

Quantifying peakflow attenuation/amplification in a karst river using the diffusive wave model with lateral flow

Jean-Baptiste Charlier, Roger Moussa, Pierre-Yann David, Jean-François Desprats

▶ To cite this version:

Jean-Baptiste Charlier, Roger Moussa, Pierre-Yann David, Jean-François Desprats. Quantifying peak-flow attenuation/amplification in a karst river using the diffusive wave model with lateral flow. Hydrological Processes, 2019, 33 (17), pp.2337-2354. 10.1002/hyp.13472 hal-02273711

HAL Id: hal-02273711 https://brgm.hal.science/hal-02273711

Submitted on 29 Aug 2019

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers. L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 Quantifying peakflow attenuation/amplification in a karst river using

2 the diffusive wave model with lateral flow

Jean-Baptiste Charlier^{*,1}, Roger Moussa², Pierre-Yann David³, Jean-François Desprats¹ ¹ BRGM, Univ. Montpellier, Montpellier, France. ² LISAH, Univ. Montpellier, INRA, IRD, Montpellier SupAgro, Montpellier, France ³ BRGM, F-76130 Mont Saint-Aignan, France. * corresponding author: e-mail: j.charlier@brgm.fr; Tel: +33 (0)4 67 15 79 77 Key words Flood; Lateral flow; Karst, Diffusive wave model; Peakflow attenuation or amplification; Surface water / groundwater interactions

21 Abstract

22 The aim of this paper is to quantify peakflow attenuation and/or amplification in a river, 23 investigating lateral flow from the intermediate catchment during floods. This is a challenge 24 for the study of the hydrological response of permeable/intermittent streams, and our 25 contribution refers to a modelling framework based on the inverse problem for the diffusive 26 wave model applied in a karst catchment. Knowing the upstream and downstream 27 hydrographs on a reach between two stations, we can model the lateral one, given information 28 on the hydrological processes involved in the intermediate catchment. The model is applied to 29 33 flood events in the karst reach of the Iton River in French Normandy where peakflow 30 attenuation is observed. The monitored zone consists of a succession of losing and gaining 31 reaches controlled by strong surface-water / groundwater (SW/GW) interactions. Our results show that, despite a high baseflow increase in the reach, peakflow is attenuated. Model 32 33 application shows that the intensity of lateral outflow for the flood component is linked to 34 upstream discharge. A combination of river loss and overbank flow for highest floods is proposed for explaining the relationships. Our approach differentiates the role of outflow 35 36 (river loss and overbank flow) and that of wave diffusion on peakflow attenuation. Based on 37 several sets of model parameterization, diffusion is the main attenuation process for most cases, despite high river losses of up to several m³/s (half of peakflow for some 38 39 parameterization strategies). Finally, this framework gives new insight into the SW/GW 40 interactions during floods in karst basins, and more globally in basins characterized by 41 disconnected river-aquifer systems.

42

43 **1** Introduction

44 In a catchment densely monitored for hydrological surveying, flood-flow generation can be 45 investigated at the network scale in reaches between two gauging stations. In such a runoff-46 runoff approach, flows from the intermediate catchment can be significant contributors of 47 flood flow at the outlet, depending on catchment geometry and hydrological conditions. 48 Although lateral flow from the intermediate catchment is well documented for low-water 49 periods, investigating for instance surface-water / groundwater (SW/GW) interactions in 50 successive reaches (Covino et al., 2011; Mallard et al., 2014), there is a lack of information on 51 how to define lateral flow during flood events.

52 As shown on Figure 1, several types of lateral flow occur during floods, depending upon 53 catchment descriptors such as climate, relief, geology, soil, etc. For highest peakflows, lateral 54 outflow (Figure 1a) is generally considered in the case of overbank flow when waters from 55 the river flow to the flood plain, without been returned to the river during the flood 56 (Jothityangkoon & Sivapalan, 2003; Moussa and Bocquillon, 2009), modifying the shape of 57 the hydrograph (Rak et al., 2016; Fleischmann et al., 2018). Another case favouring lateral 58 outflow is specific to permeable basins, where river losses infiltrate and recharge the 59 underlying aquifer (Sorman et al., 1997; Charlier et al., 2015b; Dvory et al., 2018). This is notably the case in arid/Mediterranean environment, where the importance of infiltrating 60 floodwater for aquifer recharge has been highlighted in disconnected river- aquifer systems 61 62 (Hughes & Sami, 1992; Camarasa Belmonte & Segura Beltrán, 2001; Lange, 2005; Dahan et 63 al., 2007; Vázquez- Suñé et al., 2007). Lateral inflow (Figure 1b) is often taken into account 64 in flood modelling due to its common occurrence in all types of catchments (e.g., Cimorelli et 65 al., 2014;2018). Most inflow comes from tributaries (point lateral flow) or from hillslope runoff (distributed lateral flow). In permeable basins, groundwater flooding can also cause 66 67 large lateral inflows (Finch et al., 2004; Pinault et al. 2005; Naughton et al., 2012), but this

simple typology can mask a greater complexity of lateral flows. In most cases, several
processes are spatially and temporally combined, leading to mixed (and maybe compensated)
out- and in-flows during flooding, as shown on Figure 1c.

71 Investigating flood generation in a carbonate catchment is of great interest because it is highly 72 controlled by SW/GW interactions. Following Robins and Finch (2012), a true groundwater 73 flood-where the groundwater level rises above ground surface-is distinguished from a 74 groundwater-induced flood-which occurs as an intense groundwater discharge through springs and permeable shallow horizons into surface waters. These types of flooding notably 75 76 occur in chalk terrains of Northern Europe (Finch et al., 2004; Pinault et al., 2005; Hughes et 77 al., 2011; Morris et al., 2015; Thiéry et al. 2018), where extreme events were caused by 78 exceptionally high groundwater levels in 2000-2001. Flood generation in these cases was 79 driven by saturation of the matrix porosity (Price et al., 2000), and is associated to long-80 duration floods due to the aquifer inertia.

81 When carbonate formations are highly karstified, which can be the case in chalk formations, 82 groundwaters contribute to the fast-flow component in streams (Maréchal et al., 2008; 83 Charlier et al., 2015b), playing a significant role in flooding, even for flash floods. In karst 84 basins, flooding may also occur by the cessation of rainfall infiltration because of the small 85 retention capacity of a karst massif (Maréchal et al. 2008; Fleury et al. 2013), or because the 86 infiltration capacity of the underground conduit network is exceeded (Lopez-Chicano et al. 87 2002; Bonacci et al. 2006; Bailly-Comte et al. 2009). As chalk aquifers are often karstified, 88 specific types of groundwater flooding can occur seasonally, as is reported from Ireland in 89 low-lying topographic depressions, known as turloughs, that are fed by underground flow 90 from karst aquifers (Naughton et al., 2012; Gill et al., 2013).

Another frequent feature of flood generation in karst basins is a contrasted evolution of
peakflow, i.e. amplification or attenuation, in a same reach depending upon the prior aquifer

93 saturation level (Maréchal et al., 2008; Charlier et al., 2015b). In catchments characterized by 94 a high storage potential in thick unsaturated zones of karst aquifers, peakflow attenuation is 95 common (Jourde et al., 2013; Ladouche et al., 2014; Brunet & Bouvier, 2017), which may add 96 to the attenuation due to flood-wave diffusion in the channel. Attenuation is also enhanced in 97 medium and large catchments, where diffusion is favoured due to the development of 98 drainage networks in lowland areas (e.g., Moussa & Cheviron, 2015; Trigg et al., 2009) and 99 due to the occurrence of overbank flow (see a synthesis in Bates & De Roo, 2000; Hunter et 100 al., 2007; Moussa & Bocquillon, 2009). In karst massifs, the drainage is often developed in 101 canyons that cross-cut the carbonate plateau, where the channel commonly is rougher. We can 102 suppose that this will favour a velocity decrease of the flood wave due its meandering feature. 103 Finally, there is a need in such a context to understand the respective roles of lateral outflow 104 (losses, overbank flow) and flood routing in peakflow attenuation.

105 Many modelling methods exist for investigating SW/GW interactions on groundwater 106 flooding. As the hydrogeological context of chalk aquifers can be favourable for applying 107 physically-based models, lumped approaches (Pinault et al., 2005; Upton & Jackson, 2011) 108 were completed by distributed models, such as MODCOU (Korkmaz et al., 2009), SIM 109 (Habets et al., 2010), MIKE SHE (House et al., 2016), or MARTHE (Thiéry et al., 2018). 110 However, the specifics of karst basins with their high degree of complexity prevent an 111 efficient application of such distributed models, and few such modelling approaches have 112 been used for characterizing SW/GW interactions during flood events in karst basins. To our 113 knowledge, the only published work on the application of flood-routing models to channel 114 reaches, is based on the simplified Saint-Venant equation that describes unsteady flow in 115 partially filled channels; a first modelling approach for karst areas was proposed by Bailly-116 Comte et al. (2012), using the Kinematic Wave Equation coupled with a linear underground 117 reservoir for simulating lateral inflow. Recently, Charlier et al. (2015b) tested the relevance of 118 the Diffusive Wave Equation (DWE) for assessing lateral in- and outflows in karst rivers. 119 DWE is more adapted to floods with large wavelengths, common in medium and large 120 catchments (Ponce, 1990). Moreover, Moussa (1996) extended the analytical solution of the 121 DWE under the Hayami (1951) hypothesis (Moussa & Bocquillon, 1996b) to the case of 122 uniformly distributed lateral flow. The model can calculate the temporal distribution of lateral 123 flow in a river reach by an inverse problem approach, using as input the flow in both the 124 upstream and downstream gauging stations. The solution of the inverse problem proposed is 125 part of the hydrological MHYDAS model (Distributed Hydrological Modelling of 126 AgroSystems; Moussa et al., 2002). Using this inverse problem for the DWE we can simulate 127 the global dynamics of lateral flow during floods, which provides information on the 128 hydrological processes involved, such as tracking loss and gain reaches in rivers (Charlier et 129 al., 2015b), or characterizing matrix/conduit relationships through the underground network 130 of a karst aquifer (Cholet et al., 2017); both these publications showed promises in the ability 131 of such models to quantify lateral flow in a karst basin.

