

Reporting of Stream-Aquifer Flow Distribution at the Regional Scale with a Distributed Process-Based Model

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Received: 19 December 2013 / Accepted: 13 October 2014 /
Published online: 31 October 2014
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Abstract Groundwater withdrawals can reduce aquifer-to-stream flow and induce stream-to-aquifer flow. These effects involve potential threats over surface water and groundwater quantity and quality. As a result, the description of stream-aquifer flow in space and time is of high interest for water managers. In this study, the *EauDyssée* platform, an integrated groundwater/surface water model is extended to provide the distribution of stream-aquifer flow at the regional scale. The methodology is implemented over long periods (17 years) in the Seine river basin (76 375 km², France) with a 6 481 km long simulated river network. The study scale is compatible with the scale of interest of water authorities, which is often larger than study scales of research projects. Net and gross stream-aquifer exchange flow are computed at the daily time step over the whole river network at a resolution of 1 km. Simulation results highlight that a major proportion of the main stream network (82 %) is supplied by groundwater. Groundwater withdrawals induce a reduction of net aquifer-to-stream flow (−19 %) at the basin scale and flow reversals in the vicinity of pumping locations.

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Such an integrated model provided at the appropriate regional scale is an essential tool provided to water managers for the implementation of the EU Water Framework Directive.

Keywords Surface water - groundwater interactions · Regional modeling · Distributed process-based hydrological model · Seine river basin · European water framework directive

1 Introduction

Given demographic and climatic evolutions, contemporary societies face critical issues regarding the quantity and quality of their water resources (Sophocleous 2007; de Marsily 2008). Withdrawals from surface and groundwater bodies for domestic, industrial and agricultural needs stress the water resources, that must be preserved for economical (e.g. domestic and agricultural water supply, nuclear plant cooling) and ecological reasons (e.g. wetland conservation, protection of water and terrestrial ecosystems, fisheries) (Tellam and Lerner 2009).

In the European Union, the Water Framework Directive (WFD 2000) compels member states to address qualitative and quantitative issues of water resources. Practically, this consists in setting appropriate groundwater abstraction and pollutant release licenses and conducting attenuation or remediation operations to achieve good chemical, ecological and quantitative status of water bodies. To this effect, water authorities must have a good understanding of governing hydrological processes as well as quantitative data to assess the impact of human activities or environmental change (Dahl et al. 2007; Wagener et al. 2010). In addition, policy design (River Basin Management Plan) is improved when different scenarios can be tested with appropriate modeling tools (Barthel et al. 2012; Goderniaux et al. 2009; Tellam and Lerner 2009; Yang and Wang 2010). However, several factors impede the transfer of relevant scientific data and tools for decision making (Borowski and Hare 2007; Quevauviller et al. 2005). Due to the lack of large-scale datasets and difficulties in model implementations, researchers often investigate processes at the local scale of a study site, while water managers usually need results at the catchment scale (Krause et al. 2011; Nalbantis et al. 2011).

Among the processes governing hydrosystems, the exchange flow between streams and aquifers plays a critical role and entails scientific challenges (Krause et al. 2009; Sophocleous 2007). Rivers can drain or, on the opposite, recharge aquifers. In temperate regions, streams often remain in a gaining regime and base flow from groundwater constitutes the main contribution to the river flow (Kalbus et al. 2009; Mas-Pla et al. 2013; Pinder and Jones 1969; Tóth 1963). This is the opposite in arid regions, where rivers are usually in a losing state and stream-to-aquifer flow constitutes the main recharge to the aquifers (McCallum et al. 2012a). Though a losing, or gaining regime can be dominant at the river catchment scale, the stream-aquifer interactions are usually neither permanent in time nor spatially homogeneous at the basin scale. There are temporal cycles and rivers can gain water from aquifers along certain reaches and lose water from others (Bencala et al. 2011; Jha et al. 1999). Furthermore, many aquatic and terrestrial ecosystems depend on flow dynamics and nutrient fluxes at the stream-aquifer interface (Bertrand et al. 2012). For these reasons, a deep understanding and consideration of the stream-aquifer interface is mandatory for water resources management in quantitative, qualitative, and ecological terms.

Numerous methods have been developed to quantify stream-aquifer flow (Mouhri et al. 2013). However, accurate and representative estimates at the large scale remain difficult to obtain. The direct measurement of groundwater inflows or outflows from a river reach to

an aquifer by differential gaging or seepage measurements is uncertain because of measurement errors and lack of spatial representativity (McCallum et al. 2012b; Rosenberry 2008). Many analytical solutions for stream depletion by pumping wells have been developed during the last decades (Hantush 1965; Kirk and Herbert 2002; Serrano and Workman 1998). These equations can provide an estimate of stream to aquifer flow. They are relevant at the local scale for simplified geometries but inapplicable for long term, regional scale studies (Lin and Medina 2003). Field investigations and monitoring network can be used to infer stream-aquifer flow with detailed small scale calibrated numerical models (Hunt et al. 2006; McCallum et al. 2012b; Mouhri et al. 2013; Sophocleous et al. 1988). Results from densely monitored study sites and detailed fine scale models can provide accurate data and investigate the processes of stream-aquifer interactions (Frei et al. 2009; Kikuchi et al. 2012; Mouhri et al. 2013). However, these studies imply extensive and costly monitoring techniques and their implementation is very rare at the regional scale, which is the water authorities working scale (Krause et al. 2007, 2009).

Neither field direct observations, nor analytical equations, nor detailed numerical modeling can be used to estimate stream-aquifer flow at the regional scale. As a result, regional hydrosystem modeling (Flipo et al. 2012) presents a relevant alternative. As reviewed by Sparks et al. (2013) and Flipo et al. (2014), several distributed, process-based models (DPBM) are now capable of simulating surface-subsurface hydrological processes: *InHM* (VanderKwaak and Loague 2001), *Eaudyssée* (Flipo et al. 2012), *MODHMS* (HydroGeoLogic Inc., 2006) (Panday and Huyakorn 2004), *HydroGeoSphere* (Brunner and Simmons 2012), *ParFlow* (Kollet and Maxwell 2006), and *MIKE SHE* (Refsgaard and Storm 1995). Providing spatial and temporal distribution of stream-aquifer flow to stakeholders at the regional scale is a critical need (Krause et al. 2011). However, the number of applications of such integrated models at the regional scale remains limited (Brunner and Simmons 2012; Flipo et al. 2014). In addition, the majority of studies in which stream-aquifer interactions are quantified is found to be at the local and intermediate scales (less than 4000 km²) (Flipo et al. 2014; Refsgaard et al. 1998; Said et al. 2005; Xevi et al. 1997; Zhang and Werner 2009). Some attempts have been made at limited scales in urban context (Ellis et al. 2007), but to the authors' knowledge, the description of stream-aquifer flow at the regional scale (i.e. > 10 000 km²) has not been reported in the literature yet.

