Identifying a dominant discharge for natural rivers in southern Italy

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A R T I C L E   I N F O

Article history:
Received 8 February 2011
Received in revised form 28 October 2011
Accepted 29 October 2011
Available online 4 November 2011

Keywords:
Dominant discharge
Bankfull discharge
Effective discharge
Fluvial processes
Italy

A B S T R A C T

Natural rivers are subjected to continuous adjustments in response to any change in the environment. These environmental changes may occur naturally, as in the case of climatic variation or changes in vegetative cover, or may be related to human activities including channelization, damming, bank protection, and bridge construction. Identifying the value of that discharge ('dominant' or 'effective' discharge, QD) considered responsible for the main changes operated by a river has been a subject of great challenge to scientists and engineers during the last decades. In fact, this threshold value is largely adopted for stream-management decisions, for predicting the stable slope upstream of grade-control structures, and for designing moderate-to-large-sized hydraulic structures. In this paper, a simple concept of dominant discharge, corresponding to that value of QD that accounts for at least 50% of the total suspended sediment load transported by the river, is introduced and discussed. The work is based on a long-term data set that includes measurements of monthly discharge and suspended sediment load in 27 stream gauge stations located in three different regions of southern Italy. The analysis showed that QD corresponds to values of the return time, T, ranging from 1 to 5 years, confirming previous findings by other authors in similar analyses. Also, as the values of QD were well correlated with the 1.5-year peak discharges of the same investigated rivers, an empirical approach is suggested in order to estimate QD in rivers with no sediment load measurements.

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1. Introduction

The concept of ‘dominant’ (or ‘effective’) discharge, QD, in geomorphic processes is not clearly defined yet, and it has long been debated during the last few decades (Knighton, 1998). Because river morphology depends on the magnitude and frequency of liquid and solid discharges (Wolman and Miller, 1960), the evaluation of QD is essential in predicting the dynamic quasiequilibrium of a river system (Chang, 2008). Such equilibrium depends strongly on the interaction between flow and sediment load; and identifying a threshold discharge discriminating the equilibrium condition, after which a river reach can be assumed dynamically stable, is a great challenge for scientists and engineers. In fact, this threshold value is largely adopted for stream management decisions regarding stream restoration and (or) rehabilitation (Brookes and Shields, 1996; Hey, 1998; Shields et al., 2003); it is required by some theoretical models aimed at predicting the stable slope upstream of grade-control structures such as bed slits and check-dams (Porto and Gessler, 1999; Ferro and Porto, 2011), and it is also used for designing moderate-to-large-sized hydraulic structures (Brodilov, 1996).

Inglis (1941) gave a general definition of QD as that value of discharge which controls the meander length and breadth of a river. This definition is in agreement with that proposed by Shields et al. (2003) who recognized three different approaches for determining QD: the effective discharge QEF, the bankfull discharge QBF, and the discharge that corresponds to a given return interval QTR. The effective discharge QEF, that will be described later in a graphical form, follows the definition of Wolman and Miller (1960) by which it can be considered as the increment of discharge that transports the larger sediment load over a period of years (Shields et al., 2003). The bankfull discharge QBF, following the definition of Wolman (1954), is described as the point of incipient flooding at which flow overtops the natural channel and spreads across the floodplain. The QBF is that particular value of discharge with a given return time, T, that is often assumed to be the channel-forming discharge; in general, values of T of approximately 1 to 2.5 years in the partial duration series are used to define QBF (Leopold et al., 1964; Andrews, 1980).

Dunne and Leopold (1978) described the bankfull discharge QBF as the most effective at forming and maintaining average channel dimensions. For this reason, the term ‘bankfull discharge’ has been used interchangeably with the terms ‘effective discharge’, ‘channel-forming discharge’, and ‘dominant discharge’ (Simon et al., 2004). This confusion is principally due to the fact that originally the definition of bankfull discharge was related to stable channels having regular hydraulic
geometry (Simon et al., 2004). Consequently, the attempts of identifying a value of dominant discharge in natural rivers with irregular depths and cross-sections fail if \(Q_D\) is generated via \(Q_{BF}\).

