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PAPER

Groundwater response to serial stream stage fluctuations in shallow unconfined alluvial aquifers along a regulated stream (West Virginia, USA)

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Abstract Groundwater response to stream stage fluctuations was studied in two unconfined alluvial aquifers using a year-long time series of stream stages from two pools along a regulated stream in West Virginia, USA. The purpose was to analyze spatial and temporal variations in groundwater/surface-water interaction and to estimate induced infiltration rate and cumulative bank storage during an annual cycle of stream stage fluctuation. A convolution-integral method was used to simulate aquifer head at different distances from the stream caused by stream stage fluctuations and to estimate fluxes across the stream–aquifer boundary. Aquifer diffusivities were estimated by wiggle-matching time and amplitude of modeled response to multiple observed storm events. The peak lag time between observed stream and aquifer stage peaks ranged between 14 and 95 hour. Transient modeled diffusivity ranged from 1,000 to 7,500 m^2/day and deviated from the measured and calculated single-peak stage-ratio diffusivity by 14–82 %. Stream stage fluctuation displayed more primary control over groundwater levels than recharge, especially during high-flow periods. Dam operations locally altered groundwater flow paths and velocity. The aquifer is more prone to surface-water control in the upper reaches of the pools where stream stage fluctuations are more pronounced than in the lower reaches. This method could be a useful tool for quick assessment of induced infiltration rate and bank storage related to contamination investigations or well-field management.

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Introduction

Exchange between groundwater and streams occurs according to the hydraulic gradient at their interface (Boutt and Fleming 2009). This exchange influences not only baseflow to streams but also the potential for periodic reversal of flow path, i.e. induced infiltration (Desimone and Barlow 1998). Stream stage commonly fluctuates more rapidly than the rate at which groundwater levels at distance from the stream can respond (Kelly 2001). Head variations in unconfined aquifers are influenced by pumping, stream stage fluctuations, and aquifer and streambed properties (Todd 1980; Rosenshein 1988; Welch et al. 2013). Time series of alluvial heads and adjacent stream stage often show strong correlation (Cloutier et al. 2014). A stream may be ascertained to be gaining when the groundwater stage is higher than that of the stream. Stream stage fluctuations lag groundwater fluctuations during gaining periods and the opposite for losing periods. Abrupt changes in stream stage induce changes in aquifer head but with decreased amplitude and increased lag time at increasing distance from the stream (Ferris 1952; Rosenshein 1988).

Aquifer heads can fluctuate due to lateral (stream infiltration and regional potentiometric gradient), vertical (recharge, evapotranspiration, and/or leakage), and/or pumping stresses (Ferris 1952; Hall and Moench 1972; Chen 2003; Rötting et al. 2006). Stream stage fluctuations can exert greater control over aquifer heads than well and aquifer boundary conditions in highly transmissive aquifers (Spane and Mackley 2011; Cloutier et al. 2014). Stream-induced flood waves propagate across an aquifer 2 to 3 orders of magnitude faster than typical

groundwater velocities (Jung et al. 2004; Lewandowski et al. 2009; Welch et al. 2013; Cloutier et al. 2014). Flood waves induce surface water into adjacent alluvial aquifers, which later returns to the stream as its stage is lowered; this is commonly referred to as bank storage (Squillace 1996). Bank storage does not always recede as rapidly as stream stage drops, producing a hysteresis between aquifer head and bank storage (Herrmann et al. 2013). Bank storage depends upon aquifer and streambed properties, the amplitude and duration of flood wave, pumping duration and rate, and the distance of the well from a stream (Serfes 1991; Kelly 2001; Lewandowski et al. 2009).

Jacob (1950) first examined the effects of periodic fluctuations in river stage on an aquifer and found that rate of aquifer response related to its hydraulic diffusivity. Such response has been modeled using various one-dimensional (1D) analytical solutions which estimate homogeneous aquifer properties based on matching water levels (Ferris 1952; Rowe 1960; Pinder et al. 1969; Grubb and Zehner 1973; McFadden 1983; Reynolds 1987). Such solutions have also been used to estimate bank storage and induced infiltration rate (Cooper and Rorabaugh 1963; Hall and Moench 1972; Reynolds 1987; Barlow et al. 2000; Chen and Chen 2003, Lewandowski et al. 2009; Welch et al. 2013; Cloutier et al. 2014).

Much of this research has used stream hydrographs of a single high-flow event as a forcing condition. Cooper and Rorabaugh (1963) derived solutions using a hypothetical stream boundary condition and observed that bank storage declined to 14 % of its maximum after 10 d (d = flood duration, in days). In a similar study, Chen and Chen (2003) modeled flood-induced residual bank storage and showed it declined to between 10 and 27 % of its maximum after 6 d. However, in reality, multiple flood events commonly occur over shorter timeframes than the times in these investigations. Furthermore, a key practical result of such models is how much bank storage can not only be retained but also used over a series of flood events. Therefore, a more effective application of such analytical results may require transient application over longer timeframes than for a single flood event. To the authors' knowledge, no one has studied the effects of such long-term stream stage fluctuations on bank storage and infiltration rate. This would ideally employ simultaneous measurement of stream stage and aquifer heads at different distances from the stream, in a hydrologic setting where stream fluctuations are frequent and large. Stream stage and aquifer head would act as the source function and calibration dataset, respectively, for any model employed. Ideally, a period covering both high and low flow would be useful to show seasonal differences in response.

The purpose of this study, therefore, is to investigate (1) differences in alluvial aquifer behavior along different reaches of a long stream; and (2) differences in aquifer response to stream fluctuations between low- and high-flow periods. This will employ a year-long high-frequency dataset for aquifer head at different distances from a stream undergoing frequent stream stage fluctuations. Using these data, a model will be developed and calibrated against these heads to estimate a locally homogeneous aquifer diffusivity across the full study period. Using the calibrated model, bank storage and induced infiltration rates will be estimated for the aquifer over the study period. Single-peak stage-ratio diffusivity (measured and calculated) for each well will be compared with the multiple-peak transient modeled diffusivity to determine the robustness of the model assumptions.

Study area

The study area is located in the Ohio River Valley of West Virginia (USA) with observation wells located in two public water supply (PWS) systems: Glen Dale and New Martinsville (Fig. 1). Dominant local bedrock lithologies in West Virginia are shale, sandstone, limestone, clay, and coal of Permian and Pennsylvanian age. The Pleistocene–Holocene-aged Ohio River incised those flat-lying, low-conductivity rocks to form the modern valley (Prellwitz 2004) and filled its valley with sand and gravel, silt, and clay underlying Quaternary terraces (Simard 1989; Rogers 1990). Narrow bands of Quaternary alluvium also occur as either terrace or floodplain deposits along the major tributaries. The river is regulated at a constant pool level throughout a year just upstream of a dam. The term "pool" in this context refers to a reach of a river between two consecutive dams. A pool is at close to uniform elevation, although stage declines downstream, especially along the upper pool reach. Pool levels in many locations fluctuate rapidly during and after torrential storms and spring snowmelt events—for example, the Ohio River stage near Wheeling rose 11.27 m in 2011.