This study details and discusses a practical methodology to describe the distribution of stream-aquifer flow in space and time. The *Eaudyssée* integrated modeling platform is first presented with a focus on surface-subsurface coupling. This paper continues with a case-study in a large, densely populated water basin with a temperate climate, the Seine river basin (France). The dynamics and magnitude of stream-aquifer flow estimates are detailed for long term simulations along the 6 481 km simulated river network. The impact of groundwater withdrawals over these interactions is described. The relevance and applicability of this approach as a tool for integrated planning of hydrosystems is eventually discussed.

2 Integrated Hydrosystem Modeling

2.1 The *Eaudyssée* Platform for Hydrosystem Modeling

An integrated, distributed process-based model, the *Eaudyssée* platform (Flipo et al. 2012; Saleh et al. 2011) is used in this study. This platform has successfully simulated surface and groundwater flow in many basins of various scales and different hydrogeological settings,

such as the Seine basin, 76 375 km² (Gomez et al. 2003; Ledoux et al. 2007; Saleh et al. 2011), the Rhône basin, 87000 km² (Etchevers et al. 2001), the Upper Rhine basin, 13 900 km² (Thierion et al. 2012), the Somme basin, 6433 km² (Habets et al. 2010; Korkmaz et al. 2009) and the Loire basin, 117 500 km² (Monteil 2011). This study is also based on the *Eaudyssée* platform, but we extended the method for stream-aquifer coupling and provide for the first time, practical considerations for the description of stream aquifer flow (Section 2.2). The main modules of the *Eaudyssée* platform are briefly presented hereafter.

In the *Eaudyssée* platform, the hydrosystem is conceptually divided into three components: surface, subsurface unsaturated and subsurface saturated. Separate modules handle

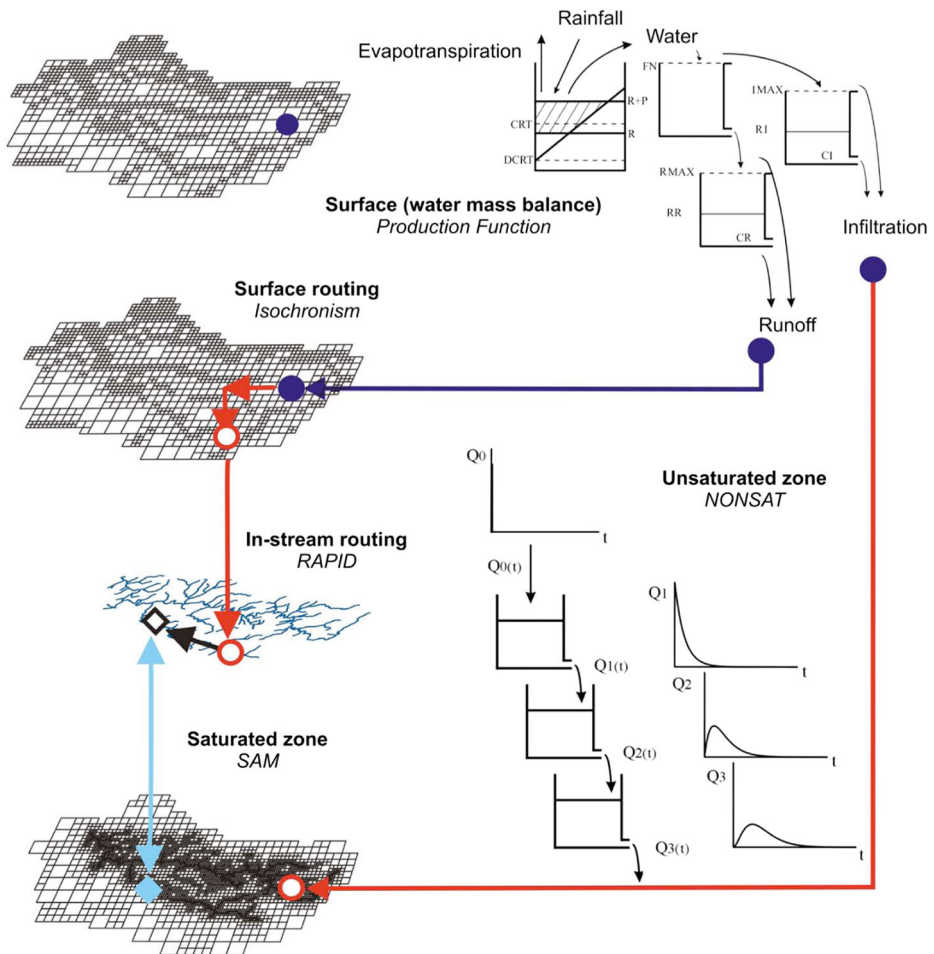


Fig. 1 Structure of the *Eaudyssée* hydrosystem modeling platform (Flipo et al. 2012). The surface water budget (top right) is based on variables R , $R + P$, RR and RI that are computed at each time step. Colored arrows display the successive steps of computation during one time step. Light blue arrow corresponds to the stream-aquifer coupling

the simulation of water flow within and between respective components: surface water budget, runoff, river flow, stream-aquifer interactions, unsaturated flow, and groundwater flow (Fig. 1).

The surface component is a conceptual model based on the soil water-budget. From meteorological input data (precipitation and potential evapotranspiration), the surface model component handles the partition of precipitation into runoff (resulting from overland flow and interflow) and computes evapotranspiration and infiltration. The influence of land use and soil cover is considered with a set of 7 parameter-models distributed in space (Deschesnes et al. 1985).

Runoff routing to the river network is based on isochronal zones (Flipo et al. 2012). This method requires the identification of the nearest river cell for each surface grid cell, which is obtained with a digital elevation model. Once the water reaches a river cell, the routing is performed with the *RAPID* module, which is based on the one-dimensional *Muskingum* scheme (David et al. 2011). To reduce the computational burden, river reaches of small upstream watersheds are not explicitly simulated. Their contribution is taken into account with surface seepage from unconfined aquifers but is not considered as stream-aquifer flow in the model.

Infiltrated water percolating below the root zone is directed to the groundwater component after a transfer through the unsaturated zone. To consider this transit time, a delay is imposed between surface infiltration and aquifer recharge using a series of identical linear reservoirs (Gomez et al. 2003; Flipo et al. 2012).

The groundwater component of *Eaudyssée* is based on the *NEWSAM* model (Ledoux et al. 1984, 1989). This model solves the diffusivity equation with a quasi-3D finite-difference scheme, which is adapted to large, layered sedimentary basins (de Marsily 1986). Similarly to MODFLOW (McDonald and Harbaugh 1988), vertical flow between aquifers is computed with a conductance model.

2.2 Surface-Subsurface Coupling

Water routed within the river network is the sum of direct flow originating from runoff and base flow originating from aquifer seepage. As rivers may be in losing or gaining regimes, the river component must be coupled with the groundwater component.

