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Use of Computer Programs STLK1 and STWT1 for Analysis of Stream-Aquifer Hydraulic Interaction

By LESLIE A. DESIMONE and PAUL M. BARLOW

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BRUCE BABBITT, *Secretary*

U.S. GEOLOGICAL SURVEY
Charles G. Groat, *Director*

For additional information write to:

Chief, Massachusetts-Rhode Island District
U.S. Geological Survey
Water Resources Division
28 Lord Road, Suite 280
Marlborough, MA 01752

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CONVERSION FACTORS

Multiply	By	To Obtain
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
cubic foot per foot (ft^3/ft)	0.0929	cubic meter per meter
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
1/foot (ft^{-1})	3.281	1/meter
foot per day (ft/d)	0.3048	meter per day
inch (in.)	25.4	millimeter
mile (mi)	1.609	kilometer
square foot per day (ft^2/d)	0.09290	square meter per day

VERTICAL DATUM

Sea Level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)--a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Use of Computer Programs STLK1 and STWT1 for Analysis of Stream-Aquifer Hydraulic Interaction

By Leslie A. DeSimone *and* Paul M. Barlow

Abstract

Quantifying the hydraulic interaction of aquifers and streams is important in the analysis of stream base flow, flood-wave effects, and contaminant transport between surface- and ground-water systems. This report describes the use of two computer programs, STLK1 and STWT1, to analyze the hydraulic interaction of streams with confined, leaky, and water-table aquifers during periods of stream-stage fluctuations and uniform, areal recharge. The computer programs are based on analytical solutions to the ground-water-flow equation in stream-aquifer settings and calculate ground-water levels, seepage rates across the stream-aquifer boundary, and bank storage that result from arbitrarily varying stream stage or recharge. Analysis of idealized, hypothetical stream-aquifer systems is used to show how aquifer type, aquifer boundaries, and aquifer and streambank hydraulic properties affect aquifer response to stresses. Published data from alluvial and stratified-drift aquifers in Kentucky, Massachusetts, and Iowa are used to demonstrate application of the programs to field settings. Analytical models of these three stream-aquifer systems are developed on the basis of available hydrogeologic information. Stream-stage fluctuations and recharge are applied to the systems as hydraulic stresses. The models are calibrated by matching ground-water levels calculated with computer program STLK1 or STWT1 to measured ground-water levels.

The analytical models are used to estimate hydraulic properties of the aquifer, aquitard, and streambank; to evaluate hydrologic conditions in the aquifer; and to estimate seepage rates and bank-storage volumes resulting from flood waves and recharge. Analysis of field examples demonstrates the accuracy and limitations of the analytical solutions and programs when applied to actual ground-water systems and the potential uses of the analytical methods as alternatives to numerical modeling for quantifying stream-aquifer interactions.

INTRODUCTION

The hydraulic interaction of aquifers and streams is important in many hydrogeologic settings. Ground-water discharge supports stream base flow and riparian ecosystems during periods of little or no precipitation, and bank storage can attenuate flood waves and dampen overall flood impacts (Winter and others, 1998). The flow of water between aquifers and streams also has implications for water quality. For example, contaminants in streams may enter ground-water systems as the result of flood-wave-induced bank storage, and chemical loading of ground-water contaminants to surface waters is affected by ground-water discharge rates as well as contaminant concentrations in ground water. Stream-aquifer hydraulic interactions are particularly important in alluvial-valley aquifers consisting of sand, gravel, and other associated sediments deposited by streams (Heath, 1984, 1988; Rosenshein, 1988; Sharp, 1988). Alluvial-valley aquifers are prevalent in areas of the United States that were covered by ice sheets during the Pleistocene Epoch; in valleys of the Mississippi River drainage system and other streams that received meltwater from Pleistocene ice sheets or mountain glaciers; and in several western and southwestern alluvial basins (Heath, 1984). In these settings, permeable aquifer sediments in contact with perennial streams allow for large and rapid exchanges of water and energy (Sharp, 1988). Because of their high permeability, potential for induced infiltration, and location along rivers that have served historically as major transportation routes and industrial corridors, these aquifers are or have the potential to be important water-supply sources (Rosenshein, 1988).

Stream-aquifer interactions have been evaluated and quantified using a variety of approaches that include field methods, analytical modeling, and numerical modeling. Field methods provide information on site- or reach-specific responses of ground-water levels and seepage rates to stream-stage fluctuations, areal recharge, and base-flow recession (the discharge of stored ground water to streams). Examples of stream-aquifer field studies are provided by Lee (1977), Sophocleous and others (1988), Dumouchelle and others (1993), Yost (1995), and Dickerman and Barlow (1997). Analytical solutions have been derived from partial differential equations of ground-water flow for several idealized stream-aquifer settings, including those in which a stream is bounded by a confined, leaky, or water-table aquifer. Barlow and Moench (1998) provide a review of several of these analytical solutions. These solutions provide a means to quantify ground-water-level fluctuations, seepage rates between a stream and adjoining aquifer, and bank storage that occur in response to stream-stage fluctuations or ground-water recharge (Bedinger and Reed, 1964; Pinder and others, 1969; Moench and Kisiel, 1970; Grubb and Zehner, 1973; Moench and others, 1974; Reynolds, 1987). Such solutions also have been used for the analysis of base-flow recession (Hall, 1968; Singh, 1969; Rutledge, 1993; and Tallaksen, 1995). Numerical models are particularly useful for complex, heterogeneous, two- and three-dimensional ground-water-flow systems. Examples of the use of numerical models are provided by Pinder and Sauer (1971), Prince and others (1989), Sophocleous and Perkins (1993), Perkins and Koussis (1996), and Whiting and Pomeranets (1997). Of the three approaches discussed here, analytical solutions often are advantageous because of their simplicity. They are more general than site-specific field experiments, yet are easier to develop for a particular site than numerical models.

Several new analytical solutions and two computer programs for their application have been developed to quantify stream-aquifer hydraulic interaction for several types of confined, leaky, and water-table aquifers (Barlow and Moench, 1998). The analytical solutions are used in combination with convolution relations (a method of superposition) to allow for analysis of continuously changing stream-stage and recharge conditions. The solutions can be applied to aquifers that are semi-infinite or of finite width and for which semipervious streambank material is present at the stream-aquifer interface. These solutions differ from previously developed analytical approaches primarily in the wide range of aquifer types to which they can be applied and by the availability of readily accessible computer programs for their use. By use of these solutions and programs, ground-water levels, seepage between an aquifer and stream, and bank storage resulting from arbitrarily varying stream stage or recharge can be quantified. In addition, aquifer hydraulic parameters can be estimated by calibration to measured ground-water-level fluctuations. Derivation of the analytical solutions and documentation of the computer programs are provided by Barlow and Moench (1998).

This report describes the use of the analytical solutions and computer programs developed by Barlow and Moench (1998) to illustrate the response of idealized aquifers to generalized stream-stage and recharge fluctuations and to demonstrate applications of the computer programs to three field settings. Application of the solutions and programs to idealized, hypothetical stream-aquifer systems illustrates the effects of aquifer type, aquifer boundaries, and aquifer and streambank hydraulic properties on stream-aquifer hydraulic interaction. Analysis of field examples from Kentucky, Massachusetts, and Iowa demonstrates the response of real stream-aquifer systems to stream-stage and recharge fluctuations, the accuracy and limitations of the analytical solutions and programs when applied to real systems, and potential uses of the solutions and programs as alternatives to numerical modeling for quantifying stream-aquifer interactions.

The authors thank Peter Schulmeyer, U.S. Geological Survey (USGS), Iowa City, Iowa; Paul Squillace, USGS, Rapid City, South Dakota; and J. Jeffrey Starn, USGS, East Hartford, Connecticut, for their assistance in providing data and other information for use in the analysis of field examples. We also thank Allen Moench, USGS, Menlo Park, California, for his assistance in the interpretation of field-example results.

COMPUTER PROGRAMS STLK1 AND STWT1

This section provides a brief background to the computer programs STLK1 and STWT1. The underlying theory of the programs and a description of their use are provided by Barlow and Moench (1998). Application of the computer programs requires specification of aquifer type and lateral extent; hydraulic properties of the aquifer, aquitard (if present), and semipervious streambank material (if present); location of an observation well or piezometer; and stream-stage and recharge stresses. These specifications describe how the stream-aquifer system is conceptualized and modeled. Simplifying assumptions about the geometry of and flow within the stream-aquifer system are needed for application of the programs. The programs require discretization of stream-stage and recharge hydrographs that are used as input stresses to the simulated stream-aquifer systems. Sample input and output files for STLK1 and STWT1 that were used in two of the three field examples are provided in the appendix.

Conceptualization of Stream-Aquifer Interaction

Water moves between hydraulically connected aquifers and adjacent streams in response to head gradients across the stream-aquifer boundary. Where ground-water levels are greater than the elevation of the stream stage, ground water discharges to the stream (gaining stream reach). Where the elevation of the stream stage is greater than ground-water levels in the immediate vicinity of the stream, seepage occurs from the stream to the aquifer (losing stream reach). The rate at which water moves between a stream and aquifer depends on the type, lateral extent, and hydraulic properties of the adjoining aquifer; the depth of penetration of the stream into the aquifer; the hydraulic properties of the streambanks and streambed; and the hydraulic gradient between the stream and aquifer.

Computer program STLK1 can be applied to confined or leaky aquifers and computer program STWT1 can be applied to water-table (unconfined) aquifers. A confined aquifer (fig. 1*A*) is one that has an overlying layer of geologic material (a confining layer) that prevents ground-water flow to or from the underlying aquifer. A leaky aquifer has an overlying layer of geologic material (an aquitard) with a much lower hydraulic conductivity than that of the underlying aquifer; the aquitard restricts but does not prevent ground-water flow (leakage) to or from the underlying aquifer. Three types of leaky aquifers can be simulated by use of program STLK1: (1) a source bed with a constant head overlying the aquitard (fig. 1*B*); (2) an impermeable layer overlying the aquitard (fig. 1*C*); and (3) a free surface or water-table (unconfined) condition within the aquitard (fig. 1*D*). Ground-water flow is assumed to be horizontal (one dimensional) in a direction perpendicular to the stream for each of the confined and leaky aquifer types. For the leaky aquifers, flow is assumed to be strictly vertical through the overlying confining layer or aquitard; thus the hydraulic conductivity of the aquitard must be small compared with hydraulic conductivity of the aquifer. A water-table aquifer does not have an overlying confining layer or aquitard and a free surface forms the upper boundary to the aquifer (fig. 1*E*). Ground-water flow in a water-table aquifer is assumed to be two dimensional (that is, horizontal and vertical) in a plane perpendicular to the stream.

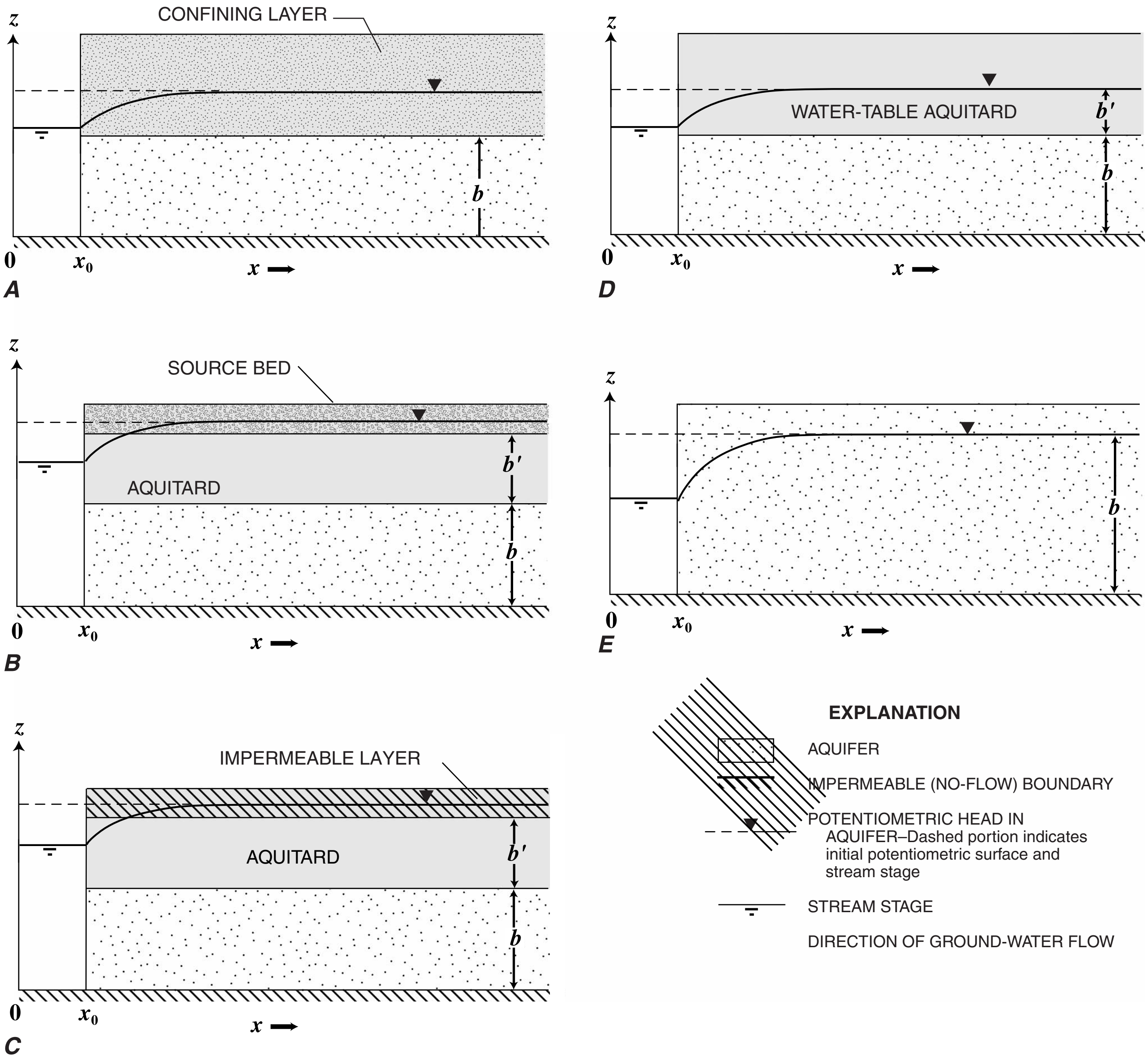


Figure 1. Types of aquifers to which computer programs STLK1 and STWT1 may be applied. (*A*) Confined; (*B*) Leaky, with a constant head overlying the aquitard; (*C*) Leaky, with an impermeable layer overlying the aquitard; (*D*) Leaky, overlain by a water-table aquitard; (*E*) Water table (unconfined). (*b*, thickness or saturated thickness of the aquifer; *b*′, thickness or saturated thickness of the aquitard.) Note: A semipervious streambank of width *d* can be simulated for each aquifer type, and the aquifer can have a finite width (x_L). (Modified from Barlow and Moench, 1998.)

Aquifers simulated by use of programs STLK1 and STWT1 may be either laterally extensive (semi-infinite) or finite in width and may have a thin layer of low-permeability material at the interface between the aquifer and stream. All simulated aquifers are assumed to be of uniform thickness and underlain by an impermeable boundary (fig. 1). In addition, the stream is assumed to fully penetrate all aquifer types, such that all seepage between the stream and aquifer occurs in a horizontal direction through the streambank. The assumption of a fully penetrating stream is discussed further in the "Assumptions" section below.

Many stream-aquifer systems lie within alluvial valleys that consist of layered and vertically stratified materials. The type of aquifer within a particular valley depends upon the history and sequence of deposition of these materials, the types of materials that were deposited, and the positions of stream stage and ground-water levels relative to the various depositional layers. Confined and leaky aquifers can occur where layers of fine-grained sediments, such as silty fine sand or silt and clay, have been deposited over coarse-grained sand and gravel. These fining-upward sequences are common along the valleys of the Mississippi, Missouri, and Ohio Rivers and other rivers and streams of the central United States where coarse-grained sediment deposited by meltwater from Pleistocene ice sheets subsequently was overlain by fine-grained alluvium deposited during and after the last phases of the retreating ice sheet (Walker, 1957; Gallaher and Price, 1966; Sharp, 1977; Heath, 1984; Rosenshein, 1988; Sharp, 1988). Confining conditions result where ground-water levels lie above the top of the coarse-grained (aquifer) materials and the fine-grained alluvium is of such low permeability that it cannot effectively transmit substantial quantities of water to or from the underlying aquifer (fig. 1*A*). These conditions commonly occur in the distant flood plains of rivers beyond the areas of present-day channel migration and deposition (Sharp, 1988).

Leaky conditions prevail where the fine-grained alluvium is permeable enough that substantial quantities of water are transmitted to and from the underlying aquifer. The several types of leaky aquifers that are modeled by STLK1 reflect various idealized hydrogeologic conditions. For example, a leaky aquifer overlain by an aquitard beneath an impermeable layer (fig. 1*C*) may correspond to cases where the relatively permeable fine-grained alluvium grades upward into glaciolacustrine or flood-plain silt and clay; the silt and clay layer may form a nearly impermeable cap over the aquitard and cause the leaky system to be under pressure. Cases where a water table is within the fine-grained alluvium correspond to a leaky aquifer overlain by a water-table aquitard (fig. 1*D*). A constant-head source bed overlying the aquitard (fig. 1*B*) may represent conditions where the aquitard underlies a ponded or flooded area near the stream. Finally, water-table conditions (fig. 1*E*) will prevail if the stream stage and ground-water levels are below the bottom of the fine-grained alluvium or if no fine-grained alluvium is present above the coarse-grained aquifer material.

Various hydraulic properties of the aquifer and semipervious streambank affect ground-water levels and seepage rates. For confined aquifers, the relevant properties are horizontal hydraulic conductivity (K_x), thickness (b), and specific storage (S_s). For leaky aquifers, the hydraulic properties of the overlying aquitard also must be considered. These properties are vertical hydraulic conductivity (K'), specific storage (S_s'), thickness (b'), and, for water-table aquitards, specific yield (S_y') and saturated thickness (b'). For water-table aquifers, the relevant properties are vertical (K_z) and horizontal (K_x) hydraulic conductivity (or K_D, the ratio of vertical to horizontal hydraulic conductivity), specific storage (S_s), and specific yield (S_y). The transmissivity (T) and storativity (or storage coefficient) (S) of an aquifer commonly are used in place of horizontal hydraulic conductivity and specific storage. Transmissivity is equal to the product of the horizontal hydraulic conductivity and saturated thickness of the aquifer ($T = K_x b$); storativity is equal to the product of the specific storage and saturated thickness of the aquifer ($S = S_s b$). Aquifer diffusivity (α), or the ratio of transmissivity to storativity (or specific yield for water-table aquifers; T/S or T/S_y), also is commonly used. The lateral width of an aquifer also may affect its response to stress at the stream-aquifer boundary. In this report, aquifer width (x_L) is defined as the distance from the center of the stream channel to an impermeable aquifer boundary. The distance from the center of the stream to the stream-aquifer boundary (x_o; fig. 1) also must be defined for computational purposes, but the actual value that is defined for x_o has no effect on the calculated aquifer response.

If semipervious streambank material impedes seepage between the stream and aquifer, it is necessary to consider the hydraulic properties of the streambank material. These properties are accounted for in programs STLK1 and STWT1 by a streambank leakance term (a)

$$a = \frac{K_x d}{K_s} , \tag{1}$$

where

d is the width of the semipervious streambank material and

K_s is the hydraulic conductivity of the streambank material in the direction perpendicular to streamflow.

The ratio K_s/d can be considered a single fluid-transfer parameter, because it is difficult to evaluate K_s and d separately. They generally are lumped together in the calibration of hydrogeologic models. As an alternative to this formal treatment, the leakance term may be interpreted loosely as accounting for constricted flow at the stream-aquifer interface, because the stream may not penetrate the full saturated thickness of the aquifer.

A hydraulic gradient between the aquifer and stream, which results in water movement across the stream-aquifer boundary, is caused by hydraulic stresses such as flood waves, ground-water recharge, ground-water recession, evapotranspiration, and leakage to an underlying aquifer. During a flood wave, stream stage increases relative to water levels in the aquifer, such that the ambient hydraulic gradient toward the stream (for a gaining stream) is reversed. Seepage occurs from the stream to the aquifer, and ground-water heads near the stream increase. Seepage that enters the aquifer adjacent to the stream is referred to as bank storage, and the volume of bank storage held by the aquifer increases until shortly after the flood peak. After the flood wave passes, stream stage falls, water in bank storage is discharged back to the stream, and ground-water heads eventually may return to pre-flood-wave conditions. A recharge event may be viewed as the opposite of a flood wave in terms of relative water levels between the aquifer and stream. It is assumed that ground-water heads everywhere in the aquifer increase relative to stream stage during the recharge event, and ground-water discharge (seepage) to the stream increases over ambient conditions. Ground-water levels and discharge rates eventually return to pre-recharge levels after the recharge ends. The response of the aquifer to evapotranspiration can be simulated in the computer programs by specifying evapotranspiration as a negative recharge event. Ground-water levels everywhere in the aquifer are assumed to decrease relative to stream stage in response to evapotranspiration.

Assumptions

Several simplifying assumptions were made in the derivation of the analytical solutions that are incorporated into the STLK1 and STWT1 computer programs. Some of these assumptions were briefly described previously. Because of their importance, the major simplifying assumptions that underlie the computer programs are summarized here for reference (Barlow and Moench, 1998):

1. Aquifers and aquitards are homogeneous and of uniform thickness.
2. Confined and leaky aquifers are isotropic and flow is strictly horizontal in a direction perpendicular to the stream. For leaky aquifers, the hydraulic conductivity of the aquitard is small compared to the hydraulic conductivity of the underlying aquifer and flow through the aquitard is strictly vertical.
3. Flow in water-table aquifers may have both horizontal and vertical components and the horizontal hydraulic conductivity may differ from the vertical hydraulic conductivity.
4. The lower boundary of each aquifer is horizontal and impermeable.
5. Hydraulic properties of aquifers and aquitards do not change with time.

6. For water-table conditions, water is released (or taken up) instantaneously in a vertical direction from (or into) the zone above the water table in response to a decline (or rise) in the elevation of the water table.
7. Changes in the saturated thickness of water-table aquifers or water-table aquitards are small compared with the initial saturated thickness.
8. The bounding stream is straight and fully penetrates the aquifer.
9. Semipervious streambank material is homogeneous, of constant thickness, and has negligible capacity to store water.
10. Initially, the water level in the stream is at the same elevation as the water level in the aquifers and aquitards (if present). At time $t = 0$, the water level in the stream is instantaneously raised (or lowered) to a new position.

Note that the analytical solutions were derived for the condition of an instantaneous step increase (or decrease) in the water level of the stream relative to the water level in the adjacent aquifer. These solutions are implemented in the computer programs for series of time-varying stream-stage and recharge inputs by use of convolution relations, which are a form of mathematical superposition. Use of the convolution relations is valid because the governing partial differential equation of ground-water flow and all of the boundary and initial conditions used in the derivation of the analytical solutions are linear. However, for linearity to hold, changes in ground-water levels must be relatively small in comparison to the saturated thickness of the aquifer.

The assumption that the stream fully penetrates the aquifer (8, above) was discussed by Hantush (1965) and, in reference to a partially penetrating pumping well, by Hantush (1964) and Neuman (1974). Hantush (1965) stated that the effect of a partially penetrating streambed on ground-water levels can be neglected at a distance $1.5b$ away from the streambank. For a partially penetrating pumping well, Hantush (1964) found that the average drawdown in an unconfined aquifer is the same as if the pumping well fully penetrated the aquifer at distances (r) greater than $1.5b\sqrt{K_z/K_x}$ from the pumping well. Similarly, Neuman (1974) found that the effect of a partially penetrating pumping well in an unconfined aquifer disappears completely at distances greater than $b/\sqrt{K_z/K_x}$ from the pumping well when time from the start of pumping exceeds $(10S_y r^2)/T$. These references indicate, and Hantush (1965) affirmed, that partial penetration of a streambed will have a small effect on ground-water-level responses to stream-stage fluctuations or recharge beyond moderate distances from the streambank. Sharp (1977) also has evaluated several of the assumptions listed above in reference to alluvial aquifers of the central United States. He argued that some of the assumptions, notably those of aquifer homogeneity and a fully penetrating stream, cannot be supported by observed hydrogeologic conditions. Nevertheless, although the assumptions require simplification of the complexities of real stream-aquifer systems, analyses based on the analytical solutions do provide useful results for a number of field applications, as is demonstrated later in this report.

Discretization of Stream-Stage and Recharge Stresses

Stream-stage and recharge stresses must be specified in the input data files for programs STLK1 and STWT1. Such stresses are illustrated by hydrographs, which represent continuously changing stream-stage (fig. 2*A*) and recharge (fig. 2*B*) conditions. Stream-stage and recharge hydrographs must be approximated for the computer programs by a time series of discrete changes in each hydrograph (Δh) during each time step Δt. The length of each time step must be uniform during each simulation and its value specified by the user. Time-step length can be determined by the density of available data, but, as with all discretization schemes, the accuracy of the solutions is improved by use of smaller time steps (Barlow and Moench, 1998).

Uniform areal recharge and evapotranspiration can be specified in the programs only for water-table conditions. Recharge is specified in the programs as a change in ground-water level of the aquifer relative to the stream stage, as shown on the right-hand axis of figure 2*B*. This increase in ground-water level resulting from recharge can be measured directly, or can be estimated by dividing the amount of recharge (R) by the specific yield of the water-table aquifer or aquitard (S_y or S'_y). The latter value, (R/S_y) or (R/S'_y), is equal to the ground-water-level increase 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 because of aquifer or aquitard heterogeneity. The change in ground-water level

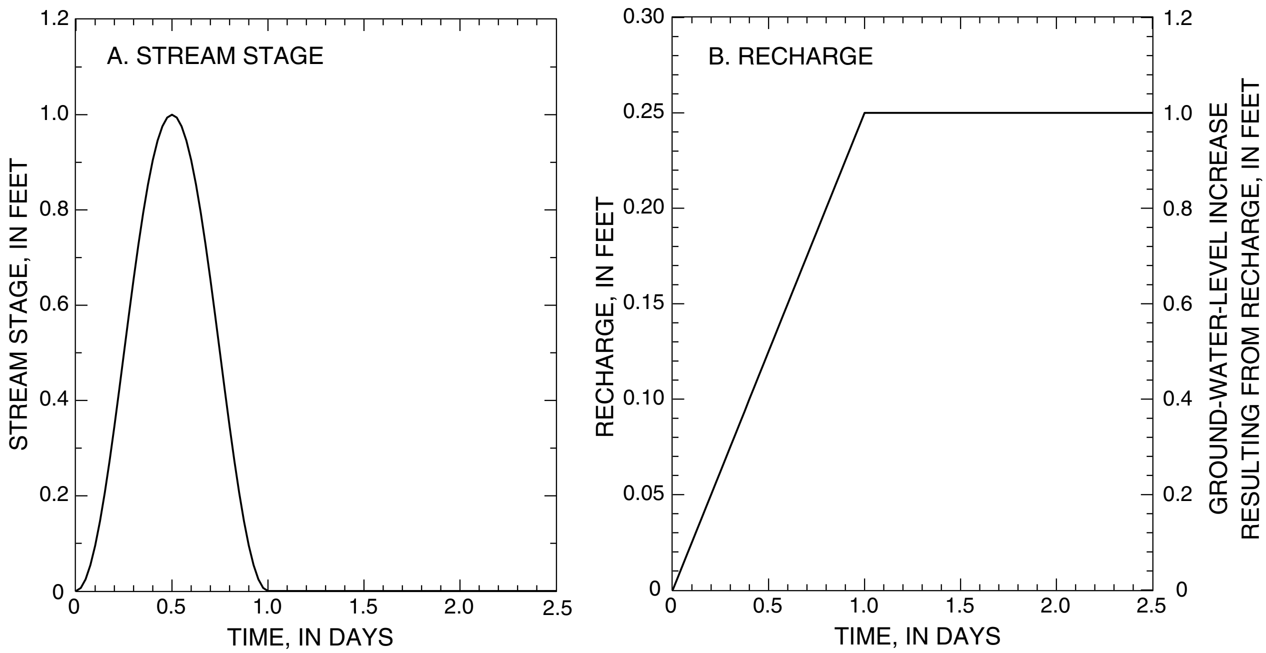


Figure 2. Stream-stage fluctuation and recharge used for simulations of hypothetical aquifers. (*A*) Sinusoidal, 1-day fluctuation in stream stage. (*B*) Linear, 1-day recharge event (specific yield equal to 0.25).

resulting from recharge is the specified stress to the system. Once recharge ends, the specified ground-water level provided as input to the programs remains at a constant value (equal to 1.0 ft in fig. 2*B*); this formulation of the recharge hydrograph is required for analytical and computational purposes (Barlow and Moench, 1998; actual ground-water levels in the aquifer would decline after recharge ends). Evapotranspiration from an aquifer also would be specified by changes (decreases rather than increases) in ground-water levels in the aquifer.

