Journal of Hydrology 230 (2000) 211–229
Journal of Hydrology
www.elsevier.com/locate/jhydrol

Aquifer response to stream-stage and recharge variations. II. Convolution method and applications

P.M. Barlow[a,*], L.A. DeSimone[a], A.F. Moench[b,1]

[a]US Geological Survey, 10 Bearfoot Road, Northborough, MA 01532, USA
[b]US Geological Survey, Bldg. 15, McKelvey Building, MS 496, 345 Middlefield Road, Menlo Park, CA 94025, USA

Received 20 May 1999; received in revised form 29 December 1999; accepted 1 February 2000

Abstract

In this second of two papers, analytical step-response functions, developed in the companion paper for several cases of transient hydraulic interaction between a fully penetrating stream and a confined, leaky, or water-table aquifer, are used in the convolution integral to calculate aquifer heads, streambank seepage rates, and bank storage that occur in response to stream-stage fluctuations and basinwide recharge or evapotranspiration. Two computer programs developed on the basis of these step-response functions and the convolution integral are applied to the analysis of hydraulic interaction of two alluvial stream–aquifer systems in the northeastern and central United States. These applications demonstrate the utility of the analytical functions and computer programs for estimating aquifer and streambank hydraulic properties, recharge rates, streambank seepage rates, and bank storage. Analysis of the water-table aquifer adjacent to the Blackstone River in Massachusetts suggests that the very shallow depth of water table and associated thin unsaturated zone at the site cause the aquifer to behave like a confined aquifer (negligible specific yield). This finding is consistent with previous studies that have shown that the effective specific yield of an unconfined aquifer approaches zero when the capillary fringe, where sediment pores are saturated by tension, extends to land surface. Under this condition, the aquifer's response is determined by elastic storage only. Estimates of horizontal and vertical hydraulic conductivity, specific yield, specific storage, and recharge for a water-table aquifer adjacent to the Cedar River in eastern Iowa, determined by the use of analytical methods, are in close agreement with those estimated by use of a more complex, multilayer numerical model of the aquifer. Streambank leakance of the semipervious streambank materials also was estimated for the site. The streambank-leakance parameter may be considered to be a general (or lumped) parameter that accounts not only for the resistance of flow at the river–aquifer boundary, but also for the effects of partial penetration of the river and other near-stream flow phenomena not included in the theoretical development of the step-response functions. Published by Elsevier Science B.V.

Keywords: Stream/aquifer interaction; Mathematical models; Confined aquifers; Unconfined aquifers; Seepage; Recharge

1. Introduction

In this second of two papers, analytical step-response functions, developed in the companion paper by Moench and Barlow (2000) for several cases of transient hydraulic interaction between a fully penetrating stream and a confined, leaky, or water-table aquifer, are used in the convolution integral

* Corresponding author. Tel.: +1-508-490-5070; fax: +1-508-490-5068.

E-mail addresses: pbarlow@usgs.gov (P.M. Barlow), ldesimon@usgs.gov (L.A. DeSimone), afmoench@dcrcamnl.wr.usgs.gov (A.F. Moench).

[1] Tel.: +1-650-329-4582; fax: +1-650-329-5590.

0022-1694/00/$ - see front matter. Published by Elsevier Science B.V.
PII: S0022-1694(00)00176-1

Nomenclature

Symbol	Definition [dimension (L, length; T, time)]
a	Streambank leakance [L]
b	Thickness of aquifer (or saturated thickness for water-table aquifer) [L]
d	Thickness of semipervious streambank material [L]
F	System input [L]
F'	Time rate of change of system input [L/T]
h	Head in aquifer [L]
h_D	Dimensionless step-response function for head in aquifer
h_i	Initial water level (or potentiometric surface) in stream–aquifer system [L]
j	Upper limit of time integration
k	Time variable of integration (time step)
K_D	Dimensionless ratio of vertical to horizontal hydraulic conductivity
K_S	Hydraulic conductivity of semipervious streambank material [L/T]
K_x	Horizontal hydraulic conductivity of water-table aquifers or hydraulic conductivity of confined and leaky aquifers [L/T]
K_z	Vertical hydraulic conductivity of water-table aquifers [L/T]
Q	Seepage rate per unit length of stream [L^2/T]
R	Recharge rate [L/T]
S	Storativity (storage coefficient) of aquifer [dimensionless]
S_i	System input [L/T]
S_O	System output [variable]
S_S	Specific storage of aquifer [1/L]
S_y	Specific yield of aquifer [dimensionless]
t	Time [T]
Δt	Time-step size
V	Bank-storage volume per unit length of stream [L^2]
x	Horizontal coordinate [L]
x_D	Dimensionless horizontal coordinate
x_L	Width of aquifer [L]
x_0	Distance from middle of stream to stream–aquifer boundary (half-width of stream) [L]
Y	System response function [variable]
z	Vertical coordinate [L]
τ	Time variable of integration (delay time) [T]

to calculate aquifer heads, streambank seepage rates, and bank storage that occur in response to time-varying stream-stage fluctuations and basinwide recharge or evapotranspiration. Two computer programs have been developed on the basis of the step-response functions and convolution integral for general application to stream–aquifer systems. The programs are used here to evaluate the response of hypothetical confined and water-table aquifers to a passing flood wave. They are then applied to the analysis of two alluvial water-table aquifers common to the northeastern and central United States. These applications demonstrate the utility of the analytical methods and computer programs for estimating aquifer and streambank hydraulic properties, recharge rates, streambank seepage rates, and bank storage.

The convolution integral has been used in surface-water studies to predict runoff from time varying rainfall input over a drainage basin, since the concept of the unit hydrograph was introduced by Sherman (1932). Venetis (1968, 1970) appears to have been one of the first to propose the use of the equation in studies of surface-water/ground-water interaction, although others (e.g. Kraijenhoff van de Leur, 1958; Bedinger and Reed, 1964) have applied a similar approach without explicit reference to the convolution integral. The integral was used to solve the inverse problem by Moench and Kisiel (1970) for ephemeral streams and to solve the direct problem by Hall and Moench (1972), among others, for perennial streams. Most applications of the convolution integral to analysis of stream–aquifer systems have been for the purpose of determining aquifer diffusivity (the ratio of transmissivity to storage) or ground-water-level fluctuations in response to a passing flood wave, for conditions in which it was assumed that semipervious streambank material was absent (Bedinger and Reed, 1964; Pinder et al., 1969; Grubb and Zehner, 1973; Reynolds, 1987; Workman et al., 1997; Serrano and Workman, 1998). Moench et al. (1974), however, applied the convolution integral to the problem of streamflow routing modified by bank storage along the North Canadian River in central Oklahoma. Their study showed that the inclusion of a streambank leakance parameter to account for the effects of resistance to seepage caused by semipervious streambank material substantially improved the simulation of streamflow hydrographs.

2. Convolution integrals

For the linear boundary-value problems discussed by Moench and Barlow (2000), the total response of a stream–aquifer system to a time series of individual stresses (stream-stage fluctuations, recharge, or evapotranspiration) can be determined by superposition (or convolution) of the system's response to the individual stresses. For the assumption of linearity to hold, hydraulic properties of the aquifer and streambank material must remain constant over time, and, for water-table aquifers, changes in head must be small in comparison with the initial saturated thickness of the aquifer. In addition, for water-table aquifers the effects of hysteresis in the unsaturated zone must be negligible.

The convolution integral can be written in general form as

$$S_o(t) = \int_0^t S_i(\tau) Y(t-\tau)\, d\tau, \qquad (1)$$

where S_o is the system output at time t, S_i the system input, Y is the system response function, t the time as measured from the onset of an event, and τ is the time variable of integration (see, e.g. Schwarz and Friedland, 1965, for a detailed treatment). The integral, written for ground-water heads $h(x,z,t)$, is

$$h(x,z,t) = h_i + \int_0^t F'(\tau) h_D(x,z,t-\tau)\, d\tau, \qquad (2)$$

where $h_D(x,z,t)$ is the dimensionless step-response function; $F'(\tau)$ is the time rate of change of the system stress; and τ is the time variable of integration. The step-response functions used in Eq. (2) are the time-domain equivalents of the Laplace transform functions given in Moench and Barlow (2000); Eq. (14) for confined and leaky aquifers and Eqs. (29), (32), or (33) for water-table aquifers.

The convolution integral also is used to calculate time-varying streambank seepage rates by (Hall and Moench, 1972)

$$Q(t) = \frac{K_x b}{x_0} \int_0^t F'(\tau) \frac{\partial h_D(x_0, z, t-\tau)}{\partial x_D}\, d\tau, \qquad (3)$$

where $Q(t)$ is seepage rate per unit length of stream from (or to) one side of the stream; K_x horizontal hydraulic conductivity of the aquifer; b saturated thickness of the aquifer; x_0 distance from the middle of the stream to the stream–aquifer boundary; and x_D is the dimensionless distance x/x_0. Head gradients at the stream–aquifer boundary $[\partial h_D(x_0, z, t)]/\partial x_D$ are the time-domain equivalents of the Laplace transform seepage rates given in Moench and Barlow (2000); Eq. (18) for confined and leaky aquifers and Eq. (36) for water-table aquifers. As used here, streambank seepage rates are positive when flow is from the aquifer to the stream (that is, positive for ground-water discharge) and negative when flow is from the stream to the aquifer.

Bank storage, $V(t)$, is defined as the cumulative volume of water per unit length of stream that has entered the aquifer (or been discharged from the aquifer) from one side of the stream over time t (Cooper and Rorabaugh, 1963, p. 349), and is calculated from

$$V(t) = -\int_0^t Q(t)\,\mathrm{d}t. \tag{4}$$

The negative sign is introduced because bank storage is taken to be a positive quantity when flow is from the stream to the aquifer. A negative value of bank storage indicates there has been a net decrease in ground-water storage over the period t.

