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Relevance of the correlation between precipitation and the 0 °C isothermal altitude for extreme flood estimation

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ABSTRACT

Extreme floods are commonly estimated with the help of design storms and hydrological models. In this paper, we propose a new method to take into account the relationship between precipitation intensity (P) and air temperature (T) to account for potential snow accumulation and melt processes during the elaboration of design storms. The proposed method is based on a detailed analysis of this P-T relationship in the Swiss Alps. The region, no upper precipitation intensity limit is detectable for increasing temperature. However, a relationship between the highest measured temperature before a precipitation event and the duration of the subsequent event could be identified. An explanation for this relationship is proposed here based on the temperature gradient measured before the precipitation events. The relevance of these results is discussed for an example of Probable Maximum Precipitation-Probable Maximum Flood (PMP-PMF) estimation for the high mountainous Mattmark dam catchment in the Swiss Alps.

The proposed method to associate a critical air temperature to a PMP is easily transposable to similar alpine settings where meteorological soundings as well as ground temperature and precipitation measurements are available. In the future, the analyses presented here might be further refined by distinguishing between precipitation event types (frontal versus orographic).

1. Introduction

The use of design storms and related design flows is a key tool for the design of hydraulic infrastructures (Niemi et al., 2015; Swain et al., 2006; Salas et al., 2014). In the case of dam design, spillways are frequently designed to withstand the probable maximum flood (PMF) (Swain et al., 2006; Salas et al., 2014). The PMF is defined as "the theoretical maximum flood that poses extremely serious threats to the flood control of a given project in a design watershed. [...]" (WMO, 2009). The PMF definition of the Bureau of Reclamation, USBR, 1974, is less severe; it is considered to be the "largest flood that can reasonably be expected to occur on a given stream at a selected point. Furthermore, the World Meteorological Organization (WMO, 2009) specified that the PMF is "converted from the probable maximum precipitation (PMP) over a design watershed". The PMP is defined as "the theoretical maximum precipitation for a given duration under modern meteorological conditions [...]" (WMO, 2009). The conversion from PMP into PMF is generally done by routing the PMP through a rainfall-runoff model (Swain et al., 2006; WMO, 2009). A key issue in PMF estimation is the choice of the modeling method and the uncertainty quantification (Jothityangkoon et al., 2013; Beauchamp et al., 2013; Lagos-Zúñiga and Vargas, 2014; Salas et al., 2014; Brigode et al., 2015; Haddad and Rahman, 2016). Depending on the availability of data, event-based or continuous simulations can be used for the rainfall-runoff routing (e.g. Jasper et al., 2002; Zeimetz et al., 2015). Data availability constraints frequently impose the choice of event-based approaches, as it is for example the case for the analysis carried out by Zhang and Smith (2003).

In snow-influenced environments, a key factor for the estimation of event runoff volumes is the aggregation state of the event precipitation (rainfall or snowfall) and the amount of event runoff that is due to the meltwater outflow from a potentially preexisting snowpack. As discussed e.g. in the work of Schaefli (2016), both quantities can be assumed to strongly depend on air temperature. A precise estimation of the aggregation state of precipitation at a given altitude (of the snowfall limit) would require knowledge of the wet-bulb temperature (e.g. Tobin et al., 2012), but air temperature has been shown to be a good proxy for hydrological purposes (Rohrer et al., 1994). Air temperature is also a dominant driver for melt processes in most alpine environments, since long wave radiation and sensible heat (two of the three

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dominant energy sources for melt) are strongly influenced by air temperature (Ohmura, 2001).

Accordingly, the relation between precipitation intensity (P) and temperature (T) and the effective runoff and snow melt are key factors for flood estimations. This P-T relationship is generally studied in the context of climate change research, where an often raised question is how strongly the precipitation intensity is likely to increase with increasing temperature (Pall et al., 2006).

Brandsma and Buishand (1997) pointed out that in Switzerland an increase of the mean daily precipitation with temperature was detectable and was also dependent upon the direction and the strength of the atmospheric flow. Lenderink and Van Meijgaard (2008), Lenderink and Van Meijgaard (2008) and Lenderink and Van Meijgaard (2010) could show similar results for Europe. They stated that precipitation extremes can raise twice as fast as expected by the Clausius–Clapeyron relation for hourly precipitation. Hardwick Jones et al. (2010) pointed out that in Australia, extreme precipitation events are more sensitive to a temperature increase and that in this world region, the P-T scaling becomes negative for temperatures beyond 20–26 °C. Moreover, they showed that the scaling depends on the duration of the precipitation event.

The large scale study of Pfahl and Niedermann (2011) led to the conclusion that at mid-latitude, the correlation between the temperature and the relative humidity of the air mass above the ocean is positive. They confirm the super-Clausius–Clapeyron relation mentioned earlier by Lenderink and Van Meijgaard (2008) for air masses over the ocean. Pfahl and Niedermann (2011) identified advection and meridional transport as main drivers for the correlation between temperature and relative humidity. However, these results cannot readily be transposed to land areas and are therefore not directly applicable in the context of flood estimation.

Later, Westra et al. (2012) performed a large scale analysis taking into account a large dataset of annual maximum precipitation across the globe that allowed them to conclude that at mid-latitudes, the association between temperature and precipitation is significant. They showed a clear correlation between the latitude and the percentage increase per degree warming for the northern hemisphere. Above 50 °N, the strength of the association becomes even stronger. Berg et al. (2009) showed that the dependence between temperature and precipitation was seasonally conditioned. In addition, Berg and Haerter (2011) analyzed the P-T scaling for different precipitation types in Germany. They concluded that for extreme hourly precipitation, super-Clausius–Clapeyron rates are possible for all precipitation types. But both Berg and Haerter (2011) and Lenderink and Van Meijgaard (2008) pointed out that the P-T scaling is less noticeable for daily precipitation.

As mentioned earlier, Hardwick Jones et al. (2010) showed that negative scaling is possible for high temperatures, which has been confirmed in the work of Lenderink et al. (2011) and Utsumi et al. (2011) for temperatures above $\sim 24\,^{\circ}$C. However, the analysis of Shaw et al. (2011) showed that for the United States, negative scalings have not been observed during the summer months. Furthermore, they stated that a super-Clausius–Clapeyron behaviour was noticeable but far less important than in the study of Lenderink and Van Meijgaard (2008), Lenderink and Van Meijgaard (2008) and Lenderink and Van Meijgaard (2010). Shaw et al. (2011) even raised the question whether temperature of the upper atmosphere can actually be characterized with the surface temperature. They stated that in some cases, the surface temperature may not represent the temperature of the upper atmosphere.

