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Characterizing aquifers when using magnetic resonance sounding in a heterogeneous geomagnetic field

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ABSTRACT
It has previously been reported that the heterogeneity of the geomagnetic field disturbs the currently-measured free induction decay signal of magnetic resonance sounding (MRS). To overcome the limitation of MRS in a non-homogeneous geomagnetic field, we adapted the spin-echo methodology usually used at the laboratory scale and in boreholes. We present examples of measurements carried out in a sandy aquifer in southern India. The 15–25 m thick sand deposit overlays a gneissic basement. Two sources of geomagnetic field heterogeneity have been identified at this site, both affecting the geomagnetic field within the sandy aquifer: the gneissic bedrock and an intruded dyke into the bedrock. Spin-echo and free induction decay signals have been recorded at six locations. We found that the groundwater content, the thickness of the saturated aquifer and its transmissivity calculated with free induction decay measurements are underestimated compared to those derived from spin-echo measurements. The closer to the dyke the higher the underestimation. Time-domain electromagnetic measurements indicate that the aquifer is rather homogeneous at the site scale, as suggested by spin-echo results. We also found that a small heterogeneity of the geomagnetic field can go unnoticed, thus leading to an unknown mis-estimate of aquifer properties when using free induction decay measurements. Thus spin-echo measurements can be used to improve the accuracy of aquifer characterization when using MRS in geological contexts where geomagnetic field heterogeneity exists.

INTRODUCTION

Throughout history, groundwater has been integral to human life and to stable agricultural production. In our changing world, declining groundwater levels and quality are often reported and improved groundwater development and resources management based on comprehensive knowledge of aquifers is required (UNESCO 2009). Especially aquifer properties such as reservoir geometry, storage-related parameters and flow-related parameters are most needed for both well settings in complex environments and for supplying groundwater flow models that are commonly used for management purposes. One of the most reliable ways to gather aquifer properties is an *in situ* survey, such as drilling exploration boreholes and conducting hydraulic tests. However, *in situ* estimates of aquifer properties are often scarce because dense field surveys are expensive in terms of time and money.

Non-invasive surface geophysical methods capable of providing rapid and dense data coverage at an affordable cost can be very useful if they provide accurate estimates of aquifer properties. Compared to other geophysical methods, magnetic resonance sounding (MRS) is selective with respect to groundwater. Since the 1990s, this distinctive feature has been used by several scientists to assess the links between MRS parameters and the hydrogeological properties of aquifers (for example, Legchenko *et al.* 2002, 2004; Vouillamoz *et al.* 2002, 2007b, 2008; Lubczynski and Roy 2003; Rubio and Plata 2005). Acceptable estimates of flow-related parameters have been obtained in numerous field studies (Vouillamoz *et al.* 2007a) although Plata and Rubio (2008) pointed out that the commonly used conversion function between MRS and flow-related parameters is not always appropriate. More importantly, no clear quantitative link has been found between MRS parameters and aquifer storage-related parameters that are very sensitive in groundwater modelling and recharge estimate (Vouillamoz *et al.* 2005; Boucher *et al.* 2009).

This paper presents a new development in the MRS method that makes it possible to estimate MRS parameters with greater accuracy in an inhomogeneous geomagnetic field, thus resulting

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in better estimates of aquifer storage and flow-related parameters. The new methodology was tested in a coastal aquifer in southern India. Our results expand areas of MRS application because perturbations of the geomagnetic field caused by magnetic rocks may exist in many geological contexts.

MRS METHOD

MRS is the field-scale implementation of the nuclear magnetic resonance (NMR) method that is used for laboratory measurements and well logging in the oil industry. To conduct field measurements, a loop of electrical wire is laid out on the ground. A pulse of alternating current $i(t) = I_0 \cos(\omega t)$ is generated in the loop. The selectivity of the MRS method with respect to water is due to the specific frequency of the nuclear precession in the geomagnetic field (Larmor frequency):

$$f_L = \omega_L / 2\pi = \gamma B_0 / 2\pi , \quad (1)$$

where γ is the gyromagnetic ratio of the nuclei and B_0 is the magnitude of the geomagnetic field. For obtaining near resonance conditions the current frequency is set as $\omega \approx \omega_L$. The energizing pulse causes a deflection of the magnetic moments of hydrogen nuclei from their equilibrium position along the geomagnetic field B_0. When the pulse is switched off, the nuclei revert to their equilibrium position, creating an alternating electromagnetic field that could be measured with the same loop (coincident loop). To carry out a sounding, signals are recorded for different values of the pulse moment $q = I_0 \tau$ (where I_0 is the current and τ is the pulse duration): the higher the moment the deeper the investigation. In common geological environments the maximum investigation depth is 50–100 m.

The magnetic resonance signal of the hydrogen nuclei is generated by the transverse component of the nuclear magnetization $M_\perp$ and its amplitude can be calculated using an integral equation (Legchenko and Valla 2002):

$$E_0(q) = \omega_L I_0^{-1} \int_V B_1 \exp(j\varphi_0(\mathbf{r})) M_\perp(q, \mathbf{r}) \theta_{MRS}(\mathbf{r}) dV , \quad (2)$$

where B_1 is the transmitting magnetic field component perpendicular to the geomagnetic field B_0, φ_0 is the phase shift of the signal caused by the electrical conductivity of the rocks, r is the coordinate vector and $0 \le \theta_{MRS} \le 1$ is the water content. The solution of equation (2) is an estimate of the groundwater content θ_{MRS}(depth) in the subsurface. In our study we assume horizontal stratification and for resolving equation (2) we apply the Tikhonov regularization method (Legchenko and Shushakov 1998).