Inglis (1941), in his definition of dominant discharge, suggested that \(Q_{BF}\) is slightly greater than the bankfull discharge \(Q_{BF}\). Wolman (1959), looking at the fluvial geometry of a small stream located in Maryland, observed that the lateral cutting of the cohesive channel banks occurred mostly in winter time following discharges that showed a recurrence interval of 8–10 times over a year. When compared to the bankfull stage, although the position of such discharges on the daily-duration curve is uncertain, these values showed a lower magnitude. In contrast, Leopold et al. (1964) observed that the flow at the bankfull is the most effective to carry sediments over time (Leopold, 1994); and based on their experience, they suggested a recurrence interval of \(Q_{BF}\) varying between 1 and 2 years. Other authors – with the aim of determining the return time, \(T\), associated with \(Q_{BF}\) – confirmed the hypothesis suggested by Leopold et al. (1964). Among them, Harvey (1969), looking at the effect of soil permeability on the hydrologic response of different rivers, observed values of \(T\) ranging from 1.8 years for gravel bed rivers to 7 years for catchments with a dominant base flow. This behavior is also documented by Roberts (1989) who reported values of \(T\) > 2 years for catchments with more permeable soils and in the range of 4–8 months for catchments with soils of very low permeability. Petit and Pauquet (1997), exploring the hydrological data set of 40 measurement stations located in 30 rivers in Belgium with a drainage area ranging in size from 4 to 2700 km², observed values of \(T\) ranging from 1.11 years for catchments with impermeable soils to 5.3 years for channel cross sections with a small width/depth ratio. More recently, Keshavarzy and Nabavi (2006), assuming the hypothesis of coincidence between \(Q_{BF}\) and \(Q_{D}\), observed for the latter mean values of \(T\) of ca. 1.11 years. This result confirmed the findings of a previous study of Erskine and Keshavarzy (1996) that obtained for some rivers located in New South Wales values of \(T\) ranging from 1.89 to 2.4 years.

However, many exceptions are well-documented in the literature. For example, Williams (1978), outlining the uncertainty associated with a correct definition of \(Q_{BF}\), obtained values of \(T\) ranging from 1 to 32 years. Emmett and Wolman (2001), looking at the hydrological response of five rivers located in the Rocky Mountains (USA), found values of \(T\) associated with \(Q_{BF}\) ranging from 1.5 to 1.7 years, in agreement with other authors. But they observed that \(Q_{BF}\), considered as the discharge responsible for most of the sediment bedload, showed lower values of \(T\) double those associated with \(Q_{BF}\), remarking on the non-coincidence between the latter and \(Q_{BF}\).

Knighton (1998) reported on the long debate related to the concept of dominant discharge and suggested that, in some cases, the discharge responsible for the most remarkable changes of fluvial forms could be associated with events of greater magnitude and then corresponding to much higher values of \(T\). The same hypothesis was formulated by Baker (1977) analyzing field data collected in Texas. In contrast, Benson and Thomas (1966), interpreting the data collected in nine natural rivers in the USA, and assuming \(Q_{BF}\) as the discharge associated with the highest value of the transported sediment, documented values of \(Q_{BF}\) lower than \(Q_{BF}\) and ranging from the mean value of annual discharge to the limit suggested by Wolman (1959). Nolan et al. (1987), looking at the suspended sediment load of five natural rivers in California, found that the discharges responsible for 50% of the transported sediment showed values of \(T\) ranging from 0.27 to 1.25 years and, consequently, these values were much lower than those associated with \(Q_{BF}\).

