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Groundwater–surfacewater interactions at the transition of an aquifer from unconfined to confined

Lensyl Urbano[a,*], Brian Waldron[b,1], Dan Larsen[a,2], Heather Shook[a,1]

[a]Department of Earth Sciences, University of Memphis, Memphis, TN 38152, USA
[b]Ground Water Institute, University of Memphis, Memphis, TN 38152, USA

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Abstract

A three-dimensional numerical model is developed to identify systematic changes in the interactions between the groundwater flow systems and streams that cross the transition zone where unconfined aquifers become confined. A generalized model is designed that represents hydrologic conditions where gently dipping aquifer-confining unit sedimentary sequences subcrop beneath a thin surficial aquifer and are transected by gaining streams flowing parallel to dip. A series of simulations are performed to investigate the effect of different hydrogeologic characteristics of the edge of the confining unit, where potential for facies changes or fluvial reworking, results in variations in the material properties of the confining unit. The results show an abrupt increase in groundwater discharge to the stream immediately upstream of the transition to confined conditions, and a corresponding, but not equal, decrease in groundwater discharge along stream reaches downstream of the edge of the confining unit. Model results were compared to measured stream discharges at a location where the Loosahatchie River of eastern Tennessee, USA, crosses an unconfined–confined transition in the upper Mississippi Embayment.

Keywords: Ground water; Surface water; Hydrogeology; Modeling

1. Introduction

Studies of ground-water/surface-water interactions in sedimentary aquifers have typically focused on the relationships between the aquifer and stream, and less on the effects of intervening confining units. In many types of sedimentary successions, finer-grained clay, mudstone, or shale deposits are interbedded with more permeable sand or limestone to create multi-layer aquifer-confining unit systems. For example, passive margin, continental-shelf deposits typically exhibit gently dipping sequences of transgressive–regressive sedimentary deposits. Laterally, the lower permeability, offshore facies thicken toward the center of the basin and grade landward into higher permeability, sandy coastal facies (Elliott, 1986) or are truncated by sandy fluvial deposits along the basin flanks (Wagoner et al., 1990; Dalrymple et al., 1994).

* Corresponding author. Fax: +1 901 678 4467.
E-mail addresses: lurbano@memphis.edu (L. Urbano), hshook@memphis.edu (B. Waldron), dlarsen@memphis.edu (D. Larsen), hshook@memphis.edu (H. Shook).
[1] Fax: +1 901 678 4467.
[2] Fax: +1 901 678 2178.

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Thus streams flowing parallel to the dip of these sedimentary sequences will first traverse the recharge areas of unconfined sandy aquifers that then become confined toward the center of the basin by lower permeability sediments. This type of stream and sedimentary-succession geometry is characteristic of shallow sedimentary basins, such as the Mississippi Embayment (Cushing et al., 1964, Fig. 1), and coastal plain deposits.

An important threshold in the groundwater flow regime is often present at the unconfined–confined transition of multi-layer aquifer-confining unit systems. The unconfined portion of the aquifer may be in direct contact with streams and surface water bodies and ground water contained within this part of the aquifer is much more likely to discharge to the surface than water in the contiguous confined aquifer. Confined water tends to remain within the aquifer and flow parallel to the direction of flow in the stream (Toth, 1963). Larkin and Sharp Jr. (1992) describe this as the difference between baseflow and interflow dominated interactions between the stream and the

