Continental hydrosystem modelling: the concept of nested stream–aquifer interfaces

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Abstract

Recent developments in hydrological modelling are based on a view of the interface being a single continuum through which water flows. These coupled hydrological-hydrogeological models, emphasising the importance of the stream–aquifer interface, are more and more used in hydrological sciences for pluri-disciplinary studies aiming at investigating environmental issues. This notion of a single continuum, which is accepted by the hydrological modellers, originates in the historical modelling of hydrosystems based on the hypothesis of a homogeneous media that led to the Darcy law. There is then a need to first bridge the gap between hydrological and eco-hydrological views of the stream–aquifer interfaces, and, secondly, to rationalise the modelling of stream–aquifer interface within a consistent framework that fully takes into account the multi-dimensionality of the stream–aquifer interfaces. We first define the concept of nested stream–aquifer interfaces as a key transitional component of continental hydrosystem. Based on a literature review, we then demonstrate the usefulness of the concept for the multi-dimensional study of the stream–aquifer interface, with a special emphasis on the stream network, which is identified as the key component for scaling hydrological processes occurring at the interface. Finally we focus on the stream–aquifer interface modelling at different scales, with up-to-date methodologies and give some guidances for the multi-dimensional modelling of the interface using the innovative methodology MIM (Measurements-Interpolation-Modelling), which is graphically developed, scaling in space the three pools of methods needed to fully understand stream–aquifer interfaces at various scales. The outcome of MIM is the localisation in space of the stream–aquifer interface types that can be studied by a given approach. The efficiency of the method is demonstrated with two approaches from the local (∼1 m) to the continental (<10 M km²) scale.
1 Introduction

The emergence of a systemic view of the hydrological cycle led to the concept of continental hydrosystem (Dooge, 1968; Kurtulus et al., 2011), which “is composed of storage components where water flows slowly (e.g. aquifers) and conductive components, where large quantities of water flow relatively quickly (e.g. surface water)” (Flipo et al., 2012, p. 1). This concept merges surface and ground waters into the same hydrological system through the stream–aquifer interface. As a key transitional component characterised by a high spatio-temporal variability in terms of physical and biogeochemical processes (Brunke and Gonser, 1997; Krause et al., 2009b), this interface requires further consideration for characterising the hydrogeological behaviour of basins (Hayashi and Rosenberry, 2002), and therefore continental hydrosystem functioning (Saleh et al., 2011).

The dynamics of water exchanges at the stream–aquifer interface is complex and mainly depends on geomorphological, hydrogeological, and climatological factors (Sophocleous, 2002; Winter, 1998). Recent eco-hydrological publications, dedicated to stream–aquifer interfaces claim the recognition of the multi-dimensionality and the complexity of the processes taking place in the interface (Ellis et al., 2007; Hancock et al., 2005; Poole et al., 2008; Stonedahl et al., 2012). Also modern landscape typologies, emerging from eco-hydrological concepts based on functionalities of morphological units, highlight the multi-dimensionality of the stream–aquifer interfaces (Bertrand et al., 2012; Dahl et al., 2007). Behind the multi-dimensionality is the notion of scales, which structures the definition, the behaviour and the functionality of the stream–aquifer interface.

Paradoxically, recent developments in hydrological modelling are based on a view of the interface being a single continuum through which water flows (Jones et al., 2006, 2008; Kollet and Maxwell, 2006; Panday and Huyakorn, 2004; VanderKwaak and Loague, 2001; Werner et al., 2006). On the one hand, this notion of a single continuum, which is accepted by the hydrological modellers, originates in the historical
modelling of hydrosystems based on the hypothesis of an homogeneous media that led to the Darcy law. On the other hand, coupled hydrological-hydrogeological models, emphasising the importance of the stream–aquifer interface, are more and more used in hydrological sciences for pluri-disciplinary studies aiming at investigating environmental issues (Ebel et al., 2009). However, these models do not explicitly consider the multi-dimensionality of stream–aquifer interfaces, as formerly highlighted by the eco-hydrological community. There is then a need to first bridge the gap between hydrological and eco-hydrological views of the stream–aquifer interfaces, and, second, to rationalise the modelling of stream–aquifer interface within a consistent framework that fully accounts for the multi-dimensionality of the stream–aquifer interfaces (Marmonier et al., 2012).

Following the attempt of Mouhri et al. (2013) aiming at rationalising the design of stream–aquifer interfaces sampling system, we first define the concept of nested stream–aquifer interfaces as a key transitional component of continental hydrosystem. Based on a literature review, we then demonstrate the usefulness of the concept for the multi-dimensional study of the stream–aquifer interface, with a special emphasis on the stream network which is identified as the key component for scaling hydrological processes occurring at the interface. Finally the paper focuses on the stream–aquifer interface modelling at various scales, with up-to-date methodologies, and gives some guidance for the multi-dimensional modelling of the interface using the MIM (Measurements-Interpolation-Modelling) methodology, which is illustrated with two examples. The first one analyses stream–aquifer interface processes from the local (∼ 1 m) to the watershed (∼ 1000 km²) scale. The second one evaluates the potential of the future space borne SWOT mission for further understanding of stream–aquifer interfaces at the regional and continental scales, which are the scales of interest for stakeholders and practitioners.
2 The concept of nested stream–aquifer interfaces

Many hydrosystem models have been developed, and especially coupled surface–subsurface hydro(geo)logical models (Loague and VanderKwaak, 2004), with no special emphasis on stream–aquifer interfaces. Based on 171 references reviewed by Flipo (2013), Table 1 synthesises practical applications of the most used Distributed Physically-Based Models (DPBMs), with a special emphasis on their spatio-temporal sizes.

During the 1970’s and 1980’s, the first sedimentary bassin’ DPBMs were developed based on the finite differences numerical scheme (Abbott et al., 1986; Freeze, 1971; Harbaugh et al., 2000; Ledoux et al., 1989; de Marsily et al., 1978; McDonald and Harbaugh, 1988; Parkin et al., 1996; Refsgaard and Knudsen, 1996). In this type of approach, the hydrosystem is divided into compartments, which exchange through interfaces.

Since the late 1990’s, new models based on finite elements numerical schemes have been developed (Bixio et al., 2002; Goderniaux et al., 2009; Kolditz et al., 2008; Kollet and Maxwell, 2006; Li et al., 2008; Panday and Huyakorn, 2004; Therrien et al., 2010; VanderKwaak and Loague, 2001; Weill et al., 2009). These models allow the simulation of the pressure head in 3-D instead of the former pseudo 3-D modelling of the piezometric head. However, it is not yet possible to straightforwardly simulate large hydro systems (> 10 000 km$^2$) with a high spatio-temporal resolution for long periods of time (a few decades) (Flipo et al., 2012). This is due to the large number of elements required to simulate such hydro systems (Gunduz and Aral, 2005), which imposes the usage of heavily parallelised codes for simulating these systems with such a spatio-temporal resolution. Only a proof of concept has recently been published by Kollet et al. (2010), who have simulated a 1000 km$^2$ basin with a high spatio-temporal resolution.

