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Models of nonlinear stream aquifer transients

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Summary The phenomenon of stream aquifer interaction was investigated via mathematical modeling using the Boussinesq equation. The effect of highly fluctuating transient stream level on the subsequent propagation of hydraulic transients in the adjacent aquifer was quantified using various solutions to the linearized and the nonlinear Boussinesq equation. The semigroup solution of the linearized equation, the spatial-partial decomposition solution of the nonlinear equation, the temporal-partial decomposition solution of the nonlinear equation, and a quasi nonlinear solution were investigated subject to a transient, large-amplitude, periodic boundary condition. The models were verified with a limited numerical solution to the nonlinear Boussinesq equation.

The results indicated that the linearized model compared well with the temporal-partial solution and with a numerical solution, whereas the spatial-partial solution did not adequately reproduce the expected damping in the water table amplitude with distance. The linearized solution was sensitive to the chosen linearized transmissivity. The quasi nonlinear solution did not exhibit the natural attenuation of the head amplitude as distance increased, but could lend itself to practical applications of stream-aquifer interaction calculations subject to discrete-time irregular flood hydrographs. The temporal-partial nonlinear solution is simpler to derive than the linearized solution.

Introduction

Alluvial valley aquifers are hydraulically connected to their adjacent channels and exchange flow through the streambed (Perkins and Koussis, 1996). The connection causes water levels in these systems to fluctuate with respect to the other. If the stream stage increases over a short period

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of time, a flow reversal between the channel and aquifer will occur as a result of a change in the hydraulic gradient (Workman et al., 1997). A flood wave is then propagated into the aquifer and increases bank storage. While the stream is returning to normal flows, the bank storage is released. Alluvial valley aquifers pose some interesting boundary conditions. One side of the flow domain is the river source/sink, which is fluctuating. As a result of the alluvial valley formation, the other side of the flow domain is the valley wall, which is a no flow boundary condition.

The ground water level rise in the great Bend Praire aquifer of Kansas area was studied by Sophocleous (1991) and has been attributed to the pressure waves from stream flooding that propagate in translatory motions through numerous high transmissivity and high hydraulic diffusivity buried channels. Whiting and Pomeranets (1997) studied the saturated flow response in an aquifer due to changes in the water level in the stream channel that cuts across the aquifer. They used the finite element method to solve the boundary value problem and have deduced that the banks can store and later release large volumes of stream water. Flood wave propagation is affected by hydraulic conductivity, specific yield, and the length of time flow reversal conditions.

Analytical solutions to the governing groundwater differential equations have been shown to accurately predict water table profiles subject to smooth boundary conditions. The Boussinesq equation is a nonlinear partial differential equation that describes horizontal, transient flow in an unconfined aquifer and its solution has important hydrological application in the study of unconfined groundwater flow, stream-aquifer interaction and bank storage effects. However, with Dupuit assumptions of negligible vertical flow components, negligible capillary pore forces, and small changes in transmissivity, the Boussinesq equation can be linearized (Van de Giesen et al., 1994; Tallaksen, 1995). The analytical solution to the linearized form of the Boussinesq equation was developed by Govindaraju and Koelliker (1994) to study the water fluxes from stream to the aquifers for any arbitrary shaped flood stage hydrograph. They showed that analytical solutions to the nonlinear equations are not very accurate as compare to the numerical solutions. Guo (1997) developed a new solution technique using the Boltzmann transformation to reduce the Boussinesq equation to a nonlinear ordinary differential equation and the flow between the reservoir and unconfined aquifer is calculated using Newton–Raphson method.

