The European mountain cryosphere: A review of its current state, trends and future challenges

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Abstract. The mountain cryosphere of mainland Europe is recognized to have important impacts on a range of environmental processes. In this paper, we provide an overview on the current knowledge on snow, glacier, and permafrost processes, as well as their past, current and future evolution. We additionally provide an assessment of current cryosphere research in Europe, and point to the different domains requiring further research. Emphasis is given to our understanding of climate-cryosphere interactions, cryosphere controls on physical and biological mountain systems, as well as related impacts. By the end of the century, Europe's mountain cryosphere will have changed to an extent that will impact the landscape, the hydrological regimes, the water resources, and the infrastructure. The impacts will not remain confined to the mountain area but also affect the downstream lowlands, entailing a wide range of socio-economical consequences. European mountains will have a completely different visual appearance, in which low and mid-range altitude glaciers will have disappeared, and in which even large valley glaciers will have experienced significant retreat and mass loss. Due to increased air temperatures and related shifts from solid to liquid precipitation, seasonal snow lines will be found at much higher altitudes, and the snow season will be much shorter than today. These changes in snow and icemelt will cause a shift in the timing of discharge maxima, and a transition of runoff regimes from glacial to nival, and from nival to pluvial. This will entail significant impacts on the seasonality of high-altitude water availability, with consequences for water storage and management in reservoirs for drinking water, irrigation and hydropower production. Whereas an upward shift of the treeline and expansion of vegetation can be expected into current periglacial areas, the disappearance of permafrost at lower altitudes and its warming at higher elevations will likely result in mass movements and process chains beyond historical experience. Future cryospheric research has not only the responsibility to foster awareness of these expected changes and to develop targeted strategies to precisely quantify their magnitude and rate of occurrence, but also to help in the development of approaches to adapt to these changes and to mitigate their consequences. Major joint efforts are required in the domain of cryospheric monitoring, which will require coordination in terms of data availability and quality. In particular, we recognize the quantification of high-altitude precipitation as a key source of uncertainty in projections of future changes. Improvements in numerical modelling, and a better understanding of process chains affecting high-altitude mass movements are the two further fields that – in our view – future cryospheric research should focus on.

1 Introduction

Ongoing climate change and the importance of its anthropogenic component have gained wide recognition (IPCC, 2013). Thereby, some regions are likely to be more vulnerable than others, both in the expected physical changes and the consequences for ways of life. Mountains are particularly subject to rapid and sustained environmental changes (Gobiet et al., 2014), and the cryosphere is the physical compartment that exhibits the most rapid ones. Changes in mountain snow, glaciers, and permafrost have resulted in significant downstream impacts in terms of the quantity, seasonality, and quality of water (Beniston et al., 2011**a**). This is particularly true for areas where snow- and icemelt represent a large fraction of streamflow. Countless studies have reported glacier retreat, permafrost warming, and snowfall decrease across mountain regions in Europe, with implications

for streamflow regimes, water availability, and natural hazards. These can in turn negatively impact hydropower generation, agriculture, forestry, tourism, and aquatic ecosystems. Consequently, downstream communities will also be under pressure, and mountain forelands with densely populated areas will be highly affected (Kaser et al., 2010, Huss et al., 2017). Both political and scientific programs are calling for better preparedness, and for the development of strategies aimed at averting

5 possible conflicts of interest, such as between economic goals and environmental protection (Beniston et al., 2014).

In the following, we provide an overview of the current knowledge on European mountain permafrost, glaciers, and snow, and the observed changes. We focus on mainland Europe, in particular the European Alps and Scandinavia, but also include – where possible – the Pyrenees and other mid-latitude European mountains. An assessment of the challenges that need to be

10 addressed in cryosphere research is provided, and we identify areas where further progress is required to improve our understanding of climate-cryosphere interactions. We argue that such improved understanding is the key for better predicting future changes and impacts, and for appropriate adaptation measures to be developed. Our views largely reflect the opinions of a body of scientists that convened during the "Riederalp Cryosphere Workshop" held in Switzerland in 2016. We therefore do not claim that all aspects of cryosphere sciences are exhaustively covered, nor that all possible elements of the cryosphere

15 are discussed (we neglect, for example, lake ice, river ice, and ice in caves), but are convinced to have summarized, in an effective manner, both the current state and the future challenges in the domain of the European mountain cryosphere research.

2. Past and future trends in European mountain cryosphere and their impacts

2.1 Changes in snow

Snow cover is the most important interface between the atmosphere and the ground, strongly influencing the surface energy

20 balance of the cryosphere. Snow affects glaciers through albedo and mass balance, and permafrost through its thermal insulation properties and melt water input. It also plays a key role for sustaining ecological and socio-economic systems in mountains, and also in the lowlands downstream of the mountain ranges. The extreme spatiotemporal variability of the snow cover remains one of the key uncertainties when quantifying the impact of climate change on the cryosphere. Snow observations are therefore a prerequisite for understanding the related processes, and for providing more reliable assessments

25 of future changes.

2.1.1 Observed changes of the snow cover

Most studies show negative trends in snow depth and snow duration over the past decades (Table 1). These negative trends are well documented in the Alps owing to the abundance of long-term observations. The changes are typically elevation dependent, with more (less) pronounced changes at low (high) elevations (Marty, 2008; Durand et al., 2009; Terzago et al., 2013).

30 Decrease in spring snow water equivalent (SWE) is equally found for the Alps (Bocchiola and Diolaiuti, 2010; Marty et al., 2017b; Fig. 1) as well as for low elevations in Norway (Skaugen et al., 2012). Only the higher and colder regions of the Fenno-

Scandinavian mountains exhibit positive trends of maximum snow depth and maximum SWE, although trends have recently become negative in these regions too (Johansson et al., 2011; Skaugen et al., 2012; Dyrrdal et al., 2013; Kivinen and Rasmus, 2015). In the Pyrenees, a significant reduction of the snowpack is reported since the 1950s (Pons et al., 2010), and also in other European mountains – where observations are less abundant – studies report declining snowpacks. The latter is particularly true for mountains in Romania (Birsan and Dumitrescu, 2014; Micu, 2009), Bulgaria (Brown and Petkova, 2007), Poland (Falarz, 2008), and Croatia (Gajić-Čapka, 2011).

The observed changes in snow depth and snow duration are mainly caused by a shift from solid to liquid precipitation (Serquet et al., 2011; Nikolova et al., 2013) and by more frequent and more intense melt (Klein et al., 2016), both resulting from higher air temperatures during winter and spring. In addition to a general warming trend, large-scale atmospheric circulation patterns such as the North Atlantic Oscillation (NAO) have been shown to influence the snow cover in Europe (Henderson and Leathers, 2010; Bednorz, 2011; Skaugen et al., 2012; Birsan and Dumitrescu, 2014; Buisan et al., 2015). For the Alps, 50% of the snowpack variability seems to be related to the establishment of atmospheric blocking patterns over Europe, although in this case the correlation between the annual snowpack variability and the NAO is weak, and limited to low elevations (Scherrer and Appenzeller, 2006; Durand et al., 2009). At higher elevations, the NAO influence can be detected a through a 'cascade' of processes that include the manner in which the positive or negative NAO modes translates into surface pressure in the Alps, Pyrenees, or Scandes, and thus on temperature and precipitation according to the resulting pressure patterns in the different European mountain regions. Together with the effect of air temperature, this determines the amount of snowfall. In recent decades, this cascade has led to an increased number of warm and dry winter days, which obviously is unfavorable for snow accumulation (Beniston et al., 2011b). In addition, the Atlantic Multi-decadal Oscillation – a natural periodic fluctuation of North Atlantic sea surface temperature – has been shown to affect the variability of alpine spring snowfall, contributing to the decline in snow cover duration (Zampieri et al. 2013).

	Time	Snow variable	Trend at low / high elevation		Source
Alps			**below 2000 m**	**above 2000 m**	
Switzerland	1958-1999	DJF snow cover duration	majority negative	no clear trend	Scherrer at al., 2004
Italy	1950-2009	DJFMA snow cover duration	majority negative	no clear trend	Valt and Cianfarra, 2010
France	1959-2005	Annual snow cover duration	majority negative	many negative	Durand et al., 2009
Scandinavia			**below 1000 m**	above 1000 m	
Norway	1961-2010	Maximum snow depth	majority negative	some positive	Dyrrdal et al., 2013
Finnland	1978-2012	Annual snow cover duration	majority negative	-	Kivinen and Rasmus, 2015
Carpathians			**below 1000 m**	above 1000 m	
Bulgaria	1931-2000	Annual snow cover duration	no clear trend	no clear trend	Brown and Petkova, 2007
Poland	1954-2001	Maximum snow depth	no clear trend	no clear trend	Falarz, 2008

Romania	1961-2003	Annual snow cover duration	no clear trend	no clear trend	Micu, 2009
Pyrenees			**below 1000 m**	**above 1000 m**	
Spain	1975-2002	Annual snow cover duration	majority negative	majority negative	Pons et al., 2010

Table 1: Recent studies of current snow cover trends in the major European mountain regions. Only significant trends are listed. Note that a direct comparison of the sites is difficult since the considered time period and snow variable can differ.

5 The observed changes in snow amounts are often abrupt. Several studies have reported a step-like change for snow depth occurring in the late 1980s (Marty, 2008; Durand et al., 2009; Valt and Cianfarra, 2010) and for snow-covered areas of the Northern Hemisphere (Choi et al., 2010). This step-like development, also observed for other compartments of the environment (Reid et al., 2016), has been suggested to be linked to atmospheric internal variability (Li et al., 2015), as well as shrinking sea-ice extent (Mori et. al. 2014). Since that step-change, the monthly mean snow covered area in the Alps has not decreased

10 significantly (Hüsler et al., 2014), and winter temperatures in large areas of the Northern Hemisphere (Mori et. al. 2014) and in the Swiss Alps (Scherrer et al., 2013) have been stagnating.

Studies analyzing high-magnitude snowfalls are rare, but indicate that extreme snow depths have decreased in Europe (Blanchet et al., 2009; Kunkel et al., 2016), with the exception of higher and colder sites in Norway (Dyrrdal et al., 2013). The

15 decrease in extreme snowfall rates is less clear, except for low elevations where the influence of increasing air temperature is predominant (Marty and Blanchet, 2012). Studies related to past changes in snow avalanche activity are scarce as well, but observations indicate that over the last decades (a) the number of days with prerequisites for avalanches in forests decreased (Teich et al., 2012), (b) the proportion of wet snow avalanches increased (Pielmeier et al., 2013), and (c) the runout altitude of large avalanches retreated upslope (Eckert et al., 2010, 2013; Corona et al., 2013) as a direct consequence of changes in snow

20 cover characteristics (Castebrunet et al., 2012).

2.1.2 Future changes of the snow cover

Regional climate model simulations show a dramatic decrease both in snow cover duration and SWE for Europe by the end of the 21st century (Jylhä et al., 2008). It has to be noted, however, that the projected increase in air temperature for coming decades is accompanied by large uncertainties in changes of winter precipitation. For mainland Europe, climate models show

25 no clear precipitation change until the 2050s, and slightly increasing winter precipitation thereafter. Projections for regional changes in snow cover are thus highly variable, and strongly depend on applied emission scenarios and considered time period (e.g., Marke et al., 2015).

For the Alps at an elevation of 1500 m above sea-level (a.s.l.), recent simulations project a reduction in SWE of 80-90% by

30 the end of the century (Rousselot et al., 2012; Steger et al., 2013; Schmucki et al., 2015a). According to the same simulations,

the snow season at that altitude would start 2-4 weeks later and end 5-10 weeks earlier than in the 1992-2012 average, which is equivalent to a shift in elevation of about 700 m (Marty et al., 2017a). For elevations above 3000 m a.s.l., a decline in SWE of at least 10% is expected by the end of the century even when assuming the largest projected precipitation increase. Future climate will most probably not allow for the existence of a permanent snow cover during summer even at the highest elevations in the Alps, with obvious implications for the remaining glaciers (Magnusson et al., 2010; Bavay et al., 2013) and the thermal condition of the ground (e.g. Marmy et al. 2016; Draebing et al. 2017, Magnin et al. 2017).

