Integrated Assessment **1:** 281–299, 2000.

Climate dynamics and extreme precipitation and flood events in Central Europe

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Past changes and possible future variations in the nature of extreme precipitation and flood events in Central Europe and the Alpine region are examined from a physical standpoint. An overview is given of the following key contributory physical processes: (1) the variability of the large-scale atmospheric flow and the associated changes of the North-Atlantic storm track; (2) the feedback process between climate warming and the water cycle, and in particular the potential for more frequent heavy precipitation events; and (3) the catchment-scale hydrological processes associated with variations in major river flooding events and that are related to land-use changes, river training measures, and shifts in the proportion of rain to snowfall. In this context an account is provided of the possible future forecasting and warning methodologies based upon high-resolution weather prediction and runoff models. Also consideration is given to the detectability of past (future) changes in observed (modeled) extreme events. It is shown that their rarity and natural fluctuation largely impedes a detection of systematic variations. These effects restrict trend analysis of such events to return periods of below a few months. An illustration using daily precipitation from the Swiss Alps does yield evidence for pronounced trends of intense precipitation events (return period 30 days), while trends of stronger event classes are not detectable (but nevertheless can not be excluded). The small detection probability for extreme events limits possible mitigation of future damage costs through an abatement of climate change alone, and points to the desirability of developing improved early forecasting/warning systems as an additional no-regret strategy.

Keywords: climate change impacts, water cycle, Runoff processes, climate trends, flood forecasting, Alpine region, Rhine

1. Introduction

Extreme precipitation events constitute an integral element of the climatic conditions of a region. Their occurrence and intensity influences the viable habitat available for humankind and ecosystems, contributes to shaping the landscape through erosion processes, and can instigate major floods and hence determine the type, location and dimensions of civil infrastructures.

In the Alpine region extreme precipitation is a major environmental factor for several related reasons. First, the region is exposed to a higher frequency of such events with heavy precipitation along the southern rim of the Alps amounting to more than ten times that over the European continental flatland [1,2]. This spatial difference is related to topographic effects such as the formation of orographic cyclones and vortices, precipitation enhancement upstream and over the topography, and orographic triggering of convection and thunderstorms (see, e.g. [3]). Second, mountainous terrain is particularly vulnerable to extreme precipitation with a multitude of adverse secondary effects that include flooding of populated valley floors, erosion of slanted farming land, and avalanches, debris flow and land slides endangering exposed settlements that are often isolated and confined to the valley floor. Moreover, excessive runoff from sustained intense precipitation can even affect and seriously damage riverbeds in remote flatland areas. Third, the Alpine region connects major European industrial centers with expensive but vulnerable transport and telecommunication networks, and gas and electric power lines. Damage to this infrastructure in remote mountain regions results in logistically challenging and costly repair, disruption to industrial production and trading activities, and a knock-on effect upon a range of other sectors.

In contradistinction to the foregoing adverse effects there are also positive effects to heavy precipitation events. Both Alpine hydroelectric power production and freshwater and ground water resources in Central Europe benefit from such events. Indeed about 40% of the average annual precipitation amounts in the Alpine region result from intense precipitation events that occur on average on only 10 days per year [1].

A systematic compilation of damages and costs from hydrological extremes is not generally available for the entire Alpine region. An indication of the impact upon the Swiss economy has been derived from a study of more than a 1000 heavy precipitation events (floods, sudden snow melt, and water triggered land slides, but excluding avalanches, hail damage and droughts) during the period 1972–1996 [4]. The country has roughly 15% of the Alpine area, and the estimate burden amounts to an average of insured, non-insured and non-insurable costs of 120 million Euro (183 million Swiss Francs) per year and a total of 112 fatalities. In the assessment of these statistics it should also be noted that most of the damage was experienced in few particularly susceptible areas, mostly communities/cantons in difficult economic situations [4], and that these regions are already burdened with

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Figure 1. Car buried in mud after the devastating flash flood of the river Saltina in the town of Brig, Switzerland on 24 September 1993. (Photo: Coffrini, Sonntags Zeitung of 26 September 1993.)

investments for the maintenance and establishment of preventive measures.

It is also noteworthy that few particularly severe events dominate the average costs. The two most damaging years (1987 and 1993) alone accounted for half of the costs from the 25 year period. One episode of heavy persistent precipitation in July 1987 affected the upper Rheuss valley and damaged the trans-Alpine transport channels across the Gotthard pass – one of the main Alpine transits. In the canton of Uri, with a population of 34 000, the event incurred a cost estimated at 340 million Euro [4]. These impacts have been further emphasized in series of recent episodes that include the following events: Vaison-la-Romaine flood in September 1992 [5]; flooding of Brig in September 1993 [6] (see also figure 1); Piedmont floods in November 1994 [7]; heavy snowfalls and avalanches in Austria and Switzerland in February 1999; and large-scale flooding in Switzerland in May 1999 [8].

The significance of extreme precipitation and flooding events for Central Europe in general and the Alps prompts consideration of changes in the frequency of their occurrence particularly in connection with anthropogenic greenhouse-gas changes to the global climate system, and the effect of changes in land-use and river hydrology. Here we assess past changes and possible future variations in the occurrence of such events in Central Europe, by reviewing the relevant physical climatic processes and illustrating some of the associated inherent limitations and uncertainties. The format followed is to overview sequentially the pertinent large-scale climate dynamics (section 2), the water-cycle feedbacks (section 3), and the hydrological processes (section 4). Thereafter the constraints upon detecting long-term trends of extremes are identified (section 5), results presented from a trend analyses of Alpine heavy precipitation (section 6), an assessment provided of possible future forecasting and warning methodologies (section 7), and finally a synthesis including the implications for policy analysis (section 8).

2. Links to the large-scale atmospheric flow

Extreme weather events that occur in the vicinity of the Alps are perforce embedded within the large-scale ambient atmospheric flow. This prompts a sequence of questions:

(i) Does the large-scale flow influence the occurrence of extreme events?

(ii) What is the physical nature of the linkage(s) between the large-scale flow and extreme events, and how is the influence manifested in terms of the climatology (frequency, intensity, spatial scale and location) of the events?

(iii) Would a global climate change modify the climatology of extreme events, and could it also spawn entirely different types of events?

The focus here is on the large-scale atmospheric flow of the Atlantic-European sector and on weather events in the Alpine region. The large-scale atmospheric flow can be viewed in terms of its time-mean (monthly or seasonally averaged) and transient components. The *time-mean* fields

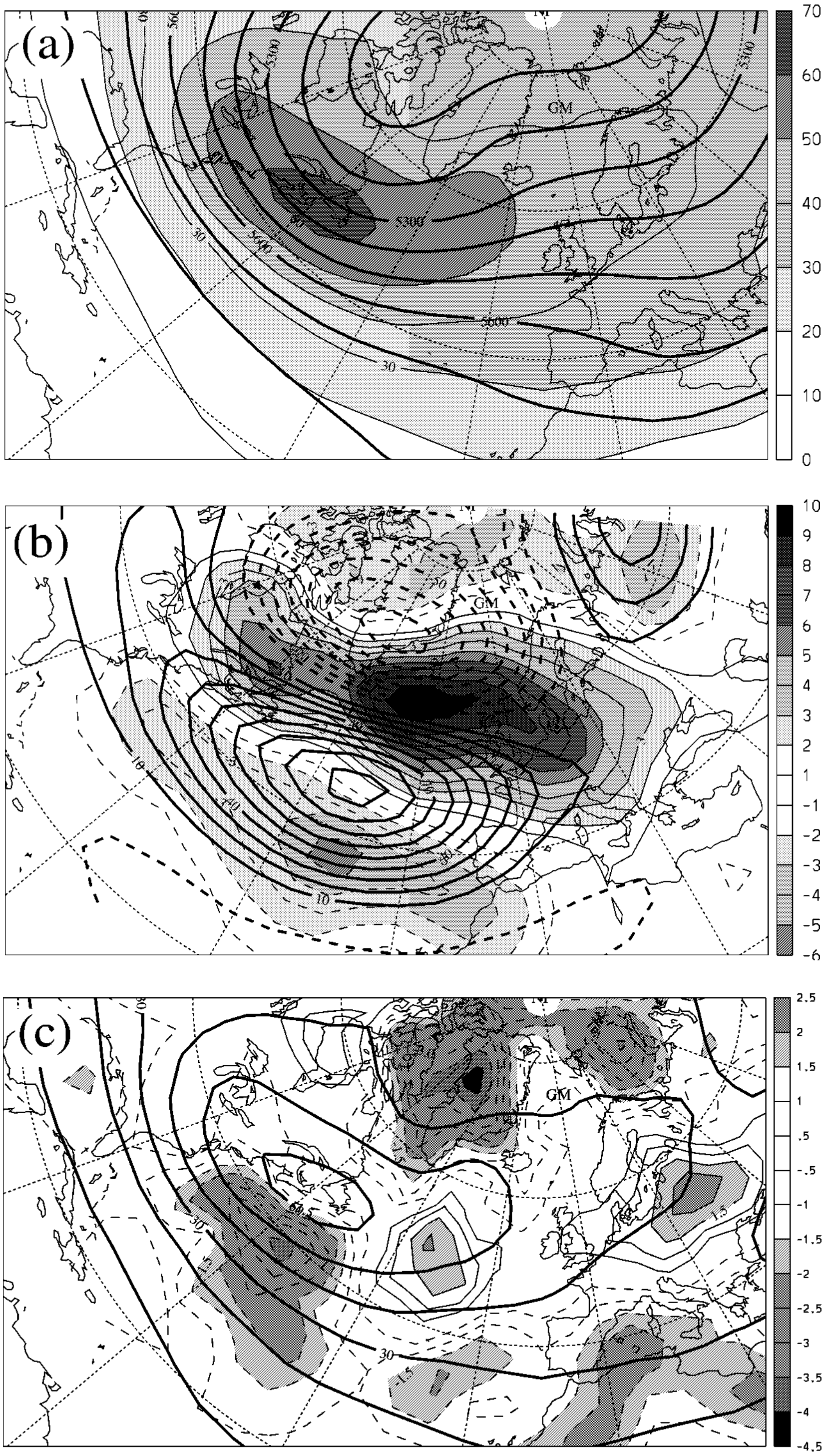

Figure 2. Depiction of the winter season time-averaged and transient atmospheric flow features for the three decades 1962–1992 at 500 hPa: (a) the mean geopotential height field and the storm track pattern; (b) an NAO-like co-variation of the height and storm track departures from the mean. In these panels the height field is portrayed with isolines and the storm track signals are shaded. (c) Trend (shaded) and time-mean (isolines) of the storm track field over the specified time-period. Storm track patterns are calculated using band-pass (2.5–8 days) filtered 500 hPa geopotential fields (see [18]). (Units: geopotential meters.)

form circumpolar patterns with superimposed wave-like variations (figure 2(a)), and the Alps are located downstream of a major trough over the Western Atlantic. The *transient* fields are dominated by low (cyclonic) pressure systems that develop off the eastern seaboard of North America and on the south side of the Alps – regions coincident with the troughs in the time-mean field. The Atlantic systems traverse eastward to form the so-called storm-track (figure 2(a)), and the Alps are located downstream and equatorward of this track.

