# MODIFICATION of STREAM FLOW ROUTING for BANK STORAGE

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*Abstract*

Bank storage is a process in which volumes of water are temporally retained by alluvial stream banks during flood events, and gradually released as baseflows. This process has implications on ground-water resource management, reducing flood peaks, and sustaining riparian vegetation. In this paper, analytical solutions are developed that describe ground water-surface water interactions and the impact of bank storage on the attenuation of stream channel discharges. In effect, the stream flow routing Muskingum method is modified for bank storage. The analysis is based on one-dimensional lateral groundwater flow in semi-infinite homogeneous unconfined aquifers, which are in hydraulic contact with streams through semipervious bed sediments. Storage in the stream reach, according to the Muskingum method, is assumed to be proportional to the reach inflow and outflow rates. The stream-reach impulse response and unit step response functions are modified for bank storage by applying the method of Laplace transformation to the coupled stream-aquifer system. Results indicate that increasing conductivity of the bank reduces the stream impulse response function at earlier times but with greater and more persisting values later. In contrast, greater aquifer conductivity decreases the unit step response at earlier time but with a diminishing effect at later times. The stream flows are routed for a hypothetical asymmetric flood hydrograph, and the results show increasingly attenuated and delayed peaks and extended tailing, with greater values of the hydraulic conductivity. The simulated stream losses to bank storage and subsequent baseflows are significant in typical alluvial sediments. Ground-water discharges are highly dependent on the retardation coefficient.

*Introduction*

The problem of stream-aquifer interactions is important to watershed management efforts aiming at mitigating hazardous flood events and optimizing surface water and ground-water resources, and has significant ecological implications. For example, urbanization increases the fraction of impervious lands in watersheds, which reduces ground-water recharge and increases surface runoff and, thus, the potential for greater stream flows during flood events. Aquifers may provide a temporary relief for increased stream flows through bank storage and, in effect, may reduce and delay flood peaks. The temporally stored volumes of water in stream banks provide

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moisture needed to sustain riparian vegetation, and when released gradually from storage sustain aquatic organisms during baseflow periods.

The lateral stream losses to bank sediments during a flood stage and the subsequent gradual releases of the volume of waters in storage to sustain baseflows between flood events is a process called bank storage (*Todd*, 1955). This problem has been the topic of numerous papers in the literature; it has been analyzed numerically by accounting for streams as time-varying boundary conditions (*Yeh*, 1970; *Mariño*, 1975), and by modeling the stream-aquifer system as two interacting dynamic systems (*Pinder and Sauer*, 1971; *Zitta and Wiggert*, 1971). Although more restrictive, practical analytical solutions were also developed for solving the linearized Boussinesq groundwater flow equation subject to fluctuating stream stages (*Cooper and Rorabaugh*, 1963; *Mariño*, 1973; *Moench et al.*, 1974; *Govindaraju and Koelliker,* 1994; *and Zlotnik and Huang*, 1999). These analytical solutions, however, do not consider the simultaneous interactions, which are inherent to a dynamically coupled stream-aquifer system; rather, their use is conditioned on *a priori* knowledge of the stream-stage fluctuations. Thus, they cannot be used for routing stream flows directly, unless iterative procedures are implemented, and their utility to watershed planning and management (e.g., evaluating the impact of watershed landscape changes) is thereby questionable. *Hunt* (1990) derived an analytical solution, which couples a linearized approximation of the kinematic wave equation in open channels to aquifer flow, and presented a dynamic model for simulating stream-aquifer interactions. The solution, however, required iterating between the groundwater solution and the flood-routing problem. *Harada et al.* (2000) introduced the concept of impulse response function to relate stream outflows to bank storage, by coupling a linear reservoir model to a semi-infinite aquifer in perfect hydraulic connection with the aquifer. The presented analytical solution, however, is limited to narrow streams and aquifer conductivities that are much greater than typically encountered in alluvial sediments. This paper presents an analytical solution to a dynamic stream-aquifer system, by modifying the stream routing Muskingum method (*Chow et al.*, 1988) for bank storage in alluvial aquifers of semi-infinite extent and separated from the streams by semi-pervious bed deposits.