Hornberger et al. (1970) used a fourth order Runge–Kutta numerical technique to solve the nonlinear Boussinesq equation. Fouss and Rogers (1992) developed an iterative computer model that used a numerical solution to the Boussinesq equation to predict subsurface water movement and water table profiles between two adjacent subsurface drains. Previous studies have indicated that the Boussinesq equation can be used to simulate the operation of drainage systems for design purposes (Skaggs, 1976, 1991; Lorre et al., 1994; Serrano and Workman, 1998). Approximate analytical solution to the nonlinear Boussinesq equation that accounts for transient boundary conditions are available in literature (Rasmussen and Anderson, 1959). However, since exact solutions do not exist, most researchers solve the flow problems using available numerical methods.

In this article a comparison of various analytical solutions of the linearized and nonlinear Boussinesq equation is conducted. For the linearized Boussinesq equation a semi group solution is obtained. A quasi nonlinear solution that allows the simulation of hydraulic transients subject to irregular flood waves in the stream is also tested. For the nonlinear Boussinesq equation, two different schemes of Adomian's method of decomposition are derived and compared with experimental data.

Decomposition is now being used to solve deterministic, stochastic, linear or nonlinear equations in various branches of science and engineering (Adomian, 1991, 1994; Srivastava and Singh, 1999; Biazar et al., 2003; Wazwaz, 2000; Wazwaz and Gorguis, 2004). It is one of the few systematic methods available to solve nonlinear differential equations. In groundwater flow problems, this method has been extensively used (Serrano, 1992, 1995a,b, 1997, 2003; Serrano and Adomian, 1996; Adomian and Serrano, 1998), to obtain analytical, and sometimes closed-form, solutions to linear, nonlinear and stochastic problems. A short introduction of the method to groundwater systems can be found in Serrano (1997, 2001). The method has been shown to be systematic, robust, and sometimes capable of handling large variances in the controlling hydrogeological parameters.

Analytical solutions to the transient groundwater flow equation

The governing equation for one-dimensional, lateral, unconfined groundwater flow with Dupuit assumptions is the Boussinesq equation (Bear, 1979):

$$\frac{1}{S}\frac{\partial}{\partial x}\left(Kh\frac{\partial h}{\partial x}\right) = \frac{\partial h}{\partial t}, \quad 0 \leqslant x \leqslant l_x,\ 0 \leqslant t \tag{1}$$

where h is the hydraulic head (L); K is the hydraulic conductivity (L T^{-1}); S is the specific yield; x is the horizontal coordinate (L); and t is the time coordinate (T); and l_x is the horizontal dimension of the aquifer (L). Without loss of generality, in this study we do not consider recharge from rainfall since our focus is on the effect of boundary conditions. The Boussinesq equation also neglects the effects of the capillary fringe and capillary hysteresis. We consider a long and thin aquifer bounded by a river on one side and an impervious boundary on the other. The boundary and initial conditions are

$$\begin{aligned} &h(0,t) = H_1(t) \\ &\frac{\partial h(l_x,t)}{\partial x} = 0 \\ &h(x,0) = h_i(x) = y_0 \end{aligned} \tag{2}$$

where $H_1(t)$ is the time-fluctuating head at the left boundary (L), such that $H_1(0) = y_0$; and y_0 is the initial head across the aquifer (L). Fig. 1 depicts the idealized cross section of the model under consideration. A smooth transient left boundary condition was considered as

$$H_1(t) = y_0 + A\sin\left(\frac{2\pi t}{B} + C\right) \tag{3}$$

where A, B, and C are constants.

