A new approach to the identification of vertical flow resistance of a partially penetrating river

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Abstract From the point of view of water protection and water conditioning, as well as sewage treatment, the question of the biological reaction capacity of natural or artificial "infiltration systems" is becoming more and more important. At present there is insufficiently precise knowledge available on the limits of the loading capacity of such systems. The answer to this question requires an integrative approach covering the vital activities and the fluxes of materials and energy. The knowledge gained so far is based on analyses of the ecosystem in the laboratory and in pseudo-natural testing facilities. Despite the limited variability of the boundary conditions, the latter are suitable to check the laboratory findings at pseudo-natural conditions so that for some aspects sufficiently precise results could be obtained. The transferability and comparability of these results depends decisively on the standardization of testing and analysing conditions.

INTRODUCTION

In order to predict the dynamic state of groundwater during development or exhaustion, the first step is to identify the hydrogeological model, including hydrogeological parameters and boundary conditions. When the boundary is a partially penetrating river, which does not completely extend to the base of the aquifer studied, and only partially extends into the aquifer or partially (or completely) into the overlying semi-confining bed, the boundary's hydrogeological parameters in the vertical direction have to be determined. Simulation with an $R-C$ network in the usual way requires repeated regulation of the vertical resistivity simulating vertical flow resistance, which is trivial and time-consuming. A new approach is given in this paper to identify vertical resistivity rapidly and precisely without any such limitation. The computation of groundwater development dynamics of Xuzhou, China is taken as an example.

AQUIFER PHYSICAL AND BOUNDARY CONDITIONS

The aquifer consists of Cambrian-Ordovician carbonates, mainly having a north-south strike, and extending about 6 km from east to west and more than 40 km from north to south. Overlying it is a loam bed with a water
table about 20 m above the base. Karstification decreases with depth, only ending at a depth of about 100 m. Along the strike, karst fissures are common in pure carbonate and sections of structural movement. Thus the aquifer is anisotropic, heterogeneous and confined. Its recharge sources are from the river and atmospheric precipitation flowing down through isolated limestone hills, while the discharge is by way of pumped abstraction.

Shown on Fig. 1 are the boundary conditions of the aquifer. The western boundary is composed of Carboniferous-Permian coal-bearing sand-shale. With 120 m water-table difference between the two sides of the boundary when exhausted by the local coal mine, it is believed to be an impermeable boundary. The eastern boundary is strata of Sinian clastic rock and seen along the north-south direction is a series of magma bodies, which make the boundary an impermeable one. Because the aquifer extends far toward south, it is assumed that the place 30 km away from the well functions as an impermeable boundary and simulation results indicate that this is the case. The northern boundary is a partially penetrating river having hydraulic continuity with groundwater, whose water table is known.

Fig. 1 Boundary conditions of the aquifer near Fenghuang Village.

STRATA BENEATH RIVER-BED AND THE RELATION BETWEEN RIVER AND GROUNDWATER

The northern boundary of the aquifer is a perennial river with plentiful water (Fig. 1). In places (Fenghuang Village) limestone is exposed in the river bed, while in other places a loam layer, 3–30 m thick, lies beneath the river bed (Fig. 2). From the coincidence between the water table of borehole D23
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on the bank and the river water level response, it is clearly seen that a close relation exists between the river groundwater near Fenghuang Village (Fig. 3). Because of the short length of the river in the district considered, and the small hydraulic gradient, it is assumed that the water table of the points along the river has nothing to do with their position and only changes with time. Furthermore, the river is a partially penetrating one, and the lithology and thickness, i.e. flow resistances, are different sizes to simulate the different flow resistances. The absolute resistivity has to be determined by identification.

METHOD COMMONLY USED IN SIMULATING VERTICAL FLOW RESISTANCE OF A PARTIALLY PENETRATING RIVER

Assume the limestone outcrops that beneath the river bed; draw an idealized section across the river. Then the discharge from the river to the limestone aquifer is:

\[ Q = \frac{H_A - H_B}{(Z_A - Z_B) / T_z + \beta_z L} = \frac{H_A - H_B}{\beta_z} \]

where:
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\( Q \) is the discharge of section \( L \) of the river (m\(^3\) day\(^{-1}\));
\( H_A, H_B \) are the head at \( A, B \) respectively (m);
\( B_z = (Z_A - Z_B)/(T_z L) \) is the vertical flow resistance coefficient (VFRC) (day m\(^{-2}\));
\( Z_A, Z_B \) are the height of \( A, B \) respectively (m);
\( T_z \) is the transmissivity along vertical flow in limestone (m\(^2\) day\(^{-1}\));
\( L \) is length represented by node beneath the bed along the river (m).

