Meteorological characterizations of extreme precipitation and floods in Switzerland

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presented by

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Abstract

Floods are amongst the most damaging natural hazards in the Alpine region. These disastrous events are often caused by long-lasting and/or heavy precipitation, which is strongly determined by synoptic to mesoscale conditions and orographic effects. Among the remaining challenges posed by heavy precipitation events, especially those influenced by topography, the estimation of the probable maximum precipitation is particularly critical for estimating hydrological risks. A better understanding of the factors governing precipitation in complex topography is therefore of primary importance.

This thesis investigates the sensitivity of extreme precipitation events with an idealized approach by artificially modifying some atmospheric parameters in high-resolution numerical weather prediction model experiments. In particular the sensitivity of the distribution and intensity of precipitation to changes of the specific humidity and temperature in the initial and boundary conditions is investigated. For this purpose, six precipitation events from both the south side and the north side of the Alps are investigated, the December 1991, September 1993, October 2000, August 2005, August 2007, and October 2011 events.

The formation of orographic precipitation is due to three main ingredients: (a) the horizontal transport of moist air masses by synoptic scale weather systems; (b) the saturation of the air mass through terrain-forced ascent governed, on the mesoscale, by the stability of the air mass; and (c) the transformation of the condensate to precipitable hydrometeors on the microscale.

Mainly the last ingredient, the microphysical aspect, shows a significant and systematic effect on the precipitation response in the sensitivity experiments. The moisture influx effect is mainly leading to an increase of the duration of periods with high moisture influx values rather than of peak values. The role of the stability is crucial for determining the type of flow (blocked or unblocked) and therefore the formation of the precipitation, but does not show a strong systematic
response in the sensitivity experiments.

The sensitivity of the microphysical processes can be roughly classified in two main categories, the unblocked and the blocked flow categories. Both categories experience on average a greater increase of the precipitation at low altitudes than at high altitudes, particularly in the specific humidity sensitivity experiments. In addition, each category has two sub-categories, one where the impact of the sensitivity experiments is mainly located upstream of the control run precipitation, and one where it is also located downstream (towards higher altitudes).

In addition, the large-scale circulation and the moisture uptake of the October 2011 event are investigated. For the first time, an important pathway for northern Alpine flooding is documented, in which the interaction of synoptic-scale to large-scale weather systems and of long-range moisture transport from the Tropics are dominant. Moreover, the trapping of moisture in a subtropical cut-off near the West African coast is found to be a crucial precursor to the observed European high-impact weather.

Finally the large-scale circulation and moisture uptakes of 36 flood events spread over a 140-year period are examined, classifying the events according to their moisture uptake regions. This novel classification shows that four classes, which are consistent in the overlapping period from the Twentieth Century Dataset and the ECMWF Interim Re-Analysis dataset (1979-present), are capable of capturing the variety of the large-scale dynamics leading to floods in Switzerland.
Résumé

Les inondations comptent parmi les risques naturels les plus dévastateurs de la région Alpine. Ces événements sont souvent causés par des précipitations intenses et/ou de longues durées, fortement déterminées par les conditions météorologiques allant de l’échelle synoptique à la méso-échelle ainsi que par les effets du relief. Parmi les défis posés par les événements précipitant intenses, en particulier ceux influencés par le relief, l’estimation des précipitations maximales probables est particulièrement importante pour déterminer les risques hydrologiques. Une meilleure compréhension des facteurs qui gouvernent les précipitations dans un relief accentué est donc d’importance primordiale.


La formation des précipitations orographiques est due à trois ingrédients principaux : (a) le transport horizontal de masses d’air humide par des systèmes météorologiques d’échelle synoptique; (b) la saturation de la masse d’air par soulèvement orographique, soulèvement qui est contrôlé, à la méso-échelle, par la stabilité de la masse d’air; et (c) la transformation des condensés en hydrométéores à la micro-échelle.

C’est principalement le dernier ingrédient, à la micro-échelle, qui produit un effet significatif et systématique sur les précipitations lors des expériences de sensibilités. L’effet sur l’apport en humidité est surtout lié à une augmentation de
la durée sur laquelle le flux d’humidité est élevé, plutôt que des valeurs de pointe elles-mêmes. Le rôle de la stabilité est crucial pour déterminer le régime d’écoulement (bloqué ou non-bloqué) et donc la manière par laquelle les précipitations se forment. Par contre, la stabilité n’est pas affectée par le type d’expériences de sensibilités effectué ici.

La sensibilité des processus microphysiques peut être classifiée grossièrement en deux catégories : bloqué et non-bloqué. Les deux catégories présentent, en moyenne, une plus grande augmentation des précipitations à basses altitudes qu’à hautes altitudes, en particulier lors des expériences de sensibilités à l’humidité. De surcroît, chaque catégorie est constituée de deux sous-catégories: une où l’impact des expériences de sensibilité est concentré en amont des précipitations de la simulation de contrôle et une autre où l’impact est aussi ressenti en aval (vers les altitudes plus élevées).

De plus, la circulation à large échelle et l’absorption d’humidité par l’atmosphère lors de l’événement d’octobre 2011 sont examinés. Pour la première fois, une importante voie d’acheminement pour une inondation du nord des Alpes est documentée. Pour cette voie, l’interaction de systèmes météorologiques à différentes échelles (de l’échelle synoptique à la méso-échelle) ainsi que le transport d’humidité sur de longues distances, depuis les Tropiques, ont une importance considérable. En outre, la captation de l’humidité par un tourbillon potentiel proche de la côte ouest de l’Afrique se révèle être un précurseur crucial pour cet événement européen extrême.

Finalement, la circulation à large échelle et l’absorption d’humidité par l’atmosphère lors de 36 événements qui conduisent à des inondations sont examinés, ceci sur une période de 140 ans. Les événements sont classifiés selon les régions où l’absorption d’humidité a lieu. Cette nouvelle manière de classifier montre que quatre classes sont capables de représenter la variété des circulations à large échelle conduisant à des inondations en Suisse. De plus, ces classes sont cohérentes lorsqu’elles sont définies par deux différents ensembles de données (Twentieth Century Dataset et ECMWF Interim Re-Analysis dataset (1979-present)).
Chapter 1.

Introduction

1.1. Motivation

Floods are amongst the most damaging natural hazards in the Alpine region. In Switzerland they represent 37% of all damage covered by insurance from 1991 to 2010 (Imhof, 2011). These disastrous events are often caused by long-lasting and/or heavy precipitation which is strongly influenced by synoptic to mesoscale conditions and orographic effects (Sénési et al., 1996; Massacand et al., 1998; Buzzi et al., 1998; Ralph et al., 2006; Martius et al., 2013). However, forecasting heavy precipitation remains a challenge for numerical weather prediction, which requires collaborative efforts of both weather services and meteorological research (Buzzi and Foschini, 2000; Richard et al., 2007; Rotunno and Houze, 2007; Rottach et al., 2009). Heavy precipitation events in Alpine regions are typically associated with a particular large-scale flow configuration but the event itself is often small in scale. This localization of the heavy precipitation is due to the impact of the steep and complex orography.

One region where orographic enhancement of heavy precipitation occurs frequently is the south side of the Alps (Frei and Schär, 1998). Several previous studies have contributed to a better understanding of these precipitation events. However, significantly fewer studies have investigated the atmospheric conditions linked to heavy precipitation events (HPEs) in the northern Alps, although an increase of flooding frequency and related impacts have been observed in this region in recent decades (Schmocker-Fackel and Naef, 2010).

Despite the remaining challenges posed by HPEs, and especially those influ-
enced by topography, estimation of the probable maximum precipitation (PMP) is critical in estimating hydrological risks. Indeed the PMP is essential to estimate the probable maximum flood (PMF) and therefore for the design of hydrological structures. Current methods to estimate PMP are described in WMO (Geneva: World Meteorological Organization, 2009.), but these methods are not well suited for small catchments located in areas with complex topography. For example a common method to estimate the PMP for a particular catchment where no observations of precipitation are available is to translate an extreme precipitation event from a nearby area in the catchment of interest. However, due to the strong variability of precipitation in mountainous regions, this may lead to a large over/underestimation of the PMP.

This thesis investigates the sensitivity of extreme precipitation events with an idealized approach by artificially modifying some atmospheric parameters. In particular the sensitivity of the distribution and intensity of precipitation to changes of the specific humidity and temperature is investigated. For this purpose, past precipitation events from both the south side and the north side of the Alps are investigated.

1.2. Orographic effects on precipitation

Precipitation as a response of the interaction between atmospheric flows and topography is an important component of the global water cycle. However mountains rarely create precipitation, but rather modify or enhance precipitation by interacting with pre-existing weather systems such as extratropical and tropical cyclones, or atmospheric rivers.

Orographic precipitation is commonly decomposed into three basic ingredients (Rotunno and Houze, 2007; Houze, 2012; Colle et al., 2013), that are examined in the following paragraphs.

(a) The horizontal transport of moist air masses by synoptic scale weather systems.

(b) The saturation of the air mass through terrain-forced ascent governed, on the mesoscale, by the stability of the air mass.
The transformation of the condensate to precipitable hydrometeors happening on the microscale.

Numerous studies have investigated the role of synoptic scale weather systems in the transport of moisture in various regions of the globe. On the south side of the European Alps, approaching upper-tropospheric troughs were found to be precursors for HPEs (Massacand et al., 1998; Massacand et al., 2001; Martius et al., 2006; Nuissier et al., 2008; Argence et al., 2008). Precipitation days above the 99% quantile in the southern Alps are in 73% linked with an upper-tropospheric trough (potential vorticity (PV) streamer) (Martius et al., 2006). A key factor for distinguishing the severity of extreme precipitation events was found to be the moisture influx.

Several studies have shown the importance of so-called Atmospheric Rivers (ARs) (Newell et al., 1992; Ralph et al., 2006) in triggering HPEs, especially over the western coast of Norway (Stohl et al., 2008; Sodemann and Stohl, 2013), for winter flooding in Britain (Lavers et al., 2011), and over the west coast of the USA (Dettinger et al., 2004; Ralph et al., 2006; Neiman et al., 2011; Cordeira et al., 2013). Lavers et al. (2011) defined ARs as elongated regions with an integrated water vapour in the atmospheric column of more than 2 cm, located mainly in the lower troposphere within the warm sector of an extratropical cyclone and within areas of strong winds (greater than 12.5 m s\(^{-1}\)). Recently Lavers and Villarini (2013) showed that ARs leading to extreme precipitation in Europe may penetrate as far inland as Poland. They found the strongest relation between ARs and annual maxima of precipitation in mountainous areas. Using a Lagrangian approach, Knippertz et al. (2013) found that about 90% of the ARs are of tropical origin and therefore described them as Tropical Moisture Exports (TMEs). TMEs are defined as tropical air masses (with a origin within 20°S and 20°N) that reach 35°N within 7 days with a water vapour flux of at least 100 g kg\(^{-1}\) m s\(^{-1}\). Moreover, using also a Lagrangian approach, Sodemann and Zubler (2010) determined that the Western Mediterranean and the North Atlantic are typical moisture source regions for Alpine precipitation events in autumn (SON). In other seasons, also local or continental sources can be dominant (Sodemann et al., 2009; Grams et al., 2014). Using a water vapour tagging method, Winschall et al. (2012) have shown the importance of North Atlantic evaporation hot spots located on the western flank of PV streamers. HPEs in Central Europe are also frequently
associated with Warm Conveyor Belts (WCBs) (Pfahl et al., 2014). WCBs are coherent air streams ascending ahead of a cold front, in the warm sector of an extratropical cyclones (Browning et al., 1973; Madonna et al., 2014). In contrast to TMEs, they rarely originate from the Tropics. Grams et al. (2014) recently highlighted the role of equatorward ascending WCBs for large-scale flooding at the Alpine north side.

An other important aspect of HPEs are the mesoscale characteristics of the air mass, and in particular its stability. Medina and Houze (2003) have shown that two events with similar synoptic-scale conditions may produced very different precipitation patterns due to their differences on the mesoscale. This mesoscale influence has been investigated in several field experiments, such as the MAP experiment in the European Alps (Bougeault et al., 2001), or more recently the IMPROVE campaigns in the US (Stoelinga et al., 2003). Characterizing the flow upstream of the Lago Maggiore region during the MAP experiment, Houze et al. (2001) have shown that during blocked flow (low Froude number computed from the Milan sounding) precipitation is enhanced 140 km or more upstream of the Alps and that the low-level flow deviates cyclonically as it approaches the Alps. In contrast, high Froude number flows are lifted over the Alps and precipitation is strongly enhanced over the lower windward slopes. HPEs are more commonly occurring in an unblocked flow situation favoured by near-moist neutral stratification. The crucial role of near-moist neutrality has been discovered and discussed early (Sawyer, 1956; Smith, 1979) for orographic precipitation. However more recently studies have highlighted its importance for HPEs, for example during the MAP observation period (Rotunno and Houze, 2007) and more recently with dropsondes observations near the coastal mountains of California (Ralph et al., 2006), but also from an idealized modelling perspective (Miglietta and Rotunno, 2006). However due to the complexity of the interaction between flows and topography both the blocked and unblocked characteristics can happen in a single event. For example, Ferretti et al. (2000) and Rotunno and Ferretti (2001) have shown that horizontal moisture gradient can lead to unblocked flow in the moistest part of the flow and to blocked flow in the dryer part.

A consequence of the different mesoscale characteristics are different transformation processes of the condensates to precipitable hydrometeors. In the two
1.2. Orographic effects on precipitation

Figure 1.1.: Conceptual model of the orographic precipitation mechanisms active in (a) stable blocked, and (b) unstable unblocked flows. The diagrams show the types of hydrometeors present in each case, along with the behaviour of the flow. The dashed box in (b) indicates the position of the embedded convective rain shower. From Medina and Houze (2003), their figure 17.

Events described by Medina and Houze (2003) the observed microphysics differ significantly in both cases. In the blocked case, the air was forced to slowly ascend over low-level blocked air and hydrometeors were formed mostly by depositional growth of snow and subsequent melting (Figure 1.1a). In the unblocked case, the low-level air experienced strong orographic lifting above the first peaks allowing the formation of cloud water and graupel by riming, strongly enhancing the precipitation below (Figure 1.1b).

Another important microphysical effect on weather systems encountering topography is the seeder-feeder effect (Bergeron, 1965). This effect occurs when precipitating particles from an upper-level cloud (seeder) fall through a lower-level orographic cloud (feeder) collecting cloud water and thus producing enhanced precipitation by coalescence or accretion. More recent studies have pointed out that the separation between the seeder and feeder cloud is not clear (Stoelinga et al., 2013). For instance gravity waves may influence the seeding cloud resulting in a self-seeding cloud (Minder et al., 2008; Smith and Barstad, 2004).

Later studies (Houze and Medina, 2005) have shown that even during blocked flow situations small cells can develop, producing pockets of high liquid water concentration, which in turn enhance precipitation by aggregation, riming or co-
alescence (Figure 1.2). Other processes can produce small-scale cellularity (Stoelinga et al., 2013), such as embedded convective instability (Fuhrer and Schär, 2005; Kirschbaum and Smith, 2008).

On a finer spatial scale, the freezing level and the wind speed can have a strong impact on the precipitation distribution. Zängl (2007b) has shown that when the freezing level is above the mountain crest and the wind is strong enough, the precipitation can be greater in the downstream valley than at the mountain crest. Indeed graupel and snow are formed in the orographic cloud above the mountain crest and, under specific geometrical and wind conditions, are advected downstream. Then, they eventually fall through the melting layer increasing their vertical speed and producing enhanced precipitation in the valley. In addition, such microphysical effects are crucial for run-off prediction since they strongly affect the precipitation distribution.

Slight modification of the ambient flow can influence the distribution and amount of the orographic precipitation. Studying bands of precipitation in the Cévennes
Cosma et al. (2002) have shown the importance of the upwind conditions and slope steepness to determine the precipitation pattern. Their idealized sensitivity experiments showed that increased steepness increases the precipitation intensity, while increased humidity modifies the precipitation band structures. These effects are caused by modification of the flow dynamics in particular the flow splitting and the gravity wave amplitude.

Also based on idealized simulations, Colle (2004) examined, among other aspects, the role of the wind speed and the freezing level. For weak wind speeds ($< 10 \text{ ms}^{-1}$), precipitation intensity is a function of the windward slope. For stronger winds, the height and width of the barrier strongly affect the precipitation distribution, since narrow and low barriers have more precipitation advected to the lee. For this moderate wind conditions the height of the freezing level was shown to be critical for the precipitation distribution. A higher freezing level favours the formation of rain through warm processes which lead to larger amount of precipitation on the windward slopes due to the higher fall velocity of rain compared to snow.

Factors governing precipitation variability on small spatial scales have been investigated by Zängl (2007a) using semi-idealized simulations. Sensitivity experiments were performed using different wind speeds, wind directions and temperature profiles. He showed that the wind direction had the strongest impact on the small scale precipitation pattern, whereas the wind speed experiments had mainly an influence on the amplitude of the precipitation and the temperature experiments tended to increase the local precipitation maxima.

Another type of sensitivity experiments has been performed by Keil et al. (2008) where the specific humidity was modified in order to account for errors in the moisture assimilated by the models. They have performed experiments where the specific humidity was modified by $\pm 10\%$ and $\pm 30\%$ ($QV +10\%$ and $QV +30\%$) at different heights. They showed that increased moisture in the boundary layer had the strongest impact, a 10% increase in this layer was comparable to a 20% increase in the mid-troposphere. Comparing stratiform and convective regions, they showed that in the $QV +30\%$ experiment the precipitation response was persistent during the whole simulation period in the stratiform region, but confined to the first hours of the simulation in the convective regions.

Similarly Schlüter and Schädler (2010) examined the effect of temperature and
relative humidity changes for two floods events, one in summer 2002 (convective) and one in winter 2003 (stratiform). They found that the impact on the convective event was larger than on the stratiform event.

Using novel airborne measurements (Schäfler et al., 2011), Schäfler et al. (2010) have compared humidity measurement with analysis fields from the European Center for Medium-range Weather Forecasts (ECMWF) in two WCBs. They reveal that the model overestimated the moisture in the inflow region of a WCB by about 14% in average, suggesting some inaccuracies in the horizontal moisture advection in the ECMWF model for this particular event.

1.3. Outline

The aim of this thesis is to investigate the sensitivity of extreme precipitations events in Switzerland to modifications of some meteorological parameters. For this purpose, artificial specific humidity and temperature changes are examined (see subsection 2.3.1 for more details). Investigating the sensitivity to specific humidity changes is motivated by its role in orographic precipitation (see above) and by the uncertainty in its measurement (Keil et al., 2008; Schäfler et al., 2011). The temperature experiments are motivated by the reproducibility of a particular event in another (warmer) season and to provide some additional information on the effect of a warmer climate on extreme precipitations events (Schär et al., 1996; Frei et al., 1998). The selection of the cases studied (6 in total) was done in collaboration with Dr. Felix Naef, a senior researcher in hydrology from the Institute for Environmental Engineering (ETH Zürich), and is explained in greater details in subsection 2.1.1.

The outline of the thesis and a short introduction to each chapter is given below.

Analysis of the August 2005 extreme precipitation event  Due to the relatively high number of case studies and to the complexity of orographic precipitation, several diagnostic tools have been developed to extract the important information from the case studies. The aim of this first chapter is to explain the different choices made and to introduce each diagnostic. The August 2005 HPE was chosen since it has been investigated already in detail by the scientific community, allowing to base our investigation on previously published work. Moreover
due to its large impact and heavy damage, this event is also relevant for a broader public.

**Analysis of the August 2005 and October 2000 events**  This second chapter continues with the investigation of the August 2005 event through sensitivity experiments and in addition introduces the October 2000 event. In this chapter the complete set of diagnostics is presented.

**Relevant sensitivity aspects of four additional events**  This chapter, focusing primarily on the sensitivity experiments, presents the important aspects of four additional events in December 1991, September 1993, August 2007, and October 2011. Although the same diagnostics have been applied to all case studies, in this chapter only the most relevant ones are shown.

**Dynamics of a Local Alpine Flooding Event**  Focusing on the large-scale circulation and the moisture uptake of the October 2011 event, this chapter documents, for the first time, an important pathway for northern Alpine flooding, in which the interaction of synoptic-scale to large-scale weather systems and of long-range moisture transport from the Tropics are dominant. Moreover, the trapping of moisture in a subtropical cut-off near the West African coast is found to be a crucial precursor to the observed European high-impact weather.

**Climatology**  To put the events investigated in this thesis in a climatological perspective, this chapter presents a short overview of the large-scale circulation and moisture uptakes of 36 events spread over a 140-year period.

**Hydrological perspective and concluding remarks**  Finally the results presented in this thesis are summarized and discussed. The discussion focuses in particular on the precipitation response of individual catchments and on relevant aspects from a hydrological perspective.
Chapter 2.

Data, tools and methods

2.1. Dataset

2.1.1. The flood dataset

The selection of relevant floods investigated in this thesis have been selected from the dataset established by Schmocker-Fackel and Naef (2010). They selected floods affecting large areas of Switzerland in order to dispose of a nation wide time series of large scale floods events (Figure 2.1). All events strongly affected the society and caused considerable damage, therefore understanding their sensitivity to meteorological parameters is particularly important. The events are separated into three main regions that correspond approximately to three different climatological regimes, the northeast region, the northwest region, and the south region (see small maps in the lowest panel in Figure 2.1). The 40 selected flood events occurred during a 160-year period from July 1850 to August 2010, but only the flood events after 1871 are considered here (36 events) as they are contained in the Twentieth Century Dataset (20CR) (details in subsection 2.1.3).

For the sensitivity experiments, 6 flood events that occurred after 1979 were selected to be simulated with the COnsortium for Ssmall-scale MOdeling (COSMO) model (details in subsection 2.2.1). This period corresponds to the ECMWF Interim Re-Analysis dataset (1979-present) (ERA-Interim) (details in subsection 2.1.2) period, which provides initial and boundary conditions for COSMO of sufficient quality. Four events were selected from the Schmocker-Fackel and Naef (2010) dataset and two others because of particularly interesting hydrological or mete-
The complete list of events investigated in this thesis is the following table:

<table>
<thead>
<tr>
<th>Event</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>21-22.12.1991</td>
<td>The event affected the Suze catchment, located in the Jura mountain, producing its largest observed run-off. A flooding of the river could potentially lead to heavy damages in the city of Bienne (Horat and Naef, 1996; Pünntener, 2013).</td>
</tr>
<tr>
<td>21-24.09.1993</td>
<td>This event affected the southern side and in particular the Saltina catchment, located in the &quot;Alpes Valaisannes&quot; (FOEN, 1994).</td>
</tr>
<tr>
<td>13-15.10.2000</td>
<td>This event heavily affected the southern side of the Alps and in particular the village of Gondo and the Saltina catchment (FOEN, 2002).</td>
</tr>
<tr>
<td>21-23.08.2005</td>
<td>This event heavily affect the northern side of the Alps and in particular the Emmental region. It is the most costly event for Switzerland (FOEN, 2008).</td>
</tr>
<tr>
<td>07-08.08.2007</td>
<td>This event mainly affect the northern side of the Alps and in particular the Jura and the Plateau region (Bezzola and Wolfgang, 2009).</td>
</tr>
<tr>
<td>09-10.10.2011</td>
<td>This event mainly affect the northern side of the Alps and in particular the Jungfrau massif. The Kander and the Lütschine catchment produced floods (Piaget et al., 2014; Rössler et al., 2014).</td>
</tr>
</tbody>
</table>

### 2.1.2. ERA-Interim

Re-analysis datasets provide a full 3D representation of the state of the atmosphere over a large period of time. A reanalysis dataset is produced by a single model, to obtain consistency over time, assimilating different available observations (radiosoundings, aircraft measurement, surface observations, satellite observations, ...). The ERA-Interim dataset is provided by the ECMWF and covers the period from 1979 to present day, assimilating all available observations during this period. The model used has a spectral T255 horizontal resolution ($\sim 75$ km) and 60 model levels in the vertical up to 0.1 hPa (Dee et al., 2011). Data are available every 6
Figure 2.1: Spatial extent of large-scale flood events in Switzerland since 1850. According to the regions affected, the flood events were classified into the three types NE, NW, and S (maps in bottom row). Additionally, for all events after 1945, the wind direction over the Alps is given from the Alpine weather statistics of MeteoSwiss (L: changing wind direction). From Schmocker-Fackel and Naef (2010), their Figure 6.
hours (from 00 UTC).

2.1.3. The 20CR dataset

The Twentieth Century Dataset covers the period from 1871 to present day at 6-hourly temporal resolution (Compo et al., 2011), with 2° horizontal resolution and 24 pressure levels. The dataset is produced by assimilating only surface pressure observations. The quality of this dataset depends strongly on the number of assimilated stations and therefore increases towards more recent periods. It is interesting to note that a large number of station data are not yet digitalized and, in principle, could be used to increase the accuracy of such a dataset.

2.2. Tools

2.2.1. High-resolution numerical weather model COSMO

The model used in this thesis to perform the sensitivity experiments is the COrrsorium for Ssmall-scale MOdeling model (Doms and Schättler, 2002; Steppeler et al., 2003). The simulations are performed in hindcast mode nested in either ERA-Interim (nested twice) or COSMO analyses (MeteoSwiss product) depending on the time period. The final horizontal resolution of the simulations is 2.2 km (0.02°) and the simulation domain is shown in Figure 2.2b. The COSMO analyses are available with an hourly temporal resolution starting in 2003 and an horizontal resolution of 7 km, covering the domain shown in Figure 2.2a. The choice was made to use the best resource available for each event, and therefore events before 2005 are simulated with ERA-Interim as boundary conditions and events afterwards are simulated using COSMO analyses as boundary conditions. The ERA-Interim based simulations are first nested using COSMO at 7 km horizontal resolution on the same domain as the COSMO analyses and then nested to the final domain and resolution. The standard model setup of the Swiss weather service is used where the microphysical parametrization is a one-moment scheme with three ice categories. Only the shallow convection is parametrized with the Tiedtke convection scheme.

A detailed schematic of the cloud microphysical processes considered in the three ice-category scheme is shown in Figure 2.3. The most important processes
Figure 2.2: a) The large domain in grey shading indicates the COSMO 7 km domain and the small one the COSMO 2.2 km domain. The colour shading represent the COSMO 7 km topography, b) same as a) but for the COSMO 2.2 km domain and topography.
Figure 2.3: Cloud microphysical processes considered in the three ice-category scheme. From Reinhardt and Seifert (2005), their Figure 1.

for this thesis are riming of snow and graupel by cloud water, depositional growth of snow particles, and the growth of rain by collision and coalescence. Rimen is the collection of supercooled cloud water by snow or graupel, and occurs during strong updrafts when the flow is forced over a mountain range or when cellularity is formed in the flow since those processes lead to high liquid water concentrations above the melting level. In contrast, the depositional growth of snow particles occurs generally during weak ascent by the direct transfer of mass from vapour to ice, favoured by the lower equilibrium vapour pressure with respect to ice than with respect to water. The collision and coalescence is a warm rain microphysical process, where the cloud water droplets collide and coalesce forming larger droplets that eventually form rain droplets. For more details and references on those microphysical processes the reader is referred to Stoelinga et al. (2013).
2.2.2. LAGRANTO

In this work, several diagnostic are defined in a Lagrangian framework. Trajectories are calculated with the Lagrangian Analysis Tool (LAGRANTO) (Wernli and Davies, 1997; Sprenger and Wernli, 2015). The method consists of three steps:

- A kinematic evaluation of the position of the air parcels (trajectories) using a predictor-corrector procedure, based on the wind field.
- An interpolation of desired diagnostic variables along the trajectories.
- A selection of the trajectories based on spatial and physical properties of the diagnostic variables along the trajectories.

2.2.3. Moisture diagnostic

In this thesis, the origin of moisture contributing to the extreme precipitation events, and the role of the different weather systems in transporting moisture towards the Alps is determined with the help of a Lagrangian moisture source diagnostic (Sodemann et al., 2008). Trajectories are calculated using the LAGRANTO trajectory tool. This diagnostic considers the changes of specific humidity within 6-hourly trajectory time steps (Figure 2.4) and associates positive changes with moisture uptakes (i.e., points 1 and 3 in Figure 2.4). The sum of all moisture uptakes, weighted with their contribution to the precipitation in the target area (i.e., point 4 in Figure 2.4), then constitutes the overall moisture source distribution. This technique has already been applied to investigate HPEs in the Mediterranean (Winschall et al., 2014), of the Central European flood in June 2013 (Grams et al., 2014), and of the Pakistan flood in 2010 (Martius et al., 2013).

