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Influence of aquifer and streambed heterogeneity on the distribution of groundwater discharge

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Abstract

The spatial distribution of groundwater fluxes through a streambed can be highly variable, most often resulting from a heterogeneous distribution of aquifer and streambed permeabilities along the flow pathways. In a previous study, observed temperature profiles in the streambed of a small stream in Germany were used to calibrate the subsurface parameters of a groundwater flow and heat transport model of the stream-aquifer system. Based on the model results, we defined four scenarios to simulate and assess the interplay of aquifer and streambed heterogeneity on the distribution of groundwater fluxes through the streambed: (a) a homogeneous low-K streambed within a heterogeneous aquifer; (b) a heterogeneous streambed within a homogeneous aquifer; (c) a well connected heterogeneous low-K streambed within a heterogeneous aquifer; and (d) a poorly connected heterogeneous low-K streambed within a heterogeneous aquifer. The results showed that the aquifer has a stronger influence on the distribution of groundwater fluxes through the streambed than the streambed itself. However, a homogeneous low-K streambed, a case often implemented in regional-scale groundwater flow models, resulted in a strong homogenization of fluxes, which may have important implications for the estimation of peak mass flows. The simulation results with heterogeneous low-K streambeds, whether or not well connected to the aquifer, were similar to the results of the base case scenario without a separate parameterization of the streambed, despite the lower permeability. We conclude that predictions of water flow and solute transport may significantly benefit from heterogeneous distributions of both aquifer and streambed properties in numerical simulation models.

1 Introduction

Groundwater fluxes at the interface between aquifers and streams can show strong variations in space and time at different scales (e.g., Ellis et al., 2007; Krause et al., 2007). This is important since the magnitude of groundwater discharge across the streambed influences the exchange with and the size of the hyporheic zone (Boano et al., 2008; Cardenas and Wilson, 2007) which plays a critical role for the functioning of stream ecosystems (Brunke and Gonser, 1997). For example, the exchange of water between aquifers and streams has important implications for the hydrochemistry of the streambed sediments (Malcolm et al., 2003), thus influencing biogeochemical nutrient cycling and habitat quality. A heterogeneous distribution of groundwater fluxes and hyporheic exchange flows leads to a patchy distribution of biogeochemical gradients and interstitial fauna (Boulton et al., 1998; Malcolm et al., 2004).

Spatial heterogeneities of groundwater fluxes through the streambed also impact the fate and transport of contaminants between aquifers and streams (e.g., Conant et al., 2004; Kalbus et al., 2007; Chapman et al., 2007). Schmidt et al. (2008) showed that the highly variable groundwater fluxes observed at a small stream resulted in a significant tailing of contaminant mass flow rates compared to the theoretical homogeneous case.

It is commonly assumed that the groundwater flux across streambeds is controlled by the heterogeneity of the connected aquifer (e.g., Wondzell and Swanson, 1996; Wroblicky et al., 1998; Storey et al., 2003; Conant, 2004). The properties of the streambed sediments may further contribute to the heterogeneous distribution of groundwater fluxes (Conant, 2004; Ryan and Boufadel, 2006, 2007). Also, geomorphologic features at different spatial scales were shown to cause variabilities of water exchange across the groundwater – surface water interface (Kasahara and Wondzell, 2003; Cardenas, 2008). Infiltrating stream water caused by streambed irregularities further leads to a very complex exchange pattern (Savant et al., 1987; Salehin et al., 2004; Gooseff et al., 2006).

