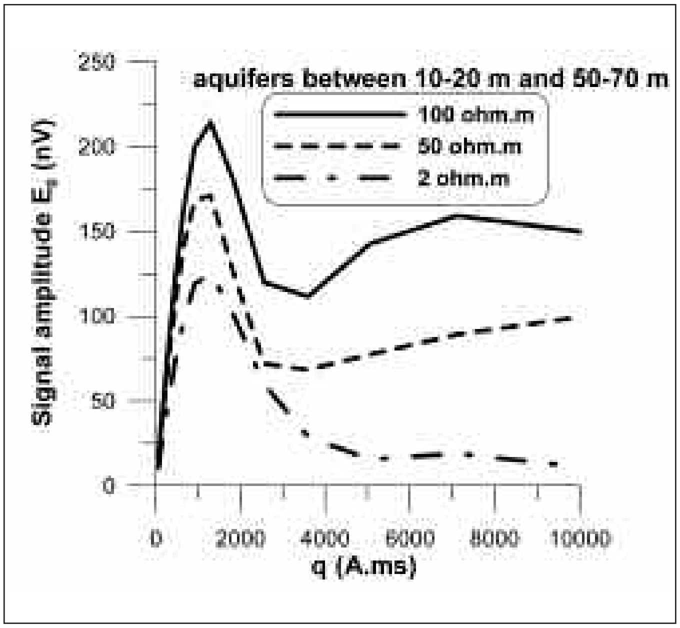

Fig. 6. The E_0(q) MRS curves for a two layers model situated between 10 and 20 m and 50 and 70 m respectively, with a 10 % in water content, for different conductive half spaces. (Modified from Trushkin *et al.,* 1995). Geomagnetic field dip 55° N; Larmor frequency 1890 Hz; antenna square loop 100 m side

Fig. 6. Curvas E_0(q) para SRM producidos por un modelo de acuífero formado por dos capas situadas entre 10 -20 m y 50-70 m respectivamente, con un 10 % de contenido en agua, en medios de diferente conductividad. (Modificado de Trushkin et al., *1995). Inclinación de campo geomagnético 55° N; frecuencia de Larmor de 1980 Hz; antena cuadrada de 100 m de lado*

tion field is much stronger than the magnetic field of the Earth, the water of the aquifer may generate a signal comparable in amplitude to a signal generated by a deeper aquifer (Legchenko, 2005; Shushakov, 2006). This effect can be seen in Figure 7, where two real examples are shown; both MRS are taken at a place where only an aquifer about 10 m thick at 5 m of depth is found. A raise of the E_0(q) curve is observed for excitations moments exciding 4000-4500 A.ms. The offset between frequencies is also present in the case of varying geomagnetic field, for any amplitude of the excitation pulse. Using an improved model for inversion may avoid an erroneous interpretation of the distortion of the MRS curves due to this effect (Legchenko, 2005).

4. The magnitude and inclination of the geomagnetic field:

The amplitude of the magnetic resonance signal e(t) depends also on the amplitude of the geomagnetic field. In the equator the intensity of the magnetic field is half than that in the poles; this makes the Larmor frequency smaller but, for the same resistivity, the skin depth is larger, allowing the electromagnetic fields travel longer in the equator. Unfortunately larger skin depth does not increase the maximum depth of penetration, because the resonance signal is pro-

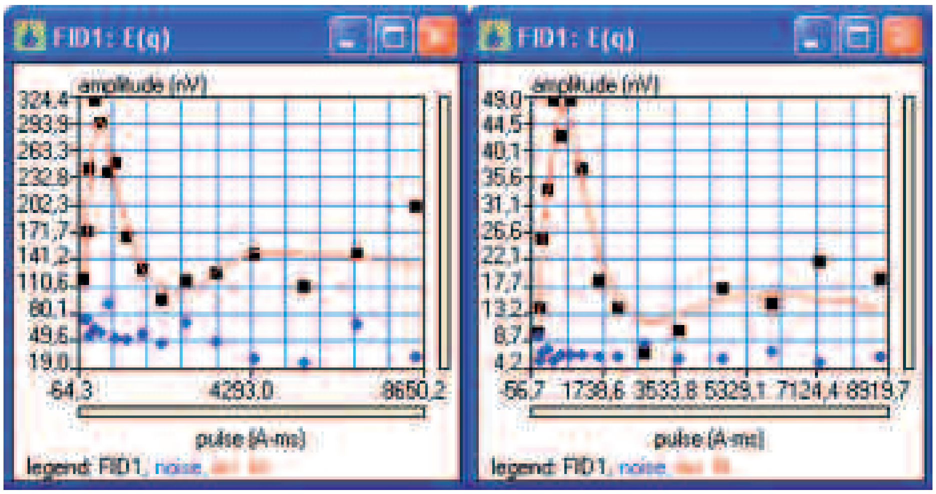

Fig. 7. MRS measured signals at two different sites, where there is only one shallow aquifer. In appearance, a deeper aquifer is also present in the measurements. Inversion of these curves without taken into account this effect, will produce a phantom deeper aquifer (From Plata *et al.,* 2004)

Fig. 7. Curvas MRS medidas en dos sitios diferentes, donde solamente existe un acuífero somero. La forma de las curvas es análoga a la que se mediría en el caso de que existiera un segundo acuífero más profundo. La inversión de estas curvas sin tener en cuenta este efecto daría lugar a interpretar un segundo falso acuífero (Tomado de Plata et al., *2004).*

portional to the square of the geomagnetic field intensity ($E_0 \approx B_0^2$), and this effect is stronger than the influence of the skin depth. Signals produced by the same aquifer can be between two and four times stronger in the poles than in the equator, the ratio increasing with depth (Legchenko *et al.,* 1997b).