2.3. Method

2.3.1. Sensitivity experiments

Specific humidity changes

Specific humidity sensitivity experiments are performed by modifying the specific humidity field of the initial and the boundary conditions at every grid point,
Figure 2.4.: Sketch of the method for identifying uptakes along a backward trajectory of an air parcel on the way from the Atlantic ocean to Greenland (black line). Time before arrival is given at the top. \( q \) (dashed line), specific humidity in the air parcel (g kg\(^{-1}\)); \( \Delta q \), changes in specific humidity of an air parcel between two time intervals; BLH, boundary layer height. Thick blue sections along the trajectory denote sections of moisture increase, and red arrows are identified evaporation locations. From Sodemann et al. (2008), their Figure 1.

at all heights. The modification is according to

\[
q^*_{i,j,k} = q_{i,j,k} \cdot f
\]  

(2.1)

where \( q \) is the specific humidity and \( f \) is the modifying factor. In this thesis, three different specific humidity experiments are performed for each HPEs, one where \( f = 1.1 \) (+10%, hereafter referred as QV +10%), a second with \( f = 1.2 \) (hereafter referred as QV +20%), and one with \( f = 1.3 \) (hereafter referred as QV +30%).

In order to obtain physically meaningful modifications, increased specific humidity cannot overcome the saturation specific humidity. This is verified directly after the calculation of the new specific humidity.
2.3. Method

Temperature changes

Modifications of the temperature field are done similarly to the specific humidity changes for two experiments, a first one where the temperature is increased in the entire model domain by $+1$ K (hereafter referred as $T+1$K) and a second one with an increase of $2$ K (hereafter referred as $T+2$K). In addition, the surface temperature is also modified. The relative humidity is kept constant and therefore the specific humidity field is modified to fulfil this criteria.

Stationary boundary conditions

Stationary boundary conditions experiments are similar to the control run except that several hours after the initialization of the simulation the boundary conditions are frozen and remain constant afterwards. The idea is to simulate the effect of stationary large-scale conditions on precipitation. The "freezing point" corresponds to the time of the maximum moisture influx at the boundaries of the COSMO (2.2 km) domain. To determine this point backward trajectories are started above the catchment of interest at the time of maximum integrated water flux into the catchment. Then the time when trajectories leave the COSMO domain is chosen as the "freezing point".

2.3.2. Control box for sensitivity experiment

To investigate the effect of the sensitivity experiments on the precipitation distribution, the temporal evolution of precipitation and hydrometeors are examined in hydrologically relevant catchments and control boxes. The control boxes are defined at lower elevations and upstream of the catchments. They allow the direct comparison of two neighbouring areas with different topography. The location of a control box is determined by the mean wind direction in the lower troposphere.

2.3.3. Lagrangian perspective on flow-mountain interaction

The interaction of the flow with the Alps is investigated from a Lagrangian perspective. For this purpose, trajectories are started every hour in a circle (green dots, Figure 2.5) around the Alps and calculated forwards until the end of the
simulation. Then the number of trajectories crossing the Alpine crest is calculated for every hour of the simulation. The Alpine crest is defined as a straight line between 7°E, 46°N to 10°E, 47°N (red line in Figure 2.5). A crossing trajectory is only counted when at least one segment, one hour time step, overlaps with the Alpine crest.
Chapter 3.

Analysis of the August 2005 extreme precipitation event

To investigate the sensitivity of extreme precipitation with respect to changes in specific humidity or in temperature a few diagnostics are needed. Diagnostics that will help to determine the characteristics of an event such as its large-scale development; its moisture uptake regions; its capability of transporting large amount of moisture; its thermodynamic stability; and its capacity to flow over a certain mountain. The aim of this chapter is to explain the different choices made and introduce each diagnostic. The methodology described here will serve as framework to analyse all cases hereafter. Here, we will apply the diagnostics only to a single case study and only to the control run. The August 2005 HPE was chosen since it has been investigated by the scientific community, allowing to base your investigation on previously published work. Moreover due to its large impact and heavy damages, this event is also relevant for a broader public.

This chapter is organised as follows. After a brief introduction (section 3.1) about the impact of the event, looking at damages, rain and run-off measurement, an overview of the sensitivity experiments performed (section 3.2) serve as motivation for the diagnostics presented later. Then the synoptic evolution of the event is presented (section 3.3) and finally the diagnostics used to study the orographic precipitation are described (section 3.4).
3.1. Introduction

3.1.1. Historical perspective

During the 19-24 August 2005, Central Europe was hit by a wide spread HPE which mainly affected Switzerland, Germany and Austria. A few days before, this synoptic situation had caused heavy precipitation and floods in Romania as well. Due to the extreme weather, villages were cut off from the rest of the world, in all three countries the transport network was strongly affected since railways had to shut down for several month, and in total 49 persons were killed. In Switzerland alone, the floods caused damage up to 3 billions CHF, the most affected canton being Bern with damage up to 805 million CHF. The August 2005 event was the most damaging flood for private property since 1972, and comparable to the 1987 flood considering damage to public infrastructures (FOEN, 2008).

From a meteorological point of view the event was exceptional, many stations broke their record precipitation amount. For example, the Meiringen station received 205 mm of precipitation in 48 h, corresponding to a return period higher than 500 years and Engelberg received 190 mm during the same period, corresponding to a 100 year return period. To give an idea of the spatial extent of the heavy precipitation, 38% of the Swiss surface received more than 100 mm during the two day period 21-22 August 2005, while 3% received more than 200 mm (MeteoSchweiz, 2006).

From a hydrological point of view, existing peak values were exceeded at a total of 26 measurement stations, including sites where new records had already been established in 1999 (FOEN, 2008). Estimated return values of the run-off peaks (Figure 3.1) show that the most impacted regions were close to the Alpine main ridge. In the region of Interlaken and Engelberg many stations have exceeded a 100 year return period. For this event, two catchments are of special interest, the Lütschine and the Schächen catchments, since they were located near the precipitation maxima and since run-off station measured values above the 200 year return period. Moreover they are confluent to the Aare (Lütschine) and the Reuss (Schächen), two rivers that were heavily affected by the event.
An important aspect of such HPEs is the quality of the forecast and of the warning process. According to MeteoSchweiz (2006), the forecast from ECMWF of the large-scale flow was quite accurate. Only small differences were observed for the 500 hPa geopotential field at 12 UTC 21 August 2005 between the 72 hours forecast (base time: 12 UTC 18 August 2005) and the analysis. However the precipitation amount predicted by the ECMWF model was largely underestimated, but other models such as the GME model of the German weather service, predicted similarly low precipitation amounts. In contrast both the COSMO ensemble (COSMO-LEPS) and the deterministic run of COSMO produced good estimates of the HPE already at 12 UTC 19 August. The deterministic run from 12 UTC 19 August 2005 showed a clear signal of heavy precipitation for the 21 August (06 UTC to 06 UTC), and at 12 UTC 20 August 2005 the distribution of the precipitation was in good agreement with observations. However the first warning was given at midday on the 21 August for a danger of level 1. The post-event analysis of the decision showed that an earlier and stronger warning should have been sent. The decision was difficult to make because only one of the models predicted a high precipitation event.
Figure 3.2.: Observed precipitation based on RhiresD (MeteoSwiss, 2011) (a) and simulated precipitation from the control run (b). The accumulated precipitation covers the 06 UTC 21 to 06 UTC 23 August 2005 period.
3.1.2. Hindcast simulation

The here simulated period covers the most intense precipitation phase from 12 UTC 21 to 12 UTC 24 August 2005 following Hohenegger et al. (2008). To assess the quality of the COSMO hindcast simulation (setup described in subsection 2.2.1), we present here a comparison with the observed precipitation based on daily precipitation dataset in Switzerland from MeteoSwiss (RhiresD). Figure 3.2a shows the observed precipitation accumulated over two days from 06 UTC 21 to 06 UTC 23 August 2005. Peak values up to 240mm can be observed, particularly in mountainous region like the Titlis (in the centre of Switzerland), the Tödi (south of the Schächten catchment) and Jungfrau range (south of the Lütschine catchment). Simulated precipitation in Figure 3.2b present a very good agreement with the observed precipitation. The simulation show higher values near the peak locations (up to 300mm) and slightly smaller values elsewhere.

The vertical structure of the atmospheric flow has to be verified as well. This is done here by comparing the simulated flow approaching the Alps against soundings in Payerne and Munich. Results are shown in Figure 3.3 for 12 UTC 21 August 2005, shortly before the start of the first heavy precipitation phase. For both locations, the COSMO sounding (dashed lines) is more humid than the radio sounding (solid lines in Figure 3.3a), which probably explains the higher equivalent potential temperature ($\theta_e$) in the COSMO sounding (Figure 3.3b). The wind direction and speed are in very good agreement between the model and the radio sounding for both locations Payerne and Munich. The wind speed differences are lower than 3 m s$^{-1}$ in the lower troposphere below 700 hPa, the Alpine crest level. Based on the precipitation fields and the soundings we can conclude that COSMO could reasonably well simulate the HPE.

To investigate the time evolution of the precipitation we will focus on the two catchments introduced above (Lütschine and Schächten, Figure 3.4). Both catchments experienced two phases of heavy precipitation, one with a duration of 12 hours starting at 12 UTC 21 August 2005 and another with a duration of 18 hours starting at 06 UTC 22 August 2005, in between a 6 hour break. While during both phases each catchment received about 100 mm of rain, the Lütschine experienced less intense precipitation (maximum of 2.5 mm (15 min)$^{-1}$) compared to the Schächten catchment (maximum of 3.5 mm (15 min)$^{-1}$). In the Lütschine catch-
Figure 3.3.: Comparison of soundings from Payerne (red) and Munich (green) with COSMO control simulation (dashed lines) at 12 UTC 21 August 2005. a) Dew point (coloured lines) and temperature (black lines) for the sounding (solid) and COSMO (dashed), b) equivalent potential temperature, c) wind direction, d) wind speed. Panels b,c,d use the same line style (solid lines for sounding and dashed lines for COSMO).
ment some snow fall can be observed (blue bars, Figure 3.4a), but none in the Schächen catchment, which lies completely below 3000 m. Because of their relative similar behaviour we will only present results from the Lütschine catchment in the following sections.

Based on those observations, three important aspect for producing the event can be distinguished: Large precipitation values near the Alpine crest highlight the important role of the orography for the rain formation; a large amount of rain in 12 hours (100 mm) indicates an anomalously strong supply of moisture; and very intense rain fall (in mm (15min)$^{-1}$) is likely related to deep convection. In the next sections these different aspects will be further investigated.
Figure 3.4.: Precipitation evolution in the Lütschine catchment (a) and the Schächen catchment (b). The bar plots show the 15 min accumulated precipitation with rain in green, snow in blue and graupel in grey from 12 UTC 20 to 12 UTC 23 August 2005. The yellow and red lines show the accumulated precipitation in the catchment and over the whole Switzerland respectively. The dashed lines show the run-off measurement (in mm(15min)$^{-1}$) for each catchment at their official FOEN station.
3.2. Overview of the sensitivity simulations

As mentioned earlier the main purpose of this thesis is to estimate the sensitivity of precipitation to variation of some atmospheric parameters. Since the diagnostics used are principally thought to investigate the different response of the sensitivity experiments, we will give here a brief overview of the sensitivity experiments conducted to motivate the need for such diagnostics. Recall that we performed two sets of experiments, one where we increased the specific humidity, and another were we increased the temperature (subsection 2.3.1).

Figure 3.5 show the total precipitation differences between each sensitivity experiments and the control run. As expected, an increase of the specific humidity (Figures 3.5a,c,e) leads to an overall increase of the precipitation. In some location the maximum increase of precipitation is above 165 mm in 72 hours (for both the QV +20% and QV +30% experiments), whereas the minimum lies around −75 mm for all experiments. Remarkably the increase of precipitation is mainly located at lower elevations, the strongest differences can be observed near the 1000 m elevation line (green line in Figure 3.5). Moreover a higher increase of the specific humidity leads to a stronger shift towards low elevations. In contrast, close to the Alpine crest (above 2000 m elevation), a decrease of the precipitation can be observed.

Figures 3.5b,d show the results for an increase of the temperature. The pattern is less clear than in the specific humidity experiments, with both low and high elevation experiencing positive and negative changes of precipitation. The maximum increase is about +105 mm (for the T +2K simulation) and the maximum decrease about −75 mm for both experiments.

To obtain a more concise image of the precipitation changes due to the sensitivity experiments, we have integrated the precipitation over the Lütschine catchment and two control boxes for each simulation (see Figure 3.6a for their location). The methodology used to define the location of the boxes is described in subsection 2.3.2. In short, since the main changes are occurring at the foot of the 1000 m line both boxes are situated in this area upstream of the Lütschine catchment. Two boxes were needed due to the rotation of the wind direction during
Figure 3.5.: Total precipitation differences between each sensitivity experiments and the control run (colour shading) for the whole simulated period (12 UTC 20 to 12 UTC 23 August 2005), blueish colours indicate an increase of precipitation in the sensitivity experiments. The green and purple line indicate the 1000m and 2000m elevation contours, respectively (taken from COSMO topography at 2 km resolution). The grey shading shows area without precipitation in the CTRLrun.
3.2. Overview of the sensitivity simulations

Figure 3.6.: Integrated precipitation (mm) (from 12 UTC 20 to 12 UTC 23 August 2005) for all experiments in the Lütschine catchment and in two different control boxes. a) shows the location of the Lütschine catchment (salmon), the north-east control box (blue), and the north control box (green). The green and purple lines indicate the 1000 m and 2000 m elevation contours, respectively (taken from COSMO topography at 2 km resolution). b) and c) show the integrated precipitation (mm) in the Lütschine catchment and in the north and north-east boxes, respectively, for each experiments. The number above each bar gives the percentage of precipitation relative to the control simulation.
the precipitation event from north-easterly to northerly. Results are displayed in Figure 3.6. Surprisingly the total precipitation change shows a very different behaviour between the catchment and each control box (Figures 3.6b,c). An increase of the specific humidity by 30% leads to a 25% increase of the total precipitation in the Lütschine catchment, but to 191% increase in the north box (Figure 3.6b), and to 83% increase in the north-east box. The large discrepancy between the low and high elevation was already shown in Figure 3.5, but is highlighted even more strongly here. An other interesting aspect of these bar plots is the non linearity of the precipitation response to the increase of the specific humidity. For example in the Lütschine catchment, both a 10% and a 20% increase lead to about 6% increase of the precipitation. The changes in precipitation due to increasing temperature are slightly smaller, a 2 K increase leads to an increase of 40% and 15% in the north and north-east boxes respectively.

Based on the sensitivity experiments, some important question arise:

- What is the main factor explaining the shift of the precipitation towards lower elevation?
- Which mechanism were important for producing rain (seeder-feeder, embedded convection, orographic lifting/blocking)?
- Does the mechanism differ between specific humidity and temperature experiments?
- Is the precipitation response similar under different atmospheric conditions?
3.3. Synoptic evolution

This section describes the synoptic evolution which led to the HPE, based on three aspects of the large-scale flow using the ERA-Interim. The first aspect is the PV on an isentropic surface describing the large-scale flow circulation, the second aspect is the total column water (TCW) describing the moist air masses and finally the moisture uptake diagnostic determining regions where the precipitable water was collected and in the same time giving an integrated picture of the time evolution of the flow. All three aspect will be used to summarize the synoptic conditions of each case presented in this work.

At 12 UTC 20 August 2005 the synoptic situation was characterized by an upper level PV streamer, which extended from Norway to France (Figure 3.7a), and two high pressure system, one over the Azores and one centred over Belarus. 12 hours later the PV streamer had evolved into a cut-off low and was situated over France and the Alps (Figure 3.7c). Co-occurring with the formation of the cut-off low, a low pressure system developed beneath the cut-off low over northern Italy (a so called Genoa cyclone). The low pressure system then followed partly the pathway of a Vb cyclone (Bebber, 1890), which is known to be linked in some cases with heavy precipitation north of the Alps (Ulbrich et al., 2003), as they may bring moisture from the Mediterranean towards the northern side of the Alps. Over the next 48 hours, the cut-off remained quasi-stationary over the Alps (Figures 3.7b-f), producing a deep cyclonic circulation and establishing a north-easterly to northerly wind towards the Alpine north side.

The TCW field shows at 12 UTC 20 August 2005 (Figure 3.8a) two air masses below the cut-off with high values of moisture. One over the western Mediterranean with more than 49 kg m\(^{-2}\) and another extending from the Black Sea over the Balkan countries towards the Alps with similar values. At 00 UTC 22 August 2005 the two air masses seem to merged in the cyclonic circulation around the Alps.

To estimate the contribution of each air mass to the precipitation in Switzerland the moisture source diagnostic is used (details of the method in subsection 2.3.1). Between 1200 UTC and 00 UTC 21 August 2005, the main moisture uptakes happened south of the Alps (Figures 3.9a,b). But starting at 12 UTC 21 August 2005
Figure 3.7.: PV (in pvu) on the 330K isentropic surface and sea level pressure (black contour, 5 hPa interval) for 12 hourly time steps starting at 12 UTC 20 August 2005.
Figure 3.8.: total column water (in kg m$^{-2}$) (colour shading) and sea level pressure (black contour, 5 hPa interval) for 12 hourly time step starting at 12 UTC 20 August 2005.
Figure 3.9.: Moisture uptake in $10^{-3}$ mm day$^{-1}$ for the next 12-hourly accumulated precipitation over the $0^\circ$W, $0^\circ$N (lower left) to $60^\circ$W, $70^\circ$N (upper right) target region (details of the method in subsection 2.3.1).
3.3. Synoptic evolution

the main uptake region shifted to the eastern side of the Alps, over land. 12 hours later and for the next 36 hours strong moisture uptake occurred over land extending as far into the continent as 40°E. This diagnostic shows that the moisture available from the Balkan countries was much more important than the moisture from the Mediterranean. Therefore the low in the Golf of Genoa seems not to have played a very important role for transporting moisture to the northern side of the Alps, as usually stated for a Vb cyclone.

Finally to summarize this, Figure 3.10 shows the mean over the whole period (from 12 UTC 20 to 12 UTC 23 August 2005) for each of the variables presented above. The quasi-stationary cut-off low, centred over the Alps, is clearly depicted in Figure 3.10a, the two moist air masses are still represented in the mean TCW field (Figure 3.10b) and the moisture uptake pattern clearly shows the important region for surface evapotranspiration and subsequent transport. Therefore this compact representation will be used to illustrate the synoptic situation of each case, rather than the detailed evolution with 12-hourly time steps.
Figure 3.10.: Mean over the period 12 UTC 20 to 12 UTC 23 August 2005 of PV (a) (vertical average between 310K and 340K), TCW (b), and the evaporative moisture sources (c).
3.4. Process analysis

3.4.1. Horizontal moisture flux

As shown in chapter 1, orographic precipitation can be decomposed in three main elements, horizontal moisture flux, mountain interaction, and conversion from condensate to precipitation. In this section, we will examine the first element based on the control simulation and compare it with climatology.

Figure 3.11 shows the IWT (intensity and direction) for 12-hourly time steps (similar to section 3.3), where IWT is calculated as:

\[
IWT = \int_0^z \rho \cdot (q_c + q_v + q_s + q_g + q_i + q_r) \cdot v \, dz \quad [\text{kg m}^{-1} \text{s}^{-1}]
\]  

(3.1)

where \(q_c, q_v, q_s, q_g, q_i, \) and \(q_r\) are the mixing ratios of cloud water, water vapour, snow, graupel, ice and rain, respectively, \(\rho\) is the air density and \(v\) is the wind speed.

At 12 UTC 20 August 2005 (Figure 3.11a) low IWT values can be observed, since Switzerland was not yet under the influence of the north-easterly wind induced by the cut-off low. 12 hours later, at 00 UTC 21 August 2005, and for the next 48 hours very high values of IWT can be seen (Figures 3.11b-e) with values above 400 kg m\(^{-1}\) s\(^{-1}\) at every time step. A maximum is reached at 00 UTC 22 August 2005 (Figure 3.11d) with values exceeding 600 kg m\(^{-1}\) s\(^{-1}\). A strong gradient is located along the Alpine crest, with high values to the north. During this 48 hours period, the wind rotated from north-easterly to northerly as shown by the arrows in Figures 3.11b-e, co-occurring with the decrease of the IWT intensity. At the end of the precipitation event, after 00 UTC 23 August 2005, the IWT intensity strongly decreased.

To analyse the link between precipitation and IWT, the latter is calculated at three points, upstream of the Lütschine catchment and of the two control boxes, and compared with the accumulated precipitation in each area (Figure 3.12). During the first 24 hours, the increasing IWT led to the occurrence of weak precipitation. In the next 12 hours, from 12 UTC 21 to 00 UTC 22 August 2005, the precipitation increased dramatically in the north-east box and the Lütschine catchment (about +80 mm in 12 hours) as the IWT intensity doubled in the north-
Figure 3.11.: Integrated water transport (IWT) (kg m$^{-1}$ s$^{-1}$) for 12 hourly step starting at 12 UTC 20 August 2005 (colour shading) and IWT component (arrows) from COSMO control simulation.
Figure 3.12: IWT [kg m\(^{-1}\)s\(^{-1}\)] at three points situated upstream of the Lütschine catchment (salmon lines), the north-east control box (blue lines) and the north control box (green lines) (black dots in Figure 3.6a denote the exact location). The dotted lines show the accumulated precipitation in the Lütschine catchment and in each of the control boxes.

one possible explanation is the wind rotation observed in Figure 3.11 towards a northerly flow which is particularly favourable to form precipitation in the Lütschine catchment, with its north facing slopes of the mountain range. However a more detailed diagnosis is needed in order to understand the differences between the two precipitation episode (details in subsections 3.4.2 and 3.4.3).

To put the simulated values into a climatological perspective, the IWT values (only calculated with \(q_v\)) for each ERA-Interim time step are shown in Figure 3.13 for the the grid point 48°N, 9°E. This grid point was chosen to characterized north-easterly flow situations over Switzerland. This graphic shows clearly that
the IWT during the 2005 event was higher than typical values for this wind orientation. All time steps of the August 2005 event are close to the 280 kg m\(^{-1}\) s\(^{-1}\) limit, which encloses almost exclusively time steps "belonging" to a HPE. For the same season, many time steps show higher IWT values but none of them has a similarly intense meridional flow component.

### 3.4.2. Mountain interaction

The next element of orographic precipitation is the mountain itself, or the interaction of the flow with it (chapter 1). When the flow encounters a mountain it will, depending on its stability and speed, flow over (unblocked flow) or around it (blocked flow). The type of interaction, flow over or flow around, strongly affects the precipitation formation, being either closer to a stratiform (blocked) or to convective (un-blocked) process (Medina and Houze, 2003). Of course due to their complexity real cases of orographic precipitation present both flow over and flow around, but this characterization of the flow is still relevant for understanding the precipitation formation. In this section we will investigate the stability and the type of interaction of the flow using both traditional values and a more innovative Lagrangian method.
First, to investigate the stability and moisture content of the air mass, a pseudosounding upstream of the heavy precipitation region (Figure 3.3, Munich) is used. The profiles reveal an almost completely saturated air mass (Figure 3.3a, Munich) which was moist unstable from the surface up to 800 hPa (Figure 3.3b). A second layer moist neutrally stratified extended from 800 hPa up to 600 hPa, and above the profile was stable. The stability of the atmospheric flow upstream of each area and shortly before each precipitation episode is shown in Figure 3.14. Prior to the first precipitation episode, at 12 UTC 21 August 2005 (Figure 3.14a), the atmospheric flow is still characterized by a moist neutral layer above 850 hPa and a north-easterly wind upstream of all areas. At lower levels, both the north and north-east control boxes exhibit a stable layer with weak winds tending to a north-westerly, westerly flow. 24 hours later, at 12 UTC 22 August 2005 (Figure 3.14b), the situation is quite similar upstream of the Lütschine catchment (salmon line). However the lowest levels upstream of the control boxes shift to a slight unstable layer up to 850 hPa, and become clearly unstable above up to 650 hPa. The wind direction upstream of each area turned slightly to a more northerly flow.

Second, the interaction with the mountain range is studied. The method used for the Lagrangian perspective is presented in subsection 2.3.3, trajectories are started every hour in a circle (green dots, Figure 3.15) around the Alps and calculated forwards until the end of the simulation.

Results are shown in Figure 3.15 as the number of trajectories per grid point.
Chapter 3. Analysis of the August 2005 extreme precipitation event

Figure 3.15: Number of trajectories per grid points (a grid point is 1600 km$^2$) every 12 hours from 12 UTC 20 to 00 UTC 23 August 2005, normalized by the total number of trajectories. Trajectories are started at each green dot every time steps of the simulation and calculated forward till the end of the simulation. The blue contours show the 2000 m height contour. The times indicate the starting times of trajectories.
every 12 hours, from 00 UTC 21 to 12 UTC 23 August 2005, normalized by the total number of trajectories at each time step. At 00 UTC 21 August 2005 (Figure 3.15a), most of the trajectories are situated at low elevation, in particular for trajectories located in Switzerland north of the Alps. The number of trajectories per grid points close the Alpine crest is lower than 0.17. At 00 UTC 22 and 12 UTC 22 August 2005 (Figures 3.15c,d), during the first and second precipitation phase, the number of trajectories close to the Alpine crest is higher, above 0.25. In Figure 3.15c, a band of high values is visible on the southern side of the Alps, at approximately 8°E, showing the flow over produced by the north-easterly flow in the previous hours. Figures 3.15e,f show higher values near the extremity of the Alps since the northerly flow is established and the flow around increases. These plots can provide some instantaneous information about the flow over/around but they are not straightforward to interpret. Therefore in order to obtain a direct comparison with the hourly precipitation evolution, a condensed view of the flow over is produced. The number of trajectories crossing a simplified Alpine crest (red line in Figure 3.6) is calculated for each hour of the simulation. Results are displayed in Figure 3.16a for trajectories started below 2000 m. Two major peaks are visible, one around 18 UTC 21 August 2005 and one around 12 UTC 22 August 2005, corresponding remarkably well with the two major phase of precipitation in the Lütschine catchment (orange line in Figure 3.16a). For both peaks the number of trajectories crossing the Alpine crest is around 800, indicating that the majority of the trajectories started north of 47°N flowed over the Alpine crest (the number of started trajectories to the north of 47°N is 1170). Therefore the flow over seems to have been crucial in the precipitation formation during this event.

The classical method to estimate if an air mass approaching a mountain range may flow over or around is the Froude number

$$Fr = \frac{U}{H \cdot N},$$

(3.2)

where $U$ is the horizontal wind perpendicular to the mountain, $H$ the mountain height and $N$ the Brunt-Väisälä frequency. When the Froude number is smaller than 1, the flow is blocked, when it is larger than 1, the flow is unblocked. A major difficulty for applying this criteria to real case studies is the estimation of
Figure 3.16.: Capacity of the air mass to flow over the mountain range. a) show the number of trajectories crossing the "Alpine crest" (simplified as a line from (7°E, 46°N) to (10°E, 47°N), red line in Figure 3.6) for each hour of the control run, the orange line show the precipitation [mm (15min)^{-1}] in the Lütschine catchment. b) show the Froude number for the starting point of each trajectory north of 46°N for each hour of the control run. (median in red, 25 and 75 percentiles in blue shading, and whiskers for max. and min. values) c) show the moist Brunt-Väisälä frequency (blue dots) for the same points as in b) with the mean value for each time step (black dots), the solid black line show the mean wind speed in the layer. All diagnostics were calculated for the 0 to 2000m layer.
the different parameters. Should it be calculated at a single pressure height, should it be done for a layer? How should the height of the mountain be defined, by the first ridge, the maximum height or a typical height of a pass? Each of these parameters depend on the event itself, different wind directions and regions of interest would lead to different parameters. Therefore we have decided to use a Lagrangian perspective to determine the type of interaction between the air mass and the mountain range. However to demonstrate the added value of the Lagrangian perspective, both methods are presented for this case study.

To estimate the Froude number several assumptions were made. First, we calculated it only for the starting points of the trajectories (green dots in Figure 3.16) and only for the starting points north of 46° N due to the persistent northerly flow during the whole event. Second, we assumed that in order for an air mass to flow over the mountain range it needs to flow, at the minimum, above one of the main Alpine passes. Therefore we fixed at 2000 m the mountain height \( H \), since it is a typical height for the main passes of the Alps. Third, we decided to estimate the Froude number for a layer extending from the surface up to 2000 m. In consequence, \( N \) was calculated as follow:

\[
N = \sqrt{\frac{g}{\theta_s} \cdot \frac{\Delta \theta}{\Delta z}}
\]  

(3.3)

with \( \theta_s \) taken at the surface, \( \frac{\Delta \theta}{\Delta z} \) calculated between the surface and 2000 m, and \( U \) taken as the mean wind speed of the layer. However this method is not very accurate in the case of a saturated flow, since \( N \) does not reflect the effect of moisture in the stability of an air mass. The results are displayed in Figure 3.16b in the form of box plots representing each time step of the control run, showing that the air mass with respect to the "dry" Froude number is mainly stable (almost all points have a Froude number below 1). However the pattern is similar to the Lagrangian method pattern, only shifted by about -6 hours since the Froude number is calculated at the starting position of the trajectories and not at the Alpine crest as for the Lagrangian method.

