THEMATIC ISSUE

A review on groundwater–surface water interaction highlighting the significance of streambed and aquifer properties on the exchanging flux

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Abstract

Quantification of groundwater (GW) and surface water (SW) interactions is crucial for effective water resource allocation and management. Immense progress has been made in the past few decades to address the different aspects of GW–SW exchanges. These have resulted in a large volume of literature. This work reviews in detail the mechanism of interaction and the applications of different field and modelling techniques. The review of flux quantification methods identifies the streambed and the aquifer beneath as two major components affecting the interactions. It is observed that the streambed is highly idealised in modelling studies, and the significance of aquifer properties in the flux quantification is found to be less emphasised. Therefore, attempts are made to highlight the potential significance of both streambed and the aquifer properties through a 2D numerical experiment. Using a superimposed GW–SW system and appropriately grouping the system parameters (as hydraulic and geometric), the experiment shows that the aquifer properties can dominate exchanging flux under certain conditions, e.g., at higher streambed conductance. The work provides suggestions to modify the widely used Darcy's approach to include aquifer properties.

Keywords GW · SW interaction · Streambed · Aquifer · Hydraulic parameters · Modelling · Field investigation

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Introduction

Groundwater (GW) and surface water (SW) are considered as a single hydrological system that interacts in different physiographic and climatic landscapes (Woessner 2000). The interaction is a complex process, and hence it is challenging to characterise the interaction with a high level of certainty. Nevertheless, the various ecological significance of GW–SW interaction has increased the research interest in this field. Studies such as Brunner et al. (2010) and Ghysels et al. (2019) characterised this interaction considering the streambed and the aquifer beneath and adjacent to the streambed as two major components of this interacting system. Other studies, e.g., Woessner (2000); Banks et al. (2011), suggest that factors such as streambed topography, hydraulic properties of GW and SW, vegetation, and climatic changes, have to be included for a holistic characterisation of this interaction. Phogat et al. (2017) identifies exchanging mechanisms and quantifying the flux across the stream among the key difficulties in characterising GW–SW interaction.

The exchanging fluxes are quantified either directly in the field using methods such as seepage meter (Lee and

Cherry 1979; Rosenberry et al. 2008) or by modelling the stream–aquifer system using various available mathematical codes (e.g., RIV Package, McDonald and Harbaugh (1988), Hydrogeosphere Therrien et al. (2010)). The field techniques are limited to local scale, whereas the numerical modelling can be extended from local to catchment scale and also to address the spatiotemporal variation of stream–aquifer properties. Extensive progress has been made in developing different numerical tools (e.g., OpenGeoSys). Overall, there exists a large volume of literature that has addressed various aspects of SW–GW interaction.

Review works, e.g., Flipo et al. (2014) reviewed in detail the GW–SW quantification methods on five different scales and introduced the concept of nested stream–aquifer interface. Brunner et al. (2017) discussed models and their working mechanism on flux quantification. Barthel and Banzhaf (2016) reviewed the interaction studies on the regional scale, identifying the key challenges and lack of modelling tools at this scale. Rosenberry et al. (2020) highlighted the development and usage of seepage meters. However, to our knowledge, the compilation of different models for different scale and its parameters, have not been sufficiently attempted in other review works.

The present review work attempts to explain GW–SW interaction by describing the mechanism of the interaction and the processes, including the methods developed to quantify the flux. The work also highlights the key parameters involved in the quantification of exchanging flux in different modelling studies. Additionally, attempts are made to highlight the significance of the governing parameters through numerical experiments using a simple two-component system. The critical parameters are to be established from the preceding review works. The purpose of the numerical work has been to gauge the importance of these (aquifer and streambed hydraulic conductivity, thickness, and width) properties with a goal to specify the future direction of research in GW–SW exchange studies.

The present paper first includes a detailed review of the flux quantifications (field methods and numerical approaches), providing an extensive compilation of the literature. Then it comprises a 2D numerical experiment to highlight the potential significance of aquifer and streambed hydraulic and geometrical properties using a superimposed GW–SW system. Finally, the conclusions drawn based on the numerical experimentations and the literature work are incorporated in the paper.

GW–SW flux processes, mechanisms, and scale dependence

Fundamentally, the interaction of GW with SW occurs through either infiltration or exfiltration from the streams. Gordon et al. (1992) categorises streams as either perennial (effluent stream), or ephemeral (influent streams), or intermittent (season depending either effluent or influent). GW interacts with SW or stream in a natural system, depending on the groundwater level, aquifer type, and hydrological linkages (Malard et al. 2002). Literature (Woessner 2000; Storey et al. 2003) suggest that the magnitude of exchanging flux is mainly governed by the streambed's hydraulic conductivities and the adjacent aquifer, the relation between stream stage and the GW gradients, and the geometry and the position of the stream.

The groundwater level defines the direction of exchanges between a stream and an aquifer. The exchange can be upward into the stream (an upwelling, gaining, or effluent stream) when the groundwater head is greater than the stream stage, whereas downwards from the stream (downwelling, loosing, or influent stream) into the aquifer when the groundwater head is less than the stream stage. Woessner (2000) describes the flow orientation as a parallel flow, which can occur when the stream stage and the groundwater head are equal (Woessner 2000). Likewise, a horizontal or flow through the reach will occur where the stream stage is less than the groundwater head on either of the banks and greater on its opposite.

The hydrological linkage between the stream and the aquifer is another crucial aspect governing the GW–SW exchange Woessner (2000). A stream is completely connected with the underlying aquifer if the GW table is above the streambed bottom. Brunner et al. (2009) show that in this case, the underlying aquifer is completely saturated, and the infiltration rate varies linearly with the GW gradients. Any decline in groundwater table due to climatic changes (Brunner et al. 2009), pumping, or evapotranspiration (Banks et al. 2011) can shift the system from being a connected one to a transition one, and eventually to a disconnected system. In contrast, Xian et al. (2017) showed that the infiltration rate becomes non-linear with the changing groundwater level as the system transforms from saturated to transition and finally unsaturated state. Apart from infiltration rates and degree of saturation, the GW–SW flux is also found to depend on the degree of connectivity compared between the homogeneous and heterogeneous streambeds (e.g., Irvine et al. 2012; Kurtz et al. 2013).