Given the relative small size of actual river width with respect to the regional subsurface mesh (1 km), water level cannot be considered as continuous at the stream-aquifer boundary (Ebel et al. 2009; Mehl and Hill 2010). As a consequence, river cells are coupled to groundwater cells with a conductance model based on a river coefficient (Ebel et al. 2009; Rushton 2007):

$$Q = RC \times L \times (H_{riv} - H_{aq}) \quad (1)$$

where Q [$\text{m}^3 \text{s}^{-1}$] is the stream-aquifer flow, counted as positive for river-to-aquifer flow, RC [m s^{-1}] is the river coefficient and H_{riv} and H_{aq} [m] are the hydraulic heads in the river and the aquifer, respectively. The nested mesh used in this model is refined along the rivers for the mesh size to be a good approximation for L , the length of the river reach in the aquifer mesh.

The river coefficient RC accounts for head losses in the geological medium (riverbed deposit and underlying rocks), river geometry, additional head losses due to converging/diverging flow that cannot be considered in the 2D horizontal model (Morel-Seytoux 2009; Liggett et al. 2012; Rushton 2007). Rushton (2007) shows that the aquifer horizontal

hydraulic conductivity can be considered as the main control of RC at the regional scale. Based on this result, RC is expressed as follows:

$$RC = f \times K_h \tag{2}$$

where K_h is the aquifer horizontal hydraulic conductivity and f is an adjustable, lumped parameter usually found within the [0.9; 1.2] interval (Rushton 2007). f can be considered as the extension of the “turning factor” mentioned by Morel-Seytoux (2009), who focused on the effect of head losses due to converging flow around the stream. In this study, parameter f also considers any other parameter likely to control stream-aquifer flow.

The conductance model (1) requires both groundwater and in-stream water levels. Contrary to the majority of models based on the conductance model, a variable river stage is used in this study. This has been shown to improve stream-aquifer flow estimate and the variability of near-stream groundwater level fluctuations (Saleh et al. 2011). However, the river network routing component (RAPID) simulates stream flow, not water level (David et al. 2011). In-stream water levels are therefore inferred from simulated stream flow values using the Manning equation (Chow 1959):

$$H_{riv} = Z_{bed} + \left[\frac{nQ}{B\sqrt{S}} \right]^{\frac{3}{5}} \tag{3}$$

with:

| | |
|-------------------|----------------------------------|
| $H_{riv}[m]$, | in-stream water level, |
| $Z_{bed}[m]$, | river bed elevation, |
| $Q[m^3 s^{-1}]$, | simulated stream flow, |
| $n[sm^{-1/3}]$, | Manning’s roughness coefficient, |
| $S[-]$, | free surface slope, |
| $B[m]$, | river cross section |

At each time step and each river mesh, the model records the values of aquifer-to-stream flow, $Q_{A \rightarrow S}$ and stream-to-aquifer flow, $Q_{S \rightarrow A}$. For a given time interval, the gross exchange flow over the whole basin, Q_{gross} is defined by:

$$Q_{gross} = \sum_i \sum_j (Q_{A \rightarrow S, (i, j)} + Q_{S \rightarrow A, (i, j)}) \tag{4}$$

where $Q_{A \rightarrow S, (i, j)}$ and $Q_{S \rightarrow A, (i, j)}$ are the values of aquifer-to-stream flow and stream-to-aquifer flow, respectively, for the i -th river cell at the j -th time step. Similarly, the net exchange flow Q_{net} , is defined by :

$$Q_{net} = \sum_i \sum_j (Q_{A \rightarrow S, (i, j)} - Q_{S \rightarrow A, (i, j)}) \tag{5}$$

Therefore, a positive Q_{net} over a time interval means that the considered river reach is overall in a gaining configuration. Q_{gross} and Q_{net} may be expressed for the whole simulated river network, in $m^3 s^{-1}$, or for a given river reach, per unit river length, in $m^3 s^{-1} km^{-1}$.

3 Implementation of the Seine River Basin Model

3.1 The Seine River Basin

The Seine river basin (76 375 km², Fig. 2), located in the north of France, is the most urbanized and industrialized French basin as it gathers ca. 15 million inhabitants, corresponding to more than 20 % of the total French population (Billen et al. 2007). Urban areas represent 5 % of the surface, they are concentrated along the main tributaries of the Seine river and its estuary, with a particularly high population density of 20 000 people per km² around the city of Paris (INSEE 2010). The Seine river basin is important for national food production as a large portion of the land surface (52 %) is devoted to agriculture. The rest of the land is covered by forest (25 %) and grass-land (18 %). Given its important population and food production, water resources in the Seine river basin are of high strategic importance : ~ 1 × 10⁹ m³ groundwater is extracted every year over the whole basin, which represents 45 % of total water withdrawals, the rest being taken from surface water (AESN and DRIEE 2013).

The Seine river and its tributaries lay in the Parisian sedimentary basin, which is one of the major French geological regions having developed since the Triassic. This sedimentary basin is a composite of carbonate and sandy formations interbedded by poorly permeable clayey units (Guillocheau et al. 2000).

The climatic regime of the Seine river basin is pluvial oceanic, modulated by seasonal variations in evapotranspiration. This regime leads to high river flow in winter and low river flow in summer that is sustained by base flow from the aquifers. In addition to these

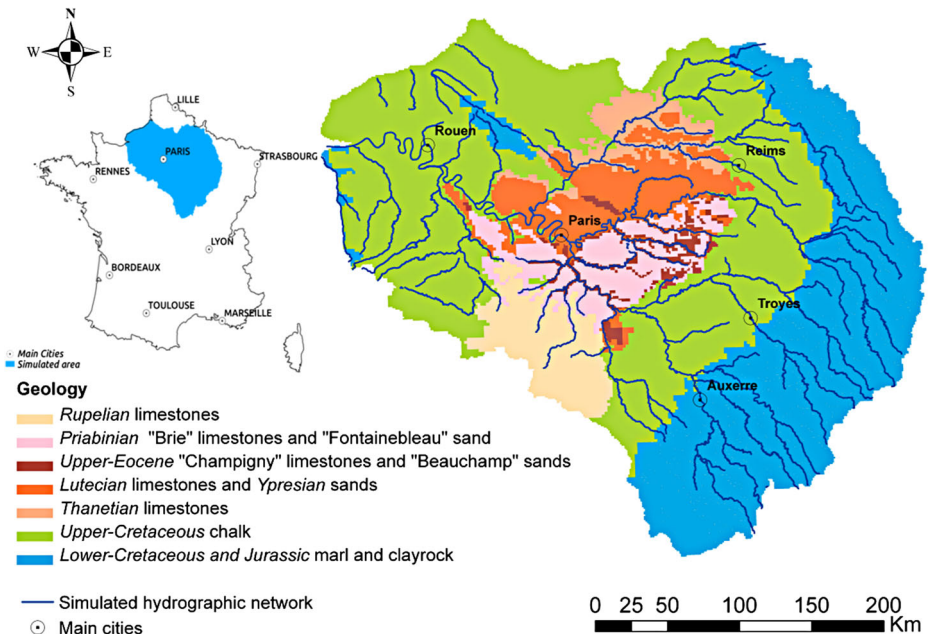


Fig. 2 The Seine river basin (76 375 km²), with the simulated river network (6 481 km) and the six aquifer layers. The simulated river network extends over the poorly permeable *Lower-Cretaceous* and *Jurassic* units where groundwater flow is not considered

