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Surface NMR within a geophysical study of an aquifer at Haldensleben (Germany)[1]

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Abstract

The surface nuclear magnetic resonance (SNMR) method has been tested at a site in Haldensleben, northern Germany, to assess the suitability of this new method for groundwater exploration and environmental investigations. More information is obtained by SNMR, particularly with respect to aquifer parameters, than with other geophysical techniques. SNMR measurements were carried out at three borehole locations, together with 2D and 1D direct current geoelectrics, as well as ground-penetrating radar, and well logging (induction log, gamma-ray log and pulsed neutron-gamma log). Permeabilities were calculated from the grain-size distributions of core material determined in the laboratory. It is demonstrated that the SNMR method is able to detect groundwater and the results are in good agreement with other geophysical and hydrogeological data. Using the SNMR method, the water content of the unsaturated and saturated zones (i.e. porosity of an aquifer) can be reliably determined. This information and resistivity data permit *in situ* determination of other aquifer parameters. Comparison of the SNMR results with borehole data clearly shows that the water content determined by SNMR is the free or mobile water in the pores. The permeabilities estimated from the SNMR decay times are similar to those derived from sieve analysis of core material. Thus, the combination of SNMR with geoelectric methods promises to be a powerful tool for studying aquifer properties.

Introduction

The first high-precision observations of nuclear magnetic resonance (NMR) signals from hydrogen nuclei were made as early as 1946 (Bloch, Hansen and Packard 1946; Purcell, Torrey and Pound 1946). Meanwhile this technique has found wide application in chemistry, physics, tomographic imaging in medicine, as well as in geophysics. Since the amplitude of the NMR signal is related to the number of hydrogen protons, the technique can be used in surface geophysics to measure the subsurface water content of rocks and soils. At present there is no other technique in

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surface geophysics to assess water content directly. Apart from water content, information about other properties, such as pore size, can also be obtained from NMR.

After it was discovered that the NMR signal is sensitive to the viscosity of the fluid (high-viscosity fluids exhibit fast relaxation), the oil industry increased their research in this field. Basic work revealed a relationship between the NMR properties of porous media, such as sandstones, and their permeability (Seevers 1966; Timur 1968, 1969a,b; Loren 1972). Current research is concentrating on the influence of pore size. In the last 10 years or so, NMR logging tools have been made available that use the CPGM pulse echo method (named after Carr and Purcell 1954; Meiboom and Gill 1958) for downhole measurement of relaxation parameters (Chandler, Kenyon and Morriss 1987; Straley *et al.* 1991). A detailed review of the use of NMR on rock cores and in boreholes is given by Kenyon (1992).

The first ideas for making use of NMR in groundwater exploration from the ground surface were developed as early as the 1960s (Varian 1962; Barringer and White 1968), but only in the 1980s was effective equipment designed and put into operation for surface geophysical exploration by scientists at the Institute of Chemical Kinetics and Combustion, Novosibirsk, Russia (Semenov, Pusep and Schirov 1982; Semenov 1987; Semenov *et al.* 1988, 1989; Legchenko, Semenov and Schirov 1990). Extensive surveys and testing have been conducted in sandy and clayey layers in Australia (Schirov, Legchenko and Creer 1991), on fractured limestone and chalk aquifers in France (Legchenko *et al.* 1995) and on fractured white chalk in France (Beauce *et al.* 1996). Tests in the USA (Lieblich *et al.* 1994) have provided some insight into the problems of sites where there is a high noise level. Some improvement in noise reduction has been achieved by using a special antenna configuration (Trushkin, Shushakov and Legchenko 1994).

Since both the phase and the amplitude of the surface NMR (SNMR) signal are affected by the electrical conductivity (Shushakov and Legchenko 1992; Trushkin, Shushakov and Legchenko 1995; Shushakov 1996), the conductivity distribution should be taken into account in the inversion of the data. A combination of the SNMR method with the time-domain electromagnetic (TDEM) method has been applied not only to detect the presence of groundwater, but also to obtain quantitative information about water content and salinity at various sites in Israel (Goldman *et al.* 1994; Gev *et al.* 1996).

In this paper, the results of the first study with SNMR in Germany are presented. The aim of the investigation is to assess the effectiveness of the method for groundwater exploration by comparing the SNMR results with groundwater data obtained by drilling and well logging. The SNMR method is also compared with other suitable geophysical methods (e.g. geoelectric methods, ground-penetrating radar and time-domain induced polarization) and is tested in combination with them for both groundwater exploration and determination of aquifer parameter values.

The basics of SNMR and the geology of the test site at Haldensleben are described below, and the data obtained using SNMR and other geophysical methods are then presented and discussed.

The SNMR method

Many atomic nuclei, including protons of the hydrogen atoms in water molecules, have a magnetic moment μ. These nuclei can be described in terms of a spinning charged particle. Generally μ is aligned with the local magnetic field B_0 of the earth. When another magnetic field is applied, the axis of the spinning proton is deflected, owing to the torque applied. When this second field is removed, the protons generate a relaxation magnetic field as they become realigned along B_0 while precessing around B_0 with the Larmor frequency,

$$\omega_0 = \gamma B_0, \tag{1}$$

where $\gamma = 0.267518$ Hz/nT, the gyromagnetic ratio for hydrogen protons.

The measurements are made using a circular or rectangular loop. An alternating current,

$$i(t) = i_0 \cos(\omega_0 t), \tag{2}$$

with a frequency ω_0 is passed through this loop for a limited time τ so that an excitation intensity (pulse moment) of $q = I_0\tau$ is achieved. After the current in the loop is switched off, a voltage is induced in the loop by the relaxation of the protons (Semenov *et al.* 1982; Legchenko *et al.* 1990; Schirov *et al.* 1991), given by

$$e(t) = E_0 \exp(-t/T_2^*) \cos(\omega_0 t + \varphi_0). \tag{3}$$

The initial amplitude E_0 is directly related to the water content as follows:

$$E_0 = \omega_0 M_0 \int_V f(r) B_\perp(r) \sin(0.5\gamma B_\perp(r) q) \mathrm{d}V, \tag{4}$$

where M_0 is the nuclear magnetization (the magnetic moment of the unit volume dV under equilibrium conditions at $t = 0$). $M_0 = 3.29 \times 10^{-3} B_0$ J/(Tm^3) for water at a temperature of 293 K. The volume fraction of water in a unit volume dV at the location $r(x,y,z)$ is given by $f(r)$. $B_\perp(r)$ is the component of the incident exciting field (normalized to 1 A) perpendicular to the static magnetic field B_0 of the earth. In a conductive medium $B_\perp(r)$ is composed of the primary field of the loop and the induced field. Note that the argument of the sine function in (4) ($\theta = 0.5\gamma B_\perp(r) q$) is the angle of deflection of the magnetic moment of the protons from the magnetic field of the earth. E_0 can be as large as a few millivolts.