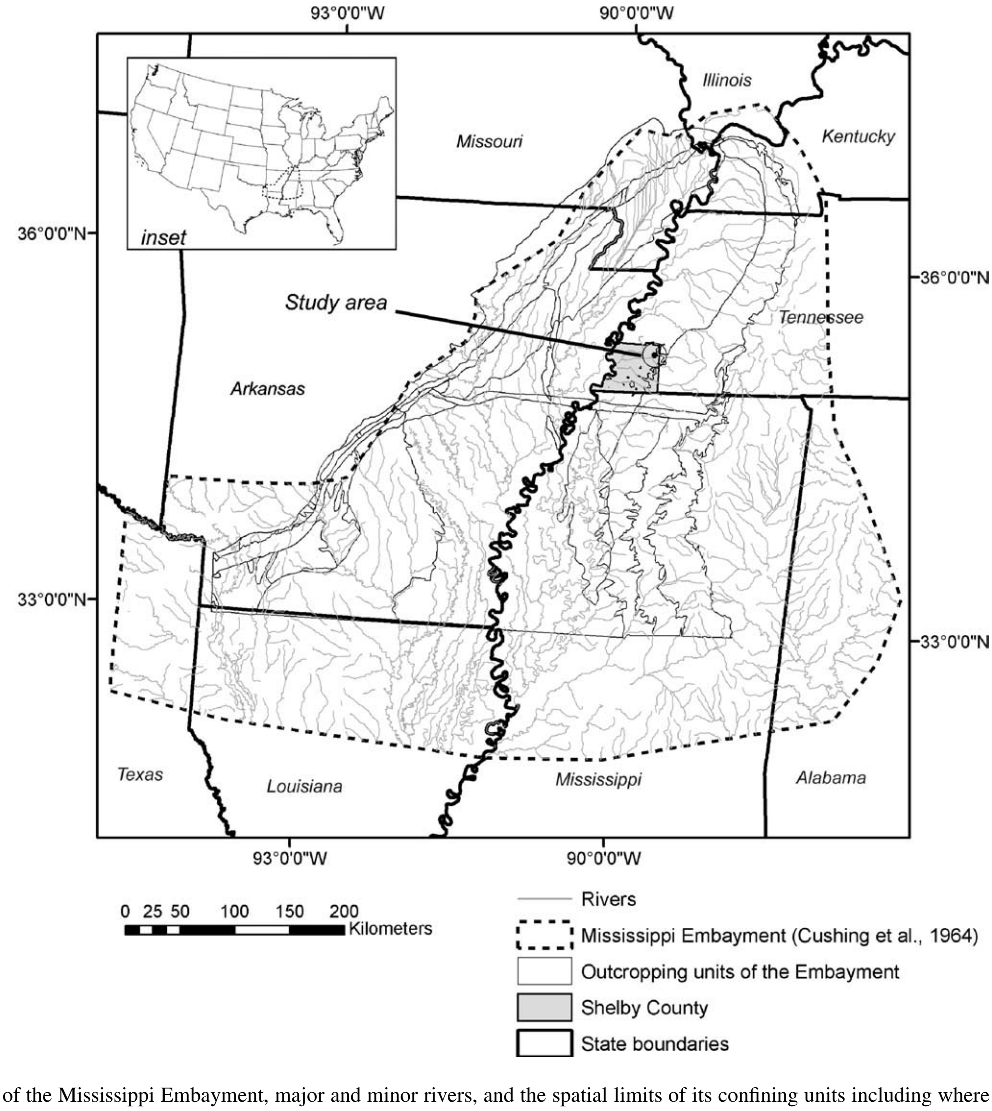

Fig. 1. Map of the Mississippi Embayment, major and minor rivers, and the spatial limits of its confining units including where the confining units do not outcrop at the surface.

particular aquifer, although their analysis did not consider the effects of confining units on these interactions.

Most previous studies of the effects of hydrogeology on groundwater–surfacewater (gw/sw) interactions have focused on specific river reaches or small basins. Vaux (1968) and Woessner (2000), for example, have shown that streambed geometry, such as the pattern of pools and riffles, is a large determinant of ground and surface water exchange within the hyporheic zone. Most theoretical studies of gw/sw interactions have been limited to two-dimensional cross section or plan view at the river reach scale (Sophocleous, 2002; Barlow et al., 2000). Regional studies, focusing on the mechanisms of gw/sw interactions, such as Sklash and Farvolden (1979) and Brunke and Gonser (1994), have shown the effects of transient gw/sw interactions in response to precipitation events, particularly the effects of bank storage and subsurface interflow (Beven, 1989; Whiting and Pomeraneto, 1997). However, the spatial and temporal extent of these effects on the groundwater system tends to be restricted to areas close to the stream. Winter (1999) and Sophocleous (2002) have written excellent recent reviews of the current state-of-the-science of gw/sw interactions.

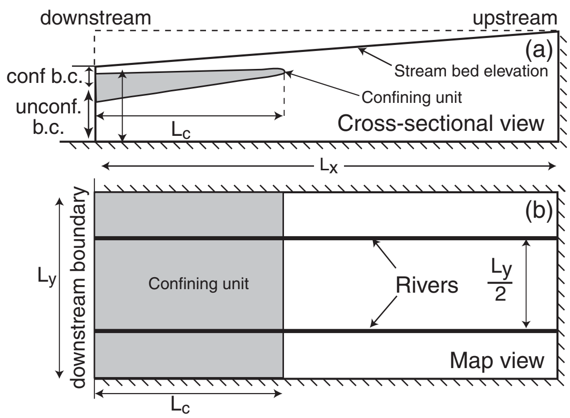

Fig. 2. Dimensions and boundary conditions of model representing the flank of a shallow sedimentary basin. Hash marks indicate no-flow boundaries. (a) Cross-section of model showing the hydrogeology of the conceptualized unconfined–confined transition zone, and the two vertical segments of the downdip boundary condition. (b) Map view showing the locations of the two stream channels.

This paper describe the development of a numerical model using the integrated finite difference code SECOFLOW_3D (Knupp, 1996), to investigate the regional pattern of groundwater discharge to streams at the unconfined–confined aquifer transition. A three-dimensional model was constructed to be representative of the hydrogeology of a gently dipping, sedimentary basin margin that is dissected by streams flowing parallel to the dip of the beds (Fig. 2). We use a simplified, constant head approximation of the downstream boundary to isolate the effects of hydrogeology on groundwater discharge to the stream and develop a theoretical model of gw/sw interactions at the unconfined–confined aquifer transition zone. We then relax the downstream boundary condition to examine the range of impacts that may result from different outflows from the confined aquifer. The theoretical spatial pattern of groundwater discharge is compared to stream flow measurements from the Loosahatchie River of western Tennessee.

2. Methods

2.1. The numerical model

2.1.1. Governing equation

The theoretical, numerical model is based on a three-layer sedimentary system consisting of a tapering confining unit that is sandwiched between two aquifers (Fig. 2). Two stream channels, located for symmetry, run parallel to the dip of the sediments, with flow in the downdip direction.

The numerical model, SECOFLOW_3D (Knupp, 1996), solves the general equation (Eq. (1)) for groundwater flow within the model;

$$\nabla K \nabla h = S_s \frac{\partial h}{\partial t} \tag{1}$$

where K is the hydraulic conductivity [LT^{-1}] and S_s the specific storage [L^{-1}] of the modeled units, while h is the hydraulic head [L], and t is time [T].

2.1.2. Upper boundary condition

SECOFLOW_3D differs from most groundwater models because it uses an adaptive numerical mesh with a free surface to determine the elevation of the water table (z_{wt}) using the kinematic (Eq. (2)) and head (Eq. (3)) boundary conditions (Knupp, 1996).

$$(K\nabla h + R)\cdot\nabla(h - z) = \omega\frac{\partial h}{\partial t} \tag{2}$$

$$h_{wt} = z_{wt} \tag{3}$$

where R is the recharge rate [LT^{-1}], z is elevation [L], and ω is the porosity at the water table (wt). The model solves the kinematic boundary condition (Eq. (2)) iteratively with the head boundary condition (Eq. (3)) to determine the elevation of the water table. The kinematic boundary condition is a statement of conservation of mass at the water table (Corbet and Knupp, 1996), whereas Eq. (3) simply states that the hydraulic head at the water table (h_{wt}) is equal to the elevation of the water table (z_{wt}). If the water table rises to the level of the land surface, in this case the bed of the stream channel, the model creates a constant head boundary (a seepage face) at the elevation of the channel bed (z_{ls}), which is expressed as,

$$h_{wt} = z_{ls} \tag{4}$$

The approach taken in SECOFLOW_3D is advantageous compared to most other groundwater models such as MODFLOW (MacDonald and Harbaugh, 1988), because it allows the water table to 'find' the land surface, in this case the streambed, instead of requiring that the gw/sw contact be pre-defined. The result is a more accurate numerical representation of areas of direct gw/sw contact and thus a more accurate upper model boundary. Free and moving boundary models are, however, more computationally intensive, and subject to numerical instability than their fixed grid counterparts.