*Stream Flow*

The Muskingum method is a widely used method for hydrologic routing in streams channels. In this method, the volume storage, $S(t)$, in a stream channel is expressed as a combination of a wedge storage, $\eta\ \xi\ [I(t) - O(t)]$, and a prizm storage, $\eta\ O(t)$ (*Chow et al.*, 1988),

$$S(t) = \eta\left[\xi\, I(t) + (1-\xi)\, O(t)\right] \tag{1}$$

in which $S(t)$ = channel storage [$L^3$]; $I(t)$ = inflow rate relative to the initial discahrge [$L^3/T$]; $O(t)$ = ouflow rate relative to the initial discharge [$L^3/T$]; $\eta$ = storage time constant for the reach [T]; and $\xi$ = a weighting factor that varies from 0 to 0.5. In natural streams, $\xi$ varies from 0 to 0.3 and averages about 0.2. The storage time constant, $\eta$, is approximately the kinematic wave travel time through the reach, and can be estimated as the time interval between the inflow and outflow peaks. If $\xi = 0$, $S(t) = \eta\ O(t)$. This storage-discharge relationship is commonly used in level pool routing.

The storage in a stream reach of length $L$ penetrating an aquifer may be described by the continuity equation (Fig. 1):

$$\frac{dS(t)}{dt} = I(t) - O(t) - 2Q(t) \tag{2}$$

in which $Q(t)$ = half the lateral flow rate to or out of the aquifer [$L^3/T$] integrated over the reach length. In the following section, we present the boundary-value problem which describes the ground water-surface water interactions, $Q(t)$. It is assumed that $S(t)$ is related to the average stream stage fluctuations, $H(t)$,

$$S(t) = L\,H(t) \tag{3}$$

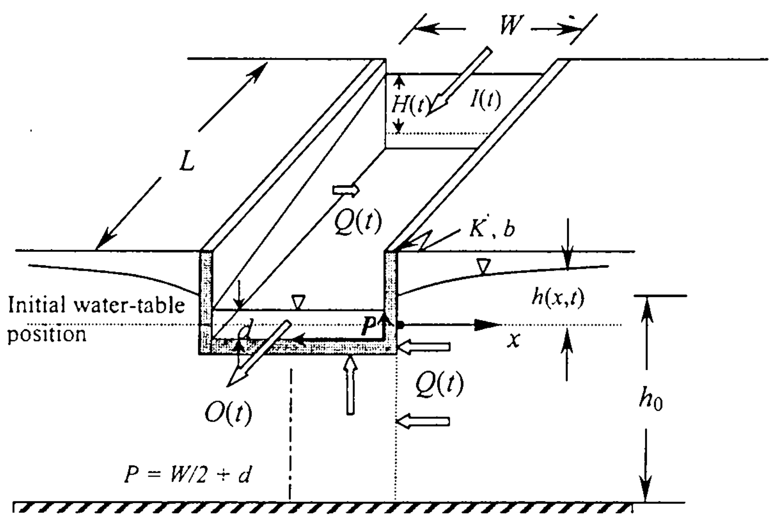


Fig. 1 Schematic diagram of the stream-aquifer system.

in which $H(t)$ = stream-stage fluctuations, relative to the initial equilibrium stage (Fig. 1), averaged over the stream-reach length, $L$. This assumption is intended to simplify the analysis. Also, it is assumed that the stream stage is initially at equilibrium with the water table in the aquifer and, thus, flow in the stream is initially uniform, i.e., $I(0) = O(0) = 0$.

