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

Martin Beniston[1,2], Daniel Farinotti[3,4], Markus Stoffel[1,5], Liss M. Andreassen[6], Erika Coppola[7], Nicolas Eckert[8,] Adriano Fantini[7], Florie Giacona[9], Christian Hauck[10], Matthias Huss[10], Hendrik Huwald[11], Michael Lehning[11,12], Juan-Ignacio López-Moreno[13], Jan Magnusson[6], Christoph Marty[12], Enrique Moran-Tejéda[14], Samuel Morin[15], Mohammed Naaim[8], Antonello Provenzale[16], Antoine Rabatel[17], Delphine Six[17], Johann Stötter[18], Ulrich Strasser[18], Silvia Terzago[19], Christian Vincent[17]

[1]Institute for Environmental Sciences, The University of Geneva, Switzerland
[2]Department of Physics, The University of Geneva, Switzerland
[3]Laboratory of Hydraulics, Hydrology and Glaciology (VAW), ETH Zurich, Zurich, Switzerland
[4]Swiss Federal Institute for Forest, Snow and Landscape Research (WSL), Birmensdorf, Switzerland
[5]Department of Earth and Environmental Sciences, The University of Geneva, Switzerland
[6]Norwegian Water Resources and Energy Directorate, Oslo, Norway
[7]Abdus Salaam International Centre for Theoretical Physics, Trieste, Italy
[8]Inst. National de Rech. sur les Technologies pour l'Environnement et l'Agriculture (IRSTEA), Saint Martin d'Hères, France
[9]Université de Haute Alsace, Mulhouse, France
[10]University of Fribourg, Department of Geosciences, Fribourg, Switzerland
[11]Ecole Polytechnique Fédérale de Lausanne, Laboratory for Cryospheric Sciences, Lausanne, Switzerland
[12]Swiss Federal Institute for Avalanche Research (SLF), Davos, Switzerland
[13]Institute for Pyrenean Ecology (IPE-CSIC), Zaragoza, Spain
[14]University of the Balearic Islands, Palma-de-Mallorca, Spain
[15]Météo France, Centre d'Etudes de la Neige, Saint Martin d'Hères, France
[16]CNR, Institute of Geosciences and Earth Resources, Pisa, Italy
[17]University of Grenoble, Laboratory of Glaciology, Saint Martin d'Hères, France
[18]University of Innsbruck, Institute of Geography, Innsbruck, Austria
[19]CNR, Institute of Atmospheric Sciences and Climate (ISAC-CNR), Turin, Italy

Correspondence to: Martin Beniston (Martin.Beniston@unige.ch)

Abstract. The mountain cryosphere is recognized to have important impacts on a range of environmental processes. This paper reviews current knowledge on snow, glacier, and permafrost processes, as well as their past, current and future evolution in mountain regions in mainland Europe. We provide a comprehensive assessment of the current state of cryosphere research in Europe and point to the different domains requiring further research to improve our understanding of climate-cryosphere interactions, cryosphere controls on physical and biological mountain systems, as well as related impacts.

We highlight advances in the modeling of the cryosphere, and identify inherent uncertainties in our capability of projecting changes in the context of a warming global climate.

1 Introduction

Ongoing climate change and the importance of its anthropogenic component have gained wide recognition (IPCC, 2013). Some regions are likely to be more vulnerable to a changing climate than others to both 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 prominent and visible changes. Changes in mountain snow, glaciers, and permafrost, moreover, have resulted in significant downstream impacts in terms of the quantity, seasonality, and quality of water (Beniston et al., 2011). This is particularly true for areas where snow- and ice-melt represent a large fraction of streamflow. Countless studies have reported glacier retreat, permafrost warming, and snowfall decrease across mountain regions in Europe and elsewhere, with implications for streamflow regimes, water availability, or natural hazards. These can in turn negatively impact hydropower, agriculture, forestry, tourism or aquatic ecosystems. As a consequence, downstream communities will also be under pressure. Both political and scientific programs are calling for better preparedness, and for the development of strategies aimed at averting conflicts of interest that can arise, for example, 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, ice, and snow, and the observed changes in these elements of the cryosphere. We focus on mainland Europe, in particular the European Alps and Scandinavia, but also address the Pyrenees and other mid-latitude European mountains undergoing large and rapid change. An assessment of the challenges that need to be 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 changes and impacts of a cryosphere responding to rapidly changing climatic conditions, and for appropriate adaptation measures to be developed. The discussions that will follow reflect the current opinions of a body of scientists that focused on a number of issues at a conference held in Switzerland in 2016. We do not claim that all aspects of cryosphere sciences are exhaustively covered, nor are all the possible elements of the cryosphere discussed (e.g., lake ice; ice in caves, etc.). However, we do believe that the elements that appear in the following text do represent much of current scientific preoccupations on this major component of mountain environments.

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

2.1 Changes in snow

Snow is an important component of hydrology, but affects also largely other cryospheric components, i.e., glaciers (through the 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 European mountains and also, quite often, in downstream lowland regions. Moreover, snow is the most important interface between the atmosphere and the ground, also influencing the other components of the cryosphere. The extreme spatiotemporal variability remains one of the key factors of uncertainty concerning the impacts of climate change on the cryosphere. Ongoing climatic change is significantly affecting the snow cover through different processes. Snow observations are thus important for understanding these processes, and for providing more reliable assessments of future changes.

2.1.1 Observed changes in snow cover

Most studies show negative trends in snow depth and snow duration over the past decades. The 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). As Fig. 1 demonstrates, the spring snow water equivalent (SWE) is clearly decreasing in the Alps (Bocchiola and Diolaiuti, 2010; Marty et al., 2017) as well as at low elevations in Norway (Skaugen et al., 2012). Only in the higher and colder regions of the Fenno-Scandinavian mountains do snow depth and SWE exhibit positive trends. However, in more recent decades trends have become mostly 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 (López-Moreno et al., 2007). In other European mountains, observations are less abundant, but declines in snowpack for mountains in Romania (Birsan and Dumitrescu, 2014; Micu, 2009), Bulgaria (Brown and Petkova, 2007), Poland (Falarz, 2008), and Croatia (Gajić-Čapka, 2011) are reported.

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 conditions (Klein et al., 2016), resulting from both higher winter and spring temperatures. 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 snow pack variability seems to be related to the establishment of blocking patterns over Europe, although in this case the correlation between the annual snow pack variability and the NAO is small and limited to low elevations (Scherrer and Appenzeller, 2006; Durand et al., 2009). The NAO influence can be detected at higher elevations through a ‘cascade’ of processes that include the NAO influence on pressure fields, and the influence of pressure fields on precipitation. Together with temperature, this determines snowfall. In recent decades, this cascade has led to an increased number of winter days with warm and dry conditions, which is unfavorable for snow accumulation (Beniston et al., 2011b). Moreover, the Atlantic Multi-decadal Oscillation (AMO), which is a natural periodic fluctuation of Northern Atlantic sea surface temperature, affects the low-frequency variability of Alpine spring snowfall (Zampieri et al. 2013) and therefore also contributes to the described decline in snow duration.

The observed changes in snow amounts are often abrupt in time, as a result of the interplay between cold temperatures and precipitation, both influenced by large-scale weather patterns. Several studies have reported a step-like change occurring in the late 1980s for snow depth (Marty, 2008; Durand et al., 2009; Valt and Cianfarra, 2010) and for snow-covered areas of the Northern Hemisphere (Choi et al., 2010), but also for other biophysical systems (Reid et al., 2016). This development is mostly the result of winter temperatures that have not risen further, neither in large areas in the Northern Hemisphere (Mori et. al. 2014) nor in the Swiss Alps (Scherrer et al., 2013). Atmospheric internal variability (Li et al., 2015), as well as shrinking sea-ice extends (Mori et. al. 2014) have been invoked as possible explanations. As a result, monthly mean (Dec-Apr) snow covered area in the Alps has not decreased significantly since the late 1980's (Hüsler et al., 2014).

Studies analyzing high-magnitude snowfall and maximal winter snow depths 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 decreasing pattern for extreme snow fall rates is less clear than for extreme snow depth, except for low elevations where the influence of increasing temperatures is predominant (Marty and Blanchet, 2012). In addition, there exist few studies related to past changes in snow avalanche activity. Over the last decades, observations indicate that (a) the number of days with prerequisites for avalanche 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 cover characteristics (Castebrunet et al., 2012).

2.1.2 Future changes in snow cover

The projected increase in temperature for coming decades is accompanied by high uncertainties in winter precipitation changes. Ensemble means show no clear precipitation change until about the 2050s, but slightly increasing winter precipitation thereafter. Projected changes in snow cover are thus highly dependent on the considered greenhouse gas emission scenario and the addressed period. Under a SRES A2 scenario, regional climate model simulations show a dramatic decrease in both the snow cover duration and SWE for Europe by the end of the 21st century (Jylhä et al., 2008).

For the Alps and at an elevation of 1500 m a.s.l., recent simulations project a reduction in SWE of 80-90% by the end of the century (Rousselot et al., 2012; Steger et al., 2013; Schmucki et al., 2015). The snow season at that altitude is projected to start 2-4 weeks later and to end 5-10 weeks earlier than today (1992-2012), which is roughly equivalent to an elevation shift of about 700 m (Marty et al., 2016). For elevations above 3000 m a.s.l., even the largest projected precipitation increase results in a decline in SWE of at least 10% by the end of the century. Future climate will most probably not see a permanent snow cover during summer even at the highest elevations, with obvious implications for glacier evolution (Magnusson et al., 2010; Bavay et al., 2013) and thermal conditions of the ground (e.g. Magnin et al. 2017, Draebing et al. 2017, Marmy et al. 2016 (Draebing et al., 2017).

Projections for Scandinavia show clear decreases for snow amount and duration for all latitudes. An exception is given by the highest mountains in the north, where strongly increasing precipitation seems to partly compensate for temperature rise, thus resulting in marginal changes only (Räisänen and Eklund, 2012). Simulations for the Pyrenees indicate declines of the snow cover similar to those found for the Alps (López-Moreno et al., 2009). Again, the dependency on future greenhouse gas emissions has to be noted: Under a high emission scenario (RCP8.5), SWE decreases by 78% at the end of the 21st century at 1500 m a.s.l. elevation, whereas a lower emission scenario (RCP6.0) shows a decline of 44%.

