Published in Journal of Hydrology 376, issues 1-2, 159-169, 2009
which should be used for any reference to this work

Spatial and temporal aspects of the transition from connection to disconnection between rivers, lakes and groundwater

Philip Brunner [a,*], Craig T. Simmons [a], Peter G. Cook [b]

[a] Flinders University, GPO Box 2100, Adelaide SA 5001, Australia
[b] CSIRO Land and Water, Gate 5, Waite Road, Urrbrae SA 5064, Australia

SUMMARY

A changing groundwater table can affect the flow regime between surface water and groundwater. Quantitative approaches are required to understand how such changes influence the flow regime. While in a fully connected regime changes in the groundwater table are related linearly to changes of the infiltration rate, in a disconnected regime the infiltration rate is essentially independent of such changes. However, transitional regimes between connected and disconnected regimes also exist and we identify different transitional pathways. We illustrate how hydrological parameters determine the pathway of transition and show that spatial variations of the infiltration flux through a surface water body strongly depends on these different pathways of transition. Moreover, the spatial distribution of seepage through a surface water body is shown to depend on the state of connection. We also show that the transition from a connected to disconnected flow regime may require a significant drop of the groundwater table. The study demonstrates that the transition zone may be of greater importance than is usually acknowledged. A comparison of lakes and rivers reveals that the latter are less likely to disconnect in response to a decrease of the regional groundwater table. We relate this behavior to differences in the build-up of a groundwater mound in 2D and 3D. Finally, we carry out some simple transient simulations for 3D systems to analyse the transient behavior of surface water groundwater interaction in the context of disconnection. We show that the state of connection is a critical variable in the dynamics of infiltration in a non-steady system.

Keywords
Surface water groundwater interaction, Losing streams, Disconnection, Infiltration rates

Introduction

Water table decline as a result of overexploitation of underground water resources has become a worldwide phenomenon (Konikow and Kendy, 2005). Clearly, these changes in water table affect surface water bodies and the interaction between the two compartments needs to be quantified. A wide body of literature has been published in the context of surface water groundwater interaction and a good review of the state of the art is given by Sophocleous (2002). However, an assumption made in many approaches in quantifying the seepage flow of a surface water compartment such as a stream, a lake or a wetland, is that the flow is saturated. As pointed out in Osman and Bruen (2002), Fox and Durnford (2003) and Brunner et al. (2009), the flow regime under the surface water body can become unsaturated. In this case, methods based on the assumption of saturated flow may lead to incorrect predictions of the seepage fluxes.

If the flow between a losing surface water body and groundwater is fully saturated, then changes in the position of the regional water table will affect the infiltration flux. This flow regime is called a *connected* regime. However, if the flow is unsaturated, then the situation can develop where the infiltration rate is effectively independent of changes of the groundwater table. This flow regime is called a *disconnected* regime. Because the different flow regimes result in different relations between infiltration flux and groundwater table position, their identification is required to understand the implications of a changing groundwater table. Even though the interaction between surface water and groundwater has been studied extensively, the subject of disconnection has received relatively little attention. Not all surface water–groundwater systems can disconnect. The first requirement is that a low permeability layer (herein referred to as the *clogging layer*) separates the surface water body from the aquifer. However, the presence of a clogging layer is a necessary but not sufficient criterion. Brunner et al. (2009) determined an exact criterion for predicting whether or not a particular system can become disconnected. The relevant variables are the ratio of hydraulic conductivity of the clogging layer to the underlying aquifer as well as the thickness of the clogging layer and the depth of the surface water body.

Previous studies e.g. Fox and Durnford (2003) have identified three stages of connectivity: connected, transitional, and

* Corresponding author. Tel.: +61 8 82012346.
E-mail address: philip.brunner@flinders.edu.au (P. Brunner).

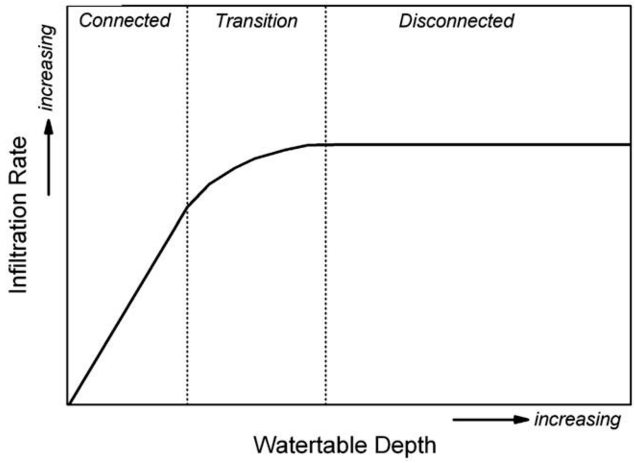

Fig. 1. Change of infiltration rate from surface water to groundwater as a function of the water table depth (e.g. measured in a nearby borehole). Three different flow regimes can be identified. Only in the connected state is the infiltration rate between the surface water body and the groundwater proportional to the head difference. In the transition zone, the flow rate is no longer a linear function of the head difference. If the groundwater table is sufficiently far below the surface water body, the infiltration rate asymptotically approaches a constant value and changes in the groundwater table no longer significantly affect the infiltration rate. In this final regime the surface water and the groundwater are disconnected.

disconnected (Fig. 1). The change from connected to disconnected is induced by lowering the groundwater table. Full disconnection occurs when the infiltration rate is no longer a function of the head difference between the surface water and the groundwater. In the connected regime, however, the infiltration flux is often approximately proportional to the head difference. It is important to note that the infiltration flux in a disconnected system is the highest possible. Desilets et al. (2008) noted that the few papers that do talk about disconnection limit their discussion to either a fully connected regime or a fully disconnected regime. By assuming that only these two states exist, it is implicitly assumed that the transition zone is small. This assumption is integrated in the architecture of commonly used modelling packages (e.g. MODFLOW) where the width of a stream or a river is associated with at most one grid cell and therefore the transition has no horizontal component. However, as we will illustrate in this study, the transition zone can be very large and cannot be neglected a priori.

Brunner et al. (2009) presented a theoretical framework to allow for the assessment of losing streams without the limiting assumption of full saturation. Infiltration beneath a losing stream will lead to the build-up of a groundwater mound. It was shown that disconnected systems can be conceptualized in terms of the height of this mound and the capillary zone above it. This conceptualization can be used to determine the critical water table position where full disconnection is reached (point of disconnection). This critical water table position can be calculated for any distance from the surface water body. In principle, a comparison of this calculated critical water table position with a measurement of the water table depth in a nearby borehole allows for the assessment of the status of connection. Brunner et al. (2009) also determined the sensitivity of the point of disconnection to relevant hydrological parameters. However, their analysis was restricted to cross sections of straight rivers. Regional flow was not considered, the analysis was steady state and homogeneous and isotropic conditions were assumed. There have been no papers dealing with disconnection of lakes. Compared to rivers, lakes are controlled by a different dimensionality and different geometrical features.

In this paper, we expand the analysis of the physics of disconnection and study spatial and temporal effects as well as the transition from connected to disconnected surface water–groundwater systems. We identify different transitional pathways for systems changing from a connected to disconnected flow regime in response to a dropping groundwater table. In analogy to the 2D (river) analysis of Brunner et al. (2009), we provide a sensitivity analysis of the point of disconnection to the relevant hydrological parameters in steady state for a 3D system (lake) and also report the critical groundwater table where the system changes from connected to transitional. Reporting both the point of disconnection and the beginning of transition allows us to quantify the sensitivity of the extent of the transition zone to the system parameters. Moreover, we illustrate how hydrological parameters determine what pathway of transition will occur in response to a decline of the groundwater table and we show that spatial variations of the infiltration flux through a surface water body strongly depend on these different pathways of transition. A comparison of circular surface water bodies with rivers reveals that circular systems are much more likely to disconnect. We relate this behavior to differences in the build-up of a mound in 2D and 3D. In order to analyse the transient behavior of surface water–groundwater interaction in the context of disconnection, we carry out some simple transient simulations for 3D systems. We illustrate the effect of a step change of the groundwater table and also simulate the influence of a changing depth of the surface water body on infiltration rates and the state of connection.

