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ORIGINAL ARTICLE

Applications of particle-tracking techniques to bank infiltration: a case study from El Paso, Texas, USA

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Abstract This paper presents results of a small scale study that utilized particle-tracking techniques to evaluate transport of river water through an alluvial aquifer in a bank infiltration testing site in El Paso, Texas, USA. The particle-tracking survey was used to better define filtration parameters. Several simulations were generated to allow visualization of the effects of well placement and pumping rate on flow paths, travel time, the size of the pumping influence zone, and proportion of river-derived water and groundwater mixing in the pumping well. Simulations indicate that migration of river water into the aquifer is generally slow. Most water does not arrive at the well by the end of an 18-day pumping period at 0.54 m^3/min pumping rate for a well located 18 m from the river. Forty-four percent of the water pumped from the well was river water. The models provided important information needed to design appropriate sampling schedules for bank filtration practices and ensured meeting adequate soil-retention times. The pumping rate has more effect on river water travel time than the location of the pumping well from the river. The examples presented in this paper indicate that operating the pumping well at a doubled distance from the river increased the time required for the water to travel to the well, but did not greatly change the capture zone.

Keywords Riverbank · Filtration · Particle tracking · Modeling · MODPATH

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Introduction

In riverbank infiltration applications wells are located on the banks to capture a portion of the river water through induced infiltration. The extraction of the groundwater near the river, particularly for agricultural areas, such as the area containing the bank infiltration site of this study, can lead to river water infiltrating into the surrounding aquifer. Often, studies of river–aquifer interaction do not focus on mechanisms controlling movement of infiltrated river water inside a nearby aquifer (for instance, Sontheimer 1980; Kim and Corapcioglu 2002; Ray 2002). However, more information on filtration parameters is required to achieve a greater level of understanding of the movement of river water inside nearby aquifers.

According to Chen (2001), most river–aquifer interaction studies have focused on the discharge losses in streams due to extraction of the groundwater by a pumping well (e.g., Hantush 1965; Chen and Yin 1999, 2001). Chen (2001) emphasized that studies on river discharge depletion must be expanded by including a determination of the following characteristics: the distance the infiltrated river water can travel into the aquifer during a pumping period, the travel time from the river–aquifer boundary to the pumping well, and the area of aquifer influenced by river water. A few studies have used analytical solutions to deal with the movement of infiltrated river water inside a nearby aquifer (for instance, Chen 2001; Chen and Yin 2001). In our study we employed numerical simulations utilizing the groundwater flow code MOFLOW (a 3-D, cell-centered, finite difference, saturated groundwater flow model developed by the USGS) to initiate particle-tracking simulations using MODPATH (a 3-D, particle-tracking code developed by USGS which computes the paths for imaginary particles of water moving through the simulated groundwater system). The simulations provided useful information on important parameters pertaining to infiltration such as pumping rate and optimal distance between riverbank and the production well. In this study travel times, pathlines, and influence zones of river water were determined between a river and a nearby pumping well for seasonal groundwater extractions. These flow/transport parameters were determined to characterize the interactions between water in the river and the alluvial aquifer. Applications of such particle-tracking techniques are important in transport studies during bank infiltration to predict the attenuation of pathogens during transport and artificial recharge. This is also important to ensure that adequate soil-retention time requirements are met for removal of human pathogens as a main goal of bank infiltration operations.

Study site

The research site, about 200 m^2 in area, is located in the Rio Bosque Wetlands Park in El Paso, Texas at the border between Texas and the Mexican State of Chihuahua (Fig. 1). The site utilizes surface water in an artificial stream (stream and river in this manuscript are used interchangeably) which meanders through the park. The stream flows approximately 4 months per year with treated water from an adjacent wastewater treatment plant. The aquifer consists of medium- to fine-grained sand, with thin discontinuous beds of clay, silt, and gravel. The entire unit is generally less than 38 m thick and was deposited by the Rio Grande during the late Pleistocene to Holocene time period (Wilson et al. 1981). No wells in the study site penetrate through the aquifer. However, the aquifer thickness can be estimated from a well 75 m from the study site that penetrated 24 m of aquifer sands before encountering a 10.5-m thick layer of shale that confines the surficial aquifer. The 10-cm diameter pumping well is located 18 m from the stream. This relatively short distance to the pumping well was chosen so that little water withdrawal occurs from the surrounding aquifer and to guarantee the larger amounts of pumped water is derived from the stream which will minimize the effects of dilution from ground water and maximize bank filtration. Also, this allows short

Fig. 1 A map of the study area. The study site is shown in the enlarged view in the inset. Inset: map showing location of wells and the geometry of the river bank filtration site

sampling campaigns. The stream is 1.5–2 m deep and 4 m wide when full. The water table during the simulation experiment was 2–2.1 m below the ground surface and 1.9 m below the stream surface, approximately at the same level as the stream bed. Although the aquifer thickness is 24 m, the bottom layer (model layer 4) was assumed to have a 50-m thickness to minimize the effect of the lower boundary condition in the simulation (Fig. 2).

As is common in arid settings, the stream is a losing reach that continually discharges into the groundwater. The water table fluctuates seasonally, being high during the late summer and fall irrigation season when the nearby Riverside Canal (Fig. 1) is full and provides an additional recharge source. Groundwater is the lowest in the winter when flow in the Rio Grande and canal are reduced.