2.1.3. Lateral and bottom model boundaries

The model domain is defined to allow three sides and the bottom of the model to be no-flow boundaries by assuming hydrogeological limits in the case of the bottom and updip boundary, and symmetry for the north and south boundaries. The downdip boundary ($x=0$) represents the confluence of the modeled streams with a large river or lake, so it is set at a constant head for most of the simulations. Large surface water bodies may not penetrate to the depth of the confined aquifer. We test the range of potential variability in a series of simulations where the downdip boundary for the upper aquifer is held constant while the boundary for the lower aquifer is changed.

2.1.4. Model simulations

SECOFLOW_3D uses a moving upper boundary, where the head or kinematic boundary condition can apply to any cell at the top of the domain depending on other hydrogeological parameters. To avoid issues of numerical stability the simulations achieve steady-state conditions using a transient simulation. In our numerical experiments, the water table is initially flat and below the level of the two stream channels. Diffuse, uniform recharge is applied to the water table and the free surface rises. In the cells beneath the stream, the water table rises until it reaches the level of the streams, at which point the free surface changes into a constant head boundary. Because the stream channels are of constant gradient and slope toward the left (constant head) boundary, the rising water table does not necessarily reach the bottom of the stream channels across the entire length of the model. In the majority of the simulations conducted during this study, the water table intersects the stream channel along approximately 90% of the channel length. The model design assumes that in the river reaches where the water table is below the streambed, there is no loss of stream flow to the water table. Thus all stream flow is derived from groundwater rather than overland flow. This assumption is appropriate at long times after precipitation events where baseflow dominates the stream hydrograph (Fetter, 1994, p. 53). In model cells away from the streams, the water-table surface remains a free surface and adjusts upward or downward until it approaches steady state.

2.2. Hydrogeologic setting

2.2.1. Hydrogeology

Stream discharge measurements made at selected locations along the Loosahatchie River in Eastern Tennessee were compared to the modeled results. Measurements were made along the stream in the area bracketing the position of the unconfined–confined transition of the Memphis aquifer in western Tennessee (Memphis Sand of Claiborne Group of Eocene age) that is inferred by 1990 (Fig. 3).

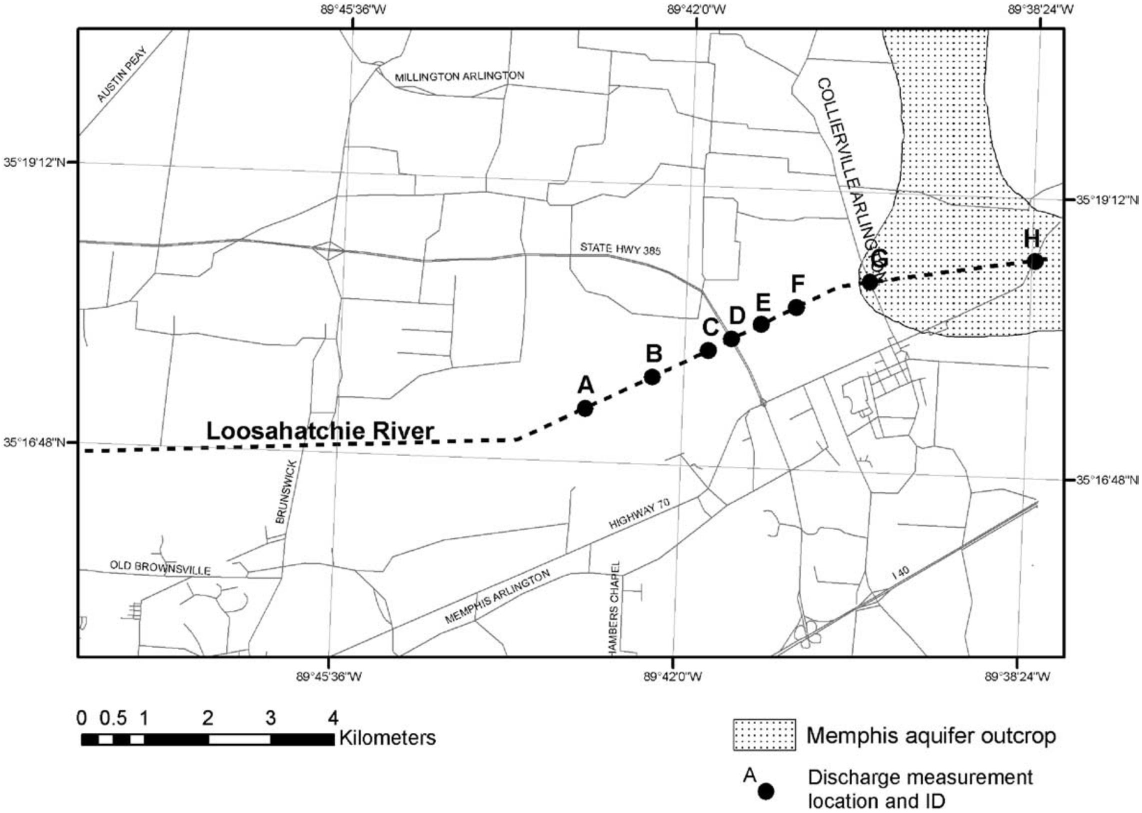

Fig. 3. Stream flow discharge measurement locations along the Loosahatchie River of western Tennessee. The unconfined area of the Memphis aquifer is shaded and the area of subcropping upper Claiborne confining unit is unshaded.