The inclination of the geomagnetic field also contributes to the maximum depth of detection, because it has an effect on the perpendicular component of the excitation field, which is the most important to modified the original orientation of hydrogen magnetic moments. The maximum amplitude of the signal is sensitive to the inclination of the geomagnetic field only when the aquifer is located at a depth less than 20-25 m. The modification of the shape of the curve $E_0(q)$ with the magnetic field inclination is more important for nonconductive rocks, with maximum $E_0(q)$ values at the equator and presents oscillations of the curve for high values of the excitation moment. (Shushakov, 1996; Legchenko *et al.,* 2002).

Due to variations of the ionosphere the external component of the geomagnetic field may change significantly during the time needed to make a full MRS (some times it can take several hours). This change produces a modification of the Larmor frequency; if the change exceeds by more than 10 Hz from the excitation pulse frequency resonance conditions of measuring fails.

5. The type of water containing rocks:

The time decay of the signal depends on the type of rocks, increasing with grain size. Only signals with a decay time greater to certain instrumental limitations (about 40 ms nowadays) can be measured. It limits the bounded water in fine grained aquifers to be detected, or to be very much attenuated. Time decay is greater for karsts limestone than for fractured sandstone (Legchenko *et al.,* 1997a,b).

The magnetic susceptibility of rocks modifies the geomagnetic field in the investigated volume. When magnetic rocks are present, the intensity and gradient of the geomagnetic field is different in different parts of the subsurface what causes the Larmor frequency be also different and hence MRS signal can not be detected. This may be the reason for some cases of failure of MRS (Legchenko *et al.,* 1998). The high magnetic field gradient also produces the decay time constant T_d to become shorter than actually measura-

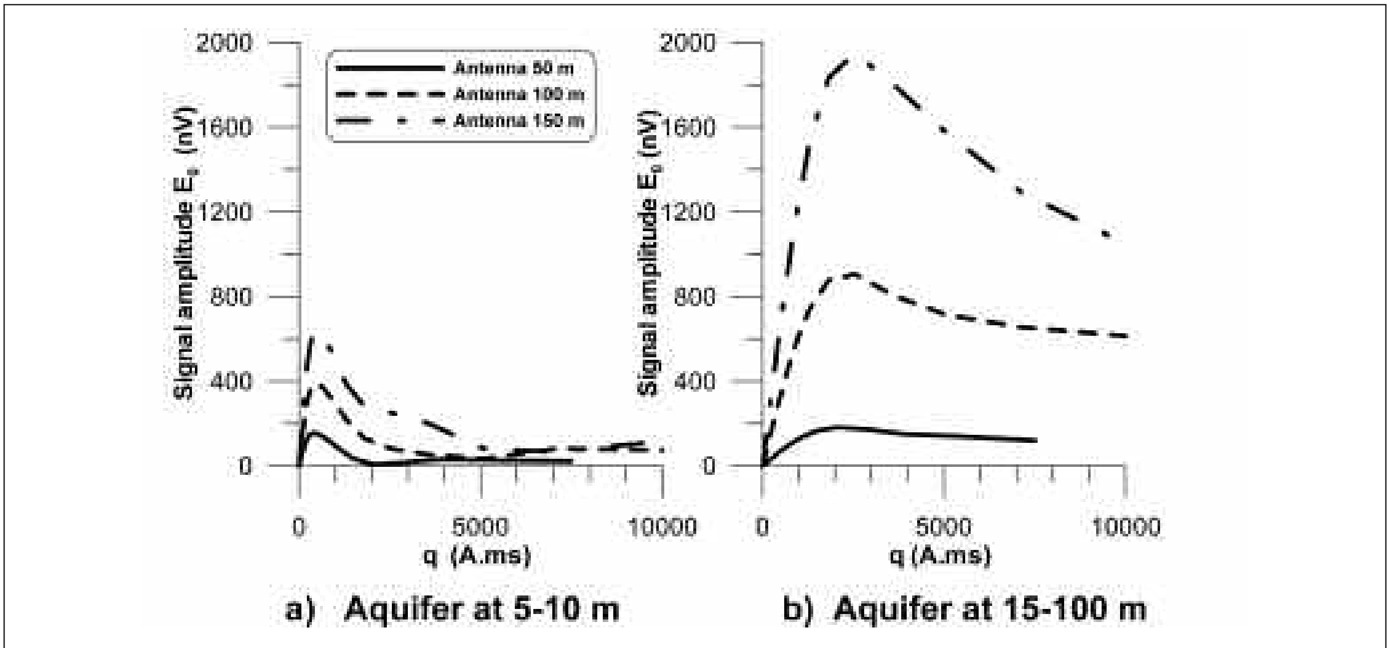

Fig. 8. Variation of the signal with the antenna size. a) Aquifer between 5-10 m. b) Aquifer between 15-100 m. (Modified from Legchenko *et al.,* 1997b). Geomagnetic field dip 55° N; Larmor frequency 1890 Hz; water content: 20%; 50 ohm.m terrain resistivity

Fig. 8. Variación de la señal con el tamaño de la antena. a) Acuífero entre 5-10 m. b) Acuífero entre 15-100 m. (Modificado de Legchenko et al., *1997b). Inclinación de campo geomagnético 55° N; frecuencia de Larmor de 1980 Hz; contenido en agua del 20%; resistividad del terreno de 50 ohm.m*

ble signals, making groundwater undetectable in this conditions (Roy *et. al,* 2006).

