Chapter 2: High Mountain Areas

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Executive Summary

The cryosphere, which includes glaciers, snow cover, permafrost, and river/lake ice, is a prominent feature in high mountain environments influencing down valley lowlands far beyond the mountains themselves. Recent and projected changes in the mountain cryosphere and associated risks and impacts on a multitude of societal needs, including adaptation responses, are synthesized in this chapter.

How has the High Mountain Cryosphere changed? Why and what have been the Impacts and Responses?

High mountain regions, as part of the global climate system, have experienced significant warming since the beginning of the 20th century and decrease in snowfall below the mean snowline elevation (*very high confidence*[1]). Temperature trends often increased with increasing elevations or were more pronounced around the mean 0°C elevation (*medium confidence*). Total precipitation often showed insignificant trends, and experienced large variability (*high confidence*). {2.2.1.1, 2.2.1.2, Box 2.1}

The snow cover amount and duration has declined in many high mountain regions since the beginning of the 20th century, especially at or below the mean snowline elevation (*very high confidence*) although subject to high variability. The decline is mostly due to atmospheric warming and associated decrease in snowfall and increase in snow melt (*very high confidence*). {2.2.2}

The vast majority of glaciers in all high mountain regions have been retreating and losing mass during the last two decades (*very high confidence*). Mass losses from the glaciers in 11 glacierized mountain regions increased from 470 ± 80 kg m^{-2} yr^{-1} in the period 1986-2005 to 610 ± 90 kg m^2 yr^{-1} between 2006-2015. Regional-scale average mass losses in units of kg m^2 yr^{-1} in the latter period were largest in the southern Andes, the low latitudes and central Europe (>900 kg m^2 yr^{-1}), and smallest in High Mountain Asia (190 kg m^2 yr^{-1}), where mass budgets have been balanced or slightly positive in some regions due to local meteorological conditions partially counteracting the effects of increasing air temperatures. {2.2.3.1}

In situ measurements in the European Alps, Scandinavia, and the Tibet Plateau show that permafrost has undergone warming, degradation and ground-ice loss in the past two decades (*high confidence*). The observed rates of change in the 21st century are higher than in the late 20th century (*medium confidence*). Other mountain regions lack in-situ observations to assess trends. {2.2.4}

Glacier shrinkage and snow cover changes have led to changes in the amount and timing of river runoff in many mountain regions during the last two decades (*high confidence*). In some regions with predominantly small glaciers (e.g., western USA and Canada), runoff from glaciers has decreased due to glacier shrinkage while in other regions typically with larger glaciers (e.g., Alaska) runoff from glaciers has increased (*medium confidence*) {2.3.1}. Runoff changes have caused significant shifts in downstream nutrients (dissolved organic carbon, nitrogen, phosphorus) and influenced water quality through increases in heavy metals, particularly mercury, and other legacy contaminants. {2.3.1.2}

The decline of the cryosphere in recent decades has affected the frequency and magnitude of some natural hazards. Retreat of mountain glaciers and thaw of mountain permafrost has decreased the stability of mountain slopes (*high confidence*). Glacier retreat has led to an increasing number and area of glacier lakes (*high confidence, medium agreement)* {2.3.2.1}. Over the past decades, there has been an increase in wet snow avalanches, and a reduction in size and run-out distance of dry snow avalanches (*medium confidence*) {2.3.2.1.2}**.** There is *high confidence* that the exposure of people and infrastructure to natural hazards in high mountain areas has increased. Cryosphere-related landslides and floods have caused severe impacts on lives, livelihoods, and infrastructure that often extend beyond the directly affected areas and that result in differentiated losses across mountain regions. {2.3.2.2}

[1] FOOTNOTE: In this Report, the following summary terms are used to describe the available evidence: limited, medium, or robust; and for the degree of agreement: low, medium, or high. A level of confidence is expressed using five qualifiers: very low, low, medium, high, and very high, and typeset in italics, e.g., *medium confidence*. For a given evidence and agreement statement, different confidence levels can be assigned, but increasing levels of evidence and degrees of agreement are correlated with increasing confidence (see Section 1.9.3 and Figure 1.4 for more details).

In terrestrial and freshwater ecosystems, biodiversity has increased overall due to changes in snow cover, permafrost thaw and degradation, and glacier retreat, although some specialist taxa have been lost (*high confidence*) {2.3.3}. Camouflage has affected wildlife with increased exposure of white coats in winter to brown snowless ground, thereby compromising animal movement and potential predation and lead to range contraction. Multiple interacting cryosphere-related challenges, including survival under a shallower and denser snowpack, have implications for high profile species, such as wolverines and alpine ungulates. {2.3.3.3}

Observed changes in the cryosphere are exerting considerable yet differentiated impacts in agriculture, hydropower, tourism and recreation activities, and other sectors in the mountains since the mid-20th century, while evidence on the long-term effectiveness of adaptation responses remains uneven and limited (*medium confidence*). Adoption of new crops and irrigation techniques have reduced vulnerability of some high mountain agricultural communities to reduced stream flow linked to glacier retreat and changes in snow amounts. Managers of hydropower facilities incorporate projections of stream flow into their planning to reduce their vulnerability to changing water amounts. Snow management, including snowmaking, has reduced the vulnerability of some mountain ski resorts to inter-annual variability and past decline of natural snow amounts. However, adaptation measures in agriculture, hydropower, tourism and other sectors are generally limited in scope, short-term focused, and fragmented. The diverse priorities, conditions and mechanisms available for the implementation and evaluation of these measures place constraints on the measures and on the assessment of these measures. {2.3.1.3.1, 2.3.1.3.2, 2.3.4.1}

How is the High Mountain Cryosphere Projected to Change, what are the Expected Consequences and Adaptation Options?

Under all considered climate scenarios for the 21st century, air temperature in high mountain regions is projected to increase, exceeding average global warming rates and driving further reductions in snowfall below the mean snowline elevation (*very high confidence*). Elevation dependent warming is projected to amplify in many mountain regions (*medium confidence*). Total precipitation is projected to show limited long-term changes, except at highest elevations where it is projected to increase (*medium confidence*). {2.2.1.1, 2.2.1.2, Box 2.1}

The mass and duration of the mountain snow cover are projected to remain highly variable, and decline by 25% [10 - 40%], below the mean snowline elevation, between the recent past period (1986-2005) and the near future (2031-2050), regardless of the Representative Concentration Pathway (RCP). For the end of the century (2081-2100), reductions up to 80% [50 - 90%] are expected under RCP8.5 and 30% [10 – 40 %] under RCP2.6. Significantly above the mean snowline elevation, steady or increased snow amounts are possible due to steady or increased total precipitation under below-freezing conditions (*medium confidence*). {2.2.2}

Glaciers in all mountain regions are projected to continue to lose mass throughout the 21st century. Projected mass reductions between 2015 and 2100 range from 29±7% for RCP2.6 to 47±10% for RCP8.5. In regions with relatively little ice cover (e.g., Central Europe, Caucasus, Low Latitudes, North Asia, Scandinavia), glaciers are projected to lose more than 80% of their current mass by 2100. {2.2.3.2}

Over the 21st century, permafrost is expected to undergo increasing thaw and degradation in response to rising air temperature (*high confidence*). Quantitative projections are scarce and limited to individual sites of some mountain regions. {2.2.4}

Changes in the high-mountain cryosphere are *likely*[2] to increase freshwater-related risks in some regions with high dependency on snow or glacier melt runoff by the end of the 21st century (*medium*

[2] FOOTNOTE: In this Report, the following terms have been used to indicate the assessed likelihood of an outcome or a result: Virtually certain 99–100% probability, Very likely 90–100%, Likely 66–100%, About as likely as not 33–66%, Unlikely 0–33%, Very unlikely 0–10%, Exceptionally unlikely 0–1%. Additional terms (Extremely likely: 95–100%, More likely than not >50–100%, and Extremely unlikely 0–5%) may also be used when appropriate. Assessed likelihood is typeset in italics, e.g., *very likely* (see Section 1.9.3 and Figure 1.4 for more details).

***confidence, high agreement*).** However, projected effects of the changes in magnitude and seasonality of runoff on hydropower, irrigation and drinking water are subject to widespread regional variation. Current capacities to explicitly account for glacier changes especially in large-scale hydrological models are limited, thus increasing uncertainty in decision making and taking adaptation measures to reduce potential vulnerabilities in certain regions. {2.3.1.1, 2.3.1.4}

There is *high confidence* that the projected retreat of mountain glaciers and thaw of mountain permafrost will continue to decrease the stability of mountain slopes, and that the number and area of glacier lakes will increase. There is also *high confidence (medium agreement)* that future cryospheric change will alter these landscape elements so that related disasters could manifest where there is no documented record of previous events {2.3.2.1}. As a consequence of ice and snow reduction on volcanoes, the hazards from floods and lahars involving melt water are projected to overall gradually diminish (*medium confidence*). Under all considered scenarios, shifts in snow avalanche activity and character are projected for the 21st century with more frequent wet snow avalanches in winter, and decline of the overall number and runout distance of dry snow avalanches in regions and elevations experiencing significant snow decline (*medium confidence*) {2.3.2.1.2}. There is *high confidence* that the exposure of people and infrastructure to high mountain natural hazards will continue further. In addition to other existing environmental and social stressors, there is *high confidence (medium agreement)* that the impacts from high mountain floods, landslides and avalanches will increase in regions where risk reduction and adaptation strategies are insufficient. {2.3.2.3}

There is *high confidence* that the structure and functioning of terrestrial and freshwater mountain ecosystems will change thereby affecting human health and well-being {2.3.5}. Key future shifts may include further upslope migration of lower elevation species and changes in the amount and timing of plant growth, increased predation of coat-color changing animals, shifts in the characteristic traits of many species, and increased potential for disturbance, e.g., increased fire and landslides, that could lead to loss or restriction in range in high mountain taxa due to the changing cryosphere {2.3.3.1}. Species extinctions may be slowed in terrestrial ecosystems by microclimate refugia (*medium confidence*) and accelerated in freshwater ecosystems due to greater variability in water resources (*high confidence*). Wide-ranging effects on large animals are projected to lead to population declines and smaller ranges where the species occur and change in behaviour (*high confidence*). {2.3.3.3}

Future cryospheric changes are projected to pose challenges to livelihoods and other economic activities in mountain regions, including agriculture, hydropower and tourism, especially under high-end climate scenarios (*high confidence)*. Existing local adaptation measures (e.g., extension of irrigation systems; current snowmaking technologies) are projected to approach their limits around 2°C of global warming since preindustrial. Moreover, vulnerabilities of mountain societies are projected to increase because of limits to their adaptive capacity (*medium confidence*) {2.3.1.4.2, 2.3.4.1}. Agriculture, hydropower and tourism activities related to the mountain cryosphere are projected to undergo major changes in the 21st century as a result of cryospheric change (*high confidence*); however these changes may also be driven by potential changes in, *inter alia*, socio-economic, technological, policy, institutional and legal aspects on access, mobility and governance of resources. {2.3.1.3.2, 2.3.4.1, 2.3.5.1}

There are limits to the adaptation capacity of socio-economic sectors under the influence of cryospheric change along with climate change (*high agreement, medium evidence*). Integrated (cross-sectoral) governance approaches hold potential in promoting socio-economic sectors' resilience and transformation, yet evidence on how these materialize to address cryosphere change in high mountain contexts remains low. Human habitability in mountain regions relies on multiple and diverse means to secure basic needs and sustain livelihood options, which are increasingly challenged by cryosphere change, induced by climate change. Recognition for and integration of indigenous knowledge and local knowledge promote resilience and adaptation in a changing climate and cryosphere environment. {2.3.6}

What are the Main Knowledge Gaps and Challenges?

Observations and projections of atmospheric conditions and associated cryospheric changes in high mountains areas, with a few exceptions, are fragmented and insufficient especially at higher elevations to adequately account for trends and spatial distribution of changes globally {2.4.1.1}. Except for past and future changes in high mountain glaciers, for which consolidated evidence is available at the global scale, there is an overall paucity of scientific studies reporting changes (past, attribution, future) in high mountain cryospheric components in a consistent and comprehensive manner globally {2.4.1.1}. Regarding impacts and adaptation responses in human systems, there is *limited evidence* of a comprehensive risk assessment approach to ascertain and systematically characterise and compare impacts across high mountain regions, particularly on risks that consider all underlying components of climate risk, including compounded risks and cascading impacts. {2.4.1.2}

2.1 Introduction

Mountain regions share common features, including rugged terrain, steep slopes, institutional and spatial remoteness that are linked to context-specific physical and social-ecological processes across vertical gradients. Due to their high elevation, mountains often feature cryosphere components, such as glaciers, seasonal snow cover and permafrost, with a significant influence on surrounding lowland areas even far from the mountains. Hence the mountain cryosphere plays a critical role in large parts of the world (Beniston, 2003). Due to the close relationship between mountains and the cryosphere, it was considered imperative to assess changes occurring in mountains and their effects in a dedicated chapter within this special report.

This chapter assesses recent and projected changes in glaciers, snow, permafrost and lake and river ice in high mountains areas, their drivers, as well as their impact on the different services provided by the cryosphere (Figure 2.1) with focus on literature published after the Fifth Assessment Report (AR5). Here high mountain areas include all mountain regions where glaciers, snow or permafrost are prominent features of the landscape, without a strict and quantitative demarcation, but with a focus on distinct regions (Figure 2.2). Many examples from specific localities in these regions are relevant to similar mountain areas. The emphasis is on changes over recent decades, rather than a perspective over a longer period, and future changes over the 21st century. Most mountain regions located in the polar regions are considered in Chapter 3. Mountain environments also change in response to climate change related effects on biodiversity or the physical environments unrelated to the cryosphere or socio-economic developments. These non-cryospheric drivers are not considered here, although unambiguous attribution can be difficult in some cases.

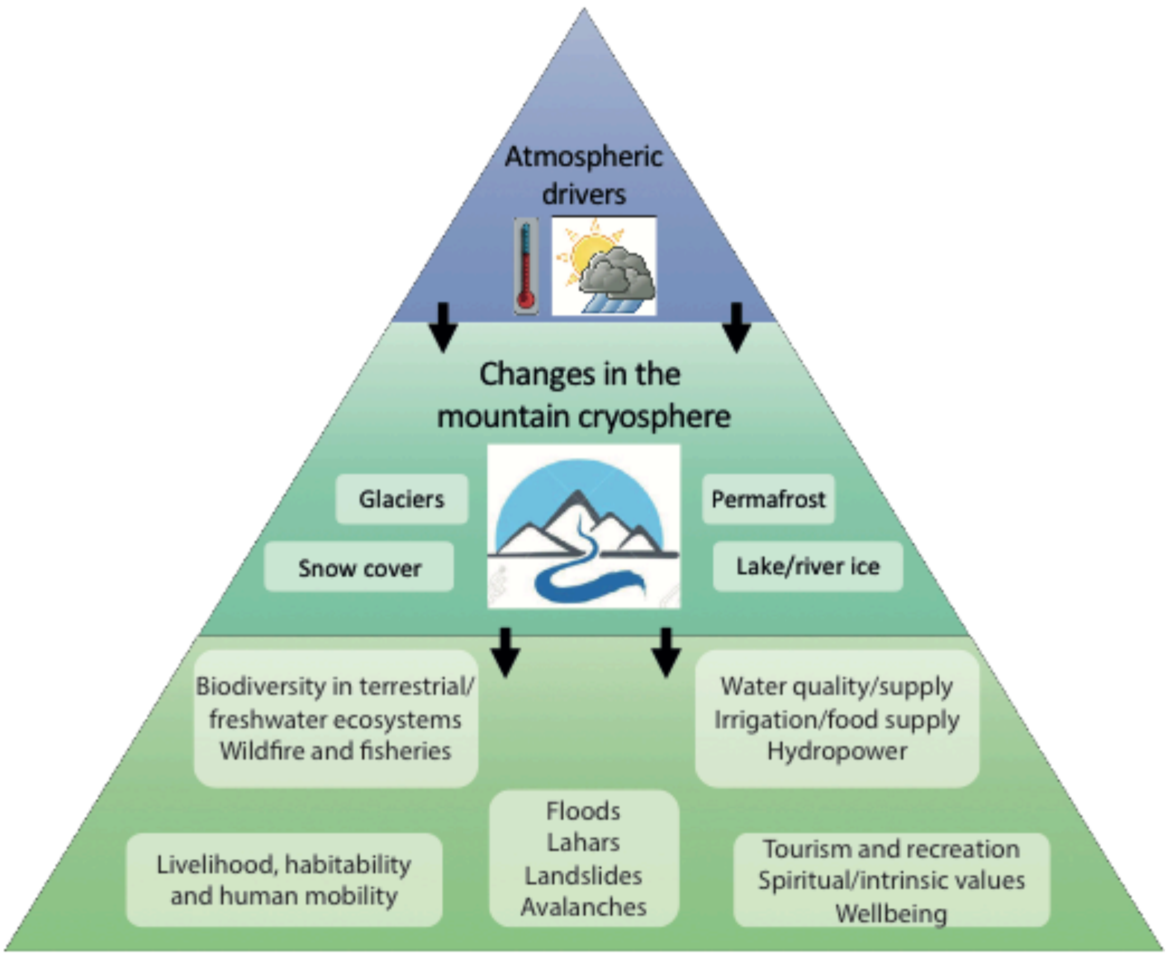


Figure 2.1: Cryospheric drivers and impacts discussed in this chapter.

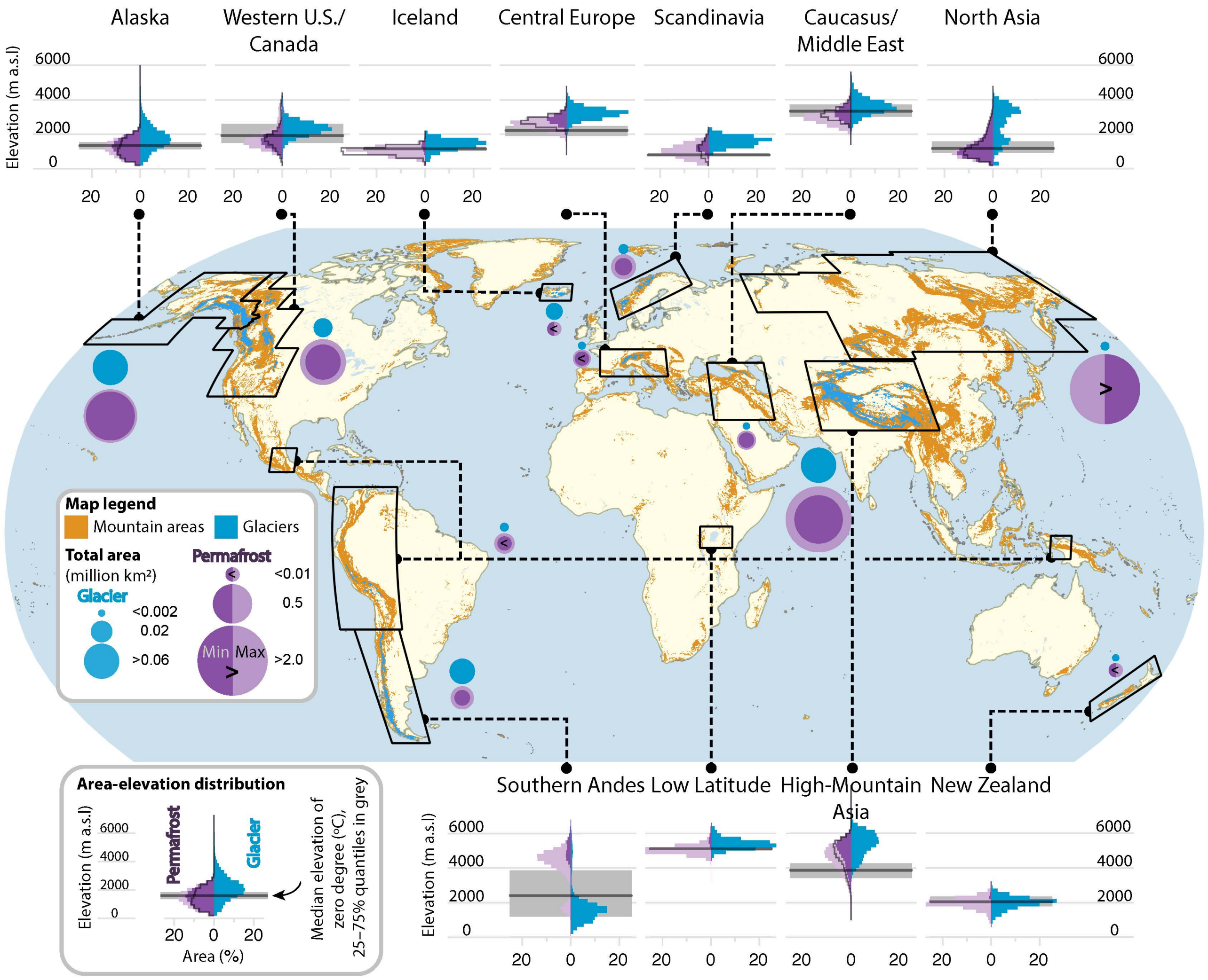


Figure 2.2: Distribution of mountain areas, glaciers and permafrost. Mountains are distinguished based on a ruggedness index (>3.5), a logarithmically scaled measure of relative relief (Gruber, 2012). Region outlines encompass 11 distinct regions with glaciers, largely matching the primary regions in the Randolph Glacier Inventory, RGI v6.0 (RGI Consortium, 2017), but cryosphere related impacts presented in this chapter are not strictly limited to these regions. Circles refer to regional glacier (RGI 6.0) and permafrost area (only mountain pixels within region boundaries). Permafrost area uncertainty is expressed as a minimum and maximum estimate (Gruber, 2012). Median ensemble of permafrst area obtained by Obu et al. (submitted) is shown for the northern hemisphere (black line). Histograms show glacier and permafrost area in 200 m elevation bins as a percentage of total regional glacier/permafrost area, respectively. Also shown is the median of the annual mean freezing level (0°C isotherm) calculated from the ERA-5 re-analysis of the European Centre for Medium Range Weather Forecasts over each region's mountain area for the period 2006 to 2015, with 25–75% quantiles in grey.

2.2 Changes in the Mountain Cryosphere

2.2.1 Atmospheric Drivers of Changes in the Mountain Cryosphere

2.2.1.1 Surface Air Temperature

Changes in surface air temperature in high mountain areas have been documented by in-situ observations (up to 5000 m a.s.l.) and regional reanalyses. Although the observation networks are often insufficiently dense (Lawrimore et al., 2011) or are not of high enough quality (Oyler et al., 2015), atmospheric warming in the 20th and early 21st century has been detected in most mountainous regions (*very high confidence*), often with a faster rate than global mean values, up to twice faster. In many regions warming trends are elevation dependent (*medium confidence*) (EDW working group, 2015; Qixiang et al., 2018). Evidence from regional-scale studies is provided in Appendix 2.A, Table 1.

Attribution studies specifically focused on other mountain regions are rare. Bonfils et al. (2008) and Dileepkumar et al. (2018) demonstrated that anthropogenic greenhouse emissions are the dominant factor in the recent temperature changes, partially offset by other human contributions (land use change and aerosol emissions for Western U.S.A and Western Himalaya, respectively). These findings are consistent with conclusions of AR5 regarding anthropogenic effects (Bindoff et al., 2013). It is thus *likely* (*limited evidence, high agreement*) that anthropogenic influence is the main contributor to surface temperature increases in high mountain areas since the mid-20th century.

Future changes are projected using global or regional models. However, there is currently no initiative, such as model inter-comparisons or coordinated model experiments, specifically addressing high mountain climate globally. The assessment of future warming rates is thus challenging, mostly due to the diversity of methodological approaches and limitations of most currently employed models in capturing mountain processes. Available evidence indicates that high mountain regions are projected to experience further increases in surface air temperature over the 21st century (*very high confidence*), consistent with global trends, but with regional differences and elevation dependent trends (*high confidence*) (Appendix 2.A, Table 1).

[START BOX 2.1 HERE]

Box 2.1: What is Elevation Dependent Warming?

The systematic and statistically significant change in atmospheric warming rate with elevation has been termed elevation-dependent warming (EDW) (EDW working group, 2015). The altitude dependency is not linear, and observational studies have often, but not always, shown amplification of warming rates with elevation, particularly in mid-latitudes and for minimum daily temperature (Diaz and Bradley, 1997; Liu and Chen, 2000; Appendix 2.A, Table 1). The amplification of warming can either increase with elevation, or be confined in a particular elevation band and decrease above it (Pepin and Lundquist, 2008; Ceppi et al., 2012; Scherrer et al., 2012). Pepin and Lundquist (2008) showed amplified warming around the 0°C isotherm, based on over 1000 high elevation stations across the globe, with a median trend (1948-2002) slightly over +0.25°C per decade, compared to +0.13°C per decade averaged over all sites. Ohmura (2012) demonstrated amplification of warming with increasing elevation in 65% of cases within 10 major mountain areas globally. Trends over the 1970-2011 period tended to be highest in the colder seasons (winter and spring) and at a higher elevation. Qixiang et al. (2018) showed both elevation and latitudinal amplification of warming rates over the period 1961-2010, reaching +0.4°C per decade in winter for stations above 500 m compared to +0.35°C per decade for stations below 500 m. For the end of the 21st century (between 1961-1990 and 2070-2099) for 13 mountain regions based on SRES scenarios range from +2.8°C (+0.25°C per decade) to +5.3°C (+0.48°C per decade) (Nogués-Bravo et al., 2007) and are expected to be more rapid in higher northern latitude mountains than elsewhere.

Several physical processes contribute to EDW, although quantifying their relative contributions has remained largely elusive. Most of the physical processes identified are similar to those explaining the observed increase in warming rate towards the polar regions. For example, the sensitivity of temperature to radiative forcing is increased at low temperatures common in polar and mountain environments (Ohmura, 2012). The relationship between specific humidity and downwelling longwave radiation is non-linear such that a given increase in specific humidity has a disproportionately large warming influence in the drier atmosphere common at high elevations and high latitudes (Rangwala et al., 2013; Chen et al., 2014b). The snow albedo feedback plays a role around the mean snowline elevation: increased surface air temperature drives a local decrease in snow cover duration, which in turn increases the absorption of solar radiation and leads to increased surface air temperature. Some processes are specific to the mountain environment. Increased latent heat release above the condensation level leads to a smaller temperature gradient in a warmer and moister atmosphere (Held and Soden, 2006). The cooling effect of aerosols, which also causes solar dimming, is most pronounced at low elevations and reduced in high elevation zones (Zeng et al., 2015). The deposition of light-absorbing particles on snow, in particular black carbon, can enhance warming rates via increased snow-albedo feedback (e.g., Ménégoz et al., 2014, see Box 2.2). Projections of EDW from global and regional climate models need to be considered carefully because of intrinsic limitations due to understanding and implementation of physical processes and often too coarse grid spacing for mountain regions (Ménégoz et al., 2014; Winter et al., 2017). EDW exhibits large regional differences because climate and synoptic conditions influence the partitioning between driving mechanisms.

[END BOX 2.1 HERE]

2.2.1.2 Solid and Liquid Precipitation

Past precipitation changes are less well quantified than for temperature and often more heterogeneous even within mountain regions (Hartmann and Andresky, 2013). Not only total precipitation amounts, but also the partitioning between liquid (rain) and solid (snow) components, exert strong forcing on the mountain cryosphere. Assessing changes in mountain precipitation is challenging because of observation scarcity and observation uncertainty (Nitu et al., accepted). Global and regional climate models have difficulties capturing mountain precipitation adequately, because of limitations due to (i) understanding and implementation of physical processes and (ii) too coarse grid spacing for mountain regions. Furthermore, regional precipitation patterns are characterized, and often dominated, by decadal variability (Mankin and Diffenbaugh, 2015) and shifts in large scale atmospheric circulation patterns (e.g., in Alaska, Winski et al., 2017). While most mountain regions do not exhibit trends in annual precipitation over the past decades (Appendix 2.A, Table 2, *medium confidence*), solid precipitation (snow) has significantly decreased due to higher temperatures, especially below the mean snowline elevation[3] (Appendix 2.A, Table 2, *high confidence*).

Future projections of precipitations indicate increases under all Representative Concentration Pathways (RCP), including the tropical Andes, the Hindu Kush Himalayas, East Asia, East Africa and the Carpathian region, and decreases in Mediterranean climate and the Southern Andes (*medium confidence,* Appendix 2.A, Table 2). Across the Himalayan-Tibetan Plateau mountains, the frequency and intensity of extreme rainfall events are projected to increase particularly during the summer monsoon season throughout the 21st century (Panday et al., 2015; Sanjay et al., 2017). This indicates a likely transition toward more episodic and intense monsoonal precipitation, especially in the easternmost part of the Himalayan chain (Palazzi et al., 2013). Projections of solid precipitation indicate a decrease, for all RCP scenarios, below the mean snowline elevation (*very high confidence*). Based on emerging literature (e.g., Kapnick and Delworth, 2013; O'Gorman, 2014) insignificant decreases, and sometimes increases, are projected for snowfall for locations significantly above the mean snowline elevation, where temperature increase is insufficient to affect the rain/snow partitioning, so that precipitation increase lead to snowfall increase (*medium confidence*).

2.2.1.3 Other Meteorological Variables

Atmospheric humidity, incoming shortwave and longwave radiation, and wind speed also influence the high-mountain cryosphere although detecting their changes and their effects on the cryosphere is even more challenging than for the dominating variables, surface air temperature and precipitation, both from an observation or modelling standpoint. Therefore, most simulation studies of cryospheric changes are mainly driven by temperature and precipitation.

Atmospheric moisture content affects latent and longwave heat fluxes with implications for the timing and rate of snow/ice ablation, and in some areas could be a significant driver of cryospheric change (Harpold and Brooks, 2018). Solar radiation affects snow/ice melt and short-lived climate forcings, such as sulphur and black carbon aerosols (You et al., 2013), influence the amount of solar radiation reaching the surface. Solar brightening caused by declining anthropogenic aerosols in Europe since the 1980s was shown to have only a minor effect on atmospheric warming at high elevation (Philipona, 2013), and effects on the cryosphere were not specifically discussed.

Wind controls preferential deposition of precipitation, post-depositional snow drift and affects melt rates of snow and glaciers through turbulent fluxes. Wind speed has decreased on the Tibetan Plateau between the

[3] FOOTNOTE : In this report, we refer to the mean snowline elevation as the average elevation of the transition between liquid and solid precipitation.