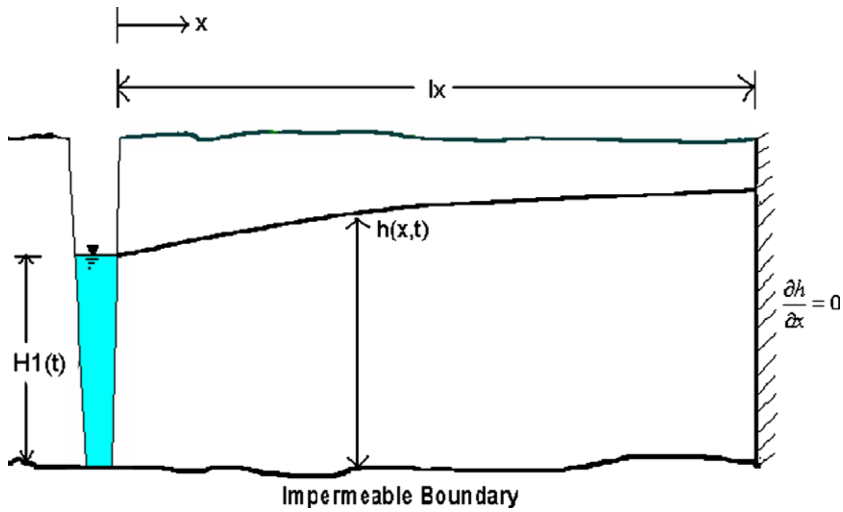

Figure 1 Idealized cross section for the mathematical model.

The semigroup solution of the linearized equation

The governing equation (1) is linearized and is expressed as

$$\frac{\partial h}{\partial t} - \frac{T}{S}\frac{\partial^2 h}{\partial x^2} = 0 \tag{4}$$

where T is the aquifer transmissivity ($L^2\,T^{-1}$) defined as the product of the hydraulic conductivity times an average saturated thickness. The solution may be expressed as the summation of two components (Serrano and Unny, 1987)

$$h(x,t) = V(x,t) + W(x,t) \tag{5}$$

where $W(x,t)$ is the primary solution, and $V(x,t)$ is the secondary solution. Substituting Eq. (5) into Eq. (4) we get

$$\frac{\partial W}{\partial t} - \frac{T}{S}\frac{\partial^2 W}{\partial x^2} = -\left(\frac{\partial V}{\partial t} - \frac{T}{S}\frac{\partial^2 V}{\partial x^2}\right) \tag{6}$$

Subject to

$$W(0,t) = H_1(t) - V(0,t) = W_0 \tag{7}$$

$$\frac{\partial W(l_x,t)}{\partial x} = -\frac{\partial V(l_x,t)}{\partial x} \tag{8}$$

$$W(x,0) = h_i(x) - V(x,0) \tag{9}$$

We choose a smooth function $V(x,t)$ such that the boundary conditions in Eq. (7) become zero:

$$V(x,t) = H_1(t) \tag{10}$$

The solution to Eq. (6) is (Serrano and Unny, 1987)

$$W(x,t) = J_t W_0 - \int_0^t J_{t-\tau}\frac{\partial H_1(\tau)}{\partial \tau}\mathrm{d}\tau \tag{11}$$

where the semigroup operator associated with Eq. (6), J_t, is given as

$$J_t f = \sum_{n=1}^{\infty} b_n(f)\phi_n(x)M_n(t) \tag{12}$$

The Fourier coefficients, b_n, are given as

$$b_n(f) = \frac{2}{l_x}\int_0^{l_x} f\sin(\lambda_n\xi)\mathrm{d}\xi \tag{13}$$

the eigenvalues, λ_n, are

$$\lambda_n = \left(\frac{2n-1}{2l_x}\right)\pi, \quad n = 1,2,3,\ldots \tag{14}$$

the basis function, $\phi_n(x)$, is given as

$$\phi_n(x) = \sin(\lambda_n x) \tag{15}$$

and

$$M_n(t) = e^{-\lambda_n^2 Tt/S} \tag{16}$$

From this solution, the solution of Eq. (6) is obtained as

$$W(x,t) = -\sum_{n=1}^{\infty} b_n(1)\phi_n(x)I_n(t) \tag{17}$$

where

$$I_n(t) = \int_0^t M_n(t-\tau)\frac{\partial H_1}{\partial \tau}\mathrm{d}\tau \tag{18}$$