132 The aim of this paper is therefore to quantify peakflow attenuation/amplification in a river, 133 and to investigate lateral flow during floods. For that, we propose a modelling framework to 134 simulate lateral flow from an intermediate catchment, using the inverse problem for the DWE. 135 The model is applied to 33 flood events in the karst reach of the Iton River in French 136 Normandy, where peakflow is attenuated. After analysing the role of groundwater on flood 137 generation, the model is used for quantifying the peakflow attenuation. The respective roles of 138 outflow and wave diffusion on peakflow attenuation are described according to various 139 parameterization strategies. Investigating the sensitivity of model parameterization, we 140 provide a new way for better understanding and quantifying flood routing through accounting 141 for lateral flow.

142 **2** Modelling approach

143 **2.1** The diffusive wave model

In order to model 1-D unsteady flow in open channels, the Saint-Venant equations can be
simplified under some assumptions (acceleration terms neglected), leading to the Diffusive
Wave Equation (DWE) (Moussa & Bocquillon, 1996a), which corresponds to the case of long
wave-length flood events generally observed in medium and large basins:

148 Eq. 1
$$\frac{\partial Q}{\partial t} + C \frac{\partial Q}{\partial x} - D \frac{\partial^2 Q}{\partial x^2} = 0$$

149 where x [L] is the length along the channel, t [T]) is the time, and the celerity C [L.T⁻¹] and 150 the diffusivity D [L² T⁻¹] are functions of the discharge Q [L³ T⁻¹]. In this case, we have a 151 simple two-parameters (C, D) model.

152 At the scale of a river reach – delimited by an input *I* and an output *O* station – the model is 153 applied to the flood component $(Q_{I_f}(t) \text{ and } Q_{O_f}(t))$ of the total discharge $(Q_I(t) \text{ and } Q_O(t))$ 154 recorded at the input and output stations, respectively. $Q_{I_f}(t)$ and $Q_{O_f}(t)$ are estimated by 155 removing the baseflow components $Q_{I_b}(t)$ and $Q_{O_b}(t)$ from $Q_I(t)$ and $Q_O(t)$.

156 The resolution of Eq. 1 is obtained using the convolution approach proposed by Moussa 157 (1996) based on the Hayami assumptions (Hayami, 1951), i.e. considering C and D as 158 constant parameters over time along a channel of length *l*:

159 Eq. 2:
$$Q_{Ifr}(t) = \int_0^p Q_{If}(t-\tau) K(\tau) d\tau = Q_{If}(t) * K(t)$$

160 Where $Q_{Ifr}(t)$ is the routed input hydrograph on the flood component, p is the time memory of 161 the system, and the symbol * represents the convolution operator. As there is no problem of 162 calculation time, the term p must be large in comparison to the travel time along a channel 163 reach. In Eq. 2, the Hayami kernel function K(t) is expressed as follows:

164 Eq. 3
$$K(t) = \frac{l}{2(\pi D)^{1/2}} \frac{exp\left[\frac{cl}{4D}\left(2 - \frac{l}{ct} - \frac{ct}{l}\right)\right]}{t^{3/2}}$$

165 Eq. 2 is then used in a direct approach to define the *C* and *D* parameters, comparing $Q_{Ifr}(t)$ 166 with $Q_{Of}(t)$. In a theoretical case without lateral flows, under diffusive wave and Hayami 167 asumptions, the routed input hydrograph is equal to the observed output one $(Q_{Ifr}(t) = Q_{Of}(t))$ 168 as shown on Figure 2a.

169 **2.2** The diffusive wave model with lateral flows

170 Several solutions exist to resolve the DWE with lateral flows (see Cimorelli et al. (2014) for a review). In our paper, the one proposed by Moussa (1996) has been selected because an 171 172 analytical solution has been proposed to solve the inverse problem, as explained herein. 173 Moreover recently, an experimental evaluation of the solution has been performed by Moussa 174 and Majdalani (2019) on a large variety of hydrograph scenarios. The model is based on 175 Hayami assumptions, considering i) diffuse lateral flows as uniformly distributed along the 176 channel reach, and ii) the two C and D parameters as constant over time during the flood. It is 177 used to route the input hydrograph to the output station accounting for lateral flows. A 178 solution to the inverse problem of the DWE is used to model lateral flows, knowing the input 179 and the output hydrographs. This last case is suitable to be used when no observations on 180 lateral flows are available. Indeed, it gives information on the potential contributions of the 181 intermediate catchment, and notably on the estimation of peakflow of the lateral hydrograph.

182 Implementing lateral flows in the DWE needs to account for the lateral flow rate q per unit 183 length [L² T⁻¹] along a channel reach, as proposed by Moussa (1996):

184 Eq. 4
$$\frac{\partial Q}{\partial t} + C\left(\frac{\partial Q}{\partial x} - q\right) - D\left(\frac{\partial^2 Q}{\partial x^2} - \frac{\partial q}{\partial x}\right) = 0$$

In the particular case of uniform distribution of *q* along the reach, under the hypotheses used
in the Hayami model (*C* and *D* constant), Moussa (1996) proposes an analytical resolution:

187 Eq. 5
$$Q_{Ifr}(t) = \Phi(t) + (Q_{If}(t) - \Phi(t)) * K(t)$$

188 Eq. 6 with
$$\Phi(t) = \frac{c}{l} \int_0^t [Q_{Af}(\lambda) - Q_{Af}(0)] d\lambda$$

189 Eq. 7 and with
$$Q_{Af}(t) = \int_0^l q(x, t) dx$$

190 where q(x, t) [L² T⁻¹] is the lateral flow rate per unit length as a function of distance along the 191 channel reach *x*. $Q_{A f}(t)$ represents the flood component of the lateral hydrograph $Q_A(t)$. As 192 illustrated on Figure 2b,c, the expression q(x, t) may be positive or negative depending upon 193 the lateral flow direction from the channel when inflow or outflow occurred, respectively.

- According to Moussa (1996), the inverse problem identifies $Q_{Af}(t)$ by knowing $Q_{If}(t)$ and Q_O 195 f(t), and according to a predetermination of the two parameters *C* and *D*. Given that,
- 196 Eq. 8 $Q_{Afr}(t) = Q_{0f}(t) Q_{Ifr}(t)$

197 Eq. 9 with
$$\Phi(t) - \Phi(t) * K(t) = Q_{Afr}(t)$$

198 Using the Laplace transforms, an approximation of the solution of Eq. 9 is

199 Eq. 10
$$\Phi(t) = Q_{Afr}(t) + Q_{Afr}(t) \sum_{i=1}^{\infty} K^i(t)$$

200 with

201 Eq. 11
$$K^{i}(t) = K * K * ... * K$$
 (*i times*)

202 $Q_{A_f}(t)$ can be easily calculated using Eq. 12, after the identification of K(t) in Eq. 3:

203 Eq. 12
$$Q_{Af}(t) = \frac{l}{c} \frac{d\phi}{dt}$$

The lateral hydrograph $Q_A(t)$ is simply calculated from $Q_{Af}(t)$ by adding $Q_{Ab}(t)$, the difference of baseflow $(Q_{Ob}(t) - Q_{Ib}(t))$ between *O* and *I* stations. 206 When no observations of $Q_A(t)$ are available to calibrate the model, this simple approach with 207 two parameters (C, D) is favoured. This choice is made in comparison with more complex 208 approaches adding a further degree of freedom, by considering additional parameters. Thus, 209 according to the principle of parsimony, we choose the most simplest and robust model, 210 considered as more reasonable. For the inverse problem, many couples of solutions (C, D)211 exist to simulate $Q_A(t)$. In our paper, the sensitivity on the various couples of solutions (C, D) 212 is then characterized to assess whether the different solutions brings some different 213 interpretations or not on the lateral contributions.

214 **2.3** Quantification of the peakflow amplification/attenuation

Based on the conceptual schemes of Figure 1, we expect that the global lateral flow at a given time is the sum of simultaneous negative and positive q values originating from various processes. Knowing this, we now must understand the causes of the attenuation or amplification of peakflow (without the baseflow component) shown on Figure 2, from the upstream station $Qx_{I f}$ [peakflow of the input hydrograph $Q_{I f}(t)$: black curve] to the downstream one $Qx_{O f}$ [peakflow of the output hydrograph $Q_{O f}(t)$: blue curve]. The difference between both peakflows is noted *E*:

222 Eq. 13
$$E = Qx_{0f} - Qx_{If} = E_D + E_A$$

with *E*>0 in the case of peakflow amplification (Figure 2b), and with *E*<0 in the case of attenuation (Figure 2c). *E* is composed of two terms linked to the hydraulic properties of the channel E_D (controlled by *D* parameter), and linked to the lateral flows (E_A). Noted that in the case of no-lateral flow component (Figure 2a), $E_A = 0$ and $E = E_D \le 0$.

