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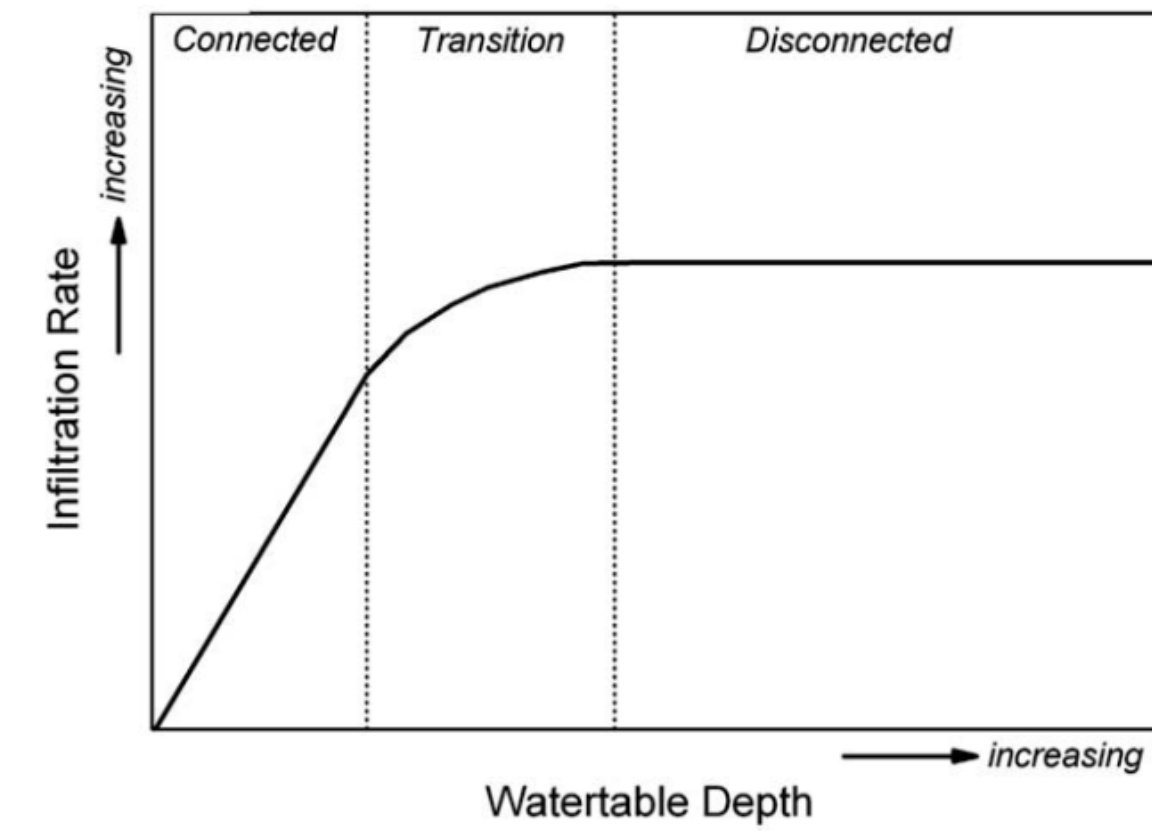
# Experimental and numerical assessment of transient stream–aquifer exchange during disconnection