In terms of h, the solution reduces to

$$h(x,t) = H_1(t) - \sum_{n=1}^{\infty} b_n(1)\phi_n(x)I_n(t) \tag{19}$$

Decomposition solution of the nonlinear equation

The spatial-partial solution

The nonlinear equation (1) subject to the boundary conditions (2) is solved using the decomposition method (Serrano and Workman, 1998). There are several decomposition schemes possible. Let us write Eq. (1) as

$$\frac{\partial^2 h}{\partial x^2} = \frac{S}{Kh}\frac{\partial h}{\partial t} - \frac{1}{h}\left(\frac{\partial h}{\partial x}\right)^2, \quad 0 \leqslant x \leqslant l_x;\ 0 < t \tag{20}$$

The first scheme to be presented is referred to as the spatial-partial decomposition method, sometimes referred to as the x-partial solution (Adomian, 1994). Let us define the operator $L_x = \frac{\partial^2}{\partial x^2}$. Applying the inverse operator L_x^{-1} to (20) we obtain

$$h = k_1(t) + k_2(t)x + L_x^{-1}\frac{1}{h}\left(\frac{S}{K}\frac{\partial h}{\partial t} - \left(\frac{\partial h}{\partial x}\right)^2\right) \tag{21}$$

where the functions $k_1(t)$ and $k_2(t)$ satisfy the boundary conditions. Using decomposition (Adomian, 1991, 1994), the solution to h may be written as the series

$$h = h_0 + h_1 + h_2 + \cdots \tag{22}$$

For the first term in the series, h_0, we choose a smooth function that satisfies the boundary and initial conditions:

$$h_0 = (H_1(t) - H_1(0))\frac{x^2}{l_x^2} - 2(H_1(t) - H_1(0))\frac{x}{l_x} + H_1(t) \tag{23}$$

From (21), the second term is obtained as

$$h_1 = L_x^{-1}\frac{1}{h_0}\left(\frac{S}{K}\frac{\partial h_0}{\partial t} - \left(\frac{\partial h_0}{\partial x}\right)^2\right) \tag{24}$$

Using (23) this becomes

$$h_1 = L_x^{-1}\frac{1}{h_0}\left(\frac{S}{K}\frac{\partial H_1}{\partial t}\left(\left(\frac{x}{l_x}\right)^2 - 2\frac{x}{l_x} + 1\right) - 4\left(\left(\frac{H_1(t) - H_1(0)}{l_x}\right)\frac{x}{l_x} - 1\right)^2\right) \tag{25}$$

Additional terms in the series may be derived in a manner described in Serrano and Workman (1998). For linear and for some nonlinear equations a pattern may be identified and a decomposition series yields a closed-form exact solution. For many nonlinear equations this is not possible. Although the algebraic difficulty in deriving additional terms in the series increases, it is known that the convergence rate of a decomposition series is usually high and that only a few terms are needed in order to assure an accurate model. From the hydrologic point of view, an accurate model is more important than a mathematical closed-form solution. For a rigorous mathematical discussion on the convergence problem of decomposition series, the reader is referred to Abbaoui and Cherruault (1994), Cherruault (1989), and Cherruault et al. (1992). It is also important to mention the rigorous mathematical framework for the convergence of decomposition series developed by Gabet (1992, 1993, 1994). He connected the method of decomposition to well-known formulations where classical theorems (e.g., fixed point theorem, substituted series, etc.) could be used. For a discussion on the convergence of decomposition series of convection–diffusion equations, including a theorem with proof, see Serrano (1998). For additional comparisons between exact and truncated decomposition solutions, see Serrano and Adomian (1996).