The headwaters of the Loosahatchie River are in the unconfined region of the Memphis aquifer and the river flows west from the headwaters for approximately 30 km before discharging into the Mississippi River. In Western Tennessee, the Memphis aquifer dips gently to the west, toward the axis of the Mississippi Embayment. The Mississippi Embayment is a syncline that plunges gently southward toward the Gulf of Mexico, and its axis parallel to the present course of the Mississippi River (Kingsbury and Parks, 1993; Hosman et al., 1968).

The subcrop of clay in the Upper Claiborne confining unit that separates the overlying watertable aquifer from the underlying Memphis aquifer marks the transition of the Memphis aquifer from unconfined to confined. The inferred unconfined region of the Memphis aquifer is indicated by the hatched area in Fig. 3 (Parks, 1990). Updip of the transition, the Loosahatchie River is in direct contact with the Memphis aquifer. Downdip of the transition, the Memphis aquifer is confined or semi-confined and the Loosahatchie River is only in direct contact with the water-table aquifer that is primarily composed of alluvial and fluvial deposits of Pleistocene through Holocene age (Fig. 3). The unconfined region of the Memphis Aquifer is indicated by the hatched area in Fig. 3 Parks (1990).

The Loosahatchie River was channelized by the US Army Corps of Engineers, and the river remains in a steep-sided engineered channel that is approximately 8 m deep through the investigation area. Under baseflow-dominated conditions the river covers most of the approximately 30 m width of the channel to an average depth of less than 40 cm along the measured reach. Alternating sand bars control flow such that the deepest part of the river, the thalweg, meanders from side to side within the straightened valley.

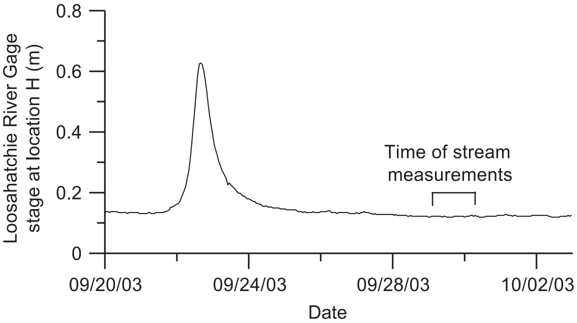

Fig. 4. Stream stage measured at the USGS gaging station on the Loosahatchie River near Arlington, TN. (location H in Fig. 3).

2.2.2. Stream flow measurements

Stream flow measurements were made at locations approximately 500 m apart along a 7 km reach of the stream that bracketed the unconfined–confined transition. Locations F through A (Fig. 4) were measured on a single day (September 30, 2003), while locations G and H were measured the following day. The US Geological Survey (USGS) gaging station on the Loosahatchie River near Arlington, Tennessee, is located at station H, and it showed no change in stream stage during the time period over which measurements were made. The gaging station measurements indicate that flow in the stream during this time was probably exclusively derived from baseflow.

Flow was measured using the velocity-area technique with standard USGS Price AA current meters integrated with Aquacalc™ data recorders. All measurements, except for that made at G, were conducted in collaboration with USGS personnel who rated measurement conditions as good, thus indicating no more than 5% standard error for each measurement. The USGS software MEASERR (Sauer and Meyer, 1992), which computes the standard error for individual discharge measurements using a root-mean-square error analysis of the individual component errors and considers measurement conditions and methodology, indicates that the standard error ranged from 4.1 to 4.8%. The pattern of stream discharge along the reach is compared to the pattern predicted by the base model.

2.3. Model parameters and base simulation

For purposes of comparison we run an initial, base simulation to identify a basic pattern of groundwater discharge to the surface at the unconfined–confined transition. The physical properties and dimensions of this model were chosen to permit a general comparison to the unconsolidated sediments of the field area in the Mississippi Embayment. The base scenario was simulated to maximize the contrast between the aquifer and confining unit by setting the hydraulic conductivity of the confining unit equal five orders of magnitude less than the conductivity of the aquifer ($K_a/K_c=10^5$). The aquifer (Fig. 2) had uniform hydraulic conductivity ($K_c=K_a=1.0\times10^{-4}$ m/s), specific storage ($S_s=1.0\times10^{-4}\ \mathrm{m}^{-1}$) and porosity ($\phi=0.25$). The storage and porosity terms are necessary because the model determines steady state by simulating transient flow over a long time (20 years). A uniform recharge rate, R, of 2.0×10^{-8} m/s was used in the model.

The model domain (Fig. 2) was created with length (L_x), width (L_y) and thickness of $12{,}000\times4000\times100$ m. It is modeled with a $60\times40\times20$ numerical mesh. Two parallel, simulated river valleys were placed 2000 m apart. The downdip ($x=0$) boundary in the base simulation is set at a constant head (50.4 m) for the entire thickness of the sediment.

Sensitivity tests were performed to examine the effects of changing the physical properties, such as hydraulic conductivity and recharge, and the numerical characteristics of the model domain. Changing physical properties affected the magnitude of the ground water discharge to the surface but did not change the relative spatial pattern of discharge. Greater mesh refinement, as well as some coarsening, did not affect the outcomes of the simulations.

3. Results

3.1. Physical properties of the confining unit

3.1.1. Effect of hydraulic conductivity of the confining unit

To examine the effect of confining unit characteristics on groundwater discharge to the stream, a series of simulations were run varying the ratio of the

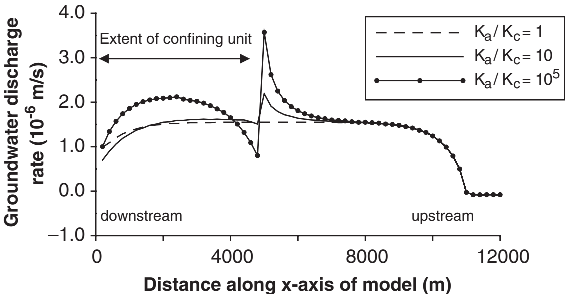

Fig. 5. Groundwater discharge to the stream with different conductivities of the confining unit (K_c). The conductivity of the aquifer is constant for all three simulations ($K_a = 10^{-4}$ *m/s*).

hydraulic conductivity of the aquifers to the confining unit (K_a/K_c). When there is no confining unit ($K_a/K_c = 1$), groundwater discharge per unit area in the stream channel increases rapidly near the stream's headwaters and then remains constant until it approaches the outlet (Fig. 5). The groundwater input to the stream remains constant because of the rectangular model watershed and the constant slope imposed on the stream. This geometry results in the size of the groundwater capture zone for each reach within the area of constant discharge being of the same size and shape. As the stream approaches the downstream boundary of the model space, groundwater discharge to the stream decreases as water in the single aquifer is captured by the constant head boundary.