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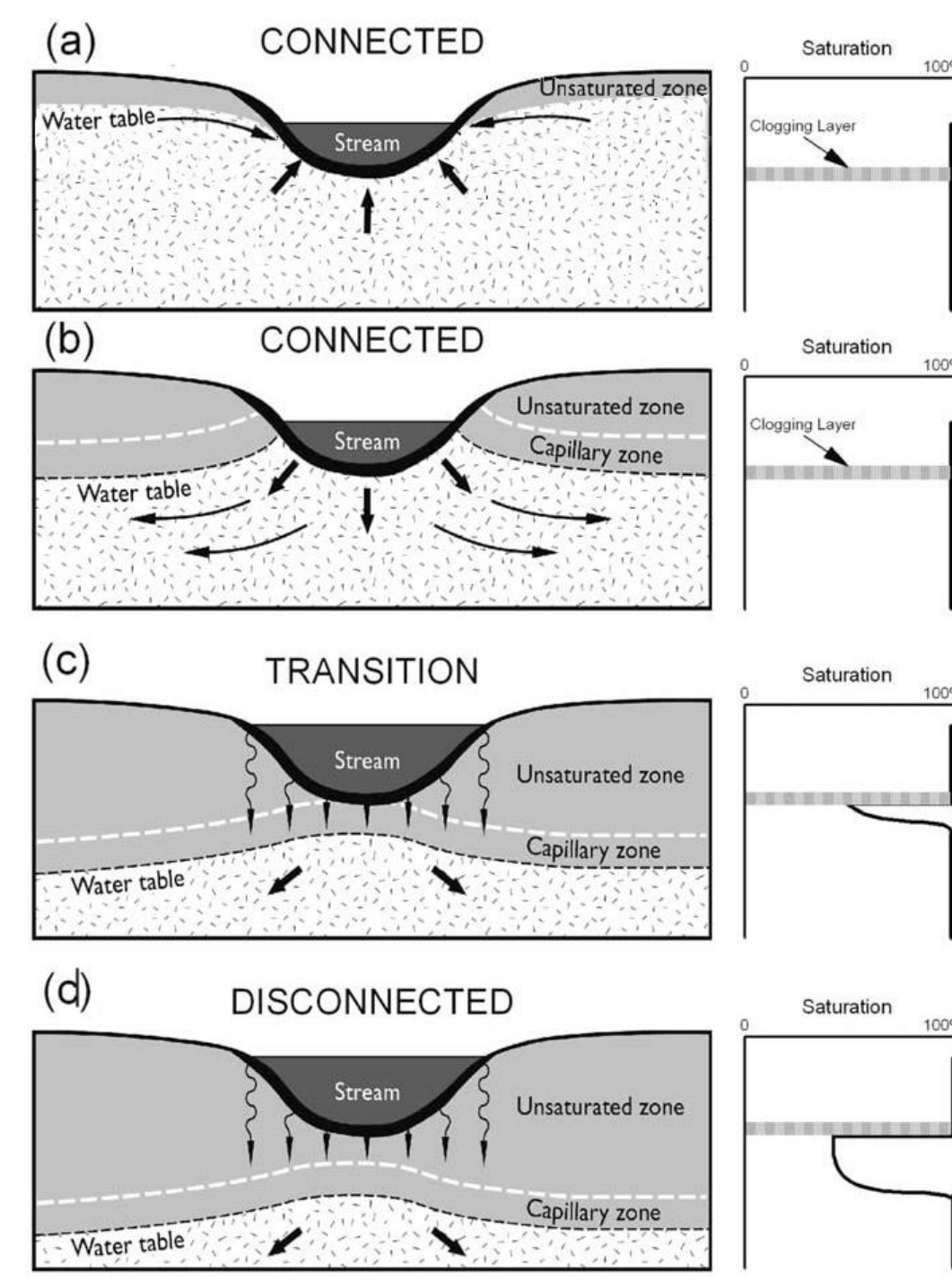
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## Motivations

- Stream–aquifer systems may fluctuate from a hydraulically connected to a disconnected state
- Natural dry conditions or intensive pumping can cause the development of an unsaturated zone below the stream bed
- Evolution of the infiltration flow between the river and groundwater :