The temporal-partial solution

We now attempt a second decomposition, defined as a temporal-partial solution scheme, sometimes referred to as the t-partial solution (Adomian, 1994). The governing equation (1) is expressed as

$$\frac{\partial h}{\partial t} = \frac{K(h_0 + h_1 + \cdots)}{S}\frac{\partial^2(h_0 + h_1 + \cdots)}{\partial x^2} + \frac{K}{S}\left(\frac{\partial(h_0 + h_1 + \cdots)}{\partial x}\right)^2 \tag{26}$$

Defining the operator $L_t = \partial/\partial t$, we can write (26) as

$$h = L_t^{-1}\frac{K(h_0 + h_1 + \cdots)}{S}\frac{\partial^2(h_0 + h_1 + \cdots)}{\partial x^2} + \frac{K}{S}L_t^{-1}\left(\frac{\partial(h_0 + h_1 + \cdots)}{\partial x}\right)^2 \tag{27}$$

Again the solution is expressed as the series $h = h_0 + h_1 + h_2 + \cdots$, where the first term satisfies

$$\frac{\partial h_0}{\partial t} - \frac{Kh_0}{S}\frac{\partial^2 h_0}{\partial x^2} = 0 \tag{28}$$

subject to

$$h_0(0,t) = H_1(t), \quad \frac{\partial h_0(l_x,t)}{\partial x} = 0, \; h_0(x,0) = h_i(x) = y_0 \tag{29}$$

The solution to (28) is the linearized solution to (1), as given by (19), when the transmissivity is taken as $Kh_0(0) = Ky_0$, that is

$$h_0(x,t) = H_1(t) - \sum_{n=1}^{\infty} b_n(1)\phi_n(x)I_n(t) \tag{30}$$

The second term is given by

$$h_1 = L_t^{-1}\left(\frac{Kh_0}{S}\left(\frac{\partial^2 h_0}{\partial x^2}\right) + \frac{K}{S}\left(\frac{\partial h_0}{\partial x}\right)^2\right) \tag{31}$$

Using Eq. (23) it becomes

$$h_1 = \frac{2K}{Sl_x^2}L_t^{-1}\left(h_0(x,t)(H_1(t) - H_1(0)) + \frac{2}{l_x^2}((H_1(t) - H_1(0))(x-1))^2\right) \tag{32}$$

Hence, a two-term approximation to Eq. (20) is given by

$$h(x,t) = H_1(t) - \sum_{n=1}^{\infty} b_n(1)\phi_n(x)I_n(t) + \frac{2K}{Sl_x^2}L_t^{-1}\left(h_0(x,t)(H_1(t) - H_1(0)) + \frac{2}{l_x^2}((H_1(t) - H_1(0))(x-1))^2\right) \tag{33}$$

Quasi nonlinear solution

Miller-VanCleave (1999) has given the quasi nonlinear solution to the Boussinesq equation to simulate the effects of the boundary condition. This solution is similar to that presented in Workman et al. (1997). The hydraulic head is conceived as a summation of various components:

$$h(x,t) = V_1(x,t) + M_1(x,t) + W_1(x,t) + W_2(x,t) \tag{34}$$

where $V_1(x,t)$ is the steady state component, $M_1(x,t)$ is the eventual steady state when the increase in the boundary head has settled in the aquifer; $W_1(x,t)$ is the transient component of the unsettled head of the previous time step (a 'correction' on $V_1(x,t)$); and $W_1(x,t)$ is the transient component caused by the new increase in the boundary head (a ''correction'' on $M_1(x,t)$). The equations for these components are respectively given as

$$V_1(x,t) = H_1(t-1) \tag{35}$$

$$M_1(x,t) = (H_1(t) - H_1(t-1))\left(\frac{l_x - x}{l_x}\right) \tag{36}$$

$$W_1(x,t) = \sum_{n=0}^{\infty}\left[\frac{2}{L}\int_0^{l_x}(h(\xi,t-1) - V_1(\xi,t))\sin(\lambda_n\xi)\mathrm{d}\xi\right] \times \sin(\lambda_n x)\mathrm{e}^{-\frac{\lambda_n T}{S}}, \quad \lambda_n = \frac{n\pi}{l_x} \tag{37}$$