3. Computer programs for implementation of analytical methods

Two computer programs were developed to implement the step-response functions and convolution integrals described in this and the companion paper. The programs calculate ground-water head at an observation-well or observation-piezometer, stream-bank seepage rates, and bank-storage volumes in response to arbitrary, time-varying changes in stream stage, recharge, or evapotranspiration. Program STLK1 is used for confined and leaky aquifers and program STWT1 is used for confined and water-table aquifers (Barlow and Moench, 1998). The three types of leaky aquifers to which program STLK1 can be applied are: (1) those in which a source bed with a constant head overlies the aquitard; (2) those in which an impermeable layer overlies the aquitard; and (3) those that are overlain by a water-table aquitard. System response to stream-stage fluctuations can be simulated for all of the confined, leaky, and water-table aquifer types. However, simulation of the response to basin-wide recharge or evapotranspiration is permitted only for water-table aquifers or leaky aquifers overlain by a water-table aquitard. For these aquifer types, recharge and evapotranspiration can be specified alone or in combination with simultaneous changes in stream stage. The programs are written in the FORTRAN-77 computer language and are available from the first author on request.

For implementation in the two programs, the integrals in Eqs. (2)–(4) are written in discretized form as

$$h(x,z,j) = h_{\mathrm{i}} + \sum_{k=2}^{j} F'(k-1)h_{\mathrm{D}}(x,z,j-k+1)\Delta t, \tag{5}$$

$$Q(j) = \frac{K_x b}{x_0}\sum_{k=2}^{j} F'(k-1)\frac{\partial h_{\mathrm{D}}(x_0,z,j-k+1)}{\partial x_{\mathrm{D}}}\Delta t, \tag{6}$$

and

$$V(j) = -\sum_{k=2}^{j} Q(k)\Delta t, \tag{7}$$

where j is the upper limit of time integration; k is the time variable of integration (time step); Δt is a constant time step; and $F'(k-1)$ is the time rate of change of the system input for time step $k-1$.

The programs require an approximation of input hydrographs (continuous records of stream-stage, recharge, or evapotranspiration) into a time series of discrete time steps of length Δt. The time rate of change of the system input for each time step is calculated by

$$F'(k-1) = \frac{F(k) - F(k-1)}{\Delta t}, \tag{8}$$

where $F(k-1)$ and $F(k)$ are the system inputs (stream stage, recharge, or evapotranspiration) at time steps $k-1$ and k, respectively. As with all discretization schemes, the accuracy of the convolution method, and therefore of the programs, is improved by use of smaller time steps.

Recharge is specified in the programs as an increase in ground-water level of the aquifer, relative to the stream stage. This increase can be measured directly, or can be estimated by dividing the amount of recharge to the water table by the specific yield of the water-table aquifer or aquitard. This estimated value is that which occurs under ideal conditions; the actual change in ground-water level resulting from a recharge event will depend on antecedent conditions, the thickness of the unsaturated zone, the height of the capillary fringe, and variations in specific yield resulting from aquifer or aquitard heterogeneity. Evapotranspiration is specified in a similar manner as recharge, but as a decrease rather than increase in ground-water levels in the aquifer.

4. Response of hypothetical confined and water-table aquifers to sinusoidal flood waves

The responses of hypothetical confined and water-table aquifers to sinusoidal-type flood waves were

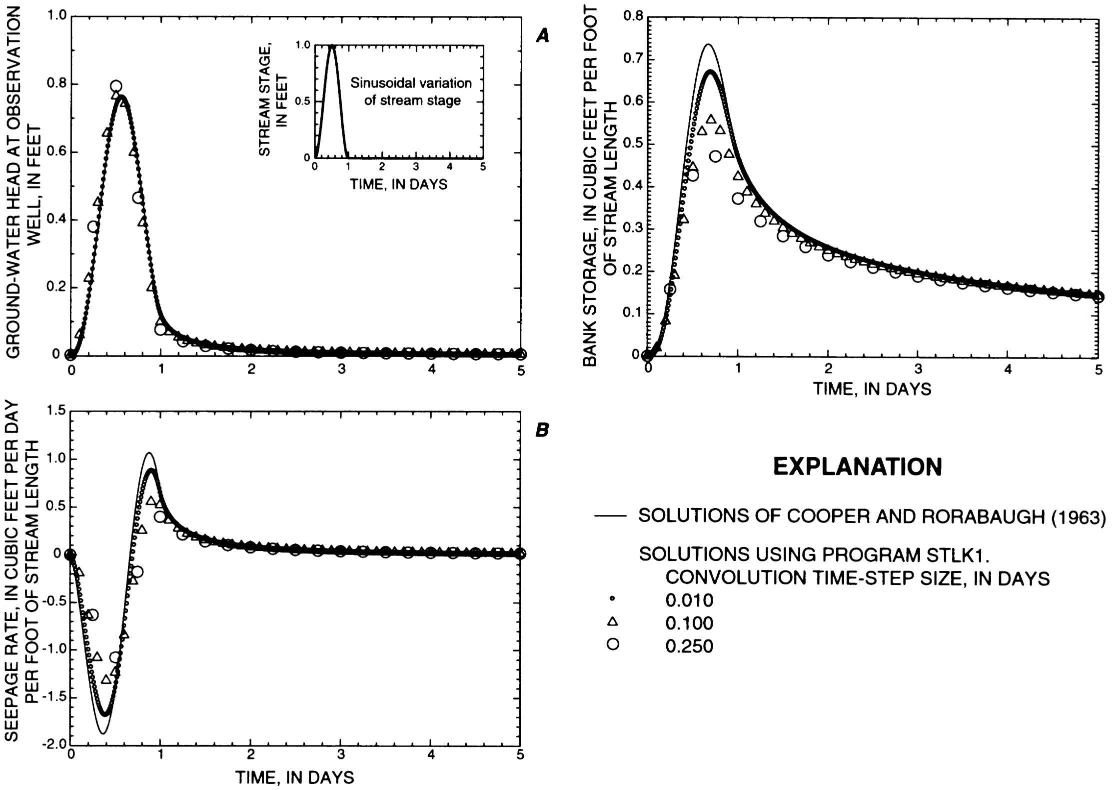

Fig. 1. (A) Ground-water head at observation-well; (B) seepage rate between stream and aquifer; and (C) bank storage in aquifer, for a 1-day sinusoidal flood wave (inset), semi-infinite confined aquifer. Observation-well is 975 ft (297.2 m) from stream–aquifer boundary. (1.0 ft = 3.1×10^{-1} m; 1.0 ft^3/day/ft = 1.1×10^{-6} m^3/s/m).

simulated by programs STLK1 and STWT1 to demonstrate the effect of time-step size on the accuracy of simulation results and the influence of the water table on ground-water-level fluctuations and bank storage. Cooper and Rorabaugh (1963) developed analytical solutions for ground-water heads, streambank seepage rates, and bank storage in a semi-infinite confined aquifer in response to a sinusoidal variation of stream stage. Their closed-form solutions are exact and therefore do not require discretization of the stream-stage hydrograph or the use of the convolution integral. For these reasons, their solutions are useful for testing the accuracy of the discretized convolution integrals used in STLK1 and STWT1.

A semi-infinite confined aquifer with a saturated thickness of 25 ft (7.6 m), hydraulic conductivity of 200 ft/day (7.1×10^{-4} m/s), and specific storage (S_S) of 1.0×10^{-5} ft^{-1} (3.3×10^{-5} m^{-1}) was simulated using STLK1 and the analytical solutions of Cooper and Rorabaugh (1963). The effects of a 1-day sinusoidal flood wave with a peak stream stage of 1.0 ft (0.3 m) were simulated over a 5-day period (Fig. 1, inset). Ground-water heads were calculated at an observation-well located 1000 ft (304.8 m) from the middle of the stream, which is 975 ft (297.2 m) from the stream–aquifer boundary. Simulations were made with STLK1 using three values of the time-step size: 0.010, 0.100, and 0.250 days.

Fig. 1 shows calculated ground-water heads, streambank seepage rates, and bank storage for the simulated conditions. The match between ground-water heads calculated using the solution of Cooper and Rorabaugh (1963) and those calculated using program STLK1 (Fig. 1A) improves as the time-step size is decreased from 0.250 to 0.010 days, as would

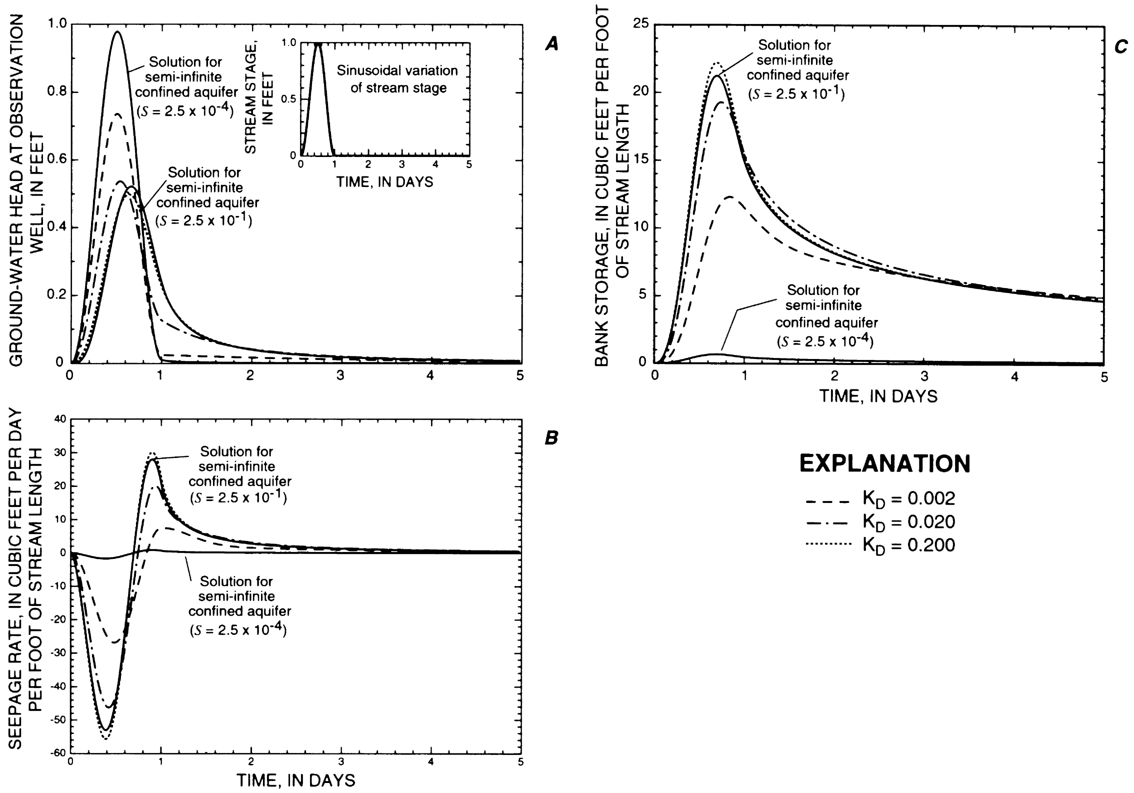

Fig. 2. (A) Ground-water head at observation-well; (B) seepage rate between stream and aquifer; and (C) bank storage in aquifer, for a 1-day sinusoidal flood wave (inset), semi-infinite water-table aquifer. Observation-well is 75 ft (22.9 m) from stream–aquifer boundary; time-step size is 0.010 days. (1.0 ft = 3.1×10^{-1} m; 1.0 ft^3/day/ft = 1.1×10^{-6} m^3/s/m).