T_2^* is the relaxation-time constant (spin-spin or transversal relaxation time). This decay time T_2^* can be of the order of a few milliseconds up to 1000 ms. It is related to the mean pore size and therefore grain size of the material. Clay, including sandy clay, usually has a decay time of less than 30 ms, whereas sand has one of 60–300 ms, gravel 300–600 ms, and pure water 600–1000 ms (Schirov *et al.* 1991).

The phase φ_0 is related to the phase of the excitation signal in (2). If the conductivity of the ground is negligible, $B_\perp(r)$ will have the same phase as the excitation current; hence, $\varphi_0 = 0$. If the ground has a high conductivity, a secondary magnetic field, superimposed on the primary field, is induced. This modifies the amplitude and phase

Figure 1. General location map of the site at Haldensleben.

of the field $B_\perp(r)$. Therefore, φ_0 is an indicator of the conductivity of the ground and the groundwater as well.

The resolution and accuracy of the method depend on $B_\perp(r)$ and decrease with depth. Higher currents I_0 and/or τ are needed to excite the protons at greater depth (as long as $\tau << T_2^\star$). By increasing q, the depth of the measurement is increased. In fact, the choice of q focuses the excitation to a certain depth range.

Investigation site

An integrated geophysical survey was carried out at a site near Haldensleben in northern Germany (Fig. 1). At the ground surface, there are mostly Quaternary sands and gravels and there is no surface drainage network. The site has little vegetation; thus, precipitation is little influenced by the root zone and percolates through the unsaturated zone as groundwater recharge.

The geology of the area in which three boreholes B7, B8 and B2 were drilled (Fig. 2) consists of Quaternary deposits: well-sorted sands interbedded with cohesive glacial till and silt. The sands and gravels form good aquifers with permeability coefficients of 10^{-4}–10^{-3} m/s. The glacial tills and silts, which act as aquicludes, may be as thick as 20 m. Since they are discontinuous, there are local hydraulic links between the aquifers.

The hydrogeological conditions at locations B7 and B8 are similar. The water table is at a depth of about 20 m and the depth of the base of the first aquifer varies between 40 and 50 m. At B7 there are interbedded impermeable tills and silts in the depth interval between 40 and 65 m. At B8 there is an impermeable till layer about 12 m thick. At both locations there is a confined second aquifer below the till. The regional aquiclude, the Rupelian clay, occurs at a depth of 75–80 m, immediately below the Quaternary. The situation at B2 is somewhat different. The water table is at a depth of 7.3 m mainly due

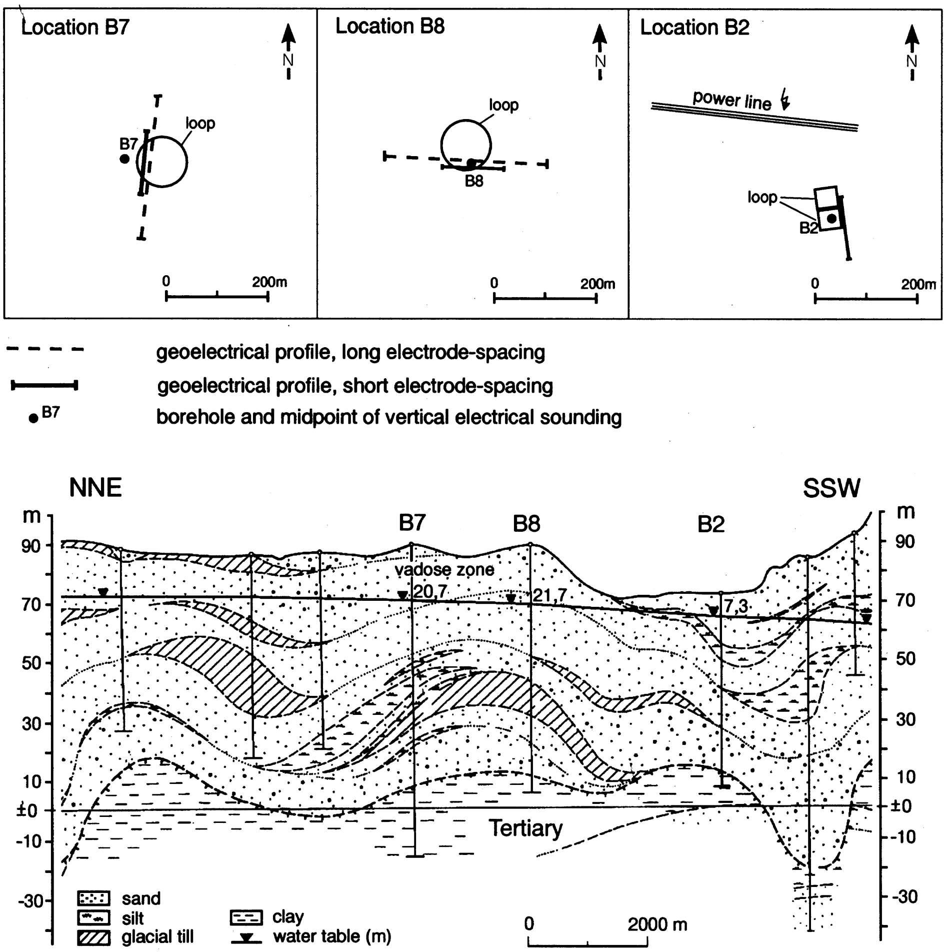


Figure 2. Measurement layouts and hydrogeological section (vertical axis exaggerated).

to the lower elevation of the site (Fig. 2). Moreover, a ponding layer 6 m thick consisting of silt is present at a depth of 16 m.

SNMR measurements

IRIS Instruments' NUMIS system, which is the only commercially available system for SNMR measurements, was employed. Standard NUMIS software was also used for processing and inversion. The measurements at locations B7 and B8 were carried out using a circular loop 100 m in diameter. A figure-eight loop shaped like two adjacent 37.5 m squares was used at location B2 (Fig. 2) in order to decrease the influence of noise, as was done by Lieblich *et al.* (1994). To determine the excitation frequency, the local magnetic field of the earth was measured using a proton magnetometer. The earth's magnetic field at B7 was $B_0 = 48\,757$ nT; this corresponds to a Larmor frequency f_0 of 2076 Hz. The duration of the excitation current was kept constant at 40 ms, corresponding to about 80 cycles. Measurements were carried out at several current strengths between 5 A and 250 A, varying the excitation intensity from 200 A.ms to ~ 8000 A.ms.

The signal amplitudes measured at B7 and B8 show a typical shape for sounding curves obtained for an aquifer at moderate depth (Figs 3a and b). Inversion shows that the water content gradually increases with depth from 5% in the unsaturated zone to 20–25% at 20 m, the depth of the water table. Below 30–35 m the water content decreases. At location B2 (Fig. 3c) signal amplitudes and water content after inversion are almost constant and very low, the maximum water content being about 10%. The noise levels were about 500–700 nV at B7, 100–300 nV at B8 and 600–1100 nV at B2. The inversions have an rms error of less than 5%, which indicates good agreement between the observed amplitudes and the amplitudes reconstructed from the model.