Later, Westra et al. (2014) argue in their detailed review that negative scalings might be due to moisture availability limitations in the case of high temperatures. Panthou et al. (2014) found similar results for Canada and could nuance that "the longer an event was, the less pronounced was the increase of extreme rainfall intensities with temperature". Similarly, Wasko et al. (2015) showed in a study conditioning the scaling on precipitation event duration that, in Australia, moisture availability limitations cannot be found for short events (1 h–2 h). In Switzerland, Molnar et al. (2015) concluded that there could be limitations of moisture availability. However, this conclusion was drawn without discussing its dependence on the precipitation duration. Busuioc et al. (2016) also found evidence for limitations in moisture availability for Romania.

According to Drobinski et al. (2016), moisture limitation might however not be the only explanation for negative scaling. They stated that the negative scaling could be overestimated by the fact that the surface temperature is not a reliable quantity to estimate the temperature at higher altitudes in arid conditions. The temperature could be highly overestimated, inducing a too pronounced negative scaling (Drobinski et al., 2016).

To summarize, it can be retained that a dependence between temperature and precipitation has been found in different studies all over the world. The nature of the P-T relationship differs according to latitudes, seasons and precipitation quantiles. There appears to be a stronger dependence for smaller time steps. Furthermore, negative scaling can occur due to moisture availability limitations, but these limitations can depend on the precipitation event duration.

At the time of this writing, the relationship between temperature and precipitation intensity *and* duration has not yet been studied for central Europe. This paper analyzes the relationship between precipitation and temperature taking into account event duration and seasonality for Switzerland.

After having introduced the PMP-PMF concept and discussed the state of the art of temperature-precipitation scaling, these scaling relations are analyzed for Switzerland and their relevance in the context of PMF estimates is addressed. The context of this study is the estimation of extreme floods within a PMP-PMF (Probable Maximum Precipitation – Probable Maximum Flood) framework for potentially snowfall-influenced environments. PMP events are usually defined independently from air temperature, which, as discussed earlier, is however essential to assess the potential effects of snowfall and melt. This paper thus proposes to use P-T scaling analysis to associate a critical air temperature to PMP events. The basic idea is hereby to identify a potential maximum temperature threshold beyond which precipitation starts decreasing with air temperature. If such a threshold does not exist, the P-T scaling cannot be directly used to associate a temperature to PMP events because PMPs are by construction far beyond observed precipitation amounts. An example of PMP-PMF estimation in the presence of snow accumulation and melt processes is included to underline the importance of a careful selection of the air temperature associated with PMP events. The overall aim of this paper is to come up with recommendations about how to select the initial air temperature for critical precipitation events of different durations.

2. Data

2.1. Precipitation data

Hourly precipitation data are provided by MeteoSuisse at 104 locations (Fig. 1). The hourly precipitation measurement started in 1981. However, not all 104 stations were operational since 1981 and some of them have been removed since then. A total of 52% of the stations have a record of over 30 years. Roughly 18% of the stations have been added during the period from 1984–2010. After 2013, 95% of the considered rain gauges were already

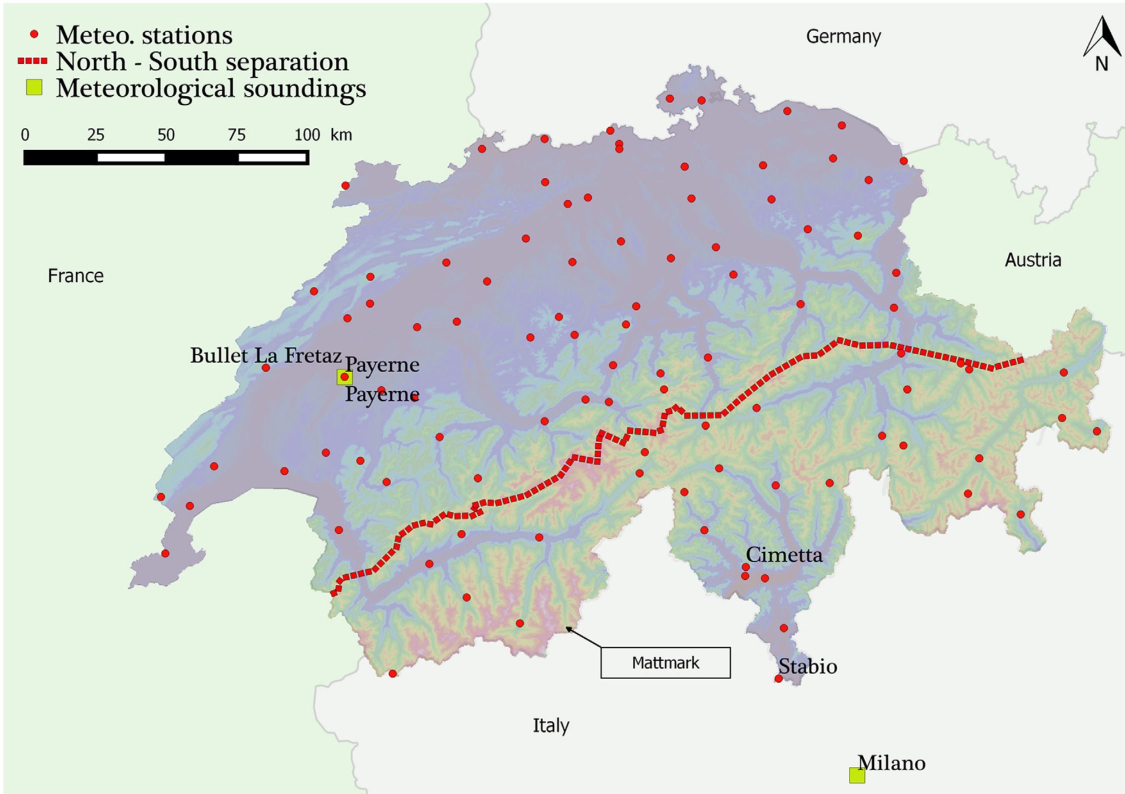

Fig. 1. Situation of the considered meteorological measurement stations and of the sounding sites on a topographic map with indications of the surrounding country boundaries.

installed. For the present study, all stations have been considered for the available data length.

2.2. Temperature: meteorological soundings

In addition to ground-based air temperature (observed at all above mentioned stations), we propose here to use air temperature observations from meteorological soundings, i.e. from weather observation balloons filled with helium that are launched twice a day (at 00:00 and 12:00 UTC) to measure air temperature through the atmosphere. The balloons are surveyed up to an altitude of 30–35 km. The use of such soundings rather than ground-based temperature observations has the main advantage of yielding a direct observation of the atmospheric conditions representative for an entire region. This avoids potentially complicated spatial interpolations of point observations and makes the proposed method more straight-forward for practical applications.