Conceptual model

For the groundwater flow simulation, a 3-D conceptual model of the bank infiltration site (Fig. 2) was constructed based on integrated hydrogeological, radar penetration, sediments, and aquifer test analyses of the site (Abdel-Fattah 2005). The model is a four-layer system, consisting of an upper unconfined layer, a leaky confining unit, a semi-confined unit, and low permeability shale at the bottom. The top and the third layer, both medium-grained sand aquifer layers, are separated by a leaky confining fine- and very fine-grained sand layer. The assumptions behind our

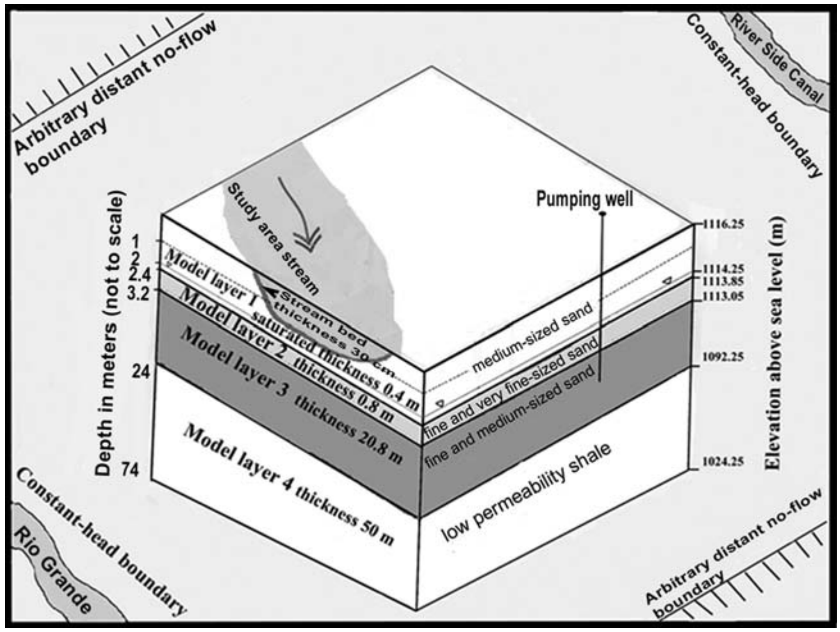

Fig. 2 A 3-D conceptual flow model for the bank infiltration site (dimensions not to scale)

model are that this semi-confined aquifer is homogeneous, isotropic, and assumes an infinite extent. The pumping well partially penetrates the semi-confined aquifer.

For the particle-tracking simulation, a 2-D conceptual model of the study site was constructed and simplified analogous to Chen's (2001) conceptual model (Fig. 3). Chen (2001) described a river–aquifer model to illustrate the migration process of river water using an analytical approach. He considered an unconfined alluvial aquifer bordered on one side by a river (Fig. 3). The unstressed gradient was nearly zero at the time of our tracer test, similarly to Chen's (2001) model. Under this circumstance, exchange of groundwater and river water does not occur (Fig. 3a). After withdrawal begins, the groundwater table around the production well declines and a cone of depression is produced (Fig. 3b). As withdrawal continues, the diameter of the cone of depression expands, and the cone eventually intercepts the river. With more pumping, the segment of influenced river may expand in upriver and downriver directions but becomes constant after steady-state conditions are reached.

In our case, the already steep hydraulic gradient from the stream is enhanced in the influenced segment. This process may take days or even weeks (Chen 2001). River water along the river–aquifer boundary ($x = 0$) begins to leak into the aquifer, and the ultimate destination of the water is the pumping well. Initially, the movement of river water in the aquifer must be very slow because the groundwater velocity is generally small near the outer limits of the cone of depression. Velocity increases as infiltrated water gets closer to the well because the gradient is much larger near the pumping well. As long as a pumping period is sufficiently long, all infiltrated water will eventually reach the pumping well. However, during a short-term pumping period, some river water may reach the well whereas other water may simply remain in the aquifer.

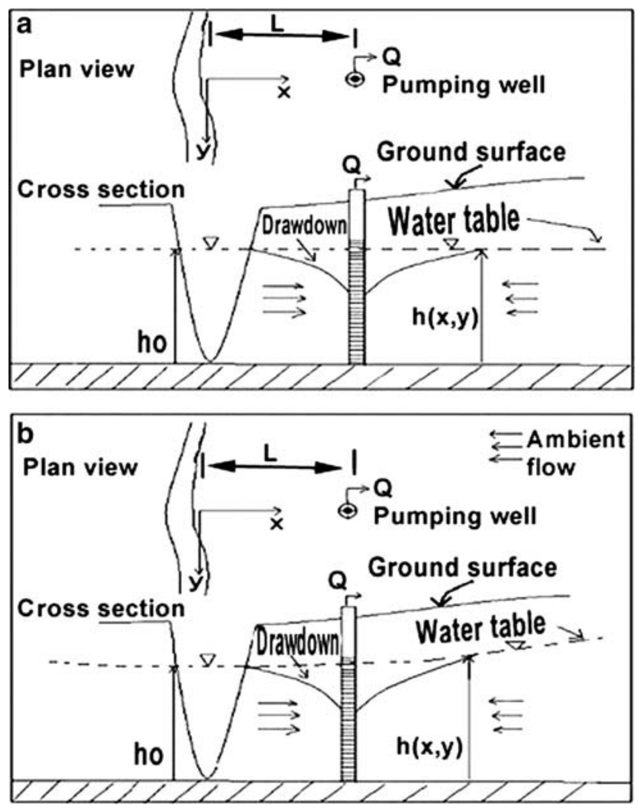

Fig. 3 Diagram showing hypothetical river-aquifer systems: **a** no hydraulic gradient between river and aquifer; **b** a regional hydraulic gradient toward the river, which gains water from the aquifer; *ho* is the water table prior to pumping, *h* is the hydraulic head at location *x*, *y*, and *Q* is the pumping rate (figures modified from Chen 2001)

The water at the location $x = 0$ and $y = 0$ (Fig. 3) requires the least time to arrive at the pumping well. Hantush (1965) termed this pathline the "meridian" line, which is the x-axis in Fig. 3.

