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The geomorphic structure of the runoff peak

R. Rigon¹, P. D’Odorico², and G. Bertoldi³

¹Dipartimento di Ingegneria Civile e Ambientale and CUDAM, Università di Trento, Via Mesiano 77, Trento, Italy

²Department of Environmental Sciences, University of Virginia, 291 McCormick Road, P.O. Box 400123, Charlottesville, VA 22904-4123, USA

³Institute for Alpine Environment, EURAC Research, Viale Druso 1, Bolzano, Italy

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Correspondence to: R. Rigon (riccardo.rigon@ing.unitn.it)

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Abstract

This paper develops a theoretical framework to investigate the core dependence of peak flows on the geomorphic properties of river basins. Based on the theory of transport by travel times, and simple hydrodynamic characterization of floods, this new framework invokes the linearity and invariance of the hydrologic response to provide analytical and semi-analytical expressions for peak flow, time to peak, and area contributing to the peak runoff. These results are obtained for the case of constant-intensity hyetograph using the Intensity-Duration-Frequency (IDF) curves to estimate extreme flow values as a function of the rainfall return period. Results show that, with constant-intensity hyetographs, the time-to-peak is greater than rainfall duration and usually shorter than the basin concentration time. Moreover, the critical storm duration is shown to be independent of rainfall return period. Further, it is shown that the basin area contributing to the peak discharge does not depend on the channel velocity, but is a geomorphic propriety of the basin. The same results are found when the effects of hydrodynamic dispersion are accounted for. As an example this framework is applied to three watersheds. In particular, the runoff peak, the critical rainfall durations and the time to peak are calculated for all links within a network to assess how they increase with basin area.

1 Introduction

A number of hydrological analyses require the evaluation of the highest peak-flow values expected to occur with a given return period. Most of the methods addressing this issue – from the simple rational method Mulvaney (1851), Doodge (1957) to the use of distributed rainfall-runoff models (e.g., Beven, 2001) – have been developed with the purpose of providing quantitative predictions of peak flows for engineering applications more than a synthesis of their dependence on the geomorphic and hydrodynamic characteristics of the watershed. To this end, this paper develops a simplified theory

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based on the concepts of geomorphologic instantaneous unit hydrograph (GIUH) and of width function (Rinaldo et al., 1991, 1995; D'Odorico and Rigon, 2003). This theory extends the results of Henderson (1963) and Myninink and Corder (1976) and complements some findings by Robinson and Sivapalan (1997). The goals of the paper include: understanding the geomorphic structure of the highest peak-flow caused by rainfall with given return period; redefining the concept of concentration time within the framework of the GIUH theory; determining the duration of the rainfall able to generate the maximum peak flow under the assigned climatic conditions; determining the time to peak as a function of rainfall and basin characteristics.

1.1 Basic concepts and results

The concept of Unit Hydrograph (Sherman, 1932) is used in the representation of the hydrograph as a sum of the responses to different rainfall inputs observed throughout an individual rainstorm

$$Q(t) = A_T \int_0^t f(t - \tau) p(\tau) d\tau \quad (1)$$

with A_T being the total contributing area, t the time, τ the time counted starting at the beginning of the rainstorm, p the intensity of effective precipitation at time τ , and $f(t)$ the instantaneous unit hydrograph (IUH). The IUH represents the travel time probability density function of a unit amount of water instantaneously injected into the basin (Gupta and Mesa, 1988); $f(t)$ can be determined either through some simple conceptual frameworks (Nash, 1957) or through the geomorphological theory (Rodriguez-Iturbe and Valdes, 1979; Gupta et al., 1980). Depending on the physical hypotheses underlying the different formulations of the IUH, $f(t)$ can be defined either within an infinite or a finite time domain. In the latter case $f(t) = 0$ for $t \geq \tau_c$, where τ_c is the concentration time of the rational method (i.e. the time at which the whole basin contributes to the discharge at the outlet). The integral:

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$$S(t) = \int_0^t f(\tau) d\tau \quad (2)$$

is known in literature as *S*-hydrograph (e.g., Doodge, 2003, p. 86) and represents the cumulative probability distribution of travel times inside the basin ($S(t) = 1$ for $t \geq \tau_c$), i.e. $S(t)$ is the ratio between contributing area at time t and basin area. In this simplified approach the rate of (effective) precipitation is assumed to be constant throughout individual rainstorms of duration t_p :

$$p(t, t_p) \equiv p H(t_p - t) H(t), \quad (3)$$

where $H()$ is the Heaviside step function (i.e. $H(t) = 1$ for $t \geq 0$ and $H = 0$ otherwise). By definition, p can be considered the expected value of the effective intensity of rainfall during the storm. In what follows we will refer to the expected value of p as a first-order approximation of the “real” storm hyetograph. This approach will allow us to obtain semi-analytical results that could be easily generalized to the case of hyetographs with non-constant intensity as suggested by D’Odorico and al. (2005) for the case of landslide-triggering precipitation. The time-to-peak, t^* , can be found either for $t = t_p$, or by solving the following equation, first derived by Henderson (1963) and hereafter called *Henderson’s equation* (see also Appendix A):

$$f(t) = f(t - t_p) \quad t_p < \tau_c. \quad (4)$$

The graphic solution of Eq. (4) is illustrated in Fig. 1. For $t > \tau_c$, $S(t)$ is a constant (hence, $f(t) = dS(t)/dt = 0$), while $S(t - t_p)$ is an increasing function of t (i.e. $f(t - t_p) > 0$). Therefore the solution, t^* , of Eq. (4) needs to be searched in the interval $[t_p, \tau_c]$. Figure 1, shows the existence of a delay, ($\Delta t = t^* - t_p$), between the end of the rainstorm and the occurrence of the flow peak. This delay depends on the characteristics of the IUH and its parameters. The delay, Δt , corresponding to the main (i.e. highest) peak can be easily determined: Fig. 2 shows how Δt is a decreasing function of t_p and becomes null as t_p approaches τ_c . Henderson (1963) and Robinson and Sivapalan (1997) solved Eq. (4) using a triangular hydrograph. However it can be solved