Other authors proposed different definitions of \(Q_{BF}\) based on their observations. Carlson (1965), for example, looking at the relationships between geometry of meanders and corresponding discharges, suggested that \(Q_{BF}\) must be defined not just as a single value but rather a wide range of flows ranging from the mean value of the monthly maximum discharge to the annual mean value. Andrews (1980) defined \(Q_{BF}\) as that discharge responsible for most of the total sediment load. Lamberti et al. (1984) considered \(Q_{BF}\) as that discharge associated with the median value of the solid/liquid discharge ratio. Crescimanno et al. (1990) gave a definition of \(Q_{BF}\): as that discharge that could be associated with the equilibrium condition suggested by Shields (1936). In their study, they found mean values of \(T\) associated with \(Q_{BF}\) in the range of 1.5–5 years. This approach, that combines channel slope, particle size, and flow magnitude in the definition of \(Q_{BF}\) could be very helpful in cases where field indicators of \(Q_{BF}\) are unreliable. Other studies exploring the effect of topography on \(Q_{BF}\) (Kilpatrick and Barnes, 1964; Lisle, 1987; Grant et al., 1990; Emmett and Wolman, 2001) have found that the recurrence interval of bankfull discharge tends to increase with an increase in channel slope; that is, high-gradient streams having coarse bed material tend to overtop their banks less frequently than low-gradient streams do (Sherwood and Huizing, 2005).

A method largely adopted in literature is that proposed by Wolman and Miller (1960) by which \(Q_{BF}\) can be considered the discharge associated with the maximum contribution of the mean value of annual sediment transport (effective discharge). In synthesis, \(Q_{BF}\) can be defined as the peak value of the curve ‘\(c\)’ depicted in Fig. 1 that is obtained by multiplying the rate of sediment transport associated with a certain discharge by the frequency of the latter.

The limit of this approach is that in their original contribution Wolman and Miller (1960) assumed that the frequency distribution of the stress operated by the flow could be represented by a log-normal law (curve ‘\(b\)’ of Fig. 1), while the relationship between shear stress and solid discharges could be described by a power law (curve ‘\(a\)’ of Fig. 1). Goodwin (2004), following this approach, proposed analytical solutions for \(Q_{BF}\) looking at several theoretical distributions of flow records and sediment transport rating curves for two catchments in the USA. His results, combined also with visual inspections of the cross sections, provided values of \(Q_{BF}\) generally lower than \(Q_{BF}\).

However, from the above considerations, clearly the concept of dominant discharge is not well defined and sometimes is unrealistic. In fact, its calculation is dependent on local factors such as the hydrologic regime (Pickup and Warner, 1976), magnitude and characteristics of sediment transport (Nolan et al., 1987; Quadir and Guo, 2009), soil permeability (Petit and Pauquet, 1997), and heavily on the type of available data (Barry et al., 2008). For example, the approach used in Fig. 1 (Wolman and Miller, 1960) that is surely the most objective of the proposed criteria, requires measurements of great detail that are not easy to get in natural rivers. Also, the assumption of a power law to indicate the relationship between shear stress and solid discharges strongly simplify the problem because it neglects the scattering that is common in this kind of measurement. Perhaps a pure empirical approach, based on a long-term data set of measured values and not derived from virtual theoretical

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**Fig. 1.** Definition of \(Q_{BF}\) according to Wolman and Miller (1960).

\[\begin{align*}
a &= \text{Rate of movement} \\
b &= \text{Frequency of occurrences} \\
c &= \text{Product of frequency and rate} \\
a &= \text{Applied stress}
\end{align*}\]
laws, can give some useful suggestions in understanding the geometric evolution of fluvial forms.

In southern Italy, where natural rivers are regulated in the upper parts to control sediment transport downstream, identifying the break between the floodplains and main channel is a great challenge. In fact, considering the dominance of flash floods in these areas, channel dimensions, including water-surface elevations of specific discharges, are changing with time. Also, because of the geology of this area, many channel reaches do not have active floodplains and consequently no bankfull discharge can be identified. In some cases though, a valley flat can be defined but it may be flooded so rarely that a bankfull discharge would not have much meaning (Williams, 1978).

For the above reason, because identifying $Q_{BF}$ is really difficult, the assumption to relate the dominant discharge to a fixed recurrence interval is much more simple and realistic to consider especially in designing hydraulic structures. In fact, because the use of grade control structures in southern Italy is crucial in regulating the sediment transport downstream, the availability of a simple method to identify $Q_D$, even in channel reaches not operated by monitoring stations, for which regional models are necessary, could be very useful. Also, for rivers subjected to restoration design this method of estimating $Q_D$ can serve as a preliminary guide for predicting a dynamic equilibrium between channel reaches, vegetation, riparian resources and infrastructures.