Free induction decay signal

In homogeneous geomagnetic fields, the transverse component of the nuclear magnetization can be measured with a single pulse sequence and is $M_\perp = M_0 \sin\left(1/2\, \gamma I_0^{-1} B_1 q\right)$ where M_0 is the macroscopic nuclear magnetization for the hydrogen nuclei. The recorded signal oscillates at the Larmor frequency and has an

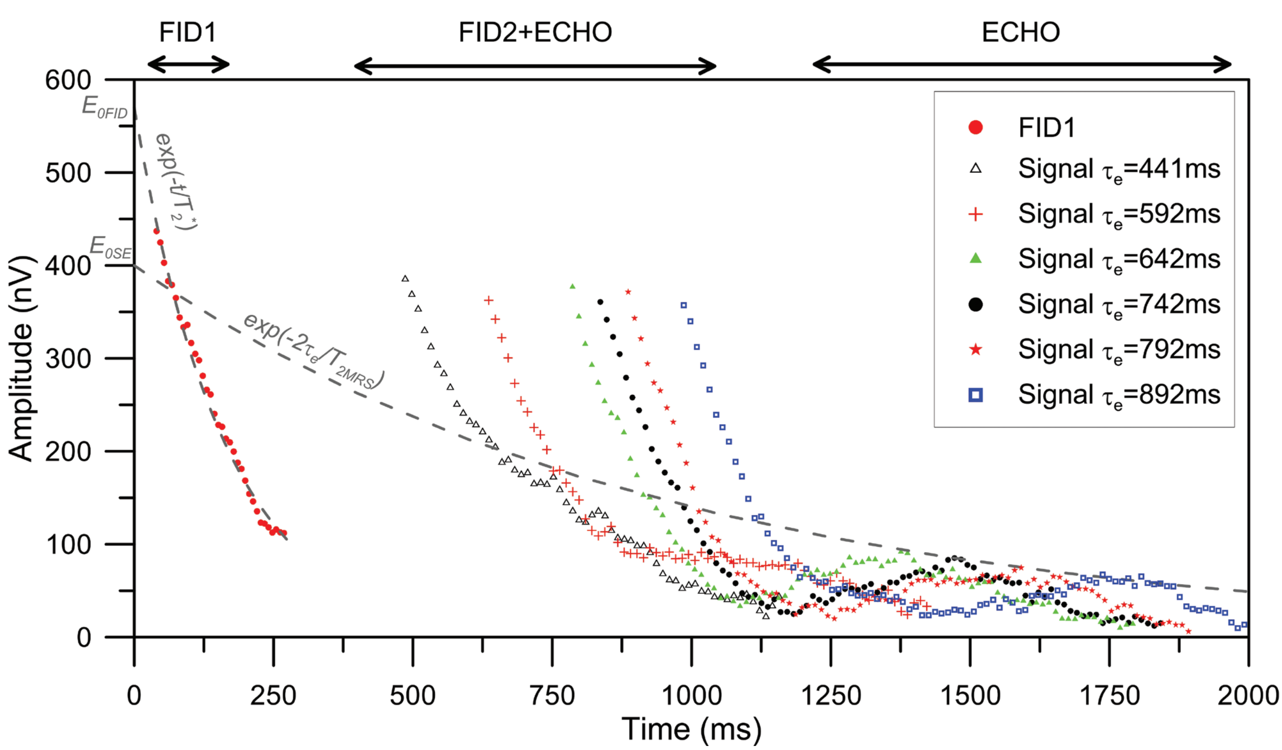

FIGURE 1

Time windows of recorded signals (q_1 = 110 Ams), site P1-sea.

exponential envelope that decays at time rate T_2^*. This so-called free induction decay (FID) signal is:

$$e(t,q) = E_{0FID}(q)\exp\left(-t/T_2^*(q)\right)\sin\left(\omega_L + \varphi(q)\right), \quad (3)$$

where E_{0FID} is the amplitude of the signal just after the energizing pulse has been turned off.

The MRS signal $E_{0mFID}(t)$ can be measured only after an instrumental delay ($\tau_{\text{dead}} \approx 40$ ms with the device we used for this study), which is needed for switching the loop from transmitter to receiver. The T_2^* value is calculated to fit measured records of the free induction decay (FID) signal and E_{0FID} is then extrapolated as (Fig. 1):

$$E_{0mFID}(q) = E_{0FID}(q)\exp\left(-\tau_{dead}/T_2^*(q)\right). \quad (4)$$

The T_2^* relaxation is caused by several phenomena:

$$1/T_2^* = 1/T_2 + \left(\gamma\Delta B_0\right) \text{ where } 1/T_2 = 1/T_{2B} + 1/T_{2S} + 1/T_{2D}. \quad (5)$$

T_2 is defined as the transverse relaxation time. $\gamma\Delta B_0$ represents the influence of the geomagnetic field heterogeneity ΔB_0 at the pore scale and it can be the dominant term of equation (5) in the presence of magnetic minerals (Roy *et al.* 2008). $T_{2\text{B}}$ is the decay time of bulk water, which is due to water molecule interactions and dependents on physicochemical properties of groundwater. T_{2S} is the surface relaxation time, which is determined by interactions between the hydrogen nuclei in water and the solid surface of the geological reservoir. $T_{2S} = \rho_2 \cdot V_{pore} / S_{pore}$ where ρ_2 is the surface relaxivity of the rock and V_{pore}/S_{pore} is the ratio of pore volume to pore surface. T_2^* is then related to the geometry of the water-filled pores, which is the basis of the use of the MRS signal to estimate groundwater flow parameters. $T_{2\text{D}}$ is the diffusion relaxation time that is related to the magnetic field gradient at the pore scale that results from the magnetic susceptibility contrast between the water and the rock. It follows from equation (5) that T_2^* is affected by the presence of magnetic field heterogeneity at all scales: the stronger the heterogeneity the shorter the value of T_2^*. When only part of the free induction decay signal has T_2^* less than about 40 ms (the dead time of the used instrumentation), the extrapolated E_{0FID} and then the groundwater content are underestimated. A slight underestimate of the water content can easily go unnoticed and a larger underestimate can be often explained by incorrect geological assumptions. At worst, if strong heterogeneity of the geomagnetic field reduces T_2^* to less than the threshold detection of the apparatus, the free induction decay signal is no longer detected although groundwater is present (Legchenko and Valla 2002; Roy *et al.* 2008).