Our focus is on the influence of heterogeneous distributions of hydraulic conductivity (K) in the aquifer and the streambed deposits on the spatial distribution of groundwater fluxes across the streambed. In a few recent modeling studies, subsurface heterogeneity was included to simulate stream-aquifer interaction. Chen and Chen (2003) considered anisotropic and layered aquifers as well as streambeds with different hydraulic

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conductivities in their simulations of stream-aquifer interactions, but did not include within-layer heterogeneity. Bruen and Osman (2004) studied the effect of spatial variabilities of aquifer K on stream-aquifer seepage flow, but did not consider a separate analysis of the influence of streambed properties. Cardenas et al. (2004) simulated the impact of heterogeneous streambed deposits on hyporheic zone geometry, fluxes, and residence time distributions, but did not include the groundwater component. Fleckenstein et al. (2006) investigated the effect of aquifer heterogeneity on the distribution of seepage on an intermediate (10^2 m) scale. In this study we look at the influence of both aquifer and streambed heterogeneity on the distribution of fluxes on the metre-scale.

2 Background

Along a 220 m reach of a small stream in Germany, streambed temperatures were mapped with high resolution by Schmidt et al. (2006). The stream is a man-made stream which partially penetrates a Quaternary alluvial aquifer. It is about 3 m wide and has an averave water depth of 0.6 m. The mean annual discharge is 0.2 $m^3 s^{-1}$ at a gradient of 0.0008 $m\,m^{-1}$. The streambed consists of a 0.6 m layer of crushed rock. The interstices of the coarse crushed rock grains are filled with allochthonous, sandy, alluvial material. The connected aquifer is unconfined with a mean saturated thickness of about 8 m and consists of sandy gravel. Further information about the study site can be found in Schmidt et al. (2006, 2008) and Kalbus et al. (2007, 2008a).

The streambed temperatures were mapped at depths between 0.1 and 0.5 m below the streambed surface by temporarily inserting a stainless steel multilevel temperature probe. The measurements were spaced at intervals of roughly 3 m. Based on the observed temperature profiles, Schmidt et al. (2006) estimated groundwater fluxes through the streambed by applying a one-dimensional analytical solution of the heat-advection-diffusion-equation. From both the temperature observations and the flux calculations, considerable spatial heterogeneity of the groundwater discharge was observed, ranging from no discharge up to 455 $L\,m^{-2}\,d^{-1}$ with a reach-average flux of

$58.2\,\mathrm{L\,m^{-2}\,d^{-1}}$. For more details on the method and the underlying assumptions the reader is referred to Schmidt et al. (2006).

The observed spatial heterogeneity was assumed by Kalbus et al. (2008a,b) to be caused by the properties of the connected aquifer. Even though observed streambed temperatures are temporally highly variable (e.g., Westhoff et al., 2007), the temperature distribution at a certain point in time is a result of the spatial distribution of subsurface permeabilities, which is constant in time, and the flow and temperature boundary conditions at the respective point in time. Focussing on the spatial variabilities, Kalbus et al. (2008a,b) simulated groundwater flow and heat transport through the streambed at the stream reach investigated by Schmidt et al. (2006). They included stochastically generated fields of K to represent the aquifer properties. The K-fields were generated from the mean and variance of the $\ln(K)$ data and from the correlation lengths in each direction. These input parameters were obtained from field data (injection logs (Dietrich et al., 2008) and pneumatic slug tests (Butler et al., 2000)). The variance of $\ln(K)$ was taken as a measure for the heterogeneity of the aquifer. However, the variance of $\ln(K)$ calculated from the field data was found to be too small to cause the observed spatial distribution of temperatures and groundwater fluxes through the streambed. By using the observed temperature distribution for calibration, the variance of $\ln(K)$ was adjusted until a value was found (σ^2=2.1) which was appropriate to cause the observed distribution of temperatures and groundwater fluxes through the streambed. From 50 realizations of K-fields used for the simulations, 10 were selected which reproduced best the field observations.

Kalbus et al. (2008a,b) assumed in their simulations that the streambed had the same properties as the aquifer and thus they did not parameterize the streambed elements in the model differently from the aquifer elements. However, it is often presumed that streambed sediments are characterized by lower permeabilities due to the deposition of fine-grained sediment and organic matter (Sophocleus et al., 1995; Su et al., 2004), which could effect the distribution of fluxes across the streambed. Moreover, a heterogeneous streambed with a parameter distribution independent of the aquifer could lead to altered discharge patterns.

