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Quantitative Hydrogeological Study of an Unconfined Aquifer by GPR along Tuul River in Ulaanbaatar

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Introduction

To assess the potential of ground penetrating radar (GPR) for detecting and monitoring groundwater movement and for estimating the hydraulic properties of an aquifer, we conducted GPR surveys at a water source area of Ulannbaatar city.

Ulaanbaatar, the capital of Mongolia, is the country's center of industry and commerce, and has experienced significant population growth in the last decade. Ulaanbaatar city is characterized by a semi-arid climate, with a hot, dry summer and a cold winter. The average annual rainfall of Ulaanbaatar station is calculated to be 243.1 mm, with nearly 74 percent of the annual precipitation falling between June and August (JICA, 1995). Water supply of Ulaanbaatar city depends solely on groundwater withdrawn from an alluvial aquifer, distributed in the Tuul River basin, which is mainly located in the southern part of the city. The water is supplied from water production wells. With the increase of population and economic development, Ulaanbaatar city is facing water shortage. Therefore, assessing the groundwater production from a well and its production capacity has become very important. However, if the groundwater level change around the production well can be observed by GPR, it will provide much more information about the aquifers. The groundwater level in the Ulaanbaatar city area is between 2 – 10 m, and the GPR technique is suitable for detecting this relatively shallow aquifer.

Field GPR surveys in Ulaanbaatar have been carried out regularly since 1997. In the investigation described in this paper, we conducted field experiments in Ulaanbaatar in October 2001and April 2002. By controlling the water production we used GPR for detecting the change of groundwater conditions around the well. This paper focuses on the practical use of GPR for groundwater monitoring, and tries to quantify the groundwater level change and to estimate the hydraulic properties by assuming a model of the aquifer system.

GPR Survey

The field GPR surveys were carried out around production well No.10 in the western site in the Central Water Source area shown in Figure 1. The Central Water Source area is around Ulaanbaatar city between the upstream and downstream basin areas. There are 70 production wells managed by USAG (Water Facilities Exploitation Department of Ulaanbaatar Municipality) in the Central Water Source. Well No.10 was drilled in1961 with an inner radius of 0.2 m and a depth of 30.7 m. The pump and the well are located in a brick pump house and the survey lines were taken around the pump house beginning from the wall of the pump house in the directions indicated in Figure 2.

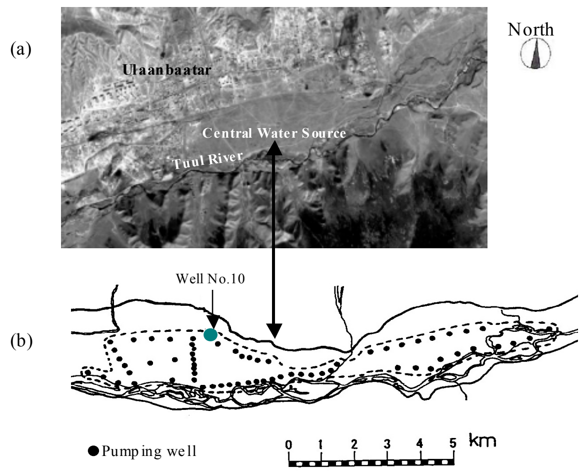

Figure 1. Central Water Source and pumping well No.10. (a) 1990 SPOT PAN image of Ulaanbaatar area. (b) Pumping well distribution at the Central Water Source area (JICA, 1995).

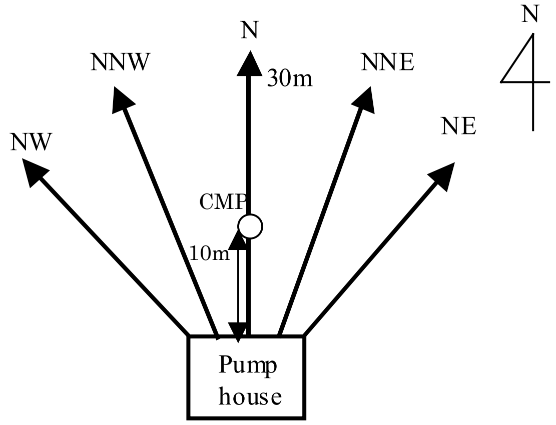

Figure 2. GPR survey lines around the pump house of well No.10.

The first experiment was carried out during 4 and 5 October 2001. The pump was stopped on 4 October when the groundwater level reached 6.45 m as measured in the well. After about 1000 minutes, on 5 October, the groundwater level was restored to 5.8 m. The groundwater level was continuously measured in the well for 3.5 hours, and it stayed at 5.8 m. From that fact, we suppose the groundwater level recovered to its quasi-steady condition. Before the pump was stopped and when the groundwater was in its static condition, GPR measurements were conducted repeatedly. CMP data were also acquired during the pumping test, which began from the midpoint at the position of 10 m with a first offset of 20 cm, and increased in a 0.1 m step to a maximum of 10 m. We show GPR data (CMP) acquired along line N here.