Materials and methods

Setting the boundary conditions

Boundary conditions for the flow model consisted of: (a) two specified head (or constant head cells), which were the two water courses located northeast (Riverside Canal) and southwest (historic Rio Grande River channel) of the study area (Fig. 2). This extension of boundaries of the small study area to these natural hydrological boundaries was done to minimize the effect of boundary conditions on the simulation results. The boundaries were specified as the stages of the water courses. The water table in the model was referenced to mean sea level (WGS 84) and was held at 1114.25 m in the small study area. This depth was on average 2 m below the ground surface and was measured in pump and observation wells prior to the test. (b) Because no natural hydrogeologic boundaries occur north and south of the study area, arbitrary no-flow boundaries were set 50 m (Fig. 2) north and south of a line drawn between the pumping well and the stream. These were far enough from the pumping well such that no effects of pumping should reach them during the modeled time (Fig. 2). Also a no-flow boundary near the deep bottom boundary (shale layer) was set. Finally, the pumping well was specified as constant flux where the pumping rate is known. Recharge from the surface was considered a constant flux boundary (almost negligible), as only a single, short raining event occurred concurrently with the test. Other recharge sources in the site were infiltration from the stream to the shallow aquifer due to the induced infiltration caused by pumping. The outflow of the system was due to stream output and was represented by a constant head boundary, and due to a specified flow boundary represented by the discharging pumping well.

Model parameter estimation

Hydraulic conductivities of model layers were estimated from analysis of aquifer test data and laboratory measurements. Vertical hydraulic conductivities of core samples extracted during well construction were measured using the constant head method (Klute 1965). Horizontal and vertical hydraulic conductivities in addition to transmissivities, storage coefficients, and leakage rates through the semi-confining unit in the aquifer were estimated from pump tests. These values were considered as initial estimates of hydraulic parameters applied to the model (Table 1). Vertical hydraulic conductivity for the bed layer underlying the stream (colmation layer) was also measured using the constant head method. The value was 3.0×10^{-6}m/s after taking the geometric mean of ten samples from different depths in the clogging layer underlying the stream.

The steady-state (unstressed) groundwater flow model and calibration

The 3-D-finite difference model was developed using MODFLOW. The model was originally developed to simulate the ground-water flow and transport of bromide and microspheres tracers to mimic transport of pathogens in a testing bank infiltration system in El Paso, Texas, USA (Abdel-Fattah 2005). The GIS coverage-based module (the conceptual model approach in GMS) was used to construct the MODFLOW model by importing a digital image of the site from the USGS Seamless Data Distribution System

Table 1 Summary of initial input parameters of aquifer layers for the groundwater flow modeling

Aquifer type	Model layer 1 unconfined	Model layer 2 confined	Model layer 3 confined	Model layer 4 confined
Top elevation (m)	1116.25	1113.85	1113.05	1092.25
Bottom elevation (m)	1113.85	1113.05	1092.25	1042.25
Thickness (m)	0.4	0.8	20.8	50
Horizontal conductivity (from aquifer test (m/s)	3.1×10^{-4}	3.3×10^{-5}	3.1×10^{-4}	3.0×10^{-5}
Vertical conductivity (from aquifer test (m/s)	1.5×10^{-4}	1.4×10^{-5}	1.5×10^{-4}	1.5×10^{-6}
Vertical conductivity (from laboratory measurements (m/s)	5.0×10^{-6}	2.0×10^{-6}	2.9×10^{-5}	2.9×10^{-6}
Net recharge rate in (m/s)	7.0×10^{-7}	n/a	n/a	n/a
Specific yield (dimensionless)	5.0×10^{-2}	9.1×10^{-5}	5.0×10^{-2}	5.0×10^{-2}
Specific storage (1/m)	1.0×10^{-2}	5.0×10^{-3}	2.0×10^{-4}	8.0×10^{-6}
Transmissivity (m^2/s)	6.4×10^{-3}	3.0×10^{-5}	6.4×10^{-4}	6.4×10^{-4}

(SDDS) as a Digital Orthophoto Quad (DOQ) created by ArcGIS (inset in Fig. 1 and the background of Figs. 4, 5, 6, 7, 8, 9, 10, 11, 12). The river in the simulation was represented by a group of nodes within the grid. The head at the river nodes was specified to be equal to the river stage. Since the pumping well represents a point of convergence in the groundwater flow and causes steep gradients in the head near the well, the grid was refined to accurately model the flow near the well to a 10-cm grid. The GIS-built model was converted to 3-D finite difference grid of the four-aquifer layers. The extended 3-D-grid containing the smaller study site covered a square region measuring 1432.64 m in the x-direction by100 m in y-direction. Length in z-direction was 74.0 m. The grid type used was cell centered, consisted of 50 rows and 187 columns.

The governing equation for ground water flow through the porous medium, which was used in the model, can be written using the following partial differential equation;

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right)-W=S_{s}\frac{\partial}{\partial t}. \tag{1}$$

Where K_{xx}, K_{yy}, and K_{zz} are the hydraulic conductivities in the three orthogonal directions (m/s or ft/day), h is the head driving the flow or is the saturated thickness of the aquifer (m or ft),W is the volumetric flux per unit volume and represents the source/sink term for water or withdrawal (m^3/s or ft^3/day).

S_s is the specific storage capacity of the porous medium (dimensionless), t is the time (s or day).

When withdrawal of water occurs, $-W = R$, where R is a general sink/source term, which is defined to be intrinsically positive to represent recharge (defines the volume of inflow to the system per unit volume of aquifer per unit of time (ft^3/ft^3/day or m^3/m^3/s).

The model equations were solved numerically with a fully implicit finite difference method using the Block Successive Over-relaxation method (Wang and Anderson 1982).

The first run of the simulation was to determine the unstressed steady-state head distribution. The calibration took the initial estimates of model parameters (mainly hydraulic conductivity) and adjusted them by trial and error until the model successfully reproduced the observed configuration of the water table prior to pumping. The model was repeatedly run until the computed solution matched field-observed values within an acceptable level of accuracy.

A solution computed with this initial model was imported for the smaller bank filtration site and the errors were analyzed. The groundwater flow model was re-calibrated in the main area of interest to improve the model's predictive power at the targeted small study site. New values for hydraulic conductivity and recharge were entered, a new solution was generated, and a new error estimate was computed. The initial input parameters (Table 1) were adjusted during calibration until the degree of fit between model simulations and field measurements was quantified by statistical means used in groundwater flow modeling: The acceptable residual should be a small fraction of the difference between the highest and lowest heads across the site and be based on: (1) the ratio of the Root Mean Squared (RMS) of error to the total head loss should be small; (2) head differential of <5% for the residual mean and standard deviation; and <10% for the ratio of the standard deviation to total head change (Anderson and Woessner 1992). For this model, the estimated RMS of error was 0.049 which is very small relative to the total head loss (0.480 m). Moreover, residual mean and standard deviation were –0.034 and 0.037 m respectively, and the ratio of the residual standard deviation to the overall range in head across the model gradient was estimated to be 0.077. Calibration statistics coupled with acceptable model parameter values indicate that a very reliable calibration was achieved. It is important to note that calibration was met with the residual mean being close to 0, and the ratio of residual standard deviation to the overall range in head is adequately less than 10%.