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analytically (Appendix B) in the case of Nash hydrograph (Nash, 1957) and with simple numerical code in the general case, as explained in the following section. If t^* is the time to peak – counted from the beginning of the rainfall – the peak flow, Q_p , is then estimated as $Q(t^*)$ using Eq. (A5) for the case of hyetographs with constant intensity:

$$Q_p = \begin{cases} \rho A_T (S(t^*) - S(t^* - t_p)) = \rho A^* & 0 \leq t_p \leq \tau_c \\ \rho A_T S(\tau_c) = \rho A_T & t_p > \tau_c \end{cases} \quad (5)$$

with t^* being a function of t_p (through Eq. 4) and $S(t^*)$ the fraction of contributing area at $t = t^*$. If the duration is smaller than the concentration time, the contributing area at $t = t^*$ is $A^* = A_T [S(t^*) - S(t^* - t_p)]$.

1.2 Extreme values of peak flows

The maximum peak flow occurring after a rainstorm with a certain return period, t_r , (hereafter referred to as extreme peak flow) can be determined by expressing ρ in Eq. (5) as a function of duration, t_p , and return period, t_r . These curves provide a statistical representation of the most severe rainfall conditions for a certain geographic location and return period. Thus, the extreme peak-flow discharge, Q_p (Eq. 5), depends on t_p also through ρ . Because for any given return period, $\rho = \rho(t_p|t_r)$ is a decreasing function of t_p and $S(t^*) = S(t_p + \Delta t)$ is an increasing function of its argument, there is a particular duration, t_p^* , which maximises the peak-flow discharge. Such a duration needs to be shorter than the concentration time, τ_c . This critical duration can be found by solving the equation $dQ_p/dt_p = dQ(t_p + \Delta t)/dt_p = 0$, where $\Delta t = \Delta t(t_p)$ is a smooth function of t_p . We will indicate with Δt^* the value of $\Delta t(t_p)$ calculated for $t_p = t_p^*$. The first order-derivative of Eq. (5a) becomes

$$\frac{dQ(t^*)}{dt_p} = A_T \left\{ \rho'(t_p^*|t_r) [S(t_p^* + \Delta t^*) - S(\Delta t^*)] + \rho(t_p^*|t_r) [f(t_p^* + \Delta t^*) (1 + \Delta'(t)) - f(\Delta t^*) \Delta'(t)] \right\} = 0 \quad (6)$$

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where $\rho'(t_p^*|t_r)$ is the first-order derivative of the $\rho(t_p^*|t_r)$ with respect to t_p and the first-order derivative of $S(t)$ has been expressed as the IUH (i.e. $S'(t_p + \Delta t) = f(t_p + \Delta t)$). Substituting Eq. (4) into Eq. (6) we obtain

$$\frac{\rho'(t_p|t_r)}{\rho(t_p|t_r)} = - \frac{f(t_p + \Delta t)}{S(t_p + \Delta t) - S(\Delta t)}. \quad (7)$$

which is equivalent to the main equation in Myninink and Corder (1976), though it is here derived for any shape of the IUH. We use intensity-duration-frequency (IDF) curves to relate rainfall intensity. Scaling models of IDF curves are often based on power laws

$$\rho(t_p|t_r) = a(t_r) t_p^{-m} \quad (8)$$

where $a(t_r)$ is a function of the return period t_r and $0 \leq m \leq 1$ is independent of t_r . Equation (8) in Eq. (7) gives

$$m = \frac{t_p f(t^*)}{S(t^*) - S(\Delta t^*)} \equiv g(t_p^*) \quad (9)$$

where $t^* = t_p^* + \Delta t^*$. Because the dependence on t_r in Eq. (8) is through $a(t_r)$ – which does not appear in Eq. (9) – Eq. (9) implies that, according to this linear theory of the hydrologic response, the critical rainfall duration (t_p^*) associated with extreme runoff peak flow values, is independent of the return period. The same result applies also to the case of self-similar design storm hyetograph (e.g., Burlando and Rosso, 1996) and to the design storm hyetograph suggested in Hershfield (1961) and commonly used in the engineering practice in the US since Bell (1969).

The solution of the system given by Eqs. (9) and (4) can be used to determine the critical rainfall duration, t_p^* , and the time to peak, t^* . Figure 3 illustrates a typical dependence of $g(t_p)$ on t_p (with $\Delta t(t^*)$ given by Eq. 4) and provides the graphical solution (i.e., t_p^*) of Eq. (9), for the case with $m = 0.63$. It can be shown that $g(0) = 1$

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and $g(\infty) = 0$, consistently with the common observation that values of m span the interval $[0, 1]$ (m ranges most commonly between 0.5 and 1). For some values of m , it is possible to have multiple solutions of Eq. (9), which correspond to local minima or maxima of discharge.

Once t_p^* is known from Eq. (9), Eq. (5) provides the maximum discharge under a rainfall of assigned return period, t_r , as

$$Q_{\text{MAX}}(t_r) = \rho(t_p^*|t_r) C(t^*, t_p^*) A_T \quad (10)$$

where $C(t^*, t_p^*) \equiv S(t^*) - S(t^* - t_p^*)$. We notice that Eq. (10) is similar to the well known rational method equation Chow et al. (1988); however, the runoff coefficient, C , depends on the effective fraction of contributing area evaluated as a function time to peak, t^* , and on the critical rainfall duration, t_p^* (Eq. 6) rather than on the concentration time. The total contributing area, A_T , can be determined as explained in the following section. We note that in the rational method the coefficient C accounts for effects of "within-storm" rainfall variability, runoff-generation (i.e., C is a runoff coefficient), and runoff routing. Our approach explains only the dependence of C on routing processes, while the effects of "within-storm" rainfall patterns are not investigated. We also note that, unlike the rational method, our theory does not assume values of contributing area and rainfall duration. Rather, both A_T and t_p^* are the outcome of the interplay of basin and climatic characteristics, and are calculated as values associated with the maximum peak flows.

