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Does the simple dynamical systems approach provide useful information about catchment hydrological functioning in a Mediterranean context? Application to the Ardèche catchment (France)

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Abstract

This study explores how catchment heterogeneity and variability can be summarized in simplified models, representing the dominant hydrological processes. It focuses on Mediterranean catchments, characterized by heterogeneous geology, pedology, and land use, as well as steep topography and a rainfall regime in which summer droughts contrast with high-rainfall periods in autumn. The Ardèche catchment (south-east France), typical of this environment, is chosen to explore the following questions: (1) can such a Mediterranean catchment be adequately characterized by simple dynamical systems approach and what are the limits of the method under such conditions? (2) What information about dominant predictors of hydrological variability can be retrieved from this analysis in such catchments?

In this work we apply the data-driven approach of Kirchner (WRR, 2009) to estimate discharge sensitivity functions that summarize the behavior of four sub-catchments of the Ardèche, using non-vegetation periods (November–March) from 9 years of data (2000–2008) from operational networks. The relevance of the inferred sensitivity function is assessed through hydrograph simulations, and through estimating precipitation rates from discharge fluctuations. We find that the discharge-sensitivity function is downward-curving in double-logarithmic space, thus allowing further simulation of discharge and non-divergence of the model, only during non-vegetation periods. The analysis is complemented by a Monte-Carlo sensitivity analysis showing how the parameters summarizing the discharge sensitivity function impact the simulated hydrographs. The resulting discharge simulation results are good for granite catchments, found to be predominantly characterized by saturation excess runoff and sub-surface flow processes. The simple dynamical system hypothesis works especially well in wet conditions (peaks and recessions are well modeled). On the other hand, poor model performance is associated with summer and dry periods when evapotranspiration is high and low-flow discharge observations are inaccurate. In the Ardèche catchment, inferred precipitation rates agree well in timing and amount with observed gauging

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obtained through this approach are simple, with a limited number of parameters that can be estimated from the available data.

Kirchner (2009) represents a catchment with a simple bucket dynamical model where system parameters are derived directly from detected streamflow fluctuations during recession periods. He based his analysis on storage–discharge relationships with one essential assumption: discharge depends only on the total water stored in the catchment. This approach allows the derivation of a first-order non-linear differential equation for simulating rainfall–runoff behavior. Until now, this approach has mostly been applied in small, humid catchments. Kirchner (2009) obtained good results for the Severn (8.70 km²) and Wye (10.55 km²) catchments at Plynlimon, in Mid-Wales. Teuling et al. (2010) also applied this approach to the prealpine Rietholzbach catchment (3.31 km²) getting good results in wet periods and poor model performance during dry periods. The recent study of Brauer et al. (2013) showed similar results for the Dutch lowland Hupsel Brook catchment (6.5 km²) where discharge results were correctly reproduced only in certain periods. Krier et al. (2012) applied the concept of doing hydrology backwards to infer spatially distributed rainfall rates in the Alzette catchment (1092 km²) in Luxembourg, and found that introducing a soil moisture threshold led to model improvement, especially under the wet conditions. However, they did not simulate hydrographs.

To our knowledge, the method has not been evaluated in a Mediterranean context, where the rainfall regime exhibits strong contrasts between dry conditions in summer and intense rainfall events, often related to stationary Mesoscale Convective Systems (Hernández et al., 1998), during autumns. The area is also characterized by heterogeneous topographical, vegetation and geology conditions. The study of the water cycle in such Mediterranean conditions, as well as a better understanding and modelling of processes triggering flash floods, are central research topics addressed in the HyMeX¹

¹www.hymex.org

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(Hydrological Cycle in the Mediterranean Experiment, Drobinski et al., 2013) program and in the FloodScale² project (Braud et al., 2014), to which this study contributes.

Our study area is the Ardèche catchment (2388 km², see location in Fig. 1), which is typical of Mediterranean catchments with highly variable rainfall, steep slopes, and heterogeneous geology and pedology. It is one of the studied catchments of the Cévennes-Vivarais Hydro-Meteorological Observatory (OHM-CV, Boudevillain et al., 2011). Previous studies in this catchment mainly focused on flood forecasting and discharge quantile estimation. Discharge time series from the Ardèche catchment were used to assess the value of new observations in estimating extreme quantiles, such as information derived from paleofloods (Sheffer et al., 2002); historical floods (Lang et al., 2002; Naulet et al., 2005) or post-flood survey peak discharge estimates (Gaume et al., 2009). Flood forecasting studies extended to the whole Cévennes-Vivarais region are numerous and include work by Sempere-Torres et al. (1992), Duband et al. (1993), Le Lay and Saulnier (2007), Saulnier and Le Lay (2009), Trambly et al. (2010), and Garambois et al. (2013). Use of distributed hydrological models for process understanding during flash floods in the Cévennes-Vivarais region is more recent. Examples of such studies are those of Bonnifait et al. (2009), Manus et al. (2009), and Braud et al. (2010). Those studies use a reductionist approach to gain insight into active hydrological processes during floods and highlight a lack of data or parameter information.

As a complementary approach to the modeling studies mentioned above, we adopt in this study the data-based approach proposed by Kirchner (2009) to estimate the hydrological water balance of the examined Ardèche catchment and to gain insight into the dominant associated processes. The idea is to later on, use this insight to propose modeling simplifications with very few parameters to learn more about hydrological functioning at the catchment scale.

In the present paper, we focus on the following questions: (1) what is the applicability of Kirchner method and what are its limitations in a Mediterranean type catchment

²<http://floodscale.irstea.fr/>

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like the Ardèche with its particular conditions (size, climate, geological and pedological heterogeneity, and use of data from operational networks)? (2) What can we learn about dominant hydrological processes using this methodology?

To answer those questions, we first estimate the discharge sensitivity function using the available discharge data. Then we assess the relevance of the obtained function by testing how well the simple model based on it can simulate observed discharge, and can retrieve rainfall. The study is complemented by examining the sensitivity of the results to variations in the parameters of the discharge sensitivity function.

2 Field site and data

2.1 The Ardèche catchment

The Ardèche catchment is located in southern France (Fig. 1). The catchment has an area of 2388 km², and the Ardèche river itself has a length of 125 km. There are two main tributaries in the Ardèche basin: the Baume and Chassezac Rivers, which join the Ardèche River close to one another. Elevation ranges from the mountains of the Massif Central (highest point: 1681 m) in the northwest, to the confluence with the Rhone River (lowest point: 42 m) in the southeast.

The main lithologies found in the Ardèche are schist, granite, and limestone (Fig. 2). Upstream, the Ardèche flows from west to east in a deep granite valley, then flows through basalt formations and schist in a north-south direction. Downstream, it flows through bedded and massive limestone before flowing into the Rhone River (see for example the description provided by Naulet et al., 2005).

Among the land use types found in the Ardèche, forest dominates throughout the basin (Corine Land Cover database³). Forest is represented by a mix of coniferous (27%), broadleaf (13%) and Mediterranean trees (17%). Shrubs and bushes are also well represented in the catchment, occupying a significant portion of the area (17%).

³<http://sd1878-2.sivit.org/>

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We also distinguish significant areas of bare soil in the central and southern part of the Ardèche, as well as a few small urban areas and areas of early and late crops.

In the Ardèche basin, there is a strong influence of the Mediterranean climate with seasonal heavy rainfall events during autumn. Historical data have shown that these events usually lead to flash floods (Lang et al., 2002). These authors mention seven rainfall events locally exceeding 400 mm during the 1961–1996 period. They also comment on the relatively quick flow response (a couple of hours) to precipitation due to the steepness of the upstream part of the catchment and presence of granitic and basaltic rocks.

Figure 3 shows the average hourly regime of the main terms of the water balance equation for all Julian days between 2000–2008. The hydrological year consists mainly of two periods. There is a rainy season (September–February) with maximum precipitation intensity in autumn, characterized by rainfall amounts greatly exceeding reference evapotranspiration ET_0 (calculated based on the SAFRAN reanalysis of Quintana-Seguí et al. (2008): see next section), and by high discharge. On the other hand, during the dry season (March–August), on average ET_0 is much larger than precipitation and runoff is low. Evapotranspiration is influenced by the seasonal cycle of the vegetation, which is particularly marked in the Ardèche catchment, which is mostly covered by forests (around 60 % of the total catchment area, with 27 % of the forest being coniferous and thus remaining green even in winter).

2.2 Available data and first data consistency analysis

2.2.1 Observations used in the study

In the Ardèche catchment, measurements of the hydrological state variables have mainly been started in the 1960s for the purpose of flood forecasting. In our study, we use hourly data of precipitation (P), reference evapotranspiration (ET_0) and discharge (Q) from the period 1 January 2000 until 31 December 2008. These data come from

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operational networks, and not from research catchments as in previous applications of the Kirchner method, which renders the study more challenging.

The analysis is mostly constrained by the availability of discharge data. The latter are obtained from the national Banque Hydro web-site (www.hydro.eaufrance.fr) and Electricité de France. For our study, we need discharge data that are not influenced by human activity, as Kirchner's method assumes mass conservation. Unfortunately, numerous dams and hydro-power stations are located in the upper parts of the Ardèche and Chassezac catchments (Fig. 1). These dams are also used to regulate the water level throughout the year, in particular to ensure a sufficient discharge in the river for recreational use in the summer period. As data to reconstruct the "natural" discharges are not available, we had to discard several gauging stations located downstream of the dams.