Transient groundwater flow simulation and calibration

Predicted water levels obtained by the calibrated steady-state simulation were used as the initial heads in the transient groundwater flow simulation of the actual 18-day pump test. A forced-gradient flow was created by operating the submersible pump inserted in the pumping well (Fig. 2) for 6 h before running the test to reach steady-state flow in the saturated zone. The pumping rate was set at 0.54 m^3/min. Changes in water level in the stream during the test were negligible. Here, transient pumping conditions were modeled to examine the effect of the pumping well discharging from layer 3 from December 16, 2003 (10:30 a.m.) until January 3, 2003 (10:00 a.m.). Since flow was transient, the velocity field was calculated each time the observed water level date was available. Therefore, 66 stress periods were simulated with 1-time step of simulation assigned to each stress period. Results of transient flow simulation were presented in the form of head contour maps at the end of each time step.

In the stressed-model calibration, a set of transient-field-observed heads from the same observation wells at the site was used. The model calibration was evaluated and accepted based on hydrographs at each target location following the same procedures described above in calibrating the unstressed steady-state model. For each transient model run, an analysis of the observed versus computed water levels was conducted to determine the accuracy of the simulation. The calibration was ultimately accepted through a series of model runs using revised parameter values that produced output that agreed reasonably with the real system observations with an estimated error of ±30 cm with a 95% confidence. For the stressed model, the estimated RMS of error was 0.13 m which is very small relative to the total transient head loss (total drawdown) which was 4.61 m. Moreover, residual mean was 0.1 m, absolute mean of residuals was 0.11 m, and the ratio of the residual standard deviation (taken her as the RMS) to the overall drawdown in the model gradient (4.61 m) was estimated to be 0.03.

Particle-tracking simulations

To estimate travel times and construct pathlines of river water in the aquifer, the particle-tracking code MODPATH was used. MODPATH (McDonald and Harbaugh 1988; Pollock 1989) is 3-D USGS particle-tracking and post-processing (display) program designed to work with MODFLOW to identify travel times and paths of particles mimicking contaminants (or tracers). The results of this program represent groundwater travel times and pathlines for advective transport only. Using a flow field computed by MODFLOW, MODPATH can track a set of fictitious particles to simulate the movement of contaminants starting from user-defined point-source locations. MODPATH assumes the validity of Darcy's law and the law of conservation of mass in the same fashion as MODFLOW. MODFLOW models provide the velocities only at the midpoints of the cell boundaries. The particle velocities are calculated by MODPATH using linear interpolation. Since the velocity component of the particle at any time are known functions of the particle coordinates, the coordinates of the particle at any future time can also be computed (Pollock 1989).

The groundwater velocity was first estimated using the calibrated groundwater flow model described above. In order to estimate the tracking times, an effective porosity value was defined for each of the cells in the grid. An effective porosity value of 0.20 was used for the top and lower two layers of the model based on results from laboratory measurements. A lower effective porosity value (0.15) was used for the less permeable model layer (layer no. 2, Fig. 2). To show the migration process of infiltrated river water in the unconfined aquifer, forward and backward particle tracking were performed using MODPATH. A set of particle starting locations were specified surrounding the cell containing the pumping well. Particles were tracked forward to track infiltrated water from the stream into the shallow aquifer and backward from the pumping well to locate the origin of the flow paths in the stream for the 18-day test period.

Results

Forward tracking from the river: travel times and pathlines determination

A forward tracking scheme was used in MODPATH to determine which particles are captured by the pumping well. Particles were placed along interface between stream and aquifer in a subdomain of the model, extending 97.46 m upriver and 113.84 m downriver from the $x = 0$ and $y = 0$ position. Figure 4a shows the pathlines of river water converging at the pumping well in plan view after 18 days of pumping and Fig. 4b shows 3-D view of the same event. The white squares in Fig. 4a (gray in the 3-D view in Fig. 4b) represent the starting locations of the particles at the river–aquifer boundary, and the white arrowed lines represent the river water pathlines terminating at the pumping well whereas the black arrowed lines represent the locations of river water pathlines that did not reach the well during the 18-day pumping period. As shown in

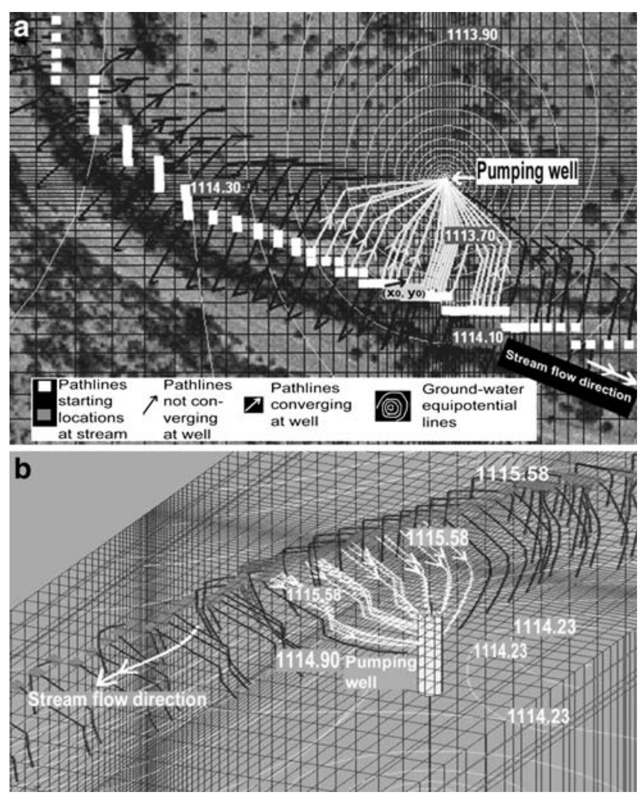

Fig. 4 Top portion (**a**) shows plan view of pathlines of induced river water that are converging and not converging at the pumping well. Bottom portion (**b**) shows 3-D view of same event