Introducing the confining unit results in a substantial change in the pattern of discharge to the stream (Fig. 5). An asymmetric peak in the groundwater discharge to the stream occurs immediately upstream of the edge of the confining unit. As the stream crosses over the subcrop of the confining unit, an asymmetric dip in groundwater discharge to the stream occurs before it rises again. In the simulation with a relatively small conductivity contrast ($K_a/K_c = 10$), discharge downstream of the transition returns to the sustained discharge rate observed updip of the transition. For higher K_a/K_c ratios, the discharge rate immediately downstream of the transition decreases more significantly before rising to a much higher sustained rate relative to the upstream rate. A decrease in discharge to the stream is again seen at the downstream edge of the model resulting from groundwater capture by the downdip constant head boundary condition.

To identify the cause of the groundwater discharge fluctuations at the edge of the confining unit, we used particle tracking to determine the recharge areas of three reaches of the stream (Cushing et al., 1964). Under steady-state conditions, the groundwater discharge to these stream reaches is directly proportional to the sizes of their capture zone, as the only fluid input to the model is spatially uniform recharge.

The peak in groundwater discharge is a result of localized groundwater mounding downdip of the transition edge. This mounding is caused by the lower permeability of the confining unit, which impedes recharge and creates a much thinner unconfined aquifer above the confining unit compared to the thicker unconfined aquifer east ($x > 5000$ m) of the confining unit. A thinner aquifer of the same hydraulic conductivity requires a greater hydraulic gradient, and thus a higher water table divide, to transport the diffuse interfluvial recharge to the streams. The water table mound results in shallow unconfined flow counter to the regional flow direction. This is observed in the capture zone for the river segment at the transition (Fig. 6(b), Point 2) where water is captured from the unconfined aquifer both updip and downdip of the transition. This capture zone can be compared to the smaller capture zone of the upstream reach (labeled (3) in Fig. 6), which is reflective of the conditions in which no confining unit is present (Fig. 5 shows a comparison of discharge with and without the confining unit).

3.1.2. Gradational change in conductivity of the confining unit

The edge of the subcropping confining unit simulated in the previous section is modeled as an abrupt edge. In many cases gradational changes exist between aquifers and confining units as a result of either intertonguing finer and coarser grained units or, incised valley or channel deposits that cut through fine-grained deposits (Castle and Miller, 2000). Fig. 7 shows the groundwater discharge patterns to the stream when the confining unit is modeled as having a transitional conductivity from that of the aquifer media (10^{-4} m/s) to that of the clay (10^{-9} m/s). Two simulations are conducted, the first where K_c is linearly decreased and the second

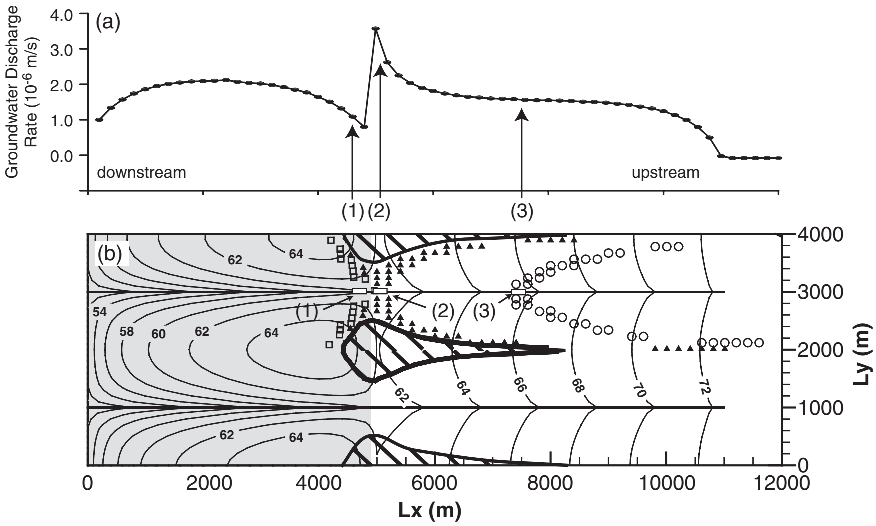

Fig. 6. Relation of groundwater discharge and capture zones along 200 m long reaches in the modeled stream. (a) Groundwater discharge along stream channel where the confining unit has a constant, low hydraulic conductivity ($K=1\times10^{-9}$ m/s). (b) Contour map of water table showing capture zones for three reaches of the stream and the confined aquifer. Reach (1) is located at the dip in groundwater discharge and its capture zone is represented by open squares. Reach (2) is located at the groundwater discharge spike and its capture zone is shown by solid triangles. Reach (3) is located in a region of relatively constant discharge, and its capture zone is represented by open circles. The recharge area for the confined aquifer is indicated by hatching.

where K_c is logarithmically decreased over a 2000 m distance, with the more conductive material present at the transition edge ($K_c=K_a$). The characteristic increase in groundwater discharge at the transition decreases by nearly 30% for the logarithmic scenario, as compared to an abrupt transition, and slightly more (36%) for the linear case. The point of discharge flexure migrates downgradient across the transition zone with the furthest longitudinal movement represented in the linear transition results. In both scenarios, the localized increase in discharge to the stream is followed by a localized decrease (Fig. 7).