1970s and the early 2000s, and stabilized or increased slightly thereafter (Yang et al., 2014; Kuang and Jiao, 2016). This is consistent with existing evidence for a decrease in wind speed on mid-latitude continental areas since the mid-20th century (Hartmann et al., 2013). Except on the Tibetan Plateau, literature on past changes of wind patterns in mountain areas is very limited. A few studies have addressed regional projections of wind speed and direction, with little emphasis on mountain regions (e.g., Najac et al., 2011). Hanzer et al. (2018) computed 21st century changes of wind speed for a high mountain catchment in Austria, finding increases in winter (on average from 0 to 0.2 m s^{-1}) and decreases in summer (from 0 to 0.2 m s^{-1}), compared to 1971-2000, with changes increasing from RCP2.6 to RCP8.5 and from near future to end of 21st century.

2.2.2 Snow Cover

Snow on the ground is an essential and ubiquitous component of the mountain cryosphere. It affects mountain ecosystems, and plays a major role for mass movement and floods in the mountains. It plays a key role in nourishing mountain glaciers, and provides an insulating and reflective cover when present at their surface. It influences the thermal regime of the underlying ground, including permafrost, with implications for ecosystems. Climate change modifies key variables driving the onset and development of the snow cover (e.g., solid precipitation), and those responsible for its ablation and melt (e.g., air temperature, incoming radiation, deposition of light absorbing particles). Seasonal snow, especially in low-lying and mid-elevation areas of mountain regions, has long been identified to be particularly sensitive to climate change, generating concern from multiple stakeholders due to projected drastic reductions.

The mountain snow cover is characterized by a very strong interannual variability, including decadal climate variability, similarly to its main driving force solid precipitation (Lafaysse et al., 2014; Mankin and Diffenbaugh, 2015). Past changes must thus be assessed over sufficiently long time periods. Long-term in-situ records are scarce and sometimes missing in some regions of the world (Rohrer et al., 2013). Satellite remote sensing provide new capabilities in monitoring snow cover on regional scales, although the record length is often insufficient to assess climate trends. Below the mean snowline elevation, there is *high confidence* that the mountain snow cover has declined since the middle of the 20th century, with regional variations (Appendix 2.A, Table 3). Well above the mean snowline elevation, snow cover trends are generally insignificant or unknown (*medium confidence*).

Systematic attribution studies for the observed changes in seasonal snow in the mountain areas are limited (Pierce et al., 2008; Rupp et al., 2013). Nevertheless, combined analysis of snow and meteorological data indicates that most of the snow cover changes can be attributed to more precipitation falling as liquid precipitation (rain) below the mean snowline elevation and increases in melt rate at all elevations. These changes are mostly due to changes in atmospheric forcings, especially increased air temperature (Marty et al., 2017), which in turn are attributed to anthropogenic forcings at a larger scale (Section 2.2.2.1). Assessing the impact on snow changes of the deposition of short-lived climate forcers is an emerging issue (Skiles et al., in press, and references therein). This concerns light absorbing particles, in particular, which include deposited aerosols like black carbon, organic carbon and mineral dust, or microbial growth, although the role of the latter has not been quantified. Due to their seasonally variable deposition flux and impact, and mostly episodic nature for dust deposition (e.g., Kaspari et al., 2014; Di Mauro et al., 2015), light absorbing particles significantly contribute to inter-annual fluctuations of seasonal snow melt rate (*medium evidence, high agreement*). However, there is only *limited evidence* (*medium agreement*) that these particles have contributed to the observed snow changes since the mid-20th century. A fraction of snow cover decline could be due to anthropogenic increase in black carbon deposition, as identified in High Mountain Asia (*limited evidence, medium agreement*). Increases in dust deposition causing accelerated snowmelt are predominant over the effect of black carbon decrease in most other regions studied (Europe including Iceland, Western North America) (*limited evidence, medium agreement*).

Projected changes of the mountain snow cover are studied based on available model experiments, either directly from GCM or RCM output, or following downscaling and the use of impact models, such as snowpack models. These projections generally do not specifically account for future changes in the deposition rate of light absorbing particles on snow (or, if so, simple approaches have been used hitherto, e.g., Deems et al., 2013), so that future changes in snow conditions are mostly driven by changes in meteorological drivers assessed in Section 2.2.1. Climate projections of future changes of mountain snow conditions generally follow projected changes in air temperature, modulated by concurrent changes in precipitation (example for Europe and High Mountain Asia (Himalaya and Hindu-Kush Karakoram), Figure 2.3 and Appendix 2.A, Table 3). Below the mean snowline elevation, the snow cover (depth or mass) is projected to decline by 25% [10 - 40%], between the recent past period (1986-2005) and the near future (2031-2050), regardless of the RCP climate scenario. By the end of the century (2081-2100), reductions of up to 80% [50 - 90%] are expected under RCP8.5, 50% [30 - 70 %] under RCP4.5 and 30% [10 – 40 %] under RCP2.6. At higher elevations, well above the mean snowline elevation, projected reductions are smaller (*high confidence*), mostly because temperature increases affect mostly the ablation component of snow cover evolution, and not the onset and accumulation component. The projected increase in snow accumulation may exceed the projected increase in melt, resulting in net increase in snow mass and duration (*medium confidence*). All elevation levels and regions of the world are projected to exhibit sustained inter-annual variability of snow conditions throughout the 21st century (*high confidence*).

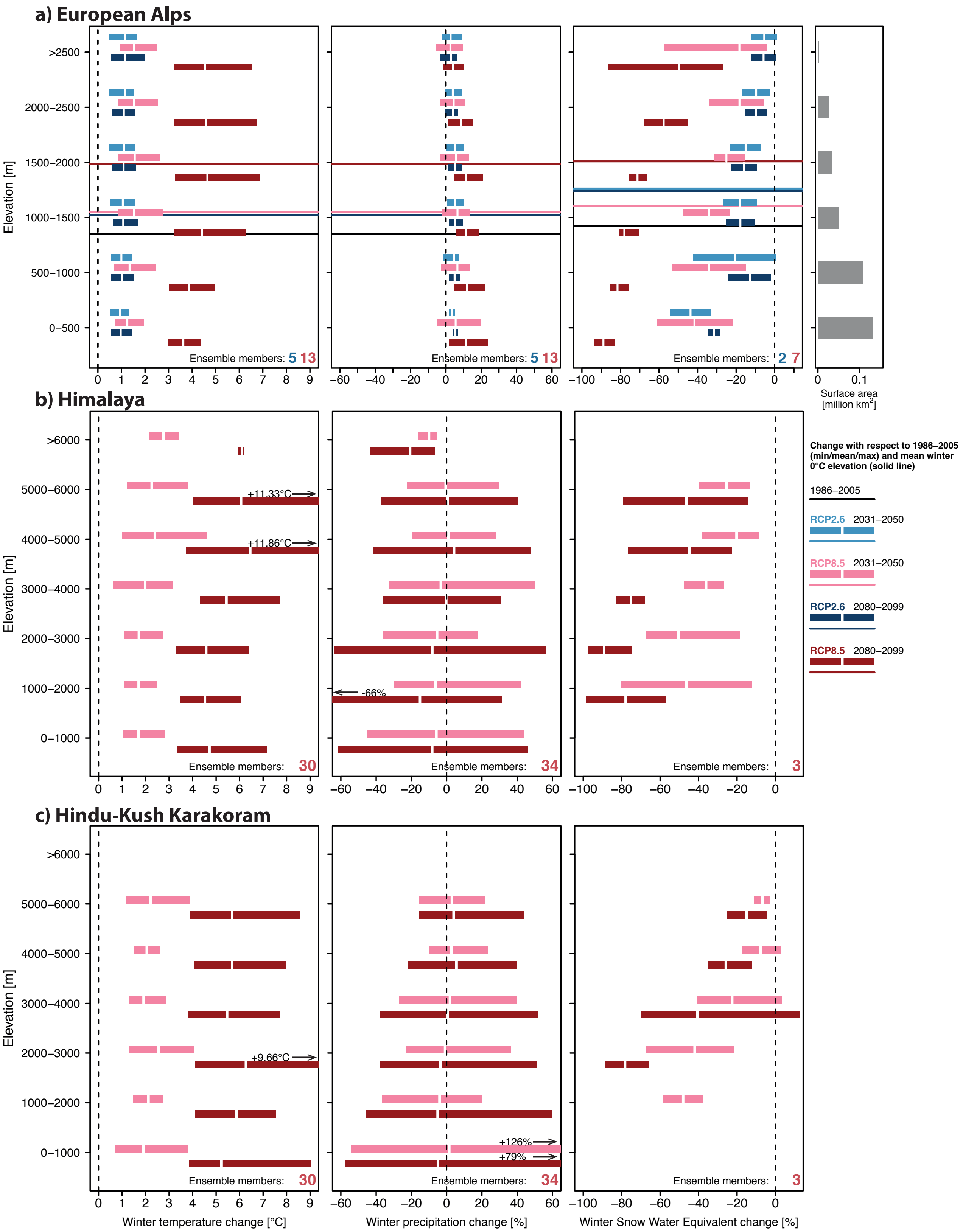


Figure 2.3 Projected elevation-dependent change (1986-2005 to 2031-2050 and 2080-2099) of mean winter (December-May) air temperature, precipitation and snow water equivalent, averaged over 500 m elevation classes in the European Alps, Hindu-Kush Karakoram and Himalaya. For Europe, changes were derived for RCP2.6 and RCP8.5 from EURO-CORDEX high-resolution (12 km) regional climate model output (Jacob et al., 2014). The numbers to the upper right of each panel reflect the number of available simulations (note that not all models provide snow water equivalent and the ensemble size is smaller for this variable). Horizontal lines: Mean elevation of the winter 0°C line as obtained by regressing climatological mean winter temperatures for the respective periods and scenarios onto grid cell elevation. Note that the winter 0°C line differs depending on the ensemble considered, with less ensemble members for which snow water equivalent data are available. For 1986-2005 time period, the same ensemble composition as RCP8.5

was used, for each panel. Rightmost panel: Area-elevation distribution of the Alpine analysis domain in one the regional climate models used (COSMO-CLM). For Hindu-Kush Karakoram and Himalaya, changes were derived for RCP8.5 from CMIP5 global climate output (Terzago et al., 2014).

2.2.3 Glaciers

The high mountain areas considered in this chapter (Figure 2.2) include ~173,000 glaciers covering an area of 252,000 km^2 (RGI Consortium, 2017), roughly 30% of the total global glacier area (not including the two ice sheets of Greenland and Antarctica). These glaciers span an altitude range from sea-level to >8000 m a.s.l. and occupy diverse climatic regions. Their mass budget is determined largely by the balance between snow accumulation and melt at the glacier surface, driven primarily by atmospheric conditions. Glaciers respond to imbalances in their mass budget by adjusting their volume, size and shape over time scales of decades or more, and are thus a robust medium-term indicator of climate change. Rapid changes of mountain glaciers have multiple impacts for social-ecological systems, affecting not only bio-physical properties such as runoff volume and sediment fluxes in glacier-fed rivers, and glacier-related hazards but also ecosystems and human livelihoods, socio-economic activities and sectors such as agriculture and tourism as well as other intrinsic assets such as cultural values.

2.2.3.1 Observed Changes and Attribution

Satellite and in-situ observations of changes in glacier area, length and mass show a globally coherent picture of continued mountain glacier recession in the last three decades (*very high confidence*) with only few exceptions (Zemp et al., 2015). The global trend is significant despite considerable inter-annual and regional variations (Medwedeff and Roe, 2017). Departures from this trend occurred in some regions, but were generally locally restricted (Section 2.2.3.2). Since AR5, several new estimates of global-scale glacier mass budgets have emerged indicating increasingly negative glacier mass budgets in most mountain regions over at least the last three decades (*high confidence*), but annual variability and regional differences are large. Based on ~20,000 geodetic glacier observations, Zemp et al. (submitted) found an increase in mass loss over all mountain regions from 470 ± 80 kg m^{-2} yr^{-1} in the period 1986-2005 to 610 ± 90 kg m^{-2} yr^{-1} in the period 2006-2015, which is considerably more than their estimated global average (i.e. also including the glaciers other than the ice sheets in the polar regions not covered in this chapter). During the latter period mass budgets were most negative in the Southern Andes (-1200 kg m^{-2} yr^{-1}) and least negative in High Mountain Asia (-190 kg m^{-2} yr^{-1}, Figure 2.4). However, due to large ice extent, the total mass loss and corresponding sea-level equivalent is largest in Alaska, followed by the Southern Andes and High Mountain Asia. All mountain regions combined contributed to sea-level at a rate of 0.41 mm yr^{-1} during 2006-2015 (See Chapter 4).

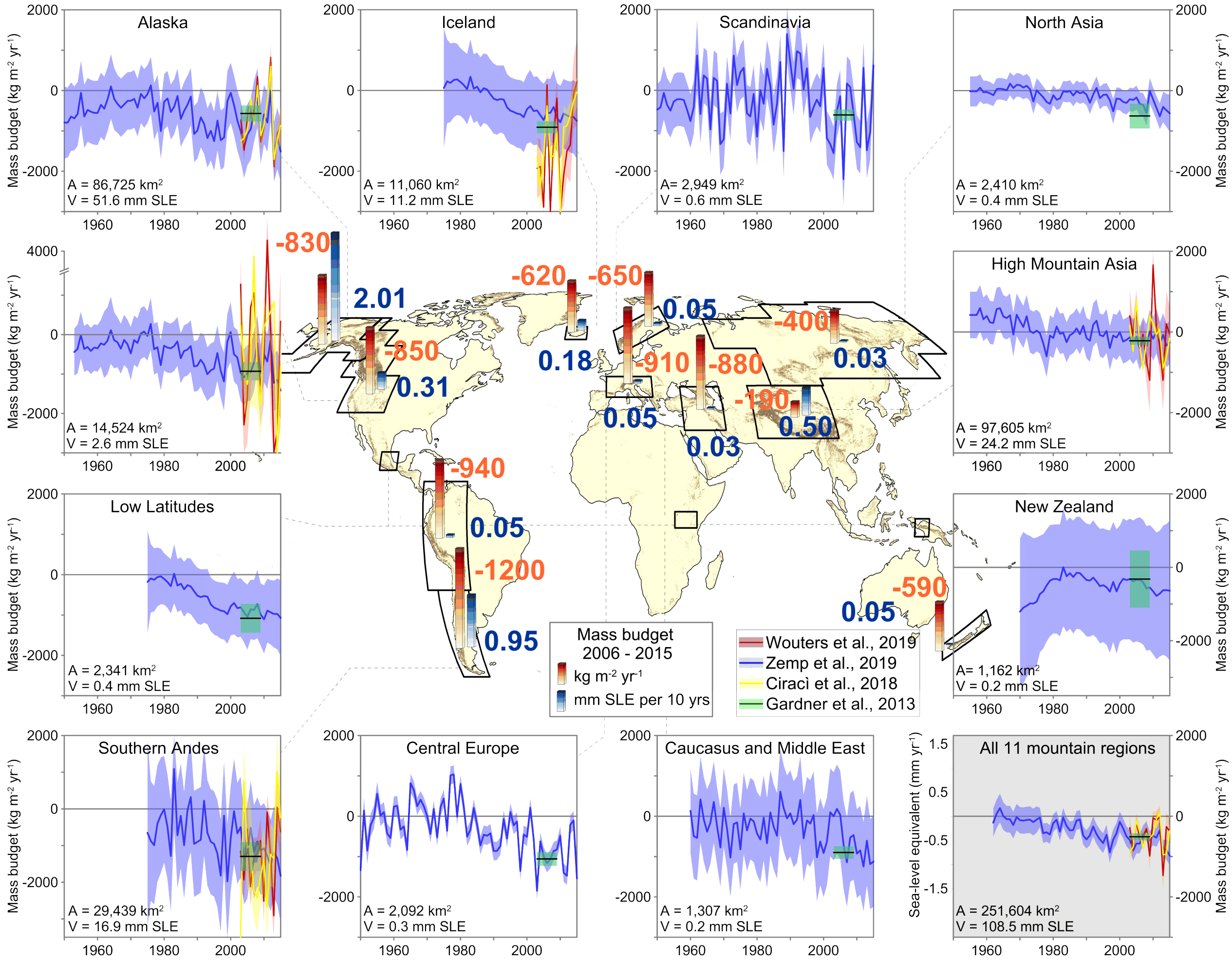


Figure 2.4: Annual glacier mass budgets in units of kg m^{-2} yr^{-1} for 11 glacierized mountain regions (Figure 2.2) and all regions combined. Shading refers to the random error of the regional mass change. Estimates by Zemp et al. (submitted) are based on extrapolation of glaciological and geodetic balances. Estimates by Ciracì et al. (submitted) and Wouters (submitted) are from the Gravity Recovery and Climate Experiment (GRACE) and only shown for the regions with glacier area > 3,000 km^2). Estimates by Gardner et al. (2013) were used in AR5. Glacier areas (A) and volumes (V) are based on (RGI Consortium, 2017) and Huss and Farinotti (2012) updated to the glacier outlines of RGI 6.0, respectively. Red and blue bars and associated numbers refer to regional budgets for the period 2006-2015 in units of kg m^{-2} yr^{-1} and mm sea-level equivalent (SLE) per decade, respectively, using the data by Zemp et al. (submitted).

It is *virtually certain* that global glacier recession during the last 100 years is primarily due to global atmospheric warming (e.g., Marzeion et al., 2014). It was estimated that the anthropogenic fraction of mass loss of all glaciers outside Greenland and Antarctica increased from 25 ± 35% during 1851–2010 to 69 ± 24% during 1991–2010 (Marzeion et al., 2015). Other factors, such as changes in meteorological variables other than air temperature or internal glacier dynamics, have modified the temperature-induced glacier response in some regions. For example, glacier mass loss over the last seven decades in the European Alps was exacerbated by increasing long-wave irradiance and latent heat due to enhanced humidity (Thibert et al., 2018). In the Tien Shan changes in atmospheric circulation in the north Atlantic and north Pacific in the 1970s resulted in an abrupt reduction in precipitation and thus snow accumulation, amplifying temperature-induced glacier mass loss (Farinotti et al., 2015). Deposition of light absorbing particles, growth of algae and bacteria and local amplification phenomena such as the enhancement of particles concentration due to surface snow and ice melt, and cryoconite holes, have been shown to enhance ice melt (e.g., Ginot et al., 2014; Zhang et al., 2017) but there is *limited evidence* and *low agreement* that long-term changes in glacier mass are linked to these processes (Painter et al., 2013; Sigl et al., 2018). Rapid retreat of the calving outlet glaciers in Patagonia was attributed to changes in glacier dynamics (Sakakibara and Sugiyama, 2014).

In Alaska 36 marine-terminating glaciers exhibited a complex pattern of periods of significant retreat and advance during 1948-2012, highly variable in time and lacking coherent regional behaviour (McNabb and Hock, 2014). These fluctuations can be explained by internal retreat-advance cycles typical of tidewater glaciers that are largely independent of climate (Brinkerhoff et al., 2017). Irregular and spatially inconsistent glacier advances in the Karakoram have been associated with surge-type flow instabilities (Bhambri et al., 2017). In contrast, larger-scale balanced or slightly positive glacier mass budgets since at least the 1970s over the Karakoram as well as the western Tibetan Plateau, the West Kunlun Shan, and Pamir mountains (Gardelle et al., 2013; Kääb et al., 2015; Azam et al., 2018) have been related to specific meteorological mechanisms countering the effects of atmospheric warming (Kapnick et al., 2014; Sakai and Fujita, 2017), and locally to an increase in irrigation intensity which in turn affects the regional climate (de Kok et al., 2018). Glacier advances in Norway in the 1990s and in New Zealand between 1983 and 2008 have been attributed to local increases in snow precipitation (Andreassen et al., 2005) and lower air temperatures (Mackintosh et al., 2017), respectively, caused by changes in atmospheric circulation. Glacier advances have also been observed in response to volcanic activity (Barr et al., 2018), and due to supraglacial deposition of mining waste (Jamieson et al., 2015).

2.2.3.2 Projections

It is *very likely* that in all mountain regions glaciers will lose substantial mass by the end of the century. Due to the pronounced current mass imbalance, many glaciers are expected to further recede to adjust their geometry to current climate conditions, even if the current climate were to remain constant (*very high confidence*) (Mernild et al., 2013; Marzeion et al., 2018). Global-scale glacier projections from six glacier models forced each by 8 to 21 Global Circulation Models (GCMs) indicate mass losses relative to 2015 of 29±7% (RCP2.6) to 47±10% (RCP8.5) by 2100 for the 11 mountain regions but relative mass reductions vary greatly between regions (Hock et al., submitted). Projected mass losses tend to be largest (>80%) in regions dominated by smaller glaciers and relatively little current ice cover (e.g., Central Europe, Caucasus, Low Latitudes, North Asia, Scandinavia).

These global-scale projections are consistent with results from local or regional-scale studies. For example, Kraaijenbrink et al. (2017) projected mass losses for all glaciers in High Mountain Asia of 64 ± 5% (RCP8.5) by the end of the century (2071-2100) compared to 1996-2015, and 36±7% for a 1.5 ºC increase in global air temperature. A high-resolution regional glaciation model including ice dynamics and surface mass balance indicated that by 2100, the volume of glacier ice in western Canada will shrink by 70±10% relative to 2005 with the maximum rate of ice volume loss, predicted to occur in 2020–2040 (Clarke et al., 2015). Trüssel et al. (2015) projected almost complete loss of >340 km^2 low-lying Yakutat Glacier in Alaska by 2070 under the RCP6.0 emission scenario, and by 2110 when the current climate was assumed stable.

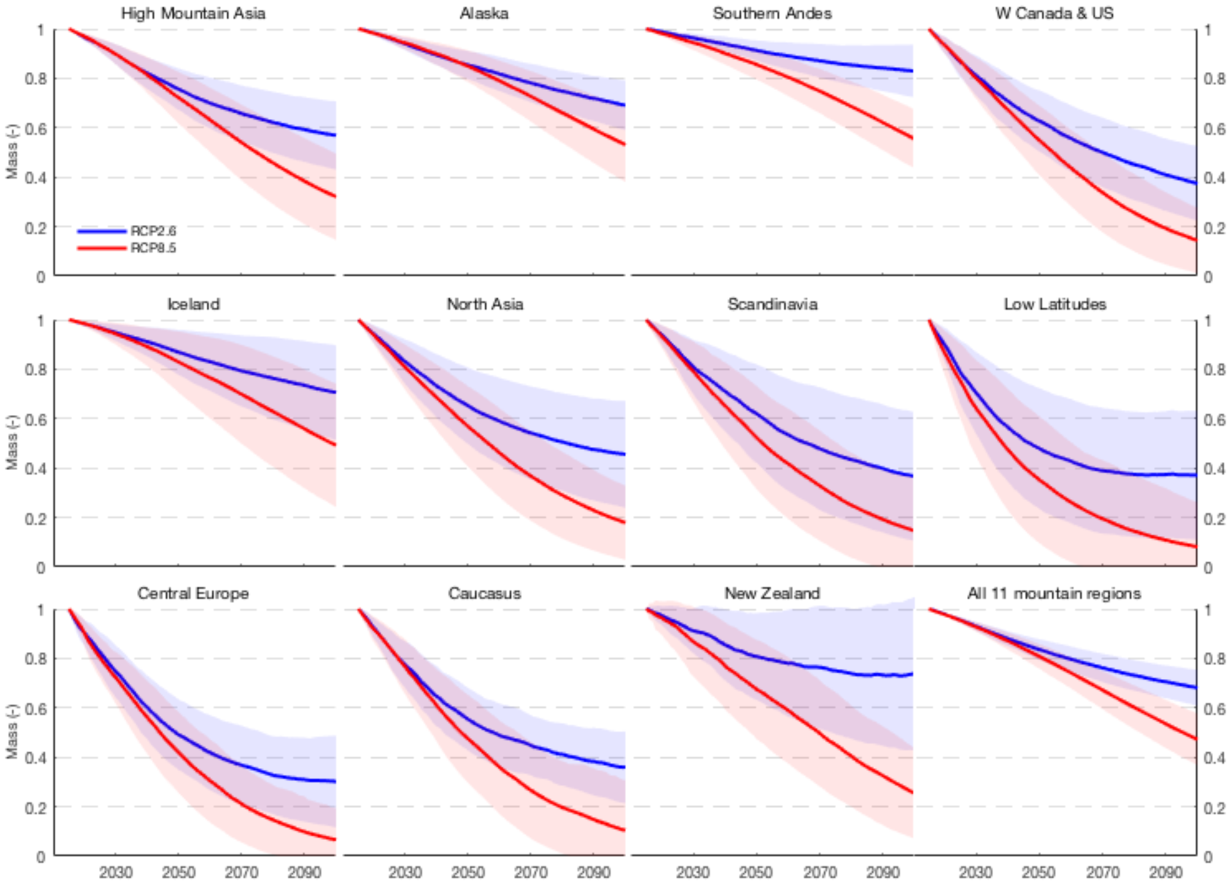


Figure 2.5 Projected glacier mass evolution between 2015 and 2100 relative to each region's glacier mass in 2015 for emission scenarios RCP2.6 and RCP8.5. Lines and shading refer the arithmetic mean ± standard deviation of 46 (RCP2.6) and 88 (RCP8.5) individual model runs from four to six glacier models forced each by data from five to 21 General Circulation Models (GCMs). Regions are sorted according to decreasing glacier area. Data are from Hock et al. (submitted) and references therein.

2.2.4 Permafrost

Overall, data on permafrost in mountains are scarce (Tables 2.1 and 2.2, PERMOS, 2016; Bolch et al., 2018) and measurement sites unevenly distributed globally and within mountain regions. Permafrost, unlike glaciers and snow, is a subsurface phenomenon that cannot easily be observed remotely. As a consequence, the understanding of its distribution and change is less developed and quantitative than for glaciers or snow. Permafrost change in many mountain areas can only be inferred indirectly (Gruber et al., 2017).

High-mountain areas as shown in Figure 2.2 comprise approximately 45% (6–8 million km^2) of the global permafrost area. The distribution of permafrost in mountains is heterogeneous as shown in detailed regional simulation studies in high-mountain regions globally (Boeckli et al., 2012; Bonnaventure et al., 2012; Westermann et al., 2015; Azócar et al., 2017; Zou et al., 2017) and a new map for the northern hemisphere (Obu et al., submitted).

Permafrost observations in the European Alps, Scandinavia, and the Tibetan Plateau show warming (Table 2.1, Figure 2.5) and degradation of permafrost at individual sites (e.g., Phillips et al., 2009) during the past two decades. Bedrock warms faster than debris or soil and permafrost close to 0ºC often warms at a lower rate than colder permafrost because ground-ice melt slows warming. Several European bedrock sites (Table 2.1) have rates of permafrost warming exceeding the rate of about 0.6°C warming inferred for the 20th century based on thermal gradients at depth in an ensemble of European bedrock sites (Harris et al., 2003; Gruber et al., 2004).

Observed active-layer thickness increased in mountains of the European Alps, Scandinavia, and the Tibetan Plateau during the past two decades (Table 2.2) and, in the European Alps, observed inter-annual variability is high (PERMOS, 2016). Electrical-resistivity monitoring in the European Alps during approximately the past 15 years revealed increasing subsurface liquid water content (Hilbich et al., 2008; Bodin et al., 2009; PERMOS, 2016), likely indicating gradual ground-ice loss. Rock glacier velocities observed in the European Alps in the 1990s were on the order of a few decimetres per year and during approximately the past 15 years they often were about 2–10 times higher (Bodin et al., 2009; Lugon and Stoffel, 2010; PERMOS, 2016) and destabilisation has been documented (Delaloye et al., 2010; Buchli et al., 2013; Bodin et al., 2016). One particularly long time series shows velocities around 1960 just slightly lower than during recent years (Hartl et al., 2016). The majority of similar landforms investigated in the Alaska Brooks Range accelerated since the 1950s, while few others slowed down (Darrow et al., 2016).

Decadal-scale permafrost warming and thaw are largely driven by air temperature and moderated by snow and soil conditions (Wu and Zhang, 2008). Periods of cooling, one or a few years long, have been observed and attributed to extraordinary snow conditions (PERMOS, 2016). Snowpack thinning and changed timing may result in temporary and localised ground cooling, even in a warming atmosphere (cf. Zhang, 2005). Extreme increases of active-layer thickness often correspond with summer heat waves (PERMOS, 2016) and permafrost degradation can be accelerated by water percolation (Luethi et al., 2017). The attribution of differences in warming rates to regional climate or local characteristics is difficult with the few long-term observations available. Similarity and synchronicity of inter-annual to decadal velocity changes of rock glaciers within the European Alps (Bodin et al., 2009; Delaloye et al., 2010) and the Tien Shan (Sorg et al., 2015), suggests common regional forcing such as summer air temperature or snow cover.

Scenario simulations for the Tibet Plateau until 2100 estimate permafrost area to be strongly reduced, for example by 20–65% for RCP2.6 and RCP8.5 and a spatial resolution of $0.5^o \cdot 0.5^o$ (Lu et al., 2017). Such coarse-scale studies (Guo et al., 2012; Slater and Lawrence, 2013; Guo and Wang, 2016), however, are of limited use in quantifying changes and informing impact studies in steep terrain due to inadequate representation of topography (Gruber, 2012). Fine-scale simulations, on the other hand, are local or regional, limited in areal extent and differ widely in their representation of climate change and frozen ground. They reveal regional and elevational differences of warming and degradation (Bonnaventure and Lewkowicz, 2011; Hipp et al., 2012; Farbrot et al., 2013) as well as warming rates differing between locations (Marmy et al., 2016) and seasons (Marmy et al., 2013). While model differences preclude a quantitative summary, these studies agree on increasing warming and thaw of permafrost for the 21st century. This is also expected when using air temperature increase as a simple proxy of 21st century permafrost loss globally (Chadburn et al., 2017). In mountains, permafrost thaw at depth can be accelerated by lateral warming (Noetzli and Gruber, 2009) and deep percolation of water (Hasler et al., 2011). Some peaks in the European Alps may lose permafrost completely by the end of the century (Magnin et al., 2017).