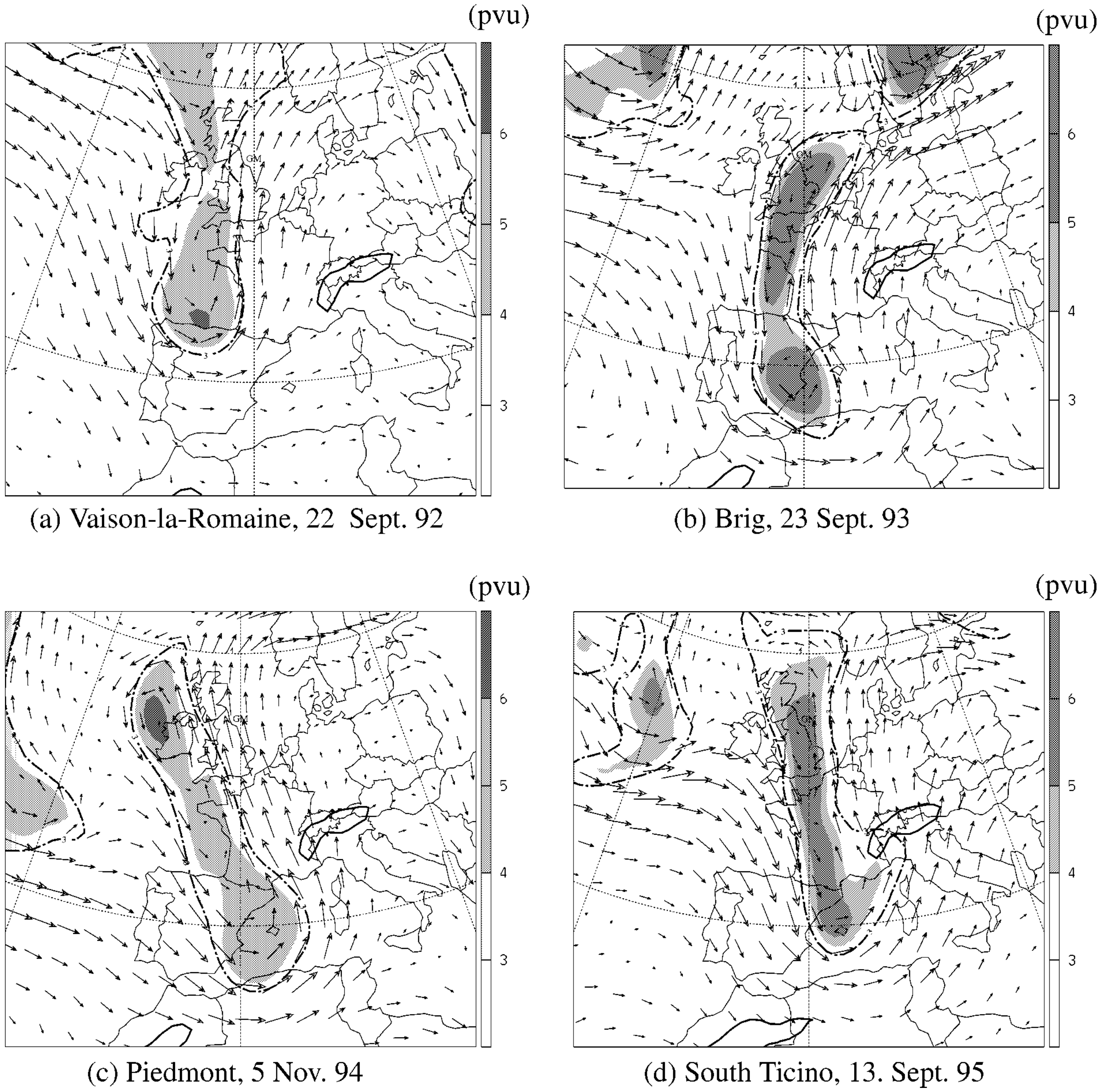

Figure 3. Illustrations of the upper-tropospheric flow structure that preceded four events of heavy precipitation on the Alpine southside. Each panel shows the wind vectors of the flow field and the presence of a streamer (shaded, potential vorticity) on a tropopause-level surface (200 hPa, about 11.5 km). (From [12].)

Both the time-mean and transient components undergo a strong seasonal cycle, e.g., the Atlantic storm track is stronger and located more poleward during winter. There are also distinctive interannual and decadal variations of the time-mean fields in the form of distinct teleconnection patterns, i.e., spatially coherent anomalies that undergo synchronous but irregular reversal in sign [9]. In the Atlantic-European sector the most prominent teleconnection is the North Atlantic Oscillation (NAO) (see, e.g. [10]). It is characterized by departures away from the time-mean that are of opposing sign and located over the Icelandic and the Azores regions (figure 2(b)). Its positive (negative) phase corresponds to reduced (increased) pressure over Iceland and a corresponding enhancement (reduction) in the strength of the westerly flow in mid-Atlantic, and a realignment and extension (reduction) of the storm track toward (away from) Scandinavia (figure 2(b)). The NAO phases correlate with the temperature (but not the precipitation) of the Alpine region.

Now we turn to consider question (i). There is abundant circumstantial evidence linking extreme European weather events with the large-scale flow. Rapid heavy rain and wind storms on the Alpine northside [11] and prolonged heavy rain events on the southside [12] are usually accompanied by strong transient flow features in the upper troposphere referred to as streamers (see figure 3). Again prolonged dry spells over the Iberian peninsula and Morocco correlate notably with sustained positive phases of the NAO with its high (anticyclonic) pressure pattern over the Azores [13,14]. Indeed the end of an extended drought over the Iberian peninsula in 1995 coincided with a notable reversal in phase of the NAO (see, e.g. [15]).

For consideration of question (ii) the foregoing examples serve to illustrate two types of physical linkages between the large-scale flow and extreme events. First localized short-time scale *storms* are linked to individual transient features in the upper-level flow. In particular a storm's intensity is linked to the amplitude of the feature, its duration is established essentially by the feature's movement across the Alpine region, and its location influenced by both the location of the feature aloft and the Alpine terrain. Second regional and sub-continental *spells* are linked to sustained anticyclonic flow features with their intensity, duration and location determined principally by the large-scale flow. Thus this simple categorization couples storms with transient features of the large-scale flow and spells to sustained features.

There are however more subtle linkages. For example the sequence of heavy Alpine snow "storms" of the 1998–1999 winter was associated with an anomalous strong and persistent north-westerly flow. Also the transient and sustained features of the large-scale flow are themselves interlinked, and there are indications that the time-mean fields in the region of cyclogenesis influence the structure of the maturing and matured cyclones downstream. In particular the appearance of a streamer over the Eastern Atlantic is associated with more anticyclonic time-mean environment in mid-Atlantic, and the streamer itself accounts for events of heavy rainstorms on the Alpine southside. In effect there is a "phenomena-chain" linking the far-field time-mean pattern to the occurrence of local extreme events.

Question (iii) relates to the impact of a *global* climate change upon *Alpine* extreme events. Here we discuss separately the *in situ* and far-field effects. *In situ* effects will result from a mere change in the temperature and humidity of the airflow and weather systems incident upon the Alps. A temperature change would yield a change in the frequency distribution of warm and cold spells, and additional moisture flux into the region coupled with orographically-induced precipitation would favor increased severity of rainstorms. Such an effect has been calibrated in model simulations of the regional climate [16]. Note however that a warmer incident atmosphere and increased moisture flux would have a contrasting effect upon snowstorms, and the net impact upon the snowfall (and its subsequent melt) is more difficult to assess.

For the *far-field* effects the impact will hinge critically upon whether climate change will manifest itself via a change in the frequency of the atmosphere's natural interannual teleconnection patterns. The increase in global mean temperatures over the last few decades can be accounted for by the observed decadal-scale strengthening of the NAO [10,17]. There has also been an accompanying change (see figure 2(c)) in the eastward extent of the Atlantic storm track towards Europe [18]. These effects are consistent with a change in the relative frequency of the phases of the NAO, rather than in a change to a different or new type of teleconnection pattern, but clearly there are implications for the occurrence and frequency of extreme events downstream over Europe.

The paradigm that climate change will manifest itself via changes in the frequency and amplitude of existing teleconnection patterns carries important ramifications. It serves to underline the desirability of enhancing our understanding of and ability to predict these patterns and the associated "phenomena chain". It also suggests that there could be a change in the frequency and intensity (but not necessarily the type) of extreme events. The issue of predicting these features of extreme events is taken up again in section 7.

3. Climate change and the intensification of the water cycle

The total vertically integrated water content of the atmosphere amounts to as little as ∼25 mm, and this represents ∼0.001% of the overall global (fresh and sea) water resources. Despite its small amount, atmospheric water (in the form of water vapor, water droplets, and ice crystals) is one of the most important factors in our climate system and participates in several key feedback processes: Water vapor is the most important greenhouse gas and thereby affects the infrared radiation budget; clouds as well as surface snow cover reflect incident sunlight and thereby affect the visible radiation budget; and finally phase transitions of water in the atmosphere closely couple the atmospheric water and energy cycles. This latter coupling is so highly efficient due to the extremely large value of the latent heat of vaporization/condensation.

