CLIMATIC CONTROL ON THE VARIABILITY OF THE FLOOD DISTRIBUTION

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ABSTRACT

The variability of the second order moments of flood peaks with respect to geomorphoclimatic basin characteristics was investigated. In particular, the behavior of the coefficient of variation (*Cv*) was analyzed with respect to its sensitivity to physically consistent quantities. We achieved results in fairly good agreement with real world observed characteristics and more light is so shed on the relationship between *Cv* and basin size. It appears that *Cv* is mainly controlled by the climate and some loss features. Many observed data, reported in literature, show a decrease of *Cv* with basin area *A*, usually ascribed to the limited spatial extent of extreme events, which leads to a decrease of *Cv* of areal rainfall intensity. An increase of *Cv* with the area at small scales is sometimes recognized as well. Such different behaviors were herein accounted for by the concurrent effect of two parameters, representative of the way rainfall losses and effective rainfall scale with the basin area, respectively, affect the *Cv*(*A*) relationship. In addition, it is shown how the same relationship may be affected by the observation that rainfall often presents practically constant skewness for durations ranging between a few hours and one or two days with a decrease in its value only for smaller duration of rainfall events. Yet, this effect is counterbalanced by the fact that small durations are significant for very small basins, where the flood peak contributing area tend to be constituted by a lumped part of the basin rather than by the sum of contributing subbasins as it is typical for larger watersheds.

KEYWORDS: Floods, Climate, Coefficient of variation, Scaling.

1. Introduction

The general context of this research is to promote a deeper knowledge of the physical mechanisms involved in the flood generation process with particular regard to its use in the framework of flood frequency analysis. Within this field, important problems may be considered still open notwithstanding the great efforts produced in the last twenty years of hydrologic research. Among those, we like to point out that most procedures for estimating the mean annual flood, in the context of the index-flood method (*Dalrymple*, 1960), are still empirical and they are often different from one region to another. Moreover, statistical regional analyses may lead to consider regions different for geology, morphology, climate, etc. as if they were homogeneous. The available methods for flood estimation, including regional analysis, usually make use of statistical procedures which essentially exploit the hydrometric and pluviometric information leaving only a marginal role to other information (geology, climate, vegetation and so on) which is crucial to the investigation of the base process whose maxima we are dealing with. Therefore, the main purpose of our research is to enhance the role of such kind of information within the estimation procedures, providing for it a major and objective role not confined to the feeling and arbitrariness of the technicians.

Nevertheless this kind of approach may lead to a deeper knowledge of the base process mechanisms as a feedback of the statistical analysis performed over annual flood maxima.

The way we follow with this aim consists in the theoretical derivation of flood frequency as initially introduced by *Eagleson* (1972) and successively rearranged by *Iacobellis and Fiorentino* (2000) and *Fiorentino and Iacobellis* (2001) as it will be briefly resumed hereafter.

2. Theoretical framework

The cited theoretical model (*Iacobellis and Fiorentino,* 2000) was developed for the analytical derivation of the flood frequency distribution. Such development was meant as a tool for flood prediction in ungauged basins as well as a contribution to the improvement of knowledge of physical processes controlling the flood generation processes. In the model, an important characteristic of the probability distribution of floods is the mean annual number Λ_q of independent floods, which, when this distribution is schematized as a compound Poisson process (e.g. EV1, GEV, TCEV), represents the distribution parameter that more strongly controls the coefficient of variation C_v.

