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Aquifer Characterisation by Magnetic Resonance Field and Laboratory Measurements

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SUMMARY

In order to evaluate the potential and reliability of surface-NMR measurements for aquifer characterisation we compare a field dataset collected on a well known test site with sandy aquifers to the laboratory-NMR and hydrogeological measurements of porosity and hydraulic conductivity using cores samples and grain size analyses. The decay time distribution derived from the QT-inversion of the surface-NMR data shows good agreement to laboratory-NMR results. A comparison of the water contents and T2* decay times achieved from different inversion approaches, i.e., using lately developed QT-Inversion and SAMOVAR show similar results. For most parts of the subsurface, the hydraulic conductivity from field and laboratory-NMR agreed very well with values estimated from grain size analyses or measured in the laboratory. Disagreements especially in parts of the first aquifer need to be further evaluated.

Introduction

The characterisation of aquifers using geophysical methods promises to complement hydrological field methods in a fast and cost-efficient way. The most important parameters in hydrogeology to characterise aquifers are porosity and hydraulic conductivity. The nuclear magnetic resonance (NMR) method has shown to have promising abilities. The amplitude of the NMR signal can be directly used to determine the water content in the subsurface. Its decay time is related to the inner surface of the medium which can be linked to pore size distribution and hydraulic conductivity. But the performance of the surface-NMR method is not fully assessed yet and its potential is not exploited. Therefore, further examinations using field results, laboratory and borehole measurements under controlled conditions are needed.

The porosity (Φ) of a medium is defined as the percentage of voids in the material. Under saturated conditions it is equal to the volumetric water content (Θ). In unconsolidated sediments where all pores are connected this water can be divided into bound (Θ_B) and mobile (Θ_M) water. Mobile water is drainable by gravity and therefore accessible to hydrological processes. In sandy aquifers this represents almost all present water. The amount of water detectable by NMR (Θ_{NMR}) depends on the decay time of the water, the dead time of the instrument and the noise conditions. For high-frequency laboratory-NMR the dead time is in the range of several μs and thus bound water with decay times typically $< 33 ms$ can be detected ($\Theta_{NMR} \approx \Theta$). For field or surface-NMR the instrumental dead time range between 12 and 40 ms dependent on the used filter properties and thus most of the bound water is undetectable ($\Theta_{SNMR} \approx \Theta_M$).

The hydraulic conductivity (K) is a fundamental hydrological parameter to describe flow processes in aquifers used e.g., to determine the production rate of a well or the propagation of contaminants. K of a porous medium depends generally on the dimension of the flow channels. In hydrogeology various empiric formulas exist to link K to grain size distributions of sandy aquifers such as after Hazen (1892)

$$K = 0.0116 \cdot d_{K10}^2, \quad (1)$$

where d_{K10} is the grain size at the intersection of the 10% ordinate with the cumulative curve. The T_2 decay time distribution obtained from NMR can be converted to the pore size distribution by the surface relaxivity of the medium (Kleinberg et al., 1994). Therefore in analogy to (1), empiric formulas have been found to calculate K-values from Φ and T_2 such as Seevers (1966)

$$K = C_K \Phi T_2^2. \quad (2)$$

The factor C_K is lithology and site specific and depends mainly on the surface relaxivity of the medium. A range of lithology specific C_K values is summarized by Mohnke and Yaramanci (2008). Using surface-NMR the measured decay time is affected by dephasing processes due to inhomogeneities of the magnetic field and thus only the shorter decay time T_2^* is determined. Legchenko et al. (2002) transferred the formula (2) from the laboratory to surface-NMR by using T_2^* . Because of the multi-exponential behaviour of NMR signals from heterogeneous or layered media a distribution of T_2 or T_2^* times can be estimated using appropriate inversion schemes. To transfer this T_2 distribution to a representative T_2 time we used the log-mean T_2 value after Straley et al. (1994).

Test site and methods

The test site Schillerslage is located north of Hannover. It has a typical lithology for the N-Germany quaternary sediment basin (Figure 1). Two sandy layers, separated by a fine-grained till layer, are overlying cretaceous marls. The first layer has a thickness of 12 to 13 m and can be subdivided in an upper part with medium sand (0-5 m) and a lower section with more coarse sand (5-13 m). The aquiclude below consist of till reaching from 12 to 16 m. The second sandy layer reaches from 16 to 21 m and consists of slightly limy medium grey sand. The water table is approximately 1 to 2 m below surface.