The coupling of the water cycle with the atmospheric circulation takes place on many levels. Associated radiative forcing and latent heating may drive, amplify or weaken atmospheric circulations such as planetary-scale circulations, low-pressure systems in the mid-latitudes; meridional and monsoonal circulations in the tropics; as well as embedded circulations on the mesoscale. The redistribution of water substance by precipitation may feed back to the ocean since precipitation and freshwater runoff from the continents provides a driving agent for oceanic circulations. In turn, such changes may also feed back to the local precipitation climate.

Overall, an assessment of climate change and its water cycle thus requires a fully integrated assessment, interactively accounting for all the relevant key feedbacks. Nevertheless – it is justified to specifically review in this section aspects of the temperature–moisture feedback, for its major potential importance in climate change. This feedback links climate change with a moistening of the atmosphere and an intensification of the hydrological cycle. Such a moistening could have major repercussion upon a wide range of cloud and precipitation processes, and contribute directly to changes in the regional hydrological response.

The basic premise of the temperature–moisture feedback is based on the observation that the atmospheric water vapor content is primarily controlled by the temperature dependence of the saturation mixing ratio, rather than by changes in relative humidity. Observed interannual and intraseasonal variations of the atmospheric moisture content lend support to this general behavior in the middle and high latitudes (e.g. [19,20]) but not in the tropics [21]. Similarly, observations show that the atmosphere has not only become warmer during this century, but also moister – at a rate which is approximately consistent with the temperature dependency of the saturation mixing ratio (e.g. [20,22]).

In agreement with the above premise, simulations undertaken with global climate models (GCMs) suggest that a global-scale climate warming could be associated with a substantial increase in the column-integrated atmospheric moisture content (e.g. [23]). The results suggest an increase of the atmospheric moisture content by about 7% per degree of warming.

The moistening of the atmosphere as a result of greenhouse gas forcing is due to several factors. First, greenhouse gas forcing implies an increase of the incoming infrared radiation at the surface, and this not only warms the boundary layer but also enhances the hydrological cycle, as much of the heating at the surface goes into evaporating surface moisture [24–26]. Second, evaporation over sea and evapotranspiration over moist land surfaces are approximately proportional to the saturation humidity deficit. Thus, even when assuming constant relative humidity in the planetary boundary layer, changes in evapotranspiration have a similar temperature dependency as the saturation mixing ratio itself. Together these two factors imply an intensification of evaporation and ultimately of the whole hydrological cycle with increasing surface energy balance and temperature. Indeed, under current climatic conditions, around 80% of the net surface energy balance are converted into evapotranspiration, and only 20% are converted into surface sensible heat fluxes [24,27]. Thus, the energetic effects of climate change associated with changes in the hydrological cycle may be more substantial than those resulting from direct thermal effects.

While the observed and simulated atmospheric moistening establishes a fairly close relationship between the atmospheric moisture content and the saturation mixing ratio except for the tropics, globally averaged evapotranspiration and precipitation rates in GCM simulations appear to increase at a substantially slower rate (see, e.g. [24,28–30]). The prime reason for this behavior is probably the dominance of oceanic evaporation, and the fact that the sea surface temperature warms at a slower rate than global surface temperature. Considerations of the atmospheric water balance then imply that the averaged residence time of the water molecules in the atmosphere is increased in current GCM experiments of climate change [26]. Such an increase in residence time might be associated with a reduction of the space/time fraction within which precipitation falls. Yet, at the same time, there is a net increase in globally averaged precipitation.

A critical question regarding the intensification of the hydrological cycle is thus its impact on the frequency-intensity distribution of precipitation events. All GCM simulations

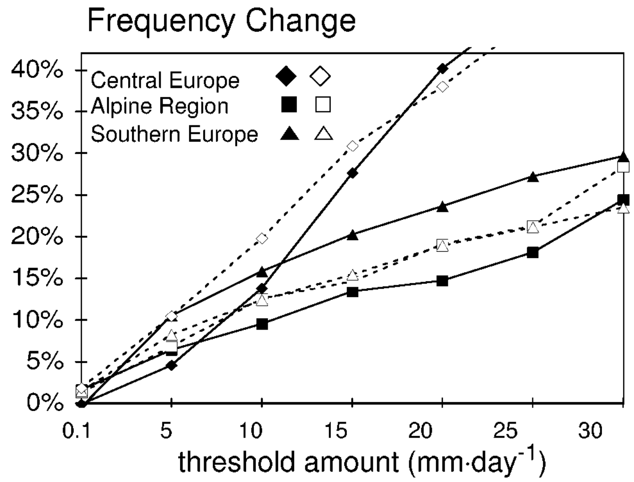

Figure 4. Change in daily precipitation statistics (frequency change as a function of threshold) simulated with a regional climate model for a surrogate climate change scenario (—) and derived from a scaling of the empirical probability density function (- - -). Results are valid for the autumn season and for three subregions in Central Southern Europe. The large-scale scenario is characterized by a uniform temperature increase by 2 degrees and a 15% increase of the atmospheric moisture (unchanged relative humidity). The climate model mimics the regional effects of "climate moistening". (From 16.)

with increased greenhouse gas concentrations suggest some enhancement of the hydrological cycle and most provide also some indication of an attendant shift towards more numerous events of heavy precipitation [31,32]. The latter effect has been interpreted in terms of a simple moisture-precipitation feedback. Since only about 20% of the precipitation comes from evaporation within a distance of about 1000 km [26], precipitation is usually more directly linked to the atmospheric moisture content than to nearby evapotranspiration. Fowler and Hennessy [33] pointed out that an associated flat percentage increase of each precipitation event (scaling of the precipitation probability density function) by $\sim$20%, which corresponds to a temperature–moisture feedback of $\sim$3 K, might ultimately lead to a strong increase in the frequency of heavy precipitation events. For midlatitude conditions and stratiform precipitation processes, this hypothesis has been confirmed by a regional climate modeling study (figure 4, [16]): The simulated changes in precipitation frequency resulting from a warming and moistening are found to be comparatively minor for weak events but exhibit a strongly progressive increase with the event size. As shown in figure 4, the simulated changes in the precipitation statistics are found to be in good agreement with the scaling hypothesis.

In addition to the moisture content, the nature of the response will depend on many additional factors such as changes in storm track dynamics, soil moisture conditions, cloud formation processes, atmospheric stratification, and the relative importance of stratiform versus convective precipitation. Since the last IPCC report [34] additional scenarios using global and regional models have become available. For equilibrium doubling of carbon dioxide, Hennessy et al. [35] find that for a given return period of 1 year there is an increase of the precipitation intensity by 10–25% in Europe, USA, Australia, and India. For a given precipitation intensity, the return periods becomes shorter by a factor 2–5. McGuffie et al. [36] confirm this conclusion when comparing the behavior of five different GCMs. Zwiers and Kharin [37] compute the changes of event size with a return period of 20 years in equilibrium integrations. In the CO_2 doubling experiment they find an increase of the likelihood for heavy precipitation almost worldwide with maximum absolute changes in event size in the tropics and over India. Jones et al. [38] use a regional model with about ∼50 km resolution driven by a global model. Over Europe, both the driving GCM and the regional model find frequency increases for events exceeding 10 mm/day by between 12 and 50% throughout the year.

In relation to precipitation in the Alpine and pre-Alpine region, the temperature–moisture feedback is likely to play an important role during the colder seasons, when the water cycle is primarily determined by sea-to-land transport, and when convective precipitation is not overly important. The associated changes in the frequency of extreme events will be modulated by changes in the synoptic scale circulation (cf. section 2) and will be particularly sensitive to weather patterns with strong warm and moist flow towards the Alps.

During the summer season, however, changes in heavy precipitation events are more difficult to assess, as these are affected by additional factors. The uncertainty comes from the fact that summer precipitation over the European continent is affected by the underlying land surface. During episodes of active convection, evapotranspiration and thus soil moisture dynamics as well as cloud-radiative feedbacks become highly relevant, thus questioning the direct applicability of the temperature–precipitation feedback.

Climate change scenarios derived from GCM experiments have indeed detected – despite the global intensification of the hydrological cycle and the associated increase of mean precipitation – an increased likelihood for the occurrence of drought conditions in semi-arid regions [39,40]. The prime reason are increases in evapotranspiration, which are efficiently reducing the summer-time soil moisture content. Surface conditions (albedo, soil moisture content, vegetation, and snow cover) determine the evolution of surface variables in response to the larger-scale synoptic forcing, and related changes may feed back to the near surface climate [41,42] and ultimately also to the precipitation response [43]. Beside numerical simulations, there is also observational evidence from lagged correlation analysis between soil-moisture conditions and subsequent precipitation for such effects to be active in the middle latitudes [44].

In relation to the European and Alpine climate, analysis of the literature suggests that with climate change the semi-arid regions of southern Europe may experience drier conditions during summer and autumn. To the extent that much of the Alpine summertime convection occurs with slightly south-easterly flow, this might yield a concomitant reduction of Alpine precipitation. However, this hypothesis is speculative at present, as it does not properly integrate changes in ambient atmospheric circulation. Additional numerical experiments of the aforementioned feedbacks are thus needed, along with model improvements of the relevant processes to assess the fate of summer precipitation over Europe.

4. Climate change, hydrological processes, and large-scale flooding

Atmospheric processes can spawn heavy/extended precipitation with the potential for flooding, but much of the resulting impact is ultimately controlled by hydrological processes at the surface and in the soils, determining the evolution and peek level of river runoff. These processes are particularly relevant for flooding events along rivers draining large-scale basins of several thousand square kilometers. Several of the major European river systems originate in the Alpine ridge (Po, Rhone, Danube, Rhine). This section explains the conditions for flood initiation and runoff processes controlling large-scale river floods and it discusses, at the example of the river Rhine, the potential role of anthropogenic influences and future prospects of large-scale floods.