227 Diffusion is responsible for a peakflow attenuation E_D of the input hydrograph expressed as 228 follows:

229 Eq. 14
$$E_D = Qx_{Ifr} - Qx_{If}$$
 with $E_D \le 0$

10

with Qx_{Ifr} , the peakflow of the routed input hydrograph $Q_{Ifr}(t)$ without lateral flow (dashed grey curve). Lateral flow is responsible for an attenuation/amplification expressed as:

232 Eq. 15
$$E_A = Qx_{Of} - Qx_{If}$$

233 Depending upon the importance of out- and in-flows, E_A may be negative or positive, 234 respectively.

Hereafter, we investigate the flood processes in a catchment favouring peakflow attenuation (E<0) according to high diffusion ($E_D<0$) and the occurrence of river losses and overbank flow ($E_A<0$).

238 **2.4 Sensitivity analysis on a virtual example**

239 To illustrate the model behaviour and its calibration, various parameterization sets of C and D 240 parameters were used to apply the inverse model on the same couple of theoretical hydrographs Q_{If} and Q_{Of} (Figure 3). Figure 3a and 3b present the simulations carried out by 241 242 fixing D and varying C. In the case of a gaining reach (Figure 3a), the results show that the 243 highest lateral peakflow is simulated for the lower C. When increasing C, lateral peakflow 244 decreased and became constant when C exceeded a threshold (here C>0.4 m/s) at the same 245 time that outflow was simulated at the start of the flood event. This illustrates a dynamics of compensation of out- and inflows during the same flood event due to a conservation of the 246 247 flood volume (total lateral flood volume was equal for all simulation tests). The model 248 behaviour is simpler in the case of a losing reach, because Figure 3b shows that the higher the 249 C, the higher and the earlier will be the lateral outflow peak. Figure 3c and 3d show a similar 250 test, but now fixing C and varying D. In the case of lateral inflow (Figure 3c), the results 251 showed that the higher the D, the higher will be lateral peakflow. Similar to Figure 3a, 252 temporal lateral outflow is simulated at the beginning of the flood when D increases. In the 253 case of a losing reach (Figure 3d), the higher the D, the higher and the earlier will be the 254 lateral outflow peak. Globally, this sensitivity analysis shows that various lateral hydrographs 255 are simulated for different couples of *C* and *D* parameters, due to equifinality in the modelling 256 approach. As observed in previous studies (Moussa & Bocquillon, 1996a; Yu et al., 2000; 257 Charlier et al., 2009; Cholet et al., 2017), *C* is more sensitive than *D*, as a variation of *C* by 4 258 in our test case generated a same range of lateral peakflow variations as a variation of *D* by 259 10,000.

260 Consequently, for a given flood event, the C and D parameters should be optimized. In our 261 case, we expect that, whatever the flood, the fast component of lateral flow will contribute to 262 peakflow at the output station. Thus, we chose to optimize C and D in order to put in phase 263 peakflows of the routed inflow and outflow hydrographs. Figure 3e and 3f show the effect of 264 the inverse model, varying D and calibrating C under these conditions. Regarding the 265 evolution of parameters, it shows that C decreases when D increases, but up to a lower limit 266 (of C = 0.22 m/s in our case for D > 150 m²/s). In the case of a gaining reach (Figure 3e), the 267 higher the D, the higher will be the lateral peakflow, but for a losing reach (Figure 3f), the 268 higher is the D, the lower will be lateral outflow peakflow. These results show that, contrary 269 to varying D and fixing C (Figure 3b), lateral outflow peakflow decreases when D increases, 270 provided the corresponding C is optimized following the hypothesis of a same routing scheme 271 for lateral out- and in-flows.

272 **2.5** Framework of the modelling approach used

We propose a framework in this paper for defining amplification/attenuation of peakflows in a river reach. Though interpretation of the results will obviously be better when confronting them with field knowledge, the model application can stand alone in order to help decipher some hydrological processes in catchments with a complex behaviour. The modelling framework has four steps: First, the base and flood components of the hydrographs at the two gauging stations
must be separated;

- Second, calibration of the DW model parameters (*C* and *D*) applying the direct approach without lateral flow (Eq. 2). We saw above how the calibration strategy may influence the results, and we thus must optimize *C* and *D* by inputting phase peakflow of the routed inflow Qx_{Ifr} and of the outflow Q_{Of} (as illustrated on Figure 1);

- Third, is the calculation of the lateral hydrograph, applying the inverse approach of the
 DW model using in Eq. 10 the pre-calibrated *C* and *D* values from Eq. 2;
- Fourth, we quantify peakflow amplification/attenuation using Equations 12 and 13.

The choice of the modelling calibration on peakflows proposed in the second step appears to be the most likely in the absence of monitoring lateral flow along the reach. In order to account for the uncertainty on this choice, different sets of simulation by varying D values (and corresponding calibrated C values; see Figure 3f) should be carried out. This is tested in the following case study.

292 **3 Case study**

293 3.1 Field site

294 **3.1.1 Basin presentation**

The Iton basin is located in Normandy, north-west France (Figure 4a). Land use consists mainly in cereal crops and grassland, and the only important urban area is Evreux (100,000 inhabitants) on the Iton in the downstream part of the basin. The topographic catchment is 1050 km² at the Normanville gauging station, 7 km downstream Evreux city (Figure 4b).

299 3.1.2 Climate

The climate is of the humid temperate oceanic type. Annual rainfall ranges between 500 and 1000 mm, with an inter-annual average of 600 to 715 mm between the upstream and downstream areas of the catchment, respectively. The intra-monthly variations are relatively 303 low, with slightly wetter months in autumn (60 to 80 mm/mo) compared to other periods of 304 the year (40 to 65 mm/mo), but in general rainfall is quite regularly distributed throughout the 305 year. The region has also been characterized in the past by exceptional rainfall in 2000-2001 306 (about 100 mm of rainfall depth in 2 to 4 days only), generating catastrophic flood events 307 enhanced probably by two previous years of accumulated wetness (Pinault et al., 2005).

308 3.1

3.1.3 Geology and hydrogeology

The geomorphological context of the basin can be described as a plateau cross-cut by the Iton River and its main tributary, the Rouloir (Figure 4b). On the plateaux, Late Cretaceous chalk formations are covered by a clayey formation associated with loess, up to several tens of metres thick. In the valley bottom, chalk formations may also be covered by alluvium. Thus, aquifers in the basin are located in the karstified chalk that is mostly covered by shallow formations, as indicated by the non-exposed karst-aquifer symbol in the extract of the World Karst Aquifer map (Chen et al., 2017) in Figure 4a.

316 The underground karst networks in the chalk are fed by diffuse infiltration waters through 317 swallow-holes developed in the (non-cohesive) shallow formations on the plateau, but also 318 from river losses where the chalk is exposed in the river bed. This can generate a drying up of 319 the stream, as in the "Dry-Iton" reach and the Rouloir tributary (Figure 4b). Outlets of these 320 karst aquifers are the springs at the foot of the hillslopes close to the river, and feeding it. The 321 use of artificial tracers (yellow arrows; Figure 4b) showed that infiltrated Iton waters bypass 322 the streambed underground, to reappear downstream in the same bed via resurgence springs 323 located near the confluence with the Rouloir (David et al, 2016).

324 3.1.4 Conceptual model of lateral flow

The conceptual model presented in Figure 4c is the result of hydrogeological and hydrological studies, highlighting surface-water / groundwater (SW/GW) interactions of various origin (Charlier et al, 2015a; David et al., 2016). The main horizontal line represents the Iton River 328 for the 75-km reach between the two gauging stations Bourth and Normanville (input and 329 output in Figure 4c; see Figure 4b for location). Surface flow is in light blue colour and 330 groundwater flow in dark blue. The dashed line represents ephemeral streams due to river 331 losses. For surface flow, the main properties of the Iton are: i) Drying-up of the drainage 332 network (as well as the Rouloir) where it crosses the karst zone; and ii) Contribution of the 333 main tributary (Rouloir) and of the two groups of springs near the confluence. Groundwater 334 flow is composed of infiltrated river losses as well as aquifer contributions via several springs. 335 Finally, in this conceptual scheme of SW/GW interactions in the Iton basin in its karst part, 336 lateral flow is defined by outflow from river losses and inflow from groundwater origin.

337 3.2 Data

338 3.2.1 Hydrological and hydrogeological time series

339 Mean rainfall and soil-humidity indices (HU2) over the Bourth and Normanville sub-basins 340 were obtained from METEO FRANCE, available on the COMEPHORE/ANTILOPE and 341 SAFRAN ISBA MODCOU (Habets et al., 2008) databases, respectively. Even if HU2 is not 342 derived from observation data sets, this index is widely used by modelers to initialize 343 hydrological models. Thus, despite the uncertainty on the model output, this is a pertinent data 344 set of soil wetness, that is used as it by end-users (forecasters). Streamflow hydrographs were 345 obtained from the "Service de Prévision des Crues" (SPC) for the Bourth (code: H9402030) 346 and Normanville (code: H9402040) stations, available on Banque Hydro (2015). Groundwater 347 levels were obtained from ADES (2015). All hydrological and hydrogeological time series were synchronized at an hourly time step over the 1999-2014 period. 348

349

Flood selection and processing for model application 3.2.2

350 For flood event analysis, the highest 33 peakflows at the Bourth gauging station (Input 351 station) were selected from the dataset (Table 1). Rainfall events ranged between 16 and 100 mm, and peakflows between 9 and 26 m^3/s at the Input station and 5 to 17 m^3/s at the 352

Output one. Minimum discharge-inducing overbank flow is 14 and 10 m³/s for the Bourth and Normanville stations, respectively. These thresholds correspond to discharge generating flows in the flood plain without rapid return towards the channel. Table 1 shows that the 10 highest flood events were partially subject to overbank flow.