be expected. Though differences in calculated heads between the two approaches are insignificant for a time-step size of 0.010 days, differences between the two methods for seepage rates (Fig. 1B) and bank storage (Fig. 1C) can be significant even when using a relatively small time-step size, particularly near times of maximum and minimum seepage rates. These results point to the necessity of using a relatively fine discretization of the stream-stage (or recharge) hydrograph for accurate calculations of seepage rates and bank storage.

In a second series of simulations, the effects of a 1-day sinusoidal flood wave on a semi-infinite water-table aquifer were simulated using STWT1. No closed-form analytical solution is available to which the results of the simulations can be compared. The simulated aquifer has an initial saturated thickness of 25 ft, horizontal hydraulic conductivity of 200 ft/day, specific storage of 1.0×10^{-5} ft^{-1}, and specific yield (S_y) of 0.25. Three separate simulations were made for anisotropic ratios of vertical to horizontal hydraulic conductivity (K_D) of 0.2, 0.02, and 0.002, respectively, and a time-step size of 0.010 days. Ground-water heads were calculated at a fully penetrating observation-well 100 ft (30.5 m) from the middle of the stream, which is 75 ft (22.9 m) from the stream–aquifer boundary. Results for the water-table aquifer conditions were compared to those for a confined aquifer with the same hydraulic properties, aquifer dimensions, and observation-well location as were used for the water-table aquifer, but using the limiting values of aquifer storativity ($S = S_S b$) equal to 2.5×10^{-4} and 2.5×10^{-1}.

Fig. 2 shows calculated ground-water heads, streambank seepage rates, and bank-storage volumes for the simulated conditions; the two solutions for

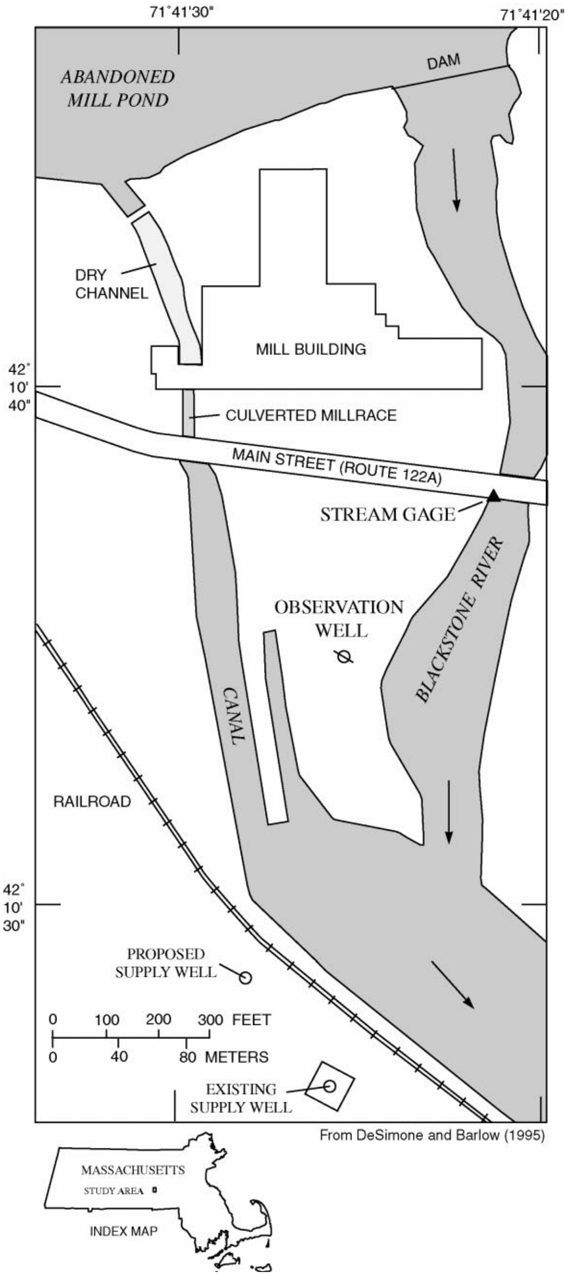

Fig. 3. Blackstone River study site, South Grafton, Massachusetts.

confined aquifers are shown as solid lines. Calculated heads for water-table conditions approach those for the confined aquifer with $S = 2.5 \times 10^{-4}$ as the value of K_D is decreased, and approach those for the confined-aquifer condition with $S = 2.5 \times 10^{-1}$ as the value of K_D is increased (Fig. 2A). Also, calculated seepage rates (Fig. 2B) and bank storage (Fig. 2C) for the water-table aquifer decrease as the value of K_D is decreased. These trends occur due to the increased resistance to vertical flow that results from the smaller values of vertical hydraulic conductivity.

Initially, very soon after the arrival of the flood wave, seepage from the stream is taken into aquifer storage by expansion of the aquifer material and compaction of the pore water (that is, elastic-storage effects). At later times, bank storage in the unconfined aquifers occurs primarily by filling of unsaturated and partially saturated pores above the water table, which is a function of the specific yield of the unconfined aquifer. During these periods, streambank seepage rates and bank-storage volumes for the unconfined aquifers (and confined aquifer with $S = 2.5 \times 10^{-1}$) are substantially larger than that for the confined aquifer with only elastic storage (that is, $S = 2.5 \times 10^{-4}$) (Fig. 2B and C).

5. Applications to field sites

The analytical methods and computer programs described in this and the companion paper have been used to evaluate hydraulic interaction of stream–aquifer systems in the northeastern and central United States (DeSimone and Barlow, 1999). Two of these applications are described in this section for field sites where contamination of hydraulically connected stream–aquifer systems has occurred.

5.1. Blackstone River stream–aquifer system, Massachusetts

Ground water in the stratified-drift aquifer near an abandoned textile mill along the Blackstone River, central Massachusetts (Fig. 3), is contaminated with trichloroethylene, 1,2-dichloroethene, vinyl chloride, and other volatile organic compounds (HMM Associates, 1993, 1994; DeSimone and Barlow, 1995). Hydraulic interaction of the aquifer and Blackstone River are of concern at this site because of possible discharge of the contaminant plume to the river and the potential for contamination of nearby municipal supply-wells through induced infiltration. Ground-water levels, which vary from 3 to 5 ft (0.9 to 1.5 m) below land surface in the stratified drift

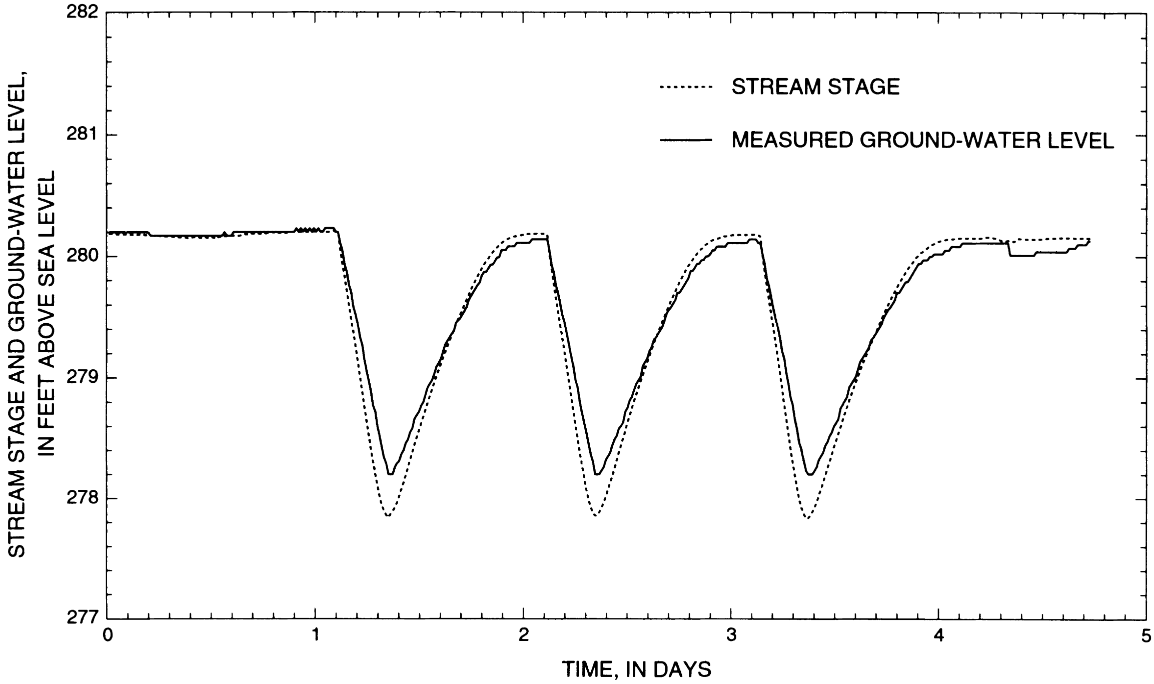

Fig. 4. Stream stage and ground-water levels measured in an observation-well located 95 ft (29 m) from the streambank in the Blackstone River stratified-drift aquifer, South Grafton, Massachusetts. (1.0 ft = 3.1×10^{-1} m).

adjacent to the river, fluctuate rapidly in response to changes in stream stage caused by operation of a small hydroelectric facility about 1 mi (1.6 km) downstream from the study site (Fig. 4). Ground-water levels and stream stage measured in September 1994 were used with STWT1 to evaluate hydrologic conditions in the aquifer and to estimate hydraulic properties of the stratified drift and streambank materials.

