Characterisation of selected extreme flash floods in Europe and implications for flood risk management

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1. Introduction

The occurrence of flash flooding is of concern in hydrologic and natural hazards science due to the top ranking of such events among natural disasters in terms of both the number of people affected globally and the proportion of individual fatalities. According to Barredo (2007), 40% of the flood-related casualties occurred in Europe in the period 1950–2006 are due to flash floods. The potential for flash flood casualties and damages is also increasing in many regions due to the social and economic development bringing pressure on land use. Furthermore, evidence of increasing heavy precipitation at continental (Groisman et al., 2004) and global scales (Groisman et al., 2005) supports the view that the global hydrological cycle is intensifying as the planet warms (Huntington, 2006). As a consequence, the flash flood hazard is expected to
increase in frequency and severity in many areas, through the im-
pacts of global change on climate, storm–weather systems and riv-
er discharge conditions.

Flash floods are associated with short, high-intensity rainfalls,
mainly of convective origin that occur locally. As such, flash flood
usually impact basins less than 1000 km², with response times of
a few hours or less. The time dimension of the flash flood response
is linked, on one side, to the size of the concerned catchments, and
on the other side, on the activation of surface runoff that becomes
the prevailing transfer process. Surface runoff may be due to differ-
ent generating processes, owing to the combination of intense
rainfall, soil moisture and soil hydraulic properties. It may be also
enhanced by land use modification, urbanization and fire-induced
alteration. Analysis of flash flood processes is important because
these events often reveal aspects of hydrological behaviour that
either were unexpected on the basis of weaker responses or high-
light anticipated but previously unobserved behaviour (Archer
et al., 2007; Delrieu et al., 2005; Borga et al., 2007). Characterising
the response of a catchment during flash flood events, thus, may
provide new and valuable insight into the rate-limiting processes
for extreme flood response and their dependency on catchment
properties and flood severity. Moreover, local flood-producing pro-
cesses are more amenable to analysis in the typical small–scale
flash flood basins than in larger catchments where the regional
combination of controls can be relatively more important (Merz
and Bloschl, 2008).

The examination of flash flood regimes across Europe shows
that space and time scales of flash floods change systematically
when moving from Continental to Mediterranean regions, while
seasonality shifts accordingly from summer to autumn months
(Gaume et al., 2009). This has several hydrological implications,
which need to be considered, for example, when examining poten-
tial effects of land use (urbanization, deforestation, afforestation)
and climate change on flash flood occurrence.

Flash flood forecasting, warning and emergency management
are, by their nature, suitable to cope with the characteristics of
flash flood risk (Drobot and Parker, 2007; Collier, 2007). Specific
difficulties with flash flood forecasting relate to the short lead
times and to the need to provide local and distributed forecast
(Norbiato et al., 2008). Attempts to characterise the flood response
to short and intense storm events is therefore central in this con-
text (Collier and Fox, 2003). However, investigating these aspects
is difficult due to lack of systematic observational data for flash
floods, encompassing data on the flood-generating rainfall at the
required space and time detail and discharge data. Flash flood
events are difficult to monitor because they develop at space and
time scales that conventional measurement networks of rain and
river discharges are not able to sample effectively (Gaume and Bor-
ga, 2008). Moreover, being flash floods relatively rare event at the
local scale, these are difficult to observe in experimental catchments.

A better characterisation of flash floods in Europe over various
time and spatial scales is sought in this work as an important as-
pect of climate and hydrologic science in general, and to improve
flood risk management in particular. The aim of this research is
threefold: (i) to summarise the data from an archive of selected ex-
treme flash flood events occurred in Europe in the period from
1994 to 2007, together with background climatic and hydrological
information, (ii) to characterise these events in terms of basins
morphology, flood-generating rainfall, peak discharges, runoff
coefficient and response time, and (iii) to use the insight gained
with this analysis to identify implications for flash flood risk
management.

The archive includes data from 25 major flash flood events oc-
curred since 1994, with 20 events occurred since 2000. Data have
been collected in several regions of Europe, even if without achiev-
ing a systematic coverage at continental scale. The hydrometeoro-
logical data include high-resolution rainfall patterns and flood
hydrographs or peak discharges. Climatic information and data
concerning morphology, land use and geology are also included
in the database. Hydrologic and hydraulic models are used to check
the consistency of the data and to enable the reconstruction of spe-
cific events for which only partial information is available.

The presentation of the paper will adopt the following outline.
Section 2 summarises prior studies on flood and flash flood charac-
terisation across Europe. The methodology adopted to develop
the archive and basic statistics are reported in Section 3. Section 4 pro-
vides a characterisation of the events in terms of climate and ba-
sins morphology. Section 5 is devoted to the analysis of the
flood-generating rainfall, peak discharges, runoff coefficient and
response time. Finally, the implications for flash flood risk manage-
ment are examined in Section 6, together with the conclusions
from the study.

2. Prior studies on flash flood characterisation across Europe

Observational difficulties of flash floods, barriers in hydrometeoro-
logical data transfer (Viglione et al., in press) and lack of a com-
prehensive archive of flood events across Europe hinder the
development of a coherent framework for analysis of flood clima-
tology, hazard and vulnerability at the pan-European scale (Barred-
o, 2007). Among the few studies with a continental view, Barredo
(2007) reports a catalogue of the major flood events from 1950
to 2006 in the European Union. In his study, Barredo characterised
major floods in terms of casualties and direct damages. Twenty-
three out of the 47 events listed in the catalogue are classified as
flash floods, and are mainly localised in Italy, Spain and southern
France. Flash flood events are also reported in Germany, Belgium
and UK. In spite of the smaller areas impacted by these events,
flash floods caused 2764 fatalities (i.e., 52 casualties per year in
average), making 40% of the overall casualties reported in the
study, largely exceeding river floods (18%), and being second only
to storm–surge floods (42%). It is worth noting that fatalities due
to storm–surge floods concentrate into three large events which
occurred from 1953 to 1962 on coastal regions of northern Europe,
whereas flash floods occurred over the whole considered period in
a number of European regions.

Gaume et al. (2009) analysed date of occurrence and flood peak
distribution of flash floods from an inventory of events that oc-
curred in selected regions of Europe over a 60 years period (from
1946 to 2007). The archive collated data from both instrumented
and ungauged basins. In contrast to Barredo (2007), the archive
used by Gaume et al. (2009) includes a substantial number of
events from central and eastern European countries. Gaume et al.
(2009) noted a peculiar seasonality effect on flash flood occurrence,
with events in the Mediterranean region (Italy, France and Catalo-
nia) mostly occurring in the autumn months, whereas events in the
inland Continental region (Romania, Austria and Slovakia) tend to
occur in the summer months, revealing different climatic forcing.
Consistently with this seasonality effect, spatial extent and dura-
tion of the events was smaller for the Continental events with re-
spect to those occurring in the Mediterranean region. Finally,
Gaume et al. (2009) outlined that the flash flood regime is gener-
ally more intense in the Mediterranean Region than in the Conti-
nental areas. The present work builds upon the investigation by
Gaume et al. (2009), by examining more closely the control of wa-
tershed physiography and channel network geometry on flood
response, and extending the analysis to the runoff coefficient and
the response time. Because of the requirement of high-resolution data,
in particular spatially-distributed rainfall, we used only a portion of
the events considered by Gaume et al. (2009), and several cases,
especially from the Mediterranean region, were not included in this study.

Parajka et al. (in press) analysed the differences in the long-term regimes of extreme precipitation and floods across the Alpine–Carpathian range (from France to Romania) using seasonality indices and atmospheric circulation patterns to understand the main flood-producing processes. The analysis was supported by cluster analyses to identify areas of similar flood processes, both in terms of precipitation forcing and catchment processes. The results allowed to isolate regions of similar flood generation processes including southerly versus westerly circulation patterns, effects of soil moisture seasonality due to evaporation and effects of soil moisture seasonality due to snow melt.