4.1. Flood initiation and flood-generating hydrological processes

Precipitation and temperature are decisive meteorological variables for the release of a flood. Besides the basin- and time-integrated precipitation amounts, the temporal evolution and spatial distribution within the catchment play an essential role for the development of and the partition into the various runoff components (see below), and for the relative timing of the peek runoff between the tributaries. Surface air temperature determines the partitioning into snow and rain during the precipitation event and – in combination with radiation and surface winds – controls the runoff from snow and glacier melt and determines the amount of re-evaporation from soils and vegetation.

Temperature and humidity conditions of the soils at the instant of the precipitation event are important initial factors to the evolution of a flood. Frozen or moisture-saturated soils have a limited infiltration and moisture-storage capacity, which tends to accelerate the runoff process. Hence the generation of a flood also critically depends on the meteorological history prior to the heavy precipitation episode.

The flood hydrograph consists of several main runoff components characterized by different runoff velocities. *Surface runoff* refers to waters routed at the surface or within the top soil layers and the river beds. Surface runoff reacts quickly to excessive water, and the response-time of a river system to precipitation is largely depending on the fraction of surface runoff. *Interflow* refers to the delayed runoff through layered soils. This component also reaches the catchment river network and contributes to the total outflow at the watershed outlet. Response times of the interflow component depend on the permeability of the soils, with less permeable soils being generally characterized by a higher flow channel density and hence faster response times and higher flood potential. *Baseflow* or *ground-water flow* refers to the water runoff into the deep soils. It is the predominant runoff component during dry weather conditions, but it has, in general, only a small portion to the hydrograph of a flooding event.

Three hydrological stages of a flood evolution can be distinguished: (1) The *generation and partition* of the various run-off components. Apart from the meteorological conditions, this process is controlled by a range of physiographic catchment characteristics, such as the topography, slope, landcover, landuse, vegetation, water storage capacities as well as soil conductivity and stratification. The flood potential is enhanced in situations that allow a high fraction of surface runoff. (2) The *runoff concentration* determines the pathways of the various runoff components through the catchment river network and subsurface flow channels to the watershed outlet. A high flood potential is achieved when surface runoff from various parts of the catchment and subsurface flows are in phase. (3) The *flood routing* finally determines the flow of water masses through the tributary network to the main river. Riverbed geometry, slope and roughness determine the speed of this runoff. The peek water level and the duration of floods along the main river will critically depend on the runoff timing between tributaries. The existence of natural retention areas, lakes, manmade storage reservoirs, and dams can moderate the peek level and flooding impacts.

4.2. Large-scale flooding in the Rhine river basin

The Rhine river basin (see the geographical map in figure 5) covers an area of 185 300 km^2. The main river length is about 1300 km. The basin represents landscapes from the mountainous parts of the Alps and numerous hilly catchments (i.e., Vosgues, Black Forest) down to the lowlands of Western Germany and the Netherlands. The Rhine basin touches nine countries and belongs to the large rivers in Europe with a relatively high discharge per unit area (mean in Cologne 14.3 l s^{-1} km^{-2} for the period 1891–1998). In contrast to other rivers, the seasonal variations of runoff are relatively small. This is due to the various runoff regimes within the different subcatchments and regions [45]. The Rhine river can be divided into six main parts (figure 5): The Alpine Rhine (from its source to the Lake of Constance) and the High Rhine (from the Lake of Constance to Basel) – both with alpine catchment character; the Upper Rhine (between Basel and Bingen, including the tributaries Neckar and Main) and the Middle Rhine (between Bingen and Bonn, including the tributary Mosel) – both characterized by hilly catchments; the Lower Rhine (from Bonn to the German-Dutch border) and the Rhine-Delta – both located within lowland.

Figure 6 shows the mean runoff regimes (mean monthly discharges) at different gauging stations, which represent the various regions of the Rhine river. The average runoff regime of the Alpine rivers is dominated by snowmelt, and

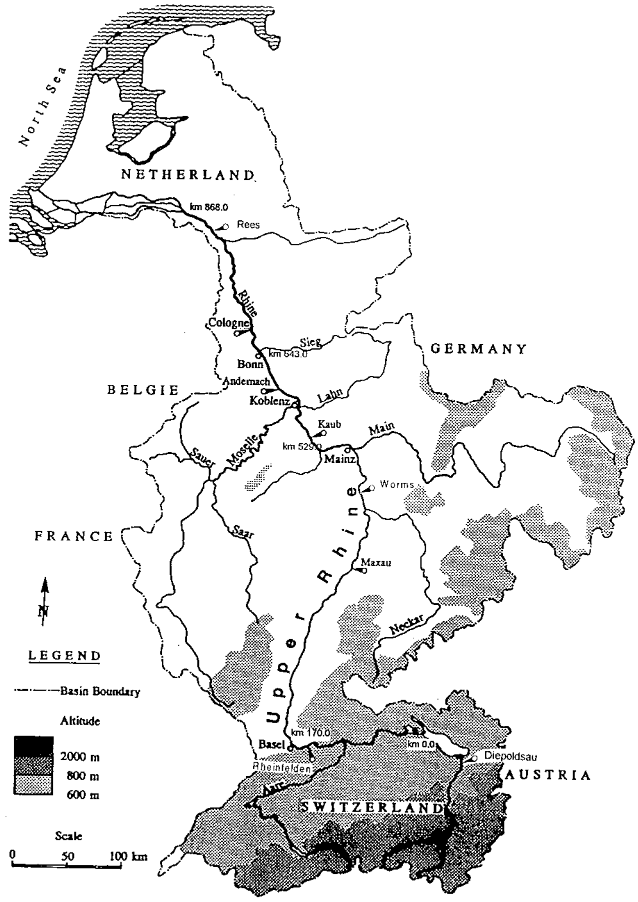

Figure 5. Map of the Rhine river basin in Central Europe, including geographical names, tributaries, and major river gauges mentioned in the text.

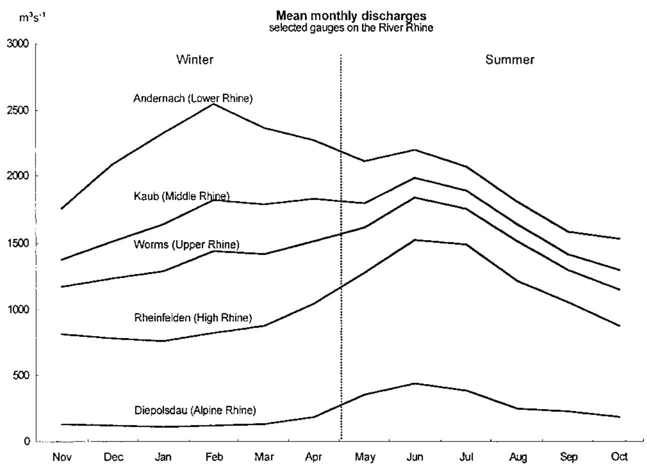

Figure 6. Annual cycle of long-term mean runoff observed at five river-gauges along the Rhine river, demonstrating the downstream variation of runoff regimes. Portions of areas for the subcatchments intermediate to the respective river gauges are: Alpine Rhine 3.3% (1.4% glacier covered), Alpine and High Rhine 18.6% (1.3% glacier covered), Upper Rhine 18.6%, Middle Rhine 18.8%, and Lower Rhine 19.5%.

from the high-altitude parts of the basin also glacier melt contributes. The discharge is governed by glacial and nival regimes with a pronounced maximum in summer and a minimum in winter. In the lower catchment parts, the maxi-

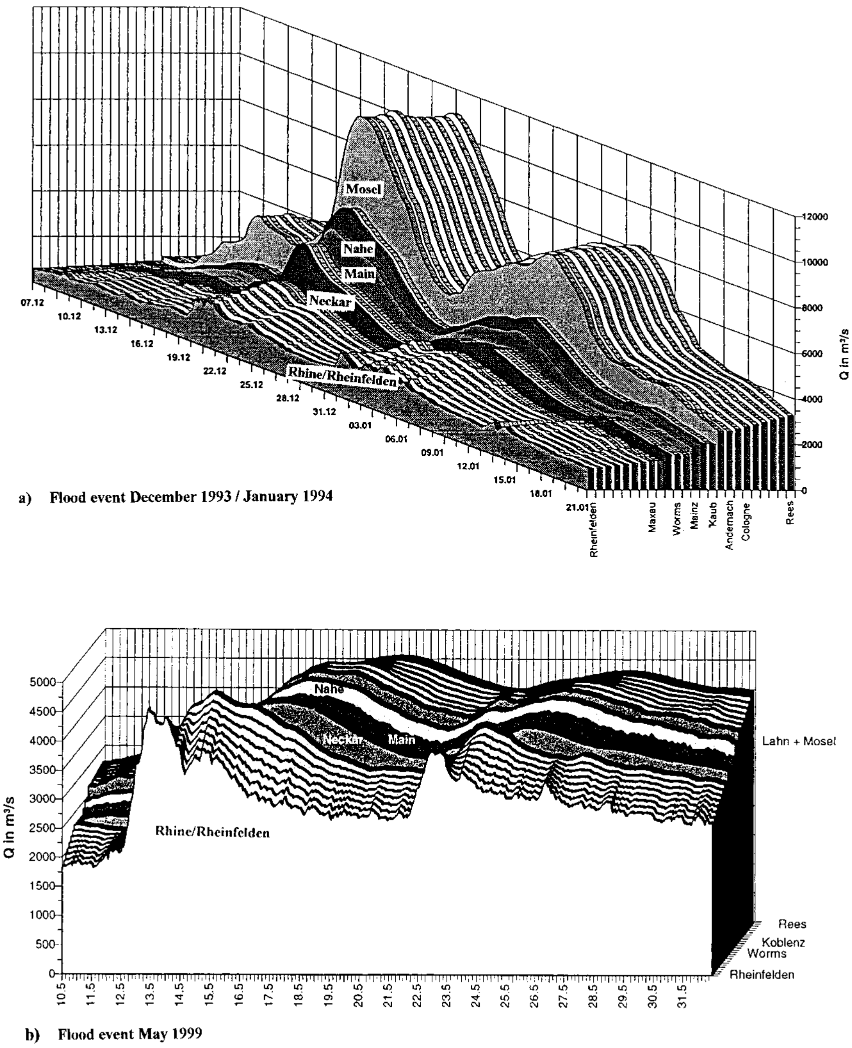

Figure 7. Hydrograph (time-evolution of river runoff) observed at various river gauges of the Rhine river, during two major flood events in (a) December/January 1993/1994 and (b) May 1999. (From [45].)

mum runoff progressively moves towards the winter months. In all tributaries of the Rhine river originating in medium-altitude mountains, the pluvial regime prevails with lower discharge in summer. In the Lower Rhine, the high winter runoff from the hilly and lowland catchments overcompensates for the low discharge from the Alpine regions [45]. The difference between runoff regimes is primarily caused by the different meteorological conditions in combination with differences in altitude, topography, landuse, vegetation, soils, and groundwater storage.