In contrast to laboratory-NMR, surface-NMR is usually applied over an inhomogeneous, mostly horizontally layered medium. Therefore a sounding technique with varying pulse length and thus penetration depth is used. An inversion scheme is necessary to separate the NMR signal from layers at different depth. To show the effect of different inversion schemes on the interpretation of surface-NMR data we compared the commonly used SAMOVAR (Iris Instruments, 2000) software with the recently developed QT-inversion scheme from Müller-Petke and Yaramanci (2010). Instead of the initial amplitude approach which fits the decay curve of each pulse moment independently, the QT-inversion scheme uses the whole available data cube consisting of the whole measured T_2^* decay curves at each pulse moment. This improves the signal quality by taking the coherence of close-by pulse moments into account. In contrast to the mono-exponential approach e.g. used by SAMOVAR, QT additionally allows for an estimation of a T_2^* distribution by fitting for a multi-exponential decay.

The presented surface-NMR sounding was done using a Numis Lite equipment with a 2 turn circular loop of 50 m diameter. The measurement was centred on the presented borehole. For the laboratory-NMR measurements we used a MARAN equipment applying a CPMG pulse sequence to measure T_2 . The used samples have a diameter of 5 cm and were extracted from bigger core samples to disturb the matrix of the sand as little as possible. Even they were still wet we had to resaturate them with groundwater. Note that enclosed air within the samples was inevitable during the resaturation process.

As references for the porosity we measured the bulk density of dried samples assuming a grain density of 2.65 g/cm^3 . K-values were estimated after Hazen (1892) using grain size analyses. Measurements of K were done on undisturbed sand samples ($6 \times 4 \text{ cm}$) using the constant head method. On core samples ($8 \times 100 \text{ cm}$) the falling head method was applied.

Comparing surface-NMR with laboratory-NMR

To verify the T_2^* distributions obtained from the QT-inversion of our surface-NMR data we compare the results with T_2 distributions from laboratory-NMR measurements (Figure 1).

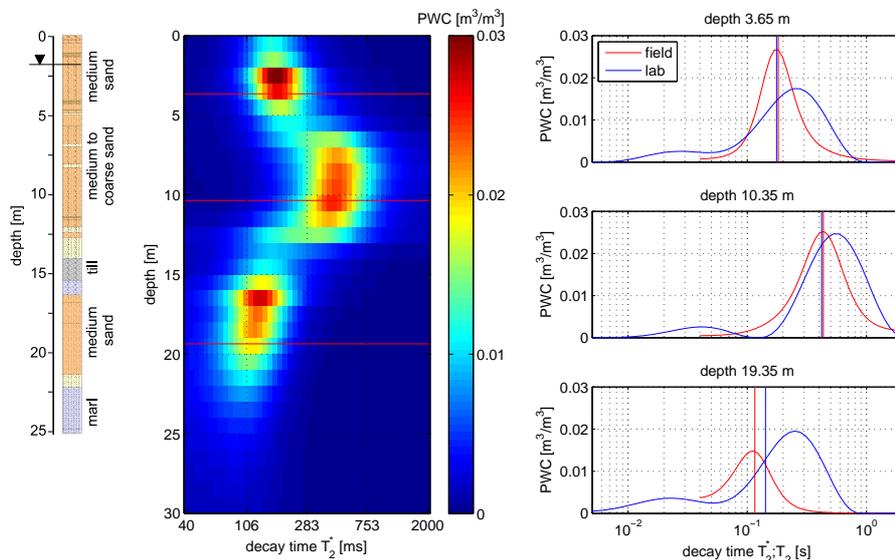


Figure 1 Borehole description (left) and partial water content (PWC) distribution of T_2^* derived from QT-inversion of surface-NMR data (centre). T_2 decay time distributions (right) from laboratory-NMR on samples (blue) compared to the QT-inversion result at respective depth (red). The distributions were fitted using approximately 25 decay times per decade. The vertical line marks the respective log-mean decay time.

The distribution of the partial water content derived from the QT-inversion divide the subsurface into

three layers with different T_2^* times. This is in good agreement to the geometry and lithology gained from the borehole showing sandy aquifers with different grain sizes. Therefore longer T_2^* times correlate qualitatively with coarse material.

The comparison of the decay time distribution from field (T_2^*) and laboratory (T_2) measurements at samples from selected depth confirm qualitatively this separation. However the maximum peaks of the T_2 time distributions seem to be shifted to longer times by a factor of approximately 2. One explanation for this is the dephasing effect due to inhomogeneities of the magnetic field which reduce T_2^* compared to T_2 . Additionally another explanation might be that the measured T_2 time is affected by the different frequencies of both NMR measurements which are $2 \cdot 10^3$ Hz for surface-NMR and $8 \cdot 10^6$ Hz for laboratory-NMR thus differing by 3 magnitudes. This effect would be in analogy to Mattea et al. (2004) who showed that the frequency dependency of T_1 measurements has the same sign and is of a similar magnitude. The fast decaying water with T_2 times between 10 to 60 ms as detected by laboratory-NMR can not be resolved by surface-NMR. This is due to the higher instrumental dead time and the worse signal-to-noise ratio of field measurements. Additionally, the observed shift of T_2^* compared to T_2 makes the detection even harder. However, the calculated log-mean of both decay time distributions fit well. The significant shift to shorter T_2 times of the log-mean compared to the maximum of the distribution measured by laboratory-NMR data is due to averaging characteristic of the log-mean and the additional water signal resolved at shorter decay times. This explains the shift but the similarity to the surface-NMR data needs to be further studied.