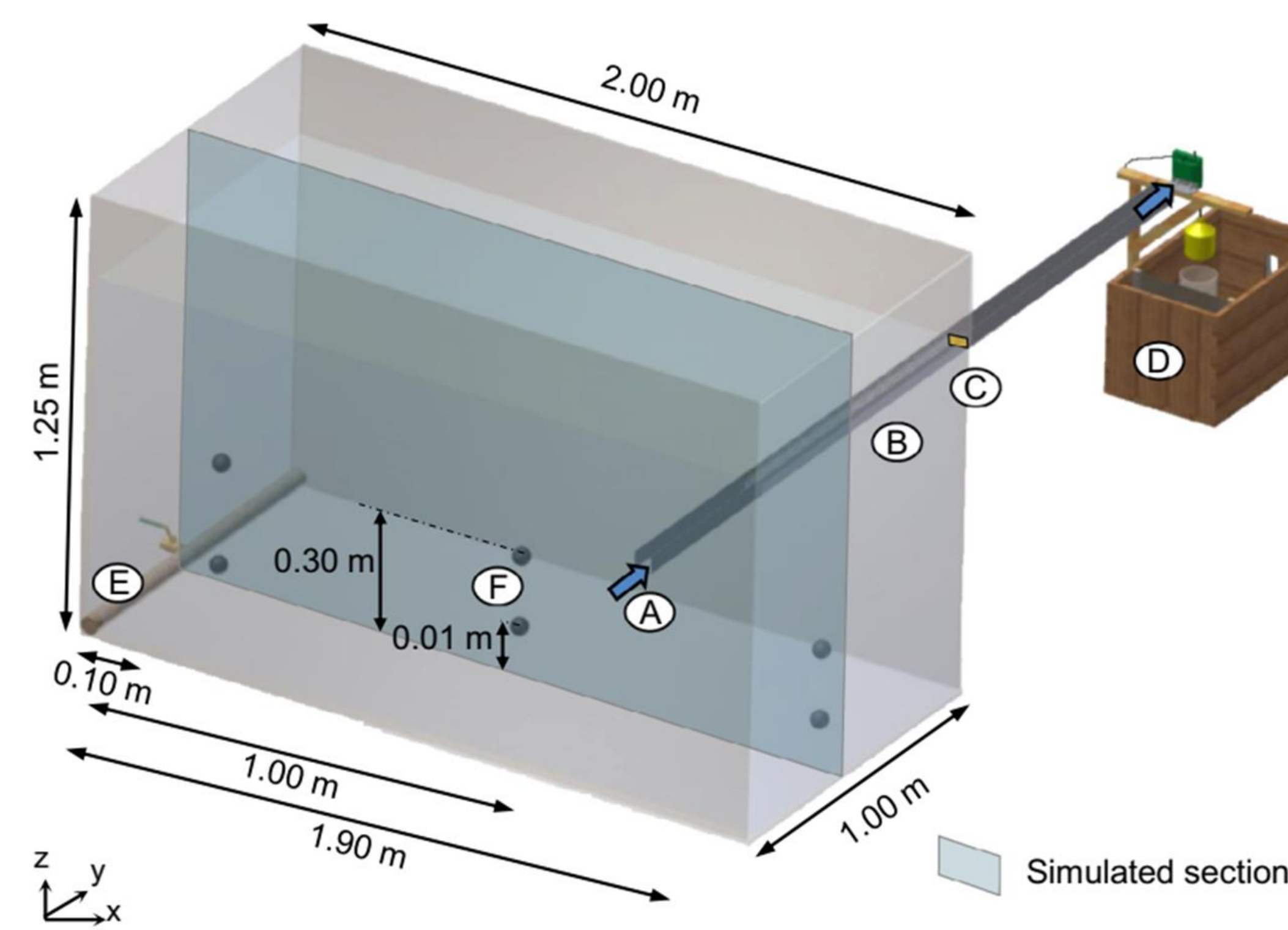
Brunner et al., 2009



Brunner et al., 2009

- The main motivation of this study is to obtain experimental data in order to investigate the transient behavior of the exchange between river and aquifer, in the light of previous numerical findings.
- The transition period between connected and disconnected regimes are specifically examined

## Laboratory experiments



### Experimental design of the sand box:

- (A) Input flow:** The input flow of the stream was controlled by electromagnetic flowmeters (MIK L303, KOBOLD) of 3.2–63 L h<sup>-1</sup> flow range and a ±2% full-scale accuracy,
- (B) Stream:** it consists of a rectangular section aluminum tube (2.5 m long, 5 cm wide and 1 mm thick). The slope of the channel is 9%. It is subdivided into three parts: a 0.5 m long upstream section designed to obtain a laminar flow, a one-metre long hole-punched central section, in contact with the aquifer and a one-metre long downstream section,
- (C) Overflow:** the water level in the stream was prescribed by a 8.5 mm high overflow located downstream, resulting in a 1.3 cm water level in the middle of the stream,
- (D) Spillway:** The spillway water level was measured by a shaft encoder level sensor (SE 200, OTT), the accuracy of which is 1 L h<sup>-1</sup>.
- (E) Drain:** at the bottom of the tank, a 5 cm diameter drain (Drain power FILTREX) was used to lower the water table. It is manufactured using pellets of resin. Its hydraulic permeability is close to 1 m s<sup>-1</sup>,
- (F) Pore water pressure sensors** (SSK, model P306AV) at two different elevations (0.01– 0.3 m) above the bottom.
- The rate of stream infiltration was estimated by computing the difference between input and output stream flows.

## Mathematical model : Ginette

- Surface flow (1 dimension) :** Manning-Strickler 's equation (Manning, 1890)

$$Q_{river} = K_{St} S R^{2/3} i^{1/2}$$

$Q$ : the stream discharge (m<sup>3</sup> s<sup>-1</sup>),  
 $K_{St}$ : the Strickler roughness coefficient (m<sup>1/3</sup> s<sup>-1</sup>),  
 $A$ : the cross-sectional zone of flow (m<sup>2</sup>),  
 $i$ : the channel slope,  
 $R_h = A/P_w$ : the hydraulic radius (m), where  $P_w$  is the wetted perimeter (m).