As shown in *Iacobellis and Fiorentino* (2000), in the case of floods generated by rain storms, Λ_q can be related to a water loss parameter f_A through the following formula:

$$\Lambda_q = \Lambda_p \exp\left(-\frac{f_A^{\ k}}{E[i_{A,\tau}^k]} \right) \tag{1}$$

In equation (1), Λ_p is the mean annual number of independent storms, $E[\cdot]$ is the expectation operator, and $i_{A,\tau}$ represents the average intensity of the maximum rainfall amount measured during the storm in a duration τ, where τ is the lag time of the basin of surface area A. Equation (1) is based on the simplified assumption that the peak discharge Q_P is related to $i_{A,\tau}$ by the following equation:

$$Q_P = \xi \; (i_{a,\tau} - f_a) \, a + q_o \qquad (2)$$

which may be considered sufficiently well suited for use in the frame of a theoretical model for deriving the flood distribution (*Gioia et el.*, 2001). In the above equation a is the variable area contributing to runoff peak, ranging from 0 to A, ξ is a routing factor less than unity, q_o is a constant base flow and f_a is the rainfall intensity threshold above which runoff is produced. It can be shown that equation (1) can be derived from equation (2) under the hypothesis that $i_{A,\tau}$ is a Weibull variate with shape parameter k. Incidentally, k equals unity when the Weibull distribution reduces to the exponential.

The time-space behavior of the involved quantities is basically controlled by the commonly observed geomorphological power-type relationship between basin lag-time τ_A and basin area A, which can be written as:

$$\tau_a = \tau_1 \, a^{\nu} \quad \text{with} \quad \tau_1 = \tau_A A^{-\nu} \qquad (3)$$

where τ_A is the lag-time of the basin and ν is a parameter that usually assumes values close to 0.5. The mean areal rainfall intensity $E[i_{A,\tau}]$ is usually found to scale with A according to the power law

$$E\left[i_{A,\tau}\right] = i_1 \; A^{-\varepsilon} \qquad (4)$$

where i_1 is rainfall intensity referred to the unit area. Also, in the model f_A is in general supposed to scale with the basin area A through a relationship of the type:

$$f_A = f_1 \, A^{-\varepsilon'} \qquad (5)$$

in which f_A represents the average water loss rate when the entire basin contributes to the flood peak. Indeed, τ_A, $E[i_A]$, f_A, ν, ε and ε’ are characteristic features of basins.

Replacing relations (4) and (5) in equation (1), after considering that

$$E\left[i_A^k\right]=\left(\frac{E\left[i_A\right]}{\Gamma(1+1/k)}\right)^k \tag{6}$$

we obtain:

$$\Lambda_q = \Lambda_p \exp\left[-\left(\frac{f_1\Gamma(1+1/k)A^{\varepsilon'-\varepsilon}}{i_1}\right)^k\right] \tag{7}$$

From the above relation it appears that the scaling relationship Λ_q-area is clearly dependent on the values assumed by the parameters ε and ε'. The first parameter depends on the slope of the IDF curve and on the exponent of the geomorphological relationship τ-area and usually assumes values around 0.3-0.4. The second one is much more variable and its value, according to *Fiorentino and Iacobellis* (2001) and *Fiorentino et al.* (2001), may be characterized by the long term climate by way of the probable state of the basin in terms of its antecedent soil moisture conditions. In particular, *Fiorentino and Iacobellis* (2001) derived the following expression:

$$f_A = \vartheta_1 c_1 A^{-\nu} + \vartheta_2 c_2 A^{-\nu/2} + \vartheta_3 c_3 \tag{8}$$

where c_1, c_2 and c_3 are coefficients mainly depending on the spatial averages of initial abstraction, characteristic sorptivity and gravitational infiltration rate respectively. In equation (8), ϑ_1, ϑ_2, and ϑ_3 are weights, ranging from 0 to 1, which are around unity in a certain climate, and approach zero elsewhere. In particular, the first of them tend to prevail in arid zones, while the last one is mainly active in hyper-humid climates. Thus, when the first term prevails, equation (8) reduces to equation (5) with the value of the exponent ε' which, according to the literature value of ν, tends to approach –0.5. In the case of prevalence of the second term, related to the infiltration rate through unsaturated soils, ε' is about –0.25. Finally, the dominance of the gravitational infiltration rate, in saturated soils, should lead to ε' approaching zero. Yet, one may acknowledge that the processes involved can confuse or prevail on each other, depending on particular conditions of soils, land coverage and initial moisture.