Groundwater in Glen Dale and New Martinsville is pumped at approximately 2,000 and 8,000 m^3/day, respectively, from the unconfined alluvial aquifer using seven wells completed near the base of the gravel aquifer. Table 1 summarizes pumping and observation wells within the study area. The Glen Dale well field has three observation wells and two production wells, pumped intermittently one well at a time with weekly rotation. The New Martinsville well field has five pumping wells pumped intermittently that are in use with two wells, N4 and N5, alternating operation. At both sites, observed well heads allow examination of hydraulic head fluctuations both close to and at a distance from the river.

Locks with accompanying dams are major facilities for navigation along the Ohio River. Hannibal Locks and Dam lies across the Ohio River from the town of New Martinsville. Gated dams maintain a relatively constant river stage (190 m) upstream of the dam and create a head difference of 6.4 m between the upstream and downstream pools (Fig. 2). Seven of the 10 observation wells used to lie adjacent to the 68-km-

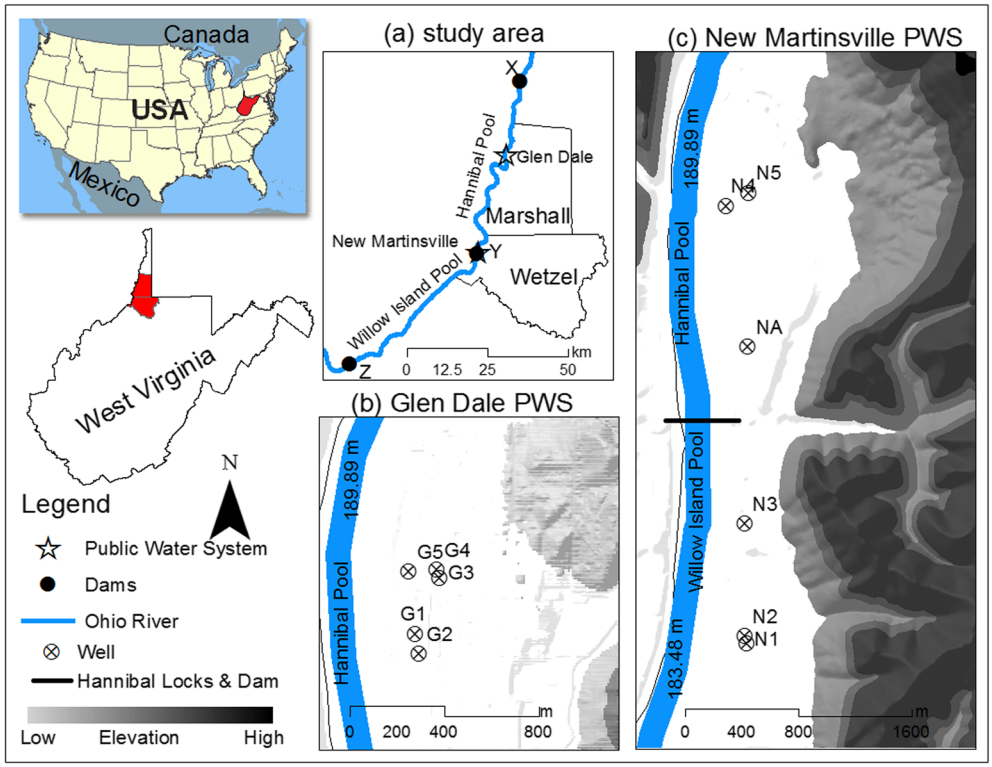

Fig. 1 Study area: *Glen Dale* and *New Martinsville* and their well fields in connection to the Ohio River. The elevations along the Ohio River are representative pool levels

long Hannibal Pool. The New Martinsville system has three wells downstream and two wells upstream of the dam. Wells N1, N2, and N3 are approximately 1,600, 1,500, and 750 m downstream and N4 and N5 are approximately 1,500 and 1,600 m upstream from the dam respectively. This setting allows examination of the effect of dam operations on groundwater flow.

Methodology

Analysis of aquifer head and stream stage

Groundwater response to stream stage fluctuations was measured at 10 wells from February 2014 to February 2015. River stages at 30-min intervals were compiled for the three US

Table 1 Characteristics of wells in the study area

PWS	Well	Pump rate (m^3/day)	Distance from river (m)	Screen length (m)	Well-top elevation[a] (m)	Well depth (m)	Water table (masl)	Casing height[b] (m)	Sensor elevation (masl)	Aquifer width (m)
Glen Dale	G1	2,071	140	7.6	199.6	24.4	190.4	0.2	184.6	1,000
	G2[c]	2,071	146	7.6	199	24.4	189.9	0.6	184.4	1,000
	G3[d]	–	250	–	204.5	24.1	193.5	0.2	189.4	1,000
	G4[c,d,e]	–	230	–	203.9	25.2	191	0	184.1	1,000
	G5[c,d]	–	110	–	197.2	9.5	190.5	0	188.7	1,000
New Martinsville	N1	1,908	270	3.7	192	17.7	181.1	1.8	180.3	950
	N2	1,635	250	3.7	192.6	18	180.9	1.8	179.8	950
	N3	2,180	90	4.6	196	22.7	182.9	0.5	178.5	300
	N4	2,589	105	6.1	195.4	22.9	189.2	0.3	186.8	1,400
	N5	2,180	270	2.7	197.5	20.4	189.1	0.3	183.8	1,400