seasonal variations, a 17-year cycle associated with the *North Atlantic Oscillations* (NAO) have been revealed in climatic variables and stream discharge (Massei et al. 2010) as well as groundwater levels (Flipo et al. 2012). Given this periodic variability, long term trend hydrological analysis of the Seine river basin is only relevant for simulation periods covering a whole number of 17-year cycles (de Fouquet 2012). Over the 1989-2006 period, mean annual rainfall in the Seine river basin was 784 mm (Vidal et al. 2010).

3.2 Model Structure, Boundary Conditions and Discretization

The surface mesh used in this study (95 560 km²) covers the whole watershed of the Seine river (76 375 km²) and extends outward to include groundwater hydraulic boundaries constituted by rivers of the adjacent basins (Somme, Meuse, Loire, Loir and Touques) (Fig. 2). This surface is discretized into a progressive nested square mesh of 35 198 cells with cell sizes varying from 1 to 8 km. Among these cells, 6 481 river cells are refined to 1 km × 1 km to represent river reaches with drainage areas equal to, or larger than 250 km². As a result, the simulated river network is 6 481 km long and presents Strahler order below 3 to 4 (Gomez et al. 2003). Rivers with a smaller drainage area are not explicitly modeled, but base flow to the main river reach from these small tributaries is considered by seepage from unconfined aquifers.

This study is based on an existing hydrogeological model which considers the main aquifers in interaction with the Seine river and its tributaries (Viennot 2009). From top to bottom, these aquifers can be summarized as follows :

1. *Rupelian* limestones
2. *Priabonian* “Brie” limestones and “Fontainebleau” sands
3. *Upper-Eocene* “Champigny” limestones and “Beauchamp” sands
4. *Lutetian* limestone and *Ypresian* sands
5. *Thanetian* limestones
6. *Upper-Cretaceous* chalk

Similarly to the surface mesh, the groundwater domain is discretized with a progressive nested square mesh (41 609 cells of 1 km² to 16 km²). The confined/unconfined aquifer state is assumed to remain the same throughout the simulation and set with the value of the storage coefficient. A no-flow boundary conditions is imposed at the groundwater mesh border, except where the mesh boundary corresponds to an adjacent river. In this latter configuration, a fixed-head boundary condition is imposed. The Upper-Cretaceous chalk aquifer forms the general stratum of the layered system. The underlying unit of poorly permeable lower-Cretaceous and Jurassic marl and clay rocks is not modeled. However, the river network developing over this lower geological unit (Fig. 2) is explicitly simulated.

Groundwater withdrawals are taken into account with a mean daily pumping rate, inferred from the annual groundwater abstraction data of the Seine basin water management agency (Agence de l’eau Seine-Normandie) for domestic, industrial and agricultural needs. The agricultural withdrawals have been supposed to take place only during the irrigation period, i.e. from June to September.

3.3 Input Data

The surface water budget component is parameterized in accordance with the land use from the European *Corine Land Cover* map (European Environment Agency, 2006) and the

French soil type map from the *Institut National de Recherche Agronomique* (INRA). Given land use and soil type distributions, the basin surface is classified into 31 *production functions* (Gomez et al. 2003) with respective set of parameters for the surface water budget component.

The climatic input data (precipitation and potential evapotranspiration) is taken from the Météo-France SAFRAN database on the 1971–2010 record (Quintana-Seguí et al. 2008; Vidal et al. 2010).

3.4 Model Calibration

Initial model parameters were taken from the calibration results of Viennot (2009). So as to improve model performances, the Seine basin hydrogeological model has been re-calibrated over the 1996–2006 period using a set of 118 stream gages (discharge and water levels) and 183 groundwater monitoring wells distributed among the six aquifer units of the model (Figs. 4 and 5). The calibration followed the stepwise iterative procedure detailed by Flipo et al. (2012) :

1. **Surface water** component calibration: *production function* parameters that control infiltration, soil water storage and evapotranspiration.
2. **Groundwater** component calibration: aquifer transmissivity, storage coefficient, and conductance coefficients between horizontal layers as well as the f coefficient in Eq. 2.

These two steps are conducted iteratively. Within each step, parameters are adjusted by manual trial and error. To facilitate the calibration operation, performance criteria (bias, *Nash* efficiency, and root mean square error) at every observation locations (gaging stations and observation wells) are automatically exported to Qgis (QGIS Development Team 2013). Surface parameters are adjusted in function of simulated river discharge bias and *Nash* efficiency at gaging stations. Aquifer transmissivity and storage coefficient were adjusted from the distribution of bias and RMSE on water levels at the observation wells. “Informative” performance criteria are preferred. In particular, the bias not only provides a quality estimate, but also informs the modeler on how to adjust aquifer transmissivity to improve model performance.

For the simulation of in-stream water levels, the Manning’s roughness coefficient, n is optimized with more than 60 gaging stations where both levels and flow records were available. Rating curves were adjusted to the Manning’s formula (3) with the root mean squared error (RMSE) as objective function. The Seine River center-lines, width and longitudinal slopes for the Manning’s equation are obtained from the digital elevation model.

3.5 Simulation Scenarios

In order to analyze the natural variability of the hydrosystem and describe the effects of groundwater abstraction, several simulation scenarios have been considered in this study (Table 1). The effects of climatic variability are investigated with 3 simulation periods, corresponding to the long term conditions (simulation LTC), a dry hydrological year (simulation DY) and a wet hydrological year (simulation WY). For each of these simulation periods, the model has been run twice: once considering actual pumping rates and once with pumping rates set to zero (natural situation without groundwater abstraction). The records of the Seine river discharge at Paris (Austerlitz) present contrasting characteristics for these three simulation periods (Fig. 3a). The long-term behavior (LTC) is characterized by an

Table 1 Simulations considered in this study present different climatic conditions: dry year (DY), wet year (WY), and long term conditions (LTC)

| Name | Configuration | Date range | Pumping | \bar{P} [mm] | \bar{Q} [m ³ s ⁻¹] |
|--------|---------------|------------|---------|----------------|---|
| CAL | Long-term | 1995-2006 | yes | 775 | 310 |
| VAL | Long-term | 1984-1995 | yes | 774 | 281 |
| TEST | Long-term | 1984-2006 | yes | 784 | 285 |
| LTC | Long-term | 1989-2006 | yes | 784 | 285 |
| LTCnoP | Long-term | 1989-2006 | no | 784 | 285 |
| DY | Dry | 1995-1996 | yes | 540 | 185 |
| DYnoP | Dry | 1995-1996 | no | 540 | 185 |
| WY | Wet | 2001-2002 | yes | 1113 | 550 |
| WYnoP | Wet | 2001-2002 | no | 1113 | 550 |