Contrarily to the atmosphere–groundwater interface (mostly the soil and the vadose zone), which was intensively studied through experimental (even with satellites facilities) and modelling approaches up to a project of a 1 km $\times$ 1 km distributed modelling
of the earth hydrological cycle (Beven and Cloke, 2012; Wood et al., 2011, 2012), the stream–aquifer interfaces have only been intensively surveyed for broadly two decades (Fleckenstein et al., 2006; Marmonier et al., 2012). Its study by the eco-hydrological community led to a re-conceptualisation of its nature from the river being seen as an impervious drain that collects the effective rainfall and transfers it to the ocean, toward a more subtle view that integrates more spatio-temporal processes in the hydrosystem functioning. Indeed, the stream–aquifer interface is now conceptualised as a filter through which water flows many times over various spatial (from centimetres to kilometres) and temporal scales (from seconds to months) before to reach the sea (Datry et al., 2008). One of the main challenges is to understand the role of the stream–aquifer interfaces in the hydro(geo)logical functioning of basins (Hayashi and Rosenberry, 2002). The multi-dimensionality of the problem at hand imposes to define the scales of interest.

The five commonly recognised scales (scale is used here for the size of the studied objects) are the local, the reach, the catchment, the regional, and the continental ones (Blöschl and Sivapalan, 1995; Dahl et al., 2007; Gleeson and Paszkowski, 2013), being defined as:

- **local scale** (or the experimental site scale) [10 cm–\(\sim\) 10 m]: this scale concerns the riverbed or the hyporheic zone (HZ, see Sect. 3.2 for more details);

- **intermediate or reach scale** [100 m–\(\sim\) 10 km]: it concerns the river reach, a pound or a small lake;

- **catchment–Watershed scale** [10 km\(^2\)–\(\sim\) 1000 km\(^2\)]: this scale connects the stream network to its surface watershed and more broadly to the hydrosystem. This is the scale from which surface-ground water exchanges are linked with the hydrological cycle and the hydrogeological processes;

- **regional scale** [10 000 km\(^2\)–\(\sim\) 1 M km\(^2\)]: this is the scale of water resources management, and the one for which the least is known about stream–aquifer exchange dynamics. For a conceptual analysis of the stream–aquifer interfaces,
the watershed and the regional scales can be merged into a single category referred to as the regional scale (Mouhri et al., 2013). Merging these two scales is consistent with the fact that a regional basin is a collection of smaller watersheds. The distinction between the two categories is only necessary to conceptualise the scaling of processes as discussed in the final section of this paper;

- continental scale (> 10 M km²): this scale is a collection of regional scale basins. The difference with the regional scale is that there is a broader range of hydro-climatic conditions, which imposes to take into account climatic circulations.

From a conceptual point of view, stream–aquifer exchanges are driven by two main factors: the hydraulic gradient and the geological structure. The hydraulic gradient defines the water pathways (Winter, 1998), whereas the geological structure defines the conductive properties of the stream–aquifer interface (White, 1993; Dahm et al., 2003). These two factors are fundamental for hydrogeologists, who derive from those subsurface flow velocities and transfer times. The time scale to be considered also varies depending on the studied object (HZ itself or a sedimentary basin functioning) (Harvey, 2002). Estimating the stream–aquifer exchanges at a sedimentary basin scale then requires the combination of various processes with different characteristic times or periods covering a wide range of temporal orders of magnitude (Blöschl and Sivapalan, 1995; Flipo et al., 2012; Massei et al., 2010): hour-day for river flow, year-decade for effective rainfall, decade-century for subsurface transit time. To address this, models are used as spatio-temporal interpolators. The final choice of model, which can be either conceptual, statistical, process-based or hybrid, is a trade off between a number of factors, such as the required accuracy, type and availability of data, available computational facilities, temporal and spatial scale. The rationale for selecting a particular stream–aquifer modelling technique is a function of the application’s objective and of the model’s suitability for modelling key aspects of the problem at hand (Saleh et al., 2011).
Mouhri et al. (2013) proposed a multi-scale framework to study stream–aquifer interfaces. Their approach is based on the observation that the two main hydrosystem components are the surface and groundwater components, which are connected by nested interfaces (Fig. 1). Stream–aquifer interfaces consist in alluvial plain at the regional and watershed scales (Fig. 1a and b), while within the alluvial plain, they consist in riparian zone at the reach scale (Fig. 1d). Within the riparian zone, they consist in the hyporheic zone at the local scale (Fig. 1c), and so on until the water column–benthos interface within the river itself (Fig. 1f). Before further developing the multi-scale framework, the various descriptions of stream–aquifer interfaces are outlined.

3 Multi-dimensionality of the stream–aquifer interface

A literature review of process-based modelling of stream–aquifer interfaces’ functioning is presented in Table 2, which synthesises 42 references. The majority of them focuses on the local scale (21), while only four consider the regional and continental scales. The remaining mostly focuses on the local-intermediate (9) and intermediate scales (7).

3.1 A multi-scale issue structured around the intermediate scale – the river

The river network is identified as being the location where flow paths mix at all scales, and therefore the location of hydrological process scaling.

Near river groundwater flow paths are mainly controlled by regional flow paths in aquifer systems (Malard et al., 2002). Indeed, the groundwater component of hydrosystem controls the regional flows towards the alluvial plains and the rivers. Such flow paths define the total amount of water that flows in the stream–aquifer interface (Cardenas and Wilson, 2007b; Frei et al., 2009; Kalbus et al., 2009; Rushton, 2007; Storey et al., 2003). This is not a new concept as the river network corresponds to drains collecting regional groundwater (Fig. 1a), which sustain the network during
low flow period (Ellis et al., 2007; Pinder and Jones, 1969; Tóth, 1963). These large scale structural heterogeneities can also generate local conditions that favour local re-infiltration of river water towards the aquifer system (Boano et al., 2010; Cardenas, 2009a, b; Fleckenstein et al., 2006). These re-infiltrations (Fig. 1b and c) can even constitute the main recharge of some peculiar local aquifer systems, as for instance some alluvial plain (Krause and Bronstert, 2007; Krause et al., 2007).