Spin-echo signal

The spin-echo pulse sequence has been developed for NMR applications in heterogeneous magnetic fields (Hahn 1950). The MRS set-up for a spin-echo measurement is the same as for a free induction decay measurement but the measuring scheme consists of a double-pulse sequence. The first pulse q_1 is generated in the loop and the second pulse q_2 is applied after a time delay τ_e measured from the first pulse. During the delay between the pulses, magnetic moments phase out with the time constant T_2^*. The second pulse causes the hydrogen nuclei to refocus at time $t = 2\tau_e$ and a so-called nuclear spin-echo signal can be detected. In our experiment, we adapted the spin-echo sequence to the MRS field scale by using a measuring scheme where the second pulse moment q_2 is twice as large as the first pulse moment q_1 (Sushakov and Fomenko 2004; Legchenko *et al.* 2009). An example of recorded signals is presented in Fig. 1. The time origin $t = 0$ is the end of the first pulse. After an instrumental dead time of ~40 ms, the free induction decay signal created by the first pulse is recorded during 240 ms (FID1). A second pulse is then used after a delay τ_e from the first pulse. For demonstration purposes, six records corresponding to six values of τ_e are presented on the same figure. For the long inter-pulse delays (τ_e>~600 ms), the free induction decay signals created by the second pulse are clearly observed (FID2), followed by bell-shaped signals that arise at time $2\tau_e$, thus confirming spin-echo signals. For the shorter inter-pulse delay (τ_e<~600 ms), the signals recorded after the second pulse is a mixture of FID2 and spin-echo.

We assume the spectra of the spin-echo signal to be symmetrical and the delay between the pulses to be long enough to complete phasing out of the nuclear magnetization. Under near resonance conditions and respecting $q_1 = q_2/2 = q$, $M_\perp$ could be calculated as $M_\perp = -M_0 \sin^3\left(1/2\,\gamma I_0^{-1} B_1 q\right)$ (Bloom 1955). The spin-echo signal E_{SE} measured at time $2\tau_e$ after the first pulse is attenuated by the relaxation as follows:

$$E_{SE} = E_{0SE}\exp\left(-2\tau_e/T_{2MRS}\right), \quad (6)$$

where T_{2MRS} is the observed transverse relaxation rate.

The time constant T_{2MRS} and the initial amplitude of the signal E_{0SE} are calculated by performing measurements of spin-echo amplitude varying $\tau_e \cdot T_{2MRS}$ is calculated by fitting an exponential curve to the envelope of measured E_{SE}, and E_{0SE} is extrapolated using equation (6) (Fig. 1). During our experiments we did not observe non-exponential echo decay and we found that the exponential fit given by equation (6) provided reasonable results.

CONVERTING MRS PARAMETERS INTO HYDROGEOLOGICAL PROPERTIES

Storage parameters of saturated rocks

Using free induction decay measurements, the groundwater content θ_{MRS_FID} is defined as the fraction of water with a sufficiently long decay time constant T_2^* (>~40 ms) measured with the actual instrumentation (Legchenko *et al.* 2002). Bound water molecules attached to rocks due to forces of molecular attraction and water molecules precessing in a heterogeneous geomagnetic field generate short free induction decay signals that are undetectable with the available MRS instrument.

Using SE measurements, θ_{MRS_FID} is defined as the fraction of water located in a heterogeneous geomagnetic field that can be measured with the actual instrumentation (due to the bandwidth of the used MRS device, the spin-echo can be recorded if $T_2^* >\sim 10$ ms). If the geomagnetic field heterogeneity concerns the entire investigated volume, then θ_{MRS_SE} allows estimating the fraction of groundwater located outside the field of rock molecular attraction. Free water that can be displaced by gravity or hydraulic head gradient is known to hydrogeologists as the effective porosity n_e and so $\theta_{MRS} \approx n_e$. Another important storage-related parameter is the specific yield S_y, which is the ratio of the volume of water that releases an unconfined aquifer by gravity forces when drained, to the volume of the drained aquifer. Specific yield and effective porosity are different parameters because a portion of the effective porosity cannot be drained by gravity forces (capillary water). n_e and S_y have comparable values for coarse grain rocks but n_e differs from S_y in fine-grained rocks and $\theta_{MRS} \geq S_y$ (Lubczynski and Roy 2007).

The free hydrostatic water column that provides a volumetric estimate of free groundwater can be calculated as $H_w = \theta_{MRS}\Delta z$, where the water content θ_{MRS} and the aquifer thickness Δz are derived from MRS measurements. Compared to groundwater content, the free hydrostatic water column has the advantage of being less sensitive to the geophysical equivalence (Legchenko *et al.* 2004).

Actually, the quantitative relationship between the specific yield estimated from pumping test analysis and θ_{MRS_FID} was found to be poor. Vouillamoz *et al.* (2005) reported that the difference between θ_{MRS_FID} and S_y was on average ±79% in weathered granite of Burkina Faso (using a linear relationship between θ_{MRS_FID} and S_y) and Boucher *et al.* (2009) found an average difference of about ±60% in sandstone of Niger (using a non-linear relationship).

Flow parameters of a saturated aquifer

The hydraulic conductivity K is the coefficient that links the hydraulic head to the flow through a porous medium. K is linked to both water and rock properties as $K = kg/v$, where k is the permeability of the rock reservoir, g the acceleration of gravity and v the kinematic viscosity of the water that is strongly temperature dependant. Another flow parameter is commonly used for hydrogeological studies because it gives a more complete characteristic of an aquifer flow that considers also the flowing area: the transmissivity $T = Ke$ where e is the saturated thickness of the aquifer. The basis of the relationship between MRS and flow-related parameters is the dependency of MRS decays on the geometry of water-filled pores: inspired by hydrogeological works that proposed relationships between the permeability and the size of the rock's grain (Fetter 1994) the basic equation used in the MRS field is (Seevers 1966):

$$k = C_k \theta_{NMR}^a T_k^b, \quad (7)$$

where θ_{NMR} is the NMR porosity, T_k is a NMR decay time parameter and C_k, a and b are parametric factors. Equation (7) has been adapted to the MRS scale of measurement and is usually used as (Legchenko *et al.* 2002):

$$T_{MRS} = K_{MRS}\Delta z = C_T \theta_{MRS}^a T_d^b \Delta z \quad (8)$$

with $a = 1$ and $b = 2$ as proposed by Seevers (1966) or $a = 4$ and $b = 2$ as proposed by Kenyon *et al.* (1988).