Conceptual model

Brunner et al. (2009) showed that disconnected systems can be conceptualized by the build-up of a groundwater mound and a capillary zone above it. The capillary zone was defined to be the region above the water table where saturation is a function of depth. Theoretically, the capillary zone is of infinite extent because the pressure and saturation above the water table approach their minimum values asymptotically. For both practical and comparative purposes, Brunner et al. (2009) defined the top of the capillary zone as the height above the water table where the pressure is within 0.1% of the minimum possible value. The system is disconnected when the top of the capillary zone is below the base of the clogging layer. A 1D analysis is sufficient to predict the extent of the capillary zone of a disconnected system. The infiltration flux that occurs in a disconnected system is required to calculate this extent. This flux can also be calculated using a 1D analysis. On the other hand, a 2D or 3D analysis is required to determine the groundwater mound.

If a fully saturated flow model is used to determine the point of disconnection, three steps are required: (1) calculate the infiltration flux that occurs at disconnection; (2) determine the extent of the capillary zone under this infiltration flux; (3) calculate the groundwater mound where the distance between its the highest point and the clogging layer equals the extent of the capillary zone. Even though in a fully saturated/unsaturated model these three steps do not have to be carried out explicitly, we described them in more detail because they are important for the subsequent analysis.

The infiltration flux of a disconnected system needs to be quantified in order to calculate the height of the capillary zone. This flux can be calculated according to methods described in Osman and Bruen (2002). In disconnected systems, the flow regime between the clogging layer beneath the surface water body and the groundwater table is unsaturated. The flux through this unsaturated zone must be equal to the infiltration flux through the clogging layer. This flux is given by the head gradient through the clogging layer and its hydraulic conductivity. Clearly, the negative pressure at the base of the clogging layer that occurs at disconnection increases the head gradient through the clogging layer. At the same time, the hydraulic conductivity of the aquifer (K_a) is reduced by the negative pressure. These two conditions (gravity flow through the aquifer and continuity of flow through the clogging layer and the aquifer) allow the calculation of the infiltration flux q_{max} of a disconnected system. q_{max} corresponds to a minimum possible value of suction at the base of the clogging layer. This minimum value of pressure is required to identify the height of the capillary zone.

Once the infiltration flux q_{max} is calculated, the height of the capillary zone can be quantified. In Brunner et al. (2009), the height above the water table where the saturation is within 0.1% of the minimum possible value (at the particular infiltration rate) was defined as the top of the capillary zone. The relation between height and pressure can be calculated by finely discretising the Richards' equation and solving it for the specific flow rate. This definition of the capillary zone should not be confused with the standard definition of the capillary fringe used in soil physics. (The capillary fringe is only the height above the water table at which the suction is equal to the air entry value.)

Fig. 2 illustrates the effect of infiltration flux on the height of the capillary zone. Profiles of saturation for sand under different infiltration fluxes are shown. Where there is no water flux (not shown), the saturation continues to decrease as the height above the water table increases. Under a constant flux, however, the saturation approaches a constant value. This value is dependent on soil parameters as well as the magnitude of the infiltration flux. As the infiltration flux increases, the height at which a constant value of saturation is reached (which defines the top of the capillary zone) decreases.

In a 1D system, the extent of the transition zone is equal to the height of the capillary zone. In two and three dimensions, a groundwater mound develops under the recharging surface water body, and therefore different states of connection can occur at different points. This manifests itself as an extended transition zone. We define a surface water body as *fully connected* if it is connected at all points (and hence saturated conditions are present at every point beneath it) and *fully disconnected* if it is disconnected at all points. Any state between fully connected and fully disconnected is called *transitional*. If the surface water body can disconnect, then the lowering of the groundwater table will cause the surface water

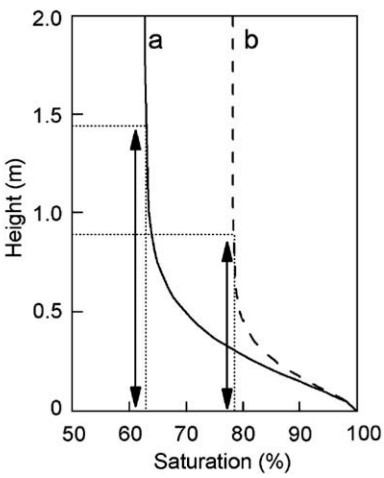

Fig. 2. Saturation versus height above the water table for a soil under vertical infiltration rates of: (a) 0.1 and (b) 0.5 m d^{-1}. The height of the capillary zone (indicated by the arrows) is defined as the height above the water table where the saturation is within 0.1% of the minimum possible value. Thus, for a flux of 0.1 m d^{-1}, the height of the capillary zone is 1.45 m, while for a flux of 0.5 m d^{-1} it is 0.88 m. The saturated hydraulic conductivity is 12.5 m d^{-1}. The van Genuchten parameters used are α = 3.6 m^{-1} and β = 1.56.

body to initially disconnect at the edge. If the water table is lowered further, the unsaturated zone will propagate towards the centre of the surface water body.

In principle, two different pathways of disconnection are possible: (1) the surface water body remains connected in the centre while the edge is disconnected and (2) both the centre as well as the edge of the system are in transition at the same time. Whether a system follows the first or the second pathway depends on its hydraulic and geometric properties. Fig. 3 shows stages between full connection and full disconnection for these two different pathways: the two principle pathways are shown in Fig. 3a–e and f–j, respectively. Let us consider Fig. 3a–e first where the centre remains connected while the edge is disconnected. Initially, the water table is beneath the surface water body, but the flow remains saturated at every point (Fig. 3a). As the water table is lowered, flow at the edge of the surface water body becomes unsaturated, but the capillary zone still intersects with the clogging layer (b). In the next stage, the edge of the surface water body is fully disconnected (i.e., the depth of the water table below the clogging layer exceeds the height of the capillary zone), but the system remains connected in the centre (c). In the fourth stage, the flow is unsaturated beneath the entire surface water body, but the capillary zone in the centre of the surface water body still intersects with the clogging layer (d). In the final stage, the capillary zone is below the base of the clogging layer throughout the lake (e), and the system is fully disconnected.

The second possible pathway is shown in Fig. 3f–j. Here, the edge is in transition while the flow in the centre becomes unsaturated (and therefore moves into transition). Whether a particular river or lake will follow stages a–e or f–j will depend largely on its size and on the height of the capillary zone. The importance of the pathways will be discussed later.

Numerical modelling

Numerical modelling is required to determine the shape of the groundwater mound beneath a losing surface water body, and hence the progression from a connected to a disconnected system. First, we use a numerical model to demonstrate on a quantitative basis the different stages of transition. The model allows us to calculate the critical water table position which defines the transition points between the different stages of connection, and also to analyse the sensitivity of these critical water table positions to the relevant hydraulic parameters. Simulations are carried out for both rivers and circular surface water bodies. The circular surface water bodies could either be a lake or a wetland. However, in the following sections we only use the term lake. We also use the model to calculate spatial variations of the infiltration flux and to simulate transient effects of changing boundary conditions on the state of connection.