*Ground-water Flow at the Interface*

Ground-water flow in a semi-infinite homogeneous unconfined aquifer may be described by the linearized Boussinesq partial differential equation (Fig. 1):

$$\frac{\partial h(x,t)}{\partial t} = D\frac{\partial^2 h(x,t)}{\partial x^2} \tag{4}$$

and subject to the initial and boundary conditions:

$$h(x,0) = 0 \tag{5}$$

$$-T\frac{\partial h(0,t)}{\partial x} = P\,K'\frac{[H(t) - h(0,t)]}{b} \tag{6}$$

$$h(\infty,t) = 0 \tag{7}$$

in which $h(x,t)$ = water-table fluctuations relative to the initial equilibrium position [L]; $D = T/S$ = aquifer diffusivity [$L^2/T$]; $T$ = K $h_0$ = average transmissivity of the aquifer [$L^2/T$]; $K$ = hydraulic conductivity [L/T]; $h_0$ = average water-table elevation above the base of the aquifer (Fig. 1) [L]; $S$ = specific yield of the unconfined aquifer; $P$ = half the wetted perimeter of the stream [L]; $K'$ = the hydraulic conductivity of the stream bed [L/T]; and $b$ = thickness of the stream bed [L]. In Eq. (4) the water-table fluctuations are assumed to be small relative to the average saturated thickness of the aquifer. Also, the third-type boundary condition (6) assumes that compressibility of the aquifer below the stream can be ignored. Thus, ground-water flow below the stream is at quasi-steady state, which may be valid only for aquifers deeply incised by the streams.

Stream losses (or baseflow) across a channel length $L$ is given by

$$Q(t) = P\,L\,K'\frac{[H(t) - h(0,t)]}{b} \tag{8}$$

In the following section, we use the above equations to derive the impulse response and unit step input response functions for streams modified for stream-aquifer interactions.

*Impulse Response Functions*

It can bee seen that the substitution of the Muskingum relationship (1) for $S(t)$ in (2) results in a linear system, which relates $O(t)$ to $I(t)$ and $Q(t)$. Thus, $O(t)$ is uniquely characterized by its impulse response function and can be described by the convolution integral:

$$O(t) = \int_0^t u(t-\tau)\,I(\tau)\,d\tau \tag{9}$$

in which $u(t)$ = impulse response function [$T^{-1}$], which describes the temporal variations of the outflow from the stream reach due to an instantaneous input of unit amount at $t$ = 0 at the upstream inflow boundary. In the Laplace transform domain, Eq. (9) is described by

$$\tilde{O}(p) = \tilde{u}(p)\,\tilde{I}(p) \tag{10}$$

where the Laplace transform of a function $f(t)$ is defined by $\tilde{f}(p) = \int_0^\infty f(t)\,e^{-pt}\,dt$. The application of the Laplace transformation to Eqs. (1-8) and comparison with (10) should yield

$$\tilde{u}(p) = \frac{1}{(1-\xi)^2\eta}\frac{1+R\sqrt{p/D}}{(R/\sqrt{D})p^{3/2}+p+\gamma(R/\sqrt{D})\sqrt{p}+[1/(\eta(1-\xi)]} - \frac{\xi}{1-\xi} \quad (11)$$

where $\gamma = 1/(\eta(1-\xi)) + (2P/W)(K'/b)$, and $R = Tb/PK'$ is the retardation coefficient [L] modified for partial penetration. In the specific case of $P \approx h_0$, we have $R \approx Kb/K'$, which is equivalent to a stream completely penetrating the aquifer. The inverse Laplace transform of (11) is given by

$$u(t) = \frac{1}{(1-\xi)^2\eta}\frac{1}{2\pi i}\int_{\alpha-i\infty}^{\alpha+i\infty}\frac{1+R\sqrt{z/D}\quad e^{zt}}{(R/\sqrt{D})z^{3/2}+z+\gamma(R/\sqrt{D})\sqrt{z}+[1/(\eta(1-\xi)]}dz - \frac{\xi}{1-\xi}\delta(t) \quad (12)$$

in which $\delta(t)$ is the Dirac delta function. Noting that the denominator in the integrand (we refer to the integrand as $F(z)$) is a multiple-valued function of the complex variable $z$, then the integral is evaluated on the complex plane by introducing a branch cut along the negative real axis (Fig. 2) with the argument of the principal branch of $\sqrt{z}$ defined from $-\pi$ to $\pi$ ($z = re^{i\theta}$ and $-\pi \le \theta \le \pi$):