Except for the two highest excitation intensities, the decay times at B7 were 140–220 ms and quite irregular. At B8, the decay times increase smoothly from 180 ms to 250 ms at a 4000 A.ms excitation intensity and then decrease slightly. The decay times at B2 change from 150 to 100 and then 160 ms with increasing excitation intensity. The range of the decay times corresponds to that of medium sand (Schirov *et al.* 1991). The phases at B7 and B8 begin with 0° and increase to 90° and 60°, respectively, indicating the existence of conductive layers at depth, in agreement with the results for water content.

In general, the SNMR results are in agreement with the borehole data and geoelectric data from sites B7 and B8, at least down to a depth of 40 m, clearly confirming the presence of the aquifer. Resolution with depth becomes poorer due to the thickness of the horizontal layers used in inversion increases. The inferred decrease in water content with depth is not necessarily reliable. At location B2, the distribution of the water content does not correspond to the known geology, but the noise level was remarkably high despite the use of a figure-eight loop. The noise was probably due to the power line nearby (Fig. 2). Usually the long dimension of a figure-eight loop is orientated parallel to power lines (Lieblich *et al.* 1994), but in this case problems of access prevented it.

Geoelectric measurements

An extensive geoelectric survey was also carried out at the test location. As the geoelectric method is the technique most commonly used to explore for groundwater, it is important to compare the geoelectric data and SNMR data and to attempt a combined interpretation, particularly with respect to the aquifer parameters. Moreover, the conductivity of the ground influences the SNMR signal amplitude and phase. High conductivity can result in a reduced investigation depth, as is the case for EM methods. A conductive layer above the aquifer causes a decrease in the SNMR amplitude; the same layer below the aquifer increases the SNMR amplitude.

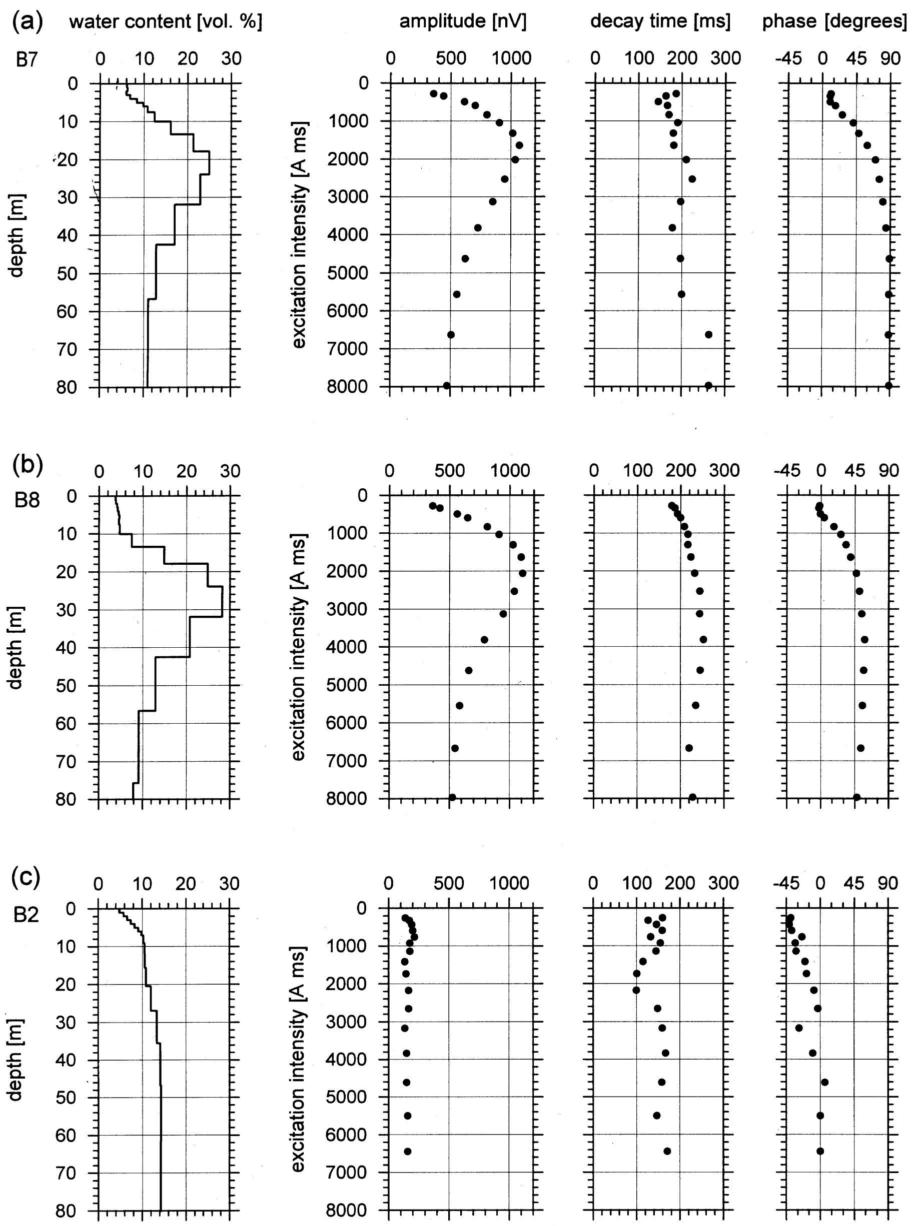
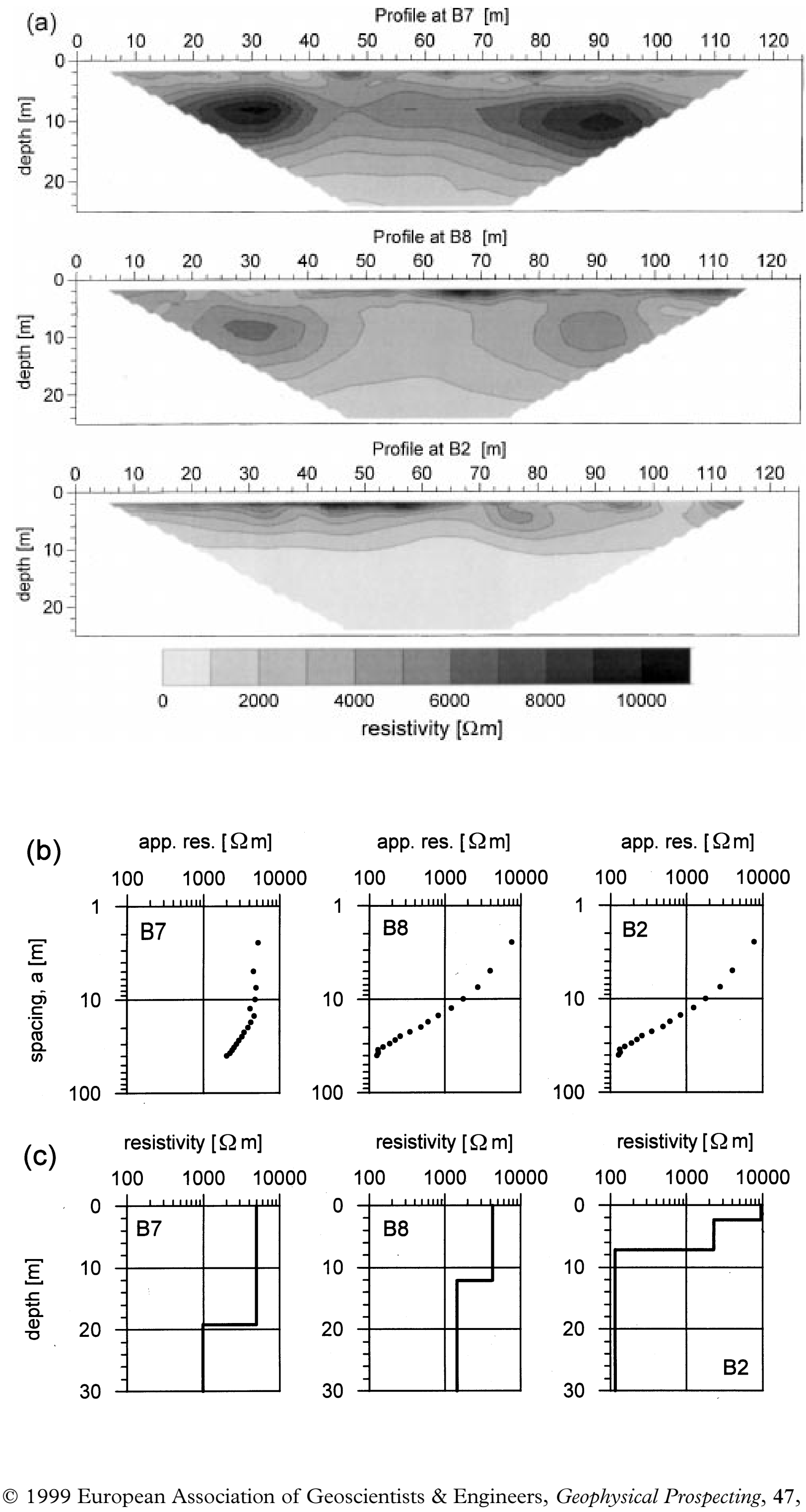