2 The geomorphological analysis of runoff peaks

The geomorphological theory of the hydrologic response (Rodriguez-Iturbe and Valdes, 1979; Gupta et al., 1980; Rinaldo et al., 1991; Rinaldo and and Rodriguez-Iturbe, 1996) provides an interpretation of the IUH, based on the basin morphology and simple dynamical assumptions. In this paper we express the Geomorphologic Instantaneous Unit Hydrograph (GIUH) through a generalization of the width function, $W(x)$,

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(e.g., Shreve, 1969; Kirkby, 1986; Gupta and Mesa, 1988; Brutsaert, 2005). $W(x)$ is the probability distribution of distances, x (measured along the network), between any point in the basin and the outlet. In recent years a number of studies have recognized the soundness of this approach, strengthened its theoretical bases (Rinaldo et al., 1991, 1995; Saco and Kumar, 2002a,b; D'Odorico and Rigon, 2003; Botter and Rinaldo, 2003) and shown its applicability and calibration to small and large catchments (e.g., Naden, 1992; Snell and Sivapalan, 1994; Franchini and OConnell, 1996; Da Ross and Borga, 1997; Naden et al., 1999; Yang et al., 2002; Brutsaert, 2005).

The basin-scale travel time distribution, $f(t)$, can be expressed as a function of $W(x)$

$$f(t) = \int_0^L W(x) f(t|x) dx \quad (11)$$

where $f(t|x)$ is the travel time distribution in a path of length x , and L is the length of the longest drainage path.

2.1 The kinematic case

When the effects of hydrodynamic dispersion are negligible, water is subject mainly to advection (kinematic wave), and the probability distribution of travel times, $f(t|x)$, for the rain falling at a distance x (measured along the flow path) from the outlet is

$$f(t|x) = u \delta \left(t - \frac{x}{u} \right) \quad (12)$$

with u being the flood wave channel celerity and $\delta()$ the Dirac delta-function. Hence, Eq. (12) in Eq. (11) gives:

$$f(t) = u W(ut). \quad (13)$$

In this framework, which generalizes Ross (1921), the concentration time is rigorously $\tau_c = L/u$, with L being the longest drainage path. When the IDF curves are expressed by Eq. (8) and $f(t)$ by Eqs. (13), (4) and (9) become:

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$$W(u t) = W(u (t - t_p)) \quad (14)$$

$$m = \frac{u t_p W(u(t_p + \Delta t))}{(S(t_p + \Delta t) - S(\Delta t))} \quad (15)$$

Equation (14) provides the lag $\Delta t = t^* - t_p$ between the end of the storm and the peak-flow occurrence, while Eq. (15) gives the critical rainfall duration, t_p^* . As noted before, t_p^* is independent of t_r . When Eqs. (14) and (15) are solved for different values of the parameter u , the drainage area contribution to the peak flow, $A^* \equiv A_T(S(t^*) - S(t^* - t_p))$, does not change. As shown in Fig. 4, $f(t)$ shrinks in width as u increases, but at the same time it increases in height, maintaining a constant area, $A^* \equiv A_T(S(t^*) - S(t^* - t_p))$.

The maximum peak discharge depends hyperbolically on rainfall durations (and channel flow velocity):

$$Q_{MAX} = a(t_r) (t_p^*)^{-m} C(t^*, t_p^*) A_T = a(t_r) (t_p^*)^{-m} A^*. \quad (16)$$

Equation (13) provides a model of IUH which depends only on the parameter u and on the basin morphology (i.e. on the shape of the width function). Because A^* is independent of u , both A^* and Q_{MAX} depend only on the form of $W()$, i.e. on the structure of the flow paths.

2.2 The effect of diffusive wave propagation

The analysis presented in the previous section accounts for the mean and most of the variance of the unit hydrograph (e.g., D'Odorico and Rigon, 2003). However, it does not include the effects of hydrodynamic dispersion which introduces a smoothing on the peak flows. In this case flood routing can be expressed by a parabolic differential equation which is obtained either as a diffusive-wave approximation of the de Saint-Venant equations, or through the assumption that water parcels are subject to Wiener dynamics, superimposed to the deterministic advection discussed in the previous sections.

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This assumption leads to the estimation of the travel time distribution as a solution of the Kolmogorov's backward Eq. (Mesa and Mifflin, 1986) with suitable boundary conditions. Thus, the probability distribution of travel times is expressed by the Gaussian inverse function (Rinaldo et al., 1991)

$$f(t|x) = \frac{x}{\sqrt{4 \pi D t^3}} \exp \left[-\frac{(x - u t)^2}{4 D t} \right] \quad (17)$$

where D is the coefficient of hydrodynamic dispersion; the kinematic case (Eq. 12) is obtained for $D \rightarrow 0$.

In this case the hydrograph can be expressed (see Eq. C2 in Appendix C) as

$$\frac{Q(t)}{A_T p(t_p, t_r)} = \begin{cases} \int_0^L dx W(x) \Theta(t|x) & \text{if } 0 \leq t \leq t_p \\ \int_0^L dx W(x) [\Theta(t|x) - \Theta(t - t_p|x)] & \text{if } t > t_p. \end{cases} \quad (18)$$

where

$$\Theta(t|x) = \frac{1}{2} \operatorname{Erfc} \left(\frac{x}{2\sqrt{tD}} - \frac{u}{2}\sqrt{\frac{t}{D}} \right) + \frac{1}{2} \operatorname{Exp} \left(\frac{ux}{D} \right) \operatorname{Erfc} \left(\frac{x}{2\sqrt{tD}} + \frac{u}{2}\sqrt{\frac{t}{D}} \right). \quad (19)$$