In second instance, the spatial distribution of the stream bed permeabilities controls the dynamics of stream–aquifer exchanges within the alluvial plain, and therefore the near-river piezometric head distribution (Calver, 2001; Fleckenstein et al., 2006; Frei et al., 2009; Genereux et al., 2008; Hester and Doyle, 2008; Kalbus et al., 2009; Käser et al., 2009; Rosenberry and Pitlick, 2009). Finally the longitudinal morphology of the river and the topography of the river bed, consisting in a pluri-metric succession of pools and riffles (Fig. 1e), also impact the stream–aquifer exchanges (Koch et al., 2011; Marzadri et al., 2010; Whiting and Pomeranets, 1997) and the river hydraulic regime (Cardenas and Wilson, 2007a; Munz et al., 2011; Saenger et al., 2005). Ultimately a very fine scale process (~ cm–dm), due to the in-stream non hydrostatic flow induced by bedform micro-topography (Fig. 1f), also increases the absolute value of the total stream–aquifer exchanges (Cardenas and Wilson, 2007a; Cardenas and Wilson, 2007b; Endreny et al., 2011; Janssen et al., 2012; Käser et al., 2013; Krause et al., 2012b; Sawyer and Cardenas, 2009; Stonedahl et al., 2010).

It is therefore important to study the stream–aquifer exchanges in the dual perspective of regional and local exchanges; the former being controlled by regional recharge and structural heterogeneities, the later by the longitudinal distribution of stream bed heterogeneities and the river morphology (Schmidt et al., 2006). These two types of
controlling factors may also generate water loops within the stream–aquifer interfaces, the river corridor being the location where these processes merge.

Ellis et al. (2007) confirmed this statement with the investigation of the spatio-temporal relevance of both data sampling density and models from the intermediate scale to the local one. They concluded that stream–aquifer exchange distributions are submitted to multi-scale controls, which influence the thickness of the HZ and the patterns of groundwater flow through the riverbed.

### 3.2 The stream–aquifer interface at the local scale – the hyporheic zone

At the local scale (plot, river cross section), the stream–aquifer interface consists in a hyporheic zone (HZ), which corresponds to an ecotone, whose extent varies dynamically in space and time. This ecotone is at the interface between two more uniform, yet contrasted ecological systems (Brunke and Gonser, 1997): the river and the aquifer. In a broad sense, the HZ is “the saturated transition zone between surface water and groundwater bodies that derives its specific physical (e.g. water temperature) and biogeochemical (e.g. steep chemical gradients) characteristics from active mixing of surface and groundwater to provide a habitat and refugia for obligate and facultative species” (Krause et al., 2009a, p. 2103). White (1993) also indicates that the HZ is located beneath the stream bed and in the stream banks that contain infiltrated stream water. Furthermore, Malard et al. (2002) identified five generic HZ configurations, that depend on the structure of the subsurface media, and especially on the location of the impervious substratum:

1. No HZ: the stream flows directly on the impervious substratum. A perennial lateral HZ can appear in the zone of significant longitudinal curvature of the stream, for instance in the case of meanders (Boano et al., 2009; Cardenas, 2009a; Revelli et al., 2008).
2. No aquifer unit: a HZ can appear due to the infiltration of the stream water towards the substratum or through the stream banks. In the former case, the substratum is located near to the stream bed sediments.

3. Existence of a HZ in a connected stream–aquifer system: the HZ is created by advective water from both the stream and the aquifer unit. The impervious substratum is located beneath the aquifer unit.

4. Existence of a HZ in a disconnected stream–aquifer system: a distinct porous media lies in between the stream bed and the aquifer unit. This porous media would not be saturated if the stream bed were impervious. In this configuration, two subcategories are to be found:
   
   a. the vertical infiltration of stream water towards the top of the aquifer unit generates a zone of mixing waters at the top of the aquifer unit, far enough below the stream bed to be disconnected from it,
   
   b. a perched HZ is formed below the stream bed due to the infiltration of stream water. In this particular case, the porous media below the stream bed is either very thick or its conductive properties are so poor that the surface water may not reach the aquifer unit.

The extent of the HZ, which depends on the hydrological settings, varies from centimetres to hundreds of meters (Brunke and Gonser, 1997; Woessner, 2000; Wroblicky et al., 1998). Even in a specific configuration, the extent and the nature of the stream–aquifer interface vary through time, depending on the hydro(geo)logical context. For instance, Conant (2004) and Storey et al. (2003) reported that the HZ is affected by the regional flow of the aquifer system, whereas Wroblicky et al. (1998) indicate that the variation of the head difference between aquifer and stream modifies the extent of the HZ.
3.3 The stream–aquifer interfaces at the regional and continental scales – the alluvial plains

The interaction between surface and subsurface waters has also been identified at the basin scale, where geological heterogeneities control the stream–aquifer exchanges, which in return can control the near river piezometric head distribution in the case of an alluvial aquifer (Boano et al., 2010; Cardenas, 2009a, b; Fleckenstein et al., 2006).

Although the usage of DPBM covers a broad range of spatial scales, only 19 publications among 183 (Tables 1 and 3) concern large river basins (> 10,000 km²) (Abu-El-Sha’s and Rihani, 2007; Andersen et al., 2001; Bauer et al., 2006; Boukerma, 1987; Christiaens et al., 1995; Etchevers et al., 2001; Golaz-Cavazzi et al., 2001; Gomez et al., 2003; Habets et al., 1999; Hanson et al., 2010; Henriksen et al., 2008; Kolditz et al., 2012; Ledoux et al., 2007; Lemieux and Sudicky, 2010; Monteil, 2011; Park et al., 2009; Saleh et al., 2011; Scibek et al., 2007) and except Monteil (2011), Pryet et al. (2013) and Saleh et al. (2011), none of them focuses on stream–aquifer exchanged water flux. Moreover, among DPBMs dedicated to stream–aquifer exchanges, no application was carried out at the regional scale (Table 3).

At this scale, most of the hydro(geo)logical models are limited to take into account local processes as the effect of near river pumping, or storage in the hyporheic zone, because they require a very fine spatial discretisation, which can be incompatible with the resolution of the model or, at most, drastically decreases the efficiency and precision of the model. Moreover, the usage of regional models to solve local issues, as well as the reverse, leads to equifinality problems (Beven, 1989; Beven et al., 2011; Ebel and Loague, 2006; Klemes, 1983; Polus et al., 2011), boundary conditions inconsistencies (Noto et al., 2008), or computational burdens (Jolly and Rassam, 2009). The usage of local models to solve regional issues also leads to the same effects (Aral and Gunduz, 2003, 2006; Wondzell et al., 2009). Moreover, neither a too simple model, nor a too complex one can provide relevant answers (Hill, 2006; Smith et al., 2004; Wondzell et al., 2009). Therefore alternative ways of modelling are needed to properly simulate...
the behaviour of stream–aquifer interfaces at the regional scale (Werner et al., 2006), especially that for a given reach of river the direction of stream–aquifer exchanges can vary longitudinally (Bencala et al., 2011).

According to Krause et al. (2011) the knowledge on processes occurring in the stream–aquifer interface and the need for knowledge by water resources managers is first inversely correlated, and second not much is known about the role of stream–aquifer interfaces at the regional scale, which is the scale of interest for practitioners. There is therefore a crucial need to develop innovative methodologies for assessing stream–aquifer exchanges at the regional scale.