Figure 3. SNMR data and results of the inversion for water content at the locations (a) B7, (b) B8 and (c) B2.


(a)
Profile at B7 [m]
depth [m]
Profile at B8 [m]
Profile at B2 [m]
resistivity [Ωm]
(b)
app. res. [Ωm]
spacing, a [m]
B7
B8
B2
(c)
resistivity [Ωm]
depth [m]
B7
B8
B2

Investigations to rather shallow depths with multi-electrode 2D resistivity measurements using a basic electrode spacing of 2.5 m were carried out initially. The directions of the profiles at the three locations are shown in Fig. 2. The resistivity distributions as calculated by a 2D inversion algorithm are shown in Fig. 4a. At all locations there is a distinct and rapid decrease of the resistivity with depth, indicating the presence of groundwater. The low resistivity at B2 is encountered at quite shallow depths. At locations B7 and B8 there appear to be some local low-resistivity inhomogeneities corresponding to discontinuous glacial till layers.

2D inversion of geoelectric data using a non-linear, least-squares algorithm which gives smooth models (in this case the RES2DINV package, Loke and Barker 1996) is not robust enough to identify sharp resistivity boundaries, as expected for groundwater in sandy formations. The inversion algorithms tend to 'smear' the resistivity data when large differences are present. Consequently, the depth to groundwater cannot be determined precisely since 2D block inversion algorithms imposing sharp boundaries are not yet widely available. Therefore, vertical sets of data were selected from the 2D data for the midpoints of the profiles and fed into a 1D inversion algorithm using a two-layer model. The results shown in Fig. 4c are somewhat ambiguous. The top layer has a mean resistivity of $5\times10^3\,\Omega$m for all locations. A low-resistivity layer, probably indicating groundwater, is detected at 19 m at B7, at 12 m at B8 and at 6.5 m at B2. The resistivity for the aquifer is found to be about $10^3\,\Omega$m at B7 and B8, but $10^2\,\Omega$m at B2. At B7 the depth of 19 m obtained for the aquifer is slightly lower than the depth of the water table measured with an electrical tape gauge, but at B8 the water level obtained by the geoelectric measurement is erroneous. At location B2 the aquifer resistivity is in agreement with the value of $10^2\,\Omega$m known from other extensive geoelectric surveys in this area, but the resistivities obtained for the aquifer at B7 and B8 are too high by an order of magnitude. These interpretations are confirmed by borehole data as discussed below.

In order to increase the depth of investigation, 2D measurements were conducted with a larger array with an electrode spacing of 10 m. The gain in depth is obtained at the expense of a decrease in resolution, particularly in the shallow range. The 2D measurements were carried out at locations B7 and B8 and the results are shown in Fig. 5a. For shallow depths the results are in agreement with those obtained using the smaller array, particularly with respect to the position of the shallow glacial till. Increasing the investigation depth to 55 m showed that the resistivity of the upper aquifer may be as low as $20\,\Omega$m. However, due to the smoothing by the 2D inversion discussed above, the abrupt change in resistivity at the water table is not visible in the inverted data.

This observation is confirmed by the Schlumberger soundings, which were

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Figure 4. (a) 2D resistivity cross-sections obtained using a short electrode spacing at locations B7, B8 and B2. (b) The rms errors for model fitting are lower than 5%. (c) Selected soundings at the centre of 2D sections inverted using 1D layered models with sharp boundaries.