The analyses carried out here are based on meteorological data from Switzerland. This relatively small country (40,000 km^2) shows several clearly distinct hydro-climatological regions (Froidevaux, 2014), which can be roughly separated into two major zones (Schiemann and Frei, 2010): north and south of the Alps (Fig. 1). For this analysis, meteorological soundings from Payerne (CH) and Milan (It) are considered. The situation of the two cities is shown on Fig. 1. The soundings of Payerne can be assumed to be roughly representative of the climatic conditions for the meteorological stations north of the Alps and the soundings from Milan for the meteorological stations situated south of the Alps. The subdivision as well as the attribution are shown on Fig. 1. At Milan, the soundings were available from 1973 to present. For the soundings at Payerne, the measurements started in 1981.

If using temperature soundings, the reconstruction of the ground temperature or the estimation of isothermal altitudes is influenced by the assumption made about how the temperature varies with altitude, i.e. by its gradient or lapse rate. As shown in this paper, the seasonal variation of this gradient is small and extreme precipitations occur mainly during warm periods, for which the gradient is relatively constant (Table 1). Therefore, the error induced by the assumption on the gradient remains relatively small.

3. Methodology

The analysis of the relationship between the 0 °C isothermal altitude and precipitation characteristics is completed here in two steps: i) an analysis of the correlation between the 0 °C isothermal altitude measured before a precipitation event and the precipitation intensity and ii) an analysis of the maximum measured 0 °C isothermal altitude before a precipitation event and the duration of the subsequent event. These two analysis steps allow for the determination of whether the precipitation intensity is reaching a maximum for a certain temperature or if a relation between the event duration and the temperature is detectable.

The knowledge of this two relations is important for extreme flood estimation, because the flood estimation is more reliable when the meteorological inputs of a design storm are coherent. Ensuring this coherence is particularly important when the design storm is not obtained with a meteorological model (which would necessarily return temperature and precipitation as a coherent couple), as it is the case for most approaches, i.e. statistical extrapolations (Meylan et al., 2008), frequency duration curves (Meylan et al., 2008), Hershfield method (Hershfield, 1961; WMO, 2009).

Table 1
Retained temperature gradients for the four validation stations and the mean relative error of the 0 °C isothermal altitude estimation.

Sounding	Meteo. station	Δ_{summer} [°C/100 m]	Δ_{winter} [°C/100 m]	ϵ_{sum} [−]	ϵ_{win} [−]
Payerne	Bullet la Fretaz	−0.6	−0.4	0.08	0.33
Payerne	Payerne	−0.65	−0.4	0.13	0.57
Milan	Stabio	−0.6	−0.4	0.08	0.36
Milan	Cimetta	−0.6	−0.6	0.07	0.16

Concerning the precipitation data, the following steps are performed. First, the rain gauges are attributed to the geographical classes in order to take into account the climatic regions mentioned above. Then the precipitation events measured at each rain gauge are determined and the precipitation volume and the event duration are derived. The events are then classified in precipitation duration classes (3.2). The next step is to separate the dataset into seasonal sets in order to take into account the meteorological seasonality. Berg et al. (2009) focused on the data of July to characterize the scaling in summer and the data of January for the winter time with the argument that the trends are the strongest for these periods. This paper follows the same approach but extends the period to 3 months in order to increase the amount of considered data and capture the full season. The chosen months for the summer are June to August and December to February for the winter. Spring and autumn are defined by the remaining months.

An essential step is the validation of the coherence of the meteorological soundings and ground-based temperature observations. This was completed by comparing the 0 °C isothermal altitude, derived from the meteorological soundings, with the isothermal altitude derived from ground temperature measurements with the temperature gradient approach. Once the meteorological soundings were validated, they were used to determine the 0 °C isothermal altitude **before** the precipitation events for each climatological zone. Next the scaling between the altitude and precipitation quantiles (as it was done by Lenderink and Van Meijgaard (2008) and Lenderink and Van Meijgaard (2010)) was analyzed for different precipitation durations. To characterize the antecedent temperature – event duration relation in rare to extreme temperature conditions, the maximum measured 0 °C isothermal altitude for the different duration classes was also derived.

3.1. Definition and determination of the precipitation events

For the purpose of this study, the following criteria were used to define a precipitation event (where i stands for precipitation intensity). An event starts at the moment when $i > 0.1$ mm/h and ends when $i < 0.1$ mm/h.

In order to analyze the sensibility of the results to the definition of the precipitation events, a third criterion was introduced. It aimed at grouping the precipitation amounts of consecutive precipitation events that were separated by less than a certain duration λ. If the end date of the first precipitation event and the start date of the next event were separated by a time interval of λ hours (where $\lambda \in \{0, 1, 2\}$), these consecutive events were considered to be a single event, with the start date of the first event and the end date of the last event. Initially, the value of λ was fixed to 0. Then the analysis was repeated twice for the other two values. A graphical illustration of this third criterion is available in the supplementary material on Fig. S1.

3.2. Classifying the precipitation events in precipitation duration classes

For this analysis, duration classes are defined. The following classes are considered (in hours): $\{[1,2],[3,4],[5,8],[9,12],[13,18],[19,24]$. The classes have been defined in order to be coherent with the available Swiss PMP maps (Hertig et al., 2005) that have been elaborated for 1 h, 3 h, 6 h, 9 h, 12 h and 24 h precipitation durations.

3.3. Analysis of the meteorological soundings

3.3.1. Determination of the 0 °C isothermal altitude

The analysis of the meteorological soundings was based on the hypothesis that the temperature varies linearly through the troposphere with an inversion at the tropopause (at approximately 10 000 m a.s.l.). The altitude of the 0 °C isothermal altitude was deduced from a linear interpolation of the data corresponding to each sounding.

In order to increase the reliability of the interpolations, the following criteria have been retained for the determination of the 0 °C isothermal altitude. (i) The data set of the meteorological soundings had to contain more than two measurements (minimum data for the definition of a line). (ii) The measured temperature during one sounding had to change sign, meaning that the balloon passes through an altitude with 0 °C. (iii) The coefficient of determination r^2 of the trendline and the soundings had to be higher than 0.8. Fig. 2 shows an example of the interpolation and the derived 0 °C isothermal altitude. It can be seen that the assumption of a linear behaviour up to 8 000 m a.s.l. was valid.