One can observe that the range covered by the ε’ values may easily lead to positive or negative values of the difference ε – ε’. Indeed such difference may give rise to ascending or descending relationships Λ_q-area.

Case study

The investigated zone consists of 32 gauged basins of three administrative regions, namely Basilicata, Puglia, and Calabria, in Southern Italy (**Errore. L'origine riferimento non è stata trovata.**).

These basins are listed in Table 1. In this area the climate is quite variable, ranging from the hot-dry Mediterranean (semiarid or dry sub-humid) type of the northeastern sector (Puglia), cold and humid type of the West-Southern sector (Basilicata and Calabria) where the orography is much more developed (Southern Apennine). The annual average rainfall goes from about 600 mm in Puglia to about 2000 mm in the highest parts of Basilicata and Calabria.

The climatic classification was based on the Thornthwaite (1948) climatic index:

$$I = \frac{h - E_p}{E_p} \qquad ()$$

In which h is the mean annual rainfall and Ep is the mean annual potential evapotranspiration, calculated here according to Turc's formula (Turc, 1961), dependent on the mean annual temperature only.

On the study area, regional statistical analyses were available on the annual maxima of daily rainfall, all based on the Two Component Extreme Value (TCEV) distribution (Rossi et al., 1984) regional model. Parameters Λ_1, θ_1, and Λ_2, θ_2 of the TCEV were estimated using a Maximum Likelihood (TCEV-ML) procedure (Gabriele and Iiritano, 1994) with hierarchical estimation of parameters (Fiorentino et al., 1987), based on the homogeneous areas found in Gabriele and Villani (1999 – rapporto nazionale). The same regional model was applied to series of annual maxima of flood peaks recorded at the stations mentioned in Table 1. Resulting values of Λ_q were agani taken from Gabriele and Villani (1999 – rapporto nazionale).

As a result of the model application to a large region of Southern Italy including Basilicata, Calabria and Puglia (*Claps et al.*, 2001), the ratios of regional estimates of Λ_q and Λ_p are shown in Figure 1 for basins of Calabria, mostly humid and permeable along with arid basins of Puglia and Basilicata (Table 1). Different behavior not dependent on the basin area is shown by humid basins of Puglia and Basilicata where the flood-rainfall yield in terms of number of events is basically controlled by the climate (*Fiorentino and Iacobellis*, 2001).

The observed pattern of the ratio Λ_q/Λ_p is explained by combining equations (4) and (5), in equation (1). In fact, the slope of the scaling relationship (4) of areal rainfall intensity is substantially homogeneous over the three regions (*Claps et al.*, 2001), and the resulting behavior of the ratio Λ_q/Λ_p depends on the scaling functions (5), shown in figure 2. In Figure

1 also appear two slightly humid basins with large basin area (Ofanto at Rocchetta S. Antonio and Basento at Menzena), whose behavior may be assimilated to the arid case.

3. Coefficient of variation of floods and basin area.

The presented theoretical framework also allows us to shed more light on the relationship between *Cv* and the basin scale (e.g. *Smith*, 1992, *Gupta and Dawdy*, 1995, *Robinson and Sivapalan*, 1997a, b; *Blöschl and Sivapalan*, 1997) just looking at the dependence of the involved parameters with basin scale. In fact, many observed datasets, reported in literature, show a decrease of *Cv* with area, usually ascribed to the limited spatial extent of extreme events which leads to a decrease of *Cv* of areal rainfall intensity. An increase of *Cv* with the area at small scales is sometimes recognized as well. For instance, the observation that rainfall presents skewness practically constant for durations ranging between a few hours and one or two days while this statistics is usually lower at smaller durations could partly account for the above mentioned increase of *Cv*.