Several authors (e.g., Trauth et al. 2013; Käser et al. 2014) have shown that the streambed geometry such as bars, dunes, riffles) impacts its conductivity and thus the resulting GW–SW exchange flux. The conductivity of the streambed and the aquifer exhibit spatial variability and affect both the magnitude and direction of exchanging flux at a small scale (< 100 m, see Conant (2004)) and intermediate scale (between 100 and 1000 m, e.g., Kalbus et al. 2006; Fleckenstein et al. 2006)) found that aquifer heterogeneity is more significant in the small scale ∼ 1 m). A similar observation

was presented at the intermediate scale (~ 100 m) in Fleckenstein et al. (2006).

The discussion above highlights the importance of the scale /level of the investigation. Further, the scale of the particular study defines the processes involved (e.g., nutrient cycling, biochemical reactions) and the method of estimation (see Sect. "Estimation of interaction flux") to be employed. Literature (e.g., Woessner 2000; Boano et al. 2014; Barthel and Banzhaf 2016) defines the following four different scales for the GW–SW interaction investigations: point scale, local (reach scale), sub-catchment, and catchment scale.

Literature such as Woessner (2000), Barthel and Banzhaf (2016) have extensively characterised *point scale*. As per these works, the point scale can be characterised by the processes in the pore space of the stream–aquifer boundary. The processes involve isolated or combined physical, chemical, and biological activities. The pressure gradient in and between the GW–SW is the major driving force. In contrast, the properties determining the exchange include distribution, geometry and size of pores, and streambed connectivity, and the underlying aquifer. Literature work shows that the flow at this scale has been generally quantified using Darcy's law.

The *reach scale* (or the local scale) is a larger version of the point scale consisting of stream cross-section, floodplain, and adjacent geological units. Fleckenstein et al. (2006), Barthel and Banzhaf (2016), among very few others provide details of this scale. The fundamental processes occurring in the interface at the reach scale remain the same as the point scale. However, due to the large number of observations required, the reach scale processes are not described in a discrete way as is with the point scale. Fleckenstein et al. (2006), Barthel and Banzhaf (2016) show that an increase in the scale (from point to the reach) increases the complexities of the problem, largely because of high spatiotemporal variability in governing parameters (GW–SW head and streambed leakage coefficient). Incorporating a large number of parameters in a model is then a challenging task. The *sub-catchment scale* covers the stream's drainage area, whereas the *regional scale* encompasses climatic, geological, geomorphological, and biological parameters of the catchment (Barthel and Banzhaf 2016). Developing a model at this scale with or without the inclusion of processes (e.g., contaminant transport, biological activities) is itself a challenge. Therefore, only very limited studies (e.g., Lamontagne et al. 2014)) can be found on the regional scale compared to research on other scales.

Hyporheic zone (HZ), a transitional zone where GW interacts with SW, has mostly been a focus (see Harvey and Bencala 1993; Boulton et al. 2010) in characterising GW–SW interactions at the local and point scales. HZ encompasses ecological, hydrological, hydrogeological, chemical, and biological aspects of the ecosystem, and therefore, a single definition of this zone is difficult to formulate.

Boulton et al. (2010) defined HZ based on the distribution of invertebrates. Triska et al. (1993) defined it with respect to the concentration of chemical tracers (e.g., Lautz and Siegel 2006) used the proportion of surface water in this zone to define HZ. Fischer et al. (2005) and Mucha et al. (2006) considered HZ as a biogeochemical filter and thus emphasised its characterisation in qualitative and quantitative GW–SW interaction studies. In general, hydrogeologists (e.g., Woessner 2000) define HZ as a water-saturated, transitional zone representing a critical interface between groundwater and surface water and agree with its importance in quantifying the GW–SW fluxes.

Depending on the scale of the study, the fluxes are either measured directly in the field or modelled using available field data. The following section presents an overview of the state-of-the-art techniques used for the flux estimation at the GW–SW interface, including in the HZ.

Estimation of interacting flux

Literature provides studies on GW–SW flux quantification considering factors such as evapotranspiration, clogging (Banks et al. 2011), stream nature, e.g., ephemeral and intermittent streams (Rau et al. 2017), flow conditions, e.g., lateral flow (Xian et al. 2017), vertical flow (Glose et al. 2019), streambed heterogeneities (Frei et al. 2009), water table fluctuations (Phogat et al. 2017). These factors have mostly been considered independent of each other. Only very limited studies have attempted to combine these factors for characterising the GW–SW fluxes. Phogat et al. (2017), for example, developed a 2D stream–aquifer interaction model that involved climatic, vegetative, and streamflow conditions under heterogeneous subsurface. The study pointed out that the consideration of geology and the hydraulic conductivity of the streambed and its vicinity is essential for quantifying exchanging flux. The following sections focus on literature works on the estimation of GW–SW exchange fluxes using field methods and numerical approaches.