Projections for Scandinavia show clear decreases for snow amount and duration. Exceptions are the highest mountains in Northern Scandinavia, where strongly increasing amounts in precipitation could compensate the temperature rise and result in marginal changes only (Räisänen and Eklund, 2012). Simulations for the Pyrenees indicate a decline of the snow cover similar to that found for the Alps (López-Moreno et al., 2009). Again, the dependency on future emissions is significant: For a high emission scenario (RCP8.5), SWE decreases by 78% at the end of the 21st century at 1500 m a.s.l. elevation are expected, whereas a lower emission scenario (RCP6.0) projects a decline of 44%.

Extreme values of snow variables are often the result of a combination of processes (e.g., wind and topographic influence for drifting snow), making predictions of their frequency highly uncertain (IPCC, 2012; 2013). By the end of the 21st century, models suggest a smaller reduction in daily maximum snowfalls than in mean snowfalls over many regions of the Northern Hemisphere (O'Gorman, 2014). An investigation for the Pyrenees (López-Moreno et al., 2011), however, finds a marked decrease in the frequency and intensity of heavy snowfall events below 1000 m a.s.l., and no change in heavier snowfalls for higher elevations. Changes in extreme snowfall and snow depth are also likely to depend on compensation mechanisms between higher temperatures (Nicolet et al., 2016), more intense precipitation, and increased climate variability, rendering any prediction of future snow storm frequency, magnitude, and timing difficult. In addition, most available studies either deal with marginal distributions or postulate stationarity (Blanchet and Davison, 2010; Gaume et al., 2013), two approaches that are questionable in the context of a changing climate.

Changes in snow cover duration and snow depth, as well as other snow properties, will have an effect on various ecosystems: Earlier snowmelt is associated with an anticipation of plant phenology (Pettorelli et al., 2007) which can potentially induce a mismatch between plant blooming and herbivore activity, similar to observations in Arctic regions (Post et al. 2009). In the case of alpine ibex, for example, snow has been found to have a dual effect: Too much winter snow limits adult survival, whereas too little snow produces a mismatch between alpine grass blooming and herbivore needs (Mignatti et al 2012). Abundance of alpine rock ptarmigan populations, on the other hand, has been shown to depend on the onset of spring snowmelt, and the timing of autumn snow cover (Imperio et al., 2013). The increasing number of hard (icy) snow layers due to higher temperatures, on the other hand, can have a significant effect on the life of plants and animals (Johansson et al., 2011).

In addition, human activities will be influenced by the anticipated changes. The reduction of the snow season duration, for example, will have severe consequences for winter tourism (Uhlmann et al., 2008; Steiger and Abegg, 2013; Schmucki et al., 2015b), water management (Laghari et al., 2012; Hill-Clarvis et al., 2014; Gaudard et al., 2014; Köplin et al., 2014) and ecology (Hu et al., 2010; Martz et al., 2016). Similarly, change in moisture content or density of snow will affect infrastructure stability under extreme loading (Sadovský and Sykora, 2013; Favier et al., 2014).

The effect of climate change on avalanche risk is largely unknown; although empirical relations between snow avalanche activity and climate exist (Mock and Birkeland, 2000), the knowledge is insufficient for sound long-term projections. With a few exceptions, existing studies on avalanche-climate interactions focus on recent decades and are very local in scope (Stoffel et al., 2006; Corona et al., 2012, 2013; Schläppy et al., 2014, 2016). Direct effects of climate change on avalanche frequency, timing, magnitude, and type mainly exist in form of changes in snow amounts, snowfall succession, density and stratigraphy as a function of elevation. The trend towards wetter snow avalanches is expected to continue, although the overall avalanche activity will decrease, especially in spring and at low elevations (Martin et al., 2001; Castebrunet et al., 2014). In contrast, an increase in avalanche activity is expected at high elevations in winter due to more favorable conditions for wet-snow avalanches earlier in the season (Castebrunet et al., 2014). Even if the expected rise of tree line elevation may reduce both avalanche frequency and magnitude, present knowledge on avalanche-forest interactions is incomplete (Bebi et al., 2009). Due to the highly non-linear nature of avalanche triggering response to snow and weather inputs (Schweizer et al., 2003) and to the complex relations between temperature, snow amounts, and avalanche dynamics (Bartelt et al., 2012; Naaim et al., 2013), it remains unclear whether warmer temperatures will indeed lead to fewer avalanches because of less snow. The most destructive avalanches, moreover, mostly involve very cold and dry snow resulting from large snowfall, but may also result from wet snow events whose frequency has increased in the past (Sovilla et al., 2010; Castebrunet et al., 2014; Ancey and Bain, 2015). Finally, mass movements involving snow often occur at very local scales, making them difficult to relate to climate model outputs, even with downscaling methods (Rousselot et al., 2012; Kotlarski et al., 2014).

2.2 Changes in glaciers

Mountain glaciers are a key indicator of rapid and global climate change. They are important for water supply as they modulate the water cycle at different temporal and spatial scales, affecting irrigation, hydropower production, and tourism. Evaluating the retreat or complete disappearance of mountain glaciers in response to climate change is important to estimate impacts on water resources (e.g. Kaser et al., 2010; Pellicciotti et al., 2014), and to anticipate natural hazards related to glacier retreat, e.g., ice avalanches or the formation of new lakes (Frey et al., 2010; Gilbert et al., 2012; Faillettaz et al., 2015; Haeberli et al., 2016a, b).

2.2.1 Observed changes in glaciers

Glaciers in mainland Europe cover an area of nearly 5,000 km^2 (Table 2) and have an estimated volume of almost 400 km^3 (Huss and Farinotti, 2012; Andreassen et al., 2015). From historical documents such as paintings and photography (Zumbühl et al., 2008), it is clear that glaciers have undergone substantial mass loss since the 19th century (Fig. 2) and that the pace of mass loss has been increasing (Zemp et al., 2015). A loss of 49% in the ice volume was estimated for the European Alps for the period 1900-2011 (Huss, 2012). Repeat inventories have shown a reduction in glacier area of 11% in Norway between 1960 and the 2000s (-0.28%/yr) (Winsvold et al., 2014), and 28% in Switzerland between 1973 and 2010 (-0.76%/yr) (Fischer et al., 2014). Periods with positive surface mass balance have, however, occurred intermittently, notably from the 1960s to the mid-1980s in the Alps, and in the 1990s and 2000s for maritime glaciers in Norway (Zemp et al., 2015; Andreassen et al., 2016). Glacier area loss has led to the disintegration of many glaciers, which has also affected the observational network (e.g., Zemp et al., 2009; Carturan et al., 2016). In the more southerly parts of Europe, glacier retreat in Pyrenees has accelerated since the 1980s and the small glaciers are currently mostly in a critical situation (López-moreno et al. 2016, Rico et al., 2017)

Country	Area (km²)	Volume (km³)	Year	Reference
Norway	2692	271	1999-2006	Andreassen et al., 2012b
Sweden	262	12	2002	Brown and Hansson, 2004
Switzerland	943	67	2008-2011	Fischer et al., 2014
Austria	415	17	2006	Abermann et al., 2009
Italy	370	18	2005-2011	Smiraglia and Diolaiuti, 2015
France (Alps)	275	13	2006-2009	Gardent et al., 2014
France-Spain Pyrenees	3	<1	2011	Marti et al., 2015
TOTAL	4960	399		

Table 2: Distribution of glacier area and volume in continental Europe and mainland Scandinavia. Years of reference and respective publications are given for the glacier area. Ice volume estimates refer to 2003 for continental Europe and Sweden (Huss and Farinotti, 2012), and to 1999-2006 for Norway (Andreassen et al., 2015). Uncertainties in ice volume are in the order of 10-20%.

Glacier retreat during the 20th century has mainly been attributed to changes in atmospheric energy fluxes and associated air temperature changes. A good correlation between air temperature and melt exist, making long-term air temperature time series the favorite option to explain 20th century glacier retreat (Haeberli and Beniston, 1998). High melt rates in the 1940s have, however, also been associated to changes in solar radiation (Huss et al., 2009). Several studies used calibrated temperature-index methods to simulate snow and ice melt responses to atmospheric forcing (Braithwaite and Olesen, 1989; Pellicciotti et al., 2005) although the appropriateness of such approaches for long-term studies has often been debated (Huss et al., 2009; Gabbi et al., 2014; Réveillet et al., 2017). Glacier response to atmospheric forcing is driven by different factors, ranging from

synoptic weather patterns to local effects enhanced by topography. The latter influences, among others, the distribution of precipitation, solar radiation, and wind. Several studies have shown that glacier mass balance can be influenced by the NAO. Glacier advances in Scandinavia during 1989-1995, for example, are attributed to increased winter precipitation linked to the positive NAO phase during that period (Rasmussen and Conway, 2005). In the European Alps, the relationship between the

5 NAO and glacier surface mass balance is less pronounced (Marzeion and Nesje, 2012; Thibert et al., 2013). This is essentially because the Alps are often a "pivotal zone" between southern and northern Europe, where the correlations between the NAO index and temperature or precipitation tend to be generally stronger (i.e., in the Mediterranean zone and in Scandinavia).

Glacier evolution during the 20th century also highlights the importance of the surface albedo feedback, as albedo governs the

10 shortwave radiation budget at the glacier surface, which is the dominant energy source for melting. The sensitivity of ablation to albedo has generally been assessed using energy-balance considerations (Six and Vincent, 2014) or degree-day approaches (e.g., Pellicciotti et al., 2005). Oerlemans et al. (2009) and Gabbi et al. (2015) investigated the influence of accumulation of dust or black carbon on melt rates for Swiss glaciers in the last decades, revealing annual melt rates increased by 15-19% compared to pure snow. Monitoring, reconstructing, or modelling the surface albedo of glaciers is challenging (Brock et al.,

15 2000) as its spatial and temporal evolution is linked to changes in surface properties (mainly snow grain size and grain shape) and to the deposition of impurities on the ice. Albedo changes are also determined by snow deposition (amount and spatial distribution), making the annual surface mass balance highly sensitive to snow accumulation (Réveillet et al., 2017). Properly quantifying the amount and distribution of accumulation over glaciers is therefore a key to better assess the glacier surface mass balance sensitivity to changes in climate, and to simulate its future evolution (Sold et al., 2013).

Glacier dynamics are influenced by numerous variables such as mass change and basal hydrology for temperate glaciers, and by ice temperature changes for cold glaciers. In temperate glaciers, ice dynamics is mainly driven by thickness changes and the basal hydrological system which in turn affects basal sliding. The large decrease in ice thicknesses over the last three decades has led to a strong reduction in ice flow velocities (Berthier and Vincent, 2012). Increased water pressure, on the other

25 hand, reduces the frictional drag and thus increases the sliding rate. Sliding velocities are low if the water under glaciers drains through channels at low pressure, and high if the water drains through interconnected cavities (Röthlisberger, 1972; Schoof, 2010). Although changes in seasonal ice flow velocities are driven by subglacial hydrology it seems that, at the annual to multi-annual time scales, the ice flow velocity changes do not depend on changes in subglacial runoff (Vincent and Moreau, 2016). A few temperate alpine glaciers, such as Belvedere glacier in Italy, have shown large accelerations due to a change in subglacial

30 hydrology (Haeberli et al., 2002), whereby the mechanisms of this surge-type movement remain unclear. In some rare cases, the reduction of the efficiency of the drainage network followed by a pulse of subglacial water triggered a catastrophic break-off event as in the case of Allalingletscher in 1965 and 2000 (Faillettaz et al., 2015).

Studies of cold glaciers in the Monte Rosa and Mont Blanc area revealed that englacial temperatures have strongly increased over the last three decades due to rising air temperatures and latent heat released by surface meltwater refreezing within the glacier (Lüthi and Funk, 2000; Hoelzle et al., 2011; Gilbert et al., 2014). A progressive warming of the ice is expected to occur and propagate downstream. As a result, changes of basal conditions could have large consequences on the stability of hanging glaciers (Gilbert et al., 2014). Such changes in basal conditions are understood to be responsible for, e.g., the complete break-off of Altelsgletscher in 1895 (Faillettaz et al., 2015).