To account for the humidity, the moist Brunt-Väisälä frequency should be used to determine the moist Froude number. This is done here using the following
definition for $N_m$ (Lalas and Einaudi, 1974; Durran and Klemp, 1982)

$$N_m^2 = \frac{g}{T} \left( \frac{dT}{dz} + \Gamma_m \right) \left( 1 + \frac{Lq_w}{RT} \right) - \frac{g}{1 + q_w} \frac{dq_w}{dz} \quad (3.4)$$

where $q_w$ is the total water mixing ratio. $L$ is the latent heat of evaporation and $R$ is the ideal gas constant for dry air. $\Gamma_m$ is the moist adiabatic lapse rate, which is given by

$$\Gamma_m = \Gamma_d (1 + q_w) \left( 1 + \frac{Lq_w}{RT} \right) \left( 1 + \frac{c_p q_v + c_w q_l}{c_p} + \frac{\epsilon L^2 q_w}{c_p RT} \left( 1 + \frac{q_w}{\epsilon} \right) \right)^{-1} \quad (3.5)$$

where $c_p$, $c_{pv}$ and $c_w$ are heat capacities of dry air, water vapour and liquid water under a constant pressure, and $\epsilon = 0.622$ is the ratio of the gas constants for dry air and water vapour. Following Jiang (2003), the total mixing ratio ($q_w$) and the liquid mixing ration ($q_l$) have been defined as follows

$$q_w = q_v + q_c + q_i + q_s + q_g \quad (3.6)$$
$$q_l = q_c + q_i + q_s + q_g \quad (3.7)$$

where $q_v$, $q_c$, $q_i$, $q_s$ and $q_g$ are the mixing ratios of vapour, cloud, ice, snow and graupel respectively. The calculation of $N_m$ using the same criteria as for $N$ shows that most of the layers are absolutely unstable (Figure 3.16c). Since the moist Froude number is based on the square root of $N_m$, it cannot be calculated for most of the layers, this is why Figure 3.16c displays $N_m$ together with the mean wind speed of each time step to mimic the Froude number. The results are again consistent with the Lagrangian and the "dry" perspective, since the two $N_m$ minima co-occur with two wind maxima at the same time as the peaks of the dry Froude number.

In conclusion, the Lagrangian perspective gives a more detailed perspective on the role of the mountain-flow interaction in producing precipitation. To obtain a relatively quick and concise overview of the flow characteristics, the "flow over" type of plot (Figure 3.16) will be the one used to describe the flows in the different simulations hereafter.
3.4.3. Precipitation formation

The last element is the conversion from the condensate to precipitable hydrometeors (chapter 1). In subsections 3.4.1 and 3.4.2 we have shown that the horizontal moisture flux and the characteristics of the flow (flow over versus flow around) can explain only part of the precipitation variability observed during the August 2005 event. The first episode was characterized by a high moisture flux directed almost perpendicular towards north-eastward facing slopes in a flow over situation. The second episode was characterized by a lower moisture flux directed towards northward facing slopes but also in a flow over situation. This section aims at understanding the different microphysical processes occurring during both episodes.

To this end, the hydrometeor distribution for the two precipitation episodes is examined with two cross-sections covering the Lütschine catchment and the two control boxes (red in Figure 3.6). Figure 3.17 shows 3-hourly averages over two 12-hour precipitation periods, starting at 12 UTC 21 August 2005 and 12 UTC 22 August 2005, for two cross-sections (one for the north control box and the other for the north-east control box). During the first episode, vertically extended large amounts of graupel and snow are visible (second panel in Figure 3.17a) during the 15-18 UTC 21 August 2005 period in the north-east cross-section, indicating that strong embedded convection took place during this first phase. Snow and graupel extend up to almost 300 hPa. 3 hours later (third panel in Figure 3.17a) several convective cells are visible throughout the cross-section as indicated by the peaks in graupel and snow concentrations. The snow layer spans now the entire troposphere. Weaker signals of embedded convection are visible in the north cross-section since during this first episode the wind direction was principally north-easterly, reducing the role of the north facing slopes. An other important aspect for this first episode was the existence of a large cold-phase cloud composed mainly of snow and ice (blue and cyan contours, Figures 3.17a,b) covering the whole area serving as seeder for the lower warm cloud (feeder).

For the second precipitation episode, very different processes are involved. A first major difference is the absence of the large cold-phase cloud (Figures 3.17(c,d)) which prevent the seeder-feeder effect. A second difference is the absence of graupel, indicating less favourable condition for embedded convection. Hydrometeors are mainly concentrated close to the highest elevations, indicating that the main process for this second phase is air mass transformation by pure orographic
Figure 3.17.: Cross-sections for the red boxes shown in Figure 3.6a, aligned on the north-east control box (a, c) and the north control box (b, d). Each small panel shows 3-hourly average fields over the width of the red box, each figure covering a 12-hourly period starting at 12 UTC 21 August 2005 (a, b) and 12 UTC 22 August 2005 (c, d). The coloured contours show various hydrometeors. Cyan stands for ice in $10\text{mgkg}^{-1}$ interval, blue for snow in $200\text{mgkg}^{-1}$ interval, grey for graupel in $100\text{mgkg}^{-1}$ interval, red for cloud water in $75\text{mgkg}^{-1}$ interval, and green for rain in $100\text{mgkg}^{-1}$ interval. The grey shading shows the averaged topography below the cross-section.

lifting. This is consistent with the precipitation evolution in both the control boxes and the Lütschine catchment (dotted lines in Figure 3.12).
3.5. Summary

In summary, the importance of the upper-level dynamics for transporting moist air masses from distant locations has been shown (section 3.3). On the mesoscale, the importance of the IWT intensity was shown to be regulated by the angle between topography and the impinging air, and the type of precipitation (subsection 3.4.1). In this case, due to its moist neutral stratification the flow was capable to flow over the topography (subsection 3.4.2). Two processes were essential for the formation of precipitation during this event (subsection 3.4.3), embedded convection and riming (seeder-feeder effect).
Chapter 4.

Sensitivity experiments for the August 2005 and the October 2000 events

4.1. The August 2005 event

4.1.1. Brief recapitulation

The impact, large-scale circulation and precipitation of the August 2005 event are described extensively in chapter 3, however we provide here a very brief recapitulation of the principal aspects, first on the synoptic evolution and then on the sensitivity experiments.

Synoptic evolution

The HPE mainly affected Switzerland, Germany and Austria with record precipitation amounts and run-off return period as well as major infrastructures damages (details in section 3.1). The upper-level flow was characterized by a quasi-stationary cut-off low over the Alps, inducing a cyclonic circulation down to the surface and triggering north-easterly to northerly flow against the Alps (section 3.3, Figure 3.10a). Extremely moist air masses situated initially over the Balkans (section 3.3, Figures 3.10b,c) were driven towards the Alps, where they were lifted producing heavy precipitation. On the regional scale (Switzerland),
the precipitation was separated in two episodes (section 3.1, Figure 3.4a). The first episode, starting at 12 UTC 21 August 2005 and lasting for 12 hours, was characterized by a strong IWT embedded in a north-easterly flow. The second episode, starting at 06 UTC 22 August 2005 and lasting for 18 hours, was characterized by a northerly flow with significantly weaker IWT.

Sensitivity overview

Two types of experiments have been performed, one where the specific humidity was increased at all levels and one where the temperature was increased. In order to analyse precisely the effect of the sensitivity experiments, control boxes were defined upstream of the Lütschine catchment (section 3.2, Figure 3.6a). Due to the rotation of the main wind direction during the event, two control boxes were needed, one to the north-east (for the north-easterly phase) and one to the north (for the northerly phase).

Specific humidity For the specific humidity experiments, simulations showed an overall increase of precipitation (Figures 3.5a,c,e). The strongest increase occurs at low elevations, especially close to the 1000 m elevation line. Surprisingly very weak increase or even a slight decrease can be observed at higher elevations. This behaviour is highlighted by the comparison of accumulated precipitation in the different experiments (Figures 3.6b,c). An increase of the specific humidity by 30% led to a +25% increase of the total precipitation in the Lütschine catchment, but to a +191% increase in the north box (Figure 3.6b), and to a +83% increase in the north-east box.

Temperature For the temperature experiments precipitation changes are weaker. Moreover no clear distinction can be made between low and high elevations, both experiencing increase and decrease. However the changes are consistent in the two control boxes, where increasing temperature leads to increasing precipitation (Figure 3.6). For example in the T+2K experiment, the north control box showed an increase by +48% and the north-east box a +17% increase of the total precipitation.
4.1. The August 2005 event

4.1.2. Process analysis

The following two main questions arising from the previous sections will be discussed here:

(a) Which process may explain the strong difference between the precipitation
response in the control boxes and in the Lütschine catchment, especially for the specific humidity experiments?

(b) Why are temperature experiments less sensitive than specific humidity experiments?

By looking more precisely at the time evolution of the precipitation changes in the different areas, we observe that changes in the Lütschine catchment are mainly happening during the second precipitation episode and more particularly during the last 12 hours of the simulations (Figure 4.1b). During the first episode, the only difference is a delayed start of the intense precipitation which grows with increasing moisture changes (see the 12 UTC 21 to 12 UTC 22 August 2005 period in Figure 4.1b), and eventually leads to a merging of the two precipitation episodes. For the north-east control box the main changes occur during two different periods. Firstly prior to the first precipitation episode, particularly during the 12 UTC 20 to 00 UTC 21 August 2005 period (Figure 4.1d), the QV +20% experiment produces +55 mm more precipitation than the CTRL experiment and similar changes can be observed in all experiments. Secondly during the second precipitation episode, and only for the specific humidity experiments, large precipitation increases occur during the last 18 hours of the simulations, about +70 mm in the QV +20% and QV +30% experiments. In contrary precipitation changes in the north control box only occur during the first precipitation episode (Figure 4.1f), when the precipitation increases by up to +100 mm in 36 hours in the QV +30% experiment. The subsequent analysis will focus on these specific episodes.

**Horizontal moisture flux**

To investigate the reason for the changes in precipitation, we will first look at the changes in the horizontal moisture flux. Upstream of the Lütschine catchment the IWT exhibits decreasing peak values with increasing specific humidity during the first precipitation episode (see peak values at 18 UTC 21 August 2005, blue lines Figure 4.1a), which explains the delay of the first precipitation episode. In contrast, the IWT during the second precipitation episode is slightly increasing with increasing specific humidity, but more interestingly remains high over a longer period. At 00 UTC 23 August 2005 the IWT is about 150 kg m\(^{-1}\) s\(^{-1}\) in the
control run and about 200 kg m\(^{-1}\)s\(^{-1}\) in the QV +30\% experiment, whereas both experiments have values of about 260 kg m\(^{-1}\)s\(^{-1}\) 12 hours earlier. Therefore the duration rather than the intensity of high IWT period is important. This also visible in the north control box (Figure 4.1e), during the period from 00 UTC 21 to 00 UTC 22 August 2005 when precipitation changes are the largest, especially for the QV +20\% and QV +30\%. During this period, the IWT remains high over a longer period in the QV +20\% and QV +30\% experiments than in the QV +10\%, T +1K and T +2K experiments, which exhibit however higher peak values.

Interestingly the strong IWT increase in the T +1K and T +2K experiments does not have a strong impact on the precipitation, especially for the north control box (yellow and red lines in Figure 4.1f). For example the IWT peak value increases by +300 kg m\(^{-1}\)s\(^{-1}\) upstream of the north control box during the 12 UTC 21 to 00 UTC 22 August 2005 period, whereas the precipitation increases by only +25 mm.

**Large-scale flow and mountain interaction**

The second aspect of orographic precipitation, flow-mountain interaction, shows less significant changes. Nevertheless, an interesting aspect of the flow index is its good correlation with the precipitation in the Lütschine (orange lines in Figure 4.2). In all experiments, increasing flow over leads to increasing precipitation in the Lütschine catchment. However, the main point is the disappearance of the gap between the two precipitation episode in the specific humidity experiments (Figures 4.2a-d). The QV +10\% experiment shows an increase of the flow over during the gap period, but the gap is still present with weak precipitation during this period. On the other hand, the QV +20\% and QV +30\% experiment show a complete disappearance of the gap linked with the merge of the two precipitation episodes. The temperature experiments however show a pattern very similar to the CTRL run.

**Precipitation formation**

As mentioned in subsection 4.1.2, a strong precipitation increase can be observed with increasing humidity in the north control box during the first precipitation episode (Figure 4.1). In order to investigate in more detail the precipitation formation due to the increasing moisture flux during this period, cross-sections along
Figure 4.2.: Capacity of the air mass to flow over the mountain, i.e. the number of trajectories crossing the "Alpine crest" (simplified as a line from (7°E, 46°N) to (10°E, 47°N), red line in Figure 3.6) for each hour of the control run, the orange line show the precipitation [mm (15min)^{-1}] in the Lütschine catchment. Each panel shows a different experiment, CTRL in (a), QV+10% in (b), QV+20% in (c), QV+30% in (d), T+1K in (e), and T+2K in (f).
the north control box are shown in Figure 4.3.

All experiments show a saturated lower troposphere (large cloud water content, red contours in Figure 4.3), however in the QV +20% and the QV +30% experiments the low-level cloud does not extend much further than 63 km, whereas in all the other experiments it extends further than 84 km. This is probably explained by the more intense convective cells, marked by the large snow and graupel concentrations (blue and grey contours in Figures 4.1c,d), which are very efficient in removing moisture from the air column. This strong precipitation efficiency explains the decrease of moisture influx upstream of the Lütschine catchment mentioned earlier. Interestingly the overall moisture content is increased in the temperature experiments but the stability of the atmosphere remains comparable to the CTRL run. Figures 4.3e,f do not exhibit significantly stronger convective cells and the graupel concentration is comparable to the CTRL run. This could explain the weaker precipitation changes in the temperature experiment compared to the specific humidity experiments.

Another important period is the 1200 UTC 22 to 00 UTC 23 August 2005 interval during which the north-east control box exhibits increasing precipitation for increasing specific humidity but shows no changes in the temperature experiments. To investigate the differences between the temperature and the specific humidity experiments, cross-sections aligned on the north-east control box are shown (Figure 4.4). In the CTRL run, the precipitation in the north-east control box is controlled by the presence of snow layers above the water cloud during the first two 3-hour period (first two panels in Figure 4.4a). The riming of snow by cloud water is therefore important for the formation of precipitation. During the last two 3-hour period, the snow layers vanish, preventing rain formation in the north-east control box. By contrast, during the specific experiments QV +20% and QV +30% (Figures 4.4c,d), the snow layers are still present during the last two 3-hour period, allowing rain formation in the north-east control box. Conversely during the temperature experiments (Figures 4.4e,f), the snow layers are only present during the first two 3-hour period. The increased IWT compared to the CTRL run is due to an increase of cloud water and to a local increase of graupel. The increase of graupel is principally located northward of 75 km, see in particular the third panel of Figure 4.3. This increase is reflected in the weak increase of precipitation in the Lütschine catchment in the T+2K experiment.
Figure 4.3.: Cross-sections along the north control box, for all experiments. 3-hourly means are shown for consecutive periods starting at 00, 03, 06, 09 UTC 21 August 2005. The coloured contours show various hydrometeors. Cyan stands for ice in $10\text{mgkg}^{-1}$ intervals, blue for snow in $200\text{mgkg}^{-1}$ intervals, grey for graupel in $100\text{mgkg}^{-1}$ intervals, red for cloud water in $75\text{mgkg}^{-1}$ intervals, and green for rain in $100\text{mgkg}^{-1}$ intervals. The grey shading shows the averaged topography.
Figure 4.4: Cross-sections along the north control box, for all experiments. 3-hourly means are shown for consecutive periods starting at 12, 15, 18, 22 UTC 22 August 2005. The coloured contours show various hydrometeors. Cyan stands for ice in 10mg kg$^{-1}$ intervals, blue for snow in 200mg kg$^{-1}$ intervals, grey for graupel in 100mg kg$^{-1}$ intervals, red for cloud water in 75mg kg$^{-1}$ intervals, and green for rain in 100mg kg$^{-1}$ intervals. The grey shading shows the averaged topography.
during this period (Figure 4.1b).

4.1.3. Summary

- The impact of the IWT is stronger on precipitation when the duration of high IWT period is increased rather than when the peak values are increased.

- The static stability does not explain the precipitation changes, but is essential to describe the type and timing of the precipitation.

- A first microphysical effect is that efficient removal of moisture by graupel in convective cells reduces the moisture available downstream, favouring precipitation at low elevations.

- Another microphysical effect is that increased snow layers during specific experiments increases precipitation by riming.

- Increased temperatures lead to slightly enhanced convective cells, but the locations of the precipitation remain usually similar to the CTRL run.
4.2. The October 2000 event

4.2.1. Synoptic evolution

The large-scale situation during the October 2000 event is typical for an autumn extreme precipitation event in Switzerland. Such events are characterized by an upper-level PV streamer that advects moist air towards the southern slopes of the Alps where it is orographically lifted producing heavy precipitation (Massacand et al., 1998; Martius et al., 2006). In this case, the PV streamer was first located over western Europe, extending from Norway to Morocco at 00 UTC 13 October 2000. Slowly moving eastward, it then evolved into a cut-off low centred over the west coast of Spain at 12 UTC 14 October 2000. The cut-off low remained then quasi-stationary over this area until 00 UTC 16 October 2000 (Figure 4.5a), producing a southerly to south-easterly wind (due to the rotation of the cut-off low) towards the Alpine south side even at the lowest levels. Driven by the upper-level circulation, warm humid air from Northern Africa (Figure 4.5b) was advected along the eastern flank of the PV streamer. Due to the quasi-stationary cut-off low, the advection of moist air was important over several days helping to produce the very large precipitation amount.

The moisture source diagnostics shows to principal uptake regions (Figure 4.5c). The main one is located in the Mediterranean south of Italy, and another is located over the North-Atlantic. The Atlantic source region may indicate the presence of Atlantic hot spots, i.e, areas of strong evaporation upstream of the PV streamer introduced above. This mechanism has been described by Winschall et al. (2012), who have shown that southern Alpine HPEs are frequently preceded by intense North Atlantic evaporation.

The precipitation observations show that the accumulated precipitation was exceptionally large (Figure 4.6a), for example the Simplon region, enclosing the Saltina catchment, experienced more than 480 mm in 48 hours.

For this event, the COSMO control run produced less precipitation than observed with maxima below 480 mm in 48 hours. However the spatial distribution is well represented (Figure 4.6b), for example the Simplon region shows a maximum in the CTRL as well.

Two catchments are especially interesting for this event. Both the Lütschine, even if situated on the northern side of the Alps, and the Saltina catchments ex-
Chapter 4. Analysis of the August 2005 and October 2000 events

Figure 4.5: Mean over the period 00 UTC 13 to 00 UTC 16 October 2000 of PV (a) (vertical average between 310K and 340K), TCW (b), and the evaporative moisture sources (c).

experienced the highest measured run-off during this event. The time evolution of the precipitation in both catchments (Figure 4.7) shows first a relatively weak and continuous precipitation period, with intensities below 1 mm (15min)$^{-1}$ in the Lütschine catchment and slightly below 2 mm (15min)$^{-1}$ in the Saltina catchment. A 12-hour period, centred on 06 UTC 15 October 2000, with intense precipitation occurred in both catchments, but the Lütschine experienced much weaker precipitation intensity (about 3 mm (15min)$^{-1}$) than the Saltina catchment (about 5 mm (15min)$^{-1}$). However the run-off measurements for the Saltina catchment indicate a second earlier peak at 06 UTC 14 October 2000, which is missed by the
4.2. The October 2000 event

Figure 4.6: Observed precipitation based on RhiresD (MeteoSwiss, 2011) (a) and simulated precipitation from the control run (b). The accumulated precipitation covers the 06 UTC 13 to 06 UTC 15 October 2000 period.

COSMO simulation, explaining the lower accumulated precipitation shown in Figure 4.6.

4.2.2. Sensitivity overview

To investigate the differences between the sensitivity experiments, two control boxes have been defined in addition to the catchments previously described. The two control boxes should account for the rotation of the main wind direction. Therefore one is situated south of the Saltina catchment (blue area in Figure 4.8a, hereafter referred as the south control box) and the other is situated to the south-east of the Saltina catchment (green area in Figure 4.8a, hereafter referred as the south-east control box).

The comparison of the total accumulated precipitation between the different simulations (Figure 4.8) shows a variable response of the sensitivity experiments. The Lütschine catchment experiences the strongest total precipitation changes during the QV +10% experiment (salmon bars in Figure 4.8) with an increase of +29%. In contrast, the QV +30% experiment produces only a slight increase of +9%. The Lütschine catchment experiences also the strongest total precipitation changes in the temperature experiments, compared to the other areas, with +25% for the T +2K experiment.

Interestingly, the Saltina catchment shows a different behaviour, even if located at a similar elevation as the Lütschine catchment and situated upstream of it. In-
Figure 4.7.: Precipitation evolution in the Lütschine (a) and the Saltina (b) catchments. The bar plots show the 15 min precipitation with rain in green, snow in blue and graupel in grey. The yellow and red lines show the accumulated precipitation in the catchment and over the whole Switzerland respectively. The dashed line shows the run-off measurement (in mm(15min)$^{-1}$) for each catchment at their official Swiss Federal Office for the Environment (FOEN) station.
4.2. The October 2000 event
deed only the QV +10% experiment produces a very weak increase of the total precipitation (aquamarine bars in Figure 4.8b) of only +3%. Both the QV +20% and QV +30% experiments produce a decrease of the total precipitation of −8% and −21%, respectively. In parallel the temperature experiments did not change the total precipitation in this catchment.

The control boxes situated upstream of both catchments show a very different response to the sensitivity experiments. The south control box shows little total precipitation changes for all sensitivity experiments, the largest increase is produced by the QV +10% and QV +20% experiments with an increase of +17%. On the other hand, the south-east control box shows a strong increase of total precipitation with increasing specific humidity. The QV +10% experiment shows an increase of +29%, and both the QV +20% and QV +30% experiments produce an increase of +58%, the largest increase for the areas considered here. The south-east control box reacts as well to the temperature changes with a maximal increase of +12% for the T +2K experiment.

As for the August 2005 event, the temperature experiments exhibit a weaker, more linear response of the total precipitation compared to the specific humidity changes.

4.2.3. Process analysis

The main questions, emerging from the previous short analysis, to be investigated now in greater detail, are:

- When and why does the precipitation decrease in the Saltina catchment with increasing specific humidity?

- Are the precipitation differences occurring during the same period in all experiments?

- Why is the south-east box reacting so strongly compared to the others areas?

By looking more precisely at the time series of precipitation (right column in Figure 4.9) in the different areas, we can determine the interesting periods when precipitation changes occur between the different experiments. For the Lütschine catchment (Figure 4.9b) the major changes occur during the 12-hour period from 00 UTC 15 to 12 UTC 15 October 2000 for all types of experiments. The difference
between the QV +10% and QV +20%/QV +30% experiments occurs during this period.

The slight increase of the total precipitation in the Saltina catchment (Figure 4.9d) during the QV +10% experiment also happens during this 12-hour period (about +10 mm compared to the CTRL run). In contrast the QV +30% experiment shows lower precipitation intensity during the whole simulation. The temperature experiments show only weak temporal variations of precipitation that compensate themselves.

The south control box (Figure 4.9f) experiences the strongest precipitation changes during short periods around 00 UTC 14 and 00 UTC 15 October 2000.
for the specific humidity experiments. The weak precipitation changes during the temperature experiments are not clearly observable.

The south-east control box (Figure 4.9h) experiences the strongest precipitation changes during the 18-hour period from 12 UTC 13 to 06 UTC 14 October 2000 (+70 mm in the QV +30% compared to the CTRL), and during the 12-hour period from 00 UTC to 12 UTC 15 October 2000 (+80 mm in the QV +30% compared to the CTRL). The precipitation changes in the temperature experiments are confined to the last 12 hours of the simulations.

**Horizontal moisture flux**

To investigate the reason for the changes in precipitation, we first investigate the horizontal moisture flux changes (left column in Figure 4.9). All experiments and all areas exhibit their largest IWT increase during the precipitation maximum around 06 UTC 15 October 2000, with a typical increase of about +100—400 kg m$^{-1}$ s$^{-1}$. This is in good agreement with the changes in precipitation presented above that mainly occur during this 12-hour period from 00 UTC 15 to 12 UTC 15 October 2000. A good example is the south-east control box (Figures 4.9g,h), where the IWT increase is the strongest, in particular in the QV +20% and QV +30% experiments, and associated with a precipitation increase. Surprisingly during the first 36 hours only the control boxes show a significant increase of the IWT, but the increase is not directly linked with the precipitation changes during this period. For example during the 12-hour period from 12 UTC 13 to 00 UTC 14 October 2000 in the south-east box (Figures 4.9g,h), the weaker increase of IWT in the QV +20% compared to the QV +30% experiment leads to a stronger precipitation increase. Furthermore despite the similar changes in IWT in the south and south-east control boxes during this period, only the south-east control box experienced precipitation increase. The IWT upstream of the Saltina catchment (Figure 4.9c) is characterized by a very weak increase during the first 48 hours. But all experiments produce a IWT peak at about 00 UTC 14 October 2000, which is rapidly decreasing afterwards returning to levels comparable to the CTRL without, however, producing significant precipitation changes. In conclusion, IWT is only partially helpful to understand the precipitation changes. High values over a relatively long period are almost always linked to precipitation, however IWT peaks are not systematically linked to precipitation peaks.
Chapter 4. Analysis of the August 2005 and October 2000 events

Figure 4.9: IWT [kg m$^{-1}$ s$^{-1}$] (left panels) and accumulated precipitation [mm] (right panels) for the Lütschine catchment (a,b), the Saltina catchment (c,d), the south control box (e,f) and the south-east control box (g,h). IWT is calculated upstream of each area (black dots in Figure 4.8a denote the exact location). The accumulated precipitation is a mean over each area. The coloured lines represent the different experiments, CTRL in black, QV +10% in light blue, QV +20% in blue, QV +30% in dark blue, T +1 K in yellow, and T +2 K in red.
Large-scale flow and mountain interaction

The second aspect investigated to analyse the precipitation changes is the mountain-flow interaction. For this case, less than 1/4 of the started trajectories (∼ 800 at each time step) can potentially flow over the Alpine crest (red line in Figure 4.8), due to the southerly flow situation and the shape of the Alps. Figure 4.10 shows that until 12 UTC 14 October 2000 only about 400 trajectories are crossing the Alps. There is then a slight increase of flow over afterwards, especially from 09 UTC 15 October 2000. However the mountain-flow interaction seems not to be affected by the sensitivity experiments, the general picture is similar in all experiments. Only the specific humidity experiments show a slight increase of flow over trajectories.

