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A coupled vertically integrated model to describe lateral exchanges between surface and subsurface in large alluvial floodplains with a fully penetrating river

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Abstract:

This paper presents a vertically averaged model for studying water and solute exchanges between a large river and its adjacent alluvial aquifer. The hydraulic model couples horizontal 2D Saint Venant equations for river flow and a 2D Dupuit equation for aquifer flow. The dynamic coupling between river and aquifer is provided by continuity of fluxes and water level elevation between the two domains. Equations are solved simultaneously by linking the two hydrological system matrices in a single global matrix in order to ensure the continuity conditions between river and aquifer and to accurately model two-way coupling between these two domains. The model is applied to a large reach (about 36 km²) of the Garonne River (south-western France) and its floodplain, including an instrumented site in a meander. Simulated hydraulic heads are compared with experimental measurements on the Garonne River and aquifer in the floodplain. Model verification includes comparisons for one point sampling date (27 piezometers, 30 March 2000) and for hydraulic heads variations measured continuously over 5 months (5 piezometers, 1 January to 1 June 2000). The model accurately reproduces the strong hydraulic connections between the Garonne River and its aquifer, which are confirmed by the simultaneous variation of the water level in the river and in piezometers located near the river bank. The simulations also confirmed that the model is able to reproduce groundwater flow dynamics during flood events. Given these results, the hydraulic model was coupled with a solute-transport component, based on advection-dispersion equations, to investigate the theoretical dynamics of a conservative tracer over 5 years throughout the 36 km² reach studied. Meanders were shown to favour exchanges between river and aquifer, and although the tracer was diluted in the river, the contamination moved downstream from the injection plots and affected both river banks.

KEY WORDS ecohydrology; river-aquifer interactions; transport; coupled model; Garonne River, 2D Saint Venant, 2D Dupuit

INTRODUCTION

River–aquifer systems are composed of dynamic spatial mosaics connected by fluxes of water, matter and organisms (Jones and Mulholland, 2000; Sophocleous, 2002). Hydrological connections between the main river channel and the adjacent aquifer in the floodplain are considered to be essential for the operation and integrity of fluvial hydrosystems (Thoms, 2003). These connections can facilitate the exchange of carbon and nutrients between the river channel and the aquifer. They take place in a subsurface zone composed of riverbed sediments called the hyporheic zone (Triska *et al.*, 1989; Findlay, 1995; Bencala, 2000). Research over the past two decades has established that this zone has a great impact on the productivity of the entire river system (see reviews in Jones and Holmes, 1996; Brunke and Gonser, 1997; Boulton *et al.*, 1998; Sophocleous, 2002). The impact of the

hyporheic zone on stream ecosystem operation is determined by biogeochemical processes and by the proportion of water fluxes exchanged between river and aquifer. In large rivers, the hydrological regime influences the degree of interaction among soils, groundwater and surface water. At low flow, rivers are primarily influenced by aquifer discharge, and surface water penetration into the floodplain is limited (Baker and Vervier, 2004). During high flow, river water penetrates far into the floodplain, the aquifer table rises and areas hydrologically connected to the river channel may extend laterally for up to several kilometres in alluvial floodplains (Stanford and Ward, 1993; Wroblicky *et al.*, 1998). In the broad floodplains of large rivers, underflow is expected to be dominated by advected channel water (White, 1993) and the resulting hyporheic zone might be primarily lateral on a scale of hundreds of metres (Ward, 1989; Stanford and Ward, 1993).

As they allow complex geometry and boundary conditions to be reproduced, numerical models such as finite-element or finite-difference techniques are frequently used to study river–aquifer interactions. This is a complex, multi-scale and inter-disciplinary problem, and a

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number of modelling methods have been proposed (see reviews by Packman and Bencala, 2000; Runkel *et al.*, 2003; Cardenas *et al.*, 2004; Boano *et al.*, 2006). One of the principal current focuses of this field of research is the method of coupling between river and aquifer equations (Kollet and Maxwell, 2006). Indeed, despite the fact that aquifer and surface water are hydraulically interconnected, they are often modelled as two separate systems and are analysed independently (Liang *et al.*, 2007). It is now well recognized that fully coupled models for river and aquifer flows are necessary to obtain a better understanding of the hydrological pathways in hydrosystems (Panday and Huyakorn, 2004; Gunduz and Aral, 2005). Fairbanks *et al.* (2001) demonstrated that the ‘fully implicit’ approach (VanderKwaak and Loague, 2001; Gunduz and Aral, 2003), in which both systems of equations are solved in a single global matrix, is the most numerically stable method to couple surface and subsurface models. Gunduz and Aral (2003) tested this solution by coupling a one-dimensional channel flow model based on the dynamic wave form of the Saint Venant equations with a two-dimensional vertically averaged saturated groundwater flow model. They concluded that this approach provides an efficient solution for the coupled flow problem formulated for both systems. Liang *et al.* (2007) used a similar approach to build a two-dimensional numerical model for predicting flood flows. The solution proposed by VanderKwaak and Loague (2001) and Gunduz and Aral (2003) can be particularly efficient for studying two-way interactions between rivers and aquifers, but examples of such applications are still rare (Langevin *et al.*, 2005). Many coupled models of variable complexity have been developed to simulate the interactions between surface and subsurface systems (VanderKwaak and Loague, 2001; Morita and Yen, 2002; Yeh and Huang, 2003; Weng *et al.*, 2003; Panday and Huyakorn, 2004; Langevin *et al.*, 2005; Loague *et al.*, 2005). For most of the models mentioned above, the spatial extent of the interactions between groundwater and surface water is assumed to be constant, or spatially and temporally transient changes in processes are considered in a rudimentary manner (Krause and Bronstert, 2007). In addition, the full three-dimensional (3D) numerical solution encounters limitations in mesoscale applications because of the high computation time required (Krause and Bronstert, 2007). Gunduz and Aral (2003) also concluded that the coupled models tend to be increasingly complex and their solution can suffer from numerical complications. It appears that the complexity of the model should thus depend on the objectives of each case.

In this context, the main objective was to propose a coupled and simultaneous solved model to quantify fluxes of water and solutes between a large river and its shallow aquifer at mesoscale (Malard *et al.*, 2002; Krause and Bronstert, 2007). The aim was to propose a model at the interface between hydrology and ecology, with an interdisciplinary and integrated approach (Gurnell *et al.*, 2000; Tetzlaff *et al.*, 2007), which could be easier to apply than 3D solvers previously developed. The model

includes the hydraulic model presented by Liang *et al.* (2007) but has been extended to compute solute transport. It requires input data that are relatively easy to obtain (stream and floodplain geometry in two dimensions, averaged impermeable bedrock elevation, mean hydraulic conductivity and porosity). It can be relevant to consider such vertically integrated two-dimensional (2D) models when lateral exchange is a dominant process between river and aquifer, or when the geomorphology of the studied site allows integration over the z direction. This model might thus be useful when complex models are not fully necessary, for example when the objective is to have an averaged quantification of the hydrology of a site in order to integrate biogeochemistry.

The model, called 2SWEM (for Surface-Subsurface Water Exchange Model), consists of a hydraulic component that describes the main hydrological characteristics of surface water and groundwater, coupled with a solute-transport component to follow the dynamics of nutrients. The hydraulic model is based on the equations system of Liang *et al.* (2007): the flow equations in the river and in the aquifer are integrated over the z direction to produce a coupled two-dimensional model in both domains.

In this paper, 2SWEM was tested by applying it to a large sector (about 36 km²) of the Garonne River, the third largest river in France and the largest river in south-west France. In its middle course, it fully penetrates the alluvial aquifer, leading to significant lateral interactions (Weng *et al.*, 2003; Baker and Vervier, 2004) that increase the surface and length of contact between streamwater solutes and active biological zones in floodplain sediments. The 2SWEM simulation results were compared with hydraulic heads measured (i) on one sampling date (30 March 2000) in the floodplain using 27 piezometers and (ii) daily over 5 months (from January to June 2000) using 5 piezometers. After model verification, the solute-transport component was added in order to simulate the injection of a conservative tracer at some locations in the alluvial aquifer and to follow its theoretical transport.

STUDY SITE

Location and hydro-geological context

The Garonne River is the third longest river in France, with a stream length of 525 km from its headwaters within the Aran Valley (Spain) in the high Pyrenees, to the Gironde Estuary, which discharges into the Atlantic Ocean (Figure 1A). The Garonne is the largest river in southern France (Strahler stream order 7 at the study site) and drains ~57 000 km² of the Pyrenees, Massif Central and Aquitaine Basin. The flow regime of the Garonne is pluvio-nival; its tributaries come from the Central Massif (pluvial regime) and from the Pyrenean Massif (nival regime). This regime can present two low water flow periods depending on accumulation of a snow mantle in the mountains areas; one in late summer (August to