[a] Well-top elevation obtained from drillers' report

[b] Casing height above ground surface

[c] Sealed loggers

[d] Observation well

[e] Barometric logger included

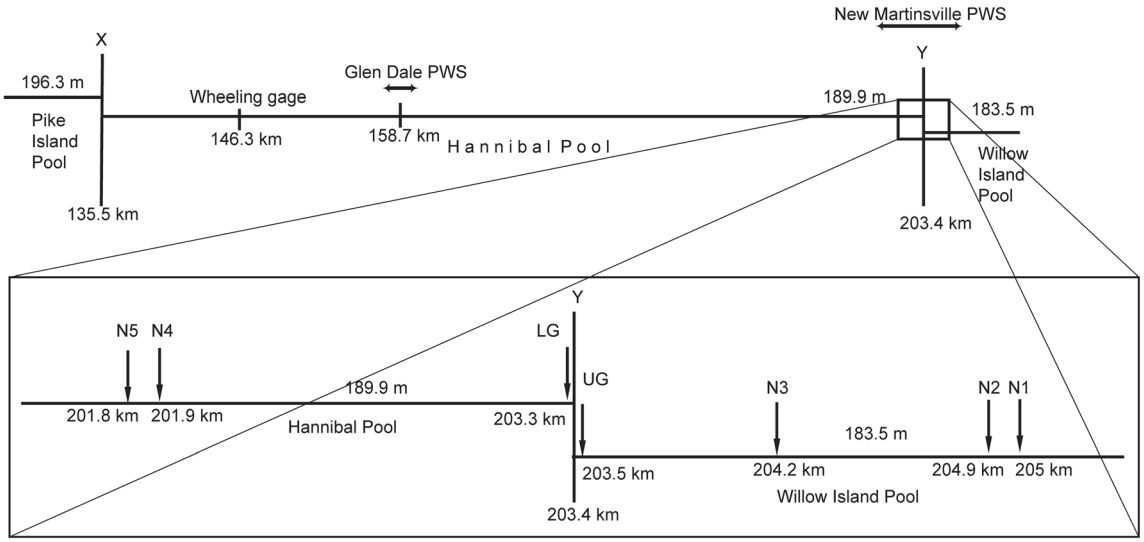

Fig. 2 Cross section of the Ohio River pools between *X* (Pike Island Locks and Dam) and *Y* (Hannibal Locks and Dam) featuring gaging stations—Wheeling, Upper Hannibal (*UG*), and Lower Hannibal (*LG*)—and *Glen Dale* and *New Martinsville PWS* with their well locations. UG is upstream of the dam at the lower reach of Hannibal Pool and LG is downstream of the dam at the upper reach of *Willow Island Pool*. The metrics below the line represent the Ohio River mileage from its origin and above the line represent pool elevation from mean sea level

Geological Survey gaging stations (Wheeling, Upper Hannibal, and Lower Hannibal) across two river pools (Hannibal and Willow Island) from October 2013 to February 2015 (USGS 2015). Pressure transducers coupled to data loggers (vented Global Water® WL-16 and sealed Onset® U020) were installed in pumping wells below their pumping water levels to collect fluid pressures at 1-min intervals. These transducers have pressure accuracy of ±0.1 and ±0.05 % of full scale, respectively. The loggers were downloaded monthly and converted to heads above mean sea level using standard techniques (Weight and Sonderegger 2001). Water column height above the sensor was measured periodically with an electric tape (±5 mm) to verify transducer readings.

For the intermittently pumping wells, daily maximum heads were employed to eliminate the effects of pumping well losses. Later, hourly heads for those wells were interpolated using cubical spline function. Lagged cross-correlation was used to estimate peak lag time between stream and groundwater levels (Sheets et al. 2002).

Analytical model of floodwave response

Figure 3 shows a conceptual model of surface/alluvial-aquifer interaction after stream stage rises from low (A) to high (B) , increasing bank storage by some volume (C). To simulate the conditions of Fig. 3, a one-dimensional solution employing the Dupuit-Forchheimer assumption was developed to estimate head across a semi-infinite aquifer bounded by the river on one side and a low-conductivity valley wall on the other.

$$\frac{\partial^2 h}{\partial x^2} = \frac{1}{D}\frac{\partial h}{\partial t} \tag{1}$$

Boundary conditions are

$$h(x,0) = h_0; \frac{\partial h}{\partial x}(L,t) = 0; \text{ and } h(0,t) = \mathrm{f}(t) \tag{2}$$

wherethe variables correspond to the following list:

$h(x,t)$	Aquifer stage at distance x from the stream at time t
D	Aquifer diffusivity
h_0	Initial water level
L	Aquifer width
t	Time since the beginning of stream stage fluctuations
$f(t)$	Time-varying stream stage

Per the Dupuit-Forcheimer assumption, Eq. (1) applies to an unconfined aquifer in which water level fluctuations are very small in comparison to the saturated thickness and all flow is horizontal (Hantush 1965). Induced aquifer head variations in response to f(t) were solved using the convolution integral by considering only surface-water fluctuations, i.e. no groundwater withdrawal, recharge, lateral inflow, or evapotranspiration (Hall and Moench 1972). Observed stream stage, only, was employed to drive aquifer head fluctuations. The solution is:

$$h(x,t) = h_0 + \int_0^t \frac{df(t)}{d\tau}\,\mathrm{erfc}\left(\frac{x}{\sqrt{4D(t-\tau)}}\right)d\tau \tag{3}$$

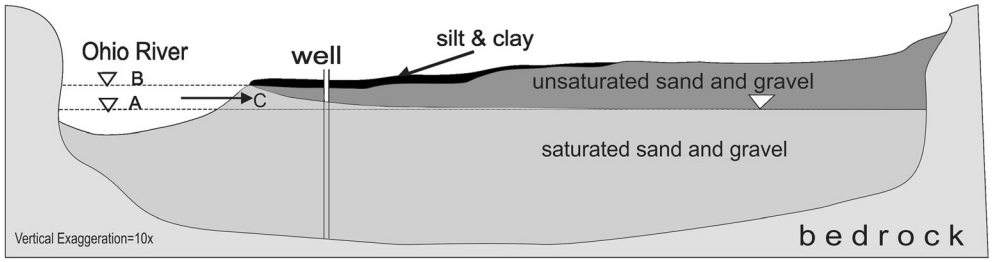

Fig. 3 Conceptual model of bank storage increase by volume *C* during stage rise from *A* to *B*. The aquifer thickness ranges between 10 and 20 m, and width ranges between 300 and 1,400 m. The Ohio River is 400–500 m wide

In numerical evaluation of Eq. (2), change in stream stage within a uniform time step were held constant. Similar to Hall and Moench (1972), the solution calculated induced infiltration rate and bank storage at each time step. The transmissivity of the aquifer was estimated from the modeled diffusivity using storage coefficient of 0.2 (assumed). The negative sign connotes aquifer outflow rate and volume, respectively.

$$Q(t) = T\frac{\partial h(0,t)}{\partial x} \tag{4}$$

$$v(t) = \int_0^t Q(t)\,dt \tag{5}$$

wherethe following terms are defined as:

T Transmissivity
Q Unit-width induced infiltration rate
v Unit-width bank storage

Model assumptions include: (1) homogeneous aquifer properties; (2) all groundwater flow perpendicular to the riverbank; (3) no pumping from wells; and (4) no recharge. Under these assumptions, stream stage fluctuations drive all aquifer response.

Simulation using Eq. (2) at hourly time steps was started at the end of a recession period, October 2013, when the stream was at the lowest stage, i.e., near steady state (Reynolds 1987). Aquifer head, induced infiltration rates, and bank storage were coded and simulated in MATLAB for the period October 2013 to February 2015. The model estimated aquifer head for a year at a time, resetting the initial condition to the lowest stream-recession stage for each year. The model was calibrated by varying aquifer diffusivity to match simulated to observed aquifer heads based on congruence of multiple hydrograph peaks associated with flood waves (Pinder et al. 1969), without exceeding the observed aquifer heads. This resulting transient diffusivity was compared to single-peak (May 2014) stage-ratio diffusivity calculated using observed peak-height ratios of well to stream, analogous to Eq. (2), assuming the stream rise was instantaneous (Carslaw and Jaeger 1959).