As the stations were not designed and managed for low-flow measurements, the low-flow time series were investigated by contacting the operational services in charge of the stations. Consequently, two stations had to be removed from further analysis due to unreliable measurements and agriculture water withdrawals in summer periods. Ultimately, four sub-catchments could be examined: the Ardèche at Meyras (#1), the Borne at Nicolaud Bridge (#2), the Thines at Gournier Bridge (#3), and the Altier at Goulette (#4); see locations in Fig. 1. These four sub-catchments are characterized by steep slopes ($> 15\%$), average altitude of around 1000 m and igneous and metamorphic rock formations. We have also computed Strahler stream order and channel length using TauDEM tools (Tarboton et al., 2009) in order to classify and measure the size of the river network. The analysis was conducted using the 25 m resolution IGN DTM and the D8 flow direction algorithm, so the resulting network statistics may only loosely resemble those that would be obtained from more accurate procedures such as field mapping. Main physiographic catchment characteristics are summarized in Table 1.

The discharge data were available at varying time intervals, and were aggregated to hourly sums. Two types of precipitation data have been examined and are used throughout the analysis. Local rain gauges at the hourly time step provided by the

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OHM-CV data base (Boudevillain et al., 2011) are used as the primary source of rainfall data for the catchment Ardèche at Meyras (#1). For the catchments Borne at Nicolaud Bridge (#2), Thines at Gournier Bridge (#3) and Altier at Goulette (#4), we use the SAFRAN reanalysis of Météo-France, based on 8 by 8 km² grids (Quintana-Seguí et al., 2008) since the local gauging station shows either lack of data and time gaps, or there is no rain gauging station in the catchment (e.g. Thines at Gournier Bridge (#3)). These data are calculated as catchment averages at hourly time steps. To compute the reference evapotranspiration ET_0 , we also use the climate variables of the SAFRAN reanalysis of Météo-France at an hourly time step. ET_0 is calculated using Penman–Monteith formula according to FAO recommendations (Allen et al., 1998).

In our study, we assumed that actual evapotranspiration is equal to potential evapotranspiration (PET) throughout the year, being defined as reference evapotranspiration ET_0 modulated by a crop coefficient depending on the nature of vegetation for each catchment (Eq. 1).

$$AET = PET = K_C \times ET_0 \quad (1)$$

We also took into account the seasonal variability of vegetation through the definition of three crop coefficient stages: initial (1 January–1 April), mid-season (15 April–15 October) and late season (1 November–31 December). Periods between initial and mid-season as well as between mid-season and late season are interpolated linearly. The values for the Ardèche catchments were obtained through the FAO database (Allen et al., 1998). For each catchment we determined the cover estimates for each vegetation type (Broad-leaf forest, Mediterranean forest, Coniferous forest, Early crops, Late crops, Shrubs and bushes and Bare soil) and we calculated a weighted average crop coefficient per sub-catchment for each stage (see Table 2).

Reference evapotranspiration ET_0 and ET_0 modulated by the crop coefficient ($K_C ET_0$) over examined period (2000–2008) are given in Table 3.

The strong hypothesis that $AET = PET$ is likely to be more relevant in winter, when there is sufficient water content in the air and soils, than in summer. Nonetheless we

use the assumption as a first rough approximation in order to assess the feasibility of such a simple modeling concept.

2.2.2 Data consistency

To further assess data quality, we tested the consistency of the local rainfall station with SAFRAN data for the Ardèche at Meyras (#1) catchment. The resulting coefficient of determination was 0.99.

For the rest of the sub-catchments, we first assumed that SAFRAN rainfall is representative of the catchment average. However, by looking at the mean annual water fluxes (Table 3) and estimated runoff coefficients, we infer that the mass balances for catchments #2, #3 and # 4 are implausible.

For these reasons, two actual evapotranspiration (AET) estimates and runoff coefficients are provided to gain useful insight about data uncertainty. The first evapotranspiration estimate comes from the water balance $AET_{WB} = P - Q$, where P is the average annual precipitation and Q the annual runoff. The second estimate corresponds to Turc (1961) annual actual evapotranspiration, which is calculated using the following formula:

$$AET_{Turc} = \frac{P}{\sqrt{0.9 + \frac{P^2}{L^2}}} \quad (2)$$

where P is annual precipitation in mm yr^{-1} and $L = 300 + 25T + 0.05T^3$ (T is the average annual temperature in $^{\circ}\text{C}$). Table 3 also presents two different runoff coefficients. The first runoff coefficient (C) is calculated as the ratio between Q and P whereas the second runoff coefficient (C_{Turc}) is calculated using the following equation:

$$C_{Turc} = \frac{P - AET_{Turc}}{P} \quad (3)$$

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where C_{Turc} is the runoff coefficient, P is precipitation (mm yr^{-1}) and AET_{Turc} is the actual Turc evapotranspiration (mm yr^{-1}). We use AET_{Turc} in this formula along with precipitation in order to estimate annual runoff coefficients in the examined catchments.

The values of the water balance components differ from catchment to catchment as illustrated in Table 3. In addition, the mass balance AET_{WB} and Turc AET estimates are only consistent for the Ardèche at Meyras (#1) catchment; at the other three sites they differ greatly, leading to inconsistent runoff coefficients for the same catchment. This suggests that either the rainfall or ET_0 (or possibly both) are not representative at the other catchments.

Discharge data uncertainty has been addressed in many works and sometimes it can be quite large, especially in catchments where high flows are seldom gauged due to safety reasons (Le Coz et al., 2010) or where low flows may be difficult to measure accurately. Nevertheless, here we decided to go ahead with the available operational discharge data, to assess if the Kirchner method can provide useful information about catchment hydrological functioning in a Mediterranean context, even in the presence of some uncertainty in the discharge data.

However, in order to apply the Kirchner method with data where water balance closure is more representative, we rescaled precipitation and $K_C \text{ET}_0$ values for catchments (#2, #3 and #4). Our re-scaling scheme (see next section for more details) assumes that the discharge data were accurate enough for the application of the Kirchner method, which relies mainly on discharge data.

2.3 Rescaling of water balance fluxes

The first step in the rescaling analysis was to obtain a robust estimate of actual evapotranspiration.

We used following equation of Fu (1981) to draw Budyko (1974) type curves for the Ardèche catchments:

$$\frac{AET}{P} = 1 + \frac{ET_0}{P} - \left[1 + \left(\frac{ET_0}{P} \right)^w \right]^{\frac{1}{w}} \quad (4)$$

where AET/P is the evapotranspiration ratio, ET_0/P is the dryness index, AET/ET_0 is the evapotranspiration efficiency, P/ET_0 is the wetness index and w is a catchment parameter.

The parameter w was empirically derived by Fu (1981) and it can have values from $[1 \sim \infty]$. Zhang et al. (2004) defined parameter w as a coefficient representing “the integrated effects of catchment characteristics such as vegetation cover, soil properties and catchment topography on the water balance”.

In our study, we drew Fu curves with parameter w ranging between 1.5 and 5 to get an insight about evapotranspiration ratios in the Ardèche. The next step was to compare those curves with mean actual annual evapotranspiration ratios obtained using the Turc (1961), Schreiber (1904), Pike (1964) and Budyko formulae (see Table 4). We note from Fig. 4 that almost all calculated AET/P ratios lie in a range between 1.7 and 3. On the other hand, the AET estimates derived using $AET_{WB} = P - Q$ (cyan color in Fig. 4) for catchments #2, #3 and #4 were found to lie outside the range of values given by the various formulae, highlighting the water balance problem. Finally, to assess and adjust our data sets (P and ET_0), we chose Turc inferred evapotranspiration as representative for future analysis since it depends only on precipitation and temperature, and not on potential evapotranspiration as other formulae do (see Table 4).

We then make the following assumptions. We assume that the long-term average Q is valid. We also assume that the “relative” day-to-day variations of $K_C ET_0$ and P are valid, but that the mean P does not reflect the whole-catchment P , and the mean $K_C ET_0$ does not reflect the mean AET. Therefore the means need to be rescaled to achieve a consistent set of measurements. As mentioned before, we assume that the Turc (1961) formula correctly describes the relationship between average AET and average

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3.1 Estimation of the sensitivity function $g(Q)$

Kirchner (2009) proposed a method for determining non-linear reservoir parameters for a simple bucket model with the assumption that discharge Q depends uniquely on total water storage S in the catchment. The analysis starts, as many parametric rainfall–runoff models do, with the water balance equation where the total catchment storage variation is estimated using:

$$\frac{dS}{dt} = P - \text{AET} - Q \quad (5)$$

where S is water storage volume [L] and P , AET, and Q are rates of precipitation, actual evapotranspiration, and discharge, respectively [L T^{-1}]. Q , P , AET and S are considered as functions of time and considered to be averaged over the whole catchment (Kirchner, 2009).

It is known that precipitation measurements are spatially variable. Rain gauges reflect precipitation on areas much smaller than the catchment itself. The same comment is valid for evapotranspiration estimates, which are typically representative of much smaller areas than the catchment.

In Eq. (5), only discharge can be considered as a state variable that characterizes the entire catchment. This observation led Kirchner (2009) to make the fundamental assumption that discharge is uniquely dependent on total water storage S in the catchment, and that therefore:

$$Q = f(S) \quad \text{or} \quad S = f^{-1}(Q) \quad (6)$$

Differentiating Eq. (6) with respect to time, one obtains:

$$\frac{dQ}{dt} = \frac{dQ}{dS} \frac{dS}{dt} = \frac{dQ}{dS} (P - \text{AET} - Q) \quad (7)$$

and differentiating Eq. (7), following Kirchner (2009), one obtains:

$$\frac{dQ}{dS} = f'(S) = f'(f^{-1}(Q)) = g(Q) \quad (8)$$

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where $g(Q)$ is the “sensitivity function” as defined in Kirchner (2009). It describes the sensitivity of discharge to changes in storage, as a function of discharge itself. This is useful because discharge is directly measurable whereas whole-catchment storage is not.