2.2.2 Future evolution of European glaciers

Over the last two decades, various studies on potential future glacier retreat in Europe have been published. These can be broadly classified into site-specific (e.g., Giesen and Oerlemans, 2010) and regional studies (e.g., Salzmann et al., 2012). Methods range from simple extrapolation of past surface or length changes, to complex modelling of glacier mass balance and ice flow dynamics. Similarly, applied models range from simple degree-day approaches (e.g., Braithwaite and Zhang, 1999; Radic and Hock, 2006; Engelhardt et al., 2015) to complete surface energy balance formulations (e.g., Gerbaux et al., 2005), or from simple parameterizations of glacier geometry change (Zemp et al., 2006; Huss et al., 2010; Linsbauer et al., 2013), over flowline models (e.g., Oerlemans, 1997; Oerlemans et al., 1998) to three-dimensional ice-flow models solving the full Stokes equations (Le Meur et al., 2004; Jouvet et al., 2011; Zekollari et al., 2014). All models indicate a substantial reduction of glacier ice volume in the European Alps and Scandinavia by the end of the century. Small glaciers are likely to completely disappear (Linsbauer et al., 2013), and even large valley glaciers, such as Great Aletschgletscher, Rhonegletscher, Vadet da Morteratsch (Switzerland), or ice caps as Hardangerjøkulen and Spørteggbreen (Norway) are expected to lose up to 90% of their current volume (Jouvet et al., 2009, 2011; Giesen and Oerlemans, 2010; Farinotti et al., 2012; Laumann and Nesje, 2014; Zekollari et al., 2014; Åkesson et al., 2017). Many glacier tongues will disappear, including for example the one from Briksdalsbreen, the outlet glacier of mainland Europe's largest ice cap Jostedalsbreen (Laumann and Nesje, 2009).

At the scale of mountain ranges, model studies relying on medium-range emission scenarios consistently predict relative volume losses of 76-97% for the European Alps, and of 64-81% for Scandinavia (Marzeion et al., 2012; Radic et al., 2014; Huss and Hock, 2015) for the 21st century. Since the mountain glaciers in Europe are far out-of-balance with the present climate (e.g., Andreassen et al, 2012a; Mernild et al, 2013), such volume losses must be expected even with strong efforts to reduce CO_2 emissions and to stabilize global warming at less than +2°C as recommended by the Paris COP-21 climate accord (Huss, 2012; Salzmann et al., 2012). Due to their limited altitudinal extent, many glaciers are unable to reach a new equilibrium with climate even if air temperatures are stabilized by the end of this century. Furthermore, ice caps in Norway that contribute to a large part of the total ice volume in Europe (Table 2), are highly sensitive to mass balance–altitude feedback due to their hypsometry and large ice thicknesses. Model experiments suggest that Hardangerjøkulen will not regrow with its present mass balance regime once it has disappeared (Åkesson, 2017). However, uncertainties in projections of future glacier evolution are

still considerable and improvements are required in both the quality of the input data and the physical basis upon which glaciological models are built (see Section 3.1).

2.3 Changes in permafrost

5 Permafrost is defined as lithospheric material with temperatures continuously below 0°C, and covers approximately 20 million km^2 of the Earth surface, with a fourth of it being located in mountainous terrain (Gruber, 2012). Although the understanding of the thermal state of permafrost has increased significantly within the recent past, knowledge gaps still exist regarding the volume of permafrost ice stored in Europe, its potential impact on future water resources, and its effect on slope stability, including processes leading to permafrost degradation and talik formation (Harris et al., 2009; Etzelmüller, 2013; Haeberli,

10 2013).

2.3.1 Observed changes in permafrost and in rock glacier flow velocities and ice volume

Permafrost borehole temperatures are monitored in many European mountain ranges (documented and available in the Global Terrestrial Network for Permafrost GTN-P database, Biskaborn et al., 2015), several of the sites being accompanied by meteorological stations and ground surface temperature measurements (Gisnås et al., 2014; Staub et al., 2016). However, as

15 mountain permafrost is usually invisible from the surface, various indirect methods need to be employed to detect, characterize and monitor permafrost occurrences. These methods include surface-based geophysical measurements to determine the physical properties of the subsurface, including water and ice content distributions (Kneisel et al., 2008; Hauck, 2013), and geodetic and kinematic measurements to detect subsidence, creep and slope instabilities (Kääb, 2008; Lugon and Stoffel, 2010; Kaufmann, 2012; Kenner et al., 2014; Arenson et al. 2016).

The longest time series of borehole temperatures in Europe started in 1987 at the Murtèl-Corvatsch rock glacier in the Swiss Alps (Haeberli et al., 1998; Fig. 3), a period that is much shorter compared to the available ones for the other cryospheric components such as snow (cf. Section 2.1) or glaciers (cf. Section 2.2). The past evolution of permafrost at centennial timescales can to some extent be reconstructed from temperature profiles in deep permafrost boreholes (e.g., Isaksen et al.

25 2007), pointing at decadal warming rates at the permafrost table in the order of 0.04°–0.07°C for Northern Scandinavia. Permafrost has been warming globally since the beginning of the measurements (Romanovsky et al., 2010; Noetzli et al., 2016; Fig 3). This warming was accompanied by an increase of the thickness of the seasonal thaw layer (hereafter referred to as the active layer; Noetzli et al., 2016). The considerable year-to-year variability can be linked to variations in the snow cover, as a reduction in snow cover thickness reduces thermal insulation. Latent heat effects associated with thawing mask the recent

30 warming trend for "warm" permafrost sites (temperatures close to the freezing point), which is otherwise clearly visible in cold permafrost (see Fig. 3).

The increasing trend in permafrost temperatures and especially the deepening of the active layer has been hypothesized to lead to an increased frequency of slope instabilities in mountain ranges, including debris flows and rockfalls (Gruber and Haeberli, 2009; Harris et al., 2009; Bommer et al., 2010; Stoffel, 2010; Fischer et al., 2012; Etzelmüller, 2013; Stoffel et al., 2014a, b). The disposition conditions and triggering mechanisms of slope instabilities can be diverse, and depend on subsurface material

5   (e.g., unconsolidated sediments versus bedrock), its characteristics (fractures and fissures, ice and water content, slope angle, geological layering), and changes of these properties with time (Hasler et al., 2012; Krautblatter et al., 2012; Ravanel et al., 2013; Phillips et al., 2017). Water infiltration into newly thawed parts of permafrost is often mentioned as a possible triggering mechanism (Hasler et al., 2012), but only few observational data are available to confirm this hypothesis. By means of tree-ring reconstructions (Stoffel et al., 2010; Stoffel and Corona, 2014), the temporal evolution of debris-flow frequencies has

10 been addressed for a series of high-elevation catchments in the Swiss Alps. These studies point to increased debris-flow activity as a result of climate warming since the end of the Little Ice Age (Stoffel et al., 2008; Bollschweiler and Stoffel, 2010a,b; Schneuwly-Bollschweiler and Stoffel, 2012) and a dependence of debris-flow magnitudes due to instabilities in the permafrost bodies at the source areas of debris flows (Lugon and Stoffel, 2010; Stoffel, 2010).

15 Several studies have documented recent events of rock slope failures in the Alps (Ravanel et al., 2010; Ravanel and Deline, 2011; Huggel et al., 2012; Allen and Huggel, 2013). Some of these failures are clearly related to deglaciation processes (Fischer et al., 2012; Korup et al., 2012; Strozzi et al., 2010). Unusually high air temperatures have additionally been associated with these processes as the penetration of meltwater from snow and ice into cleft systems results in a reduction of shear strength and enhanced slope deformation (Hasler et al., 2012). Considering the multiple factors that affect rock slope stability, however,

20 it is generally difficult to attribute individual events to a single one (Huggel et al., 2013). Improved integrative assessments are therefore necessary.

Further evidence of climatic impacts on high-mountain rock slope stability comes from the analysis of historical events. For the Alps, inventories documenting such events exist since 1990 (Ravanel and Deline, 2011; Huggel et al., 2012) and indicate

25 a sharp increase in the number of events since 1990. This makes the temporal distribution of rock slope failures resembles the evolution of mean annual temperatures. Given the fact that monitoring and documentation efforts have been intensified during the past decades, it remains unclear to which degree this correlation is affected by varying temporal completeness of the underlying datasets.

30 Data on the ice volume stored in permafrost and rock glaciers are still scarce. To date, hydrologically oriented permafrost studies have been utilizing remote sensing and meteorological data for larger areas, or have had a regionally constrained scope such as e.g. the Andes (Schrott, 1996; Brenning, 2005; Arenson and Jacob, 2010; Rangecroft et al., 2015), the Sierra Nevada (Millar et al., 2013) or Central Asia (Sorg et al., 2015; Gao et al., 2016). To our knowledge, no systematic studies exist on permafrost-hydrology interactions for the European mountain ranges to date.

In site-specific model studies, subsurface data are only available from borehole drillings and geophysical surveying. The models used often have originated from high-resolution hydrological models (e.g. GEOtop, Endrizzi et al. 2014), soil models (e.g. COUP model, Jansson 2012, Marmy et al. 2016) or snow models (such as Alpine3D/Snowpack, Lehning et al. 2006,

5   Haberkorn et al. 2017) and have been successfully extended to simulate permafrost processes. Recently, explicit permafrost models have been developed as well (e.g., Cryogrid 3, Westermann et al. 2016).

Because of its complexity, permafrost evolution cannot be assessed by thermal monitoring alone. Kinematic and geophysical techniques are required for detailed process studies. Kinematic methods are used to monitor moving permafrost bodies (e.g.,

10   rock glaciers) and surface geometry changes. Hereby, methods based on remote sensing allow for kinematic analyses over large scales (Barboux et al., 2014, 2015; Necsoiu et al., 2016) and the compilation of rock glacier inventories (e.g., Schmid et al., 2015), whereas ground-based and airborne kinematic methods focus on localized regions and on the detection of permafrost degradation over longer time-scales (Kaufmann, 2012; Klug et al., 2012; Barboux et al., 2014; Müller et al., 2014, Kenner et al., 2014, 2016; Wirz et al., 2014, 2016). Long-term monitoring of creeping permafrost bodies shows an acceleration in motion

15   during recent years, possibly related to increasing ground temperatures and higher internal water content (Delaloye et al., 2008; Ikeda et al., 2008; Permos, 2016; Scotti et al. 2016; Hartl et al. 2016). The kinematic monitoring methods mentioned above, however, cannot be used for monitoring of permafrost bodies without movement or surface deformation (e.g., sediments with medium to low ice contents, rock plateaus, gentle rock slopes etc.). Remote sensing has so far not enabled thermal changes in permafrost to be assessed.

Geophysical methods can detect permafrost, and characterize its subsurface ice and water contents (Kneisel et al., 2008; Hauck, 2013). They also provide structural information such as active-layer and bedrock depths. In recent years, repeated geoelectrical surveys have been applied to determine ice and water content changes, thus complementing temperature monitoring in boreholes (Hilbich et al., 2008a; Pellet et al., 2016). Results from such Electrical Resistivity Tomography (ERT) monitoring

25   show that permafrost thaw in mountainous terrain is often accompanied by a drying of the subsurface, as the water from the melted permafrost often leaves the system downslope and is not always substituted in the following summer (Hilbich et al., 2008a; Isaksen et al., 2011). A 15-year ERT time series from Schilthorn, Swiss Alps, shows for example a clear decreasing trend of electrical resistivity, corresponding to ice melt, throughout the entire profile below the active layer (Fig. 4). The corresponding temperature at 10 m depth is at the freezing point, and shows no clear trend. ERT is increasingly used in

30   operational permafrost monitoring networks to determine long-term changes in permafrost ice content (Hilbich et al., 2008b, 2011; Supper et al., 2014; Doetsch et al., 2015; Pogliotti et al., 2015).

2.3.2 Future evolution of European permafrost

Physically-based models of varying complexity are employed for process studies of permafrost (for a review see Riseborough et al., 2008; Etzelmüller, 2013) and specifically for the analysis of future permafrost evolution. These models should not be confused with permafrost distribution models (Boeckli et al., 2012, Gisnaas et al. 2016, Deluigi et al. 2017), which are statistical and often based on rock glacier inventories and/or topo-climatic variables such as potential incoming solar radiation and mean annual air temperature. Physically-based site-level models are used in combination with Regional Climate Models (RCM) for studies of long-term permafrost evolution (Farbrot et al., 2013; Scherler et al., 2013; Westermann et al., 2013; Marmy et al., 2016), similar to land-surface schemes used for hemispheric permafrost modelling (Ekici et al., 2015; Chadburn et al., 2015; Peng et al., 2016). Physically-based models are also used to explain the existence of low-altitude permafrost occurrences (Wicky and Hauck 2016) and to analyse the dominant processes for the future evolution of specific permafrost occurrences in the European mountains (Scherler et al., 2014; Fiddes et al., 2015; Zhou et al., 2015; Haberkorn et al., 2017; Lüthi et al. 2017). Simulations for different mountain ranges in Europe suggest an overall permafrost warming and a deepening of the active layer until the end of the century (see Fig. 5 for four examples from the Swiss Alps; similar simulations from Scandinavia are found in Hipp et al., 2012; Westermann et al., 2013, 2015; Farbrot et al., 2013).