$$\frac{1}{2\pi i}\int_{\alpha-i\infty}^{\alpha+i\infty}F(z)\,dz = -\frac{1}{2\pi i}\lim_{\substack{\varepsilon\to 0,\\ \rho\to\infty}}\left\{\oint_{C_\rho}F(z)\,dz + \int_{\Gamma_1}F(z)\,dz + \oint_{\Gamma_\varepsilon}F(z)\,dz + \int_{\Gamma_2}F(z)\,dz\right\} \quad (13)$$

The evaluation of the integrals, the details of which are not shown, yields the following integral expression for the impulse response function:

$$u(t) = \frac{1}{(1-\xi)^2\eta}\frac{4}{\pi}\frac{T\sqrt{D}}{W}\int_0^\infty\frac{y^2e^{-y^2t}}{\Lambda(y)}dy - \frac{\xi}{1-\xi}\delta(t) \quad (14)$$

where

$$\Lambda(y) = R^2y^2(\gamma-y^2)^2 + D\left[1/(\eta(1-\xi)) - y^2\right]^2 \quad (15)$$

*Harada et al.* (2000) obtained a closed-form solution for $u(t)$ for the specific case of level pool routing ($\xi = 0$), perfect hydraulic connection with aquifer ($R = 0$), and when $SKh_0/W^2 > 1/\eta$. This latter condition, however, is limited to narrow channel widths and aquifer conductivities much greater than those encountered in natural alluvial sediments. Figure 3 shows the effect of the hydraulic conductivity on the impulse response function (the Dirac-delta contribution is not shown). Notice that as $K$ increases, $u(t)$ shows increasingly smaller values at earlier time $t$ but increasingly greater values at later time $t$ and extended tailing. This implies that bank storage regulates stream discharges by attenuating their peaks due to initial lateral losses to the aquifer and continuously supplying the stream outflows by gradual releases from bank storage through

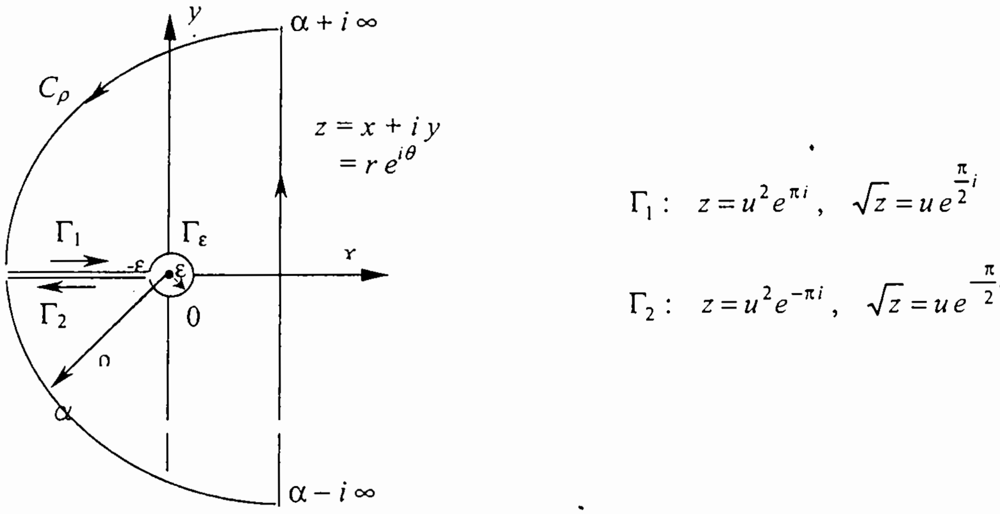


Fig. 2 Contour integration for the Laplace inverse transformation of the multiple-valued function $F(z)$.