4 Modelling stream–aquifer exchanges

4.1 Models to simulate stream–aquifer interface

Surface water groundwater exchanges, mostly through the soil or the stream–aquifer interface, are simulated with two different models (Ebel et al., 2009; Kollet and Maxwell, 2006; LaBolle et al., 2003; Furman, 2008), whatever the number of simulated spatial dimensions (Tables 2 and 3):

- A conductance model or first order exchange coefficient, for which the interface is described with a water conductivity value. The exchanged water flux is then calculated as the product of the conductivity by the difference of piezometric heads between the aquifer and the surface water body. Depending on the model, the difference of pressures can also be used. This model implicitly formulates the hypothesis of a vertical water flux between surface water and groundwater whatever the mesh size. This is the most common model for simulating stream–aquifer exchanges. There are diverse conductance’s formulations, especially in the case of disconnected aquifers and streams (Osman and Bruen, 2002). Irvine et al. (2012) advocate for the usage of the conductance model if the stream bed heterogeneities are well described, which is usually critical (Genereux et al., 2008).
However the conductance coefficient depends on the temperature because it implicitly integrates the fluid viscosity (Engeler et al., 2011). Moreover, the validity of the first order law is critical in case of a flood when water expends in the flood plain (Engeler et al., 2011).

– Continuity of pressures and fluxes at the interface. This boundary condition requires an iterative or a sequential computation, although the iterative one is more precise (Sulis et al., 2010). Sometimes the iterative process also leads to a discontinuity of the tangential component of the water velocity at the interface with the stream bed (Discacciati et al., 2002; Miglio et al., 2003; Urquiza et al., 2008). This is not a problem as this discontinuity can be interpreted as representative of the stream bed load.

Recent numerical developments allow for solving the coupled surface and subsurface equations at once with a matricial system. This method is called coupled in Tables 2 and 3, and can be used with whatever selected stream–aquifer interface model. One of the main drawbacks of this method is that it is computationally demanding and usually requires a parallelised model in order to simulate real hydrosystem.

From a conceptual point of view, the conductance model permits to better understand the hydrological processes occurring at the stream–aquifer interface (Delfs et al., 2012; Ebel et al., 2009; Liggett et al., 2012; Nemeth and Solo-Gabriele, 2003) and is equivalent to the continuity one in the case of a highly conductive interface.

### 4.2 Temperature as a tracer of the flow – the local scale

The study of heat propagation is a powerful tool for assessing stream–aquifer exchanges (Anderson, 2005; Constantz, 2008; Mouhri et al., 2013) based on the temperature used as a tracer of the flow. Coupled with in situ measurements, two methods, based on heat transport governing equations, are used to quantify stream–aquifer exchanges (Anderson, 2005):
1. Analytical models (Stallman, 1965; Anderson, 2005) are widely used to inverse temperature measurements solving the 1-D heat transport equation analytically under simplifying assumptions (sinusoidal or steady boundary conditions and homogeneity of hydraulic and thermal properties) (Anibas et al., 2009; Anibas et al., 2012; Becker et al., 2004; Hatch et al., 2006; Jensen and Engesgaard, 2011; Keery et al., 2007; Lautz et al., 2010; Luce et al., 2013; Rau et al., 2010; Schmidt et al., 2007; Swanson and Cardenas, 2011).

2. Numerical models which couple water flow equation in porous media with the heat transport equation in 2-D or 3-D. These models are divided in two categories based on the numerical scheme: finite differences (Anderson et al., 2011; Anibas et al., 2009; Constantz et al., 2002, 2013; Constantz, 2008; Ebrahim et al., 2013; Lewandowski et al., 2011; Mutiti and Levy, 2010; Rühaak et al., 2008; Schornberg et al., 2010) or finite elements (Kalbus et al., 2009; Mouhri et al., 2013). These models have the advantage of calculating spatio-temporal stream–aquifer exchanges with the capability of accounting for the heterogeneities under transient hydrodynamical and thermal conditions.

Eventually the two approaches provide estimates of the conductance coefficient that best represents the stream–aquifer interface at the local scale.

### 4.3 The conductance model at the regional scale

To the authors’ knowledge, very few DPBMs have been applied to assess stream–aquifer exchanges at the regional scale (> 10 000 km²) (see Sect. 3.3). These applications exclusively use the conductance model, for which the longitudinal distribution of the conductance along the stream network has to be calibrated (Pryet et al., 2013). To provide accurate estimates, the conductance model has to be constrained by the piezometric head below the river and the surface water elevation. Former applications used a fixed water level throughout the simulation period (Flipo et al., 2007; Gomez et al., 2003; Monteil, 2011; Thierion et al., 2012). Saleh et al. (2011) showed that this
methodology not only leads to biased assessments of stream–aquifer exchanges, but also to biased estimates of the near river piezometric head distributions.

The simulation of surface water levels is therefore of primary importance for the estimation of distributed stream–aquifer exchanges along the stream network at regional scale (Pryet et al., 2013; Saleh et al., 2011). Saleh et al. (2013) recommend the usage of local 1-D Saint-Venant based hydraulic models to build rating curves for every cell of a coarser regional model (Saleh et al., 2011) that uses simpler in-stream water routing models as RAPID (David et al., 2011). Such models are then coupled with the conductance model to simulate stream–aquifer exchanges at the regional scale along thousands of kilometres of river networks with a 1 km spatial discretisation (see for instance Pryet et al., 2013 for such an application along 4000 km of the Paris basin river network).

4.4 Conceptual needs at the continental scale

Russell and Miller (1990) achieved the first runoff calculation based on a $4^\circ \times 5^\circ$ grid mesh coupled with a Land Surface Model (LSM) and an Atmospheric Global Circulation Model (AGCM). It appears that even at this scale the river networks play an important role in the circulation models and water transfer time. Since then, few models have been developed to simulate the main river basins in the AGCMs with a grid mesh of $\sim 1^\circ \times 1^\circ$, which roughly corresponds to a 100km $\times$ 100km resolution (Oki and Sud, 1998). Geographical Information Systems (GISs) were used to derive the river networks from Digital Elevation Models (Oki and Sud, 1998). Jointly RRMs (River Routing Models) have been developed with simple transfer approaches, assuming either a steady uniform water velocity at the global scale or a variable water velocity based on simple geomorphological laws and the Manning Formula (Arora and Boer, 1999).

