

Floods in Mountain Basins

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Abstract This chapter provides a general introduction to recent research on floods in mountain catchments and reviews state-of-the-art contemporary knowledge on the topic in Poland and Switzerland. The selection of the areas illustrated in this chapter is motivated by the fact that the Swiss Agency for Development and Cooperation (SDC) had funded a research project on floods in the Polish Tatra Mountains and their forelands, to which this book is also dedicated.

Keywords Floods · Flash floods · Mountain environments · Precipitation · Climate change

1 Introduction

Mountain environments cover roughly 25 % of the land surface and are often referred to as ‘natural water reservoirs’; this is because a substantial amount of water surplus is usually transported from mountain areas to adjacent lowlands in some of the largest river systems on Earth (Viviroli et al. 2003). Mountain regions cover 52 % of Asia, 36 % of North America, 25 % of Europe, 22 % of South America, 17 % of Australia, and 3 % of Africa, as well as substantial areas of islands including Japan, New Guinea, and New Zealand (Bridges 1990).

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Floods in mountain basins are often flashy (Borga et al. 2008, 2014), and therefore differ from most other fluvial floods in that the lead time for warnings is generally very limited (e.g., often much less than two hours). Flash floods usually occur in mountain river catchments draining less than 1000 km² (Gaume and Borga 2008; Lumbroso and Gaume 2012). In these environments, direct current meter measurements are often impossible to conduct during flood peaks for safety and technical reasons (Fukami et al. 2008).

One of the greatest difficulties to characterize floods in mountain rivers is that for a given event, several processes coupling between hillslopes and channels may take place concurrently (i.e., debris flows, hyperconcentrated flow, and clear water flow), with different characteristics such as rheology or the number of phases involved (Montgomery and Buffington 1997; Bracken and Croke 2007; Bodoque et al. 2011). Mountain basins often respond rapidly to intense rainfall rates because they have high slopes and a quasi-circular morphology and, as a consequence strong connectivity (Ruiz-Villanueva et al. 2010; Youssef et al. 2011). Likewise, in these basins precipitation has an important orographic component. As a result, precipitation is very variable from a spatio-temporal perspective (Rotunno and Houze 2007). The above determines that mountainous basins are highly prone to have extreme precipitation events, both in terms of total volume and intensity. The resulting floods have a rapid hydrological response, which is characterized by rather “peaky” hydrographs (i.e., hydrographs with short lag times). They reach their flow peaks within a few hours, and thus allow for very little or even no advance warning at all to prevent flood damage (Borga et al. 2007, 2008).

Among the many processes contributing to the generation of floods in mountain rivers, three seem to be key in producing large and/or extreme floods: high-intensity convective rainfall (Gutiérrez et al. 1998; Hicks et al. 2005), extensive deep cyclones generating a few days-long, orographic rainfall (Sturdevant-Rees et al. 2001), and outburst floods resulting from the failure of either natural or artificial

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dams (Cenderelli 2011; Worni et al. 2014; Schwanghart et al. in press). These processes may generate peak discharges greatly exceeding average flood discharges. For instance, glacial-lake outburst floods in Nepal had discharges up to 60 times greater than normal floods generated by snowmelt runoff, glacier melting, and monsoonal precipitation (Cenderelli and Wohl 2003).

In mountain rivers, drainage area and channel gradient, as well as the magnitude and frequency of hillslope failures will not only affect channel morphology, but sediment supply from hillslopes will also influence total sediment load during flood (Lin et al. 2008) as well as the spatial patterns of erosion and deposition along the channel (Cenderelli and Kite 1998; Wohl 2010). Material from mass movement processes (e.g., landslides, debris flows) may not only deliver large amounts of mineral sediments, but also introduce wood debris into the channel corridor. Wood in channels can then favor the creation of temporary debris dams and subsequently produce secondary flood pulses, thereby enhancing erosion, and/or lead to the destruction of infrastructure along the channel (Wohl 2010; Ruiz-Villanueva et al. 2016a, b, c). As a result, floods in mountain rivers often differ from those in lowland environments due to the close coupling between the channel and adjacent hillslopes (Wohl 2010).