3.3.2. Validation with ground-based air temperature

Ground-based temperature data from meteorological stations was used to check that the altitude of the 0 °C isotherm (derived from soundings) could be reproduced with the use of a simple temperature gradient from a surface air temperature. This was in fact a condition to ensure that the identified precipitation-temperature relationships could be used in a standard hydrological modelling setting where the ground-based air temperature time series are used to identify the aggregation state of precipitation and the melt conditions.

For this validation, it is important to distinguish between meteorological stations located in the mountains or in the valleys because of the thermal inversions that regularly occur in winter times. Two stations have been chosen for each micro-climatic region, Payerne and Bullet La Fretaz for the North and Cimetta and Stabio for the South (see Fig. 1). The geographical coordinates of these stations are reported in Table 2 in the Swiss reference coordinate system CH1903/LV03.

The reconstruction of the 0 °C isothermal altitude with the temperature gradient method is completed as follows:

$$H_{iso}(t) = H_{stat}(t) - \frac{T_{stat}(t)}{\Delta}, \quad (1)$$

where $H_{iso}(t)$ [m a.s.l.] is the 0 °C isothermal altitude at time step t, $H_{stat}(t)$ [m a.s.l.] is the altitude of the meteorological station and $T_{stat}(t)$ °C is the measured temperature at this station. $\Delta(t)$ is the temperature gradient for the time step t. Here a single gradient is estimated for summer and a lower gradient for winter (see Section 4.1). For the purpose of the present validation, the reconstruction was based on two different gradients, one for the winter and one for the summer period, which corresponds to common practice in hydrological modelling (Schaefli and Huss, 2011).

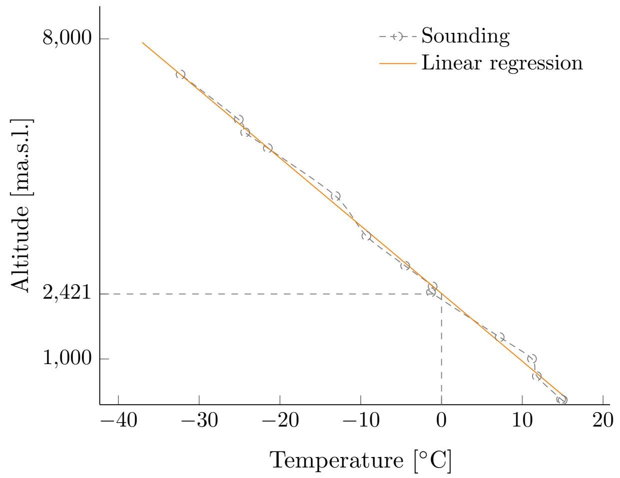

Fig. 2. Example of meteorological sounding data (Temperatures and corresponding measurement altitudes) and of the linear regression used for the determination of the 0 °C isothermal altitude.

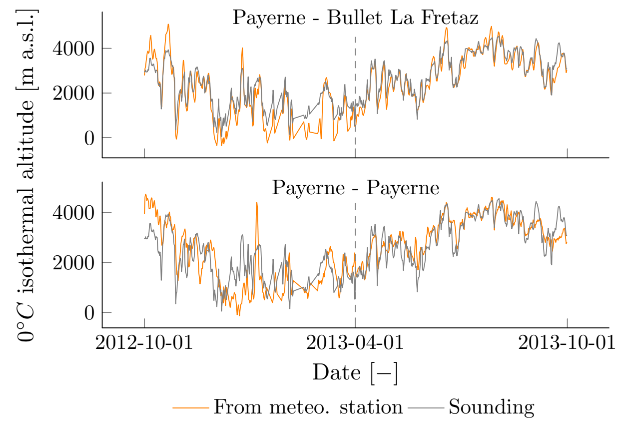

Fig. 3. Comparison between the 0 °C isothermal altitude derived from meteorological soundings of Payerne and from ground temperature measurements at Bullet la Fretaz and Payerne. The dashed line indicates the separation into winter and summer.

Table 2
Coordinates X [m], Y[m] and Z [m a.s.l.] (in Swiss national coordinate system CH1903/LV03) of the meteorological stations used for the validation of the gradient approach and of the sites where the meteorological soundings have been undertaken.

Station name	X	Y	Z
Meteorological stations			
Bullet La Fretaz	534 230	188 080	1 202
Payerne	562 150	184 855	490
Cimetta	704 370	117 515	1 672
Stabio	716 040	77 970	353
Meteorological sounding stations			
Payerne	562 200	184 800	491
Milano	743 870	43 708	103

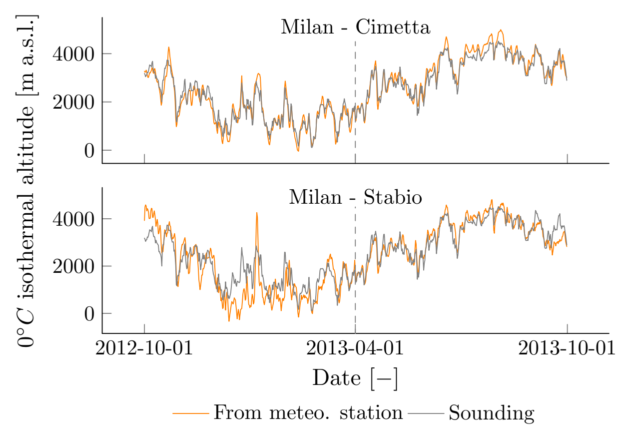

Fig. 4. Comparison between the 0 °C isothermal altitude derived from meteorological soundings of Milan and from ground temperature measurements at Cimetta and Stabio. The dashed line indicates the separation into winter and summer.

The quality of the reconstruction of the 0 °C isothermal altitude is measured in terms of the mean relative error ϵ (Eq. (2))

$$\epsilon = \frac{1}{N}\sum_{t=1}^{N}\frac{H_s(t) - H_{iso}(t)}{H_s(t)} \tag{2}$$

where N is the number of estimation-measurement couples and H_s is the isothermal altitude measured by the meteorological sounding.

4. Results and discussion

4.1. Validation of reconstruction

The identified gradients for the reconstruction of the 0 °C isothermal altitude as well as the mean relative error ϵ are shown in Table 1. The obtained gradients corresponded well to the known range of gradients in the Alps (Damm and Felderer, 2013; Rolland, 2003). It should be mentioned that these gradients have not been determined by minimizing the estimation error. The goal was only to show whether the 0 °C isothermal altitude is representative for the ground temperature. For this purpose, the gradients have been determined by trial-and-error. The results are sufficiently good to show the strong link between the ground temperature and the 0 °C isothermal altitude. The reconstruction based on Eq. (1) gave reliable results for both regions and both periods, as can be seen from the performance measure in Table 1 and from the reconstructed time series of isothermal altitudes (Figs. 3 and 4).