Comparison of aquifer parameters derived from different methods

After we verified the quality of the received T_2^* times from the QT-inversion we compared the results with data obtained from other inversion schemes and hydrogeological measurements, e.g. hydraulic conductivities measured on samples and obtained from grain size analyses (Figure 2).

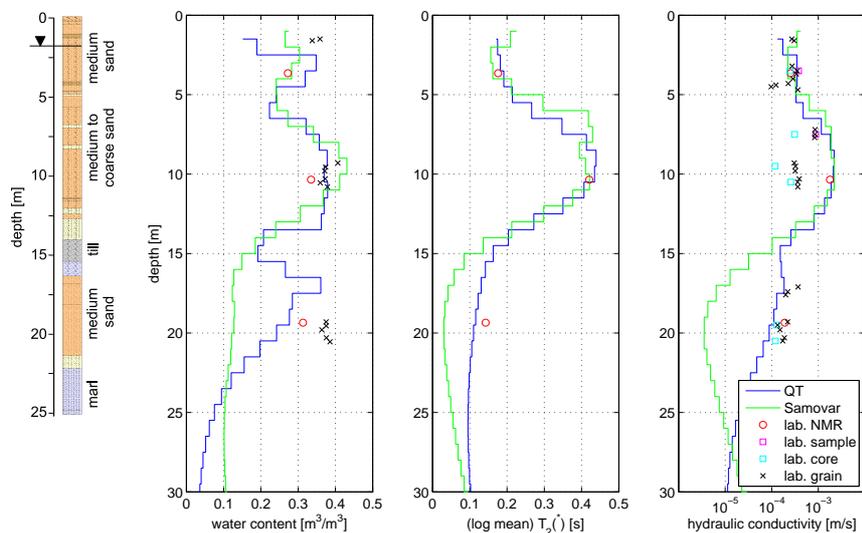


Figure 2 Comparison of total water content (left), (log-mean) T_2^* (centre) and hydraulic conductivity (right) over depth derived from different methods and inversion schemes.

For both parts of the upper aquifer down to 13 m, the vertical distribution of the water content obtained from the QT-inversion agrees very well with the result from SAMOVAR and the geometry and lithology shown by the borehole. The Θ_{SNMR} is confirmed by the porosity calculated from grain analyses. The Θ_{NMR} from selected laboratory samples seems to underestimate the water content in the aquifers which might be due to the problem of fully saturate the samples. The lower aquifer between 16 and 21 m was only detected from the surface-NMR data by the QT-inversion. The sharp lower boundary of

the aquifer to the underlying marl is smoothly resolved spreading the water over a thicker layer and thus underestimating the actual water content.

The different decay times obtained from the QT-inversion (log-mean T_2^*) and SAMOVAR (mono-exponential fitted T_2^*) agrees very well for the upper aquifer. As for the water content the lower aquifer is only resolved by the QT result. There the data from laboratory-NMR (log-mean T_2) seems to confirm the QT result as already shown in Figure 1.

Finally, the hydraulic conductivities deduced from all NMR measurement after SeEVERS (1966) using the same C_K value coincide very well with the data obtained in the laboratory. They range from 10^{-4} m/s for the upper and lower aquifer to 10^{-3} m/s for the deeper part of the first sand layer. At around 10 m depth there are some differences between calculated K-values from NMR, grain size analyses and measurements on the drill cores. Thus K seems to be overestimated by NMR by approximately one magnitude. A reason for this could be that the assumption of a constant surface relaxivity, which leads to a constant C_K , is not valid for all parts of the aquifers.

Conclusions and Outlook

In the presented field example the decay time distributions obtained from surface-NMR measurements using the QT-inversion scheme were qualitatively in a good agreement to laboratory-NMR results. This promises additional information from surface-NMR about pore radius distribution of aquifers. The observed shift between the maximum of the T_2^* and T_2 distributions were of a magnitude of factor 2.

The used QT-inversion of the surface-NMR data resolves the lower aquifer better than the commonly used amplitude inversion approach. For the shallow subsurface, the estimated water contents and T_2^* decay times from both schemes were next to coincident for the presented field example. Laboratory measurements confirmed the surface-NMR results. For most parts of the subsurface, the hydraulic conductivity from surface-NMR and laboratory-NMR measurements agreed very well with K-values estimated from grain size analyses or measured in the laboratory. Disagreements especially in parts of the second aquifer need to be further studied.

In future work we are going to gather information from additional methods such as pumping tests and hydraulic tomography, further samples as well as a comprehensive comparison to the results from Spectral Induced Polarisation.

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