In summary, permafrost in the European Alps, Scandinavia and the Tibetan Plateau has warmed during the 21st century and some observations reveal ground-ice loss and degradation (*high confidence*). During the 21st century, rates of permafrost warming in the European Alps and Scandinavia as well as rock-glacier acceleration in the European Alps exceeded values of the late 20th century (*limited evidence, high agreement,*). Observed decadal-scale permafrost changes are attributed to atmospheric warming (*very high confidence*). Lacking data precludes direct assessment of permafrost change outside the European Alps, Scandinavia, and the Tibet Plateau. Permafrost is expected to undergo increasing thaw and degradation during the 21st century (*high confidence*), although structural differences in available simulations preclude a quantitative summary. Permafrost thaw and degradation impact people via hazards (Section 2.3.2), runoff and water quality (Section 2.3.1), and greenhouse gas emissions (Section 2.2.6).

Table 2.1: Observed changes of permafrost mean-annual ground-temperature (MAGT) in mountain regions. Values are based on individual boreholes or ensembles of several boreholes reported. Underscored temperatures are averages during observation period. The MAGT refers to the last year in a period and is taken from a depth of 10–20 m unless the borehole is shallower. Region names refer to Figure 2.2.

Elevation [m asl]	Surface Type	Period	MAGT [ºC]	MAGT trend [ºC per decade]	Reference
Global					

>1000	various (43)	2006–2017		0.2 ± 0.05	(Biskaborn et al., 2018)
Central Europe					
2500–3000	debris or coarse blocks (>10)	1987–2005	> –3	0.0–0.2	(PERMOS, 2016) (Noetzli et al., 2018)
		2006–2017		0.0–0.6	
3500–4000	bedrock (4)	2008–2017	>–5.5	0.0–1.0	(Pogliotti et al., 2015) (Magnin et al., 2015) (Noetzli et al., 2018)
Scandinavia					
1500–1800		1999–2009	>–3	0.1–0.9	(Isaksen et al., 2011) (Christiansen et al., 2010)
High-mountain Asia (Tibetan Plateau)					
		2005–2016		0.3	(Noetzli et al., 2018)
3500	meadow	2003–2011	–1.2	0.38	(Liu et al., 2015)
~4650	meadow (6)	2002–2012	–1.52 to –0.41	0.08 to 0.24	(Wu et al., 2015)
~4650	steppe (3)	2002–2012	–0.79 to –0.17	0.09 to 0.18	
~4650	bare soil (1)	2003–2012	–0.22	0.15	
4500–5000	unknown (6)	2002–2011	–1.5 to –0.16	0.08 to 0.24	(Peng et al., 2015)

Table 2.2: Observed changes of active-layer thickness (ALT) in mountains. Numbers in brackets indicate how many sites are summarised for a particular surface type and area. Region names refer to Figure 2.2.

Elevation [m asl]	Surface Type	Period	ALT in last year [m]	ALT trend [cm per decade]	Reference
Scandinavia					
353–507	peatland (9)	1978–2006	~0.65–0.85	7–13	(Åkerman and Johansson, 2008)
		1997–2006		13–20	
Central Europe					
2500–2910	bedrock (4)	2000–2014	4.2–5.2	10–100	(PERMOS, 2016)
High-mountain Asia (Tibetan Plateau)					
4629–4665	meadow (6)	2002–2012	2.11–2.32	34.8–45.7	(Wu et al., 2015)
4638–4645	steppe (3)	2002–2012	2.54–3.03	39.6–67.2	
4635	bare soil (1)	2002–2012	3.38	18.9	
4848	meadow	2006–2014	1.92–2.72	15.2–54	(Lin et al., 2016)
3500	meadow	1992–2011	1.70	19	(Liu et al., 2015)

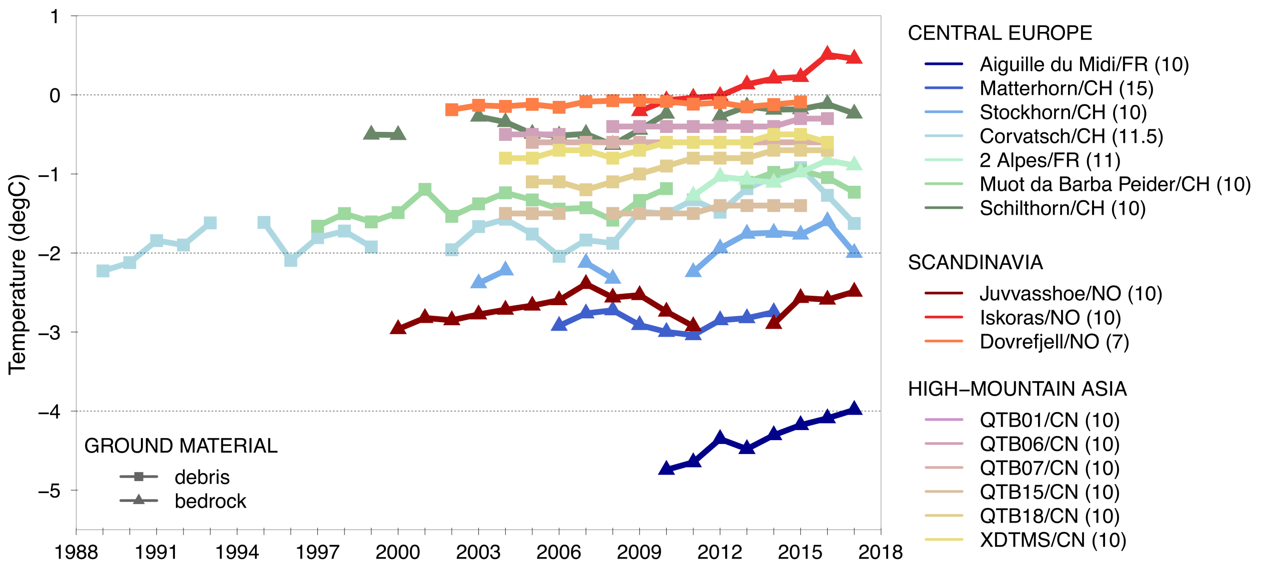


Figure 2.6: Mean annual ground temperature of permafrost boreholes in Central Europe, Scandinavia and High-Mountain Asia. The depth of measurements reported is chosen to be approximately 10 m and shown in brackets, site name and country are indicated in legend. Redrawn based on data reported in Noetzli et al. (2018).

2.2.5 Lake and River Ice

Observations of extent, timing, duration and thickness of lake and river ice rely mostly on in-situ measurements (e.g., Sharma et al., submitted), and, increasingly on remote sensing (Duguay et al., 2014) but studies focusing specifically on mountain regions are scarce. Some studies include mountain lakes as part of larger-scale investigations. For example, using microwave remote sensing, Du et al. (2017) found shorter ice cover duration for 43 out of 71 examined lakes >50 km^2 including lakes on the Tibetan Plateau regions between 2002 and 2015, but regional trends were significant ($p < 0.05$) in only five lakes, due to large interannual variability. A >570 year-long observational series for Suwa lake in Japan, indicated a trend towards earlier ice freeze date that increased from 0.19 days per decade in the period 1443–1683 to 4.6 days per decade in the period 1923–2014 (Sharma et al., 2016). Observations of a subalpine lake in Austria over the period 1921-2015 indicated later freeze dates (17.4 days per century), a significant decrease in full lake ice cover, increased interannual variability in ice cover and an increase in surface water temperature (~1–2 °C) in spring and summer and less (~0.3°C) (Kainz et al., 2017). A global analysis of in situ and satellite-derived data showed that lake summer surface water temperatures between 1985 and 2009 increased on average by 0.34°C per decade) but trends varied largely even within the same regions (O'Reilly et al., 2015).

There is *high confidence* that air temperature and solar radiation are the most important drivers to explain observed changes of lake ice dynamics (Sharma et al., submitted). However, lake characteristics such as salinity, lake morphometry and wind exposure or the amount and characteristic of inflowing water especially for lakes fed by glaciers (Kropácek et al., 2013; Song et al., 2014; Yao et al., 2016; Gou et al., 2017) explain the high variability in lake ice cover dynamics in mountain regions.

Despite high spatial and temporal variability in lake and river ice cover dynamics mountains lakes are expected to continue to experience later freezing, earlier break-up, and shorter ice cover in the future in response to further air temperature increases (Gebre et al., 2014; Du et al., 2017)). More lakes are expected to lose their seasonal ice-cover and turn to open-water systems throughout the year (Sharma et al., submitted).

River ice can trigger ice jamming and flooding, but information regarding past and future changes in frequency and character of river ice jams are lacking for the mountain regions.

Overall, there is only limited evidence on changes in lake and river ice specifically in the mountains, indicating a trend, but not universally, towards shorter lake ice cover duration consistent with increased water temperatures.

[START BOX 2.2 HERE]

Box 2.2: Local, Regional and Global Climate Feedbacks Involving the Mountain Cryosphere

The cryosphere interacts with the environment and contributes to several climate feedbacks. Particularly prominent among the climate-relevant feedbacks are ones associated with snow cover. The presence or absence of snow on the ground drives profound changes to the energy budget of land surfaces, hence influences the physical state of the overlying atmosphere. The reduction of snow on the ground, potentially amplified by aerosol deposition and modulated by interactions with the vegetation, contributes to the snow albedo feedback, which is a powerful feedback loop directly acting on elevation-dependent warming (Box 2.1). In mountain regions, this feedback mostly operates at the local scale and is seasonally variable (most visible effects on the fringes of the snow season).

Regional climate feedbacks involving the high mountain cryosphere, and in particular the snow albedo feedback, have only been detected in large mountain regions such as the Himalayas, using global and regional climate models. Feedbacks associated to light absorbing particles deposition and enhanced snow albedo feedback were shown to induce surface air warming (locally up to 2°C) (Ménégoz et al., 2014) with accelerated snow cover reduction (Ji, 2016; Xu et al., 2016), and were also suggested to influence the Asian monsoon system with associated precipitation increase (Yasunari et al., 2015). However, many of these studies focussed on so-called rapid adjustments, rather than longer-timescale, further fetching feedbacks, because they used regional or global models constrained by large scale synoptic fields. Large scale climate feedbacks involving specifically the high mountain cryosphere remain largely unexplored. Although mountain regions play a role for the global climate, because of their topography (e.g., Naiman et al., 2017), there is no evidence for global climate feedbacks specifically involving the mountain cryosphere, largely because of the limited spatial extent of changes of the mountain cryosphere.

High mountain regions include permafrost that is rich in soil organic carbon (Dymov et al., 2015; Fuchs et al., 2015; Wu et al., 2017). From 1986-2000 on the Qinghai–Tibet Plateau, land cover change in permafrost regions led to an estimated 120 Mg C/yr loss of soil organic carbon (Genxu et al., 2008). Water from thawing permafrost on mountain hillslopes drains, leading to soil aeration and vulnerability of soil carbon (Dymov et al., 2015). Emissions of carbon dioxide (CO_2) and methane (CH_4) has and is projected to continue to occur and varies depending on the extent of regional warming and the temperature sensitivity of the soil organic carbon, which is often high in mountain permafrost soils (Mu et al., 2016; Mu et al., 2017; Sun et al., 2018a). The degradation of permafrost in peatland soils that collapse following thaw is projected to lead to the most rapid loss of high elevation soil carbon (Mu et al., 2016; Mamet et al., 2017). Though warming and nitrogen fertilization increased CH_4 uptake in a wetland permafrost ecosystem, the high soil organic carbon pool and temperature sensitivity indicate potential for a significant positive feedback to climate change (Chen et al., 2017).

[END BOX 2.2 HERE]

2.3 Change in Mountain Ecosystems, their Services, Managed System and Human Responses

2.3.1 Water

Freshwater is probably the most important ecosystem service provided by the mountain cryosphere. The runoff per unit area generated in mountains is on average approximately twice as high as in lowlands (Viviroli et al., 2011) making mountains a significant source of water supporting livelihoods in and far beyond the mountain ranges themselves. The presence of snow, glaciers, and permafrost can exert a strong control on the amount, timing and biogeochemical properties of runoff (FAQ2.1). Changes to the cryosphere due to climate change may alter this provision of fresh water (*high confidence*) with direct consequences for both the upstream and the downstream populations (Barnett et al., 2005; Beniston, 2005; Munia et al., 2016).

2.3.1.1 Changes in River Runoff

Both increases and decreases in annual runoff in recent decades have been observed in glacier- and snowmelt-dominated basins. Many of these trends are attributed to changes in meltwater from ice and/or snow. Typically glacier retreat first leads to an increase in glacier runoff, until a turning point, often called "peak water" is reached, upon which runoff declines (FAQ2.1 and Figure 2.7). Positive trends in annual runoff have been observed in many glacier-dominated river basins, for example, the European Alps (Bard et al., 2015), High Mountain Asia (Chen et al., 2016) and Alaska (O'Neel et al., 2014) and attributed to increased meltwater from glaciers. Some rivers in North America have experienced a decrease in annual runoff for 1998-2011 which may be due to an increase in evapotranspiration associated with a longer growing season with less snow cover, despite the increase in peak runoff due to increased melt water (Brahney et al., 2017). Simulated glacier meltwater in La Paz glaciers in Bolivia showed a positive trend for 1998-2011 with recent warming but basin runoff showed no trend due to decreasing precipitation (Soruco et al., 2015). These increases or decreases in observed runoff imply changes in frequency and intensity of flood and drought, but evidence is limited except for glacier lake outburst floods (Section 2.3.2). A consistent shift to earlier runoff peak in recent decades has been observed in many basins influenced by snowmelt and glacier melt (Barnhart et al., 2016).

Projections show longer hydrological droughts in large parts of the snow-dominated climate regardless of RCP emission scenario, with increases in low flows, partly due to a shorter snow accumulation season and an earlier snowmelt peak (Mankin et al., 2015; Wanders et al., 2015). Projected decreases in the snow melt contribution to runoff leads to high risk in water securities in many regions including Sierra Nevada, Coast

Ranges, Rocky Mountains, Pyrenees, High Atlas, Aegean Region, Armenian Highlands, Lebanon/Antilebanon, Taurus, Zagros, Pamirs, parts of Hindu Kush-Karakoram-Himalaya region, Tibetan Plateau, Dabie mountains (*medium confidence*) (Naz et al., 2016).

Due to glacier shrinkage, glacier runoff is projected to decline in many high mountain regions by the end of the 21st century (*medium evidence, high agreement,*), indicating that peak water has already been reached or will be reached in the next decade or two (FAQ2.1 and Figure 2.7) (Bliss et al., 2014; Huss and Hock, 2018). Regional and local-scale projections show decreasing trends of glacier runoff in High Mountain Asia (Engelhardt et al., 2017a), Central Europe (Bavay et al., 2013; Uhlmann et al., 2013; Farinotti et al., 2016; Etter et al., 2017), South America (Frans et al., 2015; Ragettli et al., 2016) and North America (Beamer et al., 2016; Frans et al., 2016; Moyer et al., 2016). Part of the Indus (Lutz et al., 2014; Koppes et al., 2015) exhibit a steady increase in runoff. Increase followed by decrease in runoff until the 21st century is projected in regions such as European Alps (e.g., Farinotti et al., 2016) and western Himalaya (Engelhardt et al., 2017b).

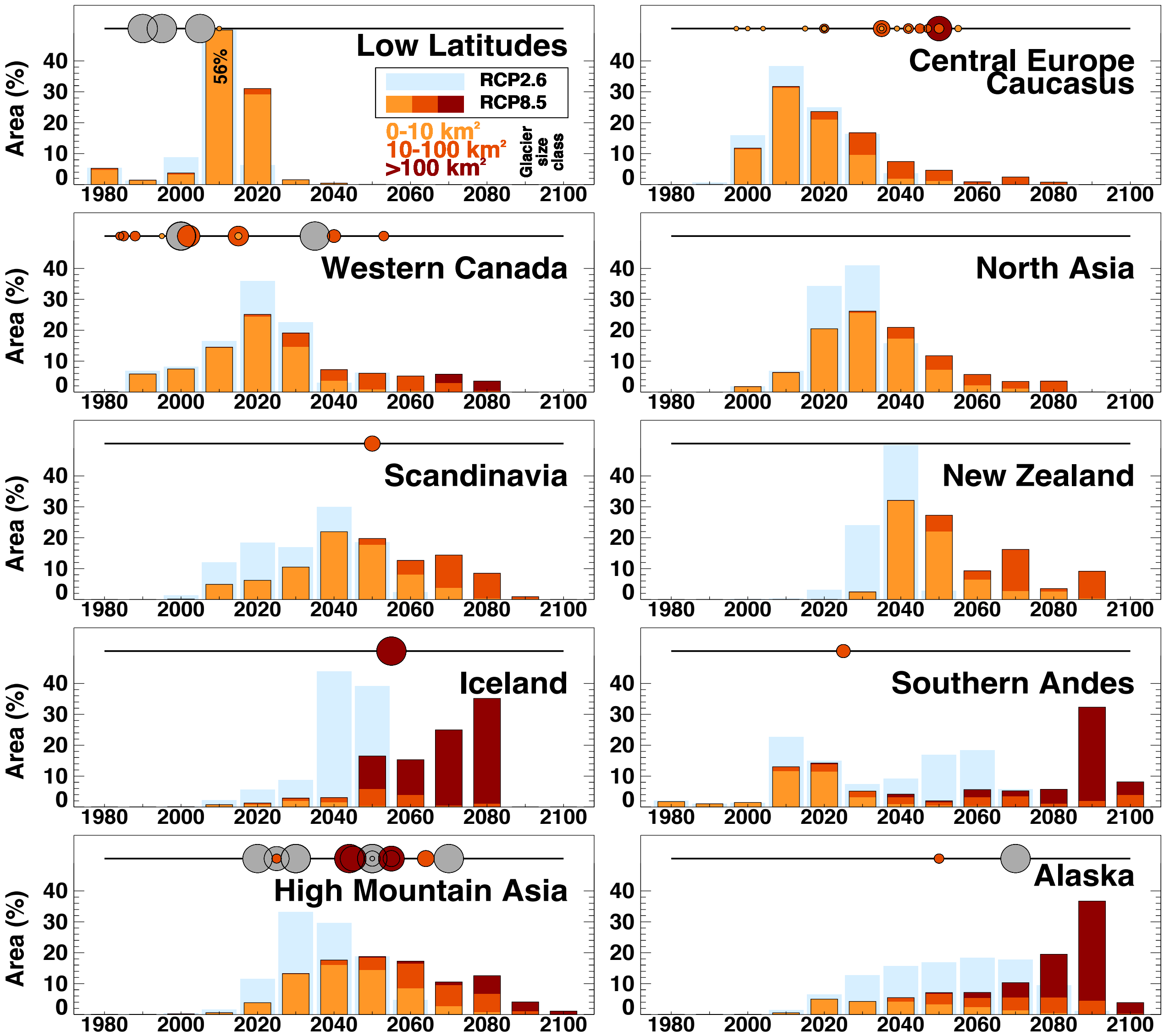


Figure 2.7: Timing of peak water from glaciers in different regions under two emission scenarios (RCP8.5 and RCP2.6). Peak water refers to the year when annual runoff from the initially glacierized area will start to decrease due to glacier shrinkage after a period of melt-induced increase. The bars are based on Huss and Hock (2018) who used a global glacier model to compute the runoff of all individual glaciers in a region until year 2100 based on 14 GCMs. Depicted is the area of all glaciers that fall into the same 10-year peak water interval expressed as a percentage of each region's total glacier area. Shadings distinguish different glacier sizes indicating a tendency for peak water to occur later for larger glaciers (only shown for RCP8.5 for better readability). Circles mark timing of peak water from individual

case studies based on observations or modelling (Appendix 2.A, Table 4); circle diameter is proportional to investigated glacier area. Grey circles show the total glacier area for studies that include a collection of glaciers of different sizes.

Catchments with large ice volumes are projected to have an increase in runoff (i.e., before peak water) but regions with smaller ice volumes have already passed peak water and presently show decreased runoff (Huss et al., 2017). In the tropical Andes, accelerated glacier melt may further increase glacier contribution to river flow until it reaches peak water (Pouyaud et al., 2005). In some basins in High Mountain Asia, contribution of ice melt to total runoff will remain relatively stable until the middle to late 21st century (e.g., Su et al. (2016), or even until the end of the 21st century Immerzeel et al. (2013)). However, it should be noted that as methods for calculating glacier runoff vary among different studies (Radić and Hock, 2014), runoff contribution from glaciers may not be directly comparable.

Permafrost warming may affect runoff by releasing water directly from ground ice or indirectly by changing hydrological pathways as permafrost degrades, however, studies quantifying the effects on runoff are scarce.

In summary, there is *high confidence* that glacier and snow cover decline have affected and will continue to affect the amounts and seasonality of river runoff in many glacier and snow-dominated river basins, however, trends in annual runoff can vary substantially among regions and can be even be opposite in sign. Changes in annual runoff and shifts in seasonality depend on glacier area and seasonal snow amount, which in turn are affected by changes in temperature and precipitation as well as the sensitivity of the catchment to these changes. The changes in runoff including diminished seasonal forecast skill in snow-dominant regions will affect downstream water management (Section 2.3.1.4).

[START FAQ 2.1 HERE]

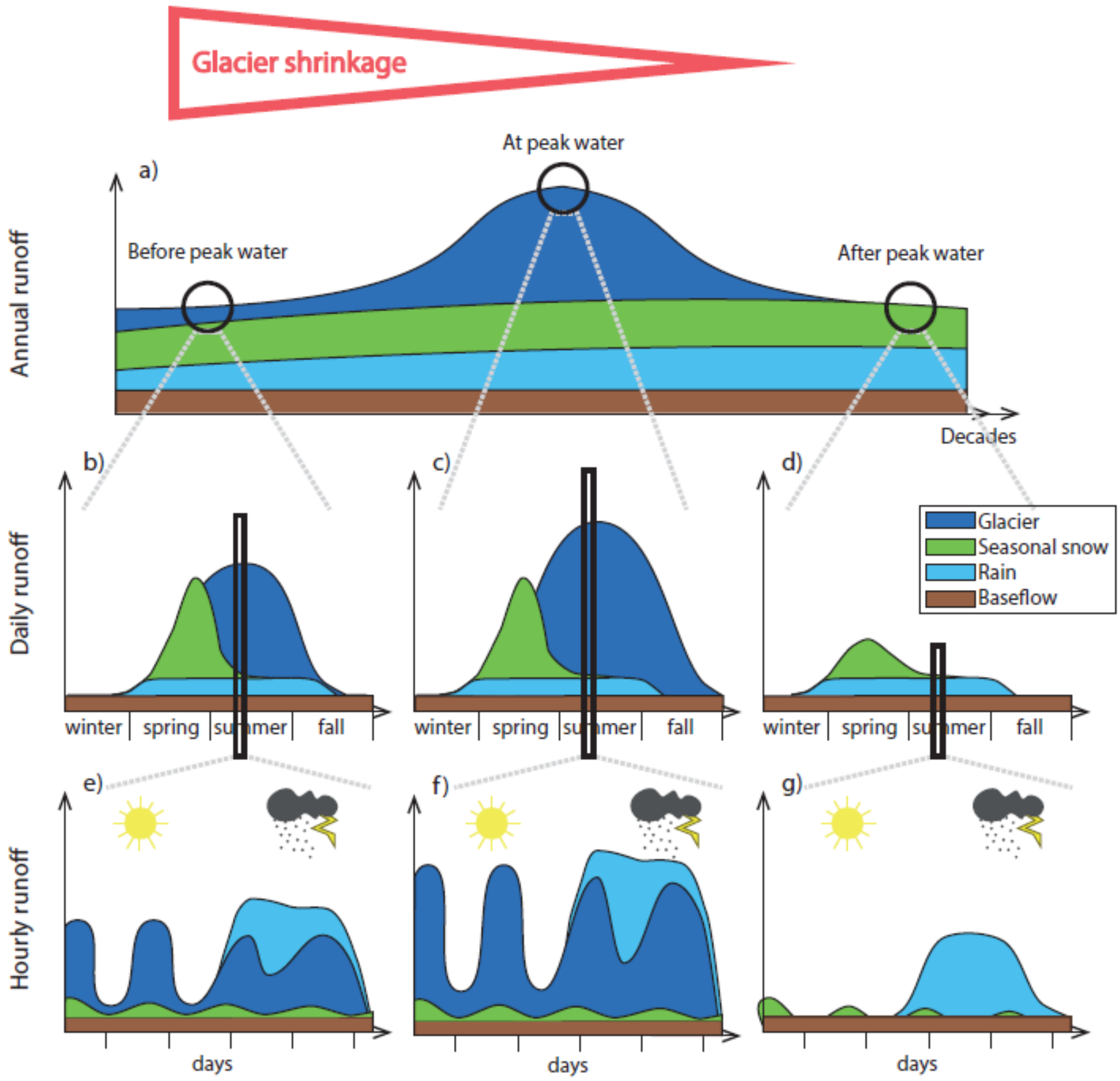
FAQ2.1: How does glacier shrinkage affect water supplies further downhill?

Shrinking glaciers can affect water supply not only close to the glacier but also far from mountain areas. Glaciers act as natural water storage, since precipitation such as rain and snow falling on the glacier can be stored temporarily as snow or ice for periods ranging from a few hours to many centuries. As glaciers shrink in response to climate change, water is released from long-term storage. Initially, glacier runoff increases, because the glacier melts faster and communities downstream experience increased water flow. However, there will be a turning point, perhaps after a few decades, often called ‘peak water’, after which glacier runoff will decrease: the glacial water store decreases steadily as the glacier continues to shrink and eventually melts completely (see FAQ2.1; Figure 1a). In highly glaciated river basins, peak water runoff can exceed the initial runoff by 50 percent or more. This extra water can be stored or used immediately in many different ways, such as for hydropower or irrigation. After the turning point, glacier runoff declines steadily depending on the climate, the speed of glacier shrinkage, and glacier size. Eventually this additional supply of water stops as the glacier disappears and communities downstream lose this valuable water source. River runoff will then depend mainly on rainfall, snow melt, and evaporation.

Furthermore, glacier decline can change the timing in the year when most water is available downhill. In mid- or high latitudes, glacier runoff is greatest in the summer, when glacier ice continues to melt once the winter snow has disappeared. Hence, glaciers can support a reliable water supply, especially during warm and dry periods, even where there is little glacier cover. After peak water, the capacity for glacier meltwater to compensate for hot and dry summers will decrease (see FAQ2.1, Figure 1b-d). In tropical areas, seasonal temperature variations are small and alternating wet and dry seasons are the main control on the amount and timing of glacier runoff throughout the year.

During the summer melt season, glacier runoff is greater during the day, when temperatures and solar radiation are highest (see FAQ2.1, Figure 1e-g). As peak water occurs, more intense glacier melt rates also increase these daily runoff maxima significantly. Moderate glacier cover (10-40% of a river basin area) regardless of peak water tends to reduce the variability of total amounts of runoff from year to year, as more meltwater during hot and dry years can compensate for the reduced rainfall.

The effects of glacier runoff further downhill depend on local characteristics or the distance between a glacier and a certain location. Close to the glaciers (e.g., several kilometres), initial increases in yearly glacier runoff until peak water followed by decreases can affect water supply, and the intensified peaks in daily runoff from the glacier can pose a flood hazard. Although with increasing distance the impact of glacier shrinkage on total river runoff becomes small or negligible, the melt water from glaciers in the mountains can be an important source of water in hot and dry periods when runoff is low, even hundreds of kilometres away from the glaciers. Other components of the water cycle such as rainfall, evaporation and snow melt can compensate or intensify the effects of changes in glacier runoff as the climate changes.



FAQ 2.1, Figure 1: Schematic overview of changes in runoff from a significantly (> 50%) glaciated river basin as glaciers shrink. (a) annual runoff from the entire basin. (b-d) close-up of runoff variations throughout one year from glacier melt, seasonal snowmelt, rain and baseflow (delayed flow of water from below the surface), (e-g) corresponding hourly variations when sunny and rainy weather during the summer (e-g). Note that seasonal and daily runoff variations are different before, during and after peak flow.

[END FAQ 2.1 HERE]

2.3.1.2 Water Quality

Cryospheric change in the mountains and associated changes in runoff and water temperature can significantly influence water quality. Ecosystem services can be affected through the release of stored anthropogenic compounds. The deposition and release of black carbon is associated with legacy pollutants, notably persistent organic pollutants (POPs), particularly polycyclic aromatic hydrocarbons, and heavy metals (Hodson, 2014). These pollutants have been released to surface waters from alpine glaciers in the nearby Gangetic Plain during the dry season (Sharma et al., 2015). Poly chlorinated biphenyls have been linked to glacial melt and although their use has declined or ceased there is lag time of release from glaciers (Li et al., 2017). Glaciers also represent the most unstable stores of dichlorodiphenyl-trichloroethane (DDT) in European and other mountain areas flanking large urban centres and glacier-derived DDT has accumulated in lake sediments downstream from glacierized mountains (Bogdal et al., 2010). Bioflocculation can increase the residence time of contaminants stored in glaciers and may reduce their overall toxicity (Langford et al., 2010).

Of the heavy metals, mercury is of particular concern and an estimated 2.5 tonnes has been released by glaciers to downstream ecosystems across the Tibetan Plateau over the last 40 years (Zhang et al., 2012). In more pristine areas, geogenic mercury contributions from sediment-rich glacier runoff can be as large or larger than the flux from melting ice (Zdanowicz et al., 2013). Both glacier erosion and atmospheric deposition contributed to the high rates of total mercury export found in a glacierized watershed in coastal Alaska (Vermilyea et al., 2017) and mercury output is predicted to increase in glacierized mountain environments (Sun et al., 2017; Sun et al., 2018b) (*medium confidence*). However, a key issue is how much of this glacier-derived mercury, largely in the particulate form, is converted to toxic methyl mercury downstream. Indeed, projections indicate future climate change will enhance the mobilisation of metals in metamorphic mountain catchments (Zaharescu et al., 2016). The release of toxic contaminants, particularly where glacial melt waters are used for irrigation and drinking water in the Himalayas and the Andes, is potentially harmful (Hodson, 2014) (*medium confidence)*.

Water originating from areas of permafrost degradation can also contribute heavy metals that exceed guideline values for drinking water quality (Thies et al., 2013). Northern hemisphere permafrost regions are estimated to store 1,656 (+/- 962) Gg of mercury of which over 50% is frozen in permafrost and thus vulnerable to mobilization with via mountain permafrost thaw (Schuster et al., 2018). In addition, permafrost degradation can enhance the release of other trace elements (e.g., aluminium, manganese and nickel) with climate change (Colombo et al., 2018).