The projected increase in permafrost temperatures is mainly due to the anticipated increase in air temperatures. The latter also causes the snow cover duration to decrease, thereby reducing the thermal insulation effect (Scherler et al. 2013, Marmy et al. 2016). In spite of similar trends in RCM-driven permafrost studies, comprehensive regional-scale maps or trends for projected permafrost changes in Europe are not available to date. This is partly because of the insufficient borehole data, but mainly because of the large heterogeneity of the permafrost in European mountain ranges. The latter strongly depends on surface and subsurface characteristics (e.g., fractures/unfractured rock, fine/coarse-grained sediments, porosity, etc), microclimatic factors (snow cover, energy balance of the whole atmosphere/active layer system, convection in the active layer, etc.) and topo-climatic factors (elevation, aspect, slope angle).

The largest and most important impacts related to permafrost thawing have yet to occur. Along with changes in precipitation, permafrost thawing is projected to affect the frequency and magnitude of mass wasting processes in mountain environments (IPCC, 2012). This is especially true for processes driven by water, such as debris flows (Stoffel and Huggel, 2012; Borga et al., 2014). Based on statistically downscaled RCM data and an assessment of sediment availability, Stoffel et al. (2011, 2014a, b) concluded that the temporal frequency of debris flows is unlikely to change significantly by the mid-21st century, but is likely to decrease during the second part of the century, especially in summer. At the same time, the magnitude of events might increase due to larger sediment availability. This is particularly true in summer and fall when the active layer of the permafrost bodies is largest, thus allowing for large volumes of sediment to be mobilized (Lugon and Stoffel, 2010). Accelerations of rock-glacier bodies might play an additional role (Stoffel and Huggel, 2012). Providing projections for future sediment

availability and release for areas that are experiencing permafrost degradation and glacier retreat is particularly important in the European Alps, where the exposure of people and infrastructure to hazards related to mass movements is high (Haeberli, 2013).

5   Finally, it should be noted that in contrast to glacier melting, permafrost thawing is an extremely slow process (due to the slow downward propagation of a thermal signal to larger depths, and additional latent heat effects). As permafrost in European mountains is often as thick as 100 m, a complete degradation is therefore unlikely within this century.

2.4 Changes in meltwater hydrology

In spring, summer, and autumn, seasonal snow and glacier ice are released as meltwater into the headwaters of the alpine water
10   systems. Because of the temporally shifted release of water previously stored as snow and ice and the significant surplus of precipitation compared to the forelands, mountains have often been referred to as "water towers" (Mountain Agenda, 1998; Viviroli et al., 2007). The meltwater contribution to streamflow is important for millions of people downstream (Kaser et al., 2010). The Alps, in particular, are the water source for important rivers that flow into the North Sea (Rhine), the Black Sea (Danube) and the Mediterranean Sea (Rhône and Po); a comprehensive overview of the major alpine water systems is given
15   in EEA (2009).

The most important seasonal runoff signal in the Alps is the melt of snow (Beniston, 2012). This is because the precipitation distribution is fairly even throughout the year, and because the amount of water retained in and released from reservoirs and lakes is only a small fraction of the total water volume (Schaefli et al., 2007; López-Moreno et al., 2014). Temperature-induced
20   changes in streamflow (such as rain-to-snow fraction, seasonal shift of snowmelt, glacier runoff contribution) are generally better understood than the ones caused by changing spatio-temporal precipitation patterns (Blaschke et al., 2011). Nevertheless, understanding long-term trends in runoff requires an accurate estimate of the amount and distribution of snow accumulation during winter (Magnusson et al., 2011; Huss et al., 2014). The response of snowmelt to changes in air temperature and precipitation is influenced by the complex interactions between climatic conditions, topography and wind redistribution of
25   snow (López-Moreno et al., 2012; Lafaysse et al., 2014).

Several national assessments have addressed the hydrologic changes in alpine river water systems, highlighting important regional differences (FOEN, 2012; APCC, 2014); these reports contain a wealth of specific literature. Regional peculiarities are the result of spatial differences in temperature and precipitation changes, although other factors such as local land-use
30   changes or river corrections may play a role as well (EEA, 2004).

Compared to snowmelt, the total ice-melt volume from glaciers in the Alps is minor. At sub-annual scales, however, contributions from glacierized surfaces can be significant not only for the headwater catchments close to the glaciers (Hanzer

et al., 2016), but also for larger basins where glacierization is small (Huss, 2011). This is particularly true during summer when specific runoff yield from glacierized areas is much higher than from non-glacierized ones (Farinotti et al., 2016). In a warming climate with retreating glaciers this also holds for annual scales, as additional meltwater is released from ice storage that has accumulated over long time periods.

In general, it can be stated that in large catchments the ice melt component in streamflow results from the contribution of many individual glaciers of varying size and setting. At the decadal or longer scale, the ice melt from these individual glaciers might both be rising or declining, depending on glacier size and climatic trend, but as a result of the superposition of the many different contributions the resulting streamflow of a regional, glacierized catchment might not show a long-term trend, as has

10   been simulated for the Rhine (KHR/CHR 2016), for example. In this latter study, glaciers were considered according to a further developed δh-method (Huss 2010), allowing for both glacier retreat and transient glacier advances as required in long-term simulations: daily fractions of the rain, snowmelt and glacier ice melt components of streamflow were determined for the Rhine basin from 1901 to 2006, with highest ice melt contributions during the periods with negative mass balances in the 1940s and 1980s.

Scenarios of changing streamflow affected by retreating glaciers in a warming climate have recently been developed in various physically-based, distributed modelling experiments (Weber and Prasch, 2009; Prasch et al., 2011; Hanzer et al., 2017). Fig. 6 illustrates future streamflow of a currently highly glacierized catchment (roughly 35% glacierization) in the Austrian Alps. Even for the moderate RCP2.6 scenario (IPCC, 2013) which corresponds roughly to the COP-21 "2°C policy", the glacier melt

20   contribution to runoff becomes very small by the end of the century. In the second half of the century, summer runoff amounts decrease strongly with simultaneously increasing spring runoff. While in the RCP2.6 scenario the month of peak runoff remains unchanged, RCP4.5 and RCP8.5 project the peak to gradually shift from July towards June. Alpine streamflow will hence undergo a regime shift from glacial/glacio-nival to nivo-glacial, i.e., the timing of maximum discharge will generally move from the summer months to spring (Beniston, 2003; Jansson et al., 2003; Collins, 2008; Farinotti et al., 2012; Prasch et

25   al., 2011; Hanzer et al., 2017). For many streams utilized for hydropower generation, this phenomenon can be superimposed by the effects of discharge regulation. The regimes in Fig. 6 indicate that (a) the effect of warming increases after the mid of the century for all scenarios, (b) the effect is of the same order of magnitude as the one of the choice of climate model, and (c) the timing of the maximum contribution of ice melt to streamflow – referred to as "peak water" – has already passed, i.e., that the effect of declining glacier area already overrides the increasing melt caused by rising temperatures. The time of occurrence

30   for "peak water" mainly depends on the size of the glacier, and can hence differ for adjacent catchments (Hanzer et al. 2017). Until the middle of the century and for large scales, the decrease of annual streamflow due to glacier retreat is expected to be small.

By 2100, the glaciers in the Alps are projected to lose up to 90% of their current volume (Beniston, 2012; Pellicciotti et al., 2014; Hanzer et al., 2017). By then, peak discharge is likely to occur 1-2 months earlier in the year (Horton et al., 2006) depending on carbon-emission scenario. In Switzerland, a new type of flow regime called "*pluvial de transition*" (transition to pluvial) was introduced to classify such newly emerging runoff patterns (SGHL/CHy, 2011; FOEN, 2012). Regime shifts have long been recognized and can be interpreted as the prolongation of observed time series – the longest one in the Alps being the recorded water level and streamflow discharge of the Rhine river at Basel since 1808. Some investigations, however, show that annual runoff totals may change only little, as the overall change resulting from reductions in snow and ice melt, changing precipitation, and increased evapotranspiration is unclear (SGHL/CHy, 2011; Prasch et al., 2011). Other studies, instead, highlight the significance of future regime shifts in headwater catchments (Pellicciotti et al. 2014). Obviously, the complex interplay of snow- and icemelt contribution to discharge in a changing climate, combined with the other processes determining the streamflow regime, and their scale dependencies are not yet fully understood. There is general consensus that only a few high-altitude regions of the Alps will continue to have a glacial regime in the long term (FOEN, 2012). Further south in Europe, a loss of importance of snowmelt runoff in the last three decades has been detected for the majority of the mountain rivers in the Iberian Peninsula (Morán-Tejeda et al., 2014). For the Scandinavian mountains, the observed increase of runoff in winter and spring and the earlier snowmelt are projected for the future (RCP8.5), with relatively small changes of total annual runoff (NCCS 2017). This holds for all scenarios, with the changes being less in the moderate scenarios RCP4.5 and RCP2.6, respectively.

More than by changes in the cryosphere, hydrolgical extremes in the alpine region are affected by the changing climate itself. Periods of persistent and exceptional dry conditions have been identified based on a simple drought index around the mid 1940s and for the late 1850s to the 1870s (van der Schrier et al., 2007). For the 21st century, Heinrich et al. (2013) found increasing temperatures and negative trends of precipitation in the ENSEMBLES multi model dataset (reference period 1961-1990), the latter mainly in the southern and eastern parts of the Alps (Brunetti et al. 2009). Drought conditions are expected to increase until the end of the 21st century due to higher air temperatures, larger evapotranspiration rates, increased water use, and less precipitation (Calanca, 2007). The regions south of the Alps are thereby particularly affected.

Flood frequency has increased during the past 30 years in 20% of the river basins in Austria, mainly in small catchments North of the main Alpine ridge (Blöschl et al., 2013). For Switzerland, Allamano et al. (2009) found increasing flood peaks over the course of the last century, caused by increasing air temperature and precipitation. However, Schmocker-Fackel and Naef (2010) showed that this increase is comparable to known past periods of increased flood frequency, and apparent changes in flood frequency can also be attributable to construction measures in the river reaches and the loss of regulation reservoirs. In the future, summer floods might occur less frequently across the entire Alpine region since the frequency of wet days is projected to substantially decrease in summer (Rajcak et al. 2013). The magnitude and frequency of winter and spring floods, however, might increase due to higher intensity of extreme precipitation events in all seasons and for most regions in the Alps

(Christensen and Christensen, 2007; Stoffel et al., 2016). In glacierized catchments, this effect adds to the regime shift with the streamflow maximum occuring earlier in spring after the moment of „peak water". Additionally, more frequent rain-on-snow (ROS) events can add to liquid precipitation if air temperatures continue to rise (Würzer et al., 2016a, b).

5   Despite the general trend towards drier summers, indications for more frequent severe flooding due to heavy or extended precipitation events in the future have been found (Christensen and Christensen, 2007; Stoffel et al., 2016). Although summer floods might occur less frequently, the magnitude and frequency of winter and spring floods might increase since more frequent rain-on-snow (ROS) events can add to liquid precipitation if air temperatures continue to rise (Würzer et al., 2016a, b). It must be noted, however, that the severity and frequency of ROS events will tend to decrease in the future, not just because of rising

10   temperatures, but when they become sufficiently warm to substantially reduce mountain snow cover and thus the potential for catastrophic consequences (Beniston and Stoffel, 2016). For the Scandinavian mountains, the recent tendency to increased frequency of rain floods is projected to continue, and their magnitude to increase. Meltwater-induced floods, however, will decrease over time (NCCS 2017), as is the deficit of soil moisture towards the end of the century. Many of the references discussed in the previous chapters 2.1 (snow), 2.2 (glaciers) and 2.2 (permafrost) are of relevance here too.