Results

Aquifer and stream hydrographs

Stream stage at the upper reaches of the two pools (Hannibal and Willow Island) showed fluctuations that were quite different from those in the lower reach of the Hannibal Pool (Fig. 4). In the upper reaches, high flows dominated from December to May and low flows dominated thereafter until November; however, in the lower reach, low flow was indistinguishable from high flow, and stream fluctuation was an order of magnitude or more lower than in the upper reaches. Stream stage at the upper reach of the Hannibal Pool rose up to 4 m above base flow level. The peak flows occurred at the same time as there was a drop in the regulated stage at the downstream limit of the pool (e.g., the dam); that is, there was poor correlation between hydrographs in different parts of the pools related to dam operations.

Hydraulic head in Glen Dale and New Martinsville wells at different distances from the Ohio River mimicked the stream hydrograph, especially during high flow (Figs. 5 and 6). At low flow, fluctuations in aquifer and stream stage were infrequent and of low amplitude. Stream stage remained significantly higher than aquifer water level during high flow and vice-versa during low flow at Glen Dale (Fig. 5) and the upper reach of Willow Island (New Martinsville; Fig. 6b); however, river stage remained very slightly higher than the aquifer water level almost year round in the lower reach of Hannibal Pool (Fig. 6a). Water level fluctuations in the two pools across the Hannibal Dam differed significantly (Fig. 6). Water levels in well N4 and the lower reach of Hannibal Pool were almost uniform year round, in contrast to water levels in wells N1, N2, and N3 and the upper reach of Willow Island Pool. Figure 6 shows the stream hydrographs vary due to dam operations altering natural runoff patterns both spatially and temporally. These variations in stream stage affect groundwater/surface-water interaction. During high flow, aquifer water level mimics the stream hydrograph, indicating that stream stage controls aquifer levels at these times; however, during low flow, aquifer water level is higher than river stage, indicating baseflow control.

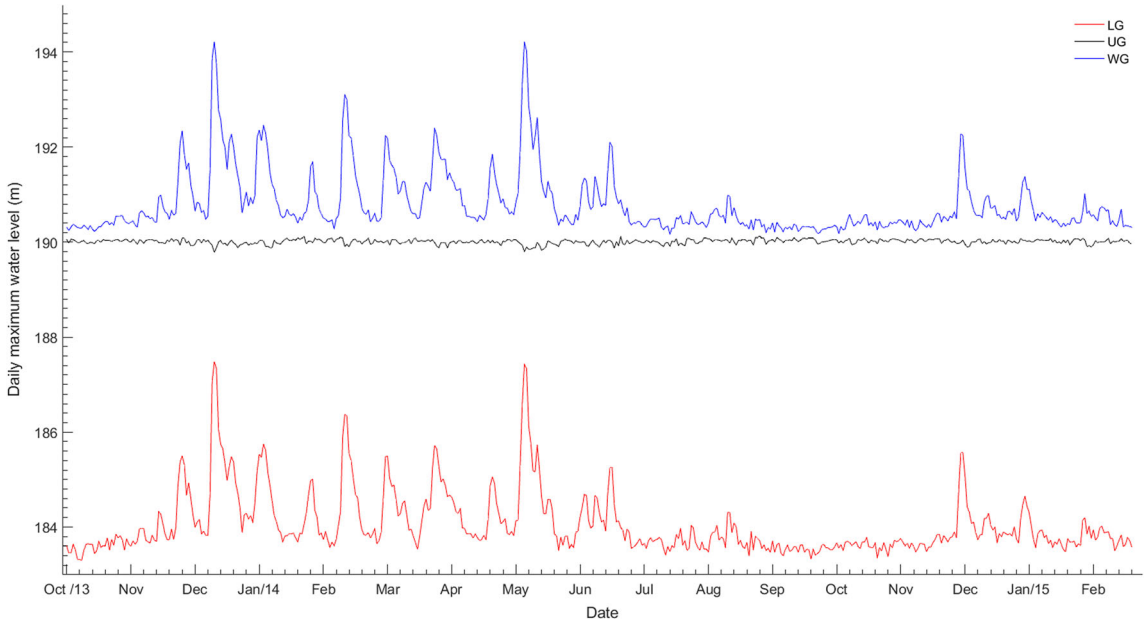

Fig. 4 Daily maximum water levels at Wheeling (*WG*), upstream of the Hannibal Locks and Dam (*UG*), and downstream of the Hannibal Locks and Dam (*LG*) from October 2013 to February 2015

Cross correlation and peak lag time during the high flow period (March–August 2014) were estimated for wells at Glen Dale and New Martinsville (Fig. 7). The time series correlated significantly well except for well N4. Aquifer water level lags river stage by 14–95 hour at the upper reaches but lags almost 3 months for wells adjacent to the lower reach. The lag time at different wells is positively correlated with distance from the river but negatively correlated with saturated thickness.

Simulated aquifer heads driven by stream fluctuations

Glen Dale site

Model results for wells G1 to G4 at Glen Dale yielded a range of diffusivities (Table 2). Following the high-flow period (December 2013 to May 2014), water levels declined until December 2014 (Fig. 8a,b). Figure 9 depicts model-estimated induced infiltration rates and bank storage in

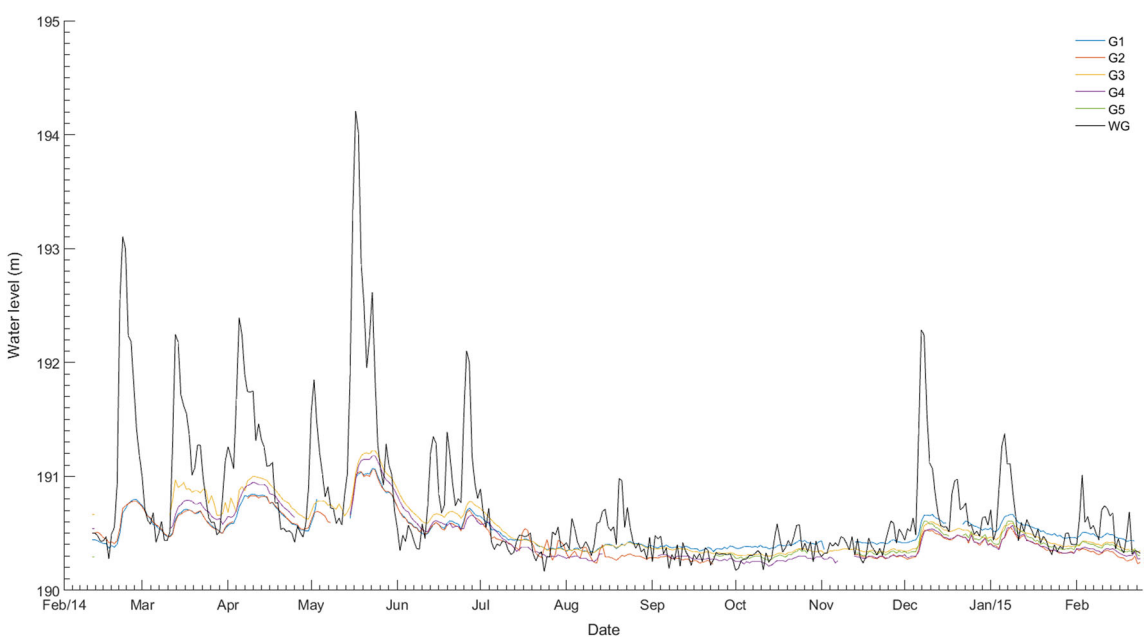

Fig. 5 Observed well heads and the Ohio River stage at the Wheeling gaging station (*WG*), Glen Dale, from February 2014 to February 2015