Using free induction decay signals, T_d is primarily T_{1_MRS} or T_2^* if T_{1_MRS} is not available (Vouillamoz *et al.* 2002). Comparison between transmissivity estimated from pumping test analysis (T_{pt}) and transmissivity calculated from equation (8) showed an average difference usually ranging in-between –50% and +100% (Vouillamoz 2007a). This uncertainty is considered as acceptable because it is about of the same order as the uncertainty on T_{pt} calculated from field measurements. C_T is usually calculated as:

$$C_T = \sum_{i=1}^{n} T_{pt\_i} \bigg/ \sum_{i=1}^{n} F_i, \quad (9)$$

with $F = \theta_{MRS} \cdot T_d^2 \cdot \Delta z$. However, C_T exhibits a wide range of values depending on the local geology, which makes equation (8) usable only after parametrization. Moreover, the C_T value seems to be also affected by geological heterogeneities within the same aquifer as reported by Plata and Rubio (2008). These authors found links between C_T and F and they proposed to calculate C_T as:

$$C_T = mF^{-n}, \quad (10)$$

using three different pairs of values (m_i, n_i) according to three ranges of the F value.

FIELD EXPERIMENT

Area of investigation

Field measurements were conducted in December 2008 on a barrier beach in south-western India, in the state of Karnataka (Fig. 2). At this site, named Sasihithlu, the subsurface consists of sand deposits that overlie a granitic-gneissic basement of Archean age. The sand deposits originated from river sediments and the littoral currents are mainly composed of medium-to-coarse sized grains of quartz and contain shell fragments (Jayappa and Subramanaya 1994). The thickness of the sand deposits ranges from 15–25 m and the static water level measured in numerous wells ranges from 0.5–2 m below the ground surface. The Sasihithlu barrier beach was selected because of the presence of two magnetic sources at the site scale: the granitic-gneissic bedrock that creates a geomagnetic field gradient at depth within the sandy aquifer and a dyke of magnetic rocks located in the south of the area that creates a geomagnetic field gradient in the south-north direction (Fig. 2).

Method and materials

The geomagnetic field was measured at 1.5 m above the ground surface with a portable G856AX magnetometer from Geometrics.

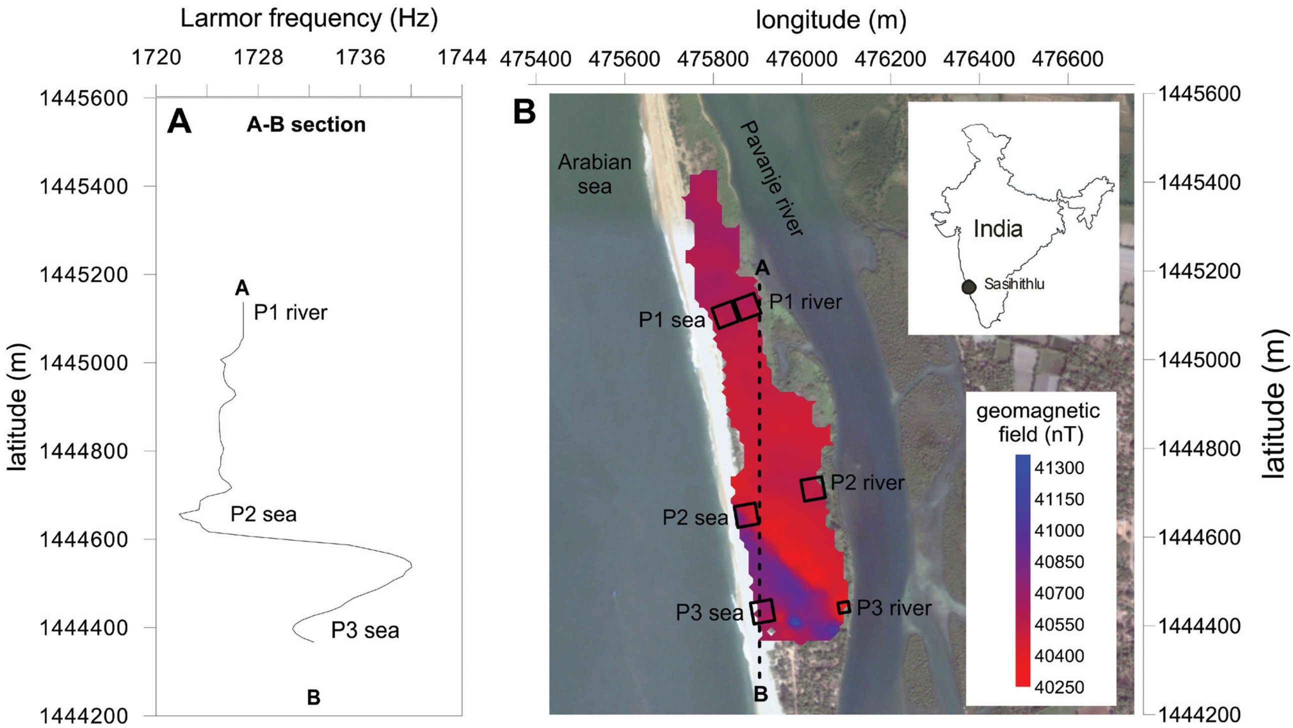

FIGURE 2
Location of test sites and geomagnetic field measurements. a) Cross-section of Larmor frequency; b) map of geomagnetic field intensity.