A more detailed inspection showed that the reconstruction of the isothermal altitude (and thus also the back-calculation from the isothermal altitude to ground temperature) for the two stations Cimetta and Stabio was good during winter (mean relative error $\epsilon_{win} \leqslant 0.36$) and very good during summer (mean relative error $\epsilon_{sum} \leqslant 0.08$), which suggests that a single gradient during each season is a good solution despite the fact the some convective summer events might show stronger gradients. For the meteorological stations La Fretaz and Payerne, the reconstruction was also very good for summer (mean relative error $\epsilon_{sum} \leqslant 0.13$) and slightly less close to the sounding-based isothermal altitude for winter (mean relative error $\epsilon_{win} \leqslant 0.57$). However, the reconstruction could also be considered very good due to the frequent temperature inversions in winter. Given the good linear relationship between the 0 °C isothermal altitude and ground temperature, the remainder of this paper reports all results in terms of the 0 °C isothermal altitude.

It is important to mention that the linear relationship between the 0 °C isothermal altitude and ground temperature might break down during local weather phenomena that typically occur in mountainous areas. This might in particular be the case during periods of thermal inversions or during Foehn conditions. The former situations typically correspond to anticyclonic conditions when no precipitation is occurring. Consequently, the errors induced by thermal inversions or Foehn events are not critical for this analysis because it focused on extreme precipitation.

4.2. Quantile scaling analysis

The precipitation events are classified into the 9 classes discussed in Section 3.2. Events of more than 72 h have been omitted here because the number of measured events was very small. For each duration class, the 0.5, 0.75, 0.90, 0.95 and 0.99 precipitation intensity quantiles have been estimated from all couples of precipitation intensity (P) and of 0 °C isothermal altitudes (H_s). Hereby, the mean intensities over the duration of the events were used.

Quantiles were only computed for sample sizes that were large enough for direct quantile identification from the sample. For example, for the quantile 0.99 the sample must have had at least 100 values. It turned out that for longer precipitation durations than those who fell in the class [19, 24], not enough data was available for returning representative results. Therefore, the quantile scaling analysis did not consider precipitation durations longer than 24 h.

Figs. 5 and 6 show examples of these $P - H_s$ scalings for the soundings from Payerne and the seasons summer and winter. The results for fall and spring as well as for the soundings from Milan are given in the supplementary material. These figures show $P - H_s$ scaling plots (for $\lambda = 0$) which, as discussed earlier, are equivalent to the more classical precipitation-temperature ($P - T$) scaling plots.

During summer, the $P - H_s$ scaling showed a clear increase in precipitation intensity with increasing 0 °C isothermal altitude for the soundings from Payerne (Fig. 5). No sign of a limitation of moisture availability could be detected up to the precipitation duration class [13,18], for which the scaling was much flatter. Wasko et al. (2015) found that moisture limitations could not be confirmed for short events. Our results are thus in line with their findings. The flatness of the scaling for the duration class [13, 18] could indicate a transition from short events not influenced by moisture limitation to longer events where these limitations can play a role. During the summer season, not enough events fit the duration class [19,24] to warrant inclusion in the analysis.

Concerning the soundings from Milan (available in supplementary material, Fig. S4), the same trend can be observed for the duration classes [1,2] and [3,4]. For the duration class [5,8], moisture availability limitations could cause the decrease of the scaling rate for the 99% quantile line for high 0 °C isothermal altitudes. The more erratic behaviour of the 99 % quantile of the duration class [9,12] made it hard to draw a conclusion. However, a clear increase was not detectable for the mentioned quantile. The scaling of the events that fitted the class [13,18] was positive over the entire range of 0 °C isothermal altitudes. For the class [19,24], not enough events had been recorded to estimate the 99% quantile. The lower quantiles indicated a nearly constant scaling above 3000 m a.s.l..

For winter, the Payerne soundings (Fig. 6) suggested the existence of moisture availability limitations for the duration classes [1,2] and [3,4] for high 0 °C isothermal altitudes. The scalings of the classes [5,8] and [9,12] showed an increasing trend without any signs of moisture limitations. The 99% quantile line of the duration class [13,18] showed a nearly constant scaling, that could also have been a sign of moisture availability limitations. The scaling for the duration class [19,24] had a decreasing trend for high 0 °C isothermal altitudes when considering the 99% quantile. Concerning the soundings from Milan (S6), the duration class [1,2] showed that an upper limit for the precipitation intensity could have been reached for a certain isothermal altitude. The negative scaling of the 99% quantile line appeared rather abruptly; however, the quantile points that led to this negative scaling were calculated based on more than 500 data points. The lower 95% quantile also indicated a negative scaling for the same 0 °C isothermal altitude. The other duration classes suggested that moisture availability limitations could have led to the decrease in scaling rates for the 99% quantiles.

To illustrate the sensitivity of the results to the precipitation event definition, the soundings from Payerne were used as an example. The results for $\lambda = 1$ and $\lambda = 2$ were very similar as illustrated in Fig. 7 for the 99% quantile. The results for the other quantiles are available in the supplementary material (Figs. S8–S11). These figures show that the similarity of the results based on the definitions $\lambda = 1$ and $\lambda = 2$ compared to those deduced from $\lambda = 0$ is not dependent on the quantiles. The sensitivity analysis did not show a significant change in the observed trend as a function of the event duration. The sensitivity analysis for the Milan soundings led to the same conclusion (results not shown).

For spring, the scaling rates were clearly decreasing for high 0 °C isothermal altitudes for the duration classes [1,2] as well as [3,4] when the Payerne soundings (Fig. S3) have been considered. However, for the duration classes [5,8] and [13,18], the scaling remained positive over the entire 0 °C isothermal altitude range. On the other hand, the duration class [9,12] seemed to lead to reduced scaling rates at high 0 °C isothermal altitudes, but the tendency was not clear. For the [19,24] class, the same reduced scaling was noticeable. Regarding the soundings from Milan (Fig. S7), a clear reduction of the scaling rates was recognizable for the classes [1,2] and [3,4]. The scaling of the events of class [9,12] was even negative for high 0 °C isothermal altitudes when looking at 95% quantiles or lower. Concerning the 99% quantile, not enough values have been measured for the very high 0 °C isothermal altitudes, thus the positive scaling was not representative. A less pronounced decrease for high 0 °C isothermal altitudes was visible for the duration class [13,18]. Unfortunately, the class [19,24] did not allow to conclude on the 99% quantile.