Field methods

Literature suggests that the two most common methods used for measuring GW–SW flux are: (a) seepage meters and (b) Darcian flux-based measurements. Although relatively simple and easily applicable in the field, these methods are marred by critical limitations. Darcian flux estimation calculates the exchanges as a product of the streambed hydraulic conductivity ($K_s [LT^{-1}]$) and the vertical hydraulic gradient($i[]$), i. e., $v_d = K_s i$. Slug tests (Conant 2004), permeameter test (Landon et al. 2001), constant rate pumping tests (Kelly and Murdoch 2003), etc., are found to be used to measure K_s. Likewise, the vertical gradient (i) in the

field is estimated using a mini-piezometer (Lee and Cherry 1979), peizomanometer (Kennedy et al. 2007), potentiomanometer (Baxter and Hauer 2003). Seepage meters, when combined with a mini-piezometer, have been employed to determine the streambed conductivity (e.g., in Landon et al. 2001). This meter consists of a cylinder with an open base and a plastic bag connected at the top. The operation of the meter requires time-dependent volumetric measurement of water in the plastic bag (Lee and Cherry 1979). The water lost from the bag indicates the losing stream; otherwise, it is gaining one. A seepage meter, though a simple instrument, has several errors associated with installation. The most critical is that it disturbs the flow field. Kelly and Murdoch (2003) showed the collecting bag's material, and the volume (empty/half-filled bag) also induce an error in the seepage measurement. Several modifications include introducing automated devices replacing the seepage bag, e.g., a heat-pulse flowmeter (Taniguchi and Fukuo 1993), acoustic-velocity sensor (Paulsen et al. 2001) to capture the temporal variability of seepage can be found in the literature. Rosenberry et al. (2020) provide a very detailed study on the seepage meters.

The application of both Darcian flux estimation and seepage meters is limited at the local scale. To capture higher temporal and spatial variability, a large number of measurements are required, which is time-consuming and laborious. Brodie et al. (2009) found that seepage meters are not optimally suitable in fast-flowing or deep stream, gravelly, hard, or weedy sediment beds; besides, they require a large duration (in days) to estimate flux in a low-flux system. Similarly, several authors (e.g., Post and von Asmuth 2013; Harvey et al. 2013) have highlighted uncertainties associated with the measurement of the streambed conductivity and the vertical head gradient using seepage meters. Kennedy et al. (2010) showed that streambed heterogeneities could lead to over 30% difference in flux measurement using seepage meters and the Darcian method.

Extensive data requirements and process-based investigation mar the studies at a regional scale. Very few studies (e.g., Lamontagne et al. 2014; Barthel and Banzhaf 2016) have addressed the GW–SW interaction at the regional scale. The hydrograph separation method is one among the applied at this scale to quantify the exchange. This method is appropriate for the mountainous region and does not represent the true scenario when applied to the plain or areas with controlled runoff (Li et al. 2020). Barthel and Banzhaf (2016) present a detailed review of regional scale GW–SW interaction quantification, its characteristics, and an overview of tools and methods.

Temperature profiling is an emerging approach for the characterisation of the GW–SW interface and estimation of the flux (Hatch et al. 2006; Krause et al. 2012). The methods using temperature profiling approaches are fast and

efficient and possess possibilities for obtaining high spatial and temporal resolution (e.g., Krause et al. 2012). They also provide possibilities to acquire time series data (e.g., Hatch et al. 2006; Keery et al. 2007; Naranjo and Turcotte 2015). Another direct method for measuring exchange flux is the StreamBed Point Velocity Probe (SBPVP, e.g., Cremeans and Devlin 2017). It is a direct push device that measures the in situ water velocities at the exchanging interface with tracer on the probe surface. These tracers are independent of the hydraulic gradient information, hydraulic conductivity and do not require a long equilibration time. A comparative study by Cremeans et al. (2020) showed that both seepage meter and SBPVP provided almost identical flow velocity. Compared to that, the Darcy method and temperature profiling showed variation upto 60%. However, this study recommended Darcy and temperature profiling methods as more suitable for preliminary studies to distinguish zones worthy for detailed investigations.

The above discussed field methods are effective measurement tools, but their applicability is limited to low spatial and temporal scales. Repeated measurements are required to generate a high density of spatiotemporal data, which increases the possibility of measurement errors and uncertainties. Therefore, for a clearer understanding of the exchanging process, the field measurements are to be complemented by mathematical/numerical modelling (indirect methods). The choice of modelling technique for numerical approaches depends on the objectives and scale of the study and the availability of the field data. In the following sections, the modelling techniques are discussed under three categories—(a) heat tracer models, b) coupled modelling techniques, and (c) Darcy-based modelling techniques.

Heat tracer models

Studies by Lautz et al. (2010), Gordon et al. (2012) found that exchanging water in the stream–aquifer system carries heat that influences the stream temperature. This implies that heat can be considered as a natural tracer. Realising this, Hatch et al. (2006) and others developed heat tracer models based on 1D heat transport equation (Eq. 1, see Carslaw and Jaeger 1959) to quantify GW–SW exchange flux:

$$\frac{\partial T}{\partial t} = K_e \frac{\partial^2 T}{\partial z^2} - \frac{n v_f}{\gamma} \frac{\partial T}{\partial z}, \tag{1}$$

where $T(t, z)$ is time (t) and space (z, depth)-dependent temperature, K_e is effective thermal diffusivity, $\gamma = \frac{\rho c}{\rho_f c_f}$, the ratio of heat capacity of the streambed material ρc to that of the saturated sediment–fluid system ($\rho_f c_f$), n is porosity, and v_f is the vertical fluid velocity (positive upwards). Considering vertical water flux, a sinusoidal temperature variation and no change in temperature along

the depth of streambed, Hatch et al. (2006) provided the following analytical solution of Eq. (1) for the amplitude ratio (Ar) and temperature phase change ($\Delta\varnothing$):

$$q_{A_r} = \frac{C}{C_w}\left(\frac{2K_e}{\Delta z}ln(A_r) + \sqrt{\frac{\alpha + v^2}{2}}\right) \qquad (2)$$

$$q_{\Delta\varnothing} = \frac{C}{C_w}\left(\sqrt{\alpha - 2\left(\Delta\varnothing\frac{4\pi K_e}{P\Delta z}\right)^2}\right), \qquad (3)$$

$$\alpha = \sqrt{v^4 + \left(\frac{8\pi K_e}{P}\right)^2}, \qquad (4)$$

where C and C_w ($Jm^3°C^{-1}$) are the heat capacity of the sediment water matrix and the water, respectively; Δz (m) is the spacing between two measurement points in the streambed; $v(ms^{-1})$ and P (s) are the thermal front velocity and the period of the temperature signal. Lautz et al. (2010) further used these solutions to study interactions at upstream and downstream of the stream with different geomorphic features. Lautz et al. (2010) study showed that the approach was able to identify the effect of vertical heterogeneities in sediment hydraulic conductivity.