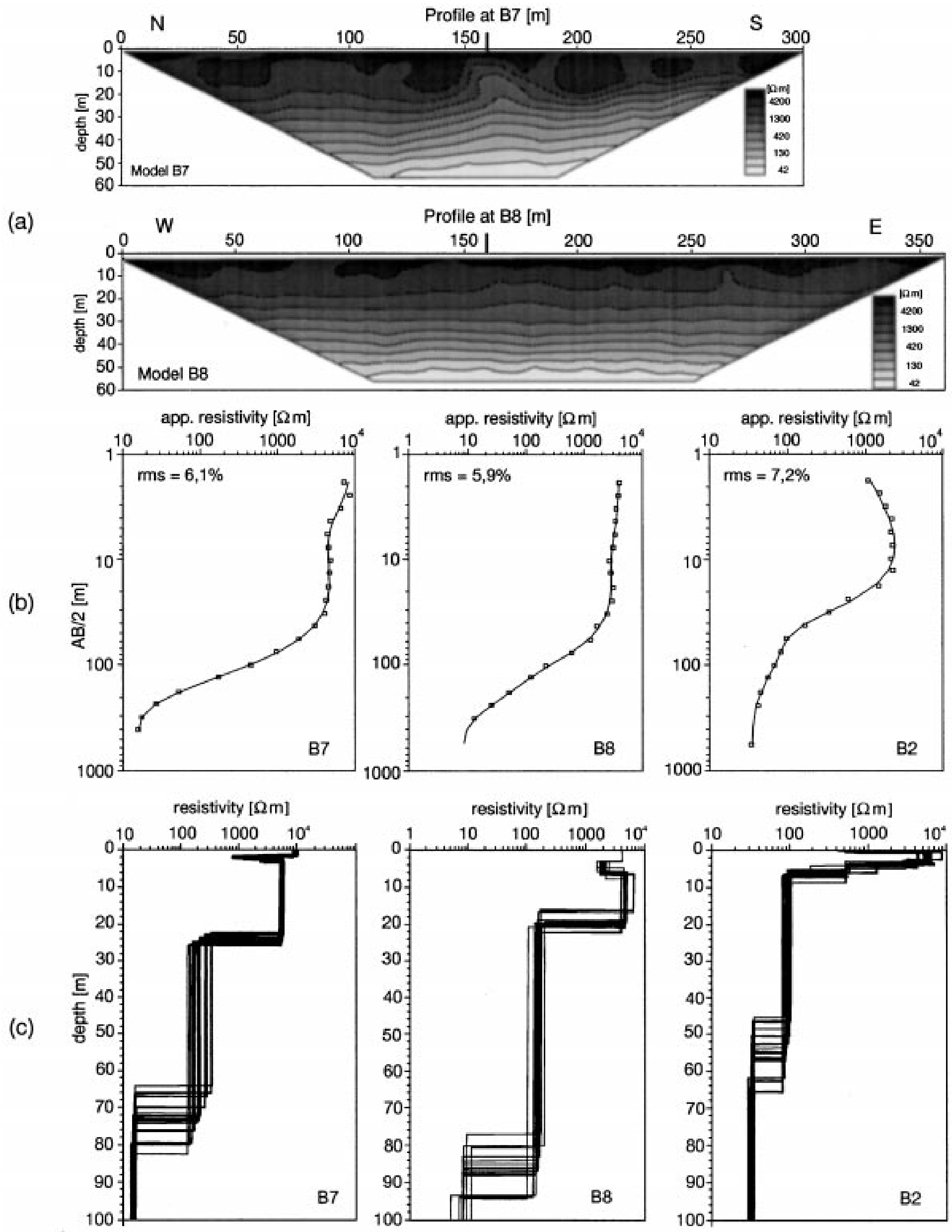

Figure 5. (a) 2D resistivity cross-sections obtained using a long electrode spacing at locations B7 and B8. (b) The rms errors for model fitting are lower than 10%. (c) In an earlier site investigation vertical electrical soundings were carried out and equivalent models developed after 1D inversion by curve fitting.

subsequently carried out at all locations with very large electrode spacings, also using 1D inversion. In the three areas B7, B8 and B2, the water table is very clearly indicated at 23, 20 and 6 m, respectively (Fig. 5b); these depths are in good agreement with borehole data. The resistivity of the unsaturated zone is 5×10^3 Ωm, and the aquifer has a resistivity of about 100 Ωm. The relatively high resistivities shown by the aquifers are due to the low concentrations of total dissolved solids in the groundwater. All the Schlumberger sounding curves record the strong resistivity difference at the water table. At B2 the relatively low resistivities near the surface are visible in the sounding curves obtained using a short spacing. In 1D inversion using interactive curve matching, no *a priori* information was used. The optimal model for all three locations consists of five layers. Several more thin layers could be derived from the shape of the curve, but the results would not be reliable if there are 2D or 3D geoelectric structures.

It should be noted that the 2D geoelectric measurements using a large spacing included time-domain induced polarization (TIP) in order to detect cohesive layers (till and silt), which borehole measurements indicated as being present between the aquifers (Fig. 6). However, inversion did not yield any anomalous TIP parameter values. In view of the high resistivities at shallow depths, it seemed feasible to map the groundwater with ground-penetrating radar (GPR), but no clear reflections were obtained from the water table, as attenuation was still high at shallow depths due to the considerable amount of capillary water and some thin, conductive, cohesive layers, as confirmed by well logging.

The discussion above illustrates the problems encountered during groundwater exploration even using relatively well-developed methods such as 2D geoelectrics and GPR.

Borehole measurements

The SNMR and geoelectric results, as well as the parameters derived from grain-size analysis of cores, were checked by well logging using the following methods: gamma-ray log (GR), induction log (IL), impulse neutron-gamma log (ING) and salinity log (Sal). The lithology, the conductivity and water content of the various layers, as well as the conductivity of the groundwater, were derived from the measured data (Fig. 6).

The ING method is sensitive to the total volume of water in the pore space. This is a fundamental difference from the SNMR method, which is sensitive only to the free water. Water in small pores of sediments like clay and silt and adhesive water, which is bound on the grain surface by strong molecular attraction, cannot be detected by SNMR (Schirov *et al.* 1991; Lieblich *et al.* 1994). Therefore, the water content given by ING logs is generally higher than that determined by SNMR. For saturated sediments the total water content equals the total porosity. The relationships between total pore space (total water content), effective porosity (free water) and adhesive water are shown in Fig. 7 for clastic rocks.

The ING logs show that the water content for boreholes B7 and B8 have approximately the same range and distribution. In the unsaturated zone down to a

(a)