In this study, field data of suspended sediment load and discharge collected in 27 monitoring stations located in southern Italy are used to explore the existing relationship between $Q_D$ – here assumed as that discharge responsible for at least 50% of the total suspended sediment load of each river – and the associated return time $T$. The analysis will give useful suggestions for the choice of a proper return time associated with the discharge to be used in designing grade-control structures in a Mediterranean environment.

2. The field data

Since 1920, the Italian Hydrographic Service (SIMI) published daily and monthly data of discharge and monthly values of suspended sediment load for many rivers in the country. In this study, long-term measurements of suspended sediment load operated by 27 rivers located in Sicily, Calabria, and Basilicata have been considered (Fig. 2). All the rivers investigated in this analysis, even if some differences in terms of geology are obvious considering the large geographic area covering these three regions, have hydrodynamic and geometric characteristics typical of the rivers belonging to the Italian Appennine mountain chain. In general, they show an intermittent rather than continuous sediment transport process and only during short periods of the year the flow occupies the entire available cross-section or the low part of the stream channel. In particular, the rivers located in Calabria and Sicily can be ascribed to the category better known as ‘fumare’ (Sabato and Tropeano, 2004) that show typical high gradients and short lengths and originated gravel bed material. The rivers located in Basilicata drain generally larger headwaters with larger proportion of flat areas and with alluvial bed material of smaller size if compared to the previous ones. However, the past activity carried out generally during the early fifties and consisting in the construction of grade-control structures (mainly bed sills and check-dams) in the upper part of these rivers, made the reaches where the monitoring stations are located morphologically stable so that the data base used for this analysis is not affected by significant change of the hydrographic system.
The choice of using suspended sediment data is derived by the hypothesis that, according to several authors, suspended sediment is strongly representative of the total sediment load (see Wolman and Miller, 1960; Benson and Thomas, 1966; Simon et al., 2004). Gordon et al. (2004), for example, documented that the majority of sediment transport is caused by the suspended load, which typically exceeds bedload by a factor of 5–50 (Allan and Castillo, 2007). Also, the higher erodibility of surface soils lying in some of the catchments examined here and documented by Cavazza (1961), suggests that even in these areas the contribution of suspended sediment can be considered

<table>
<thead>
<tr>
<th>River</th>
<th>Station</th>
<th>Code</th>
<th>Years</th>
<th>N</th>
<th>Area (km²)</th>
<th>Qₜ₅ (m³ s⁻¹)</th>
<th>T (years)</th>
<th>Qₚ₉₅:₅ (m³ s⁻¹)</th>
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<td>B1</td>
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<td>1953–1980</td>
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Fig. 3. Comparison between empirical and log-normal distributions of the maximum values of MMDD series for three stream gauge stations.
dominant. However, the absence of bed load data for these rivers makes this choice obvious although it should be recognized that the availability of total sediment load data sets could strengthen the results obtained here. Following this general assumption, the analysis to identify $Q_D$ in the rivers examined here was carried out in three different steps.

The first step is aimed at checking for the presence of an overlapping series of monthly mean daily discharge (MMDD; m$^3$ s$^{-1}$) and total monthly sediment load (TMSL; kg) for the same stream gauge station. This produced the information provided in Table 1 that lists the number of rivers and associated stream gauge stations selected for the analysis and grouped for geographic regions. The overall measurements cover the period from 1930 to 1990 with sample sizes ranging from 5 to 34 years.

For each stream gauge station, only the monthly overall suspended sediment load $T_{5,j}$ (kg), transported by the river was considered. This value was calculated by multiplying the monthly suspended sediment load concentration, $C_{5,j}$ (kg m$^{-3}$), of the $j$-th month times the monthly runoff, $V_{5,j}$ (m$^3$), of the same month:

$$T_{5,j} = C_{5,j} \cdot V_{5,j}.$$  \hspace{1cm} (1)

Because some of the data set were available at a daily scale, in these cases the monthly runoff $V_{5,j}$ (m$^3$) of Eq. (1) was calculated by multiplying the mean daily discharge $Q_{5,j}$ (m$^3$ s$^{-1}$) of the $j$-th month times the number of seconds of that month.

