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RECHARGE TO ALLUVIAL VALLEY AQUIFERS FROM OVERBANK FLOW AND EXCESS INFILTRATION[1]

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ABSTRACT: Recharge is an important parameter for models that simulate water and contaminant transport in unconfined aquifers. Unfortunately, measurements of actual recharge are not usually available causing recharge to be estimated or possibly added to the calibration procedure. In this study, differences between observed water-table elevations and water-table elevations simulated with a model based on the one-dimensional Boussinesq equation were used to identify both the timing and quantity of recharge to an alluvial valley aquifer. Observed water table elevations and river stage data were recorded during a five-year period from 1991 to 1995 at the Ohio Management Systems Evaluation Area located in south-central Ohio. Direct recharge attributed to overbank flow during and shortly after flood conditions accounted for 65 percent of the total recharge computed during the five-year study period. Recharge of excess infiltration to the aquifer was intermittent and occurred soon after large rainfall events and high river stage. Specification of constant recharge with time values in ground-water simulation models seems inappropriate for stream-aquifer systems given the strong influence of the river on water table elevations in these systems.
(KEY TERMS: groundwater hydrology; modeling; water table; boundary interactions; alluvium aquifer; recharge; hydrogeology.)

INTRODUCTION

Two important components contributing to the water budget of an alluvial valley aquifer are recharge and stream/aquifer interaction. Recharge is the portion of infiltration that reaches the water table after passing through the soil profile. As a result of evapotranspiration and matric potentials in the soil profile, many precipitation events may not recharge the aquifer. Stream/aquifer interaction is the infiltration and exfiltration of water across the streambed. Ground-water levels in alluvial valley aquifers fluctuate with changes in stage levels in the associated surface stream. The vertical magnitude and horizontal extent bank storage during and after a flood is a function of the change in stage level of the stream, the duration of the event, the hydraulic conductivity of the stream/aquifer system, and the porosity of the aquifer (Tabidian *et al.*, 1992; Govindaraju and Koelliker, 1994; Sophocleous, 1991; van de Giesen *et al.*, 1994; Workman *et al.*, 1997).

The flood of 1993 and the recent flooding in the upper Midwest have generated a renewed interest in the dynamics of stream-aquifer interaction (Sophocleous *et al.*, 1996; Job, 1994). One method of studying stream-aquifer interaction is with the use of numerical models (Jorgensen *et al.*, 1989; Sophocleous and Perkins, 1993). Because of the complexity of groundwater models and the inability to measure recharge, a uniform recharge value with time is generally assumed for the aquifer to account for the water that percolates through the profile (Nortz *et al.*, 1994).

Workman *et al.* (1997) presented an analytical model that was capable of simulating transient water levels in an alluvial valley aquifer subject to fluctuating river stage. The mathematical model was evaluated with data obtained during a study of stream/aquifer interactions in an alluvial valley in south-central Ohio over the period of August 1991 to December 1995 (Ward *et al.*, 1994; Workman *et al.*, 1991; and Jagucki *et al.*, 1995). A portion of the data, October 1991 to September 1992, was discussed in a report by Jagucki *et al.* (1995) and used by Workman *et al.* (1997) for model evaluation. These data showed rapid movement of unconfined aquifer water levels coincident with changes in stage of the Scioto River. The model was capable of predicting daily water levels at

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a position 1500 m from the river with an average absolute deviation of 0.09 m for the 1991-1992 water year.

Since the transient model was judged to be accurate in its capability to reproduce the propagation of river floodwaters into the aquifer, unexplained small differences between observed and simulated water levels could be attributed to unaccounted recharge. This is particularly true during periods when the effect of fluctuating river stage are not as pronounced. In this paper, the stream/aquifer model was used to compute the frequency and magnitude of recharge events that occurred over a five-year period (August 1991 to August 1995).

THE MATHEMATICAL MODEL

The simplest mathematical model describing transient changes in regional ground-water flow uses the linearized Boussinesq equation with Dupuit assumptions, subject to time-dependent boundary conditions. For conditions when the right boundary remains constant, the governing differential equation and boundary conditions are

$$\frac{\partial h}{\partial t} - \frac{T}{S}\frac{\partial^2 h}{\partial x^2} = \frac{1}{S} 0 \le x \le L_x, 0 < t \tag{1}$$

$$h(0,t) = h_1(t),\ h(L_x,t) = h_2,\ h(x,0) = h_0(x)$$

where h(x,t) is the hydraulic head (L); T is the average aquifer transmissivity (L^2/t); S is the aquifer specific yield; I is the mean recharge to the aquifer (L/t); x represents the space coordinate (L); t is the time coordinate; L_x is the horizontal dimension of the aquifer (L); $h_1(t)$ represents the time-dependent river stage at the left boundary (L); h_2 represents the right boundary condition, assumed constant over time (L); and $h_0(x)$ is the initial condition (L).