Model setup

The model setup is shown in Fig. 4. A circular lake (Fig. 4a) was simulated using a 30° slice of an axisymmetric system. A small number of simulations were also carried out for a river (Fig. 4b) to enable a comparison between lakes and rivers. The river setup is identical to that used by Brunner et al. (2009). For both setups, a clogging layer with thickness h_c and hydraulic conductivity K_c separates the aquifer from the surface water body. The surface water body itself has a depth d and is defined as a constant head boundary. The river model is set up with a finite difference numerical scheme and the circular surface water body with finite elements. A grid convergence analysis was carried out in both cases to ensure that the grid spacing did not affect the results. Vertical

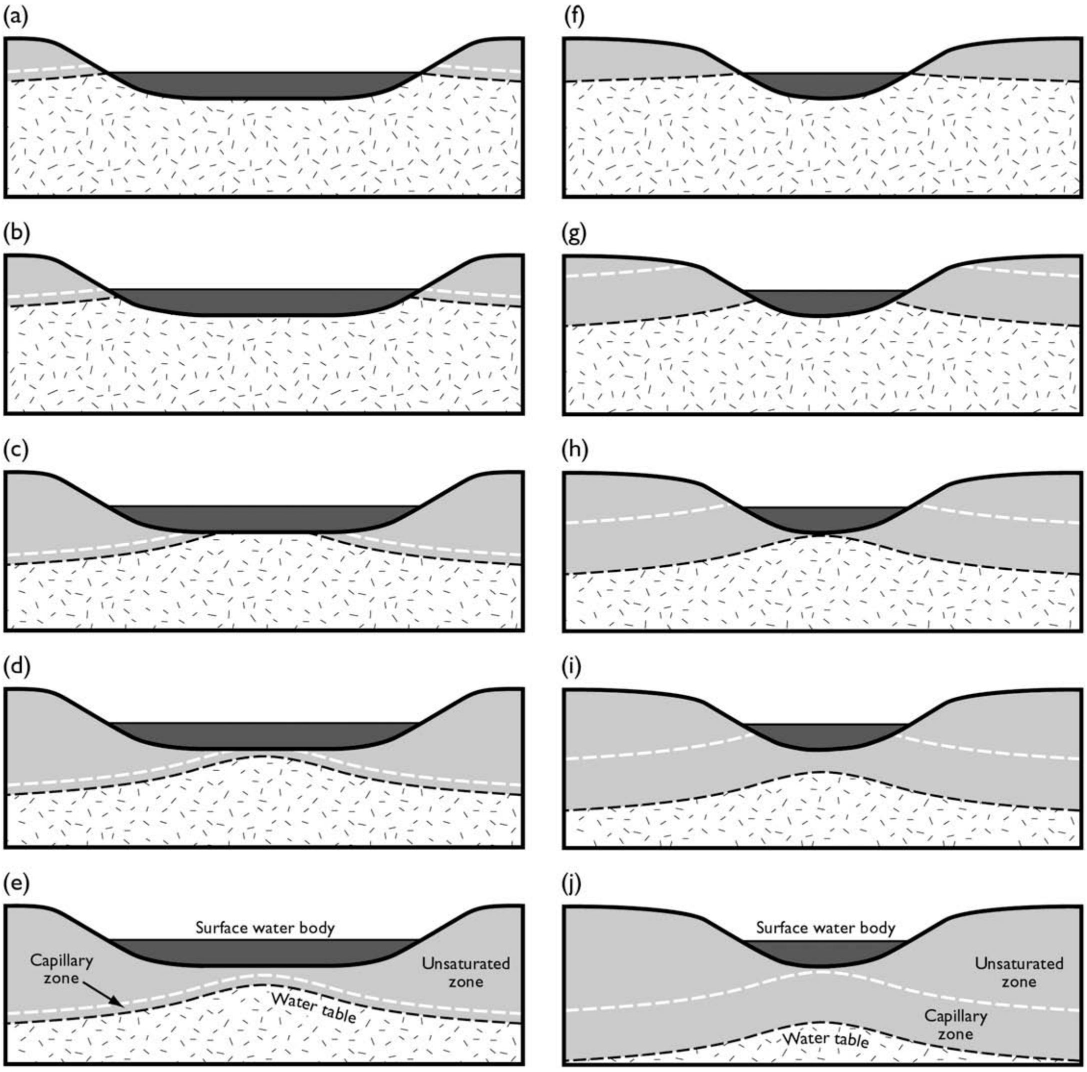

Fig. 3. Possible stages of connectivity between a surface water body and the groundwater for two different systems: (a)–(e) represent system 1, (f)–(j) system 2; (a) and (f) show connected regimes while (e) and (j) are both disconnected. The remaining diagrams indicate transitional stages. In system 1 shown in (a)–(e), the edge of the lake becomes fully disconnected while the centre remains saturated. In system 2, the centre of the surface water body is in transition before the edge becomes fully disconnected.

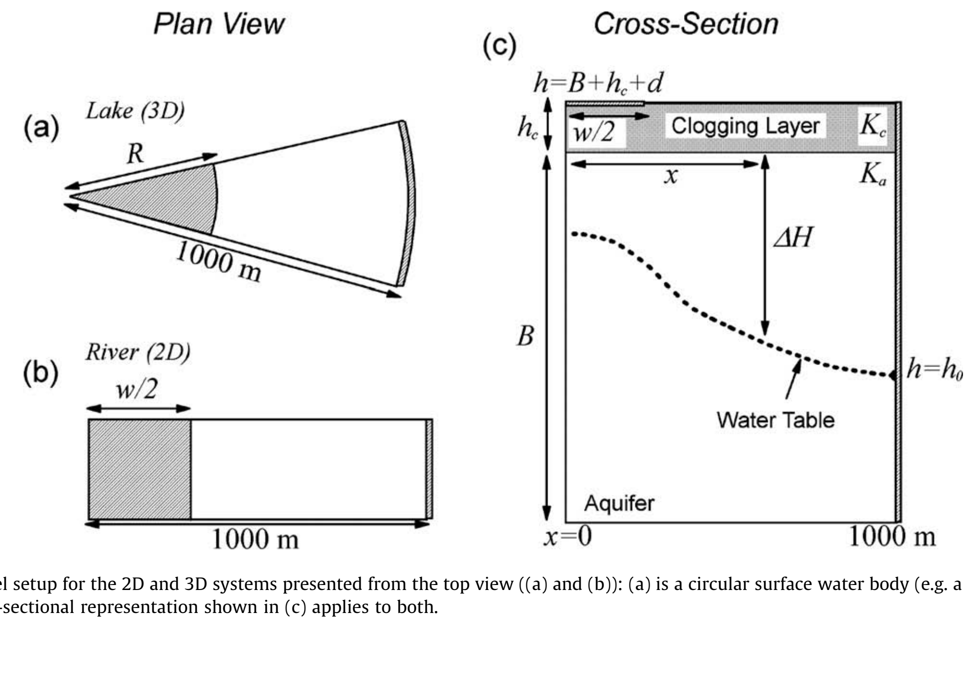

Fig. 4. Conceptual model setup for the 2D and 3D systems presented from the top view ((a) and (b)): (a) is a circular surface water body (e.g. a lake or a wetland) and (b) a straight river. The cross-sectional representation shown in (c) applies to both.

discretization must be high in the region below the clogging layer where unsaturated flow can occur. In order to simulate the point of disconnection correctly, this fine discretization must at least cover the extent of the capillary zone that occurs in full disconnection. Therefore, the extent of the capillary zone was calculated as outlined in section two and the required extent of fine discretization was identified for each simulation. A vertical discretization of at most 0.05 m was applied over the vertical distance of the capillary zone. Constant head conditions ($h = h_0$) were used for the lateral boundaries ($x = 1000$ m) and a no flow boundary was used at the base of the aquifer. By defining a constant head at the lateral boundary, we are essentially imposing a horizontal flow condition at this point. It is therefore important that the location of this boundary is far away from the infiltration layer to ensure that the vertical flow component under the mound is not significantly affected by it. We found that for the range of model parameters we tested $x = 1000$ m was a sufficient distance. All simulations were carried out using the groundwater flow model HydroGeoSphere (HGS) (Therrien et al., 2006). HydroGeoSphere is a three dimensional numerical model describing fully-integrated subsurface and surface flow and solute transport. HydroGeoSphere's capability to simulate disconnection problems was demonstrated in Brunner et al. (2009) by verifying its capability to accurately simulate the build-up of a mound as well as the capillary zone and by comparing with appropriate semi-analytical solutions.