Fig. 4, not all the river water volumes reach the pumping well after the 18-day pumping period for a 18-m river pumping well distance and pumping rate of 0.54 m^3/min (Scenario 1). But the river water in the reach close to the pumping well and extending 18 m upriver (pathline 9) and 17.7 m downriver (pathline 14′) from $x = 0$ and $y = 0$ point (white lines), reaches the well (Figs. 4, 5). Figure 5 shows the distances for the pathlines starting locations of the pathlines that only converge at the pumping well. The times required for these water particles to get to the well are different. Those particles farther from the well take much longer to reach the well. For example, the water particle at the location ($x = 0$, $y = 0$) takes 1.46 days (35 h) to reach the well. This pathline is almost straight (pathline 1, Fig. 5). The water at the location ($x = 0$, $y = 18$ m; pathline 9) takes 2.71 days (65.04 h) to reach the well. From the downriver direction (the other side of the location $x = 0$, $y = 0$), the water at the location $x = 0$, $y = 17.7$ m; pathline 14′, takes 2.55 days (61.2 h) to reach the well.

Simulations of varied pumping rates and river-pumping well distances

When the river-pumping well distance was doubled (36 m) keeping the same original pumping rate of 0.54 m^3/min (Scenario 2), none of the river water pathlines converged at the pumping well (Fig. 6). However, when both the distance to the river and the pumping rate were doubled (1 m^3/min) (Scenario 3) many pathlines now converge at the pumping well (Fig. 7). At this combined effect of doubled distance and pumping rate, the water particle at the location ($x = 0$, $y = 0$) takes 1.70 days (40.8 h) to reach the well. The new upriver and downriver starting locations that have pathlines converge at the pumping well were located at $x = 0$, $y = 24.87$ m with a travel time of 2.51 days (60.24 h) and $x = 0$, $y = 29.5$ m with a travel time of 2.79 days (66.96 h) respectively.

These simulations indicate that the pumping rate has more effect on river water travel time and influence zone than does the location of the pumping well. Using Scenario 1, the pathlines 18 m upriver and 17.7 m downriver converged at the pumping well. Whereas doubling both the distance (36 m) and the pumping rate (1 m^3/min) (Scenario 3) added more distant pathlines converging at the well (24.78 m upriver and 29.5 m downriver). This conclusion was also supported by the results obtained when using half the distance (9 m) between the river and the pumping well and keeping the same original pumping rate (0.54 m^3/min) (Scenario 4). In this scenario, the water particles at the location ($x = 0$, $y = 0$) takes 0.7 days (16.8 h) to reach the well (Fig. 8). The upriver and downriver locations with

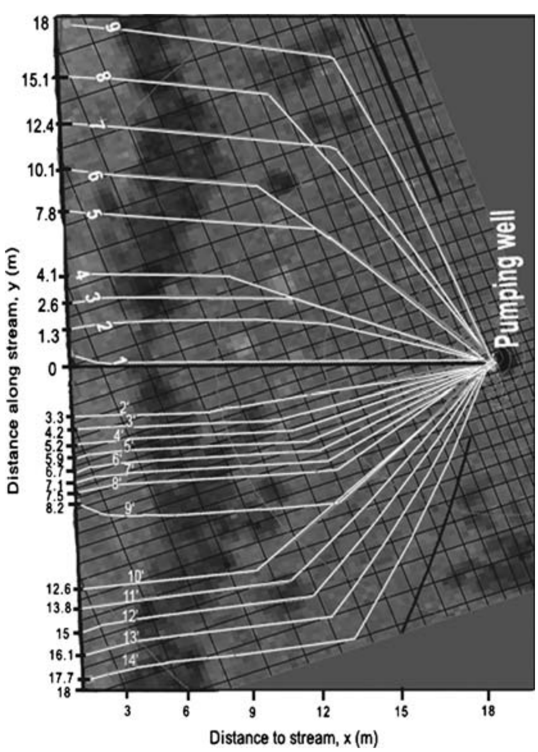

Fig. 5 Pathlines of induced river water only converging at the pumping well. Travel time along each pathline varies (Scenario 1)

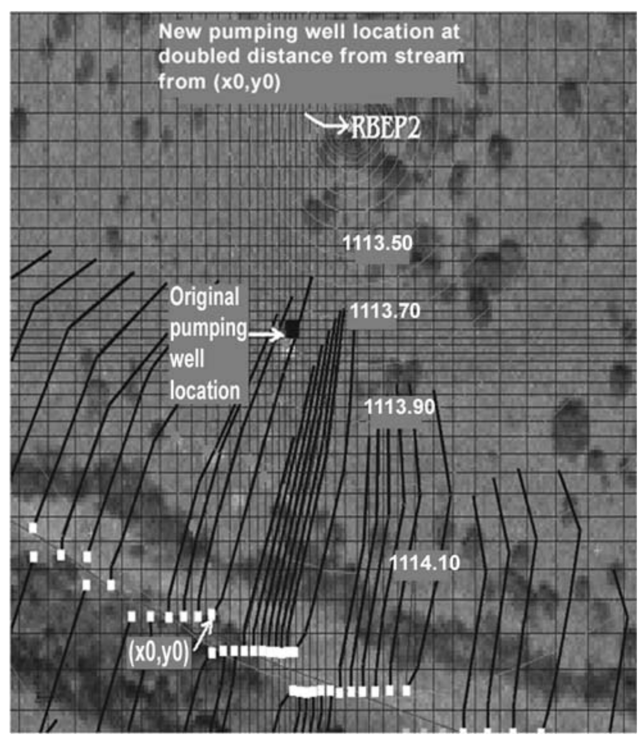

Fig. 6 Pathlines of induced river water after doubling the distance between the river and the pumping well. None is converging at the pumping well (Scenario 2)

