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GEOMORPHIC EFFECTIVENESS OF FLOODS ON TROTUŞ RIVER CHANNEL (ROMANIA) BETWEEN 2000 AND 2012

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Abstract: We determined the hydrological and hydraulic characteristics of flood events occurring on the river Trotus between 2000 and 2012 based on data recorded daily at 6 gauging stations, direct measurements in the field (channel slope, grain size, channel width, and riverbank height) and data provided by orthophotos taken before and after flood events. The parameters we derived from these data (i.e., bankfull discharge, mean annual flood, catastrophic flood, geomorphic flood, specific stream power, shear stress, and critical shear stress) allowed us to assess the degree of geomorphological impact of each flood, according to their magnitude and duration. Of the 5 flood events (i.e., 2004, 2005, 2007, 2010, and 2012) during the study period, two (the 2005 flood, with a peak discharge of 2845 $m^3 s^{-1}$, and the 2010 flood, with a peak discharge of 1567 $m^3 s^{-1}$) stood out both in terms of the particular values of the aforementioned parameters and, in particular, the major changes they generated in the Trotus river channel. The bankfull discharge flood duration (72 hours in 2005 and 222 hours in 2010) and the duration above the Miller-Magilligan threshold (45 hours in 2005 and 48 hours in 2010), can to a large extent explain the high geomorphic effectiveness of moderate-magnitude flood events that span longer time frames. While the magnitude of the 2005 flood (with a recurrence interval of 200 years) was nearly double that of the 2010 flood, their geomorphological impact was rather similar. The response of the channel to these floods consisted of discontinuous spatial adjustments in the channel width and channel bed elevation. The assessment of the channel aggradation/degradation rates was performed based on the evolution over time of average multiannual discharge levels and channel survey data (cross-sections). Prominent changes in the channel occurred in the Târgu Ocna–Căiuţi sector (30 km long, nearly 20% of the entire length of the river), where, between 2005 and 2012, the channel bed deepened by an average value of 0.85 m (0.12 cm y^{-1}). In terms of their geomorphic impact, the 2005 and 2010 flood events can be considered severe floods, even catastrophic in certain sectors.

Keywords: Floods; specific stream power; flood duration; geomorphic effectiveness; geomorphic impact

1. INTRODUCTION

The on-going, nearly two-decade-long monitoring of the Trotuş River channel has allowed for the observation of net, visible changes in the channel. Some of these significant changes have primarily included channel bed degradation and an increase in the channel width at the expense of the floodplain. Surely, these changes can be attributed to a wide range of causes (i.e., climate changes, basin-wide anthropogenic impact, neotectonics, etc.). However, the most plausible cause is the increased magnitude and frequency of geomorphically effective floods. Our observations are corroborated by studies reviewing flood events from the European continent that have occurred during the past 25 years, and it is concluded that Romania is among the most affected countries in this regard (Kundzewicz et al., 2012; Pińskwar et al., 2012).

In fluvial geomorphological research, the classic question still remains whether there is a relationship between discharge magnitude and frequency and their geomorphic effectiveness in shaping river channels and valleys (Costa & O`Connor, 1995). Attempts to provide a proper answer to this delicate problem of geomorphology resulted in an array of studies that became classics: Hack & Goodlett (1960); Wolman & Miller (1960);

Wolman & Gerson (1978). Thus, Wolman & Miller (1960) admit that the largest proportion of geomorphic work performed by rivers occurs during moderate floods, with a frequency of 1 or 2 per year, which is typical for most regions worldwide. Despite widespread acknowledgement that high-frequency moderate floods generate the largest share of the year-long geomorphic work, studies addressing this subject are far fewer than research tackling extreme events (Kale & Hiri, 2007). The explanation could reside in that the former type of floods generate low-intensity forces that have a lesser effect on landforms, whereas the latter type (i.e., extreme or catastrophic events), whose recurrence intervals range from decades to centuries, result not only in bed load transport but also in substantial changes in the channels (Magilligan, 1992; Molnar et al., 2006). Extreme floods, which commonly result in visible geomorphological changes over varying periods of time, can explain the growing interest in the study of extreme events of various intensities occurring throughout the world (Downs et al., 2013; Thompson & Croke, 2013). The first observation attested to the fact that the geomorphic impacts of similar magnitude and frequency flood events are not necessarily identical, thus rendering comparisons and extrapolations rather difficult (Fuller & Heerdegen, 2005). To determine the differences in the behaviour of similar magnitude and frequency floods, several parameters or aggregations of parameters have been employed (Molnar et al., 2006). Early on, the flood peak discharge was used for this purpose, only to be replaced by parameters such as specific or unit stream power, channel boundary shear stress, and energy expended per unit area, all of which, unlike the flood peak discharge, can provide additional information on the potential of a certain flood event (Baker & Costa, 1987). In many cases, changes occurring in the river channel during large flood events were highlighted and substantiated by using the *specific stream power* parameter, which quantifies the actual energy exerted on the channel bed and banks (Kale, 2007; Barker et al., 2009; Ortega & Heydt, 2009). However, some studies show that processes that induce changes in the channel bed during peak discharge on a local scale are rather difficult to detect by using a single hydraulic parameter (Miller, 1990; Krapesch et al., 2011), thus requiring the use of additional parameters (Bull, 1979; Beven, 1981). Nonetheless, on a large scale, this parameter provides satisfactory information on channel adjustments during flood events. Aside from the aforementioned parameters, flood duration is a significant factor because it can explain the differences in terms of impact on the channel. In other words, the geomorphic effectiveness of floods with lower values of peak discharge, shear stress or stream power but longer durations is more apparent compared to the effectiveness of higher magnitude, lower duration floods (Magilligan et al., 2015). The geomorphic effectiveness of a flood is further influenced by certain local factors (channel and valley slope gradients, confinement, degree of resistance of constituent rocks, etc.) or traits of the drainage basin (basin area, drainage network density, etc.) (Magilligan et al., 2002).

Studies that have examined extreme flood events on Romanian rivers post-2000 have either addressed synoptic conditions preceding the floods and the relationship between the amount of precipitation and the resulting peak discharge values (Romanescu & Nistor, 2011; Romanescu, 2013) or the hydrological traits of floods (Chirilă & Preda, 2006, 2007; Chirilă & Zaharia, 2010) and flood control management (Salit et al., 2013). The geomorphic impact of large flood events has largely been overlooked. In the few such studies, the topics addressed have included alluvium transport during floods (Obreja, 2012), calculation of hydrological and hydraulic parameters characteristic of floods using the HEC-RAS software (Armaş et al., 2013), and determining hydraulic parameters in relation to the bankfull discharge (Minea et al., 2011; Toroimac et al., 2012; Roşca et al., 2015).

The objectives of this study are as follows: (1) determining the hydraulic parameters characteristic of each flood; (2) determining the geomorphic effectiveness of floods in relation to their duration and magnitude; and (3) assessing the geomorphic impact associated with floods with various recurrence intervals.

2. STUDY AREA

The Trotuş drainage basin is located in the central sector of the Eastern Carpathians and forms a part of the largest drainage basin in Romania, i.e., the Siret basin (Fig. 1). In total, 57.5% of the basin area (of the total area of 4350 km^2), in the upper and middle sectors, is occupied by lithological outcrops composed of Carpathian flysch (in which sandstone, conglomerate, marl and clay are prevalent).

Until the 2000s, the Trotuş River was commonly regarded as a gravel bed river. Following the flood events that occurred between 2000 and 2010, which scoured the channel down to the in situ bedrock, the appearance of several middle and lower course gravel bed-bedrock reaches was documented.

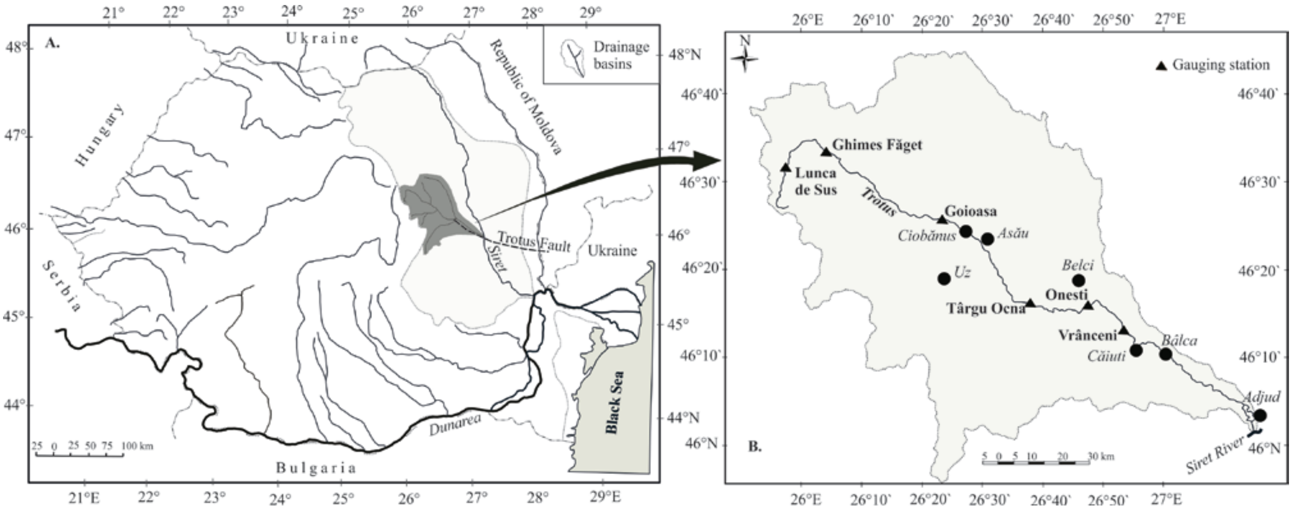

Figure 1. Location of Trotuş drainage basin in Romania and within Siret drainage basin (**A**); location of gauging stations (triangles) and settlements mentioned in the text (circles) within Trotuş drainage basin (**B**).