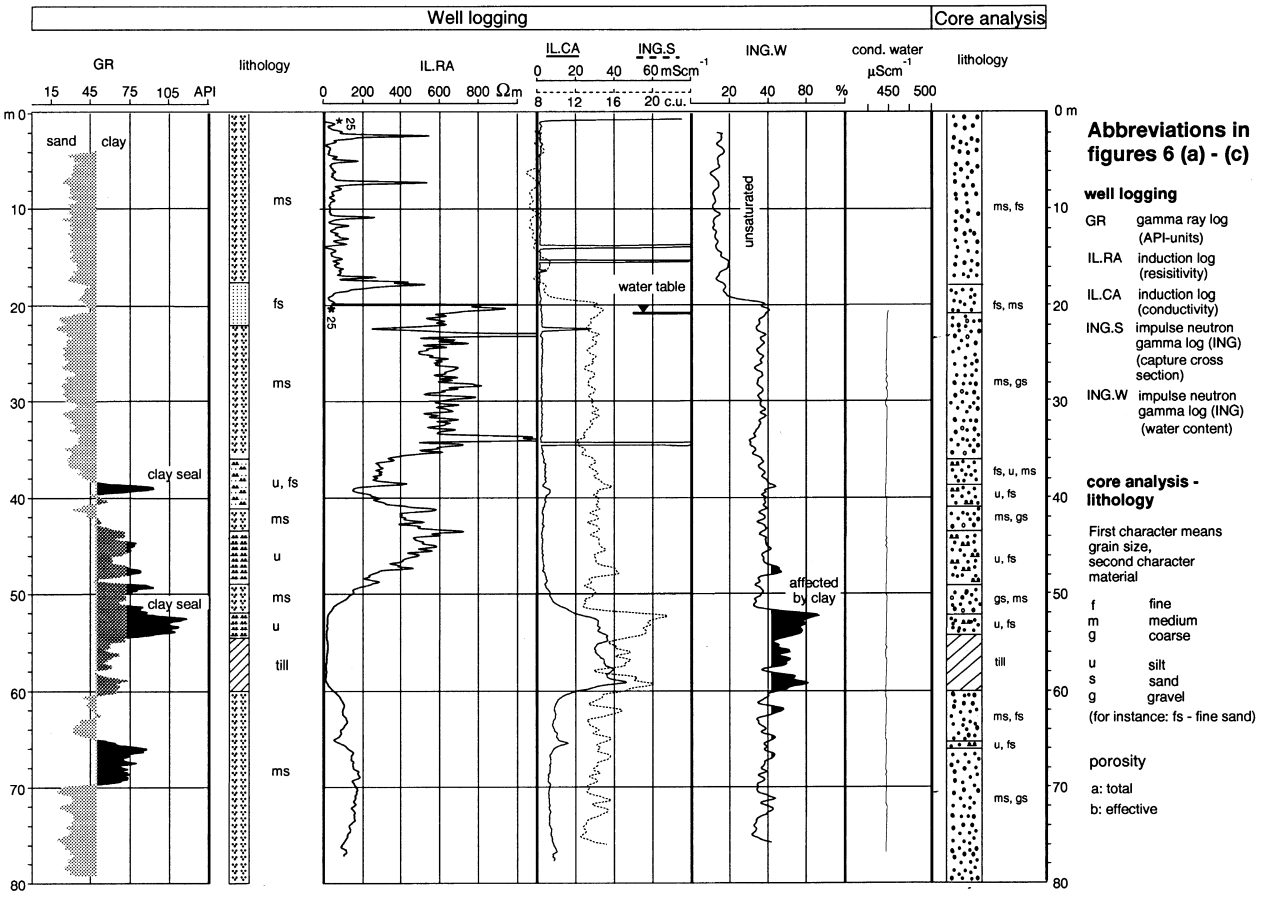

Figure 6. Well logging and drilling results at the locations (a) B7, (b) B8 and (c) B2.