Gordon et al. (2012) provided a MATLAB™-based computer code (VFLUX) to calculate the vertical flux. For the development, five temperature sensors were inserted in the streambed at various depths, and Lautz et al. (2010) approach was used for the quantification of exchanging flux. In addition to Hatch et al. (2006), the developed code also adopted an analytical solution provided by Keery et al. (2007), which quantifies flux by the amplitude and phase change of the captured temperature time series. The phase change approach of estimation only provides the magnitude of the flux, unlike the amplitude ratio approach, which provides both the magnitude and direction of the exchanging flux. Glose et al. (2019) used an updated VFLUX version (VFLUX2) to quantify exchanging flux. A bulk parameter, vertically integrated hydraulic conductivity, was proposed as an alternative to the subsurface conductivity. The proposed parameter was calculated using Darcy's law, where flux was obtained from VFLUX2 and the measured hydraulic head gradients. This bulk parameter is, however, limited to the studies where subsurface heterogeneities are not essential.

Lautz (2010) compared the non-ideal assumptions for the 1D heat transport equation and found that the non-vertical flow was the major source of error in the flux estimation, whereas temperature time series data and flux measurements proved to be more reliable than Darcy-based techniques. These and other significant studies based

on heat as a tracer for exploring GW–SW interaction are tabulated in detail (parameters used, K_s, η, K_t) in Table 1.

Heat flow models have been used to study both steady and transient cases. Schornberg et al. (2010) concluded that transient temperature boundary condition was reliable for high exchanging flux and steady state for medium-to-high fluxes. Anibas et al. (2009), in their study in Aa River, Belgium, showed that both steady and transient states compared effectively in both summer and winter seasons, whereas in transitional season (spring and fall), the study showed significant deviation in the steady-state case. The same study concluded that the steady-state condition should be used in the case of a detectable difference in the temperature between the stream surface and the GW.

The major advantage of temperature-based modelling, apart from being fast and inexpensive, is that thermal conductivities of the sediment vary within a very limited range compared to highly varying hydraulic conductivities. Generally, the GW modellers neglect the temperature dependency of the leakage coefficient (conductivity divided by thickness). Only very few studies (e. g., Engeler et al. 2011) have addressed this factor. Moreover, most heat-based studies have not identified the streambed and aquifer as two different components; instead, they have treated them as a single subsurface unit (Gordon et al. 2012). Thus, these methods are not able to delineate the significance of individual units of the GW–SW system. Coupled models, presented in the next section, are more suitable when such delineations are significant.

Numerical approaches

Coupled models

Quantification of GW–SW interaction by coupling a surface water model with a groundwater flow estimation model can be found in several works (e.g., Yuan et al. 2011; Trauth et al. 2013; Xian et al. 2017). Coupling of the two models is done so that the results from the surface water model are applied as a boundary condition for the groundwater models (e.g., Smits and Hemker 2004). Other studies (including Brunner and Simmons 2012; Tang et al. 2017) used Hydro-GeoSphere, a fully coupled model for estimating flux and the characterisation of the GW–SW. A detailed tabulation of the models and the parameters used (e.g., η, w_s, K_a, t_a) is presented in Table 1. The table comprehends various models used for the exchange flux quantification, the scale used in the study and highlights the parameters used in the study.

Darcy-based models

Darcy's method is the most widely used groundwater modelling method (Osman and Bruen 2002; Brunner et al. 2010;

Table 1 Models, their parameters and scales, used for GW–SW interaction

	Site and reference	Model used	Dimension/scale	Parameter considered*	Key feature
Heat-based models	Ischua Creek New York (Becker et al. 2004)	Current meter / Stream temp. survey	Area averaged (40 km)	η	Streamflow measurement proved to be most reliable but are practical over long stream reaches
		Heat transport modelling	1D	–	Temperature gradient is a point estimate, whereas temperature survey gives flux only when coupled with streamflow information
	Pine river Canada (Conant 2004)	Streambed temperature mapping	–	–	Developed an empirical relation relating streambed temperatures and flux quantified from mini-piezometer
	Schachtgraben, Germany (Schmidt et al. 2006)	Streambed temperature measurement	1D	t_a	Valid only for vertical flow / Could be applied as a preliminary method for identifying sites requiring detailed investigation
	Pajaro River, Central Coastal California (Hatch et al. 2006)	Time-series thermal data	1D	t_a	Insensitive to streambed scour and sedimentation / Limited to the homogeneous and saturated streambed
	Pine River Canada (Schmidt et al. 2007)	Analytical method/VS2DH	1D	η, K_s, K_t, t_a	Minimal water temperature data and thermal properties requirement / Method is limited to vertical flow in gaining stream / Quasi-steady state is justifiable only for a certain period of the year
	–(Lowry et al. 2007)	Distributed temperature sensing (DTS)	–	–	DTS complements direct flux measuring techniques / Installation limitations
	Aa River, Belgium and Lake 77, Germany (Anibas et al. 2009)	VS2DH/FEMME	1D	η, K_t, t_a	Steady state and transient compared well for summer and winter
	Synthetic Model (Schornberg et al. 2010)	VS2DH	1D	K_s, η, K_t, t_a	Heterogeneity causes larger errors as compared to structural arrangements of geologic units in the sediments / Transient temperature boundary condition is reliable for high exchanging flux and steady state for medium-to-high fluxes
	Swiss prealpine River Thur (Vogt et al. 2010)	fiber-optic distributed temperature sensing	1D	–	High spatial resolution of measurement captured discontinuities in the streambed
	Red Canyon Creek, Wyoming, USA (Lautz et al. 2010)	Heat transport modelling	1D	K_s, η, K_t	Captured the influence of streambed heterogeneity in vertical flux exchange

Table 1 (continued)