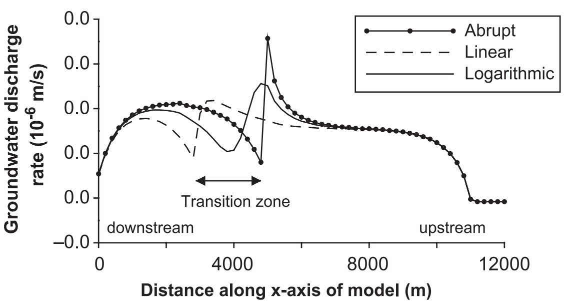

Fig. 7. Effects of a gradational change in confining unit conductivity on groundwater discharge to the stream compared to an abrupt confining unit edge. The transition zone indicates the distance over which there is a linear or logarithmic increase in the hydraulic conductivity of the confining unit, from 10^{-9} m/s at the downstream edge to 10^{-4} m/s at the upstream edge.

3.1.3. Fingering at the transition zone

A generalized representation of fingering of the aquifer into the confining unit, as is representative of river entrenchment into the confining clay is depicted in Fig. 8. The replacement material is assumed to be equal to K_a. Such fingering along the Loosahatchie

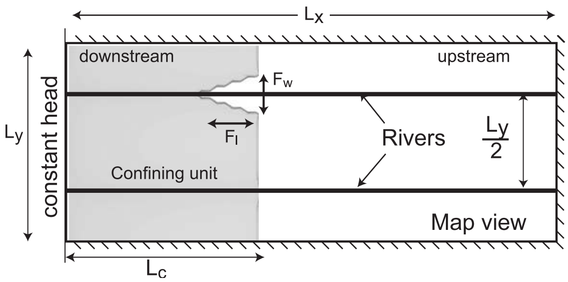

Fig. 8. Map view of a finger of aquifer intruding into the confining unit along a stream. F_l is the finger length and F_w is the finger width.

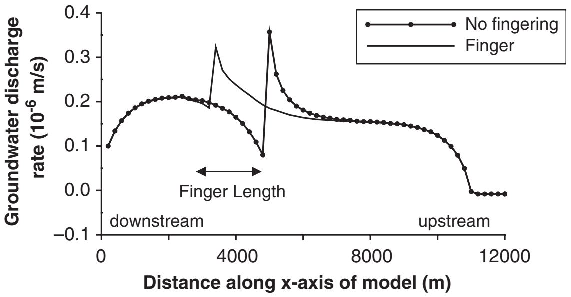

Fig. 9. Groundwater discharge pattern at the stream with fingering of aquifer material into the clay compared to the pattern without a finger.

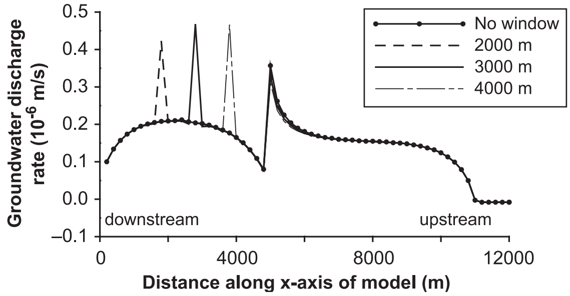

Fig. 10. The effect of breaches in the confining unit beneath the stream on the groundwater discharge pattern to the stream. Breaches are located at 2000, 3000 and 4000 m from the edge of the model. The edge of the confining unit is at 5000 m. The discharge pattern without a breach is also shown.

River may be inferred from Parks (1990) (it is indicated by the hatched area in Fig. 3).

Three simulations were performed where the length of the simulated fingers (F_l) is 1600 m and the maximum width (F_w) varied (100, 700 and 1500 m). The aquifer-confining unit transition itself is modeled as an abrupt change. The results indicate that the peak in ground water discharge, relative to the fingerless scenario, migrated to the downstream edge of the finger. The magnitude of the spike is similar for the three shapes, although the localized dip after the discharge spike becomes slightly more elevated (5%) as the width of the finger decreases. Model results did not significantly differ based upon the shape used so only the result for $F_w = 1500$ m is shown in Fig. 9.

3.1.4. Localized breaches in the confining unit

The last variation in the confining unit characteristics is the fragmentation or slight feathering of the confining unit. For this scenario, a breach in the confining unit is inserted 1600 m downgradient of the abrupt transition. This small, elliptical breach is approximately 200 m in length and 100 m wide and may be representative of localized river entrenchment, fragmentation in onlapping fine-grained, sediments or localized faulting. The hydraulic conductivity through the breach is equal to K_a.

The breach has its greatest impact on the interaction between ground and surface water when located directly beneath the river. When the breach is laterally distant from the river (1000 m) but in close proximity to the edge of the transition zone, its only impact is a slight elongation of the recovery process from the subsequent decrease after the point of transition. With the window beneath the river, the groundwater discharge noticeably increases (Fig. 10) both at the transition zone and more so at the breach (position 1000–2000 m downgradient); discharge at the breach is 31% higher than that at the transition. Therefore, as the position of the breach is migrated downstream, the magnitude of the groundwater discharge at the breach initially increases as it coincides with the rising pattern (following the characteristic dip) before decreasing (position 3000 m downgradient). Irrespective of the location of the breach, there is very little effect on discharge to the streams at the edge of the confining unit.

3.2. Downdip boundary condition

In many situations the downdip, constant head, boundary may apply only to the surface aquifer, and be disconnected from the confined aquifer. In the field area, the downdip boundary represents the Mississippi River, and there is great uncertainty regarding the degree of interaction that exists between the river and the confined aquifer. To examine the potential significance of this boundary we created two variants of the base simulation with different boundary conditions for the confined aquifer, while keeping the same constant head boundary for the water-table aquifer. In the first scenario, the head at the confined aquifer boundary is lowered by 10.4 m (to 40.0 m) from the base simulation. In the second scenario the confined aquifer boundary is made a no-flow boundary.