The objective of our study was to investigate the influence of the heterogeneity of both the aquifer and the streambed sediments on the spatial distribution of fluxes through the streambed. In numerical simulations we used different combinations of aquifer and streambed heterogeneity to evaluate which of these hydrological units has a stronger influence on the flux distribution. Focussing on spatial variations at given boundary conditions, we performed steady-state simulations since the subsurface properties are not expected to change over time. This study is a theoretical investigation of flow processes between aquifers and streams. However, we based the numerical model parameters on measured field data to obtain results in realistic orders of magnitude.

3 Methodology

Based on the study by Kalbus et al. (2008a,b), we used their model set-up and the 10 K-field realizations selected in their study as the base case for subsequent simulations. To evaluate the effect of streambed characteristics, we added different hydraulic conductivity scenarios for the streambed sediments to the model. The results were compared with the base case model results and the observed distribution of groundwater fluxes obtained by Schmidt et al. (2006) from mapped streambed temperatures.

3.1 Model set-up

A two-dimensional groundwater flow and heat transport model using the model code HEATFLOW (Molson et al., 1992; Molson and Frind, 2005) was set up according to the model used by Kalbus et al. (2008a,b). The conceptual model (Fig. 1) represents a vertical longitudinal profile along the streambed and within the underlying aquifer, and corresponds to the length of the investigated stream section and the saturated thickness of the aquifer (220 m×8 m). The upper 0.60 m hydrostratigraphic layer represents the streambed sediments. The model grid consists of 220×65 elements with a layer thickness varying from 0.20 m at the bottom to 0.05 m at the top. The system is assumed to be at steady state. The bottom and top boundaries are constant head boundaries, left and right boundaries are no-flow boundaries, leading to vertical flow through the system. Although the assumption of vertical groundwater discharge seems rigid for complex stream-aquifer systems it is commonly made for the interpretation of groundwater fluxes through the streambed (e.g., Cardenas and Wilson, 2007; Keery et al., 2007). The constant head values were chosen such that for each simulation the mean groundwater flux through the model equalled the mean flux through the streambed calculated from the observed temperature profiles ($q_z$mean=58.2 L m^{-2} d^{-1}). The temperature boundary conditions correspond to the mean stream water temperature during the mapping programme (18.4°C) at the top boundary and the constant deep groundwater temperature (10.9°C) at the bottom boundary. No energy flux is assumed across the left and right boundaries. The thermal transport properties were taken from the literature (thermal conductivity of the saturated sediments=2 J s^{-1} m^{-1} °C^{-1}; matrix specific heat=800 J kg^{-1} °C^{-1}; matrix density=2630 kg m^{-3}; specific heat of water=4174 J kg^{-1} °C^{-1}; density of water=1000 kg m^{-3}). A porosity of 0.25 was estimated from field data.

A heterogeneous distribution of hydraulic conductivity was achieved by including stochastically generated fields of hydraulic conductivity in the simulations. With the code FGEN (Robin et al., 1993), the K-fields were generated from the mean and variance of ln(K) and the correlation lengths in each direction (Table 1). These data were obtained from field observations of K, except the variance which was calibrated with the observed temperature distribution by Kalbus et al. (2008a,b). Ten realizations of the K distribution were used for the simulation of each of the scenarios explained below.

3.2 Scenarios

Base case: This is the case simulated by Kalbus et al. (2008a,b), for which the streambed was not parameterized differently from the aquifer; it was assumed to have

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the same properties as the aquifer. The base case will be used for comparison with the different streambed scenarios described in the following list.

Case A: The aquifer was assumed heterogeneous as in the base case, the streambed was assumed homogeneous with K two orders of magnitude less than the mean aquifer K.

Case B: To investigate the potential of the streambed sediment layer alone to cause a heterogeneous distribution of groundwater discharge, the aquifer was assumed homogeneous with the same mean K as in the base case and the streambed was assumed heterogeneous with the same statistical properties as the aquifer in the base case.