The measurement revealed a magnetic anomaly of about 1000 nT over a distance of 200 m (Fig. 2b). The direction of the anomaly (north 315°) is also the main direction of 100 to 200 km long lineaments reported in the area by Ganesha Raj (1994). Thus, the magnetic anomaly is probably caused by a dyke intruding into the granitic-gneisses basement.

The magnetic susceptibility of rocks was assessed with a susceptibility meter (Kappa meter KT-6®). Sand samples were collected at numerous locations at the ground surface and down to a maximum depth of 2.3 m using an auger. Measurements on the basement rock were conducted on an outcrop and on numerous cuttings of boreholes located a few kilometres from the site. All sand samples were non-magnetic (susceptibility less than 10^{-5} SI) but the granitic-gneisses basement exhibited susceptibility ranging between 10^{-4} and 10^{-2} SI. Such values of magnetic susceptibility have been previously reported to disturb the free induction decay record (Legchenko *et al.* 2002; Vouillamoz *et al.* 2007a; Roy *et al.* 2008).

MRS measurements were carried out with the Numis^{plus} apparatus from Iris Instruments. A square-shaped loop measuring 50 m long per side and with two turns was used (except at location P3-river where a square-shaped loop of 25 m long per side was used). At every location, full soundings (that is 12–16 pulses) were carried out in both free induction decay mode using a pulse duration τ_1 = 10 ms and in spin-echo mode, using a first pulse duration of τ_1 = 10 ms and a second pulse of τ_2 = 20 ms applied in anti-phase. Three full spin-echo soundings were usually carried out at every location using three inter-pulse delays. The inter-pulse delays were selected according to T_2^* and T_{2MRS} values and ranged between ~200 ms < τ_e <~800 ms (see examples Fig. 1). Free induction decay and spin-echo measurements were interpreted with Samovar software v 11.3 (Legchenko *et al.* 2009).

Rock electrical resistivity was also assessed with time-domain electromagnetic (TDEM) soundings, using the TEM-FAST 48HPC instrument from AEMR.

RESULTS

For demonstration purposes, we present detailed results of measurements located at about 500 m from the magnetic anomaly (location P1-river). We then compare storage and flow-related parameters of the six locations calculated from free induction decay and from spin-echo signals.

P1-river location

The non-homogeneity of the geomagnetic field estimated from magnetometer measurements at the scale of the MRS loops was 50 nT (which is a Larmor frequency variation of about 2 Hz). Such a variation over a distance of 60 m does not point out a significant field gradient in the horizontal plane. Thus, the field gradient created by the magnetic dyke probably does not extend up to this location.

Free induction decay signals of a good quality have been recorded (average signal-to-noise ratio of 35, Fig. 3) without any indication of geomagnetic field heterogeneity. Indeed, consider-

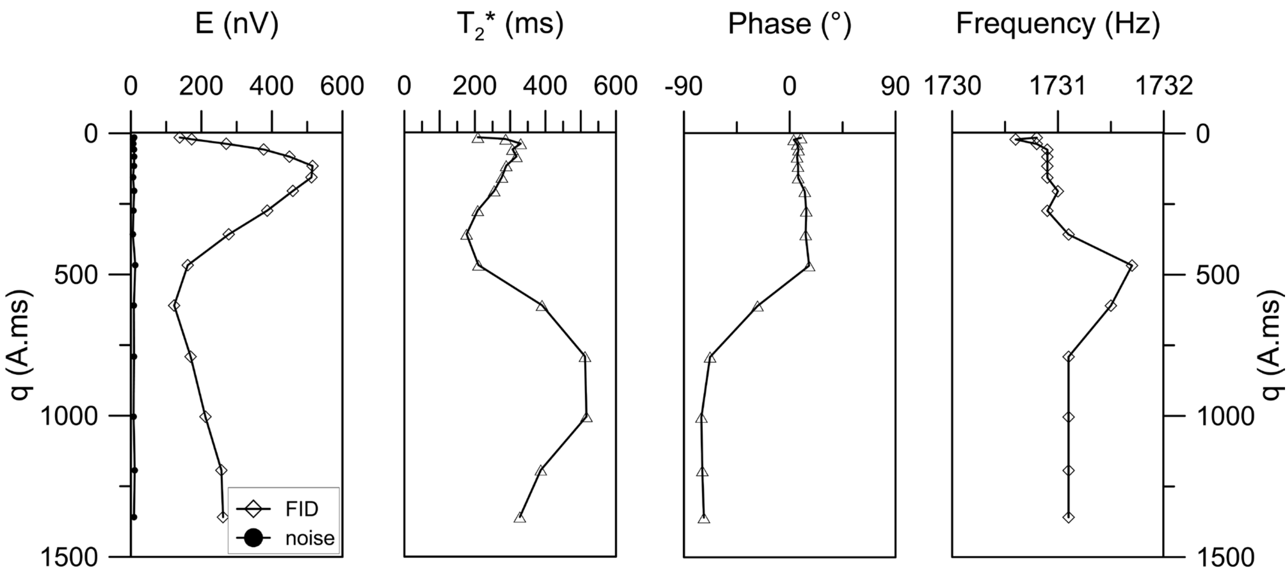

FIGURE 3
MRS-FID, site P1-river.

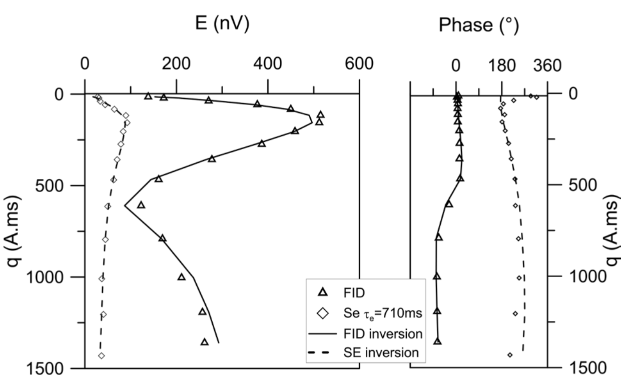

FIGURE 4
MRS and inversion fits, site P1-river.

ing data up to q ~ 600 Ams (signal recorded with q ~ 600 Ams is assumed to be generated by the shallow aquifer (Legchenko 2004)), the frequency is stable with q values and the average T_2^* of 260 ms is an expected value for medium to coarse sands. The smooth decrease of T_2^* with increasing q can easily be explained by possible lithological changes.