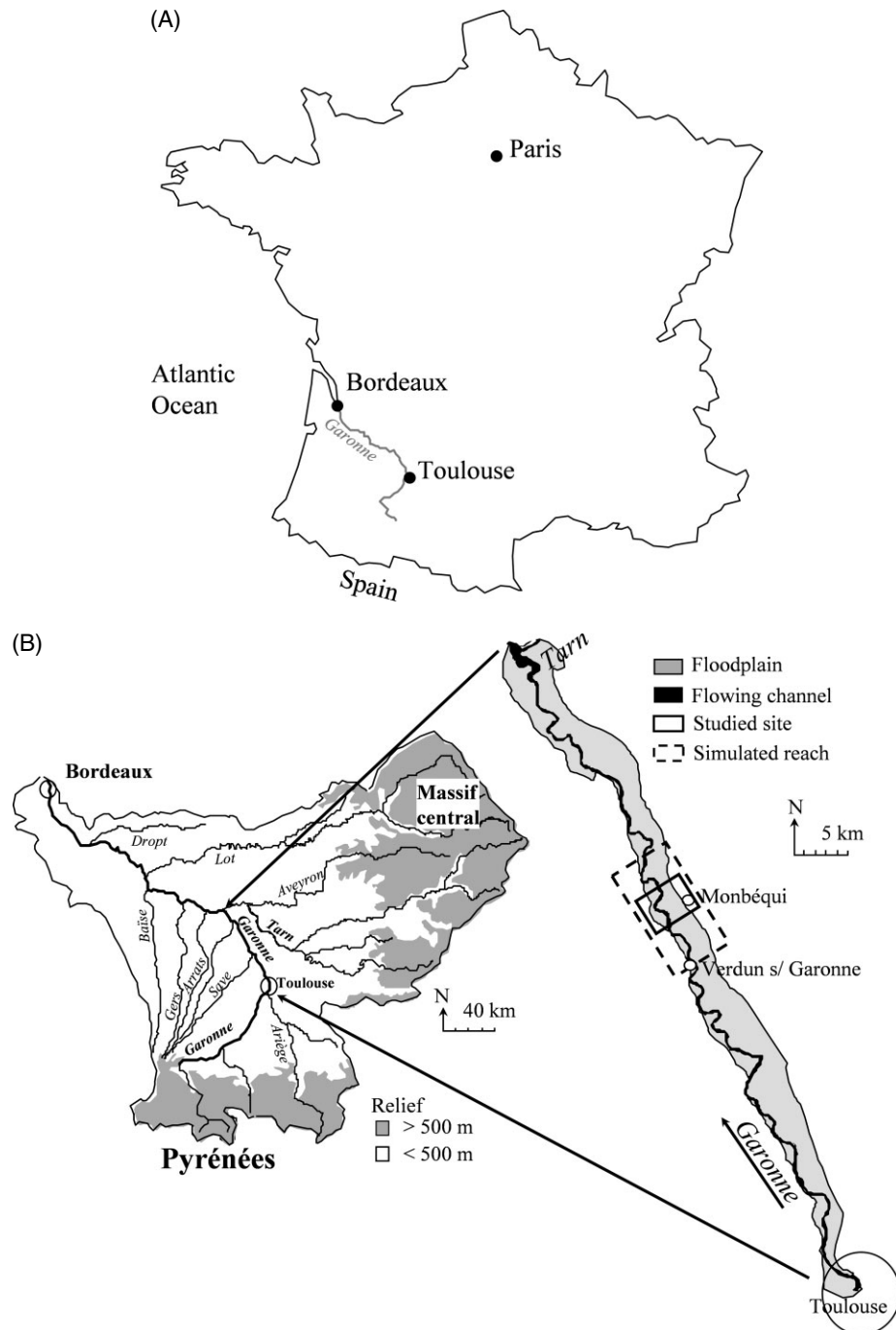


Figure 1. (A) The Garonne River, south-west France, (B) the middle Garonne (from Toulouse to the confluence with the Tarn River), the simulated reach and the studied site

October) and one less frequently in winter (January to March).

The Middle Garonne River, upstream of the confluence with the Tarn River (Figure 1B), covers a drainage basin of 32 350 km². In this sector, the valley contains a classic flight of terraces that represent episodic bedrock valley deepening, punctuated by lateral migration of deposition of sediments (Lancaster, 2005). The valley is profoundly bound by an impermeable layer of molassic substratum and the river flows and migrates in the floodplain, between more or less sloping hillsides (Figure 2). The slope of hillsides is marked by a series of breaks, which

correspond to a series of terraces, more or less developed according to the left or right bank. The floodplain is delimited on its sides by the high terrace, which is safe from flooding. The middle terrace, which is about 2 km wide, is used for agricultural purposes and is flooded only rarely (every 30–50 years). The lower terrace, which is a few hundred metres wide, is flooded about every 5 years and is devoted to poplar plantations. The river banks themselves, which are 10–100 m wide, are flooded at least once a year and are occupied by riparian vegetation, mainly black poplar and white willow (Lambs, 2004). Field expertise shows that the impermeable substratum

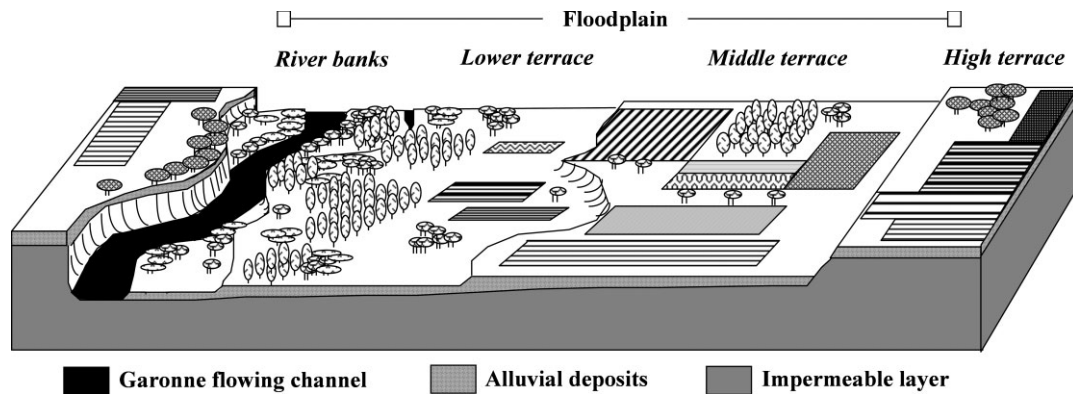


Figure 2. Cross-sectional schematic diagram of the Garonne River floodplain in the simulated reach

of the high terraces is placed above the topographical surface of the floodplain, so there is no groundwater flow coming from the edges and the aquifer in the floodplain is disconnected from the larger-scale aquifer (Lancaster, 2005).

The Garonne channel can be classified as a mixed bedrock–alluvial stream (Howard *et al.*, 1994; Howard, 1995). The floodplain contains between 4 and 7 m of quaternary sand and gravel deposits overlying an impermeable layer (mix of molasse, clay and marl). The aquifer in the alluvium is located at a depth that varies from 2 to 5 m in low water periods, but can rise rapidly up to the soil surface during floods (Weng *et al.*, 2003). As shown in Figure 2, the Garonne River fully penetrates the unconfined aquifer in the floodplain. Its bed is at, and sometimes below, the lower boundary of the aquifer, leading to dominant lateral surface/subsurface interactions compared with vertical interactions (Steiger *et al.*, 2000).

The river reach studied here (Figure 1B), between Toulouse city (860 000 inhabitants) and the Tarn River confluence, has a meandering, single-thread channel pattern. It is 12 km long and there are no major tributary inputs to the reach, which is characterized by a mean annual discharge of $200 \text{ m}^3 \text{ s}^{-1}$, an average channel width of 150 m and a mean channel slope of 0.0010 m m^{-1} . The discharge is highly variable, as it ranges from $50 \text{ m}^3 \text{ s}^{-1}$ in late summer to $4000 \text{ m}^3 \text{ s}^{-1}$ during exceptional floods. The total floodplain is about 3 km wide in this area. It is heavily cultivated (maize, sunflowers, sorghum, etc.), leading to major nitrate influx into the groundwater, where concentrations of $50\text{--}100 \text{ mg L}^{-1}$ are common (Pinay *et al.*, 1998; Sanchez-Perez *et al.*, 2003). Banks can be zones of self-purification for rivers, by creating a temporary or permanent storage zone for particular, dissolved organic matter (Vervier *et al.*, 1993), nitrates (Sanchez-Perez and Trémolières, 2003; Sanchez-Perez *et al.*, 2003), and phosphorus (Bonvallet-Garay *et al.*, 2001; Bonvallet-Garay, 2003).

Equipment and data used

In 1999, a site of 13 km^2 near the village of Monbequi (40 km north of Toulouse) was instrumented with an observation network of 27 piezometers (Figure 3). These

are distributed along a 2 km meander of the Garonne River to provide a good spatial distribution of hydrological data throughout the floodplain. The piezometers, which are made from polyvinyl chloride and are fully screened, penetrate through the sediments and extend to the underlying impermeable layer. Five piezometers (P9, P15, P19, P22 and P23, in black circles in Figure 3) are equipped with continuous water level recorders (Orphimèdes Ott, France) that provide water level measurements in the alluvial aquifer every 10 min. There is also a gauging station 3 km upstream of the study site at Verdun-sur-Garonne (Figure 1B), where water level is recorded by the regional environmental agency (DIREN).

Hydraulic heads data were used for verification of the hydraulic model. Two data sets were used in this paper: the first of these consisted of a single-occasion measurement of hydraulic heads in all piezometers on 30 March (discharge in the Garonne River = $198 \text{ m}^3 \text{ s}^{-1}$). For the second set, mean hydraulic heads were daily calculated in the five piezometers equipped with Orphimèdes from 1 January to 1 June 2000 (Figure 4). These two data sets were compared with the hydraulic heads simulated with 2SWEM under the same hydraulic conditions.