To examine the relationship between water table position and infiltration flux, we start with the lateral boundary high enough to allow the regime under the surface water body to be fully saturated (connected). The lateral head boundary is then lowered, and the steady solution calculated for each lateral boundary head. As this head value at the outer edge boundary is lowered, an unsaturated zone first develops at the edge and subsequently extends horizontally towards the centre of the surface water body. If the head boundary is lowered sufficiently, the system becomes fully disconnected. At this point, the distance between the bottom of the clogging layer and the groundwater table in the centre of the lake is equal to the height of the capillary zone. In this approach, we have used the parameter estimation model PEST (Doherty, 2002) to estimate these critical water table positions.

Quantitative illustration of transitional stages

Fig. 5 shows the effect of lowering the boundary conditions for two circular systems with different pathways of transition. We will later show that identifying what path of transition a system will take is important for understanding how the spatial distribution of infiltration fluxes will vary during transition. The mean infiltration flux beneath the surface water body, q, as well as the pressure under the clogging layer at the edge and in the centre of the surface water body is plotted in each case as a function of the vertical distance ΔH between the bottom of the clogging layer and the groundwater table at horizontal distance of $3R$ from the centre (where R is the lake radius). Clearly, ΔH is a function of the location of the observation point, and can be reported for any location between the surface water body and the point where the boundary condition is defined. ΔH will increase as the observation point is moved further away from the edge of the surface water body.

The points of greatest interest are when the system moves from full connection to transition (ΔH_c) and when it reaches full disconnection (ΔH_d). These two points define the extent of the transition zone at the particular location x where ΔH is measured. Fig. 5a and b illustrate a system that follows the transition path from

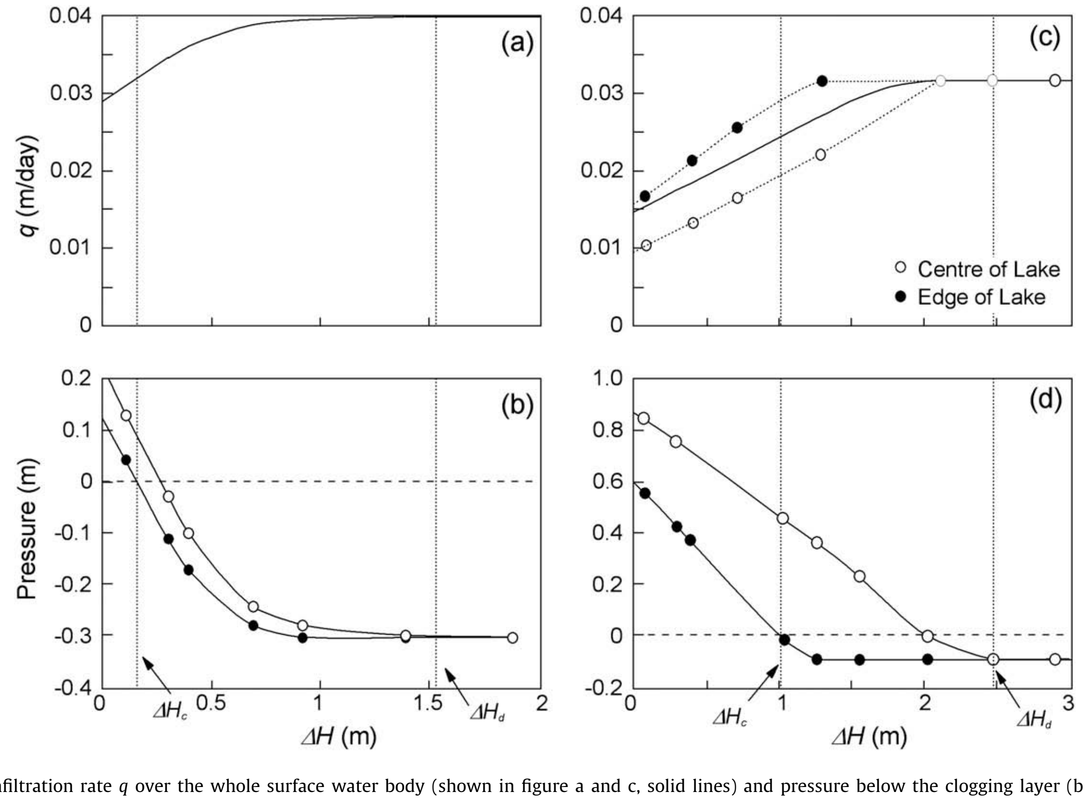

Fig. 5. Average infiltration rate q over the whole surface water body (shown in figure a and c, solid lines) and pressure below the clogging layer (b and d) at the edge (open dots) and the centre (black dots) for different two circular systems. The vertical lines in all graphs indicate when the system reaches transition (ΔH_c) or full disconnection (ΔH_d). The left hand side of the plot (a and b) shows system 2, the right hand side (c and d) system 1. The hydraulic properties for system 1 are: $K_C = 0.01$ m d^{-1}, $h_C = 0.5$ m, $d = 1$m, $K_a = 1$ m d^{-1}, $B = 40$ m, $R = 50$ m. The van Genuchten parameters are for a sand soil ($\alpha = 14.5$ m^{-1} and $\beta = 2.68$). The radius R in system 2 is 10 m and the van Genuchten parameters for the aquifer are for a loam ($\alpha = 3.6$, $\beta = 1.56$). All other parameters are identical to system 1. ΔH is evaluated at $x = 3$R. As well as the average infiltration rate, part c also shows the infiltration rate in the centre of the lake and at the edge. These are not shown in Fig. 5a because they are barely discernible from the average.

connected to disconnected shown in Fig. 3f–j, while Fig. 5c and d illustrate a system that follows the transition path shown in Fig. 3a–e. The transitional stages shown in Fig. 5c and d (corresponding to Fig. 3a–e) are likely to occur in wide systems with a steep groundwater mound and a small capillary zone. Transitional pathways with stages shown Figs. 5a, b and 3f–j are more likely to be found in narrow systems with a flat mound and a large capillary zone. Both types of transition can be found in rivers or lakes. High capillary zones are present in systems with small infiltration rates and soil properties that allow for a high capillary rise. Steep mounds are more likely to occur in aquifers with small hydraulic conductivity and a high infiltration rate.

Consider initially Fig. 5a and b. Lowering the lateral head boundary (increasing ΔH) causes the system to move from connected to transition. Transition is reached at $\Delta H_c = 0.17$ m. This corresponds to the state depicted in Fig. 3g. When ΔH increases beyond 0.27 m, the pressure in the centre drops below zero, and both the edge and centre are in transition (this state corresponds to the state shown in Fig. 3h). The system becomes fully disconnected at $\Delta H_d = 1.52$ m. The changes in pressure are also reflected in the infiltration rate. In Fig. 5a, the average infiltration rate over the lake is plotted. A stable infiltration rate is reached after the centre of the lake fully disconnects.

Consider next Fig. 5c and d. The transitional states shown here correspond to those shown in Fig. 3a–e. The pressure at the edge drops below zero at $\Delta H_c = 1.01$ m (see Fig. 3b for comparison). At this point, the system is in transition. By further lowering the lateral head boundary condition, the edge becomes disconnected (at $\Delta H = 1.25$ m). However, the centre of the lake remains fully connected. The system is fully disconnected at $\Delta H_d = 2.47$ m (pressure and flux in the centre of the lake both stabilize).

Steady state sensitivity analysis

The sensitivity of ΔH_d to relevant hydrologic parameters for river systems has been determined by Brunner et al. (2009). Here, we report the point of disconnection for circular surface water bodies. Additionally, we report the point where the system goes into transition. This allows us to estimate how changes of the parameters influence the extent of the transition zone. For one example, we vary the parameter to an extent that the pathway of transition changes.