For fall, a decreasing scaling rate was perceivable for the Payerne soundings (Fig. S2) for the duration classes [1,2], [3,4], [5,8] and [9,12]. The two other classes showed a constant increase of the scaling rate. Concerning the soundings from Milan (Fig. S5), only the class [9,12] showed a decreasing scaling rate for the 99% quantile. When looking at the 95% quantile, also the classes [1,2] and [13,18] showed a decreasing trend for high 0 °C isothermal altitudes.

The above discussion shows that the scaling plots do not lead to a clear and unique conclusion of an upper precipitation limit for a certain 0 °C isothermal altitude. The nature of the $P - H_s$ scaling is different for different duration classes and the low amount of "extreme" data leads to low confidence for the points with high isothermal altitude. Therefore a reliable conclusion on a maximum precipitation associated to a certain isothermal altitude (beyond which the intensity would decrease again) cannot be drawn. Nevertheless, signs of limitations of moisture availability can be detected for both short and long events independently of the season, but are not clearly related to the precipitation duration. Furthermore, the maximum 0 °C isothermal altitude observed within each precipitation duration class showed a clear continuous decreasing trend with precipitation duration (Fig. 8). This observation holds for all seasons and all regions and does not depend on the event definition that has been shown to have minor influence (Figs. 7 and S8–S11). The winter shows the lowest maximum isothermal altitudes for a given duration class, followed by spring, fall and summer (for Payerne and Milan). Except for winter, the maximum isothermal altitudes were slightly higher for Milan than for Payerne.

An analysis of the temperature gradient, deduced directly from the soundings (slope of regression line in Fig. 2) could lead to an explanation of the behaviour of this decreasing isothermal altitude with precipitation duration. Fig. 9 showed, in fact, that the mean value of the gradients for the three event definitions was

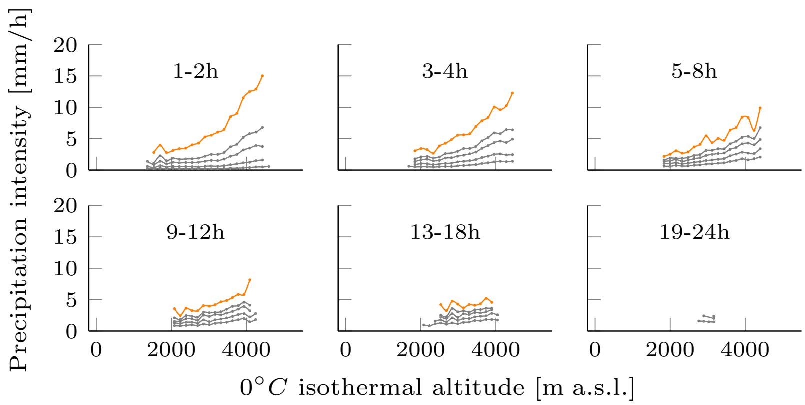

Fig. 5. 0.5, 0.75, 0.90, 0.95 and 0.99 quantiles (highlighted in orange) of the different considered precipitation duration classes south of the Alps (Payerne meteorological soundings) and for the summer period (June-August) for $\lambda = 0$.

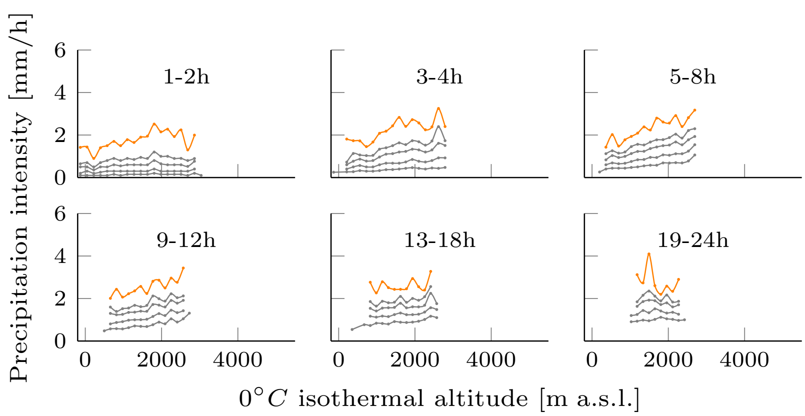

Fig. 6. 0.5, 0.75, 0.90, 0.95 and 0.99 quantiles (highlighted in orange) of the different considered precipitation duration classes south of the Alps (Payerne meteorological soundings) and for the winter period (December-February) for $\lambda = 0$.

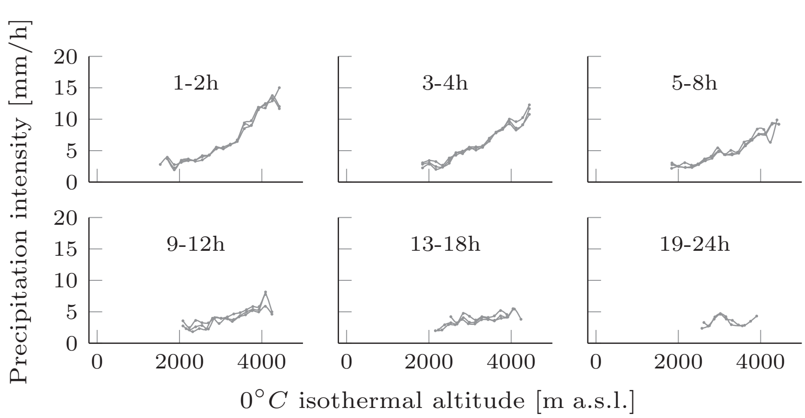

Fig. 7. Superposition of the results based on the three definitions ($\lambda = 0, 1, 2$) for the precipitation events. Represented are the results for the soundings from Payerne during the summer period for the **0.99** quantile.

decreasing with duration (the negative value was increasing). The decrease of the gradient values before a long rainfall event can be explained as follows. Long-lasting precipitation events mostly occur during warm fronts where the uplift and subsequent adiabatic cooling of the air mass is slow, and this lead to relatively low environmental lapse rate (temperature gradient). On the contrary short rainfall events occur during cold fronts or strong convective/orographic episodes, where uplift and adiabatic cooling of the air mass is fast and violent, thus the temperature gradient in that layer of the atmosphere will be larger. However, it may also be due to the air mass flux acceleration. This could induce the air mass to lift and consequently to cool down, leading to the homogenization of the air mass, inducing smaller gradients. Long rainfall events often occur on generalized degradations of the weather, and thus in colder situations (lower 0 °C isothermal altitude) than short heavy rainfall episodes. This generalized alteration of the weather tends to produce effects far in advance. Gradient decrease could be one of them.