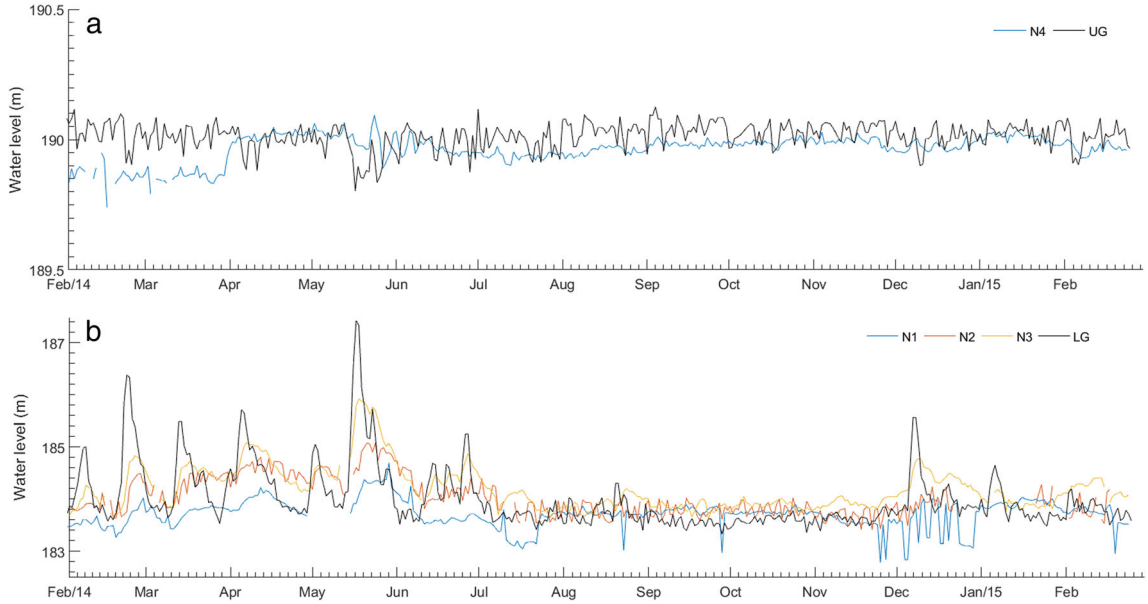

Fig. 6 Observed well heads and the Ohio River stages at Upper Hannibal (*UG*) and Lower Hannibal (*LG*) gaging stations, across the Hannibal Locks and Dam, from February 2014 to February 2015. The *y-axis* in part **a** is enlarged 5-fold in comparison to that of **b**

response to stream stage fluctuations for Glen Dale aquifer. Stream fluctuations were most frequent between December 2013 and May 2014, with maximum and minimum stages of 194.2 and 190 m respectively. Such cyclic stream fluctuations induce groundwater and surface-water exchange influencing bank storage. Cumulative bank storage reached its maximum (135 m^3/m) at the end of high flow in May 2014 and declined to 42 m^3/m by the end of the baseflow period. The maximum aquifer inflow and outflow rates (positive and negative, respectively) for Glen Dale were 24 m^3/day/m (December

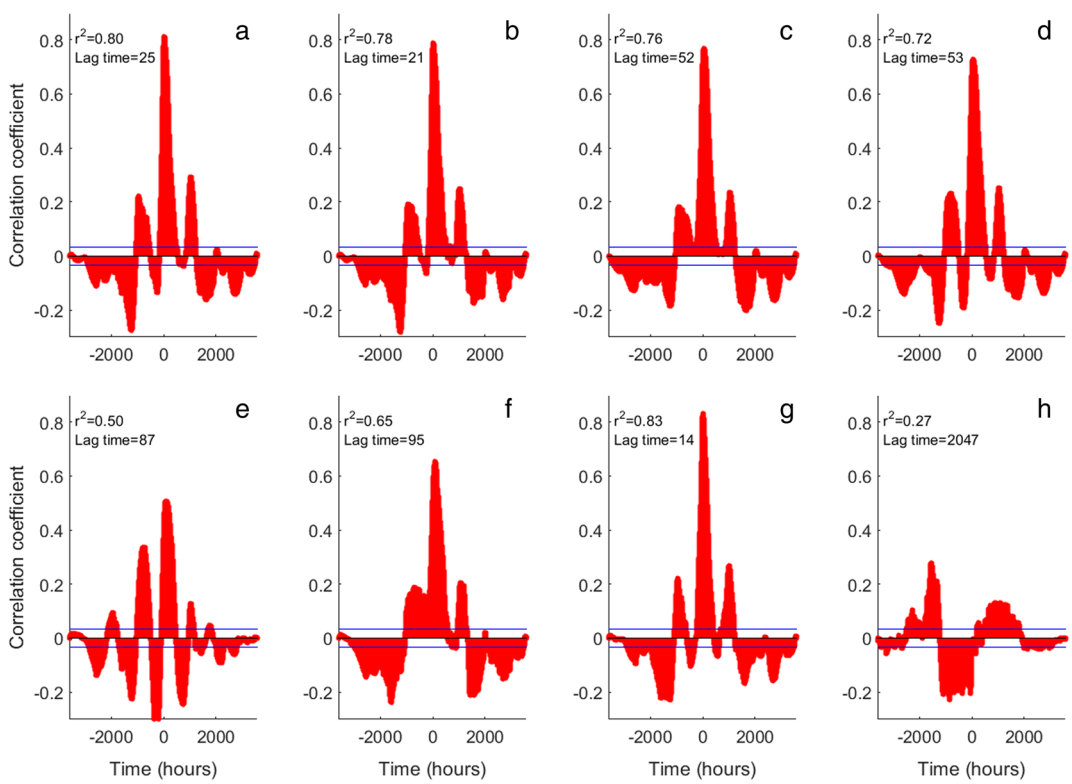

Fig. 7 Cross correlation and time lag for aquifer response to river stage fluctuations during high flow periods (March to August 2014). *Blue lines* are 95 % confidence intervals. **a**–**d** Represents wells G1, G2, G3, and G4; **e**–**h** Represents wells N1, N2, N3, and N4

Table 2 Comparison between single-peak (May 2014) stage-ratio diffusivity (measured and calculated) and multiple-peaks transient modeled diffusivity for Glen Dale and New Martinsville aquifers based on the river stage fluctuations at Wheeling gaging station (3.64 m) and Lower Hannibal gaging station (3.6 m)

Site	Single-peak ΔH_{aq} (m)	Single-peak lag time (h)	Single-peak stage-ratio diffusivity (m^2/day)	Multiple-peaks transient diffusivity (m^2/day)	Difference between the two diffusivities (%)
G1	0.64	25	5,600	1,000	82
G2	0.44	21	4,200	1,000	76
G3	0.47	52	5,400	3,400	37
G4	0.49	53	4,800	3,250	32
N1	0.64	80	6,100	7,500	23
N2	0.61	95	4,400	5,000	14
N3	1.19	14	15,500	4,000	74

2013) and 5 m^3/day/m (January and May 2014). A few flow reversals were observed during the long recession period but of lower magnitudes than those during high flow.