6. The size of the loop:

The maximum amplitude of the induced voltage e(t) is proportional to the whole number of hydrogen protons within a certain volume around the antenna. Increasing the diameter of the wire loop or antenna laid down on the surface may increase the volume of water studied, and hence the amplitude of the signal. When the target is of limited size relatively to the antenna, this will not happen. As a first single approximation, it can be considered that water is excited in an area around the antenna of approximately one and a half times its diameter. Increasing the size of the antenna increases the measured signal amplitude also because of the greater surface for electromagnetic induction produced by the response field (Legchenko *et al.,* 1997b), but it also increases the magnitude of the signal introduced by the noise. In Figure 8 a comparison is established between the theoretical $E_0(q)$ sounding curves for two aquifers of different thickness: 5 m (at 5 m of depth) and 85 m (at 15 m of depth): about the same maximum amplitude is registered from the thick aquifer and from the shallower thinner aquifer when using the smallest antenna of 50 m; using larger antennas for the thin aquifer gives bigger signals than the thick aquifer with the smallest antenna. The detection depth is really controlled by the pulse moment $q= I_0\tau$. The signal e(t) is produced by deeper water molecules for increasing values of q, but the loop size sets a limit to the penetration. For a given antenna diameter, increasing the current intensity I_0, increases the amplitude of the electromagnetic excitation field, so that it can excite deeper water protons. The maximum depth of inves-

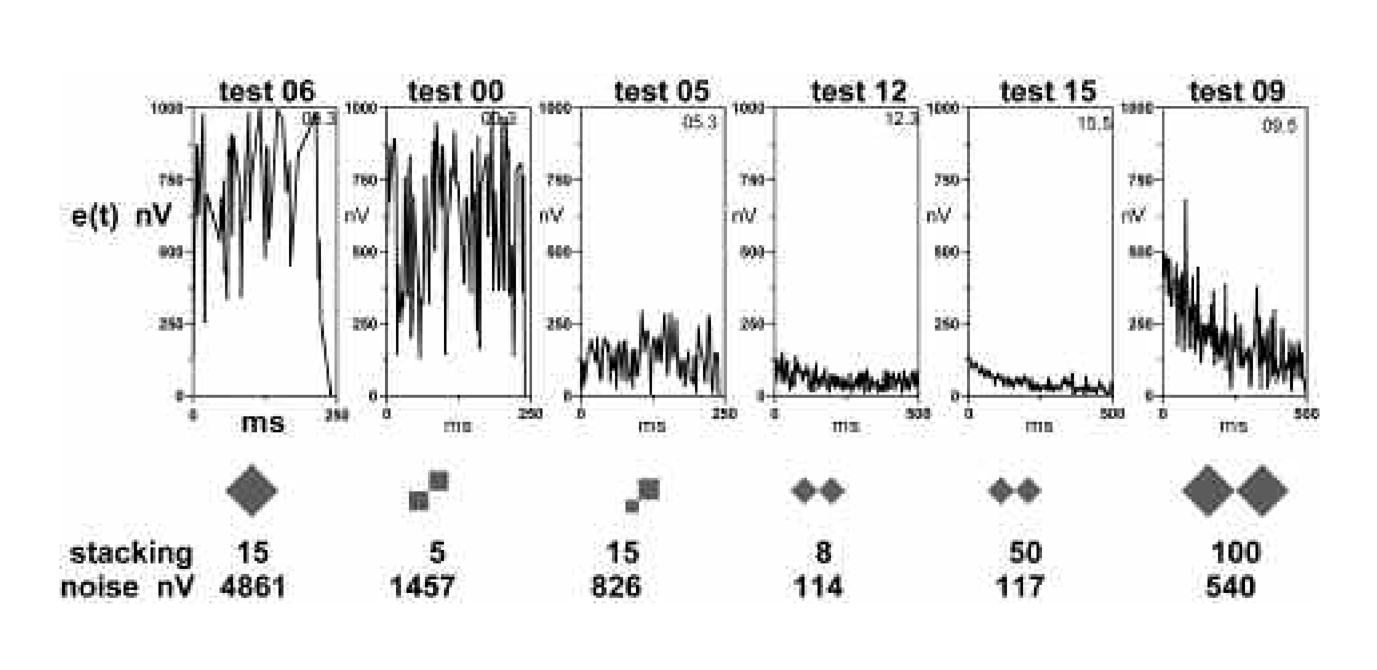

Fig. 9. Example with real data to show the improvement in the signal/noise ratio attained at the same site by modifying the shape and size of the antenna and the number of signals stacked. At the bottom of the figure an indication of the antenna shape, size and orientation, as well as the stacking number used and the mean value for the ambient noise are given. Each graph represents the values for the stacked measured signal (signal plus noise) versus recording time e(t), for the same value of the excitation moment q. Big antennas are of 75 m of side; small ones are of 37.5 m. (Modified from Plata and Rubio, 2002)

Fig. 9. Ejemplo de datos reales que muestran la mejora producida en la relación señal/ruido con mediciones efectuadas en el mismo lugar, pero modificando la forma, el tamaño, la orientación de la antena, así como el número de señales sumadas. En la parte inferior de la figura se indican la forma, tamaño, orientación de la antena, número de adiciones efectuadas en cada caso y el nivel medio de ruido ambiente obtenido. Cada gráfico representa el valor de la función e(t) (señal más ruido) medida, utilizándose en todos los casos el mismo valor del momento de excitación q. La antena de mayor dimensión tiene 75 m de lado, y la de menor es de 37.5 m (Modificado de Plata y Rubio, 2002)

tigation is considered to be less than twice the antenna diameter, and for practical applications it is usually set equal to the diameter or side of the loop.