NUMERICAL MODEL

Previous modelling work on the Garonne floodplain and 2SWEM outline

In the case of the Garonne River, two models have recently been applied to investigate the impact of river–aquifer interactions on the biogeochemical operation of the system. Sauvage *et al.* (2003) developed a 1D hydro-biogeochemical model to investigate the evolution of nitrogen dynamics in the Garonne channel during low water periods. This model has a proven ability to give good simulations for low water periods and allows estimation of fluxes between free-flowing water and epilithic biofilm. However, this model cannot be used to identify the areas of interaction between river and aquifer and to integrate the complex two-way coupling between aquifer and river during higher water periods. Weng *et al.* (2003) used the MARTHE model (MARTHE Hydrodynamic Software developed by Bureau de Recherches

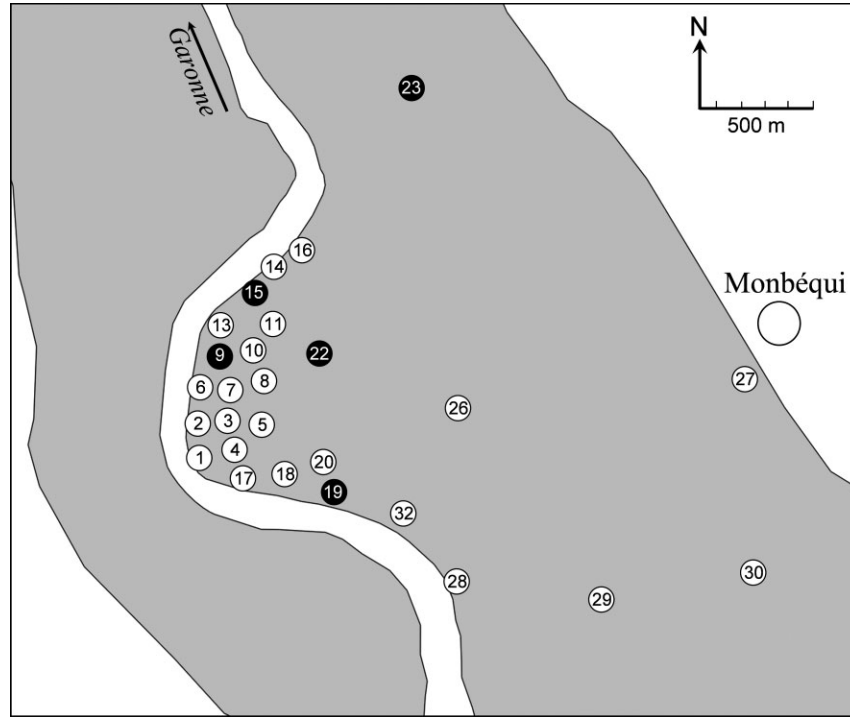


Figure 3. Location of the 27 piezometers in the studied site. The black circles designate the five piezometers equipped with continuous water level recorders

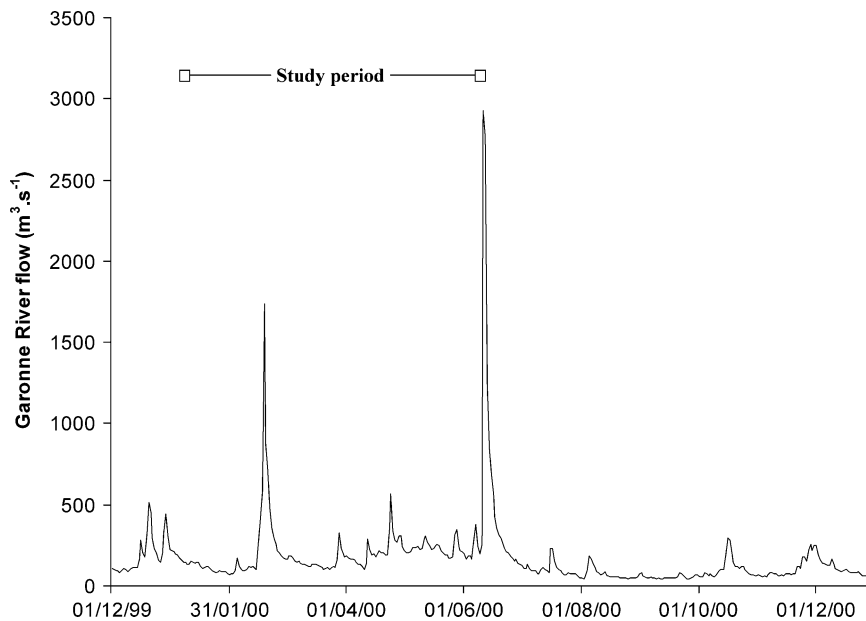


Figure 4. Garonne River discharge measured at Verdun-sur-Garonne for the year 2000. The study period from 1 January to 1 June 2000 corresponds to the temporal sampling period for water levels in piezometers at the studied site

Géologiques et Minières, BRGM, France) to quantify the buffer function of the alluvial floodplain in a meander of the Garonne River. MARTHE is a typical groundwater model, based on Darcy's equations, which takes into account the surface water as an external boundary condition. The model showed good accuracy for reproducing water levels in groundwater, both for high and low water flow in the river, and was able to simulate transport of conservative elements (Weng *et al.*, 2003). However, this model needs complex topographical and hydraulic field

measures to be used and calibrated, and it did not simulate the direct two-way coupling between surface and subsurface. These two models have brought some understanding about the lateral transfer of nutrients from the Garonne River channel to the floodplain, but relatively little work has been published on the transfer from floodplain to river channel. However, Lambs (2004) observed significant outputs of groundwater into the Garonne River. It is thus essential to be able to quantify water and solute exchanges from stream to aquifer and vice versa.

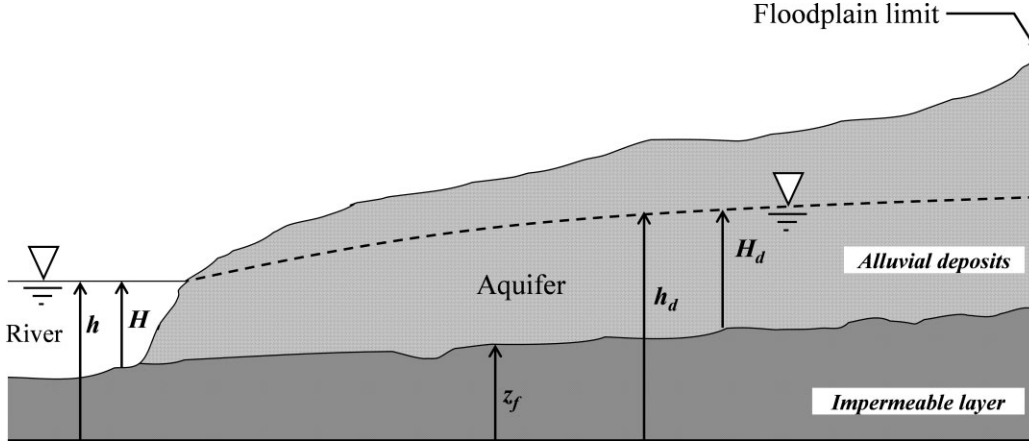


Figure 5. Cross-sectional schematic diagram of a fully penetrating river and adjacent sloping aquifer. H (in m) is the water depth in the channel, h is the level of the free surface water (in m) with respect to sea level (i.e. $h = H + z_f$, z_f is the impermeable layer elevation), and H_d (m) is the water surface elevation of the aquifer ($h_d = H_d + z_f$)

The model presented in this paper is illustrated in Figure 5. Two domains are distinguished: the river and the floodplain. It is well recognized that in fluvial environments, interactions between river and aquifer occur on different spatial scales and in the neighbourhood of the river the flow is generally three-dimensional (Findlay, 1995; Jones and Holmes, 1996; Dole-Olivier, 1998). However, in the case of rivers fully penetrating an alluvial aquifer, or when the horizontal scale is larger than the water depth, exchanges are primarily lateral and it can be relevant to consider a vertically integrated two-dimensional model. In the proposed model, the flow equations are thus integrated over the z direction to produce a coupled two-dimensional model in both domains. Channel geometry is assumed to be rectangular and the alluvial aquifer is shallow. The channel and the aquifer are bordered at their base by an impermeable layer (z_f). The hydraulic model calculates the surface water elevations (H and H_d) and the specific discharges (q and q_d) in the river and floodplain respectively. It requires input data that are relatively easy to obtain, namely stream and floodplain geometry in 2D, averaged impermeable layer elevation (z_f), mean hydraulic conductivity and porosity. For both river and floodplain equations, the vertical movements of flow are neglected. The equations are directly coupled at the common interface, where water exchange takes place.

The coupling with a transport model also allows solute exchanges between floodplain and river to be studied. It is possible to simulate the dynamics of a tracer injected into the river, and to follow its infiltration through the riverbed sediments in order to delineate the extent of the hyporheic zone. The model can also be used to follow the dynamics of a solute dispersed in the floodplain (for example pesticides in agricultural fields) and to predict its concentration in the river and in sediments downstream from the contamination point.

Hydraulic model governing equations

When the horizontal scale is larger than the water depth, the well known two-dimensional shallow water

model, which represents mass and momentum conservation, can be obtained by depth-averaging the Navier–Stokes equations. These equations are also called the two-dimensional Saint Venant equations. Neglecting the Coriolis, wind and viscous forces, the 2D Saint Venant equations are given by (Chow *et al.*, 1988):

$$\frac{\partial h}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} = 0 \quad (1.1)$$

$$\frac{\partial q_x}{\partial t} + \frac{\partial(q_x \cdot q_x/H)}{\partial x} + \frac{\partial(q_x \cdot q_y/H)}{\partial y} = \sum F_x \quad (1.2)$$

$$\frac{\partial q_y}{\partial t} + \frac{\partial(q_y \cdot q_x/H)}{\partial x} + \frac{\partial(q_y \cdot q_y/H)}{\partial y} = \sum F_y \quad (1.3)$$

where t is time, H (in m) is the water depth in the channel, h is the level of the free surface water (in m) with respect to sea level (i.e. $h = H + z_f$, z_f is the impermeable layer elevation, see Figure 5), q_x and q_y are the river specific discharges along the x - and y -axis (as the model is 2D vertically integrated, these discharges are in $\text{m}^2 \text{s}^{-1}$).