3. Data collection methodology

The aim of the data collection methodology was threefold: (i) to identify extreme flash flood events representative of different hydroclimatic European regions, (ii) to collect high-resolution data enabling the characterisation of the flood response for each event (in terms of response time and runoff coefficient), initial soil moisture status and climate, and (iii) to collect data for the characterisation of the morphological properties of the catchments, land use, soil properties and geology. These requirements led to focus on events with a high-resolution data coverage, and in particular with the availability of weather radar observations permitting rainfall estimation with fine spatial and temporal resolution and of discharge data from streamgauge stations and/or post-flood analysis (Borga et al., 2008; Marchi et al., 2009). As a consequence, attention was focused on events occurred since mid 1990s in European regions covered by weather radar systems. The preliminary identification of flash floods benefited from the collection of primary data on flash floods described by Gaume et al. (2009). The three steps of the methodology used for the set up of the archive are described in the following sections.

3.1. Events selection and severity

A definition of flash flood event was required as a working principle to develop the archive and select the events. Initially, the definition of flash flood event was based on the duration of the causative rainfall, the size of the catchment impacted by the flood and the severity of the event. Consistently with the rules adopted by Gaume et al. (2009), duration of the storm event was limited to 34 h and maximum size of the catchment area was set to 1000 km². As a follow up, these rules were slightly relaxed to include one event with a larger catchment size (Gard flood in France, 2002, with a maximum basin area of 1856 km²).

Severity of storm precipitation and flood response was a further requisite for event selection. The criterion adopted and met by all selected events is that the return period of the flood-generating rainfall exceeds 50 years at least for some rainfall duration. The selection of the 50-year recurrence interval as a threshold for event selection ensures consistency of the study archive with earlier archives of extreme rainfall events (Hershfield, 1961; Schumacher and Johnson, 2006). Rainfall was preferred to peak discharge for assessing flash flood severity because of the difficulties in providing estimates of flood peak return period in many ungauged basins. In many cases, the events were extraordinary in terms of severity. Either rainfall or peak discharge return period exceeded 50 years (and sometimes 1000 years) for the events of Fella (Italy, 2003) (Norbìato et al., 2007), Valencia (UK, 2004) (Roca and Davison, 2010), Kamp (Austria, 2002) (Götzknecht et al., 2002), Mala Svinika (Slovakia, 1998) (Gaume et al., in press), Aude (France, 1999) (Gaume et al., in press) and Gard (France, 2002) (Gaume et al., in press). The point rainfall amount recorded by one raingauge for the Weisseritz river event (Germany, 2002) represents the new all time record for the whole territory of Germany for a daily duration (James et al., 2004).

For twelve out of 25 studied flash floods, discharge data were available at more than one cross-section, and for some of these several internal watersheds could be analysed. We retained all these data but for a few events for which a disproportionately high number of internal data were available. To avoid over-representing these floods in the analysis, we removed those cases where catchment properties and catchment responses were notably similar each other. As a general rule, the number of retained nested catchments is proportional to the size of the largest catchment area impacted by the flood \( A_{\text{max}} \). One basin was retained for events with \( A_{\text{max}} \) up to 100 km², and more basins were added for the events in which this threshold (or multiples of 100) was exceeded. The maximum number of watersheds retained for each flood \( n_{\text{max}} \) thus corresponds to the ratio \( A_{\text{max}}/100 \) rounded to the upper integer:

\[
 n_{\text{max}} = \left\lceil \frac{A_{\text{max}}}{100} \right\rceil
\]

A few exceptions were applied to include in the sample watersheds with good quality hydrometeorological data and representative of particular morphoclimatic conditions. The criterion described above was not applied to contiguous, non-nested, catchments.

The final catalogue reports data from 25 flash flood events, including 60 catchments in ten European countries (Table 1). Some of the events were investigated in earlier studies (Table 1); results from these analyses and modelling efforts were taken into account during this study. Fig. 1 shows the location of the flash floods on a European map of Köppen–Geiger climate classification (Peel et al., 2007). Fig. 1 aims at describing the general climatic context of the studied flash floods. Some watersheds, especially those covering a wide range in elevation, encompass different climatic classes. The date of flood occurrence is also reported in the figure. Although the spatial coverage on Europe is not complete, the events encompass a wide range of climatic conditions. Fig. 1 shows that the selected floods are mainly located in a geographical belt crossing Europe from Catalonia (Spain) to the Black Sea in Romania, covering southern France, northern Italy, Slovenia, Austria and Slovakia, with a few more events from UK, Germany and Crete (Greece). This spatial pattern results from the requirements of using radar rainfall estimates for event analysis and to ensure that flood data originate from accurate and standardised post-event analyses. Even though this selection allows to pool extreme events from different hydroclimatic regions, some regions (notably central and eastern Mediterranean) are likely to be underrepresented in the archive. This points out to the need to extend in future works the analysis to incorporate more data from these areas.

The occurrence months generally agree with the seasonality pattern of flash flood-generating rainfall over the various European regions and with the observations from Gaume et al. (2009) and from Parajka et al. (in press). Over much of continental Europe, thunderstorms are typically a summer phenomenon (Trewartha, 1981). Frequency of flash flood-generating storms increases with the rise in surface temperatures in early summer and reaches a maximum in June in central-eastern Europe; in July over much of central and northern Europe, including Italian Alps; in September and October in the western Mediterranean and in winter over southern Italy and eastern part of the Mediterranean basin. The contribution of snowmelt is null for most of the studied flash floods, which occurred in summer or autumn in watersheds of low or moderate elevation (Table 1). For an event, occurred on 2006 on the Isarco river system (northern Italy), ice-melt was observed. However, its contribution to the runoff was negligible (Norbìato et al., 2009).
### 3.2. Rainfall data

For all the events, but three cases (Giofyros event of 1994, Almyrida event of 2006, and the Posina event of 1999), radar rainfall observations were made available for rainfall estimation together with raingauge data. The quantitative precipitation estimation (QPE) problem is particularly crucial and difficult in the context of flash floods since the causative rain events may develop at very short space and time scales [Krajewski and Smith, 2002; Bouilloud et al., in press]. The rainfall QPE problem in the context of the re-analysis of flash flood events presents a number of specificities. On the one hand, and as shown later in this paper, flash floods often occur in mountainous or hilly regions resulting in increased enhanced radar visibility problems associated with the intervening relief. On the other hand, flash floods generally result from convective rainfall which makes the visibility problem less stringent due to the extended vertical dimension of the precipitating clouds; in addition, the ice–water changes of phase, resulting for instance in bright bands, may have relatively less impact on the radar QPE compared to more stratiform precipitation conditions. Moreover, the small spatial scale of the flash floods-generating storms, combined with the density of most raingauge networks, is such that just a few raingauge data are available for checking the radar observations at fine time resolution. However, more raingauge data are available for re-analysis (particularly at the event duration scale) than for real-time applications.

A methodology was specifically devised for rainfall estimation with use of radar and raingauge data (Bouilloud et al., in press). Depending on the relative locations of the impacted regions and the radar systems available, as well as their operating protocols and maintenance, the quality of the radar datasets may vary a lot. Depending on the relative locations of the impacted regions and the radar systems available, as well as their operating protocols and maintenance, the quality of the radar datasets may vary a lot.