Notice that $d\Theta(t|x)/dt = f(t|x)$ with $f(t|x)$ given by Eq. (17); therefore, for $t \geq t_p$, Henderson's equation can be written as

$$\int_0^L dx W(x) [f(t|x) - f(t - t_p|x)] = 0. \quad (20)$$

The time t^* satisfying Eq. (20) is always larger than the precipitation duration as in the case discussed in Sect. 2. All the other results developed in the previous section can be extended to the case $D > 0$, once the function $W(u t)$ in Eqs. (13), (14) and Eq. (14) is replaced by

$$\omega(t) \equiv \int_0^L W(x) f(t|x) dx. \quad (21)$$

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Thus, when Eq. (21) is substituted into Eqs. (14) and (15), the critical rainfall time is still associated with only one value of x^* , while the area, A^* , depends on D and the shape of $W(x)$ but not on u . The peak discharge can be obtained by substituting the time to peak, t^* , into Eq. (18). Interestingly, the concept of concentration time, used in the non-diffusive (kinematic) framework, would be meaningless in this case because the domain of $\omega(t)$ is infinite. However, it can be here re-introduced as a stochastic variable, $\hat{\tau}_c$ with distribution given by Eq. (17) when $x = L$. In fact, in the case $D \neq 0$, after a time $\tau_c = L/u$ has elapsed from the beginning of the rainstorm, the furthestmost portions of the basin may still contribute to the hydrologic response with probability smaller than 1.

3 Case study

An application of the theory developed in this paper, was carried out for the cases of the Longo watershed, a small alpine catchment ($A = 10.3 \text{ km}^2$) within the Avisio basin ($A = 469 \text{ km}^2$) at Predazzo (Italy) and of the Salt River in California (USA) ($A = 2020 \text{ km}^2$).

In this application the width function was calculated accounting for the dependence on the difference in the velocities in channels in hillslopes. Different moisture conditions (Rinaldo et al., 1995; D'Odorico and Rigon, 2003) can be expressed through the quantiles, q , of a wetness index distribution (Beven and Kirkby, 1979; Barling et al., 1994), which defines the parts of the basins contributing to the hydrograph as saturation overland flow (e.g., Sivapalan and al., 1987; Beven and Wood, 1993; Woods and Sivapalan, 1997, 1999). Thus, the total contributing area, A_T , is a function of the degree of saturation of the watershed, $A_T = q A_b$, where A_b is the basin area; the width function is calculated using only the saturated part of the basins (an not the whole basin area) and depends on the channel velocity, u , and the saturated fraction, q , of the basin, while the area contributing to the peak flow, A^* , depends on q but not on u , as shown in the previous sections.

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It is observed that τ_c , does not grow linearly with increasing values of q , starting from a minimum value (i.e. the main stream length divided by u). In all these cases the variation of τ_c remains limited by the hillslope extent, and becomes negligible in the larger basin. Table 1 reports the values of the variables τ_c , t^* , t_p^* , A^* and Q^* in the Longo basin for different values of saturated fraction of the basin. It is observed that t^* is usually less than half τ_c . This difference is due to the long tails of the width function. The contributing area is about eighty per cent of the total saturated area, A_T , and this fraction remains almost constant with varying q , i.e. a first rough estimation gives:

$$A^* \sim 0.8 A_T \sim 0.8 q A_b \quad (22)$$

Excluding the lowest saturation conditions, the critical rainfall duration increases almost linearly with $q > 20\%$, while the delay Δt^* of the maximum peak increases non-linearly.

Because of its simplicity, the flow peak analysis can be easily extended to estimate the runoff peak in all the channel network links inside a basin, providing a regionalization of peak flows. Figure 5a, b and c show an example for the Avisio basin. Figure 5a shows how the maximum discharge Q_{MAX} (with rainfall return period, $t_r = 100$ years) at any link increases with the contributing area A , as:

$$Q_{MAX}(A) \approx 0.906 A^{0.984} \quad (23)$$

where the discharge is in m^3/s and the contributing area in km^2 . In this case, the parameters ($q = 30\%$, $u_c = 2$ m/s, $r = 100$) were derived from calibration on a few high-flow events measured in some sub-catchments of the Avisio. The heterogeneity of the responses shown in Fig. 5 for the small contributing areas is completely due to the different pathway lengths and their subdivision between hillslopes and channels. The maximum discharge at any link is due to rainfall of different duration as shown in Figure 5b. The critical duration is only weakly dependent on (i.e., increasing with) the area.

For smaller contributing areas the critical rainfall duration is even more affected by the variability of the hillslope length. Even though data were not available to confirm the variability of t_p^* in the smaller basins, these results are consistent with those of Wood

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and al. (1990). Figure 5c shows the delay, Δt , of the time to peak with respect to rainfall duration as a function of the contributing areas. It is found that Δt increases non-linearly with the contributing area. The flow peak is delayed with respect to the end of the rainstorm and this delay is larger in the larger subbasins.

The effect of the hydrodynamic dispersion is presented in Fig. 6a, which shows t^* as a function of t_p in a mid-size basin (Salt River – CA, 2020 km²). Notice how t^* is always larger than t_p when $t_p < \tau_c$ (in this basin $\tau_c = L/u = 11.2$ h), as opposed to the classic assumptions of the rational method that $t^* = t_p$ when $t_p < \tau_c$. Because from Eq. (5) $Q_p/(\rho A_T) = A^*/A$, Fig. 6b shows the portion of the watershed contributing to the basin response at the peak flow. It is observed that, with low values of the dispersion coefficient, the response is similar to the kinematic case and the contributing saturated source area is almost A_T (i.e. $A^* \approx A_T$) when $t_p \approx \tau_c = 11.2$ h. With relatively large values of D (hence of the variance of travel times), A^*/A_T remains smaller than 1 in a broader interval of values of t_p . The adimensional parameter $Q_p/(\rho A_T)$ is called in literature “the peakdeness” of the hydrograph (Myninink and Corder, 1976) and was found to vary in real cases between 0.5 and 1.5. Values larger than 1 must then be due to the variability of the rainfall and not to the basin geomorphology.