7. The effect of external electromagnetic noise:

Electromagnetic noise, produced by power lines, pipe lines and industrial activities, is the most severe limitation for signal detection. Different configurations for the receiver antenna can be used to reduce the noise effects (Truskin *et al.,* 1994), but it also has an effect on the signal amplitude, as is explained in Bernard (2007, this Issue). Stacking of signals may also improve the signal to noise ratio. An example of several e(t) curves measured at the same place and with different recording parameters is shown in Figure 9.

If the amount of water is enough to produce a strong signal, in many circumstances it is possible to select the appropriate recording parameters for field survey so that a reasonable signal to noise ratio can be attained. Nevertheless, this kind of noise is not stationary, and it may vary along the day. In Figure 10 some examples of these situations are given.

In conclusion, there are several factors than can

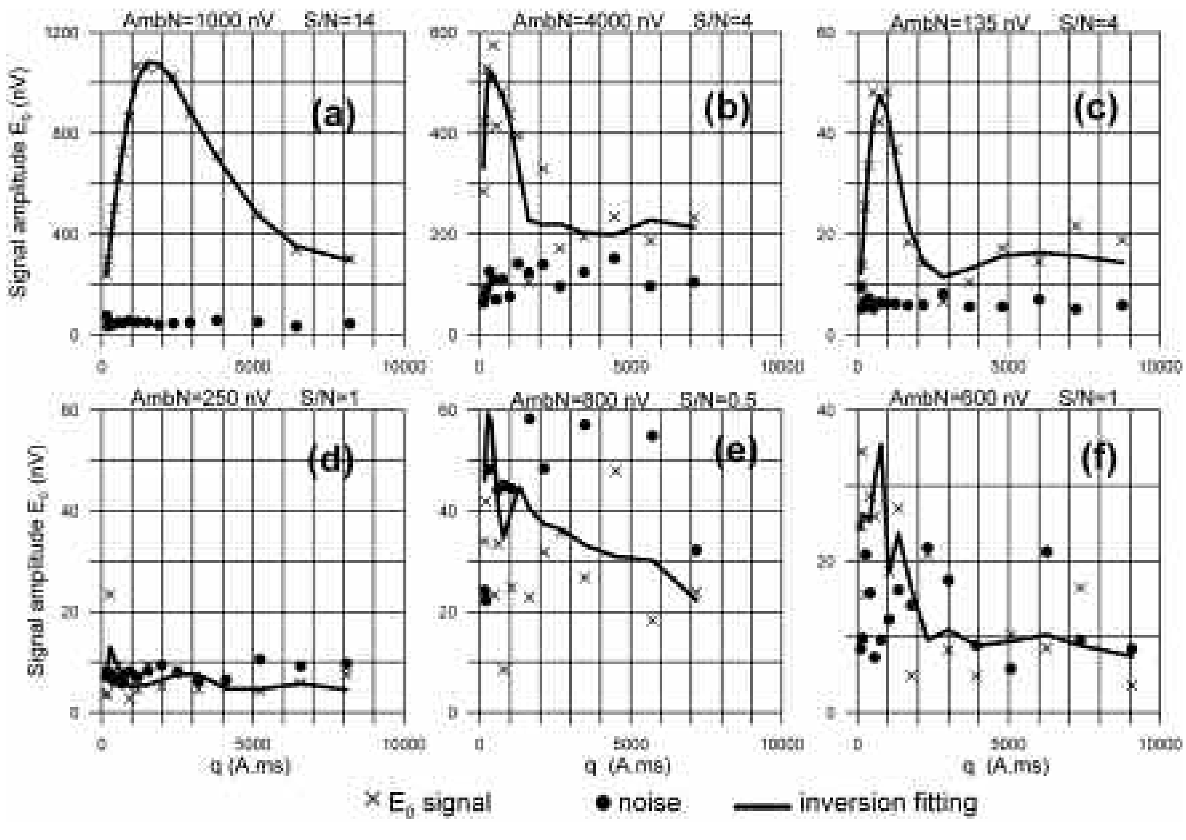

Fig. 10. Examples of several MRS with different signal/noise ratios (S/N). a) Strong signal and low noise produce good records. b) and c): The presence of noise, but with enough amplitude of the water-signal, gives acceptable records. d) No signal, no water. e) A high level of noise with a small water-signal, results in useless records. c) and f):Records made at the same site, taken at different times, showing the effect of a variation in the noise level and its behaviour. AmbN: ambient noise. (From Plata and Rubio, 2003)