In the context of the presented model such behavior could be accounted for, the way parameters like Λ_p, k, ε and others, scale with the basin area. For example, the ε value, representative of the precipitation scaling patterns as it is seen from the basin, seems to decrease with area producing higher *Cv*.

Let us now assume on a first approximation that $k = 1$, corresponding to the hypothesis of exponential distribution of rainfall intensity. In this case the model proposed by Iacobellis and Fiorentino leads to a distribution that under a very wide range of situations is not very far from a simple Gumbel distribution (EV1). Consequently, to the aim of this paper, the relationship between *Cv* and Λ_q can be confidently expressed in the closed form:

$$Cv = \frac{1.28255}{\log \Lambda_q + 0.5771} \quad (9)$$

Replacing equation (7) into (9) we get:

$$Cv = \frac{1.28255}{\log \Lambda_p - \frac{f_1}{i_1} A^{\varepsilon - \varepsilon'} + 0.5771} \quad (10)$$

which represents a simple *Cv*-*A* relationship.

In figure 3 we show, as an example, the functional relationship of equation (10) for a given set of rainfall characteristic parameters, namely Λ_p= 20, i_1 = 13 mm/h and ε = 0.33. The patterns shown in figure 3 for different values of ε' (0, 0.16, 0.33 and 0.5) highlight that the scaling relationship *Cv*-*A*, the other quantities constant, is significantly dependent on the values ε and ε' and their difference.

Commenting on figure 3 and taking into account the typical values of ε' presented in the previous section 2, we should point out that *Cv* is likely to decrease with the basin area *A* in arid basins where the prevailing runoff generation mechanism is of the Horton type and f_A tends to scale with *A* raised to the power ε' = -0.5 (for a more general comment on this topic, see also *Fiorentino and Iacobellis*, 2001). On the contrary, in humid and vegetated basins, where the prevailing runoff generation mechanism can be reconnected to soil saturation excess, and ε' tends to be zero, *Cv* may increase as the basin area increases. On the other hand, figure 3 also points out that the achieved relationship *Cv*-*A* shows a more significant sensitivity of *Cv* to *A* when ε'is greater than ε, and that this may provide an explanation for the fact that in the real world a negative scaling of *Cv* with the basin area *A* seems to be prevailing.

4. Conclusions.

In this study the behavior of the coefficient of variation *Cv* of floods was investigated with particular regard to the way it tends to scale with the basin area *A*. It was developed in the frame of the theoretical distribution of floods proposed by *Iacobellis and Fiorentino* (2001). According to this distribution, *Cv* is mainly controlled by mean annual number of floods, Λ_q, and by the shape parameter of the probability distribution of rainfall, *k*. In the model, Λ_q is in turn dependent on Λ_p, f_A and $E[i_A]$, whose definitions are provided earlier in this paper. In addition, in a region where *k* ad Λ_p are constant, $E[i_A]$ and f_A scale with the basin area by a power law with exponents ε and ε' respectively. Therefore, the relationship *Cv* – *A* depends on ε and ε'.

This result adds new arguments to the controversial question whether *Cv* should theoretically increase or decrease as the basin size becomes larger. In particular it suggests that a significant role on the control of the *Cv* behavior is played by the abstraction characteristics at the basin scale. In addition, as these features have been shown to be strongly related to the long term climate, it is also pointed out how the climate drives the scaling relationship *Cv* – *A*. In particular, a double control is recognized: the first one relates to the precipitation IDF and their scaling behavior, and the latter is strongly influenced by the basin response in terms of the rainfall-floods events ratio as determined by the rainfall threshold for runoff generation f_A.

Finally, we should acknowledge that non-linearity effects in the process of floods generation, not taken into account in this paper, may tend to add complexity to the scaling properties of *Cv*.