ANALYSIS OF STREAM-AQUIFER HYDRAULIC INTERACTION IN IDEALIZED SYSTEMS

The response of an aquifer to stream-stage or recharge stresses depends on such variables as the type, geometry, boundaries, and hydraulic properties of the aquifer and aquitard (if present) and the presence and hydraulic properties of semipervious streambank material. In this section, an analysis is made of idealized, hypothetical confined, leaky, and water-table aquifers to illustrate how stream-aquifer hydraulic interaction is affected by these variables. Stream-stage and recharge stresses are simulated in several independent model runs using STLK1 and STWT1. In the analyses, a flood wave is represented by a 1-foot, 1-day sinusoidal fluctuation in stream stage (fig. 2*A*). Recharge is represented as a 0.25-foot, linear recharge event, which corresponds to a 1-foot increase in ground-water levels (with a specific yield of 0.25), and which also lasts 1 day (fig. 2*B*). Physical and hydraulic properties of the aquifer, aquitard, and semipervious streambank (where present) used in the hypothetical simulations, unless otherwise specified, are given in table 1 for the idealized confined, leaky, and water-table aquifers.

The various aquifer types indicated in table 1 respond differently to stream-stage fluctuations because of differences in their storage capacities and the boundary condition at the top of the aquifer or aquifer-aquitard system. In general, ground-water levels rise higher and more rapidly in the simulated confined aquifer and leaky aquifer overlain by an impermeable layer than in the other types of leaky aquifers or the water-table aquifer (fig. 3*A*). However, seepage rates and bank storage for the confined aquifer and leaky aquifer overlain by an impermeable layer are much lower than those for the other leaky aquifer types or the water-table aquifer (figs. 3*B* and *C*). These differences result because of the limited storage

Table 1. Physical and hydraulic properties of idealized stream-aquifer systems and other data used in simulations of hypothetical aquifers

[Physical and hydraulic properties as defined in the text and(or) figure 1. ft, foot; ft/d, foot per day; ft^{-1}, per foot; ∞, infinity; --, not applicable]

Physical and hydraulic property	Aquifer type				
	Confined	Leaky, with constant-head over the aquitard	Leaky, with impermeable layer over the aquitard	Leaky, with water-table aquitard	Water table
Aquifer Properties					
K_x (ft/d)	200	200	200	200	200
K_z (ft/d)	--	--	--	--	20
K_D (dimensionless)	--	--	--	--	0.1
b (ft)	25	25	25	25	25
S_s (ft^{-1})	$1x10^{-5}$	$1x10^{-5}$	$1x10^{-5}$	$1x10^{-5}$	$1x10^{-5}$
S_y (dimensionless)	--	--	--	--	$2.5x10^{-1}$
x_L (ft)	∞ or 2,000	∞ or 2,000	∞ or 2,000	∞ or 2,000	∞ or 500
Aquitard Properties					
K′ (ft/d)	--	2	2	2	--
b' (ft)	--	10	10	10	--
S_s' (ft^{-1})	--	$1x10^{-4}$	$1x10^{-4}$	$1x10^{-4}$	--
S_y' (dimensionless)	--	--	--	$2.5x10^{-1}$	--
Properties of the Stream and Semipervious Streambank Material					
x_o (ft)	25	25	25	25	25
K_s (ft/d)	--	--	--	--	--
d (ft)	--	--	--	--	--
a ($K_x d/K_s$, ft)	--	--	--	--	--
Other Data					
Δt (days)	0.025	0.025	0.025	0.025	0.025
Distance from stream-channel center to well (x , ft)	125	125	125	125	125
h_o (ft)	0	0	0	0	0

capacity of the confined aquifer. Storage is available only from the compressibility of the aquifer matrix and pore water (represented by the specific storage, S_s, of the aquifer). A small amount of additional storage capacity is provided by the aquitard (S_s') for the leaky aquifer overlain by an impermeable layer, such that seepage rates and bank storage are slightly higher than in a confined aquifer, but the impermeable layer overlying the aquitard prevents hydraulic connection with any overlying system. For the other leaky aquifer types, the constant-head source bed overlying the aquitard (fig. 1*B*) and movement of the water table within the aquitard (accounted for by S_y' ; fig. 1*D*) provide additional, relatively large storage capacities as compared to that provided by aquifer-matrix and pore-water compressibility alone. Thus, for these two leaky aquifer types, ground-water-level changes are buffered by the aquitard, and more water can move into and out of the aquifer from the stream than for the cases of the confined aquifer or leaky aquifer overlain by an impermeable layer. The constant-head boundary represents an infinite source (or sink) of water for the underlying aquifer. Consequently, seepage to the aquifer is permanently lost from the aquifer-aquitard system at the constant-head boundary, and there is no drainage of bank storage back to the stream, as shown in figure 3*C*. Seepage rates and bank storage are greatest for the water-table aquifer (figs. 3*B* and *C*), in which the unsaturated materials above the water table provide a large storage capacity, and the more permeable aquifer sediments present less resistance to water flow than the sediments of the simulated aquitard in the leaky aquifer.

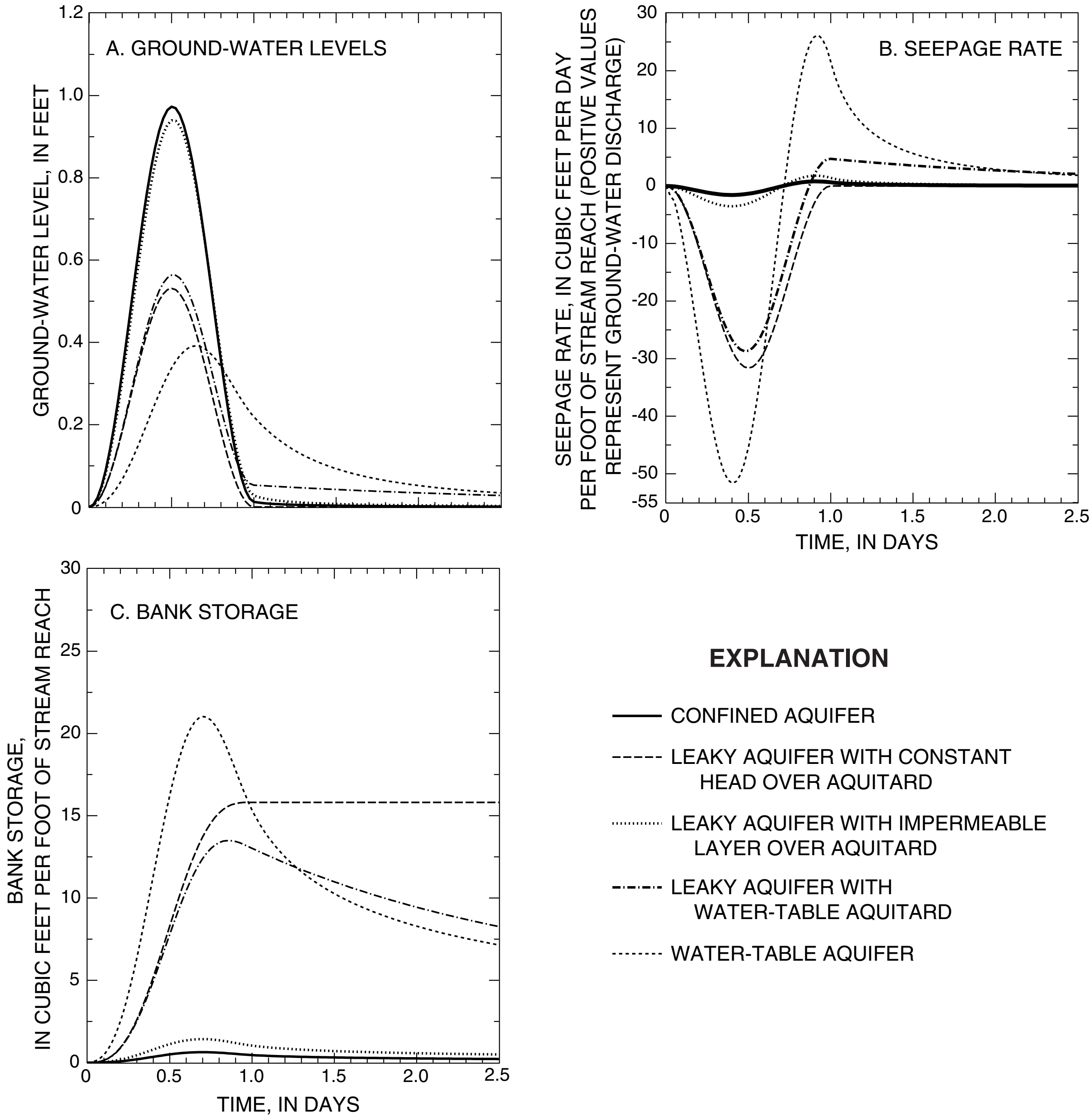


Figure 3. Effect of aquifer type on the response of hypothetical semi-infinite aquifers to a sinusoidal stream-stage fluctuation. Hydraulic properties of aquifers shown in table 1. (*A*) Ground-water levels, 100 feet from streambank. (*B*) Seepage rate. (*C*) Bank storage.

The response of ground-water levels to a stream-stage fluctuation is attenuated with distance from the streambank. At greater distances from the streambank, ground-water levels rise less rapidly, peak at lower levels, and decline more gradually for all aquifer types. Ground-water levels in the simulated semi-infinite and finite-width confined aquifers (fig. 4) are affected at relatively large distances from the stream compared to the other aquifer types (see discussion below) because the limited storage capacity of the confined aquifer allows for less attenuation of the flood wave with distance. The peak increase in ground-water level in the semi-infinite confined aquifer 100 ft from the stream is 0.97 ft, which nearly equals the 1-foot increase in stream stage (fig. 4*A*), and ground-water levels as far as 10,000 ft from the streambank were affected by the increase in head at the stream-aquifer boundary (peak at 0.09 ft; not shown). In the finite-width confined aquifer, there was little attenuation of ground-water levels with distance and there was a rapid return to initial conditions for the aquifer properties simulated (table 1; fig. 4*B*); thus, simulation of the lateral boundary 2,000 ft from the streambank substantially reduced the capacity of the aquifer to store water and energy associated with the flood wave. Consequently, seepage rates to the aquifer are lower for simulated finite-width aquifers than for a semi-infinite aquifer (fig. 5*A*) and bank storage drains more rapidly in finite-width aquifers than in a semi-infinite aquifer (fig. 5*B*). These effects also were demonstrated by Cooper and Rorabaugh (1963) and Whiting and Pomeranets (1997). Cooper and Rorabaugh (1963) demonstrated that the shapes of the response curves for seepage and bank storage are functions of the period of the stage oscillation and of the diffusivity and width of the aquifer.

Ground-water levels in water-table aquifers and aquifers overlain by a water-table aquitard are attenuated at shorter distances from the streambank than are water levels in confined aquifers and leaky aquifers without a water-table boundary. The greater attenuation in the former types of aquifers results from the large storage capacity provided by saturation of the pores as the water table rises. Also, ground-water levels are attenuated at shorter distances from the streambank in the water-table aquifer than in the leaky aquifer overlain by a water-table aquitard (fig. 6). Ground-water levels rise less than 0.05 ft in the water-table aquifer 500 ft from the streambank in response to the 1-foot flood wave, compared to a peak increase of 0.34 ft for the leaky aquifer overlain by a water-table aquitard. The storage capacities of the simulated water-table aquifer and leaky aquifer with water-table aquitard were large enough that calculated ground-water levels at nearly all distances from the streambank were essentially the same for conditions of either a semi-infinite aquifer or 2,000-foot finite-width aquifer (fig. 6). As the distance from the stream to the lateral boundary (aquifer width) decreases, the finite-width solutions for the water-table-bounded aquifer types deviate from the semi-infinite solutions similar to that for confined aquifers (fig. 4).

Hydraulic properties of the aquifer and aquitard also affect the response of the stream-aquifer system. Calculated ground-water levels, seepage rates, and bank storage for several values of aquifer diffusivity for a confined, semi-infinite aquifer are shown in figures 7 and 8. Ground-water levels at a given distance from the streambank peak at higher values with increasing values of aquifer diffusivity, α. Aquifer diffusivity increases if horizontal hydraulic conductivity (K_x) increases or specific storage (S_s) decreases. Because calculated ground-water levels depend on the diffusivity of the aquifer, simultaneous order-of-magnitude increases (or decreases) in K_x and S_s have no effect on calculated ground-water levels, because they offset each other. For example, calculated ground-water levels are the same for conditions in which K_x equals 30 ft/d and S_s equals $1x10^{-5}$ ft^{-1} as those for conditions in which K_x equals 300 ft/d and S_s equals $1x10^{-4}$ ft^{-1} (α equals $3x10^6$ ft^2/d, dash-dotted line, fig. 7). However, seepage and bank storage for these two simulations are not identical. For a fixed value of K_x, seepage rates and bank storage increase with increasing S_s, because higher values of S_s correspond to greater volumes of water released from

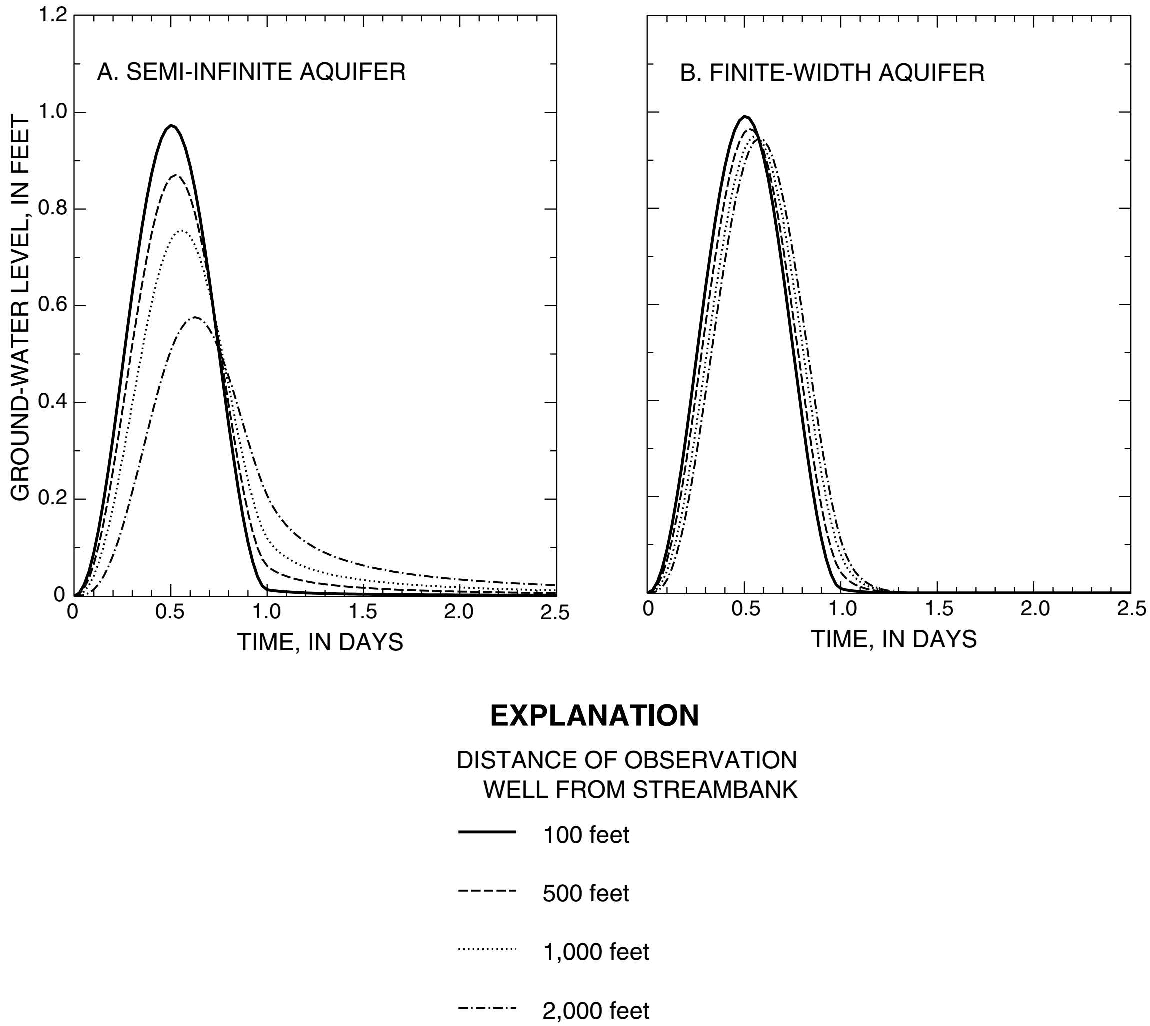


Figure 4. Effect of observation-well distance from the streambank and lateral extent of the aquifer on the response of ground-water levels in hypothetical confined aquifers to a sinusoidal stream-stage fluctuation. Hydraulic properties of the aquifers shown in table 1. (*A*) Semi-infinite aquifer. (*B*) Finite-width aquifer with lateral boundary 2,000 feet from the stream.

or taken into storage per unit change in ground-water head. Also, for any fixed value of S_s, seepage rates and bank storage increase with increasing K_x, because for a confined aquifer seepage and bank storage are a function of the product of hydraulic conductivity and aquifer diffusivity. A simultaneous order-of-magnitude increase in the value of K_x and order-of-magnitude decrease in S_s, which result in a 100-fold increase in aquifer diffusivity (solid line, figs. 8*A* and *B*), yield identical seepage rates and bank-storage volumes for a semi-infinite confined aquifer; this results from the specific solution for seepage for the confined, semi-infinite case (Hall and Moench, 1972, eqs. 7 and 10) and would not necessarily be the case for the other aquifer types.

In a water-table aquifer, vertical hydraulic conductivity (K_z) and specific yield (S_y) also affect aquifer response. As the ratio K_z:K_x (K_D) is decreased, resistance to flow in the vertical direction causes the response of a water-table aquifer to approach that of a confined aquifer. For example, as shown in figure 9, when K_D is decreased from 0.1 to 0.01 (as S_y remains constant at 2.5 x 10^{-1}), calculated ground-water levels rise more rapidly and to a higher peak value (fig. 9*A*), seepage rates between the stream and aquifer decrease (fig. 9*B*), and bank storage is reduced (fig. 9*C*). Similar effects result from lower values of S_y (with K_D remaining constant), which represents the volume of water that is released from (or taken up by) storage from movement of the water table (fig. 9).

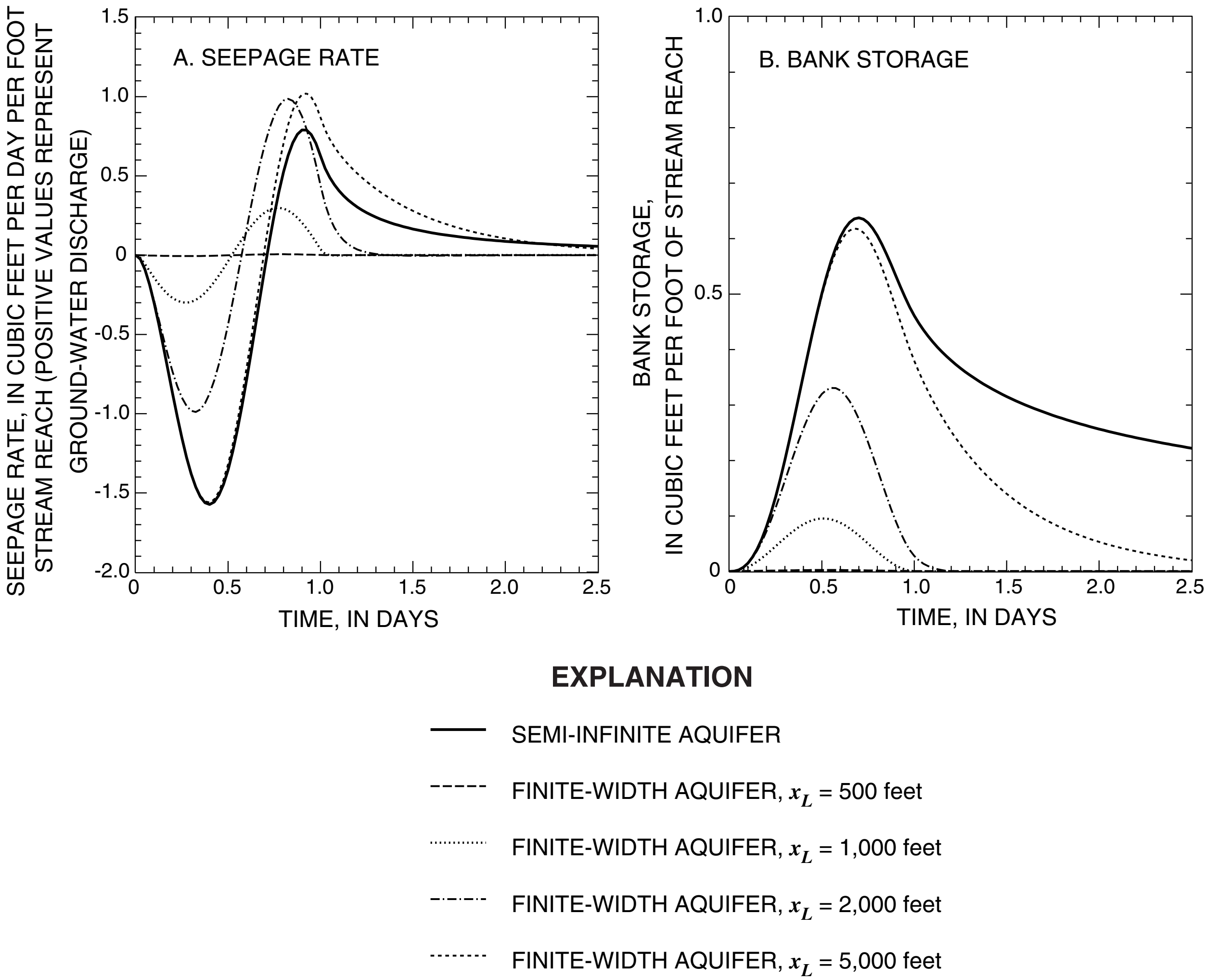


Figure 5. Effect of the lateral extent of the aquifer on seepage and bank storage in hypothetical confined aquifers in response to a sinusoidal stream-stage fluctuation. Hydraulic properties of the aquifers shown in table 1 unless otherwise specified; x_L, aquifer width. (*A*) Seepage rate. (*B*) Bank storage.

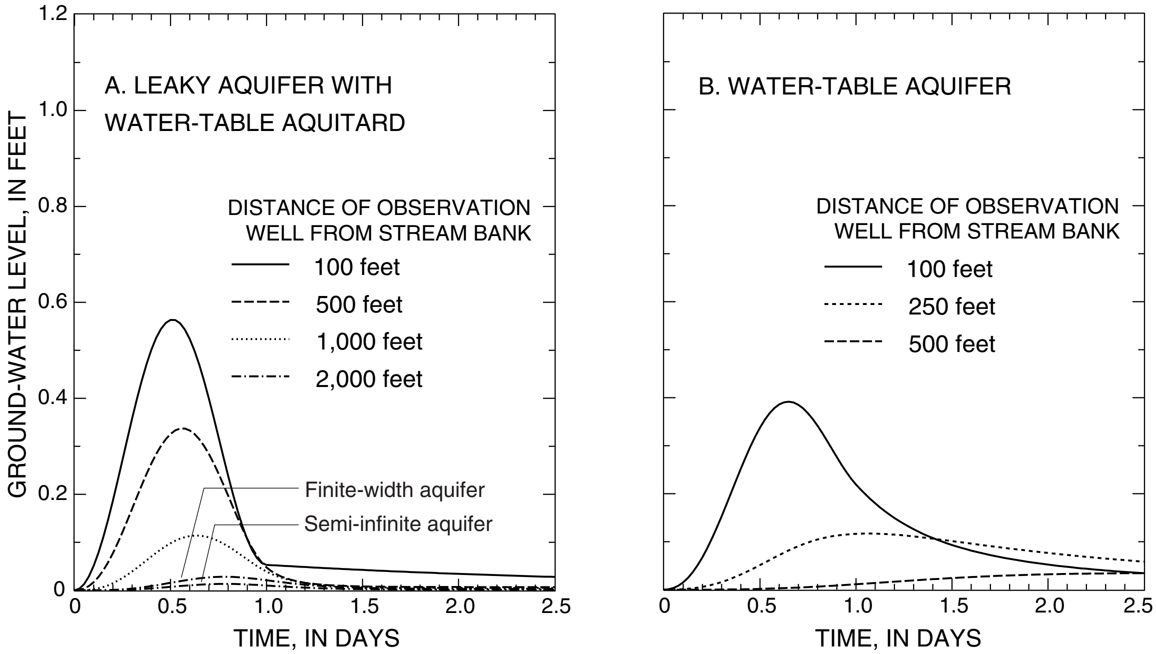


Figure 6. Effect of observation-well distance from the streambank and the lateral extent of the aquifer on the response of ground-water levels in a hypothetical leaky aquifer with a water-table aquitard and a hypothetical water-table aquifer to a sinusoidal stream-stage fluctuation. Hydraulic properties of the aquifers and aquitard shown in table 1; curves represent both semi-infinite and finite-width (2,000 feet) aquifers unless otherwise specified. (*A*) Leaky aquifer with a water-table aquitard. (*B*) Water-table aquifer.

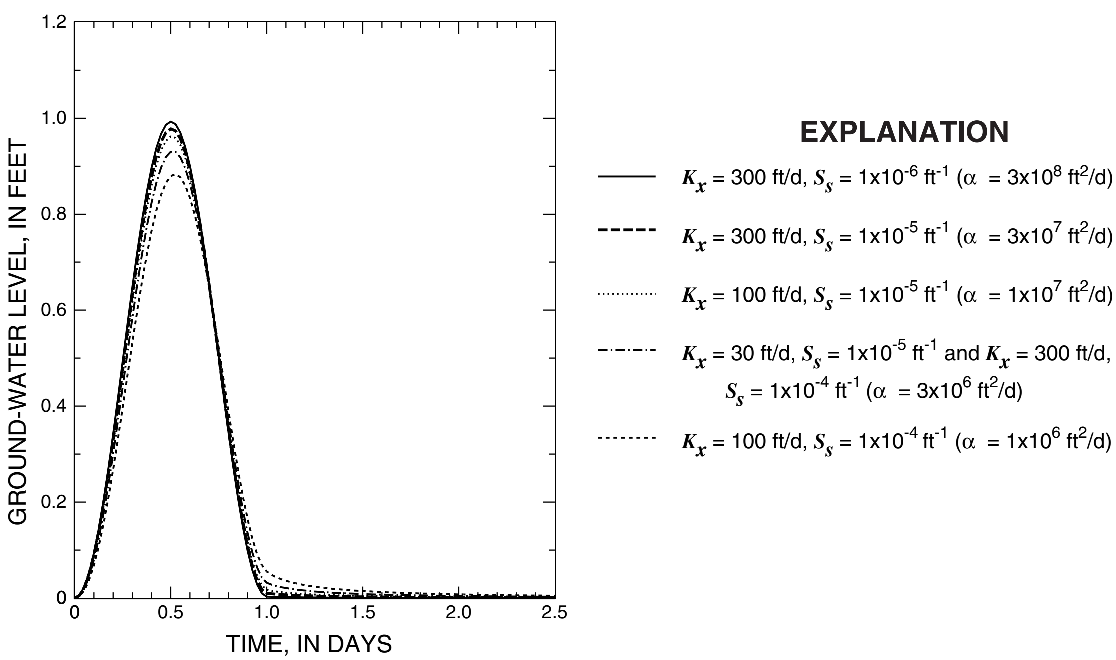


Figure 7. Effect of aquifer hydraulic properties on ground-water levels at 100 feet from the streambank in a hypothetical, semi-infinite confined aquifer in response to a sinusoidal stream-stage fluctuation. Hydraulic properties shown in table 1, unless otherwise specified. (α, aquifer diffusivity; K_x, horizontal hydraulic conductivity; S_s, specific storage; ft/d, foot per day; ft^{-1}, per foot; ft^2/d, square foot per day.)