Table 1. Characteristic discharge values on River Trotuş

Gauging station	DRM	BA	MAD	Q_{bf}	PD	YMD
Lunca de Sus	146	88	0.885	8	23.2	1984
Ghimeş Făget	127	381	3.65	38	127	1975
Goioasa	106	781	6.5	63	353	2004
Târgu Ocna	69	2091	17	170	1.490	2005
Oneşti	54	2836	23	230	2.294	2005
Vrânceni	37	4092	35	325	2.845	2005

DRM = distance to the river mouth (km); BA = basin area (km^2); MAD = mean annual discharge (m^{3}s^{-1}); Q_{bf} = bankfull discharge (m^{3}s^{-1}); PD = peak discharge (m^{3}s^{-1}); YMD = year of maximum discharge.

The total length of the River Trotuş is approx. 160 km. For nearly 130 km from the headwaters, the river forms a well-delineated single-thread channel, with an average sinuosity index of 1.5. In the lower course, i.e., the last 30 km, the channel becomes braided. The average median diameter (D_{50}) of the channel bed sediments throughout the entire 160 km length is 88.6 mm, whereas the extreme values range from 490 mm (Asău) to 20 mm (Adjud) (Dumitriu, 2007; Dumitriu et al., 2011). The average channel slope gradient ranges from 0.17 m m^{-1} in the upper course (Lunca de Sus) to 0.018 m m^{-1} in the lower course (just downstream of Căiuţi). The channel width within the sector pertaining to the sinuous type ranges from several meters in the upper course to approx. 100 m in the lower course, whereas in the braided reach the channel width can be as great as 300 m. However, this parameter is different after each significant flood event.

The average long-term precipitation amounts to approx. 800 mm basin-wide, varying by ± 200 mm in the high mountain areas compared to lower areas. The interaction between the physical geographical traits of the study area and the circulation of air masses results in deviations in the distribution of monthly and annual precipitation. Such was the case in 2005, when the precipitation recorded during 11-13 July accounted for 100-150% of the multiannual average of July (Dumitriu, 2007).

The average multiannual discharge from the River Trotuş ranges from 0.9 m^3 s^{-1} in the upper course (Lunca de Sus gauging station) to 35 m^3 s^{-1} in the lower course (Vrânceni gauging station) (Table 1). Except for the gauging sites on the upper course of Trotuş (Lunca de Sus and Ghimeş Făget), for which the highest peak discharge values were recorded between 1975 and 1985, the stations recorded the highest peak discharges in 2004-2005. Discharge values recorded in 2005 are considered exceptional historical values, with an *Annual Exceedance Probability* (AEP) of 0.5%, which results in an *Average Recurrence Interval* (ARI) of 200 years. The highest peak discharge on the River Trotuş is considered to be the value recorded in 2005 (2845 m^3 s^{-1} at the Vrânceni gauging station), rather than the 3720 m^3 s^{-1} value recorded on 29 July, 1991, because the latter was not entirely generated by natural causes and was due in part to the failure of Belci dam (Podani & Zăvoianu, 1992).

The rise in the peak discharge values post-2000 can be attributed to the increasing amount of precipitation over a very short period of time

(Pleşoianu & Olariu, 2010). This could be a valid argument for the present situation, in which, of the top four peak discharge values recorded at the Vrânceni gauging station (during 64 years), two occurred during the study period (2845 m^3s^{-1} - 2005; 1700 m^3s^{-1} - 1975; 1567 m^3s^{-1} - 2010; and 1510 m^3s^{-1} - 1988).

3. DATA AND METHODS

Data employed in this study came from three sources: *(i)* the six gauging stations in operation along the river Trotuş (Lunca de Sus, Ghimeş Făget, Goioasa, Târgu Ocna, Oneşti and Vrânceni) under the management of the "Romanian Waters" National Administration, which provides data on daily mean discharge and suspended sediment values, maximum discharge values, characteristic levels, channel cross sections; *(ii)* direct field measurements performed between 1995 and 2013, which yielded information on the following variables: slope gradient, bed material grain size, active channel width and riverbank height. These variables were determined in the field to assess the state of the channel before and after flood events, whereas for the duration of the flood, we used the information provided by the water discharge summaries *(iii)* the third data source is based on the data provided by the first two sources. By employing equations established in the literature, we computed the values of the bankfull discharge, the mean annual flood, the specific stream power, and the critical shear stress. In addition, to compare the geomorphic impact of flood events between 2000 and 2012, we also used orthophotos taken prior to the flood events in 2005, 2010 and thereafter.

3.1. Bankfull discharge

The hydrological and geomorphological literature has introduced various definitions and methods for the calculation of bankfull discharge (Q_{bf}). Nevertheless, the definitions can be grouped into two categories (Radecki-Pawlik, 2002), the first of which defines the bankfull discharge in terms of the geometry of a cross-section, whereas the second describes it in terms of the volume of water. Some of the most widely cited definitions were introduced by Williams (1978), who regards the bankfull discharge as the flow that fills the active channel to the top of its banks, and Dunne & Leopold (1978), who define the bankfull discharge as the most effective flow in forming and/or maintaining average channel dimensions and which has an average recurrence interval of 1.5 years (Ma et al., 2010). In the present study, the value of the bankfull discharge (table 1) is calculated by applying the Parker et al., (2007) equation:

$$Qbf = 3.732\, B_b H_b \sqrt{g H_b S} \left(\frac{H_b}{D_{s50}} \right)^{0.2645} \quad (1)$$

where B_b (m) is the bankfull channel width, H_b (m) is the bankfull cross-sectionally averaged channel depth, g is the gravitational acceleration equal to 9.8 m s^{-2}, and D_{s50} (mm) is the bed surface median grain size.

3.2. Mean annual flood, catastrophic floods and geomorphic stream flows

The mean annual flood (MAF) was defined as either the arithmetic average of all annual floods for the recorded gage period (or other specified time interval), the 2.33-year recurrence interval flood (Kale, 2007), or the 1-year recurrence interval flood (Tomkins et al., 2007). Erskine & Saynor (1996) showed that floods in which the flood peak discharge is 10 times greater than the mean annual flood can be termed *catastrophic floods (Cf)*. Furthermore, they are divided into floods that barely exceed the critical threshold, termed Cf_1, and larger catastrophic events (>100 year ARI), termed Cf_2.

The geomorphological effect of floods can be evidenced by the ratio between peak discharge and mean annual flood or by the *flash flood magnitude index* (FFMI) values. When the peak discharge is at least 10 times greater than the mean annual flood, the changes in the channel are significant and have long-term effects (Erskine & Saynor, 1996).

At the Vrânceni gauging station, the *geomorphic stream flow* (sensu Gaeuman et al., 2003) amounts to 300 m^3 s^{-1}. Although this value is very close to the bankfull discharge, the two should not be mistaken. The sum of the differences between the daily mean discharge in a given year that exceeds a certain threshold (commonly set as the 1.5-year recurrence interval peak discharge) and the value of the respective threshold represents the *volume of geomorphic stream flow* (Gaeuman et al., 2005). To determine this parameter, we considered two reference years, i.e., 2000 (during which the lowest peak discharge value throughout the gage period was recorded - 103 m^3 s^{-1}) and 2005 (during which the highest peak discharge was recorded - 2845 m^3 s^{-1}). In 2000, of the total annual flow of 21×10^6 m^3, the annual geomorphic flow was null. In contrast, in 2005, of the total annual flow that amounted to 76×10^6 m^3, the annual geomorphic flow accounted for 26% or 20×10^6 m^3. Of this volume, 17×10^6 m^3 (85%) was conveyed through and acted on the river channel from 11 to 14 July, 2005. In 2010, the annual geomorphic flow was 19×10^6 m^3, accounting for 25% of the total annual flow (74×10^6 m^3).

3.3. Specific stream power

As indicated by O'Connor (1993), the unit or specific stream power (SSP) is directly related to the local rates of energy expenditure and reflects the physical capability of performing geomorphic work (Benito, 1997). Specific stream power provides a good indicator of geomorphic effectiveness (Barker et al., 2009). Specific stream power is computed using the equation introduced by Bagnold (1980):

$$\omega = \Omega/w = \gamma QS/w \qquad (2)$$

where ω is the specific stream power (W m^{-2}), Ω is total stream power per unit channel length (W m^{-1}), w is the width of the water surface (m), γ is the specific weight of water (9810 N m^{-1}), and Q is the discharge (m^3 s^{-1}).

Based on the knowledge that four of the characteristics of stream power are strongly correlated with the changes occurring in the channel (Julian & Torres, 2006; Julian et al., 2012), we proceeded to calculate them. These four properties are as follows: event peak (in Wm^{-2}) = ω_{max}; magnitude (in Wm^{-2}) = $\Sigma\omega$ when $Q>Q_{2.33}$ ($Q_{2.33}$ represents the threshold above which we compared ω to channel geometry); duration (in days) = time in which $Q>Q_{2.33}$; and variability = number of individual flood events $>Q_{2.33}$.

We further computed the *critical stream power* to have a better understanding of the influence of stream power on sediment transport, based on the equations of Costa (1983) for equation (3) and Williams (1983) for equation (4):

$$\omega_{cr} = 0.009D_{50}^{1.686} \qquad (3)$$

$$\omega_{cr} = 0.79d^{1.27} \quad 10 \leq d \leq 1500 \qquad (4)$$

where ω_{cr} is the critical stream power (Wm^{-2}).

3.4. Shear stress

The shear stress can be defined, in very broad terms, as a measure of stream competency. A specific particle will be displaced only when the shear stress acting on that particle is greater than the resistance forces opposing its movement. The shear stress is calculated by employing the equation introduced by Du Boys (1879):

$$\tau = \gamma RS \qquad (5)$$

where τ is the shear stress (Nm^{-2}).

The magnitude of shear stress required to move a given particle is known as the critical shear stress (τ_{cr}). The stream competency required to displace sediments of a particular size can be quantified by using the Shields (1936) equation, hence the designation *critical Shields stress* (Knighton, 1998):

$$\tau_{cr} = \tau_{ci} g (\rho_s - \rho_w) d \qquad (6)$$

where τ_{cr} is the critical shear stress (N m^{-2}), τ_{ci} is the dimensionless critical shear stress (N m^{-2}), ρ_s is the density of the sediment (the value 2650 kg m^{-3} is usually used), ρ_w is the density of water (which is dependent on temperature and usually approx. 1000 kg m^{-3}), and d is the size of the particle of interest (m). Shields'studies have shown that in gravel bed channels of homogeneous sediment sizes and turbulent flow, the value of dimensionless critical shear stress is 0.06.