Soluble reactive phosphorus (SRP) concentrations in rivers downstream of glaciers are predicted to decrease with declining glacier coverage (Hood and Berner, 2009) as a large percentage of SRP is associated with glacier-derived suspended sediment (Hawkings et al., 2016). In contrast, dissolved organic carbon (DOC), dissolved inorganic nitrogen and dissolved organic nitrogen concentrations in pro-glacial rivers will increase with glacier shrinkage (Hood et al., 2015; Milner et al., 2017) (*medium confidence)*. However, bulk DOC bioavailability will decrease over time because glacier-derived DOC is highly bioavailable (Hood et al., 2009) and readily incorporated into downstream biota (Fellman et al., 2015). Globally, mountain glaciers are estimated to release about 0.8 Tera g yr^{-1} of DOC to downstream ecosystems (Li et al., 2018). Loss rates of DOC from glaciers in the high mountains of the Tibetan Plateau were estimated to be ~0.19 Tera g C yr^{-1}, (Li et al., 2018) suggesting that DOC is released more efficiently from Asian mountain glaciers than other regions (Liu et al., 2016). Permafrost degradation is also a major and increasing source of bioavailable DOC (Abbott et al., 2014; Aiken et al., 2014). Major ions calcium, magnesium, sulphate and nitrate (Colombo et al., 2018) are also released by permafrost degradation as well as acid drainage leaching into alpine lakes (Ilyashuk et al., 2018).

Increasing water temperature has been reported in some high mountain streams (e.g., Groll et al., 2015; Isaak et al., 2016), producing changes in water quality and species richness (Section 2.3.3). In contrast, water temperature in highly glacierized regions are expected to show a transient decline, due to an enhanced cooling effect from increased meltwater (Fellman et al., 2014). In summary changes in the mountain cryosphere will cause significant shifts in downstream nutrients (DOC, nitrogen, phosphorus) and influence water quality through increases in heavy metals, particularly mercury, and other legacy contaminants.

2.3.1.3 Cryosphere Change and Mining

In some mountain regions, glacier retreat and related processes of change in the cryosphere have afforded greater accessibility for extractive industries and development of related activities that extract materials such as minerals and metals. Accelerated glacier shrinkage and retreat have been reported to facilitate cold mountain mining activities in South America, North America, and Central Asia, which also interact with and have consequences for other social, cultural, economic, political, legal, ecological, hydrological, or nature-protection measures, where climate change impacts also play a role (Evans et al., 2016; Petrakov et al., 2016). Conversely, there is also some evidence suggesting there is a link between enhanced economic activities and development strategies, mainly in extractive industries, that have affected glaciers and the mountain space around it (Kronenberg, 2013; Khadim, 2016). For instance, mining interventions have had reported impacts by changing glacial dynamics and glacier structure, due mainly to activities such as excavation, extraction, and use of explosives, which deposit dust and other mine waste material close to or over glaciers during extraction and transportation (Kronenberg, 2013; Jamieson et al., 2015; Evans et al., 2016). These have reportedly generated slope instabilities, accelerated the retreat of glaciers due to enhanced surface melt, and affected water quality through contamination and pollution, impacting upon communities and ecosystems downstream (Kronenberg, 2013; Anacona et al., 2015; Jamieson et al., 2015; Evans et al., 2016). However, there is no published evidence on relationships and direct links between climate induced changes in the cryosphere and impacts to/from extractive activities in the high mountains.

2.3.1.4 Key Impacts and Vulnerability

2.3.1.4.1 Hydropower

Water for hydropower reservoirs and plants often originates from glacier and snowmelt, so changes in the cryospheric components can directly affect hydropower production. Hydropower comprises about 16% of electricity generation globally but contributes close to 100%, in many mountainous countries (IHA, 2017; IHA, 2018). The impact of cryosphere changes on hydropower is intrinsically related to the associated changes in glacier and snowmelt water runoff, which can both increases and decrease (Section 2.3.1.1, FAQ2.1) depending on the region. Many hydropower plants were constructed according to historical meteorological and hydrological data, and may need considerable modification to operate under a different streamflow and climate regimes. Potential increased flows in ice-fed reservoirs in the short-term will be followed by longer term total annual decrease in many areas, and the pattern can be further complicated by differing evolutions for catchments that are close together (Fatichi et al., 2013; Gaudard et al., 2013; Gaudard et al., 2014). In addition, hydropower generation can be affected by increased formation of glacial lakes and the resulting risk of damage to hydropower infrastructure due to glacier lake outburst floods (Section 2.3.2). An increase in frequency or magnitude of extreme events will increase the risk for hydropower plants and may bring risks to energy security (Jackson and Ragulina, 2014). Floods, and especially glacier lake outburst floods, can also cause a sudden increase in sediment input to a reservoir and damage hydropower plants. Hydropower reservoirs can be beneficial in managing downstream consequences of changes in the cryosphere as storing and providing freshwater in drought condition, or alleviating the effects of glacier floods (Jackson and Ragulina, 2014; Colonia et al., 2017). When glacier runoff declines, it will often be necessary to construct additional storage to maintain the same level of power production. Limitation in projecting glacier change induced runoff changes and lack of data in high mountains make it difficult to forecast the capacity of hydropower (Condom et al., 2012). Hence, despite the realisation of risks in hydropower (IHA, 2017; IHA, 2018) or efforts of hydropower agencies and regulatory bodies to quantify changes or to develop possible adaptation strategies, only a few organisations are incorporating current knowledge of climate change into their planning. The World Bank uses a decision tree approach to identify potential vulnerabilities in a hydropower project incurred from key uncertain factors and their combinations (Bonzanigo et al., 2015).

2.3.1.4.2 Agriculture

Recent research emphasizes the effects of changes in glaciers and snow cover on stream and river runoff (Section 2.3.1.1), which provides irrigation water that is important for cultivation of crops and pasture. Some areas, particularly in the Central Andes (Baraer et al., 2015), the central Himalayas (McDowell et al., 2013; Rasul and Molden, in review; Mukherji et al., in revision) and western Himalayas (Nüsser and Schmidt, 2017; Mukherji et al., in revision) and the Cascades (Frans et al., 2016) have already been observed to face reductions in water available for irrigation (*medium evidence, high agreement*), with many others, such as Central Asia, projected to do so by the end of the century (*very high confidence*) (Chen et al., 2016; Huss et al., 2017). In some areas, such as the Swiss Alps, there is no clear signal for change in total precipitation by the end of the century, but the decline in glacier and snow contributions lead to a reduction of river runoff in the summer, when it is most important for agriculture (Addor et al., 2014).

The evidence for attributing changes in irrigation supply to cryosphere processes and river runoff is based both on scientific observations and on indigenous knowledge and local knowledge, the former characterised as systematic in nature and focused on large rivers in middle and lower portions of major basins, the latter often more localized and focused on small rivers in upper portions of basins; moreover, the former tend to be more evenly distributed around the world, and the latter more evident in the Himalayas, the Karakoram and the Andes (McDowell and Koppes, 2017). There have been few efforts to compare, reconcile and integrate these two sets of observation (Ingty, 2017).

In addition, cryosphere changes may affect agriculture through their impacts on land cover and on soil (*medium confidence*). Reduction in snow cover and thawing of permafrost contribute to changes in the land cover and land use, which may affect the role of forests in protecting soil and storing water, with potential cascading effects on downstream areas (Cristea et al., 2014; Hinckley et al., 2014; Chaulagain, 2015).

In addition, increasing air temperatures will increase crop evapotranspiration, thus increasing water demand for crop production to maintain optimal yield (Beniston and Stoffel, 2014) (*high confidence*). Rising temperatures are associated with upslope movement of cropping zones, which favours some farmers in high mountain areas, who have become able to cultivate new crops, such as onions, garlic and apples in Nepal (Huntington et al., 2017; Hussain et al., 2018) and maize in Ecuador (Skarbø and VanderMolen, 2014). Agriculture in high mountain areas is sensitive to non-climate drivers (Porter et al., 2014) as well, such as market forces and political pressures (Montana et al., 2016; Sietz and Feola, 2016; Figueroa-Armijos and Valdivia, 2017) though these lie largely outside the scope of this chapter.

2.3.1.4.3 Adaptation measures for agriculture

Information about adaptation activities in mountain agriculture consist largely of case studies in specific communities, valleys or watersheds, and thus is limited in spatial and temporal scope and context (McDowell et al., submitted). Large-scale review and synthesis studies, based on complementary data sourced from remote sensing, national statistics, or other systematic methods to establish trends over space and time, remain limited. The majority of the activities are autonomous, though some are planned, or carried out with support from national governments, NGOs, or international aid organizations. Section 2.3.6.1 examines adaptation activities in mountain agriculture, discussed here, in the context of livelihood systems and emigration.

[START BOX 2.3 HERE]

Box 2.3: Local Adaptation Responses to Cryosphere Shrinkages and Water Shortage in Northwest India

Cryosphere changes have impacted water resources and livelihoods in Ladakh, a cold arid mountain region in the western Himalayas (Clouse et al., 2017). Agriculture in Ladakh depends on snow and glacier meltwater for irrigation. Since the 1980s, the region has been experiencing a shortage of water for irrigation due to recession of glaciers and reduced snow cover, particularly before the monsoon, when irrigation, is critical for agriculture (Crook and Osmaston, 1994; Bhasin, 1997; Nandargi and Dhar, 2011; Clouse, 2016). To cope with this water shortage for irrigation, villagers in the region have developed a number of adaptation measures including artificial ice reservoirs (locally known as ice stupa), frozen ponds and snow barriers, which store meltwater as ice in winter and release it in spring. The first two techniques have also been termed “ice reservoirs” (Nüsser et al., 2018).

Ice stupas hold water, frozen into conical shapes, until spring, when it melts and flows down to the fields (Clouse, 2016; Clouse et al., 2017). Frozen ponds are formed from water which is conveyed across a slope through channels and check dams to shaded surface depressions near the villages. The diverted water freezes into ice reserves, which melt in the spring and flow to fields (Vince, 2009; Clouse, 2016; Shaheen, 2016). Snow flow barrier are bands which collect wind-blown snow located near high mountains passes in the winter, which melts in the spring and meltwater directed to fields (Chalise and Khanal, 1996; Clouse, 2016).

These adaption measures and techniques, particularly snow barriers, use local materials and draw on local knowledge (construction techniques) and suit with local physical environment (Clouse, 2016; Nüsser et al., 2018). One recent study Nüsser et al. (2018) examined 14 ice reservoirs and concluded that they are suited to the physical environment and serve as appropriate "site-specific water conservation strategies". However, this study questions their usefulness as a long-term adaption stragegy, because their efficacy depends on winter runoff and freeze thaw cycles, both of which have been changing and are projected to change further.

[END BOX 2.3 HERE]

Adaptation activities within agricultural systems are found in mountain regions around the world, though they are absent from some localities. They include a variety of practices and measures, listed below. The most widely reported types are autonomous adaptations, except for the ones where government support is specifically noted.

Adaptation responses within irrigation systems include the adoption of new irrigation technologies. Water-delivery technologies which reduce loss are adopted in Chile (Young et al., 2010). Similarly, greenhouses have been adopted in Ecuador (Knapp, 2017) and Nepal (Konchar et al., 2015) to reduce evapotranspiration, and reduce frost damage, though limited access to finance is a barrier to these activities. Box 2.3 describes innovative irrigation practices in India. Local pastoral communities have responded to these challenges with techniques broadly similar to those in agricultural settings by expanding irrigation facilities for instance in Switzerland (Fuhrer et al., 2014). In addition to adopting new technologies, some water-users make investments to tap more distant sources of irrigation water. Cross-Chapter Box 2 in Chapter 1 discusses such efforts in northern Pakistan. Local institutions and embedded social relations play vital role in enabling communities to respond to the impacts. Indigenous pastoral communities who have tapped into new water sources to irrigate new areas in Peru have also strengthened the control of access to existing irrigated pastures (Postigo, 2014) and Bolivia (Yager, 2015). In an example of indigenous populations in the US, two tribes who share a large reservation in the northern Rockies rely on rivers which receive glacier meltwater to irrigate pasture, and to maintain fisheries, domestic water supplies, and traditional ceremonial practices. Tribal water managers have sought to install infrastructure to promote more efficient water use and to protect fisheries, but these efforts have been impeded by land and water governance institutions in the region and by a history of social marginalization (McNeeley, 2017).

The adoption of new crops and varieties is an adaptation response found in several regions. Farmers who rely on irrigation in the Naryn River basin in Kyrgyzstan have shifted from the water-intensive fruits and vegetables to fodder crops such as barley and alfalfa, which are more profitable. Upstream communities, with greater access to water and more active local institutions, are more willing to experiment with new crops than those further downstream (Hill et al., 2017). In other areas, crop choices also reflect responses to rising temperatures along with new market opportunities such as the demand for fresh vegetables by tourists in Nepal (Konchar et al., 2015; Dangi et al., 2018) and the demand for roses in urban areas in Peru (SENASA, 2017). Indigenous knowledge and local knowledge, access to local and regional seed supply networks, proximity to agricultural extension and support services also facilitate the adoption of new crops (Skarbø and VanderMolen, 2014).

High mountain communities have sought new financial resources from wage labour (Section 2.3.6.2) and from government sources to support adaptation activities. Local water user associations in Kyrgyzstan and Tajikistan have adopted less water-intensive crops and reorganized the use and maintenance of irrigation systems, investing government relief payments after floods (Stucker et al., 2012). Similar combinations are reported from India (Dame and Mankelow, 2010; Clouse, 2016; Nüsser and Schmidt, 2017), Nepal (McDowell et al., 2013) and Peru (Postigo, 2014). In contrast, fewer steps have been adopted in Uzbekistan, due to low levels of capital availability and rigidities in national agricultural policies (Aleksandrova et al., 2014).

Adaptation efforts in the agricultural sector are sometimes scanty or entirely absent (*medium evidence, high agreement*). Planted areas have been reduced in a number of different areas in Nepal (Sapkota et al., 2010; Gentle and Maraseni, 2012; Sujakhu et al., 2016; Bastakoti et al., 2017) and in India (Bastakoti et al., 2017).

Though local residents perceive the impacts on agriculture and livelihoods, few are able to undertake adaptation activities. Barriers to adaptation include a lack of finance and technical knowledge, weakness of community and state organizations, and ambiguous property rights. In the Cordillera Blanca in Peru, there have also been reports of declining agricultural yields, due to cryosphere changes, in the absence of adaptation activities. Irrigation water is less available in some tributaries of the Santa River, while dry spells and unseasonal frosts have also impacted agriculture, creating food insecurity (Bury et al., 2011); market pressures and shifts in water governance also put pressure on agriculture (Rasmussen, 2016).

2.3.1.4.4 Drinking water supply

Few studies provide thorough empirical assessments of the effects of cryosphere change on drinking water supply. The Andes have received the most attention, especially the major cities of Quito, Ecuador; Lima, Peru; and La Paz, Bolivia (Chevallier et al., 2011). Of these, the cryosphere impacts are most severe in La Paz (Buytaert and De Bièvre, 2012). The contribution of glacier water to the city between 1963 and 2006 was assessed at 15% (Soruco et al., 2015), though rising as high as 86% during drought months (Buytaert and De Bièvre, 2012). These studies show the vulnerability of the cities to water scarcity from cryosphere changes in the near term. Two other Andean cities Huancayo and Huaraz, in Peru, have also faced threats to their water supplies from glacier retreat. The fraction of glacier meltwater that flow into the rivers that serve these cities is 13% and 19%, respectively; these cryosphere contributions to the cities' water supplies have seen significant decline in the present day. Population growth and poor infrastructure maintenance in these cities exacerbate water scarcity (Buytaert and De Bièvre, 2012; López-Moreno et al., 2014; Somers et al., 2018).

In summary, it is *likely* that cryospheric changes will increase risks to the amounts of drinking water supply as well as quality, but *confidence is low*, due to *limited evidence*.

2.3.1.5 Water Governance and Response Measures

Changes in the cryosphere, as a response to climatic change and resultant changes in hydrological regime and river runoff (Section 2.3.1.1) are likely to bring additional challenges in water management and governance particularly in areas where cryosphere melting contributes significantly in rivers and stream flows (Barnett et al., 2005; Munia et al., 2018; Smith and Bookhagen, 2018). In glaciated river basins, the changes lead to consequences for the water availability that are important to local and other social-ecological systems at the catchment level and further downstream (Scott et al., 2018).

A changing cryosphere poses certain risks to the governance of water resources, given reported examples of conflict and tensions among key actors in high mountains and neighbouring regions (Bocchiola et al., 2017; Milner et al., 2017), particularly where legal frameworks and a lack of adequate water governance beyond glacier protection are not present (Carey et al., 2014a; Haeberli et al., 2016; Vuille et al., 2018). Moreover, there is insufficient evidence to link these regulations to the effective security and management of water resources. For example, as Hurlbert and Gupta (2016) state, even when the glacier protection law in Argentina has prevented mining development, it has not been effective in tackling the impacts of climate change to secure water resource in glacier zones. In 2018, the United Nations Human Rights Council passed a declaration to "protect and restore water-related ecosystems" in mountain areas as elsewhere from contamination by mining (UNHRC 39th Assembly General, 2018).

Coordination and mainstreaming of common transboundary adaptation strategies remains a key challenge for many mountain regions, primarily due to weak institutional capacities and limited voice of mountain communities in key decision-making policy processes at local, regional, national and transboundary water systems. For example, differences in political and other interests along transboundary river systems, are reasons for conflicts between the upstream and downstream communities in different countries in the Hindu Kush Himalayas, particularly in glacier dominated rivers (Molden et al., 2014; Rasul, 2015).

Transboundary policies that account for adaptation to climate change can address conflict management and improve co-management of transboundary resources, in the context of a changing mountain cryosphere. Adaptation policies that connect local and national adaptation efforts at the transboundary level are also receiving attention in the literature as a key consideration. For example, in Nepal, the adaptation policy in mountain areas focuses largely on forestry and soil conservation, while in Bangladesh the focus is on water and flood management (Pandey et al., 2016; Gain et al., 2017). Moreover, India has a strong domestic agricultural market and thus focuses on the resilience of its agricultural sector and related activities. Despite their common element in addressing water and its governance, the differences in how these policies frame adaptation and their intended effects make it difficult to implement as a common transboundary adaptation strategy, resulting in low scalability or transferability of policies across cases and mountain regions (Vij et al., 2017).

Recently, however, a few efforts have been made in different mountain regions for transboundary cooperation on monitoring and assessing glacier balance, river runoffs and better management of cryosphere challenges such as in the Alps (an intergovernmental joint project among Austria, France, Germany, Italy and Switzerland) (Mair et al., 2011), in the Andes (CARE 2010, a project for adaptation to the impact of glacier retreat by the World Bank) (Warner et al., 2012), Central Asia (regional collaboration among Kazakhstan, Kyrgyzstan, Tajikistan, Turkmenistan and Uzbekistan) (Komagaeva, 2017) and Hindu Kush Himalaya region (ICIMOD: an intergovernmental centre to share knowledge and research among Afghanistan, Bangladesh, Bhutan, China, India, Myanmar, Nepal and Pakistan) to facilitate transboundary cooperation in glaciated river basins (Scott et al., 2018).

2.3.2 Landslide, Avalanche and Flood Hazards

High mountains are particularly prone to mass movements such as avalanches or landslides. Snow and ice exert key control on mountain slope stability. The behaviour of ice changes dramatically when approaching and reaching 0°C, leading to changes in many mass movement processes. This section assesses knowledge gained since previous IPCC reports, in particular Chapter 3 of the Special Report on 'Managing the Risks of Extreme Events and Disasters to Advance Climate Change Adaptation' (Seneviratne et al., 2012).

Hazards covered in this section range from localised effects on mountain slopes and adjacent valley bottoms (reach of up to several kilometres) to events reaching far into major valleys and even surrounding lowlands (reach of tens to hundred kilometres), and to cascading events. Natural hazards and associated disasters are sporadic by nature, and vulnerability and exposure exhibit strong geographical variations. These characteristics require that assessments of change are based not only on direct evidence, but also on laboratory experiments, theoretical considerations, and numerical modelling.

2.3.2.1 Observed and Projected Changes

2.3.2.1.1 Unstable slopes and landslides

Permafrost thaw typically increases the rate of movement of frozen debris bodies or lowers their surface due to loss of ground ice (subsidence). These processes typically affect only structures on top of permafrost or in the immediate vicinity, and can cause problems for engineered structures such as buildings, hazard protection structures, roads, or rail lines (Phillips and Morrow, 2007; Jin et al., 2008). Slope movement and subsidence/heave are strongly related to ground temperature, ice content, and water input (Wirz et al., 2016; Kenner et al., 2017). Where massive ground ice is exposed, retrogressive thaw erosion can develop (Niu et al., 2012).

There is *high confidence* that the frequency of rocks detaching and falling from steep slopes (rock fall) increased within zones of thawing permafrost over the past half-century (Allen et al., 2011; Ravanel and Deline, 2011; Coe et al., 2017). Permafrost thaw is also in theory expected to increase the likelihood of rock fall and rock avalanches (larger volumes compared to rock falls) (Gruber and Haeberli, 2007; Krautblatter et al., 2013), further confirmed by ice being present in the detachment zone of previous events (Geertsema et al., 2006; Phillips et al., 2017; Sæmundsson et al., 2018). Summer heat waves can trigger rock instability with only short delay (Allen and Huggel, 2013). This is in line with theoretical considerations about fast thaw of frozen fractures (Hasler et al., 2011) and other climate impacts on rock stability, such as from large temperature variations (Luethi et al., 2015). Similarly, permafrost thaw can increase the frequency and volumes of landslides from frozen sediments (Wei et al., 2006; Ravanel et al., 2010; Lacelle et al., 2015).

A range of slope instability types is connected to glacier retreat (Allen et al., 2011; Evans and Delaney, 2015). Debris left behind by retreating glaciers (moraines) is typically over-steepened directly after glacier retreat and has been documented to slide or collapse, or to form fast flowing water-debris mixtures (debris flows) (Zimmermann and Haeberli, 1992). Over timescales of decades to millennia, or even over glacial cycles, rock slopes adjacent to glaciers or formerly covered by them tend to become unstable, start to deform and, in some cases, eventually collapse. Increased landslide activity in recently deglacierized zones has received increased attention (Korup et al., 2012; McColl, 2012; Deline et al., 2015; Kos et al., 2016; Serrano et al., submitted). According to Cloutier et al. (2016) more than two-thirds of the large landslides that occurred in northern British Columbia between 1973 and 2003, occurred on Little-Ice-Age-exposed cirque walls. Ice-rich permafrost environments following glacial retreat enhance slope mass movements(Oliva and Ruiz-Fernández, 2015). At lower elevations, re-vegetation and rise of tree limit are able to stabilize shallow slope instabilities. Overall, there is *high confidence* that glacier retreat in general destabilizes adjacent debris and rock slopes, but knowledge of trends remains incomplete due to scarce observations.

The acceleration of rock glaciers (frozen debris slopes slowly deforming under gravity), which is in principle their expected response to increases in ground temperatures (Kääb et al., 2007), can contribute to increased debris-flow activity by increasing material supply to debris-flow starting zones (Stoffel and Graf, 2015; Wirz et al., 2016; Kummert et al., 2017).

Ice break-off and subsequent ice avalanches are often a natural process from steep glacier fronts. Where climatic changes alter the geometry and thermal regime of steep glaciers, they may cause ice avalanche hazards to either increase, decrease, or remain unaltered, depending strongly on local conditions (Fischer et al., 2013; Faillettaz et al., 2015), so that clear global or regional trends are not expected. The few available observations are insufficient to detect trends. Several cases are known where large parts or even complete steep glaciers fail, and there is *high agreement* that for steep glaciers frozen to their bedrock such failures are promoted by an increase in basal ice temperature (Gilbert et al., 2015).

Glacier surges constitute a wide-spread form of quasi-periodic substantial increases in glacier speed over a period of a few months to years, often accompanied by glacier advance (Harrison et al., 2015; Sevestre and Benn, 2015). In a number of cases (Bevington and Copland, 2014; Round et al., 2017; Steiner et al., 2018), surge-related glacier advances dammed up rivers causing major flood hazards. In rare cases, glacier surges directly inundate agricultural land and damage infrastructure (Shangguan et al., 2016). The mechanisms involved in glacier surging do not require climatic changes to occur (Murray et al., 2003). Overall, however, there is *medium confidence*, that surging can be related to climate (Sevestre and Benn, 2015) and thus change with climate. Some glaciers and regional surge-type glacier clusters have reduced or even stopped surge activity, or are projected to do so, as a consequence of negative glacier mass balances (Eisen et al., 2001; Kienholz et al., 2017). In contrary, increased surge activity seems to occur for the region of positive glacier mass balances on and around the western Tibet plateau (Copland et al., 2009; Gardelle et al., 2012; Kääb et al., 2018). Enhanced melt-water production is also able to trigger or enhance surge-type instability, in particular for poly-thermal glaciers (i.e., glaciers that in parts contain ice below 0°C) (Dunse et al., 2015; Yasuda and Furuya, 2015). Deposition of mining debris on the ice is able to directly trigger surge-like glacier movement (Jamieson et al., 2015).

A rare type of glacier instability with large volumes (~ 10^7–10^8 m^3) and high mobility (up to 200-300 km/h), is the complete collapse of large sections of low-angle valley glaciers and subsequent ice/rock/debris avalanches. Such glacier collapses have been documented for only three cases; in the Caucasus Mountains in 2002 (Huggel et al., 2005; Evans et al., 2009), and the 2016 twin Aru events, Tibet (Kääb et al., 2018). Whereas climate changes seem not to have played a direct role in the 2002 Kolka Glacier collapse, climate-driven changes in glacier mass balance, water input into the glaciers, and partially frozen glacier beds were clearly involved in the 2016 collapses (Gilbert et al., 2018). Besides the 2016 Tibet cases, it is unknown how climate change could alter the potential for such massive and rare collapse-like glacier instabilities.

2.3.2.1.2 Snow avalanches

Snow avalanches can occur spontaneously, purely due to meteorological causes, following significant snow precipitation episodes or wet-snow conditions conducive to wet-snow avalanches. Avalanches can also be triggered accidentally or artificially, e.g., by the passage of skiers, fall of rocks or ice, explosives (control measures) or earthquakes (Schweizer et al., 2003). Changes in snow cover characteristics are expected to induce changes in spontaneous avalanche activity (Naaim et al., 2013; Steinkogler et al., 2014). Tree-rings and historical archives are used to infer longer-term changes of avalanche activity (Giacona et al., 2017). Ballesteros-Cánovas et al. (2018) reported increased avalanche activity in the Western Indian Himalaya over the past decades related to increased frequency of wet-snow conditions (i.e., presence of liquid water in snow). Correlations between avalanche activity and climate variability were identified in North and South America (García-Sellés et al., 2010; McClung, 2013). In Europe, past changes in meteorological and snow conditions are correlated with avalanche runout elevation (Eckert et al., 2013) and with avalanche activity in forested areas (Teich et al., 2012). Studies suggest that avalanche mass and run-out distance have decreased in past decades, with a decrease of avalanches with a powder part since the 1980s, a decrease of avalanche numbers below 2000 m but an increase above (Eckert et al., 2013; Lavigne et al., 2015; Gadek et al., 2017). A positive trend in the proportion of wet snow avalanche activity in December thru February was shown over the 20th century (Pielmeier et al., 2013; Naaim et al., 2016). Land use and land cover changes also contribute to changes in avalanches (García-Hernández et al., 2017; Giacona et al., in review). Mostly inconclusive results were reported by Sinickas et al. (2015) and Bellaire et al. (2016) regarding the relationship between avalanche activity, climate change and disaster risk reduction measures in North America. In summary, there is *medium confidence (medium evidence, low agreement)* that avalanche activity shifted from dry to wet snow avalanches, and the size and run-out distance of dry snow avalanches decreased.

Future projections mostly indicate an overall decrease in seasonal snow at the annual scale, but the occurrence of occasionally high snow precipitation should remain significant throughout most of the 21st century Castebrunet et al. (2014) showed that for the French Alps future climate conditions under an SRES A1B scenario for mid and end-century may favour the appearance of a wet snowpack at high elevations or earlier in the season, which could require to upgrade or modify prevention measures (Ancey and Bain, 2015). Katsuyama et al. (2017) reached similar conclusions in Northern Japan. Castebrunet et al. (2014) estimated an overall 20-30% decrease of natural avalanche activity for mid and end of the 21st century, respectively, under SRES A1B scenario, compared to the reference period 1960-1990. The overall trend in avalanche activity will depend on the regions and elevation, and in some areas may increase first due to increased wet snow conditions at high elevation while the snowpack remains deep enough, then decrease due to the continued reduction of seasonal snow (Castebrunet et al., 2014; Mock et al., 2017). In summary, there is *high confidence* that projected snow changes in mountain regions will favour major shifts in avalanche activity and character. Wet snow avalanches are projected to increasingly occur anytime during the winter, and the overall number and runout distance of dry snow avalanches to decrease in regions and elevations experiencing significant reduction in snow conditions (*high confidence*).

There is no published evidence addressing the links between climate change and accidental avalanches triggered by recreationists or workers. The resulting risk and number of casualties will continue to strongly depend on the behaviour of people travelling in avalanche terrain.

2.3.2.1.3 Floods

Glacier lakes are lakes that are dammed by glacier ice or moraines. Glacier-related floods, including floods from outbursts of such lakes, are documented for most glacierized mountain ranges and are among the most far reaching glacier hazards. They can affect areas tens to hundreds of kilometres downstream (Carrivick and Tweed, 2016). Retreating glaciers often leave behind lakes at their fronts and margins (Frey et al., 2010; Gardelle et al., 2011; Loriaux and Casassa, 2013). Lake systems often develop on top of downwasting, low-slope glaciers where they coalesce from temporally highly variable supraglacial lakes (Benn et al., 2012; Narama et al., 2017). Advancing glaciers can temporarily dam rivers, for instance through surging (Round et al., 2017), causing particularly large floods once the ice dams breach. There is *high confidence* that current global glacier shrinkage caused new lakes to form and existing lakes to grow (Loriaux and Casassa, 2013; Zhang et al., 2015; Buckel et al., 2018). There is also *high confidence* that the number and area of glacier lakes will continue to increase in the future. New lakes will develop closer to steep and potentially unstable mountain walls where lake outbursts can be triggered by the impact of landslides (Frey et al., 2010; ICIMOD, 2011; Linsbauer et al., 2016; Colonia et al., 2017). An exception are water accumulations under glaciers. Their outbursts can cause floods similar to those from surface lakes, but little is known about the processes involved and any trends under climate change (see also jökulhlaups under Section 2.3.2.1.4). How the number of glacier floods changed in the recent past is not well known (Carrivick and Tweed, 2016; Harrison et al., 2018), although a number of flood cycles have been documented, spanning decades (Geertsema and Clague, 2005; Russell et al., 2011). A decrease in glacier lake floods in recent decades could suggest a delayed response of lake outburst activity to glacier retreat (Harrison et al., 2018) but inventories might significantly underestimate the number of events (Veh et al., 2018). In contrast to the lakes themselves, lake outbursts are also not necessarily directly coupled to climatic changes as a number of other factors are important. Advancing, stagnant or retreating glaciers may all produce water bodies that are able to burst out.The thawing of permafrost and the melting of buried ice in lake dams (Fujita et al., 2013; Erokhin et al., 2017; Narama et al., 2017) have been shown to lower their stability and contribute to outburst floods.