### Table 1
Summary information on the flash floods.

<table>
<thead>
<tr>
<th>Region/catchment impacted</th>
<th>Date of flood peak</th>
<th>Country</th>
<th>Climatic region</th>
<th>No. of studied watersheds</th>
<th>Range in watershed area (km²)</th>
<th>Range in average elevation (m)</th>
<th>Storm duration (h)</th>
<th>Previous studies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Magarola River</td>
<td>June 10, 2000</td>
<td>Spain</td>
<td>Mediterranean</td>
<td>1</td>
<td>94.3</td>
<td>388</td>
<td>8.4</td>
<td>Llasat et al. (2003)</td>
</tr>
<tr>
<td>Costa Brava</td>
<td>October 13, 2005</td>
<td>Spain</td>
<td>Mediterranean</td>
<td>2</td>
<td>57.3–73.8</td>
<td>170–223</td>
<td>9</td>
<td></td>
</tr>
<tr>
<td>Almyrida River</td>
<td>October 16, 2006</td>
<td>Greece</td>
<td>Mediterranean</td>
<td>1</td>
<td>24.7</td>
<td>212</td>
<td>15</td>
<td></td>
</tr>
<tr>
<td>Posina River</td>
<td>September 20, 1999</td>
<td>Italy</td>
<td>Alpine–Mediterranean</td>
<td>1</td>
<td>116</td>
<td>1045</td>
<td>24</td>
<td></td>
</tr>
<tr>
<td>Sesia River</td>
<td>June 5, 2002</td>
<td>Italy</td>
<td>Alpine–Mediterranean</td>
<td>5</td>
<td>75–983</td>
<td>494–1512</td>
<td>22</td>
<td>Sangati et al. (2009)</td>
</tr>
<tr>
<td>Selška Sora River</td>
<td>September 18, 2007</td>
<td>Slovenia</td>
<td>Alpine–Mediterranean</td>
<td>3</td>
<td>31.9–212</td>
<td>847–992</td>
<td>16.5</td>
<td>Rusjan et al. (2009)</td>
</tr>
<tr>
<td>Trisanna River</td>
<td>August 23, 2005</td>
<td>Austria</td>
<td>Mediterranean</td>
<td>1</td>
<td>122</td>
<td>2409</td>
<td>25</td>
<td></td>
</tr>
<tr>
<td>Isarco and Passirio Rivers</td>
<td>October 3–4, 2006</td>
<td>Italy</td>
<td>Alpine</td>
<td>6</td>
<td>12–342</td>
<td>1809–2863</td>
<td>12.5</td>
<td>Nobile et al. (2009)</td>
</tr>
<tr>
<td>Kamp River</td>
<td>August 7, 2002</td>
<td>Austria</td>
<td>Continental</td>
<td>4</td>
<td>70.1–622</td>
<td>480–873</td>
<td>31</td>
<td>Gutknecht et al. (2002)</td>
</tr>
<tr>
<td>Turniansky Creek</td>
<td>June 19, 2004</td>
<td>Slovakia</td>
<td>Continental</td>
<td>1</td>
<td>70.4</td>
<td>331</td>
<td>1.25</td>
<td></td>
</tr>
<tr>
<td>Casimcea River</td>
<td>August 29, 2004</td>
<td>Romania</td>
<td>Continental</td>
<td>1</td>
<td>500</td>
<td>145</td>
<td>19</td>
<td></td>
</tr>
<tr>
<td>Feernic River</td>
<td>August 23, 2005</td>
<td>Romania</td>
<td>Continental</td>
<td>1</td>
<td>168</td>
<td>683</td>
<td>5.5</td>
<td>Roca and Davison (2010)</td>
</tr>
<tr>
<td>Ilisua River</td>
<td>June 20, 2006</td>
<td>Romania</td>
<td>Continental</td>
<td>1</td>
<td>139</td>
<td>672</td>
<td>9</td>
<td></td>
</tr>
<tr>
<td>Clit River</td>
<td>June 30, 2006</td>
<td>Romania</td>
<td>Continental</td>
<td>1</td>
<td>36</td>
<td>570</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>Grinties River</td>
<td>August 4, 2007</td>
<td>Romania</td>
<td>Continental</td>
<td>1</td>
<td>52</td>
<td>1039</td>
<td>4</td>
<td></td>
</tr>
<tr>
<td>Starzel River</td>
<td>June 2, 2008</td>
<td>Switzerland</td>
<td>Continental</td>
<td>3</td>
<td>15.1–124</td>
<td>643–797</td>
<td>8</td>
<td></td>
</tr>
<tr>
<td>Valency River, Boscastle area</td>
<td>August 16, 2008</td>
<td>United Kingdom</td>
<td>Oceanic</td>
<td>1</td>
<td>20</td>
<td>192</td>
<td>6.75</td>
<td></td>
</tr>
</tbody>
</table>
case (Pellarin et al., 2002), (3) implementation of corrections for range-dependent errors (e.g. screening, attenuation, vertical profiles of reflectivity), and (4) optimisation of the rainfall estimation procedure by means of radar-raingauge comparisons at the event duration scale (Bouilloud et al., in press). The methodology was applied consistently over most of the events, but for the cases where only radar products were made available (Valency River flood, occurred in UK in 2004, and the floods observed in Slovakia).

3.3. Discharge data

Both discharge data from streamgauge stations and from post-flood analysis were available in the study. Discharge data from streamgauge stations were available for 29 cases, whereas data from post-flood analysis were used in the remaining 31 cases. Post-event analysis methods include a range of procedures for indirect estimation of peak discharges, generally encompassing the following steps: identification of the flow process (which was categorised into the following classes: liquid flow, hyperconcentrated flow, debris flow), high-water marks identification, post-flood river geometry survey, and application of appropriate hydraulic methods for peak flood computation (Costa and Jarrett, 2008). With regard to the classification of the flow process, only liquid flows were considered in this study. Together with peak discharge values, post-flood analysis methods were used also to derive time of the raising flow, flood peak time, and rate of recession. Timing estimates were obtained based on eyewitnesses interviews and accounts. A standardised method for post-event analysis was used throughout the study (Gaume and Borga, 2008; Borga et al., 2008; Marchi et al., 2009). Estimates of flood peak for the earlier events were reviewed considering the original field notes, photographs, reports, and documentation, and conducting field visits to the flood locations. Discharge data from streamgauges were obtained based on extrapolation of rating curves from smaller observed flows. The rating curves were checked to evaluate the degree of extrapolation required and to assess the quality of the final estimates. Although great care was devoted to the various steps of discharge estimations, we should note that all the peak flood data should be regarded as affected by considerable uncertainty. An accuracy of 15–20 min has been reported for the timing estimates obtained by means of eyewitness interviews (Gaume, 2006).

The large percentage of discharge data obtained from post-event analysis underlines the importance of indirect discharge estimates in setting up catalogues of flash floods. This is particularly the case for events which impact small catchment areas. Table 2 categorises catchment areas according to the method used to derive the peak flood data (streamgauge versus post-event analysis). Discharge data from gauging stations generally concern catchments which are significantly larger (p-value smaller than 0.001 according to the Mann–Whitney U test) than those for which estimates are obtained from post-event analysis. This is not an unexpected finding; larger scale flash floods events have higher probability to be recorded by streamflow measuring stations, whereas events with smaller spatial extent generally impact ungauged basins. An implication of this finding is that systematic survey of flash floods is particularly important in the European regions where these events are climatologically characterised by smaller spatial extent, such as in the Continental areas. Without

Fig. 1. Location and climatic context of studied flash floods; the numbers indicate the months of flash-flood occurrence (map of Köppen–Geiger climate classification from Peel et al., 2007).
systematic post-event analysis, it may be unlikely to develop reliable flash flood catalogues in these areas.

3.4. Climate, annual water balance, land use and geology data

For each catchment, values concerning the mean annual estimates of precipitation, runoff and potential evapotranspiration were collected from climate atlas or computed from available long-term rainfall, runoff and temperature data. For the ungauged catchments, mean annual runoff values were derived based on data from available downstream stations and from regional runoff relationships.

Data were also collected for the characterisation of the soil moisture status at the event start; these were rainfall and runoff (when available) data over the 30 days period before the event. Corresponding long-term data were also collected over the same period.

Geographical data include digital elevation models of the watersheds and thematic information (land use, lithology and soil maps). Digital elevation models were available at grid resolution from 20 m to 100 m. Only a short summary of the information about geology, land use and soil property is reported here, since ongoing research aims to investigate the quality and the homogeneity of this information and how these catchment properties, together with rainfall characteristics, control the event runoff coefficient.