5 Combining Eqs. (7) and (8) we can express $g(Q)$ as (Kirchner, 2009):

$$g(Q) = \frac{dQ}{dS} = \frac{dQ/dt}{dS/dt} = \frac{dQ/dt}{P - AET - Q} \quad (9)$$

where the sensitivity function can be described using precipitation (P), actual evapotranspiration (AET), discharge (Q) and rate of change of discharge (dQ/dt).

10 Following the approach of Kirchner (2009), we consider periods when precipitation and actual evapotranspiration are relatively small compared to discharge, obtaining the following equation, which shows that under these conditions the discharge sensitivity function can be estimated from discharge data alone:

$$g(Q) = \frac{dQ}{dS} \approx -\frac{dQ/dt}{Q} \Big|_{P \ll Q, AET \ll Q} \quad (10)$$

15 We select hourly records for nighttime (defined as a period between sunrise and sunset) during which the total rainfall is less than 0.1 mm within the preceding 6 h and following 2 h (Krier et al., 2012). We also tested larger time windows (10 and 12 h instead of 8 h) which did not improve $g(Q)$ estimation.

20 The sensitivity function $g(Q)$ is estimated using discharge records from non-vegetation periods (from November to March) from 2000 until 2008, when vegetation and ET_0 could be considered to have a smaller impact on stream discharge. The resulting $g(Q)$ function was used for precipitation retrieval and discharge simulation during both non-vegetation and vegetation periods (April–October).

25 We avoid the vegetation period for the estimation of the $g(Q)$ function since, as Fig. 3 shows, during this period ET_0 is much larger than discharge, and the Ardèche catchments clearly respond to ET_0 forcing during the entire 24 h period.

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In addition, in the Ardèche basin, the diurnal amplitude (computed as half the difference between the daily maximum and minimum flow) often exceeds 20 % of the daily average flow.

These rainless nighttime hours are further used to determine the sensitivity function $g(Q)$ by constructing “recession plots” (Brutsaert and Nieber, 1977) of the flow recession rate $(-dQ/dt)$ as a function of discharge. Following Brutsaert and Nieber (1977) and Kirchner (2009), the flow recession rate is estimated as the difference between two successive hours as:

$$\frac{-dQ}{dt} = \frac{(Q_{t-\Delta t} - Q_t)}{\Delta t} \quad (11)$$

Then, the discharge is averaged over those two hours as $(Q_{t-\Delta t} + Q_t)/2$. Binning is then done by grouping the individual hourly data into ranges of Q and then calculating the standard and mean error for $-dQ/dt$ and Q for each bin. Following Kirchner (2009), values of $-dQ/dt \leq 0$ are also included in the binning analysis to avoid the introduction of bias. The bin size was initially set at 1 % of the logarithmic range in Q but was locally increased if necessary to bring the standard error of $-dQ/dt$ down to 50 % of the mean $-dQ/dt$ (Kirchner, 2009).

A quadratic function (Kirchner, 2009) is then fitted to the binned means leading to the following empirical equation in log space:

$$\ln(g(Q)) = \ln \left(-\frac{dQ/dt}{Q} \Big|_{P \ll Q, AET \ll Q} \right) \approx c_1 + c_2 \ln(Q) + c_3 (\ln(Q))^2 \quad (12)$$

3.2 Discharge simulation

Discharge sensitivity functions can be used to simulate discharge (Kirchner, 2009) by combining Eqs. (9) and (10), resulting in the following expression, where the quadratic function of Eq. (12) is used to describe $g(Q)$:

$$\frac{dQ}{dt} = \frac{dQ}{dS} \frac{dS}{dt} = g(Q)(P - AET - Q) \quad (13)$$

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In solving this equation, attention is paid to two details, time lags and numerical instabilities (Kirchner, 2009). A time lag is introduced to account for flow routing delays between changes in catchment storage and changes in discharge at the outlet. Changes in subsurface storage could also lag behind rainfall inputs due to the delays necessary for rainfall to infiltrate and change discharge at the outlet. However, these time lags do not affect the estimation $g(Q)$ since Q and dQ/dt are measured simultaneously at the catchment outlet.

Equation (13) indicates that dQ/dt depends on the balance between precipitation, actual evapotranspiration and discharge. However, variations in $P - AET - Q$ are mainly forced by variations in precipitation. For instance, in the Ardèche at Meyras (#1) catchment, the variance of hourly precipitation is over 15 times larger than the variance of hourly discharge and around 80 times larger than the variance of hourly evapotranspiration. In discharge simulations, lag time is not of such importance since discharge is highly auto-correlated. However, in precipitation retrieval, lag time is taken into account to enhance model performance (see Sect. 3.3 for more details) because precipitation varies more on short time scales.

In order to minimize numerical instabilities, Eq. (13) is solved using its log transform (Kirchner, 2009):

$$\frac{d(\ln(Q))}{dt} = \frac{1}{Q} \frac{dQ}{dt} = \frac{g(Q)}{Q} (P - AET - Q) = g(Q) \left(\frac{P - AET}{Q} - 1 \right) \quad (14)$$

Equation (14) is then computed using fourth-order Runge–Kutta integration, iterating on an hourly time-step. A single value of measured discharge is used to initialize the simulation. In addition, Kirchner (2009) also remarked that solution can be unstable unless the parameter C_3 of Eq. (12) is less than 0.

To estimate the AET term in Eq. (14), Kirchner (2009) originally used Penman–Monteith reference evapotranspiration and a rescaling effective parameter (k_e) that was calibrated for the entire examined period. Other authors have used slight variants of this approach: Teuling et al. (2010) used the Priestley–Taylor equation to estimate

catchment-scale evapotranspiration, defining the evaporation efficiency as a fitting parameter; Brauer et al. (2013) used a parameter f that takes into account the difference between potential and actual evapotranspiration on a monthly basis.

For the application to the Ardèche catchment, as mentioned in Sect. 2.2.1, we assumed that AET was given by Eq. (1) (computed at the hourly time step). According to the catchment, original $K_C ET_0$ (catchment #1) or rescaled $K_C ET_0$ (catchments #2, #3, #4) are used.

To show how data inconsistency problems may affect the performance of discharge simulation, we also run the model with non-rescaled values of precipitation and evapotranspiration. The corresponding performance of the model is reported in Sect. 4.5.

3.3 Rainfall retrieval based on $g(Q)$

Until recently, it was considered infeasible to infer precipitation from stream flow fluctuations. Spatial-temporal variability of precipitation is high and conventional rain-gauges can only measure precipitation over an area that is many orders of magnitude smaller than a catchment itself. We assess the relevance of the inferred storage-discharge relationship for the examined catchments in the Ardèche using the rainfall retrieval scheme (“doing hydrology backward”) as proposed by Kirchner (2009) and further tested by Krier et al. (2012).

Following Eq. (13) that describes the catchment as a simple non-linear dynamical system, we can infer temporal patterns of precipitation rates from streamflow fluctuations using the following equation as outlined by Kirchner (2009):

$$P - AET = \frac{dS}{dt} + Q = \frac{dQ/dt}{dQ/dS} + Q = \frac{dQ/dt}{g(Q)} + Q \quad (15)$$

To apply this concept, one must take account of the travel time lag between changes in discharge from the hillslope and changes in stream flow at the outlet. A time-lag l is

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used as a dimensionless model evaluation statistic indicating how well the simulated discharges fit the observations. We compute the NSE to emphasize the high flows as shown in the following equation:

$$\text{NSE} = 1 - \left(\frac{\sum_{i=1}^n (Y_i^{\text{obs}} - Y_i^{\text{sim}})^2}{\sum_{i=1}^n (Y_i^{\text{obs}} - Y_i^{\text{mean}})^2} \right) \quad (18)$$

5 where Y_i^{obs} is the i th observation of discharge data, Y_i^{sim} is the simulated discharge value for i th time step, Y_i^{mean} is the mean of all observed data and n represents the number of observations.

NSE values range between $-\infty$ and 1.0, with 1 representing the optimal value (e.g. Moriasi et al., 2007, for a recent review of performance criteria). We also computed
10 NSE on the logarithm of the discharge to give less weight to the peaks.

In addition, Percent bias (PBIAS) is also calculated as a part of the model evaluation statistics. It measures total volume difference between two time series, as Eq. (19) indicates:

$$\text{PBIAS} = \left(\frac{\sum_{i=1}^n (Y_i^{\text{obs}} - Y_i^{\text{sim}}) \cdot (100)}{\sum_{i=1}^n (Y_i^{\text{obs}})} \right) \quad (19)$$

15 where Y_i^{obs} is the i th observation of discharge data, Y_i^{sim} is the simulated discharge value for i th time step and n represents the number of observations.

The optimal value of PBIAS is 0.0 where positive values indicate model overestimation bias, and negative values indicate model underestimation bias (e.g. Gupta et al., 1999).

20 In rainfall retrieval, model performance is assessed by using the coefficient of determination (R^2) to quantify the linear correlation between observed and inferred precipitation. R^2 ranges from 0 to 1, where higher values indicate smaller error variance

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(e.g. Moriasi et al., 2007). Although the inversion formula yields individual hourly values (Eq. 14), we use daily averages to compute R^2 . This is done to reduce the effects of small discrepancies in timing that become less consequential when R^2 is calculated on a daily time step (Kirchner, 2009).

3.5 Sensitivity analysis

In this part, we performed a Monte-Carlo analysis to sample the parameter space defined by the three parameters C_1 , C_2 and C_3 and investigate further whether the values derived from stream flow fluctuations are representative, and how these parameters impact streamflow simulations. This Monte-Carlo sensitivity study was conducted for the Ardèche at Meyras (#1) catchment.