Floods originating from the combination of rapid melting of snow and intense rain precipitation, referred to as rain-on-snow events, are some of the most damaging in mountain areas (Pomeroy et al., 2016; Il Jeong and Sushama, 2018). The hydrological response of a catchment to a rain-on-snow event depends on the characteristics of the precipitation event, but also on turbulent fluxes driven by wind and humidity, which are responsible for most of the melting energy (Pomeroy et al., 2016), and the state of the snowpack in particular liquid water content (Würzer et al., 2016). An increase of the occurrence of rain-on-snow events in high elevation zones, and a decrease in occurrence at the lowest elevations was reported (western U.S.A., 1949-2003, McCabe et al. (2007); Oregon, 1986 – 2010, Surfleet and Tullos (2013); Switzerland, 1972 – 2016, Moran-Tejéda et al. (2016), central Europe, 1950 – 2010, Freudiger et al. (2014)). Several studies report an increase in the occurrence of rain-on-snow events towards high latitudes in the Northern Hemisphere (Putkonen and Roe, 2003; Ye et al., 2008; Cohen et al., 2015). In summary, rain-on-snow events have increased over the last decades at high elevations and/or in high latitude areas, particularly during transitions periods from autumn to winter and winter to spring (*high confidence*). Rain-on-snow events have decreased over the last decade at low elevation or low-latitude areas, except for the coldest months of the year (*high confidence*).

Future projections follow the trends observed in past decades. Il Jeong and Sushama (2018) projected an increase in rain-on-snow events in winter and a decrease in spring, for the period 2041–2070 (RCPs 4.5 and 8.5) in North America, with conclusions corroborated by Musselman et al. (2018). The frequency of rain-on-snow events in the Swiss Alps is projected to increase at elevations higher than 2000 m a.s.l. (SRES A1B, 2025, 2055, and 2085) (Beniston and Stoffel, 2016). This study showed that the number of rain-on-snow events may increase by 50%, with a regional temperature increase of 2°C–4°C, and decrease with a temperature increase exceeding 4°C. In summary, the frequency of rain-on-snow events is projected to increase in areas where changes in snow cover are smallest (way above the mean snowline elevation), and happen earlier in spring and later in autumn (*high confidence*). The frequency of rain-on-snow events is projected to decrease below the mean snowline elevation (*high confidence*).

2.3.2.1.4 Combined hazards and cascading events

The largest mountain disasters in terms of volume, reach, damage and lives lost that involve ice, snow and permafrost occur through a combination of processes or chain reactions (Anacona et al., 2015; Evans and Delaney, 2015). Some process chains are frequent and typical, but others are rare, specific to local circumstances and difficult to anticipate. Glacier lake outbursts are often triggered by impact waves from snow avalanches, landslides, or calving events, or by temporary blockage of surface or subsurface drainage channels (Benn et al., 2012; Narama et al., 2017). Rock slope instability and catastrophic failure along fjords cause tsunami hazards (Hermanns et al., 2014; Roberts et al., 2014). A recent landslide-generated wave at Taan Fjord, Alaska, run up 193 m on the opposite slope and then travelled more than 20 km down the fjord (Higman et al., 2018). A 2017 tsunamigenic landslide in western Greenland (Gauthier et al., 2018) may have been related to permafrost degradation. The generated wave damaged a village and caused four fatalities. Landslides in glacier environments entrain snow and ice that fluidizes, incorporating additional loose glacial sediments or water bodies, and eventually multiplying the mobility, volume and reach of such mass movements compared to those without snow and ice involved (Schneider et al., 2011; Evans and Delaney, 2015). Glaciated frozen rock walls constitute particularly complex thermal, mechanical, hydraulic and hydrologic interactions between steep glaciers, frozen rock and its ice content, and unfrozen rock sections (Harris et al., 2009; Fischer et al., 2013; Ravanel et al., 2017). From the observed and expected degradation of permafrost, shrinkage of glaciers and increase of glacier lakes it is reasonable to assume that event chains involving these could in general increase in frequency or magnitude, but there is *limited evidence.*

During eruptions of ice and snow-clad volcanoes, substantial meltwater is often produced. This typically results in floods and/or lahars (mixtures of meltwater and volcanic debris) which can be exceptionally violent, and cause large-scale loss of life and destruction to infrastructure (Barr et al., 2018). The most devastating example from recent history occurred in 1985, when the medium-sized eruption of Nevado del Ruiz volcano, Colombia, melted substantial amounts of snow and ice, and produced lahars that killed more than 23,000 people some 70 km downstream (Pierson et al., 1990). Ice and snow-clad volcanoes and associated hazards are typically focused in the Cordilleras of the Americas, with additional important locations in the Aleutian and Cascade Volcanic arcs (USA), Mexico, Kamchatka (Russia), Japan, New Zealand and Iceland (Seynova et al., 2017). In particular under Icelandic glaciers, volcanic activity and eruptions can melt large amounts of ice and cause especially large floods if water accumulates underneath the glacier (termed jökulhlaups) (Björnsson, 2003; Seneviratne et al., 2012). There is *high confidence* that over years-to-decades, as glaciers and seasonal snow-cover continue to decrease, interactions between volcanoes and the cryosphere will become less common. Thus, and despite the limited possibility for according observations, there is *medium confidence* that the overall hazard related to floods and lahars will gradually diminish (Aguilera et al., 2004; Barr et al., 2018). Though, shrinkage of glaciers may uncover steep slopes of unconsolidated volcanic sediments, thus decreasing the resistance of these volcano flanks to heavy rain fall and increasing the hazard from related debris flows. Finally, there is some, but still *limited evidence* to suggest that as glaciers diminish, 'unloading' of the mantle will trigger an increase in the frequency of large-magnitude volcanic eruptions, and associated hazards (Cooper et al., 2018). Much of the evidence to support this assertion is theoretical, rather than observational; though there is strong evidence to suggest that such behaviour occurred during past periods of deglaciation (Cooper et al., 2018). However, how volcanoes will respond to future changes in climate remains unclear, though the potential remains that large volcanic eruptions will be more likely and more frequent with continued deglaciation (Blankenship et al., 1993; Schmidt et al., 2013). In all, future climate-driven changes in snow and ice are expected to modify the character of volcanic activity and its impacts. However, this is likely to occur in complex and locally variable ways and at a variety of time-scales (Barr et al., 2018; Swindles et al., 2018).

2.3.2.2 Vulnerability, Exposure and Impacts

Climate risk has emerged as a central concept at the science-policy dialogue interface, where risk is defined as a physical event (hazard) intercepting with an exposed and vulnerable system (e.g., community or ecosystem) (IPCC SREX and Assessment Report 5, AR5). A clear distinction here is that vulnerability is linked to the inherent characteristics of a society or system, while changes in climate primarily influence hazards and related trends (Figure 2.8). Impacts are then the realisation of risk. In theory this conceptualisation provides a basis for the adaptation strategies that consider both the changes in frequency or magnitude of hazards due to climate change as well as societal dynamics that shape the exposure and vulnerability of people and social-ecological systems.

2.3.2.2.1 Drivers of exposure

There is *high confidence* that the exposure of people and infrastructure to hazards in high mountain regions has increased over recent decades, and this trend is expected to continue in the future. In some regions, tourism development is one of the drivers that have been linked to this change, where often poorly regulated expansion of infrastructure such as roads, foot-tracks, and overnight lodging bringing more visitors into remote valleys and exposed sites (Gardner et al., 2002; Uniyal, 2013). As an example for this development, many of the more than 350 fatalities resulting from the 2015 earthquake-triggered snow-ice avalanche in Langtang, Nepal, were foreign trekkers and their local guides (Kargel et al., 2016). Further, several thousand religious pilgrims were killed during the 2013 Kedarnath glacier flood disaster (State of Uttarakhand, Northern India) (Kala, 2014). The expansion of hydropower (Section 2.3.1) is another key factor, and in the Himalaya alone, up to two-thirds of the current and planned hydropower projects are located in the path of potential glacier floods (Schwanghart et al., 2016). Changes in exposure of local communities are complex and vary regionally. For example, climate change, and related threats to traditional forms of livelihoods is leading to outmigration from some mountain regions (Tiwari and Joshi, 2015), while other communities may relocate towards higher elevation, and potentially more exposed zones where they are able to maintain their crops under a warmer climate (Malla, 2009).

2.3.2.2.2 Drivers of vulnerability

Considering the wide-ranging social, economic, and institutional factors which enable communities to adequately prepare for, respond to, and recover from climate impacts (Chen et al., 2014a; Cutter and Morath, 2014), there is *medium confidence* that mountain dwelling communities, particularly within developing countries, are highly vulnerable to the adverse effects of climate change. This assessment recognizes that there are few studies that have systematically assessed the vulnerability of mountain communities(Carey et al., 2017). Coping capacities within mountain communities may be limited due to a number of reasons.

Fundamental weather and climate information is lacking to support both short-term early warning for imminent disasters, and long-term adaptation planning (Rohrer et al., 2013). Communities may be politically and socially marginalised (Marston, 2008). Incomes are typically lower and opportunities for livelihood diversification restricted (McDowell et al., 2013). Emergency responders can have difficulties accessing remote mountain valleys after disasters strike (Sati and Gahalaut, 2013). Cultural or social ties to the land can limit freedom of movement (Oliver-Smith, 1996). Conversely, there is evidence that some mountain communities exhibit enhanced levels of resilience, drawing on long-standing experience, and indigenous knowledge and local knowledge gained over many centuries of living with extremes of climate and related disasters (Gardner and Dekens, 2006). In the absence of sufficient data, few studies have considered temporal trends in vulnerability (Huggel et al., 2015a).

2.3.2.2.3 Impacts

Empirical evidence from past events shows that cryosphere-related landslides and floods can have severe impacts on lives and livelihoods, often extending far beyond the directly affected region, and persisting for several years. Glacier lake outburst floods alone have over the past two centuries directly caused at least 400 deaths in Europe, 5745 deaths in South America, and 6300 deaths in Asia (Carrivick and Tweed, 2016), although these numbers are heavily skewed by individual large events occurring in Huaraz and Yungay, Peru (Carey, 2005) and Kedarnath, India (Allen et al., 2016). National-level economic impacts from glacier floods have been greatest in Nepal and Bhutan (Carrivick and Tweed, 2016). The disruption of vital transportation corridors that can impact trading of goods and services (Gupta and Sah, 2008), and the loss of earnings from tourism can represent significant far-reaching and long-lasting impacts (Nothiger and Elsasser, 2004; IHCAP, 2017). Less tangible, but equally important impacts concern the cultural and social disruption resulting from temporary or permanent evacuation (Oliver-Smith, 1979). Over the period 1985–2014, absolute economic losses in mountain regions from all flood and mass movements (including non-cryospheric origins) were highest in the Hindu-Kush Himalaya region (USD 45 billion), followed by the European Alps (USD 7 billion), and the Andes (USD 3 billion) (Stäubli et al., 2017). Other impacts are related to drinking and irrigation water and livelihoods. Given the expected continued increase in exposure of people and assets in high mountains there is *high confidence* that the impacts from high mountain floods and landslides will increase over the coming century in regions where risk reduction and adaptation strategies prove insufficient in reducing losses.

2.3.2.3 Disaster Risk Reduction and Adaptation to Flood and Landslide Hazards

Applying an integrative socio-physical risk perspective to flood and mass movement hazards in high mountain regions paves the way for adaptation strategies that address the underlying components of hazard, exposure and vulnerability (Carey et al., 2014b; McDowell and Koppes, 2017; Allen et al., 2018; Vaidya et al., 2018). Critical scientific literature reflecting on the successes or lessons learned from adaptation projects remains scarce, though. Some degree of adaptation action has been identified in a number of countries with glaciated mountain ranges, particularly across the Andes and Himalayas, and mostly as reactive responses (rather than formal anticipatory plans) to hydrological changes (McDowell et al., submitted). Specifically for flood and landslide hazards, adaptation strategies include hard engineering solutions such as glacial lake drainage or slope stabilisation that reduce the hazard potential, nature-based solutions such as revegetation efforts to stabilise hazard-prone slopes or channels, hazard and risk mapping as a basis for land zoning and early warning systems that reduce potential exposure, and various community-level interventions to develop disaster response programmes, build local capacities and reduce vulnerability. There is a long tradition of engineered responses to reduce glacier flood risk, most notably beginning in the mid-20th century in Peru (Box 2.4) and Switzerland (Haeberli et al., 2001), and more recently in the Himalaya (Ives et al., 2010).

Early warning systems necessitate strong local engagement and capacity building to ensure communities know how to prepare and respond to any emergency, but also to ensure the long-term sustainability of any project. The need for ground-level education and communication has been demonstrated in Peru, where international cooperation led to the installation of a technologically advanced early warning system drawing on best practices from Europe (Muñoz et al., 2016), only to have the equipment destroyed following opposition and mistrust from local communities (Fraser, 2017). In some cases, local residents have played active roles in detecting and communicating flood events and in formulating responses. In both Pakistan and Chile, glacier flood warnings, evacuation and post-disaster relief have largely been community-led (Ashraf et al., 2012; Iribarren Anacona et al., 2015).

Cutter et al. (2012) highlight the post-recovery and reconstruction period as an opportunity to build new resilience and adaptive capacities. However, too often this process is rushed or poorly supported by appropriate sustainable long-term planning, as illustrated following the 2013 Kedarnath glacier flood disaster, where guest houses and even schools are being rebuilt on-site in the same exposed locations, driven by short-term economic motives (Ziegler et al., 2014). In mountain regions, there is a particular need for forward-thinking planning and anticipation of emerging risks and opportunities, as changes in the cryosphere, together with socio-economic, cultural and political developments are producing conditions beyond historical precedent (Haeberli et al., 2016).

Different types of participants, such as researchers, policy-makers, international donors and local communities do not always agree on the timing of different stages of disaster risk reduction projects and programs, impeding full coordination (Huggel et al., 2015b; Allen et al., 2018). Furthermore, several authors have called for an improved evidential basis to underpin adaptation planning, arguing for a paradigm shift, whereby transdisciplinary and cross-regional collaboration places human societies at the centre of studies, providing the basis for more effective and sustainable adaptation strategies (McDowell et al., 2014; Carey et al., 2017; Vaidya et al., 2018; McDowell et al., submitted).

The evidence and learnings emerging from mountain regions generally reconfirm the findings from the SREX, including the requirement for multi-pronged approaches customised to local circumstances, integration of indigenous knowledge and local knowledge together with improved scientific understanding and technical capacities, strong local participation and early engagement in the process, and high-level communication and exchange between all actors. Particularly for mountain regions, there is *high confidence* that integration of knowledge and practices across natural and social sciences, and the humanities, is an important prerequisite for addressing complex hydrological challenges, hazards, and risks.

[START BOX 2.4 HERE]

Box 2.4: Challenges to Farmers and Local Population Related to Shrinkages in the Cryosphere: Cordillera Blanca, Peru

The Cordillera Blanca of Peru contains most of the glaciers in the tropics. Glacier coverage in Peru's Cordillera Blanca declined significantly in the recent past (Burns and Nolin, 2014; Mark et al., 2017). Since the 1940s, glacier hazards have killed thousands (Carey, 2005; Carey, 2010) and remain threatening (Rivas et al., 2015; Emmer et al., 2016b; Somos-Valenzuela et al., 2016). Glacier wastage over time has also caused hydrologic variability, with "peak water" passed in most Cordillera Blanca basins several decades ago, resulting in a reduction in glacier runoff, particularly in the dry season (Baraer et al., 2012; Vuille et al., 2018). Residents living adjacent to the Cordillera Blanca have long recognized this glacier shrinkage, including rural populations living near glaciers and urban residents worried about glacier lake floods and glacier landslides (Carey, 2010; Bury et al., 2011; Jurt et al., 2015; Heikkinen, 2017; Walter, 2017). Glacier hazards and the glacier runoff variability increase human vulnerability and uncertainty while diminishing adaptive capacity (Rasmussen, 2016).

Cordillera Blanca residents' risk of glacier-related disasters is shaped by an assemblage of variables, with physical and societal factors intersecting to increase risk. Physical hazards include rapidly expanding glacial lakes, new lake formation, slope instability, rising temperatures, and precipitation changes (Emmer et al., 2016a; Emmer et al., 2016b; Colonia et al., 2017; Haeberli et al., 2017). Human vulnerability stems not only from physical variables but also from other factors, such as minimal access to education and healthcare, poverty, limited political influence and resources, weak government institutions, and residents' inhabitation of potential flood paths (Hegglin and Huggel, 2008; Carey et al., 2012; Lynch, 2012; Carey et al., 2014a; Heikkinen, 2017). Early warning systems have been, or are being, installed at glacial lakes Laguna 513 and Palcacocha to protect populations (Muñoz et al., 2016). Lake 513 was previously drained for outburst prevention in the early 1990s but nonetheless caused a destructive flood in 2010 (Carey et al., 2012; Schneider et al., 2014). The early warning system was subsequently installed, but some local residents destroyed it in 2017 due to political, social and cultural conflicts (Fraser, 2017). The nearby Lake Palcacocha also threatens populations (Wegner, 2014; Somos-Valenzuela et al., 2016; Heikkinen, 2017).

Vulnerability to hydrologic variability and declining glacier runoff is also shaped by intertwining human and biophysical drivers playing out in dynamic hydro-social systems (Bury et al., 2013; Carey et al., 2014b; Rasmussen et al., 2014; Drenkhan et al., 2015; Carey et al., 2017; Heikkinen, 2017; Paerregaard, 2018). Water security is influenced by both water availability (supply from glaciers) as well as by water distribution, which is affected by factors such as water laws and policies, global demand for agricultural products grown in the lower Santa River basin, energy demands and hydroelectricity production, potable water usage, and livelihood transformations over time (Carey et al., 2014b; Vuille et al., 2018). In some cases, the formation of new glacial lakes can create opportunities as well as hazards, such as new tourist attractions and reservoirs of water, thereby showing how socioeconomic and geophysical forces intersect in complex and ways (Carey, 2010; Colonia et al., 2017).

[END BOX 2.4 HERE]

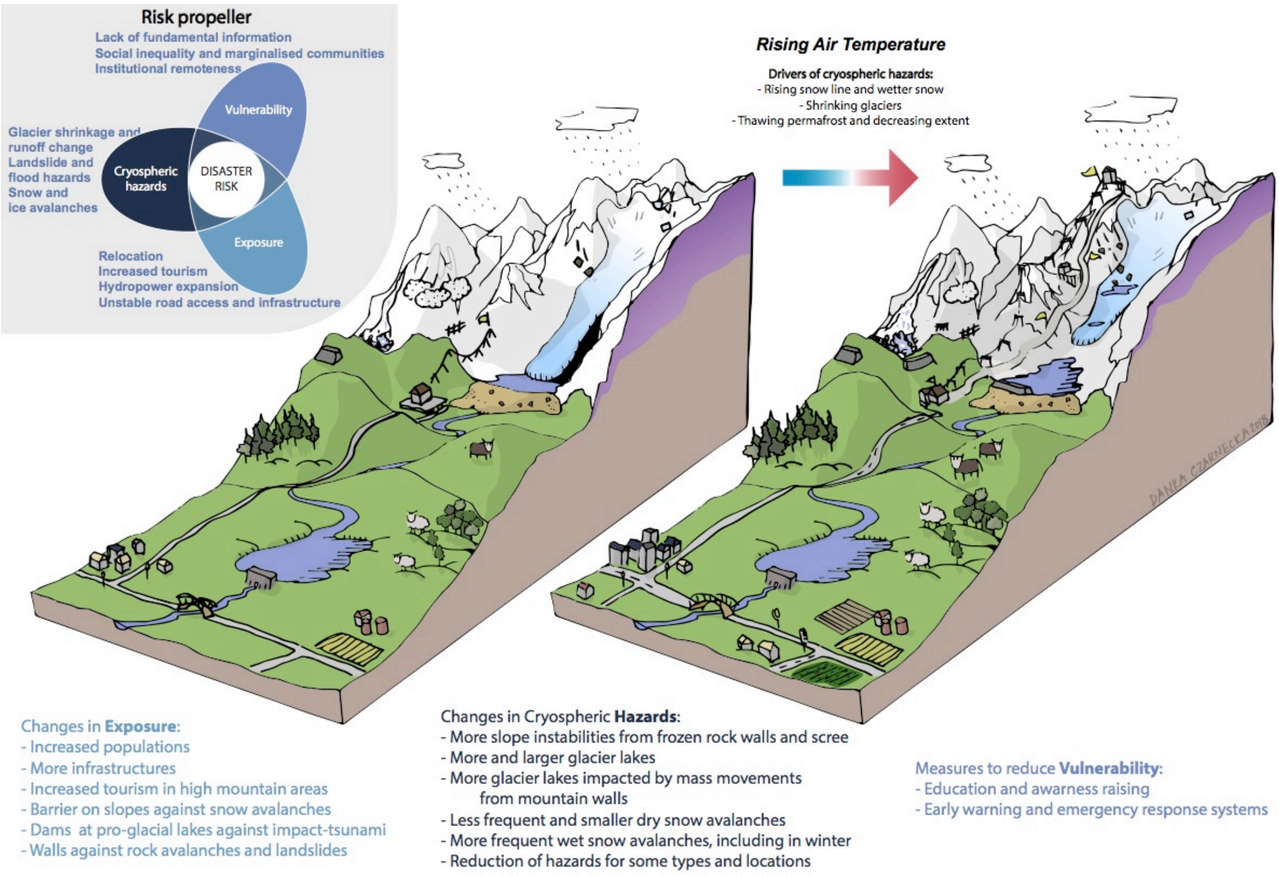


Figure 2.8: Anticipated changes in high mountain hazards under climate change, driven by changes in snow cover, glaciers and permafrost, overlay changes in the exposure and vulnerability of individuals, communities, and mountain infrastructure.

2.3.3 Biodiversity

2.3.3.1 Terrestrial Ecosystems

Cryosphere components determine plant species composition and ecosystem function in all high mountain regions (e.g., Anthelme and Lavergne, 2018) (Figure 2.2). The shrinking cryosphere, manifested by retreating glaciers, thawing permafrost, and shorter periods of seasonal snowpack, increases and decreases water availability, thereby affecting diverse ecological processes with significant consequences on phenology, productivity, and biodiversity (Bjorkman et al., 2018; Steinbauer et al., 2018; Wang et al., 2018) (Figure 2.9).

Many of the changes in plant phenology, such as earlier onset of growth and flowering (König et al., 2018; Post et al., 2018), and the upslope expansion of treeline (Lubetkin et al., 2017) are linked to patterns of snow cover (Xie et al., 2017; Winkler et al., 2018) (*high confidence*). Thus, warmer winters with less snow cover increases the time gap between snow loss and start of plant growth with consequences for energy, water and nutrient dynamics (Contosta et al., 2017; Hubbard et al., 2018). Ecosystem shifts to taller vegetation have varied impacts on snowpack and snowmelt rates, often dependent on regional climate and leaf life span, which affect energy exchange (Bjorkman et al., 2018). Years where climatic extremes occur, especially where snow depth is exceptionally low or high, may counteract the effect of land use and warming on treeline and limit upward expansion (Barros et al., 2017).

Near the upper elevation limit for plants, plant species richness has increased (*high confidence*), e.g., five-times faster across Europe during the recent decade, 2007-2016, compared to 1957-1966 (Steinbauer et al., 2018). These increases may reverse, due to loss of endemic and cold-adapted specialist species due to reduction of climatically suitable habitats (e.g., Dullinger et al., 2012) (*medium confidence*). Declines in the abundance of alpine species have been found near the warm-margins of their altitudinal and latitudinal distribution (Lesica and Crone, 2017) and through long-term experimental warming that also leads to early snowmelt (Panetta et al., 2018). Due to reduced snow duration, two mountain ecosystems are expected to be most vulnerable even in a short-term perspective, namely the nival and snowbed communities characterized by perpetual and long-lasting snow (Bjork and Molau, 2007; Pauli et al., 2014; Pickering et al., 2014) *(high confidence*). Large-scale biodiversity loss in high-mountain environments may be delayed, because high landscape heterogeneity in snow cover leads to different microhabitats (Scherrer and Körner, 2011; Graae et al., 2018), and the high longevity of most alpine species (Rosbakh and Poschlod, 2018) (*medium confidence*).

In exceptionally arid, high mountain areas, thawing permafrost contributes substantially to plant water supply, thereby promoting plant growth (e.g., Ishikawa et al., 2005; Wang et al., 2016) (*high confidence*). Though, over time, permafrost degradation may lead to shifts from wetlands to semi-deserts and steppes (Cheng and Jin, 2013) (*medium confidence*). On the Tibetan Plateau, permafrost thaw removed barriers to water flow, changing creek locations and causing the degradation of pastures (Jin et al., 2009). Other desertification processes and loss of vegetation due to grazing often result in soil warming and further permafrost thaw (Yang et al., 2010).

Recolonization velocity of plants after mountain glaciers retreat can differ considerably depending on the environmental context (*medium confidence*). Where glaciers are receding in the Central Andes, this process is slow (Zimmer et al., 2018). At lower elevation mountain regions, such as coastal Alaska, glacier recession can result in more rapid colonization by forest vegetation, with implications for terrestrial carbon storage at regional scales (Buma and Barrett, 2015). In the Himalayas, models predict forests will establish in high mountain areas currently characterized as rock or ice (Rashid et al., 2015). Climate warming may increase plant establishment following glacier retreat (Mondoni et al., 2015).

Where cryospheric changes lead to disturbance by fire (Gergel et al., 2017), landslide, flood, or an altered ecosystem state dominated by invasive species (Lembrechts et al., 2018), shifts in terrestrial vegetation may be rapid (*medium confidence*). Due to decreasing snowfall and earlier snowmelt, mountain ecosystems across the Western U.S. are experiencing an increase the severity and extent of wildfires (e.g., Westerling, 2016). Globally, climate variability that leads to fuels aridity accounts for nearly one third of burned area (Abatzoglou et al., 2018). Snowfall in years after the disturbance can influence ecosystem recovery (Wilson, 2018) and resulting exports of soil and nutrients to freshwater ecosystems (Gould et al., 2016).

2.3.3.2 Freshwater Ecosystems

As melt water from glaciers decreases with a shrinking cryosphere, particularly where glaciers are small in size, river flows will become more stochastic and water temperature and channel stability should increase (Milner et al., 2017) (*high confidence*). Riverine habitats will have less heterogeneity (Giersch et al., 2017) in mountain headwater areas, favouring more generalist communities, particularly with respect to aquatic bacteria and fungi (Fell et al., 2017) shifting towards taxa favouring warmer waters (Freimann et al., 2015). Increased primary production, dominated by diatoms and golden algae (*Hydrurus foetidus*), will probably occur as glacial runoff decreases, with increased richness, density and alpha diversity of algal species (Fell et al., 2017). However some stenotherm (cold tolerant) diatom species will be lost resulting in a decrease in gamma (regional) diversity (Fell et al., 2017). Bryophytes may become more common as channels become more stable (Milner et al., 2017) (*medium confidence)*.

A global analysis by Brown et al. (2018) indicates that predictable mechanisms govern river invertebrate community responses to decreasing glacier runoff. Whilst specialist cold species may be lost, generalists are likely to take their place resulting in an increase in functional diversity (see Glossary) as glacier cover decreases (*medium confidence*). Community assembly models demonstrated that dispersal limitation was the dominant process underlying these patterns, although environmental filtering was also evident in highly glacierized basins. Analysis of three comprehensive invertebrate datasets from equatorial (Ecuador), temperate (Italian Alps) and sub-Arctic (Iceland) regions indicated a distinct threshold of glacier cover in the number of taxa that decrease in density below 19-32% glacier cover in river basins (Milner et al., 2017) (*medium confidence*). Other global-scale studies have identified 11-38% of the regional species pool would be lost when glacier cover in the watershed falls below 5-30% (Jacobsen et al., 2012).

The identification of these transition or tipping points between alternate ecosystem states is important to aid conservation and mitigation efforts in mountain environments as these critical thresholds of glacier/snowmelt/permafrost loss can be identified (Khamis et al., 2014). These tipping points with a shift to an alternate state arises from the loss of cold stenothermic species, many of them endemic (see Glossary), as glacial runoff decrease leads to a loss of beta (turnover between reaches) and gamma diversity as glaciers retreat and switch to a regime more dominated by snowmelt. However local (alpha) diversity will increase (*high confidence*) in certain regions although this is less likely in the tropics (Jacobsen et al., 2012; Cauvy-Fraunié et al., 2016). There is also clear evidence from Europe (Pyrenees) and North America (Rockies) that glacier loss threatens the existence of endemic, cold-adapted invertebrates (Brown et al., 2007; Giersch et al., 2015; Giersch et al., 2017), likely leading to a loss of genetic diversity (Jordan et al., 2016). Beta genetic diversity within individual riverine invertebrate species in mountain headwater areas will decrease as the loss of environmental heterogeneity in headwater habitats with decreasing glacier runoff reduces the isolation of individuals and permits intermixing to a greater degree (Finn et al., 2013; Finn et al., 2016; Hotaling et al., 2018) (*medium confidence)*. An in-situ plasticity response by invertebrates to these environmental changes is the most likely mechanism for taxa persistence compared to migration (as many headwater alpine taxa are dispersal limited) or adaptation (Hotaling et al., 2017). Extinction of range restricted prey taxa may increase as more favorable conditions facilitates upstream movement of large bodied invertebrate predators (Khamis et al., 2015).