baseflows. The Laplace transform of $Q(t)$ can be shown to be

$$\widetilde{Q}(p) = \frac{T}{(1-\xi)W} \frac{\sqrt{p/D}}{(R/\sqrt{D})p^{3/2} + p + \gamma(R/\sqrt{D})\sqrt{p} + [1/(\eta(1-\xi)]} \widetilde{I}(p) \tag{16}$$

From this equation one can deduce that the impulse response function that describes ground water-surface water intearctions is given by the Laplace inverse transform of the product of the first two terms on the right-hand side,

$$u^{*}(t) = \frac{T}{(1-\xi)W} \frac{1}{2\pi i} \int_{\alpha - i\infty}^{\alpha + i\infty} \frac{\sqrt{z/D} \quad e^{zt}}{(R/\sqrt{D})z^{3/2} + z + \gamma(R/\sqrt{D})\sqrt{z} + [1/(\eta(1-\xi)]} dz \tag{17}$$

Similarly, the integration of the complex-valued function is carried along the contour lines shown in Fig. 2, and can be shown to be:

$$u^{*}(t) = \frac{2T\sqrt{D}}{W(1-\xi)\pi} \int_0^\infty \frac{e^{-y^2 t}}{\Lambda(y)} \left( y^2 - \frac{1}{\eta(1-\xi)} \right) dy \tag{18}$$

and $Q(t)$ is given by the convolution integral:

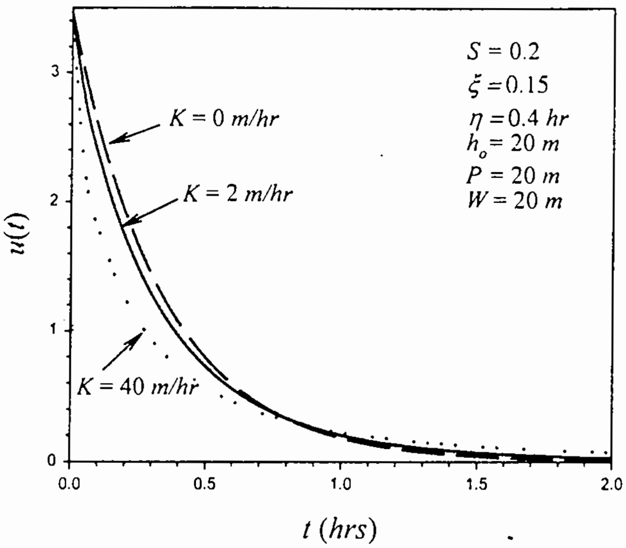


Fig. 3 Impulse response function for $K$ = 0 (no-bank storage), 2, and 40 m/hr.

$$Q(t) = \int_0^t u^*(t-\tau)\, I(\tau)\, d\tau \tag{19}$$

Figures 4(a)-4(b) display the stream outflow and ground-water flow hydrographs in response to a hypothetical asymmetric flood-inflow hydrograph and assumed hydraulic parameters. The inflow hydrograph is assumed to be of the type proposed by *Cooper and Rorabaugh* (1963), $I(t) = N\, I^* e^{-\delta t}\, [1-\cos(\omega t)]$, when $0 \le t \le t_c$, and $I(t) = 0$, $t \ge t_c$. The routed outflow hydrograph, $O(t)/I^*$ ($I^*$ is the peak of the inflow flood event), is obtained by numerical integration of (9) with the impulse response function, $u(t)$, given by (14 and 15), for $K$ = 0, 2, and 40 m/hr. As the aquifer hydraulic conductivity increases, the outflow hydrograph displays greater attenuation and a delayed peak outflow, but with extended tailing; that is, greater baseflow rates proceed after the end of the flood event. Figure 4(b) shows ground-water flux at the interface, $2 \times Q(t)/I^*$, for $K$ = 2 m/hr and $R$ = 0, 5, 20, and 50 m, and $K$ = 40 m/hr and $R$ = 0 m. It is clear that bank storage has greater effect on ground water-surface water interactions as $K$ increases. The hydraulic conductivity $K$ = 40 m/hr may be encountered in predominantly gravelly-coarse sand sediments. Also, note that increased retardation coefficient results in delayed but decreasing baseflow rates. $Q(t)/I^*$ is calculated by numerical integration of (19) with $u^*(t)$ given by (18).