CMP gathers showed in Figure 3a were acquired during pumping, i.e., the groundwater level in the well was 6.45 m. Figure 4a shows the CMP gathers collected after recovery, that is, when the water level in the well was 5.8 m. Figure 3b and Figure 4b show the velocity spectrum (Yilmaz, 1987) which were derived from the data in Figure 3a and Figure 4a. The velocity differences produced by different water level depths can be seen around 65 ns in the two figures. The difference between the relative dielectric constant of liquid water ($\varepsilon_{r,w} \simeq 81$) and those of most rock matrix materials ($\varepsilon_{r,g} = 3-5$) is large. Accordingly, it is known that the dielectric constant of most geological materials is governed by their water content. When the relative dielectric constant of the soil is ε_r, EM wave velocity (Davis and Annan, 1989) in the soil is given by

$$v = c/\sqrt{\varepsilon_r} \qquad (1)$$

where c is the velocity of light in air. Therefore, the travel time from a boundary at the depth d is given by

$$\tau = \frac{2d}{v} = \frac{2d\sqrt{\varepsilon_r}}{c} \qquad (2)$$

The velocity obtained from the velocity spectrum is the normal moveout (NMO) velocity, approximately the same as the rms velocity assuming the medium is homogeneous from the surface to the boundary, when the subsurface consists of multiple horizontal layers. Therefore, in order to estimate the relative dielectric constant of each layer, the rms velocities have to be corrected to interval velocities. The average interval velocity of the n-th layer can be calculated using the Dix formula (Dix, 1955)

$$V_n^2 = \frac{V_{RMS_n}^2 t(0)_n - V_{RMS_{n-1}}^2 t(0)_{n-1}}{t(0)_n - t(0)_{n-1}} \qquad (3)$$

where V_{RMS_n} and $t(0)_n$ denote the rms velocity and vertical reflection travel time to the n-th layer. The relative dielectric constant can be obtained by equation (1) using the interval velocities, and the depth of each layer can be obtained by equation (2).

Generally, the water content θ, porosity ψ and water saturation S_w are related as $\theta = \psi S_w$. Several cases showing the relationship between the dielectric constant and properties of θ, ψ and S_w have been reviewed by Sen et al. (1981) and Shen et al. (1985). However, it is impossible in practice to derive both porosity ψ and water content θ independently from the dielectric constant. An empirical equation derived by Topp et al. (1980) using various soil samples with different degrees of saturation showing the relation between the dielectric constant and water content is given as

$$\theta = -0.0503 + 0.0292\varepsilon_r - 5.5\times10^{-4}\varepsilon_r^2 + 4.3\times10^{-6}\varepsilon_r^3 \quad (4)$$

By using the interval velocities, we can obtain the water content of the soil at each depth through equations (1), (2) and (4). The water content was calculated by using CMP data shown in Figure 4a. The rms velocities were extracted from the velocity spectrum. The water table was defined as 5.53 m (after topographic correction). The water content (Figure 4c) varied from 6.8% to 8.9% in the vadose zone. It is 31.87% if we use one layer velocity model (suppose only one layer below water level) and 32.8% and 31.5% if two-layer model was applied (suppose two layers below water level) in the saturated zone.

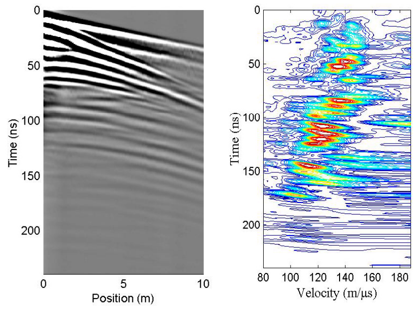

Figure 3 CMP gathers acquired along line N (midpoint = 10 m) during pumping (water level in the well was 6.45 m) in October 2001. (a) CMP gathers. (b) Velocity spectrum obtained from (a).