Fig. 10. Ejemplos de varios SRM con diversos valores de la relación señal/ruido (S/N). a) Señal de gran amplitud y ruido bajo dan lugar a buenos registros. b) y c): Cuando la señal del agua es de suficiente amplitud respecto del ruido, pueden obtenerse registros aceptables. d) Si no existe agua, no existe señal. e) En presencia de un alto nivel de ruido y una baja amplitud de la señal, se obtienen registros no válidos. c) y f): Registros obtenidos en el mismo lugar, pero en diferentes momentos, mostrando el efecto de la variación del nivel y tipo de ruido con el tiempo. AmbN: ruido ambiente (Según Plata y Rubio, 2003)

modify in quite an important way the possibility of achieving good MRS measurements, and they constitute the frame of limitations of the methodology. As in any other geophysical method, these limits have to be known and understood for the MRS users. The most favourable conditions for MRS are met when measurements are made close to the magnetic poles (high values of the amplitude and dip of the Earth's magnetic field), resistive (500 ohm.m) and not magnetic rocks, and long decay times (800 ms). The most unfavourable conditions are in the equator, low resistivity rocks (<10 ohm.m) and short decay times (fine grained materials with T_d<30 ms). Some of these limitations are nowadays object of research and development, improving the instrumentation, mathematical modelling and the field methodology.

The mathematical model of the signal

The mathematical model of the physical phenomenon and its measurement allows getting the theoretical signal produced by a defined configuration of the ground water and rocks distribution in the subsurface (modelling or direct problem), as well as deducing the aquifers configuration from the MRS signal actually measured in the field (inversion problem).

To translate the physical phenomena of MRS into mathematical equations, several steps have to be accomplished:

- The magnetic moment of the water molecules in equilibrium state.
- The electromagnetic field created in the rocks by a cable loop of different sizes and shapes on the surface of the ground when passing through it an electric current of intensity I_0 and frequency f_L, during a time τ.
- The electromagnetic field created by the water molecules after being excited by the primary field.
- The voltage induced on a cable loop at the surface by this secondary field.

The geometry, electric properties, and amount of water of the subsurface layers through where these fields travel have also to be taken into account.

A detailed description of the set of the model equations can be found in Schirov *et al.* (1991); Goldmand *et al.* (1994);Trushkin *et al.* (1995); Shushakov (1996); Legchenko *et al.* (1996); Legchenko *et al.* (1997b); Weichman and Lavely (1999); Legchenko and Valla (2002); Weichman *et al.* (2002); Legchenko (2005); Hertrich *et al.* (2005a), among others. The following is only a summary description intended for not specialists in the mathematical field.

Special attention is given in this description to the hypothesis and assumptions made in the mathematical modelling, because of its influence in the final results.

In thermal equilibrium magnetic moments of hydrogen protons are oriented parallel to the geomagnetic field B_0, giving rise, after the Curie's law, to a macroscopic magnetic moment M_0 for protons in water:

$$M_0 = n(\gamma^2 h^2 / 4kT)B_0$$

where:

- n is the number of hydrogen protons per unit volume of water
- T is the absolute temperature
- h, k and γ are constants: Planck, Boltzman's and gyromagnetic ratio of hydrogen protons respectively

Salinity affects the hydrogen density and the conductivity of the water, which affects the electromagnetic fields and the value of n. Normally groundwater has low NaCl concentrations, except in costal aquifers with sea water intrusion. Within aquifers down to 150 m of depth, the variations of n and T with the geographical coordinates is not very significant, but the variation of the geomagnetic field can be important, which significantly changes the value of M_0 (Dunn *et al.,* 2002), but for a reduced geographical area with non magnetic rocks, M_0 may be considered as a constant.

At a point located at a distance r from the surface in a coordinate reference system,the magnetic moment due to one small volume of rock dV(r) with a volumetric water content distribution w(r) (0≤w(r)≤1) will be

$$dM(r) = M_0 w(r) dV(r)$$

When an alternating current of intensity I_0 and frequency ω_L (Larmor frequency) is passed through a loop of wire (transmitter T_x) on the surface of the Earth during a time τ (excitation pulse q=$I_0\,\tau$ in A.ms), it creates an oscillating magnetic field $B_T \exp(-i\omega_L t)$ with a maximum value by unit of current intensity given by $b_T=B_T/I_0$. This field has a component $b_{T\perp}$ (r) perpendicular to the geomagnetic field B_0, and its amplitude depends on distance, rock resistivity, and geomagnetic field inclination. $b_{T\perp}$ causes the nuclear magnetization for protons dM(r) to tilt at an angle θ from its equilibrium position along B_0, given by

$$\theta(r) = 0.5\gamma b_{T\perp}(r) I_0 \tau = 0.5\gamma b_{T\perp}(r) q$$

(note the dependency on excitation pulse amplitude q) producing a component $dM_\perp(r)$ of the magnetic moment perpendicular to the geomagnetic field B_0

$$dM_\perp(r) = M_0 w(r) \sin(\theta(r)) dV(r)$$

When the pulse is terminated ($t>\tau$), the component $dM_\perp(r)$ does not disappear instantaneously, but the vector dM(r) precesses about the geomagnetic field at the Larmor frequency ω_L until it is again aligned with B_0. This rotating magnetic dipole produces an electromagnetic secondary field that induces a voltage into the receiver loop on the surface (R_x), which is the MRS measured signal. In a rotating reference coordinates system the equation for the variation of the voltage with time (free induction decay signal), for a fixed value of q is given by:

$$e(t,q) = -\int_v dM_\perp(r) \omega_L b_{R\perp}(r) \exp(-t/T_d(r))$$

$$e(t,q) = -\int_v \omega_L M_0 w(r) b_{R\perp}(r) \sin(0.5\gamma b_{T\perp}(r) q)$$
$$\exp(-t/T_d(r)) dV(r) = E_0(q) \exp(-t/T_d(r))$$

that is the envelope of the induced voltage obtained after the synchronous detector. T_d is the time decay constant of the signal, and $b_{R\perp}$ is the magnetic induction field that would be created by a unit current in the receiver antenna. The initial amplitude $E_0(q)$ is then given by

$$E_0(q) = -\int_v \omega_L M_0 w(r) b_{R\perp}(r) \sin(0.5\gamma b_{T\perp}(r) q) dV(r) \qquad [1]$$