4 Conclusions

The paper has developed a method for the evaluation of extreme peak-flows based on the theory of the instantaneous unit hydrograph and on the assumption that storm hyetographs have constant rainfall intensity. The system of two Eqs. (14) and (15) gives the maximum allowable discharge, Q_{MAX} , produced by a precipitation event with a certain return period, t_r , time to peak, $t^* = t_p + \Delta t(t_p)$ and constant intensity. When the intensity-duration dependence of extreme precipitation is expressed by a power law, t_p^* and Δt do not depend on the return period. Analytical expressions of Δt were obtained in particular for the linear reservoir and for the Nash models (in Appendix B).

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These methods were also applied to the geomorphological IUH using the framework of the width function. It was found that the area, A^* , contributing to extreme peak flows, Q_{MAX} , does not depend on channel celerity but on the saturated fraction of the basin, q , the ratio between channel and hillslope velocities, u_c/u_h , and the dispersion coefficient, D . Thus, the extreme peakflow, Q_{MAX} , is expressed through a framework that generalises and clarifies the traditional expression of the rational method.

It was also shown that the peak discharge due to surface runoff increases almost linearly with the contributing area. Moreover, both the critical rainfall duration, t_p^* , associated with maximum peakflow values, and the delay between t_p^* and the time-to-peak are increasing functions of the contributing area. The applicability of this framework is partly shown through a few basins with different sizes and morphologies. The semi-analytical character of the simplified theory allows for a fast estimation of the maximum discharge flowing in any link of the river network.

Appendix A

When $t < t_p$ and the rainfall is given by Eq. (3) the rate of flow is:

$$Q(t) = A_T \rho \int_0^t f(t - \tau) d\tau = -A_T \rho \int_t^0 f(t_1) dt_1 = \rho A_T S(t) \quad (\text{A1})$$

where the change of variable in the integral is straightforward and $A(t) \equiv A_T S(t)$ is the watershed area contributing to the flood discharge at time t as follows from the definition of width function. For $t > t_p$ we have instead:

$$Q(t) = A_T \rho \int_0^{t_p} f(t - \tau) d\tau \quad (\text{A2})$$

After the change of variable $t_1 = t - \tau$ we obtain:

$$Q(t) = -A_T \rho \int_t^{t-t_p} f(t_1) dt_1 = A_T \rho \left[\int_0^t f(t_1) dt_1 - \int_0^{t-t_p} f(t_1) dt_1 \right] \quad (\text{A3})$$

from which, we finally have:

$$Q(t) = \rho A_T [S(t) - S(t - t_p)]. \quad (\text{A4})$$

$Q(t)$ is a continuous function of t at $t = t_p$.

Thus, the basin response Eq. (1) can be expressed as

$$Q(t) = \begin{cases} \rho A_T S(t) & (0 \leq t \leq t_p) \\ \rho A_T (S(t) - S(t - t_p)) & (t > t_p), \end{cases} \quad (\text{A5})$$

where $S(t)$ is the the S-hydrograph (Doodge, 2003). Notice that $S(t)$ is a continuous function of time and $Q(t)$ has a possible discontinuous derivative at $t = t_p$. The maximum discharge is obtained at the time to peak, t^* , which is found either for $t = t_p$ or by solving the equation $dQ/dt = 0$ (Henderson, 1963):

$$\begin{aligned} \frac{dS(t)}{dt} = f(t) = 0 \quad & 0 \leq t \leq t_p \\ \frac{dS(t)}{dt} = \frac{dS(t - t_p)}{dt} \quad & t > t_p. \end{aligned} \quad (\text{A6})$$

By definition in Eq. (A6a) $f(t)$ is null for $t^* = 0$ – which represents a trivial and unphysical solution – or for $t^* = \tau_c = L/u_c$.

Thus, if $t_p > \tau_c$ the time to peak coincides with the concentration time, otherwise (i.e. for $t_p \leq \tau_c$) the solution is found by solving Eq. (A6b), which is equivalent to Eq. (4).

Appendix B

An interesting application of Henderson's equation is found for the case of the Nash IUH (Nash, 1957):

$$f(t) = \frac{1}{n!} \left(\frac{t}{k}\right)^{n-1} e^{-t/k} \quad (\text{B1})$$

where n and k are two calibration parameters. For $n = 1$ the hydrograph is a negative exponential (linear reservoir) and the peak is at t_p (i.e. $t^* = t_p$). For $n \geq 2$, Eq. (4) becomes

$$\left(1 - \frac{t_p}{t}\right)^{n-1} = e^{-t_p/k} \quad (\text{B2})$$

5 which is solved as:

$$t^* = \frac{t_p}{1 - (\exp(-t_p/k))^{1/(n-1)}}. \quad (\text{B3})$$

It is easy to observe that t^* is always greater than t_p and that t^* is an increasing function of the parameter n .