	Site and reference	Model used	Dimension/scale	Parameter considered*	Key feature
	–(Lautz 2010)	Temperature measurement combined with VS2DH	1D and 2D	K_s, η, K_t	Temperature time series is better than Darcy-based technique for capturing spatial variability of flux Non-vertical flow velocities are a significant source of error
	Aa River, Belgium (Anibas et al. 2011)	STRIVE	1D	K_t	Discharging flow conditions Higher fluxes on the banks than in the centre
	Ninemile Creek, Marcellus, New York (Gordon et al. 2012)	VFLUX	1D	K_t, η	Can capture curved hyporheic flow paths Point estimate for temperature time series
	–(Xie et al. 2015)	RiverHeat	1D	η, w_s, t_a	Simultaneously solved streamflow and heat transport Tested heat tracer method for reach scale and suggested the use of other tracer techniques for better estimation
Coupled models	Well field for water-supply, Netherlands (Smits and Hemker 2004)	Duflow-MicroFem	1D	–	Computationally light
	Red Canyon Creek (Lautz and Siegel 2006)	MODFLOW and MT3D	3D	η, K_a	Better than particle tracking as it compares well with field data
	Tame, Birmingham (Ellis et al. 2007)	FAT3D-UNSAT and MODFLOW	–	η, K_s	Flow is affected by - Large scale—urban river geometry and aquifer management Channel scale—local geomorphologic units
	Fuxianhu Lake catchment, southwest China (Zhang and Werner 2009)	WATLAC- MODFLOW (CHD) 2005	–	–	WATLAC—computational lighter than MODHSM for small field and time scales Easy spatial discretisation for coupling
	–(Fenske et al. 2009)	HEC-RAS—MODFLOW, model interface OpenMI	–	–	Two-way coupling- Heads and flow rates passed from MODFLOW to HECRAS Stream stages, conductivities, and bottom elevation passed from HECRAS to MODFLOW
	Synthetic Model (Ward et al. 2010)	COMSOL—solute transport and electric resistivity (ER) simulations	3D	η, K_a, t_a	ER method supplements exchanging flux data for slow exchanging subsurface region
	–(Banks et al. 2011)	HydroGeoSphere	2D	η, K_s, K_a	Vegetation as an important factor governing the connectivity between GW and SW Clogging layer is conceptualised

Table 1 (continued)

Site and reference	Model used	Dimension/scale	Parameter considered*	Key feature
Hardhof, Zurich Switzerland (Engeler et al. 2011)	SPRING	3D	$K_s, \eta, K\,t_a$	Temperature dependence leakage coefficient Leakage concept is inapplicable in flood events Coupled variably saturated groundwater flow and heat transport
Synthetic Model (Irvine et al. 2012)	HydroGeoSphere and GCOSIM3D	3D	K_a, t_a, K_s	Studied streambed heterogeneity in connected, disconnected, and transitional streams Homogenous streambed simulations are unsuitable for transitionally connected stream–aquifer
Synthetic model (Trauth et al. 2013)	CFD Toolbox OpenFOAM v1.7.1 MIN3P3D	3D	η, w_s, K_a, t_a	Studied the effect of stream infiltration, GW-flow, and streambed morphology in pool–riffle sequence
–(Noorduijn et al. 2014)	MODFLOW- NWT	–	$\eta, K_s, t_a, L_{rep}^{\#}$	Concept of representative stream length introduced Illustrated the importance of distance of piezometer from the stream
Pedler Creek, South Australia (Batlle-Aguilar et al. 2015)	HydroGeoSphere	2D	K_a	Calibrated and validated a 2D model for an ephemeral stream assuming isotropic and no-flow conditions at the soil surface Head difference between stream and aquifer not sufficient to determine connectedness
Synthetic model (Xian et al. 2017)	HydroGeoSphere	2D- 3D	K_a, t_a, K_s	Analysed the transition from connected to disconnected stream–aquifer system with heterogeneous streambed
Darcy-based models				
Synthetic model (Osman and Bruen 2002)	MODFLOW	2D	K_a, t_a, K_s	Modified MODFLOW to MOBFLOW represents negative pressure in losing streams with a clogging layer Aquifer properties affect exchanging flux
Speed River, Southern Ontario (Storey et al. 2003)	MODFLOW	3D	η, K_s	Amount, the extent of infiltration, and travel time of exchanging flux depend on hydraulic conductivity, GW flux, and gradient Vertical flow paths are more persistent as compared to lateral flow paths
Synthetic model (Brunner et al. 2010)	MODFLOW and HydroGeoSphere	2D	K_a, t_a, K_s	Suggested the inability to represent flow in the unsaturated zone Mismatch in streambed and stream width discretisation produces differences in the exchanging flux

Table 1 (continued)

Site and reference	Model used	Dimension/scale	Parameter considered*	Key feature
Synthetic model (Ghysels et al. 2019)	MODFLOW	2D	K_a, t_a, K_s	Modified RIV Package addressed the heterogeneity of the streambed and the lateral flow from the aquifer
Other modelling tools –(Crosbie et al. 2014)	Vertical resistivity soundings (VES)/HYDRUS	–	–	Observed that exchanging flux is more sensitive to clogging layer thickness than its conductivity
Synthetic model (Frei et al. 2009)	PARFLOW	3D	K_a	Spatial and temporal aquifer heterogeneity and specific architecture of high conducting zones are crucial factors governing GW–SW exchange in a deep GW table
Scott Creek, South Australia (Phogat et al. 2017)	HYDRUS-2D/3D Version 2.xx	2D	–	Illustrated the importance of streambed thickness and conductivity and the impact of water table on the transition from connected to the disconnected stream–aquifer system
–(Batlle-Aguilar and Cook 2012)	HYDRUS, Field Experiment	2D	K_a	Illustrated about the transient infiltration and recharge beneath streambed and also the streambank infiltration
Cosumnes River basin (Fleckenstein et al. 2006)	–	50 river-kilometres	–	Simulated river–aquifer exchange for a set of different heterogeneity models to assess how textural heterogeneity influences seepage patterns and low river flow
Lule River Sweden (Siergieiev et al. 2015)	FEFLOW 6.2, FePest	2D	η, K_s, K_a	Effect of stream–aquifer geometry and streambed hydraulic conductivity affects the bank storage on a regulated stream