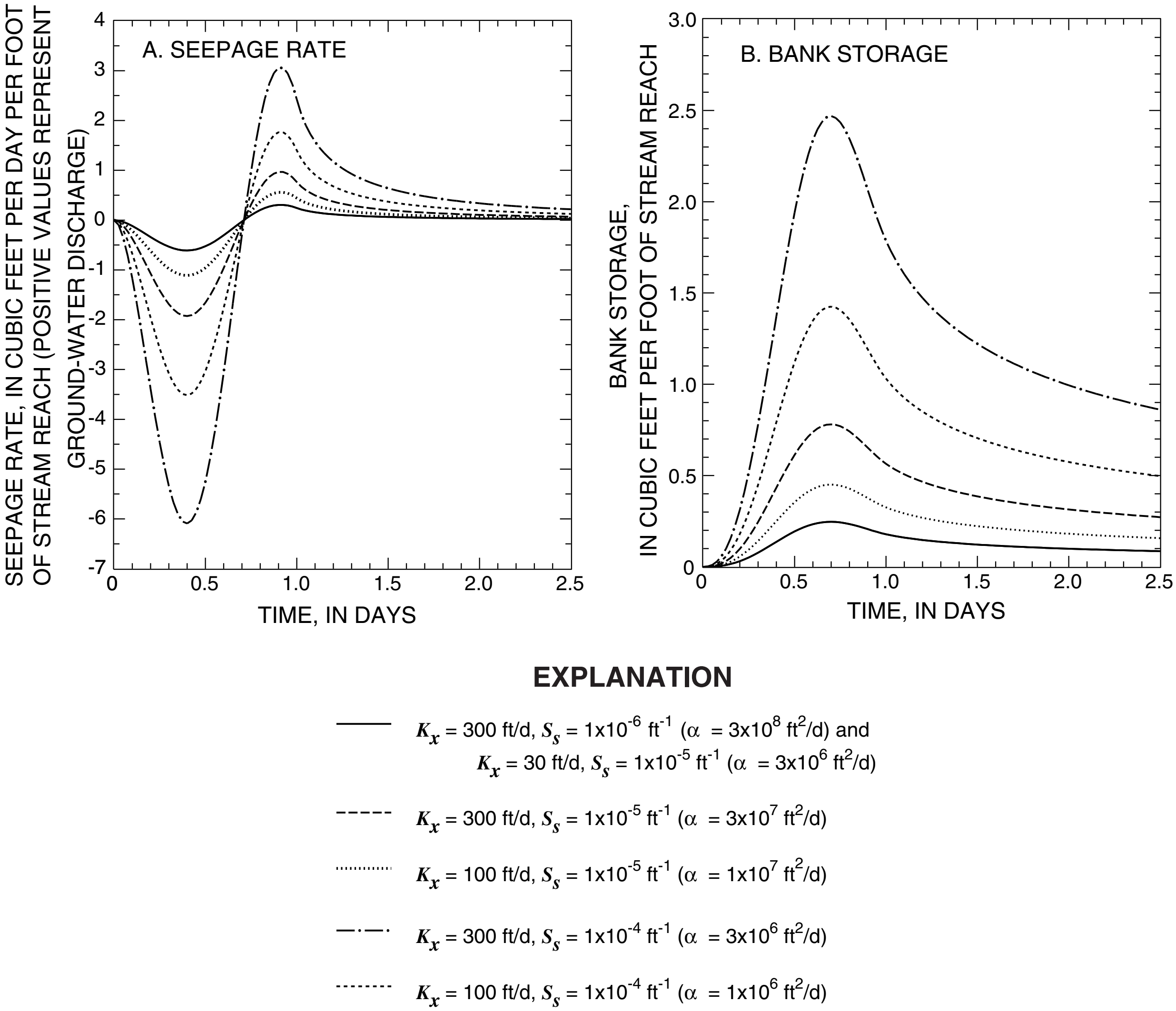


Figure 8. Effect of aquifer hydraulic properties on seepage rate and bank storage in a hypothetical, semi-infinite confined aquifer in response to a sinusoidal stream-stage fluctuation. Hydraulic properties shown in table 1, unless otherwise specified. (*A*) Seepage rate. (*B*) Bank storage. (α, aquifer diffusivity; K_x, horizontal hydraulic activity; S_s, specific storage; ft/d, foot per day; ft^{-1}, per foot; ft^2/d, square foot per day.)

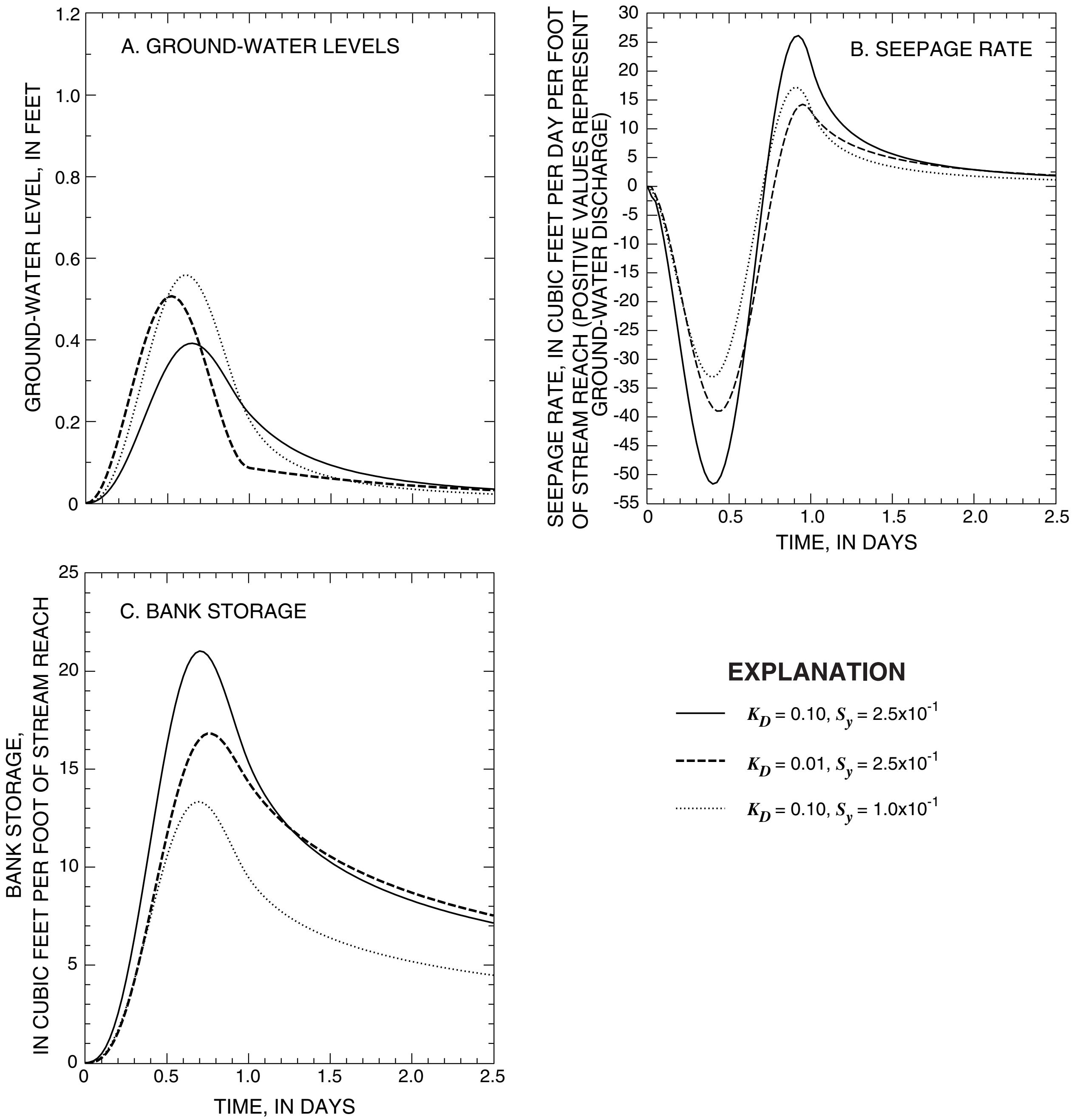


Figure 9. Effect of aquifer hydraulic properties on the response of a hypothetical, semi-infinite water-table aquifer to a sinusoidal stream-stage fluctuation. Hydraulic properties of the aquifer shown in table 1, unless otherwise specified. (*A*) Ground-water levels, 100 feet from streambank. (*B*) Seepage rate. (*C*) Bank storage. (K_D, ratio of vertical to horizontal hydraulic conductivity; S_y, specific yield.)

Lower values of vertical hydraulic conductivity (K') and specific yield (S_y') of a water-table aquitard also result in more rapidly rising and higher peak ground-water levels, lower seepage rates, and less bank storage. Lower values of K' in a water-table aquitard yield reduced flow through the aquitard and lower values of S_y' result in the release of smaller volumes of water from the aquifer per unit change in head. Also, as the thickness (b') of a water-table aquitard decreases, the response of a leaky aquifer overlain by a water-table aquitard approaches that of a water-table aquifer.

Low-permeability streambank material dampens the response of an aquifer to stream-stage fluctuations. The leakance term, a (eq. 1), accounts for semipervious streambank material in the analytical solutions and computer programs. Higher values of a can result from either increased streambank thickness or decreased streambank hydraulic conductivity (K_s relative to K_x). Thus, higher leakance values represent increased hindrance by the streambank material to the transfer of water across the stream-aquifer boundary. Higher leakance values also result in reduced ground-water levels, seepage rates, and bank storage volumes in the aquifer (fig. 10).

When recharge is simulated in addition to a stream-stage fluctuation, seepage to the aquifer and bank-storage volumes are less than, or even the reverse of, those that result from the stream-stage fluctuation alone. Recharge alone, simulated as a 1-day, linear event, results in seepage from the aquifer to the stream at rates that increase during the 1-day period of recharge and then decrease exponentially with time after the recharge ends (fig. 11*A*). Without recharge, the 1-day, sinusoidal flood wave results in an initial pulse of water moving from the stream to the aquifer during the first 0.7 days, which gradually drains from the aquifer after the flood wave passes. When recharge and the flood wave are simulated simultaneously, seepage rates at the stream-aquifer interface resulting from these two stresses are superimposed. Initially, after a brief period of ground-water discharge, seepage is into the aquifer from the stream but at lower rates than would result from the flood wave alone; subsequently, seepage is from the aquifer to the stream but at higher rates than would result from recharge alone (fig. 11*A*). The net result is that the total volume of bank storage because of the flood wave is reduced when recharge also is simulated (fig. 11*B*). The bank storage also drains more quickly from the aquifer when recharge and the flood wave are simulated simultaneously. The cumulative volume of ground water discharged to the stream (represented as negative values of bank storage, fig. 11*B*) increases until all of the recharged water has drained (for the finite-width aquifer simulated in fig. 11, this occurs after about 40 days). For a finite-width aquifer, total recharge to the aquifer is equal to the recharge rate multiplied by the width of the aquifer and the duration of the recharge event; the cumulative volume of ground-water discharge approaches this value. However, for a semi-infinite aquifer, seepage continues and cumulative ground-water discharge increases indefinitely. For this reason, the calculated bank storage for the finite-width (x_L equal to 500 ft) and semi-infinite aquifers begin to diverge after about 4 days (fig. 10*B*).

In summary, aquifer type, aquifer geometry, and hydraulic properties of an aquifer, aquitard, and semipervious streambank all are important in determining the response of the aquifer to stream stage and recharge stresses. Recharge also is important in determining net seepage rates and bank-storage volumes for a period of simultaneous stream-stage fluctuations. In some cases, similar responses to a given stream-stage fluctuation can be obtained using different combinations of aquifer boundaries and hydraulic parameters. For example, simulating an aquifer as confined with a semipervious streambank (fig. 10*A*) or as leaky with a water-table aquitard and no semipervious streambank (fig. 6*A*) can yield similar responses in ground-water levels, although slightly different hydraulic parameters may be needed in the simulation. Thus, conceptualizations and simulations of a stream-aquifer system should incorporate as much field-based hydrogeologic information as possible.

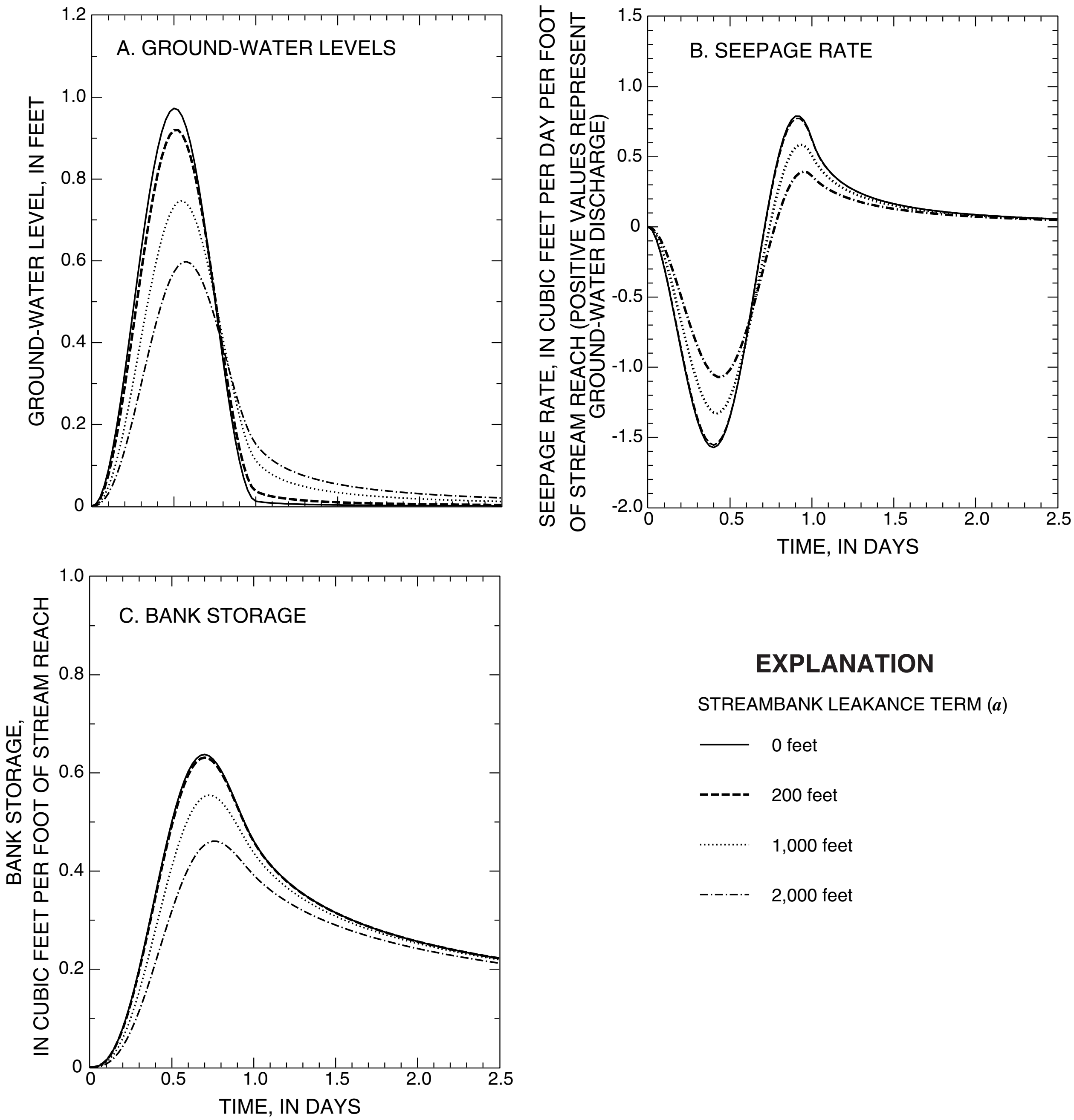


Figure 10. Effect of streambank properties on the response of a hypothetical, semi-infinite confined aquifer to a sinusoidal stream-stage fluctuation. Hydraulic properties of the aquifer and streambank shown in table 1, unless otherwise specified. (*A*) Ground-water levels, 100 feet from streambank. (*B*) Seepage rate. (*C*) Bank storage.

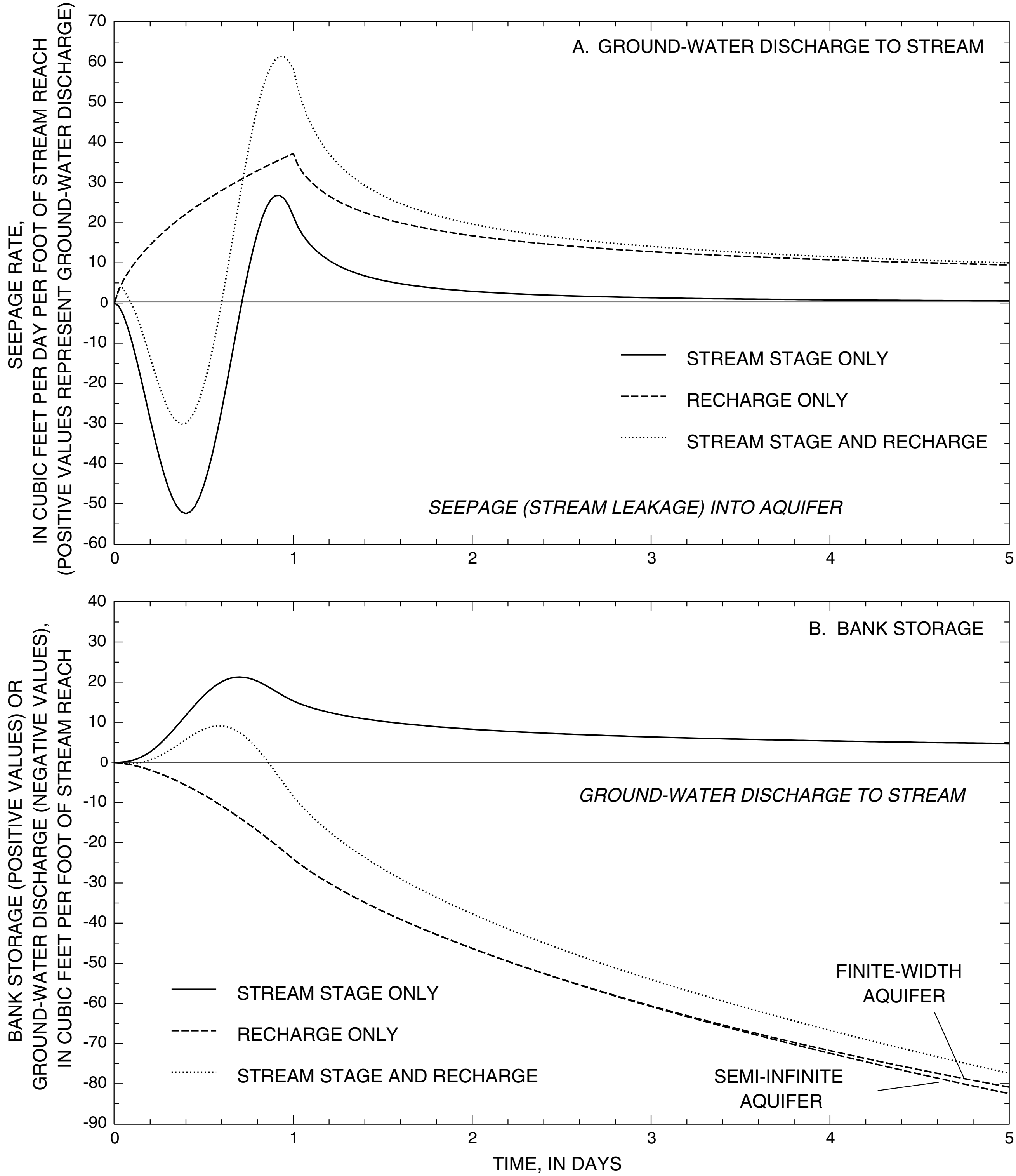


Figure 11. Response of a hypothetical water-table aquifer to a 1-day linear recharge event and a 1-day sinusoidal stream-stage fluctuation. Aquifer is semi-infinite unless indicated; hydraulic properties of aquifers shown in table 1. (*A*) Seepage rate. (*B*) Bank storage and ground-water discharge.

ANALYSIS OF STREAM-AQUIFER HYDRAULIC INTERACTION IN FIELD APPLICATIONS

Published data from alluvial and stratified-drift aquifers in Kentucky, Massachusetts, and Iowa are used to demonstrate applications of the programs to field settings. In each case, available hydrogeologic information is used to develop a conceptual model of the stream-aquifer system. The conceptual model then forms the basis for development of an analytical model of the site, using either STLK1 or STWT1. The analytical model is calibrated by matching ground-water levels calculated by STLK1 or STWT1 to measured ground-water levels.

Tennessee River Alluvial-Aquifer System, Calvert City, Kentucky

Ground water in the alluvial aquifer at a site along the Tennessee river, western Kentucky (fig. 12), is contaminated with benzene, 1,2-dichloroethane, and other volatile organic compounds (VOCs). Ground-water levels in the aquifer fluctuate in response to changes in river stage caused by releases from Kentucky Dam. During periods of high river stage, ambient hydraulic gradients toward the river are reversed. Infiltration of river water into the aquifer during floods at this site was of concern because the infiltrated water could move ground-water contaminants toward nearby municipal and industrial supply wells. Ground-water levels and stream stage measured in January and February 1988 were used in the present study to estimate hydraulic properties of the alluvial aquifer by calibration of measured ground-water levels to ground-water levels calculated with STLK1. The calibrated stream-aquifer model then was used to determine seepage rates between the alluvial aquifer and Tennessee River and bank-storage volumes in the aquifer during the 38-day study period. STLK1 also was used to evaluate how different conceptualizations of the aquifer affect calculated water levels, seepage rates, and bank-storage volumes. Information presented on the geology, hydrogeology, and interaction of the alluvial aquifer with the Tennessee River is based on the reports of Dames and Moore (1988a, 1988b, 1991) and Starn and others (1995). Stream-stage and ground-water-level data are from Dames and Moore (1988a).

Site Description

The study site is located on a stream terrace and flood plain adjacent to the Tennessee River, in an industrial complex about 2 mi north of Calvert City and 4 mi downstream of Kentucky Dam. The alluvium occurs inside a broad meander of the Tennessee River, forming a continuous, 2-mile wide band south of the river and extending about 5 mi farther south in irregular fingers (fig. 12). It overlaps and is bounded to the south and east by a permeable, Cretaceous-age deposit of sand, clay, and gravel. Weathered and fractured limestone bedrock of Cretaceous age underlies the alluvium in most areas. An intervening clay layer also is present between the alluvium and bedrock near the river in some areas.

The alluvial aquifer was formed by glacial-meltwater deposition and has been divided into a lower, middle, and upper alluvium. A hydrogeologic section perpendicular to the river near the study site is shown in figure 13. The alluvial deposits near the study site are about 100 ft thick. The lower alluvium consists of well-sorted gravelly sand and sandy gravel; the middle alluvium consists of interbedded sand, silt, and clay; and the upper alluvium consists of lake-bed silt and clay. The lower alluvium is the most permeable of the units and is considered to be the aquifer. Ground-water levels measured in wells as far as 750 ft from the Tennessee River in the alluvial aquifer indicate that river-stage changes propagate rapidly through the lower alluvium (fig. 14; Dames and Moore, 1988b). The middle alluvium probably forms a confining layer to the underlying, more permeable lower unit and, based on water-level data collected during an aquifer test, likely allows some leakage to the lower unit (Starn and others, 1995). The upper alluvium may act as an impermeable layer above the middle alluvium, but the upper layer generally is not present in the flood plain immediately adjacent to the river channel (Dames and Moore, 1991). Recharge to the lower alluvial aquifer consists of ground-water flow from the adjacent Cretaceous-age sands (fig. 12), precipitation, and infiltration from the river; precipitation recharge is likely limited by the low-permeability upper alluvium where it is present. Ground water generally flows toward the river in the direction of decreasing potentiometric head (fig. 12).

The Tennessee River is about 1,300 ft wide near the study site and penetrates about one-third to one-half of the total saturated thickness of the alluvial

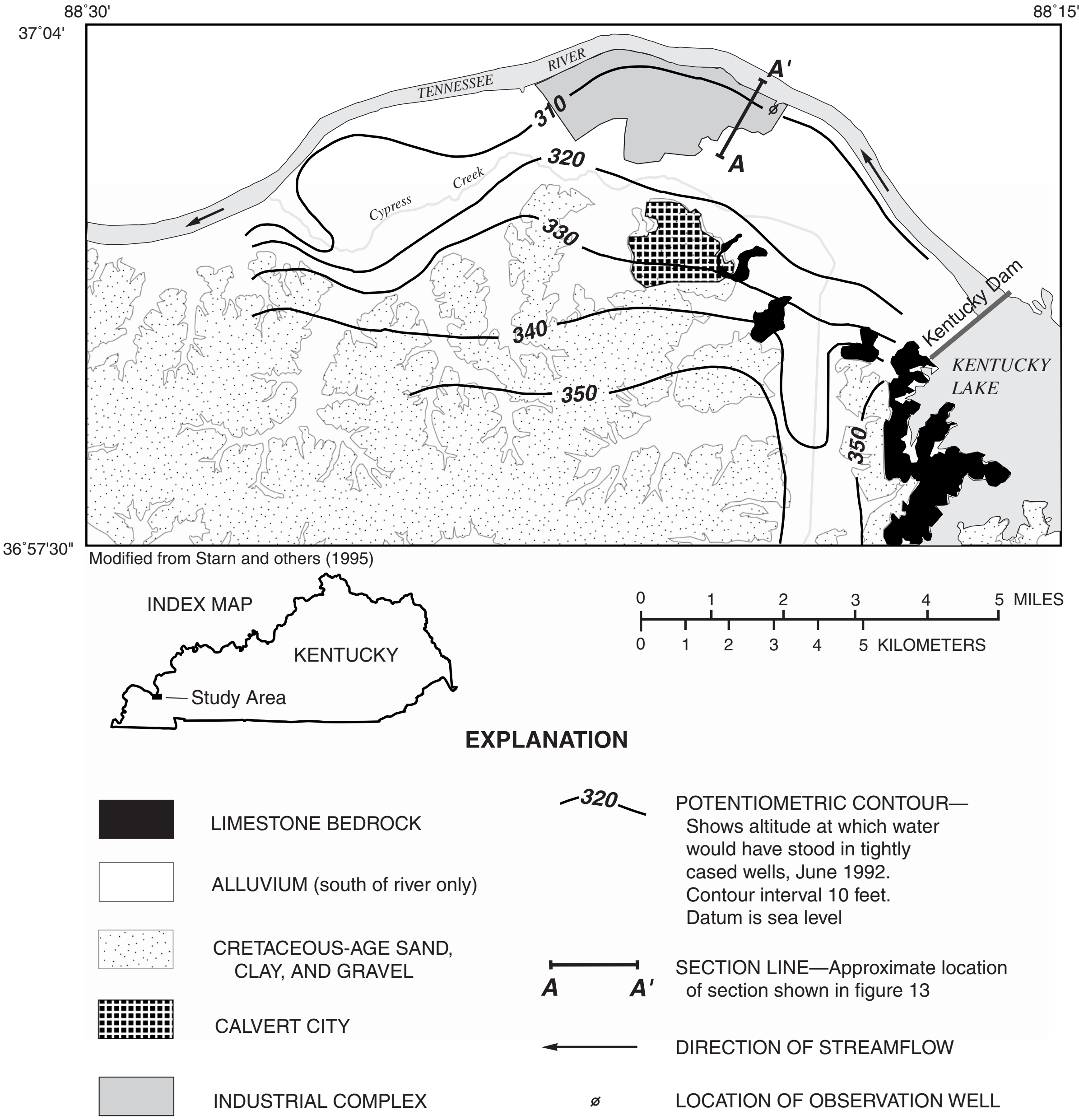


Figure 12. Tennessee River study site, extent of the alluvial aquifer, and potentimetric surface in the aquifer system, near Calvert City, Kentucky.

deposits, such that the river bottom is in the middle or lower alluvium (Dames and Moore, 1991; fig. 13). Annual mean discharge of the Tennessee River at a gage just downstream of Kentucky Dam, approximately 3 mi upstream of the study site, was 65,630 ft^3/s during the period 1965–84 (D.L. McClain, U.S. Geological Survey, personal commun., 1997). The streambed, which is about 20 ft thick in some places, consists of sands with lenses of organic-rich, fine material (Dames and Moore, 1991; Starn and others, 1995). Starn and others (1995) estimated the ratio of streambed hydraulic conductivity to streambed thickness, K_s/b, to be 0.00592 d^{-1} (a equal to 5×10^4 ft) by calibration of a two-dimensional, steady-state ground-water-flow model of the area around the study site.