The stratified drift was deposited by glacial-melt-water streams and forms a narrow valley aquifer along the length of the Blackstone River. The aquifer is bounded laterally by till and bedrock uplands and is about 0.4 mi (0.6 km) wide at the mill site. Recharge to the aquifer is from precipitation and inflow from the adjacent till and bedrock. Ground water in the stratified drift is generally under water-table conditions (Walker and Krejmas, 1986). The stratified-drift deposits at the study site consist of about 20–50 ft (6–15 m) of coarse to medium sand and gravel or coarse to fine sand and gravel with traces of silt (HMM Associates, 1993). The sand and gravel deposits are underlain by about 5–20 ft (2–6 m) of dense, sandy glacial till and granitic or schistose bedrock (HMM Associates, 1993).

The Blackstone River penetrates approximately 10–20% of the saturated thickness of the aquifer and has a semipervious streambed. Vertical hydraulic conductivity of the streambed sediments determined by seepage-meter measurements averaged 1.4 ft/day (4.9×10^{-6} m/s) (Whitman and Howard, 1990); no information is available on streambed thickness. The river averages about 150 ft (45.7 m) in width at the site. Formerly, the main stem of the river was freely connected to a pond by a canal that runs through a culvert beneath the mill building to Route 122A (Fig. 3). Currently the pond is partially dry, and the water moves through the canal only during extremely high river discharge. Backwater from the river fills the canal, south of Route 122A. Fine-grained, organic-rich streambed sediments line the canal bottom and are assumed to have negligibly small hydraulic conductivity. The river is generally gaining, except where pumping wells induce infiltration of river water into the aquifer. Annual mean discharge of the Blackstone River about 15 mi (24 km) downstream

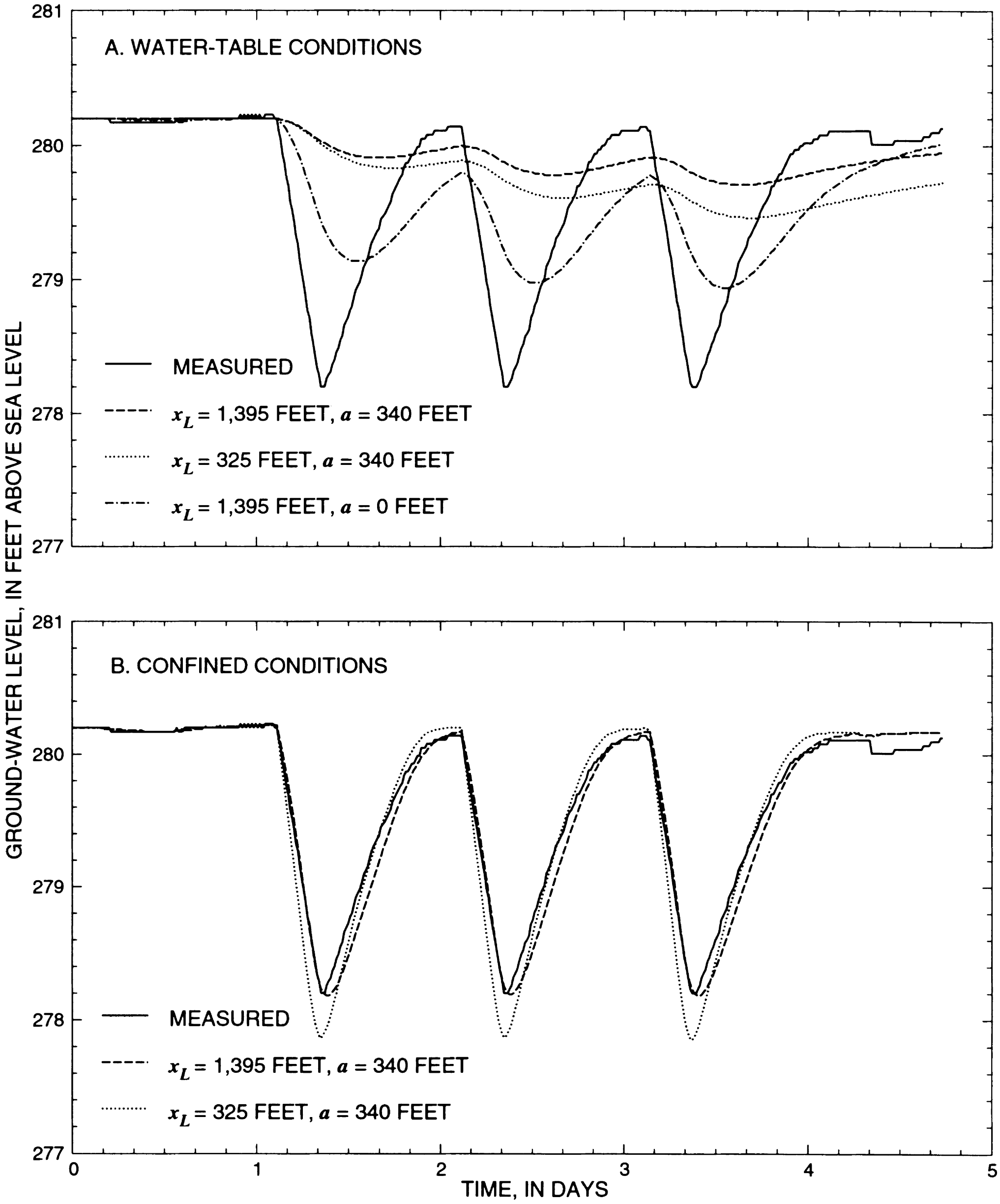

Fig. 5. Calculated ground-water levels in the Blackstone River stratified-drift aquifer, South Grafton, Massachusetts, in response to three daily stream-stage fluctuations under: (A) water-table and (B) confined conditions. Observation-well is 95 ft (29 m) from the streambank and screened near the water table. Model conditions for (A) are: horizontal hydraulic conductivity, 200 ft/day (7.1×10^{-4} m/s); ratio of vertical to horizontal hydraulic conductivity, 0.1; and specific yield, 0.2.

from the study site for the period 1929–1996 was 775 ft^3/s (22.0 m^3/s) (Socolow et al., 1997).

Based on information from Whitman and Howard (1983, 1990), BSC Engineering (1986), and HMM Associates (1994), the aquifer was initially simulated by use of program STWT1 as a finite-width water-table aquifer with semipervious streambank material. A value of horizontal hydraulic conductivity equal to

200 ft/day (7.1×10^{-4} m/s) was specified on the basis of the transmissivity estimated for the aquifer at nearby supply-wells (Whitman and Howard, 1983) and an average saturated thickness of 47.6 ft (14.5 m) determined from lithologic information at the site. Two lateral (no-flow) boundaries were tested: one at the stratified-drift/upland boundary located 1395 ft (425 m) west from the center of the stream channel, and a second at the canal located 325 ft (99 m) from the center of the stream channel. Ground-water levels were simulated at an observation-well located 95 ft (29 m) from the streambank (Fig. 3) and screened near the water table. Stream stage (at the gage shown in Fig. 3) and ground-water levels measured at 15-min intervals were available for the analysis (DeSimone and Barlow, 1995). Because precipitation was negligible during the study period, recharge was not simulated.

As a part of the calibration process, various values of the hydraulic properties of the aquifer and of the streambank material consistent with available data for the site were tested. In all cases, the amplitude and frequency of measured ground-water level fluctuations were underestimated by the analytical model when water-table conditions were assumed (Fig. 5A). However, when confined conditions were simulated, calculated water levels matched measured values closely (Fig. 5B). It was also found that the fine-grained material that lines the canal bottom effectively acts as a hydraulic barrier between the canal and underlying aquifer. Similar calibration results (not shown) were obtained using ground-water levels measured in a well also located 95 ft from the streambank but screened near the bottom of the aquifer, and in a well 250 ft (76 m) from the streambank and screened near the bottom of the aquifer. Ground-water-level fluctuations measured in the shallow and deep-wells at a distance of 95 ft from the streambank were nearly equal, which suggests that horizontal ground-water flow predominates at that location in the aquifer.

Hydraulic properties of the stratified drift and streambank material estimated by calibration of the analytical model for confined-aquifer conditions were K_x, 200 ft/day; S_s, 2.0×10^{-5} ft^{-1} (6.6×10^{-5} m^{-1}); and streambank leakance (a), 340 ft (104 m). It should be noted that it would not be possible to determine a unique value of the hydraulic conductivity or specific storage of the aquifer in the calibration process without independent knowledge of either parameter or of streambank seepage rates. In the absence of such information, it is only possible to determine an effective diffusivity for the aquifer, which, on the basis of the calibration of the analytical model, is 1.0×10^7 ft^2/day (1.1×10^1 m^2/s). The availability of independently determined estimates of transmissivity and saturated thickness at the site, however, from which an independent estimate of K_x was made, provides some assurance that the values of K_x and S_s have been uniquely determined by the calibration process.