Precipitation formation

As mention in subsection 4.2.2 the Saltina experienced a decrease of total precipitation with increasing specific humidity and subsection 4.2.3 has shown that the decrease occurs at all time steps. As both the IWT and mountain-flow interaction could not explain this decrease, the transformation from condensate to precipitation hydrometeors is probably the key factor. Figure 4.11 displays a cross-section through the Saltina catchment (westernmost red box in Figure 4.8) for the 12-hour period from 00 UTC to 12 UTC 14 October 2000. It shows that increasing specific humidity leads to a strong increase of the snow content, especially at high elevations during the first two 3-hourly periods (blue contours in Figures 4.11b-d). Likewise increasing specific humidity leads to an increasing convective activity, especially in the QV +20% and QV +30% experiments, as shown by the increasing graupel concentration (grey contours in Figures 4.11c,d). The strongest increase can be seen during the first 3 hours of the QV +30% experiment. The enhanced embedded convection is linked with a very efficient moisture removal that leads to more intense rainfall during this period. The precipitation increases due to the convective cells are also visible in the time evolution of the precipitation in the Saltina and the South box (Figures 4.9d,f) shortly after 00 UTC 14 October 2000. This is clearly visible in the two 3-hourly periods following the intense convection in QV +20% (lower panels in Figure 4.11c), when the snow content is yet lower than the snow content in the CTRL. The QV +30% experiment even expe-
Figure 4.10.: Capacity of the air mass to flow over the mountain range. The blue bars show the number of trajectories crossing the "Alpine crest" (simplified as a line from (7°E, 46°N) to (10°E, 47°N), red line in Figure 4.8) for each hour of the control run, the orange line show the precipitation [mm (15min)^{-1}] in the Saltina catchment. Each panel represents a different experiment, the control run in a), the specific humidity experiments in b-d), and the temperature experiments in e,f).
4.2. The October 2000 event

Figure 4.11.: Cross-section aligned on the south control box (blue bars) and the Sältina catchment (red bars) for a 12-hour period from 00 UTC 14 October 2000 to 12 UTC 14 October 2000. Each small panel shows 3-hourly average fields over the width of the red box. The coloured contours show various hydrometeors. Cyan stands for ice in 10 mg kg$^{-1}$ intervals, blue for snow in 200 mg kg$^{-1}$ intervals, grey for graupel in 100 mg kg$^{-1}$ intervals, red for cloud water in 75 mg kg$^{-1}$ intervals, and green for rain in 100 mg kg$^{-1}$ intervals. The grey shading shows the averaged topography. Each panel represents a different experiment, the control run in a), the specific humidity experiments in b-d), and the temperature experiments in e,f).
riences stronger embedded convection leading to a faster removal of the moisture as demonstrated by the earlier depletion of snow content already in the second 3-hourly period (upper right panel in Figure 4.11d). The strong removal of moisture is also visible in the IWT flux upstream of the Saltina catchment and the south control box (aquamarine and green solid lines in Figure 4.9). The IWT upstream of the Saltina peaks at 600 kg m\(^{-1}\) s\(^{-1}\) shortly after 00 UTC 14 October 2000 in the QV +30% experiment (+100 kg m\(^{-1}\) s\(^{-1}\) compared to the CTRL), but rapidly decreases to about 300 kg m\(^{-1}\) s\(^{-1}\) three hours later, a level similar to the CTRL. As a consequence to the reduction of the snow content, the precipitation is strongly reduced, especially at high elevation. This can be seen in Figures 4.11c,d (lower panels) with no snow at locations further than 52 km (the approximate location of the Saltina catchment). In summary, the increasing efficiency increases precipitation intensity during a short period and particularly at the lower elevations. As a consequence, the Saltina catchment experiences a decrease of the total precipitation, due to its high elevation.

To explain the differences observed in the Lütschine catchment described in sections 4.2.2 and 4.2.3, a cross-section covering the Lütschine catchment and the south-east control box is displayed in Figure 4.12 for the 12-hour period from 00 UTC to 12 UTC 15 October 2000. It shows that precipitation in the Lütschine catchment is caused by two processes, growth of snow by agglomeration (seeder-feeder effect) and the melting of snow downstream of the Jungfraujoch mountain range (the high mountain between 18 km and 36 km in Figure 4.12). The agglomeration of snow particles occurs in the feeder cloud (red contours in Figure 4.12) located over the orography. Then due to the strong wind over the Alpine crest, the light snow hydrometeors are advected further downstream over the Lütschine catchment. As the particles sediment downstream of the Jungfraujoch mountain range, they melt producing rain in the Lütschine catchment. This process has also been observed in other studies (Zängl, 2007b). Due to increasing snow content with increasing specific humidity, the effect of these two processes is enhanced producing more intense rain fall in the Lütschine catchment. The maximum is reached during the QV +10% experiment that shows the highest snow content.

In a similar manner, increasing snow and particularly graupel concentrations (blue and grey contours in Figure 4.12) lead to increasing precipitation efficiency
4.2. The October 2000 event

Figure 4.12: Cross-section aligned on the south-east control box (green bars) and the Lütschine catchment (salmon bars) for a 12-hour period from 00 UTC 15 October 2000 to 12 UTC 15 October 2000. Each small panel shows 3-hourly average fields over the width of the red box. The coloured contours show various hydrometeors. Cyan stands for ice in 20 mg kg$^{-1}$ intervals, blue for snow in 600 mg kg$^{-1}$ intervals, grey for graupel in 75 mg kg$^{-1}$ intervals, red for cloud water in 150 mg kg$^{-1}$ intervals, and green for rain in 100 mg kg$^{-1}$ intervals. The grey shading shows the averaged topography. Each panel represents a different experiment, the control run in a), the specific humidity experiments in b-d), and the temperature experiments in e,f).
above the south-east box (green bars). This becomes clear when comparing the graupel concentrations in the CTRL run and the specific humidity experiments in the first two 3-hourly panels. The graupel concentration in the CTRL run is below 75 mg kg$^{-1}$, but for example in the QV +20% experiment, the graupel concentration is above 300 mg kg$^{-1}$. This increased graupel concentration is produced during the whole period shown in Figure 4.12 and favours long-lasting strong precipitation over the south-east control box.

4.2.4. Summary

The most important aspects revealed by this detailed analysis of the October 2000 event are:

- Increase of the graupel concentration through riming is a key factor for increasing the precipitation efficiency. When the increase occurs above the low elevations it increases the precipitation (in our cases the south-east control box) and reduce the available moisture downstream (in our case the high elevations areas such as the Saltina catchment).

- In parallel an increase of snow concentrations leads to an increase of the advection of snow towards the lee slopes increasing precipitation there, in our case in the Lütschine catchment.

4.3. Summary

All experiments produced a larger precipitation response for longer high IWT periods than for shorter period with significantly higher IWT peak values. In both cases, the stability was crucial for determining the type of precipitation but was not significantly influenced by the sensitivity experiments, neither in the specific humidity experiments nor in the temperature experiments. However, for both experiments stronger embedded convection above the first peaks was a key factor in explaining the changes in the precipitation pattern. Several studies have already observed this effect (Houze et al., 2001; Medina and Houze, 2003). This effect is to reduce the precipitation at high elevations due to more efficient moisture removal above the first peaks. The increased snow content is a systematic
response in all different experiments, although particularly strong for specific humidity experiments. This increase led to increasing precipitation by riming. In August 2005, the riming occurred rather at high elevations during the CTRL, and increased above the lower elevations with increasing specific humidity. In October 2000 however, riming occurred above all elevations in the CTRL, and was intensified above the lower elevations in the specific humidity experiments.
Chapter 5.

Relevant sensitivity aspects of four additional events

5.1. The December 1991 event

5.1.1. Synoptic evolution

The synoptic situation during the December 1991 event was characterized by an upper-level ridge over the North-Atlantic and a strong high pressure system centred over the Azores. Downstream of the ridge a quasi-stationary PV streamer was extending from Denmark to North Africa (Figure 5.1a). On its western flank the PV streamer produced northwesterly winds towards the Alps. Moisture transport from the Atlantic was induced by the ridge-anticyclone system as shown by the averaged TCW (Figure 5.1b). The amount of moisture transported to Switzerland was exceptionally high (Froidevaux, 15.09.2014), both for this time of year and for the wind direction (northwest). The moisture uptake occurred mainly over the Atlantic, north of 30°N and a small part of the uptake took place in the Subtropics south of 30°N (Figure 5.1c).

The warm front beneath the western flank of the PV streamer produced strong precipitation on the relief, with some areas experiencing more than 240 mm in 48 hours, especially the northern Alps (south of the Lütschine catchment, Figure 5.2a). However due to the cold temperatures, precipitation fell almost exclusively as snow over the Alps. In contrast, over the Jura mountains (north of 47°N, 7°E) an important part of the precipitation fell as rain (Figure A.1a), with more
than 90 mm in 48 hours. This large precipitation amount resulted in the flooding of the Suze catchment (brown polygon in Figure 5.2).

The comparison of the observed precipitation (Figure 5.2a) and the precipitation produced by the CTRL run (Figure 5.2b) show that the COSMO model overestimated the overall precipitation. The largest differences are visible on the Alpine main crest with precipitation amount above 360 mm in 48 hours (compared to the 240 mm in the observations). However the precipitation over the
5.1. The December 1991 event

Figure 5.2.: Observed precipitation based on RhiresD (MeteoSwiss, 2011) (a) and simulated precipitation from the control run (b). The accumulated precipitation covers the 06 UTC 21 to 06 UTC 23 December 1991 period.

Figure 5.3.: Integrated precipitation (mm) (from 12 UTC 20 to 12 UTC 23 December 1991) for all simulations in the Suze catchment and a control box. a) shows the location of the Suze catchment (brown) and the north control box (green). The green and purple lines indicate the 1000m and 2000m elevation contours (taken from COSMO topography at 2 km resolution). The red box indicates the location of the cross-section used in subsection 5.1.3 b) shows the integrated precipitation (mm) in the Suze catchment and north control box for all simulations. The number above each bar gives the percentage of precipitation relative to the control simulation.

Jura mountains was slightly better represented with peak values of 120 mm in 48 hours (90 mm in the observations).

5.1.2. Sensitivity overview

To quantify the precipitation changes in the specific humidity and temperature experiments, a control box is defined to the northwest of the Suze catchment (green
polygon in Figure 5.3a, hereafter referred as Suze North box). Total precipitation in the Suze catchment and the Suze North box are shown in Figure 5.3 for all experiments. It shows that for this event precipitation in the Suze catchment (brown bars in Figure 5.3), does not vary significantly with specific humidity changes. The maximal increase is only +3% during the QV +30% experiment. However increasing temperature reduces the total precipitation in the catchment, a decrease of −9% occurs in the T +2K experiment. On the other hand, the control box (green bars in Figure 5.3) experiences a strong increase of total precipitation with increasing specific humidity. For example the QV +30% experiment produces a precipitation increase of +50%. Inversely, increasing temperature experiments lead to a decrease of the total precipitation also in the box, with a maximal decrease of −36% during the T +2K experiment.

5.1.3. Process analysis

From the previous section the following main questions arise:

(a) Why is the precipitation in the Suze catchment not reacting to increasing specific humidity, when precipitation in the Suze North box strongly increases?

(b) Why is precipitation reduced in the temperature experiments, both in the Suze catchment and in the Suze North box?

The first question is investigated based only on the QV +10% and QV +30% experiments. Similarly, for the second question, only the T +2K experiment is used.

First, the temporal evolution of precipitation and IWT is investigated. Focusing only on the CTRL run (black lines in Figures 5.4a-b) and the Suze catchment, it appears that precipitation is uniquely associated with the previously mentioned warm front. Indeed, precipitation is maximum during the 18-hour period from 18 UTC 21 to 12 UTC 22 December 1991, when the IWT remains constant at a very high value (about 700 kg m$^{-1}$s$^{-1}$). Understandably the IWT shows slightly higher values upstream of the Suze North box compared to the location upstream of the Suze catchment.

Comparing the CTRL run with the QV +10% and QV +30% experiments reveals no significant changes of the IWT. However, despite this fact, the precipitation increases in the Suze North box in both experiments. During the period
from 00 UTC to 12 UTC 22 December 1991, precipitation in the Suze North box increases by $\sim 30$ mm in the CTRL, but by $\sim 50$mm in the QV+30% experiment. In contrast, during the T+2K experiment the IWT is increased up to 850 kgm$^{-1}$s$^{-1}$ upstream of the Suze North box, but interestingly the total precipitation decreases by $\sim 20$ mm.

To better understand the precipitation changes, cross-sections (red box in Figure 5.3a) covering the Suze catchment and the Suze North box are shown in Figure 5.5 for the CTRL run, the QV+10%, QV+30%, and T+2K experiments. The cross-sections are shown for the period from 00 UTC to 12 UTC 22 December 1991, when the precipitation differences between the experiments are largest. The CTRL run shows an important warm cloud at low levels (red contours in Figure 5.5) throughout the whole cross-section and graupel and snow layers at higher
Chapter 5. Relevant sensitivity aspects of four additional events

Figure 5.5.: Cross-section over the red box shown in Figure 5.3a for a 12-hour period from 00 UTC to 12 UTC 22 December 1991. Each small panel shows 3-hourly average fields over the width of the red box. The coloured contours show various hydrometeors. Cyan stands for ice in 10 mg kg\(^{-1}\) intervals, blue for snow in 150 mg kg\(^{-1}\) intervals, grey for graupel in 75 mg kg\(^{-1}\) intervals, red for cloud water in 150 mg kg\(^{-1}\) intervals, and green for rain in 75 mg kg\(^{-1}\) intervals. The coloured bars indicate the location of the Suze catchment (brown) and the Suze North control box (green). The grey shading shows the averaged topography. Each panel represents a different experiment, the control run in (a), the specific humidity experiments QV +10% and QV +30% in (b,c), and the temperature experiment T +2K in (d).
levels, especially over the Suze catchment (close to the mountain range top). The cross-section shows that precipitation is mainly formed by riming and coalescence below the graupel/snow maximum, especially during the first 9 hours. The situation is similar during the QV +10% experiment with however a slightly lower cloud water concentration and a snow cloud extending further upstream (towards western boundary of the cross-section). In this case, due to the extended snow layer, the precipitation formation is increased above the Suze North catchment, in the upstream direction. Moreover during the QV +30% experiment, the snow layer is deeper, leading to a longer life time of the snow layer above the Jura mountain, and therefore precipitation over both the Suze North and the Suze catchment is maintained during the whole 12-hour period. In contrast, the T +2K experiment shows a decrease of the snow and graupel concentrations due to the higher altitude of the zero degree line. This decrease leads to a decrease of the precipitation formation, especially over the Suze North box.

5.1.4. Summary

In summary, the formation of precipitation due to the interaction of ice hydrometeors (here graupel and snow) with cloud water is crucial during this event. Modification of the ice hydrometeors concentration induces both precipitation increase and decrease. The experiments with increased specific humidity produce higher snow concentrations, inducing enhanced precipitation particularly upstream of the Suze catchment, where the snow concentration is very low in the CTRL run. On the opposite, experiments with increased temperature lead to an overall decrease of ice hydrometeors and therefore to a decrease of the overall precipitation.

5.2. The September 1993 event

5.2.1. Synoptic evolution

The flood event of September 1993 was characterized by an upper-level PV streamer (Figure 5.6a) located over France extending from the British Island to Spain (oriented north-south). During the course of the event, the PV streamer slowly evolved into a northwest to southeast orientation. The PV streamer is also visible
in the TCW field (Figure 5.6b), as a band of low values extending from the British Island down to Morocco, with high values on its eastern flank, particularly over the Mediterranean, south of Genoa. The precursor role of the upper-level structure for this heavy precipitation event, and its link with the mesoscale flow was already shown by Massacand et al. (1998), Buzzi and Foschini (2000), and Massacand et al. (2001). Interestingly the moisture uptake diagnostics (Figure 5.6c) shows that most of the moisture originated from the eastern Mediterranean in a band located to the south of Italy, extending from Algeria to Greece. The moist air mass was then advected northward on the eastern flank of the PV streamer towards the Southern Alps.

This situation was very similar to the October 2000 case described in section 4.2 and also produced heavy precipitation south of the Alps (Figures A.3b,c). The maximum of the observed precipitation (Figure 5.7a) was located near the Saltina catchment (red polygon) in the Simplon mountain range, with more than 360 mm in 48 hours. In this catchment, the majority of the precipitation fell as rain (Figure A.3a) The large precipitation amount produced a flood in Brig which caused damage for about 500 millions CHF (FOEN, 2002). The COSMO CTRL run (Figure 5.7b) produced less precipitation especially over the Ticino with peaks values above 240 mm, whereas more than 480 mm were observed. In their study of large-scale precipitation extreme in the Mediterranean, covering the period from 1979-2012, Raveh-Rubin and Wernli (2015) found that the September 1993 is the most extreme precipitation event.

5.2.2. Sensitivity overview

To investigate the sensitivity of this event, two interesting catchments were chosen. The Saltina catchment (red polygon in Figure 5.8a) where the flood occurred and the Hinterrhein catchment (brown polygon in Figure 5.8a) in the eastern part of the precipitation area, chosen because of its strong reaction in the sensitivity experiments. In addition, two control boxes are defined to the south of each catchment, hereafter referred to as Saltina South and Hinterrhein South boxes (green and blue polygons in Figure 5.8).

The Saltina catchment and Saltina South box show a very weak precipitation response for increasing specific humidity, both the QV+20% and QV+30% experiments produce an increase of only +5% (Figure 5.8b). However, in the Saltina
5.2. The September 1993 event

Figure 5.6: Mean over the period 06 UTC 22 to 06 UTC 25 September 1993 of PV (a) (vertical average between 310K and 340K), TCW (b), and the evaporative moisture sources (c).

catchment, the increasing temperature experiments produce lower total precipitation compared to the CTRL run, with a very strong decrease of $-31\%$ in the T+2K experiment. In contrary the Saltina South box experiences a slight increase of the total precipitation during the temperature experiments, with a maximal increase of $+12\%$ in the T+2K experiment.

The Hinterrhein catchment and the Hinterrhein South box behave very differently with a strong increase of total precipitation with increasing specific humid-
Chapter 5. Relevant sensitivity aspects of four additional events

Figure 5.7.: Observed precipitation based on RhiresD (MeteoSwiss, 2011) (a) and simulated precipitation from the control run (b). The accumulated precipitation covers the 06 UTC 22 to 06 UTC 25 September 1993 period.

ity (Figure 5.8c). The Hinterrhein catchment experiences an increase of +80% total precipitation both in the QV +20% and QV +30% experiments. The Hinterrhein South box experiences a stronger increase of +133% during the QV +20% experiment and of +124% during the QV +30% experiment. In addition, both areas experience an increasing total precipitation for increasing temperatures, with about +40% total precipitation during the T +2K experiment.

5.2.3. Process analysis

From the previous section, two main questions arise:

(a) What causes the large difference between the Saltina catchment and the Hinterrhein catchment in the specific humidity experiments?

(b) What causes the decrease of precipitation in the Saltina catchment during the T +2K experiment?

For simplicity reason, only the CTRL, the QV +10% and T +2K experiments will be investigated here in greater details, since they are sufficient to answer the questions above. First the temporal evolution of precipitations and IWT are shown in Figure 5.9 for the CTRL. During the precipitation phase (from 06 UTC 23 to 06 UTC 24 September 1993) IWT values are ranging from 380 kg m⁻¹ s⁻¹ to 600 kg m⁻¹ s⁻¹, the highest values being found upstream of the Hinterrhein South box, whereas the Saltina catchment experiences the lowest IWT due to its high
Figure 5.8.: Integrated precipitation (mm) (from 06 UTC 22 to 06 UTC 25 September 1993) for all simulations in the Hinterrhein and the Saltina catchments and in two different control boxes. a) shows the location of the Hinterrhein catchment (brown), the Saltina catchment (red), the Hinterrhein South control box (blue), and the Saltina South control box (green). The green and purple lines indicate the 1000m and 2000m elevation contours respectively (taken from COSMO topography at 2 km resolution.) b, c) show the integrated precipitation (mm) in the Hinterrhein catchment and the Hinterrhein south box, and the Saltina catchment and the Saltina south box, respectively, for all simulations. The number above each bar gives the percentage of precipitation relative to the control simulation.
elevation. But interestingly over the Saltina catchment and Saltina South box the precipitation efficiency seems to be significantly higher than over the Hinterrhein catchment and Hinterrhein South box. For example, despite a regularly higher moisture influx in the Hinterrhein South box compared to the Saltina South box, the total precipitation in the Hinterrhein South box (150mm) is only half the precipitation in the Saltina South box (300mm). In addition, precipitation in the Saltina catchment and in the Saltina South box is starting at 06 UTC 23 September 1993, whereas precipitation is starting 12 hours later in the Hinterrhein catchment and Hinterrhein South box.

The increase of specific humidity in the QV +10% experiment only leads to an increase of the IWT upstream of the Hinterrhein catchment and the Hinterrhein South box (Figures 5.10e-h). In the QV +10% experiment, the Hinterrhein South box experiences almost 1.5 times the moisture influx of the CTRL run (Figure 5.10g). The increase of IWT and precipitation are both stronger for the Hinterrhein South box than for the Hinterrhein catchment. However the strong
5.2. The September 1993 event

Figure 5.10: Temporal evolution of the IWT (left panels) and the accumulated precipitation (right panels) for the Saltina catchment (a,b), the Hinterrhein catchment (c-d), and the Hinterrhein South box (e-f), for the 72-hour period from 06 UTC 22 September 1993 to 06 UTC 25 September 1993. The CTRL run is coloured in black, the QV +10% experiment in light blue, and the T +2 K experiment in red.
IWT increase seems not to be enough to explain the increasing precipitation, since the T +2K experiment has similar IWT values than the QV +10% experiment in both areas (Figures 5.10e,g) but the precipitation remains significantly lower (Figures 5.10f,h).

Likewise the decrease of precipitation in the Saltina catchment during the T +2K experiment seems not to be directly related to changes in the IWT since there are almost no IWT changes compared to the CTRL run. The temporal evolution of precipitation and the IWT for the Saltina South box are not shown because they are very similar to the CTRL experiment.

In order to understand the difference of the IWT between the Saltina and the Hinterrhein areas during the QV +10% experiment, the movement of the air masses is analysed with backward trajectories. They are started in the cross-section boxes (red polygons in Figure 5.8) every 25 hPa vertically up to 700 hPa and every 10 km horizontally. The trajectories are calculated backwards for 5 hours starting at 00 UTC 24 September 1993, the middle point of the precipitation phase, retaining only trajectories that produce precipitation during the last hour (from

Figure 5.11.: 5-hour backward trajectories started at 00 UTC 24 September 1993 from the cross-section boxes (red polygons in Figure 5.8a), coloured by height [m]. Trajectories are started every 25 hPa vertically and every 10 km horizontally. The arrows show the wind field and the shading the specific humidity [g kg⁻¹] both at 850 hPa. All data are taken from the CTRL run.
23 UTC 23 to 00 UTC 24 September 1993). In the CTRL run, clear differences are observed between the two areas (Figure 5.11). The trajectories started over the Saltina area originate from low levels over the Po valley (with an averaged height of 831 m at 20 UTC 23 September 1993), whereas the trajectories from the Hinterrhein area originate from higher levels (with an averaged height of 1495 m at 20 UTC 23 September 1993) and have a more southern origin. This height difference between the trajectories from the Saltina areas and from the Hinterrhein areas explains the difference observed in the IWT values. Indeed, the increase of specific humidity had a weaker effect at lower elevation since the air was already saturated or close to saturation.

Figure 5.12.: Cross-sections centred on the Saltina catchment (red box in Figure 5.8a) covering the 12-hour period from 18 UTC 23 to 06 UTC 24 September 1993. Each small panel shows 3-hourly averaged fields over the width of the box. The coloured contours show various hydrometeors. Cyan stands for ice in 20 mg kg$^{-1}$ intervals, blue for snow in 500 mg kg$^{-1}$ intervals, grey for graupel in 200 mg kg$^{-1}$ intervals, red for cloud water in 150 mg kg$^{-1}$ intervals, and green for rain in 75 mg kg$^{-1}$ intervals. The coloured bars indicate the location of the Saltina catchment (red) and the Saltina South control box (green). The grey shading shows the averaged topography. Each line represents a different experiment, the control run (a), the specific humidity experiments QV +10% and QV +30% (b,c), and the temperature experiment T +2K (d).
The difference in the air masses approaching the Saltina and the Hinterrhein catchment can also be seen in cross-sections aligned on both catchments (Figure 5.12 and Figure 5.13). Cross-sections aligned on the Saltina catchment (Figure 5.12) are shown for the 12-hour period from 18 UTC 23 to 06 UTC 24 September 1993 during which the precipitation difference between the CTRL and the T+2K become large. The cross-sections exhibit typical cloud structures for orographically lifted air masses, coherent with the trajectories coming from low levels presented above. Cloud water is concentrated at low levels and reaches into the Saltina South box (blue bar in Figure 5.12a). Enhanced graupel concentrations are located above the first peaks, where strong precipitation occurs. The precipitation formed above the Saltina catchment is due to melting of the snow that is advected northward. The main difference of the QV+10% experiment (Figure 5.12b) compared to the CTRL run, is the larger snow concentration at high levels. However it does not have an impact on the surface precipitation; the conditions are similar to the CTRL run above the Saltina South box and precipitation formation was already very efficient above the Saltina catchment in the CTRL run. In contrast, the T+2K experiment produces a strong enhancement of convection above the first peak, particularly visible during the last two 3-hour periods by the larger graupel concentration (last two panels of Figure 5.12c). This enhance convective activity is very efficient in removing moisture from the atmosphere. Indeed, the snow concentration above the Saltina catchment in the CTRL run is not present in the T+2K experiment, leading to a strong reduction of the precipitation formation in this area.

On the other hand, cross-sections over the Hinterrhein areas show that precipitation is formed by depositional growth and melting of snow due to the presence of large snow concentrations at high levels (Figure 5.13). This is consistent with the trajectories originating from higher levels presented above. Both the QV+10% and the T+2K experiment show an increase of the snow concentration compared to the CTRL leading to strong precipitation throughout the cross-section as already highlighted above with the continuous precipitation increase in both the Hinterrhein catchment and Hinterrhein South box. In both experiments, precipitation formation is enhanced above the Hinterrhein South box due to an enhanced graupel concentration. However, both the snow and graupel con-
5.2. The September 1993 event

Figure 5.13.: Same as Figure 5.12 but for the Hinterrhein (red bar) catchment and the Hinterrhein South control box (blue bar)

Concentrations are higher in the QV +10% compared to the T+2K explaining the stronger precipitation response in the QV +10% experiment.

5.2.4. Summary

The differences between the Saltina and Hinterrhein areas were due to the differences in the mean height of the approaching air masses. Air masses approaching the Saltina areas originated from low levels and formed precipitation by convection over the first peak and spillover precipitation in the Saltina catchment, whereas the air masses approaching the Hinterrhein areas originated from higher levels and formed precipitation by depositional growth of snow and melting. This major difference led to very different responses to the sensitivity experiments. In the Saltina areas precipitation formation was already very efficient in the CTRL run and did not experienced large changes in the specific humidity experiments. However increased convective activity in the temperature experiments led to an increase of precipitation at low levels and to a decrease of the snow concentra-
tion at higher levels and therefore to a decrease of the precipitation in the Saltina catchment. In the Hinterrhein catchment, depositional growth of snow and subsequent melting were enhanced by the higher snow concentration, and precipitation above the Hinterrhein South catchment was even further enhanced by larger graupel concentrations compared to the CTRL run, for both specific humidity and temperature experiments.

5.3. The August 2007 event

5.3.1. Synoptic evolution

The synoptic situation of the August 2007 event was characterized by a quasi-stationary PV streamer located over the British Island (from 12 UTC 06 to 00 UTC 09 August 2007) that evolved into an upper-level cut-off low at 06 UTC 09 August 2007 (Figure 5.14a). Despite the presence of this PV streamer, the wind at 850 hPa remained weak during the whole period. Under the influence of first the eastern flank of the streamer and then of the cyclonic circulation induced by the cut-off low centred over the Alps, the main wind direction strongly varied from west to southerly to northwesterly. This is also reflected in the 6-hourly precipitation field from the COSMO CTRL run (Figure A.10), where precipitation affected the western part, the southern part and then the northern part of Switzerland. Concerning the moist air masses, high TCW values were located below the eastern flank of the PV streamer, in a band extending from south of Norway to Spain (Figure 5.14b). This band evolved similarly to the streamer with a slow easterly movement. The moisture diagnostic shows that uptake principally took place over land (Figure 5.14c), in a band extending from northern Germany to southern France, coherent with the high TCW values described above (Figure 5.14b). This event happened during the COPS campaign (Wulfmeyer et al., 2011, IOP 14), and Wernli et al. (2010) found that the specific humidity at 925 hPa on the 08 UTC 08 August 2007 was in the 95 percentile of a climatology from 1979 to 2006 based on the ECMWF products.

The observed precipitation (Figure 5.15a) shows a relatively homogeneous distribution with slightly higher values over the northern Alps and near Luzern (8.5°E, 47°N). This distribution is principally due to the convective characteris-
5.3. The August 2007 event

Figure 5.14.: Mean over the period 12 UTC 06 to 12 UTC 09 August 2007 of PV (a) (vertical average between 310K and 340K), TCW (b), and the evaporative moisture sources (c).
tic of the precipitation (as shown later). A similar homogeneous distribution is produced by the COSMO CTRL run, with overall slightly weaker values. In particular the precipitation maximum in the Zurich area is missing.

5.3.2. Sensitivity overview

Due to the special characteristics of this event, the analysis focus on the Lütschine catchment (salmon polygon in Figure 5.16a) and a control box situated relatively far to the northeast of the catchment (hereafter referred as the Northeast box, blue polygon in Figure 5.16a).

There are strong differences between the two areas in terms of total precipitation (Figure 5.16b). The Lütschine shows a weak decrease of the total precipitation with increasing specific humidity, the maximum decrease occurs in the QV$+30\%$ experiment with $-8\%$. In contrast, the Northeast control box shows a strong increase of precipitation with increasing specific humidity, a maximum increase of $+119\%$ occurs in the QV$+30\%$ experiment.