New Martinsville site

Upper and Lower Hannibal gage readings were used to simulate well heads in upstream and downstream aquifers from the dam, respectively (Fig. 2). As for Glen Dale aquifer, matching of simulated to observed heads yielded a range of diffusivities (Table 2). The magnitude of water level fluctuations differed between the three wells (Fig. 10a–c)—for example, the water level in well N4 was relatively static compared to wells N1 and N3. N1 showed lower amplitude response to flood events than N3, closer to the dam, whereas N1 and N3 had distinct high and low flow periods, but not N4. For wells N1, N3, and N4, the maximum error between modeled and observed water levels at high flow were 38, 70, and 10 cm, respectively, but 25, 45, and ±0.01 cm during low flow.

In contrast to Glen Dale, the stream stage in the lower reach of Hannibal Pool fluctuated little between 189.8 and 190.1 m (Fig. 11). Such minor fluctuations induced very little head difference or flow across the aquifer–stream interface. Stream stage remained nearly static at approximately 190.0 m. Maximum aquifer outflow to the stream occurred at times of peak flow at the Wheeling gage: 3 m^3/day/m (December 2013) and 2.8 m^3/day/m (May 2014). The aquifer discharged to the stream from October 2013 to May 2014 and resumed reverse-flow conditions thereafter. Bank storage near the lower reach of Hannibal Pool (N4) decreased from November 2013 to May 2014 and increased by 9 m^3/m from May–December 2014, which is the reverse of the pattern for Glen Dale (arrows in Fig. 11). This net loss of bank storage due to stream stage fluctuations was 8 m^3/m.

Stream stage fluctuations, induced infiltration rates, and cumulative bank storage at the upper reach of Willow Island Pool (well N3) were similar to that at Glen Dale (Fig. 12). Stream stage rose by up to 4 m, increasing bank storage to approximately 140 m^3/m at the end of high flow. Estimated maximum aquifer inflow and outflow rates were 46 m^3/day/m (December 2013) and 10 m^3/day/m (January and May 2014), almost twice that of Glen Dale. Unlike Glen Dale aquifer, bank storage at the end of recession was approximately −3 m^3/m.

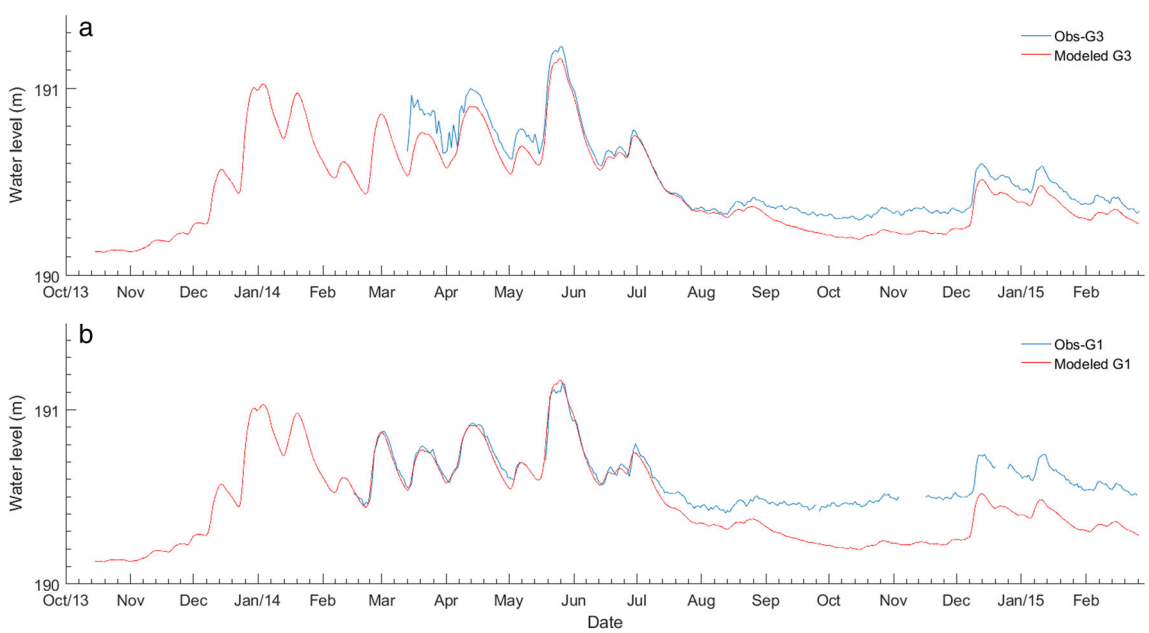

Fig. 8 Modeled versus observed daily maximum water level in **a** well G3 and **b** well G1 at Glen Dale

Fig. 9 Bank storage (*top*) and induced infiltration rate (*middle*) due to river stage (*bottom*) fluctuations near well G1 at Glen Dale

Model parameters

The multiple-peaks transient diffusivities for Glen Dale and New Martinsville wells ranged between 1,000 and 7,500 m^2/day. In comparison, the single-peak stage-ratio diffusivity for the same wells ranged between 4,200 and 15,500 m^2/day. The former values differed from the latter ones by 14–82 % with respect to the latter ones at these sites (Table 2).