- Groundwater flow (2 dimensions) in the (x,z) plane:** Richards' Equation (de Marsily, 1989)

$$\frac{\partial}{\partial x} \left[ \frac{\rho k_x k_r(S_w)}{\mu} \frac{\partial p}{\partial x} \right] + \frac{\partial}{\partial z} \left[ \frac{\rho k_z k_r(S_w)}{\mu} \frac{\partial p}{\partial z} + \rho g \right] = \rho S_w \frac{\partial S_w}{\partial p} \frac{\partial p}{\partial t} + \rho q$$

$p$ : the pore water pressure (Pa),  
 $S_w$ : the water saturation,  
 $q$ : a volumetric water source or sink (s<sup>-1</sup>),  
 $\omega$ : the total porosity,  
 $\rho$ : the density of water (kg m<sup>-3</sup>),  
 $t$ : the time (s),  
 $k_x$  and  $k_z$ : the intrinsic permeabilities along the x and z axes, respectively (m<sup>2</sup>),  
 $k_r(S_w)$ : the relative unsaturated hydraulic conductivity,  
 $g$ : the acceleration due to gravity (m s<sup>-2</sup>),  
 $z$ : the vertical coordinate taken as positive upward (m),  
 $\eta$ : the dynamic viscosity of water (Pa s).