Regarding extremes, model results suggest a smaller reduction in daily maximum snowfalls than in mean snowfalls over many regions of the Northern Hemisphere by the end of the 21st century (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. The ongoing evolution towards more wet than dry snow avalanches will 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 in winter at high elevations due to more favorable conditions for wet-snow avalanches earlier in the season (Castebrunet et al., 2014). The reduction of the snow season length will have large consequences for winter tourism (Uhlmann et al., 2008; Steiger and Abegg, 2013; Schmuki 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). Decreasing snow duration and shrinking snow depths are not the only changes with implications: especially in Fenno-Scandinavian mountains, the increasing number of hard (icy) snow layers due higher temperatures (Johansson et al., 2011) can have a significant effect on the life of plants and animals.

a. 2.2 Changes in glaciers

Mountain glaciers are recognized as a key indicator of rapid and global climate change. They are important for water resources 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., 2016), with the latter also offering opportunities in terms of water storage.

Observed changes in glaciers

Glaciers in mainland Europe cover an area of nearly 5,000 km^2 (Table 1) and have an estimated volume of almost 400 km^3 (Huss and Farinotti, 2012; Andreassen et al., 2015). Based on historical archives such as paintings and photography (Zumbühel et al., 2008), it is seen that glaciers have undergone substantial mass loss since the 19th century (Fig. 2) and 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). Repeated inventories have shown a reduction in glacier area of 11% in Norway between 1960 and the 2000s (Winsvold et al., 2014), and 28% in Switzerland between 1973 and 2010 (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). 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).

Country	**Area (km^2)**	**Volume (km^3)**	**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**			

Table 1: *Distribution of glacier surface area and estimated ice volume in continental Europe and mainland Scandinavia. Years of reference and respective publications are given for glacier area. Ice volume estimates refer to 2003 (Huss and Farinotti, 2012) for continental Europe and Sweden, 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 been attributed mainly to changes in atmospheric energy fluxes, which in turn translate to air temperature in a non-linear manner. However, good correlations 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). It has also been shown that changes in solar radiation could partly explain high melt rates in the 1940s (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). However, the relevance of these approaches over multi-decadal time periods has been little assessed. This is due to both the lack of long-term in-situ meteorological measurements close to the study sites and the temporal evolution of melt sensitivity to temperature (Huss et al., 2009; Gabbi et al., 2014; Réveillet et al., 2017). Changes in climate variables affecting glacier behavior are driven by different factors, ranging from large-scale synoptic weather patterns to regional and local effects enhanced by topography. The latter influences the distribution of precipitation, solar radiation, and wind, among others. Several studies have shown that the NAO influences glacier surface mass balance, in particular in southwestern Norway. The glacier advances in Scandinavia during 1989-1995 are attributed to increased winter precipitation linked to the positive NAO phase during this period (Rasmussen and Conway, 2005). In contrast, in the European Alps, the relationship between the NAO and glacier surface mass balance is less pronounced (Marzeion and Nesje, 2012; Thibert et al., 2013).

The glacier evolution during the 20th century also highlights the importance of the surface albedo feedback, as albedo governs the shortwave radiation budget at the glacier surface, which is the dominant energy source for melting. Snow and ice ablation sensitivities to albedo have been generally assessed using physical energy-balance considerations (Six and Vincent, 2014) or degree-day approaches (e.g., Engelhardt et al., 2015). Oerlemans et al. (2009) and Gabbi et al. (2015) investigated the influence of accumulation of dust or black carbon on melt rates on Swiss glaciers in the last decades, which revealed increasing annual melt rate by 15-19% compared to pure snow. Monitoring, reconstructing, or modeling the surface albedo of glaciers is challenging (Brock et al., 2000) as its spatial and temporal evolution is linked to changes in surface properties (e.g., snow grain size) 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).

Due to a large ice mass loss during the 20th century, alpine glacier dynamics have been strongly affected, leading to a substantial decrease in ice flow velocities (e.g., Berthier and Vincent, 2012). However, 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. Increased water pressures, on the other hand, reduce the frictional drag and thus increase 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). Although changes in seasonal ice flow velocities are driven by subglacial hydrology it seems that, at the annual to multiannual time scales, the ice flow velocity changes do not depend on changes in subglacial runoff (Vincent and Moreau, 2016). A few temperate alpine glaciers have shown large accelerations due to a change in subglacial hydrology, e.g., the Belvedere glacier in Italy (Haeberli et al., 2002), even if 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).

Several studies of cold glaciers in the European Alps 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; 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 complete break-off of Altelsgletscher in 1895, for example (Faillettaz et al., 2015).

The future of European glaciers

Estimates of future glacier changes in the European Alps and Scandinavia have been motivated by questions regarding the climate change sensitivity of glaciers but also by applied studies related to the consequences of declining ice volume for future water resources. Over the last two decades, various studies on future glacier change in Europe have been published that can be broadly classified into site-specific (e.g., Giesen and Oerlemans, 2010) and regional studies (e.g., Salzmann et al., 2012).

Projections of future glacier change necessarily involve a model for the surface mass balance response of glaciers, and a model for the response of ice flow dynamics. The applied models vary in complexity and range from simple degree-day models (e.g., Braithwaite and Zhang, 1999; Radic and Hock; 2006) to complete surface energy balance models (e.g., Gerbaux et al., 2005), and 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 disappear completely (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 by 2100 (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 famous outlet glacier of Jostedalsbreen, the largest ice cap in mainland Europe (Laumann and Nesje, 2009). At the scale of mountain ranges, model studies relying on different approaches and medium-range emission scenarios consistently predict, over the 21st century, 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).

Even with strong efforts to reduce CO_2-emissions and to stabilize global warming at around +2°C as recommended by the Paris COP-21 climate accord, ice volume losses of around 80% are expected for the European Alps (Huss, 2012; Salzmann et al., 2012). Mountains glaciers in Europe are already strongly out-of-balance with present climatic conditions, and substantial mass losses are already committed (e.g., Andreassen et al, 2012a; Mernild et al, 2013). Due to their high sensitivity to climate change and limited altitudinal extent, many European glaciers are unable to reach a new equilibrium with climate even if air temperatures are stabilized by the end of this century. Furthermore, the present-day ice caps in Norway, that contribute to a large part of the total ice volume in Europe (Table 1), 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 3.2).

2.3 Changes in permafrost

Permafrost is defined as lithospheric material with temperatures continuously below 0°C, and covers approximately 20 million km^2 of the Earth, 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 ice stored in European permafrost, 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, 2013).

2.3.1 Observed changes in permafrost

Permafrost borehole temperatures are monitored in many European mountain ranges (GTN-P database, Biskaborn et al., 2015), several of the sites being accompanied by meteorological stations and ground surface temperature measurements (Gisnaas et al., 2014; Staub et al., 2016). However, as 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 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. 2007), pointing at decadal warming rates at the permafrost table in the order of 0.04°–0.07°C for Northern Scandinavia and Svalbard. 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 (active layer thickness; 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 warming trend for "warm" permafrost sites (temperatures close to the freezing point), which is otherwise clearly visible in cold permafrost (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 rockfalls and debris flows (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; see also Section 3f). The disposition conditions and triggering mechanisms of slope instabilities can be diverse, and depend on subsurface material (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., 2016). Water infiltration into newly thawed parts of permafrost is often mentioned as a possible triggering mechanism (Hasler et al., 2012; Schneuwly-Bollschweiler and Stoffel, 2012), but only few observational data are available. On the other hand, permafrost affects hydrology through interactions with the snow cover, and by influencing infiltration, subsurface runoff and drainage (Arenson et al., 2010; Scherler et al., 2010; Langston et al., 2011; Zhou et al., 2015). Permafrost bodies and especially rock glaciers are also water reservoirs affecting runoff regimes and water availability. Data on the ice volume stored in such features remain, however, scarce. To date, hydrologically oriented permafrost studies have been based on remote sensing and meteorological data for larger areas, or have had a locally 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 European mountain ranges to date.

2.3.3 Changes in rock glacier flow velocities and ice volume

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., 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 show an acceleration in 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 cannot, however, 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 this Electrical Resistivity Tomography (ERT) monitoring 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 (Figure 4a) is at the freezing point, and shows no clear trend. ERT is increasingly used in 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 permafrost evolution

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 modeling (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 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). However, in spite of similar trends in the above-mentioned RCM-driven permafrost studies, comprehensive regional scale maps or trend values for current and projected permafrost changes in Europe are not available to date - partly, because there is an insufficient number of borehole data available, but mainly because of the large heterogeneity of the permafrost in European mountain ranges, which depends strongly 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.) in addition to classical topo-climatic factors (elevation, aspect, slope angle).

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 the next 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 systems. Because of the seasonal release of previously stored water 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 often 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 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 (Schaefli et al., 2007; López-Moreno et al., 2014). Temperature-induced changes in streamflow (rain-to-snow fraction, seasonal shift of snowmelt, glacier runoff contribution) are generally better understood than the ones caused by changing precipitation (Blaschke et al., 2011). Nevertheless, understanding long-term trends in runoff require 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 temperature and precipitation is influenced by the complex interactions between climatic conditions, topography and wind redistribution of 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 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.

Scenarios of changing streamflows affected by retreating glaciers in a warming climate have recently been presented in various physically-based, distributed modelling experiments (Weber and Prasch, 2009; Prasch et al., 2011; Hanzer et al., 2017). Figure 6 illustrates future streamflows of a currently highly glacierized catchment (roughly 35% glacierization) in the Austrian Alps. Even for the moderate IPCC RCP2.6 scenario (IPCC,) which corresponds roughly to the COP-21 “2°C Policy”, the glacier melt 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 runoffs. 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 streamflows 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 al., 2011; Hanzer et al., 2017). For many streams utilized for hydropower generation, this phenomenon can be superimposed by the effects of regulation. The regimes in Fig. 6 indicate that (a) the effect of warming increases after the mid of the century for all scenarios, (b) this 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., the effect of declining glacier area has already become larger than the increasing melt caused by the rising temperatures. This does not necessarily hold true for other headwater catchments (Hanzer et al. 2017). Until the mid-21st century and for large scales, such as the entire Austrian region, the decrease of annual streamflow is expected to be small.

By 2100, the glaciers in the Alps may lose up to 90% of their current volume (Beniston, 2012; Pellicciotti et al., 2014; Hanzer et al., 2017), and peak discharge is likely to occur 1-2 months earlier in the year (Horton et al., 2006) according to carbon-emission scenarios. In Switzerland, a new type of flow regime (called “*pluvial de transition”* i.e., 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 available since 1808 for the Rhine river. 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 ice-melt contribution to discharge in a changing climate, combined with the other processes determining streamflow regime, and their scale dependencies are not yet fully understood. There exists a general consensus, however, that only few high-altitude regions of the Alps will continue to have a glacial regime in the long term (FOEN, 2012).