There are more similarities for the temperature experiments since both areas experiences a increase of total precipitation. The Lütschine catchment shows an increase of $+22\%$ in the T$+2K$ experiment while the Northeast box experiences an increase of $+46\%$.
5.3. The August 2007 event

Figure 5.16: Integrated precipitation (mm) (from 12 UTC 06 to 12 UTC 09 August 2007) for all simulations in the Lütschine catchment (salmon polygon in a) and the Northeast control box (blue polygon in a). The green and purple lines in a) indicate the 1000 m and 2000 m elevation contours respectively (taken from COSMO topography at 2 km resolution), b) shows the integrated precipitation (mm) in the Lütschine catchment and the Northeast box for all simulations. The number above each bar gives the percentage of precipitation relative to the control simulation. The blue box shows the starting area of the trajectories shown in Figure 5.20

5.3.3. Process analysis

From the previous section, two main questions emerge:

- Why is the Northeast control box reacting so strongly in the moisture experiments compared to the Lütschine catchment?
- Why do the temperature and the specific humidity experiments lead to differing results?

A first important aspect regarding the first question is the time evolution of precipitation. In the Lütschine catchment (Figure 5.17b), the precipitation mainly increases during the last 18 hours of the simulation, from 12 UTC 08 to 12 UTC 09 August 2007, with an increase of +100 mm in the CTRL run. However the precipitation difference with the QV +30% experiment occurs mainly from 00 UTC to 12 UTC 08 August 2007. In the Northeast box precipitation mainly increases in two distinct phases, first in the 12-hour period from 00 UTC to 12 UTC 07 August 2007 (+40 mm in the CTRL run) and second in the last 12 hours from 00 UTC to 12 UTC 09 August 2007 (+30 mm in the CTRL run). In this area, the largest difference produced by the QV +30% experiment occurs during the first 12 hour period, during which the precipitation is increased by 80 mm compared
Figure 5.17.: Temporal evolution of the IWT (left panels) and the accumulated precipitation (right panels) for the Lütschine catchment (upper panels) and the Northeast box (lower panels), for the 72-hour period from 12 UTC 06 to 12 UTC 09 August 2007. IWT is calculated at the location of the black dots in Figure 5.16a. The accumulated precipitation is a mean over each area. The coloured lines indicate the different experiments, CTRL in black, QV +30% in blue, and T+2K in red.

to the CTRL run.

Regarding the time evolution of the IWT, the situation in both catchment is very similar. Both experience relatively weak values due to the weak winds, with a maximum of 200 kg m\(^{-1}\) s\(^{-1}\) in the CTRL run. The QV +30% experiment is similar to the CTRL run with slightly lower values during the period from 00 UTC to 15 UTC 07 August 2007, and the T+2K experiment shows slightly enhanced values during the whole simulation period. IWT is therefore not an important variable to explain the precipitation variations in the sensitivity experiments.

To investigate in greater details the processes leading to the precipitation changes cross-sections aligned with the Northeast control box and the Lütschine catchment are presented.

The first cross-sections cover the 12-hour period from 00 UTC to 12 UTC 07 August 2007 (Figure 5.18), during which precipitation changes are the largest in the
Figure 5.18: Cross-sections aligned on the Lütschine catchment and the Northeast control box (red box in Figure 5.16a) covering a 12-hour period from 00 UTC to 12 UTC 07 August 2007. Each panel shows 3-hourly averaged fields over the width of the box. The coloured contours show various hydrometeors. Cyan stands for ice in 10 mg kg\(^{-1}\) intervals, blue for snow in 300 mg kg\(^{-1}\) intervals, grey for graupel in 200 mg kg\(^{-1}\) intervals, red for cloud water in 75 mg kg\(^{-1}\) intervals, and green for rain in 100 mg kg\(^{-1}\) intervals. The coloured bars indicate the location of the Lütschine catchment (salmon) and the Northeast control box (blue). The grey shading shows the averaged topography. Each line represents a different experiment, CTRL (a), QV +30% (b), and T +2K (c).

Northeast box. This period is characterized by strong convective activity over the north of the Alps, particularly over the Northeast box (blue bars in Figure 5.18, as shown by the large graupel concentration throughout the troposphere. In the CTRL run, convection occurs during the first two 3-hour period (first two panels in Figure 5.18a), whereas convection occurs during the whole 12-hour period in the QV +30% experiment. Increased specific humidity has increased the convective potential. In the T +2K experiment, convection is slightly stronger during the first 3-hour period compared to the CTRL run, but stops thereafter.

Therefore despite weak northerly winds (not shown), the increasing moisture does not increase convection near the Alps, as instability is already released in the Northeast box and its surroundings.
Figure 5.19.: Cross-sections aligned on the Lütschine catchment and the Northeast control box (red box in Figure 5.16a) covering a 12-hour period from 12 UTC 08 to 00 UTC 09 August 2007. Each panel shows 3-hourly averaged fields over the width of the box. The coloured contours show various hydrometeors. Cyan stands for ice in 10 mg kg\(^{-1}\) intervals, blue for snow in 300 mg kg\(^{-1}\) intervals, grey for graupel in 200 mg kg\(^{-1}\) intervals, red for cloud water in 75 mg kg\(^{-1}\) intervals, and green for rain in 100 mg kg\(^{-1}\) intervals. The coloured bars indicate the location of the Lütschine catchment (salmon) and the Northeast control box (blue). The grey shading shows the averaged topography. Each line represents a different experiment, CTRL (a), QV +30% (b), and T+2K (c).

The second set of cross-sections cover the 12-hour period from 12 UTC 08 to 00 UTC 09 August 2007 and depict a very different situation. Indeed, in this period, strong convection also occurs south of the Alps, as shown by the large graupel concentrations in Figure 5.19a. And very interestingly, the outflows of the strong convective cells are advected northward over the Alpine crest producing strong precipitation in the Lütschine catchment. For a better characterization of this unusual situation, 7-hour backward trajectories are shown in Figure 5.20. The trajectories are started at 15 UTC 08 August 2007, in the middle of the cross-section period, from the blue box shown in Figure 5.16a. They are started every 20 km horizontally and every 25 hPa vertically from 1000 to 300 hPa. They show that at about 3 km (and above), the flow was southerly but it was north to northwesterly.
5.3. The August 2007 event

Figure 5.20: 7 hours backward trajectories started at 15 UTC 08 August 2007 from the blue box in Figure 5.16. The colouring shows the altitude in m. Winds arrows are shown at 700 hPa at the starting time of the trajectories.

at low levels.

In the QV +30% experiment (Figure 5.19b), both the timing and the location of the convective cells are different compared to the CTRL run, leading to a different setting of the snow and graupel layers above the Alpine crest. In addition, the low level water cloud is reduced in this experiment. This explains the slightly weaker precipitation produced by the QV +30% compared to the CTRL run.

In contrast, the T +2K experiment (Figure 5.19c) shows enhanced convective activity on the southern side of the Alps, especially during the first two 3-hour period. This leads to higher snow concentrations north of the Alps, in particular during the last two 3-hour periods. In addition, the low level water cloud is enhanced during this experiment. Both aspects lead then to increased precipitation on the northern side of the Alps and therefore in the Lütschine catchment (Figure 5.17b).
5.3.4. Summary

In this case, the large precipitation difference between the far upstream control box and the catchment are due to the convective nature of the precipitation. The time periods with the strongest precipitation are not the same in the Northeast box and the Lütschine catchment. Increased specific humidity leads to increased convective activity and therefore precipitation in the Northeast box. Precipitation in the Lütschine catchment is formed by the northward advection of strong convective cell outflows located over the southern slope of the Alps. Modification of the timing, location and intensity of the cells modifies precipitation in the Lütschine catchment. The specific humidity experiments do not change the intensity but rather the location and timing, producing a slight reduction of the precipitation in the catchment. On the other hand, convection is enhanced during the T+2K experiment leading to enhanced precipitation in the catchment.

5.4. The October 2011 event

5.4.1. Synoptic evolution

The flood event of October 2011 started with a cold episode that lasted several days and produced important snow fall over the Alps. This cold episode ended at 12 UTC 09 October 2011 (Figure A.11a) and was important for some catchments in the Alps, however the snow fall was not a critical aspect for the flood in the Lütschine catchment (Felix Naef, personal communication), our catchment of interest. A second phase, starting at 18 UTC 09 October 2011, was characterized by a warm front beneath the western flank of a PV streamer (Figure 5.21a), which produced strong precipitation over the Alps in the form of rain. The two phases are clearly visible in the 6-hourly precipitation from the COSMO CTRL run shown in Figure A.12. The PV streamer, extending from Sweden to Libya, was located at the eastern edge of an upper-level ridge situated over the Northern Atlantic. Strong, stable northerly winds were established on the western flank of the streamer and persisted despite the eastward movement of the streamer. An atmospheric river established below the upper-level ridge, as shown by the TCW band with high values over the Atlantic in Figure 5.21b. The moisture was brought from the Subtropics (Figure 5.21c) towards Europe around a high pres-
sure system centred over Spain. A detailed description of the synoptic situation leading to this HPE is presented in chapter 6.

Figure 5.21. Mean over the period 00 UTC 09 to 12 UTC 10 October 2011 of PV (a) (vertical average between 310K and 340K), TCW (b), and the evaporative moisture sources (c).

The strong precipitation period was concentrated on 12-18 hours starting at 18 UTC 09 October 2011. Observed precipitation is mainly located over the high peaks of the Alps (Figure 5.22a), with values up to 160 mm close to the Lütschine catchment. The CTRL run produced accurate precipitation (Figure 5.22b) with slightly lower peak values (up to 120 mm).
5.4.2. Sensitivity overview

The impact of the sensitivity experiments is investigated here for the Lütschine catchment and a control box upstream to the north (salmon and blue polygons in Figure 5.23a). Both areas show increasing precipitation with increasing specific humidity (Figure 5.23b). In the Lütschine catchment, the total precipitation is increased by $+35\%$ at most during the $QV_{+20\%}$ and $QV_{+30\%}$ experiments. The total precipitation increase seems to flatten with increasing specific humidity. In the control box the precipitation is increased by $+146\%$ during the $QV_{+30\%}$ experiment. The increase is strong, non-linear and increases with increasing specific humidity.

Both areas experience a weaker precipitation increase with increasing temperature, the maximal increase ($+11\%$) occurs in the Lütschine catchment in the $T_{+2K}$ experiment. An additional sensitivity experiment is performed for this event, where the lateral boundary conditions were kept constant, corresponding to the conditions at the time step with the highest influx of moisture. In this experiment the boundary conditions were fixed after 21 UTC 09 October 2011 until the end of the simulation (more details in section 2.3.1). This stationary boundary conditions experiment produces a increase of about $+26\%$ in both areas.

5.4.3. Process analysis

From the previous section two main questions emerge:
(a) Why is the increase of precipitation in the moisture experiments in the Lütschine significantly weaker than the increase in the Lütschine North box?

(b) What are the main differences between the CTRL and the stationary experiment?

The important precipitation period for this event is from 21 UTC 09 to 09 UTC 10 October 2011, when accumulated precipitation in the CTRL run increases from 20mm to 60mm (black line in Figure 5.24b). However the effect of increased specific humidity can be seen during two periods of precipitation. First at the end of the cold phase, from 00 UTC to 08 UTC 09 October 2011, when both the QV +20% and QV +30% experiments (blue lines in Figure 5.24b) show a similar weak precipitation increase of +10mm compared to the CTRL run. Second, during the last 3 hours of the experiments, from 09 UTC to 12 UTC 10 October 2011, when precipitation still occurs in both the QV +20% and the QV +30% experiments, in contrast to the CTRL run. Yet the effect of increased specific humidity in the Lütschine North box occurs only during the last 12 hours of the experiments, from 00 UTC to 12 UTC 10 October 2011 (blue lines in Figure 5.24d). Precipitation changes associated with the stationary experiment, in both the Lütschine catchment and the Lütschine North box, are mainly due to slightly more
Figure 5.24.: Temporal evolution of the IWT (left panels) and the accumulated precipitation (right panels) for the Lütschine catchment (a,b), and the Lütschine North box (c,d), for the 36-hour period from 00 UTC 09 October 2011 to 12 UTC 10 October 2011. The IWT is calculated at the location of the black dots in Figure 5.23a. The accumulated precipitation is a mean over each area. The coloured lines indicate the different experiments, CTRL in black, QV +20% in light blue, QV +30% in dark blue, and stationary in green.

intense precipitation during the 00 UTC to 12 UTC 10 October 2011 period and to the longer lasting precipitation (as for the QV +20% and QV +30% experiments) (green lines in Figures 5.24b,d).

From an IWT perspective, the changes are weak for all sensitivity experiments. Upstream of both the Lütschine catchment and the Lütschine North box, the IWT is slightly increased in the QV +20% and QV +30% experiments compared to the CTRL run, from 00 UTC to 18 UTC 09 October 2011. During the same period, as expected, the stationary experiment IWT is similar to the CTRL IWT. From 00 UTC 10 October 2011 onward IWT values remain in all experiments about 400 kg m$^{-1}$s$^{-1}$ upstream of the Lütschine catchment and about 500 kg m$^{-1}$s$^{-1}$ upstream of the Lütschine North box. On the other hand, the IWT during the CTRL run decreases during the same period.
5.4. The October 2011 event

Figure 5.25.: Cross-sections centred on the Lütschine catchment (red box in Figure 5.23a) covering a 12-hour period from 00 UTC to 12 UTC 10 October 2011. Each panel shows 3-hourly averaged fields over the width of the box. The coloured contours show various hydrometeors. Cyan stands for ice in 10 mg kg$^{-1}$ intervals, blue for snow in 100 mg kg$^{-1}$ intervals, grey for graupel in 75 mg kg$^{-1}$ intervals, red for cloud water in 150 mg kg$^{-1}$ intervals, and green for rain in 50 mg kg$^{-1}$ intervals. The coloured bars indicate the location of the Lütschine catchment (salmon) and the Lütschine North control box (blue). The grey shading shows the averaged topography. Each line represents a different experiment, the control run (a), the specific humidity experiments QV$+20\%$ and QV$+30\%$ (b,c), and the stationary experiment (d).

In summary, slightly higher IWT during a longer period in all sensitivity experiments increases the precipitation in both the Lütschine catchment and the Lütschine North box. However, this IWT increase produces much more precipitation in the Lütschine North box than in the Lütschine catchment.
In order to investigate this differing response, cross-sections (red polygon in Figure 5.23a) centred on the Lütschine catchment and the Lütschine North box are shown in Figure 5.25 for the different sensitivity experiments. The cross-sections cover the period from 00 UTC to 12 UTC 10 October 2011. In this blocked flow situation the precipitation is mainly formed by riming of snow close the mountain peak, which in this case is strong enough to form graupel (grey contours in Figure 5.25). As the stratiform cloud produced by the warm front grows, from 03 UTC to 09 UTC 10 October 2011, the precipitation increases in the upstream direction towards lower elevations (see first three panels of Figure 5.25a), by depositional growth of snow and subsequent melting. As the warm front continues to move eastward the snow layer decreases stopping the formation of precipitation throughout the whole cross-section (last panel of Figure 5.25a).

In the sensitivity experiments, the situation remains very similar, however, the depth and density of the snow layer is increased during the whole 12-hour period from 00 UTC to 12 UTC 10 October 2011 (Figures 5.25b,c). This thicker snow layer favours the earlier and more intense formation of rain at low levels. This effect is particularly visible during the second period (from 03 UTC to 06 UTC 10 October 2011), when comparing the CTRL run and the QV +20%, QV +30% experiments (green contours in the second panel in Figures 5.25a-c). In addition, the precipitation efficiency at high levels, above the Lütschine catchment (salmon bar in Figure 5.25), is already high in the CTRL run due to the presence of graupel. Therefore the larger snow layer does not affect the rate of precipitation formation in the catchment. Moreover since the snow layer remains above the Lütschine catchment in the QV +20% experiment during the last 3-hour period (last panel in Figure 5.25b), the precipitation in the catchment is similar to the precipitation in the QV +30% experiment.

Concerning the stationary experiment (Figure 5.25d), the longer precipitation period described above is simply explained by the longer persistence of the snow layer which remains similar throughout the whole 12-hour period. However the weak strengthening of the snow layer in the upstream direction explains the weak precipitation changes above the Lütschine North box, compared to the specific humidity experiments.
5.4.4. Summary

The increased specific humidity experiments produce a denser and larger snow layer which increases the precipitation formation above the Lütschine North box, but does not increase the already efficient precipitation formation over the Lütschine catchment. The effect of the stationary experiment is to produce a more persistent snow layer which also produces precipitation during the last 3-hour period (from 09 UTC to 12 UTC 10 October 2011), in contrast to the CTRL run. However, in this experiment, the snow layer is not increased upstream and therefore does not affect the precipitation in the Lütschine North box significantly.
Chapter 6.

Dynamics of a local Alpine flooding event in October 2011: moisture source and large-scale circulation

The current chapter, based on already published work (Piaget et al., 2014), investigates a predominantly northern Alpine HPE that occurred around 10th October 2011 and affected several small catchments on the Alpine main ridge. The strongest impact was locally confined to the Jungfrau mountain range in the Bernese Alps. For instance, the rivers of Kander and Lütschine in the northern Alps experienced discharges of more than the 100 years return period\(^1\). An overview of the observed precipitation\(^2\) in Switzerland is given in Figure 6.1a. The location marked *Jungfrau* approximately indicates the Alpine main ridge. Here accumulated precipitation exceeded 360 mm within 5 days.

Interestingly, this event occurred in two closely linked phases. First, a cold episode with locally more than 40 cm of fresh snow in 24 hours produced a layer of (wet) snow in altitudes above 1000 m. This was followed by the passage of a


warm front that approached Switzerland from the north and was associated with heavy rainfall (also at high altitudes >3000 m) of locally more than 160 mm in 12 hours (Rössler et al., 2014).

Along with the high rainfall amounts and the sudden snow melt severe flooding occurred in several catchments 3, mainly along the northern side of the Alpine main ridge, but also in some catchments on the leeward (southern) side of the first Alpine barrier, e.g. the Lötschental (Rössler et al., 2014). Damages of up to 55 millions Swiss francs were caused by the event, with hundreds of houses affected, and a shut-down of traffic on road and railways 4.

Numerical weather prediction models had difficulties in correctly predicting the local precipitation amounts for two major reasons: First, the Alpine orography led to a very complex precipitation pattern (Figure 6.1a and Rössler et al. (2014)) that global models could not resolve (Figure 6.1b). Only the MeteoSwiss very high resolution model COSMO-2 (2km horizontal resolution) adequately represented the observed precipitation (cf. Figure 4.9 in Kaufmann, 2012; Rössler et al., 2014). Kaufmann (2012) also showed that the global European Centre for Medium-range Weather Forecasts (ECMWF) high resolution model (HRES, about 16km horizontal resolution) had particular difficulties in resolving the warm frontal phase of the HPE.

Second, medium-range forecasts initialised 3 to 10 days prior to the event showed marked variations of accumulated precipitation for different forecast initialisation times, in particular for the warm frontal phase of the event (Figure 6.2).

Rössler et al. (2014) discussed in details the local aspect of this HPE. This chapter investigates the large-scale circulation leading to this HPE by addressing the following scientific questions:

1. Which primary large-scale flow features were involved in triggering this local Northern Alpine HPE?
2. Where is the source of the moisture for the HPE?
3. Which role did weather systems such as ARs, TMEs, or WCBs play in transporting moisture towards Europe?
4. What are the physical and dynamical processes establishing the particular large-scale flow situation?
5. Which are the reasons for the variations in accumulated precipitation forecasts?
Chapter 6. Dynamics of a Local Alpine Flooding Event

The chapter is organised as follows. After a brief introduction of the data and methods used (section 6.1), we provide an overview of the chain of events in the Atlantic-European region that led to the Northern Alpine flooding event (section 6.2). Subsequently, aspects of the moisture transport and the moisture sources region will be examined using a Lagrangian framework (section 6.3). The dynamics of the large-scale flow pattern leading to the precipitation event and the reasons for forecast uncertainty are explored in more detail in section 6.4 based on wave activity flux (WAF) and PV diagnostics. We conclude with a discussion (section 6.5) and the presentation of the main results (section 6.6).

6.1. Data and methods

6.1.1. Data

Six-hourly operational HRES analysis data (ECMWF) interpolated on a regular geographical grid with a 1.0° horizontal resolution are the data basis for this study. On the same grid, operational HRES forecasts initialised every 12 hours are used to assess forecast uncertainty. In addition ERA-Interim data (ECMWF Interim Re-Analysis, Dee et al., 2011) for the period 1979-2011 serve as a climatological reference.

A six-hourly precipitation reference is constructed using short-term HRES forecasts on a regular geographical grid with 0.125° horizontal resolution (full model resolution).

Similar to (Pfahl and Wernli, 2012), forecast steps from 6 to 12 hours and 12 to 18 hours are used to account for model spinup.

The high resolution gridded rain-gauge observation ⁵ as well as the station data ⁶ are provided by MeteoSwiss.

6.1.2. Rossby wave diagnostic

In section 6.4, two complementary diagnostic methods are employed to explore the physical processes that helped to establish the large-scale flow situation. The

⁵http://www.meteosuisse.admin.ch/web/fr/services/portail_des_donnees/donnees_matricielles/precip/rhiresd.html
first method is based on quasi-geostrophic, weakly-nonlinear wave theory and the second method is based on the PV perspective of baroclinic instability and downstream development.

**Wave activity flux diagnostic**

The first method builds on the conservation properties of wave activity (A) for almost-plane, linear waves in conservative flows. Under these assumptions, wave activity is redistributed by the divergence of a flux vector $\vec{F}$: $\partial A/\partial t + \nabla \cdot \vec{F} = 0$. The meridional and vertical components of $\vec{F}$ are a measure of the meridional flux of zonal momentum and the meridional heat flux, respectively. A propagating wave packet exhibits convergence of $\vec{F}$ ($\nabla \cdot \vec{F} < 0$) at its leading edge and divergence of $\vec{F}$ ($\nabla \cdot \vec{F} > 0$) at its trailing edge. The underlying weakly-nonlinear theory links $\nabla \cdot \vec{F}$ to changes in the background (or mean) flow. Specifically, convergence ($\nabla \cdot \vec{F} < 0$) is associated with a deceleration of the background flow. Here, we employ a wave flux formulation by Takaya and Nakamura (2001) that can be applied to migratory waves propagating along a slowly varying, non-zonal background state. In this study, only the horizontal flux components are considered, which emphasises the horizontal propagation of upper-tropospheric wave trains and neglects baroclinic interactions.

The wave activity flux (WAF) formalism requires the separation of the total flow into a background state and related anomalies. The anomalies then represent the wave pattern of interest. Here, we follow the separation of Nakamura and Wallace (1993) and consider both high- and low-frequency wave signals. For the high-frequency wave (hereafter referred to as fast-transients) the background state is defined by a 6-day low-pass filter. The fast-transients are extracted by a 6-day high-pass filter. The low-frequency wave is extracted by a 6-60 days band-pass filter and the associated background state is defined by a 60 day low-pass filter. To the extent that the low-frequency wave constitutes a background state for the fast-transients, divergence of the flux of fast-transients implies a modification of the low-frequency wave. Specifically, convergence of fast-transients can be interpreted as a local amplification of the low-frequency wave (Nakamura et al., 1997). When interpreting the results it has to be kept in mind that the real atmosphere does not fulfil the formal assumptions made for the formulation of the WAF concept. However, previous work (e.g. Nakamura and Wallace, 1993; Nakamura et
al., 1997) has demonstrated the utility of WAF diagnostics when applied beyond its formal validity.

**PV perspective**

The second method is a quantitative PV diagnostic based on the concepts of baroclinically coupled Rossby waves (Bretherton, 1966; Hoskins et al., 1985) and their downstream development (Simmons and Hoskins, 1979). The total PV field is partitioned into a background state (defined here as a 30-day average centred on 7th October) and lower- and upper-level PV anomalies\(^7\). Following Davis et al. (1996), 650 hPa is chosen as the separation level. Using piecewise PV inversion under non-linear balance (Davis, 1992) the wind fields associated with the PV anomalies is obtained. These wind fields are then used to calculate advective PV tendencies at upper levels (see Equation 6.1).

\[
\frac{\partial PV'}{\partial t} \approx \left. \frac{\partial PV'}{\partial t} \right|_{bt} + \left. \frac{\partial PV'}{\partial t} \right|_{bc} + \left. \frac{\partial PV'}{\partial t} \right|_{div} = -\left( v_{bt} + v_{bc} + v_{div} \right) \cdot \nabla PV
\]

(6.1)

Upper-level PV tendencies due to the upper-level PV anomalies themselves signify the quasi-horizontal propagation of the upper-level Rossby wave (hereafter referred to barotropic (bt) propagation). Upper-level PV tendencies due to lower-level PV anomalies signify baroclinic (bc) coupling. We complement this picture with upper-level advective PV tendencies due to the divergent wind (div) which is derived by Helmholtz partitioning. The divergent component serves as a surrogate for the impact of mid-tropospheric latent heat release. This framework has been employed by Riemer and Jones (2010) to analyse cases of ridge building in idealized baroclinic wave experiments.

### 6.2. Synoptic overview

In this section, we discuss the synoptic evolution in the North Atlantic-European region that preceded the high impact weather (HIW) situation in the Alps from 12 UTC 01 October to 12 UTC 10 October 2011. Furthermore, the local conditions in Switzerland in the region of severe precipitation and flooding are illus-

\(^7\)The lower-boundary \(\theta\) anomaly is part of the low-level anomalies.
Figure 6.3.: Potential vorticity (shading, pvu) on the 325 K isentropic level and mean sea level pressure (black contours every 5 hPa). Labels mark discussed weather systems: Hurricane Ophelia (O), Tropical Storm Philippe (P), PV streamer (later cut-off) (C), high pressure systems (H1,H2,H), an extratropical cyclone over Newfoundland (NC), and a trough (T). Times shown are indicated in the plot (every 36 hours, starting at 12 UTC 01 October 2011)

6.2.1. Formation of a persistent (upper-level) cut-off cyclone

Our investigation starts with the origin of a mid-Atlantic cut-off cyclone that played an important role in the transport of tropical moisture to the mid-latitudes several days before the HIW (detailed moisture transport will be discussed in section 6.3). At 12 UTC 01 October 2011, four distinct weather systems characterised the synoptic situation in the North Atlantic (Figure 6.3a): an upper-level ridge and associated surface high-pressure system (H1), a PV streamer downstream of this ridge located at 25°W (labelled C), Hurricane Ophelia (labelled O) located near 63°W, 28°N, and Tropical Storm Philippe (labelled P) located near 50°W, 25°N. Within the next 36 h, the southern part of the PV streamer (C) disconnected from the high-PV reservoir and formed an upper-level cut-off (at 00 UTC 03 October,
Figure 6.4.: Left: PV (shading, pvu) on the 325K isentropic level and mean sea level pressure (black contours with 5hPa interval) for the 12 UTC 07, 00 UTC 09 and 12 UTC 10 October 2011 respectively (a,c,e). Right: equivalent potential temperature at 850hPa, 30 mm contour of total precipitable water in red and sea level pressure (black contours with 5hPa interval) for the same dates (b,d,f). Labels in (a,c,e) refer to the same weather systems as in Figure 6.3b. This cut-off (C) persisted for the next 4 - 5 days and remained quasi-stationary near, 30°W, 35°N (Figures 6.3b - d,6.4a). The cut-off was a deep tropospheric system as indicated by the closed sea level isobars (Figures 6.3b - d). Deep southerly flow was therefore established between the eastern flank of the cut-off (C) and a high-pressure system (H2) building further east (Figures 6.3b,c). This southerly flow was crucial for the transport of tropical moisture towards Europe (see section 6.3). It is likely that the outflow of Ophelia and Philippe amplified the upper-level ridge H1 and assisted the PV streamer formation (C) as it was observed in similar situations of tropical cyclones undergoing extratropical transition (ET) (Riemer et al., 2008; Atallah and Bosart, 2003; Grams et al., 2011).
6.2. Development of a high-amplitude Rossby wave train

A second key feature for the continued transport of moisture towards the Alps was the formation of a strong and persistent surface high over the eastern North Atlantic and an elongated trough over central Europe (H and T, respectively, in Figures 6.3d and 6.4). Both features are a signature of a high-amplitude Rossby wave train that spans the North Atlantic - European region at 12 UTC 07 October (Figure 6.4a). The dynamics of this wave train will be the focus of section 6.4. Here, we note that the propagation of a wave train from North America into the North Atlantic is indicated at 12 UTC 04 and 00 UTC 06 October (Figures 6.3c-d). The remnant outflow circulation of Ophelia led to a local amplification of the wave pattern over the eastern North Atlantic around 05 October (see Figures 6.3c,d and section 6.4). WCB trajectories started at 00 UTC 03 October show a WCB developing from the vicinity of Ophelia and impacting the upper-level flow at 00 UTC 05 October (Figure 6.5a). The outflow of the trajectories were localized in the upper level ridge (H). A much more pronounced amplification of the wave train occurred due to the rapid development of an extratropical cyclone near Newfoundland (labelled NC, from 00 UTC 06 October to 12 UTC 07 October, Figures 6.3c,6.4a). Divergent flow, associated with the cyclone’s WCB, further enhanced ridge building in the North Atlantic (section 6.4) similar to Grams et al. (2011). Indeed trajectories started at 00 UTC 04 October show that, along with the cyclogenesis of NC, a WCB develop over the North Atlantic and that the outflow impacted the upper-level flow at 00 UTC 06 October (see Figure 6.5b).
At the same time the cut-off (C) decayed and H1, H2 were replaced by one pronounced North Atlantic ridge (labelled H).