A representative set of 10 000 random (C_1 , C_2 , C_3) triplets was selected using Monte-Carlo simulation. Then the discharge was simulated using the model presented in Sect. 3.2 and Eq. (14). We used the Nash Sutcliffe efficiency (ln for low flow and linear for high flows) as likelihood measures of the similarity between the simulated and observed discharge. Then we verified that the parameter set derived from data is in the range of the sets leading to the best agreement between model and observations. The number of simulations (10 000) was assumed to be adequate because the best-fit NSE did not change significantly beyond 10 000 simulations. Besides, using 10 000 simulations is considered to be acceptable in view of the relative simplicity of the parametric model. For comparison, Zhang et al. (2008) and Tekleab et al. (2011) used 20 000 simulations for a four-parameter dynamic water balance model, and Uhlenbrook et al. (1999) used more than 400 000 model runs for the much more complex HBV model with 12 parameters.

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4 Results

The results section is divided into five parts. In the first part results concerning estimation of $g(Q)$ function and its sensitivity analysis are given. Then we present the assessment of the relevance of this estimated $g(Q)$ function by examining the accuracy of the simulated discharge (Sect. 4.2) and retrieved precipitation (Sect. 4.3). In Sect. 4.4, the impact of parameter variations on the simulated hydrographs and results of the Monte-Carlo simulations are shown. Finally, the results with non-scaled original data are presented in Sect. 4.5.

4.1 $g(Q)$ estimation

Figure 5 shows an example of a recession plot for the Ardèche at Meyras (#1) catchment for the all non-vegetation periods between 2000 and 2008.

We observe that the recession plot exhibits large scatter at low discharge. This result is consistent with the findings of Kirchner (2009) and Teuling et al. (2010). They argue that this is possibly due to measurement errors, precision errors, and possible differences between the modeling concept and reality.

Table 5 provides values of the recession plot parameters for all four catchments during non-vegetation periods between 2000 and 2008. It shows one parameter set for each catchment. We observe that our choice of the non-vegetation period for estimation of $g(Q)$ gives consistent results amongst different catchments, with similar values of parameters C_1 and C_2 . We also observe that the C_3 parameter, which controls the downward/upward curving of the $g(Q)$ function, is always negative, ranging from -0.02 up to -0.2 . This is important because Kirchner (2009) obtained realistic simulated discharge only when recession plots are downward-curving on a log-log scale (meaning the C_3 parameter is negative). Eventually, these parameter sets allowed stable discharge simulation as can be seen in Sect. 4.2.

We have also tested $g(Q)$ estimation for all vegetation periods between 2000 and 2008; during these periods, the C_3 parameter tended to be positive. In this case,

when the $g(Q)$ function is extrapolated to very low discharges, very high values of $g(Q)$ are obtained, and thus, numerical instabilities appear that led to model non-functionality. This is also probably due to the distortion of the discharge time series by evapotranspiration as explained in Sect. 3.1.

4.2 Discharge simulations

Continuous discharge simulations were performed for 2000–2008. Figure 6 presents a simulation extract (year 2004) for the Ardèche at Meyras (#1) catchment. Table 6 presents a model performance summary (NSE, NSE calculated on the logarithm of the discharge and PBIAS) for each catchment and each year.

By looking at Fig. 6, we can see that discharge simulations reproduce the observed hydrograph behavior better in winter and non-vegetation periods. The low-flow (summer) periods are less well reproduced, even if the overall performance of the simulation is good. The influence of evapotranspiration in summer periods can be one of the explaining factors for that. It should be noted that high evapotranspiration influence is visible only when discharge is evaluated in log space. In linear space, evapotranspiration has a negligible influence on (already quite small) discharge, and the model runs well under dry conditions.

We note in Table 6, that the Ardèche at Meyras (#1) catchment shows satisfactory performance with $NSE = 0.68$, $NSE \log = 0.74$ and PBIAS of 7.9% for the nine-year simulation period. Unsatisfactory performance is observed for year 2005 yielding $NSE = -0.15$, $NSE \log = 0.07$ and PBIAS of 62.2%. Year 2005 in general can be characterized as a dry year with annual precipitation of 775 mm and annual reference evapotranspiration of 947 mm for this catchment. A mean annual precipitation across the examined period (2000–2008) of 1621 mm and mean evapotranspiration of 809 mm clearly confirms that year 2005 can be considered as “dry”. Furthermore, Gupta et al. (1999) show that PBIAS values for streamflow tend to vary more as compared to other performance criteria, in dry than in wet years. This could be another possible explanation to the overall poor model performance in 2005 for the Ardèche

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and #3 in summer periods (red line) and consequently for total examined period too (green line).

The optimized time lags are generally very small (less than 2 h), which confirms the very short response time in the Ardèche catchment. In order to see whether the retrieved daily rainfalls were sensitive to the lag time, we compared the results obtained with different lag times for two catchments: the Ardèche at Meyras (#1) and Altier at Goulette (#4). The Ardèche at Meyras (#1, 98 km²) has an optimized lag time of 2 h. We tested the retrieval behavior with lag times of 1 and 2 h and we observe almost no change in the performance (Table 7): we obtain the same coefficient of determination of 0.41 and a bias of 7.9 mm d⁻¹ at a lag time of 1 and 2 h. Similar results are obtained for the Altier at Goulette (#4) catchment, where we observed a slightly better precipitation modeling performance with lag time of 1 h ($R^2 = 0.72$) rather than with a lag time of 2 h ($R^2 = 0.71$).

4.4 Sensitivity analysis

4.4.1 Impact of parameter variations on the simulated hydrographs

As a first approach, a manual sensitivity analysis was done by successively varying the values of each parameter and plotting the corresponding simulated hydrographs. The results for the Ardèche at Meyras (#1) catchment (year 2004) are presented; see Figs. 8 and 9 for C_3 and C_1 parameters, respectively. The results for the parameter C_2 are not presented here since this parameter only varies slightly when estimated from non-vegetation periods in each year (see Sect. 4.1) and the results are graphically quite similar to those for the parameter C_1 (but peaks are less affected). The NSE values of log discharge are also calculated (Table 8).

We can see that C_3 seems to be influential during the low-flow summer period and also during recessions of events following low-flow periods (Fig. 8). However, it does not play a significant role in the peaks and in well-established high-flow conditions. In contrast, the C_1 parameter has an important influence on the whole hydrograph

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(Fig. 9), including the peaks. Low values of C_3 tend to flatten the model response, causing overestimated low-flow values and underestimated peaks.

From Table 8 we can also observe that the model efficiency for the parameter values that were obtained from the recession plots is close to optimal (at least for this year at this site), and cannot be substantially improved by manual parameter adjustments.

4.4.2 Exploration of parameter range using Monte-Carlo simulations

In order to complement to manual sensitivity analysis presented above, to explore the range of these parameters and to assess whether the parameters of the $g(Q)$ function-derived from data analysis are representative, we performed Monte-Carlo simulations using the model described by Eq. (14) and randomly sampling the three parameters C_1 , C_2 and C_3 . The parameters were sampled randomly from the a priori defined parameter range given in Table 9. For each simulation, the NSE and NSE log (on the log of discharge) were calculated to assess the “performance” of the parameter set. The results are presented using dotty plots for the Ardèche at Meyras (#1) catchment in Fig. 10. Table 9 also indicates the range of “behavioral” values for each parameter as derived from the dotty plots, defined as the range where NSE is higher than 0.7, along with the values derived from the recession plots.

The results show that when the parameters are calibrated to discharge simulations, their ranges are quite large. The maximum model performance appears to be around 0.8 for all three parameters and both indicators. Low-flow performance (NSE log) is not very sensitive to the variations of the parameters. Giving peak flow more weight (NSE) allows the identification of clear optima and a narrower range for the C_1 parameter. Concerning the C_2 parameter, although the initial guess of the parameter range was quite narrow (see Table 9), the final “optimized” range is almost the same, with no clear optimum. For the C_3 parameter, the final “optimized” range is found to be the half of the initial one. These two parameters appear thus to be not very sensitive, although the sign of the C_3 parameter was already identified as a key element of successful discharge simulations. Finally, the parameter values obtained from recession

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plots are in the optimized parameter range, thus suggesting that the analysis of discharge recessions is sufficiently informative and that there is no need of additional model calibration for discharge simulation. Beven and Binley (1992) have argued that having too many parameters increases the degrees of freedom beyond what data can properly deal with; this results in having different sets of parameters that give similar results (the equifinality problem). Figure 10 shows that this is not the case, suggesting that model is not over-parameterized. More importantly, however, our analysis shows that the recession plots yield parameter estimates that are consistent with (and arguably better constrained than) parameter values obtained from conventional model calibration methods.

4.5 Modeling performance with non-scaled original data

In Sect. 2.3 we introduced a rescaling technique to obtain more representative water balances for catchments #2, #3 and #4. Here, we show the consequences of foregoing this rescaling for those three catchments that showed unrealistic mass balances (Table 3). Figure 11 shows observed discharge and simulated hourly hydrographs for the Altier at Goulette (#4) catchment for the year 2000, obtained with non-scaled data, rescaling of precipitation alone, and rescaling of both precipitation and evapotranspiration.

The lack of water balance closure may contribute substantially to poor model performance, as can be seen from Fig. 11. We observe that when the original non-scaled data are used, discharge is generally underestimated. By introducing the re-scaled precipitation, better peaks reproduction can be obtained, but model performance is still poor during the vegetation period. Eventually, by using in addition the re-scaled evapotranspiration, significantly better results are obtained in both vegetation and non-vegetation periods.

The simple dynamical systems approach, like many modeling approaches, is based on conservation of mass; it is therefore unsurprising that it may perform poorly when tested against data sets that violate mass conservation.