Despite the general trend towards drier summers, a recent review of 21st century climate change in the European Alps found that severe flooding might become more frequent due to heavy or extended precipitation events in future (Gobiet et al., 2014; Stoffel et al., 2016). In addition, 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 atmospheric temperatures continue to rise (Würzer et al., 2016). However, a cutoff beyond which ROS events will decrease with increasing temperatures is expected when the amount of snow becomes significantly reduced in response to higher temperatures (Beniston and Stoffel, 2016). Concerning droughts in the Alpine region, a clear trend towards increasing occurrence and severity has been highlighted (Gobiet et al., 2014).

Finally, climate change also effects the seasonal duration of lake ice and spring break-up dates (George, 2010), with an overall trend to earlier thawing of one week per century, approximately (Livingstone, 1997). The timing of the break-up is strongly correlated with local and regional surface temperatures, determined to a large extent by synoptic-scale meteorological processes. In many regions such as, e.g., St. Moritz/Engadin (Switzerland), the lake ice coverage during the winter months is an important landscape feature for tourism.

2.5 Impacts on downstream water management

Sectors that are directly dependent on alpine headwaters will need to adapt to the changes outlined above. Different scenarios therefore need to be considered, depending on how governance will cope with water-related conflicts that may arise from changes in water demand (Nelson et al., 2007; Beniston et al., 2011).

2.5.1 Agriculture

Shifts in agricultural production are expected with climate change (Jaggard et al., 2010; Gornall et al., 2010). Most studies conducted in the alpine regions project reduced soil water content as a result of increasing evaporation. 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 droughts will affect both croplands and grasslands. In 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). 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 power markets, as both electricity production and demand are linked to meteorological variables (Apadula et al., 2012). As a consequence of earlier snowmelt and reduced water discharge from glaciers, hydropower production is expected to increase in winter 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 peak hydropower production has so far not been affected by climate change. This is possibly the result of the large existing reservoir volumes, which are able 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). Reservoir management, however, will become more challenging as a consequence of higher fluctuations in electricity demands linked to the intermittent production of new renewable energy sources (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 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 Tourism

Increasing temperatures are anticipated 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 considered as an adaptation measure 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 (Putz 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 improved water management (Roson and Sartori, 2012).

2.6 Ecosystem functioning

In high mountains, the interaction between the biotic and abiotic ecosystems components is especially strong. The demanding conditions of high elevations control the physical environment of living beings, which in turn can modify the environment to make it more suitable for their own survival. Increasing temperatures, for example, induce an upward and poleward shift of flora and fauna (Parmesan, 2006). Such modifications can affect mountain biodiversity, especially for endemic species, and species with limited dispersal capacity (Viterbi et al., 2013). Among the many controlling factors, the state of the cryosphere is a crucial driver of ecosystem functioning (see Callaghan and Johansson, 2015 for a recent review). Retreating glaciers open new bare areas for colonization, and changes in snow cover affect ecosystem dynamics in multiple ways. Earlier snowmelt is associated with an anticipation of the blooming season of alpine plants (Pettorelli et al., 2007) which could induce a mismatch between producers (plants) and consumers (herbivores), similar to what is observed in Arctic regions (Post et al. 2009).

The population of the Alpine ibex, for instance, has been monitored annually in the Gran Paradiso National Park (Italy) since 1956, and was contrasted against average winter snow depth (Jacobson et al., 2004; Fig 7). In particular, the adult ibex population density is limited by the winter snow cover in a much stronger way when the population size is large (Fig. 7d). By contrast, the dramatic decline of the ibex population after 1995 was linked to a drastic reduction in the survival of newborns in the first winter (Mignatti et al 2012). It has been suggested that the reduced snow cover and the earlier snowmelt would have led to an earlier blooming of alpine grasses, causing them to be drier and less energetic in late July, which is the period when newborns are fed with milk by their ibex mothers. Snow thus seems 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.

Abundance of Alpine rock ptarmigan has been observed to decline in abundance as well based on annual census data from northwestern Italy (Fig. 8). Population dynamics are above all controlled by the onset of spring snowmelt, and the timing of autumn snow cover (Imperio et al., 2013). Ecosystem models driven by outputs of RCMs are able to reproduce the observed changes, and project a further population decline, but results can differ considerably depending on how effects of population density are accounted for (Figs. 8d-e).

3 Challenges and issues for cryosphere research in European mountains

3.1 Observational data: access, availability, quality, spatial and temporal distribution

Quality-controlled data with high spatial and temporal resolution are essential for both the detection of past changes in components of the cryosphere, and the development and validation of numerical models that project future evolution (Lehning et al., 2016; Beniston et al., 2012; Quevauviller et al., 2012). The reliability of data used in climate-related research has been questioned in the past, particularly with respect to whether the accuracy and precision of environmental data – including temperature and precipitation, for example – are sufficient for distinguishing long-term trends from inter-annual variability. In addition to intrinsic accuracy-limitations of measuring equipment, changes of sensors, sensor location or surrounding environment can make the interpretation of non-homogenized time series very challenging (Venema et al., 2012). The methodologies applied for data collection and homogenization often differ as a result of different legislations, competences, practices or priorities - a problem particularly prominent in Europe, were mountain areas (Alps, Pyrenees, Scandinavia) are under different national and regional authorities. Data quality must thus be assured by rigorous and standardized control. International coordination and standards must be established, and compliance has to be guaranteed.

Existing measurement sites and instruments are not homogeneously distributed. Environmental observations are typically biased towards lowland and mid-elevations, mostly because of the logistical challenges in maintaining high-elevation monitoring sites. There is a clear lack of, and demand for, adequate environmental information from high elevations. Such information is essential for cryosphere related research, and pivotal for quantifying elevation-dependent warming, precipitation, snowmelt, or river runoff, amongst others. Substantial efforts and new ideas are required to improve the spatial coverage and representativeness of the variables of interest (Orlowsky and Seneviratne, 2014). Europe could take a leading role here, since despite complicated logistics, most measurement sites of interest in mountainous terrain are still reasonably accessible.

Data availability and spatial coverage is often confined by country borders or by limited competence and responsibility of the institutions collecting the data. This is the reason why, for example, studies on snow changes based on in-situ measurements and covering the entire Alps barely exist. “Administrative-borders effects” due to the relatively small size of countries comprising mountains often also influence spatially interpolated data, introducing artifacts from artificial domain limits. In the worst case, such artifacts can flaw the findings of entire studies. Rather than adhering to administrative borders, environmental data should sample regions defined on the base of geomorphological, topographical, and climatologic considerations. This is one of the (political, non-scientific) challenges that could – if resolved – significantly improve the availability and homogeneity of cryosphere related observational data.

To date, many data have restricted ownership. The consequences of such non-open-access policies include lack of data, impossibility of accessing existing data, delays in obtaining them, non-availability of real-time data, and duplication of data-collection efforts. A more liberal and open-data policy would contribute to solving part of the problem. The recent push for open-access data policy of major funding agencies (e.g., EC, ERC, NERC, ANR, DFG, SNF) has therefore to be welcomed. To be successful, however, the definition of common standards for different types of environmental data is required. As data acquisition can be related to important investments in terms of both equipment and labor costs, moreover, it is important that adequate mechanisms for rewarding groups and agencies investing in field-data collection are established. This is a particular challenge for the cryosphere since data acquisition involves substantial logistics and maintenance of sensing instrumentation in most cases.

Open-access platforms for cryospheric data are currently underdeveloped, and existing efforts are often uncoordinated. International cooperation, as well as well-defined management, sharing and archiving policies are a key to success. Platforms should be widely known, easily searchable and citable, contain quality controlled data, and provide standardized metadata to make datasets understandable for end-users. An example for such a platform is the Global Earth Observation System of Systems (GEOSS), which is a data catalogue of a partnership of more than 100 states and the European Commission. The National Snow and Ice Data Center (NSIDC) and the World Glacier Monitoring Service (WGMS) are two examples for institutions providing successful data portals in the cryospheric domain. At the national level, a noteworthy initiative is the Swiss Open Support Platform for Environmental Research (OSPER), which has set a benchmark in data provision and metadata integration for a large number of environmental datasets. A particular challenge in the centralization and archiving of environmental data is the large data volume generated by modern remote sensing techniques.

Recently, both in-situ sensing technology and (terrestrial and air/space born) remote sensing technology are rapidly developing, opening new views and understanding of the cryosphere. A particular challenge here is the combination and two-way validation of ground truth and remotely sensed data. There is also a spatial gap to close, i.e., upscale point measurements to larger areas, and vice versa, focus and validate large-scale data products with reliable, high-quality ground truth observations. This task may be accomplished in European mountain regions where observational station networks are generally relatively dense.

National meteorological services have for decades been exchanging observations for real-time forecasts and climate monitoring. Such data exchanges have been possible through the establishment of data reporting standards and data models, as well as robust exchange mechanisms, both under the umbrella of a global authoritative organization, i.e., the World

Meteorological Organization (WMO). The recent establishment of a Global Cryospheric Watch program at WMO makes it possible that cryospheric data exchange benefits from the experience acquired from the exchange of meteorological information. This is anticipated to significantly facilitate mountain cryospheric studies across borders.

3.2 Modeling of the cryosphere: spatial resolution and physical processes in complex terrain

Snow as a mostly non-permanent interface between the earth's surface and the atmosphere features a variety of small- and large-scale physical processes, which significantly influence the mass and energy exchange at the surface. Small-scale processes include water transport in snow and firn (Würzer et al., 2016a, Wever et al., 2014; Wever et al., 2016), phase changes (i.e. melt, refreeze, sublimation and condensation), drifting and blowing snow, as well as metamorphism (e.g., Aoki et al., 2011; Pinzer et al. 2012). While the mechanistic point-scale understanding of these processes is rapidly increasing (Wever et al., 2014), the challenge arises from quantifying their effect at larger scales, i.e., upscale from the point-scale to the catchment or even continental scale. To date, such up-scaling techniques are missing to a large extent. Examples of the link between large- and small-scale effects are the often significantly altered snow distribution after a storm (Lehning et al., 2008; Schirmer et al., 2009) or the change in snow albedo after a melt event. These effects (and many others) are insufficiently represented in large-scale weather and/or climate models, yet most likely would lead to significant changes in model predictions. The spatial distribution of small-scale snow properties is also essential for the correct interpretation of satellite remote sensing signals. For example, ice-lenses or liquid water in snow heavily influence the microwave backscatter (Marshall et al., 2007), which is the basis for many satellite remote-sensing products. The problem of properly up-scaling snow properties is not peculiar to the complex terrain of mountains, but exists also for less complex topography such as Antarctic sea ice (Trujillo et al., 2016), the Canadian prairies, the Siberian tundra, or the Greenland ice sheet, where the melt water amount that can be stored in snow and firn (Forster et al., 2014) and above frozen ground (cf. Section 2.3) is unknown.