Case C represents a naturally developed streambed which basically consists of the same material as the underlying aquifer, but is assumed to have experienced some infilling by fine-grained sediments or organic matter. The aquifer and streambed were both assumed heterogeneous (using the same variance and correlation lengths as in the base case), while streambed infilling was simulated by dividing the K-value of each streambed element by 100. The streambed thus has the same degree of heterogeneity as the aquifer, but the mean K is two orders of magnitude less.

Case D: The streambed properties were assumed to be independent of the aquifer, which may occur for instance in streambeds with high sediment turnover rates. The connectivity between aquifer and streambed is low, which was achieved by generating new K-fields for the streambed layers only. As in Case C, the mean streambed K was chosen two orders of magnitude less than the mean aquifer K. The other statistical parameters for the K-field generation (variance, correlation lengths) were adopted from the aquifer statistics to enable a direct comparison with Case C.

The aquifer and streambed properties used for the generation of K-fields for the simulations of the base case and the four scenarios are summarized in Table 1.

4 Results and discussion

Homogeneous low-K streambeds (Case A) significantly dampen the groundwater fluxes compared to the base case scenario and result in a relatively uniform flux distribution close to the mean (Fig. 2A). The range of fluxes is much smaller than in the base case (Fig. 3). Homogeneous low-K streambeds thus serve as homogenizing layers which eliminate the influence of the aquifer texture. This case will never occur in reality, since all naturally developed streambeds as well as artificially constructed streambeds develop some degree of heterogeneity resulting from groundwater fluxes, sediment turnover, hyporheic fluxes, or activities of the interstitial and benthic fauna. Nevertheless, homogeneous low-K streambeds are often implemented in regional-scale groundwater flow models (e.g., McDonald and Harbaugh, 1988) and in the analysis of stream flow depletion through pumping (Chen et al., 2008) where the stream-aquifer interaction is governed by a conductance term representing the resistance of the streambed (Rushton, 2007). This approach may be sufficient for evaluating water budgets on a regional scale, but for a detailed analysis of flow and transport processes it may not be appropriate. For instance, in cases of contaminated groundwater discharging to a stream, maximum contaminant mass flow rates may be underestimated since areas of high groundwater discharge contribute more mass flow than low-discharge areas. Schmidt et al. (2008) also showed that a heterogeneous distribution of groundwater discharge strongly influences the timescales of contaminant release from a contaminated streambed. Hence, for small-scale investigations of stream-aquifer interactions, a representation of the streambed in flow models using a boundary condition with a uniform conductance term is not recommended. The streambed conductance should rather be resolved on a small scale to cover the range of high- and low-permeability zones in the streambed.

In Case B, a heterogeneous streambed on top of a homogeneous aquifer leads to a wider distribution of fluxes than in Case A (Fig. 2B), but the range is still much smaller than in the base case (Fig. 3). This case will also never occur in reality, since all aquifers show some degree of heterogeneity. Nevertheless, it shows that the streambed alone does not cause the observed distribution of fluxes. The passage through the streambed, which is much shorter compared to the passage through the aquifer, seems insufficient to cause highly diverse flow paths. Larger structures are necessary to direct the flow into highly permeable zones resulting in higher flow velocities.

A heterogeneous streambed with a mean K two orders of magnitude less than the mean K of the heterogeneous aquifer (Case C) shows a similar pattern of fluxes to the base case (Fig. 2C). The high- and low-discharge zones are at the same locations and the range of fluxes is similar to the range of the base case (Fig. 3). The maximum fluxes are even higher than those of the base case. This is a result of the larger gradient which had to be implemented in the models to achieve the reach-average flux of $58.2\,\mathrm{L\,m^{-2}\,d^{-1}}$ (average hydraulic gradient=0.01; base case: 0.002). Within high-permeability zones, this higher gradient leads to increased fluxes compared to the base case with a lower gradient. When reaching the streambed, the short passage through the less permeable streambed does not have much influence on the flow velocities in these zones since the permeability is still higher than in the neighbouring low-discharge zones.