Due to the linearity of Equation (1), the solution may be split into different components: the steady state component, the transient component, and the component due to the initial condition (Workman *et al.*, 1997). The final result is obtained as

$$h(x, i) = V(x, i) + M(x, i) + W_1(x, i) + W_2(x, i) \tag{2}$$

where V(x,i) is given by

$$V(x,i) = -\frac{Ix^2}{2T} + Ax + h_1(i-1),\ A = \frac{h_2 - h_1(i-1)}{L_x} + \frac{IL_x}{2T} \tag{3}$$

which is the steady state component obtained after the left boundary head at the end of the previous day i-1, $h_1(i-1)$, has completely propagated across the aquifer. M(x,i) is given by

$$M(x,i) = (h_1(i) - h_1(i-1))\left(\frac{L_x - x}{L_x}\right) = \Delta h_1(i)\left(\frac{L_x - x}{L_x}\right) \tag{4}$$

which is an approximation of the change in the left boundary condition over the time increment Δt where $\Delta h_1(i) = h_1(i) - h_1(i-1)$. $W_1(x,i)$ and $W_2(x,i)$ are given by:

$$W(x,i) = W_1(x,i) + W_2(x,i)$$

$$W_1(x,i) = \sum_{n=0}^{\infty}\left[\frac{2}{L_x}\int_0^{L_x}\left(h(\xi,i-1) - V(\xi,i)\right)\sin(\lambda_n\xi)d\xi\right]\Bigg|\sin(\lambda_n x)e^{\frac{\lambda_n^2 T}{S}}$$

$$W_2(x,i) = \sum_{n=1}^{\infty}\frac{2SL_x^2\Delta h_1(i)}{Tn^3\pi^3}\sin\left(\frac{n\pi x}{L_x}\right)\left(e^{\frac{n^2\pi^2 T}{L_x^2 S}} - 1\right) \tag{5}$$

to describe the transient redistribution of water in the aquifer. Note that W_1 requires a space numerical integration of the difference between the ground-water head and the steady state at t = i-1.

Each of the components in the solution has physical significance. The head at the end of time period i, h(x,i), is made of an "eventual" steady function V for the previous-time boundary head, $h_1(i-1)$, a function M depicting the "eventual" steady state when the increase in the boundary head $\Delta h_1(i)$ has settled in the aquifer, a transient W_1 because of the "unsettled" head from the previous time (i.e., a "correction" on V), and a transient W_2 caused by the new increase in the boundary head $\Delta h_1(i)$ (i.e., a "correction" on M).

APPLICATION OF THE MODEL

The hydrogeology of the Ohio Management Systems Evaluation Area (OMSEA) site and the areas adjacent to the site have been extensively studied (Jagucki *et al.* 1995; Norris, 1983a, 1983b; Norris and Fidler, 1969; Nortz *et al.*, 1994). These studies found the hydraulic conductivity of the aquifer to range

from 122 to 152 m/d with a mean value of 142 m/d. The specific yield of the site was estimated to be 0.18 to 0.22 with a mean of 0.2. At the OMSEA site, the unconfined aquifer was approximately 18-20 m thick and extended across the width of the valley (approximately 2 km) (Figure 1). The vertical scale of Figure 1 was greatly exaggerated to depict the cross-section of the aquifer.

The left boundary of the problem (h_1,t) was taken to be the daily-recorded elevation of the Scioto River, which drains much of central Ohio. The transient nature of flow in the river can be seen in Figure 2. A stream gage at Higby, Ohio (approximately 21 km upstream from the OMSEA) has monitored flow in the 13,290 km^2 watershed for 60 years (Nortz *et al.*, 1994). During the 60-year period, the mean flow rate was 130 m^3/s with a minimum of 6.9 m^3/s and a maximum flow of 5012 m^3/s. The change in river stage between maximum and minimum flows was 7.4 m. The National Weather Service (NWS) operates a wire-weight gage at Piketon, Ohio (approximately 2 km upstream from the OMSEA). Since the OMSEA site is downgradient of the NWS gage, Scioto River elevations adjacent to the OMSEA site were used to develop a gradient correction factor of 1.52 m between the NWS gage site and OMSEA site (Jagucki *et al.*, 1995). The gradient correction factor was subtracted from the NWS gage values to obtain a better estimate of the Scioto River stage adjacent to the OMSEA site. The total change in stage recorded during the period of August 1991 to August 1995 was approximately 7.3 m (Figure 2). The change is similar to the long term differences measured at the Higby gage and indicates that some very large flows were observed during this study. The range between low stage and high stage is unknown for the Scioto River near the OMSEA site, but the range is likely greater than the 7.4 m measured at the Higby gage because the research site is downgradient of Higby and additional tributaries empty into the river.