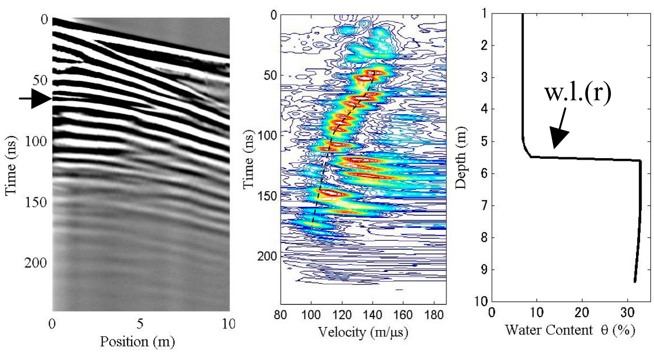

Figure 4 CMP gathers acquired along line N (midpoint =10 m) when groundwater level after recovery (water level in the well was 5.8 m) in October 2001. (a) CMP gathers. (b) Velocity Spectrum obtained from (a). (c) Water Content.

The second experiment was carried out in April 2002. This experiment differed from the earlier one (Lu and Sato, 2002) in that it used a longer survey line

(30 m instead of 15 m). In this experiment, we controlled the well production for three steps. In the first step, the water level in the well was 8.25 m when the well production was in full-working condition. The well production was then reduced by half; after about 1000 minutes, the water level in the well was restored to 8 m. At last, the water level was restored to 7.73 m after the well production was stopped for about 1000 minutes, and it stayed at 7.73 m for several hours. The flow is considered to approach a quasi-steady state in which no significant additional water level change was observed in the well. During the three periods, GPR surveys were conducted repeatedly.

Figure 5 shows the common offset profiles which were acquired along the survey line N when the well production was in the different condition. Figures 5a and 5b look very similar. The horizontal reflections can be observed at around 100-110ns in the two profiles but the reflection strength is different. The residual profile, Figure 5c shows the difference between Figure 5a and 5b. In Figure 5c, the horizontal reflection appears at around 110 ns. This difference is caused by the water level change. It shows a good correspondence with the water level observed in the pumping well. In Figure 5c, the horizontal reflections appear from position x=0m (+3m offset, the distance from the pumping well to the starting point of the survey line) to about x=23m (+3m offset). After x=23m (+3m offset), the horizontal reflections are very weak and we estimate the water level almost did not change.

In the Central area the alluvial aquifer was developed and used for the water supply system of Ulaanbaatar. The previous study by PIIINS (1977) estimated hydraulic conductivity of the aquifer. Hydraulic conductivity of the upper layer (range 10 m to 20 m) varied from 0.122 cm/s to 0.285 cm/s and averaged 0.179 cm/s. A model of the aquifer system should be assumed if we want to estimate the hydraulic properties of the aquifer.

Steady-state conditions of hydraulics of wells in unconfined aquifers are described by Fetter (2001). If a well is pumped for a long period, the water level may reach a state of equilibrium; that is, there is no further drawdown with time. The region around the pumping well where the head has been lowered is known as the cone of depression. When equilibrium has been achieved, the cone of depression stops growing because it has reached a source where the recharge rate equals pumpage. These are also known as steady-state conditions.

Figure 6 shows a well that is penetrating an unconfined aquifer [11]. In the case of steady radial flow in an unconfined aquifer, the assumptions are needed that the well is pumped at a constant rate and equilibrium has been reached. For such a system, Theim (1906) derived an equation for steady radial flow in an unconfined aquifer,

$$K = \frac{Q}{\pi(b_2^2 - b_1^2)} \ln\left(\frac{r_2}{r_1}\right) \qquad (4)$$

Where K is the hydraulic conductivity, Q is the pumping rate, and b_1 and b_2 are the saturated thicknesses at distance r_1 and r_2 from the pumping well.