- Van Genuchten (VG) model (1980) and Mualem formulation(1975)**

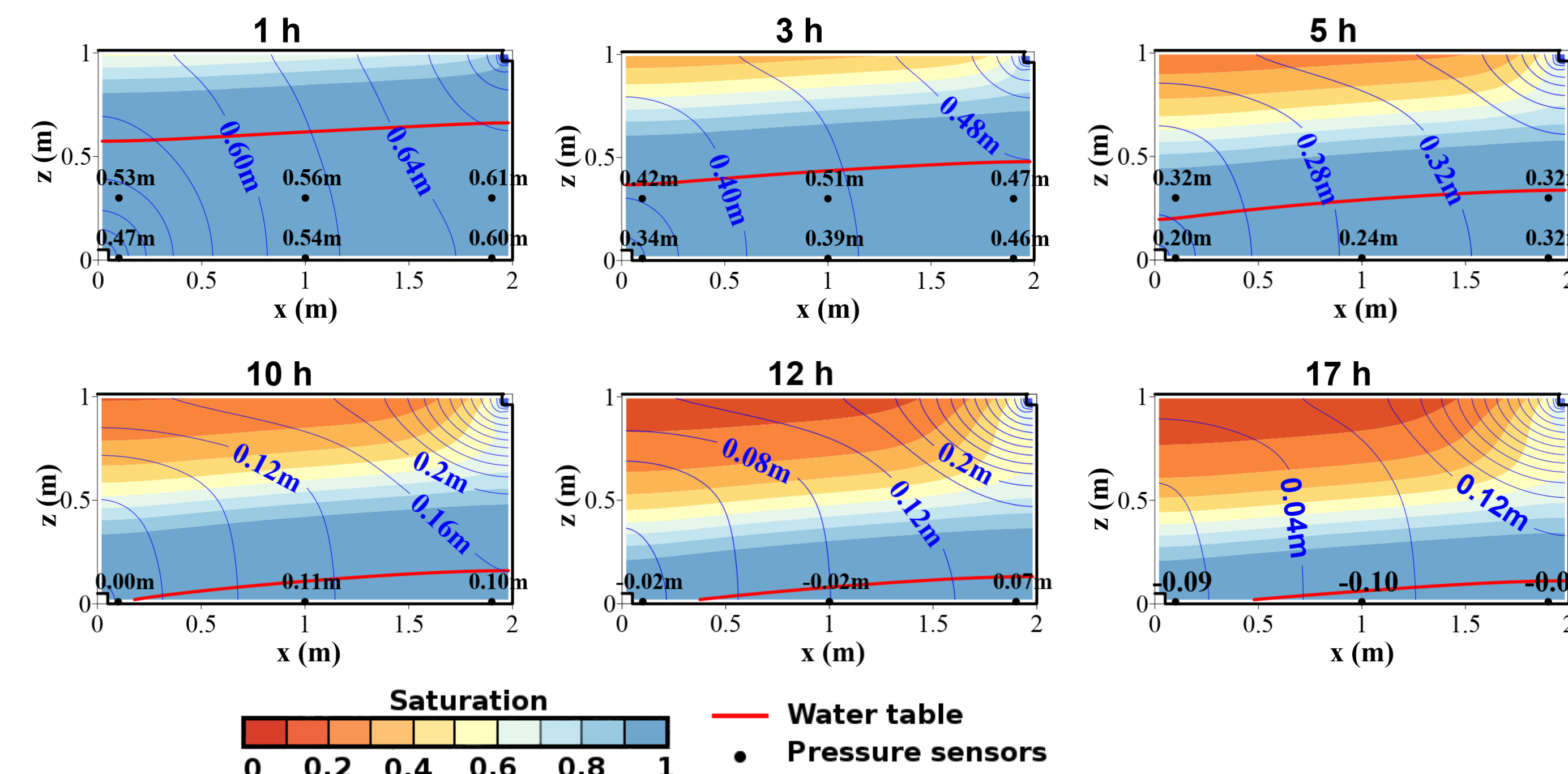
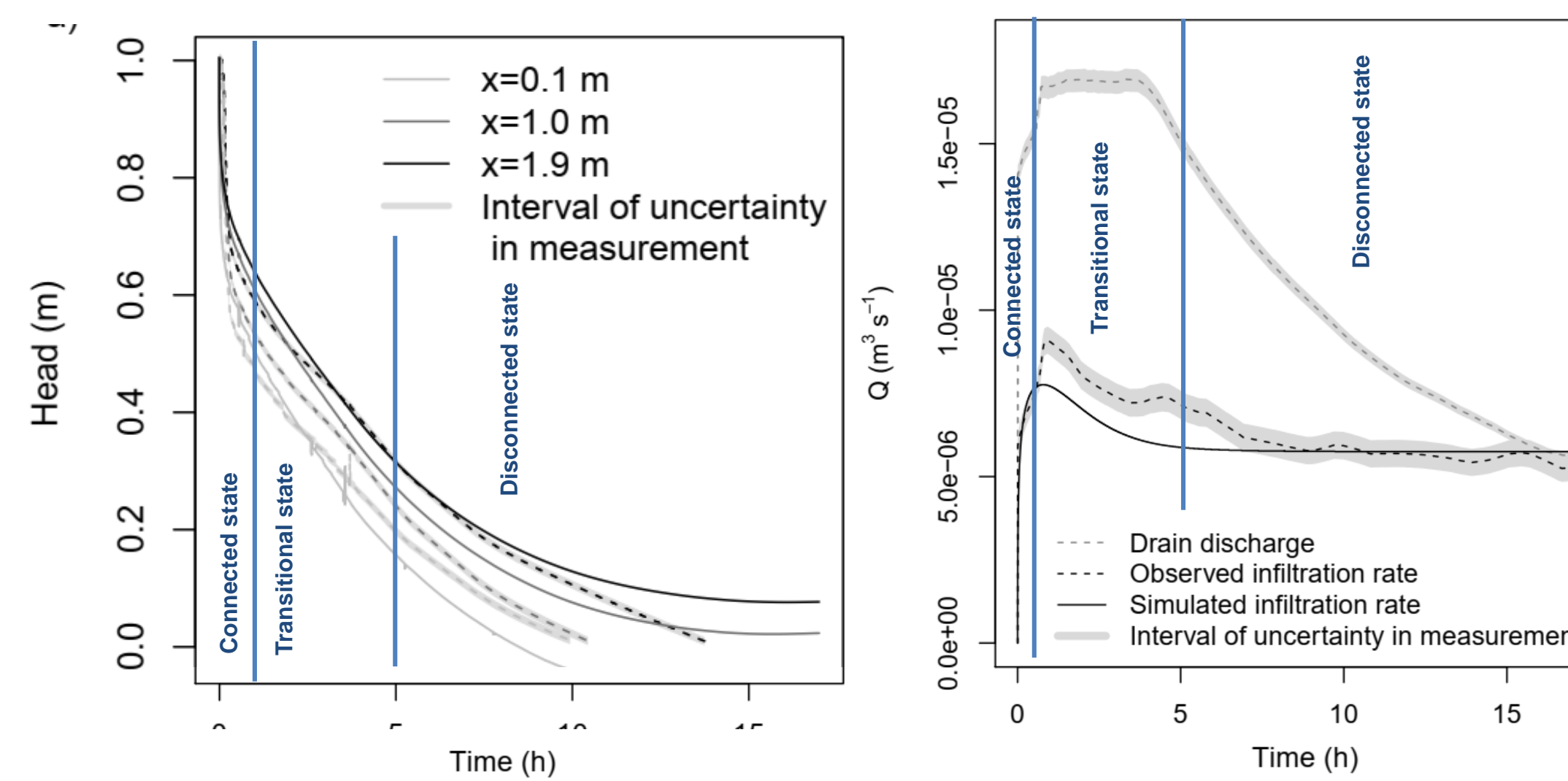
$$S_e = \frac{S_w - S_{wres}}{1 - S_{wres}} = (1 + (-\alpha P_c)^n)^{-m} \quad k_r(S_e) = S_e^{\frac{1}{2}} \left( 1 - \left( 1 - S_e^{\frac{n}{n-1}} \right)^m \right)^2$$

$S_e$ : the effective saturation,  
 $S_{wres}$ : the residual soil water saturation,  
 $P_c$ : the capillary pressure (Pa),  
 $\alpha$  (Pa<sup>-1</sup>),  $n$  and  $m$  are van Genuchten's parameters, with  $m = 1-1/n$ .