Extreme floods can be disastrous, resulting in substantial material losses and/or large numbers of fatalities, if they affect managed valley sections and inhabited valley floors. Examples can be an outburst flood caused by a landslide entering the Vaiont reservoir, Italian Alps, with 2600 deaths (Semenza and Ghirotti 2000), or a flash flood in the Arás basin, Spanish Pyrenees, resulting in 87 fatalities (Gutiérrez et al. 1998). A decisive role in the origin of the disasters must be attributed to inappropriate management decisions, unadjusted to the actual hazard, rather than to the course of natural phenomena. Such was the situation with the Vaiont flood, where water was stored in the reservoir despite the identification of a giant landslide on the valley side above the reservoir (Semenza and Ghirotti 2000) or with the flood on the Arás that destroyed a camp site located on an alluvial fan (Gutiérrez et al. 1998). More frequent are less dramatic situations when flood damages are enhanced by structures that reduce channel conveyance (e.g., Arnaud-Fassetta et al. 2005). Approximately 40 % of the flood-related deaths in Europe between 1950 and 2006 were linked to flash floods (Barredo 2007). The lack of data on flash floods (Marchi et al. 2010), and in particular the lack of accurate discharge estimates (Borga et al. 2014), can often provide an obstacle to improvements in flood forecasting, warning, planning and emergency management.

Less intense floods, especially regularly recurring ones, are key to maintain the function and integrity of aquatic and riparian ecosystems as they allow lateral exchange of water, nutrients and organisms between river channel and the connected floodplain (Junk et al. 1989) and promote recruitment of trees in the riparian areas (Hughes and Rood 2003). However, extreme floods are generally considered as threats for aquatic ecosystems (Wydoski and Wick 2011) that may dramatically though temporarily reduce the abundance of riverine biota such as fish or benthic macroinvertebrates. Because such events cannot be predicted, very few papers compared pre- and post-flood condition of river biocoenosis in mountain rivers.

While the existing studies analyzed direct action of floodwaters on biota (e.g., the effect of the extreme flood of 1997 on the Oder River, Czech Republic, on fish fauna; Lojkasek et al. 2005), no papers have described more prolonged effects of extreme flood events on the physical structure of mountain river habitats.

The Polish-Swiss research project FLORIST has focused on floods in the headwater and foreland reaches of the rivers originating in the Polish Tatra Mountains (Kundzewicz et al. 2014) and has brought together debris-flow, flash flood and mountain river flood experts from both countries. In the following, and in an attempt to set the stage for this book, we briefly summarize a few of the key findings on (flash) flood processes in both countries by reviewing some of the more recent publications on the topic.

2 Physiographic and Meteorological Conditions of Flood Occurrence in the Swiss Alps

In Switzerland, ca. 30 % of the national territory is located between 1000 and 2000 m a.s.l. for which a quick response between precipitation input (often associated with high intensity) and runoff output is the most common process. The sensitivity of these surfaces to produce floods is often exacerbated by the fact that this altitudinal band is also characterized by steep slope gradients and thin soil cover, in addition to the presence of extensive stream networks ensuring high specific discharge and abundant overland flow (Weingartner et al. 2003). Large amounts of precipitation are favored by the uplift of air masses, either in the form of rather localized, convective uplift (i.e. a vertical movement induced by vertical instability in the atmosphere) or cyclonic (also called advective) uplift in a horizontal plane of a warm over a cold air mass (Grebner et al. 2000). Under current climatic conditions prevailing in the European Alps, Schwab et al. (2001) point to upper intensity rates of up to 100 mm/h for convective rainfalls, but the ongoing increase of mean and extreme temperatures might likely lead to increasing precipitation intensities in the future (Gobiet et al. 2014). In the past, floods induced by convective storms were most frequent during the summer half-year, but recently have started to occur in all seasons as a result of the decreasing snow cover and the reduced buffering effect of snow (Beniston 2005). In the case of advective precipitation events induced by closed cyclonic fields, the areas affected can extend for several 100 km² (Grebner et al. 2000). The intensity of advective rainfalls is much smaller than that of convective showers, and typically reaches values of 10–20 mm/h north of the Alpine divide (Grebner et al. 1999).