The inverse approach of the DW model was applied to the karst portion of the Iton River from hydrographs of the Input (Bourth) and Output (Normanville) stations. Model application requires in a first step a separation of the base and flood components (Section 2.5), which was done with the BFI method (Gustard et al., 1992) using ESPERE software (Lanini et al., 2016).

361 4 Results

362 **4.1 Groundwater influence on surface flow**

363 4.1.1 Base and flood components

364 An example of the base and flood flow separation is given in Figure 5 for the 2000-2001 365 hydrological cycle at the Bourth Input station in black (Figure 5a) and the Normanville Output station in blue (Figure 5b). The input-output relationships for the base and flood 366 367 components account for a 2-day delay (Figure 5c), corresponding to the mean delay of 368 peakflows. It shows a contrasted behaviour: the baseflow increases 3 to 4 fold from input to output station, whereas flood flow decreases by several m^3/s for the highest flood events. This 369 370 means that lateral groundwater inflow contributed highly to stream flow for the base 371 component, at the same time that strong lateral outflow occurred for the flood component.

372 **4.1.2 Baseflow analysis**

Figure 6 shows the relationships between baseflow calculated at the output gauging station and groundwater levels for six piezometer wells. We observe a slight to fair correlation for wells located on the plateau, having mainly multi-year cycles (Normanville and Cierrey piezometers Figure 4b for location). Best correlations are obtained for wells with annual cycles at Nogent-le-Sec, Moisville, and Coulonges (best linear correlation with $R^2 > 0.7$). The Graveron well, with both multi-year and annual cycles, reflects an intermediate behaviour.
These results show that SW/GW interactions on the baseflow component is driven by lateral
exchanges with the karst aquifer, best shown by the Coulonges piezometer well.

381 **4.1.3** Flood analysis

382 The input-output relationships for peakflow are plotted in Figure 7 according to two factors 383 used as key indices of the catchment saturation level: the soil-humidity index (HU2, Figure 384 7a) and the karst saturation index (GW depth z in the Coulonges well; Figure 7b). Before 385 assessing the effect of such factors, it is interesting to observe that output peakflow is always 386 less than, or equal to, the input one. A peakflow attenuation is also observed for the highest flood events when $Qx_I > 15 \text{ m}^3/\text{s}$. This value corresponds to the threshold of overbank flow at 387 388 the input station, meaning that this process may explain part of the attenuation of the highest 389 flood events. In the first case, HU2 seems not to be a discriminant factor for explaining the 390 data variability, as most events are characterized by indices close to the saturation level (i.e. 391 HU2 ~60, HU2 ranging between 40 and 65). In the second case, the initial GW depth seems 392 to explain the attenuation variation for a given input peakflow: the higher the initial GW 393 depth, the higher will be output peakflow. Globally, this analysis shows that the soil saturation 394 index is not a suitable factor for differentiating flood intensity. On the contrary, these results 395 show that the peakflow attenuation is related to the antecedent groundwater level.

396 **4.2** Simulation of a lateral flood hydrograph

397 4.2.1 Model application to a mono-peak flood event

Figure 8 shows an example of the DW model application for the mono-peak flood event of 06/01/2001 with the optimized parameters C = 0.35 m/s and D = 500 m²/s. From top to bottom rainfall, soil saturation index (HU2), groundwater depth in the Coulonges well, discharge for the flood component, and total discharge are shown. On each discharge plot, four hydrographs correspond to the observed input hydrograph (Q_{If} and Q_{I} ; black curve), the 403 observed output hydrograph (Q_{Of} and Q_{O} ; blue curve), the routed input hydrograph (Q_{Ifr} and 404 Q_{Ir} ; dashed grey curve) using the direct model without lateral flows, and the simulated lateral 405 hydrograph (Q_{Af} and Q_A ; dotted red curve) using the inverse model. Both soil saturation index (HU2 >58) and groundwater levels (GW >-15 m AGL) were saturated before the 406 407 beginning of the flood. The discharge analysis shows a strong attenuation of the flood hydrograph along the reach, halving the peakflow from 26.8 m³/s (Qx_l) to 13.7 m³/s (Qx_o). 408 Before the flood, lateral inflow ($Q_A(t)$) was 3.7 m³/s; but during the flood it became negative. 409 410 This may be interpreted as a continuous contribution of lateral baseflow hidden by occasional 411 high losses in the flood component. Using the minimum values of the lateral flood component, we can quantify the maximum intensity of lateral outflow Qn_{Af} as -6.8 m³/s. The 412 peakflow attenuation due to outflow (losses+overbank flow) E_A is -6.6 m³/s and the 413 attenuation due to diffusion E_D is -9.2 m³/s, leading to a total peakflow attenuation E of -414 $15.8 \text{ m}^3/\text{s}.$ 415

416 Analysis of this single flood event shows that, despite lateral baseflow, the outflow associated 417 to the flood component can be quantified. As outflow by overbank flow may occur during the 418 highest discharge, outflow by river losses is probably continuous as long as the stream flows 419 at the input station. This suggests that losses are compensated by highest baseflow discharge 420 during recession periods, leading to under-estimating their real values. The other interesting 421 point is that we can compare peakflow attenuation by hydraulic processes (diffusion) and by 422 outflow (river loss+overbank flow), quantifying it (for a given parameterization set) equal to 423 57% and 43%, respectively.

424 **4.2.2** Distribution of model parameters

Following the above example, the model was applied to the 33 main flood events (Table 1), using the parameterization strategy presented in Section 2.5. Several values of D were selected for optimizing the C parameter, for inputting phase peakflows of the routed input

hydrograph (Q_{lr}) and of the output one (Q_0) . Figure 9 shows C distribution using boxplots for 428 five D values of 500, 1000, 2500, 5000, and 10,000 m²/s, which is the classic range for 429 430 streams and rivers (Todini, 1996), such as in our study, knowing that D increases with the size of the river. Boxplot analysis shows that C values range between 0.1 and 0.4 m/s with a 431 432 relative small variability for a given D. The higher the D, the lower will be the range of C 433 values from 0.35 to 0.11 m/s. As shown in the sensitivity analysis (Section 2.4 above), a 434 lower limit of C values to almost 0.15 m/s is observed for D values above $2500 \text{ m}^2/\text{s}$. 435 Knowing that lateral flows are highly sensitive to C values and less so to D values, various 436 parameterization sets will be tested in the following section.

437 **4.3** Quantification of peakflow attenuation

438 **4.3.1** Assessment of lateral outflow

439 In order to quantify outflow during floods, we express the maximum lateral outflow intensity 440 as a function of the input peakflow. Figure 10 presents Qn_{Af} (i.e. maximum losses as negative 441 values attributed to outflow) vs. Qx_{If} for all the 33 flood events and for five calibration 442 strategies that vary D (and the corresponding optimized C) from 500 to 10,000 m²/s; dark blue 443 colours refer to the lowest D values. The first result confirms that outflow for the flood 444 component is simulated for all flood events, regardless of input peakflow. This means that 445 lateral outflow intensities of the fast component are systematically higher than potential 446 lateral inflow values during the flood, i.e. flood flow from karst springs and tributaries, or surface runoff. As expected, the second result confirms that, globally, the lateral outflow 447 intensity is higher for a lower D. Outflow increases with increasing input peakflow, but the 448 relationship stabilizes when $Qx_{If} > 12 \text{ m}^3/\text{s}$, very close to the threshold value of 14 m³/s for 449 450 starting overbank flow when considering peakflow of the total discharge $Qx_I(Qx_I=Qx_{If}+Q_{Ib})$. For $D=500 \text{ m}^2/\text{s}$, Qn_{Af} reaches a ceiling of ~9 m³/s, against 7 m³/s for $D=1000 \text{ m}^2/\text{s}$ and 2 m³/s 451 452 for the highest D value of 10,000 m²/s.

453 The influence of the initial karst saturation level on outflow intensity has been tested, and any 454 concluding results were highlighted to validate this hypothesis. Consequently, input peakflow 455 seems to be the main driver of outflow intensity. An interesting result is that when discharge is below the overbank flow threshold ($Qx_{If} < 12 \text{ m}^3/\text{s}$), outflow is mainly due to river losses, 456 457 following a linear relationships between Qn_{Af} and Qx_{If} . Depending upon the parameterization strategy, these river losses may reach high values of up to 9 m^3/s , corresponding to half of the 458 459 peakflow at the input station. When discharge exceeds this threshold, the highest discharge 460 outflow ceiling may be explained by a limitation of the infiltration rate into the stream bed, 461 due to a ceiling of the water-level increase in the river bed when overbank flow occurs.

462

4.3.2 Factors influencing peakflow attenuation

463 Two phenomena participate in peakflow attenuation (E): hydraulic processes due to flood 464 wave diffusivity (E_D) and hydrological processes due to outflow (E_A) (cf. Eq. 13). To quantify their respective roles, Figure 11 presents the E_A vs. E_D relationships for the same five 465 parameterization strategies as in Figure 10. We see that E_D ranges between -18.0 to -0.5 m³/s 466 whereas E_A ranges between -9 to 4.0 m³/s. As expected, the higher the D (light blue colour), 467 468 the higher the $|E_D|$ values (negative in the graph since E_D is inevitably an attenuation of 469 peakflow). Except for cases with the lowest D values ($D=500 \text{ m}^2/\text{s}$), and some cases with 470 $D=1000 \text{ m}^2/\text{s}$ (dark blue colour, Figure 11), the points lie below the $E_D=E_A$ line, i.e. 471 attenuation due to diffusivity is often higher than that due to outflow. It is interesting to note 472 that when D is high and thus $|E_D|$ is high, E_A is positive (lateral flow became inflow) while 473 being below the $E_D = -E_A$ line. This means that in these cases, despite flood amplification due 474 to lateral inflow, peakflow attenuation is finally observed because the effect of high 475 diffusivity overtakes it.