Concerning droughts in the alpine region, periods of persistent and exceptional dry conditions have been identified around the mid 1940s and for the late 1850s to the 1870s (van der Schrier et al., 2007). In the future, drought conditions are expected to increase due to higher air temperatures, larger evapotranspiration rates, increased water use, and less precipitation (Calanca, 2007). The regions South of the Alps are thereby particularly affected. Further south in Europe, a loss of importance of

20   snowmelt runoff in the last three decades has been detected for the majority of the mountain rivers in the Iberian Peninsula (Morán-Tejeda et al., 2014).

2.5 Impacts on downstream water management
Montain agriculture, hydropower and tourism are directly dependent on alpine headwaters and they will need to adapt to the

25   changes in water availability and its seasonal distribution as a consequence of decreasing role of snow and glacier melting in the hydrology as outlined in section 2.4. Different scenarios therefore need to be considered, depending on how governance will cope with water-related conflicts that may arise from changes in water availability and demand (Nelson et al., 2007; Beniston et al., 2011a).

30   **2.5.1 Agriculture**

Shifts in agricultural production are expected with climate change as a consequence of higher water demand from crops and less water available due to higher temperatures and longer dry-spells, but also as consequence of earlier snowmelt and less glacier contribution (Jaggard et al., 2010; Gornall et al., 2010). Most studies for alpine regions project reduced soil water

content as a result of increasing evaporation, thinner snowpack and earlier snowmelt (Wu et al., 2015; Barnhart, 2017). This will lead to increased water demand for irrigation (Jasper et al., 2004; Schaldach et al. 2012; Riediger et al. 2014), and will add to the changes in water availability resulting from changing snow and glacier melt (Smith et al., 2014). The effects of more frequent climatic and hydrological droughts in the future (Gobiet et al., 2014) will affect both croplands and grasslands. In

5   Switzerland, the latter cover around 75% of the agricultural land and sustain domestic meat and dairy production (Fuhrer et al., 2006). The majority of crops currently cultivated in the Alps have been shown to be very sensitive to precipitation deficits in the growing season (Fuhrer et al., 2006; Smith et al., 2014). High irrigation demands will thus likely put additional pressure on rivers, especially small ones as they suffer more from inter-annual variability (Smith et al., 2012). Together with generally decreasing summer discharge, this will more frequently create low flow conditions which largely favour increasing water

10 temperatures in streams with negative consequences for water quality and the aquatic fauna. Long-term water-management strategies will be important to face these challenges and to ensure that future agricultural water needs can be met (Riediger et al., 2014).

2.5.2 Hydropower

Climate change is a key driver in electricity markets, as both electricity production and demand are linked to weather and

15 climate (Apadula et al., 2012). As a consequence of earlier snowmelt and reduced water discharge from glaciers, hydropower production potential is expected to increase in winter and spring, and to decline in summer (Hauenstein, 2005; Kumar et al., 2011). Currently, energy demand is higher in winter than in summer, but this may change as rising summer temperatures increase energy requirements for the cooling of buildings (López-Moreno et al., 2008, 2011; Gaudard and Romerio, 2014). A study conducted for the Mattmark dam in the Swiss Alps and for the Val d´Aosta, Italy (Gaudard et al., 2014) revealed that

20 peak hydropower production has so far not been affected by climate change. This is possibly the result of the large existing reservoir volumes which enable to offset seasonal changes (Farinotti et al., 2016). Indeed, it has been suggested that no urgent adaptation of the hydropower infrastructure will be required in Switzerland within the next 25 to 30 years (Haunstein, 2005). For Austria, little changes in annual hydropower production (±5%) have been proyected up to 2050 using A1B SRES, but there could be a significan reduction in summertime power generation (Wagner et al., 2017). Reservoir management, however,

25 will become more challenging as a consequence of higher fluctuations in electricity demands linked to the intermittent production of new renewable energy sources such as photovoltaic and wind power, or biogas (Gaudard and Romerio, 2014). Furthermore, the inter-annual fluctuations in water availability are expected to increase (Gaudard et al., 2014). Run-of-river power plants are expected to be less vulnerable to climate change, as they are usually installed on streamflows with small hydrological fluctuations (Gaudard et al., 2014). Hydropower plants can also be effective in attenuating floods (Harrison and

30 Whittington, 2001). Additional safety concerns include the melting of permafrost and the possibility of more frequent heavy rainfall, resulting in both more frequent slope instabilities and potential flood waves that may endanger power plants (Peizhen et al., 2001; Schwanghart et al., 2016). Increased sediment loads from deglacierized surfaces may additionally affect power generation, in particular by affecting the wear of infrastructure or the silting of storage volumes (Beniston, 2003).

2.5.3 Winter tourism

Increasing air temperatures are expected to result in shorter skiing seasons and a shift of the snow line to higher elevations (Abegg et al., 2007; Steiger, 2010). This will likely lead to smaller number of visitors and reduced revenues, and thus have important economic impacts on alpine winter tourism. Generation of artificial snow is designed to buffer the impact of inter-annual variability of snow conditions, and is increasingly deployed in alpine ski resorts (Uhlmann et al., 2009; Steiger, 2010; Gilaberte et al., 2014; Spandre et al., 2016). In Switzerland, ski slope areas employing artificial snow-making equipment have tripled (from 10 to 33%) from 2000 to 2010 (Pütz et al, 2011). In the French Alps, 32% of the ski slope area was equipped with snow-making facilities in 2014, and this proportion is likely to reach 43% by 2020 (Spandre et al., 2015). In Austria, this share is about 60%, mainly due to the lower average elevations of the Austrian ski areas, and in the Italian Alps, almost 100% of the ski areas are equipped (Rixen et al., 2011). Water consumption for tourism in some Swiss municipalities is high compared to other uses. A study focusing on three tourism destinations in Switzerland, for example, found this consumption to be equivalent to 36% of the drinking water consumption (Rixen et al., 2011). Water and energy demands of ski resorts will increase, which may in turn lead to higher prices for consumers (Gilaberte et al., 2014). Also, summer mountain resorts could be affected by water shortages in the future, thus calling for adapted water management (Roson and Sartori, 2012).

3. Challenges and issues for cryosphere research in European mountains

In the following, we present four points that we consider to be crucial for the future development of cryospheric research. The points were identified with the European mountain cryosphere in mind, but are mostly transferrable to other regions as well. The four points include (1) the numerical modelling of the cryosphere, (2) the understanding of cascading processes, (3) the quantification of precipitation and its phase changes, and (3) issues related to the acquisition, management, and sharing of observational data.

3.1 Modelling of the cryosphere: Spatial scales and physical processes in complex terrain

For simulating the past, models of snow, glaciers and permafrost critically rely on meteorological observations, either from station measurements, spatial interpolation of surface observations or reanalyses, as forcing data. For future simulations, instead, climate models are used. As their current horizontal resolutions, typically in a range between hundreds and tens of kilometres, do not allow a reliable representation of small-scale processes, climate models are generally downscaled using dynamical, statistical, or stochastic downscaling techniques. The outputs are eventually adjusted to take into account model biases in mountain regions (Terzago et al., 2017; Kotlarsky et al. 2014) and then employed to force off-line snow, glaciers and permafrost models. A key point of such approach is to quantify the errors associated to each step of the procedure and to evaluate the propagation of the uncertainties along the modelling chain.

Concerning climate models, an important source of error in the mountainous areas is the coarse representation of the topography, the interaction between topography and atmospheric processes, and the altitudinal temperature gradients, which directly translate into a crude separation of the precipitation phase (Wilcox and Donner 2007, Chen and Knutson, 2008; Wehner et al., 2010, Sillmann et al., 2013). Increased horizontal resolution (Boyle and Klein, 2010) and refinements in the representation of vertical processes – including convection and cloud microphysics (Kang et al., 2015) – are important improvements to be achieved towards a better simulation of mountain processes.

The increasing availability of field data can help in refining the understanding of such processes through direct data-based inference (Diggle and Ribeiro, 2007) and assimilation techniques (Leisenring and Moradkhani, 2011). While such approaches are common to various fields of environmental sciences (e.g., Banerjee et al., 2003), applications in cryospheric research are still rarely found. This is, amongst other, because of some specific difficulties that include the existence of embedded spatial scales (Mott et al., 2011), strong vertical gradients (e.g., temperature, wind speed, phase of precipitation, etc.), and the non-linearity linked phase transitions (Morán-Tejeda et al., 2013). To take full advantage of the ever increasing volume and variety of observational data, the development of adequate statistical models will also be required (Gilks et al., 2001; Wikle, 2003; Cappé et al., 2005). Separating space and time effects has been suggested to be the easiest way to address spatio-temporal data (Cressie and Wikle, 2011). However, temporal evolutions at small spatial scales cannot be inferred this way, so that the application of non-separable spatio-temporal covariance models, for example, seem to have great potential (Gneiting et al., 2007; Genton and Kleiber, 2015).

3.1.1 Snow modelling

The snow pack is affected by a series of small-scale processes, including water transport in snow and firn (Würzer et al., 2016b, Wever et al., 2014; Wever et al., 2016), phase changes (i.e. melt, refreezing, sublimation and condensation), drifting and blowing snow, as well as metamorphism (e.g., Aoki et al., 2011; Pinzer et al. 2012). While the mechanistic understanding of these processes at the point-scale is rapidly increasing (Wever et al., 2014), challenges comes from quantifying their effects at larger scales. Examples of the interplay between large- and small-scale effects include the altered snow distribution after a storm (Lehning et al., 2008; Schirmer et al., 2009) or the change in snow albedo after a melt event. The underlaying small-scale processes (such as snow-grain saltation and metamorphism in the case of wind redistribution and albedo changes, respectively) are reasonably understood at the point scale, but are insufficiently represented or even neglected in large-scale models due to the lack of adequate upscaling schemes. Small-scale snow properties are also essential for the correct interpretation of remote-sensing signals. Ice-lenses or the liquid water content in snow, for example, heavily influence the microwave backscatter, which could be used to infer large scale SWE (Marshall et al., 2007). We argue that the challenges in snow modelling are currently scale dependent.

For modelling snow at the point scale and with a high level of detail, for example in snow stability estimates in the context of avalanche warning, two physically-based models are mostly used: CROCUS (Vionnet et al., 2012) and SNOWPACK (Lehning et al., 1999). Nevertheless, many of the processes described in these models, such as metamorphism and mechanical properties for example, still have a high degree of empirical parameterization (Lehning et al., 2002). The main challenge for snow model

5   development at this scale is the formulation of a consistent theory for snow microstructure (Krol and Lowe, 2016) and its metamorphism.

For hydrological applications, the catchment scale is the most relevant one (Kumar et al., 2013). Snow models of different complexity are used for this (Essery, 2015; Magnusson et al., 2015), and principal challenges are to (a) distinguish between

10   uncertainties introduced by the model structure and uncertainties related to the input data (Schlögl et al., 2016), and (b) develop transferable, site-independent model formulations without the need of calibration. The latter is particularly important for reliable predictions of climate change effects (Bavay et al., 2013) and for model applications to ungauged catchments (Parajka et al., 2013).

15   Large-scale models, finally, use relatively simple, parametric snow schemes, as these require only a small set of input variables and are computationally less expensive (Bokhorst et al., 2016). Current numerical weather prediction systems generally use oversimplified single-layer snow schemes (IFS documentation, 2016; GFS documentation, 2016). Only in some rare cases do these models explicitly represent the liquid water content within the snowpack, or incorporate a refined formulation for snow albedo variability (Dutra et al., 2010; Sultana et al., 2014). Climate models generally resolve the diurnal and seasonal variations

20   of surface snow processes (i.e., surface temperature, heat fluxes) while they simplify the treatment of internal snow processes such as liquid water retention, percolation and refreezing (Armstrong and Brun, 2008; Steger et al., 2013) or the evolution of snow microstructure due to metamorphism. Future research should clarify the degree of complexity required in snow schemes when integrated in large-scale climate models (van den Hurk et al., 2016).