In case of an independent heterogeneity of the streambed (Case D), the pattern is still similar to that of a related heterogeneity as in Case C, but the locations of high- and low-discharge zones have been slightly displaced, some peaks have disappeared, while other peaks have developed (Fig. 2D). The range of fluxes is almost identical with the range of the base case (Fig. 3). Again, the higher gradient leads to increased flow velocities through the high-permeability zones of the aquifer. As opposed to Case C, however, groundwater flow from high-K zones within the aquifer may now intersect low-permeability zones in the streambed and will thus be diverted to neighbouring zones with higher permeabilities. This attenuates some of the peak flows observed in Case C and creates new peaks at other locations.

Comparing the mean (solid line) and median (dashed line) in Fig. 3, it becomes ap-

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parent that greater spatial heterogeneity mainly leads to an increase in the proportion of high fluxes. Because we assumed vertical flow through the model domain, the fluxes cannot become less than zero, but well connected high-permeability zones can lead to very high fluxes which are concentrated in small areas. This is even more evident from Fig. 4, which shows the relative contribution of the streambed area to the cumulative flux. In Cases A and B, the band representing the range between maximum and minimum fluxes of all K-field realizations is narrow and almost straight with a slope of 1:1. In these cases, a certain proportion of streambed area thus contributes a similar proportion of cumulative flux. For instance, 20% of the streambed area contributes 22% (Case A) to 30–33% (Case B) of the cumulative flux. In Cases C and D, a much smaller proportion of streambed area contributes a larger proportion of cumulative flux. For instance, in Case C, 20% of the streambed area contributes 50–74% of the cumulative flux along the modelled reach. The band is much wider in Cases C and D, indicating considerable variation between the different K-field realizations.

5 Conclusions

Previous simulations of groundwater flow and heat transport through a streambed have revealed that strong spatial variations in groundwater discharge to a stream are caused by a heterogeneous distribution of aquifer hydraulic conductivity. The influence of the streambed on the distribution of fluxes was investigated in subsequent simulations with different scenarios of aquifer and streambed hydraulic conductivity. The aquifer was found to have a stronger influence on the spatial distribution of fluxes than the streambed. However, the implementation of a homogeneous low-K streambed within a heterogeneous aquifer caused a significant homogenization of the fluxes. This behaviour should be considered when using the concept of streambed conductance in regional-scale groundwater models. A heterogeneous distribution of hydraulic conductivity only in the streambed was not sufficient to cause strong flux variations. Simulation results with heterogeneous low-K streambeds were similar to the results from the model without a distinction between aquifer and streambed properties. Thus, if streambed infilling has to be considered in a model, which leads to a reduced permeability of the streambed sediments compared to the aquifer, it is recommended to implement a heterogeneous distribution of streambed hydraulic conductivity to avoid underestimating peak flows. These results also confirm the applicability of the methodology proposed by Kalbus et al. (2008a,b) to use measured streambed temperatures for calibration of aquifer properties even without distinguishing between the aquifer and streambed.

Observed distributions of groundwater fluxes through the streambed may often be a result of both aquifer and streambed heterogeneity, with the aquifer having a stronger influence. Numerical model predictions of groundwater flow and solute transport may thus significantly benefit from heterogeneous distributions of aquifer and streambed properties.

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Table 1. Aquifer and streambed properties of all simulation cases. K=hydraulic conductivity, σ^2=variance of ln(K), λ_x and λ_z=correlation lengths in the x- and z-directions.