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Site	A (km^2)	τ_A (h)	Λ_p ($years^{-1}$)	Λ_q ($years^{-1}$)	Cv	f_A (mm/h)
Santa Maria at P.te Lucera Torremaggiore	58	2.6	44.6	2.6	0.92	6.48
Triolo at Ponte Lucera Torremaggiore	56	2.6	44.6	3.1	0.70	8.67
Salsola at Ponte Foggia San Severo	455	7.3	44.6	5.0	0.54	1.89
Casanova at Ponte Lucera Motta	57	2.6	44.6	3.7	0.81	5.62
Celone at Ponte Foggia San Severo	233	5.2	44.6	6.6	0.72	2.09
Celone at San Vincenzo	92	3.3	44.6	6.1	0.61	3.41
Cervaro at Incoronata	539	8.0	44.6	5.2	0.58	1.80
Carapelle at Carapelle	715	9.2	44.6	8.5	0.57	1.19
Venosa at Ponte Sant'Angelo	263	5.6	44.6	4.2	1.18	2.64
Arcidiaconata at Ponte Rapolla Lavello	124	3.8	44.6	4.1	0.65	3.67
Bradano at Ponte Colonna	462	4.3	21.0	4.0	0.76	2.36
Bradano at San Giuliano	1657	7.1	21.0	2.9	0.79	2.17
Basento at Menzena	1382	5.95	21.0	6.6	0.63	1.22
Ofanto at Rocchetta Sant'Antonio	1111	11.5	21.0	4.7	0.58	1.13

Table 1a. *Arid basins in Puglia and Basilicata.*

Site	A (km^2)	τ_A (h)	Λ_p $(years^{-1})$	Λ_q $(years^{-1})$	Cv	f_A (mm/h)
Esaro at La Musica	520	4.7	20	3.0	0.82	3.5
Coscile at Camerata	285	3.7	20	3.2	0.74	4.5
Trionto at Difesa	32	2.8	20	10.7	1.09	1.0
Tacina at Rivioto	79	3	10	4.0	1.27	3.2
Alli at Orso	46	3	20	4.0	0.72	5.8
Melito at Olivella	41	3	20	4.8	0.62	4.5
Corace at Grascio	182	3.8	20	4.5	0.70	3.4
Ancinale at Razzona	116	3.9	10	3.3	0.73	4.4
Alaco at Mammone	15	1.3	10	3.5	0.75	7.4
Amato at Marino	113	4.6	20	5.0	1.18	2.6
Lao at Piè di Borgo	280	3.7	34	5.5	0.59	3.9
Noce at La Calda	42.5	1.3	34	13.7	0.41	2.9

***Table 2b**. Basins in Calabria.*

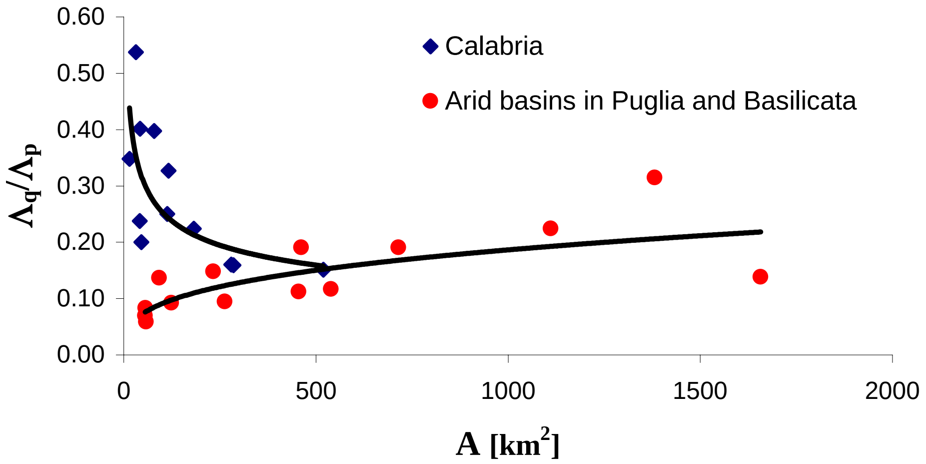

Figure 1. *Mean annual number of flood and rainfall events ratio versus basin area A in Calabria and arid basins in Puglia and Basilicata*