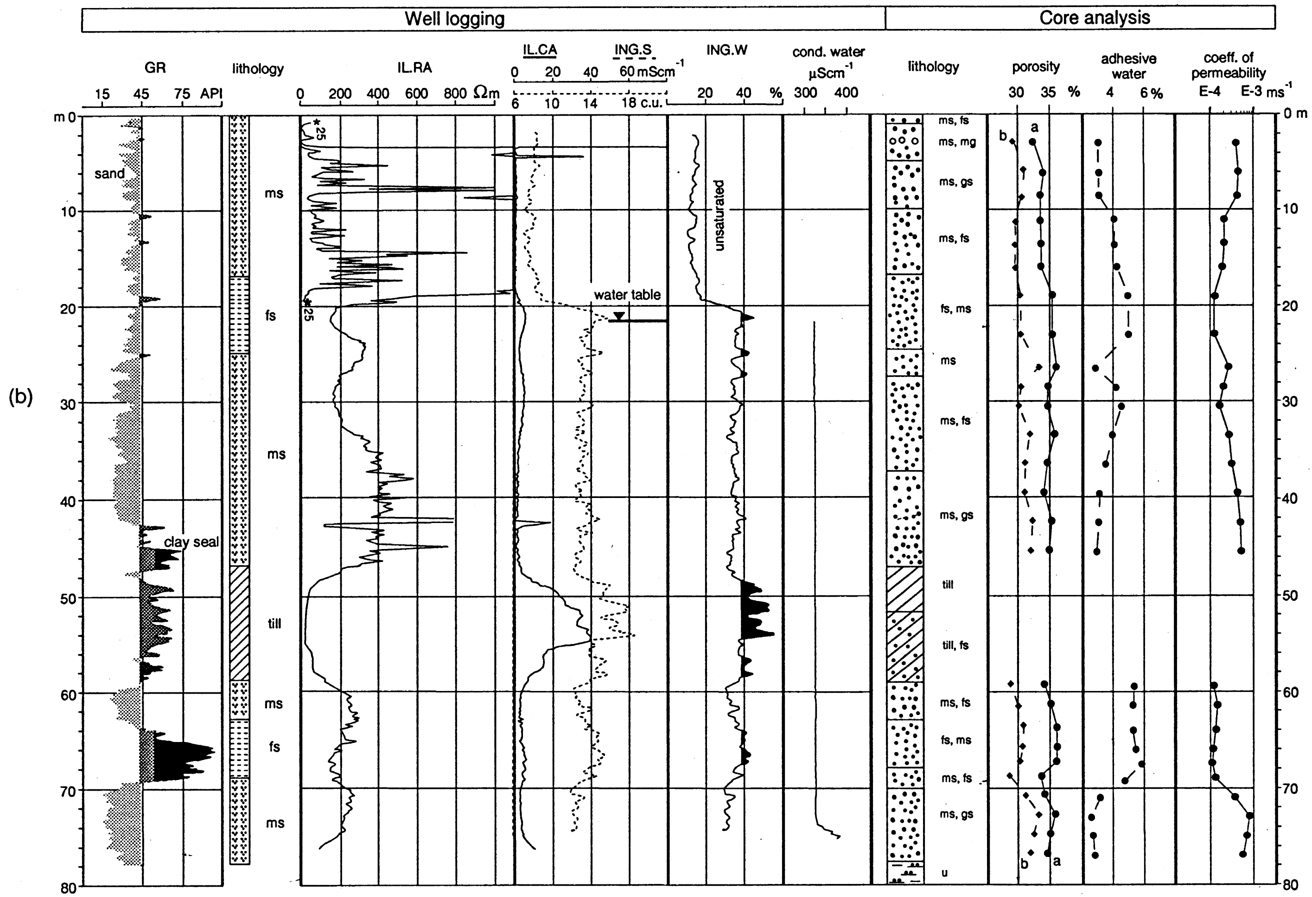

Figure 6. Continued

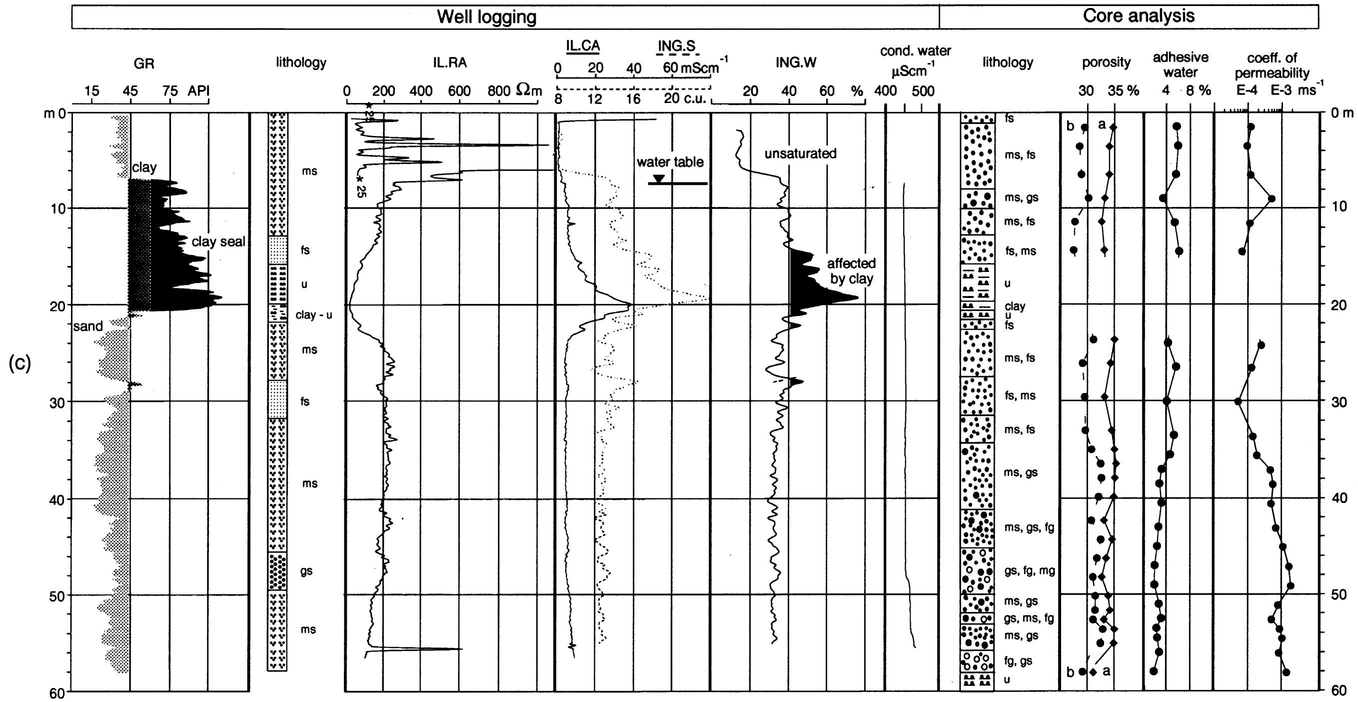

Figure 6. Continued

depth of 6 m, the total water content averages 15% and increases slightly to 20% at about the depth of the water table. Experience indicates that the amount of adhesive water in the saturated zone should be less than 8%. The rather high values are probably caused by the presence of percolation water.

The ING log (ING.W) gives the water content of the capillary fringe just above the water table as about 35% on average. Slightly higher or lower values are due to finer or coarser-grained material (see porosity curves in Figs 6b and c). Clearly visible is the effect of clay seals behind the casing, where the water content is 50–65%.

In borehole B2, the thick clay seal directly below the water table causes the water content in this zone to rise to 75%. In the lower part of the borehole the water content is 3–5% less than in B8 and B7. This is in good agreement with the porosities and permeability coefficients obtained from core analysis (Fig. 6).

Discussion

The following discussion of the SNMR results will focus on test locations B7 and B8. Location B2 is not suitable for determining aquifer parameter values due to the extremely low signal-to-noise ratio, even though a square figure-eight loop was used to reduce noise.

Comparison of the SNMR results with the ING log shows that the water content determined from the SNMR data is too low by 5–10% in the unsaturated zone and by about 10–12% in the upper part of the aquifer, i.e. between the water table and a depth of 40 m. In the deeper parts of the aquifer, the SNMR data indicate an unrealistically low water content, possibly due to an insufficient data density. However, the actual reason is not fully understood yet and further investigation of the inversion algorithm and the measuring procedure is necessary.

Previous work has already shown that SNMR measurements detect only free water (Schirov *et al.* 1991; Lieblich *et al.* 1994). Water bound on the pore surfaces has very short relaxation times and cannot be detected by SNMR yet, owing to the technically inevitable dead time of 30 ms in the measurements. Figure 7 shows empirical relationships between grain size and relaxation time (Schirov *et al.* 1991), permeability (Hölting 1992), and porosity (Davis and de Wiest 1966) for unconsolidated sediments. Clay and silt, which normally constitute aquicludes, have higher total porosities than sandy aquifers. This means that aquicludes have a higher total water content than sands when fully saturated. However, a large proportion of this water is adhesive water and this must be taken into account when interpreting SNMR results and comparing them with ING logs.

The porosity of an unconsolidated sediment depends basically on its texture, which is characterized by the particle-size distribution, d_{60}/d_{10} – the degree of sorting. This value is obtained by mechanical sieve analysis of core samples. It is plausible that sediments with a lower degree of sorting are better sorted than those with a higher degree of sorting. Sieve analysis was carried out on aquifer material from boreholes B8 and B2. Porosities and amounts of adhesive water were then estimated using

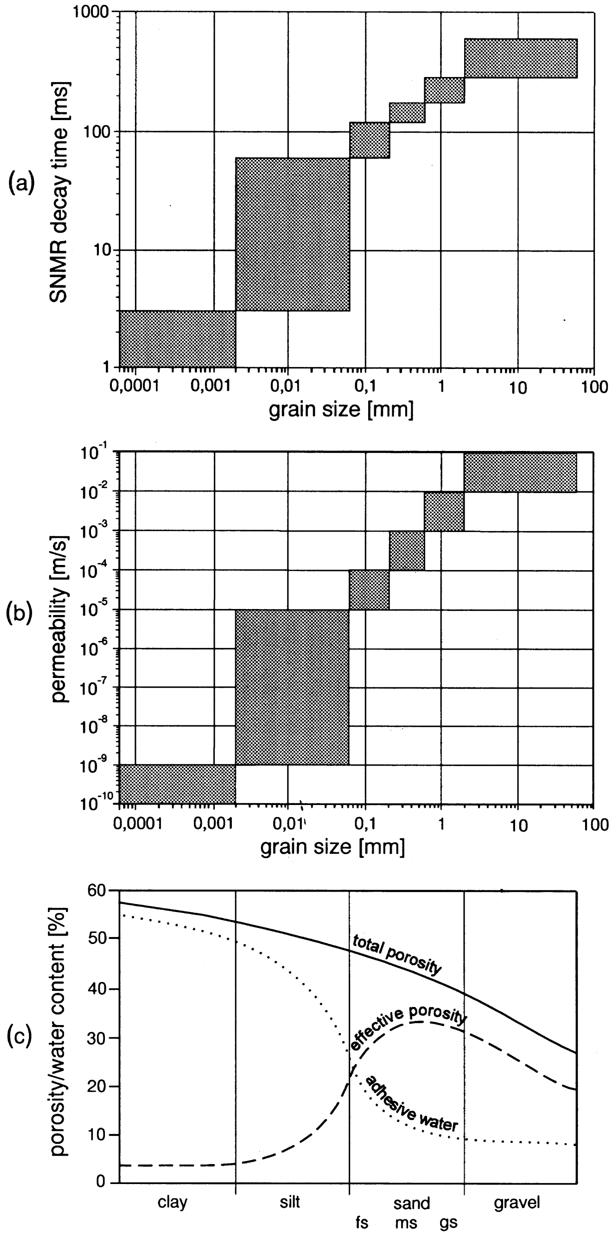

Figure 7. (a) Relationship of SNMR decay-time ranges to the grain-size ranges of unconsolidated sediments adapted from SNMR results at different sites and from corresponding core material after Schirov *et al.* (1991). (b) Relationship of permeability ranges to grain-size ranges after Hölting (1992). (c) Relationship between porosity and adhesive water for different sediments after Davis and de Wiest (1966).

conventional methods (Beyer 1964; Beyer and Schweiger 1969). The reliability of the estimated parameters is high, especially for well-sorted sands like those in the test area.