Three relevant scales can be distinguished for the modeling of mountain snow:

(1) The point scale is typically chosen for applications such as snow stability estimates in the context of avalanche warning (Durand et al., 1999; Gaume et al., 2014) or detailed studies in snow hydrology such as the analysis of rain on snow events (Beniston and Stoffel, 2016; Würzer et al., 2016b), the analysis of the impact of infiltrating melt water on the thermal regime of permafrost (Scherler et al. 2010) and the insulation effects of snow on permafrost slopes and rock walls (Lütschg et al. 2008, Marmy et al. 2016, Gisnås et al. 2014, Haberkorn et al. 2017). For snow modeling, there are two widely used physically based models, namely the French model CROCUS (Vionnet et al., 2012) and the Swiss model SNOWPACK (Lehning et al., 1999). Nevertheless, many processes – including metamorphism and mechanical properties for example – still have a high degree of empirical parameterization (Lehning et al., 2002). The main challenge for snow model development at this smallest scale is the formulation of a consistent theory for snow microstructure (Krol and Lowe, 2016) and its metamorphism.

(2) The catchment scale is mostly employed in hydrological applications (Kumar et al., 2013) and snow models of varying complexity are used for that (Essery, 2015; Magnusson et al., 2015). A principal challenge at this scale is (a) to distinguish between uncertainties introduced by the model structure and uncertainties related to the input data (Schlögl et al., 2016), and (b) to develop site-independent models that can be used without 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). A difficulty is bridging the gap between the scales, i.e., upscale scattered point measurements to the catchment scale. Recent sensing technology (e.g., airborne and terrestrial laser scanning, ALS/TLS) can provide spatially resolved data of snow covered area and snow. Other variables such as spatially distributed snow density and liquid water content, as well as local wind data are instead difficult to obtain. These quantities can be modeled, but corresponding observations should be available for validation. Furthermore, distributed data of snow albedo would be desirable, as the limitated spatial resolution of satellite data do not allow for high resolution data suitable for model validation to be obtained. A promising emerging approach here is the use of UAVs equipped with radiometers and multispectral cameras.

(3) The large scale is most relevant in weather forecast and climate models. The snow cover affects the surface energy balance through the high snow albedo and its low thermal conductivity at all scales, whilst at the large scale feedbacks to regional weather, climate, and the water budget are expected (Groisman et al., 1994; Viterbo and Betts, 1999; Immerzeel et al., 2010). Large-scale models use relatively simple, parametric snow schemes, as these require only a few input variables and are computationally inexpensive (Bokhorst et al., 2016). Current numerical weather prediction systems generally use single-layer snow schemes (IFS documentation, 2016; GFS documentation, 2016) which are known to oversimplify reality. 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 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 within the snowpack (Armstrong and Brun, 2008; Steger et al., 2013) and the evolution of the microstructure due to snow metamorphism. More complex snow schemes, with multiple snow layers and snow-water retention processes have been successfully integrated, for example, in the EC-Earth global climate model (Dutra et al., 2012; Hazeleger et al., 2012). Future research will need to clarify the degree of complexity required in snow schemes when they are integrated in large-scale climate models (van der Hurk et al., 2016). In large-scale modeling frameworks, reliable snowpack simulations are currently limited by the coarse representation of topography. This implies inaccurate representation of altitudinal temperature gradients, and a crude separation of the precipitation phase, since convective processes are inadequately represented or oversimplified (Wilcox and Donner 2007, Chen and Knutson, 2008; Wehner et al., 2010, Sillmann et al., 2013). The lack of reliable, high-resolution observations of snow water equivalent, moreover, hampers the validation of climate model outputs (Mudryk et al., 2015), especially with respect to streamflow. Future improvements in snow simulations are required in the domains of increased horizontal resolution (Boyle and Klein, 2010), snow transport and post-depositional redistribution, and refinement in the representation of precipitation processes, including convection and cloud microphysics (Kang et al., 2015).

The increasing abundance of field data helps refining our understanding 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), the specificities of cryosphere data require adaptations of the general framework of statistical modelling. Cryosphere-specific difficulties 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). Taking maximum advantage of data sets of increasing size, variety and quantity also involves proceeding in parallel with the development of adapted and comprehensive statistical models (Gilks et al., 2001; Wikle, 2003; Cappé et al., 2005). The easiest way to address spatio-temporal data is to separate space and time effects (Cressie and Wickle, 2011). However, temporal evolutions at small spatial scales cannot be inferred in this manner. Regional climate change interacting with topography has, for example, resulted in different evolutions of avalanche activity over different parts of the French Alps (Lavigne et al., 2015). On this basis, other applications of non-separable spatio-temporal covariance models (Gneiting et al., 2007; Genton and Kleiber, 2015) have great potential.

Similar challenges that need to be addressed in permafrost modelling include: (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 forced not only 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 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.) prohibit 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).

In site-specific model studies, these subsurface data are often available due to borehole drillings and geophysical surveying (cf. Section 2.3). High-resolution subsurface models can then be calibrated to the observed conditions, and used for subsequent climate impact studies, as long as model calibration can be assumed to be constant in a future climate. These models have often 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, Haberkorn et al. 2017) and have been successfully extended to simulate the complex permafrost environment. Recently, also explicit permafrost models have been developed (Cryogrid 3, Westermann et al. 2016). Model inter-comparison studies with uncalibrated model set-ups show, however, that due to the abundance of permafrost-relevant processes in atmosphere, 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 is especially due to the difficulty to simulate phase changes in permafrost and corresponding latent heat transfer correctly, which becomes even more important the larger the uncertainty about initial ground ice content is.

Challenges in regional or hemispheric permafrost modelling therefore include not only numerical aspects or process-oriented model improvements, but also data availability and up-scaling issues (Fiddes et al. 2015, Westermann et al. 2015). Most land surface schemes of current GCM's and RCM's now have soil freezing schemes included (e.g., McGuire et al. 2016), however, neither reliable ground ice content maps as input nor ground temperature maps for validation exist. In combination with the need for good snow, soil moisture and vegetation data, this lack of deeper subsurface information poses the largest uncertainties in current and future permafrost temperature and distribution estimates.

3.3 Estimating liquid/solid precipitation in complex terrain

Large uncertainties affect the estimates of solid and liquid precipitation at high elevations (Rasmussen et al., 2012). These uncertainties mainly arise from two situations: (a) The low density of precipitation gauges at high elevation (only 3% of homogenized stations worldwide are located above 2000 m a.s.l., and less than 1% above 3000 m a.s.l.; Pepin et al., 2015); while areas >3000m comprise only a small fraction of the European landscape, it is exactly these regions where many cryospheric processes are located, e.g., permafrost or perennial snow cover. (b) The large biases in precipitation observations due to under-catch at high elevations is on the order of 30% (Adams and Lettenmeier, 2003; Yang et al., 2005) and is particularly large for solid precipitation. This is because solid precipitation is particularly influenced by wind, and perturbations due to icing and riming. Efforts are currently ongoing to address these problems (e.g., Solid Precipitation Intercomparison Experiment, WMO), but reliable references for ground truth measurements are still not available.

The precise quantification of precipitation and its spatial and temporal distribution are crucial for predicting future water availability. The spatial distribution of precipitation 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. How these processes may change as a response to future modifications in local and synoptic wind patterns is presently poorly understood.

Recently, remote sensing methods such as terrestrial and airborne laser scanning or radar have been successful in quantifying solid and liquid precipitation. Recent progress in measuring snow distribution in mountains (Grunewald et al., 2010; Kirchner et al., 2014) has allowed a better understanding of typical distribution patterns of Alpine water resources (Grunewald et al., 2014) as well as making a link to precipitation (Scipion et al., 2013; Mott et al., 2014). The results have highlighted that even in highly instrumented mountain ranges such as the Alps, total precipitation is very poorly quantified.

The combination of new measurement options with more classical ones such as precipitation radar will lead to a more complete understanding of precipitation amounts in high mountains.

3.4 Glacier mass changes

3.4.1 From local to regional assessments

The temporal and spatial incompleteness of available glacier mass balance data, limits the estimates of the contribution of glacier melting to sea-level rise, water resources, and biodiversity in mountain catchments. For regional assessments of glacier mass balance, combining local studies for selected glaciers with remote sensing for larger regions is a priority. This is also a strategy within the integrative monitoring approach presented by the Global Terrestrial Network for Glaciers (e.g., Haeberli et al., 2000).

The increasing number of satellite sensors, together with their improved spatial, radiometric, and temporal resolutions, has made remote sensing essential for the monitoring of glaciers. Computing decadal glacier-volume variations at the regional scale from the differencing of digital elevation data has become a standard technique (e.g., Berthier et al., 2014). Retrieving glacier-wide annual or seasonal surface mass balance is more challenging but can be assessed by measuring the end-of-summer snow line as a proxy for the equilibrium-line elevation (e.g., Rabatel et al., 2005), or using albedo maps of the glacier surface (Dumont et al., 2012). Recent studies showed that winter and summer balances can be quantified either by integrating the albedo signal over the accumulation or ablation period (Sirguey et al., 2016), or by using seasonal snow maps derived from the SPOT-VGT sensor (Drolon et al., 2016). *In-situ* data are, however, still required to calibrate relationships and validate the methods.

New satellite sensors (e.g., Sentinel-1 and 2) provide also the possibilities to complement *in-situ* measurements for glacier surface flow velocities. This has been done for large regions and very short time intervals (5-10 days) (Dehecq et al., 2015; Kääb et al., 2016). Such data are crucial to monitor the dynamic state of glaciers, to aid inverse approaches for estimating ice thickness distribution, or for the assessment of glacier-related hazards.

3.4.2 Assessment of future changes

To increase the accuracy of future glacier projections and runoff estimates, a number of issues need to be addressed. Ice thickness distribution is an important input for dynamic glacier modeling. As it is impossible to measure ice thickness distributions of all glaciers individually, model applications are necessary. Several existing models have been compared within the Ice Thickness Models Intercomparison eXperiment (Farinotti et al., 2017), revealing that results largely depend on the quality of the input data (glacier outline, surface elevation, mass balance or velocities). New high-resolution satellite images make such input data available, thus opening the way toward improved future global estimates of glacier thicknesses.