^{n}subsurface porosity, #representative stream length, *symbols are defined in Sect. "Numerical experiment"

Ghysels et al. 2019). These groundwater models are further applied in the estimation of the GW–SW exchange. An example of this can be the extension of the most common groundwater code MODFLOW for quantifying stream–aquifer interaction using one of the following four different codes: River package, RIV (McDonald and Harbaugh 1988), Stream Package (Prudic 1989), Streamflow Routing Package, SFR1 (Prudic et al. 2004) and SFR2 (Niswonger and Prudic 2005). The RIV package represents river as a head-dependent boundary and computes exchanging flux as (see Ghysels et al. 2019) using

$$Q_s = C(h_s - h_{\text{aqu}}), \tag{5}$$

where $Q_s[L/T]$ is the exchanging flux per unit length, $C = (K_s W_s L_s)/t_s [L^2/T]$ the conductance of the streambed, $K_r[L/T], W_s[L]$, $L_s[L]$ and $t_s[L]$ are the hydraulic conductivity, width, length, and thickness of the streambed, respectively. h_s and h_{aqu} are the heads at the stream and the aquifer below the stream, respectively. In general, research works (e.g., Ghysels et al. 2019) quantifying the exchanges consider the head difference between the stream head and the head in the underlying aquifer. If the h_{aqu} falls below the riverbed bottom, head in streambed bottom (h_{sbot}) replaces h_{aqu} in Eq. (5). Several studies used the RIV package or other extension of MODFLOW or coupled MODFLOW with other surface water tools to estimate exchanges (complete details are provided in Table 1).

Brunner et al. (2010) presented a few limitations of the RIV package in representing unsaturated zone overlying aquifers. Further, this study found MODFLOW not to be the best suitable option for the losing stream–aquifer system. Osman and Bruen (2002) developed a code called MOB-FLOW to overcome the limitations of the MODFLOW-RIV package. MOBFLOW includes the effect of negative pressure gradients in the underlying aquifer. Ghysels et al. (2019) also modified the RIV package to include the effect of heterogeneity and lateral seepage from the streambed.

Overview of modelling techniques

Table 1 presents a list of GW–SW exchange flow models comprehended from the literature. The table classified the models as heat-based models, coupled models, and Darcy-based models. The tabulated list highlights the dimension/scale of the study and the key model parameters. Heat-based models used the combination of field measurement and the modelling tools, e.g., VS2DH, whereas the MODFLOW RIV package is the most commonly used Darcy-based model, used in about 20% of the presented studies. From different modelling studies, it is clear that water in GW–SW interaction exchanges laterally as well as vertically, essentially making the system 3D. However, about 80% of the

presented studies have used either 1D or 2D models. Further, very few studies have focused on the regional scales (Barthel and Banzhaf 2016).

The parameterisation of the model also varies significantly. The parameters included in most of the studies are the streambed and the aquifer hydraulic (Darcy or coupled) or thermal (heat-based) conductivity, streambed geometry (MODFLOW). The discussion provided in Sec. "Field methods" showed that the measurement of streambed properties is very challenging and marred by a high level of uncertainties. For example, in the RIV package of MODFLOW, the head is required to be measured in the aquifer just below the streambed. But it is practically impossible to delineate the streambed. Hence the streambed parameters are either estimated indirectly or used as a calibration factor. On the other hand, the measurement of the head using the field techniques is marred by a large number of errors and challenges (see Sect. "Field methods"). Also, in the natural systems, neither the streambed nor the aquifer beneath is homogenous. Most of the studies provided in Table 1 have either ignored the streambed or have simplified it to be a homogenous, isotropic layer (e.g., Brunner et al. 2010). Only very few studies have addressed the heterogeneity of the streambed (e.g., Ghysels et al. 2019).

The significance of the aquifer beneath the streambed in GW–SW interaction is highlighted in works of, e.g., Lackey et al. (2015), whereas the works of Fleckenstein et al. (2006), Crosbie et al. (2014), Siergieiev et al. (2015) highlights the significance of streambed properties—hydraulic conductivity and geometry. However, the literature compilation (Table 1) does not suggest the dominant property affecting the exchanging flux. With experience gained from literature works, this study intends to highlight the role of both streambed and the aquifer properties in GW–SW exchange. The intention here is to suggest future research directions that would provide a better understanding of GW–SW flux.

Numerical experiment

With experience gained in developing this review work, it has been attempted to identify the criticality of the aquifer and streambed properties in quantifying GW–SW flux through numerical experiments. The aim here has also been on suggesting the future direction of research works that would provide a better understanding of GW–SW flux and an improvement of approaches for field investigations for the quantification of GW–SW flux on different scales.

The conceptual model is a symmetrical 2D, steady-state homogeneous system with two components: (1) a streambed and (2) an underlying aquifer. This stream–aquifer system is assumed to be a completely connected system of a losing stream. The underlying aquifer is thus saturated.

Fig. 1 Model setup used in OpenGeoSys v6 (Kolditz et al. 2012)
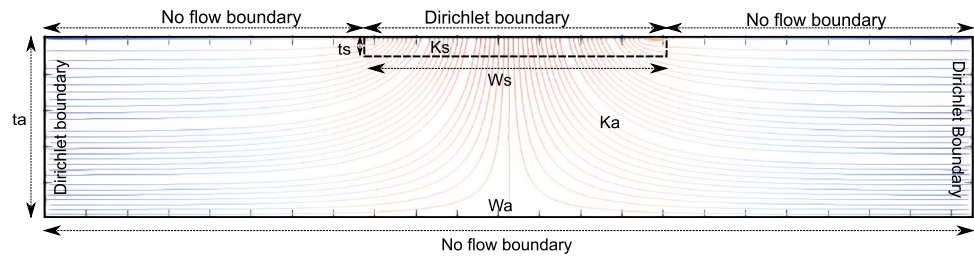
The conceptual setup can also be considered as a super-position of infiltration from stream to the GW flow in the aquifer. This allows simulating infiltration and GW flow separately by simply adjusting the boundary conditions, a major advantage of this simplified conceptual SW–GW system.