Rhine floods can become extreme events with catastrophic consequences when simultaneous flooding occurs in several subcatchments and when the associated discharge is superposed in the main river. Nevertheless, the long-term gauging record of the Rhine shows that there has never been any flood with simultaneous high intensity in all major subcatchments [45]. Figure 7 shows the hydrographs of two extreme flood events at several important gauging stations from Rheinfelden down to the German-Dutch boundary. The December-flood in 1993 was primarily generated in the hilly catchments in the middle River basin and by part in the lowland. The Neckar, Main, and Mosel rivers caused an extreme flood with damages particularly in the Middle Rhine regions. During the flood in May 1999 (figure 7(b)) a completely different situation occurred. This flood had it’s main origin in the Alps. The most serious damage occurred in Switzerland and Southern Germany, while the Middle Rhine was less affected. In this case, runoff from heavy precipitation was supported by the melting of winter snow in the Alps and in the Alpine parts of the High Rhine section. Precipitation and meltwater were combined and runoff was accelerated by saturated soils and high lake levels. During the May-flood 1999, the higher flood peak at Rheinfelden was reduced downstream due to smaller precipitation amounts in the following subcatchments. In the Middle Rhine and Lower Rhine region no extreme floods occurred.

Beginning in the 19th century, different river training measures were realized in the Upper Rhine [45,46], which inadvertently caused an increase in flood discharges and flood damages. Due to the short-cutting of river meanders and the straightening of Rhine river, the length of the main river channel was shortened and the slope, the erosion and the depth of the river bed were correspondingly increased. The sediment regime of the Rhine was also disturbed by the construction of river power plants. In this way the frequency of foreland flooding was decreased. At the same time, however, the retention area was reduced due to the construction of dikes which were constructed with the purpose to improve agricultural efficiency. The increased sealing of land surfaces and reduction of natural flood areas caused a higher proportion of rapid surface runoff in hydrographs, and a distinct increase of flood frequency. Associated with higher discharge velocities, the in-phase superposition of flood peaks from the tributaries at the main stream became more frequent. This is particularly pronounced for the situation of the Rhine downstream the estuary of the Neckar river. Some counter measures were realized in the last years in the region of the Upper Rhine.

Changes in landuse and vegetation cover conditioned by man-made activities can also have a marked influence on the frequency, magnitude, and duration of floods. The increase of urban areas and settlements at the expense of forests leads to more rapid runoff formation and higher flood levels. Such changes, however, can be considered to have a more pronounced impact in small catchments.

Besides man-made influences on runoff and flood generation in the past, the hydrological impacts of climate change may become more important in the future. Changes in precipitation and temperature regimes, in the ratio of snow to rain, changes in atmospheric circulation and the character of atmospheric fronts – all are connected to changes in runoff regimes and in flood generation characteristics. Current GCM-based climate change scenarios for the Rhine basin yield higher air temperatures throughout the year, higher winter precipitation, and somewhat smaller summer precipitation than current climatic conditions [47]. Hydrological scenario calculations [48,49] show that this might substantially affect the runoff regime of the Rhine, primarily due to the replacement of nival regimes by more pluvial regime types. Thus the mean Rhine discharge is expected to rise in the winter and to decrease in the summer in comparison to the present state. Due to the temperature increase, the sequence of snowfall, growth/melt of the snow cover is reduced in duration. The proportion of rainfall is expected to increase at the expense of snow. Overall, an increase of the flood risk in winter has thus to be anticipated. The compensation between different runoff regimes in the lower part of the Rhine river may also be reduced. In addition, an increase in the frequency of heavy events and in total amounts (cf. sections 3 and 6) will contribute in the same direction. In contrast, the mean discharge is likely to decrease during the summer, while the impacts on local floods of convective origin are difficult to assess due to current uncertainties in GCM simulations.

Changes in the seasonal runoff regime and the frequency of anomalous high and low river runoff could alter the risk for flooding damages and could affect the usage of the Rhine river as a major water way. Both the expected reduction of summer-time water conditions and an increase of winter-time peek levels could contribute to reducing the period amenable to ship navigation which in turn would have economic repercussions.

The example of the Rhine river illustrates that strategies for improving flood and low flow forecasts, require sophisticated warning and protection measures. The coupling of meteorological weather forecast models with area-distributed hydrological forecast models is a promising way towards this aim (see later section 7).

5. Detectability of trends in extreme events

The strong impacts from extreme events and the potential for changes related to long-term anthropogenic impacts (see sections 3 and 4) naturally prompt to the question if systematic changes in their occurrence can be seen in observational records of the recent past. In this section we illustrate that, although objective analyses of long-term changes (trend analyses) are technically feasible, the evidence from such analyses may be very limited as the rarity of extremes largely prevents the identification of changes, even if these would be present.

Objective inference about a long-term climate trend is generally obtained from a statistical description of the instrumental record (a statistical model) which separates its variations into a gradual component (the trend, i.e., the change of "typical" climatic conditions over the period), and a quasi-random component (the stochastic variations, i.e., the short-term fluctuations about the "typical" conditions). The stochastic fluctuations constitute an element of *noise* in the climate record from which an eventual trend (the *signal*) can be delineated only with limited accuracy. For example an anomalously high number of wind storms in later years of a wind record could be the result of either a gradual increase in storm probability (a positive trend) or the coincidental clustering of storm-rich years with a stable background storm probability. An important task of a trend analysis is to assign confidence bounds for the trend estimates which are compatible with the "noisy" record. The result of this formal procedure (statistical testing) can be summarized in the notion of *statistical significance*: A trend is said to be *statistically significant* if the climate record comprises a long-term trend which is unlikely ($<5\%$) a pure result of an inaccurate trend estimate from stable long-term conditions. *Statistical significance* is a notion that helps the interpretation of a trend estimate on the background of its uncertainty and it can be viewed as a means of statistical trend "detection". However, the notion has no implication upon the physical cause, or the future continuation of the trend.

The uncertainty of trend estimation is particularly relevant to the analyses of extremes. The small number of rare events embraced in typical climate records is responsible for a high "noise" level with the risk of a poor signal-to-noise ratio. Trends may fail to be recognized as *statistically significant* by the testing procedure. Quantitative information

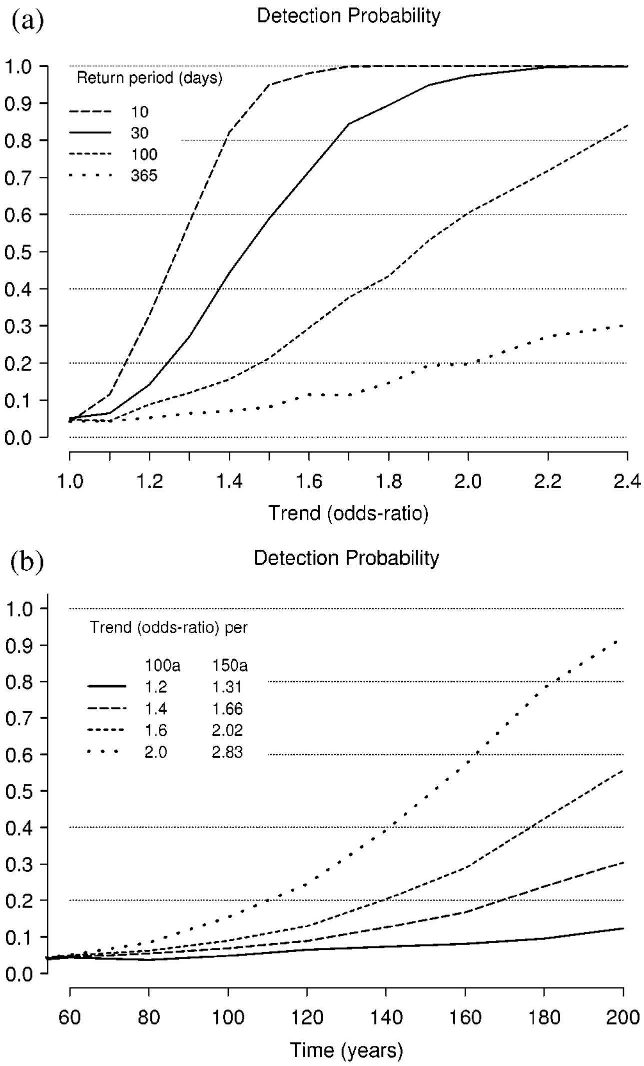

Figure 8. Probability for detecting trends in rare weather events: (a) Detection probability as a function of trend magnitude (horizontal axis, an odds ratio of 2 means a doubling of the frequency within the 100 year period) valid for seasonal counts in records of 100 years length. (b) Detection probability for extreme events (return period 365 days) as a function of record length and for four values of the trend. Trend values (relative change between the end and the beginning of the record) are listed in the inset and refer to a 100 and 150 year period. (See also [50].)

on trend detectability is therefore important a-priori knowledge for a sound interpretation of trend-analysis results.

Estimates of trend detectability for extreme events have been calculated by Frei and Schär [50] in terms of a *detection probability* expressing the chance of trend detection as a function of trend magnitude, record length and rarity of events. The calculations are based on the concept of binomial distributed counts and the statistical simulation of surrogate "trendy" and "noisy" records of extremes. (For details see [50].)