The force components, F_x and F_y , which account for gravity and roughness effects, can be written as:

$$\sum F_x = -gH \cdot \frac{\partial h}{\partial x} - \frac{n^2 g |q| q_x}{H^{1/3}} \quad (1.4)$$

$$\sum F_y = -gH \cdot \frac{\partial h}{\partial y} - \frac{n^2 g |q| q_y}{H^{1/3}} \quad (1.5)$$

where g is the gravitational acceleration (m s^{-2}), q is the discharge (in $\text{m}^2 \text{s}^{-1}$) and n is the Manning roughness coefficient (as the model is 2D vertically integrated, this parameter is in $\text{s m}^{-4/3}$). It is well known that the solutions of these equation systems may present steep fronts and shock discontinuities, which can cause numerical difficulties (Delis and Katsaounis, 2005) and model divergence. To limit such problems, an artificial viscosity term has been added to the Saint Venant equations that stabilizes the solution by making the cell Reynolds number of order unity (COMSOL User's Guide).

For an isotropic, saturated, homogeneous and unconfined aquifer, the Dupuit–Forchheimer assumption allows the three-dimensional equations to be reduced to Darcy–Dupuit 2D equations given by (Bear, 1972):

$$q_{xd} = -K_H \cdot H_d \cdot \frac{\partial h_d}{\partial x} \quad (2.1)$$

$$q_{yd} = -K_H \cdot H_d \cdot \frac{\partial h_d}{\partial y} \quad (2.2)$$

$$\begin{aligned} \phi \frac{\partial h_d}{\partial t} = & \frac{\partial}{\partial x} \left(K_H \cdot H_d \cdot \frac{\partial h_d}{\partial x} \right) \\ & + \frac{\partial}{\partial y} \left(K_H \cdot H_d \cdot \frac{\partial h_d}{\partial y} \right) + q_r \end{aligned} \quad (2.3)$$

In particular, complex 3D flows near the river/bank interface may not necessarily require the two levels to be exactly equal. Another limitation of the model, as it is currently implemented, is the fact that it does not take into account variations in river bathymetry, as the channel is rectangular and the river, which forms a boundary with the aquifer, fully penetrates the aquifer and is in direct hydraulic connection with it.

Transport model

In this paper, the transport of a conservative tracer in the river and the aquifer is described using the classical advection–dispersion equation (Bear, 1972). This equation is also vertically integrated assuming that vertical concentration gradients are negligible in the

$$\mathbf{D} = \mathbf{D}_0 + \alpha_T \left\| \frac{\mathbf{q}_d}{\phi H_d} \right\| \mathbf{I} + (\alpha_L - \alpha_T) \frac{\mathbf{q}_d \mathbf{q}_d}{\phi H_d \|\mathbf{q}_d\|} \quad (4.2)$$

with $\mathbf{D}_0 = \frac{D}{\tau} \mathbf{I}$

$$\mathbf{D} \rightarrow \left[\begin{array}{cc} \frac{D}{\tau} + \alpha_T \left\| \frac{\mathbf{q}_d}{\phi H_d} \right\| + (\alpha_L - \alpha_T) \frac{q_{dx} q_{dx}}{\phi H_d \|\mathbf{q}_d\|} & (\alpha_L - \alpha_T) \frac{q_{dx} q_{dy}}{\phi H_d \|\mathbf{q}_d\|} \\ (\alpha_L - \alpha_T) \frac{q_{dx} q_{dy}}{\phi H_d \|\mathbf{q}_d\|} & \frac{D}{\tau} + \alpha_T \left\| \frac{\mathbf{q}_d}{\phi H_d} \right\| + (\alpha_L - \alpha_T) \frac{q_{dy} q_{dy}}{\phi H_d \|\mathbf{q}_d\|} \end{array} \right] \quad (4.3)$$

where K_H is the hydraulic conductivity (m s^{-1}), H_d (m) is the water surface elevation, h_d is the level of the free water surface with respect to sea level ($h_d = H_d + z_f$), q_{dx} and q_{dy} (in $\text{m}^2 \text{s}^{-1}$ per unit width) are discharges along the x - and y -axis in the porous medium, q_r is the aquifer recharge (m s^{-1}) and ϕ is the porosity (dimensionless).

Coupling procedure

Several options are available to couple surface and subsurface hydraulic models (Panday and Huyakorn, 2004). Fairbanks *et al.* (2001) demonstrated that the ‘fully implicit approach’, also called ‘simultaneous solution’ (VanderKwaak and Loague, 2001; Gunduz and Aral, 2003), in which both systems of equations are solved in a single global matrix, is the most numerically stable method. The fully coupled solution also outperforms linked/iteratively coupled methods in terms of computational efficiency for highly interactive systems (Panday and Huyakorn, 2004). In our case, the equations in both domains are solved together: the direct coupling between surface and subsurface equations is provided by imposing continuity of the water levels and of the fluxes along the river-aquifer interface:

$$H = H_d \quad (3.1)$$

$$q_x n_x + q_y n_y = q_{xd} n_x + q_{yd} n_y \quad (3.2)$$

The resulting implemented model, 2SWEM, calculates three unknown variables for both domains: the water elevation and the two specific discharges (along the x - and y -axis). At this point, it is important to recognize that these boundary conditions impose some limitations.

In porous media, the resulting transport equation can be written as:

$$\frac{\partial \phi C}{\partial t} + \nabla \cdot \left(C \frac{\mathbf{q}_d}{H_d} \right) = \nabla \cdot (\phi \mathbf{D} \cdot \nabla C) + R + \frac{q_r c_r}{H_d} \quad (4.1)$$

where C (mg L^{-1}) is the tracer component concentration, R is the bulk reactive source/sink term for non-conservative components, c_r is the recharge concentration, q_d is the previously defined discharge in the floodplain and \mathbf{D} is the dispersion tensor ($\text{m}^2 \text{s}^{-1}$), which can be written as:

where τ is the tortuosity of the medium (dimensionless) and α_L and α_T are the longitudinal and transversal dispersivity respectively (m). It must be emphasized that it is very easy to add as many transport equations as required by the geochemical model describing the particular situation under consideration. Since the application described later in this paper involves only one chemical species, the presentation of the model is limited to this case. This transport model does not take into account the period leading to the establishment of a fully dimensional concentration field, i.e. transport through the unsaturated zone and the 3D mechanisms in the neighbourhood of the contamination zone.

Numerical implementation

The equations are solved using a finite element numerical model, COMSOL™ Multiphysics software (www.comsol.com). The linear solver used is the direct solver UMFPAK (Davis, 2004), while other solvers, including iterative solvers, are available if necessary.

UMFPACK is a set of routines for solving non-symmetrical sparse linear systems, using the Unsymmetric MultiFrontal method. Equations are solved in a fully coupled manner using the implemented Newton–Raphson scheme. All COMSOL Multiphysics solvers for time-dependent problems are iterative, the time step is adaptive and determined by local error estimates. If the error tolerances, relative or absolute, are not met, the local time step is disqualified and a new smaller time step is chosen. The relative and absolute error parameters determine the limit for the estimated error in each integration step. This means that the time steps vary automatically according to the convergence of the solution. The time step is thus chosen to be short enough to resolve discharge variations in the river and in the floodplain, and at the same time long enough to avoid unnecessary long simulation times. Limits (max/min) for this time-step control and the steps of storage of the solution can also be specified. For this paper, the absolute and relative tolerance parameters for the time-dependent solver were 0.001 and 0.01 respectively, and the solution was stored daily.

The space discretization was based on triangular finite elements. For this application, the mesh consists of 29 688 triangular finite elements that cover a total surface of about 36 km². Typical triangular element sizes discussed below are about 40 m² in the river and up to 100 m² in the floodplain.

FIELD APPLICATION

Data for the model

The 2D geometry of the Garonne River and the floodplain was created with MapInfo Professional® version 6.5 software. The limits were digitized on IGN 1/25000 maps and the resulting vector data were saved as a DXF file, which can be imported to COMSOL Multiphysics software. In the case of the Garonne River, which fully penetrates the alluvial aquifer, the bottom elevation of the impermeable layer (z_f) is continuous for both domains. The z_f value was fixed to the mean measured slopes 0.0010 m m⁻¹ and 0.0008 m m⁻¹ along directions parallel and perpendicular to the river respectively (Steiger *et al.*, 2000).

In the river, the Manning roughness coefficient was calculated by calibration of the Strickler coefficient K_s (m^{1/3} s⁻¹, with $n = 1/K_s$) with a 1D Saint Venant model in the same sector of the Garonne River (Sauvage *et al.*, 2003). Its value was fixed at 0.04 s m^{-1/3}. At the study site near Monbéqui village, hydraulic conductivity values have been estimated during pumping tests and slug tests and the values obtained vary from 10⁻² to 10⁻⁵ m s⁻¹. Weng *et al.* (2003) used these values and a modelling inverse method to obtain a map of the hydraulic conductivity for this site. For the simulations, a mean hydraulic conductivity of 10⁻³ m s⁻¹ was imposed throughout the simulated reach, while at the Monbéqui site, hydraulic conductivities corresponded to values calibrated by Weng *et al.* (2003). During the study period,

the recharge was set to 0 m s⁻¹ since the fluctuating river stage had a dominant impact on groundwater heads. The porosity of alluvial deposits in the floodplain was fixed at 0.30 based on interpretation of pumping tests throughout the simulated reach (Seltz, 2001).