The amplification of the mid-latitude flow pattern continued downstream (Figure 6.3d) and a pronounced trough (T) was located over Europe at 12 UTC 07 October (Figure 6.4a). Strong northerly flow established between this trough and the eastern flank of the North-Atlantic ridge (H). The northerly flow reached deep through the troposphere and persisted for the next four days. The elongation and subsequent anticyclonic breaking of the trough over Europe (Figures 6.4a,c,e) promoted the persistence of the northerly flow over central Europe from 12 UTC 07 October to 12 UTC 10 October. This northerly flow initially advected cool air into central Europe (Figures 6.4b,d), leading to the predecessor snow fall that occurred in the Swiss Alps from 00 UTC 08 October to 12 UTC 09 October. Due to the longevity of the North-Atlantic ridge (H) and the breaking of the trough (T), however, warm air of tropical origin was advected around the ridge and then into Europe on a northwesterly trajectory beginning on 09 October (labelled TME, Figures 6.4d,f). Trajectories extracted from the TME climatology of Knippertz et al. (2013) confirm that this feature spanning from the African west coast to the Alps is a TME (see Figure 6.6).
Figure 6.7.: (a,b) show wind speed (ms$^{-1}$) in blue and wind direction (°) in red for Interlaken and Jungfraujoch station respectively. (c,d) show the temperature at 2m (blue) and pressure (red) for both station. Accumulated precipitation data was only available at Interlaken station (c). The figure covers the period between 00 UTC 06 and 00 UTC 12 October 2011, with a 10 min. resolution.

A closer inspection of $\Theta_e$ at 850hPa reveals some details of the tropical origin of the moist, warm air mass. At 12 UTC 07 October, tropical air that had been advected poleward between the cut-off (C) and high pressure system H2 during the previous days (Figures 6.3a-d) converged between the cold front of the new cyclone NC and the new Atlantic ridge H (Figure 6.4b). The process of moisture convergence in frontogenetical region, has already been described by Cordeira et al. (2013) to be crucial for the formation of an AR (TME). This TME was subsequently advected around the Atlantic ridge into Europe (Figure 6.4d) and reached the Alps at 00 UTC 10 October as a prominent warm front (not shown). The HIW on 10 October was then triggered by the warm air impinging on the Alpine topography (cf. section section 6.3).
6.2.3. Sharp transition between distinct air masses in the Swiss Alps

Observational data from two neighbouring weather stations - Interlaken in central Switzerland at 580 m above mean sea level (amsl) and Jungfraujoch 20 km south at 3580 m amsl, representing the lower- and mid-troposphere, illustrate the vertical extent of the cool air mass (Figure 6.7). With the passage of the cold front on 7 October, temperatures at both stations dropped by 10° C and 15° C, respectively, in 12 hours (Figures 6.7c,d). The cold conditions remained over Switzerland for two days and with sustained northwesterly winds an orographic blocking situation developed at the Alpine North side. This orographic blocking led to a first, cold HPE, between 00 UTC 07 October and 00 UTC 09 October. Snow observation at sites First (2110 m), Gandegg (2717 m), and Schilthorn (2360 m) in the Lütschine catchment, recorded an increase of snow depth of 45 cm, 100 cm, and 55 cm during this 48h hours (not shown). The accumulation of precipitation in
solid form has been crucial for the flood in the leeward Lötschen valley (Rössler et al., 2014), but less relevant for the flood in the Lütschine Valley (Felix Naef, 2013, personal communication).

Observations at Jungfraujoch, which can be considered to represent the flow in the free-troposphere, show that the northwesterly flow persisted over the whole period (until 00 UTC 12 October, Figure 6.7b). Embedded in this northwesterly flow, the moist, warm air mass of tropical origin (Figures 6.4d,f and section 6.3) reached Interlaken and the Jungfraujoch between 12 UTC 09 and 00 UTC 10 October (Figures 6.7b,d). A distinct warm front is indicated by the very sharp gradient of $\theta_e$ over the British Isles in Figure 6.4d at 00 UTC 09 October. The passing of this warm front was associated with intense precipitation at Interlaken around 00 UTC 10 October 2012 (Figure 6.7c). At this time, also TME trajectories (Figure 6.6) reached the Alps. Temperature observations at Jungfraujoch indicate that the freezing point rose to well above 3000 m at the same time (Figure 6.7d).

Consequently, intense warm precipitation occurred at the north side of the Alps in the form of rain up to altitudes of more than 3000 m. Particularly high values of precipitation were measured near the crest of the Bernese Alps with some stations reporting more than 120 mm in 12 hours, (Rössler et al., 2014).

### 6.3. Transport of moisture

From the previous section it emerges that the large-scale flow over the North Atlantic region steered a TME against the Northern Alps. Here we provide an analysis of the moisture transport and uptake associated with the HPE to determine the relative importance of the different weather systems involved in the Northern Alpine HPE.

Total precipitable water (TPW) of the TME spanning along the northern flank of the North Atlantic ridge, over the British Isles into Central Europe widely exceeded 30 mm and belonged to the top 1% of the most extreme TPW (exemplary shown here for 06 UTC 10 October, Figure 6.8). Locally TPW even exceeded the 0.1% percentile namely over the British Isles and in the region of the HPE in the Northern Alps. Note that these high values of TPW over Central Europe were associated with strong northerly winds on the upstream flank of the narrowing trough (T) so that the amount of moisture transported towards the elevated orog-
raphy was exceptional (Rössler et al., 2014).

Extreme values of TPW also occurred in a strong southwesterly flow on the downstream side of the trough (T) near 40°E, 45°N over Turkey. Thus the trough (T) over Europe played an important role for transporting extremely moist air along both its up- and downstream flanks.

The excess of the 99% and even 99.9% percentile for TPW along the diagnosed TME highlights the unusual transport of moisture towards Europe. By definition a TME is rooted in the Tropics and therefore implies a poleward transport of tropical moisture. However, Bao et al. (2006) and Cordeira et al. (2013) have shown the importance of local moisture sources in the formation of a TME. To investigate the formation and development of the TME, we employed the Lagrangian moisture source diagnostics introduced by Sodemann et al. (2008, for details see section 6.1)

For this purpose, we calculated an ensemble of 10 days backward trajectories started every 30 hPa from 1000 to 500 hPa over Switzerland (45.94—47.56°N and 6.2—9.6°E), equidistantly distributed within the region (every 20 km). Only trajectories for which relative humidity exceeded 80% at their starting position over Switzerland were selected, i.e. air parcels that could potentially trigger rain over the area. In total 6159 trajectories were started at 00 UTC 10 and at 06 UTC 10 October during the peak time of the warm-frontal precipitation. From these trajectories we subsequently selected those that had lost at least 1 g kg⁻¹ specific humidity during the last 6 hours, to highlight the transport of moisture by air parcels producing more intense rain over the Alps (Figure 6.9).

A significant fraction of the trajectories were located over the African continent 10 days before reaching the Alps (47% of the trajectories crossed the red box in Figure 6.9, top). Only 20% travelled through the western Atlantic (blue box), while a majority of the trajectories were affected by the circulation of the stationary central Atlantic cut-off low (C) identified in section 6.2 (85% crossed the black box centred on the region of moisture uptake at the southeastern edge of the cut-off; see discussion below).

The mean trajectories (Figure 6.9, top) indicate that the air parcels were affected by the circulation of the cut-off low around 4-6 October 2011 (see Figures 6.3 c,d); 6 to 4 days before arriving the northern Alps.

The trajectories originating from the western Atlantic first circulated around
6.3. Transport of moisture

Figure 6.9.: 10 days backward trajectories initialized at 00 UTC and 06 UTC 10 October. Top: coloured trajectories represent mean location and specific humidity (g kg$^{-1}$) for trajectories crossing either the Atlantic box (blue) or the African box (red). Light grey dots represent six-hourly individual trajectory time steps. Labels mark days prior to arrival in Switzerland and correspond to the black lines in the lower panel. Bottom: solid lines show mean specific humidity (g kg$^{-1}$) for trajectories crossing the Atlantic (blue) and African (red) boxes. Dotted lines show mean pressure height (hPa).
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Figure 6.10: a) Relative vorticity s⁻¹ (shading) and wind m s⁻¹ at 700 hPa. Red contour indicates 2 pvu at \( \Theta = 325 \text{K} \). The African Easterly Wave is indicated by the AEW label. The former African Easterly Jet is indicated by the ex-AEJ label b) same as Figure 6.8 but at 00 UTC 04 October

the western Atlantic anticyclone H1 and subsequently around the cut-off low, reflected in the "U" shape of their mean trajectory around 35°E (Figure 6.9, top). Here they merged with the trajectory bundle emerging from West Africa and subsequently both bundles are steered around the new anticyclone H2 in the eastern Atlantic towards Europe.

The air parcels emerging from the African continent, were initially embedded in the West African monsoon system. In the begin of October 2011 a vigorous African Easterly Wave (AEW) travelling westward as a wave disturbance of the African easterly jet (AEJ) crossed the West African Coast at about 17°W, 15°N (remnants located at about 36°W, 10°N on 4 October, Figure 6.10a). A large fraction of the air parcels arriving in the Northern Alps on 10 October 2011 were initially embedded in the northern part of this AEW (corresponding to labels -10 and -8 (30 September to 2 October 2011) in Figure 6.9, see also location of air parcels at -10 days in Figure 6.11). During 2 to 4 October 2011 (-8 to -6 days, Figure 6.9, top) the AEW interacted with the central Atlantic cut-off low to the North (not shown). As a consequence of this interaction the AEJ split and was strongly deflected poleward on the eastern flank of the AEW (see also Figure 6.10).

This resulted in the advection of a very moist tropical air mass to the North and into the circulation of the central Atlantic cut-off low as shown by the percentile-
based extreme value statistics for the TPW field in Figure 6.10b. At 00 UTC 04 Oct 2011 (6 days prior to arrival), TPW exceeded the 99% percentile in a large region off the West African coast.

Compared to previous studies (e.g. Stohl et al., 2008), the trajectories also indicate that the impact of tropical cyclones for transporting moisture is rather low in this case. Only few trajectories emerged from the vicinity of Hurricane Ophelia or Tropical Storm Philippe in the western subtropical Atlantic.

Both the mean trajectory travelling through the Atlantic box and the mean trajectory of air parcels emerging from West Africa gained moisture when being affected by the circulation of the cut-off low (see colour shading, Figure 6.9, top).

The trajectories emerging from West Africa had a very high mean specific humidity of about 8 g kg\(^{-1}\) (red solid line, Figure 6.9, bottom). These trajectories gained only about 3 g kg\(^{-1}\) from 8 to 4 days prior to arrival at the Alps when transitioning from being embedded in the AEW to being affected by the circulation of the cut-off. The increase of moisture during the considered 10-day period is relatively small compared to the initial amount of moisture transported by the trajectories and compared to climatology (Sodemann and Zubler, 2010). The trajectories from the Atlantic gained moisture from 8 to 4 days prior to arrival at the Alps (red solid line, Figure 6.9, bottom) when circulating around and below the central Atlantic cut-off low. Their mean specific humidity increased by 6 g kg\(^{-1}\) to about 10 g kg\(^{-1}\) on 6 October (-4 days).

The mean height of the trajectories (dashed lines, Figure 6.9, bottom) remained at very low levels (below 800 hPa) and strong lifting occurred just when the air reached the Alps.

In summary, for both the Atlantic and the West African trajectory bundles, the highest increase in specific humidity occurs during days -8 to -4 when the majority of the air parcels circulated below the upper-level cut-off (C). The central Atlantic cut-off is hence an effective moisture collector (cf. Knippertz and Martin, 2007) and the primary weather system that helped moistening the air parcels that later reached the Northern Alps. The trajectories emerging from the African continent initially transported higher values of specific humidity and constitute around 50% of the trajectory set. Therefore we argue that they were predominantly responsible for transporting moisture to Europe.

In the following the moisture uptake as derived by the diagnostic of Sodemann
Figure 6.11.: Moisture uptake in $10^{-2}$ mm (12h)$^{-1}$ for the reduced set of trajectories (see text for details) and for the 12-hour accumulated precipitation that fell in the box centred on Switzerland (red) from 18 UTC 09 to 06 UTC 10 October. The blue dots indicate the location of each trajectory at day -10 (the ending points of the backward trajectories). Black lines show the location of cross sections shown in Figure 6.12.

et al. (2008, details in section 6.1) for the entire 10 days trajectory subset (Figure 6.11) is investigated in more detail. The air parcel locations 10 days prior to arrival in Europe (blue dots in Figure 6.11) reflect that the majority of trajectories emerge from West Africa. The major moisture uptake region is located off the west coast of Africa near 30°W, 20°N and corresponds to the southern part of the quasi-stationary central Atlantic cut-off (Figure 6.11, see also black box in Figure 6.9). Uptake values, at this location, reached more than $10 \cdot 10^{-2}$ mm (12h)$^{-1}$. A second maximum of moisture uptake is located southwest of Ireland centred around 20°W, 50°N with uptake maxima of about $5 \cdot 10^{-2}$ mm (12h)$^{-1}$. This indicates that the tropical moisture sources in the earlier phase of the TME were
subsequently reinforced by local moisture uptake along the TME (as discussed by Cordeira et al., 2013)). Moreover, uptake over land was very weak and limited to a region around the Channel and the West African coast.

Figure 6.12.: Cross-sections at the locations indicated in Figure 6.11 (black lines) through the southern portion of the cut-off in the central Atlantic at 06 UTC 05 October (top) and along the West African Coast at 00 UTC 01 October (bottom). Color shading indicates the specific humidity (2 g kg$^{-1}$ interval), the red contours the potential temperature (3 K interval) and the blue contour shows the 1.5 pvu contour (only in the top panel). Circles indicate location of trajectories from the reduced set in the cross-section (within a two degree lat/lon buffer zone), size scaled with respect to specific humidity. Black lines indicate the wind component perpendicular to the cross section. Positive/negative values (solid/dashed) in the top panel correspond to southerly/northerly flow and in bottom to westerly/easterly flow (5 m s$^{-1}$ interval). In the top panel the position of the cut-off low is indicated, as well as the northward deflection of the AEJ (ex-AEJ). In the bottom panel, the location of the AEJ is marked.
Cross-sections along the West African coast and through the main uptake region at times when most of the trajectories were located in their vicinity help to better understand the dynamics of the moisture transport and uptake (Figure 6.12). The south-north cross section along the West Africa coast (18°W) at 00 UTC 01 October, indicates a very humid, warm monsoon layer south of 18°N vertically extending up to around 600 hPa (Figure 6.12, bottom). The northern edge of the monsoon front is reflected in the strong horizontal specific humidity gradient centred around 18°N and the sloping isentropes in that region. To the North, a very shallow moist and extremely stably stratified layer at low-levels (> 950 hPa) extends over the cool ocean surface. Above this shallow cool, humid near-surface layer the Saharan Air Layer (SAL) is evident with a much drier and more neutrally stratified air mass centred around 23°N. The SAL is advected westward off the continent within a large-scale easterly flow. The AEJ has its maximum at the poleward edge of the monsoon layer (13°−18°N) with an easterly wind component of more than 12 m s⁻¹ in a layer vertically extending from 800 to 600 hPa.

Interestingly, most of the trajectories are located just above the more stably stratified part of the monsoon layer south of 18°N and just above the stable stratified near-surface layer further to the North and embedded in the rather strong AEJ. Turbulent mixing within the moist and less stably stratified upper monsoon layer, leads to a weak increase of specific humidity along the trajectories during days -10 to -8 prior to arrival in Europe (see Figure 6.9, bottom). However most importantly, the trajectories already exhibit high values of specific humidity (ranging from around 6-12 g kg⁻¹, Figure 6.12, bottom and Figure 6.9, bottom) and are transported westward with the AEJ. Moistening by turbulent mixing of the 700−900 hPa air layer above the stable stratified near-surface layer at the West African coast has previously been shown by Grams et al. (2010) in the context of the nocturnal sea breeze, which they called the "Atlantic Inflow" (their Figure 11 c,d).

A similar cross section along 21°N south of the Central Atlantic cut-off at 06 UTC 05 October helps to understand the maximum of moisture uptake in that region (Figure 6.12, top). The cut-off is evident with a lowered tropopause between 38°W and 33°W as indicated by the 1.5 pvu contour (Figure 6.12, top). The poleward deflected AEJ appears slightly to the East of the cut-off as a southerly wind maximum of more than 13 m s⁻¹ centred at 25°W, 650 hPa (labelled ex-AEJ,
Figure 6.12, top). In this region also the highest values of specific humidity are evident ranging from more than 16 g kg\(^{-1}\) near the surface to 4 g kg\(^{-1}\) at 600 hPa and reflecting the poleward transport of tropical humid air by the AEJ (Figure 6.12, top). A very humid air layer stretching from the region below the cut-off (38°W to 33°W) to the poleward deflected AEJ maximum (28°W to 24°W) and vertically extending from the surface to about 700 hPa exhibits a less stable stratification (Figure 6.12, top). Almost all trajectories are located within this layer and at altitudes around 800 hPa. From examination of near surface parameters such as wind and latent heat flux (not shown) we could exclude the role of local evaporation. Thus not evaporation from the ocean surface but strong turbulent mixing in an less stable moist air mass that has been transported in this region during the precedent days can explain the strong moisture uptake seen in this region. Thereby the interaction of the upper-level central Atlantic cut-off low with a vigorous AEW most likely produces the less stable conditions. The intensity of the uptake maximum south of the Central Atlantic cut-off can be explained by the slow movement of the trajectories during that time, the trajectories stayed almost 50% of the time in the black box.

In addition, the air parcels slowly descended to the moister lower levels (starting at about 800 hPa at day -9 and descending to 900 hPa at day -4 (Figure 6.9) bottom) which likewise results in a weak increase of specific humidity.

In summary, the moisture diagnostics revealed that a substantial fraction of the moisture involved in the HPE is of tropical origin in West Africa. Unlike to previous studies this air mass did not gain most of its moisture when travelling towards the precipitation region but started already in the Tropics with high values of specific humidity. The moderate moisture uptake along its path towards Europe was not due to local evaporation from the ocean surface but due to turbulent mixing and in a very humid and unstably stratified air layer that was predominantly linked to a Central Atlantic cut-off low. The air subsequently steered around the cut-off into the flanks of an eastern Atlantic ridge and towards Europe.

6.4. Large-scale dynamics and predictability

The ridge-trough couplet over the Atlantic evolving into a narrow trough over central Europe was crucial for finally steering the moist air towards the North-
ern Alps and triggering the Alpine flooding event. In this section we explore the physical processes that helped to establish this particular upper-level flow configuration. Two complementary diagnostic methods are employed. The first method is based on quasi-geostrophic, weakly-nonlinear wave theory. Within this framework, it will be shown that the ridge-trough pattern is part of a wave packet that extended from the West Pacific into the Atlantic and evolved with relatively low frequency (see section 6.1). Faster frequency modes then played an important role in amplifying the pattern locally.

The second method is based on the PV perspective of baroclinic instability and downstream development. The PV perspective allows a closer link of the amplification of the ridge-trough pattern to the synoptic-scale evolution of weather systems prior to the HIW event.

6.4.1. Wave activity flux diagnostic

The first method uses a wave activity flux diagnostic and considers only the horizontal flux components to emphasize the horizontal propagation of upper-tropospheric wave trains and neglect baroclinic interactions (see section 6.1).

Our diagnostic reveals that a large-scale, low-frequency wave packet constituted a precursor to the high-amplitude ridge-trough pattern. From 1st to 5th October, a low-frequency wave packet expanded over the entire North Eastern Pacific, North America, and into the central North Atlantic (see red contours in the upper panel of Figure 6.13, the envelope of the wave packet is identified by the contiguous contours). Until about the 5th of October, the location of the leading edge of the low-frequency wave packet remained persistent in the middle to eastern Atlantic and hence reduced the background flow. The persisting signal of convergence of the wave activity flux vectors (red arrows) over this region helped to initially build and amplify the central Atlantic ridge (H) during 1st to 5th October (see Figures 6.3a,b). Only after the ridge over the central Atlantic (H) was established, the low frequency wave started to propagate downstream (beginning on the 5th of October, see red arrows and contours in the middle and bottom panels of Figure 6.13).

After 4th of October, two important fast-transients wave packets emerged. On the 5th and 6th of October, strong convergence of fast-transients was found on the upstream side of the ridge (H) in the western North Atlantic (region east of the
Figure 6.13.: Upper-tropospheric wave activity flux, averaged from 250 hPa to 400 hPa for the low-frequency wave (red) and the fast-transients (grey shading). The arrows depict the direction of the flux and the contours/shading its magnitude. The contour lines start at 15 m² s⁻², with an interval of 20 m² s⁻². The low-frequency wave signal is horizontally averaged over 5 longitudes and 3 latitudes. For visual clarity, the flux fields are temporally averaged. The depicted averaged fields represent the instantaneous (noisy) fields well. Top: time average from 1st to end of 4th of October, middle: time average from 5th and 6th of October, bottom: time average from 7th and 8th of October.

dark grey shading in the middle panel of Figure 6.13). According to Nakamura et al. (1997), the convergence of the fast-transients helped to amplify and maintain the newly established low-frequency ridge (H) that gradually spanned over the entire North Atlantic basin (see Figures 6.4a,c). This interaction constitutes an important feedback from smaller to larger scales, i.e. from the fast-transients to the low frequency wave signal. Around the same time, another wave packet of fast-transient started to propagate along the mid-latitude wave guide at the northern edge of the large-scale ridge (H) over the Atlantic towards Europe (see grey
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Figure 6.14.: PV tendencies integrated over the central Atlantic ridge (H) (top panel) and the downstream trough (T) (bottom panel) from 5th to 10th October for the total, barotropic, baroclinic, and divergent flow components. Negative (positive) tendencies intensify the ridge (trough) and positive (negative) tendencies weaken the ridge (trough). Colors indicates the time of the integration in six hourly steps. Red and cyan colours span Figure 6.3 and yellow, green and blue colours span Figure 6.4 shading in the middle and bottom panel of Figure 6.13). Starting on the 7th of October, there was a strong amplification of the fast-transients over central Europe. The convergence of the fast-transients over central Europe increased also supporting the amplification of the evolving trough (T) on the downstream side of the Atlantic ridge (grey contours in the bottom panel of Figure 6.13) (see Figure 6.4a).

Our interpretation of the wave activity flux diagnostic is that a low-frequency wave packet initiated ridge (H) building and trough (T) formation over the central Atlantic and western Europe. The convergence of fast-transients in the flanks of
the ridge (H) and trough (T), respectively, then played an important role in amplifying the ridge-trough pattern locally. In the following, we adopt a PV perspective to shed some further light on this local amplification of the wave pattern.

### 6.4.2. PV perspective

Here a quantitative PV diagnostic is applied to the ridge-trough pattern regarding it as part of an upper-level Rossby wave train. The ridge (H) and the trough (T) are identified by their respective negative and positive PV anomalies on the 325K isentrope (see section 6.2). The evolution of the ridge and the trough is here described by the advective PV tendencies spatially \textit{integrated} over the area of the negative and positive PV anomaly, respectively (see Riemer and Jones (2013)). Anomalies are defined as deviations from a 30 day average background state, centred on 7th October background state (see section 6.1). For the time period from 5th - 10th October, we quantify contributions from barotropic propagation (upper-level), baroclinic feedback (low-level), and the divergent flow (latent heat release) to the temporal evolution of the integrated anomalies (Figure 6.13). To illustrate the overall evolution of the ridge and the trough in this PV framework, the tendencies due to the total flow are shown as well. As we consider the evolution of the \textit{integrated} anomalies, note that the barotropic propagation constitutes wave packet propagation with the group velocity and that the baroclinic coupling can be regarded as baroclinic growth.

The integrated PV tendencies by the total flow demonstrate that the ridge amplified strongly from the 5th to 8th and weakened afterwards (Figure 6.14a). Initially, the ridge built due to barotropic propagation (from 5th - 6th October). Subsequently, the ridge amplified by baroclinic coupling (7th - 10th October). The baroclinic growth, however, was not directly associated with the cyclonic circulation of the strong Newfoundland cyclone (NC). Only after a strong low-level warm anomaly had formed downstream of the cyclone and the high pressure system (H) (Figures 6.4a,b) had moved into the eastern North Atlantic did the baroclinic coupling amplify the ridge (H) in its eastern part (starting on 7th October, not shown). Importantly, the divergent flow had a pronounced impact on ridge building, comparable in magnitude to the barotropic and baroclinic components (Figure 6.14a). Based on the pattern of upper-level PV advection by the divergent flow (not shown), we attribute the divergent component to the outflow
of the WCB associated with the Newfoundland cyclone (NC) until 7th October
12 UTC (see section 6.2 and Figure 6.5b). The weakening of the ridge, starting on
8th October, was then again governed by barotropic propagation (Figure 6.14a).

The trough (T) amplified by barotropic propagation on the 6th and 7th October
(Figure 6.14b) and weakened by this process afterwards. Consistent with the
concept of downstream development (Simmons and Hoskins, 1979), this growth
and decay pattern occurred approximately 1 day after a similar pattern showed
up for the ridge. In contrast to the ridge (H), baroclinic coupling amplified the
trough (T) over the whole period. This persistent amplification can be attributed
to a cold anomaly upstream of the trough (not shown). The cold anomaly and the
incipient trough move in concert over the North Atlantic and into Europe without
significant change in phase. The phase-locked upper- and lower-level anomalies
amplify concurrently indicating that baroclinic instability is an important
mechanism for the trough amplification. The divergent flow on the other hand
makes an inconsistent, secondary contribution to the trough evolution.

The PV perspective corroborates the results of the wave activity flux analysis
that the ridge-trough pattern was part of a propagating upper-level wave packet.
The local amplification of the ridge (H) can be attributed to outflow from the
WCB associated with the Newfoundland cyclone (NC), and to baroclinic cou-
pling once that cyclone has formed a pronounced warm sector and displaced the
persistent high-pressure system into the eastern North Atlantic. The release of
baroclinic instability is found to be an important contribution to the amplifica-
tion of the trough (T).

Finally, the piecewise PV inversion reveals further that the barotropic flow ex-
erted a strong deformation on the upstream flank of the trough beginning on
the 8th October (quantified by a filamentation time scale as in Tsai et al. (2010)
and Riemer and Jones (2013), not shown). It is thus plausible to assume that the
high-amplitude ridge upstream of the trough was instrumental for the trough’s
evolution into an elongated feature.

6.4.3. Implications for predictability

In the introduction we noted that the medium-range precipitation forecast for
the Alpine region was highly variable prior to the event (Figure 6.2). As previ-
ously discussed it emerged that the large-scale moisture transport from the Trop-
ics within a TME and its steering towards the Northern Alps was crucial for trig-
ergating the extreme precipitation. To understand better the medium-range forecast variability we briefly explore how well the large-scale flow in the North Atlantic-European region was represented in ECMWF’s high resolution (HRES) and ensemble forecast (EPS) with particular focus on the quasi-stationary trough (T) over Europe.

The anomaly correlation coefficient (ACC, e.g. Wilks, 2011; Persson and Grazzini, 2011) is used to assess the spatially integrated HRES forecast accuracy for the large-scale mid-latitude flow. The ACC is calculated for a region centred on the ridge-trough couplet (H, T) that established the northerly flow over Central Europe (10°W to 30°E, 30°N to 65°N) and shown at forecast initialisation time every 12 hours for forecasts lead times from 48h to 144h. As meteorological quantity we use PV on the mid-tropospheric 325K isentropic level (around 500 hPa) rather than the more commonly used geopotential on 500 hPa. This choice is motivated by the fact that PV generally better accentuates synoptically important uncertainties in numerical weather forecasts compared to geopotential, as PV is a second derivative of the geopotential. The usefulness of PV in diagnosing forecast errors of the large-scale flow has recently thoroughly been investigated by (e.g. Davies and Didone, 2013; Kaufmann, 2012).