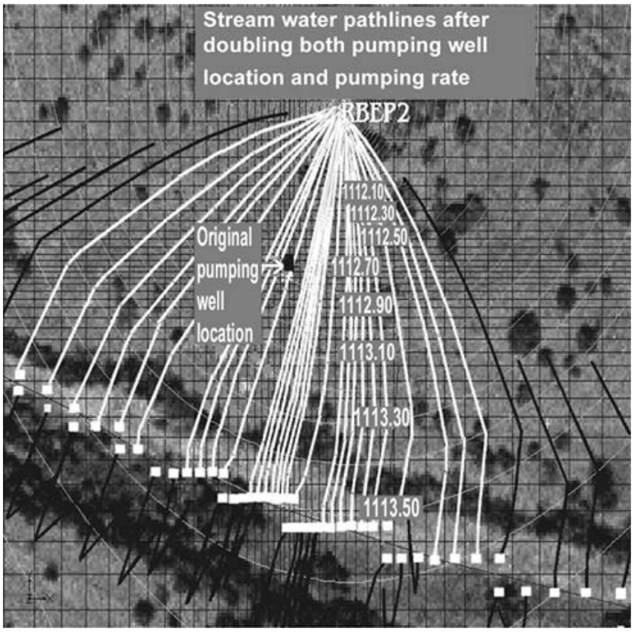

Fig. 7 Pathlines of induced river water after doubling both the pumping rate and the distance between the river and the pumping well (Scenario 3)

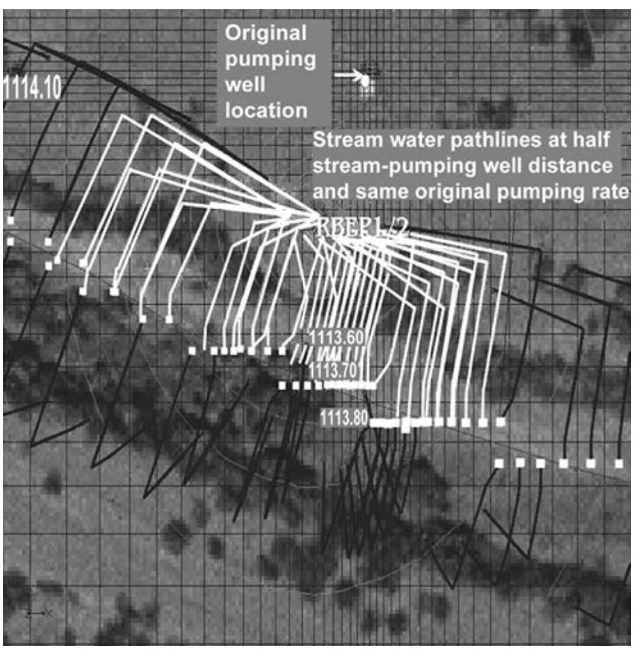

Fig. 8 Pathlines of induced river water converging at the pumping well using half the original river-pumping well distance and the same original pumping rate (Scenario 4)

pathlines converging at the pumping well were almost identical to the original simulation (Scenario 1). The upriver and downriver locations having pathlines converging at the pumping well were located at $x = 0$, $y = 17.92$ m with a travel time of 2.42 days (58.08 h) and $x = 0$, $y = 17.60$ m with a travel time of 2.44 days (58.56 h), respectively. However, when the pumping rate for this simulation of half river-well distance was doubled (Scenario 5), new upriver and downriver location with pathlines converging at the pumping well were added (Fig. 9) (29.11 m upstream and 29.67 m downstream). The shortest pathline, $x = 0$, $y = 0$, needed only 0.43 days (10.32 h) to reach the well.

Riverbank infiltration influence zone

Results from previous simulations indicated that the aquifer fraction between the river and pumping well can be replaced with infiltrated river water during a pumping period of several days. The geometry of this zone is another characteristic of interest in analyzing river–aquifer interactions incorporated in bank infiltration schemes. The river water extending 18 m upriver and 17.61 m downriver from $x = 0$ and $y = 0$ location reaches the well (white lines in Figs. 4a, b, 5) in the 18-day pumping period. The water particles traveling along the meridian line reached the well after only 2.71 days. However, those particles farther from the well take much longer to reach the well.

The river water particles moving along other pathlines remain in the aquifer, some already near the well, and others still far from the well. The area covered by the pathlines in this portion of the aquifer is filled with water from the river. Connecting the ends of these pathlines forms a curve representing equal-pumping-time locations for the induced river water (Fig. 10). The area is almost uniform as modeled because the most important influence is the 2-m head gradient between the river and the aquifer.

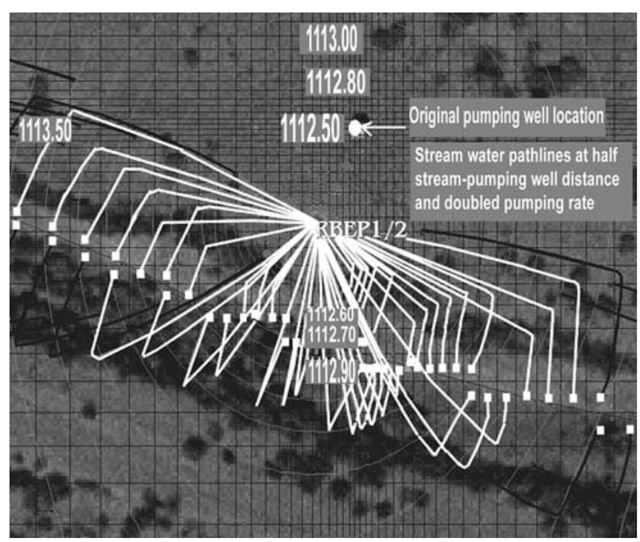

Fig. 9 Pathline of induced river water converging at the pumping well using half the original river-pumping well distance and doubling the pumping rate (Scenario 5)