The studied watersheds show a wide range of land uses. However, since most of the catchments are located in mountainous or hilly regions, urban and suburban land cover represents usually a limited fraction of the catchment surface. Conversely, urban areas are often located close to the outlet of the catchments. Agricultural areas prevail in hilly watersheds, whereas forests are widespread in mountainous areas. In Alpine and Alpine–Mediterranean watersheds, large areas without vegetation cover (bare rock and scree) are found at the highest elevations. Very different geolithological conditions characterise the basins, with karst areas reported a number of catchments. Lakes and artificial reservoirs are present in some of the largest catchments considered here. However, their drainage areas are minor, in relation to the overall catchment, and the corresponding attenuation effects on the flood hydrograph was considered to be small.

3.5. Distributed rainfall–runoff model

A distributed hydrological model was implemented over all the catchments here analysed to simulate the flood hydrographs. The purpose of the model application was twofold: (i) to ensure consistency among the disparate type of event data and (ii) to enable estimation of the event runoff coefficient (defined as the ratio of event runoff to total storm rainfall depth) for the cases where only peak flow and timing data are available. The consistency check aimed to identify possible inconsistencies in the available data, both in terms of rainfall and runoff volumes, and in terms of timing of the runoff response with respect to the space–time structure of rain fields. When inconsistencies were identified, data were scrutinised again, and, when it was the case, field visits, surveys and interviews were organised and carried out to ensure either error correction or data removal. In the latter case, other event characteristics, such as stages or other evidences of flood, were retained.

The main requirements for the choice of the model were a structure consistent with the available observations (i.e., able to use efficiently spatially-distributed rainfall data) and a limited number of calibration parameters.

In the model, the SCS-Curve Number (SCS-CN) procedure (Ponce and Hawkins, 1996) is applied on a grid-by-grid way for the spatially distributed representation of runoff generating processes. The distributed runoff propagation procedure is based on the identification of drainage paths, and requires the characterisation of hillslope paths and channelled paths. A channelisation support area \((A_h) (\text{km}^2)\), which is considered constant at the subbasin scale, is used to distinguish hillslope elements from channel elements. Discharge at any location along the river network is computed by:

\[
Q(t) = \int_A q(t, x) \, dx
\]

where \(A (\text{km}^2)\) indicates the area draining to the specified outlet location, \(q(t, x)\) is the runoff at time \(t\) and location \(x\), and \(\tau(x)\) is the routing time from the location \(x\) to the outlet of the basin specified by the region \(A\). The routing time \(\tau(x)\) is defined as:

\[
\tau(x) = \frac{L_s(x)}{v_h} + \frac{L_c(x)}{v_c}
\]

where \(L_s(x)\) is the distance from the generic point \(x\) to the channel network following the steepest descent path, \(L_c(x)\) is the length of the subsequent drainage path through streams down to the watershed outlet, and \(v_h\) and \(v_c\) (m s\(^{-1}\)) are two invariant hillslope and channel velocities, respectively.

The model includes also a linear conceptual reservoir for base flow modelling. A more complete description of the model is reported in Borga et al. (2007). The model was applied by using the available topographical information (with grid size resolution ranging from 20 m to 100 m) and at time step ranging from 15 to 30 min, depending on the event.

The model parameters were estimated over the catchments available for each event by means of a combination of manual and automatic calibration to minimize either the Nash–Sutcliffe efficiency index over the flood hydrographs (for the gauged catchments) or the mean square error over the flood peak and the timing data (rise, peak and recession) (for catchments where runoff data were provided from post-event surveys). Example of model application over the studied events are reported by Borga et al. (2007) and by Sangati et al. (2009). The combination of the SCS-CN method and of the conceptual single store for the representation of the subsurface flow dynamics provided an efficient representation of the two main flow pathways – surface and subsurface – contributing to runoff generation at the event time scale.

4. Flash floods characterisation: climate and physiographic factors

4.1. Climate and annual water balance

Climate variability strongly impacts the mechanisms of flood generation in two ways: in a direct way through the variability of storm characteristics, and indirectly through the seasonality of
rainfall and evapotranspiration that affect the antecedent catchment conditions for individual storm events (Sivapalan et al., 2005). Furthermore, climate may influence the runoff generation processes by controlling the geomorphological structure of catchments, through soil formation and erosion processes, as exemplified, for instance, by the positive relationship between drainage density and mean annual precipitation (Gregory and Gardiner, 1975).

Two main objectives drive the climatic characterisation in this section: (i) to evaluate the range of climates of the considered case studies and (ii) to identify the distribution of the events under the different climatic conditions. We applied the Budyko's climatic classification scheme (Budyko, 1974) to display the climatic characteristics of the catchments. This is achieved by presenting the specific response of each catchment on the Budyko plot (Fig. 2), which expresses \( E/P \), the ratio of average annual evapotranspiration \( E \) to average annual precipitation \( P \) as a function of \( E_p/P \), the ratio of average annual potential evapotranspiration \( E_p \) to average annual precipitation. Actual evapotranspiration \( E \) for each catchment was derived as the long-term difference between \( P \) and \( R \) (runoff) for the basins. The ratio \( E_p/P \) is a measure of the climate, and is called the dryness index (or index of dryness). Large \( E_p/P \) (>1) represents dry climate (water-limited conditions), while small \( E_p/P \) (<1) represents a wet climate (energy-limited conditions). Thus the Budyko diagram encapsulates a major climatic control on annual water balance.

In Fig. 2, as well as in the next sections of this paper, a subdivision of climate into five classes is adopted: Mediterranean, Alpine–Mediterranean, Alpine, Continental and Oceanic. With reference to the Köppen–Geiger classification (Fig. 1), Mediterranean corresponds to \( Csa \) and \( Csh \), Continental corresponds to \( Dfa \), \( Dfb \) and \( Dfc \), and Oceanic to \( Cfb \). As the consequence of the wide range in elevation of the watersheds, the two classes related to mountainous regions (Alpine–Mediterranean and Alpine), encompass various climates of the Köppen–Geiger classification, from \( Cfb \) to \( Efh \) or \( Ehf \) at the highest elevations. Each point in Fig. 2 represents a flash flood event. The figure shows that the sample of events considered here represents a wide interval of climatic controls on the annual water balance, ranging from very wet to semi-arid conditions. Twenty out of the 25 floods occurred under energy-limited conditions. Conversely, all the flash floods in the Mediterranean region occur in water limited conditions, with the notable exception of the Gard area (flash floods of Avène 1997 and Gard 2002), where large values of evapotranspiration fluxes are balanced by abundant mean annual precipitation. It is worth noting that local dry conditions characterise also one catchment under Continental climate (Casimcea, Romania), located close to the Black Sea with low mean annual precipitation. On the contrary, Alpine and Alpine–Mediterranean flash flood watersheds, as well as the Boscastle flash flood (Oceanic or Marine West Coast Climate) display typically humid climatic conditions, with values of actual evaporation \( E \) close to potential evaporation \( E_p \).

The concentration of flash flood catchments under energy-limited conditions may suggest that more humid climatic conditions affect the initial soil moisture status for individual storm events, and thus have an indirect effect on flash flood occurrence. On the other hand, this may result from a sampling effect due to the under-representation of flash flood events in the Mediterranean region. Given the available observations, we are not in a position to clarify this issue, and future efforts should be aimed to increase the observations of flash flood events in these undersampled areas.

The occurrence of most of European flash floods under energy-limited conditions is in strike contrast with observations reported for the US, where all major flash floods were reported to occur in arid or semi-arid part of western and southwestern United States (Costa, 1987). This needs to be considered when comparing flash floods events and hazards in Europe and US and when discussing flash flood forecasting models and procedures.