Based on the equation developed by Shields, several equations were developed to assess the critical shear stress in natural channels with heterogeneous substrate sizes (Fischenich, 2001). These include the equations introduced by Julian (1995), of which the following is adequate for the particular conditions of the Trotuş channel bed (gravel and cobbles Bed River):

$$\tau_{cr} = 0.06(\rho_s - \rho_w) d \mathrm{Tan}\theta \qquad (7)$$

where θ is the angle of repose of the particle.

4. RESULTS AND DISCUSSIONS

4.1. Hydraulic floods characteristics

4.1.1. Mean annual flood

Because the arithmetic average of all annual floods and the 2.33-year recurrence interval flood have relatively similar values, the mean annual floods on the Trotuş are as follows: Lunca de Sus – 11.5 m^3 s^{-1}; Ghimeş Făget – 47 m^3 s^{-1}; Goioasa - 80 m^3 s^{-1}; Târgu Ocna - 270 m^3 s^{-1}; Oneşti - 400 m^3 s^{-1}; and Vrânceni - 650 m^3 s^{-1}. The peak discharge values recorded at each gauging station indicate that the large floods that occurred between 2000 and 2012 were 2 to 6 times greater than the mean annual floods. Conversely, if we consider the 1-year recurrence interval flood, the values of the mean annual floods are as follows: Lunca de Sus - 4 m^3 s^{-1}; Ghimeş Făget –17 m^3 s^{-1}; Goioasa - 28 m^3 s^{-1}; Târgu Ocna - 68 m^3 s^{-1}; Oneşti - 100 m^3 s^{-1}; and Vrânceni -150 m^3 s^{-1}. During the 2000-2010 interval, the peak discharge at the Vrânceni gauging station was approx. 20 times higher than the mean annual flood of 2005, and 10 times higher than that of 2010, which explains the significant changes in the river channel (Fig. 2). The FFMI value (calculated according to Baker, 1977) at the Vrânceni station for the 2000-2010 interval was 0.32, which is above the worldwide average value (McMahon et al., 1992) and indicates that large floods have average variability in the case of the river Trotuş. This variability possibly results in the occurrence of catastrophic floods that can substantially alter the channel morphology.

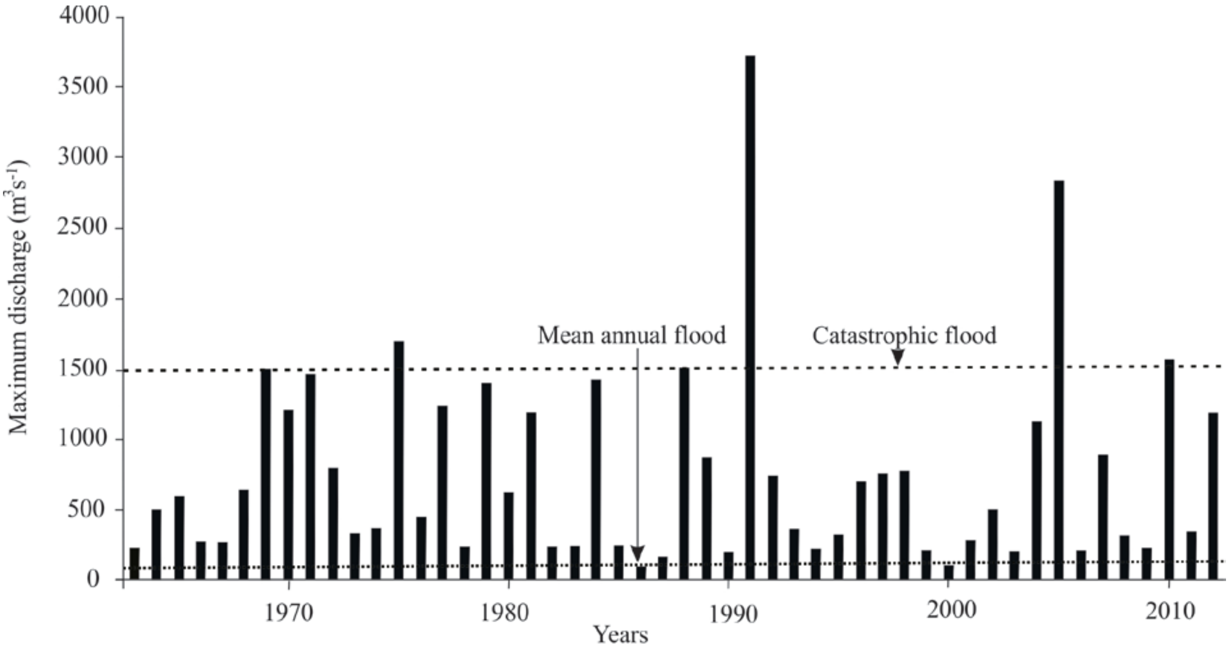

Figure 2. Mean annual flood and catastrophic flood at Vrânceni gauging station. Catastrophic floods were identified using the definition of Erskine and Saynor (1996): flood peak 10 times greater than the mean annual flood.

4.1.2. Catastrophic flood

If we approach the mean annual flood in terms of the 2.33-year recurrence interval, we can conclude that no catastrophic floods have occurred in Trotuş drainage basin. If we consider the mean annual flood to be equal to the 1-year recurrence interval discharge, we have the following situation: Lunca de Sus – no catastrophic floods; Ghimeş Făget – no catastrophic floods; Goioasa – 3 catastrophic floods in the C_1 category (2004 – 353 $m^3 s^{-1}$; 2005 - 331 $m^3 s^{-1}$; and 2010 - 296 $m^3 s^{-1}$); Târgu Ocna – 3 catastrophic floods, of which 2 rank in the C_1 category (1991 – 772 $m^3 s^{-1}$ and 2004 - 682 $m^3 s^{-1}$) and 1 in the C_2 category (2005 – 1490 $m^3 s^{-1}$); Oneşti – 2 catastrophic floods, of which 1 ranks in the C_1 category (1991 – 1290 $m^3 s^{-1}$) and 1 in the C_2 category (2005 – 2294 $m^3 s^{-1}$); and Vrânceni – 6 catastrophic floods, of which 4 are in the C_1 category (1969 – 1503 $m^3 s^{-1}$; 1975 - 1700 $m^3 s^{-1}$; 1988 – 1510 $m^3 s^{-1}$; and 2010 - 1567 $m^3 s^{-1}$) and 2 in the C_2 category (1991 – 3720 $m^3 s^{-1}$ and 2005 –2845 $m^3 s^{-1}$).

4.1.3. Specific stream power and shear stress

The geomorphic significance of the data thus obtained was determined according to certain critical threshold values cited in the literature. Because the main focus of this study is the geomorphic effectiveness of major floods, we used 300 Wm^{-2} and 100 Nm^{-2} as the minimum thresholds for specific stream power and shear stress, respectively, because these values are associated with major morphological changes in the channel as defined by Miller (1990) and Magilligan (1992). This set of values is widely used, although Meyer (2001) showed that major changes in river channels can occur at stream power values ranging from 50 to 200 Wm^{-2}.

The magnitude of channel adjustments after a flood event is closely related to a number of properties of specific stream power (Julian et al., 2012). The four properties of specific stream power were computed for the Târgu Ocna and Vrânceni gauging stations. The yielded results were as follows: ω_{max} =1550 W m^{-2} and 671 W m^{-2}, respectively; $\Sigma\omega$ =4117 Wm^{-2} and 1771 Wm^{-2}, respectively; duration = 5 and 4 days, respectively; variability = 7 (1 in 2004; 2 in 2005; 1 in 2007; and 3 in 2010) and 10 (2 in 2004; 3 in 2005; 2 in 2007; 2 in 2010; and 1 in 2012), respectively. The values of the four characteristics of stream power estimated for the Târgu Ocna and Vrânceni gauging stations, particularly the magnitude and the duration, account for the more evident changes in the midcourse channel of the Trotuş compared to the lower course. The flood magnitude (4117 W m^{-2} at Târgu Ocna and 1771 W m^{-2} at Vrânceni) is regarded as the best predictor of erosion rates, especially when banks are composed to a large extent of non-cohesive sediments (Julian & Torres, 2006). The changes appear more evident because within a relatively short period of time (i.e., 2005-2010), three large floods occurred, which prevented the channel from being restored to its pre-2005 state. Moreover, these events boosted the erosion processes. If during a certain time frame (e.g., in the case of river Trotuş, 1991-2004) there are no large floods that increase stream power, the channel can be restored to its previous size (particularly the channel width) prior to the flood event (VanLooy & Martin, 2005). The time required for the recovery of its previous size varies depending on the type of channel (gravel bed or sand-gravel bed), the climate, etc., and can range from several years to 40 years (Curtis & Whitney, 2003). According to Brunsden & Thornes (1979), an accurate indicator of the recovery period is the *transient form ratio* (TFR). When TFR is above or equal to 1, the channel in unstable because it must perpetually adjust due to high flood frequency. The lower the TFR (<1), the more stable the channel is over a longer period of time, during which instability may only occur locally (Nanson & Erskine, 1988). The TFR and FFMI are strongly correlated (Erskine, 1996). Thus, we can conclude that within the Vrânceni reach, the channel of the Trotuş River is overall stable, although large floods can result in marked local instability.

The values of stream power can be related to certain aspects of river channel dynamics (Petit et al., 2005a) or may be used as a criterion for devising a hydro-geomorphological typology of rivers (Schmitt et al., 2001). In the case of the Trotuş, the calculated values of specific stream power are comparable to or lower than those estimated for other rivers (Krapesch et al., 2011). Apparent changes occurred in the Trotuş channel in 2005 and 2010, during which the values of stream power were the highest in the entire study period (641 Wm^{-2} and 360 Wm^{-2}, respectively). In both instances, the river performed an impressive amount of geomorphological work.