Although free induction decay signals do not present any indication of geomagnetic field non-homogeneity, clear spin-echo signals have been recorded with three values of inter-pulse delay τ_e. The full spin-echo sounding carried out with τ_e = 710 ms is presented in Fig. 4. The quality of the spin-echo records is good, with an average signal-to-noise ratio of 3.3. T_{2MRS} is calculated using equation (6), considering the largest signal amplitudes of the soundings (116 Ams < q < 273 Ams). According to pulse moment values, T_{2MRS} values range in-between 900–1200 ms, with an average of T_{2MRS} = 1000 ms, which is used as a fixed value to inverse the spin-echo (SE) soundings (Fig. 4).

Both inverted amplitude and phase fit the observed data well, with an RMS of 2% for MRS-FID as well as for MRS-SE. The average water content derived from free induction decay and spin-echo soundings exhibits similar values of 22% and 24%, respectively (Fig. 5a). However, extrapolating the initial amplitude of the spin-echo signal and then calculating the water content is strongly affected by the T_{2MRS} value because the extrapolation concerns a long period of time ($2\tau_e$ = 1420 ms in our example). A simple estimation shows that assuming the range 900 ms < T_{2MRS} < 1200 ms, the MRS-SE water content varies between 21–34%. The FID initial amplitude is less sensitive to the inaccuracy of T_2^* estimation because the extrapolation is conducted over only 40 ms (dead time of the Numis apparatus): the MRS-FID water content ranges in-between 20–24%, considering 200 ms < T_2^* < 300 ms. Note that the error in water content due to the relaxation during the pulse is negligible because the pulse duration τ_q is short compared to T_2^* (T_2^* ~25τ_q , Walbrecker *et al.* 2009).

The MRS results show that the depths to the substratum obtained from FID and SE measurements are different: 12.5 m and 21 m below ground surface respectively (Fig. 5a). For both free induction decay and spin-echo measurements the maximum depth of investigation defined after Legchenko *et al.* (2008) is deeper than 22 m and hence, we consider the results as reliable. The TDEM results show that the depth to the substratum ranges between 20–22 m below ground surface (Fig. 5b), thus indicating that the aquifer thickness is underestimated by free induction decay measurements probably because of the geomagnetic field gradient created at depth by the gneissic basement. This hypothesis also explains the decrease of T_2^* with increasing q values (Fig. 3). Logically, the column of hydrostatic free groundwater $H_w = \theta_{MRS}\Delta z$ estimated from free induction decay measurements is 25% lower than that estimated from MRS-FID.

The MRS transmissivity is calculated using both free induction decay (FID) and spin-echo (SE) results:

$$T_{MRS\_FID} = 8.610^{-7} \cdot \theta_{MRS-FID} \cdot T_2^* \cdot \Delta z_{FID}$$
$$T_{MRS\_SE} = 1.210^{-8} \cdot \theta_{MRS-SE} \cdot T_{2MRS} \cdot \Delta z_{SE}. \qquad (11)$$

The values of C_T in equation (11) have been calculated at the neighbouring location P1-sea, where the aquifer transmissivity is known ($8.610^{-2} m^2/s$). At location P1-river, the transmissivity values calculated with both free induction decay and spin-echo results are similar (difference of 5%). Although Δz_{FID} is strongly underestimated, transmissivity is well estimated using free induction decays, probably because of the C_T effect in equation (11): indeed, free induction decay signals are similarly affected by the geomagnetic field gradient at the P1-river location and at the neighbouring parametrization P1-sea location (P1-sea is adjacent to P1-river, Fig. 2).

Storage and flow parameters of Sasihithlu beach barrier

Measuring both FID and SE signals at six locations gives an opportunity to compare the hydrogeological parameters estimated from these data. The transmissivities of all locations are calculated using the same values of C_T-derived at the P1-sea location (equation (11)).

The hydrostatic column of free groundwater and the transmissivity obtained from SE measurements are quite homogeneous at the site scale (average deviation of 23% and 27% respectively) but both storage and flow parameters derived from free induction decay measurements are clearly heterogeneous (average deviation of 102% and 61%, respectively, Fig. 6). The geological assumption of an homogeneous aquifer all over the investigated area is confirmed by the TDEM measurements, which indicate an homogeneous saturated layer: the depth to the bedrock is 20 ± 1.3 m below the ground surface and the resistivity of the aquifer saturated with seawater is 0.9 ± 0.05 Ωm.

Moreover, the value of the free hydrostatic column of water estimated from free induction decay measurements is linked to the distance to the magnetic dyke: the closer to the dyke the higher the underestimate (Fig. 7a). The impact of the distance to the source of the geomagnetic field heterogeneity is strong on ($\theta\Delta z$) but it is less on the transmissivity, probably because of the damping effect of the C_T parameter: the ratio of transmissivity T_{MRS_FID}/T_{MRS_SE} is about constant, except for very close to the dyke (less than 10 m, Fig. 7b).

DISCUSSION

Estimates of hydrogeological parameters

Free induction decay and spin-echo soundings located further than about 100 metres from the dyke provide similar results (Table 1): $\theta_{FID} \sim \theta_{SE}$, the free hydrostatic column of water is only slightly

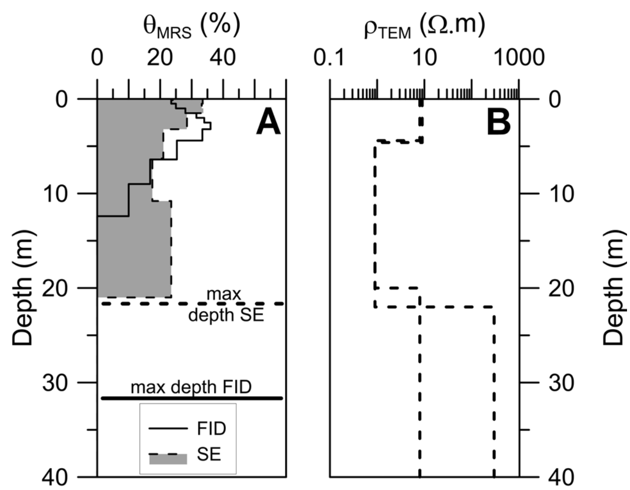

FIGURE 5
Inversion results of MRS, site P1-river. a) MRS results; b) TDEM equivalent solutions.