The sum of the 12 monthly values of $T_{5,j}$ ($10^3$ t) gives the global value of $T_{5,i}$ ($10^3$ t) related to the $i$-th year. Integrating $T_{5,j}$ ($10^3$ t) over the entire period of measurements ($N$ years), the global value of suspended sediment load $T_5$ ($10^3$ t) for that river is obtained:

$$T_5 = \sum_{i=1}^{N} \sum_{j=1}^{12} T_{5,j}.$$ \hspace{1cm} (2)

The second step of the analysis consisted in ranking each of the 27 MMDD series maintaining the association between discharge and corresponding sediment load. Following this logic, we could calculate the cumulative sediment load and its percentage associated with each value of discharge. The value of MMDD corresponding to at least 50% of the total sediment load was assumed as dominant discharge $Q_D$. Details of $Q_D$ for all the stream gauge stations considered in this study are also listed in Table 1.

The third step required the knowledge of the frequency of each discharge in order to associate the value of the return period $T$ with each value of MMDD. To do that, a plot of maximum values of MMDD as a function of recurrence interval (flood-frequency curve) based on the two-parameter log normal distribution (LN2) was prepared for each of the 27 gauged study sites. In Fig. 3, a comparison between empirical frequency distribution $F(Q)$ and the two-parameter log normal distribution $P(Q)$ is depicted for some of the maximum values of MMDD examined in the analysis.

Although Fig. 3 illustrates the results for 3 monitoring stations only (assumed here representative of each geographic area investigated), the comparison shows in general a good agreement between empirical and theoretical values of MMDD so that the LN2 law,
3. Results and discussions

The use of LN2 allowed us to associate predetermined values of the return period $T$ with the percentage of total suspended sediment load for each stream gauge station. This method can be considered very helpful in order to identify a meaningful discharge or range of discharges to compare sediment transport rates and then reduces the uncertainties related to the difficulties in looking for a form-based 'bankfull' criteria.

The relationship between the return period $T$ and the percentage of total suspended sediment load is reported in Fig. 4 for three stations of measurement (assumed here representative of each geographic region), and considering different values of the return time ($T = 1.001, 1.005, 1.01, 1.05, 1.10, 1.15, 1.2, 1.5, 2, 5, 10, 20, 50, and 100$ years). The values of $T$ from 1.001 to 1.5 have been used in order to check further the relationship between $T_3$ and the lower values of the recurrence interval. In fact, the analysis carried out by Williams (1978), aimed at exploring the relationship between bank-full discharge and return time based on 36 active floodplain stations located in the USA, documented for $T$ values from 1.01 to 200 years with the mode or peak recurrence interval of about 1.5 years on the annual maximum series. Also, previous investigations (Porto and Gessler, 1999) aimed at identifying stable conditions in some of the rivers located in the same geographic area reported here, outlined the dependence on the return time of the model used for predicting a stable slope stressing that the model efficiency improved for discharges corresponding to low values of $T$.

Fig. 4 shows that, even if there is a certain variability, the values of $T$ associated with 50% of the total suspended sediment load are very close to 1–2 years (see dashed lines). This is confirmed by the data reported in Table 1 that shows values of $T$ ranging from 1.5 to 4.9 years for Basilicata, from 1.02 to 3.1 years for Calabria, and from 1.4 to 3.5 years for Sicily. There are only two exceptions in Sicily: the stations Baiata at Sapone and San Leonardo at Monumentale show greater values of $T$ respectively equal to 6.6 and 10.2 years. A possible explanation for these higher values of $T$ can be related to the geology of the headwaters draining to these monitoring stations. In fact, regarding the station Baiata at Sapone, we can say that the catchment area is characterized by a dominant presence of calcareous rocks that because of their particular hydrological system may affect the frequency distribution of the recorded discharges. For example, because of the larger porosity of this type of rocks, a certain amount of runoff can disappear from the surface and consequently the frequency distribution curve results shifted to the higher values of $T$ affecting the final results. This explanation is even more convincing for the station San Leonardo at Monumentale. In fact, the analysis shows a value of $T$ equal to 1.4 years for the station Vicari (see Table 1) located several kilometers upstream of the station Monumentale while for the latter the value of $T$ is ca. 7–8 times greater. It is easy to notice that in its lower area where the station Monumentale is located, the catchment San Leonardo shows a large proportion of calcareous rocks (Fierotti, 1988) while the area draining to Vicari is covered by clay and this produces higher rates of suspended sediment load associated with lower values of discharge. Conversely, when the flow approaches the lower part of the catchment, some of the runoff follow the underground ways and affect the frequency distribution curve generated at the monitoring station.