Another challenge in glacier modeling is the use of approaches that explicitly consider ice dynamics for glacier evolution. 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), but such applications at the regional scale still require simplifications (Clarke et al., 2015). For estimating future glacier evolution, ice dynamics models need to be coupled to adequate representation of glacier surface mass balance.A key issue in this respect is the modeling of future surface mass balance at the mountain-range scale. The Glacier Model Intercomparison Project (Glacier-MIP, www.climate-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 changes of meteorological variables and in their downscaling at a spatial scale compatible with glaciers. Some studies use degree-day approaches for long-term simulations of glacier-wide surface mass balance (Réveillet et al., 2017), or also account for potential radiation (Hock, 1999). However, with a shift in energy fluxes at the glacier surface, calibrated degree-day factors might change in the future. Application of process-based models that are able to resolve the full energy balance are thus required (Hanzer et al., 2016), but the accuracy and resolution of the input data needs to be improved. A focus on modeling winter balance and the spatial distribution of snow accumulation is also needed (Réveillet et al., 2017). Additional studies should assess the impact of supraglacial debris and related feedbacks on the surface energy balance (Reid and Brock, 2010). This is particularly important as many glacier tongues tend to 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 represent a hazard or reduce the efficiency of hydropower plants (Lane et al., 2016).

3.5 Extreme snow events and related hazards

Heavy snow events and related phenomena such as avalanches are by definition rare. This makes them much less well understood and more difficult to forecast than “average” behaviors. This is reflected in the lack of related baseline data (IPCC, 2012; 2013). Also in mountains, 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). Snow-related extremes are often the result of a combination of different processes (e.g. wind and snow for drifting snow) making predictions of their future behavior highly uncertain. In contrast, they are important natural hazards in European mountains, where they often puts people and infrastructures at risk. For example, winter storms often hinder mobility by disrupting rail, road and air traffic. Extreme snowfall can overload buildings and cause them to collapse, and can lead to flooding due to subsequent melting. Deep snow, combined with strong winds and unstable snowpack, contributes to the formation of avalanches, and can cause fatalities and economic loss as a result of damage to property or communication routes.

3.5.1 Changes in snow extremes

Whether extreme snowfall and snow depths will decrease or not in European mountains in the future remains an unsolved question for now. This is because of the limited results available, and because of possible compensation mechanisms between warmer temperatures, more intense precipitation, and increased climate variability, that all make the future regime of snow storms (i.e. their number, magnitude, and timing) difficult to anticipate. In addition, nearly all available results concern marginal distributions, or make an assumption of stationarity (Blanchet and Davison, 2010; Gaume et al., 2013). Recent results, however, suggest that the dependence structure of extreme snowfalls may be affected by warming (Nicolet et al., 2016). Such information is important for extrapolating outside of observation points and evaluating integrated quantities.

More detailed knowledge about the evolution of extremes in snow properties will be relevant as well. The moisture content or density of snow is needed, for example, when evaluating the probability for a particular infrastructure to collapse under future extreme snow loads (Sadovský and Sykora, 2013). This topic has not yet been addressed, and developments in jointly projecting the evolution of the different variables are required. The same holds true for projections of heavy drifting snow events resulting from wind gusts. To date, the combined evolution of snow amount, type (dry, wet), and density in complex mountain topography remains virtually unknown. More generally, impact models relating socio-economic consequences to extreme snow events (e.g. roof collapse probability as a function of snow mass and roof technology, or risk to road traffic as function of snow storm magnitude) remain oversimplified. Efforts are required to combine snow-climate and vulnerability-assessment expertise (Favier et al., 2014) if realistic future projections are to be made.

3.5.2 Changes in snow avalanche activity in relation to snow and ecosystem changes

Even if empirical relations between snow avalanche activity and climate do exist (Mock and Birkeland, 2000), knowledge of long-term responses of avalanche risk to climate change remains largely insufficient. With a few exceptions, studies focus on the very recent decades, and exist only for a very restricted number of regions (Stoffel et al., 2006; Corona et al., 2012, 2013; Schläppy et al., 2014, 2016). Direct effects of climate change on the avalanche number, timing, magnitude and type mainly exist in the form of changes in snow amounts, snowfall succession, density and stratigraphy as a function of elevation. Indirect effects are linked to changes in forest locations, size, and species composition. Notably, the ongoing rise of tree lines may reduce both avalanche frequency and magnitude. This is because of the reduction of potential release areas, the reduction in triggering susceptibility (as a result of the anchoring effect of trees) and the reduction in runout for a given snow amount. However, avalanche-forest interactions remain complex processes, and are not yet fully understood, even under stationary conditions (Bebi et al., 2009). A possible general shift in elevation of avalanche activity may be hypothesized, but this is neither proven, nor generalizable at the very local level. Constituting and investigating long-term series of avalanche events (including historical and paleo-archives in addition to existing records; see Stoffel et al., 2010) will be required to test this hypothesis.

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), investigating how snow-climate controls the physics of snow avalanches remains necessary for realistic projections. In particular, it is unclear whether warmer temperatures always lead to fewer avalanches because of less snow. This is because of potentially higher instability levels in winter linked to larger climate variability (Beniston, 2005). 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 (Castebrunet et al., 2014). Recent results show that wet-snow avalanches indeed have a high damage potential due to their potentially long runouts and high impact pressures (Sovilla et al., 2010; Ancey, 2015). Hence, specific investigation of the rheology of such flows is required to realistically anticipate future changes in avalanche risk.

3.5.3 Other snow contributions to mass movements and cascading processes

In addition to snow avalanches, snow plays a role in numerous other mass movements and/or cascading processes, and understanding their temporal evolution is important. Until recently, for example, slush-flows – mixtures of water and loose snow (Hestnes, 1998) – were mostly documented in Scandinavian mountains during a rather short spring period only (Schlyter et al., 1993). Local testimonies, however, now report such events over larger areas and longer time periods. This seems to be in relation to changes in snow cover characteristics. Similarly, mixed ice and snow avalanches are expected to become more common due to the retreat of hanging glaciers and the resulting ice falls. This was observed for the Grandes Jorasses (Margreth and Funk, 1998; Vincent et al., 2015) and Taconnaz cases, for example. Also in relation to permafrost degradation and glacier recession, high snow amounts could play a role (Stoffel et al., 2014b). To which extent ongoing warming will affect the frequency of such processes in European mountains remains to be investigated in greater detail. In order to reduce expected impacts, enhanced efforts are also required to better define adaptation and mitigation strategies. These include the detection of favorable locations for infrastructure, the better prediction of avalanche timing and magnitude, and the design of efficient early warning systems.

3.6 Shifts in geomorphic risks as a function of changing cryosphere conditions

Changes in air temperatures and precipitation are likely to affect the frequency and magnitude of mass movements such as shallow landslides, debris flows, rock slope failures, or ice avalanches (Stoffel and Beniston, 2006; Stoffel et al., 2014a, b). So far, however, changes in mass-movement activity can hardly be detected in observational records, making the projection of the future evolution of such phenomena particularly challenging.

The largest and most important changes and impacts related to permafrost thawing have yet to occur. In general terms, smaller permafrost bodies with deeper freeze-thaw cycles are expected. Increasing air temperatures are obviously controlling such changes, but other factors, such as micro-climate, terrain and soil properties, as well as the onset and duration of snow cover can play important roles (Scherler et al., 2014). Understanding the interplays between these mechanisms and going beyond temperature-based projections will be a key for enhancing the reliability of future projections.

Changes in temperatures and precipitation will not only affect permafrost, but are projected to influence the frequency and magnitude of mass wasting processes in mountain environments more in general (IPCC, 2012; Gobiet et al., 2014). This is especially true for processes driven by water, such as debris flows (Stoffel and Huggel, 2012; Borga et al., 2014). A warmer climate also results in more precipitation to fall in liquid form at high elevations, thus increasing the area contributing effectively to runoff (Beniston, 2005; Stoffel and Beniston, 2006). At the same time, however, increasing air temperatures may allow vegetation to colonize higher elevations, possibly stabilizing loose material (Baroni et al., 2007). To date, this interplay between long-term vegetation evolution and various types of slope instabilities is poorly understood and loosely quantified.

The temporal evolution of debris-flow frequencies has been addressed for a series of high-elevation catchments in the Swiss Alps (Stoffel et al., 2011, 2014a, b). Based on statistically downscaled RCM data and an assessment of sediment availability, these studies 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. The magnitude of the events, however, 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 and allows for larger volumes of sediment to be mobilized (Lugon and Stoffel, 2010). The 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 remains challenging, and significant efforts are required if the associated uncertainties are to be reduced. This is particularly important in the European Alps, where the exposure of people and infrastructure to hazards related to mass movements is high (Haeberli, 2013).

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 clearly seem related to de-glaciation processes (Fischer et al., 2010; Korup et al., 2012;Strozzi et al., 2010). Extremely warm temperatures have additionally been associated with these processes as the penetration of melt water 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, it is generally difficult to attribute individual events to one single factor (Huggel et al., 2013), and improved integrative assessments are necessary.

Further evidence of climatic impacts on high-mountain rock slope stability comes from the analysis of historical events. For the Alps, inventories documenting events since 1990 exist (Ravanel and Deline, 2011; Huggel et al., 2012) and indicate a sharp increase in the number of events since 1990. Monitoring and documentation efforts for rock slope failures have been intensified during the past decades, thus introducing a certain bias as compared to the early 20th century. This is especially true for small rock-fall events. Although the documentation for large slope failures (e.g. >100,000 m^3) can be assumed to be reasonably complete, improving the homogeneity of the datasets upon which trend-analyses are built is important if correct conclusions are to be drawn. In Switzerland, for example, the temporal distribution of rock slope failures resembles the evolution of mean annual temperatures, but it is unclear to which degree this correlation is affected by varying temporal completeness of the underlying datasets. The temperature sensitivity of rock slope stability in high mountains should therefore be further investigated.

3.7 Evaluating and communicating uncertainty

As outlined throughout the manuscript, predicting the future evolution of cryospheric components is challenging. On the one hand, the challenges stem from the incomplete understanding of the processes leading to given changes, on the other, future predictions are intrinsically affected by a range of uncertainties. Adequately evaluating and communicating such uncertainties is all but a trivial task, and this is both because the interplay between individual 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.