The sensitivity of ΔH_c and ΔH_d to the model parameters is shown in Fig. 6. In this Figure, we used the term ΔH_{crit} as a term relevant to both ΔH_c and ΔH_d. All parameter variations were carried out by varying a base case shown in all plots of Fig. 6. The base case corresponds to the system shown in Fig. 5c and d. Both ΔH_c and ΔH_d appear approximately proportional to the hydraulic conductivity of the clogging layer, K_c, and also to the depth of ponding, d. Inverse relationships between ΔH_c and ΔH_d, and K_a and h_c are also apparent. These relationships are independent of the location of the observation point. Similar sensitivities were obtained for a river system by Brunner et al. (2009). The ratio $\frac{K_c d}{h_c}$ is approximately equal to the flow rate through the clogging layer for a system that is not connected, and thus the values of ΔH_c and ΔH_d are approximately proportional to this flux. The relationships between the head difference and the lake radius (R) and aquifer thickness (B) are more complex. ΔH_c and ΔH_d increase as R increases, but the rate of increase is not constant. The value of ΔH_c and ΔH_d decrease as B increases. With an increasing B, however, the rate of decrease is reduced. All proportionalities described above were independent of the location x of the observation point.

In the plot showing the variation of R, the points where the system is disconnected at the edge and where it goes into transition in the centre are also plotted. The point the two lines cross (at $R = 34$ m) indicates a change in the pathway of transition. For any R larger than 34 m, the system is fully disconnected at the edge before transition in the centre starts and the system follows the pathway shown in Fig. 5c and d. However, if the radius for this particular setup is smaller than 34 m the system enters transition in the centre before the edge is fully disconnected. This corresponds to the pathway shown in Fig. 5a and b.

The vertical difference between the points in Fig. 6 showing ΔH_c and ΔH_d represents the extent of the transition zone. Depending on the hydraulic parameters, the transition zone can be of significant extent.

Spatial variations of infiltration fluxes

The spatial distribution of seepage through a lakebed has been studied by numerous authors (e.g. Pfannkuch and Winter, 1984) but to our knowledge the state of connection has never been taken into account in these studies. In Fig. 5c, the average infiltration flux over the lake is plotted. A constant infiltration rate is reached after the centre of the lake disconnects. In addition to the average infiltration, the flux in the centre and the edge of the lake are presented. Because the edge can disconnect while the centre remains connected, the infiltration fluxes vary across the lake during transition. In the system shown in Fig. 5a and b, on the other hand, the pressures at the centre and edge are never markedly different (Fig. 5b), and therefore the infiltration flux at the centre and the edge are similar and closely follow the average infiltration. Fig. 7a shows the spatial variation of infiltration flux across the same lake which is shown in Fig. 5c and d. Lines numbered 1–5 illustrate how the infiltration flux varies between the centre and the edge of the lake for different stages of connection. The different stages of connection are calculated by lowering the lateral head boundary. These simulations are carried out in steady state. By progressively lowering the water table, the system changes from fully connected to fully disconnected. Only if the system is fully disconnected is the infiltration flux is uniform across the lake. Where the system is connected or in transition, the flux varies across the lake. However, the magnitude of the variation is greatest in the transitional stages. In any case, the infiltration flux is greatest at the edge of the lake. Fig. 7b follows the same transitional pathway as Fig. 7a (as per Fig. 3a–e), while Fig. 7c follows the alternative path (Fig. 3f–j). Systems that follow paths 3f–j usually show less spatial variability of flux. It is noteworthy that the magnitude of the spatial variation decreases as the hydraulic conductivity of the aquifer increases. This is because the groundwater mound is flatter, and so the variation in pressure at the base of the clogging layer is smaller.

Circular lakes versus straight rivers

For an idealized straight river, groundwater flow is two dimensional while for a circular lake it is three dimensional. For both two dimensional (2D, straight river) and three dimensional (3D, circular lake) systems the conceptual model employed to assess disconnection is the same (build-up of a capillary zone above a groundwater mound). One obvious difference between a river and a circular lake system is the influence of the characteristic length scale describing the extent of the surface water body. In a river, the area contributing to infiltration is proportional to its width w, while in a lake the area of the surface water body is proportional to R^2. While the extent of the capillary zone is identical for both lakes and rivers (for systems with identical hydraulic conductivity, thickness of aquifer and clogging layer as well as depth of the surface water body), the build-up of the groundwater mound is not. The height of a groundwater mound under a lake is smaller than under a river because the flux is dissipated through the additional dimension which has the effect of reducing the hydraulic gradient. Compared to straight rivers, lakes are therefore more

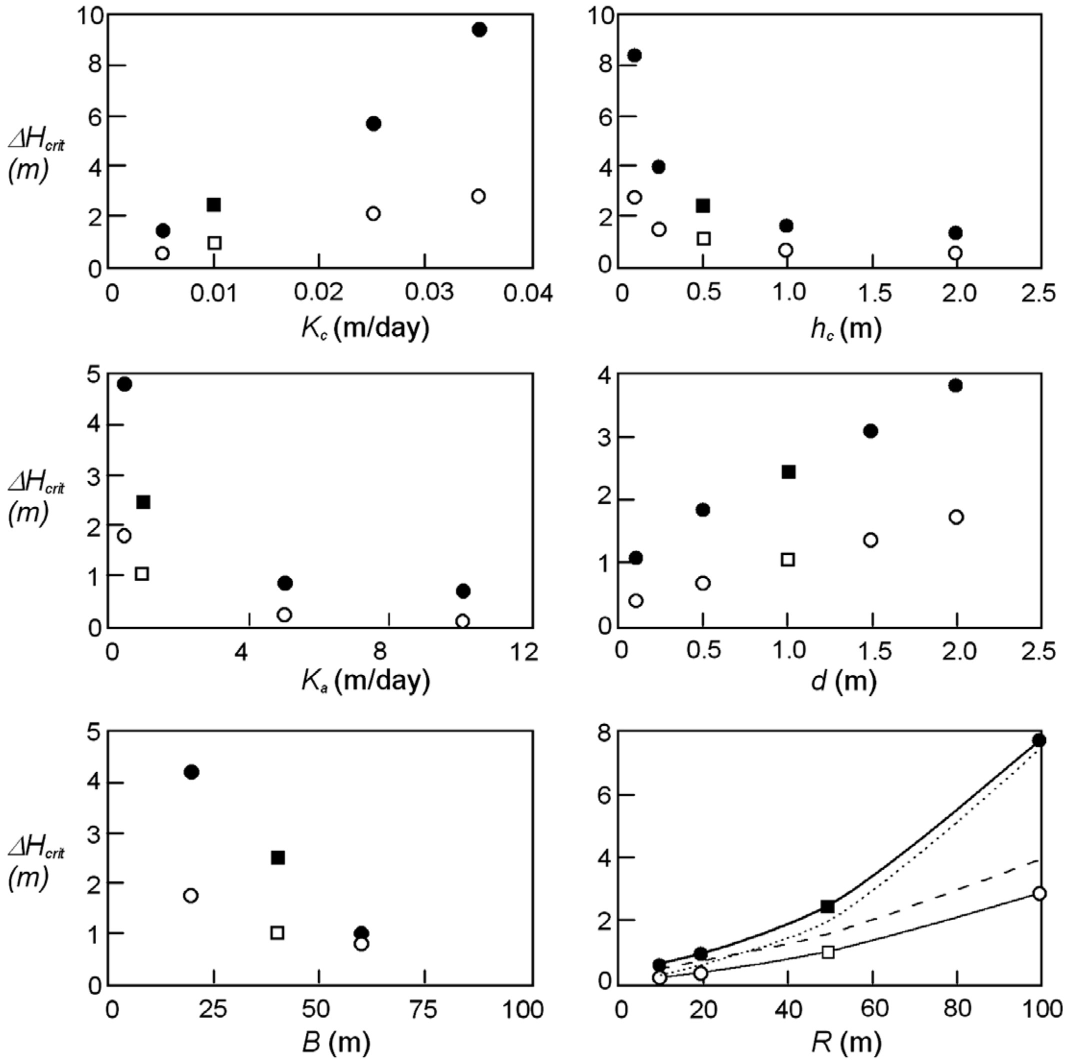

Fig. 6. Sensitivity of ΔH_C (open symbols) and ΔH_d (solid symbols) for a circular surface water body to the hydraulic conductivity of the clogging layer K_c, the thickness of the clogging layer h_c, the hydraulic conductivity of the aquifer K_a, the depth of the surface water body d, the thickness of the aquifer B and the radius of the surface water body R. The water table depth ΔH_C and ΔH_d *is* expressed as the difference between the clogging layer and the groundwater table at distance $x = 3R$ for the two different stages of disconnection. In each case, a single parameter is varied while the other parameters remain unchanged. A base case was chosen and is plotted in every figure (indicated by the squares). The base case of the simulation is $R = 50$ m, $d = 1$ m, $B = 40$ m, $h_c = 0.5$ m, $K_c = 0.01$ m d^{-1}and $K_a = 1$ m d^{-1}. All simulations use van Genuchten parameters for sand. The two additional lines plotted in the frame showing the relation between ΔH_C and R represent the point of full disconnection at the edge as well as the beginning of transition in the centre. The finely dotted line following ΔH_d (solid symbols) shows when the system at the edge become disconnected while the dashed line closely following ΔH_C illustrates when transition in the centre starts. Note that these two lines cross at $R = 34$ m.