At the Vrânceni gauging station, the temporal distribution of daily specific stream power during the entire study period (4749 days) is as follows: for 4307 days (90.7% of the study period), this parameter, ω, was below 15 W m^{-2}; for 321 days (6.7%), ω ranged from 15 to 30 W m^{-2}; for 121 days (2.5%), ω was above 30 W m^{-2}; for 17 days (0.3%) ω was above 100 W m^{-2}; for 7 days (0.1%) ω was above 200 W m^{-2}; and for 4 days (0.01%), ω was above 300 W m^{-2}. The power classes employed are identical to those introduced by Petit et al. (2005b). The prevalence of daily specific stream power values lower than 15 W m^{-2} (over 90% of the analysed period, with an average value of just 7.5 W m^{-2} for the 13-year period) indicates the channel is rather stable and less active in this reach (Brookes, 1988). Against this background, the changes that occurred during the 4-day interval when the stream power exceeded 300Wm^{-2} reveal the effectiveness of extreme flood events, which confer a certain degree of instability to some reaches of the Trotuş channel. Overall, however, the channel can be ranked in the dynamic stable channel category, in which some changes to the channel bed and banks are expected (Cotton, 1999). The values of the shear stress vary depending on the flood magnitude (Caruso et al., 2013). However, on the river Trotuş, the differences are rather small compared to other rivers. For example, at the Vrânceni gauging station, the shear stress during the 2005 flood (2845 $m^3 s^{-1}$, a 200-year flood) was just 17% higher compared to the 2010 flood (1567 m^3s^{-1}, a 20-year flood). A large share of a river's energy is employed to overcome the resistance of bedforms; thus, the τ_{cr}/τ ratio is an expression of this process, acting as an indicator of the energy loss due to the resistance of bedforms (Petit et al., 2005a). The lower the value of this ratio, the higher the extent to which the shear stress is converted to bedform shear stress, which results in a decrease in the grain shear stress and in the energy available for bed material displacement and transport (Petit, 1990). On the Trotuş, the lowest values of the τ_{cr}/τ ratio occur in the upper and lower courses (ranging from 0.33 to 0.5), whereas the highest values are recorded in the midcourse (0.86 at Goioasa). This value distribution indicates that the amount of energy available for sediment displacement and movement reaches a maximum value in this sector (Table 2).

The higher values of the τ_{cr}/τ ratio estimated for the 2010 flood compared to the 2005 event could account for the significant degradation of the channel bed in the middle course, which reached bedrock in some patches. The critical stream power and the critical shear stress can be regarded as geomorphological thresholds whose exceedance can result in sudden or gradual changes in the channel (Bull, 1979). Thus, they can be used as indicators of stability, even when the channel bed is composed of various materials whose threshold of movement varies greatly (Beyer, 1998).

Table 2. The shear stress for peak discharge.

Gauging station	S	D_{50s}	τ_{cr}	τ				τ_{cr}/τ	
				2004	2005	2007	2010	2005	2010
Lunca de Sus	0.015	87	75	-	-	82	149	-	0.50
Ghimeş Făget	0.012	96	83	-	130	-	108	0.64	0.77
Goioasa	0.007	119	103	157	139	104	120	0.74	0.86
Târgu Ocna	0.005	126	110	111	167	130	149	0.66	0.74
Oneşti	0.004	48	41	-	125	50	86	0.33	0.48
Vrânceni	0.002	83	38	-	103	71	85	0.37	0.45

S = channel slope (m m^{-1}); D_{50s}= median grain sizes of the surface (mm); τ_{cr} = critical shear stress (N m^{-2}); τ = shear stress for peak discharge (N m^{-2})

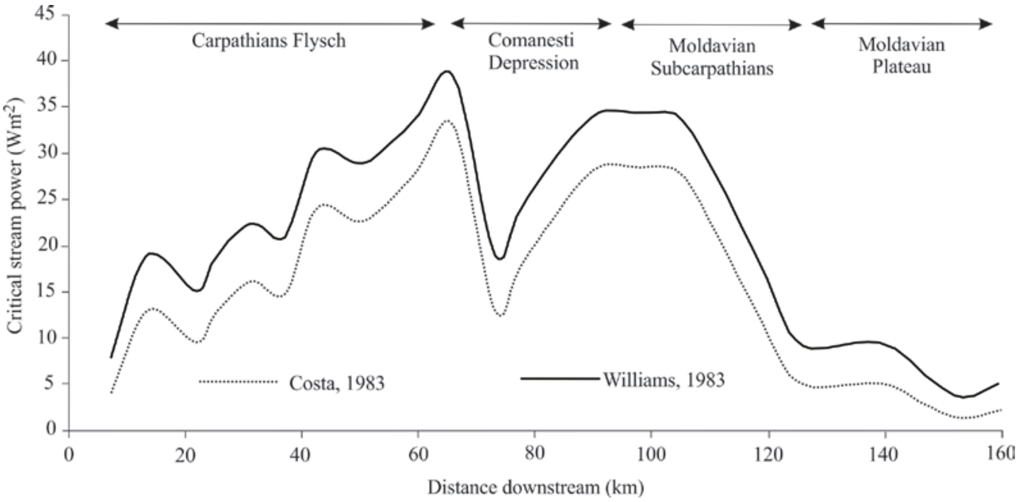

Figure 3. The critical stream power along river Trotuş computed based on the equations of Costa (1983) and Williams (1983).

When the shear stress exceeds the critical shear stress value, some of the bed material is removed, possibly resulting in changes in channel morphology, which in turn may induce some instability in the channel. However, the exceedance of these thresholds does not necessarily result in apparent changes of the channel (Graf, 1983). By comparing the effects of floods in 2005 and 2010 on the Trotuş channel, we determined that the flood in which the Q_{bf} exceedance period was greater generated the most significant changes in the channel (i.e., the 2010 flood) rather than the event with the highest peak discharge (2005).

Along the Trotuş channel, the correlation of D_{50} and the critical specific stream power highlights the influence of the main landforms traversed by the river on the traits of the bed material (Fig. 3). The peak values of this parameter occur in the gorge sectors located both upstream and downstream of the Comăneşti depression. In the reaches where the valley narrows, the hillslopes supply coarser material to the channel such that D_{50} exceeds 100 mm. Thus, local sediment sources may alter to a large extent the normal grain size distribution, particularly when the size of the supplied material differs from that of upstream source areas (Knighton, 1980). When the Trotuş is compared to rivers of the same size range, the values of the critical specific stream power are lower compared to smaller rivers and come very close the values yielded by applying the equations of Costa and Williams (Petit et al., 2000). The highest shear stress values were recorded during the 2005 flood at all gauging stations, with the exception of Goioasa, whose peak value occurred in 2004 (Table 2). The maximum value recorded throughout the basin was 167 Nm^{-2} at the Târgu Ocna streamgage.

4.2. Geomorphic effectiveness of floods in relation to their duration and magnitude

A wide range of parameters can be employed (i.e., magnitude, duration, frequency, etc.) to establish the defining elements of a flood. Among these, flood magnitude and duration appear to play major roles in defining the channel form. Moreover, duration alone can explain how floods with lower peak discharges, shear stresses or stream powers can have, within the same alluvial channel, greater geomorphological impacts compared to floods with higher instantaneous values of the same parameters (Costa & O'Connor, 1995).

On the Trotuş river, the changes in the channel were far more visible in the aftermath of the 2010 flood compared to the 2005 flood, despite the fact that the magnitude of the 2005 event was considerably greater (i.e., twice as large on average). This observation confirms that magnitude is a poor predictor of geomorphological work performed by a particular flood (Thompson & Croke, 2013). In addition, Costa & O'Connor, (1995) noted that in some cases the changes in the channel are not consistent with the peak discharge of the flood but rather with the "*time over threshold*", which better accounts for the amount of geomorphological work.

By analysing the data recorded at the gauging stations along the Trotuş River, we documented that flood duration is the variable that best explains the similar changes in the channel when flood magnitude differs greatly. Thus, the average duration of the 2005 flood was approx. 280 hours (with an average magnitude of 1410 m^3 s^{-1}), whereas the average duration of the 2010 flood was 1300 hours (with an average magnitude of 635 m^3 s^{-1}). At all 6 gauging stations, the duration of the 2010 flood was 3 up 5 times longer in the upper course, and approx. 17 times longer in the middle course compared to the 2005 flood (Fig. 4).

Some surprisingly visible effects occurred following the flood event in 2012, particularly in the middle course of the Trotuş. Although the magnitude and duration of this flood would not have predicted the extent of the changes, the relatively long duration of the event (approx. 600 hours in the middle course) accounted for its geomorphological impact. Thus, floods with very similar magnitudes and frequencies (in our case, the events in 2007 and 2012) can generate very different geomorphological effects even within the same basin (Fuller, 2008). Therefore, it is a rather difficult task to predict the likely response of the channel to a flood of a certain magnitude and frequency (Thompson & Croke, 2013).

Major geomorphological changes only occur in river channels when a certain erosional threshold is exceeded (Magilligan, 1992). If the threshold exceedance duration is large enough, even a moderate magnitude flood may cause significant changes in the channel (Dean & Schmidt, 2013). An interesting comparison can be drawn between the flood events in 2005 and 2010 in terms of these thresholds. The average duration of the exceedance of the Miller-Magilligan threshold amounted to 40 hours during the 2005 flood and 20 hours during the 2010 flood, whereas the average duration of the exceedance of the bankfull discharge (Q_{bf}) was 64 hours in 2005 and 110 hours in 2010 (Fig.4 and Tab.3). The consequences of the two flood events in terms of the major changes experienced by the channel were very similar, despite the considerably higher magnitude and the average duration of exceedance of the Miller-Magilligan threshold in 2005. Thus, it is further confirmed that the duration of a flood event can compensate for a lower magnitude (Costa & O'Connor, 1995). Moreover, under certain conditions, significant channel changes may also occur at stream power values below the Miller-Magilligan threshold (Meyer, 2001), as was likely the case during the flood in 2010.

4.3. The geomorphological impact of floods

The geomorphic effectiveness of floods has been defined in various ways (Dean & Schmidt, 2013). Thornbury (1954) and Miller (1987) suggested that this parameter is proportional to the extent to which the landscape is modified by that particular flood event, whereas Wolman and Miller (1960) defined geomorphic effectiveness as the geomorphological work performed during a flood event. Wolman & Gerson (1978) believe that the geomorphic effectiveness of a flood event is the sum of the geomorphological changes in the channel generated by the flood, to which they added the time required for the channel to revert to its pre-flood state. The hydraulic variables with which it is determined whether a flood was geomorphically effective primarily include the specific stream power and the total shear stress (Baker & Costa, 1987). The geomorphological impact of major flood events is rather variable, even along the same river channel. Therefore, this impact is commonly considered to be spatially discontinuous because reaches that are strongly affected may alternate with reaches that have undergone only minor changes (Croke et al., 2013). Basin-wide, the 2005 flood may be regarded as the event with the highest geomorphological impact throughout our study period and beyond. However, in certain reaches, the changes generated by the 2010 and 2012 flood events were more prominent compared to the effects of the previous event (2005), which confirms that in some instances there is a low correlation between flood magnitude and its geomorphological impact (Krapesch et al., 2011; Thompson & Croke, 2013).