Field expertise on the Monbéqui site and recent simulations with a soil crop model (STICS, Brisson *et al.*, 1998), allowed several agricultural parcels with great nitrate leaching that can affect the quality of aquifer and Garonne River to be identified. For the transport simulations in the present study, it was decided to simulate conservative tracer dynamics injected in two of these parcels with high nitrate concentrations. These two plots represent the theoretical input source of ‘contamination’ to be followed (Figure 11A). The velocity field for the transport model is taken from steady-state flow simulation with the mean annual discharge of the Garonne River (200 m³ s⁻¹). The mean dispersion coefficient in the Garonne River was calculated from values given by Sauvage *et al.* (2003), and fixed at 10 m² s⁻¹. Longitudinal dispersivity in the aquifer was taken to be equal to 15 m, and lateral dispersivity was 10 times smaller, 1.5 m.

Initial and boundary conditions

For the upstream boundary conditions in the Garonne River, the water elevation measured at the gauging station at Verdun-sur-Garonne during the year 2000 (Figure 4) was used. In the floodplain (Figure 6), the upstream and downstream boundary conditions were prescribed heads, which can vary with time and distance from the river, obtained from a larger-scale simulation. As explained in the ‘Location and hydro-geological context’ section (see also Figure 2), there is no groundwater flow coming from the edges and one can impose ‘no flow’ conditions on both sides of the floodplain. As boundary conditions for the transport component, continuity conditions between floodplain and river, and convective flux on the limits of the floodplain were used.

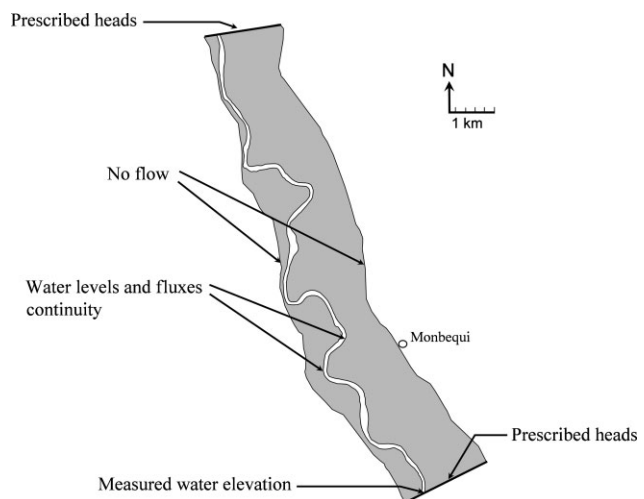


Figure 6. Simulated reach and boundary conditions for the hydraulic model

For initial conditions of the hydraulic component, $H(x, y, t = 0)$ and $H_d(x, y, t = 0)$ were taken from previous calculations in stationary mode for a constant discharge in the Garonne River that corresponds to the mean annual discharge ($200 \text{ m}^3 \text{ s}^{-1}$). For transport, the conservative tracer concentration was taken to be equal to 0 mg L^{-1} throughout the studied reach, except in the two 'polluted' plots with an initial concentration of 10 mg L^{-1} .

Parameters for model accuracy evaluation

The evaluation of the model included calculation of the absolute residual mean between measured and simulated hydraulic heads, the correlation coefficient, the normalised root mean square error (NRMSE) and the Nash–Sutcliffe coefficient (Nash and Sutcliffe, 1970). The NRMSE is a frequently used measure of the difference between values predicted by a model and the values actually observed. It is calculated by dividing the

root mean squared error by the range of observed values. The NRMSE is generally expressed as a percentage, where lower values indicate less residual variance. The Nash–Sutcliffe coefficient is also widely used to assess the predictive power of hydrological models. Generally, the Nash–Sutcliffe coefficient is used to quantitatively describe the accuracy of model predictions for discharge, but this method can also be used to describe the predictive accuracy of other outputs such as hydraulic heads, as long as there are observed data with which to compare the model results. Nash–Sutcliffe coefficient values range from $-\infty$ to 1. The closer the coefficient to 1, the more accurate the model. A value of 1 indicates a perfect match of modelled output to observed data. An efficiency of 0 indicates that the model predictions are as accurate as the mean of the observed data. A value less than zero occurs when the observed mean is a better predictor than the model.

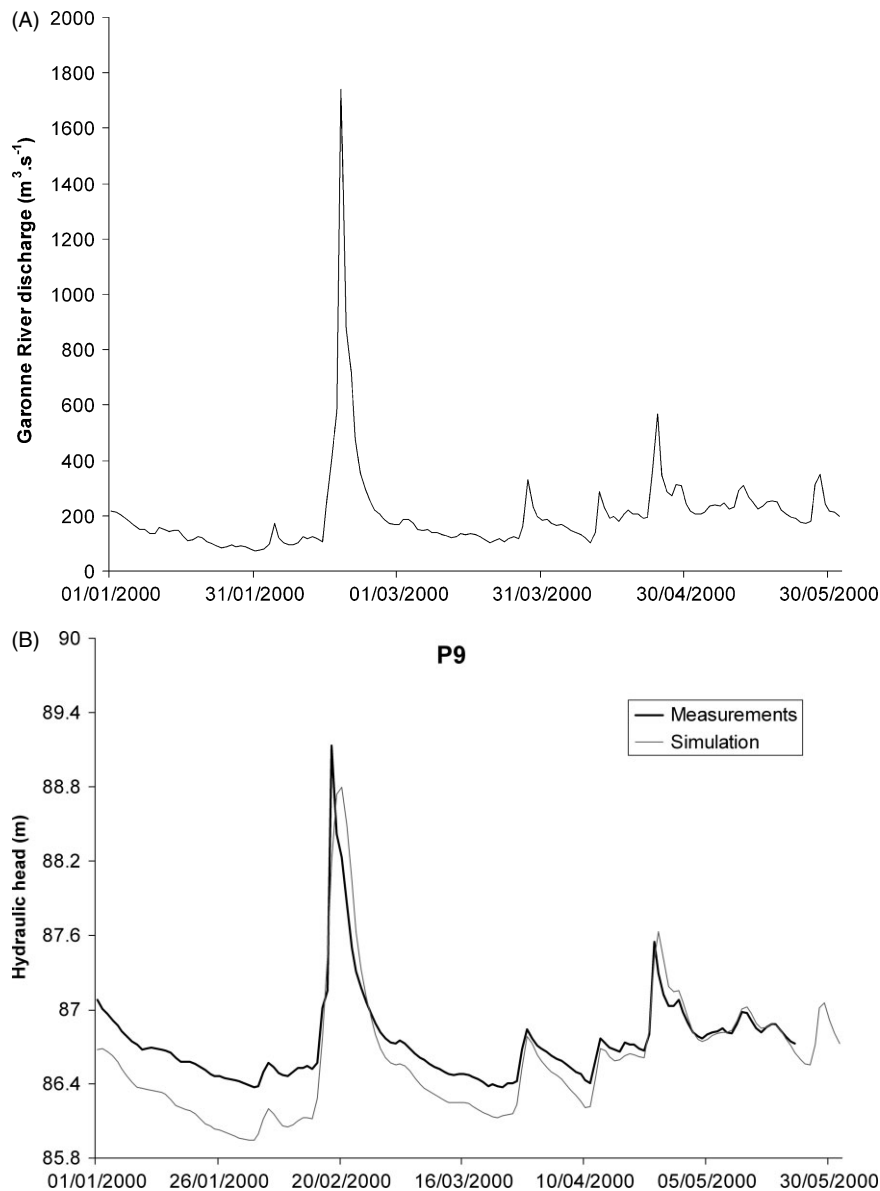


Figure 7. (A) Garonne River discharge at Verdun-sur-Garonne during the studied period from 1 January to 1 June 2000 and (B, C, D, E, F) simulated compared with measured hydraulic head in P9, P15, P19, P22 and P23, respectively