The key element driving future changes in the cryosphere is, obviously, the evolution of future climate. Uncertainties in future climate projections will thus inevitably propagate to any change derived therefrom. Increasing the awareness of the kind of uncertainties which affect projections of future climate is therefore of paramount importance. Clearly making a distinction between the concepts of “prediction” (or “forecast”) and “scenarios” (or “projection”) for example, is central: Whilst the first concept refers to the assessment of the likelihood with which a future event will happen given the evidence that is available up to a certain point in time, the second describes the consequences arising if a certain set of assumptions are to become true in the future. As an example: A meteorological forecast aims at telling what weather will occur during the upcoming days by assessing the state of a given set of variables that can be measured at the moment the forecast is issued; a climate change scenario, instead, aims at telling what the mean atmospheric conditions will be in several decades, if a given change in radiative forcing was to occur.

Another important point to be made is that measurements are also affected by uncertainties. This may seem trivial at first, but is neglected all too often outside the scientific community, where “measurement” is often interpreted as equivalent to “truth”. Climate model simulations, for example, are often validated against gridded observational datasets. Translating station-based information to gridded data products, however, requires several steps including quality control, homogenization, and interpolation for instance. Differences in grid resolution, station density, interpolation method, and sampling error add additional uncertainties which can, in case of precipitation for example, even be variable in time and space (Rudolf et al., 1994; Schneider et al., 2014). As an example, Fig. 9 shows the magnitude of the resulting differences by comparing 3 different datasets of “observed” winter and summer precipitation for central Europe. Large differences are particularly evident in regions of complex topography, such as the Alps, Norway, mountainous parts of Italy, the Carpathians, the Pyrenees, or the west coast of the British Isles.

Climate models are obviously prone to uncertainties as well, which are related to: (1) the expected climate forcing, (2) natural climate variability, and (3) internal model variability (Tebaldi and Knutti, 2007; Hawkins and Sutton, 2009). Whilst natural variability dominates uncertainty at time scales up to a few decades, scenario uncertainty is dominant on even longer time scales. Model uncertainty can be important across time scales (Latif, 2011), and is again most prominent for mountains and complex topography (Fig. 10). This is principally related to the differences in model resolution and model parameterizations.

4 Conclusions

This review has brought together a sample of the knowledge and the experience of a number of experts in the fields of mountain snow, ice and permafrost in European mountain regions, in order to convey their views on the prospects and challenges for research on the cryosphere as it responds to past, current, and future changes in climate. While we do not claim to have provided an exhaustive overview, the paper has nevertheless addressed the current state of knowledge in terms of the observed evolution of the European mountain cryosphere and associated impacts – notably on water, ecosystems and the services provided by these resources. A catalog of challenges has been identified, focusing on as-yet unresolved issues of data access, high-resolution modeling, quantification of risks and extreme events, and communication of uncertainty. These issues have an obvious effect on our capability of projecting future shifts in the mountain cryosphere, and the impacts that these shifts are likely to generate. The latter will have a bearing on the viability of a number of economic sectors, including notably hydropower, agriculture, and tourism.

In this paper, much attention has been devoted to 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 data of high quality for both understanding the functioning and evolution of the various elements in specific regions, and assess future changes in snow, ice and permafrost via modelling.

Access to state-of-the art models using high spatial and temporal resolution is essential to furthering our understanding of feedbacks between the atmosphere, the hydrosphere and the cryosphere, and the future behavior of cryospheric processes as a function of greenhouse-gas emissions. Global climate models have seen their resolution increase in recent decades, but much of the information still remains too coarse for most mountain cryosphere research. Physically-based, nested global-to-regional modeling techniques can provide adequate data for atmosphere-cryosphere studies. However, such results are highly-dependent on the initial and boundary conditions that drive the coupled models, and errors in these conditions obviously propagate into the model solutions.

Communicating research results on climate and cryospheric science is a challenge that needs careful consideration. The importance and imminence of climatic change is generally 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 photography portraying glacier evolution over time. In this sense, climate-induced changes in the cryosphere enable a unique and convincing form of communication to the public and to policymakers, and more effort should be dedicated to illustrate how these changes can impact upon water resources, mountain ecosystems, natural hazards, and thus a wide range of economic activities. Elevation-dependency and regional patterns of the phenomena related with a cryosphere adapting to rapidly changing climatic conditions is an important issue for future comparative research.

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 respond to the changes in climate in coming decades, there will clearly be an increasing need for adaptation strategies based on robust knowledge.

5 Acknowledgements

Martin Beniston and Markus Stoffel would like to acknowledge the support of the University of Geneva for the international cryosphere workshop “From Process Understanding to Impacts and Adaptation”. This meeting took place in Riederalp (Valais Alps), Switzerland, 15-19 March, 2016. Much of the material presented in this article is inspired from the presentations made at this meeting.

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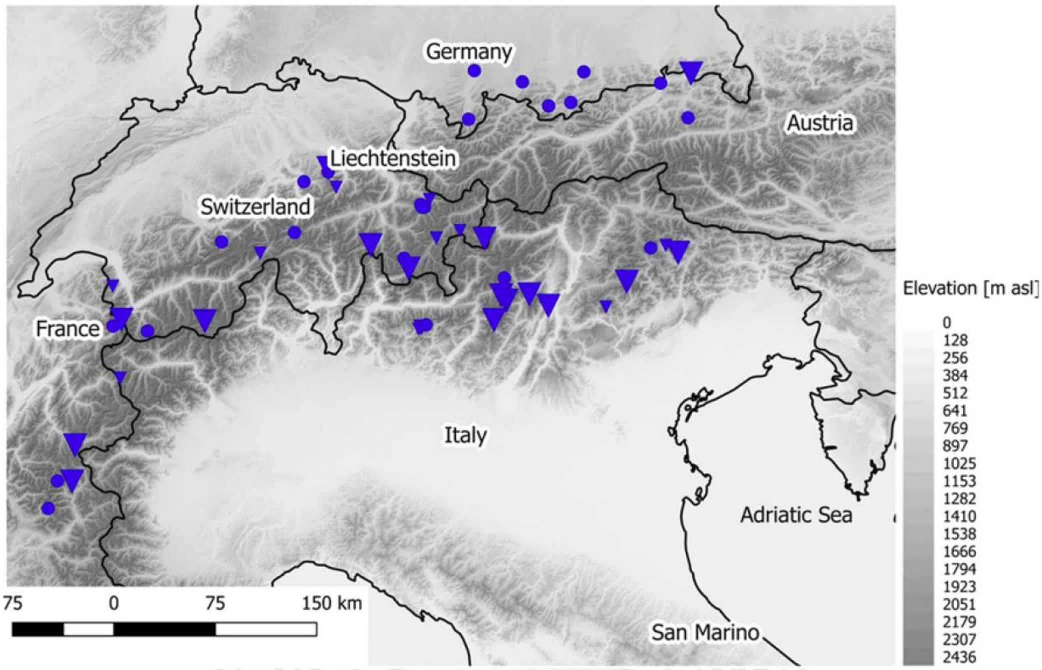


Figure 1: Geographical distribution of the 45 year trend (1968-2012) for April 1th SWE in the Alps. All stations show a negative trend. Large triangles indicate significant trends (p = 0.05) and small triangles indicate weakly significant trends (p = 0.2). Circles represent stations with no significant trend (p = 0.2). The elevation is given in gray. (Adapted from Marty et al. 2017b)

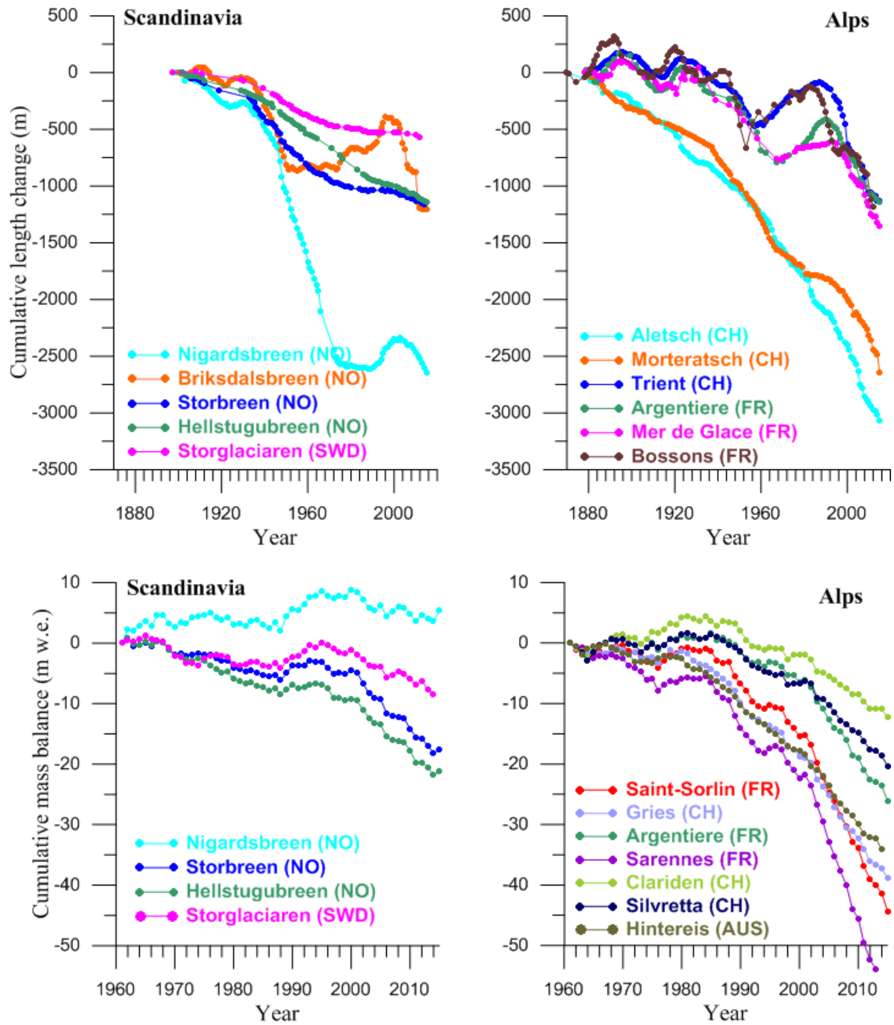


Figure 2: Length and surface mass balance changes documented with *in-situ* measurements for glaciers in Scandinavia and in the European Alps. Sources: WGMS (2015) and earlier issues with updates (Andreassen et al., 2016).