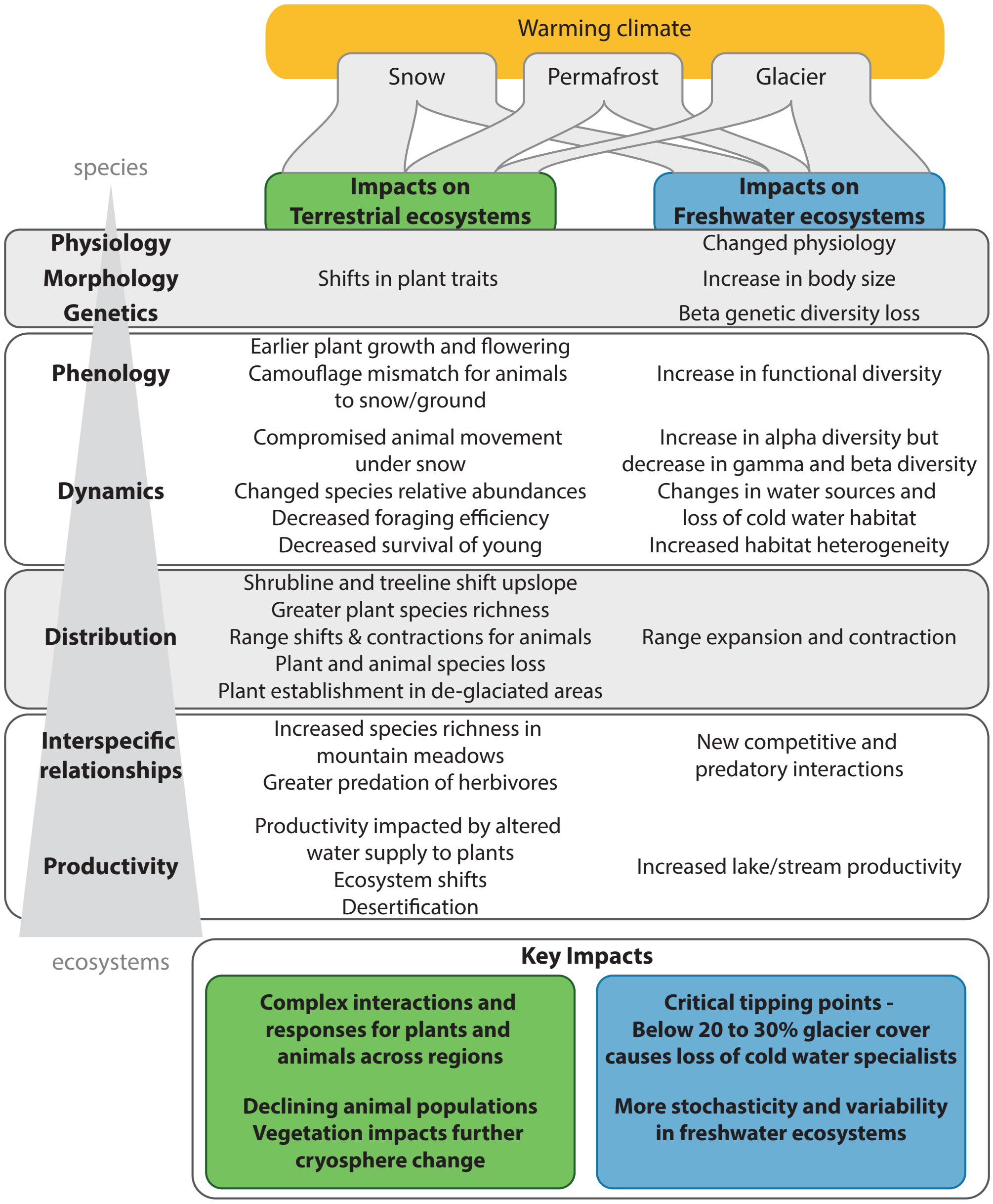


Figure 2.9 Summary of the effects of a shrinking cryosphere on terrestrial and freshwater ecosystems in high mountain areas, which occur across different scales of biological organization from species to ecosystems. The strength of the three cryospheric elements (change in snow, permafrost and glaciers) on the two ecosystems is expressed by the width of the link. The gray elements have no significance except for highlighting the different sections.

2.3.3.3 Wildlife and Fisheries

Changes in snow timing, duration, depth, and density are expected to have wide-ranging effects on animals in high mountain regions (e.g., Williams et al., 2015) (*high confidence*). Across the northern hemisphere, 21 diverse vertebrate species undergo a photoperiod-triggered seasonal molt from brown in summer to white in winter as camouflage to track the seasonal presence or absence of snow (Mills et al., 2018; Zimova et al., 2018). As winter-white animals confront more days with brown snowless ground, camouflage mismatch could increase predation rates sufficiently to cause population declines (Zimova et al., 2016; Atmeh et al., 2018) (*high confidence*) and may have already contributed to range contractions for several species, including mountain hares and ptarmigan (Imperio et al., 2013; Pedersen et al., 2017).

As a result of changing snowpack, loss of subnivean space (under snow), the insulated and thermally stable region under the snow at the soil-snow interface, will affect mountain animals that depend on this habitat (Penczykowski et al., 2017; Zuckerberg and Pauli, 2018) (*high confidence*). A compromised subnivean can negatively affect animal movement patterns (e.g., arthropods, small rodents and carnivores), survival (e.g., wood frogs; Sinclair et al., 2013), and overall abundance and community assembly (e.g., Australian alpine arthropods, Slatyer et al. (2017)) (*medium confidence*). Mountain ungulates can be negatively affected when forage becomes inaccessible due to ice formation under snow (Hansen et al., 2011).

Snow also shapes morphological and behavioural adaptations of large vertebrates in mountain regions (e.g., Gilg et al., 2012) (*high confidence*). For example, thermoregulatory dynamics, movement efficiency, foraging, and denning success in wolverines are all facilitated by presence of snow, (McKelvey et al., 2011; Webb et al., 2016) (*medium confidence*). Loss of snow patches that inhibit biting insects can influence foraging and thus, reproductive fitness (Vors and Boyce, 2009). Additionally, the phenological mismatch between birth date and vegetation growth is typically due to snowmelt timing and can lead to reduced survival of young and population mean fitness (e.g., Plard et al., 2014) (*high confidence*). Traditional Saami reindeer husbandry in northern Scandinavian mountains is threatened by rain-on-snow events, to which responses are limited (Keskitalo, 2008; Forbes and Kumpula, 2009; Eira, 2012; Mathiesen et al., 2013; Cramer, 2014). Ice layers in the snow can lead to reindeer starvation in their wintering habitats, and to massive abortion before springtime calving or calves with reduced vigour.

Semi-aquatic mammals (e.g., the water shrew (Soricinae) and desmans (Talpidae)) may be affected by the changes in the cryosphere. Reduced glacial influence may significantly impact the Iberian desman (*G pyrenaicus*) (Biffi et al., 2016) as their range is influenced strongly by the presence of aquatic invertebrate prey from rivers with low temperatures and fast-flowing, oxygenated waters. In contrast, glacier retreat has the potential to benefit amphibian species by creating more mountain river habitats with warmer waters and more abundant invertebrate prey (Ludwig et al., 2015).

Many climate variables influence fisheries through direct and indirect pathways. The key variables or drivers of interest include: changes in air and water temperature, precipitation, salinity, ocean circulation and mixing (linked to glacial runoff), nutrient levels, ice cover, glacial melt, storm frequency and intensity, and flooding (Stenseth et al., 2003). For example in Alaska where salmon are important in both commercial and sport fisheries, all species will be affected by reductions in glacial runoff from mountain glaciers over time (Schoen et al., 2017) particularly in the larger systems where migratory corridors to spawning grounds are affected (*medium confidence*). A shrinking cryosphere will affect cold mountain salmonid species, e.g., brook trout, by causing fish to migrate further upstream to find suitable habitat or in some cases become extinct (Hari et al., 2006). Within the Yanamarey watershed of the Cordillera Blanca in Peru, fish stocks have either declined markedly or have become extinct in many streams, possibly due to seasonal reductions of fish habitat in the upper watershed because of the glacier recession (Bury et al., 2011; Vuille et al., 2018). In contrast glacial recession along the mountains of the Pacific Northwest and coastal Alaska have created a large number of new stream systems which have been and will be in the future colonized by anadromous salmon that contribute to fisheries, both commercial and sport (Milner et al., 2017; Schoen et al., 2017) (*medium confidence*).

Insights into plant and animal phenological shifts and their distribution, and the rapid changes in freshwater environments in high mountain areas are critical to implement planning and adaptation. Adaptation approaches to shifts in mountain terrestrial and freshwater ecosystems may vary greatly according to the extent they are sustaining lives, livelihoods, or cultures.

2.3.4 Tourism and Recreation

The mountain cryosphere provides important aesthetic and recreational services to society (Xiao et al., 2015). The cryosphere is also an important resource for tourism, providing livelihood options to mountain communities. Changes in the cryosphere have affected mountain tourism and recreation services both negatively and positively, consequently affecting the visitor experience and revenue for local communities.

Winter sports (ski tourism), given its reliance on favourable atmospheric and snow conditions, is particularly vulnerable to climate change in mountains (Arent et al., 2015, Chapter 3, SR15). Efforts to reduce its vulnerability to occasionally unfavourable meteorological conditions include improved slope preparation, i.e. grooming; snowmaking, i.e., artificial production of snow (Steiger et al., 2017); snow farming, i.e. storage of snow over the summer season (Grünewald et al., 2018). Snow management methods have been implemented since the late 20th century, sharply rising in past decades (e.g., Spandre et al., 2015), and are routinely used to reduce the exposure of the ski tourism industry to inter-annual variability of snow conditions. Their effectiveness for adaptation to long-term climate change depends not only on their own sensitivity to atmospheric conditions (e.g., temperature thresholds for snowmaking), but also on externalities such as water and energy availability (required for snowmaking), and meeting investment and operating costs (Dawson and Scott, 2013; Hopkins and Maclean, 2014; Steiger et al., 2017). Socio-cultural perception and acceptability of high-tech measures could limit the overall adaptation potential. In general, ski resorts located at lower elevation are more exposed to climate change than higher elevation resorts, given higher interannual variability and stronger long-term decline of natural snowfall (Section 2.2.2), as well as more frequent (and increasing) periods of time when the air temperature conditions are inadequate for snowmaking. Based on studies accounting for snow management, and excluding lowest-lying areas with already often unreliable snow conditions, risks to snow reliability including snowmaking are moderate in most cases for global warming under 2°C since preindustrial (although the required water availability needs to be assessed). Higher risks are encountered above 3°C global warming (Steiger et al., 2017; Spandre et al., submitted). However, assessing climate risks in ski tourism is complex and context-specific, given local climate conditions and other factors related to the demand-side and structural adaptive capacity of ski resorts and their communities. Steiger et al. (2017) also highlight a paucity of studies relevant to ski tourism in Asia and South America.

The relationship between glaciers and tourism has been examined by a limited but growing number of studies (Moreau, 2010; Purdie, 2013; Espiner and Becken, 2014; Welling et al., 2015; Stewart et al., 2016). Glacier retreat has created challenges for local communities, tourist enterprises and government agencies that rely on and promote this type of tourism. Landscape change, increased exposure to hazards and water scarcity are common problems. Diversification across types of tourism products and services, as well as across seasons, has been a commonly advocated adaptation response. Climate change might increase or decrease tourism in high mountain areas in different regions, depending on the tourism product that is adapted or enhanced with changing conditions, for example an increase in number of sunny days in summer in the Alps may be a factor for increased visitation for some segments of the market (Pröbstl-Haider et al., 2015). Glaciers are used to support summertime skiing, particularly in the European Alps. Summer ski resorts operating on glaciers increasingly rely on snow management and snow making on the glacier itself. In recent years several resorts operating on glaciers have stopped summertime operations, due to unfavourable snow conditions and excessive operating costs (e.g., Falk, 2016). Fischer et al. (2016) reported that active snow management reduced the negative glacier mass balance from -0.78 ± 0.04 m w.e. yr^{-1} to -0.23 ± 0.04 m w.e. yr^{-1} over a ten years period on an Austrian glacier used for summer skiing. Attempts to counteract glacial retreat at scales that would cover the entire glacial area have not materialized. Beyond local management of snow on glaciers for summer ski tourism, Oerlemans et al. (2017) quantified the summer snowmaking requirements in the ablation area for limiting the retreat rate of the Morteratsch Glacier (Switzerland), and indicated that this could be effective to reduce snowmelt although further studies are warranted to assess potential side effects. In parts of the European Alps, there is evidence for growing glacier fore fields and the development of outwash fans, as well as increased floods and destabilization of slopes. As a result, in some cases, trails have been closed or re-routed, or infrastructure installed such as bridges and fixed ropes to facilitate access (Wang et al., 2010; Ritter et al., 2012). Some operators of glacier tour companies have shifted to new sites as glaciers become inaccessible after retreat, diversifying to offer other activities or simply reduced their activities (Furunes and Mykletun, 2012). Glacier retreat has also had positive impacts on tourism. For example, in some cases, it has encouraged road development resulting in the diversion of investment funds for tourism (Dangi et al., 2018), or the formation of proglacial lakes, which, despite their hazard potential (Section 2.3.2), have been advocated as a tourist and recreation attraction (Haeberli et al., 2016) or could be used as reservoirs for snowmaking. Some efforts by tourism providers have been made to use glacier retreat as a positive opportunity for awareness-raising, using glacial retreat to attract visitors as ‘last chance’ tourism or to raise awareness about climate change (e.g., in Chacaltaya, Bolivia, where a glacier that supported the world’s highest ski resort disappeared leading to the closure of this resort (Kaenzig et al., 2016). However, these efforts have been met with limited success, as noted for New Zealand (Purdie et al., 2015). In Peru, the Pastoruri Glacier was closed to tourists in 2007 due to safety concerns given its rapid retreat, resulting in reduced visitor numbers (Palomo, 2017). Since then, an improved and better regulated trail was opened (Bury et al., 2011) and labelled the “climate change route’’, to attract tourists to the area again (Palomo, 2017). The national park service and local communities jointly manage the area, which is a frequently visited area given its proximity to paved roads (Rasmussen, 2018).

Changes in the mountain cryosphere are reported to impact mountaineering and alpine climbing practices, particularly in terms of compromised safety along access routes to the mountain and mountain huts, whose foundation stability may also be compromised and therefore need reinforcing (Duvillard et al., 2015). Examples from the literature include cases where rock fall hazard are increasingly experienced and reported by mountaineers in places such as Switzerland (Temme, 2015) and New Zealand (Purdie et al., 2015). Glacier retreat has reportedly led to increased moraine wall instability resulting in the need to abandon and/or create new routes and the need to install ladders and other forms of fixed anchors on the rock to maintain and facilitate access for hikers (Mourey and Ravanel, 2017). Over the course of time and with increasingly changing conditions with retreating glaciers, these installations themselves need replacement and upgrading, consequently requiring assessment on their feasibility as an adaptation option in the long term.

Fluctuations in visitor numbers and changes in visitations patterns over peak tourism seasons are cited in some studies as a potential consequence to climatic and cryospheric change (Pröbstl-Haider et al., 2015; Purdie et al., 2015; Temme, 2015). However, in general, these may be a poor proxy to establish direct potential links to climate change and a changing cryosphere with economic feasibility for these activities, given that other key factors that also determine visitation in high mountain regions for tourism and recreation purposes can play a much more significant role.

In summary, cryospheric changes due to climate change drive changes in tourism and recreation activities in the mountains (ski, glacier tourism, mountaineering) (*medium confidence*) (Figure 2.10). Adaptation measures currently in place attempt to address mainly short-term economic losses to the tourism sector, however the extent to which they are successful in terms of economic viability and benefits to local communities more broadly, is less evident (*limited evidence, medium agreement*). Future cryospheric changes are projected to pose challenges to tourism activities in mountain regions, especially under high-end climate scenarios. Existing structural local adaptation measures (e.g., current snowmaking technologies) are projected to approach their limits around 2°C of global warming since preindustrial (*high confidence*). Tourism activities related to the mountain cryosphere are projected to undergo major changes in the 21st century (*high confidence*), however these may not only be climate change driven but also potential changes in user demand, transportation costs, and other issues related to legal aspects on access, mobility and governance of land may be important factors to consider.

2.3.5 Spiritual and Intrinsic Values, and Human Well-being

Cryosphere changes also impact spiritual and intrinsic values (Batavia and Nelson, 2017), which are held by populations in high mountains and other regions around the world; these impacts often harm human well-being (*medium evidence, high agreement*). Spiritual and intrinsic values can include aesthetic dimensions, which are also an element of tourism and recreation (Section 2.3.4), though they focus more directly on ties to sacred beings or to inherent rights of entities to exist (Daniel et al., 2012). However, they overlap, since the visual appeal of natural landscapes links with a sense of the immensity of mountain landscapes, glaciers and fresh snow (Paden et al., 2013; Gagné et al., 2014). Moreover, different stakeholders, such as local communities, tourists and policy-makers, may place different values on specific cultural services (Schirpke et al., 2016).

Spiritual and intrinsic values often, but not exclusively, rest on deeply-held religious beliefs and other local customs (*medium evidence, high agreement*). In high mountain regions, some communities understand mountains through a religious framework (Bernbaum, 2006). In settings as diverse as the Peruvian Andes, the Nepal Himalaya and the Hengduan Mountains of southwest China, local populations view glacial retreat as the product of their failure to show respect to sacred beings or to follow proper conduct; experiencing deep concern that they have disturbed cosmic order, they seek to behave in closer accord with established traditions, but fear that the retreat may continue, leading to further environmental degradation and to the collapse of natural and social orders (Becken et al., 2013; Gagné et al., 2014; Allison, 2015). In the United States of America, the glaciated peaks of the Cascades have also evoked a deep sense of awe and majesty, and an obligation to protect them (Carroll, 2012; Duntley, 2015). Similar views are found in the Alps, where villagers in the South Tyrol of Italy speak of treating glacier peaks with "respect," and state that glacier retreat is due, at least in part, to humans "disturbing" the glaciers Brugger (Brugger et al., 2013), resulting in what Albrecht et al. (2007) termed solastalgia, a kind of deep environmental distress or ecological grief (Vince and Sale, 2011; Cunsolo and Ellis, 2018).

The populations in mountain regions provide a cultural service to themselves and to wider society through their indigenous knowledge and local knowledge, which contribute to scientific understanding of glaciers and to management of water resources and hazards, and which is threatened by glacier retreat. Their knowledge of glaciers is often tied to their religious beliefs and practices, is based on direct observation (Gagné et al., 2014), stories passed down from one generation to another within community (Stensrud, 2016) and other sources. Residents of mountain areas can provide dates for previous locations of glacier fronts, sometimes documenting these locations through the presence of structures (Brugger et al., 2013). Their observations often overlap with the record of instrumental observations (Deng et al., 2012), and can significantly extend this record (Mark et al., 2010).

An additional cultural value is the contribution of glaciers to the understanding of human history. Glacier retreat has supported the increase of knowledge of past societies by providing access to archaeological materials and other cultural resources that had previously been covered by ice. The discovery of Oetzi, a mummified Bronze Age man whose remains were discovered in 1991 in the Alps near the Italian-Austrian border, marked the beginning of scientific research with such materials (Putzer and Festi, 2014). After 2000, papers began to be published, describing the finds that were uncovered in retreating glaciers and shrinking ice patches in the Wrangell-Saint Elias Range (Dixon et al., 2005), the Rocky Mountains (Lee, 2012) and Norway (Bjørgo et al., 2016). The field has matured recently, with the appearance of global synthesis of the subject (Reckin, 2013) and the establishment of the Journal of Glacial Archaeology in 2014. This field provides new insight into human cultural history and contributes to global awareness of climate change (Dixon et al., 2014). Though climate change permits the discovery of new artefacts and sites, it also threatens these objects and places, since they are newly exposed to harsh environmental conditions (Callanan, 2016).

2.3.6 Impacts on Household Economics, Residence Patterns and Habitability, and National Economies

2.3.6.1 Livelihoods

The cryosphere plays an important role in sustaining livelihoods—the basis of household economies--of mountain communities (Rasul and Molden, in review). The published literature on livelihoods in high mountain regions points to three patterns, found in most regions of the world: a strong reliance on natural resources including cryospheric resources, a diversity of livelihoods to make use of production zones at different elevations, and a strong seasonality of particular livelihoods, as a means to accommodate to the constraints imposed by short growing seasons and seasonal patterns of travel (*very high confidence: high agreement, robust evidence*). Due to high dependency on natural resources, people living in high mountain areas are highly vulnerable to cryospheric changes (McDowell et al., 2014; Carey et al., 2017; Rasul and Molden, in review). The bulk of studies on mountain livelihoods in the context of cryosphere change focus on specific livelihoods, which are discussed elsewhere in this chapter, notably agriculture (Section 2.3.1.3.1), tourism (Section 2.3.4.1) and labour migration (Section 2.3.6.2) (Figure 2.10).

The literature also contains case studies which describe the adoption of new livelihoods as part of responses to climatic and non-climatic stresses, including cryospheric changes, though there is low agreement on overall patterns of these livelihood changes. In some instances, the different livelihood strategies complement each other to support income and well-being. A study of a 2014 debris flow in Nepal found that it temporarily reduced agricultural productivity because of damage to irrigation, and households adopted other livelihood strategies including generating income through wage labour migration (van der Geest and Schindler, 2016). A community of indigenous pastoralists in Bolivia increased such migration to purchase fodder for their animals when glacier retreat reduced streamflow that supports irrigation of pastures (Yager, 2015). A review of migration in the Himalaya and Hindu Kush found that households that participated in labour migration and received remittances had improved adaptive capacity, and lowered exposure to natural hazards (Banerjee et al., 2018).

In other cases, the households and communities which seek to integrate different livelihoods encounter conflicts or incompatibilities between livelihoods, especially when they seek to diversify income through labour migration in contexts where climate-related shocks and uncertainty affect agricultural production and thus reduce income. Sustainable management of land, water and other resources is highly labour intensive, and thus labour mobility constrains and limits the adoption of sustainable practices (Gilles et al., 2013). Moreover, the labour available to a household is differentiated by age and gender, so wage labour migration, often of young males, entails either a loss of capacity to undertake specific tasks or a readjustment of the division of labour (Alata et al., 2018).

2.3.6.2 Residence Patterns and Habitability

Cryosphere changes in high mountain areas have influenced human mobility during this century, and are likely to do so through 2100, by altering water availability and inducing exposure or vulnerability to mass movements and floods and other cryospheric induced disasters (Barnett et al., 2005; Carey et al., 2017; Rasul and Molden, in review). These changes can have negative impacts on livelihoods and on settlements and infrastructure, providing greater incentives to mountain people to engage in temporary or permanent migration. (*limited evidence*, *medium agreement*).

Human mobility is a centuries-old practice in high mountain areas. Transhuman movements between summer and winter pastures of pastoral populations, participation in regional market networks, and regional labour migration are common in Asia, Europe and North and South America (Lozny, 2013). In the face of climate change, other environmental changes including shifting patterns of snow and ice, and demographic, economic, social, cultural and political drivers, patterns of movement have been changing in the recent past and present, and are projected to continue changing, at least in the near term (*high agreement, medium evidence*). However, establishing a causal relationship between cryosphere stressors and human mobility in mountain regions is extremely complex, since decision-making about mobility influenced by interaction of multiple drivers at individual, household and societal levels (*high confidence*).

It is worth noting that the research to date is based on case studies which rely on interviews; though some studies (e.g., Milan et al., 2015) use random sampling, many do not, relying on convenience or chain referral sampling instead. They also show migration on several time scales, including short-term, long-term and permanent migration. Migration is usually described as taking place within the country of origin, and sometimes within the region; however, cases of international migration are also recorded. Moreover, there are no studies that link cryosphere changes to migration based on large sample surveys or census data that tease out statistical linkages among the drivers and migration responses, or that use remote sensing data to study migration.

Though large sample surveys are lacking, some studies examine the relation of rates of emigration and dependence on glacier meltwater rather than from snowmelt, rainfall and groundwater. A study in the central Peruvian Andes examined three different elevation zones in one region, showing that migrants from the highest zone, most dependent on glacier meltwater, travelled further and remained absent longer than from the lower zones (Warner et al., 2012). In another region, the reverse relationship was noted. In the Naryn River drainage in Kyrgyzstan, labour migration, lasting months or years, is more extensive from the downstream communities than the upstream communities, even though the latter rely more directly on surface water, with its large glacier meltwater component from the Tien Shan, for irrigation; this pattern reflects more efficient water management institutions in the upstream communities, which relieves the effects of water scarcity there (Hill et al., 2017) (*medium evidence, low agreement*).

Transhumant pastoralism in high mountain areas is declining, due to climatic and non-climatic factors, including changes in snow distribution and glaciers (*limited evidence, medium agreement*). The climatic stressors include changes in snow and glaciers which adversely affect herders at their summer residences and winter camps in the Himalayas (Namgay et al., 2014) and in Scandinavian mountains (Mallory and Boyce, 2018). Erratic snowfall patterns, as well as a decrease in rainfall, are perceived by herders in Afghanistan, Nepal and Pakistan to have resulted in vegetation of lower quality and quantity (Shaoliang et al., 2012; Joshi et al., 2013; Gentle and Thwaites, 2016). Heavy snowfall incidents in winter caused deaths of a large number of livestock in northern Pakistan in 2009 (Shaoliang et al., 2012). Herders in Nepal reported of water scarcity in traditional water sources along migration routes (Gentle and Thwaites, 2016). However, rising temperatures, with associated effects on snow cover, have some positive impacts. Seasonal migration from winter to summer pastures start earlier in northern Pakistan, and residence in summer pasture lasts longer (Joshi et al., 2013), as it does in Afghanistan (Shaoliang et al., 2012).

Changing water availability, mass movements and floods are cryosphere processes which drive internal and international migration (*medium evidence, high agreement*). In most cases, climate is only one of several drivers (employment opportunities and better educational and health services in lowland areas are others). A debris flow in central Nepal in 2014 led more than half the households to migrate for months (van der Geest and Schindler, 2016). In the Santa River drainage, Peru, rural populations have declined 10% between 1970-2000, and the area of several major subsistence crops also declined (Bury et al., 2013). Research in this region suggests that seasonal emigration within subdrainages of the main Santa drainage increases as subdrainages move from a stage of peak water (Section 2.3.1.1) to later stages, with decreased dry season flow (Wrathall et al., 2014). Studies which project migration emphasize decreased water availability following glacier retreat as a driver in Kyrgyzstan (Chandonnet et al., 2016) and Peru (Oliver-Smith, 2014).

Water availability and natural hazards are cited as causes of spontaneous resettlement, a larger-scale process than labour migration, and one that indicates cryosphere-driven challenges to habitability. In southern Chile, an entire community relocated after a glacier lake outburst flood in 1977 (Iribarren Anacona et al., 2015). A village in western Nepal moved to lower elevation after decreasing snowfall reduced the flow of water in the river on which their pastoralism and agriculture depended (Barnett et al., 2005). A village in northern Pakistan moved to a lower elevation after massive debris flows disturbed irrigation channels, disrupted water supplies and damaged fields and houses (McDonald, 1989).

Two specific themes in the study of cryosphere changes and migration have emerged in the recent literature: age-specific migration patterns and the issue of habitability. In the cases in which the age of migrants is discussed, young adults are reported to migrate more often, though their specific ages are not always stated. These migrants face non-climate drivers as well. Emigration has increased in recent decades from two valleys in highland Bolivia which rely on glacier meltwater, as water supplies have declined, though other factors also contribute to emigration, including land fragmentation, increasing household needs for cash income, the lack of local wage-labour opportunities and a greater interest among the young in educational opportunities located in cities (Brandt et al., 2016). A recent study documents the inter-generational dynamics of emigration from a livestock-raising community in the Peruvian Andes, where glacier retreat has led to reduced flow in streams which support crucial dry-season pasture. Though people 50 years old or older are accustomed to living in the high pasture zones, younger people view livestock-raising as a means of accumulating capital that will facilitate their movement to towns at lower elevations. They invest in improved stock and pasture, but then later sell off their animals. Some retain a fraction of their herds, leaving then with herders who are paid in a share of the increase. The human and animal populations of the communities are shrinking (Alata et al., 2018). In Nepal, young members of high-elevation pastoral households were increasingly engaged in tourism and labour migration (Shaoliang et al., 2012).

The issue of habitability arises in the cases, mentioned above, of communities which relocate after floods or debris flows destroy houses and irrigation infrastructure, or damage fields and pastures. It occurs as well in the cases of households with extensive long-term migration, where agricultural and pastoral livelihoods are undermined by the cryospheric change (Barnett et al., 2005). In addition, the loss of cultural values, including spiritual and intrinsic values, (Section 2.3.5) can contribute to decisions to migrate (Kaenzig, 2015). Combined with the patterns of permanent out-migration, this issue of habitability raises issues of the limits to adaptation in mountain areas (Huggel et al., 2018).