4.2. Physiographic factors and characteristics of the channel network

Physiographic factors may affect flash flood occurrence in specific catchments by combination of two main mechanisms: orographic effects augmenting precipitation, and topographic relief promoting rapid concentration of streamflow. Both effects have been documented in the literature (Costa, 1987; O’Connor and Costa, 2004). Storm quasi-stationarity is a characterising feature of several flash flood-generating rainfalls, with very intense precipitation insisting on the same locations for enough time to produce heavy accumulations. One of the elements that favour the anchoring of convective system is the orography, which play an important role in regulating of atmospheric moisture inflow to the storm and in controlling storm motion and evolution (Smith et al., 1996; Davolio et al., 2006). Relief is necessary for promoting flow concentration along drainage ways, which results in high unit discharges and relevant geomorphic effects of flash floods in sloping watersheds. Heavy convective precipitation may occur also in plain areas, but the ensuing flood generally lack the kinematic component, which characterises the propagation and the hazard potential of flash floods. Collier and Fox (2003) and Collier (2007) incorporated the two relief effects in their procedure for assessing the susceptibility of catchments to flooding due to extreme rainfall. They identified two catchment morphological characteristics that may affect baseline susceptibility to flooding: catchment slope, and ratio of catchment area to mean drainage path length. In a similar vein, in this section we analyse morphometric characteristics of the studied catchments both in the vertical and in the horizontal plane. This is carried out by investigating two specific morphological relationships: basin steepness to basin size, and channel length to basin size.

The steepness of a river basin is considered in terms of the relief ratio, the ratio of the total basin relief to the total basin length. The total relief is the elevational difference between the highest and the lowest points in the basin and the total length is the length of the main channel from watershed divide to outlet. The relief ratio is a dimensionless number that has found wide use in the comparison of basins (Schumm, 1956, 1963). The relation between basin area and relief ratio is presented in Fig. 3. As expected, a decrease in basin steepness is observed with increasing basin area, following general relationships reported in the literature (Dade, 2001). Values of the relief ratio range between 0.008 and 0.15, with an average value of 0.055. The lowest values are observed in low
Power-law relations are used to relate watershed area as the average distance from basin outlet to each point in the basin. In the slope coefficient (the exponent), the line of organic correlation is reported. Maximum stream length is measured from basin outlet to the crest of the drainage divide along the stream channel (Hack, 1957); mean stream length is computed as the average distance from basin outlet to each point in the basin. Power-law relations are used to relate watershed area $A$ to $L$ and $L_m$, respectively, as follows:

\[ L = 1.514 \cdot A^{0.557} \]  
(No. of cases: 60; $R^2 = 0.86$)

\[ L_m = 0.867 \cdot A^{0.551} \]  
(No. of cases: 60; $R^2 = 0.88$)

Since the aim in the study here is to describe the functional relationship between the two variables with the primary interest in the slope coefficient (the exponent), the line of organic correlation (also called geometric mean functional regression) (Helsel and Hirsch, 2002) was preferred to ordinary least squares regression for deriving the relationships.

Eq. (4) is a version of the classical Hack’s law, following Hack (1957) who reported $L \propto A^{0.6}$ for streams in the Shenandoah Valley and adjacent mountains of Virginia. The exponent in Eq. (4) is in the range 0.55–0.59 reported by Rigon et al. (1996) for the Hack’s law in basins ranging from 50 to 2000 km$^2$. Explanations for the exponent being larger than 0.5 (implying positive allometry) emphasised the role of basin elongation as well as the fractal characteristic of river networks (Rigon et al., 1996). The statistically significant equivalence of the exponents in Eqs. (4) and (5) indicates constant scale relations between the length of the flow path along the maximum stream from the watershed divide and the average length of the flow paths in the drainage basin. This finding supports the results by Maritan et al. (2002), who defined the relation between basin area and mean stream length as a “strong version of Hack’s law and a proof of the embedded similarity in the network structure” and found that the maximum stream length is proportional to the mean stream length.

Overall, the relationships between upstream watershed area and relief ratio, maximum stream length and mean flow length point out that flash flood basins are generally characterised by a non negligible topographic relief, whereas the other physiographic features are in the range reported for basins of similar size.

5. Flash floods characterisation: rainfall and flood response

5.1. Flood-generating rainfall: amount and duration

This section examines the characteristics of the rainfall that produced the flash floods by using two variables: rainfall amount and rainfall duration. Fig. 5 plots the relations between these variables taking into account the largest watershed for each flash flood. The largest watersheds have been chosen because rainstorm duration over watersheds of different size has small variability for the studied events and the largest watersheds are more affected by the longest durations. Moreover, choosing the largest watersheds for each flood reduces the scatter of the plot and increases its readability. The combination of rainfall duration and amount is influenced by the local climate and strongly affects the runoff generation during the event. Rainstorm duration is defined here as the time duration of the flood-generating rainfall episodes which are separated by less than 6 h of rainfall hiatus.

Fig. 5a shows the relation between total event rainfall versus duration for the different climatic regions considered in this study. For these rainfall events (storms) the range of values is from 33 mm in 1 h (Turniansky Creek, June 2004) (the smallest humid region storm event) to 700 mm in 24 h (Aude River basin, November 1999). Examination of Fig. 5a shows that the events can be grouped into three different classes. The first group includes storm events which lasted up to 7 h. The corresponding rainfall amount is less than 100 mm, and almost all these events occurred under a Continental climate. The second group includes storms with duration from 7 to 22 h. The maximum rainfall amounts reported for these events is 300 mm. Almost all these events are from the Mediterranean and the Alpine–Mediterranean region. Finally, the third group is made of events of longer duration, up to 34 h, and with storm cumulated amount up to 700 mm. As such, these events are in the transition between flash flood and flood events. Most of these events are from the Mediterranean region. However, two cases are also reported from the Continental region. These correspond to the Weisseritz flash flood event close to Dresden (2002) and to the Kamp flash flood in Austria (2002), both with very high precipitation amounts. These episodes were embedded within the...
large flood event which impacted the Central European region in the summer 2002 (Kundzewicz et al., 2005).

The plot of rainstorm duration versus the ratio of total event rainfall to mean annual precipitation (Fig. 5b) affords comparison of rainstorms under different climatic conditions, taking into account the differences in the mean annual precipitation. Whereas the relative ranking of the flash floods in the various climatic regions does not show significant differences between Fig. 5a and b, inspection of Fig. 5b shows the different impacts of the flash flood-generating rainfalls on the local annual water balance for the various climates. In the Mediterranean climate, the ratio of event to annual precipitation is generally greater than 0.2, indicating a relatively large impact of these precipitations on the annual water balance. Conversely, for the Continental, Alpine and Alpine–Mediterranean regions, the ratio is less than 0.2, showing a relatively less significant influence on the annual water balance.

The spatial and temporal scales of the events are represented in Fig. 6, which plots the area of the largest watershed for each flood versus the duration of the causative rainstorm. The use of the largest watershed area to surrogate the spatial scale of the flash flood event is made on a convenience basis. A more direct way would imply the use of the radar rainfall information. However, this would require use of data from a radar network, instead of use of the radar information from the closest radar site, as it was done in this study. Anyway, the selection of the study watersheds reflects the severity of the flash floods and as such represent a suitable indicator of the spatial scale of flash floods.

The positive correlation between these two variables is especially apparent in the lower limit, which defines a minimum rainstorm duration required to trigger a flash flood in a catchment of given size. The figure also shows that flash floods are essentially associated to Mesoscale Convective Systems, accordingly with the classification of Orlanski (1975), who introduced a subdivision into space and time scales for the atmospheric processes.

Table 3 presents the range of drainage areas of the largest impacted watersheds under different climates (events under Alpine and Oceanic climates are not included owing to the small number of cases). The largest flash flood basins are preferentially located in the Mediterranean and the Alpine–Mediterranean climate, where flash floods are generated also by longer storm events generally occurring during the fall season. Smaller spatial scales characterise the events occurring in the Continental region, with generally shorter rainstorms (although with a few relevant exceptions: cf. Fig. 5), which occur preferentially during the summer season. This is consistent with the general outline of European flash floods reported by Gaume et al. (2009).