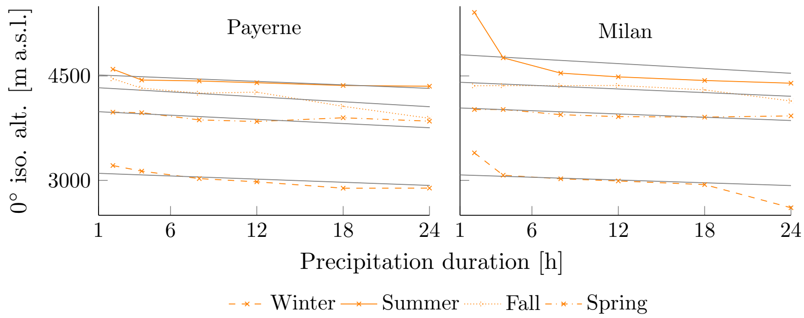

Fig. 8. Maximum 0 °C isothermal altitude plotted against the precipitation duration. The altitude values are the mean values of the three considered event durations.

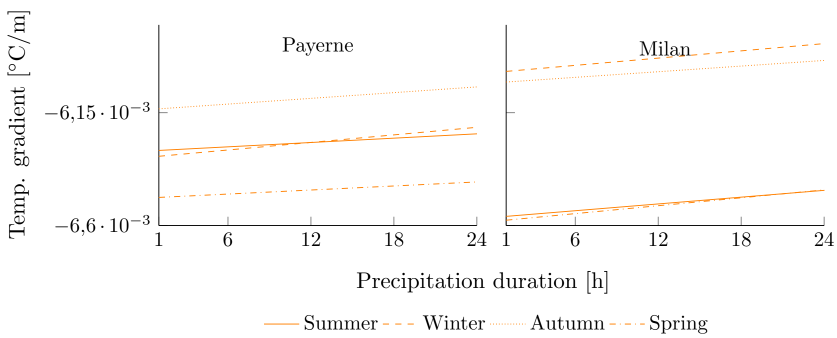

Fig. 9. Temperature gradients plotted against the precipitation duration.

5. Relevance of the results for extreme flood estimation

5.1. Discussion of the 0 °C isothermal altitude in a hydrological context

The presented scaling analysis and in particular the absence of a scale-break (decreasing scaling beyond a certain isothermal altitude) seems to suggest that in the Swiss Alps, even the highest precipitation intensities can occur with very high 0 °C isothermal altitudes, up to 4800 m a.s.l. as has been observed through meteorological soundings from Milan during summer. Comparing this value to the hypsometric curve of Switzerland (Fig. S12), showed that most of the Swiss territory does not reach such altitudes. Thus this would cause the majority of the Swiss Alpine catchments to contribute entirely to the discharge generation (no solid precipitation) and that, in presence of a pre-existing snow pack or in presence of glaciers, melt processes would play a crucial role. In the context of PMP-PMF, this result would a priori require to consider the maximum measured isothermal altitude in relation with the PMP. This assumption is conservative because it could lead to an overestimation of the snow or/and glacier melt. As PMP-PMF is used for the estimation of the safety flood for dam design, a conservative assumption is justifiable.

Analyzing the precipitation-duration-0 °C isothermal altitude scaling on a seasonal basis shows, however, two important facts that are not directly visible from a traditional precipitation-temperature scaling analysis: i) the maximum 0 °C isothermal altitude per duration class strongly depend on the season and ii) the maximum 0 °C isothermal altitude decrease with precipitation duration.

The above points imply that in certain seasons (winter, spring), high intensity precipitation cannot occur as rainfall at all elevations but will partly fall as snow. The maximum isothermal altitude was as low as 2550 m a.s.l. for a 72 h winter event in Payerne. This might have practical implications for flood estimation in catchments for which the critical situations tend to occur in winter and spring. For high Alpine hydropower reservoirs, critical situations are known to occur only in late summer and early fall when the reservoirs are full. The results presented here suggest that for this time of the year, very high 0 °C isothermal altitudes (implying a high percentage of rainfall and melt contribution) should be assumed for all precipitation durations when safety flood estimations are undertaken.

The identified decreasing trend of maximum isothermal altitude as a function of precipitation duration was 230 m/24 h (average for the 4 seasons and the two sounding stations). For small catchments where the relevant precipitation duration for PMP estimation will be short, this low gradient suggests that the choice of an 0 °C isothermal altitude should essentially depend on seasonal considerations. For larger catchments with higher relevant durations for the PMP, the 0 °C isothermal altitude decrease might become relevant. The lowering of the 0 °C isothermal altitude of up to 500 m for summer for events of 72 h might play a crucial role regarding the simulation of discharge in Alpine catchments. In the following case study, it has been assumed that PMP events occur under high temperature. Summer conditions are therefore considered for the hydrological application.

5.2. Application: the case study of the Mattmark dam catchment

The relevance of the relationship between air temperature and precipitation event duration for extreme flood event simulations under summer conditions is illustrated in this paper with the GSM-Socont hydrological model (Schaefli et al., 2005; Jordan et al., 2012; Schaefli and Zehe, 2009) used for precipitation-runoff simulation. This model is a semi-distributed conceptual hydrological model developed for mountainous catchments.

We applied the model to the catchment of the Mattmark dam in the Southern Swiss Alps (Fig. 1). Around 28% (10 km^2) of the entire catchment surface, 36 km^2, is covered by glaciers. The basin goes from 2174 m a.s.l. to around 4000 m a.s.l. Given the size of the study catchment, this model does not take into account river routing.

The considered precipitation event was a 6 h-PMP derived from the Swiss PMP maps (Hertig et al., 2005) with a mean intensity of 56 mm/h. An example of a hyetograph of this 6 h-PMP is shown in Fig. S13 for a selected point of the catchment. This temporal structure resulted from rainfall mass curves (WMO, 2009) and has been kept constant over the entire catchment. The spatial structure was given by the Swiss PMP maps (Hertig et al., 2005) with a resolution of 2 × 2 km^2. 0 °C isothermal altitude intervals starting at 3000 m a.s.l. (low 0 °C isothermal altitude for summer) and going up to 4500 m a.s.l. (high 0 °C isothermal altitude for summer) in a 500 m step, have been considered in order to assess the sensitivity of the PMF to the 0 °C isothermal altitude.