likely to become disconnected from the groundwater. This is illustrated in Fig. 8, which shows the difference between ΔH_d for straight rivers and circular lakes with otherwise identical properties (i.e., identical values for K_a, K_c, d, h_c and B). Because the infiltration flux at disconnection is identical for both systems, the thickness of the capillary zone is also identical. ΔH_d is plotted for a river and a circular lake as a function of the characteristic width of the surface water body. It is clear that the river disconnects later than the lake. In other words, if the water table is progressively lowered, then a lake would disconnect before a river of the same characteristic length scale and identical hydraulic properties (i.e., if $w/2 = R$).

Brunner et al. (2009) showed that the point of full disconnection can be calculated by combining analytical solutions (to quantify the build-up of the mound) with the Richards equation (to calculate the capillary fringe). Warner et al. (1989) provided an overview and discussion of commonly used equations to calculate the mound height. However, if the mound is calculated using common analytical solutions such as the Boussinesq equation, the error in predicting the mound height can be significant if steep gradients are present. In this context it is important to realize that compared to rivers, the gradients are lower under lakes. Because mound heights in rivers are larger compared to an equivalent lake system, the error in predicting the groundwater mound for a lake is smaller in comparison to a corresponding river.

Transient simulations

The preceding analyses have been steady state. Here, we carry out transient simulations to investigate how the state of connection influences infiltration in response to changing surface water depth and changes of the groundwater table. Changes of surface water body depth and the regional groundwater table are simulated to understand the transient behavior of connection and disconnection. To avoid repetition, these simulations are carried out for the circular lakes only, but the mechanisms illustrated for lakes are identical for rivers. Changes in the regional groundwater table are simulated by lowering the head at the lateral boundary.

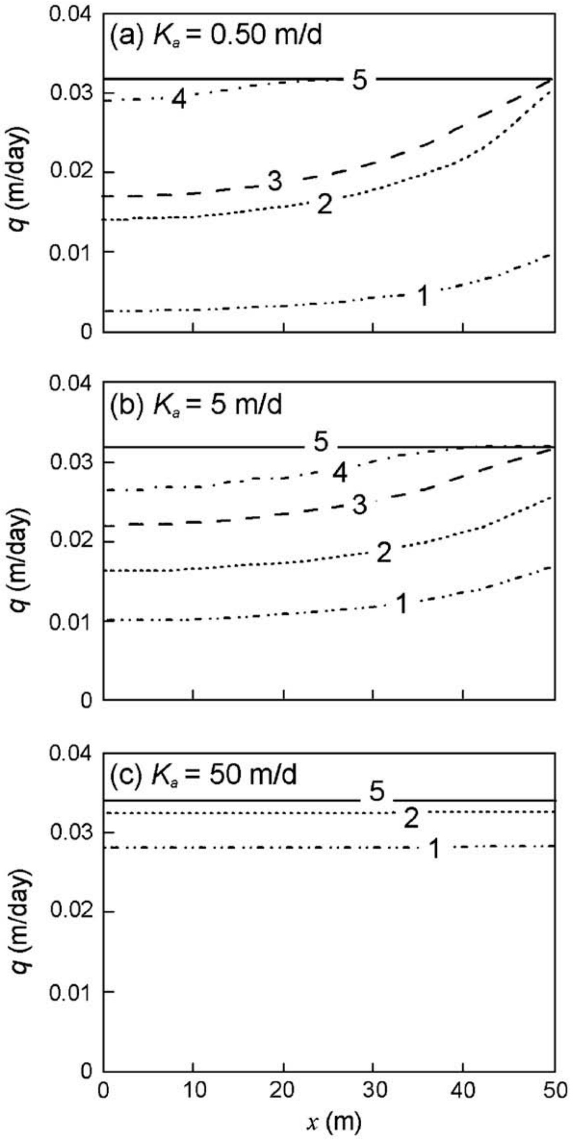

Fig. 7. Changes of infiltration rate across a lake for different states of connection. $x = 0$ is the centre of the lake, $x = 50$ the outer edge. Lines numbered 1–5 show the progression of infiltration fluxes as the water table at the lateral model boundary is lowered. In each case line 1 represents a fully connected system, line 2 is the transition point between full connection and transition, line 3 is the point at which the edge reaches full disconnection, and line 5 is full disconnection. (Line 4 shows a state in between lines 3 and 5, but does not correspond to a particular transition point.) In part c, lines 3 and 4 are not plotted, as they are barely discernible from line 5.

Changes in surface water depth are simulated by varying the constant head boundary representing the surface water body. The variations are imposed as step functions. The transient simulations are carried out until the hydraulic behavior of the systems approximates their new steady state. In a second step, the boundaries are reset to the original positions with the new steady states as initial conditions. The radius of the surface water body, the hydraulic conductivities and the thickness of the aquifer and the clogging layer are identical to the base case shown in Figs. 7 and 5c, d. The porosity is set to 0.1. However, for the simulations representing changes of the groundwater table, the distance where the lateral boundary condition is defined is moved closer to the surface water body (to $x = 200$ instead of $x = 1000$) to ensure that the step change of the boundary results in a significant temporal change under the lake.

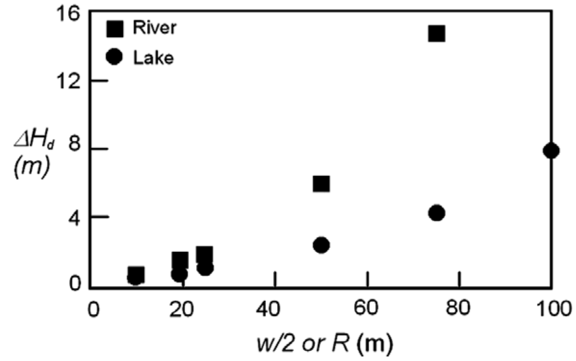

Fig. 8. ΔH_d for a river and a lake plotted versus $w/2$ (river) and R (lake). ΔH_d is measured at $x = 1.5R$ for the lake and at $x = 1.5\ w/2$. The circles indicate a circular lake, the squares the straight river. The parameters chosen correspond to the base case described in the previous section.

Fig. 9a shows a plot of ΔH versus mean infiltration flux (q) after a step variation of the depth of the surface water body from $d = 1$ m to $d = 1.5$ m. The figure also shows the steady state relations between ΔH versus q for a system with $d = 1$ m and with $d = 1.5$ m. The lateral head boundary is kept constant during this simulation and is set to a low value to ensure the system remains disconnected even after the increase of the depth of the surface water. The initial groundwater table in the observation borehole is point 1 in Fig. 9a. The increase in depth results in an almost instantaneous increase in the infiltration flux, while the depth to groundwater initially remains unchanged (point 2). However, the increased infiltration flux eventually leads to a rising groundwater table in the observation borehole. The new steady state of the system (now with a depth of the surface water body $d = 1.5$) is point 3. This steady state is the initial condition for a calculation simulating a decrease from $d = 1.5$ m to $d = 1$ m. This change again leads to an almost immediate reduction of the flux, while the water level in the borehole initially remains unchanged (point 4). Eventually, the reduced flux leads to a decrease in size of the groundwater mound, resulting in an increase of ΔH. Eventually the system stabilizes at the original starting point (1). Points 1–4 define a rectangle vertically bounded by the steady state relationships.