According to similarity, current intensity \( I \) simulates the discharge \( Q \):

\[
I = \frac{(U_A - U_B)}{R_z}
\]

where:
\( U_A, U_B \) are the electric potentials simulating \( H_A, H_B \) (V);
\( R_z \) is the resistor simulating VFRC (\( \Omega \)).

The ratios between the corresponding physical values are

\[
\lambda_1 = H/U, \quad \lambda_2 = R_2/\beta_2, \quad I/Q = \lambda_1/\lambda_2
\]

where \( \lambda_1 \) and \( \lambda_2 \) are known beforehand. \( H_A \) is the height of the water surface (known); \( H_B \) can be approximately replaced by the head measured in borehole D23 (known). Obviously, \( \beta_2 \), the object to be identified, corresponds to \( R_z \) to be determined in resistor model.

The identifying process of \( R_z \) is as follows: at first define the value of \( T_z \) empirically. Then from

\[
R_z = \lambda_2 \frac{Z_A - I_B}{T_z L}
\]

The value of \( R_z \) is calculated, which, as the initial value, is added to the corresponding position in the already installed operation network (see Fig. 4).

**Fig. 4** Operation network \((A-B)\). See text for explanation.

The various pumped (injected) water quantities are transformed into \( I (-I) \), the known boundary condition \( H \) into \( U \), and then \( I \) and \( U \) are input into the network. Repeatedly changing parameters (including \( R \)) until the simulation curve of the observation borehole (including that of borehole D23) is basically fitted with the water-table curve of the borehole measured in situ (the error should be within allowable range), the hydrogeological parameters
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(including B) identified are what is needed. From the above it is clear that $R_z$ has to be changed repeatedly in order to provide $B_z$. To make the procedure less time-consuming and less trivial, we propose another method to get $B_z$ without changing $R_z$.

A NEW APPROACH TO IDENTIFICATION OF VERTICAL FLOW RESISTANCE

The procedure is as follows:
(1) Install a new node $C$ beyond boundary node $B$, and a resistor $R$ can be connected between the two nodes (Fig. 5).
(2) At first give an initial value of $R_{zc}$ as $R_z$; try to make $R_{zc}$ smaller than the $R_z$ to be identified. This is not difficult for an experienced hydrogeologist, and even if $R_{zc} > R_z$, no trouble will occur.
(3) "Extract" current intensity $I_2$ at node $C$, which is equal in effect to a new pumping well beyond the boundary. If the assumed $R_{zc}$ is greater than $R_z$ to be identified, a current intensity $-I$ has to be put in.
(4) According to Kirchof’s law the following relation exists for current intensities at $B$: $I = I_2 + I_3$ (when $R_{zc} < R_z$) or $I = I_3 - I_1$ (when $R_{zc} > R_z$).
(5) Input various values of $I_2$ by key (ordinarily three or four times). By measuring the electrical potential $U_B$, the head at node $B$ can be obtained. When the simulated $H_B$ fits into the measured in situ $H_B$, take down the value of $I_2$ of this time, and from:

$$I_1 = (U_A - U_B) / R_{zc}$$

$I_1$ as well as $I_2$ ($= I_1 - I_2$) can be found. Then the $R_z$ value wanted can be calculated [$= (U_A - U_B) / I_3$]. Finally we can have the VFRC $B_z$ ($R_z = \lambda_z \beta_z$).

Fig. 5 Operation network (A-B-C). See text for explanation.

The idea and procedure described above only takes one node as an example. The same can be done for other nodes on the boundary. After changing $I_2$, instead of directly altering $R_z$, no regulation of $R_z$ is needed, and what we have to do is to input and change current intensity $I_2$ by key. Within several hours, at most one or two days $R_z$ can be identified, while the
old method would take several days or even several tens of days to reach the same result.