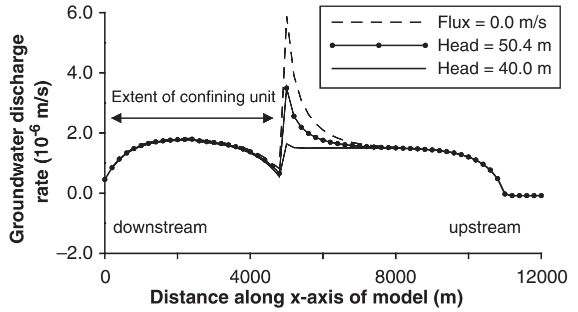

Fig. 11. The effect of the confined aquifer boundary condition on the pattern of groundwater discharge to the stream at the downdip edge of the model. In the base simulation the unconfined and confined aquifer boundaries are set at a constant head of 50.4 m. The confined aquifer boundary is changed to a constant head of 40.0 m and a no-flow.

The 10.4 m lower head boundary in the confined aquifer is equivalent to a line of extraction wells at the boundary, which increases demand for water to recharge the confined aquifer. With an extremely low-permeability confining unit the source of recharge to the confined aquifer is localized at the edge of the transition zone. The increased demand for water in the confined aquifer reduces water available for discharge to the streams, eliminating the peak in discharge at the edge of the confining unit (Fig. 11).

A no-flow boundary in the confined aquifer increases the head in the confined aquifer to the maximum possible without introducing water from outside the domain. Without an alternative exit route, all surface recharge updip of the unconfined–confined transition zone must exit the model at the streams. As a result, the discharge peak at the edge of the transition zone increases substantially above the base simulation in this scenario (Fig. 11).

3.3. Stream flow measurements

3.3.1. Stream discharge rate

All model simulations demonstrate that the presence of a confining unit results in systematic changes in groundwater discharge to the modeled stream. Stream flow can be computed by the model as the sum of upstream groundwater discharge at each point along the stream. In the model scenario where there is no confining layer ($K_a/K_c=1$) the stream gains

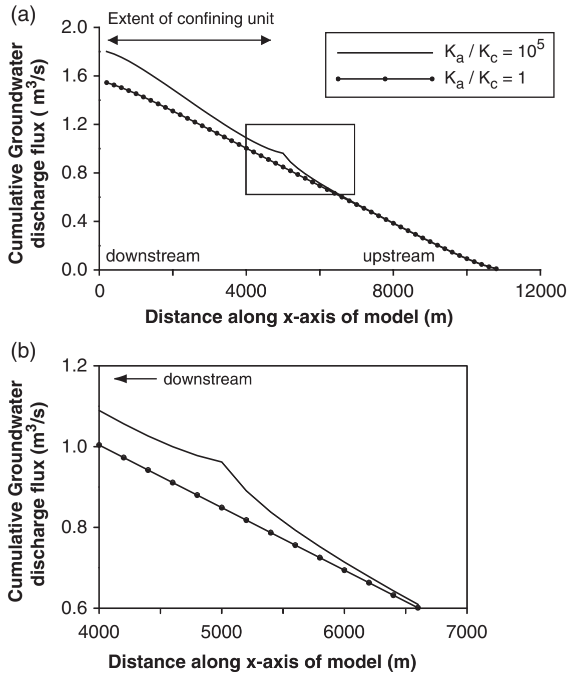

Fig. 12. (a) Modeled stream discharge with ($K_a/K_c=10^5$) and without ($K_a/K_c=1$) a confining unit. (b) Magnified view of the pattern of flow near the transition zone for qualitative comparison to measurements in the Loosahatchie River.

at a constant rate (Fig. 12). With a confining unit present ($K_a/K_c=10^5$; Fig. 5), the peak in groundwater discharge near the edge of the confining unit results in a steepening of the stream flow curve while the reduction in groundwater discharge after the peak causes a flattening of the stream discharge curve (Fig. 12). The modeled stream gains groundwater throughout.

Stream flow measurements were made at eight locations along a reach of the Loosahatchie River, bracketing the inferred subcrop of the unconfined–confined transition (Parks, 1990) of the Memphis aquifer in western Tennessee (Fig. 3). Stream flow measurements were made during September 2003 under base-flow dominated stream conditions as demonstrated by data from the USGS stream flow gage at location H (Fig. 4).

The stream flow measurements show an abrupt increase in flow that may be coincident with the approximate location of the edge of the confining unit (Fig. 13), which generally agrees with the

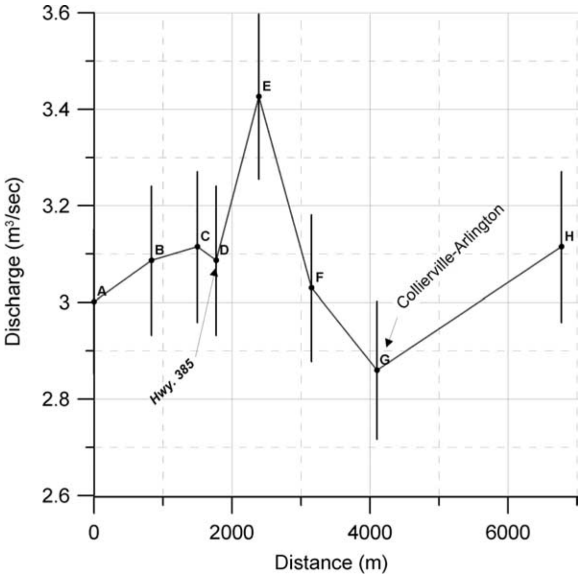

Fig. 13. Stream discharge rates measured in the Loosahatchie River in September 2003. The error bars indicate a 5% standard error range for the discharge measurements. Locations of discharge measurements are shown on Fig. 3.

model prediction. The measured discharge at location E is outside the estimated measurement error of the measurements at locations D and F. This increase in groundwater discharge is only 10% above the average discharge updip of the transition. However, downstream of this location, the Loosahatchie River returns to its previous flow rate, which implies that the stream is loosing water to the groundwater system. This loosing reach is in contrast to the base model results, which indicate a continuously gaining stream, with an increase in flow of as much as 10% above the average gain in flow before stream flow gain pattern returns to its original trajectory (Fig. 12).