Figure 8(a) depicts the detection probability as a function of trend magnitude (horizontal axis). The estimates pertain for one hundred year records (a comparatively long extent for today's daily instrumental series), and for seasonal event counts (representative for seasonally stratified trend analyses). The results point to the pronounced limitations in trend detection for very rare events. While a centennial change by a factor of 1.4 (abscissa) in the frequency of moderate events (return period 10 days) is detected with a probability of more than 0.8, this value drops to 0.4 for events with a monthly return (30 days) and below 0.1 for extremes with a return period of 365 days. For the latter category of extremes even a very strong signal in the form of a doubling in frequency over 100 years is identified as *statistically significant* only with a weak probability of 0.2. Trend detection for this category of events is possible only in presence of very extreme trend magnitudes. It should be noted that this latter category still embraces typically 25 events in a 100 year record (for each season).

While this analysis highlights the difficulty to infer gradual changes of extremes from past climate records, it is also interesting to examine the possibilities for trend detection in the future. Detectability is expected to improve due to a progress of an eventual gradual change towards larger amplitudes and due to the larger statistical sample reflecting in more accurate trend estimates. Results for four scenarios of long-term trends are displayed in figure 8(b), again for events with a return period of 365 days. Although the detection probability evidently improves with longer record length, it is still clearly insufficient to discriminate changes up to a doubling over 150 years. Hence, even with the assumption of substantial changes in the frequency of extremes into the next few decades these trends will likely remain masked by the short-term natural fluctuations and are not soon becoming amenable to a sound statistical detection.

The principal limitations for a sound observational inference of long-term trends in extreme events has implications for climate data analysis. Firstly, care should be exercised during the interpretation of results: A statistically non-significant trend is not synonymous to the absence of a trend. In the case of rare events, even trends of dramatic magnitude may be masked by the stochastic component in the record. Secondly, the pronounced decrease of detectability with the rarity of events suggests that trend analyses could profit from focussing on moderately intense events rather than damage-causing extremes. Although trends inferred from moderate cases must not necessarily be representative for extremes, they can nevertheless be suggestive of changes in the frequency distribution. Thirdly, it is to be expected that similar statistical constraints hold for inferences from climate model simulations regarding future changes of extremes. Again the statistical sample of very rare events is limited by short integration periods of climate simulations and sound conclusions on expected future variations are necessarily to be based on cases of moderate intensity, rather than very rare catastrophic events.

In addition to various sources of uncertainty in our knowledge of future climate change, these considerations point towards a fundamental uncertainty in our assessment of past long-term variations. This latter uncertainty is intrinsic to the rarity of extremes and will not significantly improve with future availability of data. As a consequence, eventual long-term changes in the probability of extremes, or the expected value of damage costs – for example as an effect of anthropogenic climate change – will reveal unambiguously only long after they are in effect and have attained a substantial magnitude, at least on a regional level. Hence over the medium term, predictions of event probabilities and expected values will be of limited value. The major challenge posed by the principal limitation to detect changes in extremes is for planning tasks with a long-term perspective. For example, the provident planning of long-lived infrastructure or the definition of construction standards will be difficult without quantitative assessment of the risk and its changes over the life-time of the construction. In essence it will be difficult to maintain security standards over long live-times. Similar repercussions can be expected for insurance sectors whose premium planning critically depends on far future risk changes.

6. Heavy precipitation trends in the Alpine region

Physical contemplation and simulations with global and regional climate models provide ample evidence for the sensitivity of the hydrological cycle (see section 3). "Climate moistening" could be associated with a substantial increase in the frequency of heavy precipitation and would be expected primarily in regions/seasons with a maritime influence, where the potential for enhanced moisture transports can be utilized from evaporation over open water surfaces. The water cycle of Central Europe could be susceptible to these mechanism: Long-range water transports from the Atlantic and Mediterranean oceans dominate as moisture source for precipitation during the winter half year [51]. Moreover, on a smaller scale, a prominent class of heavy precipitation systems in the Alpine region, frequently developing during the autumn season along the southern rim of the ridge, is characterized by intense low-level moisture advection from the Mediterranean sea (e.g. [52–54]). In this context the analysis of long-term precipitation changes in Central Europe appears of immediate interest.

As regards seasonal mean precipitation, a European-scale analyses of sparse rain-gauge data [55] and a multivariate trend analysis for a dense centennial network in Switzerland [56] have both identified increasing winter-time precipitation trends over regions north-west of the Alpine main crest. Relating the trend to the frequency of characteristic weather classes of the Alpine region, the latter study concluded, that the long-term increase is dominated by the increase of precipitation intensity within individual weather classes, rather than long-term shifts in their frequency. Over northern Italy (i.e., to the south of the Alpine ridge), the analysis of extended observation records (1833–1996, [57]), has revealed an overall precipitation decrease which is statistically significant in autumn.

As regards the changes in the frequency distribution and the occurrence of heavy precipitation there is not as comprehensive information for the Alps: Considering a carefully proofed observation record in the south-western Alpine region (Genova, Italy), Pasquale [58] found a pronounced increase of precipitation intensity (defined as the average amount per precipitation day). The increase was particularly evident in the autumn season, i.e., the season with peak activity of heavy precipitation in this region of the Alps. Another study for the long-term record of Vienna in the Eastern Alps [59] did not reveal statistically significant trends since the late 19th century, yet the type of events considered was fairly severe and trend detectability might have been limited (see section 5).

Considering daily data of the dense centennial network in Switzerland, Frei, and Schär [50] examine trends in the occurrence of rare, *intense* rainfall in Switzerland. An *intense* event is defined as a daily total exceeding the amount of a 30 day return period rainfall. While the category of *intense* events does not exclusively comprise rare extremes, relevant for adverse impacts, their modest rarity warrants a reasonable detection probability (see figure 8 and section 5). The employed statistical technique takes account of the specific nature of count data (see [50] for details).

The analysis reveals seasonally distinct results (see figure 9): For summer and spring (not shown), stations with positive and negative trend estimates are roughly balanced and trends are statistically significant in exceptional cases only. In contrast for autumn (not shown) and winter, there is a strong prevalence of stations with an increasing trend and for a large fraction (about one third) of the stations the trend estimate is statistically significant. The results provide evidence for an increase in the frequency of *intense* precipitation in the Swiss Alps during the 20th century. The increase observed in the autumn and winter seasons is substantial, amounting to 20–80% within 100 years (see figure 9 (c) and (d)).

The findings for daily precipitation in Switzerland are conform with results obtained for a range of other regions world-wide: Using daily data from 180 long-term records (since 1910) of the continental US, Karl et al. [60,61] found an increase in the proportion of precipitation from heavy events (i.e., the 95th percentile) primarily for summer and spring and for the north-eastern and western (i.e., the coastal) parts of the country. Rakhecha and Soman [62] considered annual extremes of 1–3 day precipitation amounts in India (since 1901) and noted increasing trends along the north-western coast and decreasing trends over the southern (i.e., tropical) part of the peninsula. For observations in Japan, Iwashima, and Yamamoto [63] reported that most of the record daily precipitation amounts since 1890 were observed in the second half of the 20th century. Again, Suppiah and Hennessy [64] found evidence for increasing trends of intense daily precipitation (the 90th percentile) in South-Eastern Australia during summer since 1910.

The observed increase of intense precipitation events in the Alpine region and in many other maritime regions of the middle latitudes is conform with the hypothesis of a warming-induced intensification of the water cycle (see section 3). However, the trend analyses alone do not allow for conclusive indications about the physical nature of the ob-