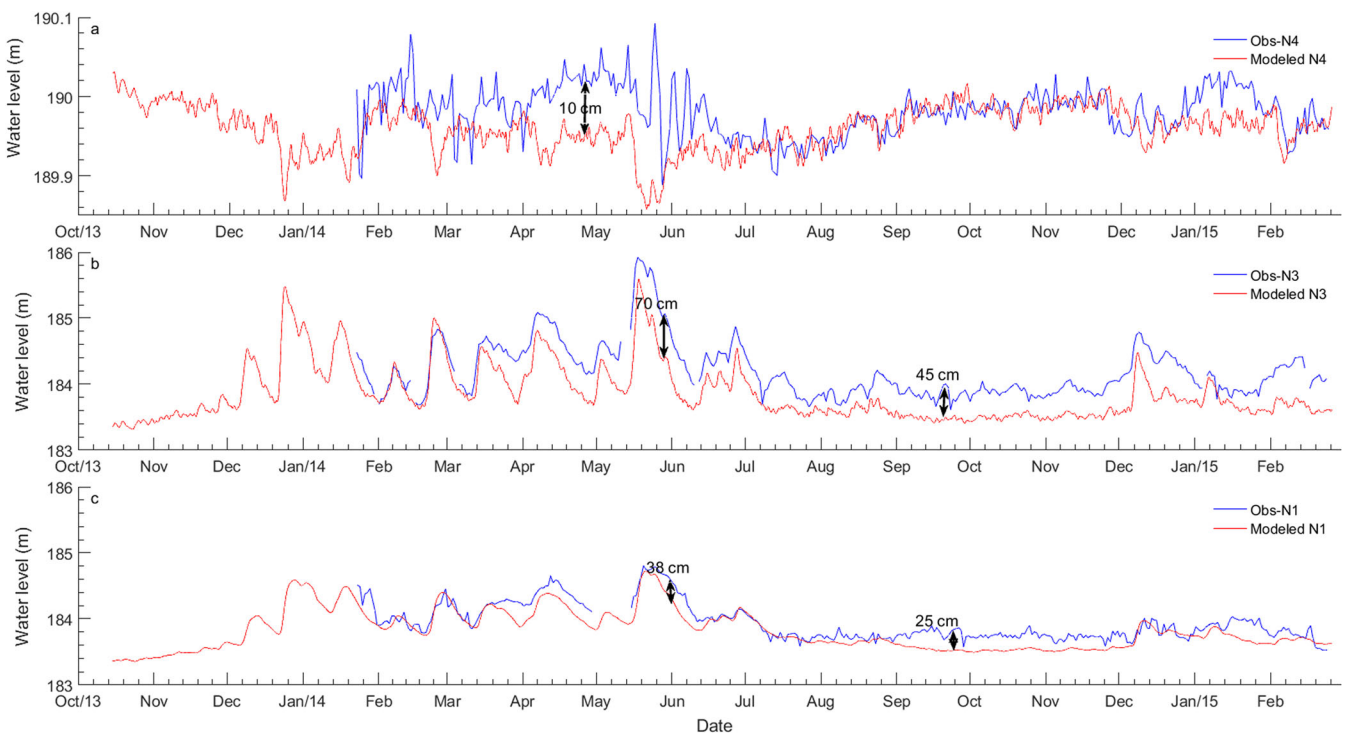

Fig. 10 Modeled versus observed water levels in wells *N1*, *N3*, and *N4* at New Martinsville. The *y-axis* in part **a** is 24-fold enlarged in comparison to that of **b** and **c**

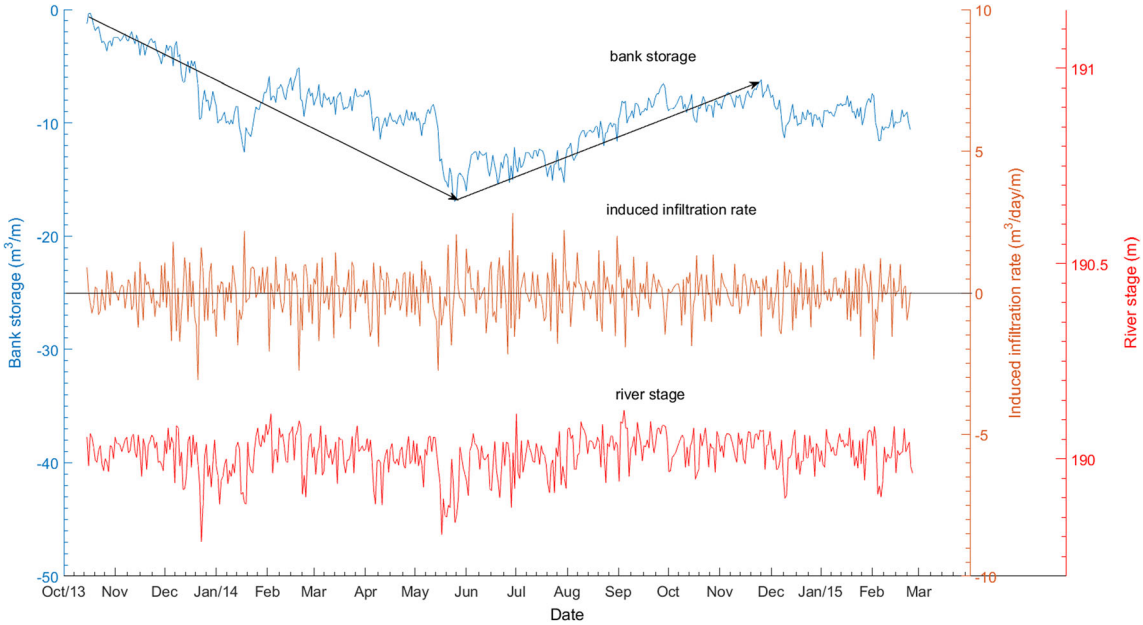

Fig. 11 Bank storage (*top*) and induced infiltration rate (*middle*) due to river stage (*bottom*) fluctuations near well N4 in the lower reach of Hannibla Pool.

Discussion

Spatial differences in groundwater-stream interaction

A year-long dataset of stream stage shows frequent peaks related to flooding events at high flow in the upper reaches of both Hannibal and Willow Island pools (Fig. 4). At the lower reach of the Hannibal Pool, however, stream stage is clearly regulated by dam operations (Fig. 4b). High and low flow periods are distinguishable for the former but indistinguishable for the latter. A fundamental difference in the observed stream forcing signal is related to, in this case, dam location and operations. Because of this difference in forcing, water levels in LG and UG, across Hannibal Dam, are inversely correlated (Fig. 6). It is likely that, at some location upstream from the dam, a transition from runoff-dominated to regulation-dominated conditions exists, suggesting that distinctly different zones with respect to groundwater/surface-water interaction prevail along the Ohio River.