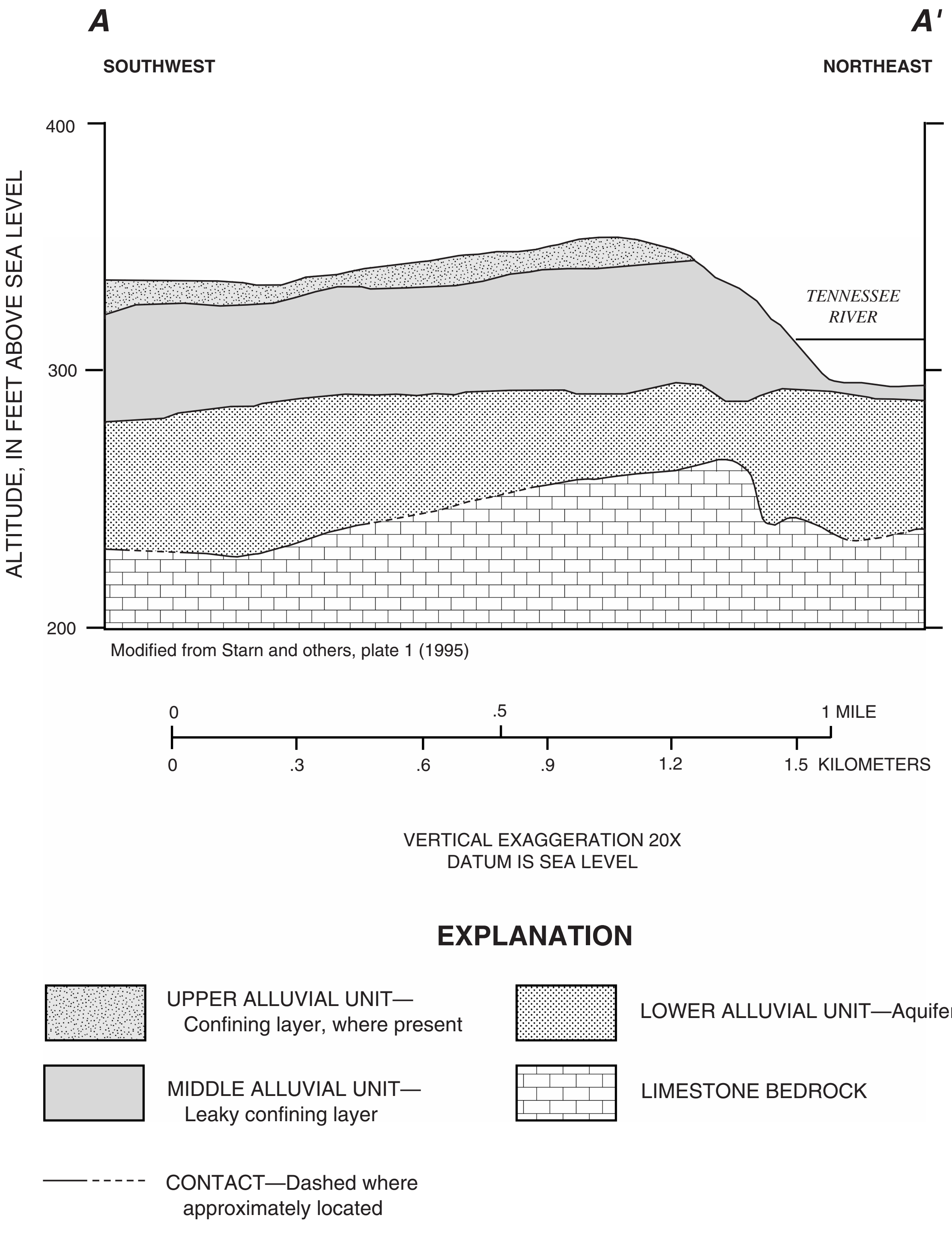


Figure 13. Hydrogeologic section through the Tennessee River alluvial aquifer near the study site. Approximate location of the section shown in figure 12.

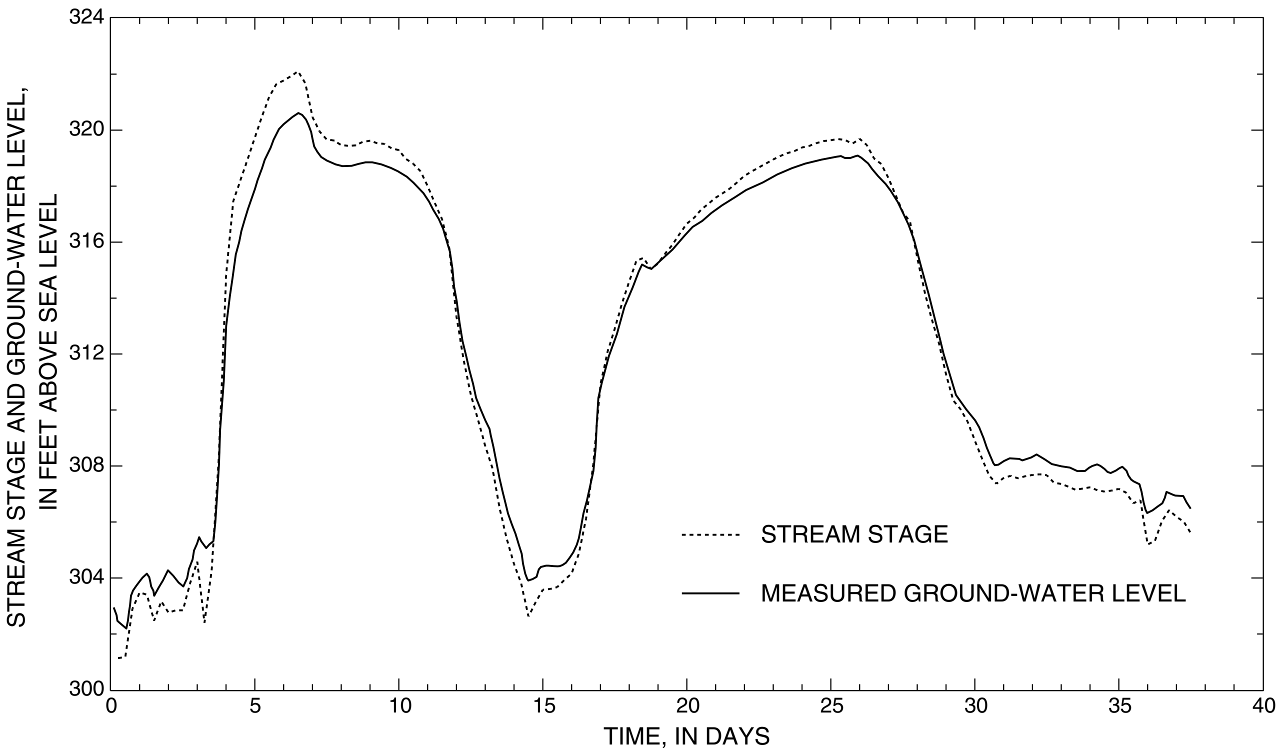


Figure 14. Stream stage and ground-water levels measured in an observation well located 125 feet from the streambank in the Tennessee River alluvial aquifer near Calvert City, Kentucky.

Analysis of Response of Stream-Aquifer System to Stream-Stage Fluctuations

The aquifer was simulated as a semi-infinite, leaky aquifer overlain by a water-table aquitard (see fig. 15); this conceptual model was based on information from Starn and others (1995), Dames and Moore (1991), and J.J. Starn (U.S. Geological Survey, oral commun., 1997). The observation well used in the present analysis was located 125 ft from a man-made indentation in the river (fig. 12) and screened at 21 to 26 ft below the ambient water table, or about 50 to 55 ft below land surface. Stream stage and ground-water levels were measured at 6-hour (0.25 day) intervals. Recharge to the lower alluvium, estimated by Starn and others (1995) to be 0.0005 ft/d, was assumed to be negligible in the present analysis. Aquifer and aquitard thicknesses were estimated from cross sections of the alluvium at and near the observation well. Aquifer thickness is about 15 to 30 ft between the observation well and the river and increases away from the river; the thickness of the aquitard is about 30 ft. Initial estimates of aquifer, aquitard, and streambank hydraulic properties used to calibrate the model were available from descriptions of geologic materials, slug tests, aquifer tests, and calibration of a steady-state numerical model. Sample input and output files used in analysis of this site with STLK1 are provided in the appendix.

The calibrated stream-aquifer model (obtained by trial and error using STLK1) that corresponds most closely with the available data and the conceptual model of the system uses values of K_x equal to 300 ft/d and S_s equal to 1×10^{-5} ft^{-1} for the aquifer; values of K' equal to 0.5 ft/d, S'_s equal to 1×10^{-4} ft^{-1}, and S'_y equal to 0.07 for the aquitard; and no semipervious streambank material (table 2). Ground-water levels calculated by STLK1 with these hydraulic properties closely match the measured ground-water levels at the well for the 38-day period (fig. 16*A*; the best-fit model is shown as the leaky aquifer with a water-table aquitard, although the other models fit well also, as discussed below). The model-calibrated values of aquifer hydraulic conductivity and thickness correspond to a transmissivity of 9,000 ft^2/d, which is between transmissivity estimates of 20,000 ft^2/d, determined with a calibrated numerical model of the alluvial aquifer for the area shown in figure 12 (Starn

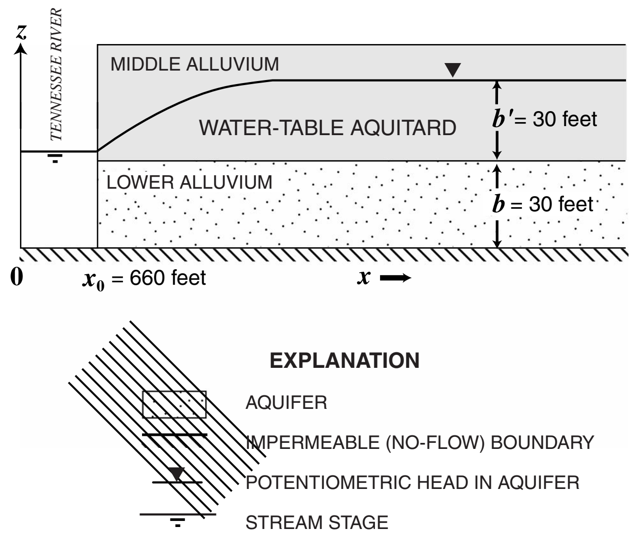


Figure 15. Conceptual model of the Tennessee River alluvial aquifer near Calvert City, Kentucky, used for simulation with computer program STLK1. (*b*; saturated thickness of the aquifer; *b*′; saturated thickness of the aquitard; and x_0, distance from the stream center to the stream-aquifer boundary.)

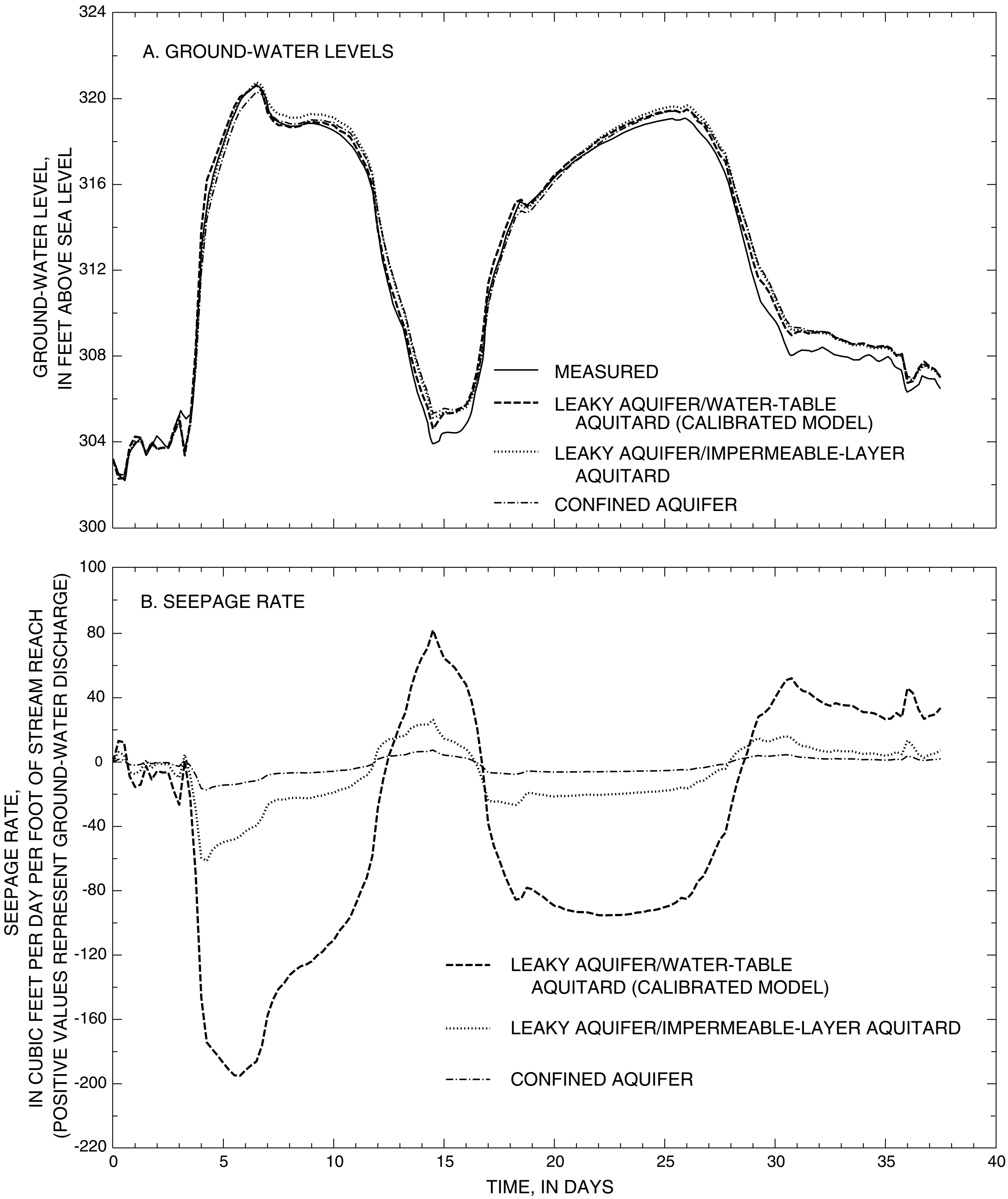
and others, 1995), and estimates of 2,940 ft^2/d and 5,000 ft^2/d, determined from an aquifer test and a calibrated numerical model, respectively, of the alluvial aquifer near the Tennessee River (Davis and others, 1973; Dames and Moore, 1991). The transmissivity value determined with the stream-aquifer analytical model represents conditions near the 125-foot observation well adjacent to the river, where the lower alluvial aquifer is thin (Dames and Moore, 1991). The estimate of Starn and others (1995) may be larger than the estimate from the analytical model because Starn and others (1995) modeled a larger area, where the lower alluvial aquifer generally is thicker. The difference between the modeled areas of the numerical model (Starn and others, 1995) and the analytical model also may explain the difference in the streambed leakance term used in the two models.

Alternative models of the stream-aquifer system, in which the aquifer was modeled in separate simulations as a leaky aquifer with an impermeable layer over the aquitard and as a confined aquifer, yielded calculated ground-water levels that agreed equally well with measured values as the best-fit model (fig. 16*A*). These alternative models use the same physical and hydraulic properties as the best-fit model (table 2), except that semipervious streambank material is simulated. The streambank-leakance term, *a*, equals 400 ft for the leaky aquifer with an impermeable layer over the aquitard and equals 2,000 ft for the confined aquifer. These values are reasonable based on the available information—values of *a* equal to 400 ft or 2,000 ft could correspond to 5-foot-thick streambed sediments with a hydraulic conductivity, K_s, equal to 4 ft/d or 0.8 ft/d, respectively. The leakance values used in the analytical models are lower (representing less impedance by the streambed to flow) than the value of 50,700 ft that corresponds to the streambed properties determined by Starn and others (1995); this discrepancy may reflect the difference between conditions in the area near the observation well, to which the stream-aquifer model was calibrated, and conditions averaged across the larger area simulated with the numerical model.

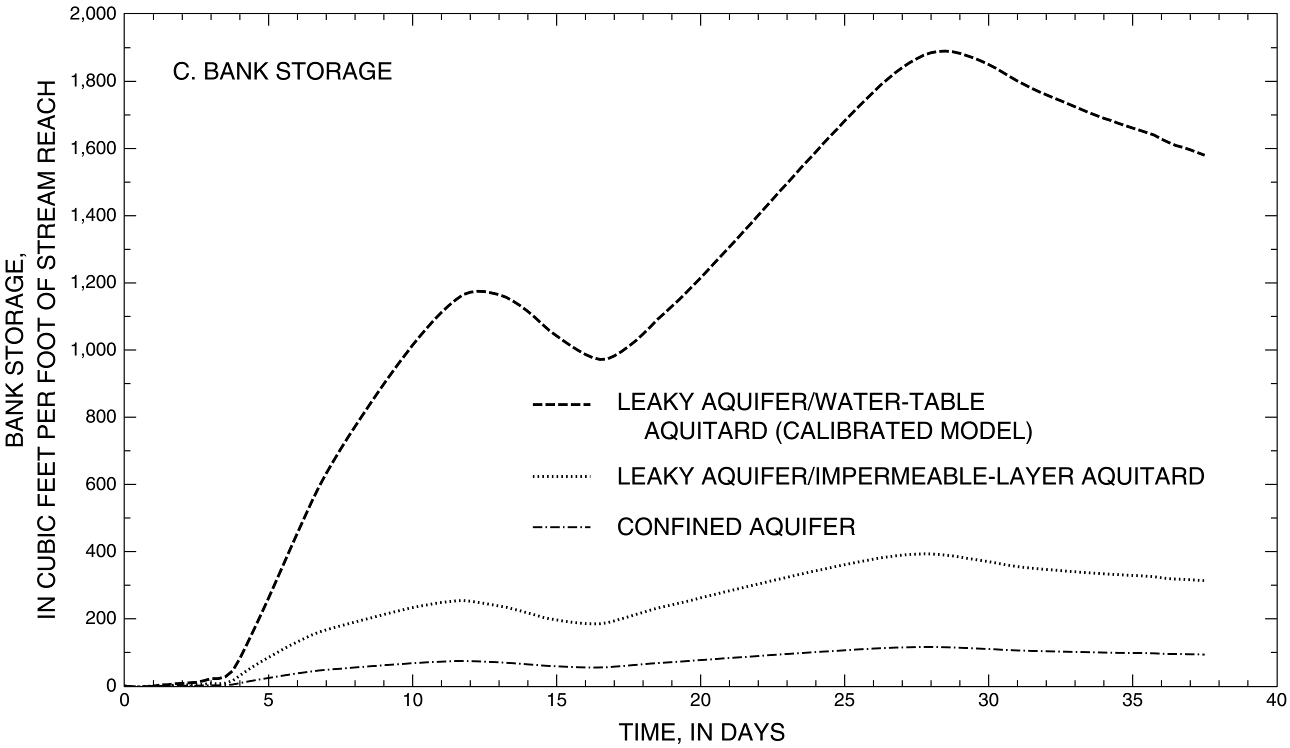
Table 2. Physical and hydraulic properties of stream-aquifer systems used in calibrated models for three alluvial and stratified-drift aquifers in Kentucky, Massachusetts, and Iowa

[Physical and hydraulic properties as defined in the text and(or) figure 1; ft, foot; ft/d, foot per day; ft^{-1}, per foot; ∞, infinity; --, not applicable]

Physical and hydraulic property	Field application: Tennessee River alluvial aquifer	Field application: Blackstone River stratified-drift aquifer	Field application: Cedar River alluvial aquifer
Aquifer Properties			
Type	Leaky, water-table aquitard	Confined	Water table
K_x (ft/d)	300	200	309
K_z (ft/d)	--	--	62
K_D (dimensionless)	--	--	0.2
b (ft)	30	47.6	30
S_s (ft^{-1})	1x10^{-5}	2x10^{-5}	3x10^{-5}
S_y (dimensionless)	--	--	0.2
x_L (ft)	∞	1,395	1,490
Aquitard Properties			
K' (ft/d)	0.5	--	--
b' (ft)	30	--	--
S_s' (ft^{-1})	1x10^{-4}	--	--
S_y' (dimensionless)	0.07	--	--
Properties of the Stream and Semipervious Streambank Material			
x_o (ft)	660	75	177
K_s (ft/d)	--	1.4	16
d (ft)	--	2.4	1.6
a ($K_x d/K_s$, ft)	0	340	31



Figue 16. Calculated ground-water levels, seepage rate, and bank storage in the Tennessee River alluvial aquifer near Calvert City, Kentucky, in response to a 38-day stream-stage fluctuation. (*A*) Calculated and measured ground-water levels. (*B*) Seepage rate.



Figue 16. Calculated ground-water levels, seepage rate, and bank storage in the Tennessee River alluvial aquifer near Calvert City, Kentucky, in response to a 38-day stream-stage fluctuation. (*C*) Bank storage—*Continued*.

Given the uncertainty generally associated with determining the streambank leakance term, identification of the stream-aquifer analytical model that most accurately describes field conditions must be based on available data and knowledge of hydrogeologic conditions at the site, such as the aquifer-test results discussed previously (p. 23). This is important because, although the three models result in similar calculated ground-water levels, the seepage rates and bank-storage volumes associated with these ground-water-level changes are very different for the three models (figs. 16*B* and *C*). Calculated seepage rates for the simulated leaky aquifer with a water-table aquitard (the calibrated model) are about 4 times larger than seepage rates calculated for the leaky aquifer overlain by an impermeable layer and about 20 times larger than seepage rates calculated for the confined aquifer. Bank-storage volumes among the three alternatives also differ substantially. These results indicate that inappropriate or inaccurate simulation of the aquifer, aquitard, or boundary conditions could lead to inaccurate conclusions about the flood-wave-induced movement of river water into the alluvial aquifer at the Tennessee River site. In this case, the most appropriate model to use is the best-fit model shown in table 2, because this model is most consistent with independent hydrologic evidence (water-level data from an aquifer test, as described previously) indicating that the alluvial aquifer behaves like a leaky aquifer overlain by a water-table aquitard.

Blackstone River Stratified-Drift Aquifer System, South Grafton, Massachusetts

Ground water in the stratified-drift aquifer at a site along the Blackstone River, central Massachusetts (fig. 17), is contaminated with trichloroethylene, 1,2-dichloroethene, vinyl chloride, and other VOCs. The VOC contamination is within 1,000 ft of existing and proposed municipal supply wells that are near the river farther downstream. Hydraulic interaction of the aquifer and stream are of concern at this site because of possible discharge of the VOC plume to the river and contamination of supply wells through induced infiltration. Ground-water levels in the aquifer fluctuate rapidly in response to changes in stream stage caused by operation of a small hydroelectric facility about 1 mi downstream from the study site. Ground-water levels and stream stage measured in September 1994 were used in the present study to evaluate hydrologic conditions in the aquifer and to determine hydraulic properties of the stratified drift by calibration to ground-water levels calculated with program STWT1. The calibrated stream-aquifer model then was used to estimate seepage rates between the aquifer and the Blackstone River and bank-storage volumes associated with the daily operation of the hydroelectric dam. The hydrogeology and interaction of the stratified-drift aquifer with the Blackstone River, as described here, is based on the reports of Whitman and Howard (1983, 1990), BSC Engineering (1986), Walker and Krejmas (1986), and HMM Associates (1993, 1994). Stream stage and ground-water-level data are from DeSimone and Barlow (1995).

Site Description

The study site is near an abandoned textile mill on the Blackstone River in South Grafton, Massachusetts (fig. 17). The stratified-drift was deposited by glacial-meltwater 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 wide at the mill site. Recharge to the aquifer is from precipitation and inflow from the adjacent till and

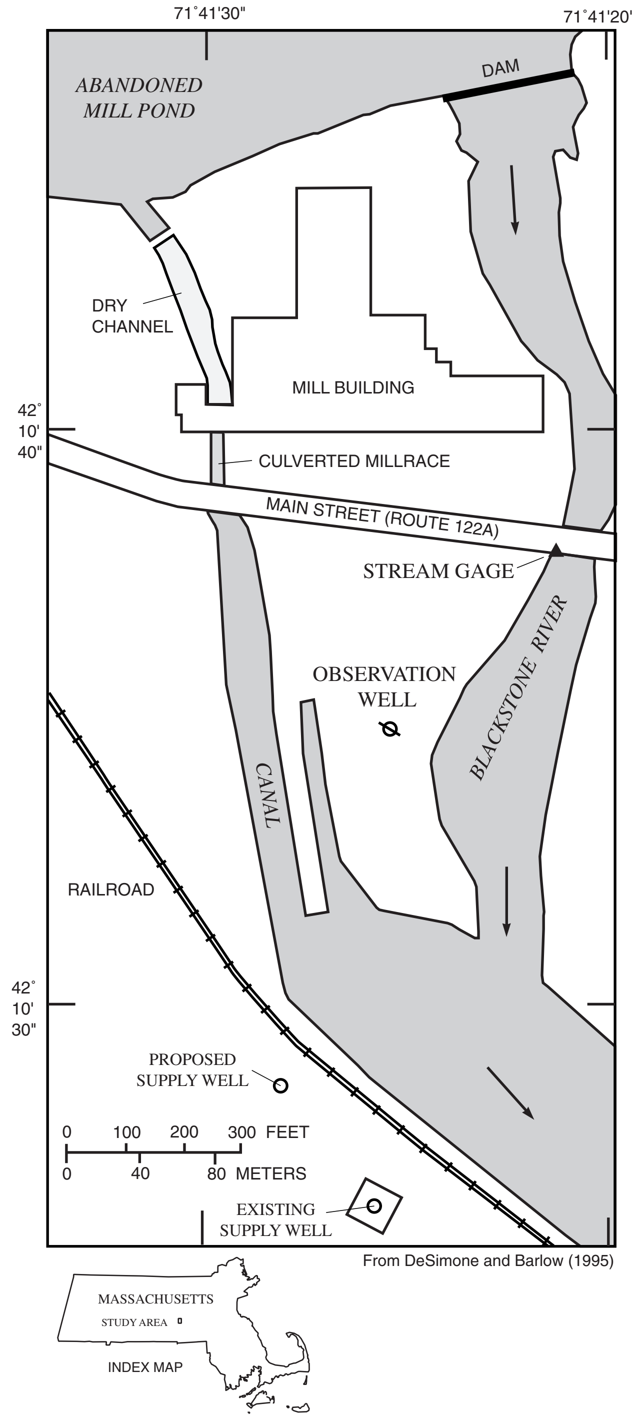


Figure 17. Blackstone River study site, South Grafton, Massachusetts.

bedrock, and 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 to 50 ft of coarse to medium sand and gravel or coarse to fine sand and gravel with traces of silt (HMM Associates, 1993). An intermittent peat layer, 2 to 8 ft thick, was observed within the coarse deposits at well-boring locations near the intersection of the canal and road (fig. 17). The sand-and-gravel deposits are underlain by about 5 to 20 ft of dense, sandy glacial till and granitic or schistose bedrock (HMM Associates, 1993).