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catchments varied from 16 km² to 103 km² and model performance was not correlated to the size of the catchment. The model performance in the relatively large catchments is as good as in the smaller catchments. Krier et al. (2012) report that when this approach is used for “doing hydrology backward” and retrieving rainfall amounts, the model performance in larger basins is as good as or sometimes even better than in smaller catchments. Kirchner (2009) also addressed this issue arguing that the approach was unlikely to work for catchments that are too big (e.g. more than 1000 km²). This is due to the lag times required for changes in discharge to reach the outlet; in such large catchments these lag times would be so long and variable that the model would be likely to fail. In addition, the theory presented here could not be expected to work in the catchments that are bigger than the scale of individual storms (Kirchner, 2009). Suggestions for how to deal with large river basins are given in Sect. 6.

5.1.2 Data quality

Our study demonstrates that data quality is particularly important for the application of this method. Concerning discharge data, the method is based on the discharge-sensitivity function $g(Q)$, and discharge disturbances consequently will lead to biases in the appraisal of the catchment functioning. In the present study, we use discharge data from operational networks. We have shown in Sect. 2.2 that there are known issues with the quality of these data for our purposes. Nevertheless, when data consistency is sufficient (e.g. Ardèche at Meyras (#1) station), a robust estimation of the $g(Q)$ function from non-vegetation periods can be obtained, leading to accurate simulation of the discharge.

Furthermore, the quality of rainfall data was questioned at the early beginning of our work, and re-scaling of precipitation was needed to obtain realistic results. The gridded SAFRAN product is known to underestimate precipitation in mountainous areas and to underestimate the occurrence of strong precipitation ($P > 20 \text{ mm day}^{-1}$, Quintana-Seguí et al., 2008; Vidal et al., 2010). Some authors tried to overcome this problem by

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interpolating the SAFRAN data across altitude bands (Etchevers et al., 2001; Lafaysse et al., 2011; Thierion et al., 2012) but these data were not available for the present study. In addition, SAFRAN re-analyses are based on existing rain gauges. In mountainous areas, rain gauges are rare and the existing stations, generally located in the plain, may not capture the increase of rainfall with altitude (Molinié et al., 2011). It would be interesting to assess the performance of the rainfall retrieval using more accurate rainfall estimates as reference. As reference daily rainfall, we propose to use the SPAZM reanalysis (Gottardi, 2009), which improves rainfall estimation in mountainous area, when it becomes available to us.

Assuming that discharge data is reliable, it was shown that when input rainfall and ET_0 consistent with the water balance closure are used, the discharge simulated using the $g(Q)$ function is much more accurate than with the original input data.

Work is currently in progress in order to quantify the rainfall–runoff data accuracy. For discharge data, this work is based on the BaRatin method (Le Coz et al., 2013) which provides an uncertainty range on the estimated discharge. The uncertainty can be propagated to the whole discharge time series and the next step will be the propagation to the hydrological water balance and the quantification of uncertainty for the annual and monthly values. This work will help quantify which of the data (rainfall, discharge or both) need to be improved.

In addition, the operational discharge measurement network has recently been complemented by research instrumentation covering nested scales (see Braud et al. (2014) for details). In particular, small catchments ranging from 0.5 to 100 km² have been monitored continuously since 2010. The data set was not long enough to be used in the present study, but these new data are expected to be of higher accuracy than the operational data used in this study, so that they can provide additional insight into the hydrological response of the catchments.

Regarding discharge uncertainty, if data have to be rescaled, an approach like the one proposed by Yan et al. (2012) should be preferred, as it allows a consolidation of the water balance at the scale of the whole Ardèche catchment, taking into account

data uncertainties on all the components, and constraining the results with the water balance equation along the river network.

The simulation results show that additional effort must be put into quantifying data uncertainty in both discharge and rainfall. The derivation of more accurate rainfall fields combining various data sources (such as radar data and in situ gauges; see, for instance, Delrieu et al., 2013) should also be encouraged. It could also be interesting to use actual evapotranspiration estimates derived from remote sensing techniques adapted to complex topography (e.g., Gao et al., 2011; Seiler and Moene, 2010) to obtain independent estimates of AET and better constrain this component in hydrological modeling.

5.2 Adequacy of Kirchner method in our catchment

The sampling strategy of Kirchner's method to derive the $g(Q)$ functions from non-vegetation periods appeared to be adequate in our case. We estimated $g(Q)$ by using the streamflow data from non-vegetation periods of the 9 year time series (2000–2008) and then used the resulting parameterization to reproduce the hydrographs (continuous simulations) for the rest of the 9 year interval. This procedure can be understood as a “differential split-sample test” (Klemes, 1986) where the 9 year-long period encompasses different seasonal precipitation variations including wet and dry periods. The results show that the information retrieved from only a fraction of the discharge time series is relevant also for periods with very different characteristics.

Independently from the data quality issues, we also showed that the Kirchner model performs better during the wet, winter periods than the dry, summer periods and dry years (see Sect. 4.2). We interpret these results as an indication that the current model is not fully adapted to the high evapotranspiration conditions of our Mediterranean catchments. The method is therefore less reliable when discharge is low, especially in summer. This is one limitation of the Kirchner method for dry catchments.

In addition, the recent study of Brauer et al. (2013) showed that the two-parameter model they used, cannot deal with complexity of hydrological processes in their

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catchment (only 39% of the hydrographs had NASH over 0.5). In the Ardèche catchments, however, the three-parameter model succeeds in capturing the catchment behavior, with quite good response of discharge to rainfall in non-vegetation periods (peaks and recession were nicely reproduced).

5.3 Catchment functioning hypotheses derived from the analysis

The most important output from our application of the simple dynamical systems approach is the validation of underlying hypotheses and information about the dominant processes that can be derived from the model parameterization.

5.3.1 General considerations

The Kirchner model is based on an underlying hypothesis that regards a catchment as a single nonlinear bucket model. The good performance of the model in each sub-catchment suggests that this theory, although it was developed for humid regions, remains valid for these Mediterranean sub-catchments. We can thus interpret that these sub-catchments do follow the model's functioning hypotheses, especially in winter and non-vegetation periods. These results are coherent with findings of Brauer et al. (2013) for the Hupsel Brook catchment, Kirchner (2009) for Plynlimon and Teuling et al. (2010) for the Rietholzbach catchment. In contrast, during the vegetation period the model seems to be less adapted to our Mediterranean setting. The catchments seem to behave differently when they are dry. This is probably due to the strong influence of the evapotranspiration conditions. Our results suggest the existence of another storage, probably more superficial than the "Kirchner" storage which could be used to supply evapotranspiration with shorter time scales, and which may be largely decoupled from groundwater seepage that sustains base flows.

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5.3.2 Links with physiographic characteristics of the catchments

The model works better in the Ardèche at Meyras (#1) and Thines at Gournier Bridge (#3) catchments, which both have the same geological formation, granite (see Fig. 2). The hypothesis of saturation excess runoff and subsurface flow makes particular sense in this geology (e.g. Cosandey and Didon-Lescot, 1989; Trambly et al., 2010). Unaltered bedrock tends to be impermeable, but flow pathways are created in the many fractures, joints and fissures of the altered horizons. During extended rainfall those flow pathways might become connected, generating rapid subsurface flow (Krier et al., 2012). Moreover the parameter values of the granite catchments are quite similar (see Table 5).

Geology thus appears to be a predictor of the hydrological variability in the Ardèche basin. This is also consistent with the contemporary literature, as geology has been invoked in numerous recent studies as a controlling factor of flood response (Gaál et al., 2012; Garambois et al., 2013; Krier et al., 2012; Vannier et al., 2013). As also discussed by Kirchner (2009), the theory is challenged by catchments with heterogeneous geology and thus with many unconnected subsurface storage reservoirs. This might explain the good modeling performance in granite catchments (see also Vannier et al. (2013) for similar conclusions using a reductionist modeling approach).

6 Conclusion and perspectives

Our study describes in detail the application of Kirchner's methodology to four catchments of the Ardèche basin ranging from 16 to 103 km², typical of the Mediterranean environment.

To have more representative water balance fluxes, we re-scaled precipitation and evapotranspiration for three sub-catchments (#2, #3 and #4). In our work we used average annual scaling coefficients for the whole time-series (for precipitation and evapotranspiration). Eventually, varying this scaling coefficient according to different seasons

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could possibly lead to a better approximation of hourly precipitation and evapotranspiration fluxes.

We calculated the discharge sensitivity functions from non-vegetation periods and performed continuous discharge simulations with an hourly time step for the period 2000–2008. We also inferred precipitation and performed sensitivity analyses of the three parameters of the discharge sensitivity function.

Our results show that good results for discharge simulation can be obtained, especially under winter humid conditions and for catchments characterized by predominantly granitic lithology. Under dry conditions, poor model performance is mainly related to the disturbed water balance terms, high influence of AET and imprecise discharge measurements. Improving AET estimation is recommended for better model performance in summer periods when evapotranspiration is high and when the unsaturated zone has a significant role in attenuating the precipitation input. Working on the quantification of data accuracy and error reduction is also recommended in order to get more robust and reliable results.

As a perspective to this study, dominant predictors of runoff variability other than geology (such as land use, soil properties, drainage density, topographic steepness etc.) still need to be explored and linked to catchment hydrological behavior. Relating the obtained parameters of the discharge sensitivity function to the catchment characteristics using different statistical classification techniques (e.g. Principal Component Analysis (PCA) and Factor Analysis of Mixed Data (FAMD) or Self-Organized Maps) could allow us to apply the method also to ungauged basins, thus contributing to the PUB initiative (Hrachowitz et al., 2013). Another step would be then to create a distributed “Kirchner type” hydrological model where a parameter set would be attributed to “regions” discretized on the basis of their physiographic characteristics. This would allow us to determine the rainfall–runoff behavior in large scale river basins by taking into account the precipitation spatial distribution and flood flow routing through the channel network. We would then be able to broaden our understanding of non-linear catchment response and travel time lags as suggested by Kirchner (2009).