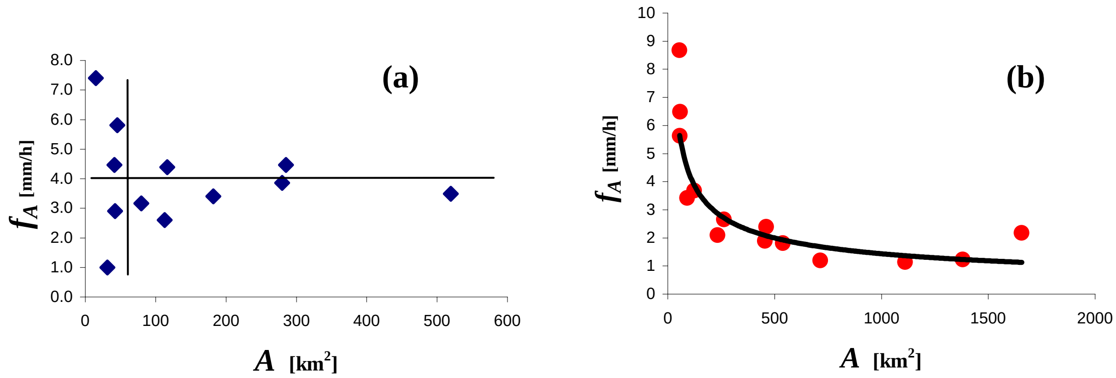

Figure 2. *Average space-time water loss intensity versus basin area A in Calabria* (**a**) *and arid basins in Puglia and Basilicata* (**b**).

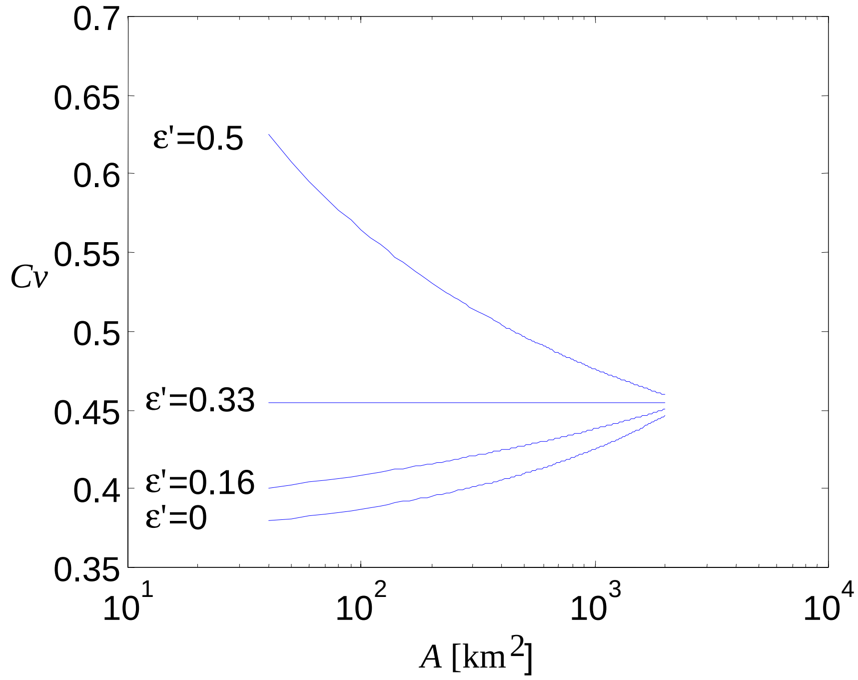

Figure 3. *Coefficient of variation of floods versus basin area A according to equation* (10).