These simulation tests replicate a wide range of classic stream and river *D* values found in the literature. Finally, we should evaluate the proportion of E_D and E_A for the theoretical range of *D*, calculated with the following formula proposed by Chow (1959) that considers simple network descriptors: $D = \overline{Q} / (2 \times slope \times width)$, where \overline{Q} is the mean flow discharge for a rectangular section. Varying the mean slope of the river from 0.0010 to 0.0015, the mean river width from 5 to 10 m, and \overline{Q} from 5 to 20 m³/s, *D* ranges between 250 and 2000 m²/s, over a quite small range of values compared to the tested ones (500 to 10,000 m²/s). In this theoretical case, E_D and E_A are roughly equal according to Figure 11, but this result should be taken with care due to the high spatial variability of channel properties along the Iton River.

In summary, these results show that, in the case of low D values, peakflow attenuation is equally due to diffusion and to outflow, but in cases with high D values, most of this attenuation is caused by diffusion. They also show that, despite lateral inflow in the case of highest D values, these contributions are compensated by a strong attenuation due to flood wave routing.

490 **5 Discussion**

491 5.1 On the interest of using a diffusive wave model to assess peakflow attenuation 492 and/or amplification

493 Although aquifer's recharge by river losses can attenuate floods in arid/Mediterranean 494 environment (Sorman et al., 1997; Hughes & Sami, 1992; Lange, 2005; Dahan et al., 2007; 495 Vázquez- Suñé et al., 2007) or in karst basins (Jourde et al., 2013; Charlier et al., 2015b; 496 Brunet and Bouvier, 2017), we cannot neglect hydraulic diffusion processes when we are 497 interested in peakflow forecasting. In the case of permeable basins, our results show that, 498 despite the presence of an infiltration zone in the river bed with significant losses (several 499 m^{3}/s), peakflow attenuation is mainly related to diffusion of the flood wave. This raises the 500 question of the mechanisms favouring such attenuation, which may be related to the 501 meandering morphology of the drainage network, as well as to the zones of temporary storage 502 for highest flood events (Moussa & Cheviron, 2015).

503 Our example shows the added value of using a diffusive model-combining direct and 504 inverse problem approaches-for better understanding and quantifying SW/GW interactions 505 during floods, and which deserves to be tested on other types of basins where significant 506 lateral losses and/or gains are observed (Martin & Dean, 2001; Ruehl et al., 2006; Payn et al., 507 2009). For instance, the inverse DWE model has also been used for investigating lateral flow 508 in underground karst conduits, and for defining the exchanges between conduits and the 509 fissured matrix (Cholet et al., 2017). In parallel to these considerations that promote the model 510 as a tool for diagnosing SW/GW exchanges at different scales, our results highlight the 511 inaccuracies that can be generated by using non-diffusive models, as is frequently the case for 512 flood modelling (see review in Singh, 2002) and for karst basins (Bailly-Comte et al., 2012; 513 Dvory et al., 2018). Our results are coherent with Naulin's work (2012) in the Cévennes 514 region (southern France), who showed that DWE was more suitable in lowland areas, 515 including karst formations, than in mountains with less permeable hard-rock formations.

516 **5.2** Surface-water / groundwater interactions in permeable basins

517 5.2.1 River losses and overbank flow

518 The relationship between lateral outflow and input peakflow has established a function of 519 river losses that improves the understanding of SW/GW interactions in permeable basins. In 520 fact, below the discharge threshold for overbank flow, outflow is mainly generated by losses that account in our case for several m^3/s during a flood. Although this process is well known 521 522 in many catchments when studying low water-level periods, it is generally not considered 523 during flood events, because inflow conceals it. Despite some works on infiltrating floodwater 524 in basins characterized by disconnected river-aquifer systems (Hughes & Sami, 1992; Lange, 525 2005; Dahan et al., 2007; Vázquez- Suñé et al., 2007), the estimation of infiltration rate 526 during the flood (i.e. at a high temporal resolution) is generally disregarded. Thus, our study

brings a relevant approach to help quantify the loss intensities as well as the recharge rate ofthe underlying aquifers.

529 Outflow due to river losses is an important process as it may represent up to half of peakflow 530 in models, depending upon the parameterization strategy. The estimated value loss of several 531 m^{3}/s is important, but not exceptional as it is coherent with observations made on other 532 ephemeral karst rivers in southern France (Ladouche et al. 2002, 2004), or in the Jura 533 Mountains (Charlier et al., 2014). The linear relationship between outflow intensity and input 534 peakflow (below the overbank flow threshold) argues for the control of loss rate by water 535 height in the river. This implies that the aquifer fed by the losses is disconnected with the 536 river, agreeing with the absence of influence of groundwater level on this relationship.

537 The ceiling of outflow intensity with the increase of discharge into the river is most probably 538 explained by the occurrence of overbank flow, knowing that flood plain attenuation can play a 539 key role on the modification of hydrograph shape (Sholtes & Doyle, 2011; Valentova et al., 540 2010; Rak et al., 2016; Fleischmann et al., 2018). Indeed in the study case, outflow on the 541 flood plain appears to be an important process of peakflow attenuation for the highest floods, 542 since the increase in infiltration with an increased input peakflow is stopped (Figure 10). 543 Another concept may explain this outflow ceiling: several studies of karst hydrology reported 544 a limitation of infiltration during rainfall events due to small void diameters or constricted 545 conduits at depth (Bonacci, 2001; Bailly-Comte et al., 2009). However, such a process 546 requires specific monitoring that fell outside the scope of our study.

547

5.2.2 Loss and gain in river reaches

The specific features of the studied river, including both loss and gain reaches, render an analysis of lateral exchanges difficult. The occurrence of both out- and in-flow in some reaches has been conceptualized as hydrologic turnover (Covino et al. 2011; Mallard et al., 2014) in simultaneous loss and gain reaches. This pattern was highlighted both in chalk catchments, where it was found that groundwater flooding consists of a combination of intermittent stream discharge and anomalous springflow (Hughes et al., 2011), and in karst rivers characterized by successive loss and gain reaches from a multi-layered aquifer in deep canyons (Charlier et al., 2015b). Improving the understanding of lateral exchange during floods, our modelling approach opens a novel way to help deciphering the various contributions of loss and gain.

558 5.3 Saturation state in a karst catchment: soil moisture vs. aquifer storage

559 An influence of the aquifer-saturation state on peakflow attenuation is observed in the karst 560 part of the catchment. This agrees with several papers on the exceptional groundwater 561 flooding of 2000-2001, in karstified chalk areas of northern Europe (Finch et al., 2004; 562 Pinault et al., 2005; Hughes et al., 2011; Morris et al., 2015; Thiéry et al. 2018). However, we 563 did not see this influence on the relationship between lateral outflow intensity and input 564 peakflow on the flood component, as might have been expected. This means that aquifer 565 saturation probably influences the baseflow component, which increases strongly in the karst 566 reach (Figure 5). Even if river losses recharge the underlying aquifer, which in turn feed the 567 river downstream, these results are not contradictory. In fact, losses are controlled by the 568 infiltration zone, which is never fully saturated, whereas baseflow is linked to groundwater 569 levels (Figure 6).

570 Comparing this pattern with classic catchment hydrology, it is interesting to note that the soil 571 moisture index HU2 (which only reflects the supposed behaviour of the soil cover) doesn't 572 influence the runoff-runoff relationships in our case, even for events for which the saturation 573 state of the aquifer is low. This is finally consistent with the fact that, with lateral 574 contributions being mainly of underground origin, it is the initial antecedent saturation of the 575 aquifer that is the best indicator of the saturation state of the catchment. This result reflects the 576 specificities of karst catchments as compared to other types of catchment. Most production functions in hydrological models are designed to consider the role of soil moisture (e.g.
Horton, 1933; Philip, 1957; SCS, 1972; Morel-Seytoux, 1978). Our results show, however,
that such models cannot be generalized for carbonate basins with significant SW/GW
interactions, when neglecting deep infiltration and groundwater storage in the bedrock.

581

5.4 Implications for flood forecasting

582 On the basis of our results and the available data, several insights can be proposed for reliable 583 flood forecasting in permeable basins including karst as well as more generally ephemeral 584 streams. As a karst aquifer is a complex hydrogeological medium, an analysis of its 585 hydrogeological behaviour and its role in runoff at the basin scale is an essential prerequisite. 586 Our first recommendation is not to neglect the influence of hydraulics (diffusivity) on flood 587 routing. For example, floodplains or karst canyons promote meandering networks that are 588 supposed to be an exacerbating factor in wave diffusion. Knowing this, the main risk in 589 forecasting is thus to significantly over-estimate peakflow when applying non diffusive flood-590 routing models. The second recommendation is to account for river loss in the modelling 591 approach if flood analysis shows significant outflow. The relationship we propose between 592 input peakflow and lateral flow intensity can serve as a basis for such an infiltration function 593 to be implemented in a model. The third recommendation is to use groundwater level as an 594 index of basin saturation for initializing hydrological models. This has to be considered in 595 preference to a soil moisture index, which appears inappropriate for such a basin with fast 596 infiltration at depth.

597 6 Conclusions

We propose a framework for quantifying peakflow attenuation and/or amplification in a river, based on defining lateral flow during floods in the case of a highly permeable basin that favours surface-water/groundwater interactions. The novelty of our research is the use of the inverse problem of the DWE proposed by Moussa (1996) to simulate a lateral flow hydrograph in a river reach draining karst formations (Normandy, France), knowing the hydrographs from both upstream and downstream gauging stations. Application of the model to several flood events of various intensity shows that, despite a high groundwater contribution to the baseflow component, the peakflow was strongly attenuated. Our approach was designed to differentiate between attenuation generated by wave diffusion and that generated by outflow related to river loss and overbank flow.