The Blackstone River partially penetrates the stratified-drift aquifer and the river has a low-permeability streambed. The vertical hydraulic conductivity of the streambed sediments was determined by seepage-meter measurements to average 1.4 ft/d (Whitman and Howard, 1990); no information was available on streambed thickness. The river generally receives ground-water discharge, except where pumping wells at this site and elsewhere may induce infiltration of river water into the aquifer. The river averages about 150 ft in width at the site. Annual mean discharge of the river at Woonsocket, R.I., about 15 mi downstream from the study site, is 775 ft^3/s (1929–96) and was above average (837 ft^3/s) in 1994, the year in which the stream-stage and ground-water-level data were collected.

Ground-water levels were measured in an observation well in an area of the aquifer immediately south of the textile mill and west of the river, about 1,000 ft downstream from the abandoned mill pond and dam (fig. 17). Formerly, the pond and main stem of the river were freely connected through a canal (which runs through a culvert beneath the mill building to Route 122A, fig. 17) about 250 ft west of the observation well. Currently (1998), the pond is partly dry and water moves through the canal only during extremely high flows. Backwater from the river fills the canal south of Route 122A. Fuel oil has seeped into the canal from leaking underground storage tanks and covers the canal banks in some areas.

Stream stage in the river is affected by operation of a small hydroelectric facility, about 1 mi downstream from the study site. Sudden releases of impounded water at the hydroelectric facility result in 1- to 2-foot fluctuations in stream stage within 24-hour periods at the site (fig. 18). Releases occurred almost daily during midweek in September 1994. Ground-water levels, which are about 3 to 5 ft below land surface in the stratified-drift aquifer adjacent to the river, respond rapidly to these sudden changes in stream stage (fig. 18).

Analysis of Response of Stream-Aquifer System to Stream-Stage Fluctuations

Based on information from Whitman and Howard (1983, 1990), BSC Engineering (1986), and HMM Associates (1994), the Blackstone River stratified-drift aquifer was initially simulated as a finite-width water-table aquifer with semipervious streambank material at the stream-aquifer boundary (fig. 19). A K_x of 200 ft/d was used based on a transmissivity value determined from an aquifer test conducted by Whitman and Howard (1983) for the nearby supply wells. Two lateral boundaries, one at the stratified-drift/upland boundary (x_L equal to 1,395 ft) and one at the canal (x_L equal to 325 ft) were tested. Ground-water levels were simulated at an observation well located 95 ft from the streambank and screened near the water table (figs. 17 and 19). The peat layer described previously, which might act as a local confining unit, was not noted at this or other observation wells in the southeastern part of the site. Stream stage (at the gage shown in fig. 17) and ground-water levels that were measured at 15-minute intervals were available for the analysis (DeSimone and Barlow, 1995). Precipitation was negligible during the study period (DeSimone and Barlow, 1995), and recharge was not simulated.

Analysis of the stream-aquifer system with program STWT1 determined that measured ground-water levels at the observation well could not be matched by calculated water levels if water-table conditions were assumed. Various values of K_x (150 to 250 ft/d), S_s ($1x10^{-5}$ to $5x10^{-4}$ ft^{-1}), S_y (0.15 to 0.30) and a (0 to 700 ft) that were consistent with the available data were tested. In all cases, the rapid decreases in ground-water levels resulting from the daily stream-stage fluctuations were underestimated when water-table conditions were simulated (fig. 20*A*). However, when confined conditions were simulated, calculated water levels closely matched measured values (fig. 20*B*; hydraulic properties as given in table 2). The best match was obtained with the aquifer boundary at the stratified-drift/upland boundary (x_L equal to 1,395 ft) rather than at the canal (x_L equal to 325 ft). Similar results were obtained with water levels measured at wells screened near the bottom of the aquifer at 95 and 250 ft from the streambank (not shown).

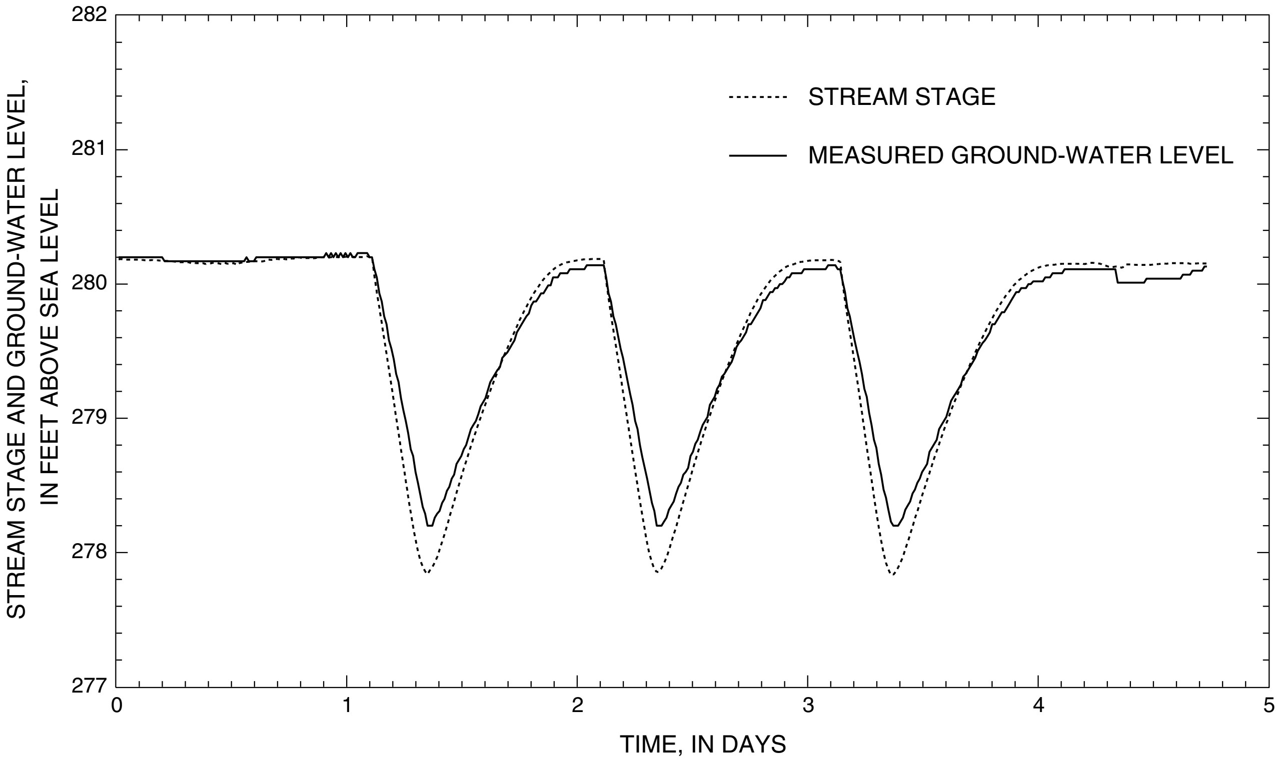


Figure 18. Stream stage and ground-water levels measured in an observation well located 95 feet from the streambank in the Blackstone River stratified-drift aquifer, South Grafton, Massachusetts.

Lithologic evidence indicates that the stratified-drift aquifer is under water-table conditions. Analysis with STWT1, however, suggests that the response of the aquifer to the rapid stream-stage fluctuations reflects the 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 about 3 to 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 the land surface (Bouwer and Rice, 1978, 1980). Narasimhan and Zhu (1993) and Gillham (1984) 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 that the effective specific yield is zero when the capillary fringe extends to the surface and that the response of the aquifer 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; A.F. Moench, U.S. Geological Survey, personal commun., 1997) and effects of delayed drainage take effect. The thickness of the capillary fringe varies with grain size--for example, from 0.5 ft in coarse sand to greater than 3 ft in coarse silt. Silt-rich layers were common in the stratified-drift aquifer at the Blackstone River site (HMM Associates, 1993). Therefore, the capillary fringe may have been fairly thick in some areas. Saturated conditions also were likely to have extended nearly to the land surface adjacent to the river, where the topography is generally flat.

Whether the stratified-drift aquifer at the Blackstone site behaves like a confined or water-table aquifer in response to the stream-stage fluctuations could have implications for contaminant transport and chemical transformation processes because of the large differences in calculated seepage rates and bank-storage volumes for the different aquifer types (fig. 21). Maximum seepage rates calculated with the

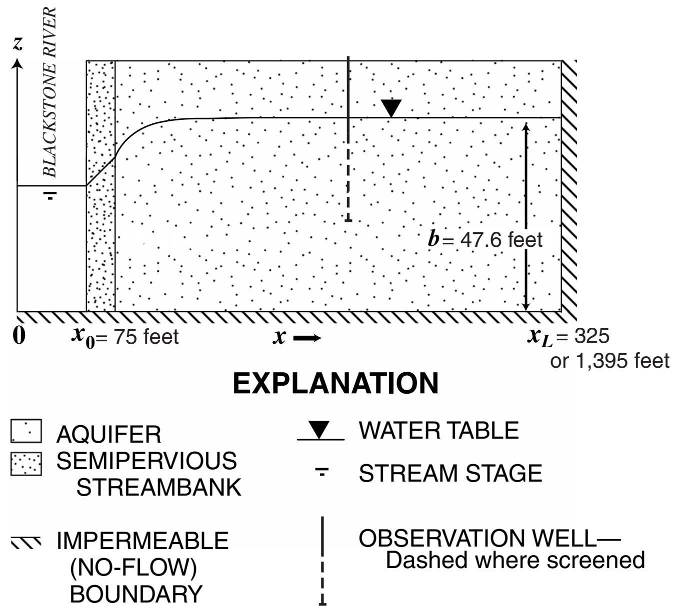


Figure 19. Conceptual model of the Blackstone River stratified-drift aquifer, South Grafton, Massachusetts, used for simulation with computer program STWT1. (*b*, saturated thickness of the aquifer; x_0, distance from the stream center to the stream-aquifer boundary; x_L, distance from stream center to lateral aquifer boundary.)

water-table solutions are nearly 200 ft^3/d per foot of stream reach, whereas maximum seepage rates calculated with the best-fit confined solutions are about 10 ft^3/d per foot of stream reach (aquifer width equal to 1,395 ft; fig. 21*A*). The large seepage rates calculated with the water-table simulations also are associated with large changes in bank storage for each daily stream-stage fluctuation (fig. 21*B*). The larger fluxes of water in the case of water-table aquifers, as compared to the confined aquifers, would result in more mixing of contaminated and uncontaminated water at the margins of the VOC plume and a more dynamic hydrologic regime near the river. This mixing and movement could promote geochemical conditions that favor the natural attenuation of some contaminants, or alternatively, could increase the distribution of other contaminants, such as light non-aqueous-phase liquids, in the saturated and unsaturated zones of the aquifer through hydromechanical dispersion or "smearing."

Note that bank storage appears to decrease with time for the water-table solutions, which suggests that an apparent net discharge of ground water results under water-table conditions that apparently does not result under confined conditions. This apparent difference reflects the different response times of the water-table and confined aquifers to the change in average stream stage. The average stage is high during the period of no fluctuation at the beginning of the simulation and decreases to a new, lower average stage during the period of fluctuating stage. Water levels in the water-table aquifer (fig. 20*A*) respond slowly to the net decrease in average stage and do not reach a new equilibrium after 3 days of stream-stage fluctuations as in the confined aquifer (fig. 20*B*).

Although the analytical models provide insight into the behavior of the stream-aquifer system at the Blackstone River site, analysis of hydraulic conditions at the site also illustrates some limitations of the analytical models. For example, because the analytical models are one-dimensional (or two-dimensional in the *x*, *z* directions for the water-table aquifer), the canal and its confluence with the stream cannot be simulated. Thus, the calculated aquifer response might not account for the effects of the fluctuating canal stage. The good agreement between calculated and measured ground-water levels that was obtained when the canal was not simulated (fig. 20*B*), however, suggests that flow in the canal is hydraulically disconnected from flow in the aquifer. Variations in aquifer thickness also cannot be simulated with STLK1 and STWT1. Hydraulic properties estimated by the calibrated model may have been affected by these simplifying assumptions about the aquifer geometry.

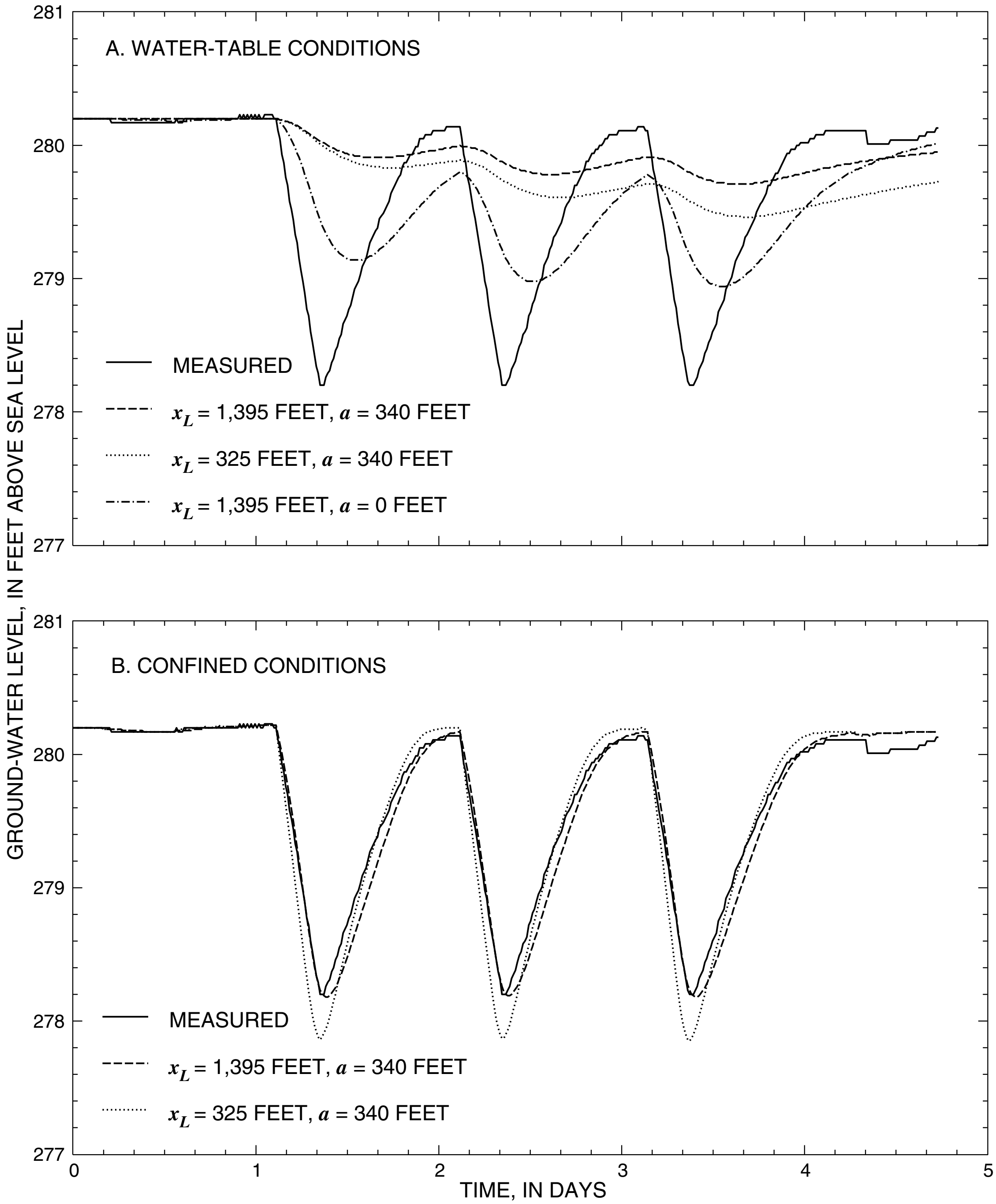


Figure 20. Calculated ground-water levels in the Blackstone River stratified-drift aquifer, South Grafton, Massachusetts, in response to three daily stream-stage fluctuations under water-table and confined conditions. Observation well is 95 feet from the streambank and screened near the water table. Hydraulic properties given in table 2 unless otherwise specified. (*A*) Water-table conditions (vertical hydraulic conductivity, 20 ft/d; ratio of vertical to horizontal hydraulic conductivity, 0.1; specific yield, 0.2). (*B*) Confined conditions. (x_L, aquifer width; a, streambank leakance term.)

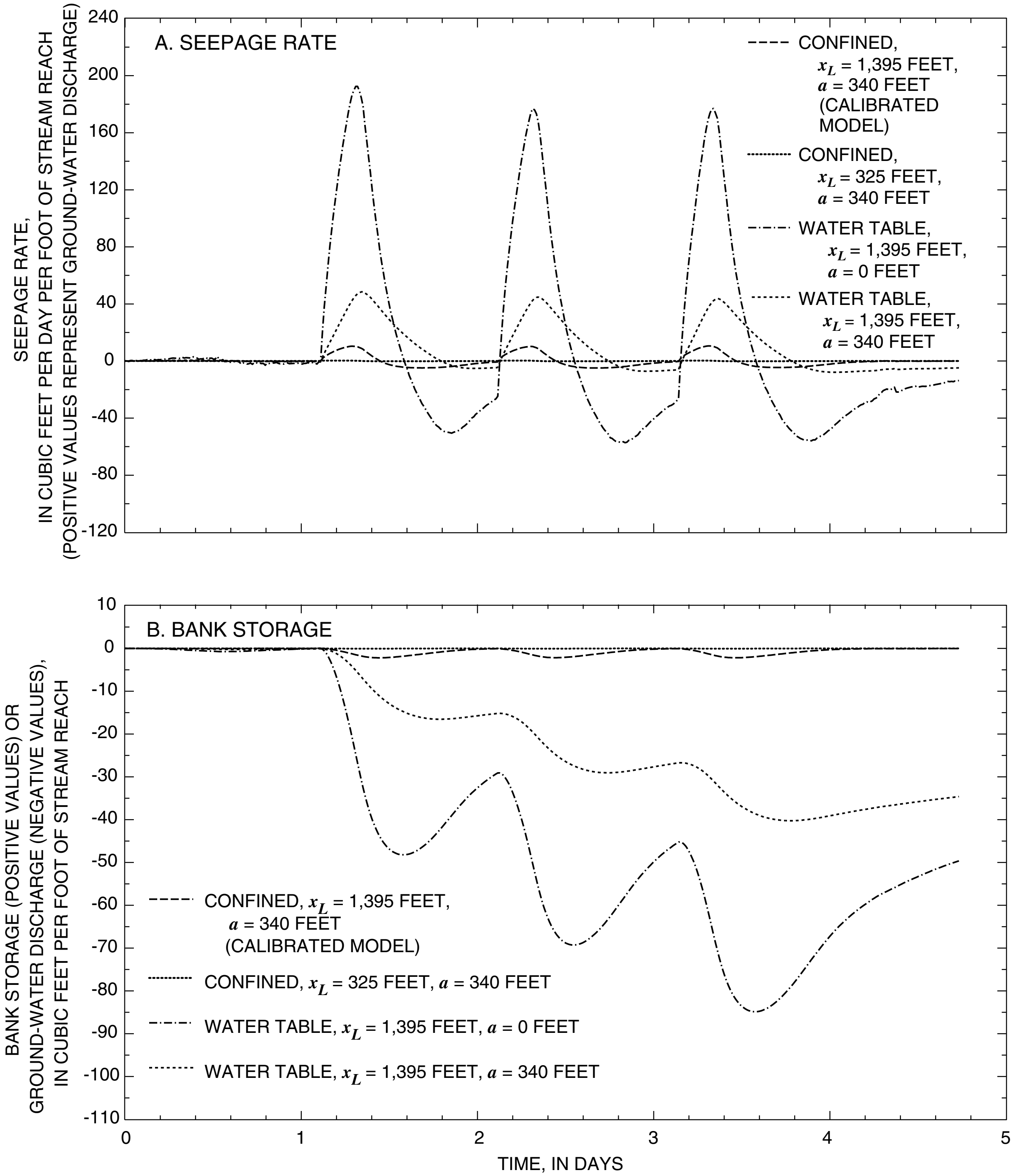


Figure 21. Calculated seepage rate and bank storage in the Blackstone River stratified-drift aquifer, South Grafton, Massachusetts, in response to three daily stream-stage fluctuations. (*A*) Seepage rate. (*B*) Bank storage. (x_L, aquifer width; a, streambank leakance term.)

Cedar River Alluvial-Aquifer System, Cedar Rapids, Iowa

Chemical and hydrologic evidence indicates that agricultural chemicals are transported into the alluvial water-table aquifer adjacent to the Cedar River in eastern Iowa (fig. 22) during periods of elevated stream stage caused by direct surface runoff (Squillace, 1996; Squillace and others, 1996). The aquifer then discharges water and chemicals to the river during declining stream stage. Stream stage and ground-water levels measured at a site about 6 mi southeast of Cedar Rapids, during a 1-day period of rapid change in stream stage in October 1988, were used in the present study to estimate hydraulic properties of the water-table aquifer and semipervious streambank material. The calibrated stream-aquifer analytical model then was used to evaluate bank storage during the 1-day flood event and during a 55-day period of simultaneous stream-stage fluctuations and recharge in March and April 1990. The geology, hydrogeology, aquifer geometry, interaction of the alluvial aquifer with the Cedar River, and movement of agricultural chemicals between the aquifer and river, as described here, are based on the reports of Squillace (1996), Squillace and others (1993, 1996), and Schulmeyer and others (1995). Stream-stage and ground-water-level data are from Schulmeyer and others (1995) and P.M. Shulmeyer (U.S. Geological Survey, written commun., 1997).

Site Description

The study site (fig. 22) is an unfarmed, wooded area on a flood plain of the Cedar River (Squillace and others, 1996). The alluvial aquifer at the site consists of glaciofluvial sediments deposited by meltwater streams during retreat of the Wisconsinan-age ice sheet. The alluvium forms a vertically heterogeneous, sand-rich, fining-upward sequence that reaches a maximum thickness of about 50 ft beneath the river and thins laterally (fig. 23). The alluvium is underlain by unfractured, low-permeability glacial till and carbonate bedrock and bounded laterally by till-covered uplands at a distance of about 1,310 ft from the streambank (1,490 ft from the stream center; Squillace and others, 1996). Recharge to the aquifer is from precipitation, leakage from ephemeral streams and ponds, bank storage from the Cedar River, and infiltration of flood water from the river. Discharge from the aquifer is by ground-water flow to streams and by evapotranspiration (Squillace and others, 1996). The Cedar River at the study site is about 350 ft wide and penetrates about one-fifth of the saturated thickness of the aquifer. Squillace and others (1996) indicate the presence of a thin layer of fine sediment along the riverbank and riverbed between the river and aquifer (fig. 23). Annual mean discharge of the Cedar River at Cedar Rapids was 3,687 ft^3/s during the period 1903–97 (May and others, 1998).

Analysis of Response of Stream-Aquifer System to a 1-Day Stream-Stage Fluctuation

Ground-water levels measured in partially penetrating observation wells near the river respond to stream-stage fluctuations in the river and to recharge to the aquifer. Water levels measured in three of these wells during a 1-day period in October–November 1989, in which the river stage rose and fell 4 ft in response to a sudden release of water from an upstream dam, were used to determine hydraulic properties of the alluvium and semipervious streambank material by calibration to ground-water levels calculated using STWT1. The rapid increase in stream stage and ground-water levels (fig. 24) followed a 6-month period of dry weather and low streamflow representing base-flow conditions (Squillace and others, 1996). Observation wells used in this analysis are 33, 98, and 164 ft from the streambank and are screened in coarse-grained sand at about 30 ft below land surface (fig. 23). Stream stage and ground-water levels at the three wells were measured at 15-minute intervals (P.M. Schulmeyer, U.S. Geological Survey, written commun., 1997). Note the change in amplitude of the ground-water levels with distance from the streambank shown in figure 24. Sample input and output files used in analysis of this site with STWT1 are provided in the appendix.

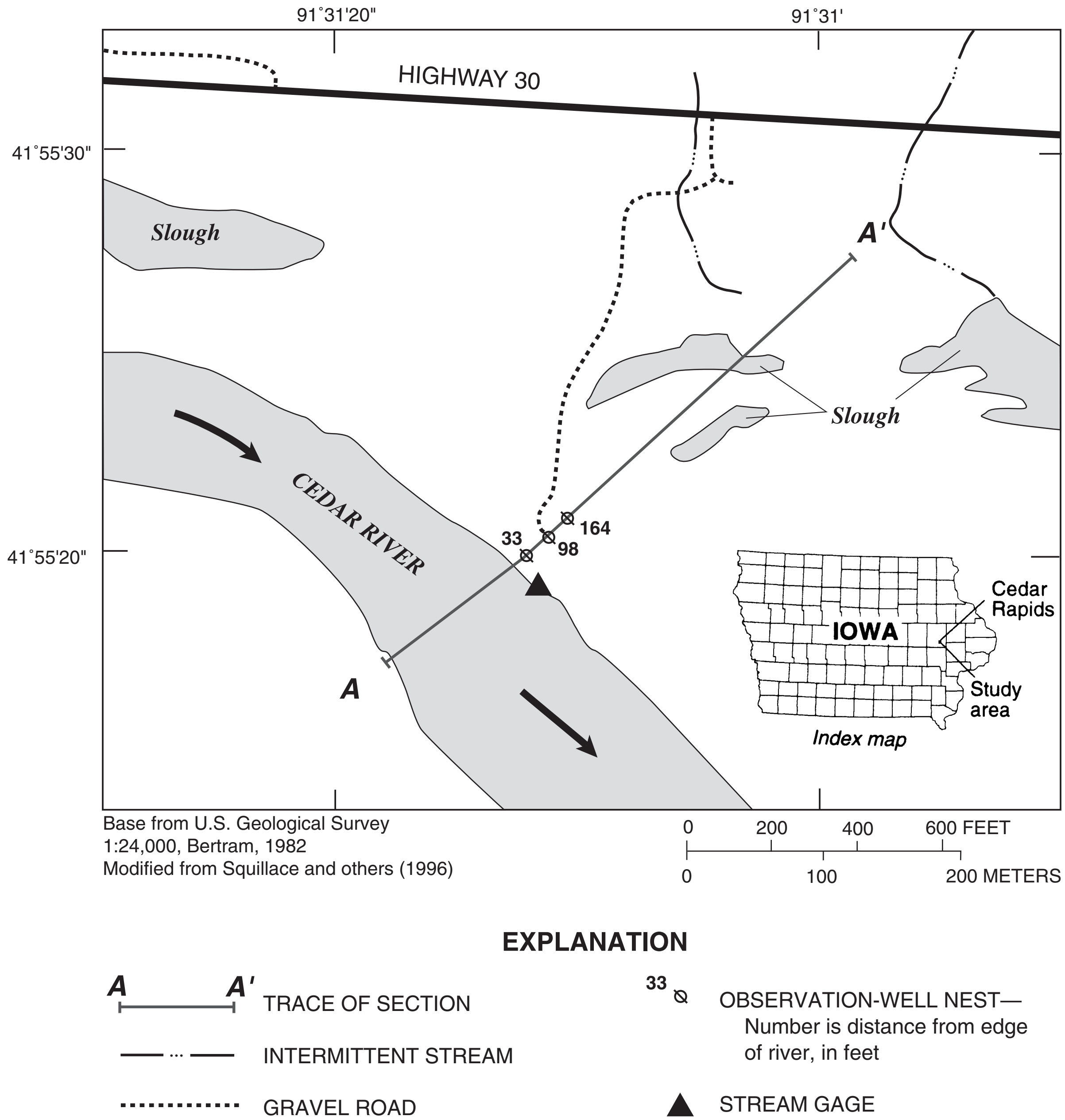


Figure 22. Cedar River study site near Cedar Rapids, Iowa.