$$W_2(x,t) = \sum_{n=1}^{\infty}\frac{2Sl_x^2\Delta H_1(t)}{Tn^3\pi^3}\sin\left(\frac{n\pi x}{l_x}\right)\left(\mathrm{e}^{-\frac{n^2\pi^2 T}{l_x^2 S}} - 1\right) \tag{38}$$

where $H_1(t-1)$ is the boundary head at time step $(t-1)$; $\Delta H_1(t)$ is the change in boundary head during the time interval $(t, t-1)$; and the rest of the terms as before. This scheme permits the calculation of the solution stepwise in time and the investigation of the propagation of hydraulic transients from natural flood hydrographs that are not as smooth and regular as Eq. (3).

Verification with a numerical solution and a linearized solution

Few exact analytical solutions to the nonlinear Boussinesq equation with time dependent boundary conditions are available for verification. As a preliminary alternative, we compared the nonlinear solutions derived in Section ''Analytical solutions to the transient groundwater flow equation'' with a standard explicit finite difference solution. Although such a numerical solution is inefficient and restrictive in terms of the allowable simulation time step, it serves our purpose. Fig. 2 shows a comparison between a two-term space-partial nonlinear solution equations (23)–(25), the temporal-partial nonlinear solution equation (33), and the numerical solution at $x = 100$ cm. Fig. 3 shows a comparison between the space-partial nonlinear solution equations (23)–(25), the temporal-partial nonlinear solution equation (33), and the numerical solution at $x = 200$ cm. The simulations suggest that the time-partial solution compares reasonably well with the numerical one, except for some minor discrepancies at the wave peaks. On the other hand, the space-partial solution does not compare well with the numerical one. It does not exhibit the natural attenuation of the hydraulic transient as it travels into the aquifer. The quasi nonlinear solution equations (34)–(37), not shown, gave a reasonable approximation of the measured hydrographs, but tended to somewhat over predict the heads peaks. This solution starts with a known head distribution at a given time and forecasts the heads at a future time step. For this reason, it is not a continuous in time analytical solution, but could lend itself to simulations in natural watersheds subject to irregular river hydrographs.

Further verification was done by comparing the nonlinear solutions with the linearized and the numerical solution. Fig. 4 illustrates one of the simulations and a comparison

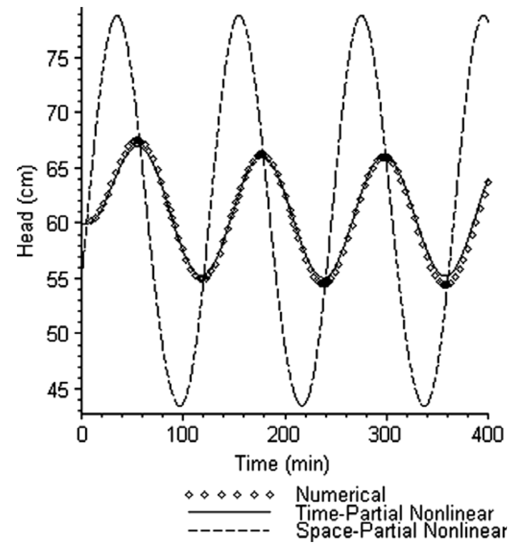

Figure 3 Comparison between nonlinear solutions and an explicit finite-difference solution to Eq. (1) $x = 200$ cm.

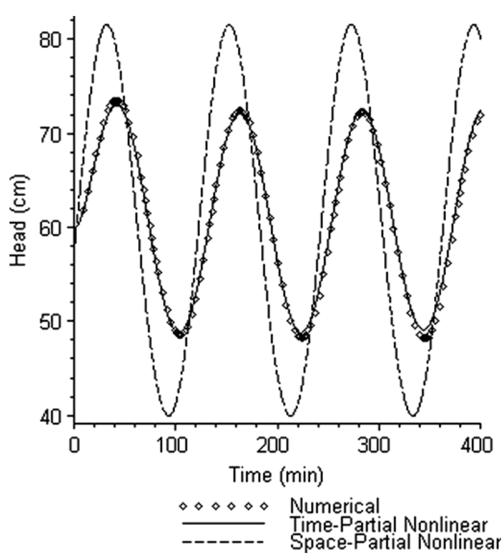

Figure 2 Comparison between nonlinear solutions and an explicit finite-difference solution to Eq. (1) $x = 100$ cm.