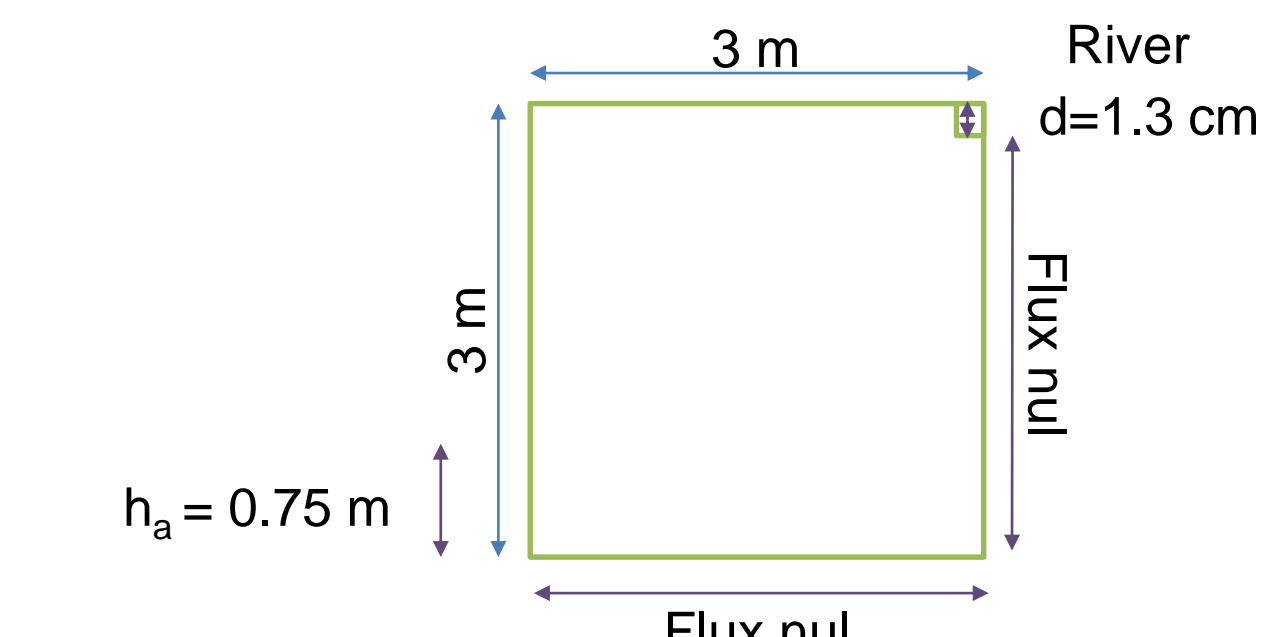
## Numerical modeling of the experiment

Parameters	A priori values	Calibrated values
$K_s$ (m.s <sup>-1</sup> )	$2.2 \cdot 10^{-4}$	$2.56 \cdot 10^{-4}$
$K_{sb}$ (m.s <sup>-1</sup> )	$10^{-7}$	$3.00 \cdot 10^{-7}$
$\omega$	0.38	0.38
$\alpha$ (m <sup>-1</sup> )	$2.6 \pm 0.15$	2.64
$n$	$3.9 \pm 0.6$	3.38
$S_{wres}$	0.02-0.13	0.02