5.2. Unit peak discharges and relation with catchment area

The dependence of peak discharge on watershed area is widely accepted, and there is a substantial body of literature describing the relationship between discharge and basin area, both for single-event peak flows and for mean annual peak flows (Smith, 1992; Gupta et al., 1996; Furey and Gupta, 2005, among the others). The relationship between the catchment area and the unit peak discharge (i.e., the ratio between the peak discharge and the upstream catchment area) was investigated for our database plotting the data in a log–log diagram and analyzing the envelope curve. When data from all the regions are grouped together, the unit peak discharges exhibit a marked dependence on area (Fig. 7). The envelope curve reported in Fig. 7 is as follows:

Table 3

<table>
<thead>
<tr>
<th>Region</th>
<th>No. of flash floods</th>
<th>Range in watershed area (km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mediterranean</td>
<td>7</td>
<td>55.0–1856</td>
</tr>
<tr>
<td>Alpine–Mediterranean</td>
<td>4</td>
<td>116–983</td>
</tr>
<tr>
<td>Inland Continental</td>
<td>11</td>
<td>12.7–622</td>
</tr>
</tbody>
</table>

where $Q_u$ is the peak unit discharge ($\text{m}^3/\text{s} \cdot \text{km}^2$) and $A$ is the upstream area ($\text{km}^2$). The relationship is the same as that reported by Gaume et al. (2009). Since the highest unit peak discharges in the two samples are the same and correspond to events from the Mediterranean region, it is not surprising that the envelope curves are overlapping. For small basin areas, the flash floods observed under Continental climate, namely in Slovakia, also attain high values of unit discharge, even though these unit peaks seem to decrease with upstream area in a faster way than for the Mediterranean events. This behaviour points out once more the different space and time scales of the generating storm events.

The exponent in the power-law relationship in Eq. (6) is smaller than that reported in the literature for extreme floods in the world (Herschy and Fairbridge, 1998) and for Europe (Herschy, 2002). Herschy and Fairbridge (1998) considered 41 maximum floods from the world to develop the following envelope equations:

$$Q_u = 970 \cdot A^{-0.44} \quad (6)$$

For the maximum floods in Europe, Herschy (2002) proposed the following equation:

$$Q_u = 850 \cdot A^{-0.643} \quad A \geq 100 \text{ km}^2 \quad (7)$$

Interestingly, the archive used by Herschy included relatively few flash flood events, reflecting more accurately the behaviour of riverine flow.

Relatively low unit peak discharges, less than $1 \text{ m}^3/\text{s} \cdot \text{km}^2$, are reported for some Alpine and Continental catchments in the area range between 50 and 500 $\text{ km}^2$. It is worth recalling that areas of moderate unit peak flows do not necessarily equate to areas of moderate flood risk, since the flood risk for any particular location is in part due to local forecasting, warning, and communication systems, existing flood protection works, as well as the community’s social, political, and regulatory setting. Some of the flash floods exhibiting moderate unit peak values resulted in high loss of human lives and large damages. These effects are exemplified by the flash flood events occurred in Romania during the 2005 summer season, which led to 76 casualties and to overall economic losses as high as 1.6% of the gross national product (Constantin-Horia et al., 2009).

Fig. 7b shows that the highest unit peak discharge were collected by means of post-flood surveys. This can be referred to the small areas of watersheds concerned, as well as to the choice of the streams investigated in post-flood surveys, which commonly focus on the areas most severely hit by the flood, and points out the unique role of post-flood survey in flash flood analysis. For watersheds areas in the range of 50–200 $\text{ km}^2$ the sample includes data from both gauging and post-flood reconstruction: the overlapping of unit peak discharge values indicates that the use of different methods for assessing peak discharge did not result in systematic differences.

5.3. Runoff coefficient

The event runoff coefficient is a key concept in hydrology and an important diagnostic variable for catchment response. Examination of runoff coefficients is useful for catchment comparison to understand how different landscapes filter rainfall into event-based runoff. Specific questions in the case of flash floods concern: (i) quantification of runoff coefficients and analysis of possible differentiations with climatic regions and (ii) investigating the relationship between runoff coefficients and antecedent moisture conditions.

Event runoff coefficients are usually estimated as the ratio of event runoff volume to event rainfall volume. This requires the separation of the event hydrograph into the two components of baseflow and event flow, and then the determination of starting time and end time of event flow. In this study, starting time was identified as the time of the first rise of discharge, and the corresponding runoff value was used to determine the baseflow. The time corresponding to the end of event runoff was estimated by separating the recession curve into the components of surface and saturated flow by plotting the recession curve in a semilogarithmic paper (Tallaksen, 1995). These two components are thought to represent different flow paths in the catchment, each characterised by different residence time, the outflow rate of groundwater flow being lower than the recession rate of the surface flow component. The separation between baseflow and event flow was carried out by continuing the baseflow until the major flood peak and then connecting with a straight line to the recession curve as defined above. For consistency, the same procedure of hydrograph analysis was applied both to recorded hydrographs and to hydrographs simulated by means of the rainfall–runoff model. The latter hydrographs were used for the ungauged basins where only the magnitude and time of the peak were reported together with time of first rise and recession. The ratio of baseflow to peakflow is low in all the floods analysed here: this implies that the uncertainties in the separation between baseflow and event flow regard essentially the definition of the end time of event flow and hence to the analysis of the recession curve. We checked that the distribution of runoff coefficients derived by means of hydrological modelling was close to that derived from hydrograph analysis. To this purpose, the analysis based on hydrological modelling.
was applied to the cases for which complete hydrograph data were available. The runoff coefficients derived in this way compared favourably with those obtained from observations, with a mean relative error of $-0.05$ (the runoff coefficients from hydrological modelling show a slightly underestimation with respect to those obtained from direct observations) and a Nash–Sutcliffe efficiency value of 0.89.

Fig. 8 shows the frequency distribution of the runoff coefficient; the mean value is 0.35, with standard deviation 0.18, median 0.37, and interquartile range 0.20–0.45.

These results compare well with the values obtained by Merz and Blöschl (2003) and Merz et al. (2006) for flash flood cases occurred in Austria. Merz and Blöschl (2003) were able to compare runoff coefficients for different flood types (they considered the following flood types: short rain floods, long rain floods, rain-on-snow floods and snowmelt floods). These authors reported that, based on their data from Austria, runoff coefficients are smallest for flash floods, and they increase, in that order, for short rain floods, long rain floods, rain-on-snow floods and snowmelt floods. Relatively low values of event runoff coefficient were also reported by Goodrich (1990) in his analysis of flash flood events on the semi-arid Walnut Gulch catchment. In five of the largest flash flood events observed in a 20 years period, the runoff coefficients range between 0.07 and 0.21. On the other hand, large values of runoff coefficient (close to 0.9) have been reported by a number of authors for extremely intense storm events characterised by wet initial conditions (Smith et al., 1996, 2005).

Table 4 compares the values of the runoff coefficients in the various climatic regions (but Oceanic climate, for which only one event is reported in the archive). Runoff coefficients are relatively higher for Mediterranean flash floods, while the smallest values are reported for the Alpine watersheds (even though the small sample size suggests that the significance of this comparison should be considered with caution). The higher runoff coefficients of Mediterranean flash floods are likely due to the type of flood-generating storm events, which are longer and characterised by larger rainfall amount than for the other cases.