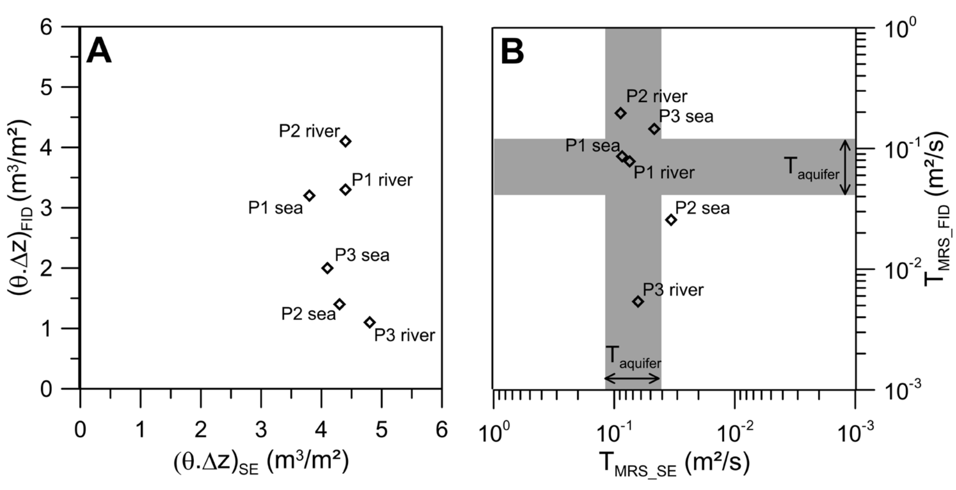

FIGURE 6
Comparison of storage parameters (a) and flow parameters (b) obtained from SE and FID soundings. The range of the $T_{aquifer}$ value is calculated at the P1-sea location.

underestimated with MRS-FID because $\Delta z_{FID} < \Delta z_{SE}$ and $T_{MRS\text{-}FID} \sim T_{MRS\text{-}SE}$ because the free induction decays are similarly affected from one site to the other and the transmissivity is calculated by a unique C_T value. Further than 100 m from the dyke, the geomagnetic field heterogeneity is small and probably results from the only vertical gradient created by the bedrock at depth. However, free induction decay and spin-echo soundings located next to the dyke provide very different results: the hydrostatic column of water is mis-estimated with MRS-FID by 80% on average and the transmissivity cannot be reliably estimated with a unique C_T constant (Table 1). Next to the dyke, the geomagnetic field heterogeneity is much stronger and probably results from both vertical and horizontal field gradients. Note that close to the dyke the transmissivity is better estimated when using several C_T , as proposed by Plata and Rubbio (2008), thus indicating that the misfit previously reported between transmissivity estimated from MRS-FID as compared to pumping tests analysis is probably also controlled by geomagnetic field heterogeneities (Table 1).

Estimates of T_{2MRS} and θ_{SE}

Considering the surveyed site, the use of SE results clearly improves the hydrogeological characterization of the aquifer compared to the use of free induction decay results. However, the uncertainty in spin-echo water content (θ_{SE}) is strongly dependent on the uncertainty in the T_{2MRS} estimate. For investigating errors in an estimate of θ_{SE} , we conduct a numerical modelling. We consider a saturated aquifer of 20% of water content lying between 1–10 metres below the ground surface. The use of a square-shaped loop of 50 metres long per side is simulated with a Larmor frequency of 2000 Hz and a geomagnetic field inclination of 50°. The maximum initial signal amplitude E_{0SE} of the modelled sounding is 150 nV at $q \sim 200$ Ams. We consider three cases corresponding to different contexts that can be met on the field: 1) no free induction decay is recorded because the geomagnetic field is strongly heterogeneous at the sounding scale and T_{2MRS} is long, that is $T_2^* <$ 40 ms and T_{2MRS} = 1000 ms, 2) the same context but shorter T_{2MRS}, that is $T_2^* <$ 40 ms and T_{2MRS} = 500 ms and 3) long free induction

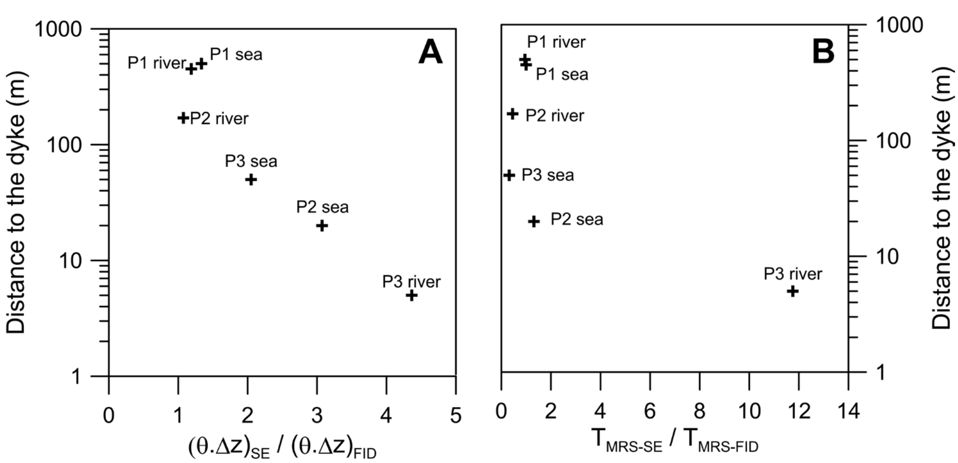

FIGURE 7
Comparison of MRS results as a function of the distance to the dyke. a) Hydrostatic column of free groundwater; b) aquifer transmissivity.