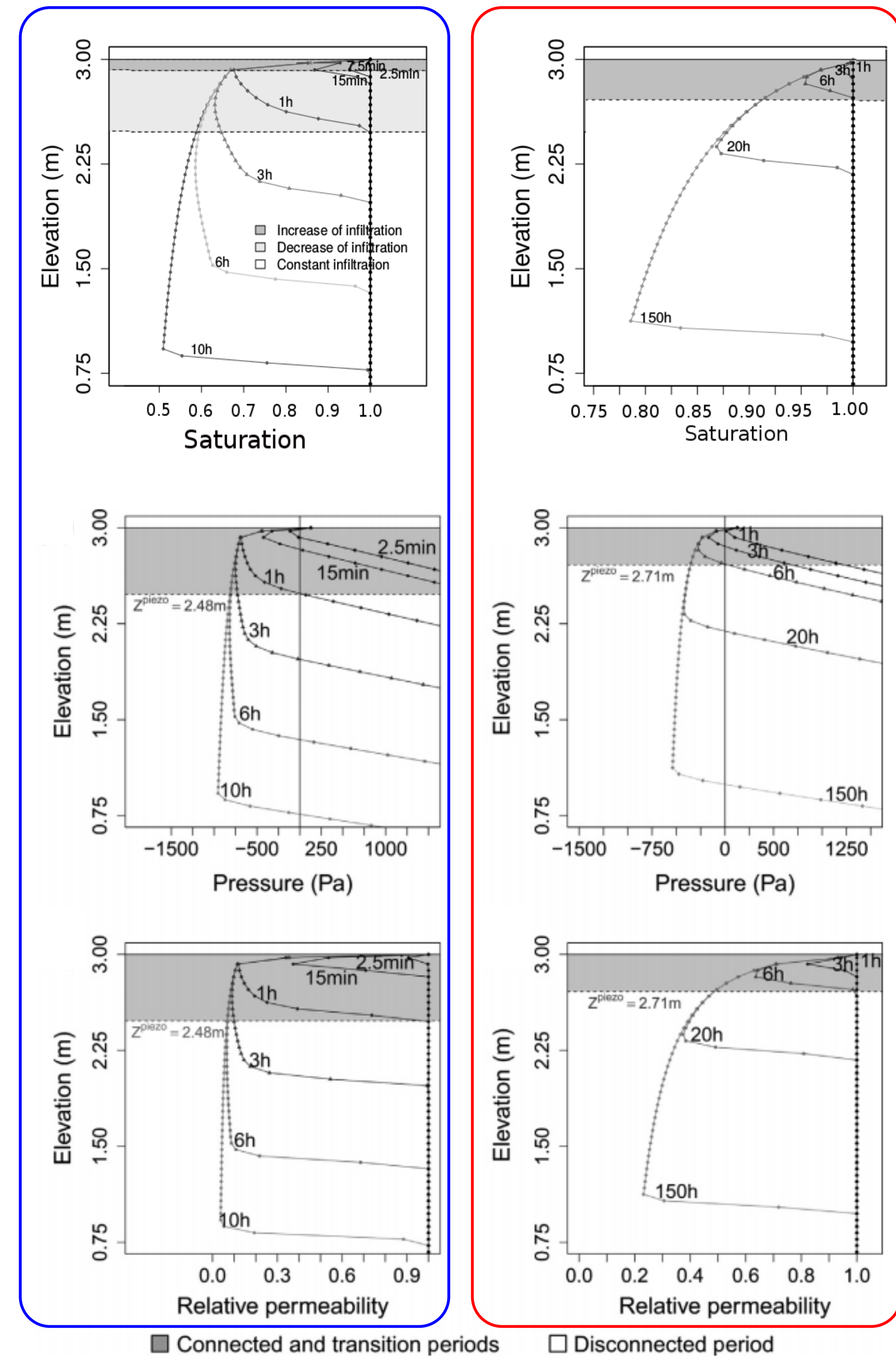
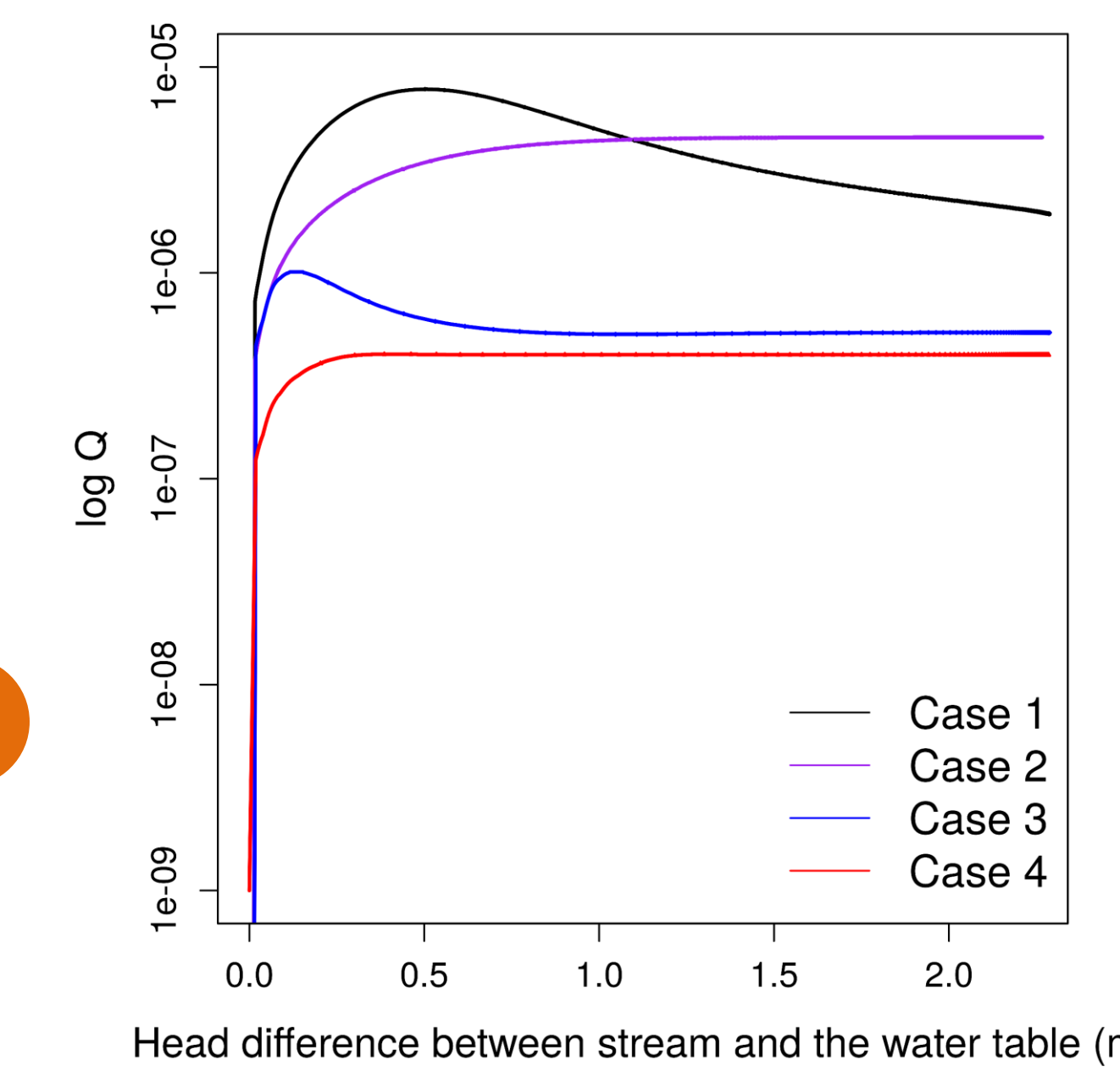
Experimental values  
Values of Soeiro (1964)  
Values of Allègre et al. (2010)