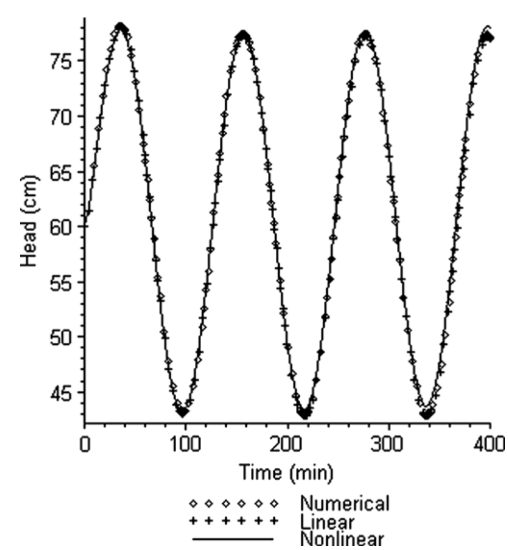

Figure 4 Comparison between linear (Eq. (19)), temporal-partial nonlinear (Eq. (33)), and numerical solutions at $x = 50$ cm.

of the linearized solution (Eq. (19)), the temporal-partial nonlinear solution (Eq. (33)), and the numerical solution at $x = 50$ cm. Fig. 5 illustrates a comparison of the linearized solution, the temporal-partial nonlinear solution, and the numerical at $x = 200$ cm. The linearized solution appears to reproduce well the damping of the hydraulic transient as distance increases, exemplified as a decrease in amplitude with distance.

As before, a two-term nonlinear spatial-partial solution equations (23)–(25) was found not to reproduce well the physical damping in the hydraulic head as x increases. In other words, the spatial-partial solution does not exhibit the significant attenuation in the amplitude of the hydraulic head as distance increases. Thus, the spatial-partial solution tended to significantly over estimate the hydraulic heads and for this reason simulations are not shown. The quasi nonlinear solution equations (34)–(37) gave a reasonable approximation of the measured hydrographs, but tended to somewhat over predict the heads. In this study we focus on one-dimensional solutions to the Boussinesq equation. Hunt (2005) has presented a solution to the two-dimensional equation.

Model output was found to be sensitive to the selected magnitude in the aquifer parameters, K and S. The results indicate that the linear model, and the temporal-partial nonlinear model compare well the numerical model, while the spatial-partial nonlinear model and the quasi nonlinear model tend to somewhat exaggerate the values of the peak hydrographs. The time-partial nonlinear model appears to be a simple model for potential applications to the investigation of the effect of high-amplitude flood hydrographs on adjacent aquifer heads. Fig. 6 shows a simulation of the spatial distribution of heads at different times. It is interesting to observe the ''whipping'' effect as the hydraulic transient propagates into the aquifer. Fig. 7 illustrates the temporal distribution of hydraulic heads damping at different locations in the aquifer.