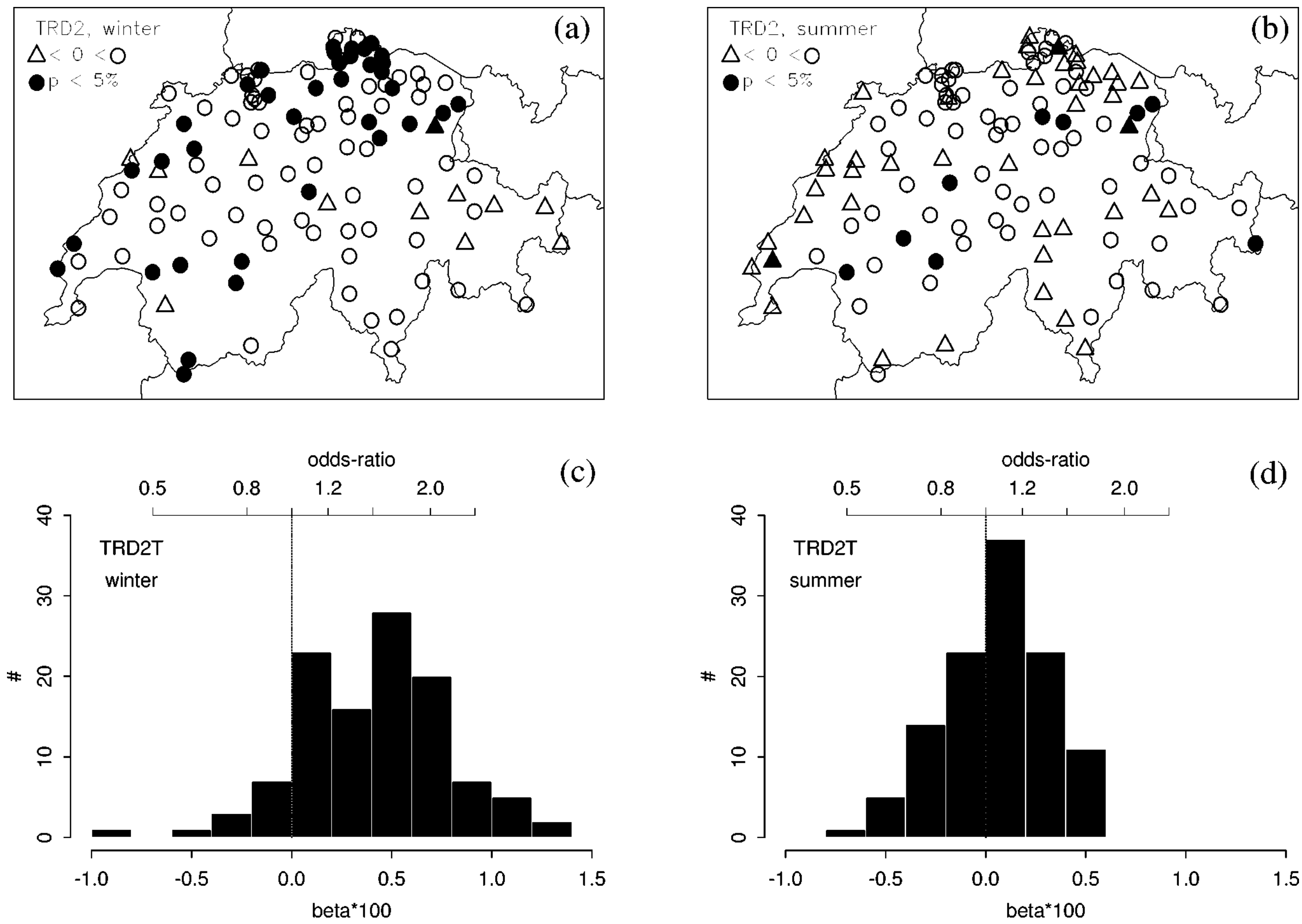

Figure 9. Observed centennial trends in the occurrence of intense precipitation events in Switzerland in winter (panels (a) and (c)) and summer (panels (b) and (d)). Symbols represent the sign of the trend for each station ((○, ●) increasing, (△, ▲) decreasing) and its statistical significance ($p < 5\%$, filled symbols). Histograms of trend estimates (panels (c) and (d)) are shown for the 113 stations (top axis represents ratio of frequencies between the end and the beginning of the period in a logarithmic scale). (From [50].)

served trends and their continuation into the future. Also the results for intense events are not necessarily representative for severe, damage causing events, although they provide indication for long-term changes in the precipitation frequency distribution.

An increase in the frequency of heavy precipitation events in the Alpine region could have serious repercussions for the natural environment and living conditions over the long-term. The development of impacts will however also depend on local vulnerabilities, changes in the hydrological response at the surface, the vagaries of interannual variations and the ability to forecast and issue early flood warnings.

7. Prospects for improved forecasting and warning

Experience suggests that damage from heavy precipitation and flooding events can substantially be reduced when appropriate warnings are available with sufficient lead time. A timely warning allows to set the respective infrastructure on alert, to evacuate particularly vulnerable regions, to close affected traffic routes such as exposed bridges and railway lines, and to optimize the management of dams and other water management infrastructure.

Current flood forecasting and warning systems employ both an atmospheric and a hydrological component. The atmospheric component provides the meteorological input about precipitation rate, precipitation type, surface temperature, and surface radiation balance. For some purposes, when it is sufficient to consider past precipitation, the atmospheric component may simply be based on observations from conventional meteorological and weather radar instruments. More generally, however, the atmospheric component is based on a numerical weather prediction model. These models operate on a computational grid and integrate the atmospheric state forward in time from the initial conditions. Typically forecasts are conducted using a series of nested models of increasing resolution. The first model in this chain is a global weather forecasting model, while subsequent steps involve limited area "meso-scale" models.

The hydrological component consists of a scheme to predict the evolution of river runoff and lake levels based upon the meteorological input and the initial state of the system (e.g., its initial soil moisture content and water levels). Such models describe the infiltration of precipitation into the soil, the formation of runoff, and the routing of runoff through streams, rivers and lakes. The underlying models can be based on lumped or distributed methodologies. In the former case, there is a bulk consideration of individual catchments and sub-catchments (see, e.g. [65]), while the latter methodology uses a computational grid not unlike those considered in atmospheric models (e.g. [66]).

A central characteristic of runoff forecasting systems is the response time of the catchment under consideration. It is defined as the time delay between a precipitation pulse and the runoff at a stream gauge exiting the catchment. For large catchments such as the Rhine, the response time may be as large as several days, while for very small mountainous catchments it may be as short as a few minutes. The response time scale of the catchment under consideration thus determines the lead time of the meteorological forecast which is needed to provide useful runoff forecasts. Figure 10 presents estimates of the relevant time-scales for the mid-latitudes. The response time is to first approximation proportional to the horizontal scale of the catchment, and yields a straight line in the logarithmic display of figure 10.

Figure 10 also includes estimates of the atmospheric predictability. Since the atmosphere is a chaotic system, there are intrinsic limits to its predictability. Currently it is believed that on the synoptic scale (the scale of individual low pressure systems) this limit is around 10–15 days. The most advanced global weather forecasting models nowadays provide useful information on this scale for about 6 days (e.g. [67]). The predictability time scale decreases with decreasing horizontal scale. At a scale of a few kilometers, for instance, predictability is very low (around 30 min), in particular during episodes of convective (thunderstorm-like) precipitation. Individual convective cells and thunderstorms have a scale of a few kilometers, and their evolution and track is predictable for very short periods only. Yet the forecasting of total precipitation in a small-scale catchment may require precise information on the behavior of individual convective cells.

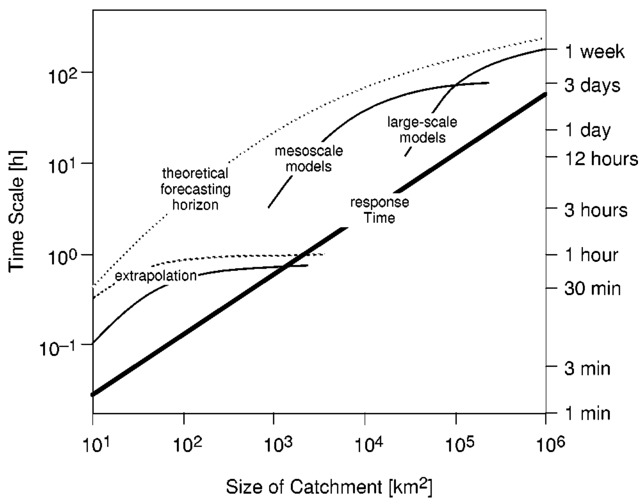

Figure 10. Catchment response-time (bold line) and atmospheric predictability time scale (thin lines) as a function of catchment area. The values provided are approximate estimates valid for mid-latitude regions.

Based on the currently available flood forecasting systems and the expected developments in the field of atmospheric and hydrological numerical modeling, three characteristic approaches may be distinguished (cf. figure 10):

(a) For major river basins (>50 000 km^2), short term flood forecasting with lead times of about 1 day can mainly rely upon conventional meteorological and hydrological observations (i.e., precipitation and runoff observations upstream) and hydrological modeling. Due to the long response time of such catchments, most of the water to be converted into runoff throughout the forecasting period is already within the hydrological system at forecast initialization time. However, results from large-scale weather forecasting models can successfully be applied to extend the forecasting time scale past 24 h [68].

(b) For intermediate-scale catchments (1000–50 000 km^2), a combination of hydrological modeling and rainfall forecast is essential. On these scales a substantial fraction of the runoff falls as precipitation within the preceding 24 h. In practice, mesoscale atmospheric models with horizontal resolutions below 50 km are needed to provide the relevant precipitation forecasts. This is particularly essential in mountainous regions, where the underlying topography has a large impact upon the resulting precipitation response.

(c) Finally, for small-scale catchments (<1000 km^2), the intrinsic forecasting limit of the atmospheric dynamics becomes relevant. At these scales forecasts are only possible over short periods of a few hours. Numerical weather prediction models are not much used at these scales, since the preparation of the forecast itself requires a time frame exceeding the forecasting time. For very short term warning (<1 h), however, weather radar information and conventional precipitation measurements (nowcasting) can be used and extrapolated in time. This may be particularly useful to provide automated warnings for very small-scale catchments (<100 km^2). At these scales it becomes relevant to account for individual thunderstorms and convective cells.

Recent research efforts associated with the spatial band (b) suggest that a substantial improvement of current operational systems appears possible by using very high-resolution atmospheric models. Numerical simulation of atmospheric processes has experienced rapid progress in recent years (see [69,70]). One key factor in this development was the computer revolution which provided – at a constant investment level – exponentially increasing computing power. As a result, it became possible to integrate the governing equations of the atmosphere with increasing resolution and accuracy (see table 1). The further development of forecasting models is well underway, and it is currently expected that the feasible horizontal resolution will go increasing at a rate similar to that in the past. Thus, limited-area numerical models will reach a resolution of ∼1 km early in the 21st century. Such a high resolution has major advantages and offers exciting prospects. In particular, with such a resolution it will become possible to

Table 1
Horizontal resolution of operational weather forecasting models.