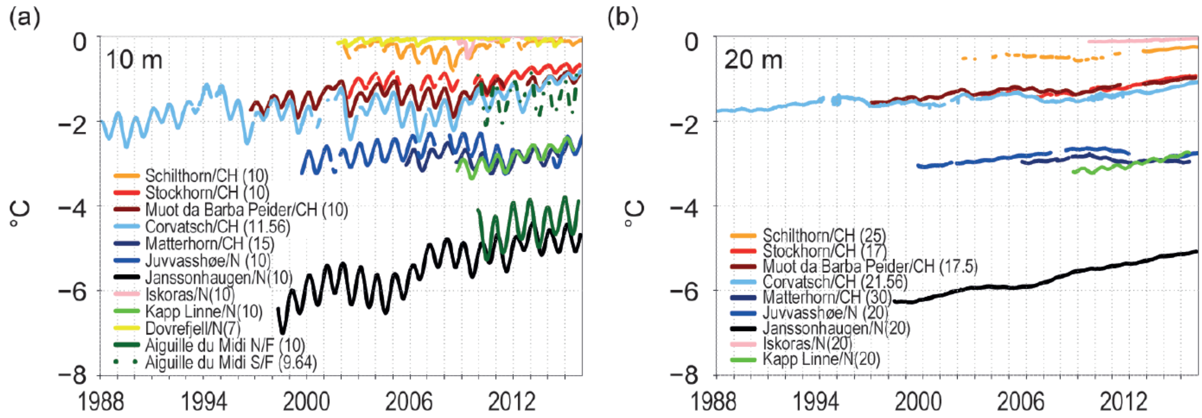


Figure 3: Temperature evolution of mountain permafrost in Norway (N), France (F), and Switzerland (CH) measured in boreholes at (a) 10m and (b) 20m depth (exact depth given in parenthesis). Figure adapted from Noetzli et al. (2016).

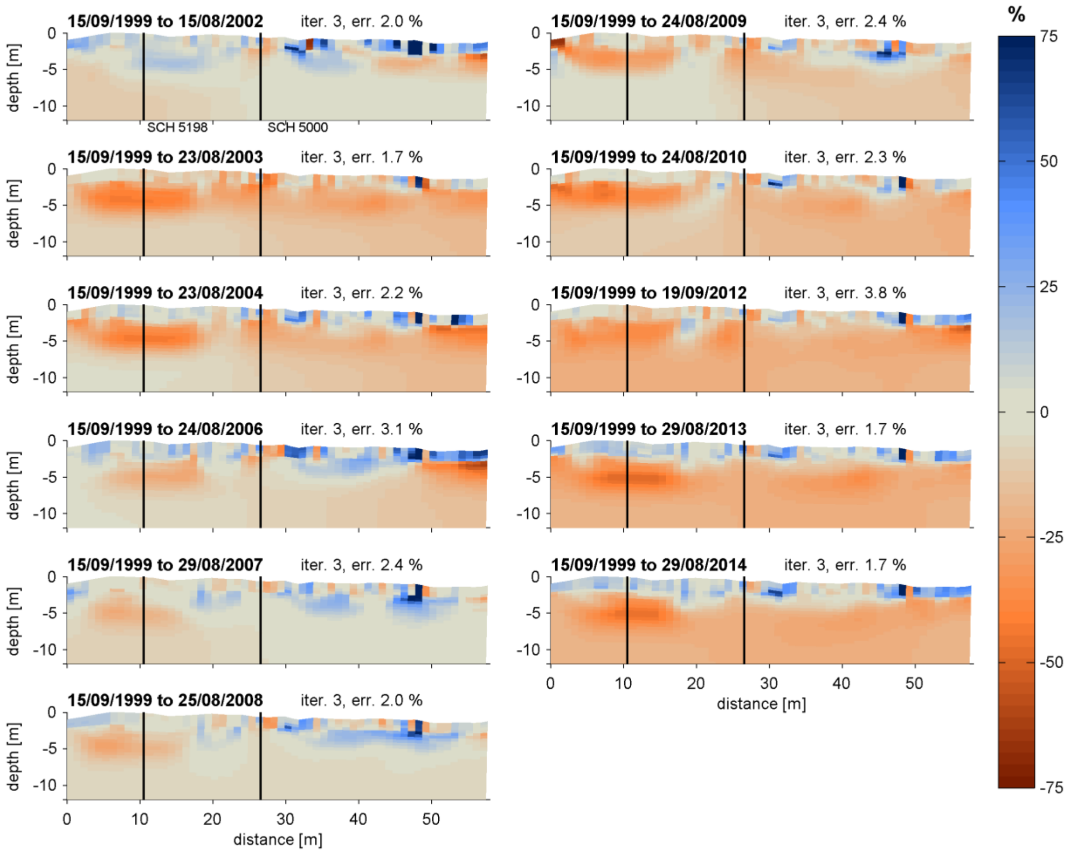


Figure 4: 15-year change in specific electrical resistivity (given as % specific resistivity change) along a 2-dimensional Electrical Resistivity Tomography (ERT) profile at Schilthorn, Swiss Alps (2900 m a.s.l.). Red colors denote a resistivity decrease corresponding to loss of ground ice with respect 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)

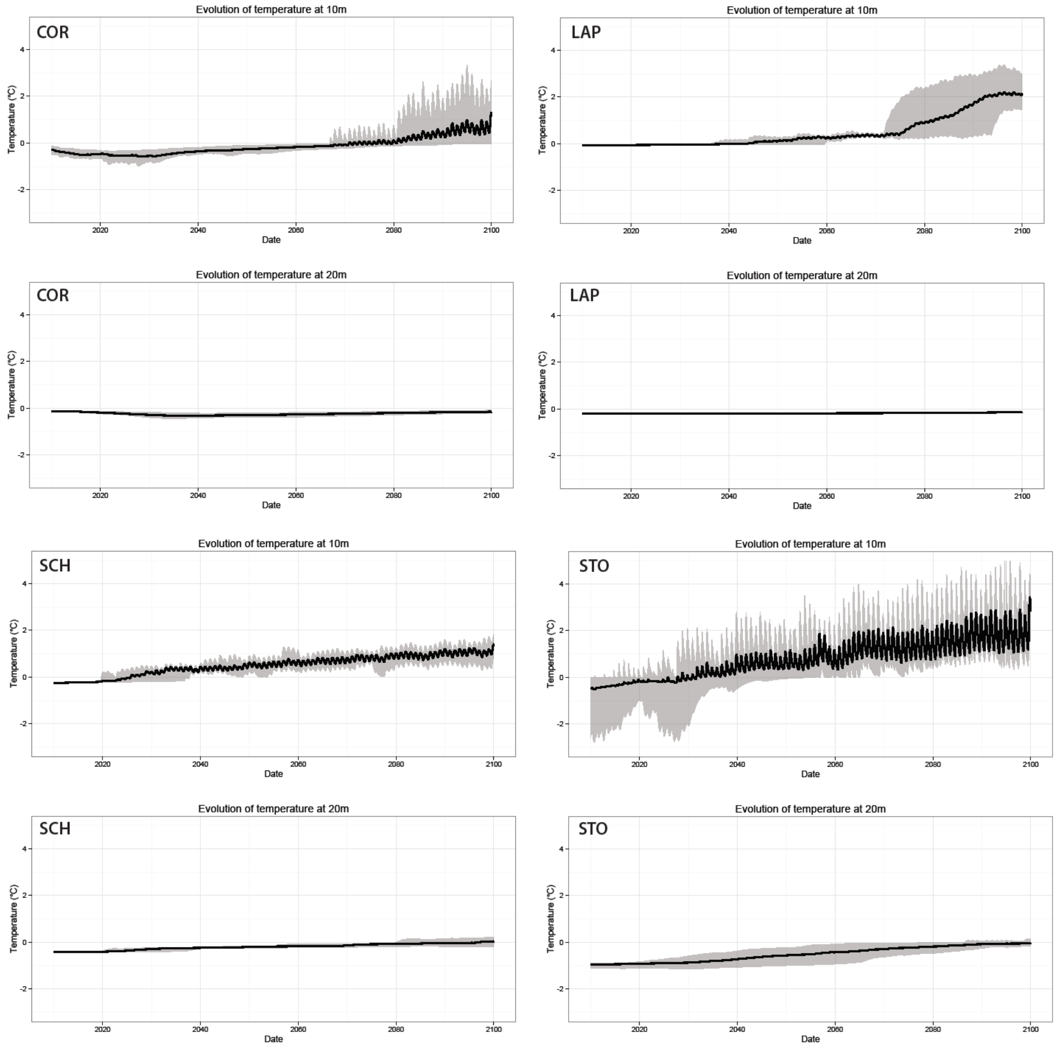


Figure 5: Modelled long-term evolution of ground temperatures at 10 m and 20 m at four different permafrost sites in the Swiss Alps (COR: Murtèl-Corvatsch, LAP: Lapires, SCH: Schilthorn, STO: Stockhorn), as simulated with the COUP model (Marmy et al., 2016). The black lines represent the median scenario 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).