Identical variations of the surface water depth were carried out for a system with the lateral head boundary high enough that the system is initially in a transitional state (point 1 in Fig. 9b). After a step variation of the depth of the surface water body from $d = 1$ m to $d = 1.5$ m the flux almost instantly increases (point 2). In contrast to point 2 in Fig. 9a, the infiltration flux is initially above the steady state solution for $d = 1.5$ m. After the initial increase of flux, however, the system approaches its new steady state (point 3). This new steady state is no longer close to a disconnected state – this is apparent from the location of point 3 of the curve plotting the steady state relationship between ΔH and q for a system with $d = 1.5$ m. The decrease from $d = 1.5$ to $d = 1$ m results in a decrease of flux that again deviates from the steady state solution (point 4) before it approximates its initial state (point 1). The reason why the infiltration flux is not bounded by the steady state solution (in contrast to the disconnected one) lies in the change of the hydraulic gradient through the system as a response to the change of the surface water depth. In a connected system, the hydraulic gradient can be defined through any point of the aquifer and the surface water body. In this case, changes in the surface water body do not immediately result in a new steady state because a certain amount of time is required until the groundwater table responds and a new steady state with the corresponding hydraulic gradient is reached. In a disconnected system, the hydraulic gradient corresponding to the infiltration rate is independent of the groundwater

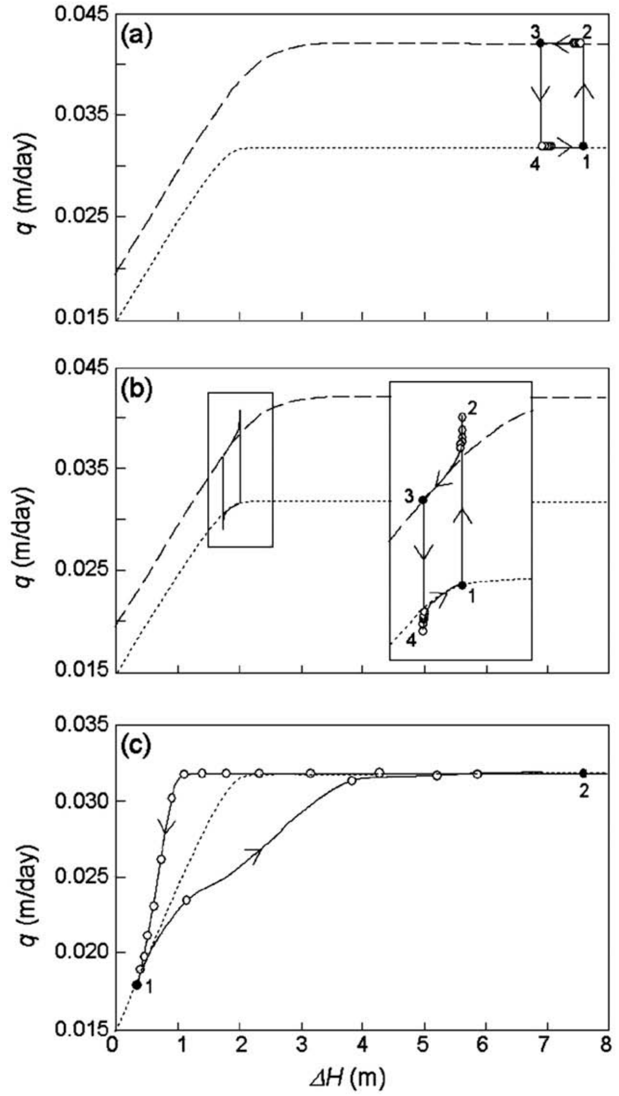

Fig. 9. Infiltration rate of the surface water body q as a function of ΔH after changing the depth d of the surface water body ((a) and (b)) and changing the ambient groundwater table (c). In all simulations, ΔH is the distance between the bottom of the clogging layer and the groundwater table 150 m away ($x = 3\ R$) from the centre of the surface water body. The dotted lines in (a)–(c) represent the steady state relations between ΔH and the infiltration rate for the base case ($d = 1$ m). The dashed lines in (a) and (b) represent the steady state relation for $d = 1.5$ m (all other parameters remain unchanged). The open dots on the solid lines connecting points 1–4 in (a) and (b) and points 1–2 in (c) are equally spaced in time. (The time difference between adjacent dots is 5, 1 and 7 days in (a), (b) and (c), respectively.)

table. It is given only by the gradient through the clogging layer. Therefore, a change of surface water depth in connected systems results initially in an infiltration rate above (in case of an increase of surface water depth) or below (in case of a decrease of the water depth) of the steady state solution.

These simulations show that the state of connection influences the transient behavior in response to a change of the surface water depth. There are key differences between connected and disconnected systems in this respect: the infiltration rate in disconnected systems shows no transient behavior, a change in surface water depth results almost immediately in a new, steady infiltration rate. The infiltration rate in connected systems, on the other hand, approaches the new steady state in a transient way. Also, in connected system an increase of surface water depth initially results in an infiltration above the steady state solution, while a drop initially results in an infiltration rate below the steady state. In disconnected system, the infiltration rate instantly reaches steady state and no deviation is possible.

In a second type of simulations, the head at the lateral boundary is changed while the depth of the surface water body is kept constant (Fig. 9c). The starting point for these simulations is point 1 in Fig. 9b. Initially, the system is in a transitional stage. The rapid drop of the head at the lateral boundary leads to a decrease of the groundwater table as well as an increased infiltration flux. However, as a result of the rapid change, the relation between the depth to groundwater and the infiltration flux deviates from the steady state solution (also plotted in Fig. 9c). The depth to groundwater is clearly larger than ΔH_d while the flux is still is below the q_{max}. If the boundary would have dropped at a much slower rate, the relation between ΔH and q would approximate the steady state function. Nevertheless, a new steady state is reached eventually (point 2). This new steady state was used as initial condition to simulate the increase in the lateral head boundary back to the initial position. In response to this variation, the groundwater table in the borehole changes before any change of the infiltration flux is observed and the relation between depth to groundwater and q initially is a straight line. As time progresses, the rise of the groundwater table propagates to the surface water body and reduces the infiltration flux. Finally, the initial system state at point 1 is reached again. These simulations show the assessment of the state of connection on the basis of the groundwater table in an adjacent borehole can be difficult if rapid drops of the ambient groundwater table are expected. Such an assessment is difficult because the water table measured in the borehole may not correspond to the state of connection in steady state.

Discussion

This is the first paper that analyses disconnection under lakes. The analysis is based on an idealized setup, where the lake is considered to be circular. The hydraulic conductivity of the clogging layer and the aquifer is homogeneous and isotropic. We did not consider the effect of hysteresis in the retention curve in this initial analysis. We have reported the critical groundwater table where such a system changes from connected to transition and from transition to disconnected and have carried out a sensitivity analysis of the relevant hydrological parameters. The sensitivity analysis showed that for a given aquifer thickness and radius, the depth to groundwater where the system disconnects is approximately proportional to the depth of the circular water body and the hydraulic conductivity of the streambed sediments and inversely proportional to the thickness of these sediments and the hydraulic conductivity of the aquifer. The proportionalities for the water table where the system changes from connected to transition are identical. Identifying the relevant parameters and their sensitivities may assist in the design of field investigations aimed at assessing disconnection status, and also assists in extrapolating results of local field studies both temporally and spatially.

The sensitivity analyses also illustrated that the transition zone can be of significant extent. This challenges the commonly made assumption that a system is either connected or disconnected. Knowing the extent of the transition zone is important for two reasons: (1) if a system with a large transition zone is simulated in a numerical model, a fine horizontal discretization of the surface water body is required to adequately reproduce the transition zone. In many commonly used numerical models, however, this is not possible. For example, in the streamflow routing package (Prudic et al., 2004) of MODFLOW the width of a river can only be related to one single grid cell. For many cases, this is appropriate but if the interaction of a river with a wide transition zone is to be modelled, this limitation should be considered in the choice of the numerical model; (2) the extent of the transition zone influences the spatial variation in infiltration flux across a surface water body. This is important in the interpretation of the spatial distribution of seepage rates measured by seepage meters (Lee, 1977) or similar devices.