¹ \bar{P} stands for mean annual rainfall over the Seine river basin and \bar{Q} for the mean annual discharge rate of the Seine river at Paris gaging station

extended flood event from November to June. The dry year (DY) is characterized by a low discharge rate all along the year, except for two consecutive, moderate flood events in January and February. In contrast, the Seine river during the wet year (WY) presents a long,

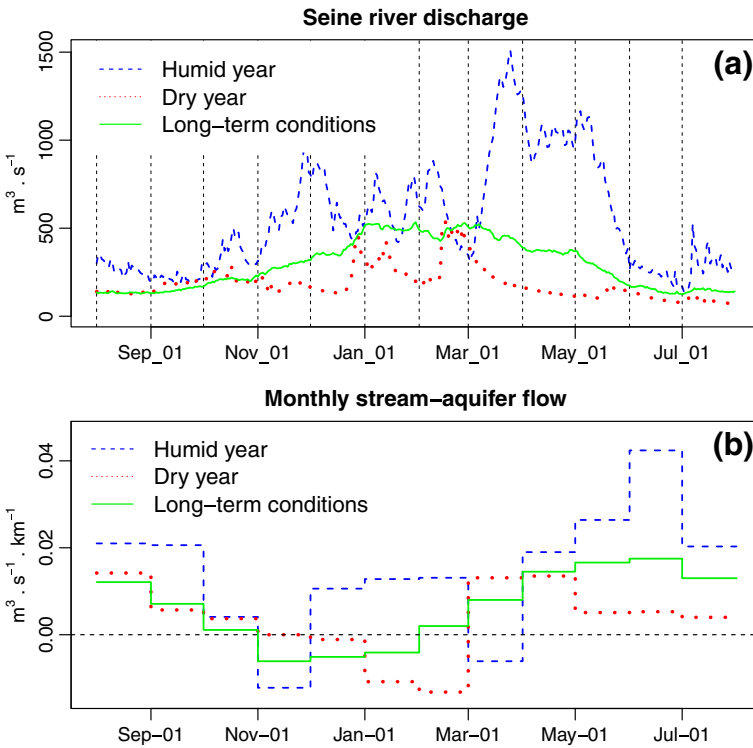


Fig. 3 a Seine river discharge at Paris gaging station and (b) net stream-to-aquifer flow over the reach outlined in Fig. 6, compared for long-term conditions (LTC), dry year (DY) and wet year (WY) (Table 1)

but moderate flood event from November to March, followed by an intense flood event from April to May.

4 Results

4.1 Model Performances

From the initial parameter set (Viennot 2009), ca. 200 model runs have been completed to obtain satisfactory model performances, dropping the overall RMSE of simulated groundwater levels from 10.64 m to 3.69 m and increasing the *Nash* efficiency at the basin outlet from 0.59 to 0.80. Calibration results are presented in Table 2 and the spatial distribution of performance criteria in Figs. 4 and 5. To assess the representativity and stationarity of calibrated parameters, validation and test runs were conducted over distinct time intervals (Table 2). Performance criteria are relatively stable for the calibration, validation and test runs, which confirms the parameter set (Flipo et al. 2012; Kurtulus and Flipo 2012; Maier and Dandy 2000).

The weak value of the bias obtained for simulated river discharge at the basin outlet (Table 2) confirms the estimate of evapotranspiration at the basin scale. The simulation of peak-flow revealed with the quality of *Nash* efficiency (0.8), at least in the main river network, validates the partition of rainfall into run-off and infiltration (Fig. 5). Stream-aquifer flow cannot be directly measured, but is indirectly validated with the quality of base flow estimates.

4.2 Stream-Aquifer Response to Contrasted Climatic Conditions

The calibrated model is used to analyze the variability of stream-aquifer interactions in the Seine river basin under different climatic contexts, with simulations LTC, DY and WY

Table 2 Model performances for groundwater levels (top) and river discharge at basin outlet (bottom)

| Groundwater Levels | | | | | | |
|---------------------------------|------------------|----------|------------------|----------|------------------|----------|
| Period: | Cal. (1995-2006) | | Val. (1984-1995) | | Test (1984-2006) | |
| Aq. | bias [m] | rmse [m] | bias [m] | rmse [m] | bias [m] | rmse [m] |
| 1 | +0.39 | 2.71 | -0.22 | 3.02 | -0.11 | 2.82 |
| 2 | -2.06 | 5.28 | -2.88 | 5.04 | -2.77 | 5.38 |
| 3 | +0.8 | 3.75 | +1.16 | 3.38 | +0.86 | 3.76 |
| 4 | -0.12 | 4.6 | -0.11 | 7.63 | -0.26 | 6.16 |
| 5 | -0.73 | 0.91 | -0.09 | 0.29 | -0.08 | 0.52 |
| 6 | +0.44 | 3.54 | -0.24 | 3.33 | +0.13 | 3.58 |
| All | +0.31 | 3.64 | -0.05 | 4.23 | +0.04 | 3.95 |
| River Discharge at Basin Outlet | | | | | | |
| Period: | Cal. (1995-2006) | | Val. (1984-1995) | | Test (1984-2006) | |
| | bias [%] | Nash [-] | bias [%] | Nash [-] | bias [%] | Nash [-] |
| | -2 | 0.81 | -2 | 0.80 | -2 | 0.80 |