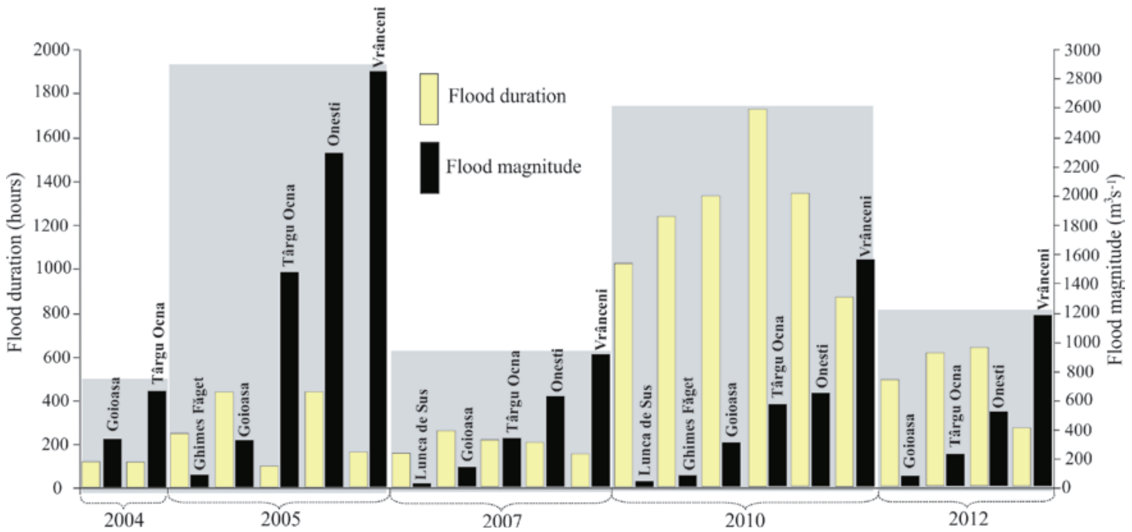

Figure 4. Flood duration versus flood magnitude on river Trotuş at the gauging stations.

Table 3. Hydraulic and geomorphic data for flood events occurring between 2000 and 2012

Year	Flood event	Gauging station	Peak discharge (m^3s^{-1})	Recurrence interval (years)	Duration above the Q_{bf} (hours)	Duration above the threshold of 300 Wm^2 (hours)
2004	1	Goioasa	353	50	17	7
	2	Târgu Ocna	682	20	12	6
2005	3	Ghimeş	90.6	50	70	30
	4	Goioasa	331	50	50	40
	5	Târgu Ocna	1490	200	63	42
	6	Oneşti	2296	200	65	42
	7	Vrânceni	2845	200	72	45
2007	8	Lunca de Sus	9.55	5	6	0
	9	Goioasa	135	5	22	3
	10	Târgu Ocna	331	5	24	9
	11	Oneşti	610	5	15	2
	12	Vrânceni	900	5	25	0
2010	13	Lunca de Sus	17.6	50	80	0
	14	Ghimeş	85.5	50	85	24
	15	Goioasa	296	50	222	48
	16	Târgu Ocna	587	10	85	30
	17	Oneşti	641	10	20	5
	18	Vrânceni	1567	20	170	11
2012	19	Goioasa	80.3	5	15	0
	20	Târgu Ocna	218	5	14	0
	21	Oneşti	513	5	46	0
	22	Vrânceni	1185	10	46	5

1= 26th – 31st July; 2= 26th – 31st July; 3=8 th-17th July; 4 = 9th-27th July; 5 = 11st-15th July; 6 = 9th-27th July; 7 = 10th -17th July; 8 = 23rd- 29th October; 9 = 21st – 31st October; 10 = 20th – 29th October; 11 = 22nd - 31st October; 12 = 21st – 27th October; 13 = 20th June -1st August; 14 = 14th May – 4th July; 15 = 18th May – 12nd July; 16 = 7th May – 20th July; 17 = 22nd June – 16 th August; 18 = 25th June – 31st July; 19 = 12nd May – 10th June; 20 = 18th May – 11st June; 21 = 23rd May – 18th June; 22 = 24th May – 4th June.

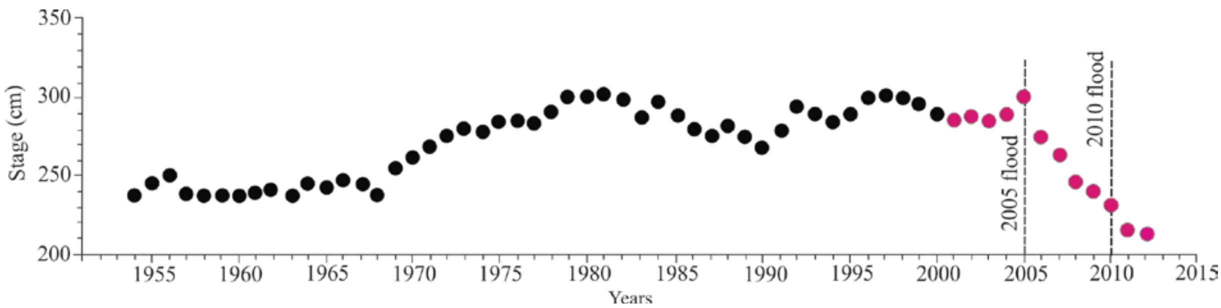

Figure 5. Variation in stream stage for mean annual discharge (35 $m^3 s^{-1}$) at Vrânceni gauging station.

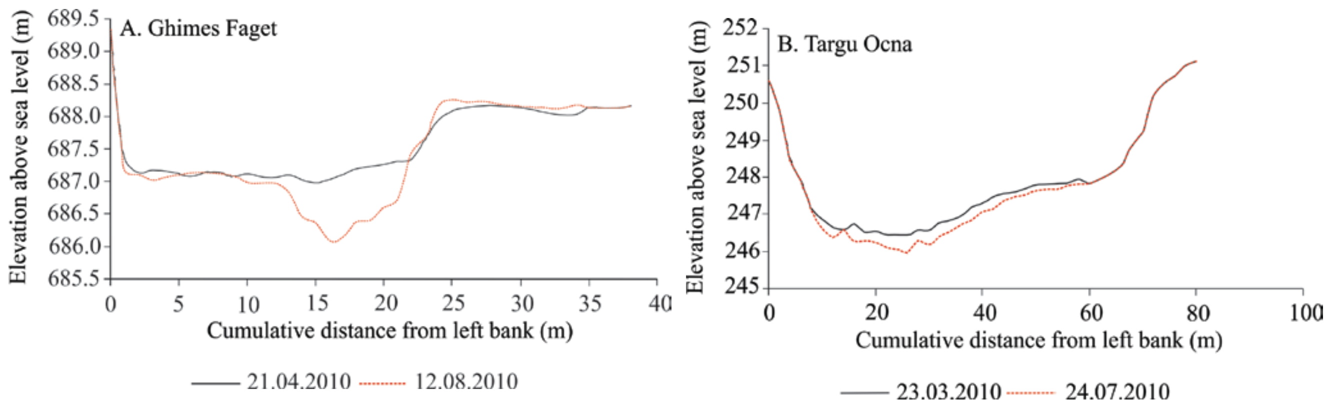

Figure 6. Monitored cross-channel sections before and after 2010 flood on the Trotuş River at Ghimeş Făget (A) and Târgu Ocna (B)

The variable geomorphological impact of the same flood or of distinct flood events in terms of their spatial and temporal manifestations can be attributed to several factors, such as the local channel configuration, the resistance of constituent rocks, and the presence or absence of vegetation (Emmett & Wolman, 2001; Fuller, 2007). In the case of the river Trotuş, the situation observed in the field was very consistent with the results obtained by data processing, both of which confirm that channel widening and degradation/aggradation are the main response of the channel to floods occurring during the study period. Moreover, examples cited in the literature commonly include these two types of processes by which the channel adjusts to post-flood events (Lenzi et al., 2006; Juracek & Fitzpatrick, 2009; Molnar et al., 2010; Bowen & Juracek, 2011; Rădoane et al., 2010, 2013). The changes in the channel bed elevation during the flood events occurring between 2000 and 2012 were determined by employing the method introduced by Juracek (2002) (Fig.5). According to this method, channel bed aggradation or degradation was inferred from changes in the stage associated with the mean annual discharge for the period of record. Also, cross-sectional profiles across the Trotuş River showing changes in channel bed elevations due to aggradation or scouring (Fig. 6).

In the upper course (where the general tendency is channel bed degradation) and the midcourse (where channel bed aggradation is prevalent), the 2005 flood resulted in very low aggradation or degradation, but in the Târgu Ocna–Căiuţi reach, in the lower course, the 2005 and 2010 floods led to significant channel bed degradation. The results obtained using this method are fully consistent with the situation documented in the field, which confirms the visible tendency towards channel bed degradation, favouring the gravel bed to gravel-bed-bedrock river transition. According to data from the Vrânceni gauging station, from 2005 to 2012 the channel bed deepened by 0.85 m (i.e., by an annual rate of 0.12 cm y^{-1}), which is in the same range as other European rivers (Rinaldi et al., 2005). Moreover, the degradation rates on the Trotuş are very similar to those reported for the Siret river after the 2010 flood (Obreja, 2012).

However, the yielded value refers exclusively to the Vrânceni streamgage. In certain reaches, the deepening of the channel bed following the 2005, 2010 and 2012 flood events was much greater (i.e., by as much as 1 m) (Fig. 7). Some of the reasons for the changes in channel bed elevation on the river Trotuş were indicated by Rădoane et al., (2013).