The resulting critical rainfall time is determined (Eq. 9) by solving:

$$10 \quad m = \frac{t_p (t^*(n-1)) e^{t^*/k}}{\Gamma(n, (t^* - t_p)/k) - \Gamma(n, t^*/k)} \quad (\text{B4})$$

where Γ is the incomplete gamma function:

$$\Gamma(a, x) = \int_0^\infty t^{a-1} e^{-t} dt. \quad (\text{B5})$$

As noted, the linear-reservoir model ($n = 1$ case), is a particular case of Eq. (B2). In this case, $\Delta t = 0$ for any t_p . In this case Eq. (9) becomes

$$15 \quad m = \frac{(t_p/k) e^{-t_p/k}}{1 - e^{-t_p/k}}. \quad (\text{B6})$$

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Appendix C

The hydrograph response is expressed by Eq. (1) with ρ given by Eq. (3) and $f(t - \tau)$ by Eq. (11):

$$Q(t) = A_T \rho(t_p, t_r) \int_0^t H(t_p - \tau) \times \int_0^L W(x) f(t - \tau|x) dx d\tau. \quad (C1)$$

5 When $f(t - \tau|x)$ is expressed by Eq. (17), the change of variable $t - \tau \rightarrow t'$ leads to

$$\frac{Q(t)}{\rho(t_p, t_r) A_T} = \begin{cases} \int_0^L dx W(x) \int_0^t f(t'|x) dt' & 0 \leq t \leq t_p \\ \int_0^L dx W(x) \int_{t-t_p}^t f(t'|x) dt' & t > t_p \end{cases} \quad (C2)$$

where

$$\Theta(t) = \int_0^t f(t'|x) dt' = \mathcal{L}^{-1} \left[\frac{\hat{f}(s|x)}{s} \right]_{t'=t}. \quad (C3)$$

10 In Eq. (C3) $\mathcal{L}^{-1}[\]$ represents the inverse Laplace-transform, while $\hat{f}(s|x)$ is the Laplace transform of Eq. (17)

$$\hat{f}(s|x) = \text{Exp} \left[x \frac{u - \sqrt{u^2 + 4 s D}}{2 D} \right]. \quad (C4)$$

Equation (C4) in Eq. (C3) gives

$$\Theta(t) = \text{Exp} \left(\frac{u x}{2 D} \right) \mathcal{L}^{-1} \left[\frac{\text{Exp} \left(-a \sqrt{s + b^2} \right)}{s} \right] \quad (C5)$$

15 where $a = x/\sqrt{D}$ and $b = u/2\sqrt{D}$. The inversion of the Laplace transform leads to Eq. (19).

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5 The model PeakFlow is freely available for download under GPL 3.0 license as part of the GIS JGrass (where it is coded as a OpenMI 1.4 component) and in the JGrasstools (<http://www.jgrasstools.org>) coded as a OMS3 component.

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Table 1. Relevant quantities calculated for the rescaled width functions of the Longo catchment ($r = 10$): q is the fraction of saturated areas; t_p^* is the duration of the rainfall which gives the largest peak discharge; t^* the time to peak; τ_c the concentration time; A^* the area contributing to the peak; Q^* is the largest discharge; A_T the total saturated area.

q	t_p^* (s)	t^* (s)	τ_c (s)	A^* (km ²)	Q^* (m ³ s ⁻¹)	A_T (km ²)
0.05	4624	5075	14 631	0.418	6.25	0.529
0.13	5834	6808	15 721	1.086	14	1.373
0.28	6063	7632	16 287	2.288	28.85	2.931
0.55	6407	8325	16 570	4.497	54.8	5.668
0.82	6763	8884	17 496	6.799	80.07	8.472
1	7292	9925	17 496	8.3171	94.012	10.33

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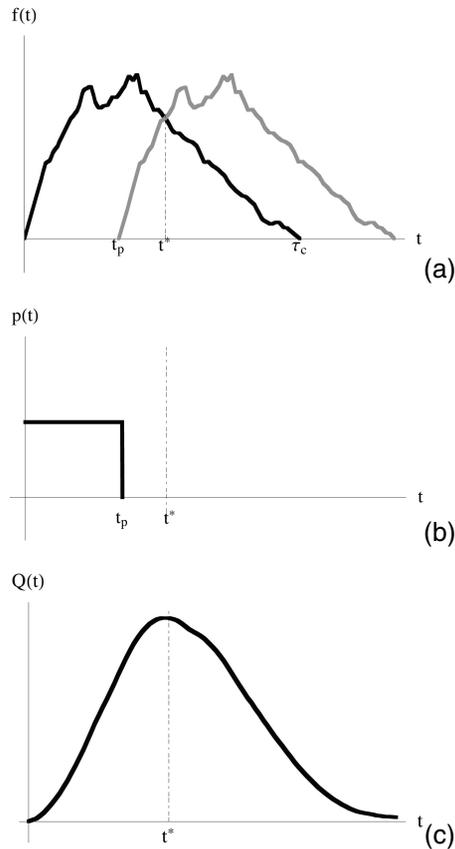


Fig. 1. The solutions of Eq. (4) are given by the crossing of the unit hydrograph $f(t)$ with another unit hydrograph, $f(t - t_p)$, shifted by a distance, t_p . The figure shows: **(a)** the graphical solution of the equation (in black $f(t)$ and in grey $f(t - t_p)$); t^* is the time to peak. **(b)** t^* is usually larger than t_p . This is true for constant intensity uniform hyetograph; **(c)** the discharge obtained by the convolution of the IUH in **(a)** with the rainfall in **(b)**.

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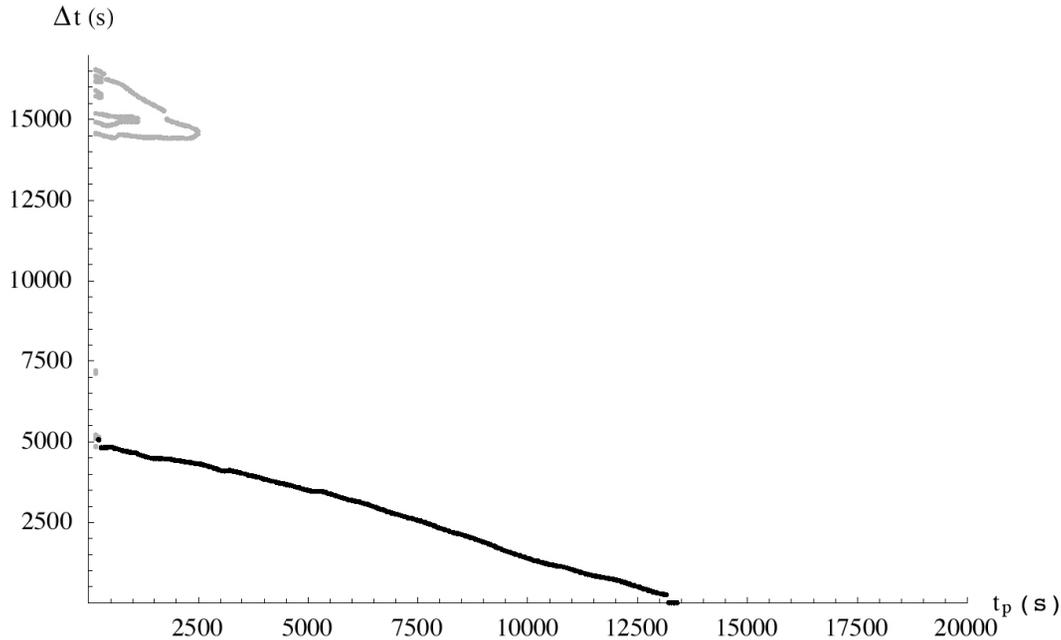