*Unit Step Response Function*

If the stream inflow rate increases from 0 to 1 at time $t$ = 0, and continues indefinitely at that rate, then the outflow rate in response to this unit step increase is given by the unit step response function $g(t)$ (*Chow et al.*, 1988). It is given by the integration of the impulse response function,

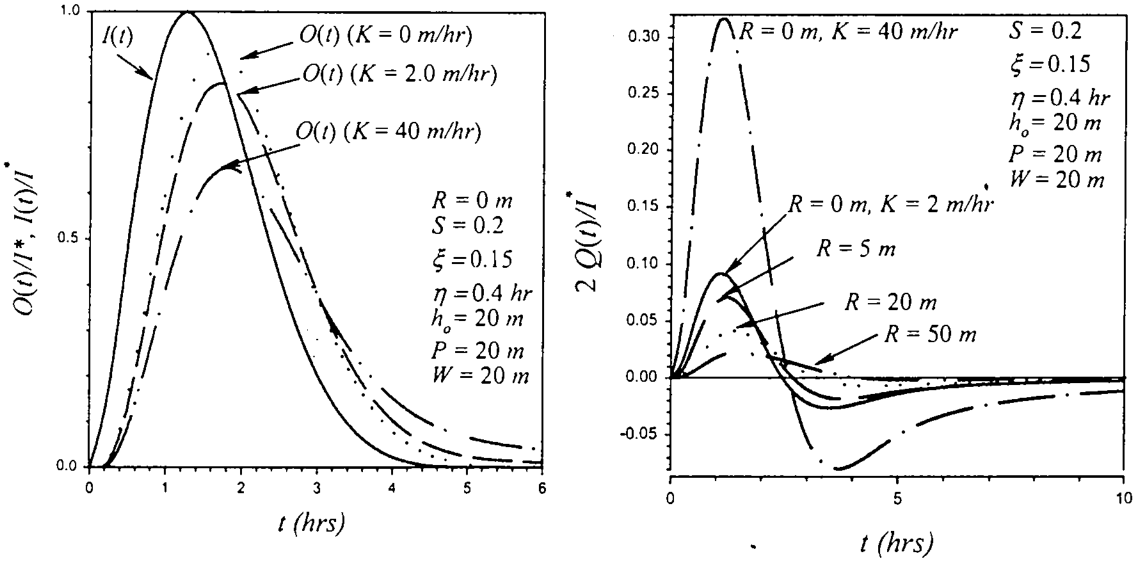


Fig. 4 (a) Outflow hydrograph ($K$ = 0, 2, and 40 m/hr), and (b) Ground-water flow hydrograph at the interface ($R$= 0 and $K$ = 40 m/hr, and $K$ = 2 m/hr with $R$ = 0, 5, 20, and 50m).

$$g(t) = \int_0^t u(\tau)\, d\tau \tag{20}$$

By substituting (14) and (15) for $u(\tau)$ and carrying out the integration one obtains:

$$g(t) = \frac{1}{(1-\xi)^2}\frac{4}{\eta\,\pi}\frac{T\sqrt{D}}{W}\int_0^\infty \frac{1-e^{-y^2 t}}{\Lambda(y)}dy - \frac{\xi}{1-\xi} \tag{21}$$

Figure 5 shows the behavior of the function $g(t)$ for $K$ = 0 (no-bank storage), 2, and 40 m/hr. It can be seen that in response to a unit increase in the inflow rate at the beginning of the flood event, increased aquifer conductivity, thereby bank storage effect, can significantly reduce the outflow from the stream channel during the earlier stages of the event. Also, significant volumes of water entering the channel will be retained in the bank sediments for extended periods of time, as $K$ increases. It can be shown that as $t \to \infty$, $g(t) \to 1$, and shows less variation with $K$. This implies that the impact of bank storage diminishes after large time, assuming that the flood event persists.