Decharme and Douville (2007) implemented the approach with a constant in-river water velocity (assumed to be 0.5 m s$^{-1}$) within the LSM, today referred to as SURFEX (Masson et al., 2013). Step by step the description of stream–aquifer exchanges was improved with:
– The introduction of a variable in-river water velocity (Decharme et al., 2008).
– A transfer time delay due to the stream–aquifer interface (Decharme et al., 2012).
– The explicit simulation with a DPBM of the worldwide largest aquifer systems coupled with the explicit simulation the river networks draining surface basins larger than 50 000–100 000 km$^2$ (Vergnes and Decharme, 2012).
– The explicit simulation of stream–aquifer exchanges based on the conductance model on a 0.5° × 0.5° grid mesh (Vergnes et al., 2012; Vergnes and Decharme, 2012) in agreement with the continental scale transfer time delay of 30 days introduced by Decharme et al. (2012).

As expected given the numerical experiments of Maxwell and Miller (2005), accounting for groundwater kinetics improves the global hydrological mass balance (Decharme et al., 2010; Alkama et al., 2010; Yeh and Eltahir, 2005). Although the explicit simulation of stream–aquifer exchanges with the conductance model slightly improves the models’ performances in terms of spatio-temporal discharge and real evapotranspiration assessments (Vergnes et al., 2012; Vergnes and Decharme, 2012), the global calibration of the conductance parameter has to take into account the multi-scale structure of the stream–aquifer interfaces, which means that a better assessment, not only of simple DEM derived river networks, but also of the transfer time in the stream–aquifer interfaces is required as well as the subgrid definition of dendritic river networks. Coupled with proper scaling procedures (see next section) these approaches seem to be less computationally demanding than the one proposed by Wood et al. (2011) and slightly less over parametrised, which should permit to better resolve the estimation of stream–aquifer exchanges at the continental scale.

### 4.5 Up and downscaling stream–aquifer exchanges

The conductance model historically assumes vertical fluxes at the stream–aquifer interface (Krause et al., 2012a; Sophocleous, 2002), so that it seems to be a proper
framework for determining up and down scaling properties of stream–aquifer interfaces (Boano et al., 2009; Engdahl et al., 2010). However this hypothesis becomes less valid at regional scale when the grid mesh is getting coarser (Mehl and Hill, 2010; Rushton, 2007). Indeed, in heterogeneous media modelling, the transmissivity field and the associated piezometric heads are highly mesh size dependent (Renard, 1997). Furthermore, in many models the calculated piezometric head corresponds to a volumetric average on the grid cells (Bear, 1972; de Marsily, 1986; Ledoux et al., 1989). In the case of a coarse grid, the calculated piezometric head at the stream–aquifer interface does not represent the piezometric head beneath the river itself (Fig. 2). The averaging process can then induce uncertainties in the assessment of the conductance parameter, which becomes scale dependent (Vermeulen et al., 2006).

The hypothesis of vertical fluxes is discussed by Rushton (2007) based on numerical experiments that showed its limit. Indeed, at the regional scale, stream–aquifer exchanges seem to be more controlled by the horizontal permeability of the aquifer unit than by the equivalent vertical permeabilities of both the river bed and the aquifer unit. Recently, this new formulation of the drivers of stream–aquifer exchanges proved to be suitable for the calibration of a regional modelling of stream–aquifer exchanges (Pryet et al., 2013).

As formulated by Rushton (2007), Pryet et al. (2013) calibrated a correction factor. To properly scale the conductance model, the correction factor should be defined analytically by linking the conductance to the vertical permeabilities of the stream bed and the aquifer unit (through the anisotropy of the near stream aquifer unit) (Morel-Seytoux, 2009). Coupled to the scaling of the conductance, proper distribution of piezometric heads has to be estimated (Vermeulen et al., 2006). A potential methodology could consist in a near stream interpolation of the regional piezometric head, which should consider the local variability of the transmissivity field (Chen and Durlofsky, 2006), verify the integrity of the regional flux (Mehl and Hill, 2002) and take into account the geometrical change of boundary conditions (Panday and Langevin, 2012).
Coupling these up and downscaling procedures of both parameters and state variables is critical for the explicit formulation of the nested stream–aquifer interface concept in a modelling framework structured around the river network, where the computational power needs to be concentrated.

5 The MIM methodology: from concepts to practice

The methodology of Mouhri et al. (2013) is hereby graphically developed, scaling in space the three pools of methods (measurements-interpolation-modelling) needed to fully understand stream–aquifer interfaces at various scales. The outcome is the MIM (Measurement-Interpolation-Modelling) methodological tool, which localises in space the type of stream–aquifer interface that can be studied by a given approach (see the five scales of interest in Fig. 3: local, reach, watershed, regional and continental scales).

5.1 Coupled in situ-modelling approaches: from local to watershed scale

Figure 4 displays the types of stream–aquifer interfaces that can be studied by the multi-scale sampling system developed by Mouhri et al. (2013), based on LOcal MOnitoring Stations (LOMOS) distributed along a 6 km river network covering a 40 km² watershed. As illustrated in Fig. 4, a single LOMOS allows the monitoring, based on water pressure and temperature measurements, of stream cross-sections ranging from 0.1–
\(\sim\) 10 m. LOMOS data are used with coupled thermo-hydro models to determine the properties of the aquifer units and the river beds (Mouhri et al., 2013), which can be used to assess the value of the conductance at the watershed scale (Mehl and Hill, 2002; Morel-Seytoux, 2009; Vermeulen et al., 2006; Rushton, 2007). Assuming that it is possible to distribute multiple LOMOS data, and the associated conductance values, along a stream network (for instance using FO-DTS – Fiber Optic Distributed Thermal Sensors), local in situ data become the basis of a broader surface–subsurface
modelling at the watershed scale (Table 1). It thus appears in the MIM space that the upscaling is structured around stream cross-sections of \( \sim 1-10 \text{m} \) (Fig. 4).

### 5.2 Space borne approaches: regional and continental scales

At the regional and continental scales, stream–aquifer interfaces can be observed, at least partially, using satellite measurements (Fig. 4). Current satellite platforms do not allow for accurate observation of stream–aquifer exchanges, but they should be able to provide valuable information in the near future (Alsdorf et al., 2007). Indeed, total water storage (e.g. surface waters and ground waters) variations can be estimated from the Gravity Recovery and Climate Experiment (GRACE) mission, launched in 2002 (Tapley et al., 2004). Ramillien et al. (2008) present an extensive review of large-scale hydrological use of the first years of GRACE data. However, these data have low spatial (300–400 km) and temporal resolution (from 10 days to 1 month) (Ramillien et al., 2012), limiting their use for continental scales. Moreover, these data have to be coupled with ancillary information to distinguish between surface waters and ground waters variations. For instance, surface water variations can be estimated by combining multi sensors measurements. Optical or Radar images are used to compute water extent (Cretaux et al., 2011) and can be combined with Digital Elevation Model (DEM) or with water elevation measurements from NADIR altimeters (Calmant et al., 2008) to derive storage changes and fluxes (Neal et al., 2009; Gao et al., 2012). The mismatch between acquisition time, repeatability, and spatial coverage of such data implies that it is difficult to use them for the assessment of stream–aquifer exchanges at the continental scale.