4. Discussion

The stream flow discharge measured in the Loosahatchie River shows significant fluctuations in groundwater discharge into the stream that is likely to be related to the unconfined–confined aquifer transition. The model simulations indicate a clear increase in groundwater discharge into the stream immediately upstream of the edge of the confining unit, which results in an overall increase in stream discharge. The modeled increase appears to correlate to the discharge peak measured at location E on the Loosahatchie River. The stream discharge data, however, differ from the model results in two ways.

The first difference between the model results and the discharge data is a small decrease in discharge that was measured upstream of the discharge spike (location G). One possible explanation for the difference is that lenses of clay from the confining unit exist beneath the river upstream of the location of the suspected primary edge of the confining unit. Presence of such a clay lenses at shallow depth could locally inhibit transfer of ground water to the stream. Another possibility for this decrease could be measurement error.

The second and more substantive difference between the model results and stream flow discharge measurements is the decrease in stream discharge downstream of location E (Fig. 13). Downstream of the inferred unconfined–confined transition zone, the stream flow measurements indicate that the stream actually loses water to subsurface. The hydrogeology proximal to the transition point along the Loosahatchie River is not fully understood, and none of the modeled confining unit configurations show a losing stream downgradient of the headwaters of the streams. One possible explanation is that the substantial groundwater pumping from Memphis Sand aquifer by the City of Memphis extracts creates a loosing stream where the aquifer outcrops, however, the analogous simulation, where the head boundary condition of the confined aquifer is 10.4 m lower than the head boundary of the unconfined aquifer, does not show this behavior. Instead the discharge peak upstream of the confining unit edge is reduced, while the lower groundwater discharge downstream of the confining unit edge remains positive and relatively unchanged (Fig. 11).

Local water table pumping near the stream was not modeled but could explain the discharge pattern measured in the Loosahatchie River, but we observed no evidence of this. No significant withdrawals from the shallow aquifer are known in this area. Land use in the local area is primarily agricultural and the stream flow measurements were made outside of the known irrigation schedule.

A more likely explanation for the discrepancy between the field measurements and model results is the simplified gw/sw interactions in the model; specifically the lack of a feedback that would reintroduce water into the groundwater system because of induced stream level changes. The stream is modeled as a constant-head boundary equal to the elevation of the bottom of the sloping stream channel. This simplified boundary condition was used to isolate the effects of the hydrogeology on the groundwater discharge pattern to the stream, but ignores the possible effects of this discharge on stream stage that may result in return flow into the groundwater system.

Recent modeling of combined ground water and surface-water flow at a drainage basin scale by Sudicky et al., (2003) shows complex patterns of influent and effluent stream behavior resulting from individual rain events and subtle topographic variations. Such complexities may contribute higher frequency temporal and spatial variations in stream discharge that were not considered in our model.

4.1. Implications of results

4.1.1. Stream geomorphology

The observed fluctuations in groundwater discharge and the changes in the groundwater flow system seen in our modeling results may affect the geomorphology of the streams. Larkin and Sharp Jr. (1992) demonstrate a correlation between stream geomorphology and the direction of groundwater flow relative to the channel slope. They determined that groundwater flow parallel to stream flow corresponds to shallower channel gradients, greater sinuosity, small width to depth ratios, and deeper river penetration into an aquifer. Given the changes in groundwater flow patterns and groundwater discharge to the surface, the unconfined–confined transition zone is likely to be an excellent place to investigate the effect of groundwater discharge and water table elevation on stream and floodplain geomorphology.

4.1.2. Recharge areas for confined aquifers

Another interesting aspect of the model results shown in Fig. 6 is that the capture zone for water entering the confined aquifer lies largely on the interfluve between the streams and is in the area close to the edge of the confining unit. Thus, for the model geometry considered here, the gw/sw interactions appear to have little bearing on water resources in the adjacent confined aquifer system. This observation may be important for considerations of source-water protection and artificial-recharge systems associated with confined aquifers adjacent to unconfined zones.

4.1.3. Subsurface geology identification

Model results show that near surface subcropping of a confining unit has a noticeable effect on groundwater discharge to a stream. Identification and mapping of the subcrop unit can be costly should nearby geologic data not be available. Conducting detailed discharge measurements along a stream in proximity to the suspected subcrop may prove to be a beneficial and cost effective means for identifying or constraining the characteristics of the tapering edge of a confining unit.

5. Conclusions

Numerical modeling of ground-water/surface-water interactions at the transition of an aquifer from unconfined and confined conditions shows an abrupt, spatially limited increase in groundwater discharge to a gaining stream. Simulations involving variable material properties and geometry of the confining unit indicate that the general pattern of the groundwater discharge peak and decline persists; however, the magnitude and lateral extent of the discharge fluctuations depend on the shape and material properties of the confining unit. Greater hydraulic conductivity contrasts between the confining unit and the aquifer result in larger groundwater discharge fluctuations, while gradational changes in confining unit properties smooth the discharge fluctuations over longer distances. Fingering of aquifer and confining-unit sediments results in a dislocation of the groundwater discharge fluctuations with a small decrease in their magnitude. Therefore, in stream reaches overlying the unconfined–confined transition, the properties and geometry of the confining unit are of equal importance to those of the aquifer in regards to ground-water/surface-water interaction.

Stream-discharge measurements from the Loosahatchie River in western Tennessee provide some evidence to support these modeled results, but indicate that further refinement of the numerical model may be required to capture the ground-water/surface-water interactions at the unconfined–confined aquifer transition. Further investigation into this phenomenon will increase our understanding of ground-water/surface-water interactions and possibly allow improved identification of aquifer transition zones beneath stream systems via the conducting of detailed discharge measurements. The measured and modeled fluctuations in discharge and the reorientation of the groundwater flow system near the edge of the confining unit may also affect the geomorphology of streams and floodplains crossing the transition zone by increasing flow in the stream and maintaining a shallow water table in the floodplain.

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