## Synthetic cases



Parameters	Case 1	Case 2	Case 3	Case 4
$K_s$ (m.s <sup>-1</sup> )	$2.56 \cdot 10^{-4}$	$2.56 \cdot 10^{-5}$	$5.79 \cdot 10^{-5}$	$5.79 \cdot 10^{-6}$
$K_{sb}$ (m.s <sup>-1</sup> )	$3 \cdot 10^{-7}$		$2.2 \cdot 10^{-7}$	
$\omega$	0.38		0.43	
$\alpha$ (m <sup>-1</sup> )	2.64		14.5	
$n$	3.38		2.68	
$S_{wres}$	0.02		0.1	



- For small values of  $n$ , a highly saturated zone is maintained below the stream bed and the maximum infiltration rate decreases. Since  $n$  describes the homogeneity of the pore size distribution, the infiltration ought to be reduced at lower values of  $n$  (more abundant small pore size).
- The parameter  $\alpha$  provides a gauging of the height of the capillary fringe, which is approximated by  $1/\alpha$ . High values of  $\alpha$  yield low values of the capillary fringe. The higher the capillary fringe height (small values of  $\alpha$ ), the higher is the occurrence of large vertical gradients of saturation for large aquifer hydraulic conductivity values. In this case, a maximum value of the infiltration rate can be obtained during the transitional stage.
- Two types of behavior were obtained. Cases 2 and 4 with lower hydraulic conductivities of the aquifer yield a curve shape similar to that obtained by Brunner et al. (2009) for steady state infiltration rate whereas cases 1 and 3 with higher hydraulic conductivities yield the same curve shape obtained in our experimental study.
- They differ by their transitional regime, which may be explained by the drainage rate leading to a rapid decline of the water table and a stratified saturation profile below the stream bed.
- The rapidity of the drawdown due to the drainage can be described using the analytical expression of the approximate time constant for an unconfined aquifer  $\tau$ , proposed by Reilly and Harbaugh (2004), which is:
 
$$\tau = \frac{\omega_d L^2}{T}$$

$\omega_d$ : the specific yield,  
 $L$ : the characteristic length of the system (m),  
 $T$ : the transmissivity (m s<sup>-2</sup>).
- Below a threshold value  $\tau_{th}$  of about 2 days, a maximum infiltration rate is obtained during transition whereas above this threshold, the plateau is the maximum value. Such an alternative behavior for the transitional stage can be of great interest in the case of a pumping near a river in transient conditions. Indeed, the equivalent hydraulic radius  $R$  of a steady-state solution for a pumping well is given by:
 
$$R = 1.5 \sqrt{T/\omega_d t}$$
- Taking a maximum value for  $T$  of  $10^{-1}$  m<sup>2</sup> s<sup>-1</sup>,  $\omega_d = 0.2$  and  $t = \tau_{th}$  yields a  $R$  value of about 500 m. This suggests that a maximum infiltration rate in the transitional phase may occur for pumping wells at a distance from the stream of less than 500 m.

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