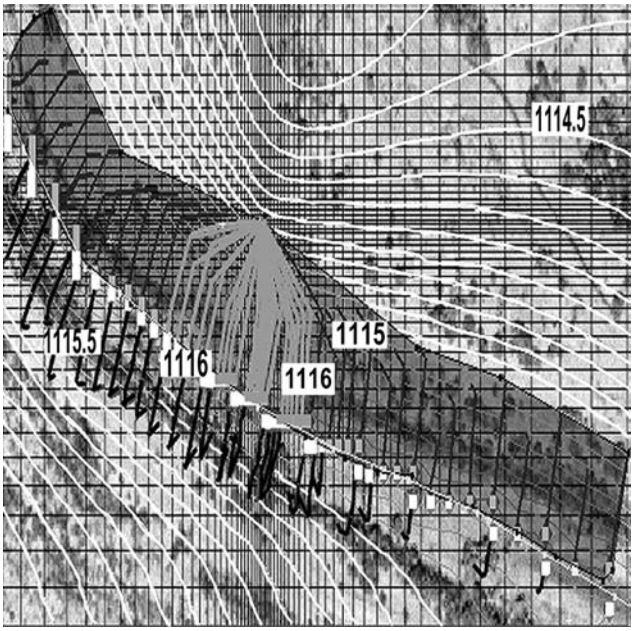

Fig. 10 Pathlines for 18-day pumping period with the area covered by the pathlines (*shaded area*) has been filled with river water infiltrated during the pumping period

This area was estimated and was found to be 2,890 m^2, which delineates the part of aquifer filled by infiltrated river water or the bank infiltration influenced zone. Note that this area represents the influenced aquifer after 18 days of continuous pumping. Given the area of bank infiltration influence and the aquifer parameters, the volume of river water discharged to the aquifer can be determined. The area of bank infiltration influence zone (A) was 2,890 m^2, aquifer thickness (b) was 24 m, and specific yield (S) was 0.05; thus the river water leaked to the aquifer is approximately equal to $A \times b \times S$, which is 3,468 m^3.

Backward tracking of particles: capture zone analysis

The delineation of backward flow paths of particles using MODPATH is shown in Fig. 11a which illustrates that water contributed from the river is mixing with water contributed from the surrounding aquifer. The 3-D nature of delineation of backward flow paths of particles is best seen in a cross sectional view (Fig. 11b).

The backward tracking option in MODPATH was used to delineate capture zones for the pumping well for given time periods. It is clear from the pumping well capture zone after 18 days of pumping (Fig. 11a) that the pathlines intersect the area covered by the river in the study site which indicates that a connection between the river and the production well occurred during the pumping period. The evolution of the capture zone with time for the pumping well is shown in Fig. 12. The area for 18-day-pumping period influence zone is 2,080 m^2. The influenced area will be smaller for a shorter pumping period as shown in Fig. 12 that illustrates seven zones for different pumping periods.

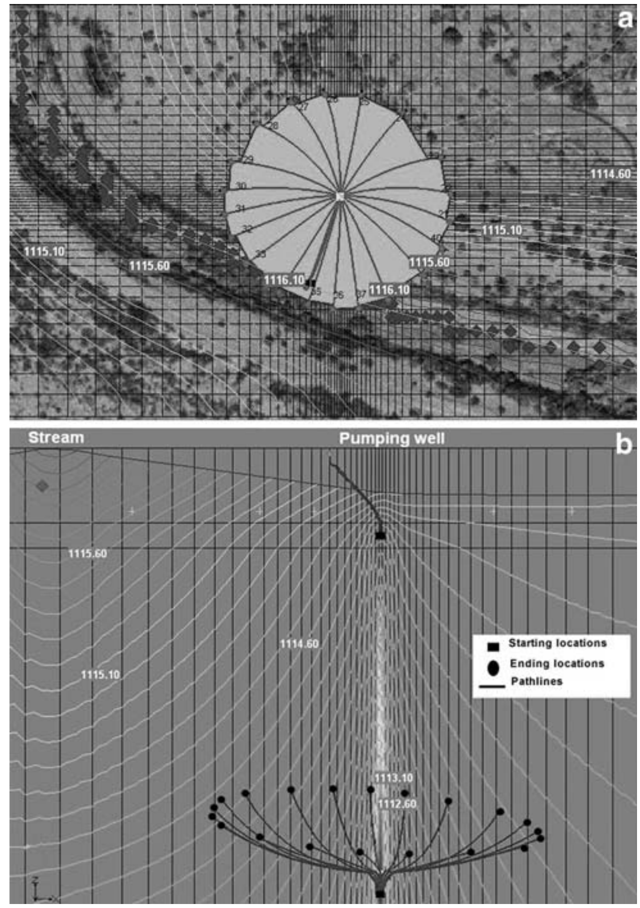

Fig. 11 Plane view (**a**) of delineation of backward flow paths of particles using MODPATH for 18-day pumping period. Capture zone illustrated by the fan-shaped polygon. The *lower part* of the figure (**b**) shows cross-section view of the same event

Mixture of groundwater and river water at the pumping well

The particle-tracking models demonstrate that river water at the study site moves slowly to the well during pumping periods. For example, for a pumping well located only 18 m from the river, it takes 1.46 days to reach the well for the shortest river-pumping well pathline, which is the straight pathline (pathline 1, Fig. 4). For wells located farther from the river, it will take longer. The pumping well will produce a mixture of river water and groundwater, thus a water sample collected from the well may therefore be a mixture of the river water and groundwater chemistry.