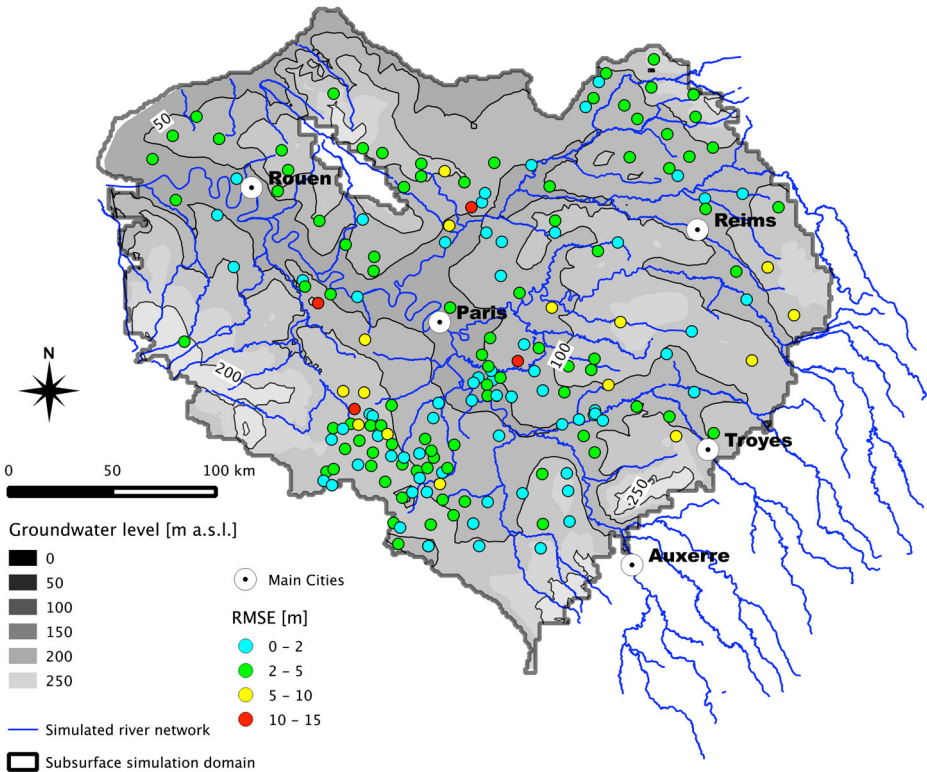


Fig. 4 Performance criteria (RMSE) for groundwater levels over the calibration period (CAL, Table 1). Contours and grey-scale map in the background present simulated mean groundwater levels for the uppermost (unconfined) aquifer during the same period

(Table 1). Such an analysis is of interest for water managers as the average behavior (simulation LTC) may not be representative of fluctuating conditions. For all of these simulations, *current* pumping rates are considered.

For the long term simulation (LTC), the majority of the river network is in a gaining configuration (Fig. 6), which means that on the whole, the aquifer system supplies water to the river network. The mean ratio of stream length in gaining configuration is 82 % (2 668 km out of the 3 254 km simulated river length in contact with an aquifer). This ratio reaches its highest value at the end of winter, from February to March (86 %) and its lower value in November (76 %). The mean ratio is weaker for the dry year (80 %) and higher for the wet year (86 %), which highlights that gaining configurations are more common in wet years.

For gaining streams, the net aquifer-to-stream flow (Q_{net} , Eq. 5) ranges between 0 and $+0.1 \text{ m}^3 \text{ s}^{-1}$ per km river length. Only few reaches are in a losing configuration, with negative Q_{net} (reddish colors). Over the whole simulated river network, the total net aquifer-to-stream flow is in average $57.56 \text{ m}^3 \text{ s}^{-1}$ (Table 3). As a result, the simulated river network contributes to ca. 10 % of the average flow rate of the Seine river at the basin outlet during the same time interval ($506 \text{ m}^3 \text{ s}^{-1}$). Due to the occurrence of stream-to-aquifer flow, the total gross exchange flow (Q_{gross} , Eq. 4) reaches $95.76 \text{ m}^3 \text{ s}^{-1}$ and is 66 % higher than Q_{net} .

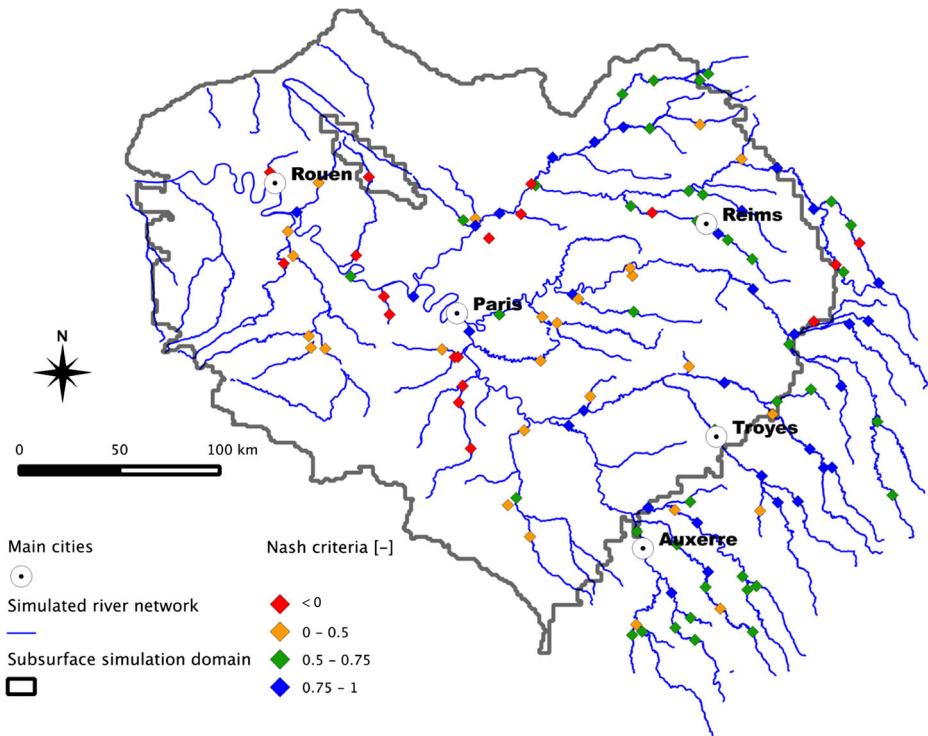


Fig. 5 Performance criteria (*Nash* efficiency) for river discharge during the calibration period (CAL, Table 1)

The temporal variability of stream-aquifer flow for a reach of the Seine river near Paris (delineated in Fig. 6) is presented in Fig. 3b. This river reach corresponds to a “river water body”, the EU Water Framework Directive unit for water management (Wasson et al. 2003). This reach remains most of the time in a gaining regime, but reversals occur at different occasions depending on climatic conditions. For the long term conditions (simulation LTC), stream-to-aquifer flow tends to occur at the beginning of high-flow (from November to January), but configurations are contrasting between dry and humid periods. Only one reversal is observed during the dry year (simulation DY), while two flow reversals can be observed during the wet year (simulation WY). Overall, the variability of stream-aquifer flow over the reach of interest is greater in humid year and the gross exchanged flow of humid year is about tenfold the value for the dry year.

For the humid period, it can be observed that stream-aquifer flow reversals (Fig. 3b) occur at the beginning of flood periods (November and March), which are both characterized by a rapid increase in river discharge (Fig. 3a). This has already been observed by Jha et al. (1999) and has been explained by Saleh et al. (2011). River stage rapidly rises during a flood event, while groundwater levels rise is slower, which induces stream-to-aquifer flow.