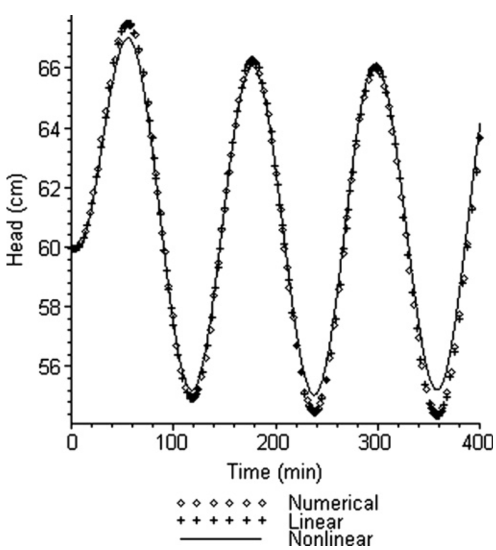

Figure 5 Comparison between linear (Eq. (19)), temporal-partial nonlinear (Eq. (33)), and numerical solutions at $x = 200$ cm.

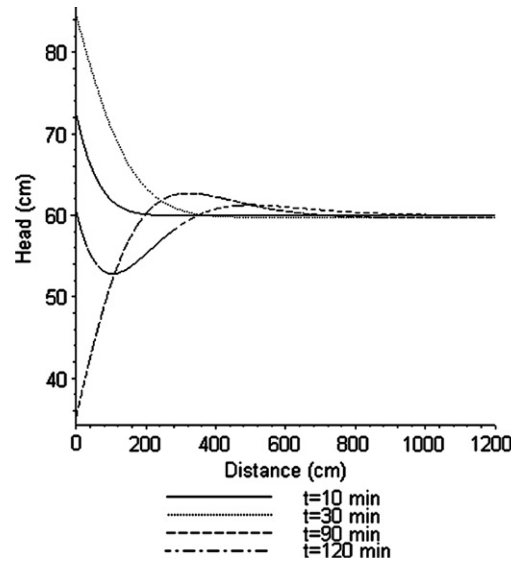

Figure 6 Simulated spatial distribution of heads at various times for the nonlinear solution given by Eq. (33).

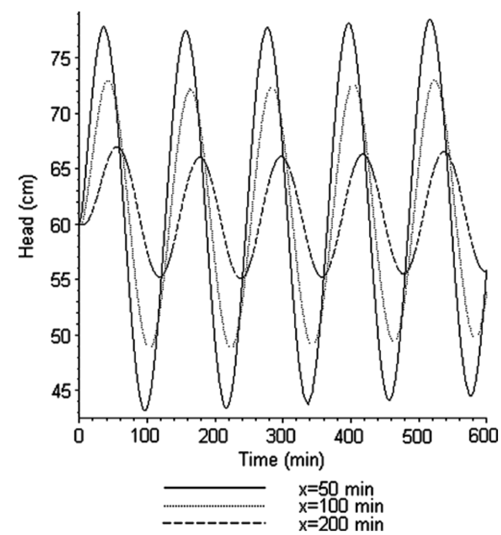

Figure 7 Simulated temporal distribution of heads at various distances from the river for the nonlinear solution given by Eq. (33).

Summary and conclusions

The phenomenon of stream aquifer interaction was investigated via mathematical modeling using the Boussinesq equation. The effect of highly fluctuating transient stream level on the subsequent propagation of hydraulic transients in the adjacent aquifer was quantified using various solutions to the linearized and the nonlinear Boussinesq equation. A periodic analytic function was used to represent the time variability and high fluctuation in the stream-aquifer boundary. The solutions derived included a semi-group solution to the linearized Boussinesq equation, two nonlinear decomposition solutions series expansion, one space-partial and one temporal-partial solution, and a quasi nonlinear solution. The solutions were verified with a standard explicit finite difference solution with reasonable agreement.

The results indicated that the linearized solution and the temporal-partial nonlinear solutions compared well with the numerical solution. The temporal-partial nonlinear solution is simpler to derive than the linearized solution. The space-partial nonlinear solution did not compare well with either the numerical solution, since it did not reproduce the wave damping as it travels into the aquifer. The quasi nonlinear solution reproduced reasonably well the numerical solution, although it tended to over estimate the peak heads in the aquifers. However, this solution could find some practical applications in real watersheds where the head distribution and boundary hydrographs are know at discrete time intervals.

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