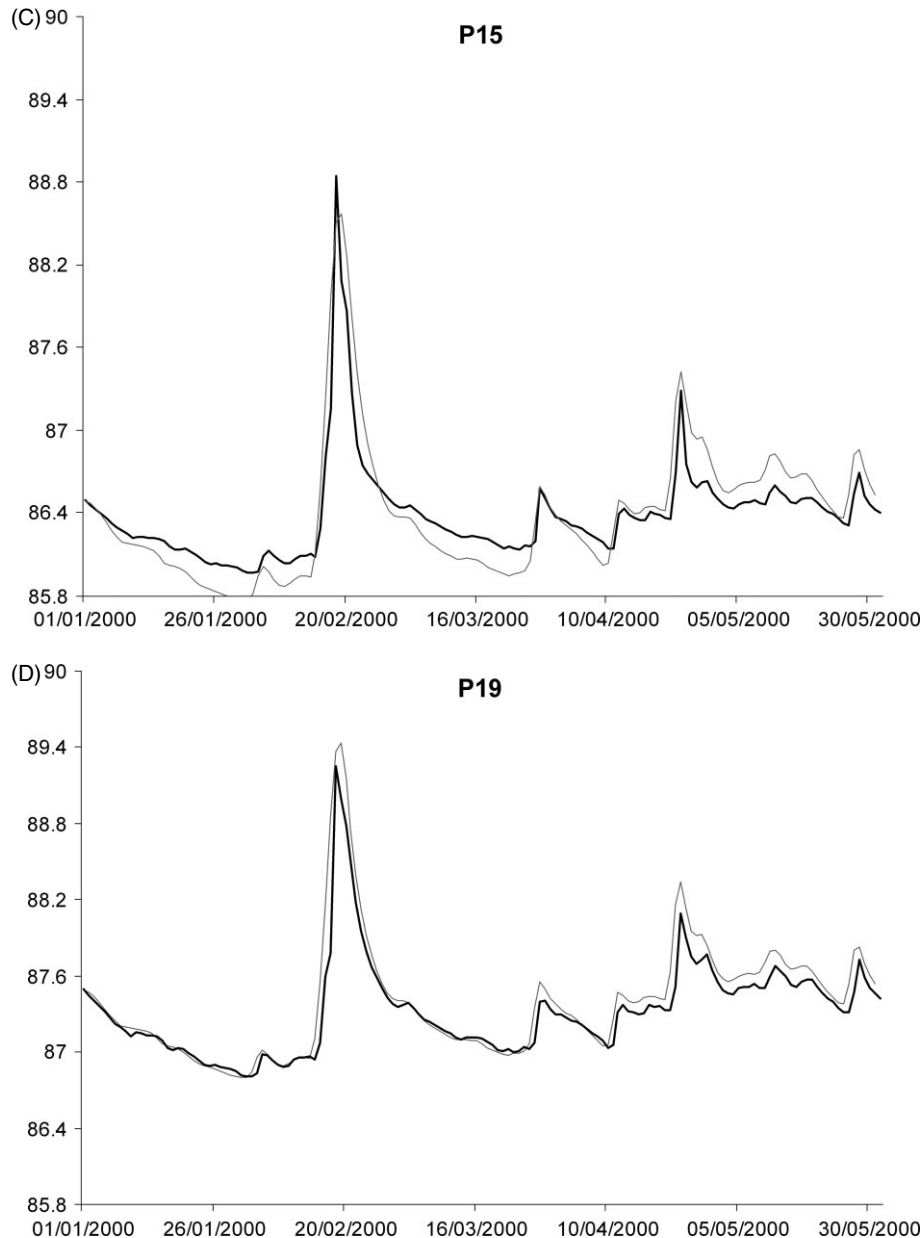


Figure 7. (Continued)

RESULTS AND DISCUSSION

Comparison between 2SWEM simulations and field measurements

Hydraulic heads measured in the Garonne floodplain were compared with those simulated with 2SWEM from 1 January to 1 June. The results are summarized in Table I. During the study period, 2SWEM accurately reproduced the strong variations in hydraulic heads, in terms of space as well as time, for the piezometers P9, P15 and P19, near the right bank of the river (Figure 7B, C and D). Far away from the river, in P22 and P23 (Figure 7E and F), the variations in hydraulic heads were slighter and tendencies were correctly simulated by 2SWEM in terms of space but not time. For P9 and P23, the hydraulic head variations were well reproduced by the model but it slightly underestimated the elevation in P9 and overestimated it in P23. This is probably

due to the fact that initial conditions of the hydraulic model for transient flow were taken from steady-state calculations for a constant discharge in the Garonne River ($200 \text{ m}^3 \text{ s}^{-1}$). As these initial conditions do not well integrate the actual variations in Garonne flow during the months preceding the period of study in 2000, this problem was expected. The results also show that the predicted increase in hydraulic heads after the passing of the flood in February occurred later than the observed increase in P22 (3 days later, Figure 7E) and P23 (5 days later, Figure 7F). The evaluation parameters for these two piezometers (Table I) confirmed that the model predictions were less accurate for them. These deviations could probably be explained by the small-scale heterogeneity of alluvial sediments, actual hydraulic conductivity and slope of the impermeable layer, which can modify local flow paths and the time transfer of

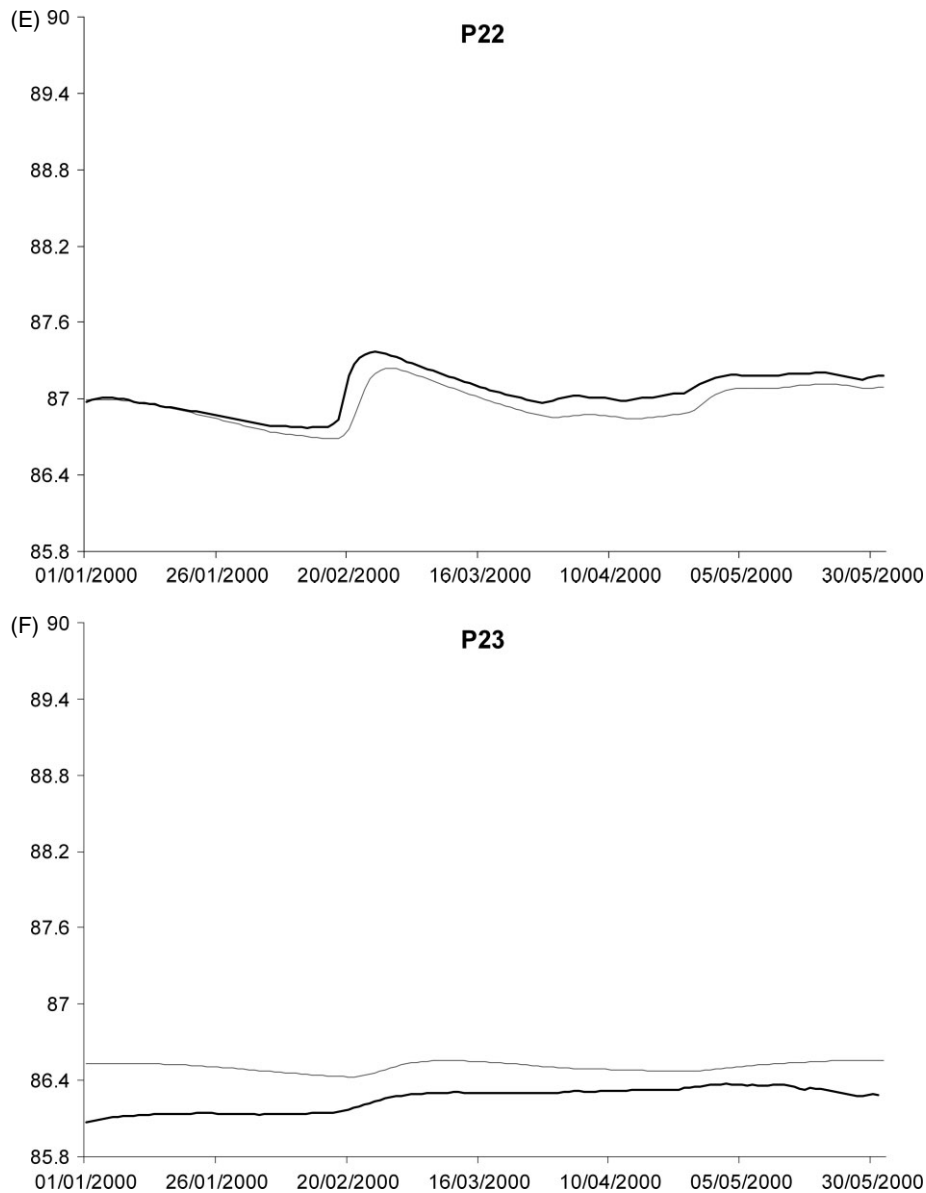


Figure 7. (Continued)

water from the river. However, it is now well established that in fluvial hydrosystems, the most important part of hydrological–biogeochemical activity occurs near the river banks (Lautz and Siegel, 2006; Lautz *et al.*, 2006). It can thus be predicted that this slight delay in time response for the piezometers far away from the river will not have significant consequences for future work on the biogeochemical operation of the system.

To verify whether 2SWEM is also able to reproduce the spatial distribution of hydraulic heads throughout the meander, the simulated heads were then compared with those measured in the 27 piezometers on 30 March (Figure 8). A clear upstream–downstream hydraulic head gradient appeared in the meander, with higher water levels upstream (P17–P20), probably due to surface-water infiltration, caused by local changes in bed morphology and consequently changes in water pressure at the bed (Vaux, 1968). This gradient indicates that on this date,

Table I. Parameters for evaluating the accuracy of 2SWEM predictions from 1 January to 1 June. The absolute residual mean (ARM, in m), the correlation coefficient (*r*), the normalised root mean square error (NRMSE) and the Nash–Sutcliffe (N-S) coefficient are calculated for each piezometer

Piezometer number	ARM (in m)	<i>r</i> (%)	NRMSE (%)	N-S coefficient
P9	0.22	90.26	13.17	0.26
P15	0.16	88.7	7.87	0.59
P19	0.09	96.3	6.02	0.85
P22	0.09	87.6	19.7	0.42
P23	0.25	0	89.72	−2.91

aquifer flows occurred from south-east to north-west, parallel to the general direction of the floodplain. The model reproduced this gradient well, the maximal absolute residual between observations and simulations being 0.36 m (for P32) and the mean absolute error only 0.15 m. The