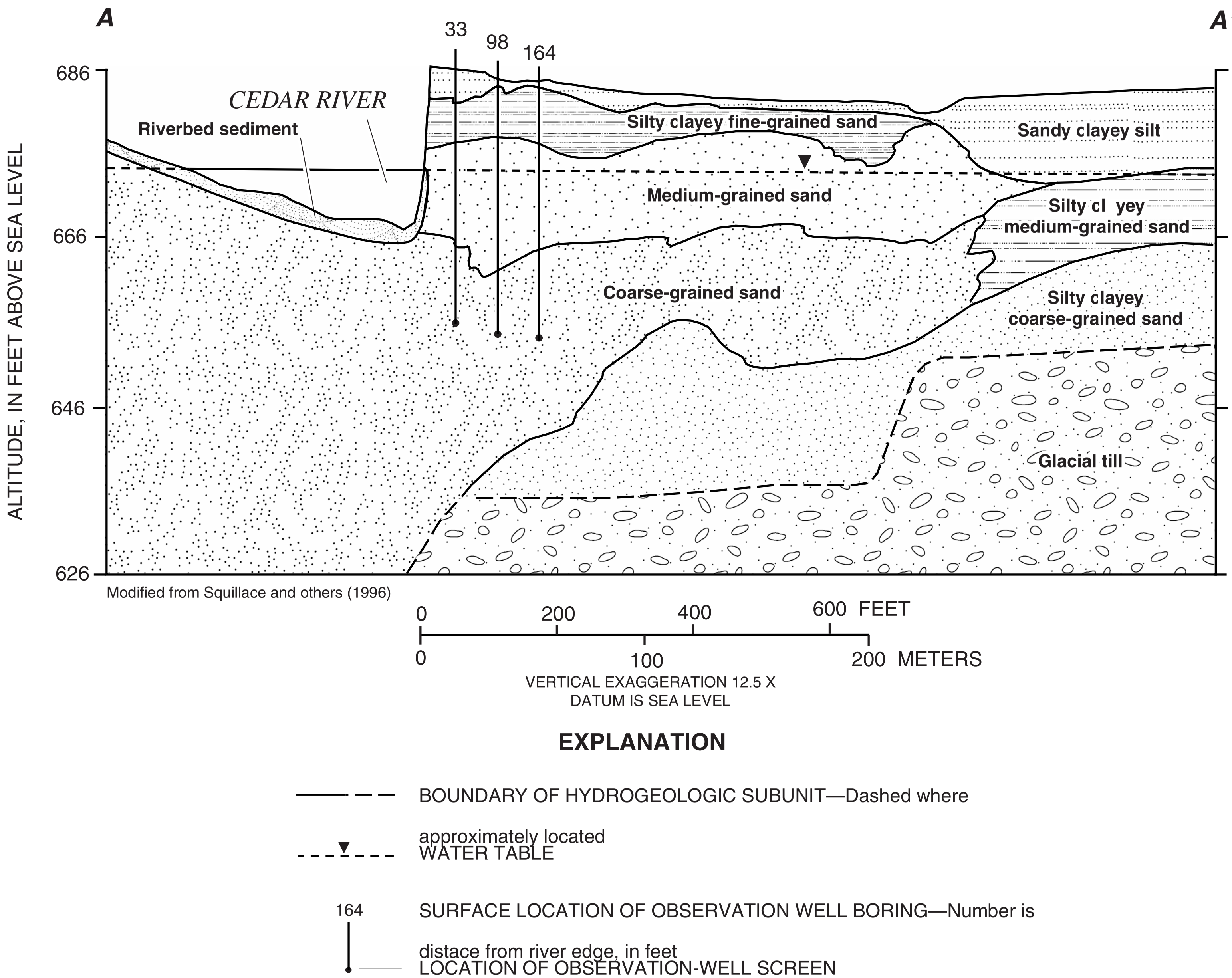


Figure 23. Hydrogeologic section of the Cedar River study site near Cedar Rapids, Iowa.

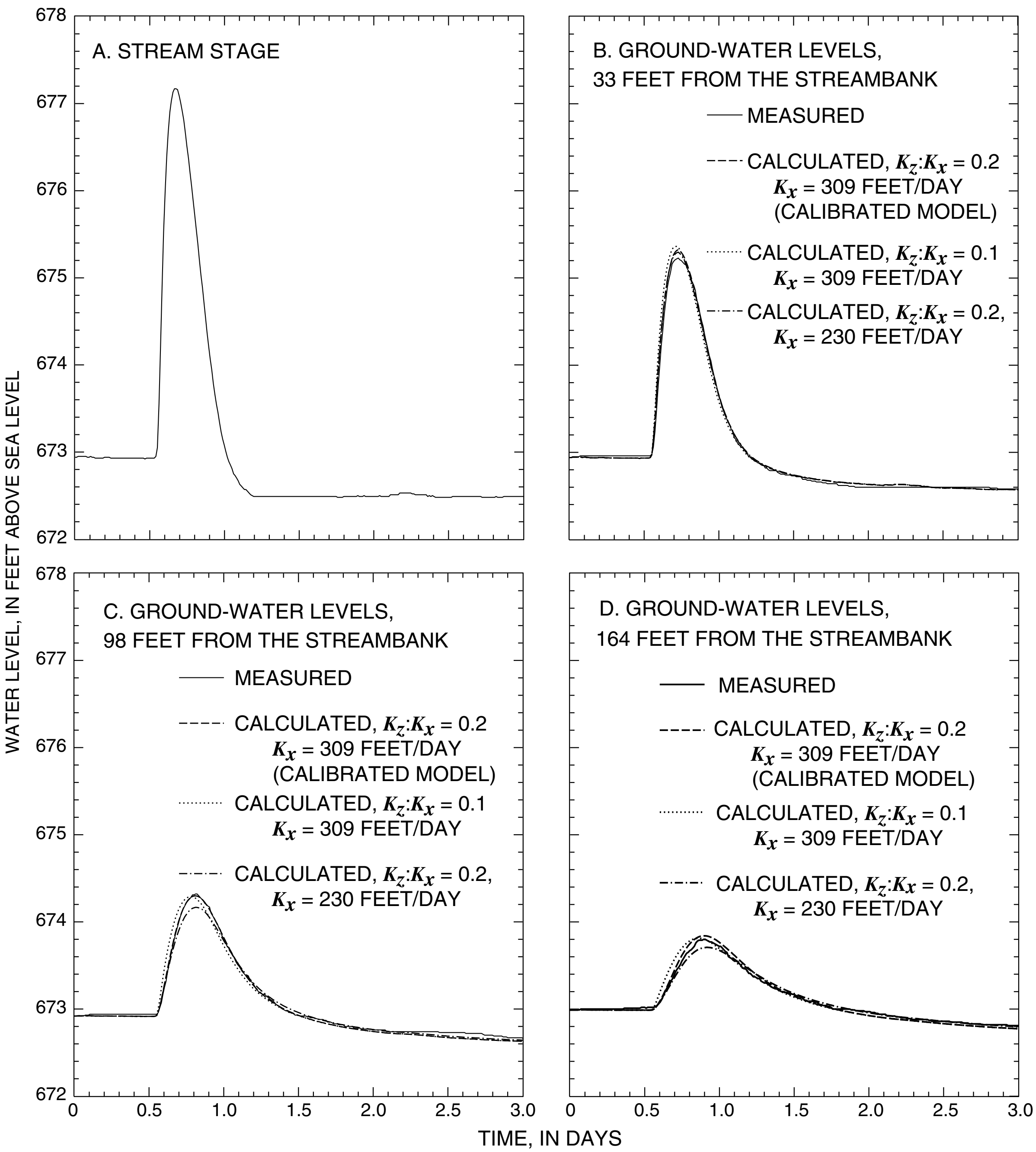


Figure 24. Stream stage and calculated and measured ground-water levels in observation wells located at three distances from the streambank in the Cedar River alluvial aquifer near Cedar Rapids, Iowa, for a 1-day stream-stage fluctuation. (*A*) Stream stage. (*B-D*) Ground-water levels at (*B*) 33 feet; (*C*) 98 feet; and (*D*) 164 feet from the streambank. (K_z, vertical hydraulic conductivity; K_x, horizontal hydraulic conductivity.)

The aquifer is simulated as a finite-width, water-table aquifer bounded by semipervious streambank material at the stream-aquifer boundary (fig. 25). 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 and others, 1996). A horizontal hydraulic conductivity of 309 ft/d (see table 2) was used for the simulated (homogeneous) 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; these units compose 70 percent of the simulated section of aquifer (fig. 23). Fine-grained sediments above the water table and near the upland end of the section, as well as the underlying glacial till (fig. 23), were not simulated using STWT1.

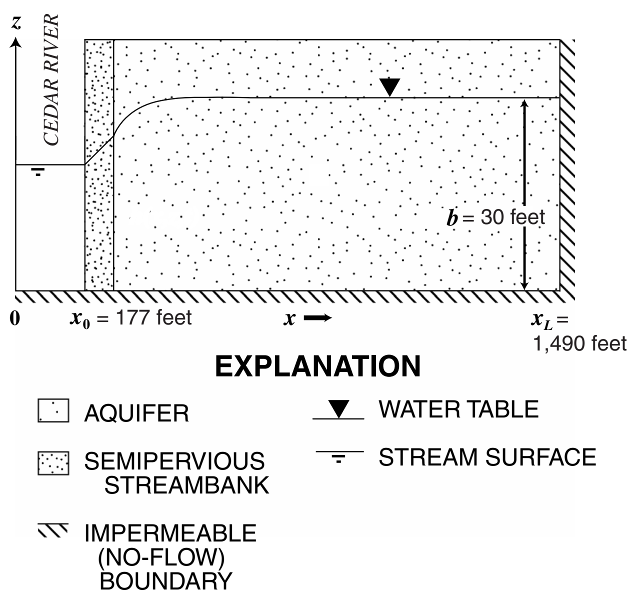


Figure 25. Conceptual model of the Cedar River alluvial aquifer near Cedar Rapids, Iowa, used for simulation with computer program STWT1. (*b*, saturated thickness of the aquifer; x_0, distance from the stream center to the stream-aquifer boundary; x_L, distance from stream center to lateral aquifer boundary.)

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. 24*B*–*D*). The best-fit calculated ground-water levels were obtained using aquifer and streambank properties (table 2) that were only slightly different from those used in the calibrated numerical model. A lower value of horizontal hydraulic conductivity (230 ft/d; K_D was held constant) that is closer to the value for the fine-grained sediments at the upland end of the section also was tested with the STWT1 program but did not yield as good a fit to measured values as did the value for the medium- and coarse-grained-sediment units (309 ft/d) (fig. 24*B*–*D*). The calibrated analytical stream-aquifer model differed from the numerical model in that the ratio of vertical to horizontal hydraulic conductivity (K_D) was higher in the stream-aquifer model (0.2) than in the numerical model (0.1). When the value of K_D equal to 0.1 was tested in the analysis with STWT1, calculated ground-water levels increased slightly more rapidly than the measured water levels in all three wells (fig. 24*B*–*D*). The higher K_D value of 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. Other properties of the aquifer and streambank, including specific yield, specific storage, and streambank (or riverbottom) thickness and vertical hydraulic conductivity, were the same as used in the calibrated numerical model (Squillace and others, 1996). The close agreement between measured ground-water levels and ground-water levels calculated with the STWT1 program and between the analytical and numerical-model parameters (the numerical model simulated the stream as partially penetrating) 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.

The 1-day, 4-foot rise in stream stage resulted in a model-estimated maximum seepage rate of 368 ft^3/d per foot of stream reach and a total bank-storage volume of 62 ft^3 per foot of stream reach along the length of the streambank. Seepage rate into the aquifer and bank storage peaked at 0.04 days before and 0.18 days after, respectively, the peak of the stream flood wave. All bank-storage water that entered the aquifer in response to the flood wave appears to have returned to the river about 1.5 days after the stream-stage peak; water continued to drain from the aquifer, however, because the stream stage continued to decrease below its initial level. Seepage rates and bank storage were only slightly less than those given above in the simulations where K_D was equal to the numerical-model value of 0.1, instead of the best-fit value of 0.2; a decrease in K_x (from 309 to 230 ft/d; K_D was held constant) had a similar, though slightly greater, effect than the decrease in K_D.

Analysis of Response of Stream-Aquifer System to a Simultaneous 55-Day Stream-Stage Fluctuation and Recharge

The aquifer and streambank hydraulic properties estimated from simulation of the 1-day stream-stage fluctuation in October–November 1989 were used to simulate the response of the alluvial aquifer to simultaneous stream-stage fluctuations and recharge during a 55-day period of precipitation and surface runoff in March and April 1990 (fig. 26). Daily measurements of stream stage, which rose a maximum of about 6 ft during the period, and of ground-water levels were available for the site (Schulmeyer and others, 1995). Also, seepage rates and bank-storage volumes calculated by the numerical model of Squillace and others (1996) were available for comparison with those calculated by STWT1.

An initial recharge rate of 0.0034 ft/d (ground-water level increase of 0.017 ft/d) was tested on the basis of results of the numerical model of the site (fig. 26). However, the closest match between measured and calculated ground-water levels at the three observation wells was obtained for a recharge rate of 0.0020 ft/d (ground-water level increase of 0.010 ft/d; fig. 27). The latter value may be more representative of actual recharge, which was determined to be about one-half of the model-calibrated value used by Squillace and others (1996) for a nearby alluvial aquifer (Hansen and Steinhilber, 1977). Ground-water levels were underestimated by STWT1 when only the stream-stage fluctuation, but not recharge, was simulated. Thus, simulation of recharge improved the match between measured and calculated ground-water-level fluctuations. (The data-input and result files provided in the appendix are for the simulation with a recharge rate of 0.0020 ft/d.)

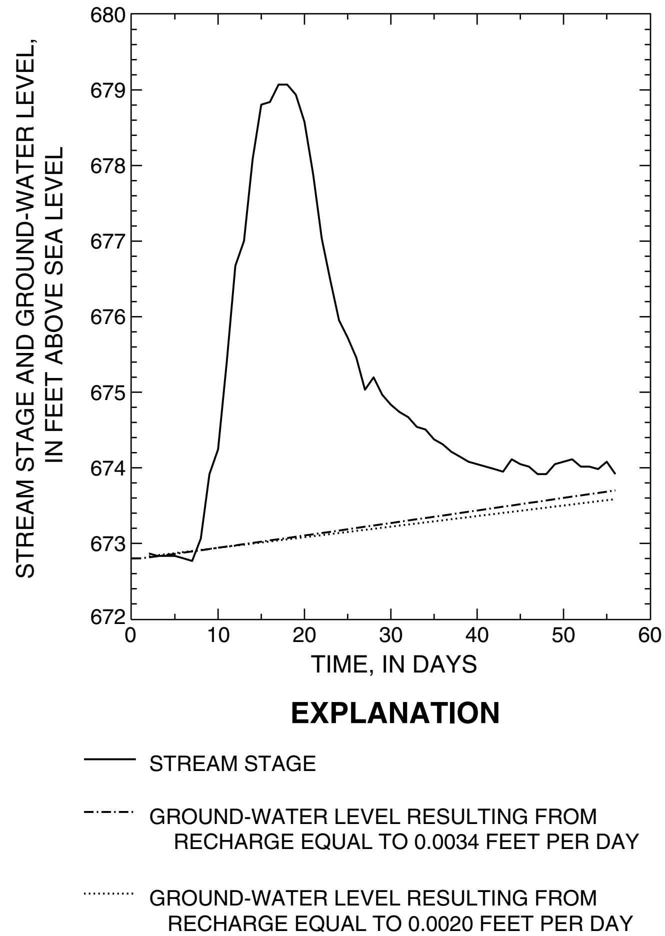


Figure 26. Stream stage and ground-water levels resulting from recharge at the Cedar River alluvial aquifer site near Cedar Rapids, Iowa, for a 55-day stream-stage fluctuation.

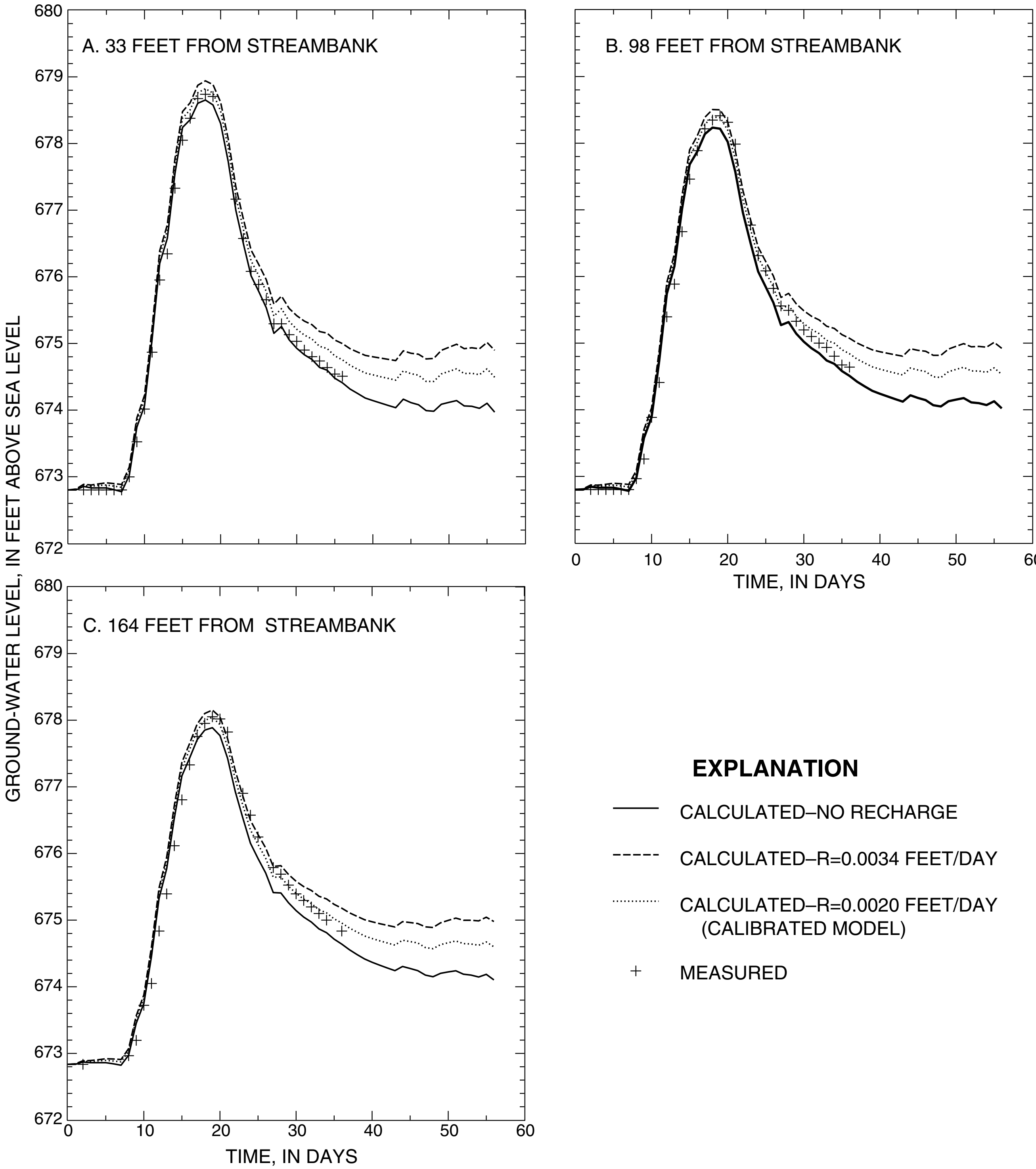


Figure 27. Calculated and measured ground-water levels in observation wells located three distances from the streambank in the Cedar River alluvial aquifer near Cedar Rapids, Iowa, for a simultaneous 55-day stream-stage fluctuation and recharge. (*A*) 33 feet from the streambank. (*B*) 98 feet from the streambank. (*C*) 164 feet from the streambank. (R, recharge rate.)

The simulated recharge resulted in increased hydraulic gradients toward the river compared to those for the condition of no recharge. Consequently, seepage rates to the river following the flood wave were greater (fig. 28*A*) and the total volume of bank storage in the aquifer was lower (fig. 28*B*) than without recharge. These simulations demonstrate that recharge can have a substantial effect on the total accumulation and the timing of release of bank-storage water. Seepage rates and bank-storage volumes calculated with STWT1 (for a recharge rate of 0.0020 ft/d) agree well with those estimated with the transient numerical model of the site. The maximum bank-storage volume calculated with program STWT1, 689 ft^3 per foot of stream reach (fig. 28*B*), is only about 10 percent greater than the maximum bank storage volume estimated with the numerical model, 624 ft^3 per foot of stream reach (Squillace and others, 1996); although it should be noted that a lower recharge rate was used in the analytical model than in the numerical model. Maximum seepage rate through the streambank was estimated by program STWT1 as 83 ft^3/d per foot of stream reach (fig. 28*A*). This value is about 50 percent greater than the maximum rate of seepage, 57 ft^3/d per foot of stream reach, estimated by the numerical model flowing through the bottom of the simulated, partially penetrating river.

Analysis of the Cedar River site demonstrates that, although several simplifying assumptions about aquifer hydraulic properties and geometry are required to simulate the stream-aquifer system, a model can be developed using the analytical solutions and computer programs that compares well with a much more complex numerical model with respect to estimated hydraulic properties and the simulated aquifer response to stream-stage stresses. Although the analytical methods cannot incorporate all the complexities of real stream-aquifer systems, they can provide useful results for a number of applications, such as the use of natural stresses to estimate hydraulic properties, preliminary estimates of hydraulic properties for numerical modeling, and the testing of hypotheses relating to stream-aquifer hydraulic interaction. For some purposes, the simplifying assumptions required for the analytical methods may be limiting. For example, an effective horizontal hydraulic conductivity that integrates the permeability of all the sediments in the modeled section is specified. This integrated value may not represent the actual heterogeneity of the aquifer adequately enough for the analysis of contaminant transport. The assumption of an impermeable lower boundary to the aquifer also limits the capability of the model to quantify flow to an underlying, low-permeability unit such as the glacial till at the Cedar River site. Analysis of the Cedar River site also illustrates the importance of accurate recharge rates, which are not necessarily easy or straightforward to determine, for the calculation of seepage rates and bank storage associated with a flood wave; accurate recharge data are necessary for the analysis of stream-aquifer hydraulic interaction by either analytical or numerical methods.

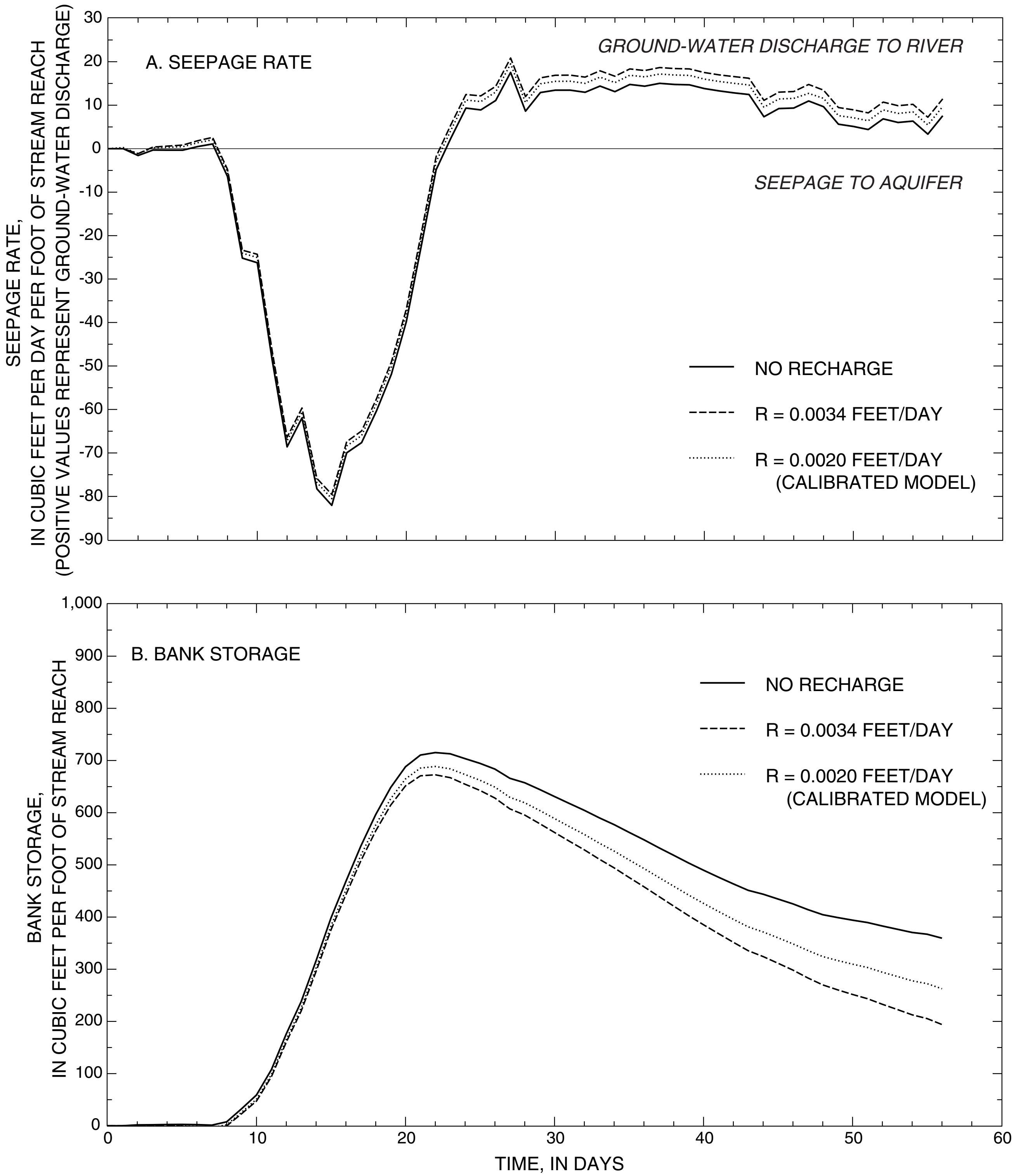


Figure 28. Calculated seepage rate and bank storage in the Cedar River alluvial aquifer near Cedar Rapids, Iowa, in response to a simultaneous 55-day stream-stage fluctuation and recharge. (*A*) Seepage rate. (*B*) Bank storage. (R, recharge rate.)

SUMMARY

The hydraulic interaction of aquifers and streams is important for water-quality and water-quantity concerns in many hydrogeologic settings. Analytical solutions to the ground-water flow equation are methods of quantifying stream-aquifer interaction that are simple, versatile, and easy to use. This report demonstrates the use of two computer programs (STLK1 and STWT1) that are based on analytical solutions to the ground-water-flow equation for stream-aquifer systems. The programs can be used to analyze the hydraulic response of confined, leaky, and water-table aquifers to stream-stage fluctuations and to ground-water recharge or evapotranspiration.