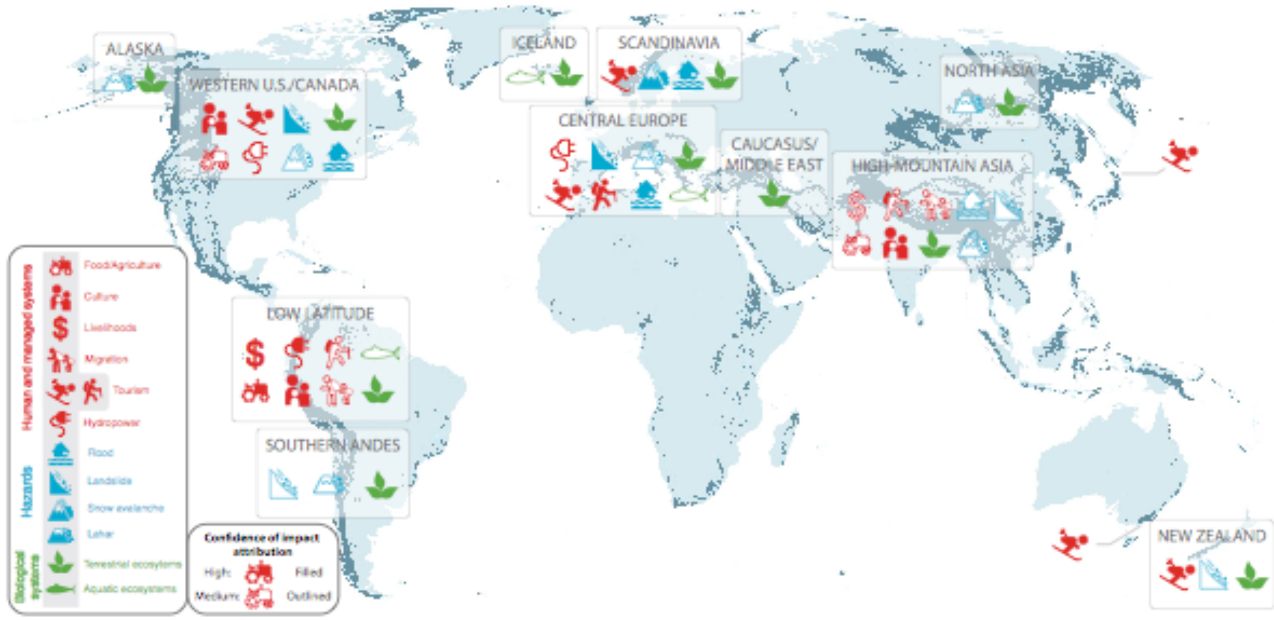


Figure 2.10: Biophysical and human sectors that have experienced an impact in the past several decades that can be partly attributed to cryospheric changes. Only impacts documented in the scientific literature are shown (see Appendix 2.A, Table 5). Confidence levels refer to confidence in attribution to cryospheric changes.

2.3.6.3 Economic Impact

Though the literature on economic loss of cryospheric changes is scanty, recently a few studies have reported the potential economic impact of cryospheric changes (Vergara et al., 2007; Gabbi et al., 2012; Gaudard et al., 2014; Sturm et al., 2017). The economic losses are incurred through two pathways - economic loss and damage due to climate induced natural disasters and through the additional risk and loss of potential opportunities or additional investment would be necessary to manage or adapt with the challenges brought about by the cryopsheric changes. Economic loss and damages due to cryospheric induced disasters are reported in many high mountain regions and they are projected further to increase which will require additional costs in risk reduction measures. For instance, the Zhangzangbo glacier outburst flood in Tibet, China, in 1981 killed 200 people and damaged infrastructure and property extensively in Tibet and Nepal, with estimated economic losses of US$ 456 million (Mool et al., 2001). Similarly, Dig Tsho flood in the Khumbu Himal of Nepal in 1985 has damaged a hydropower plant and other properties, with estimated economic losses of US$ 500 million (Shrestha et al., 2010). An adaptation cost to dig a channel in Tsho Rolpa glacier in Nepal that lowered a glacial lake costed US$ 3 million in 2002 (Bajracharya, 2010).

The energy sector is likely to suffer hugely due to additional risk and loss of potential opportunities. In Peru, the cost of glacier shrinkages for the energy sector was estimated to be about US$ 740 million annually (Vergara et al., 2007). Likewise, an additional cost of US$ 100 million will be required for the Peruvian government to purify the deteriorating water quality resulted due to cryospheric changes.

2.4 Key Knowledge Gaps, Global Policy Frameworks and Pathways to Sustainable Development

2.4.1 Key Gaps in Knowledge and Evidence

2.4.1.1 Observations, Detection and Attribution, Projections

Observations of atmospheric conditions especially at high elevations, and cryospheric elements in high mountain regions are still limited. Elevation Dependent Warming (EDW) is a key attribute and area for mountain-specific climate monitoring and observation, the understanding of which is still limited. Trends in total or solid precipitation at high altitude are highly uncertain, due to intrinsic uncertainties of in-situ observation methods, and large variability. Studies of snow cover in Asia and Andes are mostly restricted to satellite-borne measurements, which span limited time coverage and lack evaluation from in-situ observations (Section 2.2.2). Furthermore, there is limited evidence on the potential added value of indigenous knowledge and local knowledge to complement observations (instruments, models) (Section 2.2.3.1).

There is high uncertainty on projections at higher altitude due mostly to the inability of regional climate models and downscaling methods to capture the subtle interplays between large-scale climate change and local phenomena influenced by complex topography (Section 2.2.1). Coarse-scale simulations of future permafrost conditions are mostly of limited use due to coarse spatial resolution or lacking representation of topographic effects (Section 2.2.4.2). For glacier retreat, and its link to destabilizing adjacent debris and rock slopes, incomplete knowledge of past events in remote mountains and the influence of other variable local and regional factors, means incomplete evidence on how current global glacier retreat influences the frequency and magnitude of such instabilities (Section 2.3.2.1). Radiative forcing effects of light absorbing impurities and understanding their spatiotemporal dynamics is also a key knowledge gap for the attribution of changes in high mountain snow and ice and the identification of regional feedbacks.

Overall, except for recent studies on glacier mass balance and glacier runoff largely driven by a need to estimate glacier contribution to sea-level rise (e.g., Huss and Hock, 2018), very few studies have addressed the past and future evolution of cryospheric components in a homogeneous way across all high mountains of the world. This is largely due to the regional relevance of these issues, which in general do not require the development of approaches operating at the global scale. However, this limits our ability to assess and compare the intrinsic evolution of these components in a harmonized manner, and their impacts, policy relevance, and adaptation options. Improved cross-disciplinary studies bringing together current observation and modelling approaches in each specific field hold potential to address this gap in the future, especially in relationship to water resources and water-related hazards (regional climate, glacier, snow cover, surface hydrology). This also applies to impact studies for sectors under the influence of a smaller number of cryospheric elements (hazards related to a changing permafrost, ski tourism, snow avalanches, rain-on-snow events, impacts on ecosystems) which currently are mostly addressed at the regional, if not local scale, and lack global comparisons and perspectives. Detection and attribution studies specifically targeting mountain cryosphere and associated sectors are scarce, and this forms a critical knowledge gap for climate change assessments in high mountain regions.

2.4.1.2 Impacts, Vulnerability, Risk, Resilience and Adaptation

Impacts and associated disasters from a changing cryosphere, as manifestation of climate risk in high mountains, are evident through experiences with water scarcity and with changes in frequency and/or magnitude of hazards, combined with societal and ecosystems dynamics and pre-conditions which shape the exposure and vulnerability of social-ecological systems (Section 2.3.2.3). However, at global scales, few studies have taken a comprehensive risk approach to ascertain and systematically characterise and compare impacts across high mountain regions, particularly risk assessments that consider all underlying components of climate risk, including compounded risks and cascading impacts where instances of deep uncertainty in responses and outcomes may arise (Section CCB-4). Specifically for the hydropower sector, sufficient data basis, rigorous modelling of available production capacity in future and incorporation of acquired knowledge into operational procedures remain challenging and need to be addressed (Section 2.3.1.2).

While adaptation measures are reported in the literature, gaps exist in systematic and comparative ex-post assessments and evaluations of adaptation measures and their intended effects (McDowell et al., 2014; Rasul and Molden).

2.4.2. High Mountains, Global Policy Frameworks, and Climate-resilient Development Pathways

There is *limited evidence* on specificities of high mountain regions in responding to key global frameworks designed to steer and guide action on climate change and sustainable development. Relevant global frameworks include the Paris Agreement (UNFCCC, 2015), UN 2030 Agenda and its Sustainable Development Goals (SDGs) (UN, 2015), and the Sendai Framework for Disaster Risk Reduction (UNISDR, 2015).

In international climate policy, addressing the negative impacts of climate change is articulated in the Paris Agreement under Article 8, more specifically depicted as 'Loss and Damage' (UNFCCC, 2015). Despite evident impacts of climate change on the mountain cryosphere, there is *limited evidence* or references in the literature to Loss and Damage in this context, lacking considerable knowledge on how Loss and Damage in the mountain cryosphere might be conceptualized, categorized, and assessed (Huggel et al., 2018)). In high mountain regions, the already committed and unavoidable climate change due to delayed response of glaciers to climatic stimuli are relevant aspects to consider for Loss and Damage (Huggel et al., 2018)).

For monitoring and reporting on progress towards sustainable development through the implementation of the SDGs, in particular SDG 13 on Climate Action (UN, 2015), the disaggregation of data for SDG indicators and targets at subnational scales is considered a key challenge for high mountain regions, which cover areas both within country boundaries and/or across borders in transboundary contexts (Rasul and Tripura, 2016). This is a reported issue for some countries where SDG indicator data is relatively underdeveloped in these countries, requiring use of proxies that make it challenging to compare progress between mountain regions (Bracher et al., 2018). A further challenge for mountain regions, is a lack of commonly agreed definitions to delineate mountain areas, requiring countries to apply their own definitions in demarcating these areas within their boundaries, which may limit efforts to report on progress on SDGs on mountains in transboundary contexts (Bracher et al., 2018). On substance, the economic performance of livelihood enterprises, coupled with robust socioeconomic data for mountain systems, are still lacking in many parts of the world, compromising the ability for meaningful comparison and aggregation of data and knowledge for monitoring and reporting on progress on SDGs at regional or global scales (Gratzer and Keeton, 2017).

Disasters associated with natural hazards in high mountains are placing many communities and their sustainable development at risk. The Sendai Framework for Disaster Risk Reduction 2015–2030 (UNISDR, 2015) offers a guiding framework under which risks, including climate change, can be accounted for and addressed at national scales. However, there is *limited evidence* based on studies that report on the specificities of high mountains in monitoring and reporting on progress on priorities and targets set out in the Sendai Framework (Wymann von Dach et al., 2017), particularly in systematically reporting on root causes of disaster risks in high mountains and associated compounded risks and cascading impacts.

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Appendix 2.A: Supplementary Material

2.A.1 Details of Studies on Temperature Observations and Projections

Appendix 2.A, Table 1: Overview of studies providing evidence for past and projected trends in temperature, per high mountain region (as defined in Figure 2.2)

Region	Domain	Variable	Change	Time period	Scenario	Method	Reference
Alaska	N America >55°N	Annual mean temperature	+0.61 °C/ decade	1961-1990 to 2070-2099	SRES A1F1	Downscaled GCMs	Nogués-Bravo et al. (2007)
	N America >55°N	-	+0.35°C/ decade	1961-1990 to 2070-2099	SRES B1	Downscaled GCMs	Nogués-Bravo et al. (2007)
Western U.S.A./ Canada	Colorado and Pacific Northwest	Minimum daily temperature	<0.37°C/dec at low elevations >0.75°C/dec at highest elevations (>4000 m)	1979 - 2006	Past	In-situ	Diaz and Eischeid (2007)
	NE U.S.A.	Mean temperature	+0.35°C/dec Mt Washington (1905 m) +0.31°C/dec Pinkham Notch (613 m)	1970 - 2005	Past	In-situ	Ohmura (2012)
	NW U.S.A.	Mean temperature	+0.11°C/dec above 2000m	1991 - 2012	Past	In-situ	Oyler et al. (2015)
	Colorado Rockies	Spring temperature (April)	up to +1°C/dec	1995-2005 to 2045 - 2055	SRES A2	Pseudo-GW runs: RCMs	Letcher and Minder (2015)
	Whole N America	Mean temperature	+0.14°C/dec >500 m	1948-1998	Past	In-situ	Pepin and Seidel (2005)
	N America <55°N	Annual mean temperature	+0.49°C/ dec	1961-1990 to 2070-2099	SRES A1F1	Downscaled GCMs	Nogués-Bravo et al. (2007)
	N America <55°N	Annual mean temperature	+0.27°C/dec	1961-1990 to 2070-2099	SRES B1	Downscaled GCMs	Nogués-Bravo et al. (2007)
Iceland	Full domain	Mean annual temperature	+0.21 to 0.40°C/dec depending on GCM/RCM	2000 - 2100	RCP8.5	GCM/ RCM pairs	Gosseling (2017)
Central Europe	Swiss Alps	Mean annual temperature	+0.35°C/dec	1959 - 2008	Past	In situ	Ceppi et al. (2012)
	-	Autumn temperature	+0.17°C/dec	-	Past	-	-
	-	Summer temperature	+0.48°C/dec	-	-	-	-
	Switzerland	Mean annual temperature	+0.13°C/dec	1864 - 2016	Past	In-situ	Begert and Frei (2018)
	Jungfraujoch, 3580 m	Mean annual temperature	+0.43°C/dec	1970 - 2011	Past	In-situ	Ohmura (2012)
	Sonnblick, 3109 m	Mean annual temperature	+0.30°C/dec	1980 - 2011	Past	In situ	Ohmura (2012)

	European Alps	Mean annual temperature	+0.25°C/dec	1961-1990 to 2021-2050	SRES A1B	Downscaled GCMs	Gobiet et al. (2014)
		-	+0.36°C/dec	1961-1990 to 2069-2098	-	-	-
Scandinavia	Whole area	Summer or winter temperature	+0.45°C/dec low elevation (<500m) in winter, +0.27°C/dec at high elevation (~1500 m) in summer	1961-1990 to 2070-2099	SRES A1B	RCM ensembles	Kotlarski et al. (2015)
	Whole area	Mean annual temperature	+0.54°C/dec	1961-1990 to 2070-2099	SRES A1F1	Downscaled GCMs	Nogués-Bravo et al. (2007)
		-	+0.31°C/dec	1961-1990 to 2070-2099	SRES B1	Downscaled GCMs	-
Caucasus/ Middle East	Whole area	Mean annual temperature	+0.14°C/dec	1958-2000	Past	In-situ	Diaz et al. (2003)
North Asia	Whole area	Mean annual temperature	+0.76°C/dec	1961-1990 to 2070-2099	SRES A1F1	Downscaled GCMs	Nogués-Bravo et al. (2007)
	-	-	0.43°C/dec	1961-1990 to 2070-2099-	SRES B1	Downscaled GCMs	-
Southern Andes	18°S to 42°S	Mean annual temperature	-0.05°C/dec	1981-2010	Past	In-situ	Vuille et al. (2015)
	Central Andes 10°S to 25°S	Mean annual temperature in the free atmosphere (500 hPa)	+0.16°C/dec to +0.41°C/dec	1979-2008	Past	Reanalyses	Russell et al. (2017)
	Whole area	Mean annual temperature	+0.34°C/dec	1961-1990 to 2070-2099	SRES A1F1	Downscaled GCMs	Nogués-Bravo et al. (2007)
	-	-	+0.18°C/dec	1961-1990 to 2070-2099	SRES B1	Downscaled GCMs	-
Low Latitude (including tropical Andes)	Tropical Andes	Mean temperature	+0.09°C/dec	1958 - 2000	Past	In-situ	Diaz et al. (2003)
	La Paz (Bolivia)	Mean temperature	-0.70°C/dec	1985 - 2010	Past	In-situ	Ohmura (2012)
	Tropical Andes	Mean temperature	0.3°C/dec	1961 - 2000 to 2080 - 2100	RCP8.5	Downscaled GCMs	Vuille et al. (2018)
	Bolivian Andes	Mean temperature	+2.7°C to +3.2°C (i.e. 0.34°C/dec to 0.4°C/dec)	1950 - 2000 to 2040 - 2069	SRES A1B	Downscaled GCMs	Rangecroft et al. (2016)
	-	-	+4.2°C to +4.9° (i.e. 0.38°C/dec to 0.44°C/dec)	1950 - 2000 to 2070 - 2099	-	-	-

	East Africa	Mean temperature	+0.18°C/dec	1958 – 2000	Past	In-situ	Diaz et al. (2003)
	South and East Africa	Mean temperature	+0.14°C/dec >500 m	1948-1998	Past	In-situ	Pepin and Seidel (2005)
High-mountain Asia	Hindu-Kush-Himalaya	Mean temperature	+0.1°C/dec	1901 - 2014	Past	In-situ	Krishnan et al. (2018)
	-	-	+0.2°C/dec	1951 - 2014	-	-	-
	Mukteshwar (2311 m), India	Mean temperature	+0.48°C/dec	1980 - 2010	Past	In-situ	Ohmura (2012)
	Toutouhe (4535 m), China	Mean temperature	+0.02°C/dec	1970 - 2005	Past	-	-
	Himalaya	Mean temperature	+0.06°C/dec	1958 - 2000	Past	Reanalysis	Diaz et al. (2003)
	Tibetan Plateau	Mean temperature, wet season (MJJAS)	+0.4°C/dec	1979 - 2011	Past	In-situ	Gao et al. (2015)
	-	Mean temperature, dry season (Oct-Apr)	+0.54°C/dec	-	-	-	-
	Tibetan Plateau	Mean temperature	+0.69°C/dec (>3000 m)	1981 - 2006	Past	In situ	Qin et al. (2009)
	Tibetan Plateau	-	+0.55°C/dec (1000-3000 m)	-	-	-	-
	Himalaya/ Tibetan Plateau,	Minimum Temperature (Winter)	+0.32°C/dec (~1600 m) to +0.75°C/dec (~4100 m)	1971-2000 to 2071-2100	RCP8.5	CMIP5	Palazzi et al. (2017)
	Hindu-Kush Himalaya	Mean temperature	+0.6°C/dec (winter) +0.54°C/dec (summer)	1976-2005 to 2066-2095	RCP8.5	RCMs	Sanjay et al. (2017)
	Himalaya	Mean temperature	+0.57°C/dec (winter) +0.45°C/dec (summer)	1970-2005 to 2070-2099	RCP8.5	RCMs	Dimri et al. (2018)
	Tibetan Plateau peak warming 4500–5000 m	Minimum Temperatures	+0.85°C/dec (winter) +0.53°C/dec (annual)	1961-2006	Past	In situ	Liu et al. (2009)
	Tibetan Plateau >2000 m	Mean temperature	+0.16°C/dec +0.32°C/dec (winter)	1955-1996	Past	In situ	Liu and Chen (2000)
	Tibetan Plateau around 4500 m	Mean temperature	+0.65°C/dec	2006-2050	RCP8.5	Downscaled RCMs	Guo et al. (2016)
	Tibetan Plateau	-	+0.51°C/dec	2006-2050	RCP8.5	-	-

	2000-2200 m						
	Tibetan Plateau >2000 m	Mean temperature	+0.28°C/dec	1961-2007	Past	In situ	Guo et al. (2012)
New Zealand and SE Australia	New Zealand	Annual mean temperature	+0.33 °C/ decade	1961-1990 to 2070-2099	SRES A1F1	Downscaled GCMs	Nogués-Bravo et al. (2007)
	-	-	+0.17°C/dec	1961-1990 to 2070-2099	SRES B1	Downscaled GCMs	-
	Australia	Mean temperature	+0.16°C/dec (>500 m)	1948-1998	Past	In situ	Pepin and Seidel (2005)
Japan	Fuji San (3775m), Japan	Mean air temperature	+0.35°C/dec	1985 - 2005	Past	In-situ	Ohmura (2012)

2.A.2 Details of Studies on Precipitation Observations and Projections

Appendix 2.A, Table 2: Overview of recent studies providing evidence for past and projected trends in precipitation, per high mountain region (as defined in Figure 2.2)

Region	**Domain**	**Variable**	**Change**	**Time period**	**Scenario**	**Method**	**Reference**
Alaska	Alaska	Total annual precipitation	Increase from +8% to +40%, depending on the region	1949 - 2016	Past	In-situ, 18 stations	Wendler et al. (2017)
Alaska	South and Southeast Alaska	Snow day fraction	-15% to +7%	1970 - 1999 to 2040 - 2069	RCP4.5	Statistically downscaled GCMs	Littell et al. (2018)
	-	-	-25% to +4%	-	RCP8.5	-	-
	-	-	-22 % to 4 %	1970 - 1999 to 2070 - 2099	RCP4.5	-	-
	-	-	- 41 to -6 %	-	RCP8.5	-	-
Western U.S.A./ Canada	California	Winter precipitation	Insignificant	1920 - 2014	Past	Gridded dataset based on in-situ data	Mao et al. (2015)
	Western US, "Warm mountain sites"	Snowfall amount	-70% to -35 %	1950 - 2005 to 2040 - 2069	RCP8.5	Statistically downscaled GCMs	Lute et al. (2015)
	Western US, "Cold mountain sites"	-	-20 % to - 5 %	-	-	-	-
	Western US, "Warm mountain sites"	90% percentile of snowfall events	-30 %	-	-	-	-

	Western US, “Cold mountain sites”	90% percentile of snowfall events	+5 %	-	-	-	
	Southern California	Total winter snowfall	-40 % (1500 - 2000 m) -22% (2000-2500 m) -8% (above 2500 m)	1981 - 2000 to 2041 - 2060	RCP2.6	Multiple downscaled GCM	Sun et al. (2016)
	-	-	-52 % (1500 - 2000 m) -28% (2000-2500 m) -11% (above 2500 m)	-	RCP8.5	-	-
	-	-	-43 % (1500 - 2000 m) -26% (2000-2500 m) -13% (above 2500 m)	1981 - 2000 to 2081 - 2100	RCP2.6	-	-
	-	-	-78 % (1500 - 2000 m) -48% (2000-2500 m) -18% (above 2500 m)	-	RCP8.5	-	-
	Canada	Ratio of snowfall to total precipitation	Decrease, more pronounced in Western Canada	1948 - 2012	Past	In-situ	Vincent et al. (2015)
	Western Canada	Winter precipitation	+ 11%	1979 - 1994 to 2045 - 2060	RCP8.5	Downscaled GCM	Erler et al. (2017)
	-	-	+ 17%	1979 - 1994 to 2085 - 2100	-	-	-
Iceland	Whole area	Winter precipitation	Insignificant	1961 - 2000	Past	Reanalysis and in-situ	Crochet (2007)
	Whole area	Total precipitation	Insignificant	1981 - 2000 to 2081 - 2100	RCP4.5, RCP8.5	RCMs	Gosseling (2017)
Central Europe	European Alps	Total precipitation	Insignificant, dominated by internal variability	1901–2008	Past	Gridded product based on in-situ data	Masson and Frei (2016)
	Swiss Alps	Fraction of days with snowfall over days with precipitation (annual)	-20 % below 1000m, -10% to -20% between 1000m and 2000m, -5% above 2000m	1961 - 2008	Past	In-situ	Serquet et al. (2011)
	-	Fraction of days with snowfall over days with	-30 to -50 % below 1000m, -10% to -30%	-	-	-	-

		precipitation (spring)	between 1000m and 2000m, -5% to -10% above 2000m				
	Pyrenees	Total precipitation	Insignificant decrease (-0.6%/ Decade)	1910-2013	Past	In-situ	López-Moreno (2005)
	Carpathian	Number of days per year with precipitation > 20 mm	Increase (>7)	1960 - 2010	Past	In-situ	Kohler (2014)
	Carpathian mountains	Summer mean Precipitation	Decrease by up to -20 mm per month	1971-2000 to 2071 -2100	RCP8.5	RCMs	Alberton et al. (2017)
	Greater Alpine Region (GAR)	Winter precipitation	+12.3%	1971-2000 to 2071-2100	RCP4.5	5 EUROCORDEX RCMs	Smiatek et al. (2016)
		Spring precipitation	+5.7%	-	-	-	-
		Summer precipitation	-1.7%	-	-	-	-
		Fall precipitation	+2.3%	-	-	-	-
	-	Number of days with precipitation > 15 mm	+10.9%	-	-	-	-
	Alpine Region	Mean winter (DJF) precipitation	+8 %	1981 - 2010 to 2020 - 2049	RCP4.5	EUROCORDEX RCMs (0.11°)	Rajczak and Schär (2017)
	-	-	+6 %	-	RCP8.5	-	-
	-	-	+12 %	1981 - 2010 to 2070 - 2100	RCP4.5	-	-
	-	-	+17%	-	RCP8.5	-	-
	Alps	Annual solid precipitation amount	-25 %	1981 - 2010 to 2070 - 2099	RCP4.5	EUROCORDEX RCMs (0.11°)	Frei et al. (2018)
	-	-	-45%	-	RCP8.5	-	-
	Pyrenees, below 1500 m elevation	Frequency and intensity of heavy snowfall events	Decrease	1960 - 1990 to 2070 - 2100	SRES A2	Dynamically downscaled GCM	López-Moreno et al. (2011)
	Pyrenees, above 2000 m elevation	-	Insignificant except at high altitude (+30% increase)	-	-	-	-
	Pyrenees, above 2000 m elevation	-	+20-30%	-	SRES B2	-	-

Scandinavia	Finland	Annual snowfall over total precipitation ratio	Decrease (-1.9% per decade)	1909 - 2008	Past	In-situ	Irannezhad et al. (2017)
	Scandinavian mountain (high elevation)	Annual snowfall	20% Increase	1961 – 1990 to 2071 -2100	SRES A1B	Multiple RCM	Räisänen and Eklund (2012)
Caucasus/ Middle East	Greater Caucasus	Total precipitation	-9 kg m^{-2} yr^{-1}	1936 - 2012	Past	In-situ	Elizbarashvili et al. (2017)
	Adjara mountains	-	+6 kg m^{-2} yr^{-1}	-	-	-	-
North Asia	Middle and East Tian Shan	Snowfall fraction	Decrease, from 27% in 1960–1969 to 25% in 2005–2014	1960 -2014	Past	In-situ	Chen et al. (2016)
	West Tian Shan	Winter Total precipitation	+23%	1960 - 2014	Past	In-situ	-
	Northern Tian Shan	Total precipitation	+5 %	1976 - 2005 to 2070 - 2099	RCP8.5	CMIP5 GCMs	Yang et al. (2017)
	Western Tianshan and northern Kunlun Mountains	Solid precipitation	- 26.5%	-	-	-	-
Southern Andes	Chile and Argentina	Annual precipitation	General decrease (up to ~ -6 kg m^{-2} yr^{-1}) with positive values in the southwest corner of the region	1979 - 2010	Past	In-situ stations and gridded, reanalyses	Rusticucci et al. (2014)
	-	DJF precipitation trend	Insignificant (model mean)	2006-2100	RCP4.5	5 CMIP5 GCMs	Zazulie et al. (2018)
			Significantly positive only for one GCM in RCP8.5		RCP8.5	-	-
		JJA precipitation trend	Insignificant (model mean).	-	RCP4.5	-	-
		-	Significantly negative only for two	-	RCP8.5	-	-

			GCMs in RCP8.5				
Low Latitude (including tropical Andes)	Claro River (Colombian Andean Central mountain range)	Annual precipitation	Insignificant	1981 - 2003	Past	In-situ	Ruiz et al. (2008)
	47 mountain protected areas in five National Parks in the tropical belt (30S-30N, including Central America, South America, Africa, South Asia, Southeast Asia)	Annual precipitation	Insignificant, except decrease in Africa	1982 - 2006	Past	In-situ	Krishnaswamy et al. (2014)
	Tropical Andes	Annual precipitation	Geographically variable. Precipitation increase up to ~2000 m. No significant changes in eastern slope above 2000 m, decrease in the western slope above 4000 m	1961-1990 to 2071 - 2100	SRES A2, B2	Downscaled GCM	Urrutia and Vuille (2009)
	Central Andes	Annual precipitation	-19% to -33%	1961 - 2010 to 2071 - 2100	RCP8.5	Large GCM ensemble, relationship to synoptic wind	Neukom et al. (2015)
High-mountain Asia	HKK	DJFMA precipitation	Insignificant	Depending on the dataset: 1998–2010; 1979–2010; 1951–2007; 1950–2009; 1950–2009; 1979–2010	Past	Gridded observations and reanalysis	Palazzi et al. (2013)

	Himalaya	JJAS precipitation	-0.010 kg m^{-2} d^{-1} yr^{-1}	Depending on the dataset: 1951–2007	Past	Gridded observations and reanalysis	Palazzi et al. (2013)
		-	-0.021 kg m^{-2} d^{-1} yr^{-1}	1950–2009	-	-	-
	Himalaya	Summer Precipitation	+0.008 to 0.014 kg m^{-2} d^{-1} yr^{-1}	2006-2100	RCP8.5	GCM multi-member ensemble	-
	Karakoram	Winter precipitation	Significant increasing trend	1961 - 1999	Past	In-situ	Archer and Fowler (2004)
	Monsoon-dominated regions, easternmost Himalayas	Annual precipitation trend	-13.7 ± 2.4 kg m^{-2} yr^{-1}	1994-2012	Past	In-situ	Salerno et al. (2015)
	-	Precipitation during monsoon months	-9.3 kg m^{-2} yr^{-1}	-	-	-	-
	Northwestern Indian Himalayas	Snowfall fraction	Significant decreasing trend (3 out of 7 stations)	1991-2005	Past	In-situ	Bhutiyani et al. (2010)
	-	Winter precipitation trend	Increasing but statistically insignificant	1866 - 2006	-	-	-
	-	Monsoon and annual precipitation trend	Significant decreasing	-	-	-	-
	Tibetan Plateau	Annual precipitation	+1.43 kg m^{-2} yr^{-1}, large spatial variations	1960 - 2014	Past	In-situ	Deng et al. (2017)
	Northern Altai	Annual precipitation	-0.14 kg m^{-2} yr^{-1}	1966 - 2015	Past	In-situ	Zhang et al. (2018)
	Southern Altai	-	+0.89 kg m^{-2} yr^{-1}	-	-	-	
	Hengduan Mountain region	Annual precipitation	Insignificant decrease	1961 - 2011	Past	In-situ	Xu et al. (2018)
		Springtime precipitation	Insignificant increase in springtime	-	-	-	-
	Hindu Kush-Himalaya	Precipitation >95th, precipitation intensity	Insignificant	1960-2000	Past	In-situ	Panday et al. (2015)

	Eastern Himalaya	Annual precipitation	Increase of 15–27% (max. contribution in summer)	1970–1999 to 2070–2099	SRES B1, A1B, A2 and RCP8.5	CMIP3 and CMIP5 GCMs	-
	Western Himalaya-Karakoram	Annual precipitation	Increase of 1–5% (due to increase in winter precip.)	-	-	-	-
	Northwest Himalaya and Karakoram	Precipitation	-0.1% (JJAS) 7% (DJFMA)	1976 - 2005 to 2036 - 2065	RCP4.5	CORDEX RCM	Sanjay et al. (2017)
	-	-	3.5% (JJAS) 14.1% (DJFMA)	1976 - 2005 to 2066 - 2095	-	-	-
	-	-	3.7% (JJAS) 12.8% (DJFMA)	1976 - 2005 to 2036 - 2065	RCP8.5	-	-
	-	-	3.9% (JJAS) 12.9%(DJFMA)	1976 - 2005 to 2066 - 2095	-	-	-
	Central Himalaya	Precipitation	4.4% (JJAS) -0.7% (DJFMA)	1976 - 2005 to 2036 - 2065	RCP4.5	-	-
	-	-	10.5% (JJAS) 1.5%(DJFMA)	1976 - 2005 to 2066 - 2095	-	-	-
	-	-	9.1% (JJAS) -1.3% (DJFMA)	1976 - 2005 to 2036 - 2065	RCP8.5	-	-
	-	-	19.1% (JJAS) -8.8% (DJFMA)	1976 - 2005 to 2066 - 2095	-	-	-
	Southeast Himalaya and Tibetan Plateau	Precipitation	6.8% (JJAS) 3.1% (DJFMA)	1976 - 2005 to 2036 - 2065	RCP4.5	-	-
	-	-	10.4% (JJAS) 3.7% (DJFMA)	1976 - 2005 to 2066 - 2095	-	-	-
	-	-	10.2% (JJAS) 0.9% (DJFMA)	1976 - 2005 to 2036 - 2065	RCP8.5	-	-
	-	-	22.6% (JJAS) 0.6%(DJFMA)	1976 - 2005 to 2066 - 2095	-	-	-
	Tibetan Plateau	Total precipitation change	+3.2%	1961 - 2005 to 2006 - 2035	RCP2.6, RCP8.5	CMIP5 GCMs	Su et al. (2013)

	Tibetan Plateau	Total precipitation change	+6%	1961 - 2005 to 2036 - 2099	RCP2.6	-	-
	-	-	+12%	-	RCP8.5	-	-
New Zealand and SW Australi a	New Zealand	Total precipitation amount	Absence of marked trends, seasonally and geographical ly variable	1900 - 2010	Past	In-situ	Caloiero (2014); Caloiero (2015)
	SW Australia	Total annual precipitation	Reduction since 1970s	1900 - 2010	Past	In-situ	Grose et al. (2015)
	-	-	-5 % (high variability)	1950 - 2005 to 2020 - 2039	RCP2.6	Downsca led GCMs	-
	-	-	-5 % (high variability)	-	RCP8.5	-	-
	-	-	-5 % (high variability)	1950 - 2005 to 2080 - 2099	RCP2.6	-	-
	-	-	-10 % (high variability)	-	RCP8.5	-	-
Japan	Japan	Intense precipitation	+30 % per century	1898 - 2003	Past	In-situ	Fujibe et al. (2005)
		Weak precipitation	-20% per century	-	-	-	-
	Japan (Tokai region)	99th percentile of daily precipitation	From +10% to +50% in DJF	1984 - 2004 to 2080 - 2100	RCP8.5	Single dynamic ally downscal ed GCM (MRI AGCM)	Murata et al. (2016)
	Central Japan	Winter (NDJFM) Snowfall	Decrease in most parts of Japan (up to -300 kg m^{-2}) increase in the central part of northern Japan.	1950 - 2011 to 2080 - 2099	+4°C warming in 2080–2099 with respect to 1861–1880, under RCP8.5	MRI-AGCM3. 2 (dynamic ally downscal ed)	Kawase et al. (2016)
		Heavy snowfall (10 years return period)	Increase (10 mm) in the inland areas of Central and in northern Japan.	-	-	-	-

2.A.3 Details of Studies on Past and Future Changes of Seasonal Snow Cover

Appendix 2.A, Table 3: Synthesis of recent studies providing evidence for past and future changes of seasonal snow cover in high mountain areas, per high mountain region (as defined in Figure 2.2).