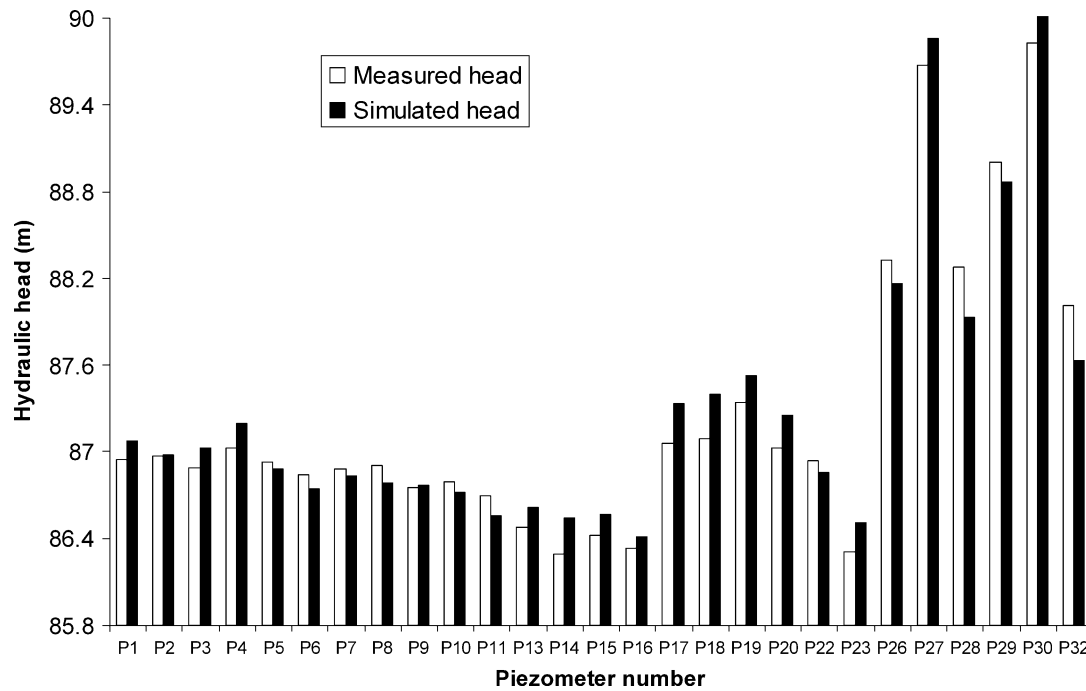


Figure 8. Histogram of simulated and measured hydraulic heads on 30 March 2000

normalized residual mean square error was 4.9% and the correlation coefficient between observed and simulated heads 98.3%. The residuals between observations and simulations were greater for P14, P15, P17 to P20 and for the inland piezometers far away from the meander (P26, P27, P29 and P30). These deviations could be explained by the fact that palaeochannels crossed the meander bend (Weng *et al.*, 2003) and may again induce local hydraulic conductivity variations and preferential flow paths that drain water downstream of the meander where the wetlands supply the river.

During the study period, aquifer flow directions were frequently modified, in particular during the passage of the flood in February. The simulated map of hydraulic heads and velocity fields (Figure 9A) shows that the model accurately reproduced the modification of flow-directions described in previous work on this site (Weng *et al.*, 2003; Lambs, 2004). When the discharge is low, aquifer flow occurs from south-east to north-west, parallel to the general direction of the floodplain, and the river drains the floodplain over a large part of the studied reach. During the falling stage, there is a reversal of flow direction and drainage of part of the alluvial aquifer into the river (Figure 9B). During the rise of the flood, Figure 10 confirms that river water infiltrates quickly into the first metres of the alluvial aquifer and the hydraulic heads near the river increase strongly. The impact of the flood decreases with the distance from the river but hydraulic heads are modified beyond a distance of at least 500 m from the river banks. From the results it was seen that temporal variations in flow direction, amount and intensity in Monbéqui meander should depend on flow variability. The flow regimes should then have implications for the biogeochemical activity in the hyporheic zone and origin and time of

residence of fluxes passing through the meander would influence its bio-reactor activity (Iribar *et al.*, 2008).

Conservative tracer dynamics simulation

Temporary storage of solutes in streams mainly depends on uptake and storage in riverbed sediments. These processes often lead to the formation of 'tails' following the passage of a solute pulse, and can account for important transformations of reactive solutes by active microbial communities fixed in riverbed sediments.

2SWEM was used to simulate the theoretical dynamics of a conservative tracer injected in the floodplain in two agricultural parcels (Figure 11A). This step is a necessary preliminary step in investigating the exchange of reactive solutes between the aquifer and the river using the future complete hydro-biogeochemical model. The tracer dynamics were followed for a 5 year period and for mean annual discharge of the Garonne River ($200 \text{ m}^3 \text{ s}^{-1}$). Two points in the river (R) and three points in the aquifer (Aq) downstream of the injection sites were investigated. Simulations showed that during the first months of simulation, the tracer progressively flowed through the sediments to the river. Downstream of the meander near Monbéqui, the concentration in the river increased significantly after 9 months (Figure 11B) and the maximum was obtained after 20 months of simulation. Owing to dilution in the channel, the maximum concentration was only about $6.13 \times 10^{-5} \text{ mg L}^{-1}$. Because of their positions relative to the two contaminated parcels, River1 and Aq1 plots received only a small quantity of tracer from the more upstream parcel. In contrast, profiles for Aq2 and Aq3 indicated that these two plots, located in large meanders, received tracer from the two parcels infiltrated on the right side of the floodplain. In Aq4, the weaker increase of tracer concentration indicated that

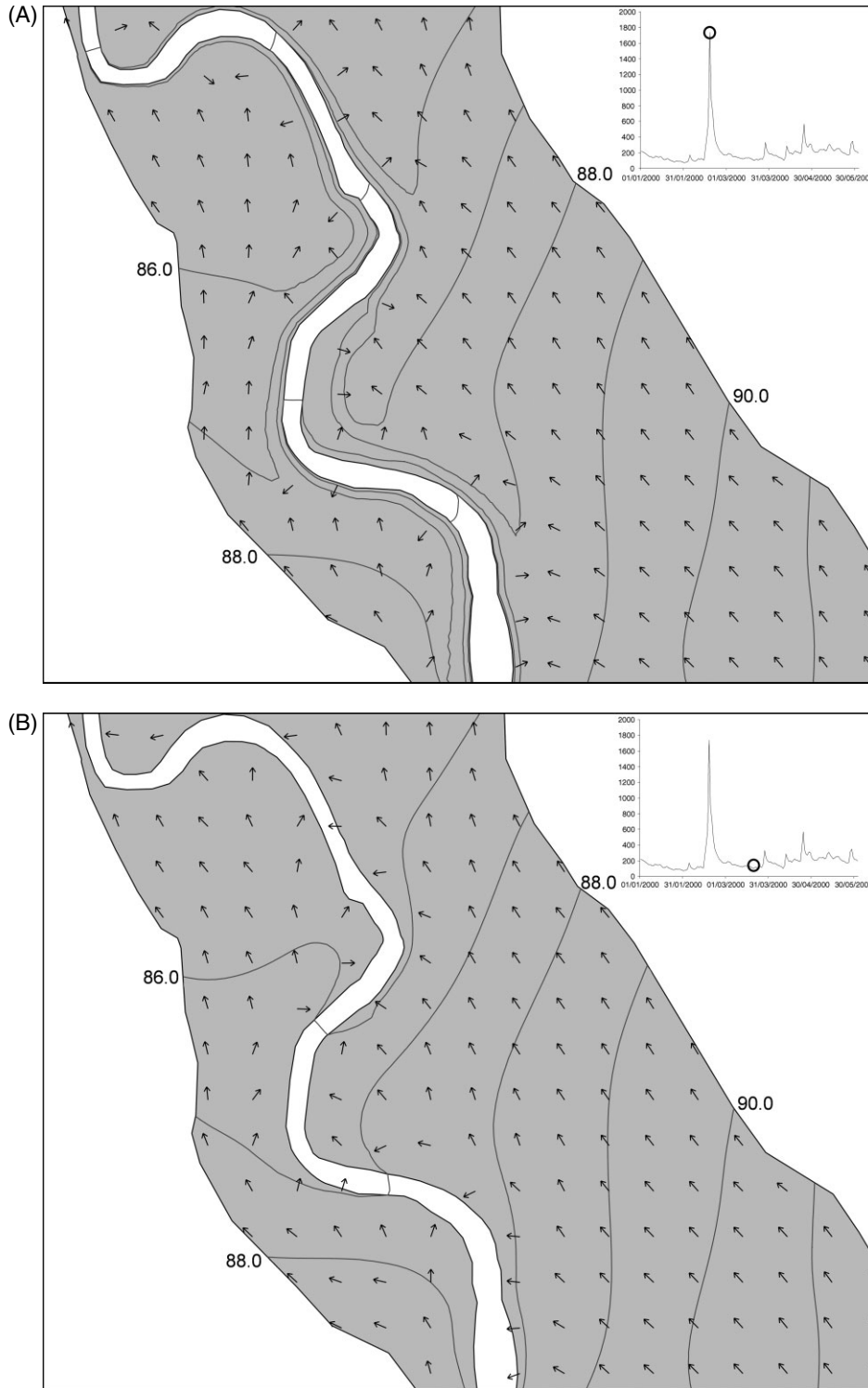


Figure 9. Groundwater flow patterns for (A) a daily mean river flow of 1740 m³ s⁻¹ at Verdun s/ Garonne (18 February 2000), and for (B) 124 m³ s⁻¹ (18 March 2000). Arrows represent velocity field, numbers are hydraulic heads elevation (in m, with respect to sea level)

exchanges were not constant throughout the reach but that preferential exchange could occur (Packman and Ben-cala, 2000; Lautz and Siegel, 2006). The main exchanges occurred in the more sinuous part of the reach, particularly in the largest meanders downstream from the studied reach. This confirms that in the Garonne River, the exchanges between channel and aquifer are largely controlled by stream channel form (Vervier *et al.*, 1993).