Workman *et al.* (1997) showed that for small, gradual changes in the right boundary condition (h_2),

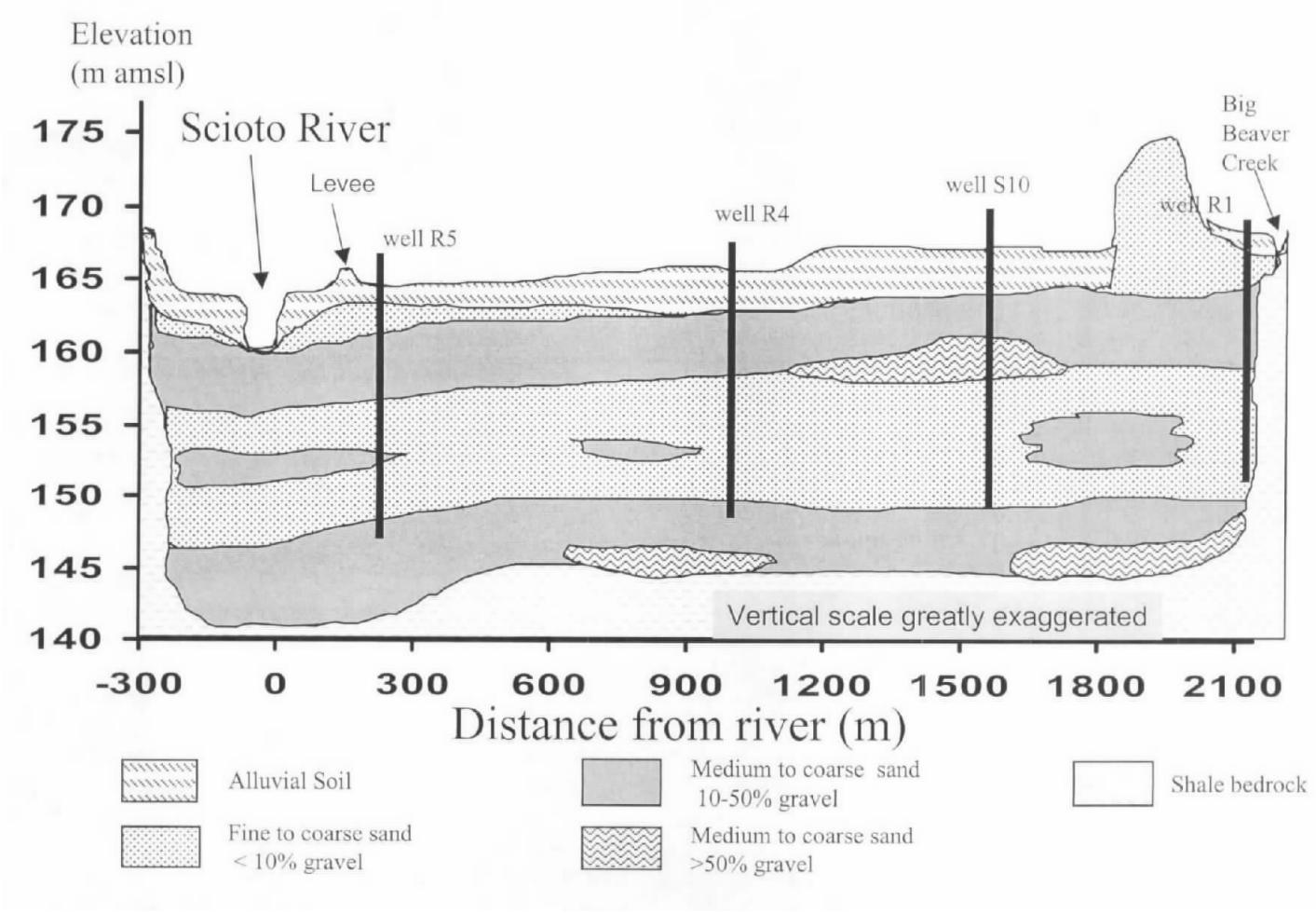

Figure 1. Cross-Section of the Scioto River Alluvial Valley Aquifer Near Piketon, Ohio.