3.1.2 Permafrost modelling

25   Challenges that need to be addressed in permafrost modelling include the development of (1) static large-scale permafrost distribution models, (2) high-resolution and site-specific permafrost evolution models and (3) transient hemispheric permafrost models or land-surface schemes of RCM/GCMs. Current state-of-the-art permafrost distribution models (e.g., Gisnås et al. 2017) are not only forced by statistical and topo-climatic variables such as mean annual air temperature and potential incoming radiation, but also by operationally gridded data-sets of daily air temperature and snow cover. Statistical distributions of snow

30   and other surface characteristics (soil type, roughness) allow for the representation of sub-grid variability of ground temperature (Gubler et al. 2011, Gisnås et al. 2014, 2016). However, the lack of spatial data on subsurface properties (thermal conductivity, porosity, ice content, etc.) prohibits a refined assessment of the permafrost distribution on catchment or local scales, at least for the discontinuous permafrost zone. Acquiring spatial data on subsurface properties as input and validation

data is hereby one of the greatest current challenges in permafrost research (e.g., Hauck, 2013; Etzelmüller, 2013; Gubler et al., 2013).

Model inter-comparison studies using uncalibrated model set-up show that due to the abundance of permafrost-relevant
5   processes in the atmosphere and at the snow surface and subsurface, a detailed simulation of permafrost processes on local scales is impossible without the availability of surface and subsurface data (Ekici et al. 2015). This results from the difficulty of correctly simulating phase changes and corresponding latent heat transfer in permafrost when the initial ground ice content is poorly known. Challenges in regional or hemispheric permafrost modelling therefore include numerical aspects and process-oriented model improvements, as well as data availability and up-scaling issues (Fiddes et al. 2015, Westermann et al. 2015).
10   Most land surface schemes of current GCM's and RCM's now include soil freezing schemes (e.g., McGuire et al. 2016). However, neither reliable ground ice content maps as input nor ground temperature maps for validation exist. Combined with the need of reliable snow, soil moisture and vegetation data, this lack of subsurface information poses the largest uncertainties in estimates of current and future permafrost temperature and spatial distribution.

3.1.3 Glacier modelling

15   To increase the accuracy of glacier mass balance and runoff estimates, the distribution of ice volume and ice thickness is of primary importance. Glacier volume can be estimated from their surface area using scaling approaches (e.g. Bahr et al., 1997). However, such approaches do not provide the ice thickness distribution, which together with the surface mass balance controls the ice dynamics and the response of the glaciers to climate forcing. As it is currently impossible to measure ice thickness distributions of all glaciers individually, model applications are necessary. Several existing glacier models have been compared
20   within the Ice Thickness Models Intercomparison eXperiment (Farinotti et al., 2017), revealing that (i) results largely depend on the quality of the input data (glacier outline, surface elevation, mass balance or velocities) and (ii) model complexity is not directly related to model performance. New high-resolution satellite images make the required input data available, thus opening the way toward improved future global estimates of glacier thicknesses.

25   Another challenge in glacier modelling is the use of approaches that explicitly consider ice dynamics. Jouvet et al. (2009) showed that 3D full-Stokes models representing ice flow without approximation can be applied if the required input data are available (e.g., glacier thickness, surface velocities). At the regional scale, however, simplified approaches are still required (Clarke et al., 2015). For estimating future glacier evolution, ice dynamics models need to be coupled to adequate representations of glacier surface mass balance. The Glacier Model Intercomparison Project (Glacier-MIP, www.climate-
30   cryosphere.org/activities/targeted/glaciermip) assesses the performance of regional to global-scale glacier models to foster the improvement of the individual approaches, and to reduce uncertainties in future projections. There are uncertainties in future meteorological variables and in their downscaling at the glacier scale. Some studies use degree-day approaches for long-term simulations (Réveillet et al., 2017), or also account for potential radiation (Hock, 1999). However, with a change in the relative

magnitude of energy fluxes at the glacier surface, calibrated degree-day factors are likely to change too. Application of models able to resolve the full energy balance are thus required (Hanzer et al., 2016). To be applicable, however, the accuracy and resolution of the corresponding input data needs to be improved. More emphasis on the modelling of winter mass balances and the spatial distribution of snow accumulation (Réveillet et al., 2017), as well as the impact of supraglacial debris and

5   related feedbacks (Reid and Brock, 2010) is required. The latter is particularly important as many glacier tongues become increasingly debris covered as they shrink. Feedback effects of black carbon and aerosols deposition on the glacier surface is also subject of study (Gabbi et al., 2015). Finally, more research on glacial sediment transport and erosion is needed as glacier retreat exposes large amounts of unconsolidated and erodible sediments that might – when entrained – represent a hazard downstream, or reduce the efficiency of hydropower plants (Lane et al., 2016).

10   **3.1.4 Modelling uncertainty**

Predicting the future evolution of cryospheric components is challenging in many respects. On the one hand, these challenges stem from the incomplete understanding of the physical processes leading to given changes, and from the lack of more comprehensive data sets. On the other hand, predictions are intrinsically affected by a range of uncertainties. Adequately evaluating and communicating such uncertainties is an all but a trivial task, both because the interplay between individual

15   systems can be complex, and because end-users of projections are typically uncertainty-adverse. Outside the scientific community, "uncertainty" and "error" are two concepts often not sufficiently distinguished. This can lead to important misunderstanding and misinterpretations. Improving the way uncertainties are communicated is especially important when presenting scientific results to policymakers or stakeholders, as this can significantly affect the level of trust assigned to a particular finding.

3.2 Cascading processes and process chains

The projected increases in air temperatures imply enhanced melting of snow and ice. These, in turn, can lead to glacier recession and glacier debuttressing, an acceleration of creep in perennially frozen talus with high ice contents, and decreasing stability of steep, deeply frozen rock walls (Haeberli, 2013). Under certain conditions, new lakes will form in depressions left

25   by receding glaciers (Haeberli and Linsbauer, 2013; Haeberli et al., 2016a). The largest and most important impacts related to ongoing glacier wasting and permafrost thawing have yet to occur, with likely and potentially drastic impacts on the frequency and magnitude of mass wasting processes (such as rockfalls, rockslides, icefalls, debris flows) (IPCC, 2012). In high mountain environments, where relief energy is substantial, a single event can lead to a chain of reactions or cascading process, thereby gratly amplifying the magnitude of the original event. In the case of new lakes forming underneath hanging glaciers or at the

30   foot of over-steepened and unstable rock slopes (Haeberli et al., 2010), for instance, mass movements into the lake could generate flood waves (Worni et al., 2014). These waves could overtop the lake dam, form a breach and cause a downstream

glacier lake outburst flood (GLOF) with discharges that could potentially be much larger than rainfall-induced floods (Schwanghart et al., 2016).

Another example of a process chain is the 2017 rockfall at Piz Cengalo (Grisons, Switzerland). In that case, the impact on a debris-covered glacier was suggested as a possible cause for the release of sufficient liquid water to transform part of the rockfall into a debris flow. The multiple debris flows, resulting from the August 23, 2017 rockfall, have reached the village of Bondo where they caused substantial damage to an area that would have remained unaffected in the absence of an amplifying chain of reactions. Other process cascades in the Swiss Alps include the assumed liquefaction of avalanche snow into debris flows in October 2011 during an intense rain-on-snow event (Morán-Tejeda et al., 2016). This resulted in massive damage in several regions of the Bernese and Valais Alps and at locations that were not affected by debris flows for decades.

The melting of glaciers and the increasing instability of permafrost bodies will possibly lead to the occurrence of mass movement events beyond historical experience, both in terms of frequency and magnitude (Stoffel and Huggel, 2012). If these processes occur as chain reactions, their consequences can be devastating for the downstream area. For the time being, however, the description of such cascades remains theoretical or anectdotical, and a general effort is required if such pricesses are to be better understood.

Models can be helpful in assessing the hazard potential of process chains, and can provide important insights that would be difficult to obtain by observations alone. Although a number of numerical models have been developed to simulate different types of extreme mass movements, such modelling efforts still face challenges stemming from a lack of process understanding, and the difficulty in measuring the necessary parameters in the field (Worni et al., 2014). In addition, current models were often conceived for the representation of single processes, and are therefore inadequate for analyzing the interactions between processes involved during cascading events. Much effort is required to improve our understanding for how climate changes will affect alpine mass movements.

3.3 Estimating liquid/solid precipitation in complex terrain

Precipitation plays a key role in all cryospheric processes. When temperatures are close to, or beneath the freezing point, precipitation determines the amount of snow accumulation, and significantly influences the energy transfer within permafrost bodies. In its liquid form, instead, it can contribute to water storage within the snowpacks and in glaciers, and steers direct water runoff.

At high elevations, large uncertainties affect the estimates of solid and liquid precipitation (Rasmussen et al., 2012). These uncertainties mainly arise from two facts: (a) The low density of precipitation gauges at high elevation (only 3% of stations worldwide are located above 2000 m a.s.l., and less than 1% above 3000 m a.s.l.; Pepin et al., 2015), which are the regions

where cryospheric processes occur, and (b) the large bias in observations due to precipitation undercatch. The latter is in the order of 30% for high elevation (Adam and Lettenmeier, 2003; Yang et al., 2005) and is particularly large for solid precipitation, since snowfall is importantly influenced by wind, and since icing and riming affect the measurements. Efforts are currently ongoing to address these problems (e.g., Solid Precipitation Intercomparison Experiment SPICE by the WMO, Nitu et al. 2014), but reliable references for ground truth measurements still are not available.

Large uncertainties also exist in the spatial and temporal distribution of precipitation. The spatial distribution, in fact, is not only determined by synoptic systems, but is also strongly affected by topography (Mott et al., 2010). For snow, post-depositional transport such as creep, saltation, suspension, and avalanching additionally influence the spatial distribution. Recent progress in measuring snow distribution in mountains through terrestrial and laser-scanning or radar, for example (Grunewald et al., 2010; Kirchner et al., 2014), has allowed for a better understanding of typical distribution patterns of alpine water resources (Grunewald et al., 2014). When combined with relations between snow depths and SWE (e.g., Jonas et al., 2009), such information can also be used to quantify winter precipitation totals (Scipion et al., 2013; Mott et al., 2014). The results have highlighted, however, that even in highly instrumented mountain ranges such as the Alps, precipitation is still very poorly quantified. The combination of weather modelling with advanced data assimilation techniques, including precipitation radar data, will lead to a more complete understanding of precipitation amounts in high mountains, and this is urgently required if future changes in precipitation are to be projected correctly. In this respect, climate modelling is called for continuing the efforts towards increased model resolution. For cryospheric research, achieving convection-resolving resolution is fundamental, since the process is fundamentally controlling the spatial distribution in high-alpine precipitation.

3.4 Observational data: Access, availability, quality, and spatial and temporal distribution

Availability of observational data is a key for detecting, understanding, and projecting any environmental process (Beniston et al., 2012; Quevauviller et al., 2012). For cryospheric processes, at least two specific issues arise. First, additional efforts should be made to promote the long-term monitoring of key processes such as glacier retreat or permafrost thaw, for which high-quality, homogeneous, long-term observations are rare. In fact, at the moment, only a few long-term monitoring programs comprising meteorological, hydrological and/or glaciological data collection exist, such as for the Oetztal Alps in Austria (Strasser et al. 2017). Second, ground-based observations of the cryosphere are affected by the difficulties in accessing the regions of interest (Klemes 1990). Logistical challenges and related costs are introducing a bias in the availability of cryosphere-related observations. Even in European countries disposing of reasonable financial and infrastructural resources, have such observations a bias towards low and mid elevations, as well as towards easily accesible regions. This also applies to related environmental variables, in particular meteorological variables (Orlowsky and Seneviratne, 2014). Substantial efforts and new technical solutions are required to improve the spatial coverage and representativeness of the variables of interest.

The high logistical and monetary costs of acquiring data at high elevations also explains why many cryosphere related data across Europe still have restricted ownership. The consequences include lack of data, limitations to accessing existing data, delays in obtaining them, and duplication of data-collection efforts. The current push for open-data policies of major funding agencies (e.g., EC, ERC, NERC, ANR, DFG, SNF) is therefore to be welcomed. Some examples for successful data portals related to cryospheric data include the Global Earth Observation System of Systems (GEOSS) – a data catalogue of a partnership of more than 100 states and the European Commission – the National Snow and Ice Data Center (NSIDC), the World Glacier Monitoring Service (WGMS), the International Network for Alpine Research Catchment Hydrology (GEWEX-INARCH) or the Swiss Open Support Platform for Environmental Research (OSPER). The recent establishment of a Global Cryospheric Watch program at the World Meteorological Organization (WMO) facilitates the exchange of cryospheric data. To be successful, however, the definition of common standards for different data types is required, as well as adequate mechanisms for rewarding and acknowledging groups and agencies investing in field-data collection.