TABLE 1
MRS parameters and hydrogeological properties obtained from free induction decay and spin-echo measurements

	Locations further than 100 m to the dyke (P1-sea, P1-river, P2-river)		**Locations closer than 100 m to the dyke (P2-sea, P3-sea, P3-river)**	
	Average	Average deviation	Average	Average deviation
θ_{SE}/θ_{FID}	1.08	2%	2.2	61%
$\Delta z_{SE}/\Delta z_{FID}$	1.3	25%	1.5	35%
T_{2MRS}/T_2^*	5.5	82%	6.3	167%
$(\theta_{SE}\Delta z_{SE}) / (\theta_{FID}\Delta z_{FID})$	1.2	9%	3.2	80%
T_{MRS_SE}/T_{MRS_FID} (single C_T, equation (9))	0.8	23%	4.5	490%
T_{MRS_SE}/T_{MRS_FID} (3 C_T values, equation (10))	1	5%	1	34%

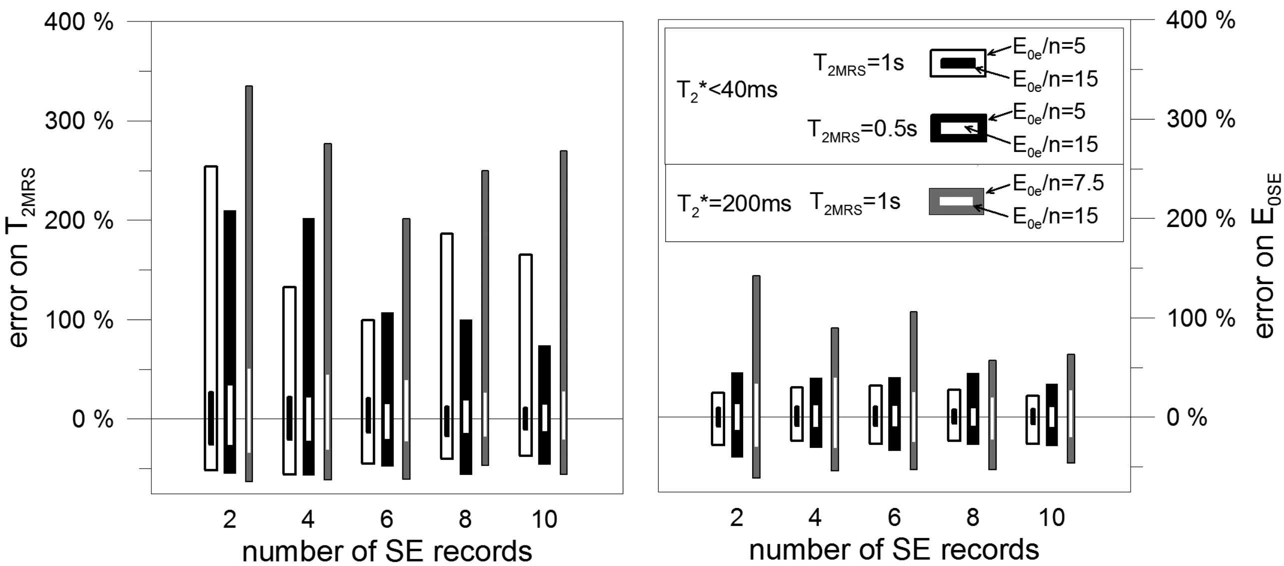

FIGURE 8
Uncertainty on T_{2MRS} and E_{0SE}.

decay signals are recorded together with spin-echo signals ($T_2^* = 200$ ms and $T_{2MRS} = 1000$ ms). This third context is critical for spin-echo measurements because the spin-echo signal can be corrupted by a free induction decay signal if the delay τ_e between the pulses is not long enough (Fig. 1).

We check the uncertainty of recovered T_{2MRS} and E_{0SE} values as a function of the number of recorded SE amplitudes corresponding to different delays between the pulses and to the signal-to-noise ratio of the records. Our simulations consider a number of spin-echo amplitudes (E_{SE},τ_e) ranging between 2–10 and signal-to-noise ratio $E_{0SE/noise}$ ranging from 3–15. Results of the simulation are presented in Fig. 8. We observe that: 1) the higher the signal-to-noise ratio the lower the uncertainty whatever the number of SE records: for a high $E_{0SE/noise}$ ratio, uncertainties on both T_{2MRS} and E_{0SE} are always fewer than ±20%; 2) for low $E_{0SE/noise}$ ratio, the higher the number of spin-echo records the lower the uncertainty; 3) uncertainty on T_{2MRS} is always higher than that on E_{0SE}.

Our survey in India corresponds to the modelled case with long T_{2MRS} decay rates, high values of ratio $E_{0SE/noise}$ and the number of spin-echo records being 2 or 3, depending on the location. The numerical modelling results indicate that the maximum uncertainties are ±18% and ±11% on T_{2MRS} and E_{0SE}, respectively. These uncertainties on MRS parameters imply uncertainties on the output hydrogeological properties, which are about ±35% on the MRS water content and ±85% on the MRS transmissivity.

CONCLUSION

We have shown that a heterogeneity in the geomagnetic field can cause severe underestimates of both storage and flow-related parameters calculated from FID measurements. During our study we adapted and tested the spin-echo technique to the MRS method. When the geomagnetic field is not homogeneous, our results suggest that the SE method provides more accurate results than the currently used FID method. However, the uncertainty on SE results can be high if the signal-to-noise ratio is low. Finally, we observed that there is not always an obvious indication of geomagnetic field heterogeneity, resulting in unnoticed mis-characterization of an aquifer using FID measurements. Thus, we recommend checking the presence of SE signals even if FID signals are recorded, when perturbations of the geomagnetic field are geologically possible as in igneous rocks, metamorphic rocks and all sedimentary rocks formed by the deposition of possibly magnetic materials.

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