However, the overall results suggest, according to several authors (Leopold et al., 1964; Andrews, 1980; Erskine and Keshavarzy, 1996; Petit and Pauquet, 1997), that $Q_D$ is associated with low values of the return period. Also, this assumption is consistent with what was found by Crescimanno et al. (1990) in six Sicilian rivers even if they used a different approach in determining $Q_D$.

Previous works (Porto and Gessler, 1999; Ferro and Porto, 2011) aimed at calculating the stable slope upstream of check-dams are based on the assumption of a 2-year peak discharge as a representative flow in designing grade-control structures. This hypothesis seems to be consistent with the data listed in Table 1 where 18 of 27 stream gauge stations show values of $T < 2$ years. Of course, considering that the data used in our study have been collected at a monthly temporal scale, the real relationship between peak flow, $Q_{PK}$, and sediment peak discharge is not available, and the effect of extreme events in determining $Q_D$ is still unknown. However, because peak flow data are available for the rivers examined here, an attempt was made to relate the values of $Q_D$ with the values of peak discharge corresponding to different values of $T$. The diagram depicted in Fig. 5A shows the correlation coefficient of this relationship against different values of $T$. From this diagram we can see that this relationship shows a maximum for $T = 1.5$ years and suggests that $Q_{PK}$ corresponding to low values of $T$ is better correlated with the fluvial form processes. Fig. 5B illustrates the relationship between $T$ and the variance of residuals. Again, this function shows a minimum for $T = 1.5$ years confirming once more this assumption.

Based on Fig. 5A, an attempt was made to check, for the available data set, the relationship between $Q_D$ and $Q_{PK}$. This relationship is shown for each of the three geographic areas in Fig. 6A–C. In all cases, we can see a clear positive relationship of the form:

$$Q_D = a Q_{PK}^{n}$$

indicating that $Q_D$ and $Q_{PK}$ are very closely related.

The relatively small number of data, especially for Basilicata (only four stations available), precludes precise statistical fitting of this equation, but the visual fits suggest that Eq. (3) provides an effective
representation of the relationship; and this is further confirmed by the $r^2$ values for the individual data sets, which assume the value of 0.45 for Basilicata (only four values), 0.98 for Calabria, and 0.80 for Sicily.

However, some further considerations may be done about the constant $a$ and the exponent $n$ of Eq. (3). In fact, if as an assumption, the data of Basilicata are not taken into account considering the small number of stations, the exponent $n$ lies in a very small range (from 0.82 to 0.99) suggesting that its value may be independent of geographic factors.

In contrast, the magnitude of the constant $a$ in Eq. (3) could be expected to vary according to the different geographic factors, and particularly in response to the theoretical model used for calculating $Q_{PK-1.5}$.

For this study site, however, the constant $a$ appears to vary relatively little between Calabria and Sicily, and a single relationship of the form represented by Eq. (3) (i.e., $Q_D = 0.188 Q_{PK-1.5}$, $r^2 = 0.85$) has been fitted by least squares to the combined data set from all data in Fig. 7.