Though lithologic evidence indicates that the aquifer is under water-table conditions, calibration of the analytical model suggests that the aquifer's rapid response to stream-stage fluctuations apparently reflects elastic-storage effects of confined conditions rather than gravity drainage associated with movement of the water table. These results may be related to the shallow water table, which is only about 3–5 ft below land surface across most of the site and is likely to be even closer to land surface near the river. When the unsaturated zone is thin, a water-table aquifer may behave like a confined aquifer if the capillary fringe, where sediment pores are saturated by tension, extends nearly to land surface (Bouwer and Rice, 1978, 1980). Gillham (1984), Sophocleous (1985), and Narasimhan and Zhu (1993) demonstrate that the effective specific yield of a water-table aquifer decreases with decreasing thickness of the unsaturated zone when near-saturated conditions exist close to the land surface. They suggest or infer that the effective specific yield is zero when the capillary fringe extends to land surface and that the aquifer's response to recharge or stream-stage stresses is determined by elastic storage only. When the water table declines under these conditions, the aquifer may behave like a confined aquifer until the head in the aquifer decreases below the air entry pressure of the sediments (Gillham, 1984) and effects of delayed drainage take effect. The thickness of the capillary fringe varies with grain size—for example, from 0.5 ft (0.2 m) in coarse sand to greater than 3 ft (0.9 m) in coarse silt. Silt-rich layers were common in the stratified-drift aquifer at the Blackstone River site (HMM Associates, 1993). Therefore, it is possible that the capillary fringe may have been fairly thick in some areas. Saturated conditions also were likely to have extended

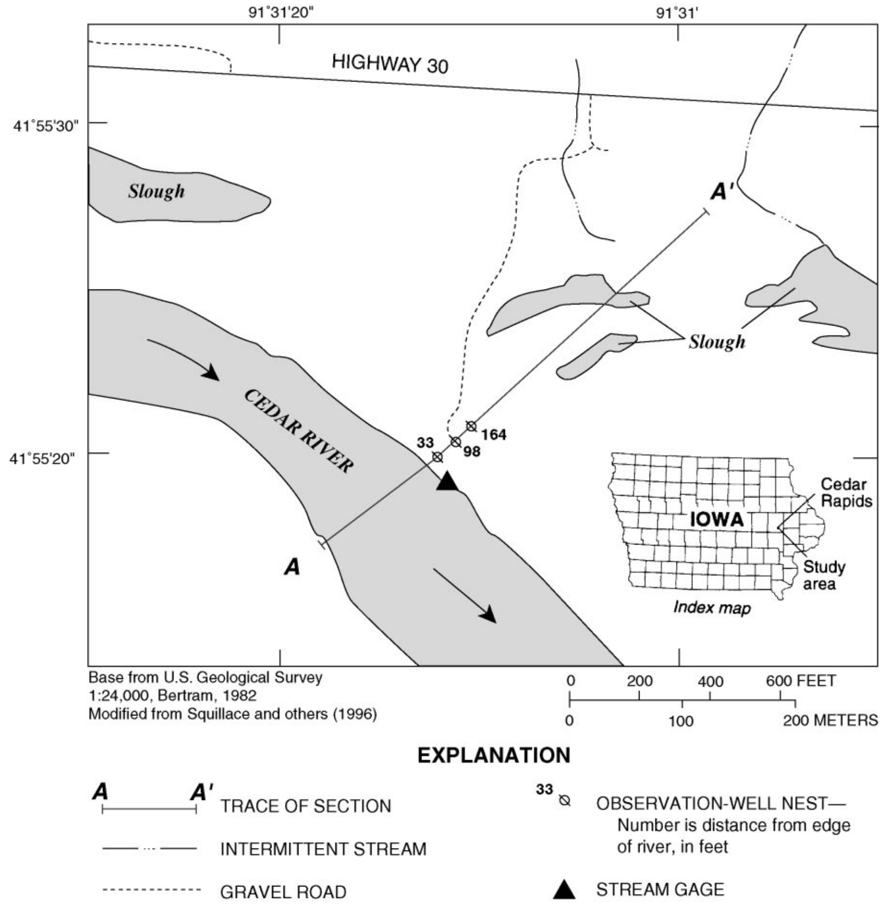

Fig. 6. Cedar River study site near Cedar Rapids, Iowa.

nearly to the land surface adjacent to the river, where the topography is generally flat.

Whether the aquifer behaves like a confined or water-table aquifer in response to stream-stage fluctuations could have implications for contaminant transport and chemical transformation processes at the site because of large differences in calculated seepage rates and bank-storage volumes that occur for the different aquifer types. Maximum seepage rates determined for the period of analysis for water-table conditions were nearly 2.0×10^2 ft^3/day per foot of stream length (2.2×10^{-4} m^3/s per meter of stream length), whereas those calculated for the calibrated confined-aquifer model were about 1.0×10^1 ft^3/day per foot of stream length (1.1×10^{-5} m^3/s per meter of stream length). The large seepage rates calculated for water-table conditions also are associated with large changes in bank storage for each daily stream-stage fluctuation. The smaller seepage rates that occur for confined conditions would result in less mixing of contaminated and uncontaminated water at the margins of the plume and a less dynamic hydrologic regime near the river than would occur for water-table conditions.

5.2. Cedar River stream–aquifer system, Iowa

Chemical and hydrologic evidence indicate that atrazine and other agricultural chemicals are transported into the alluvial water-table aquifer adjacent

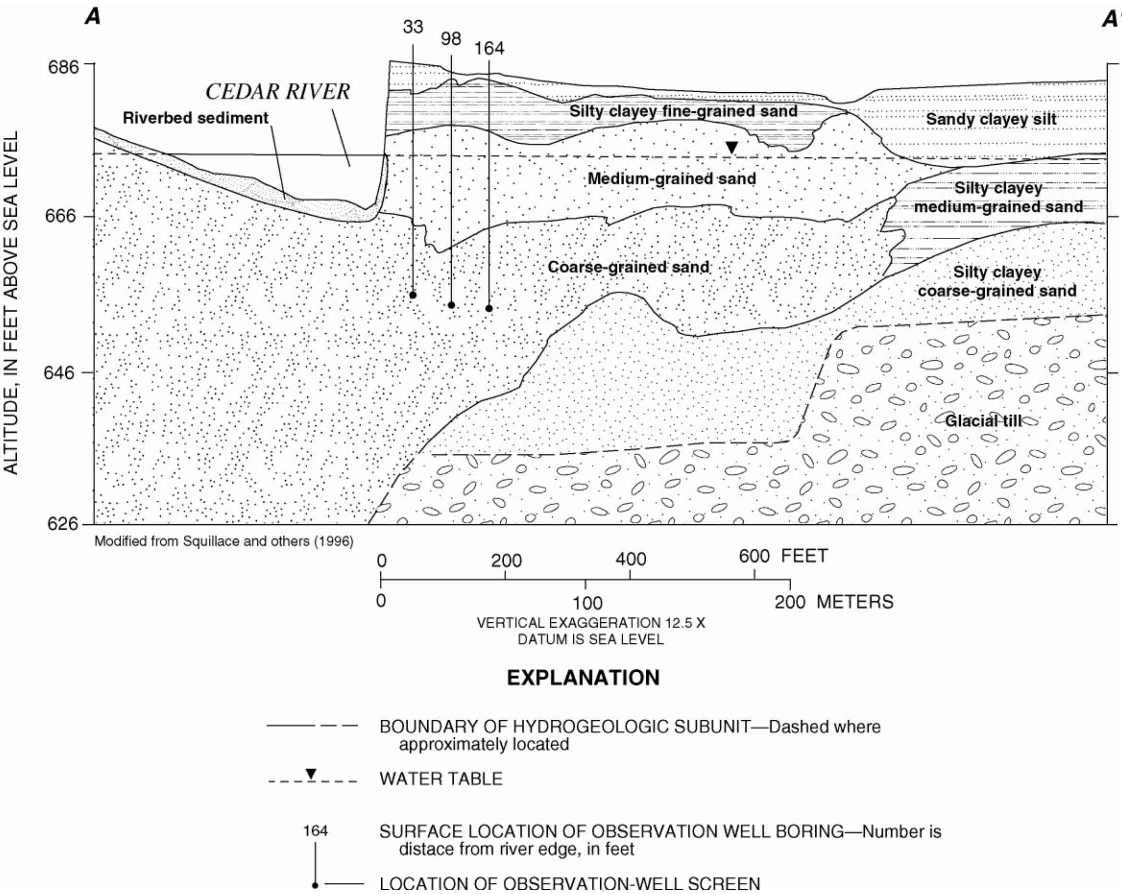

Fig. 7. Hydrogeologic section of the Cedar River study site near Cedar Rapids, Iowa. (1.0 ft $= 3.1 \times 10^{-1}$ m).

to the Cedar River in eastern Iowa (Fig. 6) during periods of elevated stream stage caused by direct surface runoff (Squillace, 1996; Squillace et al., 1996). The aquifer then discharges water and chemicals to the river during declining stream stage. River stage and ground-water levels measured at a site about 6 mi (10 km) southeast of Cedar Rapids during a 1-day period of rapid change in river stage were used to estimate hydraulic properties of the aquifer and semipervious streambank material by calibration of an analytical model of the site. The calibrated model was then used to estimate aquifer recharge rate, streambank seepage rates, and bank-storage volumes during a 55-day period of simultaneous river-stage fluctuations and recharge in March and April 1990.

The study site (Fig. 6) is a wooded area on a flood plain of the Cedar River (Squillace et al., 1996). The alluvial aquifer at the site consists of glaciofluvial sediments that form a vertically heterogeneous, sand-rich, fining-upward sequence. The alluvium reaches a maximum thickness of about 50 ft (15 m) beneath the river, and thins laterally (Fig. 7). The aquifer is underlain by unfractured, low-permeability glacial till and carbonate bedrock and is bounded laterally by till-covered uplands at a distance of about 1310 ft (399 m) from the streambank. Recharge to the aquifer is from precipitation, leakage from ephemeral streams and ponds, bank storage from the Cedar River, and infiltration of floodwater from the river. Outflow from the aquifer is by ground-water discharge to streams and by evapotranspiration. The Cedar River at the study site is about 350 ft (107 m) wide and penetrates about 20% of the saturated