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Table 1. Physiographic characteristics of the four examined Ardèche sub-catchments. Strahler stream order, channel length and drainage density are calculated from the 25 m IGN DTM using TauDEM tools (Tarboton et al., 2009).

Catchment ID	#1	#2	#3	#4
River and catchment name	Ardèche at Meyras	Borne at Nicolaud Bridge	Thines at Gournier Bridge	Altier at Goulette
River name	Ardèche	Borne	Thines	Altier
Drainage area (km ²), <i>A</i>	98.43	62.6	16.73	103.42
Average altitude (m)	898.54	1113	892.75	1149.13
Average slope (%)	23.43	20.13	16.72	17.13
Forest cover (%)	68	68	51	42
Strahler stream order	4	3	3	5
Channel length (km), <i>L</i>	94.31	59.26	13.51	97.38
Drainage density (km km ⁻²), <i>D = L/A</i>	0.96	0.95	0.81	0.94

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Table 2. Weighted average crop coefficient for each examined catchment per growing stage.

Catchment name	Crop coefficient (K_C)		
	$K_{c_initial}$	$K_{c_mid_season}$	$K_{c_late_season}$
The Ardèche at Meyras (#1)	0.74	0.94	0.79
Borne at Nicolaud Bridge (#2)	0.73	0.96	0.80
Thines at Gournier Bridge (#3)	0.68	0.94	0.75
Altier at Goulette (#4)	0.62	0.97	0.75

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Table 3. Hydro-climatic characteristics of the four examined Ardèche sub-catchments (2000–2008).

Catchment ID	#1	#2	#3	#4
Catchment name	Ardèche at Meyras	Borne at Nicolaud Bridge	Thines at Gournier Bridge	Altier at Goulette
Precipitation (mm yr^{-1}), P	1621	1633	1892	1176
Streamflow (mm yr^{-1}), Q	1057	1579	970	932
Runoff coefficient, C	0.65	0.97	0.51	0.79
Actual Evapotranspiration (mm yr^{-1}), $\text{AET}_{\text{WB}} = P - Q$	564	54	922	244
ET_0 SAFRAN (mm yr^{-1})	809	792	860	775
$K_C \text{ET}_0$ (mm yr^{-1})	731	729	762	699
Turc Actual evapotranspiration (mm yr^{-1}), AET_{Turc}	609	505	571	475
Runoff coefficient, C_{Turc}	0.62	0.69	0.70	0.60
Temperature ($^{\circ}\text{C}$), T	11.2	8.0	9.9	7.7
P_{Turc} (mm yr^{-1})	–	2084	1541	1407
Scaling P coefficient	–	1.27	0.81	1.2
Scaling AET coefficient	–	0.69	0.75	0.68
New runoff coefficient, C_n	0.65	0.76	0.63	0.66

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Table 4. Description of different empirical formulas for estimating mean annual actual evapotranspiration (AET is actual evapotranspiration [mm yr^{-1}], P is precipitation [mm yr^{-1}], ET_0 is potential evapotranspiration [mm yr^{-1}], and T is mean air temperature [$^{\circ}\text{C}$]).

Equation	Reference
$AET = P \left[1 - \exp \left(-\frac{ET_0}{P} \right) \right]$	Schreiber (1904)
$AET = \frac{P}{0.9 + \left(\frac{P}{L}\right)^2}$, where $L = 300 + 25T + 0.05T^3$	Turc (1961)
$AET = P / \left[1 + \left(\frac{P}{ET_0} \right)^2 \right]^{0.5}$	Pike (1964)
$AET = \left[P \left(1 - \exp \left(-\frac{ET_0}{P} \right) \right) ET_0 \tanh \left(\frac{P}{ET_0} \right) \right]^{0.5}$	Budyko (1974)

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Table 5. Parameter values for the examined catchments for all non-vegetation periods (2000–2008).

Catchment name (ID)	Non-Vegetation period		
	C_1	C_2	C_3
The Ardèche at Meyras (#1)	−3.74	0.65	−0.2
Borne at Nicolaud Bridge (#2)	−4.08	0.74	−0.15
Thines at Gournier Bridge (#3)	−3.71	0.72	−0.13
Altier at Goulette (#4)	−3.80	0.82	−0.02

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Table 6. Summary statistics of computed NSE, NSE log and PBIAS for each examined catchment in the Ardèche basin.

Year	The Ardèche at Meyras (#1)			Borne at Nicolaud Bridge (#2)			Thines at Gournier Bridge (#3)			Altier at Goulette (#4)		
	NSE linear	NSE log	PBIAS (%)	NSE linear	NSE log	PBIAS (%)	NSE linear	NSE log	PBIAS (%)	NSE linear	NSE log	PBIAS (%)
2000	0.60	0.85	-20.7	0.76	0.83	5.02	0.49	0.86	-18.14	0.53	0.70	-1.58
2001	0.61	0.85	5.7	0.59	0.74	33.56	0.27	0.85	-1.43	0.67	0.62	1.86
2002	0.82	0.82	-1.2	0.63	0.53	-12.77	0.68	0.83	-15.05	0.65	0.44	-17.88
2003	0.76	0.72	13.	0.73	0.63	5.78	0.79	0.82	14.27	0.89	-0.19	12.43
2004	0.69	0.86	5.1	-0.07	0.37	-35.28	-0.26	0.78	-18.38	0.42	0.05	-11.09
2005	-0.15	0.07	62.2	0.66	0.64	18.16	0.21	0.53	48.22	0.70	-0.86	0.04
2006	0.51	0.71	19.6	0.68	0.58	0.58	0.36	0.72	17.67	0.18	-0.61	6.90
2007	0.11	0.67	21.8	0.51	0.28	-23.67	0.30	0.71	24.47	-1.22	0.34	-14.48
2008	0.76	0.85	8.2	0.75	0.43	-9.35	0.69	0.79	-6.89	0.83	0.62	6.04
2000–2008	0.68	0.74	7.9	0.67	0.61	0.75	0.55	0.78	0.98	0.74	0.18	-0.29

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Table 7. Model performance of inferred vs. measured daily rainfall in four sub-catchments for all non-vegetation periods 2000–2008.

Gauging station	R^2	Mean Bias [mm day ⁻¹]	Slope	Time lag [h]
Ardèche at Meyras (#1)	0.41	7.9	1.1	2 (optimized)
Ardèche at Meyras (#1)	0.41	7.9	1.1	1
Borne at Nicolaud Bridge (#2)	0.56	7.4	1.01	2 (optimized)
Thines at Gournier Bridge (#3)	0.61	4.7	1.22	2 (optimized)
Altier at Goulette (#4)	0.71	2	1.09	2
Altier at Goulette (#4)	0.72	2	1.09	1 (optimized)

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Table 8. NSE values of log discharge for the Ardèche at Meyras (#1) catchment, illustrating sensitivity to changes in the C_1 and C_3 parameters.

C_1 parameter [-]	NASH on log of discharge	C_3 parameter [-]	NASH on log of discharge
-4	0.81	-0.3	0.68
-3.8	0.85	-0.25	0.79
-3.74 (from data)	0.86	-0.21	0.85
-3.7	0.86	-0.2 (from data)	0.86
-3.6	0.86	-0.19	0.86
-3.5	0.86	-0.17	0.86
-3.4	0.85	-0.16	0.85
-3.3	0.83	-0.15	0.83
-3.2	0.81	-0.1	0.45
-3	0.71	-0.09	0.26

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Table 9. Comparison of the chosen parameter range and parameters obtained from non-vegetation periods for the Ardèche at Meyras (#1) catchment.

Parameters	Lower/upper bound		
	C_1 [-]	C_2 [-]	C_3 [-]
Parameter range	[-1]–[-6]	[0.1–1]	[-0.001]–[-0.5]
The range of “behavioral” values	[-3.5]–[-4.5]	[0.1–0.9]	[-0.001]–[-0.25]
Reference (from recession plots)	-3.74	0.65	-0.2

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Table 10. Model performance for three examined catchments over the whole examined period (2000–2008), comparing the original operational data and rescaled precipitation and evapo-transpiration data.

Performance	Operational	Rescaled P	Rescaled P and AET
Catchment	Nicolaud Bridge (SAFRAN rain)		
NASH	0.45	0.65	0.67
NASH log	0.58	0.70	0.61
PBIAS [%]	42	14.2	0.75
Catchment	Gournier Bridge (SAFRAN rain)		
NASH	0.36	0.50	0.55
NASH log	0.79	0.62	0.78
PBIAS [%]	−13.8	22	0.98
Catchment	Goulette (SAFRAN rain)		
NASH	0.54	0.79	0.74
NASH log	−4.90	−2.99	0.18
PBIAS [%]	49	23.65	−0.29

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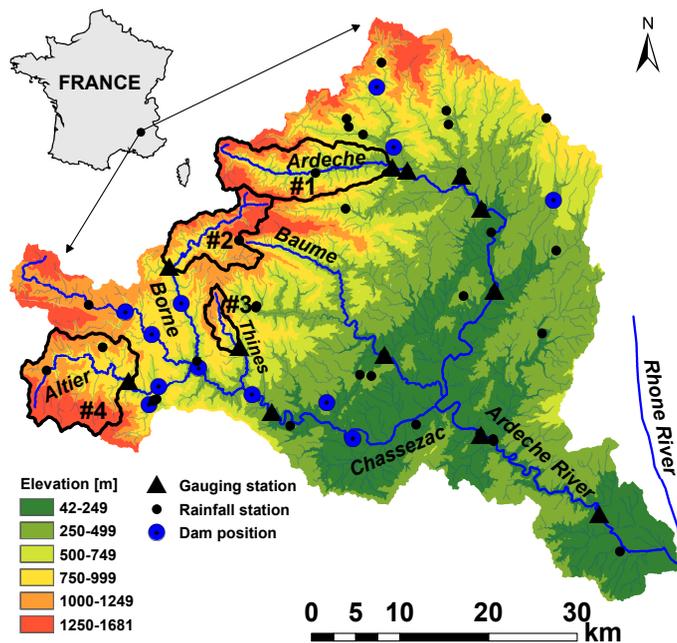


Figure 1. Map of the Ardèche catchment with gauging and rainfall stations, dam locations, and catchments that were examined (in bold): #1. Ardèche at Meyras; #2. Borne at Nicolaud Bridge; #3. Thines at Gournier Bridge; #4. Altier at Goulette.