608 Our results provide new insight in flood routing processes in a karst context and more 609 generally in permeable basins favouring ephemeral streams and disconnected river-aquifer 610 systems. First, the model restituted the global dynamics of lateral flow, given information on 611 the hydrological processes involved. Second, we could propose a relationship quantifying 612 outflow intensity as a function of peakflow discharge at the upstream gauging station. Based 613 on previous experimental work investigating the hydrological processes at the origin of loss 614 and gain in rivers, we could highlight the importance of river losses and then of overbank 615 flow for highest flood events. Third, as lateral flow is characterized for unsteady-state 616 conditions, the relative contribution of outflow compared to attenuation due to diffusion was 617 characterized for several sets of model parameterization, allowing interpretations according to 618 parameter sensitivity.

In a more global way, our approach deserves to be tested as a diagnostic tool before applying hydrological models for flood forecasting in permeable—karst—basins. The conclusions provided by our model can help modellers in selecting the best tool in terms of hydrological processes to be simulated as well as of parameterization strategy.

623 Data Availability Statement

624 The datasets used in this article can be obtained by contacting Jean-Baptiste Charlier625 (j.charlier@brgm.fr).

626 Acknowledgments

We warmly thank Cédric Zaniolo and Stéphane Piney of the "Service de Prévision des Crues (SPC)" at Rouen for fruitful discussions that helped improving the paper. The work was funded by the French Governmental Administration for Risk Prevention (DGPR), the Service Central d'Hydrométéorologie et d'Appui à la Prévision des Inondations (SCHAPI), and the French Geological Survey (BRGM).

632

633 **References**

- ADES (2015). portail ADES, http://www.ades.eaufrance.fr/ (data exported in July 2015).
- Bailly-Comte, V., Jourde, H., & Pistre, S. (2009). Conceptualization and classification of
 groundwater-surface water hydrodynamic interactions in karst watersheds: Case of the karst
 watershed of the Coulazou River (southern France), *Journal of Hydrology*, *376*(3–4): 456–
 462, doi:10.1016/j.jhydrol. 2009.07.053.
- Bailly-Comte, V., Borrell-Estupina, V., Jourde, H., & Pistre, S. (2012). A conceptual
 semidistributed model of the Coulazou River as a tool for assessing surface water–karst
 groundwater interactions during flood in Mediterranean ephemeral rivers, *Water Resources Research*, 48, W09534, doi:10.1029/2010wr010072.
- Banque Hydro (2015). http://www.hydro.eaufrance.fr/ (data exported in February 2015).
- Bates, P.D., & De Roo, A.P.J. (2000). A simple raster-based model for flood inundation
 simulation. *Journal of Hydrology 236*, 54–77
- Bonacci, O. (2001). Analyses of the maximum discharge of karst springs, *Hydrogeological Journal*, 9, 4, 328–338.

- Bonacci, O., Ljubenkov, I., & Roje-Bonacci, T. (2006). Karst flash floods: An example from
 the Dinaric karst (Croatia), *Natural Hazards and Earth System Sciences*, 6, 195–203,
 doi:10.5194/nhess-6-195-2006.
- Brunet, P., & Bouvier, C. (2017). Retour d'expérience sur la crue du 12 septembre 2015 à
 Lodève (Hérault, France) : influence du karst sur les débits de pointe de crue. *La Houille Blanche, 3*: 39-46. DOI :10.1051/lhb/2017020.
- Camarasa Belmonte, A.M., & Segura Beltrán, F. (2001). Flood events in Mediterranean
 ephemeral streams (ramblas) in Valencia region, Spain. *Catena*, 45, 229–249.
 doi.org/10.1016/S0341-8162(01)00146-1
- Charlier, J.-B., Moussa, R., Cattan, P., Cabidoche, Y.M., Voltz, M. (2009). Modelling runoff
 at the plot scale taking into account rainfall partitioning by vegetation: application to
 stemflow of banana (Musa spp.) plant. *Hydrology and Earth System Sciences*, *13*, 2151–2168,
 doi: 10.5194/hess-13-2151-2009.
- 661 Charlier, J.-B., Desprats, J.-F., & Ladouche, B., (2014). Appui au SCHAPI 2014 Module 1
- 662 Rôle et contribution des eaux souterraines d'origine karstique dans les crues de la Loue à
- 663 Chenecey-Buillon, BRGM/RP-63844-FR report, 109p. http://infoterre.brgm.fr/rapports/RP-
- 664 63844-FR.pdf (accessed April 2018).
- Charlier, J.-B., David, P.-Y., Lanini, S. & Desprats, J.-F. (2015a). Appui au SCHAPI 2015 –
 Module 1 Rôle et contribution des eaux souterraines d'origine karstique aux crues des
 bassins normands de l'Iton et de l'Orbiquet. *BRGM RP-65570-FR report*, 84 p.
 http://infoterre.brgm.fr/rapports/ RP-65570-FR.pdf (accessed April 2018).
- 669 Charlier, J.-B., Moussa, R., Bailly-Comte, V., Danneville, L., Desprats, J.-F., Ladouche, B., &
- 670 Marchandise, A. (2015b). Use of a flood-routing model to assess lateral flows in a karstic
- 671 stream: implications to the hydrogeological functioning of the Grands Causses area (Tarn

- 672 River, Southern France), *Environmental Earth Sciences*, 74: 7605–7616, DOI
 673 10.1007/s12665-015-4704-0.
- 674 Chen, Z., Auler, A. S., Bakalowicz, M., Drew, D., Griger, F., Hartmann, J., Jiang, G.,
- 675 Moosdorf, N., Richts, A., Stevanovic, Z., Veni, G., & Goldscheider, N., (2017). The World
- 676 Karst Aquifer Mapping project: concept, mapping procedure and map of Europe.
- 677 *Hydrogeological Journal*, 25: 771–785. DOI 10.1007/s10040-016-1519-3
- Cholet, C., Charlier, J.-B., Moussa, R., Steinmann, M., & Denimal, S., (2017). Assessing
 lateral flows and solute transport during floods in a conduit-flow-dominated karst system
 using the inverse problem for the advection-diffusion equation, *Hydrology and Earth System Sciences*, *21*: 3635-3653. DOI: 10.5194/hess-21-3635-2017.
- 682 Chow, V. T. (1959). Open-Channel Hydraulics. McGraw-Hill, New York. 680p.
- 683 Cimorelli, L., Cozzolino, L., Della Morte, R., & Pianese, D. (2014). Analytical solutions of 684 the linearized parabolic wave accounting for downstream boundary condition and uniform 57-76, 685 lateral inflows. Advances in Water Resources, 63. DOI: 686 /10.1016/j.advwatres.2013.11.003.
- 687 Cimorelli, L., Cozzolino, L., D'Aniello A., & Pianese, D. (2018). Exact solution of the Linear
 688 Parabolic Approximation for flow-depth based diffusive flow routing, *Journal of Hydrology*689 563,620–632, DOI: /10.1016/j.jhydrol.2018.06.026.
- 690 Covino, T. P., McGlynn, B. L., & Mallard, J. M. (2011). Stream-groundwater exchange and
 691 hydrologic turnover at the network scale, *Water Resources Research*, 47, W12521,
 692 doi:10.1029/2011WR010942
- Dahan, O., Shani, Y., Enzel, Y., Yechieli, Y., & Yakirevich, A., (2007). Direct measurements
- 694 of floodwater infiltration into shallow alluvial aquifers. J. Hydrol. 344, 157–170.

- 695 David, P.-Y., Meire, B., Cary, L., Charlier, J.-B., Lemesnil, M., Ladouche, B., Richard, M., &
- 696 Yecora Zorzano, L. (2016). Étude du fonctionnement et de la vulnérabilité du bassin versant
- de l'Iton État des connaissances, analyse des données et premiers résultats. *BRGM/ RP-*65618-FR report, 401 p. http://infoterre.brgm.fr/rapports/ RP-65618-FR.pdf (accessed April
 2018).
- Dvory, N. Z., Ronen, A., Livshitz, Y., Adar, E., Kuznetsov, M., & Yakirevich, A. (2018).
 Quantification of groundwater recharge from an ephemeral stream into a mountainous karst
 aquifer. *Water*, *10*(1), 79. doi:10.3390/w10010079
- Finch, J. W., Bradford, R. B., & Hudson, J. A. (2004). The spatial distribution of groundwater
- flooding in a chalk catchment in southern England. *Hydrological Processes*, 18: 959–971.
 doi:10.1002/hyp.1340
- Fleischmann, A. S., R. C. D. Paiva, W. Collischonn, M. V. Sorribas, & P. R. M. Pontes
 (2016), On river- floodplain interaction and hydrograph skewness, *Water Resour. Res.*, 52,
 708 7615–7630, doi:10.1002/2016WR019233.
- 709 Fleury, P., Maréchal, J. C., & Ladouche, B. (2013). Karst flash-flood forecasting in the city of
- 710 Nîmes (southern France). *Engineering Geology*, *164*: 26–35. doi:10.1016/j.enggeo.2013.06.007
- Gill, L. W., Naughton O., & Johnston, P. M. (2013). Modelling a network of turloughs in
 lowland karst. *Water Resources Research*, 49(6):3487–3503.
- 713 Gustard, A., A. Bullock, & J.M. Dixon. (1992). Low flow estimation in the United Kingdom.
- 714 Report no. 108. Wallingford, Oxfordshire, UK: Institute of Hydrology
- 715 Habets F., A. Boone, J. L. Champeaux, P. Etchevers, L. Franchisteguy, E. Leblois, E. Ledoux,
- 716 P. Le Moigne, E. Martin, S. Morel, J. Noilhan, P. Q. Segui, F. Rousset-Regimbeau, & P.