The high correlation shown in Fig. 7 suggests that the use of a design discharge corresponding to low values of $T$ in previous investigations appears to be reasonable (see Porto and Gessler, 1999; Ferro and Porto, 2011). Also, assuming that long-term measurements of $Q_{PK}$ are available for a certain cross section of a river, Eq. (3) can be used (as a first attempt) to predict $Q_D$. However, in the absence of $Q_{PK}$ measurements, a simple method to predict the dominant discharge is still necessary. For this reason, an attempt was made to relate $Q_D$ with $Q_{PK}$ values estimated by a regional hydrological model [Two Component Extreme Value (TCEV) Distribution — see Rossi et al., 1984] and with the catchment area, the latter being very effective in representing bankfull stage conditions (see Petit and Pauquet, 1997). Both relationships, depicted in Fig. 8A,B, show evidence of a link between $Q_D$ and $Q_{PK-1.5}$ or the catchment area.

In particular, Fig. 8A demonstrates that if we use for $Q_{PK-1.5}$ the new set of values estimated by the regional TCEV model for all 27 series, the correlation between $Q_D$ and $Q_{PK-1.5}$ does not change substantially. Also, the same degree of correlation ($r = 0.91$) is documented by Fig. 8B where on behalf of $Q_{PK-1.5}$ we use catchment area as an independent variable. In the end, both equations depicted in Fig. 8A, B can be used, as a first attempt, to estimate $Q_D$ as defined in this investigation.

4. Conclusions

One of the possible approaches for determining the dominant discharge $Q_D$ of a natural river requires the knowledge of the discharge with a given return interval $T$. Because recurrence interval relations are intrinsically different for channels with flashy hydrology than for those with less variable flows (Shields et al., 2003), many studies have concluded that recurrence interval approaches produce poor estimates of $Q_D$ (Pickup and Warner, 1976; Doyle et al., 1999). Consequently, assuming a priori a value of $T$ to generate $Q_D$ should be avoided, although it may be useful at times to serve as a first estimate of the dominant discharge in stable channels (Shields et al., 2003; Doyle et al., 1999). However, although exceptions are well
documented in the literature, in general, low values of \( T \) (of approximately 1 to 2.5 years) in the partial duration series are used to define \( Q_0 \) (Leopold et al., 1964; Andrews, 1980; Erskine and Keshavarzy, 1996; Petit and Paquet, 1997).

In the study reported here, although the 27 rivers examined in this analysis belong to three different geographic regions (Basilicata, Calabria, and Sicily), the value of the dominant discharge \( Q_0 \) empirically deduced as that value of discharge responsible for at least 50% of the total suspended sediment load, was shown to be very much related to discharges corresponding to low values of return time \( T \). This result is consistent with what other authors found in similar contexts where it was assumed that fluvial forms are regulated by a discharge with a recurrence interval of 1–2 years. The same assumption is made in a Mediterranean environment where \( Q_0 \) is required for predicting the stable slope upstream of grade-control structures. The results obtained in this analysis showed that a \( Q_0 \) value of first attempt can be deduced using predicted values of peak discharge with \( T = 1.5 \) years \( (Q_{PK-1.5}) \) if a suitable regional model is available. Also, the information illustrated in Fig. 8 shows that a reasonable value of \( Q_0 \) can also be deduced using catchment area as an independent variable. These findings, considering the simplicity of the proposed approach, can be seen as a useful tool, at least in a Mediterranean environment, for stream-management decisions.

For example, the search of a design discharge for predicting the stable slope upstream of grade-control structures, such as bed sills and check-dams, largely used to stabilize natural rivers in these areas, would be very much easier if a simplified method for determining \( Q_0 \) is available.

However, these results must be seen as preliminary because they are related to a limited number of rivers and also they are dependent on suspended sediment load only. Further analysis involving additional catchments and total sediment load measurements or estimates can be useful to check the strength of the results obtained here.

![Fig. 8. Relationship between \( Q_0 \) and \( Q_{PK-1.5} \) (A) and between \( Q_0 \) and catchment area (B).](image)

**Acknowledgements**

The study reported in this paper was supported by a grant from MIUR RDb 2010. Both authors set up the research, analysed the results, and participated in writing the paper. The referees and the Guest Editor are thanked for their important and helpful comments.

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