The last very large flood in the Swiss Alps, Prealpine foreland and plateau occurred in August 2005, causing damage in the order of 3 billion Swiss Francs (DETEC 2008). The circulation pattern triggering the devastating event is referred to as a *Vb situation* and is quite common in this part of Europe. Situations comparable to that of August 2005 occur several times per year, but the large volume of rainfall and the duration of the event over such a large area should be considered as

quite unusual. The meteorological event of August 2005 is comparable with the heavy precipitation events of June 1910, July 1977, August 1987 and May 1999. Prolonged and intensive periods of precipitation may also be expected in Switzerland in the future—possibly more frequently than in the past due to global climate change (DETEC 2008).

In terms of process activity in small, Alpine catchments, information is much scarcer and mostly relies on archival records and/or reconstructions. Along with flash floods, debris flows represent one of the most common processes in high mountain areas and are commonly initiated by the mobilization of sediment stored in channels or by shallow landslides, in addition to sudden input of large amounts of water, such as rainstorms, rapid snow melt, rain-on-snow events, or the sudden release of water from glaciers or from dammed lakes. Most commonly, however, debris flows in the Alps have been triggered by high-intensity, short-duration rainstorms or low-intensity, long-duration precipitation events, typically during the summer half-year (Stoffel et al. 2011; 2014a, b; Schneuwly-Bollschweiler and Stoffel 2012; Toreti et al. 2013). Projected changes in mean and extreme temperatures and precipitations are likely to influence the temporal frequency and magnitude of mass wasting in mountain environments (Gobiet et al. 2014). This is especially true for debris flows, where changes in rainfall intensity and duration, in combination with higher temperatures, are thought to lead to enhanced process activity, provided that sediment is not limited and that the occurrence of events is driven primarily by water input above a certain threshold (Stoffel and Huggel 2012; Borga et al. 2014). A warmer climate also results in higher 0 °C isotherms, thus allowing for more precipitation to fall in liquid form even in the uppermost portions of mountain catchments, thereby increasing the area contributing effectively to runoff (Beniston 2005; Stoffel and Beniston 2006). In the case of the Zermatt valley (Valais, Swiss Alps), projections of future precipitation have been realized in the framework of the EU-FP7 ACQWA Project (Beniston and Stoffel 2014). Based on point-based downscaled climate scenarios and for the periods 2001–2050 and 2051–2100, analysis of temperature and rainfall changes above specific thresholds (10–50 mm) as well as the duration of precipitation events (1–3 days) reveals a drying tendency for future summers and more precipitation during the shoulder seasons (Stoffel et al. 2014a, b). At the same time, despite the general decrease in precipitation sums in summer, an increase in the occurrence of heavy (>40 mm) 1-day precipitation events is observed in the region. In conclusion, the drier conditions in future summers and the wetting of springs, falls and early winters are likely to have significant impacts on the behavior of debris flows. Based on the current understanding of the debris-flow systems and their reaction to rainfall inputs, one might expect only slight changes in the overall frequency of events by the mid-21st century, but possibly an increase in the overall magnitude of debris flows due to larger amounts of sediment delivered to the channels and an increase in extreme precipitation events. In the second half of the 21st century, the overall absolute number of days with conditions favorable for the release of debris flows will likely decrease, especially in summer. The anticipated increase of liquid rainfalls during the shoulder seasons (March, April, November, December) is not

expected to compensate for the decrease in future heavy summer rainfalls over 2 or 3 days in absolute terms, but magnitudes, in contrast, can be expected to increase in the study area. The volume of entrained debris from the source areas tends to be larger in summer and fall when the active layer of the permafrost bodies is largest and allows for larger volumes of sediment to be mobilized (Lugon and Stoffel 2010), but the situation has been shown to depend also on the stability and climate change-related accelerations of rock-glacier bodies (Stoffel and Huggel 2012). Along with the occurrence of more extreme precipitation events, these rock-glacier instabilities could lead to debris flows without historic precedents in the future.