To overcome these issues a new space-borne mission, the Surface Water and Ocean Topography (SWOT) mission, is currently being developed by NASA, CNES (French Spatial Agency) and CSA (Canadian Space Agency), for a planned launch around 2020. SWOT will provide maps of distributed water elevations, water extents and water slopes on two swaths of 50 km coverage each. It will enable the observation of rivers wider than 100 m and surface areas larger than \( 250 \times 250 \text{m} \) (Rodríguez, 2012).
Accuracies on water elevation and water slope will be around 10 cm and 1 cm km$^{-1}$, respectively, after averaging over 1 km$^2$ water area (Rodríguez, 2012). From these requirements, Biancamaria et al. (2010) estimated that SWOT should be able to provide useful information to compute discharge for river reaches with drainage areas above 70 000 km$^2$. This preliminary assessment was recently refined by Andreadis et al. (2013), who estimate that rivers with a bank full width of 100 m have drainage area ranging from 1050 to 50 000 km$^2$. Although the database contains errors (reported errors on river width range from 8 to 62 %), it provides the order of magnitude of minimum drainage area that should be sampled by SWOT. Thanks to the two swaths and its ~20 day repeat orbit, the instrument will observe almost all continental surfaces in between 78° S and 78° N, allowing the sampling of all drainage areas above 50 000 km$^2$. Therefore SWOT is a valuable tool to localise in the MIM space (Fig. 4). SWOT data consist of raw cloud data, which appear on the measurement axis, and reach averaged data to reduce uncertainties (see the Interpolation axis in Fig. 4). Both products can be coupled with regional or continental hydrosystem models. To achieve such coupled applications, it will be necessary to use downscaling methods (Aires et al., 2013) and/or assimilate SWOT observations in stream–aquifer interface models like the one used by Pryet et al. (2013), Saleh et al. (2011) and Vergnes and Decharme (2012).

5.3 Further challenges

Albeit being a breakthrough in terms of surface coverage SWOT requirements impose restrictions on observable stream–aquifer interfaces, which can be visualised in the MIM space (Fig. 4). Unfortunately, it appears in the MIM space that SWOT applications do not completely overlap other methodologies as the one proposed to scale processes between the local and the watershed scales. To overcome this issue, an incoming airborne campaign, called AirSWOT, with a main payload similar to the one of SWOT, but with higher spatial resolution (metric), will (i) help to determine whether regular airborne campaigns can provide a valuable tool to connect the watershed scale to the
regional/continental one with the help of multi-scale modelling tools (cf. Sect. 4.5) and (ii) permit to design new in situ monitoring stations derived from the LOMOS defined by Mouhri et al. (2013) but dedicated to the watershed/regional scale, which means for river cross-sections larger than a few decametres.

6 Conclusions

The systemic view of hydrosystems makes us reformulate the stream–aquifer interface as a cascade of nested objects. These nested objects depend on the scale of interest. At the watershed, regional and continental scales, they consist in alluvial plains, while within the alluvial plan itself (intermediate-reach scale), they consist in riparian zones. Within the riparian zone (local scale), they consist in HZ, and so on until the water column–benthos interface within the river itself.

Estimating stream–aquifer exchanges therefore requires to combine the modelling of various processes with different characteristic times. Depending on the refinement of the modelling at the regional scale (i.e. number of processes taken into account), the estimation of stream–aquifer exchanges may vary significantly. As stakeholders need more detailed information at the regional scale, which is the scale of water resources management, it is crucial to develop modelling tools which can precisely simulate stream–aquifer exchanges at the reach scale within a regional basin. These innovative modelling tools should be multi-scale modelling platforms, which implement the concept of nested stream–aquifer interfaces as the core of the coupling between regional and local models: the former simulating the basin, the latter the alluvial plains. To achieve this, it was shown that processes scaling should be performed around the river network.

To fully estimate stream–aquifer exchanges, this multi-scale modelling tool has to be coupled with observation devices. The MIM methodology provides a powerful framework to jointly develop observation infrastructures and modelling tools, allowing the localisation of the global structure in the scale space. The main result of the first analysis
is that airborne campaigns, as well as regional in situ systems, will have to be ratio-
ualised to connect the watershed to the regional and continental scales, which will be
sampled by the SWOT mission.

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version of the paper.

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Table 1. Coupled surface–subsurface hydrological DPBM. References to be found in Flipo (2013).

<table>
<thead>
<tr>
<th>Model</th>
<th>SW</th>
<th>GW</th>
<th>Cg</th>
<th>Δx</th>
<th>Δt</th>
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<tr>
<td>CATHY</td>
<td>Muskingum-Cunge</td>
<td>RE</td>
<td>P</td>
<td>0.01 ha–</td>
<td>hours–</td>
</tr>
<tr>
<td></td>
<td>1-D FD</td>
<td>3-D FE</td>
<td></td>
<td>690 km²–</td>
<td>decades</td>
</tr>
<tr>
<td>CaWaQS</td>
<td>SV m</td>
<td>DE</td>
<td>K</td>
<td>2500 km²–</td>
<td>decades</td>
</tr>
<tr>
<td></td>
<td>1-D FD</td>
<td>pseudo 3-D FD</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>EauDyséeb</td>
<td>Muskingum</td>
<td>DE</td>
<td>K</td>
<td>1000 km²–</td>
<td>days–</td>
</tr>
<tr>
<td></td>
<td>RC + 1-D FD</td>
<td>pseudo 3-D FD</td>
<td></td>
<td>10 000 km²</td>
<td>century</td>
</tr>
<tr>
<td>HydroGeoSpherea</td>
<td>DW dw</td>
<td>RE</td>
<td>K</td>
<td>1 m²–</td>
<td>hours–</td>
</tr>
<tr>
<td></td>
<td>2-D FE</td>
<td>3-D FE</td>
<td></td>
<td>25 M km²</td>
<td>300 000 yr</td>
</tr>
<tr>
<td>InHM</td>
<td>DWm</td>
<td>RE</td>
<td>K</td>
<td>15 m²–</td>
<td>hours–</td>
</tr>
<tr>
<td></td>
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<td>3-D FE</td>
<td></td>
<td>100 km²</td>
<td>century</td>
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<tr>
<td>MIKE SHE</td>
<td>SV</td>
<td>BE</td>
<td>K</td>
<td>10 km²–</td>
<td>hours–</td>
</tr>
<tr>
<td></td>
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<td>3-D FD</td>
<td></td>
<td>375 000 km²</td>
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<tr>
<td>MODCOU</td>
<td>isochronism</td>
<td>DE</td>
<td>K</td>
<td>100 km²–</td>
<td>days–</td>
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<tr>
<td>MODFLOW</td>
<td>Coupling dependent</td>
<td>pseudo 3-D FD</td>
<td></td>
<td>100 000 km²</td>
<td>century</td>
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<td>RE</td>
<td>K</td>
<td>0.8 km²–</td>
<td>hours–</td>
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<td>K</td>
<td>3 ha–</td>
<td>hours–</td>
</tr>
<tr>
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<tr>
<td>ParFlowa</td>
<td>KWm</td>
<td>RE</td>
<td>P</td>
<td>3 ha–</td>
<td>hours–</td>
</tr>
<tr>
<td></td>
<td>2-D FE</td>
<td>3-D FE</td>
<td></td>
<td>10 000 km²</td>
<td>decades</td>
</tr>
<tr>
<td>SHE</td>
<td>SV</td>
<td>BE</td>
<td>K</td>
<td>5 ha–</td>
<td>days–</td>
</tr>
<tr>
<td>SHETRAN</td>
<td>DW</td>
<td>BE</td>
<td>K</td>
<td>3 ha–</td>
<td>hours–</td>
</tr>
<tr>
<td></td>
<td>1-D FD</td>
<td>3-D FD</td>
<td></td>
<td>16 000 km²</td>
<td>years</td>
</tr>
<tr>
<td></td>
<td>1-D FD</td>
<td>3-D FD</td>
<td></td>
<td>20 000 km²</td>
<td>millenium</td>
</tr>
</tbody>
</table>