Scenario	Aquifer properties	Streambed properties
Base Case	Heterogeneous K_{mean}=2.1E-04 m s^{-1} σ^2=2.1 λ_x=6.0 m λ_z=1.5 m	Same as aquifer
Case A	As in base case	Homogeneous K=2.1E-06 m s^{-1}
Case B	Homogeneous K=2.1E-04 m s^{-1}	As in base case aquifer
Case C	As in base case	As in base case but each streambed element K divided by 100 (K_{mean}=2.1E-06 m s^{-1})
Case D	As in base case	Heterogeneous, independent K-fields with K_{mean}=2.1E-06 m s^{-1} σ^2=2.1 λ_x=6.0 m λ_z=1.5 m

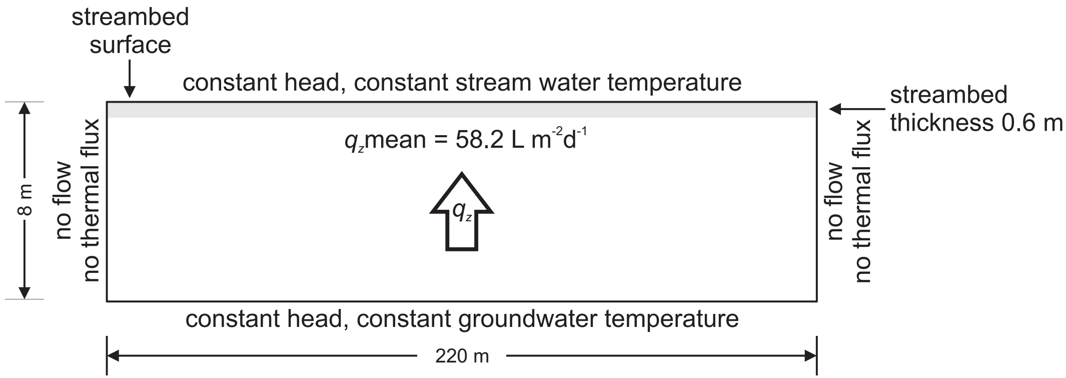

Fig. 1. Model definition and boundary conditions.

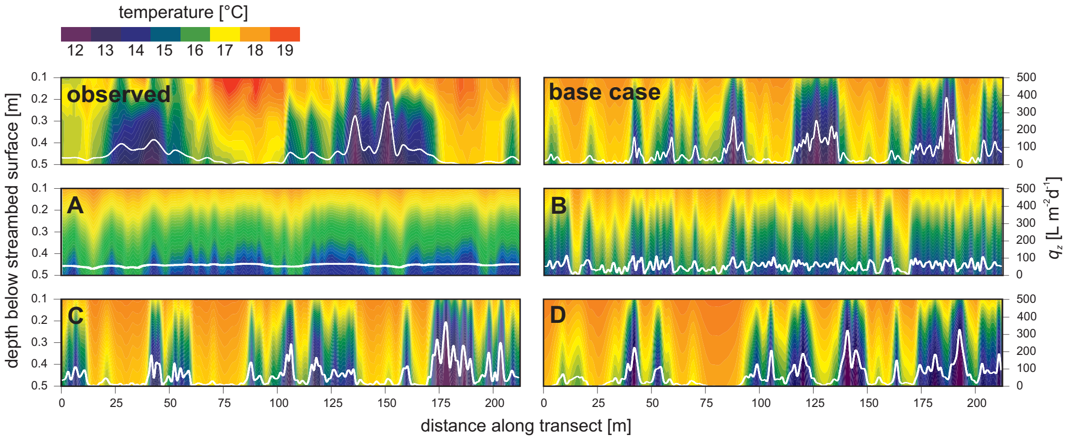

Fig. 2. Observed (top left; after Schmidt et al., 2006) and simulated (base case and Cases A–D) results showing temperature (colour maps) and flux distributions (white curves) in the streambed (represented by the upper grey zone in Fig. 1). Temperature data are shown at streambed depths between 0.1 and 0.5 m corresponding to the observations. Simulated results are shown from one example out of ten K-field realizations (the same realization is shown in all scenarios). Vertical exaggeration is approx. 100×.