3 Physiographic and Meteorological Conditions of Flood Occurrence in the Polish Carpathians

The Carpathians cover 6 % of the territory of Poland. They are relatively low mountains in comparison to the Swiss Alps and elevations above 2000 m a.s.l. occur only in the Tatra Mountains shared by Poland and Slovakia (areas of 22.3 and 77.7 %, respectively), with the highest peaks: Rysy (2499 m) in the Polish part and Gerlach (2655 m) in Slovakia. The Polish part of this high-mountain massif represents less than 1 % of the area of the Polish Carpathians. In the remaining part of the Polish Carpathians, almost entirely underlain by flysch, the highest peak reaches 1725 m a.s.l. and relative elevations between the bottoms of intramontane basins and mountain summits vary between 500 and 1200 m. This physiographic setting is clearly reflected in the character of geomorphic hazards. Debris flows occur relatively frequently only in the Tatra Mountains (Kotarba 1992) where, however, they can hardly be considered a threat to man in this uninhabited and undeveloped area. Floods are the hazards most damaging to human property and dangerous to life in the region. Their timing and principal drivers vary along the west-east gradient (Chełmicki et al. 1999). The eastern part of the Polish Carpathians, with lower elevations and more continental climate, is characterized by frequent occurrence of moderate, snow-melt floods in early spring and rare occurrence of large floods caused by summer rainfall. Snow-melt floods are typical of the south-eastern part of the Upper Vistula Basin, and in its Carpathian portion they mainly occur in the foothill parts of the San and Wisłok catchments. Their occurrence depends on the snow-cover duration and the thickness of snow pack at the end of winter. A weak decreasing trend in snow-cover duration and a more significant trend in the snow-pack thickness were observed in the second half of the 20th century (Falarz 2004). Favourable conditions for a snow-melt flood on the Wisłok River were recorded in Rzeszów after the snowy winter 1963/1964 with 52 cm-thick snow pack in March. The last such thick snow pack in Rzeszów (53 cm) was recorded in March 2005. The duration of snow cover at this location during the last warm period 1988-2015 was shorter by 10 days than in the more cooler years 1951–1987. Recently possibility of the occurrence of snow-melt floods has been very low. During the last three winters 2013/2014–2015/2016 with the duration of snow cover between 20 and 28 days, the maximum thickness of snow

pack was lower than 10 cm. In higher parts of the Carpathians, snow pack is relatively thick, but snow-melting period extends from April at lower altitudes to early June in the higher parts of the Tatras. As a result, snow-melt floods in these parts of the Carpathians are very rare.

In the western part of the Polish Carpathians, with higher elevations and more oceanic climate, floods are caused by summer rainfall. Summer floods differ in duration and extent according to the triggering rainfall (Starkel 1996). A few days-long rainfall with the total sum of precipitation amounting to a few hundred millimeters, sometimes preceded by a longer period of increased precipitation, that saturates shallow soil, results in floods encompassing the whole or a considerable part of the Polish Carpathians and even the whole Upper Vistula Basin. The four largest floods in the Polish Western Carpathians occurred in July 1934, 1970, 1997 and in May 2010 and were triggered by a few days-long rainfall with average intensity of 8–10 mm h⁻¹. Such precipitation was always caused by cyclones coming along the Vb track from the Mediterranean region to Central-Eastern Europe where they became partly stationary. On the western side of such cyclones, humid air masses are moving from N or NE perpendicularly to the Carpathian chain and produce prolonged, orographically strengthened rainfall. On the northern slopes of the mountains, daily totals of precipitation reach 100–250 mm (Cebulak 1992) and three-days totals amount to 300–500 mm—see Niedźwiedź and Łupikasza (this volume) for more details about these synoptic situations. In the western part of the Upper Vistula Basin (west of the Dunajec), the floods caused by prolonged rainfall are more frequent than in the eastern part (Cebulak 1992).