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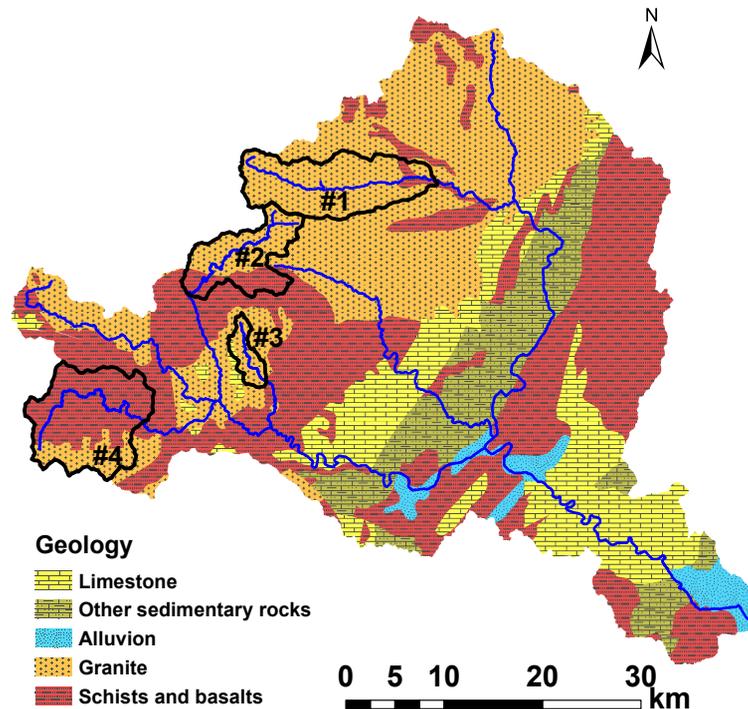


Figure 2. Geological map of the Ardèche catchment (extracted and processed from geological map of France 1 : 1 000 000 issued by BRGM (6th edn., 1996).

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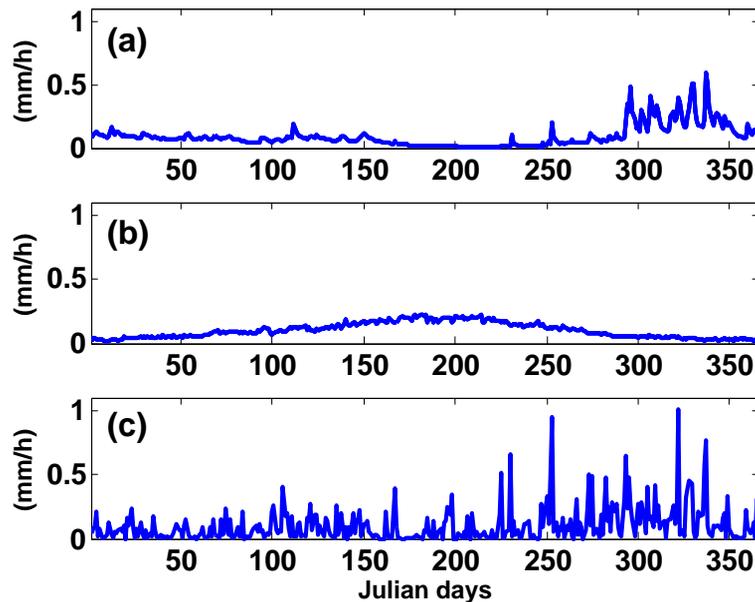


Figure 3. Average hourly discharge (a), reference ET_0 (b) and rainfall (c) in $[\text{mm h}^{-1}]$ at the Ardèche outlet for all Julian days between 2000–2008. (b) and (c) are calculated from the SAFRAN reanalysis.

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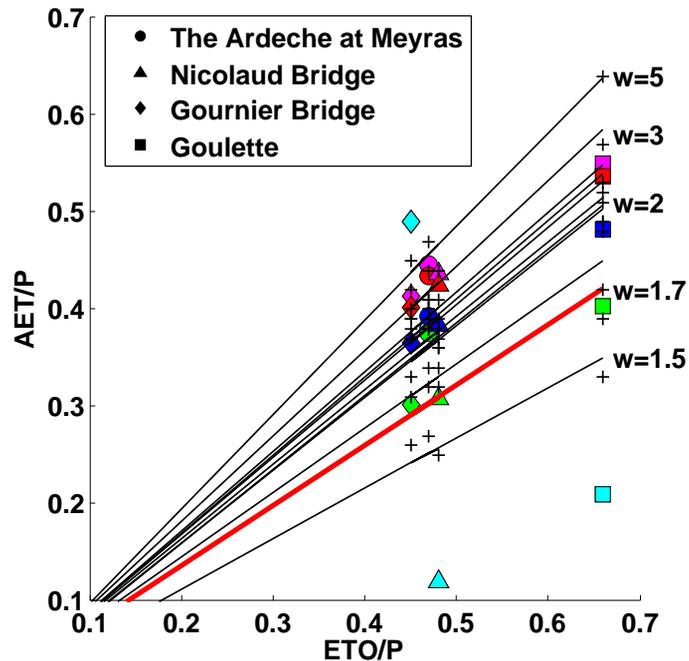


Figure 4. Mean annual evapotranspiration ratio (AET/P) as a function of index of dryness (ET_0/P) for different values of parameter w , using Fu (1981) curve and different formulas (Turc, Schreiber, Pike, Budyko; see Table 4). Colors correspond to different formulas (cyan = original data; green = Turc, blue = Schreiber, pink = Pike, red = Budyko) and shapes represent different examined catchments.

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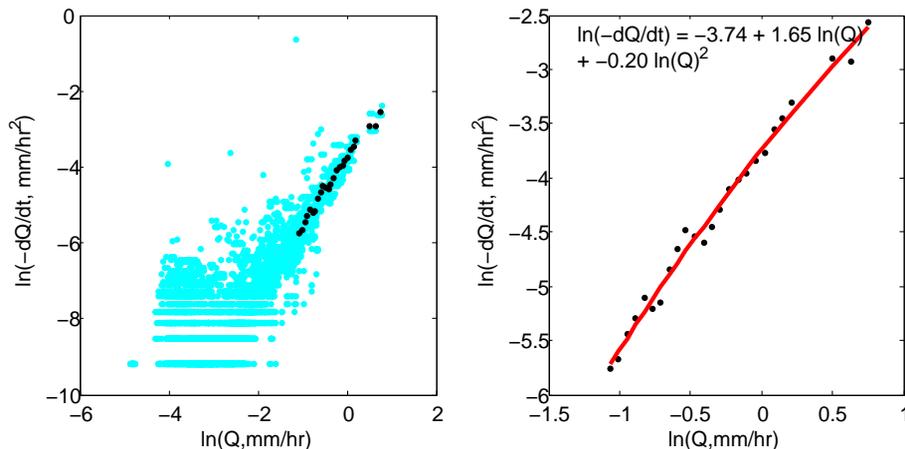


Figure 5. Recession plots for the Ardèche at Meyras (#1) catchment for all non-vegetation periods between 2000 and 2008; (left) Flow recession rates ($-dQ/dt$) as a function of flow (Q) for individual rainless night hours (blue dots) and their binned averages (black dots). (right) Quadratic curve fitting with binned means.

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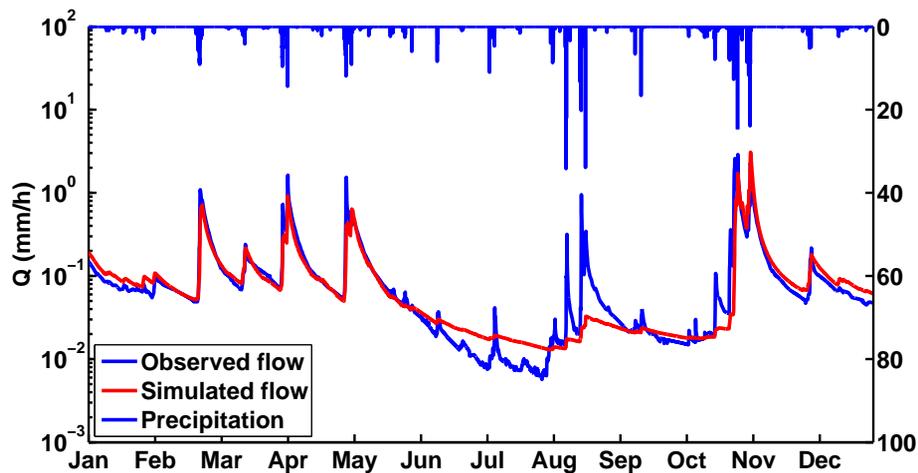


Figure 6. Series of simulated hourly hydrographs (red) for the Ardèche at Meyras (#1) catchment for the year 2004, compared with observed discharge (blue).