SW: Surface water; GW: Groundwater; Cg: Coupling DE: Diffusivity Equation; BE: Boussinesq Equation; RE: Richards Equations SV: Saint-Venant; DW: Diffusive Wave; KW: Kinematic Wave; RC: Rating Curves FD: Finite Differences; FE: Finite Elements; K: Conductance model; P: Pressure continuity,

a Parallelised code,
b Can be parallelised by tasks,
m Friction is calculated with the Manning formulae,
dw Friction is calculated with the Darcy–Weisbach formulae.
Table 2. Physically-based modelling of stream–aquifer exchanges.

<table>
<thead>
<tr>
<th>Ref</th>
<th>exch</th>
<th>Spec</th>
<th>Resolution</th>
<th>Scale</th>
<th>CS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brunner et al. (2009a, b)</td>
<td>K</td>
<td>2-D V LAT</td>
<td>[1–100] m, [≤ 0.05] m</td>
<td>perm</td>
<td>loc-int</td>
</tr>
<tr>
<td>Brunner et al. (2010)</td>
<td>K</td>
<td>2-D V LAT</td>
<td>[1–10] m, [0.1–10] m</td>
<td>perm</td>
<td>loc-int</td>
</tr>
<tr>
<td>Cardenas et al. (2004)</td>
<td>K</td>
<td>3-D</td>
<td>0.25 m–0.25 m, 0.04 m</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Cardenas and Wilson (2007b); Cardenas and Wilson (2007c)</td>
<td>P</td>
<td>2-D V LON</td>
<td>0.01 m–0.01 m^a</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Cardenas (2009a)</td>
<td>P</td>
<td>2-D H</td>
<td>NS (80 m–45 m)</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Discacciati et al. (2002)</td>
<td>P</td>
<td>3-D</td>
<td>[0.5–5] m–[0.5–5] m, [0.3–1.5] m^a</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Ebel et al. (2009)</td>
<td>K</td>
<td>3-D</td>
<td>[1–20] m–[1–20] m, [0.05–0.25] m</td>
<td>adapt</td>
<td>loc-int</td>
</tr>
<tr>
<td>Engeler et al. (2011)</td>
<td>K</td>
<td>3-D</td>
<td>[1–50] m–[1–50] m, [1.6–40] m</td>
<td>900 s</td>
<td>int</td>
</tr>
<tr>
<td>Fleckenstein et al. (2006)</td>
<td>K</td>
<td>3-D</td>
<td>200 m–100 m, [5–40] m</td>
<td>3 h</td>
<td>int</td>
</tr>
<tr>
<td>Frei et al. (2009)</td>
<td>P</td>
<td>3-D</td>
<td>20 m × 50 m × 0.5 m</td>
<td>min</td>
<td>int</td>
</tr>
<tr>
<td>Frei et al. (2010)</td>
<td>K</td>
<td>3-D</td>
<td>0.1 m × 0.1 m × 0.1 m</td>
<td>adapt</td>
<td>loc</td>
</tr>
<tr>
<td>Gooseff et al. (2006)</td>
<td>K</td>
<td>2-D V LON</td>
<td>0.20 m–[0.3–0.5] m</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Hester and Doyle (2008)</td>
<td>K</td>
<td>2-D V LON</td>
<td>3 m–[0.1–0.25] m</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Irvine et al. (2012)</td>
<td>K</td>
<td>3-D</td>
<td>0.5 m–[0.5–2.6] m, [0.03–0.7] m</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Kalbus et al. (2009)</td>
<td>K</td>
<td>2-D V LON</td>
<td>1 m–[0.05–0.2] m</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Kasahara and Wondzell (2003)</td>
<td>K</td>
<td>3-D</td>
<td>[0.3–0.5] m–[0.3–0.5] m, [0.15–0.3] m</td>
<td>perm</td>
<td>loc-int</td>
</tr>
<tr>
<td>Kasahara and Hill (2006)</td>
<td>K</td>
<td>3-D</td>
<td>[0.6–3.5] m–[0.2–0.5] m–[0.15–0.15] m</td>
<td>perm</td>
<td>loc</td>
</tr>
<tr>
<td>Koch et al. (2011)</td>
<td>K</td>
<td>3-D</td>
<td>NS (1.7 km–200 m–0.5 m)</td>
<td>1 h</td>
<td>int</td>
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<tr>
<td>Lautz and Siegel (2006)</td>
<td>K</td>
<td>3-D</td>
<td>0.5 m–0.5 m–[0.6–2] m</td>
<td>perm</td>
<td>loc-int</td>
</tr>
<tr>
<td>Marzadri et al. (2010)</td>
<td>K</td>
<td>3-D</td>
<td>[0.19–1.88] m–[0.06–0.5] m–[0.1] m</td>
<td>perm</td>
<td>loc-int</td>
</tr>
<tr>
<td>Marzadri et al. (2011)</td>
<td>K</td>
<td>3-D</td>
<td>NS (16.9 m–2.6 m–1.6 m)</td>
<td>perm</td>
<td>S</td>
</tr>
<tr>
<td>Miglio et al. (2003)</td>
<td>P</td>
<td>3-D</td>
<td>[0.2–0.5] m–[0.2–0.5] m–[0.05–0.15] m</td>
<td>600 s</td>
<td>loc</td>
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<tr>
<td>Mouhti et al. (2013)</td>
<td>P</td>
<td>2-D V</td>
<td>[0.01–0.1] m–[0.01–0.1] m</td>
<td>min</td>
<td>loc</td>
</tr>
<tr>
<td>Munz et al. (2011)</td>
<td>K</td>
<td>3-D</td>
<td>0.5 m–0.5 m–[0.1–2.48] m</td>
<td>1 h</td>
<td>loc</td>
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<tr>
<td>Osman and Bruen (2002)</td>
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<td>2-D V LAT</td>
<td>NS (360 m–21 m)</td>
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<td>S</td>
</tr>
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<td>Pryet et al. (2013)</td>
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<td>2-D H</td>
<td>1 km–1 km</td>
<td>adapt</td>
<td>int-rég S</td>
</tr>
<tr>
<td>Revelli et al. (2008)</td>
<td>K</td>
<td>2-D H</td>
<td>NS ([0.22–4.4] km–[0.19–3.8] km)</td>
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<td>int</td>
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<td>Rushton et al. (2007)</td>
<td>K</td>
<td>2-D V LAT</td>
<td>20 m–0.2 m</td>
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</tr>
<tr>
<td>Saenger et al. (2005)</td>
<td>K</td>
<td>V LON</td>
<td>0.1 m–0.02 m</td>
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<td>loc</td>
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<td>Saleh et al. (2011)</td>
<td>K</td>
<td>2-D H</td>
<td>[1–4] km–[1–4] km–[–] m</td>
<td>1 j</td>
<td>reg</td>
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<tr>
<td>Sawyer and Cardenas (2009)</td>
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<td>2-D V LON</td>
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<td>loc</td>
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<td>Storey et al. (2003)</td>
<td>K</td>
<td>3-D</td>
<td>[1–8] m–[1–8] m–[0.25–0.42] m</td>
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<tr>
<td>Sulis et al. (2010)</td>
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<td>[1–80] m–[1–80] m–[0.0125–0.5] m</td>
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</tr>
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<td>Tonina and Buffington (2007)</td>
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<td>loc</td>
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<td>2-D V LON</td>
<td>0.5°–0.5°–[–] m</td>
<td>1 d</td>
<td>reg</td>
</tr>
<tr>
<td>Vergnes and Decharme (2012)</td>
<td>K</td>
<td>2-D V LON</td>
<td>0.5°–0.5°–[–] m</td>
<td>1 d</td>
<td>con</td>
</tr>
<tr>
<td>Wondzell et al. (2009)</td>
<td>K</td>
<td>3-D</td>
<td>[0.125–2] m–[0.125–2] m–[0.16–0.4] m</td>
<td>perm</td>
<td>loc</td>
</tr>
</tbody>
</table>