Further channel adjustments generated by discharge values with recurrence intervals below 200 years between 2000 and 2012 included the widening of both the wetted channel and the active channel. To illustrate this, we selected a 1 km-long reach located at the Bâlca stream –Trotuş confluence (Fig. 1). Before 2005, the maximum width of the wetted channel was approx. 60 m, and the active channel was approx. 200 m (Fig. 8A). After the 2005 flood, the former amounted to as much as 160 m, and the latter 250 m (Fig. 8B). Following the 2010 flood, the two parameters rose to 210 m and 300 m, respectively (Fig. 8C). Thus, after the two flood events, the wetted channel widening amounted to approx. 71% and the active channel widening to approx. 33%. The area of bars nearly doubled during the study period. Between 2005 and 2012, four major floods occurred in the lower Trotuş course, which prevented the channel from being restored to its initial state due to the insufficient recovery time (Harvey, 2007). The changes in the channel after the 2010 and 2012 flood events were also similar to the ones reported after the 2005 flood. Between pre-flood 2005 and post-flood 2010, the channel migrated towards the right bank of the analysed reach by as much as 150 m and towards the left bank by as much as 100 m (Fig. 8D).

Prior to the 2005 flood, the maximum distance between the road and the wetted channel bank was approx. 275, and the minimum distance was 125, whereas post-flood 2010, these had been reduced to 175 m, and 75 m, respectively. Channel widening, regarded as a response to major flood events, appears to be an adjustment common to all rivers. However, channel widening rates following flood events differ greatly between rivers. Fuller (2008) showed that a flood with a 100-year recurrence interval resulted in a wetted channel widening of 171% and an active channel widening of 500% on Kiwitea Stream (New Zealand). Increases in channel width ranging from 4% to 300% have also been reported by Wallick et al., 2007 and Toone et al., 2014.

Considering the geomorphological impact of the 2005 and 2010 flood events on the Trotuş River channel (particularly in the lower course), both floods can be considered severe impact events (Miller, 1990), despite the fact that they differed greatly in terms of magnitude and recurrence interval. The literature reports similar cases, in which the geomorphological impact of floods with smaller recurrence intervals is similar to or even exceeds the impact of floods with longer recurrence intervals (Hickin & Sichingabula, 1988; Bryant & Gilvear, 1999). In these instances, as well as in the case of the Trotuş River, it appears that flood duration leaves a stronger mark on the geomorphological changes than flood magnitude.

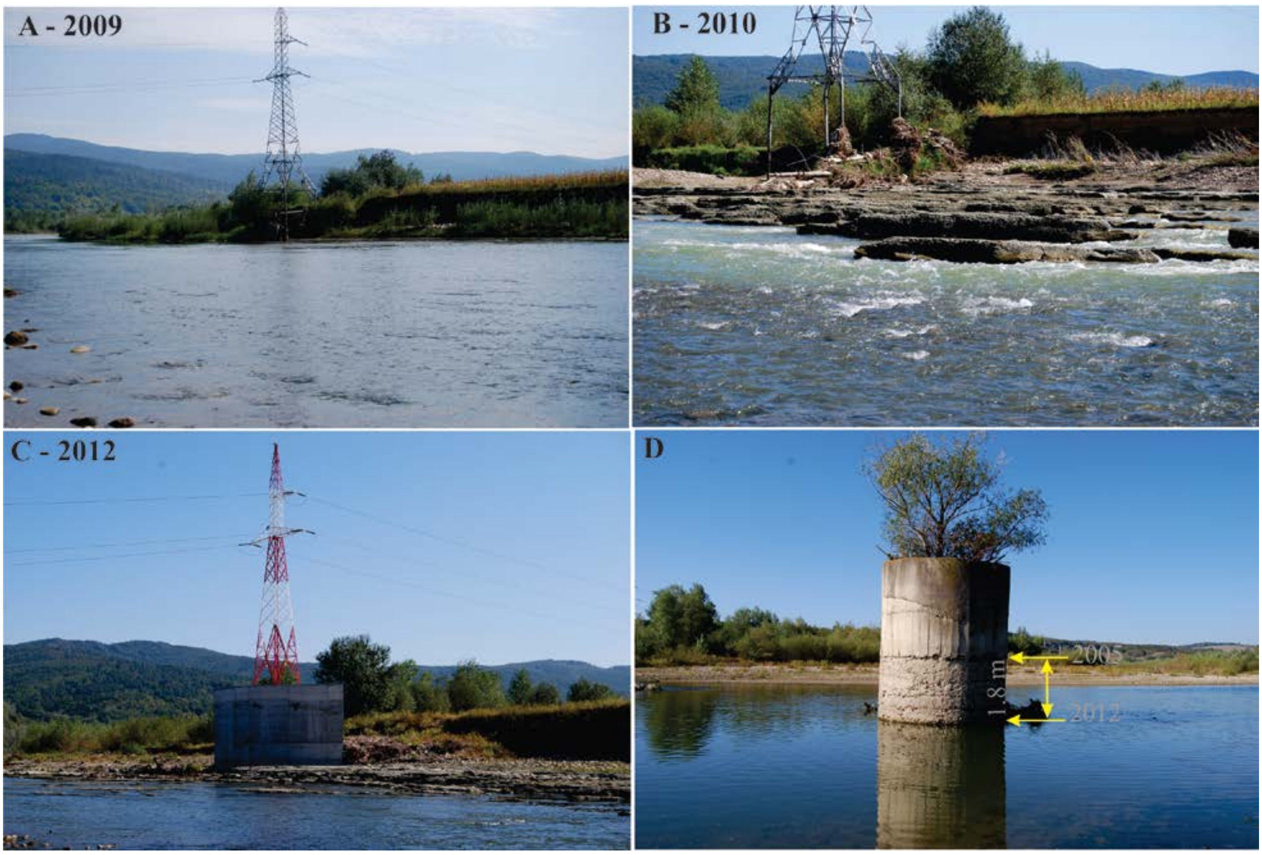

Figure 7. Channel bed degradation and the gravel bed – gravel-bed-bedrock river transition on Trotuş river at Căiuţi between 2009 and 2012 (**A, B** and **C**); River Trotuş 35 km upstream of the confluence with River Siret, incision of 1.8 m (**D**) (photo D. Dumitriu).

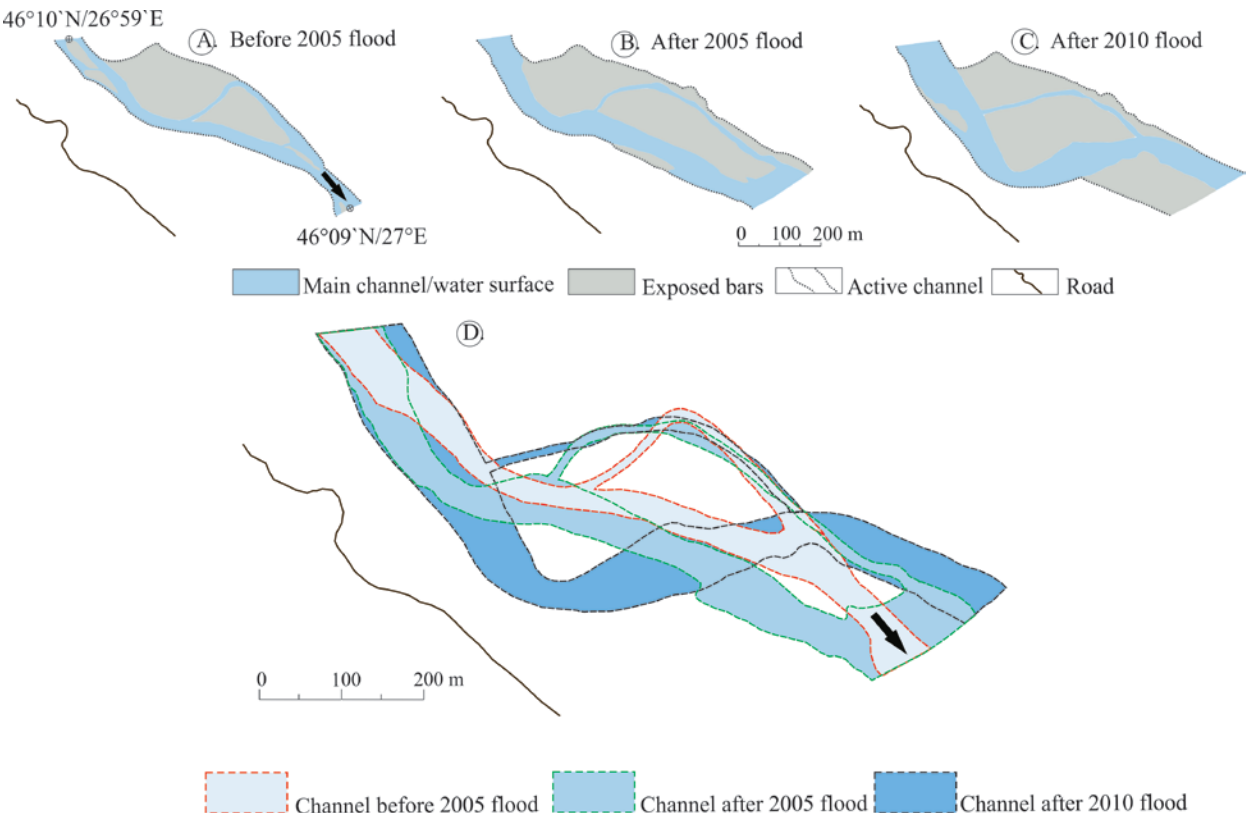

Figure 8. Channel planform configuration of Bâlca reach before and after the 2005 and 2010 floods.

5. CONCLUSIONS

The specific objectives of this study were to determine the hydraulic parameters of floods occurring between 2000 and 2012 and assess the changes they generated in the Trotuş River channel. Of the 6 catastrophic floods (*sensu* Erskine & Saynor, 1996) recorded between 1950 and 2012 at the Vrânceni gauging station, two (among which the 2845 $m^3 s^{-1}$ historical discharge value in 2005, with a 200-year recurrence interval) occurred during the study period and within a short time span, which accounts for the major changes in the channel. The hydraulic parameters of the flood events recorded between 2000 and 2012 largely explain the adjustments in the channel. Thus, the peak values of the daily specific stream power and the shear stress recorded during the 2005, 2010 and 2012 floods were above the Miller-Magilligan threshold, the exceedance of which can result in major geomorphological changes. The geomorphological impact of floods consisted primarily in channel widening and channel bed aggradation or degradation. In the upper and middle courses, floods resulted in very low levels of aggradation or degradation, whereas in the Târgu Ocna–Căiuţi reach, the floods led to significant channel bed degradation. Although they differed greatly in terms of magnitude, the 2005 and 2010 floods can both be ranked as severe events (*sensu* Miller, 1990) due to their geomorphological impact.