The total porosities derived from sieve analysis show good agreement with the ING logs in Figs 6b and c, where the parameter ING.W for saturated material is a function of the total porosity. The water content derived from SNMR is lower than that obtained from the ING log by an amount approximately equal to the adhesive water (3–6%).

The water content itself gives no clear indication of whether the soil is saturated or unsaturated. Therefore, it is not a direct measure of the yield of a well. In the unsaturated zone many isolated pockets of capillary water may represent a considerable volume of free water, which, however, is not exploitable groundwater. A complete evaluation of water in a soil requires a knowledge not only of the amount of water in the soil, but also of its energy status. This is described by the soil water retention curve (Wilson, Everett and Cullen 1995). The SNMR results for the test area show a relatively large amount of water in the unsaturated (vadose) zone. This is in good agreement with the ING log, which shows 12–15% total water on average (Figs 6a–c).

Geoelectric methods are the geophysical techniques most widely used in groundwater exploration. The electrical resistivity measured depends to a large extent on water content, but the relationship between water content and resistivity is not unambiguous. The resistivity also depends on the salinity of the water, on pore structure, and, to a large extent, on clay content. The resistivity of unsaturated sand is described by the well-known Archie relationship (1942),

$$\rho = \rho_w \phi^{-m} S^{-n}, \tag{5}$$

where ρ denotes the resistivity of the rock, ρ_w denotes the water resistivity, ϕ denotes the porosity, m is the Archie exponent, S denotes the degree of saturation, and n is the saturation index. Using V_w, V_{por} and V for the volumes of the water, the pores and the rock, respectively, the porosity is given as $\phi = V_{por}/V$, the degree of saturation $S = V_w/V_{por}$ and the water content $G = V_w/V = \phi S$. For a saturated rock, i.e. $S = 1$ and $G = \phi$, (5) reduces to

$$\rho_0 = \rho_w \phi^{-m}. \tag{6}$$

Two parameters that express the influence of pores on the resistivity are the formation factor $F = \phi^{-m} = \rho_0/\rho_w$ and the saturation index $I = S^{-n} = \rho/\rho_0$.

Using the aquifer resistivity $\rho_0 = 100\,\Omega$m, obtained from geoelectric sounding, the resistivity of the water $\rho_w = 29\,\Omega$m, obtained from the salinity log, and the porosity $\phi = 0.35$, obtained from borehole logging, it is found that $F = 3.5$, which is within the usual range, and $m = 1.2$, which is slightly low. Substituting these values in (5) gives $\rho = 100S^{-n}$, which can be used to estimate the degree of saturation, and hence the water content, in the unsaturated zone using a resistivity of $5 \times 10^3\,\Omega$m. For most sandy material $n = 2$, but even if n is taken as ranging from 1.7 to 2.3, the degree of saturation obtained is 0.1–0.18, yielding a water content of 3.5–6.3%. This is less than the water content derived from core analysis and SNMR by a factor of up to 3. Using the observed water content of 15% for the unsaturated zone yields a resistivity of about $4{-}7 \times 10^2$ Ωm. The observed resistivity is much higher than this, probably due to enclosed air and isolated water in the sands. In this particular case, geoelectrics is not very useful for estimating the water content of the unsaturated zone. Moreover, values are not always available for (5).

In order to obtain permeabilities from SNMR measurements, the empirical relationship between decay time and average grain size observed in many SNMR surveys (Schirov *et al.* 1991) can be used (Fig. 7a). The relationship between grain size and permeability (Fig. 7b) often used in hydrogeology (Hölting 1992) can be used to derive an expression for the permeability coefficient k (in m/s) as a function of decay time T (in s),

$$k = 1.1\, T^{4.14}. \tag{7}$$

The decay times of around 100–200 ms from the Haldensleben survey yield permeability coefficients of $\sim 0.8 \times 10^{-4}$ to 1.4×10^{-3} m/s, which are in very good agreement with those derived from the core material (Fig. 6). Therefore, it is suggested that (7) be used to estimate permeabilities from decay times in future studies. Some modification of this equation may be needed in the future with an increase in the database.

Conclusions

The use of SNMR for groundwater exploration is still in an experimental stage. Nevertheless, the results already obtained with this method and those described in this paper show that the SNMR method has the potential to advance to a valuable tool for groundwater exploration and aquifer characterization.

There is only a 1D inversion method available for SNMR data, but it is suitable for the configuration used. Future developments have to take 2D configurations and inversions into account, as there may exist distinct 2D features whose effect may not be documented fully using a 1D interpretation. It also generally appears to be very important to include conductivity data in the interpretation, but not at this particular site since the resistivities of the unsaturated zone and aquifers are unusually high.

The 1D inversion of SNMR data may also be ambiguous, since different regularizations in the inversion impose a certain degree of smoothness upon the distribution of water content (Legchenko and Shushakov 1998; Yaramanci, Lange and Knödel 1998) and may lead to differing results, making it impossible to decide which is the most realistic model without supplementary information. The rms error, usually used for any geophysical inversion, is not necessarily sufficient for assessing the fit of a model to the observed data (Yaramanci *et al.* 1998). The inversion models for SNMR data presented here yield values for the variance that are so large that the water content values can only be considered to be rough estimates.

At sites where no *a priori* information is available, SNMR should always be carried out in conjunction with electrical methods, i.e. direct current geoelectric, electromagnetics and even GPR. This will help to decrease ambiguity in the results and will also allow hydrogeological parameters to be estimated, as has been demonstrated in this paper. But despite all the difficulties, the quality of geophysical exploration for groundwater and aquifer properties will have an increased degree of reliability by using SNMR as a direct indicator of water and soil properties.

The results obtained at the test locations show that the potential of SNMR is not restricted to determining the water content of aquifers, but can also determine free water in the unsaturated zone, which is not available groundwater. This is important for environmental research, because this water affects contaminant transport in several ways. In order to evaluate the potential of the unsaturated zone to protect groundwater resources, detailed knowledge of the water content and permeability of this zone is required. The SNMR results, together with the geoelectric data obtained at the test locations, show that the SNMR method is a promising tool for this purpose. In fact, SNMR may turn out to be the only suitable non-intrusive tool for examining the degree of saturation in the unsaturated zone under certain soil conditions.

The importance of the SNMR method lies in its ability to measure water content directly. In this respect it is unique, since all other geophysical methods measure water content indirectly via resistivity, seismic velocity, etc. Using SNMR in combination with other geophysical methods, the problem of salinity in determining the water content from the resistivity can be resolved. Moreover, properties of the pore network (e.g. as expressed by the Archie exponent) can be deduced if the water content is known and salinity is sought, or vice versa.

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