The relation between event runoff coefficient and the event cumulated rainfall is reported in Fig. 9. Overall, there is a slight dependence of runoff coefficient on the water input and the scatter is very large, with values of the runoff coefficients distributed in a wide interval over the range of precipitation depth. There are several possible explanations for the wide scatter reported in Fig. 9. Considering the meteorological factors, for flash flood events it is not only the rainfall depth but also rainfall intensity that likely controls the magnitude of the runoff coefficient. A further potential source of scatter is the variability of the initial soil moisture conditions which combines with the subsurface water storage capacity to determine the amount of water input that needs to be stored in the soil before event runoff may start, i.e., the so-called initial losses. Wood et al. (1990) note that for large floods the role of antecedent soil moisture for flood response should decrease with increasing return interval. However, a number of studies (Sturdevant-Rees et al., 2001; Gaume et al., 2004; Borga et al., 2007) showed considerable impact of initial soil moisture conditions on runoff from extreme flash floods.

We developed an antecedent precipitation index to assess the impact of initial soil moisture conditions on runoff coefficients. Since we are considering different events occurring under various climates, the index needs to account also for the climatic variability. The index was computed as the ratio of the precipitation in the 30 days before the flash flood event to the long-term 30 days average for the same period. The number of years considered for the long-term mean varies from site to site, ranging generally from 20 to 30. We chosen a 30 days antecedent period because in several ungauged basins only monthly long term-average rainfall data are available. Moreover, use of 30 days (instead of shorter duration) helps to reduce the sampling spatial error associated to the use of point rainfall in order to represent basin average precipitation. Three classes of antecedent saturation were considered: Dry (index of antecedent saturation $<0.5$), Normal (from 0.5 to 1.5), and Wet ($>1.5$). The index capability to evaluate the initial soil moisture conditions compared well with the predictions from a continuous soil moisture accounting hydrological model (Norbio...
et al., 2008). Snowmelt is not considered in the index: however, snowmelt was a significant contribution to basin saturation only for one event (Sesia River basin, northern Italy, June 2002), which was also characterised by abundant precipitation in the 30 days before the event. Data for assessing antecedent saturation were not available for two cases; the analysis was thus carried out on a sample of 58 basins. Table 5 compares runoff coefficients in the three classes of antecedent precipitation index. Values of the runoff coefficient increase with moving from Dry to Normal and Wet antecedent conditions, even though the variability of the runoff coefficient within the same class is relatively high. The variability is highest within the “Dry” class, as expected, and decreases with increasing the initial soil moisture. Application of the Mann–Whitney test shows that differences are statistically significant (at $p = 0.07$) when comparing the “Dry” and “Wet” classes. This shows that antecedent moisture conditions can play a significant role in determining land-surface response to extreme rainfall events.

It is interesting to note that only 11 cases were characterised by “Wet” conditions, whereas 17 cases occurred at the end of periods characterised by precipitation significantly lower that the long-term monthly average.

The influence of the initial soil moisture status on triggering of flash floods was investigated by counting the number of cases in the Dry, Normal and Wet conditions stratified in the various climatic regions (Table 6). This comparison shows that usually the events in the Mediterranean climatic region occur with Dry and Normal antecedent conditions, whereas the events in the Continental region occur with Normal and Wet conditions. The Alpine–Mediterranean region is characterised by occurrences over all the different antecedent conditions. Although the small sample size affords only preliminary considerations, examination of these results suggests that abundant and relatively long-lasting precipitation in Mediterranean regions may trigger flash floods overcoming low initial wetness, whereas in Continental areas moderate or high initial saturation is required for flash flood occurrence. This could be referred to prevailing short duration and low rainfall amounts of summer storms that cause flash floods under Continental climate.

### 5.4. Response time

The specific problem of flash flood risk management is that these floods interact with social organisation at space and time scales that imply unusually short warning lead times (Creutin et al., 2009). This section is devoted to the quantification of the response time for the considered events and to relate the response time to morphological parameters such as the catchment area.

As a measure of the response time, we used the concept of lag-to-peak (Dingman, 2002) or lag time in the following. In this study, we defined the lag time as the duration between the time of the centroid of the generating rainfall sequence and the time of the discharge peak. We used the centroid of rainfall, instead of the mean used threshold to distinguish flash floods from slow-rising floods (Georgakakos, 1986). Average lag time is 4.98 with a standard deviation of 4.19; the median value is 3.09.

Fig. 10 shows the frequency distribution of lag time. While the range of values covers a relatively large interval of times (up to 16 h), in 37 out of 50 cases lag time is less than 6 h, which is an often used threshold to distinguish flash floods from slow-rising floods (Georgakakos, 1986). Average lag time is 4.98 with a standard deviation of 4.19; the median value is 3.09.

Fig. 11 reports the relationship between lag time and watershed area, with indications of the climate regions. The examination of this relationship requires consideration of the sampling process used to collect the data for the individual catchments. Indeed, for some flash floods, multiple catchments are reported, with very similar storm duration and similar time to peak but with different watershed areas. For these cases, the lag time is almost constant with varying the watershed areas. Owing to this reason, we derived the lower bounded curve enclosing all lag time values in the plot for each value of watershed area to represent the relationship between lag time and watershed area.

We used power-law relationships to represent the envelope curves defining the lower limit of lag time $T_L$ (h) versus basin area $A$ (km$^2$) (Robinson and Sivapalan, 1997). The relationship exhibits a break point in slope at a basin area equal to 350 km$^2$ when plotted on a log–log plot (Fig. 11):

$$T_L = 0.08 \cdot A^{0.55} \quad \text{for } A \leq 350 \text{ km}^2$$

$$T_L = 0.003 \cdot A^{1.10} \quad \text{for } A > 350 \text{ km}^2$$

The relationships show that the lag time amounts to 45 min, 1 h and 6 h for basin areas equal to 50 km$^2$, 100 km$^2$ and 1000 km$^2$, respectively. The higher value of the exponent in Eq. (10) for larger

<table>
<thead>
<tr>
<th>Table 5</th>
<th>Summary statistics of runoff coefficient for different antecedent saturation conditions.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Saturation class</td>
<td>No. of cases</td>
</tr>
<tr>
<td>Dry</td>
<td>17</td>
</tr>
<tr>
<td>Normal</td>
<td>30</td>
</tr>
<tr>
<td>Wet</td>
<td>11</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Table 6</th>
<th>Antecedent saturation conditions under different climatic regions (no. of basins). Alpine and Oceanic regions are not reported owing to the small number of cases.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dry</td>
<td>Normal</td>
</tr>
<tr>
<td>Mediterranean</td>
<td>7</td>
</tr>
<tr>
<td>Alpine–Mediterranean</td>
<td>4</td>
</tr>
<tr>
<td>Continental</td>
<td>0</td>
</tr>
</tbody>
</table>
Eqs. (5) and (9), this velocity is equal to 3 m s$^{-1}$ for basin areas less than 350 km$^2$, implying that for the events and the catchments this similarity indicates that for the events and the catchments. Both these assumptions can be reasonably met with flash flood events.

Lag time, i.e., mean residence time in the basin, depends on two factors. One factor is the distance between the basin outlet and the geometrical center of mass of the event runoff (Woods and Sivapala, 1999). With the assumption of spatial homogeneity for rainfall and runoff, this factor may be related to the mean stream length, which is determined by the planar geometry of the catchment and its associated network. The other factor is the velocity of travel, which is determined by the hydraulics of flow, governed by stream slope, roughness and the elevation geometry of the channel network, and both at-a-site and downstream hydraulic geometry variations. In general, stream velocity could vary in space and time, governed by the spatial and temporal variability of rainfall and catchment characteristics. It is interesting to note that the exponent in the relationship between mean stream length and basin area (Eq. (5)) is very similar to the exponent in Eq. (9). This similarity indicates that for the events and the catchments considered here, the lag time and the stream length increase in the same way for basin areas less than 350 km$^2$, implying that the velocity of travel is remarkably constant in the given range of areas (Sivapalan et al., 2002). When using the values reported for Eqs. (5) and (9), this velocity is equal to 3 m s$^{-1}$. This value represents the characteristic velocity of the flash flood process (at basin scales less than 350 km$^2$), defined as the ratio of characteristic length (mean river length) and time (response time or lag time), and it is related to the celerity with which the flood wave moves through the catchment.