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REFERENCES

Armaş, I., Gogoaşe-Nistoran, D.E., Osaci-Costache, G. & Braşoveanu, L., 2013. *Morpho-dynamic evolution patterns of Subcarpathian Prahova River (Romania)*. Catena 100, 83-99.

Bagnold, R.A., 1980. *An empirical correlation of bedload transport rates in flumes and natural rivers*. Proceedings of the Royal Society of London 372A, 453– 473.

Baker, V. R., 1977. *Stream channel response to floods with examples from central Texas*. Geol. Soc. Am. Bull. 88, 1057– 1071.

Baker, V.R. & Costa, J.E., 1987. *Flood power*, in: Mayer, L., Nash, D. (Eds.), Catastrophic Flooding, Allen and Unwin, London, pp. 1–21.

Barker, D. M., Lawler, D. M., Knight, D.W., Morris, D. G., Davies, H. N. & Stewart, E. J., 2009. *Longitudinal distributions of river flood power: the combined automated flood, elevation and stream power (CAFES) methodology*. Earth Surf. Proc. Land. 34, 280-290.

Benito, G., 1997. *Energy expenditure and geomorphic work of cataclysmic Missoula flooding in the Columbia River gorge, USA*. Earth Surf. Proc. Land. 22, 457–472.

Beven, K. J., 1981. *The effect of ordering on the geomorphological effectiveness of hydrological events.* Inter. Assoc. Sci. Hydr. Pub. 132, 510-526.

Beyer, P.J., 1998. *Spatial variability in stability thresholds, Verde River, Arizona*. Middle States Geographer 31, 1-14.

Bowen, M. & Juracek, K.E., 2011. *Assessment of the geomorphic effects of large floods using streamgage data: the 1951 floods in Eastern Kansas, USA*. Phys. Geogr. 32, 52-77.

Brookes, A., 1988. *Channelized rivers, perspectives for environmental management*. John Wiley and Sons, New York, 326 p.

Brunsden, D. & Thornes, J.B., 1979. *Landscape sensitivity and change*. Trans. Inst. Brit. Geogr. 4, 463-484.

Bryant, R.G. & Gilvear, D.J., 1999. *Quantifying geomorphic and riparian land cover changes either side of a large flood event using airborne remote sensing: River Tay, Scotland*. Geomorphology 29, 307–321.

Bull, W. B., 1979. *Threshold of critical power in streams*. Geol.Soc. Am. Bull. 90, 453-464.

Caruso, B., Ross, A., Shuker, C. & Davies T, 2013. *Flood hydraulics and impacts on invasive vegetation in a braided river floodplain*, New Zealand. Environ. Nat. Resour. Res. 3, 92-110.

Cotton, G.K., 1999. *Hydraulic design of flood control channels*, in: Mays, L.W. (Ed.), Hydraulic desing handbook, McGraw-Hill, Chpt. 16, pp.16.1-16.42.

Costa, J. E., 1983. *Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range*. Geol. Soc. Am. Bull. 94, 986–1004.

Costa, J.E. & O'Connor, J.E., 1995. *Geomorphically effective floods*, in: Costa, J.E., Miller, A.J., Potter, K.W., Wilcock, P. (Eds.), Natural and Anthropogenic Influences in Fluvial Geomorphology, Monograph. 89, Am. Geophys. Union, pp. 45–56.

Chirilă, G. & Preda, A., 2006. *Aspects regarding flash floods and floods in the Trotus river hydrographic basin. Conference on water observation and information system for decision support*, Balwois, Ohrid, Republic of Macedonia, may 23-26.

Chirilă, G. & Preda, A., 2007. *Aspecte privind viiturile şi inundaţiile produse în anul 2005 în bazinul hidrografic Trotuş- Aspects regarding 2005 flash floods and floods in the Trotus river basin* . Comunicări de Geografie XI, Bucureşti, 255-260.

Chirilă-Neculau, G. & Zaharia, l. 2010. *Maximum flow variability and flood potential in the Trotus* *catchment*. Stud. Univ. Babes-Bol., Geogr. LV, 87-98.

Croke, J., Todd, P., Thompson, C., Watson F., Denham R. & Khanal G., 2013. *The use of multi temporal LiDAR to assess basin-scale erosion and deposition following the catastrophic January 2011 Lockyer flood,* SE Queensland, Australia. Geomorphology 184, 111–126.

Curtis, J.A. & Whitney, J.W., 2003. *Geomorphic and hydrologic assessment of erosion hazards at the Norman municipal landfill, Canadian River floodplain central Oklahoma*. Environ. Eng. Geosci. 9, 241–252.

Dean, D.J. & Schmidt, J.C., 2013. *The geomorphic effectiveness of a large flood on the Rio Grande in the Big Bend region: Insights on geomorphic controls and post-flood geomorphic response*. Geomorphology 201, 183-198.

Downs, P.W., Dusterhoff, S.R. & Sears, W.A., 2013. *Reach-scale channel sensitivity to multiple human activities and natural events: Lower Santa Clara River, California, USA*. Geomorphology 189, 121–134.

Du Boys, M.P., 1879. *Etudes du régime du Rhône et de l'action exercée par les eaux sur un lit à fond de graviers indéfiniment affouillable.* Annales des Ponts et Chaussées 5, 141– 195.

Dumitriu, D., 2007. *Sediment system of the Trotuş drainage basin (in Romanian)*. Editura Universităţii, Suceava, 197 p.

Dumitriu, D., Condorachi, D. & Niculiţă, M., 2011. *Downstream variation in particle size: a case study of the Trotuş River, Eastern Carpathians (Romania)*. An. Univ.Oradea, Geogr., t. XXI (2), 222-232.

Dunne, T. & Leopold, L.B., 1978. *Water in Environmental Planning*. Freeman, San Francisco, 818 p.

Emmett, W.W. & Wolman, M.G., 2001. *Effective discharge and gravel-bed rivers*. Earth Surf. Proc. Land. 26, 1369-1380.

Erskine, W.D., 1996. *Response and recovery of a sand-bed stream to a catastrophic flood*. Z. Geomorphol. 40, 359-383.

Erskine, W.D. & Saynor, M.J., 1996. *Effects of catastrophic floods on sediment yields in southeastern Australia*, in: Erosion and Sediment Yield: Global and Regional Perspectives, Proceedings of the Exeter Symposium, July 1996, IAHS Publ. 236, 381-388.

Fischenich, C., 2001. *Stability thresholds for stream restoration materials. EMRRP Technical Notes Collection (ERDC TNEMRRP- SR-29)*. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Fuller, I.C., 2007. *Geomorphic work during a "150-year" storm: Contrasting behaviors of river channels in a New Zealand catchment*. Ann. Assoc. Am. Geogr. 97, 665-676.

Fuller, I.C., 2008. *Geomorphic impacts of a 100-year flood: Kiwitea Stream, Manawatu catchment, New Zealand*. Geomorphology 98, 84–95.

Fuller, I.C. & Heerdegen R.G., 2005. *The february 2004 flood in the Manawatu, New Zealand; hydrological significance and impact on channel morphology*. J. Hydrol. (NZ) 44, 75-90.

Gaeuman, D.A., Schmidt, J.C. & Wilcock, P.R., 2003. *Evaluation of in-channel gravel storage with morphology-based gravel budgets developed from planimetric data*. J. Geophys. Res.108, F1, 6001, doi:10.1029/2002JF000002.

Gaeuman, D.A., Schmidt, J.C. & Wilcock, P.R., 2005. *Complex channel responses to changes in stream flow and sediment supply on the lower Duchesne River, Utah*. Geomorphology 64, 185–206.

Graf, W.L., 1983. *Flood-related channel change in an arid region river*. Earth Surf. Proc. Land. 8, 125 – 139.

Hack, J.T. & Goodlett, J.C., 1960. *Geomorphology and forest ecology of a mountain region in the central Apalachians*. U.S. Geol. Surv., Prof. Pap. 347, 66p.

Harvey, A.M., 2007. *Differential recovery from the effects of a 100-year storm: significance of long-term hillslope channel coupling; Howgill Fells, northwest England*. Geomorphology 84, 192–208.

Hickin, E.J. & Sichingabula, H.M., 1988. *The geomorphic impact of the catastrophic October 1984 flood on the planform of Squamish River, southwestern British Columbia*. Can. J. Earth Sci. 25, 1078–1087.

Julian, J.P. & Torres, R., 2006. *Hydraulic erosion of cohesive riverbanks*. Geomorphology 76, 193–206.

Julian, J.P., Thomas, R.E., Moursi, S., Hoagland, B.W. & Tarhule, A., 2012. *Historical variability and feedbacks among land cover, stream power, and channel geometry along the lower Canadian River floodplain in Oklahoma*. Earth Surf. Proc. Land. 37, 449-458.

Juracek, K.E., 2002, *Historical channel change along Soldier Creek, northeast, Kansas: U.S.* Geological Survey Water-Resources Investigations Report 02-4047, 23 p.

Juracek, K.E. & Fitzpatrick, F.A., 2009. *Geomorphic applications of stream-gage information*. River Res. Appl. 25, 329-347.

Kale, V.S., 2007. *Geomorphic effectiveness of extraordinary floods on three large rivers of the Indian Peninsula*. Nat. Hazards 85, 306–316.

Kale, V.S & Hire, P.S., 2007. *Temporal variations in the specific stream power and total energy expenditure of a monsoonal river: The Tapi River, India*. Geomorphology 92, 134–146.

Knighton, A.D., 1980. *Longitudinal changes in the size and sorting of stream-bed material in four English rivers*. Bull. Geol. Soc. Am. 91, 483–502.

Knighton, A.D., 1998. *Fluvial forms and processes: A new perspective*. Oxford Univ. Press Inc., New York, 383 p.