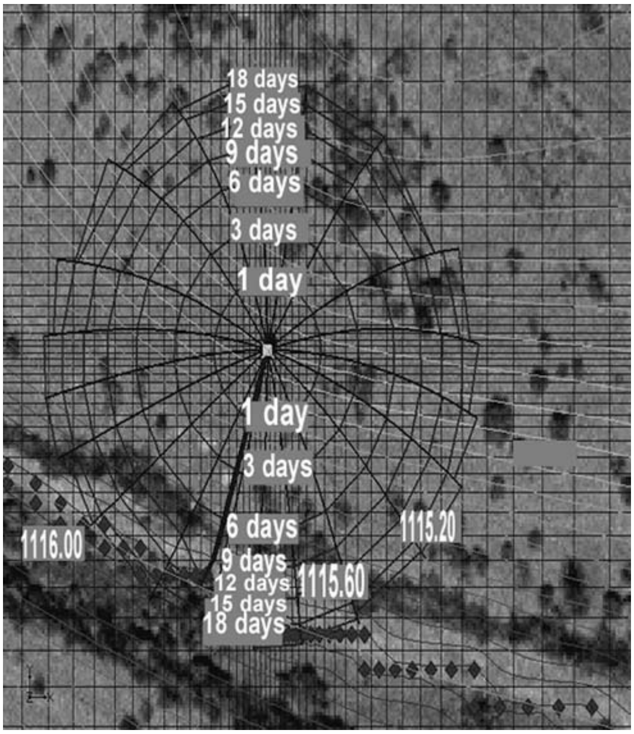

Fig. 12 Evolution of the capture zone with time for the pumping well after 1, 3, 6, 9, 12, 15, and 18 days of pumping

Chen (2001) described a method to determine the percentage of river water at the pumping well using the geometry of the river-water influenced zone around the well. The area of the aquifer contained within the farthermost upstream and downstream pathlines located above and below the meridian pathline (the one with particle's location starts at $x = 0$, $y = 0$, Fig. 5) and both converging in the pumping well is expressed in terms of the angle created by the two pathlines. The streamwater migrated along these two above- and below-pathlines has arrived at the pumping well while the streamwater migrated in the area outside the two pathlines has not. This angle is measured and compared to a 360°-angle (complete circle, which represents the capture zone of the well). As an example, according to the area of aquifer contained within the two pathlines of 17.81 m upriver; (pathline 9 in Fig. 5, of 2.71 days) and 17.65 m down-river; (pathline 14′ of 2.55 days) both located above and below the meridian pathline and form a measured angle of 160° around the well, the percentage of river water is estimated to be 44% (or the ratio of 160°/360°) of the total water flowing into the well. Because the pathlines are curved around the well, the angles should be measured near the well (for detailed description of the method, readers are referred to Chen's 2001 article).

This percentage was achieved after 18 days of pumping. However, as the pumping continues, the percentage of river water will increase in the well and the measured converging pathlines' angle will increase. Note that the distance of the well to the river is relatively short for this study. Most bank filtration wells, for example irrigation wells, are located far from rivers. Although an irrigation season generally lasts 3 months (Chen 2001), most often, several days of continuous pumping are followed by a number of days without pumping. Therefore, in reality, infiltrated river water moves even more slowly to the pumping well.

Discussion and conclusion

Particle tracking, pathlines, travel times, and influence zones are important characteristics in understanding river–aquifer interactions occurring during bank infiltration due to groundwater extraction. These techniques can be important in transport studies during bank infiltration to predict the attenuation of pathogens during transport and artificial recharge. This is especially important to ensure that adequate soil-retention time requirements are met for removal of human pathogens and other pollutants. The simulations of this study indicate that the pumping rate has more effect on river water travel time than the location of the pumping well from the river. Operating the pumping well at a doubled distance from the river increased the time required for the water to travel to the well, but did not greatly change the capture zone. The water along the meridian line ($x = 0$, $y = 0$) only traveled the pumping time period allocated for the simulation (18 days) and more time was needed to reach the well. However, doubling the pumping rate greatly expanded the influence zone and reduced the travel time of particles. When the distance between the river and the pumping well was halved and the original pumping rate was used, the travel time for particles was reduced, but the capture zone from the stream was not changed. When the pumping rate for this half river–well distance scenario was doubled, more upriver and downriver locations with pathlines converging at the pumping well were added to the influence zone.

The examples considered in this study are for partially penetrating streams. As it was shown in the earlier sections of this study, when a river partially penetrates the aquifer, the cone of depression created by the pumping well expanded into the aquifer beneath the river and even to the other side of the river (Figs. 4a, b, 6, 7, 8, 9). Also, the well was drawing from the aquifer on the other side of the river as the river, which was represented by a specified head in this simulation, worked as a water divide. Sophocleous et al. (1995) and Chen and Yin (1999) showed that a river discharges less water when it partially penetrates the aquifer. The discharge rate increases with greater river penetration. It is apparent that influence zones are smaller for partially penetrating rivers when compared with results from fully penetrating river (for instance Chen 2001).

The understanding of the relative importance of well location and pumping rate can also be used to help design appropriate sampling schedules for bank filtration applications. The pathlines and travel times of the river water, as well as the bank infiltration influence zone, can be used to predict the time when the induced river water gets into the well and the ratio of river water to groundwater at the well. The best dates and times to take water samples that better represent the infiltrated river water at the pumping well or other observation point locations could be determined more accurately. Samples taken too early may contain no infiltrated river water at all because infiltrated river water has not yet reached the well. It is recommended that pathlines and travel times be taken into account in scheduling water samples from the pumping well. Moreover, the ratio of river water to groundwater flowing into the well was determined using particle tracking. This provides an idea about groundwater dilution and mixing ratios, which is an important determining factor in assessing bank filtration removal capability.

The results of these numerical simulations agreed well with findings of Chen’s (2001) analytical model. Discharge of river water into the aquifer creates an influence zone that may have distinct water chemistry from other parts of the aquifer. If the river water is contaminated and the bank infiltration mechanisms were effective in removing certain contaminant, then the area of the aquifer that contains river water should have enhanced groundwater quality. Generally, a pumping episode in a bank filtration practice generates only a narrow influence zone in the direction perpendicular to the river, and this zone is widest along the meridian line and its width decreases rapidly upriver and downriver. The influence zone can grow toward the pumping well and upriver and downriver with multiple periods of pumping and/or increasing the rate of pumping.

The results from the simulations in this study suggest that migration of infiltrated river water into the nearby aquifers is generally slow and most infiltrated river water does not arrive at the pumping well at the end of an 18-day pumping period at a pumping rate of 0.54 m^3/min. Infiltrated river water may remain in the aquifer for several days or months before arriving at the pumping well.

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