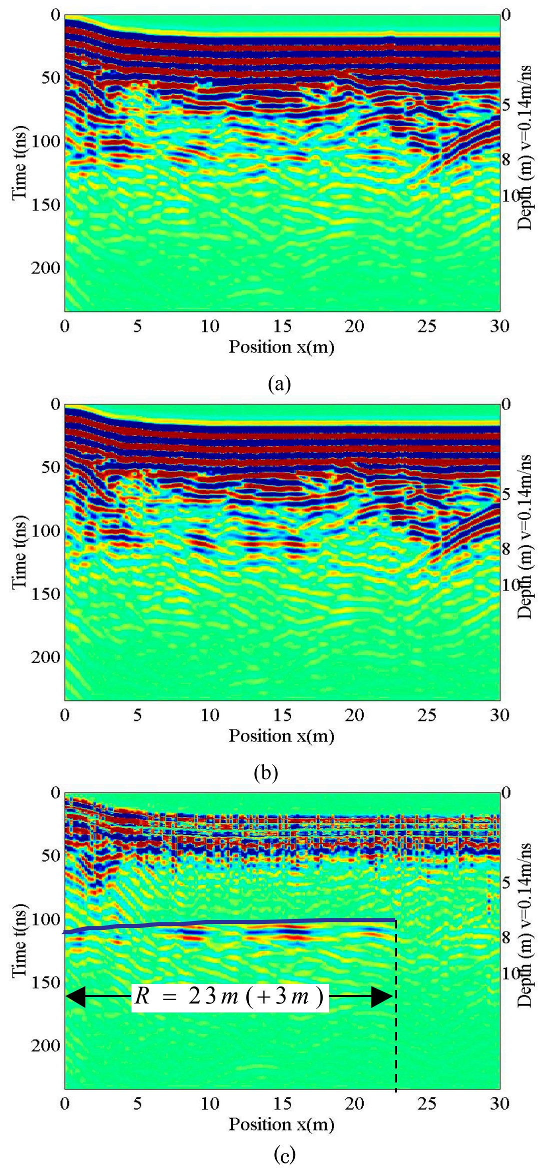

Figure 5. Common offset profiles along the survey line N in April 2002. (a) The well was in full-working condition (water level in the well was 8.25m). (b) The well production was stopped for 1000 minutes (water level in the well was 7.73m). (c) Residual profile of (a) and (b); R=26m, is the radius of influence of the well.

In our case, before the pumping operation of No.10 well was stopped, it had been pumped continuously for a long time for the Ulaanbaatar city water supply. The well was pumped at a constant rate of $75.5\ m^3/h$. The water table in the well was 8.25m. In this instance, we assumed the pump had a steady radial flow in an unconfined aquifer with the assumptions described in the previous section.

Figure 5c, the residual profile shows that the horizontal reflection at around 110ns represents the water level change. The horizontal reflection appears from x = 0 m (+ 3 m offset) to x = 23 m (+ 3 m offset). We estimate this distance represents the radius of influence of the well, that is, the drawdown can not be observed and the water table is 7.73m at this position. CMP analysis defined the water table at 8m at x=15m (+3m offset) when the well production was in the full-working condition. The electrical survey reveals the aquifer thickness to be 54m. The radius of the well No.10 (used as r_1) is 0.2m. Combining the parameters obtained from GPR data with the hydrogeologic data, the hydraulic conductivity can be estimated by equation (4). The estimated hydraulic conductivities are 0.01cm/s, 0.0623cm/s and 0.131cm/s, the average is 0.068cm/s.

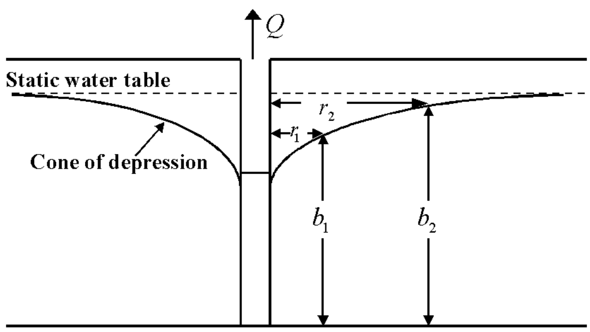

Figure 6. A well in an unconfined aquifer.

Discussion

The model of the aquifer system used in this study to estimate the hydraulic properties combined the GPR data is an ideal model, not a realistic depiction. The major advantages of using this ideal model are that (a) the required inputs are simple, since the model does not require detailed information with hydrogeological structure; and (b) the formulation is easy to compute.

However, this model has some problems. First, because the model of the aquifer system is an ideal model, it is defined using the radially symmetric, horizontal, steady flow. But the correlation is weak between the ideal model and an actual aquifer system. This may account for some of the uncertainties in the results. Additionally, the radius of influence of the well was defined as 26 m, which means the water table depth at this position should be the same as in the well (7.73 m) in the steady-state condition. There are some possible explanations for why the water level change extended that far in Figure 5c: (1) the water level change was very small beyond that distance; (2) it might be visible beyond that distance but the length of the survey line limited it.

Conclusions

To further examine the potential applicability of GPR to hydrogeological applications, we performed GPR surveys by controlling the pumping operation. The GPR technique successfully yielded quantitative information about water level change, and the hydraulic properties could be estimated by combining GPR data and hydrogeologic data.

The results of this study indicate that in this case useful information can be derived from the combination of the common-offset data and the CMP data. The groundwater level change could be quantitatively estimated by comparing the two sets of GPR data acquired under different conditions. The CMP data and velocity analysis provide information on locating the water table. A hydraulic model of the aquifer has been proposed in this study to relate the GPR data to hydraulic properties. Combining GPR data with hydraulic data, the estimation of hydraulic properties showed encouraging results. Quantitative information extracted from the GPR data made GPR a good tool for estimating the hydraulic properties of the aquifer.

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