This equation is established for a resistive earth model (free space) and homogeneous magnetic static field; in its mathematical development some assumptions are made, and in particular it is considered that the duration of the excitation pulse τ is much shorter than the time decay T_d, and the amplitude of the magnetic excitation field smaller than the Earth's magnetic field.

As the excitation pulse uses an alternating current with frequency ω_L, the free induction decay signal measured in a fixed reference system, considering a phase shift φ between the EM field in the subsurface and the induced current in the loop, becomes:

$$e(t,q) = E_0(q) \sin(\omega_L t + \varphi) \exp(-t/T_d(r))$$

For a given value of the pulse q, the function e(t) is linear with regard to the water content w(r). The q dependence is nonlinear (sine function). The maximum signal contribution will come from a definite depth range. The depth increases with q, making it possible to make a depth sounding varying the value of q. Equation [1] allows calculating the theoretical signal produced by a certain water distribution, or direct problem.

In order to calculate the boundaries of the water-saturated layers from the really measured $E_0(q)$ function (inverse problem), equation [1] must be solved. For the case of horizontal infinite layers, it can be rewritten as the convolution of a kernel function K(q,z) (response of a thin water layer at depth z) with the function of the water concentration with depth,

$$E_0(q) = \int_0^\infty K(q,z) w(z) dz \qquad [2]$$

with

$$K(q,z) = \omega_L M_0 \int_{x,y} b_{R\perp}(z) \sin(\theta) dx dy \qquad [3]$$

In as much as the kernel takes into account the influence of the parameters that affect the signal $E_0(q)$, the inversion of equation [2] will give the correct solution for w(z). Nevertheless, this problem belongs to the category of ill-posed ones, which means that the solution is not unique. There are different functions w(z) with the same value for the integrated water quantity (sum of the product of the water content by the thickness of each layer) that provide a theoretical curve $E_0(q)$ that fits the observations (equivalence principle). A more complete description of the inversion procedure can be found in Legchenko and Shushakov (1998); Mohnke and Yaramanci (2002); Weichman *et al.* (2002); Guillen and Legchenko (2002); Braun and Yaramanci (2003); Legchenko (2003a, b); Legchenko (2005), among others, and is also presented in Yaramanci and Hertrich (2007, this Issue) and Legchenko (2007, this Issue).

Basic ideas about the decay time T_d

The signal e(t) has an exponential decay governed by the factor $\exp(-t/T_d)$, and completely vanishes when the magnetic moments of hydrogen protons are again aligned with the geomagnetic field B_0. The time decay constant T_d is the time at which the signal has diminished approximately to one third of its initial value (0.367 times) and its value reflects how effectively the magnetic energy of the hydrogen protons is transferred to or from its surroundings. The decay time constant is a property of the protons system and its environment. The precession rotating vector dM(r) has one component in the direction of the geomagnetic field B_0 and another in a plane perpendicular to B_0, so that the decay time T_d is not the same for both directions. The longitudinal one is called T_1 and the transversal T_2; approximately $T_1 \approx 1.5\ T_2$. Since the measurable effect is linked to the transverse magnetization, the decay time constant of the observed relaxation is the transverse one T_2. In porous media the values of T_1 and T_2 are proportional to the mean pore size, and are affected by the chemical and physical properties of water in the rock, the uniformity of the geomagnetic field, the magnetic susceptibility of rocks and fluids, the fluid viscosity and diffusion, and the paramagnetic impurities ions on the pore walls or in the fluids (Dunn *et al.,* 2002), being T_1 less influenced by magnetic heterogeneities than T_2. In fact, instead of T_2 the observed decay time (free induction decay time) is somewhat a shorter one T_2^*, because of natural inhomogeneities in the geomagnetic field magnitude, due to a nonzero magnetic susceptibility of rocks.

The correlation between the decay relaxation time and grain size is discussed in Semenov (1987).

An empirical correlation between decay time and pore size is given in Table 1. Though it is better to use T_1 because it is more reliable than T_2^*, not enough field data are yet available for values of the longitudinal decay time in different lithologies.

The free water and the bounded water (see Mejias and Plata, 2007, this Issue) are differentiated by the decay time, which is much shorter for the bounded water (less than 20-30 ms) (Schirov et al., 1991). The excitation pulse $q=I_0\,\tau$ is supposed to be much shorter than the decay time of the signal ($\tau << T_2^*$). For signals with decaying time less than 200-300 ms this assumption is not valid and should be taken into account in the mathematical model (Legchenko et al., 1997b).