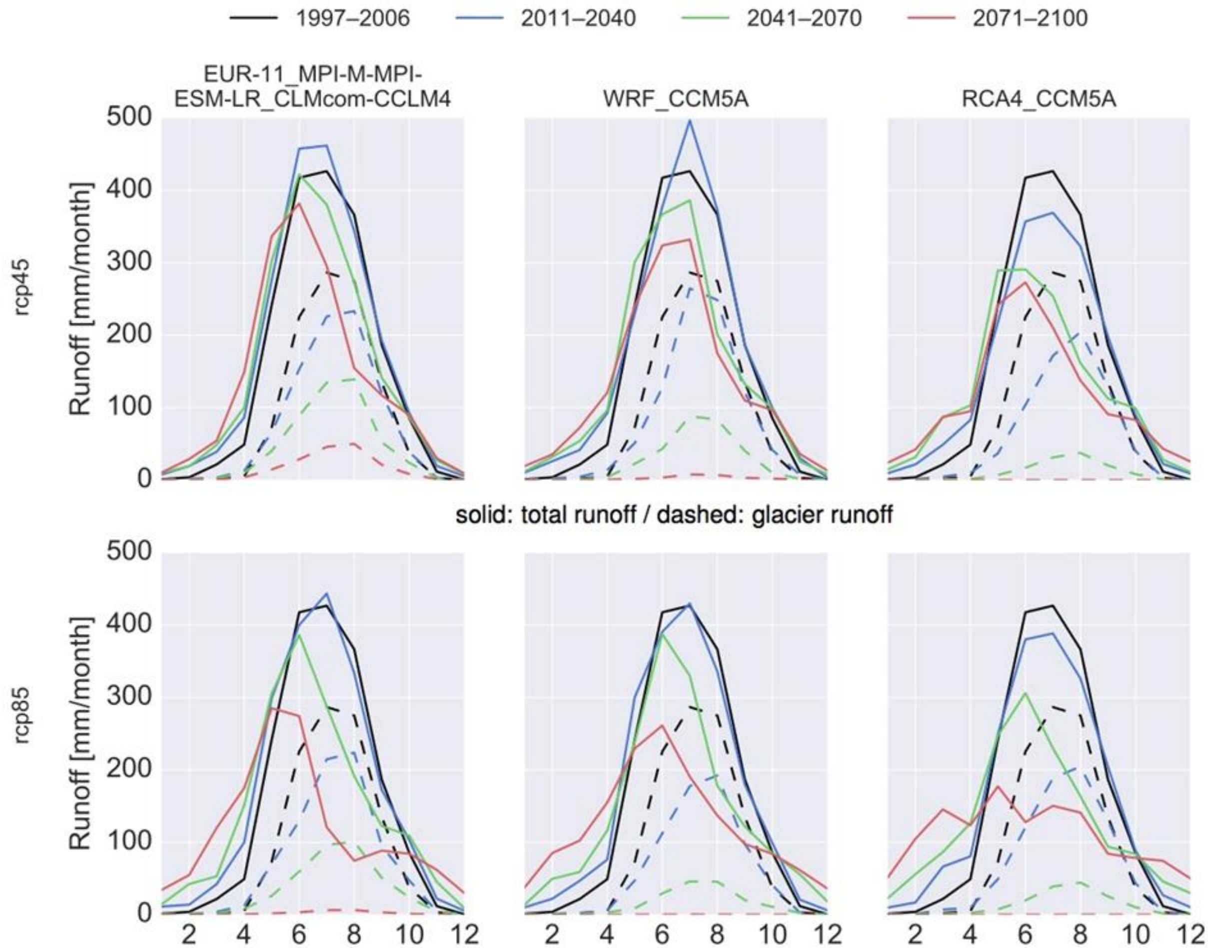


Figure 6: Shifts of streamflow regimes for the Rofenache catchment (Austrian Alps, 1891–3762 m a.s.l., 98 km², ~35 % glacierization as of 2006) as simulated with the AMUNDSEN model using downscaled EURO-CORDEX projections for the RCP2.6, RCP4.5 and RCP8.5 scenarios. Solid and dashed lines indicate the multi-model mean total and ice melt runoff, respectively, and shaded bands indicate the climate model uncertainty shown as ± 1 standard deviation. Figure adapted from Hanzer et al. (2017).

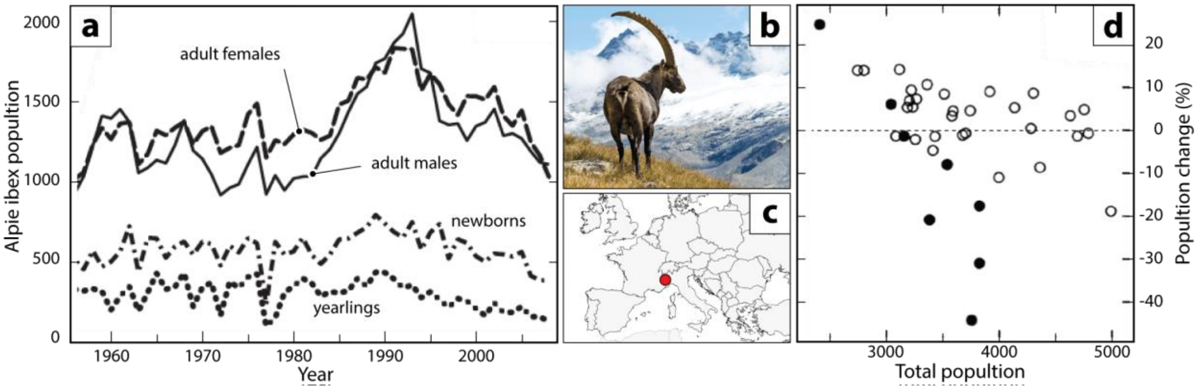


Figure 7: (a) Total number of adult Alpine ibex (b) counted at Gran Paradiso National Park, Italy (c). (d) Relative population change against population size. Solid circles indicate that the winter snow depth was more than half a standard deviation above the long-term average. Panels (a) and (b) are adapted from Jacobson et al. (2004).

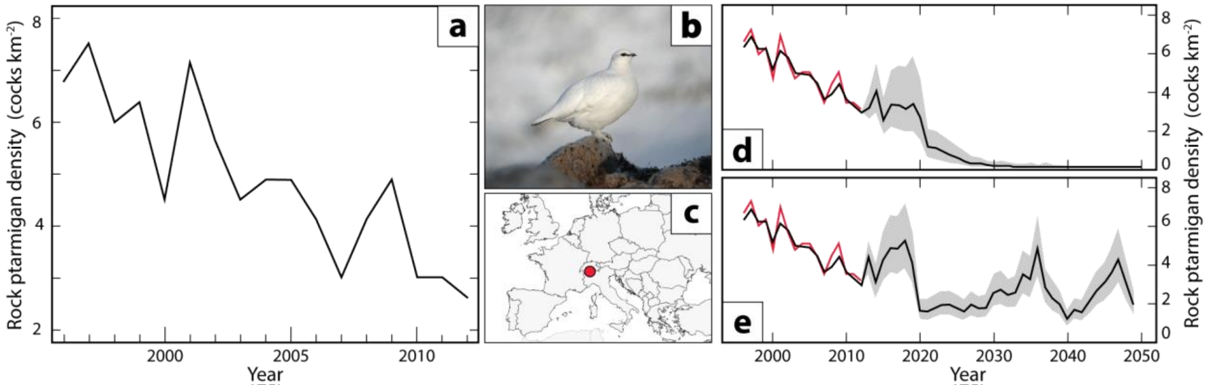


Figure 8: (a) Observed density of rock ptarmigan cocks (b) at the Veglia Devero protected area, Italy (c). (d+e) Reconstructed (red) and projected (black) rock ptarmigan density from two population dynamics models including (d) snow drivers only, and (e) snow and delayed density dependence. Panels (a), (d) and (c) are adapted from Imperio et al. (2013)

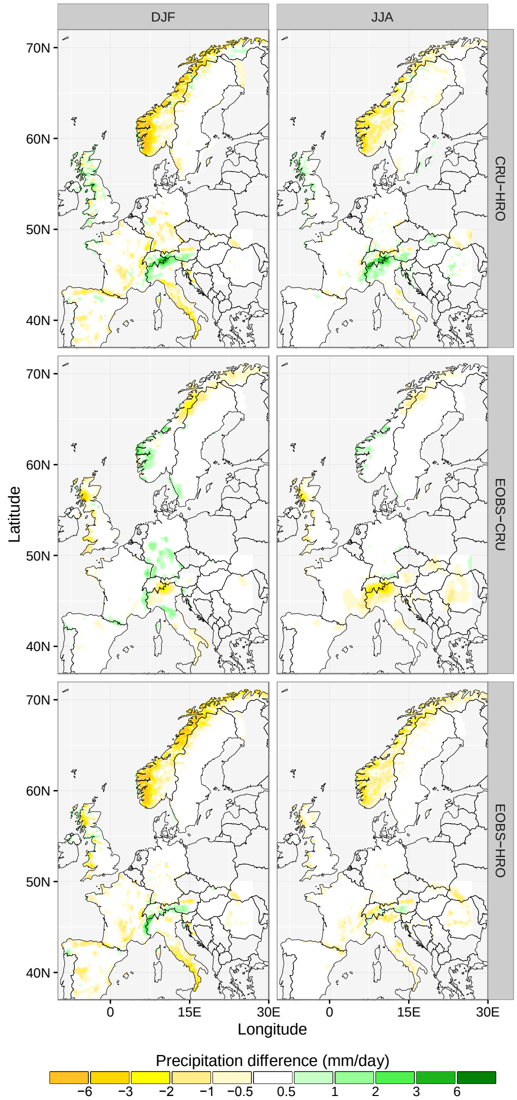
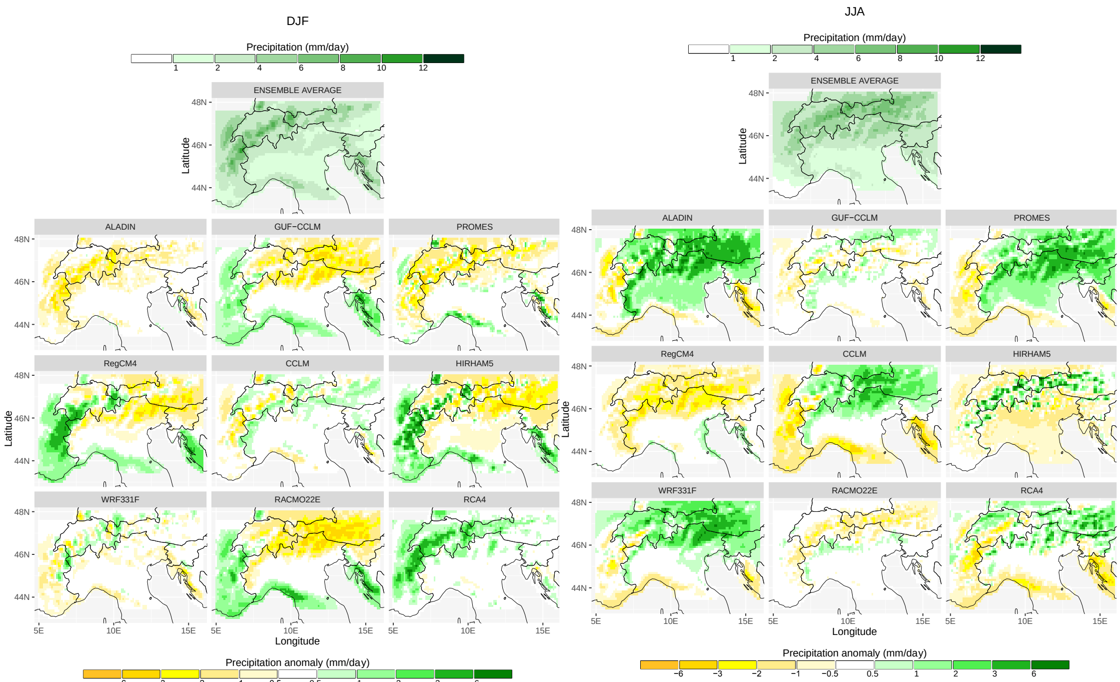


Figure 9: Seasonal average precipitation differences for December-January-February (DJF; left) and June-July-August (JJA; right) between CRU and HRO (first row), E-OBS and CRU (second row), and E-OBS and HRO (third row).



a b

Figure 10: DJF (a) and JJA (b) precipitation as derived from 9 regional climate models. The average of the model ensemble is shown in the top panel at the center.