Krapesch, G., Hauer, C. & Habersack, H., 2011. *Scale orientated analysis of river width changes due to* *extreme flood hazards*. Nat. Hazards Earth Syst. Sci. 11, 2137–2147.

Kundzewicz, Z.W., Pińskwar, I. & Brakenridge, G.R., 2012. *Large floods in Europe, 1985–2009*. Hydrolog. Sci. J. DOI:10.1080/02626667.2012.745082.

Lenzi, M.A., Mao, L. & Comiti, F., 2006. *Effective discharge for sediment transport in a mountain river: computational approaches and geomorphic effectiveness*. J. Hydrol. 326, 257–276.

Ma, Y., Huang, H.Q., Xu, J.X., Brierley G.J. & Yao, Z., 2010. *Variability of effective discharge for suspended sediment transport in a large semi-arid river basin*. J. Hydrol.388, 357–369.

Magilligan, F. J., 1992. *Thresholds and the spatial variability of flood power during extreme floods*. Geomorphology 5, 373–390.

Magilligan, F. J., Gomez, B., Mertes, L. A. K., Smith, L. C., Smith, N.D., Finnegan, D. & Garvin, J. B., 2002. *Geomorphic effectiveness, sandur development, and the pattern of landscape response during jökulhlaups: Skeiöarársandur, southeastern Iceland*. Geomorphology 44, 95–113.

Magilligan, F. J., Buraas, E.M. & Renshaw, C. E., 2015. *The efficacy of streampower and flow duration on geomorphic responses to catastrophic flooding*. Geomorphology 228, 175–188.

McMahon, T. A., Finlayson, B. L., Haines, A. T. & Srikanthan, R., 1992. *Global Runoff*. Catena Paperback, 166 p.

Meyer, G.A., 2001. *Recent large-magnitude floods and their impact on valley-floor environments of northeastern Yellowstone*. Geomorphology 40, 271–290.

Miller, A.J., 1987. *What does it take to make a geomorphically effective flood? Some lessons from the November 1985 flood in West Virginia*, in: Kite, J.S. (Ed.), Research on the Late Cenezoic of the Potomac Highlands, West Virginia, Open File Rep. OF8801, West Virginia Geol. and Econ. Survey, Morgantown, pp. 3–30.

Miller, A. C., 1990. *Flood hydrology and geomorphic effectiveness in the central Appalachians*. Earth Surf. Proc. Land. 15, 119–134.

Minea, G., Toroimac, G.I., Zaharia, L., Zarea, R. & Borcan, M., 2011. *Spatial variations of stream power in the Buzău and Ialomița river catchments (Romania)*. Lucrările Conferinţei "Apa și Aerul – Componente ale Mediului", Cluj-Napoca, 18-19 march, 2011.

Molnar, P., Anderson, R., Kier, G. & Rose, J., 2006. *Relationships among probability distributions of stream discharges in floods, climate, bed load transport, and river incision*. J. Geophys. Res. 111, F02001, doi: 10.1029/2005JF000310.

Molnar, P., Densmore, A.L., McArdell, BW., Turowski, J.M. & Burlando, P., 2010. *Analysis of changes in the step-pool morphology and channel profile of a steep mountain stream following a large flood*. Geomorphology 124, 85–94.

Nanson, G.C. & Erskine, W.D., 1988. *Episodic changes of channels and floodplains on coastal rivers in New South Wales*, in: Warner, R.F., (Ed.), Fluvial Geomorphology of Australia, Academic Press, Sydney, pp. 201-221.

Obreja, F.G., 2012. *The sediment transport of the Siret River during the floods from 2010*. Forum geografic, Studii şi cercetări de geografie şi protecţia mediului 11, 90-99.

O'Connor, J.E., 1993. *Hydrology, hydraulics, and geomorphology of the Bonneville flood*. Geol. Soc. Am. Special Paper 274, 1-84

Ortega, J. A. & Heydt, G. G., 2009. *Geomorphological and sedimentological analysis of flash-flood deposits. The case of the 1997 Rivillas flood (Spain)*. Geomorphology 112, 1–14.

Parker, G., Wilcock P.R., Paola C., Dietrich W.E. & Pitlick J., 2007. *Physical basis for quasi-universal relations describing bankfull hydraulic geometry of single-thread gravel bed rivers*. J. Geophys. Res. 112 (F4), doi:10.1029/2006JF000549.

Petit, F., 1990. *Evaluation of grain shear stress required to initiate movement of particle in natural rivers*. Earth Surf. Proc. Land. 15, 135– 148.

Petit, F., Perpinien, G. & Deroanne, C., 2000. *Détermination des puissances spécifiques critiques dans des rivières à charge de fond caillouteuse*. Revue Géographique de l'Est 40, 2-11.

Petit, F., Gob, F., Houbrechts, G. & Assani, A.A., 2005a. *Critical specific stream power in gravel-bed rivers*. Geomorphology 69, 92-101.

Petit, F., Hallot, E., Houbrechts, G. & Mols, J., 2005b. *Évaluation des puissances spécifiques de rivières de moyenne et de haute Belgique*. B. Soc. Géogr. de Liège 46, 37-50.

Pińskwar, I., Kundzewicz, Z. W., Peduzzi, P., Brakenridge, G.R., Stahl, K. & Hannaford, J., 2012. *Changing floods in Europe*, in: Kundzewicz, Z. W. (Ed.), Changes in Flood Risk in Europe, Special Publication 10, IAHS Press, Wallingford, Oxfordshire, pp. 83-95.

Pleşoianu, D. & Olariu, P., 2010. *Monitoring data proving hydroclimatic trends in Siret hydrographic area*. Present Environment and Sustainable Development 4, 327-338.

Podani, M. & Zăvoianu, I., 1992. *Causes and effects of the July 1991 floods in Moldova (in Romanian)*. Studii şi cercetări de geografie, XXXIX, 61-68.

Radecki-Pawlik, A., 2002. *Bankfull discharge in mountain streams: theory and practice*. Earth Surf. Proc. Land. 27, 115-123.

Rădoane, M., Pandi, G. & Rădoane, N., 2010. *Contemporary bed elevation changes from the Eastern Carpathians*. Carpathian Journal of Earth and Environmental Sciences, 5, 49–60.

Rădoane, M., Obreja, F., Cristea, I. & Mihăilă, I., 2013. *Changes in the channel-bed level of the eastern Carpathian rivers: Climatic vs. Human control over the last 50 years*. Geomorphology 193, 91-111.

Rinaldi, M., Wyzga, B. & Surian, N., 2005. *Sediment* *mining in alluvial channels: physical effects and management perspectives*. River Res. Appl. 21, 805–828.

Romanescu, Gh., 2013. *Floods in the Siret and Prut Basins*, in: Lóczy, D. (Ed.), Geomorphological impacts of extreme weather. Case studies from Central and Eastern Europe, Springer, pp. 99-120.

Romanescu, Gh. & Nistor, I., 2011. *The effects of the July 2005 catastrophic inundations in the Siret River's Lower Watershed, Romania*. Nat. Hazards 57, 345 – 368.

Roşca, S., Bilaşco, Ș., Petrea D., Fodorean, I. & Vescan, I., (2015). *Bankfull discharge and stream power influence on the Niraj River morphology*. Carpathian Journal of Earth and Environmental Sciences, 10 (1), 133 – 146.

Salit, F., Zaharia, L. & Beltrando, G., 2013. *Assessment of the warning system against floods on a rural area: the case of the lower Siret River (Romania).* Nat. Hazards Earth Syst. Sci. 13, 409–416.

Shields, A., 1936. *Anwendung der Aehnlichkeitsmechanik und der Turbulenzforschung auf die Geschiebebewegung. Mitteilungen der Preussischen Versuchsanstalt für Wasserbau und Schiffbau*, Heft 26, Berlin.

Schmitt, L., Maire, G. & Humbert, J., 2001. *La puissance fluviale: définition, intérêt et limites pour une typologie hydro-géomorphologique des rivières*. Z. Geomorphol. 45, 201- 224.

Tomkins, K.M., Humphreys, M.T., Wilkinson, M.T., Fink, D., Hesse, P.P., Doerr, S.H., Shakesby, R.A., Wallbrink, R.A. & Blake, W.H., 2007. *Contemporary versus long-term denudation along a passive plate margin: the role of extreme events*. Earth Surf. Proc. Land. 32, 1013–1031.

Thompson, C. & Croke, J., 2013. *Geomorphic effects, flood power, and channel competence of a catastrophic flood incofined and unconfined reaches of the upper Lockyer valley, southeast Queensland, Australia*. Geomorphology 197, 156-169.

Thornbury, W.D., 1954. *Principles of Geomorphology*. John Wiley and Sons, New York, 618 p.

Toone, J., Rice, S.P. & Piégay H., 2014. *Spatial discontinuity and temporal evolution of channel morphology along a mixed bedrock-alluvial river, upper Drôme River, southeast France: Contingent responses to external and internal controls*. Geomorphology 205, 5–16.

Toroimac, G.I., Zaharia, L., Minea, G. & Borcan, M., 2012. *Channel typology based on stability criteria. Case study: watersheds (Romania)*. Procedia Env. Sci. 14, 177 – 187.

VanLooy, J.A. & Martin, C.W., 2005. *Channel and vegetation change on the Cimarron River, southwestern Kansas, 1953–2001*. Ann. Assoc. Am. Geogr. 95, 727–739.

Wallick, J.R., Grant, G., Lancaster S.T., Bolte, J.P. & Denlinger, R.P., 2007. *Patterns and controls on historical channel change in the Willamette River, Oregon, USA*, in: Gupta, A. (Ed.), Large Rivers: Geomorphology and management, John Wiley & Sons, pp. 491-516.

Williams, G.P., 1978. *Bankfull discharge of rivers*. Water Resour. Res. 23, 1471–1480.

Williams, G. P., 1983. *Paleohydrological methods and some examples from Swedish fluvial environments, I. Cobble and boulder deposits*. Geogr. Ann. A 65, 227–243.

Wolman, M. G. & Miller, J. P., 1960. *Magnitude and frequency of forces in geomorphic processes*. J. Geol. 68, 54– 74.

Wolman, M. G. & Gerson, R., 1978. *Relative scales of time and effectiveness of climate in watershed geomorphology*. Earth Surf. Proc. Land. 3, 189–208.

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