- 717 Viennot (2008). The SAFRAN-ISBA-MODCOU hydrometeorological model applied over
- 718 France, Journal of Geophysical Research., 113, D06113, doi:10.1029/2007JD008548.
- 719 Habets F., Gascoin S., Korkmaz S., Thiéry D., Zribi M., Amraoui N., Carli M., Ducharne A.,
- 720 Leblois E., Ledoux E., Martin E., Noilhan J., Ottl'e C. & Viennot P. (2010). Multi-model
- 721 comparison of a major flood in the groundwater-fed basin of the Somme River (France).
- 722 *Hydrology and Earth System Sciences*, 14:99–117.
- Hayami, S., (1951). On the propagation of flood waves. *Disaster Prevention Research Institute Bulletin*, 1: 1-16.
- Horton, R., (1933). The role of infiltration in the hydrologic cycle. *American Geophysical Union Transactions*, 14, 446–460
- House, A. R., Thompson, J. R., Sorensen, J. P. R., Roberts, C., & Acreman, M. C. (2016).
 Modelling groundwater/surface water interaction in a managed riparian chalk valley wetland. *Hydrological Processes*, 30: 447–462. doi: 10.1002/hyp.10625.
- Hughes DA, & Sami K (1992). Transmission losses to alluvium and associated moisture
 dynamics in a semiarid ephemeral channel system in southern Africa. *Hydrol. Process.* 6:45–
 53
- Hughes, A. G., Vounaki, T., Peach, D.W., Ireson, A.M., Jackson, C.R., Butler, A.P.,
 Bloomfield, J.P., Finch, J. & Wheater, H.S. (2011). Flood risk from groundwater: examples
 from a Chalk catchment in southern England. *Journal of Flood Risk Management, 4*: 143–
- 736 155. doi: 10.1111/j.1753-318X.2011.01095.x
- Hunter, N. M., Bates, P.D., Horritt, M.S., & Wilson, M.D., (2007). Simple spatiallydistributed models for predicting flood inundation: a review. *Geomorphology* 90, 208–225.

- Jothityangkoon, C., & M. Sivapalan (2003). Towards estimation of extreme floods:
 Examination of the roles of runoff process changes and floodplain flows, *J. Hydrol.*, 281(3),
 206–229. doi.org/10.1016/S0022-1694(03)00237-3
- Jourde, H., Lafare, A., Mazzilli, N., Belaud, G., Neppel L., Doerfliger, N., & Cernesson, F.
 (2013). Flash flood mitigation as a positive consequence of anthropogenic forcings on the
 groundwater resource in a karst catchment. *Environmental Earth Sciences*, 71:573–583.
 doi:10.1007/s12665-013-2678-3.
- Korkmaz, S., Ledoux, E., & Onder, H. (2009). Application of the coupled model to the
 Somme river basin. *Journal of Hydrology*, 366 (1–4): 21-34,
 https://doi.org/10.1016/j.jhydrol.2008.12.008.
- Ladouche, B., Dörfliger, N., Pouget, R., Petit, V., Thiéry, D., Golaz, C., Bakalowicz, M.,
 Lachassagne, P., Pinault, J.-L., Durand, V., & Cubizolles, J., (2002). Caractérisation du
 fonctionnement des systèmes karstiques nord-montpelliérains Rapport du Programme
 Buèges 1999-2001. *BRGM/RP-51584-FR report, http://infoterre.brgm.fr/rapports/RP-51584- FR.pdf* (accessed April 2018).
- Ladouche, B., Dörfliger, N., Izac, J.-L., Cubizolles, J., Le Strat, P., Du Couedic, C., Aunay,
 B., & Thomson, P., (2004). Évaluation des ressources en eau des Corbières. Phase 1 :
 Synthèse de la caractérisation des systèmes karstiques des Corbières Orientales. Vol. 2 –
 Caractérisation géologique et hydrogéologique du système karstique du "synclinal du BasAgly". *BRGM/RP-52919-FR report, http://infoterre.brgm.fr/rapports/RP-52919-FR.pdf*(accessed April 2018).
- Ladouche, B., Maréchal, J.-C., & Dörfliger, N., (2014). Semi-distributed lumped model of a
 karst system under active management. *Journal of Hydrology*, 509, 215–230.
 doi:/10.1016/j.jhydrol.2015.02.003.

- Lange, J. (2005). Dynamics of transmission losses in a large arid stream channel, *Journal of Hydrology*, 306 (1-4):112-126.
- Lanini, S., Caballero, Y., Seguin, J.-J., & J.-C. Maréchal, (2016). ESPERE—A MultipleMethod, *Groundwater*. 54(2): 155-156, doi: 10.1111/gwat.12390.
- 767 Lopez-Chicano, M., Calvache, M.L., Martin-Rosales, &W., Gisbert J (2002). Conditioning
- factors in flooding of karstic poljes—the case of the Zafarraya polje (South Spain). *Catena*49(4):331–352.
- 770 Mallard, J., B. McGlynn, & T. Covino (2014). Lateral inflows, stream-groundwater exchange,
- and network geometry influence stream water composition, Water Resources Research, 50,
- 772 4603–4623, doi:10.1002/2013WR014944.
- Maréchal, J.C., Ladouche, B., Dörfliger, N. (2008). Karst flash flooding in a Mediterranean
 karst, the example of Fontaine de Nîmes. *Engineering Geology*, 99(3–4):138–146.
 doi:10.1016/j.enggeo.2007.11.013.
- Martin, J. B. & Dean, R.W. (2001). Exchange of water between conduits and matrix in the
 Floridan aquifer, *Chemical Geology*, 179: 145–165, https://doi.org/10.1016/S00092541(01)00320-5.
- Morel-Seytoux, H., 1978. Derivation of equations for variable rainfall infiltration. *Water Resources Research*, 14 (4), 561–568.
- Morris, S.E., Cobby, D., Zaidman, M. & Fisher, K. (2015). Modelling and mapping
 groundwater flooding at the ground surface in Chalk catchments. *Journal of Flood Risk Management*. doi: 10.1111/jfr3.12201
- Moussa R. (1996). Analytical Hayami solution for the diffusive wave flood routing problem
 with lateral inflow. *Hydrological Processes* 10(9): 1209-1227.

- Moussa, R. & Bocquillon, C. (1996a) Criteria for the choice of flood routing methods in
 natural channels, *Journal of Hydrology*, 186: 1–30.
- Moussa R., & Bocquillon C. (1996b). Algorithms for solving the diffusive wave flood routing
 equation. *Hydrological Processes* 10 (1): 105–124.
- 790 Moussa R., & Bocquillon C. (2009). On the use of the diffusive wave for modelling extreme
- flood events with overbank flow in the floodplain. *Journal of Hydrology*, 374: 116-135,
 doi:10.1016/j.jhydrol.2009.06.006.
- Moussa R., & Majdalani S. (2019). Evaluating lateral flow in an experimental channel using
 the diffusive wave inverse problem. *Advances in Water Resources* 127: 120-133.
 Doi:10.1016/j.advwatres.2019.03.009.
- Moussa, R., Voltz M., & Andrieux, P., (2002). Effects of the spatial organization of
 agricultural management on the hydrological behaviour of a farmed catchment during flood
 events. *Hydrological Processes* 16: 393-412. doi: 10.1002/hyp.333.
- Moussa, R., & Cheviron, B. (2015). Modeling of Floods—State of the Art and Research
 Challenges. In: Rowiński P., Radecki-Pawlik A. (eds) *Rivers Physical, Fluvial and Environmental Processes. GeoPlanet: Earth and Planetary Sciences. Springer, Cham.* doi:
 10.1007/978-3-319-17719-9_7.
- Naughton, O., Johnston, P.M., Gill. L.W., (2012). Groundwater flooding in Irish karst: The
 hydrological characterisation of ephemeral lakes (turloughs). *Journal of Hydrology*. 470–471:
 805 82-97.
- Naulin, J.-P. (2012). Modélisation hydrologique distribuée pour la prévision des coupures de
 routes par inondation, application au département du Gard. *PhD Thesis, Ecole Centrale de Nantes*, 290p.