Thus, the sensitivity to magnetization of rocks limits the reliability of T_2^* as a parameter for estimating hydraulic conductivity and encourages the application of T_1. With the actual instrumentation T_1 may be approximated by measuring two T_2^* delayed values, and it is named after T_1^*. A new equation for $E_0(q)$ is

T_2^* ms	lithology
< 30	Sandy clays
30-60	Clayey sands, very fine sands
60-120	Fine sands
120-180	Medium sands
180-300	Coarse and gravely sands
300-600	Gravel deposits
600-1500	Surface water bodys

Table 1. Correlation between relaxation time T_2* and lithology (after Schirov *et al.,* 1991)
Tabla 1. Valores del tiempo de relajación T_2 para diversas litologías. (Según Schirov* et al.*, 1991)*

needed, from where to estimate T_1 (Legchenko, 2003a) using two excitation pulses with a delay between them:

$$E_{20}(q,\tau_p) = \omega_L \int_V \sum_k [B_{1k}(r)\exp(j(\varphi_{0k}(r)+\Pi)$$
$$M_{\perp k}(r,\tau_p)/I_{ok}]w(r)x(r)dV(r)$$

where

$$x(r) = 1 - \exp(-\tau_p / T_1^*(r))$$

is the unknown function, τ_p is the time delay between pulses, and $M_{\perp k}(r,\tau_p)$ the component perpendicular to the geomagnetic field of the spin magnetization after the second pulse. Nevertheless, in rocks with a high internal magnetic field gradient (like basaltic gravel highly magnetized) the signal from free water T_2^* is shorter (about 10 ms) than the currently measurable signal length, what makes water undetectable, either with one or two excitation pulses.

The relationship between the measured relaxation time and pore geometry in an aquifer is the physical basis for establishing a relationship between the hydrogeological parameters and proton magnetic resonance ones, and, as the magnetic susceptibility (and other factors) may significantly vary from one site to another, no universal relationship is to be expected and some regional calibration is mandatory (Legchenko and Valla, 2002), as is explained in Lubczynski and Roy (2007,this Issue). By replacing measurements of T_2^* by T_1^* for estimating hydraulic conductivity more reliable results will be attained.

Some considerations on the applicability of the model

In the model used to describe the physical phenomenon of MRS and its measurement in mathematical equations, quite a few simplifications and assumptions are or can be made. Some of them are mandatory; otherwise the physical basis of the method is not fulfilled, like to assume the geomagnetic field to be homogeneous all over the investigated volume, which means that the subsurface must be non-magnetic.

Some of the possible simplifications that can be made are:

- To assume horizontal stratification and topography.
- To use the same antenna as transmitter and receiver.
- To neglect the effect of the high conductive layers.
- Not to take into account the phase behaviour for large excitation pulses in the case of shallow aquifers.
- To assume the mean size of the pores within each water saturated layer to be homogeneous.

The only commercially available inversion software up today is based in a 1D model of the underground, with electrically resistive homogeneous layers parallel to the surface, using the same antenna for transmitter and receiver, homogenous pore size inside each layer, and do not look at the effects of shallow layers and phase shifts, though the geoelectrical model of the underground is considered. If MRS measurements are taken at locations where the assumptions made in the model do not fit the real situation, inversion using the inappropriate model may lead to errors in interpretation. When the lithology of investigated area is uniform, and the variations in the signal are due only to the variations in the water content, the errors produced by the use of an inappropriate model will be the same for all the MRS stations performed in the area, and the result can still be used for qualitative mapping of the aquifers (Legchenko *et al.,* 1997b; Rubio and Plata, 2005).

Topographical effects depend on the loop size, especially in the presence of shallow aquifers. For gentle slopes (less than 10 degrees) the errors introduced ignoring topography in the model are within the uncertainty of the measurements produced by the EM noise (Girard *et al.,* 2006). When using different transmitter and receiver antennas, the effect may be more important. An extended model has been presented for 2D and 3D that can take into account the topography (Rommel *et al.,* 2006).

When the same antenna is used as transmitter and receiver the signal will be in phase with the transmitter pulse, in resistive terrains, assuming excitation field amplitude smaller than geomagnetic field. However if the antennas are separated, in most locations there will be a nonzero out of phase component. The use of separated loops provides supplementary resolution to shallow aquifers. The observed signal is not only determined by the amount of separation of the antennas but also by the orientation of the geomagnetic field. Extended mathematical formulation of the response signal from MRS allowing the treatment of individual transmitter and receiver loops is given by Hertrich *et al.* (2005a,b); Legchenko (2003a,b) and Legchenko (2005), among others.

The MRS mathematical theory was first developed in free space, where the excitation field has no phase variation, and there is no difficulty in determining the effect of its active component, that is the one perpendicular to the static field. In the simplified model given by equation [3] for coincident transmitter and receiver, only amplitude inversion is used for conductive rocks, and an error may be produced. If conductive structures exist above or in the aquifer, the inversion with the simplified model will result into an underestimation of the water content. When the conductive layers are below the aquifer, an overestimation of the water content will be produced. In general, if conductivity effect is neglected in the inversion process, geometry and amount of water of the first aquifer may be erroneously approximated because of the phase shift, and the deeper aquifer may be not deduced from the $E_0(q)$ curve because of the screening effect (Trushkin *et al.,* 1995). The error may be more or less significant depending on the resistivity values and aquifer conditions, and the simplified model fits well and can be used quantitatively when the signal is long (gravel, coarse sand); when it is shorter (fine materials) only qualitative results can be obtained, and a more complex model should be used for inversion of complex signals (Legchenko, 2006).