The ACC for PV at 325K reveals an important forecast degradation for initial times between 12 UTC 5 October to 12 UTC 6 October (Figure 6.15). During this period the ACC for lead times of more than 4 days dropped below the em-

![Figure 6.15.](image-url)
empirical threshold for meaningful forecast skill of 0.6. This indicates poor skill for the prediction of the upper-level PV pattern during the HPE. For example the forecast initialized at 00 UTC 6 October had an ACC of only 0.52 for a 96h forecast valid at 00 UTC 10 October. Thus there is a severe misrepresentation of the upper-level flow and the associated PV pattern that was crucial for the sustained northerly flow against the Alps and the passage of the warm front from 12 UTC 09 to 12 UTC 10 October. As the PV trough at this time is a truly synoptic-scale feature (Figures 6.4b,c), the significant degradation of ACC signifying the misrepresentation of this trough may be regarded as a severe forecast bust.

To get more insight in the reasons for the reduced forecast skill for initialization times between 5 and 7 October 2011 we exemplarily explore the EPS forecast initialized at 00 UTC 6 October and note that similar results were found for other initialization times during that period.

Forecast verification based on analysis data is only possible a posteriori. Ensemble forecast systems are effective tools to assess the uncertainty of forecasts in real-time. Thereby the spread amongst the individual forecasts, so-called ensemble members, is a measure for the uncertainty of the forecast evolution. Increased spread as measured in terms of the standard deviation amongst the different members, indicates times and locations where the evolution of the atmospheric flow exhibits large intrinsic uncertainty.

Figure 6.16 shows a longitude-time (Hovmoeller) diagram of the standard deviation of PV on the 320K isentropic level zonally averaged from 30°N-65°N for the EPS forecast initialized at 00 UTC 06 October. This approach is particularly useful to illustrate the variability of mid-latitude troughs and ridges within the scenarios given by the ensemble (Anwender et al., 2008). For the initialisation time discussed here the HRES forecast substantially underestimated the warm frontal precipitation over the Northern Alps (Figure 6.1b).

The standard deviation of PV at 320K allows a clear attribution of two major plumes of high ensemble spread in the Atlantic region (70°W-30°E) to the discussed weather systems (Figure 6.16). The first plume emerges over the Western Atlantic (60°W) slightly after forecast initialisation time and is linked to the elongation of the upper-level trough associated with the Newfoundland cyclone (NC), the subsequent building of ridge (H) and presumably the ET of Philippe. More prominently, a second plume of high ensemble spread emerges over Cen-
6.4. Large-scale dynamics and predictability

Figure 6.16.: Hovmoeller diagram for the standard deviation of PV at 320K zonally averaged from 30°N-65°N for the EPS forecast initialised at 00 UTC 06 October. Line A denote the plume associated with the Newfoundland cyclone (NC), and line B the plume associated with the trough (T).

Central Europe (0°W) on 7 October when the northerly flow established. The second plume emerges just 12-18h after the first and about 60° downstream, which hints on a downstream propagation of the forecast uncertainty. The spread with this second plume strongly increases until 13 October while propagating eastward. This indicates that there are severe problems in correctly predicting the location and amplitude of the narrowing trough (T) over Central Europe. However, the northerly flow on the upstream flank of this trough was essential for steering the moist, warm tropical air against the Northern Alps and thus in triggering the HPE. Therefore we argue that the misrepresentation of this upper-level trough was responsible for the high variability of precipitation forecasts during the warm phase prior to the event.

Beyond the uncertainty for the Alpine flooding around 10 October the evolution of the trough (T) has also consequences on the upper-level flow further to the East (30°E-60°E) as reflected in the continuing high values of standard deviation until 14 October.
6.5. Summary and Discussion

This chapter presents a detailed investigation of a Northern Alpine HPE in October 2011 that led to severe, but local flooding. A particular focus is put on the preceding large-scale circulation in the Atlantic-European region, moisture transport and sources, and the dynamical processes establishing the particular large-scale flow situation.

A synoptic analysis showed that the event occurred in two distinct phases. First, the passage of a cold front led to moderate precipitation and the accumulation of wet snow in altitudes above 1000 m. This was followed by extreme precipitation associated with the passage of a warm front. In contrast to other studies of Alpine flooding that focused on the Alpine South side and identified southwesterly flow towards the Alps on the downstream side of a trough as the main precipitation trigger (e.g. Martius et al., 2008), here the advection of the moist, warm air mass towards the Northern Alps occurred in an unusually stationary northerly flow on the upstream flank of a narrow trough. The synoptic analysis further indicated that the moist, warm air was transported from the Tropics towards Europe along the flanks of a prominent ridge.

A Lagrangian moisture diagnostic sheds more light on the origin of the moisture. Unusual moisture transport occurred within a TME emerging primarily from Tropical West Africa. Total precipitable water exceeded the climatological 99th percentile during both the formation and the landfall of the TME. An upper-level cut-off low in the subtropical Central Atlantic was identified as a key weather system that helped to initiate the TME and acted as a moisture collector. Below the cut-off the majority of the air parcels that were later associated with the heavy warm frontal precipitation in the Northern Alps converged and gained additional moisture. Unlike to other cases this moisture uptake was not due to local evaporation from the ocean surface but due to turbulent mixing in a very humid air layer in the lower troposphere. About 20% of the air parcels reaching the Alps emerged from the western Atlantic and eastern North America, starting with rather low values of specific humidity and being substantially moistened when circulating below the cut-off. More importantly, about 50% of the air parcels emerged from West Africa. They were initially transported westward with the African Easterly Jet (AEJ), and embedded in the northern part
of a vigorous African Easterly Wave (AEW). When the AEW interacted with the cut-off the AEJ was deviated poleward and advected a substantial amount of moisture into the circulation of the cut-off. Subsequently, the two major bundles of air parcels (from the Atlantic and from West Africa) merged in the cut-off and were steered into the flanks of a ridge over the eastern Atlantic and in the upstream flank of the trough over Central Europe. Eventually, the moist, warm air mass of the TME reached the Northern Alps in a persistent northerly flow; it was orographically lifted and triggered heavy precipitation at the Alpine main ridge.

The physical processes involved in the formation of the eastern Atlantic ridge and central European trough were investigated in more detail using a wave activity flux (WAF) and advective PV tendency diagnostic. The ridge formed at the leading edge of a Rossby wave-train, which extended from the North Pacific over North America into the western North Atlantic. A significant ridge amplification occurred due to divergent outflow of a WCB (WCB) associated with a cyclone off the east coast of North America and due to baroclinic coupling with the developing warm sector downstream of this cyclone. The ridge itself then contributed to the amplification of a trough further downstream over central Europe, which then established the strong and persistent northerly flow against the Northern Alps on its upstream flank. Operational numerical weather forecasts had major uncertainties regarding the position and amplitude of the trough, which led to errors in the forecast of the location and intensity of the HPE.

As discussed in the following, our study revealed important new aspects of the role of the large-scale flow situation in triggering Northern Alpine flooding events. An important synoptic feature for the initialization of the event was the stationary upper-level cut-off low (Figures 6.3b-d) in the Subtropics. In fact, it may be hypothesized that the subtropical upper-level cut-off serves as an initial moisture collector at the early stage of the TME, as discussed by Knippertz and Martin (2007). It is striking that the specific humidity of air parcels increased by 3 to 4 g kg$^{-1}$ while the air parcels circulate below the cut-off 8 to 4 days prior to the HPE (Figure 6.9). Subsequently, the transport of moisture occurred at low levels, which is in agreement with the results of Knippertz et al. (2013). Such persistent and stationary sub-tropical weather systems may be a general precursor to European HIW, in particular when interacting with midlatitude weather systems.
Cordeira et al. (2013) recently presented a detailed case study of (AR) involved in a HPE at the North American west coast. In contrast to Knippertz et al. (2013) and the results presented here, their study indicated that the moisture transport from the Tropics occurred at mid- to upper-levels, while moisture convergence along the AR was important at lower levels. This suggests that there may be differences in the AR/TME structure depending on the individual case and/or the ocean basin. Similar to Cordeira et al. (2013), we also see WCBs embedded in the TME (Figure 6.17). This confirms that also in the Atlantic case presented here local moisture convergence in a frontogenetic region strongly contributed to TME maintenance at a later stage. However, the WCB did not play a direct role in the transport of moisture for this case.

The moisture diagnostics (section 6.3) emphasised the moisture uptake below the subtropical cut-off for this HPE. Unlike to other cases, however, the majority of the air parcels involved in this event emerged from West Africa and were already very humid. Moisture uptake along the trajectories only played a secondary role for these air parcels. Previous studies documented that typical moisture source regions for Alpine heavy precipitation during the SON season are located in the Western Mediterranean and the North Atlantic (Sodemann and Zubler, 2010). Their study is based on a 7-year period from January 1995 to August 2002 and showed that the mean source latitude for the Northern Alps is northward of 42°N. Moisture uptake by turbulent mixing in a humid lower tropospheric layer for the October 2011 event discussed here, was mainly confined to tropical and sub-tropical latitudes south of 30°N (Figure 6.11). This underlines the unusual and extreme character of this HPE. Furthermore the large-scale syn-
optic situation in Europe with northerly flow against the Alps associated with moisture transport from low latitudes is very unusual.

The importance of TME-like features to produce heavy precipitation and flooding in midlatitudes has been shown by other studies, for the North American west coast (Ralph et al., 2006; Neiman et al., 2011; Cordeira et al., 2013) and coastal areas in Europe (Lavers and Villarini, 2013), and more specifically for the British Isles (Lavers et al., 2011) and the Norwegian Coast (Stohl et al., 2008; Sodemann and Stohl, 2013). However, none of these studies had a particular focus on the potential role of TMEs in triggering heavy precipitation further inland at the west-east oriented Alps. Some studies focusing on Alpine heavy precipitation have shown similar situations without specifically looking at the TME/AR mechanism. Stucki et al. (2012), e.g., examined 24 floods in Switzerland between 1868 and 2005 and classified these in 5 weather pattern categories. Their zonal flow (ZOF) category representing a near zonal upper-level jet stream, accompanied by a small band of enhanced $\Theta_e$ at low levels, is somehow comparable to the large-scale flow situation seen in this study. However, for the October 2011 event the zonal upper-level jet was shifted northward and the humid air reached the Alps from the North due to the blocking ridge in the Atlantic. Earlier studies have examined the synoptic situation of extreme precipitation events in Switzerland with similar results (W. Wehry, 1967; Grebner and Roesch, 1998). This underlines the importance of this specific large-scale flow situation for triggering floods over the Northern Alpine region.

The WAF diagnostics showed that a low-frequency wave (> 6 days) initiated the ridge-trough couplet over the North-Atlantic European region (Section 6.4). Subsequently the convergence of fast-transient wave packets of small temporal scale (< 6 days) in that region helped to maintain and amplify this upper-level flow pattern. Other studies also showed the importance of this feedback mechanism (Enomoto et al., 2006). The WAF diagnostic is consistent with the PV diagnostic (see Figure 6.14a) which showed the importance of the barotropic and divergent flow for the evolution of the ridge. The divergent flow contribution consisted of diabatic processes, as manifested in the trajectories and the strong divergent upper-level flow of a strong cyclone’s WCB (thus a weather system of small temporal scale) and was particularly important for the amplification of the ridge-trough couplet.
Unlike previous studies focussing on extratropical transition (ET), here the two subsequent ETs of Hurricane Ophelia and Tropical Storm Phillipe played only a minor role in triggering the Alpine HIW. Studies focusing on the ET of tropical cyclones (TC, e.g. Riemer et al. (2008), Atallah and Bosart (2003), Grams et al. (2013), and Archambault et al. (2013)) discussed the importance of the diabatic TC outflow for the ridge building downstream of the ET system and the modification of the Rossby wave pattern further downstream. It is thus plausible to argue that in a similar way the WCB outflow was crucial in amplifying the North Atlantic ridge which then was instrumental for the evolution of the trough over Europe into an elongated feature. Also the WCB embedded in the TME and maintaining it at a later stage (Figure 6.17) was associated with the ET of Phillipe.

Furthermore, using the PV-based metrics, we were able to show that major medium-range forecast uncertainties were related to the location and intensity of the trough (subsection 6.4.3) and then translated into a highly uncertain precipitation forecast. Such uncertainties in smaller scale details of the upper-level flow have been shown to play a key role in determining the local precipitation pattern (Schlemmer et al., 2010; Fehlmann et al., 2000).

6.6. Conclusion

This study examined a HPE on the north side of the Swiss Alps in October 2011 that led to severe flooding in two Alpine catchments. Several aspects of this event were examined: the origin and transport of moisture, the development of the large-scale circulation pattern that was crucial for the moisture transport, and predictability issues. The following key results were obtained.

• The heavy precipitation occurred in two distinct phases of which the second warm frontal phase was most important. The moist, warm air mass originated primarily from West Africa and was transported within a TME. Total precipitable water content during the formation of the TME in the Tropics and later during the HPE at the northern Alps was above the climatological 99th percentile.

• Moisture uptake occurred predominantly within a subtropical cut-off low that effectively served as a moisture “collector”. Here air from the western
Atlantic region and West Africa merged, gained moisture, and the moisture transport into the mid-latitudes within the TME was initiated.

- Subsequently, the moisture was transported poleward along the flank of a prominent ridge in the eastern North Atlantic. Following the anticyclonic circulation, the TME was then steered towards central Europe and into a northwesterly flow that established on the eastern side of the ridge and the upstream side of an amplifying trough.

- The formation of this elongated trough over Europe was crucial to steer the moist, warm air towards the Northern Alps.

- This prominent ridge-trough couplet was part of a large-scale Rossby wave train that could be traced back into the North Pacific. More locally, the North Atlantic ridge amplified significantly due to divergent diabatic outflow from a WCB associated with a cyclone off the American east coast and due to baroclinic coupling with the developing warm sector downstream of this cyclone. Baroclinic growth was of primary importance also for the amplification of the trough over Europe.

- PV based error metrics pointed to problems in the representation of the position and intensity of the upper-level trough in medium-range numerical weather forecasts explaining a high variability of the precipitation forecast with lead times 3 to 10 days prior to the event.
Chapter 7.

Climatology

7.1. Introduction

This chapter aims at classifying the extreme events identified by Schmocker-Fackel and Naef (2010) from a meteorological point of view. As a brief recall, the events are defined as HPEs leading to regional floods in Switzerland, and 36 of them occur during the 20CR period. The classification made here is based on the moisture uptake regions of each event as identified with the Lagrangian model developed by Sodemann et al. (2008). The reason for using the moisture uptake regions is that a combination of two important ingredients for precipitation events, the moisture flux and the large-scale dynamics are implicitly described by the moisture uptake pattern. Therefore it provides concise and meaningful summary of the conditions leading to HPEs. The chapter is organised as follows. First, a classification is made using ERA-Interim in section 7.2 and then used to classify the 36 flood events. Second, the characteristics of each class are examined in more details (section 7.3).

7.2. Classification

The work presented in this section is largely based on work done by Franziska Scholder-Aemisegger, who investigated HPEs using ERA-Interim. The HPEs are defined as 3-day period for which the accumulated precipitation is above the 95 percentile leading to 330 events for the time period from 1979-2011. For each event, the moisture uptake regions are determined using 20 days backward trajec-
Chapter 7. Climatology

Figure 7.1.: Moisture uptake regions used for the classification.

...tories and the moisture uptake diagnostic tool (details in subsection 2.2.3). The moisture uptakes are then attributed to one of the 11 regions shown in Figure 7.1. The 11 dimensional phase space built with the 11 regions and the 330 events is further analysed with a K-mean algorithm forced to identify 4 clusters.

The resulting four classes are presented in Figure 7.2. The classes are ordered such that continental contribution are increasing. Class 1 (Figure 7.2a) represents events where the largest part of the moisture uptake (approximately two-third) is located in the North Atlantic and with a substantial part (13%) located in the Subtropical Atlantic (below 30°N). Class 2 (Figure 7.2b) represents events where approximately 50% of the moisture uptake is located in the North Atlantic but where the Subtropical Atlantic part is low. Class 3 (Figure 7.2c) represents events where a third of the moisture uptake is located in the Mediterranean Sea and with equal parts from the European continent and the North Atlantic (25%). Class 4 (Figure 7.2d) represents events where the majority of the moisture uptake is located over the European continent.

Similarly as for the ERA-Interim HPEs, the moisture uptake regions for the 36 flood events are determined with the 20CR dataset. Using the same clustering method as for all ERA-Interim HPEs, four classes are defined (Figure 7.3). Despite the much less robust results due to the small sample of events used for the clustering, the classes are in surprisingly good agreement with the ERA-Interim classes. The main difference are the larger contribution from the Subtropical Atlantic and African continent regions in respectively, class 1 and class 3. This suggests a more important southern origin of the moisture when flood events only
7.2. Classification

The results of the classification of the 36 events based on the ERA-Interim classes are shown in Table 7.1 and compared, for the overlapping period, with the ERA-Interim HPEs. They show that each class is represented during the 134-year period, however a larger sample of events is needed to analyse the temporal variability. For the events in the ERA-Interim period, the comparison of the classes obtained with the moisture uptakes calculated with ERA-Interim shows

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**Figure 7.2.** Classes based on the moisture uptake regions of 330 HPEs using ERA-Interim. Each class shows the fraction of moisture uptake in 6 regions, the North Atlantic (blue), the African continent (orange), the European continent (green), the Mediterranean Sea (red), the Subtropical Atlantic (violet) and all others regions (yellow). The exact definition of the regions is given in Figure 7.1.
Table 7.1.: Classification of 36 flood events in Switzerland since 1873, based on Schmocker-Fackel and Naef (2010), according to the ERA-Interim classes. The left column shows the results using moisture sources based on 20CR, the right column shows results based on ERA-Interim. Green numbers signify that both dataset are in agreement, whereas the red numbers denote disagreement.
very good agreement. The only difference is the class for the 25 September 1987 event, which is examined in further details in subsection 7.3.2.

To put the results of the 36 events into a climatological perspective, the moisture uptake regions and classes for each day of the period from 1 January 1871 to 31 December 2010 are calculated. The results, presented in Figure 7.4, show that there is significant variability in the temporal evolution of the classes. As expected, class 2 shows a higher frequency than the other classes. This is due to
the large-scale westerly flow, which is the most frequent situation in Switzerland. This is also consistent with the low frequency of the class 4, for which the European continent contributes moist to the moisture source, since either a weak large-scale flow or an easterly flow configuration are probably needed to obtain this class. Another interesting aspect is the apparent anti-correlation of classes 1 and 3. For example, during the period from 1900 to 1940, the frequency of class 1 is above 90 (days per year) and the frequency of class 3 is below 80. In contrast, after 1940 the frequency of class 3 remains above 80 whereas the frequency of class 1 stays almost exclusively below 80.

### 7.3. Description of the classes

To illustrate more specifically the characteristics of each class, this section describes each class with an event that occurred in the ERA-Interim period.
7.3. Description of the classes

7.3.1. Class 1, 16 February 1990

To illustrate the characteristics of class 1, the 16 February 1990 event is used. This event affected heavily the Jura region and to some extent the western and northeastern part of Switzerland (Figure 2.1).

As previously mentioned the first class (Figure 7.2a) describes events where the moisture originates mainly from the North Atlantic and to a non-negligible part from the Subtropical Atlantic. In this case, the North Atlantic represents slightly more than two-third (68%) of the moisture uptake regions using ERA-Interim (Figure 7.5a) and slightly more than 80% using 20CR (Figure 7.5b). The Subtropical Atlantic part is significantly larger using ERA-Interim with 22% than using 20CR with 8%.

To further investigate the differences of the moisture uptake regions, the moisture uptake patterns of both datasets are presented in Figures 7.5c,d for the 96-hour period from 00 UTC 14 to 00 UTC 18 February 1990. In both cases, the moisture uptakes occur mainly west of 20°W with, however, a larger southern component using ERA-Interim (Figure 7.5c), whereas there is a maximum north of 40°N and to the west of 40°W when using 20CR.

In order to examine the link between the moisture uptake and the large-scale dynamics, PV on 310 K and TCW, averaged over the same 96-hour period as for the moisture uptake, are shown in Figures 7.5e,f. The PV field shows that a ridge was located over the North Atlantic, as well as a weak PV trough on its western flank (centred near 40°W). A band of high TCW values is located below the PV ridge, and follows it towards Europe. The moisture uptake maximum (Figure 7.5c) is located south of the upper-level trough at the tip of a TCW band emerging from the subtropics. The occurrence of such a TCW band, emerging from the Subtropics towards Europe, can be observed in many class 1 events such as the events on 14 February 1877 (Figure B.1c), 10 March 1896 (Figure B.2c), 20 January 1910 (Figure B.3b), 26 February 1957 (Figure B.6a), and 14 May 1999 (Figure B.8c).

7.3.2. Class 2, 25 September 1987

To illustrate the second class (Figure 7.2), the 25 September 1987 event is used. This event heavily affected the southern part of Switzerland and the Jura region; in addition the central part of Switzerland was also affected (Figure 2.1). In this
Chapter 7. Climatology

Figure 7.5.: Moisture sources (a,b), moisture uptakes $[10^{-3} \text{ mm day}^{-1}]$ (c,d), PV [pvu] at 310K (e), and TCW [kg m$^{-2}$] (f) averaged over the 96-hour period from 00 UTC 14 to 00 UTC 18 February 1990 using ERA-Interim (a,c,e,f) and 20CR (b,d).
7.3. Description of the classes

(a) Moisture sources using ERA-Interim

(b) Moisture sources using 20CR

(c) Moisture uptakes using ERA-Interim

(d) Moisture uptakes using 20CR

(e) PV at 320K

(f) TCW

Figure 7.6.: Moisture sources (a,b), moisture uptakes [$10^{-3}$ mm day$^{-1}$] (c,d), PV [pvu] at 320K (e), and TCW [kg m$^{-2}$] (f) averaged over the 96-hour period from 00 UTC 23 to 00 UTC 27 September 1987 using ERA-Interim (a,c,e,f) and 20CR (b,d).
case, the class 2 identified using ERA-Interim differs from the class 1 identified using 20CR.

As previously mentioned class 2 represents events where moisture uptakes are still located for a large part in the North Atlantic, but where the European continent accounts for approximately a quarter of the overall moisture sources. Using ERA-Interim (Figure 7.6a), the North Atlantic part account for approximately 50\% of the overall regions, whereas its account for approximately 60\% using 20CR (Figure 7.6b). The European continent and Mediterranean parts are larger when using ERA-Interim (20\% and 11\%) than using 20CR (14\% and 3\%).

To further investigate the differences in the moisture uptake regions, the moisture uptake patterns are presented in Figures 7.6c,d using ERA-Interim and 20CR averaged over the 96-hour period from from 00 UTC 23 to 00 UTC 27 September 1987. They show that moisture uptake occurs mostly over Spain and further west over the North Atlantic. Using ERA-Interim (Figure 7.6c), there is a maximum centred over the north-eastern part of Spain, explaining both the larger contribution from the European continent and the Mediterranean Sea. Indeed, despite the good overall agreement of the moisture uptake pattern using 20CR, no maximum over Spain is diagnosed (Figure 7.6d). This disagreement explains the different classes identified when using ERA-Interim and 20CR. This event is therefore a representative of events that have characteristics at the interface of two classes.

The large-scale aspects of the event are shortly presented here using PV (at 320 K) and TCW averaged over the same 96-hour period as for the moisture uptake in Figures 7.6e,f. No particular upper-level feature can be distinguished from the PV field, but both the moisture uptake pattern and the TCW field suggest the presence of a south-westerly flow.

### 7.3.3. Class 3, 15 October 2000

To illustrate the third class the October 2000 event is used. As described in details in section 4.2, this event heavily affected the southern part of Switzerland (Figure 2.1).

As previously mentioned this class (Figure 7.2c) describes events where a third of the moisture uptakes is located in the North Atlantic, the European continent and the Mediterranean. For this event, a large part of the moisture originates
7.3. Description of the classes

(a) Moisture sources using ERA-Interim

(b) Moisture sources using 20CR

(c) Moisture uptakes using ERA-Interim

(d) Moisture uptakes using 20CR

(e) PV at 330K

(f) TCW

Figure 7.7.: Moisture sources (a,b), moisture uptakes [10⁻³ mm day⁻¹] (c,d), PV [pvu] at 330K (e), and TCW [kg m⁻²] (f) averaged over the 96-hour period from 00 UTC 13 to 00 UTC 17 October 2000 using ERA-Interim (a,c,e,f) and 20CR (b,d).
from the African continent, with approximately 30% using both datasets (Figures 7.7a,b). Among the other class 3 events, only the 23 September 1993 event (Figure B.8d) shows a similarly large part from the African continent. However, also for this event roughly a third originates from the Mediterranean, with 24% using ERA-Interim and 31% using 20CR. The North Atlantic part is larger using ERA-Interim (27%) than 20CR (18%), and the European continent part is approximately 10% for both datasets.

To further investigate the moisture uptake region differences, the moisture uptake patterns using both datasets are presented in Figures 7.6c,d averaged over the 96-hour period from 00 UTC 13 to 00 UTC 17 October 2000. The moisture uptake maximum centred over Sardinia is stronger using 20CR and explains the larger portion of the Mediterranean. The larger contribution of the North Atlantic using ERA-Interim is due to the band extending to the west of Spain along 40°N, which is almost not present using 20CR.

The large-scale dynamics of the event are summarized with the PV and TCW fields (Figure 7.7) averaged over the 96-hour period from 00 UTC 13 to 00 UTC 17 October 2000. The quasi-stationary PV streamer located over France is clearly visible (Figure 7.7e), as well as the northward advection of very moist air masses over Italy (Figure 7.7f). As mentioned in section 4.2, the North Atlantic uptakes may indicate the presence of Atlantic hot spots, i.e., areas of strong evaporation upstream of the PV streamer. This would also explain the difference between both dataset, since the lower resolution of the 20CR dataset both vertically and horizontally may reduce the capability of the model to represent such features.

### 7.3.4. Class 4, 23 August 2005

To illustrate the fourth class the August 2005 event is used. As described in details in chapter 3, this event heavily affected northern Switzerland (Figure 2.1).

As previously mentioned this class (Figure 7.2d) describes event where the largest portion of the moisture uptake is located above the European continent. In this case, an even higher proportion of the uptakes occurred over the European continent (Figures 7.8a,b) than indicated by the mean values of class 4. Indeed, for both datasets the contribution from the European continent is approximately 75%. In addition, the part of the North Atlantic is approximately 10% for both datasets, but the part of the Mediterranean is larger using 20CR (10%) than using ERA-
### 7.3. Description of the classes

#### Moisture sources using ERA-Interim

(a) Moisture sources using ERA-Interim

(b) Moisture sources using 20CR

(c) Moisture uptakes using ERA-Interim

(d) Moisture uptakes using 20CR

(e) PV at 330K

(f) TCW

**Figure 7.8**: Moisture sources (a,b), moisture uptakes [10⁻³ mm day⁻¹] (c,d), PV [pvu] at 330K (e), and TCW [kg m⁻²] (f) averaged over the 96-hour period from 00 UTC 21 to 00 UTC 25 August 2005 using ERA-Interim (a,c,e,f) and 20CR (b,d).
Interim (3 %).

To further examine the moisture uptakes, the uptake patterns of both datasets are shown in Figures 7.8c,d averaged over the 96-hour period from 00 UTC 21 to 00 UTC 25 August 2005. They show that a large part of the moisture uptake occurred over Eastern Europe until 40° E. The eastern extent of the moisture uptake can be also observed in other class 4 event such as the 14 June 1876 event (Figure B.1b) or the 15 June 1910 event (Figure B.3c).

The large-scale dynamics is shortly presented here using averaged PV and TCW over the 96-hour period from 00 UTC 21 to 00 UTC 25 August 2005 (Figures 7.8e,f). As previously described in chapter 3, the upper-levels are characterized by the presence of a cut-off low that is quasi-stationary over the Alps (Figure 7.8e). The cyclonic circulation induced by the cut-off at low-levels transported the high TCW values over the Balkan and Eastern Europe towards Switzerland.

7.4. Conclusion

The moisture uptake classification of the flood events is capable of capturing the large-scale dynamical characteristics of an event as shown by the case studies for each class presented above and the summary plots of the 36 flood events (Appendix B).

Moreover, the classes of the events that occurred in the period when ERA-Interim and 20CR overlap show very good agreement between the datasets. In addition, when the classes are examined in more details, the repartition of the moisture uptake in the different regions shows that similar values are obtained for both datasets. This gives good confidence that the 20CR dataset can be used to examine long-term variability of large-scale conditions leading to floods using the moisture source diagnostic tool.

Further interesting questions arise from this brief analysis that deserve more detailed investigation:

- Do we obtain similar classes when 20CR moisture uptake regions are used to determine the classes?
- How does the repartition and temporal evolution of the 20CR classes change when only days with extreme precipitation are considered?
• Is there variability in the seasonal frequency of the classes?

• Are there trends in the classes over the 20th century?
Chapter 8.

Hydrological perspective and concluding remarks

The aim of this chapter is to discuss implications of the previous meteorological investigations for hydrology and to summarize the key results presented in this thesis.