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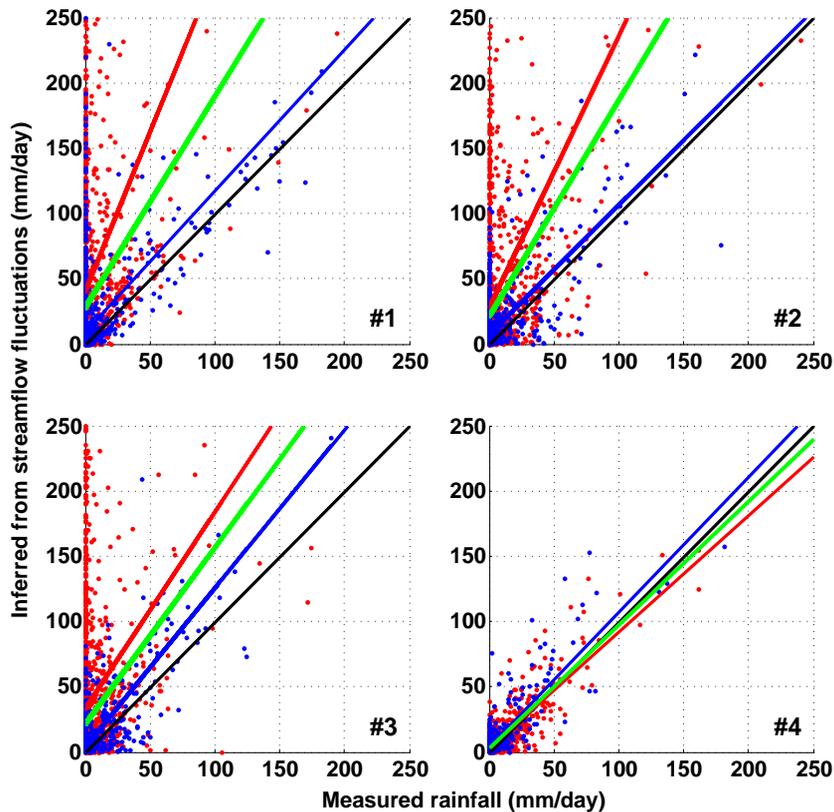


Figure 7. Inferred vs. measured daily precipitation for examined catchments: #1. Ardèche at Meyras; #2. Borne at Nicolaud Bridge; #3. Thines at Gournier Bridge; #4. Altier at Goulette. Blue dots correspond to the inferred daily totals from non-vegetation periods; red points correspond to the inferred daily totals from vegetation periods; blue line is correlation for non-vegetation period, red line for vegetation period and green line for total examined period.

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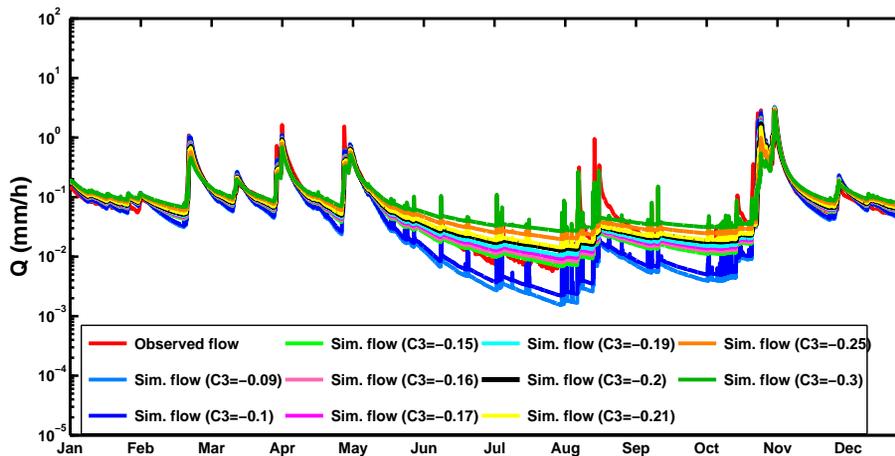


Figure 8. Observed vs. simulated hydrograph for the Ardèche at Meyras (#1) catchment (year 2004), with C_3 parameter variations (C_1 (-3.74) and C_2 (0.65) values are kept constant).

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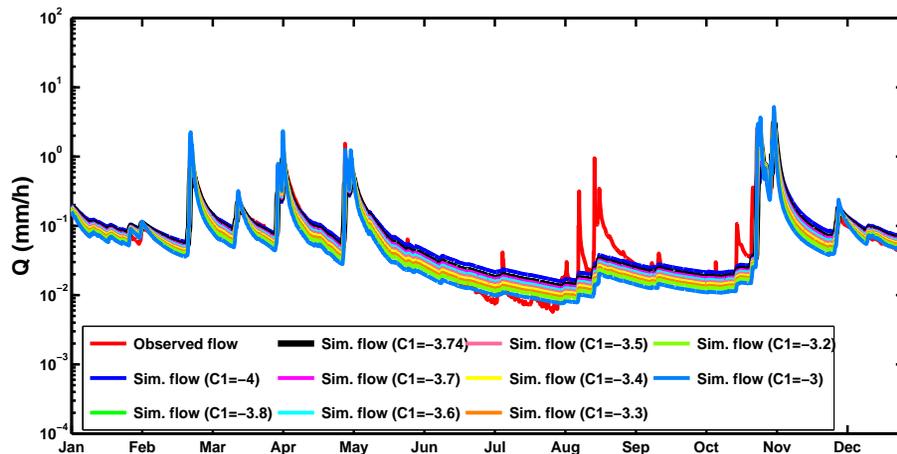


Figure 9. Observed vs. simulated hydrograph for the Ardèche at Meyras (#1) catchment (year 2004) with C_1 parameter variations (C_2 (0.65) and C_3 (-0.2) values are kept constant).

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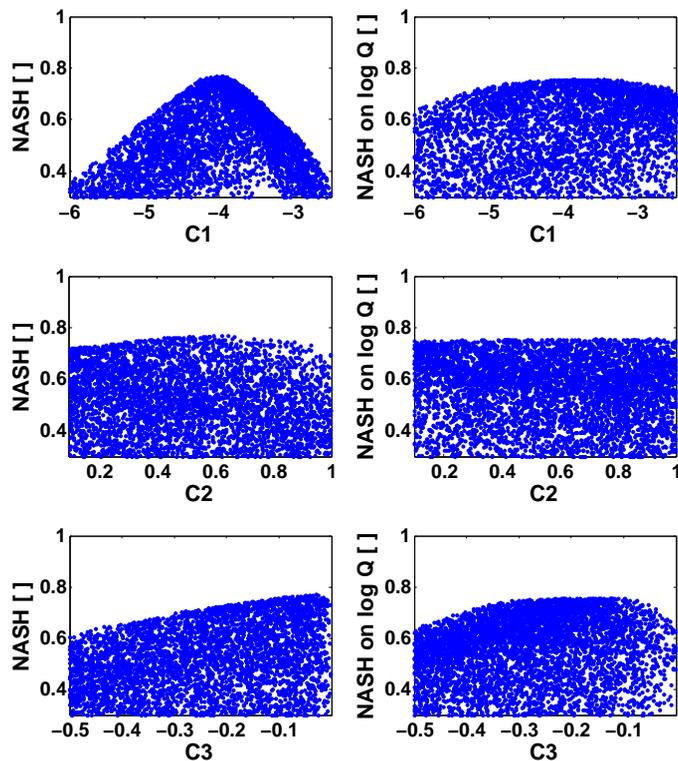


Figure 10. Dotty plots for the Ardèche at Meyras (#1) catchment (left: plots with NASH efficiencies; right: plots with NASH efficiencies calculated on $\log Q$).

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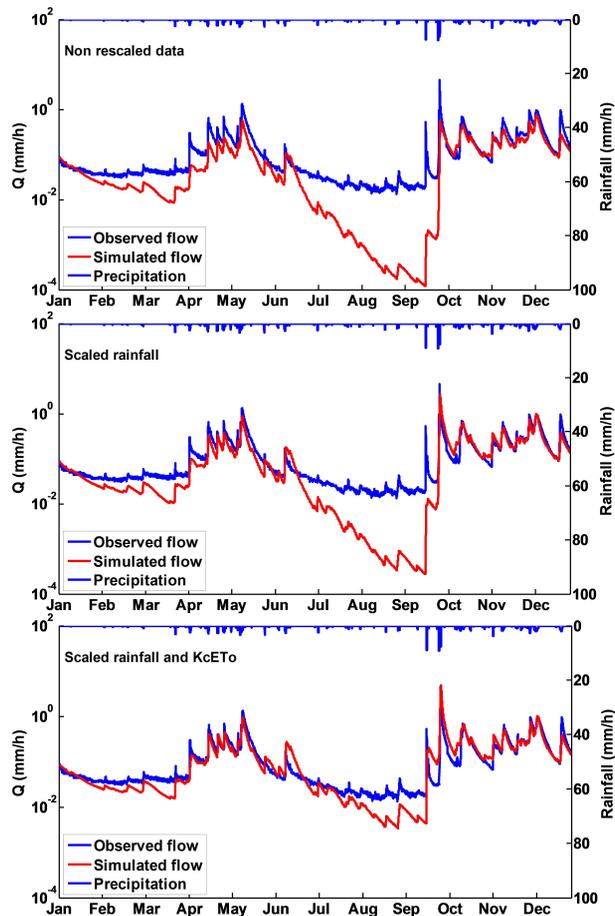


Figure 11. Series of simulated hourly hydrographs (red) for Altier at Goulette (#4) catchment for the year 2000 and its comparison with observed discharge (blue), using original non-scaled data (top), with re-scaled P only (middle), and re-scaled P and $K_{C}ET_0$ (bottom).