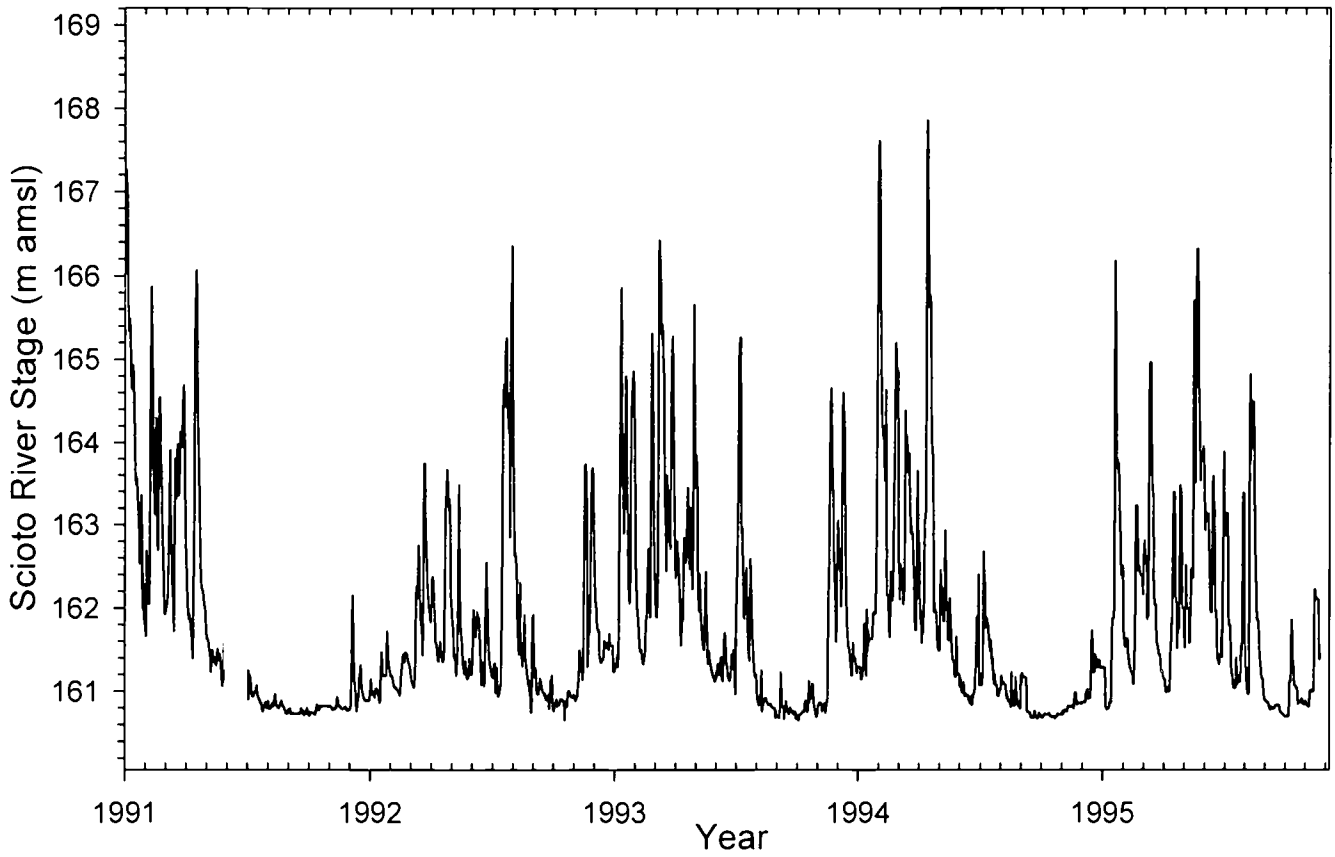

Figure 2. Daily Recorded Stage of the Scioto River Over the Five-Year Study Period.

insertion of a variable right boundary condition in Equation (3) marginally improved the overall performance of the model. Since the purpose of this paper was to use simulated and measured water-table elevations as a means to determine recharge, the right boundary was allowed to vary according to intermittent stage measurements of Big Beaver Creek located on the right boundary. Big Beaver Creek drains an area composed predominantly of the poorly permeable bedrock uplands from which runoff is rapid and bank storage is minimal (Jagucki *et al.*, 1995). These conditions produce rapid stage fluctuations in the small creek during storm events but only intermittent flow during the dry summer months.