Year	Global models	Regional models
1985	200 km	50 km
1999	50 km	15 km
200?	20 km	1 km

- realistically resolve major valleys and massifs of the underlying topography (cf. figure 11), better account for land–sea contrast and coastal effects, and thereby better simulate the atmospheric flow and precipitation response over complex orography;
- improve the prediction of precipitation including the occurrence of heavy events by explicitly resolving moist convection and thunderstorms (this will replace the questionable parameterization of convective precipitation in today's models); and

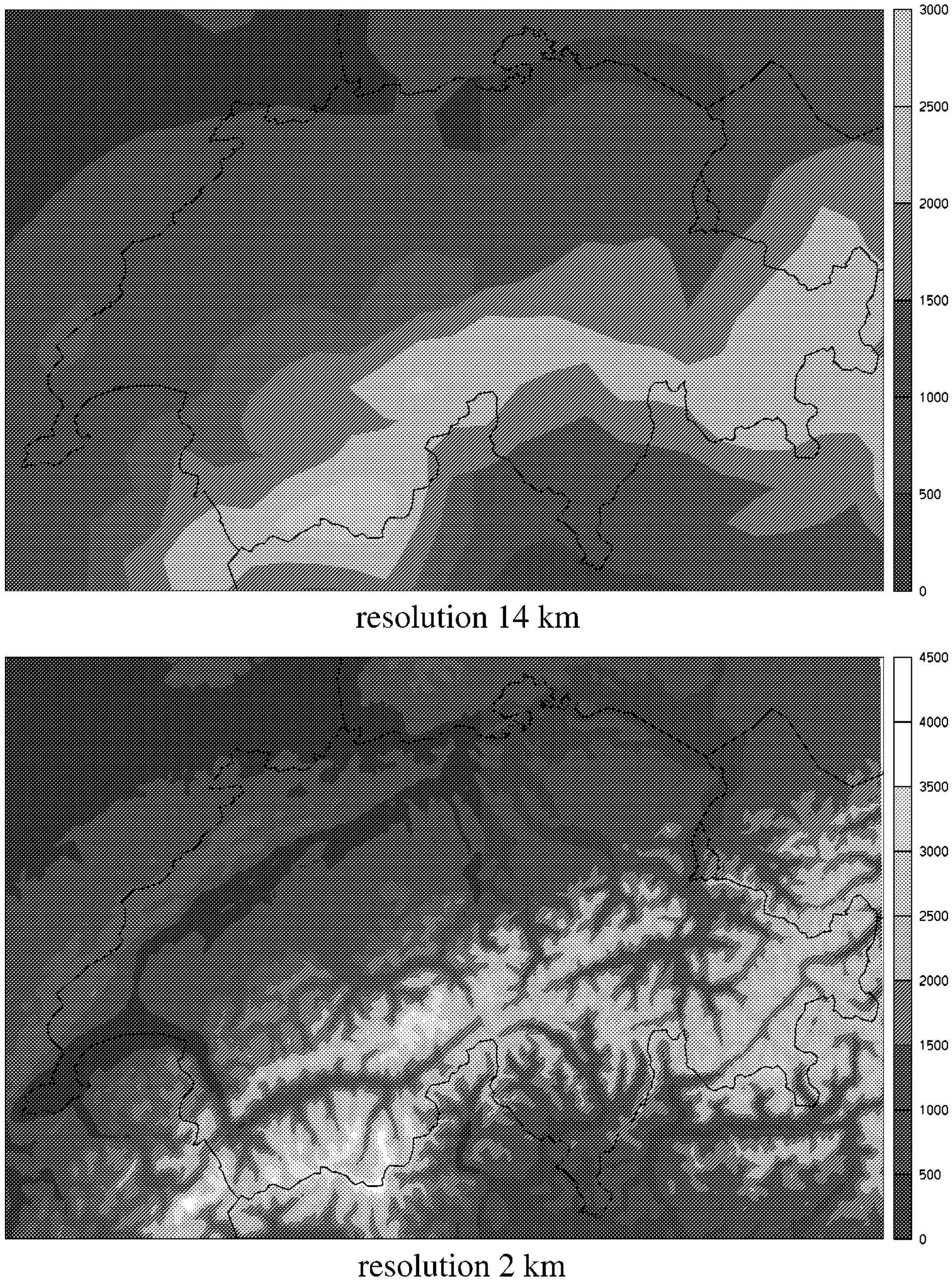

Figure 11. The topography of Switzerland at the resolution of current numerical weather prediction models (14 km, top panel) and that of future numerical models (2 km, bottom panel).

- for the first time to resolve intermediate-scale catchments and thereby provide high-resolution data for the improvement of runoff predictions.

These developments are considered to be of high economic value. In the Alps for instance, flooding events that occurred during the last decade have inflicted loss of life and large property damage. The Brig flash flood (September 1993) and the Piedmont flood (November 1994) together resulted in 75 casualties and caused property damage of $\sim$10 000 million Euro [71]. The timely numerical prediction of heavy precipitation and runoff events in combination with real-time information from *in situ* and remotely sensed observations could substantially mitigate these effects. An improved understanding of heavy precipitation events is also of considerable interest for the current research on anthropogenic climate change, since the frequency of heavy precipitation events is one of the most pertinent, influential and sensitive characteristics of our climate [16,34].

The foregoing addressed development of high-resolution weather prediction models will require substantial input from the scientific community. First, the development of high-resolution models requires the availability of accurate high-resolution data sets for validation purposes. This is a difficult challenge since the current operational data in the free atmosphere has – even over populated areas such as Europe – a resolution of only a few hundred kilometers. Second, the understanding of atmospheric, microphysical and hydrological processes at scales between $\sim$1 and $\sim$100 km needs to be improved to enable the development of realistic model formulations of these processes. Third, a closer interaction between atmospheric scientists and hydrologists is highly desirable to define integrated flood forecasting, warning and response systems, since the predictability limits may in many cases (e.g., the Brig flood mentioned above) be dictated by the hydrologic rather than atmospheric component. One programme that is addressing many of these aspects is the Mesoscale Alpine Programme (MAP). It involves research efforts using high-resolution atmospheric and hydrological models, as well as a large-scale field campaign to provide high-resolution data sets for validation and research purposes [71–73].

Despite the limited atmospheric predictability in the spatial band (c), where the timing and position of individual convective cells quickly eschews deterministic predictability after which the cells become randomly organized, the use of high-resolution atmospheric models might become attractive due to the better representation of the underlying precipitation physics. However, in this spatial band the forecasting strategy must be geared to provide probabilistic rather than deterministic information. Thus, particular attention should be devoted to a quantitative assessment of predictability. It is now well established that the high degree of nonlinearity introduces chaotic components into the atmospheric dynamics, which pose intrinsic limits to deterministic predictability [74]. The use of ensemble techniques (parallel model simulations for the same situation using slightly different initial conditions) has helped to delineate these limits for the atmospheric circulation (e.g. [75]), and related techniques are now operationally applied at many forecasting centers to estimate the degree of atmospheric predictability and its day-to-day variations. This research has largely focused on the synoptic scale, but its potential value for predicting extreme weather events now becomes more apparent [76]. The main advantage of ensemble forecasting systems is their ability to provide probability forecasts rather than merely a deterministic prediction with unknown probability. Only a probability forecast allows an a priori analysis of cost/loss ratio, and can serve to objectively issue a warning with a predefined false alarm rate.

8. Synthesis and conclusions

The foregoing sections have discussed some central aspects related to heavy precipitation and flooding in mid-latitude regions, and to the sensitivity of such events with respect to climate change. Several factors may affect the future evolution of the frequency of extreme events in Central Europe: First, changes in the general circulation of the atmosphere may affect the preferred track of Atlantic storms, and this relates the frequency of extreme events with planetary-scale teleconnection pattern of the mid-latitude atmospheric flow (section 2). Second, there is evidence for a global-warming-induced moistening of the atmosphere, both from theoretical and observational grounds. From an energetic point of view one can argue that climate change refers at first approximation to "climate moistening" rather than "climate warming". Both global and regional climate change scenarios and sensitivity studies suggest that there may be pronounced accompanying shifts in the frequency of heavy precipitation events. The underlying dynamics suggests that the sensitivity of this effect increases with the return period of the event, i.e., comparatively small increases (or even decreases) in the frequency of weak and/or moderate precipitation events may be accompanied by pronounced increases in the frequency of heavy events (see section 3). Third, in addition to changes in precipitation characteristics, changes in runoff formation will occur. With increasing winter temperatures, there will be a growing proportion of rainfall at the expense of snowfall, which will accelerate the runoff formation process (cf. section 4). This factor might be particularly relevant for large catchments such as the Rhine, which did benefit in the past from the compensating effects of different runoff regimes in the various subcatchments.

The vulnerability of our society to changes in the frequency of extreme events is calling for a careful monitoring of their occurrence. However, the detectability of trends in extreme events is very limited, as is directly implied by their rarity. The "detection probability" as introduced in section 5 provides a quantitative description of this difficulty. Several important conclusions can be drawn from the respective analysis: First, trends in the frequency of events with return periods exceeding a few years are not detectable with statistical significance – even when unanticipated large changes were to occur. For such events, the notion "a statistically significant trend was not detected" is meaningless, as the detection probability is anyway negligible, irrespective of the statistical methodology under consideration. Second, the analysis demonstrates that the detection probability rapidly increases with decreasing return period, thus suggesting to focus trend analysis on event categories with a return period of one or a few months. When applying such a methodology to intense precipitation events in the Alpine region (with return periods of 1 month), pronounced changes in frequency of such events were detected, with increases in autumn and winter of as much as $\sim$30% in 100 years (section 6).

The impacts of climate change through modifying the frequencies of extreme events will however not only depend on climate- and weather-related characteristics, but also on the further development of human settlements and infrastructure, as is well exemplified by past man-made changes of the Rhine catchment (section 4). In addition, susceptibility to capital damage and loss of life will crucially depend upon the presence of appropriate warning and response systems. With regard to heavy precipitation events and episodes of flooding in large and intermediate-scale catchments, there appears to be a high potential to improve current warning methodologies by the use of high-resolution atmospheric and hydrological models (section 7).

Changes in the frequency of extreme events have for long been one of the key topics in the institutional and public discussion about greenhouse gas reduction/stabilization measures and policies. It is likely that the small "detection probability" of trends in extreme events that was discussed in section 5 will ultimately reduce this motivation. A small detection probability implies that it will be difficult or even impossible to unambiguously demonstrate that anthropogenic climate change and associated changes in the frequency of extreme events indeed result in the high costs as, e.g., estimated by the IPCC [77]. Clearly, the situation is incompatible with the idea of an investment (in the form of global abatement policies) that returns a profit (in the form of a reduced number of extreme events). This lack of a convincing investment-profit feedback – which is a direct result of the low detection probability alluded to above – suggests that the motivation for implementing strict abatement policies on the level of CO_2 reduction might – regrettably – become even lower than anticipated today. In fact, for an individual country it is economically more attractive to invest in reducing the negative impacts of extreme events, irrespective (and not knowing) whether these are of anthropogenic or natural origin. Related early warning systems for flood events are currently in development in many industrialized countries, and it can be expected that such systems will in the future substantially reduce the negative impacts of heavy precipitation and flooding events. If implemented on a global scale (and in particular in the tropics where disastrous events of this kind are very common), such systems might ultimately contribute towards reducing the damage associated with extreme events below current levels, even when the frequency of associated events should increase.

Acknowledgement

Research of this paper was supported by the Swiss National Science Foundation (contract SPP-U 5001-044602).

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