Region	Domain	Variable	Change	Time period	Scenario	Method	Reference
Alaska	Whole area	Duration	Decrease	20th century	Past	Remote sensing	Brown et al. (2017)
	-	SWE	Decrease	20th century	Past	-	-
	Mountainous Alaska	Snow well above mean snowline elevation	Increase	1840 - present	Past	Indirect evidence from glacier accumulation	Winski et al. (2017)
	Mountainous Alaska	SWE	-10 to -30%	1970 - 1999 to 2040 - 2069	RCP8.5	Multiple RCMs	Littell et al. (2018)
	-	SWE	-40 to -60%	1970 - 1999 to 2070 - 2099	-	-	-
Western U.S.A./ Canada	Western U.S.A.	Springtime SWE	Decrease for 92% stations	1955 - present	Past	In-situ observations	Mote et al. (2018)
	-	April 1st SWE	-15 to -30%	1955 - present	-	-	-
	Canada	Duration	2 to 12 days per decade	1950 - 2012	Past	In-situ observations	DeBeer et al. (2016)
	Western U.S	April 1st SWE	- 50%	1965 - 2005 to 2010 - 2040	RCP8.5	Multiple RCMs	Naz et al. (2016)
	-	Duration	-10 to -100 days	1976 - 2005 to 2071 - 2100	RCP8.5	-	Musselman et al. (2018)
Iceland	Whole area	Duration	0 to 10 days per decade	1980 - 2010	Past	Remote sensing	Brown et al. (2017)
	Low elevation	Snow depth	-100%	1981 - 2000 to 2081 - 2100	RCP8.5	Single RCM	Gosseling (2017)
	Top of central Vatnajökull	Snow depth	+20%	1981 - 2000 to 2081 - 2100	-	-	-
Central Europe	European Alps and Pyrenees	Snow depth	Decrease below mean snow line elevation, step decrease in late 1980s	Mid 20th - present	Past	In-situ, reanalyses	Beniston et al. (2018) Reid et al. (2016)
	European Alps	SWE	Decrease below mean snow line elevation, step decrease in late 1980s	Mid 20th - present	Past	In-situ	Marty et al. (2017b)
	European Alps	Duration	Insignificant trend, decrease at low elevation (700 to 900 m) in the SE and SW Alps	1985-2011	Past	Optical remote sensing	Hüsler et al. (2014)
	Swiss Alps	Onset date	12 days later	1970 – 2015	Past	In-situ	Klein et al. (2016)
	-	Melt-out date	26 days earlier	-	-	-	-

	Austrian Alps (500 to 2000 m elevation)	Snow cover days	-13 to -18 depending on the region	1950-1979 to 1980-2009	Past	Modelling based on in-situ observations	Marke et al. (2018)
	Austrian Alps (2000 to 2500 m)	-	-12 to -14 depending on the region	-	-	-	-
	Austrian Alps (above 2500 m)	-	-20 (central Austria)	-	-	-	-
	Franch Alps, 1800 m altitude	Snow cover duration	-24 days	1958-2009	Past	Local reanalysis	Durand et al. (2009)
	French Alps	Melt onset	2 weeks earlier above 3000 m	1980 - 2015	Past	In-situ	Thibert et al. (2013)
	-	Melt intensity	15% stronger above 3000 m	-	-	In-situ / modelling	-
	European Alps	Winter SWE	- 40 % below 1500 m elevation	1971 - 2000 to 2020 - 2049	SRES A1B	Multiple RCM	Steger et al. (2012) Gobiet et al. (2014) Beniston et al. (2018)
	-	-	- 70% below 1500 m elevation	1971 - 2000 to 2070 - 2099	-	-	-
	-	-	- 10% below 1500 m elevation	1971 - 2000 to 2020 - 2049	-	-	-
	-		- 40% below 1500 m elevation	1971 - 2000 to 2070 - 2099	-	-	-
	French Alps	Winter mean snow depth	- 20%	1986 - 2005 to 2030 - 2050	RCP2.6	Multiple RCM	Verfaillie et al. (2018)
	-	-	- 30 %	-	RCP8.5	-	-
	-	-	- 30 %	1986 - 2005 to 2080 - 2100	RCP2.6	-	-
	-	-	- 80 %	-	RCP8.5	-	-
	European Alps	Similar results as above and strenghtening of the asymmetrical seasonal snow decline pattern (stronger trend for reduced snow cover duration in spring than in fall).					Marty et al. (2017a); Terzago et al. (2017); Hanzer et al. (2018)
	Pyrenees, below 1000 m	Snow cover duration	Decrease in majority of stations	1975-2002	Past	In-situ	Pons et al. (2010); Beniston et al. (2018)
	Pyrenees, above 1000 m	-	Decrease in majority of stations		-	-	-
	Pyrenees, Andorra (1645 m elevation)	Number of days with snow depth above 5, 30 and 50 cm	Increase until about 1980 then decrease (not	1935-2015	Past	In-situ	Albalat (2018)

			stastitically significant, high variability)				
Scandinavia	Norway	Snow depth and SWE	Decrease at low elevation	20th century	Past	In-situ	Skaugen et al. (2012); Dyrrdal et al. (2013); Beniston et al. (2018)
	-	-	Increase at higher elevation	20th century	-	-	-
	Northern Scandinavia	Duration and SWE	Further decrease at low elevation, marginal changes at high elevation	1971 -2000 to 2010 - 2100	A1B	GCM downscaled using RCM	Räisänen and Eklund (2012); Beniston et al. (2018)
Caucasus/ Middle East	Lack of long term analysis in Causaus and Middle East mountains, based on observations for past changes and model projections for the future.						
North Asia	Lack of long term analysis in North Asian mountains, based on observations for past changes and model projections for the future.						
Southern Andes	Whole area	Snow covered area	Decrease (but too short period, high variability)	2000-2015	Past	Optical remote sensing	Malmros et al. (2018)
	Whole area	Mean SWE	-13%	1980 - 2010 to 2035 - 2065	RCP4.5	Multiple RCM	López-Moreno et al. (2017)
	-	-	-17%	-	RCP8.5	-	
	-	Duration	7 days	-	RCP4.5	-	
	-	-	10 days	-	RCP8.5	-	
	Limarí river basin, north-central Chile	Peak SWE	-32 % above 5000 m -82% between 2500 m and 3000 m -100% between 2000 and 2500 m	1961 - 1990 to 2071 - 2000	B2	Single GCM/RCM pair	Vicuña et al. (2011)
	-	-	-41 % above 5000 m -96% between 2500 m and 3000 m -100% between 2000 and 2500 m	-	A2	-	-

Low Latitude (including tropical Andes)	Whole area	Compared to mid and high latitude mountain areas seasonal snow cover has limited relevance in the tropical Andes and other tropical areas, except in the immediate vicinity of glaciers. Satellite-based observations are too short to address long-term trends.					Saavedra et al. (2018)
	Whole area	No specific literature on climate projections of snow cover in the Andesor other tropical areas.					Vuille et al. (2018)
High-mountain Asia	Himalaya and Tibetan Plateau	Snow covered area	Significant interannual variability	2000 - 2015	Past	Optical remote sensing	Tahir et al. (2015) Gurung et al. (2017) Li et al. (2018) Bolch et al. (2018)
	Himalaya	SWE	-10.60 kg m^{-2} yr^{-1} for areas above 500 m elevation	1987 - 2009	Past	Passive microwave remote sensing	Smith and Bookhagen (2018); (Wang et al., 2018)
	Hindu-Kush Karakoram	Winter snow depth (December to April)	-7 %	1986 - 2005 to 2031 - 2050	RCP8.5	Analysis of GCM output	Terzago et al. (2014)
	-	-	-28 %	1986 - 2005 to 2081 - 2100	-	-	-
	Himalaya	-	-25 %	1986 - 2005 to 2031 - 2050	-	-	-
		-	-55%	1986 - 2005 to 2081 - 2100	-	-	-
New Zealand and SW Australia	SW Australia	SWE	Reduction, especially in springtime	Mid-20th century - present	Past	In-situ	Fiddes et al. (2015); Di Luca et al. (2018)
	-	Duration	Reduction, especially in springtime	-	-	-	-
	New Zealand	Too limited observation records to ascertain long-term trends					
New Zealand and SW Australia	Australia	SWE	Reduction, especially below 1000 m	1980-1999 to 2030-2049	SRES A1B	Multiple downscaled GCMs	Hendrikx et al. (2013)
	Australia	SWE	-15 %	1990 - 2009 to 2020 - 2040	SRES A2	Multiple downscaled GCMs	Di Luca et al. (2018)
	-	-	-60 %	1990 - 2009 to 2060 - 2080	-	-	-
	New Zealand	SWE	-3% to -44 % at 1000 m	1980 - 1999 to 2030 - 2049	SRES A1B	Multiple downscaled GCMs	Hendrikx and Hreinsson (2012)
	-	-	-8 % to -22 % at 2000 m	-	-	-	-

	-	-	-32% to -79% at 1000 m	1980 - 1999 to 2080 - 2099	-	-	-
	-	-	-6% to -51 % at 2000 m	-	-	-	-
Japan	Japan	Lack of long term trend analysis based on in-situ records.					
Japan	Japan	Winter snow depth	-50 % at low elevation	Base: 1990s Future: time period corresponding to 2°C warming.	+2°C global warming (from SRES A1B)	Multiple downscaled GCMs (time sampling)	Katsuyama et al. (2017)
	-	-	-10 % at high elevation	-	-	-	
	Japan, mountain catchment	SWE	- 36%	1981 - 2000 to 2046 - 2065	SRES A1B	Multiple downscaled GCMs	Bhatti et al. (2016)

2.A.4 Details of Studies on Peak Water

Appendix 2.A, Table 4: Overview of studies providing estimates for peak water in individual glaciers or glacierized basins that are plotted in Figure 2.7. Peak water is the approximate year derived from observations (past) or modelling (future) when annual runoff in glacierized basins reaches a maximum due to glacier shrinkage.

Region	Glacier/basin name	Study type	Peak Water (year)	Glacier area (km^2)	Glacieri-zation (%)	Reference	Remarks
Alaska	Copper River basin	regional	~2070	~13,000	~21	Valentin et al. (2018)	
	Wolverine	single	~2050	17	67	Van Tiel et al. (2018)	No clear peak
Western Canada	Hood	single	~2015	~9	100	Frans et al. (2016)	Runoff from glacier area
	Bridge	single	~2015	73	53	Moyer et al. (2016)	Qualitative statement: At / close to peak water (no future modelling)
	Mica basin	regional	~2000	1,080	52	Jost et al. (2012)	Already past peak water, no actual maximum detected
	Bridge	single	~2000	73	53	Stahl et al. (2008)	Already past peak water, no actual maximum detected
	Hoh	single	1988	18	100	Frans et al. (2018)	Runoff from glacier area
	Stehekin	single	1985	19	100		
	Cascade	single	1984	12	100		
	Hood	single	1995	5	100		
	Thunder	single	2040	32	100		
	Nisqually	single	2053	18	100		
	Several basins in Western Canada	regional	≈2000	150		Fleming and Dahlke (2014)	"Peak Water already over" (qualitative statement); runoff data analysis, no modelling of future glacier runoff
	Western Canada, coastal Alaska	regional	≈2035	26,700	100	Clarke et al. (2015)	Runoff from glacier area; varying with GCM and RCP used (between ~2022 and 2065)
Iceland	S-Vatnajökull, Langjökull, Hofsjökull	single/regional	≈2055	≈5000	100	Björnsson and Pálsson (2008)	

Scandi-navia	Nigardsbreen	single	≈2050	45	70	Van Tiel et al. (2018)	No clear peak
Central Europe	Gries	single	2020	5	49	Farinotti et al. (2012)	
	Silvretta	single	2015	5	5		
	Rhone	single	2042	18	46		
	Gorner	single	2035	51	63		
	Aletsch	single	2050	117	59		
	Trift	single	2045	17	43		
	Zinal	single	2047	11	65	Huss et al. (2008)	
	Moming	single	2039	6	63		
	Weisshorn	single	2035	3	39		
	Morteratsch	single	2020	16	15	Huss et al. (2010)	
	Forno	single	2042	7	34		
	Albigna	single	2020	6	30		
	Plaine Morte	single	2055	8	100	Reynard et al. (2014)	
	Findel	single	2035	16	74	Uhlmann et al. (2013)	
	Findel	single	~2050	16	74	Huss et al. (2014)	Peak water between 2035-2065 depending on climate model input
	Swiss Alps	>100 single g.	1997	<0.05	100	Huss and Fischer (2016)	
	Swiss Alps	>100 single g.	2000	0.05-0.125	100		
	Swiss Alps	>100 single g.	2004	0.125-0.5	100		
High Mountain Asia	Chon Kemin basin	regional	~2025	112	11	Sorg et al. (2014)	
	Largest Rivers of China	regional	~2070	≈30,000		Su et al. (2016)	
	Hailuogou	single	~2050	45	36	Zhang et al. (2015)	No clear peak; declining glacier runoff after 2050, stable overall runoff
	Kakshaal basin	regional	~2020	740	4	Duethmann et al. (2016)	Runoff from glacier area
	Sari-Djaz basin	regional	~2030	2,580	20		
	Naryn basin	regional	~2030	1,160	2	Gan et al. (2015)	Peak water around 2050 for RCP8.5
	Urumqi	single	2020	2	52	Gao et al. (2018)	
	Yangbajing basin	regional	~2025	312	11	Prasch et al. (2013)	Peak water between 2011 and 2040
	Headwaters of Brahmaputra, Ganges, Indus	regional	~2050	~30,000		Lutz et al. (2014)	
		regional	~2050	~30,000			
	All glaciers	regional	~2030	~90,000	100	Kraaijenbrink et al. (2017)	Peak Water around 2050 for RCP8.5
		regional	~2030	~90,000	100		
		regional	~2030	~90,000	100		
	Chhota Shigri	single	2025	16	46	Engelhardt et al. (2017)	No clear maximum runoff
	Hypothetical	single	2055	50	1	Rees and Collins (2006)	Runoff from glacier area
	Hypothetical	single	2064	50	1		
	Langtang	single	2045	120	100	Immerzeel et al. (2013)	
	Baltoro	single	2044	520	100		

	Langtang	single	2055	120	34	Ragettli et al. (2016)	
Low Latitudes	Rio Santa basin	regional	~2005	200	2	Carey et al. (2014)	"Peak water already over" (qualitative statement)
	Zongo	single	2010	3	21	Frans et al. (2015)	
	Cordillera Blanca	regional	~1995	480		Polk et al. (2017)	"Peak water already over" (qualitative statement); no modelling of future glacier change and runoff
	Sub-basins of Rio Santa		~1990	200	2	Baraer et al. (2012)	Runoff data analysis, no modelling of future runoff
South Andes	Juncal	single	2025	34	14	Ragettli et al. (2016)	

2.A.5 Details of Studies on Impacts Attributed to Cryosphere Changes

Appendix 2.A, Table 5: Overview of studies showing regions where biophysical or human sectors have experienced an impact in the past several decades that can be partly attributed to changes in the cryosphere. Only impacts documented in the scientific literature are shown. Confidence levels are given for confidence in detection and confidence in attribution to cryosphere changes. Only confidence in attribution is shown in the associated Figure.2.10 in the main text.

Region	**Location**	**Impact general**	**Specific Impact**	**Detection confidence**	**Climate driver**	**Attribution confidence**	**Reference**
Global	global	Terrestrial ecosystems	Ecosystem change (ex. shrubline, treeline, bamboo shifts)				Wang et al. (2018) Steinbauer et al. (2018) Bjorkman et al. (2018)
	global (tundra)	Terrestrial ecosystems	Plant functional traits	*very high*	warming	*High*	Bjorkman et al. (2018)
	global	Terrestrial ecosystems	Altered phenology	*very high*	multiple drivers	*High*	Post et al. (2018)
Alaska	Alaska	Lahars	Lahars from ice and snow-clad volcanoes (decreasing intensity and size)	*medium*	Warming	*Medium*	
Western Canada and USA	W. USA and W. Canada	Floods	Flood	*high*	Warming	*High*	Musselman et al. (2018)
	W. USA and W. Canada	Snow avalanche	Snow avalanche	*medium*	Warming	*Medium*	McClung (2013) Sinickas et al. (2015) Bellaire et al. (2016)

	Canada	Landslides	Landslides (increasing number)	*high*	Warming	*High*	
	W. USA and W. Canada	Tourism	Ski tourism	*high*	Warming	*High*	Steiger et al. (2017)
Western Canada and USA	British Colombia, Canada	Hydropower		*high*	Warming, Reduced/increased precipitation/ snow, change .	*High/medium*	Jost and Weber (2012) Warren and Lemmen (2014) Lee et al. (2016)
	Nelson-Churchill watershed, Canada	Hydropower		*high*	Warming	*Medium*	Manitoba Hydro (2014)
	Nelson-Churchill watershed, Canada	Hydropower		*high*	Reduced/increased precipitation/ snow, change in snow cover	*Medium*	Manitoba Hydro (2014)
	Colorado	Terrestrial ecosystems	Specialist/ endemic species loss	*high*	Warming	*High*	Panetta et al. (2018)
	Cascades	Culture	Culture	*medium*	Warming	*High*	Duntley (2015)
	Rocky Mountains/ Cascades	Food/ Agriculture	Agriculture	*moderate*	Warming	*Medium*	Frans et al. (2016); McNeeley (2017)
	Pacific Northwest, USA	Hydropower		*high*	Warming	*High*	DOE (2013) Kao et al. (2015) Reclamation (2016) Tarroja et al. (2016)
	Pacific Northwest, USA	Hydropower		*high*	Reduced/ increased precipitation/ snow, change in snow cover	*Medium*	DOE (2013) Kao et al. (2015) Reclamation (2016) Tarroja et al. (2016)
	Northern California, USA	Hydropower		*high*	Warming	*High*	
	Northern California, USA	Hydropower		*high*	Reduced/increased precipitation/ snow, change in snow cover	*Medium*	

	Upper Colorado River, USA	Hydropower		*high*	Warming	*High*	Kopytkovskiy et al. (2015)
USA	Upper Colorado River, USA	Hydropower		*high*	Reduced/increased precipitation/ snow, change in snow cover	*Medium*	Kopytkovskiy et al. (2015)
Canada	Quebec	Hydropower		*high*	Warming	*High*	Warren and Lemmen (2014)
	Quebec	Hydropower		*high*	Reduced/increased precipitation/ snow, change in snow cover	*Medium*	Warren and Lemmen (2014)
South America	Tropical Andes	Tourism	Mountain tourism	*medium*	Warming	*Medium*	Kaenzig et al. (2016)
	Andes	Landslides	Landslides (increasing number)	*medium*	Warming	*High*	
	Andes	Lahars	Lahars from ice and snow-clad volcanoes (decreasing intensity and volume)	*medium*	Warming	*Medium*	
	Ecuador	Riverine biodiversity	Riverine macroinvertebrate biodiversity	*high*	Warming	*Medium*	Milner et al. (2017)
	Central Andes/ tropical mountains	Livelihoods	Livelihoods	*very high*	Warming	*High*	Warner et al. (2012); Yager (2015); Carey et al. (2017)
	Central Andes	Culture	Culture	*high*	Warming	*High*	Allison (2015); Carey et al. (2017)
	Central Andes/ tropical mountains	Food/ Agriculture	Agriculture	*high*	Warming	*High*	McDowell et al. (2014); McDowell et al.)
South America	Central Andes/tropical mountains	Migration	Migration	*high*	Warming	*Medium*	Warner et al. (2012); Kaenzig (2015); Brandt et al. (2016); Alata et al. (2018)

	Cordillera Blanca, Peru	Hydropower			Warming	*High*	Mark et al. (2010); Chevallier et al. (2011); Baraer et al. (2012); Condom et al. (2012)
	La Balsa, Peru	Hydropower		*medium*	Warming	*High*	Vergara et al. (2007)
Iceland	Iceland	Riverine biodiversity	Riverine-macro-invertebrate biodiversity	*high*	Warming	*Medium*	Milner et al. (2017)
	Sandá í Þistilfirði, Iceland	Hydropower		*high*	Warming, Reduced/increased precipitation/snow, change .	*Medium*	Einarsson and Jónsson (2010)
	Austari-Jökulsá, Iceland	Hydropower		*high*	Warming, Reduced/increased precipitation/snow, change .	*Medium*	Einarsson and Jónsson (2010)
Scandinavia	Scandinavia	Floods	Flood	*high*	Warming	*High*	
	Scandinavia	Snow avalanche	Snow avalanche	*medium*	Warming	*Medium*	
	Scandinavia	Tourism	Ski tourism	*high*	Warming	*High*	Steiger et al. (2017)
	Scandinavia	Tourism	Mountain tourism	*high*	Warming	*High*	
Central Europe	European Alps, Pyrenees	Floods	Flood	*high*	Warming	*High*	(Moran-Tejéda et al., 2016) (Freudiger et al., 2014)
Central Europe	European Alps, Pyrenees	Snow avalanche	Snow avalanche	*medium*	Warming	*Medium*	Eckert et al. (2013); Teich et al. (2012); Lavigne et al. (2015); Gadek et al. (2017); Pielmeier et al. (2013); Naaim et al. (2016)
	European Alps, Pyrenees	Tourism	Ski tourism	*high*	Warming	*High*	Steiger et al. (2017)

	European Alps, Pyrenees	Tourism	Mountain tourism	*high*	Warming	*High*	Mourey and Ravanel (2017); Moreau (2010)
	Trentino, Italy	Hydropower		*medium*	Warming/precipitation change/snow cover change	*Medium*	
	Carpathians, Eastern Europe	Hydropower		*medium*	Warming/precipitation change/snow cover change	*High*	Alberton et al. (2017)
	Sava River Basin, SE Europe	Hydropower		*medium*	Warming/precipitation change/snow cover change	*Medium*	
	Austrian Alpine region	Hydropower	cooling water	*high*	Warming/precipitation change/snow cover change	*High*	Wagner et al. (2015)
	HPP Löntsch, Switzerland	Hydropower		*high*	Warming/precipitation change/snow cover change	*High*	Hänggi and Weingartner (2011)
	HPP Oberhasli, Switzerland	Hydropower		*high*	Warming/precipitation change/snow cover change	*Medium*	Stähli et al. (2011)
	Göscheneralp reservoir, Switzerland	Hydropower		*medium*	Warming, Reduced/increased precipitation/snow, change .	*Medium*	Stähli et al. (2011)
	HPP Gougra, Switzerland	Hydropower		*high/ medium*	Warming, Reduced/increased precipitation/snow, change .	*Medium*	Raymond Pralong et al. (2011)
	Prättigau, Switzerland	Hydropower		*high*	Warming, Reduced/increased precipitation/snow, change .	*High*	Hänggi and Weingartner (2011)
Central Europe	Upper Rhone, Switzerland	Hydropower		*high*	Warming	*Medium*	Fatichi et al. (2013); Gaudard et al. (2013); Clarvis et al. (2014); Gaudard et al. (2014)
	Alps	Culture	Culture	*medium*	Warming	*High*	Brugger et al. (2013)

	Swiss Alps	Food/ Agriculture	Food	*high*	Warming	*Medium*	Beniston and Stoffel (2014)
	Italian Alps	Riverine biodiversity	Riverine macroinvertebrate biodiversity	*high*	Warming	*Medium*	Milner et al. (2017)
	French Pyrneees	Riverine biodiversity	Riverine macroinvertebrate biodiversity	*high*	Warming	*Medium*	Khamis et al. (2015)
	Austrian Alpas	Riverine biodiversity	Diatom biodiversity	*high*	Warming	*Medium*	Fell et al. (2017)
	Austrian Alps	Riverine biodiversity	Microbial biodiversity	*high*	Warming	*Medium*	Wilhelm et al. (2013)
	Alps	Landslides	Landslides (increasing number)	*high*	Warming	*High*	
Central Europe	European Alps	Health	Reduced water quality	*high*	Warming	*Medium*	Thies et al. (2007); Ilyashuk et al. (2018)
	continent wide	Terrestrial ecosystems	increased plant species richness	*very high*	warming	*high*	Steinbauer et al. (2018)
North Asia	Kamchatka	Lahars	Lahars from ice and snow-clad volcanoes (decreasing violence and volume)	*medium*	Warming	*Medium*	
High Mountain Asia	continent wide	Terrestrial ecosystems	increased plant species richness	*very high*	warming	*high*	Steinbauer et al. (2018)
	Himalaya and Tibetan Plateau	Floods	Flood	*medium*	Warming	*Medium*	
	Himalaya and Tibetan Plateau	Snow avalanche	Snow avalanche	*medium*	Warming	*Medium*	Ballesteros-Cánovas et al. (2018)

	High Mountain Asia	Landslides	Landslides (increasing number)	*medium*	Warming	*High*	
	Himalayas/ Karakoram/ High Mountain Asia	Livelihoods	Livelihoods	*high*	Warming	*Medium*	Milan et al. (2015); van der Geest and Schindler (2016)
	Himalayas/ Hengduan Mountains/ High Mountain Asia	Culture	Culture	*high*	Warming	*High*	Allison (2015)
	Himalayas/ Karakoram/ Hindu Kush/Tien Shan	Food/Agriculture	Agriculture	*high*	Warming	*Medium*	McDowell et al. (2014); McDowell et al.)
	Himalayas/ Karakoram/ Tien Shan/High Mountain Asia	Migration	Migration	*high*	Warming	*Medium*	Shaoliang et al. (2012); Milan et al. (2015); Gautam (2017); Hill et al. (2017)
	Himalaya and Tibetan Plateau	Tourism	Mountain tourism	*medium*	Warming	*Medium*	Dangi et al. (2018)
	Ganges Basin	Hydropower		*medium*	Warming	*High*	Lutz et al. (2015)
	Ganges Basin	Hydropower		*medium*	Reduced/ increased precipitation/ snow, change in snow cover	*Low*	Lutz et al. (2015)
High Mountain Asia	Indus Basin	Hydropower		*medium*	Warming	*High*	Lutz et al. (2015) Lutz et al. (2016) Shrestha et al. (2015)
	Indus Basin	Hydropower		*medium*	Reduced/ increased precipitation/ snow, change in snow cover	*Low*	Lutz et al. (2015) Lutz et al. (2016) Shrestha et al. (2015)
	Langtang, Nepal	Hydropower		*medium*	Warming, Reduced/increased precipitation/ snow, change .	*Medium*	Lutz et al. (2015)
	Tamakoshi Basin, Nepal	Hydropower		*medium*	Warming, Reduced/increased precipitation/ snow, change .	*Medium*	

	Kulekhani Basin, Nepal	Hydropower		*low*	Warming, Reduced/increased precipitation/ snow, change .	*Medium*	Shrestha et al. (2014)
	Tajikistan	Hydropower		*medium*	Warming, Reduced/increased precipitation/ snow, change .	*Low*	
Japan	Japan	Tourism	Ski tourism	*high*	Warming	*High*	Steiger et al. (2017)
New Zealand	New Zealand	Landslides	Landslides (increasing number)	*medium*	Warming	*Medium*	
	New Zealand, Australia	Tourism	Ski tourism	*high*	Warming	*High*	Steiger et al. (2017)

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