The stream in the conceptual system is represented by a constant head boundary making it the only water source in the system. The fluvial plain, representing the location between the edge of the stream and the domain edge, is a no-flow boundary. The aquifer bottom is a no-flow boundary, whereas the identical constant head boundary at the left and right of the model makes the model aquifer a non-flowing one (see Fig. 1). Thus, the difference of head between the stream (higher) and the domain edge ensures that the stream is the only water source in the model.

The aquifer width or lateral extent of the modelled aquifer is the point where the head in the aquifer is being assigned (e.g., as in Brunner et al. 2010; Ghysels et al. 2019). If the aquifer head boundary is close to the stream edge, the flux will have the influence of both vertical stream infiltration and horizontal groundwater flow. The flowline near the stream will hence not be horizontal (see Fig. 1). On the other hand, if the extent is large enough, the flux's vertical component is negligible at the edge boundaries, i.e. flowlines are horizontal. Thus, the model extent has been set based on different numerical experiments in a manner that the model width does not significantly influence the flux. The vertical component is related to the GW–SW flux and is dominant for infiltrating stream. Thus, it can also be considered as a measure of the impact of the GW–SW flux on the shape of the flow field as a function of the distance to the stream.

OpenGeoSys v6 (Kolditz et al. 2012) is used for simulations. Based on Brunner et al. (2010), the considered stream widths as 7.5 m, 15 m, and 30 m. Likewise, the considered thickness of the streambed, based on Xian et al. (2017), are varied from 0.001 m to 0.75 m. The conductivity of the streambed ranges between 1E-7 and 1E-5 m/s (fine silt to the silty sand range), and that for aquifer ranges between 1E-5 and 1E-4 m/s (fine sand to medium sand). Due to the strong vertical gradient across the streambed, a finer unstructured mesh of optimal mesh density 0.3 is used near and in the streambed, and a coarser mesh density of 0.5 is provided in the aquifer. The discretisation is based on the percentage

difference between the simulated flux at different mesh densities with respect to simulation time.

Based on the preceding review works, the numerical experiment considers the following five parameters: hydraulic conductivity of stream (K_s) and aquifer (K_a), stream (t_s) and aquifer thickness (t_a) and stream width (w_s). These parameters can be grouped as geometrical parameters (t_s, t_a, and w_s) and hydraulic parameters (K_s and K_a). Also, they can be classified on the basis of the ease of measurement in the field. t_a, K_a and w_s, (t_a and K_a) can be quantified with very high accuracy; whereas, t_s and K_s can be considered as difficult parameters as accurately quantifying them is not easy. Further, the difficult parameters are grouped as a K_s/t_s ratio referred to hereafter as specific conductance (C_s).

The results of the numerical modelling are shown in Fig. 2, wherein the flux (Q_s) (normal scale) versus streambed conductance (C_s) (logarithmic scale) have been plotted for various combinations of the parameters. All curves have a logistic function like behaviour. For very low values of streambed conductance, the flux tends to zero. This corresponds to the expectation that low conductance of the streambed leads to a sealing of the stream towards the aquifer. If the flux is also plotted on a logarithmic scale (log–log plot), then the linear relation between the conductance and the flux (Eq. 5) is valid for the initial parts of the curves, i.e. lower values of the streambed conductance. But all the curves show that the linear relation is not valid for higher values of the streambed conductance. The curves approach a limit that depends on the curve parameter. That means the conductance of the streambed, in the case of high values, is not controlling the flux. In Fig. 2a, it can be seen that the flux limit is increasing with the width of the stream. This is reasonable because for high values of the streambed conductance and for the given aquifer parameters, the flux is controlled by the size of the interface, i.e. the width of the stream (also seen in Ghysels et al. 2019). Figure 2b and c illustrates the impact of the thickness and the hydraulic conductivity on the flux limit for a given stream width and if the streambed conductance is high. Also, this behaviour can easily be explained. The case of high streambed conductance corresponds to the case where there is no specific streambed layer. That means that the aquifer properties now control the flux. An increased hydraulic

(a)

(b)