Eleven water-table wells were constructed over a 260 ha area surrounding the OMSEA site (Jagucki *et al.*, 1995). The wells were constructed with 152-mm diameter PVC casing. A 6.1-m long, 2.54-mm slotted, PVC screen was positioned to bracket the highest and lowest expected water-table elevations in each well (Jagucki *et al.*, 1995). All of the water-table wells were instrumented with shaft encoders and electronic dataloggers that recorded hourly water levels. The three wells (R5, R4, and S14) shown in Figure 3 lie on a flow path from the eastern edge of the OMSEA site to the Scioto River (Jagucki *et al.*, 1995). These wells are located 215, 975, and 1525 m from the Scioto River, respectively. Well R1 was located at the eastern boundary of the aquifer approximately 2000 m from the Scioto River.

For purposes of testing the model, measured values from the previous characterization of the aquifer were used for each of the parameters in the mathematical model. The hydraulic conductivity was 142 m/d, the aquifer thickness was 18.3 m, and the specific yield was 0.2. The aquifer was assumed to be at steady state at the beginning of the simulation. Initial water levels across the aquifer were computed from Equation (3).

The good agreement between the observed and simulated water-table elevations in Workman *et al.* (1997) indicated that the model accurately predicted the transient changes in the aquifer subject to fluctuations in river stage during a period when recharge was negligible. In this paper, deviations between observed and simulated water-table elevations on any particular day were assumed to be the direct result of recharge to the aquifer. Because of the