8.1. Hydrological perspective

Apart from the meteorological perspective, it is interesting to examine the precipitation response of the different catchments (Figure 8.1) for each event. This approach can shed some light on the PMP in the considered catchments by looking at the individual precipitation sensitivity to variations in the atmospheric flow conditions. The PMP question is very complex and cannot be resolved easily, and therefore only a specific portion of this question is addressed here. The following aspects will be examined in this section:

- Is there a marked difference between the precipitation response of the catchments?
- Which catchments are reacting the strongest, and for what type of events?

To answer these questions, the total precipitation is examined in six different catchments for all sensitivity experiments (Figure 8.2), and a summary of the characteristics of each catchment is presented in Table 8.1, where the maximum pre-
precipitation obtained in one of the CTRL runs is listed and compared with the maximum precipitation in all sensitivity experiments. A first distinct characteristic is the large variability of the precipitation response of the individual catchments in the different events. For example, the Reuss catchment reacts strongly during the October 2000 event, with an increase of +140 mm in the QV +30% experiment (Table 8.1). But, in contrast, during the September 1993 event, the precipitation only weakly increases in all sensitivity experiments (Figure 8.2e), despite the similarities with the October 2000 event in the large-scale flow. The precipitation responses of the Suze catchment (Figure 8.2h) during the August 2005 and August 2007 events are also interesting since in both cases the maximum values obtained are higher than during the December 1991 event, the main flood event in this catchment. Despite the different temperature evolution prior to these events, which is very important, since a frozen ground is an important ingredient for triggering floods in the Suze catchment (Felix Naef, personal communication 2014),
Figure 8.2: Total precipitation (accumulated over 72 hours except for the October 2011 event where it is accumulated over 36 hours, in mm) for all nine catchments (Figure 8.1) and sensitivity experiments in all six events. The name of the catchment is given below each panel. The coloured bars indicate the different experiments, in yellow the CTRL, in light, medium and dark blue the QV +10%, QV +20%, and QV +30%, in orange the T+1K, and in red the T+2K.
Table 8.1.: Maximum precipitation (first two columns) in any CTRL run (left) and any sensitivity experiment (right) for each catchment. Maximum increase (last two columns) during any specific humidity (left) and temperature (right) experiment compared to the CTRL run. Below each value the event is specified.

<table>
<thead>
<tr>
<th>Catchments</th>
<th>max. precipitation</th>
<th>max. increase</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CTRL</td>
<td>all</td>
</tr>
<tr>
<td>Allenbach</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>143</td>
<td>200</td>
</tr>
<tr>
<td></td>
<td>Dischma</td>
<td></td>
</tr>
<tr>
<td></td>
<td>154</td>
<td>187</td>
</tr>
<tr>
<td></td>
<td>Hinterhein</td>
<td></td>
</tr>
<tr>
<td></td>
<td>165</td>
<td>348</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Luetschine</td>
<td></td>
</tr>
<tr>
<td></td>
<td>266</td>
<td>327</td>
</tr>
<tr>
<td></td>
<td>Reuss</td>
<td></td>
</tr>
<tr>
<td></td>
<td>224</td>
<td>309</td>
</tr>
<tr>
<td></td>
<td>Saltina</td>
<td></td>
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<tr>
<td></td>
<td>444</td>
<td>456</td>
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<tr>
<td></td>
<td></td>
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</tr>
<tr>
<td></td>
<td>Schaechen</td>
<td></td>
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<tr>
<td></td>
<td>218</td>
<td>238</td>
</tr>
<tr>
<td></td>
<td>Suze</td>
<td></td>
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<tr>
<td></td>
<td>130</td>
<td>180</td>
</tr>
<tr>
<td></td>
<td>Ticino</td>
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<tr>
<td></td>
<td>242</td>
<td>362</td>
</tr>
</tbody>
</table>

This represents an important precipitation increase, in particular in August 2005 with an increase of +38% compared to December 1991 (Table 8.1). Another interesting aspect is the spatial variation of the precipitation response for a single event. For example, during the December 1991 event, the Allenbach experienced a +37% response in the QV+20% experiment (Figure 8.2a), whereas the maximum increase for any other catchment is below +15%.

The catchment that has the highest total precipitation, all sensitivity experiments considered, is the Saltina catchment with 456 mm (October 2000). However this catchment experienced also the highest total precipitation in the CTRL
run (444 mm, October 2000), and therefore the increase is only +3%. In contrast, the catchment that shows the highest increase during a specific humidity experiment is the Hinterrhein catchment with an increase of +183 mm (October 2000). This represents a doubling of the maximum CTRL run precipitation (165 mm), the highest maximum precipitation experienced in this catchment, all events included. The maximum increase for a temperature experiment occurs in the Lütschine catchment with an increase of +61 mm in December 1991 (+23%).

8.1. The Suze catchment in August 2005 and August 2007

![Graphs](image)

Figure 8.3: IWT [kg m\(^{-1}\) s\(^{-1}\)] (solid lines) and accumulated precipitation [mm] (dashed lines) for the Suze catchment in the August 2005 (a) and August 2007 (b). IWT is calculated upstream of the catchment (black dots in Figure 5.3a denote the exact location). The coloured lines represent the different experiments, CTRL in black and QV +30% in blue.

We pointed out, in the previous section, that the highest CTRL run precipitation in the Suze catchment occurred in December 1991, but the strongest increase in the sensitivity experiments occurs in the August 2005 event. In addition, the total precipitation in the CTRL run is higher in the August 2007 event than in the August 2005 event, but the August 2007 event shows a weaker increase in the sensitivity experiments. In this section we briefly examine the precipitation in both the August 2005 and August 2007 events and look at the differences between them.

A first major difference is the values of the incoming IWT upstream of the Suze catchment (Figure 8.3). Interestingly the IWT in August 2005 is larger than 300 kg m\(^{-1}\) s\(^{-1}\) for a 36-hour period from 00 UTC 21 to 12 UTC 22 August 2005, whereas it stays below 300 kg m\(^{-1}\) s\(^{-1}\) during the whole event in August 2007.
However, the IWT increase produced in the QV+30% experiment is stronger in the August 2005 event than in the August 2007 event.

Figure 8.4.: Cross-sections (red box in Figure 5.3a) aligned on the Suze catchment for the 12-hour period from 00 UTC to 12 UTC 21 August 2005 (a,b) and the 12-hour period from 00 UTC to 12 UTC 09 August 2007 (c,d). Each small panel shows 3-hourly average fields over the width of the red box. The coloured contours show various hydrometeors. Cyan stands for ice in 10mgkg$^{-1}$ intervals, blue for snow in 250mgkg$^{-1}$ intervals, grey for graupel in 50mgkg$^{-1}$ intervals, red for cloud water in 100mgkg$^{-1}$ intervals, and green for rain in 50mgkg$^{-1}$ intervals. The coloured bar indicates the location of the Suze catchment (brown). The grey shading shows the averaged topography. The control runs are shown in (a,c) and the specific humidity experiments QV+30% in (b,d).

To analyse in more detail the processes leading to precipitation in the Suze catchment, cross-sections aligned on the catchment are shown in Figure 8.4 for a 12-hour period when precipitation changes between CTRL and QV+30% are large. The cross-sections cover the 12-hour periods from respectively 00 UTC to
12 UTC 21 August 2005 and 00 UTC to 12 UTC 09 August 2007. The CTRL run precipitation during those periods is similar in both cases (25 mm). In both cases, precipitation is formed by the seeder-feeder mechanism and by melting of snow, and therefore the precipitation intensity depends largely on the snow concentrations. For both cases, in the QV +30% experiment, the higher snow concentration increases the precipitation intensity and extends the lifetime of the snow layer and the precipitation period. However, the precipitation response is stronger in August 2005 than in August 2007. Indeed, as already pointed out by the IWT analysis, the major difference in the QV +30% experiments comes from the strong increase of the snow concentration in August 2005.

8.1.2. The Reuss and Hinterrhein catchment in September 1993 and October 2000

September 1993 and October 2000 are two other interesting events with similar large-scale flows producing precipitation centred on the southern side of the Alps. During these events, there are marked differences between the precipitation responses of the Hinterrhein and the Reuss catchments (Figures 8.2c,e). The Hinterrhein catchment shows precipitation increases in the sensitivity experiments for both events. The precipitation increase is stronger in October 2000 with +183 mm in the QV +30% than in September 1993 with +112 mm. In contrast in the Reuss catchment there is a significant precipitation increase only in October 2000, with +114 mm in the QV +30% experiment.

To briefly analyse the main difference between the two events, we present the total precipitation for the CTRL run and the QV +30% experiment of both events in Figure 8.5. In September 1993 during the CTRL run, the area with the highest precipitation is located at the foot hills of the Alps, mostly at low elevations (Figure 8.5a). A band with high precipitation values extends southward near 8.7°E with a maximum over the Northern Apennines. Both the Hinterrhein (pink contours) and the Reuss catchment (light purple) are located at the boundary of the area with maximum precipitation. In the QV +30% experiment (Figure 8.5c), the maximum precipitation area extends eastward, to about 10°E. The precipitation band extending southward moves also eastward and is now located...
near 9.3°E. Precipitation is strongly increased in the whole band. As the precipitation increases mainly at low elevations, the Reuss catchment receives similar precipitation in the QV +30% experiment than in the CTRL run. However, the Hinterrhein catchment located further south is now located in the maximum precipitation area of the QV +30% experiment.

In October 2000, the area with the maximum precipitation extends to higher elevations than in September 1993, as observable near the Saltina catchment (red contours in Figures 8.5a,b). The precipitation extends also further west to 7°E. Both the Hinterrhein and the Reuss catchment are located at the boundary of the area with maximum precipitation. In the QV +30% experiment Figure 8.5d), precipitation extends eastward, similar to the September 1993 event. Both the Hinterrhein and Reuss catchment are now located in the area with the maximum precipitation.
In summary, in both events, the QV +30% experiment produces a precipitation increase at low elevations and particularly to the east of the CTRL run maximum. The different precipitation responses observed in the Hinterrhein and Reuss catchments are due to the extent of the area with intense precipitation. In the October 2000 event, the precipitation extends further north (towards higher altitudes) than in the September 1993 event and therefore covers the Reuss catchment.

8.1.3. Conclusion

In conclusion, we see clearly that the precipitation responses due to increased specific humidity, ranging from +18 to +183 mm (Table 8.1), are stronger than the increase in the temperature experiments, ranging from +22 to +61 mm (Table 8.1), consistent for all catchments. In addition, whereas some catchments seem to have reached their PMP in the CTRL run (e.g., the Saltina catchment in October 2000), other catchments in the same event experiences a strong increase of their maximum precipitation (e.g., Hinterrhein catchment in October 2000). Therefore, the investigation of the precipitation sensitivity, in a PMP perspective, should be conducted individually for each catchment and should take great care of the role of the small scale variability for the formation of precipitation.

8.2. Concluding remarks

The results from the sensitivity experiments show that from the three main ingredients for orographic precipitation introduced in chapter 1,

(a) The horizontal transport of moist air masses by synoptic scale weather systems.

(b) The saturation of the air mass through terrain-forced ascent governed, on the mesoscale, by the stability of the air mass.

(c) The transformation of the condensate to precipitable hydrometeors on the microscale.

Mainly the last ingredient, the microphysical aspect, shows a significant and systematic effect on the precipitation response. The moisture influx effect is mainly
leading to an increase of the duration of periods with high IWT values rather than of peak values. The role of the stability is crucial for determining the type of flow (blocked or unblocked) and therefore the formation of the precipitation, but does not show a strong systematic response in the sensitivity experiments.

The sensitivity of the microphysical processes can be roughly classified in two main categories, the unblocked (Figure 8.6) and the blocked (Figure 8.7) flow categories, which is coherent with the model presented by Medina and Houze (2003) (chapter 1).

The first category (Figure 8.6) describes the mechanisms governing the precipitation response in principally unblocked flow situations, when embedded convection occurs. Embedded convection often develops over the first peak of a moun-
tain range producing precipitation by riming, since high liquid water concentration and graupel are formed by the strong updrafts (Figures 8.6a,c). Two subtypes can be determined, one which mainly concerns the low elevations (hereafter referred as type "unblocked low") and one which affects precipitation at high altitudes (type "unblocked high"). Typically for the type unblocked low, convection already occurs in the CTRL run and produces precipitation above the lower elevations (Figure 8.6a). In this case, the main impact of the sensitivity experiments is to reinforce the convective activity above the first peak and therefore to increase principally the graupel concentration (Figure 8.6b). The graupel then efficiently removes moisture from the atmosphere and reduces the available moisture downstream (towards high elevations). The effect is to increase the precipitation at low elevations without influencing precipitation at high altitudes. This type of response occurs in the August 2005 (chapter 3) and August 2007 (section 5.3) events.

In the unblocked high type, embedded convection also occurs over the first peak, but precipitation is also formed downstream by melting of snow (Figure 8.6c). Similar to the unblocked low type, embedded convection is also enhanced over the first peak during the sensitivity experiments through the increased the graupel concentration (Figure 8.6b). However, the enhanced precipitation efficiency above the first peak reduces the snow concentration downstream and therefore reduces the precipitation at high altitude. The effect is to reduce precipitation at high altitudes while precipitation at low altitudes is increased. This type of response occurs principally in the September 1993 (section 5.2) and October 2000 (section 4.2) events.

The second category (Figure 8.7) describes the mechanisms governing the precipitation response during principally blocked flow situations, when precipitation formation is mostly due to deposition of vapour on ice particles or riming in small cells. The mechanisms described by this category are more relevant for the specific humidity experiments which produce a stronger differences in the snow layer than the temperature experiments. In addition, two subtypes can be defined, one which affects mainly the low altitudes (hereafter referred as type "blocked low"), and one which affects the high altitudes (type "blocked high"). In type blocked low, snow is formed by vapour deposition through the moderate mesoscale ascent over the mountain range (Figure 8.7a), graupel is formed mainly
Chapter 8. Hydrological perspective and concluding remarks

Figure 8.7.: Schematic of the microphysical processes involved in the precipitation formation for blocked flow. The left columns display the situation in the CTRL run, the right column the situation in sensitivity experiments with increased specific humidity or temperature. Type “low” depicts situations with precipitation responses at low altitudes and type “high” depicts situations with precipitation responses at high altitudes.

close to the highest peaks where updrafts are stronger and where cloud water is advected above the melting level, and rain is formed below the graupel maxima by melting of graupel and snow. In the sensitivity experiments, the snow and to some extent graupel concentrations are increased enhancing precipitation formation in the direction upstream of the mountain range (Figure 8.7b), towards the lower altitudes. In the same time, at high elevations, precipitation remains similar since precipitation efficiency is already high in the CTRL run. The effect is to increase precipitation at low altitude while precipitation at high altitude remains unchanged. This type of response occurs in the December 1991 (section 5.1), September 1993 (section 5.2), October 2000 (section 4.2), August 2005 (chapter 3), and October 2011 (section 5.4) events.
Figure 8.8: Total precipitation at each grid point for each sensitivity experiments of all events. Results are grouped by sensitivity experiments and altitudes classes (defined every 500 m from 500 m to 3500 m) and displayed in the form of a box plot. Each box plot extend from the lower to the upper quartile of the total precipitation, the median is marked by the dark line, and the whiskers cover the 5 to 95 percentile interval. The precipitation difference between the median value of the CTRL run and the QV +30% experiment is displayed in percent at the top of each altitude class. The colours represent the different experiments, CTRL in cream, QV +10% in light blue, QV +20% in blue, QV +30% in dark blue, T +1K in yellow, and T +2K in red.

In type blocked high, snow and graupel are formed similarly to the blocked low situation, but the lifting occurs typically further downstream due to the presence of, for example, a barrier jet, an aspect that has not been investigated in detail in this thesis. In the sensitivity experiments, the snow concentration is increased and advected to the lee side of the main peak where it melts and increases precipitation in this region. The effect is to principally increase precipitation in the lee of the mountain at high elevations. This type of response occurs in the September 1993 (section 5.2), October 2000 (section 4.2), and October 2011 (section 5.4) events.

In summary, it appears that the mechanisms described above lead, on average, to a stronger increase of the precipitation at low elevations than at high elevations, particularly in the specific humidity sensitivity experiments. This is summarized in Figure 8.8, which shows the distribution of total precipitation at each grid point.
for each event grouped by elevation and sensitivity type. For example, there is an increase in precipitation between the CTRL run and the QV +30% experiment of +63% in the lowest bin (500–1000m), whereas there is only a +13% increase in the highest bin (3000–3500m). Interestingly for the three middle bins (1000–3000m), where most of the important catchments in Switzerland are located, the precipitation changes are linearly following the specific humidity increases. The precipitations changes in the temperature sensitivity experiments are overall weaker and do not vary with elevation; in all bins the precipitations change is approximately +15% for a warming of T +2K.

However, despite those clear general results, care is needed when analysing individual catchments and events, since the small scale variability is large in a region with such a complex topography and very sensible to small variations of the atmospheric conditions.

8.3. Critical reflection and outlook

8.3.1. Critical reflection

This study used a particular approach to investigate how changes in environmental conditions can affect Alpine precipitation. However, there are also fundamentally different research methods to address this complex question. For example, a very important research area is the analysis of the changing character of precipitation due to climate change (Trenberth et al., 2003). Studies in this research area are based on statistical analyse of climate simulations for long periods, typically more than 10 years. Commonly used diagnostics are the highest percentiles of hourly and daily precipitation, which are used to investigate changes in the frequency and intensity of the precipitation episode (Ban et al., 2015). Among other things, these studies have shown that the hydrological cycle will change under a warmer climate (Allen and Ingram, 2002), and that in particular precipitation extremes are projected to become more frequent and intense (Christensen and Christensen, 2003; Fischer and Knutti, 2015; Ban et al., 2015).

The main advantage of the case study approach used in this thesis is the capacity to examine precisely the involved physical processes. Indeed, due to the relatively small number of case studies and simulations, it is possible to examine the rele-
vant large-scale, mesoscale and microscale aspects for each case. This then allows an in-depth understanding of the physical processes that determine the precipitation response for each individual case and catchment. Another advantage of the approach used here, namely the 3 ingredients approach, is to reduce the complexity of the orographic precipitation and to allow for a systematic and similar analysis of the individual sensitivity experiments.

However at the same time, this case study approach implies that the results are only valid for a limited set of cases and therefore cannot be generalized for all precipitation events. In particular, the results obtained during this thesis are based on extreme events and may significantly differ for events with averaged precipitation. Another caveat is due to the fact that a single model was used with a single microphysical scheme. This could affect the results as they are particularly dependent on cloud microphysics. Moreover the four concepts derived in this thesis are not based on a totally objective method, since they are based on a qualitative analysis of figures rather than on a quantitative evaluation of the model output. Finally the choice of the atmospheric parameters that are modified in this study to perform the sensitivity experiments can be questioned, in particular in a future climate perspective, for which other parameters might be more relevant.

One important achievement of this work is the demonstration of the complexity of the precipitation response in complicated topography and in particular the stronger precipitation response at low elevation than at high elevations. This shows the importance of understanding the individual characteristics of the catchments for estimating the PMF.

8.3.2. Outlook

They are three main areas where the research performed in this thesis could be meaningfully extended:

- This work aimed at characterizing the atmospheric response to synthetic changes in specific humidity and temperature. However in order to better understand the characteristics of floods and extreme precipitation events, the analysis of the sensitivity of the precipitation intensity and spatial distribution needs to be extended to other atmospheric parameters. For example, another crucial ingredient for orographic precipitation is the wind. There-
fore, the analysis of the precipitation sensitivity to wind speed and direction is highly relevant to quantify the PMP in very complex topographical region such as the Alps.

• As mentioned above, the four concepts derived from the results of this work are not reliable enough for a general purpose. Further investigations using semi-idealized 2D simulation, where a special focus is made on the microphysical processes, might be required.

• In order to obtain results that are statistically significant, there is a need to perform this type of experiments in a ensemble framework, where a larger number of cases could be simulated. This type of analysis could also be a good option to examine the sensitivity of the precipitation events with averaged precipitation.
Appendix A.

Case summary

This appendix presents further precipitation figures for all 6 events described in this thesis. Each section contains, a precipitation evolution diagram for the main catchment of the event and coloured according to the type of precipitating hydrometeors. Then the accumulated precipitation for RhiresD and the COSMO CTRL run are shown for the whole simulation period. Finally the 6-hourly accumulated precipitation is shown for the whole simulation period, based on the COSMO CTRL run.

A.1. December 1991

Overview
Appendix A. Case summary

Figure A.1.: (a) Precipitation evolution in the Suze catchment. The bar plots show the 15 min accumulated precipitation with rain in green, snow in blue and graupel in grey from 12 UTC 20 to 12 UTC 23 December 1991. The yellow and red lines show the accumulated precipitation in the catchment and over the whole Switzerland respectively. The dashed line shows the run-off measurement (in mm (15min)$^{-1}$) at the official FOEN station. (b) Observed precipitation based on RhiresD (MeteoSwiss, 2011) and (c) simulated precipitation from the control run. The accumulated precipitation covers the 06 UTC 21 to 06 UTC 23 December 1991 period.
6-hourly precipitation

Figure A.2.: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 12 UTC 20 to 12 UTC 23 December 1991. Continued on the next page.
Figure A.2.: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 12 UTC 20 to 12 UTC 23 December 1991.
Figure A.3.: (a) Precipitation evolution in the Saltina catchment. The bar plots show the 15 min accumulated precipitation with rain in green, snow in blue and graupel in grey from 06 UTC 22 to 06 UTC 25 September 1993. The yellow and red lines show the accumulated precipitation in the catchment and over the whole Switzerland respectively. The dashed line shows the run-off measurement (in mm (15 min)$^{-1}$) at the official FOEN station. (b) Observed precipitation based on RhiresD (MeteoSwiss, 2011) and (c) simulated precipitation from the control run. The accumulated precipitation covers the 06 UTC 22 to 06 UTC 25 September 1993 period.
6-hourly precipitation

Figure A.4.: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 06 UTC 22 to 06 UTC 25 September 1993. Continued on the next page.
Figure A.4.: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 06 UTC 22 to 06 UTC 25 September 1993.
A.3. October 2000

Overview

Figure A.5.: (a) Precipitation evolution in the Saltina catchment. The bar plots show the 15 min accumulated precipitation with rain in green, snow in blue and graupel in grey from 00 UTC 13 to 00 UTC 16 October 2000. The yellow and red lines show the accumulated precipitation in the catchment and over the whole Switzerland respectively. The dashed line shows the run-off measurement (in mm (15min)$^{-1}$) at the official FOEN station. (b) Observed precipitation based on RhiresD (MeteoSwiss, 2011) and (c) simulated precipitation from the control run. The accumulated precipitation covers the 06 UTC 13 to 06 UTC 15 October 2000 period.
6-hourly precipitation

Figure A.6: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 00 UTC 13 to 00 UTC 16 October 2000. Continued on the next page.
Figure A.6.: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 00 UTC 13 to 00 UTC 16 October 2000.
A.4. August 2005

Overview

Figure A.7.: (a) Precipitation evolution in the Luetschine catchment The bar plots show the 15 min accumulated precipitation with rain in green, snow in blue and graupel in grey from 12 UTC 20 to 12 UTC 23 August 2005. The yellow and red lines show the accumulated precipitation in the catchment and over the whole Switzerland respectively. The dashed line show the run-off measurement (in mm (15min)$^{-1}$) at the official FOEN station. (b) Observed precipitation based on RhiresD (MeteoSwiss, 2011) and (c) simulated precipitation from the control run. The accumulated precipitation covers the 06 UTC 21 to 06 UTC 23 August 2005 period.
Appendix A. Case summary

6-hourly precipitation

Figure A.8: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 12 UTC 20 to 12 UTC 23 August 2005. Continued on the next page.
Figure A.8.: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 12 UTC 20 to 12 UTC 23 August 2005.
A.5. August 2007

Overview

![Figure A.9:](image)

(a) Simulated precipitation, Luetschine
(b) Observed precipitation, RhiresD
(c) Simulated precipitation, COSMO

Figure A.9.: (a) Precipitation evolution in the Luetschine catchment The bar plots show the 15 min accumulated precipitation with rain in green, snow in blue and graupel in grey from 12 UTC 06 to 12 UTC 09 August 2007. The yellow and red lines show the accumulated precipitation in the catchment and over the whole Switzerland respectively. The dashed line show the run-off measurement (in mm (15min)$^{-1}$) at the official FOEN station. (b) Observed precipitation based on RhiresD (MeteoSwiss, 2011) and (c) simulated precipitation from the control run. The accumulated precipitation covers the 06 UTC 07 to 06 UTC 09 August 2007 period.
6-hourly precipitation

Figure A.10.: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 12 UTC 06 to 12 UTC 09 August 2007. Continued on the next page.
Figure A.10.: 6-hourly accumulated precipitation from the control run. The panels cover the 72-hour period from 12 UTC 06 to 12 UTC 09 August 2007.
A.6. October 2011

Overview

Figure A.11.: (a) Precipitation evolution in the Luetschine catchment The bar plots show the 15 min accumulated precipitation with rain in green, snow in blue and graupel in grey from 00 UTC 09 to 12 UTC 10 October 2011. The yellow and red lines show the accumulated precipitation in the catchment and over the whole Switzerland respectively. The dashed line show the run-off measurement (in mm (15min)$^{-1}$) at the official FOEN station. (b) Observed precipitation based on RhiresD (MeteoSwiss, 2011) and (c) simulated precipitation from the control run. The accumulated precipitation covers the 06 UTC 09 to 06 UTC 10 October 2011 period.
6-hourly precipitation

Figure A.12.: 6-hourly accumulated precipitation from the control run. The panels cover the 36-hour period from 00 UTC 09 to 12 UTC 10 October 2011.
Appendix B. 

Climatology

This appendix show TCW [kg m\(^{-2}\)] (left panels), moisture uptake [10\(^{-3}\) mm day\(^{-1}\)] (middle panels), and the moisture uptake regions (right panels) for all 36 floods events introduced in subsection 2.1.1. Each colour in the moisture uptake regions figures represents a region, the North Atlantic in blue, the African continent in orange, the European continent in green, the Mediterranean Sea in red, the Subtropical Atlantic in violet, and all others regions in yellow. The date of the event, and its class based on the ERA-Interim classification (details in chapter 7) are noted bellow each panel row.
Appendix B. Climatology

Figure B.1. (a) 01 August 1874, Class 2
(b) 14 June 1876, Class 4
(c) 14 February 1877, Class 1
(d) 31 August 1881, Class 2

Moisture uptake regions
Figure B.2.
Appendix B. Climatology

Moisture uptake regions

(a) 26 August 1900, Class 3

(b) 20 January 1910, Class 1

(c) 15 June 1910, Class 4

(d) 23 September 1920, Class 3

Figure B.3.
Figure B.4.
Figure B.5. Moisture uptake regions

(a) 10 August 1948, Class 2

(b) 08 August 1951, Class 2

(c) 26 June 1953, Class 3

(d) 22 August 1954, Class 2
Figure B.6.
Appendix B. Climatology

Moisture uptake regions

(a) 31 July 1977, Class 4

Moisture uptake regions

(b) 08 August 1978, Class 2

Moisture uptake regions

(c) 19 July 1987, Class 2

Moisture uptake regions

(d) 24 August 1987, Class 3

TCW

Moisture uptake

Figure B.7.
Figure B.8.

(a) 25 September 1987, Class 1

(b) 16 February 1990, Class 1

(c) 14 May 1999, Class 1

(d) 23 September 1993, Class 3

Moisture uptake

Moisture uptake regions

TCW

TCW

TCW

TCW
Appendix B. Climatology

Moisture uptake regions
(a) 15 October 2000, Class 3

Moisture uptake regions
(b) 12 August 2002, Class 4

Moisture uptake regions
(c) 23 August 2005, Class 4

Moisture uptake regions
(d) 08 August 2007, Class 4

Figure B.9.
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Bibliography


Wernli, H., Pfahl, S., Trentmann, J., and Zimmer, M. 2010. How representative were the meteorological conditions during the COPS field experiment in sum-


Acronyms

$\theta_e$ equivalent potential temperature.

20CR Twentieth Century Dataset.

AR Atmospheric River.

COSMO COnsortium for Small-scale MOdeling.

ECMWF European Center for Medium-range Weather Forecasts.

ERA-Interim ECMWF Interim Re-Analysis dataset (1979-present).

FOEN Swiss Federal Office for the Environment.

HPE heavy precipitation event.

IWT integrated water transport.

LAGRANTO Lagrangian Analysis Tool.

PMF probable maximum flood.

PMP probable maximum precipitation.

PV potential vorticity.

RhiresD daily precipitation dataset in Switzerland from MeteoSwiss.

TCW total column water.

TME Tropical Moisture Export.

WCB Warm Conveyor Belt.
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