Analysis of idealized, hypothetical stream-aquifer systems is used to show how aquifer type, boundaries, and hydraulic properties, and the inclusion of recharge, affect aquifer response to stresses. In general, ground-water levels respond less quickly and with lower amplitude in aquifer types with a water-table boundary (water-table aquifer and leaky aquifer overlain by a water-table aquitard) than in aquifer types with an impermeable upper boundary (confined aquifer or leaky aquifer with an impermeable layer over the aquitard). However, seepage rates and bank storage volumes generally are much greater for the aquifers bounded by a water table than for the other aquifer types. This reflects the greater storage capacity (because of specific yield) represented by movement of the water table as compared with storage available from the compressibility of aquifer materials and pore water. Changes in ground-water levels are attenuated more rapidly as distance increases from the streambank in the water-table-bounded aquifers than in the confined or other types of leaky aquifers. Aquifer hydraulic properties that affect the modeled aquifer response to stream-stage fluctuations include hydraulic conductivity, specific storage, specific yield, and streambank leakance. These responses also are affected by the hydraulic properties of an aquitard (if present). The presence of semipervious streambank material dampens the response of the aquifer to a stream-stage fluctuation, such that increased streambank thickness or decreased streambed hydraulic conductivity, results in lower ground-water levels, seepage rates, and bank-storage volumes. Finally, when recharge and a stream-stage fluctuation are simulated simultaneously, the flow rates at the stream-aquifer boundary from these two separate stresses are superimposed. Seepage rates from the stream to the aquifer and the bank-storage volume resulting from the flood wave are less than would result in the absence of recharge, and bank storage drains more quickly.

Published data from alluvial and stratified-drift aquifers in Kentucky, Massachusetts, and Iowa are used to demonstrate application of the programs to field settings. Analytical models of the three stream-aquifer systems are developed as follows: the available hydrogeologic information is used to develop a conceptual model for each stream-aquifer system; stream-stage fluctuations and recharge are applied to the systems as hydraulic stresses; and the models are calibrated by matching ground-water levels calculated with computer program STLK1 or STWT1 to measured ground-water levels. In all three cases, excellent matches were obtained between measured and calculated water levels using specifications of aquifer type and hydraulic properties of the aquifer, aquitard, and streambank that were consistent with the available data.

Three alternative models of the stream-aquifer system at the Tennessee River site in western Kentucky yielded calculated ground-water levels that agreed well with the measured values. The aquifer was modeled as (1) a leaky aquifer with a water-table aquitard, (2) a leaky aquifer with an impermeable layer over the aquitard, and (3) a confined aquifer. Hydraulic properties among the three models differed only in the value of the streambed-leakance term. Because the streambed leakance term often is difficult to measure (and thus generally does not provide a good basis for distinguishing among models), identifying the stream-aquifer model that best describes the real system is important and requires independent knowledge of hydrologic conditions in the aquifer. This is perhaps best illustrated by the analysis of the Tennessee River data, in which calculated seepage rates and bank-storage volumes associated with the simulated flood wave differed widely among the three models.

Analysis of the stratified-drift aquifer adjacent to the Blackstone River in central Massachusetts suggested that ground-water levels in the aquifer responded to rapid, daily stream-stage fluctuations as if

the aquifer were under confined conditions. The available lithologic information, however, indicated that water-table conditions were present in the aquifer. The response of the aquifer to the stream-stage stress may have resembled that of a confined aquifer because of the thin unsaturated zone at the site. Seepage rates and bank-storage volumes calculated with the best-fit, confined stream-aquifer model were low.

In the analysis of the alluvial aquifer adjacent to the Cedar River in eastern Iowa, hydraulic properties used in the best-fit analytical stream-aquifer model were similar to those used in a calibrated two-dimensional numerical model of the site by Squillace and others (1996). In the numerical model, the river was simulated as partially penetrating and the aquifer as heterogeneous. The close agreement in results with respect to hydraulic properties and the similarity of calculated seepage rates and bank-storage volumes obtained with STWT1 and with the numerical model demonstrates that a stream-aquifer analytical model can be developed that compares well with a more complex numerical model. This results in spite of the need for simplifying assumptions about aquifer hydraulic properties and geometry in order to simulate the system with the analytical model. Analysis of the aquifer at the Cedar River site also showed that simulation of recharge can improve an already good match between measured and calculated ground-water-level fluctuations, and that recharge can have a substantial effect on the total accumulation and the timing of the release of bank-storage water.

In summary, the computer programs STLK1 and STWT1 were applied to a variety of hypothetical and field settings. Results illustrate the effects of aquifer type, boundaries, and hydraulic properties on the hydraulic interaction of aquifers and streams and provide insight into hydrologic processes and important factors at three sites. Results were obtained that were comparable to those produced by numerical models. The programs make the analytical solutions available for easy application to field problems. Although the analytical methods and computer programs cannot incorporate all of the complexities of real systems, they can provide useful results for a number of applications.

REFERENCES CITED

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

Barlow, P.M., and Moench, A.F., 1998, Analytical solutions and computer programs for hydraulic interaction of stream-aquifer systems: U.S. Geological Survey Open-File Report 98-415A, 85 p.

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

Bouwer, H., and Rice, R.C., 1978, Delayed aquifer yield as a phenomenon of delayed air entry: Water Resources Research, v. 14, no. 6, p. 1068–1074.

_____1980, Reply to “Comment on ‘Delayed aquifer yield as a phenomenon of delayed air entry’ by Herman Bouwer and R.C. Rice” by Michael Price: Water Resources Research, v. 16, no. 1, p. 254.

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

Dames and Moore, 1988a, Report—Remedial investigation, B. F. Goodrich/Airco site, Calvert City, Kentucky, March 14, 1988: Cincinnati, Ohio, 85 p.

_____1988b, Report—Evaluation of aquifer response during flooding, B.F. Goodrich/Airco site, Calvert City, Kentucky, July 28, 1988: Cincinnati, Ohio, 16 p.

_____1991, Report, Volume 1—Remedial environmental investigation, groundwater assessment, B.F. Goodrich Company, Calvert City, Kentucky: Cincinnati, Ohio, October 29, 1991, 67 p.

Davis, R.W., Lambert, T.W., and Hansen, A.J., Jr., 1973, Subsurface geology and ground-water resources of the Jackson Purchase Region, Kentucky: U.S. Geological Survey Water-Supply Paper 1987, 66 p.

DeSimone, L.A., and 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: U.S. Geological Survey Open-File Report 95-425, 19 p.

Dickerman, D.C., and Barlow, P.M., 1997, Water-table conditions and stream-aquifer interaction in the Hunt–Annaquatucket–Pettaquamscutt aquifer, central Rhode Island, October 7–9, 1996: U.S. Geological Survey Water-Resources Investigations Report 97-4167, 1 plate.

Dumouchelle, D.H., Schalk, C.W., Rowe, G.L., and de Roche, J. T., 1993, Hydrogeology, simulated ground-water flow, and ground-water quality, Wright–Patterson Air Force Base, Ohio: U.S. Geological Survey Water-Resources Investigations Report 93-4047, 152 p.

Gallaher, J.T., and Price, W.E., Jr., 1966, Hydrology of the alluvial deposits in the Ohio River Valley in Kentucky: U.S. Geological Survey Water-Supply Paper 1818, 80 p.

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

Grubb, H.F., and Zehner, H.H., 1973, Aquifer diffusivity of the Ohio River alluvial aquifer by the flood-wave response method: Journal of Research of the U.S. Geological Survey, v. 1, no. 5, p. 597–601.

Hall, F.R., 1968, Base-flow recessions—a review: Water Resources Research, v. 4, no. 5, p. 973–983.

Hall, F.R., and Moench, A.F., 1972, Application of the convolution equation to stream-aquifer relationships: Water Resources Research, v. 8, no. 2, p. 487–493.

Hansen, R.E., and Steinhilber, W.L., 1977, Geohydrology of Muscatine Island, Muscatine County, Iowa: Iowa Geological Survey Water Supply Bulletin No. 11, 60 p.

Hantush, M.S., 1964, Hydraulics of wells, *in* Chow, V.T., ed., Advances in Hydroscience, volume 1—1964: New York, Academic Press, p. 281–432.

_____1965, Wells near streams with semipervious beds: Journal of Geophysical Research, v. 70, no. 12, p. 2829–2838.

Heath, R.C., 1984, Ground-water regions of the United States: U.S. Geological Survey Water-Supply Paper 2242, 78 p.

_____1988, Hydrogeologic setting of regions, *in* Back, William, Rosenshein, J.S., and Seaber, P.R., eds., Hydrogeology: Boulder, Colo., Geological Society of America, The Geology of North America, v. O-2, p. 15–23.

HMM Associates, Inc., 1993, Site assessment and remediation support services contract, investigation of ground-water contamination, final report, Kaltsas/Omni-Duralite site, South Grafton, Massachusetts: Concord Mass., Site No. 2-0206, TAD No. 2021, March 1993, 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: Concord Mass., Site No. 2-0206, TAD No. 2021, February 1994,variously paged.

Lee, D.R., 1977, A device for measuring seepage flux in lakes and estuaries: Limnology and Oceanography, v. 22, no. 1, p. 140–147.

May, J.E., Gorman, J.G., Goodrich, R.D., and Miller, V.E., 1998, Water resources data, Iowa, water year 1997, volume 1: surface water--Mississippi River Basin: U.S. Geological Survey Water-Data Report IA-97-1, 395 p.

Moench, A.F., and Kisiel, C.C., 1970, Application of the convolution relation to estimating recharge from an ephemeral stream: Water Resources Research, v. 6, no. 4, p. 1087–1094.

Moench, A.F., Sauer, V.B., and Jennings, M.E., 1974, Modification of routed streamflow by channel loss and base flow: Water Resources Research, v. 10, no. 5, p. 963–968.

Narasimhan, T.N., and Zhu, Ming, 1993, Transient flow of water to a well in an unconfined aquifer—applicability of some conceptual models: Water Resources Research, v. 29, no. 1, p. 179–191.

Neuman, S.P., 1974, Effect of partial penetration on flow in unconfined aquifers considering delayed gravity response: Water Resources Research, v. 10, no. 2, p. 303–312.

Perkins, S.P., and Koussis, A.D., 1996, Stream-aquifer interaction model with diffusive wave routing: Journal of Hydraulic Engineering, v. 122, no. 4, p. 210–218.

Pinder, G.F., Bredehoeft, J.D., and Cooper, H.H., Jr., 1969, Determination of aquifer diffusivity from aquifer response to fluctuations in river stage: Water Resources Research, v. 5, no. 4, p. 850–855.

Pinder, G.F., and Sauer, S.P., 1971, Numerical simulation of flood wave modification due to bank storage effects: Water Resources Research, v. 7, no. 1, p. 63–70.

Prince, K.R., Reilly, T.E., and Franke, O.L., 1989, Analysis of the shallow groundwater flow system near Connetquot Brook, Long Island, New York: Journal of Hydrology, v. 107, p. 223–250.

Reynolds, R.J., 1987, Diffusivity of a glacial-outwash aquifer by the floodwave-response technique: Ground Water, v. 25, no. 3, p. 290–299.

Rosenshein, J.S., 1988, Region 18, Alluvial valleys, *in* Back, William, Rosenshein, J.S., and Seaber, P.R., eds., Hydrogeology: Boulder, Colo., Geological Society of America, The Geology of North America, v. O-2, p. 165–175.

Rutledge, A.T., 1993, Computer programs for describing the recession of ground-water discharge and for estimating mean ground-water recharge and discharge from streamflow records: U.S. Geological Survey Water-Resources Investigations Report 93-4121, 45 p.

Schulmeyer, P.M., Barnes, K.K., and Squillace, P.J., 1995, Hydrologic data from the lower Cedar River Basin, Iowa, 1989–91: U.S. Geological Survey Open-File Report 94-711, 116 p.

Sharp, J.M., Jr., 1977, Limitations of bank-storage model assumptions: Journal of Hydrology, v. 35, p. 31–47.

_____ 1988, Alluvial aquifers along major rivers, *in* Back, William, Rosenshein, J.S., and Seaber, P.R., eds., Hydrogeology: Boulder, Colorado, Geological Society of America, The Geology of North America, v. O-2, p. 273–282.

Singh, K.P., 1969, Theoretical baseflow curves: American Society of Civil Engineers, Journal of the Hydraulics Division, v. 95, no. HY 6, p. 2029–2048.

Sophocleous, Marios, and Perkins, S.P., 1993, Calibrated models as management tools for stream-aquifer systems--the case of central Kansas, USA: Journal of Hydrology, v. 152, p. 31–56.

Sophocleous, Marios, Townsend, M.A., Vogler, L.D., McClain, T.J., Marks, E.T., and Coble, G.R., 1988, Experimental studies in stream-aquifer interaction along the Arkansas River in central Kansas--field testing and analysis: Journal of Hydrology, v. 98, p. 249–273.

Squillace, P.J., 1996, Observed and simulated movement of bank-storage water: Ground Water, v. 34, no. 1, p. 121–134.

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

Squillace, P.J., Thurman, E.M., and Furlong, E.T., 1993, Groundwater as a nonpoint source of atrazine and deethylatrazine in a river during base flow conditions: Water Resources Research, v. 29, no. 6, p. 1719–1729.

Starn, J.J., Arihood, L.D., and Rose, M.F., 1995, Geohydrology and simulation of ground-water flow in the aquifer system near Calvert City, Kentucky: U.S. Geological Survey Water-Resources Investigations Report 94-4239, 52 p.

Tallaksen, L.M., 1995, A review of baseflow recession analysis: Journal of Hydrology, v. 165, p. 349–370.

Walker, E.H., 1957, The deep channel and alluvial deposits of the Ohio Valley in Kentucky: U.S. Geological Survey Water-Supply Paper 1411, 25 p.

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

Whiting, P.J., and Pomeranets, Mikhail, 1997, A numerical study of bank storage and its contribution to streamflow: Journal of Hydrology, v. 202, p. 121–136.

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

_____1990, Zone II delineation and water quality assessment at proposed well site 7-81, South Grafton, Massachusetts, February 1990: Wellesley, Mass., variously paginated.

Winter, T.C., Harvey, J.W., Franke, O.L., and Alley, W.M., 1998, Ground water and surface water--a single resource: U.S. Geological Survey Circular 1139, 79 p.

Yost, W.P., 1995, Data on ground-water levels and ground-water/surface-water relations in the Great Miami River and Little Miami River valleys, Southwestern Ohio: U.S. Geological Survey Open-File Report 95-357, 57 p.

APPENDIX—Input And Output Files For Selected Simulations

Input file for program STLK1, Tennessee River study site, stream-stage fluctuation only.

Output file from program STLK1, Tennessee River study site, stream-stage fluctuation only.

Input file for program STWT1, Cedar River study site, simultaneous stream-stage fluctuation and recharge.

Output file from program STWT1, Cedar River study site, simultaneous stream-stage fluctuation and recharge.

Input file for program STLK1, Tennessee River study site, stream-stage fluctuation only (see Barlow and Moench, 1998, for explanation of input variables and data formats).

```
Tennessee River alluvial aquifer, Starn and others (1995)
Leaky aquifer/water-table aquitard. Well is 125 ft from streambank.
    0    2.5D-1  0
    0      3     0
 6.6D2   0.0D0   0.0D0  1.0D3
 3.0D2   1.0D-5  3.0D1
 5.0D-1  1.0D-4  3.0D1  7.0D-2
 7.85D2  3.032D2   0.0D0
    8
   151
0.00     3.02234D2     0.000
0.25     3.01141D2     0.000
0.50     3.01184D2     0.000
0.75     3.02905D2     0.000
1.00     3.03485D2     0.000
1.25     3.03404D2     0.000
1.50     3.02474D2     0.000
1.75     3.03169D2     0.000
2.00     3.02762D2     0.000
2.25     3.02849D2     0.000
2.50     3.02848D2     0.000
2.75     3.03810D2     0.000
3.00     3.04582D2     0.000
3.25     3.02402D2     0.000
3.50     3.04275D2     0.000
3.75     3.08342D2     0.000
4.00     3.14806D2     0.000
4.25     3.17464D2     0.000
4.50     3.18214D2     0.000
4.75     3.18964D2     0.000
5.00     3.19737D2     0.000
5.25     3.20494D2     0.000
5.50     3.21195D2     0.000
5.75     3.21649D2     0.000
6.00     3.21772D2     0.000
6.25     3.21920D2     0.000
6.50     3.22102D2     0.000
6.75     3.21673D2     0.000
7.00     3.20469D2     0.000
7.25     3.19948D2     0.000
7.50     3.19663D2     0.000
7.75     3.19628D2     0.000
8.00     3.19462D2     0.000
8.25     3.19438D2     0.000
8.50     3.19455D2     0.000
8.75     3.19580D2     0.000
9.00     3.19621D2     0.000
9.25     3.19527D2     0.000
9.50     3.19506D2     0.000
9.75     3.19346D2     0.000
```

10.00	3.19283D2	0.000
10.25	3.18965D2	0.000
10.50	3.18796D2	0.000
10.75	3.18536D2	0.000
11.00	3.17999D2	0.000
11.25	3.17390D2	0.000
11.50	3.16818D2	0.000
11.75	3.15793D2	0.000
12.00	3.13335D2	0.000
12.25	3.11797D2	0.000
12.50	3.10532D2	0.000
12.75	3.09650D2	0.000
13.00	3.08714D2	0.000
13.25	3.07921D2	0.000
13.50	3.06490D2	0.000
13.75	3.05380D2	0.000
14.00	3.04498D2	0.000
14.25	3.03798D2	0.000
14.50	3.02629D2	0.000
14.75	3.03133D2	0.000
15.00	3.03582D2	0.000
15.25	3.03610D2	0.000
15.50	3.03678D2	0.000
15.75	3.03936D2	0.000
16.00	3.04162D2	0.000
16.25	3.04876D2	0.000
16.50	3.06115D2	0.000
16.75	3.08057D2	0.000
17.00	3.10909D2	0.000
17.25	3.12093D2	0.000
17.50	3.12912D2	0.000
17.75	3.13815D2	0.000
18.00	3.14592D2	0.000
18.25	3.15335D2	0.000
18.50	3.15434D2	0.000
18.75	3.15029D2	0.000
19.00	3.15239D2	0.000
19.25	3.15591D2	0.000
19.50	3.15907D2	0.000
19.75	3.16295D2	0.000
20.00	3.16655D2	0.000
20.25	3.16868D2	0.000
20.50	3.17161D2	0.000
20.75	3.17370D2	0.000
21.00	3.17586D2	0.000
21.25	3.17746D2	0.000
21.50	3.17915D2	0.000
21.75	3.18137D2	0.000
22.00	3.18365D2	0.000
22.25	3.18522D2	0.000
22.50	3.18666D2	0.000
22.75	3.18809D2	0.000
23.00	3.18950D2	0.000
23.25	3.19076D2	0.000
23.50	3.19179D2	0.000

23.75	3.19253D2	0.000
24.00	3.19379D2	0.000
24.25	3.19434D2	0.000
24.50	3.19536D2	0.000
24.75	3.19607D2	0.000
25.00	3.19640D2	0.000
25.25	3.19677D2	0.000
25.50	3.19645D2	0.000
25.75	3.19514D2	0.000
26.00	3.19688D2	0.000
26.25	3.19465D2	0.000
26.50	3.18987D2	0.000
26.75	3.18792D2	0.000
27.00	3.18285D2	0.000
27.25	3.17680D2	0.000
27.50	3.17070D2	0.000
27.75	3.16731D2	0.000
28.00	3.15459D2	0.000
28.25	3.14258D2	0.000
28.50	3.13262D2	0.000
28.75	3.12405D2	0.000
29.00	3.11282D2	0.000
29.25	3.10344D2	0.000
29.50	3.10036D2	0.000
29.75	3.09583D2	0.000
30.00	3.08914D2	0.000
30.25	3.08266D2	0.000
30.50	3.07661D2	0.000
30.75	3.07368D2	0.000
31.00	3.07576D2	0.000
31.25	3.07653D2	0.000
31.50	3.07560D2	0.000
31.75	3.07636D2	0.000
32.00	3.07680D2	0.000
32.25	3.07705D2	0.000
32.50	3.07673D2	0.000
32.75	3.07405D2	0.000
33.00	3.07362D2	0.000
33.25	3.07263D2	0.000
33.50	3.07150D2	0.000
33.75	3.07195D2	0.000
34.00	3.07244D2	0.000
34.25	3.07138D2	0.000
34.50	3.07089D2	0.000
34.75	3.07123D2	0.000
35.00	3.07184D2	0.000
35.25	3.07055D2	0.000
35.50	3.06670D2	0.000
35.75	3.06777D2	0.000
36.00	3.05200D2	0.000
36.25	3.05319D2	0.000
36.50	3.06033D2	0.000
36.75	3.06430D2	0.000
37.00	3.06187D2	0.000
37.25	3.06023D2	0.000
37.50	3.05605D2	0.000

Output file from program STLK1, Tennessee River study site, stream-stage fluctuation only.

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 *         ****  U.S. GEOLOGICAL SURVEY  ****     *
 *                                                *
 *          **** STLK1: PROGRAM OUTPUT ****       *
 *                                                *
 * ONE-DIMENSIONAL MODEL OF STREAM-AQUIFER HYDRAULIC *
 *                                                *
 *    INTERACTION FOR CONFINED AND LEAKY AQUIFERS *
 *                                                *
 *       BOUNDED BY A FULLY PENETRATING STREAM    *
 *                                                *
 *           VERSION CURRENT AS OF 09/01/98       *
 *                                                *
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    Tennessee River alluvial aquifer, Starn and others (1995)
    Leaky aquifer/water-table aquitard. Well is 125 ft from streambank.


                   SUMMARY OF INPUT DATA
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  STRESS TYPE (ISTRESS):              0 (stream-stage fluctuations)
  TIME-STEP SIZE (DELT):              0.250D+00 (units of time)
  PRINTING CODE (IPRINT):             0 (stress data not printed)

 AQUIFER AND STREAMBANK CHARACTERISTICS (INPUT LINES 4 AND 5)
  AQUIFER EXTENT (IXL):               0 (semi infinite)
  AQUIFER TYPE (IAQ):                 3 (leaky, with water-table aquitard)
  STREAMBANK CODE (IXA):              0 (semipervious streambank absent)
  STREAM HALF WIDTH (XZERO):          0.660D+03 (units of length)
  LENGTH OF STREAM (XSTREAM):         0.100D+04 (units of length)

 AQUIFER PROPERTIES (INPUT LINE 6)
  HYDRAULIC CONDUCTIVITY (AK):        0.300D+03 (units of length per time)
  SPECIFIC STORAGE (AS):              0.100D-04 (units of inverse length)
  SATURATED THICKNESS (AB):           0.300D+02 (units of length)

 AQUITARD PROPERTIES (INPUT LINE 7)
  HYDRAULIC CONDUCTIVITY (AKT):       0.500D+00 (units of length per time)
  SPECIFIC STORAGE (AST):             0.100D-03 (units of inverse length)
  SATURATED THICKNESS (ABT):          0.300D+02 (units of length)
  SPECIFIC YIELD (ASYT):              0.700D-01 (dimensionless)

 OBSERVATION WELL DATA AND INITIAL CONDITIONS (INPUT LINE 8)
  DISTANCE TO OBSERVATION WELL (X):  0.785D+03 (units of length)
  INITIAL HEAD AT WELL (HINIT):      0.303D+03 (units of length)
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  START TIME OF SIMULATION (TINIT):  0.000D+00 (units of time)

PROGRAM SOLUTION VARIABLES (INPUT LINE 9)
  NUMBER OF STEHFEST TERMS (NS):                  8


                  SUMMARY OF STRESS DATA
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  NUMBER OF SPECIFIED TIME STEPS (NT):            151
  STRESS DATA NOT PRINTED


            DIMENSIONLESS PARAMETERS (CALCULATED BY PROGRAM)
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  DIMENSIONLESS DISTANCE TO WELL (XD):                      0.119D+01
  DIMENSIONLESS DISTANCE TO STREAMBANK (XZEROD):            0.220D+02
  DIMENSIONLESS WIDTH OF AQUIFER (XLLD):                     INFINITE
  DIMENSIONLESS STREAMBANK LEAKANCE (XAAD):                 0.000D+00
  DIMENSIONLESS RATIO OF AQUITARD TO AQUIFER
     STORATIVITY (SIGMA1):                                  0.100D+02
  DIMENSIONLESS RATIO OF AQUITARD TO AQUIFER HYDRAULIC
     CONDUCTIVITY (GAMMA1):                                 0.898D+00
  DIMENSIONLESS RATIO OF AQUIFER STORATIVITY
     TO AQUITARD SPECIFIC YIELD (SIGMAP):                   0.429D-02