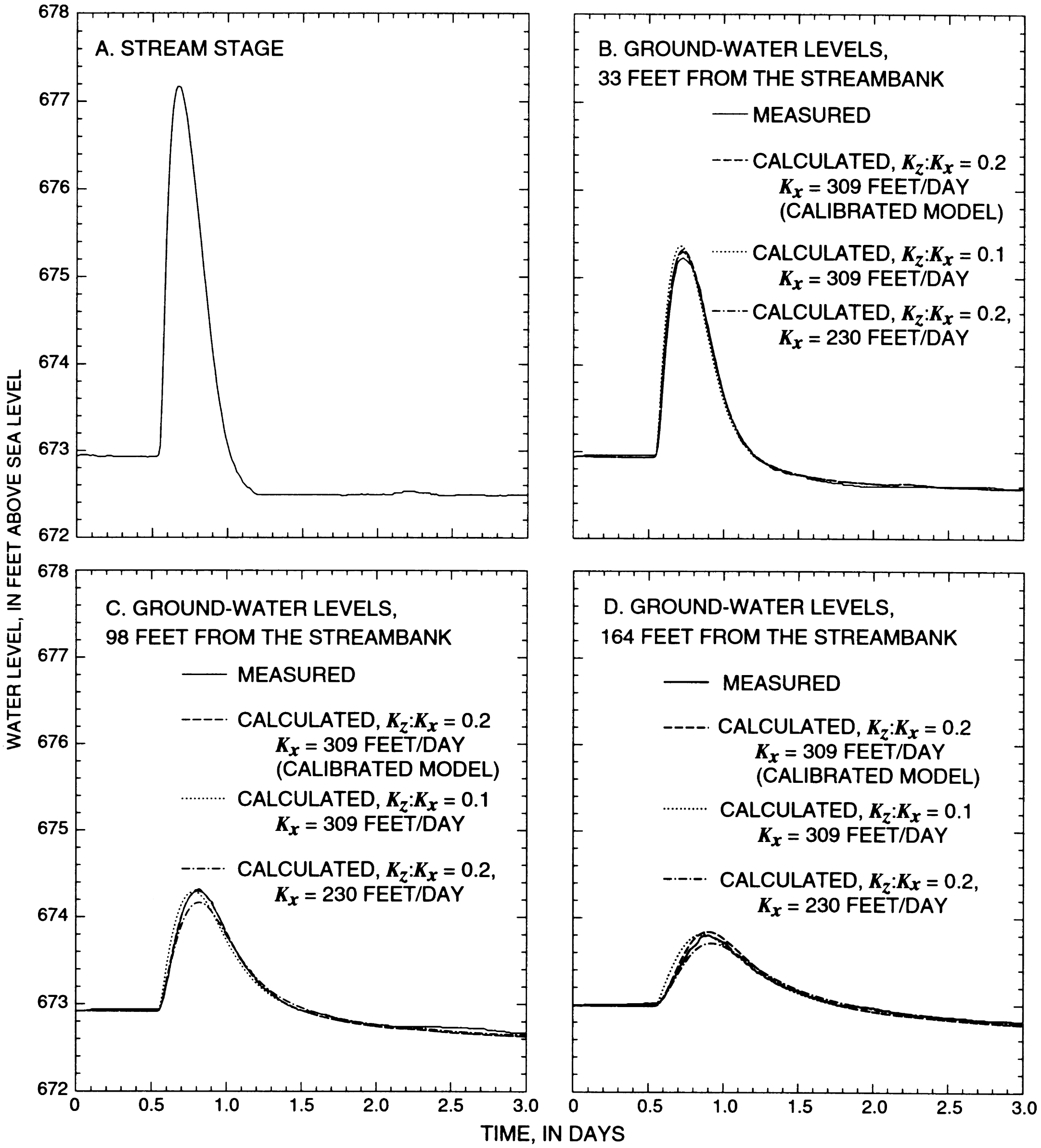

Fig. 8. Stream stage and calculated and measured ground-water levels in observation-wells located three distances from the streambank in the Cedar River alluvial aquifer near Cedar Rapids, Iowa, for a 1-day stream-stage fluctuation. (A) Stream stage; (B–D) ground-water levels, (B) 33 ft (10 m); (C) 98 ft (30 m); and (D) 164 ft (50 m) from the streambank. (1.0 ft = 3.1×10^{-1} m; 1.0 ft/day = 3.5×10^{-6} m/s).

thickness of the aquifer. Squillace et al. (1996) identified a thin layer of sediment along the riverbank and riverbed (Fig. 7) that has grain-size characteristics that are generally finer than the medium- and coarse-grained sand units of the aquifer. Annual mean discharge of the Cedar River at Cedar Rapids was 3687 ft^3/s (104 m^3/s) for the period 1903–1997 (May et al., 1998).

The hydraulic connection between the river and aquifer at the study site is indicated by ground-water levels measured in partially penetrating observation-wells that fluctuate in response to river-stage

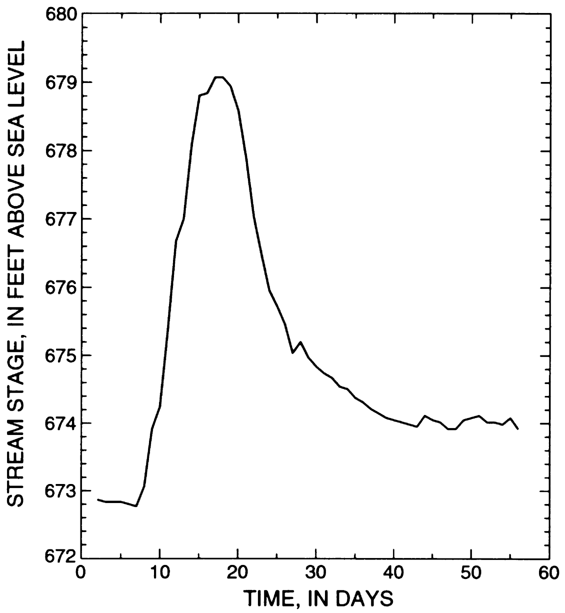

Fig. 9. Stream stage of the Cedar River at the study site, March–April 1990. (1.0 ft = 3.1×10^{-1} m).

fluctuations. This situation occurred during a period in October–November 1989 when the river stage rose and fell 4 ft (1.2 m) in response to a sudden release of water from an upstream dam (Fig. 8). The rapid increase in river stage and ground-water levels followed a period of dry weather during which base-flow conditions occurred (Squillace et al., 1996). Ground-water levels shown in Fig. 8 were measured at 15-min intervals (P.M. Schulmeyer, US Geological Survey, written communication, 1997) in observation-wells located 33, 98, and 164 ft (10, 30, and 50 m) from the streambank and screened in coarse-grained sand at about 30 ft (9 m) below land surface.

The aquifer was simulated using STWT1 as a finite-width water-table aquifer bounded by semipervious streambank material. Initial estimates of aquifer and streambank properties were available from slug-test analyses and calibration of a multi-layered, cross-sectional numerical model of the site (Squillace et al., 1996). A horizontal hydraulic conductivity of 309 ft/day (1.1×10^{-3} m/s) was used for the simulated aquifer. This value is a depth-averaged horizontal hydraulic conductivity of the medium- and coarse-grained-sediment units adjacent to the river near the wells (Fig. 7); these units contain about 70% of the simulated section of aquifer. Fine-grained sediments above the water table and near the upland end of this section, and also the underlying glacial till were not simulated using STWT1. The initial saturated thickness was taken to be 30 ft, based on geohydrologic sections presented in Squillace et al. (1996). The streambed leakance (a) was calculated to be 31 ft (9.5 m) by using values of streambank thickness (d) of 1.6 ft (0.5 m), streambank hydraulic conductivity (K_S) of 16.0 ft/day, and horizontal hydraulic conductivity (K_x) of 309 ft/d. The values for streambank thickness and hydraulic conductivity are equal to those used for the calibrated numerical model of the site (Squillace et al., 1996). The relatively small value of a used in the analytical model (31 ft) indicates that the riverbank and riverbed materials appear to cause little resistance to flow across the stream–aquifer interface.

The best-fit, calculated ground-water levels closely matched those measured at the well 98 ft from the streambank and slightly overestimated the peak water levels measured at wells 33 and 164 ft from the streambank (Fig. 8B–D). The best-fit ground-water levels were obtained using values of aquifer and streambank properties that were equal to, or only slightly different from, those used in the calibrated numerical model. Calibrated hydraulic properties of the analytical model are: K_x, 309 ft/day; K_D, 0.2; a, 31 ft; S_y, 0.2; and S_S, 3.0×10^{-5} ft^{-1} (9.8×10^{-5} m^{-1}). These values are in good agreement with those determined for the coarse-grained layers by the numerical model. The calibrated stream–aquifer model differed from the numerical model in that the ratio of vertical to horizontal hydraulic conductivity was higher in the stream–aquifer model (0.2) than in the numerical model (0.1). When a value of $K_D = 0.1$ was tested in the analytical model, calculated ground-water levels increased more rapidly than measured water levels in all three wells (Fig. 8B–D). A higher value of $K_D = 0.2$ may have been required in the analytical model because the numerical model was calibrated to conditions in which the water table may have moved into the overlying fine-grained sediments, which had a lower vertical hydraulic conductivity than the coarse-grained sediments.

As shown in a hydrogeologic section of the study site (Fig. 7), the Cedar River penetrates only about one-fifth to one-third of the coarse-grained aquifer.