High-intensity rainfall with 50–150 mm of precipitation in one to a few hours is connected to convective thunderstorms (Cebulak and Niedźwiedź 1998) and results in flash floods of local extent (Bryndal 2014), which however can be highly damaging. Such was a flash flood that occurred in July 2006 in the upper Wisłoka catchment; the intensive rain caused water stage on a tributary to the Wisłoka to increase rapidly by 3.5–4 m and the flood destroyed infrastructure in the valley and caused a few fatalities (Izmańłow et al. 2006). Several heavy downpours (>8 mm h⁻¹) or even torrential rains (>44 mm h⁻¹) of 0.5–2 h duration were recorded in different parts of the Polish Carpathians (Starkel 2011). Some of them are worthy of note: 7 June 1985 in Szymbark, 25 July 2001 in Maków Podhalański, 14 July 2002 in Muszyna, 26 July 2005 in Baligród, 5 June 2007 in the Western Tatra Mountains, and many others.

Finally, the occurrence of flysch bedrock in most of the Polish Carpathian area causes landslides to be a frequent type of geomorphic hazard that in some cases results in considerable damage to buildings or infrastructure. Landslides are frequently triggered by the same rainfall events which induce major floods in the region, but may also be activated in wet years with relatively high annual sum of precipitation dispersed over several months (Froehlich and Starkel 1995). Such extremely rainy season in the Polish Carpathians was summer 1913 (Starkel 2011).

The high Tatra Mountains feature abundant precipitation, higher than elsewhere in Poland. The highest daily precipitation totals in the Polish Tatra Mountains reach 300 mm (Polish record, overall) observed on 30 June 1973 in Hala Gąsienicowa during a northern cyclonic situation Nc. However, the occurrence of thick and highly porous slope covers and deep, karstic circulation result in relatively low

flashiness of runoff from the Tatras (Kundzewicz et al. 2014). For instance, the runoff-irregularity coefficient (the ratio of the highest and the lowest discharge on record) for the Koniówka gauging station on the Czarny Dunajec River equals 453. The flysch Carpathians are typified by low water storativity, and thus runoff from this area is more flashy, despite lower sums of precipitation (Kundzewicz et al. 2014). The most extreme case seems the upper part of the Biała catchment with only shallow, slope aquifers, for which the runoff-irregularity coefficient at the Grybów station amounts to 7500.

4 Recent Advances in Flood Research in the Polish Carpathians

Over the few past decades a change in the seasonality of floods in the Polish Carpathians has occurred (see Ruiz-Villanueva et al. 2016b who demonstrated such a change in rivers in the northern foreland of the Tatra Mountains). Major rain-induced floods, previously typically occurring between June and August, started to occur also in May (2010, 2014) and September (1996, 2007).

Major floods cause substantial modification of channel morphology (Gorczyca et al. 2013; Hajdukiewicz et al. 2016) and result in considerable economic losses. The flood of July 1934 was the largest one in the Dunajec catchment, especially in the northern foreland of the Tatra mountains (Fig. 1), and inundated vast areas in the Upper Vistula Basin. The flood occurred in the final part of a long historical period typified by low forest cover and intense agricultural and pastoral use of hillslopes in the Polish Carpathians. As a result, disastrous character of the flood was largely related to the deposition of huge amounts of coarse bed material in the channels and on valley floors (Fig. 1). In the last decades, large floods originating in the Polish Carpathians occurred in 1997 and 2010, but they were borne in largely reforested mountain catchments and hit the rivers typified by sediment deficit. They resulted in material damage of the order of 1 % of gross national product of the country (Kundzewicz et al. 2012). A considerable proportion of the damage occurred in the valleys of Carpathian tributaries to the Vistula (Biedroń et al. 2011) as a result of the erosional action of the flood flow (Fig. 2); downstream of mountain reaches, the huge masses of floodwaters propagated along the Vistula valley, causing a failure of embankments and inundation of extensive areas at several locations, even 400 km away from the mountains (Kozak et al. 2013).