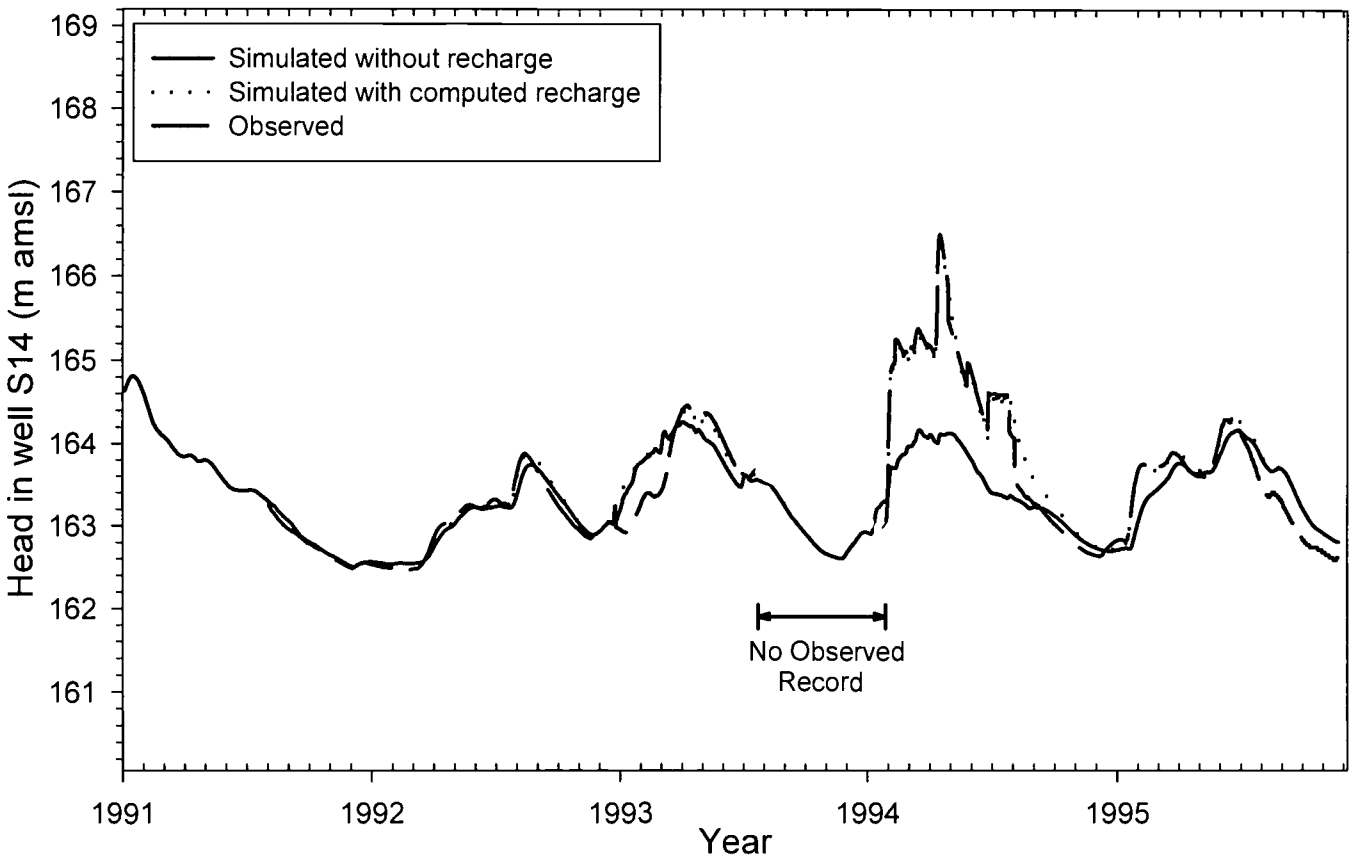

Figure 3. Observed Water Table Elevations and Simulated Water Table Elevations for Well S14 Using the Stream Aquifer Model. Observations for late summer and fall of 1993 were unavailable.

aquifer material (sand and gravel) and depth from the soil surface to the water table (2-5 m), upward flux from the water table to meet evapotranspiration demands were assumed to be negligible.

Observed water levels at well S14 were used in the model to compute recharge to the aquifer. When the simulated water level was not equal to the observed water level in well S14, a recharge value (I) was computed as:

$$I = (Observed\ WT - Simulated\ WT) * S \qquad (6)$$

where S is the specific yield of the aquifer. The simulated water levels were adjusted to the observed values and the simulation was continued. Recharge from excess infiltration occurs when positive values are computed in Equation (6). The positive infiltration, I_p, was computed as:

$$I_p = \sum_{i=1}^{n} when\ (I > 0) \qquad (7)$$

where n is the number of days. Output from the model included daily water levels across the aquifer and I_p.

RESULTS AND DISCUSSION

The simulation was started on January 1, 1991, with the initial water table across the valley assumed to be at steady state with the right and left boundaries. Water-table observations began in August 1991 and were continuous except for the late summer and fall of 1993 because of battery failure (Figure 4). The steady state solution (Equation 3) gave a good approximation of the water levels across the valley at the start of the simulation. Except for the first half of 1994, there was good agreement between the simulated and observed water table elevations assuming no recharge. The largest differences occurred when the simulated water table was much lower than the observed water table, which is an indication of unaccounted recharge.