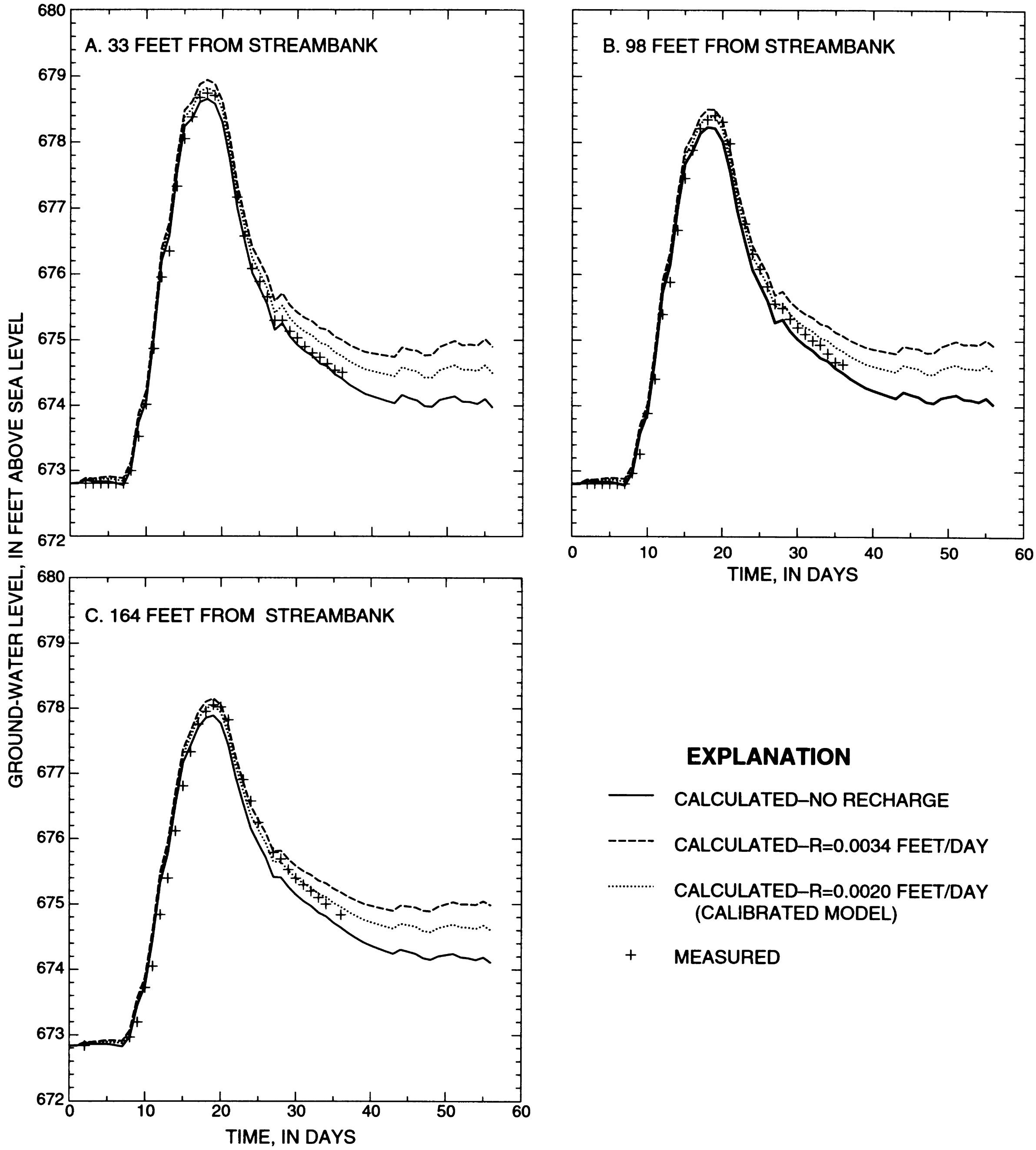

Fig. 10. Calculated and measured ground-water levels in observation-wells located three distances from the streambank in the Cedar River alluvial aquifer near Cedar Rapids, Iowa, for a simultaneous 55-day stream-stage fluctuation and recharge. (A) 33 ft (10 m) from streambank; (B) 98 ft (30 m) from streambank; (C) 164 ft (50 m) from streambank. (1.0 ft = 3.1×10^{-1} m; 1.0 ft/day = 3.5×10^{-6} m/s).

Though the numerical model of the site simulated the river as partially penetrating the aquifer, the analytical model developed here, by necessity, simulated the river as penetrating the full thickness of the aquifer. The close agreement between measured and model-calculated ground-water levels indicates that the assumption that the river fully penetrates the aquifer appears to be reasonable for the analysis of ground-water level fluctuations at the site. However, the good agreement may also be due, at least in part, to the use of the streambank-leakance parameter in the analytical model. This parameter may be considered a 'lumped' parameter that accounts not only for the increased resistance to flow at the streambank caused by semipervious streambank material but also for the partial penetration of the stream. Consequently, the calibrated value of streambank leakance in the analytical model may not be truly representative of the hydraulic properties of the streambank material.

Aquifer and streambank hydraulic properties estimated for the aquifer from simulation of the 1-day river-stage fluctuation in October–November 1989 were then used to simulate the response of the alluvial aquifer to simultaneous river-stage fluctuations and recharge that occurred during a 55-day period of precipitation and surface runoff in March and April 1990 (Fig. 9). Ground-water levels at the three observation-wells rose nearly 6 ft (2 m) in response to the precipitation and surface runoff (Fig. 10; data from Schulmeyer et al., 1995). Initially, a recharge rate of 3.4×10^{-3} ft/day (1.2×10^{-8} m/s), which equals the rate used in the numerical model of the site (Squillace et al., 1996), was specified. However, the match between measured and calculated ground-water levels at the three observation-wells was improved using a recharge rate of 2.0×10^{-3} ft/day (7.1×10^{-9} m/s) (Fig. 10). This lower recharge rate may be more representative of actual recharge, which was determined for a nearby alluvial aquifer (Hansen and Steinhilber, 1977) to be about one-half of the value used in the numerical model. Calculated ground-water levels were underestimated when only the river-stage fluctuation, but not recharge, was simulated. Thus, simulation of recharge improved the match between measured and calculated ground-water-level fluctuations.

The simulated recharge resulted in increased hydraulic gradients toward the river compared to those for the conditions of no recharge. As a result, ground-water discharge rates to the river following the flood wave were greater (Fig. 11A) and the total volume of bank storage in the aquifer was lower (Fig. 11B) than without recharge. Seepage rates and bank-storage volumes calculated with STWT1 (for a recharge rate of 2.0×10^{-3} ft/day) agree well with those estimated with the numerical model of the site, though it should be noted that a lower recharge rate was used in the analytical model than in the numerical model. The maximum bank-storage volume calculated with STWT1, 689 ft^3 per foot of stream reach (64 m^3 per meter of stream reach) (Fig. 11B), is only about 10% greater than the maximum bank storage volume estimated with the numerical model, 624 ft^3 per foot of stream reach (58 m^3 per meter of stream reach) (Squillace et al., 1996). Maximum seepage rate through the streambank was estimated by STWT1 as 79 ft^3/day per foot of stream reach (8.7×10^{-5} m^3/s per meter of stream reach) (Fig. 11A). This value is about 50% greater than the maximum rate of seepage, 57 ft^3/day per foot of stream reach (6.3×10^{-5} m^3/s per meter of stream reach), estimated by the numerical model. Differences in seepage rates calculated by the analytical and numerical models may be due to differences in the recharge rates used in the two models or to use of the streambank-leakance term in the analytical model to account for partial penetration of the stream.

6. Summary and conclusions

Analytical step-response functions, developed in a companion paper for several cases of transient hydraulic interaction between a fully penetrating stream and a confined, leaky, or water-table aquifer, are used in the convolution integral to calculate aquifer heads, streambank seepage rates, and bank storage in response to stream-stage fluctuations and basinwide recharge or evapotranspiration. Two computer programs developed on the basis of these step-response functions and the convolution integral are applied to the analysis of two alluvial stream–aquifer systems that are characteristic of those common in the northeastern and central United States. The program designated STLK1 is for confined and leaky aquifers, and the program designated STWT1 is for confined

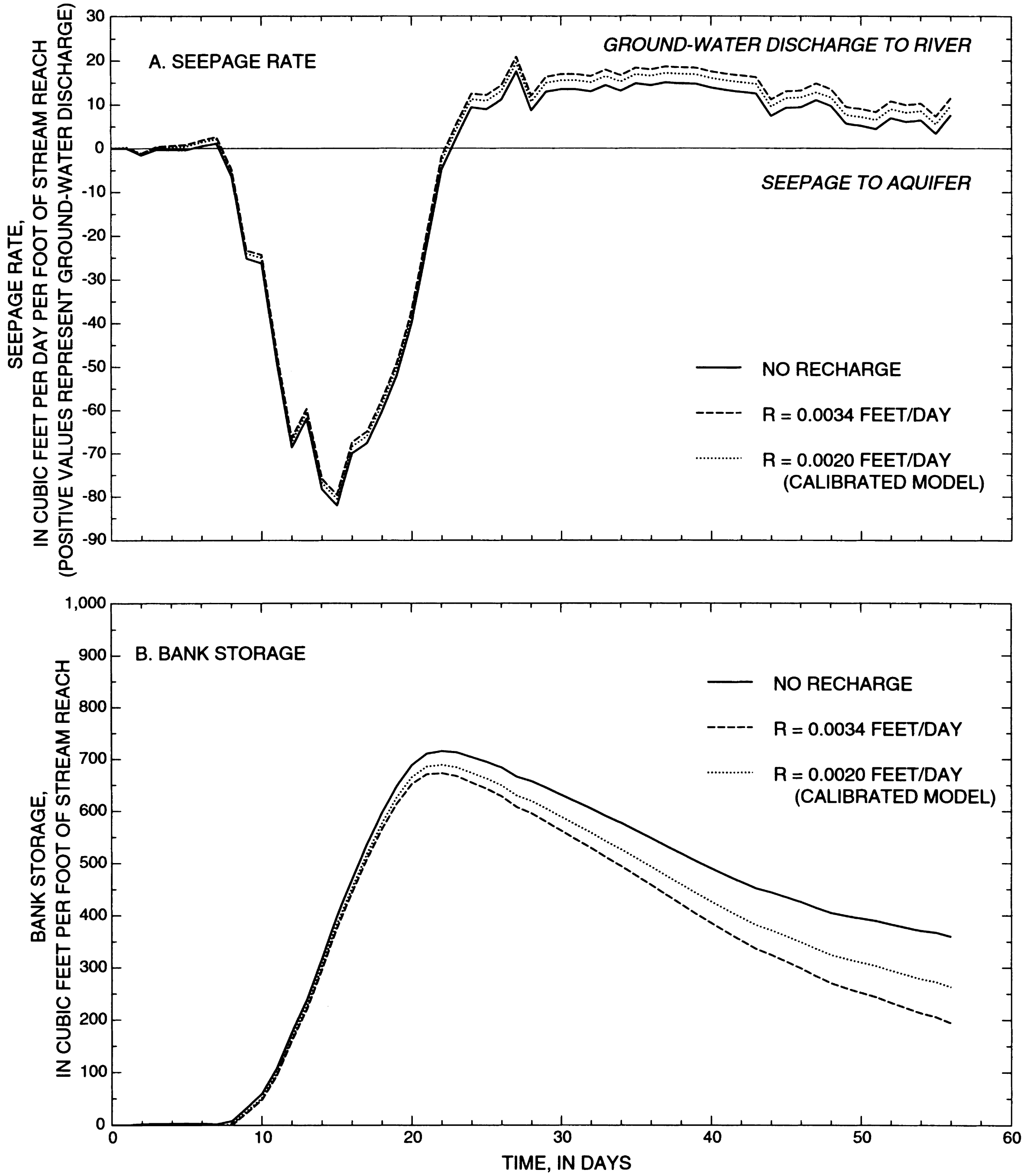

Fig. 11. Calculated seepage rate and bank storage in the Cedar River alluvial aquifer near Cedar Rapids, Iowa, in response to a simultaneous 55-day stream-stage fluctuation and recharge. (A) seepage rate; (B) bank storage. (1.0 ft = 3.1×10^{-1} m; 1.0 ft/day = 3.5×10^{-6} m/s; 1.0 ft^3/day/ft = 1.1×10^{-6} m^3/s/m).

and water-table aquifers. The analytical methods described in this and the companion paper differ from previous applications of the convolution integral primarily by: (1) the wide range of aquifer types to which the methods can be applied; (2) allowing for simultaneous simulation of stream-stage fluctuations and basinwide recharge or evapotranspiration for water-table aquifers and leaky aquifers overlain by a water-table aquitard; and (3) their implementation in readily available computer programs.