Fig. 2. Delay, Δt , of time to peak with respect to the end of the rainstorm, as a function of rainfall duration, t_p . In this case the rescaled width function approach (D’Odorico and Rigon, 2003) was used to account for the lower flow velocity in the hillslopes, $u_h \ll u$. The analysis refers to the rescaled width function of the Longo watershed (Italy) with $r = u/u_h = 10$ and $q = 1$. For values $t_p \geq \tau_c$, the delay is null. For smaller t_p there are multiple peak flows. The grey points correspond to secondary peak flows and the black points to the largest peak flow.

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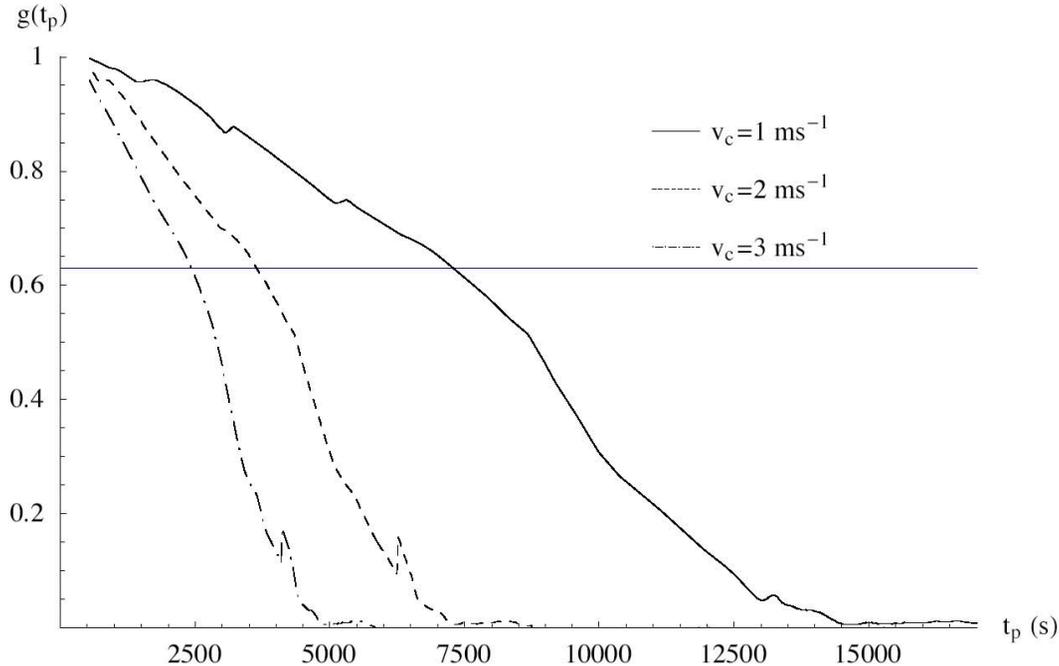


Fig. 3. Graphical solution of Eq. (9) for three values of channels celerity v_c . The value of $m = 0.63$ (horizontal line) is found by analysing the extreme precipitation in the area of the Longo watershed (Italy); the other curves represent the function $g(t_p)$ given by Eq. (9) with different channel velocities. By increasing the velocity the time to peak decreases and the peak discharge increases. However the area contributing to the peak flow remains constant in all of the three cases.

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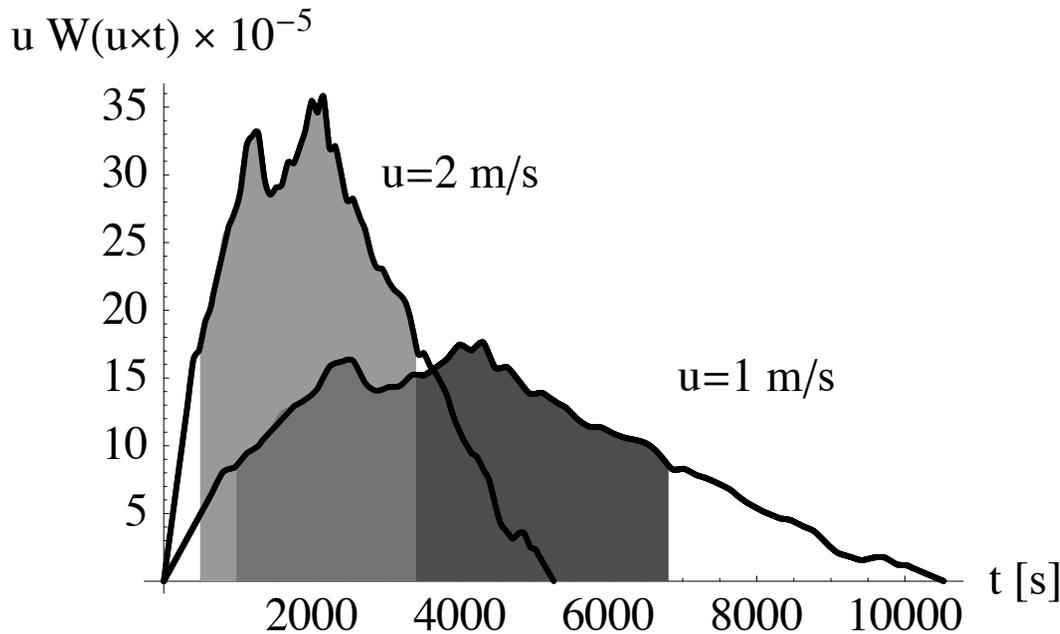