Exch (stream–aquifer exchanges’ model); K: conductance model; P: Pressure continuity; V: vertical; LAT: lateral; LON: longitudinal; H: horizontal.
Resolution: NS: not specified (total extension between parenthesis); a cell size not specified in the paper.
Spec (Specificities) \(\Delta x\) (spatial); \(\Delta t\) (temporal): perm: steady state; adapt: adaptive time step.
Scale: loc: local; int: intermediate; reg: regional; con: continental.
CS (Case Study): S: synthetical; L: lab experiment; R: real.

### Table 3. Other DPBMS for intermediate and watershed scales – complementary with Table 1.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Reference Model</th>
<th>Spatial Size</th>
<th>Time Period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bitteli et al. (2010)</td>
<td>DW 2-D FD RE 3-D FD K seq</td>
<td>3 ha</td>
<td>1 yr</td>
</tr>
<tr>
<td>Dawson (2008)</td>
<td>NS 1-D FE RE 2-D FE P seq</td>
<td>2-D vertical</td>
<td>30 min</td>
</tr>
<tr>
<td>Gunduz and Aral (2005)</td>
<td>SV 1-D FD DE 2-D FE L sim</td>
<td>1800 km²</td>
<td>3 months</td>
</tr>
<tr>
<td>Hussein and Schwartz (2003)</td>
<td>KW 1-D FD DE 3-D FD P seq</td>
<td>256 km²</td>
<td>100 yr</td>
</tr>
<tr>
<td>Kim et al. (2012)</td>
<td>SV 2-D FV DE 3-D FD K seq</td>
<td>64 km²</td>
<td>200 h</td>
</tr>
<tr>
<td>Liang et al. (2007)</td>
<td>NS 2-D BE L sim</td>
<td>8–40 ha</td>
<td>2–3 min</td>
</tr>
<tr>
<td>Peyrard et al. (2008)</td>
<td>SV 2-D FE BE 2-D FD P sim</td>
<td>36 km²</td>
<td>5 yr</td>
</tr>
<tr>
<td>Qu and Duffy (2007)</td>
<td>DW 2-D FV RE 2-D FV K sim</td>
<td>0.2 ha</td>
<td>1 month</td>
</tr>
<tr>
<td>Spanoudaki et al. (2009)</td>
<td>NS 3-D FD RE K sim</td>
<td>2.5 ha–25 km²</td>
<td>30 h</td>
</tr>
<tr>
<td>Shen and Phaniyukumar (2010)</td>
<td>DW 2-D FV DE 3-D FD K seq</td>
<td>12 ha–1169 km²</td>
<td>5 h–7 yr</td>
</tr>
<tr>
<td>Singh and Bhallamudi (1998)</td>
<td>SV 1-D FD RE 2-D FD P seq</td>
<td>0.6 m²</td>
<td>15 min</td>
</tr>
<tr>
<td>Yuan et al. (2008)</td>
<td>NS 2-D FD BE 2-D FD P sim</td>
<td>160 km²</td>
<td>30 min</td>
</tr>
</tbody>
</table>

SW: Surface water; GW: Groundwater; Cg: Coupling; Sg: Solving; DW: Diffusive Wave; KW: Kinematic Wave; NS: Navier–Stokes; SV: Saint-Venant; RE: Richards Equations; BE: Boussinesq Equation; DE: Diffusivity Equation; FD: Finite Differences; FE: Finite Elements; FV: Finite Volumes; K: conductance model; P: Pressure continuity; L: Horizontal Darcy law; sim/seq: The system of equations is solved simultaneously/sequentially; a: pseudo 3-D.
Fig. 1. Nested stream–aquifer interfaces: (a) watershed-basin scale (b) intermediate-reach scale in an alluvial plain (c) cross section of the stream–aquifer interface (d) meandered reach scale (e) longitudinal river-HZ exchanges (f) water column-sediment scale. Inspired by Stonedahl et al. (2010).
Fig. 2. Scaling effects on averaged near river piezometric heads.

Legend:
- Real piezometric head
- Calculated piezometric head on fine grid mesh
- Calculated piezometric head on coarse grid mesh
Fig. 3. MIM methodological space. Axis in logarithmic scale.
Fig. 4. Localisation of two approaches in the MIM methodological space. In yellow: upscaling methodology from the local to the watershed scale based on LOMOS coupled with DPBM. In blue: regional to continental scales covered by the SWOT space borne approach. Axis in logarithmic scale.