An effort towards rigorous open-data policies for the mountain cryosphere seems particularly important, since mountain topography often introduces sharp administrative boundaries such as country borders. The latter often impose artificial limitations in spatial coverage. Large-scale studies on snow cover changes based on in-situ measurements, for instance, barely exist because such measurements are typically acquired from individual institutions or authorities. These cause the data to be subject to different jurisdictions, competences, practices or priorities, all of which hamper free data exchange. Rather than adhering to administrative borders, environmental data should sample regions defined on the base of geomorphological, topographical, and climatologic considerations. We call the European community working in the domain of the mountain cryosphere to be an example when it comes to open data policies.

4. Conclusions

This review has summarized prospects and challenges for future research in the European mountain cryosphere. It additionally provided an overview of past, current, and future changes in the fields of mountain snow, glaciers and permafrost in the European mountains. The paper has attempted to address the current state of knowledge in terms of observed evolution and associated impacts. Within a few decades, these impacts will have a bearing on ecosystems and the services provided by them, and on a number of economic sectors, including hydropower, agriculture, and tourism. The management of natural hazards will equally be affected, and a need for adaptation to climate change will arise.

A catalog of challenges for future cryospheric research has been identified, focusing on knowledge gaps, high-resolution modelling, understandable communication of uncertainty, cascading mass movements and process chains, quantification of the spatio-temporal distribution of precipitation, as well as the availability and access of observational data. All these issues

have effects on the capability of projecting future changes in the mountain cryosphere, and the impacts that these changes are likely to generate.

The principal challenges within the cryosphere domain includes both issues specific to a sub-discipline, as well as issues generally valid to several of them. The biggest overall challenge, however, is that the European mountain cryosphere is clearly expected to undergo drastic changes. European mountains will have a completely different appearance, most glaciers will have disappeared and even the largest valley glaciers will have experienced significant retreat and mass loss. Seasonal snow lines will be found at much higher altitudes, and the snow season will be much shorter than today. The changes in snow and icemelt will cause a shift in the timing of discharge maxima and a transition towards pluvial regimes. This will in turn have significant impacts on the seasonality of water availability, and consequences for water storage and management in reservoirs for drinking water and hydropower generation. Whereas an upward shift of the treeline and expansion of vegetation can be expected into current periglacial areas, the disappearance of permafrost at lower altitudes and its warming at higher elevations will likely result in mass movements and process chains beyond historical experience.

For future cryospheric research, the scientific community can build upon the awareness of these expected changes. Targeted strategies to precisely quantify the occurring changes, as well as approaches capable to address and mitigate these changes are now required. Efforts are not only required in the monitoring of the associated processes, but also in the numerical modelling of them. Current climate change scenarios predict the evolution of the components of the European cryosphere relatively reliably until 2050. In the second half of the century, however, the development is much more uncertain. This is linked, amongst other, to the uncertainties affecting future precipitation evolution. Shifts in air temperature and precipitation will clearly be significant drivers for future changes in the mountain cryosphere, but so will changes in shortwave and longwave radiation processes (related to greenhouse-gas forcing, changes in moisture and clouds, etc.) that are also key drivers for glacier melting and permafrost thawing.

Besides the emphasis on challenges in the modelling domain and cascading mass movement processes and process chains, this paper has highlighted data issues. Indeed, there are numerous limits to data availability, related to spatial and temporal sparseness, and restricted access. Financial and institutional barriers, as well as non-harmonized data policies add to the problem. Mountain cryosphere research urgently needs open access data of high quality for both understanding the functioning of the various elements and building reliable models for projecting their evolution.

Continuous improvement of physical models and free access to high spatial and temporal resolution simulations are essential to furthering our understanding of feedbacks between the atmosphere, the hydrosphere and the cryosphere, and to translate greenhouse-gas emissions scenarios into cryospheric impacts. The spatial resolution of global climate models is rapidly increasing, but much of the information still remains too coarse for mountain cryosphere research. Physically-based, nested

global-to-regional modelling techniques are expected to provide adequate data in the future but simulation results will remain highly dependent on the models' initial conditions. Providing an adequate observational data basis for that is therefore of paramount importance.

5 Finally, communicating research results on climate and cryospheric science is a challenge in itself. In general, the imminence of climatic change is more convincing to a lay audience when changes become visible. A prominent example is the retreat of mountain glaciers, which can convincingly be brought to the public through repeated photography. In this sense, climate-induced changes in the cryosphere enable a unique and effective form of communication. More effort should be dedicated to illustrate how these changes can affect water resources, mountain ecosystems, natural hazards, and a wide range of related
10 economic activities.

By highlighting the impacts of a changing cryosphere as climate evolves, this review has attempted to emphasize the central role of the cryosphere as a key element of environmental change in high mountains. To develop appropriate adaptation strategies to respond to the changes in climate, we anticipate an an increasing need for knowledge on the cryosphere.