4.3 Impact of Groundwater Withdrawals

So as to investigate the effects of groundwater withdrawals over stream-aquifer interactions, the model has been run twice for each climatic conditions: once considering current

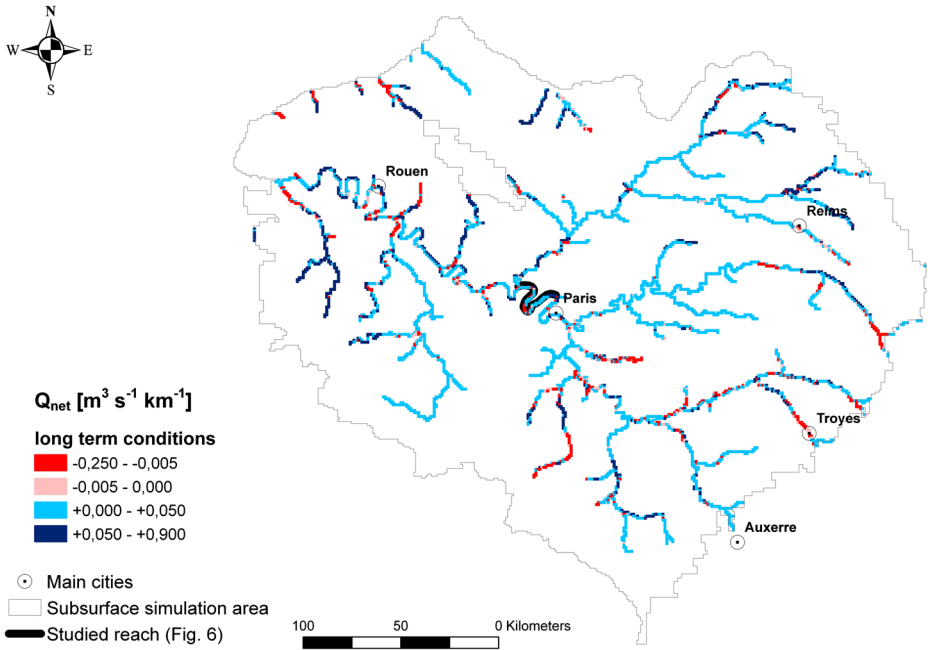


Fig. 6 Distribution of stream-aquifer flow in the Seine river basin for long term conditions (LTC simulation, Table 1). The reach bounded by *black lines* upstream Paris was used for the temporal analysis presented in Fig. 3

pumping rate and once with pumping rates set to zero (Table 1). The effects of groundwater withdrawals can be described with the difference of exchange flow obtained from the simulations with and without considering current pumping rates. For example, for the long term simulation LTC, the variation of net aquifer-to-stream flow attributable to groundwater withdrawals, ΔQ_{net}^{LTC} , reads:

$$\Delta Q_{net}^{LTC} = Q_{net}^{LTC} - Q_{net}^{LTCnoP} \tag{6}$$

where Q_{net}^{LTC} and Q_{net}^{LTCnoP} are the net aquifer-to-stream flow for simulations LTC and LTCnoP, respectively.

At the basin scale, for the long term simulation, groundwater withdrawals reduce the total aquifer-to-stream flow ($-8.24 \text{ m}^3\text{s}^{-1}$, -10%) and increase the total stream-to-aquifer flow ($+2.98 \text{ m}^3\text{s}^{-1}$, $+15 \%$) (Table 3). The associated reduction of net aquifer-to-stream flow, ΔQ_{net}^{LTC} , is $-11.22 \text{ m}^3\text{s}^{-1}$ (-19%). This corresponds to approximately one third of the average total pumping rate over the Seine basin ($32 \text{ m}^3\text{s}^{-1}$). The reduction of net aquifer-to-stream flow is approximately the same for the long term, dry and wet climatic conditions (Table 3).

In the Seine river basin, the large majority of pumping stations is located in the vicinity of the main stream network (Fig. 7). As expected, the effects of pumping over stream-aquifer flow are concentrated in the vicinity of the main pumping locations. The reduction of net stream-to-aquifer flow can reach $0.1 \text{ m}^3\text{s}^{-1} \text{ km}^{-1}$. As a result, groundwater withdrawals can be at the origin of the negative net stream-to-aquifer flow observed around the cities of Paris and Rouen (Fig. 6). This may present a threat to groundwater quality in these areas.

Table 3 Total stream-aquifer flow for the whole simulated river network in the Seine river basin. With $Q_{A \rightarrow S}$, aquifer-to-stream flow, $Q_{S \rightarrow A}$, stream-to-aquifer flow, Q_{gross} , gross exchange flow and Q_{net} , net aquifer-to-stream flow

| Simulation | Stream-aquifer flow [$m^3 s^{-1}$] | | | |
|------------------------|--------------------------------------|-----------------------|-------------|-----------|
| | $Q_{A \rightarrow S}$ | $Q_{S \rightarrow A}$ | Q_{gross} | Q_{net} |
| LTC | 76.66 | 19.10 | 95.76 | 57.56 |
| LTCnoP | 84.90 | 16.12 | 101.02 | 68.78 |
| $\Delta(LTC - LTCnoP)$ | -8.24 | +2.98 | -5.26 | -11.22 |
| DY | 64.83 | 14.52 | 79.35 | 50.31 |
| DYnoP | 72.59 | 11.76 | 84.35 | 60.83 |
| $\Delta(DY, DYnoP)$ | -7.76 | +2.76 | -5.00 | -10.52 |
| WY | 111.70 | 25.13 | 136.83 | 86.57 |
| WYnoP | 119.92 | 22.35 | 142.27 | 97.57 |
| $\Delta(WY, WYnoP)$ | -8.22 | +2.78 | -5.44 | -11.00 |

¹The effects of groundwater withdrawals are quantified from the differences of exchange flows between simulations considering pumping rates and simulations with pumping rates set to zero

At the basin scale, the average ratio of the total length of river in gaining configuration is lower for the simulation *LTC*, considering actual pumping rate (82 %), than for the simulation *LTCnoP* with pumping rates set to zero (97 %). This reveals that groundwater