Applications of the analytical methods and computer programs to two field sites where contamination of hydraulically connected stream–aquifer systems has occurred demonstrates their utility for estimating aquifer and streambank hydraulic properties, recharge rates, streambank seepage rates, and bank storage. In the first application, analysis of the water-table aquifer adjacent to the Blackstone River in Massachusetts suggests that the very shallow depth to water table and the associated thin unsaturated zone at the site cause the aquifer to behave like a confined aquifer (negligible specific yield). This finding is consistent with previous studies that have shown that the effective specific yield of an unconfined aquifer approaches zero when the capillary fringe (where sediment pores are saturated by tension) extends to land surface, and the aquifer's response is determined by elastic storage only. Hydraulic properties of the aquifer and streambank at this field site, determined by use of the analytical methods, are a diffusivity equal to 1.0×10^{7} ft^2/day (1.1×10^{1} m^2/s) and streambank leakance equal to 340 ft (104 m).

In the second application, estimates were made of the hydraulic properties and recharge rate to an alluvial water-table aquifer adjacent to the Cedar River in eastern Iowa. The estimated values are: horizontal hydraulic conductivity, 309 ft/day (1.1×10^{-3} m/s); ratio of vertical to horizontal hydraulic conductivity, 0.2; specific yield, 0.2; specific storage, 3.0×10^{-5} ft^{-1} (9.8×10^{-5} m^{-1}); and recharge, 2.0×10^{-3} ft/day (7.1×10^{-9} m/s). Streambank leakance at the site was determined to be 31 ft (9.5 m). These estimated values, determined by use of the analytical methods, were in close agreement with those determined by use of a more complex, multi-layer numerical model of the aquifer, as were calculated streambank seepage rates and bank storage. The analysis indicates that streambank-leakance parameter may be considered to be a general (or lumped) parameter that accounts not only for the resistance of flow at the river–aquifer boundary, but also for the effects of partial penetration of the river and other near-stream flow phenomena not included in the theoretical development of the step-response functions. In some cases, use of the streambank-leakance parameter to account for partial penetration of a stream may affect the accuracy of streambank seepage rates and bank-storage volumes. The analysis also illustrates that aquifer recharge can affect streambank seepage rates and bank storage associated with a flood wave.

References

Barlow, P.M., Moench, A.F., 1998. Analytical solutions and computer programs for hydraulic interaction of stream–aquifer systems. US Geological Survey Open-File Report 98-415A (85pp).

Bedinger, M.S., Reed, J.E., 1964. Computing stream-induced ground-water fluctuation. US Geological Survey Professional Paper 501B, 177–180.

BSC Engineering, 1986. Preliminary site assessment relative to oil and hazardous materials. 120 Main Street, Grafton, MA, October 1986. Boston, MA, 11pp.

Bouwer, H., Rice, R.C., 1978. Delayed aquifer yield as a phenomenon of delayed air entry. Water Resources Research 14 (6), 1068–1074.

Bouwer, H., Rice, R.C., 1980. Reply to Comment on Delayed aquifer yield as a phenomenon of delayed air entry by Herman Bouwer, R.C. Rice, by Michael Price. Water Resources Research 16 (1), 254.

Cooper Jr, H.H., Rorabaugh, M.I., 1963. Ground-water movements and bank storage due to flood stages in surface streams. US Geological Survey Water-Supply Paper 1536-J, 343–366.

DeSimone, L.A., Barlow, P.M., 1995. Water-quality and hydrologic conditions at a site of ground-water contamination by volatile organic compounds, South Grafton, Massachusetts, September and October 1994. US Geological Survey Open-File Report 95-425 (19pp).

DeSimone, L.A., Barlow, P.M., 1999. Use of computer programs STLK1 and STWT1 for analysis of stream–aquifer hydraulic interaction. US Geological Survey Water-Resources Investigations Report 98-4212 (61pp).

Gillham, R.W., 1984. The capillary fringe and its effect on water-table response. Journal of Hydrology 67, 307–324.

Grubb, H.F., Zehner, H.H., 1973. Aquifer diffusivity of the Ohio River alluvial aquifer by the flood-wave response method. Journal of Research of the US Geological Survey 1 (5), 597–601.

Hall, F.R., Moench, A.F., 1972. Application of the convolution equation to stream–aquifer relationships. Water Resources Research 8 (2), 487–493.

Hansen, R.E., Steinhilber, W.L., 1977. Geohydrology of Muscatine Island, Muscatine County, Iowa. Iowa Geological Survey Water Supply Bulletin 11, 60.

HMM Associates, Inc., 1993. Site assessment and remediation support services contract, investigation of ground-water contamination. Final report, Kaltsas/Omni-Duralite site, South Grafton, MA, Site No. 2-0206, TAD No. 2021, March 1993. Concord, MA, variously paged.

HMM Associates, Inc., 1994. Site assessment and remediation support services contract, supplementary ground-water investigation. Draft report, Kaltsas/Omni-Duralite site, South Grafton, Massachusetts, Site No. 2-0206, TAD No. 2021, February 1994. Concord, MA, variously paged.

Kraijenhoff van de Leur, D.A., 1958. A study of non-steady groundwater flow with special reference to a reservoir-coefficient. De Ingenieur 70 (19), 87–94.

May, J.E., Gorman, J.G., Goodrich, R.D., Miller, V.E., 1998. Water resources data, Iowa, water year 1997, volume 1: surface water—Mississippi River Basin. US Geological Survey Water-Data Report IA-97-1 (395pp).

Moench, A.F., Barlow, P.M., 2000. Aquifer response to stream-stage and recharge variations: I. Analytical step-response functions. Journal of Hydrology 230, 192–210.

Moench, A.F., Kisiel, C.C., 1970. Application of the convolution relation to estimating recharge from an ephemeral stream. Water Resources Research 6 (4), 1087–1094.

Moench, A.F., Sauer, V.B., Jennings, M.E., 1974. Modification of routed streamflow by channel loss and base flow. Water Resources Research 10 (5), 963–968.

Narasimhan, T.N., Zhu, M., 1993. Transient flow of water to a well in an unconfined aquifer-applicability of some conceptual models. Water Resources Research 29 (1), 179–191.

Pinder, G.F., Bredehoeft, J.D., Cooper Jr, H.H., 1969. Determination of aquifer diffusivity from aquifer response to fluctuations in river stage. Water Resources Research 5 (4), 850–855.

Reynolds, R.J., 1987. Diffusivity of a glacial-outwash aquifer by the floodwave-response technique. Ground Water 25 (3), 290–299.

Schulmeyer, P.M., Barnes, K.K., Squillace, P.J., 1995. Hydrologic data from the lower Cedar River Basin, Iowa, 1989–1991. US Geological Survey Open-File Report 94-711, 116.

Schwarz, R.J., Friedland, B., 1965. Linear systems, McGraw-Hill, New York (521pp).

Serrano, S.E., Workman, S.R., 1998. Modeling transient stream/aquifer interaction with the non-linear Boussinesq equation and its analytical equation. Journal of Hydrology 206, 245–255.

Sherman, L.K., 1932. Streamflow from rainfall by the unit-graph method. Engineering News Record 108, 501–505.

Socolow, R.S., Murino Jr, D., Casey, R.G., Ramsbey, L.R., 1997. Water resources data, Massachusetts and Rhode Island, water year 1996. US Geological Survey Water-Data Report MA-RI-96-1 (367pp).

Sophocleous, M., 1985. The role of specific yield in ground-water recharge estimations: a numerical study. Ground Water 23 (1), 52–58.

Squillace, P.J., 1996. Observed and simulated movement of bank-storage water. Ground Water 34 (1), 121–134.

Squillace, P.J., Caldwell, J.P., Schulmeyer, P.M., Harvey, C.A., 1996. Movement of agricultural chemicals between surface water and ground water, Lower Cedar River Basin, Iowa. US Geological Survey Water-Supply Paper 2448 (59pp).

Venetis, C., 1968. On the impulse response of an aquifer. Bulletin of the International Association of Scientific Hydrology 13, 136–139.

Venetis, C., 1970. Finite aquifers–characteristic responses and application. Journal of Hydrology 12, 53–62.

Walker, E.H., Krejmas, B.E., 1986. Water resources of the Blackstone River Basin, Massachusetts. US Geological Survey Hydrologic Investigations Atlas 682, 2 sheets.

Whitman and Howard, Inc., 1983. Report on continuous pumping test, South Grafton, MA, October 1993. Wellesley, MA, variously paginated.

Whitman and Howard, Inc., 1990. Zone II delineation and water quality assessment at proposed well site 7-81, South Grafton, MA, February 1990. Wellesley, MA, variously paginated.

Workman, S.R., Serrano, S.E., Liberty, K., 1997. Development and application of an analytical model of stream/aquifer interaction. Journal of Hydrology 133, 233–257.