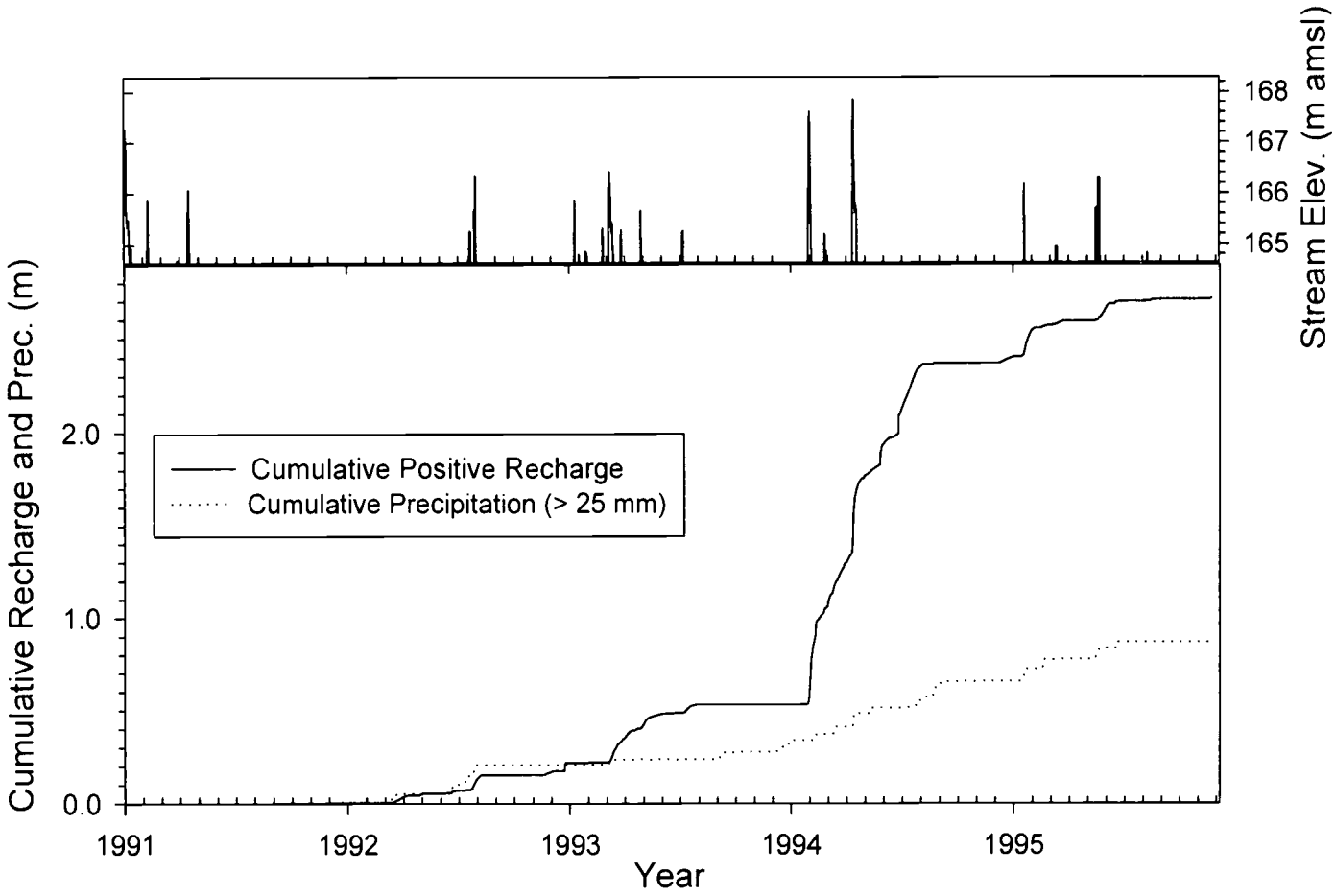

Figure 4. Cumulative Recharge as Computed by the Stream Aquifer Model and Cumulative Precipitation Greater than 25 mm. Scioto River stage is presented for instances when a flow reversal could occur between the Scioto River and Well S14.

The good agreement between the stream-aquifer model with no recharge and the observed water table elevations indicates that the boundary conditions play a more important role in water flow within the aquifer than recharge. Consistent with simulations of water and chemical transport within the aquifer modeled with the MODFLOW model, water flow was from the right hand boundary (Big Beaver Creek) towards the Scioto River (Nortz *et al.*, 1994, Finton, 1995). Throughout most of the simulation period, the water levels decreased from the right boundary to the Scioto River. An exception to the trend occurred from February to May in 1994. The two largest flood events observed during the study occurred within this 4-month period. Neither event was large enough to cause water to flood the surface at well S14, however floodwater did reach a point approximately 150 m west of well S14. Although a small levee between well R5 and the river limited over bank flow to extreme events, the large floods inundated much of the valley. The left boundary condition shifted to a point closer to well S14. Over bank flow occurred during these two flood events and added a significant amount of recharge to the aquifer.

The cumulative total of recharge computed by the model was approximately 2.6 m (Figure 4). Most of the recharge, approximately 1.6 m, occurred in 1994 during and after the two large flood events when the soil profile above the water table became saturated. Although sharp increases in recharge are noticeable during the flood event, there were also increases in predicted recharge for 1 to 2 months following the events.

Closer inspection of the data revealed that a ground-water divide existed near well S14 during the early part of 1994. Intermittent stage measurements of Big Beaver Creek were 1 to 2-m lower than water levels at well S14. The lower values for the right-hand boundary caused the model to simulate a faster decline in ground-water levels than was observed. The model continually self corrected the water levels to maintain a higher simulated ground-water level (Figure 4). A distinct soil-gravel interface existed 3 to 4 m below the soil surface in the upper terrace near

well S14. Permeability of the alluvial soil material was several orders of magnitude less than the gravel material. Once the soil profile became saturated, changes in ground-water levels were much slower because of the slow percolation of water from the soil profile and the greater potential of infiltration exceeding the storage capabilities of the soil profile.

Although 65 percent of the computed recharge occurred during the first half of 1994, there were similarities between those recharge events and the other recharge events depicted in Figure 4. Recharge was not constant and occurred at discrete times throughout the five-year period. Timing of computed recharge was coincident with most of the large rainfall events (Figure 4). In addition, the recharge events normally occurred when the stage of the Scioto River exceeded 165-m (above mean sea level). Since the water table elevation at well S14 was generally below 165-m, any stage in the Scioto River above this elevation would cause a flow from the river to the aquifer to occur. The effects of flow from the river to the aquifer on well S14 would depend on the length of time that these conditions occurred, the difference in elevation between the river and the aquifer, and the aquifer and streambed properties. Flood stage in the river, flow reversals between the river and the aquifer, and large precipitation events tend to occur together in alluvial valley systems.