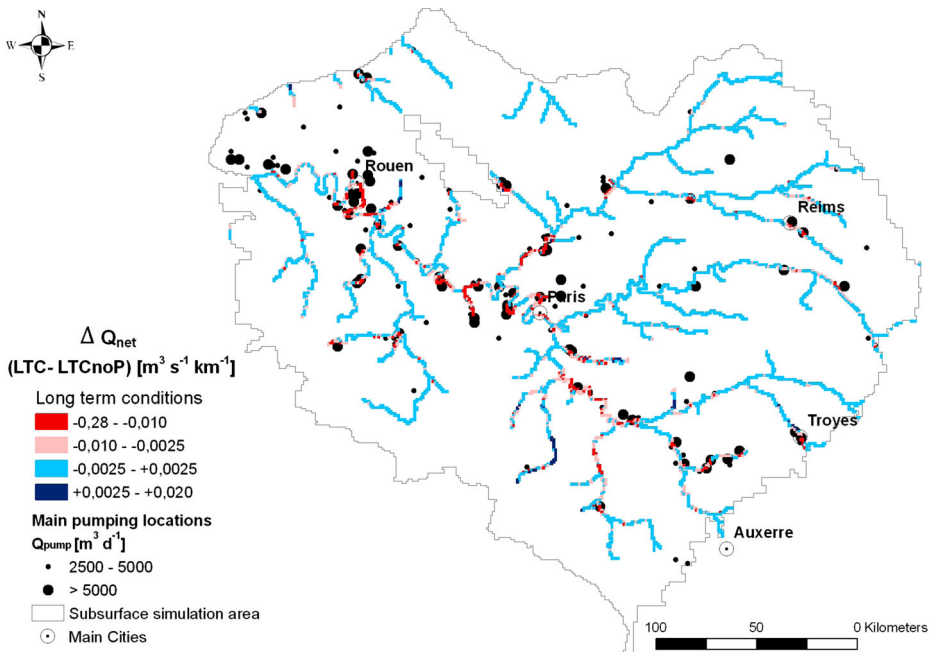


Fig. 7 Difference in net stream-to-aquifer flow (ΔQ_{net}^{LTC} , Eq. 6) between long term simulations LTC and LTCnoP (with and without considering groundwater withdrawals, respectively). Reddish colors indicate that groundwater withdrawals induce a reduction of net stream-to aquifer flow. This is particularly visible in the vicinity of the main pumping locations (black filled circles)

withdrawals induce stream-aquifer flow reversals in 15 % of the simulated river network (i.e. 500 km river length).

5 Discussion

In this study, the *EauDyssée* platform has been extended and implemented in the Seine river basin. It has successfully contributed to the identification of areas where streams are in gaining or losing conditions, to the description of variability at a daily time step and to the quantification of the effects of groundwater withdrawals. The regional model provides quantitative data on the vulnerability of surface water bodies with regard to groundwater bodies and allows prospective simulations to analyze the potential impact of climate change or different pumping scenarios. Though it is essential to water authorities and stakeholders (Tellam and Lerner 2009), this type of approach is not often reported in the literature (Flipo et al. 2014; Krause et al. 2011). We detail hereafter the principal hypothesis at the basis of the model and the main issues, which may impede the implementation of the methods in other contexts.

The implementation of an integrated model at the regional scale is valuable, since it is the scale of interest for water authorities. But to become realistic, regional modeling requires geometrical and conceptual simplifications. In particular, the continuity of charge at the stream-aquifer interface is hardly applicable in regional models with relatively coarse computational grids (Ebel et al. 2009; Mehl and Hill 2010; Rushton 2007). This justifies the use of a conductance model to estimate stream-aquifer flow. This model is based on the hypothesis that stream-aquifer flow can be inferred from the head difference between the stream and the aquifer with a stationary, linear relation. In the present study, the conductance model was improved with the use of variable in-stream water levels, which is usually not the case. This improves the estimate of stream-aquifer flow and groundwater level fluctuations in the vicinity of rivers (Saleh et al. 2011). However, stream-aquifer flow may deviate from a linear behavior in heterogeneous environments and when the stream is disconnected from the aquifer (Brunner et al. 2010; Engeler et al. 2011; Kalbus et al. 2009; Fleckenstein et al. 2006). Disconnected streams are uncommon in temperate hydrosystems such as the Seine river basin. But the assumption of an homogeneous geological medium and simplified river geometry may alter the estimate (Brunner et al. 2010; Rushton 2007). In order to improve stream-aquifer flow estimates in areas of specific interest, current regional model could be used in association with local, detailed, fine-grid models. The regional model may then provide sub-model boundary conditions and starting values for calibrated parameters (Mehl and Hill 2002; Vermeulen et al. 2006).

The GIS database built up for this study includes land use and soil type maps for surface water budget computations, a digital elevation model, a georeferenced river network for surface water routing and a geological model for the groundwater flow component. The availability of distributed climatic data (daily precipitation and potential evapotranspiration) is also critical. In addition, the availability of a detailed database of groundwater withdrawals is essential. The implementation of the *Eaudyssée* platform was made possible with the availability of a large and reliable dataset. Without such a dataset, regional hydrosystem modeling may be inapplicable. To this effect, the collection and archiving of GIS databases at the basin scale should be promoted by water authorities.

Distributed process-based model present the critical advantage of providing results distributed in space, which is essential for water resources management. However, this

implies the use of numerous distributed parameters that must be calibrated. This operation requires long-term records of river discharge and groundwater levels, collected with a unified and reliable observation network covering the whole basin. Such a dataset is available in the Seine river basin, which made the calibration of model parameters possible. To this effect, the monitoring of river discharge rates and groundwater levels should also be promoted.

In this study, trial and error manual calibration was facilitated with the implementation of an efficient post-processing procedure. However, the calibration has required a substantial amount of time. Calibration with optimization techniques require simulation to be sufficiently fast to afford numerous runs in a reasonable amount of time. To date, a run of the Seine river model over the 10-year calibration period lasts about 90 min, which makes automatic calibration hardly applicable. Simulation time could be reduced with optimized algorithms and parallelized-code, which could make optimization techniques applicable to the regional models in a near future.

It is of the highest importance for water authorities to have indicators related to their administrative units, such as the EU “water bodies” (Tellam and Lerner 2009; Krause et al. 2011). However, the representation of stream-aquifer flow aggregated over these units may conceal the actual variability of stream-aquifer flow (Fig. 6). This study highlights that water authorities should be informed of within-unit variability and potentially consider the definition of sub-units to handle specific issues revealed from the detailed, distributed results.

6 Conclusions

The *Eaudyssée* platform, an integrated surface-subsurface model was extended to quantify stream-aquifer flow in space and time at the regional scale. The model successfully simulated the Seine river basin (76 375 km²), where a majority (82 %) of streams are in gaining conditions. The results reveal that groundwater withdrawals induce a decrease of net stream-to-aquifer flow over the Seine basin (−19 %), which corresponds to the third of the average total pumping rate. The model results also highlight that pumping are at the origin of stream-aquifer flow reversals, which may cause water quality issues. This data provided at the basin scale is essential for the local water authority, the *Agence de l'Eau Seine Normandie*.

The implementation of such a regional model is conditioned by the availability of a large and reliable dataset constituted by model geometry, climatic records, and the availability of long-term observed records for the calibration procedure. To this effect, long-term monitoring of surface and subsurface water levels and the constitution of reliable basin-wide unified databases should be promoted by water authorities. As society moves toward more integrated management, this regional modeling platform aims to be coupled with groundwater and in-stream transport capabilities. This would provide water authorities a powerful tool to address the challenges related to the quantitative and qualitative issues affecting water resources.

Acknowledgements This project was conducted on the request of the *Agence de l'Eau Seine Normandie* which participated substantially to the project funding. Funding was also supported by the CNES TOSCA SWOT project and the workpackage “Stream-Aquifer Interfaces” of the PIREN Seine research program. We kindly thank the BRGM for providing the DEM and aquifer geometries.

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