                                RESULTS
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                                             TOTAL        BANK     BANK-STORAGE
     TIME         HEAD         SEEPAGE      SEEPAGE      STORAGE      VOLUME
      (T)          (L)        (L**2/T)     (L**3/T)      (L**2)      (L**3)
     ----         ----       ----------   ----------    -------      ------
 0.000000D+00  0.30320D+03  0.0000D+00  0.0000D+00  0.0000D+00  0.0000D+00
 0.250000D+00  0.30228D+03  0.1318D+02  0.2636D+05  -.3295D+01  -.6590D+04
 0.500000D+00  0.30231D+03  0.1226D+02  0.2453D+05  -.6361D+01  -.1272D+05
 0.750000D+00  0.30376D+03  -.8828D+01  -.1766D+05  -.4154D+01  -.8307D+04
 0.100000D+01  0.30425D+03  -.1552D+02  -.3104D+05  -.2735D+00  -.5471D+03
 0.125000D+01  0.30419D+03  -.1409D+02  -.2817D+05  0.3248D+01  0.6497D+04
 0.150000D+01  0.30341D+03  -.2479D+01  -.4959D+04  0.3868D+01  0.7736D+04
 0.175000D+01  0.30400D+03  -.1082D+02  -.2164D+05  0.6573D+01  0.1315D+05
 0.200000D+01  0.30366D+03  -.5594D+01  -.1119D+05  0.7971D+01  0.1594D+05
 0.225000D+01  0.30373D+03  -.6506D+01  -.1301D+05  0.9598D+01  0.1920D+05
 0.250000D+01  0.30373D+03  -.6321D+01  -.1264D+05  0.1118D+02  0.2236D+05
 0.275000D+01  0.30455D+03  -.1776D+02  -.3551D+05  0.1562D+02  0.3124D+05
 0.300000D+01  0.30521D+03  -.2656D+02  -.5312D+05  0.2226D+02  0.4451D+05
 0.325000D+01  0.30337D+03  0.4713D+00  0.9426D+03  0.2214D+02  0.4428D+05
 0.350000D+01  0.30495D+03  -.2221D+02  -.4443D+05  0.2769D+02  0.5539D+05
 0.375000D+01  0.30840D+03  -.7061D+02  -.1412D+06  0.4535D+02  0.9069D+05
 0.400000D+01  0.31389D+03  -.1465D+03  -.2930D+06  0.8197D+02  0.1639D+06
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0.425000D+01	0.31619D+03	-.1743D+03	-.3487D+06	0.1256D+03	0.2511D+06
0.450000D+01	0.31688D+03	-.1785D+03	-.3571D+06	0.1702D+03	0.3404D+06
0.475000D+01	0.31757D+03	-.1827D+03	-.3654D+06	0.2159D+03	0.4317D+06
0.500000D+01	0.31828D+03	-.1871D+03	-.3742D+06	0.2627D+03	0.5253D+06
0.525000D+01	0.31898D+03	-.1913D+03	-.3826D+06	0.3105D+03	0.6209D+06
0.550000D+01	0.31964D+03	-.1947D+03	-.3894D+06	0.3591D+03	0.7183D+06
0.575000D+01	0.32008D+03	-.1951D+03	-.3902D+06	0.4079D+03	0.8159D+06
0.600000D+01	0.32024D+03	-.1916D+03	-.3832D+06	0.4558D+03	0.9117D+06
0.625000D+01	0.32043D+03	-.1886D+03	-.3771D+06	0.5030D+03	0.1006D+07
0.650000D+01	0.32064D+03	-.1861D+03	-.3722D+06	0.5495D+03	0.1099D+07
0.675000D+01	0.32033D+03	-.1763D+03	-.3527D+06	0.5936D+03	0.1187D+07
0.700000D+01	0.31936D+03	-.1576D+03	-.3152D+06	0.6330D+03	0.1266D+07
0.725000D+01	0.31897D+03	-.1477D+03	-.2955D+06	0.6699D+03	0.1340D+07
0.750000D+01	0.31876D+03	-.1410D+03	-.2820D+06	0.7052D+03	0.1410D+07
0.775000D+01	0.31877D+03	-.1375D+03	-.2750D+06	0.7395D+03	0.1479D+07
0.800000D+01	0.31867D+03	-.1325D+03	-.2650D+06	0.7727D+03	0.1545D+07
0.825000D+01	0.31868D+03	-.1294D+03	-.2589D+06	0.8050D+03	0.1610D+07
0.850000D+01	0.31873D+03	-.1270D+03	-.2540D+06	0.8368D+03	0.1674D+07
0.875000D+01	0.31887D+03	-.1259D+03	-.2518D+06	0.8682D+03	0.1736D+07
0.900000D+01	0.31893D+03	-.1239D+03	-.2478D+06	0.8992D+03	0.1798D+07
0.925000D+01	0.31888D+03	-.1203D+03	-.2406D+06	0.9293D+03	0.1859D+07
0.950000D+01	0.31889D+03	-.1178D+03	-.2355D+06	0.9587D+03	0.1917D+07
0.975000D+01	0.31879D+03	-.1136D+03	-.2272D+06	0.9871D+03	0.1974D+07
0.100000D+02	0.31876D+03	-.1108D+03	-.2215D+06	0.1015D+04	0.2030D+07
0.102500D+02	0.31852D+03	-.1050D+03	-.2099D+06	0.1041D+04	0.2082D+07
0.105000D+02	0.31840D+03	-.1011D+03	-.2022D+06	0.1066D+04	0.2133D+07
0.107500D+02	0.31820D+03	-.9631D+02	-.1926D+06	0.1090D+04	0.2181D+07
0.110000D+02	0.31776D+03	-.8830D+02	-.1766D+06	0.1112D+04	0.2225D+07
0.112500D+02	0.31726D+03	-.7967D+02	-.1593D+06	0.1132D+04	0.2265D+07
0.115000D+02	0.31679D+03	-.7174D+02	-.1435D+06	0.1150D+04	0.2301D+07
0.117500D+02	0.31594D+03	-.5857D+02	-.1171D+06	0.1165D+04	0.2330D+07
0.120000D+02	0.31387D+03	-.2851D+02	-.5701D+05	0.1172D+04	0.2344D+07
0.122500D+02	0.31256D+03	-.1041D+02	-.2082D+05	0.1175D+04	0.2349D+07
0.125000D+02	0.31148D+03	0.3925D+01	0.7850D+04	0.1174D+04	0.2347D+07
0.127500D+02	0.31072D+03	0.1327D+02	0.2654D+05	0.1170D+04	0.2341D+07
0.130000D+02	0.30991D+03	0.2304D+02	0.4608D+05	0.1165D+04	0.2329D+07
0.132500D+02	0.30922D+03	0.3084D+02	0.6167D+05	0.1157D+04	0.2314D+07
0.135000D+02	0.30799D+03	0.4615D+02	0.9230D+05	0.1145D+04	0.2291D+07
0.137500D+02	0.30703D+03	0.5718D+02	0.1144D+06	0.1131D+04	0.2262D+07
0.140000D+02	0.30625D+03	0.6520D+02	0.1304D+06	0.1115D+04	0.2230D+07
0.142500D+02	0.30563D+03	0.7085D+02	0.1417D+06	0.1097D+04	0.2194D+07
0.145000D+02	0.30460D+03	0.8204D+02	0.1641D+06	0.1077D+04	0.2153D+07
0.147500D+02	0.30499D+03	0.7279D+02	0.1456D+06	0.1058D+04	0.2117D+07
0.150000D+02	0.30534D+03	0.6455D+02	0.1291D+06	0.1042D+04	0.2085D+07
0.152500D+02	0.30533D+03	0.6167D+02	0.1233D+06	0.1027D+04	0.2054D+07
0.155000D+02	0.30536D+03	0.5841D+02	0.1168D+06	0.1012D+04	0.2024D+07
0.157500D+02	0.30556D+03	0.5300D+02	0.1060D+06	0.9990D+03	0.1998D+07
0.160000D+02	0.30572D+03	0.4816D+02	0.9633D+05	0.9870D+03	0.1974D+07
0.162500D+02	0.30630D+03	0.3762D+02	0.7524D+05	0.9776D+03	0.1955D+07
0.165000D+02	0.30733D+03	0.2108D+02	0.4216D+05	0.9723D+03	0.1945D+07
0.167500D+02	0.30896D+03	-.3425D+01	-.6850D+04	0.9731D+03	0.1946D+07
0.170000D+02	0.31137D+03	-.3818D+02	-.7636D+05	0.9827D+03	0.1965D+07
0.172500D+02	0.31238D+03	-.5181D+02	-.1036D+06	0.9956D+03	0.1991D+07
0.175000D+02	0.31308D+03	-.6070D+02	-.1214D+06	0.1011D+04	0.2022D+07

0.177500D+02	0.31386D+03	-.7036D+02	-.1407D+06	0.1028D+04	0.2057D+07
0.180000D+02	0.31453D+03	-.7825D+02	-.1565D+06	0.1048D+04	0.2096D+07
0.182500D+02	0.31518D+03	-.8552D+02	-.1710D+06	0.1069D+04	0.2139D+07
0.185000D+02	0.31529D+03	-.8485D+02	-.1697D+06	0.1091D+04	0.2181D+07
0.187500D+02	0.31497D+03	-.7815D+02	-.1563D+06	0.1110D+04	0.2220D+07
0.190000D+02	0.31516D+03	-.7909D+02	-.1582D+06	0.1130D+04	0.2260D+07
0.192500D+02	0.31548D+03	-.8171D+02	-.1634D+06	0.1150D+04	0.2301D+07
0.195000D+02	0.31577D+03	-.8384D+02	-.1677D+06	0.1171D+04	0.2342D+07
0.197500D+02	0.31612D+03	-.8680D+02	-.1736D+06	0.1193D+04	0.2386D+07
0.200000D+02	0.31644D+03	-.8936D+02	-.1787D+06	0.1215D+04	0.2431D+07
0.202500D+02	0.31664D+03	-.9009D+02	-.1802D+06	0.1238D+04	0.2476D+07
0.205000D+02	0.31691D+03	-.9179D+02	-.1836D+06	0.1261D+04	0.2522D+07
0.207500D+02	0.31711D+03	-.9245D+02	-.1849D+06	0.1284D+04	0.2568D+07
0.210000D+02	0.31732D+03	-.9320D+02	-.1864D+06	0.1307D+04	0.2614D+07
0.212500D+02	0.31748D+03	-.9327D+02	-.1865D+06	0.1330D+04	0.2661D+07
0.215000D+02	0.31764D+03	-.9347D+02	-.1869D+06	0.1354D+04	0.2708D+07
0.217500D+02	0.31785D+03	-.9432D+02	-.1886D+06	0.1377D+04	0.2755D+07
0.220000D+02	0.31807D+03	-.9524D+02	-.1905D+06	0.1401D+04	0.2802D+07
0.222500D+02	0.31822D+03	-.9529D+02	-.1906D+06	0.1425D+04	0.2850D+07
0.225000D+02	0.31837D+03	-.9521D+02	-.1904D+06	0.1449D+04	0.2898D+07
0.227500D+02	0.31851D+03	-.9514D+02	-.1903D+06	0.1473D+04	0.2945D+07
0.230000D+02	0.31865D+03	-.9507D+02	-.1901D+06	0.1496D+04	0.2993D+07
0.232500D+02	0.31878D+03	-.9483D+02	-.1897D+06	0.1520D+04	0.3040D+07
0.235000D+02	0.31889D+03	-.9434D+02	-.1887D+06	0.1544D+04	0.3087D+07
0.237500D+02	0.31897D+03	-.9354D+02	-.1871D+06	0.1567D+04	0.3134D+07
0.240000D+02	0.31910D+03	-.9340D+02	-.1868D+06	0.1590D+04	0.3181D+07
0.242500D+02	0.31916D+03	-.9242D+02	-.1848D+06	0.1614D+04	0.3227D+07
0.245000D+02	0.31927D+03	-.9206D+02	-.1841D+06	0.1637D+04	0.3273D+07
0.247500D+02	0.31935D+03	-.9134D+02	-.1827D+06	0.1659D+04	0.3319D+07
0.250000D+02	0.31939D+03	-.9021D+02	-.1804D+06	0.1682D+04	0.3364D+07
0.252500D+02	0.31944D+03	-.8916D+02	-.1783D+06	0.1704D+04	0.3408D+07
0.255000D+02	0.31943D+03	-.8733D+02	-.1747D+06	0.1726D+04	0.3452D+07
0.257500D+02	0.31934D+03	-.8437D+02	-.1687D+06	0.1747D+04	0.3494D+07
0.260000D+02	0.31950D+03	-.8518D+02	-.1704D+06	0.1768D+04	0.3537D+07
0.262500D+02	0.31933D+03	-.8119D+02	-.1624D+06	0.1789D+04	0.3578D+07
0.265000D+02	0.31894D+03	-.7425D+02	-.1485D+06	0.1807D+04	0.3615D+07
0.267500D+02	0.31879D+03	-.7094D+02	-.1419D+06	0.1825D+04	0.3650D+07
0.270000D+02	0.31837D+03	-.6395D+02	-.1279D+06	0.1841D+04	0.3682D+07
0.272500D+02	0.31787D+03	-.5600D+02	-.1120D+06	0.1855D+04	0.3710D+07
0.275000D+02	0.31736D+03	-.4820D+02	-.9641D+05	0.1867D+04	0.3734D+07
0.277500D+02	0.31708D+03	-.4390D+02	-.8780D+05	0.1878D+04	0.3756D+07
0.280000D+02	0.31600D+03	-.2845D+02	-.5690D+05	0.1885D+04	0.3770D+07
0.282500D+02	0.31498D+03	-.1431D+02	-.2862D+05	0.1889D+04	0.3778D+07
0.285000D+02	0.31413D+03	-.3029D+01	-.6057D+04	0.1890D+04	0.3779D+07
0.287500D+02	0.31340D+03	0.6284D+01	0.1257D+05	0.1888D+04	0.3776D+07
0.290000D+02	0.31243D+03	0.1857D+02	0.3713D+05	0.1883D+04	0.3767D+07
0.292500D+02	0.31162D+03	0.2829D+02	0.5658D+05	0.1876D+04	0.3752D+07
0.295000D+02	0.31134D+03	0.3018D+02	0.6035D+05	0.1869D+04	0.3737D+07
0.297500D+02	0.31094D+03	0.3380D+02	0.6761D+05	0.1860D+04	0.3720D+07
0.300000D+02	0.31035D+03	0.3996D+02	0.7993D+05	0.1850D+04	0.3701D+07
0.302500D+02	0.30978D+03	0.4572D+02	0.9145D+05	0.1839D+04	0.3678D+07
0.305000D+02	0.30924D+03	0.5084D+02	0.1017D+06	0.1826D+04	0.3652D+07
0.307500D+02	0.30897D+03	0.5208D+02	0.1042D+06	0.1813D+04	0.3626D+07
0.310000D+02	0.30912D+03	0.4729D+02	0.9458D+05	0.1801D+04	0.3603D+07

0.312500D+02	0.30916D+03	0.4427D+02	0.8853D+05	0.1790D+04	0.3580D+07
0.315000D+02	0.30905D+03	0.4341D+02	0.8681D+05	0.1779D+04	0.3559D+07
0.317500D+02	0.30910D+03	0.4056D+02	0.8112D+05	0.1769D+04	0.3538D+07
0.320000D+02	0.30911D+03	0.3822D+02	0.7643D+05	0.1760D+04	0.3519D+07
0.322500D+02	0.30911D+03	0.3619D+02	0.7239D+05	0.1751D+04	0.3501D+07
0.325000D+02	0.30907D+03	0.3494D+02	0.6988D+05	0.1742D+04	0.3484D+07
0.327500D+02	0.30882D+03	0.3660D+02	0.7319D+05	0.1733D+04	0.3465D+07
0.330000D+02	0.30877D+03	0.3551D+02	0.7102D+05	0.1724D+04	0.3448D+07
0.332500D+02	0.30866D+03	0.3516D+02	0.7031D+05	0.1715D+04	0.3430D+07
0.335000D+02	0.30855D+03	0.3501D+02	0.7001D+05	0.1706D+04	0.3413D+07
0.337500D+02	0.30857D+03	0.3298D+02	0.6596D+05	0.1698D+04	0.3396D+07
0.340000D+02	0.30859D+03	0.3098D+02	0.6197D+05	0.1690D+04	0.3381D+07
0.342500D+02	0.30849D+03	0.3094D+02	0.6187D+05	0.1683D+04	0.3365D+07
0.345000D+02	0.30843D+03	0.3022D+02	0.6044D+05	0.1675D+04	0.3350D+07
0.347500D+02	0.30845D+03	0.2854D+02	0.5708D+05	0.1668D+04	0.3336D+07
0.350000D+02	0.30848D+03	0.2660D+02	0.5321D+05	0.1661D+04	0.3322D+07
0.352500D+02	0.30836D+03	0.2703D+02	0.5406D+05	0.1654D+04	0.3309D+07
0.355000D+02	0.30802D+03	0.3054D+02	0.6108D+05	0.1647D+04	0.3294D+07
0.357500D+02	0.30810D+03	0.2803D+02	0.5607D+05	0.1640D+04	0.3280D+07
0.360000D+02	0.30675D+03	0.4593D+02	0.9186D+05	0.1628D+04	0.3257D+07
0.362500D+02	0.30683D+03	0.4284D+02	0.8568D+05	0.1618D+04	0.3235D+07
0.365000D+02	0.30742D+03	0.3273D+02	0.6547D+05	0.1609D+04	0.3219D+07
0.367500D+02	0.30774D+03	0.2677D+02	0.5353D+05	0.1603D+04	0.3206D+07
0.370000D+02	0.30752D+03	0.2869D+02	0.5738D+05	0.1596D+04	0.3191D+07
0.372500D+02	0.30737D+03	0.2960D+02	0.5921D+05	0.1588D+04	0.3176D+07
0.375000D+02	0.30700D+03	0.3357D+02	0.6714D+05	0.1580D+04	0.3160D+07

Input file for program STWT1, Cedar River study site, simultaneous stream-stage fluctuation and recharge (see Barlow and Moench, 1998, for explanation of input variables and data formats).

```
Cedar River alluvial aquifer. Schulmeyer and others (1995)
Water-table aquifer. Mar-Apr '90 flood event. Well is 33 ft from streambank.
    2    1.0D0   0
    1      1     1
 1.77D2   1.49D3   3.09D1  1.0D3
 3.09D2   2.0D-1   3.0D-5  0.2D0   3.0D1
 2.10D2     0      11.6D0  14.1D0  0.0D0
 72.802D0  0.0D0
    8     1.0D-10  2.0D1
   57
0           72.802      0.000
1           72.802      0.010
2           72.867      0.020
3           72.835      0.030
4           72.835      0.040
5           72.835      0.050
6           72.802      0.060
7           72.769      0.070
8           73.064      0.080
9           73.917      0.090
10          74.245      0.100
11          75.394      0.110
12          76.673      0.120
13          77.002      0.130
14          78.084      0.140
15          78.806      0.150
16          78.839      0.160
17          79.069      0.170
18          79.069      0.180
19          78.937      0.190
20          78.576      0.200
21          77.887      0.210
22          77.034      0.220
23          76.477      0.230
24          75.952      0.240
25          75.722      0.250
26          75.459      0.260
27          75.033      0.270
28          75.197      0.280
29          74.967      0.290
30          74.836      0.300
31          74.738      0.310
32          74.672      0.320
33          74.541      0.330
34          74.508      0.340
35          74.377      0.350
36          74.311      0.360
37          74.213      0.370
38          74.147      0.380
```

39	74.081	0.390
40	74.049	0.400
41	74.016	0.410
42	73.983	0.420
43	73.95	0.430
44	74.114	0.440
45	74.049	0.450
46	74.016	0.460
47	73.917	0.450
48	73.917	0.460
49	74.049	0.470
50	74.081	0.480
51	74.114	0.490
52	74.016	0.500
53	74.016	0.510
54	73.983	0.520
55	74.081	0.530
56	73.917	0.540

Output file for program STWT1, Cedar River study site, simultaneous stream-stage fluctuation and recharge.

```
***************************************************
*                                                 *
*         ****  U.S. GEOLOGICAL SURVEY  ****      *
*                                                 *
*          **** STWT1: PROGRAM OUTPUT ****        *
*                                                 *
* TWO-DIMENSIONAL MODEL OF STREAM-AQUIFER HYDRAULIC *
*                                                 *
* INTERACTION FOR CONFINED AND WATER-TABLE AQUIFERS *
*                                                 *
*       BOUNDED BY A FULLY PENETRATING STREAM     *
*                                                 *
*           VERSION CURRENT AS OF 09/01/98        *
*                                                 *
***************************************************


     Cedar River alluvial aquifer. Schulmeyer and others (1995)
     Water-table aquifer. Mar-Apr '90 flood event. Well is 33 ft from strea


                      SUMMARY OF INPUT DATA
                      ---------------------

   STRESS TYPE (ISTRESS):              2 (stream-stage and recharge/leakage)
   TIME-STEP SIZE (DELT):              0.100D+01 (units of time)
   PRINTING CODE (IPRINT):             0 (stress data not printed)

 AQUIFER AND STREAMBANK CHARACTERISTICS (INPUT LINES 4 AND 5)
   AQUIFER EXTENT (IXL):               1 (finite width)
   AQUIFER TYPE (IAQ):                 1 (water table)
   STREAMBANK CODE (IXA):              1 (semipervious streambank simulated)
   STREAM HALF WIDTH (XZERO):          0.177D+03 (units of length)
   WIDTH OF AQUIFER (XLL):             0.149D+04 (units of length)
   STREAMBANK LEAKANCE (XAA):          0.309D+02 (units of length)
   LENGTH OF STREAM (XSTREAM):         0.100D+04 (units of length)

 AQUIFER PROPERTIES (INPUT LINE 6)
   HORIZONTAL HYDRAULIC
    CONDUCTIVITY (AKX):                0.309D+03 (units of length per time)
   RATIO OF VERTICAL TO HORIZONTAL
    HYDRAULIC CONDUCTIVITY (XKD):      0.200D+00 (dimensionless)
   SPECIFIC STORAGE (AS):              0.300D-04 (units of inverse length)
   SPECIFIC YIELD (ASY):               0.200D+00 (dimensionless)
   SATURATED THICKNESS (AB):           0.300D+02 (units of length)

 OBSERVATION-WELL DATA (INPUT LINE 7)
   DISTANCE TO OBSERVATION WELL (X):   0.210D+03 (units of length)
```

```
  TYPE OF OBSERVATION WELL (IOWS):     0 (partially penetrating)
  POSITION OF BOTTOM OF SCREENED
   INTERVAL OF OBSERVATION WELL (Z1):  0.116D+02 (units of length)
  POSITION OF TOP OF SCREENED
   INTERVAL OF OBSERVATION WELL (Z2):  0.141D+02 (units of length)

INITIAL CONDITIONS (INPUT LINE 8)
  INITIAL HEAD AT WELL (HINIT):        0.728D+02 (units of length)
  START TIME OF SIMULATION (TINIT):    0.000D+00 (units of time)

PROGRAM SOLUTION VARIABLES (INPUT LINE 9)
  NUMBER OF STEHFEST TERMS (NS):                     8
  RELATIVE ERROR FOR NEWTON-RAPHSON (RERRNR):        0.100D-09
  FACTOR DETERMINING NUMBER OF TERMS IN THE
   FINITE SUMS FOR HEAD AND SEEPAGE (XTRMS):         0.200D+02


                  SUMMARY OF STRESS DATA
                  ----------------------

  NUMBER OF SPECIFIED TIME STEPS (NT):             57
  STRESS DATA NOT PRINTED


            DIMENSIONLESS PARAMETERS (CALCULATED BY PROGRAM)
            ------------------------------------------------

  DIMENSIONLESS DISTANCE TO WELL (XD):                      0.119D+01
  DIMENSIONLESS DISTANCE TO STREAMBANK (XZEROD):            0.590D+01
  DIMENSIONLESS WIDTH OF AQUIFER (XLLD):                    0.842D+01
  DIMENSIONLESS STREAMBANK LEAKANCE (XAAD):                 0.175D+00
  BETA0:                                                    0.696D+01
  RATIO OF STORATIVITY TO SPECIFIC YIELD (SIGMA):           0.450D-02
  DIMENSIONLESS POSITION OF BOTTOM OF SCREENED
   INTERVAL OF OBSERVATION WELL (ZD1):                      0.387D+00
  DIMENSIONLESS POSITION OF TOP OF SCREENED
   INTERVAL OF OBSERVATION WELL (ZD2):                      0.470D+00


                                RESULTS
                                -------

                                           TOTAL        BANK      BANK-STORAGE
     TIME          HEAD         SEEPAGE     SEEPAGE     STORAGE       VOLUME
     (T)           (L)         (L**2/T)    (L**3/T)      (L**2)       (L**3)
     ----          ----       ----------  ----------    -------       ------
 0.000000D+00  0.72802D+02  0.0000D+00  0.0000D+00  0.0000D+00  0.0000D+00
 0.100000D+01  0.72810D+02  0.2428D+00  0.4856D+03  -.2428D+00  -.4856D+03
 0.200000D+01  0.72873D+02  -.1164D+01  -.2327D+04  0.9207D+00  0.1841D+04
 0.300000D+01  0.72859D+02  0.2146D+00  0.4293D+03  0.7061D+00  0.1412D+04
 0.400000D+01  0.72868D+02  0.3142D+00  0.6283D+03  0.3919D+00  0.7839D+03
 0.500000D+01  0.72877D+02  0.4441D+00  0.8882D+03  -.5215D-01  -.1043D+03
```

0.600000D+01	0.72859D+02	0.1368D+01	0.2736D+04	-.1420D+01	-.2841D+04
0.700000D+01	0.72840D+02	0.2049D+01	0.4098D+04	-.3469D+01	-.6938D+04
0.800000D+01	0.73093D+02	-.5348D+01	-.1070D+05	0.1879D+01	0.3757D+04
0.900000D+01	0.73827D+02	-.2404D+02	-.4808D+05	0.2592D+02	0.5183D+05
0.100000D+02	0.74157D+02	-.2504D+02	-.5008D+05	0.5096D+02	0.1019D+06
0.110000D+02	0.75162D+02	-.4737D+02	-.9474D+05	0.9833D+02	0.1967D+06
0.120000D+02	0.76313D+02	-.6725D+02	-.1345D+06	0.1656D+03	0.3311D+06
0.130000D+02	0.76697D+02	-.6053D+02	-.1211D+06	0.2261D+03	0.4522D+06
0.140000D+02	0.77677D+02	-.7678D+02	-.1536D+06	0.3029D+03	0.6058D+06
0.150000D+02	0.78382D+02	-.8052D+02	-.1610D+06	0.3834D+03	0.7668D+06
0.160000D+02	0.78507D+02	-.6840D+02	-.1368D+06	0.4518D+03	0.9036D+06
0.170000D+02	0.78763D+02	-.6600D+02	-.1320D+06	0.5178D+03	0.1036D+07
0.180000D+02	0.78823D+02	-.5864D+02	-.1173D+06	0.5764D+03	0.1153D+07
0.190000D+02	0.78758D+02	-.5017D+02	-.1003D+06	0.6266D+03	0.1253D+07
0.200000D+02	0.78489D+02	-.3809D+02	-.7618D+05	0.6647D+03	0.1329D+07
0.210000D+02	0.77929D+02	-.2071D+02	-.4141D+05	0.6854D+03	0.1371D+07
0.220000D+02	0.77207D+02	-.3052D+01	-.6103D+04	0.6885D+03	0.1377D+07
0.230000D+02	0.76710D+02	0.4309D+01	0.8618D+04	0.6842D+03	0.1368D+07
0.240000D+02	0.76242D+02	0.1119D+02	0.2238D+05	0.6730D+03	0.1346D+07
0.250000D+02	0.76020D+02	0.1081D+02	0.2162D+05	0.6622D+03	0.1324D+07
0.260000D+02	0.75781D+02	0.1301D+02	0.2601D+05	0.6491D+03	0.1298D+07
0.270000D+02	0.75410D+02	0.1950D+02	0.3899D+05	0.6296D+03	0.1259D+07
0.280000D+02	0.75523D+02	0.1059D+02	0.2119D+05	0.6191D+03	0.1238D+07
0.290000D+02	0.75332D+02	0.1490D+02	0.2980D+05	0.6042D+03	0.1208D+07
0.300000D+02	0.75214D+02	0.1547D+02	0.3095D+05	0.5887D+03	0.1177D+07
0.310000D+02	0.75126D+02	0.1550D+02	0.3099D+05	0.5732D+03	0.1146D+07
0.320000D+02	0.75067D+02	0.1503D+02	0.3005D+05	0.5582D+03	0.1116D+07
0.330000D+02	0.74956D+02	0.1646D+02	0.3293D+05	0.5417D+03	0.1083D+07
0.340000D+02	0.74924D+02	0.1520D+02	0.3040D+05	0.5265D+03	0.1053D+07
0.350000D+02	0.74814D+02	0.1687D+02	0.3375D+05	0.5096D+03	0.1019D+07
0.360000D+02	0.74755D+02	0.1650D+02	0.3299D+05	0.4931D+03	0.9863D+06
0.370000D+02	0.74672D+02	0.1717D+02	0.3433D+05	0.4760D+03	0.9519D+06
0.380000D+02	0.74614D+02	0.1692D+02	0.3384D+05	0.4590D+03	0.9181D+06
0.390000D+02	0.74557D+02	0.1684D+02	0.3368D+05	0.4422D+03	0.8844D+06
0.400000D+02	0.74529D+02	0.1600D+02	0.3200D+05	0.4262D+03	0.8524D+06
0.410000D+02	0.74502D+02	0.1545D+02	0.3090D+05	0.4108D+03	0.8215D+06
0.420000D+02	0.74476D+02	0.1501D+02	0.3003D+05	0.3957D+03	0.7915D+06
0.430000D+02	0.74450D+02	0.1465D+02	0.2929D+05	0.3811D+03	0.7622D+06
0.440000D+02	0.74589D+02	0.9543D+01	0.1909D+05	0.3715D+03	0.7431D+06
0.450000D+02	0.74547D+02	0.1143D+02	0.2286D+05	0.3601D+03	0.7202D+06
0.460000D+02	0.74525D+02	0.1156D+02	0.2312D+05	0.3486D+03	0.6971D+06
0.470000D+02	0.74430D+02	0.1270D+02	0.2541D+05	0.3359D+03	0.6717D+06
0.480000D+02	0.74430D+02	0.1155D+02	0.2310D+05	0.3243D+03	0.6486D+06
0.490000D+02	0.74544D+02	0.7575D+01	0.1515D+05	0.3167D+03	0.6335D+06
0.500000D+02	0.74582D+02	0.7116D+01	0.1423D+05	0.3096D+03	0.6192D+06
0.510000D+02	0.74620D+02	0.6424D+01	0.1285D+05	0.3032D+03	0.6064D+06
0.520000D+02	0.74549D+02	0.8909D+01	0.1782D+05	0.2943D+03	0.5886D+06
0.530000D+02	0.74553D+02	0.8115D+01	0.1623D+05	0.2862D+03	0.5723D+06
0.540000D+02	0.74532D+02	0.8439D+01	0.1688D+05	0.2777D+03	0.5555D+06
0.550000D+02	0.74619D+02	0.5449D+01	0.1090D+05	0.2723D+03	0.5446D+06
0.560000D+02	0.74494D+02	0.9696D+01	0.1939D+05	0.2626D+03	0.5252D+06