Genereux and Bandopadhyay (2001) have shown that the infiltration flux to a gaining lake will be greatest at the edge of the lake and will decrease exponentially with distance from the shoreline. For a losing lake, McBride and Pfannkuch (1975) concluded that seepage rates are greatest at the lake shore and decrease with increasing distance from the shore. Our simulations also showed that for connected and transitional losing systems, the infiltration flux will be greatest at the edge and that the magnitude of the variation in flux will depend on the hydraulic properties of the system. However, for disconnected systems the infiltration rate is uniform across the surface water body. The magnitude of spatial variation during transition is related to the pathway of transition a given system follows. In systems where both the edge and the centre can be in transition at the same time, spatial variations of infiltration will be small. To the best of our knowledge it has not previously been demonstrated that the spatial distribution of seepage critically depends on the state of connection and on how the system behaves during transition. Of course, surface water bodies will often have spatially variable hydraulic properties (Calver, 2001) and river beds are usually heterogeneous. A spatial variation of hydraulic conductivity will change the infiltration rates shown in our simulations. However, it is important to recognize that even in heterogeneous flow fields a systematic variation depending on the state of connection may be present.

The comparison between rivers and lakes with equivalent hydraulic properties showed that lakes will disconnect more easily. This is important for large scale water management. For example, consider a river that flows through a floodplain that also contains lakes. A reduction in the regional groundwater table will affect the river and the lake differently. The lake will be disconnected before the river. The fact that infiltration rates of circular systems are more sensitive to changes of the groundwater table has not been noted or explicitly demonstrated in previous literature. This finding also illustrates that identified states of connection for a surface water body cannot be applied to surface water systems with a different geometry. Dimensionality and geometry are clearly critical controls on disconnection that have been elucidated in previous work on this topic.

The transient simulations of variations in water table depth showed that the change of the infiltration rate depends on the state of connection. In connected systems, the infiltration rate initially deviates form the resulting steady state infiltration rates. In disconnected systems, however, the infiltration flux almost immediately equals the new steady state infiltration. The variation of the ambient groundwater table results in a relation between ΔH versus q that deviates from the steady state solution. Transient behavior of the system can therefore pose limitations to the assessment of the state of connection on the basis of borehole observations. If rapid changes of the system are expected, a time series of groundwater level in the observation bore is required.

Our analysis is for a highly idealized system: a horizontal, isotropic, homogeneous aquifer, and a stream of constant depth underlain by a homogeneous isotropic clogging layer of constant thickness and we do not consider regional groundwater flow. However, these simplifications are required to understand the key processes and variables. Further work is needed to examine the effect of irregular 2D or 3D geometry of the clogging layer and aquifer materials, and stream morphology on the system behavior.

Conclusions

In this study, we identified and described different types of transition behavior from connected to disconnected surface water bodies and examined critical spatial and temporal causes and effects. We have based our analysis on numerical simulations using a code capable of simulating fully coupled saturated-unsaturated flow. We did this for idealized geometries and boundary conditions and identified basic principles and mechanisms rather than solutions for specific field problems. The key findings of this study are:

- A significant drop of the groundwater table may be required to change the flow regime from connected to disconnected and we have shown how variations of hydrogeological parameters affect the extent of the transition zone. This challenges the commonly made assumption that a system is either connected or disconnected. If a river with a significant transition zone is simulated in a numerical model, a model should be chosen that allows sufficient horizontal discretization of the river.
- We have shown that the spatial distribution of seepage through a surface water body critically depends on the state of connection and the extent of the transition zone. In a disconnected system, there is no spatial variation of infiltration. For connected and transitional systems, a large transition zone results in an increased spatial variation in infiltration across the surface water body. We have identified two different pathways of transition from connected to disconnected and illustrated that in systems where the edge and the centre are in transition at the same time, the spatial variation of infiltration is small during transition. If the edge becomes disconnected while the centre remains connected, larger spatial variations of infiltration are expected.
- Lakes and wetlands are more likely to disconnect than rivers because of differences in the build-up of a groundwater mound in 2D and 3D. Therefore, lakes are more sensitive to changes of the regional groundwater table. This is of significant importance in assessing the impact of a changing groundwater table on surface water bodies.
- The state of connection is a critical variable for dynamic systems. In a disconnected system, changes of the surface water body result almost immediately in a new steady infiltration rate while in connected systems the new steady state is approached with time. In connected systems, a reduction of surface water depth results initially in an infiltration rate below the steady state solution while an increase of water depth results in an infiltration flux above the steady state solution. If the regional groundwater table changes rapidly, the assessment of the state of disconnection on the basis of a borehole can be biased.

Acknowledgments

This work was partly funded by the South Australian Centre for Natural Resource Management, Flinders Research Centre for Coastal and Catchment Environments, and CSIRO Land and Water. We acknowledge support for this work through funding provided by the National Water Commission, Australia. The authors would like to thank René Thérrien for ongoing and truly unconditional assistance in numerical modelling using HydroGeoSphere. We also want to thank the two anonymous reviewers for their feedback.

We acknowledge collaboration and support provided through a United States National Science Foundation Grant – NSF Project EAR 0609982.

References

Brunner, P., Cook, P.G., Simmons, C.T., 2009. Hydrogeologic controls on the disconnection between surface water and groundwater. Water Resources Research 45, W01422, doi:10.1029/2008WR006953.

Calver, A., 2001. Riverbed permeabilities: information from pooled data. Ground Water 39 (4), 546–553.

Desilets, S.L.E., Ferré, T.P.A., Troch, P.A., 2008. Effect of steam–aquifer disconnection on local flow patterns. Water Resources Research. 44, W09501.

Doherty, J., 2002. PEST – model-independent parameter estimation. Watermark Numerical Computing.

Fox, G.A., Durnford, D.S., 2003. Unsaturated hyporheic zone flow in stream/aquifer conjunctive systems. Advances in Water Resources 26 (9), 989–1000.

Genereux, D., Bandopadhyay, I., 2001. Numerical investigation of lake bed seepage patterns: effects of porous medium and lake properties. Journal of Hydrology 241 (3–4), 286–303.

Konikow, L.F., Kendy, E., 2005. Groundwater depletion: a global problem. Hydrogeology Journal 13 (1), 317–320.

Lee, D., 1977. A device for measuring seepage flux in lakes and estuaries. Limnology and Oceanography 22 (1), 140–147.

McBride, M.S., Pfannkuch, H.O., 1975. The distribution of seepage within lakes. Journal of Research of the US Geological Survey 3, 505–512.

Osman, Y.Z., Bruen, M.P., 2002. Modelling stream–aquifer seepage in an alluvial aquifer: an improved losing-stream package for MODFLOW. Journal of Hydrology 264 (1–4), 69–86.

Pfannkuch, H.O., Winter, T.C., 1984. Effect of anisotropy and groundwater system geometry on seepage through lakebeds 1. Analog and dimensional analysis. Journal of Hydrology 75 (1–4), 213–237.

Prudic, D.E., Konikow, L.F., Banta, E.R., 2004. A new stream-flow routing (SFR1) package to simulate stream–aquifer interaction with MODFLOW-2000. US Geological Survey Open-File Report 2004-1042.

Sophocleous, M., 2002. Interactions between groundwater and surface water: the state of the science. Hydrogeology Journal 10 (1), 52–67.

Therrien, R., McLarren, R.G., Sudicky, E.A., Panday, S.M., 2006. HydroGeoSphere. Groundwater Simulations Group, University of Waterloo.

Warner, J.W., Molden, D., Chehata, M., Sunada, D.K., 1989. Mathematical-analysis of artificial recharge from basins. Water Resources Bulletin 25, 401–411.