During the two last decades attention has been paid to the recruitment, transport and deposition of large wood by floods and its role as an agent of flood hazard in Polish Carpathian rivers. Kaczka (1999) was the first who indicated in Polish literature that during floods large numbers of fallen trees are delivered to mountain stream flowing through forested corridors, and the same was next demonstrated with respect to wide mountain rivers (Wyżga and Zawiejska 2005). Wood accumulations causing clogging of bridge cross-sections appeared to be an important



Fig. 1 View of Bystry Stream in its reach within the town of Zakopane just after the catastrophic flood of July 1934. The flood was the largest one on record in this region, with peak discharges of streams flowing from the Tatra Mountains up to three times higher than the largest ones recorded in the second half of the 20th century. Visible are destroyed houses in the riparian area and huge amounts of coarse bed material deposited practically at the level of the developed valley floor. Currently the stream flows in 6 m-wide, stone-lined, artificial channel. The photo no. 1-G-4647-12 of the National Digital Archive of Poland

cause of flood damage during floods (Hajdukiewicz et al. 2016) and research aimed to better understand large wood dynamics in mountain watercourses was carried out using different methods such as wood inventories (Wyźga et al. 2015), numerical modelling (Ruiz-Villanueva et al. 2016c, d, e) or tracking experiments with tagged wood (Wyźga et al. 2016b).

Successful management of flood risk requires proper recognition of the frequencies and magnitudes of flood events, characterization of hydraulic parameters of flood flows and innovative approaches to river management. Small headwater streams are typically ungauged in the Polish Carpathians, but at the same time floods borne in their catchments may provide significant risk to downstream, developed valley reaches because of short concentration times of the flood waves reflected in their flashy nature. Using dendrogeomorphic techniques, Ballesteros-Cánovas et al. (2015) identified 47 flood events that occurred on four streams in the Tatra Mountains over the last 148 years, and discussed synoptic situations leading to their triggering. In turn, Ballesteros-Cánovas et al. (2016) demonstrated that reconstruction of paleodischarges of the floods occurring on these small, headwater streams allows not only to extend existing flow records but also to improve the estimation of flood frequency distributions. Bryndal (2014)



Fig. 2 Damage to infrastructure caused by the 80-year flood of June 2010 on the Biała River. A significant increase in forest cover in the catchment after World War II and widespread in-channel gravel mining conducted in the past decades resulted in sediment starvation of the river and thus the flood action was mainly directed on erosion of the channel boundaries

analysed physiographic parameters of small catchments in the Polish Carpathians that were affected by flash floods and proposed a method allowing for identification of catchments especially prone to the generation of such floods. Wyżga et al. (2016a) analysed variability in the hydraulic importance of channel incision with increasing river size by comparing changes in the frequency of valley floor inundation at gauging stations on the Dunajec River characterizing catchments with an area ranging from 34.5 to 6735 km². The analysis indicated that the relative increase in channel capacity and the resultant loss of floodwater storage on valley floors, associated with a given absolute (i.e. expressed in metres) amount of channel incision, will be most severe in the upper river courses where the channels initially had a relatively small capacity. Czech et al. (2016) analysed the impact of river restoration on the conditions for flood flows by comparing hydraulic parameters of flood flows between closely located unmanaged and channelized cross-sections of the Biała River. They indicated that despite a short time since the beginning of the river restoration, it already brings beneficial effects for flood risk management, reducing flow energy and shear forces exerted on the bed and banks of the channel in unmanaged river reaches. Mikuś et al. (2016) analysed geomorphic responses of the multi-thread Czarny Dunajec River to floods of different magnitude and demonstrated that protection of a local road from erosion does not require river channelization but can be attained through flow re-activation in side braids.

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