CONCLUSIONS

A constant recharge value in time as assumed for many ground-water models does not seem to be appropriate for the Scioto River alluvial valley aquifer. Recharge computed by the stream aquifer model was intermittent throughout the five-year study. Significant overbank flow during large flood events overshadowed all instances of recharge that could be attributed to infiltrating water. Direct recharge associated with the two flood events accounted for 65 percent of the total recharge computed. Although less of the valley was flooded during some of the smaller flood events, a combination of direct recharge and infiltration raised the water table during these events. Recharge was not uniformly distributed over the five-year period and seemed to occur near times of large precipitation events, flood stage in the river, and periods of flow reversal between the aquifer and river. Additional work is needed to account for overbank flow and a dynamic width of the river as a boundary condition.

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LITERATURE CITED

Finton, C. D., 1994. Simulation of Advective Flow and Attenuation of Two Agricultural Chemicals in an Alluvial Valley Aquifer, Piketon, Ohio. Masters Thesis, The Ohio State University, 318 pp.

Govindaraju, R. S. and J. K. Koelliker, 1994. Applicability of Linearized Boussinesq Equation for Modeling Bank Storage Under Uncertain Aquifer Parameters. J. Hydrol. 157:349-366.

Jagucki, M. L., C. D. Finton, A. E. Springer, and E. S. Bair, 1995. Hydrogeology and Water Quality at the Management Systems Evaluation Area near Piketon, Ohio. U.S. Geological Survey Water-Resources Investigations Report 95-4139, 117 pp.

Job, C. A., 1994. The Midwest Flood of 1993: The Role of Groundwater and a Continuing Disaster. Ground Water 32(2):178-179.

Jorgensen, D. G., D. C. Signor, and J. L. Imes, 1989. Accounting for Intracell Flow in Models With Emphasis on Water Table Recharge and Stream-Aquifer Interaction: 2. Procedure. Water Resources Research 25(4):677-684.

Norris, S. E., 1983a. Aquifer Tests and Well Field Performance, Scioto River Valley near Piketon, Ohio: Part I. Ground Water 21(3):287-292.

Norris, S. E., 1983b. Aquifer Tests and Well Field Performance, Scioto River Valley near Piketon, Ohio: Part II. Ground Water 21(4):438-444.

Norris, S. E. and R. E. Fidler, 1969. Hydrogeology of the Scioto River Valley Near Piketon, South Central Ohio. U.S. Geological Water-Supply Paper, 1872, 70 pp.

Nortz, P. E., E. S. Bair, A. Ward, and D. White, 1994. Interactions Between an Alluvial-Aquifer Wellfield and the Scioto River, Ohio, USA. Applied Hydrogeology 4:23-34.

Sophocleous, M. A., 1991. Stream-Floodwave Propagation Through the Great Bend Alluvial Aquifer, Kansas: Field Measurements and Numerical Simulations. J. Hydrol. 124:207-228.

Sophocleous, M. and S. P. Perkins, 1993. Calibrated Models as Management Tools for Stream-Aquifer Systems: The Case of Central Kansas, USA, Vol. 152, pp. 31-56.

Sophocleous, M. A., A. J. Stern, and S. P. Perkins, 1996. Hydrologic Impact of the Great Flood of 1993 in South-Central Kansas. J. Irrig. Drain. Eng. 122(4):203-210.

Tabidian, M. A., D. Pederson, and P. A. Tabidian, 1992. A Paleovalley Aquifer System and Its Interaction with the Big Blue River of Nebraska During a Major Flood. *In:* The Future Availability of Ground Water Resources. American Water Resources Association, pp. 165-172.

van de Giesen, N. C., J. Y. Parlange, and T. S. Steenhuis, 1994. Transient Flow to Open Drains: Comparison of Linearized Solutions With and Without the Dupuit Assumption. Water Resources Research 30(11):3033-3039.

Ward, A. D., J. L. Hatfield, J. A. Lamb, E. E. Alberts, T. J. Logan, and J. L. Anderson, 1994. The Management Systems Evaluation Areas Program: Tillage and Water Quality Research. Soil and Tillage Research 30:49-74.

Workman, S. R., S. E. Serrano, and K. Liberty, 1997. Development and Application of an Analytical Model of Stream/Aquifer Interaction. J. Hydrol. 200:149-163.

Workman, S. R., S. E. Nokes, A. D. Ward, and N. R. Fausey, 1991. Overview of the Ohio Management Systems Evaluation Area. *In:* Proceedings of ASCE Irrigation and Drainage Conference, W. F. Ritter (Editor), Honolulu, Hawaii, July 22-26. ASCE, New York, New York, pp 725-731.