5.2.1. Hydrological model set up

The snow accumulation has been simulated with a linear transition from snowfall to rainfall at temperatures between 0 °C and 2 °C (which fits well with observed snow and rainfall data (Rohrer et al., 1994). The snowmelt has been computed with a degree-day approach (Hock, 2003). Melt water leaves the snowpack only if a certain liquid-to-solid threshold (set to $\theta_{cr} = 0.1$) is reached and can refreeze during periods of negative temperatures. The melt and rainwater (in case of rain-on-snow events) that leaves the snowpack is assumed to infiltrate into the subsoil. If the soil is saturated, no infiltration occurs. The melt and rainwater are evacuated as runoff. The catchment-scale runoff resulting either from snowpack outflow or from direct rainfall on snow-free areas is computed via a two reservoir approach (fast and slow component). Runoff from glacier-covered areas is computed with a separate ice-melt module that uses also a degree-day approach and a linear reservoir to transform melt water into runoff.

The initialization of the hydrological model is done for summer conditions as the PMP maps are admitted to represent summer PMP values. The model has been initialized with the median summer snow height (3 mm equivalent water height) and the median summer soil moisture (0.1 m equivalent water height corresponding to 40 % of the infiltration capacity). The snow pack is considered to be saturated.

5.3. The role of the 0 °C isothermal altitude for summer PMF

To illustrate the effect of the choice of a 0 °C isothermal altitude on the probable maximum flood (PMF) estimation of a typical high Alpine Swiss catchment, Fig. 10 shows the results of the hydrological simulation for Mattmark (see Section 5.2) under several assumptions of the 0 °C isothermal altitude. Given the high elevation range of this catchment and the glacier cover, the increase of the simulated flood discharge as a function of isothermal altitude was very strong, which emphasized the importance of a detailed analysis of the 0 °C isothermal altitude for flood estimation in similar catchments. The simulation results illustrate that, under summer conditions, the influence of the isothermal altitude on the flood peak is important. Thus, in the context of safety flood estimations, the possibility of very high 0 °C isothermal altitudes should not be neglected.

The simulations showed that the main difference between the flood discharges for the different 0 °C isothermal altitudes came from the precipitation aggregation state. Solid precipitations for areas higher than the 0 °C isothermal altitudes reduced significantly the flood discharge if the 0 °C isothermal altitude was low. For the warmest scenario (0 °C isothermal at 4500 m a.s.l.), the initial snow cover melted down during the first 40 min of the event. The glacier discharge of this worst case does not contribute significantly to the flood discharge. In fact, only 1.75% of the total discharge are generated by the glacier. The rather low difference between the simulations that considered 4000 m a.s.l. and 4500 m a.s.l. as 0 °C isothermal altitude can be explained by the fact that the catchment area above 4000 m a.s.l.. is very small. In general, the sensitivity of the PMF discharge is less pronounced for catchments with less topographical variability. For catchments that are too low to experience snowfall during summer, the PMF sensitivity to the 0 °C isothermal altitude would only depend on the spatial distribution of the pre-existing snowpack. The limiting case are lowland areas with zero probability of showing a pre-existing snowpack during summer.

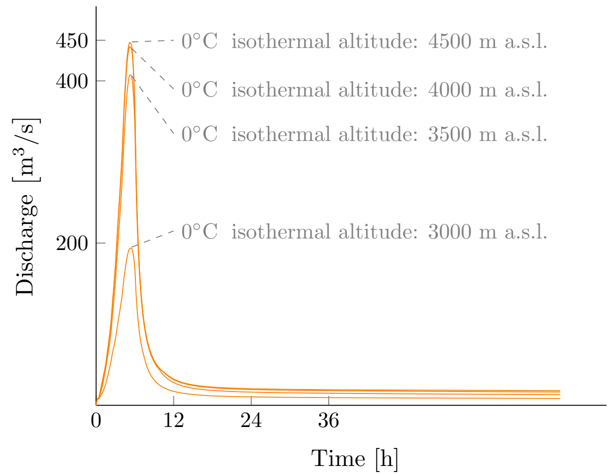

Fig. 10. A ratio of 4.3 can be observed between the peak discharges of the PMF hydrographs for an 0 °C isothermal altitude varying from 3000 m a.s.l. to 4500 m a.s.l..

6. Conclusions

This paper presented an analysis of the scaling of precipitation intensity as a function of the 0 °C isothermal altitude and of precipitation duration for Switzerland. In the context of potentially snow-influenced extreme flood estimation, the 0 °C isothermal altitude has been shown to be very closely related to air temperature, whose link to precipitation intensity is more classically analyzed in the literature.

Three conclusions can be drawn. The first one is that the observed precipitation data did not show any clear upper limit for the increase of precipitation intensity with the 0 °C isothermal altitude (or, equivalently, with air temperature). However, in agreement with earlier findings, moisture availability limitations were noticeable. Concerning these limitations, we found that the southern and northern part of Switzerland had a different behaviour and that they could not be clearly related to the precipitation duration or the season. For all the duration classes, positive and negative (or constant) scaling could be found. The largest number of classes without moisture availability limitations were detected during the warmest period.

The second conclusion element is that the maximum isothermal altitude per precipitation duration class strongly depended on the season. The maximum 0° isothermal altitudes for the winter season were similar for the south and the north of Switzerland. On the other hand, the maximum 0° isothermal altitudes for the spring, summer and autumn period were higher in the southern part of Switzerland.

The third outcome is that the maximum isothermal altitude is approximately linear with precipitation duration. It has been shown that this relation can approximately be described by a linear regression. The analysis of the temperature gradient helped to explain this relation. In particular, the homogenization of the air mass temperature due to more pronounced fluxes before longer rainfall events could lower the 0 °C isothermal altitude before long duration events.

These findings should be considered for event-based rainfall-runoff simulation of rare events with the help of PMPs in settings where snowfall and snow and ice melt processes might play a role. As illustrated for the high Alpine Mattmark catchment (36 km^2), even an increase of a couple of hundred meters of the 0 °C isothermal altitude for a short duration precipitation event in summer might have a significant impact on flood estimation in a comparable high elevation catchment.

The presented results suggest that seasonal considerations might play a crucial role for the choice of 0 °C isothermal altitudes in PMP-based flood estimation studies. These seasonal considerations might reduce the estimated flood peak for settings where critical events are known to occur during winter and spring when 0 °C isothermal altitudes are lower than during summer and fall. It is noteworthy, however, that available PMP maps often only apply to summer precipitation conditions (as is the case for Switzerland, where the maps have been derived for warm conditions).

Future research should focus on the evolution of the air temperature during precipitation events. Furthermore, a precipitation event type distinction (frontal or convective) could be performed to quantify the influence on the presented scaling results and namely on the relationship between precipitation duration and the maximum 0 °C isothermal altitude.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.jhydrol.2017.05.022.

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