These results also indicate that a point source of contamination injected in a very restricted part of the floodplain can flow and disperse over a wide area of the aquifer downstream of the contamination source. Water and solute exchanges between surface water and aquifer zone are very complex and depend on hydrological and local morphological conditions (Pretty *et al.*, 2006). Near the stream and especially in

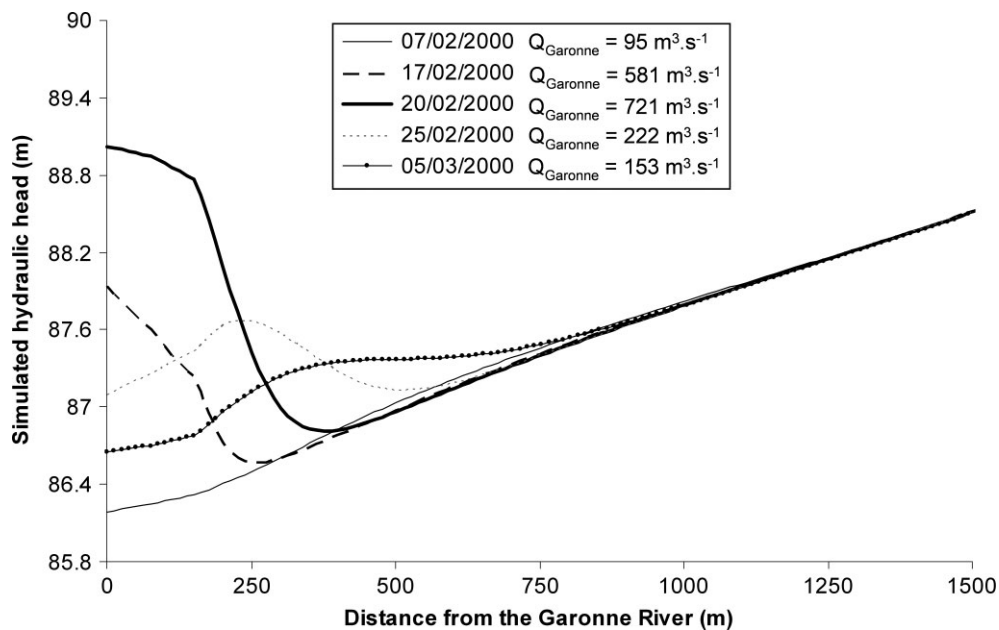


Figure 10. Variations of the hydraulic heads along a transect perpendicular to the Garonne River (from P6 to P23) before, during and after the flood in February 2000

meanders, aquifer–river exchanges can be reversed and flow lines then exhibit bypass phenomena as shown by Weng *et al.* (2003) at a scale of a meander in the Garonne River. This, in turn, can contribute to a dilution, transformation and elimination of contaminants from aquifers, such as contaminants from agricultural fields. For example, dissolved organic carbon fluxes from surface water can be used in macroporous media to complex and/or contribute to eliminating nitrates (from the aquifer) by denitrification (Sanchez-Perez *et al.*, 2003; Pfeiffer *et al.*, 2006). Contaminants from groundwater could be diluted and transformed by mixing with surface water and the resulting concentration in surface water, even if low ($\mu\text{g L}^{-1}$), could be toxic for drinking water ($5 \mu\text{g L}^{-1}$ for cadmium before treatment) and toxic for organisms at lower concentrations ($2 \mu\text{g L}^{-1}$ for cadmium, Mouchet *et al.*, 2006).

CONCLUSIONS

This paper describes the development and application of a model, 2SWEM, to investigate interactions between the Garonne River and its adjacent shallow alluvial aquifer with preferential lateral exchanges. A two-dimensional river model is fully coupled with a two-dimensional vertically averaged aquifer flow model. 2SWEM takes into account the strong physical connection between the river and the aquifer better than earlier models applied to the Garonne River system. It can be used at mesoscale and for steady or transient flow conditions. Comparisons between simulations and observations demonstrate that the proposed model can satisfactorily reproduce the exchanges between the Garonne River and its alluvial aquifer. Some practical applications

of 2SWEM can include (1) investigating the effects of aquifer inputs in the river; (2) tracing contaminants from river to aquifer or vice versa; (3) estimation of groundwater recharge from rivers or vice versa; (4) estimation of fluxes of water and solutes between river and aquifer; and (5) prediction of the impact of pumping wells, *etc.*

The 2SWEM model requires input data that are relatively easy to obtain, and it is easier to apply than more difficult 3D solvers that have been used previously. It was applied to a site where the river fully penetrates a shallow aquifer. 2SWEM might be useful when complex models are not fully necessary, when the objective is, for example, to have an averaged quantification of the hydrology of a site in order to integrate biogeochemistry. When there are dominant lateral exchange processes between river and aquifer, or when the geomorphology of the studied site allows integration over the z direction, it can also be relevant to consider such vertically integrated two-dimensional models.

Future steps in this work will consist of quantifying the fluxes of water and solutes between the river and the aquifer under different hydrological conditions. It will then be possible to integrate biochemical transformations into the transport model to follow reactive nutrients dynamics between the river and the aquifer. 2SWEM will also be coupled with the agronomic model STICS to predict nitrate leaching in vertical and horizontal directions, and nitrate transport to the aquifer and to the river. The complete hydro-biogeochemical model could be used as a tool for better understanding the hydrological and biological responses of floodplains, and could be the starting point to support water policy and integrated water management for environmental support systems.

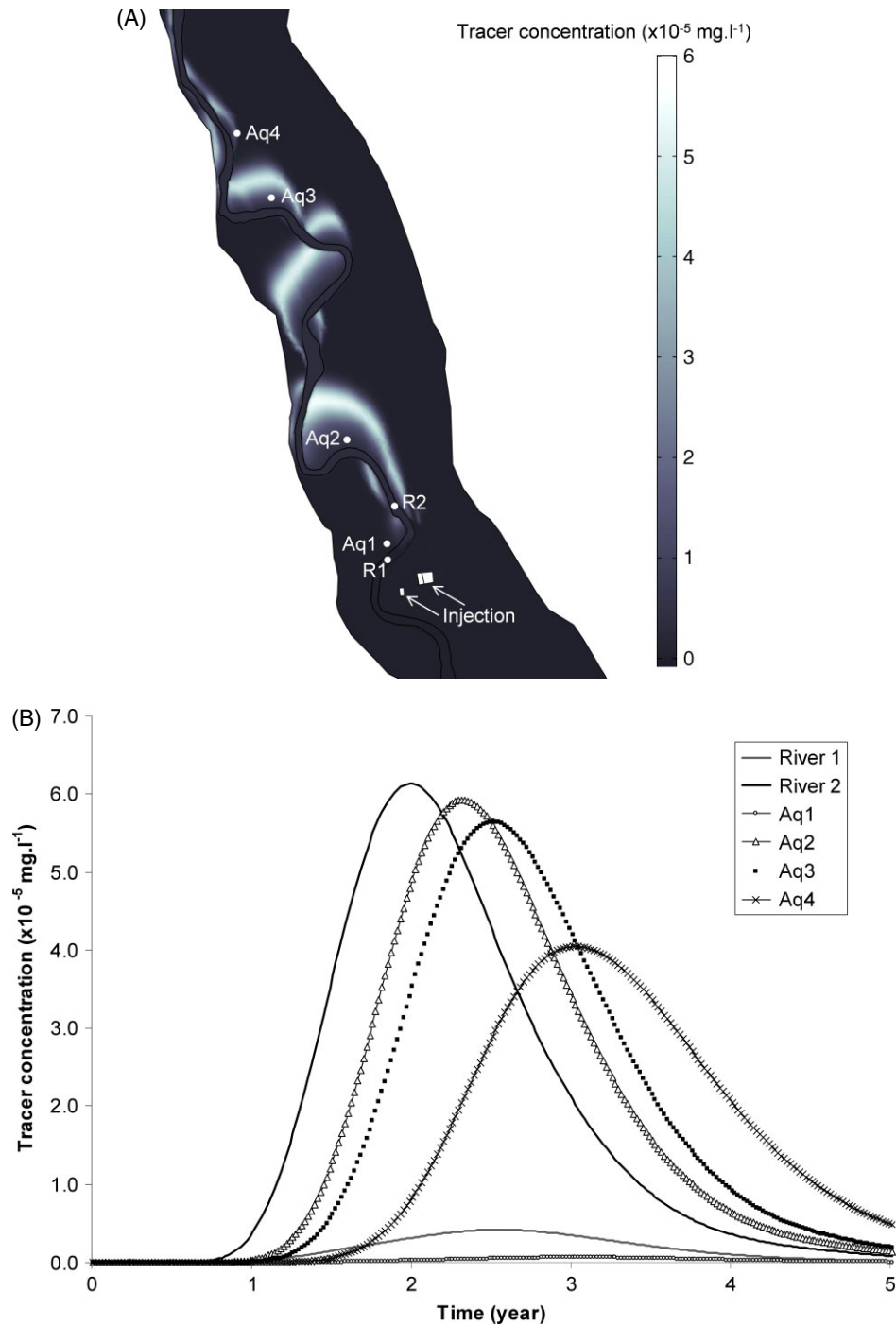


Figure 11. (A) Simulated distribution map of a conservative tracer injected in the floodplain in two plots (initial concentration = 10 mg L^{-1}) after 4 years, and (B) theoretical concentration of the tracer in Aquifer (Aq) and River (R) downstream of the injection plots

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LATERAL EXCHANGES BETWEEN SURFACE AND SUBSURFACE IN LARGE ALLUVIAL FLOODPLAINS

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