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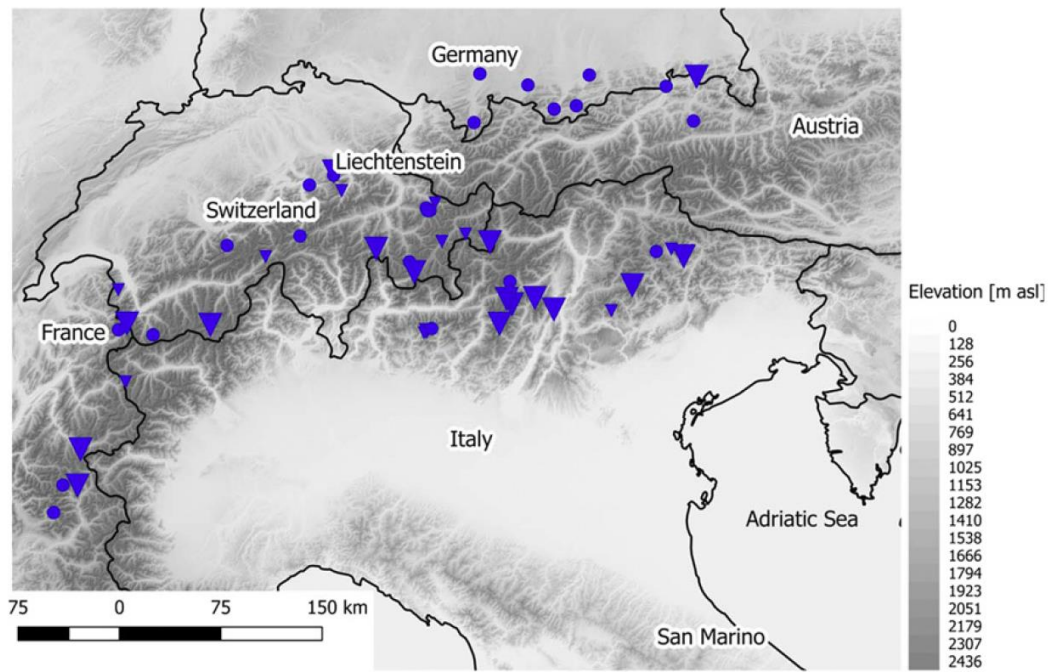
Figure 1: Geographical distribution of the 45 year trend (1968-2012) for April 1th SWE in the Alps. All stations show a negative trend. Large triangles indicate significant trends (p = 0.05) and small triangles indicate weakly significant trends (p = 0.2). Circles represent stations with no significant trend (p = 0.2). The elevation is given in gray. (Adapted from Marty et al. 2017b).
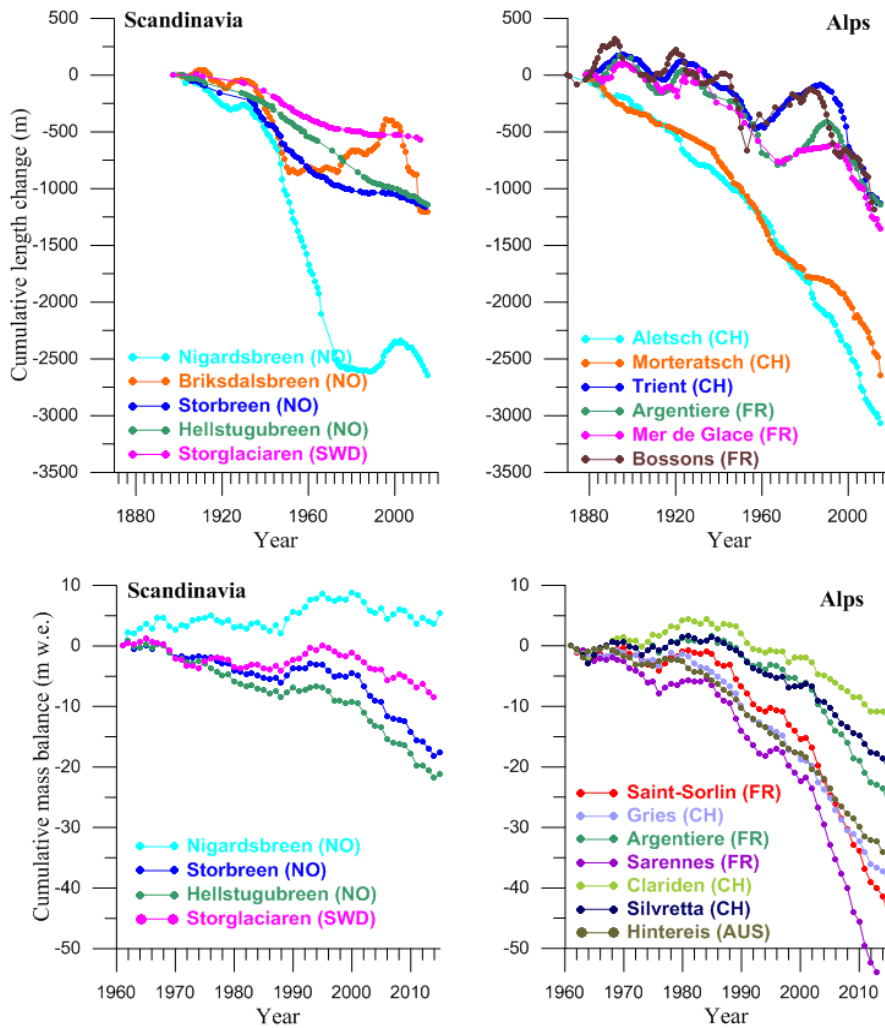
Figure 2: Length and surface mass balance changes documented with *in-situ* measurements for glaciers in Scandinavia and in the European Alps. Sources: WGMS (2015) and earlier issues with updates (Andreassen et al., 2016).
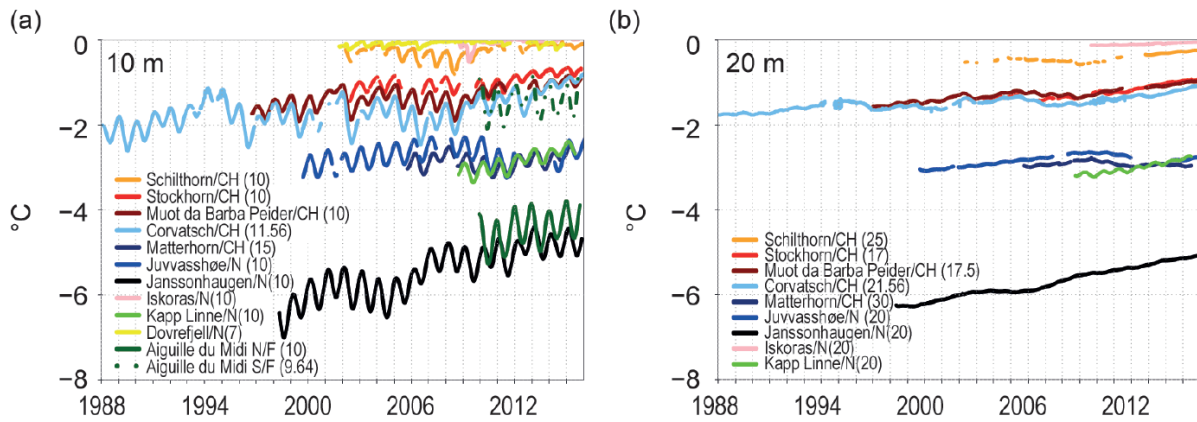
Figure 3: Temperature evolution of mountain permafrost in Norway (N), France (F), and Switzerland (CH) measured in boreholes at 10m (a) and 20m (b) depth (exact depth given in the parentheses). Adapted from Noetzli et al. (2016).
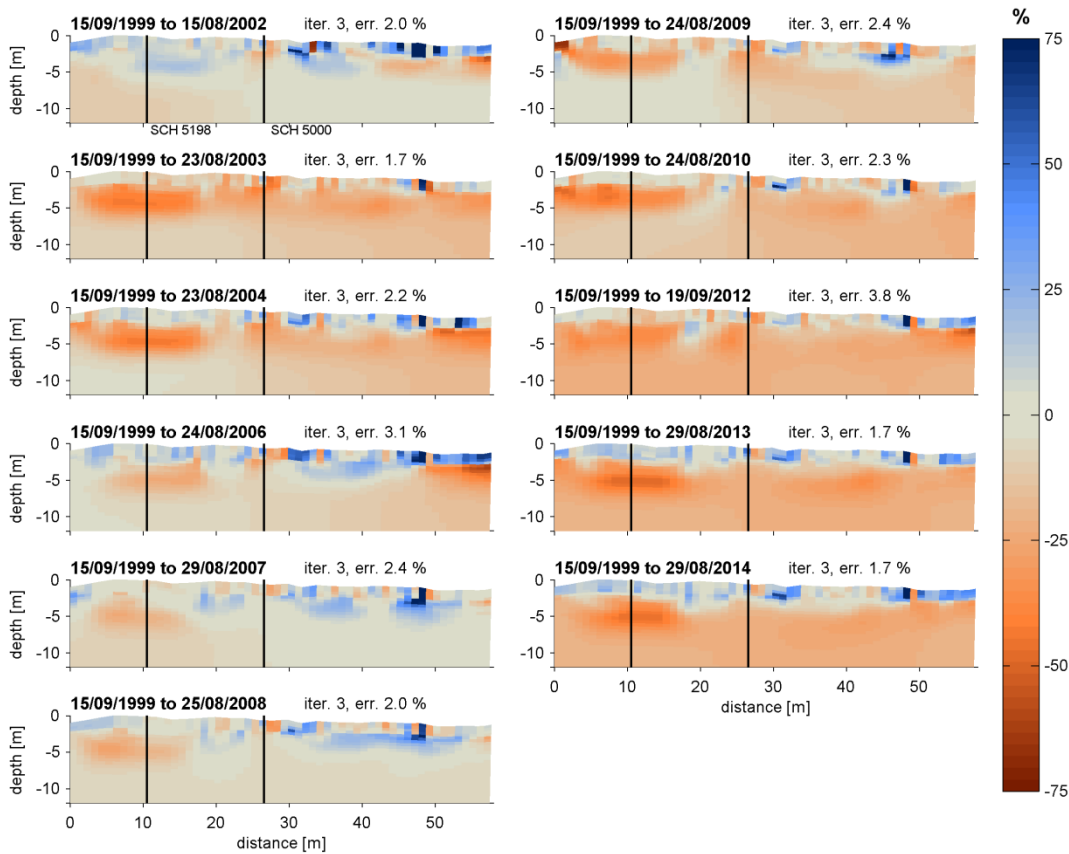
Figure 4: 15-year change in specific electrical resistivity (given as % specific resistivity change) along a 2-dimensional Electrical Resistivity Tomography (ERT) profile at Schilthorn, Swiss Alps (2900 m a.s.l.). Red colors denote a resistivity decrease corresponding to loss of ground ice with respect to the initial measurement in 1999 (see Hilbich et al. 2008a, 2011 for more details on ERT monitoring in permafrost). The black vertical lines denote borehole locations (modified after Permos, 2016).
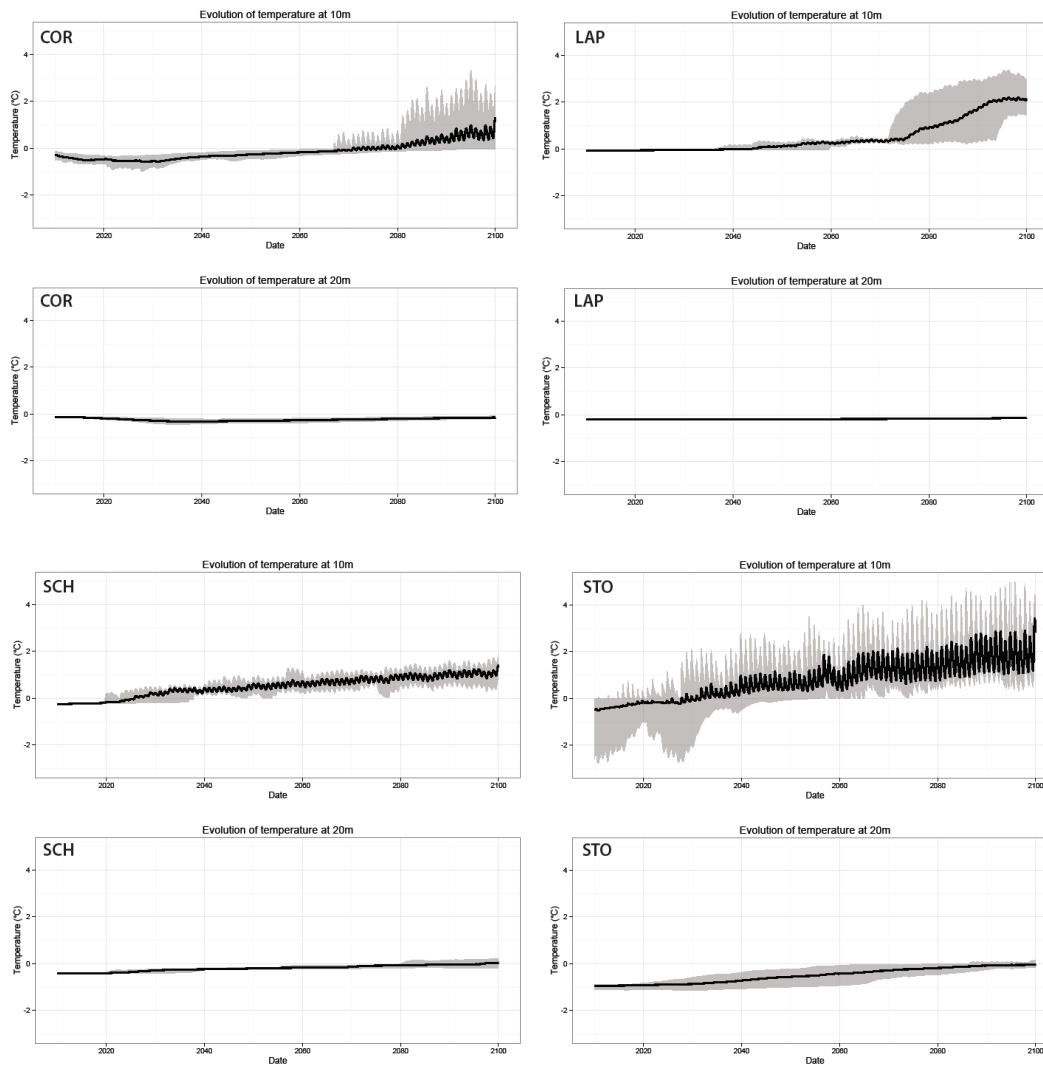
Figure 5: Modelled long-term evolution of ground temperatures at 10 m and 20 m at four different permafrost sites in the Swiss Alps (COR: Murtèl-Corvatsch, LAP: Lapires, SCH: Schilthorn, STO: Stockhorn), as simulated with the COUP model (Marmy et al., 2016). The black lines represent the median scenario and the grey zone the range of the 13 GCM/RCM chains which were used to drive the simulations. Modified after Marmy et al. (2016).
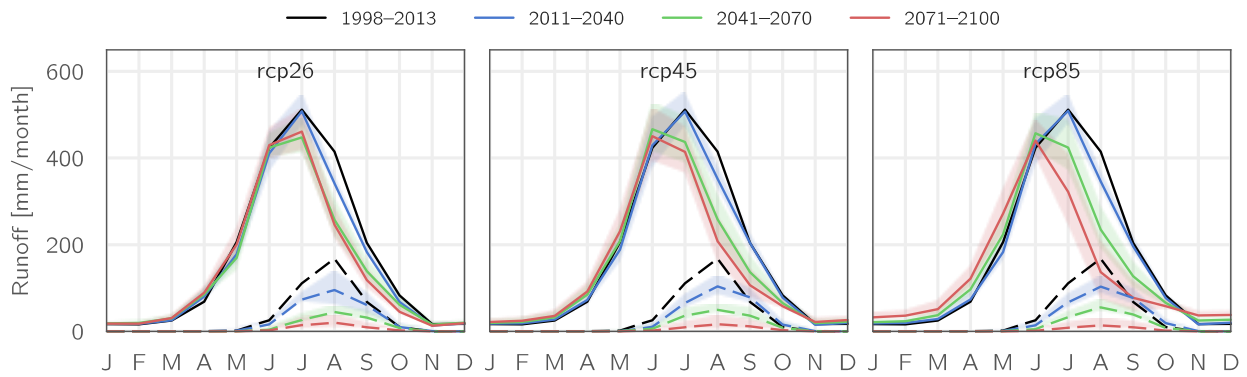
Figure 6: Shifts of streamflow regimes for the Rofenache catchment (Austrian Alps, 1891−3762 m a.s.l., 98 km², ~35 % glacierization as of 2006) as simulated with the AMUNDSEN model using downscaled EURO-CORDEX projections for the RCP2.6, RCP4.5 and RCP8.5 scenarios. Solid and dashed lines indicate the multi-model mean total and ice melt runoff, respectively, and shaded bands indicate the climate model uncertainty shown as ± 1 standard deviation. Adapted from Hanzer et al. (2017).
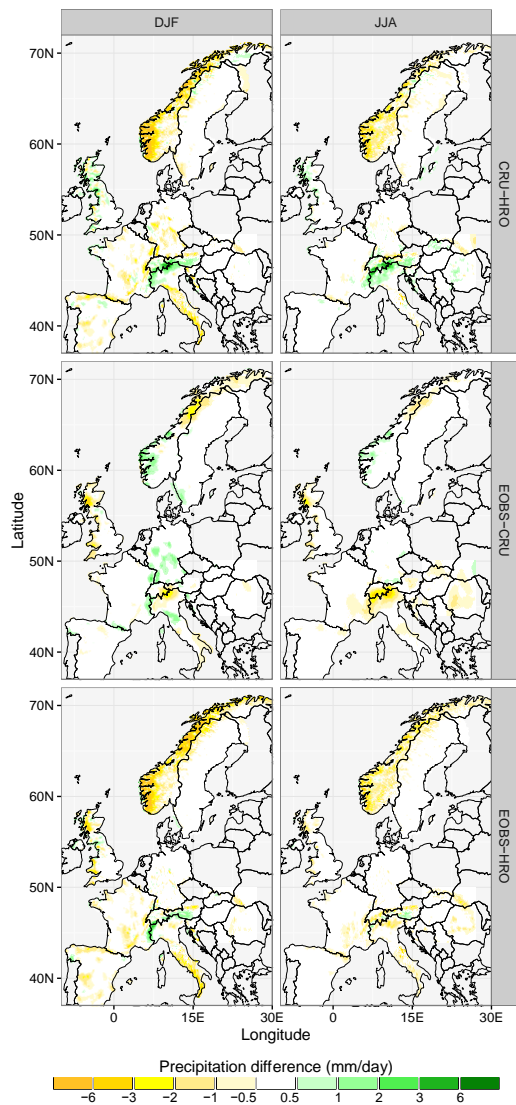
Figure 7: Seasonal average precipitation differences for December-January-February (DJF; left) and June-July-August (JJA; right) between CRU and HRO (first row), E-OBS and CRU (second row), and E-OBS and HRO (third row). Acronyms are as follows: CRU = Climatic Research Unit of the University of East Anglia; E-OBS = ENSEMBLES daily gridded observational dataset; HRO = High Resolution Observations (based on a network of local obervations).