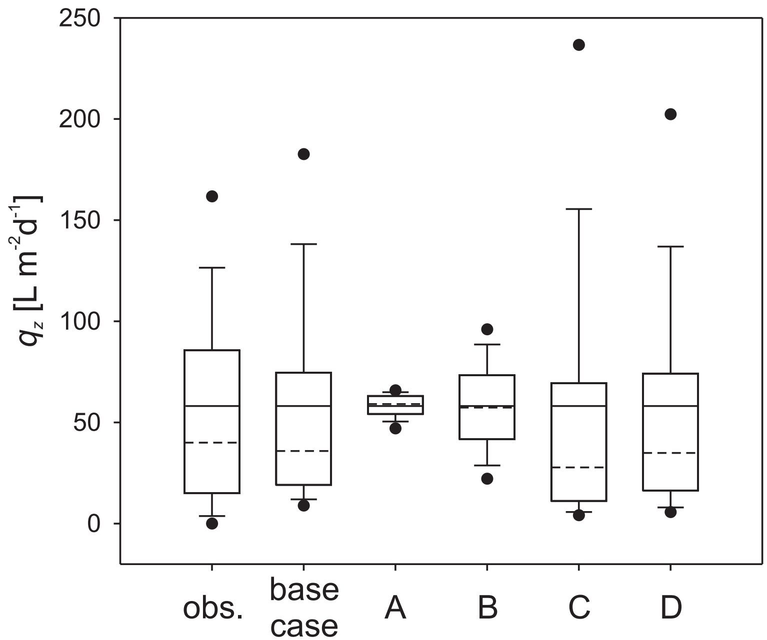

Fig. 3. Box plots of the groundwater discharge through the streambed showing 95th and 5th percentile (dots), 90th and 10th percentile (error bars), 75th and 25th percentile (box), arithmetic mean (solid line), and median (dashed line). Observed data are complete data of the mapping programme (n=140), simulated data are the complete data set from all 10 realizations (n=2200) for each case.

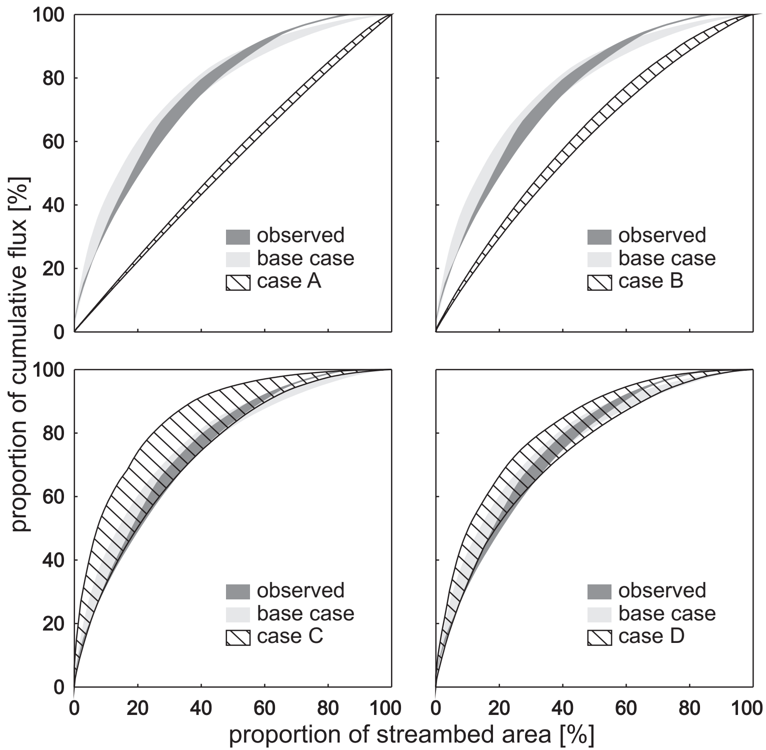

Fig. 4. Distribution of groundwater fluxes through the streambed in relation to the streambed area. Bands show the full range between maximum and minimum values of observations and modelling results, respectively.