Over conductive medium, the excitation field induces eddy currents, resulting in a new secondary field that has also to be considered in the model, taking into account the effect of the elliptically polarised field on the rotation of the magnetic moments of the protons to obtain the correct derivation of the magnetic resonance signal produced. When the conductivity of the rocks is large enough, such that the skin depth at the Larmor frequency is of the same order or smaller than the measurements depth, there are diffusive retardation time effects in the electromagnetic fields producing a nonzero out of phase component, even with coincident loops, and equation [1] should be modified:

$$E_0(q) = -\int_V \omega_L M_0 w(r) b_{R\perp}(r)\sin(0.5\gamma b_{T\perp}(r)q)$$
$$\exp(i(\varphi^{T(r)} + \varphi^{R(r)})dV(r)$$

where $\varphi^{T(r)}$ and $\varphi^{R(r)}$ are the difference in the phase between the excitation and the induced electrical currents at the receiver and transmitter (Valla and Legchenko, 2002). The exponential term makes explicit that the initial amplitude is complex and phase shifted with respect to the transmitter pulse. Conductive ground attenuates the initial voltage response at the receiver for several reasons (Weichman *et al.,* 2000): the intensity of the applied field of the transmitter and of the generated field by the protons is attenuated with depth, and the phase of the excitation field varies at different locations resulting in phase-shifted signal contribution that, when integrated, form an interference pattern at the receiver. This attenuation limits the depth at which water can be detected (Hunter and Kepic, 2003). For some authors, the inclusion of the imaginary part of the response significantly stabilizes the inversion (Weichman and Lavely, 1999), though for others real data inversion of complex signals will not stabilize the inversion, but destroy the results (Legchenko, personal communication).

It has been shown (Valla and Legchenko, 2002) that working with 1D model, elliptical polarization must be taken into account only for combined conditions: the subsurface is very conductive (<1 ohm.m), the inclination of the geomagnetic field is close to zero (equator), and groundwater is shallow. Computation results show that for low resistivities in 2D environment the elliptical polarization has to be taken into account. When resistivity of the rocks is over 5-10 ohm.m the approximate equation can still be used with linear polarization, the introduced errors being quite small. So, at least by now, in general the use of inversion amplitudes should be used, avoiding the inversion of complex signals. Examination of the phases provides additional information for estimating the reliability of the inversion result (Braun and Yaramanci, 2003).

Electrically conductive layers, variations in the geomagnetic field (when long time of measurements) and higher harmonics of the pulse can cause shifts in MRS signal, than can lead to erroneous results if inversion of the complex signal is carried out by taking into account phase variations caused only by electromagnetic shifts (Legchenko, 2003a). Water in deep aquifers mostly responds to the first harmonic of the excitation field, and the rest can be neglected. But in water close to the surface the tilt angle caused by higher harmonics is significant and must be taken into account. In consequence in case of exact resonance and resistive rocks the signal generated by deep aquifers may be real; in other circumstances the complex signal must be considered (Legchenko, 2003a). At large pulse moments water in shallow aquifers may generate a signal comparable in amplitude to the one produced by a deeper aquifer. When shallow aquifers are present, the interpretation using the simplified model can produce serious errors, because the model is not sufficiently accurate for describing the amplitude and phase behaviour of MRS signal for large pulse moments. For one aquifer system, interpretation using the simplified model may reveal an artefact that appears to be a deeper non existing aquifer. For multi layer systems the signal from a shallow aquifer may offset the signal from deeper ones, reducing the depth of exploration. The depth resolution could be significantly improved by considering not only the amplitude but also the phase of the signal. To improve the accuracy on MRS modelling in the presence of shallow aquifers, a revised mathematical model, which takes into account the higher harmonics of the transmitted pulse and a non zero frequency offset between the Larmor frequency and the pulse frequency is used, and equation [1] will become

$$E_0(q) = \omega_L \int_V \sum_k [B_{1k}(r)\exp(j\varphi_{0k}(r))$$
$$M_{\perp k}(r)/I_{0k}]w(r)dV(r)$$

where $B_{1k}(r)$ is the k^{th} harmonic of the alternating transmitted magnetic field component perpendicular to the geomagnetic field, φ_{0k} is the phase shift and $M_{\perp k}(r)$ is the transversal component of the magnetic moment of water molecules (Legchenko, 2003a; Legchenko, 2005; Shushakov, 2006). Thus, accuracy and reliability of MRS results can be improved by considering the frequency offset between the pulse frequency and the Larmor frequency and the few first harmonics of the pulse for modelling the MRS response. Using the enhanced model, the complex signal can be correctly inverted when relaxation times are long, because relaxation during the pulse is neglected what can produce errors when decay times are short (Legchenko, 2003b).

Regarding the assumption of mean pore size of water saturated rocks to be homogeneous within each water bearing layer, in reality each volume of the rocks (j) has its own process of relaxation of the hydrogen protons magnetization, with a different time T_j, according the pore dimensions. So the MRS signal is in fact the sum of the signals produced by each of these volume fractions:

$$e(t,q) = \sum E_{0j}(q)\sin(\omega_L t + \varphi)\exp(-t/T_j)$$

resulting in one multi exponential function (Dunn *et al.,* 2002; Mohnke and Yaramanci, 2005; Roy and Lubczynki, 2005, Plata and Rubio, 2005).

Acknowledgements

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