The values of response time defined here may be used to derive recommendations concerning the time resolution and the spatial density of the observation network required for monitoring the flood-generating storms. Based on Berne et al. (2004) and considering a minimum catchment area of 50 km$^2$, the required temporal resolution ranges between 10 and 15 min, with a spatial raingauge density of 1 station per 20–35 km$^2$. If weather radar observations are used to derive the rainfall fields, Sangati et al. (2009) recommends a maximum element size of 2 km to avoid biases in rainfall volume estimation.

The break point arising in the slopes of the envelope curves (Eqs. (9) and (10)) is likely to depend on the transition in topographic and hydraulic properties of the floodplain and channel system with increasing basin size. In almost all the cases considered here, the peak discharge exceeded the riverbank’s holding capacity, with water spreading out into the floodplain. When the peak discharge exceeds river capacity and the floodplain is inundated, a channel-dominated flow transitions into a valley bottom-dominated flow. This transition, which we speculate, may be associated to a basin area of 350 km$^2$, tends to attenuate the flood wave propagation and thus increase lag time, depending on floodplain topography and surface roughness. Flood peak attenuation is largely a result of storage (or greatly reduced velocity) of a portion of the runoff on overbank surfaces (Woltemade and Potter, 1994; Jothityangkoon and Sivapalan, 2003). This storage and the later release of a portion of the total flood volume produce flood hydrographs that are delayed, low and broad compared to those of similar watersheds that lack floodplain storage, such as gullies or mountain streams (Turner-Gillespie et al., 2003).

Detailed analyses by Woltemade and Potter (1994) reveal that moderate-magnitude floods (5- to 50-year recurrence interval) with relatively high peak-to-volume ratios are attenuated most, since the storage of a relatively small volume of water can significantly reduce the peak discharge. In contrast, both small and large floods are attenuated relatively little. Our results contrast with this view and shows that even extreme flash floods may be strongly attenuated with the transition to a valley bottom-dominated flow, probably because the runoff volume is anyway reduced with respect to that characterising large riverine floods. Ongoing investigation is aimed to combine data collection and hydraulic modelling effort to provide more insight into this issue.

Fig. 11b reports data accordingly with the method used to derive the peak time: from streamgauge data or from observations collected and checked during the post-event analysis. Examination of the figure shows that lag times for basin areas less than 100 km$^2$ are mostly derived from post-event analyses. Ruin et al. (2008) show that, during the September 2002 storm in the Gard region, almost half of the casualties occurred on watersheds less than 100 km$^2$. Understanding basin response times and warning lead times in these situations compels for a more extensive and systematic effort on collection of timing data about peak flow during post-event analysis. This is the only resource for documentation of flood dynamics at small spatial scale during these extreme events (Marchi et al., 2009).

6. Conclusions and implications for flood risk management

High-resolution hydrological data concerning 25 major flash floods occurred in Europe in the period 1994–2008 were collected to explore the properties of the impacted catchments, the storm...
characteristics and the features of the rainfall to runoff transformation, including the peak discharges and the runoff volumes. Data from 60 catchments located in a belt from Spain to the Black Sea, and with few cases from Germany, UK and Crete, were analysed in the study. The main results are summarised below.

- Data for more than half of the catchments considered in this study, and around 80% of the data for basins less than 100 km², were provided by post-event analysis. These proportions identify the observational problem which characterises flash floods. On one hand this means that effective analysis of flash flooding requires a systematic effort aimed to carry out in a routine way the program of field visits and post-event analysis after each flash flood event. On the other hand, this implies that new techniques for flash flood hazard should be developed which benefit from the availability of this kind of data. This is exemplified by the work of Gaume et al. (in press), which provides a method for using major flash flood events occurred at ungauged catchments to reduce the uncertainties in estimating regional flood quantiles. The method is based on standard regionalization methods assuming that the flood peak distribution rescaled by a site dependent index flood is uniform within a homogeneous region. Other avenues may be represented by the ‘causal information expansion’ advocated by Merz and Bloschl (2008) to use hydrological understanding of the local flood producing factors to improve the flood frequency estimation at-a-site. Causal information expansion is particularly important in small catchments, both because fewer and shorter records tend to be available than in larger catchments and because the flood processes are more amenable to analysis than in larger catchments where the regional combination of controls can be relatively more important.

- Examination of data shows a peculiar seasonality effect on flash flood occurrence, with events in the Mediterranean and Alpine–Mediterranean regions (Catalonia, Crete, France, Italy and Slovenia) mostly occurring in autumn, whereas events in the inland Continental region (Austria, Romania and Slovakia) commonly occur in summer, revealing different climatic forcing. Consistently with this seasonality effect, spatial extent and duration of the events is generally smaller for the Continental events with respect to those occurring in the Mediterranean region. Furthermore, the flash flood regime is generally more intense in the Mediterranean Region than in the Continental areas. Differences in the spatial and temporal scales of the events should be taken into account in the design of flash flood forecasting and warning systems. Models and procedures specifically fitted to a specific climatic setting may not work equally well in different settings.

- A distinctive morphological features of flash flood catchments is represented by their steepness. Catchments do not need to be particularly steep to favour flash flooding. However, relief is important since it may affect flash flood occurrence in specific catchments by combination of two main mechanisms: orographic effects augmenting precipitation and anchoring convection, and topographic relief promoting rapid concentration of streamflow. This result agrees with earlier scoring procedures proposed by Collier and Fox (2003) and by Collier (2007) to assess flooding susceptibility to extreme rainfall. Since flash flooding results from unique combinations of meteorological and hydrological conditions, more efforts should be devoted to identify specific morphological catchment characteristics affecting this susceptibility. Limitations in Numerical Weather Prediction models and monitoring systems will inevitably limit our ability to forecast flash floods. Owing to this reason, decision support systems are needed to provide timely information and aid those who have to make key decisions at critical times under pressure.

- The event runoff coefficients of the extreme flash floods considered here are rather low, with a mean value of 0.35. This agrees with earlier results obtained by Merz and Bloschl (2003) who reported that, based on their data from Austria, runoff coefficients are smallest for flash floods. There are two important implications for this result. On one hand, it shows the need to account for hydrological conditions in the forecasting of flash floods. On the other hand, it indicates the potential effects of land use change on runoff generation for these events. Land cover, typically, is a local phenomenon, so the impact of any disturbance is likely to strongly decrease with catchment size. However, if the disturbed watershed is impacted by a flash flood–generating storm, the potential for extreme runoff generation is greatly enhanced.

- The influence of antecedent saturation conditions on runoff coefficient has been analysed by considering the ratio of the precipitation in the 30 days before the event to the long-term average precipitation in the same period. Analysis of these results shows a significant impact of the antecedent conditions on event runoff coefficients. These results challenge the common wisdom that antecedent soil moisture is of little importance in determining the magnitude of extreme flash floods. Hence, accounting for antecedent soil moisture conditions is paramount for operational flash flood forecasting. In the typical data-poor conditions which characterise flash flood forecasting and warning, surrogate indexes which can take implicitly into account the soil moisture initial conditions are often extremely useful. This is the case of the Flash Flood Guidance, which tags rainfall accumulations needed to produce a flood of a given magnitude accordingly with the current soil moisture conditions. Norbiato et al. (2009) have shown how Flash Flood Guidance may be applied in ungauged basins with operationally useful results.

- The runoff formation and propagation displays short response times (generally less than 6 h) and a relationship between upstream watershed area and response time was derived. For catchment areas less than 350 km², the response time and the mean stream length increase in the same way, implying that the velocity of travel (equal to 3 m s⁻¹) is remarkably constant in the given range of areas. This value represents the characteristic velocity of the flash flood process (at basin scales less than 350 km²), defined as the ratio of characteristic length (mean river length) and time (response time or lag time), and it is related to the celerity with which the flood wave moves through the catchment.

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