The unit step response function $g(t)$ is useful in routing stream flows when the inflow hydrograph is of a general form and measured at discrete points rather than continuously in time, as shown in Fig. 6. In this case, the outflow hydrograph ($O_n = O(t_n)$) at a discrete point in time, $t_n$, can be obtained by breaking the convolution integral in (9) into summation of integrals over

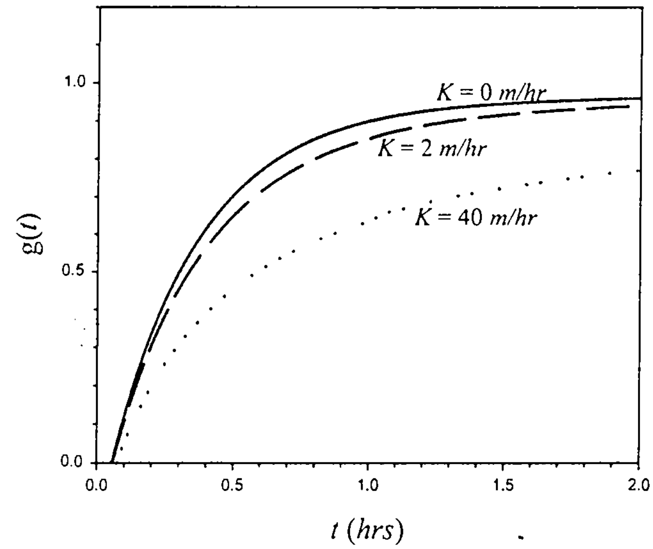


Fig. 5 Unit step response function $g(t)$ as a function of time ($K$ = 0, 2, 40 m/hr).

time increments $\Delta t_m = t_m - t_{m-1}$, between successive discrete measurements of $I(t)$, $I_m$, $0 \leq m \leq M$, as shown in Fig. 6, and integrating (following *Chow et al.*, 1988):

$$O_n = \sum_{m=1}^{n\leq M} P_m \frac{g(t_n - t_{m-1}) - g(t_n - t_m)}{\Delta t_m} \tag{26}$$

where

$$\begin{aligned} P_m &= \int_{t_{m-1}}^{t_m} I(\tau)\, d\tau \\ &= \frac{1}{2}(I_{m-1} + I_m)\, \Delta t_m \end{aligned} \tag{27}$$

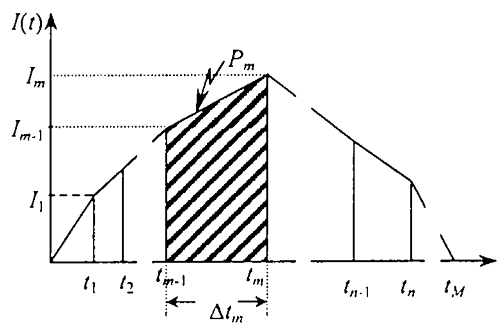


Fig. 6. Illustration of a discrete inflow hydrograph.

*Summary*

Stream-aquifer interactions are important for ground-water resource management, flood control, and sustaining vegetation and healthy ecological conditions in riparian zones. An analytic methodology is presented, which modifies the Muskingum hydrologic routing method for bank storage in streams cutting through alluvial aquifers. Integral expressions are obtained for the impulse response functions, which allow for routing continuous-time inflow hydrographs in streams, and estimating lateral losses to the surrounding bank sediments, and subsequent baseflows. An analytical expression is presented for the unit step response function, which is useful for routing general and discrete inflow hydrographs. Simulation results showed that with K typical to alluvial sediments, bank storage can significantly reduce flood peaks and sustain baseflows. The results also indicated that ground-water fluxes decrease significantly with increasing retardation coefficient, thus minimizing the effect on regulating stream outflows.

*Notice*: This paper has been reviewed in accordance with the U.S. Environmental Protection Agency's peer and administrative review policies and approved for presentation and publication. This research reported in this paper is part of Project 4940-H of the Agricultural Experiment Station of the University of California, Davis.

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16. ABSTRACT
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