Fig. 4. Two examples unit hydrographs derived from the width function with $u = 2$ m/s (with higher peak on the left) and $u = 1$ m/s (with smaller peaks on the right). The area contributing to the maximum discharge, as derived from the Henderson's equation and Eq. (17), is 80% of the total contributing area for each of the hydrograph and marked in grey (darker for the case $u = 1$ and of an intermediate tone for both the cases). In the $u = 1$ m/s case, the rainfall duration which gives $(Q_1)_{MAX} = 14 \text{ m}^3/\text{s}$ is $(t_1)_p^* = 5828 \text{ s}$; is also $(t_1)^* = 6808 \text{ s}$ (the right limit of the grey). In the case $u = 2$ m/s it is: $(Q_2)_{MAX} = 21.2 \text{ m}^3/\text{s}$, $(t_2)_p^* = 2914 \text{ s}$, $(t_2)^* = 3404 \text{ s}$.

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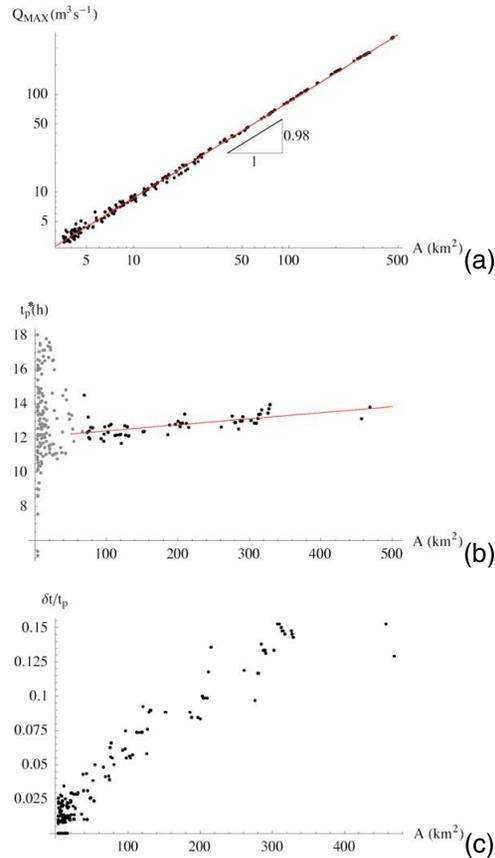


Fig. 5. (a) Plot of the peak discharge for all the links within the Avisio basin (Italy) as a function of contributing area. The parameters ($q = 30\%$, $u_c = 2 \text{ m/s}$, $r = u/u_h = 100$) were obtained from calibration on some real event in a few subcatchments and using IDF curves for a return period $t_r = 100$ years. (b) Critical rainfall duration, t_p^* , of links ends inside the Avisio basin (Italy). (c) Delay of the time to peak with respect to the rainfall duration as a function of the contributing areas.

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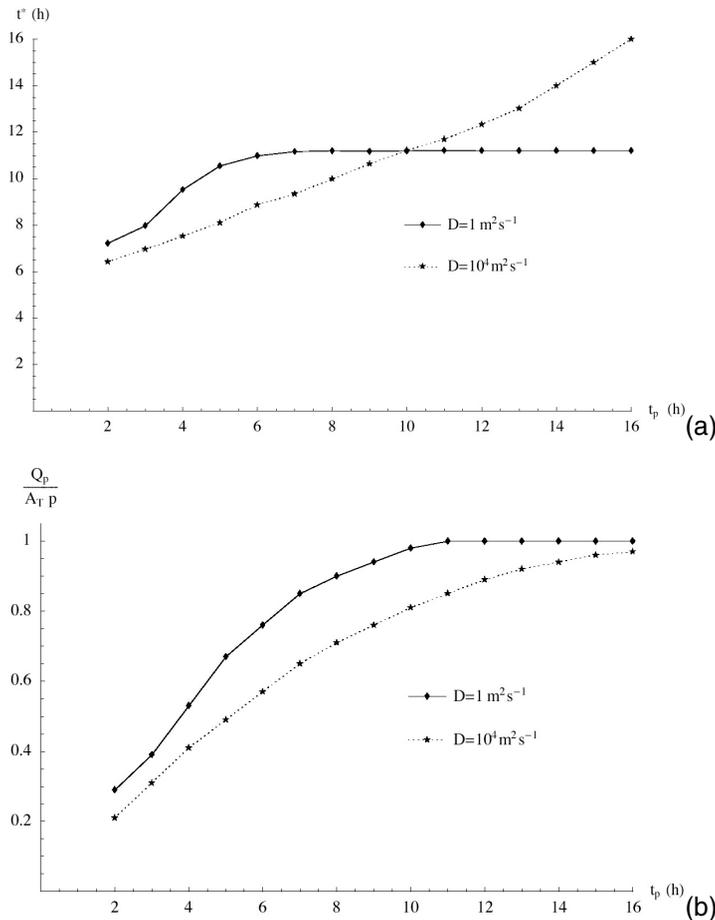


Fig. 6. (a) Time to peak (t^*) and **(b)** normalized peak flow values as a function of the rainstorm duration (t_p) in the Salt River (CA). Notice how $Q_p/(\rho A_T) = A^*/A_T$ does not depend on channel flow velocity, since Eq. (5) remains valid also in this more general case.

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