(c)
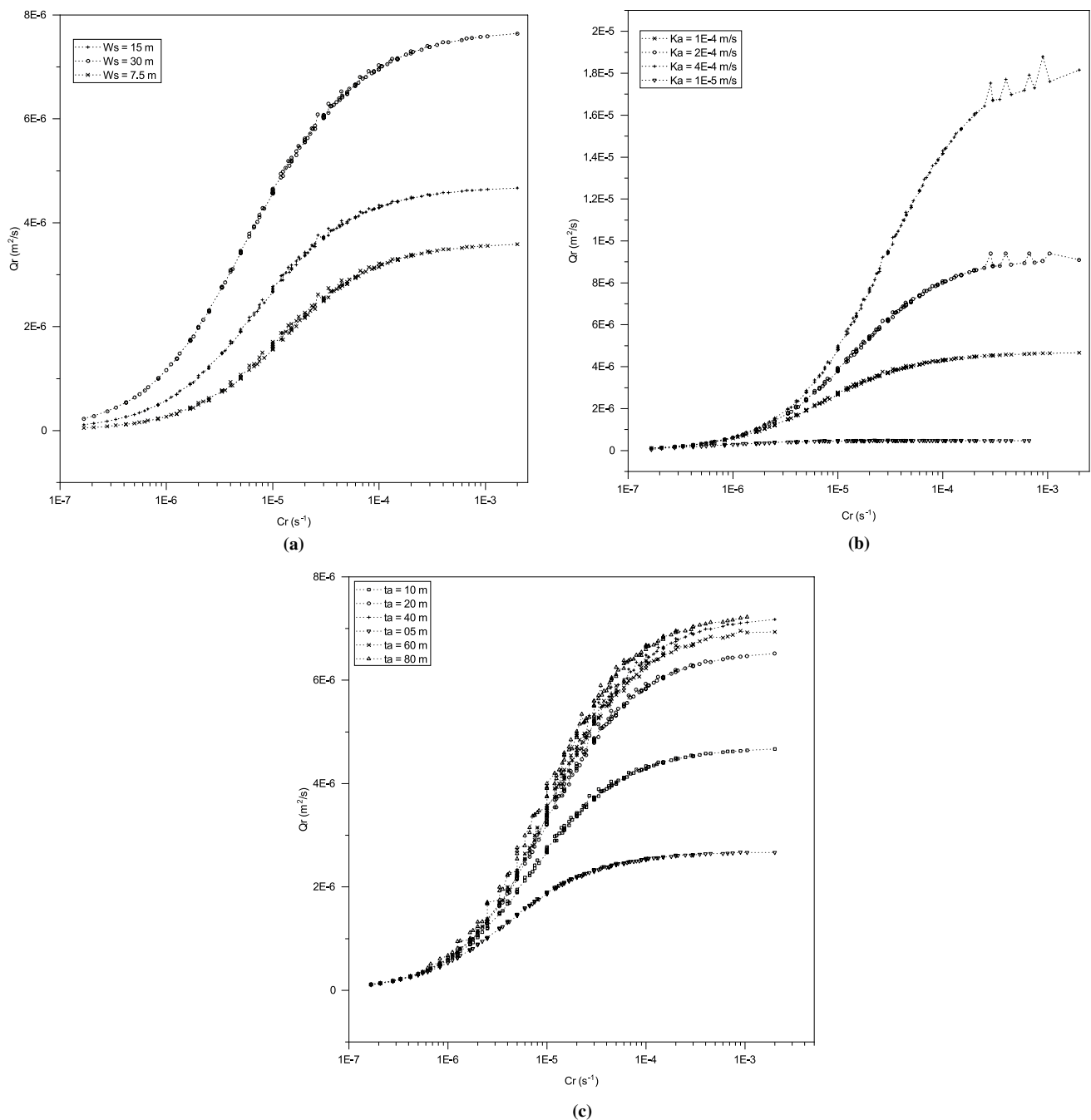
Fig. 2 a Variation of flux with streambed width, $W_a = 45\text{m}, t_a = 10\text{m}, \Delta h = 0.1\text{m}$ and $K_a = 1E - 4$ m/s (Fig. created using Origin 2021). **b** Variation of flux with aquifer hydraulic conductivity $W_a = 45$ m, $w_s = 15$ m, $\Delta h = 0.1$ m and $t_a = 10$ m (Fig. created using Origin 2021). **c** Variation of flux with aquifer thickness $W_a = 45$ m, $w_s = 15$ m, $\Delta h = 0.1$ m and $K_a = 1E - 4$ m/s (Fig. created using Origin 2021)

conductivity, as well as an increased aquifer thickness, will cause an increase in the flux. However, the experiment shows the deep aquifers (here > 40 m) have very little influence on the exchanging flux.

Conclusion

GW–SW interaction has become an important research area due to its multifarious significance in the riverine ecosystem. A better understanding of the interacting processes requires comprehensive knowledge of the interacting mechanism and flux quantification. The field

quantification methods, e.g., seepage meters, are limited to local scale and do not capture spatial variations of the streambed and the aquifer properties. The numerical modelling methods are more extensive and can be extended to capture spatial as well as temporal variations of the streambed and the aquifer properties. These methods are either based on heat tracer-based methods (40% of 37 cases), Darcy-based methods (22%), or coupled models (35%). MODFLOW RIV package and HydroGeoSphere being the most common codes used in Darcy and coupled models, respectively.

The scale of the study area defines the factors governing the flux quantification in a numerical model. Most of the studies, 97% of the 37 presented studies, focused on the local and the reach scale, and only 3% focused on the regional scale. Among the local and reach scale studies, 36% were 1D modelling studies, 28% as 2D, and only 19% as 3D studies. 1D models are mostly spatially limited, whereas 3D models have addressed spatial heterogeneities of the stream–aquifer system. However, at all scales and dimensions, the literature identifies the streambed and the underlying aquifer as two major components of the GW–SW interaction system, and their hydraulic and geometrical properties being the key parameter for the flux quantification. The significance of aquifer properties has been highlighted in only 2 out of 37 presented studies. Hence numerical experiments were performed to gauge the significance of individual aquifer properties.

Numerical experiments on a 2D, two-component, symmetrical setup of connected losing stream showed that the flux (normal scale) versus specific streambed conductance (normal scale) follows a logistic-like function. Accordingly, the flux becomes constant at higher specific streambed conductance, indicating the control of aquifer parameters on the flux at high specific conductance of the streambed ($> 1\text{E-}5 \text{ s}^{-1}$ in this study). Similarly, for the low specific conductance of the streambed ($< 1\text{E-}6 \text{ s}^{-1}$), the aquifer properties only have minimal impact on the flux. Other parameters in the logistic-like function (W_s, t_a and K_a) governs the magnitude of the limiting flux. This highlights the significance of the aquifer properties in the GW–SW interaction quantification. Therefore, the modification of the widely used approach (Eq. 5) to incorporate the aquifer properties (t_a and K_a) is suggested. The idealised numerical setup presented in this work can be extended to incorporate the effects of spatial variability of hydraulic properties of the streambed as well as the aquifer properties. The effect of different boundary conditions representing various GW flow conditions like—unequal heads on the left and right edge of the model domain could be explored using this setup. A 3D extension of the present setup could be employed to study the GW–SW interaction in different GW flow orientations.

Furthermore, the model approach can be used for a better understanding of significant properties—streambed width, aquifer hydraulic conductivity, and thickness, which controls the flux between surface water and groundwater.

Declarations

Conflicts of interest None.

Code availability Not applicable.

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