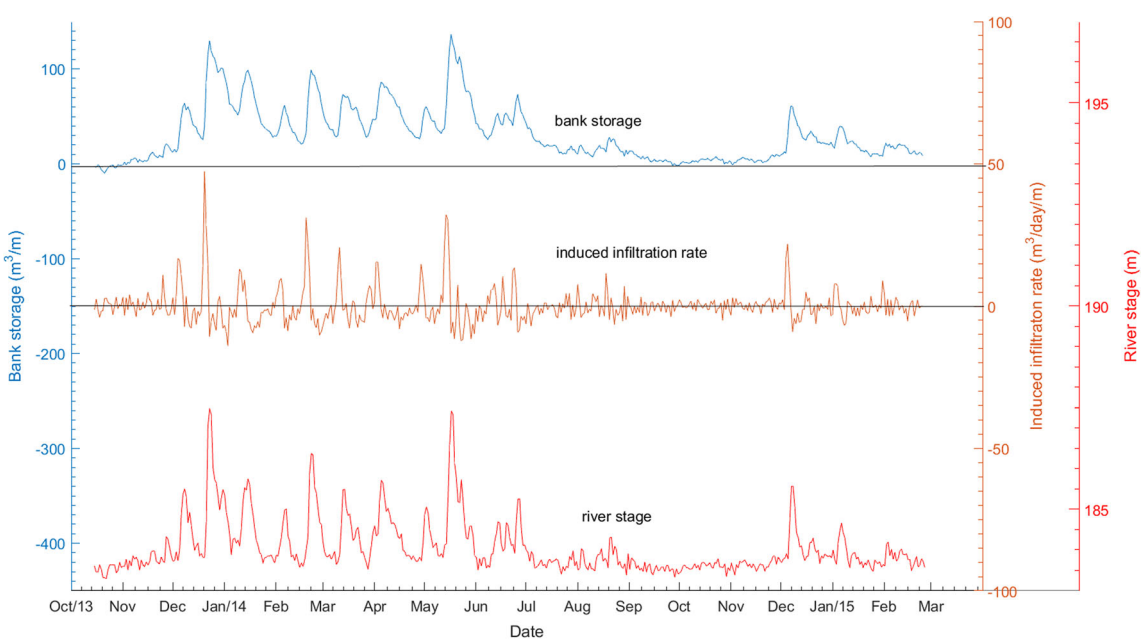

Fig. 12 Bank storage (*top*) and induced infiltration rate (*middle*) due to river stage (*bottom*) fluctuations near well N3 in the upper reach of Willow Island Pool

Wells in Glen Dale and New Martinsville at different distances from the stream responded differently to stream stage fluctuations, especially when it was higher than that in the aquifer (Figs. 5 and 6). Generally, the wells closest to the stream had higher-amplitude head fluctuations than those at distance, as would be expected. During low flow, aquifer and stream stage fluctuations were infrequent and of small amplitude, showing recession with the hydraulic gradient toward stream, i.e. baseflow conditions.

Temporal differences in groundwater–stream interaction

Model-simulated aquifer heads at Glen Dale and New Martinsville (Figs. 8 and 10) in most cases matched observed heads well during high-flow periods, but tended to deviate from congruence later in the recessional period. An exception is well N4 which showed a better fit in summer through fall 2014, when flows were actually rising due to stream regulation (note that well N4 head is at greatly larger scale compared to other wells, Fig. 10). The match tended to be much poorer later in the year during recessional periods, and the observed heads at these times tended to be underestimated by the model consistently. A major cause for this may be inherent bias in the method of calibration, which focused on wiggle-matching of hydrograph peaks that were more common in high flow periods.

The differences at low flow between observed and simulated water levels might be ascribed to boundary conditions (pumping, recharge, etc.) not incorporated into the analytical model, in addition to error in model assumptions. However, the relatively good match between modeled and observed water levels during high-flow periods in all wells but N4 suggests that prominent stream flood-event forcing is sufficient to mask recharge during these periods. Therefore, this analytical method using longer periods of actual stream and aquifer data seems to be most appropriate to fitting and parameter estimation using frequent stream high-flow events.

The variation of aquifer diffusivities from the peak-matching model suggests heterogeneity within and between the two sites (Table 2). This is due to lack of independent sources of storage parameter estimates which could vary over time and across thickness of unconfined aquifers (Pool and Eychaner 1995); or perhaps wells being at different distances from the stream (Noorduijn et al. 2014); or due to subsurface structures (Welch et al. 2014). Distant wells tended to display higher aquifer diffusivities than those close to the stream. This analytical model used simplified aquifer geometry and neglected pumping and recharge that could well play an important role in aquifer behavior, especially at low recessional flow. However, drawdown recovery for those wells were nearly instantaneously after pumping, indicating that low-rate intermittent pumping had a minimal effect on aquifer hydraulic heads. Further, numerical and chemical modeling is required to account for other complex boundary conditions; however, simulated aquifer heads tend to agree with measured heads within the distance of bank storage gain created by induced infiltration, as demonstrated by the model.

Effects of dam operation

Dam operations in this study caused the upper reaches of pools to behave as a fluctuating stream and the lower reaches to behave more like a non-fluctuating lake, maintaining an almost constant stage year-round. Such behavior has implications for groundwater/surface-water exchange, with upper-reach aquifers gaining water during high flow periods and losing it during low flow. However, lower-reach aquifers (as in the lower Hannibal Pool) showed bank storage losses during spring and gains during summer, both at a much lower rate than upper pools (Fig. 11). The head difference across the dam promotes groundwater to flow parallel to the stream not perpendicular, which violates the assumptions of the analytical model. The well N3 displayed higher head fluctuations as well as greater deviation between simulated and observed heads than that of well N1 (Fig. 10). The hydraulic gradients from the Hannibal to Willow Island pools, parallel to the river, varied from 0.001 to 0.002 during high and low flow periods. The higher hydraulic gradient infers higher rate of groundwater flow.

Conclusions

Groundwater response to stream stage fluctuations was studied in shallow unconfined alluvial aquifers along the regulated Ohio River. A year-long dataset of stages from three stream gages in two pools were collected, as well as 10 wells in two PWS systems near these gages. Field data and modeling results show well heads close to the river fluctuate more than those farther away from the stream–aquifer boundary.

Bank storage gains were induced by stream fluctuation, especially during high-flow periods and at the upper reaches of the pools. Very minor variations in stream stage were observed in the lower reaches of these pools near dams, an order of magnitude less than in the upper reaches of pools. The exchange rate for lower reaches was small and out of phase (i.e., in the opposite direction) with respect to the upper reaches of the same pool. Therefore, groundwater/surface-water interaction appears to be spatially and temporally variable along the regulated stream.

Aquifer diffusivities estimated by wiggle matching simulated peaks using the transient analytical model to

observed peaks showed gross agreement in single-peak stage-ratio diffusivity. Irregularity amplitudes of groundwater fluctuations and a wide range of estimated aquifer diffusivity values for different wells in the same well field suggest heterogeneous aquifer conditions which could not be assessed using the homogeneity-based analytical model. Further, numerical and chemical modeling are required to account for other complex boundary conditions. During high-flow periods, simulated aquifer heads using stream stage fluctuations without recharge matched observed heads reasonably well at the upper reaches of pools. However, modeled heads during baseflow deviated substantially from measured heads, almost certainly due to unaccounted boundary conditions present in reality but unaccounted for in this model.

Actual induced infiltration rate and bank storage could differ significantly from those estimated by this model where significant recharge, partial penetration, regional gradient, or heterogeneity are present. Nonetheless, this method confirms that locations along streams where enhanced groundwater/surface-water exchange may be induced. This method seems to give meaningful results in an area where large flood peaks create apparent reversals of flow into adjacent high-conductivity alluvium and could be a useful tool for assessment of induced infiltration and bank storage changes related to investigations of contamination or well field management.

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