- Payn, R. A., Gooseff, M. N., McGlynn, B. L., Bencala, K. E., & Wondzell, S. M. (2009).
 Channel water balance and exchange with subsurface flow along a mountain headwater
 stream in Montana, United States, *Water Resources Research*, 45, W11427,
 https://doi.org/10.1029/2008WR007644.
- Philip, J., 1957. The theory of infiltration: 4. Sorptivity and algebraic infiltration equations. *Soil Science* 84 (3), 257–264.
- 815 Pinault, J.-L., N. Amraoui, & Golaz C. (2005). Groundwater-induced flooding in macropore-
- 816 dominated hydrological system in the context of climate changes, *Water Resources Research*,
- 817 41, W05001, doi:10.1029/2004WR003169.
- 818 Ponce, V.M., (1990). Generalized diffusive wave equation with inertial effects. *Water*819 *Resources Research* 26 (5), 1099–1101.
- Price, M., Low, R.G., & McCann, C. (2000). Mechanisms of water storage and flow in the
 unsaturated zone of the Chalk aquifer. *Journal of Hydrology* 233:54-71.
- Rak, G., Kozelj, D. & Steinman, F. (2016). The impact of floodplain land use on flood wave
 propagation. *Nat Hazards*, 83: 425. doi.org/10.1007/s11069-016-2322-0
- Robins N. S., & Finch J. W. (2012). Groundwater flood or groundwater-induced flood?
 Quarterly *Journal of Engineering Geology and Hydrogeology*, 45:119–122. DOI:
 10.1144/1470-9236/10-040.
- 827 Ruehl, C., Fisher, A. T., Hatch, C., Huertos, M. L., Stemler, G., & Shennan, C. (2006).
- B28 Differential gauging and tracer tests resolve seepage fluxes in a strongly-losing stream,
 B29 *Journal of Hydrology*, 330:235–248, https://doi.org/10.1016/j.jhydrol.2006.03.025.
- 830 Sholtes J.S., & Doyle M.W. (2011) Effect of channel restoration on flood wave attenuation. J
- 831 Hydraul Eng 137(2):196–208

- 832 Singh, V. P. (2002). Is hydrology kinematic?, *Hydrological Processes*, 16, 667–716, doi:
 833 10.1002/hyp.306.
- Soil Conservation Service-USDA, (1972). Estimation of direct runoff from storm rainfall, *National Engineering Handbook. Section 4-Hydrology*, 10.1–10.24.
- 836 Sorman, A.U., Abdulrazzak, M.J., & Seytoux, M.H.J. Groundwater recharge estimation from
- ephemeral streams. case study: Wadi Tabalah, Saudi Arabia. *Hydrol. Process.* 1997, 11,
 1607–1619.
- 839 Thiéry, D., Amraoui, N., & Noyer, M.-L., (2018). Modelling flow and heat transfer through
- 840 unsaturated chalk Validation with experimental data from the ground surface to the aquifer.
- 841 Journal of Hydrology, 556: 660-673. doi.org/10.1016/j.jhydrol.2017.11.041
- 842 Todini E. (1996). The ARNO rainfall-runoff model. *Journal of Hydrology*, 175: 339-382.
- 843 Trigg, M.A., Wilson, M.D., Bates, P.D., Horritt, M.S., Alsdorf, D.E., Forsberg, B.R., & Vega,
- 844 M.C., (2009). Amazon flood wave hydraulics. *Journal of Hydrology* 374(1):92–105
- 845 Upton, K. A., & Jackson, C. R., (2011). Simulation of the spatio-temporal extent of

groundwater flooding using statistical methods of hydrograph classification and lumped

- parameter models. *Hydrological Processes* 25: 1949–1963. DOI: 10.1002/hyp.7951.
- 848 Valentova J, Valenta P, & Weyskrabova L (2010) Assessing the retention capacity of a
- 849 floodplain using a 2D numerical model. J Hydrol Hydromech 58(4):221–232

846

- 850 Vázquez- Suñé, E., Capino, B., Abarca, E., & Carrera, J. (2007). Estimation of Recharge
- from Floods in Disconnected Stream- Aquifer Systems, Groundwater, 45(5):579-589.
- 852 Yu, B., Sombatpanit, S., Rose, C. W., Ciesiolka, C. A. A., & Coughlan, K. J. (2000).
- 853 Characteristics and Modeling of Runoff Hydrographs for Different Tillage Treatments, Soil
- *Science Society of America Journal*, 64, 1763–1770, doi: 10.2136/sssaj2000.6451763x.

856 Figure captions

Figure 1: Lateral outflow (a) and inflow (b) during floods; two examples of combined cases are also given (c); dark and light blue colours are used to differentiate water levels before and during the flood, respectively, in surface water and groundwater (dashed lines).

860 Figure 2: Diffusive wave model on a reach without lateral flow (a), and with uniformly 861 distributed lateral flows along a channel reach according to two cases: gains (b) and losses (c). 862 The black curve Q_{If} depicts the input hydrograph, and the blue curve Q_{Of} the output one at the 863 end of the reach. The direct approach of the DWE is used to route Q_{If} at the end of the reach 864 without lateral exchanges Q_{Ifr} (dashed grey curve). In the case of lateral flows, the inverse 865 approach is used to simulate lateral flow Q_{Af} (dotted red curve), which is positive for lateral 866 inflow into the reach (b), or negative for lateral outflow from the reach (c). Terms E, E_D and 867 E_A are attenuation and/or amplification terms explained in the text (Section 2.3).

868 Figure 3: Sensitivity analysis of the inverse problem of the DWE to simulate lateral flow (Q_{Af} 869 - red dotted line) for various parameterization sets of C (in m/s) and D (in m^2/s). The analysis 870 is based on two theoretical mono-peak flood events used as input (Q_{If} - black curve) and 871 output (Q_{Of} - blue curve) on a 500-m-long reach and at a computed time step of 120 s. For 872 gaining and losing reaches, respectively, a) and b) show the effect of varying C and fixed D; 873 c) and d) the effect of varying D and fixed C; and e) and f) the effect of varying D and 874 calibrating C so that peakflow time of the routed input $(Q_{Ifr}$ – dashed grey line) is in phase 875 with the output $(Q_{Of}$ - blue curve) one.

Figure 4: a) Location of the Iton River in France (karst aquifers map from Chen et al., 2017);
b) Hydrogeological map of the Iton basin, and c) Scheme illustrating the main lateral surface
flows (light blue colour) and lateral groundwater flows (dark blue colour) along the karst
reach of the Iton River (adapted from Charlier et al., 2015a; David et al., 2016).

Figure 5: Daily input $Q_I(t)$ (a) and output $Q_O(t)$ (b) time series for the 2000-2001 hydrological cycle along the reach delimited by the two gauging stations at Bourth and Normanville. (c) Input-output relationships for the base component Q_b (grey squares) and flood component Q_f (red circles) are shown for a 2-day delay, corresponding to the mean peakflow delay.

Figure 6: Baseflow at the Normanville gauging station *vs.* groundwater level at daily time steps for six piezometer wells in the Iton catchment, showing a best correlation for Coulonges piezometer (right bottom).

Figure 7: Effect of (a) initial soil humidity and (b) of initial karst saturation on the inputoutput peakflow relationships ($Qx_O vs. Qx_I$, respectively). Soil saturation is expressed from the HU2 index (n.d.) and karst saturation from the groundwater depth 'z' below ground level (m BGL) in the Coulonge piezometer well; circle size is proportional to the initial saturation value. The discharge threshold for overbank flow is indicated for each station.

Figure 8: Hydrological time series and simulated lateral flow (C = 0.35 m/s and D = 500 m²/s) during the flood event of 06 January 2001. From top to bottom, rainfall P (at input station P_I and for lateral catchment P_A), soil humidity index HU2, groundwater depth in the Coulonges piezometer well, discharge for the flood component, and total discharge.

896 Figure 9: Boxplot of the *C* parameter calibrated for various diffusivity *D* values (n=33 flood
897 events)

Figure 10: Outflow intensity (Qn_{A_f}) *vs.* input peakflow (Qx_{I_f}) of the flood components, for different diffusivity values.

Figure 11: Peakflow attenuation generated by diffusion E_D vs. peakflow amplification or attenuation generated by lateral exchanges E_A . Positive E_A values indicate amplification due to lateral inflow, but compensated by a highest attenuation due to diffusion E_D when circles are below the E_D =- E_A line.

| Symbols | Dimensi | Definitions |
|--|------------------|---|
| | on | |
| * | - | convolution operator |
| С | $[L.T^{-1}]$ | flood wave celerity |
| D | $[L^2.T^{-1}]$ | flood wave diffusivity |
| E | $[L^3.T^{-1}]$ | Difference of peakflows Qx_{If} - Qx_{Of} |
| $E_{A,}E_{D}$ | $[L^3.T^{-1}]$ | Difference of peakflows linked to the lateral flows, and to the |
| | | hydraulic properties of the channel, respectively |
| HU2 | - | Soil humidity index |
| Ι | - | Input station |
| Κ | - | Hayami kernel function |
| l | [L] | length of the channel |
| 0 | - | Output station |
| Р | [L] | total rainfall |
| р | [T] | time memory of the system |
| <i>q</i> | $[L^2.T^{-1}]$ | lateral flow per unit length |
| Q, Q_b, Q_f | $[L^3.T^{-1}]$ | discharge, base, and flood components of discharge |
| \bar{O} | $[L^3.T^{-1}]$ | mean flow discharge for a rectangular section |
| O_{I}, O_{Ih}, O_{If} | $[L^{3}.T^{-1}]$ | discharge, base, and flood components at the input station <i>I</i> , |
| | | respectively |
| $Q_{Ifr,} Q_{Ir}$ | $[L^3.T^{-1}]$ | routed Q_{If} , and routed Q_{I} , respectively |
| Qn_{Af} | $[L^3.T^{-1}]$ | maximum intensity of lateral outflow |
| $Q_{O}, Q_{O b}, Q_{O f}$ | $[L^3.T^{-1}]$ | discharge, base, and flood components at the output station O, |
| 5 | | respectively |
| $Q_{A}, Q_{Ab}, Q_{Ab}, Q_{Af}, Q_{Afr}$ | $[L^3.T^{-1}]$ | discharge, base, and flood components of lateral exchanges, |
| v v | | respectively |
| Q_{Afr} | $[L^3.T^{-1}]$ | routed Q_{Af} , |
| $Qx_{Af}, Qx_{I}, Qx_{If}, Qx_{If}, Qx_{Ifr},$ | $[L^3.T^{-1}]$ | peakflow of $Q_{A\beta}$, $Q_{I\beta}$, $Q_{I\beta}$, Q_{Ifr} , Q_{O} , and of $Q_{O\beta}$ respectively |
| Qx_{O}, Qx_{Of} | | |
| t | [T] | time |
| X | [L] | length along the channel |
| Z. | [L] | Groundwater depth |
| λ | [L] | time |
| ϕ | - | function related to C and Q_{Afr} |
| τ | [T] | time |
| | | |
| | | |